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HYDROLOGY AND SIMULATION OF GROUND-WATER FLOW IN JUAB VALLEY, JUAB COUNTY, UTAH

By Susan A. Thiros, Bernard J. Stolp, Heidi K. Hadley, and Judy I. Steiger
U.S. Geological Survey

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CONTENTS

Abstract 1
Introduction 1
Purpose and scope 4
Previous studies 4
Acknowledgments 4
Description of the study area 4
Geology 5
Climate 8
Hydrology 11
Surface water 11
Streams 11
Distribution systems 16
Reservoirs 18
Ground water 19
Occurrence 19
Recharge 20
Seepage from streams 20
Seepage from unconsumed irrigation water and distribution systems 20
Infiltration of precipitation 24
Subsurface inflow 24
Flow 25
Discharge 25
Wells 25
Springs and seeps 28
Evapotranspiration 30
Subsurface outflow 30
Hydrologic properties 30
Water-level fluctuations 34
Long-term fluctuations 34
Seasonal fluctuations 34
Chemical characteristics of water 41
Chemical composition 41
Isotopes 48
Geochemical analyses 50
Simulation of the ground-water system in the unconsolidated basin-fill deposits of Juab Valley 53
Model construction 54
Discretization 54
Boundary conditions and data requirements 56
Recharge 57
Discharge 65
Hydrologic properties 72
Model calibration 73
Calibration parameters 73
Steady-state calibration 82
Transient-state calibration 85
Model sensitivity 93
Model limitations 96

CONTENTS—Continued

Need for future study 97
Summary 98
References cited 99

PLATE

[Plate is in pocket]

Plate 1. Map showing location of selected hydrologic-data sites and rock-sample site, Juab Valley, Utah, 1935-94

FIGURES

1. Map showing location of Juab Valley study area, Utah 2
2. Diagram showing numbering system used in Utah for hydrologic-data sites 3
3. Map showing generalized geology of the Juab Valley study area, Utah 6
4. Generalized cross section through the southern part of Juab Valley, Utah 9
5. Graphs showing annual precipitation and cumulative departure from the average annual precipitation, 1931-93, at Nephi, Utah 10
6. Map showing discharge-measurement sites on Salt Creek, on selected irrigation canals and ditches, and in the West Creek area, Juab Valley, Utah 14
7. Graphs showing estimated average seepage gain or loss for reaches of Salt Creek and selected irrigation canals and ditches, Juab Valley, Utah 17
8. Map showing approximate potentiometric surface in unconsolidated basin-fill deposits in Juab Valley, Utah, March and April 1993 26
9. Map showing areal distribution of transmissivity estimated from specific-capacity values, slug tests, and aquifer tests for unconsolidated basin-fill deposits in Juab Valley, Utah 32
10. Hydrographs showing March or April water levels measured during 1942-94 in four wells in Juab Valley, Utah, and cumulative departure of the annual maximum water content of snow from the 1942-94 average at the Payson Ranger Station snow course 35
11. Hydrographs showing long-term water-level fluctuations during 1935-94 in eight wells in Juab Valley, Utah 36
12. Map showing approximate change in water levels in unconsolidated basin-fill deposits from March-April 1965 to March-April 1993 in Juab Valley, Utah 38
13. Hydrographs showing seasonal water-level fluctuations during 1993-94 in four wells in Juab Valley, Utah 40
14. Hydrographs showing seasonal water-level fluctuations during 1937-38, 1965-66, and (or) 1993-94 in four wells in Juab Valley, Utah 42
15. Map showing dissolved-solids concentration in water from the Juab Valley study area, Utah, 1951-94 44
16. Graphs showing relation between specific conductance of water and streamflow at gaging stations on Salt Creek, Chicken Creek, and Currant Creek, Juab Valley, Utah 47
17. Graph showing relation between $\delta^{18}O$ and δD values for water sampled from selected hydrologic-data sites in Juab Valley, Utah 49
18. Graph showing relation between log calcium and log sulfate activity for water sampled along the hypothesized Chicken Creek drainage flow path in Juab Valley, Utah 52
19. Map showing model grid and location of active cells in the ground-water flow model of Juab Valley, Utah 55
20. Generalized geology and the four layers used in the ground-water flow model of Juab Valley, Utah 56

FIGURES—Continued

21. Graph showing depth from the estimated top of the saturated unconsolidated basin-fill deposits, the amount of 1963-93 average annual ground-water discharge to pumped wells in each depth interval, and corresponding model layers in the ground-water flow model of Juab Valley, Utah 57

22-29. Maps showing:

22. Location of specified-flux boundary cells that simulate recharge from nonirrigation-season streamflow, unconsumed irrigation water and distribution-system losses, and irrigation-season streamflow not included in the unconsumed irrigation water and distribution-system losses component in the ground-water flow model of Juab Valley, Utah 59
23. Location of specified-flux boundary cells that simulate recharge from subsurface inflow and seepage from ephemeral streams and discharge by subsurface outflow in the ground-water flow model of Juab Valley, Utah 66
24. Location of specified-flux boundary cells that simulate discharge to pumped wells in the ground-water flow model of Juab Valley, Utah 67
25. Location of head-dependent flux boundary cells that simulate discharge to flowing wells, springs, seeps, streams, and reservoirs in the ground-water flow model of Juab Valley, Utah ... 69
26. Location of head-dependent flux boundary cells that simulate discharge by evapotranspiration in the ground-water flow model of Juab Valley, Utah 71
27. Final distribution of horizontal hydraulic conductivity for layers 1 through 3 of the ground-water flow model of Juab Valley, Utah 76
28. Final distribution of transmissivity for layer 4 of the ground-water flow model of Juab Valley, Utah 78
29. Final distribution of vertical hydraulic conductivity for layers 1 through 4 of the ground-water flow model of Juab Valley, Utah 81

30. Hydrographs showing water level in selected wells in Juab Valley, Utah 83
31. Map showing potentiometric surface for model layer 3 contoured from water levels computed from the steady-state simulation for 1949 and the difference between model-computed and measured water levels in observation wells, Juab Valley, Utah, March 1950 84
32. Graph showing estimated annual ground-water discharge to pumped wells, length of stress periods, and average annual pumpage for each stress period of the multi-year transient-state simulation, Juab Valley, Utah 86
33. Hydrographs showing model-computed and measured water-level change in selected wells in Juab Valley, Utah, for the multi-year transient-state simulation 88
34. Map showing potentiometric surface for model layer 3 contoured from water levels computed from the transient-state simulation for 1964 and the difference between model-computed and measured water levels in observation wells, Juab Valley, Utah, March 1965 90
35. Map showing potentiometric surface for model layer 3 contoured from water levels computed from the transient-state simulation for 1992 and the difference between model-computed and measured water levels in observation wells, Juab Valley, Utah, March 1993 91
36. Map showing computed water-level difference between model layers 1 and 4 for the 1949 steady-state simulation and the 1992 transient-state simulation in the ground-water flow model of Juab Valley, Utah 92
37. Hydrographs showing model-computed and measured water-level change in selected wells in Juab Valley, Utah, for the seasonal transient-state simulation 94

TABLES

1. Estimated average annual streamflow for drainages on the east side of Juab Valley, Utah 13
2. Estimated ground-water budget for unconsolidated basin-fill deposits under average conditions, Juab Valley, Utah 21
3. Estimated recharge from unconsumed irrigation water and distribution-system losses to unconsolidated basin-fill deposits, Juab Valley, Utah 23
4. Specified ground-water recharge and discharge and model-computed discharge for the 1949 steady-state simulation, Juab Valley, Utah 60
5. Specified ground-water recharge and discharge and model-computed discharge and change in storage for the multi-year transient-state simulation, Juab Valley, Utah 61
6. Specified ground-water recharge and discharge and model-computed discharge and change in storage for the seasonal transient-state simulation, Juab Valley, Utah 63

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

Multiply	By	To obtain
acre	0.4047	square hectometer
	4,047	square meter
acre-foot (acre-ft)	0.001233	cubic hectometer
	1,233	cubic meter
acre-foot per year (acre-ft/yr)	0.00003907	cubic meter per second
cubic foot per day (ft^3/d)	0.02832	cubic meter per day
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
foot (ft)	0.3048	meter
foot per year (ft/yr)	0.3048	meter per year
foot per day (ft/d)	0.3048	meter per day
foot squared per day[1] (ft^2/d)	0.0929	meter squared per day
gallon per minute (gal/min)	0.06308	liter per second
inch (in.)	25.4	millimeter
	0.0254	meter
inch per year (in/yr)	0.0254	meter per year
mile (mi)	1.609	kilometer
square mile (mi^2)	2.59	square kilometer

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

$$°F = 1.8\ (°C) + 32.$$

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

Chemical concentration and water temperature are reported only in International System (SI) units. Chemical concentration in water is reported in milligrams per liter (mg/L), micrograms per liter (µg/L), moles per liter (mol/L), or millimoles per liter (mmol/L). These units express the solute weight per unit volume (liter) of water. A liter of water is assumed to weigh 1 kilogram, except for brines or water at high temperatures because of significant changes in the density of the water. For concentrations less than 7,000 milligrams per liter, the numerical value is about the same as for concentrations in parts per million (ppm). One thousand micrograms per liter is equivalent to 1 milligram per liter, one thousand millimoles per liter is equivalent to 1 mole per liter, and one thousand parts per billion (ppb) is equivalent to 1 part per million. A mole of a substance is its atomic or formula weight in grams. Concentration in moles per liter can be determined from milligrams per liter by dividing by the atomic or formula weight of the constituent, in milligrams. Specific conductance is reported in microsiemens per centimeter at 25 degrees Celsius (µS/cm). Stable isotope concentration is reported as per mil, which is equivalent to parts per thousand. Tritium concentration in water is reported as tritium units (TU). The ratio of 1 atom of tritium to 10^{18} atoms of hydrogen is equal to 1 TU or 3.2 picocuries per liter.

[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

Plans to import water to Juab Valley, Utah, primarily for irrigation, are part of the Central Utah Project. A better understanding of the hydrology of the valley is needed to help manage the water resources and to develop conjunctive-use plans.

The saturated unconsolidated basin-fill deposits form the ground-water system in Juab Valley. Recharge is by seepage from streams, unconsumed irrigation water, and distribution systems; infiltration of precipitation; and subsurface inflow from consolidated rocks that surround the valley. Discharge is by wells, springs, seeps, evapotranspiration, and subsurface outflow to consolidated rocks. Ground-water pumpage is used to supplement surface water for irrigation in most of the valley and has altered the direction of ground-water flow from that of pre-ground-water development time in areas near and in Nephi and Levan.

Greater-than-average precipitation during 1980-87 corresponds with a rise in water levels measured in most wells in the valley and the highest water level measured in some wells. Less-than-average precipitation during 1988-91 corresponds with a decline in water levels measured during 1988-93 in most wells. Geochemical analyses indicate that the sources of dissolved ions in water sampled from the southern part of the valley are the Arapien Shale, evaporite deposits that occur in the unconsolidated basin-fill deposits, and possibly residual sea water that has undergone evaporation in unconsolidated basin-fill deposits in selected areas. Water discharging from a spring at Burriston Ponds is a mixture of about 70 percent ground water from a hypothesized flow path that extends downgradient from where Salt Creek enters Juab Valley and 30 percent from a hypothesized flow path from the base of the southern Wasatch Range.

The ground-water system of Juab Valley was simulated by using the U.S. Geological Survey modular, three-dimensional, finite-difference, ground-water flow model. The numerical model was calibrated to simulate the steady-state conditions of 1949, multi-year transient-state conditions during 1949-92, and seasonal transient-state conditions during 1992-94. Calibration parameters were adjusted until model-computed water levels reasonably matched measured water levels. Parameters important to the calibration process include horizontal hydraulic conductivity, transmissivity, and the spatial distribution and amount of recharge from subsurface inflow and seepage from ephemeral streams to the east side of Juab Valley.

INTRODUCTION

The hydrology of Juab Valley in central Utah (fig. 1) was studied from 1992 through 1994 by the U.S. Geological Survey in cooperation with the Central Utah Water Conservancy District and the East Juab Water Conservancy District. The saturated basin-fill deposits form the principal ground-water system in Juab Valley. Hydrologic data were collected from numerous sites and the system for numbering these sites is explained in figure 2. As part of the Central Utah Project, a plan approved by Congress to transport water from the Colorado River drainage basin in Utah to areas generally along the west side of the Wasatch Range, irrigation water will be imported to Juab Valley. A better understanding of the hydrology of Juab Valley is needed by the cooperators to help manage the water resources and to develop plans for the optimal use of the imported surface water and ground and surface water in the area (conjunctive use).

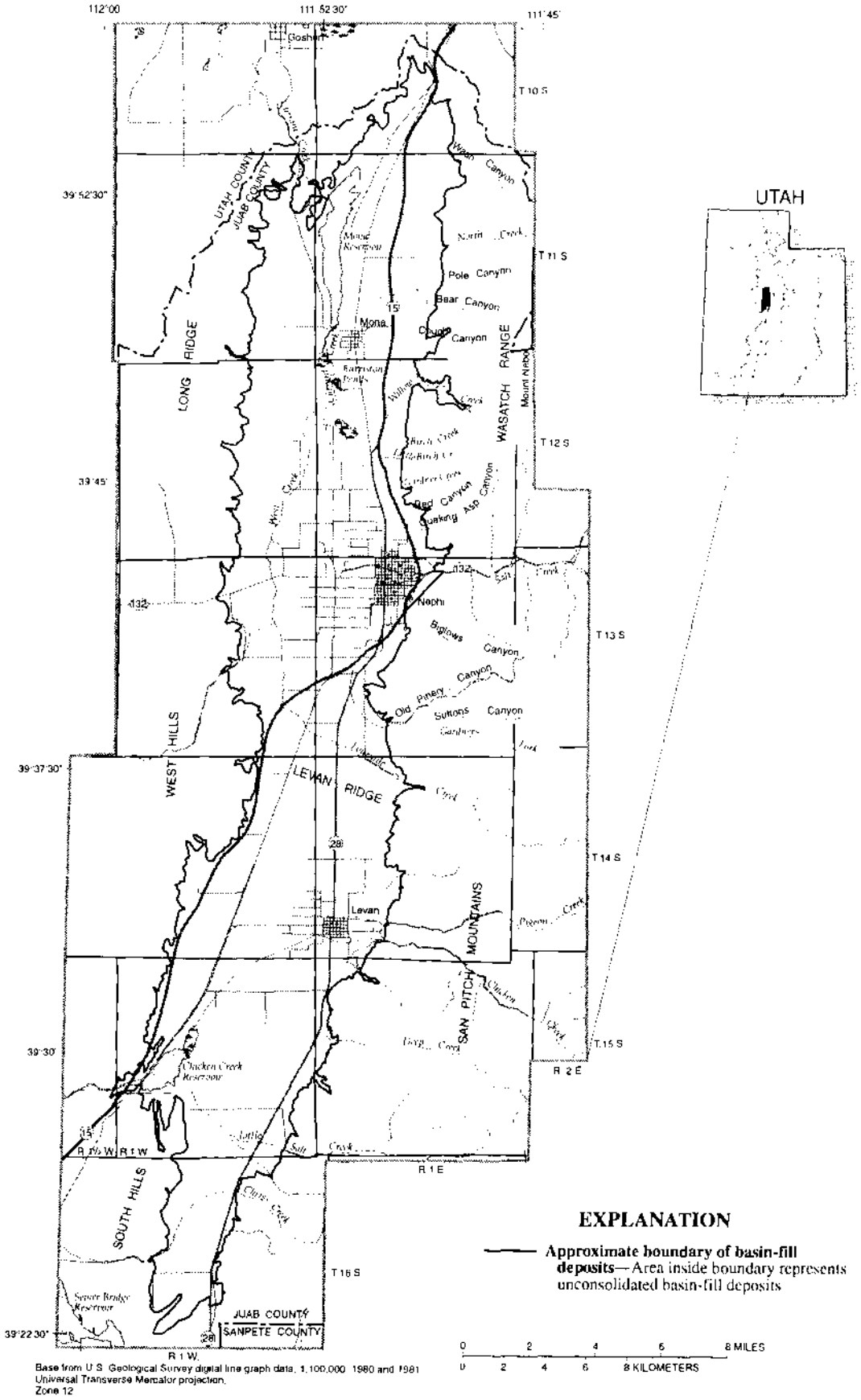
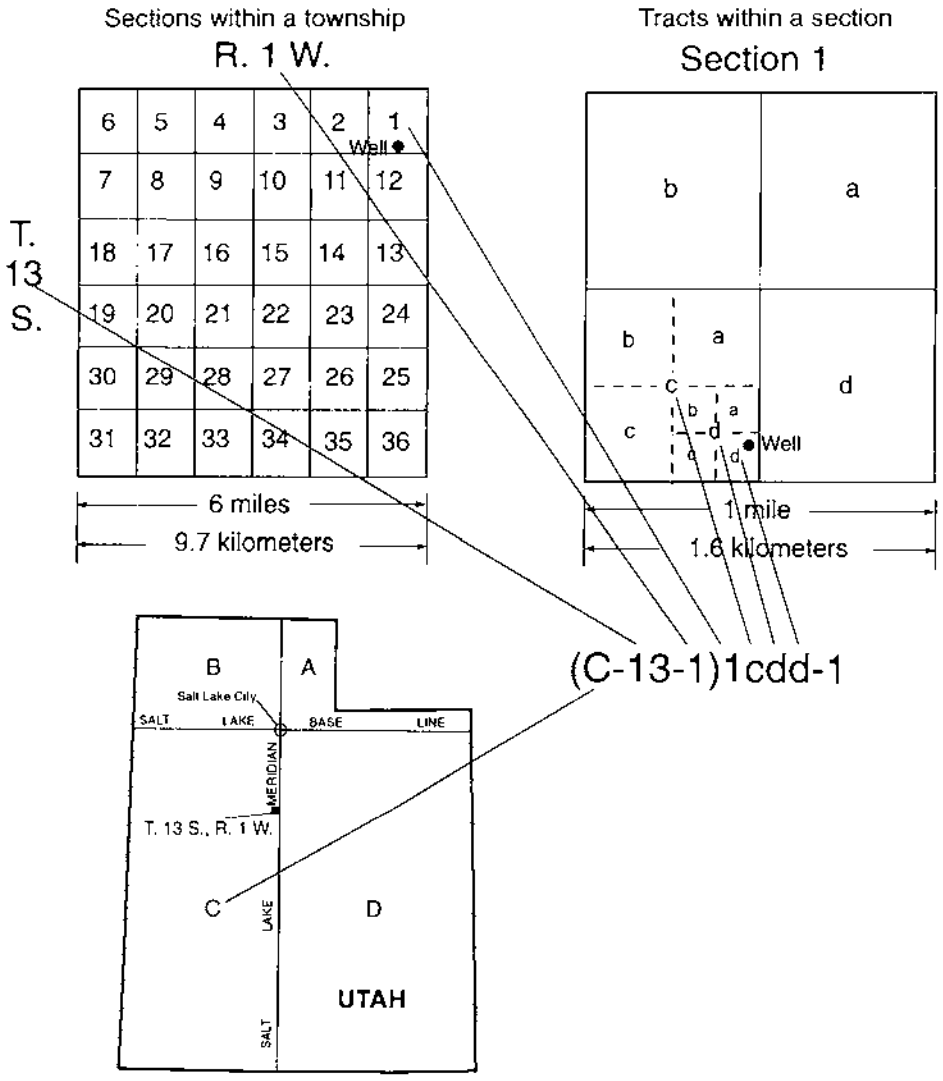


Figure 1. Location of Juab Valley study area, Utah.

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 upper- case 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 acre tracts 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. A number having all three quarter designations but without the letter S and serial number indicates a miscellaneous data site other than a well or spring, such as outflow from a group of springs. Thus, (C-13-1)1cdd-1 designates the first well constructed or visited in the southeast 1/4 of the southeast 1/4 of the southwest 1/4 in section 1, T. 13 S., R. 1 W.



[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 Utah for hydrologic-data sites.

The objectives of the study are to:

1. Assess current ground-water conditions and document changes since previous studies made during 1962-66;
2. Better define and quantify hydrologic components of the ground-water flow system in the valley by estimating recharge, flow, discharge, hydrologic properties of the unconsolidated basin-fill deposits, and chemical characteristics of water;
3. Evaluate current and potential sources of dissolved solids to ground water in the southern part of the valley; and
4. Test and confirm the conceptual ground-water flow model and provide a tool to estimate the effects of changes in ground-water discharge to pumped wells and recharge on water levels and natural ground-water discharge.

Purpose and Scope

This report describes the hydrology of Juab Valley and presents a tool, in the form of a numerical model that simulates ground-water flow in the valley, to estimate the effects of changes in water availability and water-management practices on the ground-water flow system. Reported aspects of the hydrology include descriptions of surface- and ground-water budget components, ground-water flow, hydrologic properties of the ground-water system, and chemical characteristics of ground water in the valley.

Previous Studies

The ground-water hydrology of Juab Valley was first discussed by Meinzer (1911, p. 67-74) as part of a reconnaissance of ground water in Juab, Millard, and Iron Counties, Utah. Ground-water resources in the northern part of Juab Valley were studied by Bjorklund (1967) and in the southern part by Bjorklund and Robinson (1968).

Water levels in selected wells have been periodically measured since 1935 by the U.S. Geological Survey. Streamflow-gaging stations have been in operation on three streams for various intervals from 1925 to 1995. Water-level and streamflow data are available in the files of the U.S. Geological Survey Utah District Office in Salt Lake City, Utah. Hydrologic data collected as part of this study were compiled by Steiger (1995).

Acknowledgments

The assistance and cooperation of local land owners, irrigation-company representatives, and U.S. Department of Agriculture Natural Resources Conservation Service (formerly called the Soil Conservation Service) personnel greatly aided in the collection of data for this study. Information, time, and access to hydrologic-data sites supplied by these individuals are appreciated.

Description of the Study Area

Juab Valley is a north-trending, long, narrow valley in Juab County in the central part of Utah (fig. 1). The valley consists of about 172 mi^2 of unconsolidated basin-fill deposits bounded by mountain areas composed of consolidated rocks. Levan Ridge is an east-west trending topographic divide with an altitude of about 5,220 ft that separates the valley into northern and southern parts.

The northern part of Juab Valley is bounded by the southern Wasatch Range and San Pitch Mountains on the east and the West Hills and Long Ridge on the west. It is about 22 mi long and 4 to 6 mi wide. Altitudes range from 4,860 ft at Mona Reservoir to more than 5,600 ft on some alluvial fans. Mount Nebo, at the southern end of the Wasatch Range, reaches 11,877 ft above sea level or about 7,000 ft above the valley floor. Relatively small, steep, and rocky canyons drain water from the southern Wasatch Range into Juab Valley. Salt Creek, east of Nephi, separates the southern Wasatch Range from the San Pitch Mountains. A pass separates the West Hills from Long Ridge west of Nephi.

The southern part of Juab Valley is bounded by the San Pitch Mountains on the east and the West and South Hills on the west. This part of the valley is about 16 mi long and 2 to 6 mi wide. Altitudes range from 5,050 ft at the Chicken Creek Reservoir spillway to about 5,600 ft along the boundary between consolidated rocks and unconsolidated basin-fill deposits at the mountain front. Surface water from the San Pitch Mountains generally drains toward Chicken Creek Reservoir, except at the south end of the valley, where Chriss Creek cuts across the South Hills.

Geology

Consolidated rocks in the study area range from Precambrian to Tertiary age and crop out in the mountains that surround Juab Valley. In general, the west side of Juab Valley is bounded by volcaniclastic and sedimentary rocks of Tertiary age (Oligocene to Paleocene) and the east side of the valley is bounded by rocks of Precambrian through Mesozoic age (fig. 3). Unconsolidated deposits of Quaternary age generally occur in the valley and along mountain streams.

The southern Wasatch Range is composed mainly of interbedded limestone, sandstone, and quartzite of Late Mississippian to Permian age. This assemblage was thrust eastward over rocks of Mesozoic age during the Late Cretaceous Sevier orogeny. Rocks of the allochthon are overturned at the southern end of the range and dip northwestward near Nephi (Witkind and Weiss, 1991). The strata is vertical east of Mona.

Rocks of Paleozoic age crop out at the northern and southern ends of Long Ridge and at the northern end of the West Hills. Witkind and Weiss (1991) infer from these outcrops that a sheet of thrusted rocks of Paleozoic age unconformably underlies Long Ridge and the West Hills. These rocks likely are fractured.

The Arapien Shale of Middle Jurassic age crops out at the southern end of the Wasatch Range near Nephi and in the San Pitch Mountains (fig. 3). The Arapien Shale is not exposed on the west side of the study area but is underneath a normal stratigraphic section of rocks of Tertiary through late Mesozoic age in the northern part of the West Hills (Zoback, 1992, p. E4). The Arapien Shale is a marine evaporite composed mainly of interbedded calcareous mudstone, gypsiferous siltstone, and micrite with much halite, gypsum, and anhydrite minerals (Biek, 1991). Lenses of rock gypsum and selenite have been mined primarily near where Salt Creek and Chicken Creek enter Juab Valley. The Arapien Shale in the study area is deformed because it typically was the decollement horizon for Sevier orogeny thrust faults. This deformation has caused a complex interrelation with the underlying marine Twin Creek Limestone.

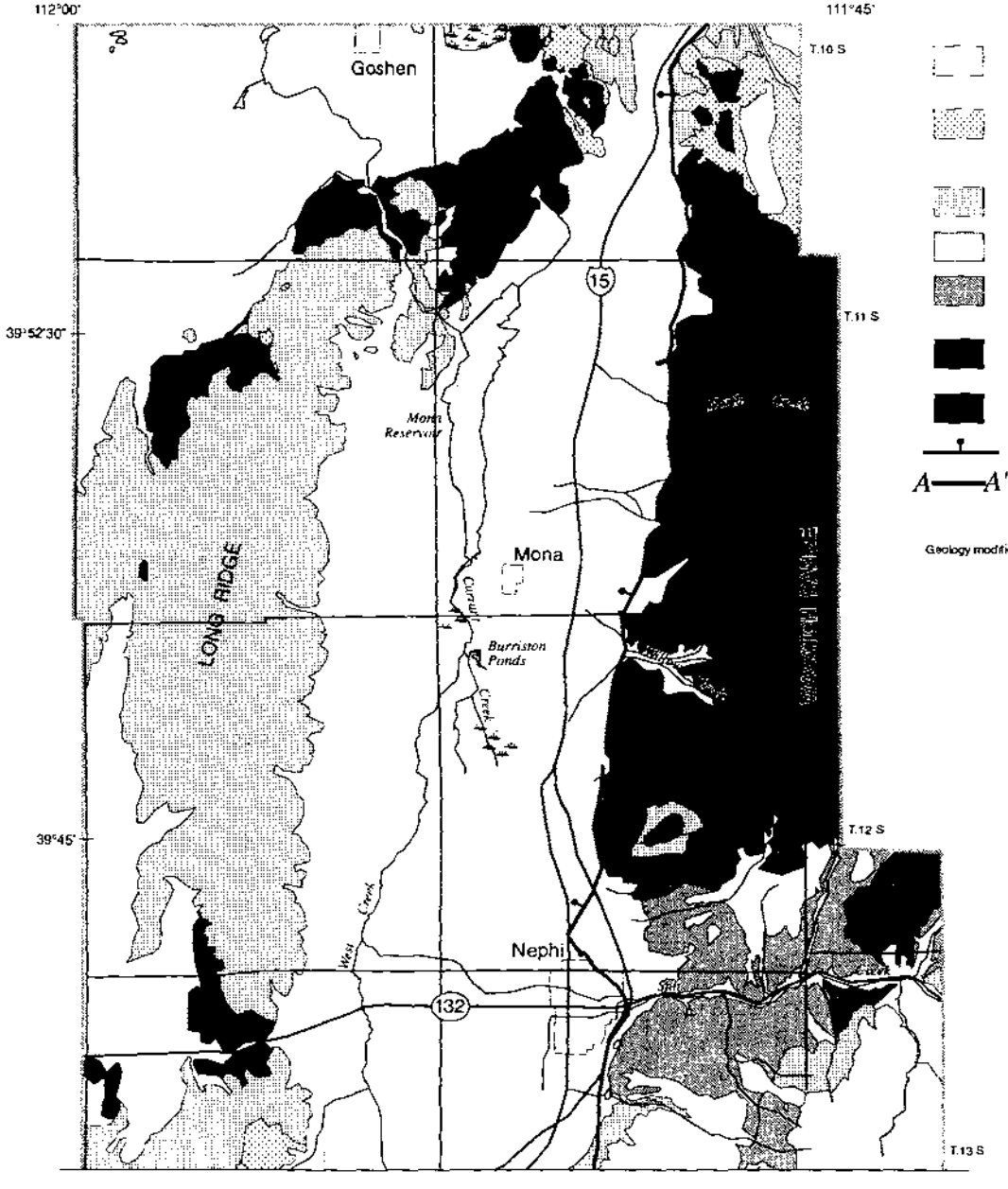
The Twist Gulch Formation of Middle Jurassic age and the Indianola Group of Cretaceous age are exposed in the San Pitch Mountains east of Juab Valley. The Twist Gulch Formation is composed mainly of reddish brown, thin to medium-bedded, fine-grained, marine sandstone, siltstone, and shale. The Indianola Group is mainly a reddish brown and gray, thick-bedded fluvial conglomerate. Indianola Group strata generally dip to the east.

Sedimentary rocks of Late Cretaceous through Tertiary age mapped in the study area are the Price River and North Horn Formations of fluvial origin and the Flagstaff Limestone and Colton and Green River Formations, primarily of lacustrine origin. The sequence of strata from the Indianola Group through the Colton Formation is present at the higher altitudes of the San Pitch Mountains.

The volcaniclastic rocks in the study area are Oligocene to Eocene in age (Tertiary Period) and consist of ashflow and welded tuffs, lava flows, and stream-deposited conglomerate and sandstone. Volcaniclastic rocks are exposed on Long Ridge, the northern part of the West Hills, in the Salt Creek drainage, along the base of the San Pitch Mountains, and at the northern end of the study area (fig. 3). Volcaniclastic and (or) sedimentary rocks of Tertiary age may underlie the unconsolidated to semiconsolidated basin-fill deposits in Juab Valley (Witkind and Weiss, 1991).

Juab Valley is part of the Basin and Range Physiographic Province (Fenneman, 1931), which has undergone intraplate extension since late Cenozoic time. Movement along the Wasatch Fault, a high-angle normal fault, has occurred since Quaternary time and has downfaulted Juab Valley relative to the southern Wasatch Range and the San Pitch Mountains (fig. 3). Geologic interpretation of seismic data indicates that the valley is bounded on the west side by smaller normal faults, which intersect the Arapien Shale at depth (Zoback, 1992, p. E11). Because the inferred fault zone on the west side of the valley is buried by alluvium along much of its length, surface evidence is limited.

Unconsolidated to semiconsolidated material of Quaternary and Tertiary age in the study area includes basin-fill, landslide, colluvial, and stream-channel deposits. The basin-fill deposits in Juab Valley are unconsolidated at land surface and become semiconsolidated with increasing depth. The thickness of these deposits was estimated to be as much as 5,250 ft thick (Zoback, 1992, p. E12) on the basis of interpretation of gravity data and a seismic-reflection profile across the valley near Nephi. A prominent and continuous reflector exists at a depth of about 1,200 ft below land surface at the west end of the profile (about 0.9 mi east of West Creek) and about 2,300 ft below land surface at the east end of the profile (near the west side of Nephi) (Zoback, 1992, p. E11). The layer is bounded by normal faults and may represent a change in the physical properties



EXPLANATION

Unconsolidated and semiconsolidated sediment—Quaternary and Tertiary

Sedimentary rocks—Upper Cretaceous and Tertiary, includes the Price River Formation, North Horn Formation, Flagstaff Limestone, Colton Formation, and Green River Formation

Volcanic and intrusive igneous rocks—Tertiary

Indianola Group—Upper Cretaceous

Twist Gulch Formation, Arapien Shale, and Twin Creek Limestone— Middle Jurassic

Other Mesozoic rocks

Sedimentary rocks—Precambrian and Paleozoic

Fault—Bar and ball on downfaulted side

A——*A'* **Line of generalized cross section shown in figure 4**

Geology modified from Witkind, Weiss, and Brown (1987) and Witkind and Weiss (1991)

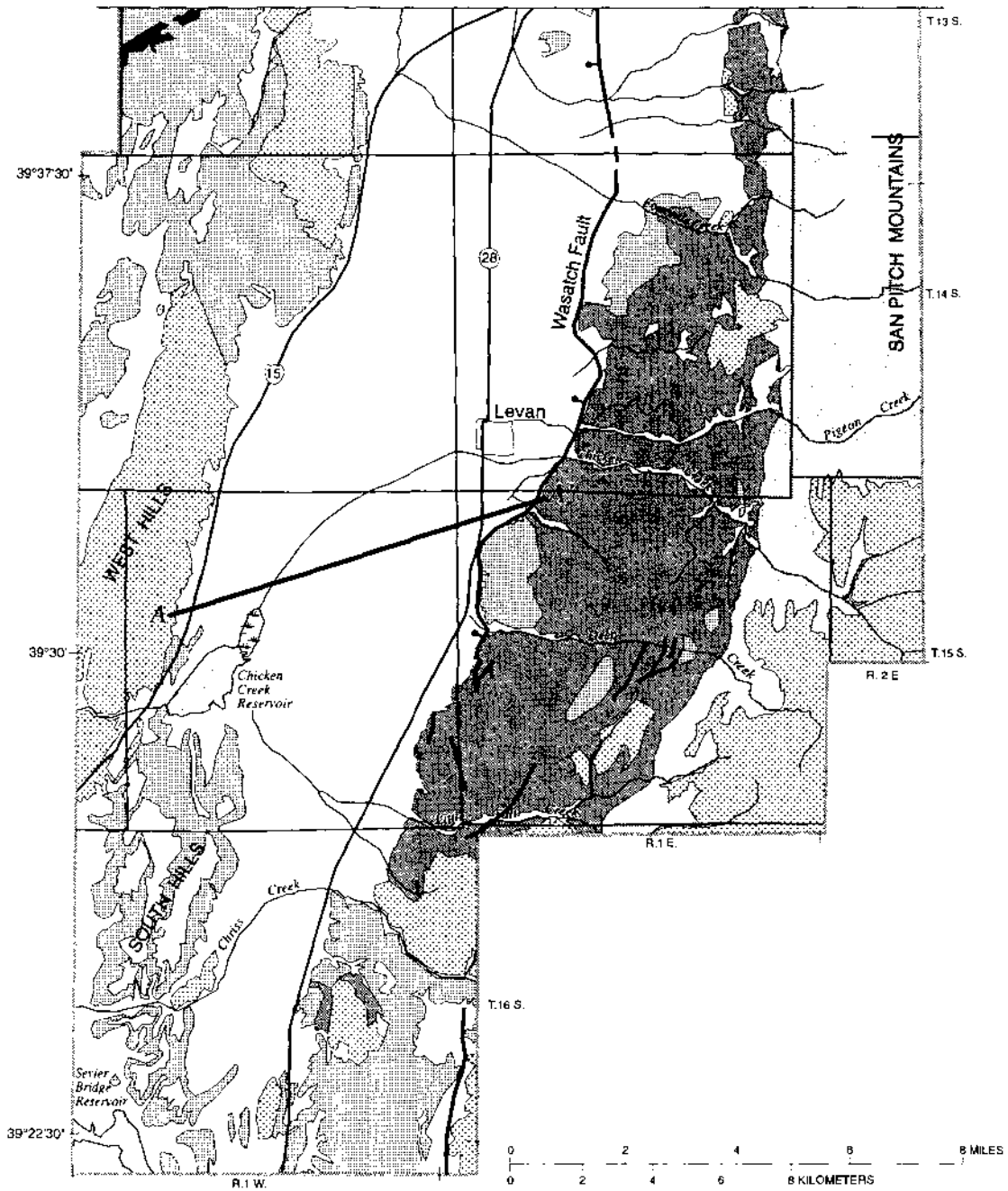


Figure 3. Generalized geology of the Juab Valley study area, Utah.

of the basin-fill deposits, such as the degree of consolidation, composition, or density of the material.

The unconsolidated basin-fill deposits that form the ground-water system discussed in this report consist only of the uppermost part of the basin-fill deposits in Juab Valley. The thickness of the unconsolidated basin-fill deposits in the valley is not known but may be indicated by the continuous reflector across the valley near Nephi. Bouguer gravity data indicate that the unconsolidated and semiconsolidated basin-fill deposits are thickest in the area west of Nephi and are less thick in the southern part of Juab Valley (Zoback, 1992, p. E5). The unconsolidated deposits include alluvial-fan, fluvial, and lacustrine deposits, primarily of Quaternary age. The fill generally is coarser grained near the mountain fronts and along stream channels and finer grained near the lower-altitude areas of the valley. Unsorted fine and coarse-grained material are present in debris-flow deposits at the mountain fronts. The variable grain size and interbedded nature of the unconsolidated basin-fill deposits in the southern part of Juab Valley are shown in figure 4. The potentiometric surface shown on the figure was determined from water levels listed in the drillers' logs and represents the water-level altitude of the confined part of the ground-water system in that area. Water-level altitudes near the eastern mountain front probably represent unconfined conditions in the ground-water system.

Climate

Average annual mean air temperature from 1961 to 1990 was 50.5 °F (10.3 °C) at Nephi and 48.9 °F (9.4 °C) at Levan (Ashcroft and others, 1992, p. 56 and 66). Winter temperatures in Juab Valley can fall below 20 °F (-7 °C) and summer temperatures can exceed 90 °F (32 °C). The average freeze-free season lasts 138 days in Nephi on the basis of 46 years of record and 129 days in Levan on the basis of 96 years of record (Ashcroft and others, 1992, p. 99-100).

Average annual precipitation from 1961 to 1990 was 14.5 in. at Nephi and 15.2 in. at Levan (Ashcroft and others, 1992, p. 56 and 66). Average annual precipitation from 1931 to 1993 was 14.0 in. at Nephi and 14.2 in. at Levan. Annual precipitation and cumulative departure from the average annual precipitation during this period at Nephi are shown in figure 5. Cumulative departure from average annual precipitation is calculated by adding the departure or difference from average precipitation for the current year to the sum of departure values for all previous years in the period of record. It can be used to identify periods of greater- and less-than-average precipitation.

Average annual precipitation at Nephi was exceeded during 1943-47, 1963-65, and 1980-87. Annual precipitation was less than average during 1931-35, 1948-50, 1958-62, 1974-77, and 1988-91. Annual precipitation at Levan generally followed the pattern noted at Nephi. Most precipitation falls in March and April with June and July generally being the driest months.

Snow depths are available for the Payson Ranger Station snow course, about 11 mi northeast of Mount Nebo in the southern Wasatch Range, and the Ree's Flat snow course, about 8 mi southeast of Levan in the San Pitch Mountains. The annual accumulation of snow in the mountains surrounding Juab Valley is the primary source of surface water to the valley. The Payson Ranger Station snow course, at an altitude of 8,050 ft, has records from 1942 to the present (1995). The average annual maximum water content of snow (in inches of water) measured at the snow course during 1942-94 is 22.7 in. (U.S. Department of Agriculture, Soil Conservation Service, written commun., 1993, and Ray Wilson, U.S. Department of Agriculture, Soil Conservation Service, oral commun., 1994). Water content determined from snow depth measured at the Payson Ranger Station snow course was 35.5 and 41.1 in. on April 15, 1983, and April 15, 1984, respectively.

The Ree's Flat snow course, at an altitude of 7,300 ft, has records from 1956 to the present (1995). The average annual maximum water content of snow measured at the snow course from 1956 to 1994 is 12.9 in. (U.S. Department of Agriculture, Soil Conservation Service, written commun., 1993, and Ray Wilson, U.S. Department of Agriculture, Soil Conservation Service, oral commun., 1994). Water content determined from snow depth measured at the Ree's Flat snow course was 24.1 and 22.0 in. on March 31, 1983, and April 24, 1984, respectively.

The annual freshwater evaporation estimated for the area north of Chicken Creek Reservoir was 53.8 in. (U.S. Department of Agriculture, 1969, p. 22). The average annual pan evaporation measured from 1948 to 1992 at Utah Lake near Lehi, about 30 mi north of Juab Valley, was 56.23 in. (Ashcroft and others, 1992, p. 96).

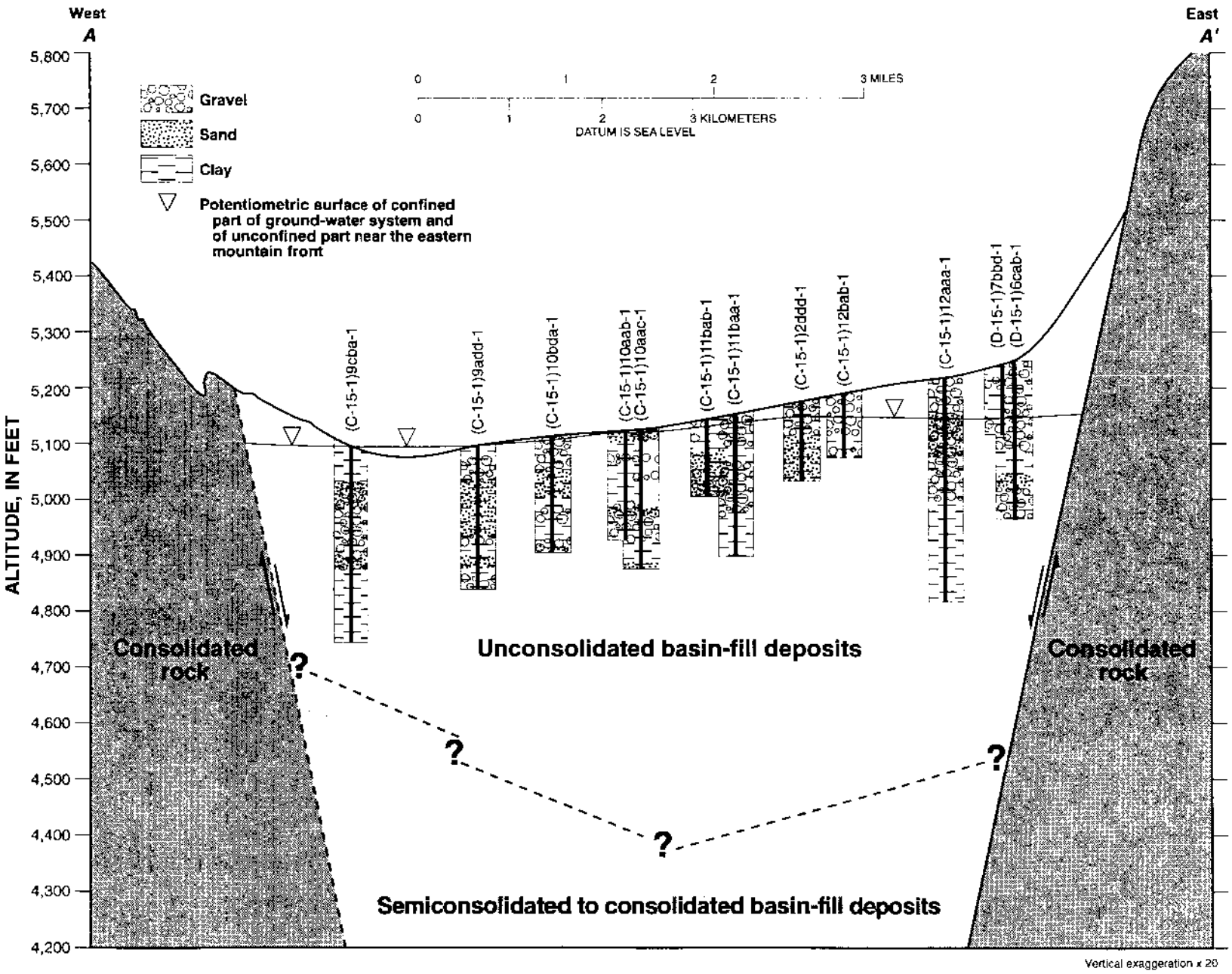


Figure 4. Generalized cross section through the southern part of Juab Valley, Utah (line of generalized cross section shown in fig. 3).

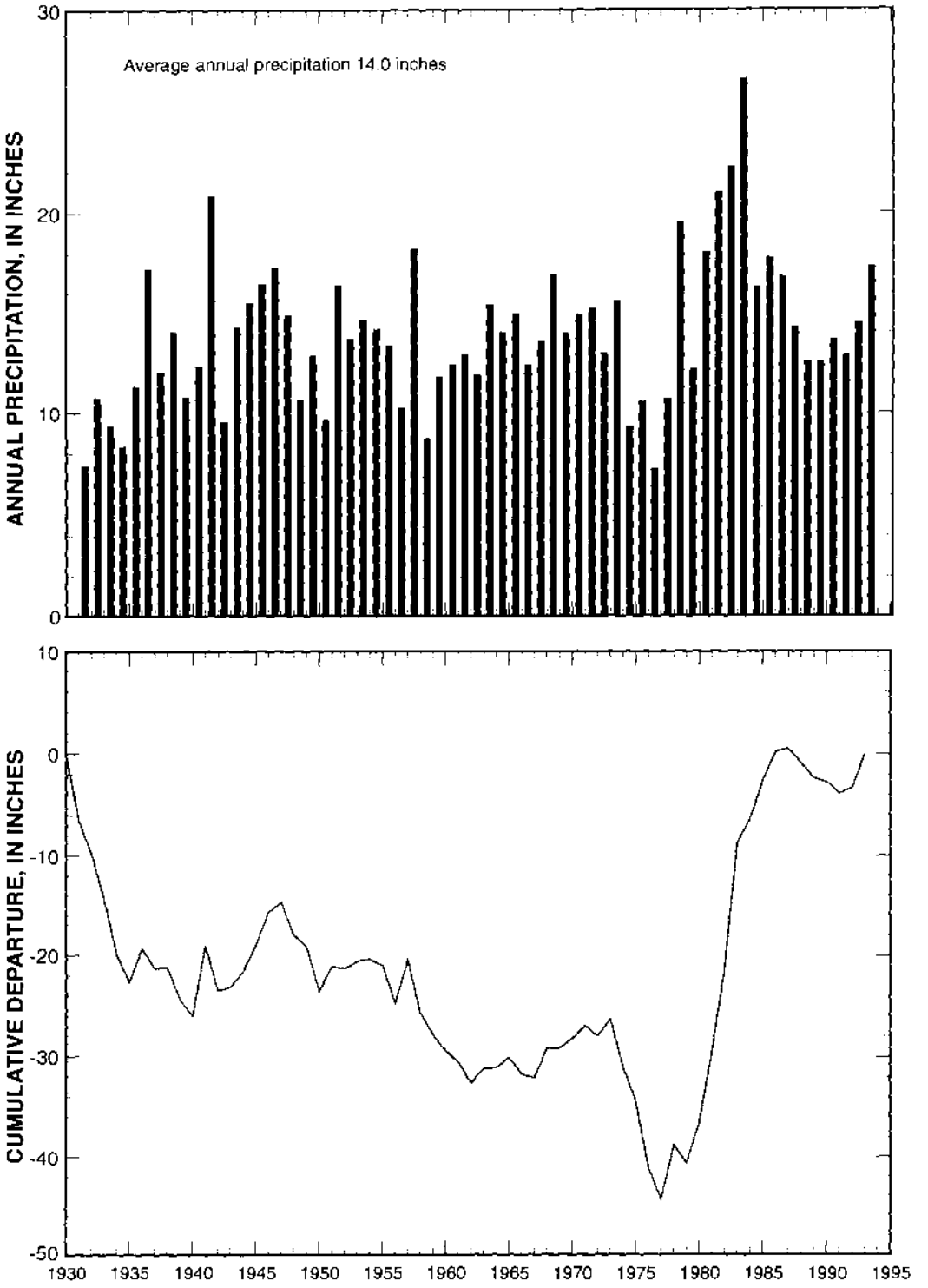


Figure 5. Annual precipitation and cumulative departure from the average annual precipitation, 1931-93, at Nephi, Utah.

HYDROLOGY

Most of the surface water that enters Juab Valley originates as precipitation on the southern Wasatch Range and the San Pitch Mountains. Ground water in Juab Valley occurs in the unconsolidated basin-fill deposits and is recharged primarily by infiltration of surface water and precipitation.

Surface Water

Surface water in Juab Valley is used primarily for irrigation. Much of the streamflow that enters the valley on the east side is diverted into irrigation distribution systems. Water also is stored in reservoirs in Juab Valley for irrigation outside of the valley.

Streams

Perennial and ephemeral streams drain the southern Wasatch Range and the San Pitch Mountains, and ephemeral streams drain Long Ridge and the West Hills. Most streamflow is diverted for irrigation at the mountain front. Only a small amount of surface water enters the valley from the west side, probably only after intense rainstorms and after the melting of greater-than-average winter snow accumulations.

Salt Creek, east of Nephi, is the largest stream that enters Juab Valley. It drains an area of 95.6 mi^2 that includes the east side of Mount Nebo in the southern Wasatch Range and the northern end of the San Pitch Mountains. Average annual streamflow for Salt Creek near Nephi (U.S. Geological Survey gaging station 10146000) (pl. 1) for water years 1952-80 was about 19,600 acre-ft. A water year is the 12-month period from October 1 through September 30 and is designated by the calendar year in which it ends. Peak flow typically occurs in May as a result of snowmelt runoff with an average monthly streamflow of about 85 ft^3/s. The average monthly streamflow from October to February ranged from 11.0 to 11.5 ft^3/s. Much of this flow is ground water discharged from consolidated rocks and channel-fill deposits upstream from the gaging station. Streamflow in Salt Creek is routed through a distribution system in Juab Valley in all but high flow situations. This water ultimately moves to the axis of the valley, west and northwest of Nephi.

Chicken Creek drains the San Pitch Mountains east of Levan with a drainage area of 27.9 mi^2. Average annual streamflow for Chicken Creek near Levan (U.S. Geological Survey gaging station 10219200) (pl. 1) for water years 1963-93 was about 5,800 acre-ft. Snowmelt runoff typically occurs in May with an average monthly streamflow of about 36 ft^3/s. Average monthly streamflow during the winter months varied from 2.6 ft^3/s in October to 1.8 ft^3/s in December to 2.2 ft^3/s in February. The average annual streamflow for water years 1982-86 was more than the period-of-record average because of greater-than-average precipitation. Average annual streamflow during 1987-92 was less than during any other recorded water years, except 1977, because of less-than-average precipitation.

Parts of the Salt Creek and Chicken Creek drainages contribute flow to springs that is piped to Nephi and Levan for public supply and irrigation. Bradley Spring, (D-13-2)5cbd-S1 in the Salt Creek drainage (pl. 1), has an estimated average annual discharge of about 3,000 acre-ft. About one half of the annual discharge occurs from April to September with generally more discharge from July to September (Roy Tolley, Nephi City, written commun., 1993). About 40 percent of the spring discharge is used for irrigation, and the remainder is used for public supply (Bjorklund, 1967, p. 45). Cobble Rock, Tunnel, and Rosebush Springs in the Chicken Creek drainage (pl. 1) have an estimated combined annual discharge of about 400 acre-ft. Water from these springs is used for public supply. The average annual streamflow from Salt and Chicken Creeks would be about 22,600 and 6,200 acre-ft, respectively, if spring discharge were not diverted.

