PREDICTIVE ANALYSES OF GROUND-WATER DISCHARGES IN THE WILLOW CREEK WATERSHED, NORTHEAST NEBRASKA

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations 78-67

Prepared in cooperation with
Lower Elkhorn Natural Resources District and
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16. Abstracts Ground-water discharge to Willow Creek in northeast Nebraska was predicted with a digital model of the ground-water/surface-water system. Recharge and irrigation requirements were determined with a model of the soil zone. The regional aquifer is Pliocene and Pleistocene sands and gravels. Water in the regional aquifer is unconfined in the western part of the watershed and confined in the eastern part. The confining layer is Pleistocene eolian silts with very fine sand interbeds overlying a basal clay. Where the regional aquifer is unconfined, perennial flow of Willow Creek is sustained by ground-water discharge. Where it is confined, the low hydraulic conductivity of the confining beds isolates the regional aquifer from Willow Creek. Adequate agreement between simulated and observed streamflows and water levels during 1975 and 1976 was obtained by modifying initial estimates of hydraulic conductivity and specific storage. The future perennial flow of Willow Creek was simulated by superimposing six patterns of ground-water withdrawals upon variations in recharge for a monthly climatic sequence identical with the period 1931-34. These analyses showed that the perennial monthly flows would be less than 12 cubic feet per second at least 50 percent of the time.

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July 1978

UNITED STATES DEPARTMENT OF THE INTERIOR

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GEOLOGICAL SURVEY

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The International System (SI) is a consistent system of metric units adopted by the Eleventh General Conference of Weights and Measures in 1960. Selected factors for converting U.S. customary units used in this report to SI metric units are given below.

Multiply U.S. customary units	By	To obtain SI units
acre	0.0040	square kilometer (km^2)
acre-foot (acre-ft)	1233	cubic meter (m^3)
acre-foot per year (acre-ft/yr)	1233	cubic meter per year (m^3/yr)
foot (ft)	.3048	meter (m)
foot per day (ft/d)	.3048	meter per day (m/d)
foot per year (ft/yr)	.3048	meter per year (m/yr)
$foot^3$ per second (ft^3/s)	28.3162	liter per second (L/s)
inch (in)	25.4	millimeter (mm)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
mile (mi)	1.609	kilometer (km)
$mile^2$ (mi^2)	2.590	$kilometer^2$ (km^2)
degree Fahrenheit (°F)	(°F -32) 5/9	degree Celsius (°C)

PREDICTIVE ANALYSES OF GROUND-WATER DISCHARGES IN THE WILLOW CREEK WATERSHED, NORTHEAST NEBRASKA

By Jack T. Dugan and Eric G. Lappala

Abstract

Ground-water discharge to Willow Creek, which drains a 204 mi^2 watershed in northeast Nebraska was predicted for six combinations of conditions of climate and ground-water development. A digital model of the ground-water/surface-water system was stressed with recharge and withdrawal functions determined from a linear reservoir model of the soil zone.

The geohydrologic system includes a regional aquifer which contains sand and gravel ranging in age from Pliocene through Pleistocene. The regional aquifer is unconfined in the western part of the watershed and confined in the eastern part. The confining layer, or blue clay, in the east is of Pleistocene age and comprises principally saturated eolian silts with very fine sand interbeds overlying a basal lacustrine clay. Where unconfined conditions exist in the regional aquifer, perennial flow of Willow Creek is sustained principally by ground-water discharge. Where confined conditions exist, the low hydraulic conductivity of the blue clay effectively isolates the regional aquifer from Willow Creek.

The stream-aquifer models were tested by comparing simulated against measured water levels and streamflow during 1975 and 1976. Agreement between simulated and observed values was obtained by modifying initial estimates of aquifer hydraulic conductivity and specific storage.

Future ground-water discharge to Willow Creek (base flow) was simulated by super-imposing different patterns of ground-water withdrawals upon variations from average recharge for a monthly climatic sequence identical with the period 1931 to 1954. Frequency analyses of the simulated monthly base flows showed that they would be less than 12 $feet^3$ per second at least 50 percent of the time under all but completely undeveloped conditions. Simulated water-level declines in the regional aquifer at the end of the pumping season after a drought period similar to that of the 1930's were less than 5 feet where unconfined conditions exist and were more than 80 feet where confined conditions exist.

Introduction

Purpose and scope

This study was made to determine the future streamflow of Willow Creek, which has a surface drainage of 204 mi^2 in northeast Nebraska (fig. 1). Results of this study will help to evaluate the feasibility of constructing a recreational reservoir southwest of the town of Pierce, Nebraska. The proposed reservoir would have a normal storage capacity of 14,215 acre-ft and would cover approximately 1,075 acres. Preliminary engineering estimates by Hoskins-Western-Sonderegger, Inc. (Written commun., undated) showed that an average base flow of 15 ft^3/s for 3 years would fill the reservoir to a pool elevation of 1,630 ft and that a base flow of 10 ft^3/s would maintain this level.

Objective

The objective of the study was to predict the base flow of Willow Creek upstream from the proposed reservoir under variable climatic conditions and ground-water withdrawals for irrigation.

Acknowledgments

This study was a cooperative effort of the U.S. Geological Survey, the Lower Elkhorn Natural Resources District, and the Conservation and Survey Division of the University of Nebraska. Extensive use was made of geologic data and interpretations supplied by D. W. Thomsen of Hoskins-Western-Sonderegger, Inc. (Written commun., undated). Cooperation of landowners during collection of field data is gratefully acknowledged. Illustrations were drafted by personnel of the Conservation and Survey Division of the University of Nebraska.

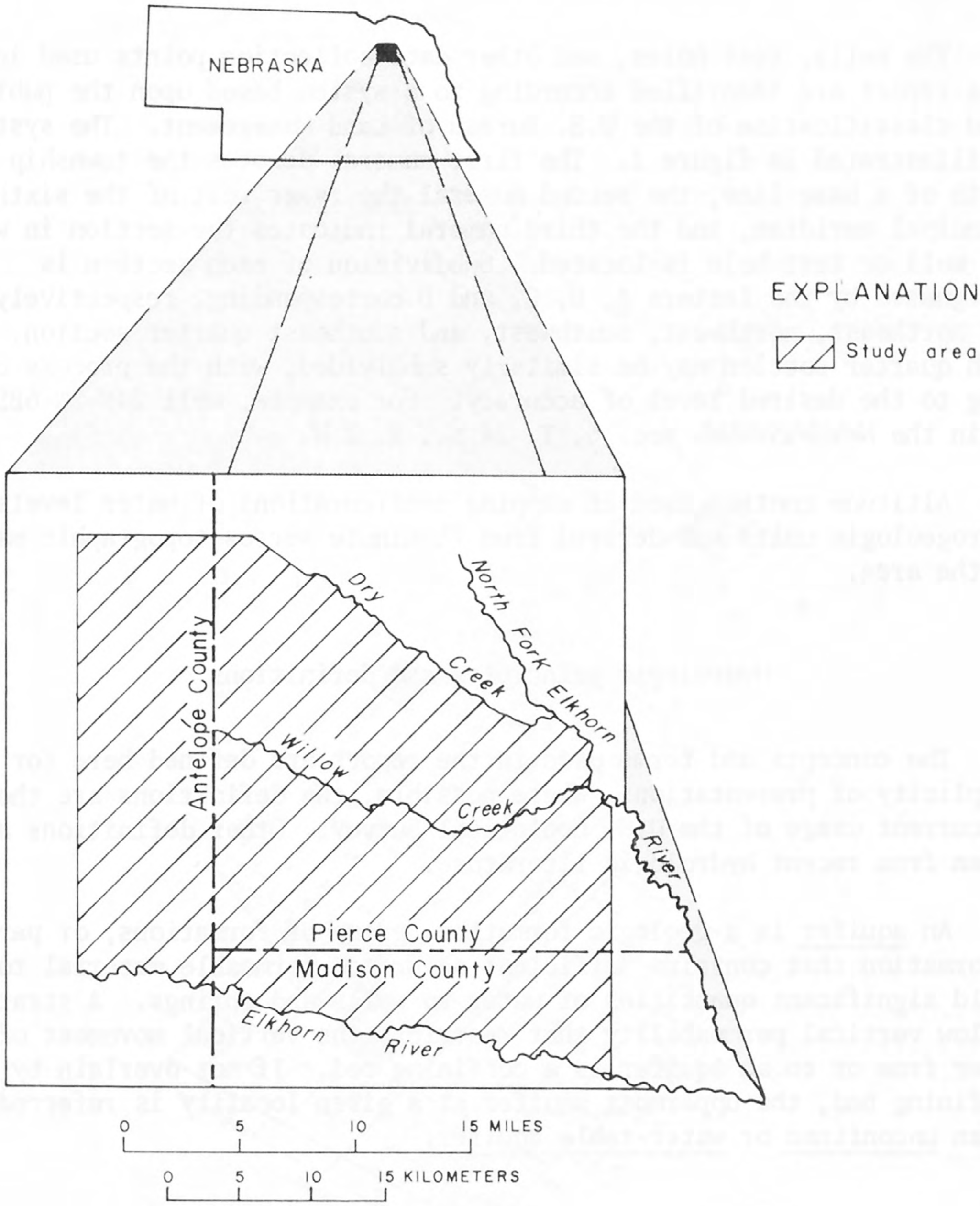

Figure 1.--Location of the Willow Creek watershed study area.

System used to identify data points

The wells, test holes, and other data-collection points used in this report are identified according to a system based upon the public land classification of the U.S. Bureau of Land Management. The system is illustrated in figure 2. The first numeral denotes the township north of a base line, the second numeral the range west of the sixth principal meridian, and the third numeral indicates the section in which the well or test hole is located. Subdivision of each section is designated by the letters A, B, C, and D corresponding, respectively, to the northeast, northwest, southwest, and southeast quarter section. Each quarter section may be similarly subdivided, with the process continuing to the desired level of accuracy. For example, well 24N-2W-6BBBB is in the NW¼NW¼NW¼NW¼ sec. 6, T. 24 N., R. 2 W.

Altitude control used in mapping configurations of water levels and hydrogeologic units was derived from 7½-minute series topographic maps of the area.

Hydrologic principles and definitions

The concepts and terms used in the report are defined here for simplicity of presentation. Where possible, the definitions are those of current usage of the U.S. Geological Survey. Other definitions are taken from recent hydrologic literature.

An <u>aquifer</u> is a geologic formation, group of formations, or part of a formation that contains sufficient saturated permeable material to yield significant quantities of water to wells and springs. A stratum of low vertical permeability that restricts the vertical movement of water from or to an aquifer is a <u>confining bed</u>. If not overlain by a confining bed, the uppermost aquifer at a given locality is referred to as an <u>unconfined</u> or <u>water-table aquifer</u>.

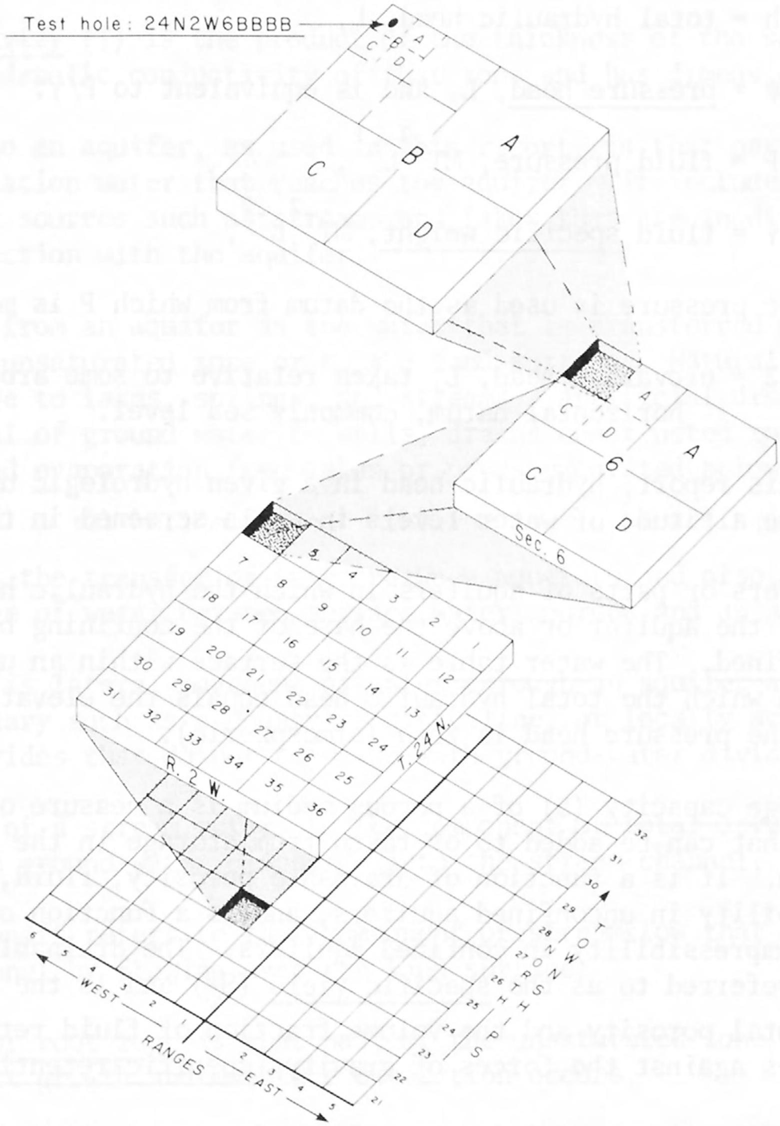

Figure 2.--System used to identify data points.

Hydraulic head (h) is an expression for the potential energy of a fluid and, when used in discussing isothermal and isohaline fluids in porous media, is expressed as:

$$h = \psi + Z,$$

where h = total hydraulic head, L,

ψ = pressure head, L, and is equivalent to P/γ.

where P = fluid pressure, $MT^{-2}L^{-1}$,

γ = fluid specific weight, $MT^{-2}L^{-2}$,

Atmospheric pressure is used as the datum from which P is measured.

Z = elevation head, L, taken relative to some arbitrary horizontal datum, commonly sea level.

In this report, hydraulic head in a given hydrologic unit is equivalent to the altitude of water levels in wells screened in the unit.

Aquifers or parts of aquifers in which the hydraulic head is above the top of the aquifer or above the base of the confining bed are said to be confined. The water table is the surface within an unconfined aquifer on which the total hydraulic head equals the elevation head, or at which the pressure head is zero (atmospheric).

Storage capacity (S) of a porous medium is a measure of the amount of fluid that can be added to or taken from storage in the pore space of the medium. It is a function of drainable porosity, fluid, and aquifer compressibility in unconfined aquifers, and is a function of fluid and aquifer compressibility in confined aquifers. The drainable porosity is commonly referred to as the specific yield (S_y) and is the difference between total porosity and the volume fraction of fluid retained in the pore spaces against the forces of gravity (specific retention).

Specific storage (S_s) is the volume of water derived from fluid expansion and aquifer compressibility per unit volume of aquifer for a unit decline in hydraulic head. The artesian storage coefficient (S) is the product of specific storage and aquifer thickness and thus is a measure of the water derived from storage over the entire aquifer thickness.

Hydraulic conductivity (K) is a measure of the volume of fluid that will move in unit time under a unit gradient in total hydraulic head through a unit area at right angles to the direction of flow. It is a property of both the porous media and the fluid and has the dimensions of velocity (LT^{-1}). In the unsaturated zone, hydraulic conductivity is also a function of moisture content.

Transmissivity (T) is the product of the thickness of the saturated zone and the hydraulic conductivity of that zone and has dimensions L^2T^{-1}.

Recharge to an aquifer, as used in this report, is that part of downward percolation water that reaches the aquifer. It includes seepage from free water sources such as streams and lakes that are in direct hydraulic connection with the aquifer.

Discharge from an aquifer is the water that is transferred from the aquifer to the unsaturated zone or to the land surface. Natural discharge includes seepage to lakes, springs, and streams. Artificial discharge includes removal of ground water by wells, drains constructed to intercept the aquifer, and evaporation from lakes or ponds excavated below the water table.

Leakage is the transfer of water between aquifers and also, in this report, exchange of water between surface-water sources and an aquifer.

Underflow is lateral movement of water through an aquifer across an arbitrary boundary such as a county or state line, or locally across topographic divides that do not coincide with ground-water divides.

Base flow of a stream refers to the component of total streamflow attributable to ground-water discharge into the stream channel.

Surface runoff refers to that component of streamflow that enters the stream channel by flowing over the land surface.

The soil or root zone is that part of the unsaturated zone in which most active root growth and moisture extraction occurs.

Field capacity (F_c) is an arbitrary value of specific retention corresponding to a pressure head of minus one-third of an atmosphere or 11.31 ft of water (Linsley, Kohler, and Paulhus, 1958). At moisture contents less than field capacity, no significant drainage of water by gravity is assumed to occur.

Wilting point (W_p) is an arbitrary value of moisture content corresponding to a pressure head of minus 15 atmospheres or 508.9 ft of water (Linsley, Kohler, and Paulhus, 1958). Permanent wilting and consequent cessation of transpiration by most plants occurs at moisture contents less than the wilting point.

Available water capacity is the difference between field capacity and wilting-point moisture content.

Evapotranspiration (ET) is the combined processes of evaporation from free water and bare soil surfaces and transpiration by plants. Evapotranspiration is a consumptive use of water at a given locality because the water vapor is usually removed from the immediate area by atmospheric circulation.

Potential evapotranspiration (PET) is defined by Linsley, Kohler, and Paulhus (1958) as the amount of water that would be evaporated from bare soil and transpired by plants if neither is under moisture stress. That is, water availability is not a constraint on the amount of potential evapotranspiration. As such, PET is a measure both of the energy available to drive the evaporative processes and of the efficiency of the mechanisms that remove evaporated water molecules from the immediate vicinity of an evaporative surface such as leaf stomata or bare soil.

Consumptive-irrigation requirement (CIR) is the amount of water required to meet ET demand after soil moisture and infiltrated precipitation have been drawn upon.

Infiltration (I) is that part of precipitation and applied surface water that enters the soil zone.

Deep percolation is water that leaves the soil zone and enters into the underlying part of the unsaturated zone.

Previous studies

A report on the Elkhorn River basin by Bentall and others (1972) included a useful overview of the study area within the context of a larger region. Geologic cross sections, water-table maps, potential well yields, and streamflow regimes were presented.

The Platte River basin Level B Study (Missouri River Basin Commission, 1975) modeled the hydrologic system of the Elkhorn River basin, which includes Willow Creek basin. The report predicted water-level declines in excess of 40 ft by the year 2020 for the northwestern parts of the study area, under conditions of complete irrigation development by 2000. Large declines in base flows of the Elkhorn and North Fork Elkhorn Rivers were also predicted. Future streamflow in Willow Creek was not discussed in the report.

