SIMULATED WATER-LEVEL AND WATER-QUALITY CHANGES IN THE BOLSON-FILL AQUIFER, POST HEADQUARTERS AREA, WHITE SANDS MISSILE RANGE, NEW MEXICO

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CONVERSION FACTORS

In this report, values for measurements are given in inch-pound units only. The following table contains factors for converting these units to metric (International System) units.

Multiply inch-pound unit	By	To obtain metric unit
acre-foot	1,233	cubic meter
foot	0.3048	meter
gallon	3.785	liter
gallon per minute	0.06309	liter per second
mile	1.609	kilometer
square mile	2.590	square kilometer
inch	25.40	millimeter
cubic feet per second	28.32	liters per second
foot per day	0.3048	meter per day
foot squared per day	0.0929	meter squared per day

Chemical concentrations are given only in metric units (milligrams per liter) or unitless values (parts per million).

Sea level: In this report "sea level" refers to the 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 "Sea Level Datum of 1929."

GLOSSARY

Aquifer -- "A formation, group of formations, or part of a formation that contains sufficient saturated permeable material to yield significant quantities of water to wells and springs" (Lohman and others, 1972).

Area of diversion -- The area throughout which ground water moves toward a discharging well (Brown, 1963). Except for the condition of a flat water table, this area will be smaller than the area of influence of that same well.

Area of influence -- The area throughout which water levels have declined measurably due to discharge from a well. Theoretically, the effects extend to the boundary of the aquifer.

Bolson fill -- Unconsolidated and semiconsolidated sediments that fill an extensive, flat basin surrounded by mountains from which drainage has no surface outlet.

Brine -- Water containing a dissolved-solids concentration greater than that of sea water (about 35,000 milligrams per liter).

Coefficient of molecular diffusion -- Coefficient used in solute-transport models to represent the process by which solutes move according to concentration gradients.

Dispersivity -- Coefficient used in solute-transport equation to represent the degree to which solutes can move through the aquifer at velocities different from the average ground-water velocity.

Freshwater -- Water containing a dissolved-solids concentration less than 1,000 milligrams per liter.

Homogeneity -- "A material is homogeneous if its hydraulic properties are identical everywhere" (Lohman and others, 1972).

Hydraulic conductivity -- "The hydraulic conductivity of the medium is the volume of water at the existing kinematic viscosity that will move in unit time under a unit hydraulic gradient through a unit area measured at right angles to the direction of flow" (Lohman and others, 1972).

Isotropy -- "That condition in which all significant aquifer properties are independent of direction" (Lohman and others, 1972). Properties that are dependent upon direction are said to be anisotropic.

Milligrams per liter -- Concentration units of a substance in water expressed as mass of solute in milligrams divided by the volume of solution in liters. Relates to concentration expressed in parts per million (mass of solute in milligrams divided by mass of solution in kilograms) as: parts per million equals milligrams per liter divided by the specific gravity of the solution. Essentially equivalent to concentration expressed in parts per million for solutions containing less than about 7,000 milligrams per liter of solutes.

Reentrant -- "A prominent, generally angular indentation into a landform" (Gary and others, 1977). The Post Headquarters area is situated in a reentrant formed by the Organ Mountain fault.

Salt-water encroachment -- Displacement of fresh ground water by the advance of salt water (Gary and others, 1977).

Specific conductance -- Electrical conductance of a volume of unit length and unit cross section at a specified temperature. Units are microsiemens per centimeter at 25 degrees Celsius. Specific conductance of water is related to the quantity and type of dissolved solids, therefore providing a relative measure of its salinity.

Specific yield -- The volume of water yielded from water-bearing material by gravity drainage that occurs when the water table declines (Lohman, 1972).

Steady-state flow -- The condition when at any point in the ground-water system, the magnitude and direction of flow are constant with time (Lohman and others, 1972).

Storage coefficient -- "The volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head" (Lohman and others, 1972).

Transient flow -- The condition when at any point in the ground-water system, the magnitude or direction of flow changes with time.

Transmissivity -- "The rate at which water of the prevailing kinematic viscosity is transmitted through a unit width of the aquifer under a unit hydraulic gradient" (Lohman and others, 1972).

Zone of dispersion -- A zone of mixing between the saline water and salt water formed by dispersion of freshwater flow, molecular diffusion, and displacements of the interface by external influences such as tides, recharge, and ground-water withdrawals (Todd, 1959, p. 282). Also called zone of diffusion or transition zone.

SIMULATED WATER-LEVEL AND WATER-QUALITY CHANGES IN THE BOLSON-FILL AQUIFER, POST HEADQUARTERS AREA, WHITE SANDS MISSILE RANGE, NEW MEXICO

By Dennis W. Risser

ABSTRACT

The quantity of freshwater available in the Post Headquarters well field is limited, and its quality is threatened by saltwater encroachment. A three-dimensional, finite-difference, ground-water flow model and a cross-sectional, density-dependent, solute-transport model were constructed to evaluate possible future water-level declines and water-quality changes in the Post Headquarters well field.

A six-layer flow model was constructed for the upper 600 feet of saturated aquifer using hydraulic-conductivity values that ranged from 0.1 to 10 feet per day, a specific yield of 0.15, and an average recharge of about 1,590 acre-feet per year. Water levels simulated by the model closely matched measured water levels for 1948-82, except in the area of rising water levels near Old Supply Well 12. This well, and possibly the southern part of the reentrant, may not be well connected hydraulically to other parts of the well field. Possible future water-level changes for 1983-2017 were simulated using rates of ground-water withdrawal of 1,033 and 2,066 acre-feet per year and wastewater-return flow of 0 or 30 percent of the ground-water withdrawal rate. The maximum simulated drawdown for 35 years of withdrawal at 2,066 acre-feet per year was 50 feet. When wastewater returns were not simulated, water levels declined as much as 40 feet near the wastewater-treatment plant, which may hasten the encroachment of saline water toward the well field. Decreasing ground-water withdrawals to 1,033 acre-feet per year for 1983-2017 caused water levels to rise as much as 17 feet in the well field. In areas around the well field, water levels declined throughout the simulation.

The cross-sectional, solute-transport model indicated that the freshwater (less than 1,000 parts per million dissolved solids) is about 1,500 to 2,000 feet thick beneath the well field. Transient simulations indicate that solutes most likely will move laterally toward the well field rather than from beneath the well field. A measured increase of as much as 200 milligrams per liter of dissolved solids in water from Supply Wells 11, 13, 16, and 20 probably is caused by movement of solutes through individual sand lenses that were not included in this generalized model. Dissolved-solids concentrations could increase at least 500 parts per million in the well field during 1983-2017 if individual sand lenses are connected to saline water in the eastern part of the modeled area.

INTRODUCTION

White Sands Missile Range is located in the Tularosa Basin of south-central New Mexico (fig. 1). The study area encompasses about 190 square miles in the vicinity of the Post Headquarters well field, part of which extends onto the neighboring Fort Bliss Military Reservation to the south.

Ground water from bolson-fill deposits at the Post Headquarters well field provides about 98 percent of the water used at the White Sands Missile Range. From 1948 to 1982, about 63,000 acre-feet of water was withdrawn from the freshwater part of the bolson-fill aquifer, which is bounded by the San Andres, San Augustin, and Organ Mountains on the west (fig. 1) and by saline ground water on the east. The large volume of withdrawals and proximity to saline water have caused concern that the freshwater supply will be threatened by saltwater encroachment. During the last 30 years, dissolved-solids concentrations have increased in water from seven supply wells. Therefore, the U.S. Geological Survey, in cooperation with White Sands Missile Range, U.S. Department of the Army, began a study in 1983 of the hydrology in the vicinity of the Post Headquarters area with emphasis on evaluating possible future water-level and water-quality changes.

Purpose and Scope

This report describes the results of a study to evaluate the potable water resources available to wells in the Post Headquarters and adjacent areas. Because the availability of potable water is partly dependent on the rate of saltwater encroachment, an evaluation of possible water-quality changes is included. The investigation was conducted by simulating water-level declines and water-quality changes with ground-water flow models. Hydrologic data used to construct flow models were collected by previous investigators. The scope of the investigation is focused on the Post Headquarters well field. The study area extends to the Small Missile Range and to just north of the Soledad Canyon area (figs. 1 and 2) to allow for adequate treatment of boundary conditions in the mathematical ground-water flow model.

Geographic Setting

The study area is in the Mexican Highland section of the Basin and Range physiographic province (Fenneman, 1931). On the western edge of the area are the San Andres, San Augustin, and Organ Mountains, which rise to a maximum height at the Organ Needle of 8,916 feet above sea level. The plains of the Tularosa Basin extend to the east.

There are no major perennial streams in the study area. After summer thunderstorms, water flows down ephemeral streams onto alluvial fans. During large storms, the ephemeral streams carry water onto playa flats about 4 miles east of the Post Headquarters, creating ephemeral playa lakes (fig. 2).

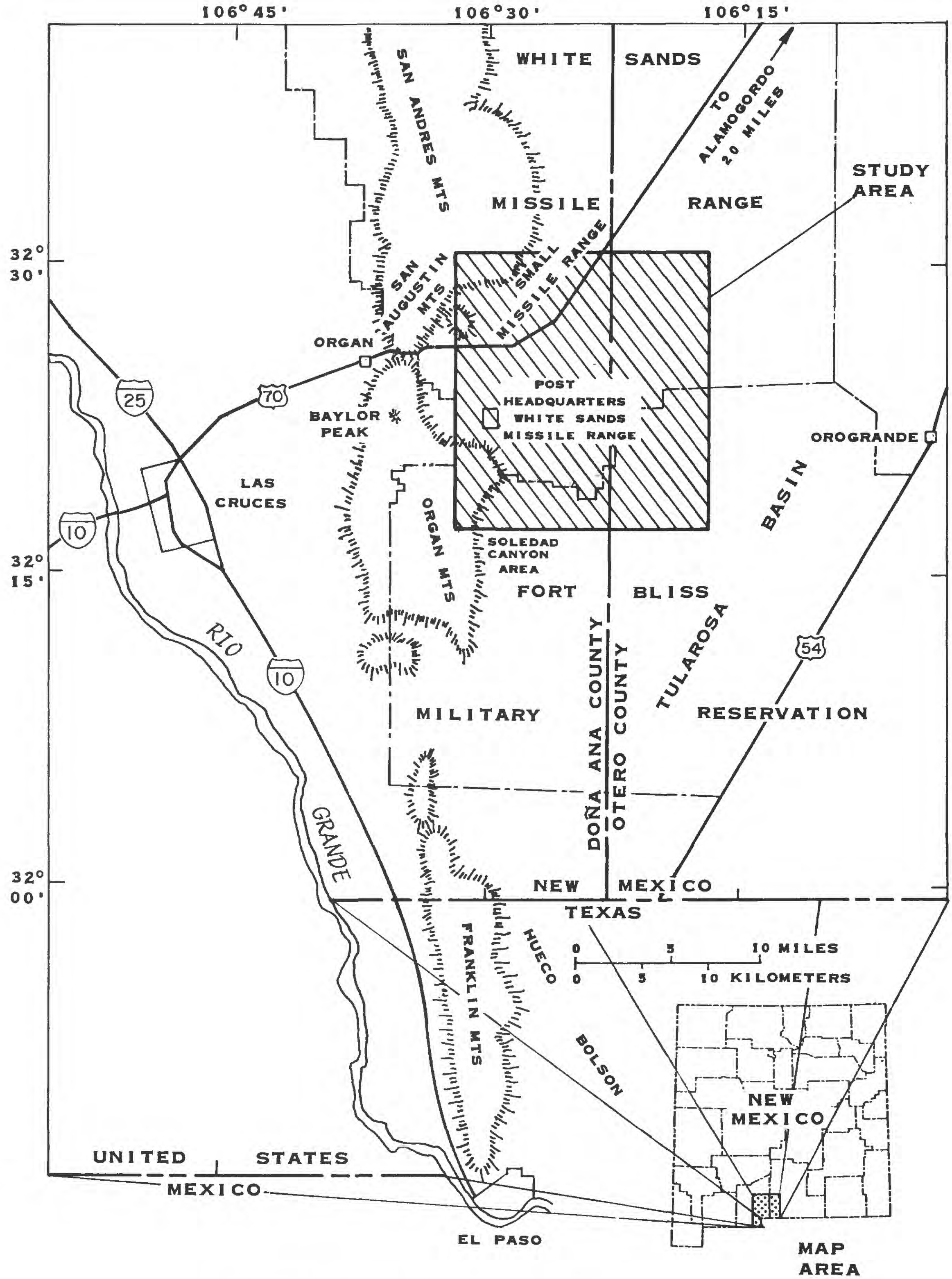


Figure 1.--Location of study area.

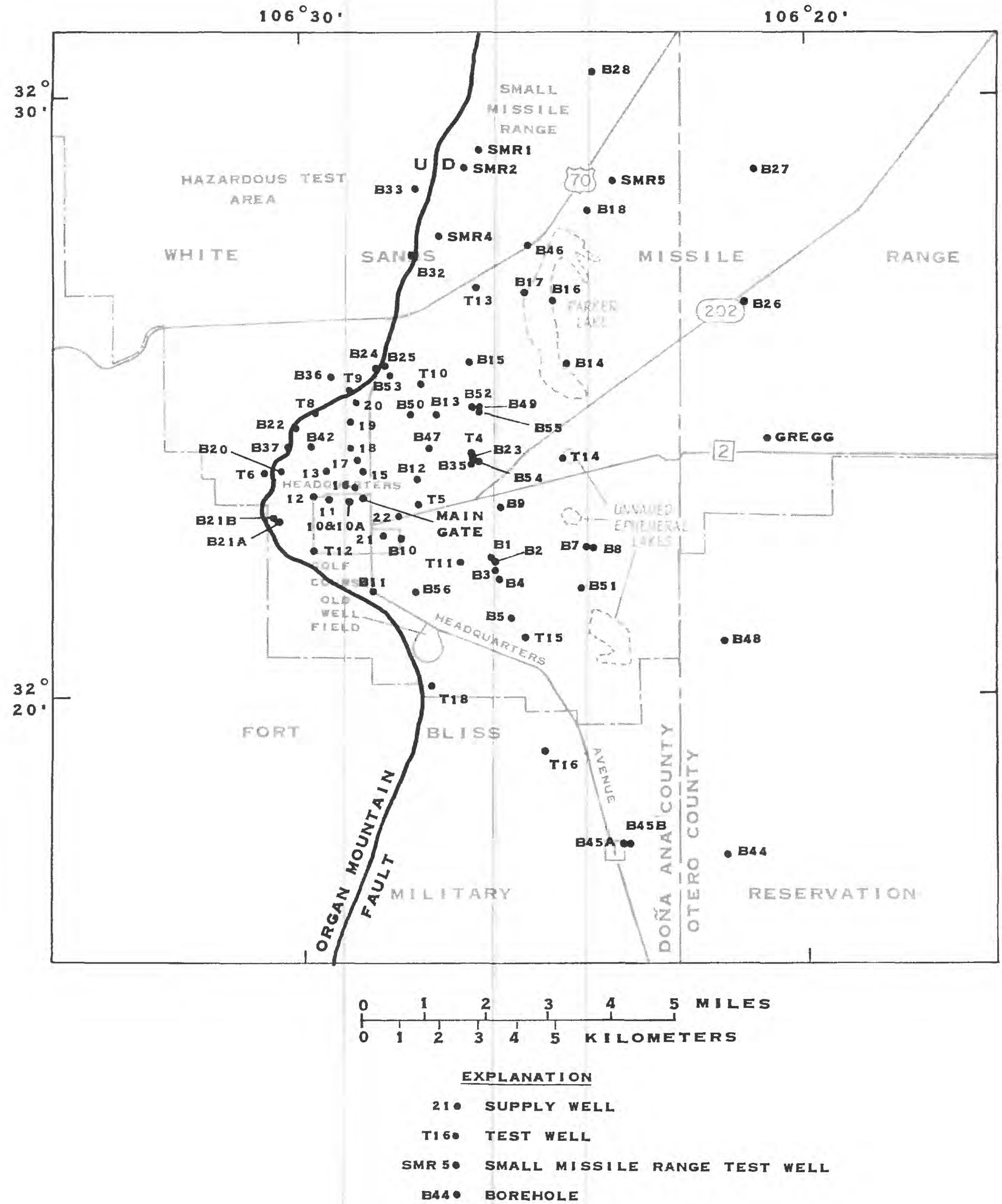


Figure 2.--Location of selected supply wells, test wells, and boreholes.

Main access to the Post Headquarters is by U.S. Highway 70, which connects Las Cruces and Alamogordo (fig. 1). The resident population of the base in 1980 was about 3,000. An additional work force of about 10,000 commutes to the base (Cunniff, 1980, p. 2).

Previous Investigations

Numerous hydrologic-data reports have been compiled. Many were published by the U.S. Geological Survey, which has conducted a hydrologic testing and surveillance program since 1955. These data reports are listed in the references as "Annual Water-Resource Reviews" and well-drilling and aquifer-testing reports.

Reports that summarize hydrologic data and include some hydrologic interpretations for the Post Headquarters area have been compiled by several authors. Hood (1963) summarized ground-water investigations for July 1960 through June 1962. Davis and Busch (1968) summarized all U.S. Geological Survey investigations at White Sands Missile Range completed at that time. Cooper (1970) made a summary of supply- and test-well records in the Post area. Kelly (1973) summarized available ground-water data and included a discussion of water-level changes, hydraulic properties, and ground-water chemistry. Kelly's report was used in a report by White Sands Missile Range (1975), which is a planning document to guide future water development.

Investigations that were primarily interpretive in nature also have been completed. Herrick (1960b) made a study of the Post Headquarters area that included a resistivity survey, an estimate of ground water in storage, estimates of recharge rates, and suggestions for locating new supply wells. Ballance and Basler (1967) studied runoff from a small paved watershed as a potential method to increase the available potable water supply. Sedillo (1969) investigated recharge to ground water from sewage effluent. Scott (1970) made estimates of runoff from the Post Headquarters watershed that were improved (Scott, 1976) after more data were collected. Kelly and Hearne (1976) made a two-dimensional, finite-difference, ground-water flow model of the Post area. They used the model to simulate potential ground-water declines through 1996.

GENERALIZED GEOLOGY

The San Augustin and Organ Mountains mainly are composed of Tertiary quartz monzonite and Precambrian granite. The San Andres Mountains consist of pre-Tertiary sedimentary rocks and Precambrian granite. The mountains were uplifted relative to the Tularosa Basin along the Organ Mountain fault in late Tertiary time, although scarps that cut recent alluvial fans indicate that some faulting may have been active as recently as 1,100 years ago (Ruhe, 1967, p. 17). Total displacement along the Organ Mountain fault may be greater than 10,000 feet (Seager and Brown, 1978, p. 148), and the dip of the fault is steep, probably between 50 and 60 degrees (Zohdy and others, 1969, p. 27; Seager and Brown, 1978, p. 147). The position of the fault at depth (fig. 3) is not well known but has been inferred using a gravity model (Seager and Brown, 1978, p. 147) and gravity, seismic, and electrical-resistivity measurements (Zohdy and others, 1969, p. 26).

Sediments washed from the mountains during Tertiary and Quaternary time have filled the basin with a heterogeneous mix of sand, gravel, silt, and clay. Lacustrine deposits of clay and gypsum occur within the alluvial sediments in the center of the Tularosa Basin. The thickness of the bolson-fill sediments may be as much as 8,000 feet (fig. 3). Test Well 14, about 4 miles east of the Post Headquarters well field, reportedly penetrated 6,005 feet of unconsolidated sediments without encountering bedrock (Doty and Cooper, 1970).

The physical character of the bolson fill varies laterally and vertically in the study area. Generally, coarse-grained, poorly sorted sediments were deposited near the mountain fronts. These deposits grade into fine-grained, well-sorted sediments toward the center of the basin. Kelly (1973, p. 20) compiled a map that shows the percentage of sand between 500 and 1,000 feet below land surface. The map indicates that sediments in the zone along the mountain fronts contain as much as 80 percent sand compared to sediments east of Headquarters Avenue that contain less than 40 percent sand. In general, sediments further from the mountain fronts contain a greater percentage of clay and gypsum than those near the mountain fronts.

In addition to lateral heterogeneity, characteristics of the bolson sediments vary with depth. Kelly and Hearne (1976, p. 29 and 59) identified a clay unit about 1,000 feet below land surface that they believed to be continuous and to be an effective barrier to ground-water flow. Whether the clay zones, indicated by several well logs, represent a continuous layer or discontinuous lenses is open to interpretation. The best record of changes in lithology with depth is from Test Well 14. Unfortunately, this well is located several miles east of the bolson sediments that contain freshwater. Major lithologic changes in Test Well 14 are shown in figure 3. The sand and silty sediments at the bottom of the section may be old alluvial-fan deposits that were formed when the mountain fronts were further east than their present position or they may indicate that older sediments eroded from the mountains were coarser grained than more recent sediments (Doty and Cooper, 1970, p. 25). An increase in seismic velocity of sediments in Test Well 14 from about 6,600 feet per second near the surface to 15,200 feet per second at the bottom of the hole (Doty and Cooper, 1970, fig. 2d) indicates a significant increase in compaction and cementation of sediments at depth.

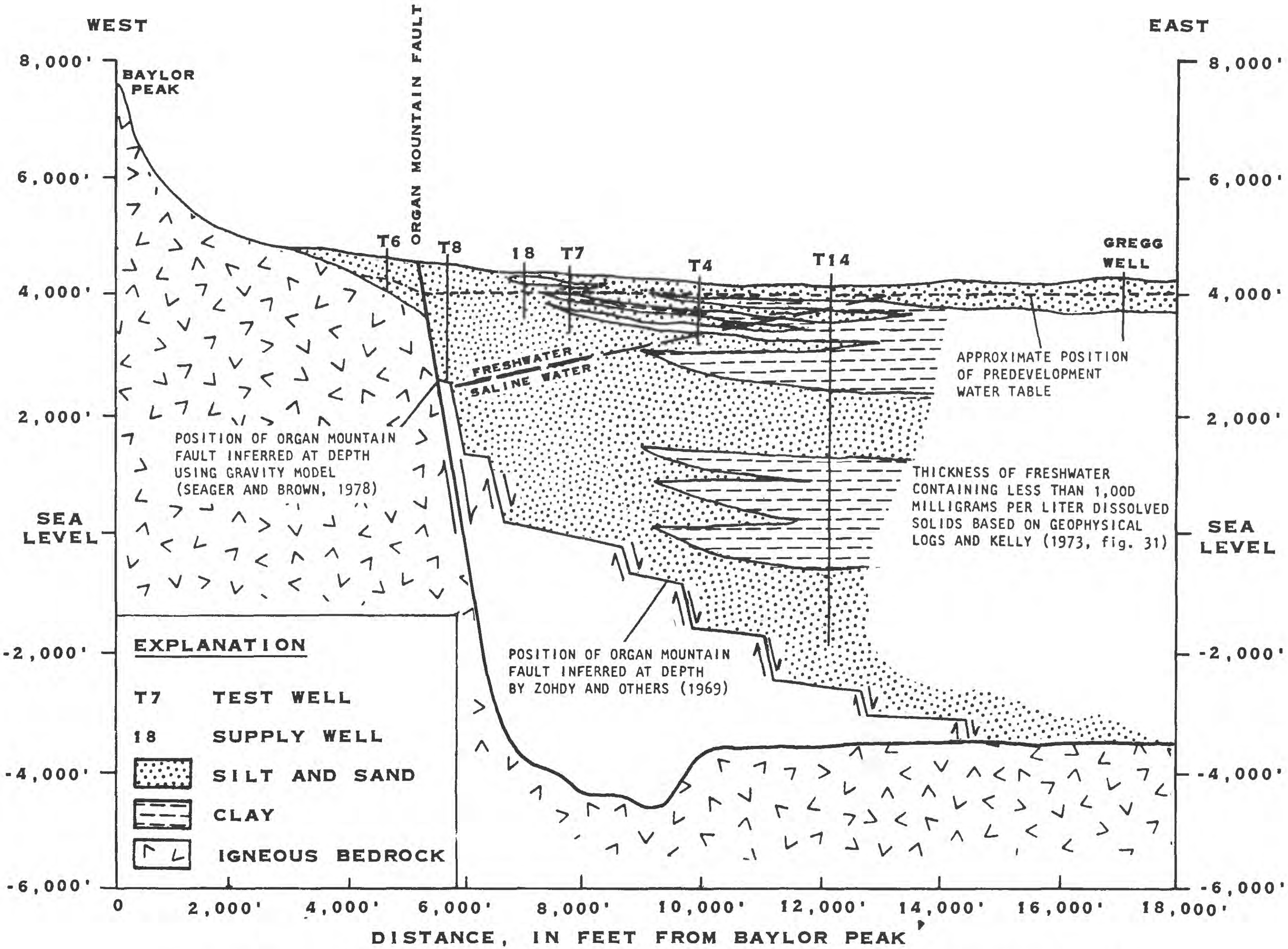


Figure 3.--Generalized hydrologic section showing approximate position of interface between fresh and saline water.

GROUND WATER

Source and Occurrence

Potable water is withdrawn by the Post Headquarters wells from bolson-fill sediments in a reentrant along the eastern side of the Organ, San Augustin, and San Andres Mountains (fig. 1). The freshwater occurs in a zone that extends 3 to 5 miles from the mountain front (Kelly, 1973, fig. 31). The freshwater part of the bolson fill is bounded to the west by granite and quartz monzonite bedrock and to the east by saline water. The thickness of the freshwater beneath the reentrant is not well known. Kelly (1973, fig. 31) showed the thickness of freshwater beneath the Post Headquarters well field to be at least 1,500 feet, mainly on the basis of results of electrical soundings by Zohdy and others (1969, fig. 17). According to McLean (1970, p. 49), freshwater containing less than 1,000 milligrams per liter dissolved solids is about 1,800 feet thick beneath the well field. Calculations using the Ghyben-Hertzberg relation (Todd, 1959, p. 278) indicate that freshwater could extend several thousand feet beneath the well field. This technique, however, assumes a hydrostatic condition and uniform properties of the aquifer, which are not strictly met in the study area.