Average annual streamflow for most of the streams that enter the east side of Juab Valley was estimated from a relation between streamflow measured in the two gaged streams near the valley margins and a regression equation developed for a region of the Colorado River Basin (Christensen and others, 1986, p. 8). The region includes all areas in the Colorado River Basin in Utah where streamflow is predominantly from snowmelt. The regression equation was considered appropriate for use because most of the streamflow in the study area is derived from snowmelt. The regression equation is

$$Q = 1.39 \times 10^{-5} A^{1.06} E^{4.67} \qquad (1)$$

where

Q = average annual streamflow, in ft^3/s;
A = contributing drainage-basin area, in mi^2; and
E = mean basin altitude, in thousands of feet.

Drainage-basin areas of the streams used to develop the regression equation ranged from 3 to

660 mi^2, and mean drainage-basin altitudes ranged from 7,560 to 10,960 ft. Most of the streams that enter Juab Valley from the San Pitch Mountains, including Salt and Chicken Creeks, have mean drainage altitudes less than those used to develop the regression equation. Average annual streamflow calculated from the regression equation using areas and mean altitudes for the Salt and Chicken Creek drainages was about 64 percent of the average annual streamflow determined from gaging-station records and spring discharge estimates. Flow in ungaged streams that enter Juab Valley from the east was estimated using the regression equation and then modified to better represent the area by dividing that value by 64 percent. The major drainages that enter the valley from the east and the estimated streamflow for these drainages are listed in table 1.

Flow in streams that drain the steep west side of the southern Wasatch Range is influenced by ground-water discharge from consolidated rocks. Lithology and geologic structure control ground-water discharge to these streams by channeling snowmelt runoff into fractures. This water is discharged to the mountain streams throughout the year. The regression equation modified by the gaging-station streamflow data method of estimating streamflow may not be valid for these streams. Flow in selected streams that drain the southern Wasatch Range and the San Pitch Mountains and the percentage of flow that occurs during the irrigation season were estimated from observations made by irrigation-company representatives and from the amount of water required to meet the demand of crops and irrigation losses (table 1). Streamflow from streams that drain the San Pitch Mountains estimated using the regression equation modified by its relation to gaged flow was similar to the amount estimated by observation. The field observations validate the use of the regression-equation/gaging-station estimate for most of the streams that enter the southern part of the valley.

Streamflow from Wash, Pole, and Couch Canyons in the southern Wasatch Range is mainly from springs that discharge from consolidated rocks. Discharge from these drainages and from Clover Creek Spring, (D-12-1)3bbc-S1 (pl. 1), is reported to increase in the late summer relative to the early spring. Discharge from most consolidated-rock springs is included in streamflow estimates.

Currant Creek and West Creek, in the northern part of Juab Valley, receive water discharged from springs and seeps in the valley, streamflow that has entered the valley at the mountain front during periods of greater-than-average snowmelt runoff, and surface runoff from snowmelt and rain from surrounding valley areas. Currant Creek begins about 3 mi south of Mona and flows north to Mona Reservoir. Bjorklund (1967, p. 40-42) estimated streamflow in Currant Creek in 1965-66 to be about 5,500 acre-ft/yr at a site that became U.S. Geological Survey gaging station 10146400 and an additional 3,200 acre-ft/yr between the gaging station and Mona Reservoir. These estimates were determined from a seepage study on Currant Creek in March 1966 (Bjorklund, 1967, p. 42) and relations between flow in the creek at that time and discharge measured from spring (D-12-1)6ddc-S1 at Burriston Ponds from May 1965 to March 1966. Virtually all of the flow in Currant Creek during 1965-66 (8,700 acre-ft/yr) is believed to be derived from ground-water discharge to springs and seeps in the area, which is referred to as base flow. Streamflow during this period likely was not affected by residual ground-water discharge resulting from greater-than-average precipitation and surface-water inflow because it was preceded by 9 years of generally less-than-average precipitation.

Gaging-station records are available for Currant Creek near Mona (U.S. Geological Survey gaging station 10146400) from 1979 through the present (1995). Average annual streamflow for water years 1979-93 is about 25,400 acre-ft, which is much greater than the discharge measured by Bjorklund (1967) during 1965-66. Greater-than-average precipitation during this period resulted in an increase in water stored in and discharged from the unconsolidated basin-fill deposits and an increase in streamflow from mountain streams. Average annual streamflow for Currant Creek at U.S. Geological Survey gaging station 10146400 for water years 1990-93 was about 7,300 acre-ft. Average annual precipitation measured at Nephi during this period was less than the 1931-93 average.

West Creek begins west of Nephi and flows north to Currant Creek. Most of the water in West Creek originates as streamflow diverted from Salt Creek into an irrigation distribution system or as local snowmelt runoff. Water that has seeped from the irrigation distribution system and irrigated fields to the ground-water system moves toward the axis of the valley where it is discharged as diffuse ground-water seepage to West Creek. A reconnaissance of streamflow in West Creek was done in December 1993. Selected discharge-measurement sites are shown in figure 6 and the measurements are presented by Steiger (1995, table 10).

During periods of high flow in the spring, water in the Salt Creek channel in excess of the distribution-

Table 1. Estimated average annual streamflow for drainages on the east side of Juab Valley, Utah

[—, no data or not applicable]

Drainage name	Drainage area (square miles)	Mean altitude (feet)	Estimated average annual streamflow (acre-feet per year) determined from: Regression equation and gaging-station records	Observations or estimated consumptive use	Estimated average annual streamflow assumed to be more representative of actual conditions (acre-feet per year)	Estimated flow occurring during the: Nonirrigation season (percent)	Nonirrigation season (acre-feet)	Irrigation season (percent)	Irrigation season (acre-feet)
Northern part									
Wash Canyon	1.6	—	—	1,500	[1]1,500	33	500	67	1,000
North Creek	4.3	8,810	1,200	3,000	3,000	33	1,000	67	2,000
Pole Canyon	4.5	8,820	1,300	500	[1]500	50	250	50	250
Bear Canyon	2.1	—	—	2,200	2,200	43	950	57	1,300
Couch Canyon	1.8	—	—	2,200	[1]2,200	50	1,100	50	1,100
Clover Creek Spring	—	—	—	2,200	[1]2,200	50	1,100	50	1,100
Willow Creek	6.9	8,530	1,700	2,200	2,200	43	950	57	1,300
Birch and Little Birch Creeks	2.6	—	—	500	500	33	160	67	340
Salt Creek	95.6	7,380	[2]22,600	—	[2]22,600	[3]22	[4]4,300	[3]78	[4]15,800
Biglows Canyon	3.8	6,900	540	—	540	33	180	67	360
Old Pinery and Suttons Canyons	7.3	7,180	1,300	—	1,300	33	430	67	870
Fourmile Creek	13.1	7,700	3,300	2,200	2,200	43	950	57	1,300
Total (rounded)					41,000		12,000		27,000
Southern part									
Pigeon Creek	13.9	7,400	2,900	2,900	2,900	16	460	84	2,400
Chicken Creek	27.9	7,410	[5]6,200	—	[5]6,200	[3]16	[6]930	[3]84	[6]4,900
Deep Creek	11.1	7,520	2,500	2,200	2,500	16	400	84	2,100
Little Salt Creek	19.5	7,150	3,600	3,300	3,600	16	580	84	3,000
Total (rounded)					15,000		2,400		12,400

[1]Most of annual streamflow is from spring discharge.

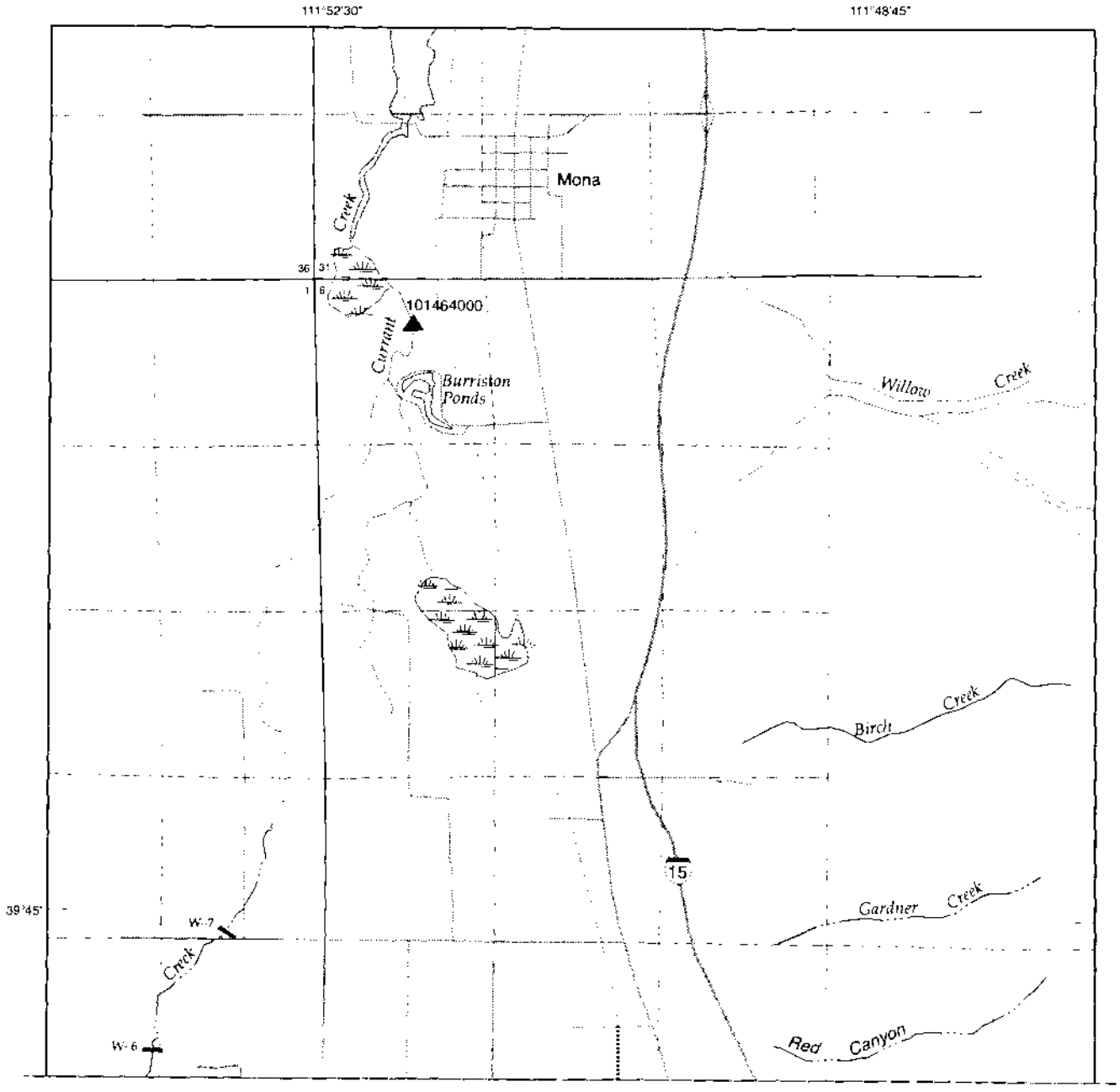
[2]Determined from gaging-station records. Includes 3,000 acre-feet per year estimated to discharge from Bradley Spring. During the irrigation season, an estimated 60 percent of Bradley Spring discharge is used for public supply (900 acre-feet per year) and 40 percent for irrigation (600 acre-feet per year).

[3]Estimated percent of flow in Salt and Chicken Creeks during the nonirrigation and irrigation seasons determined from gaging-station records.

[4]The quantity of Salt Creek streamflow occurring during the nonirrigation season is calculated from an average annual streamflow of about 19,600 acre-feet per year. During the irrigation season, flow from Bradley Spring used for irrigation is added to the average annual streamflow.

[5]Determined from gaging-station records. Includes 400 acre-feet per year estimated to discharge to springs used for public supply.

[6]The quantity of Chicken Creek streamflow occurring during the nonirrigation and irrigation seasons is calculated from an average annual streamflow of about 5,800 acre-feet per year.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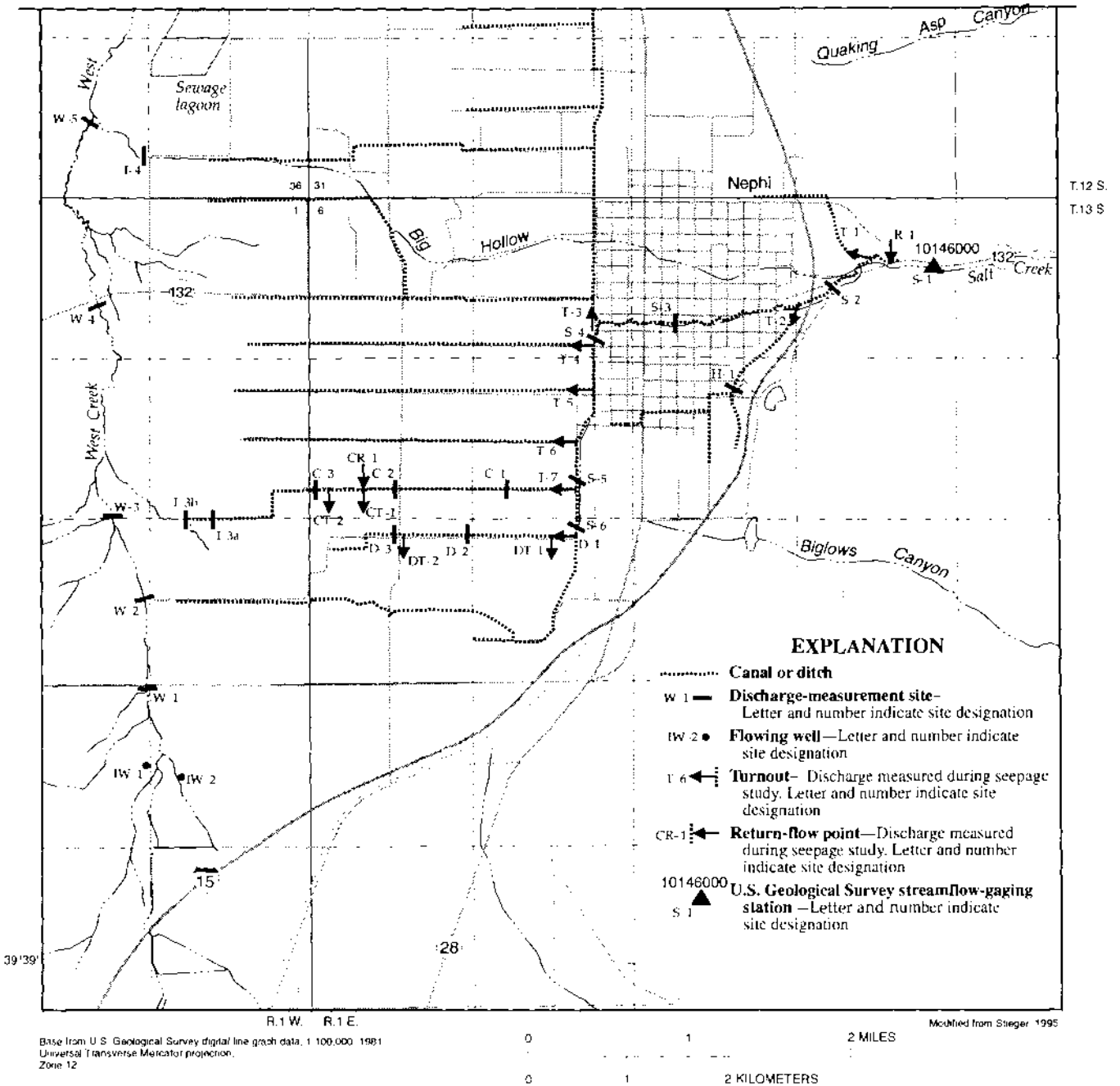


Figure 6. Discharge-measurement sites on Salt Creek, on selected irrigation canals and ditches, and in the West Creek area, Juab Valley, Utah.

system capacity is routed to a flood-control channel called Big Hollow that terminates at West Creek (fig. 6). The addition of this surface water (fig. 6, site I-4) resulted in almost 15 times more streamflow measured in West Creek at (C-12-1)24cdd (fig. 6, site W-7) on May 24, 1993, than on December 14, 1993, during base-flow conditions. Ditches were observed transporting water from West Creek and from seeps and springs downstream from site W-7 during the reconnaissance. Water from West Creek was followed to ditches in the seepage area south of Burriston Ponds (fig. 6, site W-8). Total discharge measured in the ditches was 324 gal/min or about 500 acre-ft/yr, which is assumed to drain to Currant Creek. West Creek intersects Currant Creek just downstream from Burriston Ponds, according to the U.S. Geological Survey 1:24,000 topographic quadrangle map of the area. Streamflow in West Creek is included in the estimate of streamflow in Currant Creek at U.S. Geological Survey gaging station 10146400.

Chicken Creek, in the southern part of Juab Valley, receives ground water discharged to springs and seeps in unconsolidated basin-fill deposits near Chicken Creek Reservoir. Most of this discharge is removed by evapotranspiration. Streamflow in Chicken Creek from the mountain front typically does not reach Chicken Creek Reservoir except during high-runoff years because of diversions for irrigation.

Distribution Systems

Canals, ditches, and pipelines are used to distribute water from near the mountain fronts to downgradient irrigated fields in Juab Valley. These systems are generally lined near the mountain fronts, except for the Salt Creek channel through Nephi. The distribution systems carry surface water, ground water discharged from wells, or both. The potential for seepage to the ground-water system exists where the canal or ditch is unlined or is in disrepair and the hydraulic gradient between it and the unconsolidated basin-fill deposits it crosses is downward. Gains to the canal or ditch from the ground-water system can occur where the hydraulic gradient is upward.

Streamflow seepage studies done in 1993 on Salt Creek and selected diversions from the creek in the Nephi area indicated that both gains to the stream and losses to the ground-water system occur. The stream was measured at six sites from near the mountain front to the end of the south main diversion (fig. 6, sites S-1 through S-6), a distance of about 3 mi, on August 9-11 and November 3-5, 1993. Two distribution ditches, one concrete lined (fig. 6, sites T-7 and C-1 through C-3) and the other unlined (fig. 6, sites D-1 through D-3), also were studied for seepage on August 9-11, 1993. Three sets of discharge measurements were made during each period at selected sites, including turnouts (diversions of water from the stream, canal, or ditch) and return-flow points (diversions of water to the stream, canal, or ditch), and are listed by Steiger (1995, table 9). Continuous water-stage records for selected reaches generally indicate that adjustments for fluctuations in streamflow throughout time are not necessary.

Net gains and losses for the reaches determined from selected streamflow measurements made during the seepage studies are shown graphically in figure 7. The streamflow that would be expected at the downstream end of the reach was computed by subtracting the flow measured at the turnouts and adding the flow measured at the return-flow points within the reach to the streamflow measured at the upstream end of the reach. This calculated value was subtracted from the value measured at the downstream end of the reach to determine the seepage gain or loss between the upstream and downstream ends of the reach. These calculations were made for each set of measurements. Computed seepage values were plotted as a function of reach length. A dashed line was fitted through the plotted points for each reach, and the rate of gain or loss was estimated from this line.

Within a given reach, the seepage gain or loss varied in each set of discharge measurements. This variation is shown by the scatter of the plotted points in figure 7. The scatter is attributed to one or more of the following: (1) poor measuring conditions, (2) changes in the rate of seepage, and (3) the possibility that a water user changed the volume of flow in his turnouts or return-flow points during the time of the discharge measurements.

Streamflow measurements made on Salt Creek and its south main diversion in August took place during thunderstorms which created unstable flow and thus variable discharge measurements. The scatter in this set of discharge measurements precludes its use to estimate seepage. The seepage study made in November was not affected by precipitation and no inflow from return-flow points was measured because irrigation wells that supplement flow in Salt Creek were not being pumped. The average of the three discharge measurements made in November was 15.4 ft^3/s at site S-2 (Salt Creek on the east side of Nephi) and 1.58 ft^3/s at site S-6 (the south end of the south main diversion). Generally, the

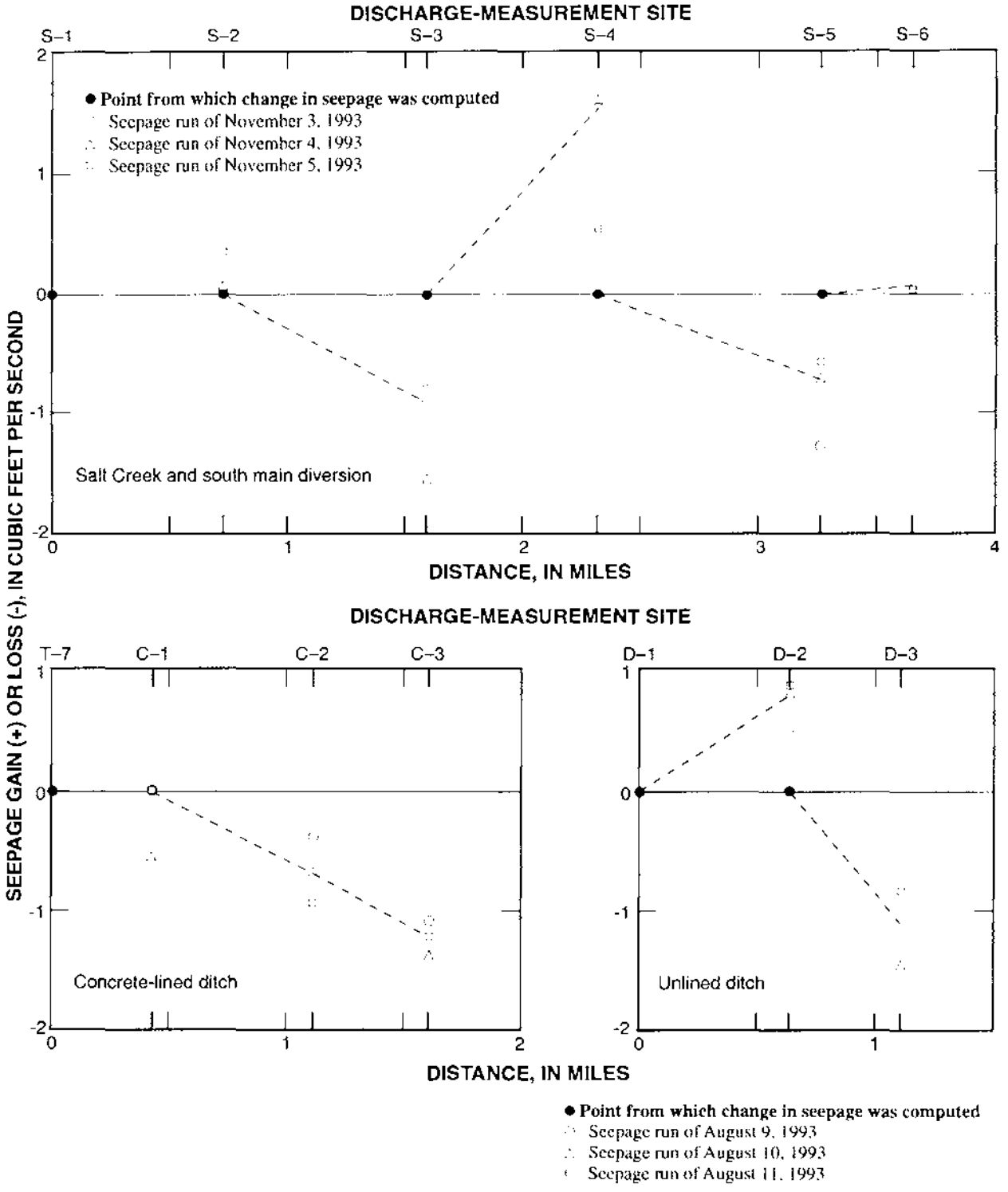


Figure 7. Estimated average seepage gain or loss for reaches of Salt Creek and selected irrigation canals and ditches, Juab Valley, Utah. Discharge-measurement site locations are shown in figure 6.

measurements were rated as being within 5 to 8 percent of the actual discharge. Many of the reaches had calculated gains or losses of the same order of magnitude as the measurement error. A graphical determination of the change in flow, based on at least three sets of measurements, reduces the effects of variation in flow caused by measurement error.

Measurements of streamflow in Salt Creek on November 3-5, 1993, indicated no net change in flow for reach S-1 to S-2 (0.75 mi), a net loss of about 0.9 ft^3/s to the ground-water system for reach S-2 to S-3 (0.84 mi), and a net gain of about 1.5 ft^3/s to the stream for reach S-3 to S-4 (0.74 mi). The net loss in flow in reach S-2 to S-3 might be caused by coarse-grained material underlying the streambed and a downward hydraulic gradient that would allow part of the streamflow to enter the ground-water system. The net gain in flow in reach S-3 to S-4 might be discharge from a shallower part of the ground-water system that is recharged by lawn and garden watering. Reach S-4 to S-5, part of the south main diversion from Salt Creek, had a net loss of about 0.7 ft^3/s in 0.93 mi. The loss in flow in this reach is probably the result of seepage to the underlying coarse-grained material and lesser amounts of upgradient unconsumed irrigation water seeping into the ditch. Reach S-5 to S-6 had an estimated gain of about 0.2 ft^3/s in 0.31 mi. This gain in flow is attributed to seepage to the ditch from upgradient unconsumed irrigation water. No change in flow was indicated in reach T-2 to H-1, a concrete-lined ditch on the east side of Nephi.

Irrigation water is distributed from the south main diversion (S-4 to S-6) westward through four major ditches, most of which are lined with concrete. Streamflow measurements made in two ditches (fig. 6, sites T-7 to C-3 and D-1 to D-3) in August 1993 indicated net losses in flow. In the southernmost concrete-lined ditch (T-7 to C-3), no change in flow was calculated for reach T-7 to C-1 (0.43 mi). A loss of about 1.2 ft^3/s in 1.19 mi was calculated for reach C-1 to C-3. Broken sections of concrete allow seepage to underlying coarse-grained material. In the southernmost unlined ditch (D-1 to D-3), a gain in flow of about 0.8 ft^3/s in 0.64 mi was calculated for reach D-1 to D-2 and a loss in flow of about 1.2 ft^3/s in 0.46 mi was calculated for reach D-2 to D-3. The gain in flow in the upper reach of the ditch (D-1 to D-2) is probably the result of seepage to the ditch from upgradient unconsumed irrigation water. The loss in the lower part of the ditch (D-2 to D-3) is probably the result of seepage to underlying coarse-grained material.

Reservoirs

Mona and Chicken Creek Reservoirs store water in Juab Valley for use outside of the valley. Both reservoirs are at the lowest parts of Juab Valley and are underlain by fine-grained material. All surface water flowing out of the valley, except for that at the southernmost end, must pass through one of these reservoirs.

Mona Reservoir is used to store water for irrigation in Goshen Valley, about 4 mi northwest of Juab Valley. The reservoir receives water from Currant Creek, flowing wells along its eastern edge, and springs and seeps. Mona Reservoir has an estimated average surface area of about 640 acres (1 mi^2) (Bjorklund, 1967, p. 42). Streamflow in Currant Creek near Goshen (U.S. Geological Survey gaging station 10146500), downstream from Mona Reservoir (pl. 1), was monitored from 1953 to 1960. The monthly streamflow at the gaging station for this period ranged from 1.91 ft^3/s in March to 57.4 ft^3/s in May and represents regulated streamflow from Mona Reservoir. Annual streamflow ranged from about 19,400 acre-ft in 1954 to about 13,500 acre-ft in 1960. The average annual streamflow from Mona Reservoir during this period was about 15,800 acre-ft. Direct precipitation on the reservoir is about 750 acre-ft/yr (14.0 in/yr x 640 acres) and evaporation from the reservoir is about 3,000 acre-ft/yr (56.2 in/yr measured at Utah Lake x 640 acres). As a result, the average net water loss from Mona Reservoir is about 18,000 acre-ft/yr.

Chicken Creek Reservoir stores ground-water discharge and local precipitation runoff in the southern part of Juab Valley. The water is used for irrigation in the Mills area, about 4 mi south of Juab Valley. The reservoir has an estimated average surface area of about 300 acres (Bjorklund and Robinson, 1968, p. 39). The average streamflow from Chicken Creek Reservoir is estimated to be about 2,500 acre-ft/yr, although less water has been available for release in recent years (Max Williams, Juab Lake Irrigation Co., oral commun., 1994). Direct precipitation on the reservoir is about 350 acre-ft/yr (14.2 in/yr x 300 acres) and evaporation from the reservoir is about 1,350 acre-ft/yr (53.8 in/yr x 300 acres). As a result, the average net water loss from Chicken Creek Reservoir is about 3,500 acre-ft/yr.

Ground Water

Occurrence

The consolidated rocks that surround Juab Valley contain and transmit variable amounts of ground water. Permeability, through pore spaces and fractures, controls the rate of ground-water movement through the rock. The southern Wasatch Range contains fractured carbonate rocks with bedding planes that dip to the west, toward Juab Valley. This area receives large amounts of precipitation relative to the rest of the study area. The potential exists for water to move through fractures and discharge below land surface to unconsolidated basin-fill deposits in Juab Valley. The rocks that make up the San Pitch Mountains generally are less fractured, dip to the east (away from Juab Valley), and receive less precipitation because they are at a lower altitude than those of the southern Wasatch Range. Ground-water flow through fractures and discharge to springs in the study area is less from these rocks than from rocks in the southern Wasatch Range. Consolidated rocks bounding the west side of the valley receive less precipitation than those bounding the east side, which results in less ground-water recharge. Estimates of spring discharge from consolidated rocks east of Juab Valley are included in streamflow estimates for the major stream drainages (see "Streams" section of this report).

The saturated unconsolidated basin-fill deposits form the principal ground-water system in Juab Valley. Ground water generally occurs in the more permeable layers of sand and gravel that are separated by less permeable layers of clay and silt and is under unconfined and confined conditions. The alluvial fans on the east side of the valley consist of unsorted debris-flow deposits of relatively low permeability adjacent to the mountain front and of interbedded silt, sand, gravel, and cobble deposits of relatively high permeability in the valley. Poor sorting generally decreases the permeability of coarse-grained material. The alluvial-fan deposits are coarser grained near the mountain fronts and finer grained in the lower areas of the valley. Ground water is generally unconfined near the mountain front, where several hundred feet of unsaturated material overlie the water table. Ground water in lower areas of the valley also is unconfined and is at or near land surface. Unconfined conditions typically occur between land surface and the first confining layer in the subsurface. These confining layers are generally within 50 ft of land surface, although their depth, thickness, and presence in the valley is variable.

Fine-grained material that can confine ground water was deposited in lacustrine depositional environments in the lower-elevation areas of Juab Valley, primarily at the northern and southern ends. Thin, discontinuous layers of clay, interbedded with coarser-grained material, were deposited in fluvial environments. The areas west of Nephi and near Levan (figs. 3 and 4) contain relatively thin confining layers in the subsurface. Ground water in these areas generally is confined.

Ground water in the unconsolidated basin-fill deposits in Juab Valley can flow from unconfined to confined conditions if the hydraulic gradient is downward and from confined to unconfined conditions if the hydraulic gradient is upward. Generally, recharge to the ground-water system occurs in areas where a downward hydraulic gradient exists and discharge from the system occurs in areas where an upward hydraulic gradient exists. The area delineated as wetlands by the U.S. Department of Agriculture, Soil Conservation Service (written commun., 1993) (pl. 1) is oriented generally along the axis of the valley and corresponds to where ground water is discharged by evapotranspiration. The ground-water divide that separates the ground-water system into the northern and southern parts in Juab Valley is believed to be near the topographic divide within the valley, Levan Ridge. Ground water recharged at the ground-water divide flows away in both directions, north and south.

The depth to the bottom of the unconsolidated basin-fill deposits that form the ground-water system in Juab Valley is not known. The continuous reflector determined from seismic-reflection data at about 1,200-2,300 ft below land surface (see "Geology" section of this report) may represent the boundary between unconsolidated and semiconsolidated basin-fill deposits in the valley. In other similar basins in Utah, the permeable deposits that yield water readily to wells occur in the upper 700 to 1,500 ft of basin fill (Gates, 1987, p. 79). Basin-fill deposits at depth likely become semiconsolidated to consolidated and are much less permeable than shallower unconsolidated deposits. Because there is little well data available in Juab Valley for depths greater than 1,000 ft, the discussion of the ground-water flow system in the valley generally applies only to the upper 1,000 ft of unconsolidated basin-fill deposits. Recharge, flow, and discharge discussed in this report are assumed to occur only in the upper 1,000 ft of unconsolidated basin-fill deposits.

The ground-water budget estimated for unconsolidated basin-fill deposits in the northern and southern parts of Juab Valley under average conditions is presented in table 2. Average conditions include 1963-93 precipitation and ground-water pumpage data and various periods of record for U.S. Geological Survey gaging stations in the study area. Average conditions represent the period of established ground-water pumpage from wells. The methods and computations used to derive the individual budget components are described and quantified in the following sections.

Recharge

Recharge to the unconsolidated basin-fill deposits is by seepage from streams, unconsumed irrigation water, and distribution systems; infiltration of precipitation; and subsurface inflow from consolidated rocks that surround the valley. Recharge varies annually but was estimated to be about 42,000-46,000 acre-ft/yr in the northern part and about 12,000 acre-ft/yr in the southern part, under average conditions.

Seepage from Streams

All of the perennial streams that enter Juab Valley from the east have been connected to a distribution system for irrigation. Streamflow during the irrigation season (April to September) was estimated to be some percentage of the estimated average annual streamflow from the drainage on the basis of gaging-station records and observations from representatives of local irrigation companies (table 1). Irrigation-season streamflow is about 27,000 acre-ft in the northern part and about 12,400 acre-ft in the southern part of Juab Valley. Because most of the flow during this period is diverted for irrigation, recharge to the ground-water system from streamflow is from losses during transmission or as unconsumed irrigation water (see "Seepage from unconsumed irrigation water and distribution systems" section of this report). Ephemeral streamflow to the east side of Juab Valley is assumed to be small, and seepage from ephemeral streams to the ground-water system is included with the estimate of recharge from subsurface inflow from consolidated rocks to the east side of the valley.

Streamflow during the nonirrigation season (October to March) generally is transmitted through pipelines, canals, and ditches to the valley. Streamflow from North and Willow Creeks and from Pole, Bear, and Couch Canyons is directed back into the stream channel near the mountain front in the winter. Streamflow that occurs during the nonirrigation season is assumed to recharge the ground-water system at a rate of 100 percent because of low evapotranspiration rates; the permeable nature of the alluvial-fan material; and the observation that none of the surface water that enters the valley at the mountain front, with the exception of Salt Creek, flows to the center of the valley. Nonirrigation-season streamflow is about 12,000 acre-ft in the northern part of Juab Valley and about 2,400 acre-ft in the southern part (tables 1 and 2).

Ephemeral streams that enter Juab Valley on the west side from the West Hills and Long Ridge recharge the ground-water system only after periods of greater-than-average precipitation and intense rainstorms. Measurements or observations of streamflow in this area were not made during this study. Recharge to the ground-water system from streams that enter the valley on the west side and from subsurface inflow from consolidated rocks on the west side was estimated to be 10 percent of the estimated annual streamflow that enters Juab Valley on the east side. This estimate is based on the lower drainage-basin altitudes and the corresponding smaller amounts of precipitation received. Additional data are needed to refine this rough estimate. Recharge from subsurface inflow and seepage from ephemeral streams to the west side of Juab Valley is about 4,200 acre-ft/yr in the northern part and about 1,500 acre-ft/yr in the southern part (table 2).

Seepage from Unconsumed Irrigation Water and Distribution Systems

An estimated 36,000 acre-ft/yr of surface water was diverted and an average of about 20,000 acre-ft/yr of ground water was pumped during 1963-93 for irrigation in Juab Valley. The amount of unconsumed irrigation water and distribution-system losses assumed to recharge the ground-water system was estimated to be about 30 percent of the surface water diverted and about 10 percent of the ground water pumped for irrigation. This is about 8,700 acre-ft/yr in the northern part and about 4,300 acre-ft/yr in the southern part of the valley (tables 2 and 3). About 20 percent of the irrigation water applied to the fields is estimated to be lost by water moving off of the fields and by evaporation. The remaining water, about 50 percent of the diverted surface water and 70 percent of the ground water pumped for irrigation, is available to meet the consumptive-use requirements of the crops.

Table 2. Estimated ground-water budget for unconsolidated basin-fill deposits under average conditions, Juab Valley, Utah

[—, not applicable]

Budget component	Flow (acre-feet per year)	Percent of total	Flow (acre-feet per year)	Percent of total
	Northern part		Southern part	
Recharge				
Seepage from nonirrigation-season streamflow	12,000	26-28	2,400	20
Seepage from unconsumed irrigation water and distribution systems	8,700	19-21	4,300	36
Seepage from irrigation-season streamflow not included in the unconsumed irrigation water and distribution-system losses component	750	2	0	0
Infiltration of precipitation	3,800	8-9	2,600	22
Subsurface inflow and seepage from ephemeral streams to east side of valley	12,500-16,500	30-36	1,200	10
Subsurface inflow and seepage from ephemeral streams to west side of valley	4,200	9-10	1,500	12
Total recharge (rounded)	42,000-46,000	100	12,000	100
Discharge				
Wells				
Pumped for irrigation and public-supply use	14,500	—	5,300	—
Pumped for domestic and stock use	150	—	50	—
Flowing wells	1,700	—	900	—
Well total (rounded)	16,300	36-39	6,300	52
Springs and seeps				
Seepage to Currant Creek upstream from gaging station 10146400[1]	5,500	—	—	—
Seepage to Currant Creek downstream from gaging station 10146400	3,200	—	—	—
Seepage to Mona Reservoir	5,800	—	—	—
Palmer Spring	—	—	700	—
Seepage to Chicken Creek Reservoir	—	—	1,100	—
Springs and seeps total (rounded)	14,500	31-35	1,800	15
Evapotranspiration	10,200	22-24	3,900	33
Subsurface outflow from consolidated rock/unconsolidated basin-fill deposit boundary	1,000-5,000	2-11	0	0
Total discharge (rounded)	42,000-46,000	100	12,000	100

[1] Includes an estimated 500 acre-feet per year of seepage to West Creek and 1,200 acre-feet per year of discharge to the southeast spring at Burriston Ponds, (D-12-1)6ddc-S1.

The partitioning of recharge to the ground-water system that is derived from surface-water streamflow and ground-water pumped from wells is based on an analysis of the average monthly consumptive use of water by alfalfa in the Levan area (U.S. Department of Agriculture, 1969, table 9). Water used for irrigation was assumed to be applied in accordance with the pattern of potential crop consumptive use and ground water was assumed to be used only to supplement streamflow. Under average conditions, the total amount of water available to meet crop demands during the irrigation season in the northern part of Juab Valley is about 22,000 acre-ft (table 3); of this amount, about 12,000 acre-ft is derived from streamflow (equivalent to 50 percent of the surface water diverted for irrigation) and about 10,000 acre-ft is derived from ground water pumped from wells (equivalent to 70 percent of the ground water pumped for irrigation). About 40 percent of the potential crop consumptive use occurs from April to June, and 40 percent of the surface and ground water available to the crops is about 8,800 acre-ft. Because ground water is a supplemental source for irrigation, it is not assumed to be used until all streamflow has been used. Because the 8,800 acre-ft/yr is about 70 percent of the streamflow available for crop use, no ground water was assumed to be needed for irrigation during April to June. The remaining 30 percent of the irrigation-season streamflow available for crop use and 70 percent of the ground water pumped from wells are used to satisfy the consumptive-use requirements of the crops during July to September.

The difference between the estimated amount of water available for crop use (22,000 acre-ft) and the estimated amount of water consumed by the crops (23,000 acre-ft) calculated for the northern part of the valley is less than 5 percent. This indicates that enough water is applied to meet the consumptive-use requirements of the crops.

The estimated amount of surface water used for irrigation in the northern part of Juab Valley does not include streamflow during the irrigation season from Biglows, Old Pinery, and Suttons Canyons and Fourmile Creek south of Nephi. Water from these drainages probably is used to irrigate some land at the mountain front, but these areas were not shown on recent maps of irrigated areas in the valley (Bookman-Edmonston Engineering, Inc., written commun., 1993). About 30 percent of the irrigation-season streamflow from these drainages (750 acre-ft) (table 2) is estimated to recharge the ground-water system.

Canals and ditches are the primary method used in the valley to distribute water to irrigated fields. Losses to the ground-water system are likely where the diverted water crosses permeable material, such as along alluvial fans, and where a downward hydraulic gradient exists between the canal and the underlying unconsolidated basin-fill deposits. Distribution systems in these areas generally are lined with concrete or pipe to reduce losses to the ground-water system and to evaporation. A seepage study done on a concrete-lined ditch and an unlined ditch southwest of Nephi indicated some net seepage to the ground-water system from both ditches (see "Distribution systems" section of this report). The concrete-lined ditch was broken in several places, resulting in the potential for losses. Water used for irrigation in Juab Valley was not differentiated by distribution-system lining or irrigation method because of a lack of data and to simplify the process of estimating seepage from unconsumed irrigation water and distribution systems.

Seepage from distribution systems to the ground-water system during transmission to irrigated fields is estimated to be greater in the spring when streamflow rates are high and evapotranspiration rates are low in comparison to the rest of the irrigation season. Transmission losses to the ground-water system in the summer months are assumed to decrease because less water is available for diversion. Water also is lost by moving off of the irrigated fields as tail-water runoff and from evaporation and therefore is not available to the crops or to the ground-water system. These losses were estimated to be 20 percent of the water diverted from streams and pumped from wells for irrigation. More water is lost as tail-water runoff from flood-irrigated fields than from sprinkler-irrigated fields, and a greater percentage of the water lost in the spring is to tail-water runoff. In the summer, losses to evaporation increase as losses to tail-water runoff decrease.

Field mapping of irrigated acreage done during 1992-93 (Bookman-Edmonston Engineering, Inc., written commun., 1993) included areas that had been irrigated in the past. Streamflow during the irrigation season and ground water pumped by irrigation wells can be applied to about 11,400 acres in the northern part and about 7,400 acres in the southern part of Juab Valley (table 3). A representative net crop-consumptive use rate of 24 in/yr (2.0 ft/yr) was estimated on the basis of crop type and crop acreage in Juab Valley (Fred Barnes, Franson-Noble and Associates, Inc., written commun., 1993). The amount of irrigation water needed by crops is approximated by multiplying this rate by irri-

Table 3. Estimated recharge from unconsumed irrigation water and distribution-system losses to unconsolidated basin-fill deposits, Juab Valley, Utah

[—, not applicable]

Location	Ground-water pumpage for irrigation (acre-feet per year)	Streamflow used for irrigation (acre-feet per year)	Recharge from water diverted for irrigation (acre-feet per year)	Percent of water diverted for irrigation that recharges ground-water system	Irrigation water available for crop use (acre-feet per year)	Irrigated area (acres)	Crop consumptive use (acre-feet per year)	Difference between water available for crop use and crop consumptive use (acre-feet per year)
Northern part[1]								
1963-93 average pumped for irrigation								
North of Mona Reservoir	200	1,000	320	27	630	500	1,000	-370
North Creek area	3,000	2,000	900	18	3,100	1,500	3,000	+100
Mona area	1,000	5,000	1,600	27	3,200	2,200	4,400	-1,200
Nephi area	[2]10,300	16,100	5,900	22	15,200	7,200	14,400	+800
Total (rounded)	14,500	24,000	8,700	—	22,000	11,400	23,000	-700
1990-92 average pumped for irrigation								
North of Mona Reservoir	400	1,000	340	24	780	500	1,000	-220
North Creek area	3,300	2,000	930	18	3,300	1,500	3,000	+300
Mona area	1,400	5,000	1,600	25	3,500	2,200	4,400	-900
Nephi area	[2]12,900	16,100	6,100	21	17,000	7,200	14,400	+2,600
Total (rounded)	18,000	24,000	9,000	—	25,000	11,400	23,000	+1,800
Southern part								
1963-93 average pumped for irrigation								
Levan area (rounded)	5,300	12,400	4,300	24	9,700	7,400	15,000	-5,300
1990-92 average pumped for irrigation								
Levan area (rounded)	8,400	12,400	4,600	22	12,000	7,400	15,000	-3,000

[1]Does not include irrigated areas downstream from Biglows, Old Pinery, and Suttons Canyons and Fourmile Creek south of Nephi. The estimated irrigation-season streamflow from these drainages is about 2,500 acre-feet per year.

[2]Includes ground water pumped from a well for public supply in Nephi to supplement water from Bradley Spring.

gated area. Irrigated crops in Juab Valley require the diversion of about 38,000 acre-ft/yr of water.

The irrigated area determined from aerial photographs and field surveys was about 13,500 acres in 1966 in the northern part of the valley (Hyatt and others, 1969, table 27) and about 3,900 acres in the early 1960's in the Levan area (U.S. Department of Agriculture, 1969, table 9). A comparison of land irrigated in the 1960's and in the early 1990's indicates that irrigated area has decreased slightly in the northern part of the valley and has almost doubled in the southern part. The addition of 12 irrigation wells since 1963 in the Levan area corresponds with the increase in irrigated area.

Because the increase in water pumped from wells in the southern part of Juab Valley is a result mainly of an increase in irrigated area, ground-water pumpage in this area was averaged for the 1990-92 period to correspond with the irrigated areas mapped during 1992-93. Recharge to the ground-water system from unconsumed irrigation water and distribution-system losses using a 1990-92 average pumping rate of about 8,400 acre-ft/yr is about 4,600 acre-ft/yr (table 3). The difference between the estimated amount of water available for crop use (12,000 acre-ft/yr) and the estimated amount of water consumed by the crops (15,000 acre-ft/yr) is a deficit of 3,000 acre-ft/yr (table 3). This shortage may be caused by an overestimation in the area being irrigated and in the amount of water lost to the crops by evaporation, tail-water runoff, and recharge to the ground-water system. A limitation of using the 1990-92 period is the effect of less-than-average precipitation on streamflow from mountain drainages and the resulting increase in ground-water pumped from wells to meet crop requirements.

Infiltration of Precipitation

The average annual precipitation at Nephi and Levan from 1931 to 1993 was about 14.0 in. (1.2 ft). The northern part of Juab Valley contains about 65,880 acres (103 mi^2) and the southern part about 43,850 acres (69 mi^2) of unconsolidated basin-fill deposits. Recharge to the ground-water system by infiltration of direct precipitation was estimated to be 5 percent of the average annual precipitation or about 3,800 acre-ft/yr in the northern part and 2,600 acre-ft/yr in the southern part of the valley (table 2). This estimate of recharge from precipitation is based on a range of values derived from previous studies done on other alluvial basins in Utah. These studies indicate that 1 to 25 percent of precipitation infiltrates to the ground-water system in areas receiving 8 to 16 in. of precipitation annually (Razem and Steiger, 1981, p. 13; Hood and Waddell, 1968, p. 24; and Feth and others, 1966, p. 43). Most infiltration occurs after the winter snow melts.

Subsurface Inflow

The amount of ground water that enters Juab Valley on the east side from consolidated rocks at the boundary with unconsolidated basin-fill deposits was calculated as the difference between estimated total discharge from the ground-water system and other forms of recharge to the ground-water system. This method assumes that the ground-water system is in equilibrium: the amount of water that enters the system is the same as the amount that leaves the system. Seepage from ephemeral streams on the east side of the valley is included in the estimate for subsurface inflow from consolidated rocks to the east side of the valley. Recharge from subsurface inflow to the west side of Juab Valley is included in the estimate for seepage from streams on the west side of the valley.

In the northern part of Juab Valley, recharge from subsurface inflow to the east side of the valley is estimated to be from 12,500 to 16,500 acre-ft/yr (table 2). The effect of ground-water pumpage from wells on water levels indicates that most of this water enters the ground-water system where Salt Creek enters the valley. For example, the average annual discharge from six wells in sections 4 and 5, Township 13 South, Range 1 East during 1963-93 was about 6,900 acre-ft. Although the potentiometric surface in this area is fairly flat and only a limited number of data points are available, no cone of depression is evident (fig. 8). Recharge from subsurface inflow to the area may compensate for the ground-water pumpage.

The difference between recharge and discharge components for the ground-water system in the southern part of Juab Valley, except for recharge from subsurface inflow and seepage from ephemeral streams, is about 1,200 acre-ft/yr. The assumption that the ground-water system is in equilibrium in this part of the valley may not be valid because of ground-water pumpage from wells in the Levan area. On the basis of a limited number of data points, the direction of ground-water flow has been reversed near the mountain front east of Levan (fig. 8). Some of the residual water therefore may be water removed from storage and some may be recharge from subsurface inflow from consolidated rocks on the east side of the valley.

Flow

The approximate potentiometric surface during March and April 1993 in the unconsolidated basin-fill deposits in Juab Valley is shown in figure 8. The potentiometric-surface map was made using water levels in wells (Steiger, 1995, table 1), generally from 100 to 300-ft deep, that were finished in unconsolidated basin-fill deposits. This interval generally represents confined conditions in the ground-water system. The direction of ground-water flow is perpendicular to the potentiometric contours in an isotropic aquifer. Ground water generally flows from the mountain-front recharge areas (unconfined conditions) toward discharge areas along the axis and near the lowest parts of the valley (confined conditions). The ground-water flow direction is generally to the north in the northern part of the valley and to the south in the southern part. Water-level data are not available in the area of the topographic divide (Levan Ridge) that separates the northern and southern parts of Juab Valley, but the ground-water divide is assumed to be near the topographic divide.

Ground-water pumpage from wells in areas near and in Nephi and Levan has altered the direction of ground-water flow from that of pre-ground-water development time. In the spring of 1993, the direction of flow for water in unconsolidated basin-fill deposits west of Nephi was from the southwest and the southeast to the north (fig. 8). In the spring of 1950, most of the ground water in this area is believed to have come from the east (Bjorklund, 1967, pl. 4). In the spring of 1993, the direction of ground-water flow on the west side of Levan was generally to the east (fig. 8) because of ground-water pumpage from wells in the area. During 1963-64, the direction of flow in the area was generally to the west, although pumpage affected potentiometric contours during this period also (Arnow and others, 1964, fig. 28).

Prior to the construction of irrigation-distribution systems, most of the streamflow that entered Juab Valley from the mountains on the east infiltrated the coarse-grained alluvial-fan deposits and was recharged to the ground-water system. A letter written by a Mona area resident in 1891 indicated that high streamflow rates resulting from snowmelt runoff in the spring and the coarse-grained nature of the deposits near the mountain front prevented the construction of a distribution system for water from North Creek. Because most of the streamflow in this area occurs in the spring, recharge to the ground-water system would be the greatest at that time also. Water levels in wells completed in unconsolidated basin-fill deposits under confined conditions near Mona, downgradient from North Creek, were reported to rise from June to August in 1891. This lag in water-level response to a recharge pulse of about 2 months probably is caused by the time of travel through unsaturated deposits near the mountain front.

Three flow paths that were hypothesized from the potentiometric surface and from the occurrence of recharge and discharge areas are shown in figure 8. A flow path in the southern part of Juab Valley is hypothesized from west of Levan to the ground-water discharge area near the northeast end of Chicken Creek Reservoir. In the northern part of the valley, a flow path extending downgradient from where Salt Creek enters the valley to the Burriston Ponds area and a flow path that represents ground-water flow from the southern Wasatch Range north of Nephi to the Burriston Ponds area are hypothesized. These flow paths are substantiated by the water chemistry, isotope, and geochemical data analysis described in the "Chemical characteristics of water" section of this report.