Brogden and others (1976) discussed the water resources of Pierce County in a report that included maps of the water table, depth to water, and transmissivity of the principal aquifer. Their report, although too generalized for use in this study, concluded that Willow Creek basin has potential for additional irrigation development.

A report by Hoskins-Western-Sonderegger, Inc. (undated) provided a preliminary feasibility study of the proposed Willow Creek Dam. The study analyzed the soils, geology, climate, and hydrology of the watershed and the damsite. A detailed examination was made of seepage and evaporation losses from the proposed reservoir. Surface-runoff characteristics in the basin were analyzed on the basis of rainfall intensity and frequency. It was concluded that a minimum of 2.5 in of rainfall in a 24-hour period was necessary to produce appreciable runoff.

Description of the study area

The study area covers approximately 382 mi^2 in northeast Nebraska (fig. 1) and includes parts of southwestern Pierce County, east-central Antelope County, and extreme northern Madison County. It includes the basin of Willow Creek upstream from that stream's point of effluence to its mouth and is bounded on the north by Dry Creek and on the south by the Elkhorn River.

Physiographically, the area lies in a transitional zone between sandhills to the west and dissected loess and till plains to the east and north. The soils, geology, and hydrology reflect the complexity of this transition. The natural vegetation is mid-to-tall grass prairie. Currently, most of the area is planted to corn. Irrigated agriculture has increased substantially in the project area during the last 20 years. By spring 1976, 280 irrigation wells had been installed in the area (fig. 3), a mean density of 0.7 wells per $mile^2$.

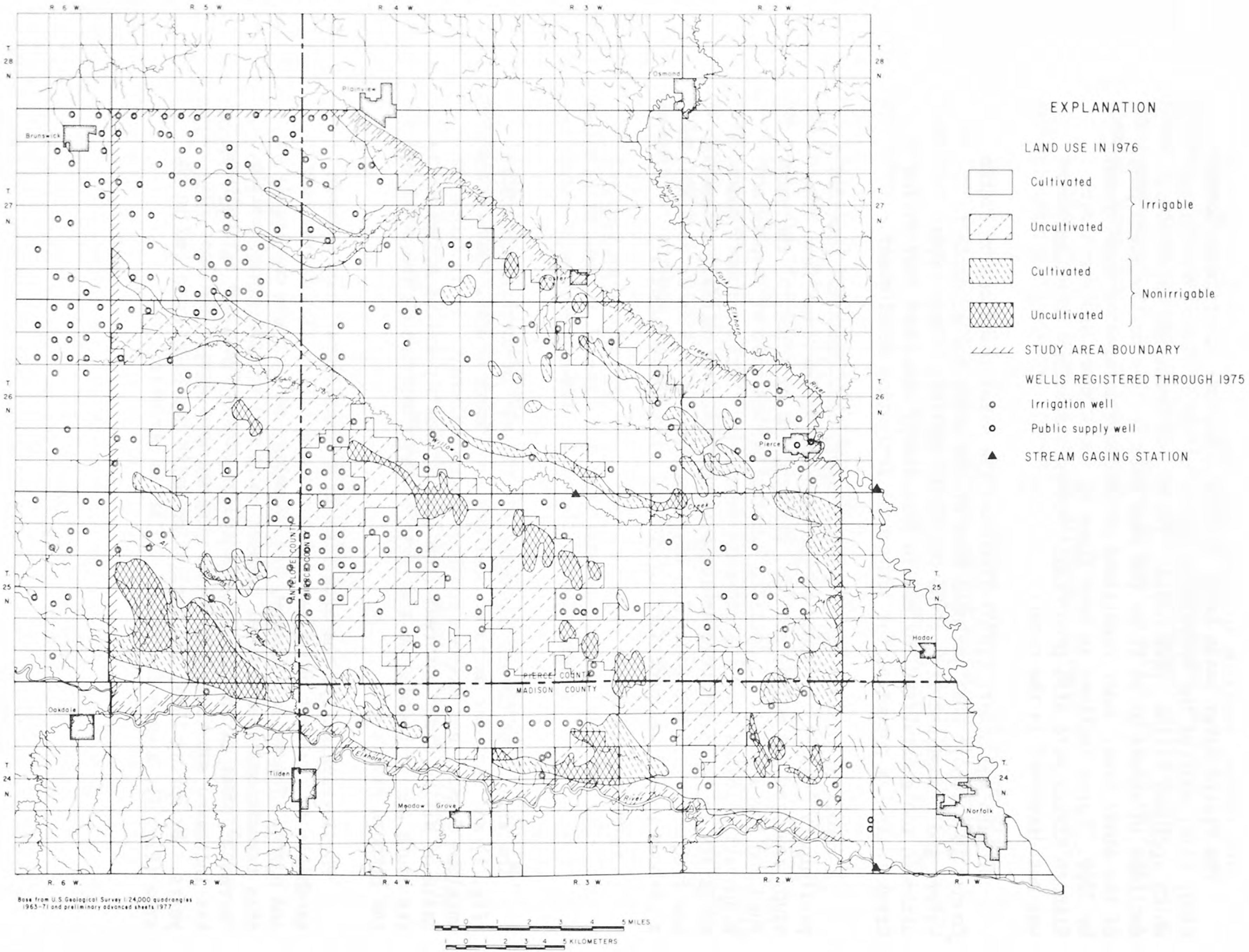

Figure 3.--Land use in 1976 and locations of registered irrigation and public-supply wells.

The area is drained primarily by Willow Creek, Dry Creek, and the Elkhorn River. Some parts of the area, however, exhibit no surface drainage. Sandy soils that cover most of the area inhibit overland runoff. Much of the streamflow is derived from ground-water discharge.

Topography and geomorphology

The study area lies along the eastern margin of the High Plains Section of the Great Plains Province (Fenneman, 1931). A portion of the area is in the sandhills and the remainder is dissected loess plains. Immediately to the east is the Dissected Till Plains Section of the Central Lowland.

The sandhills part of the area is an eastern outlier of the main sandhills region in north-central Nebraska. The hills are generally smaller in size than those in the main sandhills region and range from low hummocks to somewhat higher and larger round-topped sand dunes. The interdune valleys slope gently and have only slight relief. The drainage network is poorly developed in the sandhills area with little surface drainage. The high permeability of the sandy soils readily allows infiltration of precipitation to the zone of saturation.

The eolian deposits of the dissected loess plain have been eroded by fluvial processes. Loess deposits in the study area are as much as 75 ft thick. Relief on the loess is somewhat greater than in the sandhills area, and the topography can be classified as rolling to steep. The silts and clays comprising the loess are more easily eroded than the sandy materials. Runoff is greater from areas underlain by loess because the material is less permeable and the drainage network is better developed.

Transitional areas exhibiting characteristics of both the sandhills and the dissected loess plains exist where loess and sand are interbedded. Eolian sands have encroached on the loess deposits, creating a surface of sandy loams and loamy fine sands. These areas are less rolling and have higher infiltration rates than the dissected loess plains. Subsequent analysis of the soil zone indicates that from 10 to 15 percent of the precipitation received annually infiltrates to the zone of saturation in these transitional areas.

A small portion of the study area may be classified as bottom land, which includes alluvial plains along the streams and wet meadows in the interdune valleys. The bottom land is generally highly permeable, allowing high infiltration to the zone of saturation. Often, however, the bottom land has a shallow water table that allows ground water in storage to be returned to the atmosphere by evapotranspiration.

Soils

The soils of the study area are very complex, reflecting the transitional nature of the physiography. There are three major upland soil groups: (1) the Valentine and Thurman group, which is typical of sandhills soils; (2) the Nora group, which is principally loessial soils; and (3) the Boelus-Lorretto group, which comprises soils formed in interbedded loess and sand. The latter usually consists of a sand mantle over loess or loesslike silts. The bottom-land and terrace soils reflect the kind of materials of the adjacent uplands, ranging in texture from sandy to silty clay loam. Table 1 indicates some of the significant hydrologic properties of the major soils in the basin (Schulte and others, 1976).

Figure 4 shows the spatial distribution of the soils (adopted from Schulte and others, 1976, Hayes and others, 1924, and Goke and others, 1920). The complex arrangement of the three major upland soils and their terrace and bottom-land counterparts is apparent. There seems to be a slight predominance of sandy soils (Valentine and Thurman) over finer textured and transitional soils (Nora and Boelus-Loretto).

The typical soil of the basin is highly permeable, with a moderate to low field capacity (table 1). Steep slopes are significant in producing runoff only on the very fine textured soils (Nora and Crofton), which are limited to small areas of the eastern and southwestern parts of the study area. A preliminary study by Hoskins-Western-Sonderegger, Inc. (undated) concluded that most soils in the area are highly absorbent with runoff from precipitation and irrigation limited to isolated areas and to time periods when soils are frozen or rainfall rates are very high.

Table 1.--Selected characteristics of the major soils in Willow Creek basin[1/]

Soil	Texture of solum	Topographic position	Slope (percent)	Available water capacity of soil (volumetric fraction)	Saturated hydraulic conductivity (ft/d)
Nora	Silt-loam	Uplands	0-17	0.19	1.2-4.0
Crofton	Silt-loam	Uplands	2-60	0.22	1.2-4.0
Boelus-Loretto	Loam	Uplands	0-11	0.18	1.2-12.0
Thurman	Loamy sand	Uplands	0-20	0.11	4.0-40.0
Valentine	Sand	Uplands	0-60	0.08	12.0-40.0
Hord	Silt loam	Terrace	2-5	0.21	1.2-4.0
Ortello	Sandy loam	Terrace	2-11	0.14	4.0-40.0
Lawet	Silty clay loam	Bottom land	0-3	0.19	0.4-4.0
Loup-Orwet-Elsmere	Loamy sand	Bottom land	0-3	0.11	1.2-40.0

[1/] From Soil Survey of Pierce County, Nebr. (Schulte and others, 1976).

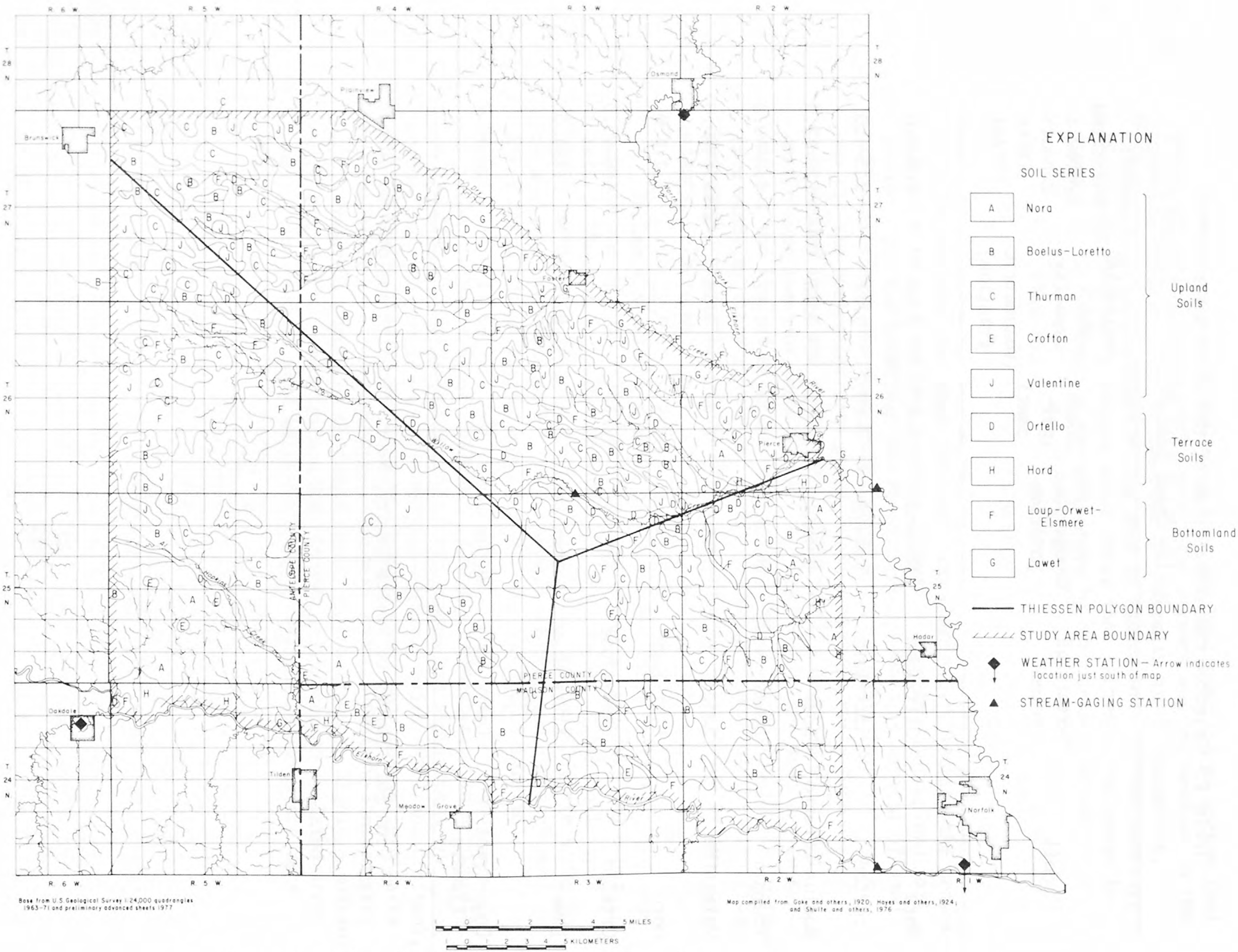

Figure 4.--Distribution of soils and boundaries of Thiessen polygons.

Vegetation and crops

The natural vegetation of the area is mainly mid-to-tall grasses, with some woody growth along watercourses and in areas of rough topography. The native grass types of the sandy areas are the Sandhills-Prairie grasses (big and little bluestem, sandreed, and needlegrasses). The Tallgrass-Bluestem-Prairie grasses (big and little bluestem, switchgrass, and indiangrass) predominate on areas underlain by the finer textured soils. Cottonwood, willow, elm, and ash grow adjacent to the streams. In isolated areas of the dissected uplands north of the Elkhorn River some outliers of the Eastern Deciduous Forest (oak and elm) occur (Kaul, 1975). Cultivation and overgrazing have removed most of the original vegetation. In overgrazed areas, shorter grasses and herbaceous plants have replaced the taller grasses. The advent of center-pivot irrigation systems has enabled cultivation of grassland that was too sloping or too sandy to be irrigated with gravity systems.

As figure 3 shows, approximately 60 percent of the agricultural land in the study area is under cultivation (Nebraska Department of Agriculture, 1970-76). In 1975, 73 percent of the tillable cropland of the study area was planted to row crops, principally corn and soybeans; 15 percent to small grains, mainly rye and oats; and 12 percent to alfalfa.

The water requirements of the different types of native vegetation vary considerably. Most of the grasses are cool-season species that have maximum water demand in spring and fall. The warm-season species have a higher water demand during the hottest summer months. Water demands by most native species are at minimum in winter, late fall, and early spring. The average growing season is approximately 150 days. Cultivated crops differ greatly in their moisture requirements. Water demand of row crops (corn and soybeans) is highest during June, July, and August. Water demand of the small grains is highest in spring and early summer. Fall-sown small grains have a secondary moisture demand during that season. The dormant season for most plants in the region is approximately 7 months.

Climate

The general climate of the study area can be described as subhumid continental or Dfa in the Köppen classification (Trewortha, 1955). However, climatic extremes are more hydrologically significant than mean conditions. The climate from year to year may range from humid to semiarid. At Norfolk, just south of the area, the mean annual precipitation from 1931 to 1975 was 24.32 in, with a standard deviation of 4.38 in. During this period, precipitation exceeded 30 in for 3 years and was less than 20 in for 10 years. The greatest amount, 37.05 in, was in 1944 and the smallest amount, 15.81 in, was in 1953. Under mean climatic conditions, 75 percent of the precipitation occurs from April through September.

The composite mean annual temperature for three climatic stations (fig. 4) near the area is 48.3°F. January, the coldest month, averages about 19.2°F and July, the warmest month, 75.4°F. The frost-free season is approximately 150 days. May 5 is the mean date for the last freeze in spring and October 1 is the mean date for the first freeze in fall. However, the last freeze in spring has occurred as early as April 10 and as late as May 29; and the first freeze in fall has occurred as early as September 6 and as late as October 28.

Table 2 summarizes several monthly climatological parameters used in hydrologic analyses for the only first order weather station near the study area at Norfolk, Nebr.

Mean daily solar radiation ranges from 143 langleys (1 langley = 1 gram-cal/cm^2) in December to 585 langleys in July. Mean annual solar radiation is about 358 langleys per day. Percent of possible sunshine for the nearest observation station at Sioux City, Iowa, ranges from a mean of about 77 percent in July and August to about 53 percent in December.

Although Norfolk is the only first-order weather station near the study area, precipitation and temperature data from Osmond and Oakdale were also used to compute recharge rates and irrigation requirements as discussed in the section on the soil-zone model. Areas assumed to be represented by data from each weather station were bounded by the Thiessen polygons shown in figure 4.

Table 2.--Climatological summary for Norfolk, Nebr., 1931-75

Month	Mean temperature (°F)	Mean precipitation (in)	Lowest precipitation (in)	Highest precipitation (in)	Mean sunshine (percent)[1/]	Mean daily solar radiation (langleys)[2/]
January	18.9	0.62	0.10	2.66	59	166
February	23.9	.78	.06	3.18	57	238
March	32.8	1.37	.25	5.14	57	333
April	48.5	2.15	.27	6.01	64	464
May	60.0	3.69	.69	8.75	65	531
June	69.9	4.88	.44	12.22	71	578
July	75.7	3.18	.18	10.83	77	585
August	73.8	2.66	.37	6.92	77	519
September	63.2	2.41	.26	8.13	73	418
October	52.5	1.33	.00	4.57	68	296
November	36.3	.62	.00	3.72	56	188
December	24.2	.63	.03	2.39	53	143
ANNUAL	48.3	24.32	15.81	37.05	77	358

[1/] From nearest observation station at Sioux City, Iowa.

[2/] Computed from method by Fritz and others (1949).

Three continuous gaging stations are located in or near the study area: The North Fork Elkhorn River near Pierce; the Elkhorn River near Norfolk; and the Elkhorn River at Neligh, west of the area. Perennial flow of Willow Creek is not well documented because streamflow has been measured only on an irregular basis. Monthly streamflow information for the gaging station on Willow Creek near Foster from November 1975 through December 1977 is given in table 3.