Potable ground water is withdrawn by wells from numerous sandy units that comprise a heterogeneous mix of bolson-fill sediments. Movement of water toward the wells and release of water from storage are complicated because the aquifer is heterogeneous and the sandy units in which supply wells are completed are not continuous between wells. These discontinuities could be caused by depositional processes or created by recent faulting within the bolson fill.

The source of water to wells and aquifer response are determined by the time period being considered. For short periods measured in hours, the separate sandy units probably act as individual confined aquifers. The almost instantaneous water-level changes recorded in Old Supply Well 15 in response to withdrawal changes in the Post Headquarters well field exemplify this condition. However, for longer periods measured in days or years, leakage between sandy units may allow stresses caused by withdrawals to be integrated until the collection of sandy units responds as a homogeneous aquifer system. The integrated aquifer-system concept is adopted in this study as a working hypothesis to investigate water-level declines and water-quality changes in the bolson-fill aquifer. In some instances, this conceptual model is not adequate to accurately evaluate aquifer response. These instances are tested and discussed in the modeling section of the report.

Recharge

The potable ground-water resource in the Post Headquarters area is replenished mainly by seepage from arroyos on alluvial fans that issue from the Organ and San Andres Mountains. Generally, these arroyos only contain water during and for a short time after storms. Precipitation that does not concentrate in arroyo channels probably is lost through evapotranspiration. Snowmelt during the early spring, when evapotranspiration rates are low, may provide an additional minor amount of recharge.

The rate at which potable ground water is recharged in the Post area has been estimated by two approaches: (1) Calculating ground-water flow across a section near the mountain front using a form of Darcy's Law, and (2) calculating average annual runoff from the Post Headquarters watershed and assuming that all runoff recharges the aquifer. Both of these approaches are flawed to some extent. To use Darcy's Law, transmissivity of the aquifer must be known. However, aquifer tests indicate that transmissivity values may vary by an order of magnitude in the Post Headquarters well field (Kelly and Hearne, 1976, p. 31). The assumption that all runoff recharges the aquifer is not correct. Runoff from large storms is known to pond in playa lakes about 4 miles east of the Post Headquarters. Water-level measurements in boreholes located in the vicinity of the playas (boreholes 14, 16, and 51 in fig. 2) indicate that very little recharge occurs from water ponded in these lakes. Evidently, most of the water evaporates because seepage through the clayey playa bottoms is restricted.

Herrick (1960a) computed recharge to the bolson fill by estimating the quantity of ground-water flow through a cross section of the aquifer west of the Post Headquarters. Applying a form of Darcy's Law:

$$Q = T\ W\ (dh/dl) \qquad (1)$$

where Q = ground-water flow rate (L^3/T),

T = transmissivity (L^3/T),

W = width of section (L), and

dh/dl = hydraulic gradient (L/L),

Herrick estimated ground-water flow to be about 6,000 acre-feet per year across a section 4.5 miles long, given a hydraulic gradient of 80 feet per mile and a transmissivity of 2,000 feet squared per day.

Herrick also used the estimate of Sayre and Livingston (1945, p. 72) that about 25 percent of precipitation falling on the Franklin and Organ Mountains recharges the Hueco Bolson sediments south of White Sands Missile Range. Applying this ratio to the 40-square-mile Post Headquarters watershed and assuming an average annual precipitation of 13 inches, he estimated ground-water recharge to be about 7,000 acre-feet per year.

White Sands Missile Range (1975) estimated recharge on the basis of Herrick's assumptions, but used a shorter cross section. They estimated a recharge rate of about 4,000 acre-feet per year.

W.C. Ballance and S.M. Longwill (U.S. Geological Survey, written commun., 1968) also estimated recharge by computing flow across a section of aquifer. On the basis of a cross section 4.5 miles long east of Headquarters Avenue, a hydraulic gradient of 40 feet per mile, and a transmissivity of 1,270 feet squared per day, a recharge rate of about 1,900 acre-feet per year was calculated. On the basis of a 4-mile-long section located 1 mile west of the highway, a hydraulic gradient of 200 feet per mile, and a transmissivity of 280 feet squared per day, a recharge rate of about 1,900 acre-feet per year was calculated. Ballance and Longwill also noted that 5 percent of their estimate of average annual precipitation (16 inches per year) on the watershed (40 square miles) would provide recharge of about 1,700 acre-feet per year.

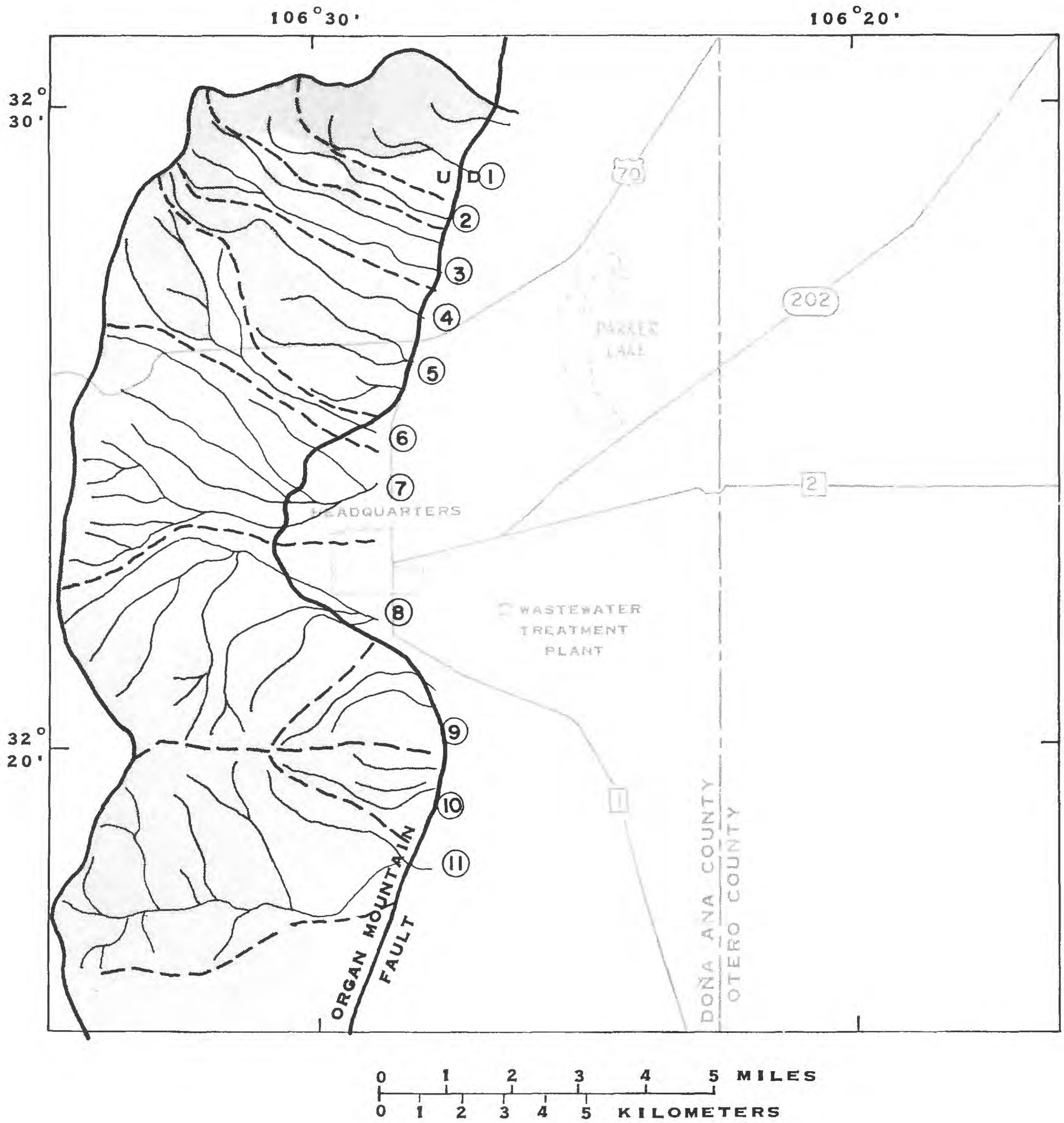
Runoff from a small watershed was measured by Ballance and Basler (1967). They found the runoff to be about 3 percent of the total precipitation on the watershed. Kelly and Hearne (1976) assumed that if 3 percent of the 16 inches of average annual precipitation runs off the 52-square-mile watershed at the Post area, the total runoff would be about 1,300 acre-feet per year. They assumed that all 1,300 acre-feet of runoff would infiltrate to the aquifer.

For this study, recharge was estimated by calculating the average runoff from the Hazardous Test Area, Post Headquarters, and Soledad watersheds using physical characteristics of each watershed as described by Borland (1970). These estimates are summarized below and in figure 4.

Watershed	Area, in square miles	Estimated annual runoff, in acre-feet	Estimated runoff, in acre-feet per square mile
Hazardous Test Area	18	390	22
Post Headquarters	46	1,140	25
Soledad	21	500	24

Assuming that all runoff infiltrates to the water table, recharge in the Post area is about 1,140 acre-feet per year according to this method. This estimate is similar to Kelly and Hearne's (1976) estimate of 1,300 acre-feet per year calculated for a slightly larger watershed area.

Scott (1976) estimated annual runoff from an 81-square-mile watershed in the Post Headquarters area. On the basis of measured streamflow collected during 8 years, he estimated runoff to be about 240 acre-feet per year. This estimate is much lower than any of the other runoff computations. The location of the gaging stations along the Post Headquarters access road probably affected the results. The gages were located far enough downslope on the alluvial fans that a considerable quantity of runoff probably seeped into the ground upstream from the gages. Another possible explanation why gaged discharge from arroyos differed so greatly from estimated average runoff could lie in the relative importance of infrequent floods.



EXPLANATION

BEDROCK OUTCROP

WATERSHED BOUNDARY

SUBWATERSHED BOUNDARY

WATERSHED	SUB-WATERSHED NUMBER	AREA, IN SQUARE MILES	RUNOFF, IN ACRE-FEET
HAZARDOUS TEST AREA	1	5.3	120
	2	3.2	66
	3	5.6	130
	4	4.0	74
POST HEADQUARTERS	5	4.6	77
	6	6.5	120
	7	16.4	370
	8	16.4	490
	9	1.8	80
SOLEDAD	10	2.6	130
	11	18.0	370

Figure 4.--Estimated average annual runoff at mountain fronts.

On August 19, 1978, a large flood ripped through the Post Headquarters area. Rainfall at Cox Ranch was about 10 inches during a 9-hour period, which was considerably greater than the previous maximum rainfall recorded during the 80-year record at the station (U.S. Army Corps of Engineers, written commun., 1978). On the basis of a map showing areas of equal precipitation during the storm, the total rainfall on the Post Headquarters watershed was estimated to be about 16,000 acre-feet. In 1980, another large flood of somewhat less magnitude than the 1978 flood occurred in the Post Headquarters area. After the 1978 flood, water levels began to rise in wells along the mountain front. From 1978 to 1982, Test Wells 6 and 9, boreholes 36, 37, 42, and 56, and Old Supply Well 12 had sharp increases in water levels (fig. 5). The water-level change from March 1978 through March 1982 is shown in figure 6. During this 49-month period, a total of about 8,400 acre-feet of water was withdrawn from the aquifer. The change in aquifer storage (an increase of 1,000 acre-feet) during this period was determined from figure 6 assuming a specific yield of 0.15. Therefore, total recharge for 1978-82 was about 9,400 acre-feet, which represents 29 percent of the precipitation during the two large storms. Although the storms did not contribute all of the recharge during 1978-82, the hydrographs (fig. 5) indicate that the storms may have been a significant source of recharge.

The significance of large storms on ground-water recharge also can be shown prior to the 1978 flood. The water-level change in Old Supply Well 12 compares closely to the cumulative departure from the mean maximum monthly precipitation (fig. 7) at the White Sands Missile Range stations. Mean maximum monthly precipitation was calculated by summing the largest monthly precipitation in each year from 1948 to 1982, then dividing by the number of years in that period. The water-level increase in Old Supply Well 12 may be a response to increased ground-water recharge caused by a greater than average number of large storms from 1957 to 1982. If the water-level fluctuations are a response to variations in recharge, recharge was not a steady process but probably occurred by pulses (fig. 7). If the 17-foot water-level rise measured in Old Supply Well 12 from 1978 to 1981 corresponds to 9,400 acre-feet of recharge from the 1978 and 1980 floods and if the relation is linear, the 60-foot rise from 1952 to 1982 corresponds to 33,000 acre-feet of recharge during the 31 years. This total indicates a recharge rate of about 1,100 acre-feet per year.

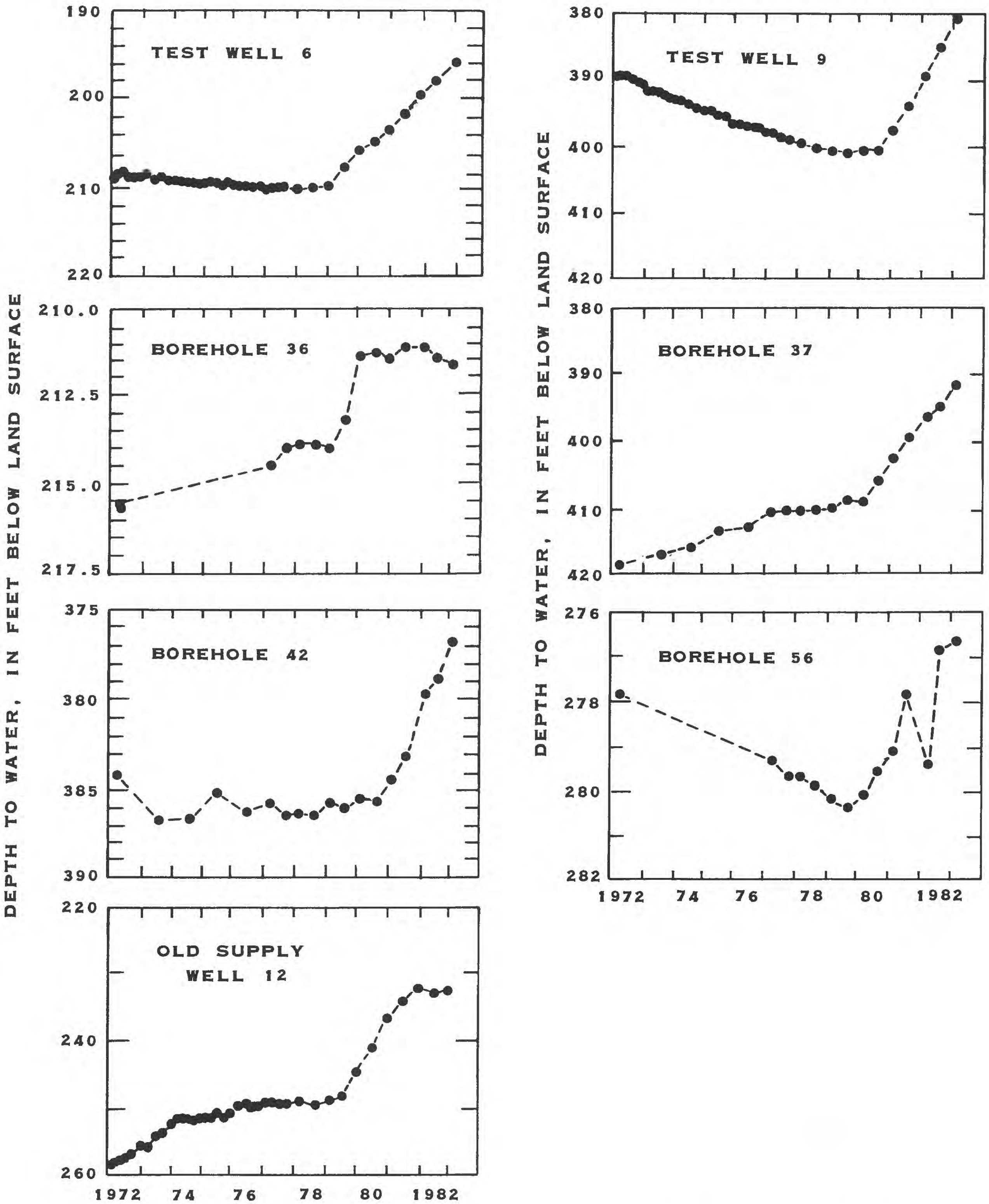


Figure 5.--Water levels in selected observation wells, 1972-83.

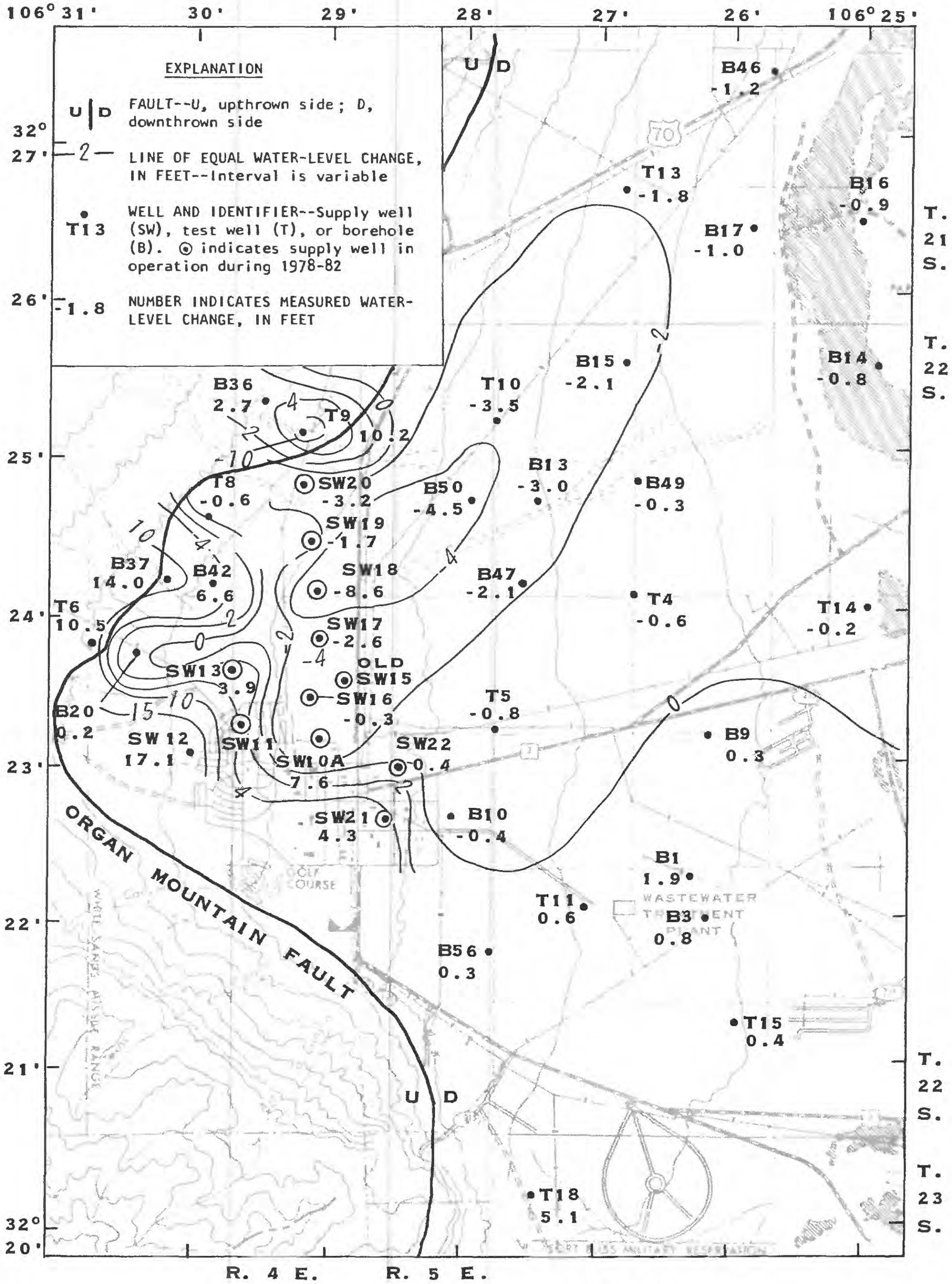


Figure 6.--Change in measured water levels, March 1978 through March 1982.

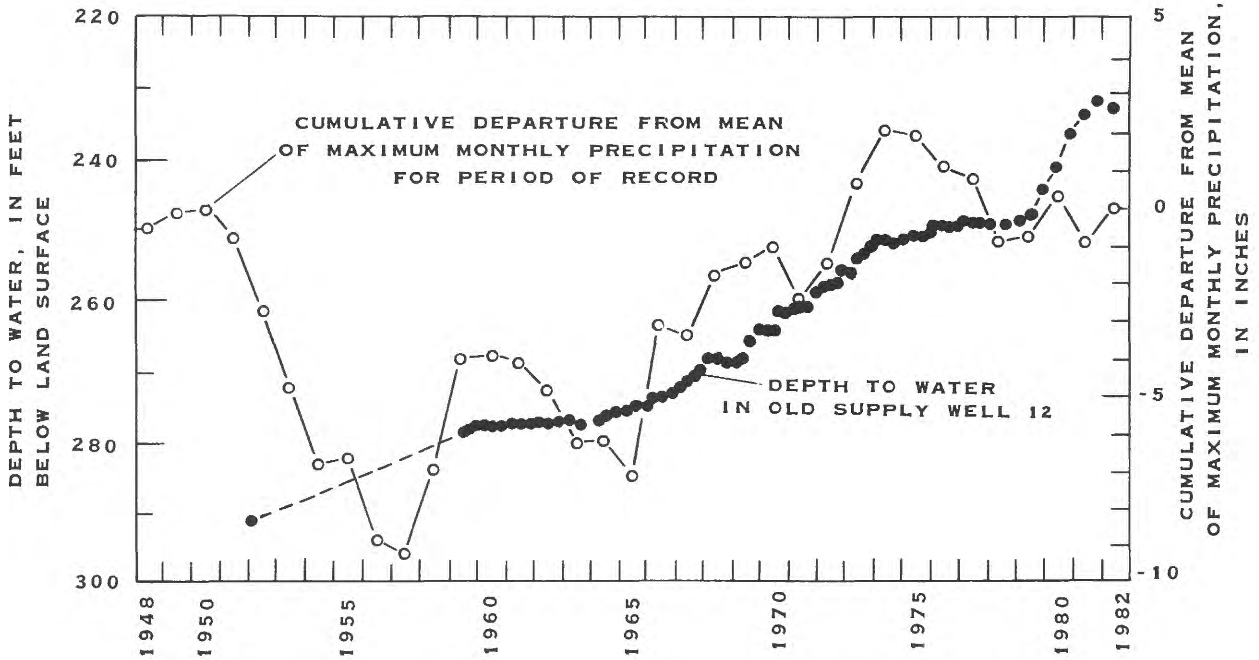


Figure 7.--Relation between water levels in Old Supply Well 12 and cumulative departure from the mean of maximum monthly precipitation, 1948-82.

