GEOHYDROLOGY OF THE HIGH PLAINS AQUIFER,

WESTERN KANSAS

By Lloyd E. Stullken, Kenneth R. Watts, and Richard J. Lindgren

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CONVERSION FACTORS

For those readers interested in metric units, the inch-pound units used in this report can be converted to the International System of Units (SI) using the following factors:

Multiply inch-pound unit	By	To obtain SI unit
inch	25.4	millimeter
foot	0.3048	meter
mile	1.609	kilometer
square foot	0.09290	square meter
acre	4,047	square meter
square mile	2.590	square kilometer
acre-foot	1,233	cubic meter
inch per year	25.4	millimeter per year
foot per second	0.3048	meter per second
foot per day	0.3048	meter per day
foot per mile	0.1894	meter per kilometer
square foot per day	0.09290	square meter per day
acre-foot per year	1,233	cubic meter per year
gallon per minute	0.06309	liter per second
gallon per day	3.7854	liter per day
cubic foot per second	0.02832	cubic meter per second

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ABSTRACT

The High Plains aquifer underlies 174,050 square miles of eight states (Colorado, Kansas, Nebraska, New Mexico, Oklahoma, South Dakota, Texas, and Wyoming) and contains approximately 3.3 billion acre-feet of water in storage. The High Plains aquifer in Kansas consists of stream- and wind-laid deposits of unconsolidated clays, sands, and gravels of the Ogallala Formation and similar associated Tertiary and Quaternary deposits that underlie 30,500 square miles of western and south-central Kansas. The deposits were laid down on an erosional bedrock surface, which was formed on consolidated rocks of Permian to Cretaceous age. Saturated thicknesses within the aquifer are as great as 600 feet near the southern border of southwest Kansas.

The aquifer is replenished primarily by infiltration from precipitation. Average precipitation at the Garden City Experiment Station is 18.93 inches per year. Ground-water flow is generally from west to east under unconfined conditions. Hydraulic connection with subcropping consolidated aquifers allows ground water to flow vertically in minor quantities.

The aquifer is depleted primarily by irrigation. The number of irrigation wells has increased exponentially from less than 500 during 1940 to about 20,000 during 1980. During 1980 there were over 100 irrigation wells per 36 square miles in some areas of west-central and southwest Kansas.

Hydraulic-conductivity estimates from 1,612 lithologic logs had an average value of 75 feet per day, with a standard deviation of 35 feet per day. Hydraulic conductivities estimated from specific-capacity tests of the High Plains aquifer in Sherman County were about one-fourth of the lithologic estimates of hydraulic conductivity. Specific yields estimated from the same lithologic logs had a mean of 0.17 and a standard deviation of 0.047. A spatial analysis indicated that correlation of individual point values is poor and adds support to the description of the Ogallala Formation as being homogeneous in its heterogeneity.

Water from the High Plains aquifer in Kansas generally is suitable for human and animal consumption and irrigation of crops. Typically, it is a calcium bicarbonate type water, with concentrations of total dissolved solids ranging from 250 to 500 milligrams per liter. The quality of water in the aquifer deteriorates toward the east due to mixing with recharge water containing dissolved minerals leached from the overlying soil and unsaturated zones and mineralized water from adjacent bedrock units. The

result is a water containing greater concentrations of dissolved solids and a change to a calcium sulfate or sodium chloride type water. Quality of water from the aquifer degrades with depth in parts of Meade and Seward Counties and in the eastern one-half of south-central Kansas due to mixing of saline water from underlying Permian rocks.

Steady-state simulations of the High Plains aquifer in northwest and southwest Kansas were developed using a U.S. Geological Survey finite-difference modeling code for two-dimensional ground-water flow. Nodes were located 15,000 feet apart in a grid pattern.

The simulated water budget for the steady-state model of predevelopment (pre-1950) conditions in the High Plains aquifer in northwest Kansas showed that annual recharge to the aquifer from infiltration of precipitation was 87,000 acre-feet per year and from boundary inflow, 21,000 acre-feet per year. Annual discharge from the aquifer was 108,000 acre-feet per year, including 81,000 acre-feet per year from leakage to streams, 23,000 acre-feet from outflow at the boundaries of the aquifer, and 4,000 acre-feet from municipal and industrial pumpage.

The simulated water budget for southwest Kansas for predevelopment (pre-1950) conditions showed that annual recharge from precipitation was 104,500 acre-feet and from boundary inflow, 32,500 acre-feet. Annual discharge from the aquifer was 137,000 acre-feet, including 58,000 acre-feet from net leakage to streams, 14,000 acre-feet from leakage to the underlying Lower Cretaceous sandstone aquifer, and 65,000 acre-feet from outflow at the boundaries of the aquifer.

INTRODUCTION

The agricultural economy of eight states in the High Plains is greatly affected by the capacity of the High Plains aquifer to sustain water withdrawals. The aquifer contains approximately 3.3 billion acre-feet of water in storage (Weeks and Gutentag, 1981), but water is being withdrawn for irrigation in excess of the rate of natural replenishment. Pabst and Gutentag (1979) reported a tenfold difference between estimated annual recharge and the amount of water withdrawn in southwest Kansas. Responding to a need for regional analysis of the aquifer, the U.S. Geological Survey began a study of the High Plains Regional Aquifer System during 1978 to provide (1) hydrologic information needed to evaluate the effects of continued ground-water development, and (2) computer models to predict aquifer response to changes in ground-water development. The plan of study for the High Plains Regional Aquifer System Analysis is described by Weeks (1978). This report is a summation of the data collection, hydrologic analysis, and computer modeling completed for the Kansas part of the regional study.

Purpose and Extent

Previous studies of the hydrology of the High Plains have been limited by political boundaries. The High Plains Regional Aquifer System Analysis provides a regional description of the water resources and operation of the hydrologic system consistent with the natural boundaries of the High Plains aquifer. The extent of the High Plains regional aquifer system is shown by shading in figure 1. The High Plains aquifer underlies 174,050 square miles in eight states, of which 30,500 square miles are in Kansas. This report is for the Kansas part of the aquifer.

The High Plains aquifer in Kansas is found in five distinct areas (fig. 1). The northwest area consists of all the aquifer north of the Smoky Hill River. The west-central area consists of the aquifer south of the Smoky Hill River in Wallace, Greeley, Wichita, Scott, Lane, and Ness Counties. The southwest area consists of the aquifer south of the Scott-Finney County line and west of central Ford County. The south-central area comprises the aquifer from central Ford County east to central Reno County (this area is also called the Great Bend Prairie or Big Bend Prairie). The Equus Beds area consists of that part of the aquifer east of central Reno County.

Methods of Investigation

The methods used to conduct this investigation may be categorized into data collection, compilation, and analysis and digital simulation. Data collection for water-use computations was a major work item during 1978-80 and included measurements of well discharge along with time-of-operation metering of irrigation wells, measurements of irrigated acreage, and inventories of crops and wells. Other measurement activities included determination of channel-geometry characteristics to estimate aquifer-streamflow statistics at many ungaged sites.

The Ground Water Site Inventory (GWSI) contained in the U.S. Geological Survey's WATSTORE computer data base was a primary repository and information source. A well inventory was made in northwest Kansas to update the GWSI data base. In southwest and south-central Kansas, records of water-right applications from the Division of Water Resources, Kansas State Board of Agriculture, and from Groundwater Management Districts 1, 3, 4 and 5 were used to locate all known irrigation wells in the High Plains of Kansas. When the year of well installation was not known, it was presumed to be the year the application was filed. This inventory was added to the GWSI data base. Considerable care was taken to identify and avoid entries for nonexistent wells or duplicate entries for the same well.

Geohydrologic data for the High Plains aquifer that were reported in publications of the U.S. Geological Survey, the Kansas Geological Survey, the Kansas Water Office (formerly Kansas Water Resources Board), and the Kansas Department of Health and Environment were compiled. Some unpublished data were found in the files of the U.S. Geological Survey office in Garden City, Kansas. These data sources provided the information needed to map water-table contours, the altitude of the base of the aquifer, the bedrock geology, the hydraulic conductivity, the specific yield, and the saturated thickness.

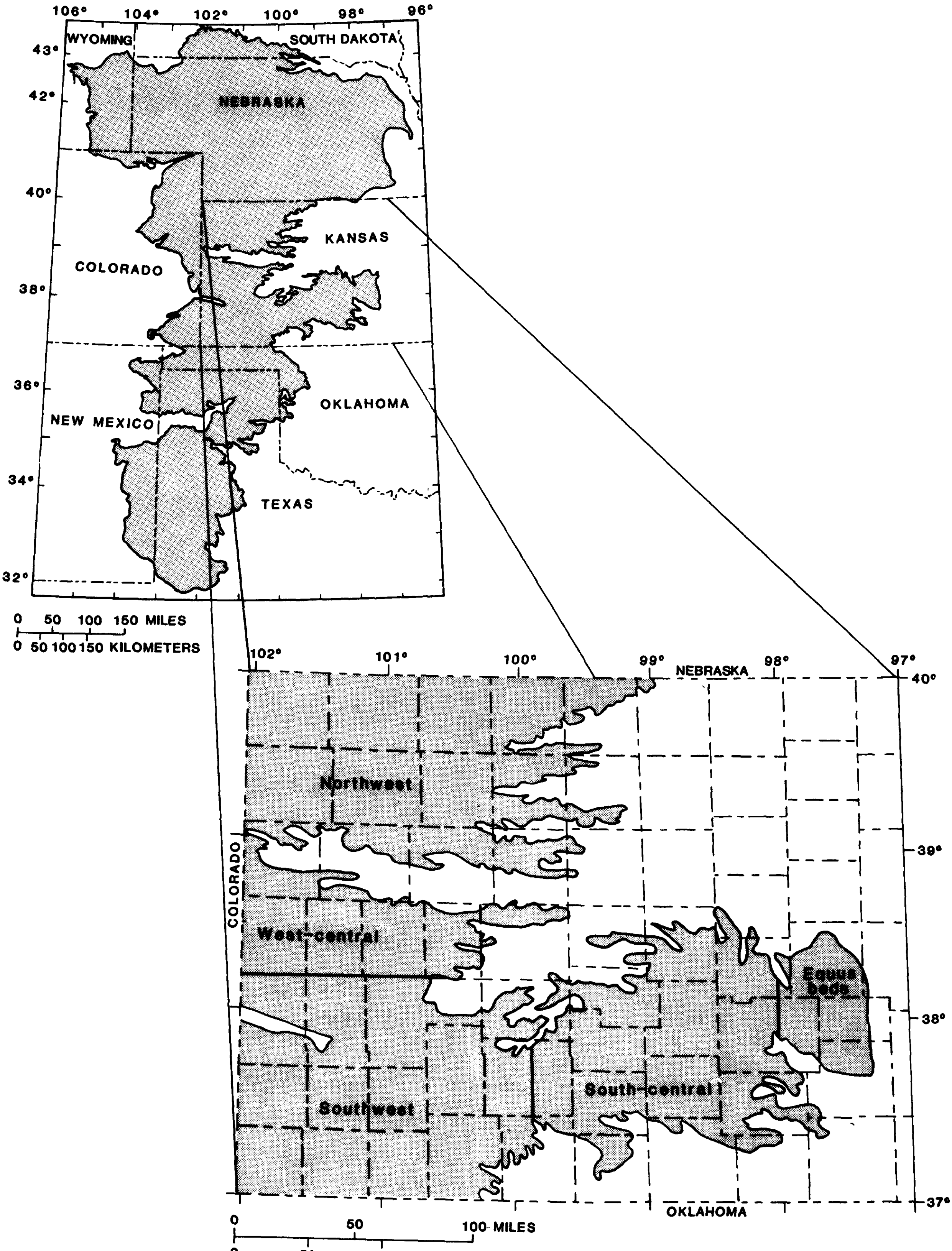


Figure 1.--Location of High Plains aquifer (shaded) and study area.

Flow from the High Plains aquifer to streams was estimated using winter streamflow data from the U.S. Geological Survey WATSTORE computerized data base. Flow from ephemeral streams to the aquifer was estimated using flow statistics derived from channel-geometry measurements.

A finite-difference model for aquifer simulation in two dimensions was used, as described by Trescott and others (1976). The model was calibrated to simulate aquifer conditions as they were prior to 1950, presumably before any significant withdrawals from the ground-water reservoir. Simulations of the High Plains aquifer in northwest and southwest Kansas were considered to be steady-state calibrations because they represented the aquifer at an instant in time when none of the flow components were changing.

Well-Numbering System

Wells in this report are numbered according to a modification of the U.S. Bureau of Land Management's system of land subdivision. In this system, the first set of digits of a well number indicates the township south of the Kansas-Nebraska State line; the second set, the range east or west of the Sixth Principal Meridian; the third set, the section in which the well is situated. The first letter of the third set denotes the quarter section or 160-acre tract within the section; 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, and D in a counterclockwise direction beginning in the northeast quadrant. The last two digits are the sequential order, beginning with "01," in which the wells in the same 10-acre tract were inventoried. In figure 2, well number 24S 32W 03DAC 01 in Finney County is in the SW1/4 NE1/4 SE1/4 sec. 3, T. 24 S., R. 32 W. For convenience, when there is only one well inventoried in the same 10-acre tract, the trailing two-digit designation is dropped.

Acknowledgments

Data collection and compilation for this project were aided by the cooperation of many people and agencies. Particular recognition is given to the many irrigators who allowed discharge measurements, including time-of-operation metering of their wells; the Division of Water Resources of the Kansas State Board of Agriculture and Groundwater Management Districts Nos. 1, 3, 4, and 5, who allowed access to their files of water-right applications; and the Kansas Department of Health and Environment who allowed access to their files of water-well completion forms.

Special recognition is due to P. R. Jordan, U.S. Geological Survey, for his base-flow analysis and to E. R. Hedman and W. R. Osterkamp, U.S. Geological Survey, for their channel-geometry, mean-annual-flow studies, which were made especially for this investigation.

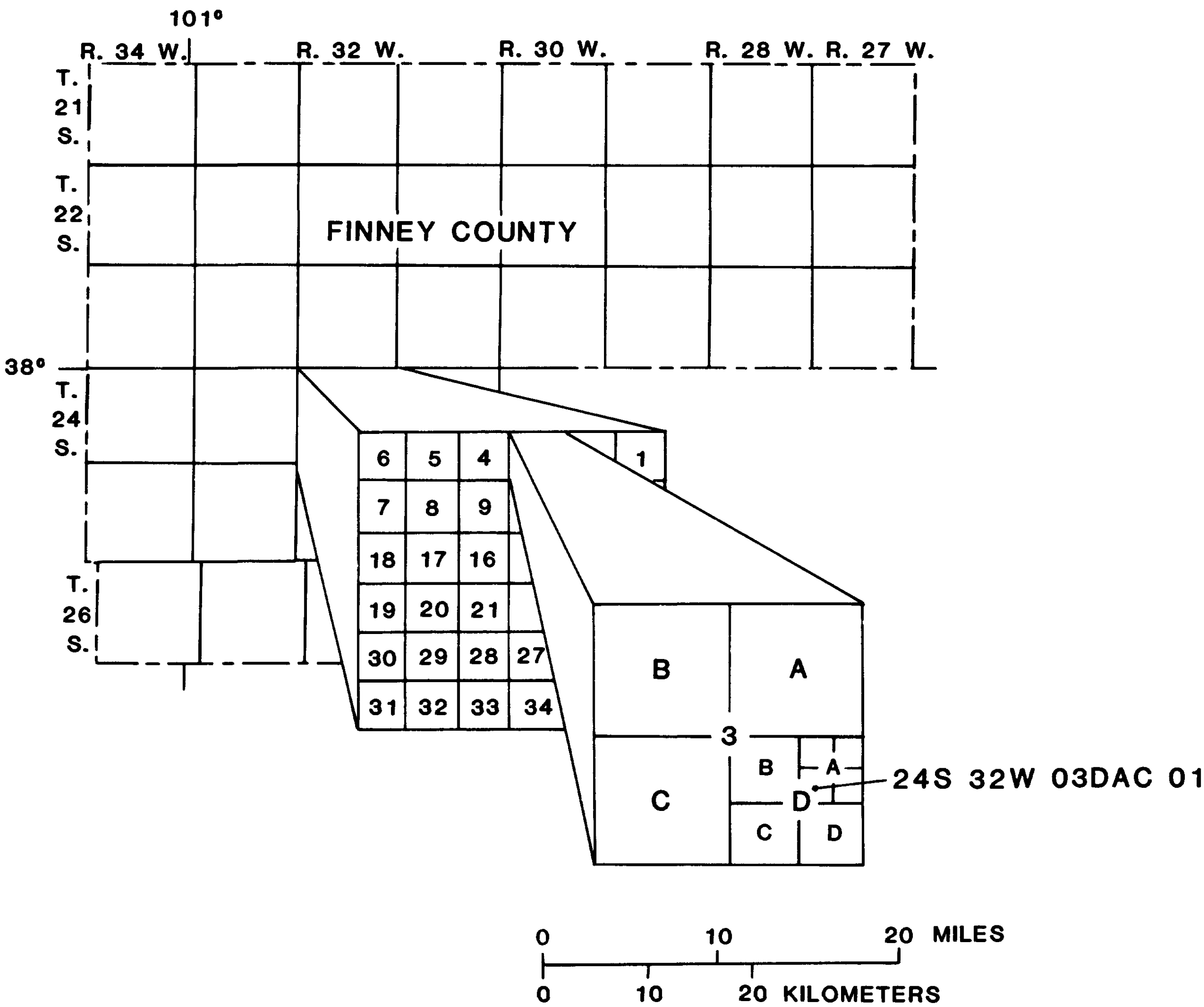


Figure 2.--Well-numbering system.

GEOLOGY

Stratigraphy

The stratigraphic interval considered in this report includes rocks from the Paleozoic, Mesozoic, and Cenozoic Eras. Only those strata that directly affect the hydrogeology of the High Plains aquifer are described. Detailed descriptions of the stratigraphy of Kansas can be found in Zeller (1968) and Merriam (1963) and in other publications of the Kansas Geological Survey. The thickness, physical character, and hydrologic properties of the rocks that affect the High Plains aquifer are summarized in table 1.

Table 1.--*Generalized section of geologic formations and their water-bearing properties*[1/]

[Modified from Gutentag and others, 1981; Pearl and others, 1970; McClain and others, 1975; Gutentag and Stullken, 1976; Fader and Stullken, 1978; Merriam, 1963; and Zeller, 1968]

Erathem	System	Series	Stratigraphic unit	Maximum thickness (feet)	Physical character	Water-supply potential
CENOZOIC	Quaternary	Holocene	Alluvium	80	Stream-laid deposits ranging from silt and clay to sand and gravel that occur along the principal stream valleys.	Yields to wells range from 500 to more than 1,000 gallons per minute.
		Pleistocene	Dune sand	75	Fine to medium quartzose sand with small amounts of clay, silt, and coarse sand formed into mounds and ridges by the wind.	The sand has a high infiltration rate and is important as an area of ground-water recharge in southwest and south-central Kansas.
			Loess	90	Silt with subordinate amount of very fine sand and clay deposited as windblown dust.	Lies above the water table and does not yield water to wells. High specific retention and capillarity.
			Undifferentiated deposits	550	Sand, gravel, silt, clay, and caliche overlie the Ogallala Formation when both formations are present; composite of stream-laid and wind-laid deposits.	The sand and gravel of the undifferentiated Pleistocene deposits and the Ogallala Formation when present are the principal water-bearing deposits in the

(continued on next page)

Table 1.--*Generalized section of geologic formations and their water-bearing properties*--Continued

Erathem	System	Series	Stratigraphic unit	Maximum thickness (feet)	Physical character	Water-supply potential
CENOZOIC	Tertiary	Miocene	Ogallala Formation	500	Poorly sorted clay, silt, sand, and gravel, generally calcareous; when cemented by calcium carbonate, forms caliche layers or mortar beds. "Algal" limestone at top north of the Arkansas River.	southwest and south-central areas. Yields range from 100 to 3,100 gallons per minute.
MESOZOIC	Cretaceous	Upper Cretaceous	Pierre Shale	1,600	Fissile dark-gray shale with local thin weathered zone of orange clay, "ochre" at the top.	Very small yields (1 to 3 gallons per minute) of "poor" quality water in outcrop areas. An "impermeable" lower boundary of the High Plains aquifer in the western two-thirds of northwest Kansas.
		COLORADO	Niobrara Chalk	250	Upper part (Smoky Hill Chalk Member) is yellow to orange-yellow chalk and light- to dark-gray beds of chalky shale. Lower part (Fort Hays Limestone Member) is a white to yellow massive chalky limestone; contains	Initially (1968-72), yielded 500 to 2,500 gallons per minute to wells in northern Finney and eastern Kearny Counties where the Fort Hays Limestone Member has

Table 1.--*Generalized section of geologic formations and their water-bearing properties*--Continued

Erathem	System	Series		Stratigraphic unit	Maximum thickness (feet)	Physical character	Water-supply potential
MESOZOIC	Cretaceous	Upper Cretaceous	COLORADO GROUP	Niobrara Chalk--Continued		thin beds of dark-grey chalky shale.	fractures and solution openings. Because of increased irrigation development, yields have been reduced by as much as 2,000 gallons per minute.
				Carlile Shale	330	Upper parts consists of a dark-gray to blue-black noncalcareous to slightly calcareous shale that locally is interbedded with calcareous silty very fine-grained sandstone. Lower part consists of very calcareous dark-gray shale and thin gray interbedded limestone layers.	Sandstone in upper part locally may yield 5 to 10 gallons per minute to wells.
				Greenhorn Limestone	200	Chalky light yellow-brown shale with thin-bedded limestone. Dark-gray calcareous shale and light-gray thin-bedded limestone; contains layers of bentonite.	Not known to yield significant quantities of water to wells.

Table 1.--*Generalized section of geologic formations and their water-bearing properties*--Continued

Era-them	System	Series	Strati-graphic unit	Maximum thickness (feet)	Physical character	Water-supply potential
MESOZOIC	Cretaceous	Upper Creta-ceous	COLORADO: Graneros Shale	130	Dark-gray calcareous shale inter-bedded with black calcareous shale; contains thin beds of bentonite. Also contains thin-bedded gray limestone and fine-grained silty sandstone layers.	Not known to yield sig-nificant quantities water to wells.
		Lower Creta-ceous	Dakota Formation	300	Varicolored claystone, mudstone, sandy shale, and siltstone with interbedded and lenticular sand-stone. Sandstones are friable to indurated white to brown, fine to medium grained, and cross-bedded to massive. In some areas the formation includes beds of lignite.	An important aquifer in parts of southwest Kansas. Yields of more than 1,000 gallons per minute are re-ported in Ford and Finney Counties. Locally in hy-draulic connection with the overlying High Plains aquifer. May contain radio-nucleids and iron above recommended levels in some areas.
			Kiowa Shale	150	Light-gray to black shale, with thin interbedded sandstones. The Kiowa Shale Member is the upper member of the Purgatoire Formation in Colorado and Okla-homa.	May yield small quanti-ties of water to wells in localized areas.

Table 1.--*Generalized section of geologic formations and their water-bearing properties*--Continued

Era-them	System	Series	Strati-graphic unit	Maximum thickness (feet)	Physical character	Water-supply potential
MESOZOIC	Cretaceous	Lower Creta-ceous	Cheyenne Sandstone	300	Massive to cross-bedded white to buff to light-gray, fine-to medium-grained sandstone, with lenses of gray to black shale, and locally a basal conglomer-ate. The Cheyenne Sandstone Member is the lower member of the Purgatoire Formation in Colorado and Oklahoma.	Locally yields water to wells for domestic and stock use. In south-eastern Colorado is used as a source of irrigation water.
	Jurassic and Triassic	Upper Jurassic and Upper Triassic	Undiffer-entiated rocks	350	Dark-gray shale; interbedded with grayish-green and bluish-green calcareous shale. Contains very fine to medium-grained silty sandstone and some thin lime-stone beds at the base.	In Morton and Stanton Counties, sandstone beds yield water in combination with the overlying Lower Cretaceous units. In the northernmost counties where the aquifer is deep-est, the water is mineral-ized.
PALEOZOIC	Permian	Upper Permian	Big Basin Formation	160	Brick-red to maroon siltstone and shale; contains very fine grained sandstone.	Where not highly miner-alized, may yield small quantities of usable water for domestic and stock wells.

