GROUND-WATER FLOW AND SOLUTE TRANSPORT IN THE EQUUS BEDS AREA, SOUTH-CENTRAL KANSAS, 1940-79

By Joseph M. Spinazola, J. B. Gillespie, and R. J. Hart

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CONVERSION FACTORS

Inch-pound units of measurement used in this report may be converted to the International System of Units (SI) using the following factors:

Multiply inch-pound unit	By	To obtain metric unit
inch	25.4	millimeter
foot	0.3048	meter
mile	1.609	kilometer
acre	4,047	square meter
square mile	2.590	square kilometer
acre-foot (acre-ft)	1,233	cubic meter
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
foot per day (ft/d)	0.3048	meter per day
foot squared per day (ft^2/d)	0.0929	meter squared per day
foot per day-foot (ft/d-ft)	1	meter per day-meter
acre-foot per year (acre-ft/yr)	1,233	cubic meter per year
inch per year (in/yr)	25.4	millimeter per year
inch per year per square mile [$(in/yr)/mi^2$]	9.8	millimeter per year per square kilometer
million gallons per day (Mgal/d)	0.04381	cubic meter per second
barrel	280.2	liter
degree Fahrenheit (°F)	[1]	degree Celsius (°C)

[1] °F = 1.8 °C + 32 and °C = 5/9 (°F-32).

DEFINITION OF TERMS

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 or springs.

Confined aquifer - An aquifer that contains water under pressure significantly greater than atmospheric. Its upper limit is the bottom of a bed of distinctly lower hydraulic conductivity than that of the aquifer material itself.

Evapotranspiration - Amount of water that is lost to the atmosphere by transpiration from vegetative growth and by evaporation from the soil.

Hydraulic conductivity - Amount of water at the existing kinematic viscosity that will move through a porous medium in unit time under a unit hydraulic gradient through a unit area measured at right angles to the direction of flow.

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

Hydraulic head - 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.

Hydrodynamic dispersion - Tendency for a solute to spread beyond the path determined strictly by convective flow in an aquifer. Hydrodynamic dispersion is caused by mechanical mixing and by diffusion.

Leakance - Vertical hydraulic conductivity of a confining bed divided by the thickness of the confining bed.

Longitudinal dispersivity - Component of hydrodynamic dispersion parallel to the direction of flow in an aquifer.

Perennial stream - Stream that flows throughout the year and has a channel that generally is below the water table.

Potentiometric surface - A surface that represents the levels to which water will rise in tightly cased wells.

Saturated thickness - Thickness of material in which all openings are filled with water under pressure greater than or equal to atmospheric.

Solute - Inorganic or organic constituents dissolved in a fluid.

Specific storage - Volume of water released from or taken into storage per unit volume of the porous medium per unit change in hydraulic head.

Specific yield - Ratio of the volume of water that the saturated material will yield by gravity drainage per unit volume of the material.

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

Transient state - Nonequilibrium conditions when hydraulic heads and the volume of water in storage change significantly with time.

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

Transverse dispersivity - Component of hydrodynamic dispersion perpendicular to the direction of flow in the aquifer.

Unconfined aquifer - An aquifer that has a water table.

Water table - A water surface in an aquifer defined by the levels at which water stands in wells that penetrate the aquifer just far enough to hold standing water. The pressure at the water surface is atmospheric.

GROUND-WATER FLOW AND SOLUTE TRANSPORT IN THE EQUUS BEDS AREA, SOUTH-CENTRAL KANSAS, 1940-79

By

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ABSTRACT

Water levels have declined about 30 feet from 1940 to 1980 in part of the Equus beds aquifer in south-central Kansas where the city of Wichita operates a well field. The aquifer is unconfined and consists of unconsolidated deposits of silt, clay, sand, and gravel of Pleistocene and Pliocene age. Saturated thickness of the aquifer, which underlies an area of 1,406 square miles, ranged from 0 to about 300 feet during 1980. Withdrawal by wells from the aquifer was about 130 million gallons per day during 1980. Total water demand, as projected by the Kansas Water Office, will be 39 percent, or 181 million gallons per day, greater during 2035 than during 1980. Ground water may provide a large part of the projected demand.

The Wellington aquifer is separated from the overlying Equus beds aquifer by about 250 feet of shale. The Wellington aquifer is confined and resulted from the dissolution of evaporite deposits that formed solution cavities or led to collapse that formed permeable rubble zones in consolidated Permian rocks.

The study was conducted to increase the understanding of the hydrology in the Equus beds area. A three-dimensional, finite-difference, ground-water flow model was developed to: (1) Reproduce hydrologic conditions in the flow system between the Equus beds aquifer and the underlying Wellington aquifer from 1940 to 1980 and (2) simulate the effect that future withdrawals could have on water supply in the Equus beds aquifer and on relationships between water levels in the Equus beds aquifer and the Wellington aquifer from 1980 to 2020. The model was developed using distributions based on descriptions of aquifer properties and on estimated rates for recharge and withdrawal by wells between 1940 and 1980. The model favorably reproduced both measured water levels and streamflow gains in the Equus beds aquifer for 1971 and 1980, and measured water levels in the Wellington aquifer.

The flow model then was used to simulate the effects of five pumping alternatives based on rates of withdrawal by wells from 1971-79. For the first alternative, withdrawal rates were decreased by one-half. Projected saturated thickness in the aquifer and streamflow gain were the greatest among the five alternatives. For the second alternative, withdrawal rates for 1971-79 were continued. Compared to results from the first alternative, saturated thickness was projected to decline from 10 to 40 feet in some areas, and streamflow gains were maintained for the Arkansas and Little Arkansas Rivers. For the three other alternatives, withdrawal rates were increased proportionally until rates were twice the 1971-79 rates. Pro-

jected saturated thickness decreased by as much as 80 feet in some areas for the doubled withdrawal rates; streamflow losses were projected for all three alternatives. Simulated water levels in the Equus beds aquifer were higher than those in the Wellington aquifer for all five alternatives.

A two-dimensional, finite-difference, solute-transport model was developed to: (1) Reproduce the movement of chloride ion in part of the Equus beds aquifer, including the Wichita municipal well field, from 1940 to 1980 and (2) simulate the effect that future withdrawal rates could have on the concentration of chloride ion from 1980 to 2020. Sources of the chloride ion were oilfield brine disposed from 1932-43 that is moving toward the well field and water in the Arkansas River. The model generally reproduced the distribution of measurements made during 1980.

The transport model then was used to simulate three pumping alternatives based on one-half, continued, and doubled rates of withdrawal by wells for 1971-79. Each simulation projected an increase in the concentration of chloride ion in the Wichita well field. The minimum projected increase was from about 20 to 40 milligrams per liter with one-half the withdrawal rates. The maximum projected increase was from about 90 to 440 milligrams per liter with the doubled withdrawal rates. The projections indicated that a continuous 1,000-milligram-per-liter source of chloride ion in streamflow losses from the Arkansas River had a greater effect on increasing chloride-ion concentrations in the Wichita well field than did the movement of residual oilfield brine.

INTRODUCTION

The physical and economic well-being of inhabitants in the Equus beds area, south-central Kansas, depends in part on the ability to anticipate the effect of an increasing demand on the quantity and quality of water in the Equus beds aquifer. The Equus Beds aquifer is the primary source of water available beneath an area of about 1,400 square miles for municipal, agricultural, industrial, and domestic uses. The aquifer is composed of saturated silt, clay, sand, and gravel that were deposited during Pleistocene and Pliocene time. Saturated thickness of the aquifer ranged from 0 to about 300 feet during 1980.

The city of Wichita historically has been the largest single user of water from the aquifer and pumped about 40 Mgal/d during 1980. Agricultural and industrial users of the aquifer pumped an additional 90 Mgal/d during 1980. The availability of ground water is likely to become increasingly important as total water demand increases. The Kansas Water Office (1984, p. 39) has projected that total water demand in the area will be 39 percent, or 181 Mgal/d, greater by 2035 than the demand during 1980.

There are three potential sources of contamination that threaten the water quality of the Equus beds aquifer. The first is brine that was disposed as part of former oilfield activities, the second is mineralized water in the Arkansas River, and the third is mineralized water in the underlying Wellington aquifer.

This study was conducted in cooperation with the Kansas Geological Survey to advance the understanding of the hydrology of the Equus beds area for the management of the water resources of the Equus beds aquifer. The purpose of this study was to: (1) Describe the flow of water in the Equus beds aquifer and between the Equus beds aquifer and the underlying Wellington aquifer, and (2) describe the movement of chloride ion in the Equus beds aquifer. The report is intended for water-resource managers, water-resource scientists, and the scientifically informed public.

Purpose and Scope

This report presents: (1) A geohydrologic description of the Equus beds and Wellington aquifers in the study area; (2) a description of the digital-modeling techniques used to represent the flow system in the Equus beds aquifer, flow between the Equus beds and Wellington aquifers, and the movement of chloride ion in part of the Equus beds aquifer; and (3) a discussion of the modeling results. Supplemental information of model data and results are available on magnetic tape from the U.S. Geological Survey office in Lawrence, Kansas.

Description of Study Area

The study area consists of 1,406 square miles in parts of Harvey, Marion, McPherson, Reno, and Sedgwick Counties, in south-central Kansas (fig. 1). The area is underlain by unconsolidated deposits, the so-called Equus beds, and it is locally referred to as the Equus beds area. The major cities in the study area are Hutchinson and Wichita. Other towns include Burrton, Halstead, McPherson, Newton, and Valley Center (pl. 1).

The Arkansas River is the major stream in the area and flows in a southeasterly direction between Hutchinson and Wichita. Alta Mills, Halstead, and Valley Center are situated along the Little Arkansas River, which joins the Arkansas River at Wichita. Sun, Sand, Black Kettle, and Emma Creeks are tributaries to the Little Arkansas River. The Smoky Hill River flows just north of the study area. Paint, Sharps, and West Kentucky Creeks are tributaries to the Smoky Hill River. Numerous other creeks are tributaries to the major streams.

The study area is in the McPherson Lowland and the eastern part of the Great Bend Lowland of the Arkansas River Lowlands physiographic region identified by Merriam (1963, p. 164-165). The land surface is flat within the Arkansas River valley flood plain, which is about 12 miles wide in the study area. The topography is gently rolling in the uplands that make up the remainder of the study area. Areas of wind-blown dune sand are present to the east of Hutchinson and along parts of the Little Arkansas River (pl. 1). The highest point in the area is about 1,650 feet above sea level in a sand-dune area near Hutchinson. The lowest point is about 1,290 feet above sea level near Wichita.

Williams and Lohman (1949, p. 26) described the climate of the area as being "... characterized by moderate precipitation, a wide range of temperature variations, moderately high average wind velocity, and comparatively rapid evaporation." The description continues to be apt. Normal annual precipitation for 56 years preceding 1940 was 30.37 inches at the Wichita weather station (fig. 1) of the National Oceanic and Atmospheric Administration (formerly the U.S. Weather Bureau). Average annual temperature at Wichita was 57.9 °F in 1940 (Williams and Lohman, 1949, p. 26). Normal annual precipitation for 1951-80 was 30.58 inches, and normal temperature for the period was 56.6 °F at the Wichita weather station (National Oceanic and Atmospheric Administration, 1980).

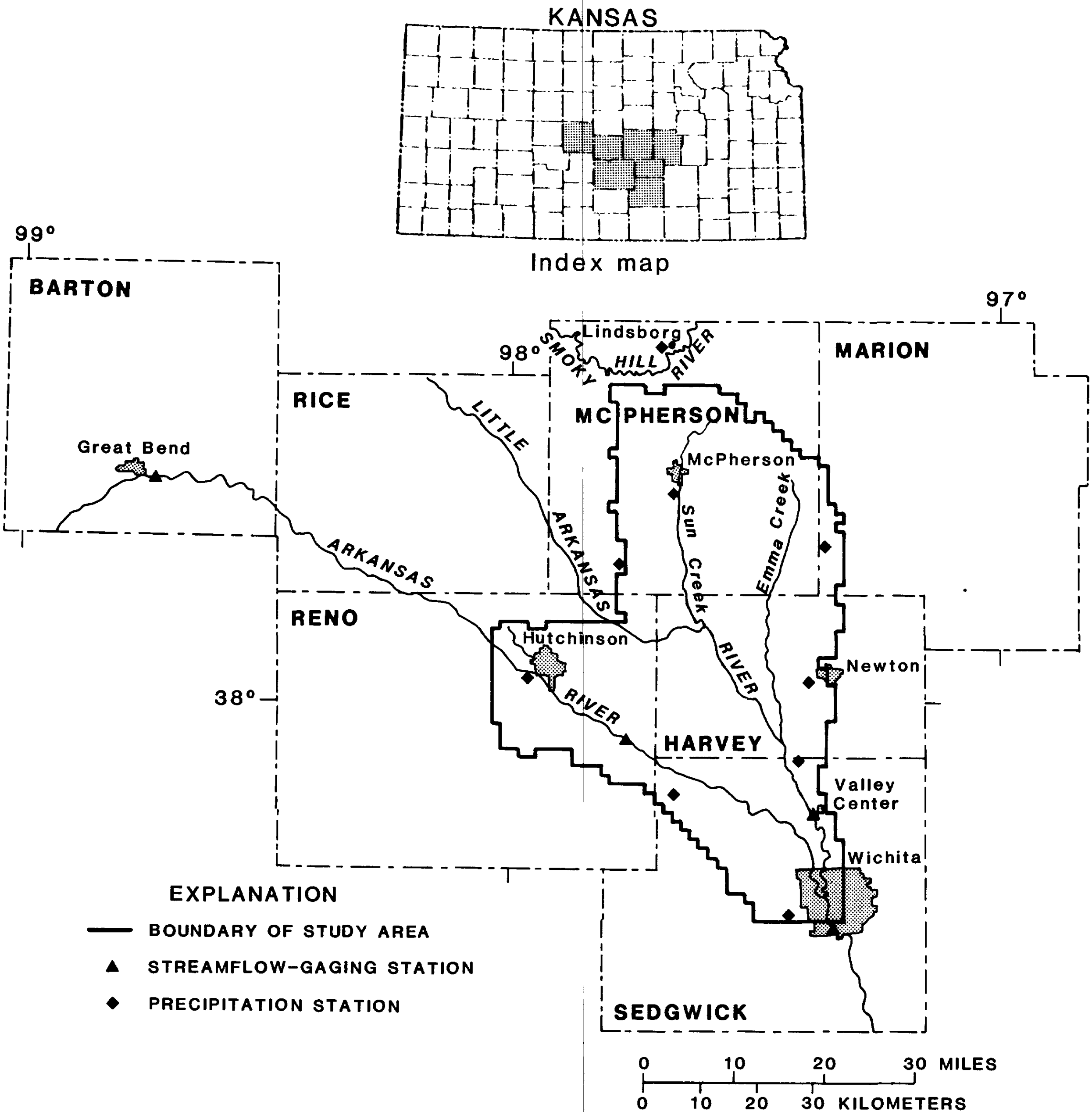


Figure 1.--Location of study area, streamflow-gaging stations, and precipitation stations.

Previous Studies

There have been many previous studies conducted in the Equus beds area. Williams and Lohman (1949) presented the most comprehensive treatise on the general geology and hydrology of the area. Other reports that concentrated on geohydrology include Stramel (1956, 1962a, 1962b, 1967), Williams (1946), Petri and others (1964), and Lane and Miller (1965). Reports that described geochemistry in relation to geohydrology include Leonard and Kleinschmidt (1976) and Hathaway and others (1981). Ground-water flow models for parts of the Equus beds area were prepared by Richards and Dunaway (1972), Green and Pogge (1977), and McElwee and others (1979). A water-quality modeling study of the Wichita well-field area was conducted by Sophocleous (1983). The geohydrology and a ground-water flow model of the Wellington aquifer was presented by Gogel (1981).

Methods of Study

A review of the existing literature was conducted to develop a conceptual representation of the ground-water flow system in the Equus beds area. Additional data were collected and analyzed to expand the understanding of the system. This phase was followed by the application of digital ground-water-flow and solute-transport models and by the analysis of model results.

Data from previous reports were used to define the geology and hydrology for the initial determination of aquifer properties. The bedrock map of the Equus beds aquifer (Williams and Lohman, 1949, pl. 7) was updated in Sedgwick County by Lane and Miller (1965, pl. 3) and also in this study by incorporating well logs supplied by well drillers to the Kansas Geological Survey (Lawrence, Kansas). Land-surface altitude was obtained from U.S. Geological Survey 1:62,500-scale topographic maps.

Hydraulic conductivity of the Equus beds aquifer was mapped using transmissivity data from Richards and Dunaway (1972) augmented with interpretations of aquifer-test data and well logs. The water-table map for the Equus beds aquifer for 1940 was prepared using data from the map prepared by Williams and Lohman (1949, pl. 1). Water-table maps for the Equus beds aquifer for 1971 and 1980 were compiled using data from WATSTORE, the U.S. Geological Survey's computer-based repository of geohydrologic data. Ground-water-withdrawal records on file with the Kansas State Board of Agriculture, Division of Water Resources (Topeka, Kansas), were compiled by use of water in each section to define ground-water withdrawal in the study area through time. Values of recharge estimated by previous studies in the area and in adjacent areas were reviewed and evaluated. Streamflow records maintained by the U.S. Geological Survey (Lawrence, Kansas) were reviewed to determine extremes of streamflow gains and losses between streamflow-gaging stations and to determine chloride-ion concentrations in river water.

The initial volume of brine that was disposed into part of the Equus beds aquifer was estimated from oil-production records on file with the Kansas Geological Survey (Lawrence, Kansas). The distribution of chloride-ion concentrations in part of the Equus beds aquifer was identified by Williams and Lohman (1949, pl. 29) for conditions during 1940-44, by Lane and Miller (1965, pl. 4), by Leonard and Kleinschmidt (1976, p. 14) for 1971, and by Sophocleus (1983, p. 9) for 1981.

Data compiled for the study were used to develop a conceptual model of the flow system in the Equus beds area. Digital computer models were applied to represent and to refine the conceptual model. Results from the digital models were compared to onsite measurements. When a satisfactory correspondence was achieved between model results and onsite measurements, the models were used to simulate the effects that specific management alternatives could have on storage and flow in the aquifer system, and the movement of chloride ion in part of the Equus beds aquifer. The computer models, model data, and model results for this study are available from the U.S. Geological Survey, Lawrence, Kansas.

Well-Numbering System

The system for numbering wells and test holes in this report is based on the U.S. Bureau of Land Management's system of land subdivision. The first number indicates the township south of the 40th parallel; the second indicates the range east (E) or west (W) of the Sixth Principal Meridian; and the third indicates the section in which the well is located. The first letter following the section number denotes the quarter section or 160-acre tract; the second, the quarter-quarter section or 40-acre tract; and the third, the quarter-quarter-quarter section or 10-acre tract. The letters are designated A, B, C, or D in a counterclockwise direction beginning in the northeast quarter of the section. Where there is more than one well in a 10-acre tract, consecutive numbers are added in the order in which the wells are inventoried. For example, 25-5W-2CBB indicates a well in the northwest quarter of the northwest quarter of the southwest quarter of sec. 2, T. 25 S., R. 5 W. (fig. 2).

Acknowledgments

Appreciation is expressed to Thomas Bell and Michael Dealy, the former and current (1985) managers, respectively, of the Equus Beds Groundwater Management District No. 2, for water-level measurements and other data. Appreciation is expressed to the Kansas State Board of Agriculture, Division of Water Resources, for access to water-use information, and to the Kansas Geological Survey for their continuing cooperation.

GEOHYDROLOGY

This section presents the geologic setting and hydrologic conditions that represent the framework and dynamics of the ground-water flow system in the Equus beds area. In this report the flow system is the combination of the Equus beds aquifer, the Wellington aquifer, and the intervening shale. The definitions of and relationships between the two aquifers are developed in the following sections.

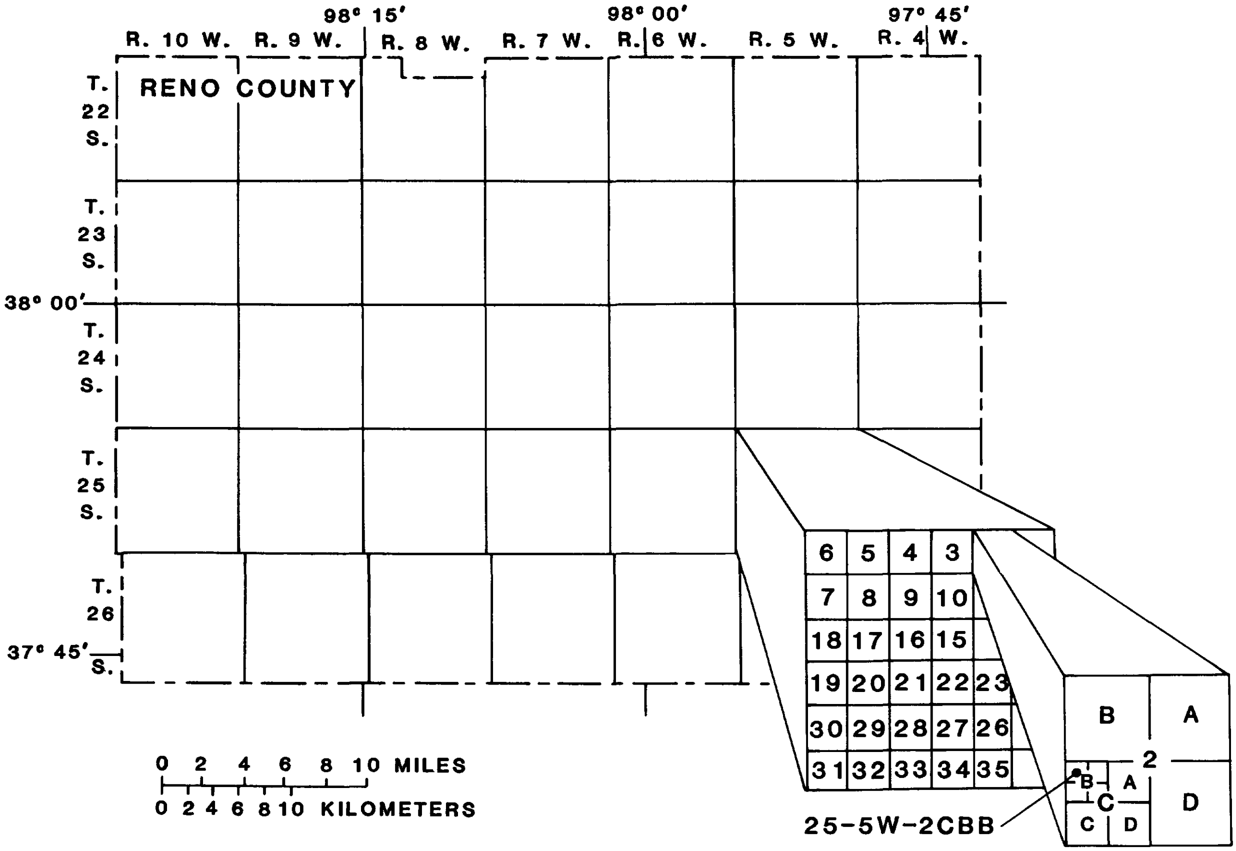


Figure 2.--Well-numbering system.

Geologic Setting

Unconsolidated Pleistocene and Pliocene deposits crop out throughout most of the study area (pl. 1). These unconsolidated deposits range in thickness from 0 to about 350 feet and are composed of heterogeneous, interfingered lenses of silt, clay, sand, and gravel of fluvial and eolian origin. Part of the unconsolidated deposits contain fossil bones and teeth of horses (generic name Equus) and are known locally as the Equus beds. The unconsolidated deposits are an important source of ground water and form the Equus beds aquifer.

Since Cretaceous time subsidence has occurred in the area. The subsidence created basins that affected the location and development of Pleistocene and Pliocene streams, and deposition by the streams (Gogel, 1981, p. 12). The basins are illustrated in the map showing the configuration of the bedrock surface below the unconsolidated deposits (fig. 3).

A major basin is present between Lindsborg (fig. 1), about 2 miles to the north of the study area, and Halstead, to the south (fig. 3). This trough in the bedrock surface is called the McPherson channel. Deposits that filled the McPherson channel probably were deposited by relatively slow-moving streams that meandered southward across the Equus beds area during late Pliocene and early Pleistocene time. In late Pleistocene time,

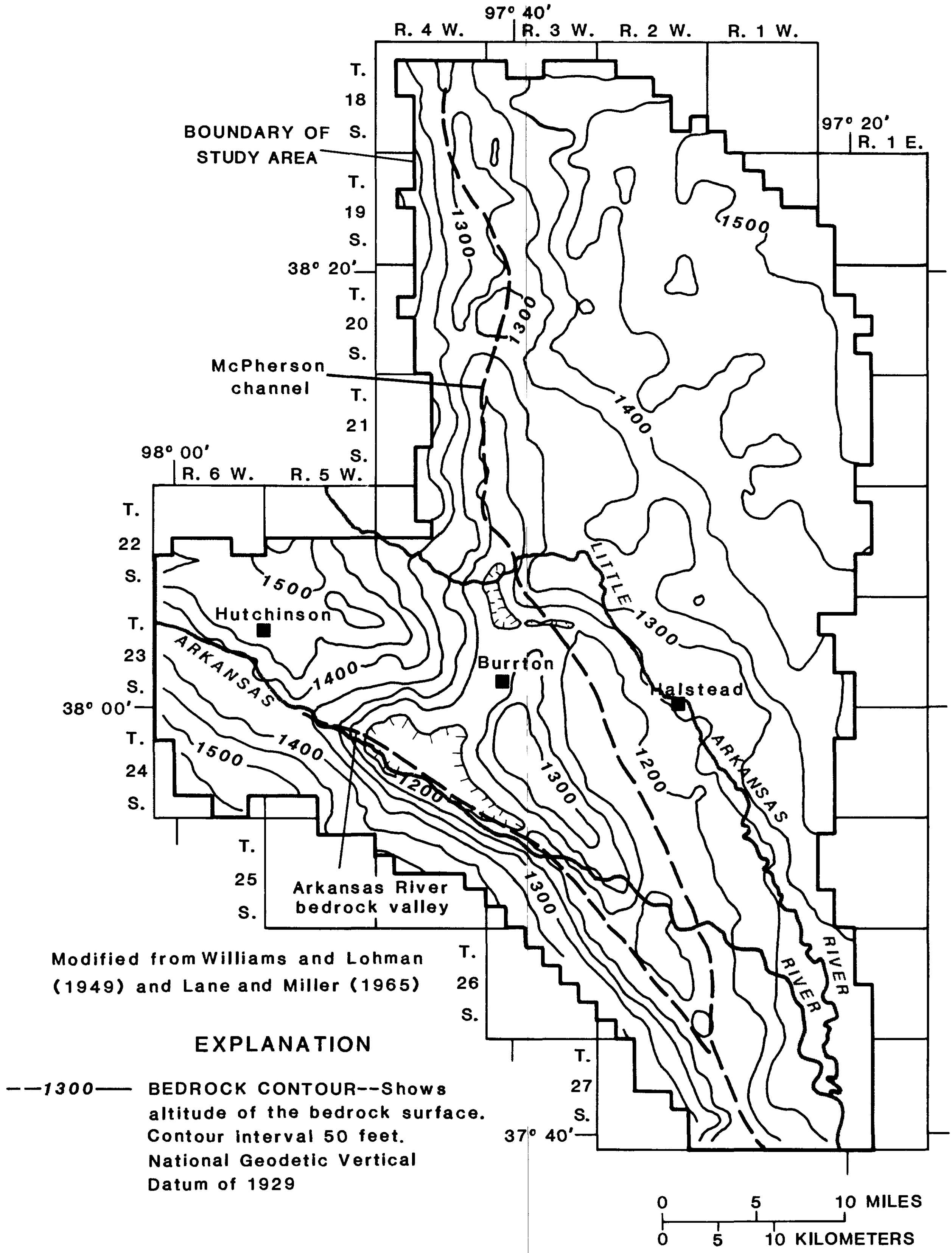


Figure 3.--Configuration of bedrock surface below the Equus beds aquifer.

the river that flowed within the McPherson channel was captured by the Smoky Hill River to the north. During this time, the ancestral Arkansas River became the dominant stream in the area.

