APPRAISAL OF THE WATER RESOURCES OF THE BIG SIOUX AQUIFER, CODINGTON AND GRANT COUNTIES, SOUTH DAKOTA

By LARRY D. PUTNAM and RYAN C. THOMPSON

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations Report 96-4275

Prepared in cooperation with the

SOUTH DAKOTA GEOLOGICAL SURVEY,

EAST DAKOTA WATER DEVELOPMENT DISTRICT, and

CODINGTON AND GRANT COUNTIES

Rapid City, South Dakota
1996

U.S. DEPARTMENT OF THE INTERIOR
BRUCE BABBITT, Secretary

U.S. GEOLOGICAL SURVEY
Gordon P. Eaton, Director

For additional information write to:

District Chief
U.S. Geological Survey
1608 Mt. View Rd.
Rapid City, SD 57702

Copies of this report can be purchased from:

U.S. Geological Survey
Branch of Information Services
Box 25286
Denver, CO 80225-0046

CONTENTS

ILLUSTRATIONS

TABLES

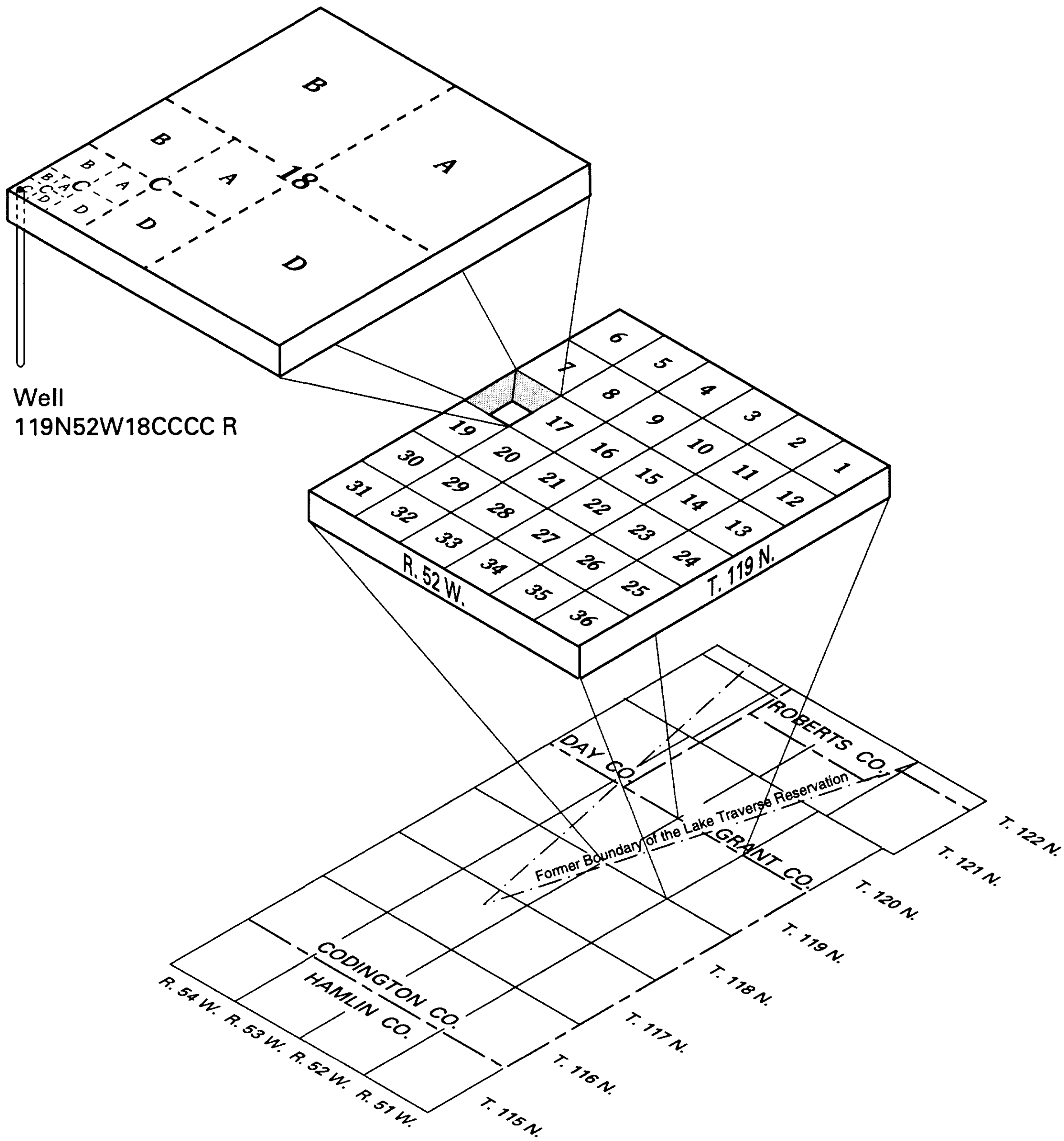
CONVERSION FACTORS AND VERTICAL DATUM

Multiply inch-pound unit	By	To obtain metric unit
acre	4,047	square meter
acre	0.4047	hectares
acre-foot (acre-ft)	1,233	cubic meter
cubic foot per second (ft^2/s)	0.02832	cubic meter per second
foot (ft)	0.3048	meter
foot per acre (ft/acre)	0.7531	meter per hectare
foot per day (ft/d)	0.3048	meter per day
foot per mile (ft/mi)	0.1894	meter per kilometer
mile (mi)	1.609	kilometer
inch	25.4	millimeter
inch per year (in/yr)	25.4	millimeter per year
square foot per day (ft^2/d)	0.09290	square meter per day
square mile (mi^2)	2.590	square kilometer

Sea level: In this report "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD of 1929)--a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, formerly called Sea Level Datum of 1929.

WELL-NUMBERING SYSTEM

Observation wells and test holes are numbered according to the Federal land survey system. This system uses township, range, and section number. Number 119N52W18CCCC indicates a well in T. 119 N., R. 52 W., section 18. The last four-letters show location within the section as shown below. When a nest of wells is drilled, sequential numbers identify each well in the nest and appear after the four-letter code. Wells located within the former boundary of the Lake Traverse Indian Reservation are given a suffix of "R."



Appraisal of the Water Resources of the Big Sioux Aquifer, Codington and Grant Counties, South Dakota

