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GROUND-WATER CONDITIONS IN SALT LAKE VALLEY, UTAH, 1969-83, AND PREDICTED EFFECTS OF INCREASED WITHDRAWALS FROM WELLS

by

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Prepared by
the United States Geological Survey
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Division of Water Rights

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CONVERSION FACTORS

For use of readers who prefer to use metric units, conversion factors for terms used in this report are listed below:

Multiply	By	To obtain
inch	25.40	millimeter
	2.54	centimeter
foot	0.304	meter
mile	1.609	kilometer
acre	4,047	square meter
	0.004047	square kilometer
square mile	2.590	square kilometer
acre-foot	1,233	cubic meter
cubic foot per second	0.02832	cubic meter per second
square feet per second	0.0929	square meter per second
square feet per day	0.0929	square meter per day

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U.S. Geological Survey

ABSTRACT

This report was prepared in cooperation with several organizations in the Salt Lake Valley and with the Central Utah Water Conservancy District to present results of a study to determine changes in the ground-water conditions in Salt Lake Valley, Utah, from 1969 to 1983, and to predict the aquifer response to projected withdrawals. The average annual recharge and discharge from the ground-water reservoir in Salt Lake Valley, Utah, during 1969-82 were estimated to be about 352,000 and 353,000 acre-feet per year. Withdrawals from wells increased from 107,000 acre-feet per year during 1964-68 to 117,000 acre-feet per year during 1969-82. The greatest increase in use was for public supply and institutions which increased from 35,000 acre-feet per year during 1964-68 to 46,700 acre-feet per year during 1969-82.

From 1969 to 1983 water levels declined from 5 to 15 feet in the southeast part of the valley where pumpage from large public-supply wells was greater during 1969-82 than during previous years. From February-March 1969 to February-March 1983 the quantity of ground water in storage in Salt Lake Valley increased by about 33,000 acre-feet.

A digital-computer model was calibrated to simulate, in three-dimensions, the ground-water flow in the principal and shallow-unconfined aquifers in Salt Lake Valley. Simulations were made to project the response to continuing withdrawals through 2020. Alternative pumping rates used were (1) the 1982 rate of pumpage and (2) increasing the 1982 rate of pumpage by 65,000 acre-feet. The simulation at the increased rate of pumpage indicated that drawdowns would reach 40-60 feet in the area east of Sandy. About 75 percent of the increased withdrawal was salvaged from water that otherwise would have been discharged to the Jordan River and its tributaries.

INTRODUCTION

This report was prepared by the U.S. Geological Survey in cooperation with the following organizations that contributed to the investigation through the Utah Department of Natural Resources: Salt Lake County Water Conservancy District, Central Utah Water Conservancy District, Granger-Hunter Improvement District, Magna Water Co. and Improvement District, City of Midvale, City of Murray, Salt Lake City Department of Public Utilities, City of Sandy, City of South Salt Lake, Taylorsville-Bennion Improvement District, City of West Jordan, Holladay Water Co., and White City Water Co. The period of study on which this report is based is July 1981 to December 1983, but the period of record covered by the report is 1969-83.

A detailed study (Hely and others, 1971) of the hydrologic system in the Salt Lake Valley (fig. 1) provided a comprehensive description of the ground-water system and predictions of the effects of future development based on an analog model. The purposes of this study were to determine changes in ground-water conditions in the Salt Lake Valley since the study of Hely and others (1971) and to predict the response of the ground-water system to continued or increased withdrawals. Annual estimates of the components of recharge to and discharge from the ground-water reservoir were determined for 1969-82. Estimates were made from data gathered during field investigations in 1981-83, from the results of seepage studies on major canals during 1982-83 (Herbert and others, 1984), and from continuous records of the flow of the Jordan River at various sites. Estimates for some elements of the water budget were not modified during this study, and values presented by Hely and others (1971, p. 119, and 135) were used for the 1969-82 water budget. Some of these estimates were revised during calibration of the digital model.

This report is the second of five planned reports. The first report (Seiler and Waddell, 1984) from this study described the results of an investigation of the shallow-unconfined aquifer during 1982-83. Subsequent planned reports will describe sources of contamination to the ground water and chemical-quality changes during 1969-83, including predicted effects of increased withdrawals on the chemical quality of the ground water; will document the digital-computer model; and will present the hydrologic data collected from 1969 to 1985.

WELL- AND SPRING-NUMBERING SYSTEM

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. By the land-survey system, the State is divided into four quadrants by the Salt Lake base line and meridian, and these quadrants are designated by the uppercase letters A, B, C, and D, indicating the northeast, northwest, southwest, and southeast quadrants, respectively. Numbers designating the township and range (in that order) follow the quadrant letter, and all three are enclosed in parentheses. The number after the parentheses indicates the section, and it is followed by three letters indicating the quarter section, the quarter-quarter section, and the quarter-quarter-quarter section--generally 10 acres[1]; the 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 denotes a spring. If a well or spring cannot be located within a 10-acre tract, one or two location letters are used and the serial number is omitted. Thus, (D-2-1)34acb-1 designates the first well constructed or visited in the NW¼ SW¼ NE¼ sec. 34, T. 2 S., R. 1 E. The numbering system is illustrated in figure 2.

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[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.

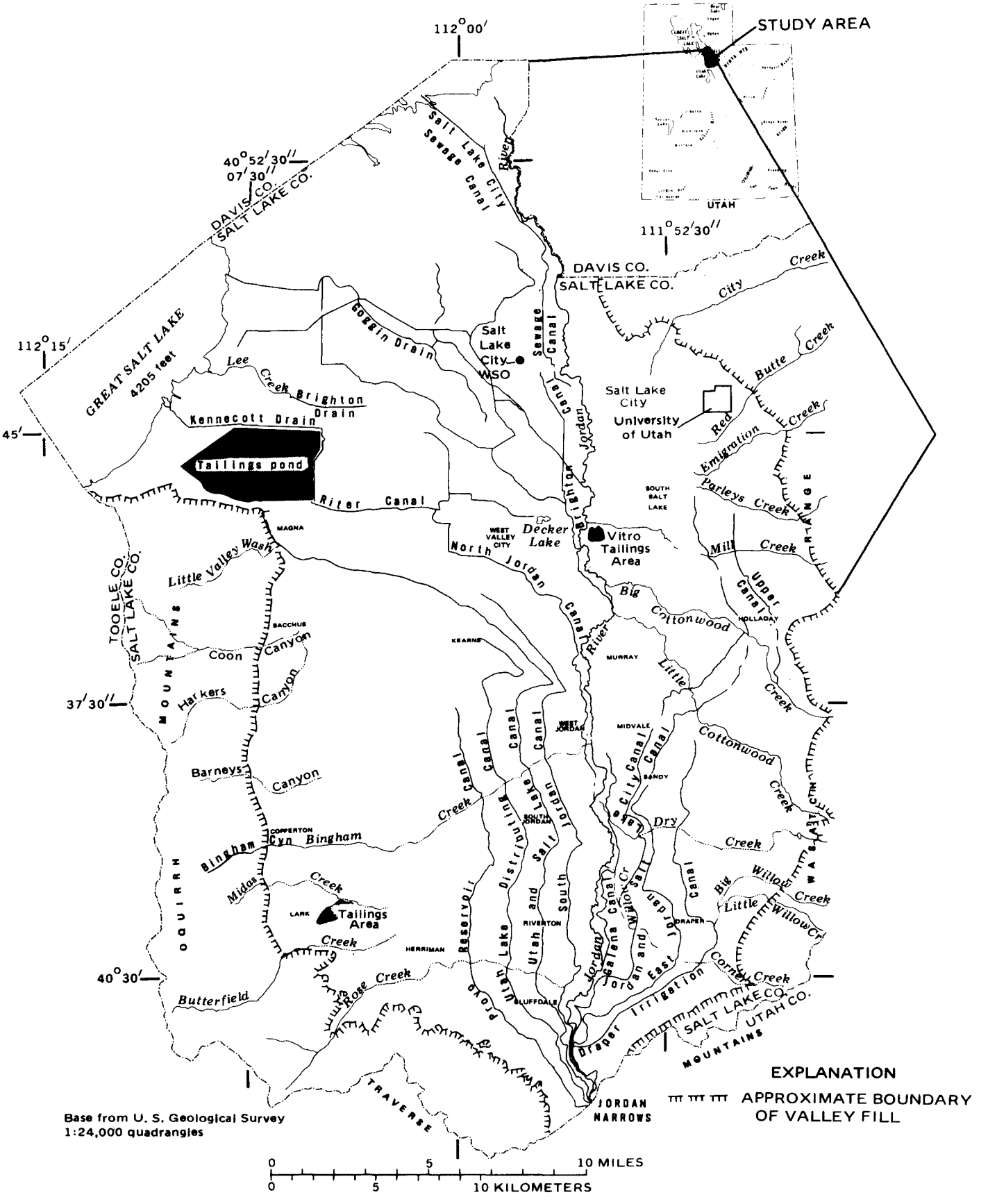

Figure 1.—Location of the study area.

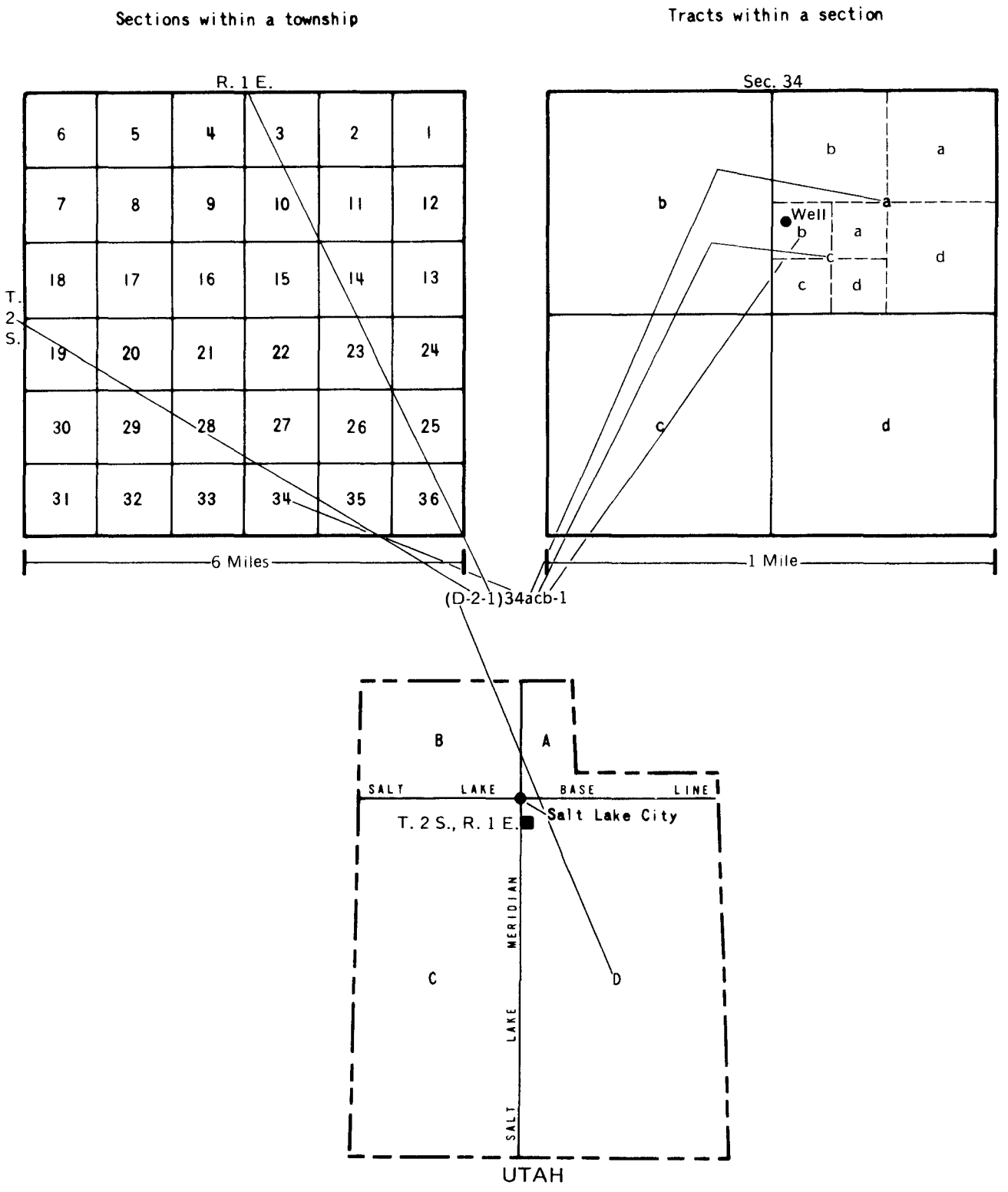

Figure 2.—Well - and spring - numbering system used in Utah.

GROUND-WATER CONDITIONS

Ground water in Salt Lake Valley (Hely and others, 1971, p. 107-111) occurs in valley fill in (1) a confined (artesian) aquifer, (2) a deep-unconfined aquifer between the artesian aquifer and the mountains, (3) a shallow-unconfined aquifer overlying the artesian aquifer, and (4) locally in unconfined-perched aquifers (fig. 3). All the aquifers consist of unconsolidated materials of Quaternary age that are connected hydraulically to some degree; thus, together they compose the ground-water reservoir in Salt Lake Valley. The "principal aquifer" in the valley consists of the confined and the deep-unconfined aquifers.

The confined aquifer attains a maximum thickness of more than 1,000 feet in the northern part of the valley. Underlying the confined aquifer are relatively impermeable semiconsolidated and consolidated rocks of Tertiary and pre-Tertiary age (Arnow and others, 1970). Within the confined aquifer relatively thin beds or lenses of fine-grained material, which may be as much as 20 feet thick but usually are not more than a few feet thick, tend to confine water in beds of sand or gravel. The fine-grained material is slightly to moderately permeable and is discontinuous, therefore, there is appreciable movement of water between the more permeable beds of sand and gravel. The hydraulic connection between different water-bearing beds in the confined aquifer has been demonstrated many times during aquifer tests (Hely and others, 1971, p. 109).

The principal aquifer generally yields water readily to wells. The most productive wells are in the deep-unconfined aquifer near the mountains where the aquifer consists of thick, coarse-grained deposits.

The confined aquifer is overlain by relatively impermeable deposits of clay, silt, and fine sand, which collectively act as a confining bed that ranges in thickness from about 40 to 100 feet. This confining bed, however, is either absent or above the potentiometric surface in a band of varying width adjacent to the mountains at the edges of the valley. Much of the water that reaches the confined aquifer first passes through the deep unconfined zone at each side of the valley.

The shallow-unconfined aquifer overlies the confining bed that overlies the confined aquifer. The shallow aquifer is composed principally of clay, silt, and fine sand; and in some parts of the valley, this aquifer has permeability only slightly greater than that of the underlying confining bed. Thus, the exact thickness of the shallow aquifer is unknown, but the maximum thickness probably is about 50 feet. The shallow-unconfined aquifer has a smaller areal extent than the principal aquifer, but it is underlain everywhere by the principal aquifer (fig. 4). The shallow-unconfined aquifer is recharged by leakage upward from the confined aquifer through the confining bed as well as by downward infiltration from precipitation, canals, irrigated lands, and streams. Because of the poor chemical quality of the water that it contains and its small yield to wells, the shallow aquifer seldom is used for water supply.

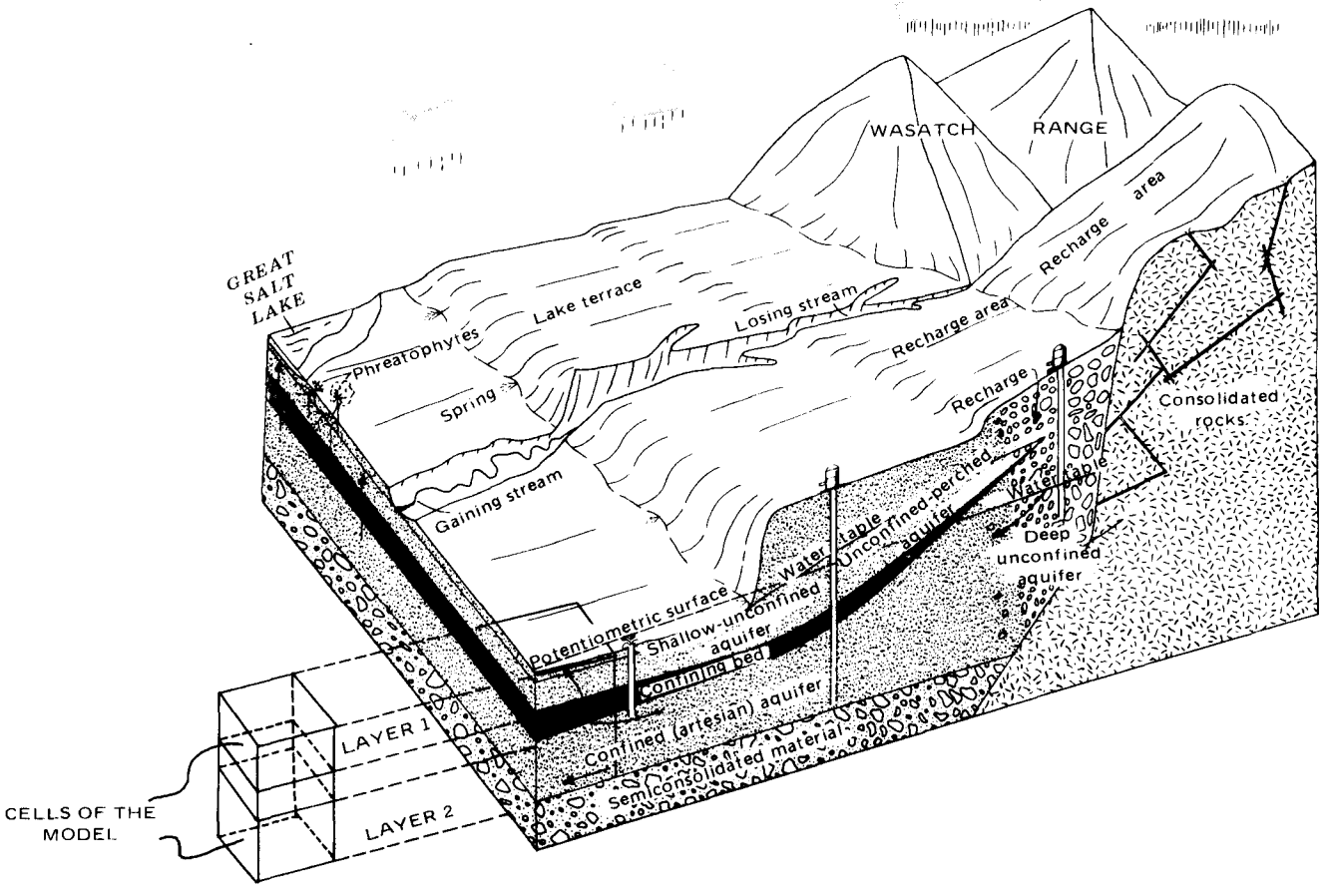

Figure 3.—Part of the ground-water reservoir in Salt Lake Valley and the relation of cells of the model to the physical system. (Modified from Hely, Mower and Harr, 1971, fig. 57.)

The perched aquifers are in areas where the bottom of the confining bed lies above the deep water table (fig. 3). Thus, an unsaturated zone exists between the deep water table and the perched water above it. The principal area of perched water is east of Midvale (fig. 4), but smaller, localized perched bodies of water are scattered around the valley. The perched aquifers supply water to only a few stock wells.

Recharge

Estimates of average recharge to the ground-water reservoir are summarized in table 1. The annual estimates of recharge during 1969-82 for the various sources are shown in figure 5. Some of the components of recharge shown in table 1 and figure 5 were derived during calibration of a digital model. The calibration of the model is discussed in section "Digital Model of Ground-Water Reservoir".

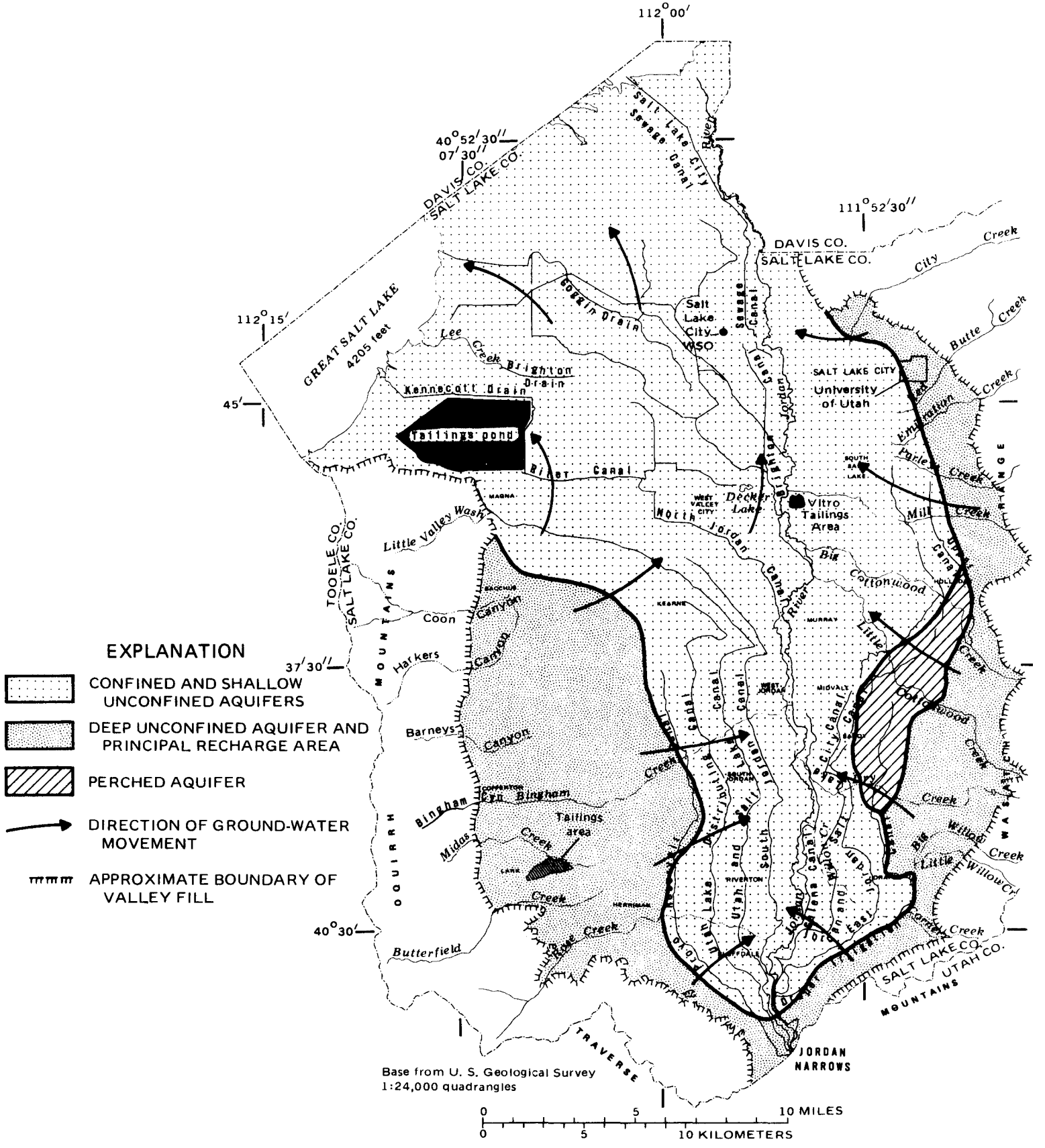

Figure 4.—Approximate areas in which ground water occurs in confined, shallow unconfined, deep unconfined, and perched aquifers in Salt Lake Valley (modified from Hely, Mower, and Harr, 1971, fig. 58).