The relatively faster moving ancestral Arkansas River scoured the older fluvial deposits from the Arkansas River valley and deposited much coarser grained material in their place. A ridge and saddle in the bedrock surface along a line from north of Hutchinson to south of Burrton separate the McPherson channel from the Arkansas River bedrock valley. For a further explanation of the geology in the Equus beds area, see Lohman and Williams (1949) and Lane and Miller (1965).

The bedrock surface beneath the unconsolidated deposits is formed on Lower Cretaceous and Lower Permian rocks. These rocks crop out around about one-half of the perimeter of the study area (pl. 1). Lower Cretaceous rocks consist of shale and sandstone and form the bedrock surface in the northeast part of the study area. The Lower Permian rocks consist of shale and evaporite deposits (salt, anhydrite, and gypsum) and form the bedrock surface for most of the rest of the study area. Formations in Lower Permian rocks of the Sumner Group include the Ninnescah Shale and underlying Wellington Formation. The Wellington Formation is divided into three units: the upper unnamed member, the Hutchinson Salt Member, and the lower unnamed member. Thickness of the Wellington Formation averages 750 feet in the study area. Lower Permian rocks of the Chase Group underlie the Wellington Formation. Rocks of the Chase Group consist of alternating sequences of limestone and shale. The stratigraphic relationship among Lower Permian rocks and unconsolidated deposits is illustrated by the geologic section on plate 1. For additional information on the geology of Lower Cretaceous and Lower Permian rocks in the area, see Williams and Lohman (1949) and Gogel (1981).

Equus Beds and Wellington Aquifers

In this report the Equus beds aquifer refers to the saturated part of most of the unconsolidated deposits of Pleistocene and Pliocene age in the study area. The aquifer was considered to be unconfined in this study. Some of the dune-sand deposits were not considered to be part of the aquifer. Stramel (1962b) indicated that the area of dune sand in T. 23 S., R. 3 W. (pl. 1), is underlain by about 40 feet of clay. Below the clay are about 200 feet of unconsolidated deposits in the McPherson channel (fig. 3). Water levels measured in wells completed in the dune sand were as much as 15 feet higher than those measured in wells completed in unconsolidated deposits below the clay. Therefore, these deposits of dune sand were considered to represent a local, perched aquifer that is related to, but generally separate from, the aquifer in the underlying unconsolidated deposits. In T. 23 S., R. 4 W. and R. 5 W., the clay is absent, and the saturated part of the dune sand was considered part of the Equus beds aquifer.

The configuration of the water table gives a general indication of the direction of water movement and the transmissivity of the aquifer. The configuration for the aquifer shown in figure 4 indicates that ground water moved from the northwest to the southeast in the Arkansas River valley during 1980. Ground water generally moved toward the river in the vicinity of the Little Arkansas River. Within the McPherson channel (fig. 3), ground water moved toward the center of the channel from the west and east, then moved either to the north toward the Smoky Hill River or to the south toward the Little Arkansas River. Along a flow path, widely spaced contours generally indicate areas of relatively greater transmissivity in the aquifer than in areas where contours are closely spaced.

Areas of greatest saturated thickness are associated with areas of subsidence (fig. 4). As much as 250 feet of saturated deposits were present in the southward-trending McPherson channel during 1980. Nearly 300 feet of saturated deposits were present in the Arkansas River bedrock valley (fig. 3). The Equus beds aquifer is contiguous with the Great Bend alluvial aquifer to the west of the Arkansas River valley, the aquifer in the Arkansas River alluvium to the extreme south of the study area, and the aquifer in the Smoky Hill River alluvium to the north. The aquifer naturally thins to zero thickness to the southwestern and eastern boundaries of the study area, as well as west of the McPherson channel.

The subsidence that affected deposition during Pleistocene and Pliocene time was caused by the dissolution of evaporite deposits by circulating ground water mainly in the Hutchinson Salt Member of the Wellington Formation. In areas where the evaporite deposits have been dissolved, subsidence and collapse of the overlying geologic units replaced the evaporite deposits with a zone of rubble many times more permeable than the adjacent undisturbed shale. This rubble zone along with associated solution cavities is called the Wellington aquifer (Gogel, 1981, p. 3). The thickness of the Wellington aquifer is variable but is thinner than the Wellington Formation in any given location in the study area. The extent of the Wellington aquifer in the study area is shown on plate 1. The Wellington aquifer extends beyond the boundary of the study area by about 30 miles to the north and by about 40 miles to the south. The aquifer was considered to be confined in this study. The Wellington aquifer is separated from the overlying Equus beds aquifer by shale that averages 250 feet thick. The shale is mainly in the upper member of the Wellington Formation and is the confining bed between the Wellington aquifer and the Equus beds aquifer. For further information on the Wellington aquifer, see Gogel (1981).

Recharge to Aquifers

Recharge from precipitation is water that reaches the water table through the unsaturated zone and adds water to storage in an aquifer. Recharge from precipitation occurs after the evapotranspiration demand in the unsaturated zone above the water table has been met. Recharge constitutes about 20 percent of precipitation in the Equus beds area (Williams and Lohman, 1949, p. 215).

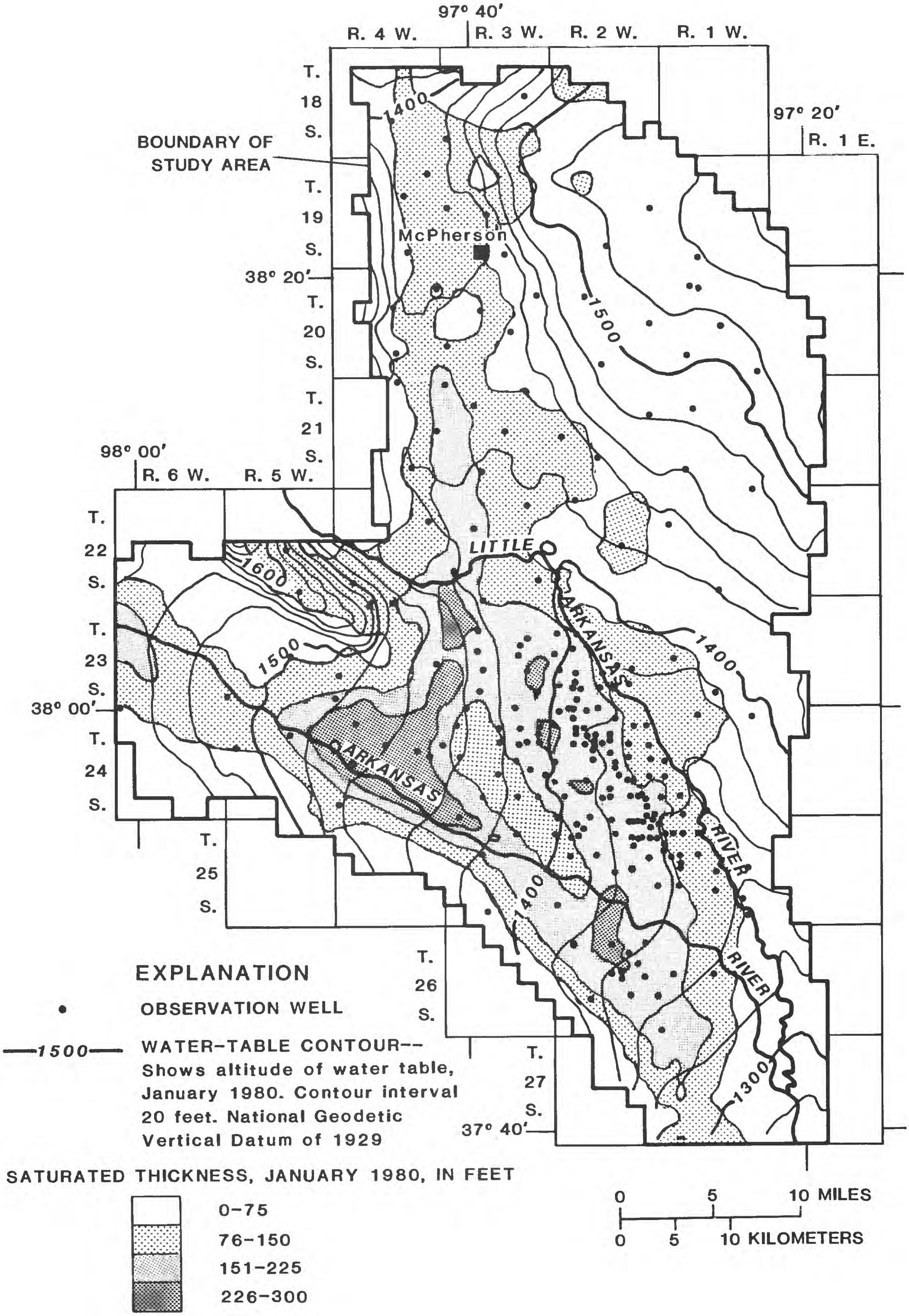


Figure 4.--Water-table configuration and saturated thickness of the Equus Beds aquifer, January 1980.

Stramel (1956, p. 41-42) determined that recharge from precipitation to the Equus beds aquifer in the Wichita well field varied from 3.75 to 8.8 in/yr, depending on the quantity of precipitation that occurred between 1940 and 1955. Using parameter-estimation techniques, Sophocleous (1983, p. 23) computed recharge values of 1.6 and 6.4 in/yr for different parts of the Wichita well field.

Other sources of recharge to the Equus beds aquifer in the study area are inflow from other aquifers and infiltration of streamflow. Ground water moves into the Equus beds aquifer laterally from the west as underflow where the Equus beds aquifer meets the Great Bend alluvial aquifer. When the altitude of the water table in the Equus beds aquifer is lower than the altitude of a stream, the stream loses water to the aquifer.

Recharge to the Wellington aquifer is the downward movement of water from the Equus beds aquifer (Gogel, 1981, p. 26-29). The potentiometric surface in the Equus beds aquifer is higher than the potentiometric surface in the Wellington aquifer in most of the study area. The difference in the potentiometric surfaces provides the hydraulic potential necessary for the movement of water through the confining shale into the Wellington aquifer. Also, wells have been used for injecting brine into the Wellington aquifer. About 1 million barrels of brine were injected into this aquifer between 1968 and 1975 (Gogel, 1981, p. 18).

Discharge from Aquifers

Natural discharge from the Equus beds aquifer is by underflow, by evapotranspiration, and by seepage to streams. Water moves as underflow from the Equus beds aquifer to the alluvium in the Arkansas River valley to the south. Underflow also occurs from the Equus beds aquifer to the alluvium in the Smoky Hill River valley at the northern end of the aquifer. Discharge also occurs through the process of ground-water evapotranspiration whenever the water table in the aquifer is close enough to the land surface that water can evaporate directly through soil pores or whenever plant roots can intercept ground water. These conditions are presumed to be present in the Arkansas River valley and in other parts of the study area where depths to the water table are 10 feet or less. When the altitude of the water table in the aquifer is higher than the altitude of a through-flowing stream, that stream gains water from the aquifer.

At the northernmost edge of the Equus beds area, Pleistocene and Pliocene deposits thin appreciably approaching the Smoky Hill River. The thinning of the deposits is due to decreasing land-surface altitude in relation to a relatively constant bedrock altitude. For example, the altitude of the land surface decreases from 1,430 feet in sec. 6, T. 18 S., R. 3 W., to 1,320 feet in sec. 31, T. 17 S., R. 3 W., in the Smoky Hill River valley. This represents a vertical drop of 110 feet over a length of 1.5 miles. Comparatively greater amounts of natural vegetative growth were observed in this area during October 1984. Aerial photographs (Rott, 1983) along the Smoky Hill River show slightly darker shading in a strip generally within 1 mile on either side of the river. Williams and Lohman (1949, p. 133) mentioned the occurrence of seeps along the banks of the Smoky Hill River. Based on these observations, this strip was considered to be a seepage front where discharge from the aquifer supplied additional water to streamflow and to evapotranspiration.

Withdrawal by wells was the greatest source of discharge from the Equus beds aquifer by 1980. Between 1900 and 1940 water was withdrawn mainly for municipal and industrial uses in the vicinities of Hutchinson, McPherson, and Wichita. The water-table map for 1940 (Williams and Lohman, 1949, pl. 1) does not show the formation of a cone of depression in these areas, indicating that storage in the aquifer was not appreciably affected by withdrawal from wells prior to 1940. Withdrawal for municipal use increased rapidly from 1940 to 1952; from 1952 to 1979, this withdrawal increased only slightly (fig. 5). Except for 1940 and 1969, withdrawal for municipal use has been greater than that for industrial use. Withdrawal for industrial use doubled from pre-1955 rates between 1955 and 1967.

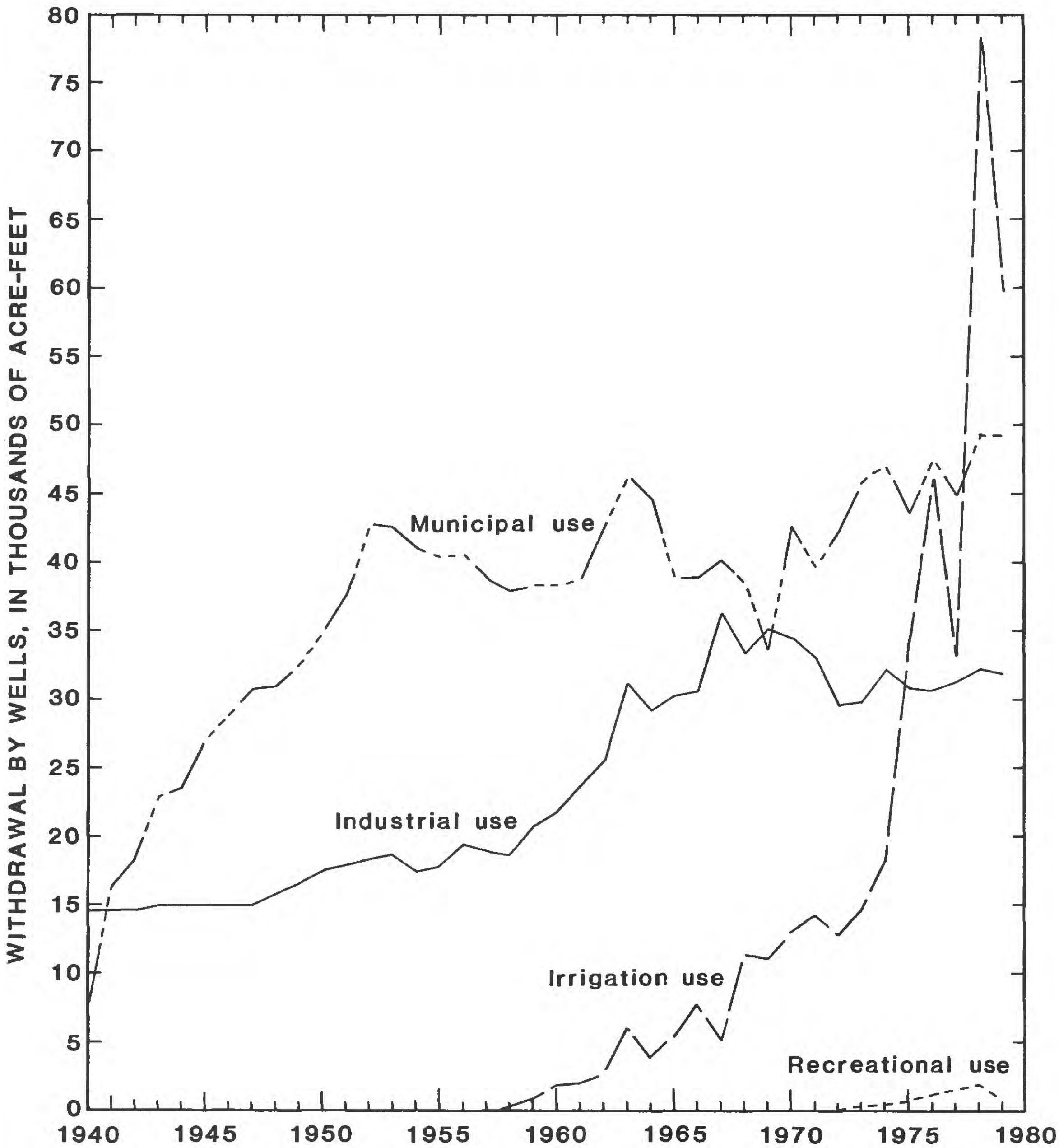


Figure 5.--Annual water withdrawal by wells from the Equus beds aquifer for various uses, 1940-79 (data from Kansas State Board of Agriculture, Topeka, Kansas).

Withdrawal for irrigation use has increased more than six-fold between 1972 and 1978. Withdrawal for irrigation use surpassed that for industrial use during 1975 and that for municipal use during 1978. The rate of withdrawal for all uses was more than three-times greater between 1970 and 1978 than between 1940 and 1970 (fig. 6).

Discharge from the Wellington aquifer occurs at locations beyond the boundaries of the study area to the north and south. Gogel (1981, pl. 1) identified an area of potential discharge corresponding to the bedrock low

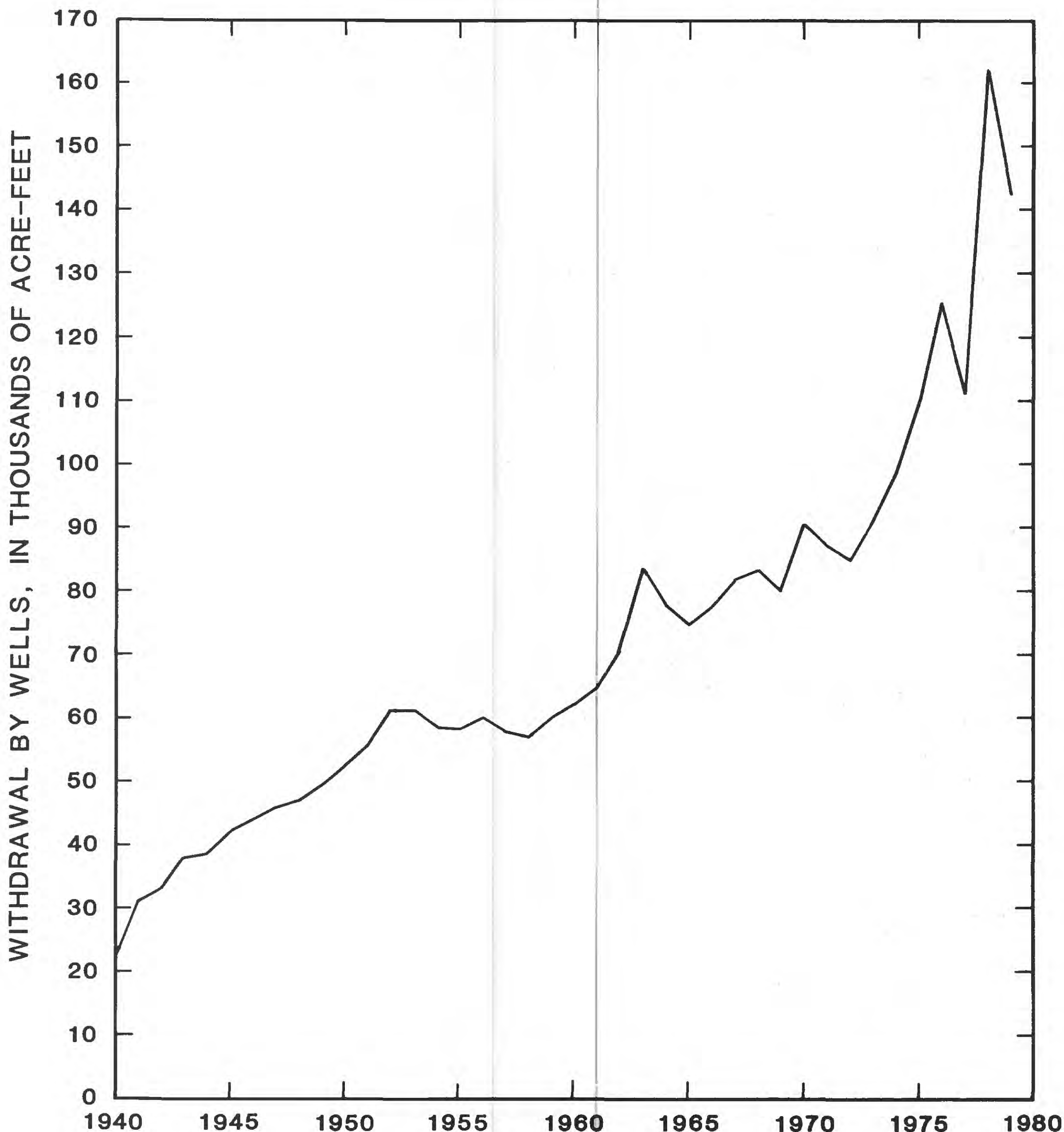


Figure 6.--Total annual water withdrawal by wells from the Equus beds aquifer, 1940-79 (data from Kansas State Board of Agriculture, Topeka, Kansas).

beneath the Arkansas River valley between Hutchinson and Burrton. In this area, the potentiometric surface in the Wellington aquifer was higher than the potentiometric surface in the Equus beds aquifer. However, the relatively impermeable nature of the confining bed between the aquifers restricted upward flow (Gogel, 1981, p. 28-29). No discharge of water from the Wellington aquifer to the Equus beds aquifer has been identified at the present time (1985).

Relationships Between Ground and Surface Water

The Arkansas and the Little Arkansas Rivers are maintained by discharge from the Equus beds aquifer during low flow (Williams and Lohman, 1949, p. 130-33; Lane and Miller, 1965, p. 47-49). The characteristics of selected streamflow hydrographs were analyzed to obtain a further indication of the interaction between the ground and surface water. Comparisons of hydrographs from streamflow-gaging stations in the study area (fig. 1) during selected periods of flow are shown in figure 7. The difference between the sum of the discharges at the upstream stations (near Hutchinson and Valley Center) and discharge at the downstream station (Wichita) indicates that the gain to both the Arkansas and Little Arkansas Rivers from the aquifer during concurrent periods of low flow ranged from about 5 to 50 ft^3/s. The individual gain for either of the two rivers could not be identified separately based on the existing data.

Sharp's Creek, Paint Creek, West Kentucky Creek, Sun Creek, and Emma Creek are ungaged. Flow was measured in each of these creeks at least once between 1968 and 1981, as shown in the following table:

Creek	Streamflow gain (cubic feet per second)	Measurement date (month-day-year)
Emma	5.64	11-21-68
	5.50	11-13-69
	6.71	12-08-69
	4.11	02-18-70
	7.74	04-15-71
	.55	10-01-71
	1.01	11-08-78
Sun	[1] 6.33	11-21-68
	[1] 4.97	11-13-69
	[1] 3.70	02-20-70
	[1] 2.19	09-10-70
	[1] 3.89	10-01-71
	[1] 5.67	11-08-78
West Kentucky	.1	03-18-81
Paint	.7	03-18-81
Sharps	2.74	03-18-81

[1] Includes effluent from municipal wastewater-treatment plant; ground water is a source of wastewater processed by treatment plant.

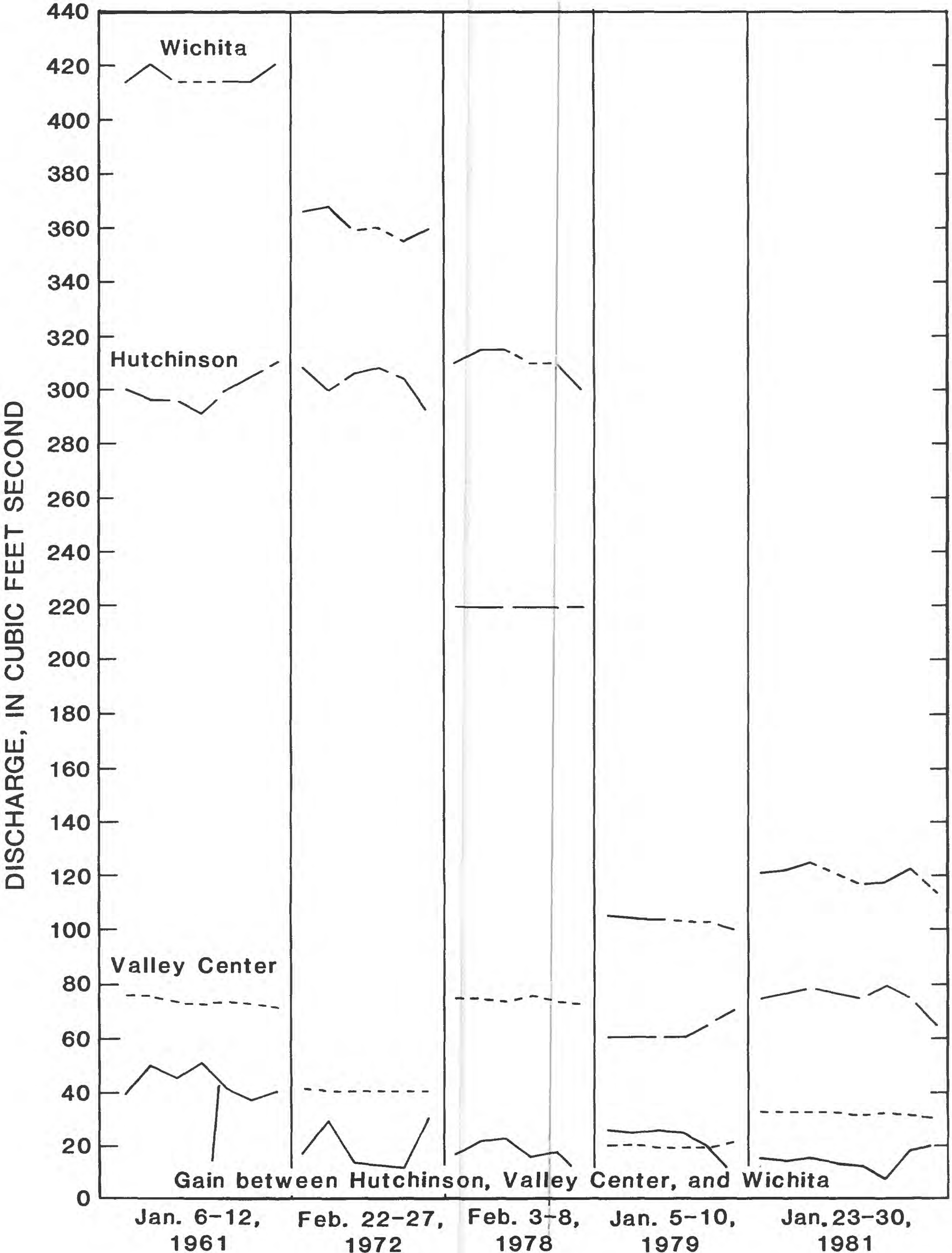


Figure 7.--Comparison of hydrographs for streamflow-gaging stations near Hutchinson and Valley Center and at Wichita during selected periods of low flow.