By Larry D. Putnam *and* Ryan C. Thompson

ABSTRACT

The Big Sioux aquifer in Codington and Grant Counties is a 150-square-mile, predominantly unconfined aquifer that is hydraulically connected to the Big Sioux River, Lake Kampeska, Pelican Lake, and Still Lake. The average thickness of the Big Sioux aquifer is 24 feet, with a maximum thickness of 54 feet, and is underlain by glacial till.

A digital computer model, with 172 rows and 60 columns of cells 1,320 feet on a side, was calibrated using historical water levels and stresses for 1978-85. The hydraulic conductivity of the aquifer ranged from 50 to 500 feet per day, but for most cells was 350 feet per day. A uniform specific yield of 0.14 was used. Riverbed hydraulic conductivity ranged from 0.05 to 1.0 square foot per day. The steady-state recharge rate was 5.53 inches per year. The steady-state evapotranspiration rate was 34.71 inches per year when the water level was at land surface. Evapotranspiration was decreased linearly to zero as the depth of the water table below land surface approached the extinction depth, generally 5 feet.

The average absolute difference between simulated and observed water levels at 17 wells during the steady-state simulation was 1.20 feet. During the transient simulation, the model was re-calibrated on a monthly basis using hydrologic data for 1984 and 1985. The average absolute difference between simulated and measured water levels was 1.72 feet for the transient simulation.

A 2-year hypothetical simulation was completed using 1993 municipal and rural-water-system pumpage, hypothetical maximum permitted irrigation pumpage, and dry conditions (1980-81). During this simulation, 22 of the 37 cells containing irrigation wells went dry. The average drawdown in the aquifer at the end of the second hypothetical irrigation season was approximately 1.5 feet, with a maximum drawdown in the aquifer of 16.2 feet. Another condition was simulated in which the irrigation pumpage was one-half of the permitted pumpage. For this condition, 15 of the 37 cells containing irrigation wells went dry. These simulations indicate that the Big Sioux aquifer probably is unable to support extensive irrigation during dry periods such as those that occurred during 1980 and 1981, or in 1976.

