SIMULATION OF THE FLOW SYSTEM OF BARTON SPRINGS AND ASSOCIATED EDWARDS AQUIFER IN THE AUSTIN AREA, TEXAS

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METRIC CONVERSIONS

For readers preferring to use metric units rather than inch-pound units, the conversion factors for the International System of Units (SI) and abbreviations for terms are listed below:

From	Multiply by	To obtain
acre-foot (acre-ft)	1,233	cubic meter
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
foot (ft)	0.3048	meter
foot per day	0.3048	meter per day
foot squared per day	0.09290	meter squared per day
gallon per day (gal/d)	0.003785	cubic meter per day
million gallons per year (Mgal/yr)	10.36	cubic meter per year
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer

DEFINITION OF TERMS

Anisotropic aquifer - An aquifer having one or more hydraulic properties that are not the same in all directions.

Aquifer - 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.

Artestian aquifer - A confined aquifer.

Confined aquifer - (artestian aquifer) An aquifer which is overlain by a relatively impermeable layer so that the water is under hydrostatic pressure. The water in an artesian well will rise above the top of the aquifer to the level of the potentiometric surface.

Constant-flux boundary - A model boundary condition that has a fixed value of volumetic flow rate per unit area (discharge) across the boundary.

Constant-gradient boundary - A model boundary condition that changes flow rate across the boundary as saturated thickness changes. The hydraulic gradient across the boundary remains constant with changes in saturated thickness.

Constant-head boundary - A model boundary condition that has a fixed value of static head, which is the height above a standard datum of the surface of a column of water that can be supported by the static pressure at a given point.

Digital model - A simplified mathematical representation of a complex system. A computer program to solve ground-water flow equations.

Discharge - Flow of water expressed as a volume per unit of time.

Ephemeral stream - A stream that flows briefly only in direct response to precipitation in the immediate locality and has a channel that is, at all times, above the water table.

Evapotranspiration - Volume of water lost through transpiration of plants and evaporation from the soil.

Geophysical log - A record obtained by lowering an instrument into a borehole or well and recording continously on a meter at the surface some physical properties of the rock material being logged.

Hydraulic conductivity - Volume of water a medium will transmit in unit time at the prevailing viscosity through a cross section of unit area, measured at right angles to the direction of flow, under a hydraulic gradient of unit change in head through a unit length of flow.

Hydraulic gradient - Rate of change of hydraulic head per unit of distance of flow at a given point and in a given direction.

Hydraulic head - Height of the free surface of a body of water above a given subsurface point.

Isotropic aquifer - An aquifer having one or more hydraulic properties the same in all directions.

Perennial stream - A stream that flows throughout the year.

Potentiometric surface - A surface representing the hydrostatic head. In an unconfined aquifer, the surface coincides with the water table. In a confined aquifer, the surface is defined by the levels to which water stands in tightly cased wells above the water body in the aquifer.

Recharge - Amount of water added to the zone of saturation.

Saturated thickness - Amount of water-bearing material filled with water under pressure greater than atmospheric.

Specific capacity - Rate of discharge of water from a well divided by the drawdown of water in the well. If constant except for the time variation, it is approximately proportional to the transmissivity of the aquifer.

Specific yield - Ratio of the volume of water that the material after being saturated will yield by gravity to the volume of the material.

Steady state - Equilibrium water levels of heads; water levels do not vary significantly with time.

Storage coefficient - Volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head.

Transient state - Nonequilibrium water levels of heads; water levels do vary significantly with time.

Transmissivity - rate at which water of the prevailing kinematic viscosity is transmitted through a unit width of an aquifer under a unit hydraulic gradient.

Unconfined aquifer - Aquifer in which a water-table body exists.

Water table - Surface in an unconfined water body at which the pressure is atmospheric. It is defined by the levels at which water stands in wells that penetrate the water body just far enough to hold standing water.

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ABSTRACT

A digital model of two-dimensional ground-water flow was used to estimate the hydraulic properties of the Edwards aquifer in a 151-square-mile area near Austin, Texas. The transmissivity, hydraulic conductivity, and specific yield were estimated for the part of the aquifer that discharges at Barton Springs in Austin. The aquifer is composed of the Edwards and overlying Georgetown Limestones of Cretaceous age and ranges in thickness from about 100 to about 450 feet.

More than 60 years of discharge measurements and 5 years of gaged discharge for Barton Springs were used to adjust springflow for the simulations. Barton Springs accounts for about 96 percent of springflow from the study area and 90 percent of the total discharge. The remaining discharge was pumpage from wells which was entered in the model. Four years of gaged recharge were used in the simulations. The potentiometric surfaces used by the models were constructed from water-level measurements in as many as 75 wells.

The transmissivity was calibrated through steady-state simulations that used the mean value of recharge and mean potentiometric surface to represent average conditions for the aquifer. The transmissivities vary from about 100 feet squared per day in the western part of the aquifer to more than 1 million feet squared per day near Barton Springs. Specific yield was calibrated through transient-state simulations for 5 consecutive months using time-dependent data for recharge, discharge, and water levels. The mean specific yield for the aquifer is 0.014 and ranges from 0.008 to 0.064. Additional aquifer properties used in the simulations include storage coefficient, altitudes of the base and top of the aquifer, and hydraulic conductivity.

A simulation for the year 2000 using projected pumping rates for municipal, industrial, agricultural, and domestic supplies indicates that the aquifer would be dewatered in the southwestern part of the study area and have large declines in the southeastern part of the study area. Another simulation of projected conditions using potential recharge enhancement predicts a rise in the potentiometric surface of about 50 feet in the southwestern part of the aquifer and moderate water-level declines in the southeastern part of the aquifer.

INTRODUCTION

Barton Springs is located in Zilker Park near the center of Austin and is a major recreational attraction for the city. The Austin Parks and Recreation Department reports more than 300,000 paid attendances annually to the Barton Springs swimming pool the last few years. Discharge from Barton Springs sustains flow in Barton Creek and enters Town Lake, which serves as a source of drinking water for the city of Austin.

As of 1981, the Edwards aquifer from which Barton Springs issues provides water for municipal, industrial, agricultural, and domestic uses for about 15,000 people in the study area. Many communities, such as Sunset Valley, San Leanna, Manchaca, Hays, Buda, and Kyle, depend solely on the Edwards aquifer for their water supply.

Much of the land within the aquifer area is being rapidly developed. Recent population projections by the city of Austin indicate that by the year 2000, about 86,000 more people will live in the aquifer area, many of whom will depend on the Edwards aquifer for water. The water supply of the aquifer is sufficient to handle current (1985) rates of pumpage. However, as the aquifer use increases, the resulting pumpage could decrease the availability of ground water and could decrease or even stop the flow of Barton Springs. Determination and evaluation of the hydrologic, geologic, and hydraulic properties of the aquifer need to be made in order to predict and evaluate the regional effects of future stresses on the aquifer.

Purpose and Approach

This report is a summary of a study to quantify the areal distribution of selected hydraulic properties of the Edwards aquifer by use of a mathematical model. The mathematical model was used to test the conceptual model of the aquifer and to simulate two possible plans of aquifer management being considered by local governing officials. One simulation indicates the effect of projected pumpage on the present potentiometric surface, and another simulation indicates the combined effect of projected pumpage and potential recharge enhancement on the potentiometric surface.

Pertinent hydrogeologic information was used to design and develop a conceptual and mathematical model of the Edwards aquifer. Much of the hydrologic data collected for this study is published in annual reports by Slade and others (1980, 1981, 1982, 1983, 1984) and Gordon and others (1985). Information concerning the wells and test holes are given in the aforementioned reports. A brief geologic description and generalized information concerning the frequency, period of record, and explanation of the hydrologic data collected and used for the model are given in the "Hydrogeologic Framework" section of this report.

Models simulating the flow in the aquifer were developed using: maps showing the altitude of the base and the top of the Edwards aquifer; potentiometric-surface maps; long-term mean and short-term periodic discharges of Barton Springs; location of major well fields and rates of pumpage from the Edwards aquifer; and location of recharge areas and rates of surface recharge to the aquifer. A grid system was developed for the aquifer, and all the above data were coded for appropriate cells of the grid. All of the hydrologic, hydraulic, and geologic data required by the model are collectively referred to as "model data" and are explained within this report.

Steady-state flow in the aquifer was simulated assuming a long-term mean discharge of 50 ft^3/s. For this simulation, the mean discharge and recharge were assumed to occur continuously and to result in the average potentiometric surface. Transmissivities were estimated, entered in the model, and adjusted until the potentiometric surface computed by the model reasonably matched the actual potentiometric surface.

Transient-state flow in the aquifer was then simulated for a 164-consecutive-day "dry" period during which recharge volumes were small. Time-dependent data for recharge, discharge, and ground-water levels were used in the simulation. Because most of the discharge during this period came from storage, specific yields, to which the simulation was most sensitive, were estimated and adjusted until a reasonable agreement between measured and simulated water levels and discharges was obtained.

Population-increase projections for 2000 were obtained from the city of Austin, along with their estimate of which areas would be using water from the Edwards aquifer at that time. The estimated pumpage for that year was entered into the calibrated steady-state model, and the potentiometric surface was computed.

Local governing officials are studying a proposal to build a runoff-control structure on Onion Creek, just upstream from the recharge zone. A projection of the effect of this structure on the mean recharge for Onion Creek was made. This "enhanced" recharge, along with the projected water demands for 2000, were entered into the calibrated model, so that the combined effect on the potentiometric surface could be determined.

Location and Extent of the Study Area

The Edwards aquifer, which provides water for at least nine counties in central Texas, exists north and south of the study area. Many large springs discharge from hydrologically independent areas of the aquifer. Barton Springs and the associated Edwards aquifer are located in and around the metropolitan area of Austin, Texas (fig. 1). The study area extends from Kyle on the south to the Colorado River on the north, and includes parts of Hays and Travis Counties where the Edwards aquifer contains water having less than 1,000 mg/L (milligrams per liter) dissolved solids. The study area is about 155 mi^2, of which all but 4 mi^2 coincides with that part of the Edwards aquifer that discharges at Barton Springs.

The northern boundary of the study area is a no-flow boundary formed by the Colorado River (Town Lake). Water levels in the Edwards aquifer north and south of the Colorado River decline with proximity to the river, and many springs near or innundated by the river discharge from the aquifer. The western boundary is the westernmost extent of the Edwards aquifer. The southern boundary is a ground-water divide. South of this divide, ground water dis-

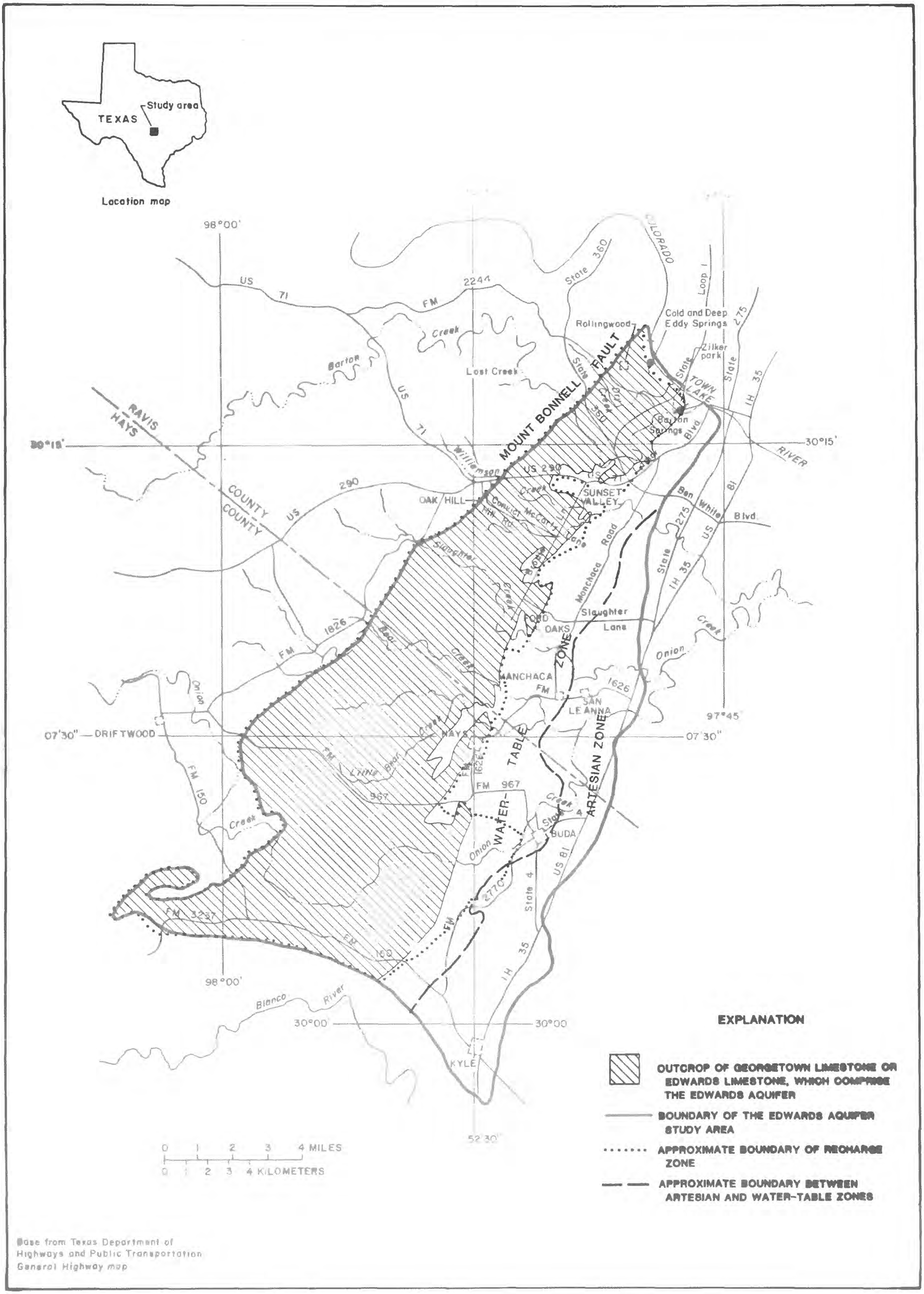

Figure 1.--Location of study area.

charges to San Marcos Springs, about 8 mi south of Kyle (Petitt and George, 1956, p. 3). The eastern boundary of the aquifer is the locally referred to "bad-water" line where water circulation to Barton Springs is decreased substantially. East of the line, the dissolved-solids concentration of the water is greater than 1,000 mg/L. The line is well defined and fairly stable. Leakage from east of the line into the aquifer is minimal and only detectable during extreme low-flow conditions.

HYDROGEOLOGIC FRAMEWORK

Geology

The Edwards aquifer is composed of the Edwards Limestone and Georgetown Limestone of Cretaceous age that dip eastward and strike northward[1/]. The aquifer crops out in the western part of the study area (fig. 1). A summary of the characteristics of the geologic units of the Edwards aquifer and adjoining formations is presented in table 1.

The Edwards aquifer is underlain and bound on the west by Cretaceous rocks that are older than the aquifer. These rocks include, from youngest to oldest, the Walnut Formation, which is as much as 60 ft thick, and the Glen Rose Limestone, which is 500 to 900 ft thick. The Walnut Formation yields little or no water in the study area. The Glen Rose Limestone yields small quantities of water that is chemically distinct and more saline than water from the Edwards aquifer.

Cretaceous rocks younger than the Edwards aquifer overlie the aquifer and extend eastward at the land surface. These formations include, from oldest to youngest, the Del Rio Clay and the Buda Limestone. Neither formation is known to yield water in the study area. The Del Rio Clay, which is 60 to 75 ft thick, is relatively impermeable and forms an upper confining layer of the Edwards aquifer. The approximate divide between the confined and unconfined zones of the aquifer are shown in figure 1. The Buda Limestone, which is 35 to 50 ft thick, is not known to yield water in the study area.

Geophysical well logs, lithologic descriptions of well logs, surface geology, and drillers' logs were used to determine the position of the Edwards aquifer in the subsurface. Baker and others (in press) presented two hydrogeologic sections along the dip of the aquifer and one hydrogeologic section along strike, as well as maps showing the altitude of the base and top of the aquifer. Large variations in the angle of dip as well as variations in the altitude and depth to the top occur locally within short distances due to faulting. The fault blocks are normally stair-stepped downward to the east. There are vertical displacements of as much as 200 ft along faults. However, no evidence has been presented to indicate that the aquifer is discontinuous, thus, ground-water flow probably is not greatly impeded by the faults. The aquifer is eroded where it outcrops; in the subsurface, the aquifer thickness varies from about 400 ft at the northern boundary of the study area to about 450 ft at the southern boundary.

1/ The stratigraphic nomenclature and descriptions used in this report are adapted from Rodda and others (1970), Garner and Young (1976), and Brune and Duffin (1983).

Table 1.--Characteristics of geologic units

System	Series	Group	Formation and member		Hydrogeologic unit	Thickness (feet)	Lithology
Cretaceous	Comanchean	Washita	Buda Limestone			35-50	Gray to tan, hard, resistant, glauconitic shell-fragment limestone and a lower marly, nodular, and less resistant limestone.
			Del Rio Clay		Confining bed	60-75	Dark gray to olive-brown, calcareous fossiliferous clay containing selenite and pyrite.
			Georgetown Limestone		Edwards aquifer	40-100	Thin interbeds of gray to tan, fine-grained, fossiliferous limestone with layers of marly limestone and marl.
		Fredericksburg	Edwards Limestone	Member 4		40±	Hard, dense, thick to thin-bedded, fine-grained limestone; soft dolomitic limestone and solution collapse zone near middle.
				Member 3		10-15±	Soft, nodular marly limestone and marl interbedded locally with flaggy limestone.
				Member 2		40±	Fine- to medium-grained, hard, thick- to thin-bedded limestone. Lower beds folded and fractured as a result of collapse in member 1.
				Member 1		200-250±	Porous dolomite and dolomitic limestone. Nodular chert common. A solution collapse zone within this member creates cavernous and vugular porosity.
			Walnut Formation		Confining bed	15-60	Hard, fine- to medium-grained fossiliferous limestone with layers of fine-grained marl, marly limestone, and nodular limestone.
		Trinity	Glen Rose Limestone	Upper member	Upper Trinity aquifer	500-900	Alternating beds of limestone, dolomite, and marl. Some anhydrite and gypsum.
				Lower member	Upper part of Middle Trinity aquifer		Massive, fossiliferous limestone and dolomite at base grading upward into thin beds of limestone, shale, marl, and gypsum. Corbula martinae bed at top.