Flow was observed by the authors in all of these creeks during all seasons of the year between 1978 and 1984. These streams were considered perennial for this study, and base flow was considered to be maintained in these streams by discharge from the Equus beds aquifer when there was no overland runoff.

Sources of Mineralized Water

During the 1930's and early 1940's, surface "evaporation pits" and the Wellington aquifer were used to dispose brine, a by-product of oil production (Williams and Lohman, 1949, p. 173). Instead of evaporating from the surface pits, much of the brine seeped downward through permeable, sandy soil into the Equus beds aquifer. Brine disposal into the Wellington aquifer increased the potential of contaminating the Equus beds aquifer with brine through leaky well casings. An additional hazard was the possibility of dissolving the evaporite deposits present in the Permian rocks, thereby increasing the potential for collapse of overlying rocks and, possibly, creating paths for the upward migration of mineralized water from the Wellington aquifer into the Equus beds aquifer. By the mid-1940's, the problems with these two disposal methods became obvious, and most of the brine disposal in the study area was restricted to deep-well injection into Cambrian and Ordovician rocks.

The reported average concentration of chloride ion in five brine samples from the active oilfield near Burrton was 120,000 mg/L (milligrams per liter) (Schoewe, 1943, p. 55). Concentrations of chloride ion in the uncontaminated part of the Equus beds aquifer were as small as 20 mg/L. The area where brine disposal took place was about 8 miles northwest of the Wichita municipal well field. Water samples from observation wells indicated that the concentration of chloride ion in the Equus beds aquifer between the brine-disposal area and the Wichita well field increased between 1940 and 1980.

Mineralized water in the Arkansas River, which is considered to originate from sources outside the study area (Williams and Lohman, 1949, p. 170), can be induced into the Equus beds aquifer when the altitude of the river is higher than the altitude of the water table in the aquifer. Streamflow and water-quality records for the Hutchinson streamflow-gaging station, on file with the U.S. Geological Survey (Lawrence, Kansas), indicate that concentrations of dissolved solids in Arkansas River water varied with discharge in the river and ranged from about 200 to 2,000 mg/L between 1959 and 1980. Chloride-ion concentrations were measured at the same station between 1961 and 1978 and ranged from 363 to 907 mg/L.

Water in the Wellington aquifer is very mineralized. The concentration of chloride ion averaged about 150,000 mg/L in 15 water samples from the aquifer (Gogel, 1981, p. 39). The source of mineralized water in the aquifer was attributed to the natural dissolution of evaporite deposits in Lower Permian rocks and to the injection of oilfield brine. To compensate for the effect of density, water levels in the Wellington aquifer are represented in this study by equivalent freshwater levels, as described by Gogel (1981, p. 21). The use of equivalent freshwater levels for the Wellington aquifer was not considered to have an appreciable effect on computations of flow (discussed below). Data from Gogel (1981, p. 28-29) indicate that recharge through the confining bed to the Wellington aquifer is less than 0.01 (in/yr)/mi^2 and that nothing but brine was injected into the aquifer. Neither of these sources was considered to change the density of water in the Wellington aquifer.

GROUND-WATER FLOW MODEL

A modular, three-dimensional, finite-difference, ground-water flow model (McDonald and Harbaugh, 1984) was used to simulate the response of the flow system in the Equus beds area to imposed stresses and to simulate flow in and between the Equus beds and the Wellington aquifers. This model is referred to as the flow model in the remainder of this report.

Theory

Flow in an aquifer at any location for any time can be described by the following continuous partial-differential equation (McDonald and Harbaugh, 1984, p. 7):

$$\frac{\partial}{\partial x}\left(K_{xx}\frac{\partial h}{\partial x}\right) + \frac{\partial}{\partial y}\left(K_{yy}\frac{\partial h}{\partial y}\right) + \frac{\partial}{\partial z}\left(K_{zz}\frac{\partial h}{\partial z}\right) - W = S_s\frac{\partial h}{\partial t} \qquad (1)$$

where

x, y, and z are cartesian coordinates;

K is hydraulic conductivity, in feet per day;

h is hydraulic head, in feet;

W is volumetric flux per unit volume for sources or sinks or both, in $days^{-1}$;

S_s is specific storage of the aquifer, in $feet^{-1}$; and

t is time, in days.

The flow equation cannot be solved directly for any but the simplest of situations; therefore, the flow model used the strongly implicit, finite-difference numerical method to approximate the solution of equation 1. This method discretizes the flow equation into increments of time and space and transforms the continuous partial-differential equation into a matrix of linear algebraic equations that then are solved simultaneously by the flow model. For more details on this subject, see McDonald and Harbaugh (1984).

In general, the flow model is a simplified mathematical solution that is analogous to a group of complex natural processes. In some cases, the mathematical equivalent of a particular process has yet to be developed. In others, the model can be used to simulate fairly complicated processes. However, the availability of the detailed data required to represent a particular process with the model may be a limiting factor. In any case, several assumptions were made about the flow system in order to apply the flow model to the flow system in the Equus beds area. These assumptions were:

(1) Flow was horizontal in the Equus beds and Wellington aquifers;

(2) each aquifer was homogeneous and isotropic;

(3) the effects from stress on the flow system beyond artificial boundaries used in the model were minimal;

(4) the density of water was uniform and was that of freshwater;

(5) temperature in each aquifer was constant;

(6) flow components in each aquifer were parallel to the planes of the axes of the model grid;

(7) flow across the confining bed is perpendicular to flow in the aquifers; and

(8) there is no change in storage in the confining bed.

One cannot overemphasize the importance that the assumptions made about the flow system are reasonable. When assumptions made about the flow system for the model are valid, model results can be interpreted with some degree of confidence. However, if the assumptions are invalid, model results warrant little confidence.

Equus Beds Aquifer

The Equus beds aquifer was represented by the upper layer of the flow model. The finite-difference grid used to represent the aquifer is shown in figure 8. Each square within the grid is referred to as a cell. The cell was the smallest subdivision of the model to which aquifer properties were attributed. Each cell represented a surface area of 1 square mile. Cells were either active, inactive, or represented a boundary condition. In an active cell, the hydraulic head was allowed to vary in response to recharge to and discharge from the ground-water system. The properties of boundary conditions are described in the next section.

Boundary Conditions

The boundary conditions that were used in the flow model were no flow, constant head, and general head. The location of these boundaries are shown in figure 8. The model required that the grid representing the aquifer be surrounded by no-flow boundaries. As the name indicates there is no flow through a no-flow boundary. The majority of no-flow boundaries represent the physical edge of the aquifer.

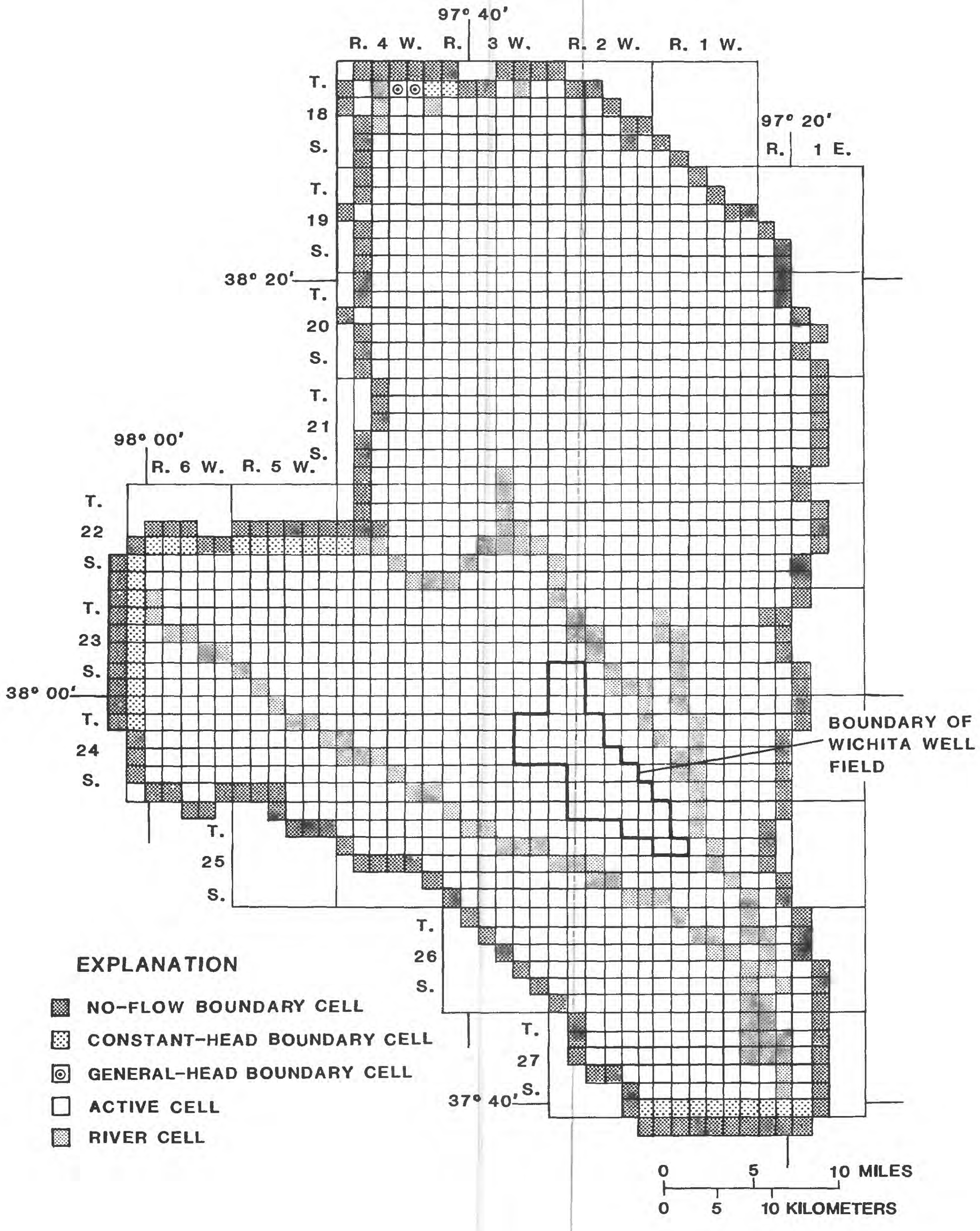


Figure 8.--Finite-difference grid, model boundary conditions, and active and river cells representing the Equus beds aquifer in the ground-water flow model.

Constant-head boundaries were used in the model to represent a hydraulic connection between laterally adjacent aquifers. These boundaries were located to the west where the Equus beds aquifer is contiguous to the Great Bend alluvial aquifer, to the south at the continuation of the aquifer in the Arkansas River alluvium, and to the north where the aquifer is contiguous to the aquifer in the Smoky Hill River alluvium. No change in the water level can occur at constant-head cells. However, recharge from or discharge to these cells was not limited by the model.

General-head boundaries were used to represent a seepage front at the northern part of the aquifer. At a general-head boundary cell, discharge through the cell was computed by the flow model as a function of the difference in the hydraulic head between the boundary cell and a fixed hydraulic head adjacent to the boundary. Constant-head and general-head boundary cells were monitored closely throughout the modeling effort to insure that discharges computed by the flow model agreed with discharges computed analytically.

Recharge

Recharge was assumed to be a function of precipitation, the nature of the land surface, and the nature of the unsaturated material between the land surface and the water table. Precipitation varied spacially as well as temporally in the study area. Precipitation was averaged from data collected at nine precipitation stations maintained by the National Oceanic and Atmospheric Administration. The stations are shown in figure 1.

Soil type and unsaturated-clay thickness were assumed to be the two factors that controlled the proportion of precipitation that became recharge. Soil type was assumed to affect the infiltration of precipitation and evapotranspiration by vegetation. Soil type was compiled from data available from U.S. Soil Conservation Service county publications (Rockers and others, 1966; Hoffman and Dowd, 1974; Penner and Wehmueller, 1979; Rott and others, 1983). These data indicated ranges of soil properties for infiltration rates and available moisture in the soil column for the soil associations present in the study area. The thickness of unsaturated clay between the land surface and the water table was compiled from well-drilling data. Unsaturated clays were considered to retard the vertical movement of water from the land surface, thereby extending the time that water remained available for evapotranspiration from the unsaturated zone. The combination of infiltration rates and unsaturated clay thicknesses were used to define a recharge factor for each cell in the grid. Products of the recharge factor and precipitation were used to define recharge for the initial-condition and transient-model simulations.

Ground-Water Evapotranspiration

Ground-water evapotranspiration from the aquifer was simulated for cells where the depth of the water table was less than 10 feet below land surface. The evapotranspiration rate for those cells was prorated by the flow model as an inverse relationship between the maximum evapotranspiration rate and the depth to the water table below land surface. No quantitative data on ground-water evapotranspiration in the study area were known to exist. An initial value of 1.2 in/yr for the maximum evapotranspiration rate was derived from a ground-water modeling study in southwestern Kansas (Barker and others, 1983, p. 39). The final value of 3.5 in/yr used for the maximum ground-water evapotranspiration rate in the Equus bed area was derived by trial-and-error, described later in the section "Discussion of Flow-Model Results."

River Leakage

River-aquifer interaction was controlled by the relationship among hydraulic head in the river, hydraulic head in the aquifer, and riverbed conductance for each active river cell in the model grid (fig. 8). Riverbed conductance was the product of the vertical hydraulic conductivity of the riverbed and length and width of the river reach for the cell, divided by the thickness of the riverbed. Hydraulic head in the river was fixed at 1 foot above riverbed altitude, and hydraulic head in the aquifer was calculated by the flow model. Lengths of river reaches were planimetered from topographic maps. Average widths for each river were determined from onsite observations. Thickness of the riverbed was arbitrarily chosen as 1 foot. Uniform values for vertical hydraulic conductivity and active river width were assigned for each river or creek as follows:

River or creek	Vertical hydraulic conductivity (feet per day)	Active width (feet)
Arkansas River	1.0	20
Little Arkansas River	1.0	20
Emma Creek	.5	10
Sun Creek	.8	10
Indian Creek	.5	10
Sharps Creek	.5	10

No data for vertical hydraulic conductivity of the riverbed were available for the Equus beds area. The values chosen for vertical hydraulic conductivity were near or within the measured range of 0.78 to 2.67 ft/d cited by Barker and others (1983, p. 31) for the Arkansas River in southwestern Kansas and eastern Colorado.

Hydraulic Conductivity

Various data were used for determining the hydraulic-conductivity values for the Equus beds aquifer used in the model:

1. Transmissivity map prepared by Richards and Dunaway (1972) and the saturated-thickness map presented in figure 4 of this report; calculated hydraulic-conductivity values ranged from 10 to 350 ft/d.

2. Aquifer-test analyses presented by Reed and Burnett (1985); calculated hydraulic-conductivity values ranged from 15 to 455 ft/d.

3. Laboratory analysis of lithologic samples from 12 test holes in the Wichita well field presented by Williams and Lohman (1949, p. 101); calculated hydraulic-conductivity values ranged from 28 to 330 ft/d.

4. Results of a study of the deposits in the Arkansas River valley in southwestern Kansas presented by Barker and others (1983, p. 25); the hydraulic-conductivity value for these deposits, which are similar to those in the study area, was 800 ft/d.

5. The depositional history previously discussed in this report relating to the McPherson channel, the Arkansas River bedrock valley, and the dune-sand deposits north and east of Hutchinson.

The values of hydraulic conductivity assigned to active cells in the model ranged from 5 to 750 ft/d (fig. 9). Hydraulic-conductivity values of 5, 10, and 25 ft/d were combined into one interval shown in figure 9. The actual distribution of hydraulic conductivity probably is much more diverse than the distribution used in the model analysis.

Wellington Aquifer

The Wellington aquifer was represented by the lower layer of the flow model. The finite-difference grid used to represent active cells and boundary conditions for the aquifer are shown in figure 10. No-flow boundaries were used around all the active cells in the grid. Most of the no-flow boundaries represented the physical edge of the aquifer. General-head boundaries were used to represent flow out of the study area to the north and south.

The distribution of transmissivity and vertical leakance used to represent the Wellington aquifer in Gogel's (1981) digital model analysis also were used in this study. Transmissivity of the aquifer ranged from 8.64 to 2,592 ft^2/d (fig. 11). The simulation of hydrologic conditions in the Equus Beds and Wellington aquifers was considered more important than simulating conditions in the confining bed between the aquifers. Therefore, vertical flow through the confining bed was simulated as a function of the vertical leakance between the two aquifers (see McDonald and Harbaugh, 1984, p. 58). The vertical leakance of the confining bed ranged from 0.6×10^{-8} to 9.68×10^{-7} ft/d-ft (fig. 12).

Initial-Condition Simulation

The initial condition in the flow system was simulated for 1940, a year that preceded large-scale withdrawal by wells from the Equus beds aquifer. Withdrawal by wells prior to 1940 was assumed to have been negligible. Withdrawal rates for the Equus beds aquifer for the initial-condition simulation were obtained from records on file with the Kansas State Board of Agriculture, Division of Water Resources (Topeka, Kansas). Recharge (fig. 13) was defined as the product of the recharge factor, discussed earlier in the report, and 1940 normal precipitation (National Oceanic and Atmospheric Administration, 1940) averaged for the precipitation stations in the area. The rate of brine injection into the Wellington aquifer was that used in the digital-model analysis by Gogel (1981).

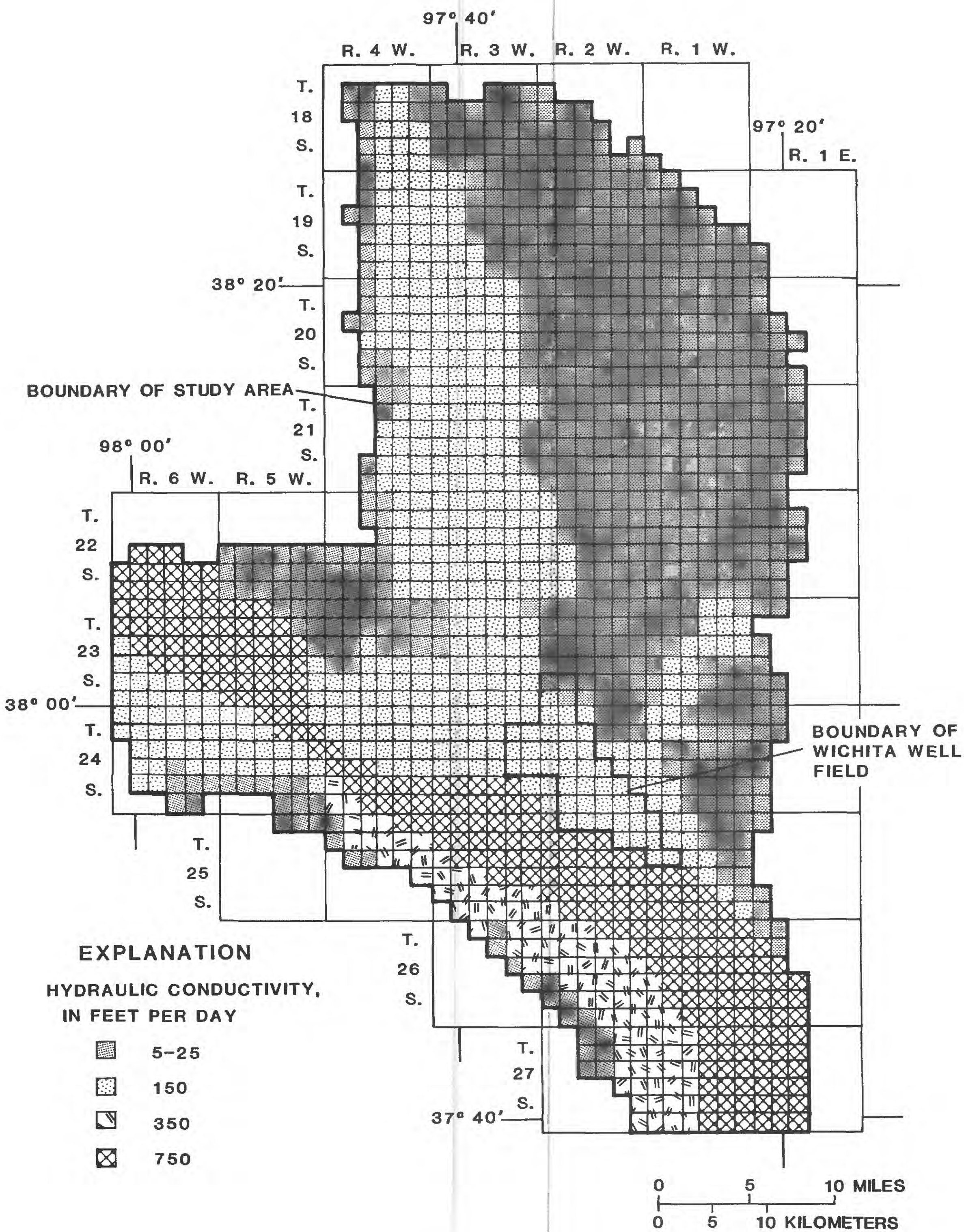


Figure 9.--Distribution of hydraulic conductivity used in the ground-water flow model for the Equus beds aquifer.

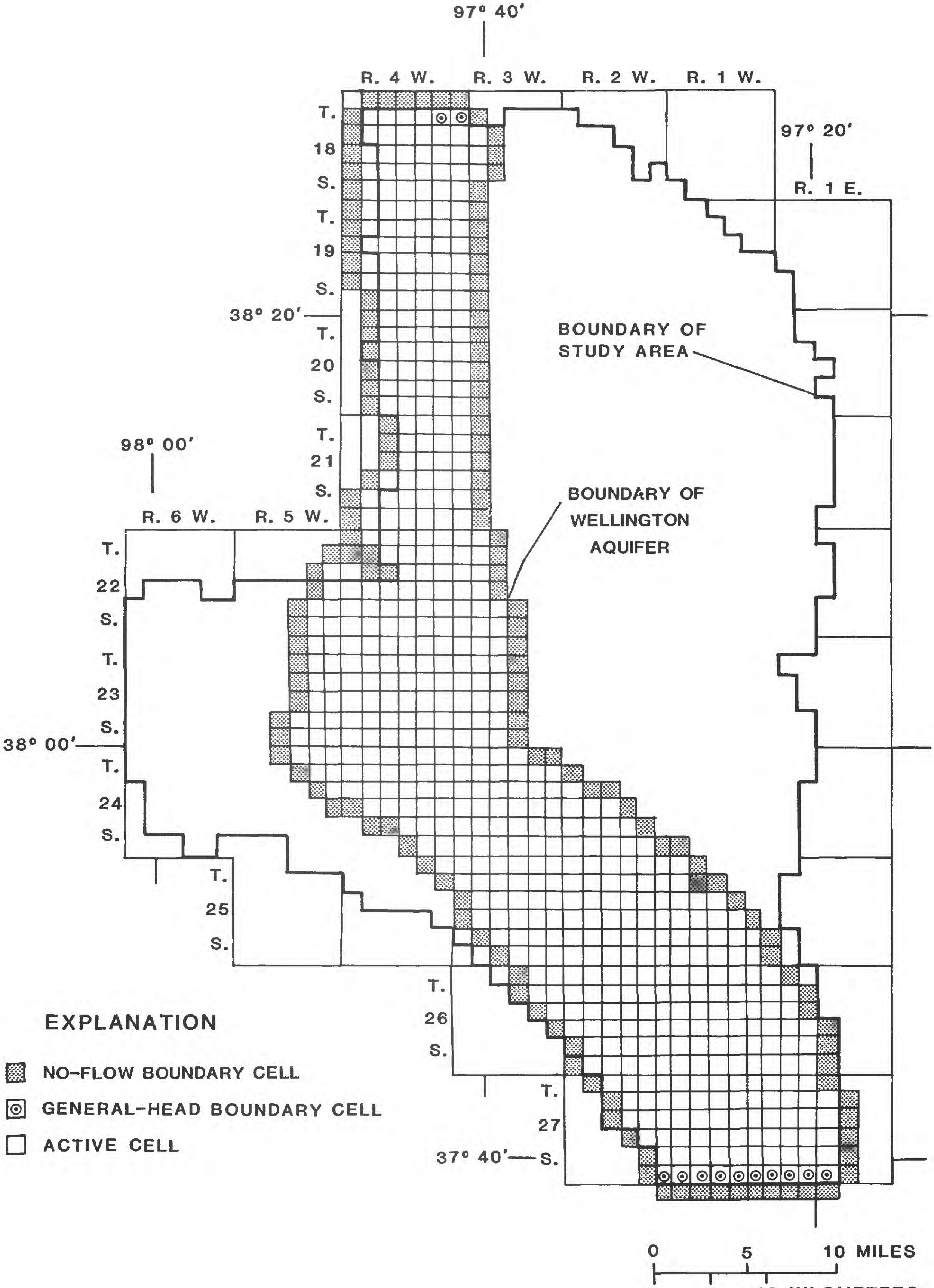


Figure 10.--Finite-difference grid, model boundary conditions, and active cells representing the Wellington aquifer in the ground-water flow model.