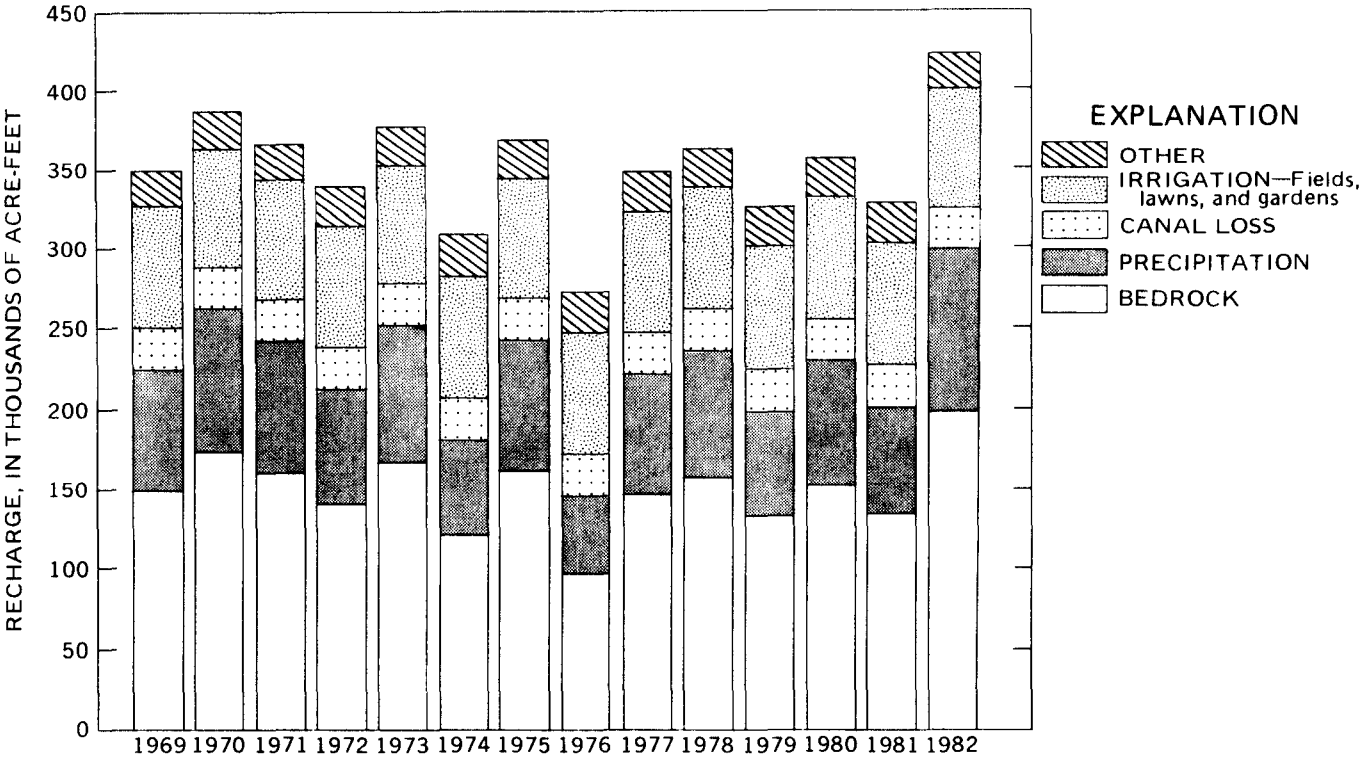

Figure 5.—Sources of ground-water recharge in Salt Lake Valley, 1969-82.

Seepage from Precipitation on the Valley Floor

During 1969-82 the average recharge from precipitation on the valley floor was estimated to be 71,000 acre-feet per year. To raise simulated water-levels in the shallow-unconfined aquifer, estimates presented by Hely and others (1971, p. 127-129), were increased during steady-state calibration of the digital model. Annual estimates were made from annual precipitation during 1969-82 (National Oceanic and Atmospheric Administration, 1970-83).

Table 1.--Ground-water recharge, in acre-feet per year, as reported from prior study, from data collected during 1969-83, and specified in or computed by the digital model

Source	Estimated for 1964-68 (modified from Hely and others, 1971, table 21)	Estimated for 1969-82 from data collected during 1969-83	Specified in or computed by digital model	
			For 1968, steady-state calibration	For 1969-82, average from transient calibration
Seepage from precipitation on the valley floor	60,000	------	70,000	71,000
Seepage from bedrock	135,000	------	154,000	157,000
Seepage from major canals	48,000	28,000	24,000	24,000
Seepage from irrigated fields	81,000	70,000	48,000	48,000
Seepage from lawns and gardens	17,000	30,000	28,000	28,000
Seepage from creek channels	20,000	------	16,000	16,000
Seepage from tailings pond north of Magna	2,400	------	0	0
Underflow in channel fill	1,500	------	1,500	1,500
Underflow at Jordan Narrows	2,500	------	2,500	2,500
Reinjection from air conditioning	2,000	4,000	2,000	4,000
Total (rounded)	369,000	------	346,000	352,000

Hely and others (1971, p. 127) computed an average recharge of 60,000 acre-feet per year during 1964-68. They computed recharge as the difference between the balance of precipitation available for evapotranspiration and ground-water recharge (454,000 acre-feet) and evapotranspiration of precipitation (394,000 acre-feet), or recharge from precipitation = 454,000 - 394,000 = 60,000 acre-feet. Only a 5 percent error in either value would cause the estimate of recharge to be in error by 20,000 acre-feet.

During this study, the value of 60,000 acre-feet was used as the initial estimate for 1968 for the steady-state calibration of the digital model. The value was increased to 70,000 acre-feet because the simulated water levels in the shallow-unconfined aquifer were several feet below the observed levels. As a result, evapotranspiration as computed by the digital model was considerably less than reported by Hely and others (1971, p. 135). Most of the additional recharge from precipitation was added to the flat lands in the northern part of the valley where less surface runoff was assumed to indicate more potential for recharge to the shallow-unconfined aquifer.

Estimates of recharge from precipitation for yearly time intervals were made by assuming that the recharge determined for the steady-state calibration varied directly with the factor computed from the ratio of annual precipitation (Py) to average annual precipitation for 1969-82 (P1969-82). Thus, the recharge (RPy) from precipitation (Py) for year (y) was computed as:

$$RPy = \frac{Py(RP1968)}{P1969\text{-}82} \qquad (1)$$

where RP1968 was 70,000 acre-feet as determined from the steady-state model calibration (table 1, column 4).

The ratios of annual precipitation, Py/P1969-82, were computed from records at the Salt Lake City WSO (International Airport) and Silver Lake Brighton (fig. 6). The Salt Lake City WSO (International Airport) is in the northern part of the Salt Lake Valley, and it was considered representative of the valley floor. The average annual precipitation at Salt Lake City WSO during 1969-82 was 17 inches. Silver Lake Brighton is in the Wasatch Range, which forms the eastern boundary of the study area. The average annual precipitation at Silver Lake Brighton during 1969-82 was 44 inches. A mountain site was included because the factor was also used to estimate seepage from bedrock into the valley fill, and there was not sufficient difference between the ratios at the two sites to justify using separate factors. So an average ratio, Py/P1969-82, based on annual precipitation at Salt Lake City WSO and Silver Lake Brighton was used to estimate annual recharge from precipitation (RPy) during 1969-82.

The average ratio of annual precipitation to average 1969-82 precipitation, as computed at the two sites, ranged from 0.56 in 1976 to 1.35 in 1982. Annual precipitation ranged from 44 percent below to about 35 percent above the average 1969-82 annual precipitation. Seepage from precipitation, as computed using equation (1) ranged from 40,000 to 97,000 acre-feet per year and averaged 71,000 acre-feet per year during 1969-82.

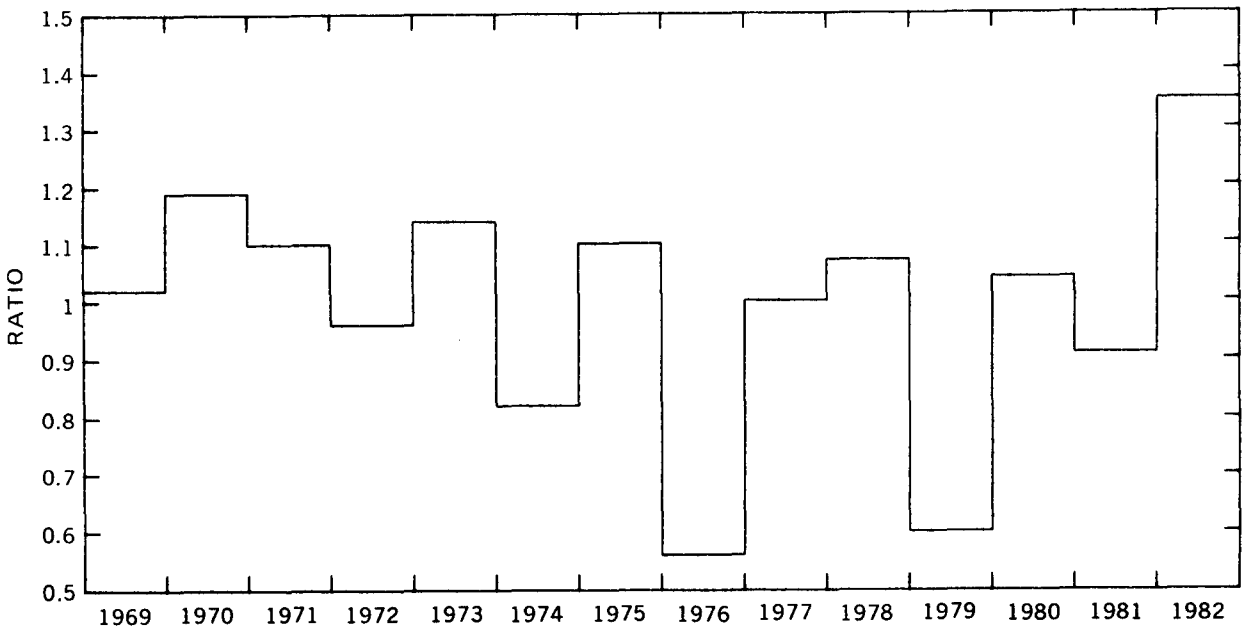

Figure 6.—Ratio of annual precipitation to average annual precipitation for 1969-82. Ratios computed from records of National Oceanic and Atmospheric Administration at Salt Lake City WSO (International Airport) and Silver Lake Brighton.

Seepage from Bedrock

During 1969-82 the average recharge to the valley fill as seepage through bedrock was estimated to be 157,000 acre-feet per year. During the steady-state calibration, recharge from seepage from bedrock was assumed to be the quantity of water necessary to maintain water levels at the boundary between valley fill and bedrock. Similar to the method used for estimating recharge from precipitation, annual estimates of bedrock recharge were made from precipitation records during 1969-82. Annual recharge from bedrock (RBy) was assumed to vary directly with the factor computed from the ratio of annual precipitation (Py) to average 1969-82 annual precipitation P1969-82, or

$$RBy = \frac{Py(RB1968)}{P1969-82} \qquad (2)$$

where RB1968 was 154,000 acre-feet of recharge determined during the steady-state calibration (table 1, column 4). Seepage from bedrock, as computed from equation (2), ranged from 85,000 to 205,000 acre-feet per year and averaged 157,000 acre-feet per year for 1969-82. Additional discussion of the procedures for computing the seepage from bedrock is presented in the section "Calibration of Model".

Seepage from Major Canals

The seepage from major canals and ditches in Salt Lake Valley was determined by Herbert and others (1984) during 1982-83 to be 28,000 acre-feet per year. Most of this seepage was from the Provo Reservoir, Utah Lake Distributing, and Utah and Salt Lake Canals in the southwest part of the valley. For 1964-68, Hely and others (1971, p. 119) estimated seepage losses from canals to be 48,000 acre-feet, or about 20,000 acre-feet more than the value determined by Herbert and others (1984). Hely and others (1971, p. 124) measured seepage from only one canal and extrapolated the losses over five other major canals. The seepage that was determined during 1982-83 was believed to be representative of a lower limit for canal losses because the measurements were made during years of above normal precipitation when the water levels in the shallow-unconfined aquifer were probably higher than normal throughout most of the valley. Even though the 1982-83 measurements were made during years that were conducive to minimizing seepage losses, the measurements probably provide more accurate estimates than those of Hely and others (1971, p. 125). During calibration of the digital model, the seepage losses from canals were reduced to 24,000 acre-feet per year.

Seepage from Irrigated Fields and Lawns and Gardens

The recharge from seepage from irrigated fields was estimated from data presented by Hely and others (1971, p. 119 and fig. 77) for 1964-68 and from the change in irrigated acreage by 1981 (University of Utah Research Institute, 1982). The total irrigated land in 1964-68 was 70,000 acres and in 1981 (fig. 7) the irrigated acreage was 52,000 acres, which amounts to a decrease of 18,000 acres. The seepage from irrigated fields during 1964-68 was 81,000 acre-feet per year (Hely and others, 1971, p. 119). By assuming that the distribution and type of crops, irrigation practices, and the rate of evapotranspiration during 1981 was the same as during 1964-68, the seepage for 1981 can be estimated by a factor proportional to the decrease in irrigated acreage, as follows:

$$\text{Seepage from irrigated fields during 1981} = \frac{81{,}000 \text{ acre-feet} \times 52{,}000 \text{ acres}}{70{,}000 \text{ acres}}$$
$$= 60{,}000 \text{ acre-feet}$$

Then, the average recharge rate from irrigated fields for 1969-82 was computed as the average of 81,000 and 60,000, or 70,000 (rounded) acre-feet per year (table 1, column 2).

During steady-state calibration of the digital model, water levels in several irrigated areas in the southern part of the valley were simulated to be much higher than observed water levels, and it was obvious that too much recharge was being applied in the model. Thus, estimated seepage from irrigation was reduced until simulated water levels were in agreement with observed levels. The final quantity of seepage from irrigation used in the model was 48,000 acre-feet per year. For the transient simulations, seepage from irrigation was assumed to be constant throughout 1969-82.

The recharge from seepage from lawns and gardens during 1981 was assumed to be the 17,000 acre-feet estimated for 1964-68 by Hely and others (1971, p. 119) plus the amount that would have been contributed from new urban areas

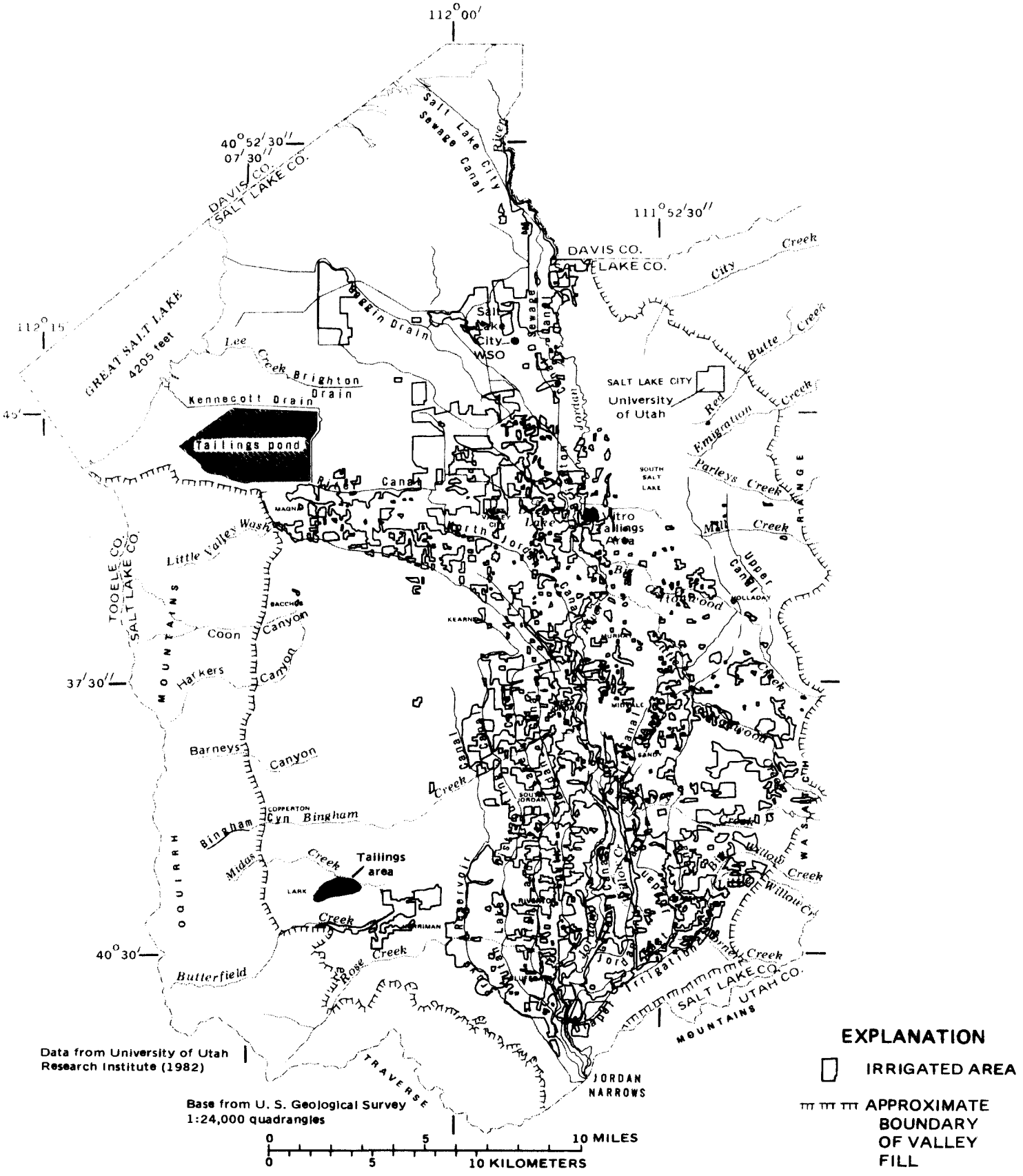

Figure 7.—Irrigated areas in Salt Lake Valley, 1981.

developed on land that was formerly used for irrigation and on formerly vacant land. Following the assumption of Hely and others (1971, p. 126) that seepage from lawns and gardens is about the same as from irrigated fields, the increase of seepage from lawns and gardens on formerly irrigated fields and vacant land was estimated. The reduction of irrigated lands between 1964-68 and 1981 represents a proportional increase in new urban areas. Thus, seepage from lawns and gardens in new urban areas in 1981 is equal to the reduction in seepage from irrigated areas between 1964-68 and 1981, 81,000-70,000 acre-feet, or 11,000 acre-feet.

The extent of urbanization on formerly vacant or nonirrigated land was computed from land-use data for Salt Lake County (Wasatch Front Regional Council, 1982), and the additional seepage from this area was estimated to be about 4,000 acre-feet during 1980 and to average about 2,000 acre-feet per year during 1969-82. Thus, the total seepage from lawns and gardens during 1969-82 was 17,000+11,000+2,000 acre-feet or 30,000 acre-feet per year.

During steady-state calibration of the digital model, estimated seepage from lawns and gardens was reduced to 28,000 acre-feet per year. For transient simulations, seepage from lawns and gardens was assumed to be a constant amount of 28,000 acre-feet per year.

Seepage from Creek Channels, Tailings Pond, and Underflow in Channel Fill, and at Jordan Narrows

During 1969-82, no new data were collected to evaluate recharge from seepage from creek channels, tailings pond, or from underflow in channel fill in the mountain canyons. The quantities of recharge reported by Hely and others (1971, table 21) for these sources are shown in column 2 of table 1. During the steady-state calibration of the digital model, seepage from creek channels was reduced by 4,000 acre-feet per year, seepage from tailings pond was eliminated, and underflow in the channel fill and at the Jordan Narrows remained the same. During the transient simulation, the seepage from creek channels and the underflow in channel fill was assumed to be constant from year-to-year during 1969-82.

Movement

Hely and others (1971, p. 129-131) provided a detailed description of the movement of ground water in Salt Lake Valley. Water-level contours for 1983 (fig. 8) indicate that the general pattern of ground-water movement in the principal aquifer during 1983 was about the same as reported by Hely and others (1971, plate 1). During 1981-82, however, additional data were collected to improve the description of vertical movement through the confining layers into the shallow-unconfined aquifer and of horizontal movement in the shallow-unconfined aquifer.

Seiler and Waddell (1984, pl. 1) show water-level contours and direction of flow in the shallow-unconfined aquifer during December 1982. The general direction of flow in the shallow aquifer is toward the Jordan River, except in the extreme northwest part of the valley where it moves toward the Great Salt Lake. The difference in hydraulic head between the shallow-unconfined aquifer and the confined aquifer during 1981-83 caused an upward movement of water from the confined aquifer to the shallow-unconfined aquifer.