INTRODUCTION

The Big Sioux River Basin has a drainage area of about 9,000 mi^2 in eastern South Dakota, southwestern Minnesota, and northwestern Iowa (fig. 1). The basin is approximately 210 mi long and 65 mi wide at its widest sections. In South Dakota, the basin extends from southern Marshall County to southern Union County. The Big Sioux aquifer in Codington and Grant Counties is an alluvial aquifer covering the entire length of the Big Sioux River and limited mostly to its flood plain.

The Big Sioux Hydrology Study began in 1982 and includes a comprehensive county-by-county investigation of the water resources within the basin. The purpose of the study is to provide hydrogeologic information and analytical tools needed for effective management of the ground-water resources in the Big Sioux River Basin. This was to be achieved through

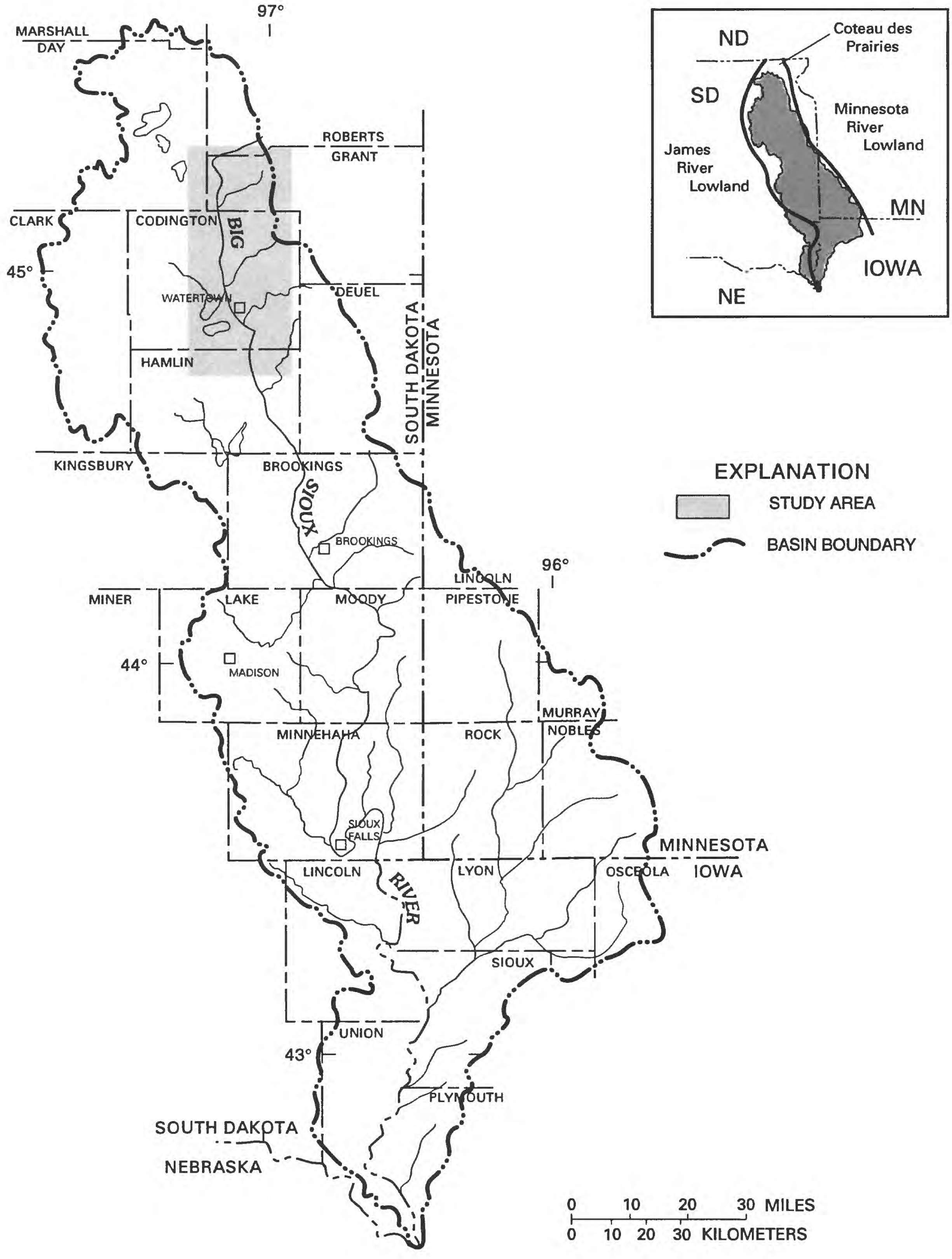


Figure 1. Location of study area.

the development of a series of digital-computer models of the Big Sioux aquifer. Each of the models was developed using a consistent set of techniques so as to be compatible with other models in the Big Sioux Hydrology Study.

