THE GROUND-WATER SYSTEM AND SIMULATED EFFECTS OF GROUND-WATER WITHDRAWALS IN NORTHERN UTAH VALLEY, UTAH

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CONVERSION FACTORS

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

Multiply inch-pound units	By	To obtain metric units
acre	0.4047	square hectometer
acre-foot	0.001233	cubic hectometer
cubic foot per second	0.02832	cubic meter per second
foot	0.3048	meter
foot squared per second	0.0929	meter squared per second
foot squared per day	0.0929	meter squared per day
inch	25.40	millimeter
	2.540	centimeter
mile	1.609	kilometer
square mile	2.590	square kilometer

National Geodetic Vertical Datum of 1929 (NGVD of 1929): A geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called mean sea level.

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ABSTRACT

The effects of withdrawals from the principal ground-water reservoir in northern Utah Valley, Utah, were projected by means of a three-dimensional, finite-difference, digital-computer model, which was constructed to aid understanding of the ground-water system and to simulate ground-water flow. The model was compared with water levels measured in 1947, and observed water-level changes from 1947-83. The model was used to evaluate ground-water data presented in previous reports, to simulate varying quantities of ground-water withdrawal and recharge, and to estimate water-level changes for 1980-2000. The average annual rate of recharge for the area is assumed to be 190,000 acre-feet per year, and the average annual discharge from wells at the end of transient-state calibration was assumed to be 50,100 acre-feet per year. Water-level declines of as much as 25 feet are projected for the 20-year period if the average recharge rate is assumed and discharge from wells is as much as 91,400 acre-feet per year. During transient-state calibration, changes in recharge to the principal ground-water reservoir were shown to be a major cause of the variations in water levels.

INTRODUCTION

The U.S. Geological Survey, in cooperation with the Utah Department of Natural Resources, Division of Water Rights, evaluated the ground-water resources of northern Utah Valley, Utah, during 1980-82. As part of that study, Clark and Appel (1985) updated the definition of the ground-water system, previously described by Hunt and others (1953) and Cordova and Subitzky (1965). Northern Utah Valley is one of the fastest growing areas in the United States, as reflected by an increase in urban population from about 72,000 in 1960 to about 164,000 in 1980. In order to meet the water needs of this expanding population, annual ground-water withdrawals for public supply increased from about 5,000 acre-feet during 1963 to about 20,000 acre-feet during the late 1970's (Clark and Appel, 1985, p. 73).

Purpose and Scope

The purpose of this report is (1) to improve estimates of hydraulic properties and recharge to and discharge from the ground-water system in northern Utah Valley; and (2) to project effects of potential increases in withdrawals from the principal ground-water reservoir on ground-water levels. An important tool used to accomplish this purpose was a three-dimensional, finite-difference, digital-computer model, and the construction and calibration of the model is descibed in this report. The model will be useful to the State's Division of Water Rights in dealing with water-allocation problems. A listing of the data used in the model is available in the files of the U.S. Geological Survey, Salt Lake City, Utah.

The system of numbering wells in Utah is based on the cadastral land-survey system of the U.S. Government. The number, in addition to designating the well, 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 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 within the 10-acre tract. Thus, (D-5-1)23dab-3 designates the third well constructed or visited in the NW1/4NE1/4SE1/4 sec. 23, T. 5 S., R. 1 E. The numbering system is illustrated in figure 1.

Topography and Climate

Northern Utah Valley is part of a north-trending elongate basin about 40 miles long and 10 to 20 miles wide, at the eastern edge of the Basin and Range physiographic province in north-central Utah (fig. 2). The valley is bounded by the Wasatch Range on the east, the Traverse Mountains on the north, and the Lake Mountains on the west. The mountains that adjoin the valley lowlands are bounded by benches (terraces) formed by glacial Lake Bonneville, which extend toward the center of the valley and Utah Lake. The altitude of the valley floor ranges from less than 4,500 feet near Utah Lake to 5,200 feet near the mountains. The highest point in the Wasatch Range is Mt. Timpanogos with an altitude of 11,750 feet, whereas the Lake and Traverse Mountains attain maximum altitudes of approximately 7,600 and 6,600 feet.

The climate of the area is generally temperate and semiarid with a typical frost-free season from late April to mid-October. Precipitation increases across the valley and on the adjoining mountains as the altitude increases, varying from less than 12 inches per year near Utah Lake to more than 50 inches per year at the crest of the Wasatch Range (U.S. Weather Bureau, 1963).

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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.

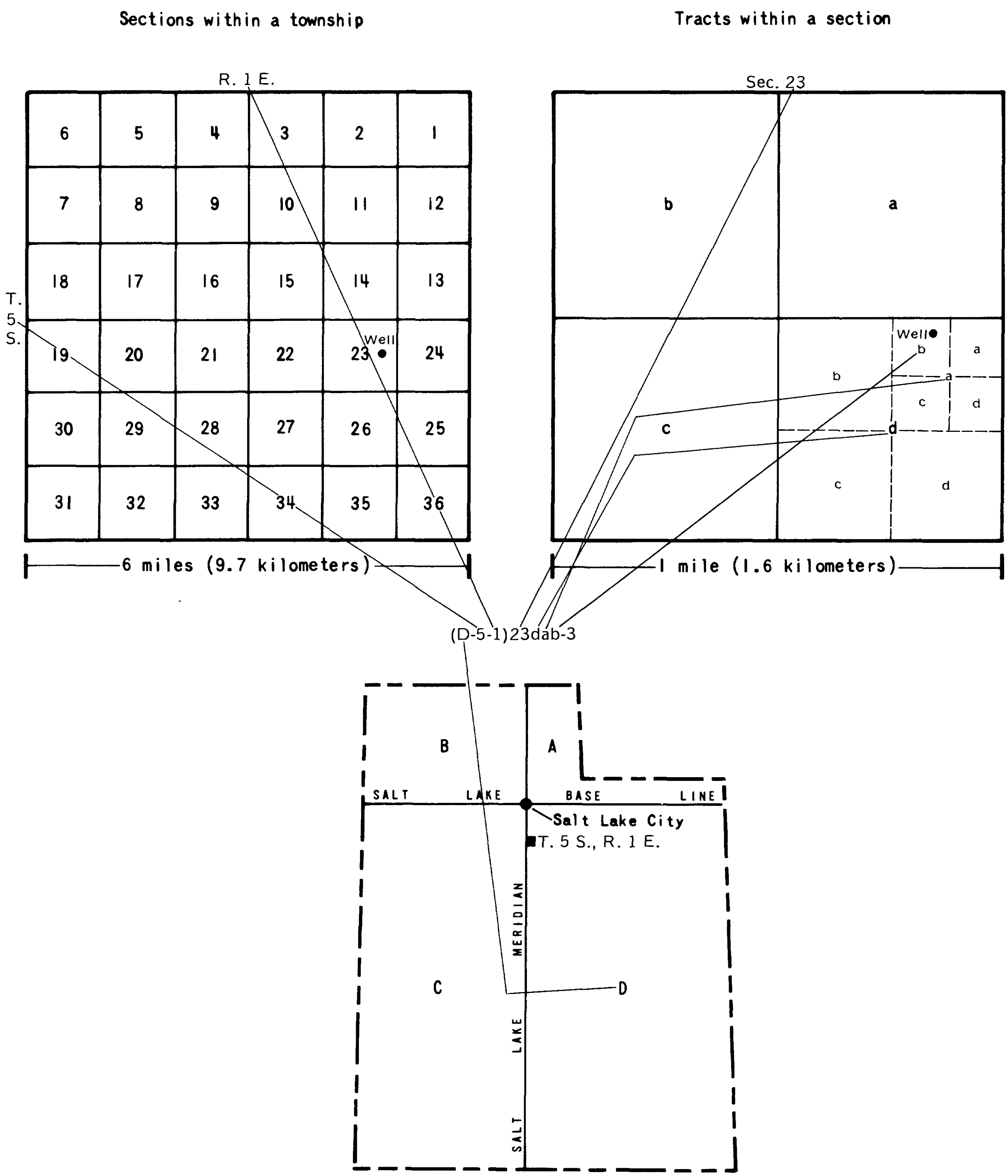


Figure 1.—Well-numbering system used in Utah.

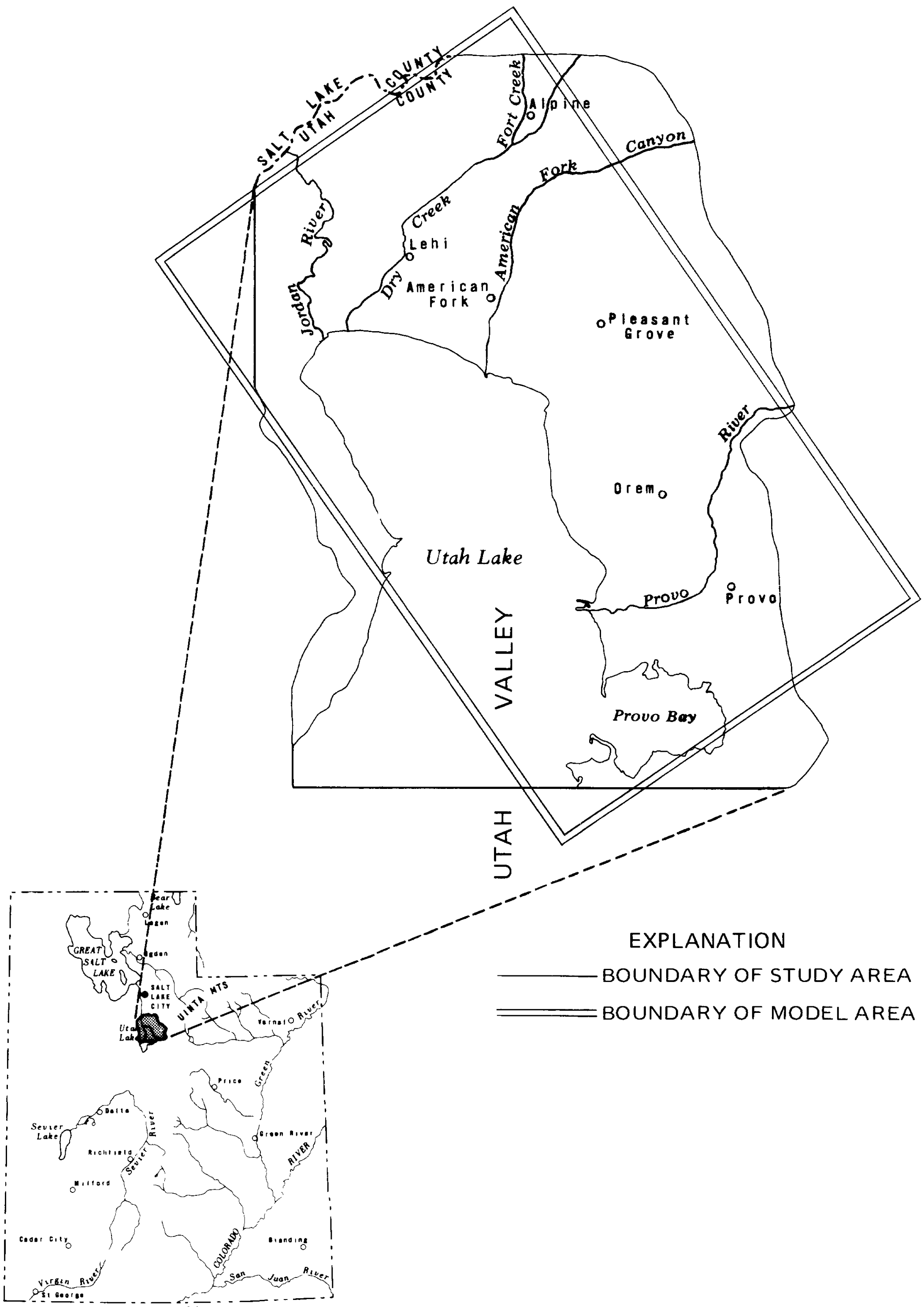


Figure 2.—Study area and model boundary used in the digital-computer model of the ground-water reservoir of northern Utah Valley.

Hydrogeologic Setting

Utah Valley is a graben formed by normal faulting during Tertiary and Quaternary time. The mountain block east of the graben has a total displacement of as much as 7,000 feet along the Wasatch fault zone (Hunt and others, 1953, p. 38). Erosion of the mountains provided the sediment that filled the graben and formed the ground-water reservoir in northern Utah Valley. The consolidated rocks that form the mountains are of Precambrian to Tertiary age and are predominately limestones and quartzites, consequently debris from these rocks also predominate in the basin fill. The fill consists mostly of unconsolidated, interbedded, lascustrine, alluvial-fan, and fluvial deposits of Quaternary and possibly Tertiary age. Coarse-grained materials, which predominate in the major aquifers in the fill, are thickest near the mountains and extend farthest into the valley along river channels. Fine-grained sediments, which predominate in confining layers or zones, are thickest in the basin center (fig. 3). The maximum thickness of the basin fill is unknown, however, the deepest known water well bottomed in fill at a depth of about 1,200 feet near the U.S. Steel Co., Geneva Works.

Surface-water inflow in major streams to northern Utah Valley is a principal source of ground-water recharge and also the primary source of water for irrigation. This inflow averaged approximately 390,000 acre-feet per year during 1963-82, which was greater than the long-term average primarily because of greater than normal precipitation during that period. An average of 78 percent of the total inflow during 1963-82 was in the Provo River and nearly 90 percent was in the American Fork and the Provo River combined. In addition, small intermittent and ephemeral streams contribute an estimated 10,000 acre-feet per year of inflow (Clark and Appel, 1985, p. 13-19). The seasonal fluctuation of surface flow is extremely large, with the greatest flow resulting from the spring snowmelt.

PRINCIPAL GROUND-WATER RESERVOIR

The principal ground-water reservoir in northern Utah Valley is in the basin fill, and it includes an unconfined (water-table) aquifer and three confined (artesian) aquifers. An unconfined aquifer in pre-Lake Bonneville deposits along the mountain fronts correlates laterally with the confined aquifers farther from the mountains. Near the mountains, sediments are generally coarse grained and fine-grained confining layers are thin or absent; whereas toward the center of the valley fine-grained sediments predominate. Thus, ground water becomes confined as it moves from the mountains toward Utah Lake. Unconfined ground water also occurs locally in perched water-table aquifers, in flood-plain deposits along stream channels, and in the valley lowlands within a few feet of the land surface. Although these deposits may be minor sources of recharge to or areas of discharge from the underlying aquifers, they are not considered to be part of the principal ground-water reservoir.

The aquifers are separated by confining layers which consist of fine-grained beds that are at least several feet thick. Vertical movement of ground water from deeper confined aquifers toward the discharge areas at the land surface is a result of a substantial pressure gradient between the aquifers. The hydrostatic pressure generally increases with depth. The vertical gradient, or head difference, is as great as 50 feet between aquifers

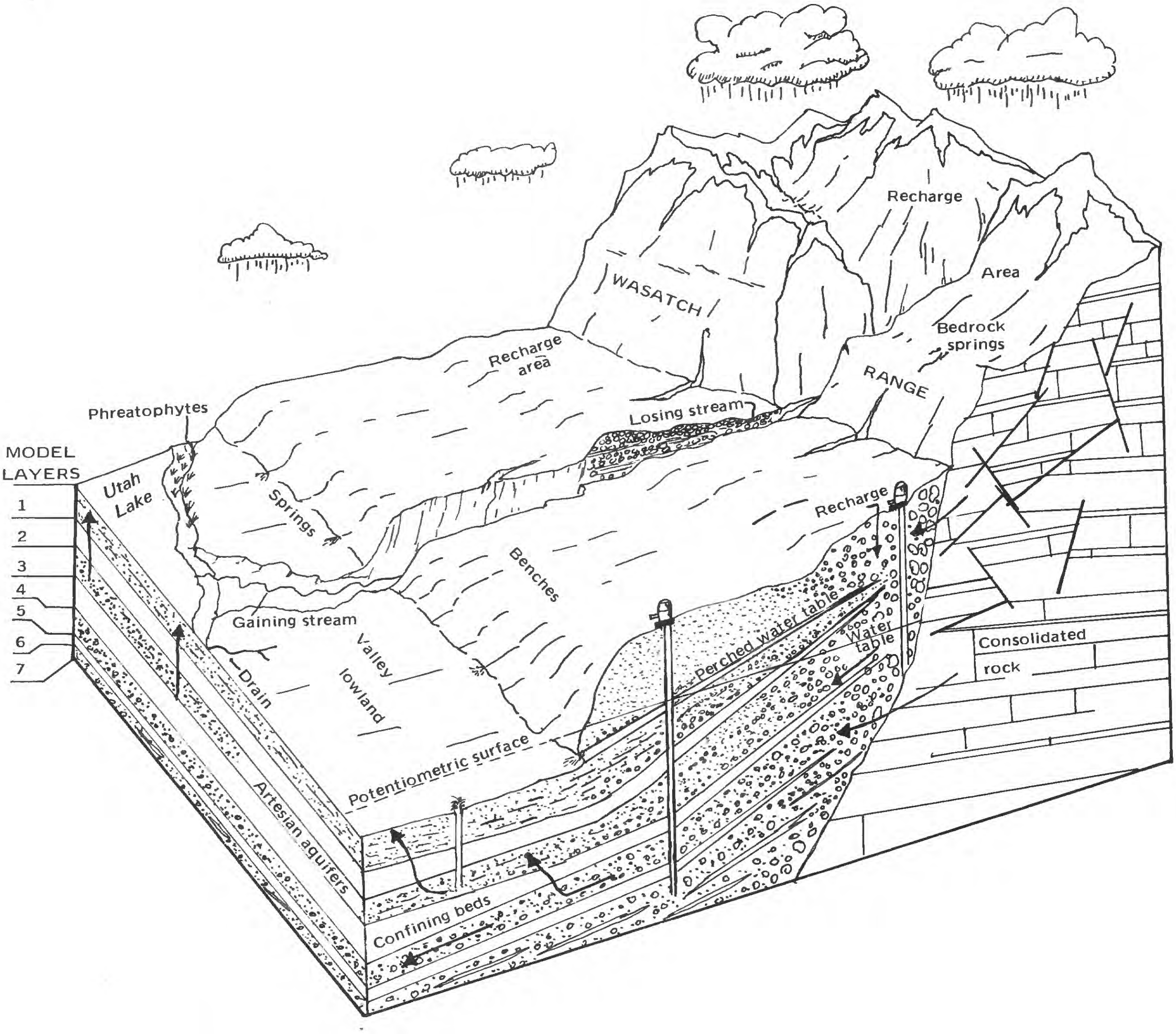


Figure 3.—Generalized block diagram showing model layers, probable ground-water movement at several points, sources of recharge, points of discharge, and generalized lithology.

in some areas. The three confined aquifers are generally the same as those described by Hunt and others (1953). Although the confined aquifers can be separated locally, their thickness, continuity, and lithology varies, making it difficult to correlate them across the valley. This has been illustrated in lithologic cross sections in previous reports (Hunt and others, 1953, pl. 4; Clark and Appel, 1985, figs. 17-19).

DESIGN AND CONSTRUCTION OF THE GROUND-WATER FLOW MODEL

The finite-difference, three-dimensional, digital-computer model used to simulate the ground-water system and flow in northern Utah Valley was developed by McDonald and Harbaugh (1983).

Data used in the construction and calibration of the model are the result of hydrologic studies in northern Utah Valley that span almost 50 years. Data were obtained from Taylor and Thomas (1939) and Hunt and others (1953) for steady-state calibration; from Cordova and Subitzky (1965) for transient-state calibration; and from Clark and Appel (1985) and Appel and others (1982) for various aspects of the model construction.

Subdivision of the Ground-Water Reservoir into Layers

The ground-water reservoir consists of a complex, interconnected, multiple-aquifer system that was first described by Hunt and others (1953, p. 79-85) as consisting of three artesian aquifers and a shallow water-table aquifer. This four-aquifer definition also was used by Cordova and Subitzky (1965) and Clark and Appel (1985), with slight changes in depth to and extent of the aquifers. The three artesian aquifers are lateral extensions of a deep water-table aquifer along the mountain front (fig. 3), and there is a substantial upward vertical gradient between aquifers, with head differences as great as 50 feet. As illustrated in lithologic cross sections in previous reports (Hunt and others, 1953, pl. 4; Clark and Appel, 1985, figs. 17-19), the aquifers vary in thickness, continuity, and lithology.

The model consists of seven layers, which are illustrated in figure 3. Layer 1 represents a shallow water-table aquifer in the discharge area; layers 2, 4, and 6 represent confining layers; layer 3 represents the shallow artesian aquifer in deposits of Pleistocene age; layer 5 represents the deep artesian aquifer in deposits of Pleistocene age; and layer 7 represents the artesian aquifer in deposits of Quaternary or Tertiary age.