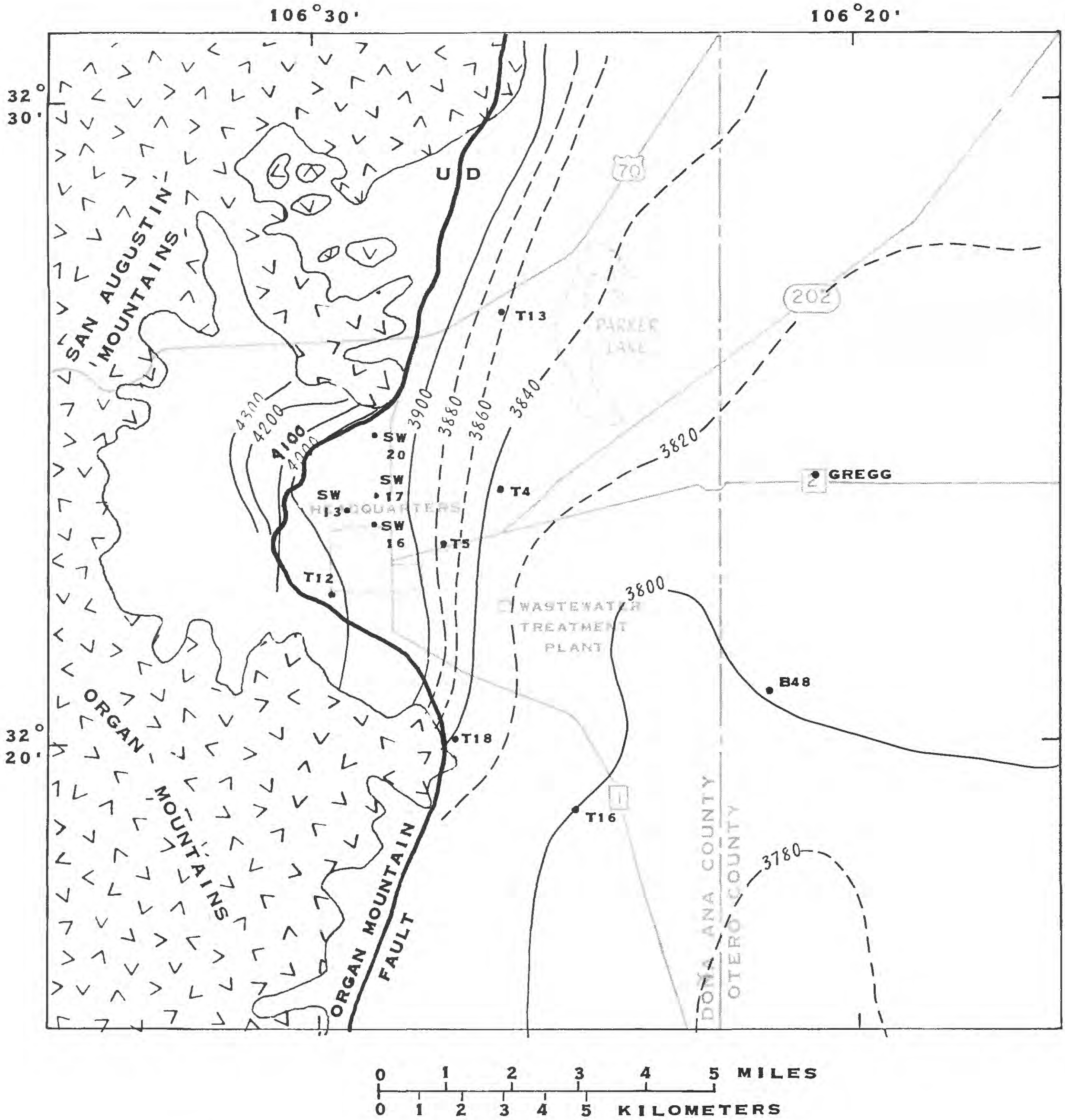
Another source of recharge to the bolson-fill aquifer is effluent from a wastewater-treatment plant. Treated effluent is discharged from a plant located about 3 miles southeast of the Post Headquarters. The effluent presently (1984) seeps into the ground through the banks of an unlined channel. Sedillo (1969, p. 84) estimated that 98 percent of the effluent infiltrates into the ground. He also estimated that 815 acre-feet of wastewater effluent seeped into the disposal channel in 1969, which is about 31 percent of the total volume of water pumped from the Post Headquarters well field in 1969. During 1976-78, between 25 and 34 percent of the total volume of water pumped was returned to the aquifer as wastewater according to records from the Facilities Engineering Directorate (written commun., 1985). Assuming that 30 percent of the total volume of water pumped from the Post Headquarters well field was returned to the ground by seepage from wastewater effluent, an average of 663 acre-feet of effluent per year may have recharged the aquifer from 1958 to 1982. Therefore, total recharge by wastewater effluent during the past 25 years could have been as large as 16,000 acre-feet.

Movement

In the study area, fresh ground water moves eastward near the mountain front. Then, because of the difference in density between fresh and saline water, the freshwater either continues to move eastward and mixes with the saline water or moves southward along the mountain front. Because intensive ground-water development has taken place since 1946, the water table today (1984) is different than under predevelopment conditions. The estimated predevelopment water table shown in figure 8 was modified from W.C. Ballance and S.M. Longwill (U.S. Geological Survey, written commun., 1968). The steep hydraulic gradient away from the mountain front indicates that ground water is recharged by streams emerging from the mountains. The hydraulic gradient flattens as the thickness of sediments and the cross-sectional area enlarge. The actual direction of ground-water flow may not be readily determined from the water-table map (fig. 8) if lenticular sand deposits along the mountain front create anisotropy with respect to hydraulic conductivity in the horizontal direction.

As freshwater moves away from the mountain fronts, some of it mixes with the more saline water in the Tularosa Basin. The mixing is caused by the movement of freshwater through dispersive aquifer sediments and by seasonal variations in the rate of freshwater recharge to the aquifer. The mixing precludes the existence of a sharp interface between fresh and saline water and instead produces a zone of dispersion between the different-density fluids.

Electric logs of test wells show a zone of dispersion across the fresh and saline water (Doty, 1968a; Lyford, 1970), indicating that mixing is occurring at the present time (1984). The width of the zone of dispersion is controlled by dispersive properties of the sediments and chemical diffusion and probably has been enhanced by intensive ground-water withdrawals. However, logs of wells in the Soledad Canyon area that are far removed from the withdrawals also show a zone of dispersion (Wilson and Myers, 1981). The dispersion may be caused by annual variations in recharge or long-term paleoclimatic changes.



EXPLANATION

(symbol)	OUTCROP AREA OF IGNEOUS BEDROCK
(symbol)	AREA OF BOLSON-FILL AQUIFER
U/D	FAULT--U, upthrown side; D, downthrown side
3800	APPROXIMATE WATER-TABLE CONTOUR--Shows approximate altitude of predevelopment water table. Contour interval is 100 feet with dashed supplemental 20-foot contours. Datum is sea level
T16 •	SELECTED WELL AND IDENTIFIER--Shown for orientation purposes; wells do not represent location for measurement of predevelopment water levels

Figure 8.--Approximate altitude of predevelopment water table.

Discharge

Prior to withdrawals from the Post Headquarters well field, there were no large quantities of ground-water discharge in the study area. A few small springs and shallow wells were used by Indians, miners, and early settlers (Kelly and Hearne, 1976, p. 20), but most ground water flowed through the study area to the south and east. This water moved into the Hueco Bolson and probably discharged along the Rio Grande, about 40 miles south of the study area, or continued to flow underground into Mexico.

Large-scale ground-water development from the freshwater aquifer began in 1946 with the construction of a series of closely spaced wells less than 400 feet deep in sections 31 and 32, T. 22 S., R. 5 E. (Kelly and Hearne, 1976, p. 22). This well field, called the old well field, was abandoned by 1954 (table 1 and fig. 2). Supply well 10, the first of the newer wells in the center of the reentrant, was drilled in 1948. Since that time, 13 additional supply wells (wells 10A through 22) were drilled. A summary of withdrawals from the Post Headquarters wells is shown in table 1 and figure 9. In 1982, about 2,170 acre-feet of water was withdrawn by 10 supply wells in the Post area. These wells, less than 1,000 feet deep, are completed less than 500 feet below the water table. The potential yield of water from the aquifer below 1,000 feet is largely unknown, but probably is less than at shallow depths due to compaction of the sediments at depth.

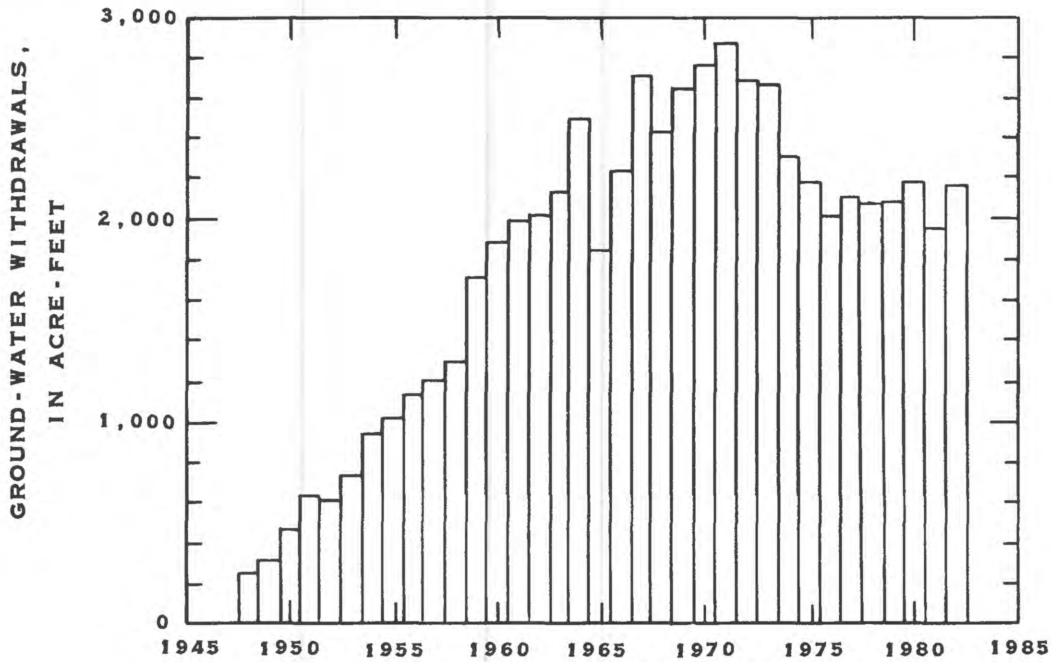


Figure 9.--Ground-water withdrawals from Post Headquarters well field, 1948-82.

Table 1.—Ground-water withdrawals from the Post Headquarters well field, 1948-82

[All withdrawals are reported in thousands of gallons]

Year	Old well field	Supply well [1]10A	11	12	13	14	15	16	17	18	19	20	21	22	Totals
1948	70,322	11,978	-	-	-	-	-	-	-	-	-	-	-	-	82,300
1949	65,905	38,612	-	-	-	-	-	-	-	-	-	-	-	-	104,517
[2]1950	65,000	65,000	22,700	-	-	-	-	-	-	-	-	-	-	-	152,700
1951	45,320	79,584	81,649	-	-	-	-	-	-	-	-	-	-	-	206,553
1952	8,781	65,072	69,653	27,823	29,892	-	-	-	-	-	-	-	-	-	201,221
1953	4,334	77,986	76,965	43,723	37,960	-	-	-	-	-	-	-	-	-	240,968
1954	6,394	70,913	50,324	52,388	41,840	64,169	21,158	-	-	-	-	-	-	-	307,186
1955	-	26,713	32	4,968	18,717	127,379	104,067	49,911	-	-	-	-	-	-	331,787
1956	-	59,535	31,228	3,414	54,060	77,267	81,149	61,604	-	-	-	-	-	-	368,257
1957	-	62,645	47,797	-	39,365	86,777	88,388	69,801	-	-	-	-	-	-	394,773
1958	-	59,948	55,075	-	62,149	21,761	124,651	98,223	-	-	-	-	-	-	421,809
1959	-	75,274	43,368	-	66,860	126,454	128,088	120,785	-	-	-	-	-	-	560,829
1960	-	74,236	75,448	-	62,970	119,866	150,482	131,297	-	-	-	-	-	-	614,299
1961	-	36,707	71,542	-	42,992	107,598	143,740	134,485	114,587	-	-	-	-	-	651,652
1962	-	16,352	56,776	-	40,511	55,094	119,819	201,362	169,352	-	-	-	-	-	659,266
1963	-	41,390	65,798	-	51,473	72,578	107,126	193,082	165,051	-	-	-	-	-	696,498
1964	-	192,828	56,640	-	29,008	47,599	142,293	186,324	149,656	11,225	-	-	-	-	815,574
1965	-	114,694	33,321	-	20,093	-	28,349	43,488	36,609	45,059	117,223	162,653	-	-	601,489
1966	-	102,964	35,502	-	25,434	-	44,319	57,244	57,992	87,484	129,369	189,602	-	-	729,910
1967	-	116,912	59,387	-	25,422	-	91,104	96,292	82,234	103,158	141,710	167,294	-	-	883,513

Table 1.—Ground-water withdrawals from the Post Headquarters well field, 1948-82—Concluded

Year	Old well field	Supply well [1]10A	11	12	13	14	15	16	17	18	19	20	21	22	Totals
1968	-	85,379	60,504	-	27,603	-	77,886	95,909	95,695	96,969	116,810	135,781	-	-	792,536
1969	-	84,540	84,637	-	41,944	-	80,528	115,375	121,469	120,053	70,443	144,002	-	-	862,991
1970	-	145,010	63,002	-	17,253	-	30,061	111,005	128,425	94,360	188,643	122,092	-	-	899,851
1971	-	163,786	67,379	-	43,357	-	7,940	136,763	150,673	97	188,610	180,413	-	-	939,018
1972	-	140,862	59,442	-	11,295	-	4,493	116,805	126,826	28,689	193,769	192,009	-	-	874,190
1973	-	142,023	81,826	-	12,234	-	3,171	95,304	114,796	80,749	182,216	158,543	-	-	870,862
1974	-	127,944	56,339	-	443	-	38,986	49,672	107,802	81,362	150,755	148,933	-	-	757,226
1975	-	97,916	61,160	-	7,335	-	61,318	74,878	104,489	71,427	87,813	148,244	-	-	714,580
1976	-	129,558	55,073	-	20,691	-	60,870	63,388	66,130	60,025	112,861	89,539	-	-	657,655
1977	-	23,067	53,220	-	43,143	-	52,305	63,023	98,515	87,110	183,899	99	13,873	18,885	688,039
1978	-	38,845	41,987	-	20,323	-	50,503	58,536	77,788	53,170	124,447	70,492	39,317	105,011	680,414
1979	-	92,505	37,212	-	19,094	-	47,821	65,230	101,772	6,530	136,474	88,178	2,113	84,867	681,796
1980	-	127,867	42,251	-	21,755	-	17,381	37,052	101,591	51,026	163,384	98,052	26,635	2,055	712,909
1981	-	100,598	51,416	-	21,066	-	-	59,498	87,415	80,117	139,620	45,754	498	52,979	638,961
1982	-	77,822	73,252	-	18,102	-	-	73,497	112,085	80,039	124,538	102,482	12,143	32,675	706,653

[1] Reported withdrawals were from well 10 from 1948 to 1962, from wells 10 and 10A combined for 1963, and from well 10A from 1964 to 1982.
[2] 1,950 withdrawals are estimated.

Since 1948, about 63,000 acre-feet of ground water has been withdrawn from the bolson-fill aquifer at the Post Headquarters well field. These withdrawals represent a new discharge from the hydrologic system that previously was in a state of dynamic equilibrium where recharge equaled discharge. The present result of this new discharge is a decline in ground-water levels as water is removed from aquifer storage. Water levels will continue to decline until a new equilibrium can be established by increasing the natural recharge or decreasing the natural discharge. The principles required to establish a new equilibrium in the Alamogordo area as described by Garza and McLean (1977, p. 19) apply equally well to the Post Headquarters well field:

"Steady-state conditions in a pumping well field begin when the cone of depression has reached the recharge or natural discharge areas, and when the recharge is increased or the natural discharge is decreased an amount equal to that pumped. In the Tularosa basin, however, the natural recharge cannot be increased by lowering the water table and the natural discharge cannot be decreased since the discharge area is a great distance from the pumping well field. Thus, steady-state conditions cannot be achieved and the effects of pumping on water levels in and near the well field are independent of the natural recharge and natural discharge. The natural recharge of freshwater is important, however, as it would otherwise be replaced by saline water moving eastward into the well field."

In the Post area, the water table is recharged from arroyos that are several hundred feet above the water table. Therefore, lowering the water table cannot induce any additional recharge from these arroyos. However, natural discharge from the Tularosa Basin will, in time, be decreased by an amount equal to the amount of water pumped. That time is so long, however, that for practical purposes, ground water will continue to be withdrawn entirely from storage. For example, assuming an ideal unbounded aquifer with a transmissivity of 1,000 feet squared per day, a specific yield of 0.15, and a withdrawal rate of 1,700 acre-feet per year, a drawdown of 1 foot at the Rio Grande about 40 miles to the south would take more than 1,000 years to achieve. Different assumptions would produce different answers, but this example illustrates the magnitude of the time involved before the natural discharge would be affected and new equilibrium could begin to be established.

Water-Level Fluctuations

Most supply and test wells in the Post Headquarters area are completed throughout tens or hundreds of feet of saturated sediments. Therefore, water-level fluctuations in these wells represent a composite change in hydraulic head throughout the units in which the well is completed. This composite change may be greater or less than the change in water-table altitude depending upon which units are contributing water to the wells. However, water-level changes in most test wells completed throughout a wide range of sediments are similar to changes observed in nearby boreholes that are completed near the water-table surface. The correlation indicates that water-level changes in most test wells probably are fairly representative of the change in water-table altitude at those locations.

Water levels in the study area have, in general, declined since large-scale ground-water withdrawals began in 1946. Kelly (1973, p. 46 and 48) showed the change in water levels from 1949 to 1964 and from 1964 to 1972 in the Post area. By 1964, water levels had declined more than 75 feet near Old Supply Well 15 and Supply Well 16, although the area of influence created by the pumping was fairly limited. Kelly's maps for 1964-72 show that water levels continued to decline and the area of influence expanded except in the southern part of the reentrant.

The water-level change from 1972 to 1978 is shown in figure 10. Because additional boreholes were available after 1972, a more complete picture of the water-level change could be compiled than that compiled by Kelly (1973). The same general trend in water-level fluctuations continued during this period with declines in nearly all wells except in the southern part of the reentrant, where water-level increases occurred. Water levels increased in Test Well 11 and in boreholes 1-5 because of recharge by effluent from the nearby wastewater-treatment facility.

The rising water level in the southern part of the reentrant prior to the 1978 flood is clearly shown at Old Supply Well 12. Water levels in Supply Wells 11 and 13 have shown some long-term increase, but the trend is less apparent. The rising water levels might be explained by a larger than normal amount of recharge from precipitation, return flow from leaky sewer lines, and lawn watering in the Post Headquarters area. Also, the hydraulic connection could be poor between the sediments in which Old Supply Well 12 is completed and the remainder of the reentrant or the well could be plugged at depth and only be open to the shallow part of the aquifer. More test wells are needed in the southern part of the reentrant to better understand the connection between water-bearing zones.

With more complete well control after 1972, the large area of influence of the well-field withdrawals can be seen (fig. 10). However, this area does not correspond to the area throughout which water moves toward the well field at the water table. The area of diversion to the well field is much smaller (fig. 11) because of the sloping surface of the predevelopment water table. If the predevelopment water table had been flat, the area of influence would coincide with the area of diversion. A complication to this simplified concept of ground-water flow in the vicinity of the well field arises if sand lenses in which the supply wells are completed are not well connected hydraulically. The hydraulic head in sands isolated vertically from other parts of the bolson fill could be significantly different than the hydraulic head at the water table. Therefore, ground-water movement in these sands could be significantly different than is indicated in figure 11.

Water-level changes from 1978 to 1982 were somewhat different than in previous periods: for example, the rise in water levels along the mountain front in Test Wells 6 and 9, boreholes 36, 37, 42, and 56, and Old Supply Well 12 (figs. 5 and 6). The water-level rise probably is the result of recharge from floods.

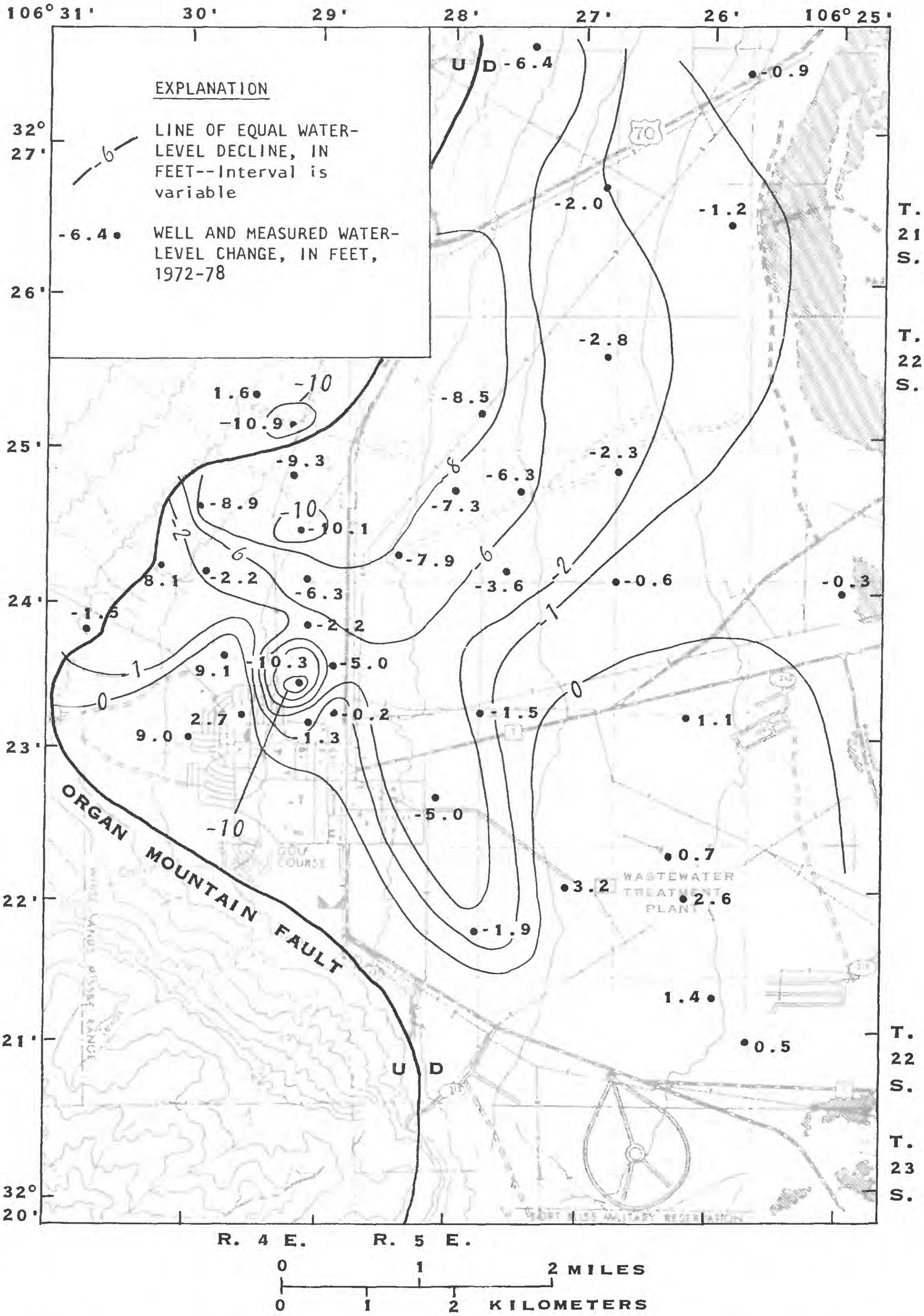


Figure 10.--Decline in measured water levels, March 1972 through March 1978.

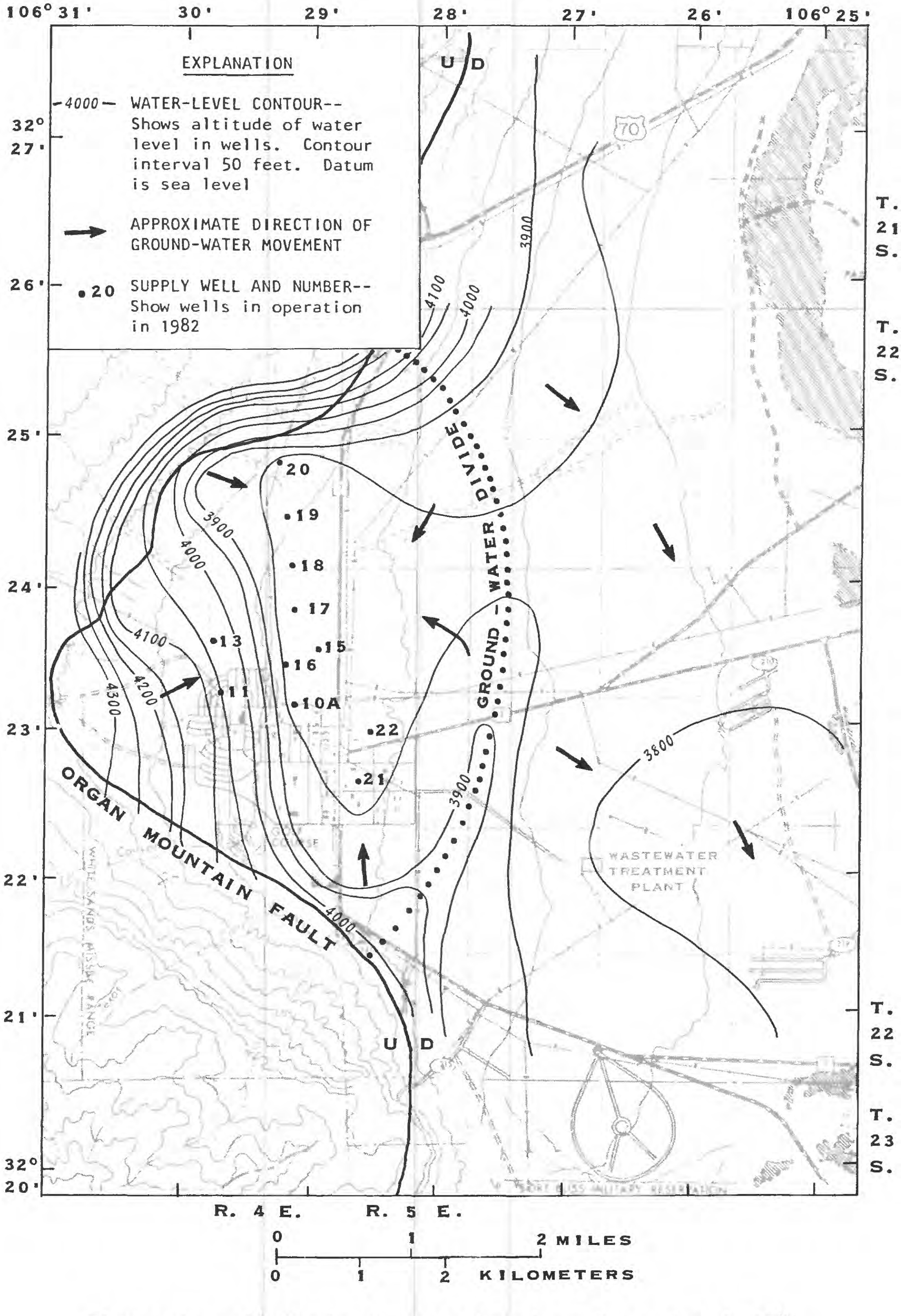


Figure 11.--Altitude of water-level surface, March 1982.