Modified from Brune and Duffin (1983, table 1).

Irregular dissolution of the limestones has created porosity, which greatly affects the hydraulic properties of the aquifer. Significant porosity along particular bedding planes was created through dissolution by meteoric water during an interval of subaerial exposure at the close of the Edwards Limestone period of deposition (Abbott, 1976). Vertical zones of greater porosity are a result of steep-angle normal faulting that began during the Miocene Epoch. Ground water undersaturated with respect to calcite and dolomite dissolved these minerals and increased the porosity, a process that may still be occurring. The vertical zones of greater porosity along faults in the outcrop also allow surface water to readily enter and move through the unsaturated zone to the water table.

Interpretation of geophysical and drillers' logs for 79 wells in the study area indicates that most wells which penetrated cavities are located very close to known faults (R. M. Slade, Jr., and others, written commun., 1985). This supports the idea that many of the cavities within the aquifer are associated with faults. The surface traces of the mapped faults in the study area (fig. 2) generally are parallel to the western boundary of the aquifer and trend northeast; thus, many cavities probably are aligned in this direction. As expected, the wells that penetrated cavities had the maximum specific capacities of all wells completed in the aquifer. Greater values of transmissivity are, therefore, probably prevalent along the direction of the faults. However, there are many faults that cross the main fault direction at acute angles (fig. 2), thus, the principal direction of greatest transmissivity may vary locally.

Hydrology

Potentiometric Surface

From 1978 to 1982, water levels in about 75 wells completed in the Edwards aquifer were measured annually. Beginning in 1979, about 19 wells were measured monthly, and from December 1982 to September 1983, 24 wells were measured monthly. These water levels and water levels measured by the Texas Department of Water Resources were used to construct maps of the potentiometric surface of the Edwards aquifer for various rates of spring discharge.

The annual water-level measurements in January 1981 were made when Barton Springs was flowing at its long-term mean discharge of 50 ft^3/s. The potentiometric surface at this time was used for the steady-state simulation (fig. 3). During January 1981, ground-water flow generally was to the east in the western part of the aquifer and to the northeast, toward Barton Springs, in the eastern part of the aquifer.

Water-level fluctuations between high- and low-flow periods vary considerably between wells. The maximum measured water-level fluctuations for all observation wells completed in the aquifer with more than 5 years of data are given in figure 4. As of 1982, no temporal trends of ground-water declines had been identified as a result of annual pumpage increases; fluctuations are thought to be caused by variations in recharge and periodic variations in withdrawals. Generally, the maximum fluctuations in the western, central, and eastern parts of the aquifer are, respectively, about 1 to 10 ft, 10 to 50 ft, and 40 to 119 ft. The greatest fluctuations occur where water in the aquifer is confined. The confined zone is about 21 percent of the study area.

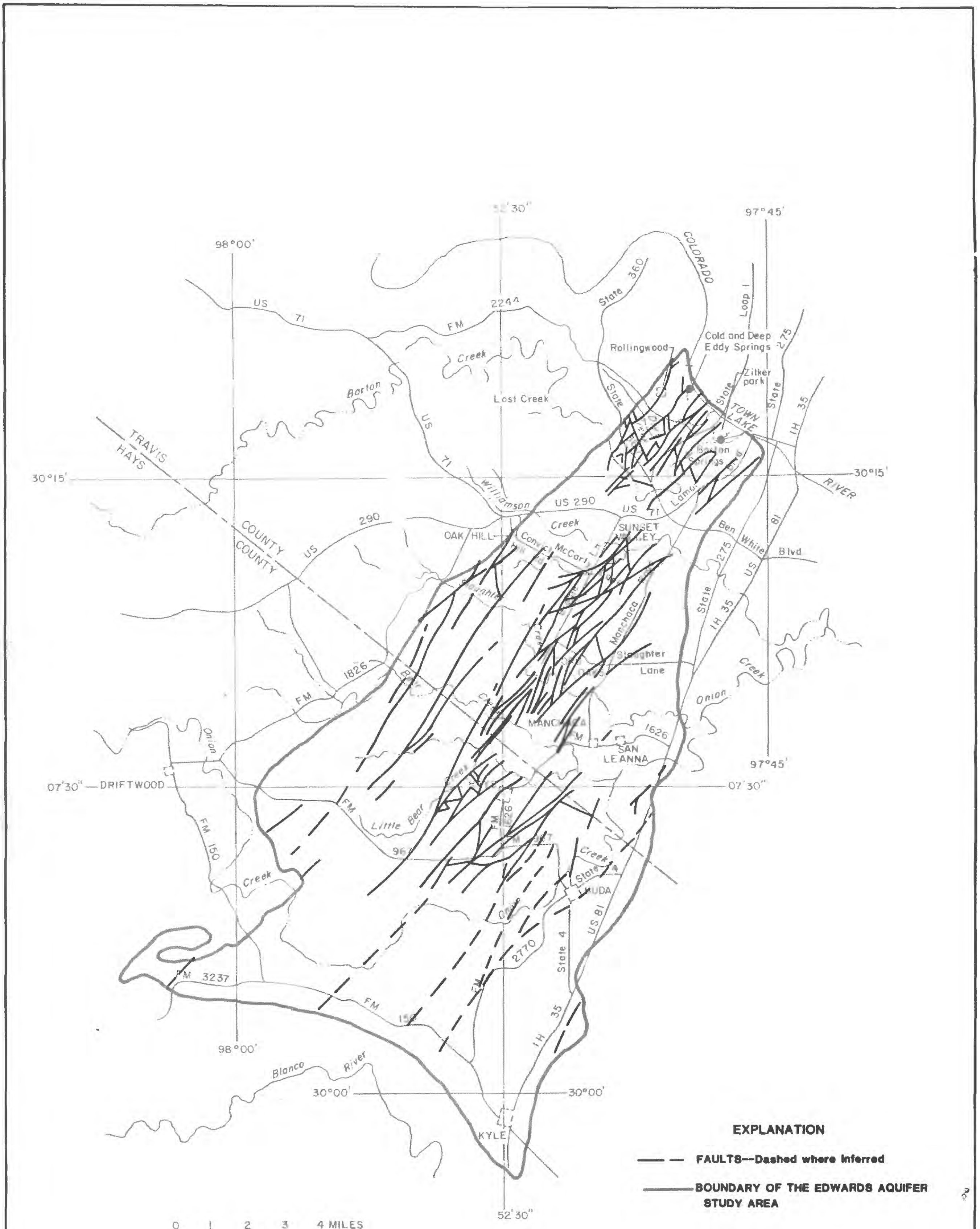

Figure 2.--Traces of major faults.

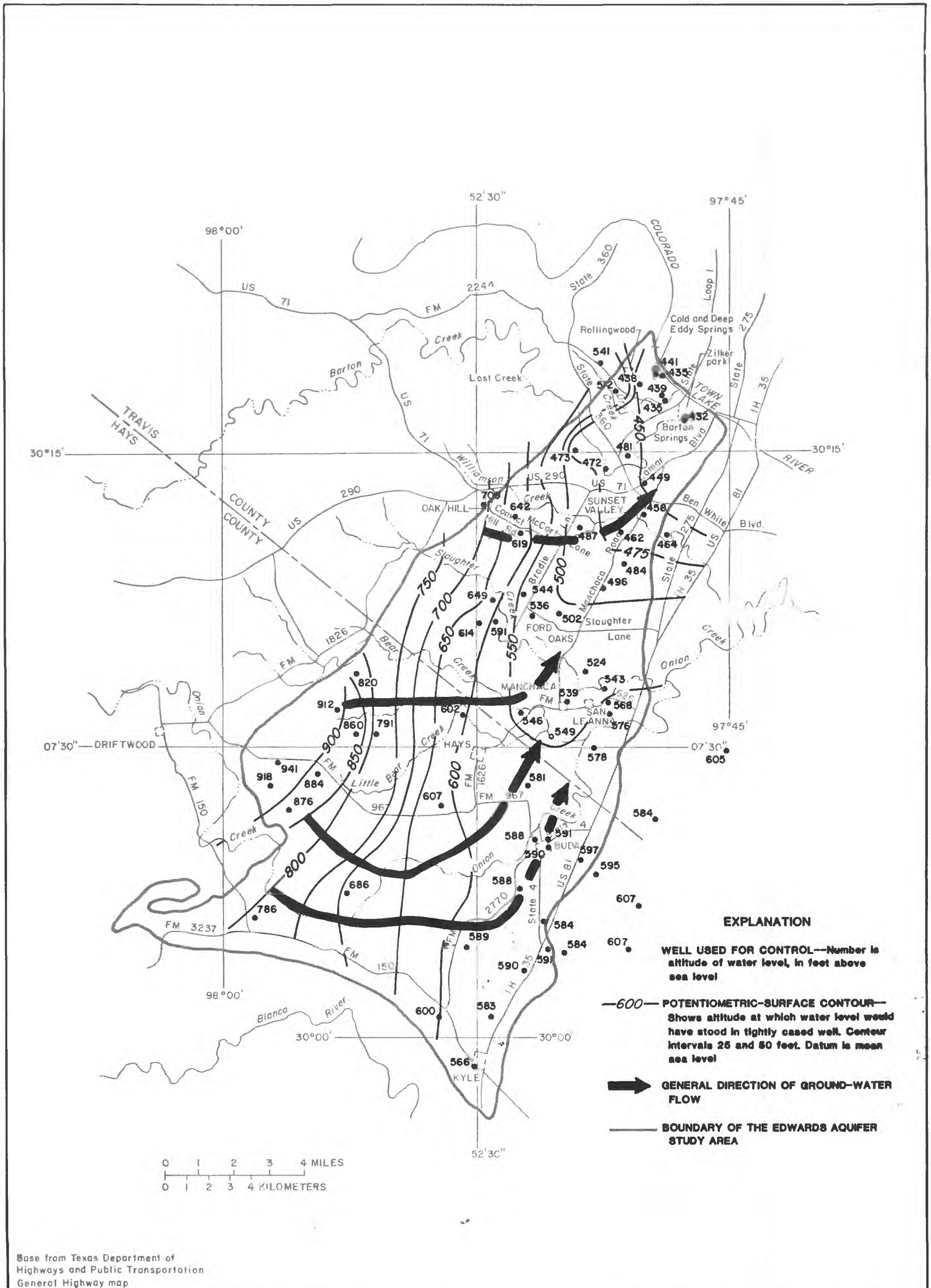

Figure 3.--Potentiometric surface, January 1981.

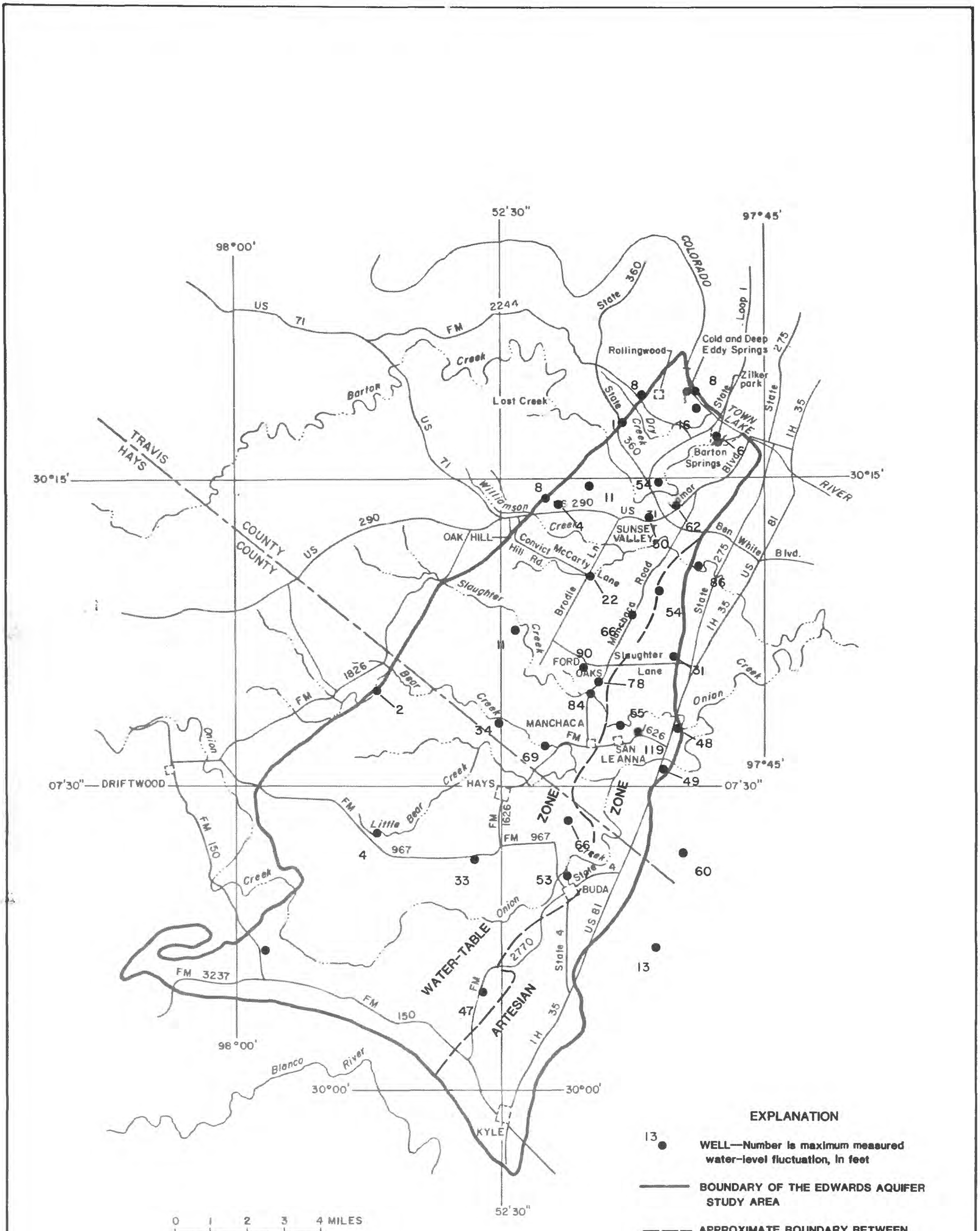

Figure 4.--Maximum measured water-level fluctuations for selected wells.

Recharge to the Aquifer

Recharge to the Edwards aquifer is from the infiltration of runoff in six watersheds that cross the recharge zone. Flow-loss studies were conducted on five of these creeks in order to determine the boundaries of the recharge reach for each creek and the quantity and location of the flow losses. By July 1979, streamflow-gaging stations were installed at the upstream and downstream boundary of the recharge reach on each creek, so that flow-loss volumes could be calculated. The method of estimating recharge to the Edwards aquifer is presented by Garza (1962). Recharge consists of the infiltration of streamflow plus direct infiltration of runoff in the interstream areas. A water-balance equation was used to estimate recharge in each watershed. Recharge within a watershed is the difference between gaged streamflow upstream and downstream from the recharge area plus the estimated runoff in the intervening area. The intervening area is the drainage area within the recharge zone between the two streamflow-gaging stations in each watershed. Runoff from that area is estimated on the basis of unit runoff from the area upstream from the recharge zone.

Based on data from the streamflow-gaging stations, about 85 percent of the total recharge in the watersheds occurs along the main channels of the 6 creeks, and the remaining 15 percent occurs along channels of tributaries and by diffuse infiltration between these creeks within the recharge zone. The recharge zone has an area of about 90 mi^2 (fig. 1). The drainage area of the watersheds upstream from the recharge zone is about 264 mi^2.