Model Grid

A node-centered grid with variable spacing was used to model the ground-water reservoir. The grid consists of 36 rows and 19 columns. The largest active nodes, which cover 0.85 square mile, generally are in areas where data are sparse. The smallest nodes, which cover 0.25 square mile, generally are in areas where there are numerous wells, large ground-water withdrawals, or historic water-level measurements. Of the total of 4,788 nodes, 2,573 are active. Layers 4-7 each have 441 active nodes, layer 3 has 321 active nodes, and layers 1 and 2 each have 244 active nodes. Figure 4 illustrates the areas of inactive nodes.

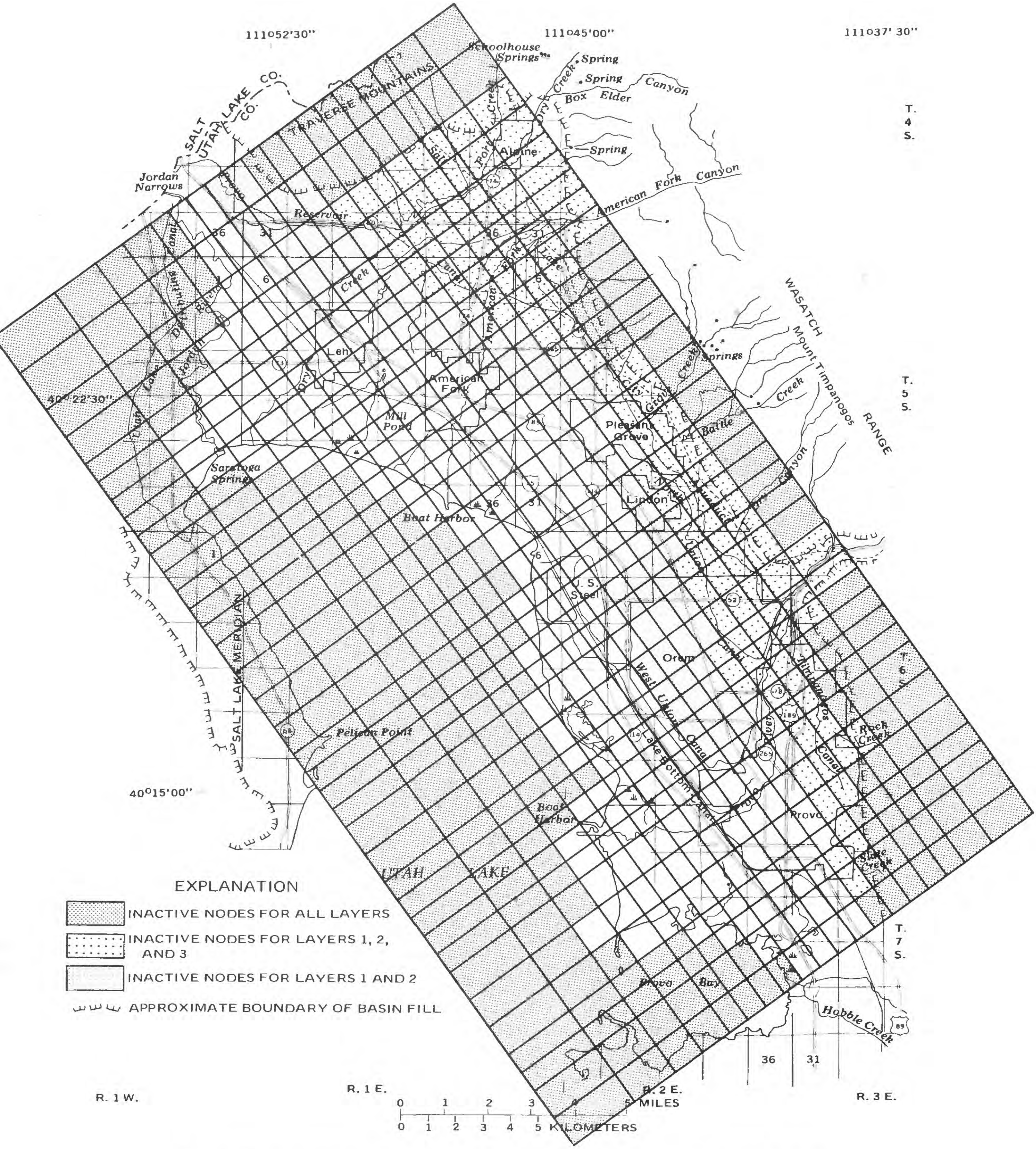


Figure 4.—Locations of inactive nodes in the model for northern Utah Valley.

A generalized block diagram of the model grid is shown in figure 5. Layers 4 through 7 represents the water-table aquifer along the mountain front and individual aquifers of confining layers elsewhere. Layer 3 is modeled only where it is saturated; therefore, it does not extend as far east as the underlying layers. Layers 1 and 2 are modeled only in the areas where they receive recharge by upward leakage from the artesian aquifers.

Boundary Conditions

The inactive nodes illustrated in figure 4 are modeled with transmissivities of zero; and they act as a no-flow, impermeable boundary, generally surrounding the active nodes. On the east, this boundary coincides with the approximate location of the contact between the basin fill and the consolidated rocks of the Wasatch Range. On the west, the no-flow boundary is approximately one node lakeward from the shoreline of Utah Lake. A no-flow boundary also was placed beneath layer 7, on the assumption that there is no upward leakage from below.

Simulations for steady-state calibration, which were made prior to transient-state simulations, used slightly different boundary conditions. Constant-head nodes were placed in layers 5 and 7 on the east border to simulate partial recharge from the consolidated rocks of the Wasatch Range. The initial-head values for these nodes were assumed to be similar to or higher than present head values. After steady-state calibration, the constant-head nodes were changed to recharge nodes. The flow rates calculated by the model were input as constant-recharge rates.

At the western edge of the model grid, constant-head nodes were placed one node out into Utah Lake and along the Jordan River to simulate ground-water discharge into the lake and river by upward leakage from the artesian aquifers. Near the lake, these discharge nodes were placed in layers 1, 2, and 3 near the shoreline of the lake where most discharge is by upward leakage by springs and at drains in swampy areas at about lake level. Dustin and Merritt (1980, p. 5-7) suggest that discharge into Utah Lake by subsurface springs generally occurs near the shoreline. The initial heads for these constant-head nodes were assumed to be similar to steady-state water levels for layer 3 and approximately equal for layers 1 and 2 to the altitude of the lake surface at the compromise level of 4,489 feet above sea level. These constant-head discharge nodes were used during steady-state (fig. 6) and transient simulations.

Simulations were made using general head-boundary nodes instead of constant-head nodes along the western border of the model, so that heads in those nodes could vary. The results of both methods were compared for steady-state and transient-state calibration, with the results being nearly equal. However, the constant-head nodes seemed a better approximation of the system and required one less unknown--the conductance of the interface of the aquifer-cell boundary--than did the general head boundary. Thus, constant-head nodes were used.

At the northern boundary of the model grid, no-flow nodes were placed everywhere except near the Jordan Narrows where constant-head nodes were placed to simulate outflow from the ground-water reservoir through the narrows. The southern boundary of the modeled area was considered to mark a

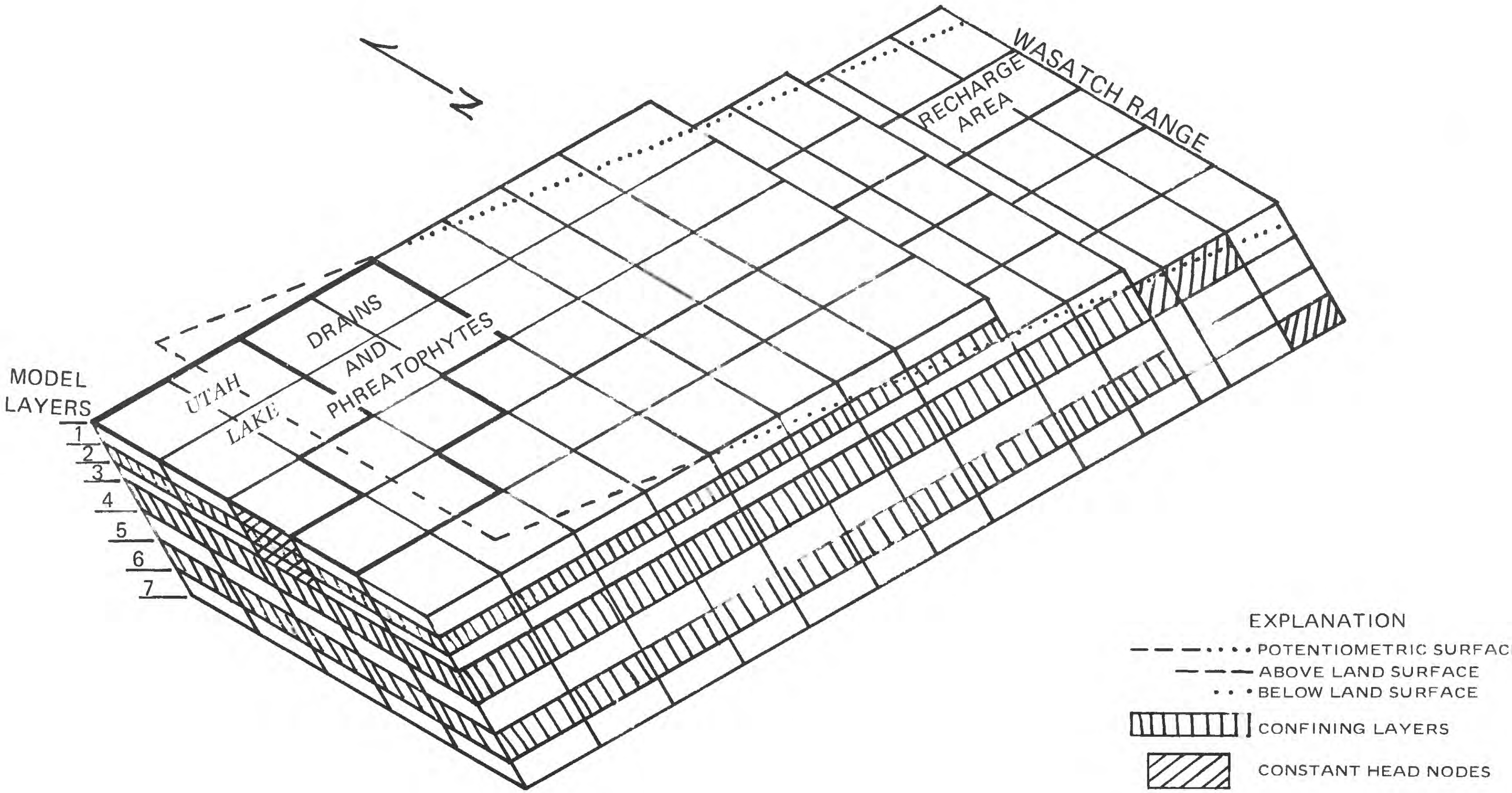


Figure 5.—Generalized block diagram showing three-dimensional model grid, areas of recharge, discharge, constant-head nodes, and model layers.

ground-water divide; thus no water discharges to the south from the model area. This assumption was based on the configuration of the potentiometric surfaces reported by Clark and Appel (1985, figures 23 and 24).

DATA INPUT

Initial Conditions

Ground-water withdrawal from the artesian aquifers in northern Utah Valley began about 1885 when water flowed from a 1.25-inch well driven to a depth of 75 feet near Lehi (Hunt and others, 1953, p. 63). By 1940, the ground-water withdrawal from wells was estimated to average about 30,000 acre-feet per year (Hunt and others, 1953, p. 73), or about 15 percent of the total discharge from the system. Ground-water withdrawals probably did not significantly affect water levels prior to 1947; thus, water levels for 1947 for the three artesian aquifers (Hunt and others, 1953, pl. 3) were used in the model as initial water levels for steady-state simulation. For those parts of the study area for which water levels were not available for 1947, more recent data from 1948-82 (Appel and others, 1983) were used for initial water levels in the model.

Recharge

Clark and Appel (1985, table 5) estimated total recharge to the principal ground-water reservoir to be about 190,000 acre-feet per year in the area simulated by the model. This recharge includes seepage from streams, irrigation canals, irrigated fields, lawns, and gardens; infiltration of precipitation; and subsurface inflow from consolidated rocks. The primary recharge area is a narrow strip of land adjacent to the mountain fronts (Clark and Appel, 1985, fig. 9) which is not underlain by fine-grained material that impedes downward movement of water. Recharge to the area simulated by the model occurs in the primary recharge area by seepage or infiltration and across the contact of basin fill and consolidated rock by subsurface inflow (fig. 3). Recharge to the area simulated by the model was input by the use of recharge nodes.

Total seepage from stream channels and irrigation canals was estimated using streamflow records for the major tributaries and from seepage estimates for some of the tributaries and canals (Clark and Appel, 1985, p. 22-29). Recharge from this seepage was based on the average annual streamflow during 1963-82, where records were available. The annual seepage losses estimated for the stream channels and their associated irrigation canals are: the Provo River--45,000 acre-feet; American Fork--13,400 acre-feet; Fort Creek--2,100 acre-feet; Dry Creek--5,500 acre-feet; Rock Creek--2,000 acre-feet; Slate Creek--1,500 acre-feet; Grove and Battle Creeks--500 acre-feet each; and other small streams--3,000 acre-feet. These rates were entered in recharge nodes as close as possible to the actual locations.

Seepage losses from the American Fork and associated canals were calculated on the basis of measurements and estimates made during 1981-82 and from records of daily discharge. Seepage losses in the natural channel in the first 1.25 miles downstream from the mouth of the canyon ranged from 100 percent, when the discharge was less than 20 cubic feet per second, to 35 percent, when the discharge was 200 cubic feet per second.

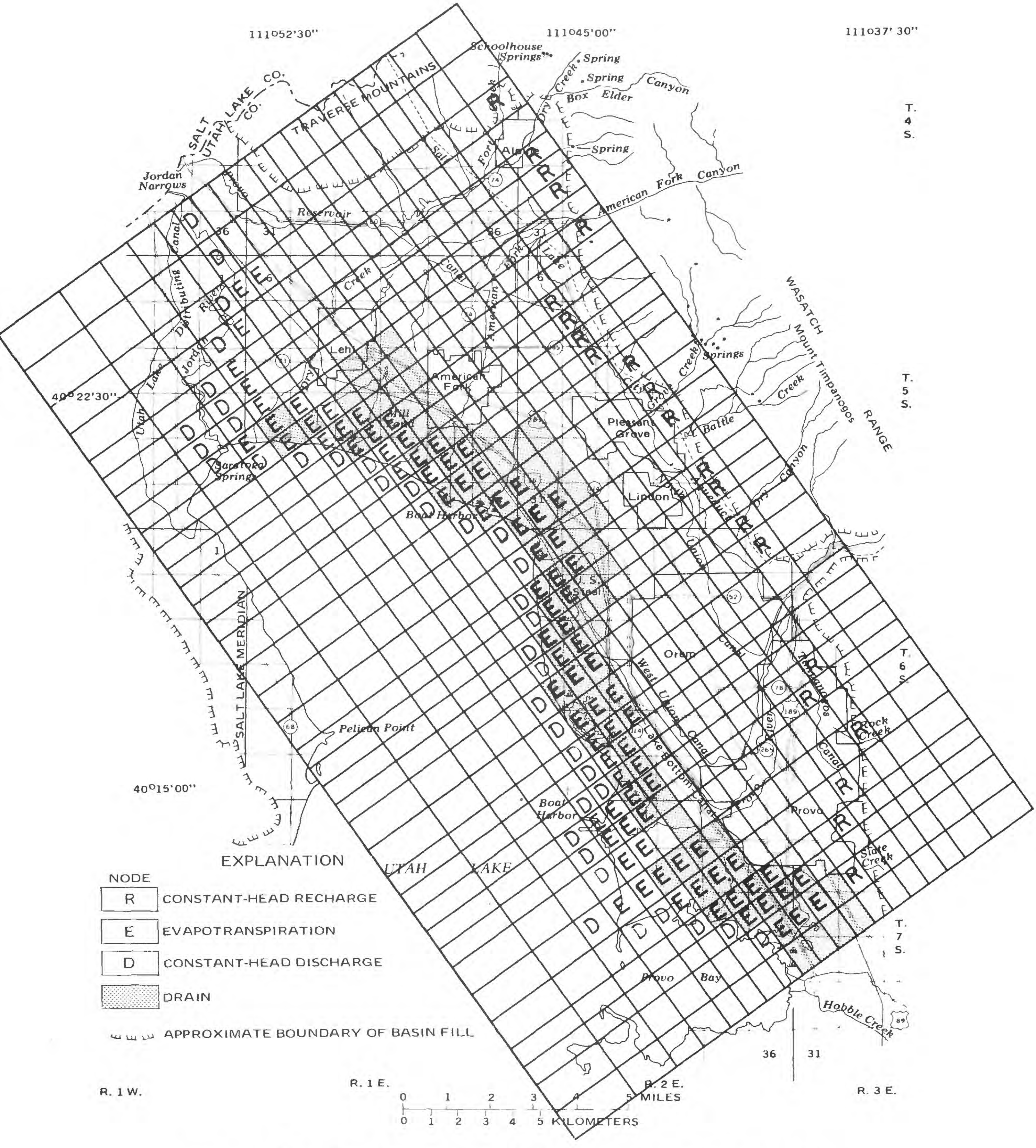


Figure 6.—Locations of nodes for constant-head recharge, evapotranspiration, constant-head discharge, and drains used for steady-state simulations in the model for northern Utah Valley.

Seepage from the natural channel of the Provo River was based primarily on two studies, one conducted during 1962 and the other from 1967-77. The results of these studies indicate that of the total loss of 30,000 acre-feet, most occurs within 2 miles of the mouth of the canyon. Seepage losses of an additional 15,000 acre-feet per year were estimated to occur from irrigation canals which receive water from the Provo River.

Evidence of the magnitude of recharge to the principal ground-water reservoir by seepage from streams was substantiated by correlations of discharge of the Provo River, the American Fork, and Dry Creek with fluctuations of water levels in wells near the mouths of the canyons where the streams enter the valley. Water levels in wells drilled through predominately coarse-grained sediments near the American Fork and Dry Creek rose about 50 feet during 1982 in response to above-average streamflow.

Recharge in the primary recharge area by seepage from irrigated fields, lawns, and gardens and by direct precipitation was calculated to be approximately 15,000 acre-feet per year. Seepage from about 5,000 acres of irrigated fields was estimated to be about 8,000 acre-feet per year based on the following: quantity of water applied, consumptive use of the crops, permeability of the soils, and the method of application. Seepage from lawns and gardens was estimated to be 2,000 acre-feet per year based on land use, municipal water records, and an assumed loss of one-third of the water applied. Annual recharge by infiltration of direct precipitation was estimated to be 5,000 acre-feet based on the area underlain by permeable soils, an average precipitation rate of 16.5 inches per year, and an assumed infiltration loss of 20 percent. The 15,000 acre-feet of recharge from these sources was input as a uniform rate over the primary recharge area for the steady-state and transient-state calibrations of the model.

Recharge from subsurface inflow occurs primarily as direct movement of water in bedrock in the adjoining mountains through fractures, bedding planes, and solution channels into the basin fill. Most of the inflow is from the Wasatch Range, which contains great thicknesses of limestone that is deformed and fractured and generally dips southwestward toward Utah Valley. Recharge by subsurface inflow to the area simulated by the model was estimated to be a minimum of 100,000 acre-feet per year by Clark and Appel, (1985, p. 31-38). Total subsurface inflow from the bedrock to the basin fill was calculated using a variation of the Darcy equation for 12 areas along the mountain fronts with similar hydraulic characteristics (Clark and Appel 1985, fig. 9). Detailed studies were made in the vicinity of the mouths of the American Fork and Dry Creek to provide more accurate estimates of subsurface inflow based on the following: seasonal variations of discharge through saturated sediments, volume of sediments saturated by large water-level rises, and seepage losses based on discharge measurements. Subsurface inflow was simulated in the model by the use of constant-head nodes (fig. 6) until steady-state calibration was complete. Total recharge from these nodes was calculated by the model to be about 100,000 acre-feet per year, a close approximation of the minimum estimate of Clark and Appel (1985, p. 31-38). The flow rates for each of these nodes along the model boundary were input as recharge rates as the last part of steady-state calibration.