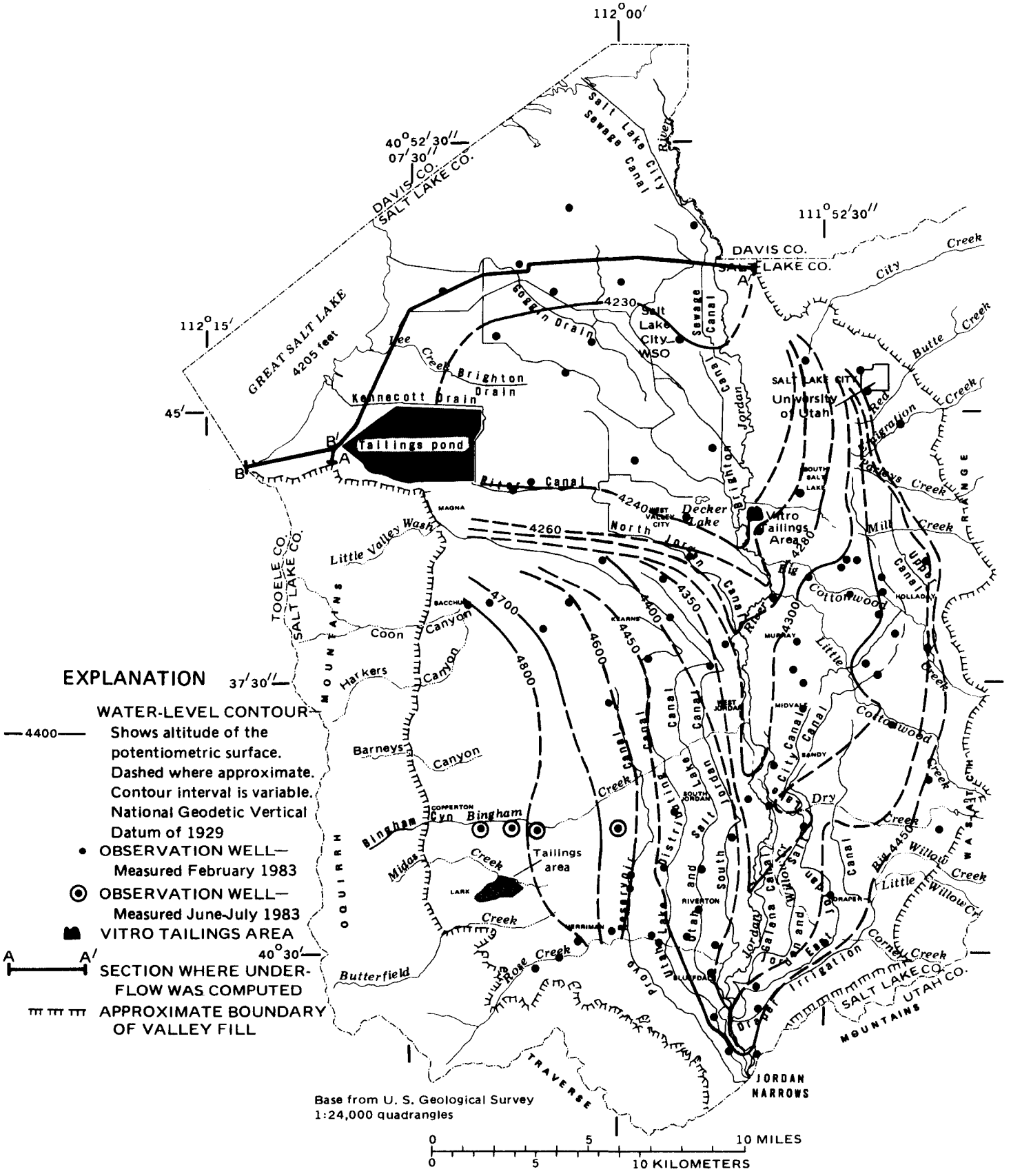

Figure 8.—Water-level contours for the principal aquifer, 1983.

During 1983, the vertical hydraulic gradient was determined from water levels in wells in sections 25 and 26 of Township 1 South and Range 1 West, near the Vitro tailings area (fig. 1). The altitude of the potentiometric surface was found to increase with the depth of the perforated zone, which is consistent with the concept of upward movement of water. The vertical gradient of the potentiometric surface increased sharply at a depth of about 58 feet (fig. 9). Specific conductance to a depth of about 38 feet exceeded 14,000 microsiemens per centimeter at 25° Celsius. Below a depth of about 58 feet, the specific conductance was less than 2,000 microsiemens per centimeter at 25 $^{\circ}$C. The slope of the curves were used to delineate the shallow-unconfined aquifer, the confining layer, and the top of the principal aquifer.

If it is assumed that the bottom of the shallow-unconfined aquifer coincides with the depth at which the specific conductance of the water began to decrease, then the bottom of the shallow-unconfined aquifer can be approximated as being at the midpoint between the depth to more saline and less saline water, or about 50 feet. Also, assuming that the top of the confining bed occurs where the change of head gradient is the greatest, the top can be approximated from the midpoint between the depths of 38 and 58 feet, or about 50 feet. This is the same depth as determined from the water-quality data.

Another change in the head gradient at the Vitro tailings area occurs between 70 and 120 feet (fig. 9). This change was attributed to the presence of permeable material in the principal aquifer; thus, the point of change approximates the bottom of the confining layer. It was assumed, for lack of more definitive information, that the bottom of the confining layer is at the midpoint of this range in depth, or at 95 feet. Thus the effective thickness of the confining bed is about 45 feet.

Water-level Changes during 1969-83

Water-level changes in the principal aquifer from February-March 1969 to February-March 1983 are shown in figure 10. Water levels declined from 5 to 15 feet in the southeast part of the valley, where pumpage from large public supply wells was greater during 1969-82 than during 1964-68. Downgradient from that area, water levels declined as much as 5 feet in a band extending toward the northwest.

Water levels rose as much as 12 feet in the northeastern part of the valley during 1969-83. The reasons for these rises are not clear. Pumpage records for wells (D-1-1)4add-1 and (D-1-1)4cbd-1 indicate that withdrawal more than doubled between 1969 and 1982, whereas water levels in well (D-1-1)5aaa-1 showed an overall increase of about 8 feet between 1969 and 1983. Obviously, recharge to the area, perhaps from precipitation or from water applied to lawns and gardens, has exceeded the increase of withdrawals.

Water levels also generally rose in the southwest part of the valley as much as 12 feet during 1969-83. West of the Jordan River rises as much as 6 feet probably are partly due to seepage from canals (Herbert and others, 1984). The rises of 6-12 feet in a large area east of the Oquirrh Mountains where there is relatively little withdrawal of water from wells probably resulted from above average precipitation during 1969-83.

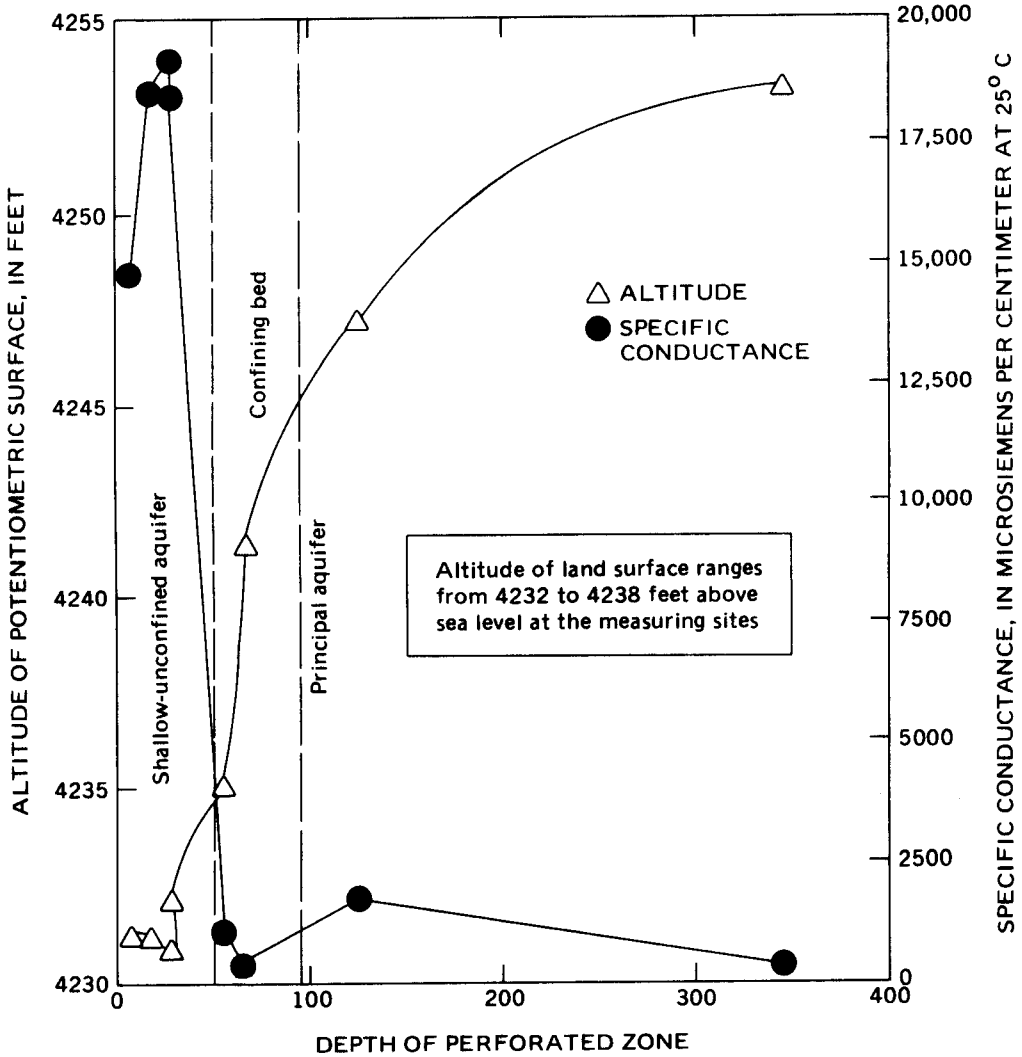

Figure 9.—Relation of depth of perforated zone to altitude of potentiometric surface and specific conductance of water in wells near the Vitro tailings area. (See fig. 1 for location.)

Storage

From February-March 1969 to February-March 1983, the quantity of water stored in the principal aquifer in Salt Lake Valley increased by about 33,000 acre-feet. The increase of storage is due primarily to rises of water levels in the unconfined part of the principal aquifer in the southwest and northeast parts of the valley (fig. 10). The change in storage was determined by dividing the valley into 11 areas and then computing the average water-level change during 1969-83 and the storage coefficient for each area. The total change in storage then was calculated from the sum of the changes in storage computed for each area. Data were not adequate to determine changes of storage in the shallow-unconfined aquifer. Comparison of maps showing average depth to water in 1968 (Hely and others, 1971, fig. 80) and in 1982 (Seiler and Waddell, 1984, plate 2) indicate that water levels were within the same ranges during 1968 and 1982.

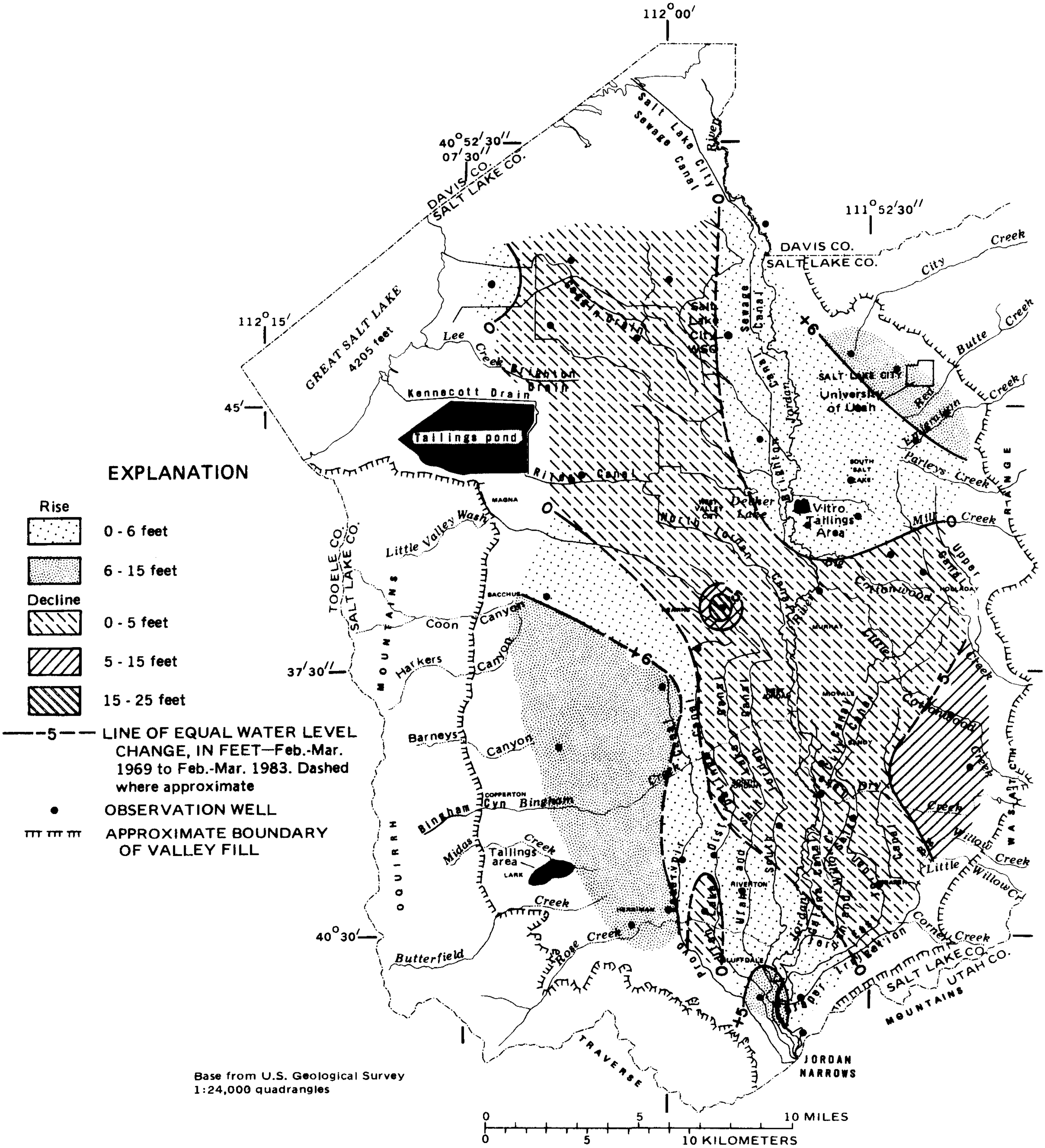

Figure 10.—Change of water levels in the principal aquifer from February-March 1969 to February-March 1983.

Hely and others (1971, p. 131-134) reported that 60,000,000 acre-feet of ground water was in storage in the valley in 1969. However, they also pointed out that the quantity of water in storage is much greater than the quantity that is readily available for withdrawal by means commonly in use. A graph prepared by Hely and others (1971, fig. 65) showed that the volume of water in storage decreased by about 13,000 acre-feet per foot of water-level decline at 1969 levels, but the decrease in volume of water in storage per foot of water-level decline is greater as water levels decline. The storage change per foot of water-level decline increases as water levels decline because an increasingly larger part of the aquifer becomes unconfined. Through use of their graph, it was estimated that if the quantity of water in storage were depleted by 130,000 acre-feet, the average water level across the valley would decline by 10 feet, and if depleted by 1,500,000 acre-feet the decline would be about 100 feet.

The change in storage computed by the model is the difference between recharge and discharge, or -1,000 acre-feet per year (table 2). The difference between the computed depletion in storage (-1,000 acre-feet per year) and the observed increase in storage (2,300 acre-feet per year) is 3,300 acre-feet per year and represents the overall error, which is less than 1 percent of the annual discharge from the valley. The error results from the approximations and generalizations involved in the estimates of recharge and discharge.

Discharge

Estimates of ground-water discharge are summarized in table 3. The annual estimates of discharge during 1969-82 from the various sources are shown in figure 11. Some of the components of discharge shown in table 3 and figure 11 were derived during calibration of a digital model. The calibration of the model is discussed in section "Digital Model of Ground-Water Reservoir".

Table 2.--Ground-water budget for Salt Lake Valley, 1969-82

	Mean annual quantity in acre-feet
Recharge (column 5, table 1)	352,000
Discharge (column 5, table 3)	353,000
Computed change in storage	-1,000
Measured change in storage	2,300

Table 3.--Ground-water discharge, in acre-feet per year, as reported from prior study, from data collected during 1969-83, and specified in or computed by the digital model

Mode of discharge	Estimated for 1964-68 (Hely and others, 1971, table 22)	Estimated for 1969-82 from data collected during 1969-83	Specified in or computed by digital model	
			For 1968, steady-state calibration	For 1969-82, average from transient calibration
Wells	107,000	------	102,000	117,000
Seeps, springs, and drains				
Inflow to Jordan River and tributaries	170,000	155,000	146,000	143,000
Major canals	------	13,000	10,000	10,000
Inflow to drains	5,000	------	5,000	5,000
Spring flow diverted for use	19,000	------	19,000	19,000
Thermal springs	2,000	------	2,000	2,000
Evapotranspiration	60,000	------	54,000	54,000
Subsurface outflow to Great Salt Lake	4,000	3,100	7,200	2,600
Total (rounded)	367,000	------	346,000	353,000

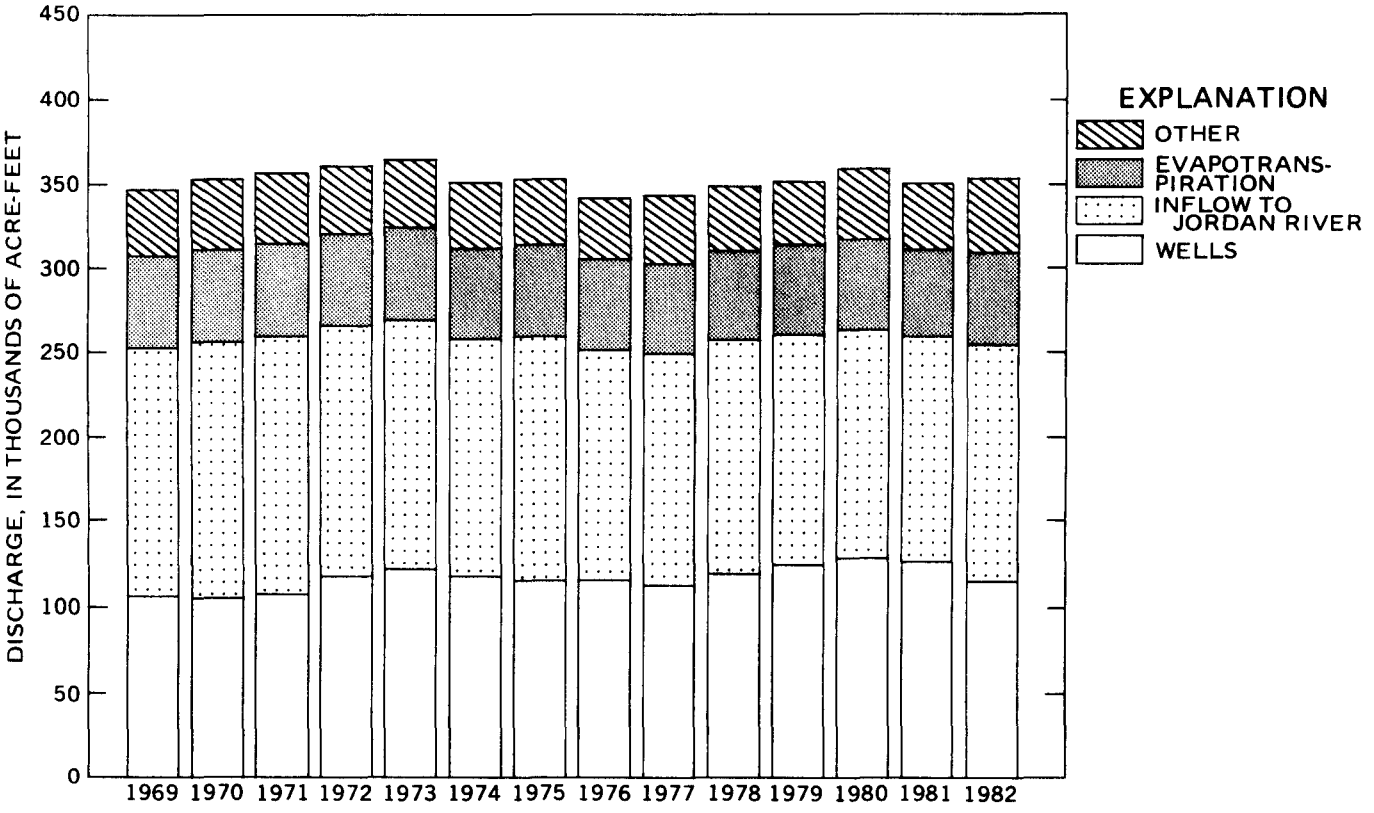

Figure 11.—Ground-water discharge from Salt Lake Valley, 1969-82.

Wells

During 1969-82, annual withdrawal by wells ranged from 105,000 to 129,000 acre-feet and averaged 117,000 acre-feet. Previously published values for annual withdrawal from wells in Salt Lake Valley were reviewed and revised as part of this investigation. The revision is discussed in Seiler and others (1985). Hely and others (1971, p. 140-141 and fig. 66) presented a summary of annual ground-water withdrawal from wells during 1931-68. The summary indicated a range from 38,000 acre-feet in 1931 to 118,000 in 1966. Withdrawals began to level off about 1964, and averaged 107,000[1] acre-feet per year during 1964-68. Average withdrawals during 1969-82 were about 10,000 acre-feet per year greater than during 1964-68.

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[1]The average withdrawals for 1964-68 were revised by Bolke and others (1973, table 3) to 110,000 acre-feet; however, the withdrawals were not tabulated by type of use, thus, Hely and others (1971) value of 107,000 acre-feet were used for purposes of this report.

The change in total withdrawal during 1969-82 was accompanied by changes in withdrawal for different uses. A comparison of average withdrawals for four types of use during 1969-82 with that of 1964-68 are shown in the following summary:

Use	Average withdrawal in acre-feet per year 1964-68	1969-82
Domestic and stock	30,000	27,500
Public supply and institutions	35,000	46,700
Industry and air conditioning	37,000	38,300
Irrigation	5,000	4,400
Total (rounded)	107,000	117,000

Withdrawals for public supply and institutions have shown the largest increase. The withdrawal for industry and air conditioning increased slightly during the 1970's (fig. 12) while withdrawal for other uses declined. Also, data in table 4 show that 50 wells, or 7 percent of the 694 wells completed during 1970-81, were for public supply and institutions. Prior to 1970, only 397 wells, or 3 percent of 11,823 wells, were used for these purposes. The use of ground water for public supply in areas where small-diameter domestic wells once served individual households is the primary factor for the changing trend in ground-water usage during 1969-82.

Records of withdrawals for public supply, institutions, industry, air conditioning, and irrigation are compiled annually by the Utah Department of Natural Resources, Division of Water Rights, and the U.S. Geological Survey, but no records are kept for withdrawals from domestic and stock wells. During 1982-83, estimates of withdrawal for domestic and stock use were made from field inventories of the number of wells in selected areas that were previously inventoried by Marine and Price (1964) and from consideration of the refinements made by Hely and others (1971). Marine and Price (1964, p. 49), using a 1957 inventory of 12 areas in Salt Lake Valley, estimated the discharge of small diameter wells to be 35,000 acre-feet per year. Hely and others (1971, p. 140 and fig. 66) refined the earlier estimates of Marine and Price and estimated 32,000 acre-feet in 1957 and 30,000 in 1968. Inventory of 4[1] of the 12 areas inventoried by Marine and Price (1964, fig. 26) indicated that there were about 15 percent fewer wells in use in the sampled areas in 1983 than in 1968. Thus, estimated withdrawal from domestic and stock wells during 1982-83 was 25,500 acre-feet per year and during 1969-82 averaged 27,500 acre-feet per year.

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[1]The areas inventoried were sections 21 and 34 of Township 1 South, Range 2 East, and Sections 12 and 22 of Township 1 South, Range 1 East.