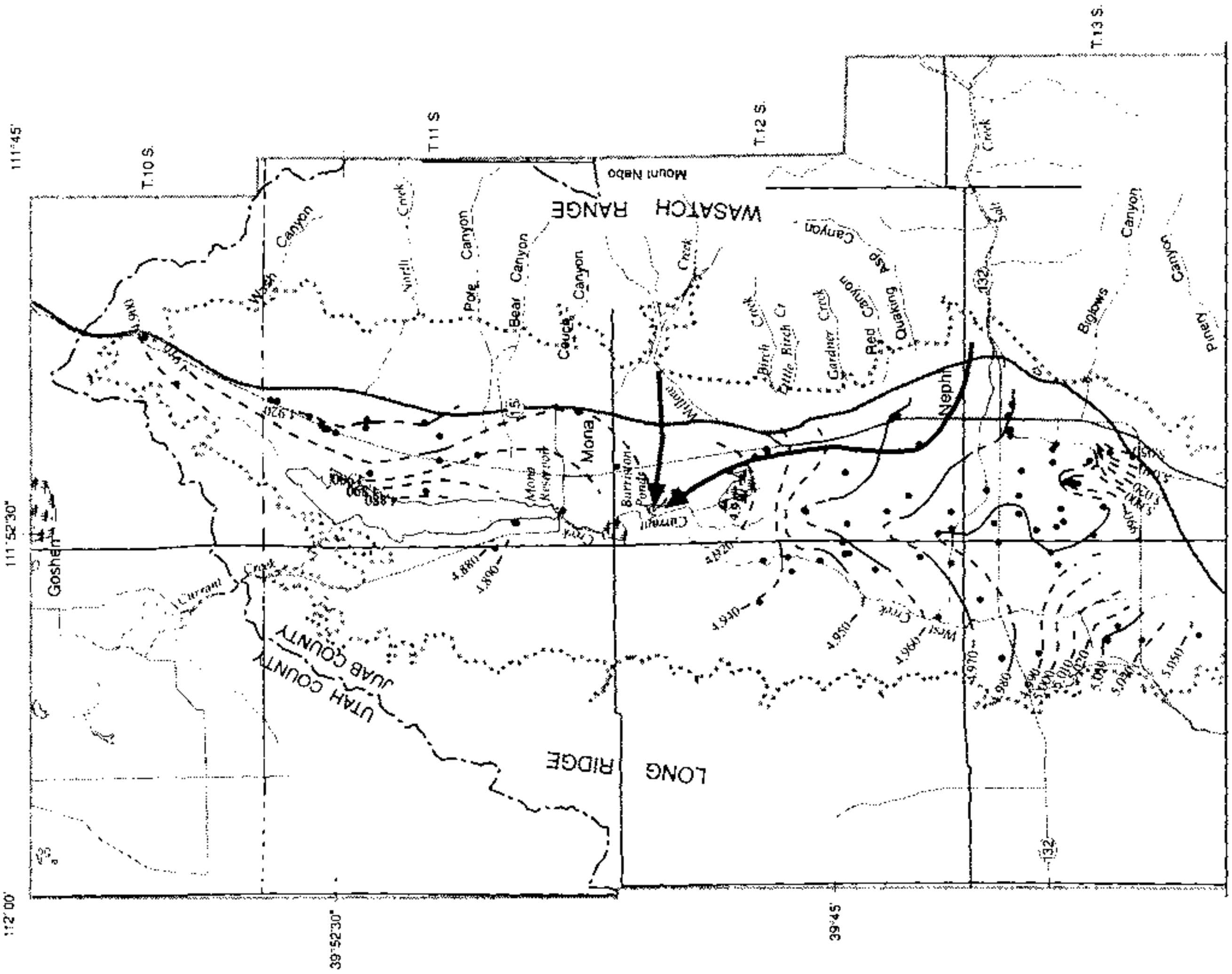
Discharge

Discharge from the unconsolidated basin-fill deposits in Juab Valley is by wells, springs, seeps, evapotranspiration, and subsurface outflow to consolidated rocks. Discharge varies annually, but was estimated to be about 42,000-46,000 acre-ft/yr in the northern part and about 12,000 acre-ft/yr in the southern part of the valley, under average conditions (table 2).

Wells

Ground-water discharge to wells completed in unconsolidated basin-fill deposits in Juab Valley is a major discharge component of the ground-water budget. Large-scale ground-water development for irrigation began in 1947, when the first well with a relatively large yield was drilled. Ground water in most of the valley is used to supplement surface water for irrigation, and wells capable of discharging large quantities of water generally begin pumping after the spring snowmelt runoff ends. More surface water is available for irrigation when precipitation is greater than average; consequently, pumpage from wells decreases.

The average annual ground-water discharge to wells in Juab Valley during 1963-93 was estimated to be about 23,000 acre-ft, on the basis of data collected



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Burriston Ponds
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Bear Canyon
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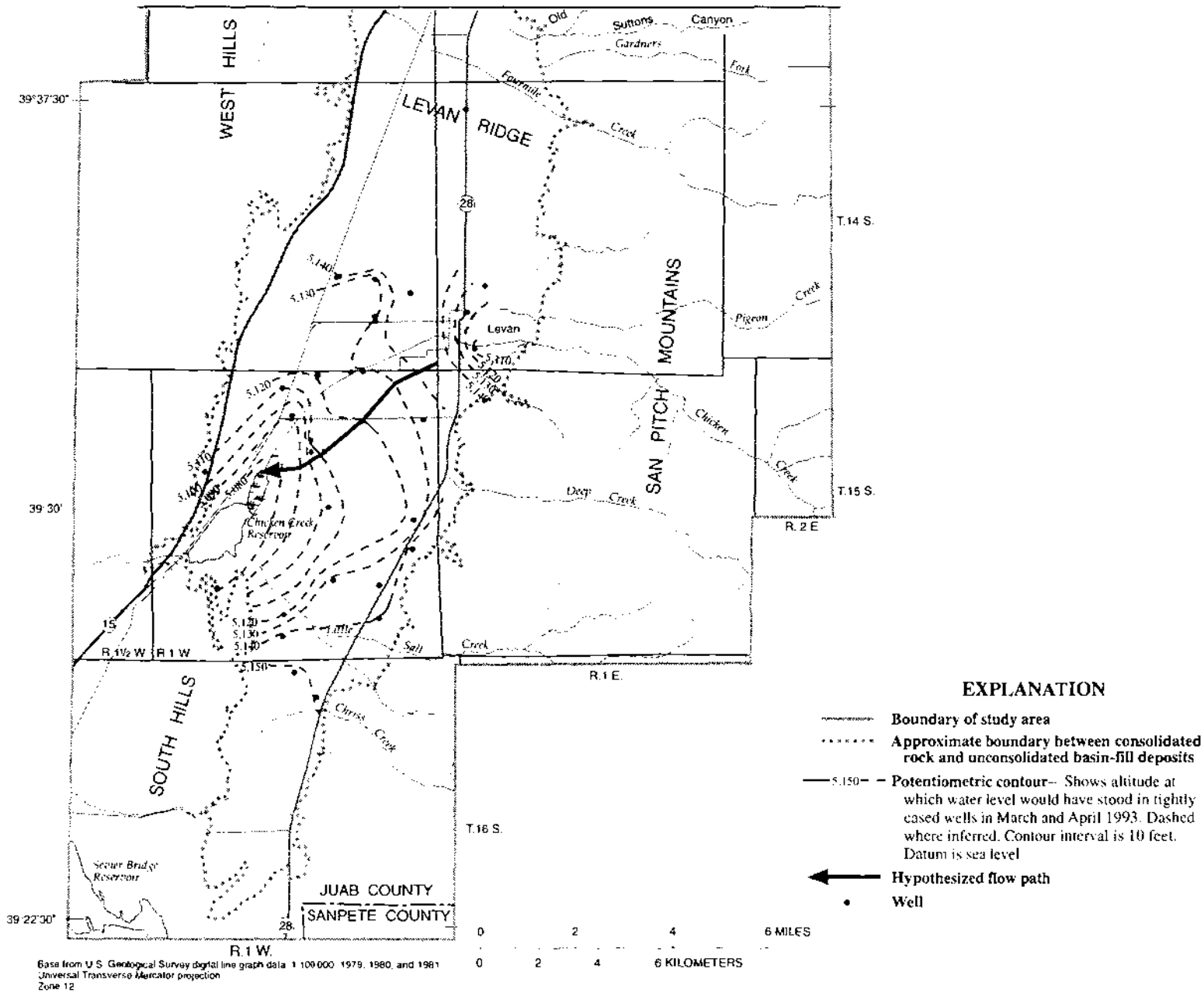


Figure 8. Approximate potentiometric surface in unconsolidated basin-fill deposits in Juab Valley, Utah, March and April 1993.

by the U.S. Geological Survey and summarized in a series of annual reports on ground-water conditions in Utah (Allen and others, 1994, p. 38). These values were modified to include ground-water discharge to flowing wells. During this period, about 16,300 and 6,300 acre-ft/yr is estimated to discharge to wells in the northern and southern parts of the valley, respectively (table 2).

Ground-water discharge to pumped wells in Juab Valley was estimated to be about 18,800 acre-ft in 1993. Most of this water, about 11,000 acre-ft pumped from 24 wells in the northern part and about 7,100 acre-ft pumped from 16 wells in the southern part, was used for irrigation in the valley. An estimated 200 acre-ft/yr was pumped from wells in Juab Valley for domestic and stock use, with about 75 percent of the total from the northern part. Ground water pumped from wells in 1963 was reported to be 14,600 acre-ft in the northern part and 2,600 acre-ft in the southern part of the valley (Arnow and others, 1964, p. 51 and 55). The 1963-93 average annual ground-water discharge to pumped wells for irrigation and public supply in the northern and southern parts of the valley is about 14,500 and 5,300 acre-ft/yr, respectively (table 2).

Discharge to flowing wells in 1993 was estimated to be 970 acre-ft in the northern part and 700 acre-ft in the southern part of the valley. Most of this water flows from large-diameter wells near Mona and Chicken Creek Reservoirs and is stored in the reservoirs for irrigation outside Juab Valley. The remaining discharge to flowing wells is used for stock watering in the valley. Ground-water discharge to flowing wells in 1963 was about 2,500 acre-ft in the northern part and 1,100 acre-ft in the southern part. The flowing-well discharge measured or estimated in 1963 and 1993 was assumed to be the average for each year. Another assumption made is that flowing wells in the valley flow throughout the year. The decrease in flowing-well discharge from 1963 to 1993 is probably the result of continued ground-water pumpage in the valley coupled with less-than-average precipitation during 1988-91. Flowing-well discharge was assumed to linearly decrease from 1963 to 1993, and the average discharge during the 1963-93 period was estimated to be 1,700 acre-ft/yr in the northern part and 900 acre-ft/yr in the southern part of the valley (table 2).

Springs and Seeps

Ground-water discharge from unconsolidated basin-fill deposits occurs as flow from distinct springs and as diffuse seepage, mainly in the low-lying areas of Juab Valley where the hydraulic head of the ground water is above land surface. Discharge to springs and seeps generally has decreased as pumpage from wells has increased in order for the ground-water system in the valley to approach a new state of equilibrium. Average annual discharge to springs and seeps in the unconsolidated basin-fill deposits was estimated to be about 14,500 acre-ft in the northern part and about 1,800 acre-ft in the southern part of Juab Valley (table 2). This estimate is based on data available for 1963-94.

Most of the springs in the valley north of Nephi discharge from the top of a continuous clay layer near a land-surface altitude of about 4,930 ft. This layer extends from north of Nephi to north of Mona Reservoir in Juab Valley and is described in several drillers' logs as a blue clay that is as much as 30 ft thick. Exposure of the blue clay layer at about this elevation by erosion has resulted in ground-water discharge at land surface. Ground water in the valley that moves downgradient along flow paths from the south and the east intersects the blue clay layer at land surface and discharges through springs in the vicinity of Burriston Ponds (fig. 8), which were first described in 1776 by the Dominguez-Escalante expedition as a "copious running spring of good water" (Chavez, 1995, p. 74).

A northern and a southern group of ponds form Burriston Ponds. Discharge measured from the outlet for the northern set of ponds was 1.6 ft^3/s in March 1966 (Bjorklund, 1967, p. 42). The spring at the southeast corner of the Burriston Ponds complex, (D-12-1)6ddc-S1 (pl. 1), discharged 2.7 ft^3/s in March 1966, which was about 54 percent of the streamflow measured at the same time at the outlet for the southern set of ponds (5.0 ft^3/s) (Bjorklund, 1967, p. 42). The remainder of flow measured at the outlet for the southern set of ponds originates at other springs and seeps in the area. On the basis of correlation of four sets of flow measurements made during 1965-66 at the southeast spring at Burriston Ponds and at Currant Creek near Mona (U.S. Geological Survey gaging station 10146400), discharge to the spring is about 22 percent of streamflow at the gaging station, or about 1,200 acre-ft/yr. Discharge to springs at Burriston Ponds is included in the streamflow estimated for Currant Creek upstream from gaging station 10146400 (see "Streams" section of this report).

Streamflow data for Currant Creek near Mona (U.S. Geological Survey gaging station 10146400) for water years 1990-93 was used to estimate ground-water

discharge to springs and seeps in the area. The 1990-93 average monthly streamflow in November, December, and January was about 9 ft^3/s. Streamflow during this period is primarily base flow and represents ground-water discharge to springs and seeps. Average monthly streamflow increased in February and was about 24 ft^3/s in March. The increased flow results from both valley and mountain snowmelt runoff. The 1990-93 average monthly streamflow in July, August, and September (the period of peak ground-water pumpage and evapotranspiration) ranged from about 5 to 6 ft^3/s. If undisturbed base-flow conditions are assumed to apply from October through May (9 ft^3/s for 243 days is 4,340 acre-ft) and pumping-influenced base-flow conditions last from June through September (5.5 ft^3/s for 122 days is 1,330 acre-ft), then the average ground-water discharge component of flow at gaging station 10146400 is about 5,700 acre-ft/yr. This rate is similar to that estimated for 1965-66 (see "Streams" section of this report) and probably represents average annual ground-water discharge to Currant Creek upstream of the gaging station from springs and seeps.

Ground-water discharge to springs and seeps at Mona Reservoir was estimated using water-budget information presented by Bjorklund (1967, p. 42-43). The outflow from Mona Reservoir, recorded for water years 1954-60 at U.S. Geological Survey gaging station 10146500, averaged about 15,800 acre-ft/yr. Direct precipitation to the reservoir was about 750 acre-ft/yr, and evaporation from the reservoir was about 3,000 acre-ft/yr (see "Reservoirs" section of this report). Flowing wells along the east side of Mona Reservoir were estimated to discharge about 2,000 acre-ft/yr into the reservoir during this period, and Currant Creek contributed about 8,700 acre-ft/yr (see "Streams" section of this report). An estimated 1,500 acre-ft/yr of water entered the reservoir from local snowmelt runoff, on the basis of 1990-93 streamflow measurements made at the Currant Creek near Mona gaging station. As a result, discharge to springs and seeps at Mona Reservoir is estimated to be 5,800 acre-ft/yr (table 2).

Base flow to West Creek consists mostly of spring and seep discharge from a shallow part of the ground-water flow system and is estimated to be about 500 acre-ft/yr. This flow is included in the streamflow estimate for Currant Creek upstream from gaging station 10146400 (see "Streams" section of this report).

In the southern part of Juab Valley, springs mainly occur northeast of Chicken Creek Reservoir, in the area of sections 15 and 16, Township 15 South, Range 1 West. The combined discharge to springs and flowing wells in section 16 was measured in a channel near Chicken Creek Reservoir at 1.2 ft^3/s (830 acre-ft/yr) in April 1994. An estimated 310 acre-ft/yr of this amount is attributed to inventoried flowing wells. Discharge to other flowing wells in the area not included in the above measurement is reported to be about 390 acre-ft/yr. Other spring- or seep-fed channels are evident on aerial photographs of the area, but flow was not observed or measured.

Palmer Spring (pl. 1), (C-15-1)16cdb-S1, is a relatively large spring with a distinct orifice south of the main spring area. The spring discharged about 1.0 ft^3/s of water from unconsolidated basin-fill deposits in April 1994, which is equivalent to about 720 acre-ft/yr. The relatively large amount of flow downgradient from an area with relatively little recharge may indicate some consolidation of ground-water recharged from a larger area than springs to the north are recharged from. This could require movement through a greater thickness of unconsolidated basin-fill deposits than springs that discharge downgradient from the Chicken Creek drainage require. Discharge to Palmer Spring flows to Chicken Creek Reservoir.

The amount of ground-water discharge to other springs and seeps in the Chicken Creek Reservoir area that flows to the reservoir was estimated by computing a water budget for the reservoir. About 3,850 acre-ft is estimated to leave the reservoir annually, 2,500 acre-ft as outflow to the Mills area and 1,350 acre-ft by evaporation from the reservoir (see "Reservoirs" section of this report). An estimated 350 acre-ft/yr is added by direct precipitation and about 1,000 acre-ft/yr enters the reservoir from local snowmelt runoff, on the basis of 1990-93 streamflow measurements made at Currant Creek near Mona. The residual amount of water after accounting for discharge to flowing wells (700 acre-ft/yr), Palmer Spring (720 acre-ft/yr), and other measured springflow (520 acre-ft/yr) is about 550 acre-ft/yr. An estimated total of about 1,800 acre-ft/yr discharged to springs and seeps in the southern part of Juab Valley in 1993 (table 2).

Discharge from some springs in the southern part of the valley has decreased since the 1950's, when ground-water withdrawal from wells for irrigation began. The amount of water estimated to be discharged to springs and seeps in 1993 therefore is likely less than it was before ground-water development began. Data are not available to estimate discharge to springs and seeps prior to ground-water development or in 1963.

Evapotranspiration

Ground-water discharge from unconsolidated basin-fill deposits by evapotranspiration occurs where the water table is near enough to land surface to support phreatophyte growth, generally in lower-altitude areas of the valley. A phreatophyte is a plant that depends on water from at or below the water table. The principal phreatophytes in Juab Valley are meadow grasses, salt grass, and greasewood with stands of salt cedar, cottonwood, and willow. About 6,615 acres (4,791 acres in the northern part of the valley and 1,824 acres in the southern part) were mapped in 1988 as wetlands (U.S. Department of Agriculture, Soil Conservation Service, written commun., 1993) (pl. 1). Evapotranspiration in irrigated areas is accounted for in the estimated amount of water consumed by the crops and evaporated from irrigated fields.

The average annual evapotranspiration rate for the area north of Chicken Creek Reservoir is 25.5 in. (2.1 ft) and monthly rates range from 0.2 in. in January to 5.6 in. in July (U.S. Department of Agriculture 1969, table 10). Bjorklund (1967, p. 43) used a rate of 30 in/yr (2.5 ft/yr) for the northern part of Juab Valley. At an average annual evapotranspiration rate of 25.5 in., evapotranspiration is estimated to be about 10,200 and about 3,900 acre-ft/yr for the northern and southern parts of Juab Valley, respectively (table 2). Discharge by evapotranspiration has decreased in the valley as pumpage from wells has increased in order for the ground-water system to approach a new state of equilibrium.

Subsurface Outflow

Some ground water probably moves out of Juab Valley through unconsolidated basin-fill deposits in Currant Creek Canyon, north of Mona Reservoir, and through unconsolidated deposits south of Chicken Creek Reservoir. The amount of subsurface outflow through these unconsolidated basin-fill deposits is probably minor compared with other estimated discharge components because the deposits in these locations are thin and narrow. No data regarding subsurface outflow to consolidated rocks in the southern part of Juab Valley were collected during this study, but subsurface outflow is believed to be small relative to other components of the ground-water budget.

Subsurface outflow from the northwest side of the valley to the consolidated rocks of Long Ridge is indicated by water levels measured in the area (fig. 8) and by analysis of the ground-water budget for unconsolidated basin-fill deposits in the Goshen area (Brooks and Stolp, 1995). The amount of subsurface outflow from the northwest part of Juab Valley was estimated using Darcy's Law, which may be expressed as:

$$Q = T\,I\,L \qquad (2)$$

where

- Q = subsurface outflow from the ground-water system, in ft^3/d;
- T = transmissivity, in ft^2/d;
- I = horizontal hydraulic gradient, dimensionless; and
- L = horizontal length of the contact between consolidated rocks and unconsolidated basin-fill deposits, in feet.

The length of contact where subsurface outflow was estimated to occur extends from the north end of Mona Reservoir to the north end of Juab Valley, about 3 mi (15,800 ft). The hydraulic gradient determined from the potentiometric surface of the area (fig. 8) ranges from about 10 ft/0.3 mi (0.006) north of Mona Reservoir to about 10 ft/0.1 mi (0.02) east of Mona Reservoir. The average transmissivity at five wells near the northeast end of and north of Mona Reservoir (fig. 9) is about 1,200 ft^2/d. If five wells east of the reservoir are included (fig. 9), the average transmissivity is about 2,000 ft^2/d. At these transmissivity values, subsurface outflow from the northwest side of Juab Valley to consolidated rocks is estimated to range from about 1,000 to 5,000 acre-ft/yr (table 2).

Hydrologic Properties

Transmissivity values for unconsolidated basin-fill deposits at 76 wells in Juab Valley were estimated from specific-capacity values using a method described by Theis and others (1963). The specific capacity of a well is the ratio of its discharge to its total drawdown (static water level minus pumping water level). Specific-capacity values for pumped wells were determined from information listed on drillers' logs and from data collected by Bjorklund (1967). Transmissivity values estimated from specific-capacity values ranged from 40 to 80,000 ft^2/d (fig. 9). The largest values are generally near the east side of the valley where coarse-grained material has been deposited and sorted by mountain-front streams.

Transmissivity values estimated from specific-capacity values are typically representative of relatively permeable layers in the subsurface because wells drilled for production purposes are commonly finished

in coarser grained, more productive zones and many of the smaller-diameter wells penetrate only a small part of the saturated deposits. Extrapolation of data to the entire saturated thickness may result in an overestimation of transmissivity. Well efficiency affects specific capacity and generally results in an underestimation of transmissivity (Theis and others, 1963). Results of aquifer tests near Nephi indicate a much higher transmissivity value than the values determined from specific-capacity data. However, transmissivity values determined from an aquifer test and from specific-capacity data for wells southwest of Levan are similar. Generally, aquifer tests are considered more reliable for estimating transmissivity.

Hydraulic conductivity for coarser-grained intervals of the unconsolidated basin-fill deposits was calculated by dividing the estimated transmissivity values by the total open interval of each respective well. Values of hydraulic conductivity derived from these data ranged from 0.60 ft/d at well (D-11-1)29cba-2 to 1,200 ft/d at well (D-12-1)31aab-1. The average value was 100 ft/d.

Slug tests were done on four wells in Juab Valley to estimate the hydraulic conductivity and transmissivity of the unconsolidated basin-fill deposits. Slug-test data analyzed with the method described by Cooper, Bredehoeft, and Papadopulos (1967) were used to estimate transmissivity values for the ground-water system of about 12,400, 150, 3,550, and 2,400 ft^2/d at wells (D-12-1)5bab-1, (D-14-1)6dbb-1, (C-13-1)12acc-1, and (C-15-1)17bbb-1, respectively (fig. 9). Data for two of these wells were analyzed using the Bouwer and Rice method (Bouwer and Rice, 1976, and Bouwer, 1989) and resulted in estimated hydraulic-conductivity values of 2.4 and 53 ft/d at (D-14-1)6dbb-1 and (C-15-1)17bbb-1, respectively.

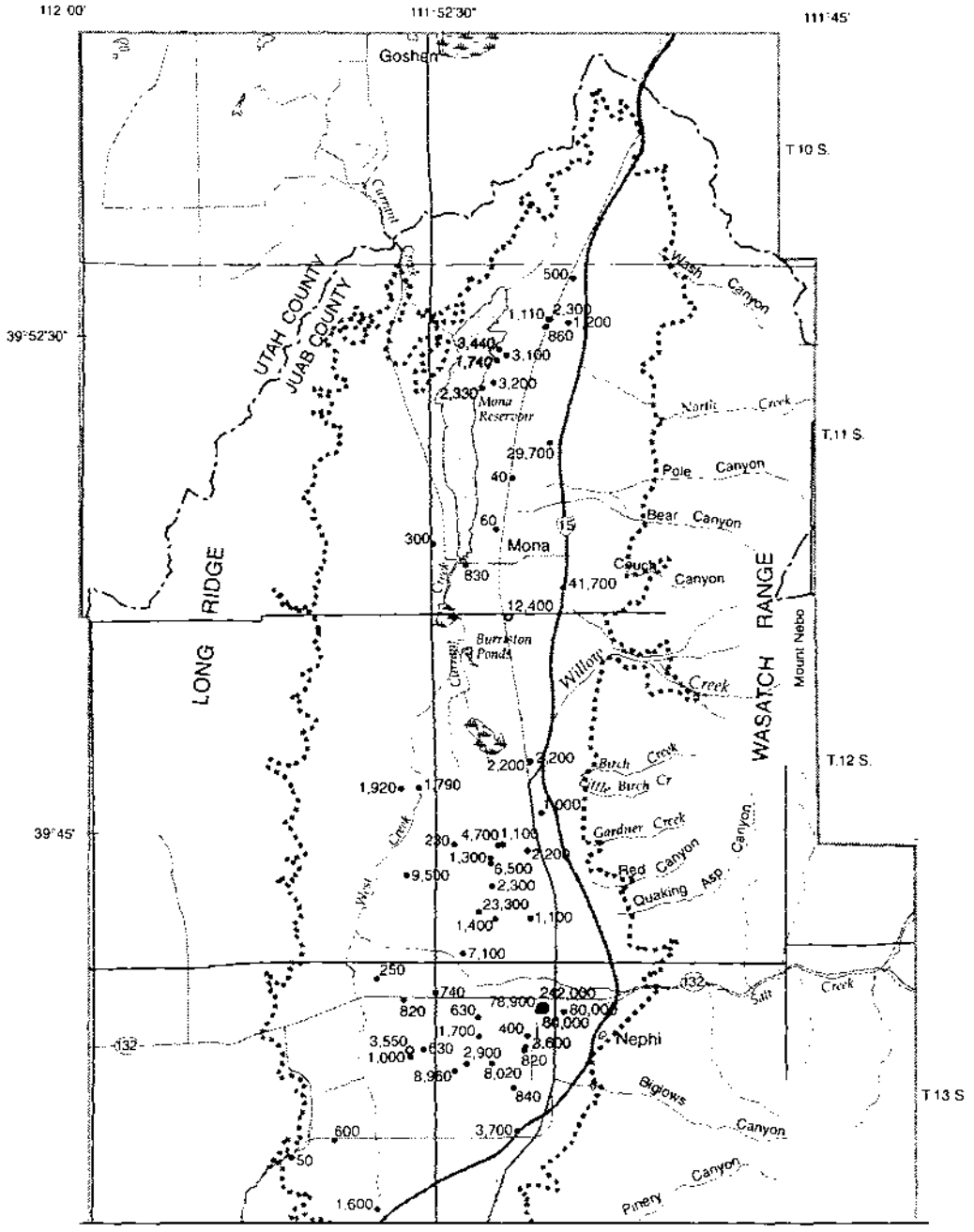
The slug-test method has limitations in estimating the hydraulic conductivity and transmissivity of a ground-water system. The value determined is representative only of the saturated material near the open interval of the well being tested and is influenced by the grain size of the material in the disturbed area around the well casing. Fine-grained material around the open intervals of the well can impede the flow of water into and out of the well if it has not been properly completed and developed. The thickness of the developed zone is typically unknown. Both the Cooper, Bredehoeft, and Papadopulos (1967) and the Bouwer and Rice (1976) methods assume that the ground-water system is isotropic and that leakage from above or below the developed zone is not occurring.

Values of transmissivity and storage coefficient for the unconsolidated basin-fill deposits were estimated from multiple-well aquifer tests at sites in Nephi and southwest of Levan (fig. 9). Water-level changes measured in wells (D-13-1)5dda-1 and 5ddb-2 in response to pumping well (D-13-1)5ddb-3, on the west side of Nephi, were analyzed using the modified Hantush method (Hantush, 1960). Transmissivity and storage-coefficient values determined from the test were about 242,000 ft^2/d and 8×10^{-5}, respectively. Bjorklund (1967, p. 36) estimated the transmissivity in this general area to be about 200,000 ft^2/d on the basis of an aquifer test. The unconsolidated basin-fill deposits in this area probably have been deposited and sorted by Salt Creek, resulting in a large transmissivity relative to the rest of the valley, and the ground-water system likely is unconfined.

In the southern part of Juab Valley, an aquifer test was done by pumping well (C-15-1)10acc-1 and monitoring water-level changes in nearby wells. The ground-water system in this area is confined and consists of interbedded fine- and coarse-grained unconsolidated basin-fill deposits. The transmissivity of the ground-water system, calculated using water-level data from the pumped well and the straight-line method described by Cooper and Jacob (1946), is about 2,300 ft^2/d. Water levels measured in four observation wells at distances ranging from 1,480 to 2,214 ft from the pumped well were analyzed using the modified Hantush method (Hantush, 1960). A transmissivity value of about 4,000 ft^2/d and a storage-coefficient value ranging from about 5×10^{-5} to 5×10^{-4} were computed using this method.

Leakage from confining layers in the unconsolidated basin-fill deposits to the pumped interval is indicated by the water-level response measured in the observation wells. The vertical hydraulic conductivity of confining layers adjacent to the pumped interval is estimated at about 2 ft/d using the modified Hantush method (Hantush, 1960). Data were not available to determine vertical hydraulic-conductivity values of the unconsolidated basin-fill deposits in other areas of Juab Valley, but clay layers throughout the valley impede vertical ground-water flow relative to horizontal flow. Reported ratios of horizontal to vertical hydraulic conductivity for heterogeneous, anisotropic material can be on the order of 100:1 or larger (Freeze and Cherry, 1979, p. 34).

Specific-yield (storativity under unconfined conditions) of the unconsolidated basin-fill deposits was estimated by comparing a compilation of specific yields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Mount Nebo
Mona
Mona Reservoir
Burraston Ponds
Nephi
Willow Creek
Birch Creek
Little Birch Cr
Gardner Creek
Red Canyon
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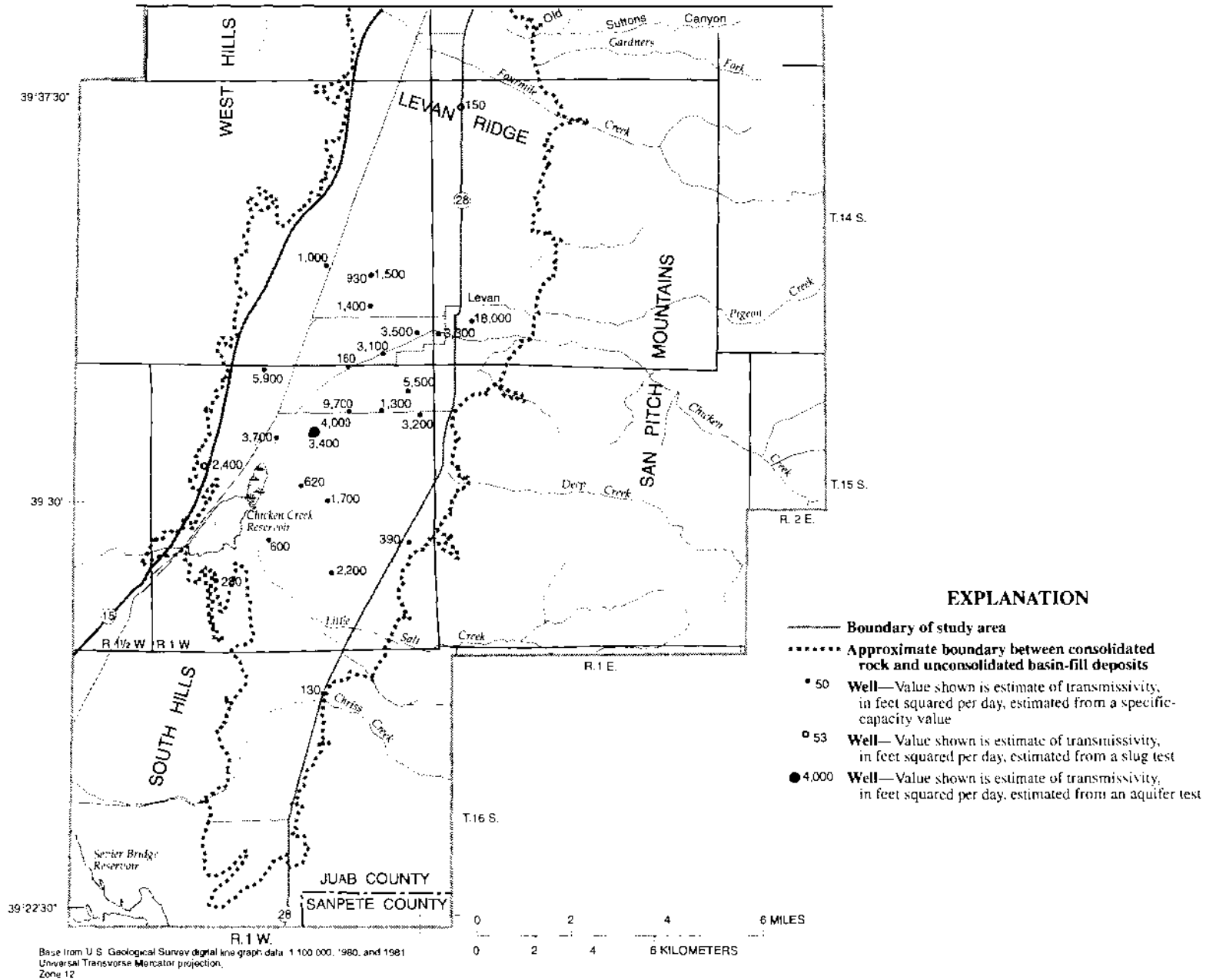


Figure 9. Areal distribution of transmissivity estimated from specific-capacity values, slug tests, and aquifer tests for unconsolidated basin-fill deposits in Juab Valley, Utah.

for various materials in unconfined aquifers (Johnson, 1967, table 29) and descriptions of materials reported in drillers' logs. Values of specific yield in the valley ranged from about 0.03 for clay to 0.25 for gravel. The storage coefficient (storativity under confined conditions) of the unconsolidated basin-fill deposits is estimated to range from 5×10^{-5} to 5×10^{-4} on the basis of aquifer tests completed in the valley and typically ranges from 5×10^{-5} to 5×10^{-3} (Freeze and Cherry, 1979, p. 60).

Water-Level Fluctuations

Water levels were measured in 71 wells in Juab Valley during 1962-66 and 1993 and in 122 wells in March and September 1993 (Steiger, 1995, table 3) to determine long-term and seasonal fluctuations. Measurements also are available for selected wells from the 1930's. Water-level fluctuations measured in most of the wells resulted from variations in precipitation in the valley and surrounding mountains.

Long-Term Fluctuations

Long-term water-level fluctuations generally follow climatic trends for the area. The relation between water levels in selected wells measured in March or April 1942-94 to the cumulative departure of the annual maximum water content of snow from the 1942-94 average at the Payson Ranger Station snow course is shown in figure 10. The annual maximum water content of snow measured at the snow course includes most of the precipitation that falls at the mountain site during the winter and is assumed to be representative of annual precipitation in the study area. Water levels measured in March or April, before ground-water pumpage from wells begins for the year, are largely affected by precipitation that occurs the previous winter.

Greater-than-average precipitation during 1980-87 corresponds with a rise in water levels measured in most wells in the valley and the highest water level measured in some wells. Less-than-average precipitation during 1988-91 corresponds with a decline in water levels measured during 1988-93 in most wells. In many wells, the lowest measured water level on record is in 1992 or 1993. This water-level decline also is affected by increased ground-water pumpage for irrigation to supplement a decrease in the surface-water supply. Water levels measured in well (D-11-1)17cdd-1, north of Mona, fluctuated almost 47 ft from a high of 19.94 ft below land surface in September 1984 to a low of 66.88 ft below land surface in September 1992 (data available for 1975-94). Water levels measured in well (D-13-1)4ccb-2 in Nephi fluctuated more than 61 ft from a high of 112.73 ft below land surface in September 1984 to a low of 174.10 ft below land surface in September 1992 (data available for 1954-93) (fig. 11). Water levels measured in well (D-14-1)30add-1 in Levan fluctuated almost 93 ft from a high of 109.10 ft below land surface in September 1984 to a low of 202.03 ft below land surface in May 1993 (data available for 1963 and 1976-94).

The effects of seasonal pumping for irrigation can be seen on water levels measured in most wells in the valley throughout the period of ground-water development; however, long-term water-level fluctuations reflect climatic trends and a decline caused by increased ground-water withdrawal from wells generally is not apparent. Water levels measured in wells (D-11-1)8add-1, (D-12-1)19cdc-1, (D-12-1)31cca-1, (D-13-1)4ccb-2, (D-13-1)7bbd-2, (C-14-1)25bdd-1, and (C-15-1)11bab-1 (fig. 11) generally are higher in March, before the start of the irrigation season, and lower in September, at the end of the irrigation season. Water levels measured in well (C-14-1)14bad-1 (fig. 11), about 2.5 mi northwest of Levan, do not seem to be influenced by seasonal withdrawals or small changes in annual precipitation, although greater-than-average precipitation in the 1980's does correspond with a steep water-level rise in the well. Well (C-14-1)14bad-1 is in an area with normally little recharge to and discharge from the ground-water system.

Water levels measured in wells in Juab Valley during March-April 1993 generally were lower than levels measured during March-April 1965 (fig. 12). Less-than-average precipitation during 1988-91 and more ground-water discharge to wells in 1992 than the 1963-93 average amount affected water levels measured in 1993 in irrigated areas. Water levels rose from March-April 1965 to March-April 1993 in areas along the axis of the valley downgradient from irrigated fields. An increase in recharge to the ground-water system in irrigated areas from unconsumed irrigation water probably is the cause of this rise.

Seasonal Fluctuations

The seasonal effects of ground-water pumpage and variations in precipitation are evident on water levels from selected wells in the valley (fig. 13). Water levels in wells (C-12-1)24dba-1, (C-13-1)11bbc-2, and (D-14-1)30dcb-1 are generally lowest in August, Sep-

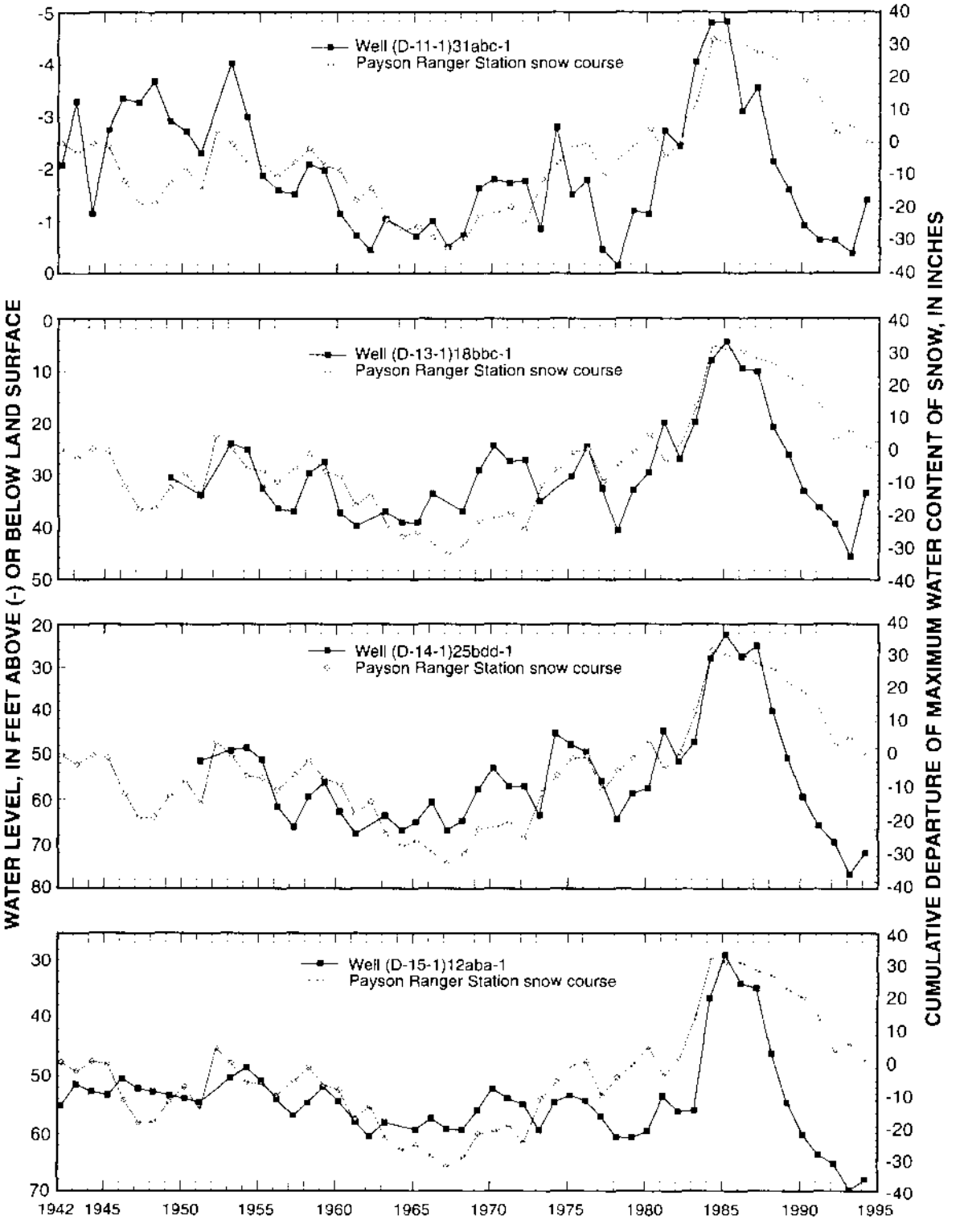


Figure 10. March or April water levels measured during 1942-94 in four wells in Juab Valley, Utah, and cumulative departure of the annual maximum water content of snow from the 1942-94 average at the Payson Ranger Station snow course.

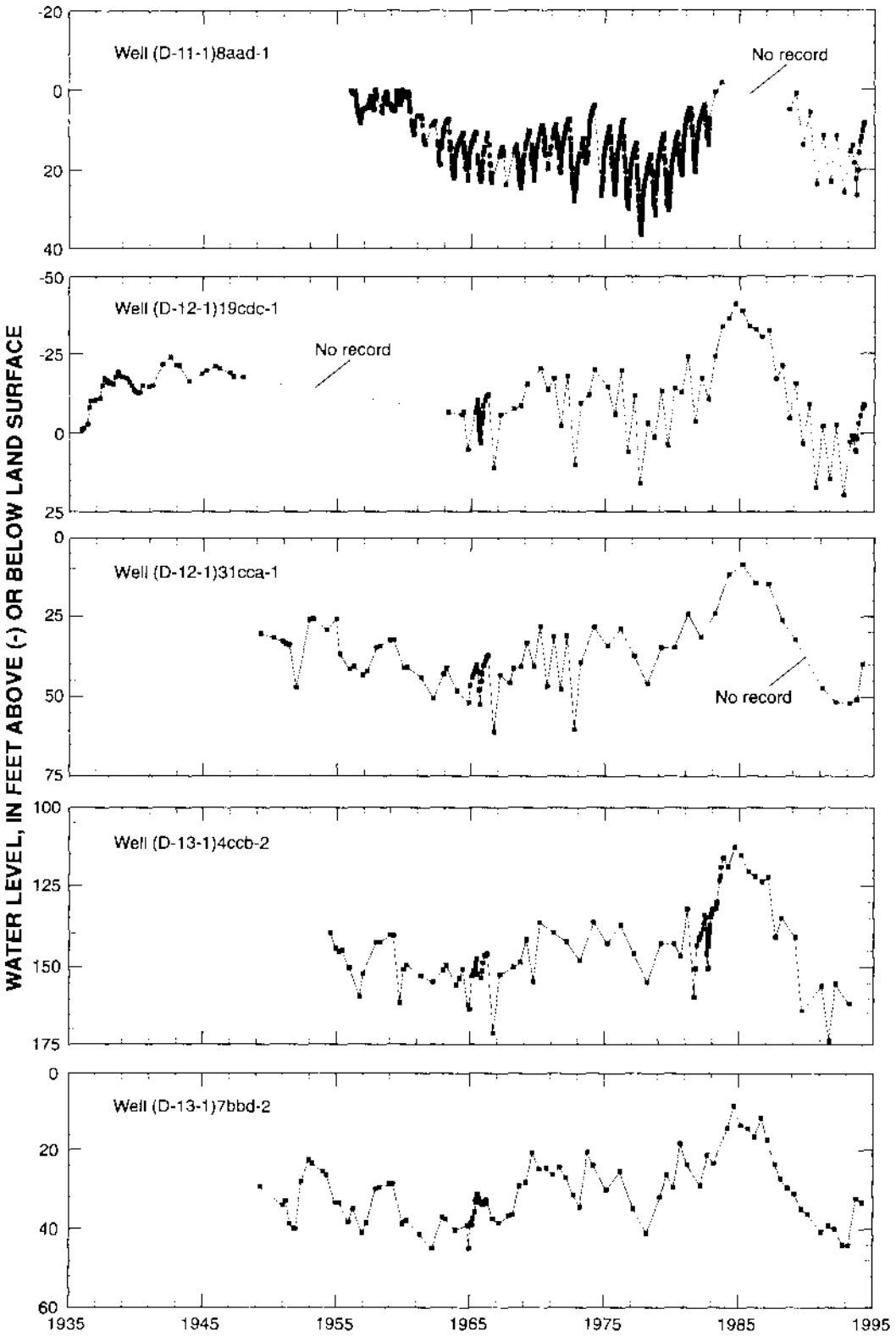


Figure 11. Long-term water-level fluctuations during 1935-94 in eight wells in Juab Valley, Utah.