Table 3.--Monthly streamflow of Willow Creek near Foster, from November 1975 through March 1977

Year	Month	Daily discharge (ft^3/s)			Coefficient of variation[1/]
		Mean	Maximum	Minimum	
1975	November	[2/]6.66	[2/]15	[2/]4.6	0.35
	December	7.53	11	4.6	.24
1976	January	8.71	11	5.0	.18
	February	18.2	30	5.2	.48
	March	17.3	29	11	.31
	April	11.7	15	9.6	.11
	May	8.53	15	6.5	.25
	June	6.67	10	5.4	.15
	July	5.00	7.0	3.5	.18
	August	3.77	4.9	2.6	.19
	September	4.01	5.1	2.8	.14
	October	4.27	4.9	3.8	.07
	November	4.69	5.7	3.6	.13
	December	4.78	6.0	3.5	.15
1977	January	4.25	5.0	3.7	.09
	February	4.55	5.4	3.7	.13
	March	7.82	13	5.0	.24

[1/]Coefficient of variation = standard deviation of daily flows/ mean daily flow.

[2/]Record estimated for first 12 days of month.

The gaging station on the North Fork Elkhorn River is below the confluences of Willow and Dry Creeks with the North Fork Elkhorn River. The mean discharge for 15 years (1960-75) at this station was 88.5 ft^3/s or 64,120 acre-ft/yr. It is estimated that 15 percent of the flow can be attributed to Willow Creek and 6 percent to Dry Creek.

The Elkhorn River, which borders the southern edge of the study area, is the largest perennial stream in the study area. Discharge measurements of the river at Neligh and Norfolk indicate a gain from ground-water discharge in the southern part of the study area. The mean discharge during the period 1966-75 was 242 ft^3/s at Neligh and 391 ft^3/s at Norfolk, which is 7 mi above the confluence of the Elkhorn with the North Fork Elkhorn River. Thus, the increase in flow of the Elkhorn River from Neligh to Norfolk was 62 percent during this period.

Much of the increase in flow of the Elkhorn River in the reach from Neligh to Norfolk can be attributed to flow from seven streams that enter from the south. Another stream, Al Hopkins Creek, enters from the north. The gain not accounted for by surface inflow is approximately 58,000 acre-ft annually. Ground-water discharge from the study area to this reach is estimated to be 9,600 acre-ft/yr, including the base flow of Al Hopkins Creek.

Dry Creek, which constitutes part of the northern boundary of the study area, begins perennial flow near the Antelope-Pierce County boundary. A ground-water divide exists between Dry Creek and Willow Creek. The divide is only 2 or 3 mi north of Willow Creek, therefore much of the drainage goes to Dry Creek. Estimated ground-water discharge to Dry Creek from the study area is 4,800 acre-ft/yr.

Willow Creek begins perennial flow approximately 1 mi west of the Pierce-Antelope County line and joins the North Fork Elkhorn River near Pierce. The stream is approximately 27 mi in length from point of effluence to its confluence with the North Fork Elkhorn River. Data from discharge measurements conducted in October 1975, April 1976, and November 1976 are illustrated in figure 5.

The data indicate that Willow Creek gains approximately two-thirds of its flow in the first 10 mi. The middle reach of the stream has no significant gain or loss; and in the last 8 mi of the stream's course, the remaining one-third or more of the flow is gained. The middle reach crosses an area that is underlain by a distinct hydrogeologic unit locally called "blue clay." This blue clay restricts the discharge of water from the regional aquifer to the stream.

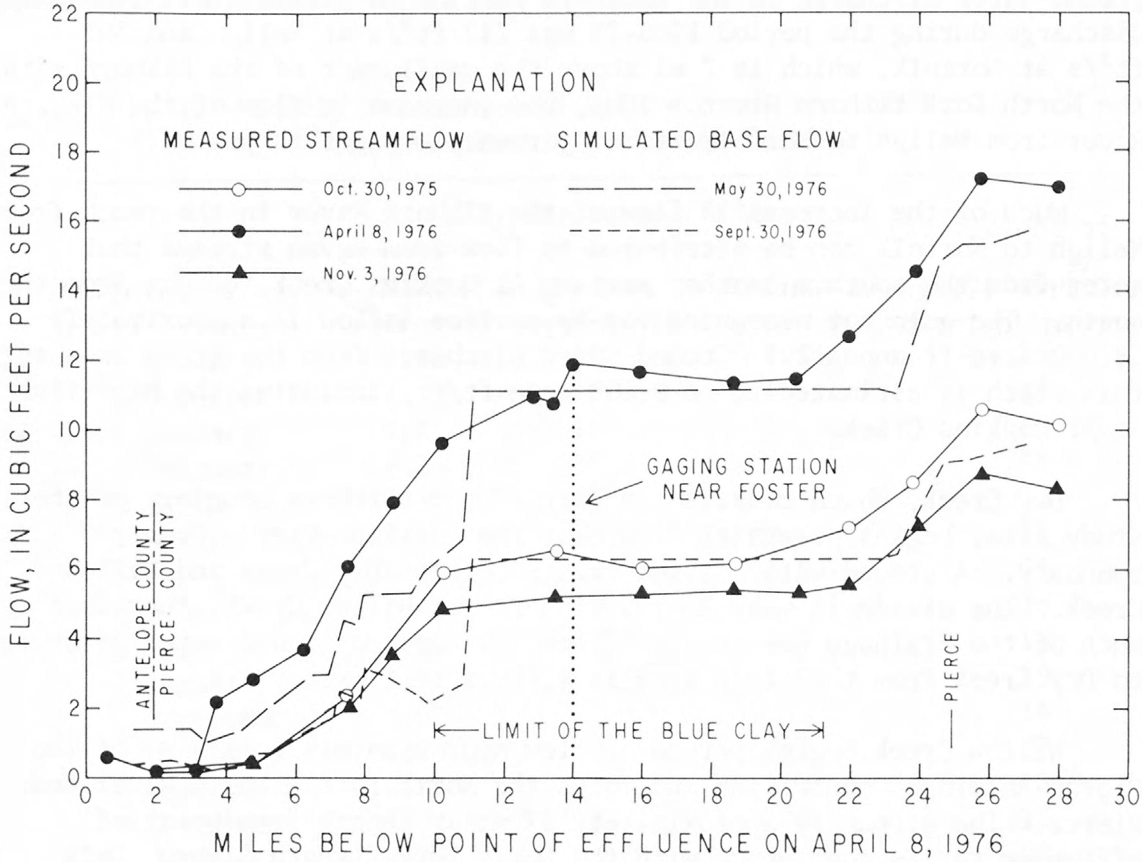

Figure 5.--Measured streamflow and computed base flow of Willow Creek.

The hydrologic system

To meet the objectives of this study required quantification of water movement between the saturated zone and the soil zone. Analysis of the saturated zone included (1) the connection between ground water and surface water, (2) amounts of recoverable water in storage, (3) movement through the saturated zone, and (4) the effects of climatic variations and ground-water withdrawals by wells on natural aquifer discharge and water in storage. Analysis of the soil zone provided values of recharge to the saturated zone and required ground-water withdrawals for irrigation as functions of the variations in climate, soils, and vegetation.

The saturated zone acts as a storage reservoir for ground water during its movement from areas of recharge to areas of discharge. The materials occurring in the saturated zone include both the permeable sands, sandstones, and gravels that comprise aquifers and the less permeable silts and clays that restrict movement of water between aquifers. These materials range in age from Pliocene through early Pleistocene. For this study the saturated zone is separated into two units based upon similar hydrologic properties rather than upon geologic age. These units are (1) the regional aquifer, which underlies the entire study area, and (2) a fine-grained unit, locally called blue clay, which overlies the regional aquifer in the eastern half of the study area. Figure 6 shows schematically the hydrologic relationship between these two units.

The blue clay

The blue clay is a hydrologic unit which overlies the regional aquifer in the eastern half of the study area. Where present, the unit is the upper confining layer for the regional aquifer. The blue clay unit comprises lacustrine clays overlain by eolian silts with local interbeds of fine sand.

The basal clay is either fluvioglacial or glacial-lacustrine in origin. The bluish color and presence of incompletely decayed organic matter in the clay indicate an anaerobic depositional environment. Figure 7 shows the configuration of the base of the blue clay and indicates that it was deposited on a gently sloping surface. The

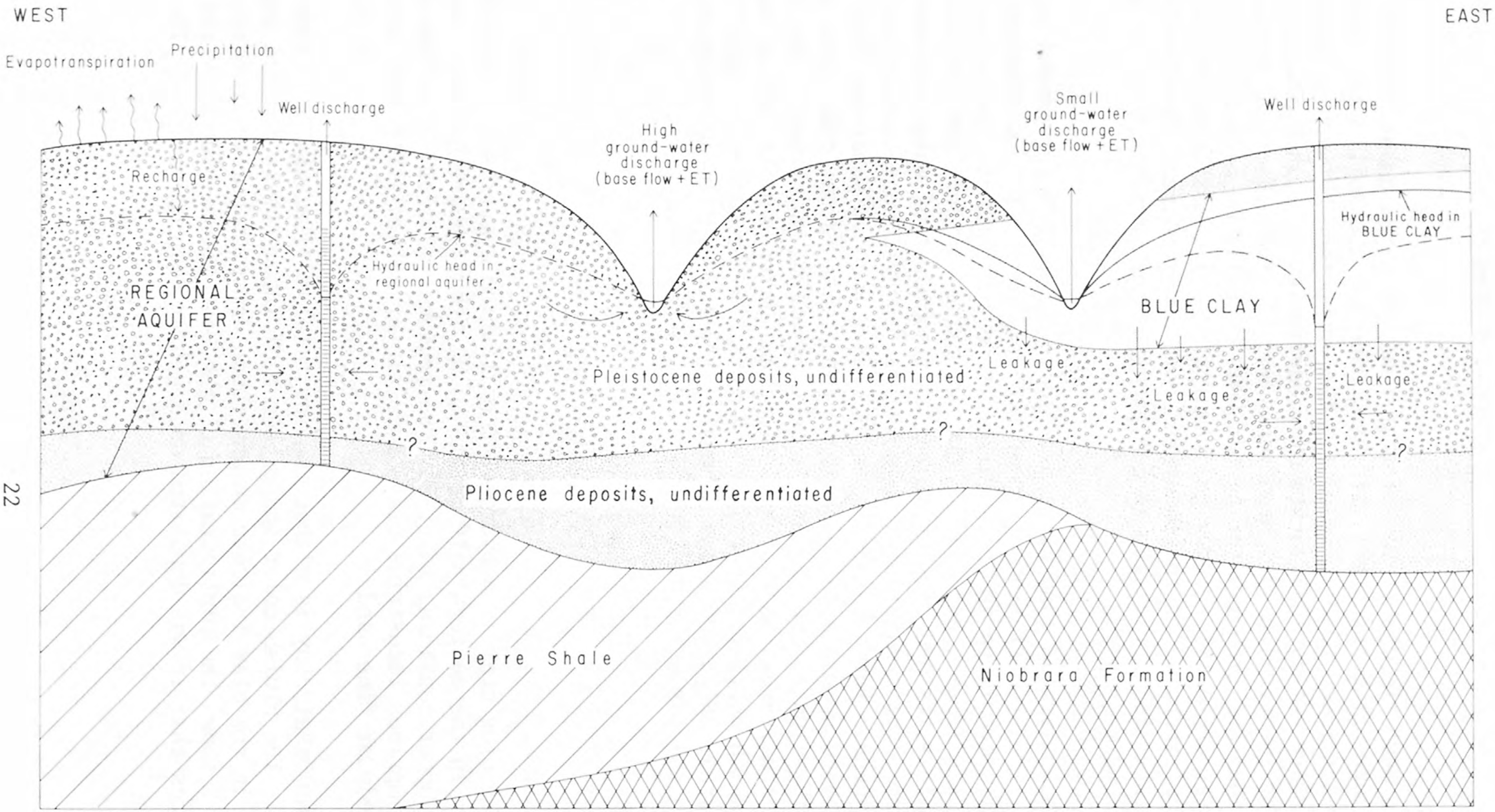

Figure 6.--Hydrologic relationship between the regional aquifer and the blue clay.

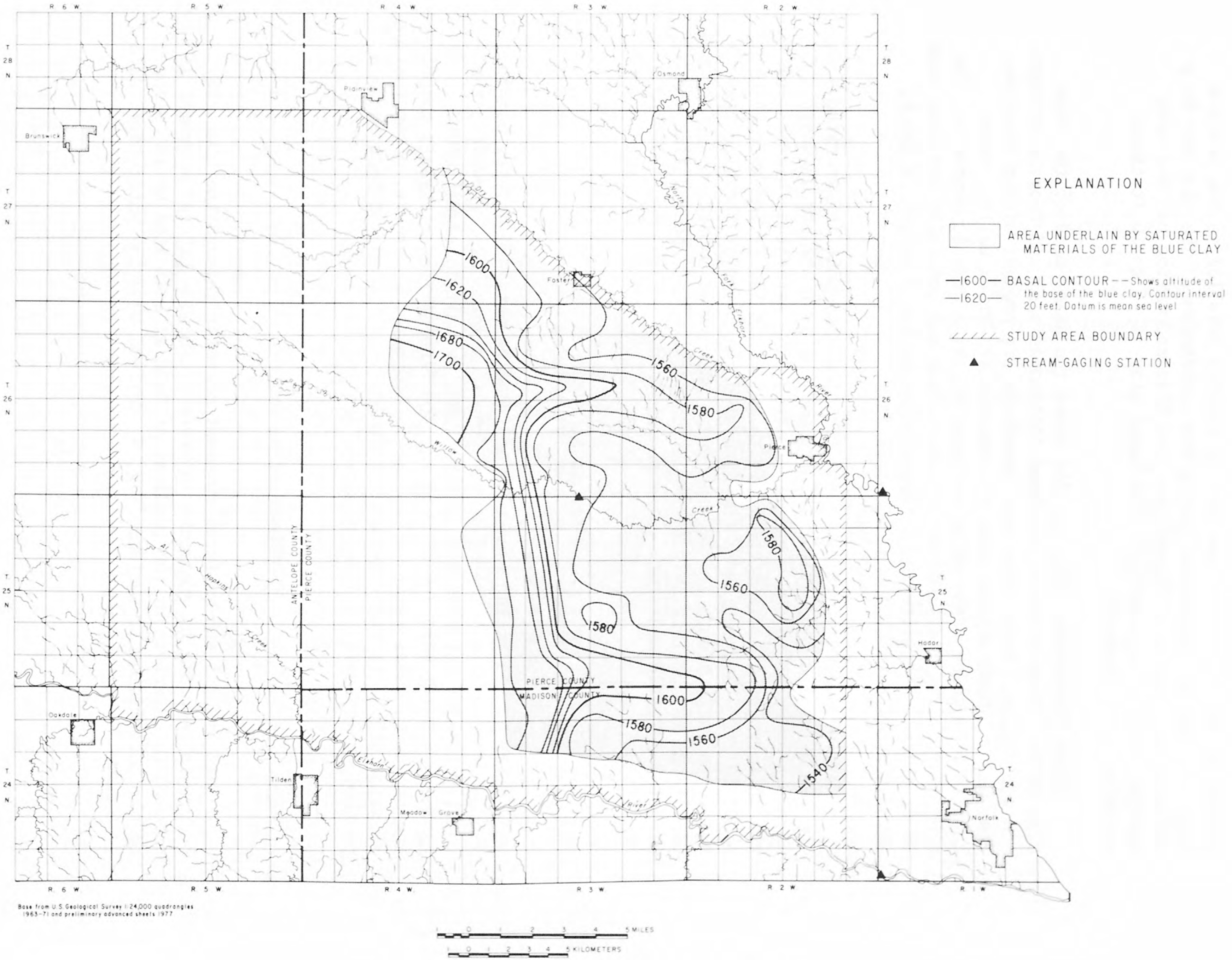

Figure 7.--Configuration of the base of the blue clay.

eastern boundary of the blue clay is roughly parallel to the western limits of Kansan Glaciation during the Pleistocene. The clay may be a periglacial deposit of Kansan age, but this assumption has not been substantiated by age dating. The basal clay is 40 to 50 ft thick wherever it is present.

Later Pleistocene deposits lie above the basal clay of the blue clay unit. These comprise Loveland (Illinoian age) and Peorian (Wisconsin age) silts. Local fine sand interbeds are present within the blue clay unit. These supply water to domestic and stock wells. Water quality from these sands often reflects the high organic content of the blue clay unit. The sands are discontinuous and generally are less than 10 ft thick. The silts and interbedded sands are 5 to 160 ft thick, with an average thickness of about 80 ft.

Most of the total thickness of the blue clay unit is saturated. The low hydraulic conductivity of the basal clay restricts downward movement of water to the regional aquifer. The saturated thickness of the unit ranges from less than 5 ft to more than 160 ft and averages about 80 ft. Figure 8 shows the saturated thickness of the blue clay. Water generally occurs in the unit under unconfined conditions. Confined conditions may be locally present in some of the interbedded fine sands used for domestic and stock water supplies.

Water in storage

Water stored in the blue clay unit can be obtained only by pumping from wells that are screened in the fine sand interbeds. However, the total volume of water stored in the blue clay unit is significant. Assuming a total porosity of 0.45 for the fine-grained materials comprising the unit, total water in storage is 2.9 million acre-ft. Computuation of water stored in the fine sand interbeds is not possible because data on their thicknesses and areal extent are insufficient.