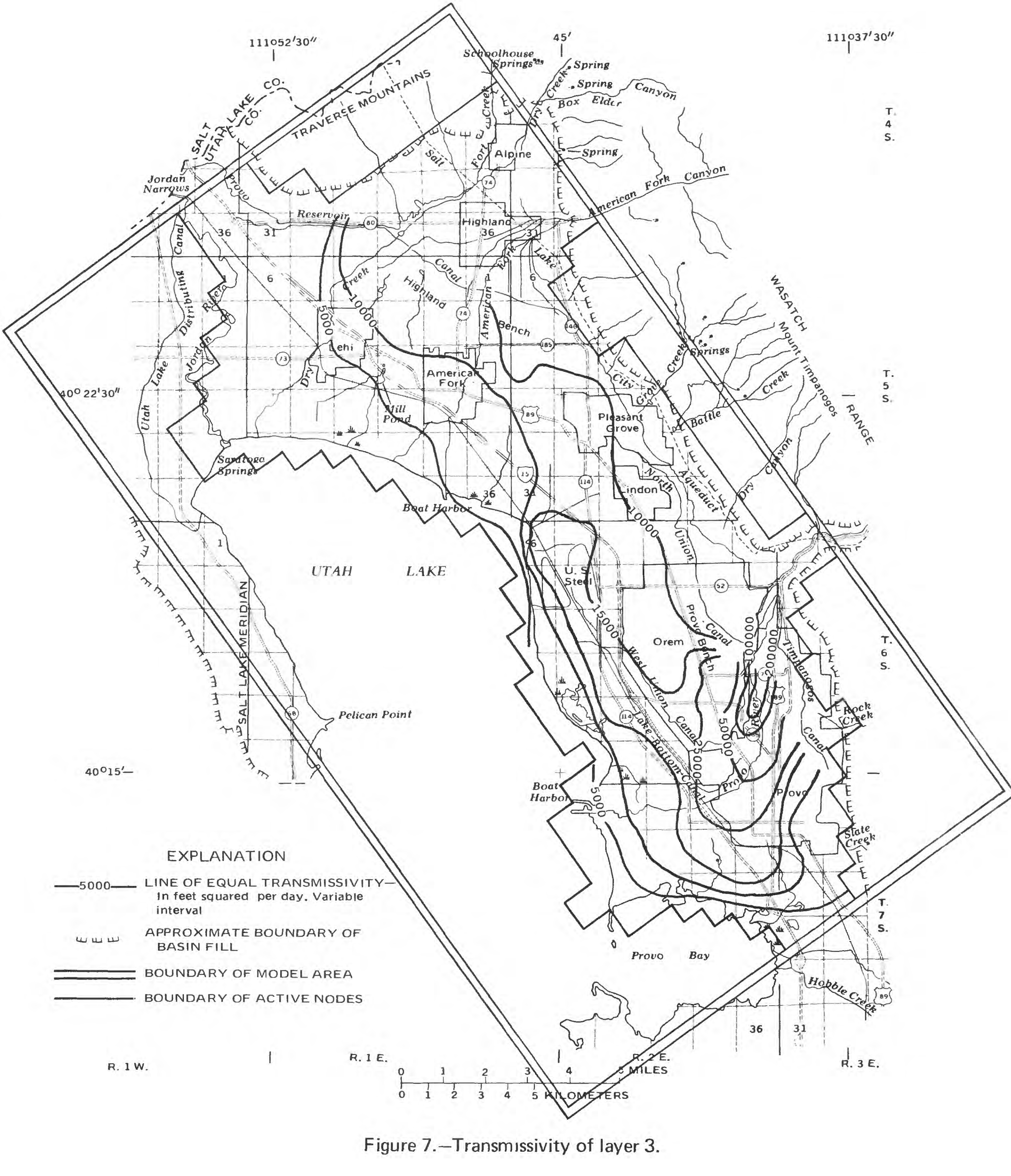
Hydraulic Properties of Aquifers and Confining Layers

The horizontal permeabilities of the three artesian aquifers and their lateral extensions to the water-table aquifer, were estimated in part using transmissivities *T*, reported by Clark and Appel (1985, table 9). The *T*, determined from aquifer tests ranged from more than 200,000 feet squared per day (in model layers 5 and 7) in coarse-grained sediments near the mountain front and in alluvial channels to about 1,000 feet squared per day (in model layer 3) in fine-grained sediments in the basin center.

Transmissivity also was estimated by multiplying the thickness of sediments described in drillers' logs by the average hydraulic conductivity *K*, for that sediment type. Values for hydraulic conductivity were derived from values reported by Clark and Appel (1985, table 9) and from aquifer tests conducted in the adjoining southern Utah Valley, which is hydrologically similar (Cordova, 1970). The average hydraulic conductivity *K*, for unconsolidated materials in Utah Valley as calculated from the aquifer tests at about 50 wells are listed below:

Material	Average hydraulic conductivity *K*, (feet per day)
Coarse gravel, cobbles, and boulders	>500
Gravel	450
Sand and gravel	200
Sand	100
Cemented conglomerate	25

The transmissivity data were used to prepare *T*, maps for the three artesian aquifers (figs. 7, 8, 9).



EXPLANATION

—5000— LINE OF EQUAL TRANSMISSIVITY—In feet squared per day. Variable interval

APPROXIMATE BOUNDARY OF BASIN FILL

BOUNDARY OF MODEL AREA

BOUNDARY OF ACTIVE NODES

Figure 7.—Transmissivity of layer 3.

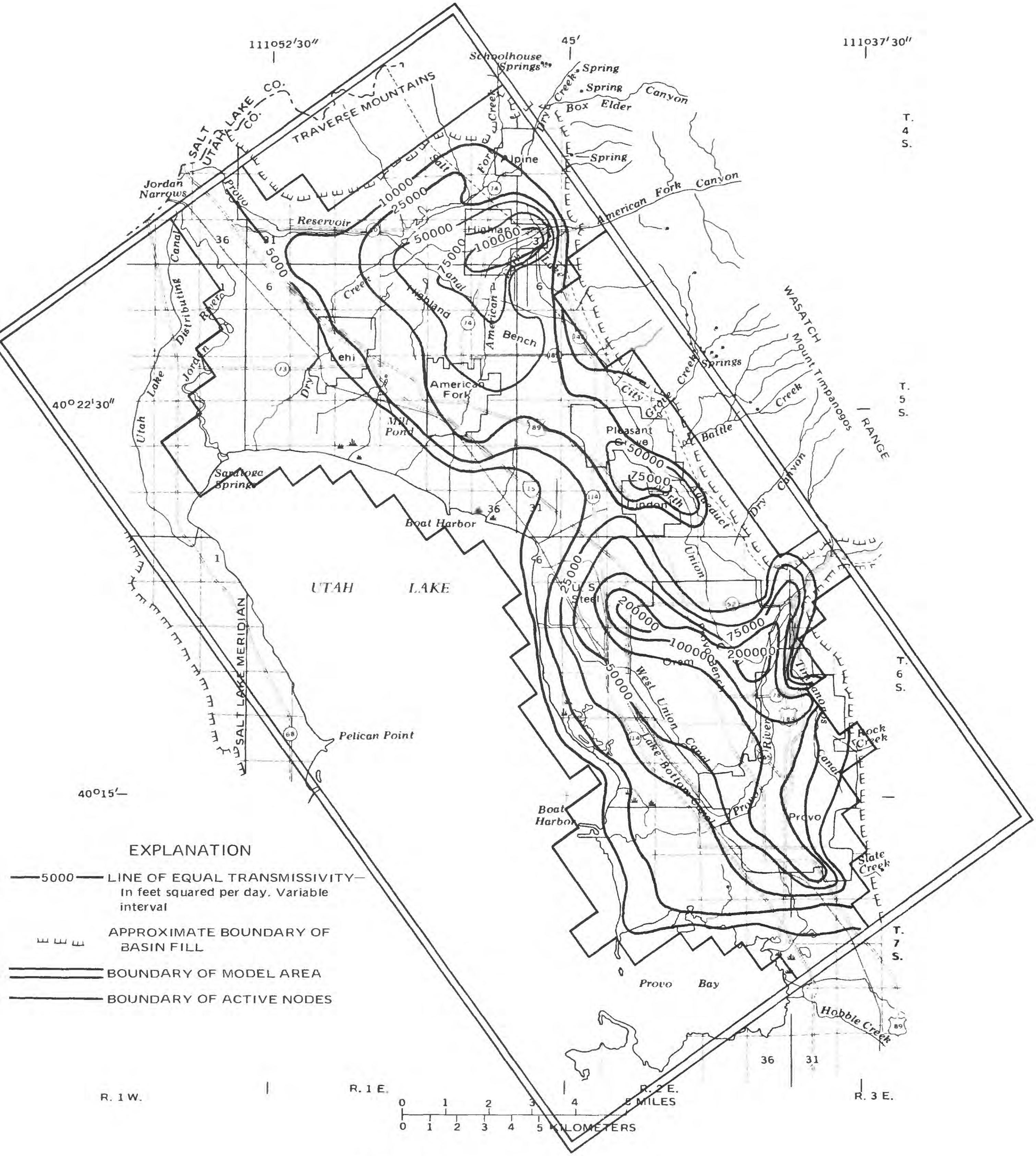


Figure 8.—Transmissivity of layer 5.

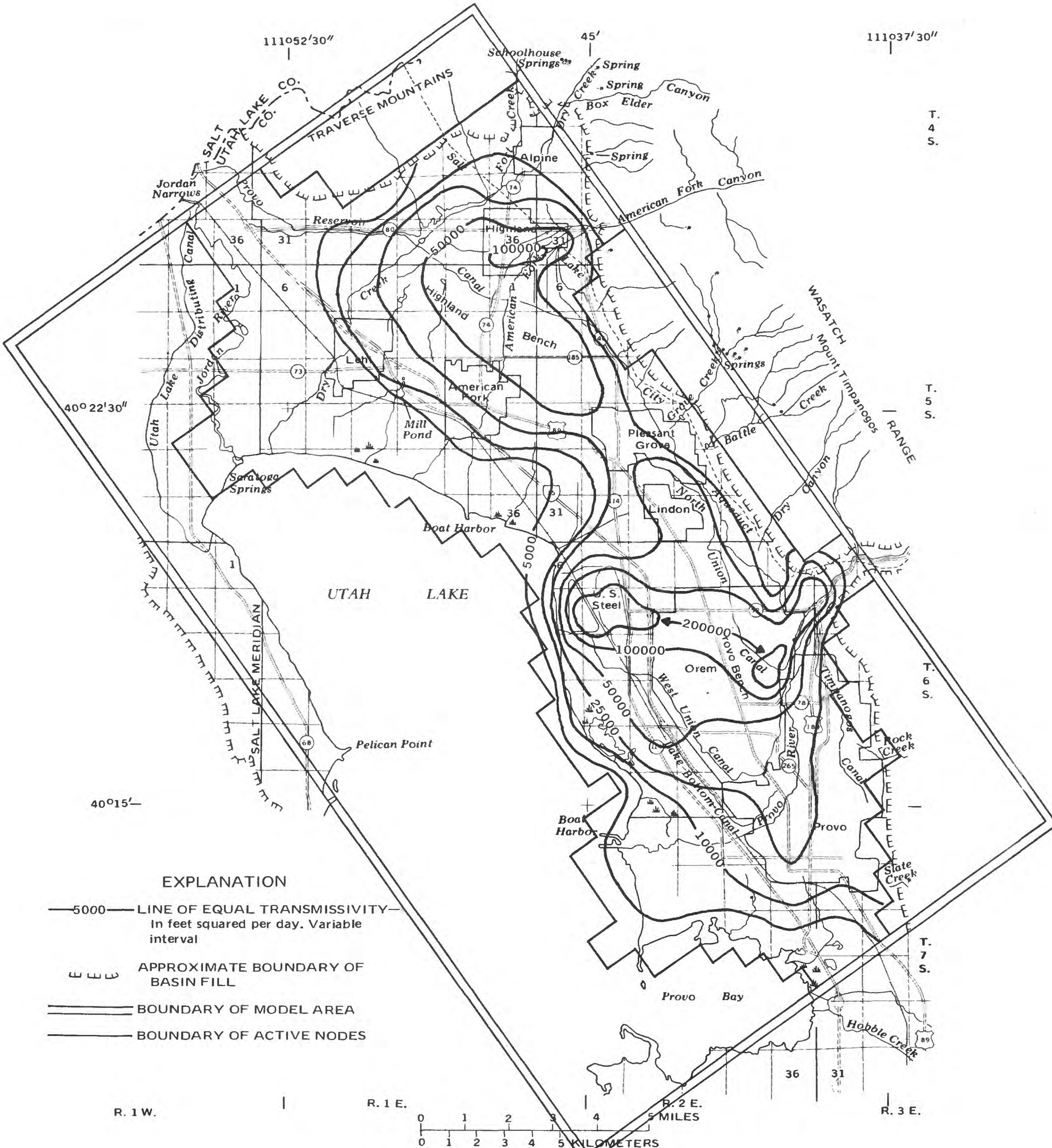


Figure 9.—Transmissivity of layer 7.

The horizontal hydraulic conductivities for the shallow water-table aquifer (layer 1) were estimated from drillers' logs to range from about 1 to 12 feet per day. The *T* is for the confining beds (layers 2, 4, and 6) were estimated from drillers' logs to range from about 1,500 feet squared per day in the basin center to 90,000 feet squared per day in layers 4 and 6 along the mountain fronts.

Vertical hydraulic conductivity was calculated from an aquifer test in the study area to be approximately 1×10^{-3} feet per day (Clark and Appel, 1985, p. 47) using the ratio method (Newman and Witherspoon, 1972 p. 1284). The initial vertical hydraulic conductivities (VCONT) used in the model were estimated by dividing that value by the approximate thickness of the bottom one-half of a model layer plus the top one-half of the underlying layer. VCONT was increased from the initial value in areas where ground water is discharged by springs and into drains and along the mountain front where confining layers are thin or nonexistent. After these adjustments were made, trial-and-error methods were used during steady-state calibration to determine the final values for VCONT, which ranged from about 1×10^{-3} to 1×10^{-1} feet per day.

Storage coefficients were not used during steady-state simulations. The values used for transient-state simulations were generally in the range of about 1×10^{-3} to 6×10^{-6} as determined from aquifer tests for the confined aquifers (Clark and Appel, 1984, table 9). The smallest storage coefficients generally are for the deepest artesian aquifer (layer 7) near the basin center where the sediments generally are fine grained and the thickness of overlying sediments is large. Storage coefficients for the shallow water-table aquifer (layer 1) were unknown and for modeling purposes were assumed to be equal to a specific yield of 1×10^{-1}. The same storage coefficient was used for layer 4 to simulate water-table conditions along the mountain front. Storage coefficients typical of the confined systems were used in layers 5-7 along the mountain front.

Discharge

Discharge from the principal ground-water reservoir was estimated for 1972-81 to be 220,000 acre-feet per year (Clark and Appel, 1985, table 11). That included discharge by drains and springs, wells, evapotranspiration, subsurface inflow into Utah Lake, seepage to the Jordan River, and subsurface outflow through the Jordan Narrows.

Discharge from the principal ground-water reservoir by drains and springs averaged about 100,000 acre-feet per year (Clark and Appel, 1985, table 15). This estimate was based on discharge measurements at 42 sites (Clark and Appel, 1985, p. 76-78). The drain matrix used in the model was an approximation of the area drained at the 42 sites. The discharge was simulated in the model by use of the DRAIN subroutine, which requires information for the altitude of the water in the drain and the conductance of the interface between the aquifer and the drain. The altitudes used for the water in the drains were similar to that of the shallow water table in the same area. Conductance is defined as the product of the hydraulic conductivity and the area of the material in the interface, divided by the thickness of the material.

The conductance used for each node was derived from trial-and-error procedures during steady-state calibration. Conductance was varied until the discharge rates calculated by the model were similar to those estimated from discharge measurements made in the field. The conductance for most nodes then was decreased so that the computed discharge still approximated the measured discharge, but the discharge from these nodes now would fluctuate directly with simulated water-level changes for those nodes. Figure 10 shows a typical relationship between calculated discharge and change in conductance. In this example, any conductance from 0.75 to 75 feet squared per second resulted in a computed discharge that closely approximates the measured discharge of 2.1 cubic feet per second. Therefore, the minimum value in the range (0.75) was used so that simulated discharge in drains could fluctuate with changes in water levels in those nodes caused by stress on the system, such as an increase or decrease in pumpage or recharge. Another reason for using the minimum value of conductance that simulates the measured discharge is that it is a unique number representing the break in slope of calculated discharge (fig. 10). Thus it can be replicated.

At the completion of the steady-state calibration, the discharge into drains and by springs from the principal ground-water reservoir was calculated by the model to be about 108,000 acre-feet per year. Figure 6 shows the location of nodes used for the similation of discharge into drains and by springs.

Of the estimated 4,000 wells in northern Utah Valley, most are flowing wells drilled before 1963. Data from Hunt and others (1953, p. 73) for 1938-40, which included ground-water withdrawals from individual sections, were used to assign initial well-discharge data for the steady-state calibration. Estimates of flowing-well discharge and records of withdrawal for pumped irrigation, public supply, and industry were used for transient-state calibration (1947-82). The quantity of water pumped for irrigation decreased during that period while the water withdrawn for public supply increased. The discharge from all wells within the same node and model layer were combined and simulated as a constant rate for that node.

Discharge upward from the artesian aquifers to layer 1 which was eventually discharged by evapotranspiration was estimated to be 8,000 acre-feet per year (Clark and Appel, 1985, table 11). Evapotranspiration from layer 1 was simulated by the model using a head-dependent option, which assumes a linear change between a maximum evapotranspiration rate when the water level is at or above land surface to no evapotranspiration when the water level is at or below the specified extinction depth. After steady-state calibration, evapotranspiration was calculated by the model to be about 9,200 acre-feet per year. The nodes where evapotranspiration was simulated are shown in figure 6.

Subsurface inflow to Utah Lake by subsurface springs was estimated to be between 25,000 and 36,000 acre-feet per year (Cordova and Subitzky, 1965, p. 19) and to average 30,000 acre-feet per year. The subsurface inflow probably varies with the hydrostatic head in the artesian aquifers, and as the estimate was made when lake and ground-water levels were low, it is probable that the estimate is too small. In addition, approximately 7,000 acre-feet of ground-water annually enters the lake by diffuse seepage through lake-bottom sediments from the artesian aquifers under the lake (Clark and Appel, 1985, p.

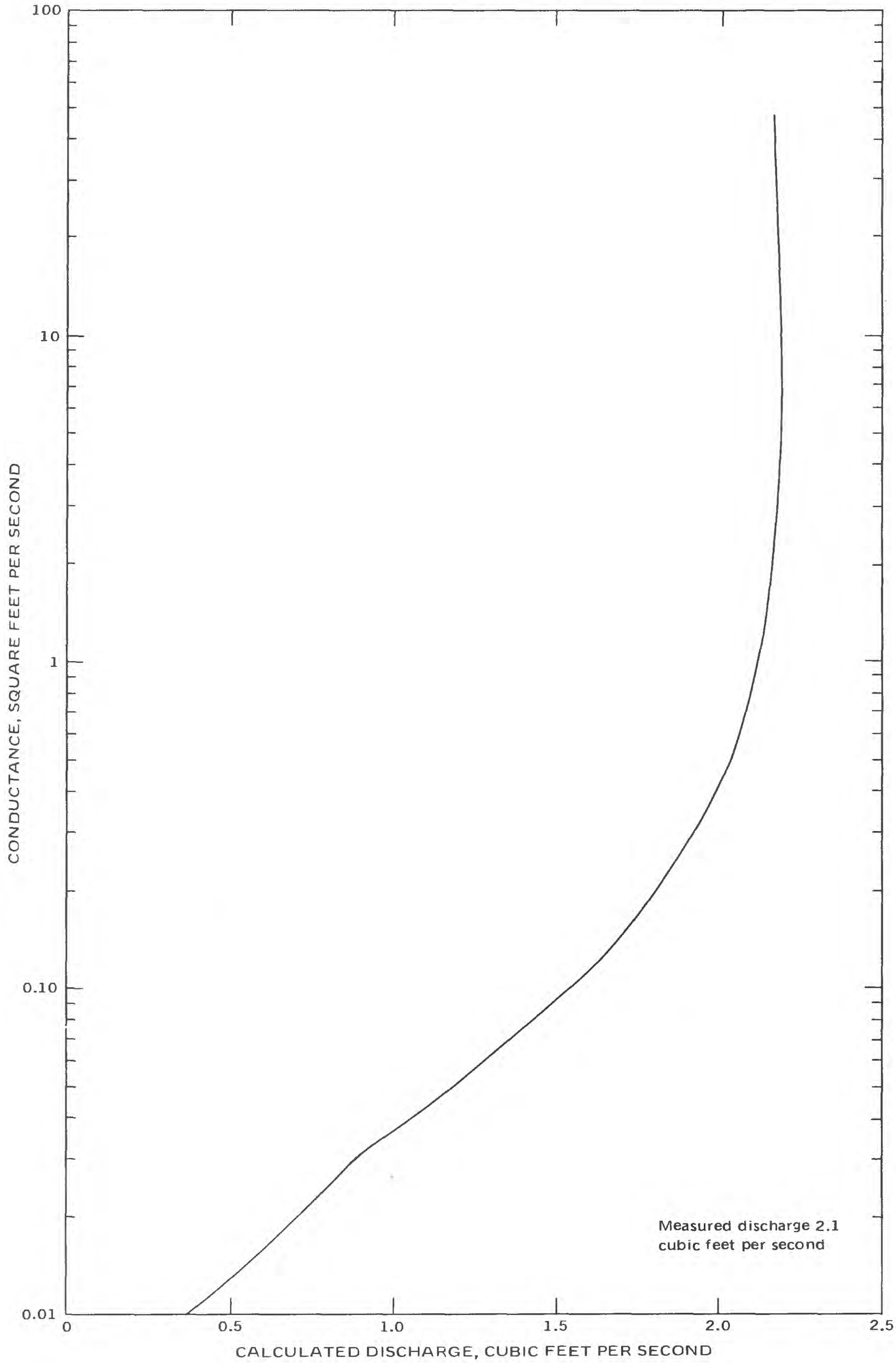


Figure 10.—Variations in discharge from drains with changes in conductance calculated by the digital-computer model.