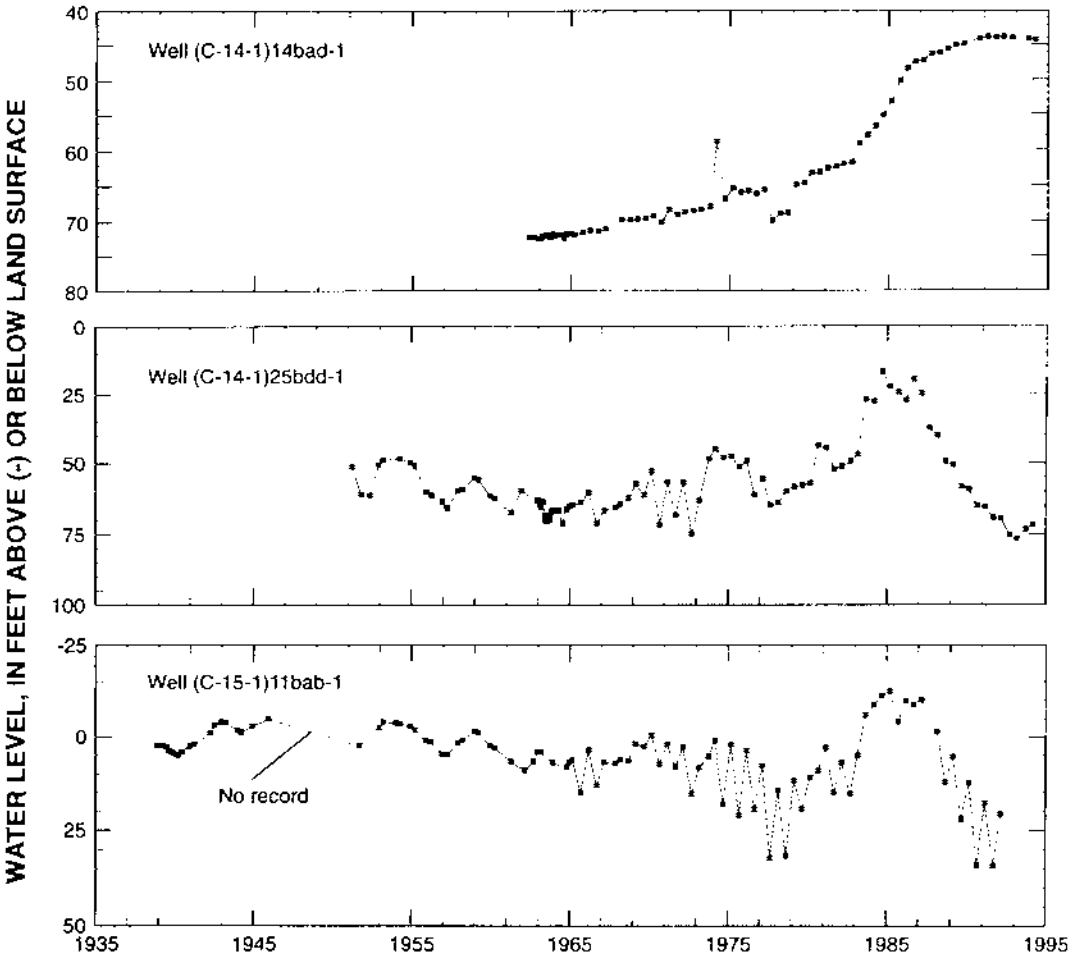
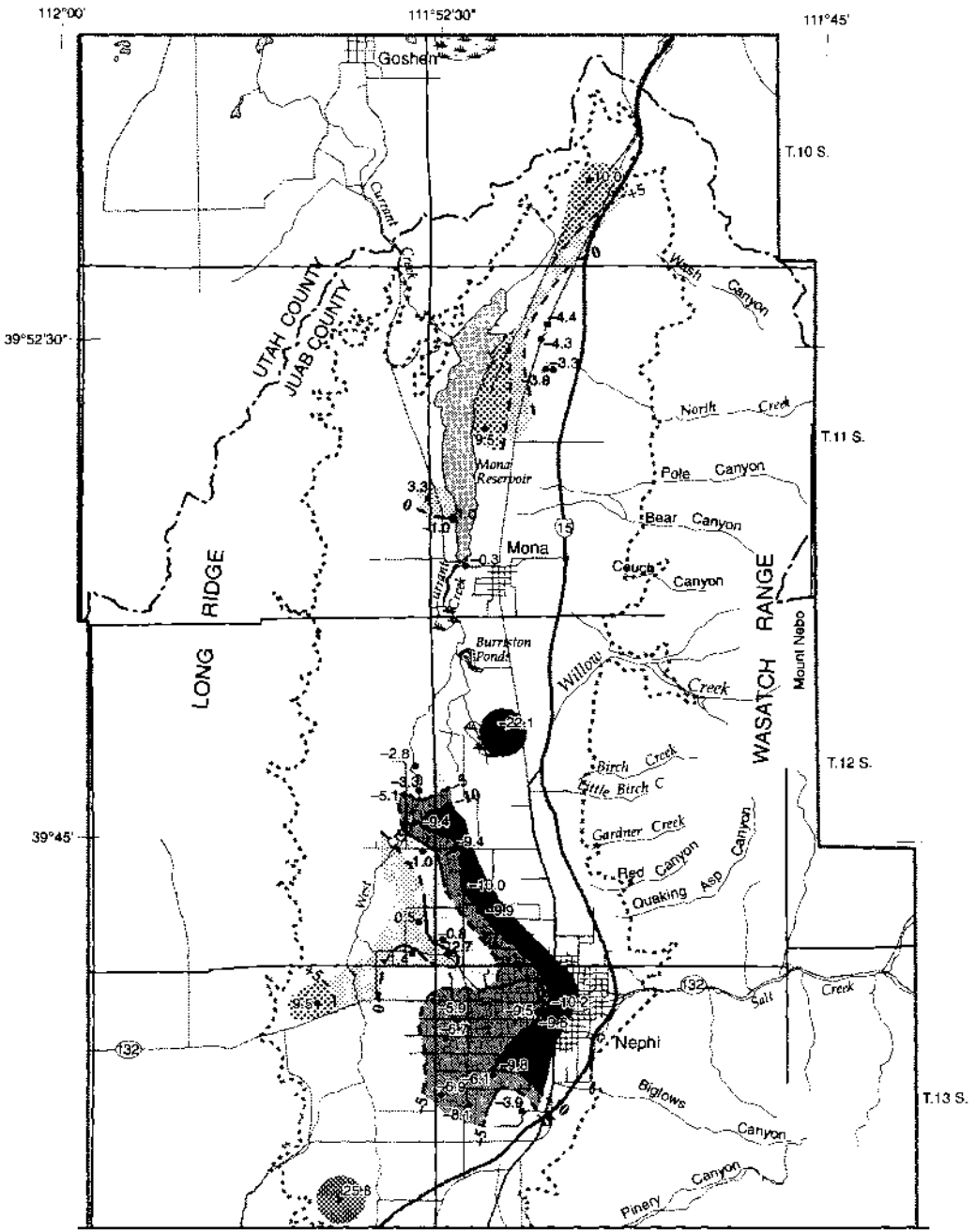


Figure 11. Long-term water-level fluctuations during 1935-94 in eight wells in Juab Valley, Utah—Continued.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LONG RIDGE
WASATCH RANGE
Mount Nebo
Currant Creek
Willow Creek
Birch Creek
Little Birch C
Gardner Creek
Red Canyon
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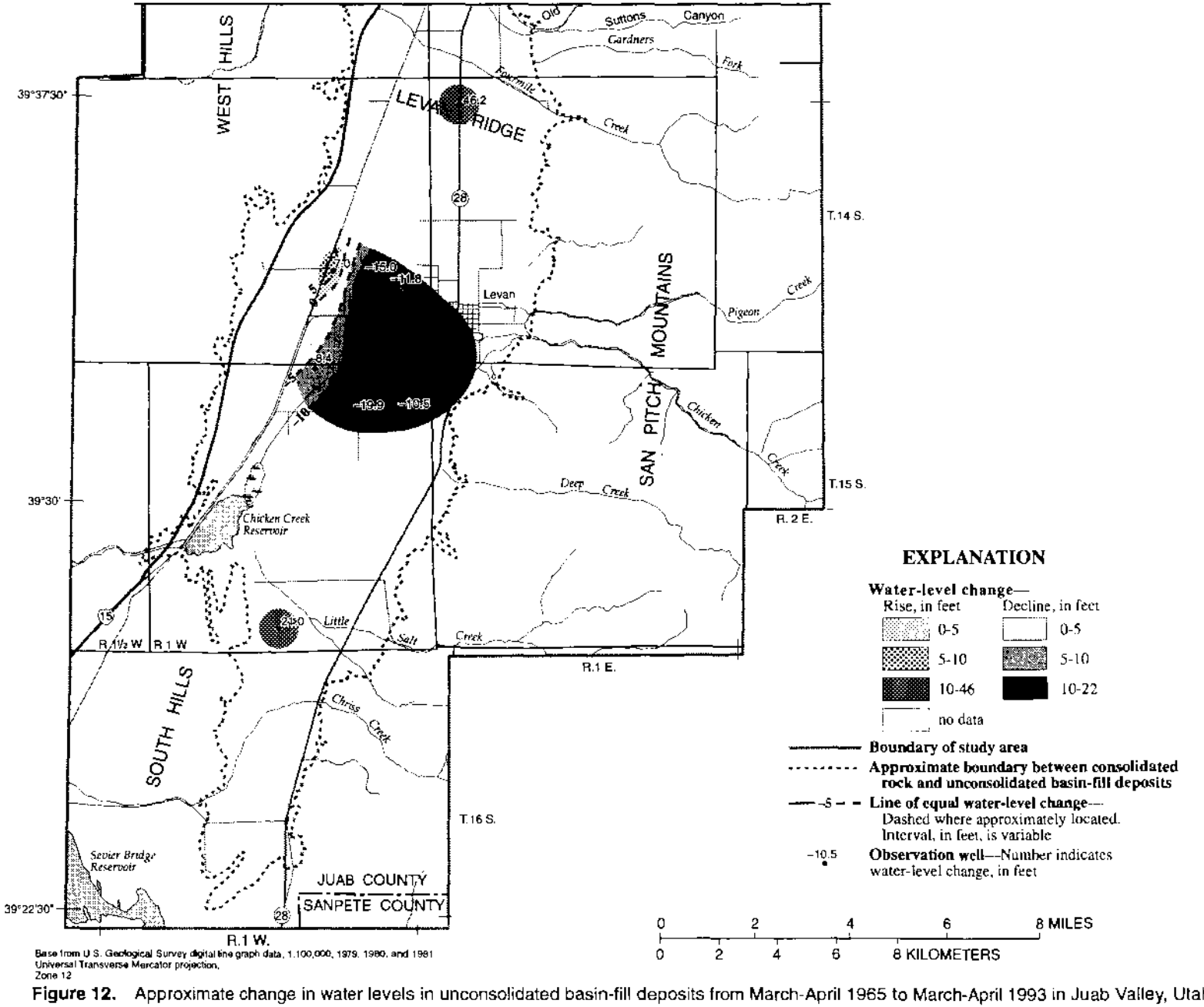

Base from U.S. Geological Survey digital line graph data, 1:100,000, 1979, 1980, and 1981
Universal Transverse Mercator projection,
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Figure 12. Approximate change in water levels in unconsolidated basin-fill deposits from March-April 1965 to March-April 1993 in Juab Valley, Utah.

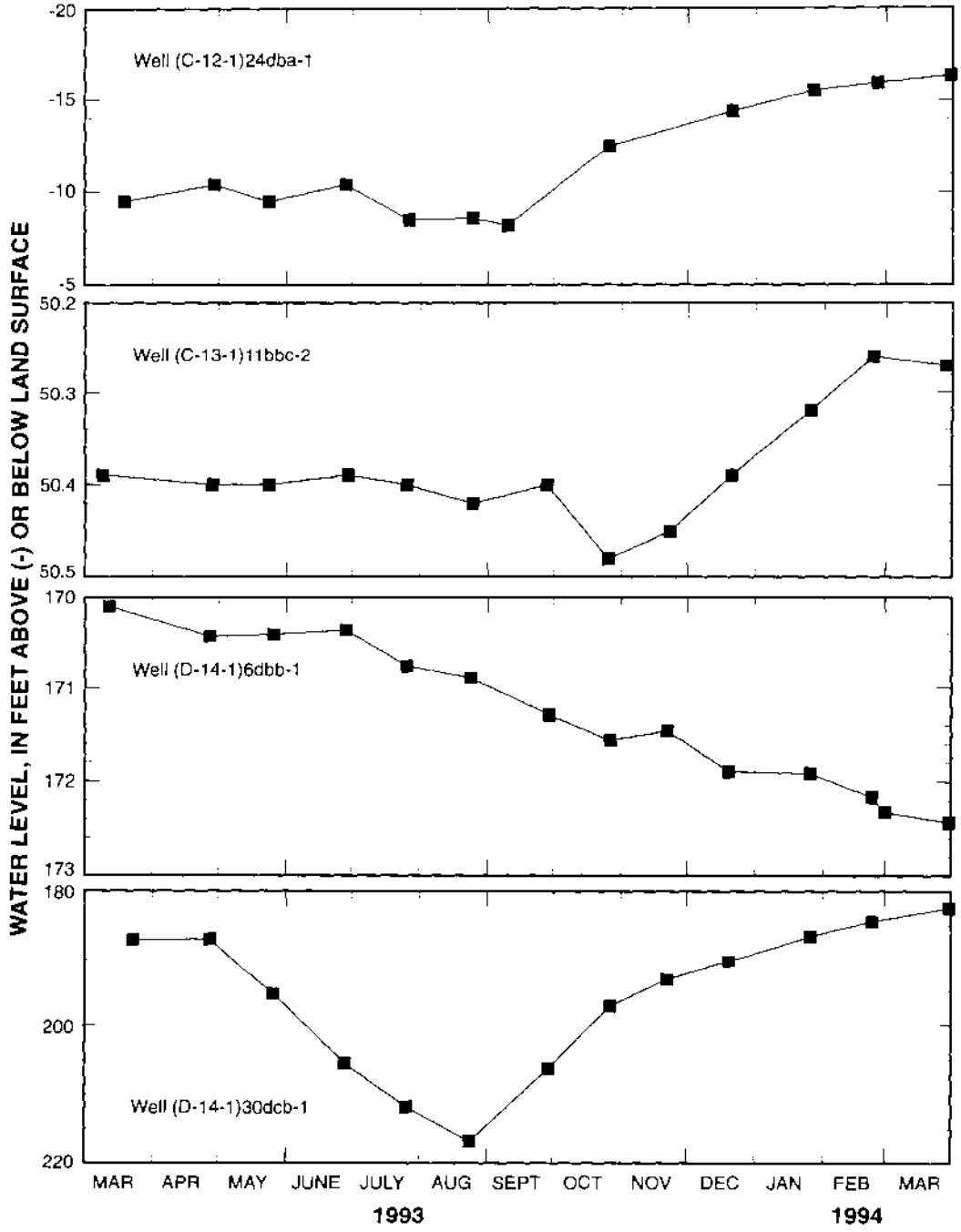


Figure 13. Seasonal water-level fluctuations during 1993-94 in four wells in Juab Valley, Utah.

tember, or October and highest in February and March. Well (C-12-1)24dba-1 is in an area where the ground water is confined and is about 2 mi from any large irrigation wells. Water levels in this well rose after September, when pumpage for irrigation in the northern part of Juab Valley stopped. Water levels measured in well (D-14-1)6dbb-1 showed a steady decline of about 2.4 ft from March 1993 to March 1994 (fig. 13). This well is near the ground-water divide that separates the northern and southern parts of the valley, where no pumping for irrigation takes place. This water-level decline probably was caused by less-than-average precipitation. Water levels in well (D-14-1)30dcb-1 in Levan declined about 30 ft from April to August 1993 (fig. 13), mostly as a result of ground-water pumpage for irrigation. The water levels recovered fully between August 1993 and March 1994.

Comparisons of seasonal water-level fluctuations in wells (D-12-1)19cdc-1, (D-13-1)6cbc-1, and (C-15-1)12aba-1 show the changes in response between 1937-38 (prior to ground-water development in the valley) and periods with ground-water pumpage (1965-66 and (or) 1993-94) (fig. 14). Seasonal water-level response measured in well (D-12-1)19cdc-1 changed from an annual peak in the early summer of 1937 to a probable peak in the spring of 1965 and 1993. Water levels affected by ground-water withdrawals declined to an annual low in August 1993 and in September 1965. Although the pattern of seasonal water-level fluctuation measured in well (D-13-1)6cbc-1 did not change from 1937-38 to 1965-66, water levels did drop from 2 to 4 ft between the two periods. Water levels measured in well (C-15-1)12aba-1 were probably not affected by ground-water development until sometime after 1966. A low water level measured in August 1993 corresponds to peak ground-water pumpage for irrigation in the area. The pattern of seasonal water-level fluctuation measured in well (D-11-1)8aad-1 during 1993-94 is similar to that of 1965-66 (fig. 14).

Chemical Characteristics of Water

Water-chemistry data for samples collected from 88 ground-water sites and 7 surface-water sites in the study area are listed by Steiger (1995, tables 6 and 7). Most of the water samples from these sites were collected during 1963-65 and 1992-94. Analysis of the chemical and isotopic composition of the water and analysis of associated geochemical data was used to better define the chemical characteristics of ground water in the valley and to determine sources of ground-water recharge, directions of ground-water flow, and water-rock interactions occurring in the ground-water system.

Chemical Composition

The chemical composition of water in the valley is controlled primarily by the composition of the rock with which it comes in contact because of water-rock interactions. The chemical composition of ground water in the unconsolidated basin-fill deposits changes from the south to the north primarily because of changes in the type of rock that makes up the basin fill. Ground water that has moved through unconsolidated basin-fill deposits derived from the Arapien Shale, a marine evaporite, along the Chicken and Pigeon Creek drainages contains calcium and sulfate as the predominant ions. Ground water in basin-fill deposits derived from and downgradient from the limestone, sandstone, and quartzite in the southern Wasatch Range has a lower dissolved-solids concentration, with calcium and bicarbonate as the predominant ions in ground water from unconsolidated basin-fill deposits in Juab Valley north of Burriston Ponds. These rocks are more resistant to weathering; therefore, less water-rock interaction occurs. Water sampled from Salt Creek near the boundary between the consolidated rock and unconsolidated basin-fill deposits (U.S. Geological Survey gaging station 10146000) and from well (D-13-1)1cab-1 (finished in the channel-fill deposits of Salt Creek) contains relatively high dissolved-solids concentrations, with sodium and chloride as the predominant ions. The channel-fill deposits are derived partly from the Arapien Shale that crops out in the canyon that separates the southern Wasatch Range from the San Pitch Mountains. Predominant ions in water sampled from Bradley Spring, (D-13-2)5cbd-S1 discharging from the Indianola Group at the northern end of the San Pitch Mountains, and from Clover Creek Spring, (D-12-1)3bbc-S1 discharging from limestone at the base of the southern Wasatch Range, are calcium and bicarbonate. Water that has reacted with the channel-fill deposits of Salt Creek, with rocks of the Indianola Group, and with limestone in the southern Wasatch Range may mix to form the calcium, sodium, chloride, and bicarbonate water type sampled from wells completed in unconsolidated basin-fill deposits in the Nephi area of the valley.

The predominant ions in water that discharges from Government Spring, (C-13-1)3bbc-S1 on Long Ridge, and from Orme Spring, (C-13-1)33cac-S1 in the

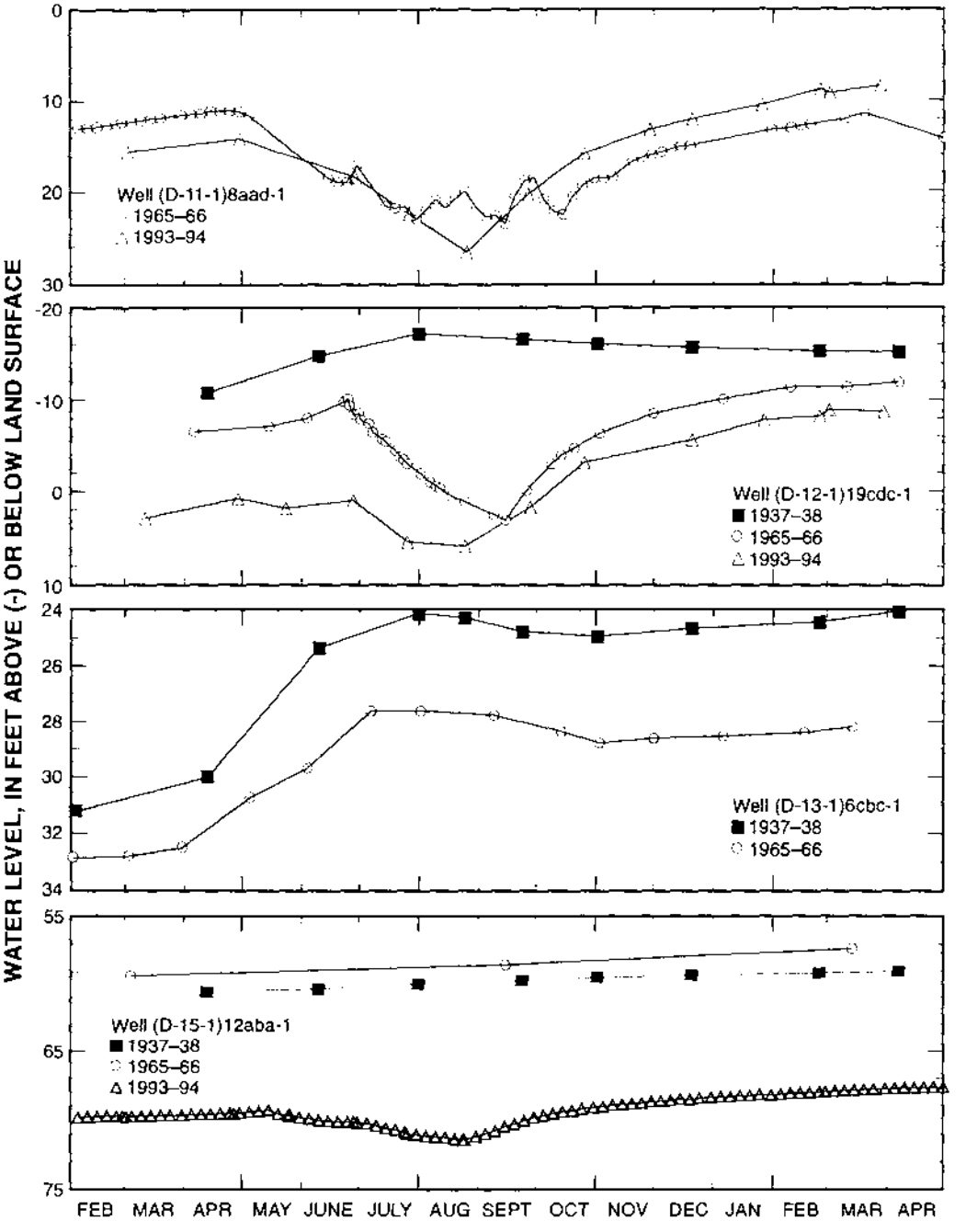


Figure 14. Seasonal water-level fluctuations during 1937-38, 1965-66, and (or) 1993-94 in four wells in Juab Valley, Utah.

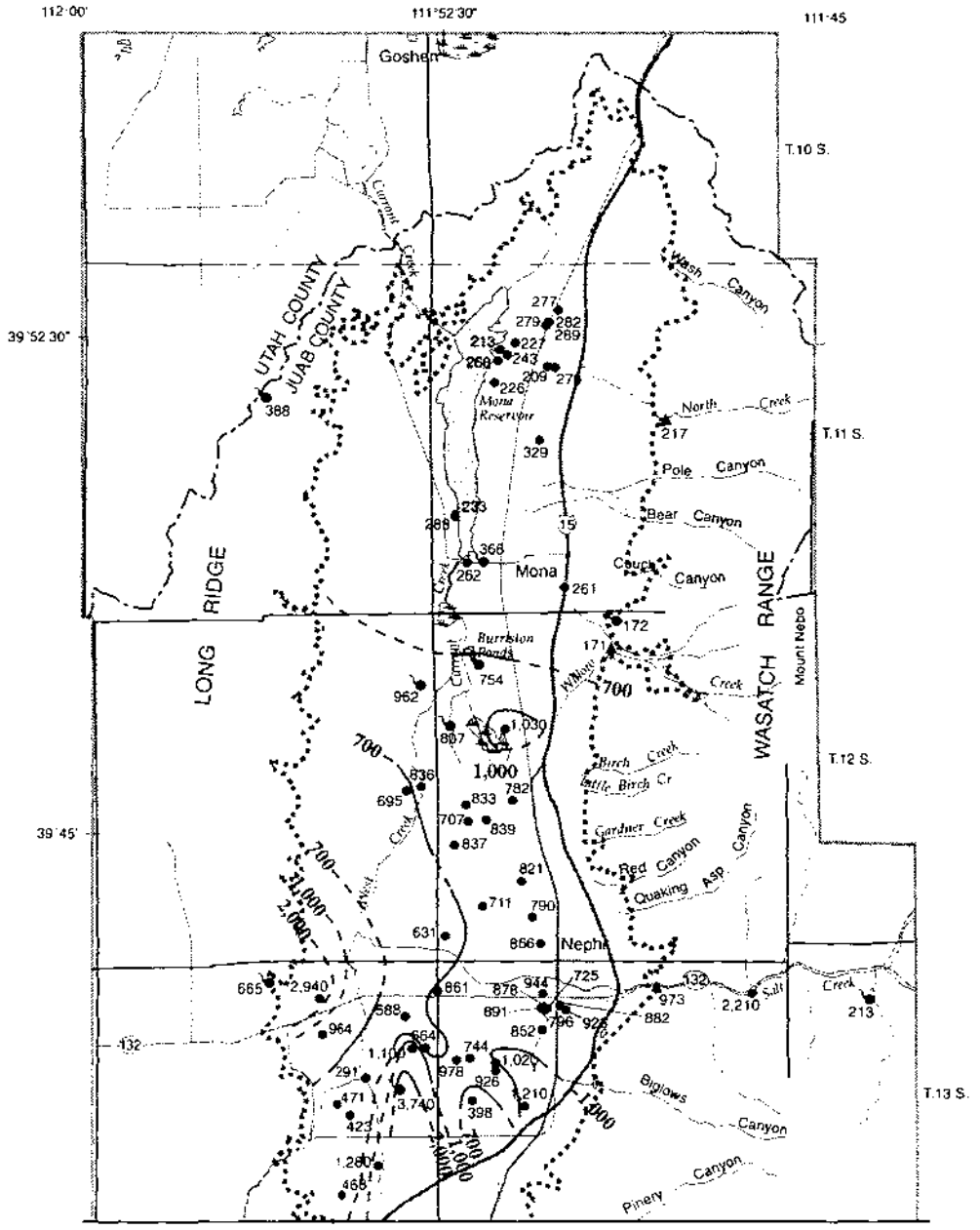
West Hills, are calcium and chloride. These springs discharge from volcaniclastic rocks of Tertiary age. Predominant ions in ground water in unconsolidated basin-fill deposits in Juab Valley downgradient from Orme Spring generally are magnesium and bicarbonate. Precipitation is probably a primary source of recharge to the valley in this area because water from the volcaniclastic rocks to the west and from the unconsolidated basin-fill deposits to the east are more mineralized than water from this area and do not include magnesium as a dominant ion. The unconsolidated basin-fill deposits in this area are probably derived from the North Horn Formation, which crops out below Orme Spring but has limited exposure in other parts of the study area. Minerals derived from the North Horn Formation might be the source of magnesium in the ground water in the unconsolidated basin-fill deposits.

Dissolved-solids concentrations in ground water sampled from unconsolidated basin-fill deposits in the northern part of Juab Valley during 1992-93 ranged from 261 mg/L at well (D-11-1)33cab-1 to 3,740 mg/L at (C-13-1)13bac-S1 (fig. 15). Concentrations in ground water in the valley north of Burriston Ponds do not exceed 400 mg/L. Dissolved-solids concentrations in ground water sampled from the southern part of the valley during this period ranged from 623 mg/L at well (D-14-1)30add-1 to 3,980 mg/L at well (C-14-1)22ddc-1 (fig. 15). A lobe of ground water with dissolved-solids concentrations less than 1,000 mg/L extends toward Chicken Creek Reservoir from about where Chicken Creek enters the valley. Dissolved-solids concentration in surface water sampled during this study ranged from 141 mg/L for Chicken Creek at (D-15-1)11aba to 1,340 mg/L for Currant Creek near Mona (U.S. Geological Survey gaging station 10146400).

Chicken and Pigeon Creeks are two of the largest perennial streams in the southern part of Juab Valley and are primary sources of recharge to the ground-water system in that area (pl. 1). The unconsolidated basin-fill deposits along Chicken Creek downgradient from the mountain front are probably coarser grained and better sorted than in other areas. High transmissivity and, therefore, more ground-water flow through the unconsolidated basin-fill deposits along this flow path would result in the more rapid removal of soluble salts, such as halite, throughout time. Material along the probable flow path for water that discharges from Palmer Spring, south of the Chicken Creek flow path, is probably not as transmissive and the recharge not as large as what occurs along the Chicken Creek flow path. The high dissolved-solids concentration in water from Palmer Spring likely is caused by a longer contact time between the water and the unconsolidated basin-fill deposits and the presence of soluble salts in the deposits.

Ground water sampled from eight sites in the northern part and five sites in the southern part of Juab Valley was analyzed for dissolved-solids concentration during 1963-65 and 1992-93. A change in dissolved-solids concentration of more than 20 percent between these periods was measured in water sampled from wells (D-13-1)4cca-1 and (C-16-1)3cdd-1. Well (D-13-1)4cca-1 was drilled in 1963 and has been pumped for irrigation water every summer since then, except for 1983 and 1984. The increase in dissolved-solids concentration may be a result of ground water flow from different areas to the well in 1993 as compared to 1964. The increase measured at well (C-16-1)3cdd-1 likely is the result of the well being deepened in 1974 from 352 to 405 ft. The ground water at depth had a higher dissolved-solids concentration.

Specific conductance provides an indication of the concentration of dissolved ions in water. Specific-conductance values measured during this study for water from selected ground- and surface-water sites are listed by Steiger (1995, tables 1, 5, 6, 7, 9, and 10). Water from Willow Creek and Clover Creek Spring in the southern Wasatch Range and from wells completed in unconsolidated basin-fill deposits east of Burriston Ponds had a lower specific-conductance value than water from wells south of the ponds. Spring water discharging to Burriston Ponds had specific-conductance values similar to those of ground water sampled from the east and from the south. Water measured from the outlet for the northern set of ponds in March 1966 had a specific-conductance value of 420 µS/cm. Water from the east side of the northernmost pond had a specific-conductance value of 455 µS/cm on September 13, 1994. These values are similar to those of water sampled to the east. Specific conductance measured at the outlet for the southern set of ponds was 985 µS/cm in March 1966 and averaged about 1,010 µS/cm for monthly measurements made from July 1993 to March 1994. The spring at the southeast corner of the Burriston Ponds complex, (D-12-1)6ddc-S1 (pl. 1), had a specific conductance of 1,450 µS/cm in March 1966. The specific conductance of water from the spring on June 24, 1993 was 1,440 µS/cm. These values are similar to those of ground water sampled to the south along the hypothesized flow path originating near Salt Creek (fig. 8).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Burrston Ponds
Nephi
Wash Canyon
North Creek
Pole Canyon
Bear Canyon
Canyon
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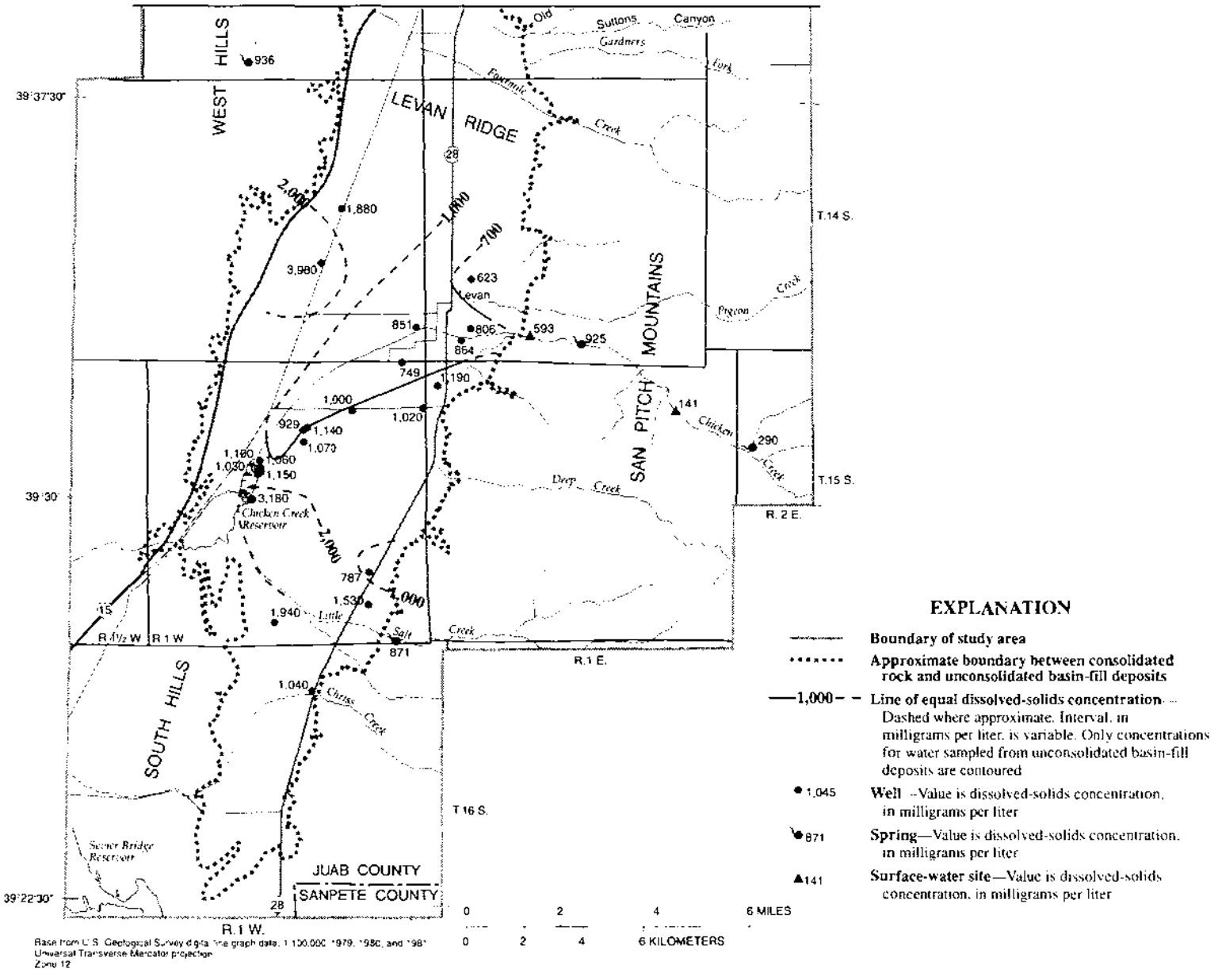


Figure 15. Dissolved-solids concentration in water from the Juab Valley study area, Utah, 1951-94.

Analysis of streamflow and specific-conductance data collected during the Currant Creek seepage study in March 1966 (Bjorklund, 1967, p. 42) indicates that flow in Currant Creek (2.8 ft^3/s), upstream from gaging station 10146400 and not attributable to outflow from Burriston Ponds, had a calculated specific-conductance value of about 3,880 µS/cm on the basis of known streamflow and specific-conductance values. Much of this water is thought to be diverted from West Creek on the basis of observations made during this study (see "Streams" section of this report). This calculated specific conductance is similar to measured values of 4,480 and 4,610 µS/cm for water diverted from West Creek to ditches in the area of (D-12-1)18ca (fig. 6, site W-8), south of Burriston Ponds.

The specific conductance of water in West Creek at (C-13-1)2dbd (fig. 6, site W-4) in December 1993 was 9,500 µS/cm. This relatively large value might be the result of the dissolution of salts that were precipitated by evapotranspiration during the summer months or by the inflow of ground water with high concentrations of dissolved ions to West Creek. Dissolved-solids concentrations in ground water sampled from the unconsolidated basin-fill deposits in this part of the valley generally are an order of magnitude less than those of water from West Creek. Specific conductance of water in West Creek at (C-12-1)24cdd (fig. 6, site W-7) from April 1993 to March 1994 ranged from 1,000 µS/cm in May 1993 (diluted by inflow from Big Hollow) to 7,100 µS/cm in April 1993. The higher specific-conductance value measured during the early spring, before snowmelt runoff peaks, might be caused by the dissolution of surface salts deposited during the previous summer by capillary evaporation of ground water and evaporation of surface water or from ground-water inflow with high dissolved-solids concentrations.

Water sampled from spring (C-13-1)13bac-S1 and from a cement pipe at (C-13-1)12ccd (fig. 6, site I-3a) had specific-conductance values of 4,600 and 3,000 µS/cm, respectively. These sites are west of Nephi near West Creek, downgradient from a large irrigated area where seepage to the ground-water system has been measured (see "Distribution systems" section of this report). The range in specific-conductance values of surface water diverted for irrigation and ground water sampled from wells in the area is from about 1,000 to 2,000 µS/cm. The relatively high values of specific conductance for water from (C-13-1)13bac-S1 and 12ccd might be caused by the circulation of irrigation water through a shallow part of the ground-water system (see "Springs and seeps" section of this report) that results in evapoconcentration and the dissolution of surface and near-surface salts.

Analysis of streamflow measured at U.S. Geological Survey gaging stations on Salt and Chicken Creeks, which enter Juab Valley at the mountain front, and Currant Creek, which begins in the valley, shows different relations between streamflow and specific conductance of the water (fig. 16). At Salt and Chicken Creeks, specific conductance of the water generally decreased as streamflow increased, primarily because of the diluting effect of less-mineralized runoff that is added to the base flow in the stream. At Currant Creek, specific conductance of the water generally increased as streamflow increased. At flow rates less than about 10 ft^3/s, the specific-conductance value of water in Currant Creek was about 1,000 µS/cm (fig. 16). At flow rates greater than about 10 ft^3/s, the specific-conductance value of the water ranged from about 1,000 to 3,500 µS/cm, with little correlation between streamflow and specific conductance. This lack of correlation is caused mainly by the mixture of base flow and other sources of inflow to the stream, such as water from West Creek (see "Streams" section of this report). Inflow to Currant Creek other than the base-flow component can have a large range in specific conductance because of differences in water sources and time of year.

Water with relatively high sodium and chloride concentrations was sampled from wells greater than 240 ft deep on the west side of Juab Valley. A chloride concentration of 1,400 mg/L was measured in water from wells (C-13-1)3dad-1 and (C-14-1)22ddc-1. Water temperatures in these wells and of other ground water sampled from the west side of the valley, (C-12-1)12aab-S1, 12aac-S1, and (C-15-1)33acd-1, were greater than 19.0 °C. The average temperature of ground water in the rest of the valley is about 13.0 °C. The warmer water may be caused by upward movement along a north-south trending fault on the west side of the valley.

Immunoassay screen tests for triazine and 2,4-D herbicides were done on water samples collected from 50 sites in Juab Valley. The screen test measures the combined concentration of 13 individual triazine compounds and six individual 2,4-D compounds but does not specify which compounds are present in the water samples. Commonly, the triazine compound measured is atrazine, a season-long weed control used with corn, sorghum, and other crops. Compounds of 2,4-D are used on grasses, wheat, barley, oats, rangeland pasture, turf, and certain other crops for postemergent control of

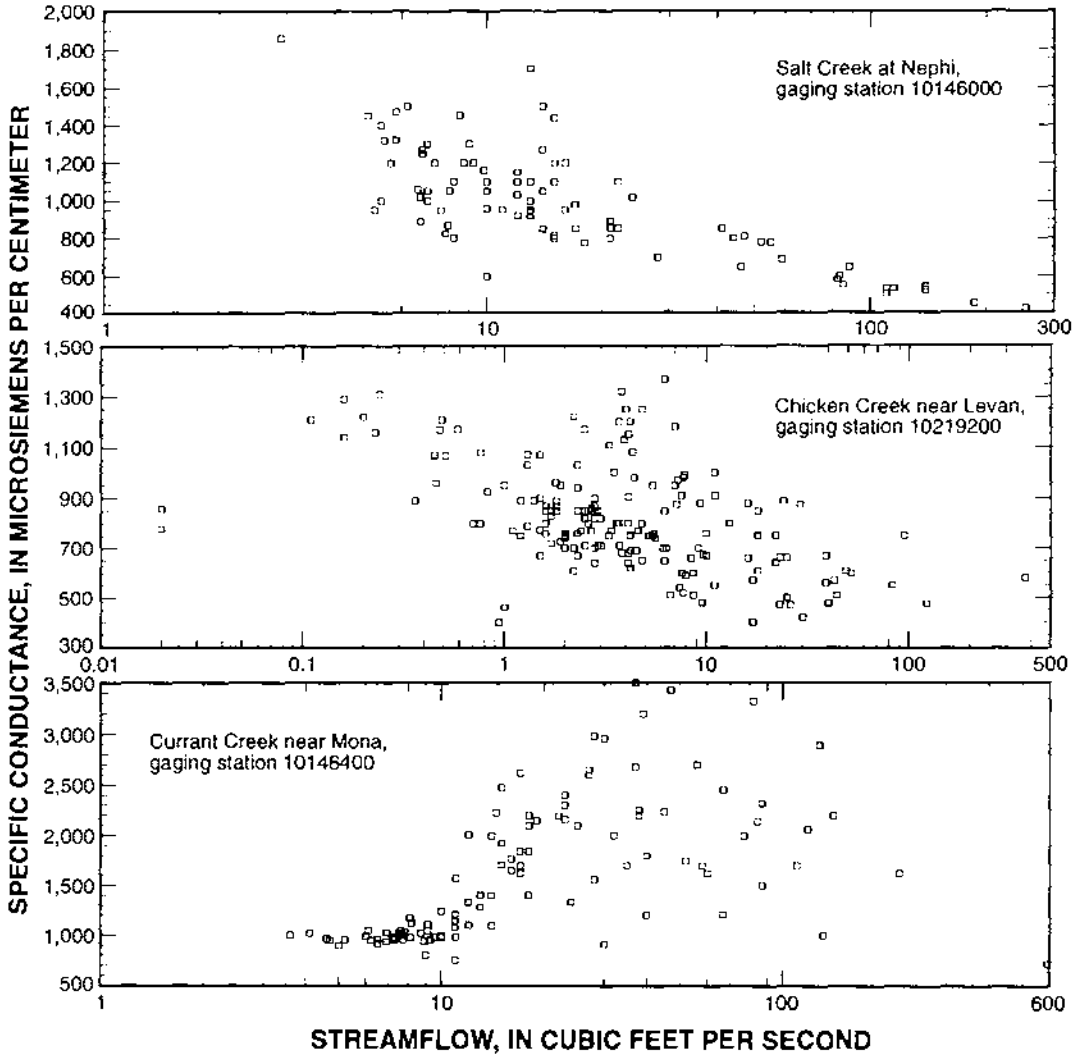


Figure 16. Relation between specific conductance of water and streamflow at gaging stations on Salt Creek, Chicken Creek, and Currant Creek, Juab Valley, Utah.

weeds. The maximum contaminant level for atrazine and 2,4-D in drinking water is 3 and 70 µg/L, respectively (Safe Drinking Water Hotline, U.S. Environmental Protection Agency, oral commun., 1995). Results of the screen tests on water samples from the valley indicate concentrations much lower than the maximum contaminant levels for atrazine and 2,4-D. For triazine, one sample had 1.1 µg/L and the remainder had less than the detection limit of 0.7 µg/L. For 2,4-D, one sample had 0.8 µg/L and the remainder had less than the detection limit of 0.5 µg/L.

Nitrogen concentrations, as nitrate and nitrite, in water sampled from the study area in 1992-94 ranged from less than 0.050 mg/L at spring (C-15-1)36cdc-S1 to 21 mg/L at (C-14-1)14cac-1, a 120-ft-deep stock well. The nitrite species is unstable in aerated water and is seldom present in concentrations high enough to influence ionic balanees. The drinking-water standard for nitrate is a maximum contaminant level of 10 mg/L (Safe Drinking Water Hotline, U.S. Environmental Protection Agency, written commun., 1994). The nitrogen concentration, as nitrate and nitrite, in water from (C-13-1)23add-1, a stock well perforated from 81 to 99 ft below land surface, and from (D-15-1)6cab-1, an irrigation well perforated from 120 to 315 ft, are 13.0 and 14.0 mg/L, respectively. Concentrations, as nitrate and nitrite, in water from four wells in Nephi used for irrigation ranged from 4.9 to 8.4 mg/L. Possible sources of

nitrate to ground water in the valley are the leaching of fertilizers from irrigated land, farm drainage, septic tank drainage, and natural accumulation of nitrate by evaporation during the formation of playas (the bottom of an undrained desert basin).

Selenium concentrations in water generally were less than 10 µg/L in Juab Valley. The drinking-water standard is a maximum contaminant level of 50 µg/L (Safe Drinking Water Hotline, U.S. Environmental Protection Agency, written commun., 1994). Ground water sampled from wells (C-14-1)14cac-1 and 22ddc-1, perforated from 160 to 409 ft below land surface, contained 38 and 62 µg/L of selenium, respectively. Well (C-14-1)22ddc-1 is used for industrial purposes and does not provide drinking water. Ground water sampled from unconsolidated basin-fill deposits in the northern part of the valley had selenium concentrations that ranged from 1 to 6 µg/L, except for well (C-13-1)3dad-1, which had less than 1 µg/L, and spring (C-13-1)13bac-S1 and well (C-13-1)23add-1, which had 9 and 14 µg/L, respectively. Well (C-13-1)23add-1 is perforated from 81 to 99 ft below land surface and is used to water stock. Selenium concentrations in water along a hypothesized flow path in the southern part of Juab Valley (fig. 8), from near Levan to the ground-water discharge area north of Chicken Creek Reservoir, ranged from 1 to 3 µg/L.

Isotopes

The stable isotopes of oxygen (^{18}O) and hydrogen (^{2}H), also called deuterium (D), vary naturally in water from the effects of topography and local climate. Stable isotopes can be used to help determine sources of ground-water recharge because they are conservative in low-temperature ground-water systems and therefore are unaffected by chemical processes. Water samples from 36 hydrologic-data sites in the study area were analyzed for ^{18}O and D (Steiger, 1995, table 8). The hydrogen isotopic ratio ($D/^{1}H$) and the oxygen isotopic ratio ($^{18}O/^{16}O$) in a water sample are reported in delta (δ) units per mil (parts per thousand or 0/00) deviations from a reference standard known as Standard Mean Ocean Water (SMOW). The δ values are determined from the following equation:

$$\delta R = [(R_{sample} - R_{standard})/R_{standard}] \times 1{,}000 \qquad (3)$$

where

δR = δD or $\delta^{18}O$ value for the water sample;

R_{sample} = ratio of D to ^{1}H or ^{18}O to ^{16}O in a water sample; and

$R_{standard}$ = ratio of D to ^{1}H or ^{18}O to ^{16}O in the reference standard.

The global variation of $\delta^{18}O$ and δD in meteoric water that results from changes in variables such as altitude and temperature of an area are shown in the global meteoric water line (Craig, 1961). Values of $\delta^{18}O$ and δD for 41 water samples taken from the study area are plotted in figure 17, along with the global meteoric water line. Because ^{18}O and D abundances decrease with increasing altitude, the altitude of the area where the precipitation occurred affects placement of the data points. Water sampled from two streams that drain the west side of the southern Wasatch Range, Willow and North Creeks, has δD and $\delta^{18}O$ values that plot to the left of the global meteoric water line and most values for water from the San Pitch Mountains and from Juab Valley (fig. 17). This indicates a higher-altitude recharge area relative to that of water sampled from other parts of the study area. Points that represent water that originates as precipitation on the San Pitch Mountains plot to the right of the global meteoric water line (fig. 17) and indicate a lower-altitude recharge area. Water from Salt Creek sampled in April 1994 and ground water sampled along or near the hypothesized flow path downgradient from where Salt Creek enters Juab Valley (fig. 8) has δD and $\delta^{18}O$ values that plot between values for water that originates as precipitation on the southern Wasatch Range and the San Pitch Mountains (fig. 17). This is because Salt Creek drains both the southern Wasatch Range and the San Pitch Mountains.

Water that has undergone evaporation becomes enriched in the heavy isotopes of oxygen and hydrogen and clusters along a trend with less slope than that of the meteoric water line. Evidence of evaporation is observed in the stable isotope data for Salt Creek. The water sampled from Salt Creek in August 1993 (base-flow conditions) is enriched in ^{18}O and D ($\delta^{18}O$ = -14.34 per mil and δD = -115.0 per mil) relative to a sample collected in April 1994 ($\delta^{18}O$ = -16.34 per mil and δD = -120.2 per mil) from the same site during peak snowmelt runoff.

Five water samples and one gypsum rock sample (Steiger, 1995, table 8) collected from sites along the hypothesized Chicken Creek drainage flow path (fig. 8) were analyzed for the isotopic concentration of sulfur to determine sources of dissolved sulfate in the water. The isotopic composition of sulfur is characterized by the ratio of sulfur-34 to sulfur-32, where $\delta^{34}S$ is the ratio referenced to a standard. Evaporites from the

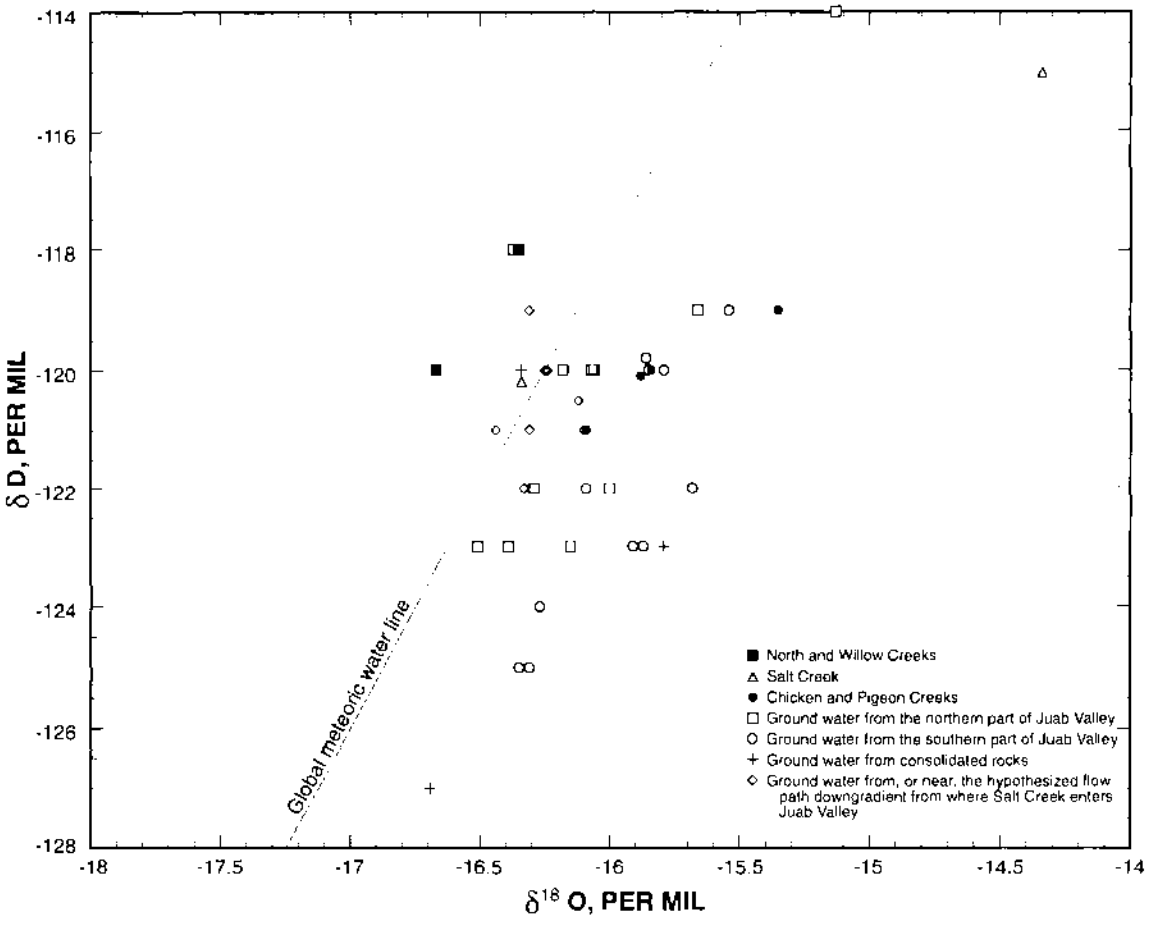


Figure 17. Relation between $\delta^{18}O$ and δD values for water sampled from selected hydrologic-data sites in Juab Valley, Utah.

Middle Jurassic Period have $\delta^{34}S$ values that range from about 15 to about 18 per mil (Holser, 1977). The gypsum rock sample collected from an outcrop in the Arapien Shale (Middle Jurassic Period) near the Chicken Creek gaging station had a $\delta^{34}S$ value of 17.4 per mil. The Arapien Shale is likely the source of much of the dissolved sulfate in ground water sampled from the southern part of Juab Valley.

Cobble Rock Spring, (D-15-2)18bab-S1, discharges from the Indianola Group in the upper part of the Chicken Creek drainage. Water from the spring likely represents the sulfur content of infiltrating precipitation and had a $\delta^{34}S$ value of 7.45 per mil. This $\delta^{34}S$ value is similar to the mean value of 7.5 per mil for ground water sampled from carbonate rocks in the central Wasatch Range (Mayo and others, 1992, p. 247). Water in Juab Valley that originated as precipitation and has dissolved gypsum from the Arapien Shale is expected to have a $\delta^{34}S$ value between 7.45 and 17.4 per mil.

Streamflow sampled from Chicken Creek at U.S. Geological Survey gaging station 10219200 (pl. 1) during base-flow conditions (August 1993) had a $\delta^{34}S$ value of 13.1 per mil. Water sampled from well (C-15-1)1baa-1, finished in unconsolidated basin-fill deposits about 2.5 mi from where Chicken Creek enters the valley, had a $\delta^{34}S$ value of 14.6 per mil. Water sampled from well (C-15-1)16bad-1 and spring (C-15-1)16bad-S1, in the discharge area of the flow path, had values of 15.9 and 14.5 per mil, respectively. The increase in $\delta^{34}S$ values in most of the water sampled along the flow path indicates the dissolution of gypsum in unconsolidated basin-fill deposits derived from the Arapien Shale. Water sampled from well (C-15-1)16bad-1, perforated from 165 to 192 ft below land surface is isotopically more enriched than water from spring (C-15-1)16bad-

S1, possibly because of movement along a deeper flow path between the mountain front and the discharge area. This would result in more contact with the unconsolidated basin-fill deposits and subsequently more reaction between the deposits and the water.

Tritium is a radioactive isotope of hydrogen that can occur as part of the water molecule. The relatively short half-life of tritium (12.43 years) makes it useful for dating recent water, particularly because large quantities were added to the atmosphere from above-ground nuclear-weapons testing beginning in 1952. Tritium in water that entered the ground-water system prior to 1952 would have decayed to concentrations of less than 1 TU by 1994, assuming no mixing with other sources of water. Concentrations measured in precipitation in Salt Lake City, Utah, by the U.S. Geological Survey peaked in 1963 at 8,230 TU, about three orders of magnitude greater than estimated concentrations in precipitation prior to nuclear-weapons testing. Concentrations in precipitation have decreased since 1963 and were about 10 to 15 TU in 1992 (R.L. Michel, U.S. Geological Survey, oral commun., 1992).

The concentration of tritium was determined for three ground-water samples from Juab Valley (Steiger, 1995, table 8) to approximate when the water entered the unconsolidated basin-fill deposits. Water from (D-11-1)8cba-1, a flowing well finished from 283 to 580 ft below land surface, had a tritium concentration of 1.1 TU, indicating that the water likely was recharged to the ground-water system prior to 1952. The error range associated with this value is plus or minus 0.3 TU. Ground water from this well likely has moved from the southern Wasatch Range into the unconsolidated basin-fill deposits (fig. 8) and has become confined by clay layers identified in the subsurface (Steiger, 1995, table 2).