Movement of water

Both lateral and vertical movement of water through the blue clay occurs in response to gradients in hydraulic head. Lateral movement is assumed to be driven by the gradient of head as indicated by the slope of the potentiometric surface (fig. 9) derived by using water-level data

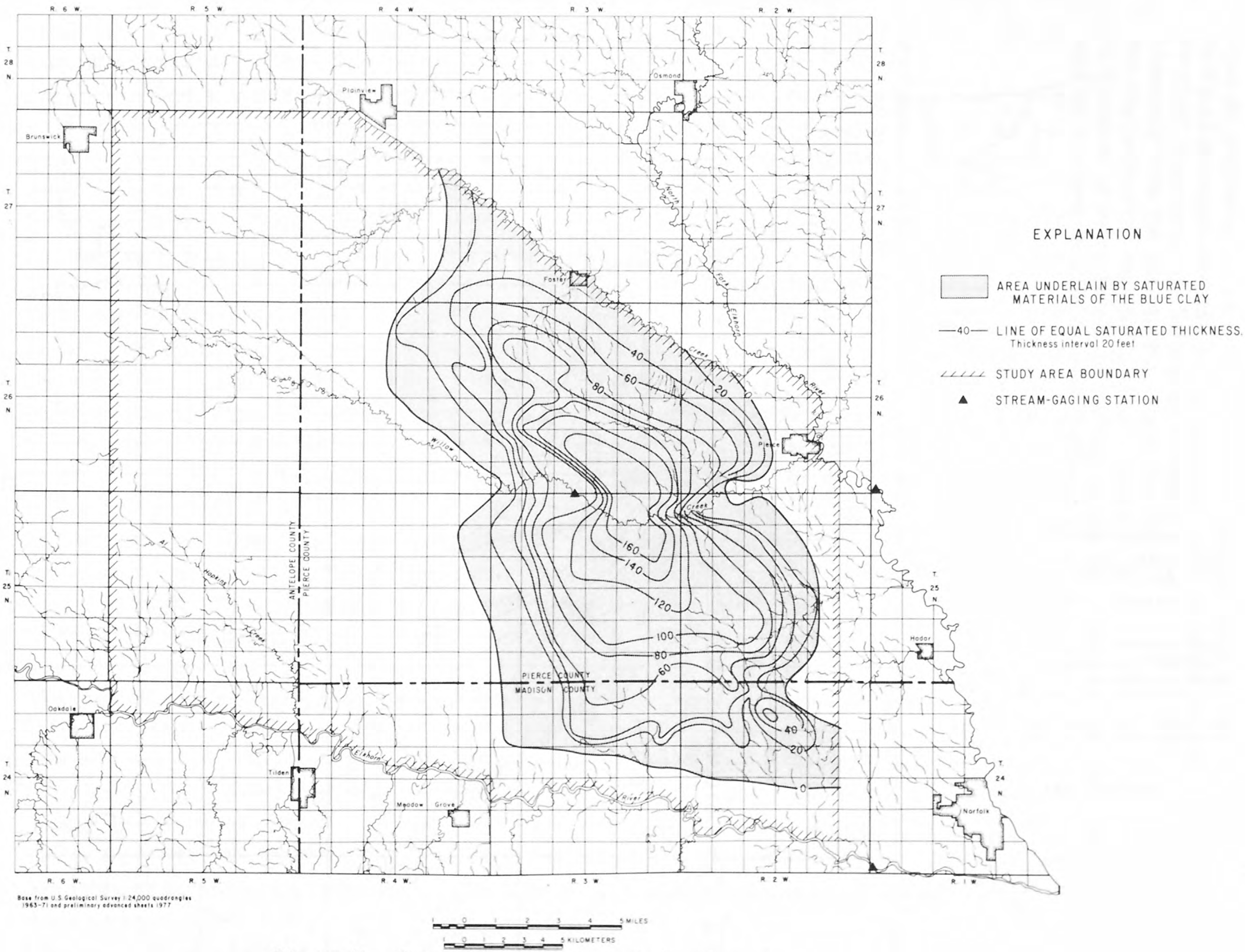

Figure 8.--Saturated thickness of the blue clay.

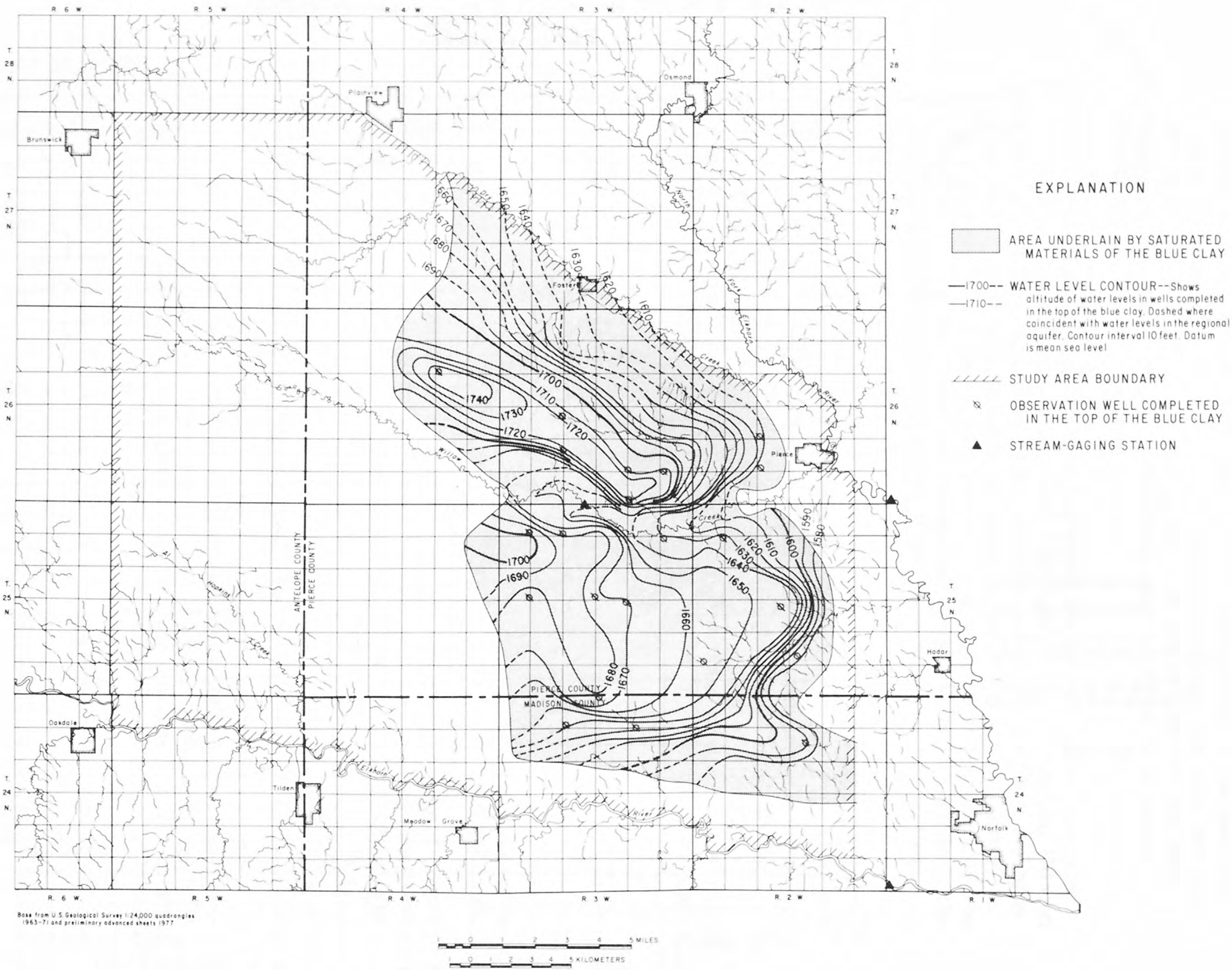

Figure 9.--Water levels in the top of the blue clay.

from 22 shallow wells installed by Hoskins-Western-Sonderegger, Inc. Vertical movement of water in the blue clay unit is caused by the vertical components of the hydraulic-head gradients. These vary continuously from the distribution shown for the top of the unit to those for the underlying regional aquifer. Vertical gradients as indicated by differences in water levels in the top of the blue clay and the regional aquifer are greater than lateral gradients by a factor of 10 to 100.

Water balance

The blue clay is recharged exclusively by deep percolation of precipitation that exceeds evapotranspiration and soil-moisture requirements. Lateral inflow and outflow are insignificant because of the low hydraulic conductivity of the unit. Similarly, discharge to perennial streams is small. This is evidenced by the lack of gain in flow of Willow Creek where it crosses the blue clay (fig. 5), even though hydraulic-head gradients toward the stream are high (fig. 9).

Most of the discharge from the blue clay occurs as evapotranspiration of ground water in shallow water-table areas and by leakage to the underlying regional aquifer. Leakage from the blue clay to the regional aquifer is about 10,900 acre-ft/yr or about 1.2 in/yr. (Computed by methods given on p. 35.) The water balance for the blue clay is considered to be: Recharge from deep percolation -ET from ground water = leakage to the regional aquifer.

The regional aquifer

The regional aquifer comprises sands and gravels of Pliocene through Pleistocene age. The upper part of the aquifer is generally fine sand to medium gravel. These materials form a fairly distinct unit over the entire study area. Figure 10 shows the north-south and east-west continuity of these materials. Locations of the sections are shown on figure 11. The thickness of the upper sand and gravel part of the aquifer ranges from 50 to 100 ft and averages about 75 ft. The lower part of the regional aquifer comprises materials that generally are finer grained than the upper part of the aquifer. These materials are interbedded silty sands, silts, and sandstones of Pliocene age. However, an identifiable sand and gravel unit about 20 ft thick occurs at the base of the aquifer in much of the study area (fig. 10).

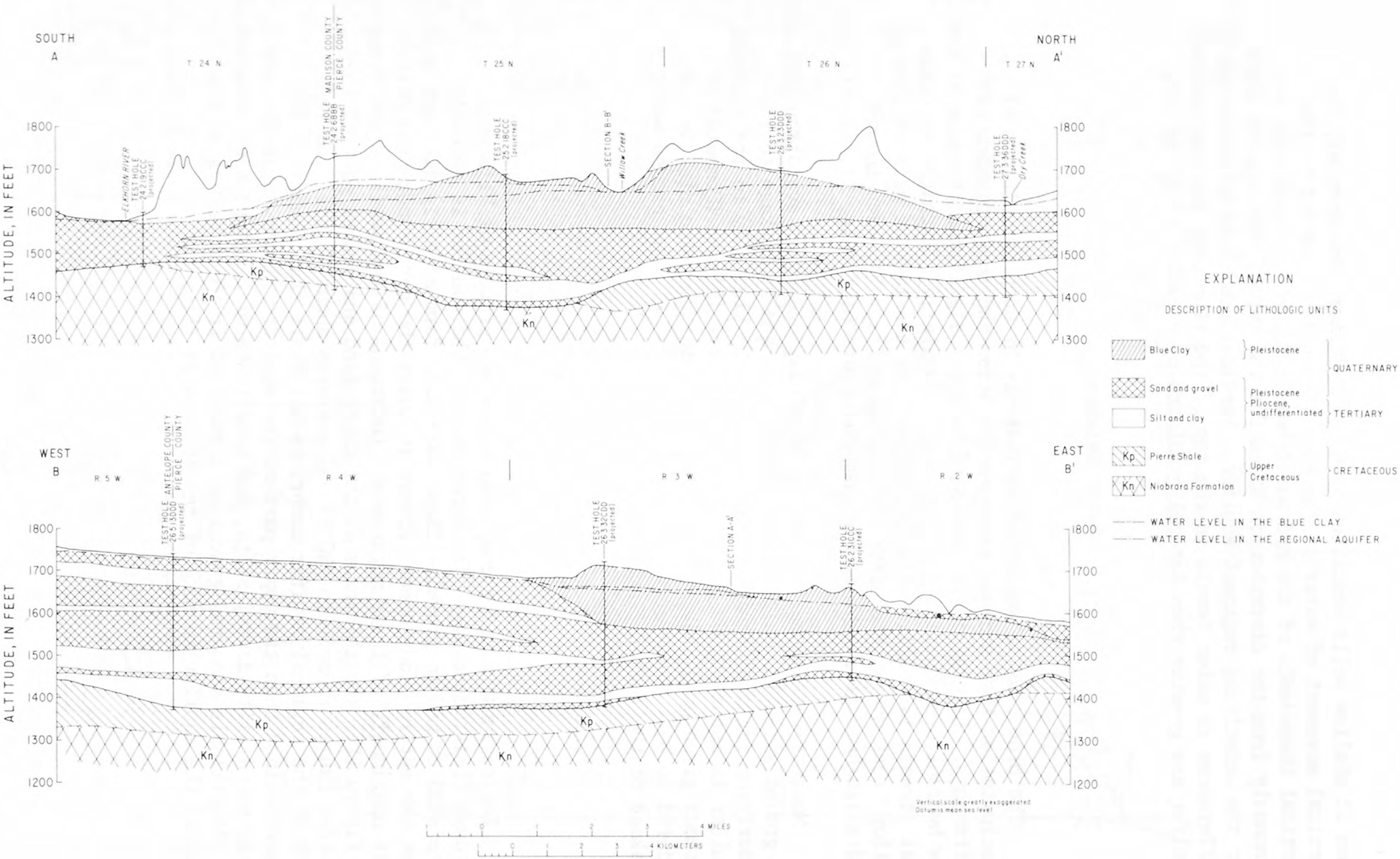

Figure 10.--Hydrogeologic sections A-A' and B-B'.

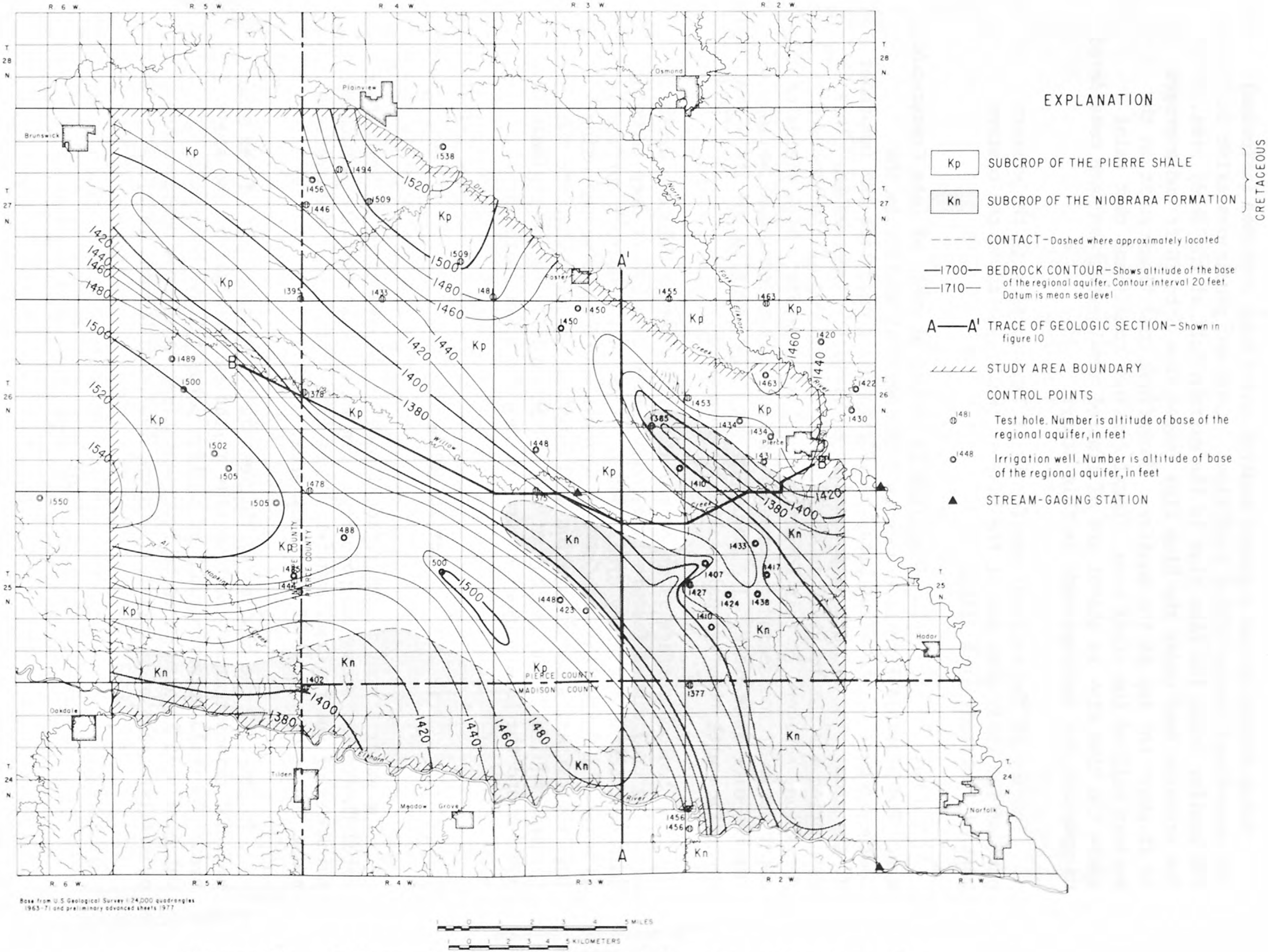

Figure 11.--Geology of the bedrock underlying the regional aquifer.

Water occurs in the regional aquifer under both confined (artesian) and unconfined (water-table) conditions. Confined conditions exist in the aquifer under the blue clay in the eastern half of the study area. The artesian head under the blue clay ranges from 5 to 70 ft and averages 40 ft above the top of the aquifer. Unconfined conditions exist in the western half of the study area. Local confined conditions that exist where the blue clay is absent are of limited areal extent and are considered unimportant for analyses made in this study.

The top of the regional aquifer is at land surface in the western part of the study area and at the base of the blue clay in the eastern part of the study area (fig. 7).

The base of the regional aquifer is shale and chalk of Late Cretaceous age. The Niobrara Formation underlies the regional aquifer in the southeast part of the study area, and Pierre Shale underlies the remainder of the area. The bedrock is assumed to be virtually impermeable. This assumption may not be valid in the area of the subcrop of the Niobrara Formation, however, as this formation supplies moderately high yields of water derived from secondary porosity of the chalk to municipal wells for the city of Norfolk. The amount of secondary porosity and its connection with the regional aquifer is unknown.

The bedrock surface is an erosional surface cut by Pliocene drainage. The predominant feature is a channel more than 100 ft deep trending southeast across the study area. Total relief on the bedrock surface is as much as 150 ft. The base of the regional aquifer was mapped using logs of test holes and irrigation wells that completely penetrate the regional aquifer. The geology of the bedrock underlying the regional aquifer and the locations of control points are shown in figure 11.

Water in storage

The water stored in the regional aquifer is a function of its saturated thickness, area, and porosity. Figure 12 shows that the saturated thickness of the regional aquifer ranges from 80 to 400 ft and averages about 200 ft over most of the study area.

In general, the aquifer is thickest in the western half of the study area and thinnest under the blue clay in the eastern half of the study area. No field determinations of aquifer porosity have been made. A value of 0.35 is assumed to be representative, based upon measurements of similar types of materials (Johnson, 1967, and Lohman, 1972).