Table 1.--*Generalized section of geologic formations and their water-bearing properties*--Continued

Era-them	System	Series	Stratigraphic unit	Maximum thickness (feet)	Physical character	Water-supply potential
PALEOZOIC	Permian	Lower Permian	Day Creek Dolomite	80	White to pink anhydrite and gypsum; contains interbedded dark-red shale.	Solution cavities have yielded large quantities (300 to 1,000 gallons per minute) of high sulfate water.
			Whitehorse Formation	350	Red to maroon fine-grained silty sandstone, siltstone, and shale.	Fresh to highly mineralized water. Not known to yield significant quantities of water to wells in Kansas.
		NIPPEWALLA GROUP	Dog Creek Formation	60	Maroon silty shale, siltstone, very fine sandstone, and thin layers of dolomite and gypsum.	Not known to yield significant quantities of water to wells in the area.
			Blaine Formation	150	Generally consists of four gypsum and anhydrite beds separated by red shale; contains bedded halite at some sites.	Not known to yield significant quantities of water to wells. Highly mineralized.
			Flowerpot Shale	180	Reddish-brown gypsiferous shale and silty shale with a few thin beds of sandstone and shale.	

Table 1.--*Generalized section of geologic formations and their water-bearing properties*--Continued

Era-them	System	Series	Group	Strati-graphic unit	Maximum thickness (feet)	Physical character	Water-supply potential
PALEOZOIC	Permian	Lower Permian	NIPPEWALLA GROUP	Cedar Hills Sandstone	200	Reddish shale, siltstone, silty shale, and sandstone.	Sandstone may contribute highly mineralized water to the High Plains aquifer where hydraulic connection occurs.
				Salt Plains Formation	300	Reddish-brown sandy siltstone and fine-grained sandstone.	May contribute highly mineralized water to the High Plains aquifer where a hydraulic connection occurs.
				Harper Sandstone	250	Brownish-red siltstone and silty shale with a few thin beds of silty sandstone.	Water may be of poor chemical quality. May yield no water or as much as 100 gallons per minute to wells in eastern part of the area.

Table 1.--*Generalized section of geologic formations and their water-bearing properties*--Continued

Erathem	System	Series		Stratigraphic unit	Maximum thickness (feet)	Physical character	Water-supply potential
PALEOZOIC	Permian	Lower Permian	SUMNER GROUP	Stone Corral Formation	20	White and light-gray anhydrite and dolomite.	Not known to yield significant quantities of water to wells in the area.
				Ninnescah Shale	450	Red and grayish-green shale, siltstone, and very fine grained silty sandstone.	May yield water of fair to poor chemical quality to wells in the outcrop areas.
				Wellington Formation	Greater than 700	Calcareous gray and blue shale containing several thin beds of limestone, gypsum, and anhydrite. The Hutchinson Salt Member, when present, is near the middle of the formation.	Not known to yield significant quantities of water to wells in the area.

1 The classification and nomenclature of the stratigraphic units used in this report are those of the U.S. Geological Survey and differ somewhat from those of the Kansas Geological Survey.

Paleozoic Rocks

The only Paleozoic rocks in contact with the base of the High Plains aquifer in Kansas are those of the Permian System although older Paleozoic strata are present in the subsurface. Rocks of the Lower Permian Sumner and Nippewalla Groups and overlying Lower and Upper Permian rocks underlie the aquifer in parts of south-central and southwest Kansas.

The Sumner Group includes more than 1,000 feet of strata, chiefly gray and silty shale with thick beds of salt in the subsurface and minor beds of red, maroon, and green shale, dolomite, limestone, gypsum, and anhydrite (O'Connor and others, 1968). The lowermost formation of the Sumner Group, the Wellington Formation, contains more than 700 feet of bedded salt in the subsurface in Clark County (Kulstad, 1959, p. 241-247). The uppermost formation of the Sumner Group, the Stone Corral Formation, is composed of anhydrite, dolomite, gypsum, and salt and is one of the most easily recognized "marker beds" on geophysical logs from wells in western Kansas.

The Lower Permian Nippewalla Group, which overlies the Sumner Group, is chiefly "red beds," which consist of red shales, siltstones, very fine sandstones, and bedded salt in the subsurface. Total thickness of the Nippewalla Group in outcropping areas is about 930 feet, excluding the salt beds (O'Connor and others, 1968, p. 51). The thickness of salt within the Nippewalla was estimated to be 250 feet, based on maximum displacement of the Blaine Formation across the Bear Creek and Crooked Creek-Fowler fault zones (Gutentag and others, 1981, p. 9).

The overlying Lower and Upper Permian rocks are predominantly red sandstone, siltstone, gypsum, and dolomite and have a total thickness of about 590 feet (Gutentag and others, 1981, table 1).

Mesozoic Rocks

Rocks of Mesozoic age underlie all of northwest and west-central, most of southwest, and parts of south-central Kansas. A small outcrop of Triassic sandstone in southwestern Morton County is the only reported outcrop of pre-Cretaceous Mesozoic strata in Kansas.

The presence of Jurassic and Triassic rocks in the subsurface is indicated on drillers' and geophysical logs for parts of western Kansas. The rocks identified as Triassic in T. 34 S., R. 42 W. (locally known as Point of Rocks) in Morton County, southwest Kansas, are assigned to the Dockum Group. The extent of the Triassic rocks in the subsurface is poorly defined because of their similarity to the underlying Permian and overlying Jurassic rocks. The predominant lithologies are red sandstones and varicolored shales that attain a maximum thickness of 320 feet (McLaughlin, 1942, p. 71).

Rocks of Jurassic age do not crop out in Kansas but have been identified in the subsurface. Maximum thickness of the undifferentiated rocks of Jurassic age is about 350 feet (O'Connor, 1968, p. 53-54). However, there is some question as to identification of the Jurassic and Triassic rocks because of their similar lithologies and the absence of diagnostic paleontological evidence.

Rocks of Cretaceous age are in contact with the base of the High Plains aquifer throughout most of western Kansas. Rocks of the Lower Cretaceous Series crop out in Barber, Clark, Comanche, and Kiowa Counties and are recognized in the subsurface in much of western Kansas. The basal unit, the Cheyenne Sandstone, is a gray, brown, or white fine- to medium-grained sandstone with interbedded dark shale. Maximum thickness of the Cheyenne Sandstone is about 300 feet (O'Connor, 1968b, p. 54-58). The Cheyenne Sandstone is overlain by the Kiowa Shale, a gray to black illitic shale with fine sandstone stringers. The Kiowa Shale ranges in thickness from 60 to 150 feet.

The Dakota Formation, which overlies the Kiowa Shale, consists of varicolored claystones, mudstones, shales, siltstones, and lenticular sandstones with some lignite beds. Dakota lithology is variable in both horizontal and vertical directions. Typically, the formation consists of a basal sandstone, a middle shale, and an upper sandstone unit. The Dakota Formation is a time-transitional unit with the time boundary between Early and Late Cretaceous age occurring within the formation at stratigraphically higher positions towards the west (King and Beikman, 1976, p. 57). Thickness of the Dakota Formation in Kansas is 200 to 300 feet (Franks, 1966, as cited in O'Connor, 1968b, p. 55).

Rocks of the Upper Cretaceous Series in Kansas consist of light- to dark-gray calcareous and noncalcareous shale, limestone, shaley limestone, and chalk. Aggregate thickness of the Upper Cretaceous is about 2,600 feet. The strata dip to the northwest with stratigraphically lower (older) rocks subcropping to the east and southeast. The formations included in this section are the Graneros Shale, Greenhorn Limestone, Carlile Shale, Niobrara Chalk, and Pierre Shale. All of these formations except the Pierre Shale are included in the Colorado Group. The general character of these rocks and the maximum thicknesses of the formations in Kansas are given in table 1. Generally these rocks are considered to be impermeable or confining beds. Locally, a sandstone near the top of Carlile Shale and limestones with solution-enlarged fractures near the base of the Niobrara Chalk may be aquifers. Detailed descriptions of the stratigraphy of the Upper Cretaceous Series in Kansas may be found in Hattin (1962; 1965a; 1965b; 1975; 1982). Generalized descriptions of the stratigraphy and geologic history of these rocks may be found in Merriam (1963) and Zeller (1968).

Cenozoic Rocks

The Cenozoic rocks of western Kansas contain thick unconsolidated continental deposits of late Tertiary and Quaternary age. The maximum thickness of the Cenozoic deposits in Kansas is about 800 feet and occurs in southwest Seward County near the Oklahoma State line. The Cenozoic rocks of western Kansas consist primarily of alluvial deposits of clay, silt, sand, and gravel but also include extensive eolian deposits of silt (loess) and dune sand. Descriptions of the stratigraphy and geology of the Cenozoic rocks of western Kansas are given by Zeller (1968) and Merriam (1963). More detailed information on the Pleistocene geology of Kansas is given in Frye and Leonard (1952), and information on the stratigraphy of the Tertiary Ogallala Formation of northern Kansas is given by Frye and others (1956). Additional information about the Cenozoic rocks of Kansas is contained in reports of the geology and ground-water resources of individual counties and in the Irrigation Series published by the Kansas Geological Survey and in the Hydrologic Atlases and Water-Supply Papers of the U.S. Geological Survey that are listed in the "References."

Tertiary Rocks

The only stratigraphic unit of Tertiary age identified in western Kansas is the Ogallala Formation of the Miocene Series. Rocks of the Paleocene, Eocene, and Oligocene Series have been identified in other parts of the High Plains but have not been identified in Kansas. The Ogallala Formation is a heterogeneous sequence of unconsolidated deposits of clay, silt, sand, and gravel that are principally of alluvial origin. Minor but stratigraphically significant beds of freshwater limestone and lentils of volcanic ash occur in the formation. The type and degree of cementation within the formation are variable. Lime-cemented beds of silty sand and gravel (mortar beds) and sandy silt (caliche) occur throughout the formation and form ledges or caprock at the outcrop. Sandstones in the Ogallala Formation of northern Kansas may have siliceous cements. A pisolitic limestone with siliceous cement, referred to as the "algal limestone" for its appearance, occurs at the stratigraphic top of the Ogallala Formation in northwest and west-central Kansas.

The thickness of the Ogallala Formation in Kansas is variable because the Ogallala was deposited on an erosional surface, and in parts of Kansas the upper surface of the formation has been removed by erosion. Maximum thickness of the Ogallala Formation in northwest Kansas is about 320 feet (Pearl and others, 1972), about 400 feet in west-central Kansas (McClain and others, 1975), about 500 feet in southwest Kansas (Gutentag and others, 1981), and about 65 feet in south-central Kansas (Fader and Stullken, 1978).

The Ogallala Formation was deposited primarily by easterly flowing aggrading streams carrying debris from the Rocky Mountains to the west. A vast plain of braided streams and coalesced alluvial fans was formed. Distribution of sediment types within the Ogallala Formation is largely random (Breyer, 1975). The Ogallala Formation commonly is considered to be "homogeneous in its heterogeneity," a statement attributed to F. C. Foley of the University of Kansas (Lawrence) in class lecture notes for "Groundwater Geology" during 1955 (E. D. Gutentag, U.S. Geological Survey, written commun., 1984).

Quaternary Rocks

The Quaternary rocks of western Kansas are of Pleistocene and Holocene age. Considerable thicknesses of both alluvial and eolian deposits occur at the surface of the High Plains in Kansas. Quaternary alluvium (stream-laid clay, silt, sand, and gravel) is the predominant type of Cenozoic rock in most of southwest and south-central Kansas. Pleistocene loess mantles much of the upland areas in western Kansas, and Pleistocene and Holocene dune sands cover most of the south-central and a significant part of the southwest areas of the State.

The alluvial deposits of the High Plains area in Kansas consist of both Pleistocene and Holocene age valley-fill, terrace, and flood-plain deposits. These undifferentiated deposits are similar in lithologic character to the Ogallala Formation from which they are partially derived. Sources of sediment for these alluvial deposits were fresh detritus from the Rocky Mountains and materials locally derived from erosion of the Ogallala Formation and older rock units. Locally derived cobbles and gravel commonly are found near the base of the alluvial deposits. Caliche and mortar beds also may occur within these unconsolidated deposits. Maximum thickness of the Quaternary alluvial deposits is about 550 feet in southwest Kansas (Gutentag and others, 1981) and about 85 feet in northwest and most of west-central Kansas. Undifferentiated deposits of Quaternary age are up to 200-feet thick in a north-south trending trough in central Scott County, west-central Kansas (Gutentag and Stullken, 1976). Most of the Cenozoic rocks in south-central Kansas are Quaternary alluvial deposits with a maximum thickness of about 300 feet (Fader and Stullken, 1978). Maximum thickness of the Holocene alluvial deposits is about 80 feet.

Eolian deposits of silt (loess) and sand of Pleistocene age mantle most of the upland areas in northwest and west-central and parts of southwest and south-central Kansas. Maximum reported thickness of loess is about 90 feet in northwest Cheyenne County (Frye and Leonard, 1952). Loess from 20- to 40-feet thick underlies most upland areas in northwest and west-central Kansas. Maximum reported thickness of loess in southwest Kansas is 45 feet (Gutentag and others, 1981). Prill (1977) reported that loess from 5- to 30-feet thick underlies about 60 percent of an 11-county study area in southwest Kansas. The loess is a uniform, homogeneous, generally fossiliferous, calcareous, and massive silt of eolian origin. Sources of silt in the loess are thought to have been wind-swept flood-plain deposits from rapidly alluviating streams during the Pleistocene age (Frye and Leonard, 1952).

The dune sand consists of well-sorted, moderately to well-rounded, generally noncalcareous arkosic sand that occurs at the land surface on about 3,000 square miles in south-central and southwest Kansas (Frye and Leonard, 1952). The largest dune tracts occur south of the big bend in the Arkansas River in south-central Kansas and extend along the south side of the river to the Colorado State line. Dune tracts also are associated with the Cimarron River in southwest Kansas. The maximum reported thickness of dune sand in Kansas is about 75 feet (Gutentag and others, 1981). The sediment source for the sand dunes was the wind-blown flood-plain deposits from the rapidly alluviating streams that migrated to the north during Pleistocene time (Frye and Leonard, 1952).

Bedrock Underlying High Plains Aquifer

The rocks underlying the High Plains aquifer (fig. 3) are predominantly Upper Cretaceous shale and limestone north of the Arkansas River, except in south-central Kansas where they are Lower Cretaceous. In southwest Kansas, south of the Arkansas River, the bedrock consists of progressively older strata to the south and east--the Lower Cretaceous sandstones and shales, the Jurassic and Triassic sandstones and shales, and the Lower and Upper Permian red sandstones, siltstones, gypsum, and dolomite.

In the western one-half of south-central Kansas, the bedrock is principally Lower Cretaceous rocks. In the eastern one-half of south-central Kansas, the bedrock surface is formed on rocks of the Lower Permian Sumner and Nippewalla Groups.

Configuration

The geology of the pre-Ogallala rock has a significant effect on the configuration of the bedrock surface. The configuration of the bedrock surface (plate 1) is a composite of subaerial erosional surfaces of several ages (Merriam and Frye, 1954, p. 36). This surface also has been affected by structural movement and by subsidence associated with the solution of evaporites from Permian rocks (Gutentag and others, 1981, p. 14). The pre-Ogallala surface south of the Arkansas River has been modified by post-Ogallala erosion. This erosion removed most of the Ogallala Formation in south-central Kansas east of longitude 100° west and, in parts of southwest Kansas, cut into the underlying Mesozoic and Paleozoic rocks (Merriam, 1963, p. 32).

The general slope of the bedrock surface in northwest Kansas is to the east-northeast from about 3,800 feet above sea level in northwest Wallace County to about 1,900 feet above sea level in northwest Phillips County. The average slope of the surface is about 12.5 feet per mile (a gradient of 0.0024). The irregular bedrock surface in southwest Kansas between the Bear Creek and the Crooked Creek-Fowler faults generally slopes at about 13.5 feet per mile (a gradient of 0.0026) to the east-southeast from 3,500 feet above sea level near the Colorado State line in southwest Stanton County to about 2,000 feet above sea level near the town of Meade in Meade County. The bedrock surface in south-central Kansas slopes generally to the east-southeast from about 2,300 feet above sea level in western Pawnee County to about 1,200 feet above sea level in Sedgwick County, providing an average slope of about 8 feet per mile (a gradient of 0.0015).

Solution Features

The dissolution and removal of evaporitic minerals from Permian rocks and carbonates from Upper Cretaceous rocks have affected and continue to modify the configuration of the bedrock surface in Kansas. The Bear Creek and Crooked Creek-Fowler faults in southwest Kansas are attributed to dissolution of halite and gypsum from the Blaine Formation and Flowerpot Shale of the Lower Permian Nippewalla Group. Maximum displacement on the bedrock surface across these features is about 250 feet (Gutentag and others, 1981, p. 9; Holdoway, 1978). The Meade Salt Sink and the Jones Ranch Sink

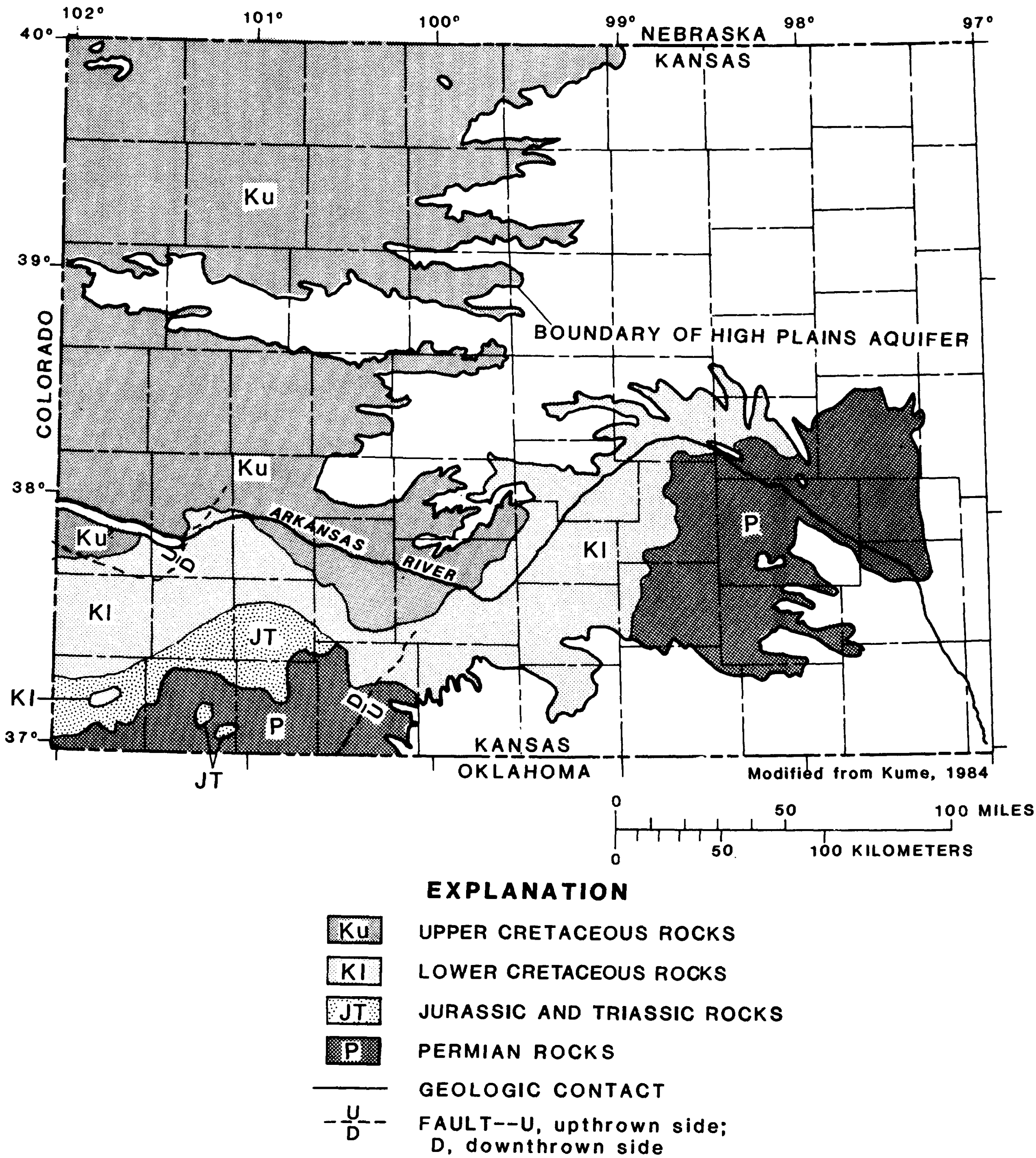


Figure 3.--Bedrock underlying High Plains aquifer.

in Meade County are believed to be associated with the Crooked Creek-Fowler fault. The Big Basin, the Little Basin, and the Ashland Basin in Clark County are associated with the dissolution of salt and gypsum from Permian rocks within 1,000 feet of the surface. The Coolidge Sink, which formed during 1926, is one of a linear series of sinkholes that occur along the Bear Creek fault.

An area of buried sinkholes occurs in central Ford County, southwest Kansas, and is indicated by hachured contours on plate 1. These buried sinkholes are believed to be an extension of the zone of solution associated with the Crooked Creek-Fowler fault. Other buried sinkholes are indicated by closed contours on a larger-scale map of the base of the High Plains aquifer prepared by Watts and Stullken (1981).

Some features on the bedrock surface previously identified as erosional are associated with the solution of evaporites from Permian rocks and the subsidence of overlying strata. The north-south trending buried valley that extends from central McPherson County to Sedgwick County was formed during the Pleistocene age. The underlying bedrock is the Wellington Formation of the Lower Permian Sumner Group. The buried valley is flanked on the west by a north-south line of sinkholes on the surface that are thought to coincide with the eastern edge of salt beds in the Wellington (Williams and Lohman, 1949). The north-south trending trough in central Scott County, west-central Kansas, is identified as a Pleistocene erosional feature (Gutentag and Stullken, 1976). Borehole geophysical logs of oil and gas wells in this area indicate that 200 to 300 feet of evaporites have been dissolved from the Lower Permian Nippewalla Group, which is separated from the Pleistocene rocks by about 1,500 feet of Permian and Cretaceous rocks. Thickness of the evaporitic rocks in the Nippewalla Group increases by 200 to 300 feet west of this feature.

Solution features formed in carbonate rocks of Late Cretaceous age have had a less dramatic impact on the configuration of the bedrock surface. Some irrigation wells in northwest Finney, Kearny, and Scott Counties, western Kansas, are completed in limestone and chalk of the Niobrara Chalk. These wells, locally known as "crack wells," extract water from fractures and solution cavities thought to be hydraulically connected to the overlying High Plains aquifer. A badlands topography is indicated on the bedrock surface in southwest Wallace County near the Colorado State line. This surface may be the result of subaerial erosion, but the formation of sinkholes during historical times in the Cretaceous rocks now exposed in the Smoky Hill River valley indicates the possibility of solution collapse as a contributing factor in the shaping of the bedrock surface.

High Plains Aquifer

The High Plains aquifer, which consists of "one or more hydraulically connected geologic units of late Tertiary or Quaternary age" (Gutentag and Weeks, 1980), is considered to be a single hydrologic unit. In Kansas, the included geologic units are, in ascending stratigraphic order, the Ogallala Formation of late Tertiary age and the alluvial and eolian deposits of Quaternary age. The geologic framework of the aquifer consists of a heterogeneous sequence of clay, silt, sand, and gravel that is predominantly of alluvial origin. Dune sands deposited during late Quaternary time in south-central Kansas also are considered a part of the aquifer because they are hydraulically connected to underlying deposits. Unsaturated Upper Quaternary loess and dune sands overlie the aquifer in most of western Kansas and affect recharge to the aquifer.

North of the Arkansas River in west-central and northwest Kansas the High Plains aquifer consists principally of the Ogallala Formation. Quaternary alluvium generally is restricted to stream valleys and to the north-south trending trough in central Scott County. The low-permeability rocks underlying the base of the aquifer consist of Upper Cretaceous limestones and shales and are considered to be a no-flow boundary. The aquifer is overlain by unsaturated loess from 20- to 30-feet thick over most of this area.

In southwest Kansas, the aquifer includes the Ogallala Formation, which was deeply eroded during the Quaternary age, and overlying thick Quaternary alluvial deposits. The bedrock at the base of the aquifer includes rocks from the Upper and Lower Cretaceous Series, the Jurassic, Triassic, and Permian Systems. The High Plains aquifer and the Lower Cretaceous and Jurassic and Triassic sandstones are hydraulically connected where they subcrop. The aquifer is overlain by unsaturated eolian deposits in most upland areas. Loess covers about 60 percent of the upland area in southwest Kansas, and dune sand covers about 20 percent.