Aquifer Properties

The hydraulic properties of the bolson-fill aquifer control its ability to transmit and store water. Because of the heterogeneity of the sediments, the hydraulic properties vary greatly. Aquifer tests have been conducted at 28 wells to determine transmissivity of the upper 500 feet of saturated bolson fill (Kelly, 1973, p. 12; Wilson and others, 1978). Because of the short duration of most tests and the large difference between vertical and horizontal hydraulic conductivity of the aquifer, the transmissivity values calculated from these tests probably represent the transmissivity of sandy zones in which the wells were completed. Horizontal hydraulic-conductivity values were calculated by dividing the transmissivity determined from aquifer tests by the screened interval of each well. Horizontal hydraulic conductivity in the study area has a median value of 6.8 feet per day and ranges between 0.1 and 208 feet per day. Warren and Price (1961) suggested that the geometric mean of hydraulic-conductivity values for a heterogeneous material may be the best "average" number to approximate the equivalent hydraulic conductivity of a homogeneous material. The geometric mean of the 28 hydraulic-conductivity values is 1.6 feet per day.

The spatial variability of horizontal hydraulic conductivity is shown in figure 12. In general, the smallest values of hydraulic conductivity are closest to the mountain front and the largest values are on the northern and southern ends of the reentrant. The distribution of hydraulic conductivity does not correspond well to the percent-sand map of Kelly (1973, p. 20). The poor correlation probably is caused by variable degrees of sorting of the sand, poor along the mountain front and better with greater distance from the source area.

The long-term storage of the aquifer is represented by specific yield. A specific yield of 0.15 was estimated from water-level decline maps by Herrick (1960b, p. 57), W.C. Ballance and S.M. Longwill (U.S. Geological Survey, written commun., 1968), and Kelly and Hearne (1976, p. 37). Water-level declines for 1972-78 also indicate that the specific yield of the upper 500 feet of saturated sediments is about 0.15. This estimate is based on ground-water withdrawals of about 13,800 acre-feet and an estimated volume of dewatered sediments of 66,600 acre-feet. Return flow from the wastewater-treatment plant was estimated to be about 30 percent of the ground-water withdrawals. Therefore, net ground-water withdrawals were about 9,700 cubic feet, which is almost 15 percent of the volume of dewatered sediments.

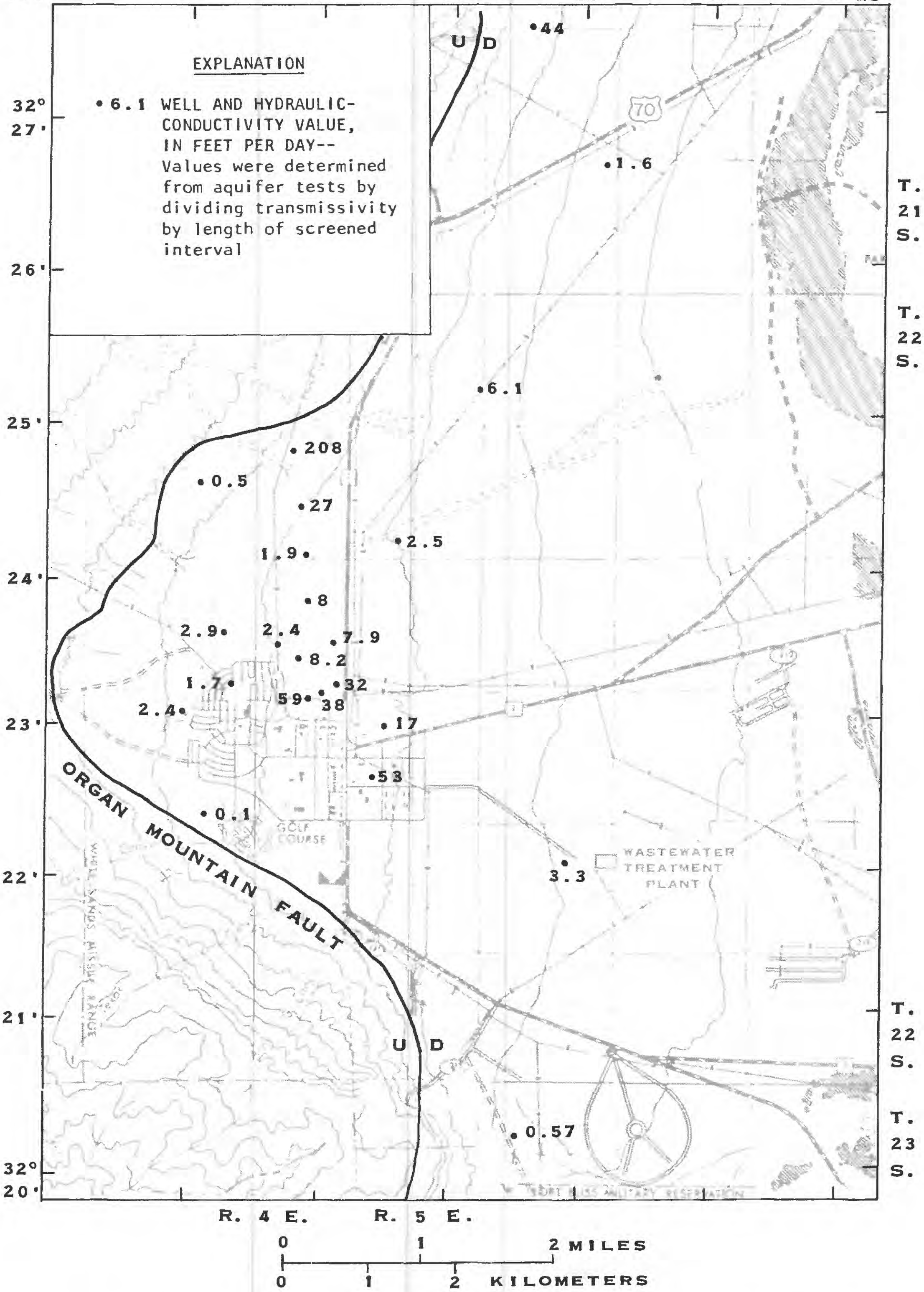


Figure 12.--Hydraulic-conductivity values determined from aquifer tests.

Hydraulic properties of the bolson fill at depths greater than about 500 feet below the water table are largely unknown. Kelly and Hearne (1976, p. 36) discussed the existence of a continuous clay layer at a depth of about 1,000 feet that they interpreted to be the bottom of the freshwater aquifer. Whether this clay exists as a continuous, effective barrier to ground-water flow is unknown. Seismic data indicate that the bolson-fill sediments probably are more compacted and cemented at depth and therefore are less permeable. An estimated empirical relation between seismic velocity and hydraulic conductivity is shown in figure 13. Seismic velocities measured from 0 to about 6,000 feet below land surface at Test Well 14 were used to estimate the approximate decrease in hydraulic conductivity with depth (fig. 14). The effect of increasing ground-water temperatures with depth on hydraulic conductivity was not evaluated. An increase of water temperature from 50 to 100 degrees Fahrenheit would approximately double the hydraulic conductivity.

The hydraulic conductivity of the bolson fill in the vertical direction is unknown. Because clay and silt layers are interbedded horizontally with sand lenses, the vertical hydraulic conductivity would be less than in the horizontal direction. Hearne (1980, p. 19) estimated vertical to horizontal anistropy with respect to hydraulic conductivity in similar geologic materials to be between 0.1 and 0.001.

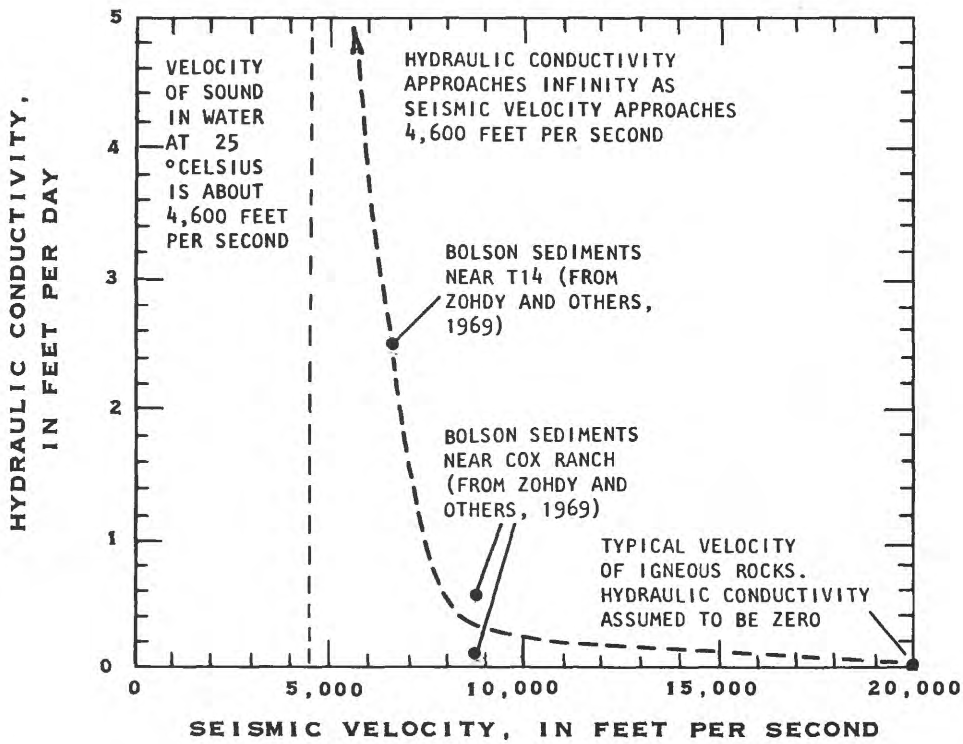


Figure 13.--Approximate relation between seismic velocity and hydraulic conductivity.

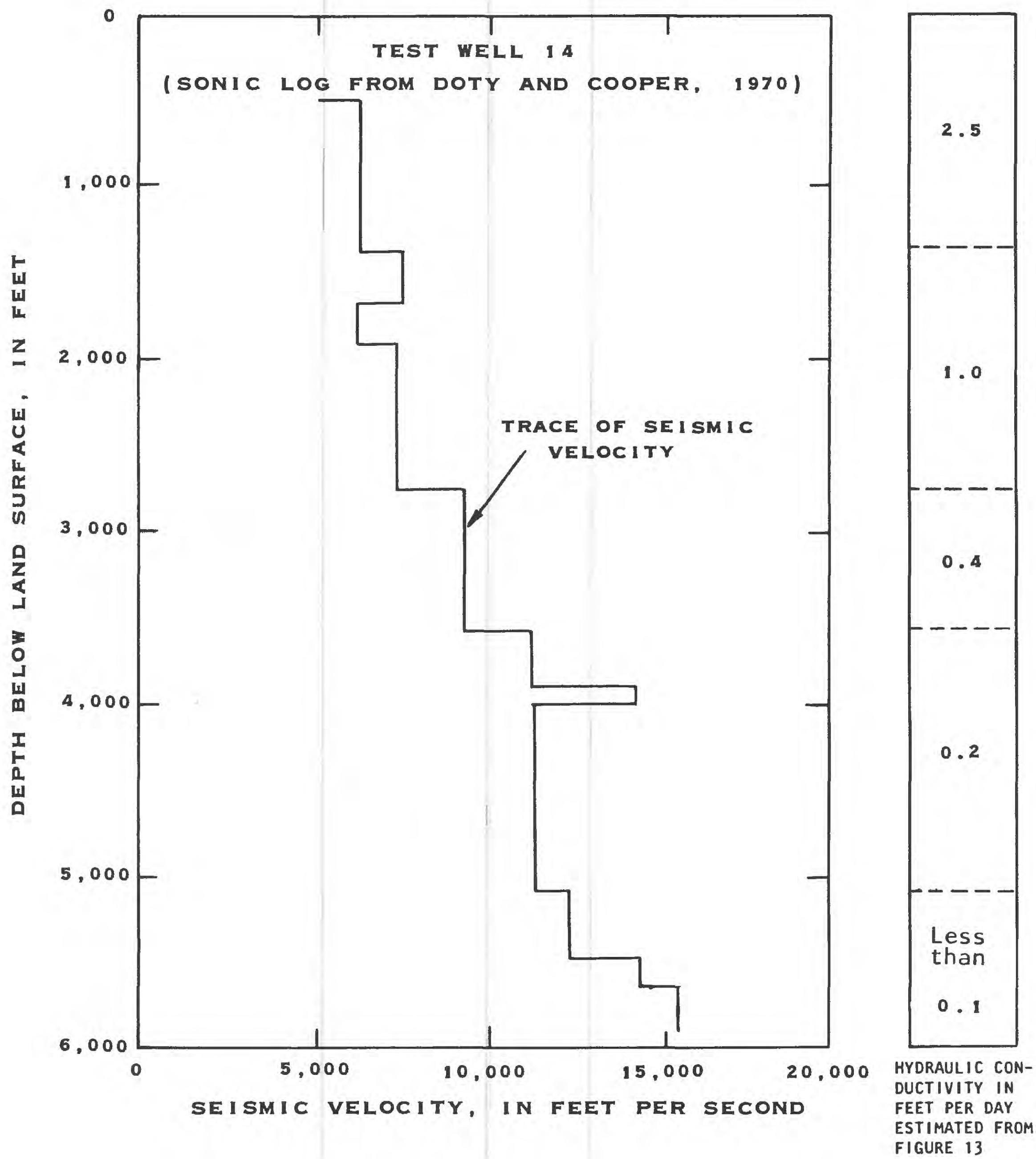


Figure 14.--Seismic velocity and estimated hydraulic conductivity of geologic material penetrated by Test Well 14.

Water Quality

The quality of ground water varies from freshwater having a dissolved-solids concentration of less than 300 milligrams per liter to brine containing greater than 100,000 milligrams per liter. The freshwater roughly parallels the mountain front as mapped by Kelly (1973, pl. 31). The thickness of freshwater below the Post Headquarers well field was estimated to be 1,500 to 1,800 feet (McLean, 1970, p. 49; Kelly, 1973, fig. 31). Kelly and Hearne (1976, p. 82) speculated that ground water probably contains large concentrations of dissolved solids below the clayey sediments present at a depth of about 1,000 feet.

The water quality in the upper 500 feet of saturated bolson fill can be divided into three general water types as shown in figure 15. Ground water containing calcium and bicarbonate as the principal ions occurs in the reentrant in a mitten-shaped zone. The calcium bicarbonate water contains the smallest concentration of dissolved solids and probably represents the area where most ground-water recharge takes place. The shape of the calcium bicarbonate zone corresponds to the pathways of major drainages emerging from the mountain front. Surrounding the calcium bicarbonate water is a zone of water containing a greater concentration of dissolved solids and a sulfate-to-bicarbonate ratio greater than 0.5. This calcium bicarbonate sulfate water probably is caused by increased dissolution of gypsum in areas where the residence time of ground water in the aquifer is long (Freeze and Cherry, 1979, p. 241) or by dissolution of pyrite from intrusive rocks and adjacent alluvium along the mountain front. Water along the mountain front in the vicinity of the fault near Test Wells 8, 9, and 12 contains greater concentrations of sulfate, which could indicate that the majority of freshwater recharge occurs east of the fault. Water in the easternmost part of the study area is a sodium chloride sulfate brine characteristic of most of the Tularosa Basin.

Changes in the quality of water in the Post Headquarters well field have been indicated by specific-conductance measurements and water-quality analyses. The largest increase in specific conductance has occurred at Test Well 4 and Supply Wells 11, 13, 16, and 20 (fig. 16). The change in quality corresponds to an increase in dissolved-solids concentration of as much as 200 milligrams per liter. Smaller increases were measured at Test Well 9, Supply Well 10A, and Old Supply Wells 12 and 15 (fig. 16). Chemical analyses of water plotted on Piper diagrams indicate that the water is becoming enriched in sulfate, as shown by samples collected from Supply Well 13 (fig. 17). An addition of sodium chloride brine to water withdrawn at the supply wells is not indicated.

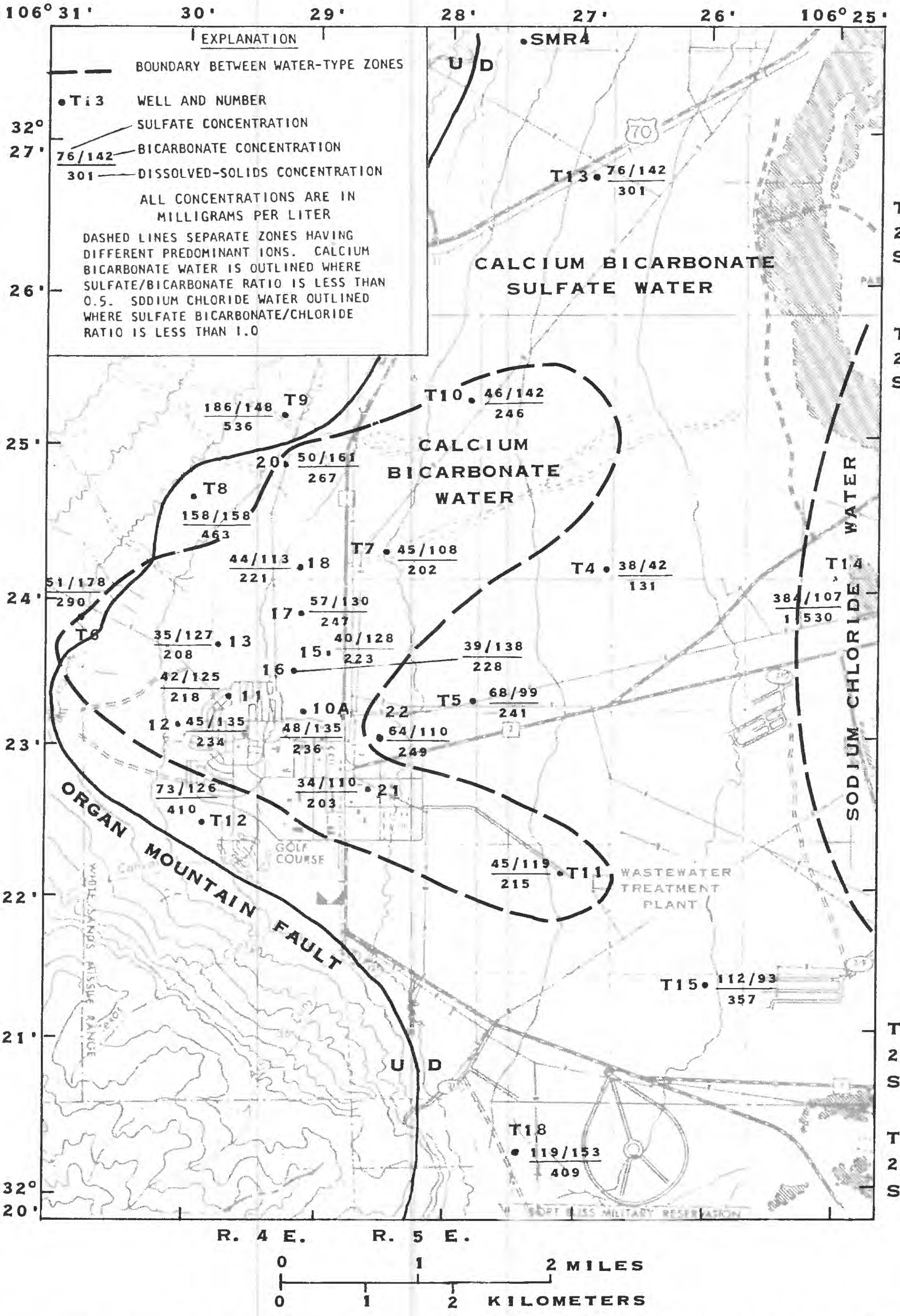


Figure 15.--Sulfate, bicarbonate, and dissolved-solids concentrations of water from selected wells.

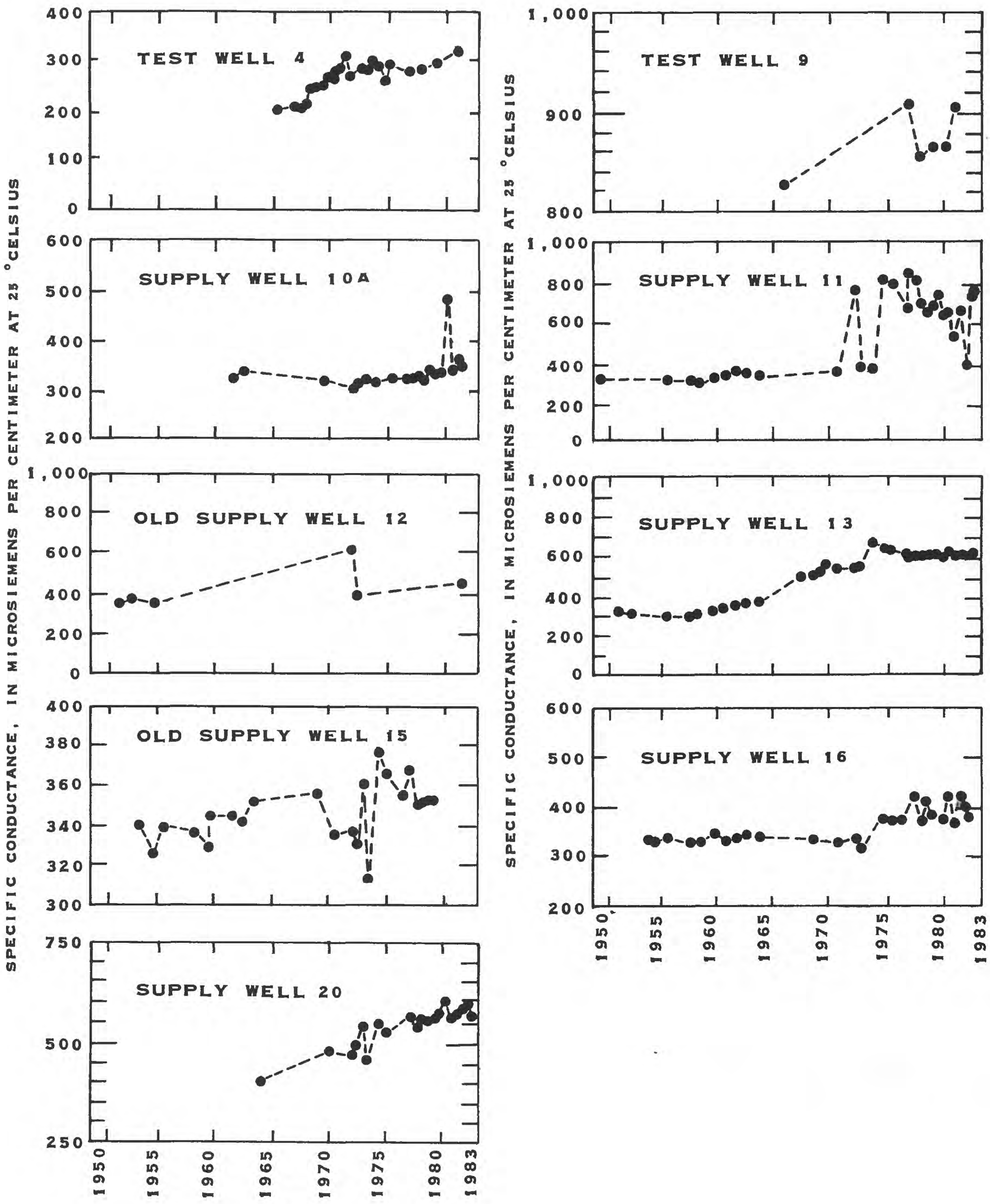


Figure 16.--Specific conductance of water from selected wells, 1950-83.