Monthly volumes of recharge were determined for each watershed for July 1979 to December 1982 and are given in table 2. These determinations are believed by the authors to be accurate to within 15 percent of actual values. Based on these data in table 2, the contribution to total recharge by watershed is:

Watershed	Percent of total recharge to aquifer
Barton Creek	28
Williamson Creek	6
Slaughter Creek	12
Bear Creek	10
Little Bear Creek	10
Onion Creek	34

The flow-loss studies also provided the means to define the distribution of recharge within reaches of each creek. Maximum recharge capacities were estimated from the flow-loss studies and the records of streamflow at the gaging stations. The maximum recharge capacity for the main channels of the creeks during sustained flow conditions has been computed or estimated as follows:

Table 2.--Recharge by watersheds and Barton Springs discharge, July 1979-December 1982

Year	Month	Monthly recharge by watershed (acre-feet) Barton Creek	Williamson Creek	Slaughter Creek	Bear Creek	Little Bear Creek	Onion Creek	Total recharge (acre-feet)	Barton Springs discharge (acre-feet)
1979	July	1,020	198	654	496	452	1,190	4,010	6,030
	Aug.	652	67	304	433	380	595	2,430	5,730
	Sept.	151	99	65	138	120	278	850	4,980
	Oct.	41	2	15	90	78	231	460	4,220
	Nov.	33	0	2	50	44	128	260	3,250
	Dec.	37	30	2	68	59	119	310	2,800
1980	Jan.	90	10	1	50	44	151	350	2,370
	Feb.	114	58	12	80	70	184	520	2,110
	Mar.	493	64	116	229	196	271	1,370	2,170
	Apr.	2,000	246	203	296	258	765	3,770	2,490
	May	3,010	1,300	3,850	1,080	1,710	3,170	14,120	3,850
	June	1,020	641	207	526	461	2,510	5,370	4,230
	July	54	0	12	56	50	190	360	3,490
	Aug.	8	0	0	8	7	79	100	2,560
	Sept.	309	119	48	236	242	912	1,870	2,200
	Oct.	1,710	27	52	613	536	1,940	4,880	2,840
	Nov.	1,510	20	96	233	204	1,370	3,430	2,580
	Dec.	3,260	80	399	585	512	2,730	7,570	3,060
1981	Jan.	1,800	45	313	366	320	2,070	4,910	2,980
	Feb.	1,150	240	362	611	535	2,390	5,290	2,930
	Mar.	4,460	1,500	2,010	1,540	1,780	3,570	14,860	4,070
	Apr.	1,330	258	339	683	597	3,170	6,380	3,780
	May	793	1,220	425	460	402	1,960	5,260	3,540
	June	1,390	1,980	3,970	2,580	2,580	3,470	15,970	4,830
	July	1,190	503	546	883	773	3,570	7,470	6,270
	Aug.	710	10	36	201	176	957	2,090	5,770
	Sept.	220	40	23	178	156	794	1,410	5,110
	Oct.	2,830	40	208	484	422	2,580	6,560	5,270
	Nov.	1,650	10	341	244	212	2,000	4,460	4,960
	Dec.	832	.5	129	148	130	1,110	2,350	4,580
1982	Jan.	504	.1	75	99	87	698	1,460	3,700
	Feb.	241	.1	43	56	49	480	870	2,910
	Mar.	262	20	39	50	44	368	780	2,830
	Apr.	855	347	339	124	109	579	2,350	2,560
	May	3,370	400	2,370	1,010	1,080	2,780	11,010	3,790
	June	2,020	40	538	460	486	1,680	5,220	4,050
	July	319	0	63	151	132	533	1,200	3,480
	Aug.	44	50	.9	19	17	104	230	2,690
	Sept.	6	2	.2	3	3	127	140	2,140
	Oct.	5	.1	0	15	13	82	120	2,030
	Nov.	6	64	1	110	97	141	420	2,020
	Dec.	11	60	25	98	85	208	490	2,520
Total		41,510	9,790	18,230	15,840	15,710	52,230	153,310	149,770

Creek	Maximum potential recharge (cubic feet per second)
Barton	30 to about 70
Williamson	13
Slaughter	52
Bear	33
Little Bear	About 30
Onion	About 120

Maximum recharge capacities during floodflows probably are greater than these values because of increased stream depths and larger areas with faults or other openings in contact with recharge water. Maximum recharge rates during floods cannot be accurately determined from discharge measurements or gaged flow, because the flow is unsteady. Total maximum instantaneous recharge is estimated to be about 350 to 400 ft^3/s. Runoff that exceeds the maximum potential recharge rate is rejected. The flow occurring at the gages at the downstream end of the recharge zone is runoff that is rejected.

There is evidence that subsurface leakage to the Edwards aquifer from the underlying upper Trinity aquifer (table 1) occurs at specific locations (R. M. Slade, Jr., and others, written commun., 1985). The inorganic-chemical characteristics of water from the Edwards aquifer differ considerably from the inorganic-chemical characteristics of water in the upper Trinity aquifer. Analyses of water from wells completed in the Edwards aquifer indicate that water from the upper Trinity aquifer leaks into the Edwards aquifer at certain locations, probably along faults. Only 13 of the 140 wells completed in the Edwards aquifer for which water-quality data are available indicate evidence of this leakage. Also, since permeability probably is much greater along faults, it is possible that water from the upper Trinity aquifer is induced into the Edwards aquifer by stresses from pumpage; thus that water is not necessarily in hydrologic circulation within the Edwards aquifer. Evidence concerning subsurface leakage indicates it to be limited in magnitude, if occurring at all. A further discussion of this subject is presented in the section "Limitations and Future Simulation Studies."

Discharge from the Aquifer

Discharge from the Edwards aquifer in the study area is by springflow and pumpage from wells. Except at the springs mentioned below, water levels throughout the aquifer are well below the land surface, thus no known evapotranspiration occurs from the aquifer. Cold and Deep Eddy Springs, located about 1.5 mi northwest of Barton Springs, consistently flow between 3 and 4 ft^3/s (Brune, 1975). These springs drain only about 4 mi^2 of the extreme northwestern part of the aquifer. This part of the aquifer is recharged by Dry Creek and is believed to be hydraulically independent of the rest of the aquifer, therefore, it was excluded from the simulation model.

The remaining 151 mi^2 of the Edwards aquifer supplies water to Barton Springs and several intermittently flowing springs. The intermittently flowing springs are in the bed of Barton Creek between State Highway 360 and Barton Springs. These springs flow less than 30 percent of the time, when ground-water levels are above the bottom of the creek at these locations, and account for only about 4 percent of total springflow.

Barton Springs accounts for the remaining 96 percent of springflow. In March 1978, a water-level recorder was installed in well YD-58-42-903, located about 200 ft from the main springs. Springflow measurements are correlated to the water levels in the well. The recorder on the well makes hourly water-level measurements that are used, along with the water level-discharge relationship, to compute the daily-mean discharges for Barton Springs. The monthly discharges of Barton Springs for July 1979 through December 1982 are given in table 2.

The monthly mean values of discharge for Barton Springs for 1917-82 have been computed based on 725 discharge measurements made during 1917-78 and on the daily-mean flows for 1979-83. Based on these monthly mean flows, the mean discharge of Barton Springs is 50 ft^3/s, and the median discharge is 46 ft^3/s. The maximum and minimum measured flows were 166 and 10 ft^3/s, respectively.

Several hundred wells in the study area supply water for municipal, industrial, domestic, and agricultural (livestock and irrigation) users. As of 1981, the total pumpage from all wells completed in the Edwards aquifer was about 3,800 acre-ft per year, which is about 10 percent of total long-term mean discharge. There are 21 identified major well fields pumping water from the aquifer, all of which report their annual pumpage to the Texas Department of Water Resources. The total pumpage from these wells, which represents the total municipal and industrial use, is about 879 Mgal/yr (2,900 acre-ft), or 76 percent of total pumpage. Most of these wells do not have metering systems to determine pumpage, thus those volumes are estimated and the accuracy is uncertain. The remaining 293 Mgal/yr (900 acre-ft) is the estimated pumpage from several hundred users for domestic and agricultural supplies (Comer Tuck, Texas Department of Water Resources, oral commun., 1984). The municipal, industrial, domestic, and agricultural uses are 43, 33, 20, and 4 percent of total pumpage, respectively. The name and pumpage rate for each major well field are given in table 3, and the location of each is shown in figure 5.

Water-Budget Analysis

A water-budget analysis was computed for that part of the aquifer discharging to Barton Springs. December 1979 to July 1982 was chosen for this analysis because the potentiometric surfaces at the beginning and end of this period are similar, thus the change in storage is minimal and total recharge should equal total discharge. Recharge and discharge occurring through adjacent aquifers are unknown, and thus not included, therefore the analysis may not necessarily be complete with respect to total recharge and discharge. However, flow through adjacent aquifers is believed to be relatively small. Recharge used for this analysis was from the six watersheds discussed in the previous section of this report. Discharge from the aquifer was computed by adding the total springflow to the total pumpage. Total springflow was estimated by increasing the discharge of Barton Springs by 4 percent to account for the flow from the intermittent springs in Barton Creek. The cumulative recharge and discharge for

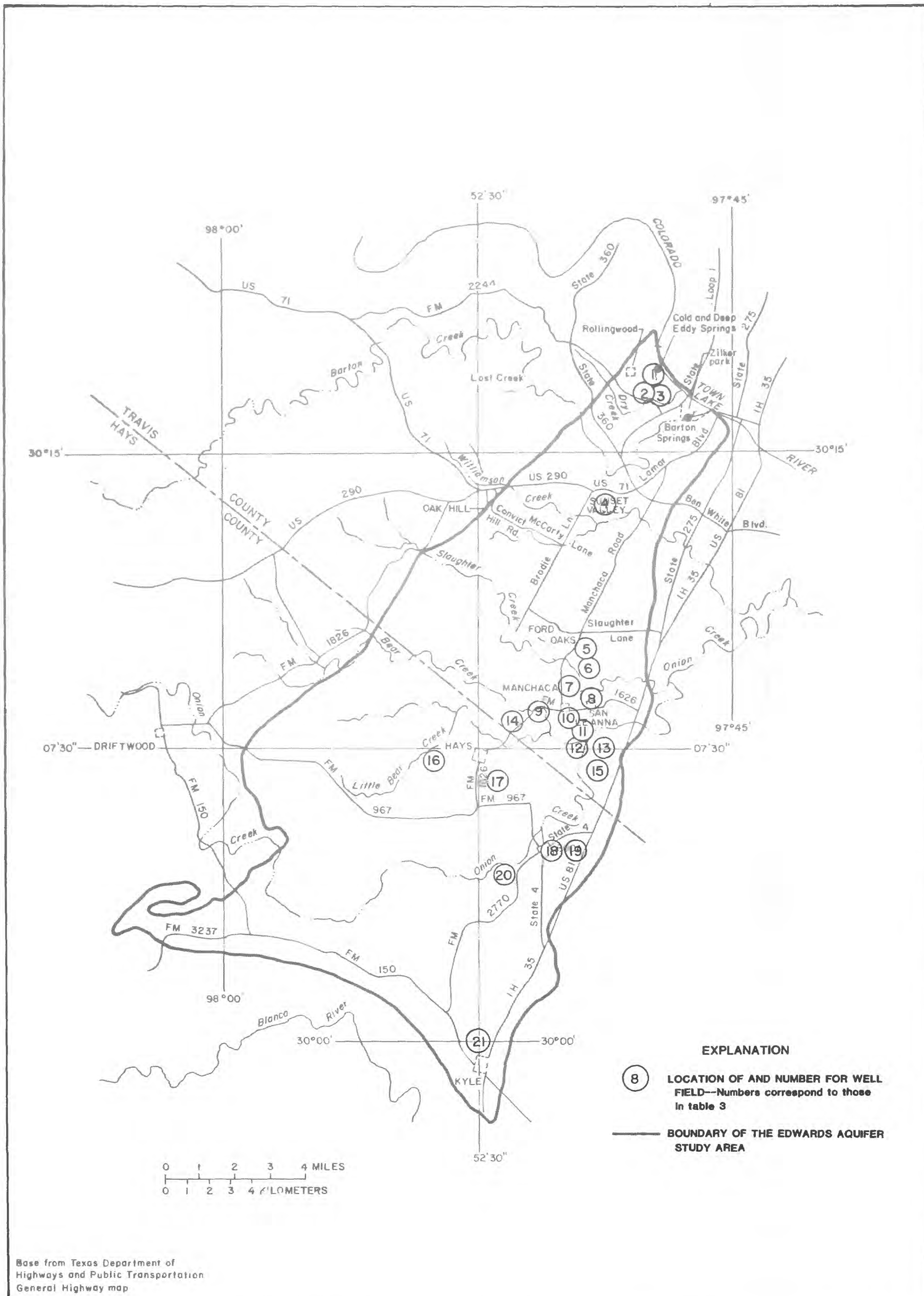

Figure 5.--Location of major well fields.

Table 3.--Major well fields and pumpage, 1981

Well field number (figure 5)	Name of well field	Pumpage (million gallons per year)	Daily mean pumpage (cubic feet per second)
1	Ridgewood Village	10.38	0.04
2	G&J Water Supply	2.83	.01
3	Dellano Hills	1.42	.01
4	City of Sunset Valley	6.60	.03
5	Malone Addition	3.30	.01
6	Slaughter Creek Acres	6.60	.03
7	Mooreland	2.12	.01
8	City of San Leanna	4.95	.02
9	Bear Creek Park	7.08	.03
10	Twin Creek Park	4.72	.02
11	Creedmor Maha Water Supply Co.	81.37	.35
12	Arroyo Doble	22.88	.10
13	Mystic Oaks	4.01	.02
14	Marbridge Foundation Ranch	50.71	.22
15	Onion Creek Meadows	12.26	.05
16	Chaparal Park	10.38	.04
17	Leisurewood	11.32	.05
18	City of Buda	44.81	.19
19	Goforth Water Supply Co.	41.27	.17
20	Centex Materials	404.01	1.71
21	City of Kyle	146.23	.62
	TOTAL	879.25	3.37

this period is about 144,000 and 128,000 acre-ft, respectively. Recharge exceeds discharge by only 12 percent, which is within the accuracy of the calculated volumes, thus surface recharge probably is in dynamic equilibrium with springflow and pumpage. Because the surface sources of recharge and discharge are comparable, subsurface sources probably are comparable also. Geochemical and hydrologic evidence indicate that subsurface recharge is limited in magnitude if it occurs at all; the magnitude of subsurface discharge probably is comparable. (See "Recharge to the Aquifer" section.)

Because surface recharge and discharge probably are in equilibrium, the long-term mean discharge of Barton Springs (50 ft^3/s or about 36,000 acre-ft per year) probably approximates the long-term mean recharge. The 1981 estimated pumpage withdrawals of 3,800 acre-ft per year thus represent about 10 percent of the long-term mean recharge.

DESCRIPTION OF THE MATHEMATICAL MODEL

Mathematical Model and Solution Techniques

The two-dimensional, finite-difference ground-water flow model described by Trescott, Pinder, and Larson (1976) was used to simulate the ground-water flow. The partial-differential equations governing ground-water flow are approximated by the model. Pertinent hydraulic and hydrologic characteristics of the aquifer are used to compute water levels. The strongly implicit procedure was used to solve the ground-water flow equations for the hydraulic-head values.

The steps in applying the finite-difference approach to ground-water movement are as follows:

1. A finite-difference grid is superimposed upon a map showing the extent of the aquifer, thus for computational purposes the continuous aquifer is simulated by a set of discrete blocks (cells).
2. Pertinent hydraulic and hydrologic characteristics of the aquifer are coded for appropriate cells within the grid.
3. The governing partial-differential equation is written in finite-difference form for each of the discrete cells.
4. The resulting set of linear finite-difference equations are solved numerically for hydraulic head at the center of each cell (or node) by a digital computer.

Assumptions

The digital-computer model of an aquifer is a mathematical representation of the system and is not as complex as the real system. Simplifications of the aquifer characteristics are necessary in order to represent the aquifer system. The assumptions used in the model formulation include the following:

1. The movement of water in the aquifer is assumed to occur only in two dimensions. This assumption is invalid where appreciable vertical gradients exist. On a regional basis, however, the vertical gradient In the Edwards aquifer is sufficiently small so that model results are probably not significantly affected by this assumption.

2. The hydraulic properties are constant throughout the area of each cell of the finite-difference grid. This assumption may contribute large errors to the modeling results. The hydraulic characteristics of the aquifer are extremely variable because the porosity and transmissivity are formed by irregular dissolution of the limestones (see "Geology" section). Wells only a short distance apart commonly penetrate parts of the aquifer that have hydraulic characteristics that differ by a factor of 100 or more. The values of the hydraulic data used for the model are considered "average" values for the entire cell area. The model is accurate on a regional basis, but will not necessarily be accurate for specific areas where the hydraulic data vary considerably from the values used in the model.

3. Within each cell, the values for specific yield and hydraulic conductivity are assumed to be constant through the entire saturated thickness. Because much of the water is in cavities, this assumption is a source of error for certain simulations. The model has been calibrated to simulate water levels based on current aquifer stresses of recharge and pumpage. Simulations involving large deviations from the calibrated water levels are subject to error because the vertical homogeneity of the hydraulic characteristics cannot be verified.

4. The simulations assume all recharge to the aquifer occurs within the recharge zone and no subsurface leakage occurs into or from the aquifer. A water-budget analysis described in the "Water-Budget Analysis" section of this report, showed that this probably is a valid assumption.

5. Stress, in the form of pumpage or recharge, is assumed to be evenly distributed throughout the area of the cells. The hydraulic heads in the model are computed at each node.

ther specific assumptions, made for the simulations presented within this report, are presented in the appropriate sections.

DESIGN AND DEVELOPMENT OF THE MODEL

Finite-Difference Grid

A grid system was developed to represent the aquifer in the simulation model (fig. 6). Each cell in the grid is referenced by the column and row in which it is located. The northwest corner of the study area is the origin. The grid is oriented along the western boundary of the aquifer and has 638 cells (21 columns and 29 rows). Only 318 of the 609 cells in the grid are within the study area and thus are active in the simulation. The smallest cell spacing is about 0.28 mi, and the largest spacing is about 1.5 mi. The smallest cell area is about 0.11 mi^2, and the largest cell area is about 1.42 mi^2. Larger cell spacings are used in the eastern part of the study area because, in that area, the hydraulic gradient is less and the saturated thickness is more consistent.

A no-flow boundary was assumed to occur at the outer boundary of the entire grid system which represents the study area. This assumption keeps the study area hydrologically independent by not allowing water to cross the boundaries of the digitized study area. All recharge to and discharge from the aquifer was assumed to occur internal to the boundaries and is accounted for as explained in the next section.