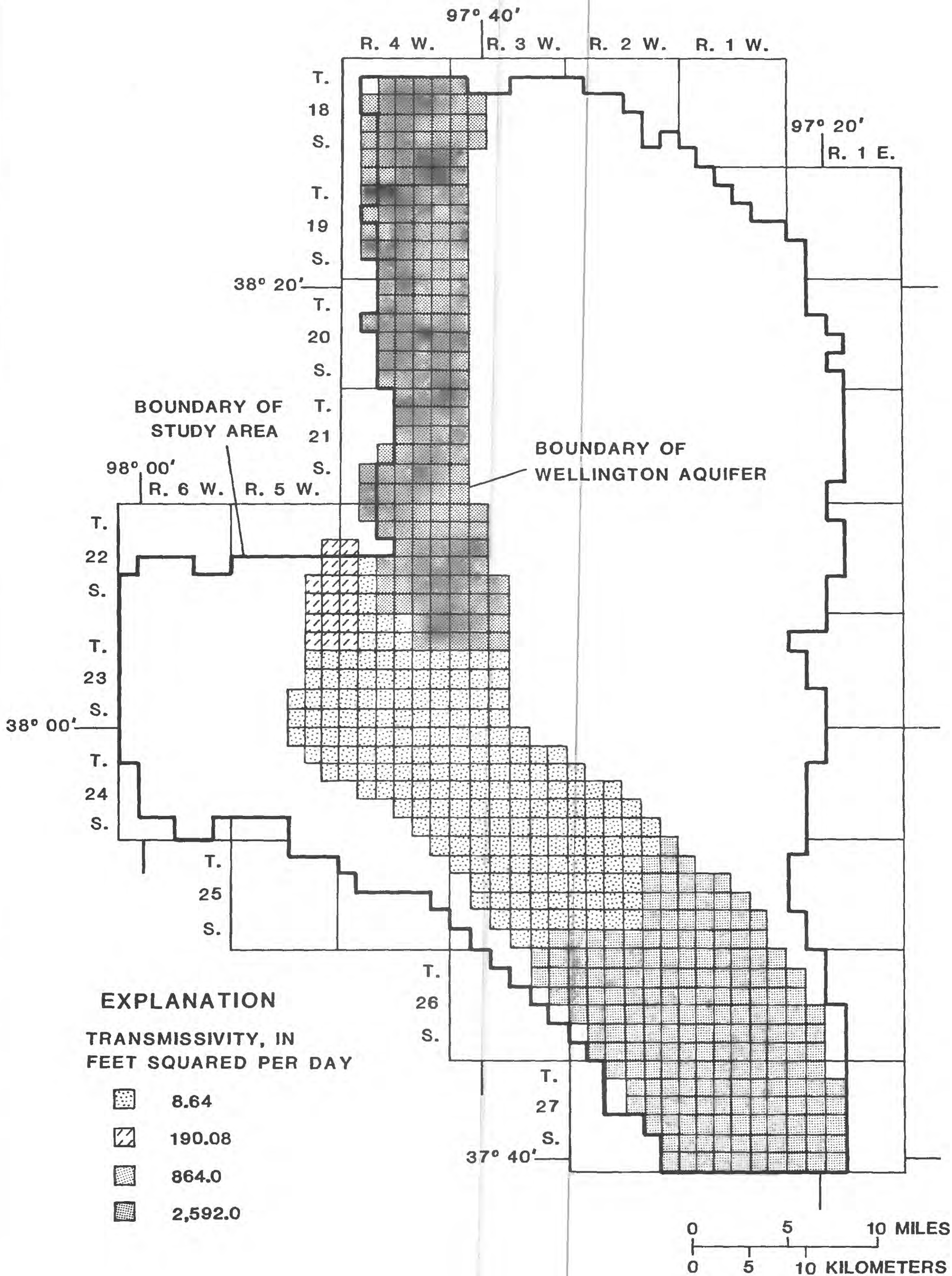


Figure 11.--Distribution of transmissivity used in the ground-water flow model for the Wellington aquifer.

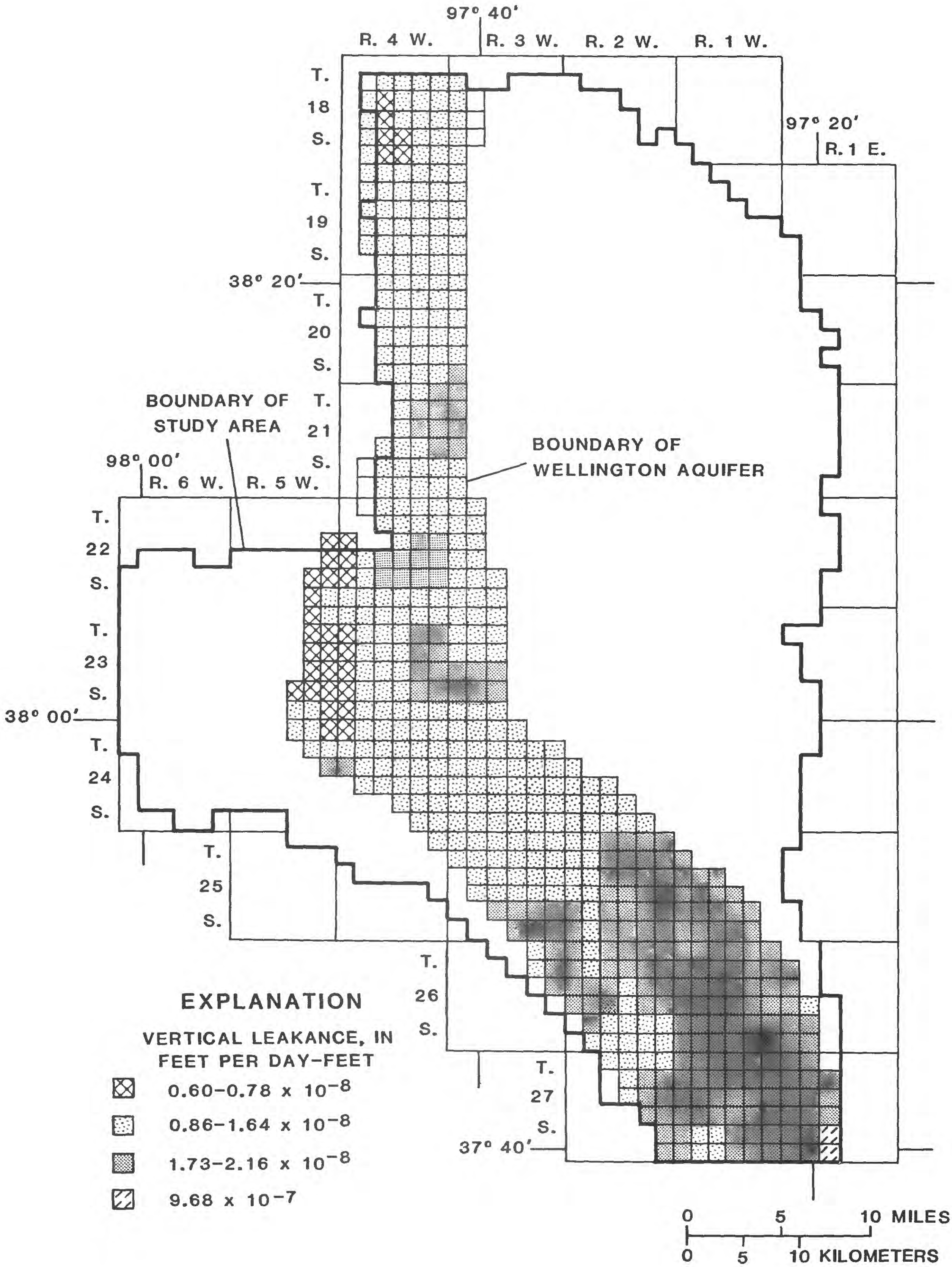


Figure 12.--Distribution of vertical leakance used in the ground-water flow model for the confining bed above the Wellington aquifer.

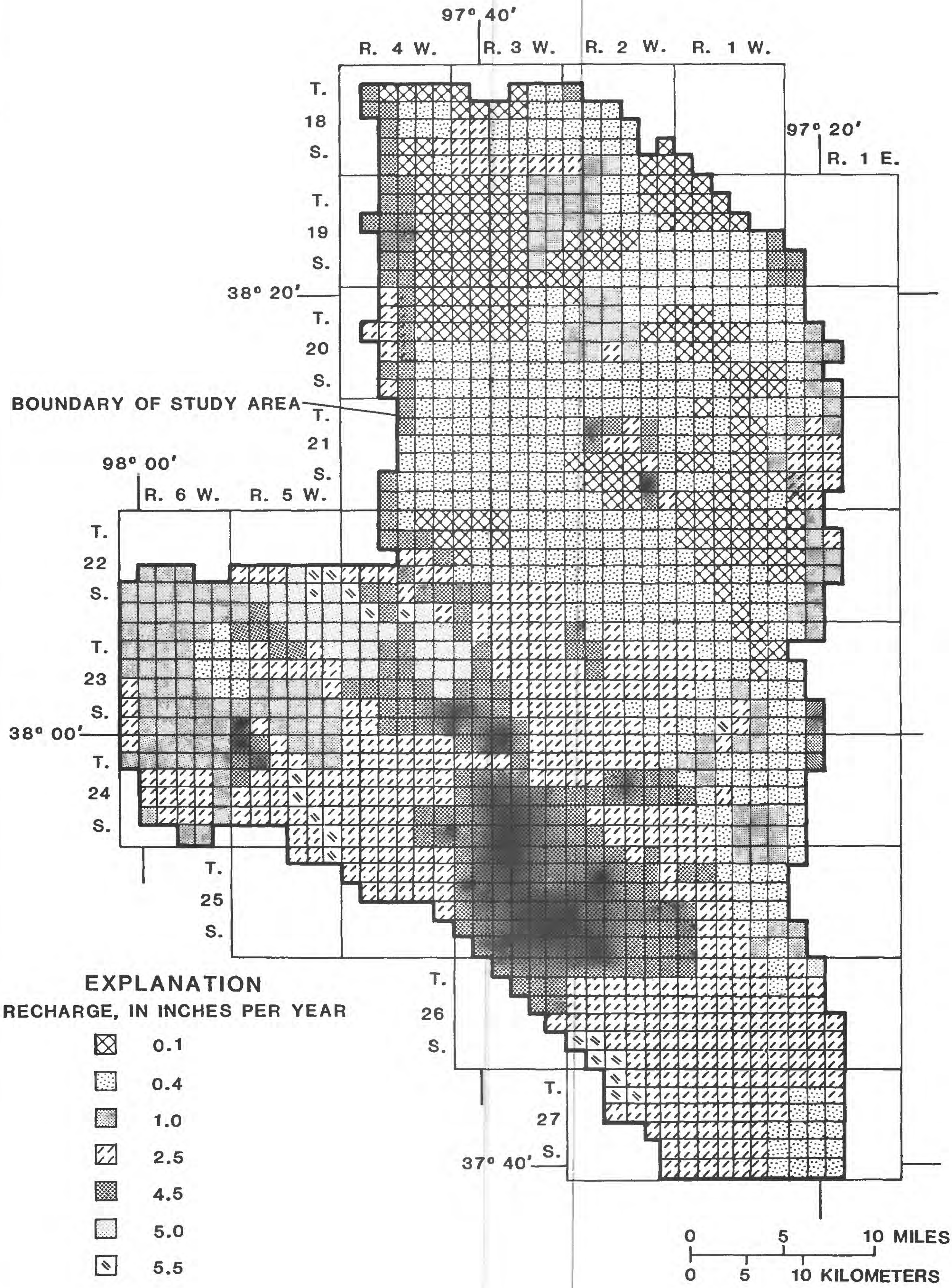


Figure 13.--Distribution of recharge to the Equus beds aquifer for initial-condition simulation in the ground-water flow model.

Transient Simulation

The object of the transient simulation was to use the flow model to reproduce a known set of hydrologic responses as recharge to and discharge from the flow system changed with time. Aquifer properties and boundary conditions described in the initial-condition simulation were used to represent the flow system in the transient simulation. The transient simulation represented conditions from 1940 to 1979. Results from the transient simulation were compared to measurements for 1971 and 1980.

Unlike the presumption of the initial-condition simulation, storage volume changed as recharge added water to the flow system or discharge removed water from the flow system in the transient simulation. A uniform specific yield of 0.15 was used to affect the storage change in the Equus beds aquifer. Reported specific yield in the aquifer ranged between 0.08 and 0.34 (Williams and Lohman, 1949, p. 94-99). A uniform storage coefficient of 0.0001 was assumed to effect the storage change in the Wellington aquifer.

Ground-Water Withdrawal By Wells

Withdrawal by wells was the major factor that affected storage changes in the ground-water flow system by 1980. Annual withdrawal from the Equus beds aquifer by wells was tabulated from records on file with the Kansas State Board of Agriculture, Division of Water Resources. Distributions of withdrawal from the Equus beds aquifer for five stress periods between 1940 and 1979 are shown in figure 14. Each stress period corresponded to a period of uniform trend in withdrawal from the aquifer. Withdrawal by each well was averaged for the length of the stress period. Averages were summed for each section to represent withdrawal from the corresponding cell. Each stress period represented a subdivision of the total transient-simulation period in which withdrawal by wells and recharge were considered constant.

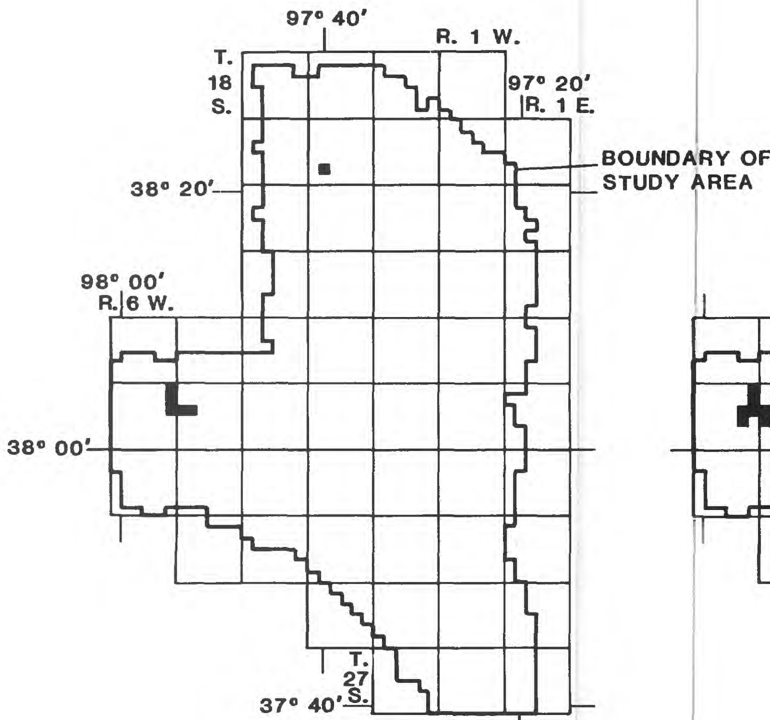
Recharge

Recharge changed for each stress period as the product of the recharge factor and average preciptation during the stress period. (Derivation of the recharge factor was described earlier.) Average precipitation for each stress period was computed from annual precipitation data available for the nine precipitation stations in or near the study area (fig. 1). The maximum, minimum, and average annual precipitation from these stations are shown in figure 15.

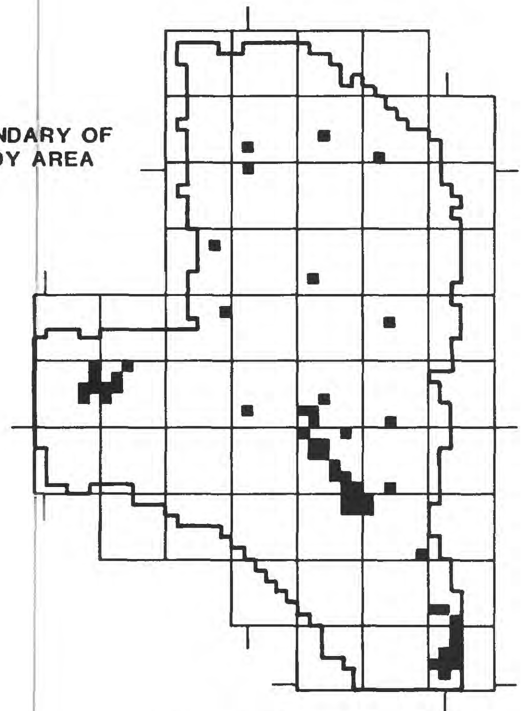
Fluid injection by wells added water to the Wellington aquifer. The distribution and total injection rate of 102 acre-ft/yr were those used in a previous digital model analysis of the aquifer (Gogel, 1981). The distribution of injection into the aquifer is shown in figure 14.

Discussion of Flow-Model Results

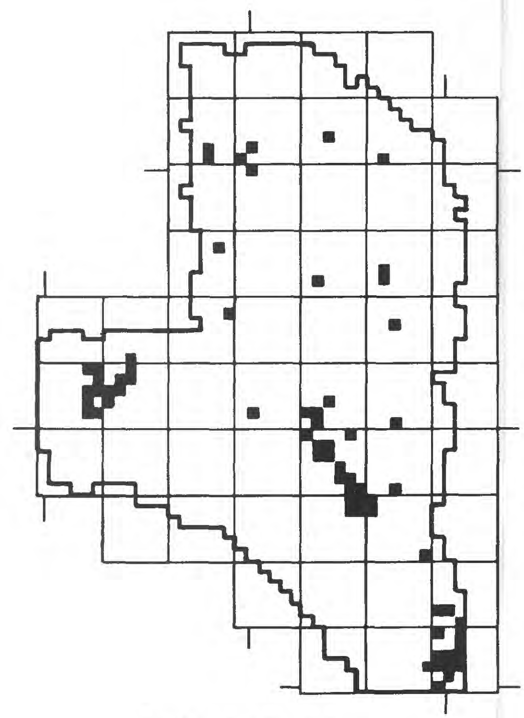
Aquifer characteristics used to define the flow model were tested by comparing maps showing hydraulic heads simulated by the flow model to maps



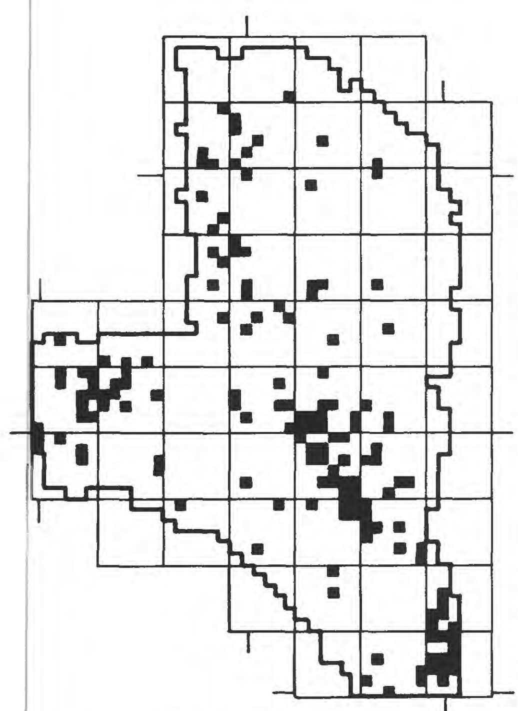
(A) Before 1940



(B) 1940–52



(C) 1953–58



(D) 1959–63

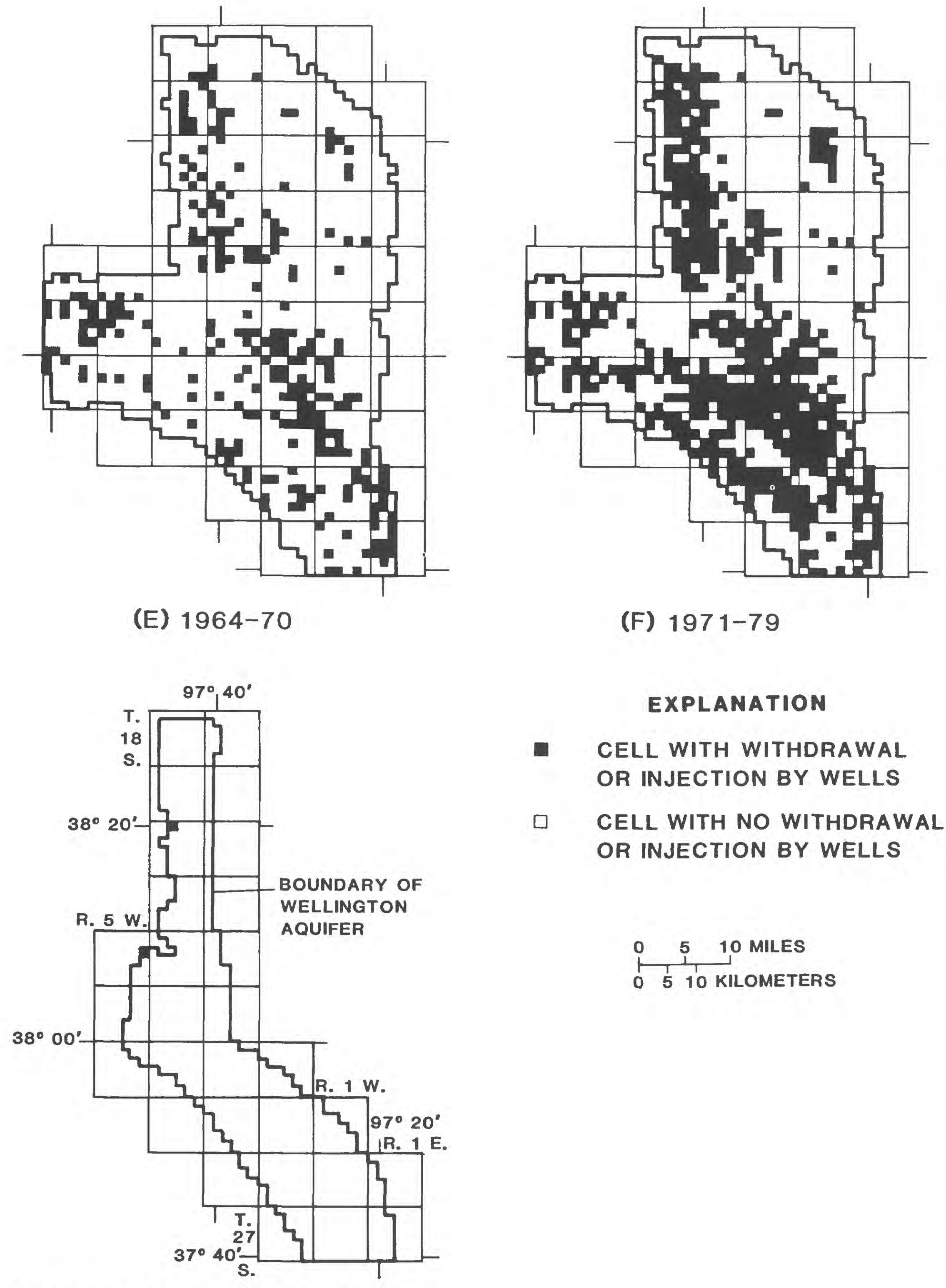


Figure 14.-- Distribution of cells used in the ground-water flow model to represent withdrawal by wells completed in the Equus beds aquifer (data from Kansas State Board of Agriculture, Division of Water Resources, Topeka, Kansas) and injection by wells completed in the Wellington aquifer (data from Gogel, 1981) for selected stress periods.

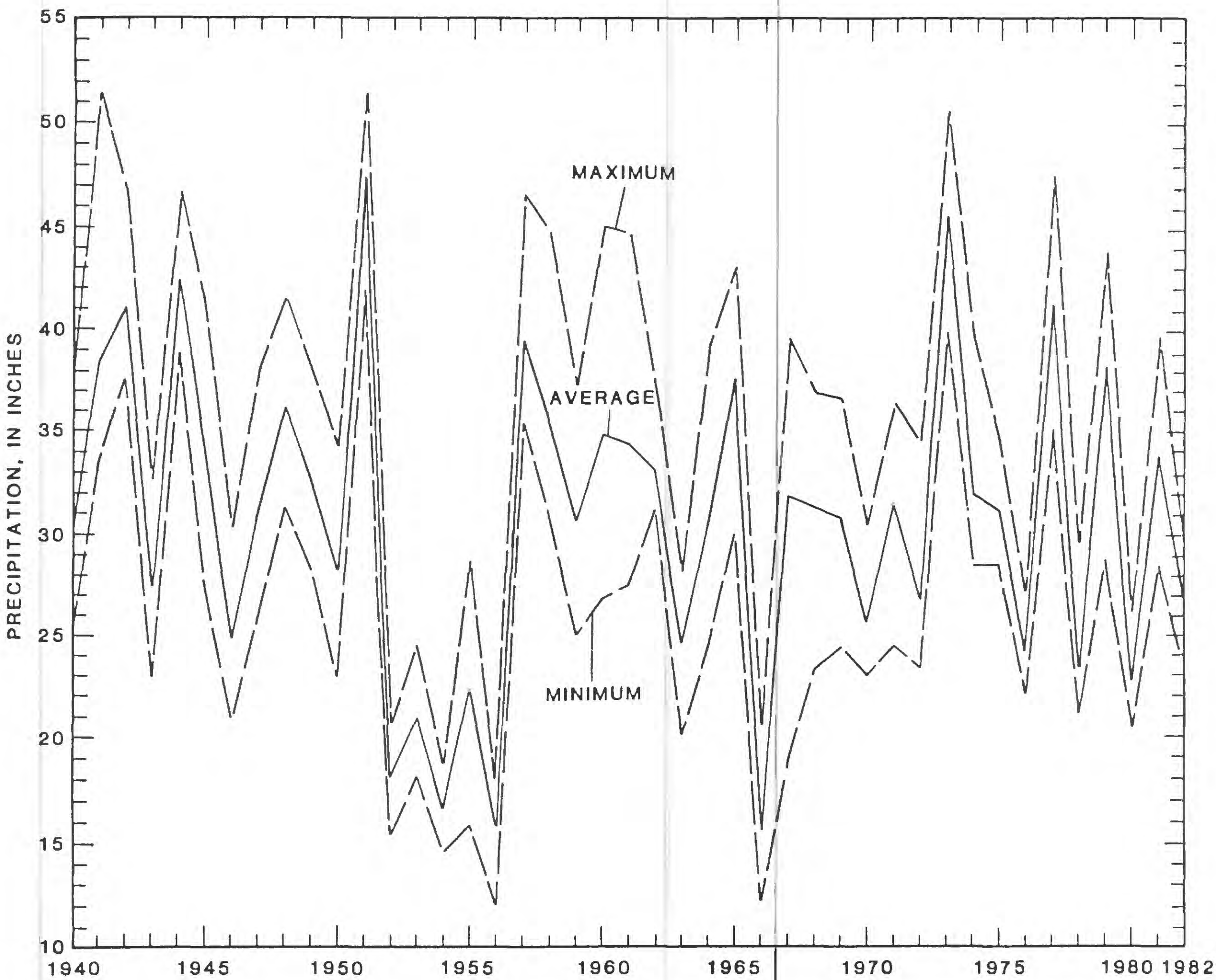


Figure 15.-- Maximum, minimum, and average annual precipitation from precipitation stations in or near study area, 1940-82.

of hydraulic head from previous reports or to maps prepared using water-level measurements from wells. Water-level measurements used in preparing maps were obtained from data on file with the U.S. Geological Survey (Lawrence, Kansas).