CHEMICAL CONSTITUENTS

SO_4 - Sulfate
Cl - Chloride
Ca - Calcium
Mg - Magnesium
Na+K - Sodium + Potassium
CO_3+HCO_3 - Carbonate + Bicarbonate

EXPLANATION

SAMPLE NUMBER	DATE OF COLLECTION	DISSOLVED-SOLIDS CONCENTRATION, IN MILLIGRAMS PER LITER
1	3-11-52	225
2	5-3-56	208
3	3-16-59	216
4	8-15-60	224
5	7-13-61	233
6	6-4-62	240
7	5-20-63	250
8	6-18-64	254
9	5-15-68	341
10	5-13-69	359
11	6-16-71	396
12	8-17-77	402
13	9-6-78	391
14	2-17-81	424

■ WATER FROM SMALL MISSILE RANGE (SMR4)

▲ BRINE FROM TEST WELL 14

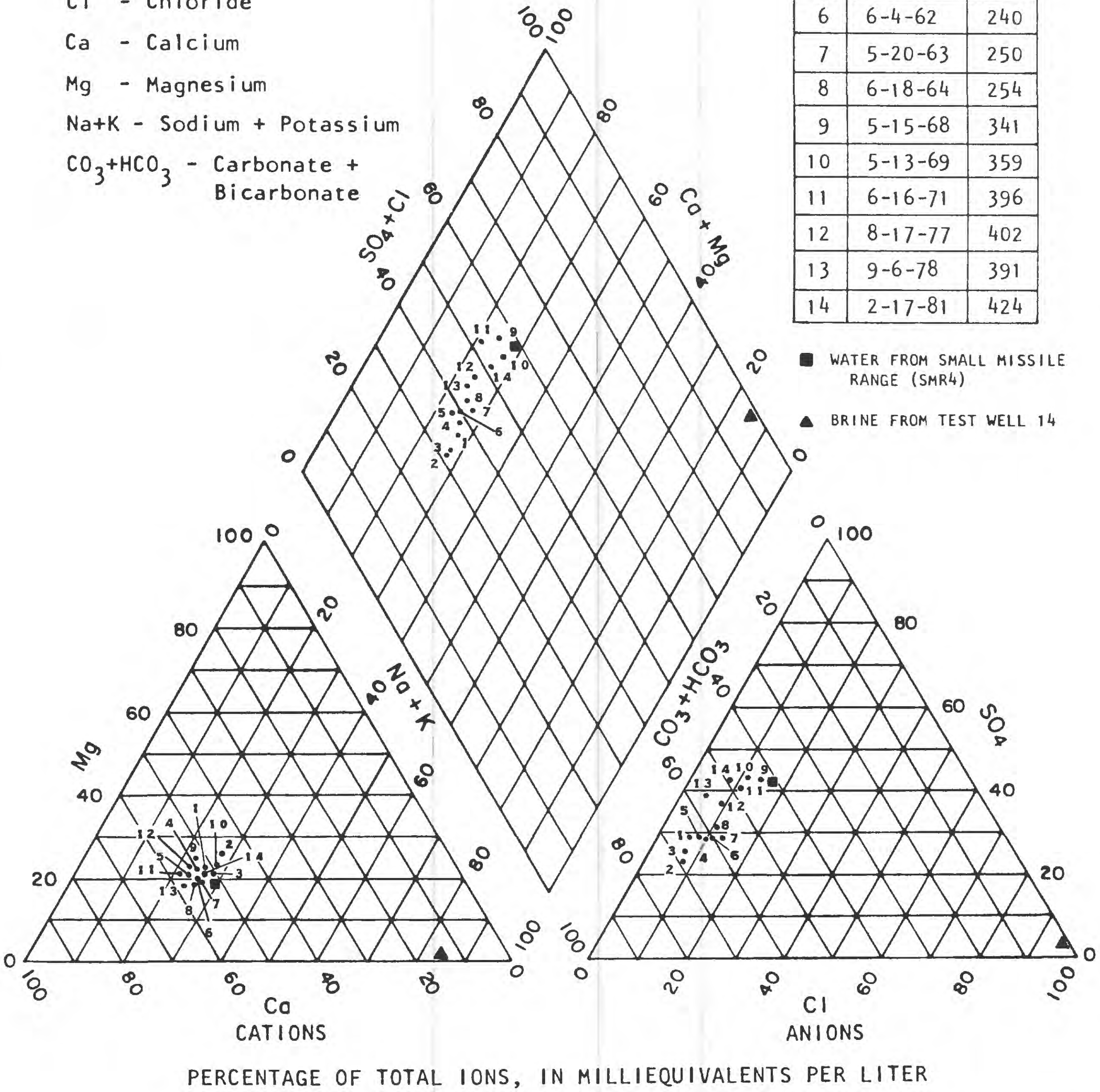


Figure 17.--Change in chemical composition of water from Supply Well 13, 1952-81.

The change in water quality could be caused by the addition of water containing greater concentrations of sulfate from several sources. The most likely source is lateral movement of ground water from the adjacent calcium bicarbonate sulfate zone as exemplified by the composition of water from the Small Missile Range at well SMR4 (figs. 15 and 17). The water-table map indicates that water is moving radially toward the well field from this zone. Leakage of water from clay layers interbedded within the saturated sands also may be a source of sulfate-rich water. However, if this leakage were taking place, the water chemistry probably would show an increase in the relative concentration of sodium with respect to calcium because of ion-exchange reactions occurring within the clay layers. Because a relative increase in sodium has not been observed, the clay layers probably are not the primary source of the more mineralized water unless the exchange sites on the clays are already full of divalent cations. Although the quality of water below the well field is unknown, sulfate-rich water could be moving upward from below the well field.

Water-quality changes at Test Well 14 can be explained by mechanisms other than encroachment of more mineralized water. Samples from Test Well 14 show that the specific conductance of water at 200 feet below land surface changed from about 500 to 1,500 microsiemens during 1972-80, whereas water sampled at 300 feet changed from about 2,800 to 1,500 microsiemens during 1970-82 (fig. 18). This change possibly was caused by mixing of water with different densities in the wellbore, which might have taken place gradually each time the logging tool was lowered into the well.

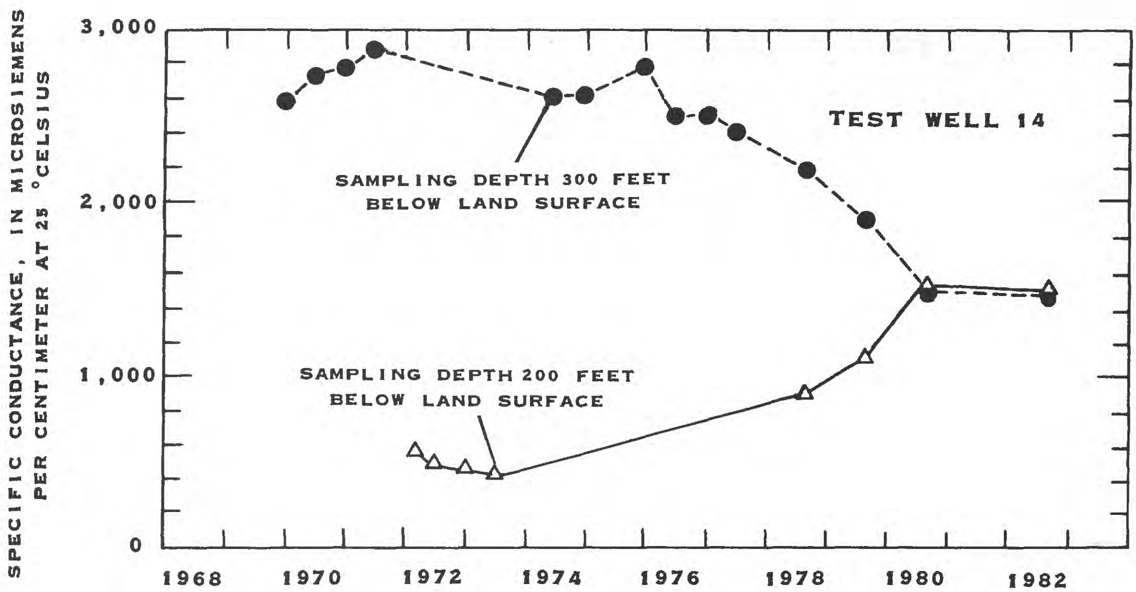


Figure 18.--Specific conductance of water in Test Well 14.

SIMULATED WATER-LEVEL CHANGES

Introduction and Limitations of the Ground-Water Flow Model

A ground-water model is a simplified analog that represents the behavior of a real hydrologic system. The ground-water model used in this report is composed of a computer program that solves finite-difference equations that describe ground-water flow and a conceptual model that includes the boundary conditions and hydraulic properties that represent the essence of the hydrologic system. The fact that the computer program, the conceptual model, and the combination of the two are all called models can be confusing. In this report, the term ground-water model refers to the combination of the computer program and conceptual model.

A carefully constructed ground-water model can be useful in investigating the physical properties of the hydrologic system. By comparing simulated and measured aquifer response, a model sometimes can be used to determine which conceptual models and estimated aquifer properties are incompatible with the measured response. The investigator usually proceeds in a trial-and-error manner to adjust the conceptual model until an acceptable concept is developed that is not rejected by observations of the real system. In this way, the model can be used to estimate hydraulic properties that are poorly known. However, in practice, generally so few hydraulic properties are known with any confidence that a large number of very different conceptual models can be made to closely simulate measured response of an aquifer. An acceptable match to the measured aquifer response does not imply that a uniquely correct model has been constructed. There are many "acceptable" models that can be made using hydraulic properties and boundary conditions that are physically absurd. Therefore, during the trial-and-error adjustment procedure, the working hypotheses are restricted to the use of "reasonable values" for all physical parameters. Because the exact physical properties of the hydrologic system are not well known, the definition of reasonable values rests to a great extent on the experience and judgment of the modeler.

Predicting future aquifer response also is an important use of ground-water models. However, for most models, the accuracy of the prediction is unknown. Fortunately, a subjective degree of confidence can be obtained from comparison of simulated and measured aquifer response during the adjustment stage of the modeling procedure.

Model Construction

The ground-water flow model is constructed by assigning boundary conditions and hydraulic properties to the computer program that represent an initial concept of the ground-water flow system. This process involves selecting an appropriate computer program, dividing the study area into finite-difference cells, setting boundary conditions that define the physical geometry and flow of water into and out of the model, and assigning hydraulic properties to each cell.

Computer Program

The flow of ground water in three dimensions can be described by the partial differential equation (McDonald and Harbaugh, 1984, p. 7):

$$\frac{\partial}{\partial x}\left(K_{xx}\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_{yy}\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_{zz}\frac{\partial h}{\partial z}\right) - W = S_s\frac{\partial h}{\partial t} \qquad (2)$$

where x, y, and z = cartesian coordinates aligned along the major axes of hydraulic conductivity Kxx, Kyy, Kzz;

h = the potentiometric head;

W = a volumetric flux per unit volume of water (source and sink terms);

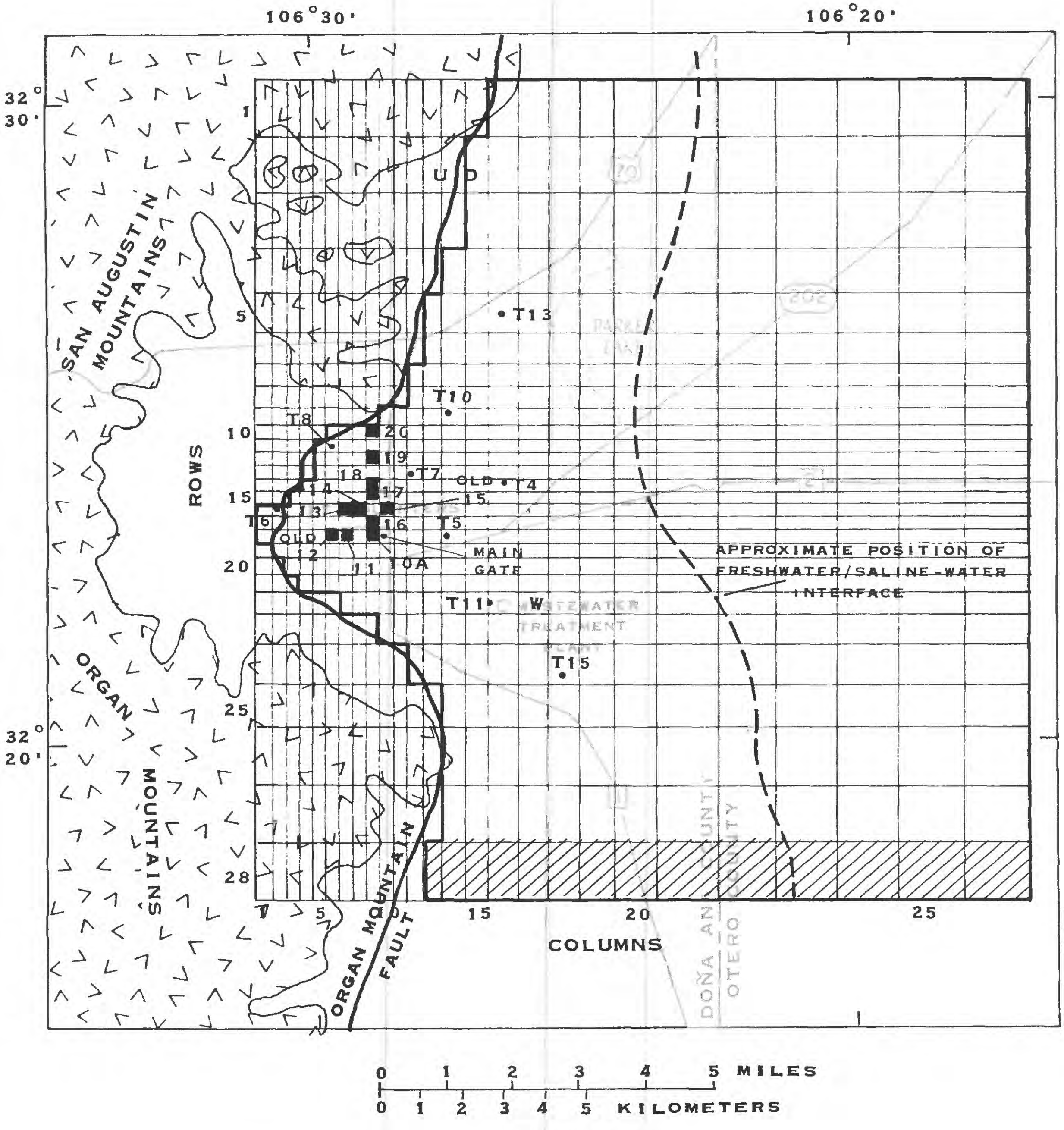
Ss = specific storage; and

t = time.

The derivatives in the differential equation of ground-water flow can be approximated by finite differences. The resulting algebraic equations are then solved simultaneously using a computer program documented by McDonald and Harbaugh (1984). The strongly implicit method was used by the program as the solution algorithm. The finite-difference approach necessitates that the study area be divided into discrete rectangular cells at which the physical properties of the aquifer are assigned.

Finite-Difference Grid

The study area was divided into a finite-difference grid that has 28 rows and 26 columns with rows oriented east-west and columns oriented north-south (fig. 19). The orientation was chosen to parallel the suspected principal directions of anisotropy with respect to hydraulic conductivity. The horizontal dimensions of the cells were varied so that the smallest cells, 1,042 feet by 1,042 feet, were in the vicinity of the Post Headquarters well field where the largest water-level declines have occurred. The cell dimensions were increased away from the well field to save computation time while allowing arbitrary boundaries to be far removed from the well field. The study area was divided vertically into six layers (fig. 20). The layers were chosen more to show the three-dimensionality of the flow system than to represent distinct geologic units. The upper three layers, each 200 feet thick, roughly correspond to the thickness of the aquifer simulated in two dimensions by Kelly and Hearne (1976). All supply wells are completed in the upper three layers (fig. 20). The location of specific cells is designated by layer, row, and column position. For example, cell 3-21-10 is located in layer 3, row 21, column 10.



EXPLANATION

- IGNEOUS ROCKS
- BOLSON-FILL AQUIFER
- CONSTANT-HEAD BOUNDARY (STEADY STATE), CONSTANT-FLUX BOUNDARY (TRANSIENT)
- NO-FLOW BOUNDARY--No ground-water flow simulated outside of this boundary
- ■ SUPPLY WELL--Shows cell location where simulated and measured water levels were compared
- • TEST WELL--Shows cell location where simulated and measured water levels were compared
- **W** CELL WHERE WASTEWATER-RETURN FLOW WAS SIMULATED

Figure 19.--Model grid, hydrologic boundaries, and location of supply and test wells.

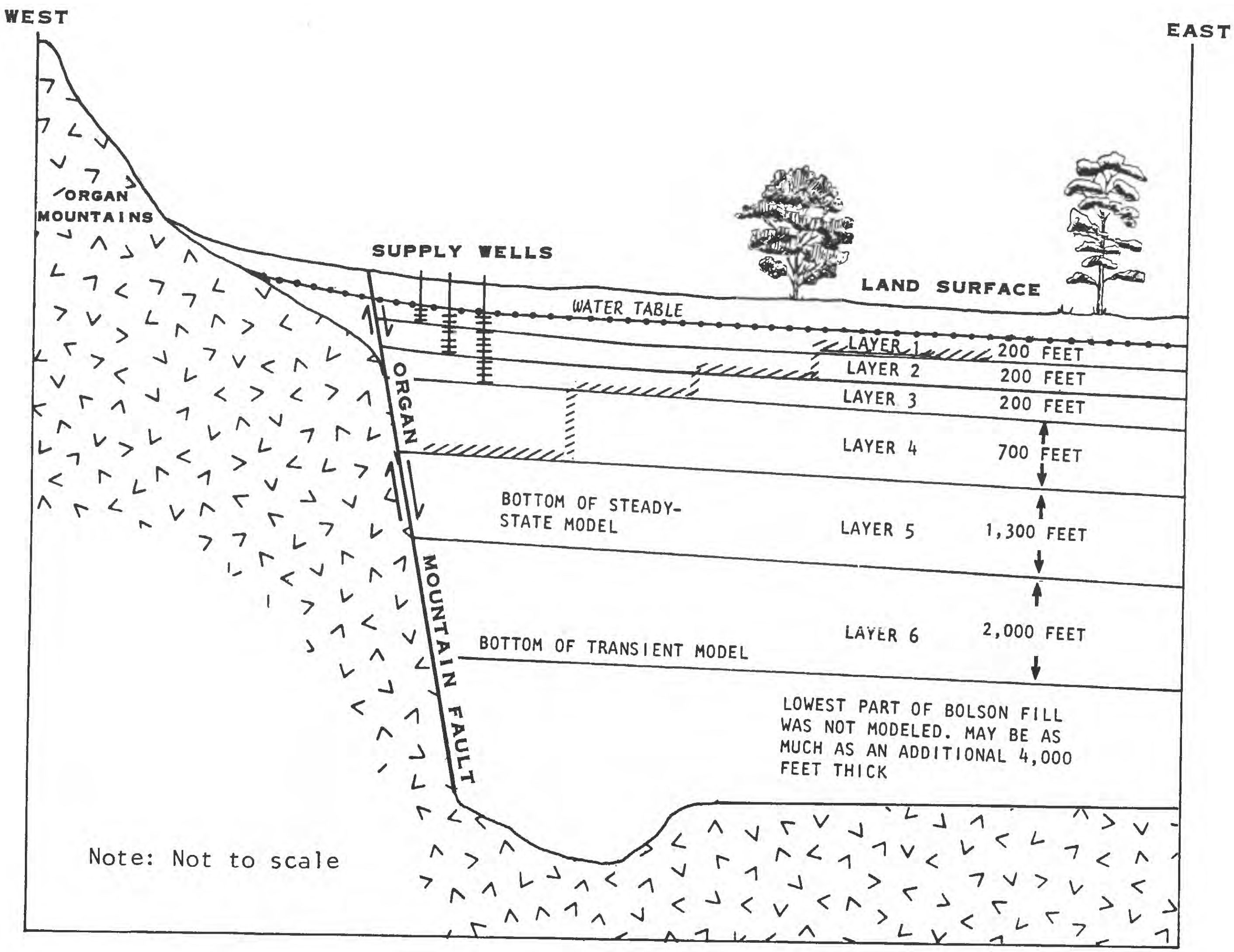


Figure 20.--Schematic diagram of layers used in the model.

Physical Geometry and Boundary Conditions

The physical geometry of the bolson-fill aquifer is specified by inactive or active cells within each layer. If parts of the aquifer represented by a layer do not exist in certain areas, the appropriate cells are specified as inactive. No flow of water can take place between active and inactive cells.

Two types of boundary conditions were used in this model: constant head and constant flux. At a constant-head boundary cell, the hydraulic head does not change from the specified value. However, as the hydraulic head in adjacent cells changes, the flow of water to and from the constant-head cell varies. At a constant-flux boundary cell, the net flow of water from the cell is specified and is not dependent upon the hydraulic head in adjacent cells. A no-flow boundary is represented by a constant-flux cell with flow set equal to zero.

The western side of the model is bounded by igneous rocks that are assumed to be impermeable. Thus, the igneous rocks were represented by a no-flow boundary (fig. 19). According to qualitative maps of the predevelopment water table, water probably flowed into the study area from the mountain front and discharged to the south. Constant-flux cells were used to simulate mountain-front recharge. To represent the outflow, constant-head cells were assigned along the southern boundary of layer 1. A no-flow boundary was used along a probable stream line as the northern model boundary. Because the north and south ends of the model do not correspond to real physical boundaries, the effect of these imaginary boundaries was tested by changing between constant-head, constant-flux, and no-flow cells during the model-adjustment procedure. To the east, a physical boundary exists that is caused by the difference in density between freshwater along the mountain front and saline water of the Tularosa Basin. This boundary was simulated in two different ways. For steady-state simulations, only the freshwater part of the aquifer was simulated using layers 1 through 4 (fig. 20). The position of the boundary between fresh and saline water was estimated and treated as a constant-head boundary. For transient simulations, water must be allowed to move from saline water to freshwater parts of the aquifer. Therefore, all six layers were used as shown in figure 20. Because the model assumes that the density of water is uniform, some error will be introduced in the simulation of transient ground-water flow. However, because the ground-water development is restricted to the freshwater part of the aquifer, the error in simulating water-level change near the well field probably will be slight. For transient simulations, the eastern no-flow boundary was located approximately along the flow line that divides the eastern and western halves of the Tularosa Basin.

The top of layer 1 corresponds to the estimated predevelopment water-table surface. Thus, the thickness of all layers is measured relative to the predevelopment surface. The no-flow boundary at the base of layer 6 represents the bottom of the aquifer at a depth of about 4,600 feet below the predevelopment water table. Although the aquifer may be as much as 4,000 feet thicker, sensitivity analyses indicate that eliminating this material does not affect simulated water levels in the well field.

Hydraulic Properties

The hydraulic properties required by the model include: hydraulic conductivity, transmissivity, vertical hydraulic conductivity, specific yield, and storage coefficient. For the initial simulation, all cells within a layer were assigned the same hydraulic properties. The hydraulic conductivity was assumed to be 2 feet per day throughout layer 1, which is approximately equal to the geometric mean of hydraulic-conductivity values determined from aquifer tests (fig. 12). The transmissivity of layers 2 and 3 was assumed to be equal to 2 feet per day times the thickness of the layer. The transmissivity of layers 4 through 6 was estimated using the relation between hydraulic conductivity, seismic velocity, and depth shown in figures 13 and 14. Vertical hydraulic conductivity of a layer was assumed to equal its horizontal hydraulic conductivity times 0.004. Specific yield of the aquifer determined from measurements of water-level decline was about 0.15. The storage coefficient was estimated as 0.000001 times the thickness of the layer (Lohman, 1972, p. 8). The hydraulic properties selected for each model layer are summarized below.

Layer number	Thickness (feet)	Horizontal hydraulic conductivity (feet per day)	Transmissivity (feet squared per day)	Vertical hydraulic conductivity (feet per day)	Specific yield	Storage coefficient
1	200	2.0	--	0.0080	0.15	--
2	200	--	400	.0080	--	0.0002
3	200	--	400	.0080	--	.0002
4	700	--	700	.0040	--	.0007
5	1,300	--	650	.0020	--	.0013
6	2,000	--	200	.0004	--	.0020

Model Adjustments

Model adjustment is a process in which hydraulic properties are changed until the model adequately simulates the measured behavior of the real hydrologic system. This process, which is commonly called model calibration, is not intended to establish with certainty the accuracy or precision of model predictions. Rather, the process is used to refine concepts and test hypotheses until an adjusted model is constructed that mimics the real hydrologic system. The model was adjusted by matching as closely as possible water-level changes measured in the observation wells in the Post Headquarters area.

The ground-water model was adjusted in two steps using transient and steady-state simulations. Generally, steady-state simulations of the predevelopment hydrologic system are conducted first. However, in this study, the predevelopment conditions of the hydrologic system were not well known. Because the estimated predevelopment water-level altitude (fig. 8) was only a rough approximation, a superposition model was constructed first to simulate transient changes in water levels caused by withdrawals. A model of transient water-level changes could be better evaluated than a steady-state model because ground-water withdrawals and water-level changes have been measured since 1948. After adjustments were made to the transient model, a steady-state model was constructed using the hydraulic properties found to be appropriate in the transient simulations. The steady-state simulations were conducted solely to test whether estimated recharge rates were reasonable. The recharge rates were tested by comparing the steady-state water-table gradient simulated using estimated recharge rates to the measured predevelopment water-table gradient (fig. 9).

Transient Simulations

The transient superposition model was constructed by assigning boundary conditions and hydraulic properties described in the previous sections. Then, the model was adjusted to match the aquifer response measured in observation wells. The superposition approach assumes that the sum of several hydraulic stresses acting separately is equivalent to the effect of the same stresses acting simultaneously. The validity of the superposition approach was tested by comparing its results to the results of the equivalent nonsuperposition problem with all stresses included. The results were nearly identical.

Model adjustments were conducted by comparing measured and simulated water-level declines for 1948-82, which were divided into 18 time periods. The first period was 1 year long and the subsequent 17 periods were each 2 years long. At the beginning of each time period, the rate and location of ground-water withdrawals could be changed. At the end of each period, the water-level change was computed. The rate and location of ground-water withdrawals were determined from pumping records (table 1). For supply wells that were completed in more than one layer, the total measured withdrawal rate was divided equally between the appropriate layers. The layer, row, and column location of ground-water withdrawals is shown in table 2 and figure 19. Ground-water recharge was simulated at cell 1-22-17 (table 2), which represents the return flow of water from the wastewater-treatment plant. The rate of recharge was estimated to be about 30 percent of the total withdrawal from the Post Headquarters wells. Wastewater returns were simulated in the model for 1959-82.

Table 2.--Location of simulated ground-water withdrawals and wastewater-return flow

[Amount of withdrawals are listed in table 1]

Recharge or discharge source	Layer numbers	Row-column location
Supply Wells 10 and 10A	1, 2, 3	18-9
Supply Well 11	1	18-7
Supply Well 12	1, 2	18-6
Supply Well 13	1	16-7
Supply Well 14	1, 2	16-8
Supply Well 15	1, 2, 3	16-10
Supply Well 16	1, 2, 3	17-9
Supply Well 17	1, 2, 3	15-9
Supply Well 18	1, 2	14-9
Supply Well 19	1, 2	12-9
Supply Well 20	1, 2	10-9
Supply Well 21	1, 2	20-12
Supply Well 22	1, 2	19-12
Wastewater-return flow	1	22-17

The initial simulation using hydraulic properties and boundary conditions described in the model-construction section did not closely match measured water-level changes. In general, the area of influence of the ground-water withdrawals extended too far east and west but not far enough north and south. Simulated water-level declines were about 400 percent too small at Supply Wells 10A, 18, and 19; about 700 percent too small at Test Well 11; about 300 percent too great at Test Wells 4 and 5; and about 500 percent too great at Test Well 6.