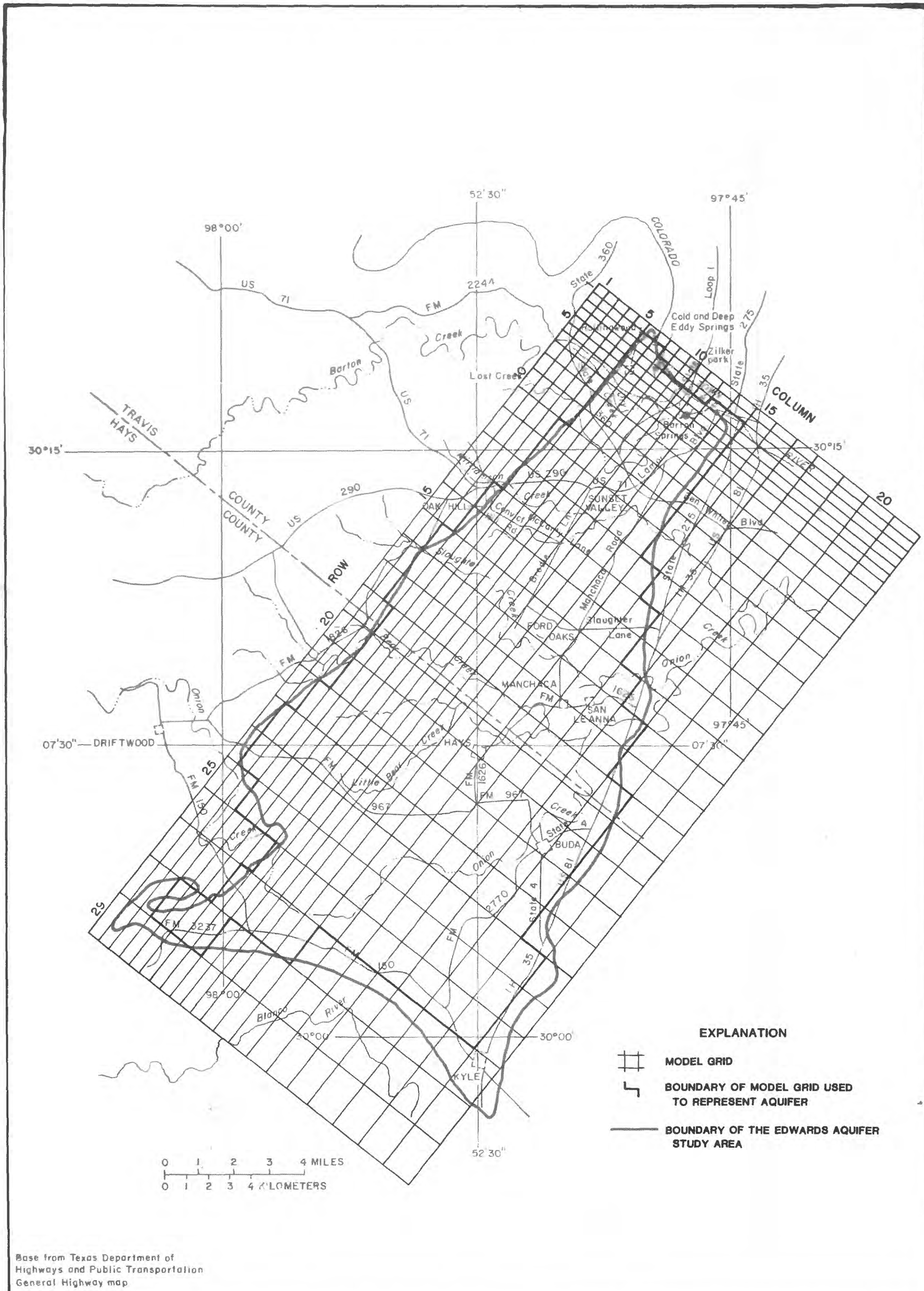

Figure 6.--Finite-difference grid for the digital-computer model.

Steady-State Simulation

A steady-state simulation of "average" hydrologic conditions of the aquifer was developed first. The purpose of this simulation was to estimate the values of transmissivity, which are treated as the "unknown" property. For this simulation, the long-term mean rates of recharge and discharge were assumed to occur simultaneously with the long-term mean potentiometric surface. Values for hydraulic head, recharge, discharge, and transmissivity were estimated for each of the finite-difference cells. The hydraulic and hydrologic data used for this simulation are discussed here.

Aquifer Conditions

Hydraulic head

The mean monthly water level was calculated for each of 19 observation wells measured during Decembeo 1979 to July 1982. During this period, the mean discharge of Barton Springs was approximately equal to the long-term mean value of 50 ft^3/s. For each of the observation wells, the mean water level was similar to the water level measured in January 1981 (fig. 3) when Barton Springs was discharging about 50 ft^3/s. Water levels throughout the aquifer are interrelated and correlate with discharges for Barton Springs (R. M. Slade, Jr., and others, written commun., 1985); thus, the January 1981 water levels are indicative of long-term mean water levels. Seventy-five wells were measured in January 1981, and those water levels were used to construct a potentiometric-surface map for the steady-state simulation. The finite-difference grid was superimposed over the potentiometric-surface map, and a "mean water level" was determined for each cell. Included in the simulation are 318 cells--the 75 cells overlying the measured water levels are considered to have measured hydraulic heads, whereas the other 243 cells have estimated hydraulic heads.

Recharge

The total recharge for this simulation was 50 ft^3/s. Of the total recharge, about 15 percent, or 8 ft^3/s, represents that recharge occurring on tributaries to the 6 main creeks or by diffuse infiltration, thus, that recharge was uniformly distributed in the cells of the recharge zone. The remaining 42 ft^3/s was distributed to the cells containing the main channels of the 6 watersheds that provide recharge by using the tabulated data in the "Recharge to the Aquifer" section. For example, 28 percent of the 42 ft^3/s, or about 12 ft^3/s of recharge, was distributed to the cells containing the main channel of Barton Creek. Recharge within the reaches of each creek was distributed by information derived from the flow-loss studies.

Discharge

Discharge is from Barton Springs and pumpage from the major well fields completed in the aquifer (fig. 5 and table 3). The lowest altitude in the potentiometric surface of the Edwards aquifer occurs at Barton Springs. Water levels at well YD-58-42-903, about 200 ft south of Barton Springs, fluctuate only about 6 ft between extreme high- and low-flow conditions. Because the water-level variations near Barton Springs are so small, the cell containing the springs was assigned a constant hydraulic head of 433 ft. The cell containing Barton Springs was treated as a hydraulic-head controlled flux, thus allowing discharge to occur at the cell to satisfy hydraulic continuity. Each of the 21 major well fields in the aquifer (fig. 5, table 3) was located by cells, and an estimated pumpage rate was assigned as outflow from each cell.

Pumpage rates were estimated by converting the 1981 annual pumpage withdrawals to an assumed uniform distribution throughout the year. These well fields represent about 2,900 acre-ft, or 76 percent, of the 3,800 acre-ft of total annual pumpage. The remaining 900 acre-ft of estimated annual pumpage is from several hundred private wells and was not included in the simulation. This domestic and agricultural usage represents 24 percent of total pumpage, but only 2 percent of total discharge from the aquifer, thus was not considered to significantly affect the results of the simulation. This volume was not included in the simulation because the locations of this use are unknown.

Transmissivity

No known aquifer-test data exist for the study area. Estimates of transmissivities were calculated, based on methods presented by Bentall (1963, p. 338-340), for all wells where specific-capacity determinations were available (fig. 7). Because of missing data for some of the specific-capacity determinations, some of the transmissivities are expressed with a maximum or minimum limiting value. Based on the 60 values available, the transmissivities vary from about 3 to about 47,000 ft^2/d. As expected, most of the wells with the largest specific-capacity values also penetrated cavities within the saturated zone. These values were used to construct a map of transmissivities, which was used to estimate a transmissivity value for each cell.

A model simulation was run using these transmissivities, and many simulations also were run using modifications of these transmissivities. Large disagreements between the simulated and measured hydraulic heads indicated that the transmissivities developed by this method were inadequate for the model. Because of the heterogeneous nature of the aquifer, the hydraulic characteristics commonly vary tremendously in short lateral distances (fig. 7). It was obvious that another method for estimating this property would have to be developed.

Because estimates of transmissivity from specific capacities were inadequate to provide a starting value for transmissivity in the model, transmissivity values for the cells were estimated based on the following procedure. Ferris and others (1962, p. 73) showed the relationship between transmissivity and flow through a form of Darcy's law in the expression:

$$Q = TIL$$

where Q = flow, in gallons per day;
T = transmissivity, in gallons per day per foot;
I = hydraulic gradient, in feet per foot; and
L = width, in feet, of the cross section through which the flow occurs.

Solving this equation for T yields:

$$T = \frac{Q}{IL} .$$

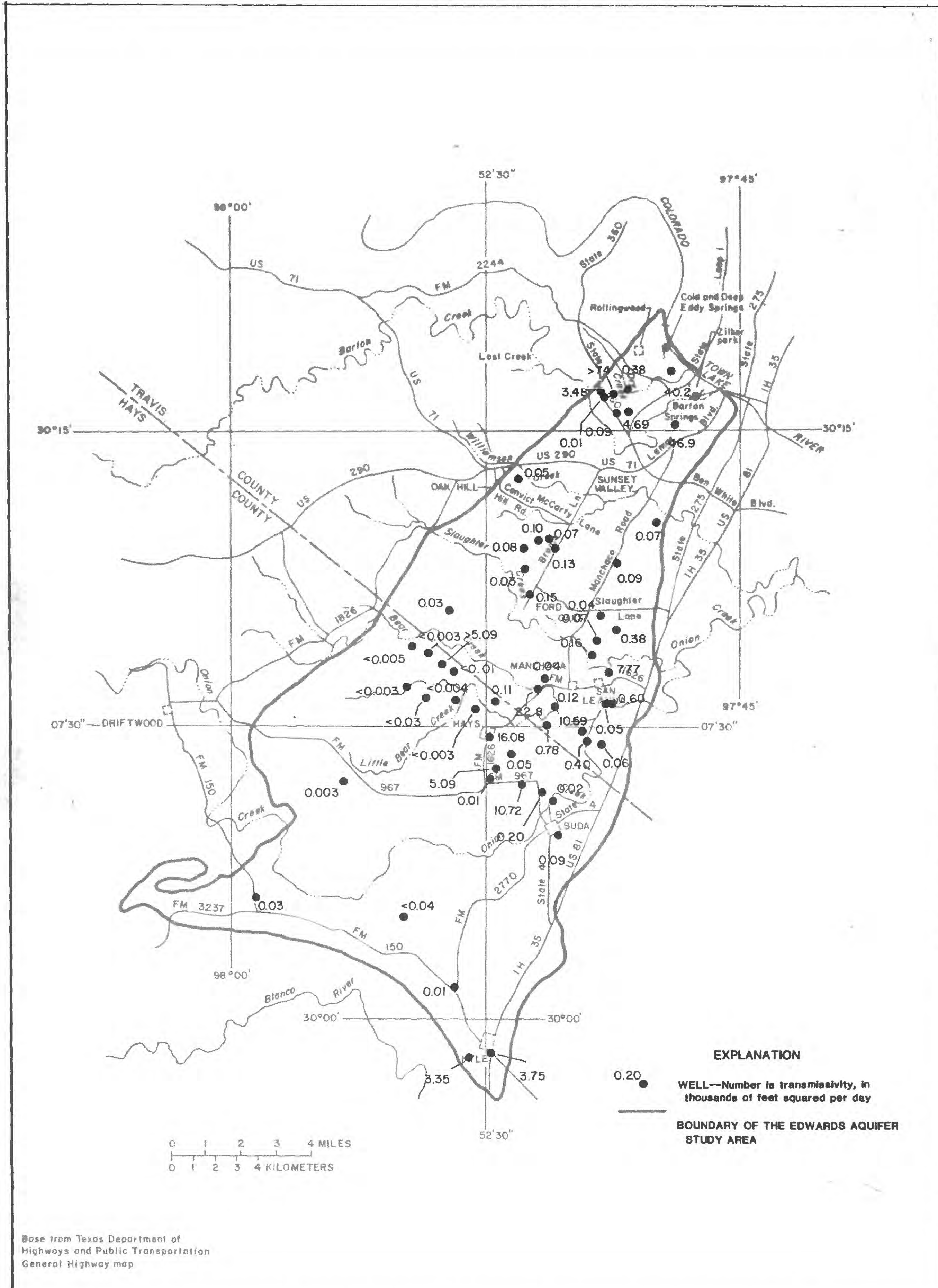

Figure 7.--Transmissivity values calculated from specific-capacity determinations of wells.

A flow-net analysis (Lohman, 1972, p. 45) was done for the recharge. Flow lines for the 42 ft^3/s of recharge occurring on the main channels of the creeks were determined from the potentiometric surface, and the approximate total flow passing through each cell was computed. Based on the January 1981 potentiometric surface, a hydraulic gradient was estimated for each cell, and the width of each cell was used as "L" in the above equation. The estimated transmissivities were calculated by this method, entered into the model, and adjusted as explained in the next section.

Calibration of the Model

Calibration refers to the process of adjusting selected data in the model until the differences between the simulated results and measured values are within acceptable limits. The transmissivities calculated from the flow net analysis were used for the initial simulation. Hydraulic heads were simulated by the model for each cell and compared to the measured hydraulic heads. The transmissivities were changed, and the model simulations were repeated until the simulated hydraulic heads reasonably matched the measured hydraulic heads. A reasonable match between these hydraulic heads could be achieved using isotropic values of transmissivity.

Transmissivities of the Edwards aquifer in the study area are not necessarily isotropic. However, the aquifer was simulated on a regional basis in this study and the model cells are large enough so that the hydrologic effects of local anisotropy may not be detectable. The transmissivities that produced the best simulation are given in table 4. A pattern of the distribution of transmissivity values indicates that the cells directly under the main channels of the creek beds have larger values than the areas between the creeks. This may be because cavities in the aquifer have formed under the creek beds where the larger volumes of recharge occur or because the creeks themselves occupy zones of structural weakness, which also cause increased permeability in the aquifer.

Many simulations were made in which anisotropic values of transmissivity were entered into the model (see "Geology" section). The transmissivity values were changed to 2, 3, 5, and 10 times greater in the northeast direction (along aquifer strike and column direction of the model grid) than in the southeast direction (along the rows). The computer program was modified so that transmissivity was treated as anisotropic only in cells containing many faults and as isotropic in the other cells. None of the simulations involving anisotropic transmissivities produced an improved match between the simulated and measured hydraulic heads.

The data in table 4 also indicates that transmissivities increase with proximity to Barton Springs. This is evident from reviewing the locations and volumes of recharge, along with a potentiometric-surface map of the aquifer (fig. 3). As water moves downgradient to Barton Springs, it converges to the outflow point, thus the flow moving through a unit width of aquifer increases with proximity to Barton Springs. Furthermore, the hydraulic gradient decreases with proximity to the springs. The second equation in the "Transmissivity" section of this report shows the relationship of transmissivity to flow and hydraulic gradient. Transmissivities must increase with increased flow or with decreased hydraulic gradient when the other factors in the equation are

Table 4.--Transmissivity values for the steady-state simulation

[Values in thousands of feet squared per day]

Row number	Column number 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21
1																					
2						0.1	0.1														
3						.3	.8	1.6					0.8	0.8							
4						.2	.2	.8	0.8	1.6	1.6	1150	1.6	.8							
5						.1	.2	.3	.8	1.6	2.5	1150	82	16							
6						.1	.1	.2	.5	1.6	2.5	1110	120	25							
7						.1	.1	.1	.2	1.6	2.5	940	160	33							
8						.2	.2	.2	.7	1.6	8.2	820	200	41							
9						.3	.6	1.6	7.4	7.4	41	660	250	82							
10						1.6	7.4	25	25	35	49	490	290	120							
11					0.1	.1	.5	4.1	16	21	19	290	330	160							
12					.1	.1	.2	.8	1.6	3.3	8.2	120	330	200							
13					.8	2.0	2.0	2.0	2.0	4.1	8.2	82	250	200	41						
14				0.1	.1	.1	.1	.1	.1	.1	.7	7.4	82	200	82	4.1					
15				.1	.2	.2	.3	.3	.4	.7	1.2	5.8	13	160	120	16					
16			0.3	.5	.7	.8	1.1	1.1	2.1	2.1	2.1	4.9	12	120	120	25	4.1				
17			.4	.4	.4	.5	.5	.6	.7	1.1	2.1	4.1	9.8	66	110	41	25	8.2			
18			.2	.2	.2	.2	.2	.5	.8	1.0	2.0	2.0	3.3	25	90	49	41	16			
19				.2	.3	1.2	2.0	2.0	3.0	4.9	4.9	5.9	5.9	20	74	66	45	25	8.2		
20				.8	.8	.8	.8	.8	.8	1.0	1.6	3.3	3.3	16	66	62	49	29	12		
21				.1	.1	.1	.1	.2	.2	.2	.7	1.2	1.3	3.3	16	62	49	33	12		
22			.1	.1	.1	.1	.1	.1	.2	.2	1.0	1.3	1.6	4.9	8.2	58	53	25	8.2		
23		0.1	.1	.1	.2	.8	1.0	1.2	1.3	1.5	1.6	1.8	2.0	2.5	8.2	41	58	16	5.8		
24		.1	.1	.1	.2	.3	.3	.2	.2	.3	.3	.7	.8	4.9	16	33	53	12	4.1	0.8	
25				.1	.2	.4	.8	.7	.8	1.0	1.2	3.0	7.4	18	20	25	41	8.2	2.5	.8	
26								2.5	2.5	4.9	8.2	9.0	15	16	20	16	1.6	.8	.8	.8	
27							.2	1.3	1.6	4.1	6.6	9.8	9.8								
28				.1	.2	.2	.7	1.2													

constant. Therefore, the increase of transmissivity with proximity to the springs is necessary.

The hydraulic heads simulated by the model were contoured and are presented in figure 8, along with the measured hydraulic heads. The differences between the simulated and measured hydraulic heads can be inferred from figure 8, which is a measure of how accurately the model is calibrated. Throughout all but a small part of the aquifer, the simulated hydraulic heads at the measured hydraulic-head locations are within 20 ft of the measured hydraulic head values. In much of the aquifer, the simulated hydraulic heads are within 10 ft of measured hydraulic-head values.