In south-central Kansas, the aquifer consists principally of alluvial deposits of Quaternary age and erosional remnants of the Ogallala Formation. Upper Quaternary dune sand overlying the alluvial deposits also is considered to be part of the High Plains aquifer. The bedrock at the base of the aquifer consists of Lower Cretaceous sandstone and shale and Upper and Lower Permian rocks of various lithologies. Hydraulic connection occurs between the permeable bedrock units and the High Plains aquifer in some areas, which is indicated by the presence of highly mineralized water near the base of the High Plains aquifer. The Upper Quaternary eolian deposits, which overlie the aquifer, are unsaturated and consist of dune sand. Recharge to the aquifer is significant in the areas covered by dune sands. Although loess is present in parts of this area, it is relatively thin when compared to the loess in southwest, west-central, and northwest Kansas.

HYDROLOGY

Hydrologic Setting

The High Plains aquifer is a reservoir that collects and stores water from precipitation, infiltration of streamflows, underground flow from adjacent areas or aquifers, and seepage of irrigation return flows. Water that infiltrates the soil to a depth below the root zone of plants and is in excess of the field capacity (the maximum amount of water a soil will hold before allowing downward percolation) of the soil continues to percolate downward through the unsaturated part of the aquifer until reaching the water table. Water in the High Plains aquifer discharges either by evaporation and evapotranspiration, through springs and streams, or by withdrawals from wells.

The depositional history of the alluvial deposits of Tertiary and Quaternary age, which comprise the High Plains aquifer, govern the water-bearing characteristics of the deposits. There is good evidence from modern streams, when compared to ancient streams, to indicate that braiding was an important factor in the deposition of the High Plains aquifer. The major identifying feature of braided streams is the occurrence of both the coarsening- and fining-upward of alluvial deposits (Gutentag and others, 1984). This process of coarsening and fining of the alluvial deposits gives a random distribution of sediments in the High Plains aquifer, suggesting that the aquifer is homogeneous on a regional scale. Test holes drilled within a 160-acre tract often show a predominance of clays and silts at one site and of sand and gravel nearby. Commonly, several test holes need to be drilled in a 160-acre tract prior to selecting a location for a large-capacity production well.

The High Plains aquifer is generally considered to be an unconfined aquifer, although some localized areas do contain confined ground water. One example of this is in Meade County west of the Crooked Creek-Fowler fault in an area of about 40 square miles. Here, flowing wells and springs have occurred (Frye, 1942).

Hydraulic Properties of High Plains Aquifer

Hydraulic properties of an aquifer include hydraulic conductivity, transmissivity, storage coefficient or specific yield, and saturated thickness. Hydraulic conductivity is defined by Lohman and others (1972, p. 4) as "...the volume of water at the existing kinematic viscosity that will move in unit time under a unit hydraulic gradient through a unit area measured at right angles to the direction of flow," and is given in units of length per time (L/T). Transmissivity is defined as "...the rate at which water of the prevailing kinematic viscosity is transmitted through a unit width of the aquifer under a unit hydraulic gradient... ." "It is equal to an integration of the hydraulic conductivities across the saturated part of the aquifer perpendicular to flow paths" (Lohman and others, 1972, p. 13). The storage coefficient is defined as "the volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head...in an unconfined water body the storage coefficient is virtually equivalent to specific yield" (Lohman and others, 1972, p. 13). Specific yield is defined as "...the ratio of (1) the volume of water which the rock or soil, after being saturated, will yield by gravity to (2) the volume of rock or soil" (Lohman and others, 1972, p. 12). Saturated thickness of an aquifer is the vertical thickness of that part of the aquifer where the interstices are filled with water.

Results of 145 aquifer tests in western and south-central Kansas were examined and are listed in table 9 at the end of this report. Hydraulic properties of the High Plains aquifer are summarized in table 2. If hydraulic conductivity was not reported, it was calculated by dividing the reported transmissivity by the reported or estimated saturated thickness of sand and gravel. Hydraulic conductivity of the High Plains aquifer in Kansas, as determined by previous investigators (table 9), ranged from 12

Table 2.--*Summary of hydraulic properties for High Plains aquifer in Kansas*

[Results for 11 wells, which partially penetrated the aquifer, were excluded from the summary]

Hydraulic property[1]	Sample size	Mean	Standard deviation	Minimum	Maximum
Hydraulic conductivity (feet per day)	119	190	260	20	1,600
Transmissivity (square feet per day)	124	15,500	14,400	1,400	78,900
Saturated or effective thickness (feet)	117	131	88	17	350

[1] R. W. Stallman (U.S. Geological Survey, oral commun., 1966) reviewed the aquifer-test results in southwestern Kansas and suggested that some of the observation wells may have been installed improperly. The observation wells installed during the early 1960's used 3-foot screens opposite only the thickest coarse-grained material. The pumped wells were installed open to all water-yielding material in the saturated section, and then the borehole was gravel packed. Stallman indicated that the aquifer-test results using a 3-foot screen only represented the screened interval and not the entire aquifer. Most tests using observation wells open to the total aquifer showed unconfined values for the storage coefficient; however the tests listed in table 9 at the end of this report, using partially screened observation wells, indicate confined values for the storage coefficients. Because of these deficiencies, the test values of storage coefficient were not included in table 2.

to 1,600 feet per day. Transmissivity of the aquifer ranged from 1,400 to 78,900 square feet per day. The reported storage coefficient ranged from 0.00011 to 0.22, indicating a range from confined to unconfined conditions. However, as R. W. Stallman (U.S. Geological Survey, oral commun., 1966) indicated, most of the test results were from observation wells that may have been improperly installed, as discussed in the footnote of table 2. The reported saturated or effective thicknesses for these aquifer tests ranged from 17 to 350 feet.

Because the number of available aquifer-test data and their distributions were insufficient to define spatial variation in hydraulic properties, drillers' lithologic logs, which are plentiful, were used to estimate the hydraulic conductivity and specific yield of the High Plains aquifer in Kansas. Although drillers' logs are a highly subjective source of data, there are enough well-described logs for an adequate spatial distribution of estimate. A random sample of 1,612 well logs was chosen from a total population of about 24,000 well logs located in the study area.

A principal objective of the High Plains aquifer study was to develop a computer model to simulate the aquifer. Before a computer model was developed it was necessary to determine whether the vertical distribution of sediment types within the aquifer is random or non-random. A method was developed by Gutentag and Weeks (1981) to evaluate vertical variability using estimates of hydraulic conductivity and specific yield based on drillers' logs. The method requires statistical computations that describe the vertical distribution of aquifer characteristics determined from drillers' logs.

The method consists of calculating the center of gravity and the variance of the distribution of transmissivity (product of hydraulic conductivity and thickness) and storage depth (product of specific yield and thickness) relative to the base of aquifer (Gutentag and others, 1984). The center of gravity indicates the position of the centroid of distribution of the aquifer characteristic within the geologic section; the variance is a measure of variability of the aquifer characteristic about the center of gravity. Thus, a center of gravity of 0.5 (one-half the saturated thickness) indicates that the aquifer characteristic is evenly distributed about the centroid of the vertical section. If the mean center of gravity determined from a group of drillers' logs is not significantly different from 0.5, then the aquifer characteristic (and the sediments that control the characteristic) is distributed randomly in the vertical section. The statistics on the vertical distribution of transmissivity for the 1,612 drillers' logs in Kansas indicated a mean center of gravity of 0.48 and a standard deviation of 0.10; the mean center of gravity for storage depth was 0.48, with a standard deviation of 0.07 (see table 4). The centers of gravity were considered as hydrologically significant of a random distribution where all sediment types are equally likely to occur at any position in the vertical section.

Based on these results, it was concluded that the sediments that comprise the aquifer are distributed randomly in the vertical section. Therefore, the aquifer can be modeled in two dimensions using vertically arranged values of hydraulic conductivity and specific yield.

Hydraulic-conductivity values were assigned to lithologic descriptions from drillers' logs based on data listed in table 3. The hydraulic-conductivity data were modified from a table of values prepared for the High Plains aquifer in Nebraska (Lappala, 1978). The values are for commonly described lithologic terms used by drillers on the High Plains. Differences between a geologist's lithologic descriptions and typical drillers' descriptions were compensated for in table 3. Similar lithologic types can be described differently by different drillers, so that in table 3 similar hydraulic-conductivity values are shown for multiple-descriptions of lithologic terms. The specific-yield values were derived from aquifer-test data in a study by Johnson (1967). The hydraulic characteristics shown in table 3 were listed for use on the High Plains by Gutentag and others (1984).

Table 3.--*Driller's lithologic descriptions and assigned values of hydraulic conductivity and specific yield*

[Modified from Gutentag and others, 1984]

Driller's lithologic description	Assigned hydraulic conductivity (feet per day)	Assigned specific yield (dimensionless)
Clay	5	0.03
Silty clay	5	.03
Sandy clay	10	.05
Clay with sand and gravel	15	.08
Clay and sandstone	10	.05
Limestone (caliche)	10	.05
Limestone and sand	30	.10
Sand	70	.23
Fine sand	50	.21
Fine-medium sand	70	.23
Fine-coarse sand	80	.25
Medium sand	80	.25
Medium-coarse sand	85	.25
Coarse sand	90	.24
Clayey sand	50	.20
Cemented sand or loose sandstone	30	.10
Silty sand	50	.20
Sand and gravel	160	.25
Cemented sand and gravel	30	.10
Clayey sand and gravel	100	.17
Silty sand and gravel	100	.17
Sandstone and sand	50	.20
Tight sandstone	10	.05
Silt	10	.05
Clayey silt	5	.03
Sandy silt	10	.05
Silt with gravel	15	.08
Soil, overburden, and road-fill	5	.03

The lithologic descriptions, thicknesses, and assigned values of hydraulic conductivity and specific yield, well location, depths to water and the base of the aquifer, and production data (discharge rate and pumping water levels) when available were coded for processing with a computer program (Gutentag and others, 1981). The printed results from the computer analyses included a listing of the data, the thickness-weighted averages of the hydraulic-conductivity and specific-yield estimates, the estimated transmissivity from lithologic interpretation and (when applicable) from production data, and statistical analyses of the relative vertical distribution (first moment or center-of-gravity) and variance (second moment) of the estimated transmissivity and specific yield.

The thickness-weighted averages of hydraulic conductivity and specific yield were calculated according to the following general equation, equation 1:

$$XBAR = [X_i \cdot d_i + X_{(i+1)} \cdot d_{(i+1)} + \ldots + X_n \cdot d_n]/D, \tag{1}$$

where XBAR = the weighted average value of the Xi's;
X = the assigned value of hydraulic conductivity (or specific yield) of interval i;
i = the interval number, from 1 to n;
d = the saturated thickness of interval i; and
D = the total saturated thickness.

The relative vertical distribution (center-of-gravity) of transmissivity and specific yield can be used to define vertical trends in the aquifer. A program was developed to test for presence of a basal layer of high permeability (basal gravel) indicated by a center-of-gravity offset towards the bottom of the saturated thickness. For most of the logs used in this investigation, the center-of-gravity of estimated transmissivity was near the middle of the saturated thickness, indicating that the estimated transmissivity of the High Plains aquifer in Kansas is uniformly or randomly distributed in the vertical dimension.

A summary of the lithologic estimates of hydraulic conductivity, specific yield, and transmissivity for the High Plains aquifer from 1,612 lithologic descriptions is given in table 4. The total sample and discrete subsets (which represent different areas within the study area) indicate that there is little variation in the hydraulic properties of the aquifer among these areas. Hydraulic-conductivity estimates tend to increase to the east and south corresponding with a change in geologic units among the areas represented by the subsets. The aquifer in northwest and west-central Kansas is predominantly the Ogallala Formation of Tertiary age, with thin alluvial deposits of Quaternary age along major stream valleys. In southwest Kansas (including Ford County), the aquifer generally consists of the Ogallala Formation overlain by thick Quaternary alluvial deposits, and in south-central Kansas (excluding Ford County), the aquifer is predominantly alluvial deposits of Quaternary age.

Estimates of the transmissivity of the aquifer in Sherman County, northwest Kansas, were made from an analysis of specific-capacity data from 136 large-capacity production wells (discharge rate greater than 100

Table 4.--*Summary of lithologic estimates of hydraulic conductivity, specific yield, and transmissivity for High Plains aquifer in Kansas*

Area or subset	Geologic source	Sample size	Hydraulic conductivity (feet per day)		Specific yield (dimensionless)				Transmissivity (square feet per day)			
			Mean	Standard deviation	Mean	Standard deviation	Mean center of gravity	Standard deviation of mean center of gravity	Mean	Standard deviation	Mean center of gravity	Standard deviation of mean center of gravity
South-central	Quaternary	324	83.2	36.8	0.167	0.0509	0.480	0.0741	8,018	6,382	0.469	0.1104
South-west	Tertiary-Quaternary	437	79.6	38.7	.167	.0495	.494	.0752	12,526	10,093	.494	.1065
West-central	Tertiary	133	70.0	34.5	.170	.0502	.480	.0785	4,030	3,556	.468	.1025
North-west	do.	359	69.5	33.2	.167	.0493	.479	.0747	4,485	3,981	.470	.1082
Sherman County	do.	359[1]/	66.3	24.0	.156	.0329	.473	.0466	9,083	4,040	.480	.0750
Total area	Tertiary-Quaternary	1,612	74.3	34.5	.165	.0468	.482	.0702	8,361	7,325	.478	.1017

1 The 359 sites for Sherman County were not included in the sample for northwest Kansas nor were any of the sites for Sherman County in the northwest Kansas sample included in the Sherman County sample.

gallons per minute). The estimation method balanced equation 139 of Lohman (1972, p. 52) for transmissivity and included effects of the length of the test, radius of the well, an assumed storage coefficient equal to the lithologic estimate of specific yield, and Jacob's correction to drawdown (Jacob, 1963). Only tests for wells that fully penetrated the aquifer were analyzed. Well losses were not considered. The transmissivity estimates then were divided by the saturated thickness to provide an estimate of the average hydraulic conductivity of the aquifer. The average value of the 136 sites for the lithologic-log estimates of hydraulic conductivity was about 69 feet per day and for the specific-capacity estimates of hydraulic conductivity was about 18 feet per day, a ratio of about 4 to 1.

Estimates of vertical hydraulic conductivity were based on the same lithologic data used to estimate the horizontal hydraulic conductivity of the saturated part of the aquifer in Sherman County. The effective vertical hydraulic conductivity may be estimated with a modification of equation (1), where the total saturated thickness is divided by the sum of the individual quotients of each lithologic thickness divided by its representative hydraulic conductivity as described by Freeze and Cherry (1979, p. 34, eq. 2.31). The mean value for estimated vertical hydraulic conductivity at the 136 sites in Sherman County was about 25 feet per day. The ratio of the means of estimated vertical hydraulic conductivity to estimated horizontal hydraulic conductivity is about 1 to 3.

Spatial variation of hydraulic conductivity and specific yield of the aquifer in the study area also was based on estimates made from the interpretation of 1,612 drillers' logs for wells and from logs of test holes drilled by the Kansas Geological Survey and the U.S. Geological Survey for previous investigations. Because of the size of the area for which the estimates were to be made, the spatial density of the data initially was picked at about one log for every 25 square miles. Neighboring estimates did not follow a uniform trend as desired for mapping. A one-county area was selected to test the effect of a greater density of data on the spatial correlation between estimates. A total of 359 logs (an average of about one log for every 3 square miles) was selected from Sherman County data to make estimates of the average hydraulic conductivity and specific yield. In many cases the distance between sites (wells and test holes) in the one-county area was less than 1 mile.

The data subsets for northwest Kansas and Sherman County were analyzed with a computer program written by Skrivan and Karlinger (1980), which is based on the theory of regionalized variables (kriging) described by Matheron (1971). The variance of the lithologic estimates of hydraulic conductivity was not a regionalized variable (spatially correlated) in either subset. This implies that the thickness-weighted averages of hydraulic conductivity are from a random "normal" distribution. Since no spatial correlation could be established for the Sherman County subset, which had a high density of data points, it is likely that selecting a greater number of logs would not improve the overall estimate of the hydraulic conductivity of the aquifer or result in a significant definition of spatial variability.

Spatial variation in hydraulic conductivity of the High Plains aquifer, as estimated from lithologic logs, is shown in figure 4. Estimates for about 1,200 sites were used in preparation of this map. The interval patterns are indicative of the relative spatial variation in hydraulic conductivity; the aquifer consists of deposits associated with braided streams, meandering rivers, and wind so there is a "...characteristic textural variability that causes much heterogeneity in the distribution of hydraulic properties" (Freeze and Cherry, 1979, p. 147).

Spatial variation in specific yield of the High Plains aquifer is shown in figure 5. The same lithologic logs (about 1,200) were used as in estimating hydraulic conductivity. The statistical summary of lithologic estimates given in table 4 shows that there is very little variability in the mean of estimates of specific yield from one area to the next.

In an unconfined aquifer, specific yield corresponds to storage coefficients determined from aquifer tests of relatively long duration where gravity drainage is complete. Specific yield is a major factor in evaluating the production life of an aquifer. Deposits of finer grained materials, which have small specific yields, would show more rapid declines in water levels for a given draught than would coarse-grained deposits. Referring to tables 3 and 4, the mean specific yield for the High Plains aquifer corresponds to a driller's description of clayey or silty sand and gravel.

Definition of the High Plains aquifer using terms of homogeneity and isotropy is somewhat dependent upon the scale of the investigation and the intended use of the definition. The aquifer is composed of a random sequence of alluvial sediments, which imparts heterogeneity to the aquifer. Lateral and vertical trends of similar recurring lithologic types were not indicated at a regional scale from lithologic data examined during this study. This type of random heterogeneity may be classified as homogeneous according to the definition of Greenkorn and Kessler (1969) as cited in Freeze and Cherry (1979, p. 31): "They redefine a homogeneous formation as one in which the probability density function of hydraulic conductivity is monomodal. That is, it shows variation in hydraulic conductivity but maintains a constant mean hydraulic conductivity through space."

Recharge to High Plains Aquifer

Sources of recharge to the High Plains aquifer are precipitation, infiltration from streamflow, horizontal ground-water movement into the aquifer from outside areas, vertical ground-water movement into the aquifer from adjacent aquifers, and seepage from irrigation return flows. Definitive values for recharge throughout the High Plains aquifer are not available. In many ground-water studies, this study included, recharge is considered as one of the lesser known quantities in the hydrologic system.

Precipitation

Recharge to the High Plains aquifer is predominantly from precipitation. Water from precipitation recharges the aquifer by infiltrating

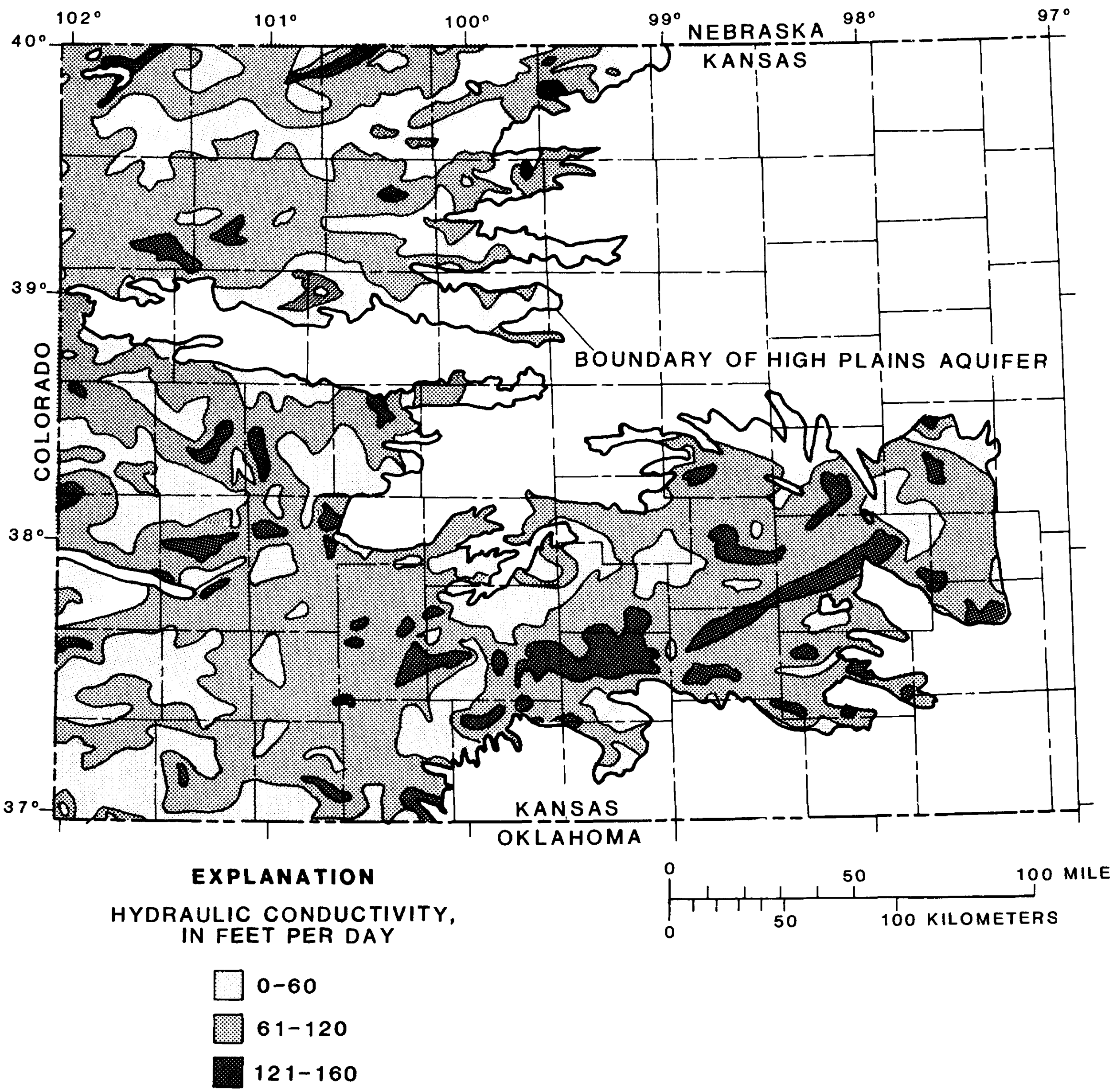


Figure 4.--Estimated hydraulic conductivity of High Plains aquifer, based on interpretation of lithologic logs.

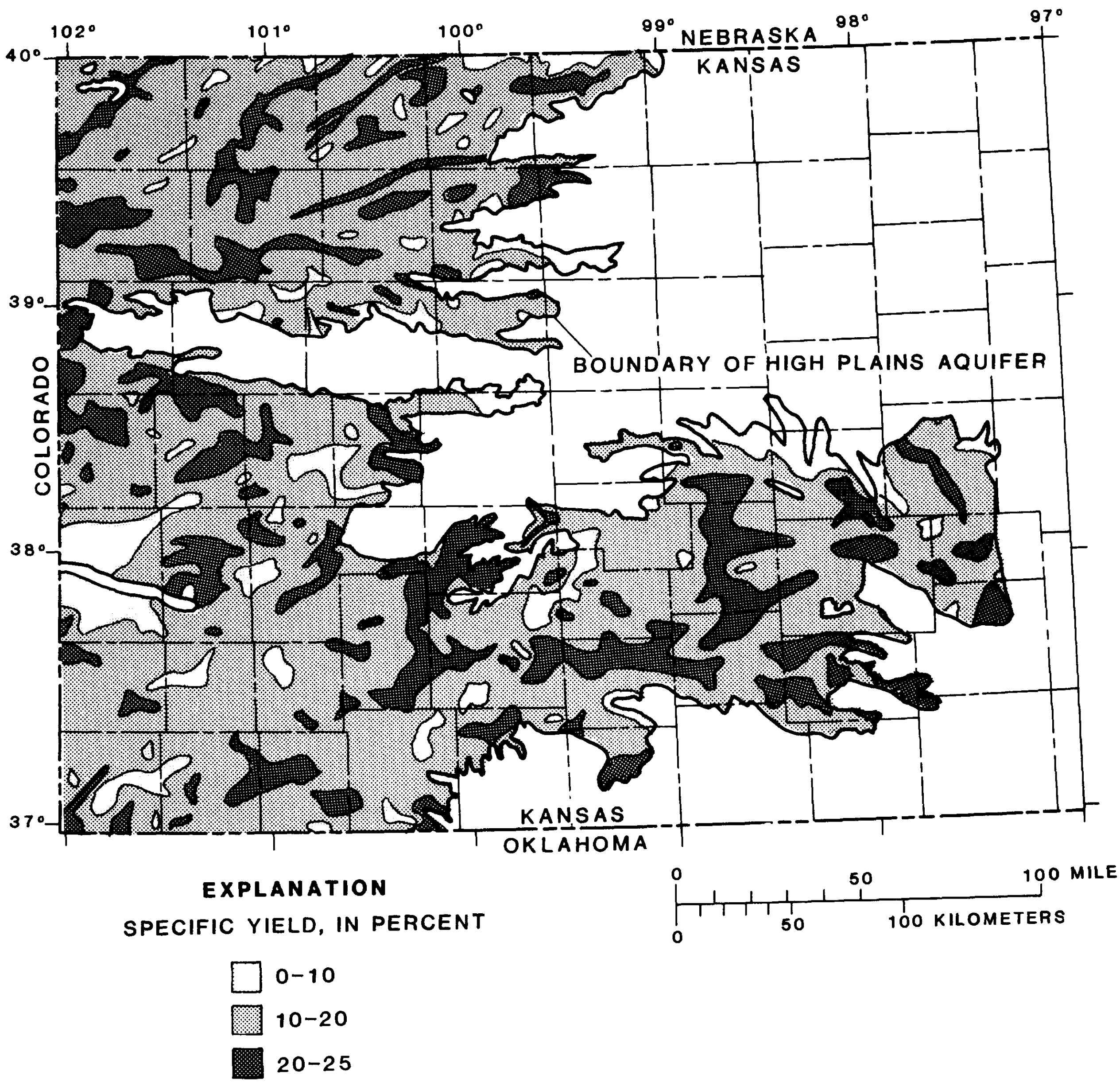


Figure 5.--Estimated specific yield of High Plains aquifer, based on interpretation of lithologic logs.

the soil, which increases the soil's moisture content beyond its field (holding) capacity. The excess moisture then percolates downward to the aquifer. Sandy areas where soils are thin and vegetation is sparse afford greater opportunity for recharge than do areas where soils are thick and vegetation is abundant.