Of the hydraulic properties assigned to the model, the horizontal hydraulic conductivity, transmissivity, and vertical hydraulic conductivity were thought to be known with the least confidence. Therefore, these properties were adjusted to provide a closer match between simulated and measured water-level changes. With the exception of the wastewater-return flow, all stresses were believed to be accurately known.

It was apparent from the initial transient simulations that adjustments to hydraulic conductivity and transmissivity in several areas would be needed to improve the model. Vertical hydraulic conductivity was varied by assuming that it was always equal to 0.004 times the horizontal hydraulic conductivity. The hydraulic conductivity and transmissivity needed to be increased in the northern part of the reentrant and decreased to the west, south, and east of the supply wells. Several simulations were conducted using different values and combinations of hydraulic conductivity and transmissivity. Hydraulic-conductivity values in the upper 200 feet of saturated aquifer (layer 1) that provided the best match between simulated and measured water-level changes ranged from 0.1 to 10 feet per day (fig. 21). Transmissivity values in layers 2 and 3 also were changed such that the transmissivity was equal to the layer thickness times the hydraulic-conductivity value shown in figure 21. Transmissivity values in layers 4 through 6 were not changed.

The hydraulic properties used in the transient model that best simulated measured water-level declines in observation wells are listed below.

Layer number	Thickness (feet)	Range of horizontal hydraulic conductivity (feet per day)	Range of transmissivity (feet squared per day)	Vertical hydraulic conductivity (feet per day)	Specific yield	Storage coefficient
1	200	0.1-10	--	0.0080	0.15	--
2	200	--	20-2,000	.0080	--	0.0002
3	200	--	20-2,000	.0080	--	.0002
4	700	--	700	.0040	--	.0007
5	1,300	--	650	.0020	--	.0010
6	2,000	--	200	.0004	--	.0020

Using the range of hydraulic-conductivity and transmissivity values in layers 1-3, the model closely simulated measured water-level changes in most wells (figs. 22 and 23). The simulated change throughout the Post Headquarters area for 1948-82 is shown in figure 24. The water-level rises measured in Old Supply Well 12, Supply Well 13, and Test Wells 6 and 8 were not accurately simulated, possibly because there were hydrologic stresses other than pumping that had not been included in the model, as described later in this report. Near the wastewater-treatment plant, water-level rises were measured at Test Wells 11 and 15. The model simulates the rise at Test Well 11 fairly well, but the match at Test Well 15 is poor. The poor fit may exist because hydraulic connection between the well and aquifer was not good until drilling fluid was cleaned from the well in 1982 or because unsaturated flow from the wastewater plant was simulated as direct recharge to the water table by the model. Also, the coarse cell size of the model near the wastewater plant does not permit very accurate simulation of changes in hydraulic head in that area.

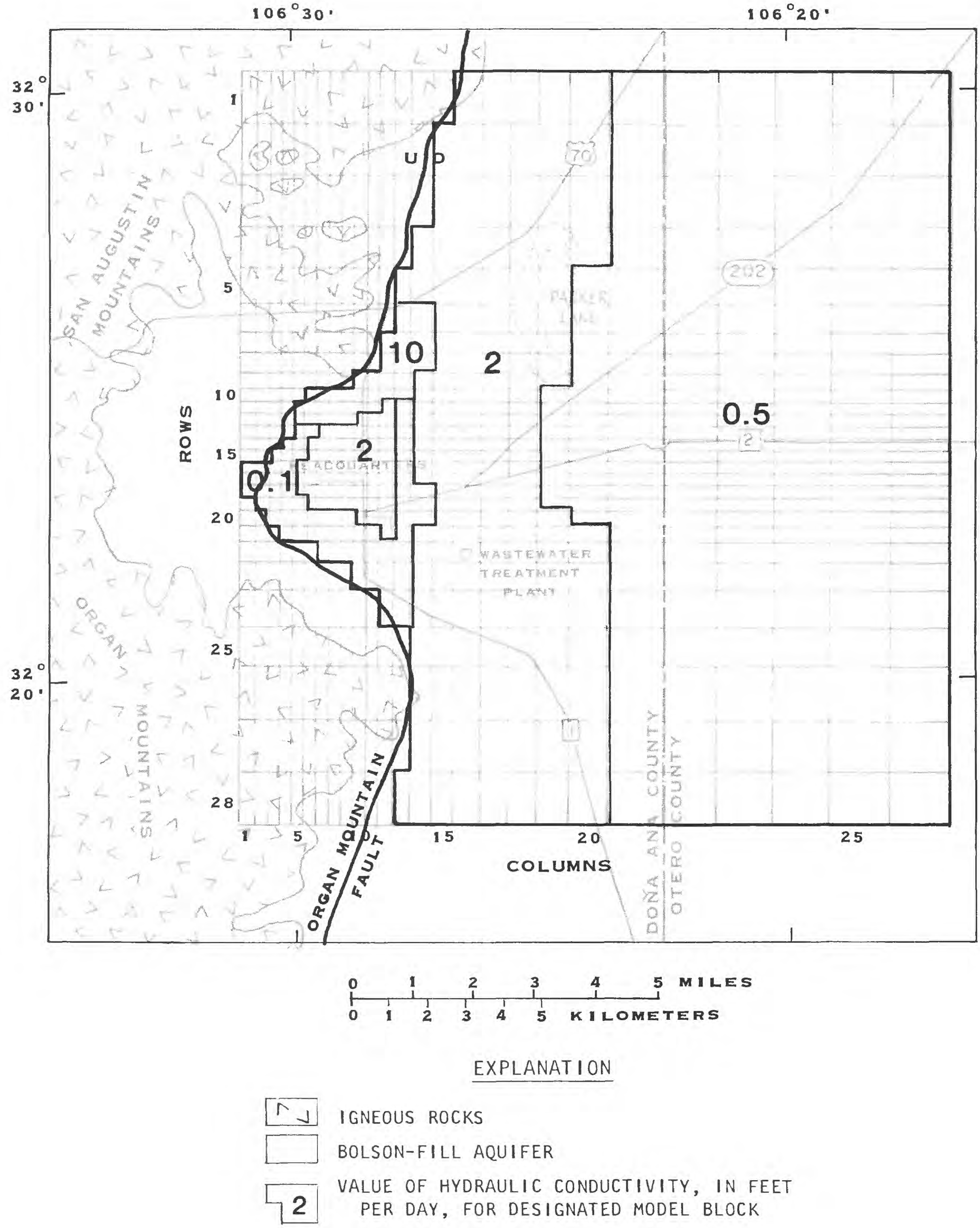


Figure 21.--Hydraulic-conductivity values assigned to model layer 1.

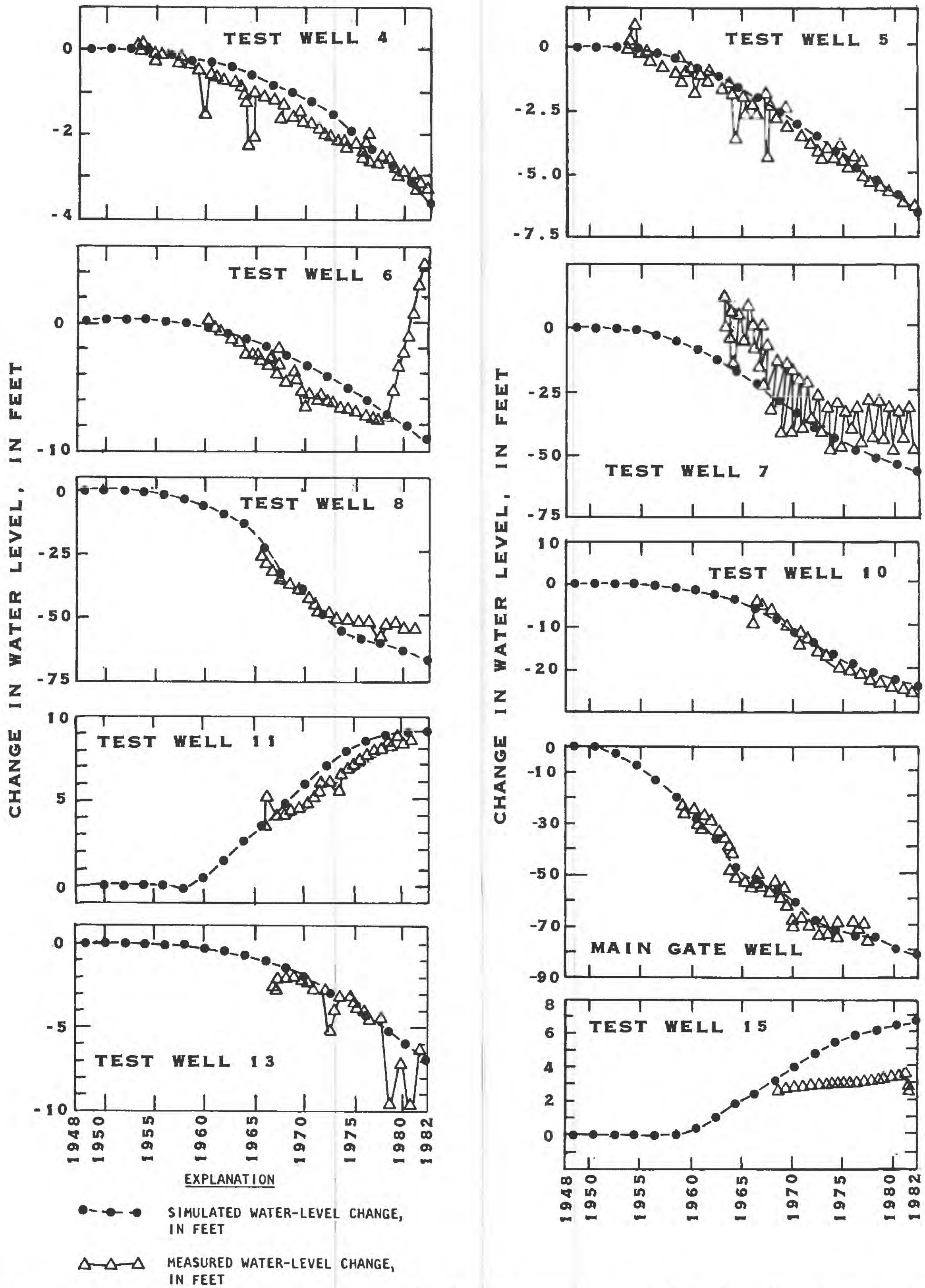


Figure 22.--Comparison between simulated and measured water-level changes in the Main Gate Well and selected test wells, 1948-82.

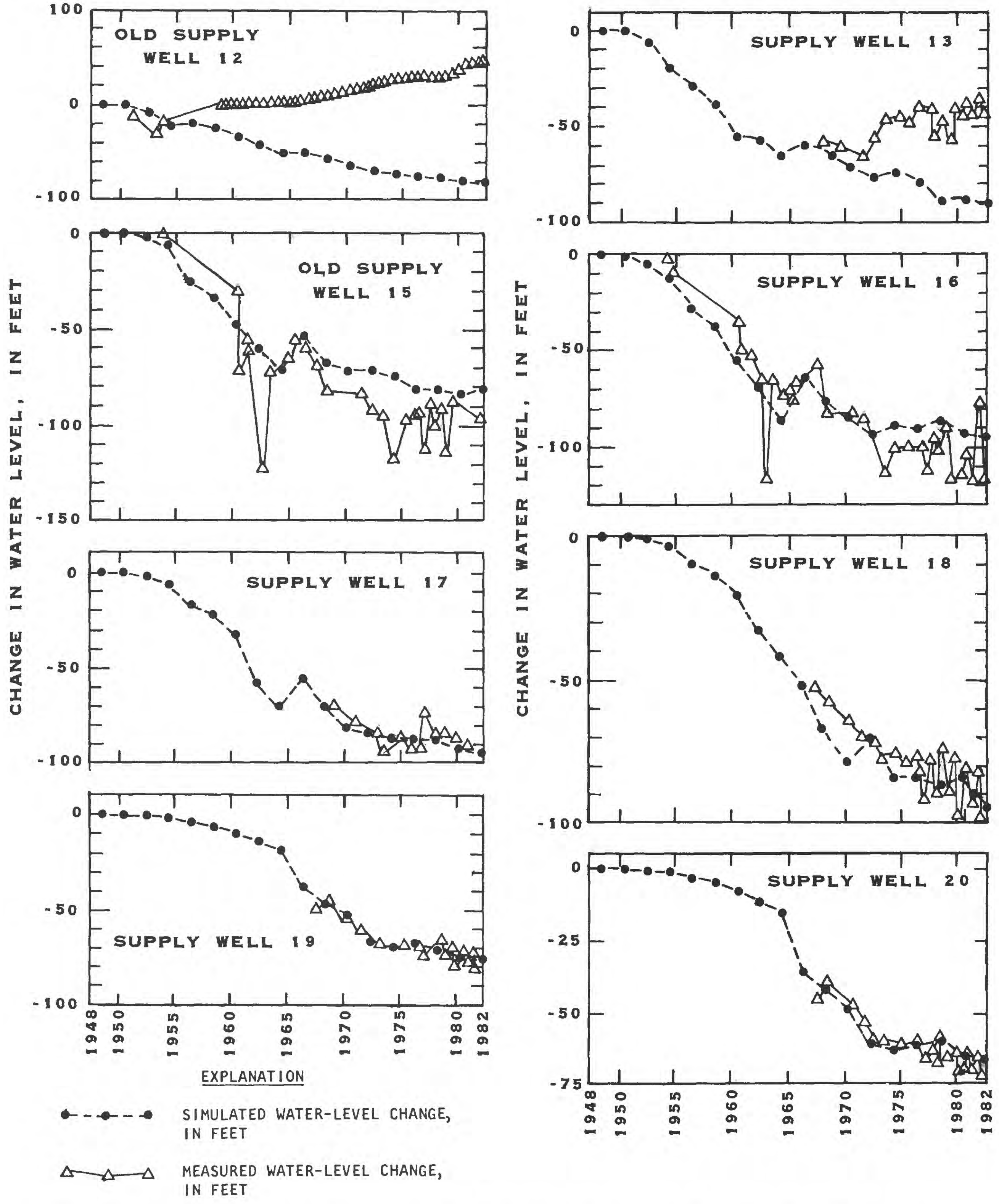


Figure 23.--Comparison between simulated and measured water-level changes in selected supply wells, 1948-82.

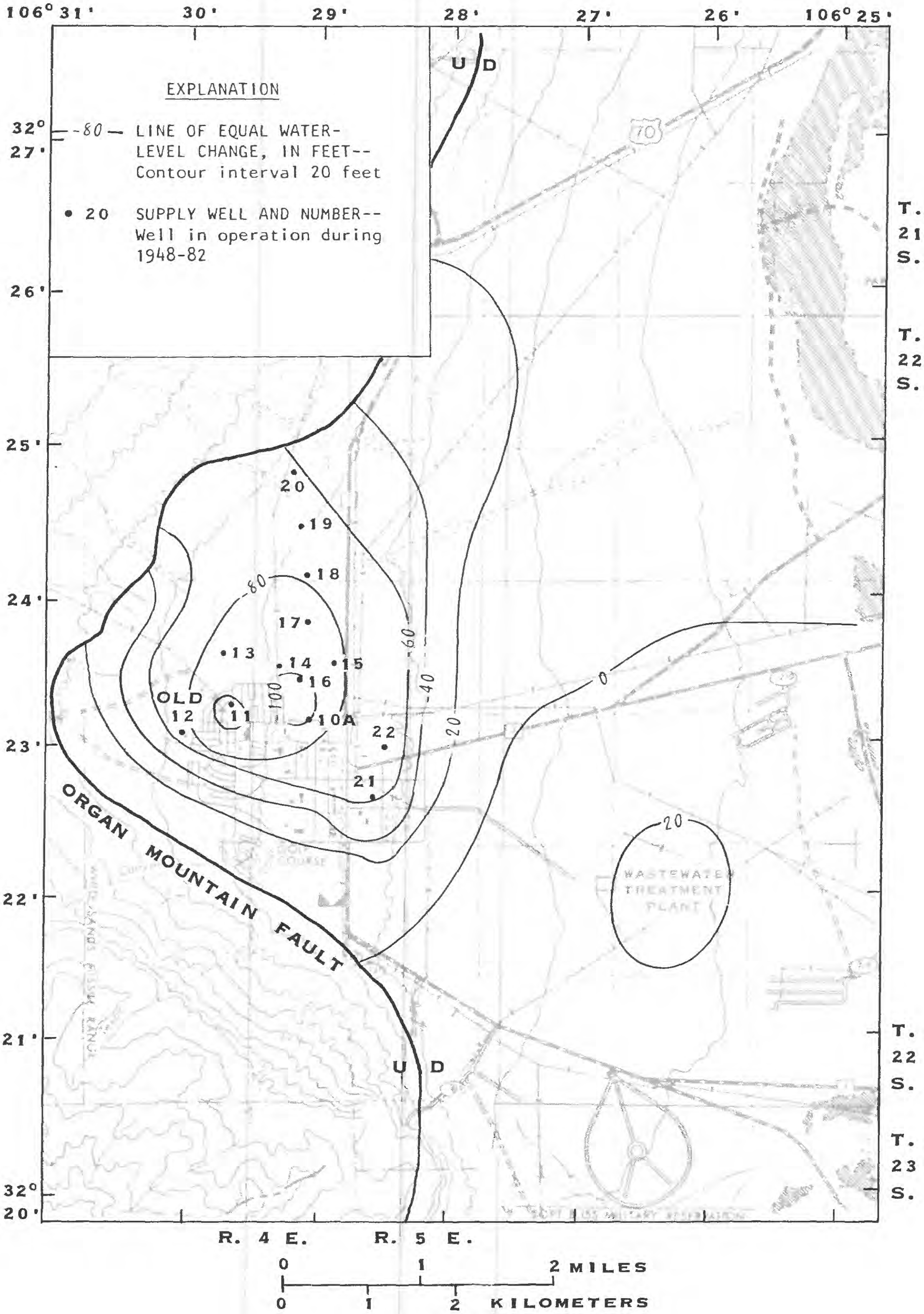


Figure 24.--Simulated water-level change for 1948-82.

A zonė of material with small hydraulic conductivity (0.1 foot per day) about 1 mile east of the well field (fig. 21) was simulated. This partial barrier to flow was needed to simulate the observed aquifer response in Test Wells 4 and 5. The apparent zone of small hydraulic conductivity could be caused by stratigraphic changes in the bolson fill or by a fault.

Many adjustments were made to test different arrangements of hydraulic properties. The concept investigated most thoroughly was that the hydraulic conductivity is greater along the mountain front than away from the mountain front. This concept was tested by allowing hydraulic conductivity in the model to be anisotropic in the horizontal plane. Anisotropy ratios from 1:2 to 1:15 were tested. The anisotropy allowed the area of influence caused by the well field to extend to the north and south to a greater extent than to the east and west. However, the simulated water-level declines were not close to measured declines at several of the observation wells. The anisotropy allowed the wastewater returns to drastically affect water levels at Test Wells 4 and 5. Also, water-level declines along the western part of the reentrant at Test Wells 6 and 8 could not be simulated accurately. Even using anisotropy ratios to achieve a good match between measured and simulated water levels, a variable distribution of hydraulic conductivity similar to that shown in figure 21 was needed. Therefore, attempts to adjust the model using anisotropy in the horizontal plane were abandoned.

Different hypotheses on the cause of rising water levels in the southwest part of the reentrant were investigated using the model. Since 1952, the water level in Old Supply Well 12 has risen about 60 feet. A less obvious rise of about 20 feet has occurred in Supply Well 13 since the early seventies.

The hypothesis that an increase in recharge caused the water-level rise was tested by simulating additional recharge at the golf course, in the major arroyo south of the housing area, and behind the levy west of the housing area. Additional recharge of 75 acre-feet per year during the past 35 years was applied at cell 1-21-7 to simulate return flow to the water table caused by watering the golf course. The model indicated that this additional recharge would have an insignificant effect on water levels as far away as Old Supply Well 12 and Supply Well 13. An increase in recharge from the arroyo that drains the southern part of the reentrant (watershed 8 in fig. 4) also was tested. Additional recharge of 400 acre-feet per year was divided between cells 1-18-1, 1-19-2, 1-20-3, 1-20-4, 1-21-6, and 1-22-7 for the entire simulation for 1948-82. The recharge decreased the simulated drawdown at Old Supply Well 12 only about 6 feet. This large volume of increased recharge clearly was not sufficient to account for the measured rise in water level. Another area where recharge may have been increased in recent years is behind the levy that protects the Post housing from floodwaters. The U.S Army Corps of Engineers' (1978) report on the flood of August 19, 1978, indicated that sheetwash was impounded behind the levy. By impounding water, the levy could have caused an increase in ground-water recharge. An increased recharge rate of 360 acre-feet per year for 20 years at cells 1-16-6, 1-17-5, 1-18-5, 1-19-5, 1-20-5, and 1-21-6 located behind the levy was insufficient to cause the measured water-level increase at Old Supply Well 12 and Supply Well 13.

The changing distribution of ground-water withdrawals within the Post Headquarters well field also was investigated as a possible explanation for the rising water levels. Even though the ground-water model as constructed accounts for the change in location and volume of withdrawals from the supply wells, the measured water-level rise in the southwestern part of the reentrant could not be duplicated. However, for one simulation where all pumping wells were mistakenly turned off in 1964, a nearly perfect match of the rising water levels was generated. This simulation could indicate that at some time in the early 1960's the hydrologic stress in the well field caused by withdrawals may have stopped affecting the water levels in Old Supply Well 12 and Supply Well 13. If, for example, corrosion of well casings or pumping of different wells caused water to be extracted from sand lenses that were not connected hydraulically to the units that were penetrated by wells in the southwestern part of the reentrant, water levels in these units could recover. Also, an increase in recharge could cause water levels to rise if the southwestern part of the reentrant is not hydraulically connected to other parts of the well field.

Conceptualizing the bolson-fill aquifer as an integrated aquifer system may not be adequate to account for rising water levels in the southwestern part of the reentrant. A more detailed knowledge of the hydraulic connection between Old Supply Well 12 and Supply Well 13 and the other supply wells is needed to help clarify the observed aquifer response. Additional test wells completed at various depths in the southern part of the reentrant would provide the data needed to better understand aquifer response in this area.

Steady-State Simulations

Prior to ground-water development in the Post Headquarters area, the hydrologic system was assumed to be in an equilibrium (steady-state) condition for which the rates of ground-water inflow and outflow were equal and appreciable changes in ground-water storage were not occurring. However, this assumption is not strictly valid over all chosen time periods. As described previously, an important mechanism of recharge may be infrequent slugs of water from very large storms that cause water levels to rise and fall as aquifer storage fills and drains. Therefore, the assumed steady-state condition must be thought of as a dynamic equilibrium that holds for long-term average values of recharge and discharge.

Several steady-state simulations were conducted for the purpose of evaluating recharge to the bolson fill. Recharge rates determined using the characteristics of the drainage basins shown in figure 4 were included in a steady-state model. If the water-table gradient simulated by the model is similar to the predevelopment gradient shown in figure 8, some measure of confidence can then be accorded the recharge estimate. However, if steady-state simulations produce a gradient significantly different from the predevelopment water table, recharge estimates probably need to be reevaluated.

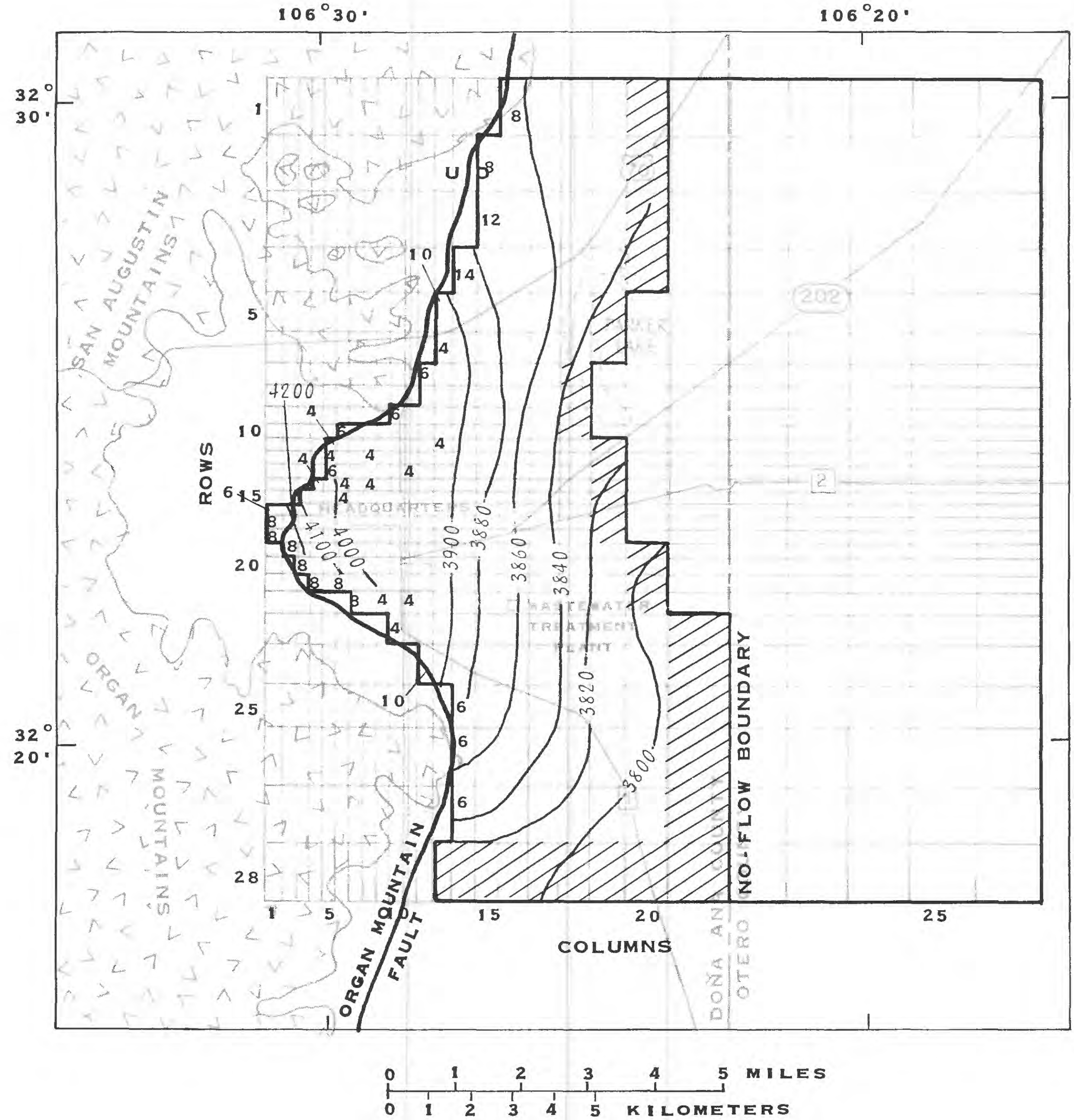
Before steady-state simulations were made, some model boundaries were changed. The changes were necessary because freshwater, owing to its lesser density, flows over the top of the saline water, which creates a separation between the freshwater and saline-water flow systems. Therefore, the no-flow boundary on the east side of the transient superposition model was changed to a constant-head boundary and moved westward for each layer to the approximate position of the interface between fresh and saline water (fig. 20). The constant-head boundary was assigned the value of the water-table altitude at the saline-water boundary. The flow simulated from these constant-head cells represents the movement of freshwater to the east where it overrides and mixes with the denser saline water.