The water-table map for the Equus beds aquifer from the initial-condition simulation (fig. 16) was compared to the water-table map during 1940-44 modified from Williams and Lohman (1949, pl. 1). The water-table map for the Equus beds aquifer from the transient simulation at the end of 1970 (fig. 17) was compared to a map prepared from water-level measurements for January 1971. A water-table map for the transient simulation at the end of 1979 (fig. 18) was compared to a map prepared from water-level measurements for January 1980.

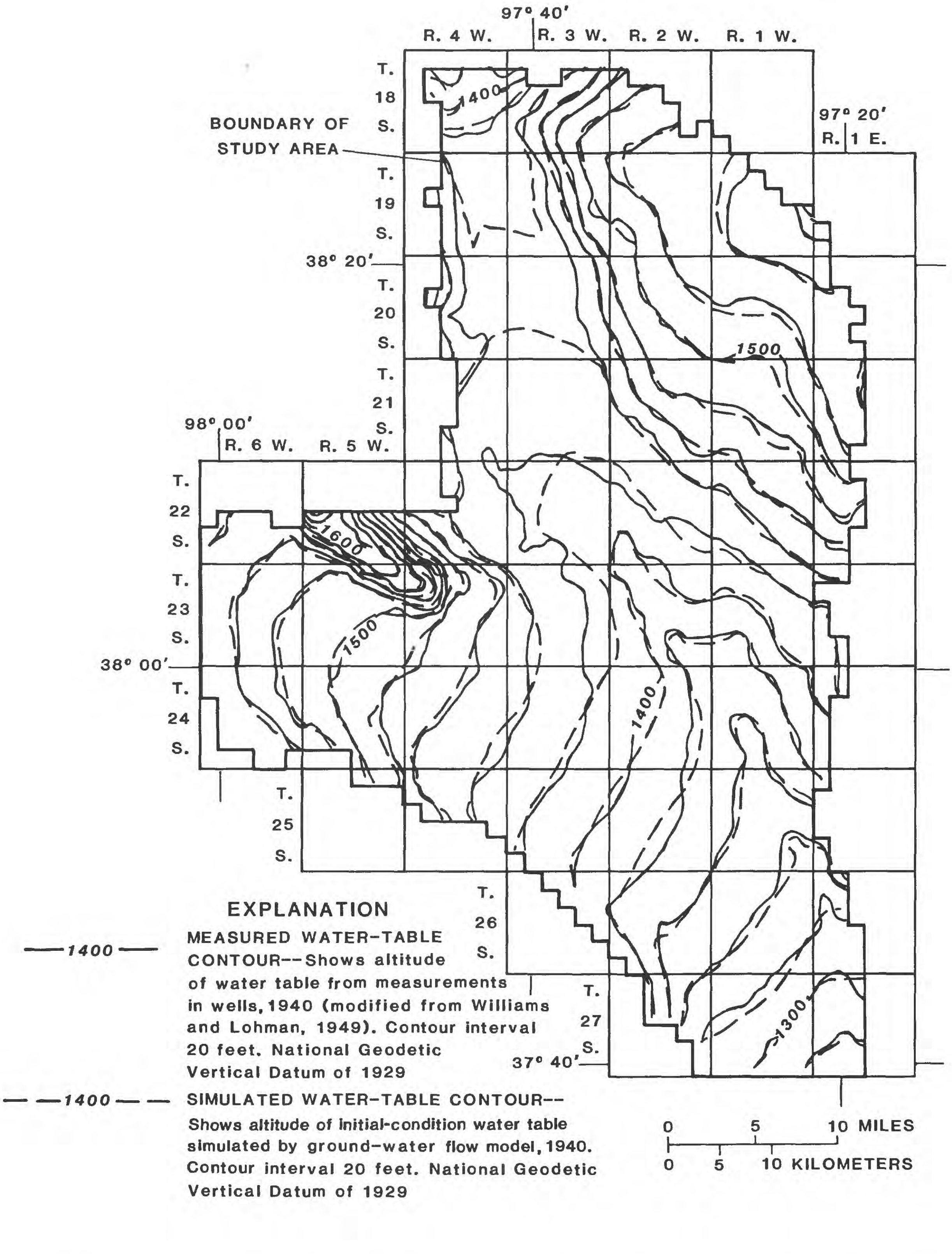


Figure 16.--Measured and simulated water table in the Equus beds aquifer for initial condition, 1940.

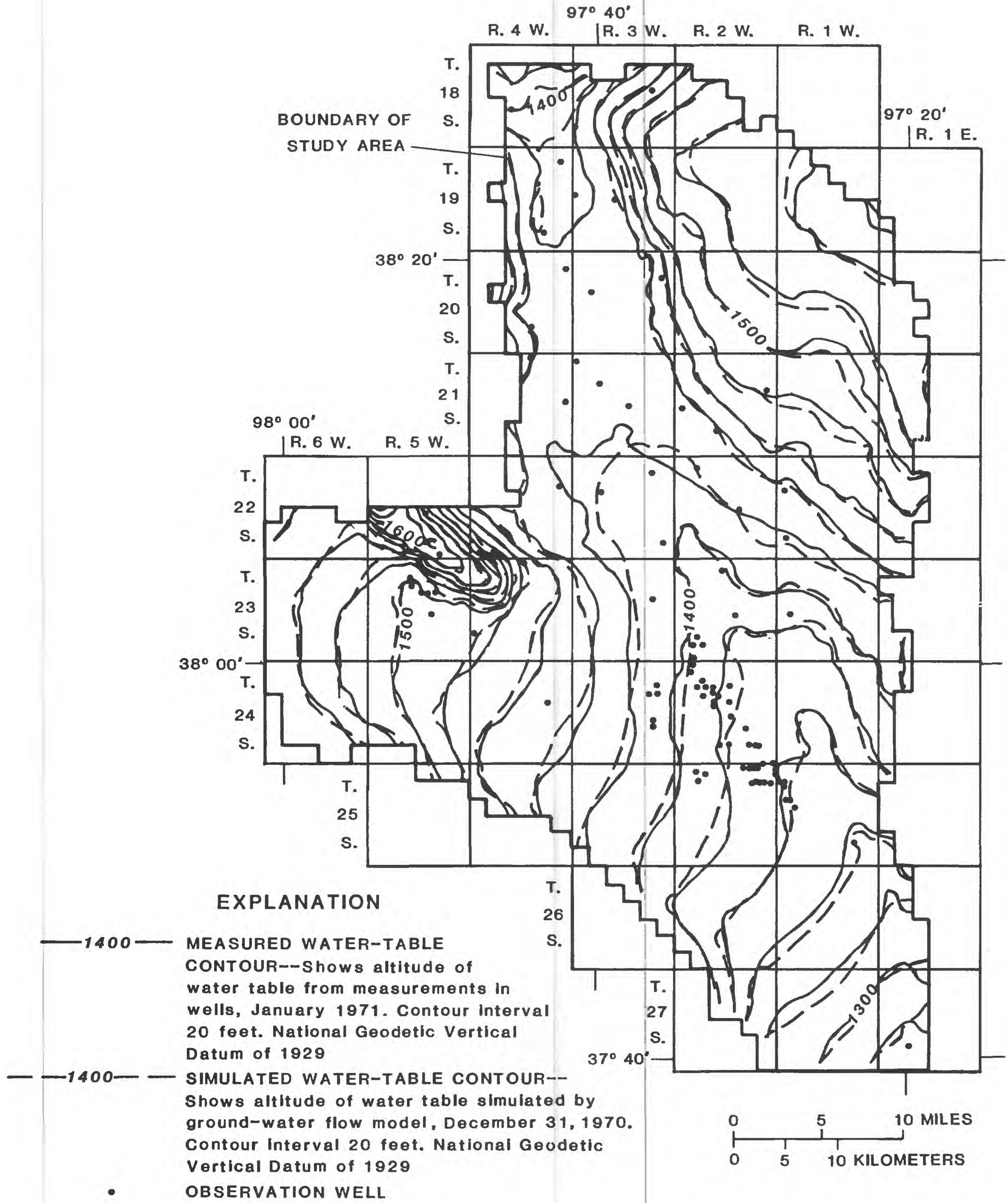


Figure 17.--Measured and simulated water table in the Equus beds aquifer, January 1971.

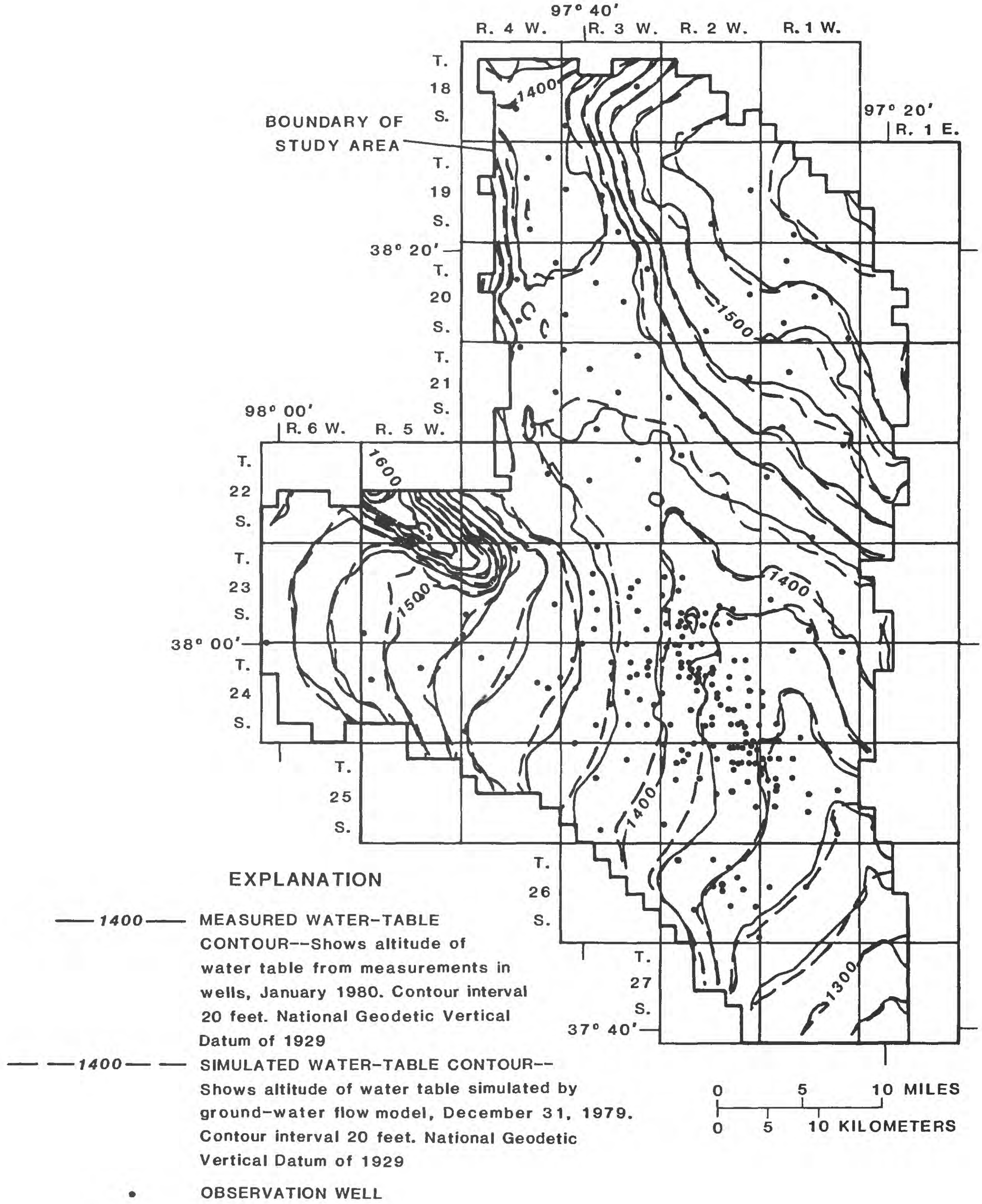


Figure 18.--Measured and simulated water table in the Equus beds aquifer, January 1980.

The potentiometric-surface map for the Wellington aquifer from the initial-condition simulation (fig. 19) was compared to the potentiometric-surface map for 1977 (Gogel, 1981, pl. 4). Water-level measurements for the Wellington aquifer were available only for 1977. Water levels simulated for the Wellington aquifer at the end of 1970 and 1979 did not differ appreciably from the initial-condition simulation and are not illustrated in this report.

If the flow model were to perfectly represent the ground-water flow system, the contours shown in figures 16-19 would coincide exactly. Because this was not the case, the differences in the contours can be ascribed to errors. Possible sources of errors include, but are not limited to, variations in aquifer characteristics unaccounted for in the model, errors involved in preparing the maps based on water-level measurements, or model assumptions made about the aquifer system that were not reasonable. Whatever the source of the differences, results from the model compared reasonably well to measurements from the Equus beds aquifer for 1971 and 1980, as well as to the measurements available for the Wellington aquifer.

The hydraulic heads in the Equus beds aquifer were higher than the hydraulic heads in the Wellington aquifer at the end of the transient simulation for all cells in the model grid. However, local field conditions could show the opposite. Dissolution, subsidence, and collapse associated with the Wellington aquifer could affect the confining bed between the Wellington and the Equus beds aquifers. The hydraulic heads for the Wellington aquifer were based on nine water-level measurements available in the study area (Gogel, 1981, pl. 4). Until more observation wells are available to provide additional information on the nature of the confining bed between the Wellington and Equus beds aquifers and on hydraulic heads in the Wellington aquifer, it is reasonable to assume that a localized potential for upward flow into the Equus beds aquifer could exist.

In addition to the comparisons between measured and simulated water levels, simulated streamflow gains were related to the computed low-flow gains for the Arkansas and Little Arkansas Rivers shown in figure 7. The following table shows simulated gains to the Arkansas River and the Little Arkansas River:

Year	Simulated gain (cubic feet per second)
1940	129
1952	105
1958	75
1963	76
1970	56
1979	35

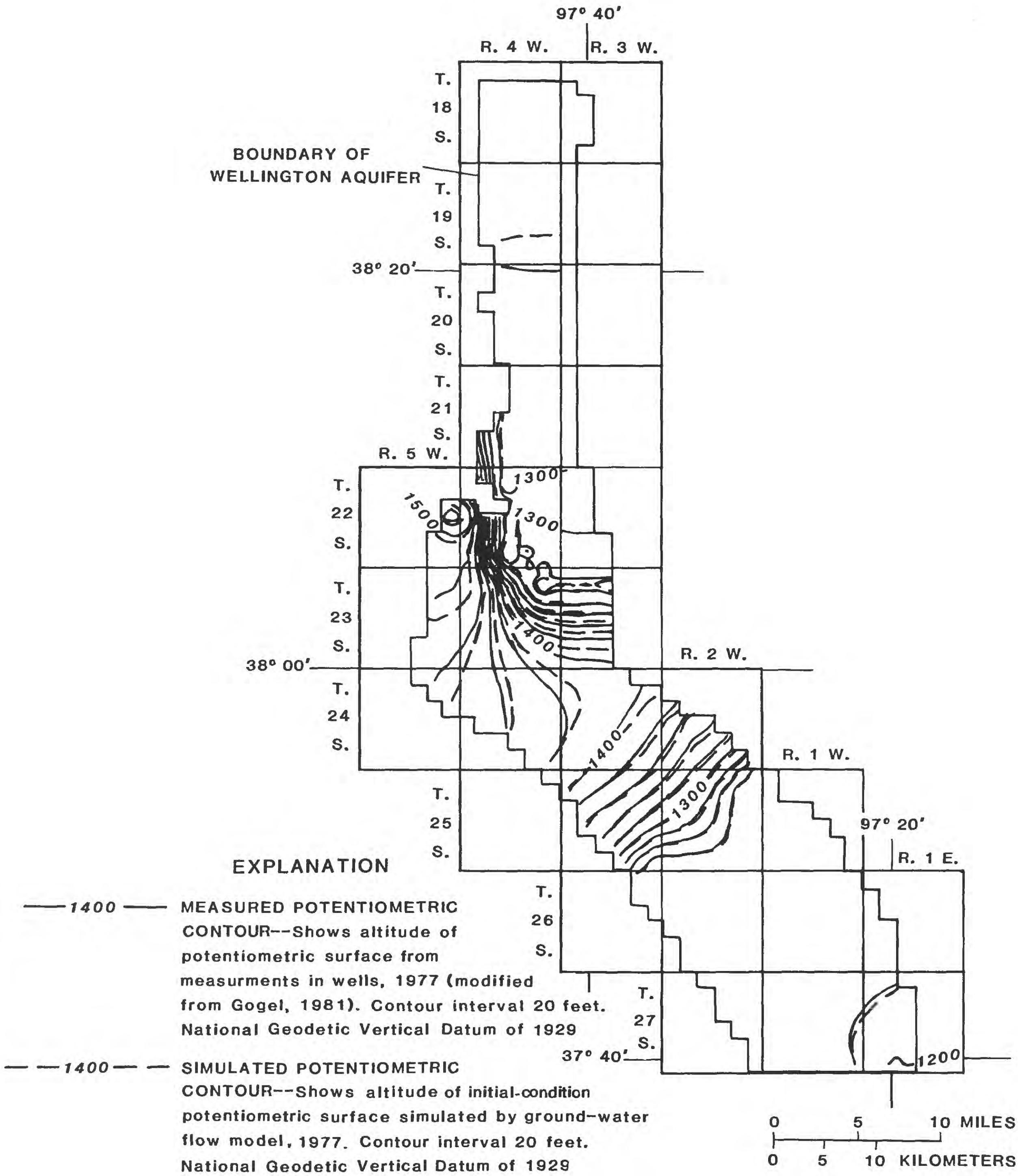


Figure 19.--Measured and simulated potentiometric surface in the Wellington aquifer for initial condition.

A direct comparison in time between measured low-flow gains and simulated gains was not considered practical due to the general nature of the available data. The analysis of the relationship between ground and surface water in the model could be improved if streamflow were to be measured independently for the Arkansas River and the Little Arkansas River.

For the creeks represented in the model, simulated gains were compared to corresponding streamflow measurements cited earlier in the section "Relationships between Ground and Surface Water." Simulated gains at the end of 1970 and 1979 for the creeks were:

Creek	Simulated gain (cubic feet per second)	
	1970	1979
Emma	0.04	0.42
Sun	3.83	1.75
West Kentucky	.54	.50
Paint	.41	.15
Sharps	1.22	.92

For the flow model, the values of some aquifer characteristics were assumed at two different levels of reliability. Boundary flow, ground-water injection and withdrawal, precipitation, the measured water-level maps, maps of aquifer geometry, and aquifer properties by Gogel (1981) for the Wellington aquifer were assumed at a high level of reliability. Hydraulic conductivity of the Equus beds aquifer, recharge factor, maximum ground-water evapotranspiration rate, and vertical hydraulic conductivity of the riverbed between streams and the Equus beds aquifer were assumed at a low level of reliability. The high reliability-level characteristics were accepted as input to the model without changes. The low reliability-level characteristics were adjusted by trial-and-error until a satisfactory comparison was achieved between measured and simulated water-level maps or streamflow gains. The values for the low reliability-level characteristics used to obtain the simulated results discussed earlier in this section are described in other sections of this report.

Water Budget

A water budget was prepared based on the calculation made by the flow model (table 1). The table compares results at the end of the 1940 initial-condition simulation and periods representing 1964-70 and 1971-79 of the transient simulation. Between 1940 and 1964-70, withdrawal by wells increased 1,630 percent. Streamflow gain decreased by 54 percent, whereas streamflow loss increased by 760 percent. Between 1964-70 and 1971-79, withdrawal by wells increased by about 42 percent. Although recharge increased by about 13 percent during this period, the rate of change of decrease in storage in the flow system was about 26 percent. The decrease in storage resulted from a decline in water-table altitudes in the Equus

Table 1.--*Simulated water budget for the Equus beds aquifer and the Wellington aquifer flow system*

[Values are given in cubic feet per second]

Aquifer	Budget term	1940 Initial-condition simulation Recharge	1940 Initial-condition simulation Discharge	1964-70 Transient simulation Recharge	1964-70 Transient simulation Discharge	1971-79 Transient simulation Recharge	1971-79 Transient simulation Discharge
Equus Beds	Lateral boundary flow	39.96	25.02	42.52	23.56	42.42	23.00
	Recharge	176.41	--	174.77	--	197.67	--
	Streamflow loss	1.96	--	14.90	--	23.37	--
	Streamflow gain	--	142.70	--	76.98	--	63.63
	Ground-water evapotranspiration	--	42.38	--	28.33	--	25.34
	Withdrawal by wells	--	7.94	--	129.46	--	184.36
	Leakage to Wellington aquifer 1/	--	.24	--	.26	--	.25
	Change in storage	0	0	27.34	.91	35.96	2.77
Wellington	Lateral boundary flow	--	.38	--	.40	--	.39
	Leakage from Equus beds aquifer 1/	.25	--	.26	--	.25	--
	Injection by wells	.14	--	.14	--	.14	--
Totals		218.72	218.66	259.93	259.90	299.81	299.74
Net		.07	--	.03	--	.07	--

beds aquifer. Lower water-table altitudes in the aquifer in relation to water-level altitudes in streams caused discharge from the Equus beds aquifer to streams to decrease by about 21 percent while loss from streams to the aquifer increased by about 57 percent. Declining water levels also resulted in a small decrease in ground-water evapotranspiration and boundary flow.

Sensitivity Analysis

There is no unique solution offered by a ground-water flow model. Identical results can be obtained using a variety of combinations for aquifer properties to represent the flow system. However, only a small number of combinations can be considered realistic based on the known limits of these properties. The final combination of aquifer properties used in the transient simulation was assumed to represent the flow system realistically, as well as mathematically. However, there is always a degree of uncertainty about the final choice of aquifer properties. Therefore, to determine the relative effect that individual aquifer properties exerted on the solution to the flow model, a sensitivity analysis was conducted.

Recharge, hydraulic conductivity, and specific yield were assumed to be the properties that had the greatest effect on flow in the Equus beds aquifer. These properties were varied independently to ascertain their relative importance on the solution calculated by the flow model. The results of six simulations were compared to results from the accepted transient simulation. For each simulation, one of the properties was increased or decreased uniformly for the entire model grid by one-half of the value used in the accepted transient simulation without changing the other properties.

A comparison of measured and simulated water levels (fig. 20) indicates that the combination of aquifer properties used in the transient simulation came closest to reproducing measured water levels in the aquifer than any of the other combinations of aquifer properties used in the sensitivity analysis. The cell used in figure 20 represents an area in the Wichita well field where water levels have declined by about 30 feet between 1940 and 1980, which is the greatest measured water-level decline in the Equus beds aquifer.

The frequency distribution (fig. 21) indicates the range of differences between measured and simulated water levels for all active cells in the Equus beds aquifer at the end of the accepted transient simulation for 1979. The flow model simulated water levels within plus-or-minus 10 feet of measured water levels for 99 percent of the model area; plus-or-minus 5 feet for 87 percent of the area; and to the nearest foot for about 14 percent of the area. The differences between measured and simulated water levels were normally distributed (fig. 21).

The mean difference between measured and simulated water levels and the standard deviation of the mean summarizes the information shown in figure 21 and compares results among the accepted transient and sensitiv-

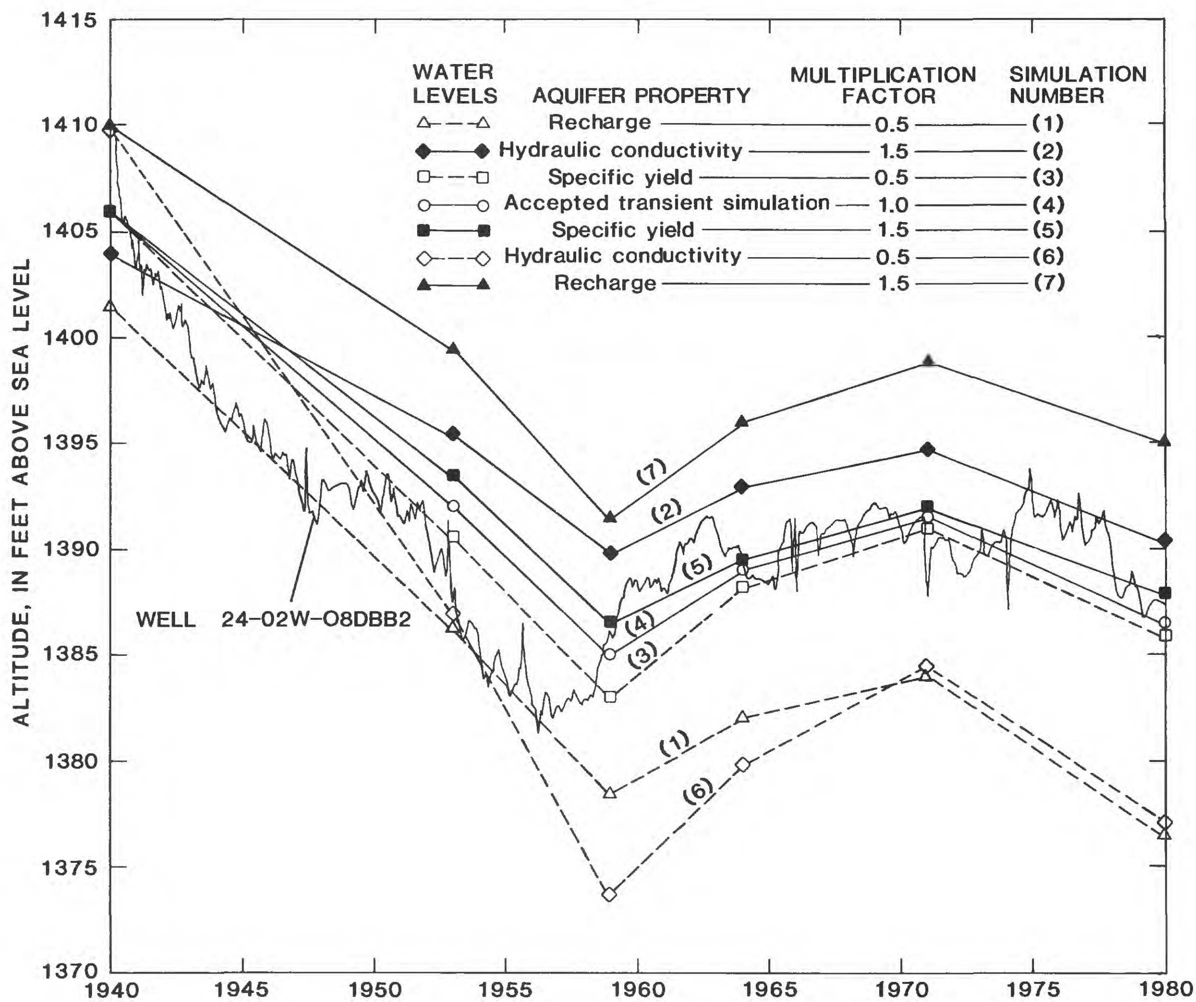


Figure 20.--Sensitivity of flow-model results to variations in selected aquifer properties at a single cell corresponding to well 24-02W-08DBB2.

ity simulations for 1979 (fig. 22). Results from the accepted transient simulation yielded the smallest mean difference and standard deviation between measured and simulated water levels compared to the sensitivity simulations. The sensitivity analysis indicated that model results for the Equus beds aquifer were affected mostly by the choice for recharge. The choice for specific yield had the least effect on model results.