Eolian deposits of late Quaternary age overlie the High Plains aquifer in most of the study area. Though generally unsaturated, these deposits greatly affect the amount of recharge that reaches the aquifer. Loess, which covers most upland areas in western Kansas, has a large field capacity of about 26 to 29 percent by weight (Meyer and others, 1953). Thick deposits of unsaturated loess, though capable of absorbing large volumes of water, will significantly reduce the quantity of recharge leaking downward because of their large field capacity. Dune sands, which cover about 3,000 square miles of south-central and southwest Kansas, generally are highly permeable and have small values of field capacity. Dune sand from Finney County is reported to have a field capacity of about 6 percent by weight (Meyer and others, 1970, p. 32). The downward leakage of recharge through dune sand is much greater than for loess under natural conditions.

In northwest Kansas, Jenkins and Pabst (1975, p. 19) estimated recharge from precipitation at 0.25 inch per year. A much lower estimate of long-term precipitation recharge of 0.05 inch per year was calculated by Meyer and others (1970, p. 84) for loessial soil in Finney County, southwest Kansas. Williams and Lohman (1949, p. 129) estimated recharge from precipitation on sandy soils in south-central Kansas to be about 6 inches per year.

Infiltration from Streamflow

Streams that flow across the area at an altitude greater than the water table lose a part of their flow to the aquifer by the same infiltration process as occurs with precipitation. Bear Creek, which flows into Kansas from Colorado, is an example of a stream that loses the majority of its flow to the aquifer. Except for extreme floods, Bear Creek flows generally are lost before they reach Kearny County. Other streams of varying size throughout the western part of Kansas lose water to the aquifer in the same manner.

Streamflow loss is best determined by operation of full-time gaging stations along the streams. Because such gages are expensive to operate and interpretation of streamflow measurements requires historical record, a more immediate method of determining streamflow volumes is desirable. A method of calculating mean annual flow from channel-geometry measurements has been devised in which mean annual flow is related to the width of the active channel (Hedman and Osterkamp, 1982). Measurements of channel geometry were made in those channels suspected of being significant contributors to the High Plains aquifer. These measurements provided estimates of recharge that were similar to those at gaged streams. Estimated recharge from streamflow losses in southwest Kansas ranged from 0.08 to 0.14 inch per year (E. R. Hedman, U.S. Geological Survey, written commun., 1982).

Ground-Water Inflow

Ground water flows into the study area from adjacent parts of the High Plains aquifer. The High Plains aquifer in Kansas extends into and is in hydraulic connection with the adjacent states of Colorado, Nebraska, and Oklahoma. Ground-water flow into Kansas occurs primarily along the Colorado-Kansas border in northwest and southwest Kansas. The High Plains aquifer does not extend westward to the Rocky Mountain front in Colorado and does not, therefore, receive water from the mountains except by streamflow.

Water in the aquifer flows in a downgradient direction perpendicular to the contours of the potentiometric surface and in a quantity proportional to the gradient of that surface, the hydraulic conductivity of the aquifer, and the cross-sectional area through which the water must pass (Darcy's law). Including 13,870 acre-feet per year of subsurface inflow from Arkansas River alluvium in Kearny County (Barker and others, 1983), the total ground-water inflow to the aquifer is estimated to be about 65,000 acre-feet per year.

Leakage from Adjacent Aquifers

The characteristics of the bedrock underlying the High Plains aquifer may significantly influence ground-water flow in the High Plains aquifer by allowing or inhibiting vertical flow between the High Plains aquifer and the bedrock aquifers. If hydraulic heads in the underlying aquifer are greater than those in the High Plains aquifer, water will attempt to move from the underlying aquifer to the overlying aquifer. Defining the quantity of flow requires a knowledge of the hydraulic-head differential and the average hydraulic conductivity of the zone through which water is flowing.

Rocks of Permian age in southwest and south-central Kansas are, in places, in hydraulic connection with the High Plains aquifer, as indicated by the presence of mineralized water in the bottom part of the High Plains aquifer (Gutentag and others, 1981; Fader and Stullken, 1978; Krothe and Oliver, 1982). The mineralized water leaves the High Plains aquifer by way of the Cimarron River (Gutentag and others, 1981, p. 43 and table 4) and several streams in the eastern part of the aquifer. The High Plains aquifer north of the Arkansas River rests on relatively impermeable Upper Cretaceous beds, and no leakage of significance occurs there.

Very limited information is available on hydraulic heads in, and the conductivities of, underlying formations. For this reason, leakage from or into the underlying aquifer is also a poorly defined quantity. In southwest Kansas, leakage to the underlying Lower Cretaceous rocks was determined by trial and error during modeling of the High Plains aquifer and is estimated to be 14,000 acre-feet per year. In south-central Kansas, Fader and Stullken (1978, p. 11) identified the Cedar Hills Sandstone as a source of natural leakage into the High Plains aquifer at an estimated 5,000 to 10,000 acre-feet per year. The Cedar Hills Sandstone is beneath and hydraulically connected to the High Plains aquifer along a north-south line at about 98° 45' W. longitude.

Irrigation Return Flow

Irrigation water in excess of crop requirements and soil-moisture holding capacity moves downward to the water table. Recharge from irrigation return flow is considered to be the difference between the pumpage and crop requirements. Values for irrigation return flow may exceed 50 percent of the pumpage where a flood-irrigation system is used on extremely permeable soil or may be as little as zero (for example, an irrigation system that is too small to supply crop requirements). Gutentag and Stullken (1976) calculated a water budget for Scott County using irrigation return flow as 20 percent of the withdrawal by wells. Irrigation practices of recent years (water scheduling based on soil moisture and sprinkler distribution) have reduced return flow significantly. Dunlap (1980, p. 5) reported very little irrigation return flow (about 3 percent) in a 12-square-mile study area in Wichita County. The amount of return flow, therefore, may vary widely with pumpage, location, and irrigating practices. Prior to 1950, irrigation return flow was an insignificant part of recharge to the High Plains aquifer because pumpage for irrigation was small.

Discharge from High Plains Aquifer

Discharges from the High Plains aquifer are evapotranspiration, seeps, springs, and base flow to streams, boundary outflow, vertical leakage to adjacent aquifers, and withdrawals by wells. These discharges are discussed below.

Evapotranspiration

Water may be discharged from the aquifer by the combined processes of evaporation and transpiration by plants. When considered as a single component, the process is called evapotranspiration. Considerable amounts of water can be lost by this process in areas where the water table is near the land surface. Very few plant species have a root structure that will penetrate the soil to depths of 10 feet in search of water; therefore in most upland areas where the depth to water is much greater than 10 feet, there is little or no evapotranspiration loss from the aquifer.

Discharge from the aquifer by evapotranspiration is negligible throughout most of the study area because the water table is too far below the land surface. In some areas of south-central Kansas the water table is near the land surface but the amount of evapotranspiration by plants from the aquifer is unknown.

Flow to Seeps, Springs, and Streams

Sustained or fair-weather streamflow is called base flow, and in most streams exiting the boundaries of the study area, it is almost entirely composed of discharge from the High Plains aquifer. For this discharge to occur, the land surface must be eroded to a level below that of the adjacent water table. When the hydraulic head in the High Plains aquifer is higher than the stage in the stream or the adjacent land surface, ground water emerges as base flow to streams or as a seep or spring. Gaging stations on streams and rivers throughout the study area record a highly variable streamflow. The ground-water or base-flow component of streamflow during spring to fall often is masked by large storm runoff and diminished by evapotranspiration from vegetation along the stream valleys.

All streams carrying discharge from the High Plains aquifer tend to flow eastward. Base-flow data from gaging stations near the eastern boundary of the aquifer represent, approximately, the ground-water discharge from the High Plains aquifer to streams. Winter records are presumed to be unaffected by evapotranspiration because vegetation is in a dormant stage. P. R. Jordan (U.S. Geological Survey, written commun., 1979) has estimated mean base flow for each winter period of record available for streamflow stations nearest the eastern periphery of the aquifer in Kansas (table 5). Because some of these stations are located beyond the edge of the aquifer and the estimates vary widely from year to year, these data were used only as guidelines for estimating the volume of streamflow discharge computed by the aquifer model discussed later in this report.

The estimated mean aquifer discharge (base flow) to streams in northwest Kansas was about 65 cubic feet per second (47,060 acre-feet per year), based on streamflow data for 1945-77 shown in table 5. In southwest Kansas, the Arkansas River, Cimarron River, and Crooked Creek were the only gaining streams leaving the aquifer boundaries during the period of record. Based on data shown in table 5, estimated mean discharge to the Cimarron River from the High Plains aquifer was about 55 cubic feet per second (39,820 acre-feet per year) and about 14 cubic feet per second (10,136 acre-feet per year) to Crooked Creek. Streamflow records on file with the U.S. Geological Survey in Lawrence, Kansas, indicate a 41 cubic feet per second (29,684 acre-feet per year) increase in mean flow from the Arkansas River at Garden City gage to the Arkansas River at Dodge City gage for 1946-50. Because this figure includes some storm runoff, the discharge from the aquifer to the Arkansas River was estimated to be less than 40 cubic feet per second (less than 28,960 acre-feet per year).

Ground-Water Outflow

Ground-water outflow can occur on those sides of the study area (High Plains aquifer in Kansas) that are not physical boundaries of the aquifer. Ground-water outflow occurs at the Kansas-Nebraska State line on the north, through the Arkansas River alluvium near Wichita on the east, and at the Kansas-Oklahoma State line on the south.

Ground-water outflow through the eastern boundary was estimated by Darcy's law to be 5,000 acre-feet per year. Flow through the northern boundary to Nebraska was estimated by modeling the aquifer to be 4,000 acre-feet per year and through the southern boundary to Oklahoma to be 33,000 acre-feet per year.

Table 5.--*Estimated mean winter base flow at selected streamflow-gaging stations, 1945-77*1/

[Average winter flow (January to April and October to December) is given in cubic feet per second]

Year	Streamflow-gaging station number2/								
	Northwest Kansas						Southwest Kansas		
	06846500	06848500	06863500	06871000	06871500	06873000	07156900	07157000	07157500
1945	--	3/2.1	--	--	--	3/22	--	39	12
1946	--	17	--	--	--	3/26	--	56	26
1947	9.8	16	6.8	--	--	18	--	55	19
1948	1.0	6.3	3.1	--	--	4.7	--	58	14
1949	11	11	13	--	--	3/28	--	69	26
1950	7.6	15	7.6	--	--	23	--	63	18
1951	15	25	14	--	9.7	46	--	70	17
1952	13	16	11	--	9.1	35	--	71	13
1953	1.0	9.6	3.6	14	5.2	15	--	58	9.5
1954	.1	6.6	1.9	11	3.0	11	--	58	11
1955	0	1.9	1.4	5.5	3.3	4.0	--	57	8.6
1956	0	1.2	1.5	3.6	2.2	6.4	--	48	6.6
1957	2.0	1.9	5.3	3.5	2.2	12	--	57	9.6
1958	9.2	6.2	9.6	11	6.0	37	--	57	13
1959	5.3	5.3	12	9.8	6.3	31	--	54	13
1960	10	7.5	15	15	3/9.5	3/38	--	54	11
1961	.5	4.4	16	7.0	5.7	32	--	58	13
1962	1.6	6.7	17	14	7.3	52	--	51	16
1963	8.5	12	10	9.3	5.2	30	--	45	3/14
1964	3.6	3/6.5	4.7	5.3	3.2	15	--	50	14

Table 5.--*Estimated mean winter base flow at selected streamflow-gaging stations, 1945-77*[1]--Continued

Year	Streamflow-gaging station number[2]								
	Northwest Kansas						Southwest Kansas		
	06846500	06848500	06863500	06871000	06871500	06873000	07156900	07157000	07157500
1965	[3]19	--	8.2	15	7.2	[3]20	--	[3]52	14
1966	[3]12	--	9.2	13	5.0	[3]17	69	--	13
1967	[3]1.9	--	3.9	6.4	3.6	5.8	67	--	11
1968	.9	--	3.7	[3]5.8	[3]3.8	7.2	59	--	13
1969	.3	--	6.5	[3]13	8.5	[3]9.8	62	--	17
1970	0	--	5.5	8.5	6.5	10	63	--	14
1971	0	--	5.3	4.6	[3]3.7	7.4	59	--	14
1972	.1	--	4.7	[3]3.1	[3]3.8	3.2	65	--	14
1973	.01	--	21	13	10	[3]30	63	--	24
1974	.9	--	24	12	11	23	53	--	18
1975	0	--	11	8.6	6.4	7.7	56	--	14
1976	0	--	12	7.4	4.6	11	--	--	[3]11
1977	0	--	8.5	[3]2.9	[3]2.8	[3]5.5	--	--	[3]9.3

1 P. R. Jordan, U.S. Geological Survey, written commun., 1979.
2 Streamflow-gaging stations and numbers are located on plate 1.
06846500 is Beaver Creek at Cedar Bluffs.
06848500 is Prairie Dog Creek near Woodruff.
06863500 is Big Creek near Hays.
06871000 is North Fork Solomon River near Glade.
06871500 is Bow Creek near Stockton.
06873000 is South Fork Solomon River above Webster Reservoir.
07156900 is Cimarron River near Forgan, Oklahoma.
07157000 is Cimarron River near Mocane, Oklahoma.
07157500 is Crooked Creek near Nye.
3 Average flow is based on partial year's record.

Vertical Leakage to Adjacent Aquifers

Vertical leakage from the High Plains aquifer to underlying aquifers occurs where the Lower Cretaceous sandstones and Permian formations subcrop under the High Plains aquifer. This includes much of the area south of latitude 38° S and east of longitude 99° W. Owing to the low-permeability bedrock formations (Upper Cretaceous) in northwest Kansas, there is virtually no vertical leakage to or through those formations. Chemical analyses of water from the Dakota Formation in areas near its subcropping indicate that ground-water outflow occurs from the High Plains aquifer to the Lower Cretaceous Dakota Formation in southern Finney County (Meyer and others, 1970, p. 105), Kearny County, and northern Haskell and Grant Counties.

Withdrawal by Wells

Large-capacity wells are defined as wells that withdraw water at a rate greater than 100 gallons per minute. In the High Plains, most of the large-capacity wells are used for irrigation and public supply. Industrial uses are minor, primarily consisting of livestock feeding and processing and as a water supply for an electric-generating plant.

Prior to 1950, large-capacity wells were few with little impact on the High Plains aquifer. Jenkins and Pabst (1975, p. 18) estimated that in northwest Kansas during 1950 there were only about 10,000 acres, mostly in stream valleys, irrigated by about 100 wells. Records of the Division of Water Resources, Kansas State Board of Agriculture, indicate that there were only about 420 irrigation wells in southwest Kansas in January 1946 (Gutentag and others, 1981, p. 58). The amount of water used by these wells was unreported. Irrigators were only beginning to develop ground water. Public supply (municipal) was the only major use of ground water at that time, but reported data were scant. A rate of 100 gallons per day per capita was used to estimate municipal use in northwest Kansas.

The Kansas Water Appropriation Act of 1945 provided that each ground- or surface-water user could submit a numbered application for a permit to beneficially use some specified amount of water (water right). Although these permits could cover more than one well prior to 1978, most do not, and they are one indicator of growth in the number of irrigation wells.

The number of large-capacity wells has increased significantly since 1950 (fig. 6). The rate of growth in number of wells and consequently the rate of growth in quantity of ground water used closely approximate the rate of growth in the number of applications.

Distribution of irrigation wells is of as much interest as the growth in numbers. A map showing the density of irrigation wells throughout the High Plains aquifer in Kansas as of 1980 is given in figure 7. Nine townships (36 square miles each) in southwest and west-central Kansas have from 109 to 144 irrigation wells, or an average of more than 3 wells per square mile. The heaviest development of irrigation centers around west-central and the northern part of southwest Kansas.

Estimates of pumpage from the entire High Plains aquifer during 1980 for irrigation were made by Heimes and Luckey (1983). They estimate that 4,215,000 acre-feet of ground water was pumped from the High Plains aquifer in Kansas during 1980. Data on irrigation demand and irrigated acreage for 5-year intervals from 1949-78 were aggregated in blocks of 10 minutes longitude by 10 minutes latitude and can be retrieved from the U.S. Geological Survey computer system as described in Luckey and Ferrigno (1982).

Changes in Ground-Water Storage

The altitude and configuration of the water table in the High Plains aquifer for 1980 is shown on plate 1. Water-level measurements were obtained during the winter (January 1980) when the effects of seasonal pumping for irrigation were at a minimum (Pabst, 1980). These water-level measurements were used to determine the altitude and configuration of the water table, to compare water-level changes from earlier years (Pabst and Stullken, 1981, 1982a, b, c; Stullken and Pabst, 1981, 1982), and to determine saturated thickness (thickness between water table and bedrock). Saturated thickness is as great as 600 feet near the southern border of southwest Kansas (Weeks and Gutentag, 1981; Luckey and others, 1981).

The water table represents the upper surface of the ground-water reservoir in the High Plains aquifer. The shape of this surface may be interpreted to provide knowledge of aquifer and ground-water flow characteristics. The water table generally slopes eastward across the study area at about 8 feet per mile. Ground water moves in a downgradient direction perpendicular to the contours of the water table. The velocity of ground-water movement probably is about 2 feet per day or less.

Widely spaced contours represent a flattening of the water table and, therefore, a reduced gradient. In southern Finney County, ground water

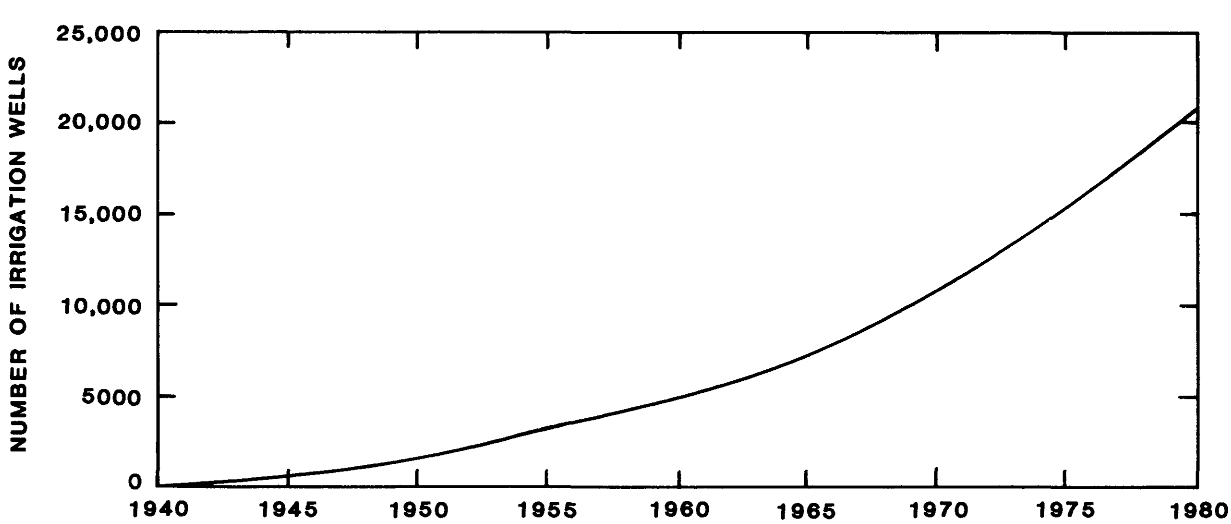


Figure 6.--Cumulative number of irrigation wells constructed, 1940-80.

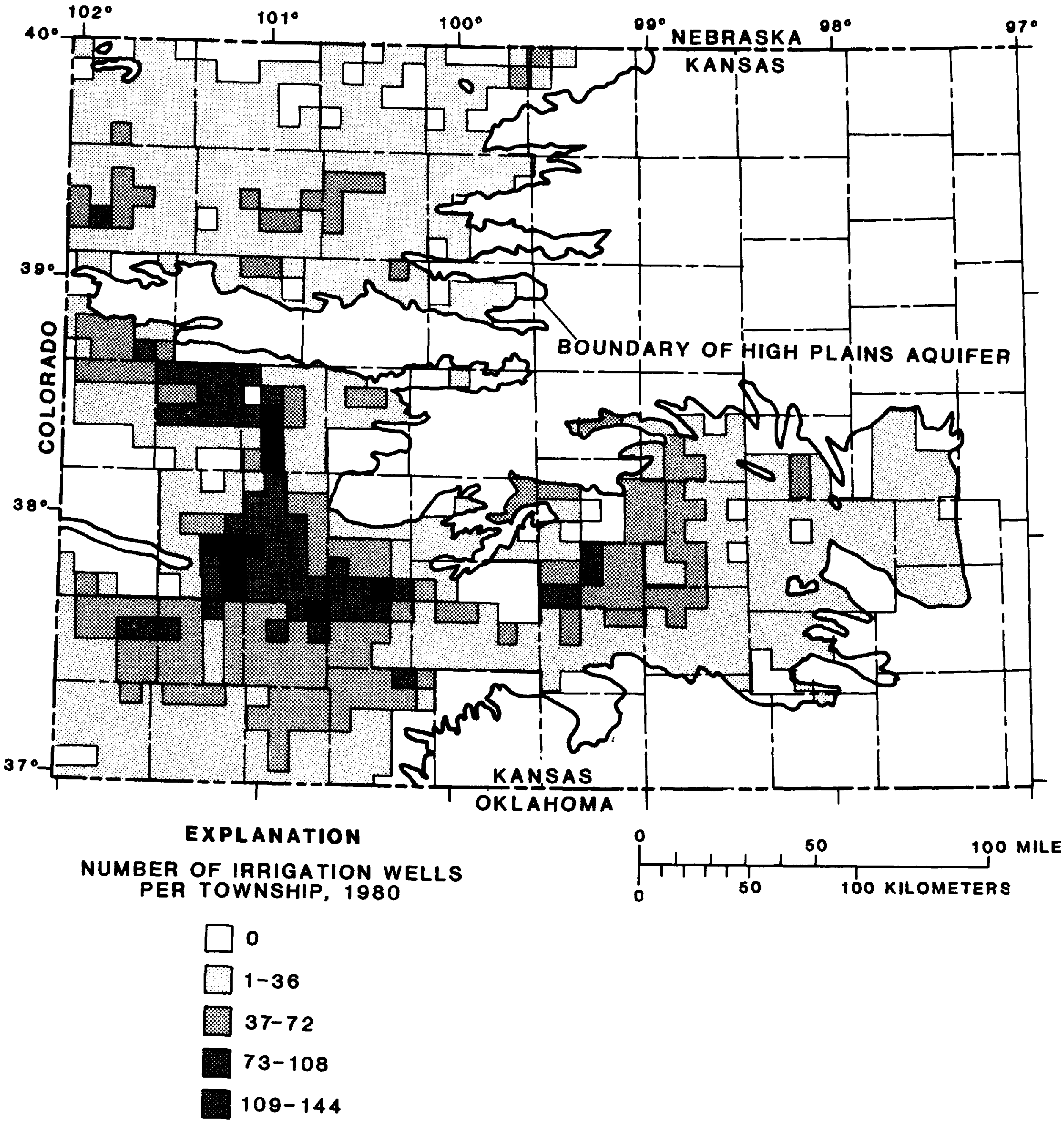


Figure 7.--Density of irrigation wells in High Plains aquifer, 1980.

moves into areas of greater saturated thickness that require less water-table gradient to maintain the flow rates. In eastern Scott County, the saturated thickness of the aquifer is less than the area upgradient from it, and the flattened water table indicates a reduction in ground-water flow.