Sensitivity Analysis

There is some degree of uncertainty associated with the data used in a ground-water flow model. The purpose of sensitivity analysis is to determine to what degree model calibration is affected by variations in the data used in a model. For the steady-state simulation, hydraulic head, recharge, springflow, and pumpage are considered "known" data and are the basis on which the calibration is evaluated. The "unknown" data are transmissivity values, thus sensitivity analysis is tested for that property. Even though the domestic and agricultural pumpage (900 acre-ft) was not included in the simulation, the model results are not sensitive to subsequent simulations that used an even distribution of that pumpage.

A simple method of model-sensitivity analysis is to uniformly vary the tested property, run the simulation model, and determine the change in the simulated hydraulic heads. Two simulations were done with the transmissivities changed; one with the values uniformly increased 10 percent, and one with the values uniformly decreased 10 percent. The simulated hydraulic heads for each of the two "sensitivity" simulations were compared to the hydraulic heads simulated by the calibrated model, and the differences between those hydraulic heads were determined. The results of these comparisons were expressed as a change in the simulated potentiometric surface between the calibrated and the "sensitivity" simulations; those results are shown in figure 9 for the increased transmissivities and in figure 10 for the decreased transmissivities.

Increasing the transmissivities tends to lower the simulated water levels throughout the aquifer (fig. 9). The decline in water levels varies from about 45 ft in the western part of the aquifer in the Little Bear Creek watershed to no change near Barton Springs. As expected, decreasing the transmissivities tends to raise simulated water levels throughout the aquifer (fig. 10). This rise varies from about 50 ft in the Little Bear Creek watershed to no change near Barton Springs. This analysis shows that the simulated hydraulic heads are more sensitive to changes in transmissivity in the western and southern parts of the aquifer than in the area near Barton Springs. The simulated hydraulic-head changes due to the sensitivity simulation are greatly minimized at the constant-head node representing Barton Springs. This effect decreases with increased distances from this node.

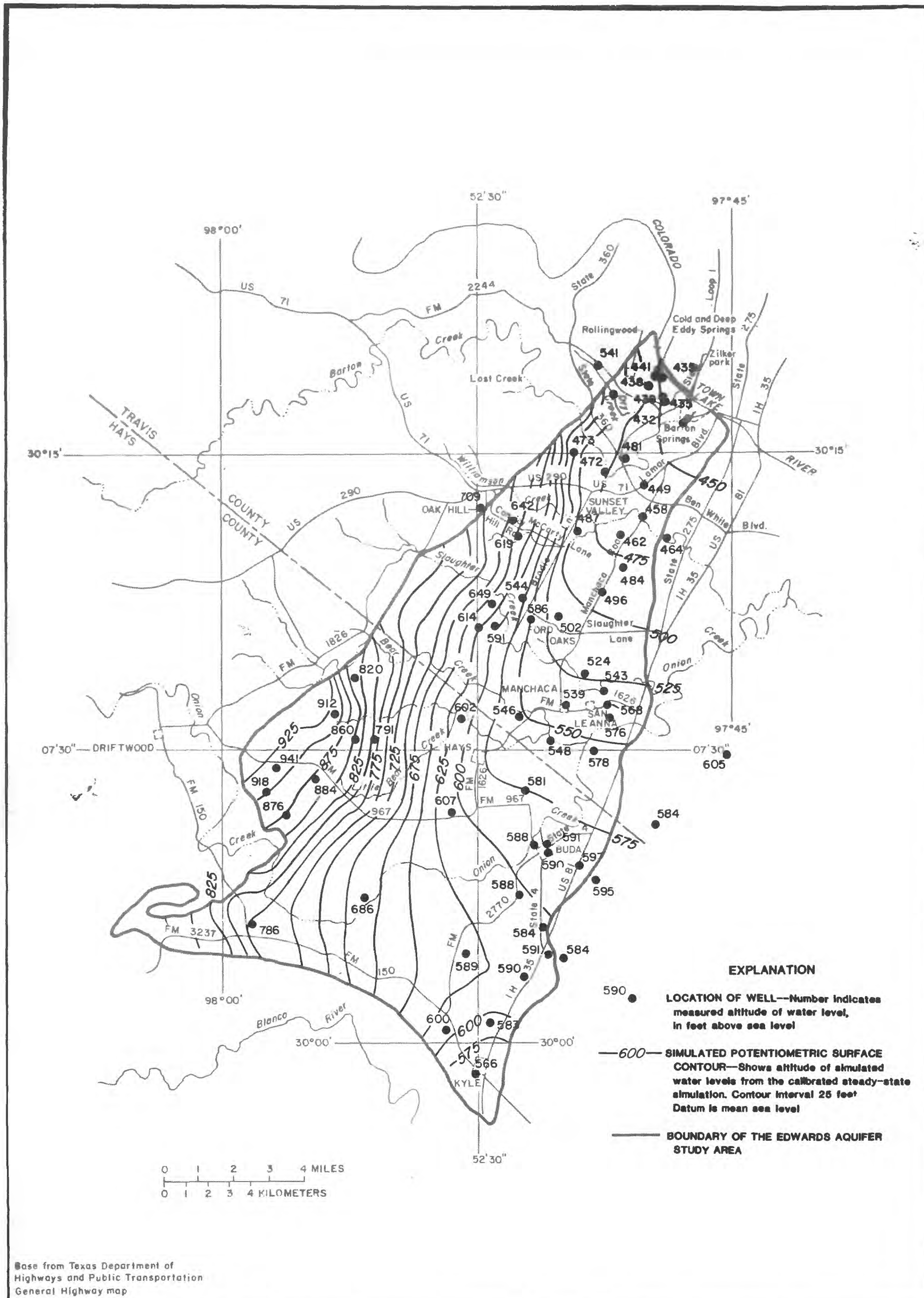

Figure 8.--Simulated steady-state potentiometric surface, January 1981.

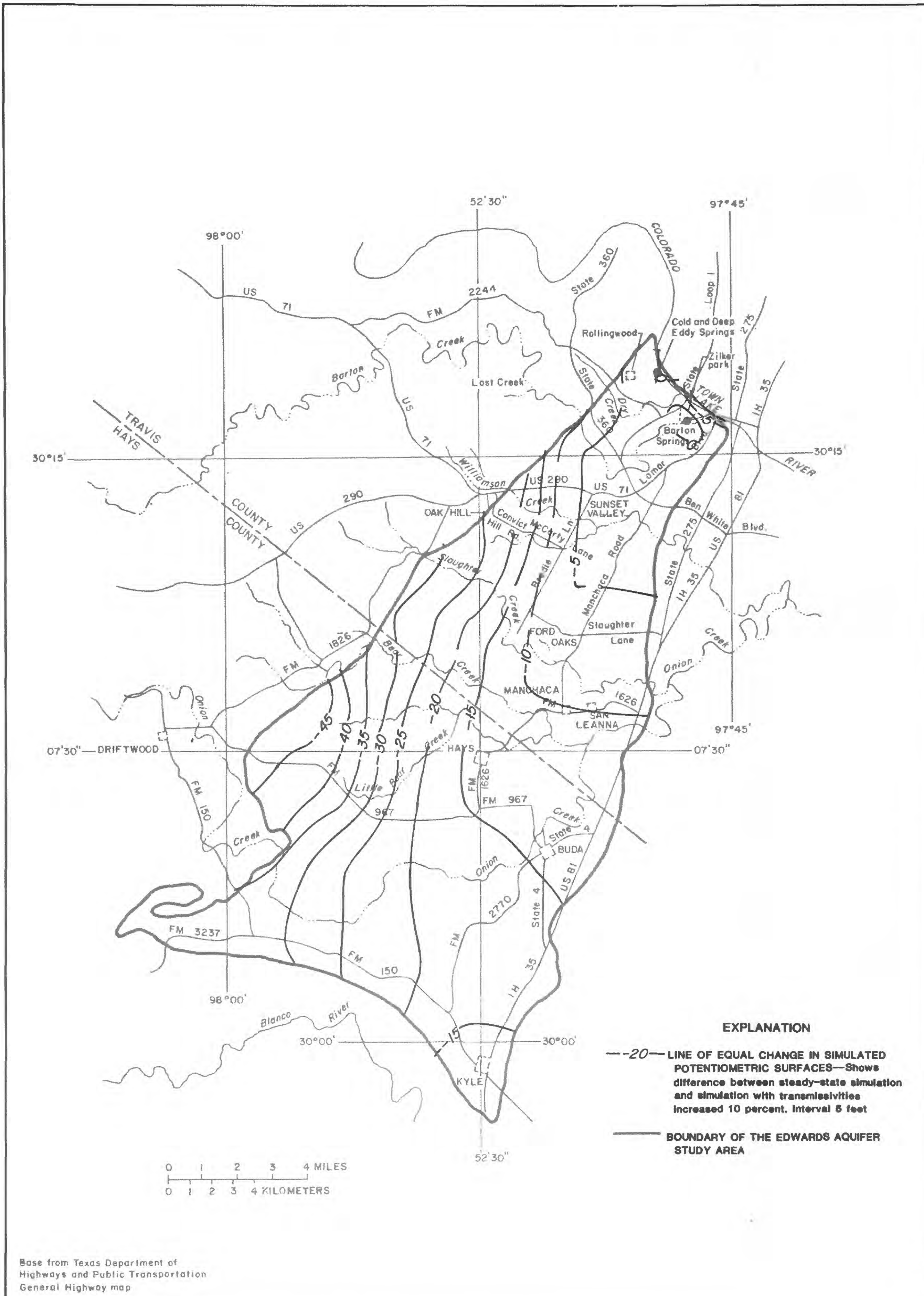

Figure 9.--Changes in simulated potentiometric surface caused by increasing modeled transmissivities by 10 percent.

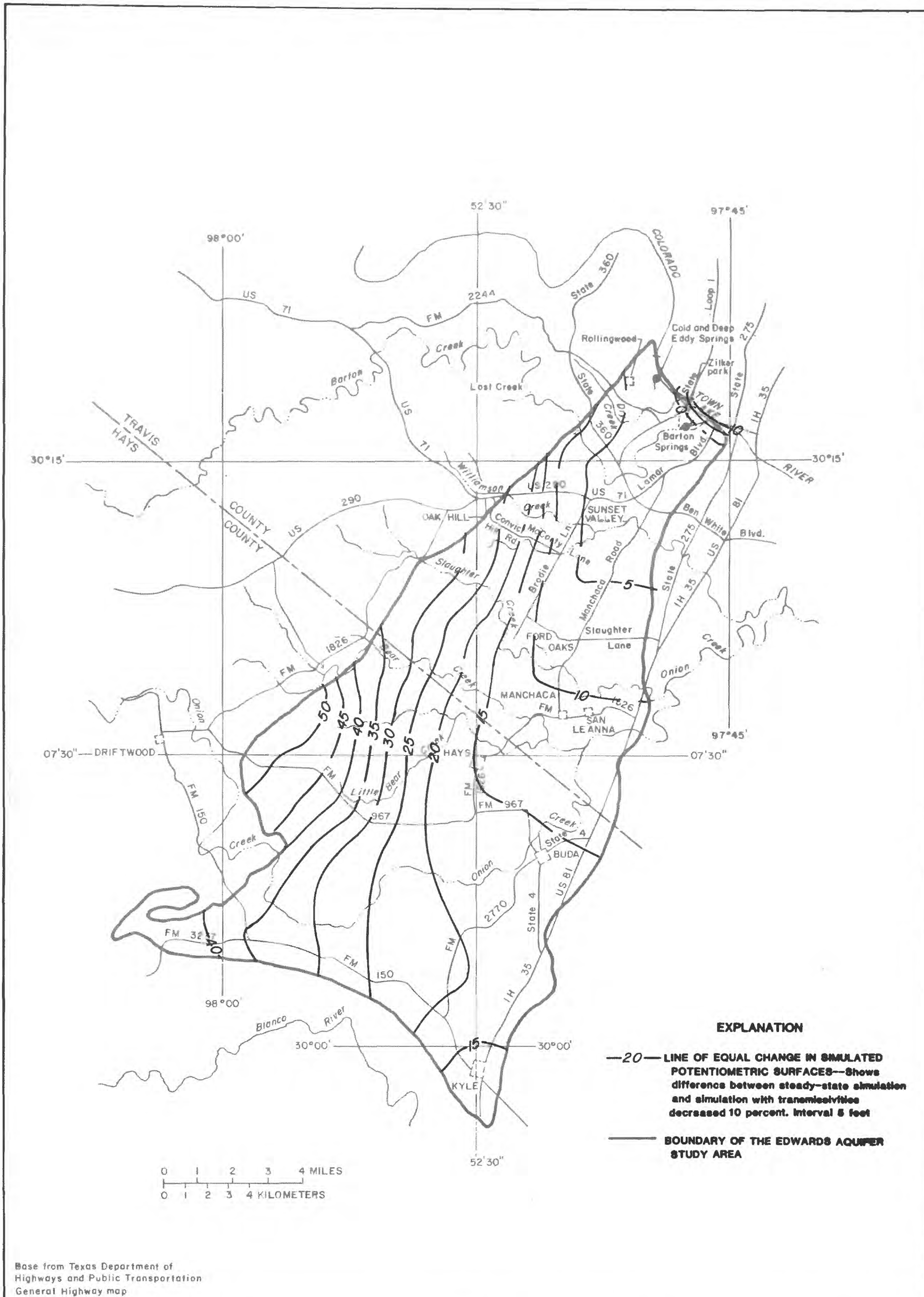

Figure 10.--Changes in simulated potentiometric surface caused by decreasing modeled transmissivities by 10 percent.

Transient-State Simulation

In transient-state simulations, simulated hydraulic head is a function of changes in recharge, discharge, and storage; therefore, storage coefficients are incorporated into the model. A transient-state simulation was made for 164 consecutive days using daily recharge and discharge data. The purpose of this simulation is to calibrate the values of specific yield and verify the hydraulic conductivities of the steady-state simulation. During the 164 days simulated by the model, the discharge averaged 67.7 ft^3/s, of which 64 ft^3/s was measured from Barton Springs and 3.7 ft^3/s was pumpage from the major well fields. The gaged mean recharge was only 14 ft^3/s during that period, thus, 54 ft^3/s was the mean contribution to discharge from storage. Because most of the discharge during this period came from storage, specific yields were sensitive to this simulation.

Values for hydraulic head, recharge, discharge, hydraulic conductivity, storage coefficient, and specific yield were entered into the model, along with the altitudes of the base and top of the aquifer. Specific yield was treated as the "unknown" property and those values were adjusted to calibrate this simulation. Discussions of the data used for this simulation follow.

Aquifer Conditions

Hydraulic head

The period used in the transient-state simulation began August 8, 1979, and ended January 18, 1980. The observation wells were measured at the beginning and end of this period and monthly during the period. The water levels measured on August 8, 1979, from these 19 wells provided the starting hydraulic heads for the simulation. Monthly water-level measurements were compared with simulated water levels from the model.

Recharge

The transient-state simulation was divided into six time steps. The average recharge for each time step was entered into the model. The actual recharge occurring during each time step and the approximated recharge as entered into the model are shown in figure 11. Recharge was distributed to the six creeks based on gaged streamflow-loss data for each creek, and within reaches of the creeks based on information derived from the flow-loss studies.

Discharge

The discharge for the transient-state simulation was computed by the model in the same way as in the steady-state simulation. A constant hydraulic head was assigned to the cell containing Barton Springs, and discharge at that cell was computed periodically during the simulation. Gaged daily discharges (fig. 11) for the springs were compared to the simulated discharges and used to adjust the specific yield in the model. Pumpage was distributed as described for the steady-state simulation.