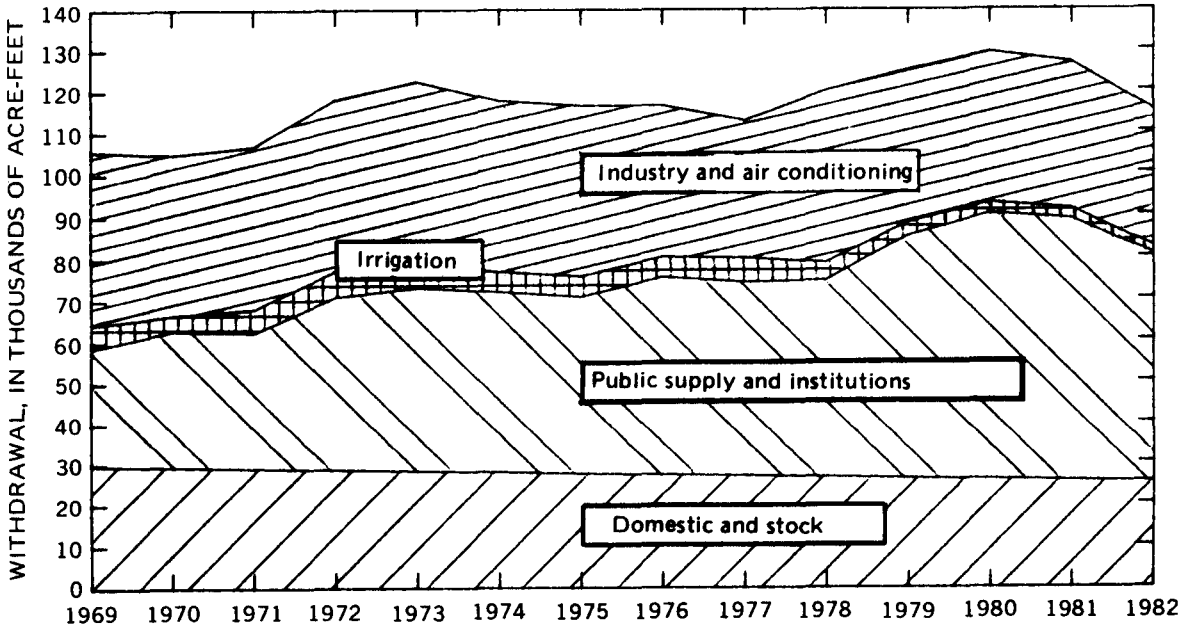


Figure 12.—Annual withdrawal of ground water by wells, 1969-82.

For the digital model, the quantity of discharge from the principal aquifer from all wells in a given model node for a specific year was combined and simulated for each year during 1969-82 (fig. 13). No changes were made in the estimated well discharge during calibration of the model.

Inflow to Jordan River and Tributaries

The ground-water inflow to the Jordan River between the Jordan Narrows and 2100 South Street, including the downstream reaches of Little Cottonwood, Big Cottonwood, and Mill Creeks, was estimated by procedures similar to those described by Hely and others (1971, p. 84). They computed the gross monthly gains for November to March, and averaged the two smallest monthly gains in each year to estimate ground-water inflow. Errors in the computation of ground-water inflow can be increased by irrigation return flows, evapotranspiration, and runoff from local storms and snowmelt. During November to March, the overall errors are minimized and reasonable estimates of ground-water inflow can be made.

For this study, the gain for each month from November to February was used to compute a maximum and minimum measured gain for each year during 1969-82 (fig. 14). During 1969-82, the average annual ground-water inflow to the Jordan River and the three tributaries was estimated to range from 85,000 to 195,000 acre-feet and to average 155,000 acre-feet.

Table 4.—Classification of wells in Salt Lake Valley
[Based on records of the Utah Department of Natural Resources, Division of Water Rights]

	Number of wells		
	Completed before 1970	Completed during 1970-81	Total as of 1981
Use:			
Domestic (D)	934	196	1,130
Stock (S)	1,354	13	1,367
Irrigation (I)	692	63	755
Combined (D), (S), and (I)	5,831	295	6,126
Industry	285	34	319
Institutions	42	1	43
Public supply	355	49	404
Unused, unknown, and plugged	2,330	43	2,373
Total reported	11,823	694	12,517
Depth (ft):			
Less than 100	2,321	22	2,343
100-200	2,973	294	3,267
201-300	2,586	207	2,793
301-400	998	63	1,061
401-500	312	28	340
501-1,000	284	71	355
More than 1,000	18	8	26
Depth unknown	2,331	1	2,332
Total reported	11,823	694	12,517
Diameter (in):			
Jetted, driven, or drilled:			
2 or less	6,298	41	6,339
2 ¼ -3	2,874	3	2,877
3 ¼ -4	836	32	868
4 ¼ -6	422	375	797
More than 6	730	242	972
Diameter unknown	590	1	591
Dug	73	0	73
Total reported	11,823	694	12,517

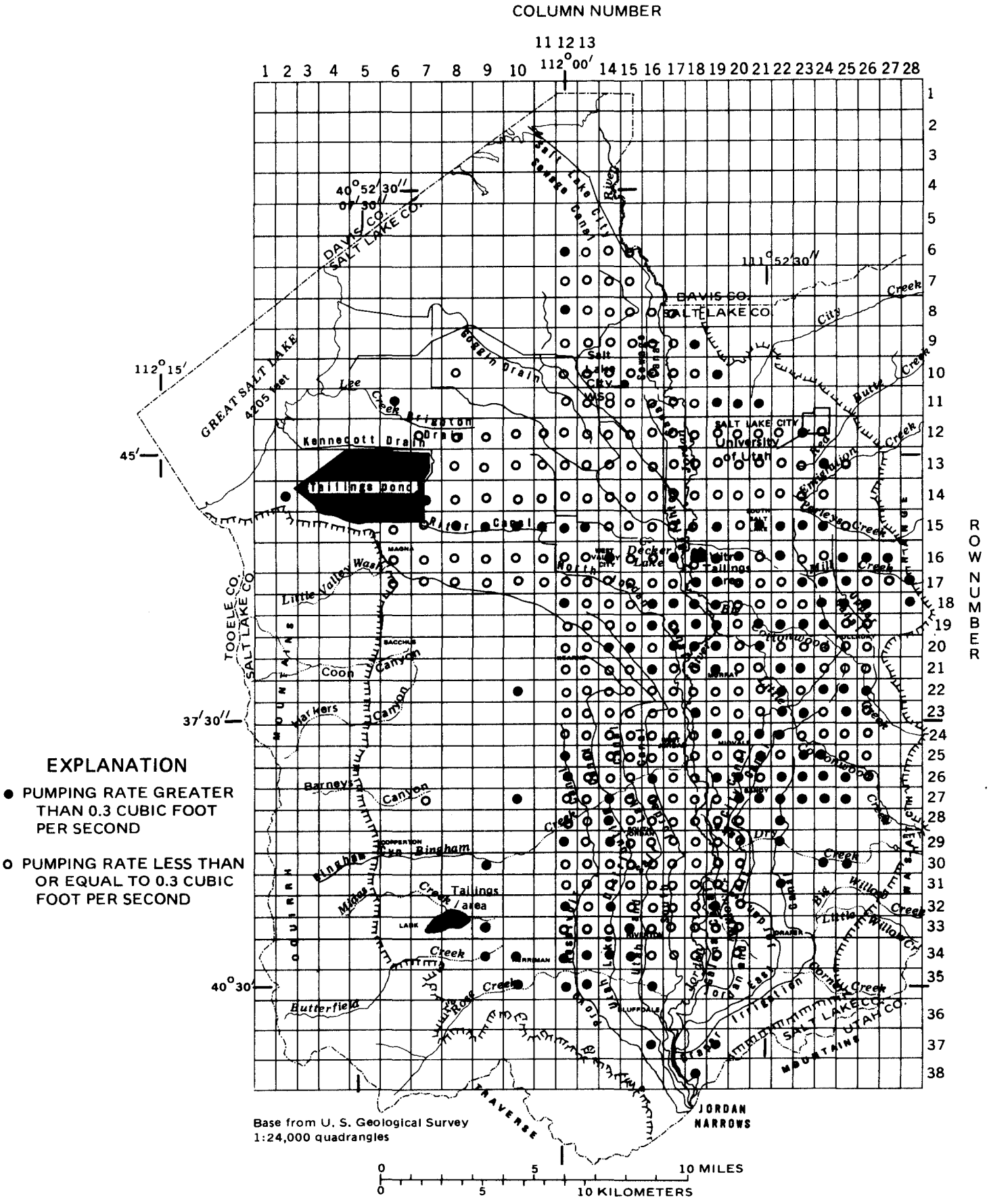


Figure 13.—Location of the centers for withdrawal of ground water used in the digital model for 1969-82. A center consists of one or more flowing or pumped wells, and it represents a point in the digital model where withdrawal was simulated.

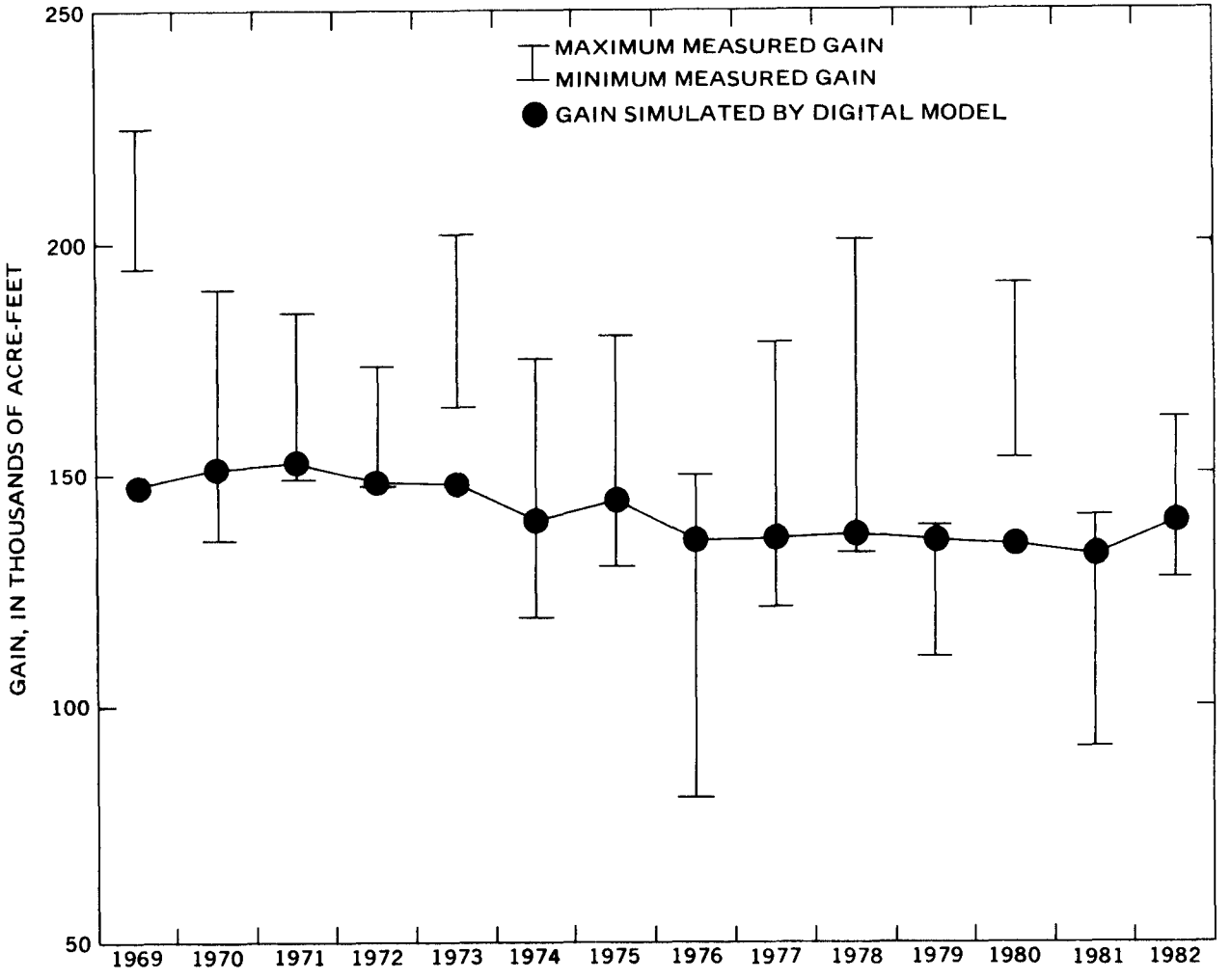


Figure 14.—Measured and simulated gain from ground-water inflow to the Jordan River between the Jordan Narrows and 2100 South Street, 1969-82. The measured gain was estimated from the four smallest monthly gains in each year which was extrapolated through the remainder of the year.

Inflow to the Jordan River between the Jordan Narrows and 2100 South Street includes effluent from eight sewage plants in addition to the discharge in Little Cottonwood, Big Cottonwood, and Mill Creeks. Unmeasured inflow to the river in the same reach includes runoff from local storms and snowmelt and return flows from irrigation. Outflow in the reach includes nine major diversions upstream from 9400 South Street plus the Brighton Canal downstream from 9400 South Street. Thus, the estimated ground-water gain was influenced by the error of all records involved.

The average annual ground-water inflow to the Jordan River between the Jordan Narrows and 2100 South Street as computed by the digital model for 1969-82 was 143,000 acre-feet, or 12,000 acre-feet less than that estimated from streamflow records. Figure 14, however, shows that the agreement between the discharge computed by the model and the discharge determined from streamflow records is poor during some years. Because of the large number of factors involved in computing the ground-water inflow from streamflow records the error associated with the computation could be large.

Additional confirmation of the amount of gain to the river as predicted by the model was made by comparing the gain to the response of observed water levels in a well near the river. The amount of upward leakage from the principal aquifer to the river is directly related to the pressure in the aquifer. Thus, the gains in river discharge should respond in the same way as water levels in the principal aquifer near the river. Figure 15 compares the river discharge as computed by the model imposed upon a plot of water-level changes in well (C-3-1)12ccb-1, which is about 100 feet from the river. The generally downward trend of water levels during 1969-82 is similar to the trend in computed gains to the river.

Inflow to Major Canals

Herbert and others (1984) reported ground-water seepage into several major canals during 1982-83. The seepage gains were considered to be discharge from the shallow-unconfined aquifer into the canals, and during 1983 the seepage was estimated to be 13,000 acre-feet. Most of these gains were in the Draper Irrigation, East Jordan, and Jordan and Salt Lake City Canals in the southeast part of the valley. The shallow-unconfined water table is less than 10 feet below land surface near most of the gaining reaches of the canals, and because 1983 was a wet year, it was concluded that recharge to the shallow-unconfined aquifer was greater than normal and the gains were not representative of years with near normal precipitation. Thus, during calibration of the digital model, the seepage to the canals was reduced to 10,000 acre-feet per year. This was the maximum total discharge that could be obtained from the shallow-unconfined aquifer in the area of the gaining canal reaches without causing the shallow aquifer to become dry during simulations of the model. A constant annual value of 10,000 acre-feet per year was used for 1969-82.

Inflow to Drains, Spring Flow Diverted for Use, and Inflow from Thermal Springs

Discharge to drains near Garfield and Magna, spring flow diverted for public supply and industrial purposes, and discharge of thermal springs were described by Hely and others (1971, p. 135-136). Their estimates of total discharge are shown in column 2 of table 3 to be 5,000, 19,000, and 2,000 acre-feet per year, respectively. No new data were collected during this study, and no revisions were made during the calibration of the digital model. The discharges were assumed constant during 1969-82.

Evapotranspiration

Hely and others (1971, p. 135, 179-188) estimated that the annual discharge of ground water by evapotranspiration (ET) was about 60,000 acre-

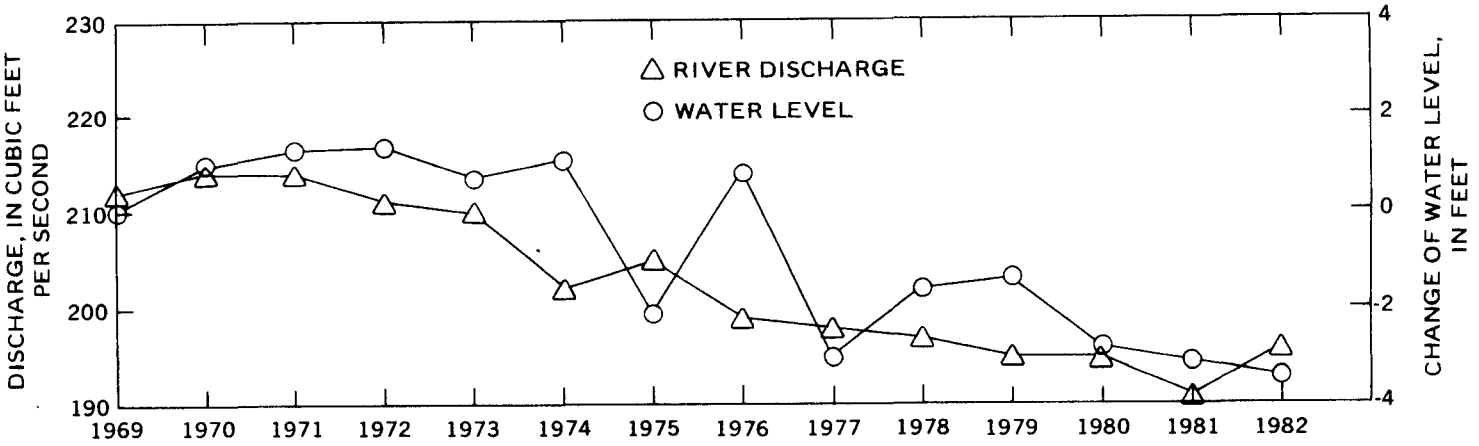


Figure 15.—Observed change of water level in well (C-3-1)12ccb-1, about 100 feet from the Jordan River, and computed gain in river discharge, 1969-82.

feet. They indicated that although most of this discharge was from the shallow-unconfined aquifer, most of the water was replaced by water moving upward from the confined aquifer. Hely and others (1971, fig. 79) prepared a map showing five major categories of land use in the Salt Lake Valley, one of which was phreatophytes. They report (p. 186) that about 43,000 acre-feet per year, or 70 percent of the ET was from the phreatophyte areas. The other 17,000 acre-feet of ET was from waterfowl-management, urbanized, cultivated, and undeveloped areas.

ET was computed by Hely and others (1971, p. 184) by means of the Blaney-Criddle formula (Blaney and Criddle, 1962), which requires that the vegetative type and density, the total area of coverage, and the depth to the water table be known. During the present study, the only new data obtained pertained to depth to the water table (Seiler and Waddell, 1984). The average depth to the water table was found to be about a foot less in 1983 than during 1964-68.

The volume of ET for five areas of land use presented by Hely and others (1971, table 33) were used to compute rates of ET. The rates of ET vary from 1.75 feet per year in the south part of the valley along the Jordan River to 0.5 feet per year in the north part of the valley near Great Salt Lake. The variation of rates are largely due to differences in land use and depths to water. Hely and others (1971, fig. 79) presented a map showing land use and Seiler and Waddell (1984, plate 2) determined the average depth to water for the shallow-unconfined aquifer. The areas of land use and associated rates of ET were duplicated as closely as possible to simulate ET in the digital model.

The model, as calibrated, computes the rate of ET as a linear interpolation between a specified maximum rate at the land surface and a value of zero at 30 feet below the land surface. The maximum rate of ET usually occurs when the water surface is nearest the land surface. A large percentage of the ET computed by Hely and others was from areas where the depth of water was less than 5 feet below the land surface and where a maximum rate of ET would be expected. Also, considering that the water levels as computed by the model for the shallow-unconfined aquifer are only accurate within ± 3 feet, it was assumed that the maximum rate of ET was equal to the rates of ET computed from the data of Hely and others (1971, table 33).

For the steady-state calibration for 1968, as well as the transient calibration for 1969-82, the ET computed by the model was 54,000 acre-feet, or 10 percent less than the average computed by Hely and others for 1964-68. During calibration of the model, attempts to increase ET to the amount of 60,000 acre-feet that was determined by Hely and others (1971, p. 135, 179-188) resulted in a poorer match between observed and computed water levels in both the shallow-unconfined and principal aquifers.

Subsurface Outflow to Great Salt Lake

Ground water in the Salt Lake Valley moves northwest toward Great Salt Lake, and most of the discharge to the lake is by upward leakage resulting from artesian pressure. Mower (1968, p. D71-D74) computed part of the subsurface flow across the north end of the valley (A-A' in fig. 8 of this report) to be 8,000 acre-feet per year. Hely and others (1971, p. 136-137) then subtracted the quantity of water discharged by evapotranspiration and wells between line A-A' and the shoreline of the lake, leaving 3,300 acre-feet per year of outflow to the lake. A shoreline at 4,205 feet above sea level was used for the calculation. Using similar procedures, Hely and others (1971, p. 137) computed that the additional subsurface outflow to the lake across a narrow strip of the valley north of the Oquirrh Mountains was about 750 acre-feet per year (B-B' in fig. 8 of this report). Thus, the total outflow to Great Salt Lake was about 4,000 acre-feet per year.

The same procedures were used for this study except for the computation of flow across line B-B'. The hydraulic gradient across line A-A' was revised based on water-level data for February-March 1983, and the total discharge across line A-A' was computed to be 6,000 acre-feet per year. Subtracting estimates of discharge by evapotranspiration and wells (3,500 acre-feet), the subsurface outflow to Great Salt Lake was computed to be 2,500 acre-feet per year. Assuming that the discharge across line B-B' decreased in proportion to the decrease across line A-A', the discharge across line B-B' was estimated to be about 600 acre-feet per year. Thus, the total subsurface outflow to Great Salt Lake during February-March 1983 was estimated to be 3,100 acre-feet per year. Subsurface outflow computed by the digital model by the steady-state calibration for 1968 was 7,200 acre-feet per year and the average annual subsurface outflow computed for 1969-82 was 2,600 acre-feet per year, or 500 acre-feet per year less than estimated from the observed data.

Hydraulic Properties

During 1983, new data pertaining to the hydraulic properties of the shallow-unconfined aquifer, of the confining layers, and of the principal aquifer were obtained from aquifer tests near the Vitro tailings area (fig. 8). Also, during calibration of the digital model, revisions were made to estimates of the hydraulic properties in various parts of the valley.

The hydraulic conductivity of the shallow-unconfined aquifer was determined to be 20 feet per day from an aquifer test at well (C-1-1)26dba-4, which is in an area for which Hely and others (1971, fig. 61) estimated the transmissivity to be 2,700 square feet per day. Thus, using an estimated thickness of 50 feet for the shallow-unconfined aquifer would give a hydraulic conductivity of about 55 feet per day.

Data from an aquifer test at well (C-1-1)26dba-5 and 6 nearby observation wells finished within and below the confining layers were used to determine hydraulic properties of the principal aquifer in the Vitro tailings area. Methods developed to analyze leaky artesian systems with the release of water from storage in the confining beds include the "Hantush modified method" described by Lohman (1972, p. 32), and the "Ratio method" described by Neuman and Witherspoon (1972, p. 1284). The Hantush modified method was used to obtain values of transmissivity and storage coefficient for the deep artesian aquifer and both methods were used to obtain an estimate of the vertical hydraulic conductivity and specific storage of the confining bed.

The vertical hydraulic conductivity of the confining layer determined from the test at well (C-1-1)26dba-5 was 0.124 foot per day. Hely and others (1971, p. 118), using a form of Darcy's equation, estimated values for vertical hydraulic conductivity of 0.016 and 0.049 foot per day for two areas in the valley. Thus, the test value of 0.124 foot per day was about 10 times larger than the smallest value reported by Hely and others.