79). The total subsurface inflow to Utah Lake, therefore is, estimated to be about 37,000 acre-feet per year. Discharge to the lake was simulated in the model by means of constant-head nodes, as explained in the section on "Boundary Conditions". Discharge to Utah Lake was calculated by the model during steady-state calibration to be about 40,000 acre-feet per year.

Seepage to the Jordan River from the principal ground-water reservoir was estimated by Clark and Appel (1985, table 15) to be between 3,500 and 5,600 acre-feet per year. This seepage was simulated in the model with the use of constant-head nodes, and after steady-state calibration, the calculated discharge to the same area was about 6,000 acre-feet per year.

Subsurface outflow from the principal ground-water reservoir through the Jordan Narrows was estimated to be at least 2,000 acre-feet per year by Clark and Appel (1985, p. 83). This outflow was simulated in the model by means of constant-head nodes and was calculated to be about 7,000 acre-feet per year. However, it was not possible to match steady-state water levels in the area; therefore, the value of 7,000 acre-feet per year is not considered to be reasonable.

MODEL CALIBRATION

The model was calibrated to steady-state conditions, which were assumed to exist in 1947. Transient-state simulations were then made using the final water levels obtained from the steady-state calibration and ground-water withdrawals and water-level fluctuations from 1947-83.

Steady-State Calibration

The calibration of the model to steady-state conditions involved comparison of water levels for the three artesian aquifers, as computed by the model, with water levels that were measured primarily during April 1947. In areas where there were few or no wells in the principal ground-water reservoir in 1947, such as on the benches, later data were used as initial water levels. Ground-water withdrawals prior to 1947 primarily were from flowing wells in the basin center. The only annual data available is for 1938-40, when the total well discharge ranged from about 27,000 to 32,000 acre-feet (Hunt and others, 1953, p. 73). Water levels in the area generally rose from 1935-46 when most were at their highest recorded levels. They then remained relatively stable until 1952 (Clark and Appel, 1985, fig. 38). It was assumed for the purpose of the model that steady-state conditions existed during 1947 and that ground-water withdrawals prior to that time had virtually no effect on water-level fluctuations. To test this assumption after steady-state calibration was complete, the total discharge from wells (about 26,000 acre-feet per year) was decreased substantially, and little or no change in water levels resulted.

During steady-state calibration, some values were adjusted on a trial-and-error basis to obtain the best results. The values that were adjusted most were vertical hydraulic conductivities (VCONT) for all the layers and head values for the constant-head recharge nodes along the mountain fronts. The transmissivity matrix for the aquifers and the 1947 water levels were not adjusted because they included the most reliable data.

Computed water levels were compared to measured water levels for 1947 in the three artesian aquifers. The comparison included 63 water levels in layer 3, 48 in layer 5, and 9 in layer 7. Figures 11 and 12 show comparisons for 1947 between water-level contours based on measurements in wells and contours generated by the model during the steady-state calibration. Most of the water levels generated by the model are within 5 feet of those measured in wells; however, in isolated areas, the difference may be as much as 10 feet.

At the conclusion of the steady-state calibration, the flow rates for the constant-head nodes along the mountain front, as calculated by the model, were input as recharge rates and the constant-head nodes were eliminated. These rates, which represent ground-water recharge by subsurface inflow, total about 100,000 acre-feet per year. Figure 13 shows the primary recharge area for the principal ground-water reservoir and the constant recharge rate for all sources of recharge that were applied to the individual nodes in the area at the end of steady-state calibration. The total recharge from these nodes, including the recharge as calculated by the model from constant-head nodes, is 190,000 acre-feet per year.

Transmissivity *T*, and vertical hydraulic conductivity (VCONT) were varied and the results compared in order to test their sensitivity in the model. The sensitivity analysis involved changing one hydraulic property and comparing the resultant water levels to water levels from the calibrated steady-state model. The analysis included model runs that used constant-head nodes at the eastern boundary resulting in a variable total recharge, or recharge nodes at the eastern boundary with total recharge set at 190,000 acre-feet per year. The amount and direction of water-level change varies depending on the model layer, the location in the model grid, and the factor used (table 1). When constant-head nodes were used and T and VCONT were changed by the same factor, the model generally computed equal but opposite water-level changes, ranging from -35 to +35 feet in the discharge area and -20 to +32 feet in the recharge area. When constant-head nodes were replaced at the eastern boundary with recharge nodes the recharge remained constant, and the range of water-level changes was much larger (-35 to +150 feet), even though the factors used were smaller (table 1).

Transient-State Calibration

Transient-state calibration primarily consisted of calibration for nine pumping periods using well discharge data for 1947-82 and water-level data for 1947-83. The last two pumping periods, 1981 and 1982, are discussed separately because they represented periods of hydrologic extremes. Additional calibration was done using data from an aquifer test.

Calibration Using Well Discharge Data for 1947-80 and Water-Level Data for 1947-81

An initial transient-state calibration was done by simulating discharge from wells for 1947-80 and comparing the computed water levels with water levels measured during 1947-81. Water-level changes were computed for seven pumping periods during the 34-year period, starting from steady-state conditions. The pumping periods, which were 1947-50, 1951-55, 1956-62, 1963-65, 1966-73, 1974-77, and 1978-80, represent intervals of time when total discharge from wells was fairly constant.

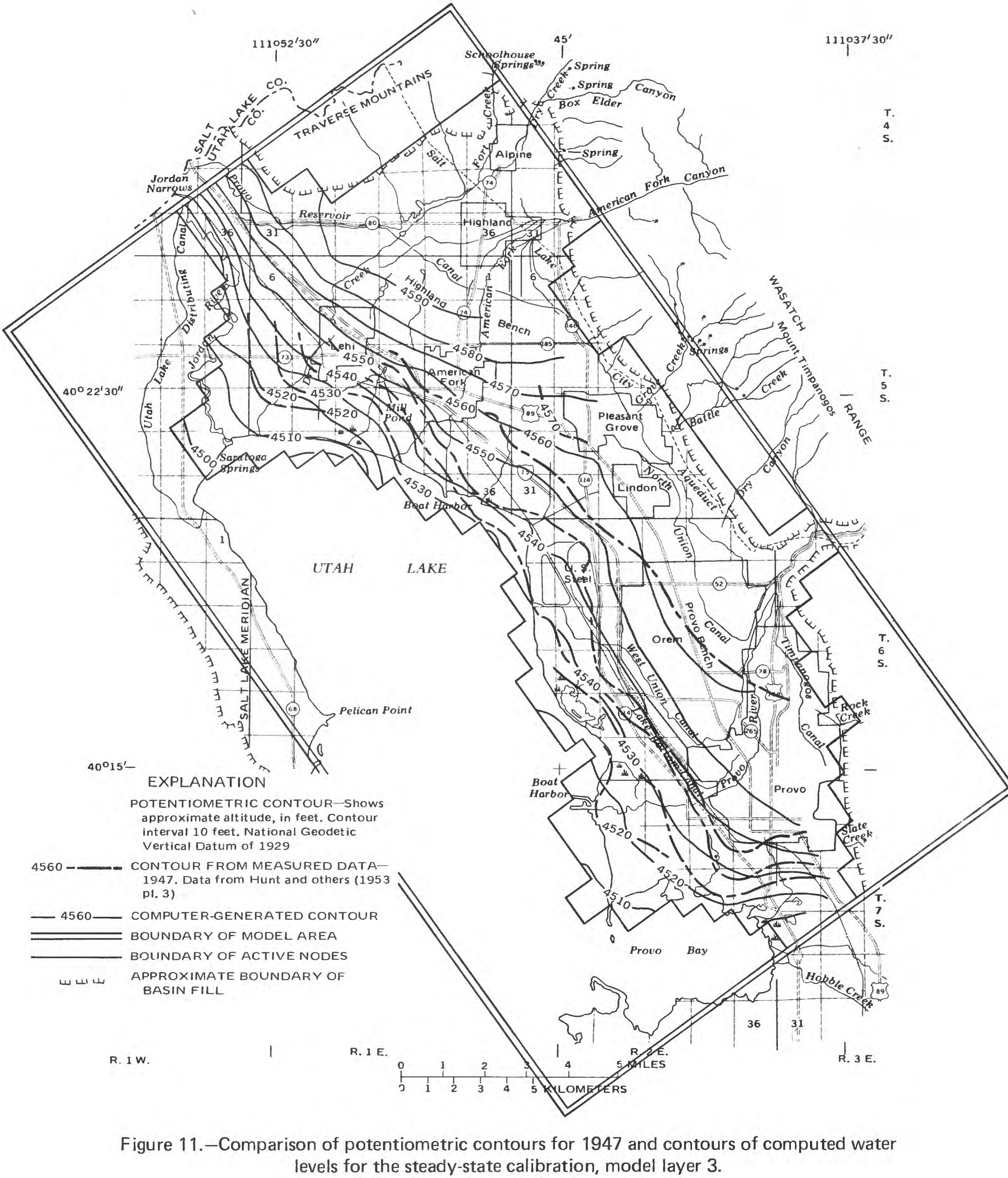


Figure 11.—Comparison of potentiometric contours for 1947 and contours of computed water levels for the steady-state calibration, model layer 3.

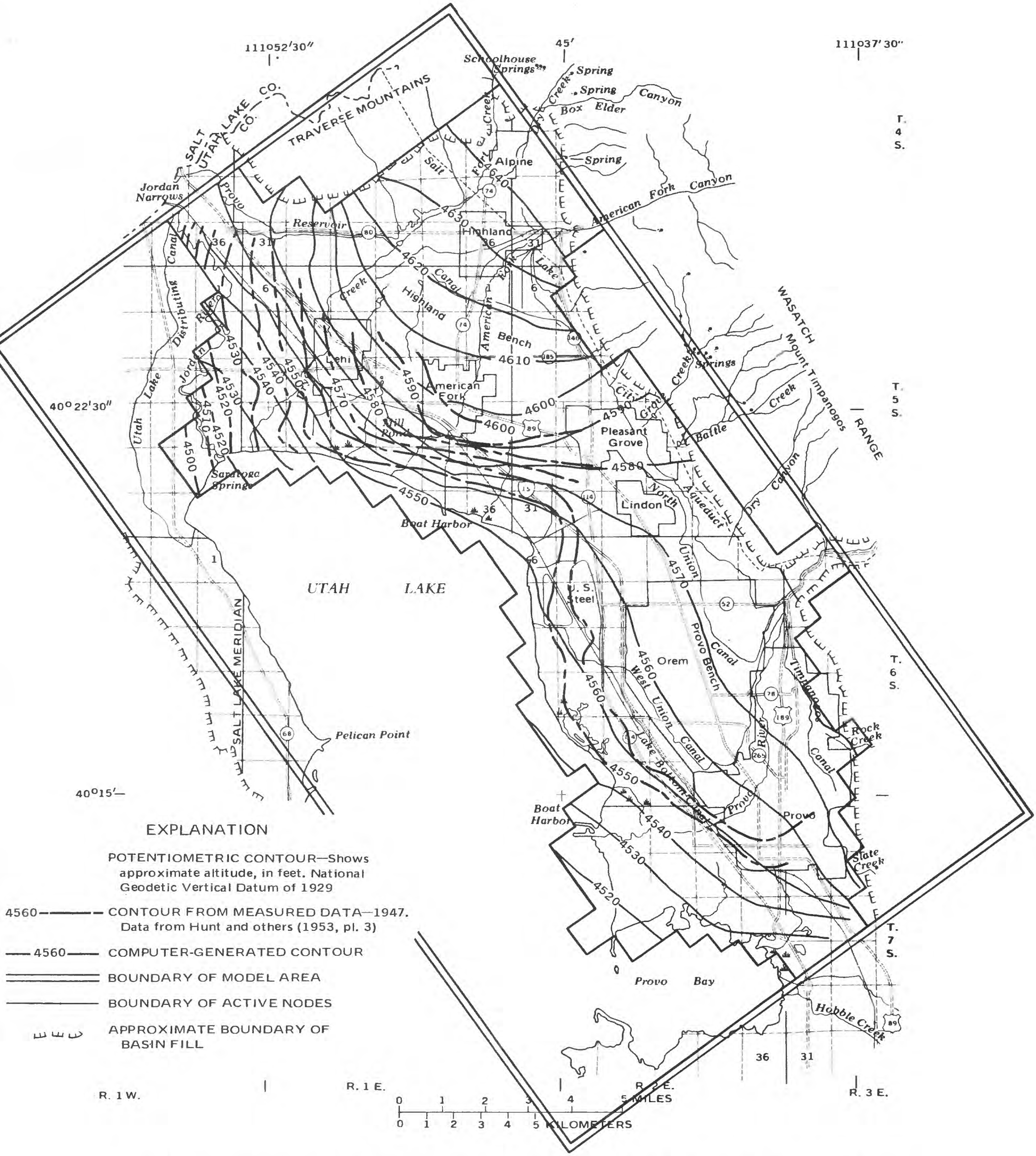


Figure 12.—Comparison of potentiometric contours for 1947 and contours of computed water levels for the steady-state calibration, model layer 5.

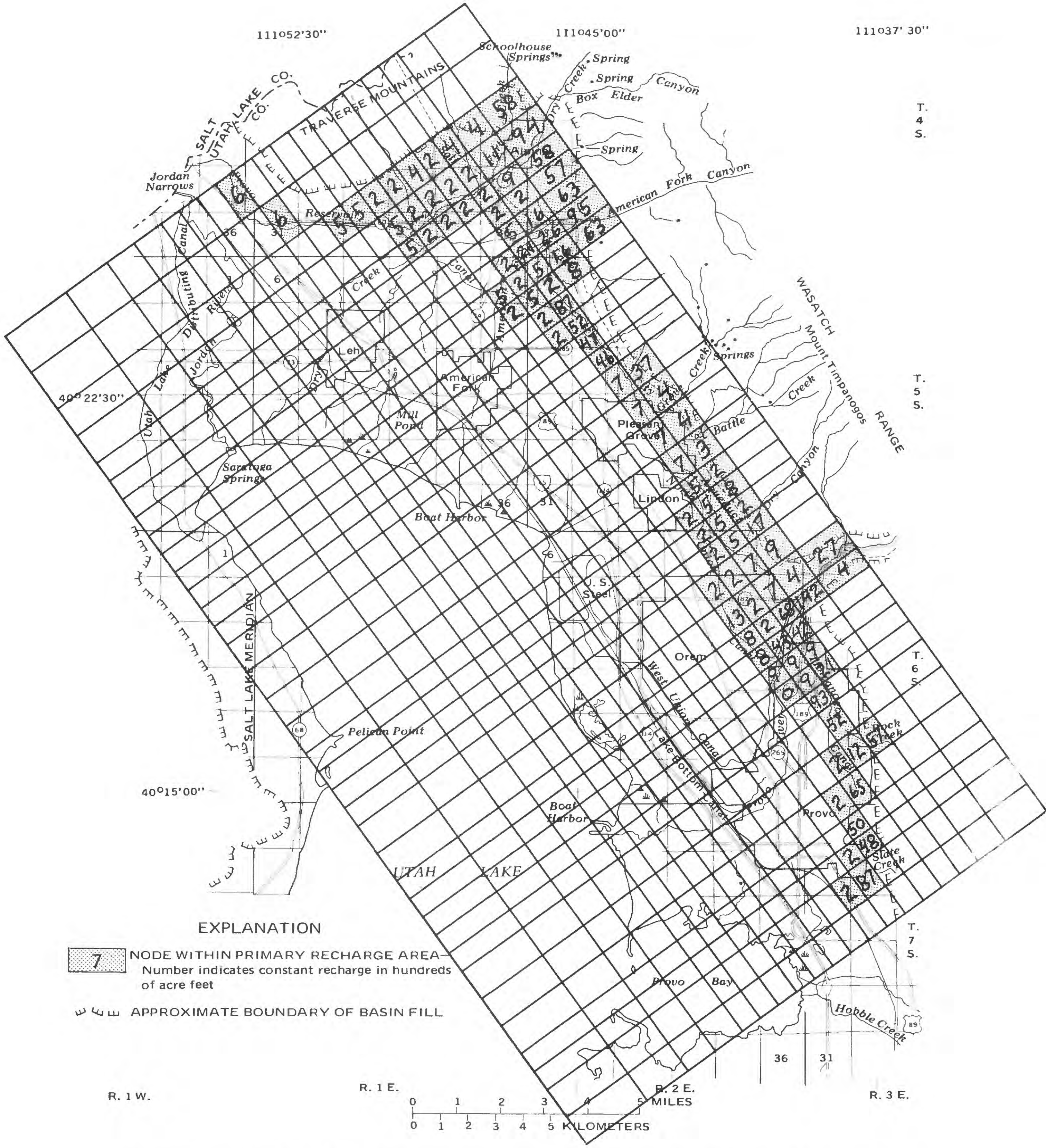


Figure 13.—Location of the primary recharge area in northern Utah Valley and the quantity of constant recharge at the end of steady-state calibration.

Table 1.--Sensitivity analysis for steady-state calibration

[General locations of discharge and recharge areas are shown in figure 3. Hydraulic property: T,transmissivity; VCONT, vertical hydraulic conductivity. Factor: /, divided by; X, multiplied by]

Hydraulic property	Factor	Total recharge (ft3/s)	Water-level change from calibrated steady-state model (ft): Layer 3, Discharge area	Layer 3, Recharge area	Layers 5 and 7, Discharge area	Layers 5 and 7, Recharge area
			Eastern boundary includes constant-head nodes			
T	X 1.5	313	+2 to +4	-1 to -5	+5 to +10	0 to -3
VCONT	X 1.5	225	-2 to -3	+1 to +4	-4 to -10	0 to -1
T	/ 1.5	212	-3 to -5	+1 to +4	-5 to -10	0 to +3
VCONT	/ 1.5	224	0 to +2	-1 to -4	+4 to +9	0 to +2
T	X 2.5	407	+4 to +7	-3 to -10	+10 to +20	+1 to -3
VCONT	X 2.5	347	-4 to -7	+1 to +9	-10 to -20	-1 to +3
T	/ 2.5	173	-7 to -13	+3 to +10	-10 to -20	0 to +14
VCONT	/ 2.5	182	+2 to +4	-2 to -11	+10 to +20	0 to +5
T	X 5.0	547	+5 to +10	-7 to -20	+15 to +30	+1 to -7
VCONT	X 5.0	425	-5 to -15	+1 to +14	-15 to -30	-1 to -5
T	/ 5.0	147	-13 to -20	+7 to +15	-15 to -35	+1 to +32
VCONT	/ 5.0	161	+3 to +6	-5 to -20	+16 to +35	+1 to +10
			Eastern boundary using recharge nodes			
T	X 1.5	262	0 to -5	-10 to -20	0 to -12	-10 to -35
VCONT	X 1.5	262	-4 to -8	-4 to -8	-7 to -17	-7 to -17
T	/ 1.5	262	0 to +6	+15 to +30	0 to +15	+15 to +40
VCONT	/ 1.5	262	+5 to +10	+5 to +10	+10 to +20	+10 to +20
T	[1]X 2.0	--	--	--	--	--
VCONT	X 2.0	262	-6 to -13	-6 to -13	-10 to -24	-10 to -24
T	/ 2.0	262	-1 to -11	+25 to +55	-3 to -33	+30 to +80
VCONT	[1]/ 2.0	--	--	--	--	--
T	/ 3.0	262	-2 to -20	+48 to +100	-5 to +50	+50 to +150
VCONT	X 3.0	262	-9 to -16	-9 to -16	-17 to -34	-17 to -34

[1] Failed to converge.