Purpose and Scope

This report contains the results of model development for the Big Sioux aquifer in Codington and Grant Counties. The model area was extended into parts of Day and Roberts Counties to the north, and into Hamlin County to the south. These areas were included in the model to adequately represent naturally occurring boundaries and to include data from an additional streamflow-gaging station. The model was constructed to be used as a tool to analyze the hydrology of the system and to provide an improved, quantitative understanding of the system. This report evaluates the effects of hypothetical drought stress and additional pumpage on water levels in the Big Sioux aquifer. These stresses include decreased precipitation, increased evapotranspiration, decreased streamflow, and increased pumpage.

Previous Investigations

Hansen (1990, 1994) investigated the water resources of Codington and Grant Counties. The major sources of surface water, Lakes Kampeska and Pelican, and seven glacial and two bedrock aquifers were described. The recharge and discharge for these sources of water were discussed, as well as the quality of surface water and ground water.

Barari (1971) discussed the hydrology of Lake Kampeska and nearby Lake Pelican. The recharge-discharge relation between the lakes and underlying aquifer, the aquifer itself, and the Big Sioux River were discussed. A list of geologic logs for the area was compiled.

Acknowledgments

The authors acknowledge the cooperation of residents and municipal officials of Codington and Grant Counties for providing information concerning the water wells they own or manage. Test-hole information provided by local drilling companies used for this study also is appreciated.

HYDROGEOLOGY OF THE STUDY AREA

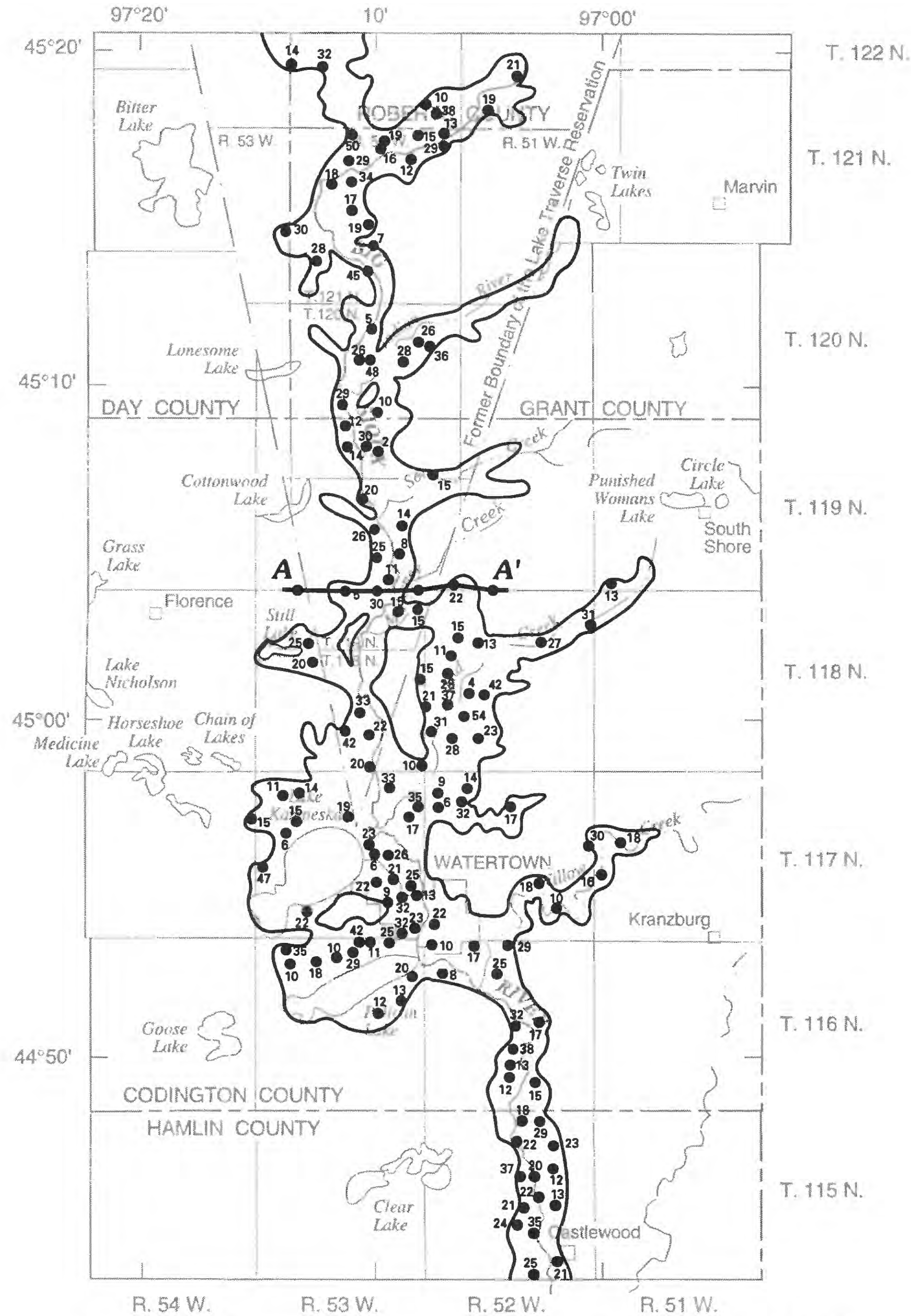
The study area is within the Coteau des Prairies, a highland plateau between the Minnesota River Lowland to the east and the James River Lowland to the west (fig. 1). The Coteau des Prairies is composed of bedrock formations overlain by unconsolidated glacial drift (outwash and till). The shallowest bedrock in the study area is the Pierre Shale of Cretaceous age, which is overlain by as much as 660 ft of unconsolidated glacial drift.