A recharge rate of 1,590 acre-feet per year (1,140 acre-feet in the Post Headquarters watershed) was used for the steady-state simulation (fig. 4). The water-table gradient simulated using this recharge rate (fig. 25) was similar to the predevelopment gradient (fig. 8), which indicates that the recharge rate was reasonable for the aquifer properties determined from the transient-flow model.

Aquifer Response to Future Withdrawals

The adjusted version of the ground-water flow model was used to simulate water-level change during 1983-2017. Four separate plans of future ground-water development were simulated. All plans assume that ground-water withdrawals take place from Supply Wells 10A, 11, 13, 16, 17, 18, 19, 20, 21, and 22 and Old Supply Well 15 at the cell locations listed in table 2. These were the same wells used in the transient adjustment simulation. The four plans differ in the total quantity of ground-water withdrawals and wastewater-return flow as shown below.

	Plan 1	Plan 2	Plan 3	Plan 4
Ground-water withdrawals, in acre-feet per year	2,066	2,066	1,033	1,033
Wastewater-return flow, in acre-feet per year	620	0	310	0



EXPLANATION

IGNEOUS ROCKS

BOLSON-FILL AQUIFER

4 CELL WITH SIMULATED RECHARGE--Number is rate in cubic feet per second times 0.01

CONSTANT-HEAD CELL

—3800— WATER-TABLE CONTOUR--Shows altitude of simulated water-table surface, in feet above sea level. Contour interval is variable

Figure 25.--Simulated steady-state water-table surface and recharge rates.

Plan 1.--This plan specified that total ground-water withdrawals from the Post Headquarters well field are a constant 2,066 acre-feet per year for 1983-2017. This rate represents the average measured withdrawal rate during 1981-82 (table 1). The return flow of wastewater was assumed to be 30 percent of the total ground-water withdrawals. Therefore, 620 acre-feet of water per year was applied at cell 1-22-17 to represent this recharge.

The simulated water-level change from 1983 to 2017 is shown in figure 26. Water levels generally continued to decline as more water was removed from storage. The largest decline of 49 feet was simulated at Supply Well 17. A rise in water level due to continued recharge from the wastewater-treatment plant was simulated for about 9 years at the row and column location of Test Well 11. After 9 years, water levels began to decline as the effect of withdrawals from the supply wells expanded.

Plan 2.--This simulation assumed the same location and rate of ground-water withdrawals as in Plan 1; however, the return flow from the wastewater-treatment plant was not simulated. This plan shows the effects on water levels if the treated effluent is not permitted to recharge the aquifer.

The simulated water-level change from 1983 to 2017 is shown in figure 27. Near the supply wells, the simulated water-level declines are nearly the same as in Plan 1. A maximum drawdown of 50 feet was simulated at the locations of Supply Wells 17 and 10A. Near the wastewater-treatment plant, water levels declined as much as 40 feet due to the effects of withdrawals and dissipation of the ground-water mound (fig. 24) caused by previous years of wastewater-return flow. Without the wastewater-return flow, water-level declines will be greater near the boundary between fresh and saline water, which could expedite encroachment of saline water toward the well field.

Plan 3.--This simulation assumed that the ground-water withdrawal rate was decreased 50 percent at all wells. Therefore, total withdrawals from the Post Headquarters well field were 1,033 acre-feet per year for 1983-2017. The wastewater-return flow was simulated as 30 percent of the new withdrawal rate, or 310 acre-feet per year. The reduced withdrawal pumping rate from the Post Headquarters wells could represent a realistic future if part of the water needs of White Sands Missile Range were met by water from another source.

The simulated water-level change for 1983-2017 is shown in figure 28. Because withdrawals were decreased 50 percent, water levels increased near the pumping centers as much as 17 feet. However, hydrographs of water levels at the locations of old Supply Well 15 and Test Well 8 indicate that water levels did not increase throughout the entire 35-year simulation. The hydrographs show that water levels increased for about 20 years (1983-2003) and then began to decline. The water-level rise near the supply wells is caused by the smaller withdrawal rate, but because water is being withdrawn exclusively from storage within the aquifer, water levels eventually begin to decline. For any plan that proposes decreased withdrawals, this general pattern of rising and falling water levels can be expected. However, the magnitude of the water-level increase is dependent upon how much the withdrawals are decreased. Away from the wells, water levels continue to decline even though withdrawals have been reduced. The declines occur because water is still being withdrawn from storage.

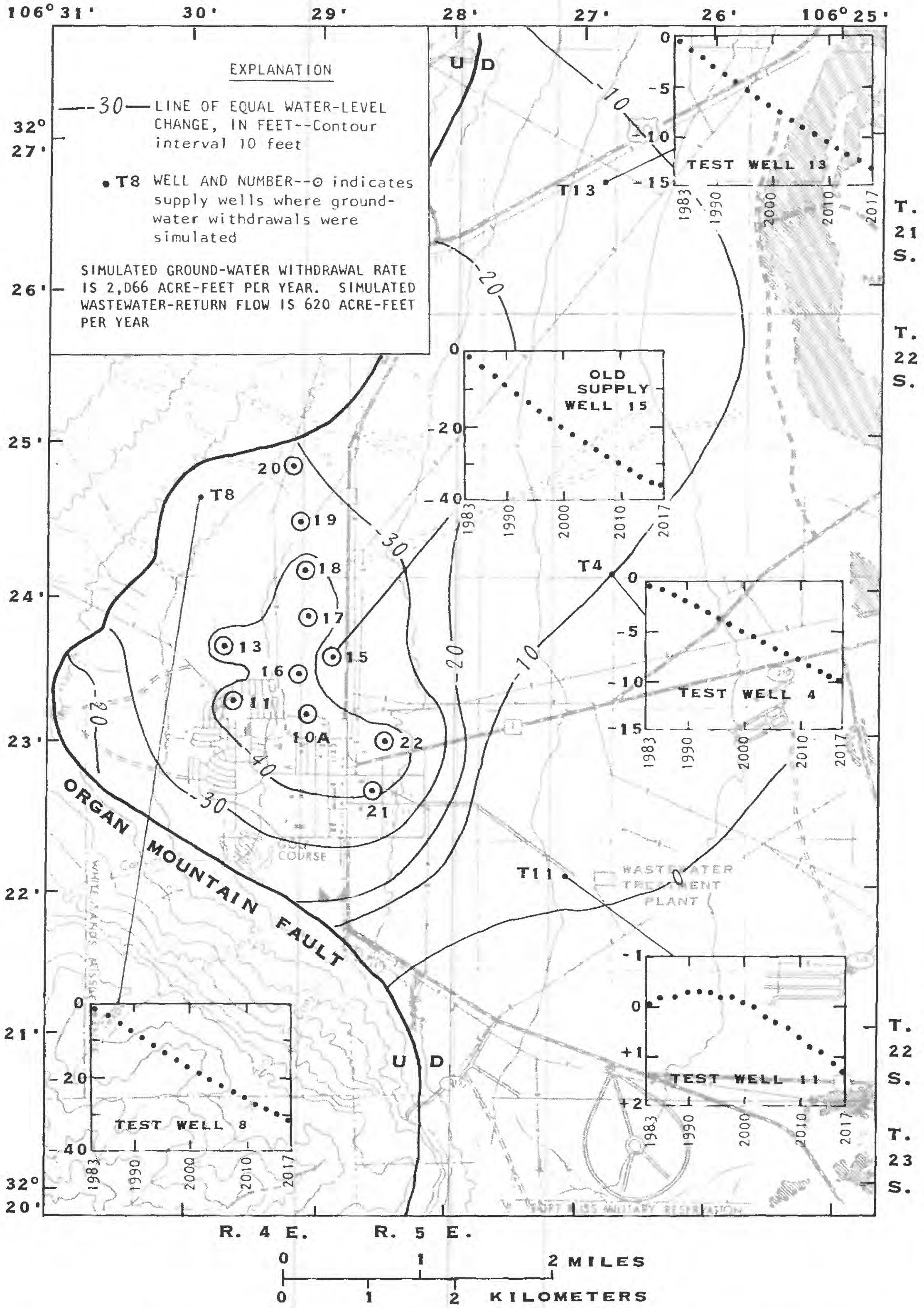


Figure 26.--Simulated change in water-table surface from 1983 to 2017 for withdrawal plan 1.

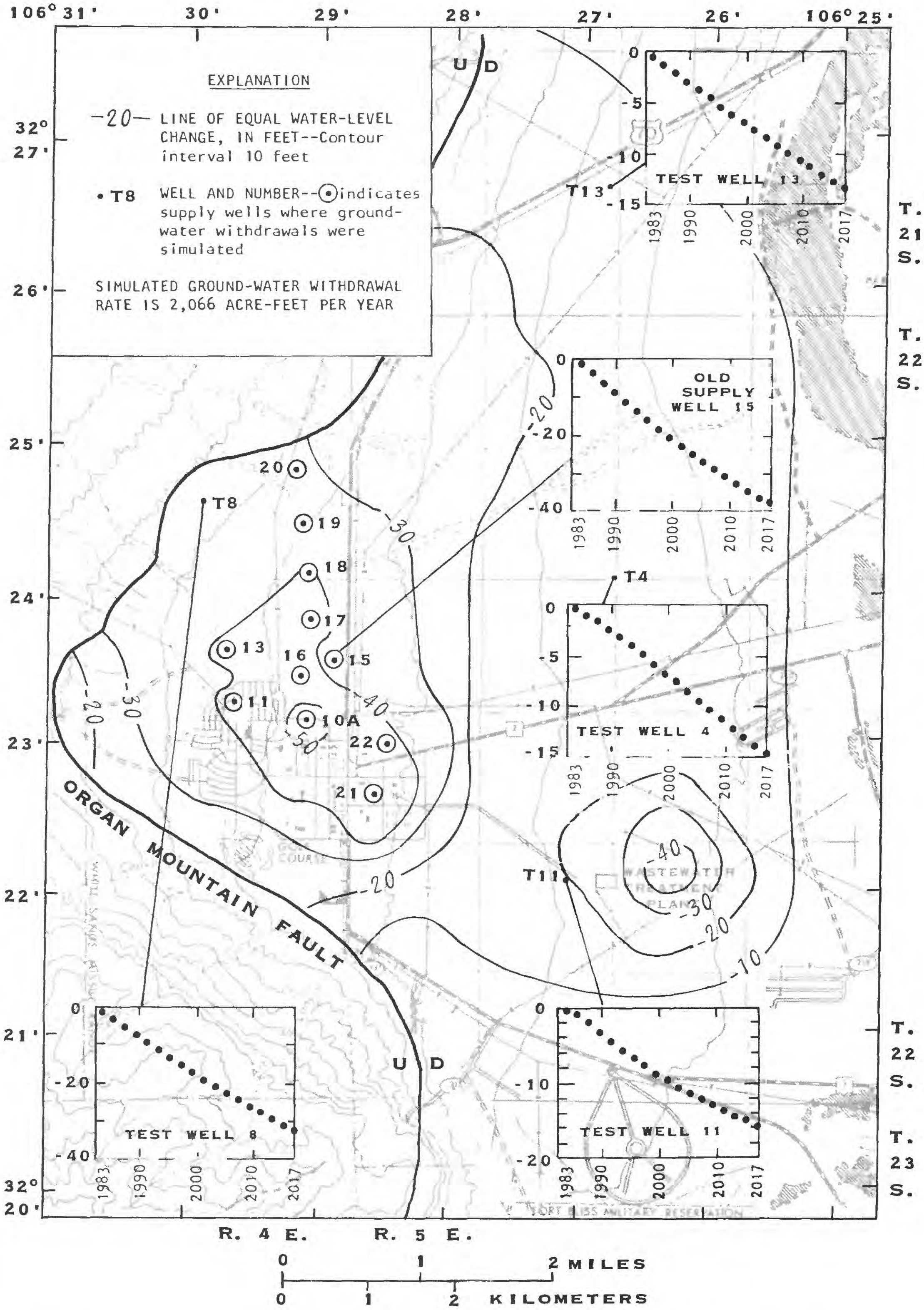


Figure 27.--Simulated change in water-table surface from 1983 to 2017 for withdrawal plan 2.

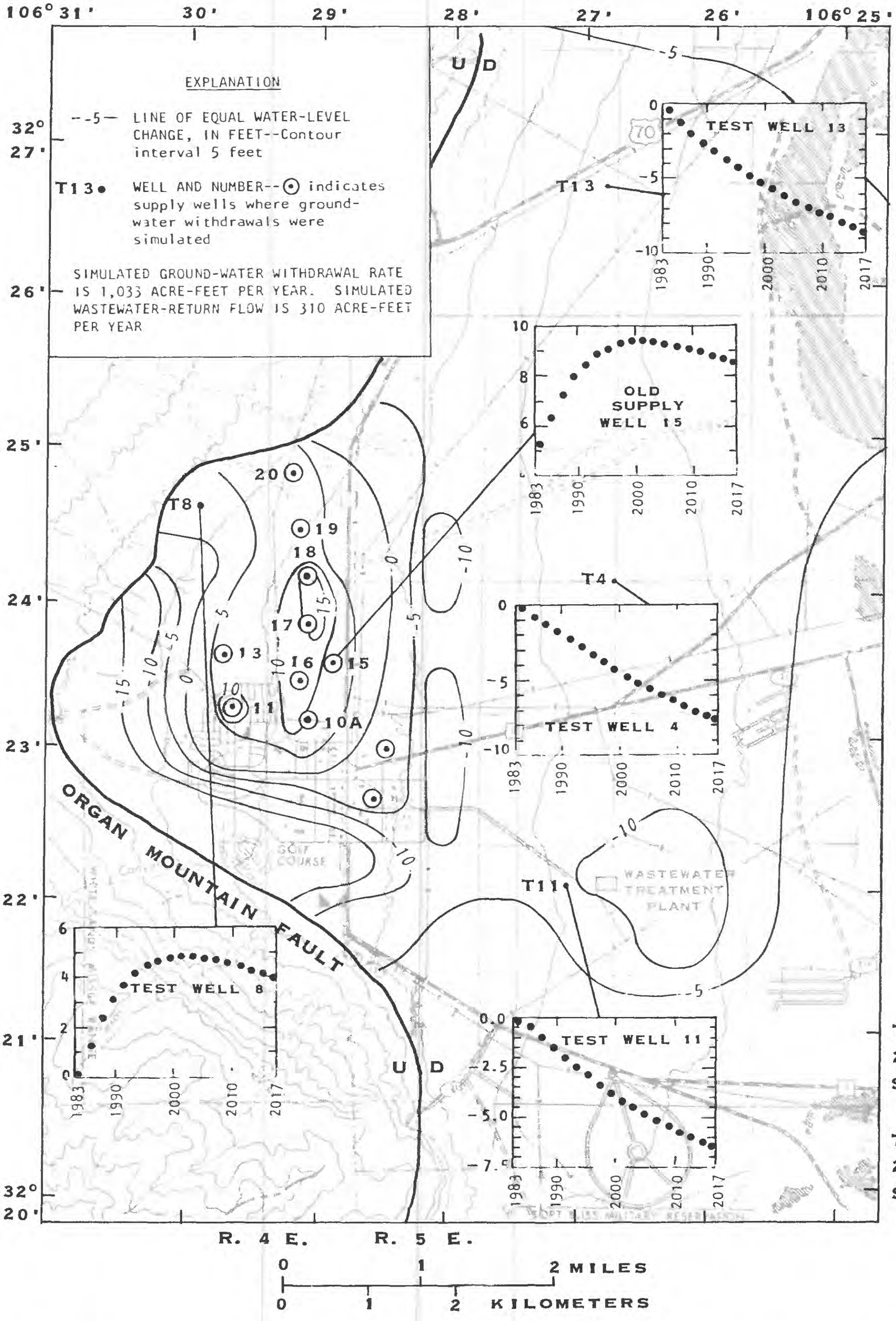


Figure 28.--Simulated change in water-table surface from 1983 to 2017 for withdrawal plan 3.

Plan 4.--This is the same simulation as Plan 3 except that return flow from the wastewater-treatment plant was not simulated. The simulated water-level change from 1983 to 2017 is shown in figure 29. Eliminating the wastewater return has little effect on the water-level rise near the supply wells. However, drawdown near the treatment plant is greater than 20 feet. Without the wastewater-return flow, water levels will decline more rapidly near the freshwater and saline-water interface, as discussed for Plan 2.

SIMULATED WATER-QUALITY CHANGES

Dissolved-solids concentrations in ground water vary from less than 300 to greater than 100,000 milligrams per liter in the Post Headquarters area. Withdrawing large quantities of freshwater from the aquifer eventually will draw denser mineralized water toward the supply wells. A model that simulates density-dependent ground-water flow and transport of solutes was used to gain some insight into the occurrence and movement of fresh and saline water in the bolson-fill aquifer near the well field. The model was used to investigate: (1) the extent and thickness of potable water prior to ground-water development, (2) the hydraulic characteristics used in the freshwater flow model, (3) the relative importance of upconing to lateral movement of saline water, and (4) water-quality changes caused by continued ground-water withdrawals.

Model Construction

A two-dimensional, finite-element model was used to simulate ground-water flow and solute movement in a vertical cross section through the Post Headquarters well field. The model was constructed using a computer code called SUTRA (Saturated-Unsaturated-TRAnsport), which was developed and documented by Voss (1984). The SUTRA code is capable of simulating ground-water flow and solute transport in saturated variable-density systems. In the following discussion, concentrations are expressed in parts per million when referring to the model. Concentrations that are the measured results of chemical analyses, however, are expressed in milligrams per liter.

An east-west section (fig. 30) was chosen through the well field along a flow line determined from the simulated predevelopment water-level map (fig. 8). The 1-meter-thick section is a representative slice of the bolson-fill aquifer that includes the freshwater and saline-water flow systems. The model is made up of 744 square elements in 12 rows and 62 columns that correspond to a section of the aquifer 37,200 feet long and 7,200 feet deep (fig. 30).

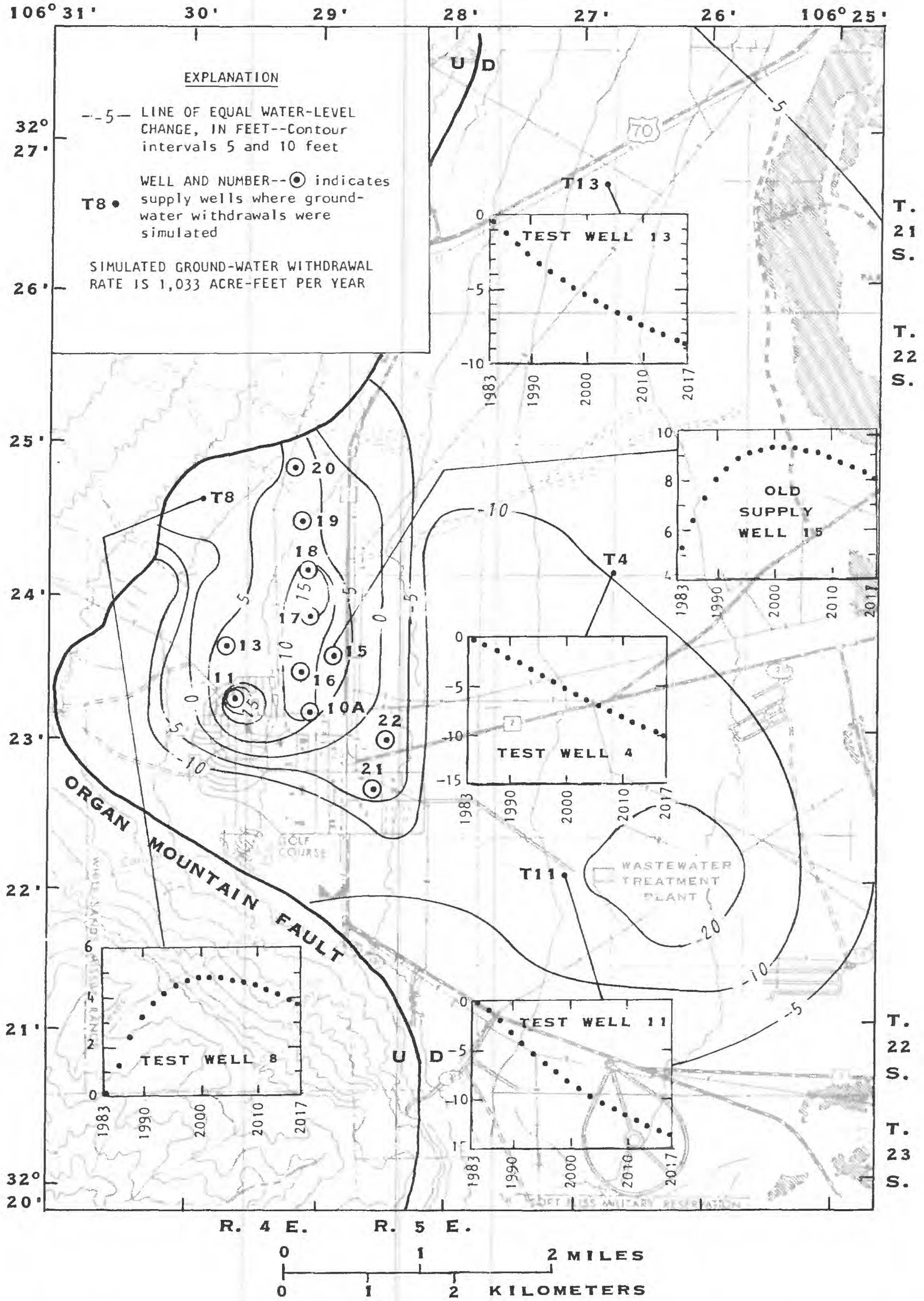
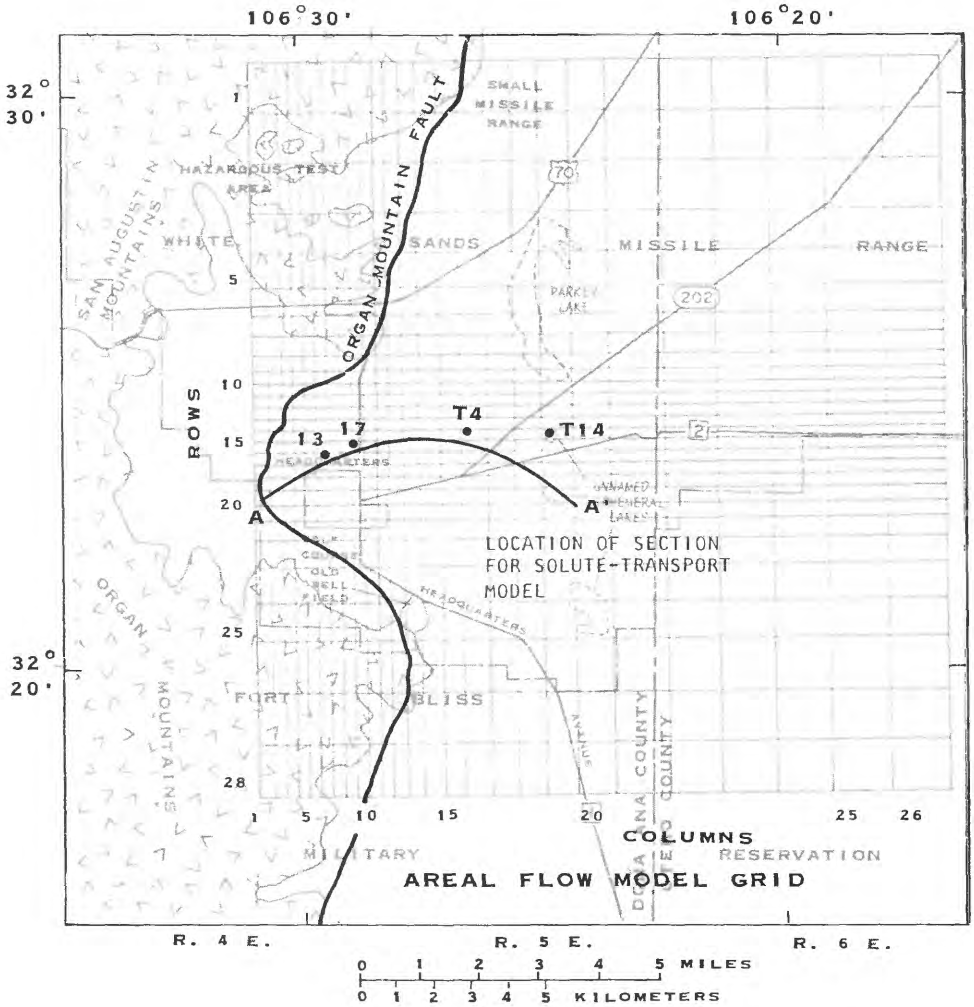


Figure 29.--Simulated change in water-table surface from 1983 to 2017 for withdrawal plan 4.