Most of the records for well discharge were compiled on the basis of use or the method of withdrawal--pumping or flowing. Records for withdrawals are fairly complete for public supply, industry, and pumped irrigation after 1962. Prior to 1963, records were less complete but withdrawal for these uses was not a significant percentage of the total well discharge. For the purpose of this report, the term "flowing wells" does not include those flowing wells used for public supply or industry for which there are records of discharge. Records for flowing-well discharge are much less complete or accurate, but this type of discharge is the major part of the total discharge for most pumping periods.

The matrix used to simulate flowing-well discharge during steady-state calibration also was used for the first three pumping periods because of a lack of additional data. Locations and discharge were taken from Hunt and others (1953, p. 73). Data from drillers' logs for about 1,200 wells greater than 2 inches in diameter in the flowing-well area were used to construct a new data base for the pumping periods after 1962. This new data base included a better definition of where and from which aquifer flowing wells were discharging. The total discharge from the flowing wells was estimated to average about 30,000 acre-feet per year from 1963-82 (Clark and Appel, 1985, table 13) and it was distributed uniformly among the 1,200 wells and then totaled for each node according to the grid location and the layer in the model.

The annual discharge from flowing wells varied between pumping periods depending primarily on the recharge to the ground-water reservoir during the pumping period. An estimate of the total discharge from flowing wells for each pumping period was made by correlating discharge from pumped irrigation wells with total surface-water inflow to the study area. When total surface-water inflow (the major source of irrigation water) is below average, the quantity of pumped irrigation water is above average. This relationship also is assumed to be true for flowing wells, most of which are used for irrigation. Flowing-well discharge, therefore, was varied according to pumped irrigation discharge for the different pumping periods.

Figure 14 shows the location of the nodes for flowing wells and wells for pumped irrigation, industry, and public supply. All the wells were used in at least one pumping period, but they probably were not used during all pumping periods. The discharge used in each of the pumping periods during the transient-state calibration is shown in table 2 by method of withdrawal or use.

During the transient-state calibration, it was evident that changes in discharge made from one pumping period to the next were not resulting in the measured water-level changes which ranged from declines of 20 feet to rises of 20 feet during this time. An attempt was made, therefore, to correlate measured water-level changes with potential changes in recharge.

Variations of annual recharge were assumed to be proportional to changes in total surface inflow to northern Utah Valley. The initial rate of recharge obtained from the steady-state calibration was 190,000 acre-feet per year. This was multiplied by a proportionality constant of about one-half of the percentage change from the average surface-water inflow during a given time period. For example, if surface-water inflow for a given time period was

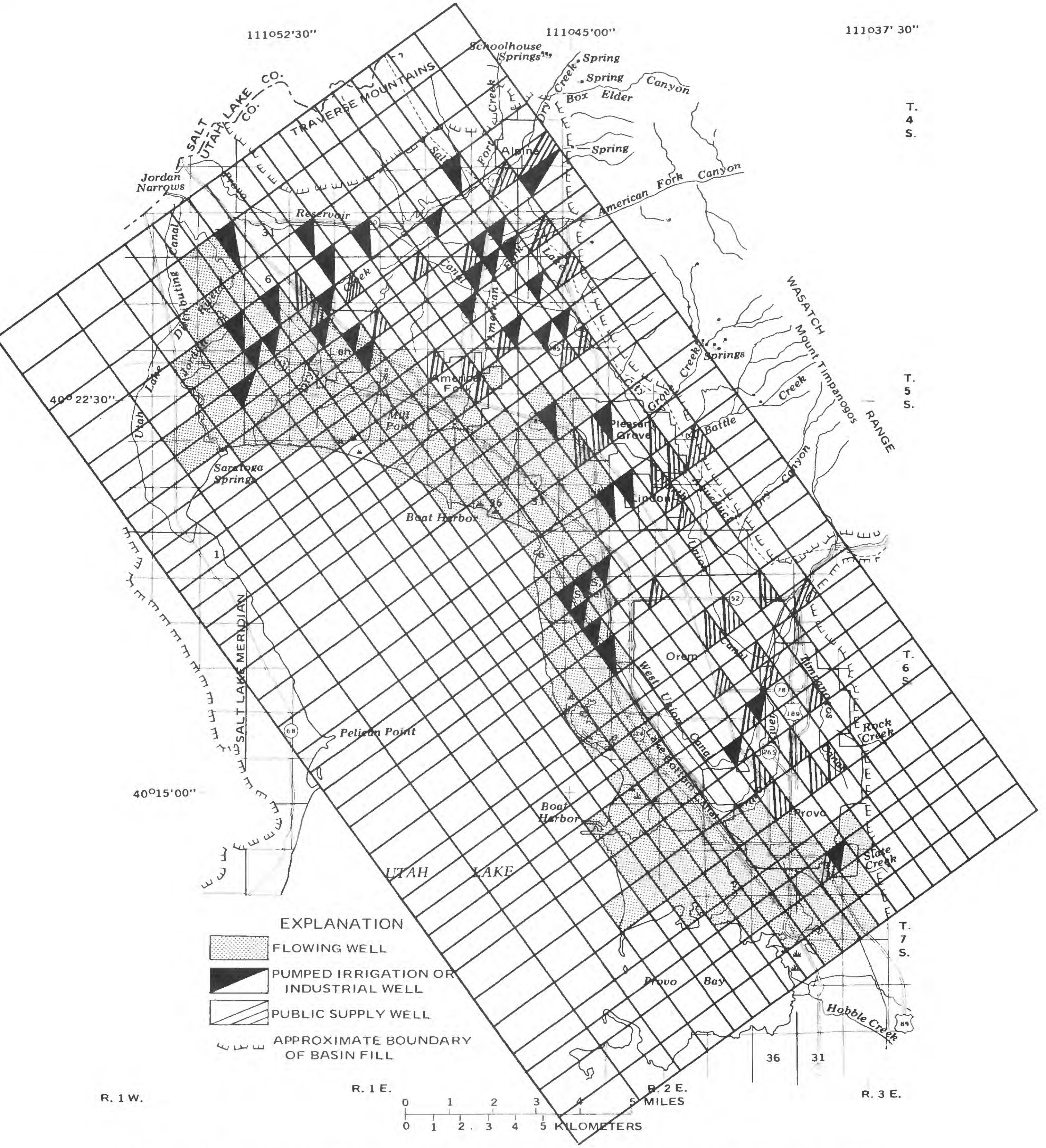


Figure 14.—Location of nodes for flowing wells, and wells for pumped irrigation, public supply, or industrial use.

Table 2.--Pumping periods and well discharge used in the transient-state calibration for 1947-83

Pumping period	Discharge (acre-feet per year)		
	Flowing wells	Pumped irrigation, public supply, and industry wells	Total
1947-50	31,200	5,800	37,000
1951-55	34,900	17,600	52,500
1956-62	35,300	14,400	49,700
1963-65	25,000	19,500	44,500
1966-73	27,200	22,300	49,500
1974-77	35,500	35,800	71,300
1978-80	22,000	30,800	52,800
1981	25,900	33,500	59,400
1982	23,200	23,700	46,900

20 percent above average, then recharge for that time period was assumed to be 10 percent above the initial rate. The proportionality constant was determined by varying the constant during repeated runs of the model until the difference between computed and measured water levels were minimized. Best results were obtained when only the last few years of a pumping period were used to calculate change from average surface-water inflow. Table 3 shows the total recharge rates used during transient-state calibration. Recharge to the principal ground-water reservoir may change significantly from one year to the next, and such changes are a major cause for variations of water levels in northern Utah Valley.

The components for the total ground-water budget used for the steady-state and transient-state calibrations are shown in figure 15, which indicates the changes in total recharge and the various components of discharge for 1947-82. A comparison of the ground-water budget as calculated by the model and the budget calculated from field data (Clark and Appel, 1985) is shown in table 4.

Figure 16 shows the measured and computed water-level changes for 16 observation wells with data for some of or all the pumping periods. For most of the observation wells, the computed water levels are close to the measured levels. At wells where the computed levels did not match the measured levels, the magnitude of the water-level changes from one pumping period to the next were usually about the same. In general, the computed levels were closer to the measured levels in the northern half of the area than in the southern half near Utah Lake.

At the completion of the transient-state calibration, contour maps of the potentiometric surface were constructed for the three artesian aquifers (model layers 3, 5, and 7) comparing the computed water levels with the water levels measured by Clark and Appel (1985, figs. 23, 24, and 25) during 1981. (See figures 17, 18, and 19.) In most parts of the study area, the two sets of contours show a reasonably close approximation.

Calibration Using Withdrawal Data for 1981-82 and Water-Level Data for 1982-83

The last two pumping periods of the transient-state calibration were made using data from 1981-82, which were years of hydrologic extremes. The surface inflow into northern Utah Valley was about 20 percent less than average during 1981 and more than 50 percent greater than average during 1982. This resulted in a large difference in recharge to the ground-water reservoir, which in turn resulted in large changes in water levels during a short period of time. This provided a opportunity to see if the model could simulate such marked changes accurately.

The recharge rates used for the 1982 and 1983 simulations are shown in table 3. Maps comparing the computed water-level changes for 1981 to 1982 and 1982 to 1983 with the corresponding measured changes for the same time periods (Holmes and others, 1982, p. 28-36; Appel and others, 1983, p. 32-42) are shown in figures 20-23.

Table 3.--Pumping periods and total recharge rates used in the transient-state calibration for 1947-83

Pumping period	Recharge rate (acre-feet per year)
1947-50	197,500
1951-55	172,800
1956-62	169,000
1963-65	216,500
1966-73	208,900
1974-77	169,000
1978-80	214,600
1981	172,800
1982	237,300

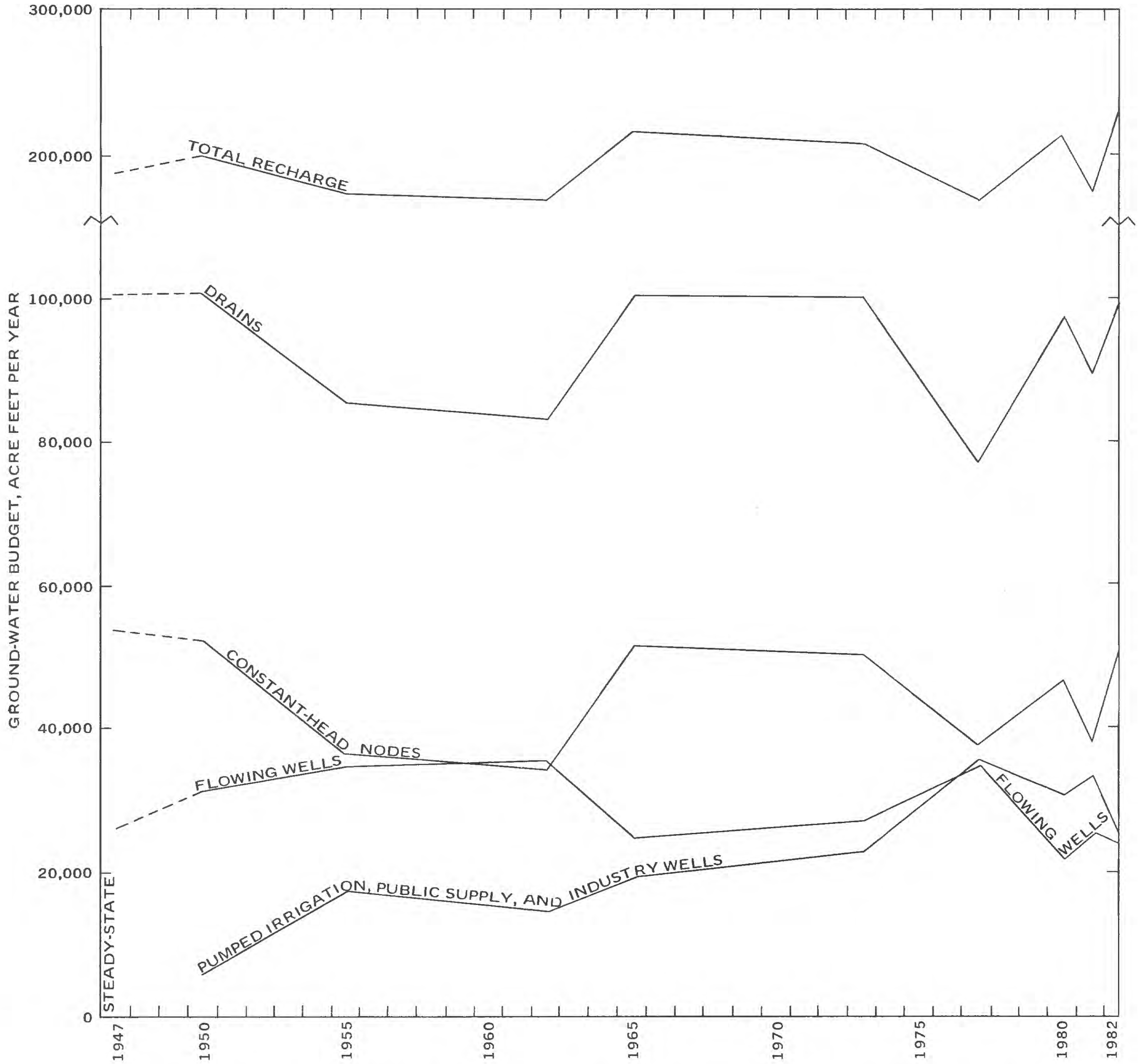


Figure 15.—Ground-water budget used for steady-state and transient-state calibrations.

Table 4.--Ground-water budget for steady-state and transient-state calibration, computed by the digital model compared to budget based on field data, in acre-feet per year

[Based on field data: Data are from Clark and Appel (1984, tables 5, 11, 13, and 15)]

Budget element	Steady-state (1947)	Transient-state (end of 1978-80 pumping period)	Based on field data
Recharge			
Seepage from waterways, irrigated fields, lawns and gardens, and direct precipitation	82,000	--	88,000
Subsurface inflow	108,000	--	[1]104,000
Total	190,000	214,600	192,000
Discharge			
Wells			
Pumped irrigation, public supply, and industry	--	30,800	31,400
Flowing irrigation stock, and domestic	26,000	22,000	36,000
Drains and springs	104,000	98,000	100,000
Discharge to Utah Lake including diffuse seepage	37,400	38,000	37,000
Outflow to Jordan River	6,000	4,300	4,600
Evapotranspiration	9,200	9,200	8,000
Outflow through Jordan Narrows	7,000	4,800	2,000
Storage	--	7,500	--
Total	190,000	214,600	220,000

[1] Does not include an additional 8,000 acre-feet calculated for areas outside the model boundary.

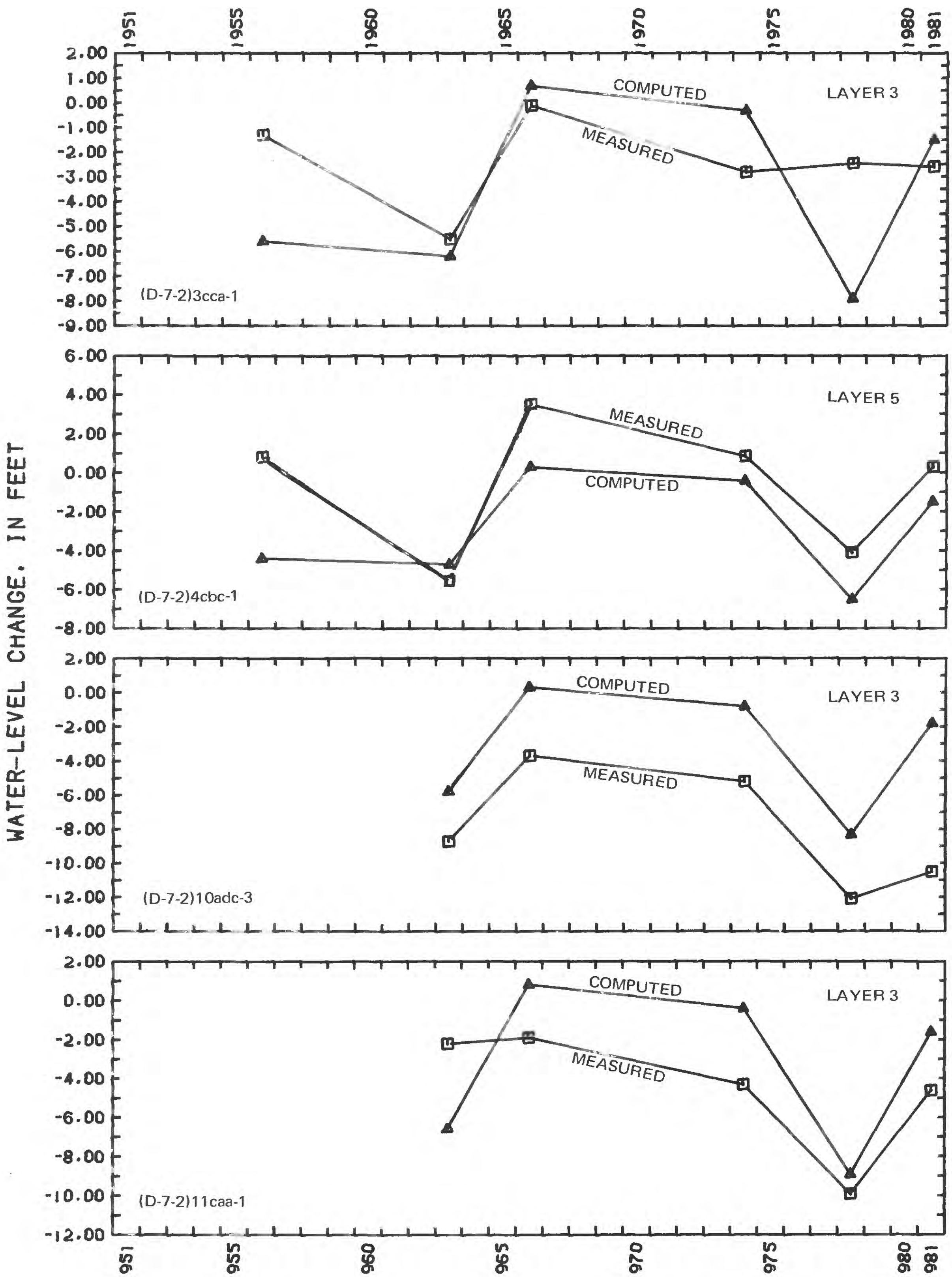


Figure 16.—Measured and computed water-level changes from 1951-81 for 16 observation wells.—Continued.

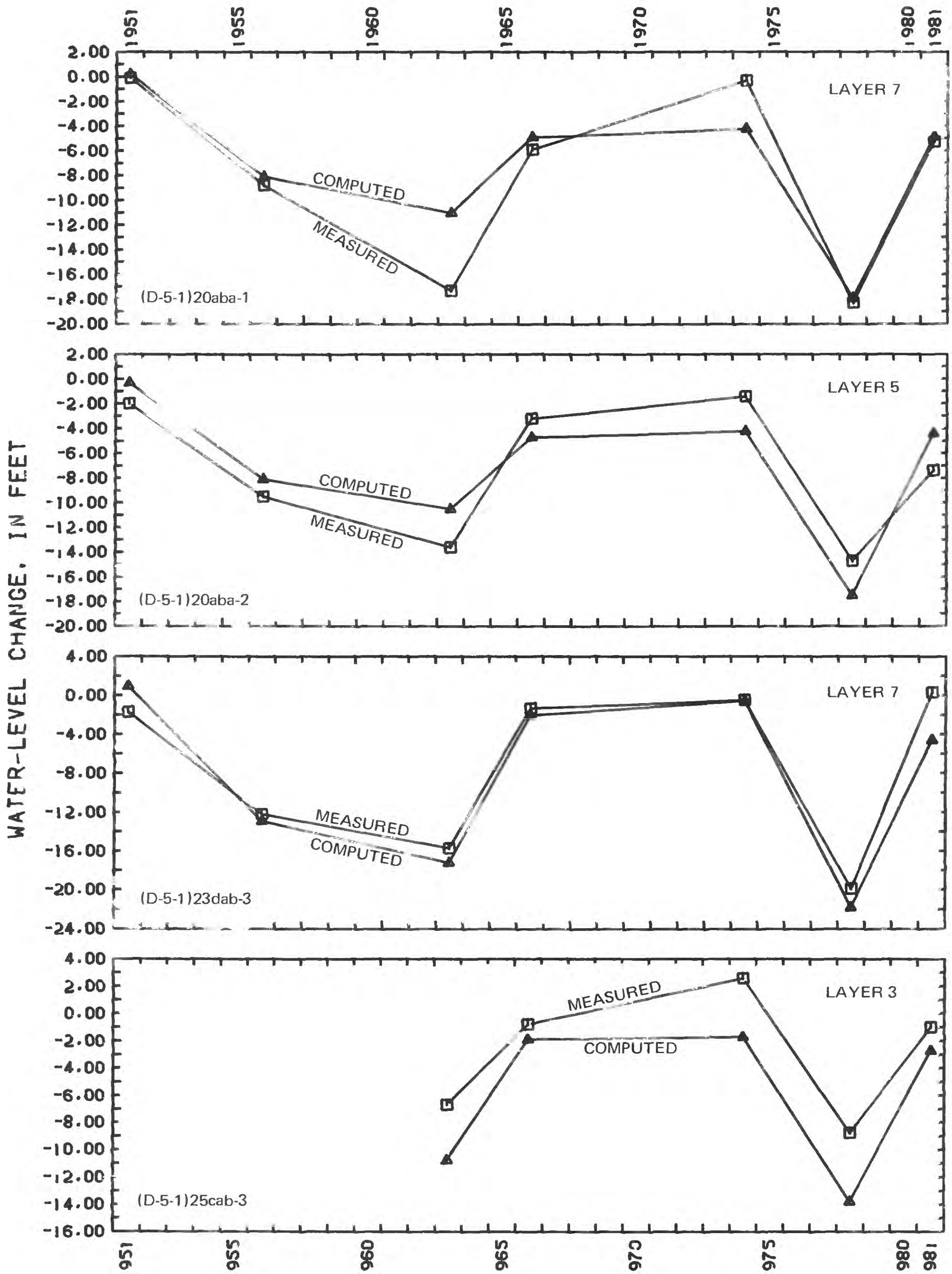


Figure 16.—Measured and computed water-level changes from 1951-81 for 16 observation wells.—Continued.