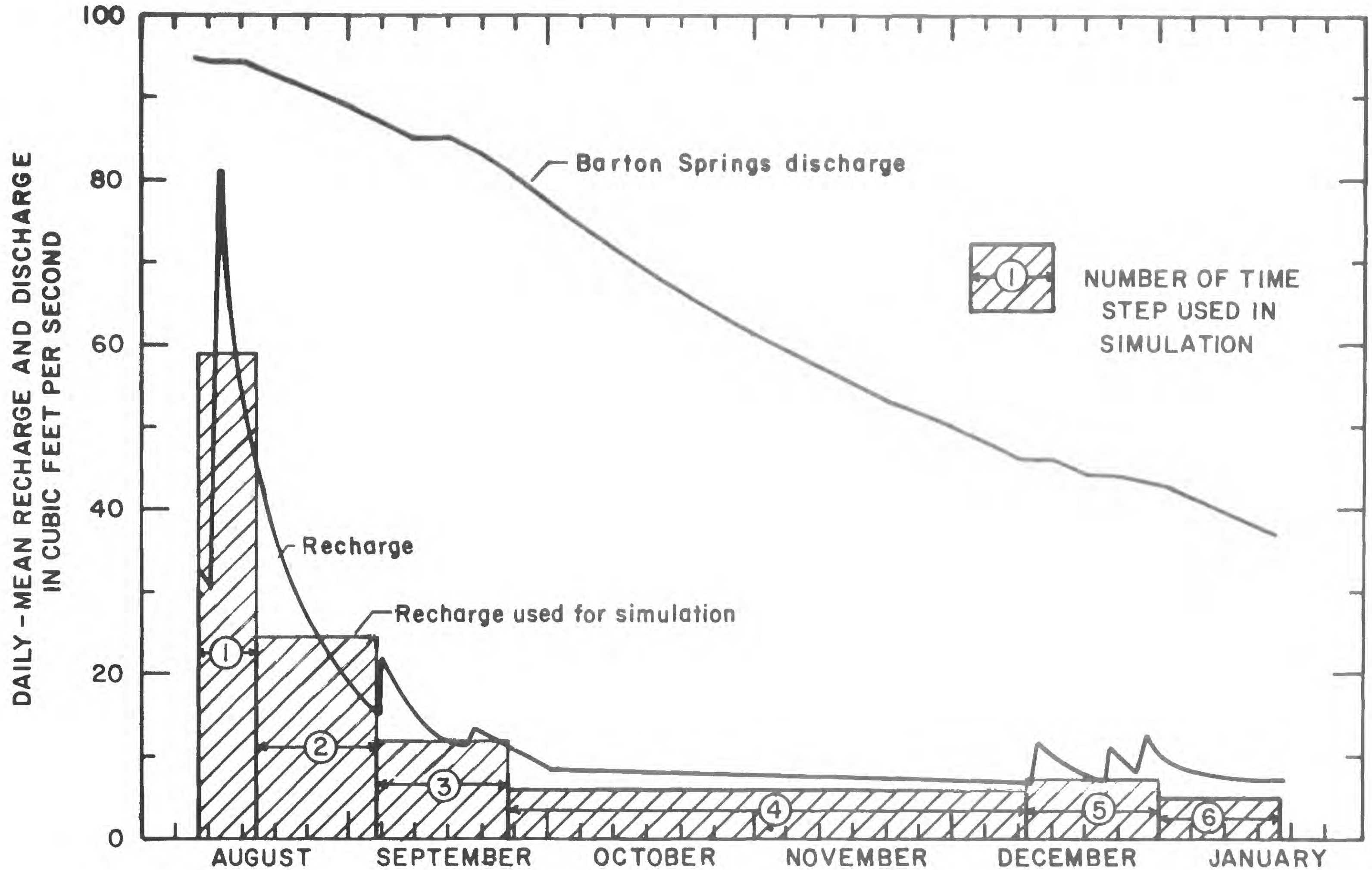

Figure 11.--Hydrographs of Barton Springs discharge and recharge for the transient-state simulation, August 8, 1979, to January 18, 1980

Aquifer Properties

Hydraulic conductivity

The hydraulic heads varied as much as 60 ft in the unconfined zone of the aquifer during this period, so hydraulic conductivities were used in place of transmissivities. Hydraulic conductivities for each cell were computed based on the transmissivities and potentiometric surface from the steady-state simulation by the following: (1) The saturated thickness for the water-table part of the aquifer was computed by subtracting the altitude of the base of the aquifer from the potentiometric surface, or for the artesian part of the aquifer, subtracting the altitude of the base of the aquifer from the altitude of the top of the aquifer, and (2) the transmissivity was divided by the saturated thickness. The computed hydraulic conductivities are presented in table 5. The hydraulic conductivities are subject to errors because of lack of data for determining the altitude of the base of the aquifer. Large vertical displacements along many faults cause large variations in the altitude of the base of the aquifer within short distances.

Storage coefficient

The storage coefficient for the confined part of the aquifer varies with the porosity and thickness of the aquifer and probably ranges from about 3×10^{-5} to 6×10^{-5} (R. M. Slade, Jr., and others, written commun., 1985). A value of 5×10^{-5} was assigned to all the cells in the confined zone, which is about 21 percent of the aquifer.

Altitude of aquifer base and top

Contour maps of the altitude of the base and top of the Edwards aquifer were prepared from drillers' logs and geophysical logs (Baker and others, in press). The model compares, for each cell, the water level throughout the simulation with the altitude of the top of the aquifer to determine whether confined or unconfined conditions apply.

Specific yield

Specific yields were estimated and used as an "unknown" property in the simulation. They were adjusted as explained in the next section.

Calibration of the Model

Values for the specific yield were adjusted in a series of model simulations until a reasonable calibration was obtained. Water levels and discharges for Barton Springs were computed at specific time steps during the 164-day simulation and were compared to the measured values. The lengths of the time steps are shown in figure 11.

Increasing the specific yields used in the simulation tends to increase the computed springflow and lower the simulated hydraulic heads. The single value of specific yield which provided the best calibration of discharge was

Table 5.--Hydraulic conductivity values for the transient-state simulation

[Values in feet per day]

Row number	Column number 1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21
1																					
2						0.5	0.5														
3						2.3	5.7	10					2.3	2.0							
4						1.8	1.7	4.1	4.7	8.2	7.4	4180	4.8	2.0							
5						.6	13	2.3	4.9	8.6	12	4060	240	41							
6						.6	.6	1.3	3.2	9.4	12	3870	370	62							
7						.6	.6	.7	1.2	10	13	3330	490	82							
8						1.4	2.0	2.2	5.8	12	43	3000	690	100							
9						3.7	5.8	20	75	67	200	2450	850	200							
10						28	210	470	330	290	220	1880	1030	310							
11					1.4	2.7	25	220	330	170	82	1090	1190	420							
12					.4	.8	12	55	12	16	36	500	1190	520							
13					12	33	39	39	18	20	38	340	950	550	100						
14				1.2	1.2	1.5	1.7	1.2	9.1	.5	3.1	33	310	560	210	9.8					
15				1.2	2.4	5.0	6.7	6.0	4.4	2.8	5.9	26	46	480	300	40					
16			3.4	5.2	5.5	9.8	16	18	17	9.8	10	24	41	340	300	59	10				
17			6.5	3.8	3.7	4.9	8.6	1.4	5.0	5.2	11	21	42	170	250	100	61	20			
18			3.4	3.2	1.6	1.8	4.3	16	4.9	5.2	9.9	10	16	64	220	120	100	40			
19				5.5	5.5	12	27	46	20	31	28	31	34	75	200	160	110	59	20		
20				46	59	8	11	14	7.1	7.6	12	22	20	89	210	160	120	67	30		
21				3.3	2.7	.7	.7	2.0	1.5	2.4	6.1	11	10	24	60	180	140	80	29		
22			.6	.6	.7	.5	.4	.8	1.2	2.0	9.8	15	16	41	37	220	150	56	19		
23		4.8	4.8	3.3	5.5	14	5.3	7.1	10	15	19	24	26	26	47	240	160	46	13		
24		3.0	3.0	3.0	5.8	12	2.2	1.9	2.2	3.3	4.3	11	13	55	140	270	170	28	9.0	1.8	
25				1.0	2.0	5.1	10	8.1	10	13	19	49	120	270	230	270	130	20	5.2	1.7	
26								37	44	140	160	140	180	190	240	190	5.6	2.0	1.8	1.7	
27							6.2	33	71	120	150	160	140								
28				8.2	16	25	36	41													
29																					

0.014. This value produced a simulated mean springflow equal to the gaged mean springflow of 64 ft^3/s.

Specific yield probably varies throughout the aquifer. Specific yield increases with increased porosity, and because of the karstic nature of the aquifer, specific yield also increases with hydraulic conductivity. As shown in table 5, hydraulic conductivities generally increase with proximity to Barton Springs, and thus, a similar pattern for distribution of specific yield probably exists. Many ranges and combinations of specific yield were used in the simulation using a distribution pattern similar to that used for the hydraulic-conductivity values. The specific yields that produced the best calibration results are presented in table 6. Selected water levels representing different parts of the aquifer and discharges calculated for this simulation are shown with the measured values in figure 12. The simulated water levels at the end of the simulation for those cells which contain observation wells are presented in figure 13.

Sensitivity Analysis

Sensitivity analysis was performed on the hydraulic conductivity and the specific yield. Values for these properties were changed, and the resultant changes in the hydraulic heads and springflow were determined.

Two simulations were done with the hydraulic-conductivity values changed while the other properties were not: one with the values uniformly increased 20 percent and one with the values uniformly decreased 20 percent. Selected water levels representing different parts of the aquifer and discharges calculated by the "sensitivity" simulations were compared to the simulated values from the calibrated simulations. The results of the sensitivity simulations are shown in figure 14 for the increased hydraulic conductivities and in figure 15 for the decreased hydraulic conductivities. Also shown in these illustrations are the measured values for selected water levels and Barton Springs discharge.

Increasing the hydraulic conductivities tends to lower the simulated water levels throughout the aquifer and increase the simulated discharge of Barton Springs (fig. 14). Simulated water levels changed only a few feet and, thus, were not very sensitive to this change. The mean discharge calculated by the simulation was 13 percent greater than the mean discharge from the calibrated simulation. Decreasing the hydraulic conductivities tends to raise simulated water levels and decrease simulated Barton Springs discharge (fig. 15). Simulated water levels were not very sensitive to this change either. The mean discharge calculated by this simulation was 14 percent less than the mean discharge from the calibrated simulation. Simulated springflow is more sensitive than simulated water levels to changes in the hydraulic conductivity.

Two simulations also were done with the specific yields changed while the other properties were not: one with the values uniformly increased 25 percent and one with the values uniformly decreased 25 percent. The results of these sensitivity simulations are shown in figure 16 for increased specific yields and in figure 17 for the decreased specific yields. The measured values for selected water levels and Barton Springs discharge also are shown.

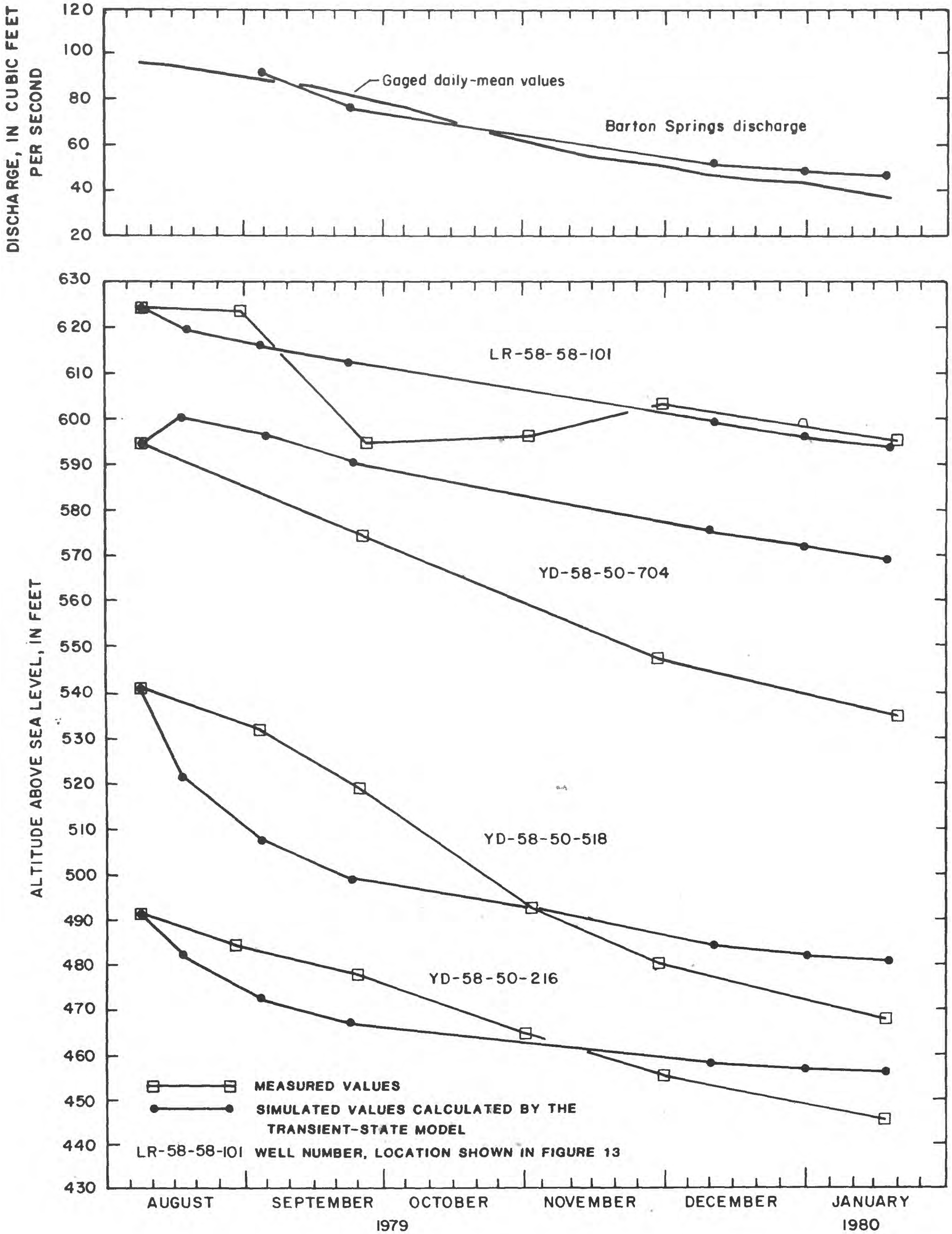

Figure 12.--Hydrographs of measured and simulated Barton Springs discharge and selected ground-water levels for the transient-state simulation.

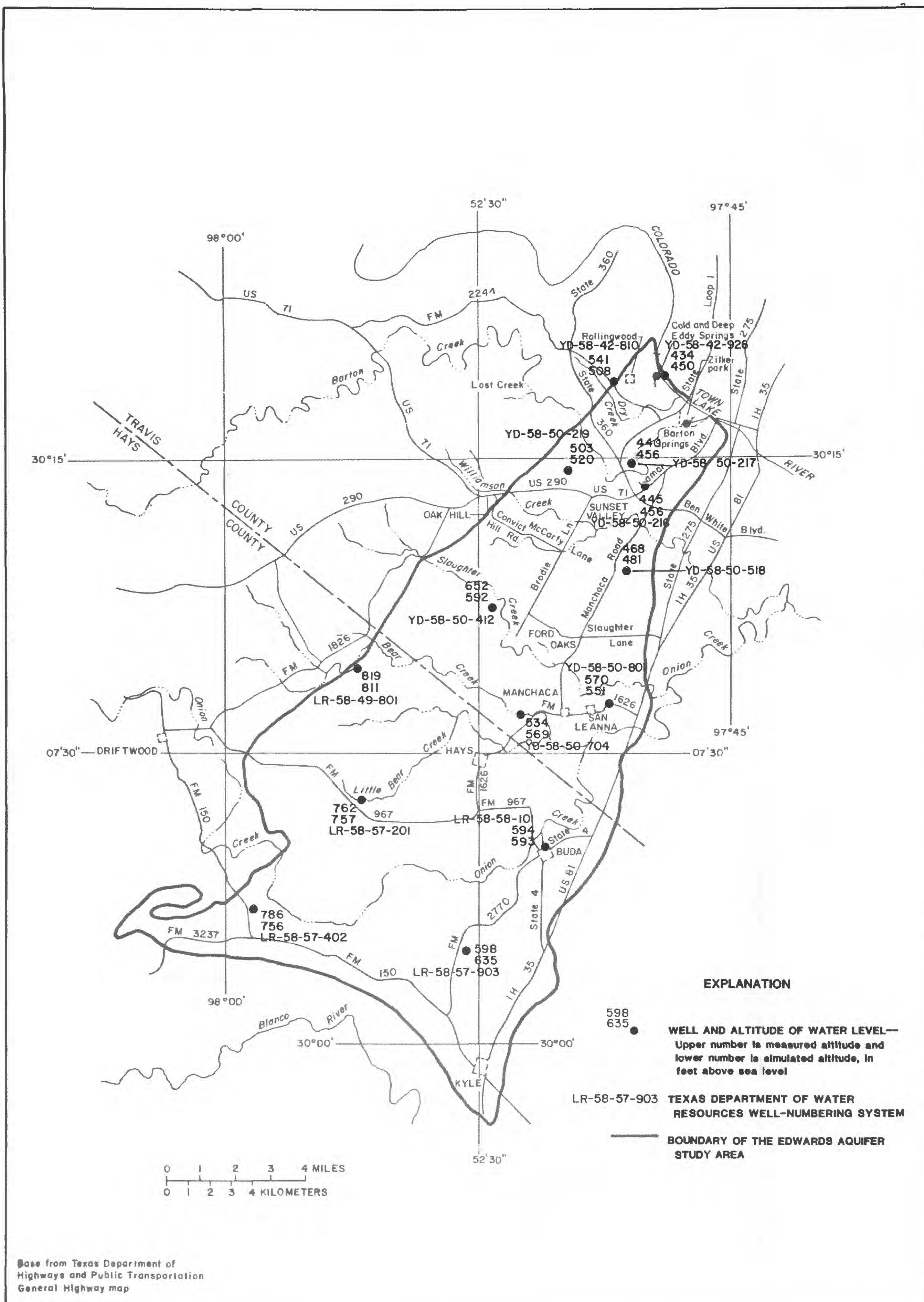

Figure 13.--Measured and simulated water levels for the end of the transient-state simulation, January 18, 1980.