The transmissivity of the principal aquifer in the Vitro tailings area was determined to be about 3,860 square feet per day. Hely and others (1971, fig. 59), however, estimated that the transmissivity in the Vitro tailings area was about 10,000 square feet per day. Storage coefficient for the principal aquifer was determined from the pumping-test data to be 4×10^{-4}, which is almost the same as reported by Hely and others (1971, p. 115).

Hydraulic properties for layers 1 and 2 were provided for each active cell within the grid. Also, a value to allow leakage between layers 1 and 2 was provided everywhere that layer 1 (representing the shallow-unconfined aquifer) occurs. Except in areas where new data had been collected, the initial values used in the model were taken from Hely and others (1971, p. 111-118). Revisions were made to some of the values on a node-by-node basis during calibration, so that agreement between observed and computed water levels and between measured and computed flow rates could be attained. During the calibration, revision to values of hydraulic properties were made only to the extent that the values remained physically reasonable.

Numerous revisions were made to the hydraulic conductivity of the shallow-unconfined aquifer during the calibration. After the calibration, the values ranged from 0.000011 to 0.001 foot per second (fig. 16). Assuming an average thickness of about 50 feet for the shallow-unconfined aquifer, the transmissivities would range from about 50 to 4,000 square feet per day. Hely and others (1971, fig. 61) reported a range from 1,300 to 4,000 square feet per day.

The vertical hydraulic conductance of the confining bed, which is the vertical hydraulic conductivity divided by the thickness of the confining bed (McDonald and Harbaugh, 1984, p. 144) was represented in the digital model. The initial values were computed by dividing the vertical conductivities computed by Hely and others (1971, p. 118) and the test value in the Vitro tailings area by an average thickness of 50 feet. During steady-state calibration, larger values were used in areas of known or suspected upward leakage from the confined zones, such as springs and swampy areas. Such areas occur along the Jordan River and the downstream reaches of tributaries. It is not known if the vertical hydraulic conductivity is actually higher in these areas or if the confining bed is considerably thinner—either property could have the same effect on the leakage rates. Changes were also made to improve the comparison between simulated and observed differences between the potentiometric surfaces in the shallow-unconfined and principal aquifers. The final values used in the digital model are shown in figure 17.

During calibration, most revisions of transmissivity of the principal aquifer were made for the south part of the valley where water levels are affected by seepage from canals and from irrigated lands. Some of the aquifer tests made by Hely and others (1971) in that part of the valley may have been affected by recharge from surface seepage, which could have resulted in an unrealistically large value for transmissivity. Some of the transmissivity values in this area were reduced by as much as a factor of 10. For the remainder of the valley, however, the values used for the digital model (fig. 18) were similar to those reported by Hely and others (1971, fig. 59).

Predicted Effects of Increased Withdrawals

The digital-computer model can be used to predict the effects of changes in ground-water withdrawals on water levels and the water budget. (See section on "Digital Model of Ground-Water Reservoir" for details of model calibration.) The effects of such changes vary depending upon the placement of new wells and the location of existing wells at which pumpage is increased. So, for simulations made in this study, it was assumed that the increased withdrawals would be where existing well discharges exceed 0.3 cubic foot per second (fig. 13).

A simulation was made with the digital model by maintaining the 1982 rate of withdrawals from wells (115,000 acre-feet) constant until 2020; another simulation was made by doubling the pumpage from all wells that had a discharge greater than 0.3 cubic foot per second during 1982 (fig. 13) and holding that constant until 2020. The second simulation had the effect of increasing the 1982 withdrawals by 65,000 acre-feet for a total rate of 180,000 acre-feet per year. The average recharge rate of 352,000 acre-feet per year computed for 1969-82 was used for both simulations.

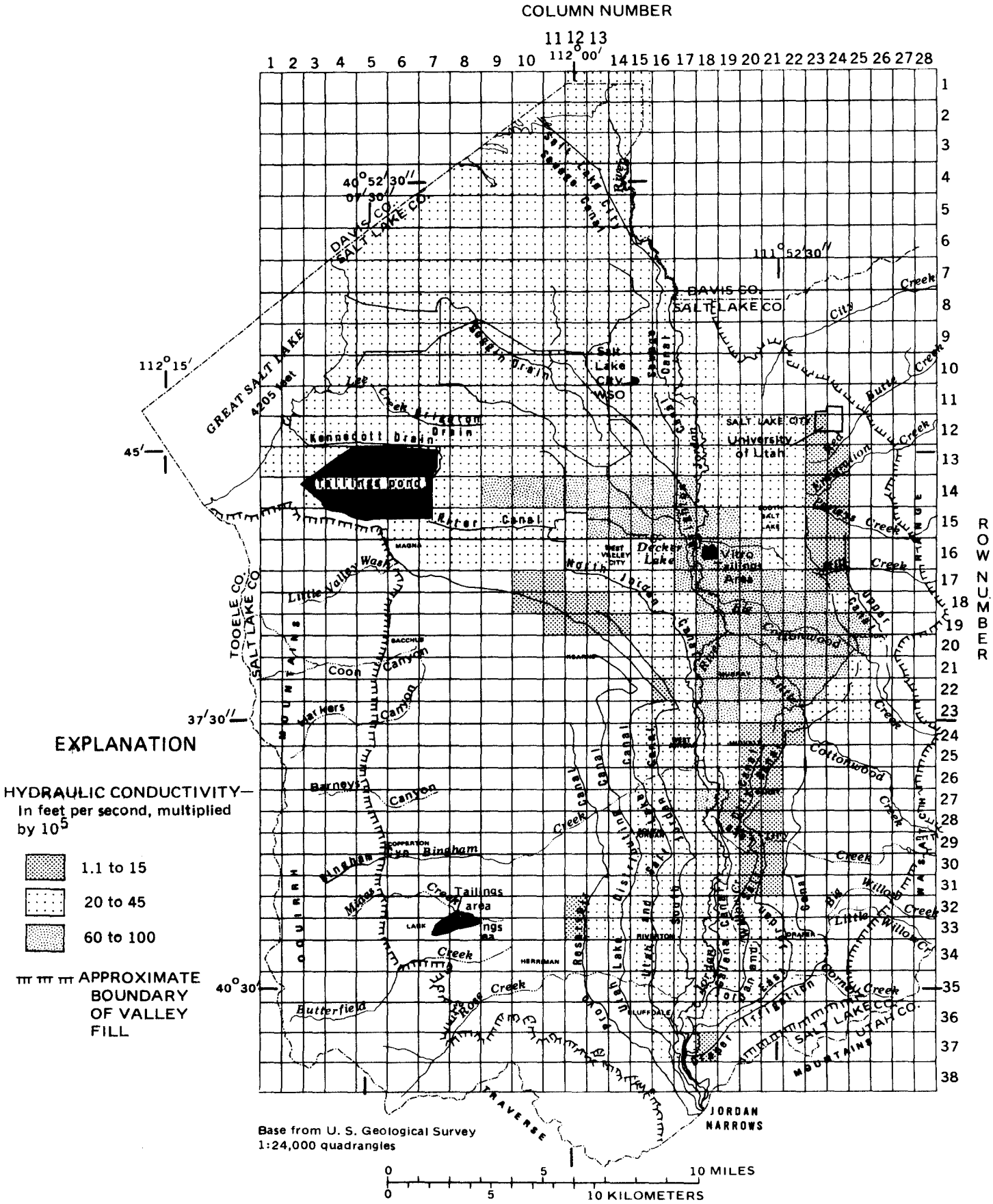


Figure 16.—Hydraulic conductivities used for the shallow-unconfined aquifer in the digital model for Salt Lake Valley.

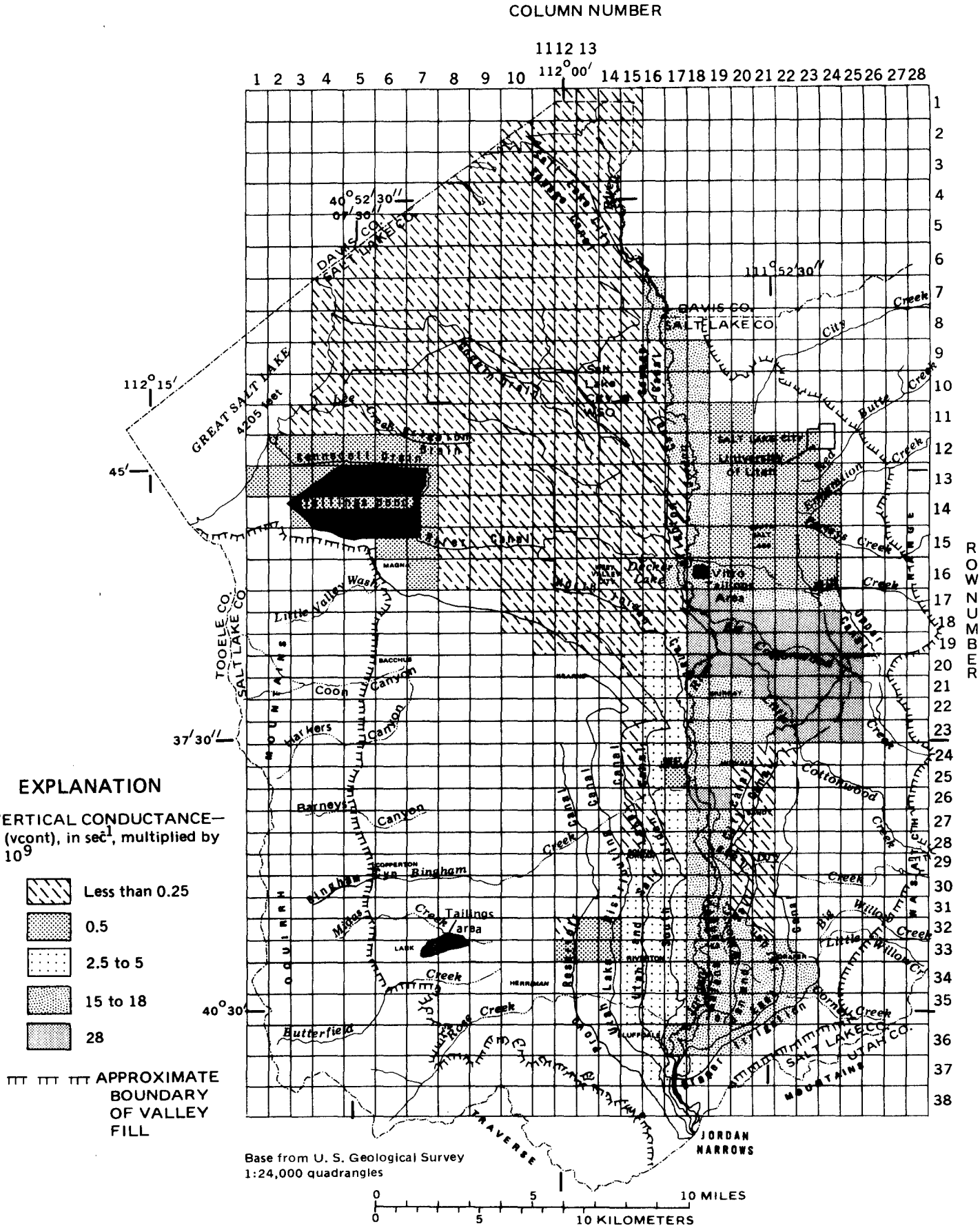


Figure 17.—Vertical conductances used for the confining layer in the digital model for Salt Lake Valley.

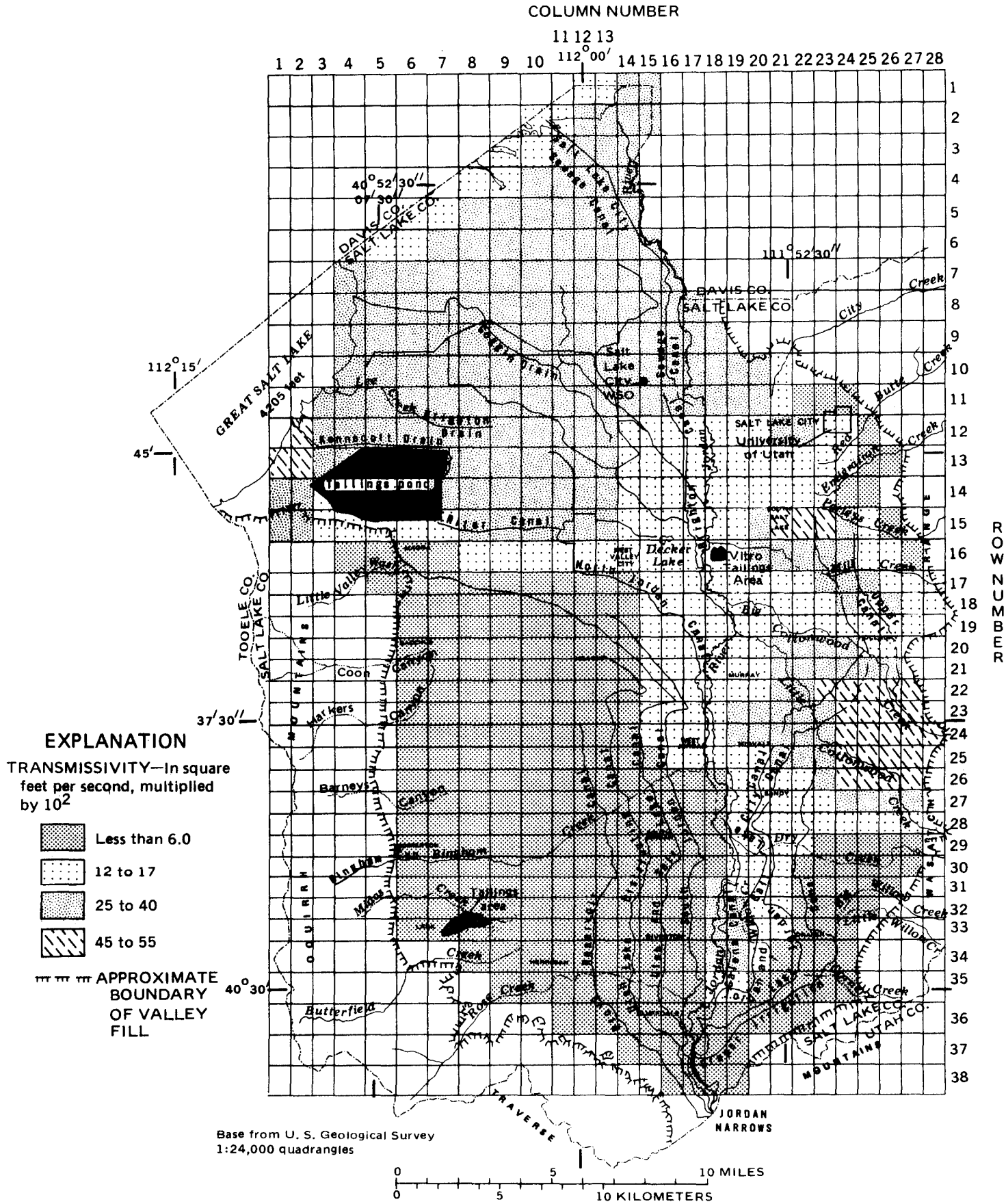


Figure 18.—Transmissivities used for the principal aquifer in the digital model for Salt Lake Valley.

The projected changes in water level resulting from the two simulations are shown in figures 19 and 20. The maximum drawdown of water levels for the first simulation was 18 feet southeast of Salt Lake City for both years 2000 and 2020. The drawdowns for the second simulation were larger, however, reaching 40-60 feet in the area east of Sandy, where a large portion of the increase of withdrawals was simulated, and exceeding 80 feet in a localized area east of Copperton. The reasons for the 10-feet rises in the southwestern part of the valley are not known.

The increase of withdrawal of ground water in the second simulation involves water that otherwise would have been discharged to streams, by evapotranspiration, or by subsurface outflow to the Great Salt Lake or would have remained in storage in the shallow-unconfined or principal aquifers. In the second simulation, the discharge to the Jordan River and tributaries in the year 2020 was 49,000 acre-feet less than in the first simulation (table 5). Thus, about 75 percent of the increased withdrawal of ground water was simulated as salvaged from water that otherwise would have been discharged to the river. An additional 5 percent of the increased withdrawal (3,000 acre-feet) was simulated as salvaged from evapotranspiration. The change in subsurface outflow to Great Salt Lake was negligible. Therefore, the remaining 20 percent of the increased withdrawal represents depletion of storage in the shallow-unconfined and principal aquifers.

DIGITAL MODEL OF GROUND-WATER RESERVOIR

A digital-computer model was calibrated to simulate, in three-dimensions, the ground-water flow in the principal and the shallow-unconfined aquifers in Salt Lake Valley. The model was used to predict water-level and water-budget changes that would be caused by simulated well discharges.

Type of Model

The modular-finite-difference model developed by McDonald and Harbaugh (1984) was selected to simulate the ground-water-flow system because it is well documented and has the flexibility to adapt to a wide variety of ground-water systems. The modular structure consists of a main program and a series of independent subroutines, which are grouped into packages. Each package deals with a specific feature of the hydrologic system which is being simulated. This permits the user to modify or examine specific hydrologic features without affecting other modules or parts of the system.

Model Construction

Construction of the model began by establishing a model grid, boundary conditions, interval of time or stress period, calibration period, and data base. The calibration period selected for steady-state conditions was 1968 and for transient conditions, 1969-82, and the stress conditions were allowed to vary annually. This required that recharge and discharge data be compiled for each year from 1969 to 1982. The initial values for hydraulic properties such as transmissivity and storage coefficient were extracted from maps given by Hely and others (1971, figs. 59 and 60) and then revised during calibration.

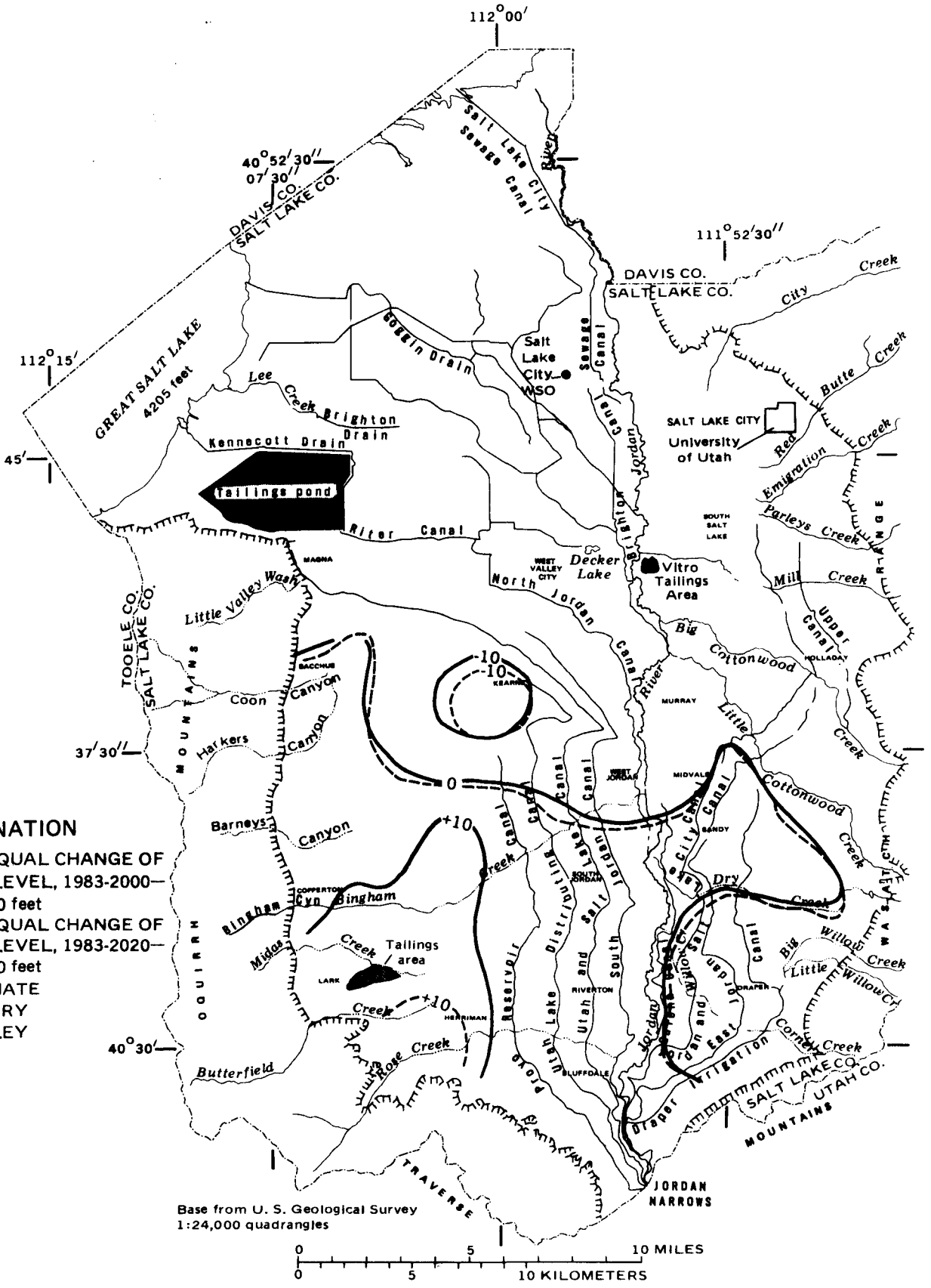


Figure 19.—Simulated change in water level for pumpage after 1982 at the 1982 rate.