Water from the southeast spring at Burriston Ponds, (D-12-1)6ddc-S1, had a tritium concentration of 21.7 TU. This value indicates a component of the ground water was recharged after 1952, most likely after 1963 when atmospheric thermonuclear testing peaked. If water discharged from spring (D-12-1)6ddc-S1 is derived solely from the hypothesized flow path that begins near Salt Creek (fig. 8), then the average flow velocity along the flow path is 3 ft/d. This velocity is based on a recharge date of 1963 and a flow path length of about 6 mi and is higher than what transmissivity data and hydraulic-head gradients would indicate. Some water from another source recharged after 1952 is necessary to account for the tritium concentration measured in water from spring (D-12-1)6ddc-S1.

Water from spring (C-15-1)16bda-S1, in the discharge area of the hypothesized Chicken Creek area flow path (fig. 8), had a concentration of 12.8 TU. This water might be a mixture of water that recharged the unconsolidated basin-fill deposits both before (flow through a deeper part of the ground-water system) and after 1952. It also might be water that recently recharged the ground-water system and has flowed through relatively shallow depths with a present-day tritium concentration that has not undergone much radioactive decay. On the basis of $\delta^{34}S$ values, which indicate that water sampled from the spring has reacted less with the basin-fill deposits than water sampled from relatively deep wells in the area, the shallower, higher velocity flow path is more likely.

Geochemical Analyses

Geochemical modeling was used to study flow paths and the sources of dissolved ions in water sampled from the Juab Valley study area. Limitations of geochemical models are imposed by the reliability and completeness of the chemical, mineralogical, and isotopic data. A modified version of the chemical-equilibrium model WATEQF (Plummer and others, 1976) available in NETPATH (Plummer and others, 1991) was used to determine the state of saturation in water of selected minerals. The state of mineral saturation can be used to evaluate chemical constraints on a ground-water system. All water samples collected in the study area were undersaturated with respect to gypsum and halite and most of the samples were near saturation or were supersaturated with respect to calcite. This means that, if a source is available, gypsum and halite will dissolve in and calcite will likely precipitate from ground water in the study area.

The computer program NETPATH (Plummer and others, 1991) was used to model possible geochemical mass-balance reactions that occur in ground water as it travels along two hypothesized flow paths in the valley. The NETPATH models are also constrained by the reliability of being able to choose an actual flow path.

Along the hypothesized Chicken Creek drainage flow path, water sampled from well (C-15-1)1baa-1 was assumed to react with minerals in the unconsolidated basin-fill deposits to form the water sampled from well (C-15-1)16bad-1. The mineral phases calcite, gypsum, and halite were assumed to occur in the basin-fill deposits, and concentrations of sulfur, calcium, sodium, and chloride were used to constrain the model calculations. The resulting geochemical mass-

balance model calculated for these conditions indicates the dissolution of 2.29 and 2.94 mmol/L of gypsum and halite, respectively, and the precipitation of 2.04 mmol/L of calcite.

The mass balance of sulfur isotopic ratios and sulfate concentration in water from the flow path and for the gypsum rock sample confirm the calculated quantity of dissolved gypsum. The mass-balance equation is:

$$C_g = (\delta^{34}S_d \, x \, C_d - \delta^{34}S_u \, x \, C_u)/\delta^{34}S_g \qquad (4)$$

where

C_g = the change in sulfate concentration (2.29 mmol/L) in water from the upgradient and downgradient wells;

$\delta^{34}S_d$ = $\delta^{34}S$ value (15.9 per mil) for water from the downgradient well, (C-15-1)16bad-1;

C_d = sulfate concentration (5.00 mmol/L) in water from the downgradient well;

$\delta^{34}S_u$ = $\delta^{34}S$ value (14.6 per mil) for water from the upgradient well, (C-15-1)1baa-1;

C_u = sulfate concentration (2.71 mmol/L) in water from the upgradient well; and

$\delta^{34}S_g$ = $\delta^{34}S$ value (17.4 per mil) for the gypsum rock sample.

The activity of an ion is an idealized concentration that is the product of the measured concentration and a conversion factor called the activity coefficient. The relation between ions can be used to evaluate chemical controls on a ground-water system. The trend in log calcium and sulfate activity in water from unconsolidated basin-fill deposits along the hypothesized Chicken Creek drainage flow path generally is 1 to 1, except for water from the discharge area where there is more sulfate relative to calcium (fig. 18). This is consistent with the geochemical reactions modeled where calcium ions derived from the dissolution of gypsum are removed from ground water by the precipitation of calcite.

The possibility that the southeast spring at Burriston Ponds, (D-12-1)6ddc-S1, may discharge ground water from a hypothesized flow path that extends downgradient from where Salt Creek enters Juab Valley and from a hypothesized flow path that represents ground water moving downgradient from the southern Wasatch Range (fig. 8) also was tested with geochemical modeling. Chemical analyses of water sampled from wells (D-12-1)19acb-1 and (D-11-1)33cab-1 were assumed to be representative of water from the two flow paths, respectively. The mineral phases calcite, gypsum, and halite were assumed to occur in the ground-water system, and sulfur, calcium, sodium, and chloride concentrations were used to constrain the model calculations. The results of the NETPATH model indicate a mixture of about 70 percent water from well (D-12-1)19acb-1 (water from the south) and about 30 percent water from well (D-11-1)33cab-1 (water from the east). The mass-balance model also indicates that the dissolution of 1.12 and 0.12 mmol/L of gypsum and halite, respectively, and the precipitation of 0.87 mmol/L of calcite is necessary to form the water sampled from spring (D-12-1)6ddc-S1.

The computer program SNORM is a geochemical model that uses water-chemistry data to calculate the idealized equilibrium (normative) assemblage of mineral salts that would precipitate from an evaporated water sample at surface conditions, also referred to as the salt norm of water (Bodine and Jones, 1986). The salt norm is quantitatively equivalent to the solute concentrations in the water. Characterization of water compositions as salt norms was used to help determine sources of solutes in ground water sampled from Juab Valley.

The normative assemblage of mineral salts for ground water sampled along the hypothesized flow path in the Chicken Creek drainage (fig. 8) generally consists of calcium sulfate (about 55 percent by weight), magnesium carbonate (about 25 percent by weight), and sodium chloride (about 16 percent by weight). Calcium sulfate, as the predominant member of the assemblage, reflects the dissolution of evaporitic gypsum ($CaSO_4 \cdot 2H_2O$) present in the unconsolidated basin-fill deposits derived from the Arapien Shale. The predominant normative salts computed for water sampled from Palmer Spring, (C-15-1)16cdb-S1, are calcium sulfate (about 37 percent by weight), sodium chloride (about 33 percent by weight), and magnesium sulfate (about 15 percent by weight). These salts also likely are dissolved from material deposited in an evaporite setting, possibly a playa in the southern part of Juab Valley, because of the large percentage of halite.

Data analysis using SNORM indicates that the water sampled from spring (C-13-1)3bbc-S1 and wells (C-13-1)3dad-1, (C-13-1)23add-1, (C-14-1)14cac-1 and (C-14-1)22ddc-1, generally on the west side of Juab Valley, contains substantial percentages of magnesium chloride (more than 15 percent by weight) in addition to sodium chloride and calcium sulfate salts. A diagnostic feature of sea water is the presence of magnesium chloride at about 6.5 percent by weight. The normative salt assemblage for sea water that has under-

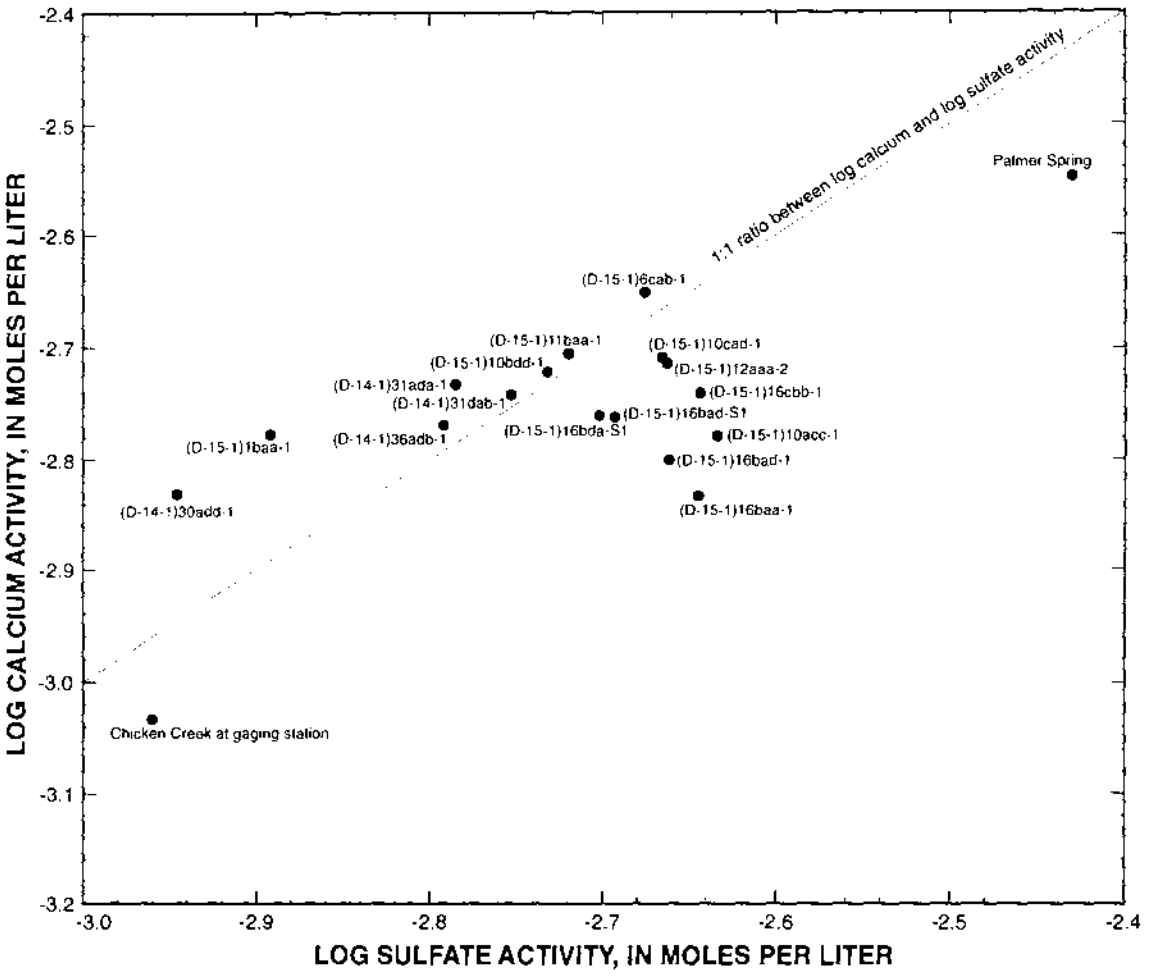


Figure 18. Relation between log calcium and log sulfate activity for water sampled along the hypothesized Chicken Creek drainage flow path in Juab Valley, Utah.

gone evaporation has less sodium chloride and calcium sulfate (because of mineral precipitation) and more magnesium chloride salts than unaltered sea water (Bodine and Jones, 1986, p. 40). If calcium and sulfate are added to sea water, then the salt norm will have an excess of calcium sulfate relative to unaltered sea water; therefore, water from these sites might be a mixture of residual sea water that has undergone evaporation and meteoric water that has dissolved gypsum from an evaporitic source, such as the Arapien Shale or basin-fill deposits derived from it. Normal faulting on the west side of the valley may have provided a route for water from a deeper consolidated-rock source to move into the unconsolidated basin-fill deposits. Water from wells (C-13-1)3dad-1 and (C-14-1)22ddc-1 is warmer than most ground water in the valley, possibly because of the geothermal gradient.

The ratio of bromide to chloride concentrations in water can be used to determine the sources of chloride in the water. The bromide/chloride (Br/Cl) ratio for meteoric water can be assumed to be similar to the ratio for ocean water, which is about 0.00347. Sodium and chloride can precipitate in the form of halite from water undergoing evaporation and leave the residual water concentrated in bromide relative to chloride. Water with a Br/Cl ratio that is near 0.003 and with high concentrations of chloride (relative to other ground water in the valley that is known to be meteoric in origin) contains a saline or brine component. Such water likely has been concentrated with bromide relative to chloride by the precipitation of halite. Water that has a

Br/Cl ratio that is much less than 0.003 probably represents meteoric water that became enriched in chloride relative to bromide as it flowed through evaporite deposits after it recharged the ground-water system (White and others, 1963, p. F13). Dissolution of the evaporite salts results in an increased concentration of sodium, chloride, calcium, and sulfate ions but a relatively small increase in bromide concentration in the water.

Ground water sampled from the area north of Burriston Ponds had Br/Cl ratios that ranged from 0.00267 to 0.00357, attributable mainly to a meteoric water source and not much water-rock interaction. The basin-fill deposits in this area are derived from rocks in the southern Wasatch Range, which contain relatively small amounts of chloride. Water sampled from wells and springs along a hypothesized flow path from Nephi to Burriston Ponds (fig. 8) had much smaller Br/Cl ratios, which ranged from 0.00044 to 0.00079, indicative of some halite dissolution.

The ratio of bromide to chloride in water sampled from wells and springs near the hypothesized flow path representing the Chicken Creek drainage (fig. 8) ranged from 0.00018 for water collected from Chicken Creek at U.S. Geological Survey gaging station 10219200 in September 1992 to 0.00290 for water collected from Cobble Rock Spring, (D-15-2)18bab-S1, in the San Pitch Mountains. The increase in chloride relative to bromide concentration in water from these two sites is a result of the dissolution of halite by meteoric water as it flows in the Chicken Creek channel. The Br/Cl ratio for water from Chicken Creek derived from snowmelt runoff during the spring should be closer to that of Cobble Rock Spring because of dilution.

Water from wells finished in unconsolidated basin-fill deposits near the mountain front in the Levan area had Br/Cl ratios that ranged from 0.00107 to 0.00162. In the discharge area of the hypothesized flow path representing the Chicken Creek drainage, the Br/Cl ratio in water from springs (C-15-1)16bad-S1 and 16bda-S1 is 0.00220 and 0.00120, respectively. These ratios are indicative of a meteoric water source in which some halite was dissolved as it traveled along the flow path. The Br/Cl ratio of water from relatively deep wells in the area, (C-15-1)16baa-1 and 16bad-1, is 0.00050 and 0.00038, respectively. These ratios are indicative of a meteoric water source in which more halite is dissolved relative to water from the springs. These wells are perforated from about 165 to 192 ft below land surface. The relation between Br/Cl ratio and depth also is supported by Br/Cl ratios in water sampled from wells (C-15-1)10bdd-1 (0.00100) and (C-15-1)10acc-1 (0.00033), perforated from 120 to 140 ft and from 132 to 350 ft, respectively. Although Br/Cl ratios determined for ground water sampled from the southern part of the valley indicate the dissolution of evaporite material as a source of chloride, water from deeper intervals in and near the discharge area has more dissolved chloride because the water has traveled along a longer flow path and has had more time to react with the basin-fill deposits.

Relatively high concentrations of sodium and chloride in water sampled from wells (C-14-1)14cac-1 and 22ddc-1 on the west side of the valley might be caused by meteoric water mixing with residual sea water that has undergone evaporation in unconsolidated basin-fill deposits in the area. The ratios of bromide to chloride in water from the two wells (0.00269 and 0.00236, respectively) is similar to that of meteoric water and indicates that the increase in chloride ions is not caused by halite dissolution. This area of the valley is near the ground-water divide that separates the northern and southern parts of Juab Valley and receives a relatively small amount of recharge from direct precipitation and runoff from two small ephemeral drainages in West Hills. A minor amount of ground water also may move into this area from the east, but it is not enough to sufficiently dilute any water from a deep source that may be present in the area.

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

A numerical model of the ground-water system in the unconsolidated basin-fill deposits of Juab Valley was constructed using the U.S. Geological Survey modular, three-dimensional, finite-difference, ground-water flow model (McDonald and Harbaugh, 1988). The numerical model was calibrated to simulate the steady-state conditions of 1949, multi-year transient-state conditions during 1949-92, and seasonal transient-state conditions during 1992-94.

The boundary conditions of the numerical model were chosen to simulate the conceptual model of the ground-water system presented in previous sections of this report. The objectives of the simulations were to test and refine the conceptual model and to provide a tool that can be used to estimate the effects of changes in ground-water discharge from pumped wells and ground-water recharge on ground-water levels and in a qualitative sense, on natural ground-water discharge.

A description of the construction and calibration of the numerical model is presented in the following sections of this report. The amount and complexity of the required input data make it impractical to present or reference all the information required to completely reconstruct this model. A copy of the ground-water flow model for Juab Valley and all associated data sets can be obtained from the U.S. Geological Survey at the Utah District office in Salt Lake City, Utah.

Model Construction

The construction of a numerical model to simulate three-dimensional ground-water flow in the unconsolidated basin-fill deposits of Juab Valley is based upon data and assumptions made to create a conceptual model of the area. These data and assumptions are distributed both laterally and vertically through the model in order to best represent the ground-water system.

Discretization

The numerical model of the ground-water system in Juab Valley is horizontally discretized by a rectangular grid consisting of 31 rows and 157 columns. The grid is oriented along the axis of Juab Valley, and all model cells in the grid are 1/16 mi^2 (40 acres) (fig. 19). The ground-water flow equations and associated hydrologic properties used to calculate flow for each model cell are formulated at the center point or node of the cell. In the numerical model, values are assigned to the node, and those values represent the average characteristics of the entire model cell. The selected cell size is small enough so that steep hydraulic gradients and areas where larger amounts of data are available or where hydraulic stresses are concentrated can be represented numerically.

The lateral boundaries of the numerical model are shown in figure 19; they generally correspond to the lateral extent of the saturated unconsolidated basin-fill deposits in Juab Valley. Vertically, the numerical model is discretized into four layers (fig. 20). This number of layers is adequate to simulate unconfined and confined conditions and the vertical movement of ground water, and to distribute the hydrologic stresses caused by ground-water discharge to pumped wells. The contacts between model layers are referenced in feet below the estimated top of the saturated unconsolidated basin-fill deposits. Although the conceptual ground-water system is generalized to include the upper 1,000 ft of unconsolidated basin-fill deposits, the actual thickness is not known. The bottom boundary used in the numerical model is assumed to be impermeable and represents the contact between unconsolidated and semiconsolidated basin-fill deposits. This is a simplification of the present (1995) conceptualization of the ground-water system, which is that the contact between unconsolidated and semiconsolidated basin-fill deposits is gradational, and although ground-water flow may occur in the semiconsolidated deposits, they are much less permeable than the unconsolidated deposits. If additional subsurface and hydrologic-property information indicates more permeable deposits at depth, then the ground-water flow model described in this report may need to be updated and recalibrated to include this deeper part of the system before using the model for projected changes caused by increased pumping.

An estimate of the top of the saturated unconsolidated basin-fill deposits was made on the basis of average water levels, reservoir and land-surface altitudes, and water-level contours. An average water level at each of 191 wells was used to estimate the top of the saturated deposits at model nodes that correspond to well locations. The averages are computed from the available water-level data for each well (Steiger, 1995, tables 1 and 3). At model nodes that represent Mona and Chicken Creek Reservoirs, the top of the saturated deposits is estimated to be equal to the altitude of the water surface of the reservoir determined from the appropriate U.S. Geological Survey 1:24,000 topographic quadrangle map. The top of the saturated deposits for the remaining model nodes was estimated on the basis of two generalized water-level contour maps. The first map is of water-level contours for the northern part of Juab Valley during the spring of 1950 (Bjorklund, 1967, pl. 4). The second map shows water-level contours for all of Juab Valley during the fall, winter, and spring of 1963-64 (Arnow, 1964, p. 53). These contour maps were used because they represent, respectively, water levels in the northern and southern parts of Juab Valley prior to large-scale ground-water development. The water-level contours were modified by (1) extending them to the approximate boundary of the unconsolidated basin-fill deposits, (2) adjusting them to match in the area where the northern and southern maps overlap, and (3) equating them to the altitude of land surface in areas where flowing wells exist.

Model layer 1 represents unconfined ground-water conditions in Juab Valley. The bottom of model layer 1 was set at 50 ft below the estimated top of the

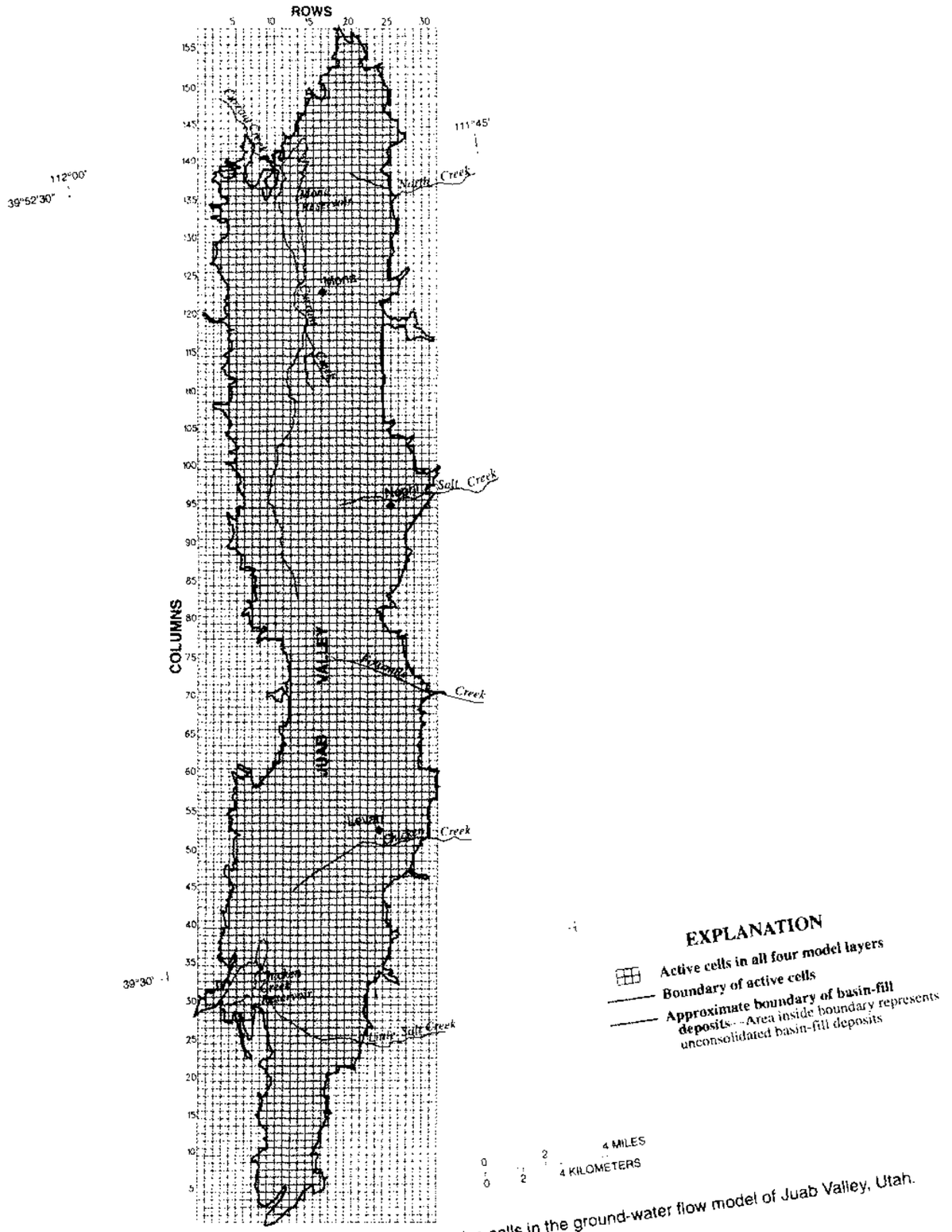


Figure 19. Model grid and location of active cells in the ground-water flow model of Juab Valley, Utah.

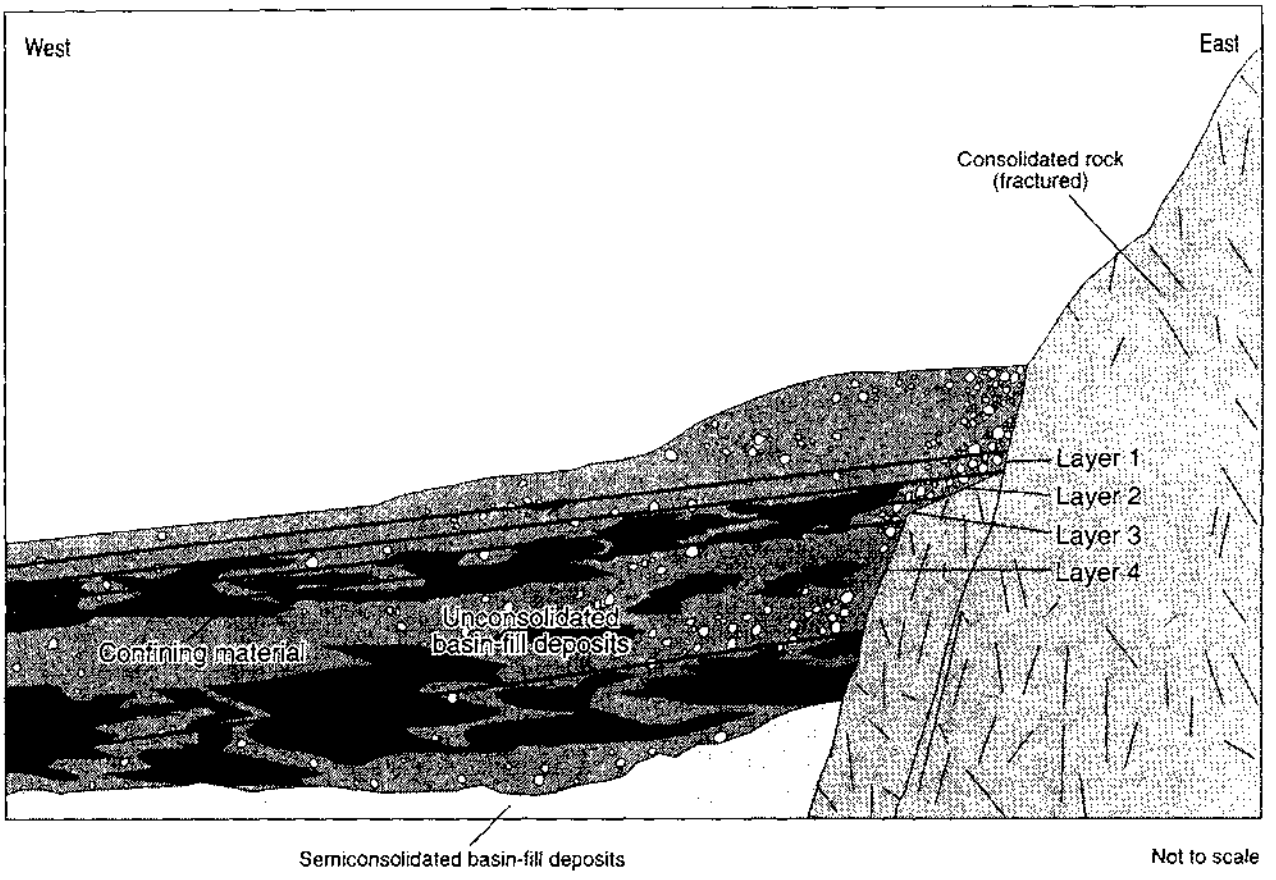


Figure 20. Generalized geology and the four layers used in the ground-water flow model of Juab Valley, Utah.

saturated unconsolidated basin-fill deposits. This corresponds with the conceptualization that unconfined conditions exist in the ground-water system in Juab Valley in about the upper 50 ft of saturated deposits. Changes in simulated ground-water recharge and discharge can cause the saturated thickness of model layer 1 to vary.

Model layers 2, 3, and 4 represent confined ground-water conditions in Juab Valley. The thickness of layers 2 and 3 are 60 ft each, on the basis of the depth at which the average annual (1963-93) ground-water discharge to pumped wells occurs. The average annual (1963-93) ground-water discharge to pumped wells in Juab Valley (about 20,000 acre-ft) was subdivided into amounts removed from 10-ft sections of the saturated unconsolidated basin-fill deposits on the basis of screened or perforated intervals (fig. 21). The average annual ground-water pumpage from the first 50 ft of saturated deposits (layer 1) is about 3,700 acre-ft. The remaining 16,300 acre-ft is evenly distributed to model layers 2, 3, and 4 (fig. 21). The depth at which water is withdrawn by a well corresponds to the perforated interval of the well as measured from the top of the saturated unconsolidated basin-fill deposits. If perforated interval information is not available for a well, the depth at which water is withdrawn corresponds to the depth of the well, again as measured from the top of the saturated deposits.

No thickness was assigned to layer 4. This layer is assumed to extend to a depth below which only small amounts of ground-water flow occur (see "Ground-water occurrence" section of this report). Some pumped wells in the valley withdraw water from depths exceeding 400 ft (fig. 21); therefore, a minimum thickness for model layer 4 is 230 ft, which is substantially greater than the thicknesses of model layers 1, 2, and 3.

Boundary Conditions and Data Requirements

Boundary conditions are used to simulate recharge to and discharge from the ground-water sys-

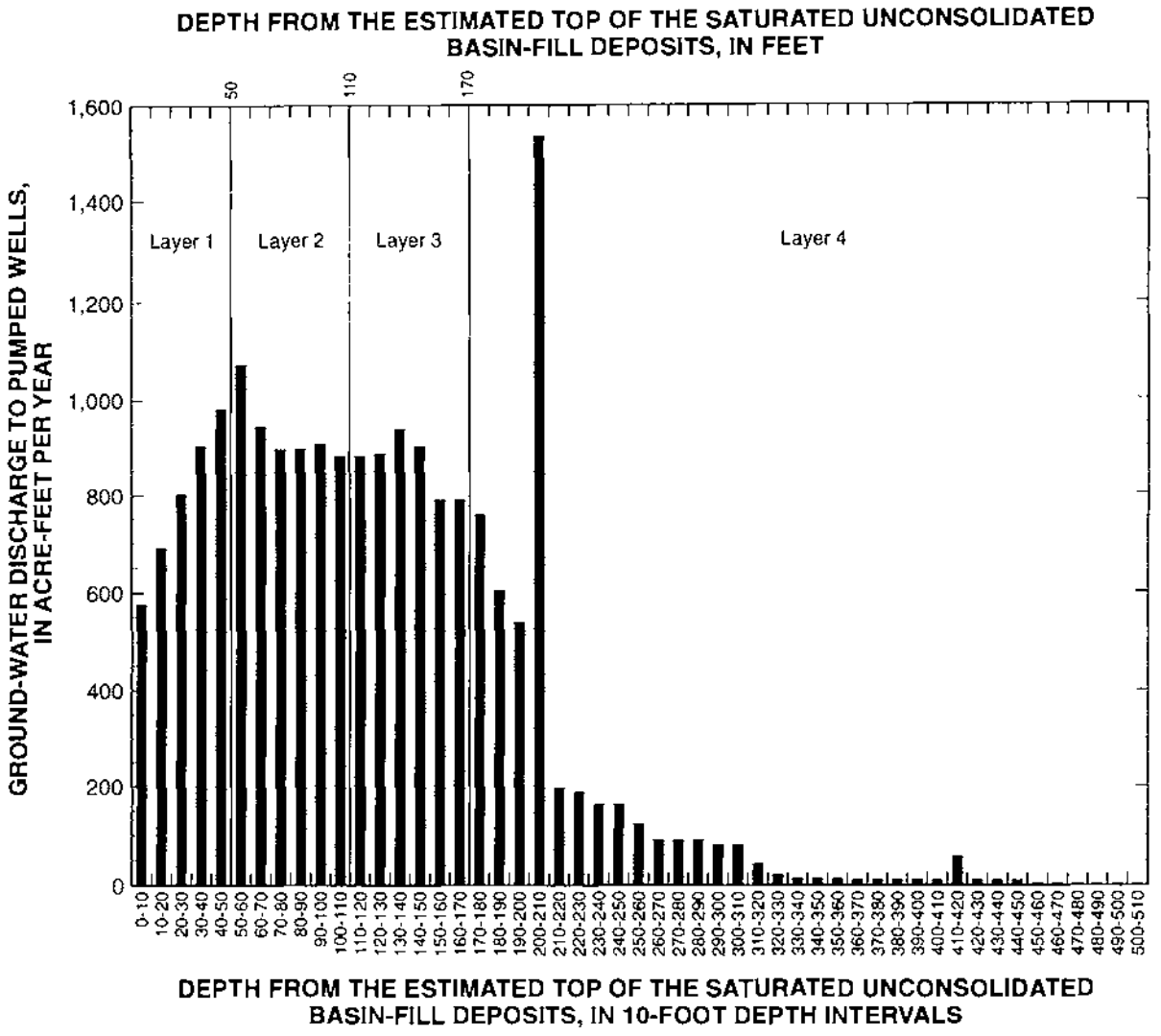


Figure 21. Depth from the estimated top of the saturated unconsolidated basin-fill deposits, the amount of 1963-93 average annual ground-water discharge to pumped wells in each depth interval, and corresponding model layers in the ground-water flow model of Juab Valley, Utah.

tem in Juab Valley and to define the mathematical limits of the numerical model. Head-dependent flux and specified-flux boundaries are the two general types of model boundaries used to describe the process by which recharge to and discharge from the ground-water system occur. At head-dependent flux boundaries, simulated flow across a boundary is proportional to the difference in model-computed and assigned water levels on adjacent sides of the boundary. These boundaries simulate recharge and discharge that is dependent on local water levels. Flow across specified-flux boundaries is assigned by the user and remains constant regardless of model-computed water levels and stresses. Specified-flux boundaries are used to simulate recharge and discharge that is independent of local water levels. On the basis of available data, the boundaries chosen in this numerical model are thought to be a reasonable mathematical representation of how the modeled ground-water system interacts with the surrounding unmodeled hydrological system. All other model boundaries are impermeable and water does not move across them.

Recharge

Recharge to the ground-water system is represented in the numerical model by seepage from streams, unconsumed irrigation water, and distribution

systems; infiltration of precipitation; and subsurface inflow. These components of recharge are simulated with specified-flux boundaries formulated by the Recharge and Well Packages (McDonald and Harbaugh, 1988, p. 7-1 and p. 8-1), and the amount of recharge is considered to be independent of local ground-water level.

Seepage from streams is separated into seepage from streams that enter Juab Valley on the east side during the nonirrigation season, seepage from streams that enter Juab Valley on the east side during the irrigation season, and seepage from streams that enter the valley on the west side. A specified-flux formulation is used to simulate recharge from nonirrigation-season streamflow because the recharge must first move through an unsaturated zone. Local water levels are not thought to affect this recharge. Recharge from nonirrigation streamflow is evenly applied to layer 1, generally at model nodes that correspond to the irrigation areas listed in table 3 and shown in figure 22. An exception is recharge during the nonirrigation season from North Creek, Willow Creek, Couch Canyon, and Bear Canyon, which is applied only to the respective stream channels near the mountain front. Data are not available to estimate recharge from nonirrigation-season streamflow directly, and the amount of recharge assumed for the simulations is varied for each stress period on the basis of the amount of recharge estimated for average conditions, 14,400 acre-ft/yr (the sum of values presented for the northern and southern parts of Juab Valley in table 2). This amount is multiplied by the ratio of actual recorded precipitation during the stress period to the average (1931-93) annual precipitation at Nephi and Levan. Recharge that is specified from seepage from nonirrigation-season streamflow is applied on the basis of the estimated amount for each stream modified by the yearly ratio. This method of estimation assumes that changes in precipitation adequately describe the variability of nonirrigation-season streamflow. For the seasonal transient-state simulation, ground-water recharge by seepage from nonirrigation-season streamflow is applied equally during the stress periods that simulate October through December and January through March. This amount is determined from the relation described previously. At the time of model calibration, annual precipitation was not known for 1994 and the nonirrigation-season streamflow for January through March 1994 was estimated on the basis of precipitation in 1993. The amounts of recharge assumed for the simulations that result from the seepage of nonirrigation-season streamflow are listed in tables 4, 5, and 6.

Recharge from unconsumed irrigation water and distribution-systems losses includes the irrigation-season component of seepage from streams that enter the valley on the east side and is simulated as a specified-flux boundary using the Recharge Package (McDonald and Harbaugh, 1988, p. 7-1). Model nodes in layer 1 were assigned to specific irrigation areas by overlaying the model grid and the mapped extent of the irrigation areas listed in table 3. Recharge for each irrigation area is distributed evenly over the entire irrigation area (fig. 22). Recharge that is specified from unconsumed irrigation water and distribution-system losses depends on the amount of irrigation-season streamflow and the amount of ground water pumped from wells for irrigation purposes in each irrigation area. Data are not available to estimate irrigation-season streamflow directly. Streamflow amounts are varied for each stress period on the basis of the amount of irrigation-season streamflow estimated for average conditions, 36,000 acre-ft/yr (the sum of values presented for the northern and southern parts of Juab Valley in table 3). This amount is multiplied by the ratio of actual recorded precipitation during the stress period to the average (1931-93) annual precipitation at Nephi and Levan. For the seasonal stress periods, this ratio is based on annual precipitation. This method of estimation assumes that the variability is adequately described by precipitation. The amount of ground-water pumpage from wells was determined for each stress period using unpublished data in the files of the U.S. Geological Survey Utah District office in Salt Lake City, Utah. Recharge from unconsumed irrigation water and distribution-system losses was assumed to be 30 percent of the irrigation-season streamflow and 10 percent of the ground water pumped from wells for each irrigation area and for each stress period.

For the seasonal transient-state simulation, recharge from unconsumed irrigation water and distribution-system losses is applied during the April through June and July through September stress periods. Seventy percent of the recharge from unconsumed irrigation water and distribution-system losses that is derived from streamflow is applied during April through June, and the remaining 30 percent is applied during July through September. All of the recharge from unconsumed irrigation water and distribution-system losses that is derived from ground water pumped from wells is applied during July through September. The amounts of specified recharge resulting from

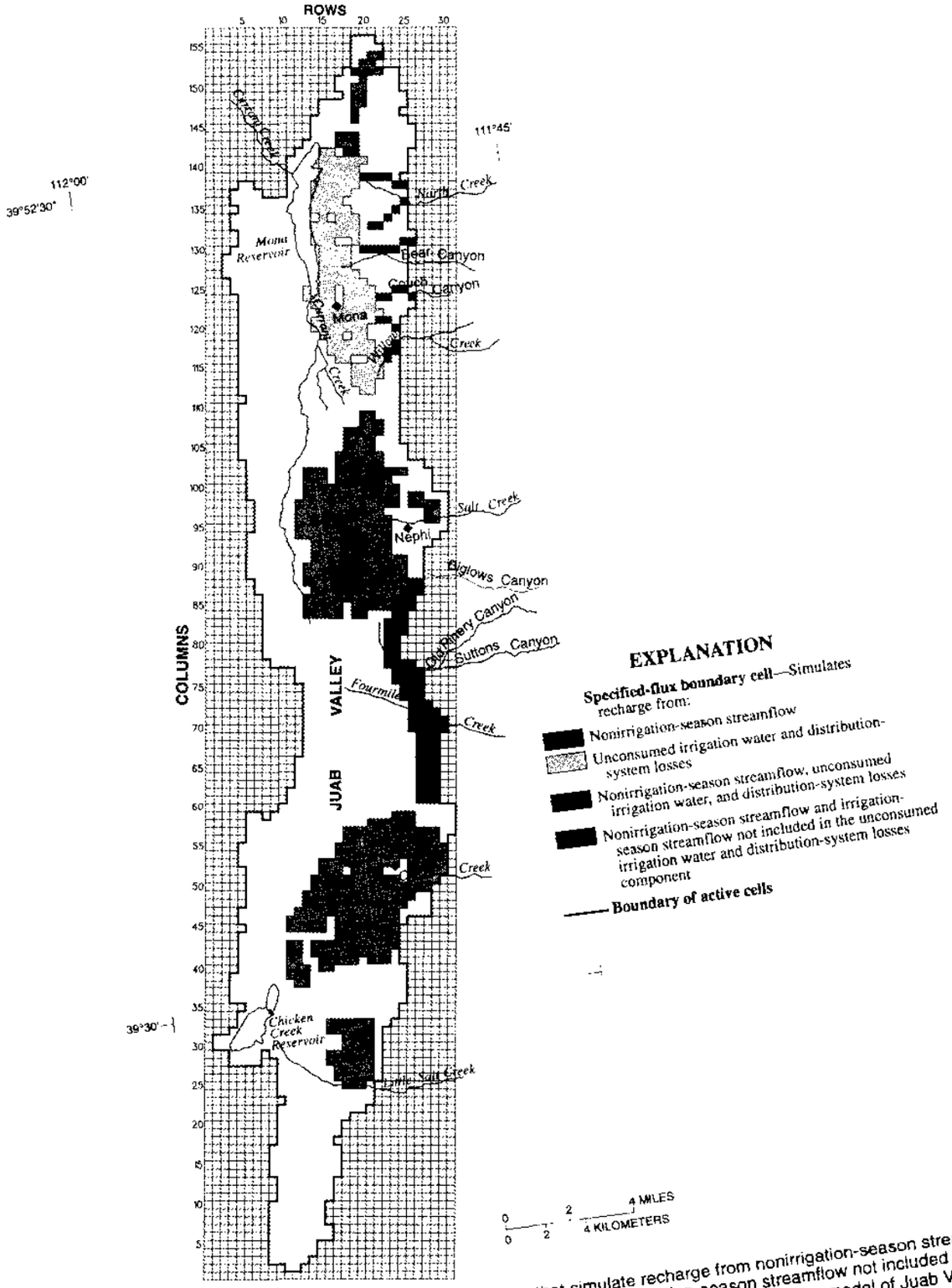


Figure 22. Location of specified-flux boundary cells that simulate recharge from nonirrigation-season streamflow, unconsumed irrigation water and distribution-system losses, and irrigation-season streamflow not included in the unconsumed irrigation water and distribution-system losses component in the ground-water flow model of Juab Valley, Utah.

Table 4. Specified ground-water recharge and discharge and model-computed discharge for the 1949 steady-state simulation, Juab Valley, Utah

[— , not applicable]

Budget component	Flow, in acre-feet per year (rounded) Northern part[1]	Southern part[1]
Specified recharge		
Seepage from nonirrigation-season streamflow	11,240	2,250
Seepage from unconsumed irrigation water and distribution systems	6,900	3,560
Seepage from irrigation-season streamflow not included in the unconsumed irrigation water and distribution-system losses component	710	0
Infiltration from precipitation	3,560	2,520
Subsurface inflow and seepage from ephemeral streams to east side of Juab Valley	10,480	1,620
Subsurface inflow and seepage from ephemeral streams to west side of Juab Valley	3,980	1,420
Total specified recharge	**36,870**	**11,370**
Specified discharge		
Wells pumped for irrigation and public supply	400	90
Subsurface outflow to consolidated rock/unconsolidated basin-fill deposits boundary	950	0
Total specified discharge	**1,350**	**90**
Model-computed discharge		
Flowing wells	30	300
Springs and seeps		
Currant Creek upstream from gaging station 10146400	10,570	—
Currant Creek downstream from gaging station 10146400	3,510	—
Mona Reservoir	9,770	—
Palmer Spring	—	370
Chicken Creek Reservoir	—	4,020
Evapotranspiration	13,300	4,650
Total model-computed discharge	**37,180**	**9,340**

[1]The division of Juab Valley into northern and southern parts is based on a topographic divide and not a ground-water divide. In 1949, the simulated flow from the southern to the northern part of Juab Valley was about 2,000 acre-feet per year. The difference between recharge and discharge amounts is mainly the simulated flow from the southern to the northern part of Juab Valley.

unconsumed irrigation water and distribution-system losses for the steady-state and transient-state simulations are listed in tables 4, 5, and 6.

Recharge from irrigation-season streamflow not included in the unconsumed irrigation water and distribution-losses component of recharge comes from streamflow during the irrigation season from Biglows, Old Pinery, and Suttons Canyons and Fourmile Creek. The recharge is applied to 88 model nodes in layer 1, adjacent to the areas where these streams enter Juab Valley (fig. 22). As described previously, specified recharge from streamflow is varied for each stress period on the basis of the amount of irrigation-season streamflow estimated for average conditions (750 acre-ft/yr, table 2) and precipitation. For the seasonal transient-state simulation, irrigation-season streamflow is varied on the basis of annual precipitation. Seventy percent of the specified recharge is applied during April through June and 30 percent is applied during July through September. The amounts of specified recharge that result from irrigation-season streamflow not included in the unconsumed irrigation water and distribution-losses component for the steady-state and transient-state simulations are listed in tables 4, 5, and 6.

Recharge from infiltration of precipitation is simulated as a specified-flux boundary using the Recharge

Package (McDonald and Harbaugh, 1988, p. 7-1). Recharge from infiltration of precipitation is distributed evenly to all active model nodes in layer 1 and is assumed to be 5 percent of the precipitation that occurred at Nephi and Levan during the stress period. The amounts of specified recharge from precipitation for the steady-state and transient-state simulations are listed in tables 4, 5, and 6.

Recharge from subsurface inflow from consolidated rocks and seepage from ephemeral streams to the west and east sides of Juab Valley is simulated as a specified-flux boundary using the Well Package (McDonald and Harbaugh, 1988, p. 8-1). Subsurface inflow is a head-dependent process that is a function of the hydraulic conductivity of the boundary between the consolidated rock and unconsolidated basin-fill deposits and the hydraulic gradient across the boundary. Data are not available to define these properties or the quantity of seepage from ephemeral streams; therefore, a simplistic specified-flux approach was used to simulate these components of recharge to the unconsolidated basin-fill deposits. The specified-flux formulation does not account for the variability in the amount of subsurface inflow caused by water-level changes in the consolidated rocks and unconsolidated basin-fill deposits.

Recharge is applied using injection wells placed in model layers 1, 2, 3, and 4 at the model nodes that correspond to the western and eastern extent of the saturated unconsolidated basin-fill deposits in the valley. This component of recharge is distributed evenly at all nodes and model layers along the west side of the valley to simulate seepage from ephemeral streamflow from both small and large drainages and from subsurface inflow (fig. 23). It is represented by 272 injection wells in the northern part of Juab Valley and 156 injection wells in the southern part. At the start of the calibration process, recharge from subsurface inflow and seepage from ephemeral streams to the west side of Juab Valley was set at 4,200 acre-ft/yr for the northern part and 1,500 acre-ft/yr for the southern part (table 2).

Recharge from subsurface inflow and seepage from ephemeral streams to the east side of the valley is represented by 288 injection wells in the northern part and 208 injection wells in the southern part and is distributed equally to the four model layers. At the start of the calibration process, recharge from subsurface inflow and seepage from ephemeral streams to the east side of the valley was set at 12,000 acre-ft/yr for the northern part and 1,000 acre-ft/yr for the southern part. The estimated average amount of recharge from subsurface inflow and seepage from ephemeral streams to the east side of the valley ranges from 13,700 to 17,700 acre-ft/yr (the sum of values presented for the northern and southern parts of Juab Valley in table 2).

For the seasonal transient-state simulation, the annual amount of recharge to the sides of the valley was partitioned as 25, 20, 30, and 25 percent, respectively, for the January through March, April through June, July through September, and October through December stress periods. These percentages are based on reported seasonal variations in discharge to Bradley Spring and selected springs and spring-fed drainages in the southern Wasatch Range that discharge from consolidated rocks. The estimate of recharge from subsurface inflow and seepage from ephemeral streams was determined to be the residual of other estimated recharge and discharge components of the ground-water budget for Juab Valley. The amount of recharge from subsurface inflow and seepage from ephemeral streams to the valley and the spatial distribution of the recharge to the east side were among several parameters adjusted during the calibration process until a satisfactory calibration was achieved (see "Calibration parameters" section of this report).

Discharge

Ground-water discharge to wells, springs, seeps, evapotranspiration, and subsurface outflow is represented in the numerical model with head-dependent flux and specified-flux boundaries. Discharge to pumped wells is the only ground-water budget component that has regularly been measured through time. Little data are available on the other budget components with which discharge can be estimated under steady-state and transient-state conditions.

Discharge to pumped wells is represented as a specified-flux boundary using the Well Package (McDonald and Harbaugh, 1988, p. 8-1). Only wells that pump 50 acre-ft/yr (30 gal/min) or more, generally those used for irrigation, are simulated. Required input data for the Well Package are the model cells and layers that correspond to the location and depth of the perforated intervals of the wells (fig. 24). If a well is screened in more than one layer, the amount of discharge assigned to each model layer equals the total discharge multiplied by the fraction of the perforated interval in that layer. The perforated interval is measured as the distance from the top of the first perforations to the bottom of the last perforations. If perforation data are not available, well discharge is simulated at the model layer that corresponds to the

Table 5. Specified ground-water recharge and discharge and model-computed discharge and change in storage for the multi-year transient-state simulation, Juab Valley, Utah

Budget element	Flow, in acre-feet per year, for each stress period (rounded)[1]							
	1949-53	1954-59	1960-67	1968-69	1970-77	1978-82	1983-86	1987-92
Northern part of Juab Valley[2]								
Specified recharge								
Seepage from nonirrigation-season streamflow	11,650	10,470	12,300	13,680	10,540	15,160	15,800	10,700
Seepage from irrigation-season streamflow not included in unconsumed irrigation water component	730	660	770	860	670	950	990	680
Seepage from unconsumed irrigation water and distribution systems	7,210	6,900	9,020	10,050	8,310	10,590	10,100	8,550
Infiltration from precipitation	3,680	3,320	3,870	4,300	3,310	4,790	4,970	3,380
Subsurface inflow and seepage from ephemeral streams to east side of Juab Valley	10,840	9,750	12,660	17,780	9,810	23,300	25,660	9,970
Seepage from ephemeral streams and subsurface inflow to west side of Juab Valley	4,120	3,710	4,810	6,760	3,730	8,850	9,750	3,790
Total	**38,230**	**34,810**	**43,430**	**53,430**	**36,370**	**63,640**	**67,270**	**37,070**
Specified discharge								
Withdrawal from wells pumped for irrigation and public supply	1,050	5,050	12,150	12,560	18,770	13,280	4,570	17,220
Subsurface outflow to consolidated rock/ unconsolidated basin-fill deposits boundary	980	880	1,150	1,610	890	2,110	2,320	900
Total	**2,030**	**5,930**	**13,300**	**14,170**	**19,660**	**15,390**	**6,890**	**18,120**
Model-computed discharge								
Flowing wells	130	520	530	580	290	610	1,180	600
Springs and seeps								
Currant Creek upstream from gaging station 10146400	10,290	9,640	9,310	9,590	7,720	9,890	12,230	9,480
Currant Creek downstream from gaging station 10146400	3,480	3,340	3,270	3,370	2,930	3,600	4,160	3,390
Mona Reservoir	9,580	8,770	8,370	8,860	6,650	9,670	12,320	8,800
Springs and seeps total	23,350	1,750	20,950	21,820	17,300	23,160	28,710	21,670
Evapotranspiration	13,010	12,060	11,330	11,710	7,380	11,690	13,970	11,340
Total	**36,490**	**34,330**	**32,810**	**34,110**	**24,970**	**35,460**	**43,860**	**33,610**
Water going into (+) or out of (-) storage[3]	+1,370	-3,750	-1,050	+6,760	-6,790	+14,120	+17,820	-13,490

Table 5. Specified ground-water recharge and discharge and model-computed discharge and change in storage for the multi-year transient-state simulation, Juab Valley, Utah—Continued

Budget element	Flow, in acre-feet per year, for each stress period (rounded)[1]							
	1949-53	1954-59	1960-67	1968-69	1970-77	1978-82	1983-86	1987-92
Southern part of Juab Valley[2]								
Specified recharge								
Seepage from nonirrigation-season streamflow	2,320	2,100	2,460	2,730	2,100	3,030	3,150	2,140
Seepage from irrigation-season streamflow not included in unconsumed irrigation water component	0	0	0	0	0	0	0	0
Seepage from unconsumed irrigation water and distribution systems	3,720	3,410	4,150	4,630	3,950	5,290	5,300	4,180
Infiltration from precipitation	2,620	2,350	2,740	3,050	2,350	3,400	3,530	2,390
Subsurface inflow and seepage from ephemeral streams to east side of Juab Valley	1,670	1,510	1,960	2,750	1,510	3,600	3,960	1,540
Seepage from ephemeral streams and subsurface inflow to west side of Juab Valley	1,470	1,320	1,720	2,410	1,330	3,160	3,480	1,350
Total	**11,800**	**10,690**	**13,030**	**15,570**	**11,240**	**18,480**	**19,420**	**11,600**
Specified discharge								
Withdrawal from wells pumped for irrigation and public supply	510	1,050	2,360	3,120	6,270	5,090	3,270	7,990
Subsurface outflow to consolidated rock/ unconsolidated basin-fill deposits boundary	0	0	0	0	0	0	0	0
Total	**510**	**1,050**	**2,360**	**3,120**	**6,270**	**5,090**	**3,270**	**7,990**
Model-computed discharge								
Flowing wells	890	870	1,130	1,150	1,000	1,110	1,220	1,030
Springs and seeps								
Palmer Spring	340	340	340	340	320	330	350	330
Chicken Creek Reservoir	3,640	3,550	3,530	3,600	3,250	3,570	3,830	3,320
Springs and seeps total	3,980	3,890	3,870	3,940	3,570	3,900	4,180	3,650
Evapotranspiration	4,280	3,990	3,750	3,860	2,610	3,270	4,130	2,740
Total	**9,150**	**8,750**	**8,750**	**8,950**	**7,180**	**8,280**	**9,530**	**7,420**
Water going into (+) or out (-) of storage[3]	+220	-1,060	+80	+1,710	-3,870	+3,630	+5,180	-5,230

[1]Flow is based on rate computed for the last time step in the stress period.