Closely spaced contours represent a steepening of the water table and, therefore, an increased gradient for lateral ground-water movement. An example of this is the area in central Meade County where the aquifer thins rapidly (Crooked Creek-Fowler fault). The steepening gradient increases ground-water discharge through springs, seeps, into streams and artesian boils.

Upstream flexures in the water-table contours near a stream indicate that ground water is being supplied to a gaining stream, as along the Cimarron River in southeastern Seward County. Broad downgradient flexures of the water-table contours in the sand-dune areas of southern Edwards and Stafford Counties indicate areas of increased recharge. Centers of intensive withdrawals from the aquifer will distort the contours to an arcuate or circular pattern (coalesced cones of depression) around their location, as in central Finney County.

The shaded areas shown on plate 1 within the High Plains boundary, but not contoured, may contain some thin saturated thicknesses (1 to 10 feet). The thinly saturated water table in these areas follows bedrock contours very closely. In areas where the bedrock has some permeability, the water table may even be below the overlying unconsolidated deposits. In both cases, water-table contours are not representative of flow in the regional aquifer.

Water-level fluctuations in observation wells respond to seasonal climatic differences. In an area where all discharge from the aquifer is natural and steady (predevelopment conditions), the water table will fluctuate about a long-term average level. During times of drought, recharge is reduced, and the water table usually declines. During periods of high precipitation, recharge would be expected to be greater than discharge, and water levels will rise.

Representative hydrographs on plate 1 show fluctuations of ground-water levels at eight locations throughout the High Plains aquifer in Kansas. These fluctuations were determined by long-term measurements of water levels in wells penetrating and screened in the High Plains aquifer. Recharge to the aquifer is shown on the hydrographs as an increase in water level at the observation well. Except for extremely shallow water tables, water levels in observation wells do not respond to recharge as rapidly as they do to withdrawals. In areas where the water table is responding to nearby withdrawals, recharge may be evident only as a reduction in the amount of annual decline. Short-term fluctuations in the hydrograph are usually similar only in nearby wells, while long-term fluctuations in the water table may be observed over a broader area to indicate regional trends. The hydrographs on plate 1 representing the water table in south-central Kansas (Kiowa and Stafford County graphs) are the only ones that do not show a general trend of declining water levels.

Depths to water in much of the High Plains aquifer are great enough to lengthen and moderate the effects of seasonal climatic changes. More immediate responses to climate are produced by seasonal withdrawals for irrigation. In areas of withdrawal for irrigation, water levels around the wells will decline during pumping periods and recover during nonpumping periods. In a typical irrigation pattern of spring pre-watering and summer irrigation, the water table will commonly rise during the winter, undergo a small decline in the spring, attempt to recover (rise) during early summer, decline to an extremely low level in late summer, and then again attempt to recover to the previous high level during the next fall and winter.

Commonly however, pumpage exceeds recharge, and the water table never fully recovers. The result is a long-term decline in water levels. These long-term declines have been a major concern of water managers throughout the State. Local ground-water management districts have been established to manage this mining of ground water. Declines in excess of 150 feet are documented for wells in Grant and Stanton Counties (Pabst and Dague, 1984). The hydrograph on plate 1 representing southwest Kansas (Grant County, 30S-37W-20BC) indicates the severity of declines in this area. The declines represent "mining of ground water" (pumpage in excess of recharge) by existing irrigation development. The rate of decline relates primarily to such variable factors as climate and agricultural economics. The lithology of the aquifer in the position of the water table can also be a source of variability in the rate of decline. The extent of water-level declines is shown on a regional map by Luckey and others (1981).

Annual precipitation for the Garden City Experiment Station (T. 24 S., R. 32 W., Finney County) for 1940-80 is plotted in figure 8. Also shown are a hydrograph of ground-water levels in well 24S-32W-03DAC in Finney County and the hydrograph for the mean annual streamflow in the Arkansas River at Dodge City. Extended periods of drought occurred during 1952-56, 1959-64, and 1973-76. During each drought period, both ground-water levels and streamflow declined severely.

Periods of extremely high streamflow often correspond to times of above-average precipitation in the upstream drainage area and are the most obvious features on the streamflow hydrograph. Extended periods of low flow, however, correlate more directly with ground-water levels. Long-term trends in ground-water-level declines also can be seen in the low-flow part of the streamflow hydrograph after 1972. An interpretation of this would be that the reach of the Arkansas River upstream from Dodge City received water from the High Plains aquifer prior to 1973. As the water table declined, ground-water flow to the river also declined and eventually reversed, with water moving from the river to the aquifer.

Chemical Quality of Water

The principal use of water-quality information in this study was to determine the suitability of the water for the intended use. Chemical, physical, and bacterial characteristics were used to define the quality of the water and its suitability for agricultural, municipal, industrial, commercial, and domestic water supplies. Recommended concentrations of

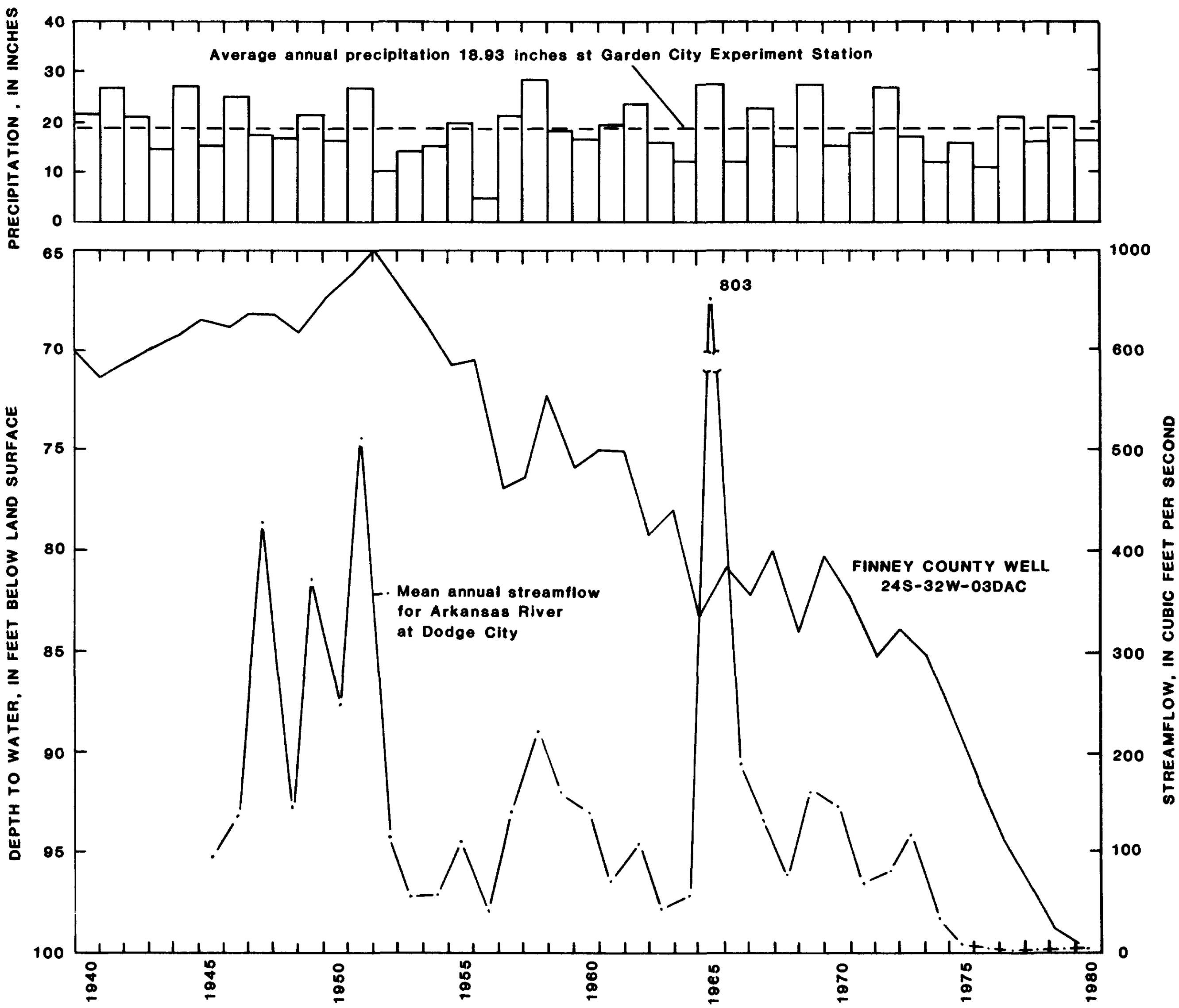


Figure 8.--Annual precipitation at Garden City Experiment Station, water levels in well 24S-32W-03DAC, and mean annual streamflow in Arkansas River at Dodge City, 1940-80.

dissolved solids and selected chemical constituents vary according to the intended use of the water. The U.S. Environmental Protection Agency (1976; 1977) has established maximum limits for concentrations of various chemical constituents in drinking water. Gough and others (1979) provide guidelines for the appraisal of toxicity hazards to plants, animals, and man of certain natural and manmade element concentrations that are of environmental concern. The suitability of ground water for irrigation was evaluated using criteria and procedures developed by the Salinity Laboratory Staff, U.S. Department of Agriculture (1954).

The chemical quality of irrigation water from the High Plains aquifer in Kansas has been described by Hathaway and others (1975; 1977; 1978a;

1978b; 1979), and Spruill (1983) has provided statistical summaries of selected chemical constituents in Kansas ground-water supplies, including those from the High Plains aquifer. Additional information about the quality of water from the High Plains aquifer in Kansas can be obtained from reports of areal studies by State and Federal agencies listed in the "References" at the end of the report. The bacterial quality of water from the High Plains aquifer in Kansas has not been well documented and is not discussed in this report.

A regional view of the chemical characteristics of water from the High Plains aquifer is given by Krothe and others (1982) and Feder and Krothe (1981). Sulfur isotope composition and chemistry of water from the High Plains aquifer in southwest Kansas and the Oklahoma Panhandle are described, and the implications of ground-water flow from underlying bedrock aquifers are discussed in a report by Krothe and Oliver (1982).

The chemical and physical quality of water from the High Plains aquifer is the result of complex geochemical, physical, and biologic interactions between the water and rocks and minerals in its flow path. The major sources of water entering the aquifer (recharge from precipitation, return flow from irrigation, streambed infiltration, and inflow from bedrock aquifers) also affect the quality of the ground water.

Precipitation, the most important source of recharge to the High Plains aquifer, may contain a wide range of dissolved substances. The chemical composition of precipitation may vary with time and geographic location. Total dissolved solids in rainwater typically range from 5 to 10 mg/L (milligrams per liter), which is very low compared to the dissolved solids in water from the High Plains aquifer. Precipitation is slightly acidic in nonurban and nonindustrialized areas. The hydrogen-ion activity (pH) of nonpolluted precipitation is about 5 to 6 (Drever, 1982).

The chemistry of the water changes as it comes in contact with the earth's surface. Much of the acidity in precipitation is neutralized, and as the water flows through the soil zone, it acquires solutes from dissolution of minerals. Some solutes, notably nitrogen and phosphate compounds, may be extracted from the water by plants. Water removed from the soil by evapotranspiration tends to concentrate dissolved solids in the soil, which result in mineral-enriched soil zones. The chemical composition of water infiltrating the soil and unsaturated zones is a function of the chemical composition of the rocks and minerals, the contact time with the soil and material in the unsaturated zones, and biologic and manmade influences.

There is very little information available on the chemical composition of return flow from irrigation in the High Plains of Kansas. Prill (1977) has reported changes in the chemical quality of water from the aquifer as a result of an artificial-recharge experiment. Concentrations of selected chemical constituents in water from the well used as a recharge source (well 23S-34W-26CCC) and in water from a shallow well at the edge of the recharge pond (well 23S-34W-26CDC) are given in table 6. The chemical composition of water from the shallow well was initially similar to the water from the deeper well used as the source of recharge water. Water from the shallow well collected at the end of a 157-day period after ponding and drainage shows a significant change in the chemical concentrations.

Table 6.--*Selected chemical analyses of ground and surface water from High Plains in Kansas*

[Values are in milligrams per liter, except as indicated; µS/cm = microsiemens per centimeter at 25 degrees Celsius (°C); NR = not reported]

Location number	Hydro-logic unit	Silica (SiO_2)	Calcium (Ca)	Magne-sium (Mg)	Sodium (Na)	Potas-sium (K)	Bicarb-onate (HCO_3)	Sulfate (SO_4)	Chlor-ide (Cl)	Fluor-ide (F)	Nitrate (NO_3)	Dis-solved solids	pH units	Temper-ature (°C)	Specific conductance (µS/cm)	Predominant Cation	Predominant Anion
23S 43W 26BAB1/	Arkansas River	13	380	200	580	14	NR	2,500	190	0.9	NR	4,290	8.1	19.0	4,700	Mixed	SO_4
23S 43W 26BAB2/	do.	9.4	150	75	180	6.9	NR	890	53	0.8	NR	1,540	8.2	22.0	1,840	Mixed	SO_4
23S 07W 01ABA3/	Quater-nary de-posits	18	61	9.4	251	2.8	288	62	284	.7	29	849	7.5	15.0	1,525	Na	Cl
23S 12W 25CDC3/	do.	26	98	8.4	102	3.6	256	36	168	.5	16	599	7.4	14.5	1,040	Mixed	Mixed
28S 17W 15DDB3/	do.	28	56	5.6	14	2.8	184	18	8.7	.4	20	234	7.5	17.0	365	Ca	HCO_3
34S 31W 30BBB	do.	25	70	17	41	3.2	212	77	55	.4	9.6	403	7.7	16.0	700	Ca	HCO_3
23S 34W 26CCC4/	Quater-nary-Ter-tiary deposits	20	250	94	123	11	181	932	113	.9	10	1,643	7.8	NR	2,130	Mixed	SO_4
23S 34W 26CDC5/	do.	7.9	222	107	146	12	154	920	167	1.2	.9	1,660	7.5	NR	2,200	Mixed	SO_4
23S 34W 26CDC6/	do.	22	482	348	240	20	185	2,320	417	1.1	24	3,960	8.0	NR	4,630	Mixed	SO_4
24S 33W 28DDA3/	do.	17	49	11	21	2.7	184	56	5.3	.8	5.3	260	7.5	18.0	560	Ca	HCO_3
28S 41W 31BDD3/	do.	14	48	16	14	4.3	158	61	10	1.1	25	261	7.6	18.5	430	Ca	HCO_3
35S 34W 10BBB3/	do.	30	100	37	61	4.2	198	76	200	.9	26	612	7.7	16.0	1,060	Ca	Cl
03S 28W 32BCA3/	Tertiary	62	46	15	19	8.1	225	21	9.9	0.9	6.5	287	7.9	16.0	430	Ca	HCO_3
08S 33W 18CCA3/	do.	48	40	19	24	6.1	218	24	11	1.5	14	179	7.6	15.0	445	Ca	HCO_3
08S 40W 18DBB3/	do.	53	26	11	40	4.0	203	20	3.9	2.1	9.5	262	8.1	16.0	380	Mixed	HCO_3
17S 35W 27CCC3/	do.	49	42	22	33	6.1	190	82	20	2.1	13	383	7.6	17.0	560	Mixed	HCO_3
34S 31W 30BBB	do.	11	150	50	670	12	95	200	1,300	1.1	2.9	2,420	8.1	NR	4,500	Na	Cl
19S 31W 16CBA	Niobrara Chalk	26	75	12	7/29	NR	254	40	29	.4	9.7	362	7.8	15.0	510	Ca	HCO_3
27S 33W 05CCC	Greenhorn Limestone	NR	87	20	15	NR	287	9.1	30	.3	53	358	NR	15.5	NR	Ca	HCO_3

Table 6.--*Selected chemical analyses of ground and surface water from High Plains in Kansas*--Continued

Location number	Hydrologic unit	Silica (SiO_2)	Calcium (Ca)	Magnesium (Mg)	Sodium (Na)	Potassium (K)	Bicarbonate (HCO_3)	Sulfate (SO_4)	Chloride (Cl)	Fluoride (F)	Nitrate (NO_3)	Dissolved solids	pH units	Temperature (°C)	Specific conductance (μS/cm)	Predominant Cation	Predominant Anion
16S 29W 11CCC	Undifferentiated Lower Cretaceous rocks	9.2	5.2	3.6	2/380	NR	734	39	110	8.0	0.7	932	NR	18.0	NR	Na	HCO_3
12S 34W 33BBD	Dakota Formation	9.2	5.6	.5	2/430	NR	722	0	230	6.0	.4	1,030	8.2	25.0	1,770	Na	HCO_3
20S 17W 22DCC3	do.	7	61	23	150	7	290	145	145	NR	.1	685	7.7	14.5	1,850	Na	Cl
23S 20W 16DDD	do.	9	42	24	79	NR	298	88	22	2.4	1.5	415	NR	15.0	NR	Na	HCO_3
24S 33W 19DBD8/	do.	14	43	12	43	2.8	NR	81	8.8	NR	.02	252	8.5	NR	NR	Mixed	?
20S 17W 22DCC2	Cheyenne Sandstone	9	796	709	17,400	60	349	3,700	28,200	3.2	.1	51,000	7.4	14	60,000	Na	Cl
34S 34W 17DDD	Jurassic-Triassic rocks	29	70	23	32	3.2	188	143	15	.5	15	423	8.4	16.7	660	Ca	Mixed
20S 17W 22DCC	Undifferentiated Permian rocks	11	776	838	16,200	64	327	3,360	26,200	3.2	.1	47,400	7.4	18.5	60,000	Na	Cl
34S 30W 27BBB	do.	10	1,500	470	10,800	38	78	2,800	18,200	1	.3	33,800	7.6	NR	57,300	Na	Cl
34S 31W 30BBB	Whitehorse Formation	33	1,300	640	5,000	320	140	1,100	11,000	.4	1.0	19,500	7.7	NR	NR	Na	Cl
32S 29W 27AAB	Big Basin Formation	8.3	320	120	87	4.8	66	1,400	16	.9	1.4	1,960	7.4	NR	1,300	Ca	SO_4
27S 11W 32ABA2	Salt Plain Formation	NR	820	340	13,100	66	93	3,410	20,300	NR	NR	38,000	7.2	NR	65,000	Na	Cl

1 Analysis of Arkansas River water at streamflow-gaging station 07137500 near Coolidge, May 24, 1982, (base flow during irrigation season).
2 Analysis of Arkansas River water at streamflow-gaging station 07137500 near Coolidge, July 6, 1982, (released from John Martin Reservoir, Colorado).
3 Data used for Stiff diagram in figure 9.
4 Recharge water from 309-foot deep well on November 20, 1967, at beginning of recharge test (Prill, 1977).
5 Initial water from 89-foot deep well at edge of recharge pond on November 20, 1967, at beginning of recharge test (Prill, 1977).
6 Final water from 89-foot deep well after 157-day drainage period (April 9, 1968) at end of recharge test (Prill, 1977).
7 Reported as sodium plus potassium (Na + K).
8 Analysis by Burns and McDonnell, Kansas City, Mo.

Water from the shallow well increased in dissolved solids from 1,660 mg/L on November 20, 1967, to 3,960 mg/L on April 9, 1968. The increase in concentrations of dissolved solids was caused by leaching of soluble salts from the soil and loess and depended on the quality of water applied, the extent of previous leaching, and the saturation percentage of the material through which the water was moving (Prill, 1977).

In southwest Kansas, a better quality water occurs in the aquifer under dune-covered areas south of the Arkansas River than occurs under the loess-covered areas north of the river (Hathaway and others, 1977). This difference may be a result of faster flow through the unsaturated zone and the lithology of the dune sand.

A major stream entering the study area that contributes significant amounts of recharge to the aquifer is the Arkansas River. Chemical analyses of water from the Arkansas River near Coolidge at U.S. Geological Survey gaging station 07137500 for the period from November 19, 1981, to August 30, 1982, showed that the water contained dissolved solids ranging from 1,360 to 4,370 mg/L and is a mixed cation sulfate type water (Geiger and others, 1983, p. 191). Selected chemical analyses of water from the Arkansas River at this location are given in table 6 for May 24 and July 6, 1982.

The quality of water in ephemeral streams of western Kansas is not documented. However, the chemical quality of water from the aquifer near the Cimarron River in parts of Morton and Grant Counties and near the terminii of Bear Creek in Kearny and Grant Counties and White Woman Creek in Scott County is similar to the type of water associated with the alluvium of the Arkansas River valley in Hamilton, Kearny, Finney, and Gray Counties. Salts, which have accumulated in the unsaturated zone during periods when there is insufficient moisture, are carried to the water table during intermittent periods of flooding.

Waters from some bedrock aquifers in Kansas have concentrations of dissolved solids greater than that of seawater, which is about 35,000 mg/L. Analyses of water from selected bedrock aquifers also are listed in table 6. Mixing of water from the bedrock aquifers degrades the quality of water in the lower parts of the High Plains aquifer in some areas of Meade and Seward Counties in southwest Kansas and the eastern part of south-central Kansas. The High Plains aquifer in these areas is hydraulically connected with rocks of the Permian System that contain soluble evaporites and brines. Solution of the evaporites from Permian rocks was discussed previously in the section on "Configuration of the Bedrock Surface." Krothe and Oliver (1982) state that the dissolved solids in water from the High Plains aquifer increase from west to east in southwest Kansas. Upward leakage of water has occurred naturally from the Permian Whitehorse Formation in southern Seward and Meade Counties into the overlying High Plains aquifer. The chemical characteristics of water from well 34S-31W-30BBB, drilled into the upper one-third (250-foot depth, Quaternary deposits) of the High Plains aquifer, the lower one-third (470-foot depth, Tertiary deposits, same site), and into the Whitehorse Formation (705-foot depth, same site) are included in table 6. Based on chloride concentrations, the water from the lower part of High Plains aquifer at this location is similar to a mix of 11 percent Whitehorse Formation water and 89 percent water from the Quaternary deposits at the 250-foot depth.

Mixing of water from permeable sandstones of the Lower Cretaceous Series and from the Jurassic and Triassic Systems with water from the High Plains aquifer probably occurs in some parts of southwestern and south-central Kansas. In northwest and west-central Kansas, where the High Plains aquifer overlies rocks of the Upper Cretaceous Series, the chemical quality of water from the High Plains aquifer is not appreciably affected by mixing of water from bedrock formations. Concentrations of fluoride in water obtained from wells located in this area range from 1.1 to 1.5 mg/L and exceed the recommended concentration levels in more than 30 percent of the samples (Spruill, 1983). Chemical analyses of water from pre-Tertiary formations in Kansas are given by Keene and Bayne (1977) for Lower Cretaceous rocks; by Kume (1984) for Jurassic and Lower Cretaceous rocks; and by Gutentag and others (1981) for Upper and Lower Cretaceous, Jurassic, and Permian rocks. Additional analyses are available in reports of local areas listed in the "References."

The chemical and physical properties of water from the High Plains aquifer have been determined at numerous sites in Kansas. Chemical analyses of water from selected wells completed in the High Plains aquifer are shown in table 6. These analyses were selected from tables in the referenced publications to be representative of waters from the aquifer. Since the water in the aquifer is the result of mixing from various sources and geochemical reactions occurring along the flow path, there is a wide variation in the type of water and the concentrations of dissolved constituents. However, some general characteristics of the aquifer's water quality are relatively constant for large areas within the study area. Dissolved solids in water from the aquifer generally increase from the west to the east, and the predominant ions change from calcium and bicarbonate to sodium and chloride (Krothe and others, 1982). Water from the High Plains aquifer in Kansas generally is suitable for human and animal consumption and irrigation of crops.