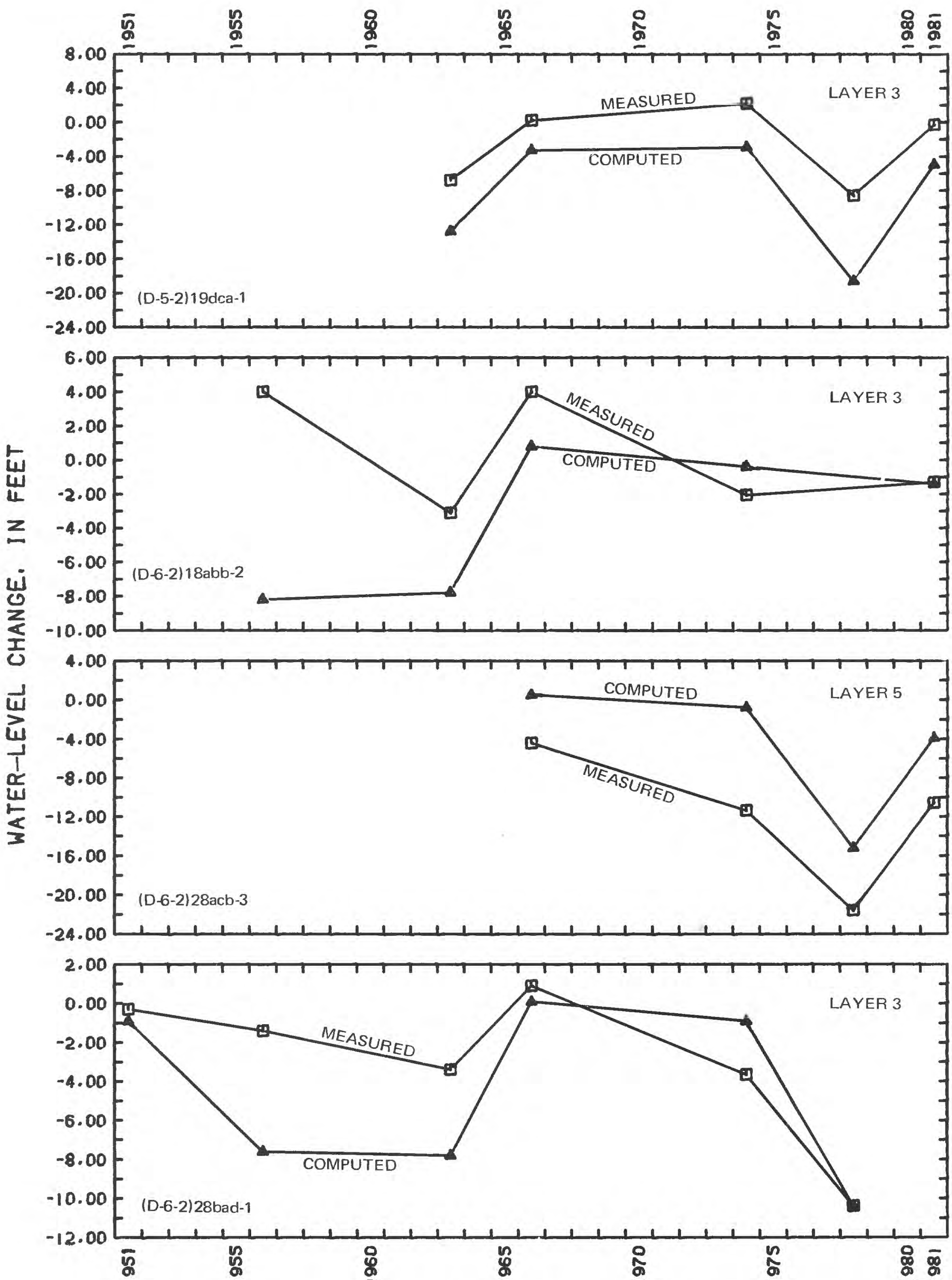


Figure 16.—Measured and computed water-level changes from 1951-81 for 16 observation wells.—Continued.

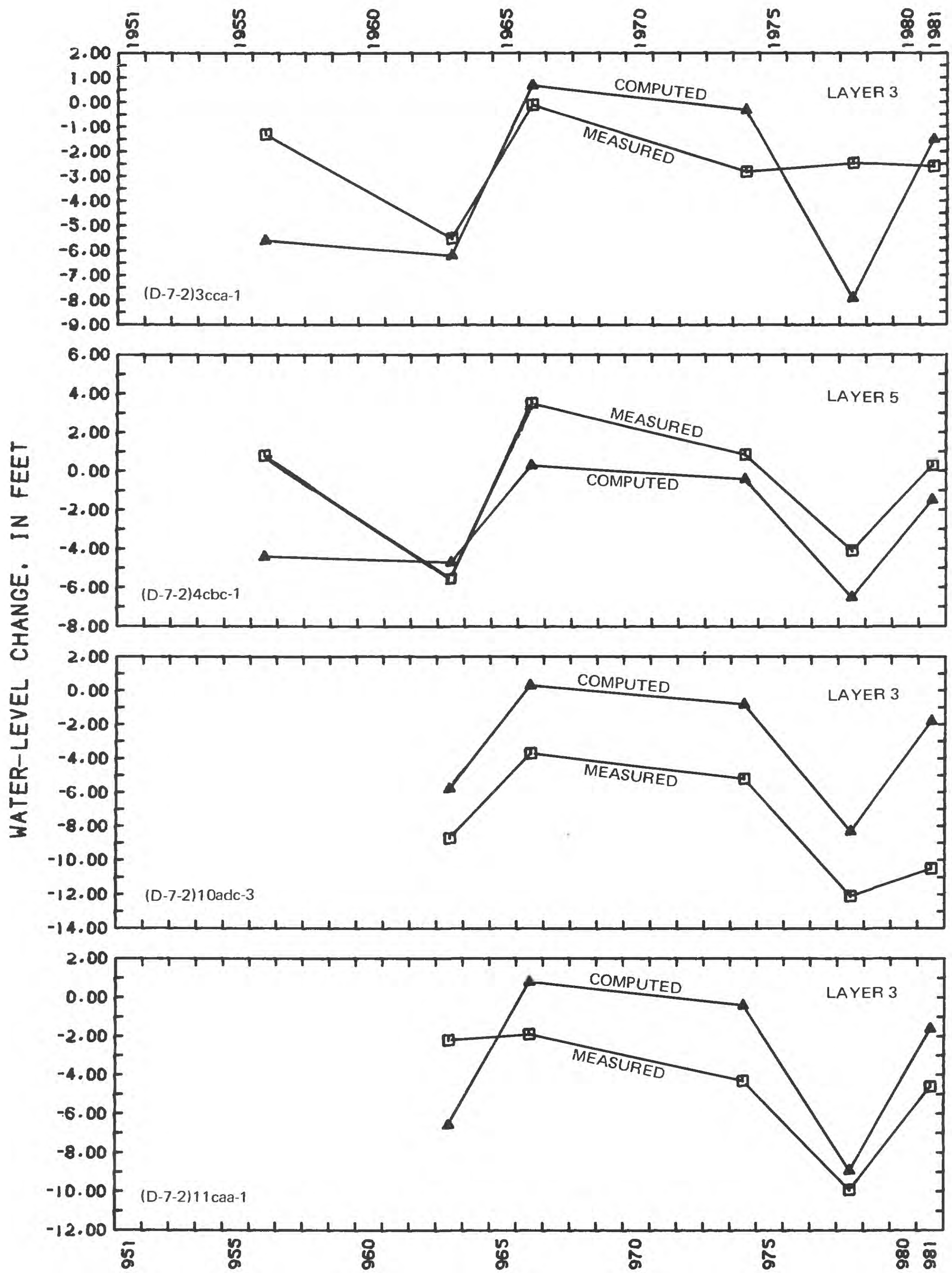


Figure 16.—Measured and computed water-level changes from 1951-81 for 16 observation wells.—Continued.

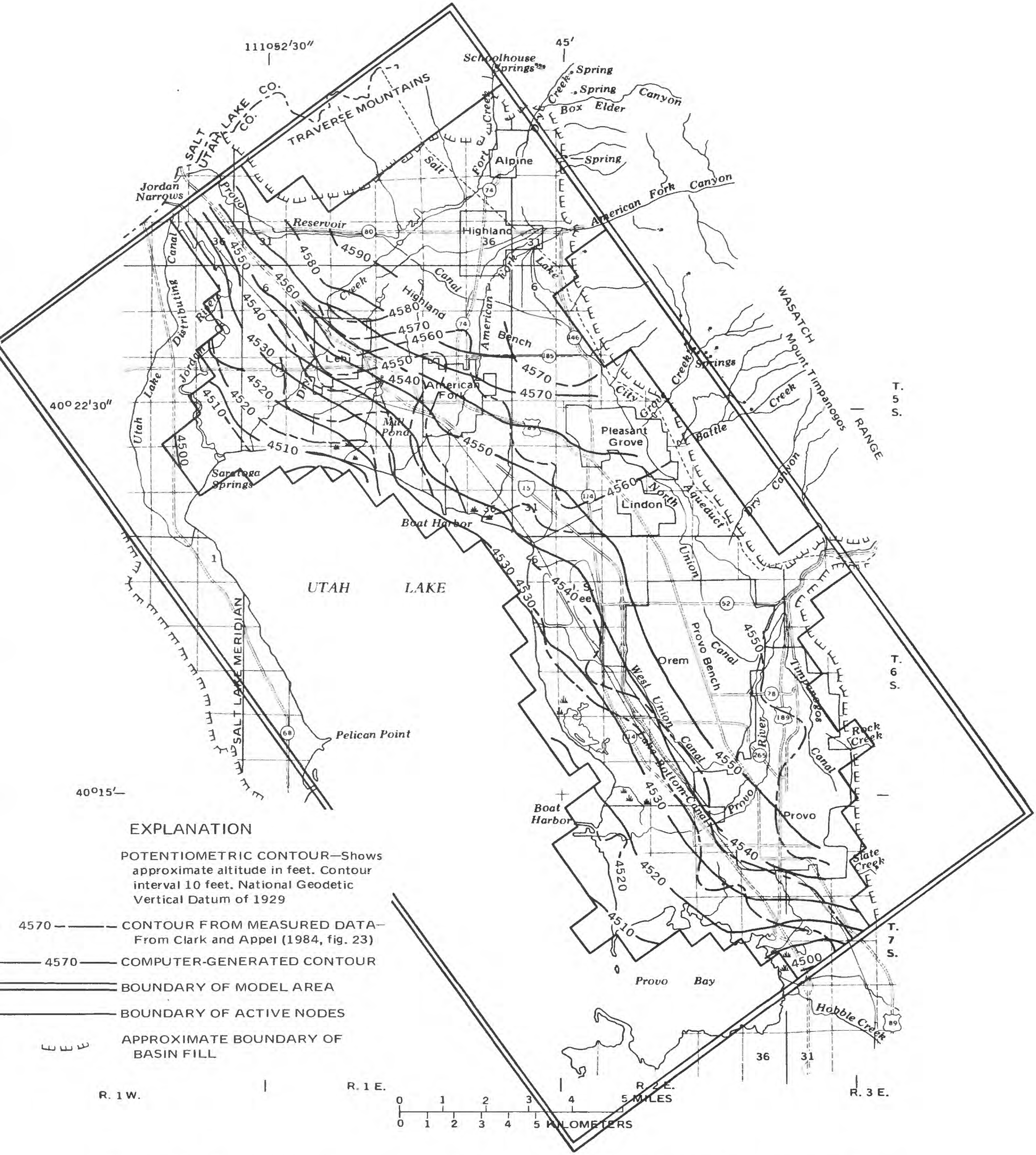


Figure 17.—Comparison of potentiometric contours for 1981 and contours of computed water levels, model layer 3.

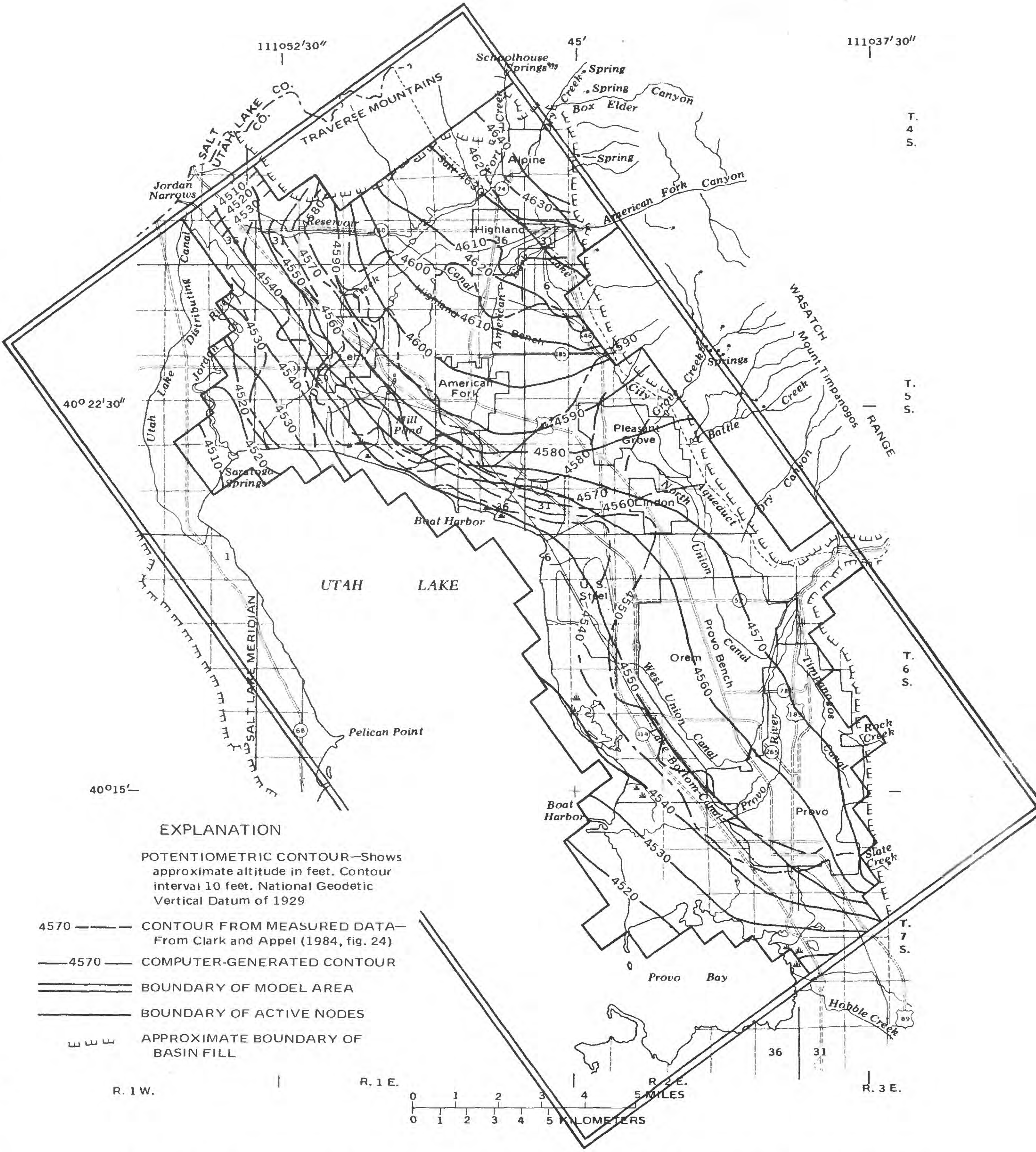


Figure 18.—Comparison of potentiometric contours for 1981 and contours of computed water levels, model layer 5.

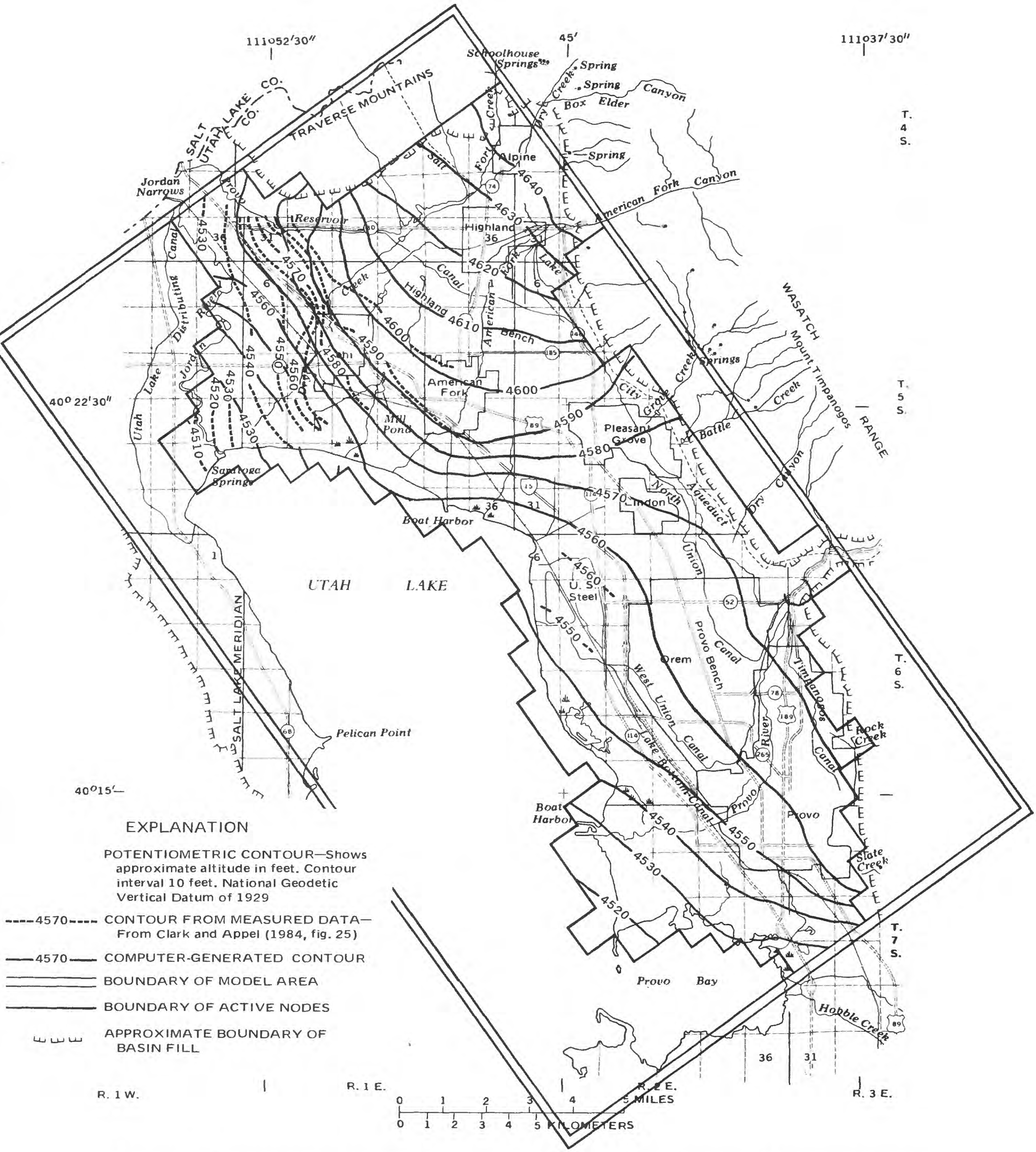


Figure 19.—Comparison of potentiometric contours for 1981 and contours of computed water levels, model layer 7.