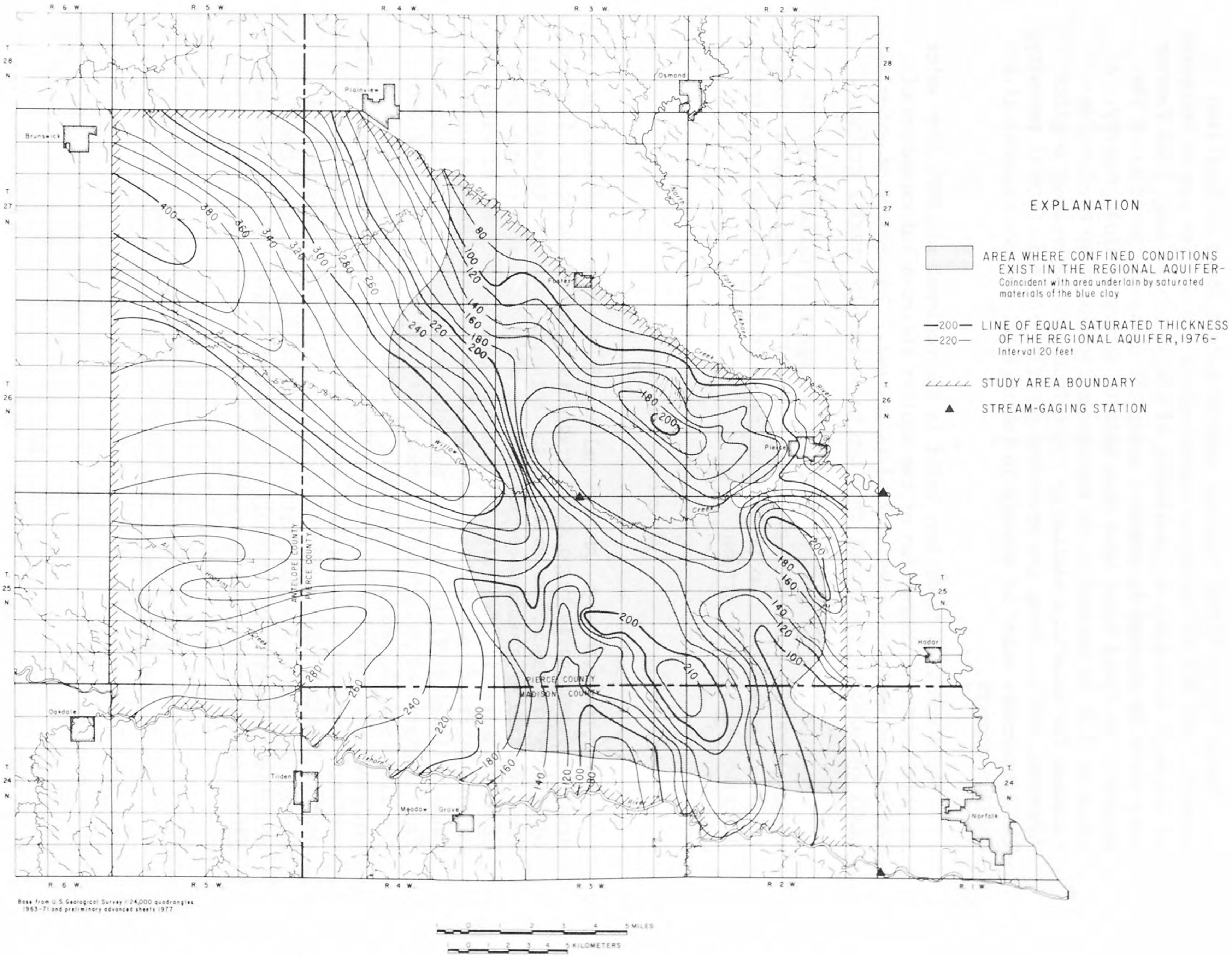

Figure 12.--Saturated thickness of the regional aquifer.

Water stored in the regional aquifer totals about 20.2 million acre-ft. Not all of the water stored within the aquifer can be recovered by drainage subsequent to dewatering of the saturated zone. The recoverable water in storage is computed using the drainable porosity of the aquifer. No field tests have been made to determine this property. A value of 0.20 is assumed to be representative, based upon values determined for materials similar to those found in the regional aquifer (Johnson, 1967). Using the drainable porosity instead of total porosity, the recoverable water in storage in the regional aquifer is about 11.6 million acre-ft.

Where confined conditions exist in the regional aquifer, some water is obtained from compression of the aquifer skeleton and expansion of the aquifer when the artesian head is lowered. This source of water is small compared to the amount that can be derived by gravity drainage.

No field measurements of specific storage are available for the regional aquifer. An initial value of 1.0×10^{-6} ft^{-1} was assumed (Lohman, 1972). This value was modified to 1.0×10^{-5} ft^{-1} based upon model testing described in a later section.

Movement of water

Water moves through the regional aquifer from areas of recharge to areas of discharge in response to gradients in hydraulic head. Isotropic conditions are assumed to exist in the regional aquifer. Consequently, the direction of ground-water movement is at right angles to the lines of equal hydraulic head (fig. 13). Forces causing movement of water are opposed by frictional forces caused by fluid viscosity and the tortuous paths followed by fluid elements through the aquifer. The resistance to flow is expressed as hydraulic conductivity and is a function of both the fluid and the medium. Hydraulic conductivities were estimated by applying an unpublished empirical method (Lappala, 1977) to rotary drill cuttings from test holes drilled in and adjacent to the study area. The empirical method, developed over many years by the Conservation and Survey Division of the University of Nebraska, has been consistently reliable in determining hydraulic-conductivity values for most unconsolidated deposits throughout Nebraska.

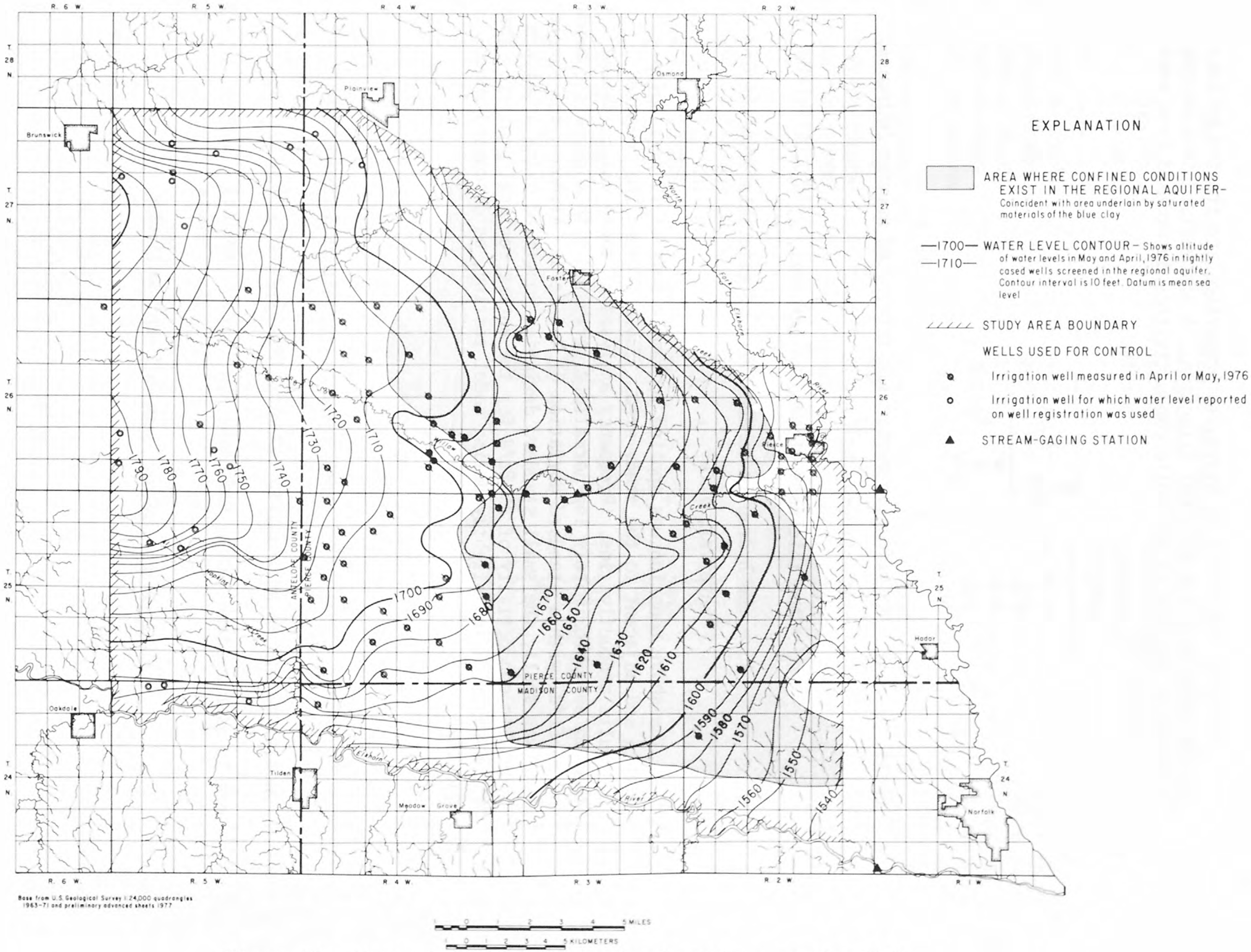

Figure 13.--Water levels in the regional aquifer, spring 1976.

Vertical variations in hydraulic properties are treated by integrating them over the vertical direction. A discretized form of this integration applied to hydraulic conductivities determined from test-hole samples is:

$$\hat{K} = \frac{\sum_{j=1}^{N} b_j k_j}{\sum_{j=1}^{N} b_j} \qquad (1)$$

where $\hat{K}$ = an effective vertically integrated average hydraulic conductivity, LT^{-1},

b_j = the thickness of sampled interval, j, L,

k_j = the hydraulic conductivity of materials from sampled interval, j, LT^{-1}.

The above equation was applied to sample logs of 19 test holes within and adjacent to the study area. Table 4 gives the computed hydraulic conductivities. Frequency analyses of these values showed them to be approximately log-normally distributed with a geometric mean of 25 ft/d, with a standard deviation of 4.6 ft/d. Although the sample size was small, the small standard deviation justified using a uniform hydraulic conductivity of 25 ft/d for the entire regional aquifer. This value was modified slightly during model testing, as discussed subsequently.

Table 4.--Test-hole data in and adjacent to the study area

Test-hole location	Depth to bedrock (ft)	Saturated thickness (ft)	Hydraulic conductivity (ft/d)
24N-2W-6BBBB	352	329	25.8
24N-4W-6BBBB	295	275	30.8
25N-2W-18CCCC	317	276	26.6
25N-4W-6BBBB	317	314	31.0
25N-5W-24AAAA	310	284	29.7
25N-6W-6BBBB	420	303	34.8
26N-2W-4BAAA	168	164	22.4
26N-2W-19BBBB	212	200	29.1
26N-3W-23DDDD	280	259	25.5
26N-3W-32DCCC	337	295	22.8
26N-5W-13DDDD	351	300	20.6
26N-7W-24AAAA	376	367	27.9
27N-3W-12DDDD	183	164	19.4
27N-4W-19BBBB	297	283	34.2
27N-5W-1AAAA	265	260	20.2
27N-5W-36DDDD	355	338	24.7
27N-6W-19BBBB	450	383	25.2
27N-7W-36DDDD	393	305	31.4
28N-7W-36DDDD	403	332	29.2

Recharge

The regional aquifer is recharge from:

1. Deep percolation of water originating as precipitation,
2. Leakage through the blue clay in the eastern half of the study area,
3. Underflow from the west through the aquifer.

Recharge from deep percolation through the soil zone is a function of precipitation, evapotranspiration, and soil water-holding capacities. The method used to compute deep percolation is described in the subsequent section on modeling the soil zone. Table 5 shows the recharge computed under mean climatic conditions for the soils and crops found in the study area. The very sandy soils (Valentine, Loup-Orwet-Elsmere, and Thurman) have the highest potential recharge. Recharge is lowest under alfalfa and highest under bare or fallow ground. The amount of fallow land in the study area is insignificant. About 85 percent of the land area is devoted to row crops and pasture which have a combined mean recharge of about 2.94 in/yr. The remaining 15 percent of the land is chiefly small grains and alfalfa and has a mean recharge of 4.06 in/yr. Recharge to the regional aquifer from deep percolation occurs under an area of about 109,000 acres. This area excludes the blue clay area and areas where shallow water-table conditions result in a net discharge of ground water by evapotranspiration.

Leakage from the blue clay to the regional aquifer is caused by a vertical hydraulic-head gradient toward the regional aquifer. Steady leakage, under nonpumping conditions, is computed by the following form of Darcy's law:

$$Q_l = \sum_{i=1}^{NL} \frac{A_i \, K_{BC}(H_{BCi} - H_{Ri})}{b_{BCi}} \qquad (2)$$

where Q_l = total steady leakage, L^3T^{-1},

i = an index on the subareas used for computation,

NL = number of subareas where leakage occurs,

A_i = area of subarea in which leakage occurs, L^2,

K_{BC} = vertical hydraulic conductivity of the blue clay, LT^{-1},

b_{BCi} = thickness of the blue clay in subarea, i, L,

H_{Ri} = hydraulic head in the regional aquifer in subarea, i, L,

H_{BCi} = hydraulic head in the blue clay in subarea, i, L.

Table 5.--Average annual recharge to the regional aquifer under nonirrigated conditions, 1931-75
[In inches]

Soil	Crop type: Row crops	Alfalfa	Small grain	Pasture	Fallow	Mean	Mean of culti-vated crops	Weighted mean for crops	Weighted mean for crops and pasture
Nora	0.05	0.00	3.35	0.00	13.63	3.41	1.13	0.51	0.31
Loretto-Boelus	2.31	.76	5.92	1.60	11.46	4.41	3.00	2.63	2.22
Thurman	6.59	3.47	8.82	4.71	15.12	7.74	6.29	6.54	5.81
Ortello	5.39	3.31	8.48	4.34	15.08	7.32	5.38	5.63	5.11
Crofton	.01	.00	2.29	.00	16.34	3.73	.77	.33	.20
Loup-Orwet-Elsmere	7.90	4.39	8.82	4.73	16.65	8.50	7.04	7.62	6.46
Lawet	1.06	.11	4.70	.21	16.39	4.49	1.96	1.46	.96
Hord	.34	.00	4.61	.12	11.35	3.28	1.65	.90	.59
Valentine	8.28	4.83	9.15	5.16	16.71	8.83	7.42	8.00	6.86
MEAN	3.55	1.87	6.24	2.32	14.75	5.75	3.89	3.73	3.17

About 106,200 acres is underlain by the blue clay in the study area. A constant value of 8.6 x 10^{-4} ft/d was used for K_{BC} over this entire area. This value is the mean of laboratory measurements by Hoskins-Western-Sonderegger, Inc., of cores taken from test holes 26N-eW-23DDD and 26N-3W-32DCCC. Total steady leakage under nonpumping conditions is about 10,900 acre-ft/yr.

Underflow of ground water through the regional aquifer into the study area is computed from the following form of Darcy's law:

$$Q_u = \sum_{i=1}^{N} K_i b_i W_i \frac{\partial h}{\partial n_i} \qquad (3)$$

where Q_u = underflow along a boundary, L^3T^{-1},

i = an index on segments of the boundary used for computation,

N = number of segments,

K_i = average horizontal hydraulic conductivity over the saturated thickness of the aquifer along segment, i, LT^{-1},

b_i = saturated aquifer thickness along segment, i, L,

W_i = length of segment, i, L,

$\frac{\partial h}{\partial n_i}$ = average hydraulic gradient at right angles to the boundary, dimensionless.

Underflow into the study area occurs principally along the western project boundary in townships 26 and 27 north. Some minor underflow into the study area occurs along the southeastern boundary in T. 24 N., R. 1 W. Areas of underflow into the study area can be identified on figure 13 as parts of these boundaries having a hydraulic-head gradient toward the interior of the study area. Total underflow into the study area through the aquifer is about 11,100 acre-ft/yr as computed with equation 3.

Discharge

Water is discharged from the regional aquifer as (1) ground-water underflow, (2) evapotranspiration in shallow water-table areas, and (3) base flow of perennial streams that are hydraulically connected to the regional aquifer. Underflow of ground water through the regional aquifer out of the study area occurs along the northern and eastern project boundaries. Using equation 3 along these boundaries, underflow out of the study area is about 13,000 acre-ft/yr.

In shallow water-table areas, evapotranspiration (ET) from ground water is appreciable. The evapotranspiration rate is governed by the ability of plants to extract water from the capillary fringe, the capacity of the soil to transmit water to the surface under upward pressure head gradients, and the capacity of the atmosphere to evaporate water from the surface.

The net ET rate of native plants that grow in shallow water-table areas of the Willow Creek watershed is approximately equal to consumptive use under conditions not limited by soil moisture. These plants are assumed to be capable of extracting water at this rate where the water table is less than 10 ft below the land surface. This rate is defined as consumptive-irrigation requirement (CIR) in this report. The method used to compute CIR values is given in the subsequent section on modeling the soil zone. CIR values for pasture and alfalfa on Valentine sand are 0.88 and 0.95 ft/yr, respectively, for the 1931-75 period. Assuming these crops and soil to be predominant in the shallow water-table areas shown in figure 14, the average ET rate from ground water would be 0.9 ft/yr where depth to water is less than 10 ft. Ground-water discharge by ET from the regional aquifer is therefore about 13,050 acre-ft/yr within the Willow Creek ground-water basin part of the area and 13,050 acre-ft/yr in the remainder of the area, a total of 26,100 acre-ft/yr.

Base flow of perennial streams in hydraulic connection with the regional aquifer is maintained by the hydraulic-head gradients toward the streams. The degree of hydraulic connection and the amount of base flow are determined by the thickness and hydraulic conductivity of the materials between the stream and the aquifer. Where hydraulic conductivity is significantly low, base flow is determined by the vertical hydraulic gradient between the stream and the aquifer rather than by lateral gradients.

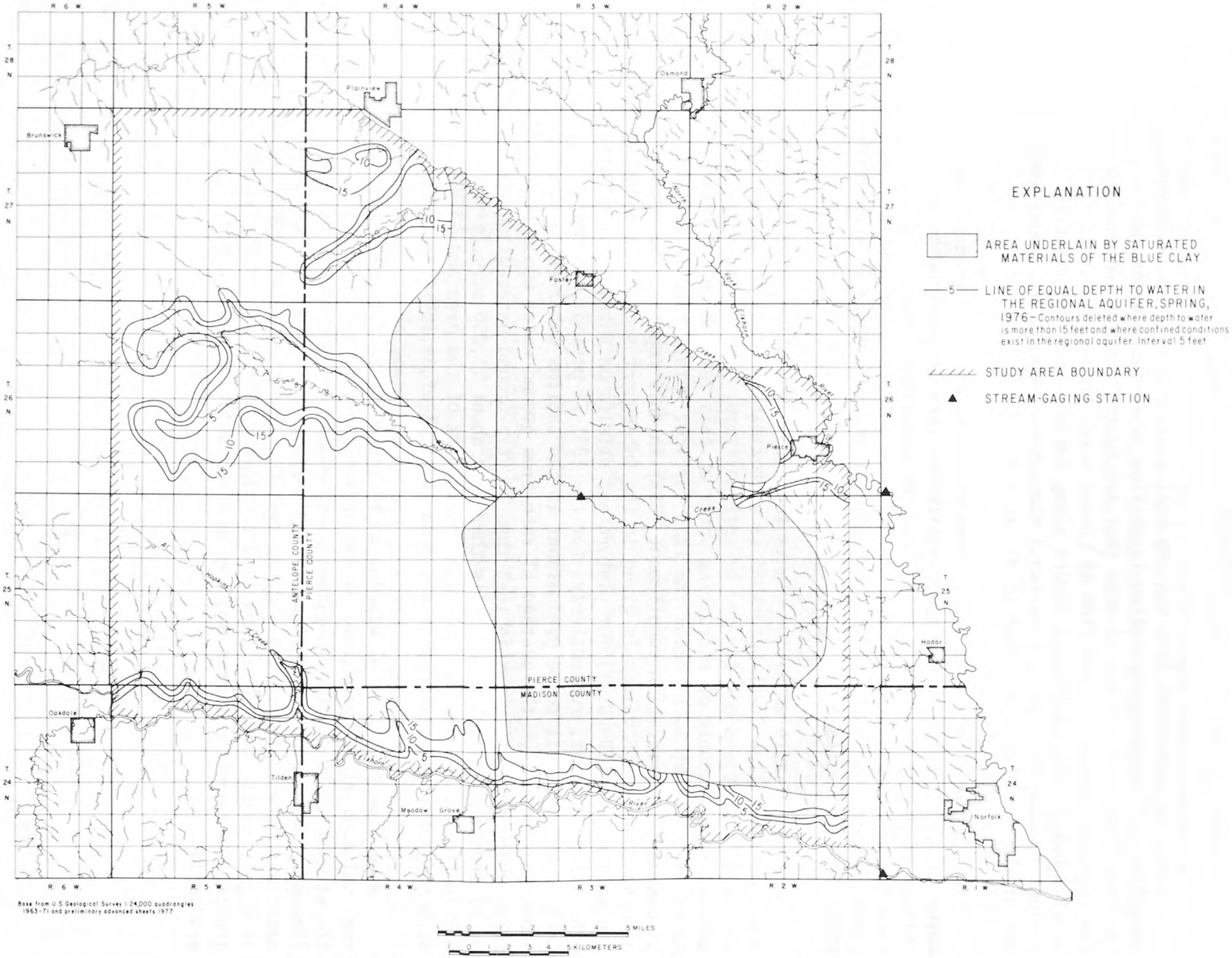

Figure 14.--Depth to water in the regional aquifer.