The concentrations of the major anions and cations in the water of the High Plains aquifer in Kansas are depicted by means of Stiff diagrams shown in figure 9. Stiff diagrams (Stiff, 1951) are used to depict the concentrations of the selected ionic species present in the water. The Stiff method uses four parallel horizontal axes extending on each side of a vertical zero axis. Concentrations of four cations, one on each axis, can be plotted on the left side of the zero axis, and likewise four anion concentrations can be plotted, one on each axis, to the right of the zero axis. The ions are always plotted in the same sequence. The end points of the horizontal axis are connected to form irregular polygonal shapes that are relatively distinctive. The width of the shape is an indication of the total ionic content of the water. Concentrations of the ions are given in milliequivalents per liter. The concentrations of ions for the selected analyses shown in table 6 are given in milligrams per liter and may be converted to milliequivalents per liter units, as used in figure 9, by dividing the milligram per liter concentration by the combining weight of the appropriate ion.

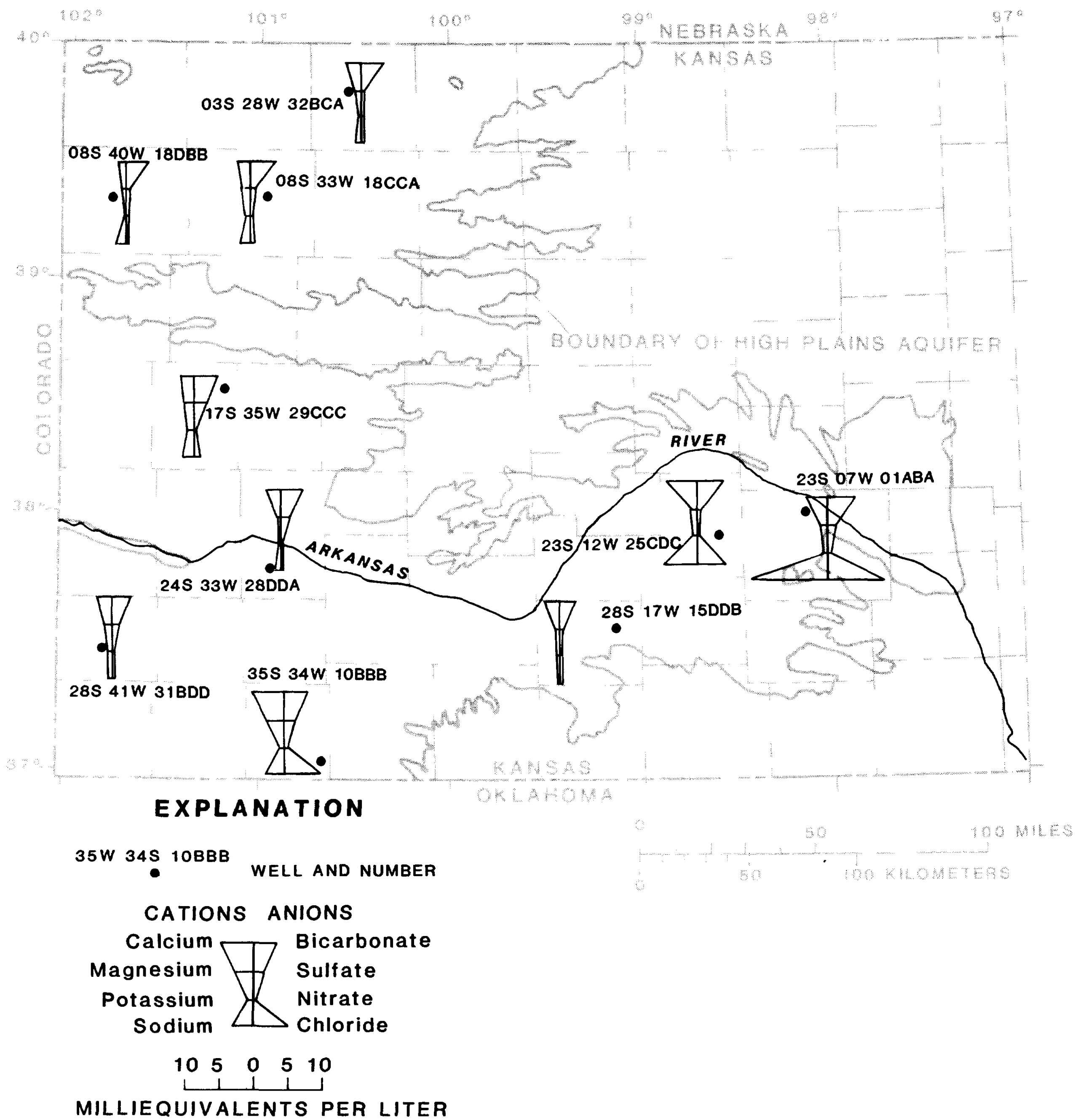


Figure 9.--Concentrations of major ionic constituents dissolved in water from selected wells completed in High Plains aquifer.

Water from the High Plains aquifer in northwest and west-central Kansas is typically a calcium bicarbonate to calcium magnesium bicarbonate water in upland areas and a mixed cation sulfate water from wells in the shallow alluvium of the major stream valleys and in the thick alluvial basin at the terminus of Whitewoman Creek in Scott County. Concentrations of dissolved solids in water from northwest and west-central Kansas increase to the east and south and towards the alluvium of the major stream valleys. Dissolved-solids concentrations are less than 250 mg/L along the Colorado-Kansas State line in Sherman and Cheyenne Counties, between 250 and 500 mg/L in most of the upland areas, and more than 500 mg/L in the alluvium of the valleys and in the Whitewoman basin. Sodium concentrations in the water are less than 25 mg/L in most of northwest Kansas and between 25 and 50 mg/L in west-central Kansas (Krothe and others, 1982).

Water from the High Plains aquifer in southwest Kansas is typically a calcium bicarbonate water in upland areas south of the Arkansas River and a mixed type water north of the Arkansas River. The mixed type water may result from the use of surface water for irrigation, the longer history of irrigation, and the leaching of salts from the unsaturated zone. Water from the alluvium of the Arkansas River valley west of central Gray County is a mixed cation sulfate water. Water from the aquifer at the terminus of Bear Creek (a poorly drained area), along the North Fork of the Cimarron River, and north of the Cimarron River in southwest Haskell and northern Seward Counties is a mixed cation sulfate or mixed cation, mixed anion type water.

Concentrations of dissolved solids in water from the High Plains aquifer in southwest Kansas are highly variable. The calcium bicarbonate waters generally contain less than 500 mg/L dissolved solids. The mixed cation, mixed anion type waters, north of the Arkansas River and along the Cimarron River and Lakin Draw, have dissolved-solids concentrations of 500 to 1,000 mg/L. The concentrations of dissolved solids in the mixed cation sulfate type waters from the alluvium along the Arkansas River range from about 3,700 mg/L in Hamilton County to about 1,500 mg/L in central Gray County. The sodium chloride type waters in the southeast part of southwest Kansas have dissolved-solids concentrations of more than 500 mg/L.

Water from the High Plains aquifer in south-central Kansas is a calcium bicarbonate type water in the western one-half of the area and a sodium chloride water in local areas in eastern Stafford and northern Reno Counties. Mixed cation sulfate type waters tend to occur along the Arkansas River valley. The concentrations of dissolved solids increase from 250 to 500 mg/L in Kiowa County to more than 1,000 mg/L in some areas east of about longitude 99° W. Water from the alluvium of the Arkansas River in south-central Kansas generally contains concentrations of between 500 and 1,000 mg/L for dissolved solids.

MODELING OF HIGH PLAINS AQUIFER

The model of the High Plains aquifer, as referred to in this report, is a set of hydrologic data and a ground-water-flow simulation program that, when used together will reproduce hydraulic heads (representing water-table altitudes) similar to measured values. The worth of such a model is its ability to compare the completeness, accuracy, and compatability of data that describe the various properties of the aquifer. The two simulations reported here, one for northwest Kansas and one for southwest Kansas, consider the data during predevelopment (prior to 1950) when all flow components were assumed to be in equilibrium (no change in storage of ground water was occurring). The results of the simulations represent at least one set of compatible, hydrologic conditions, although not necessarily the only possible set.

An aquifer model simulates flow patterns under a given set of conditions. Historical pumpage data to calibrate the northwest and southwest simulations for the transient state (changing conditions through time) were not available; therefore, no attempt was made to refine the model using transient-state conditions.

Ground-water-flow models have been developed for other parts of the High Plains aquifer in Kansas. Part of the west-central area has been modeled and is described by Dunlap (1980). In southwest Kansas, Barker and others (1983) developed a finite-element model of the Arkansas River alluvium between the Kansas-Colorado State line and the Bear Creek fault (an area adjacent to but not included in the High Plains aquifer) to investigate stream-aquifer interaction. L. E. Dunlap and others (1985) have developed a three-dimensional model of the High Plains in Kearny and Finney Counties (a follow-up to the Barker and others, 1983 modeling). Havens and Christenson (1983) have calibrated a model of the High Plains aquifer in Oklahoma.

According to P. M. Cobb and S. J. Colarullo (Cobb and others, 1983), a finite-difference model of the High Plains aquifer for a 5,400-square-mile area of south-central Kansas indicated that ground-water withdrawals, estimated at 270,000 acre-feet for 1975, have not had a perceptible effect on the potentiometric surface of the aquifer. Simulated ground-water withdrawals resulted in a 16-percent reduction in boundary outflow (base flow and ground-water outflow) for 1975 conditions when compared to 1955 (steady-state) conditions.

A conceptual model is a description of what the investigator believes is the hydrologic framework of the aquifer and how the hydrologic system works within that framework. Prior to modeling, the investigator must use his knowledge, experience, and skills to compile or make logical estimates of the boundary conditions and aquifer fluxes (flows). Previous sections of this report describe the framework as a homogeneous, unconfined aquifer and report flow quantities as they were perceived prior to modeling. These perceptions were revised and refined using a two-dimensional, finite-difference, ground-water-flow modeling program as described in the following paragraphs.

This investigation used the U.S. Geological Survey computer program developed by Trescott and others (1976) to simulate two-dimensional ground-water flow in the aquifer. Due to the complexity of the High Plains aquifer, the partial-differential equation for ground-water flow in two dimensions is not directly solvable and therefore is replaced in the program by finite-difference approximations. "The continuous derivatives...are replaced by finite difference approximations at a point (the node at the center of the block)" (Trescott and others, 1976).

Use of a two-dimensional model for simulation of the High Plains aquifer on a regional scale is justified by the fact that the aquifer, as previously discussed, appears to be "homogeneous in its heterogeneity" or, in a statistical sense, appears to be monomodal (one mean value independent of location or size of the sampled area and with uniform variance). Also, definition of layering and trends within the aquifer, required for three-dimensional analyses, has not been adequately determined through detailed analysis of drillers' logs and logs of test holes.

The basic assumptions incorporated in the finite-difference model are: (1) the aquifer properties are uniform within the blocks, and (2) the Cartesian coordinates are co-linear with the principal components of the hydraulic-conductivity tensor. When steady-state conditions are assumed, the storage term in the flow equation is set to zero. That is, there is no change in storage in the aquifer.

The computer code given in Attachment VII of Trescott and others (1976) was modified to allow program use on a minicomputer. Additional modifications to the code affected only the printing of the results of the program and consisted of (1) a listing of constant-head fluxes, (2) a listing of leakage fluxes, and (3) a summation of leakage flux for each stream. Code was added to provide explicit designation of nodes with leakage to or from a stream node. Further modification allowed the completion of the simulation even if a constant-flux node (discharging well) went dry. This was accomplished by setting the offending well discharge to zero, prior to stopping the program. This option was not necessary during final model simulations as initial data were revised to moderate computed drawdowns.

A requirement of the Trescott finite-difference model is that the hydrologic properties (aquifer characteristics and vertical fluxes) must be defined as homogeneous values for finite areas or blocks. These blocks must be rectangles, and if the aquifer is anisotropic, the axes of the finite-difference grid should be aligned along the principal directions of hydraulic conductivity of the aquifer. The grid axes may be specified in any convenient direction if the aquifer is isotropic in the horizontal plane, as is assumed here for the High Plains aquifer in Kansas.

The grids for both simulations developed as a part of this investigation utilized a common east-west base line. Parallels and perpendiculars to this base line were constructed to form a grid with intervals of 15,000 feet. Each grid block represents an area of 225,000,000 square feet, or about 8.07 square miles.

The computer program specifies the first and last (outside) rows and columns of the model grid as no-flow nodes. Other no-flow nodes may be specified by setting their aquifer hydraulic conductivity to zero. Flows in or out of the modeled area must be allowed by either assigning a constant flux or constant head at the appropriate boundary node. Constant-flux boundaries are assigned by locating a recharging (positive flux) well or a discharging (negative flux) well in each appropriate node. Constant-flux nodes may be assigned to represent any source or sink thought to be an unchanging rate of recharge or discharge from the aquifer. Constant-head boundaries are assigned by coding a negative storage coefficient at each appropriate node. The program considers each node with a negative storage coefficient to have an infinite capacity for providing or accepting whatever water is needed for equilibrium of flow with the adjacent nodes under the gradients produced by the specified constant head.

The northwest and southwest simulations used data from a common source. Each grid block that represents a physical part of the aquifer is called an active node and must be assigned initial values for water level, storage coefficient, hydraulic conductivity, altitude of bottom of the aquifer, specific yield, recharge rate, X and Y grid spacing, and, for nodes with vertical leakage, the hydraulic conductivity and thickness of the confining bed and altitude of the water level on the other side of the confining bed. The water level for each node was determined from the predevelopment (pre-1950) water-level surface (Stullken and Pabst, 1982). The initial hydraulic conductivity for each node was determined from a larger-scale version of the hydraulic-conductivity map shown in figure 4 of this report. Altitudes of the aquifer base were determined from a base-of-aquifer map by Watts and Stullken (1981), shown in lesser detail on plate 1 of this report. Initial values for recharge rate and vertical-leakage components, along with boundary conditions, are described under the individual simulation descriptions.

Leakage between the aquifer and the stream in specified river nodes occurs through the streambed and is simulated by:

$$Q = \frac{kl}{bl} (hs - ha) A , \tag{2}$$

where

- Q = rate of leakage, in cubic feet per second;
- kl = streambed hydraulic conductivity, in feet per second;
- bl = streambed thickness, in feet;
- hs = altitude of stream stage, in feet;
- ha = altitude of water table, in feet; and
- A = area of the grid block, in square feet.

The ratio kl/bl is known as leakance and is reported in units of one per time unit. Since the area of the streambed is less than the area of the grid block in which it occurs, the ratio of streambed hydraulic conductivity to thickness was reduced proportionately (streambed thickness increased) based on the area of the stream reach in the grid block in order to make the calculated amount of leakage realistic. Only nodes in which vertical leakage is to occur are given confining-bed thicknesses of greater than zero.

Steady-State Simulation of High Plains Aquifer in Northwest Kansas

Description

The finite-difference grid used in the steady-state simulation of the High Plains aquifer in northwest Kansas consisted of a matrix of nodes with 31 rows (north to south) and 51 columns (west to east) and is shown in figure 10. The boundaries of the aquifer in northwest Kansas were simulated either as no-flow or constant-head nodes. No-flow nodes represent areas in which the aquifer is absent or is not continuous with the main part of the aquifer. No-flow nodes are not shown outside the boundaries of the model (fig. 10). Constant-flux nodes were used to represent pre-1950 municipal and industrial (railroad) use of ground water although these fluxes were relatively insignificant in the model. Ground-water contributions (leakage) to the base flow of perennial streams draining northwest Kansas were simulated by river nodes (fig. 10).

Constant or assigned hydraulic heads were based on the interpretation of the map of predevelopment water-table altitudes (Stullken and Pabst, 1982) and published geologic and hydrologic maps. If the water-level contours indicated a gradient with some component perpendicular to the boundary of the aquifer and the node represented an area at the physical limit of the High Plains aquifer, the node was designated as a constant-head node. The constant-head boundaries used in this model represent the physical limit of the aquifer with the exception of the constant heads along parts of the western boundary (column 2, rows 7 through 30, fig. 10) and the northern boundary (row 2, columns 7 through 30, fig. 10). These are artificial boundaries that represent areas in which the High Plains aquifer in northwest Kansas is in direct hydraulic connection with the High Plains aquifer in adjacent states.

The initial estimates for hydraulic conductivity (fig. 4) ranged from about 5.8×10^{-5} to 1.9×10^{-3} feet per second (5 to 160 feet per day). The average value for the model was about 7.5×10^{-4} feet per second (68 feet per day). The streambed hydraulic conductivity was initialized as a uniform value equivalent to the average value of the aquifer hydraulic conductivity, about 7.5×10^{-4} feet per second (68 feet per day). The hydraulic head on the river side of the streambed in river nodes was set equal to the aquifer head for that node. Therefore, no flow was specified initially across the streambed. The calculation of leakage is the ratio of the streambed hydraulic conductivity (K') to the streambed thickness (m) multiplied by the area of the streambed (A) and is the head difference between the hydraulic head in the aquifer and the stage in the stream (Δh).

Calibration

The calibration of a ground-water-flow model is a subjective process with many of the required variables being poorly defined. Estimates of the probable range of lesser-known hydraulic properties, such as the hydraulic conductivity of the aquifer or the areal recharge, were made. The hydrologic and geologic data required for solution of the ground-water-flow simulation of the High Plains aquifer in northwest Kansas were ranked from most to

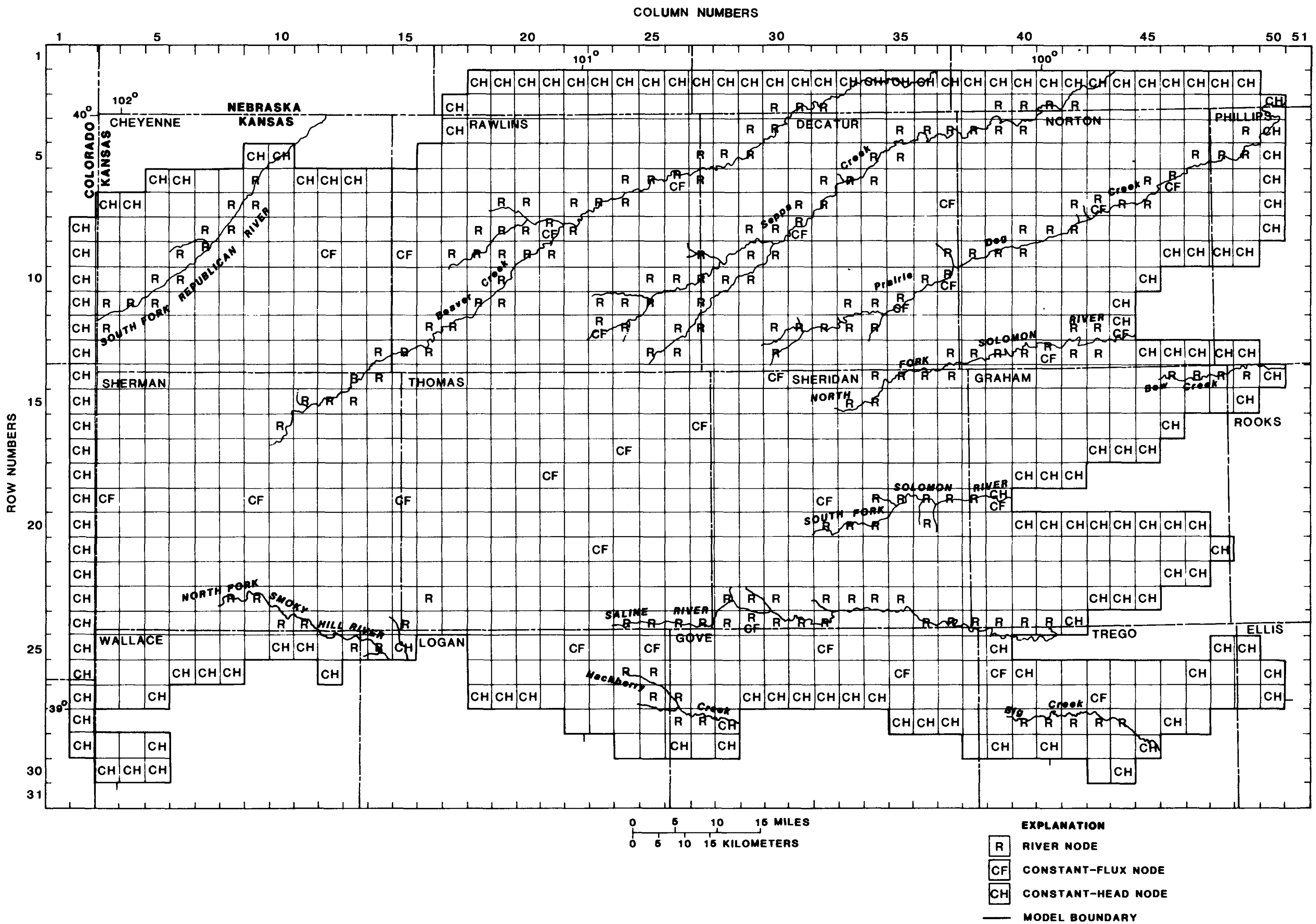


Figure 10.--Model grid and boundary conditions for steady-state simulation of High Plains aquifer, northwest Kansas.

least reliable, based on a subjective evaluation. These rankings, in descending order of reliability were: (1) hydraulic head, (2) aquifer geometry (base and sides of aquifer), (3) base-flow estimates, (4) pumpage, (5) hydraulic conductivity, (6) streambed leakance, and (7) areal recharge (precipitation).

The initial set of data did not produce a steady-state solution within the maximum specified number of iterations (100). Excessive drawdowns (observed, less computed head) that occurred during initial calibration simulations were eliminated by increasing the areal-recharge rate at the upgradient (west) end of the model. However, this caused excessive buildup in the simulated hydraulic-head surface on the downgradient end of the model and leakage to the stream nodes of about three times the estimated base flow of these streams. The increased recharge rate averaged about 1.8×10^{-9} per second, about three times the value estimated by Jenkins and Pabst (1975). Simulated leakage to streams was about 250 to 300 cubic feet per second. Numerous nodes had simulated hydraulic heads in excess of 25 feet from the observed values. The variable hydraulic-conductivity nodal values were "smoothed" by recomputing them as the average of a nine-block area. This eliminated much of the difference between measured and simulated heads.

Discrepancies between the estimates of areal recharge (precipitation) and leakage to streams (base flow) were resolved by incrementally reducing the hydraulic conductivity of the aquifer and the streambed until a satisfactory match was achieved between estimated and simulated areal recharge and between estimated and simulated base flow. The reduction of the hydraulic conductivity of the aquifer from 7.4×10^{-4} to about 2.5×10^{-4} feet per second was supported by the results of comparison of lithologic estimates with production-test estimates of hydraulic conductivity from the 136 sites in Sherman County. The ratio of the lithologic estimates to the production-test estimates of hydraulic conductivity was about four to one. A final revision of the model found that using a uniform hydraulic conductivity was more simplistic and equally workable. The streambed thicknesses and the value for uniform streambed hydraulic conductivity were revised further to produce the expected simulated hydraulic heads and leakage rates to streams. During final calibration, the areal recharge was modified on a node-by-node basis where the difference between observed and simulated heads was greater than 15 feet.