The topography of the Coteau exhibits a rough linearity in directions nearly parallel with the scarplike margins of the highland (inset, fig. 1). This results from the presence of several nearly parallel end moraines that lie along both margins. The moraines were built along the lateral margins of two lobes of glacier ice held apart by the highland between them (Flint, 1955).

The Big Sioux River Basin was formed when the James and Des Moines lobes of the Wisconsin glacier flanked the area, leaving the basin to drain meltwater and sediment from more than 400 mi of ice front. This carved out the present-day course of the Big Sioux River. The large amount of sand and gravel outwash deposited along the Big Sioux drainage overlays glacial till. As the velocity of meltwater flows began to decrease, the coarse material such as sand and gravel settled out first. Much of the finer sediments, such as silt and loam, were carried farther downstream, leaving little clay in the sand and gravel deposits.

The Big Sioux aquifer is a fluvial aquifer consisting of poorly to well-sorted surficial outwash ranging from medium sand to medium gravel. The thickness and extent of the aquifer (fig. 2) in Codington and Grant Counties is from Hansen (1994), which is slightly revised from Hansen (1990) using additional data. The extension of the aquifer boundary into Hamlin County is modified from Kume (1985). The extension of the study into Roberts County is based on data available from a water-resources investigation that was in progress in 1996.

The general direction of water movement in the Big Sioux aquifer is to the south and locally towards the Big Sioux River (Hansen, 1990). The gradient of the water-table surface generally is about 6 to 10 ft/mi, with gradients at the north edge of Grant County near 10 ft/mi. River flow depends on seasonal variations in precipitation, evapotranspiration, and ground-water storage. Small creeks hydraulically connected to the Big Sioux aquifer generally flow only during spring



Base from U.S. Geological Survey
State base map, 1961; edited 1984

0 2 4 6 MILES

0 2 4 6 KILOMETERS

EXPLANATION

AQUIFER BOUNDARY

A ——— A' LINE OF GEOLOGIC SECTION--Shown in figure 4

26 TEST HOLE--Number is thickness of aquifer, in feet

Figure 2. Thickness of sand and gravel in the Big Sioux aquifer (modified from Hansen, 1994).

and early summer because of snowmelt and rainfall runoff and discharge of ground water. Creeks generally do not flow during late fall and winter because of limited runoff and lack of ground-water discharge.

CONCEPTUAL MODEL OF THE BIG SIOUX AQUIFER

Before a ground-water system may be modeled, there must be a basic understanding of its nature. The various aspects of the system must be known well enough to ensure that they are adequately represented in the model. The physical geometry, hydraulic properties, and recharge-discharge relations are discussed in the following paragraphs.