[2]The division of the northern and southern parts of Juab Valley is based on a topographic divide and not a ground-water divide. Depending on the hydrologic conditions, simulated flow from the southern to the northern part of Juab Valley varies from about 1,000 to about 2,000 acre-feet per year.

[3]Water going into (+) storage is treated as discharge by the numerical model; water going out of (-) storage is treated as recharge by the numerical model. The difference between recharge, discharge, and storage amounts for a stress period is approximately the simulated flow from the southern to the northern part of Juab Valley.

Table 6. Specified ground-water recharge and discharge and model-computed discharge and change in storage for the seasonal transient-state simulation, Juab Valley, Utah

Budget element	Flow, in acre-feet, for each stress period (rounded)[1]				
	January 1 to March 31, 1993	April 1 to June 30, 1993	July 1 to September 30, 1993	October 1 to December 31, 1993	January 1 to March 31, 1994
Northern part of Juab Valley[2]					
Specified recharge					
Seepage from nonirrigation-season streamflow	7,290	0	0	7,290	7,290
Seepage from irrigation-season streamflow not included in unconsumed irrigation water component	0	640	270	0	0
Seepage from unconsumed irrigation water and distribution systems	0	6,200	3,800	0	0
Infiltration from precipitation	2,120	920	710	830	1,030
Subsurface inflow and seepage from ephemeral streams to east side of Juab Valley	5,230	4,180	6,270	5,230	5,230
Seepage from ephemeral streams and subsurface inflow to west side of Juab Valley	1,990	1,590	2,380	1,990	1,990
Total	**16,630**	**13,530**	**13,430**	**15,340**	**15,540**
Specified discharge					
Withdrawal from wells pumped for irrigation and public supply	0	0	11,440	0	0
Subsurface outflow to consolidated rock/unconsolidated basin-fill deposits boundary	470	380	570	470	470
Total	**470**	**380**	**12,010**	**470**	**470**
Model-computed discharge					
Flowing wells	170	180	130	160	180
Springs and seeps					
Currant Creek upstream from gaging station 10146400	2,730	2,530	2,420	2,660	2,820
Currant Creek downstream from gaging station 10146400	900	880	860	890	910
Mona Reservoir	2,240	2,300	2,140	2,200	2,260
Springs and seeps total	5,870	5,710	5,420	5,750	5,990
Evapotranspiration	0	2,960	2,840	0	0
Total	**6,040**	**8,850**	**8,390**	**5,910**	**6,170**
Water going into (+) or out of (-) storage[3]	+10,400	+4,590	-6,610	+9,210	+9,140

Table 6. Specified ground-water recharge and discharge and model-computed discharge and change in storage for the seasonal transient-state simulation, Juab Valley, Utah—Continued

Budget element	Flow, in acre-feet, for each stress period (rounded)[1]				
	January 1 to March 31, 1993	April 1 to June 30, 1993	July 1 to September 30, 1993	October 1 to December 31, 1993	January 1 to March 31, 1994
Southern part of Juab Valley[2]					
Specified recharge					
Seepage from nonirrigation-season streamflow	1,460	0	0	1,460	1,460
Seepage from irrigation-season streamflow not included in unconsumed irrigation water component	0	0	0	0	0
Seepage from unconsumed irrigation water and distribution systems	0	3,200	2,080	0	0
Infiltration from precipitation	1,500	650	500	590	730
Subsurface inflow and seepage from ephemeral streams to east side of Juab Valley	810	650	970	810	810
Seepage from ephemeral streams and subsurface inflow to west side of Juab Valley	710	570	850	710	710
Total	**4,480**	**5,070**	**4,400**	**3,570**	**3,710**
Specified discharge					
Withdrawal from wells pumped for irrigation and public supply	0	0	7,110	0	0
Subsurface outflow to consolidated rock/unconsolidated basin-fill deposits boundary	0	0	0	0	0
Total	**0**	**0**	**7,110**	**0**	**0**
Model-computed discharge					
Flowing wells	310	290	280	300	310
Springs and seeps					
Palmer Spring	110	90	80	100	110
Chicken Creek Reservoir	960	890	860	930	970
Springs and seeps total	1,070	970	940	1,030	1,080
Evapotranspiration	0	910	750	0	0
Total	**1,380**	**2,180**	**1,970**	**1,330**	**1,390**
Water going into (+) or out of (-) storage[3]	+2,770	+2,560	-5,060	+1,940	+2,030

[1]Flow is based on rate computed for the last time step in the stress period.

[2]The division of the northern and southern parts of Juab Valley is based on a topographic divide and not a ground-water divide. Depending on the hydrologic conditions, simulated flow from the southern to the northern part of Juab Valley varies from about 1,000 to about 2,000 acre-feet per year.

[3]Water going into (+) storage is treated as discharge by the numerical model; water going out of (-) storage is treated as recharge by the numerical model. The difference between recharge, discharge, and storage amounts for a stress period is approximately the simulated flow from the southern to the northern part of Juab Valley.

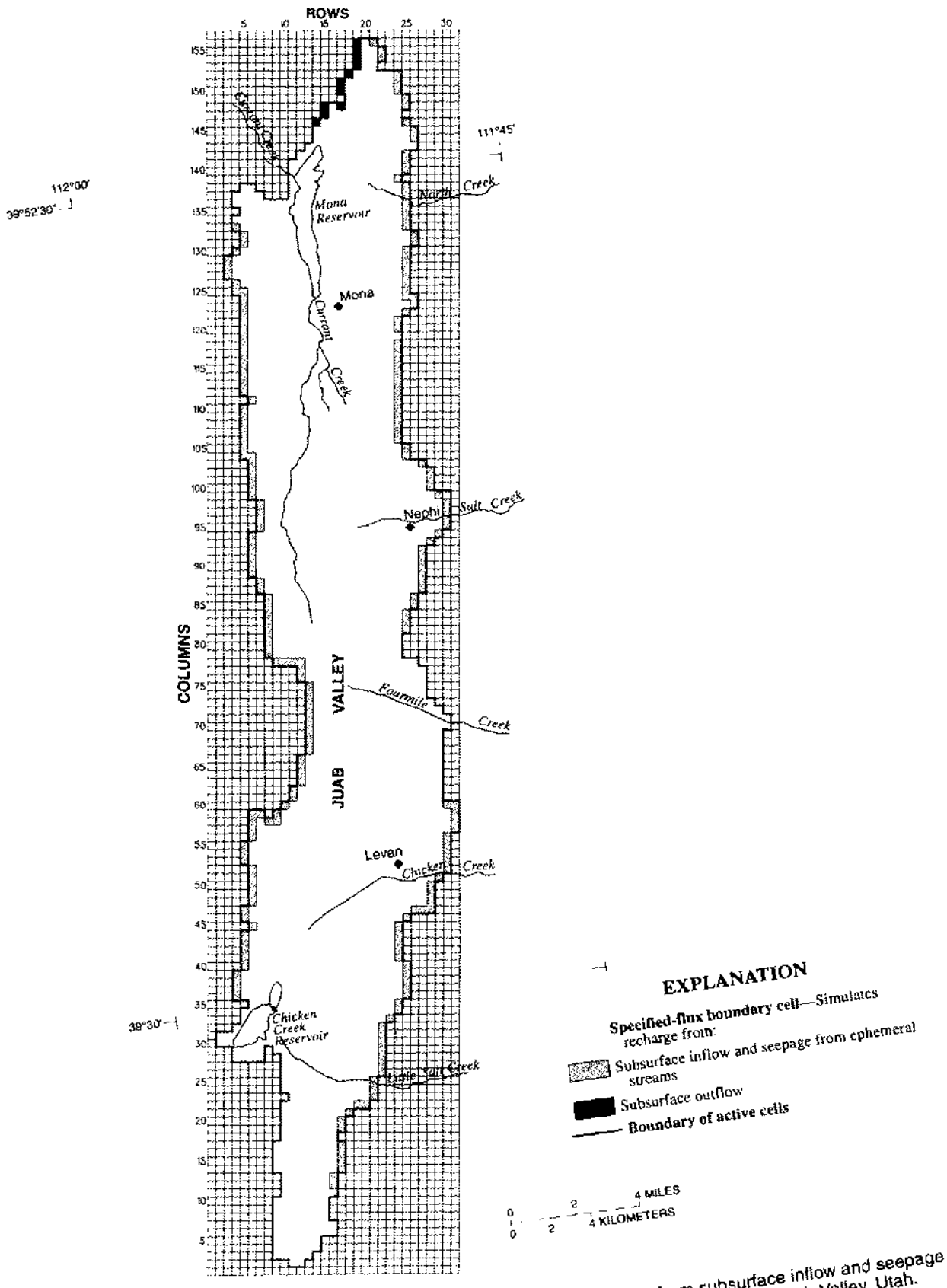


Figure 23. Location of specified-flux boundary cells that simulate recharge from subsurface inflow and seepage from ephemeral streams and discharge by subsurface outflow in the ground-water flow model of Juab Valley, Utah.

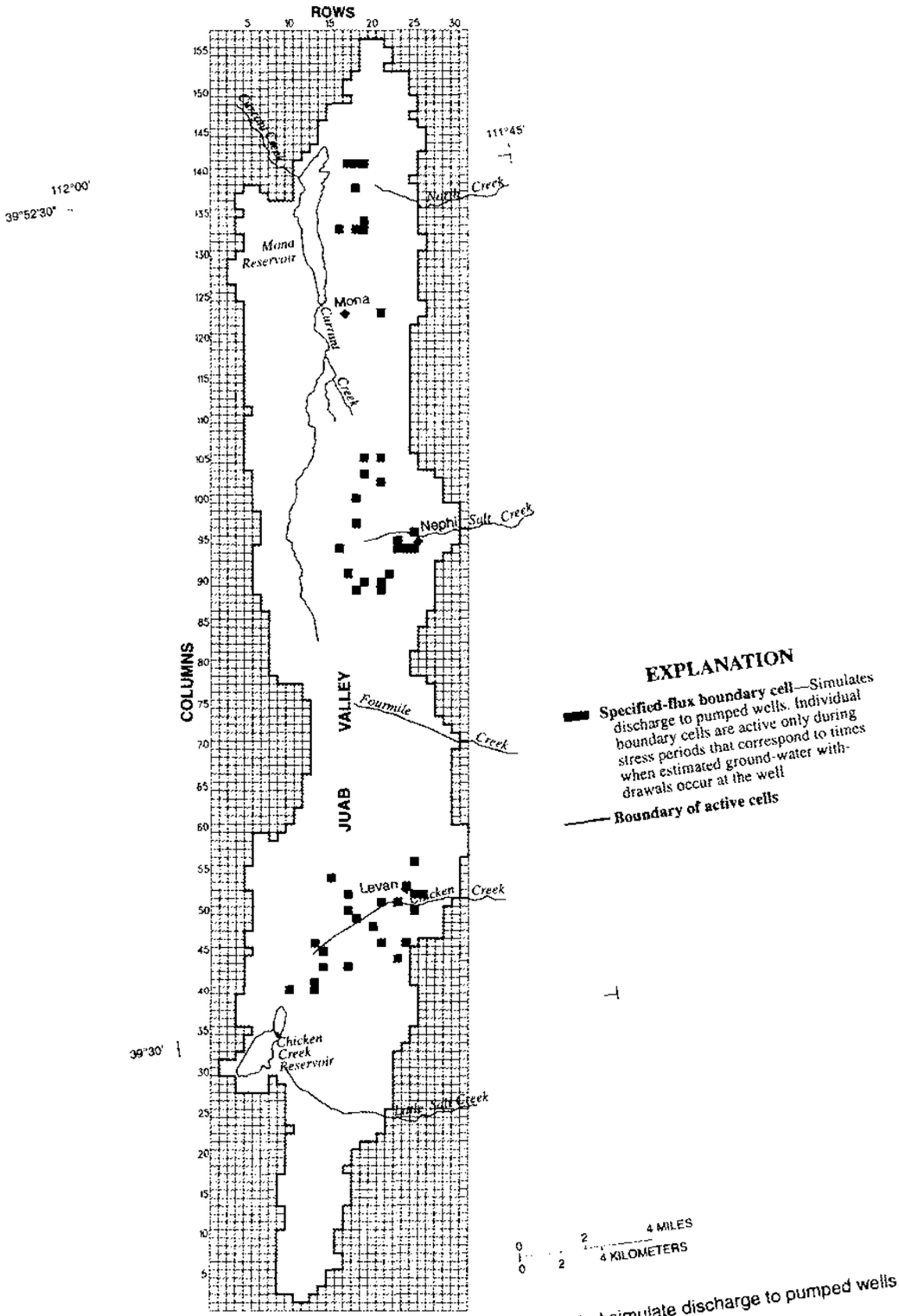


Figure 24. Location of specified-flux boundary cells that simulate discharge to pumped wells in the ground-water flow model of Juab Valley, Utah.

depth of the well. If depth information is not available, the well discharge is assigned in equal parts to all model layers. Well discharge data for each stress period were compiled and estimated from unpublished data in the files of the U.S. Geological Survey Utah District office in Salt Lake City, Utah. For the seasonal transient-state simulation, all ground-water pumpage is assumed to occur during the July through September stress period (see "Seepage from unconsumed irrigation water and distribution systems" section of this report). Well discharge for the steady-state and individual stress periods of the transient-state simulations is listed in tables 4, 5, and 6.

Ground-water discharge to flowing wells is simulated as a head-dependent flux boundary with the Drain Package (McDonald and Harbaugh, 1988, p. 9-1). Flowing-well discharge is directly proportional to the difference between the altitude of the top of the well and the altitude of the head in the well. As with pumped wells, only wells with an estimated discharge greater than 50 acre-ft/yr (30 gal/min) are simulated. Flowing wells are simulated at model cells and layers that correspond to the location and depth extent of the perforated intervals of the flowing wells (fig. 25). The perforated interval of a flowing well is measured in the same manner as for a pumped well. The Drain Package computes flow across the boundary as a function of the head difference between the model-computed water level at the center of the model cell, the drain altitude, and the drain conductance. The drain altitude assigned to flowing wells is the land-surface altitude of the center of the model block in layer 1 that corresponds to the location of the well and was determined from the appropriate U.S. Geological Survey 1:24,000-scale topographic map. This allows model-computed discharge to a flowing well to cease when the model-computed water level at the boundary is at or below land surface. Drain conductance is a calibration parameter, and final values were determined during model calibration (see "Calibration parameters" section of this report). At the start of the calibration, drain conductance was arbitrarily assumed to be 1×10^5 ft^2/d.

Discharge to springs and seeps is simulated as a head-dependent flux boundary with the Drain Package (McDonald and Harbaugh, 1988, p. 9-1). A spring or seepage area is an expression of the water table at land surface, and the amount of discharge and areal extent of the seepage area depends on local water levels. Simulated discharge to drains varies with changes in model-computed water levels, and no discharge is simulated when the water level at the drain node declines below the specified drain altitude.

The seepage area south of Burriston Ponds is represented by drains placed at model nodes in layer 1. Drains representing discharge to Burriston Ponds were placed at model nodes in layer 1 that correspond to the location of the ponds and at underlying model nodes in layers 2 and 3. These drains are simulated in the upper three model layers to account for ground water that has ponded at land surface and water that has moved through the deeper parts of the ground-water system. Gravity data indicate a gravity high in the area of Burriston Ponds (Zoback, 1992, p. E5). This change in the geologic structure of the valley may result in flow along a preferential path to the springs. The discharge to Burriston Ponds and the seepage area south of Burriston Ponds is referred to as "Currant Creek upstream from gaging station 10146400" in tables 2, 4, 5, and 6. The location of these drains is shown in figure 25 as springs and seepage areas in the northern part of Juab Valley. The drain representing Palmer Spring, a surface feature in the southern part of the valley, is placed in model layers 3 and 4 to simulate water discharging from a deeper part of the unconsolidated basin-fill deposits. The west side of Juab Valley is bounded by an east-dipping fault zone. Ground water that has moved downgradient from the east side of the valley could intersect the fault zone at depth and flow along it toward land surface and Palmer Spring. Simulating these springs as drains in multiple layers is analogous to simulating flowing wells as drains with perforated intervals in multiple layers.

Drain altitude and drain conductance are calibration parameters and their final values were determined during model calibration. The drain altitude was initially set to 5 ft below the land-surface altitude of the center of the model block and was determined from the appropriate U.S. Geological Survey 1:24,000-scale topographic maps. This was done to compensate for channelization within the seepage area or depressions near the spring. The initial estimate of drain conductance was arbitrarily assumed to be 1×10^5 ft^2/d.

Discharge to Currant and West Creeks is simulated as a head-dependent flux boundary using the Streamflow Package (Prudic, 1989). Seepage studies show that a hydrologic connection exists between the creeks and the underlying ground-water system. The placement of streamflow boundaries corresponds to the locations of Currant and West Creeks and is shown in figure 25. The Streamflow Package computes flow across the boundary as a function of the head difference

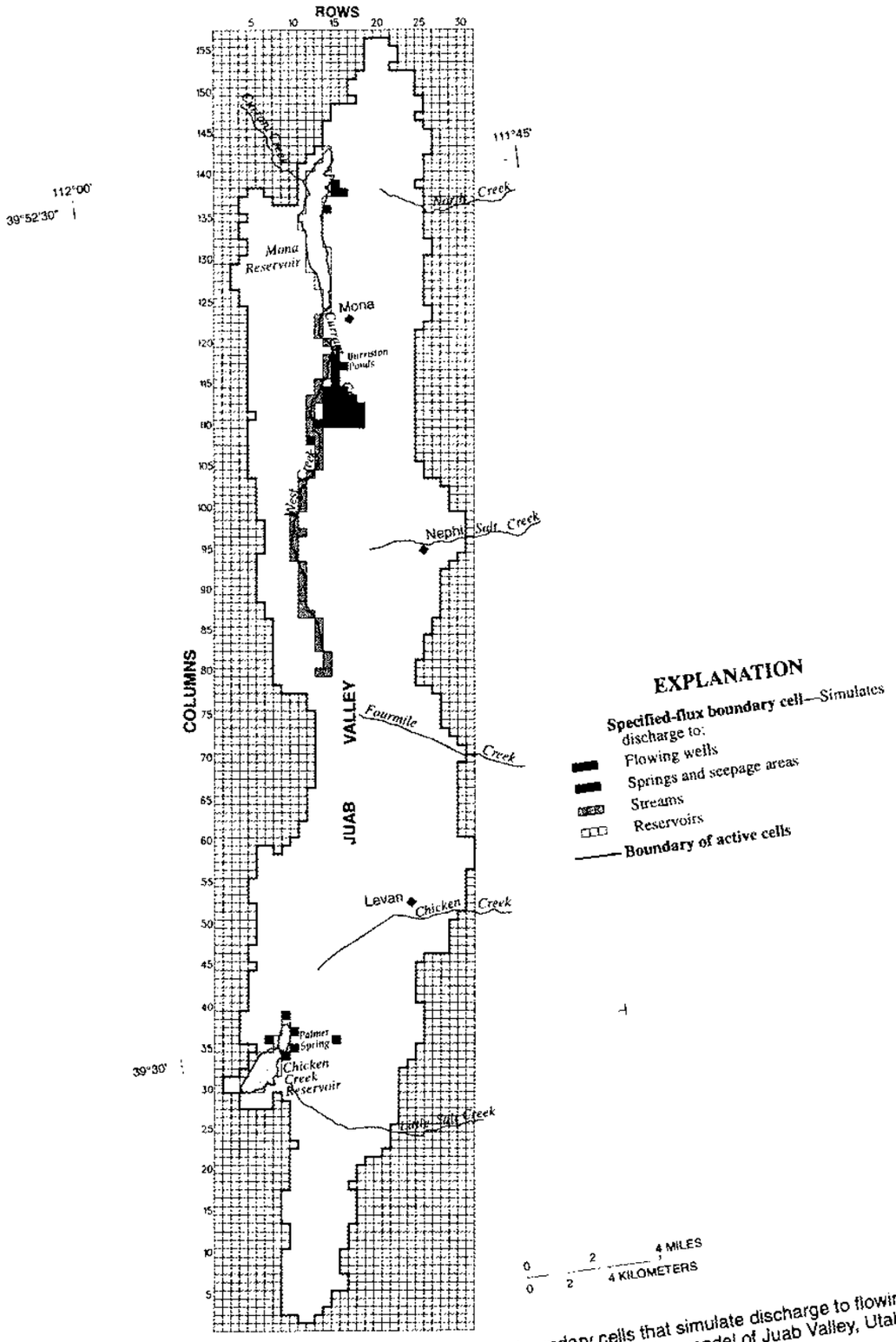


Figure 25. Location of head-dependent flux boundary cells that simulate discharge to flowing wells, springs, seeps, streams, and reservoirs in the ground-water flow model of Juab Valley, Utah.

between the model-computed water level at the center of the model cell and the water-surface altitude of the stream (hereafter referred to as stream stage) and the streambed conductance. The package accounts for changes in streamflow caused by seepage to and from the stream, tributary inflows, and diversions and can compute stream stage as a function of streamflow. The stream-stage option was not used, however, because data are not available to describe the relation between streamflow and stage. At boundaries formulated by the Streamflow Package, discharge occurs when model-computed water levels rise above stream stage. Required input data for the Streamflow Package include the following physical characteristics: altitude of the top and bottom of the streambed, stream stage, streambed width and length, and hydraulic conductivity of the streambed.

Average streambed altitude was determined from the appropriate U.S. Geological Survey 1:24,000-scale topographic maps, which on the basis of contour interval, are considered to be accurate to plus or minus 10 ft. The top of the streambed was set initially at 10 ft below land surface to compensate for stream channelization. Average streambed altitude is a calibration parameter. Stream stage for Currant and West Creeks is arbitrarily set at 3 and 1 ft above the top of the streambed, respectively. The bottom of the streambed is arbitrarily set at 1 ft below the top of the streambed; data are not available to determine this depth quantitatively. In the Streamflow Package, when model-computed ground-water levels are below the bottom of the streambed, the stream is no longer in direct contact with the ground-water system and water from the stream moves through the streambed to the ground-water system at a unit gradient (McDonald and Harbaugh, 1988, p. 6-6). On the basis of visual observations, the width of Currant and West Creeks was assumed to be 20 and 10 ft, respectively. Streambed length within individual model cells for West Creek was approximated to equal the length of the cell, 1,320 ft. Streambed length within individual model cells for Currant Creek was determined from the appropriate U.S. Geological Survey 1:24,000-scale topographic maps. Hydraulic conductivity of the streambed defines the ease with which water can move through the streambed. This property has not been measured for these streams and is a calibration parameter (see "Calibration parameters" section of this report). Hydraulic conductivity of the streambed was set at 10 ft/d for Currant Creek and 1 ft/d for West Creek at the start of the calibration process.

Discharge to springs and seeps at Mona and Chicken Creek Reservoirs is simulated as a head-dependent flux boundary by assuming that reservoir stage remains constant throughout the simulations. Discharge to the reservoirs is dependent on the difference between water levels in the ground-water system underlying the reservoir and the water-surface altitude in the reservoir. The Constant-Head Node option (McDonald and Harbaugh, 1988, p. 4-2) is assigned to model nodes in layer 1 that correspond to the areal extent of Mona and Chicken Creek Reservoirs (fig. 25). Discharge occurs across the boundary if water levels in underlying model nodes are higher than the water level in the reservoir, and recharge occurs if water levels are lower than the water level in the reservoir. The altitude assigned to the constant-head nodes representing Mona and Chicken Creek Reservoirs is 4,877 and 5,050 ft, respectively.

Discharge by evapotranspiration is simulated as a head-dependent flux boundary using the Evapotranspiration Package (McDonald and Harbaugh, 1988, p. 10-1). Discharge by evapotranspiration is dependent on depth to the water table and is, therefore, a head-dependent process. Cells simulating evapotranspiration are shown in figure 26 and correspond to areas mapped as wetlands during 1988 (pl. 1). The simulation of evapotranspiration is based on the assumption of a linear change between a maximum evapotranspiration rate when the water level is at or above land surface, to an evaporation rate of zero when the water level is below a specified extinction depth.

Input parameters for the Evapotranspiration Package are the evapotranspiration surface altitude, the extinction depth, and the maximum evapotranspiration rate. The evapotranspiration surface altitude is equal to land surface at the boundary. The extinction depth is set, subjectively, to 10 ft below land surface and represents an average extinction depth for the principal phreatophytes in Juab Valley. The estimated maximum evapotranspiration rate is set at 3.3 ft/yr for all phreatophytes in Juab Valley, which is about 160 percent of the estimated average evapotranspiration rate of 2.1 ft/yr discussed in the "Evapotranspiration" section of this report. The rate of 3.3 ft/yr is the average of the maximum rates for meadow grasses and salt grass, 2.6 ft/yr and 4.0 ft/yr, respectively (Robinson, 1958, p. 18 and 75). For the seasonal transient-state simulation, the evapotranspiration rate is set at zero for the October through December and January through March stress periods. This assumes that no plant growth occurs during the fall and winter months.

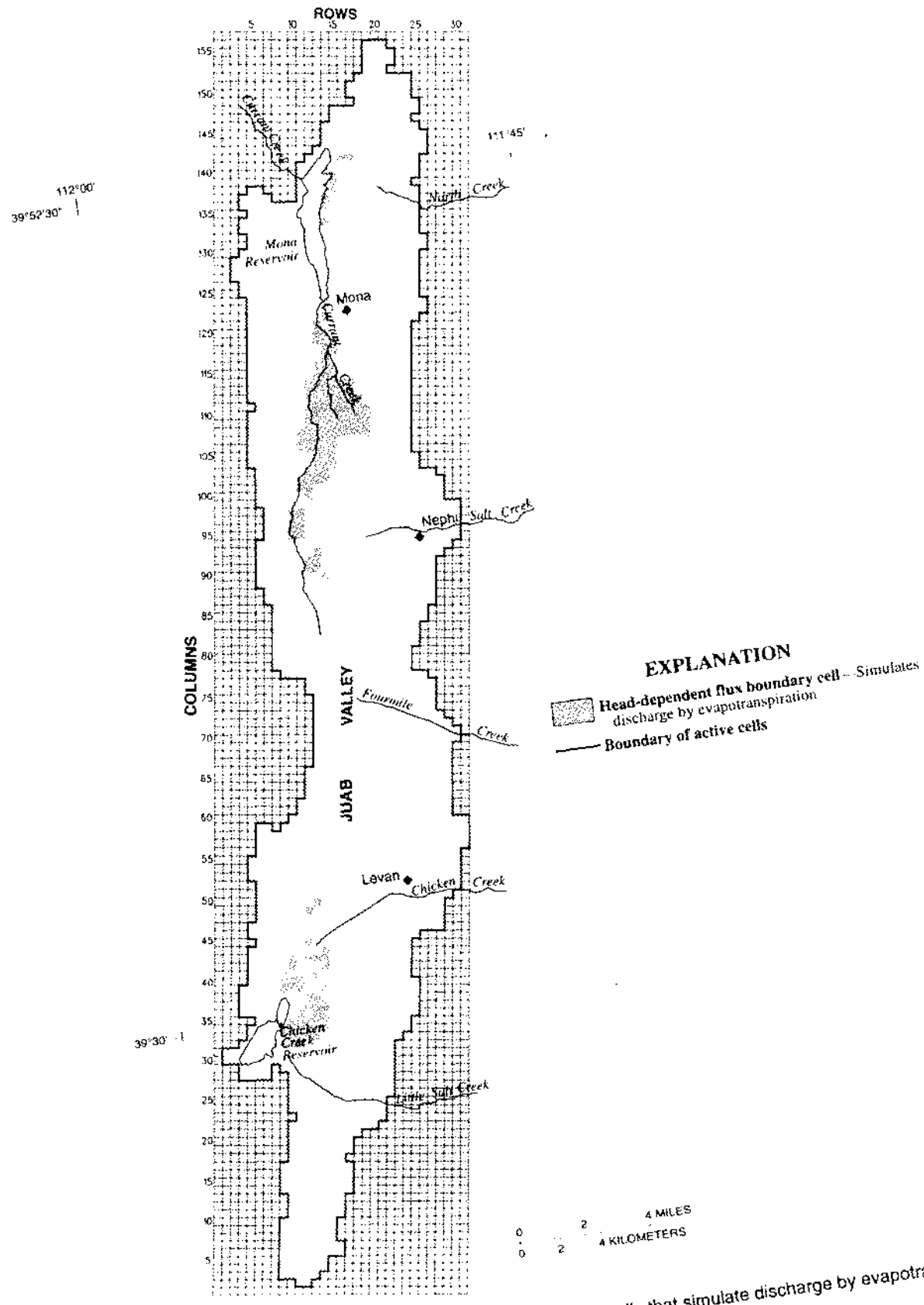


Figure 26. Location of head-dependent flux boundary cells that simulate discharge by evapotranspiration in the ground-water flow model of Juab Valley, Utah.

Discharge by subsurface outflow is simulated as a specified-flux boundary using the Well Package (McDonald and Harbaugh, 1988, p. 8-1). Data are not available to quantify subsurface outflow from the unconsolidated basin-fill deposits, so a simplistic specified-flux approach was used to simulate this component of discharge. Subsurface outflow is limited to the consolidated rocks along the northeastern part of Long Ridge. This discharge is represented by 40 pumping wells placed in model layers 1, 2, 3, and 4 at the model nodes along the western extent of the saturated unconsolidated basin-fill deposits of Juab Valley north of Mona Reservoir (fig. 23). Discharge by subsurface outflow is evenly distributed to the pumping wells. At the start of the calibration process, subsurface outflow was set at 1,000 acre-ft/yr (table 2). For the seasonal transient-state simulation, the annual specified discharge from subsurface outflow is partitioned as 25, 20, 30, and 25 percent, respectively, for the January through March, April through June, July through September, and October through December stress periods. These percentages are based on reported seasonal variations in discharge from Bradley Spring and selected springs and spring-fed drainages in the southern Wasatch Range that discharge from consolidated rocks. The amount of subsurface outflow is a calibration parameter and was determined during model calibration (see "Calibration parameters" section of this report).

Hydrologic Properties

Model parameters that simulate the hydrologic properties of the saturated unconsolidated basin-fill deposits are horizontal hydraulic conductivity, transmissivity, vertical leakance, primary storage coefficient, and secondary storage coefficient. Transmissivity and vertical leakance each incorporate model-layer thickness and hydraulic conductivity into a single term. Storage coefficients, depending on the model-layer specifications, are either equated to specific yield or to the product of specific storage and model-layer thickness. These hydrologic parameters, along with the boundary conditions, determine the amount and pattern of model-computed ground-water flow between nodes and across boundaries of the numerical model.

Horizontal hydraulic conductivity and transmissivity are two model parameters that help control simulated horizontal ground-water flow in the saturated unconsolidated basin-fill deposits. Estimates determined from specific-capacity values, slug tests, and aquifer tests for selected areas of the valley (see "Hydrologic properties" section of the conceptual system in this report) were used to limit the range of hydraulic-conductivity and transmissivity values used in the model. A horizontal hydraulic-conductivity value is required for all active model nodes in model layer 1. At the start of the calibration process, hydraulic conductivity was set to 100 ft/d near the mountains and 20 ft/d in the center of the valley. The larger initial estimates of conductivity near the mountains reflect the generally coarse-grained nature of the unconsolidated basin-fill deposits found in these areas on the basis of lithologic descriptions in drillers' logs. Model layer 1 simulates unconfined ground-water flow. Transmissivity, the product of horizontal hydraulic conductivity and model-layer thickness, is used to compute confined horizontal ground-water flow and is required model input for layers 2, 3, and 4. These model layers simulate confined ground-water flow but are allowed to become locally unconfined when the water level in the model cell drops below the top of the layer. At the start of the calibration process, the transmissivity for each of model layers 2 and 3 is the product of the initial horizontal hydraulic-conductivity value assigned to model layer 1 multiplied by the thickness of the respective layer.

The thickness of unconsolidated basin-fill deposits in Juab Valley, and therefore layer 4, is not known. On the basis of the structure of other valleys in the Basin and Range Province, the unconsolidated deposits of Juab Valley are assumed to be thicker in the center than along the margins of the valley. As a result of consolidation, deeper deposits in the interior part of the valley are assumed to have a smaller hydraulic conductivity than deposits closer to the mountain fronts. Thus, as the thickness of layer 4 increases away from the mountain fronts and the hydraulic conductivity decreases, the transmissivity would tend to be relatively constant throughout the interior part of the valley. Near the edge of the valley, transmissivity can be large because of the well-sorted coarse-grained deposits near the canyon mouths. At the start of the calibration process, transmissivity values for model layer 4 were 25,000 ft^2/d near the mountains and 5,000 ft^2/d in the center of the valley. The final distribution and values of horizontal hydraulic conductivity and transmissivity were determined during model calibration (see "Calibration parameters" section of this report).

Vertical leakance is the model parameter that helps control ground-water flow between the model layers that represent the saturated unconsolidated basin-fill deposits. Vertical leakance is part of the con-

ductance term for vertical ground-water flow and is determined from the individual vertical hydraulic-conductivity value divided by the thickness of all geohydrologic units that exist between vertically adjacent model nodes (McDonald and Harbaugh, 1988, p. 5-11). Vertical conductance is vertical leakance multiplied by the model cell area. Vertical hydraulic conductivity of the unconsolidated basin-fill deposits in Juab Valley has not been quantified except for a site in the southern part of the valley (see "Hydrologic properties" section of the conceptual system in this report). For this simulation, it is assumed that vertical hydraulic conductivity is the same throughout the total thickness of the saturated unconsolidated basin-fill deposits. At the start of the calibration process, all vertical-leakance values were based on vertical hydraulic-conductivity values of 1 ft/d near the mountains and 0.1 ft/d in the center of the valley. The final distribution and values of vertical leakance were determined during model calibration (see "Calibration parameters" section of this report).

Primary and secondary storage coefficients are model parameters used to simulate the amount of water released from and placed into ground-water storage as a result of water-level changes. The primary storage coefficient for model layer 1 is equivalent to specific yield. For model layers 2, 3, and 4, the primary storage coefficient is equivalent to the confined storage coefficient. Because model cells in layers 2, 3, and 4 may become unconfined, a secondary storage coefficient is required for these layers. This is equivalent to specific yield and is used only when the cells become unconfined (when the water level in a model cell drops below the top of the cell). At the start of the calibration process, the primary storage coefficient for model layer 1 was assigned a value of 0.20. The initial value for the primary storage coefficient for model layers 2 and 3 is 6×10^{-5} and is based on a specific storage of 1×10^{-6} ft^{-1} (specific storage is defined as the storage coefficient divided by model-layer thickness). The initial value for the primary storage coefficient for model layer 4 is 2.5×10^{-4}. The secondary storage coefficient for model layers 2, 3, and 4 was assigned an initial value of 0.20, which is identical to the specific yield of layer 1. The initial values for primary and secondary storage coefficients were set arbitrarily, and the final distributions and values were determined during model calibration (see "Calibration parameters" section of this report).

Model Calibration

The numerical model was calibrated to observed ground-water levels to simulate the steady-state conditions of 1949, multi-year transient-state conditions during 1949-92, and seasonal transient-state conditions during 1992-94. The 1949 steady-state period represents the hydrologic system in Juab Valley prior to the start of large-scale ground-water withdrawal from wells. The result of the 1949 steady-state simulation is the initial condition for the transient-state simulation. The 1949-94 transient-state conditions represent a period of extensive ground-water development and other large changes in hydrologic stress. Initially, the numerical model was roughly calibrated to the estimated ground-water budget listed in table 2 and the average water level at 191 wells. The average water level is computed from the water-level data available for 1935-94 for each well. Data are not available to estimate individual components of discharge for 1949-94 except for ground-water pumped from wells; therefore, the steady-state and transient-state simulations could not be calibrated to match discharge from known sources. Calibration parameters were adjusted until the numerical model obtained a reasonable match with water levels measured at steady-state conditions in 1949 and transient-state conditions during 1949-94.

Calibration Parameters

The parameters adjusted during the calibration process, listed in general order of importance, are horizontal hydraulic conductivity and transmissivity, spatial distribution and amount of recharge from injection wells that simulate subsurface inflow and seepage from ephemeral streams to the east side of Juab Valley, vertical hydraulic conductivity, drain conductance, streambed hydraulic conductivity for Currant and West Creeks, specific yield, the amount of recharge from injection wells that simulates subsurface inflow and seepage from ephemeral streams to the west side of Juab Valley, the amount of discharge from subsurface outflow, and storage coefficient. Generally, horizontal hydraulic conductivity was adjusted within a range determined by specific-capacity values, slug tests, and aquifer tests to attain a reasonable match between measured and model-computed water levels. The spatial distribution of recharge from subsurface inflow and seepage from ephemeral streams to the east side of Juab Valley was adjusted to match measured and model-computed water levels in local areas where large

amounts of ground-water pumpage occurred. Vertical hydraulic conductivity was adjusted to allow model-computed flows in lower layers to discharge at head-dependent flux boundaries in model layer 1. Drain conductance and streambed hydraulic conductivity were adjusted to best simulate measured water levels near flowing wells, springs, seeps, and streams. The amount of recharge from subsurface inflow and seepage from ephemeral streams to the east and west sides of Juab Valley, the amount of discharge from subsurface outflow, specific yield, and storage coefficient were adjusted to best simulate measured water-level changes through time.

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

During the 1949-94 transient-state period, the amount of ground water pumped from wells varied from about 500 acre-ft/yr in 1949 to an average of 26,000 acre-ft/yr for 1989-93. Ground water generally is used to supplement the supply of surface water for irrigation. A change in the amount of natural discharge, water in storage, and (or) recharge must have occurred in order for the system to approach a new state of equilibrium. Water-level changes observed in selected wells show no large declines associated with ground-water development in the valley (figs. 10 and 11). The values assigned to horizontal and vertical hydraulic conductivity, drain conductance, and streambed hydraulic conductivity were set so that relatively small changes in model-computed water levels correspond to relatively large variations in model-computed discharge to springs, seeps, and evapotranspiration. This was done to ensure that simulated increases in ground-water pumpage primarily would affect natural discharge from the system and would affect model-computed water levels to a lesser degree.

Horizontal hydraulic conductivity of model layers 1, 2, and 3 was varied spatially but not vertically from 1 to 200 ft/d during model calibration. The transmissivity of model layer 4 was varied from 200 to 70,000 ft^2/d. The corresponding range of transmissivity for the simulated unconsolidated basin-fill deposits (model layers 1, 2, 3, and 4) is about 400 to 105,000 ft^2/d. This is consistent with the range of transmissivity values presented in the "Hydrologic properties" section of the conceptual system in this report. The transmissivity of model layers 1, 2, and 3 can be computed by multiplying horizontal hydraulic-conductivity values by a thickness of 50, 60, and 60 ft, respectively. The actual thickness of layer 4 is not known, but it is assumed that horizontal hydraulic conductivity decreases as the thickness of layer 4 increases. The final distribution of horizontal hydraulic conductivity for model layers 1, 2, and 3 varies from 1 to 125 ft/d and is shown in figure 27. The final distribution of transmissivity for model layer 4 varies from 280 to 36,000 ft^2/d and is shown in figure 28. Horizontal hydraulic conductivity and transmissivity were determined mainly by adjusting values until a reasonable match between measured and model-computed water levels was obtained.

To match water levels near the mountains, horizontal hydraulic-conductivity values were assigned that generally range from 1 to 10 ft/d. The unconsolidated basin-fill deposits in these areas generally are coarse grained, and the relatively smaller horizontal hydraulic conductivity reflects poor sorting of those deposits. In areas near Nephi and Levan, the unconsolidated basin-fill deposits are simulated with horizontal hydraulic-conductivity values that were varied from 60 to 125 ft/d. Values are larger in these areas because the unconsolidated basin-fill deposits have probably been reworked and sorted by Salt Creek and Chicken Creek. Horizontal hydraulic conductivity and transmissivity values between major ground-water recharge areas near the mountain front and natural discharge areas in the lower parts of the valley were increased to create a good hydraulic connection. This was done to get a large variability in model-computed discharge to springs, seeps, and reservoirs as a result of ground-water withdrawal from wells near Nephi and Levan and also to provide an acceptable agreement between measured and model-computed water levels.

The amount of recharge from subsurface inflow and seepage from ephemeral streams to the east side of Juab Valley was varied from about 10,000 to 15,000 acre-ft/yr during the steady-state calibration (the minimum value for estimated average conditions is 13,700 acre-ft/yr, the sum of values presented for the northern and southern parts of Juab Valley in table 2) to match measured water levels representing the steady-state conditions assumed for 1949. The amount established during model calibration for 1949 is 12,100 acre-ft/yr (the sum of values presented for the northern and southern parts of the valley in table 4). This amount is considered to be about 95 percent of estimated average

conditions, in accordance with the ratio of precipitation in 1949 to the 1931-93 average precipitation. With this method, recharge from subsurface inflow and seepage from ephemeral streams to the east side of the valley under estimated average conditions is calculated to be 12,700 acre-ft/yr and is used as the basis for the transient-state simulation. For the multi-year and seasonal transient-state simulations, recharge from subsurface inflow and seepage from ephemeral streams to the east side of the valley was varied by the ratio of precipitation during the given stress period to the 1931-93 average precipitation.

To match the trend and to better approximate the magnitude of measured water-level changes in both the multi-year and the seasonal transient-state simulations, it was necessary to increase recharge from subsurface inflow and seepage from ephemeral streams during periods of greater-than-average precipitation. This is based on the assumption that the percentage of precipitation that recharges the consolidated-rock ground-water system increases with an increase in precipitation and is described by the equation

$$S_{stress\ period} = S_{average}\{[(P_{stress\ period} / P_{average}) - 1] \times 4 + 1\} \quad (5)$$

where

S = recharge from subsurface inflow and seepage from ephemeral streams and

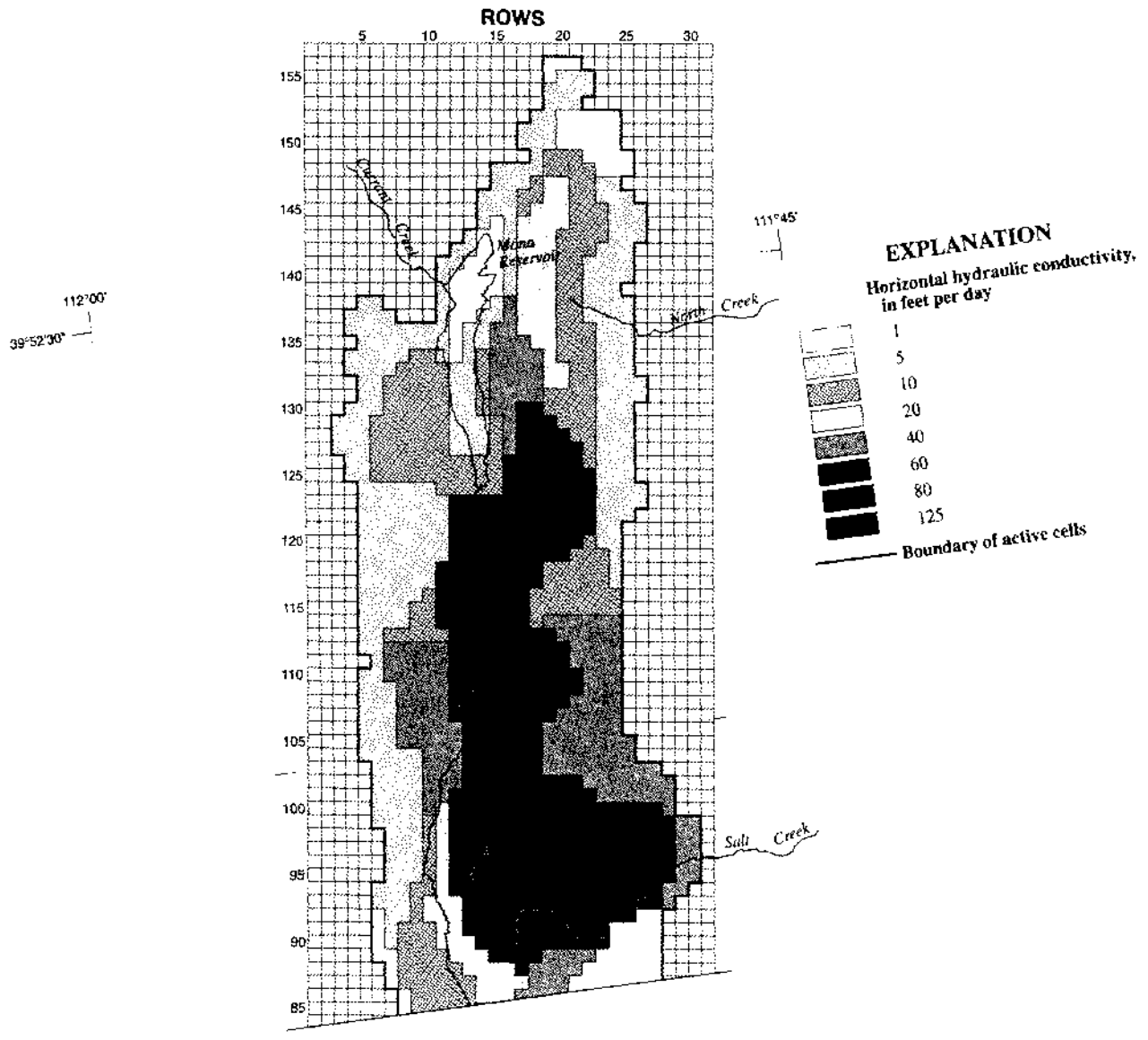
P = precipitation.

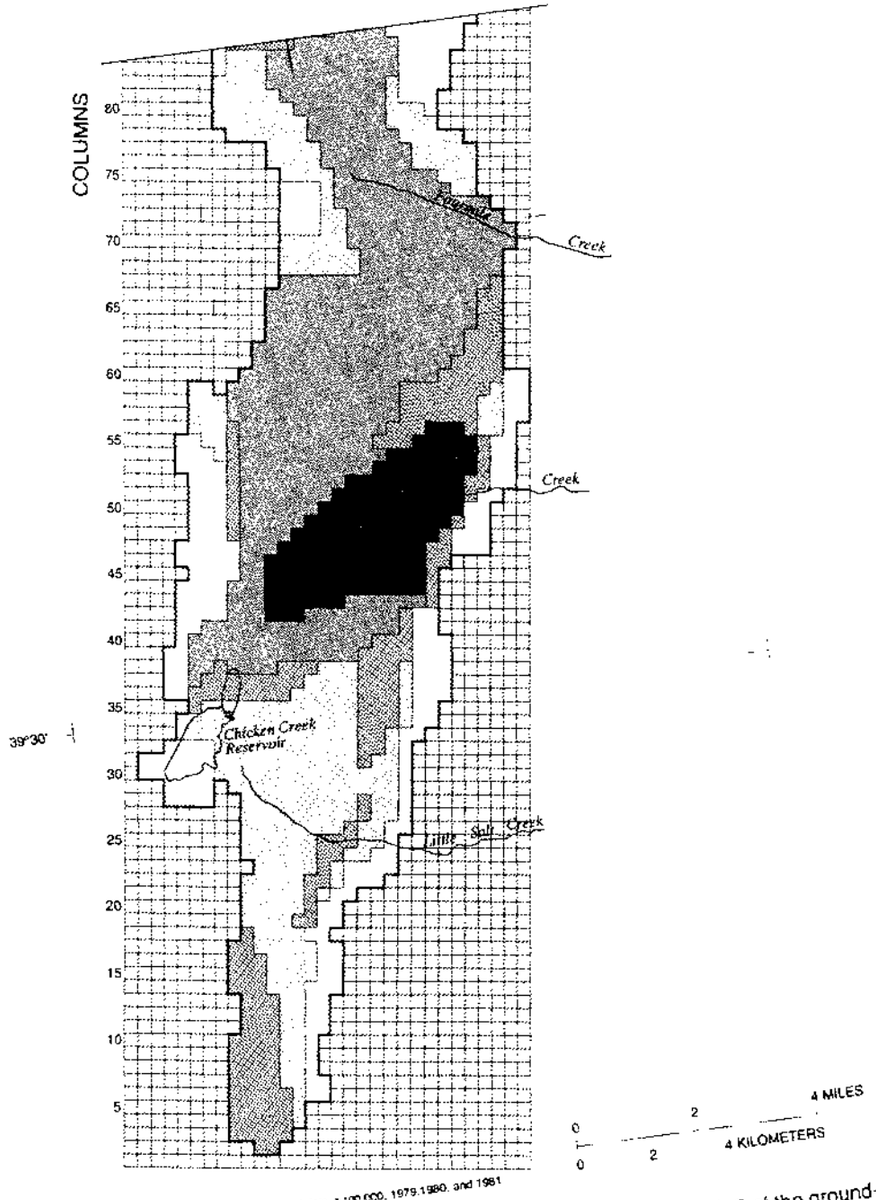
The coefficient equal to 4 in this equation was determined during the calibration process. Recharge has been varied as a function of the ratio of annual precipitation to average annual precipitation and a coefficient in other areas of the State (Holmes and Thiros, 1990, p. 47, and Lambert, 1995, p. 83).