Base flow can be considered equivalent to total streamflow for Willow Creek above the gaging station near Foster during most of the year. Because 1975-76 was a drier-than-normal period, discharges for this period (table 3) are considered equivalent to base flow. Total base flow above the mouth is estimated to be about 9,600 acre-ft/yr for years similar to 1975 and 1976. Seasonal base-flow fluctuations of Willow Creek from the point of effluence to the mouth are indicated in figure 5. The reduced flow during summer months is due to increased evapotranspiration of ground water, pumpage by wells, and direct pumpage out of the stream.

Ground-water discharge was estimated using steady-state modeling methods which will be discussed in a subsequent section. Total ground-water discharge to Dry Creek, the Elkhorn River, and Al Hopkins Creek was estimated to be about 14,400 acre-ft/yr during 1975-76.

Discharge to perennial streams--Willow Creek, Dry Creek, Elkhorn River, and Al Hopkins Creek--varies with fluctuations in the hydraulic head caused by changes in natural recharge and discharge. Analysis with the ground-water model, using climatic data for the 1931-75 period, indicates that base flows during 1975-76 were about 78 percent of the long-term mean. Therefore, the average rate of ground-water discharge to streams is about 30,800 acre-ft/yr.

Hydrologic balance

Over a given period of time, all recharge to the regional aquifer must be balanced by discharge and(or) changes in storage. Recharge and discharge, except withdrawals by irrigation wells, have been discussed in the previous section. The hydrologic balance under average (1931-75) conditions without pumping wells is shown in table 6 and is assumed to be representative of quasi-steady-state conditions with no change in storage. The recharge given in table 6 was computed as the residual of all other components of the hydrologic balance and is equivalent to 4.16 in/yr in areas excluding those underlain by the blue clay and where the water table is less than 10 ft below land surface. Soils in these areas are sandy, and the recharge rate of 4.16 in/yr compares well with the rates given in table 5.

Table 6.--Average hydrologic balance for the regional aquifer, 1931-75

Inflow, in acre-ft/yr		Outflow, in acre-ft/yr	
Ground-water underflow.....	11,100	Ground-water underflow.....	13,000
Leakage through the blue clay................	10,900	Evapotranspiration of ground water.............	26,100
Recharge from deep percolation	47,900	Base flow of perennial streams..................	30,800
TOTAL	69,900		69,900

The water balance for the regional aquifer under pumping conditions is simulated by the aquifer model. All water pumped from wells is derived from (1) ground-water storage, (2) salvage of evapotranspiration, (3) stream depletion, and(or) (4) leakage through the blue clay. For the 1976 irrigation season, a mass balance determined from the aquifer model was used to determine contributions to pumpage from these sources. The simulated pumpage was 133 acre-ft per well pumping for 90 days from each of the 280 wells present in 1975.

Source of water pumped from wells[1/]	Contribution to total pumpage, in acre-ft/yr	Percent of pumpage
Ground-water storage	28,670	77
Stream depletion	2,610	7
Leakage through the blue clay	5,960	16
Total pumpage	37,240	100

[1/] Evapotranspiration salvage was not included in the analysis because few wells exist in the areas where the water table is shallow enough to permit salvage.

Hydrologic models

In order to meet the objective of this study, water balances are needed in more temporal and spatial detail than is given in the previous sections. This detail is supplied by making analyses with mathematical models of the soil and saturated zones.

Soil zone

This model determines areal and temporal distributions of deep percolation and consumptive-irrigation demands. These are used as recharge and net-withdrawal functions, respectively, in the model of the saturated zone. Recharge and irrigation requirements are determined for combinations of different soil and vegetation types under variable temporal and spatial distributions and potential evapotranspiration.

The soil zone is modeled as a linear reservoir, subject to the following form of the continuity equation:

$$P - RO - DP - ET = \frac{L\Delta\theta}{\Delta\tau} \qquad (4)$$

where P = precipitation, LT^{-1},

RO = surface runoff, LT^{-1},

DP = deep percolation, LT^{-1},

ET = actual evapotranspiration, LT^{-1},

L = thickness of soil zone, L,

$\Delta\theta$ = change in volumetric moisture content of the soil zone, dimensionless,

$\Delta\tau$ = time interval, T (1 month used in this study).

The quantity $P-RO$ or infiltration is considered independent of soil-moisture storage and is computed by using rainfall-runoff relationships developed for experimental watersheds south of Hastings, Nebr. (Lappala, 1977). Although these relationships are empirical, they are assumed to be representative of conditions in the study area.

Operating rules are defined for the linear reservoir to determine releases from ET and DP as functions of soil-moisture storage. These rules require one to assume that the soil moisture below field capacity and above the permanent wilting point is available only to evapotranspiration and that soil moisture above field capacity is available only to deep percolation under the force of gravity. Available moisture values for each soil used are given in table 1. The operating rule used for deep percolation simulated drainage below the root zone of all soil moisture above field capacity during the 1-month time interval used. The operating rule for limiting evapotranspiration was a linear reduction from ET at field capacity to zero when soil moisture fell below a certain arbitrary value. The arbitrary value used for this study was the permanent wilting point.

Evapotranspiration at field capacity $(ETFC)$ was computed using the following:

$$ETFC = (CC)(PET) \tag{5}$$

where CC = a crop coefficient between 0 and 1 which is a function of crop type and growth stage (Lappala, 1977),

PET = potential evapotranspiration which is a function of climatic variables was computed by the Jensen-Haise method (Jensen and others, 1969):

$$PET = RsC(T-T_p)$$

where Rs = total monthly solar radiation, expressed as evaporation equivalent,

C = a correction factor to standard temperature and pressure based upon the adiabatic lapse rate and the saturation vapor pressure during the warmest month (units vary based upon those used for pressure and temperature),

T = mean monthly air temperature (°F or °C),

T_p = a temperature correction factor, °F or °C.

Solar radiation (Rs) was computed for this study using the empirical relationship with percent possible sunshine developed by Fritz and MacDonald (1949) and Rosenberg (1964):

$$Rs = C_2 d R_o (0.35+0.615*S) \qquad (6)$$

where C_2 = a unit conversion from langleys (gm-cal/cm^2) to evaporation equivalent,

R_o = daily solar radiation on cloudless days, langleys,

d = number of days per month,

S = mean monthly percent possible sunshine, expressed as a fraction.

Values of R_o were taken from published maps by Fritz (1949) for Norfolk. Mean daily values of percent possible sunshine for each month at Sioux City, Iowa (National Oceanic and Atmospheric Administration, 1931-75) were used as representative of the study area.

The moisture-balance model represented by equation 4 also was used to compute consumptive-irrigation requirements. The computed *CIR* was the amount necessary to maintain the soil-moisture level at one-half of the available moisture capacity. This level is an arbitrary value, but considered reasonable to minimize crop stress (D. Watts, oral commun., 1975). These values are given in table 7. Net withdrawals from the saturated zone were the difference between the 1931-75 average recharge rate under nonirrigated conditions (table 5) and the computed consumptive-irrigation requirement (table 7). This is equivalent to the water required to convert from dryland to irrigated conditions. The separation is necessary to evaluate changes in the saturated zone from the initial (1975) condition which was assumed to be maintained by the recharge under nonirrigated conditions. The consumptive-irrigation requirement does not include excess water that may be required for leaching of excess salts or that may be necessary under gravity systems to adequately irrigate an entire field.

Long-term average dryland recharge to the saturated zone is considered equal to deep percolation. Average deep percolation values under nonirrigated conditions for the period 1931-75 are shown by soil and crop type in table 5. Computed values of recharge for the three weather stations were averaged to prepare the table. Actual values were used in applying these to the saturated-zone model.

Table 7.--Computed mean consumptive-irrigation requirements, 1931-75
In inches

Soil	Crop type			Mean	Weighted mean of crops[1/]
	Row crop	Alfalfa	Pasture		
Nora	6.51	11.67	9.48	9.22	7.03
Loretto-Boelus	6.50	11.34	9.37	9.07	6.99
Thurman	7.54	12.34	10.47	10.12	8.02
Ortello	7.99	12.23	10.35	10.19	8.41
Crofton	6.31	13.84	10.42	10.19	7.06
Loup-Orwet-Elsmere	6.66	11.30	10.47	9.48	7.12
Lawet	4.29	9.36	8.31	7.32	4.80
Hord	4.78	10.51	8.23	7.84	5.36
Valentine	6.66	11.41	10.58	9.55	7.14
MEAN	6.36	11.56	9.74	9.22	6.88

[1/] Based on 90 percent row crops and 10 percent alfalfa.

Saturated zone

The saturated zone consists of the saturated parts of both the regional aquifer and the blue clay. The water balance for the regional aquifer is expressed by the following equation which combines the continuity equation with Darcy's law:

$$K\frac{\partial}{\partial X_i}\left[(t-b)\frac{\partial h}{\partial X_i}\right]=\left[S_s(t-b)+S_y\ \delta_{sy}(X_i-\overline{X})\right]\frac{\partial h}{\partial \tau} \qquad (7)$$

$$+\ \frac{K_{BC}}{b_{BC}}(H_{BC}-h)\ \delta_l(X_i-\overline{X})+\frac{K_s}{b_s}(H_s-h)\ \delta_s(X_i-\overline{X})$$

$$+\ Q_r\ \delta_r(X_i-\overline{X})+Q_w\ \delta_w(X_i-\overline{X})$$

$$i = 1,\ 2$$

subject to the initial condition (with no wells)

$$K\frac{\partial}{\partial X_i}(t-b)\frac{\partial h_o}{\partial X_i} = \qquad (7a)$$

$$\frac{K_{BC}}{b_{BC}}(H_{BC}-h_o)\ \delta_l\ (X_i-\overline{X})$$

$$+\ \frac{K_s}{b_s}(H_s-h_o)\ \delta_s\ (X_i-\overline{X})$$

$$+\ Q_{r_o}\ \delta_r\ (X_i-\overline{X})$$

and the boundary conditions

$$h = h_c \text{ on specified head boundaries} \qquad (7b)$$

$$\frac{\partial h}{\partial n} = \frac{\partial h_o}{\partial n} \text{ on arbitrary boundaries not occupied by streams} \qquad (7c)$$

where h = hydraulic head in the regional aquifer, L

K = hydraulic conductivity of the regional aquifer, LT^{-1} (considered homogeneous and isotropic)

t = top of the regional aquifer, L (water table in west part of study area and base of the blue clay in the east)

b = base of the regional aquifer, L

X_i = coordinate directions, (in this study X_1 = north-south and X_2 = east-west)

S_s = specific storage of the regional aquifer, L^{-1}

S_y = specific yield of the regional aquifer, dimensionless

τ = time, T

$\delta_{sy}(X_i - \overline{X})$ = Dirac delta function for specific yield (when integrated, this function has the value of 1 when $X_i = \overline{X}$ and 0 elsewhere. This mathematically allows the western part of the aquifer to be modeled as water table and the eastern part as artesian)

K_{BC} and b_{BC} = hydraulic conductivity (LT^{-1}) and thickness (L) of the blue clay

H_{BC} = hydraulic head in the blue clay, (considered a constant equal to the initial value for all simulations)

$\delta_l(X_i - \overline{X})$ = Dirac delta function for leakage through the blue clay

K_s, b_s, and H_s = hydraulic conductivity (LT^{-1}), thickness of stream bed (L), and hydraulic head (L) in streams connected to the aquifer

$\delta_s(X_i - \overline{X})$ = Dirac delta function for leaky streams

Q_r = recharge from deep percolation, LT^{-1}

$\delta_r(X_i - \overline{X})$ = Dirac delta function for recharge (no direct recharge is simulated under the blue clay; recharge under the blue clay is equivalent to leakage)

Q_w = withdrawal from wells, computed as using the consumptive-irrigation requirement applied to 133 acres/well

$\delta_w(X_i - \overline{X})$ = Dirac delta function for wells.

The water balance for the blue clay was not modeled explicitly in this study. It is incorporated in the model of the regional aquifer through the leakage terms in equation 7.

The above equations were solved by a published computer program (Trescott and others, 1976) modified for this study to (1) allow for stream depletion accounting at individual points along stream networks and (2) automatic computation of the flux required to maintain boundary condition 7c.

Use of this method requires subdividing the study area into a number of rectangular cells. (See fig. 15.) The flow equation is solved by writing an implicit finite-difference analog for each of these cells. The resulting set of algebraic equations is solved for values of hydraulic head (h) in the regional aquifer at specified points in time using the Strongly Implicit Method (Trescott and others, 1976). These head distributions are used to compute discharge from the aquifer to hydraulically connected streams, thus determining temporal variations in perennial flow of these streams.

Input data required for the model of the saturated zone are average or representative values of all the parameters in equation 7. Methods of determining these parameters have been discussed previously. For transient simulation, values also are needed for the initial hydraulic-head distribution and the recharge required to generate that distribution.

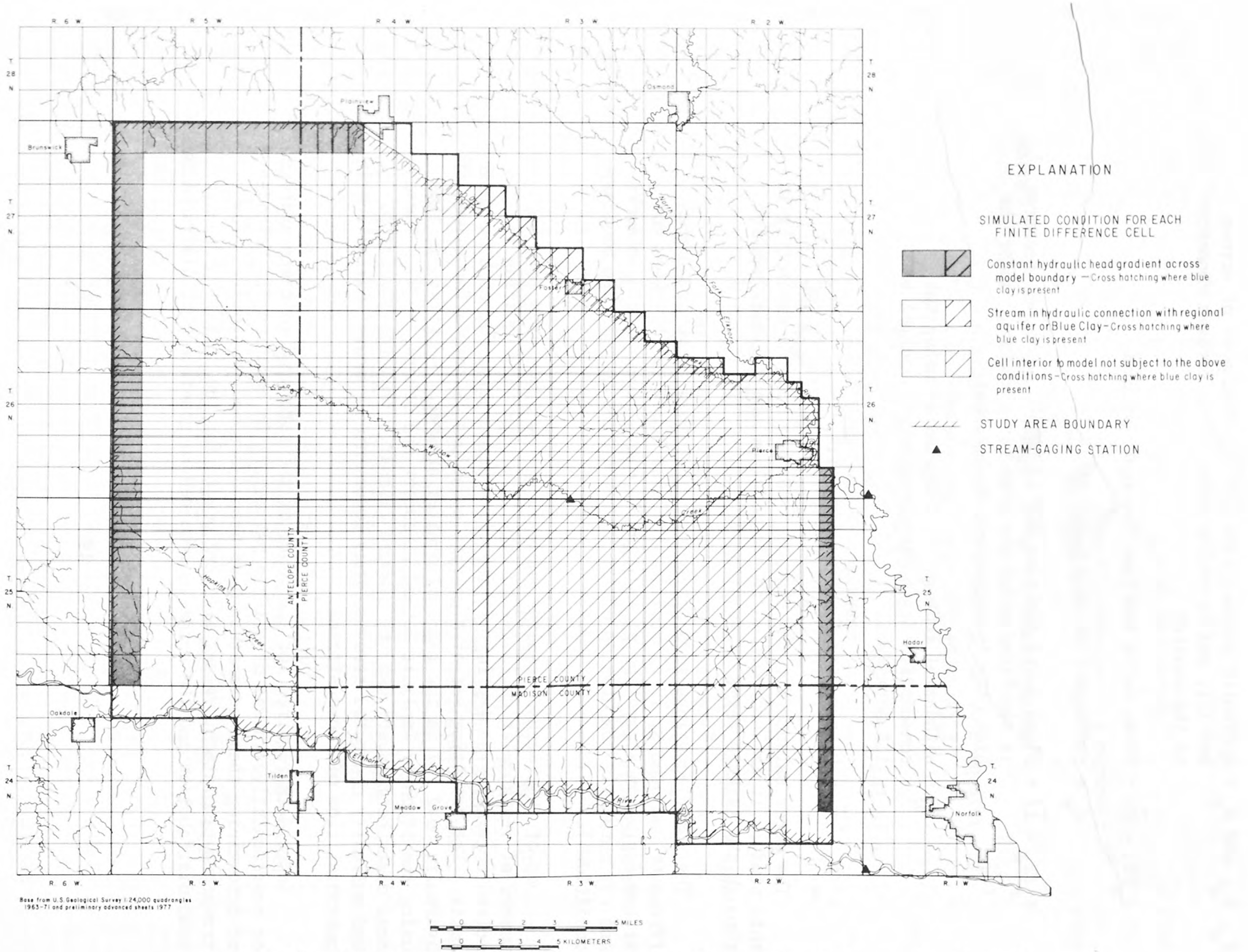

Figure 15.--Finite-difference grid used for the model of the regional aquifer.

For this study, the initial head distribution was generated as follows:

In the western part of the study area where the regional aquifer is unconfined, the initial heads were forced to the measured heads in 1975 by computing the recharge and discharge Q_{r_o} required to maintain that head distribution. This was accomplished by a direct finite-difference solution to

$$K \frac{\partial}{\partial X_i}\left[(h-b)\frac{\partial h}{\partial X_i}\right] = Q_{r_o} \qquad i=1,2 \qquad (8)$$

where the variables are as previously defined. For a further discussion of the use of this equation, see Lappala (1977).