Physical Geometry

Well and test-hole data for Codington, Grant, and parts of Roberts and Hamlin Counties were obtained from the South Dakota Geological Survey, private drillers, and other sources. The well and test-hole data provided information on the thickness, extent, and composition of the aquifer and overlying material. The Big Sioux aquifer underlies approximately 150 mi^2 of Codington and Grant Counties and is located primarily in the flood plain of the Big Sioux River. The thickness of the aquifer material ranges from 2 to 54 ft, with an average of 24 ft (Hansen, 1994) (fig. 2). The Big Sioux aquifer is primarily unconfined, with the unsaturated zone ranging in thickness from 0 to 12 ft.

The aquifer is effectively separated from underlying aquifers by relatively impermeable glacial till. The altitude of the bottom of the aquifer (fig. 3) was determined from test-hole and drillers' logs. In areas where no test holes were available, altitudes were estimated based on known values in the surrounding areas. The bottom of the aquifer within the study area ranged from 1,920 ft above sea level at the north end to 1,660 ft at the south end. A generalized cross section is shown in figure 4.

Hydraulic Properties

Twenty aquifer tests have been conducted in the Big Sioux aquifer in Moody, Brookings, and Minnehaha Counties (Ellis and Adolphson, 1969; Koch, 1980). Aquifer tests conducted on 35 wells in the Sioux Falls city well field yielded transmissivity and saturated thickness values from which hydraulic conductivity can be determined. Transmissivity is the product of hydraulic conductivity and saturated thickness. Hydraulic conductivity is the rate of flow of water through a unit cross-sectional area under a unit hydraulic gradient. Most values of hydraulic conductivity were in the range of 300 to 800 ft/d. Because the sediments north of Sioux Falls were deposited by the same glacial event and in a similar manner, the hydraulic conductivity values in the study area are assumed to be within a similar range.

The storage coefficient represents the volume of water that an aquifer releases from or takes into storage from a unit surface area of the aquifer per unit change in head. For unconfined aquifers, the coefficient is dominated by specific yield and represents the draining or filling of the pore space in the soil matrix. Koch (1980) reports specific yields, determined from aquifer tests in the Big Sioux aquifer in Brookings, Deuel, and Hamlin Counties, ranging from 10 to 17 percent. Koch chose specific yields of 0.10 (1980) and 0.20 (1982) in modeling other areas of the Big Sioux aquifer. Hansen (1988) also used a specific yield of 0.20 to model the Big Sioux aquifer in Moody County.

Specific yield of the aquifer in the study area was determined using the neutron method (Jones and Schneider, 1969; Meyer, 1962). Because specific yield is the ratio of the volume of water a saturated material will yield by gravity to its own volume, measuring the change in moisture content between fall and spring can be used to estimate specific yield. The change in water content determined with the neutron-moisture probe was used to estimate specific yield. Aluminum access tubes were placed in the aquifer at nine locations in the study area and the change in soil-moisture content was measured with a neutron probe during times of pumping or seasonal water-level change. The total water-level change during the measurement period from October 1987 to June 1988 ranged from 0.5 to 0.8 ft at the nine sites. From the neutron-probe measurements, the moisture content at the nine sites ranged from 38 percent in the saturated zone to 12 percent in the unsaturated zone. The range of specific yields determined from the neutron-probe analysis was 10 to 17 percent. The values may be smaller than the true specific yield because of incomplete drainage from the zone of water-level change.