Analysis of water budgets estimated for the mountain drainages east of Juab Valley indicates that sufficient water is available to satisfy the amount of recharge from subsurface inflow and seepage from ephemeral streams simulated during a period of greater-than-average precipitation, such as 1983-86. On the basis of measured water-level changes, recharge to the ground-water system can increase drastically during periods of greater-than-average precipitation but does not seem to be significantly less than average during periods of less-than-average precipitation. For periods of less-than-average precipitation, the actual ratio of precipitation during the given stress period to the 1931-93 average precipitation was used. Recharge to the ground-water system is probably much greater than average during periods of greater-than-average precipitation because evaporation, consumptive use of water by plants, and soil-moisture retention do not increase beyond certain values depending on the area and the plant and soil types. The method used to calculate recharge from subsurface inflow and seepage from ephemeral streams to the east side of Juab Valley accounts for this variability in recharge. Recharge specified in the model from subsurface inflow and seepage from ephemeral steams to the east side of Juab Valley ranges from 11,300 acre-ft/yr in 1954-59 to 29,700 acre-ft/yr in 1983-86 (the sum of values presented for the northern and southern parts of the valley in table 5) and was evenly distributed to each of the four layers. There is insufficient data available to verify these amounts and significant uncertainty exists with this component of recharge.

About 35 percent of the simulated recharge from subsurface inflow and seepage from ephemeral streams to the east side of the northern part of Juab Valley is distributed to the area east of Mona Reservoir. This amount of recharge is required to simulate measured and estimated discharge to pumped wells in the area, to Currant Creek upstream and downstream from gaging station 10146400, and to Mona Reservoir, without causing water-level declines much greater than measured declines. The mountains east of Mona mainly consist of fractured carbonate rocks and potentially can be a large source of subsurface inflow. About 50 percent of the simulated recharge from subsurface inflow and seepage from ephemeral streams to the east side of the northern part of Juab Valley is distributed to the area directly east of Nephi, near where Salt Creek enters the valley. The largest measured transmissivity values in Juab Valley occur in this area, and it is also an area with large amounts of ground-water pumpage. To match water levels measured in this area, a large amount of recharge from subsurface inflow was required. About 75 percent of the simulated recharge from subsurface inflow and seepage from ephemeral streams to the east side of the southern part of Juab Valley is distributed to the area east of Levan, near where Chicken Creek enters the valley. Again, this is an area with large amounts of ground-water pumpage and this amount of recharge from subsurface inflow was required to match measured local water levels.

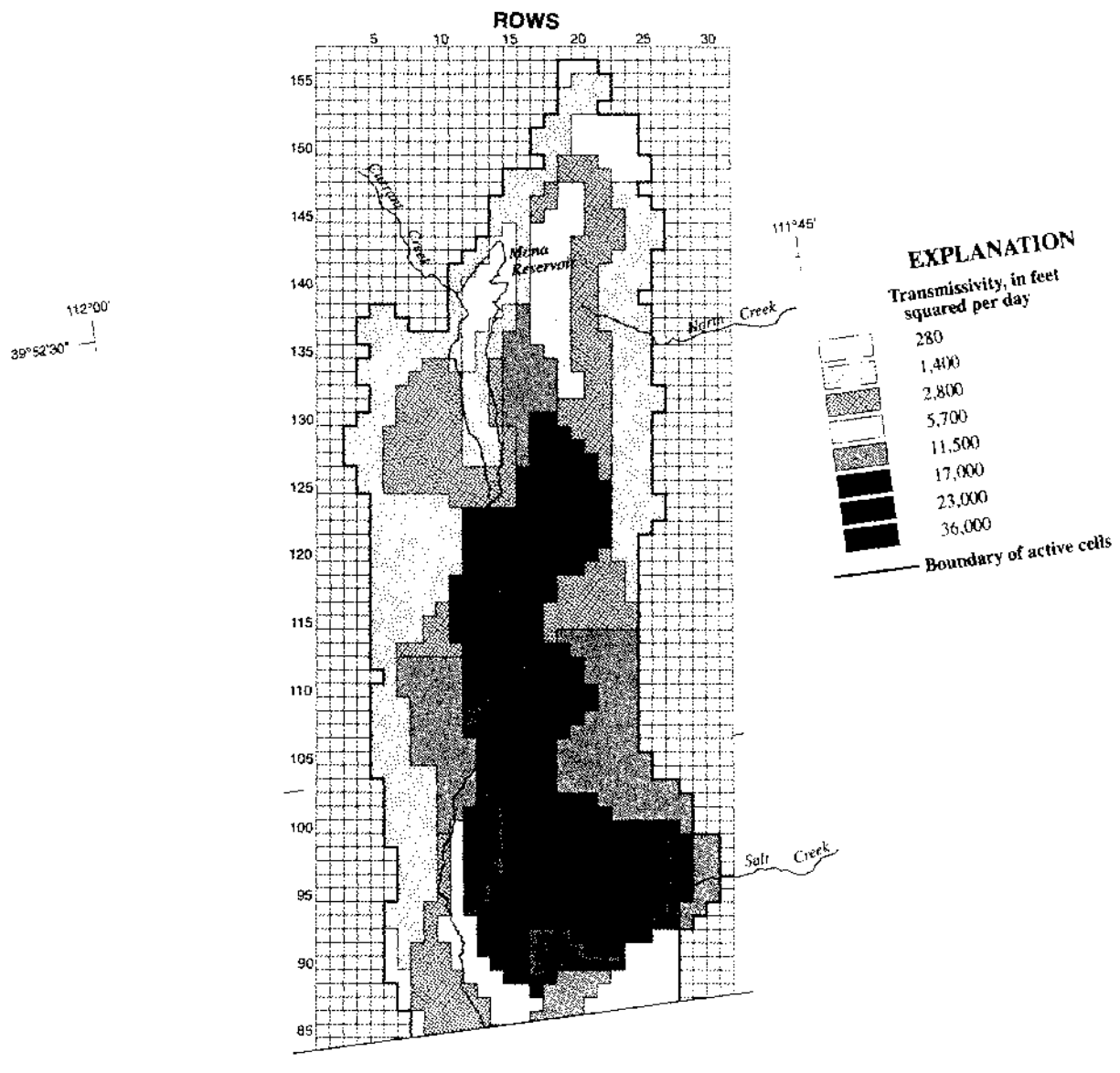
Vertical hydraulic conductivity was varied from 0.001 to 2 ft/d during model calibration. The final distribution of vertical hydraulic conductivity varies from 0.1 to 1 ft/d and is shown in figure 29. A vertical hydraulic-conductivity value of 1 ft/d was assigned to areas near the mountains and in most of the southern

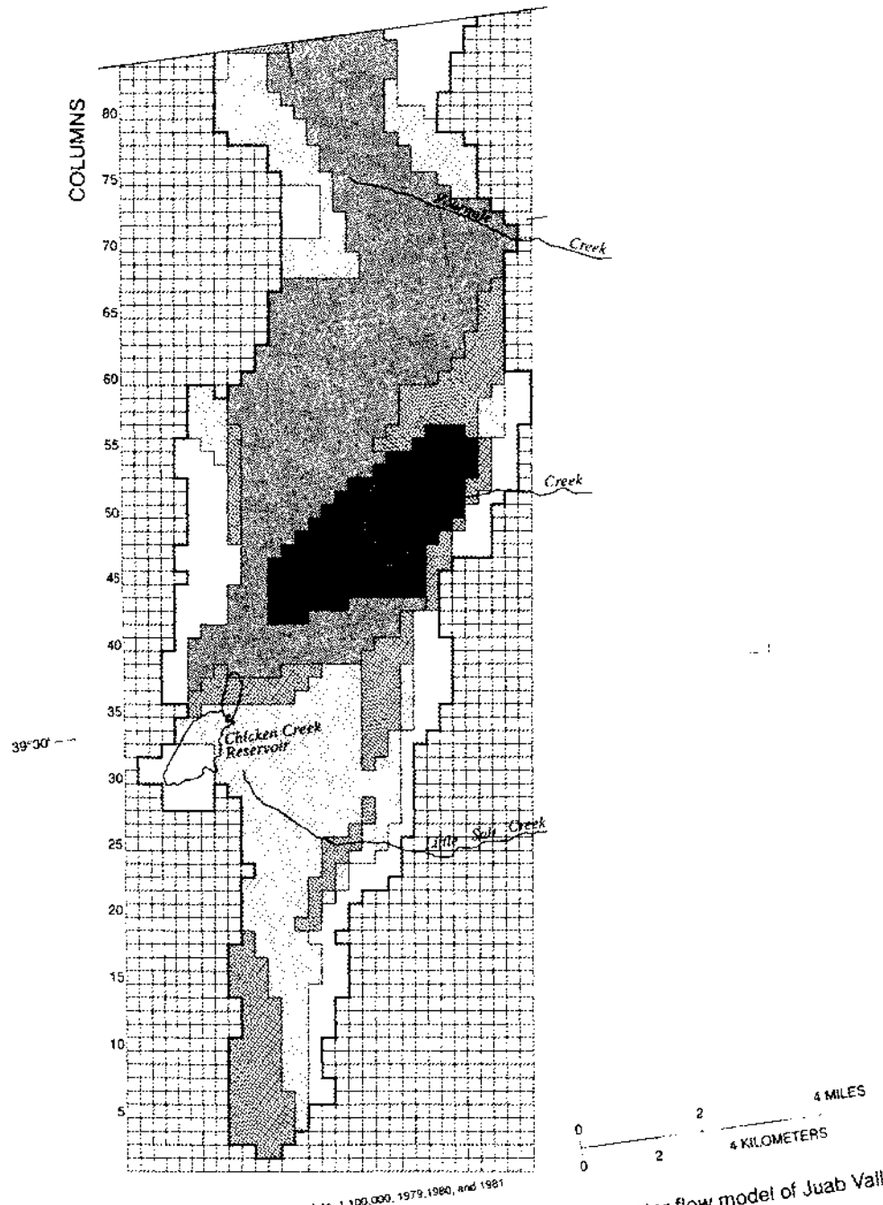

ROWS
5
10
15
20
25
30
155
150
145
140
135
130
125
120
115
110
105
100
95
90
85
Cottonwood Creek
Nicna Reservoir
North Creek
Salt Creek
112°00'
39°52'30"
111°45'
EXPLANATION
Horizontal hydraulic conductivity, in feet per day
1
5
10
20
40
60
80
125
Boundary of active cells



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

Figure 27. Final distribution of horizontal hydraulic conductivity for layers 1 through 3 of the ground-water flow model of Juab Valley, Utah.


ROWS
5
10
15
20
25
30
155
150
145
140
135
130
125
120
115
110
105
100
95
90
85
Cedar Creek
Mona Reservoir
North Creek
Salt Creek
112°00'
39°52'30"
111°45'
EXPLANATION
Transmissivity, in feet squared per day
280
1,400
2,800
5,700
11,500
17,000
23,000
36,000
Boundary of active cells



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

Figure 28. Final distribution of transmissivity for layer 4 of the ground-water flow model of Juab Valley, Utah.

part of Juab Valley where confining layers are thin or not present. Values of 0.1 and 0.5 ft/d were assigned in the north-central part of Juab Valley (west and north of Nephi), and to the lower parts of the southern part of Juab Valley. This was done to simulate the effects of confining layers in the area. A vertical hydraulic-conductivity value of 1 ft/d was assigned to the area underlying Mona Reservoir so that ground water can discharge easily to the reservoir.

Drain conductance was varied from 1×10^4 to 1×10^7 ft^2/d during model calibration. The range was arbitrarily chosen because data are not available to estimate the hydraulic conductivity of the interface between the unconsolidated basin-fill deposits and the flowing wells, springs, and seeps modeled as drains. Drain-conductance values of 5×10^4 and 8×10^5 ft^2/d were assigned to model cells that simulate flowing wells that are estimated to discharge less than and more than 80 acre-ft/yr, respectively. A drain conductance of 1×10^6 ft^2/d was assigned to model cells that simulate Burriston Ponds, the seepage areas south of Burriston Ponds, and Palmer Spring. These values provide an acceptable agreement between measured and model-computed water levels and estimated average discharge and model-computed discharge to flowing wells and springs. Ground-water pumped from wells near Nephi and Levan and large drain conductances result in a large variability during 1949-93 in simulated discharge to springs, seeps, and reservoirs. Drain altitude was adjusted up to plus or minus 10 ft, the accuracy of the contours from which the drain altitude was initially determined.

Hydraulic conductivity of the Currant and West Creek streambeds was varied from 1 to 1,000 ft/d during model calibration. Data are not available to determine the hydraulic conductivity of the streambeds directly, and the range is based on average values for streambed material composed of silty sand to clean sand (Freeze and Cherry, 1979, p. 29). Streambed hydraulic-conductivity values of 30 ft/d for Currant Creek and 1 ft/d for West Creek provide a reasonable match between measured and model-computed water levels and between the estimated average discharge and the model-computed discharge to these streams. Average streambed altitude was varied up to plus or minus 10 ft, the accuracy of the contours from which the average streambed altitude was initially determined.

Specific yield in layer 1, and in layers 2, 3, and 4 when water levels drop below the top of the layers, was varied from 0.01 through 0.3 during model calibration. This range agrees with commonly estimated values of specific yield for unconsolidated basin-fill deposits (Lohman, 1979, p. 8, and Freeze and Cherry, 1979, p. 61) and is slightly larger than the range estimated on the basis of descriptions of materials reported in drillers' logs (see "Hydrologic properties" section of conceptual system in this report). Generally, specific-yield values were assigned on the basis of horizontal hydraulic-conductivity values determined during calibration. Specific yield was set at 0.05 at model nodes where horizontal hydraulic conductivity was 1 or 5 ft/d, 0.1 at model nodes where horizontal hydraulic conductivity was 10 or 20 ft/d, and 0.2 at model nodes where horizontal hydraulic conductivity was greater than 20 ft/d. Exceptions were made in two areas. The specific yield west of Levan, in the area where horizontal hydraulic conductivity is 40 ft/d, was set at 0.1 to better match measured water-level changes. The specific yield at Nephi, in a selected area where the hydraulic conductivity is greater than 40 ft/d, was set at 0.3 to prevent large water-level declines at individual model nodes where discharge to pumped wells was simulated.

The amount of recharge from subsurface inflow and seepage from ephemeral streams to the west side of Juab Valley was set at 5,400 acre-ft/yr for 1949 (the sum of values presented for the northern and southern parts of the valley in table 4). This is about 95 percent of estimated average conditions (the sum of values presented for the northern and southern parts of the valley is 5,700 acre-ft/yr, table 2), in accordance with the ratio of precipitation in 1949 to the 1931-93 average precipitation. For the multi-year transient-state simulation, recharge to the west side of the valley was varied as the ratio of precipitation during the given stress period to the 1931-93 average precipitation. For the seasonal transient-state simulation, recharge to the west side of the valley was varied using annual precipitation and was partitioned seasonally as described in the "Recharge" section of the modeled system in this report. This component of recharge was increased during periods of greater-than-average precipitation as described for recharge from subsurface inflow and seepage from ephemeral streams to the east side of Juab Valley (equation 5). Recharge to the west side of Juab Valley from subsurface inflow and seepage from ephemeral streams ranged from 5,000 acre-ft/yr in 1954-59 to 13,300 acre-ft/yr in 1983-86 (the sum of values presented for the northern and southern parts of the valley in table 5). As with recharge to the east side of the valley from subsurface inflow and seepage from ephemeral streams, a significant degree of uncertainty exists with this component of recharge.

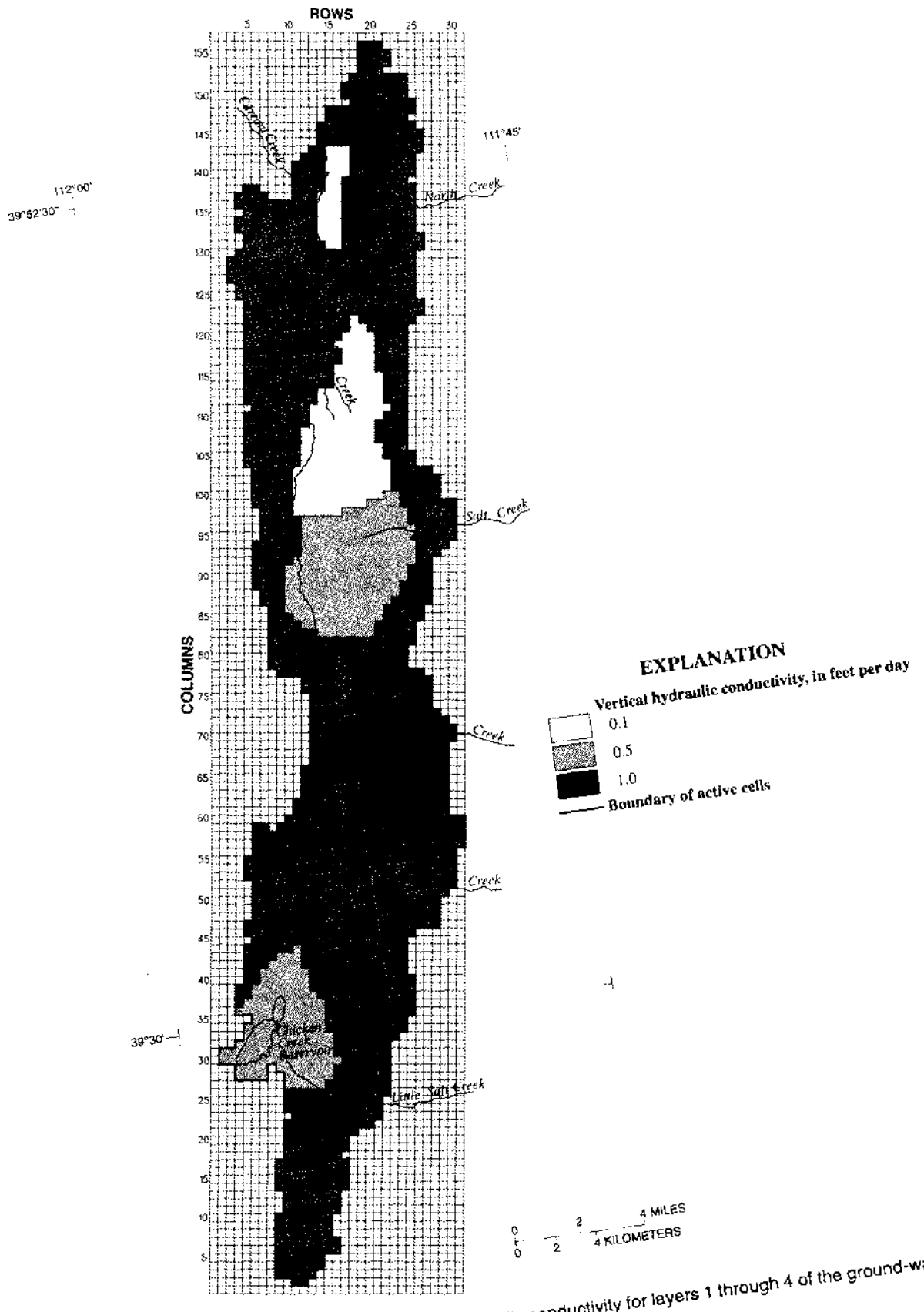


Figure 29. Final distribution of vertical hydraulic conductivity for layers 1 through 4 of the ground-water flow model of Juab Valley, Utah.

The amount of ground-water discharge by subsurface outflow from Juab Valley was varied from about 500 to 2,500 acre-ft/yr during the steady-state calibration (the minimum value for estimated average conditions is 1,000 acre-ft/yr, table 2). The amount of subsurface outflow established for 1949 was 950 acre-ft/yr (table 4), about 95 percent of the estimated average, in accordance with the ratio of precipitation in 1949 to the 1931-93 average precipitation. For the multi-year transient-state simulation, subsurface outflow was varied as the ratio of the precipitation during the given stress period to the 1931-93 average precipitation. For the seasonal transient-state simulation, subsurface outflow was varied using annual precipitation and was partitioned seasonally, as described in the "Discharge" section of the modeled system in this report. This component of discharge was increased during periods of greater-than-average precipitation as described for recharge from subsurface inflow and seepage to ephemeral streams to the east side of the valley (equation 5). Discharge by subsurface outflow ranged from 880 acre-ft/yr in 1954-59 to 2,300 acre-ft/yr in 1983-86 (table 5). Model limitations result from using a specified-flux boundary to simulate recharge and discharge from subsurface flow. The amount of flow must be specified and is not controlled by water-level changes in a transient-state simulation. Increased ground-water pumpage near the specified-flux boundaries causes the simulated discharge to subsurface outflow or recharge from subsurface inflow to likely be more or less, respectively, than what should actually occur.

Specific storage of model layers 2 and 3 was varied spatially but not vertically from 1×10^{-6} to 5×10^{-5} ft^{-1} during model calibration. The final value of specific storage assigned to all model nodes in layers 2 and 3 is 1.5×10^{-6} ft^{-1}. The storage coefficient of the unconsolidated basin-fill deposits is estimated to range from 5×10^{-5} to 5×10^{-4} on the basis of aquifer tests done in the valley. Typically, storage-coefficient values range from 5×10^{-5} to 5×10^{-3} (Freeze and Cherry, 1979, p. 60). The final storage coefficient assigned to all model nodes in layers 2 and 3 is 9×10^{-5}. This value was calculated by multiplying the specific storage of the layer by its thickness. The storage coefficient for model layer 4 was varied from 1.4×10^{-4} to 1.4×10^{-2} during model calibration. The final storage coefficient assigned to all nodes in layer 4 is 4.3×10^{-4}.

Steady-State Calibration

The numerical model was calibrated to the steady-state conditions that were assumed to have existed in the ground-water system of Juab Valley during 1949. Ground-water development in the valley was minor and most surface-water diversions were in place at this time. Therefore, recharge to and discharge from the ground-water system was dependent on the amount of precipitation in the area and not on changes in water-management practices. Examination of measured water-level changes in the valley indicate that 1949 was a period of relatively small change (fig. 30). This indicates that recharge and discharge to the ground-water system were about equal and that there was little change in the amount of ground water in storage. The conditions that are specified for the 1949 steady-state simulation are listed in table 4. Recharge and discharge amounts were determined using the methods explained in the "Boundary conditions and data requirements" section of this report. Calibration parameters were adjusted to obtain a reasonable match between model-computed water levels and water levels at seven wells measured during March and April, 1950. The steady-state simulation was not calibrated to match discharge components computed by the model because of a lack of data.

The potentiometric surface generated from model-computed water levels for layer 3 and the difference between model-computed and measured water levels in seven wells is shown in figure 31. The largest difference occurs near the north end of Mona Reservoir, where the model-computed water level is about 8 ft lower than the corresponding measured level at well (D-11-1)9bbb-4. For the hydrologic conditions estimated for 1949, the model-computed ground-water divide between the northern and southern parts of Juab Valley is located approximately at Levan (fig. 31).

The amount of specified ground-water discharge from pumped wells estimated in 1949 is about 490 acre-ft/yr (the sum of values presented for the northern and southern parts of Juab Valley in table 4). This is only about 2 percent of the 1963-93 average because major ground-water development in the valley had just started in 1947. The amount of model-computed discharge from flowing wells in 1949 is about 330 acre-ft/yr. This amount is much less than the 1963-93 average because most of the wells that contribute to the average had not yet been constructed. The small amount of discharge to wells, along with specified recharge in 1949 that is about 95 percent of the recharge

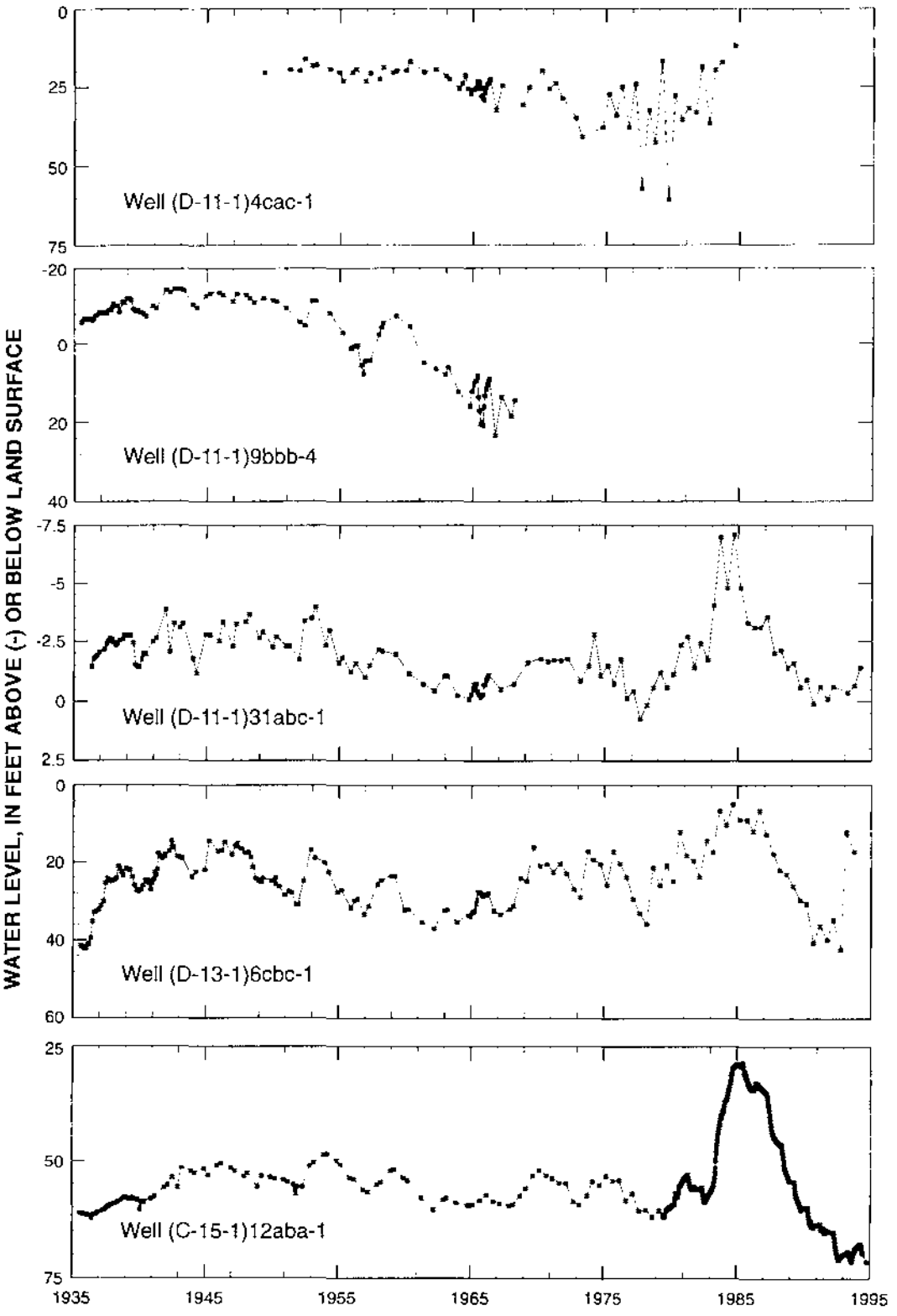


Figure 30. Water level in selected wells in Juab Valley, Utah.

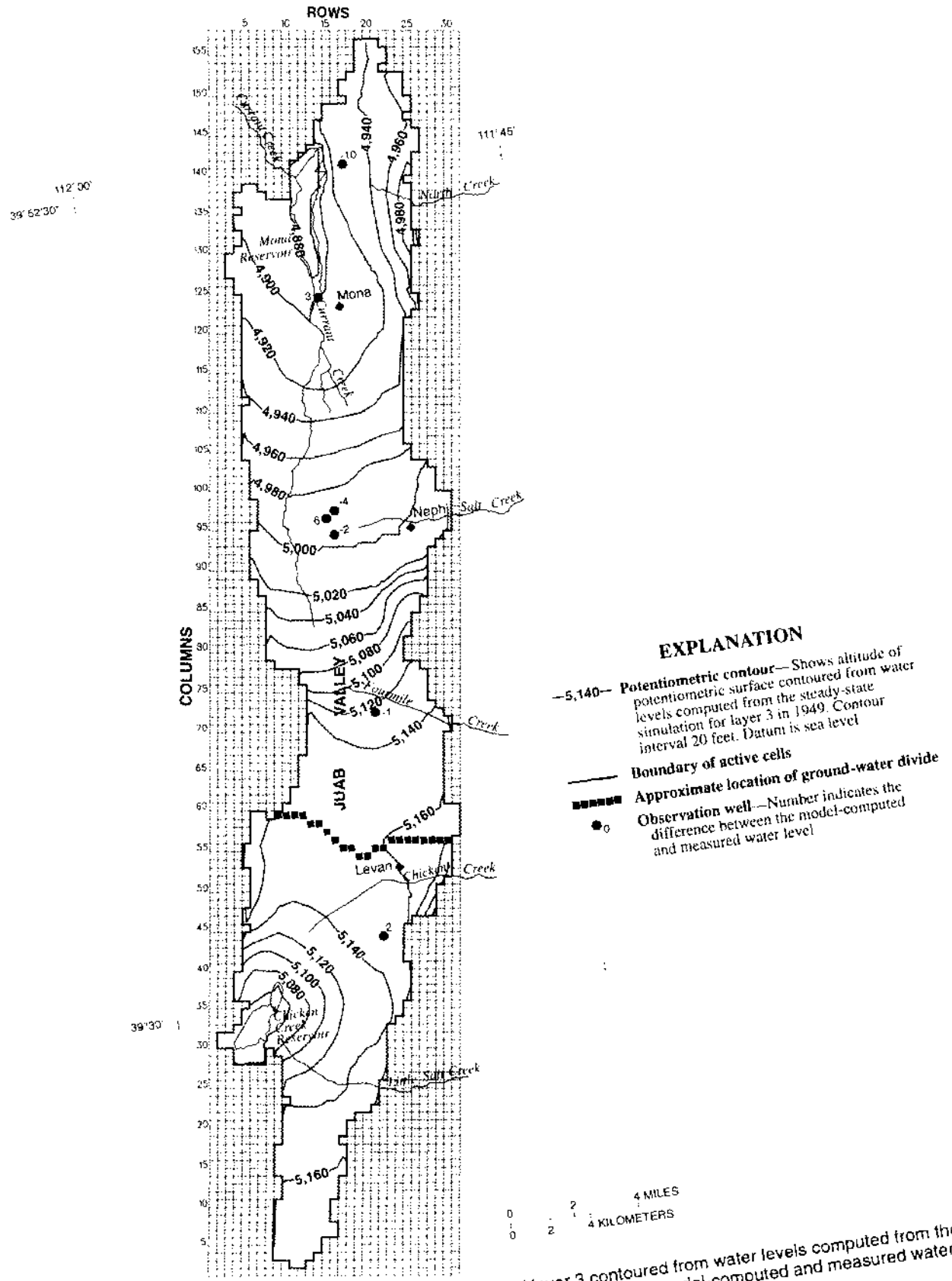


Figure 31. Potentiometric surface for model layer 3 contoured from water levels computed from the steady-state simulation for 1949 and the difference between model-computed and measured water levels in observation wells, Juab Valley, Utah, March 1950.

estimated to occur under average conditions, results in model-computed discharges for 1949 that are significantly greater than average. Model-computed discharge for the 1949 steady-state simulation is listed in table 4. The model-computed discharge to reservoirs is about 200 percent of average conditions. The sum of model-computed discharge to Currant Creek upstream and downstream from gaging station 10146400 is about 160 percent of average conditions. The model-computed discharge by evapotranspiration is about 130 percent of average conditions. The discharge specified and computed by the model for 1949 is equal to specified recharge from seepage from unconsumed irrigation water and distribution systems that is about 80 percent of average conditions and recharge from other sources that is 95 percent of average conditions. This results in the steady-state conditions simulated in 1949.

Transient-State Calibration

The numerical model was calibrated to the transient-state conditions that existed in the ground-water system of Juab Valley during 1949-94. The numerical model simulates multi-year transient-state conditions for 1949-92 and seasonal transient-state conditions for 1992-94. Initial conditions for the 1949-92 multi-year transient-state calibration are the results of the 1949 steady-state simulation.

Eight stress periods that vary from 2 to 8 years in length were used to simulate 1949-92. A multi-year approach was used because available data do not warrant yearly stress periods. Although water levels are collected on an annual basis, many other components of the model were estimated on the basis of limited data and may be better represented as averages for multi-year periods. The length of each stress period was based on a visual inspection of the estimated annual ground-water pumpage from wells in Juab Valley (fig. 32). Ground-water pumpage from wells is divided into periods where the annual withdrawals appear to be similar and can be adequately described by the average for the period. Total ground-water discharge to pumped wells simulated in each stress period equals the sum of the annual pumpage during the period. Each stress period is divided into time steps of 1-year duration.

The conditions that are specified for each of the stress periods in the 1949-92 multi-year transient-state simulation are listed in table 5. Seepage from unconsumed irrigation water and distribution systems, seepage from irrigation-season streamflow not included in the unconsumed irrigation water and distribution-systems component, seepage from nonirrigation-season streamflow, infiltration of precipitation, and discharge to wells pumped for irrigation and public supply were estimated for each stress period using the methods explained in the "Boundary conditions and data requirements" section of this report. The use of average stresses applied to the ground-water system for multi-year periods during 1949-92 results in much simplification of the conceptual model. Annual water-level changes cannot be simulated with average stresses but long-term trends can be. A numerical model could more accurately simulate the conceptual model if more data were available, which would allow use of shorter stress periods. Calibration parameters were adjusted to obtain a reasonable match between model-computed and measured water-level changes and between model-computed water levels and water levels at wells measured during March 1965 and March 1993.

Water levels computed by the model are formulated at the center point of each cell and represent the average value for the cell area. To simplify comparison of model-computed to measured water levels, measured values are assumed to represent the entire cell area. The comparison of model-computed to measured water-level changes at selected wells in Juab Valley for the 1949-92 multi-year transient-state simulation is shown in figure 33. During the 44-year period of the transient-state simulation, model-computed water-level changes generally match the overall trend of measured changes during the period but do not match yearly changes. This is a result of model calibration using stress periods that range from 2 to 8 years in length. Hydrologic stresses do not change during a stress period and are averages for the length of the period. Any variation from the average during individual years or during periods shorter than a year cannot be reproduced.

Model-computed water levels reasonably matched water levels measured from 1949 to 1967. Model-computed water levels declined during 1970-77 in contrast to measured water levels which fluctuated but did not show an overall decline. The decline in model-computed water levels was caused by increased ground-water withdrawals from wells and less-than-average precipitation that resulted in decreased recharge from subsurface inflow and seepage from ephemeral streams during this period. Model-computed water levels rose during 1978-86, which is generally in agreement with measured water levels. The peak in measured water levels occurred during 1984-85. Model-computed water levels reached a similar peak in

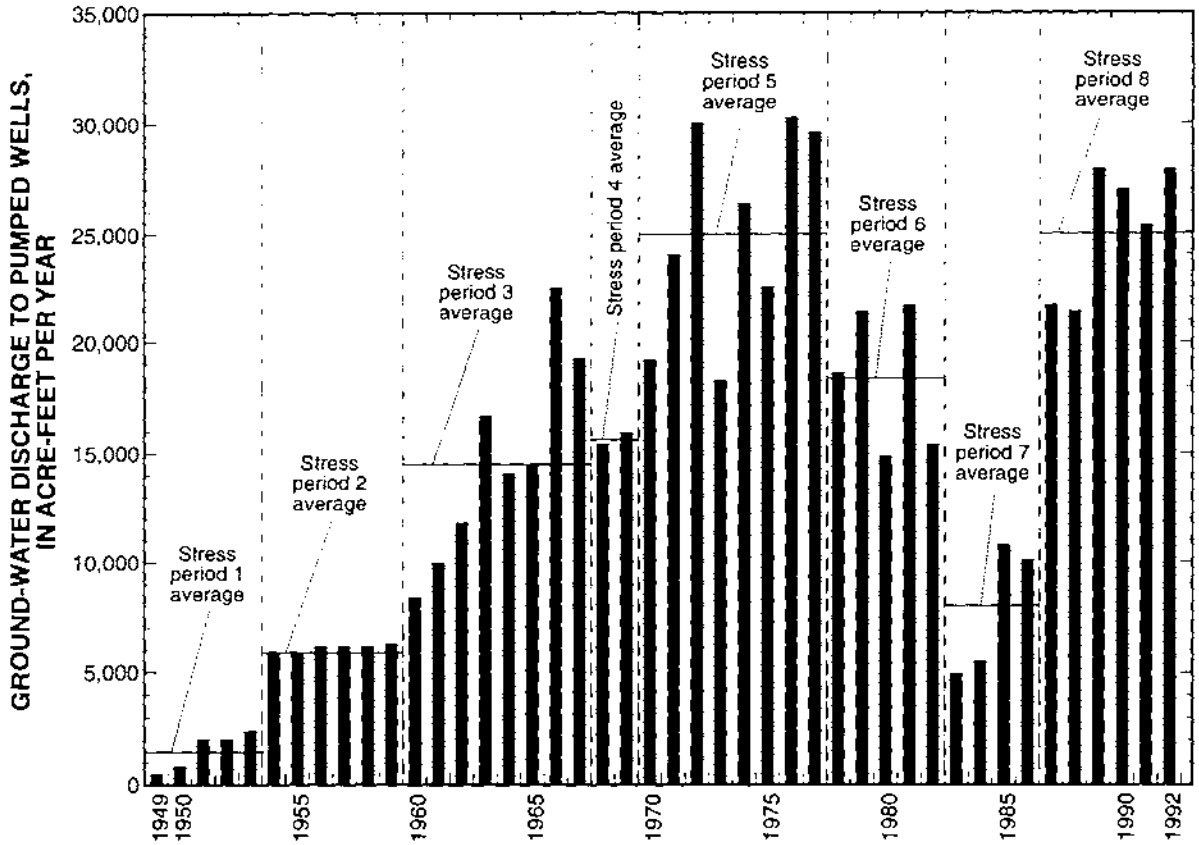


Figure 32. Estimated annual ground-water discharge to pumped wells, length of stress periods, and average annual pumpage for each stress period of the multi-year transient-state simulation, Juab Valley, Utah.

March 1987, lagging behind the measured peak by several years. This is a result of recharge and discharge components being averaged throughout the 1983-86 stress period. Both model-computed and measured water levels declined during 1987-92. In the southern part of Juab Valley, the model-computed water levels did not rise as much as water levels measured during the early 1980's; however, the overall trend is reproduced. Recharge from subsurface inflow and seepage from ephemeral streams to the east side of Juab Valley is not a large component of recharge in the southern part of the valley. It is this specified recharge component that creates large water-level changes in the model because during periods of greater-than-average precipitation, a coefficient is applied that increases the proportion of recharge to precipitation.

The potentiometric surface determined from model-computed water levels in layer 3 at the end of the 1964 time step and the difference between model-computed water levels and water levels measured in March 1965 are shown in figure 34. The potentiometric surface determined from model-computed water levels in layer 3 at the end of the 1992 time step and the difference between model-computed water levels and water levels measured in March 1993 are shown in figure 35. In general, model-computed water levels are within 10 ft of measured water levels. The mean of the differences between model-computed and measured water levels is the bias in the distribution of positive and negative values. The mean of the absolute differences is considered the mean error. The mean of the differences and the mean of the absolute differences for 88 wells measured in March 1965 are 0.4 and 8.9 ft, respectively. For 100 wells measured in March 1993, they are 1.4 and 9.0 ft, respectively.

The largest difference between model-computed water levels and water levels measured in March 1965

occurs east of Mona Reservoir, where the model-computed water level is about 42 ft higher than the corresponding measured level at well (D-11-1)21bbb-1. The largest difference between model-computed water levels and water levels measured in March 1993 occurs just north of the topographic divide that separates the northern and southern parts of Juab Valley, where the model-computed water level is about 47 ft lower than the corresponding measured level at well (D-14-1)6dbb-1. This is the only well with water-level data available near the topographic divide, and little other data is available for the area. The model-computed ground-water divide between the northern and southern parts of Juab Valley controlled by the hydrologic conditions in 1965 and 1993 moves progressively farther north from the model-computed divide controlled by the hydrologic conditions in 1949 (figs. 31, 34, and 35). This means that ground-water discharge to wells in the southern part of the valley has caused water that previously flowed to the northern part to flow to the southern part.

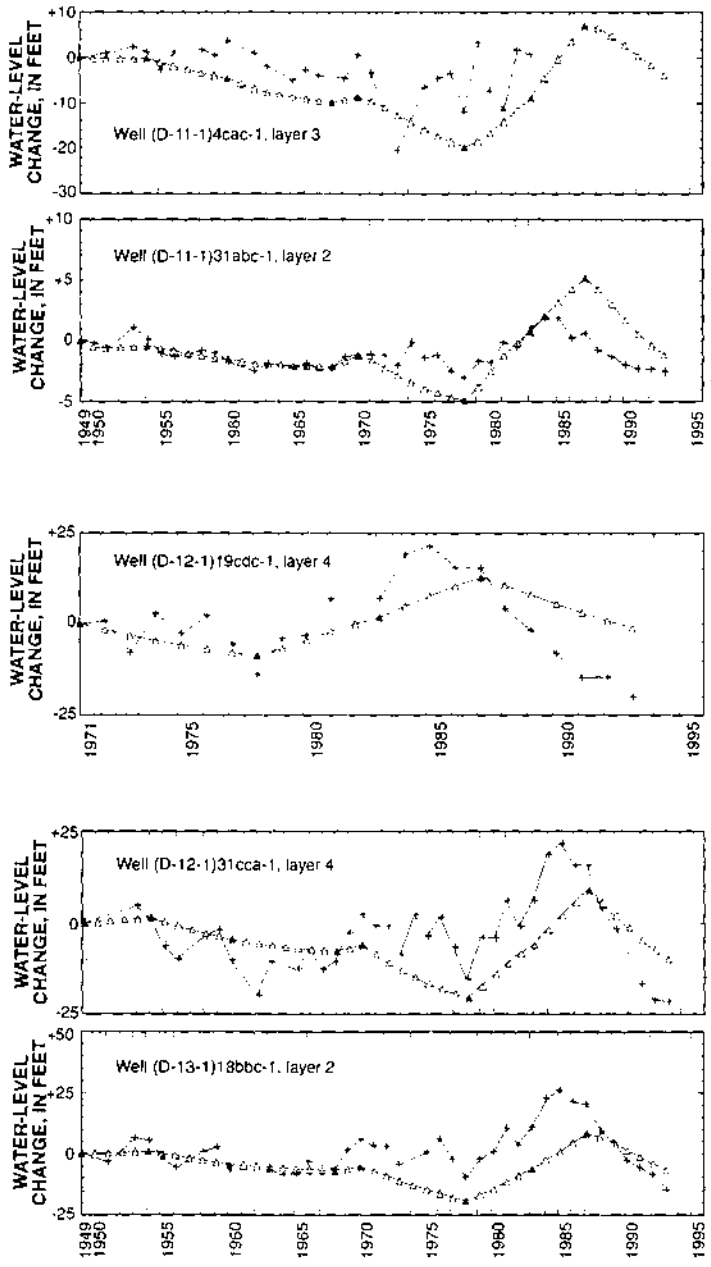
Model-computed discharge and changes in storage are listed in tables 4, 5, and 6. Initially, model-computed discharge was adjusted to match the general proportions of discharge under average conditions (table 2) and was then varied until a reasonable match was obtained between model-computed and measured water levels near the discharge areas during specific stress periods. Total model-computed discharge ranged from 32,200 acre-ft/yr during the 1970-77 stress period to 53,400 acre-ft/yr during the 1983-86 stress period (the sum of values presented for the northern and southern parts of Juab Valley in table 5).

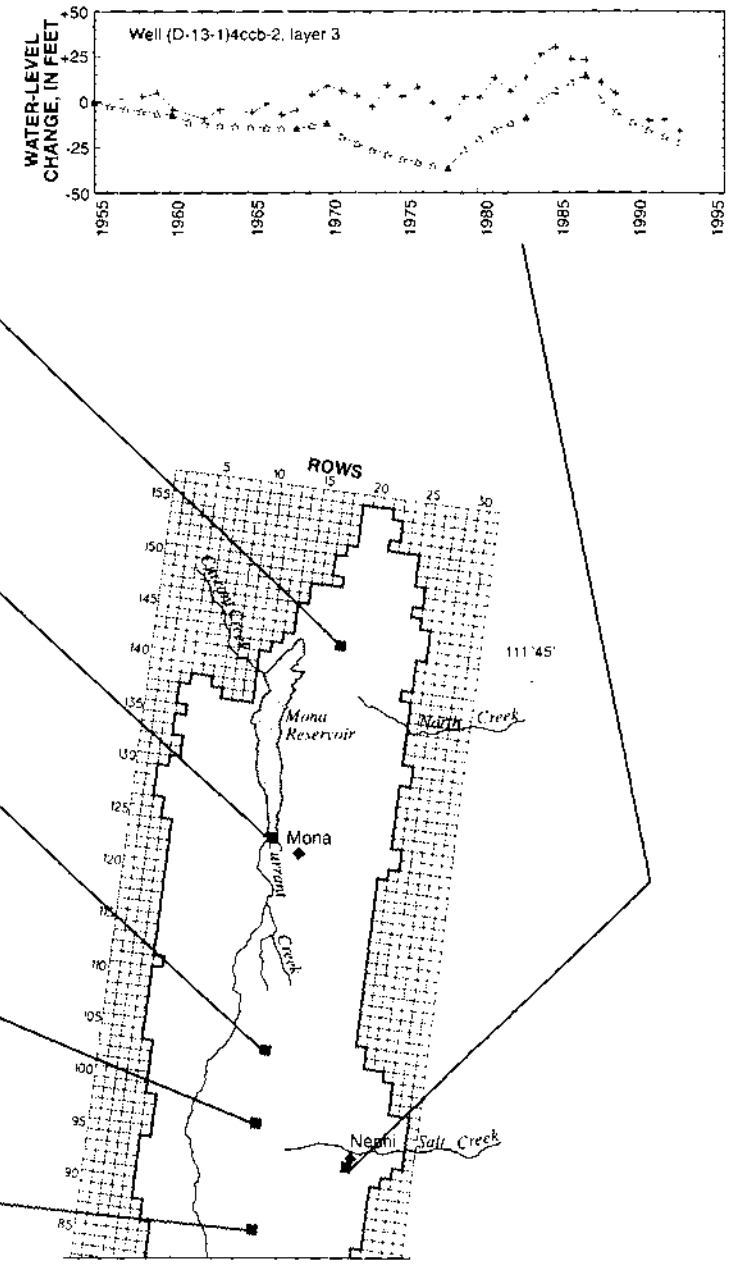
The general location where the difference between model-computed water levels for model layers 1 and 4 is zero for 1949 and 1992 is shown in figure 36. The area interior to the line is where the model-computed water levels for model layer 4 are greater than for model layer 1, indicating the direction of flow is upward. Generally, this area can be thought of as the ground-water discharge area for Juab Valley. The model-computed ground-water discharge area decreased in size during the 1949-92 transient period.

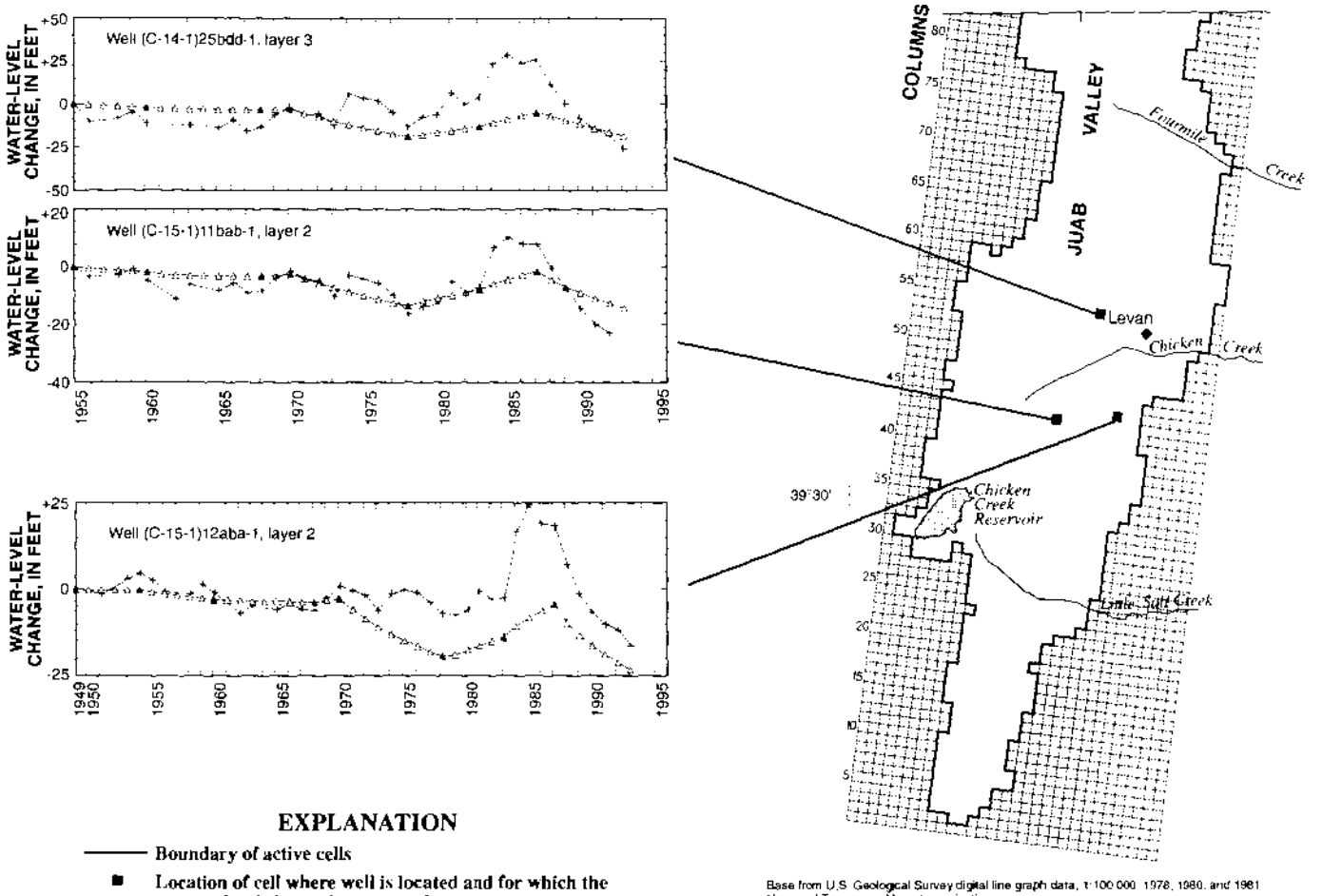
The numerical model also was calibrated to simulate seasonal stress conditions for 1992-94. The seasonal transient-state simulation begins in January 1992. The ending water levels for the 1992 time step of the 1949-92 transient-state model, which actually represent March 1992 water levels, are considered adequate as the initial water levels for the seasonal simulation. The results of the model for the first several stress periods have limited value because these stress periods represent the change from initial conditions computed using hydrologic stresses that were averaged for multi-year periods to conditions using seasonal hydrologic stresses. The seasonal transient-state simulation was divided into nine stress periods, each of 3-months duration. Each stress period is divided into monthly time steps. Three-month-long stress periods allow for the simulation of the most significant seasonal changes, which are the use of streamflow for irrigation in the spring and ground-water pumpage from wells for irrigation in the summer. In general, recharge and discharge data for the seasonal transient-state simulation are not directly measured but are converted from annual data. Use of a shorter stress period would require data that are more detailed than are currently available. Without these more detailed data, a shorter stress period would not improve the calibration of the seasonal simulation. The annual rate of recharge from subsurface inflow and seepage from ephemeral streams to the east and west sides of the valley, and discharge to subsurface outflow, is adjusted for 1992 and 1993 as explained in the "Calibration parameters" section of this report.