Transmissivity of the Wellington aquifer and vertical leakance of the confining bed above the Wellington aquifer were assumed to be the two aquifer properties that had the greatest effect on flow in the Wellington

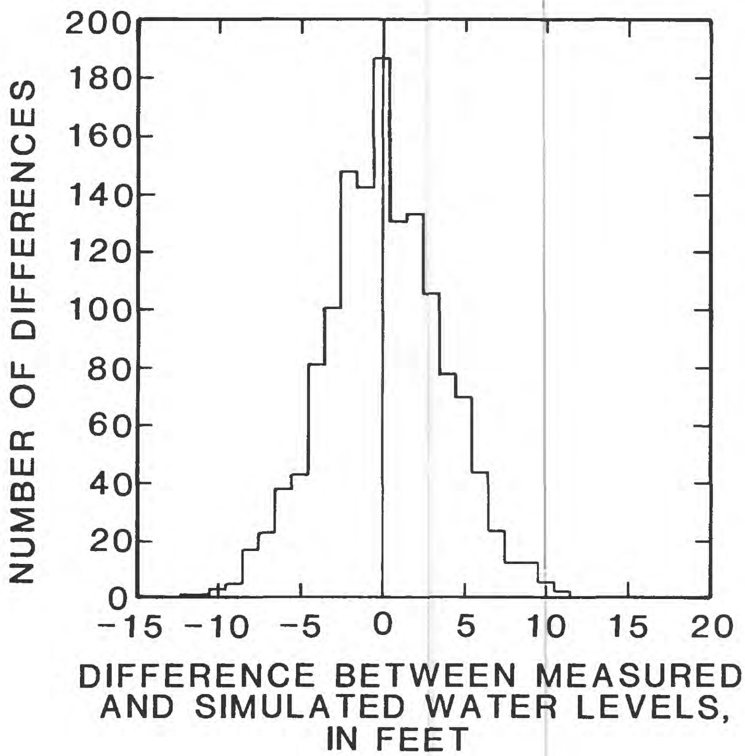


Figure 21.--Frequency distribution of difference between measured and simulated water levels for all active cells in the Equus beds aquifer at end of accepted transient simulation, 1979.

aquifer. Four sensitivity simulations were conducted by increasing or decreasing one aquifer property uniformly for the entire model grid by one-half of the value used in the accepted transient simulation without changing the other property. Results among the sensitivity simulations were compared to results of the accepted transient simulation.

The frequency distribution (fig. 23) of the differences between measured and simulated water levels for all active cells in the Wellington aquifer at the end of the transient simulation for 1979 indicates that the differences are not normally distributed. Modes at 1, 3, and 5 feet probably are the result of inadequate definition of the potentiometric surface or of aquifer properties in T. 22 S. and T. 23 S., R. 3 W., R. 4 W., and R. 5 W. where the greatest differences between measured and simulated water levels occur (fig. 19). Differences for the rest of the aquifer shown in figure 19 are minimal. The flow model simulated water levels within 8 feet of measured water levels for the entire aquifer and within 5 feet of measured water levels for about 90 percent of the aquifer.

The median, 25th, and 75th percentiles of the differences between measured and simulated water levels summarize the characteristics of the distribution shown in figure 23 for the accepted transient simulation and compare the summarized results of each of the sensitivity simulations

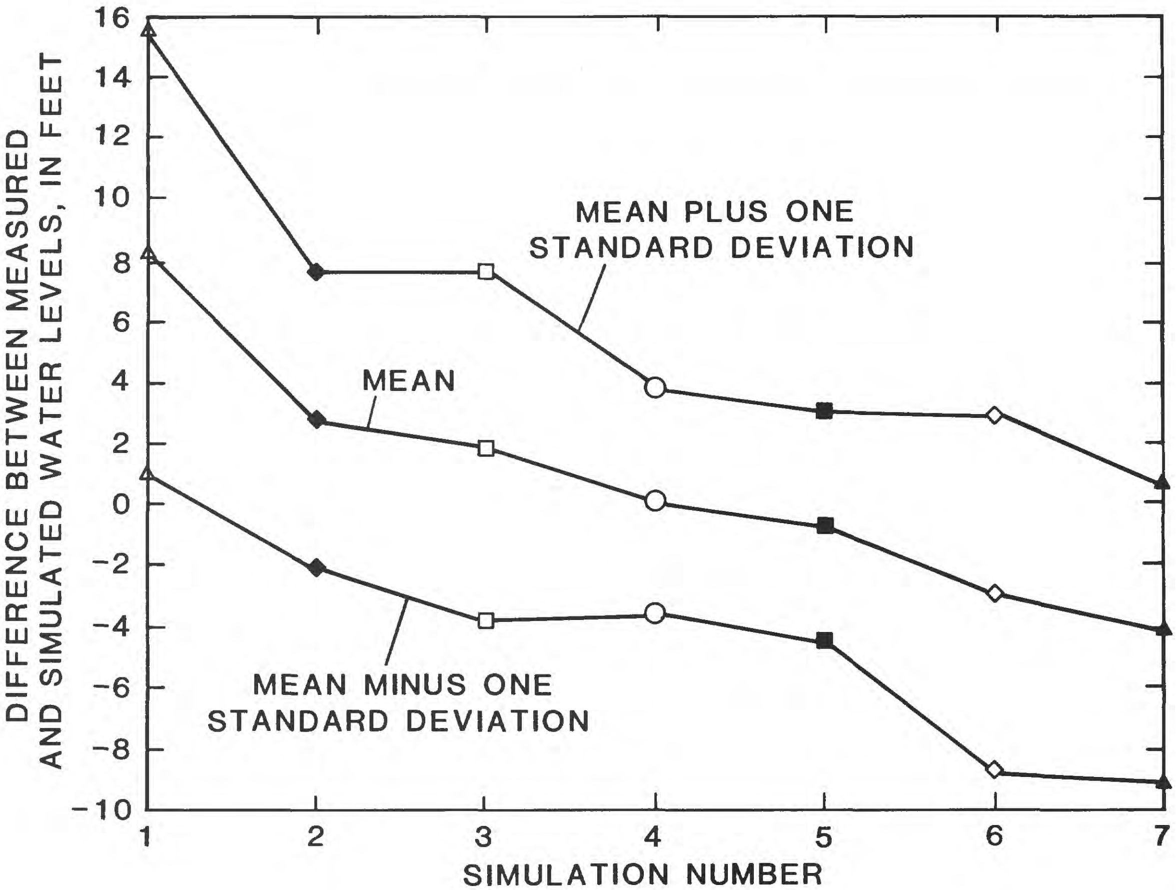


Figure 22.--Mean and standard deviations of differences between measured and simulated water levels for all active cells in the Equus beds aquifer at end of accepted transient simulation and each of six sensitivity simulations, 1979.

(fig. 24). Sensitivity simulations illustrated in figures 24 and 25 are identified as:

Simulation number	Aquifer property	Percent of accepted transient simulation
1	Transmissivity	-50
2	Vertical leakance	+50
3	Accepted transient simulation	--
4	Vertical leakance	-50
5	Transmissivity	+50

Results from the accepted transient simulation (3) yielded the smallest median and smallest range between the 25th and 75th percentiles for the difference between measured and simulated water levels compared to the other sensitivity simulations.

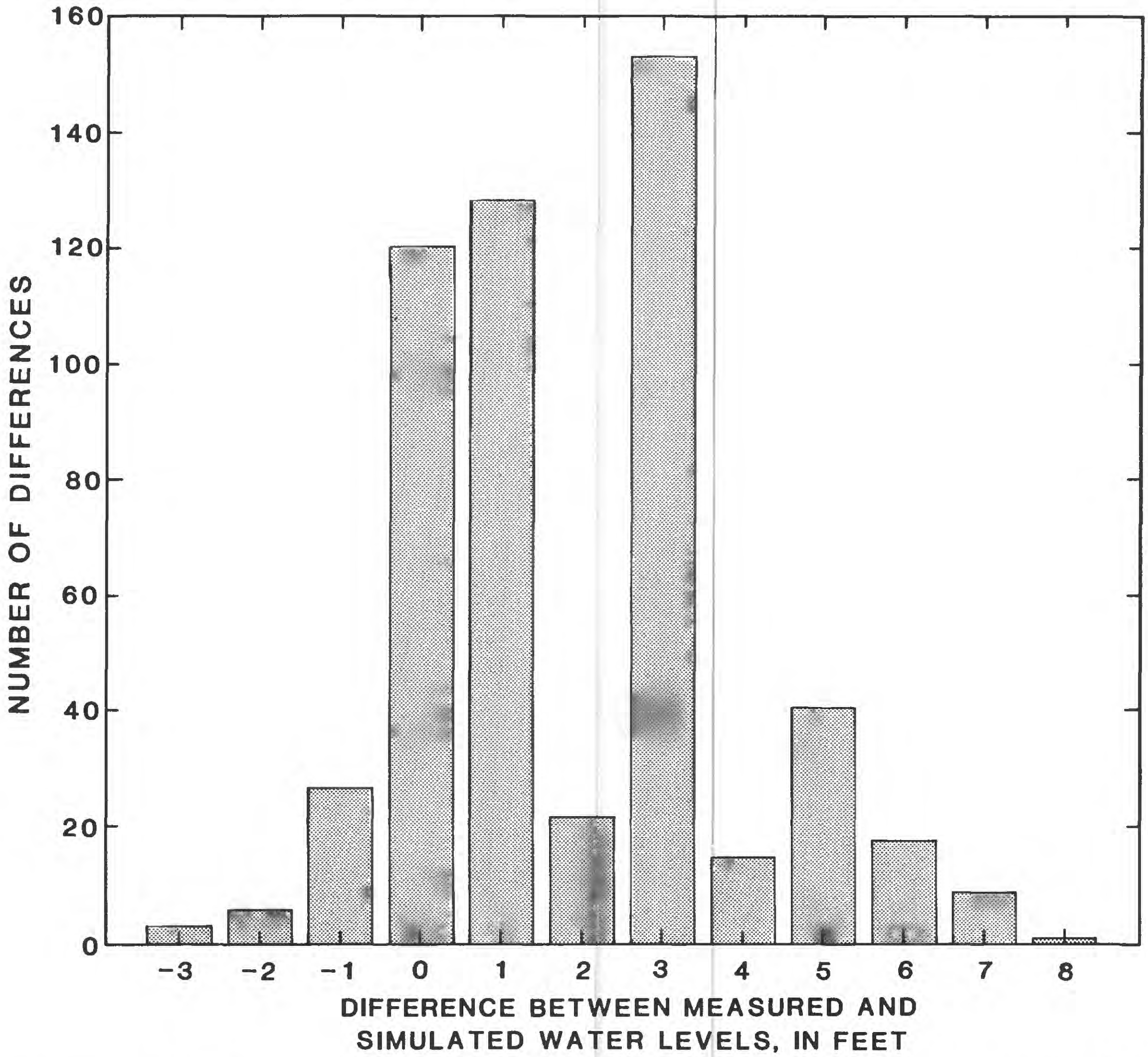


Figure 23.--Frequency distribution of difference between measured and simulated water levels for all active cells in the Wellington aquifer at end of accepted transient simulation, 1979.

The sensitivity of leakage through the confining bed above the Wellington aquifer (calculated by the model) to changes in aquifer properties is shown in figure 25. The sensitivity analysis indicated that differences between measured and simulated water levels in the Wellington aquifer and leakage to the Wellington aquifer were affected more by a change in transmissivity of the Wellington aquifer than by a change in the vertical leakance of the confining bed above the Wellington aquifer.

Flow-Model Projections

The flow model was used to project changes in saturated thickness in the Equus beds aquifer, in combined streamflow gains in the Little Arkansas and Arkansas Rivers, and in the relationship between water levels in the Equus beds aquifer and the Wellington aquifer from 1980 to 2020. The 40-year projection simulation was divided into eight 5-year stress periods. Recharge for each stress period was applied in the same manner that was used in the transient simulation. Annual average precipitation for 1940-80

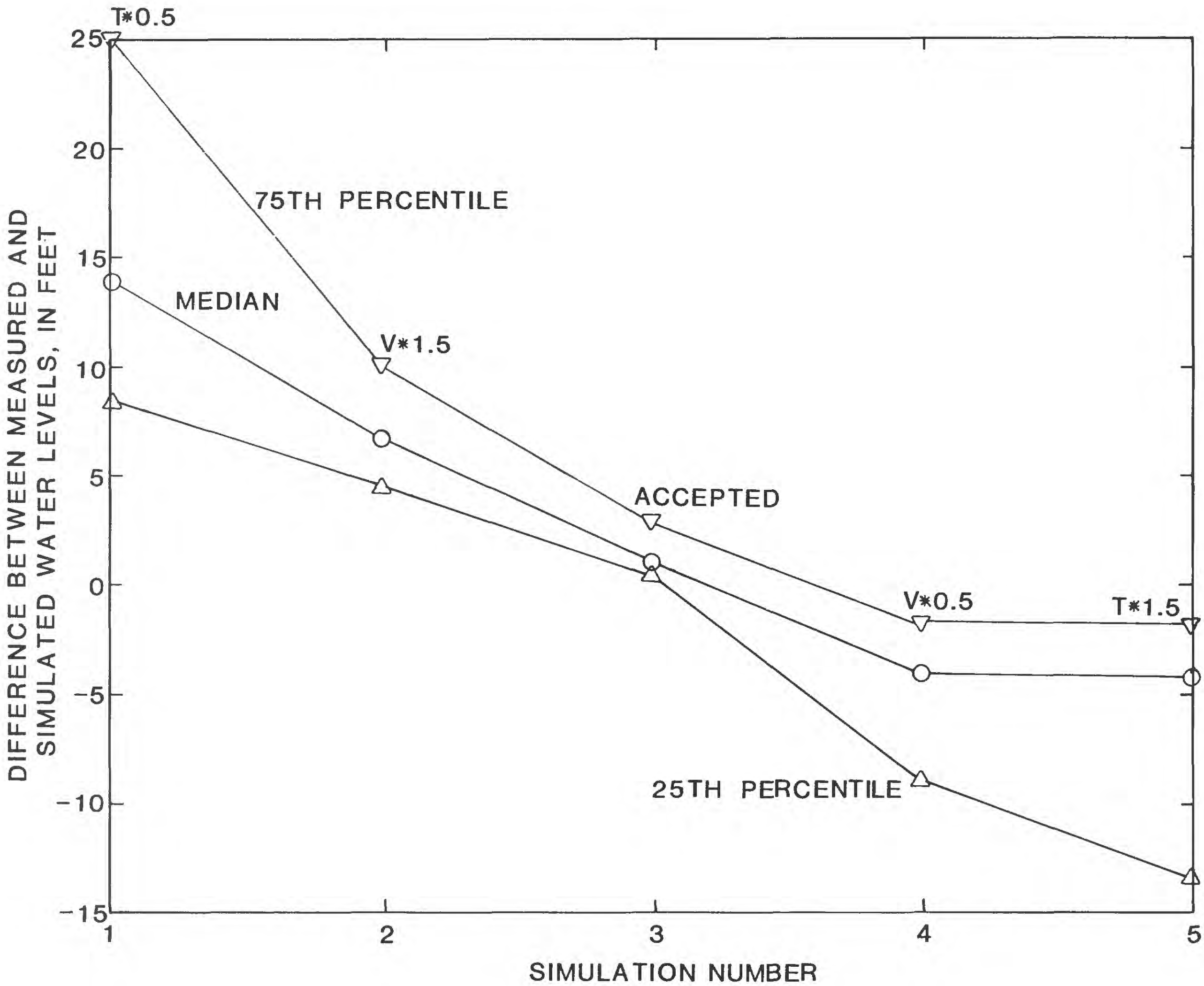


Figure 24.--Median, 25th, and 75th percentiles of differences between measured and simulated water levels for all active cells in the Wellington aquifer at end of accepted transient simulation and each of four sensitivity simulations, 1979.

was used for the projections. This period of precipitation was extrapolated to 2020 based on historical precipitation patterns. Water-table altitudes simulated by the flow model for 1980 were used as initial altitudes for the projection simulations. All other aquifer properties were the same as those used in the transient simulation.

Five pumping alternatives were represented by varying the 1971-79 ground-water-withdrawal rates used in the transient simulation as follows: (1) Rates decreased by one-half, (2) rates continued at 1971-79 level, (3) rates increased by one-third, (4) rates increased by two-thirds, and (5) rates doubled. In addition, the effect of a hypothetical well field was included for each of the alternatives. The location of the hypothetical well field is shown in figure 26. Withdrawal rates applied to cells in the hypothetical well field ranged from 30 to 1,600 acre-ft/yr for alternative 2. Rates were applied to cells in the hypothetical well field based on rates applied to surrounding cells in the transient simulation for 1971-79.

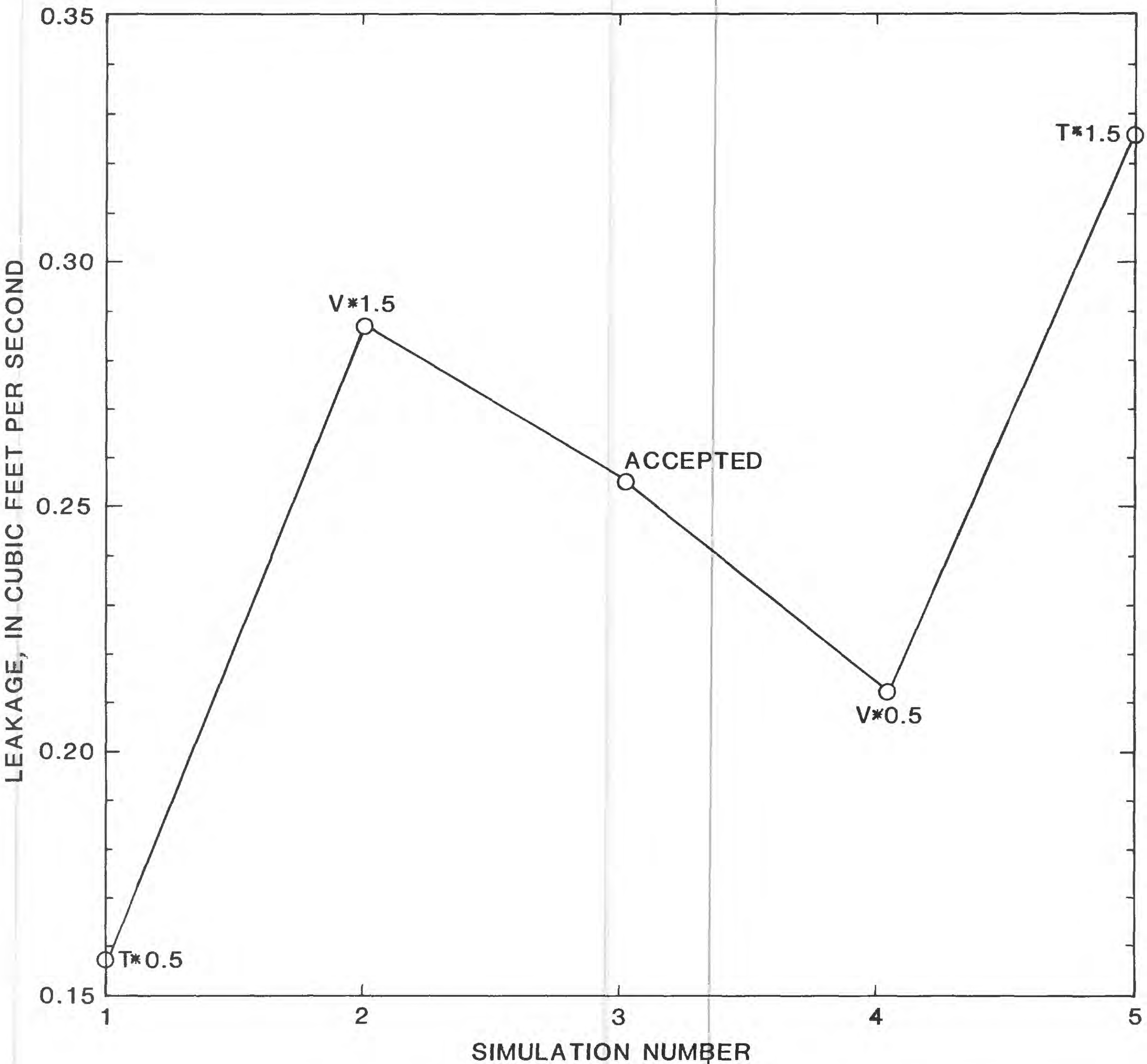


Figure 25.--Leakage through confining bed above the Wellington aquifer computed by the flow model at end of accepted transient simulation and each of four sensitivity simulations, 1979.

Rates for all cells were kept constant during the entire simulation for each of the five alternatives. The aquifer was allowed to dewater in cells during a simulation. Once a cell was dewatered, the cell became inactive for the remainder of the simulation.

The effects of the five pumping alternatives on saturated thickness by 2020 are summarized for selected cells in figure 27A. The cells selected for this figure are located in areas of special interest (cells A through D) or in areas of intensive development by wells (cells E through G). Cell A is near the city of McPherson; B, in the hypothetical well field; and C, north of the Little Arkansas River from the Wichita well field. Cells D through D' contain the reach of the Little Arkansas River north of the Wichita well field. Cells E and G are in or near areas of irrigation use. Cell F is in the Wichita well field.

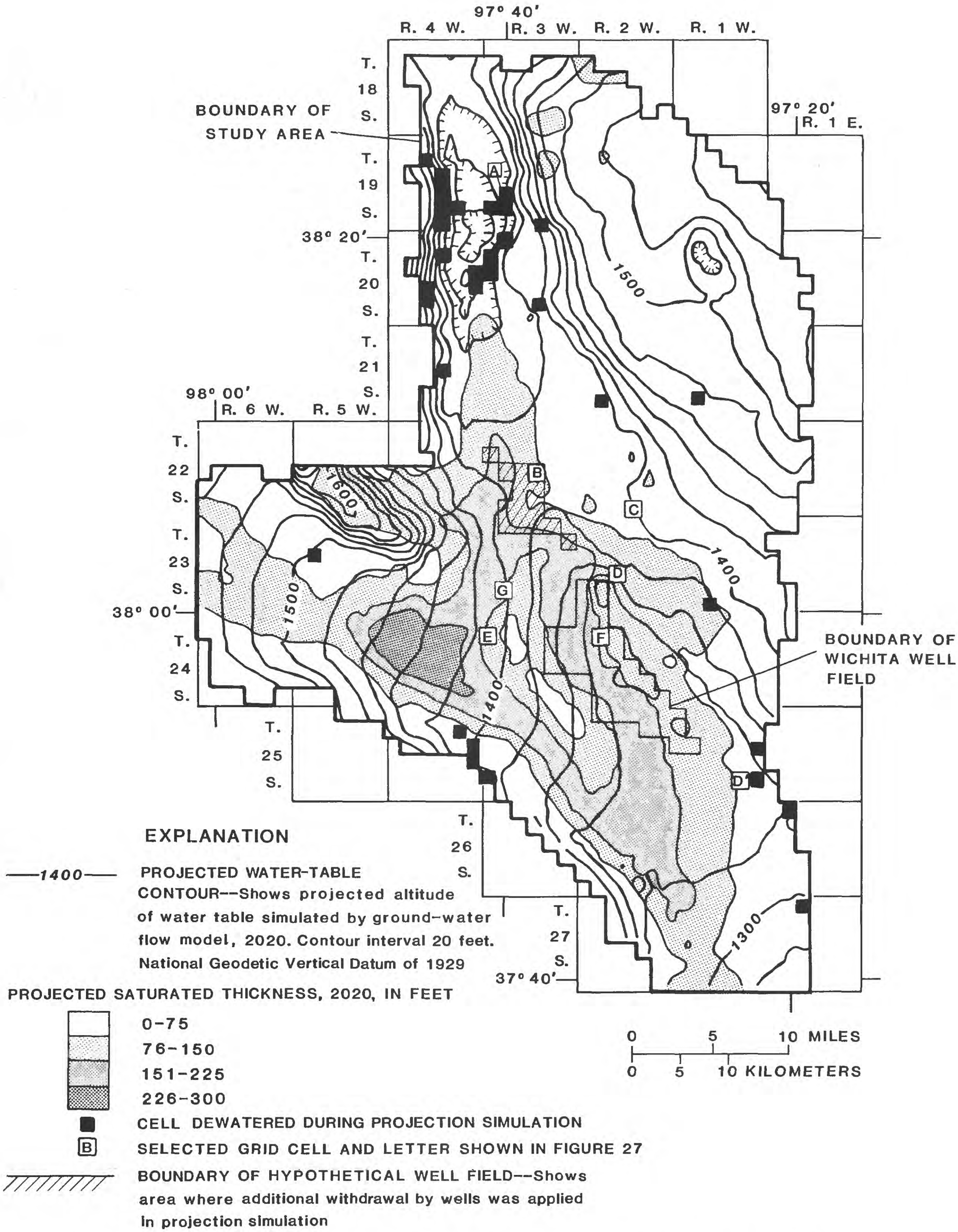


Figure 26.--Projected water-table configuration and saturated thickness in the Equus beds aquifer, 2020, using ground-water-withdrawal rates double those for 1971-79.