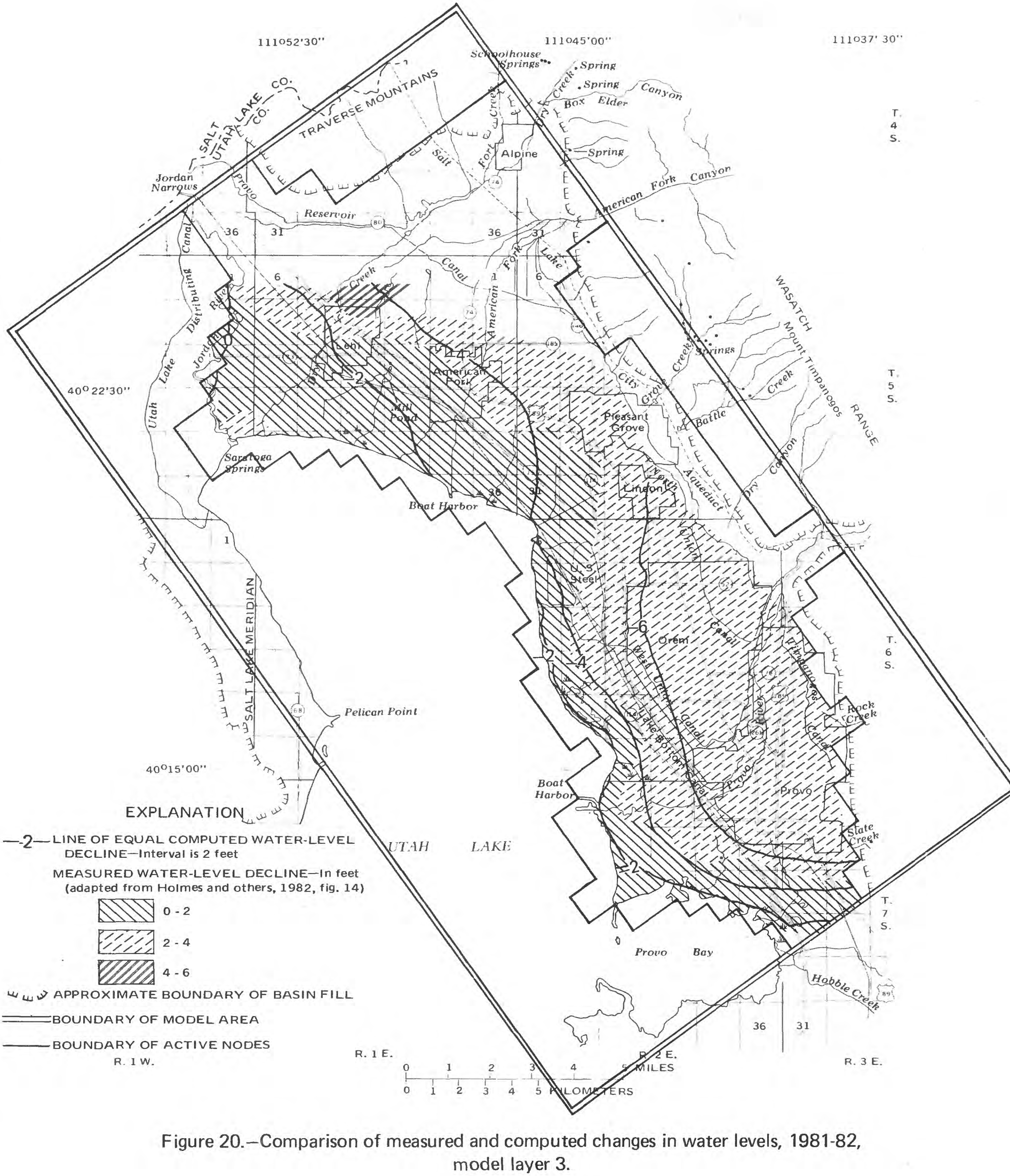


Figure 20.—Comparison of measured and computed changes in water levels, 1981-82, model layer 3.

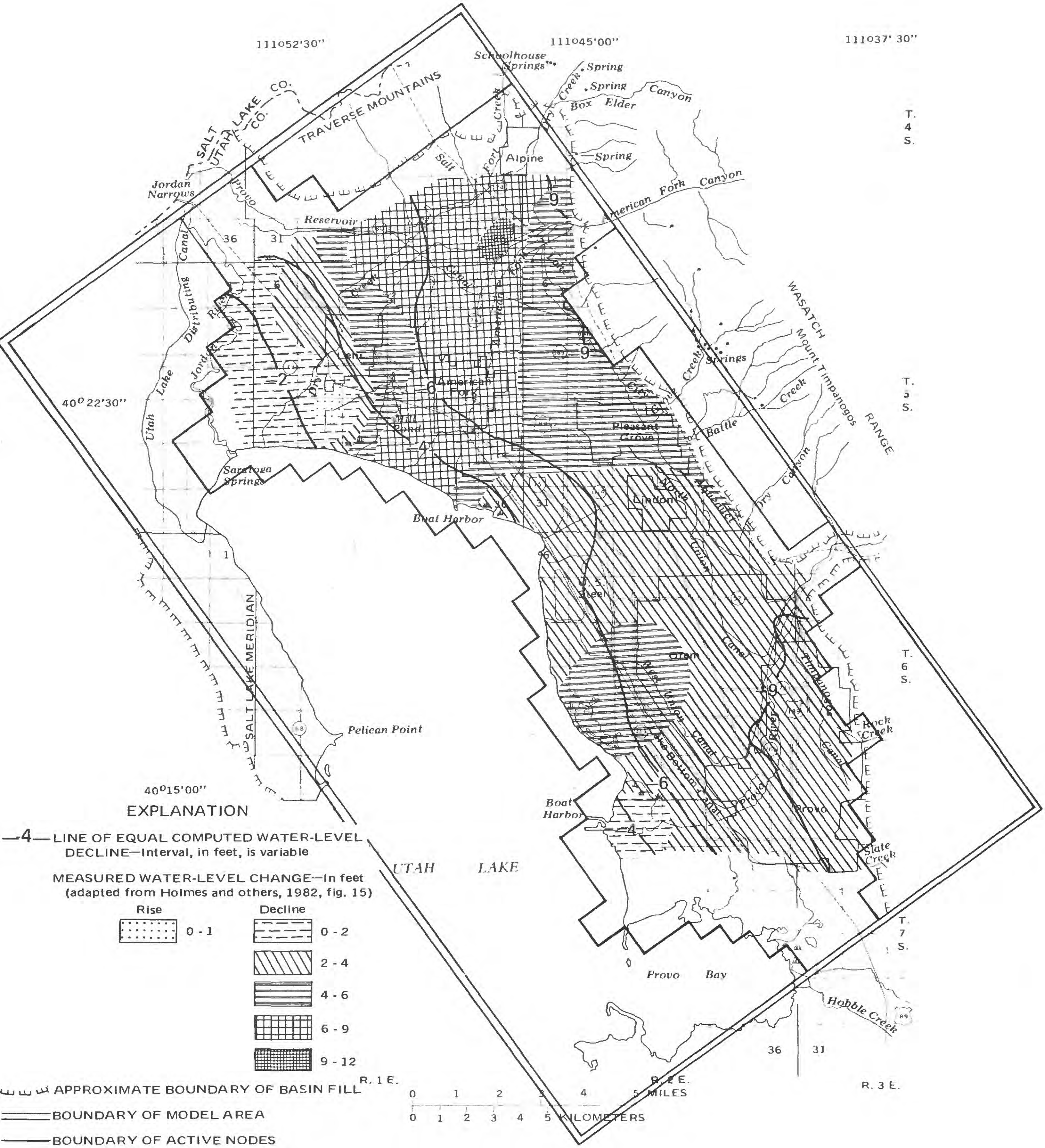


Figure 21.—Comparison of measured and computed changes in water levels, 1981-82, model layer 5.

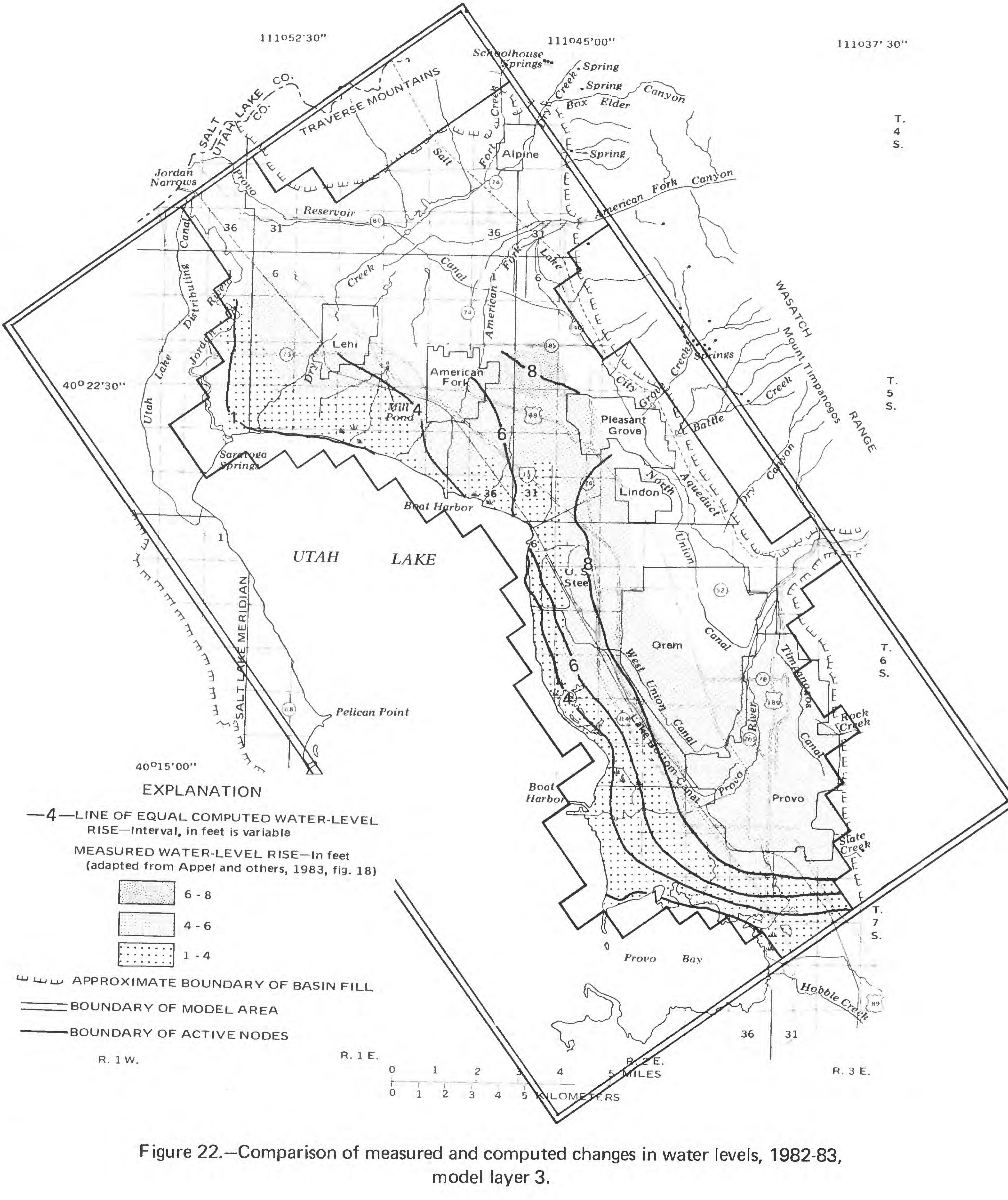


Figure 22.—Comparison of measured and computed changes in water levels, 1982-83, model layer 3.

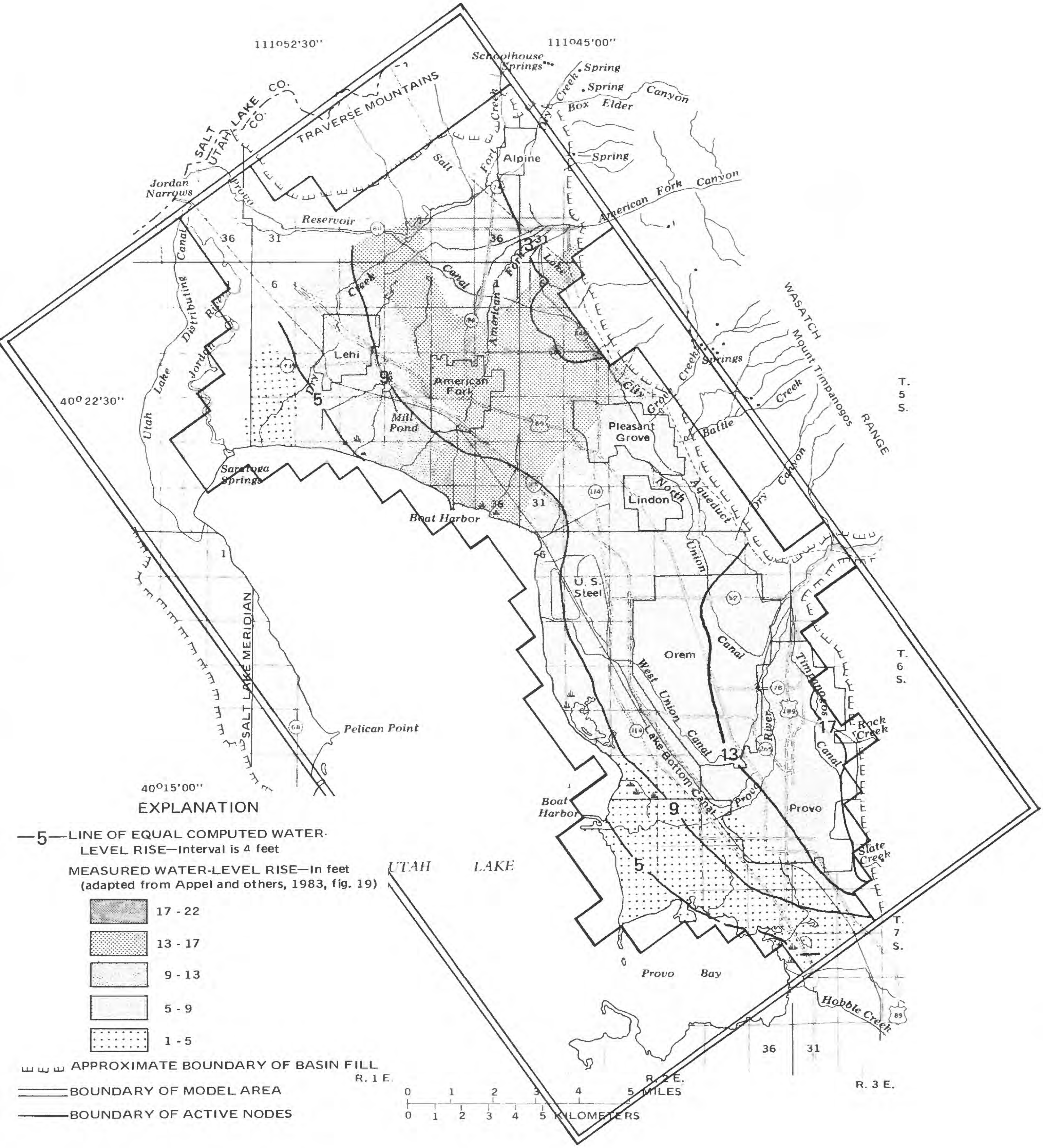


Figure 23.—Comparison of measured and computed changes in water levels, 1982-83, model layer 5.

For 1981-82, the computed drawdowns generally are greater than the measured drawdowns in layer 3 for most of the area modeled (fig. 20); but the computed drawdowns generally were less than the measured drawdowns for layer 5 in the northern half and more than the measured drawdowns for layer 5 in the southern half of the area (fig. 21). From 1982-83, when water levels rose to near record-high levels in many wells, the computed rises generally were larger than the measured rises for layer 3 (fig. 22). For layer 5, however, the computed rises generally were less than the measured rises in the northern half of the area, especially near the recharge area, but the computed rises were generally greater than the measured rises in the southern half of the area (fig. 23). Computed water-level changes at 12 observation wells in layer 7 for both time periods were a close approximation of the measured changes, ranging in difference from 0.1 to 5 feet.

The rate of ground-water discharge into drains and by springs and out of constant-head nodes used in the transient-state calibration varied with changes in recharge and discharge from wells. Table 5 shows the discharge of these sources during the various pumping periods.

Calibration Using Aquifer Test

Additional calibration of the model was done by simulation of an aquifer test near Lehi wherein a well finished in the deepest aquifer (model layer 7) was pumped for 21 hours and drawndown was measured in wells finished in each of the aquifers (layers 3, 5, and 7). The primary purpose of the simulation was to check the values for vertical hydraulic conductivity that were used for that area.

Results of the simulation of the aquifer test shown in table 6 indicate that the computed drawdowns for layers 3 and 7 are a close approximation of the measured drawdowns, whereas computed drawdowns for layer 5 do not match as well. The comparative results are reasonable considering that drawdowns are computed for the entire area of a node rather than for a point as represented by a well.

Sensitivity Analysis

Sensitivity analysis during the transient-state calibration was conducted by increasing or decreasing the values for storage coefficient by as much as an order of magnitude. Decreasing the storage coefficient an order of magnitude resulted in no water-level changes from the calibrated transient-state model during stress periods 1, 3, and 5. For the remaining stress periods, the range of water-level changes was -8 to +10 feet. Decreasing the storage coefficient generally resulted in larger water-level fluctuations between stress periods than was computed in the calibrated model. Increasing the storage coefficient an order of magnitude resulted in water-level changes in all stress periods ranging from -8 to +15 feet. This increase in the storage coefficient generally dampens the large changes in water levels between stress periods, because of the increased quantity of water in storage. To illustrate the effect of changing the storage coefficient by an order of magnitude, data for a specific node in layer 7 in the recharge area are presented in table 7.

Table 5.--Pumping periods and total discharge out of constant-head nodes and into drains and by springs used in the transient-state calibration for 1947-83

	Discharge (acre-feet per year)	
Pumping period	Constant-head nodes	Drains and springs
1947-50	52,200	102,500
1951-55	36,300	85,300
1956-62	34,500	83,200
1963-65	51,700	103,300
1966-73	50,700	102,800
1974-77	27,900	77,200
1978-80	47,000	98,000
1981	38,200	89,200
1982	52,400	102,900

Table 6.--Comparison of measured versus computed water-level drawdowns from an aquifer test near Lehi

Well location Model node Layer	Row	Column	Measured drawdown (feet)	Computed drawdown (feet)
3	5	6	1.15	1.8
3	6	7	1.63	1.6
3	6	8	.48	1.3
3	7	6	.72	1.3
7	5	8/9[1]	15.2	40.5/6.9
7	5	5	4.65	4.0
7	4	8	5.1	8.4
5	5	5	.67	2.9
5	6	8	1.60	5.2
5	8	6	.9	2.2
5	8	7	.6	2.4

[1] Well is near boundary of columns 8 and 9; computed drawdown is shown for both columns.

Table 7.--Sensitivity analysis for storage coefficient for transient-state calibration

Stress period	Calibrated model		Storage coefficient x 10		Storage coefficient / 10	
	Drawdown (feet)	Water-level change (feet)	Drawdown (feet)	Water-level change (feet)	Drawdown (feet)	Water-level change (feet)
5	+2.6		-2.8		+3.0	
		-28.8		-6.9		-36.7
6	-26.2		-9.7		-33.7	
		+23.7		+4.5		+40.8
7	-2.5		-5.2		+7.1	

PREDICTIVE SIMULATIONS

After transient-state calibration was complete, simulations were made with varying quantities of well discharge and recharge in order to estimate effects on water levels. Starting hydraulic heads for the predictive simulations were taken from the final computed heads at the end of the seventh pumping period (1978-80). During the predictive simulations, discharge from flowing wells, pumped irrigation wells, and industrial wells were assumed to remain constant at a withdrawal rate of 36,300 acre-feet per year, while only discharge from public-supply wells was varied. Illustrations for the predictive simulations are representative of layers 5 and 7 (deep artesian aquifer and Quaternary or Tertiary artesian aquifer). Predictive simulations for layer 3 (shallow artesian aquifer) generally resulted in projected drawdowns of about 1 to 4 feet less than in layers 5 and 7 primarily because of smaller withdrawals for public supply from this layer.

A simulation was made with the recharge rate of 190,000 acre-feet per year and a discharge rate from wells of 50,100 acre-feet per year including 13,800 acre-feet per year for withdrawals for public supply. Figure 24 shows that the computed drawdowns for 20 years (1980-2000) are less than 5 feet in most of the area.