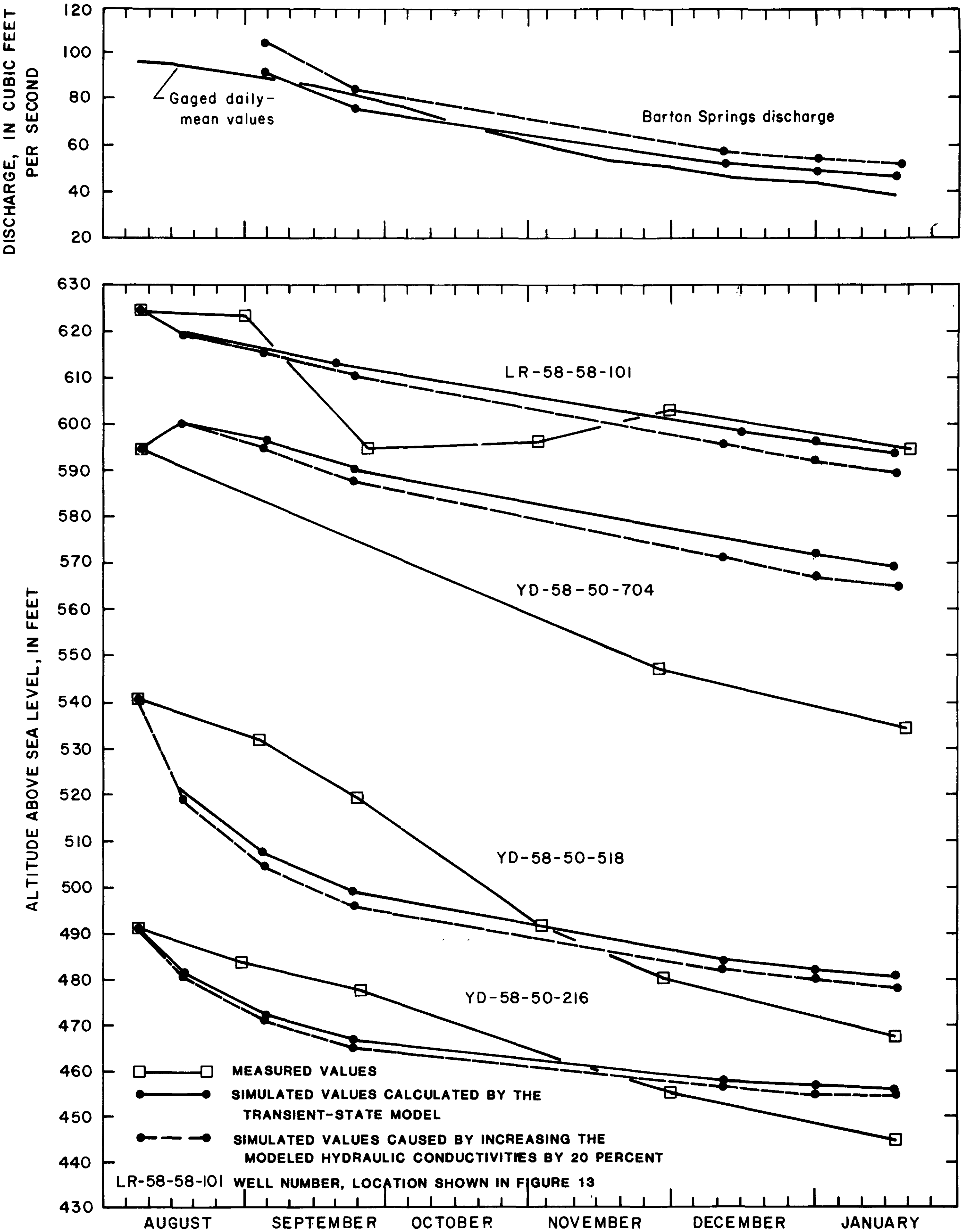

Figure 14.--Hydrographs of simulated Barton Springs discharge and simulated ground-water levels in selected wells caused by increasing calibrated hydraulic conductivities by 20 percent.

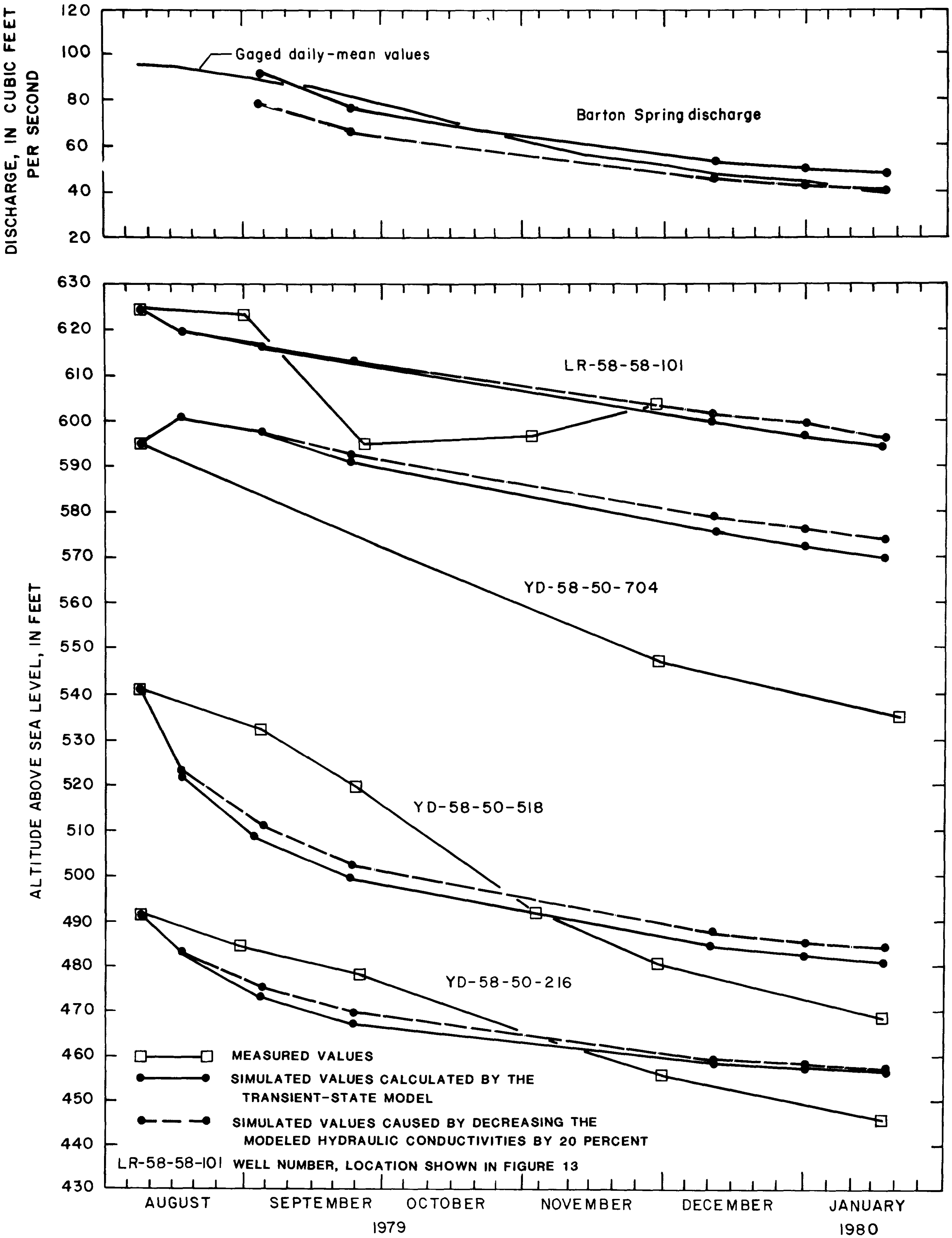

Figure 15.--Hydrographs of simulated Barton Springs discharge and simulated ground-water levels in selected wells caused by decreasing calibrated hydraulic conductivities by 20 percent.

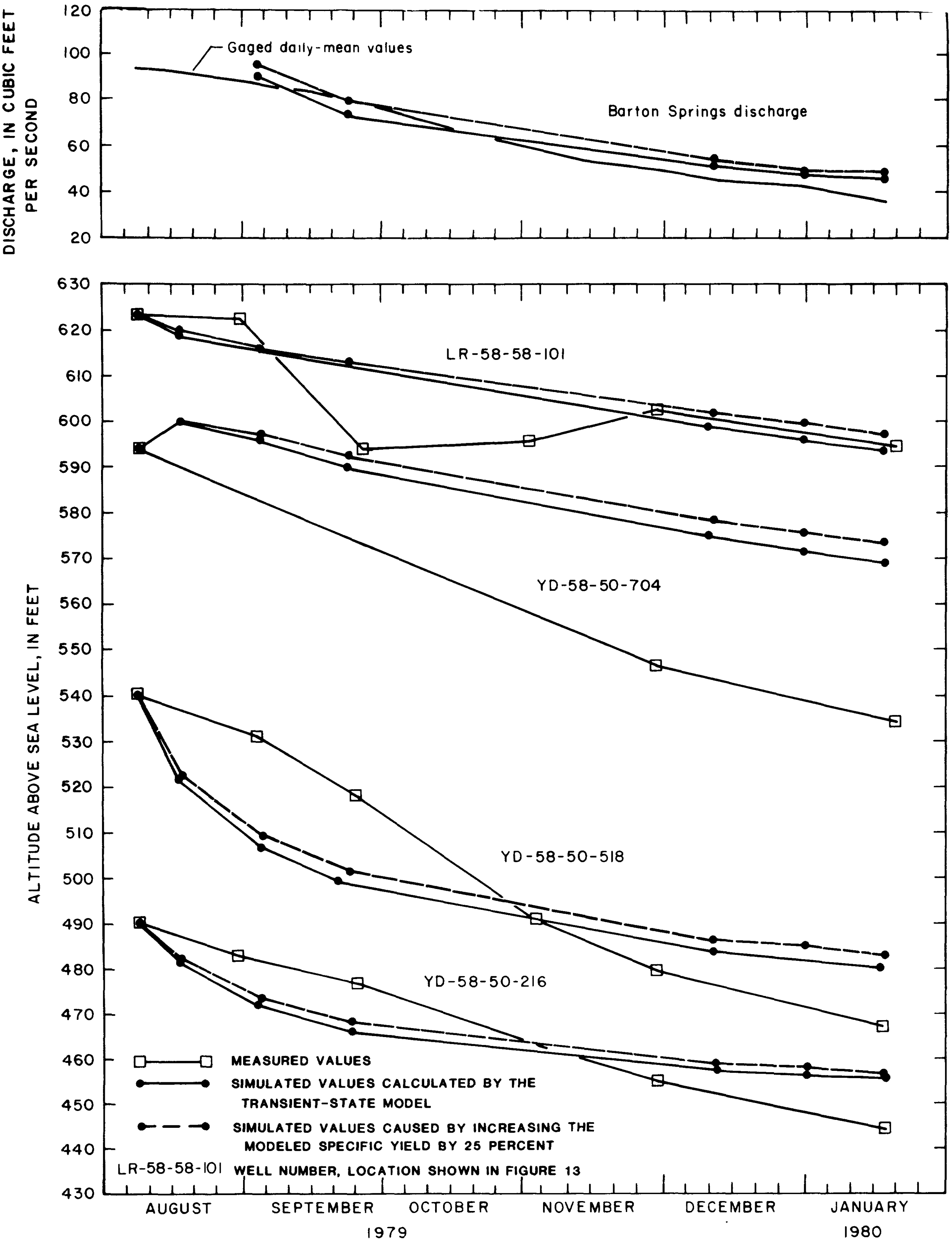

Figure 16.--Hydrographs of simulated Barton Springs discharge and simulated ground-water levels in selected wells caused by increasing calibrated specific yields by 25 percent.

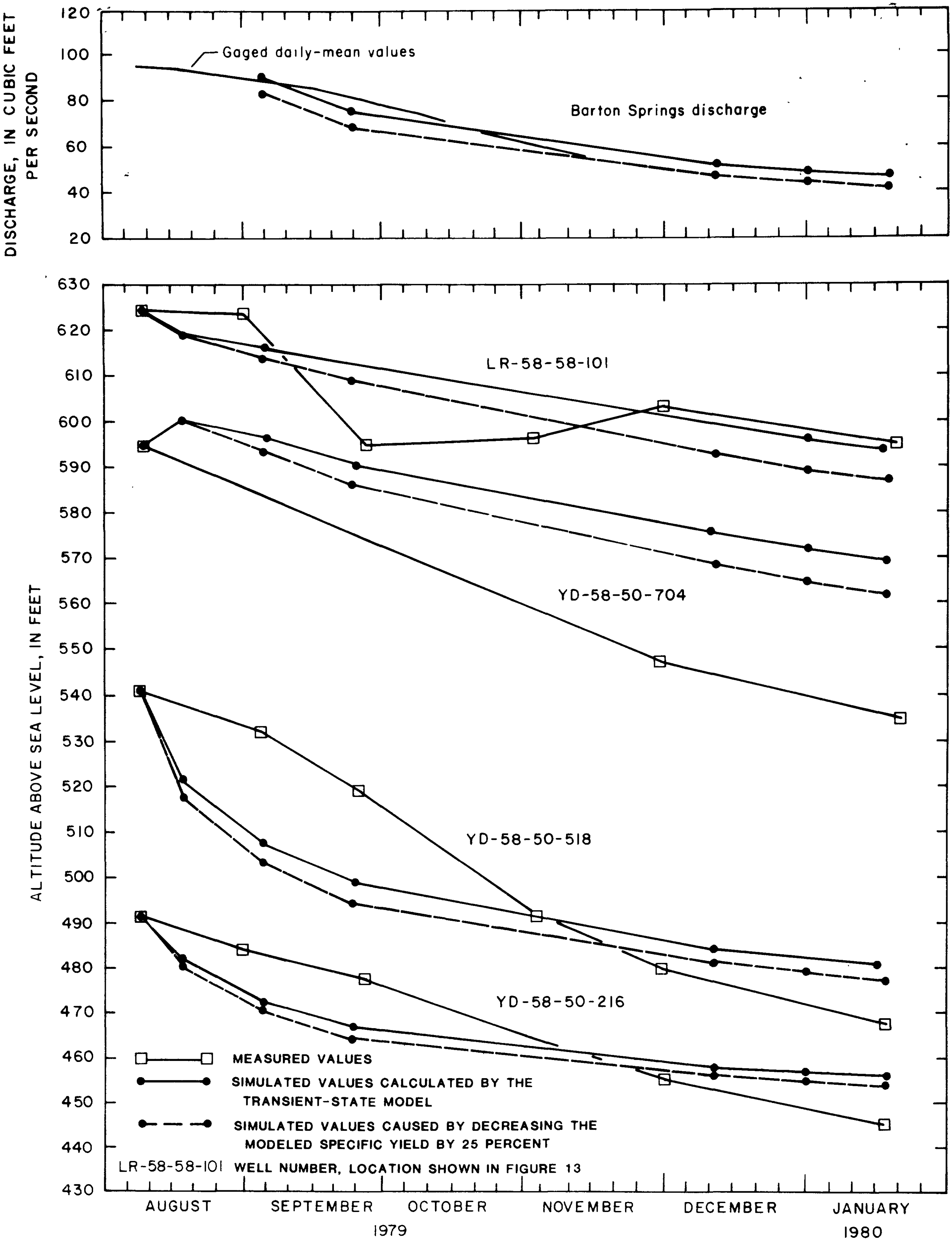

Figure 17.--Hydrographs of simulated Barton Springs discharge and simulated ground-water levels in selected wells caused by decreasing calibrated specific yields by 25 percent.

Table 6.--Specific yield values for the transient-state simulation

Column number

Row number	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21
1																					
2						0.008	0.008														
3						.008	.008	0.008													
4						.008	.008	.008	0.008	0.008	0.012	0.064									
5						.008	.008	.008	.008	.008	.012	.056	0.012								
6						.008	.008	.008	.008	.008	.012	.056	.012								
7						.008	.008	.008	.008	.008	.012	.048	.016								
8						.008	.008	.008	.008	.008	.012	.048	.016								
9						.008	.008	.008	.008	.008	.012	.040	.020	0.012							
10						.008	.008	0.012	0.016	0.016	.016	.032	.028	.012							
11					0.008	.008	.008	.008	.008	.008	.012	.020	.032	.016							
12					.008	.008	.008	.008	.008	.008	.008	.016	.032	.020							
13					.008	.008	.008	.008	.008	.008	.008	.012	.032	.020							
14				0.008	.008	.008	.008	.008	.008	.008	.008	.008	.020	.028	0.012						
15				.008	.008	.008	.008	.008	.008	.008	.008	.008	.016	.028	.020						
16			0.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.012	.020	.020						
17			.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.012	.016	.020	.012	0.012				
18			.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.012	.020	.016	.012				
19				.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.012	.020	.016	.012				
20				.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.012	.016	.016	.016	0.012			
21				.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.012	.016	.016	.012			
22			.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.012	.016	.016	.012			
23		0.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.012	.016	.016	.012			
24		.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.012	.016	.016	.012			
25				.008	.008	.008	.008	.008	.008	.008	.008	.008	.008	.012	.012	.012	.016	.012			
26								.008	.008	.008	.008	.008	.008	.012	.012	.012	.012				
27							.008	.008	.008	.008	.008	.008	.008								
28				.008	.008	.008	.008	.008													
29																					

Increasing the specific yield tends to raise the simulated water levels throughout the aquifer and increase the simulated discharge of Barton Springs (fig. 16). Simulated water levels changed only a few feet. The mean discharge from this simulation was 7 percent greater than the mean discharge from the calibrated simulation. Decreasing the specific yield tends to lower simulated water levels throughout the aquifer and decrease simulated Barton Springs flow (fig. 17). Simulated water levels changed only a few feet. The mean discharge from this simulation was 8 percent less than the mean discharge from the calibrated simulation. Simulated springflow and water levels are relatively insensitive to changes in specific yield.

Simulation of Future Water Levels

A steady-state simulation was done for projected conditions in 2000 using population-increase projections from the city of Austin. The projections indicate that the population will increase by about 86,000 people within the study area by 2000.

Austin city planners also have projected that all of the area north of and within the Bear Creek watershed will be served by water from the Colorado River by 2000, with the possible exception of the two incorporated cities within that area. Those two cities, Sunset Valley and San Leanna, presently use ground water from the Edwards aquifer, and the assumption was made that they would continue to do so. All the area within the aquifer south of the Bear Creek watershed also was assumed to use water from the aquifer for this simulation. The population grid and the modeling grid used for this simulation are presented in figure 18. The population projections were adapted to the model based on the following:

(1) For each population-increase cell, the population increase was assumed to occur at an equal density throughout the cell.

(2) The modeling cells were assigned the same population-increase density as the population-increase cells within which they lie.

(3) The municipal and domestic water use was assumed to be 150 gal/d per person.

(4) Industrial and agricultural pumpage, which is presently 37 percent of total pumpage (59 percent of municipal and domestic pumpage), was assumed to remain at the same proportion for the simulation.

The projected pumpage used in the simulation represents only the estimated demand to supply the water needs for population increases occurring within the aquifer study area. As of 1984, at least 9,000 people living east of the aquifer study area were being supplied by at least four water-supply companies that withdraw water from within the aquifer study area. Probably about 25 to 35 percent of total withdrawals from the study area are used by people living outside of the study area. The projected population living outside the aquifer study area that will be using water from the study area cannot be reasonably estimated, thus that pumpage is not included in the simulation.