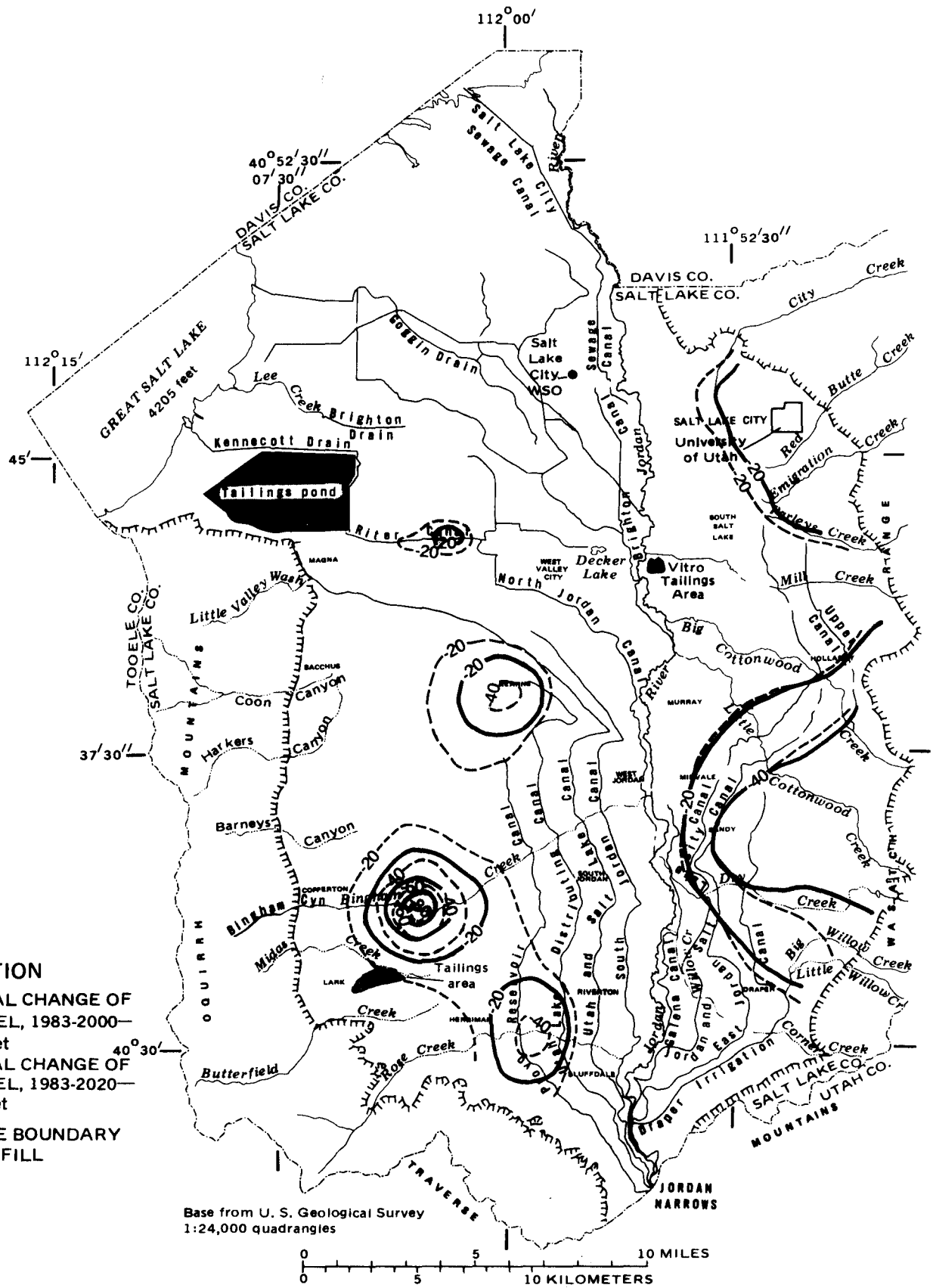


Figure 20.—Simulated change in water level for pumpage after 1982 at approximately 160 percent of the 1982 rate.

Table 5.--Effect of simulated withdrawals on ground-water discharge during 1983-2000 and 2001-2020

			1983-2000	
	Withdrawal from wells		Amount lost to streams	Amount lost to evapotranspiration
	115,000		138,000	54,000
	180,000		94,000	51,000
Increase of withdrawals	65,000			
Decrease of loss to streams and evapotranspiration caused by an increase of withdrawals		(acre-feet)	44,000	3,000
		(percent of withdrawals)	70	5

			2001-2020	
	Withdrawal from wells		Amount lost to streams	Amount lost to evapotranspiration
	115,000		138,000	54,000
	180,000		90,000	51,000
Increase of withdrawals	65,000			
Decrease of loss to streams and evapotranspiration caused by an increase of withdrawals		(acre-feet)	48,000	3,000
		(percent of withdrawals)	75	5

Model Grid

A block-centered, finite-difference grid with variable spacing was used to divide the principal aquifer into two layers of rectangular cubes called cells. The rectangular grid, which divided the study area into 38 rows and 28 columns had a grid spacing ranging from 0.7 to 1.0 miles (fig. 21). Smaller grid spacing was used in areas where there were a large number of wells, steep hydraulic gradients, or large changes in transmissivity. Figure 3 shows the relation of cells of the model to the physical ground-water system for one cell in each layer. Layer 1 was used to simulate the shallow-unconfined aquifer, which lies above the confining bed. Layer 2 was used to simulate the principal aquifer, which includes the confined aquifer and the deep unconfined aquifer, which is between the confined aquifer and the mountain block.

Boundary Conditions

Boundaries represented in the model are shown in figure 21. No-flow boundaries were placed around the modeled area as a computational convenience. Finite-flow boundaries (B in fig. 21) were specified inside the no-flow boundary in layer 2 to represent the assumed recharge from bedrock in the Wasatch Range or Traverse and Oquirrh Mountains. Constant-head boundaries (C in fig. 21) were specified along the north and northeast boundaries in layers 1 and 2 to represent the assumption that hydraulic heads are controlled by the head in Great Salt Lake. Stream boundaries (R in fig. 21) were specified along the Jordan River, lower reaches of tributaries, and the Surplus Canal to represent seepage from layers 1 and 2. The areas where evapotranspiration are specified are indicated by the stippled pattern in figure 21.

Calibration of Model

Simplifications are needed to approximate a complex three-dimensional flow system using a digital model. The errors resulting from lack of knowledge about the system and from simplification for modeling need to be evaluated by comparing model-generated values with observed values for important parameters. Water levels and independently estimated flow rates can be used to calibrate a model. The digital model was calibrated using both steady- and transient-state conditions.

Steady-State Conditions

Because of the relatively constant pumpage and small changes in storage during 1968 and preceeding years, it was assumed that recharge was about the same as discharge during 1968, and was representative of steady-state conditions. During 1968, withdrawals from wells were 105,000 acre-feet, or only 2,000 acre-feet less than the average for 1964-68. Changes in storage were less than 2,000 acre-feet in 1968 and averaged about 3,000 acre-feet during 1964-68. The overall data base for recharge, discharge, and water levels during 1968 was much superior to that of years prior to 1964, when the ground-water system may have been nearer a natural steady-state equilibrium. The steady-state calibration was useful for estimating transmissivity, vertical conductivity of the confining bed, recharge from direct seepage of precipitation, and movement through bedrock into the valley fill.

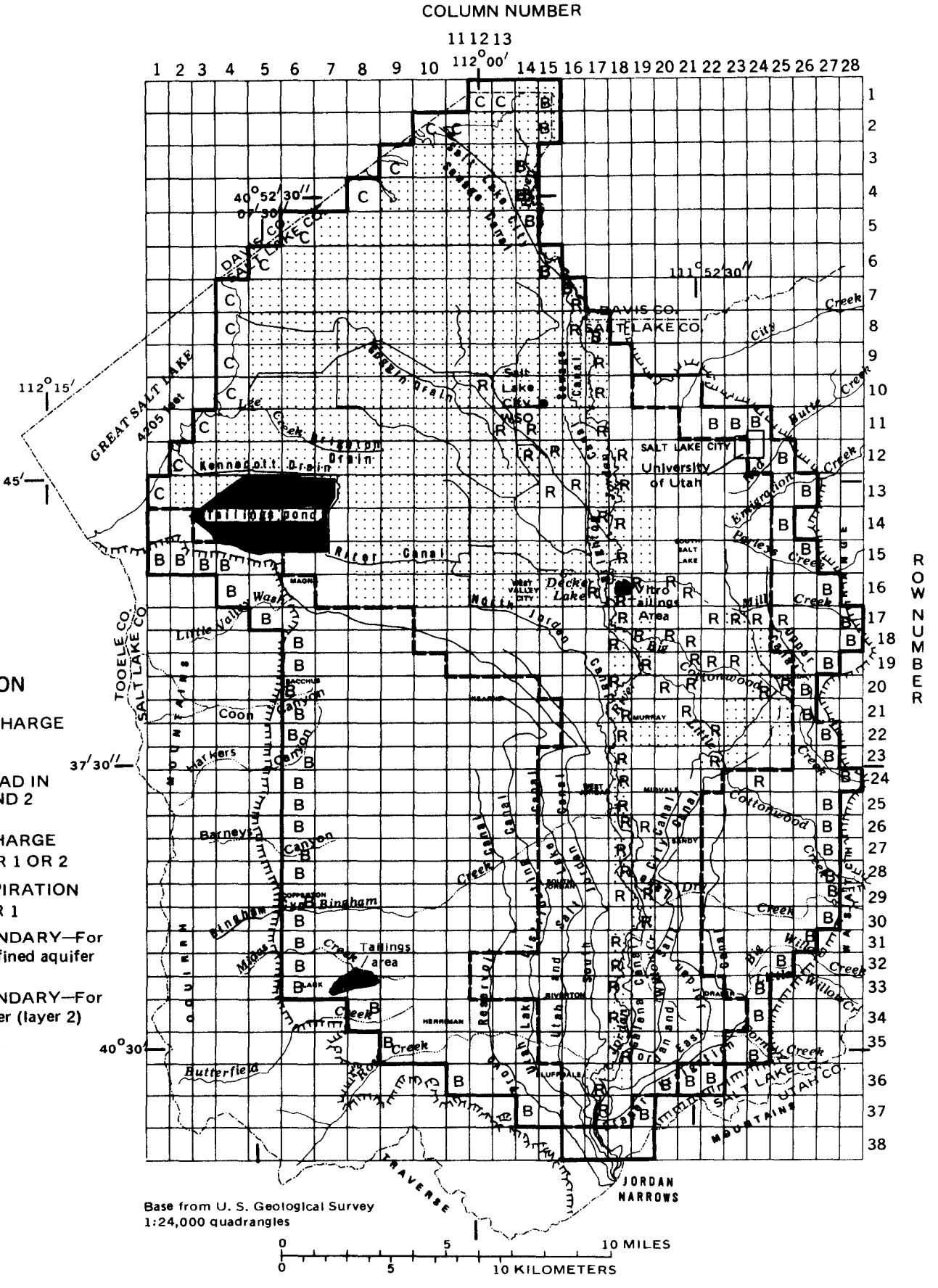


Figure 21.—Grid used in the digital model, model boundaries, and cells with bedrock recharge, evapotranspiration, or constant-head, or stream-discharge.

The steady-state calibration was begun by using recharge and discharge, as presented in columns 2 and 3 of tables 1 and 3, and hydraulic properties of the principal aquifer as given by Hely and others (1971, p. 111-118). Ground-water inflow to the Jordan River and tributaries was held constant by placing discharging wells equivalent to that inflow in layer 2. Constant-head nodes were placed just inside the model boundaries and observed heads were used in the nodes so that recharge from bedrock into the valley fill would be computed by the model during calibration.

Because of the interdependence of the variables involved in the digital flow model, calibration is an iterative process whereby one variable is revised while holding the others constant. During calibration of the hydraulic properties, all of the recharge and discharge values except for seepage through bedrock and discharge from ET were held constant.

Seepage to the Jordan River is a large portion of the ground-water discharge from the valley. Estimates of this seepage provide criteria for calibrating the model (column 3, table 3). Initial efforts in the model calibration were focused around the hydraulic parameters that controlled seepage to the river.

After suitable values for hydraulic parameters were attained along the river, revisions were made in cells throughout the study area with the objective to maintain the correct water levels and head gradients in the shallow-unconfined and confined aquifers and maintain adequate water in the shallow-unconfined aquifer.

Revision of the hydraulic parameters requires some constraints on the limits that revisions can be made to the values. Efforts were made to keep the values within at least the range of observed values in the valley and in doing this it became obvious that some of the recharge and discharge values shown in columns 2 and 3 of tables 1 and 3 would have to be revised so that observed and computed water levels would be in better agreement in some areas. Estimates of recharge from seepage, from irrigation, precipitation on the land surface, and discharge from ET were considered the least accurate of the values used as criteria for evaluating the calibration and because of this were revised to reflect better agreement among other more reliable criteria such as water levels.

Precipitation estimates were revised in conjunction with water levels observed in the shallow-unconfined aquifer (Seiler and Waddell, 1984, plate 1) and evapotranspiration as given by Hely and others (1971, table 22) and shown in column 2 of table 3. During initial runs of the model, computed water levels in the shallow-unconfined aquifer were lower than the observed values and evapotranspiration was considerably less than reported by Hely and others (1971, table 22). In order to raise water levels, increase evapotranspiration, and maintain the head gradient between layers 1 and 2, it was necessary to increase recharge from precipitation in the flat areas in the northern part of the valley.

After a satisfactory match between computed and observed water levels was attained for the principal aquifer (fig. 22) and for the shallow-unconfined aquifer (fig. 23) provisions were made to allow for variable seepage to the Jordan River and through the bedrock to the valley fill. The wells that had

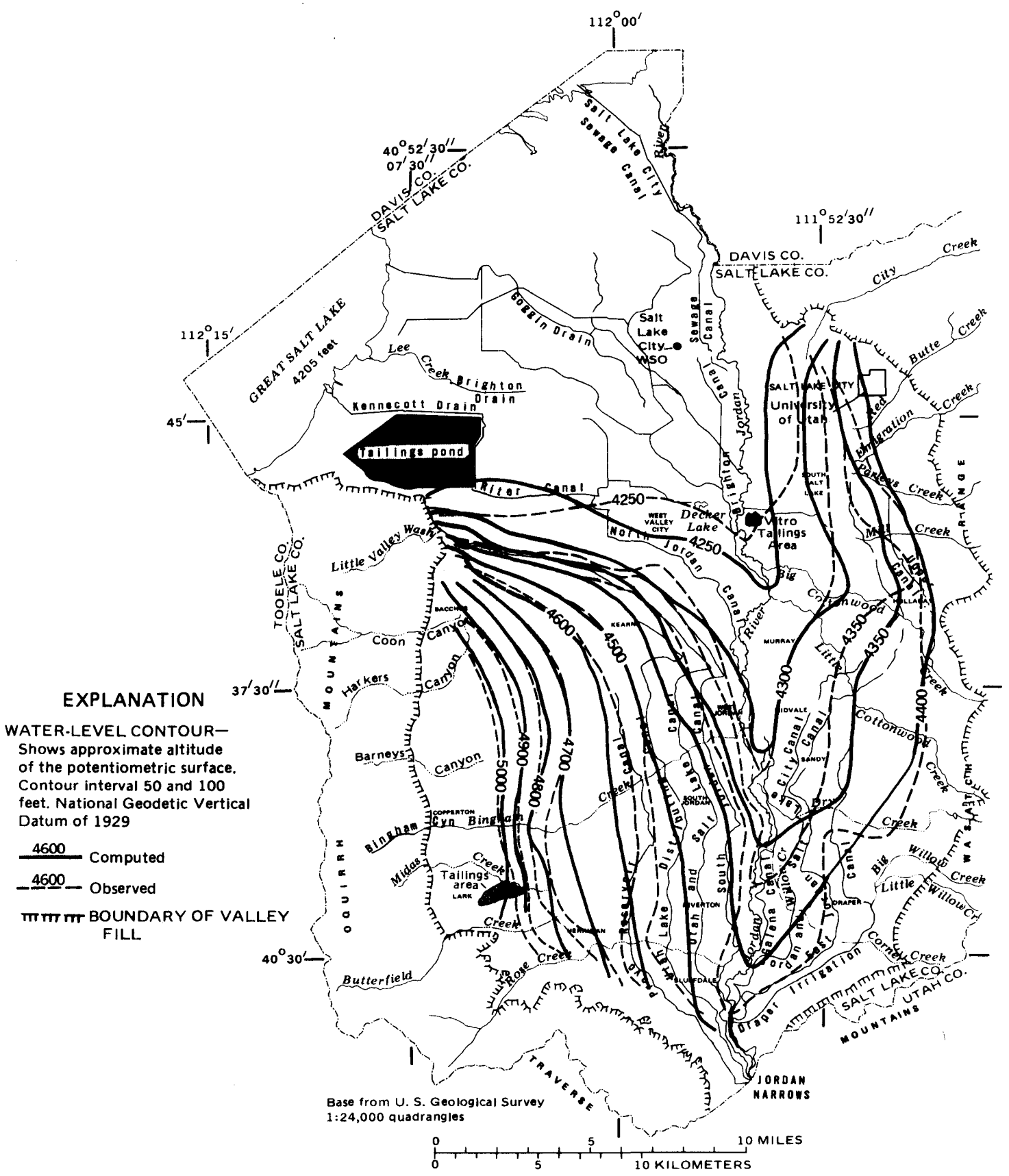


Figure 22.—Comparison of water levels in February-March 1969 to water levels computed by the digital model for the 1968 steady-state simulation for the principal aquifer.

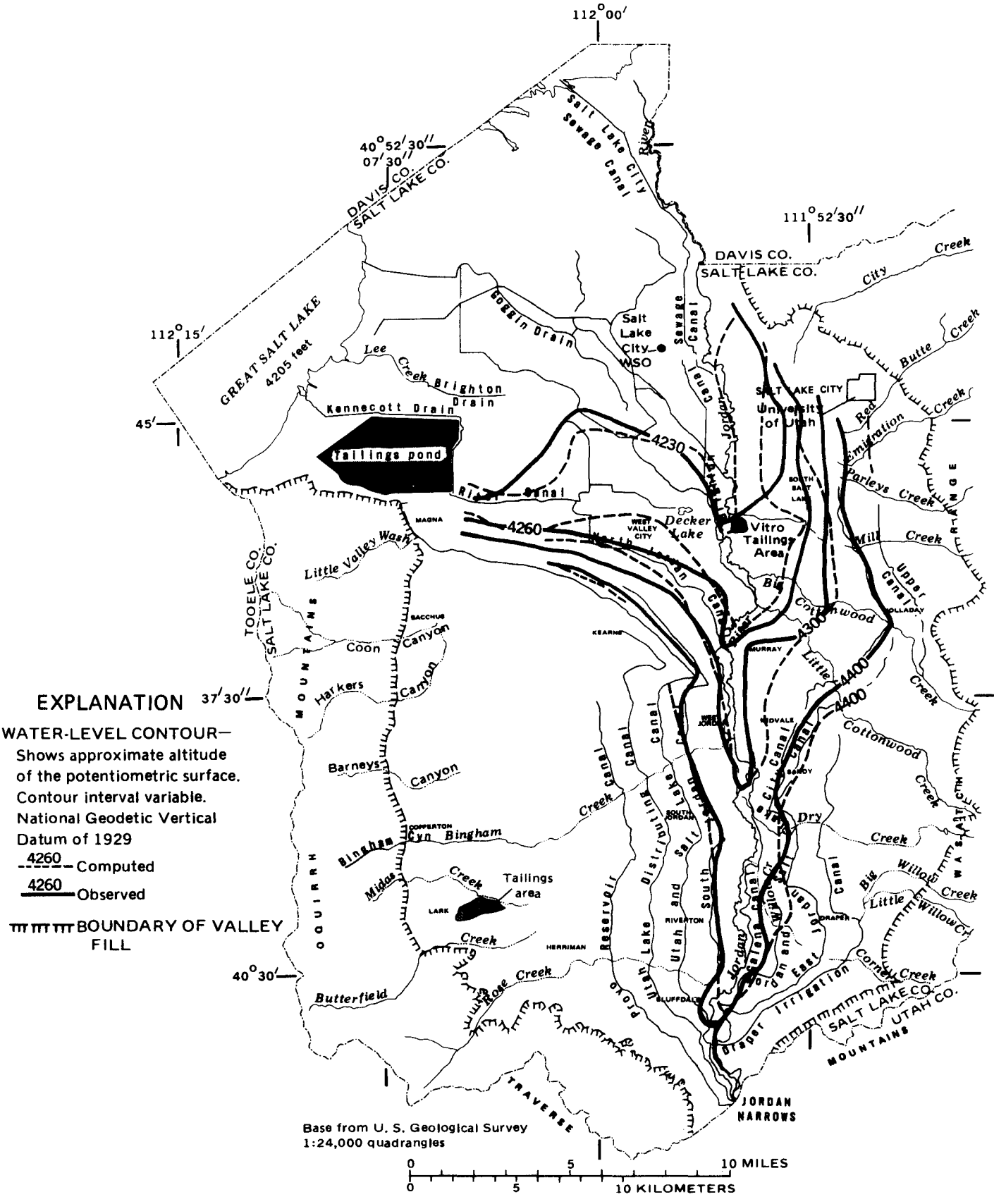


Figure 23.—Comparison of water levels in December 1982 to water levels computed by the digital model for the 1968 steady-state simulation for the shallow-unconfined aquifer.

been used to represent a constant seepage to the river were replaced with the "River Package", a model feature (McDonald and Harbaugh, 1984, p. 209) that expresses seepage as a function of a specified river-surface elevation and of a variable aquifer head. This required calibration of a conductance term for each river cell (R in fig. 21) to allow the correct amount of water to enter the river. For the steady-state calibration, the correct amount was the same as had been simulated by the wells. Verification for variable seepage to the river was deferred to the transient calibration and is discussed in the section "Transient-State Conditions". The flows through the constant head cells that had been used to simulate seepage through the bedrock were recorded and placed into the "Recharge Package" (McDonald and Harbaugh, 1984, p. 241). The "Recharge Package" was utilized because it was convenient to annually vary recharge from seepage through bedrock as well as precipitation on the land surface. The verification of the annual variations are discussed in the section "Transient-State Conditions".

The potentiometric surface computed by the model for the principal aquifer in February-March 1969 and for the shallow-unconfined aquifer in November 1983, are compared in cross sections in figure 24 with observed water levels. An offset of 1 year between model-simulation periods and water-level-measurement periods occurs because February-March water levels of a given year were used to represent water-level conditions on December 31 of the prior year. The water levels for the shallow-unconfined aquifer for November 1983 were used in place of levels for February-March 1969 because of insufficient data for 1969. Although there are no long-term records to compare water levels in 1969 with levels in 1983 it was determined that the average depth to the water table was about a foot less in 1983 than during 1964-68. The agreement between the observed and computed water levels for both aquifers was considered acceptable for purposes of the steady-state calibration. The final values of recharge and discharge for the steady-state calibration are given in column 4 of tables 1 and 3.

Transient-State Conditions

The transient-state calibration, which was representative of ground-water conditions during 1969-82, was made by simulating annual ground-water withdrawals during 1969-82 (fig. 12). The water budget, which was compiled for each year, was used as input to the model. Then the water levels that were computed for 1-year intervals were evaluated by comparing them with observed water levels for the principal aquifer.

No additional changes were made to the transmissivity of the principal aquifer or to the hydraulic conductivity of the confining bed during the transient-state calibration. However, some trial-and-error adjustments to the storage coefficient were made along the east and northeast boundaries where the principal aquifer is unconfined. The adjustments were made because the computed water levels were consistently lower than the observed water levels in these areas. The initial storage coefficients, which were taken from Hely and others (1971, fig. 60), ranged from 0.01 to 0.15 in the areas where adjustments were made. Following the adjustments, the values ranged from 0.01 to 0.10. In the confined part of the principal aquifer, trial-and-error adjustments to the storage coefficients had little effect on computed water levels.