In the eastern part of the study area where leakage through the blue clay provides recharge to the aquifer, the initial head was computed using the aquifer model to solve

$$K \frac{\partial}{\partial X_i}\left[(t-b)\frac{\partial h}{\partial X_i}\right] = \frac{K_{BC}}{b_{BC}}(H_{BC}-h)\ \delta_l(X_i-\bar{X}) \qquad i=1,2 \qquad (9)$$

Initial base flows of perennial streams were derived by solving equation 8 for discharge from the uppermost hydrologic unit everywhere in the study area. The parameters and heads used were those for the regional aquifer in the west and the blue clay in the east. Values of Q_{r_o} were summed in an upstream direction to obtain the base flow at each point along the stream networks.

Testing the model of the saturated zone

The conceptual model and the mathematical quantification of the hydrologic system are subject to many possible errors resulting from data interpretations and required assumptions. Consequently, the model must be tested to determine if it can adequately reproduce existing measures of the hydrologic response of the real system. Measures of the real system are water levels and water-level changes and base flows and base-flow changes.

The model was tested for adequacy in reproducing the initial (1975) hydraulic-head distribution, for aquifer discharge resulting from that distribution, for adequacy in reproducing changes in hydraulic head, and for aquifer discharge during the 1976 irrigation season. Due to the forced fit of the initial water levels in the western part of the study area, the computed 1975 head distribution gave information only on model adequacy under the blue clay. This test was used to determine if the constant value of 25 ft/d for hydraulic conductivity (K) of the regional aquifer would give a reasonable match to the 1975 head distribution.

As a result of this test, the value for K was increased to 31 ft/d for the entire aquifer. This value resulted in computed steady-state heads that were everywhere within 10 ft of the observed 1975 head. Leakage parameters (K_{BC} and b_{BC}) were not adjusted during this test. Since little contribution to base flow of Willow Creek occurs where it crosses the blue clay, base flows computed from simulated steady-state heads are essentially the same as those computed using values of Q_{r_o} from equation 8 for the hydrologic unit in contact with Willow Creek. Base-flow values shown in figure 5 for spring 1976 are computed with the simulated steady-state heads.

Simulation of transient head changes required values for the storage properties of the aquifer S_s and S_y. Both of these values were tested by simulating head changes over the 1976 irrigation season (June-August). The simulation required an estimate of the net withdrawal by irrigation wells during 1976. The measured water-level decline from spring to fall 1976 is shown in figure 16. Most measurements in the fall of 1976 were made in a 2-week period around the end of September. Climatic data for the test period were not available to use in the soil model; consequently, a uniform value of 1 ft was arbitrarily used for net withdrawal during the test period. Each well present during the test period was assumed to supply 1 ft of water over 133 acres. The model was tested by simulating the June-August pumping period followed by a 1-month recovery period.

Tests using $1.0\text{x}10^{-6}$ ft^{-1} as the initial value of S_s resulted in a simulated water-level decline under the blue clay at the end of the 1-month recovery period. Increasing S_s to $1.0\text{x}10^{-5}$ ft^{-1} resulted in an adequate fit to the measured water-level change (fig. 17). The value of specific yield (0.20) used for the unconfined parts of the regional aquifer was not changed during testing. Simulated changes in aquifer discharge to Willow Creek during the 1976 test period were reasonably close to those observed during both fall 1975 and fall 1976 (fig. 5).

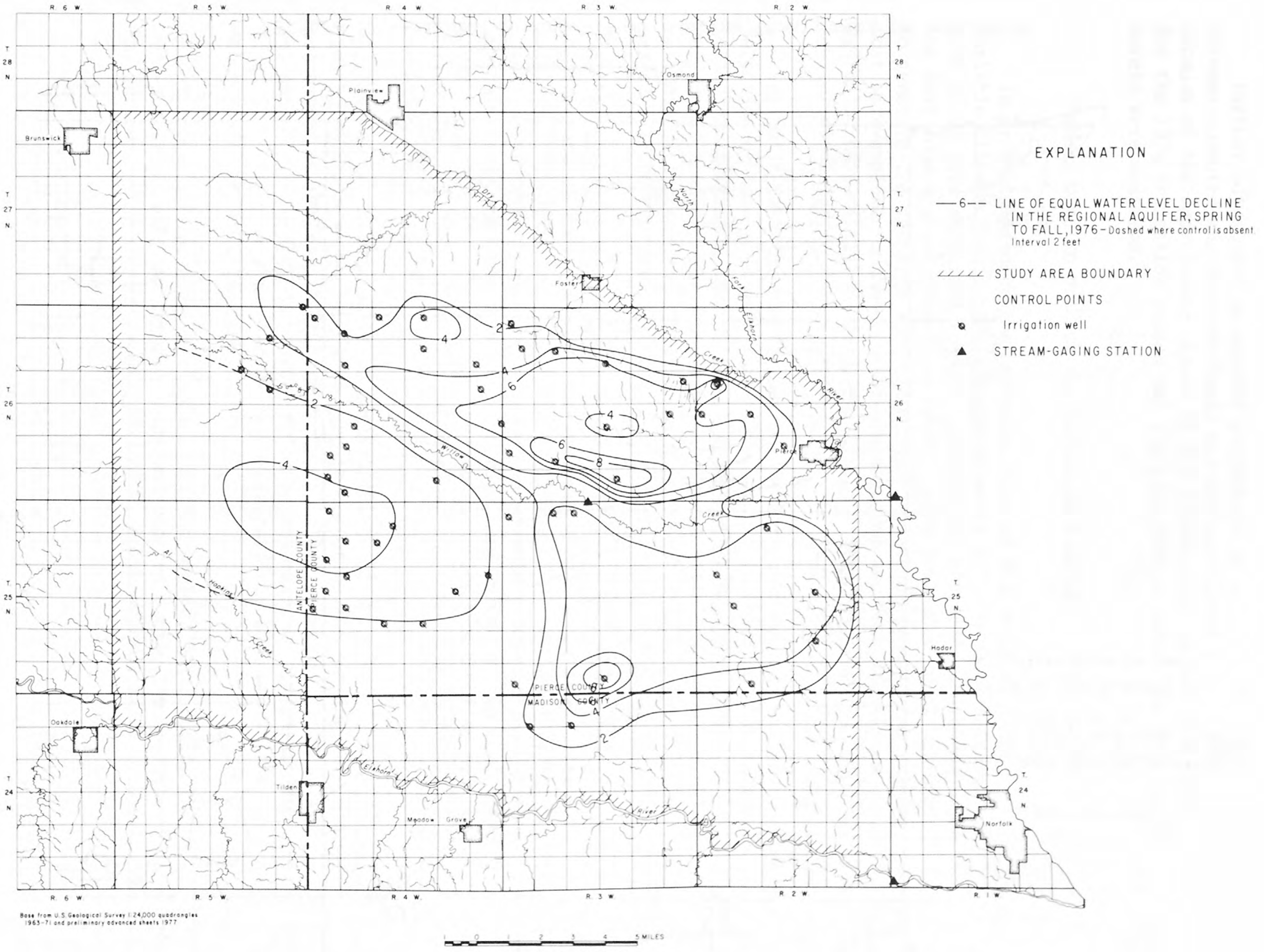

Figure 16.--Measured water-level decline from spring to fall 1976.

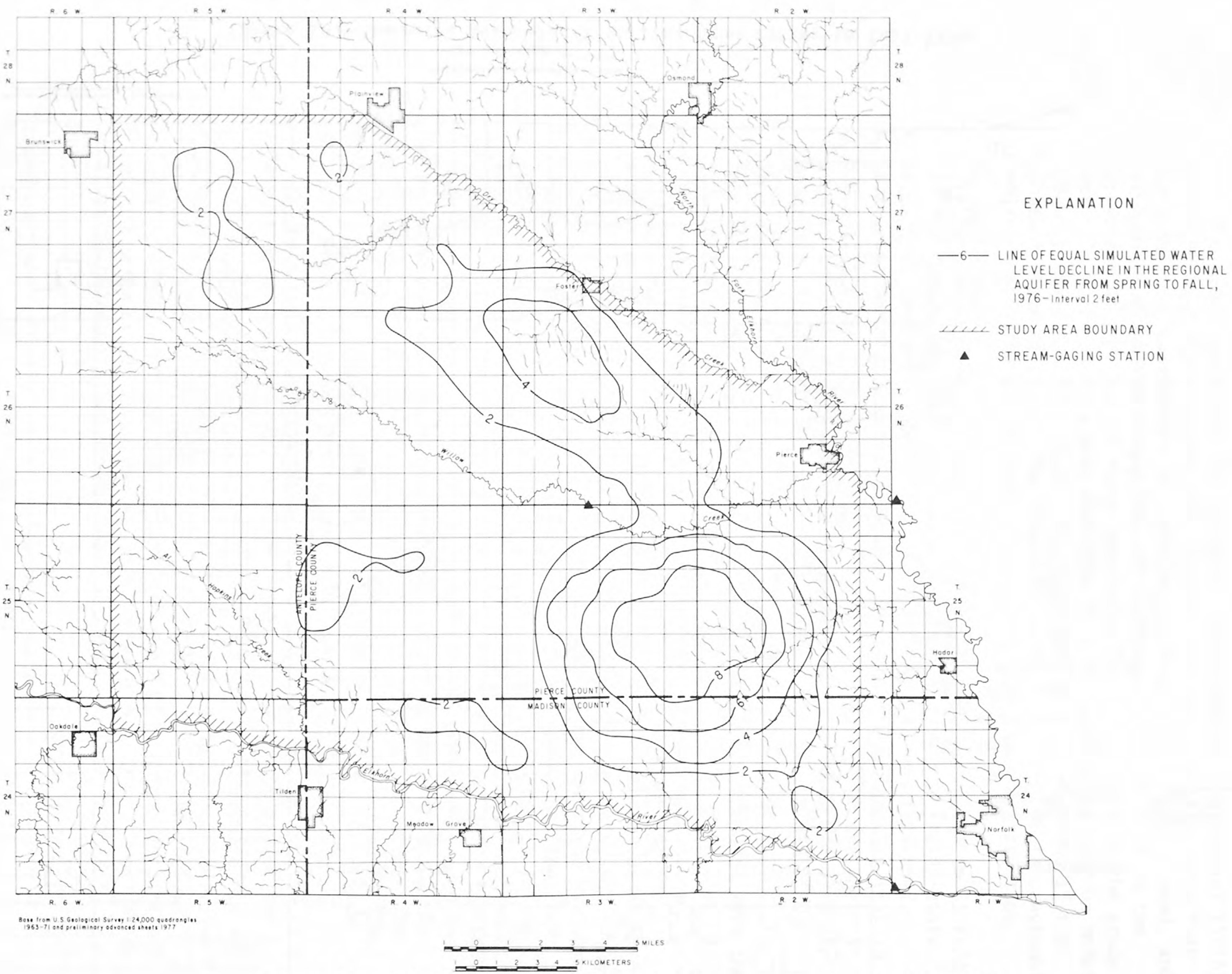

Figure 17.--Simulated water-level decline from spring to fall 1976.

Further adjustment in aquifer parameters to improve the match between computed and observed heads was not considered justifiable because of the approximate nature of the stress used to test the model for the 1976 irrigation season and the time span in which the fall water levels were measured.

Models used to predict the perennial flow of Willow Creek

In order to predict the perennial flow of Willow Creek under variable climatic conditions and ground-water withdrawals for irrigation, base flows were simulated for six combinations of these conditions using the soil-zone and saturated-zone models. The soil-zone model was used to generate the recharge and consumptive-irrigation requirement values that were applied to the model of the regional aquifer. Simulated hydraulic-head distributions using these stresses were used to compute aquifer discharge to Willow Creek.

Climatic data for the period 1931-54 were used for all analyses. Selection of this period was arbitrary, and it includes one major dry period and one major wet period. The assumption for using this particular period was that climatic conditions in recent years have been equivalent to those of the early thirties and that conditions during the 1977-90 period will be equivalent to the 1931-54 period.

The number of future irrigation wells, their prospective locations, and the rate of their installation were required for the predictive analyses. By agreement with the Lower Elkhorn NRD, new wells were simulated as being randomly emplaced on irrigable tracts of land at the approximate initial rates of 30, 40, and 80 new wells per year (alternatives 1, 2, and 3). The rate of well installation was about 40 per year during the 1972-76 period. Figure 18 shows the cumulative number of simulated wells for each alternative.

The number of new wells in a given finite-difference cell was limited so that the maximum area used in computing the pumping rate did not exceed 90 percent of the area of each cell. Otherwise, each well was assumed to supply 133 acres. Wells were limited to irrigable land (fig. 3). The land simulated as being irrigated by 1990 under this scheme is shown in figure 19.

Pumpage and recharge were applied to the aquifer model during the June-August period, and recharge only was applied during the September-May period.

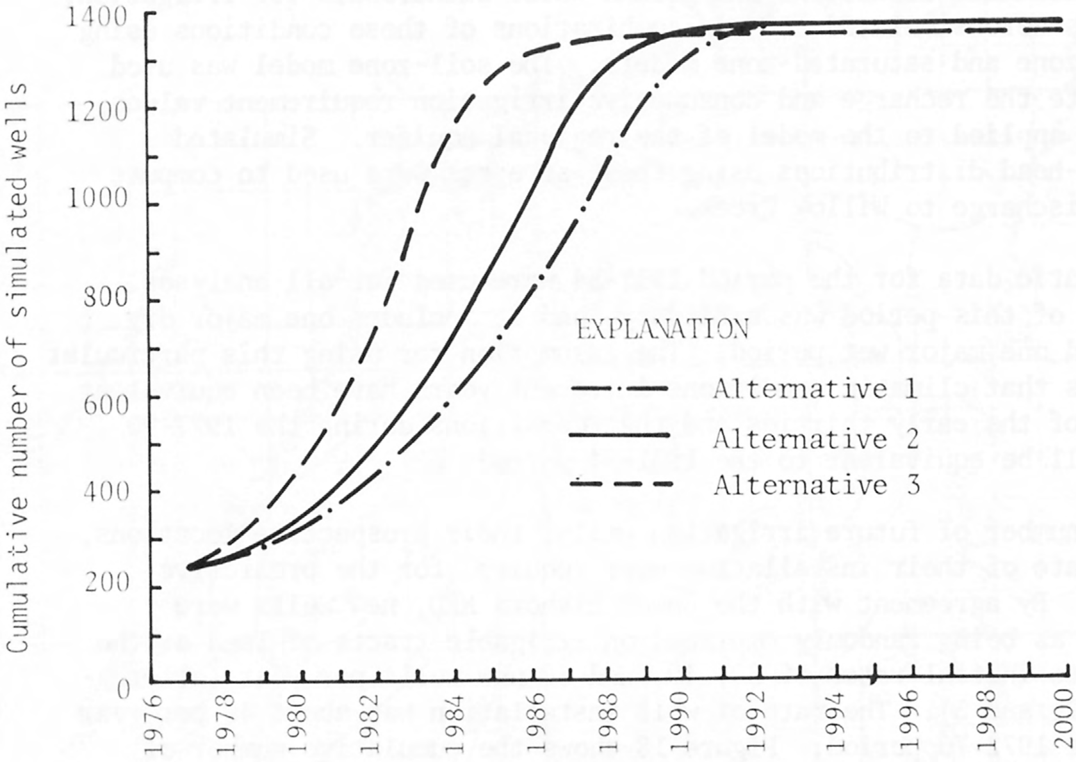

Figure 18.--Cumulative number of simulated wells for alternatives 1,2, and 3.

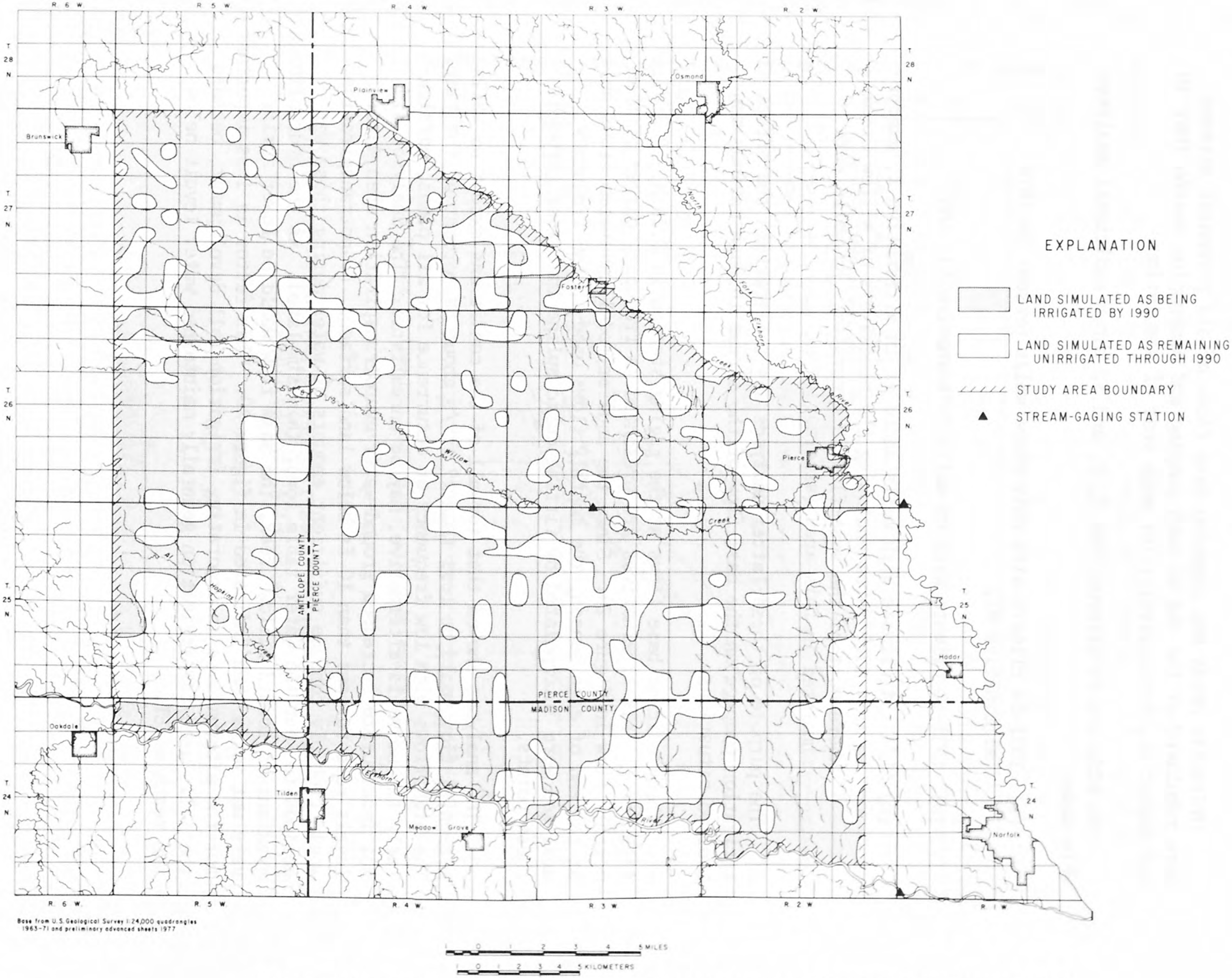

Figure 19.--Land simulated as being irrigated with ground water by 1990.