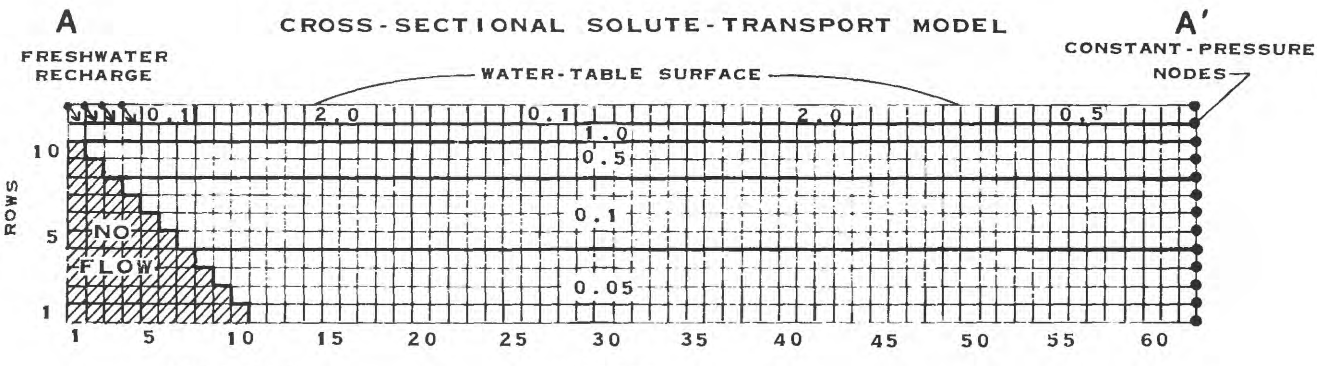
EXPLANATION

IGNEOUS ROCKS

BOLSON-FILL AQUIFER

T14 • TEST WELL

13 • SUPPLY WELL



EXPLANATION

2.0 HORIZONTAL HYDRAULIC CONDUCTIVITY, IN FEET PER DAY

0 3,000 6,000 9,000 FEET
0 1,000 2,000 METERS

Figure 30.--Location, boundaries, and horizontal hydraulic conductivity for a cross-sectional, finite-element solute-transport model.

The cross-sectional model was constructed using, as nearly as possible, the same boundaries and hydraulic properties employed in the three-dimensional flow model. The bottom of the aquifer was simulated using a no-flow boundary. The thickness of the aquifer (fig. 30) was determined from the gravity interpretation of Seager and Brown (1978, p. 147). Values of horizontal hydraulic conductivity used in the cross-sectional model are shown in figure 30. They differ from the areal flow model slightly because of the difference in cell sizes and depth of aquifer simulated. The full thickness of the bolson-fill sediments was simulated in the cross section to investigate saline-water flow paths. The ratio of vertical to horizontal hydraulic conductivity of 0.004 determined from the flow model was used. Freshwater recharge of 0.00024 cubic foot per second was added along the mountain front (fig. 30). The recharge rate was based on the 1,140 acre-feet of recharge used in the steady-state flow model divided by the 4-mile-wide section of the reentrant across which it flowed.

Chemical properties of the water also were required to simulate density-dependent flow and solute transport. All freshwater flow into the model was assumed to have a dissolved-solids concentration of 300 parts per million. This corresponds to approximately the average concentration of freshwater in the well field. Water could flow out of the model only at the eastern boundary where the pressure was held constant on the basis of the hydrostatic load of water containing 70,000 parts per million dissolved solids. This concentration was selected as a representative average value for the brine on the basis of samples collected from Test Well 14 (Doty and Cooper, 1970). All saline water that entered the model had this dissolved-solids concentration and a density of 1.049 grams per cubic centimeter.

In addition to chemical and hydraulic parameters that correspond to real physical properties, the solute model required a value of dispersivity. The dispersivity term allowed the model to simulate mechanical dispersion of solutes without simulating the exact pathways of every water particle. Reasonable values of dispersivity are dependent on the scale of the problem and probably are about equal to the length of the largest inhomogeneity in the system (Voss, 1984, p. 54). A horizontal dispersivity of 600 feet and a vertical dispersivity of 60 feet were used in the cross-sectional model. The coefficient of molecular diffusion was assumed to be insignificant compared to dispersivity.

Steady-State Simulations

The cross-sectional model was used to simulate steady-state dissolved-solids concentrations. The steady-state simulations show the predevelopment dissolved-solids concentrations in the aquifer. Using the hydraulic and chemical characteristics described in the previous section, the position of lines of equal concentration for 1,000 and 35,000 parts per million dissolved solids is shown in figure 31 as the standard simulation. The simulation indicates that freshwater is about 1,500 to 2,000 feet thick beneath the Post Headquarters well field. Water containing less than 1,000 parts per million dissolved solids is considered to be fresh and water containing more than 35,000 parts per million is brine. The zone between the lines for 1,000 and 35,000 parts per million represents the transition zone between freshwater and brine. This zone compares closely with the approximate position of the transition zone determined from geophysical evidence (McLean, 1970, fig. 29; Kelly, 1973, fig. 31). The fact that a reasonable steady-state solute-transport model could be developed without significantly changing the hydraulic properties used in the areal flow model lends credibility to the chosen values of horizontal hydraulic conductivity, vertical hydraulic conductivity, and recharge.

The flow of ground water in the section is greatly affected by its density. Brine containing 70,000 parts per million dissolved solids moves from east to west deep in the aquifer in contrast to the west-to-east flow of water in the upper part of the aquifer (fig. 31). As the brine moves toward the mountain front, it mixes with fresher water and moves back toward the center of the basin. Movement of the brine is very slow in the deep parts of the basin. For example, about 98 percent of water in column 30 flowed through the upper 1,800 feet of the aquifer. Flow velocities of brine at the bottom of column 30 were on the order of about 0.001 foot per year.

Sensitivity of the simulated dissolved-solids concentrations to changes in model characteristics was tested by varying aquifer thickness, dispersivity, hydraulic conductivity, brine density, and freshwater recharge rate (table 3). The sensitivity of the model is represented by the change in position of the transition zone (fig. 32). The simulated steady-state concentrations are relatively insensitive to the decreases in aquifer thickness (fig. 32A) and in dispersivity (fig. 32B) tested. The position of the transition zone between freshwater and brine is very sensitive to values of brine density and freshwater recharge rate and less sensitive to hydraulic conductivity (fig. 32).

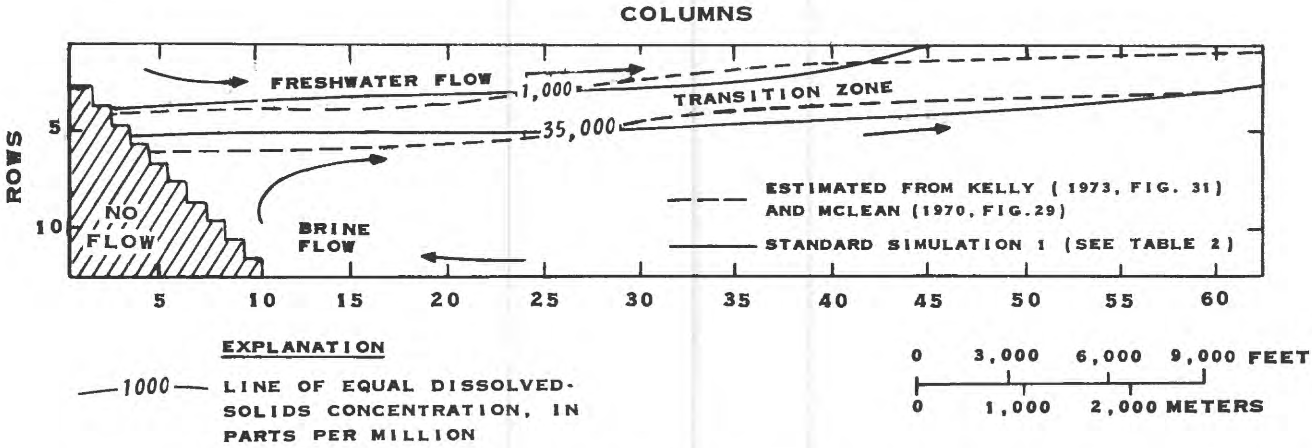


Figure 31.--Standard steady-state simulation of dissolved-solids concentration.

Table 3.--Characteristics of steady-state solute-transport model varied during sensitivity testing

Simulation number	Aquifer thickness	Horizontal dispersivity (feet)	Vertical dispersivity (feet)	Horizontal hydraulic conductivity (feet per day)	Ratio of vertical to horizontal hydraulic conductivity	Brine density (grams per liter)	Fresh-water recharge rate (cubic feet per second)
1 (standard)	deep	600	60	0.1-10	0.004	1,049	0.00024
2	shallow[1]	600	60	.1-10	.004	1,049	.00024
3	deep	[1]400	[1]40	.1-10	.004	1,049	.00024
4	deep	600	60	[1]2	[1]1	1,049	.00024
5	deep	600	60	.1-10	.004	[1]1,070	.00024
6	deep	600	60	.1-10	.004	[1]1,034	.00024
7	deep	600	60	.1-10	.004	1,049	[1].00048

[1] Indicates characteristic changed from standard conditions for that simulation.

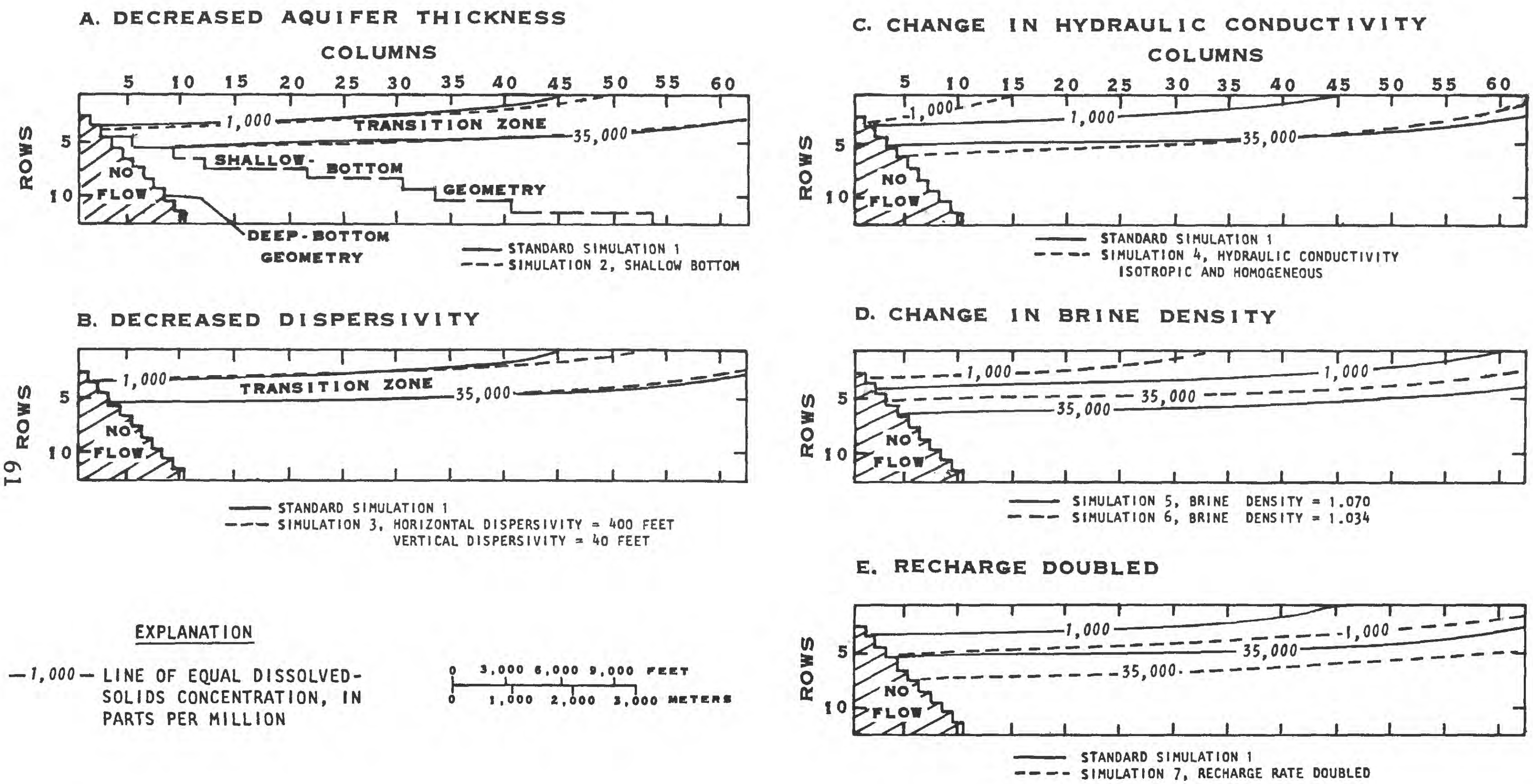


Figure 32.--Sensitivity of the standard steady-state simulation to changes in selected physical and chemical model parameters.

Transient Simulations

Transient simulations using the cross-sectional solute-transport model illustrate possible changes in dissolved-solids concentrations in the aquifer as ground water is withdrawn from storage. The withdrawal of water from storage was simulated as a specified-pressure node (fig. 33). Supply wells in the Post Headquarters well field (fig. 2) were drilled mainly in a north-south line across the reentrant. The symmetry of the stress is similar to a line sink. Therefore, the linear symmetry of the cross-sectional model was used rather than converting to a model using radial symmetry. Node 208 was assigned a specified pressure that decreased in steps every 36.5 days during each simulation. A specific storage of 0.0000016 per foot was assigned to every element in the model.

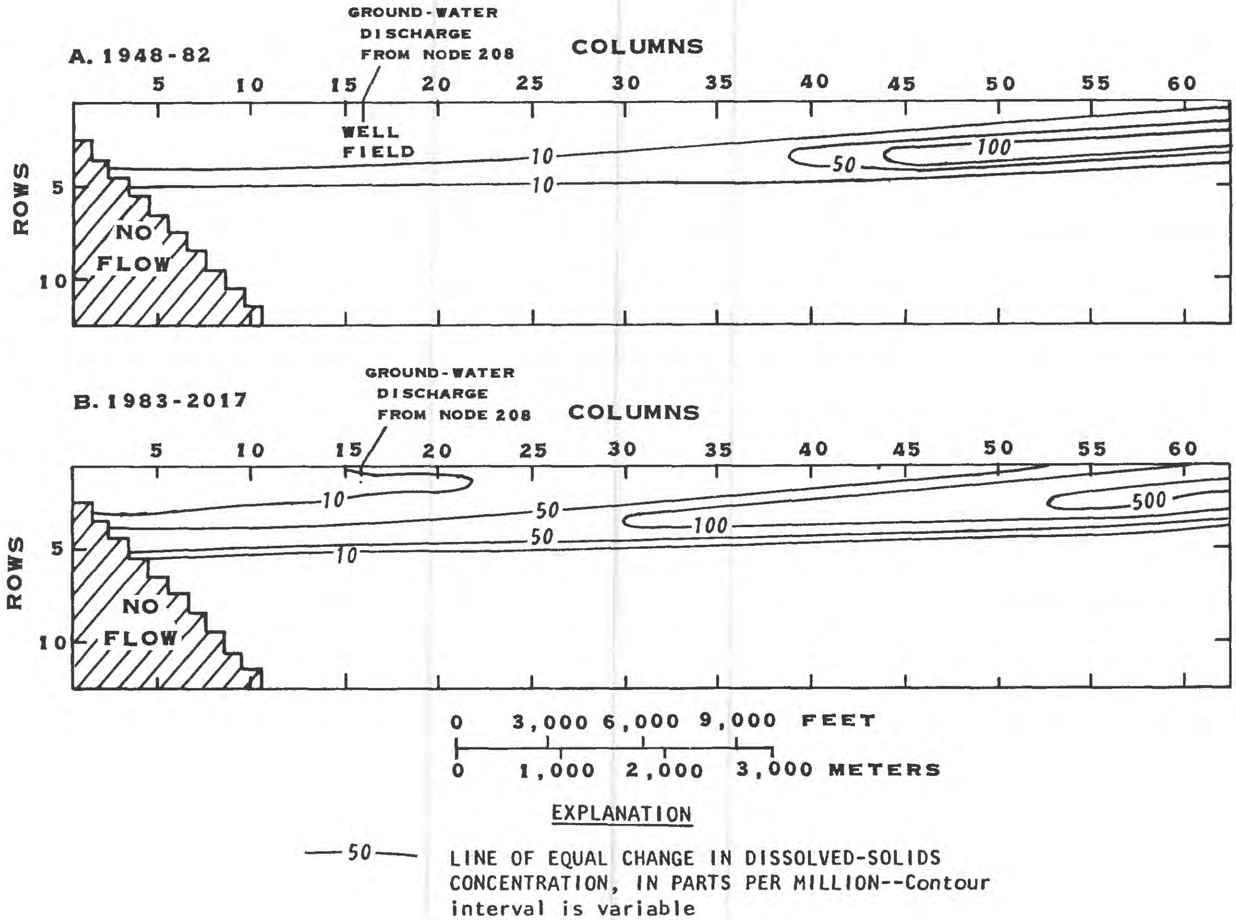


Figure 33.--Simulated change in dissolved-solids concentrations for 1948-82 and 1983-2017.

The model was used to simulate changes in dissolved-solids concentrations for 1948-82. During this time, dissolved-solids concentrations measured at supply wells increased as much as 200 milligrams per liter. The simulated change in dissolved-solids concentrations in the upper 2,000 feet of the aquifer for 1948-82 is shown in figure 33A. In the eastern part of the cross section, near the brine in the center of the Tularosa Basin, the change was greater than 100 parts per million; however, a change of less than 10 parts per million was simulated for the well field at node 208. The discrepancy between measured changes of 200 milligrams per liter and the simulated change of less than 10 parts per million for 1948-82 indicates that the measured change in water quality probably was not caused by a well-fieldwide movement of solute across the dispersion zone between freshwater and brine. The larger measured change probably indicates that solutes are moving preferentially through individual sand lenses that are connected to more mineralized water. In that respect, the model is not a complete representation of the real hydrologic system. To accurately simulate solute transport to the well field would require precise knowledge of the interconnection between individual sand lenses, which may be impossible to determine in an aquifer created from alluvial-fan deposits.

The model also was used to simulate concentration changes for 1983-2017 assuming that ground-water withdrawals and recharge remain constant, as previously described for the transient areal flow model (Plan 1). The simulated dissolved-solids concentrations (fig. 33B) increased throughout the upper 2,000 feet of the aquifer, especially in the zone of dispersion between freshwater and brine. This simulation indicates that dissolved-solids concentrations will increase by more than 500 parts per million in the eastern part of the aquifer. The simulation predicts only about a 10-parts-per-million increase in dissolved solids near the well field; however, if individual sands in the well field are connected to the eastern part of the area modeled, changes greater than 500 parts per million could occur.

Although the model does not accurately simulate water-quality changes at an individual well, it does show the relative change in water quality that can be expected in the aquifer as a result of ground-water withdrawals. Given that the ratio of vertical to horizontal hydraulic conductivity used in the model is reasonable, the lateral movement of more mineralized water from the east is more likely than upconing of water from beneath the well field. Therefore, observation wells that are located east of the well field and completed in the upper 600 feet of saturated aquifer probably will be useful in monitoring saline-water encroachment.

SUMMARY AND CONCLUSIONS

The nearness of the Post Headquarters well field to brine of the Tularosa Basin has caused concern that continued drawdown of the water table will allow saline water to move into the freshwater part of the aquifer. This study was conducted to investigate possible future water-level declines and water-quality changes in response to water withdrawals from the Post Headquarters well field. The major results and conclusions are summarized below.

1. Ground water is supplied to wells entirely from storage within the aquifer. Storage will continue to be depleted by an amount equal to the rate of ground-water withdrawal until the natural discharge from the aquifer begins to decrease. Rough calculations indicate that more than 1,000 years will be required before the natural discharge is affected by ground-water withdrawals. Therefore, the aquifer is not expected to reach an equilibrium condition regardless of whether the withdrawal rate is less than the average natural ground-water recharge rate.

2. The average ground-water recharge rate calculated from basin characteristics of the 46-square-mile Post Headquarters watershed is about 1,140 acre-feet per year. This rate produced a reasonable steady-state water table in the three-dimensional areal flow model and a reasonable freshwater thickness in the cross-sectional solute-transport model. However, the average recharge rate may be misleading because most ground-water recharge probably occurs in pulses after major storms. Analysis of water-level changes indicates that about 9,400 acre-feet of water was recharged from floods in 1978 and 1980.

3. The quality of water in the upper 500 feet of saturated bolson-fill sediments can be characterized by the major ions in solution. In the well field, where the aquifer is being recharged by freshwater from the mountain front, calcium and bicarbonate are the predominant ions. Surrounding this zone, ground water is relatively enriched in sulfate because of the solution of gypsum. The existence of sulfate-enriched water along the western margin of the reentrant probably indicates that most recharge to the aquifer occurs east of the Organ Mountain fault. Further east of the fault, the water is a sodium chloride sulfate brine typical of the Tularosa Basin.

An increase in dissolved-solids concentrations of as much as 200 milligrams per liter has been measured at Supply Wells 11, 13, 16, and 20. The change in quality appears to be caused by the addition of water from the zone of sulfate-enriched water surrounding the well field. Noticeable quantities of sodium chloride sulfate brine have not reached the supply wells.

4. A six-layer, finite-difference ground-water flow model was constructed to estimate effects of withdrawals on water levels. The hydraulic properties that produced the best fit between simulated and measured water levels are listed on the following page.

Layer number	Thickness (feet)	Horizontal hydraulic conductivity (feet per day)	Transmissivity (feet squared per day)	Vertical hydraulic conductivity (feet per day)	Specific yield	Storage coefficient
1	Variable	0.1-10	--	0.0080	0.15	--
2	200	--	20-2,000	.0080	--	0.0002
3	200	--	20-2,000	.0080	--	.0002
4	700	--	700	.0040	--	.0007
5	1,300	--	650	.0020	--	.0013
6	2,000	--	200	.0004	--	.0020

The model was able to reproduce measured ground-water declines for 1948-82 at most wells except in the southern part of the reentrant. Simulations indicate that Old Supply Well 12, where water levels have risen about 60 feet since 1952, or the entire southern part of the reentrant may not be hydraulically connected to other parts of the well field. Additional observation wells in this area would help clarify the connection between water-bearing zones.

Simulations of possible water-level changes for 1983-2017 were conducted assuming four future plans of ground-water withdrawal and wastewater-return flow shown below.

	Plan 1	Plan 2	Plan 3	Plan 4
Ground-water withdrawals, in acre-feet per year	2,066	2,066	1,033	1,033
Wastewater-return flow, in acre-feet per year	620	0	310	0

Water-level declines of as much as 50 feet were simulated near Supply Wells 10A and 17 in Plans 1 and 2 for 1983-2017. Eliminating wastewater-return flow in Plan 2 made little difference in simulated ground-water declines in the well field. However, near the wastewater-treatment plant, water levels declined more than 40 feet, which could make the well field more vulnerable to saltwater intrusion. Water levels increased in the well field during about the first 20 years of Plans 3 and 4 due to decreased ground-water withdrawals. After 20 years, water levels declined. At the end of 35 years, the net rise in water levels was still as much as 17 feet near Supply Well 17. Water levels in test wells furthest from the ground-water withdrawals declined throughout the 35-year simulation. This general pattern of water-level change will occur for any major reduction in ground-water withdrawals although the magnitude and timing of the change will depend upon the amount of reduction.

5. A two-dimensional finite-element model was used to simulate ground-water flow and solute movement in a vertical cross section through the Post Headquarters well field. Steady-state simulations using the recharge rates and hydraulic properties from the six-layer ground-water flow model and a brine density of 1.049 grams per cubic centimeter indicated that freshwater containing less than 1,000 parts per million dissolved solids is about 1,500 to 2,000 feet thick beneath the well field. The modeled result compares closely with geophysical evidence and lends support for the choice of hydraulic properties used in both models.

Transient simulations of the cross-sectional model indicate that solutes will move laterally toward the well field more rapidly than from below. The model could not duplicate measured increases in dissolved-solids concentrations at supply wells, which probably indicates that the measured increase of as much as 200 milligrams per liter at some supply wells is caused by movement of solutes through individual sand lenses with large hydraulic conductivity that could not be included in the model. Transient simulations for 1983-2017 indicate that if ground-water withdrawals and wastewater-return flow continue at 1982 rates, dissolved-solids concentrations will increase more than 500 parts per million in the eastern part of the well field. If individual sand lenses with large hydraulic conductivity connect supply wells to this area, dissolved-solids concentrations could change more than 500 parts per million at the supply wells.

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