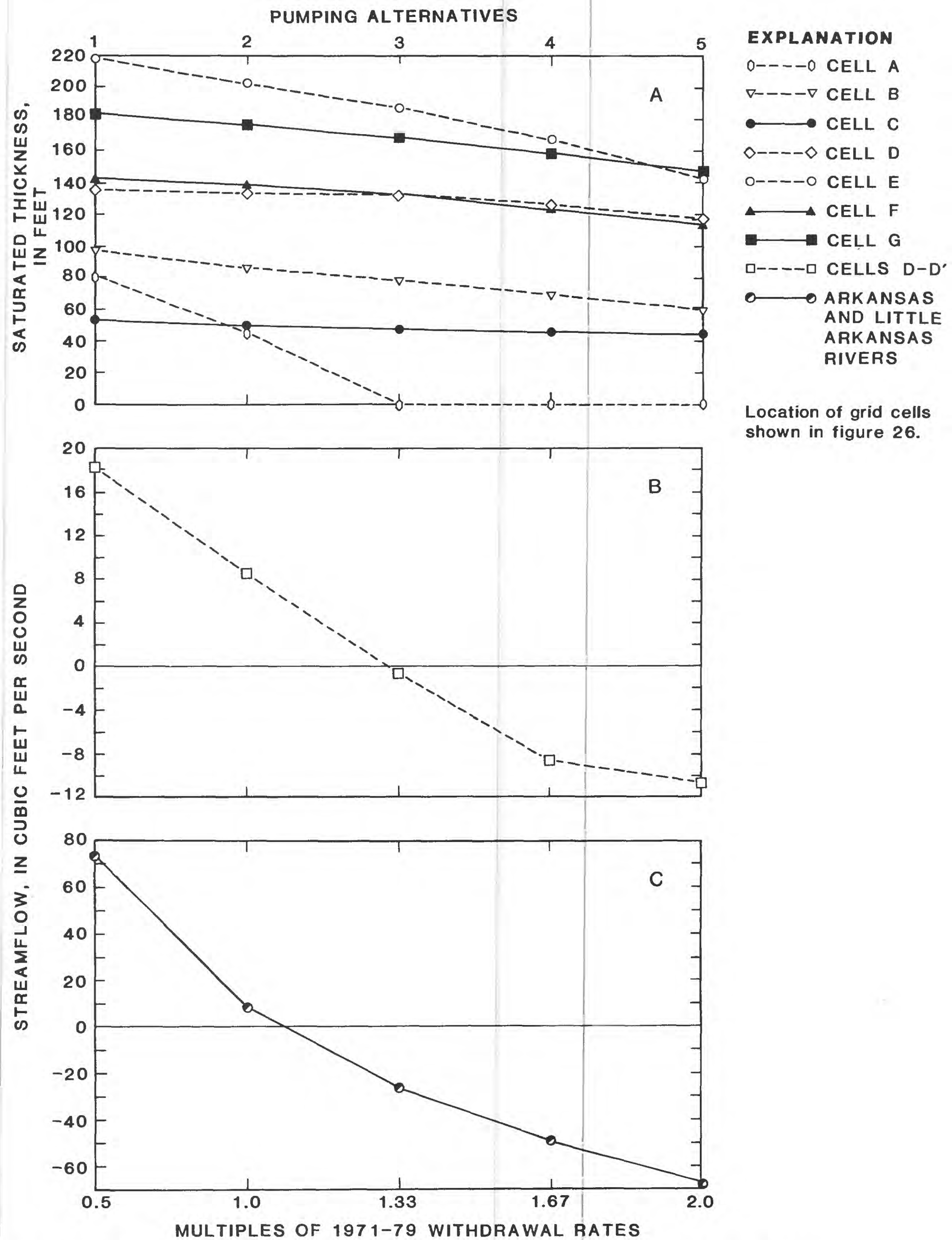


Figure 27.--Effects of alternative ground-water-withdrawal rates on projected (A) saturated thickness for selected grid cells, (B) streamflow for grid cells D-D', and (C) streamflow in Arkansas and Little Arkansas Rivers, 2020. [Negative streamflow indicates loss from stream(s).]

As withdrawal rates were increased, the saturated thickness at all of the selected cells was projected to decrease. The greatest saturated thickness in 2020 was projected for alternative 1 (one-half 1971-79 withdrawal rates) at all the selected cells. Results for alternative 2 (continued 1971-79 withdrawal rates) indicated a decrease in saturated thickness of about 40 feet compared to alternative 1 for cell A near McPherson; a decrease of about 10 feet was projected for cell B in the hypothetical well field; and about 15 feet of decrease was projected for cell E located between the Wichita municipal well field and the Arkansas River. For alternatives 3-5, cell A was projected to dewater the aquifer.

The smallest projected difference between alternatives 1 and 5 was about 5 feet for cell C, north of the Little Arkansas River from the Wichita municipal well field. The small difference indicated that withdrawal by wells in the municipal well field at the five projected rates will have little effect on saturated thickness north of the Little Arkansas River in the vicinity of cell C by 2020. Differences of about 20 to 40 feet of saturated-thickness decline were projected between alternatives 1 and 5 for cell B, located in the hypothetical well field; cell F, located in the Wichita municipal well field; and cell G, located near Burrton. The moderate projected decline in saturated thickness for this area could be the result of greater recharge in proportion to withdrawal in these areas compared to the greater declines projected for cells A and E. The water levels in the Equus beds aquifer were higher than water levels in the Wellington aquifer, indicating downward flow in all cells in the model grid at the end of the simulation for every alternative.

The effects of pumping alternatives on projected streamflow gains in a reach of the Little Arkansas River and in the Arkansas and Little Arkansas Rivers in the model area for 2020 are shown in figures 27B and 27C. The projection simulations presumed that water always was available to losing streams. Projected streamflow losses (negative streamflow, figs. 27B and 27C) must be offset by surface-water supplies that enter the streams from across the boundary of the study area. Model results are invalid if this condition is not met.

The reach illustrated in figure 27B is adjacent to the Wichita well field (D-D', fig. 26). Streamflow gains were maintained for this reach when pumping alternative 2 (1971-79 withdrawal rates continued) or pumping alternative 1 (one-half the 1971-79 withdrawal rates) was used in the projection simulation. Streamflow losses were projected in this reach for greater withdrawal rates (alternatives 3-5).

The overall gain from the aquifer to the Arkansas and Little Arkansas Rivers was projected to decrease by 2020 in proportion to the increased withdrawal for all pumping alternatives except the first (fig. 27C). For alternative 1, projected streamflow gain by 2020 increased to about 75 ft^3/s from 35 ft^3/s simulated by the model for 1980. Projected streamflow gain was maintained at about 10 ft^3/s for alternative 2, but streamflow loss was projected for any of the other alternatives.

The simulated water-table and saturated thickness in the aquifer for 2020 under the conditions of pumping alternative 5 are shown in figure 26. Many cells in the aquifer have been projected to dewater in the McPherson area under the conditions of this alternative. Saturated thickness was small at the beginning of the simulation for most of the other cells in the aquifer that were projected to dewater for alternative 5. The data in figure 27 can be used in conjunction with the data in figure 26 to approximate the effects of different pumping alternatives on hydrologic conditions in different parts of the aquifer.

SOLUTE-TRANSPORT MODEL

A two-dimensional solute-transport model (Konikow and Bredehoeft, 1978) was applied to simulate the movement of chloride ions in solution from 1940 to 1979 and to project the effects of pumping alternatives on chloride-ion movement in part of the Equus beds aquifer. Because no upward flow from the Wellington aquifer was projected by the flow model, past oilfield activities in the Burrton area and mineralized water from the Arkansas River were the only sources of chloride ion represented in the model. The model is referred to as the transport model in the remainder of this report.

Theory of Mathematical Model

Convection and hydrodynamic dispersion were the two mechanisms that were considered to move chloride ions through the aquifer; however, chloride ions were considered to move primarily by convection along flow paths in the aquifer. Convective flow is related to the average seepage velocity and can be defined by the equation:

$$V_1 = -\frac{Ki_1}{n} \qquad (2)$$

where

V_1 is the average seepage velocity in the 1 direction, in feet per day;

K is hydraulic conductivity, in feet per day;

i_1 is the hydraulic gradient in the 1 direction, in feet per foot; and

n is the effective porosity of the aquifer (dimensionless).

The flow equation was solved by the transport model using the alternating-direction implicit numerical method (Konikow and Bredehoeft, 1978, p. 5).

The dispersion and transport of a nonreactive, dissolved chemical species in two dimensions is given by the equation:

$$\frac{\partial(Cb)}{\partial t} = \frac{\partial}{\partial x_1}\left(bD_{1m}\frac{\partial C}{\partial x_m}\right) - \frac{\partial}{\partial x_1}(bCV_1) - \frac{C'W}{n} \qquad 1, m = 1, 2 \qquad (3)$$

where

C is the concentration of the dissolved chemical species, in milligrams per liter;

D is the coefficient of hydrodynamic dispersion, in feet squared per day;

b is the saturated thickness of the aquifer, in feet; and

C' is the concentration of the dissolved chemical in a fluid source or sink, in milligrams per liter.

Hydrodynamic dispersion describes the spreading of solute in addition to the movement of solute attributed to convection.

The method of characteristics approximated the solution of the partial differential equation for solute transport in the model. This method represents the solute as particles and solves for the transport of particles in the aquifer in proportion to ground-water velocity. Particle transport is described by a set of characteristic curves that represents location and concentration of the particles with respect to time. For further description of this numerical method, see Konikow and Bredehoeft (1978, p. 5-11).

The assumptions that were made about the Equus beds aquifer in order to apply the transport model were given by Konikow and Bredehoeft (1978, p. 4):

1. Darcy's law is valid and hydraulic-head gradients are the only significant driving mechanism for fluid flow.

2. The porosity and hydraulic conductivity of the aquifer are constant with time, and porosity is uniform in space.

3. Gradients of fluid density, viscosity, and temperature do not affect the velocity distribution.

4. No chemical reactions occur that affect the concentration of the solute, the fluid properties, or the aquifer properties.

5. Ionic and molecular diffusion are negligible contributors to the total dispersive flux.

6. Vertical variations in head and concentration are negligible.

7. The aquifer is homogeneous and isotropic with respect to the coefficients of longitudinal and transverse dispersivity.

Modifications to Model

The model was modified to calculate changes in transmissivity as a function of changing saturated thickness during the simulation. This modification was considered necessary because the Equus beds aquifer is unconfined and water levels have declined between 1940 and 1979. Hydraulic heads computed by the modified transport model were compared to hydraulic heads from the flow model for a separate test problem that considered withdrawal and injection by wells and recharge. The calculated water budgets for the test problem were in perfect agreement, as were the hydraulic-head distributions.

Transient Simulation

Simulations made with the transport model can be considered extensions of simulations made with the transient flow model. The transport model solves for flow and generates a velocity field in order to solve for the movement of chloride ions. The transport model does not solve for flow between the aquifer and a river or ground-water evapotranspiration as does the flow model. For this reason, results from the flow model were used extensively in the development of the transport model. Similarities between the flow and transport models are described in the following sections.

The transport model was used to reproduce the movement of chloride ions from 1940-79. Stress periods used in the transport model were identical to those used in the flow model. Stress periods were subdivided into time steps in order to minimize mathematical errors associated with calculating the transport of the solute. The initial time step for each stress period was set at 1,000 seconds. Subsequent time steps were 1.25 times the preceding time step.

Boundary Conditions

The transport model was applied to part of the Equus beds aquifer simulated using the flow model (fig. 28). The grid used in the transport model contained 436 active cells, about one-third the number of cells in the upper layer of the flow model. Cell size and aquifer properties assigned to corresponding cells in both models were identical. However, in order to represent flow in the aquifer with the transport model, a different set of boundary conditions was required.

No-flow and specified-flux boundaries were used in the transport model. No-flow boundaries encircled the model area and represented the physical edge of the aquifer for about 25 percent of the perimeter of the area. Specified-flux boundaries were used to represent flow through the aquifer along the remaining perimeter of the model. Flow computed by the flow model for each specified-flux cell was added to the appropriate cell in the transport model as well withdrawal or injection for each stress period. Specified-flux cells were assigned a concentration based on the 1940 chloride-ion concentration for the cell (fig. 29).

Recharge and Discharge

Recharge and discharge in the aquifer were represented for each stress period in the simulation. A value for net recharge was assigned to each cell in the model area. Net recharge was the sum of recharge and ground-

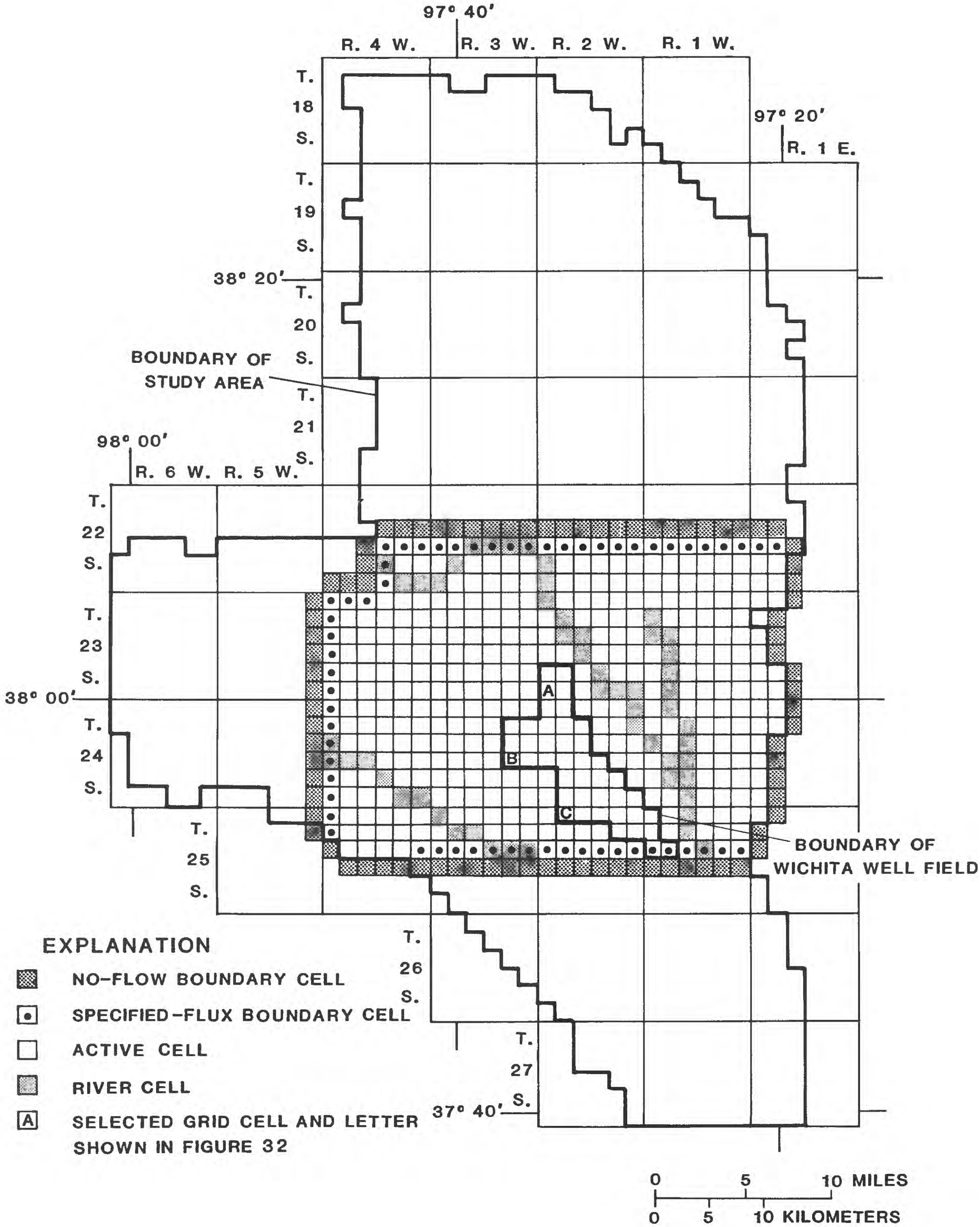


Figure 28.--Finite-difference grid, model boundary conditions, and active and river cells representing part of the Equus beds aquifer in transport model.

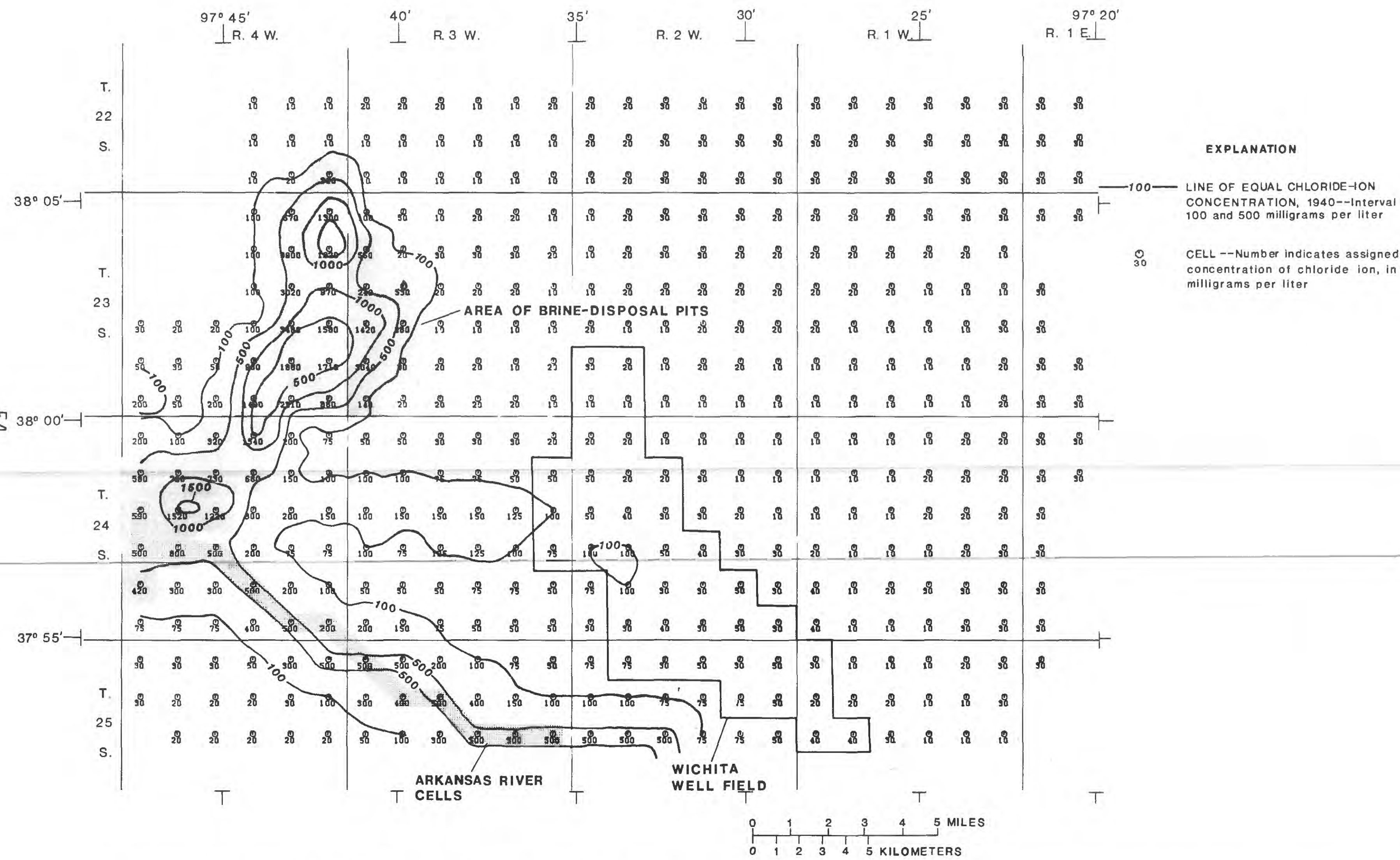


Figure 29.--Area of brine-disposal pits and chloride-ion concentrations in part of the Equus beds aquifer, 1940.

water evapotranspiration computed by the flow model for each cell. Net recharge was assigned a chloride-ion concentration of 10 mg/L. Withdrawal by wells was applied to each cell in the same manner described for the flow model.

Flow between the aquifer and river was computed by the flow model for each river cell for each stress period. That flow was added to or subtracted from comparable river cells in the transport-model grid to represent river-aquifer flow.

Porosity and Hydrodynamic Dispersion

Convection of solute in the aquifer is related to the porosity of the media (eq. 2). Laboratory analysis of porosity from samples of the aquifer materials ranged from 24.1 to 60.2 percent (Williams and Lohman, 1949, p. 94-99). The value of porosity used in the model was varied by trial-and-error to determine the effect on the simulated concentration of chloride ion. Because there is an inverse relationship between porosity and seepage velocity (eq. 2), solute moves more rapidly in less porous media. In addition, the dilution of solute in the aquifer is directly proportional to the porosity. All other things being equal, the use of a smaller value for porosity for the simulation resulted in a wider distribution of solute and in a larger concentration of solute at cells in the aquifer than did the use of a larger value for porosity. A value of 25 percent for porosity was selected as most consistent with the measured values and simulated results.

Hydrodynamic dispersion also affected the movement of solute in the aquifer (eq. 3). Hydrodynamic dispersion was related to values of longitudinal dispersivity and the ratio of transverse to longitudinal dispersivity for modeling purposes. Analysis of two breakthrough curves plotted from data in the area indicated that longitudinal dispersivity could range from about 400 to about 1,300 feet. Values of 100 feet for longitudinal dispersion and 0.3 for the ratio of transverse to longitudinal dispersion resulted in a best-fit between model results and measured data (Sophocleous, 1983, p. 43). These last two values were used in this study.

Chloride-Ion Concentration, 1940

The initial concentration of chloride ions in part of the study area during 1940 was calculated using a mass-balance approach based on oil production in the Burrton Petroleum Field from 1932-43. Total production from the oilfield was compiled from records on file with the Kansas Geological Survey (Lawrence, Kansas). The total volume of brine was determined by multiplying the total oil production from the oilfield by a representative brine-to-oil ratio. The brine-to-oil ratio is the volume of brine per unit volume of oil produced. The brine-to-oil ratios used for this study are shown in table 2. The ratios reflect the general condition that more brine than oil was produced as development of the oilfield continued. Also, as development of the oilfield continued, more brine was disposed into deep zones in rocks of Cambrian and Ordovician age through injection wells. By 1944, 95 percent of the brine was disposed into the deep zones

(Williams and Lohman, 1949, p. 177). The total volume of brine produced from the oilfield is summarized in table 2.

The total volume of brine was distributed on a percentage basis to cells where surface-disposal "evaporation pits" were used. Location of the pits were determined from aerial photography flown during the time when the pits were active. The total volume of brine was assigned a chloride-ion concentration of 120,000 mg/L based on the average of 5 brine analyses from the oilfield given by Schoewe (1943, p. 55). The concentration of chloride ion in the aquifer during 1940 was calculated by mixing the volume of brine allocated to each cell where brine pits were used with the volume of water in storage beneath the section. The chloride-ion concentration outside of the area where brine disposal took place was obtained from Williams and Lohman (1949, pl. 29). The area where evaporation pits were used for brine disposal in the model area and the chloride-ion concentration during 1940 are shown in figure 29.

Water Quality in Arkansas River

The chloride-ion concentration associated with the median streamflow at the Hutchinson streamflow-gaging station was the concentration applied to the rate of river loss. The flow model computed the river loss for each stress period for cells that represented the Arkansas River. The chloride-ion concentration was derived by regression between discharge and dissolved chloride ion in streamflow at the Hutchinson station. Streamflow at the Hutchinson station between 1940-57 was synthesized using long-term streamflow records from the Great Bend and Wichita stations (fig. 1). Chloride-ion concentrations were measured at the Hutchinson station between 1961 and 1978 and ranged between 363 and 907 mg/L. The applied concentrations of chloride ions for use in the transport model ranged from 459 to 606 mg/L between 1940 and 1979.

Table 2.--*Calculated brine production from the Burrton Petroleum Field, 1932-43*

Year(s)	Oil production (barrels)	Brine-to-oil ratio	Brine production (barrels)	Percentage of brine production disposed into evaporation pits	Brine disposed into evaporation pits (barrels)
1932-37	21,400,000	2	42,800,000	90	38,520,000
1938	3,500,000	3	10,500,000	60	6,300,000
1939	3,100,000	5	15,500,000	40	6,200,000
1940	2,600,000	6	15,600,000	30	4,680,000
1941	2,500,000	6	15,000,000	20	3,000,000
1942	2,000,000	6	12,000,000	10	1,200,000
1943	3,300,000	6	19,800,000	5	990,000
				Total	60,890,000

Discussion of Transport-Model Results

The simulated hydraulic-head distribution for 1980 generated by the transport model was, in most of the modeled area, identical to that generated by the flow model. However, the water levels computed by the transport model were as much as 3 feet lower than the water levels computed by the flow model near areas of large withdrawal by wells in the center of the Wichita well field. This difference would result in the computation of slightly faster velocities between cells and slightly larger concentrations of chloride ions. These effects were considered to overestimate the concentration of chloride ions to a small degree in this area.

Chloride-ion concentration in well water and the contoured chloride-ion distribution for the end of the transport-model simulation are shown in figure 30. The location of the simulated chloride-ion front (the 100-mg/L line) west of the Wichita well field corresponds closely with measured concentrations. North of Burrton, measured concentrations were significantly smaller than simulated concentrations. In this area, the concentration of chloride ion used to begin the simulation during 1940 may have been overestimated or the ion may presently be undetected due to the location or construction of wells in the area. Simulated concentrations exceeded measured concentrations along the Arkansas River in the southwestern part of the area, but the simulated concentrations were smaller than the measured values along the river further east. Because the river was considered the only source of chloride in this area, the larger measured values could represent either the result of underestimating the concentration of chloride ion in the river as applied in the model or the presence of an unknown source of chloride ion near the river.

There were several reasons for discrepancies between measured and simulated concentrations. The most apparent reason was that the exact volume of brine disposed in the Burrton area was unknown. This was considered a significant problem because indirect methods were used to derive the concentration of chloride in the aquifer at the beginning of the simulation. The streamflow loss from the Arkansas River was simulated as a source of chloride ion to the Equus beds aquifer but was not measured directly. Because streamflow in the Arkansas River has not been measured independently of streamflow in the Little Arkansas River, the rates of streamflow gain or loss between the Arkansas River and the Equus beds aquifer also were determined by indirect methods. Similarly, representative values for the concentration of chloride ion in the Arkansas River were determined by indirect methods.

Measurements of chloride-ion concentrations from wells varied considerably within short distances in some parts of the area during 1980 (fig. 30). Heterogeneity of deposits in the aquifer that affected the vertical movement of the solute could have contributed to the local variation in measured concentrations. Additional sources of solute, such as localized spills, also could have affected the measurements. The location of the simulated contours generally was representative of the measured distribution of chloride ions in the Equus beds aquifer. However, a precise simulation of the distribution of chloride ions would be unlikely given the inherent uncertainty of the chemical data, the nature of the aquifer, and the scale of the solute transport model.