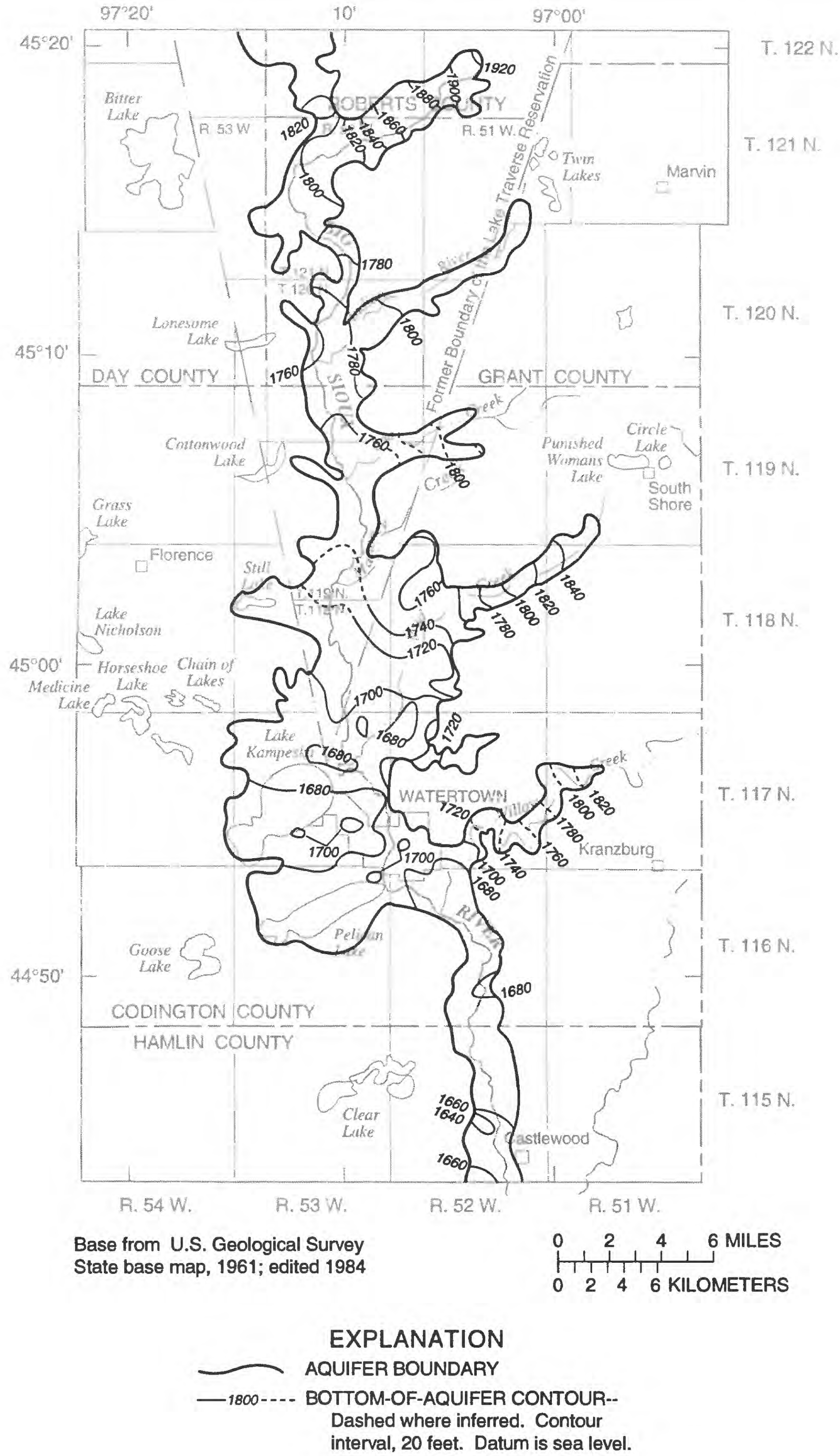


Figure 3. Altitude of the bottom of the Big Sioux aquifer in study area.