Hydraulic heads and computed base flows of all perennial streams were tabulated at the end of each pumping and nonpumping season (May 30 and August 31, respectively) for each year of simulation.

In addition to alternatives 1, 2, and 3, three additional analyses were made:

(a) 1931-54 climate with only those wells present in 1976 (alternative 4),

(b) 1931-54 climate with no wells (alternative 5), and

(c) 1931-54 climate with wells added as in alternative 2, except in a 34-mile^2 area enclosing the headwaters of Willow Creek where most of the base flow is generated. No wells were simulated in this area (alternative 6).

The purpose of these latter analyses was to ascertain the effect of irrigation pumpage on the base flow of the stream through various controlled situations.

The predicted base flows at the gaging station in T.25 N., R.3 W., 4ABB for the six alternatives are presented as frequency curves of mean monthly flow in figure 20. Since the aquifer model tabulated flows only at the end of each season, the monthly flows used in the frequency analyses were interpolated by fitting cubic spline functions through these points.

The analyses showed that for all of the conditions except no wells perennial flow would be less than 12 ft^3/s about 50 percent of the time. The high flows with low frequency of occurrence for all plans, particularly the no wells alternative, may be unrealistic. The analyses did not incorporate increased ground-water evapotranspiration when water levels rose to less than 10 ft below land surface. Evapotranspiration would probably consume sufficient amounts of water during these wet periods to keep the water table lower than simulated. As the hydraulic gradient is toward Willow Creek, the base flow would be lower during very wet years than indicated in figure 20. This factor is less important for the alternatives incorporating irrigation-well development, because the wells used for irrigation generally maintained water levels below the reach of ET.

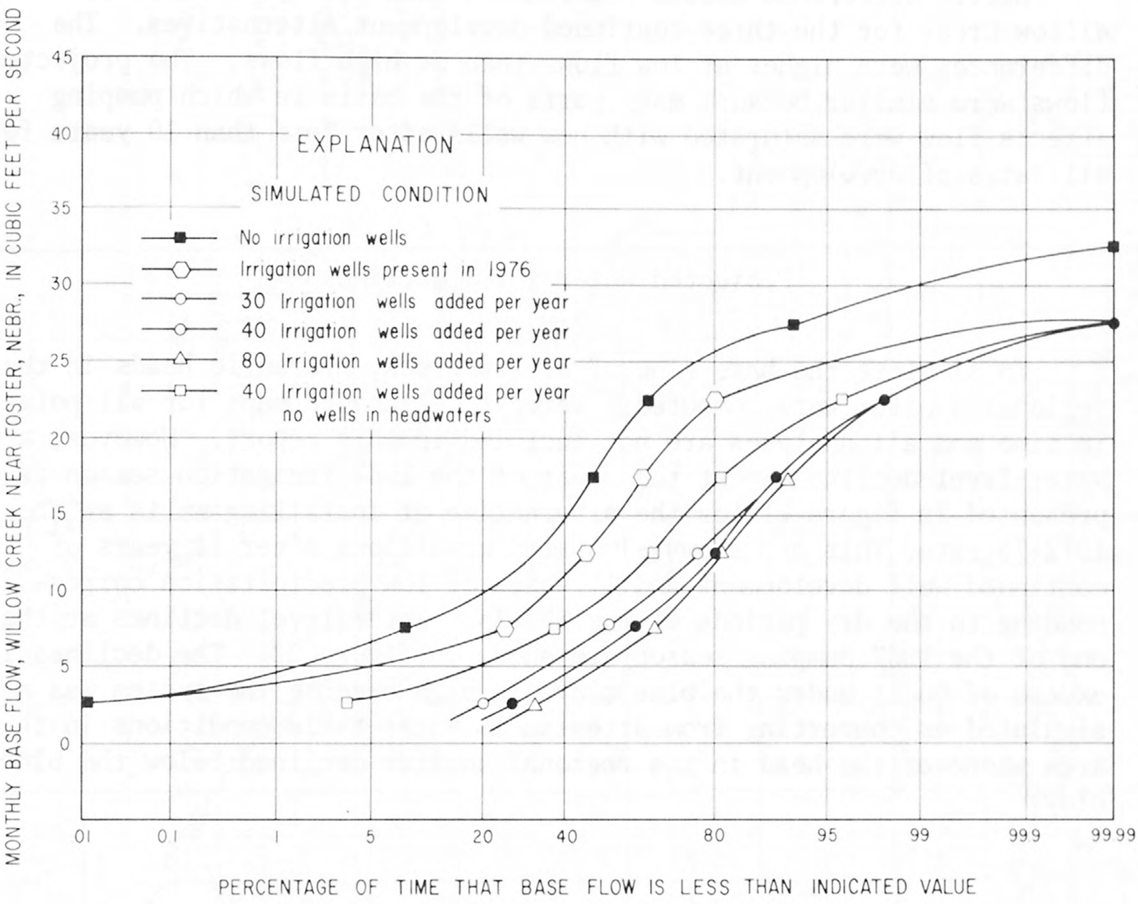

Figure 20.--Simulated perennial flow of Willow Creek under six alternative futures.

The analysis made without wells in the upper part of the basin showed less flow than did the analysis using only the existing wells. Future wells around the area in which no pumping was simulated provided sufficient declines to account for the lower flows.

Little difference exists between the simulated perennial flows of Willow Creek for the three continued development alternatives. The differences were higher at low flows than at high flows. The projected flows were similar because many parts of the basin in which pumping affects flow were saturated with new wells after less than 10 years for all rates of development.

Projected water-level declines

To simulate the base flow of Willow Creek, hydraulic heads in the regional aquifer were computed. Water-level-change maps for all points in time and all analyses are not included in this report. However, a water-level-decline map at the start of the 1987 irrigation season is presented in figure 21 for the alternative of installing wells at the 1972-76 rate. This map shows simulated conditions after 11 years of continued well development and 11 years of low precipitation corresponding to the dry periods of the 1930's. Water-level declines at the end of the 1987 pumping season are shown in figure 22. The declines in excess of 60 ft under the blue clay are high because the system was not simulated as converting from artesian to water-table conditions in this area whenever the head in the regional aquifer declined below the blue clay.

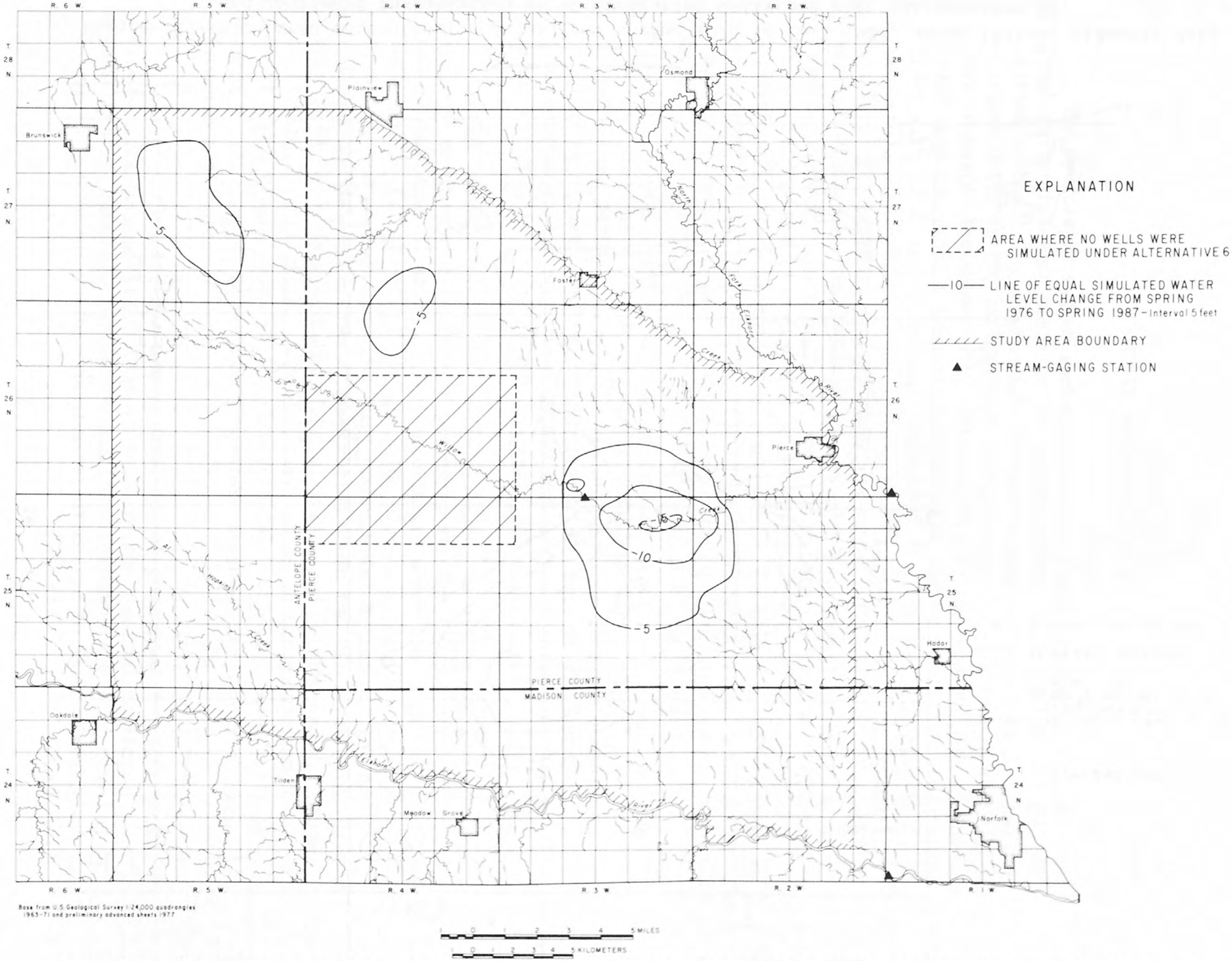

Figure 21.--Simulated water-level changes from spring 1976 to spring 1987, using 1931-41 climatic data and continued installation of 40 irrigation wells per year (alternative 2).

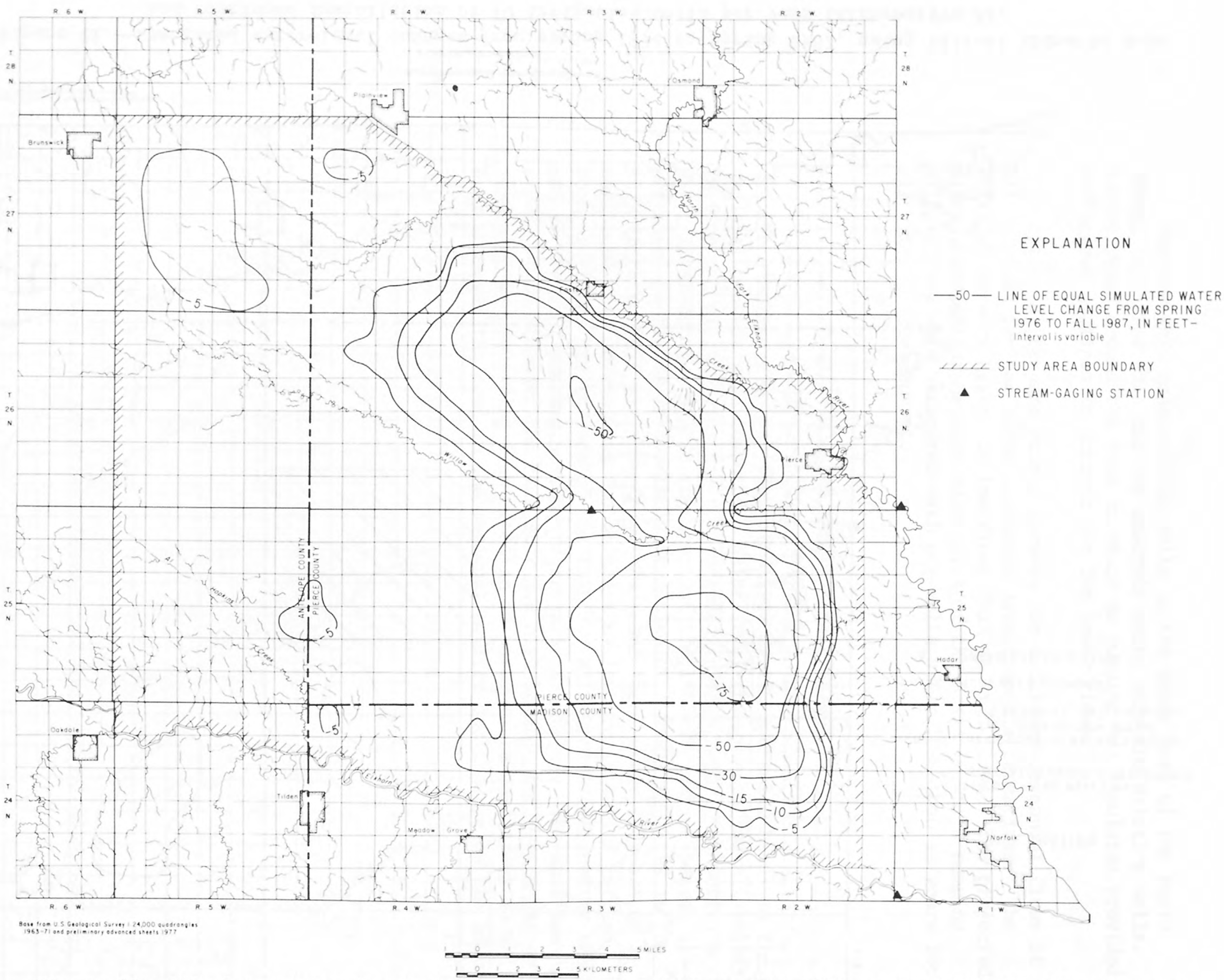

Figure 22.--Simulated water-level changes from spring 1976 to fall 1987, using 1931-41 climatic data and continued installation of 40 irrigation wells per year (alternative 2).

Summary and conclusions

The hydrologic system including most of the watershed of Willow Creek in northeastern Nebraska was quantitatively described by using digital models. The ground-water system is complex, consisting of two distinct units in the eastern half of the study area and one in the western half. The upper unit, or blue clay, in the eastern part of the study area comprises a basal lacustrine clay overlain by silts and silty clays with local very fine sand interbeds. Much of the blue clay is saturated, with hydraulic heads above those of the underlying regional aquifer. The regional aquifer comprises sands and gravels of Pliocene through Pleistocene age. Water occurs under artesian or confined conditions in the eastern part of the basin where the regional aquifer is overlain by the blue clay and occurs under unconfined or water-table conditions in the western part of the basin. The most permeable sands and gravels of the regional aquifer range in thickness from 50 to 100 ft and average about 75 ft. Hydraulic conductivity determined from test-hole sample descriptions and refined by model testing is about 31 ft/d. Specific storage determined by model testing is about 1.0×10^{-5} ft^{-1} and specific yield is 0.20. Under average conditions assumed to be represented by the 1975-76 period, the regional aquifer receives recharge from deep percolation of precipitation (47,900 acre-ft/yr) and from leakage through the blue clay (10,900 acre-ft/yr). Lateral inflow is about 11,100 acre-ft/yr and outflow is 13,000 acre-ft/yr.

Under average conditions (1975-76), water is discharged from the regional aquifer as base flow of perennial streams (30,800 acre-ft/yr) and by evapotranspiration (26,100 acre-ft/yr) where the water table is less than 10 ft below land surface. The blue clay receives recharge from deep percolation and discharges ground water by leakage to the regional aquifer and by evapotranspiration where the water table is less than 10 ft below land surface. Lateral inflow and outflow and discharge to streams is negligible. The blue clay ranges in thickness from 5 to 80 ft and averages about 40 ft. Its vertical hydraulic conductivity is about 9.0×10^{-4} ft/d, based upon laboratory measurements of cores taken from two test holes.

The blue clay is important in analyses of ground-water discharge to Willow Creek because it effectively isolates the regional aquifer. Consequently, pumping from the regional aquifer under the blue clay has little direct effect on the ground-water discharge to Willow Creek.

The hydrologic system described above was quantified by using a digital model of the regional aquifer. Stresses applied to the aquifer from deep percolation and net withdrawals for irrigation were computed using a linear reservoir model of the soil zone. Recharge and withdrawal values were thus generated that were functions of monthly climatic variations and different soils and crops. The model of the regional aquifer was tested for its ability to reproduce 1975 water levels and aquifer discharge. Water levels and base flow in 1975 were assumed to represent quasi-steady-state conditions. The only adjustment made to achieve a reasonable match with the 1975 regional heads was to increase the value of hydraulic conductivity from 25 ft/d to 31 ft/d for the entire regional aquifer.

The model was tested under transient conditions by simulating the water-level changes measured over the 1976 irrigation season. An arbitrary value of 133 acre-ft/well was applied during June, July, and August. To achieve a reasonable match between observed and simulated water-level changes, the specific storage was increased to 1.0×10^{-5} ft^{-1} from its initial value of 1.0×10^{-6} ft^{-1}. Base flow at the end of the pumping season agreed very well with seepage measurements made in the fall of 1976.

The tested model was used to determine the perennial flow of Willow Creek under (1) three levels of continued irrigation-well installation, (2) no further well development, (3) no wells, and (4) no wells in a 34-square-mile area in the upper part of the basin. The analyses showed that the base flow of Willow Creek probably would be less than 12 ft^3/s at least 50 percent of the time under all but completely undeveloped conditions. Water-level declines with development continued at the present rate, corresponding to the end of a long drought period similar to the 1930's, showed declines of more than 80 ft under the blue clay during the irrigation season and less than 10 ft by the start of the following irrigation season. Declines in the western part of the basin where water-table conditions exist were less than 7 ft at the end of the irrigation season mentioned above and less than 5 ft by the start of the following irrigation season.

The model in this report describes the hydrologic system as well as possible with the existing data. The following additional data are necessary to improve the reliability of the model:

1. Storage properties of the regional aquifer

2. Vertical permeability of the blue clay

3. Storage properties of the blue clay

4. Detailed documentation of pumpage, return flow, and discharge by ET.

Further analyses might include augmenting streamflow either by pumping from beneath the blue clay or by salvaging evapotranspiration in shallow water-table areas in the western part of the basin.

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