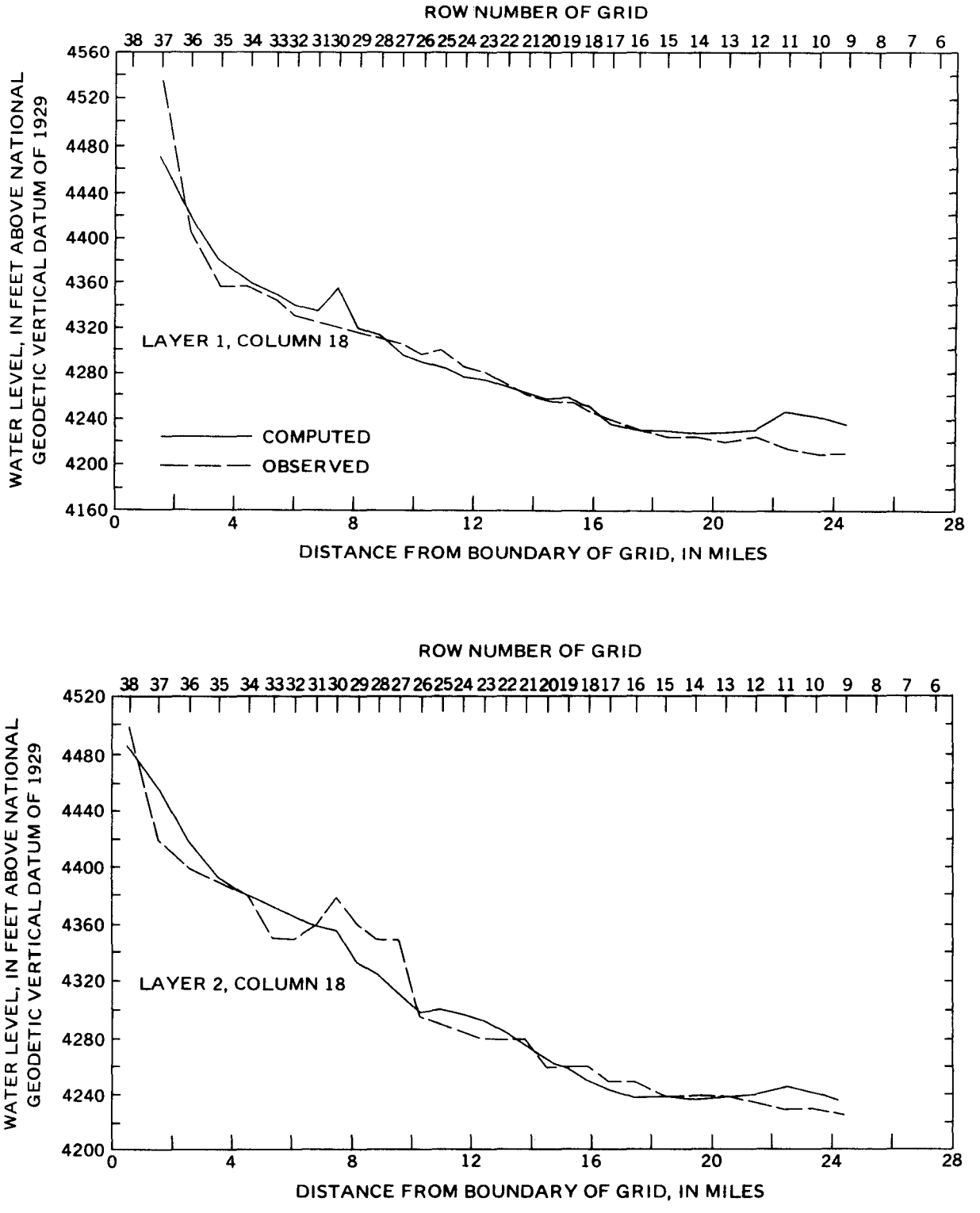
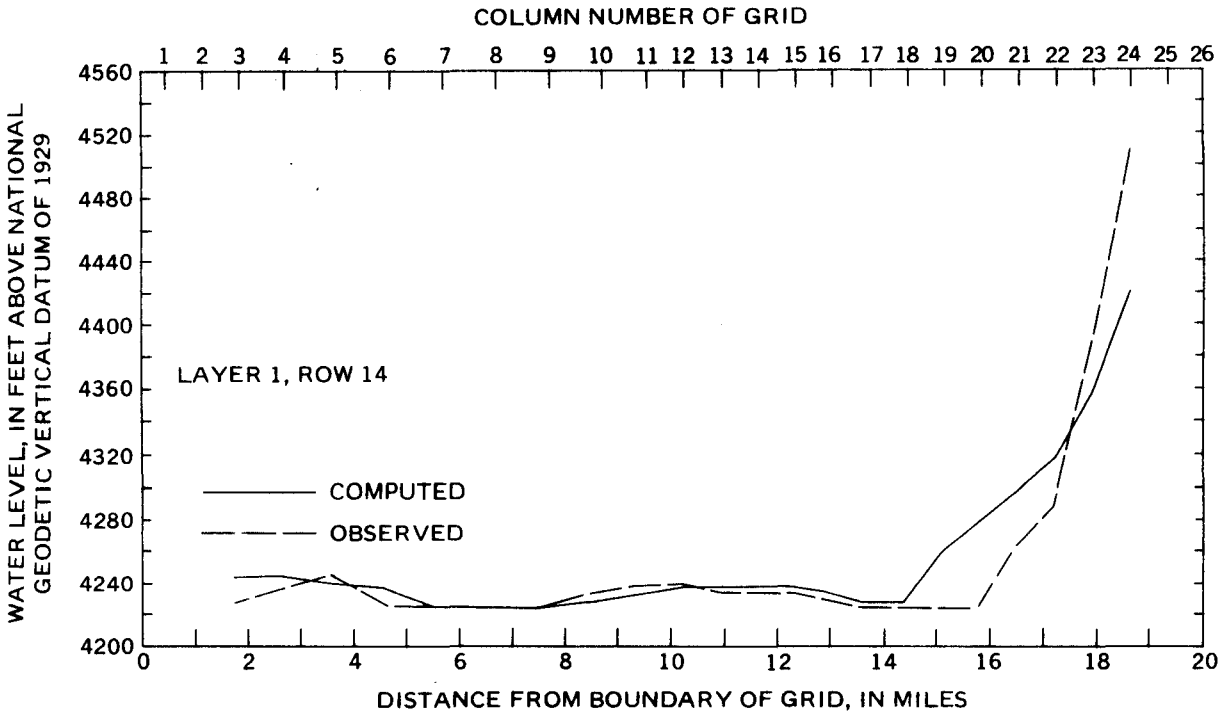


Figure 24.—Cross sections along selected rows and columns of model grid showing comparison of observed and computed water levels for steady-state calibration. (See fig. 21 for location of cross sections.)



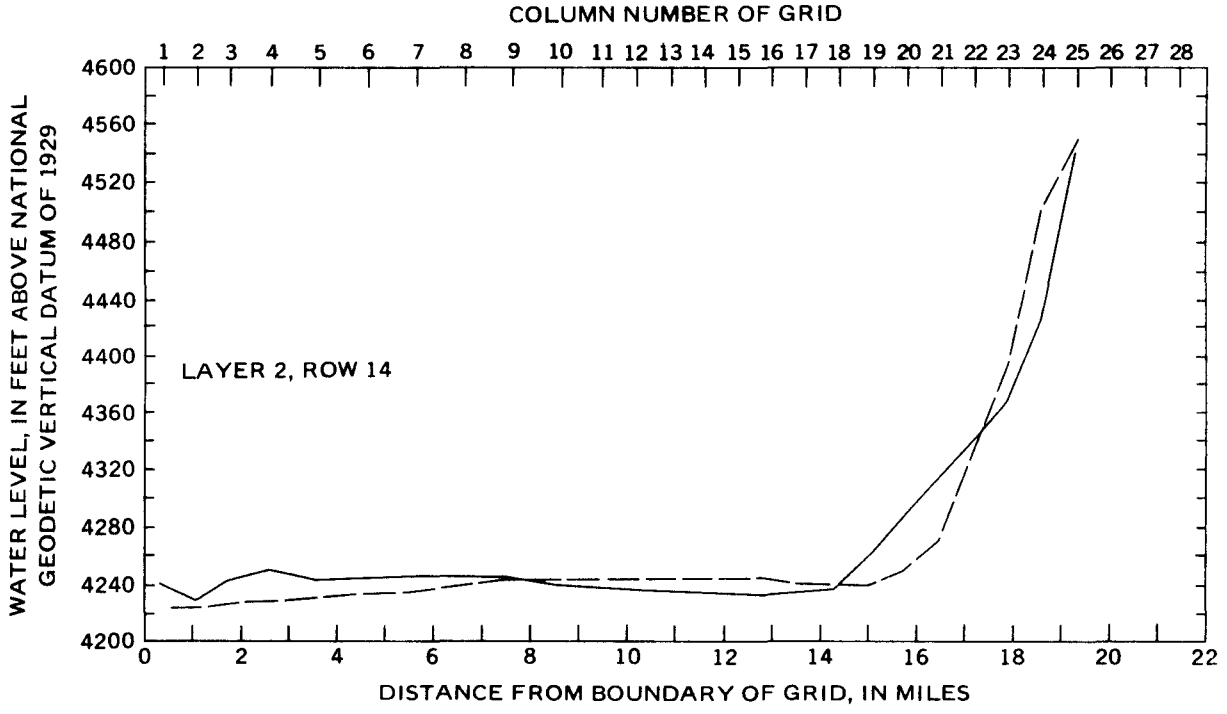
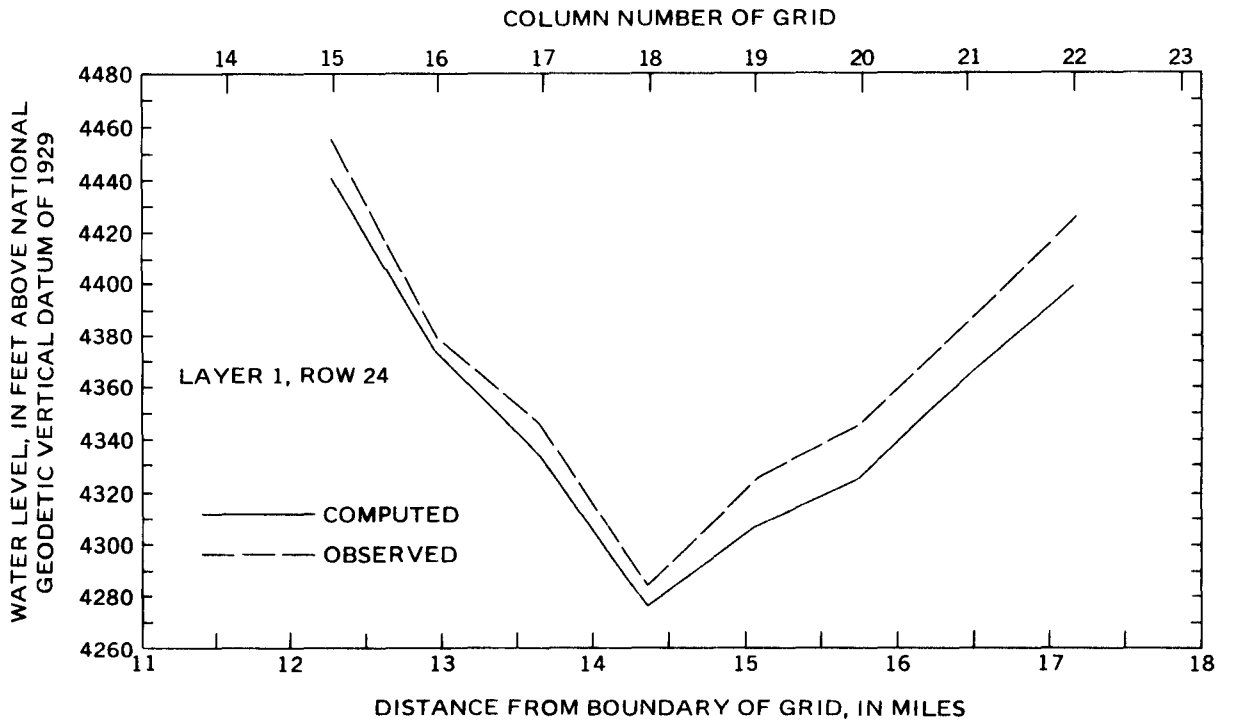


Figure 24.—Cross sections along selected rows and columns of model grid showing comparison of observed and computed water levels for steady-state calibration. (See fig. 21 for location of cross sections.)—Continued



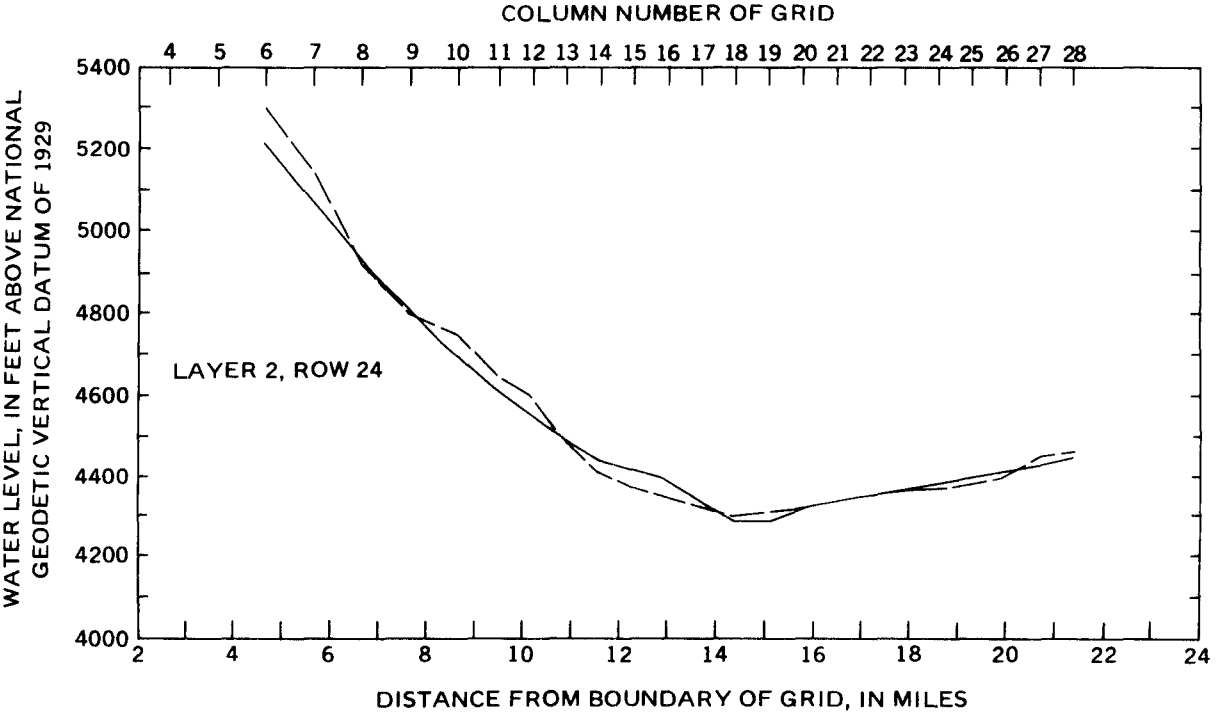
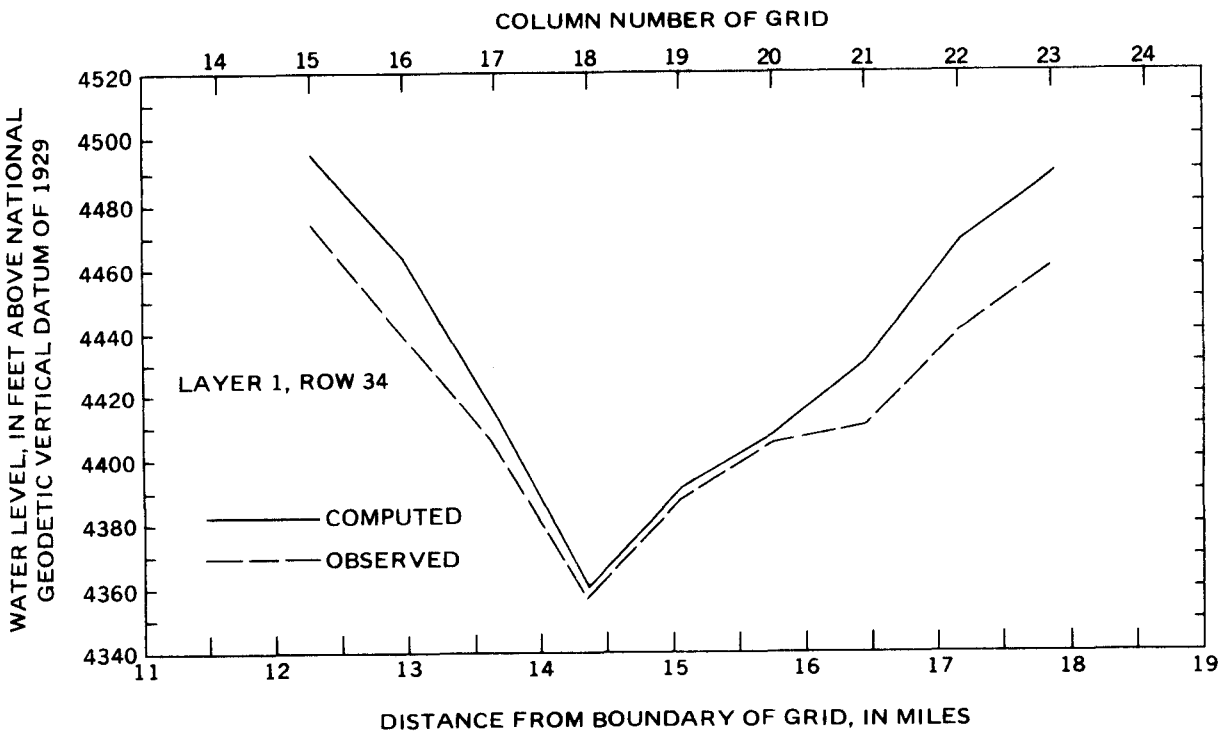


Figure 24.—Cross sections along selected rows and columns of model grid showing comparison of observed and computed water levels for steady-state calibration. (See fig. 21 for location of cross sections.)—Continued



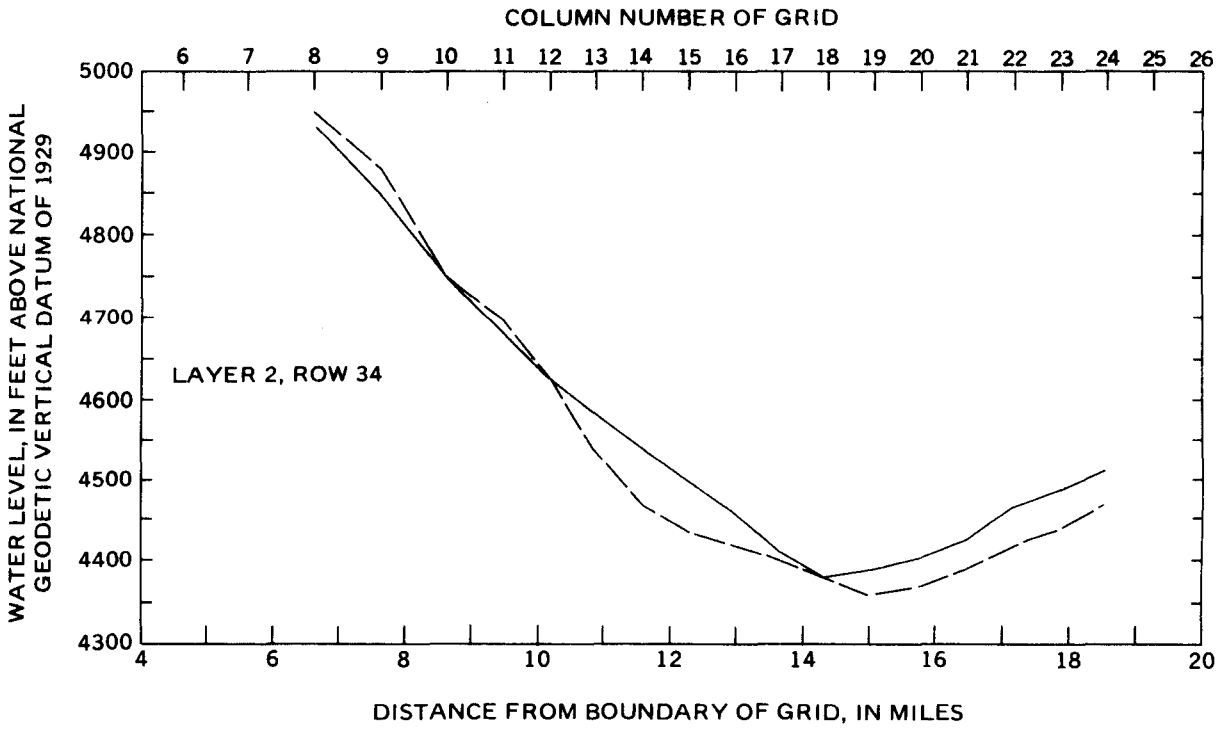


Figure 24.—Cross sections along selected rows and columns of model grid showing comparison of observed and computed water levels for steady-state calibration. (See fig. 21 for location of cross sections.)—Continued

The validity of the assumption that annual recharge from precipitation and seepage from bedrock varies directly with changes in annual precipitation was tested by comparing hydrographs prepared from observed and computed water levels in the recharge areas along the east boundary (fig. 21). The test was made by using a variable-recharge rate (see section on "Recharge") and by using a constant-recharge rate for 1969-82. The results of the test are shown in figure 25 where water levels observed in well (D-2-1)34acb-1 are compared with the water-levels computed from variable- and constant-recharge rates. The hydrograph for the constant-recharge rate was smoother and generally had smaller annual changes than did the hydrograph for the variable-recharge rate. This was particularly apparent in 1977 and 1983, after years of below- and above-average precipitation.

The generally good agreement among the three hydrographs suggests that over a long period of time the water-level trend would be about the same using either a variable- or constant-recharge rate. During any year when precipitation deviates from average by more than 20 percent, however, the predicted water levels based on a variable-recharge rate would be closer to the observed levels. The final version of the model used variable-recharge rates.

A comparison between the observed and computed change of water levels after the final calibration of the model is shown for 16 wells in figure 26. Except in the areas affected by seepage of water to the principal aquifer from irrigation and from canals, agreement between computed and observed water levels generally was satisfactory in most of the valley and this indicates that the model should be reliable for making predictions.

Agreement between computed and observed water levels was poor for many wells at which water levels are affected by seepage from irrigation (see hydrographs for wells (C-3-1)9ccc-1, (C-3-1)33aab-1, and (C-4-1)15bdc-2 in figure 26). The poor agreement resulted from the lack of sufficient data to permit computation of variable-recharge rates from irrigation.

Water levels in well (C-2-1)9ccc-1 prior to 1980 (fig. 26) probably were affected by seepage from a canal that crosses gravel outcrops near the well. The canal was blocked south of the well in about 1980, and water levels began to decline. In the model, however, constant recharge was used for seepage from canals. Thus, the model was not sensitive to changes in canal recharge, and a large deviation between observed and computed water levels occurred after 1980. Agreement between computed and observed water levels in wells (D-1-1)5aaa-1 and (D-1-1)10cac-1 was satisfactory until about 1976-77, when they began to show considerable deviation. The reasons for these deviations were not determined.