The conditions that are specified for each of the stress periods in the 1992-94 transient-state simulation are listed in table 6. Ground-water recharge from seepage from unconsumed irrigation water and distribution systems, seepage from irrigation-season streamflow not included in the unconsumed irrigation water and distribution-systems component, seepage from nonirrigation-season streamflow, infiltration of precipitation, and ground-water discharge to wells pumped for irrigation and public supply were estimated for each stress period using the methods explained in the "Boundary conditions and data requirements" section of this report.

The comparison of model-computed to measured water-level changes at selected wells in Juab Valley for the 1992-94 seasonal transient-state simulation are shown in figure 37. The timing and trend of model-computed water levels generally follow the timing and trend of measured water levels during 1993-94. Ground-water discharge to wells is the major cause of seasonal water-level changes. These water-level changes are measured at a well, and in most cases, the magnitude of the changes cannot be accurately simulated with the scale of discretization, both spatial and temporal, used in this model. Model-computed water-level changes are typically smaller than measured water-level changes because the effects of discharge at a single well are averaged across an entire model cell.


Well (D-11-1)4cac-1, layer 3
Well (D-11-1)31abc-1, layer 2
Well (D-12-1)19cdc-1, layer 4
Well (D-12-1)31cca-1, layer 4
Well (D-13-1)18bbc-1, layer 2
WATER-LEVEL CHANGE, IN FEET


Well (D-13-1)4ccb-2, layer 3
WATER-LEVEL CHANGE, IN FEET
ROWS
112°00'
39°52'30"
111°45'
Mona Reservoir
North Creek
Mona
Nephi
Salt Creek



EXPLANATION

——— Boundary of active cells

■ **Location of cell where well is located and for which the water-level change is computed**

Hydrographs showing water-level change—The Y-axis of each hydrograph shows the water-level change, in feet; scale is variable. Dashed where estimated

+ Measured water-level change

△ Model-computed water-level change

▲ Model-computed water-level change at beginning of stress period

Base from U.S. Geological Survey digital line graph data, 1:100,000, 1978, 1980, and 1981
Universal Transverse Mercator projection,
Zone 12

0 2 4 MILES
0 2 4 KILOMETERS

Figure 33. Model-computed and measured water-level change in selected wells in Juab Valley, Utah, for the multi-year transient-state simulation.

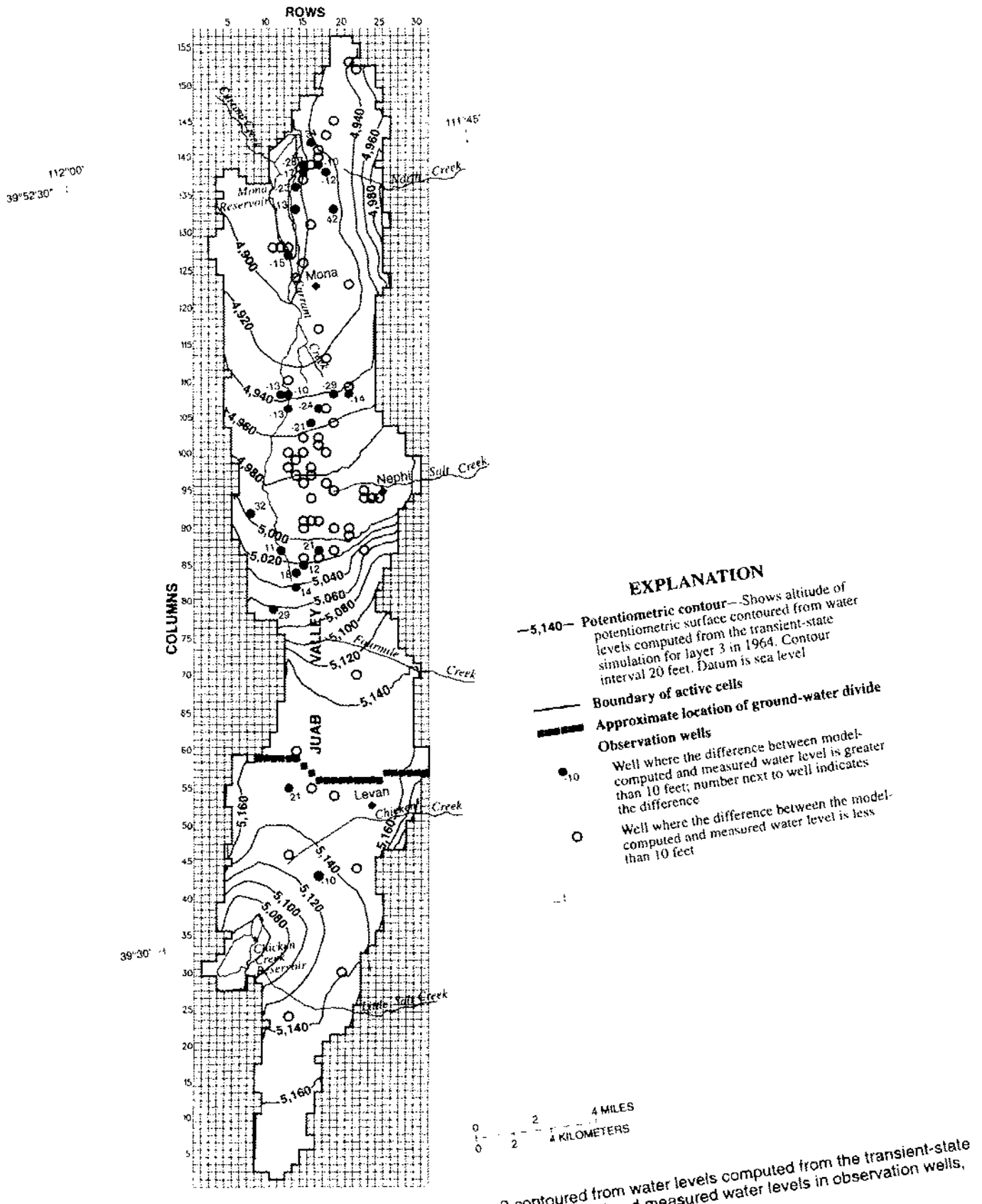


Figure 34. Potentiometric surface for model layer 3 contoured from water levels computed from the transient-state simulation for 1964 and the difference between model-computed and measured water levels in observation wells, Juab Valley, Utah, March 1965.

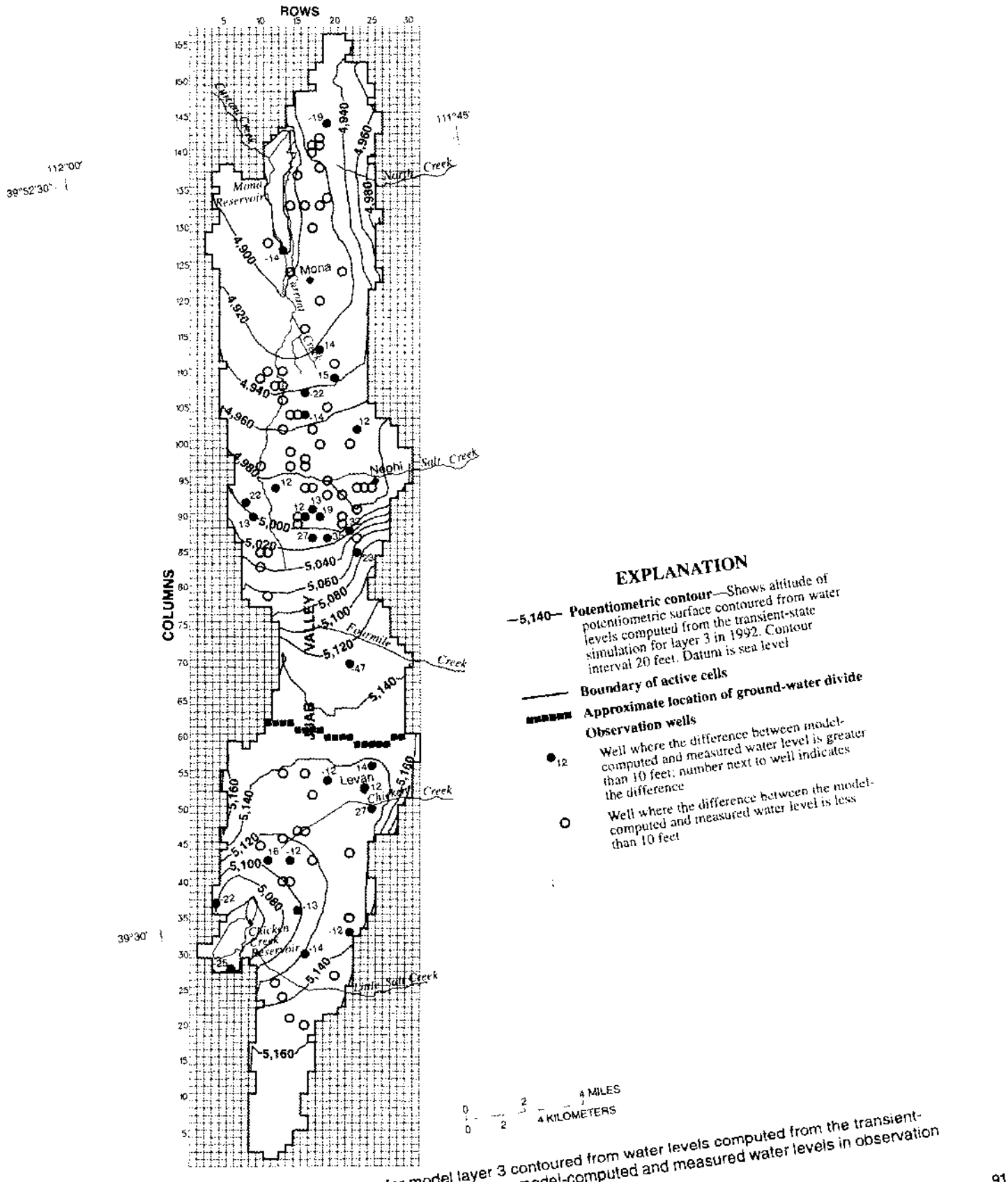


Figure 35. Potentiometric surface for model layer 3 contoured from water levels computed from the transient-state simulation for 1992 and the difference between model-computed and measured water levels in observation wells, Juab Valley, Utah, March 1993.

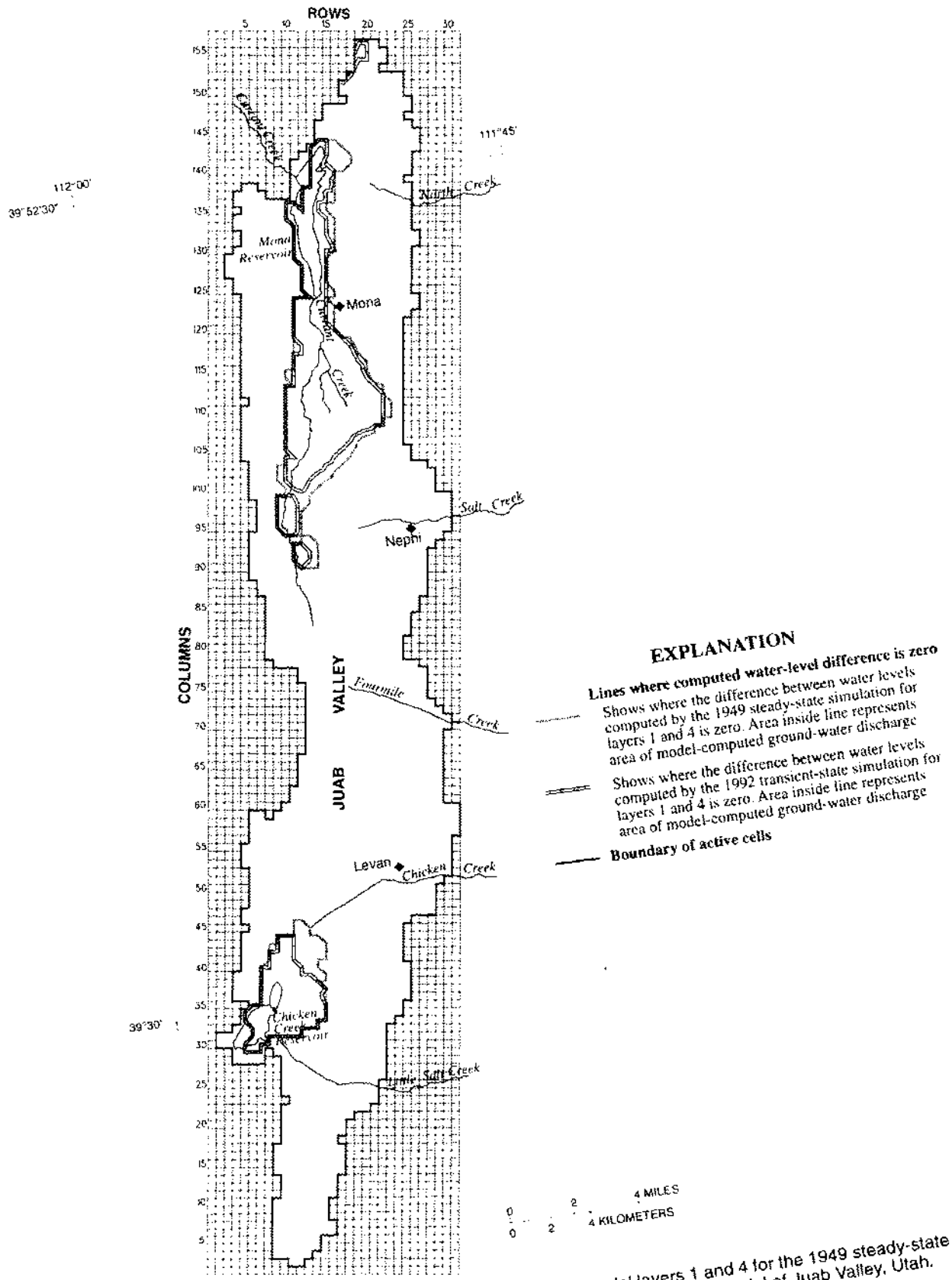


Figure 36. Computed water-level difference between model layers 1 and 4 for the 1949 steady-state simulation and the 1992 transient-state simulation in the ground-water flow model of Juab Valley, Utah.

The exception is near Levan at well (C-15-1)12aba-1, where model-computed water-level changes are greater than measured changes. This may be caused by horizontal hydraulic-conductivity values assigned to model nodes in the vicinity of this well that are too low. Ground-water discharge to wells is reported as an annual total, and the assumption that all withdrawals occur during the July through September stress period may not be completely accurate at all wells. The timing of model-computed water-level declines in the southern part of Juab Valley lags behind the measured water-level declines by about 2 months, and the timing between model-computed and measured water-level changes in the northern part of the valley generally correspond. The measured water levels indicate that ground-water pumpage from wells may begin earlier in the southern part of Juab Valley. The timing of ground-water discharge to wells in the seasonal transient-state simulation was not varied independently for the northern and southern parts of Juab Valley.

Model-computed and measured waters levels in March 1993, September 1993, and March 1994 were compared to determine the mean of the differences and the mean of the absolute differences. For 100 wells measured in March 1993, they are 2.0 and 8.7 ft, respectively. For 105 wells measured in September 1993, they are 0 and 9.3 ft, respectively. For 38 wells measured in March 1994, they are -2.1 and 9.6 ft, respectively.

The largest differences between model-computed water levels and water levels measured in March 1993, September 1993, and March 1994 occur near the topographic divide that separates the northern and southern parts of Juab Valley. The model-computed water level is about 46 ft lower than the corresponding measured level at well (D-14-1)6dbb-1 for all three comparisons.

As mentioned with respect to the multi-year transient-state simulation, discharge amounts for specific stress periods in the seasonal transient-state simulation are not known. The seasonal transient-state simulation therefore could not be calibrated to match discharge to springs or evapotranspiration. The conceptual model could be more accurately simulated if more data on the hydrologic stresses were available.

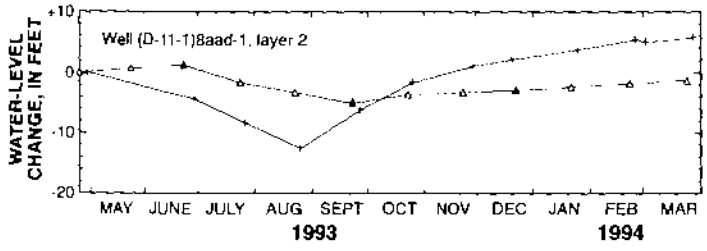
Model Sensitivity

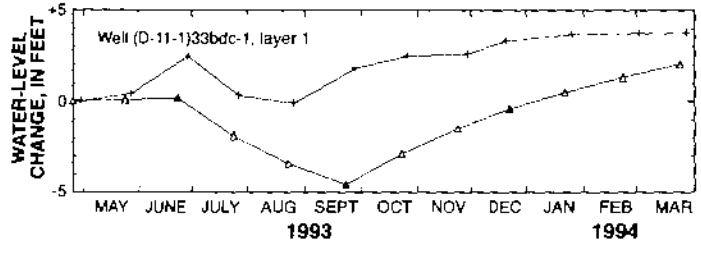
The sensitivity of model-computed water levels and flow to changes in hydrologic properties was qualitatively observed during model calibration. Once the numerical model was calibrated, selected boundary conditions and hydrologic stresses were changed, and the resulting changes in water level and flow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

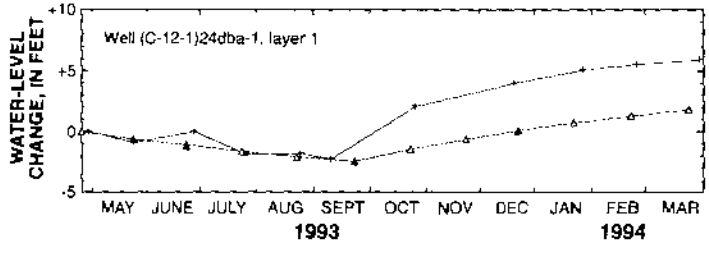
Water levels throughout the modeled area and in all layers are sensitive to the specific yield specified for layer 1. Model-computed water levels are not sensitive to changes in specific storage and storage coefficient. An increase in specific yield causes a decline in model-computed water levels. Specifically, water levels in model layer 1 and the stability of the numerical model are sensitive to specific yield and horizontal hydraulic conductivity in areas where large ground-water withdrawals from wells are simulated. This is especially true in the seasonal transient-state simulation, where all ground-water withdrawals from wells are simulated in the July through September stress period. A decrease in specific yield in areas with large withdrawals of water can cause model-computed water levels in layer 1 to decline below the bottom of the layer.

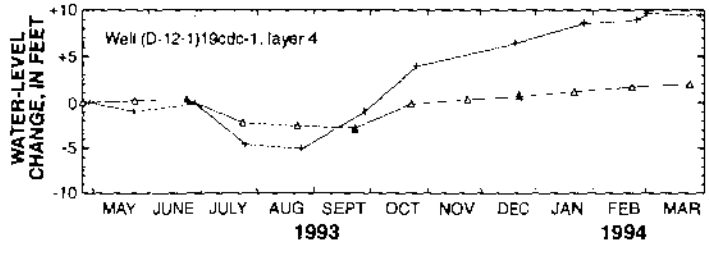
Discharge to Mona and Chicken Creek Reservoirs is sensitive to both horizontal and vertical hydraulic conductivity at and near the reservoirs. A decrease in horizontal and vertical conductivity near the reservoirs causes a decrease in model-computed discharge to the reservoirs and a rise in model-computed water levels near the reservoirs. This also causes an increase in model-computed discharge to flowing wells, springs, seeps, and evapotranspiration. Model-computed discharge by evapotranspiration is not sensitive to an increase in extinction depth. In general, an increase in extinction depth causes a small decrease in model-computed discharge to springs and seeps and a small increase in model-computed discharge by evapotranspiration. This is not surprising because evapotranspiration is simulated in areas where springs and seeps occur. Model-computed discharge to Currant Creek upstream and downstream from gaging station 10146400 is sensitive to a decrease in the corresponding streambed conductance values. A decrease in these conductances causes an increase in model-computed discharge to Mona Reservoir.

Model-computed water levels in layer 1 are not sensitive to a decrease in the altitude of the bottom of the layer. Model-computed water levels for layer 1 are controlled mainly by the altitude and the amount of water moving toward the model boundaries where discharge occurs and not by the thickness of model layer 1. Water levels throughout the modeled area and in all layers are sensitive to changes in the transmissivity of









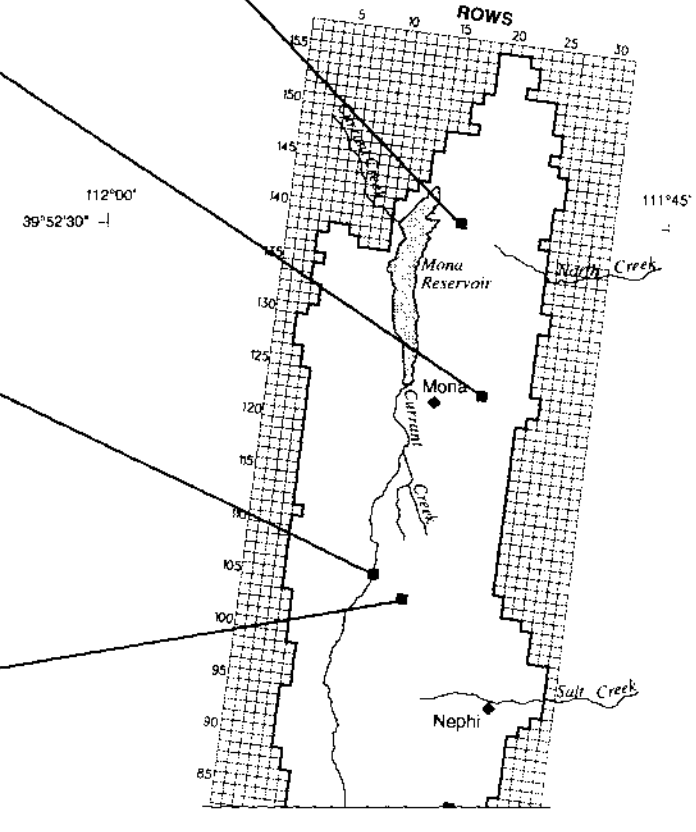
EXPLANATION

Boundary of active cells

Location of cell where well is located and for which the water-level change is computed

Hydrographs showing water-level change—The Y-axis of each hydrograph shows the water-level change, in feet; scale is variable

- Measured water-level change
- Model-computed water-level change
- Model-computed water-level change at beginning of stress period



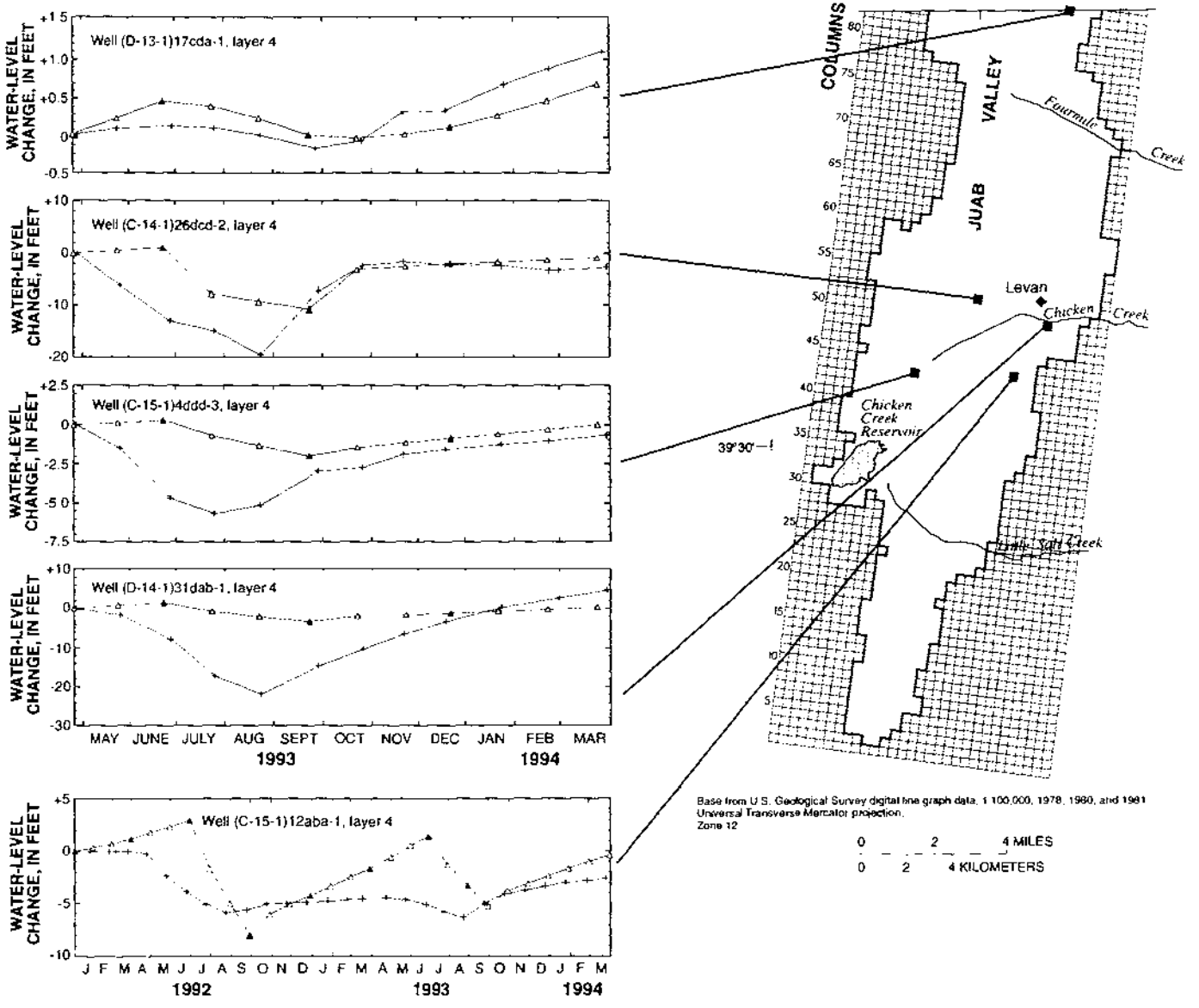


Figure 37. Model-computed and measured water-level change in selected wells in Juab Valley, Utah, for the seasonal transient-state simulation.

layer 4. The thickness of unconsolidated basin-fill deposits in the valley and hydraulic conductivity of the deposits at greater depths, parameters that control the transmissivity of layer 4, are not well defined. An increase in the transmissivity causes a decline in model-computed water levels, and a decrease in transmissivity has the opposite effect.

Water levels throughout the modeled area and in all layers are sensitive to recharge from subsurface inflow. The amount of recharge from subsurface inflow and seepage from ephemeral streams is not known with a reasonable degree of certainty. It is increased nonproportionally during periods of greater-than-average precipitation in order to better approximate the magnitude of water-level changes in certain areas of the valley. A proportional relation between precipitation and this recharge component for the 1983-86 stress period results in a decrease in recharge of about 40 percent from the nonproportional amount and a water-level decline of about 2 to 5 ft in the Nephi area. About one-third of the change in recharge is incorporated in a decrease in model-computed discharge, and the remainder is accounted for in a decrease in storage.

Model Limitations

This numerical model represents a simplification of the ground-water system in Juab Valley and is based on estimates of recharge, discharge, and hydrologic properties made from available data. The choice of specific boundaries to simulate recharge and discharge, the methods used to discretize time and to vertically discretize the simulated unconsolidated basin-fill deposits, and the specified objectives of the simulations affect the solutions obtained from the numerical model. Use of the numerical model to explore objectives other than the ones listed in the simulation introduction and stresses outside of the range of conditions used in the steady-state and transient-state simulations should be considered unreliable.

Data are not available to estimate individual components of discharge, such as to springs; therefore, the numerical model could only be calibrated to match water levels in discharge areas. The accuracy of model-computed flow for individual components of discharge is not known. Use of the model to estimate variability of individual discharge components as they are related to recharge should be done with caution.

In the topographically lowest areas of Juab Valley, a few model-computed water levels for layer 1, which represents unconfined conditions, are above the estimated altitude of land surface. This reflects errors associated with model-computed water levels. During the steady-state simulation, most of these water levels were less than 10 ft above land surface. During periods of increased recharge, the number of cells where water levels were computed above land surface increased. This is a limitation of the model in the lowest areas of Juab Valley. The vertical head difference in areas around Mona Reservoir could not be accurately simulated because of complex hydraulic gradients. Water levels in these areas could not be matched without causing some model-computed water levels to be above land surface.

The seasonal transient-state simulation assumes that all annual ground-water pumpage from wells occurs during one 3-month long stress period. This causes the numerical model to be sensitive to the amounts and location of simulated ground-water pumpage from model layer 1. An increase in pumpage from this layer may cause model-computed water levels near the simulated withdrawals to oscillate and prevent the numerical model from converging to a solution. Averaging pumpage across several model cells may prevent this from occurring.

The stress-period length used in the transient simulations affected model calibration. Model-computed water levels were calculated from stresses averaged over multi-year periods. The model-computed water levels were compared to water levels, which were measured on a yearly basis and are dependent on annual stresses. A measured water level is a single measurement at a given point in time and does not represent an annual average. Generally, the trend and the magnitude of measured changes are simulated by the model.

Recharge from subsurface inflow and seepage from ephemeral streams and discharge by subsurface outflow used in the transient simulations have a larger degree of uncertainty than do other forms of recharge or discharge that are specified in the model. These stresses were modified during model calibration to better approximate the magnitude of measured water-level change in the valley. Therefore, recharge from subsurface inflow and seepage from ephemeral streams and discharge by subsurface outflow, in addition to other calibration parameters, compensate for errors in the amount of other sources of recharge applied and the amount of discharge computed by the numerical model. Model results should be evaluated with care if future management scenarios include periods of greater-than-average precipitation. It is conceivable that more water

can be specified as recharge from subsurface inflow and seepage from ephemeral streams than is physically available from precipitation if the ratio between precipitation during the period and the long-term average is large enough.

The use of injection wells to simulate recharge from subsurface inflow and seepage from ephemeral streams and pumping wells to simulate discharge by subsurface outflow may exaggerate the effects of hydrologic stresses on model-computed water levels near the edges of the modeled area. The amount of these recharge and discharge components is specified and is, therefore, independent of any changes in the hydraulic gradient across the boundary between the consolidated rock and unconsolidated basin-fill deposits. If additional ground-water discharge to wells is simulated near the edges of the modeled area, recharge from subsurface inflow does not increase and discharge by subsurface outflow does not decrease to compensate for the withdrawals. If recharge from subsurface inflow and discharge by subsurface outflow is dependent on water levels in the unconsolidated basin-fill deposits, then computed water-level declines near the edges of the modeled area may be larger than actual declines. The simulation of increased pumpage in Juab Valley would likely show slightly exaggerated drawdowns in the system because of the use of specified-flux boundary conditions.

Model-computed water levels declined during the 1970-77 stress period, a period of increased ground-water pumpage from wells and overall less-than-average precipitation. Water levels measured during this period fluctuate but do not show an overall decline. This may be caused by subsurface inflow from consolidated rocks and subsurface outflow to consolidated rocks that is dependent on water levels in the unconsolidated basin-fill deposits rather than on a specified amount based on precipitation.

The absolute value of a model-computed water level is affected by the assumptions and simplifications of the model, so the model should be used to compare relative values rather than to determine absolute values. For example, the model could be used to estimate the rates of pumping in a well field that would result in a 20-ft decline in the water level of adjacent wells if the water levels were not affected by a specified-flux boundary condition. The model should not be used to estimate the rates of pumping in a well field that would drop the water level in adjacent wells below an altitude of 5,200 ft. When comparing two model-computed scenarios not affected by a specified-flux boundary condition, however, the relative difference in value between the scenarios is valid because both scenarios are affected by the same assumptions and simplifications. Therefore, absolute water-level altitudes cannot be predicted but changes in both water-level and water-budget components can be.

Because the thickness of the saturated unconsolidated basin-fill deposits in Juab Valley is not known, the thickness of model layers was in part based on the depth at which average ground-water pumpage from wells occurs. Changes in the estimation of model-layer thickness affects transmissivity, vertical-conductance, and storage-coefficient values, if the basin-fill deposits are assumed to be homogeneous with increasing depth.

The set of hydrologic properties, stresses, and boundary conditions used in the numerical model is only one possible combination of parameters that could result in model-computed water levels that are similar to measured ones. Simplifications and assumptions made to the conceptual model of the ground-water system because of a lack of data affect the reliability of the numerical model. Additional data on recharge and discharge components are needed to improve this ground-water flow model. Despite the limitations described in this section, the ground-water flow model can be used to better understand the ground-water system in Juab Valley. It is a useful tool to estimate the effects of changes in ground-water pumpage from wells and recharge on water levels and natural discharge.

NEED FOR FUTURE STUDY

More detailed information is needed to refine recharge and discharge estimates used in the ground-water flow model and to better define the aquifer geometry in Juab Valley. Additional seepage studies on ephemeral and perennial streams in the area are needed to better quantify recharge to the ground-water system during spring runoff and the nonirrigation season. Continuous-recording gaging stations are needed on Salt Creek and a representative stream draining the Wasatch Range to provide long-term streamflow record. More streamflow data coupled with pumpage information already being collected would help to estimate the amount of recharge that occurs from irrigation.

Seepage runs and discharge measurements also are needed to better quantify ground-water discharge to streams, springs, and seeps. Annual discharge measurements may indicate the relation between discharge to streams, springs, and seeps; and annual precipitation and recharge. Discharge from selected springs could be

monitored for an entire year in order to determine any seasonal fluctuations and to better estimate the total annual discharge from the ground-water system. Continuous-recording gaging stations just below Mona and Chicken Creek Reservoirs also would provide data needed to refine estimates of discharge to springs. Measurement of evapotranspiration rates for phreatophytes in Juab Valley is needed to better estimate the quantity of discharge by evapotranspiration.

A refined estimate of the ground-water recharge and discharge components listed previously would result in a ground-water budget residual that better represents recharge from subsurface inflow. The integration of this additional data with long-term precipitation, streamflow, and pumping-well discharge data could allow the use of shorter stress periods in the numerical model that would better represent the ground-water system in Juab Valley.

SUMMARY

Plans to import water to Juab Valley, Utah, primarily for irrigation, are part of the Central Utah Project. A better understanding of the hydrology of Juab Valley is needed to help manage the water resources and to develop conjunctive-use plans. The hydrology of the valley was studied by the U.S. Geological Survey in cooperation with the Central Utah Water Conservancy District and the East Juab Water Conservancy District from 1992 through 1994.

Average annual streamflow for Salt Creek near Nephi (1952-80) and Chicken Creek near Levan (1963-93) is about 19,600 and about 5,800 acre-feet, respectively. The saturated unconsolidated basin-fill deposits form the ground-water system in Juab Valley. Recharge to the unconsolidated basin-fill deposits in the valley is by seepage from streams, unconsumed irrigation water, and distribution systems; infiltration of precipitation; and subsurface inflow from consolidated rocks that surround the valley. Discharge is by wells, springs, seeps, evapotranspiration, and subsurface outflow to consolidated rocks. Recharge and discharge amounts vary annually, but were estimated to be about 42,000-46,000 acre-feet per year in the northern part and about 12,000 acre-feet per year in the southern part of Juab Valley, under average conditions. Ground-water pumpage in areas near and in Nephi and Levan has altered the direction of ground-water flow from that of pre-ground-water development time. The average annual ground-water discharge to wells in Juab Valley during 1963-93 is about 23,000 acre-feet per year.

Values of transmissivity and storage coefficient for unconsolidated basin-fill deposits were estimated from aquifer tests incorporating multiple wells at a site in Nephi (about 242,000 ft^2/d and 8×10^{-5}, respectively) and a site southwest of Levan (about 4,000 ft^2/d and from about 5×10^{-5} to 5×10^{-4}, respectively). Greater-than-average precipitation during 1980-87 corresponds with a rise in water levels measured in most wells in the valley and the highest water level measured in some wells. Less-than-average precipitation during 1988-91 corresponds with a decline in water levels measured during 1988-93 in most wells.

Dissolved-solids concentrations in ground water from Juab Valley sampled during 1992-93 ranged from 261 to 3,980 milligrams per liter. The increase in the sulfur-34 to sulfur-32 ratio in most of the water sampled along a hypothesized flow path in the southern part of the valley indicates the dissolution of gypsum derived from the Arapien Shale. Geochemical analyses indicate that the sources of dissolved ions in water sampled from the southern part of the valley are primarily the Arapien Shale, evaporite deposits, and possibly residual sea water that has undergone evaporation in unconsolidated basin-fill deposits in selected areas.

A numerical model of the ground-water system in the unconsolidated basin-fill deposits of Juab Valley was constructed using the U.S. Geological Survey modular, three-dimensional, finite-difference, ground-water flow model. The objective of the simulations was to test and refine the conceptual flow model and to provide a tool that can be used to estimate the effects of changes in ground-water discharge to pumped wells and ground-water recharge on ground-water levels, and in a qualitative sense, natural ground-water discharge. Horizontal hydraulic-conductivity values used in the model and determined during model calibration ranged from 1 to 125 ft/d, and vertical hydraulic-conductivity values ranged from 0.1 to 1 ft/d. Drain-conductance values ranged from 1.5×10^6 to 6×10^6 ft^2/d. Specific yield varied from 0.05 to 0.3 and generally is based on the horizontal hydraulic conductivity assigned to the model node. The storage coefficient of all active model nodes in layer 4 is 4.3×10^{-4}.

The numerical model was calibrated to the steady-state conditions that existed in the ground-water system in Juab Valley during 1949, multi-year transient-state conditions for 1949-92, and seasonal transient-state conditions for 1992-94. Steady-state conditions include estimated annual recharge for 1949 and measured water levels at seven wells during March and April 1950. The largest difference between model-

computed and measured water levels occurs near the north end of Mona Reservoir, where the model-computed level is about 10 feet lower than the corresponding measured water level. For the multi-year transient-state simulation, calibration parameters were adjusted to obtain a reasonable match between model-computed and measured water-level changes and between model-computed water levels and water levels at wells measured during March 1965 and March 1993. Parameters important to the calibration process include horizontal hydraulic conductivity, transmissivity, and the spatial distribution and amount of recharge from subsurface inflow and seepage from ephemeral streams to the east side of Juab Valley.

Model-computed water-level changes are sensitive to specific yield. Model-computed discharge to Mona and Chicken Creek Reservoirs is sensitive to both horizontal and vertical hydraulic conductivity at and near the reservoirs. Water levels throughout the modeled area and in all layers are sensitive to changes in the transmissivity of layer 4 and recharge from subsurface inflow. This numerical model represents a simplification of the ground-water system in Juab Valley and is based on estimates of recharge, discharge, and hydrologic properties made from the available data.

REFERENCES CITED

Allen, D.V., Garrett, R.B., and others, 1994, Ground-water conditions in Utah, spring of 1994: Utah Division of Water Resources Cooperative Investigations Report No. 34, 89 p.

Arnow, Ted, and others, 1964, Developing a State water plan, ground-water conditions in Utah, spring of 1964: Utah Water and Power Board Cooperative Investigations Report No. 2, 104 p.

Ashcroft, G.L., Jensen, D.T., and Brown, J.L., 1992, Utah climate: Utah Climate Center, Utah State University, Logan, 127 p.

Bick, R.F., 1991, Provisional geologic map of the Nephi quadrangle, Juab County, Utah: Utah Geological Survey Map 137.

Bjorklund, L.J., 1967, Ground-water resources of northern Juab Valley, Utah: Utah State Engineer Technical Publication No. 17, 69 p.

Bjorklund, L.J., and Robinson, G.B., Jr., 1968, Ground-water resources of the Sevier River basin between Yuba Dam and Leamington Canyon, Utah: U.S. Geological Survey Water-Supply Paper 1848, 79 p.

Bodine, M.R., Jr., and Jones, B.F., 1986, The SALT NORM: A quantitative chemical-mineralogical characterization of natural waters: U.S. Geological Survey Water-Resources Investigations Report 86-4086.

Bouwer, Herman, 1989, The Bouwer and Rice slug test - an update: Ground Water, v. 27, no. 3, p. 304-309.

Bouwer, Herman, and Rice, R.C., 1976, A slug test for determining hydraulic conductivity of unconfined aquifers with completely or partially penetrating wells: Water Resources Research, v. 12, no. 3, p. 423- 428.

Brooks, L.E., and Stolp, B.J., 1995, Hydrology and simulation of ground-water flow in southern Utah and Goshen Valleys, Utah: Utah Department of Natural Resources Technical Publication No. 111, 96 p.

Chavez, Fray Angelico, translator, 1995, The Dominguez-Escalante Journal: edited by T.J. Warner, University of Utah Press, Salt Lake City, 153 p.

Christensen, R.C., Johnson, E.B., and Plantz, G.G., 1986, Manual for estimating selected streamflow characteristics of natural-flow streams in the Colorado River Basin in Utah: U.S. Geological Survey Water-Resources Investigations Report 85-4297, 39 p.

Cooper, H.H., Jr., and Jacob, C.E., 1946, A generalized graphical method for evaluating formation constants and summarizing wellfield history: Transactions of the American Geophysical Union, v. 27, no. 4, p. 526-534.

Cooper, H.H., Jr., Bredehoeft, J.D., and Papadopulos, I.S., 1967, Response of a finite-diameter well to an instantaneous charge of water: Water Resources Research, v. 3, no. 1, p. 263-269.

Craig, H., 1961, Isotopic variations in meteoric waters: Science, v. 133, p. 1702-1703.

Fenneman, N.M., 1931, Physiography of the western United States: New York, McGraw-Hill, 534 p.

Feth, J.H., Barker, D.A., Moore, L.G., Brown, R.J., and Veirs, C.E., 1966, Lake Bonneville: Geology and hydrology of the Weber Delta District, including Ogden, Utah: U.S. Geological Survey Professional Paper 518, 76 p.

Freeze, R.A., and Cherry, J.A., 1979, Groundwater: Prentice-Hall, Englewood Cliffs, New Jersey, 604 p.

Gates, J.S., 1987, Ground water in the Great Basin part of the Basin and Range Province, western Utah, *in*

Kopp, R.S., and Cohenour, R.E., eds., Cenozoic geology of western Utah, sites for precious metal and hydrocarbon accumulations: 1987 Symposium and Field Conference, Utah Geological Association Publication 16, p. 75-89.

Hantush, M.S., 1960, Modification of the theory of leaky aquifers: Journal of Geophysical Research, v. 64, p. 1043-1052.

Holmes, W.F., and Thiros, S.A., 1990, Ground-water hydrology of Pahvant Valley and adjacent areas, Utah: Utah Department of Natural Resources Technical Publication No. 98, 64 p.

Holser, W.T., 1977, Catastrophic chemical events in the history of the ocean: Nature, 267, p. 403-408.

Hood, J.W., and Waddell, K.M., 1968, Hydrologic reconnaissance of Skull Valley, Tooele County, Utah: Utah Department of Natural Resources Technical Publication No. 18, 57 p.

Hyatt, M.L., Skogerboe, G.V., Haws, F.W., and Austin, L.H., 1969, Hydrologic inventory of the Utah Lake drainage area: Utah Division of Water Resources and Utah Water Research Laboratory, 138 p.

Johnson, A.I., 1967, Specific yield—Compilation of specific yields for various materials: U.S. Geological Survey Water-Supply Paper 1662-D, 74 p.

Lambert, P.M., Mason, J.L., and Puchta, R.W., 1995, Hydrology of the Sevier-Sigurd ground-water basin and other ground-water basins, central Sevier Valley, Utah: Utah Department of Natural Resources Technical Publication No. 103, 181 p.

Lohman, S.W., 1979, Ground-water hydraulics: U.S. Geological Survey Professional Paper 708, 70 p.

Mayo, A.L., Nielsen, P.J., Loucks, Mark, and Brimhall, W.H., 1992, The use of solute and isotope chemistry to identify flow patterns and factors which limit acid mine drainage in the Wasatch Range, Utah: Ground Water, v. 30, no. 2, p. 243-249.

McDonald, M.G., and Harbaugh, A.W., 1988, A modular three-dimensional finite-difference ground-water flow model: Techniques of Water-Resources Investigations of the U.S. Geological Survey, book 6, chap. A1.

Meinzer, O.E., 1911, Ground water in Juab, Millard, and Iron Counties, Utah: U.S. Geological Survey Water-Supply Paper 277.

Plummer, L.N., Jones, B.F., and Truesdell, A.H., 1976, WATEQF--A FORTRAN IV version of WATEQ, a computer program for calculating chemical equilibria of natural waters: U.S. Geological Survey Water-Resources Investigations Report 76-13, 61 p.

Plummer, L.N., Prestemon, E.C., and Parkhurst, D.L., 1991, An interactive code (NETPATH) for modeling NET geochemical reactions along a flow PATH: U.S. Geological Survey Water-Resources Investigations Report 91-4078, 225 p.

Prudic, D.E., 1989, Documentation of a computer program to simulate stream-aquifer relations using a modular, finite-difference, ground-water flow model: U.S. Geological Survey Open-File Report 88-729, 113 p.

Razem, A.C., and Steiger, J. I., 1981, Ground-water conditions in Tooele Valley, Utah: Utah Department of Natural Resources Technical Publication No. 69, 95 p.

Robinson, T.W., 1958, Phreatophytes: U.S. Geological Survey Water-Supply Paper 1423, 84 p.

Steiger, J.I., 1995, Selected hydrologic data for Juab Valley, Utah, 1935-94: U.S. Geological Survey Open-File Report 95-101, 85 p.

Theis, C.V., Brown, R.H., and Meyer, R.R., 1963, Estimating the transmissibility of aquifers from the specific capacity of wells, *in* Bentall, Ray, compiler, Methods of determining permeability, transmissibility, and drawdown: U.S. Geological Survey Water-Supply Paper 1536-I, p. 331-340.

U.S. Department of Agriculture, 1969, Water budget analysis, Sevier River basin, Utah, Appendix IV, 92 p.

White, D.E., Hem, J.D., and Waring, G.A., 1963, Chapter F, Chemical Composition of Subsurface waters *in* Fleisher, Michael, ed., Data for Geochemistry, Sixth Edition: U.S. Geological Survey Professional Paper 440-F, 67 p.

Witkind, I.J., and Weiss, M.P., 1991, Geologic map of the Nephi 30' x 60' quadrangle, Carbon, Emery, Juab, Sanpete, Utah, and Wasatch Counties, Utah: U.S. Geological Survey Miscellaneous Investigations Series Map I-1937, scale 1:100,000.

Witkind, I.J., Weiss, M.P., and Brown, T.L., 1987, Geologic map of the Manti 30' x 60' quadrangle, Carbon, Emery, Juab, Sanpete, and Sevier counties, Utah: U.S. Geological Survey Miscellaneous Investigations Series Map I-1631, scale 1:100,000.

Zoback, M.L., 1992, Superimposed late Cenozoic, Mesozoic, and possible Proterozoic deformation along the Wasatch fault zone in central Utah, *in* Gori, P.L. and Hays, W.W., eds., Assessment of regional earthquake hazards and risk along the Wasatch Front, Utah: U.S. Geological Survey Professional Paper 1500-A-J, p. E1-E20.

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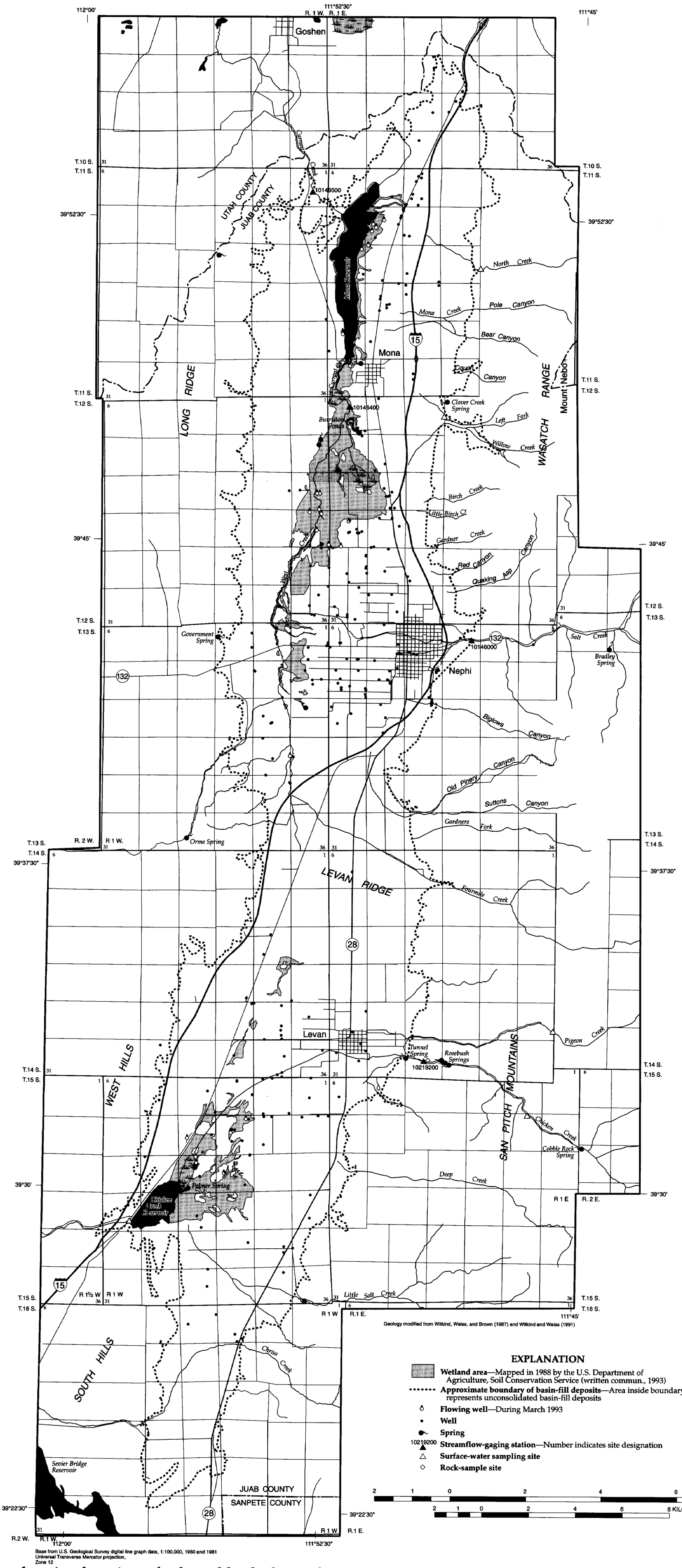
CENTRAL UTAH WATER CONSERVANCY DISTRICT
AND THE EAST JUAB WATER CONSERVANCY DISTRICT

TECHNICAL PUBLICATION NO. 114
Hydrologic-data and rock-sample sites—PLATE 1
Thiros, S.A., and others, 1996, Hydrology and simulation of ground-water flow in Juab Valley, Juab County, Utah



Map showing location of selected hydrologic-data sites and rock-sample site, Juab Valley, Utah, 1935–94

By
Judy I. Steiger
1996