Results

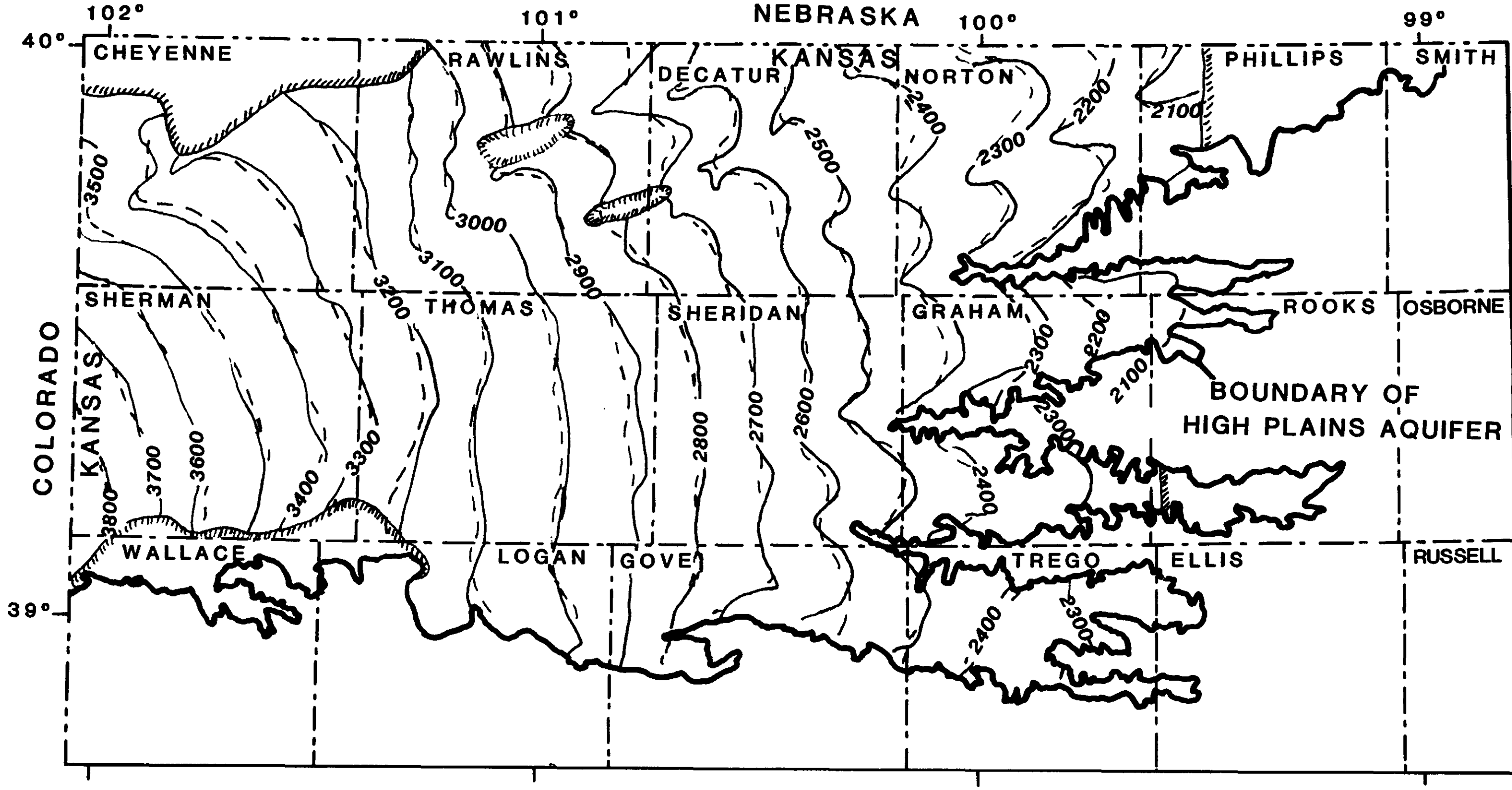
The steady-state simulation provides a refinement to some previous understandings of the ground-water-flow system in northwest Kansas. The simulation results suggest that the aquifer may be considered homogeneous (comparing one 15,000-foot square area to another), that recharge from precipitation (areal recharge) is variable, that the principal aquifer discharge is to streams, and that the thinly saturated parts of the aquifer cannot be ignored when considering boundary flow. During calibration the simulation was modified to use a uniform instead of a variable hydraulic conductivity. The best calibration attained using a variable hydraulic conductivity in the simulation resulted in a mean difference between observed and simulated hydraulic-head values of -2.28 feet, with a standard deviation of 9.25 feet. Using a uniform hydraulic conductivity allowed refinement to a mean hydraulic-head difference of -2.00 feet (also a buildup) and a standard deviation of 8.12 feet. Simulated hydraulic-head difference in 88 percent of the nodes was within 10 feet of observed values.

Recharge, previously estimated by Jenkins and Pabst (1975) to be about 0.25 inch per year, was varied node to node from 0 to 0.8 inch per year to best fit simulated conditions. The mean recharge value of 0.20 inch per year (5.28×10^{-10} feet per second) supports Jenkin's and Pabst's estimate and is used for the majority of the modeled area. No significant pattern was apparent. A uniform aquifer hydraulic conductivity of 22 feet per day (2.5×10^{-7} feet per second) used over the modeled area was considerably lower than the 68 feet per day estimated from lithology and used in preliminary calibrations. Calculated hydraulic-head values were kept near observed values when increasing or decreasing overall recharge by proportionately increasing and decreasing the uniform hydraulic conductivity. The final areal recharge varied from 0 in some nodes to 0.79 inch per year in others. Recharge was adjusted during calibration to govern changes in hydraulic head; however, the amount of recharge also was adjusted to produce the correct amount of ground-water discharge from the aquifer. Because the principal discharge from the aquifer was to streamflow, recharge adjustments could be made only within the range that would produce values within the expected limits of the hydraulic head and stream discharge. Thinly saturated erosional remnants in the eastern part of the aquifer were a significant factor in ground-water discharge from the aquifer. Constant-head nodes used to simulate flow into these areas transmitted 5,500 acre-feet per year or 24 percent of the boundary outflow.

The simulated hydraulic head (water level) computed by the calibrated model closely approximated the measured predevelopment water surface. Estimated streamflow gains also were simulated closely by the model. Both the measured and simulated water-level contours are shown in figure 11. The close approximation of the measured water-level surface does not in itself insure a model that adequately simulates all conditions since there is an almost infinite set of conditions that will give similar results. It does, however, indicate that the ratio of recharge rate to hydraulic conductivity (2.1×10^{-6} to 1) may be proportionately correct. Some degree of confidence is added also by closely approximating streamflow gains.

In some respects the model grid (15,000 feet x 15,000 feet) for northwest Kansas is relatively fine and does not lend itself to use of data aggregated by county. However, the 15,000-foot grid is too coarse to recognize local heterogeneities in the aquifer. The grid size also aggregates rather large areas of bedrock surface into one representative altitude. Computations of flow in areas where the bedrock or water-table surface has considerable relief, as near streams, should be considered more of an approximation than those areas of low relief, where a single value can better represent an entire grid block.

Ground-water-flow models of northwest Kansas could be improved by extending the model boundaries to the edge of the aquifer in Colorado and to the Republican River in Nebraska (a major aquifer drain). Reducing the grid interval would improve resolution around the stream valleys; however,



EXPLANATION

—— 3000 —— MEASURED WATER-LEVEL CONTOUR--Shows altitude of predevelopment water-level surface. Contour interval 100 feet. National Geodetic Vertical Datum of 1929

- - - 3000 - - - SIMULATED WATER-LEVEL CONTOUR--Shows altitude of predevelopment water-level surface. Contour interval 100 feet. National Geodetic Vertical Datum of 1929

WATER LEVEL AT OR NEAR BEDROCK--Area not contoured

Figure 11.--Measured and simulated predevelopment (pre-1950) water-level contours for High Plains aquifer, northwest Kansas.

no resultant improvement or change would be expected in the simulated water budget for the area. The most usable improvement would be computation of water use by small areas and by periods in history matched by water-level maps. The water-use computations would allow additional improvements in the model.

The simulated water budget of the High Plains aquifer in northwest Kansas, given in table 7, is from the final simulation using a uniform value of hydraulic conductivity. The simulated budget shows that under predevelopment conditions 81 percent of the total recharge into the aquifer in northwest Kansas came from precipitation and that 75 percent of the discharge from the aquifer was by leakage to streams. Boundary inflow was about 10 percent less than boundary outflow.

Table 7.--*Water budget from steady-state simulation of High Plains aquifer, northwest Kansas*

[Values are given in acre-feet per year]

	Recharge	Discharge
Precipitation	87,000	---
Boundary flow	21,000	23,000
Stream leakage	---	81,000
Pumpage	---	4,000
Totals	108,000	108,000

Steady-State Simulation of High Plains Aquifer in Southwest Kansas

Description

The model grid (fig. 12) for the High Plains aquifer in southwest Kansas consisted of 32 rows (north to south) and 52 columns (west to east) of 15,000-foot square blocks (nodes). Boundary conditions also are shown in figure 12. Constant-head values were specified for all sides of the model, except for parts of the southern boundary and a small area in southeastern Gray County. Parts of the aquifer in southwest Kansas have small and discontinuous saturated thicknesses and were not simulated (inactive, no-flow nodes). The boundary in eastern Meade County (southeastern boundary of the simulated area) is located along the Crooked Creek-Fowler fault. Saturated thicknesses to the east of the fault (upthrown side) are small and difficult to simulate because of limited data. Therefore, the area to the east of the fault also was excluded from the southwest simulation.

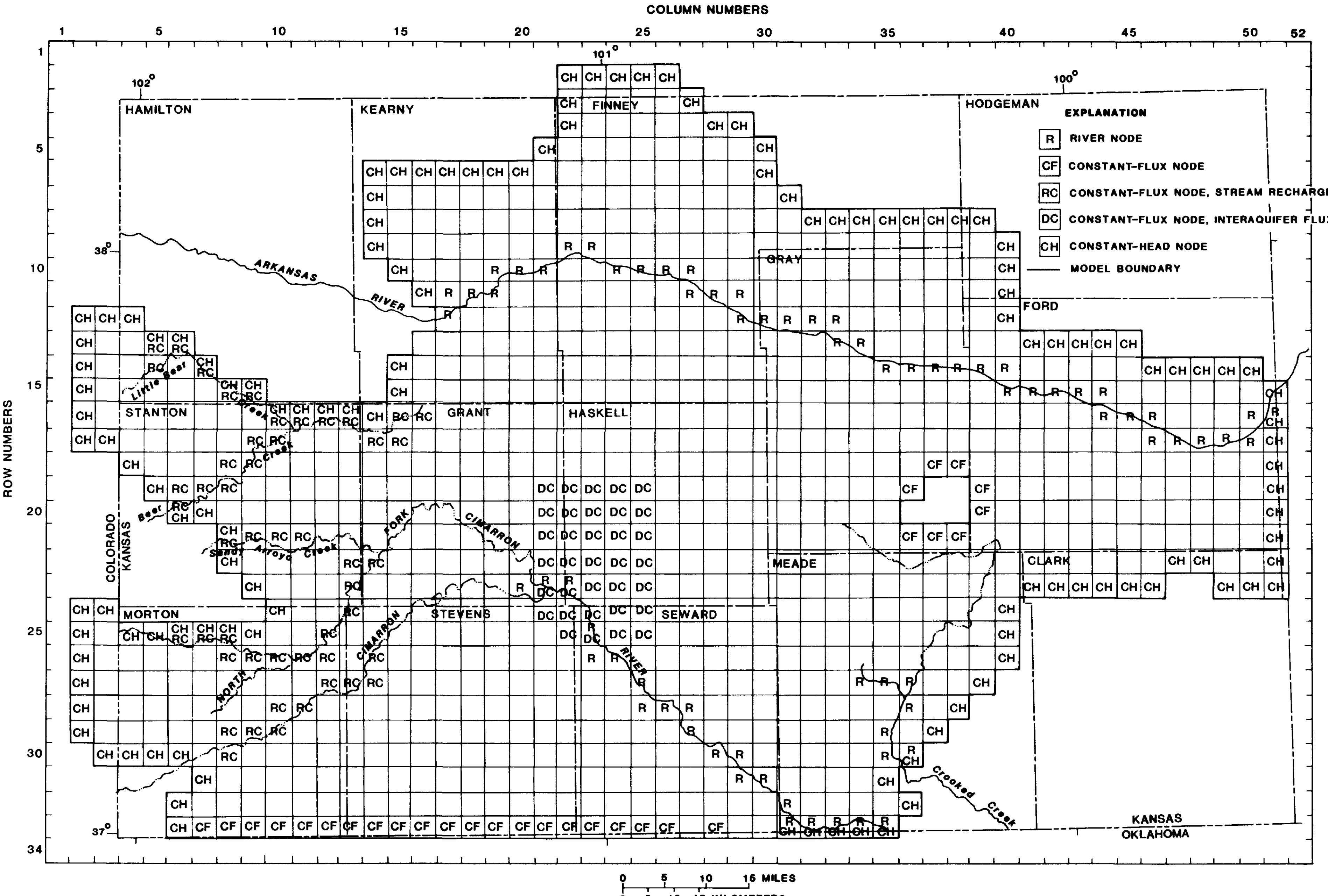


Figure 12.--Model grid and boundary conditions for steady-state simulation of High Plains aquifer, southwest Kansas.

Constant-flux boundaries represented the effects of a constant rate of water movement in or out of the aquifer. Predevelopment (pre-1950) water-level contours indicated that ground water flowed from Kansas southward into Oklahoma along much of the southern boundary (Stullken and Pabst, 1982). Constant-flux values were specified for nodes in columns 7-26 and 28 along the southern boundary by assigning discharge wells to the appropriate nodes (fig. 12). The constant flux specified for each boundary node was calculated using Darcy's law and a pre-1950 water-table contour map (Stullken and Pabst, 1982). The components of flow at right angles to the model boundary were calculated and used as the boundary flux for each node.

A constant-flux boundary also was specified around a bedrock high in southeastern Gray County (fig. 12). A small flow rate was calculated to represent recharge draining from the area outward to adjacent areas of the High Plains aquifer. This flow was assigned to recharge wells around the boundary (8 nodes totaling 1.6 cubic feet per second). Grid blocks at row 19, columns 37 and 38 and row 20, columns 36-38 in southeast Gray County (fig. 12) represent the bedrock high and are not active in the model. Also, a few nodes along the model boundary where water-table contours were perpendicular to the model boundary were not assigned any flux or head, resulting in no outward flow of ground water in these nodes (row 33, columns 27, 29, and 30 and row 13, column 16).

An areally variable hydraulic conductivity was used in the simulation. The mean value for hydraulic conductivity, which ranged from 3 to 51 feet per day over the 1,028 active nodes, was 23.2 feet per day (fig. 4). Attempts at calibration with a uniform hydraulic conductivity were unsuccessful due to the complexities of the area.

Recharge from precipitation indicated by the simulation ranged from 0 to 2 inches per year. The greatest recharge occurred in the area south of the Cimarron River and in the area between the Cimarron River and Crooked Creek where a large percentage of the surface is dune sand. The dune-sand soils have little or no runoff and allow much of the incident water to percolate to the aquifer. The calibrated recharge from precipitation for the Arkansas River valley and dune-sand areas to the south was 0.25 inch per year. Total calibrated recharge to the aquifer from precipitation was 104,500 acre-feet per year (144 cubic feet per second). Recharge from precipitation over a large part of southwest Kansas was negligible presumably because of low infiltration rates and limited precipitation. Recharge was considered as one of the least-known variables and was therefore one of the first to be changed in calibrating the simulation.

Subsurface (boundary) inflow enters southwest Kansas from the west and northwest. Boundary inflow from the Arkansas River alluvial aquifer west of the Bear Creek fault zone was simulated by two constant-flux recharge wells (row 12, column 17 and row 13, column 16, fig. 12), totaling 19 cubic feet per second (14,000 acre-feet per year) (Barker and others, 1983). The total simulated inflow was 32,500 acre-feet per year (about 45 cubic feet per second).

Streams and creeks in the western part of southwest Kansas do not penetrate the water table (ephemeral), but they do recharge the aquifer by infiltration during storm runoff and were simulated by constant fluxes (recharge wells). The streams and creeks that were simulated with recharge wells were Big Bear, Little Bear, and Sand Arroyo Creeks, the North Fork Cimarron River, and the western reach of the Cimarron River (west of column 20, fig. 12). The recharge for each node was proportional to the length of stream reach occurring in each node. Fifty-one recharge wells with a total recharge rate of 22.2 cubic feet per second (16,000 acre-feet per year) were used to simulate streams and creeks in the western part of the southwest model area (fig. 12).

A number of streams and creeks are present in the southwest modeled area. The Arkansas River, the eastern reach of the Cimarron River, and Crooked Creek were treated in the simulation as partially-penetrating streams with a leaky streambed (vertical leakage). The streams served as both a source of recharge and discharge to the aquifer, but the net leakage in each case was from the aquifer to the river (discharge from the aquifer). The simulated recharge to the aquifer owing to leakage from the Arkansas River, Crooked Creek, and the eastern reach of the Cimarron River was 19,940 acre-feet per year (about 28 cubic feet per second).

Subsurface (boundary) outflow occurs through constant-head and constant-flux nodes along the southern, eastern, and northeastern boundaries of the model area. The total simulated boundary outflow was 65,000 acre-feet per year (90 cubic feet per second).

Movement of water from the High Plains aquifer to the underlying Lower Cretaceous sandstone aquifer was simulated in southwest Kansas in parts of Grant, Haskell, Stevens, Seward, and Meade Counties. The discharge of water from the High Plains aquifer to the Lower Cretaceous sandstone aquifer was represented by discharge wells with a rate totaling 14,000 acre-feet per year (about 19 cubic feet per second). This quantity was determined by trial and error in the model-calibration process. Discharge of water from the High Plains aquifer also occurred as springs in parts of Meade County near the Crooked Creek-Fowler fault (Frye, 1942). Spring discharge was not differentiated from discharge from the High Plains aquifer to the Lower Cretaceous sandstone aquifer because it was of minor importance and difficult to quantify.

The Arkansas River, Cimarron River, and Crooked Creek were gaining streams in the area. Although in some reaches the flow gradient was from the stream to the aquifer, in most reaches flow was from the aquifer to the stream, resulting in a net discharge from the aquifer. The simulated base flow, defined as cumulative contributions from the aquifer to the stream, for (1) the Arkansas River at the eastern boundary of the southwest Kansas simulated area (row 16, column 51) was 32 cubic feet per second (23,000 acre-feet per year); (2) the Cimarron River at the southeastern boundary of the area (row 33, column 35), 56 cubic feet per second (41,000 acre-feet per year); and (3) Crooked Creek at the southeastern boundary of the area (row 30, column 36), 14 cubic feet per second (10,000 acre-feet per year). The simulated base flows are close to the estimated base flows of the Arkansas and Cimarron Rivers and of Crooked Creek. The total net simulated leakage from the aquifer to the streams in the southwest area was 58,000 acre-feet per year (80 cubic feet per second).

A value of streambed leakance (hydraulic conductivity divided by streambed thickness) of 1.34 per day (before the adjustment for streambed area and assuming a streambed thickness of 1 foot) was used in the simulation for the river nodes representing the Arkansas and Cimarron Rivers. This same leakance value was used for the Arkansas River in southwest Kansas in studies by Barker and others (1983) and by Dunlap and others (1985). Streambed and flow conditions for the Cimarron River are similar to those for the Arkansas River; therefore, the same leakance value was used for both. The channel width of Crooked Creek is much narrower than that of the Arkansas and Cimarron Rivers. Because reducing the streambed-leakance value has the same effect as reducing the streambed area (which cannot be changed in the model), the streambed leakance was reduced for Crooked Creek to 25 percent of the Arkansas River value.

Calibration

During the steady-state calibration, a number of factors were adjusted within limits of expected values until the calculated hydraulic heads approximated the measured hydraulic heads. The most significant factors adjusted were: (1) recharge from precipitation, (2) hydraulic conductivity, (3) recharge from streams in the western part of the simulated area, and (4) discharge from the High Plains aquifer to the underlying Lower Cretaceous sandstone aquifer.

The primary factor adjusted during the calibration was recharge to the High Plains aquifer from infiltration of precipitation. In areas where simulated hydraulic heads were less than measured heads, recharge was increased up to a maximum of 2 inches per year until the simulated heads approximated the measured hydraulic heads. In general, areas requiring the greater recharge to match hydraulic heads also had more permeable soils.

The initial values of hydraulic conductivity were varied uniformly (by a common factor) until the best overall match between measured and simulated hydraulic heads was obtained. Simulation attempts with hydraulic conductivities lower than the values used in the final simulation resulted in hydraulic heads that were generally too high, while attempts at simulation with hydraulic conductivities greater than these values resulted in hydraulic heads that were generally too low in the western part of the southwest area and streamflows that were significantly greater than measured historic flows in Crooked Creek and the Cimmaron River. Simulation attempts with hydraulic conductivities greater than the final values also resulted in hydraulic heads that were too large along the southeastern boundary of the simulated area (along the Crooked Creek-Fowler fault).

Initial values of recharge from streams in the western part of the southwest area were adjusted in some cases to improve the match between measured and simulated hydraulic heads. During calibration, the initial values of streambed thickness were adjusted for a few river nodes to more nearly match the measured historic streamflow distribution by reaches along the Cimarron River. Also, the initial calculated fluxes for some of the constant-flux nodes along the southern boundary were adjusted to facilitate a balance of the flux out of the southwest simulated area to the south with a reasonable amount of recharge in the area south of the Cimarron River and streamflow in the Cimarron River.

Results

During the steady-state calibration of aquifer conditions before 1950, a number of factors were adjusted within limits of expected values until the simulated hydraulic heads approximated the measured hydraulic heads. The arithmetic average difference between the measured and simulated heads for all 1,028 active nodes was -1.08 feet (a buildup) with a standard deviation of 10.5 feet. Simulation results suggest that recharge from precipitation (areal recharge) to the High Plains aquifer in the modeled area is variable and that the major aquifer discharge prior to 1950 was to streams.

Recharge, previously estimated at 0.05 to 6.0 inches per year, was varied from node to node to best fit the simulated conditions. The mean recharge value of 0.24 inch per year is slightly greater than the amount used for the simulation in northwest Kansas (0.20 inch per year). The final areal recharge varied from 0 in some nodes where the aquifer is covered by very low permeability soils, such as loess, to 2.0 inches per year in others where the aquifer is covered by dune sand. Recharge was adjusted during calibration to govern hydraulic-head decline or buildup; however, the amount of recharge also was adjusted to produce the correct amount of ground-water discharge to streams. Recharge adjustments could be made only within the range that would produce results within the limits of hydraulic head and stream base flow.

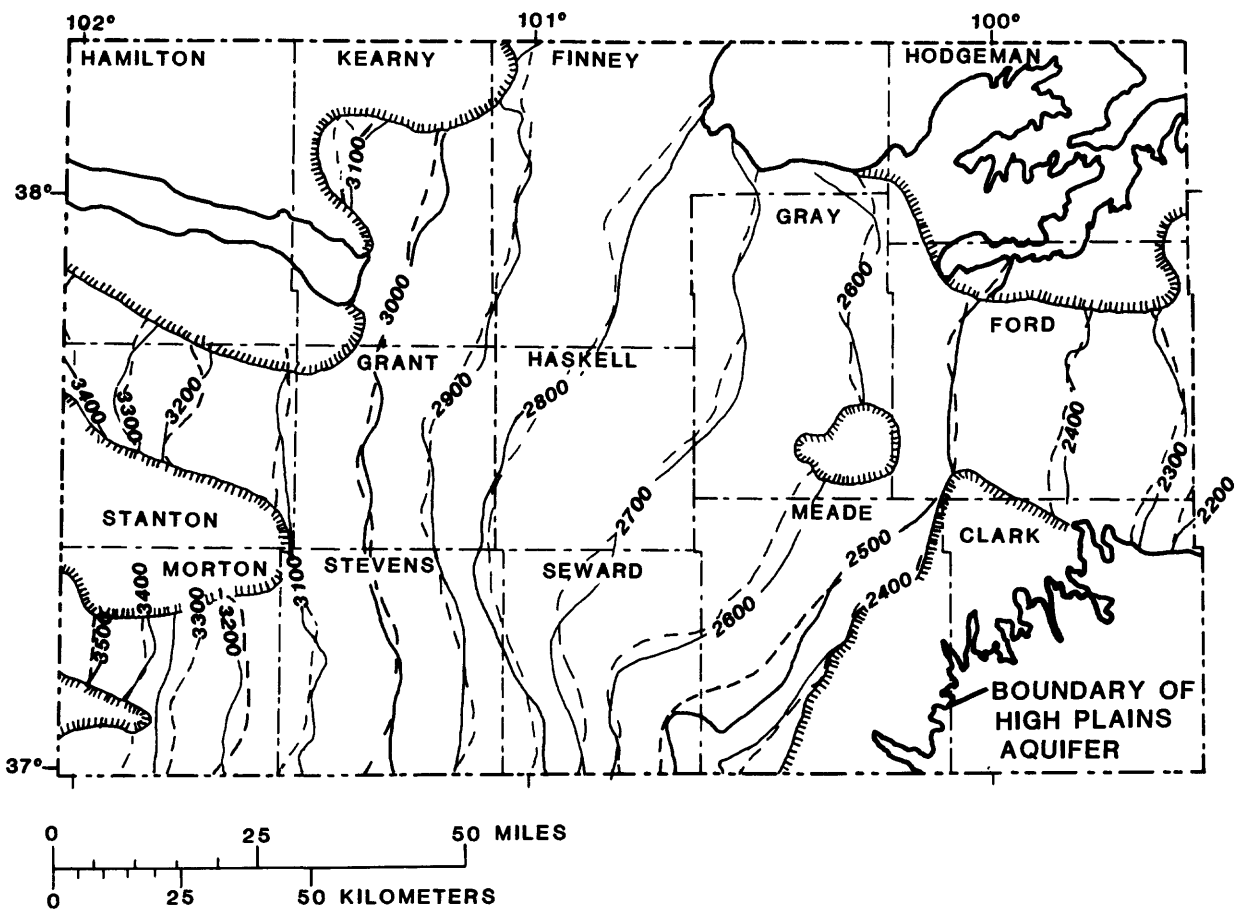
During calibration, the initial values of streambed thickness were adjusted for a few river nodes to more nearly match the measured streamflow distribution by reaches along the Cimarron River. Also, the initial calculated fluxes for some of the constant-flux nodes along the southern boundary were adjusted to facilitate a balance between the flux from the southwest area towards the south, a reasonable amount of recharge in the area south of the Cimarron River, and streamflow in the Cimarron River.