The projected pumpage was added to the 1981 pumpage to determine total projected pumpage. The total pumpage for the simulation was 12.3 ft^3/s, or about 25 percent of the long-term mean recharge. This total pumpage was entered into the calibrated steady-state model and the potentiometric surface was com- computed. The change between the January 1981 and projected potentiometric

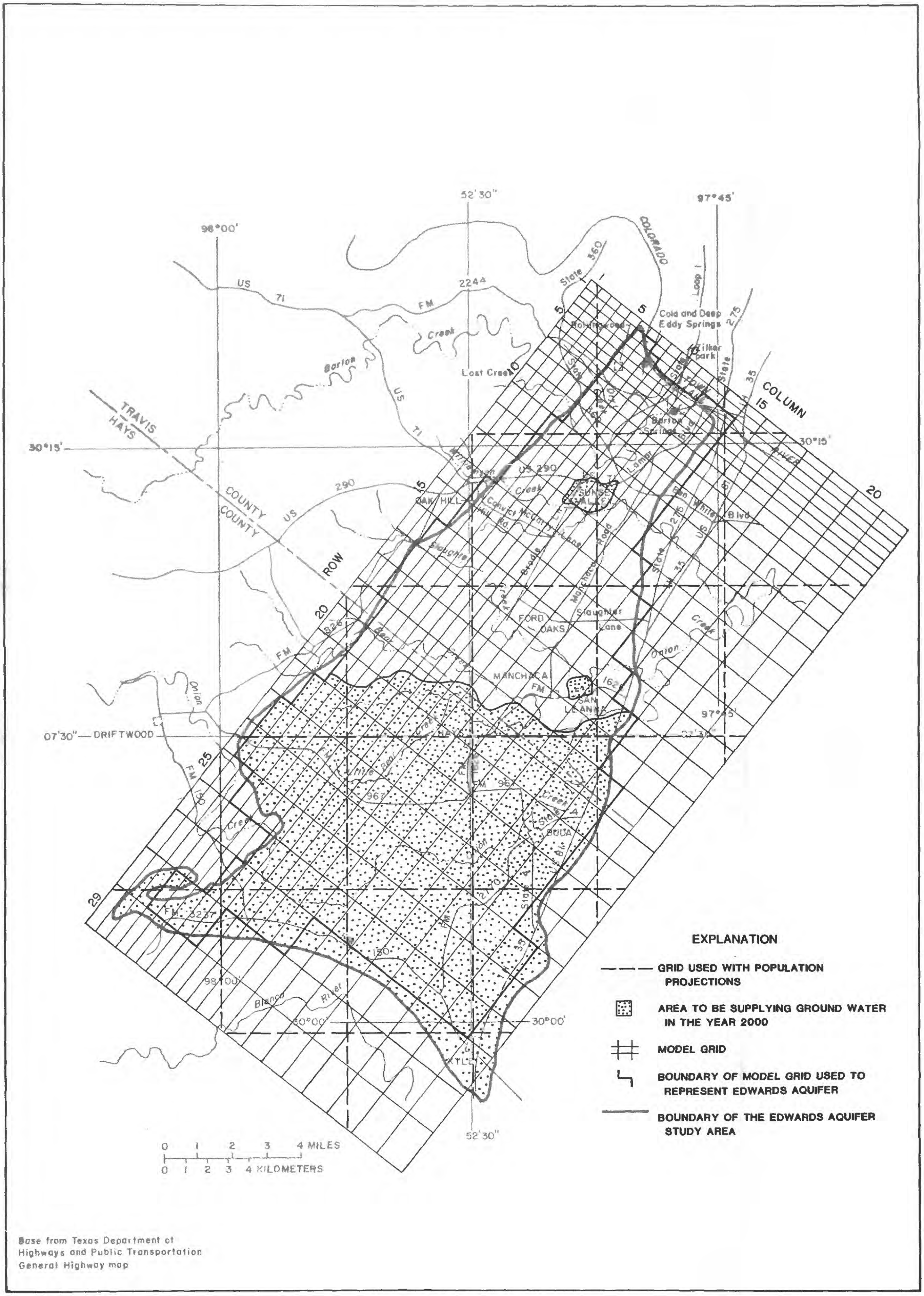

Figure 18.--Population-increase projection grids and areas projected to provide ground water by 2000.

surface in 2000 was calculated and contoured (fig. 19). The saturated thickness of the aquifer for the 1981 potentiometric surface also is shown in figure 19.

The potentiometric surface will be lower in 2000 than in January 1981. Water-level declines will be less than the saturated thickness in the eastern part of the aquifer. In the eastern area, declines will vary from zero near Barton Springs to more than 100 ft near Kyle to the south. However, the aquifer in the western one-half of the area will be dewatered. These water-level declines are considered to be minimum values for the projected population increase because pumpage from the aquifer in the study area that will be supplying increased population outside the study area is not included.

Simulation of Future Water Levels with Potential Recharge Enhancement

The creation of runoff-control structures can be very effective for increasing recharge (R. M. Slade, Jr., and others, written commun., 1985). Local governing officials from cities in the study area are studying a proposal to build such a structure--a large reservoir on Onion Creek near the upstream end of the recharge zone. The reservoir would impound tens of thousands of acre-feet of water and could be used for recharge enhancement or as a direct source of water for the area. The reservoir could significantly increase recharge volumes if managed for this purpose. During July 1979 to December 1982, about 52,000 acre-ft of recharge occurred in the Onion Creek watershed (table 2). However, during this time, almost 88,000 acre-ft of runoff occurred at the downstream end of the recharge zone. This volume represents the rejected recharge for Onion Creek during that period. If only one-half of this runoff were converted to recharge, the total mean recharge could be increased about 30 percent. This increase would exceed the pumpage projected for 2000, which is 25 percent of recharge.

Enhanced recharge from a hypothetical runoff-control structure on Onion Creek was simulated. The assumption was made that one-quarter of the mean runoff occurring at the downstream end of the recharge zone on Onion Creek would be converted to recharge, a conservative estimate. This potential recharge increase, which would average 8.5 ft^3/s, was then added to the natural mean recharge from Onion Creek (17 ft^3/s). The total actual and potential recharge was distributed on the basis of the flow-loss study. The January 1981 potentiometric surface, transmissivities, and projected pumpage used previously were combined with the potential enhanced recharge to compute the potentiometric surface for 2000. The changes between the simulated January 1981 and the 2000 potentiometric surfaces were calculated and contoured (fig. 20). In figure 20, positive values indicate the area where the water levels are projected to rise (mainly in the southwest corner of the aquifer) and negative values indicate the areas where the water levels are projected to decline.

The effect of the potential recharge enhancement on the future potentiometric surface is determined by comparing figures 19 and 20. The potential enhanced recharge will raise the projected water levels as much as about 120 ft along the western boundary of the aquifer along Onion Creek and about 40 ft along Onion Creek near Buda. Between Buda and Barton Springs, the effect of the potential enhancement on water levels decreases.

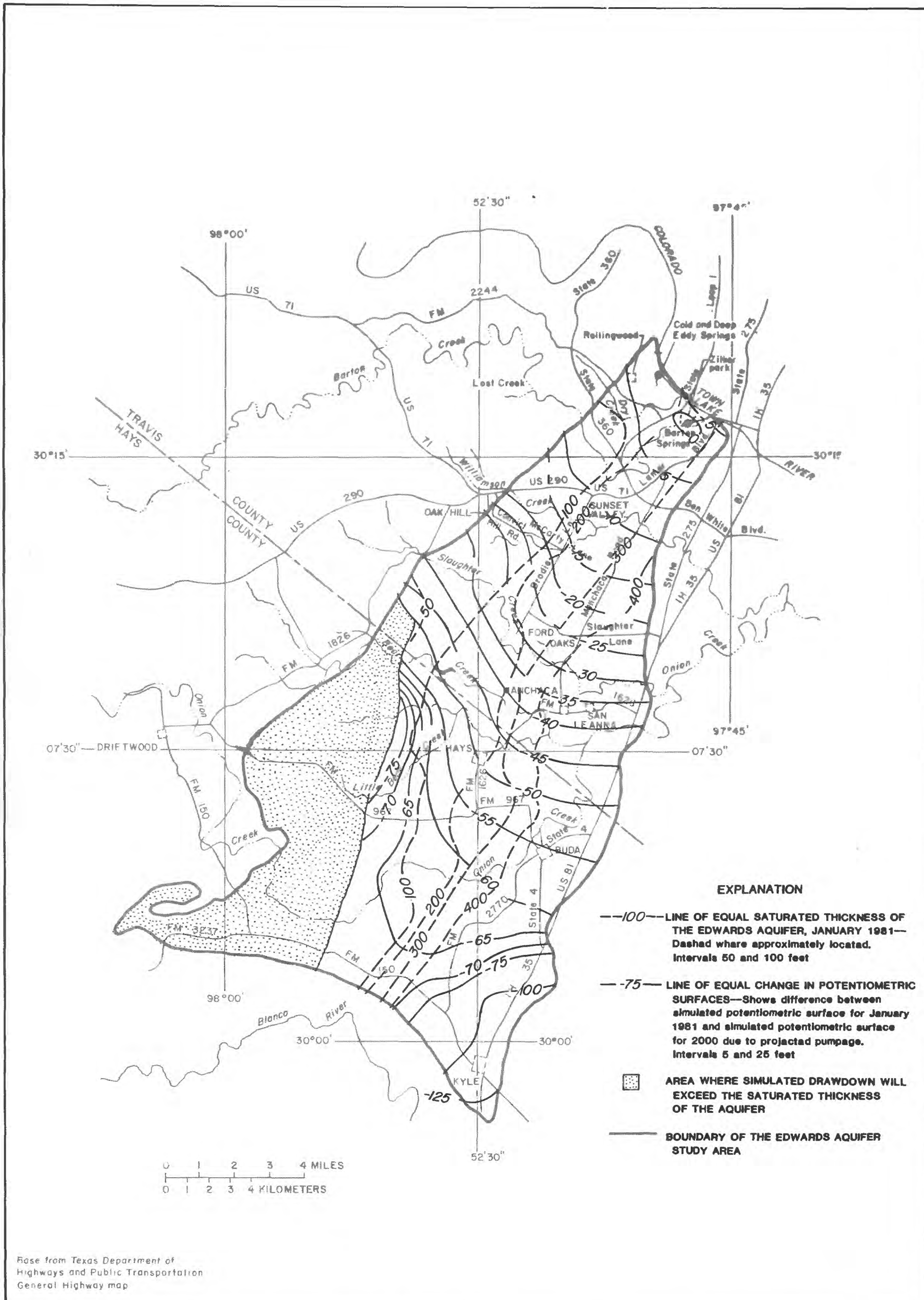

Figure 19.--Change in potentiometric surface between steady-state simulations for January 1981 and 2000 due to projected pumpage.

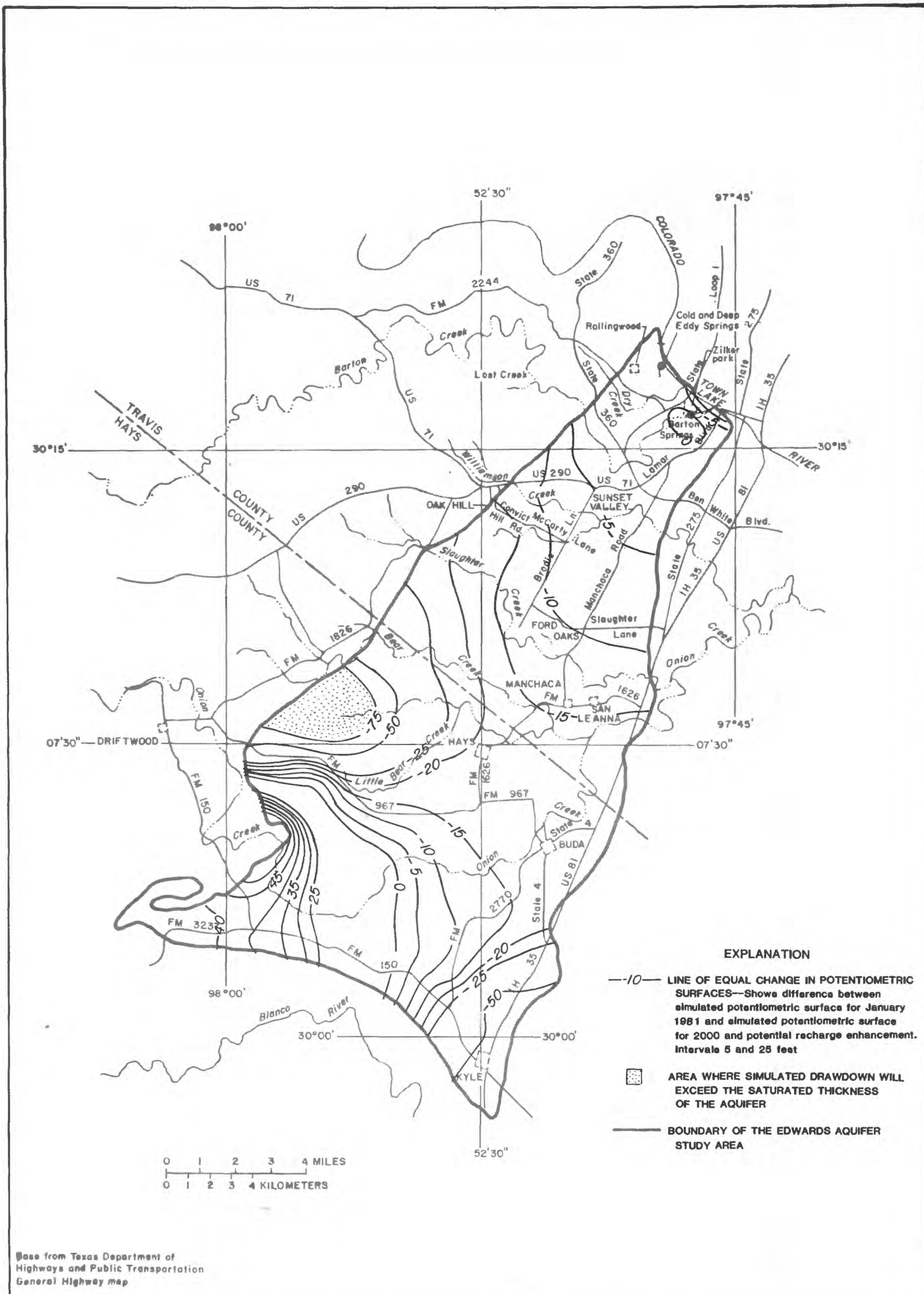

Figure 20.--Change in potentiometric surface between steady-state simulations for January 1981 and 2000 due to projected pumpage and potential recharge enhancement.

LIMITATIONS AND FUTURE SIMULATION STUDIES

Many assumptions for the model and the data used in the model, and qualifications of the simulations have been presented in this report. However, other information concerning the limitations and reliability of this model needs to be addressed. This modeling activity represents an initial effort to integrate all hydrologic data that affect the potentiometric surface and to determine net effects of projected withdrawals on a regional scale. The model used was appropriate at the time of the investigation, and all available hydrologic data pertinent to the study were included.

Nonetheless, certain assumptions underlying the use of the model were not totally met or cannot be substantiated by field conditions. For example, the hydrologic data were calibrated under present hydrologic conditions, which assumes no known subsurface recharge. However, the projected simulations can only be substantiated by future field data. Boundary conditions can only be approximated by the model, and the effect of water-level declines on water movement into the aquifer from the bad-water zone to the east, or across the ground-water divide from the south, cannot be determined at this time. Also, there are insufficient data to assess the effect of these declines on leakage into the aquifer from the underlying Upper Trinity aquifer.

Another limitation of the modeling effort involves the calibration procedure for the steady-state simulation. Other possible combinations of values for transmissivity, particularily among adjacent modeling cells, could produce similarly calibrated hydraulic heads. Assuming that transmissivity is anisotropic in certain areas could produce other combinations that also might yield the same calibrated hydraulic heads.

Studies that could improve the prediction capabilities of the projected simulations include:

1. An inventory of the number of wells and the withdrawal occurring from the aquifer. Only a few of the major well fields have a metering system for determining pumpage, so those volumes generally were estimated. An accurate accounting of this use would be needed before projections concerning the total available resources of the aquifer can be made.
2. A study of the extent and quantity of subsurface leakage from the upper Trinity aquifer into the Edwards aquifer.
3. A study of the effects of increased rates of pumpage on the ground-water divide that forms the southern boundary of the aquifer study area.

As pumpage from the aquifer increases, these studies could further the understanding of subsurface leakage created by water-level declines.

CONCLUSIONS

1. The hydraulic characteristics determined from an individual well are not necessarily representative of the aquifer on a local basis.
2. The transmissivities derived from the steady-state simulation vary from about 100 ft^2/d in the western part of the aquifer to more than 1,000,000 ft^2/d near Barton Springs.

3. The simulated hydraulic heads are more sensitive to changes in transmissivity in the western and southern parts of the aquifer than in the area near Barton Springs.

4. The mean specific yield for the aquifer derived from the model is 0.014.

5. Specific yields in the aquifer derived from the model vary from 0.008 in the western part of the aquifer to about 0.06 near Barton Springs.

6. The hydraulic heads and the springflow simulated by the model are relatively insensitive to changes in specific yield.

7. A simulation for 2000, using projected pumping rates, indicates that the aquifer would be dewatered in the southwest part of the aquifer and have large declines in the southeast.

8. Another simulation for 2000 was done using projected pumping rates and potential recharge enhancement. This simulation indicates that the potential enhanced recharge would raise future water levels as much as 120 ft in the western part of the aquifer and about 50 ft in the southeast.

9. Studies of subsurface leakage are needed in order to verify the prediction capabilities of projected simulations.

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