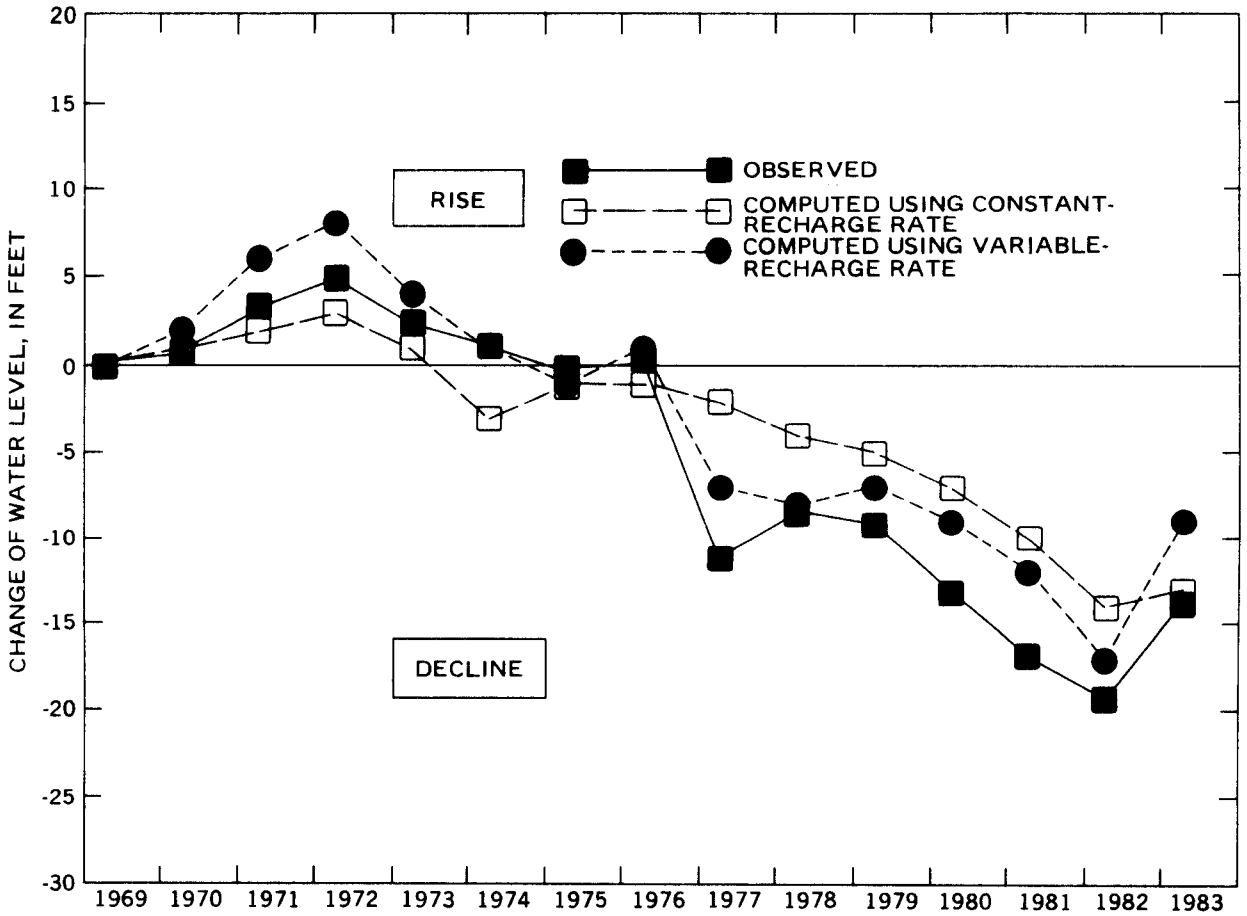


Figure 25.—Comparison of change of water levels computed from variable-and constant-recharge rates with observed change of water level in well (D-2-1)34acb-1, 1969-83.

SUMMARY

During 1969-82, the average annual recharge and discharge were estimated to be about 352,000 and 353,000 acre-feet. Withdrawal from wells averaged about 117,000 acre-feet per year, or about 10,000 acre-feet more than in 1964-68. Withdrawals for public supply and institutions increased by about 11,700 acre-feet per year during 1969-82, withdrawal for industry and air conditioning increased slightly, whereas withdrawal for other uses decreased. Water-level declines ranged from 5 to 15 feet in the southeast part of the valley where pumpage from large public supply wells was greater during 1969-82 than during the previous years. The largest rises of water levels, which were as much as 12 feet, occurred in the northeast and southwest parts of the valley. From February-March 1969 to February-March 1983, the quantity of ground water in storage in Salt Lake Valley increased by about 33,000 acre-feet.

A digital-computer model was calibrated to simulate, in three dimensions, the ground-water flow in the principal and shallow-unconfined aquifers in Salt Lake Valley. Simulations were made to evaluate the effects of projecting the 1982 rate of pumpage and increasing the 1982 discharge by 65,000 acre-feet, to the year 2020. The simulation at the increased rate of pumpage indicated that drawdowns would reach 40-60 feet in the area east of Sandy where most of the increase of withdrawals was simulated. About 75 percent of the increased withdrawals was salvaged from water that otherwise would have been discharged to the Jordan River and its tributaries.

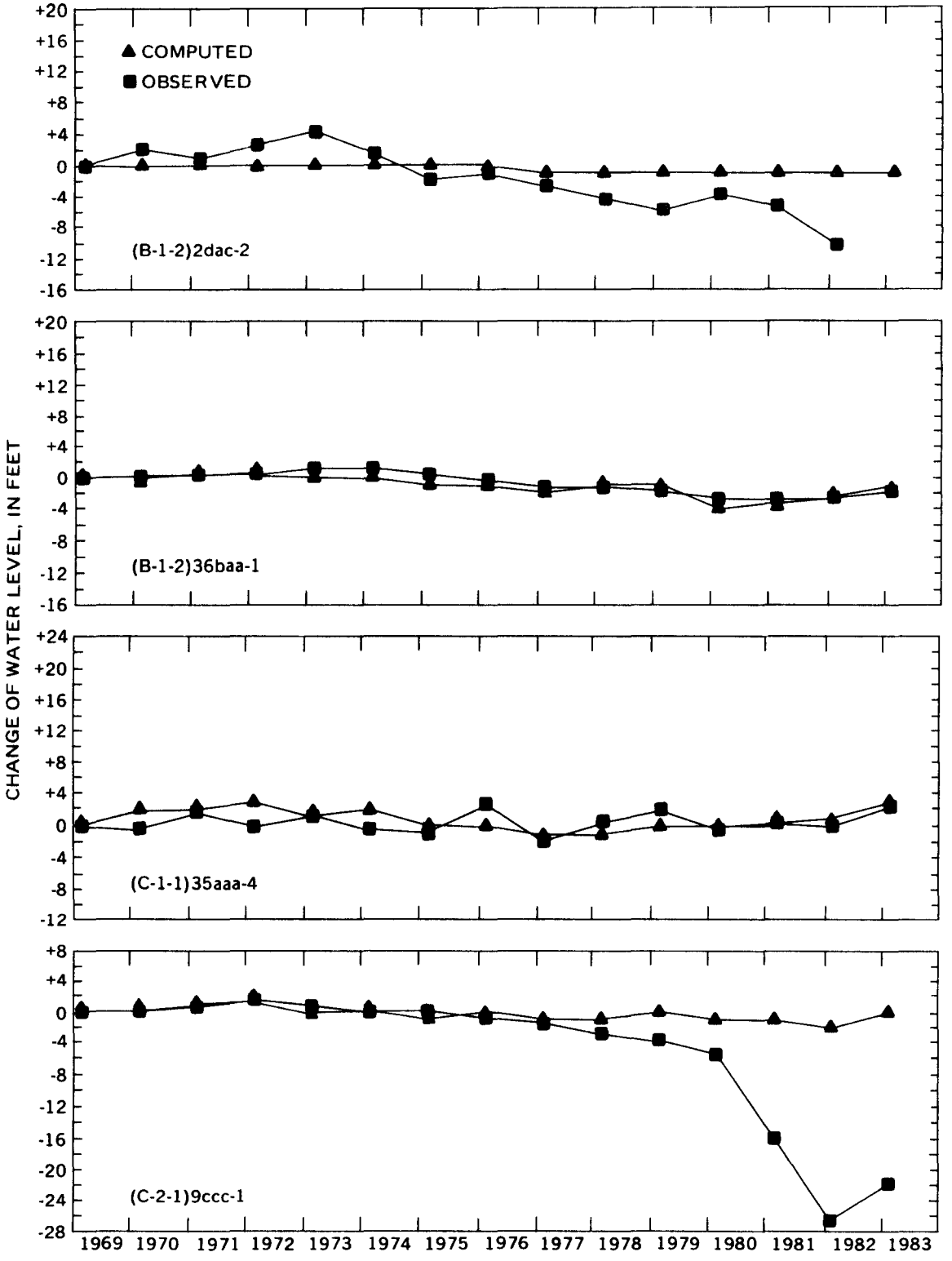


Figure 26.—Hydrographs of 16 wells showing observed and computed change of water levels, 1969-83.

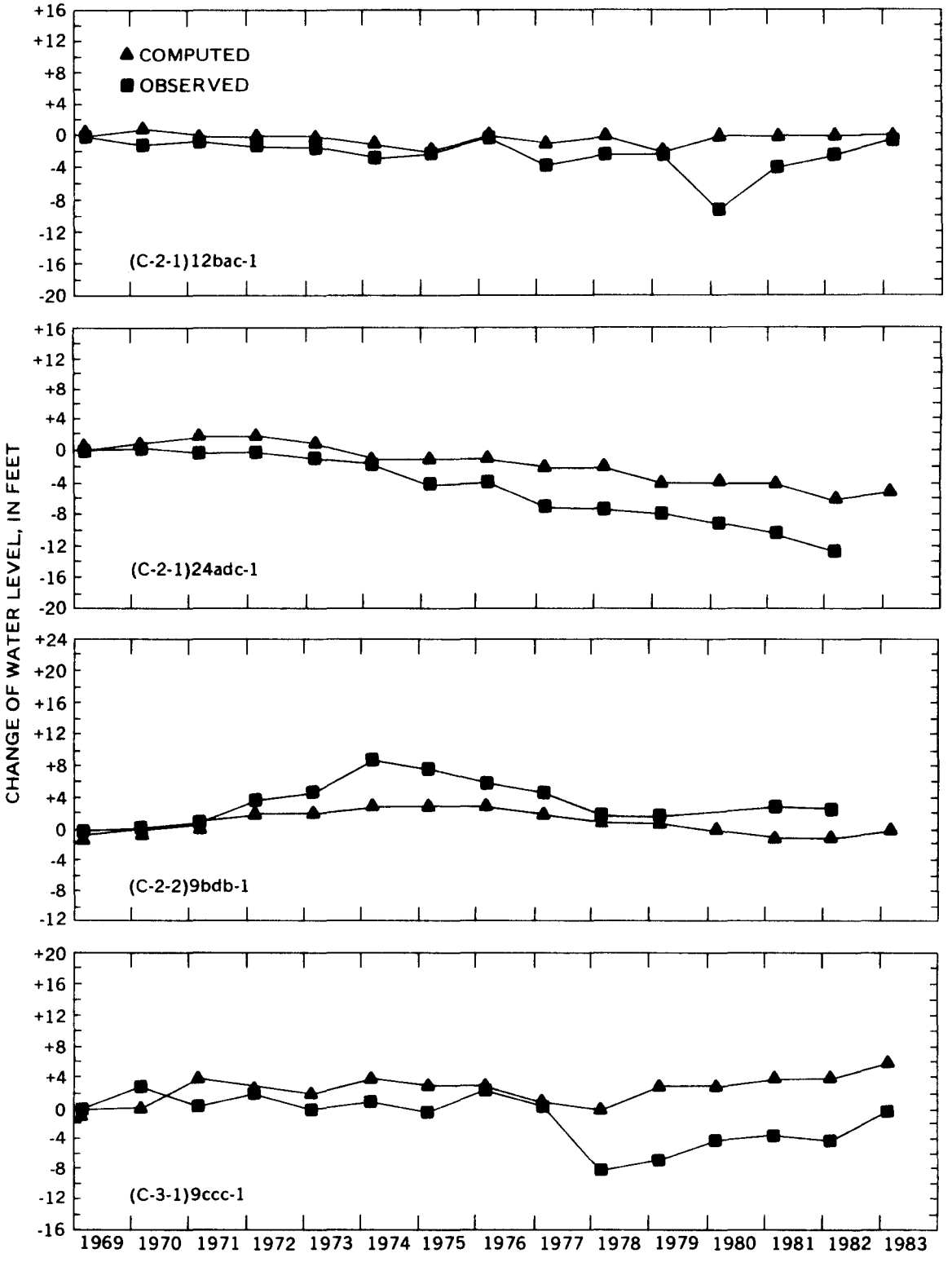


Figure 26.—Hydrographs of 16 wells showing observed and computed change of water levels, 1969-83—Continued.

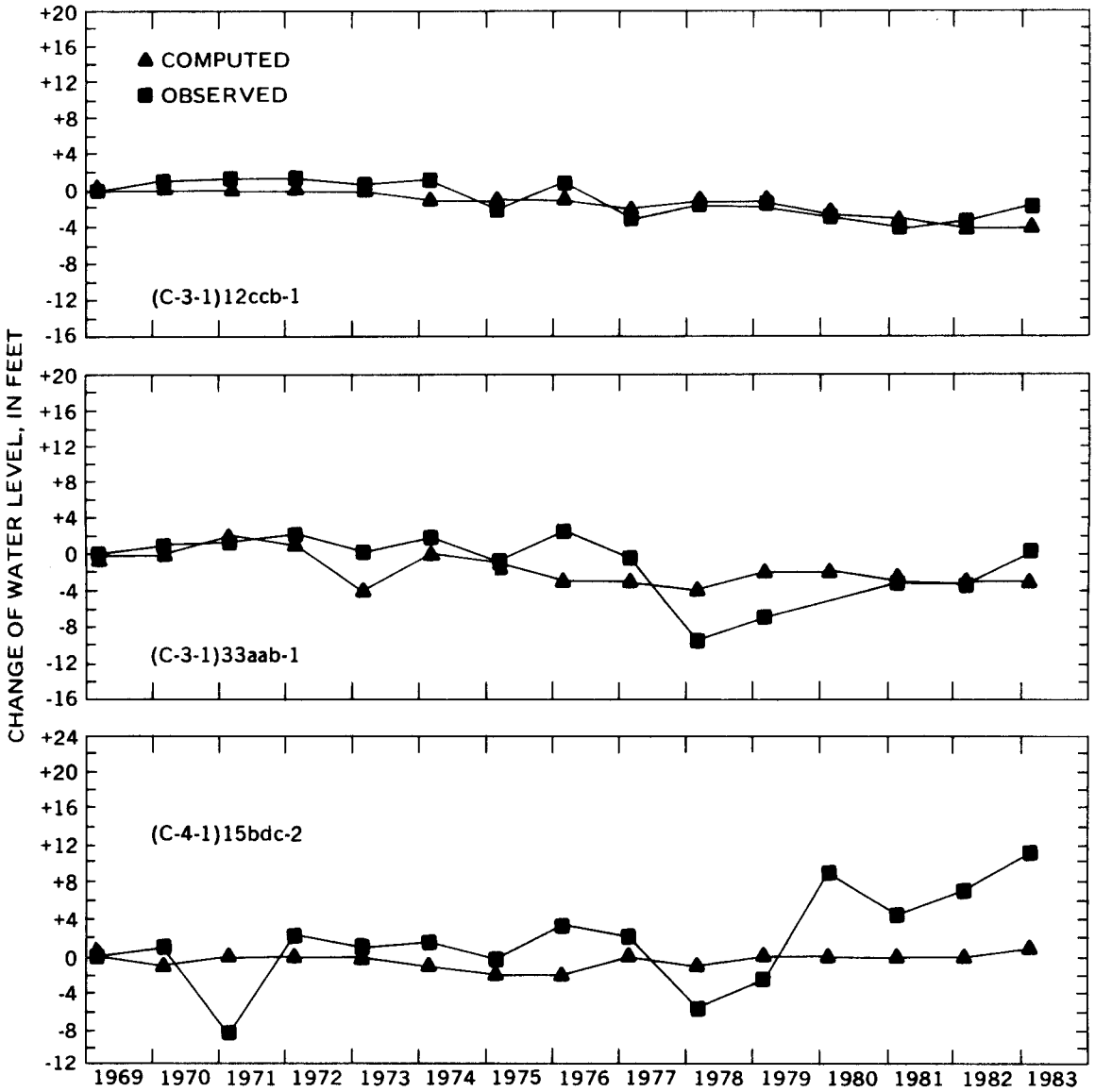


Figure 26.—Hydrographs of 16 wells showing observed and computed change of water levels, 1969-83—Continued.

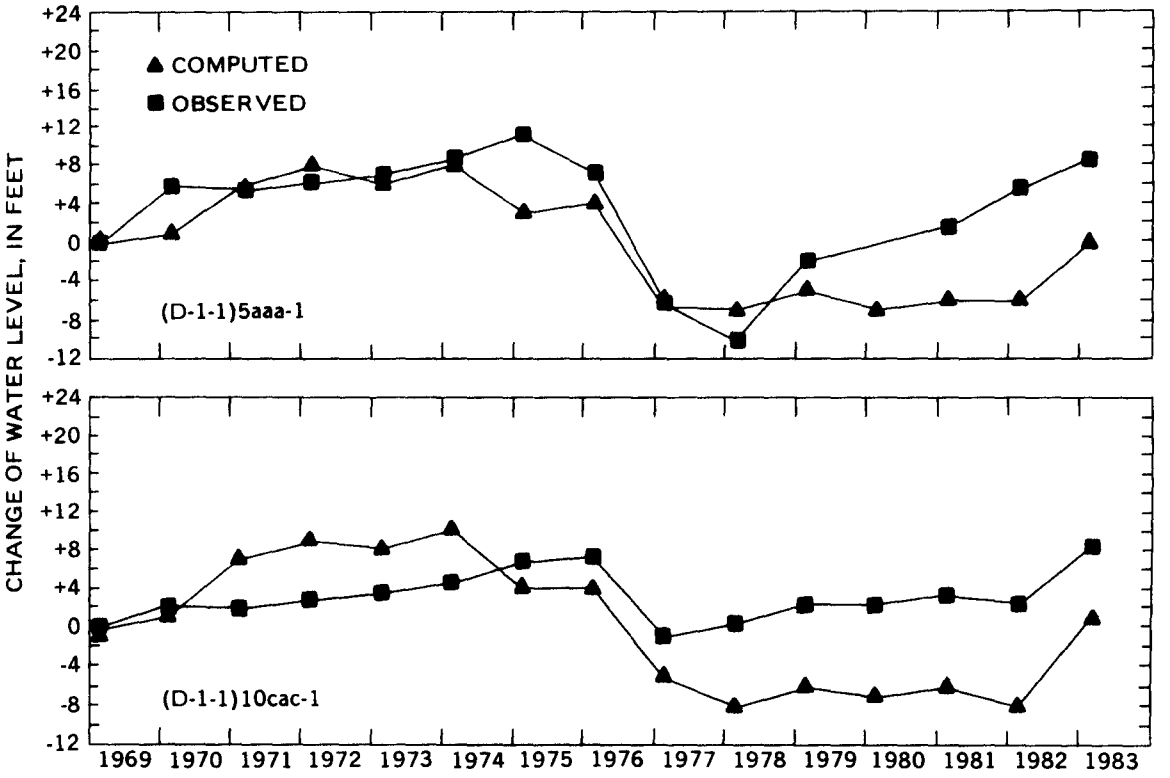


Figure 26.—Hydrographs of 16 wells showing observed and computed change of water levels, 1969-83—Continued.

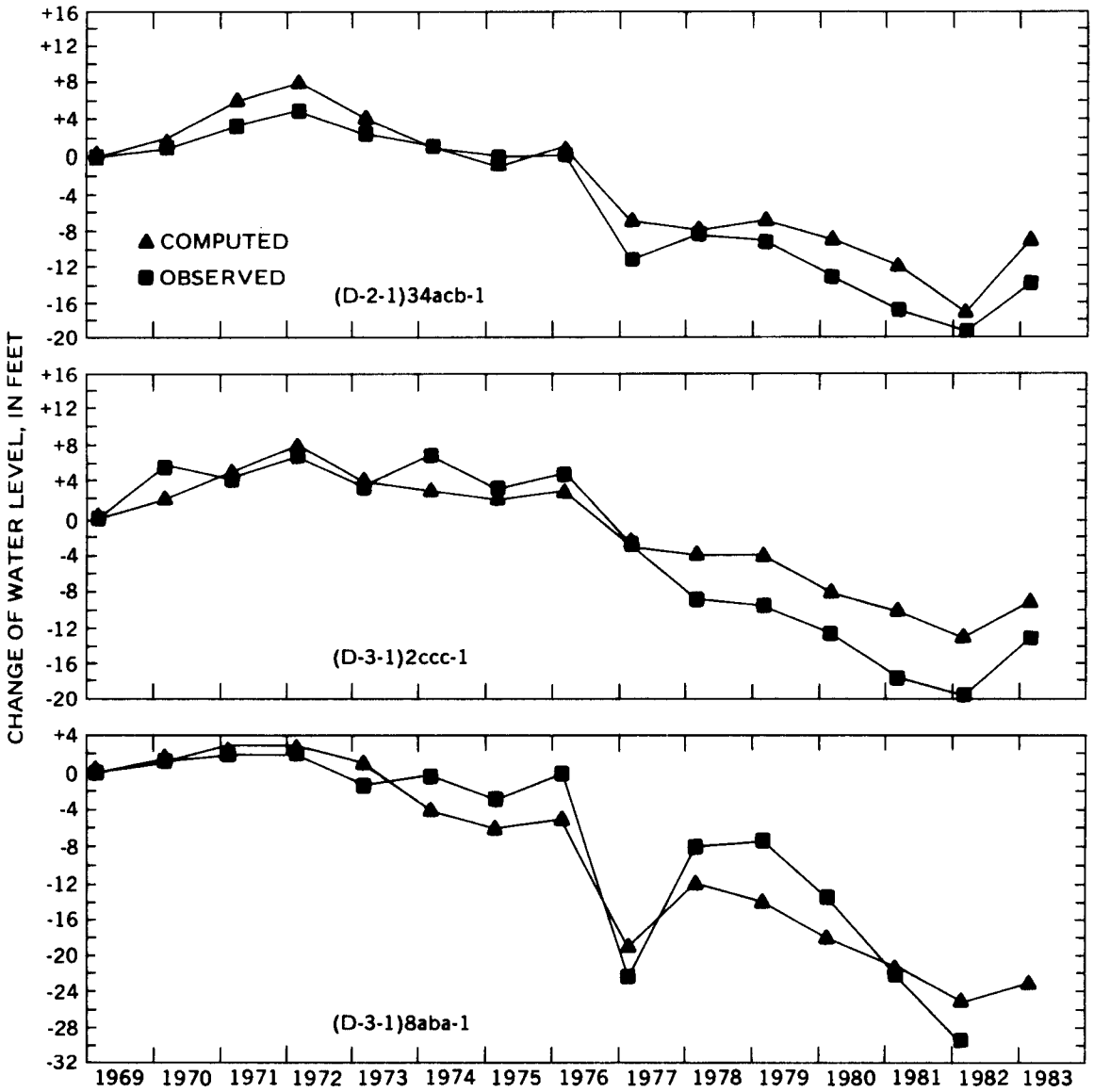


Figure 26.—Hydrographs of 16 wells showing observed and computed change of water levels, 1969-83—Continued.

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