Based on results of the northwest Kansas simulation, a uniform hydraulic conductivity was tried and found less satisfactory than using a variable hydraulic conductivity. The flow characteristics of the area between the Bear Creek and the Crooked Creek-Fowler faults were much more complex than previously expected. The area between the two faults is complicated by leakage to the underlying Lower Cretaceous sandstone aquifer, leakage to adjacent streams, areas of high recharge, zones of rapidly changing permeability (Crooked Creek-Fowler fault), and a probability of slumping near sinkholes. The aquifer hydraulic conductivity ranged from 3 to 51 feet per day, with a mean of 23.2 feet per day and a standard deviation of 8.48 feet per day.

The measured predevelopment water-table surface and the simulated surface are shown in figure 13. They are closely correlated.

In some respects, the model grid used in the southwest Kansas simulation is relatively fine and does not lend itself to use of data aggregated by county. On the other hand, the 15,000-foot grid is too coarse to recognize local heterogeneities of the aquifer. The grid size also aggregates rather large areas of bedrock into a single uniform block. Computations of flow in areas where the bedrock or water-table surface has considerable relief, as near streams, are more approximate than where a single value represents a hydraulic property for an entire grid block.

Ground-water-flow models of southwest Kansas could be improved by extending the model boundaries further into Colorado and Oklahoma and north to the Smoky Hill River. Reducing the grid interval would improve resolution



EXPLANATION

—— 3000 —— MEASURED WATER-LEVEL CONTOUR--Shows altitude of predevelopment water-level surface. Contour interval 100 feet. National Geodetic Vertical Datum of 1929

- - - - 3000 - - - - SIMULATED WATER-LEVEL CONTOUR--Shows altitude of simulated predevelopment water-level surface. Contour interval 100 feet. National Geodetic Vertical Datum of 1929

WATER LEVEL AT OR NEAR BEDROCK--Area delineated not contoured

Figure 13.--Measured and simulated predevelopment (pre-1950) water-level contours for High Plains aquifer, southwest Kansas.

around the stream valleys; however, no improvement would be expected in the simulated water budget. The most usable improvement would be to obtain water-use data by small areas and by periods in history to match the water-level maps. The water-use data can be used to calibrate a transient model.

The calibrated steady-state simulation for southwest Kansas provided a simulated water budget for the High Plains aquifer in the modeled area, as well as the simulated hydraulic heads. The simulated water budget is given in table 8, which shows that, under steady-state conditions, 60 percent of the total recharge to the aquifer came from precipitation, 19 percent came from boundary inflow, and 21 percent came from leakage of streams. Also, 38 percent of the total discharge from the aquifer went to boundary outflow, 8 percent went to the underlying Lower Cretaceous sandstone aquifer, and 54 percent went to streams. The net leakage from the aquifer to the streams in the southwest area was 58,000 acre-feet per year (80 cubic feet per second).

Table 8.--*Water budget from steady-state simulation of High Plains aquifer, southwest Kansas*

[Values are given in acre-feet per year]

	Recharge	Discharge
Precipitation	104,500	---
Boundary flow	32,500	65,000
Stream leakage	36,000	94,000
Loss to Lower Cretaceous sandstone aquifer	---	14,000
Total	173,000	173,000

SUMMARY

The High Plains aquifer underlies 30,500 square miles in western and south-central Kansas. Onsite measurements performed during this investigation included water-use determinations (well discharges, time-of-operation metering of irrigation wells, irrigated acreage, and cropping patterns) and channel-geometry measurements to determine streamflow at ungaged sites.

The geologic units of interest range from Paleozoic to Cenozoic in age. Paleozoic rocks of the Permian System underlie all of the High Plains aquifer in the study area and contain up to 700 feet of bedded salt in places. Mesozoic rocks underlie all but the extreme southwestern and eastern parts of the aquifer where Permian rocks subcrop on the bedrock surface. Included in the Mesozoic rocks is a Lower Cretaceous sandstone aquifer. Cenozoic deposits of western Kansas contain a considerable thickness of unconsolidated continental deposits of alluvial clay, silt, sand, gravel, and eolian (loess and dune sand) deposits. They include the Ogallala Formation of late Tertiary age and deposits of Pleistocene and Holocene age.

The High Plains aquifer system, which consists of one or more hydraulically connected geologic units of late Tertiary or Quaternary age, is considered to be a single hydrologic unit (an aquifer), with saturated thickness as great as 600 feet near the southern border of southwest Kansas. The High Plains aquifer stores water from precipitation, which percolates downward, and for the most part releases water from storage to downgradient rivers, springs, and seeps and to wells and plants. Downward leakage of recharge through overlying dune sand is much greater than through loess, which retains much of the infiltrating precipitation. Downward leakage of water from the High Plains aquifer to the underlying Lower Cretaceous sandstone aquifer occurs in some areas.

The High Plains aquifer is considered to be unconfined although confined conditions do occur locally (notably along the Crooked Creek-Fowler fault in Meade County). The water table as measured during 1980 slopes generally eastward at about 8 feet per mile. Water levels in the High Plains aquifer exhibit short-term fluctuations and long-term declines due to irrigation pumpage. Short-term fluctuations produced by seasonal pumpage for irrigation develop into long-term water-level declines in those areas where pumpage exceeds recharge.

The hydraulic properties of the High Plains aquifer in the study area are highly variable. Reported hydraulic conductivity ranges from 1 to 1,600 feet per day, and reported storage coefficients range from 0.001 to 0.22 (dimensionless). An analysis based on 1,612 lithologic (drillers') logs of the High Plains aquifer in the study area indicates that the mean estimated hydraulic conductivity of the aquifer is 75 feet per day, with a mean estimated specific yield of 0.17. The regional value of hydraulic conductivity for large blocks of aquifer material used in the steady-state simulation of northwest Kansas was about 22 feet per day and in the simulation of southwest Kansas ranged from 3 to 51 feet per day with an average of about 23 feet per day.

Recharge to the High Plains aquifer in Kansas is highly variable, occurring primarily from precipitation but also from infiltration of flow in ephemeral streams, ground-water inflow from adjacent parts of the aquifer, upward leakage from bedrock formations, and seepage from irrigation return flows. Estimates of recharge from streams in southwest Kansas using channel-geometry methods ranged from 0.08 to 0.14 inch per year in the reaches of the losing streams. Ground-water inflow from Colorado, Nebraska, Oklahoma, and the Arkansas River alluvium was estimated at about 65,000 acre-feet per year. Leakage from the aquifer to bedrock formations may be about 14,000 acre-feet per year in southwest Kansas and as much as 10,000 acre-feet per year of water flows upward from the bedrock formations to the High Plains aquifer in south-central Kansas.

Discharge from the High Plains aquifer in the study area occurs by evapotranspiration, flow to seeps, springs, and streams, outflow to adjacent parts of the aquifer, downward leakage to underlying formations, and withdrawals by wells. Evapotranspiration from the High Plains aquifer in Kansas is thought to be negligible throughout much of the study area because the water table is too far below the surface. Ground-water outflow to Nebraska, Oklahoma, and the Arkansas River alluvium was about 42,000 acre-feet per year. Leakage to the underlying formations occurs mostly south of latitude 38° S. and east of longitude 99° W. where the aquifer is hydraulically connected with the Lower Cretaceous sandstone aquifer. Withdrawals by wells have increased from a minor part of the ground-water discharge prior to 1950 to one of major significance. Withdrawals were estimated to be 4,215,000 acre-feet per year by 1980. Nine townships (16 square miles each) in southwest and west-central Kansas have from 109 to 144 irrigation wells for an average of more than 3 wells per square mile. The heaviest development of irrigation centers around west-central and the northern part of southwest Kansas.

Water from the High Plains aquifer in Kansas generally is suitable for human and animal consumption and irrigation of crops. Typically, it is a calcium bicarbonate type water with concentrations of total dissolved solids ranging from 250 to 500 mg/L. The quality of water in the aquifer deteriorates toward the east due to mixing with recharge water containing dissolved minerals leached from the overlying soil and unsaturated zones and mineralized water from adjacent bedrock units. The result is a water containing greater concentrations of dissolved solids and a change to a calcium sulfate or sodium chloride type water. Quality of water from the aquifer worsens with depth in parts of Meade and Seward Counties and in the eastern one-half of south-central Kansas due to mixing of saline water leaking upward from the underlying Permian rocks.

Computer simulations of steady-state flow during predevelopment time (before 1950) in northwest and southwest Kansas were calibrated as a part of this study. A U.S. Geological Survey two-dimensional, finite-difference computer program was used for both simulations with a uniform grid interval of 15,000 feet.

Results for the northwest Kansas simulation suggest that the aquifer may be considered "homogeneous" on a regional scale. That is hydraulic properties (average values) are comparable among model nodes (15,000 x 15,000 feet) although large variations may occur locally within a nodal area. A uniform hydraulic conductivity of about 22 feet per day produced a better "fit" with measured hydraulic heads than did values varying according to lithologic-log estimates. Areal recharge (precipitation), however, was highly variable, ranging from 0 to 2 inches per year depending on soil types and averaging 0.20 inch per year over the area. Stream base flow was the principal discharge from the aquifer in the study area (81,000 acre-feet per year). Hydraulic-head difference between the simulated and observed water levels averaged -2.00 feet per node (a buildup), with a standard deviation of 8.12 feet.

The southwest Kansas simulation used a variable hydraulic conductivity ranging from 3 to 51 feet per day and averaging 23.2 feet per day. The flow characteristics of the area between the Bear Creek and the Crooked Creek-Fowler faults were found to be highly variable and complex due in part to lithologic discontinuities as a result of historic slumping near sink-holes. Areal recharge from precipitation was varied from 0 to 2 inches per year depending on the soil types and averaged 0.25 inch per year over the area (104,500 acre-feet per year). Recharge from ephemeral streams in the western part of southwest Kansas was simulated with recharge wells at a total rate of 16,000 acre-feet per year. Net leakage to streams from the High Plains aquifer underlying southwest Kansas was 58,000 acre-feet per year. The downward leakage to the Lower Cretaceous sandstone aquifer was 14,000 acre-feet per year.

Several characteristics of the hydrology in northwest and southwest Kansas were suggested by the simulations. The aquifer materials underlying northwest Kansas, at the grid size used, probably can be considered more homogeneous than the aquifer material underlying southwest Kansas. In both simulations, the hydraulic conductivity was lower than the values indicated by previous investigations. Areal recharge was slightly lower in northwest Kansas than in southwest Kansas and highly variable in both simulations. Areas of thin saturated thickness in northwest Kansas were found to be significant because part of the ground water leaves the modeled area through these thinly saturated areas.

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Table 9.--*Reported hydraulic properties from wells completed in High Plains aquifer in Kansas*

[Geologic source codes: QU = undifferentiated Quaternary rocks; TO = Ogallala Formation, Tertiary age; QT = undifferentiated Quaternary and Tertiary rocks. When saturated or effective thickness is followed by an "e," the thickness was estimated. NR = not reported]

Well number	Hydraulic conductivity (feet per day)	Transmissivity (square feet per day)	Storage coefficient or specific yield (dimensionless units)	Saturated or effective thickness (feet)	Geologic source	Reference
01S 29W 30BDD	273	13,400	0.2	49e	QU	Hodson (1969)
03S 29W 21BAD	400	16,000	.2	40e	QU	Do.
04S 42W 26BAD	735	23,400	NR	32e	QU	Prescott (1953a)
05S 28W 05DCD	261	9,400	.2	36e	QU	Hodson (1969)
07S 39W 20BAD	442	53,000	NR	120	TO	Prescott (1953b)
08S 28W 09AB	60	6,800	NR	114	TO	Bayne (1956)
08S 40W 12DBA	38	4,800	NR	127	TO	Prescott (1953b)
11S 30W 27DC	43	2,400	NR	55.4	TO	Hodson and Wahl (1960)
11S 32W 03BD	46	6,600	NR	142	TO	Johnson (1958)
11S 33W 14BC	69	5,800	NR	84	TO	Do.
11S 34W 24CA	31	2,600	NR	83	TO	Do.
11S 35W 05BB	154	6,300	NR	41	TO	Do.
12S 23W 20CCC	137	6,300	.0035	46	QU	Hodson (1965)
12S 23W 30ACC	314	16,000	.013	51	QU	Do.
14S 42W 23DBB	97	22,700	.007	233	TO	Hodson (1963)
15S 39W 23CBC	148	20,100	.0006	136	TO	Do.
16S 35W 20CCC	170	12,900	NR	76	TO	Prescott and others (1954)
16S 35W 31DA	20	2,000	NR	102	TO	Bradley and Johnson (1957)
16S 41W 17CCC	144	9,500	.09	66	TO	Slagle and Weakly (1976)
17S 27W 28ACC	69	2,400	.032	35	TO	Gutentag and Stullken (1976)

Table 9.--*Reported hydraulic properties from wells completed in High Plains aquifer in Kansas*--Continued

Well number	Hydraulic conductivity (feet per day)	Transmissivity (square feet per day)	Storage coefficient or specific yield (dimensionless units)	Saturated or effective thickness (feet)	Geologic source	Reference
17S 30W 20BBC	92	3,500	0.01	38	TO	Gutentag and Stullken (1976)
17S 34W 02DC	59	5,300	NR	90	TO	Bradley and Johnson (1957)
17S 38W 35DAA	100	5,700	.08	57	TO	Slagle and Weakly
17S 39W 19DBB	174	5,400	.001	31	TO	Do.
18S 16W 16BBB	221	13,400	.036	61e	QU	McNellis (1973)
18S 17W 28BCC	235	9,400	.031	40e	QU	Do.
18S 09W 21CDB	382	13,400	.14	35e	QU	Do.
18S 28W 18ACC	130	5,700	NR	44	TO	Gutentag and Stullken (1976)
18S 32W 07AC	105	17,400	NR	165	QT	Bradley and Johnson (1957)
18S 32W 29BD	163	6,200	NR	38	QT	Waite (1947)
18S 33W 35DD	114	10,600	NR	93	QT	Do.
18S 33W 36BA	26	1,400	NR	54	QU	Do.
18S 34W 23CCD	176	3,000	.18	17	TO	Gutentag and Stullken (1976)
18S 35W 34ABB	904	48,800	NR	54	TO	Prescott and others (1954)
18S 35W 36BCB	722	36,100	NR	50	TO	Do.
18S 36W 05DBB	197	12,000	.18	61	TO	Slagle and Weakly (1976)
18S 38W 31DBC	442	16,800	NR	38	TO	Prescott and others (1954)
19S 33W 12CC	38	2,300	NR	60	QU	Waite (1947)
20S 32W 20ADB	44	3,200	.001	72	QT	Gutentag and Stullken (1976)
20S 33W 18ABC	100	6,000	.14	60	TO	Do.

Table 9.--*Reported hydraulic properties from wells completed in High Plains aquifer in Kansas*--Continued

Well number	Hydraulic conductivity (feet per day)	Transmissivity (square feet per day)	Storage coefficient or specific yield (dimensionless units)	Saturated or effective thickness (feet)	Geologic source	Reference
20S 33W 26CAA	309	9,900	0.001	32	QT	Gutentag and Stullken (1976)
21S 21W 21BC	1,600	61,500	NR	38e	QU	Fishel (1952)
21S 21W 35CC	388	18,600	NR	48e	QU	Do.
21S 21W 35BA	207	12,700	NR	61e	QU	Do.
21S 32W 08AB	175	11,200	.0014	64	QU	Meyer and others (1970)
21S 33W 02AC	198	8,300	.00063	42	QU	Do.
22S 22W 23BC	83	6,800	NR	82e	QU	Fishel (1952)
22S 31W 32DB	148	9,000	.048	61	QT	Meyer and others (1970)
23S 02W 29BBBB	30	6,100	NR	205	QU	Williams and Lohman (1949)
23S 02W 29CB	33	7,100	NR	212	QU	Do.
23S 02W 29CDDD	40	8,300	NR	209	QU	Do.
23S 02W 30DDDD	33	7,000	NR	210	QU	Do.
23S 02W 32CBBB	43	9,200	NR	216	QU	Do.
23S 02W 32CCCC	57	13,400	NR	235	QU	Do.
23S 22W 11CC	79	6,000	NR	76e	QU	Fishel (1952)
23S 34W 36DC	54	8,600	NR	160	QT	Latta (1944)
23S 35W 26CB	33	10,600	NR	320e	QT	McLaughlin (1943)
23S 37W 36BAA	43	3,300	.11	76	QT	Gutentag and others (1972)
24S 02W 06AD	41	9,800	NR	239	QU	Williams and Lohman (1949)
24S 02W 08BA	46	11,400	NR	246	QU	Do.
24S 02W 08BB	110	25,400	NR	231	QU	Do.

Table 9.--*Reported hydraulic properties from wells completed in High Plains aquifer in Kansas*--Continued

Well number	Hydraulic conductivity (feet per day)	Transmissivity (square feet per day)	Storage coefficient or specific yield (dimensionless units)	Saturated or effective thickness (feet)	Geologic source	Reference
24S 02W 08BD	162	34,900	NR	216	QU	Williams and Lohman (1949)
24S 02W 08DB	74	17,000	NR	230	QU	Do.
24S 02W 09AD	82	14,700	NR	179	QU	Do.
24S 02W 09CB	53	11,700	NR	222	QU	Do.
24S 02W 09DDDD	49	8,800	NR	178	QU	Do.
24S 02W 16BBBB1/	331	29,800	NR	90	QU	Do.
24S 02W 16CB1/	408	44,100	NR	108	QU	Do.
24S 02W 16DA	70	12,400	NR	176	QU	Do.
24S 02W 17AB	71	16,500	NR	233	QU	Do.
24S 02W 22CD	62	9,000	NR	145	QU	Do.
24S 02W 26CCCC1/	578	37,600	NR	65	QU	Do.
24S 02W 26DC1/	402	26,500	NR	66	QU	Do.
24S 02W 27AAAA	64	8,500	NR	133	QU	Do.
24S 02W 35AD	53	10,100	NR	190	QU	Do.
24S 02W 35DDDD1/	248	20,600	NR	83	QU	Do.
24S 02W 36DC	93	16,400	NR	176	QU	Do.
24S 30W 32CA	24	3,900	NR	165	TO	Stramel and others (1958)
24S 31W 27CBB	53	6,350	NR	120	QT	Latta (1944)
24S 31W 27DBA	31	3,610	NR	116	QT	Do.
24S 32W 16AB	83	9,600	0.0017	115	QT	Meyer and others (1970)
24S 32W 25BDA	43	3,900	.00062	90	QT	Do.
24S 33W 07BA	95	5,300	.00057	56	QT	Do.
24S 34W 01DDB	355	18,800	.14	53	QT	Do.
24S 35W 09CC	37	13,100	NR	350e	QT	McLaughlin (1943)
24S 35W 28CC	74	25,100	NR	340e	QT	Do.

Table 9.--*Reported hydraulic properties from wells completed in High Plains aquifer in Kansas*--Continued

Well number	Hydraulic conductivity (feet per day)	Transmissivity (square feet per day)	Storage coefficient or specific yield (dimensionless units)	Saturated or effective thickness (feet)	Geologic source	Reference
24S 39W 35CCB	517	21,700	0.14	42	QU	Gutentag and others (1972)
24S 40W 17D	1,200	36,000	.09	30	QU	Lobmeyer and Sauer (1974)
24S 41W 03BB	25	1,500	NR	60e	QU	McLaughlin (1943)
24S 41W 04BB	1,218	73,100	NR	60e	QU	Do.
25S 29W 14AB	41	8,200	NR	200	TO	Stramel and others (1958)
25S 29W 28BD	24	1,600	.0003	66	QT	Do.
25S 29W 28CA	823	18,100	.09	22	QU	Do.
25S 30W 21BD	31	1,600	.0002	52	TO	Do.
25S 30W 22AC	591	13,000	.13	22	QU	Do.
25S 30W 25DA	672	24,200	.17	36	QU	Do.
25S 36W 18ACC	193	26,200	.06	136	QT	Gutentag and others (1972)
26S 01E 29BA	760	NR	NR	NR	QU	Williams and Lohman (1949)
26S 13W 19BDB	NR	20,000	.0004	NR	QU	Layton and Berry (1973)
26S 27W 17BDDD	139	20,600	NR	148	QT	Latta (1944)
26S 28W 10BB	138	5,100	NR	37	QU	Do.
26S 32W 31CC	87	12,000	.22	138	QT	Meyer and others (1970)
26S 33W 12CA	122	18,700	.0012	153	QT	Do.
26S 37W 21DDD	35	6,300	.0006	180	QT	Gutentag and others (1972)
26S 42W 22CD	48	7,500	.0002	155	QU	Lobmeyer and Sauer (1974)
27S 13W 21ACA	NR	21,000	.001	NR	QU	Layton and Berry (1973)

Table 9.--*Reported hydraulic properties from wells completed in High Plains aquifer in Kansas*--Continued

Well number	Hydraulic conductivity (feet per day)	Transmissivity (square feet per day)	Storage coefficient or specific yield (dimensionless units)	Saturated or effective thickness (feet)	Geologic source	Reference
27S 34W 16DDD	183	26,000	0.18	142	QT	Gutentag and Stullken (1974)
27S A4W 28DAA	69	9,000	.01	130	QT	Do.
27S 36W 15DD	88	20,500	.00014	234e	QT	Fader and others (1964)
27S 37W 29CC[1/]	32	7,000	.00012	220e	QT	Do.
27S 38W 15DA[1/]	40	8,500	.00023	210e	QT	Do.
27S 38W 19CD	293	78,900	.0048	269e	QT	Do.
27S 38W 22CB	96	21,300	.00035	222e	QT	Do.
27S 38W 23CA	49	9,500	.00021	193e	QT	Do.
27S 38W 32BB	97	25,100	.0024	258e	QT	Do.
27S 40W 25CB	73	18,300	.0048	249e	QT	Do.
28S 13W 26DCB	NR	21,000	.002	NR	QU	Layton and Berry (1973)
28S 31W 02CBB	226	35,000	.20	155	QT	Gutentag and Stullken (1974
28S 32W 18BBB	200	40,000	.15	200	QT	Do.
28S 36W 11BA	133	28,800	.00022	217e	QT	Fader and others (1964)
28S 38W 12CB	23	6,800	.00028	300e	QT	Do.
28S 38W 15CB	50	15,900	.0006	320e	QT	Do.
28S 38W 27BA	57	16,700	.00021	292e	QT	Do.
28S 39W 12AC	39	5,400	.00011	140e	QT	Do.
28S 39W 20BD	118	25,100	.00095	212e	QT	Do.
28S 39W 24CC	217	62,200	.0094	286e	QT	Do.

Table 9.--*Reported hydraulic properties from wells completed in High Plains aquifer in Kansas*--Continued

Well number	Hydraulic conductivity (feet per day)	Transmissivity (square feet per day)	Storage coefficient or specific yield (dimensionless units)	Saturated or effective thickness (feet)	Geologic source	Reference
28S 41W 14AA	320	47,100	0.059	147e	QT	Fader and others (1964)
29S 35W 15AB[1]	75	17,900	.00038	240e	QU	Do.
29S 38W 35DB[1]	19	6,000	.00094	323e	QT	Do.
29S 39W 24DD[1]	32	7,800	.0011	246e	QT	Do.
30S 37W 02BA[1]	12	4,000	.00014	326e	QT	Do.
30S 37W 19AA[1]	31	7,500	.00029	244e	QU	Do.
30S 37W 26DA	67	19,400	.0032	290e	QT	Do.
30S 38W 30AC	146	45,100	.00044	308e	QT	Do.
31S 07W 18BAC	NR	12,200	NR	NR	QU	Bayne (1960)
32S 06W 04DCC	NR	7,100	NR	NR	QU	Do.
32S 07W 12CCA	NR	7,200	NR	NR	QU	Do.

1 Well does not fully penetrate the aquifer.