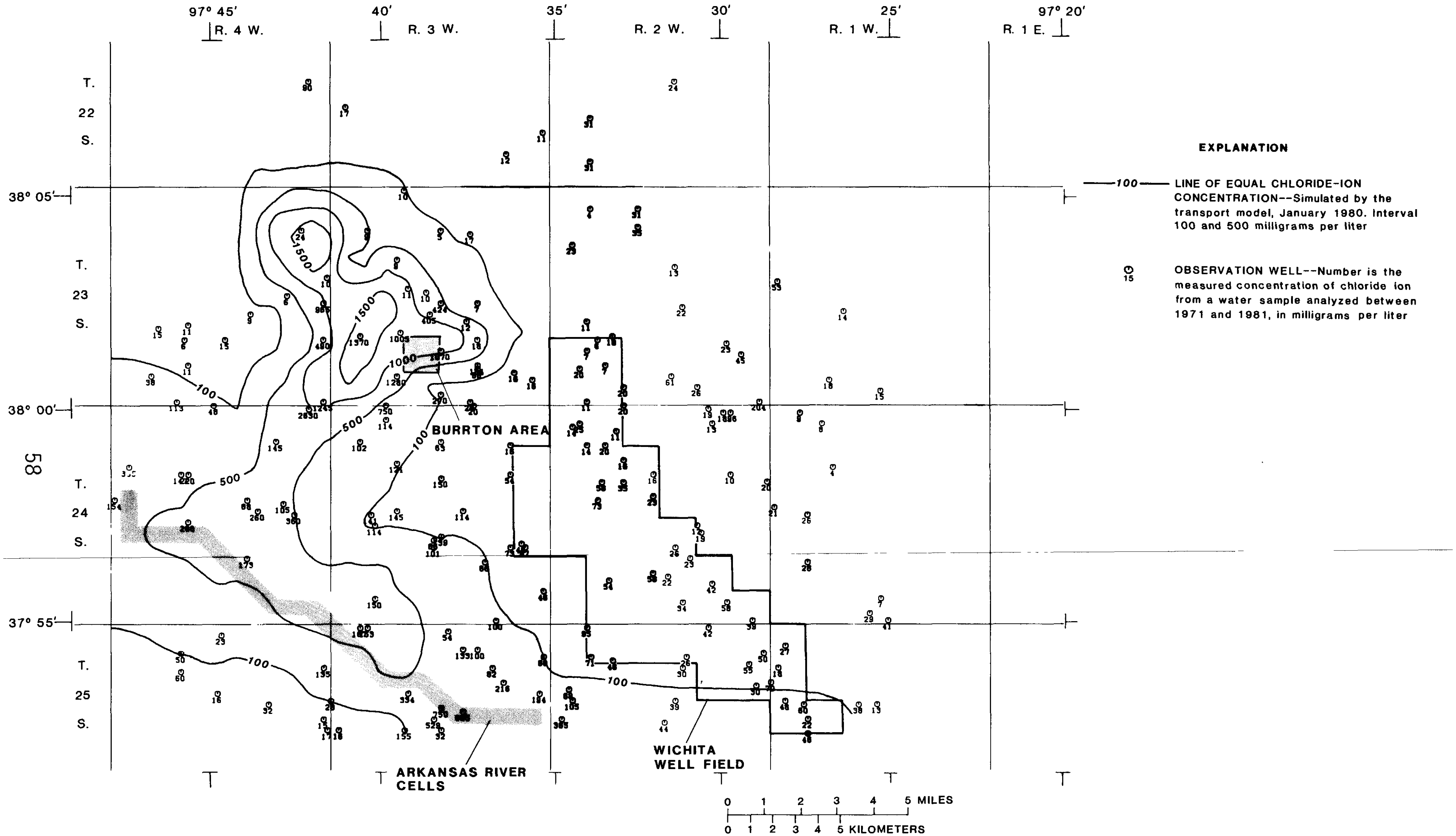


Figure 30.--Measured and simulated chloride-ion concentrations in part of the Equus beds aquifer, January 1980.

Sensitivity Analysis

The aquifer properties used in the transport model were evaluated in the section describing the flow model and were considered representative of the actual flow system. Sensitivity analysis for the transport model was limited to porosity and values representing hydrodynamic dispersion. For each sensitivity simulation one of the aquifer properties was increased or decreased uniformly for the entire transport-model grid by a proportional amount from the value used in the accepted simulation without changing the other properties. Results from the sensitivity simulations were compared to results from the accepted simulation.

The frequency distribution (fig. 31A) indicates the range of differences between "measured" and simulated chloride-ion concentrations for the model grid at the end of the accepted simulation. "Measured" concentrations were computed at each cell in the grid using kriging, a mathematical interpolation technique (Skrivan and Karlinger, 1980), based on the 188 measurements shown in figure 30. The transport model simulated chloride-ion concentrations within plus-or-minus 50 mg/L for 70 percent of the area; plus-or-minus 100 mg/L for 83 percent of the area; and plus-or-minus 200 mg/L for 87 percent of the area. The larger chloride-ion concentrations simulated by the transport model compared to measured concentrations north of Burrton (fig. 30) are the probable reason for a greater number of differences on the positive side of the frequency distribution.

The mean and standard deviation of the differences between "measured" and simulated chloride-ion concentrations summarize the characteristics of the distribution shown in figure 31A for the accepted simulation and compares the summarized results from each of the sensitivity simulations (fig. 31B). Sensitivity simulations illustrated in figure 31B are identified as:

Simulation number	Aquifer property	Percent of accepted simulation
1	Porosity	-50
2	Longitudinal dispersivity	+1,000
3	Transverse dispersivity/ longitudinal dispersivity	+50
4	Accepted simulation	--
5	Transverse dispersivity/ longitudinal dispersivity	-50
6	Longitudinal dispersivity	-1,000
7	Porosity	+50

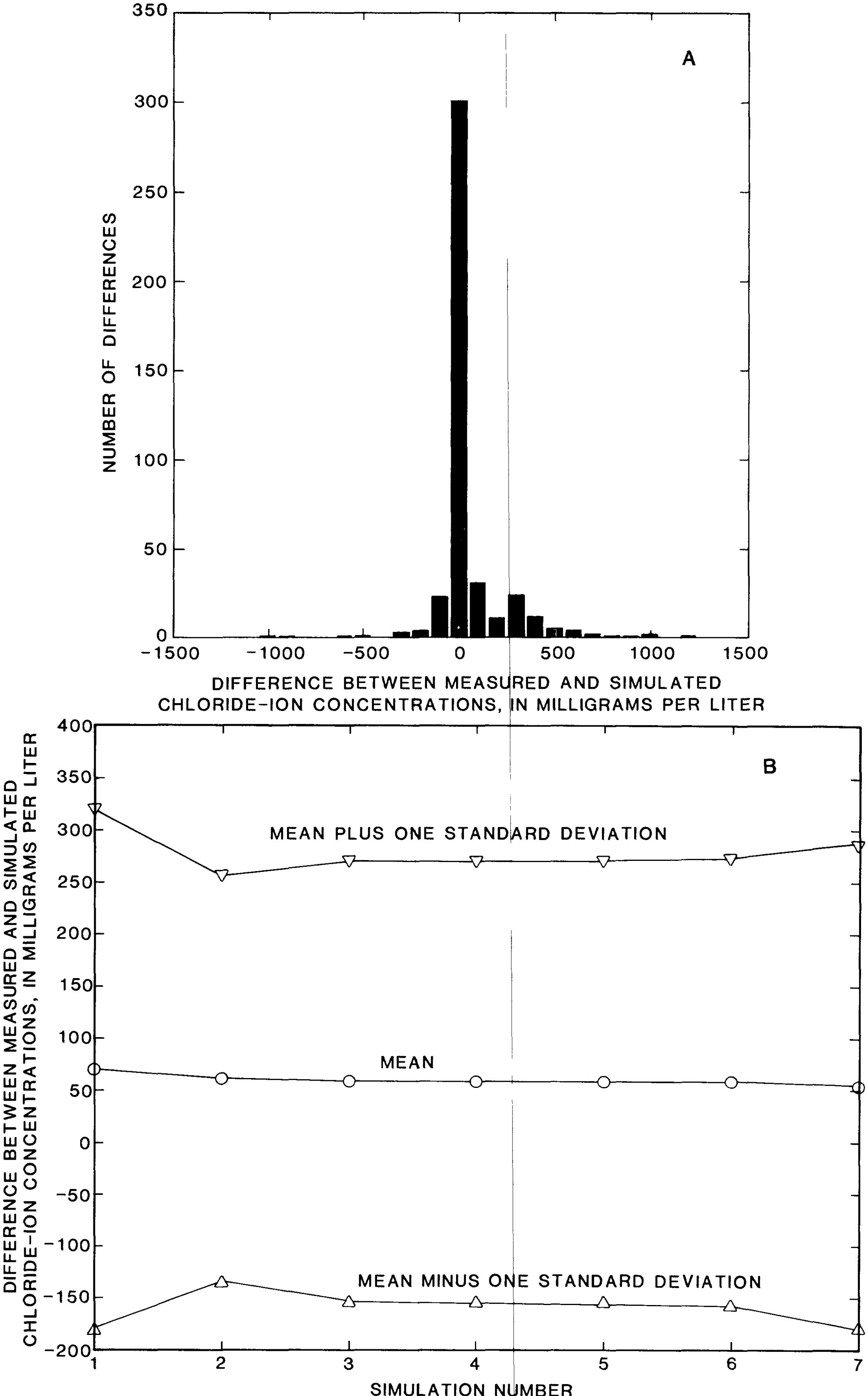


Figure 31.--Sensitivity of transport-model results to variations in selected aquifer properties. (A) Frequency distribution of difference between measured and simulated chloride-ion concentrations at all cells in transport-model grid at end of accepted simulation, 1979, and (B) mean and standard deviations of differences between measured and simulated chloride-ion concentrations at all cells in transport-model grid at end of accepted simulation and each of six sensitivity simulations, 1979.

Longitudinal dispersivity was varied by a factor of 10 to evaluate the potential range for this property. Variation by 50 percent of the accepted value was considered within the physical range for porosity, cited earlier in this report. Results from the sensitivity analysis indicate a relative insensitivity to the difference between measured and simulated chloride-ion concentrations among the ranges of the values for the selected aquifer properties used in the sensitivity analysis.

Transport-Model Projections

The transport model was used to project the effect of three pumping alternatives on the movement of chloride ion between 1980 and 2020. Projections using the transport model corresponded to flow-model projections in terms of time, hydrologic conditions, and aquifer properties. The distribution of chloride ion simulated by the transport model for 1980 was used as the initial concentration for all transport-projection simulations. A constant chloride-ion concentration of 1,000 mg/L was applied to losing reaches of the Arkansas River based on the premise that the concentration would be 10 percent greater than the greatest measured concentration should streamflow in the river decrease in the future. The three pumping alternatives corresponded with flow-model pumping alternatives of: (1) 1971-79 ground-water-withdrawal rates decreased by one-half, (2) 1971-79 withdrawal rates continued, and (3) 1971-79 withdrawal rates doubled.

Results from each of the three pumping alternatives indicated that the chloride-ion concentration at cells in the Wichita well field would increase by 2020 in direct proportion to the projected withdrawal rates. Changes in the projected chloride-ion concentration in the northern and western parts of the well field (fig. 32, cells A and B, respectively) were relatively small for pumping alternatives 1 and 2. Alternative 1 projected concentrations that doubled at the northern part from about 20 to 45 mg/L and barely changed in the western part of the well field mainly from 2015 to 2020. For alternative 2, the concentration in the northern part of the well field was projected to stay steady at about 20 mg/L from 1980 through 2005, then increase to about 100 mg/L by 2020. The concentration in the western part of the well field was projected to increase from about 90 mg/L during 1980 to 110 mg/L during 1995, gradually increasing to about 135 mg/L by 2020. For alternative 3, concentrations in the northern part of the well field were projected as stable from 1980-2000, then gradually increased to about 260 mg/L by 2020. Concentrations in the western part of the well field were projected to be about 50 mg/L larger for 2020 than those projected by alternative 2.

The concentrations of chloride ion in the southern part of the well field (cell C) were projected to increase by 1995 for each of the three pumping alternatives. The projected concentration ranged from about 90 mg/L during 1980 to about 215 mg/L by 2020 for alternative 1 and from about 90 mg/L during 1980 to about 450 mg/L by 2020 for alternative 3. The greater increase in chloride-ion concentration at the southern part of the well field for all simulated projections indicated that the continuous 1,000-mg/L source of chloride ion assumed for streamflow losses from the Arkansas River had a greater effect on increasing the chloride-ion concentration in the well field than did residual oilfield brine at the northern and western parts of the well field.

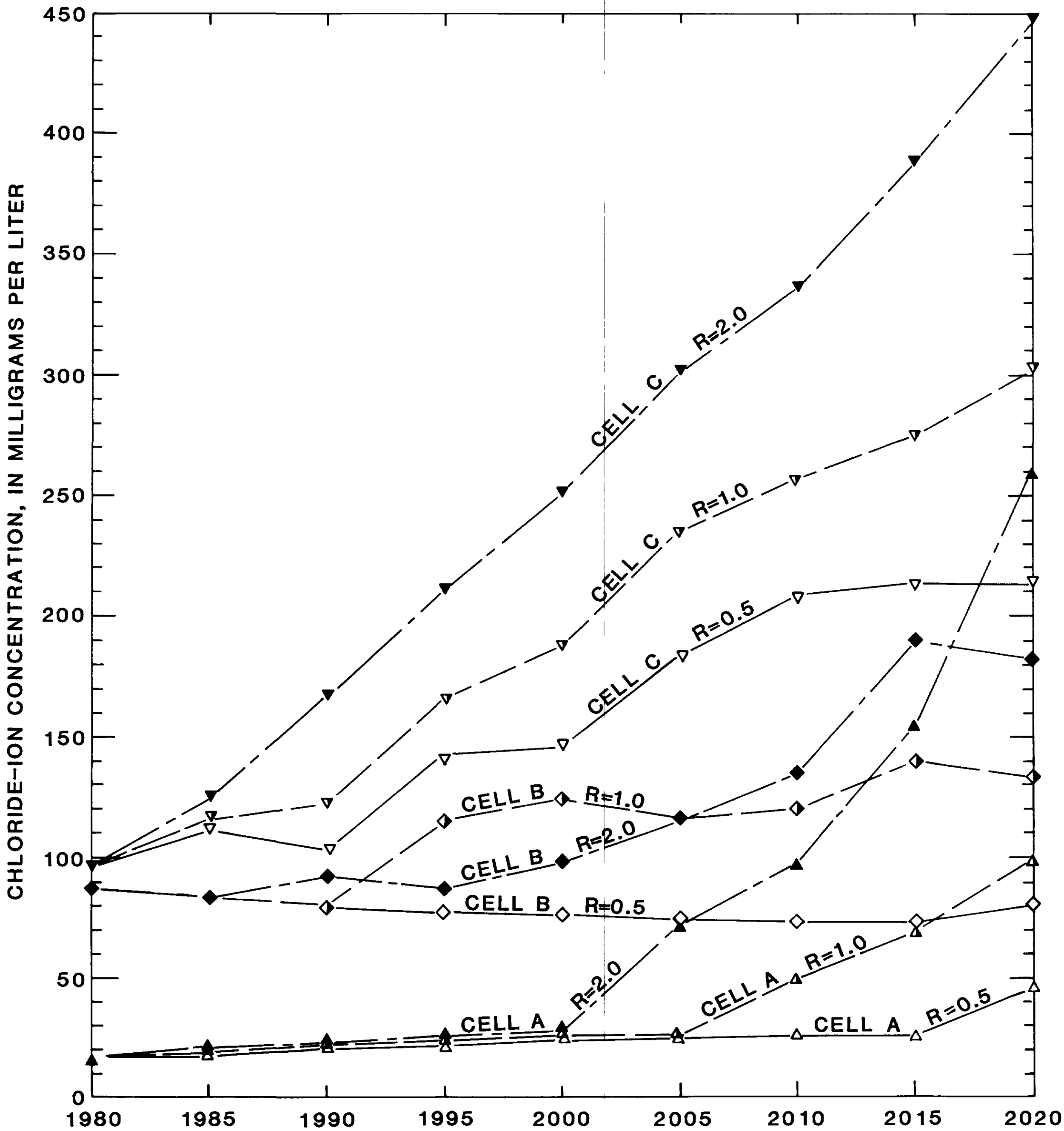


Figure 32.--Effects of alternative withdrawal rates on chloride-ion concentrations for selected grid cells shown in figure 28. (Withdrawal rates, R, are multiples of 1971-79 rates. For pumping alternative 1, R = 0.5; for alternative 2, R = 1.0; for alternative 3, R = 2.0.)

The projected chloride-ion concentration in part of the aquifer for 2020 for pumping alternative 3 is shown in figure 33. This figure indicates that the source of the projected increase in concentration in both the western and the southern parts of the Wichita well field was the mineralized water from the Arkansas River; whereas the source of the projected increase in concentration in the northern part of the well field was oilfield brine from the Burrton area. Figure 33 can be used in conjunction with figure 32 to estimate the effects of different pumping alternatives on chloride-ion concentrations in parts of the Equus beds aquifer near the Wichita well field. For example, in the vicinity of cell A, the chloride-

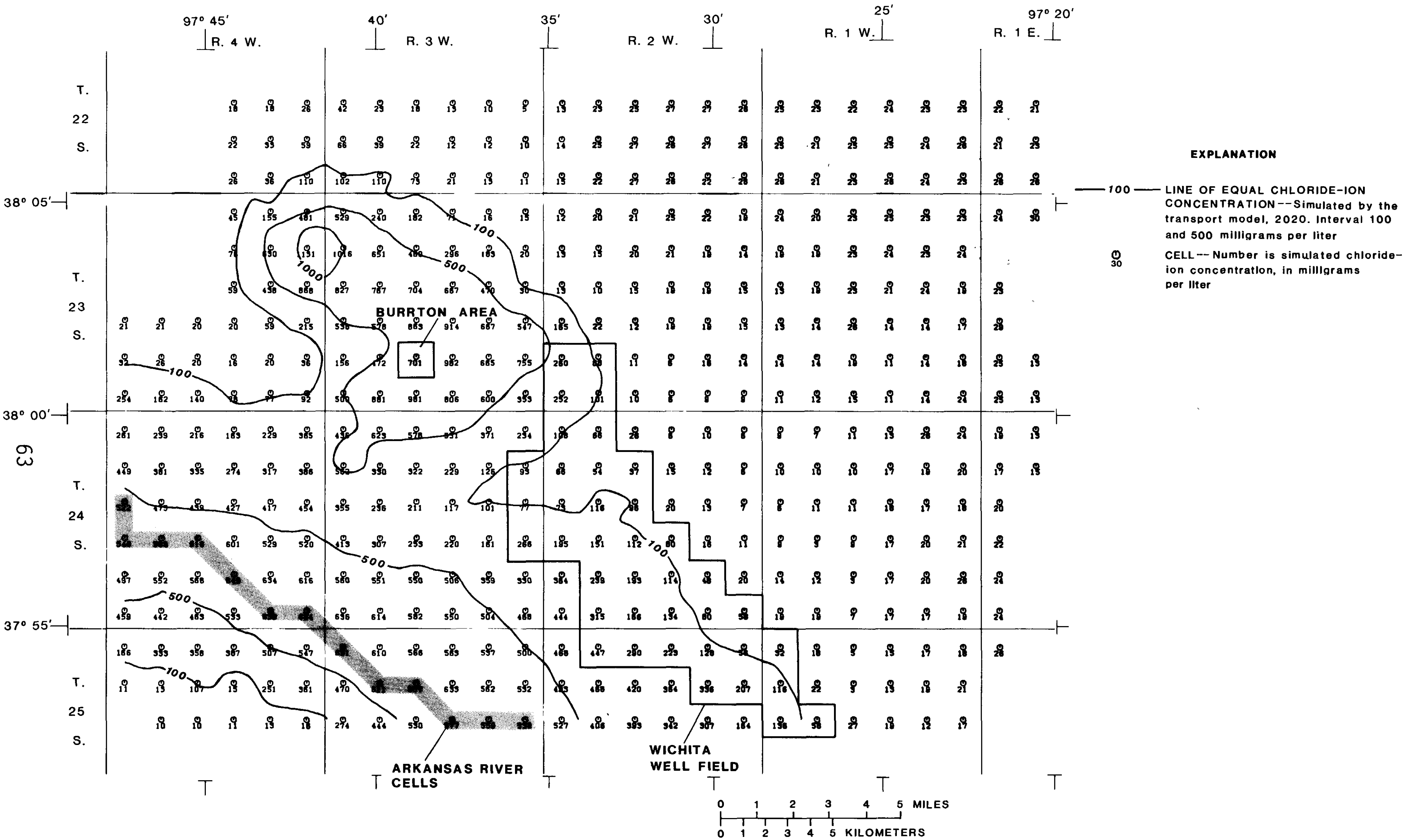


Figure 33.--Projected chloride-ion concentrations in part of the Equus beds aquifer, 2020, using ground-water-withdrawal rates double those for 1971-79.

ion concentration would be approximately one-fifth of the concentration shown in figure 33 by 2020 were 1971-79 withdrawal rates decreased by one-half.

SUMMARY

The Equus beds aquifer is the principal source of ground water underlying about 1,400 square miles in a part of south-central Kansas. About 130 Mgal/d were withdrawn by wells from the aquifer during 1980 for municipal, industrial, and irrigation uses. Withdrawal by wells has resulted in a decline in the water table in the aquifer between 1940 and 1980. The maximum decline of 30 feet has occurred where the city of Wichita maintains a well field. Total water demand in the region was projected by the Kansas Water Office to be 39 percent greater during 2035 than during 1980. Ground water is likely to provide a large part of this projected demand.

The Arkansas and the Little Arkansas Rivers cross the area and are maintained by discharge from the Equus beds aquifer. Streamflow in several perennial creeks in the area also are maintained by discharge from the aquifer.

The Equus beds aquifer is composed of silt, clay, sand, and gravel with saturated thickness ranging from 0 to about 300 feet. Reported specific yield in the aquifer ranges from 0.08 to 0.34; specific yield used in the flow and transport models was 0.15. Hydraulic conductivity used in the models ranged from 5 to 750 ft/d. Normal measured precipitation in the Equus beds area is about 30 in/yr.

The Wellington aquifer is present below about one-third of the Equus beds aquifer. The aquifer was formed by the dissolution of evaporite deposits and the collapse of overlying rocks. Storage coefficient used in the transient simulation of the ground-water flow model for the aquifer was 0.0001. Transmissivity ranged from 8.64 to 2,562 ft^2/d. The concentration of dissolved chloride in the aquifer averages about 150,000 mg/L. Leakance between the Wellington aquifer and the Equus beds aquifer ranged from 0.6×10^{-8} to 9.68×10^{-7} ft/d-ft. The aquifer is separated from the Equus beds aquifer by shale that averages 250 feet in thickness in the area.

The Equus beds aquifer and the Wellington aquifer comprise a ground-water flow system in the area. Ground-water flow in the system was simulated with a three-dimensional, finite-difference, digital-computer model. Equivalent freshwater hydraulic heads were used to represent hydraulic heads in the Wellington aquifer. The model simulated flow in the system in response to ground-water withdrawal by wells and recharge from 1940 to 1979. Water levels computed by the model for the Equus beds aquifer at the end of 1970 and 1979 compared favorably to measurements from wells for January 1971 and January 1980 in the aquifer. Water levels computed by the model for the Wellington aquifer also compared favorably to measurements available from the aquifer. Streamflow gains simulated by the model were comparable to measured gains. At the end of the transient simulation, the simulated water levels in the Equus beds aquifer were above simulated water levels in the Wellington aquifer everywhere in the study area.

The flow model was used to project the effect of five pumping alternatives on ground-water flow in the system. Model projections were made for 1980 to 2020. Recharge to the system during this time was extrapolated by continuing the historical pattern of recharge during 1940-80 to 2020. The five pumping alternatives were multiples of the 1971-79 withdrawal rates of the wells in the area. Additional withdrawal was represented by a hypothetical well field. For the first alternative, 1971-79 withdrawal rates were decreased by one-half. Results showed that saturated thickness was greatest in the Equus beds aquifer at the end of the projection for this alternative compared to all others. For the second alternative, withdrawal rates were continued at 1971-79 levels. Results of this projection indicated a decline in saturated thickness in some areas of about 10 to 40 feet compared to the results of the first alternative. The remaining three projections increased the withdrawal rates by multiples of one-third, two-thirds, and two-times the 1971-79 rates. With rates of withdrawal increased by one-third, some parts in the aquifer were projected to dewater. Declines in the projected saturated thickness became greater as the rates increased. With two-times the withdrawal rates, projected declines in saturated thickness in some areas were about 80 feet more than those projected by halving the withdrawal rate. Water levels projected for all pumping alternatives were higher in the Equus beds aquifer than those in the Wellington aquifer for all sections in the study area.

Projected streamflow gain was maintained for a reach of the Little Arkansas River adjacent to the Wichita municipal well field for continuing 1971-79 withdrawal rates. Projected streamflow loss resulted from increased withdrawal rates. Projected streamflow gain in the Arkansas and Little Arkansas Rivers for 2020 increased to 75 ft^3/s when 1971-79 withdrawal rates were decreased by one-half. Projected streamflow gain was maintained at about 10 ft^3/s when 1971-79 withdrawal rates were continued. Streamflow losses were projected for the withdrawal rates of one-third, two-thirds, and two-times more than the 1971-79 rates. Model results are not valid if actual surface-water supplies from outside the study area boundary are not available to offset simulated streamflow losses.

Oilfield brine was disposed into the Equus beds aquifer in the past. About 60,890,000 barrels of brine were calculated to have been disposed into the aquifer by 1943. Chloride-ion concentrations of the brine averaged 120,000 mg/L. Measured chloride-ion concentrations in the unaffected part of the aquifer were as low as 20 mg/L. Mineralized water in the Arkansas River may flow into the aquifer when the direction of the hydraulic gradient is from the river to the aquifer. Chloride-ion concentration in contributions of water from the Arkansas River was calculated as between 459 and 606 mg/L from 1940 to 1979, based on river discharge.

The movement of chloride ion in part of the Equus beds aquifer was represented by a two-dimensional, finite-difference, solute-transport model for 1940-79. By 1979, the measured and projected chloride-ion distributions were generally similiar although there were some differences between measured and simulated values at certain locations.

The transport model was used to project the effects that three pumping alternatives could have on the chloride-ion concentration in the Equus beds aquifer by 2020. The alternatives were based on withdrawal rates of one-half the 1971-79 rates, continued 1971-79 rates, and doubled 1971-79 rates. For all projection simulations, the chloride-ion concentration applied to river losses from the Arkansas River was assumed to be 1,000 mg/L. Each projection showed that the concentration of chloride ion in parts of the Wichita well field would increase by 2020. Projected concentration increases were proportional to the withdrawal rate. The minimum projected increase was from 20 to 45 mg/L between 1980 and 2020 in the northern part of the well field for the halved rates. The maximum projected increase was from 90 to 450 mg/L in the southern part of the well field for the doubled rates. The projections indicated that a continuous 1,000-mg/L source of chloride ion in streamflow losses from the Arkansas River had a greater effect on increasing chloride-ion concentrations in the Wichita well field than did the movement of residual oilfield brine.

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