Three simulations were made where the initial discharge of 13,800 acre-feet per year for public supply was doubled for a 10-year period (1980-1990) and then quadrupled for an additional 10-year period (1990-2000) and all other withdrawals remained constant at 36,300 acre-feet per year. These rates were used because discharge for public supply increased at approximately the same rate during the previous 20 years (Clark and Appel, 1985, table 13). Figure 25 shows computed water-level declines of as much as 25 feet when recharge is maintained at 190,000 acre-feet per year for the 20-year period. Figure 26 shows that computed declines are generally less than 20 feet when recharge is increased by 5 percent, and figure 27 shows that computed declines exceed 30 feet after 20 years when recharge is decreased by 5 percent. Greater changes in the rate of recharge, of course, would result in greater changes in water levels.

LIMITATIONS OF MODEL

A lack of complete geohydrologic data made it necessary to make some basic assumptions in the construction of the model in order to simulate field conditions. Boundary conditions and the total thickness of the principal ground-water reservoir were approximated in some areas. Discharge to Utah Lake was simulated with constant-head nodes near the shoreline of the lake, but some discharge probably occurs farther to the west in the lake. During predictive simulation, altitudes of a few of these constant-head nodes were higher than the altitudes of adjacent active nodes causing water to enter the constant-head nodes from the west. This might actually happen if large drawdowns occur in this area.

No water-level data were available for a large part of the study area; therefore, estimated water levels were used during steady-state calibration, particularly along the mountain fronts. The altitudes of the constant-head nodes used for recharge during steady-state calibration also were estimated.

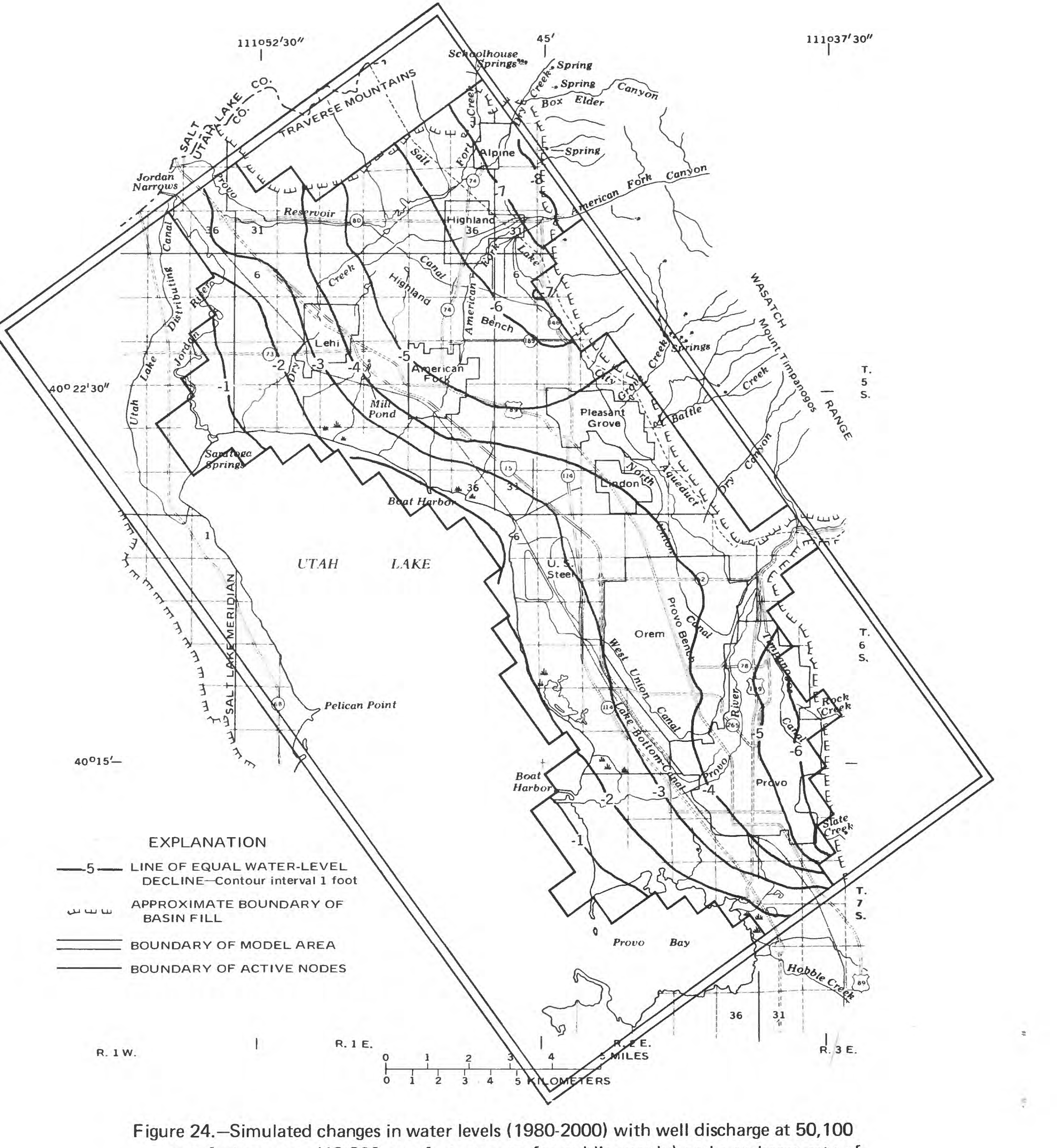


Figure 24.—Simulated changes in water levels (1980-2000) with well discharge at 50,100 acre-feet per year (13,800 acre-feet per year for public supply) and a recharge rate of 190,000 acre-feet per year.

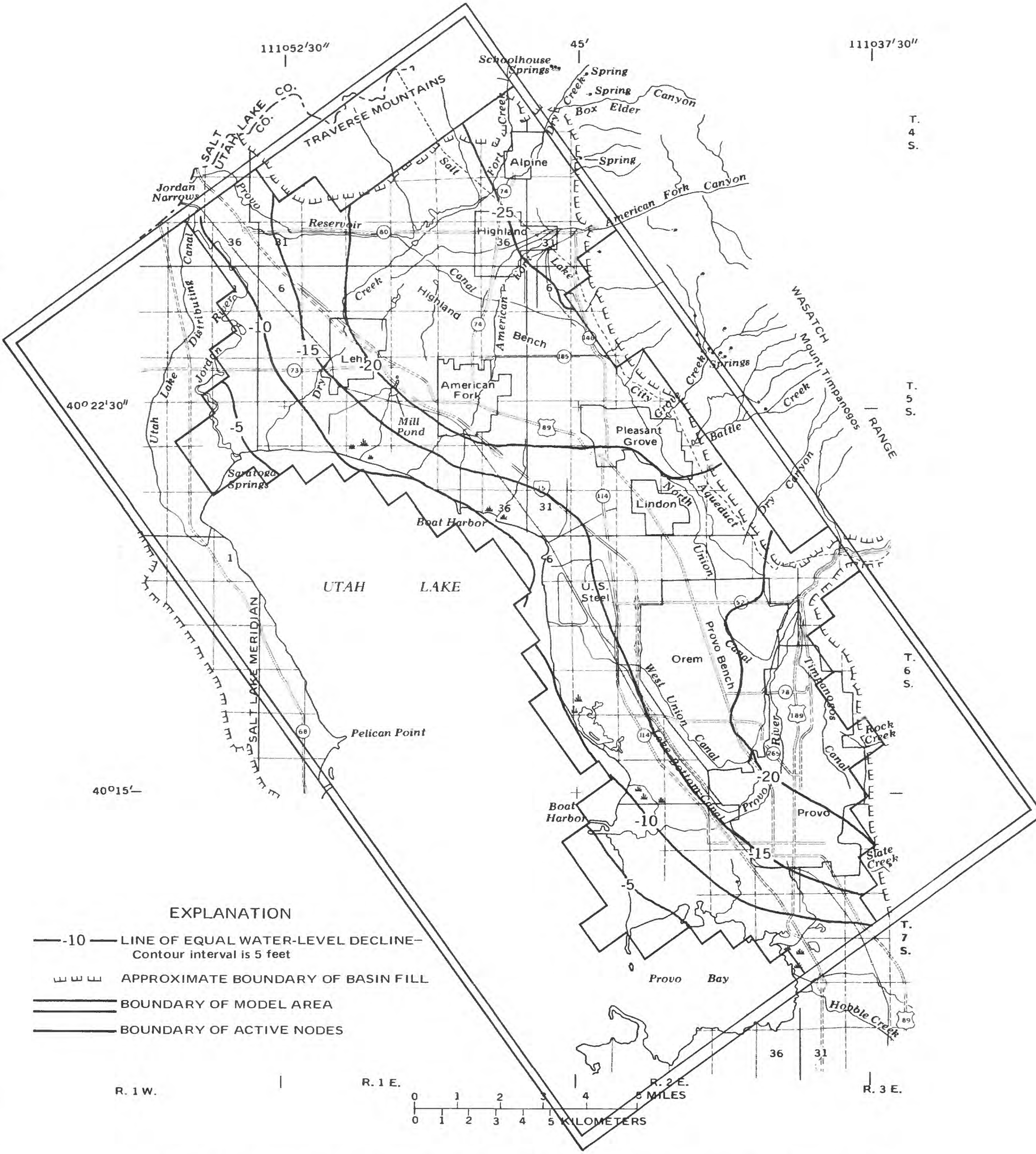


Figure 25.—Simulated changes in water levels (1980-2000) with annual well discharge of 63,900 acre-feet (27,600 acre-feet for public supply) for 10 years (1980-1990) and 91,400 acre-feet (55,100 acre-feet for public supply) for the next 10 years (1990-2000) and a recharge rate of 190,000 acre-feet per year.

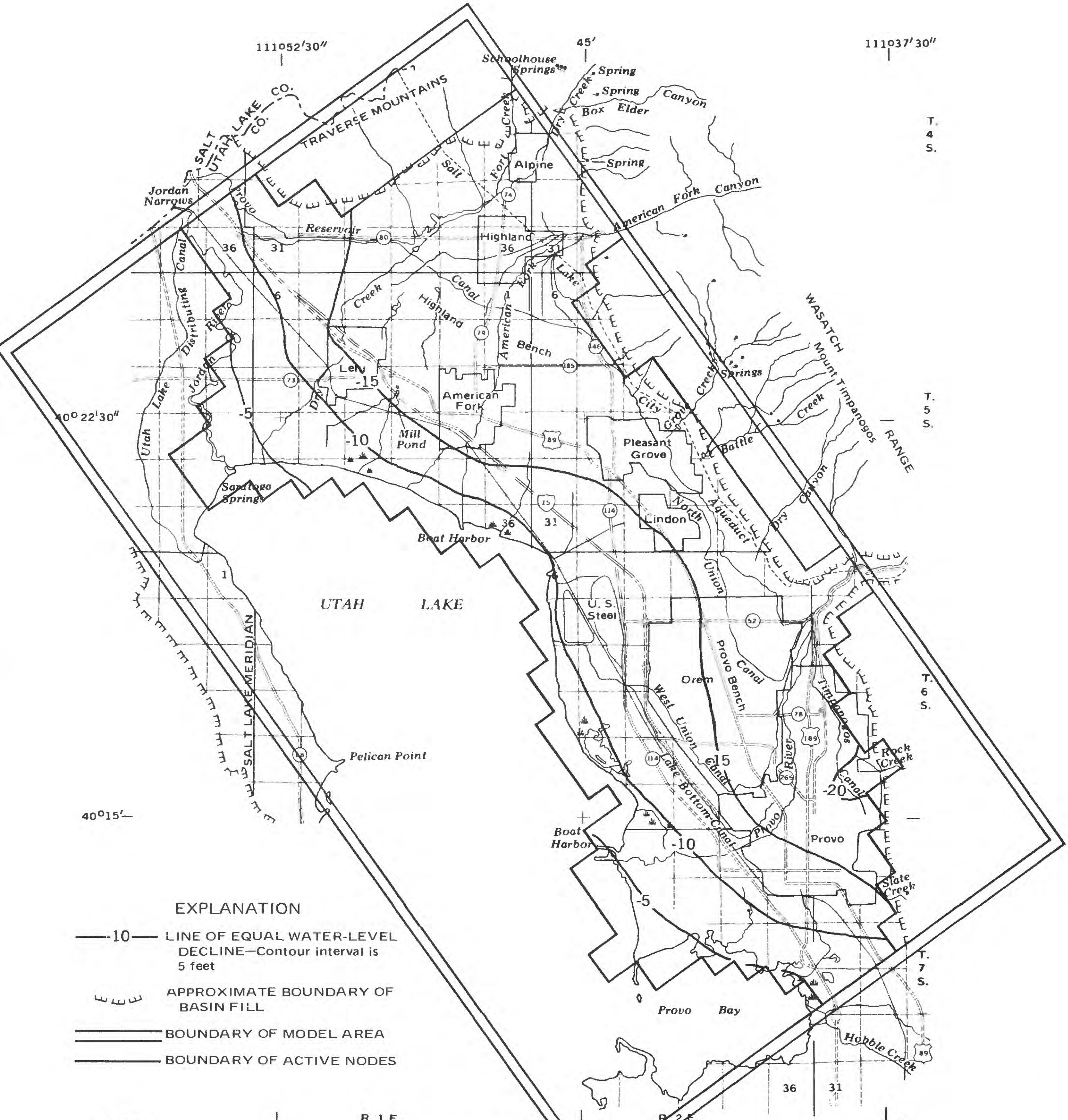
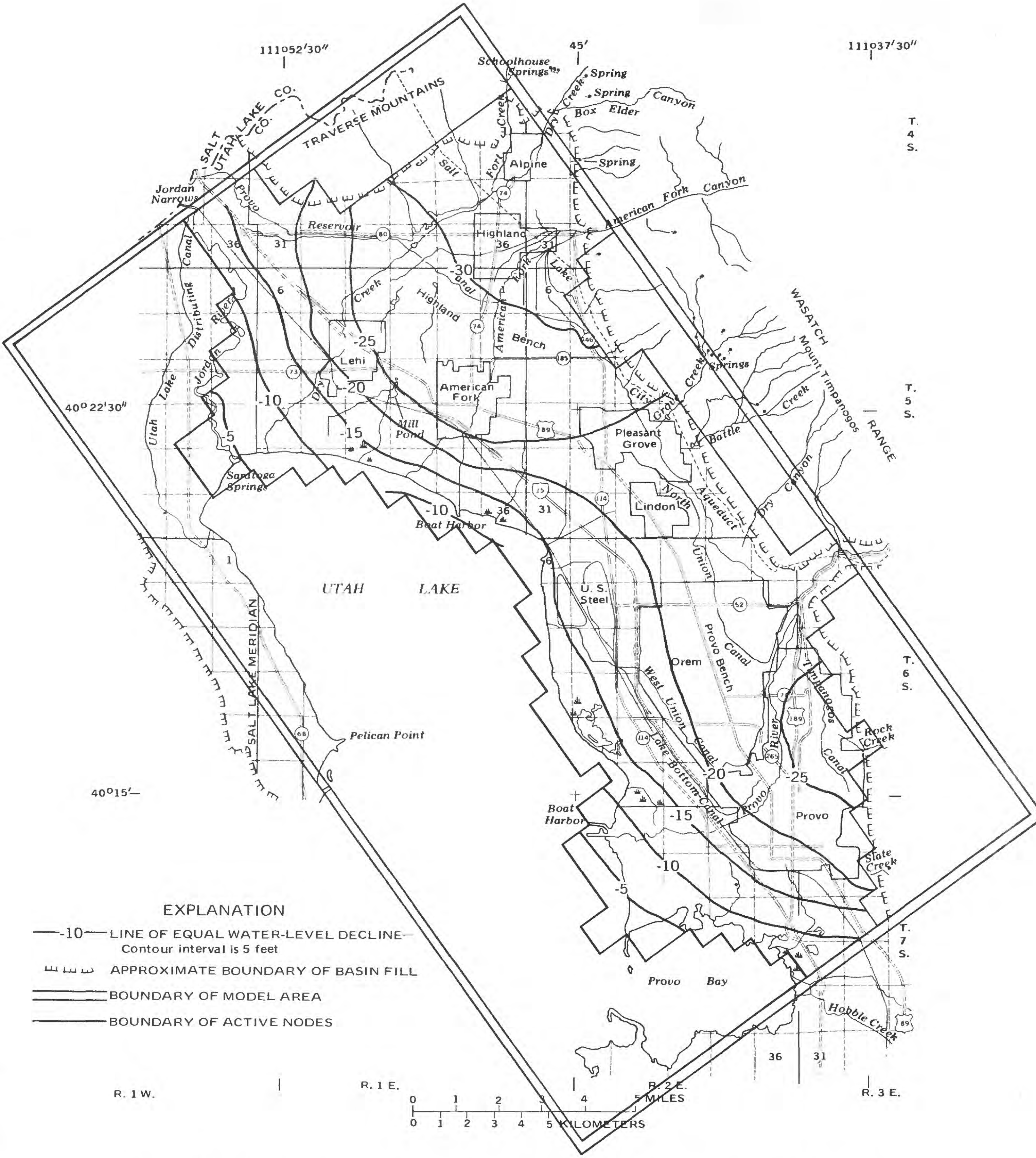


Figure 26.—Simulated changes in water levels (1980-2000) with annual well discharge of 63,900 acre-feet (27,600 acre-feet for public supply) for 10 years (1980-1990) and 91,400 acre-feet (55,100 acre-feet for public supply) for the next 10 years (1990-2000) and a recharge rate of 200,000 acre-feet per year.



EXPLANATION

—-10— LINE OF EQUAL WATER-LEVEL DECLINE—Contour interval is 5 feet

APPROXIMATE BOUNDARY OF BASIN FILL

BOUNDARY OF MODEL AREA

BOUNDARY OF ACTIVE NODES

Figure 27.—Simulated changes in water levels (1980-2000) with annual well discharge of 63,900 acre-feet (27,600 acre-feet for public supply) for 10 years (1980-1990) and 91,400 acre-feet (55,100 acre-feet for public supply) for the next 10 years (1990-2000) and a recharge rate of 180,000 acre-feet per year.

Most of the transient-state calibration involved changing the amount of recharge to the ground-water reservoir on the basis of changes in the total surface-water inflow to the study area. The assumption that recharge changes in direct proportion to surface-water inflow probably is valid, but to what extent this occurs is not fully predictable.

The values for vertical hydraulic conductivity used in the model were generally obtained by adjusting initial values. Therefore, while the areal predictive simulations probably are valid, simulated water-level changes in model layers overlying or underlying a pumped layer may not be simulated accurately at specific sites. For example, if large withdrawals are predicted from layer 5 for a certain model node, the resultant water-level changes for that node in model layers 3 and 7 may not be valid; but changes throughout the area probably would be realistic.

The discharge from flowing wells was estimated. The actual discharge could be greater or less than that used; therefore, simulated water-level changes in the flowing-well areas may not always be simulated accurately.

Despite the limitations, the model results for transient-state calibration were good. Predictive simulations, therefore, should yield satisfactory results for areas where future ground-water withdrawals are increased substantially during the next 20 years.

SUMMARY AND CONCLUSIONS

The U.S. Geological Survey evaluated the ground-water resources of northern Utah Valley, Utah, during 1980-82. As part of that study, a three-dimensional, finite-difference, digital-computer model was used along with ground-water data presented in previous reports to refine concepts of the ground-water system and to project effects on water levels of increases in withdrawals from the principal ground-water reservoir. The reservoir, simulated by the model as a seven-layer system, is composed of three artesian aquifers and their lateral extension, a deep water-table aquifer near the mountains. The model was calibrated for steady-state conditions using water levels measured in 1947 and for transient-state conditions using water-level changes from 1947-83.

Hydrologic conditions evaluated during calibration of the model included: transmissivity, vertical hydraulic conductivity, storage coefficient, subsurface inflow from bedrock, variations in recharge with time, and ground-water discharge to drains, springs, flowing wells, and Utah Lake. Values of vertical hydraulic conductivity and storage coefficient for aquifers and confining layers in the study area were derived from the model calibration. As part of the transient-state calibration of the model, changes in recharge to the principal ground-water reservoir during 1947-82 were correlated directly to changes in total surface-water inflow into the study area during that period. As part of transient-state calibration, ground-water discharge to Utah Lake and to drains and springs was correlated directly to changes in total recharge and withdrawal from wells. The model can be used to project ground-water discharge to Utah Lake under various potential changes of ground-water recharge and discharge.

Projections of water-level changes for 1980-2000 were made with varying rates of ground-water withdrawal for public supply and total ground-water recharge. The projections indicated that an average rate of recharge and an increase in withdrawal for public supply that was the same as the rate of increase during 1960-79 would result in water-level declines of as much as 25 feet.

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