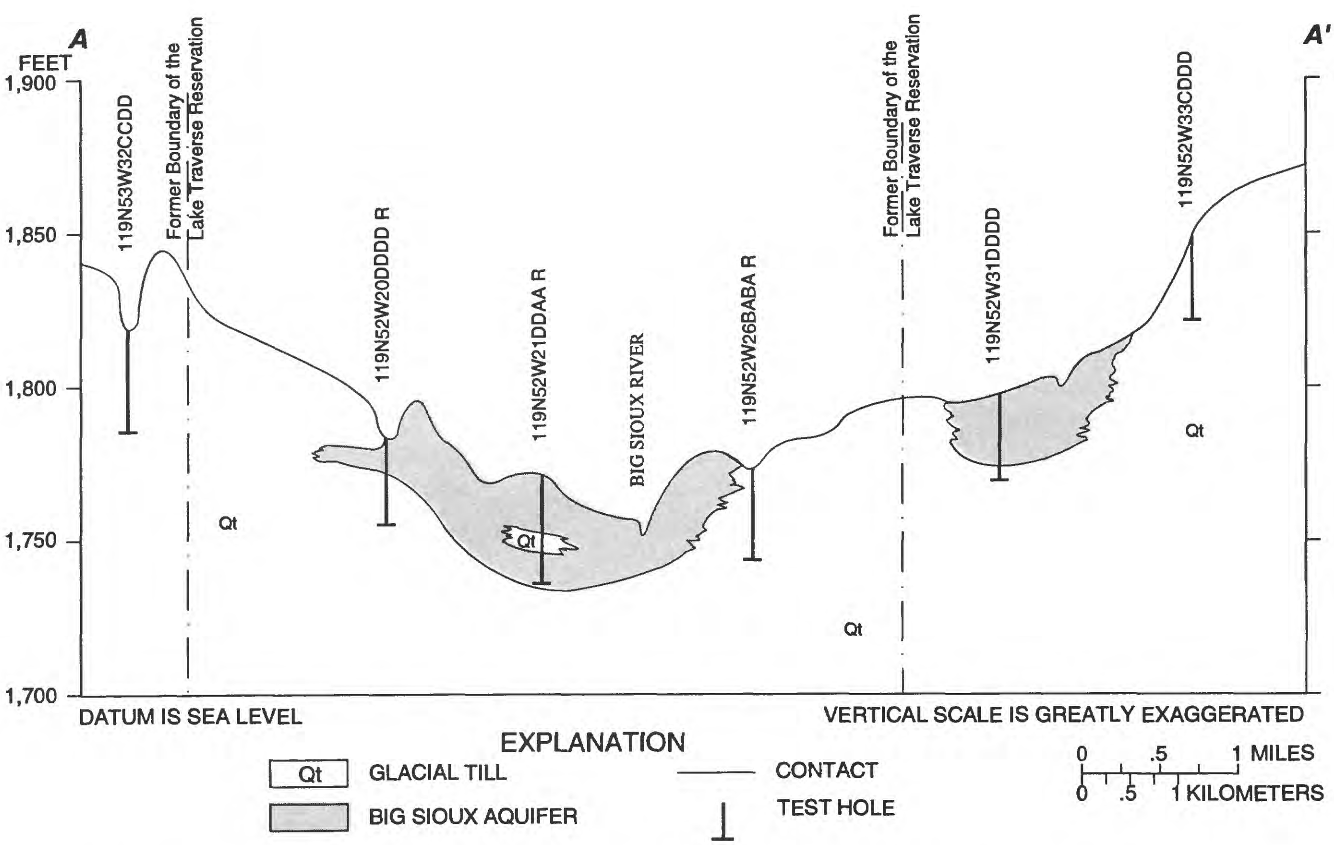


Figure 4. Geologic section A-A' showing the Big Sioux aquifer (Location of section A-A' is shown in fig. 2).

Riverbed hydraulic conductivity controls the movement of water between the aquifer and the Big Sioux River. Jorgensen and Ackroyd (1973) determined riverbed hydraulic conductivities, from three aquifer tests in the Big Sioux aquifer in Minnehaha County, that ranged from 0.5 to 1.0 ft/d. This hydraulic conductivity can be maintained where the riverbed is naturally scoured during spring runoff. Where fine sediments are deposited on the riverbed, the resistance to vertical flow is greater.

Recharge and Discharge

Recharge to the aquifer is predominantly by infiltration and subsequent percolation of rainfall and snowmelt through the overlying topsoil and by infiltration from lakes, ponds, and streams when stages are higher than water levels in the aquifer. Water-level data obtained from the South Dakota Department of Environment and Natural Resources provided information on historical water levels in the aquifer. In 1984 and 1985, additional observation wells were monitored on a monthly basis at as many as 80 sites to provide more detailed information on water-level fluctuations in the aquifer. Records of water-level fluctuations in well 118N52W21BBCB show the correlation with trends in precipitation (fig. 5). Water levels generally rise from February through June when recharge from snowmelt and spring rainfall is greater than discharge. Water levels generally decline from July through January when discharge from wells, discharge to streams, and evapotranspiration is greater than recharge. Water-level rises generally correspond with above-normal precipitation, and water-level declines correspond with below-normal precipitation. Water levels in the aquifer immediately adjacent to the Big Sioux River also may fluctuate in direct response to rises in the stage of the river. This bank storage (Freeze and Cherry, 1979) is transient in nature and returns to the river soon after the stage returns to normal. By plotting the river stage and water levels in wells within

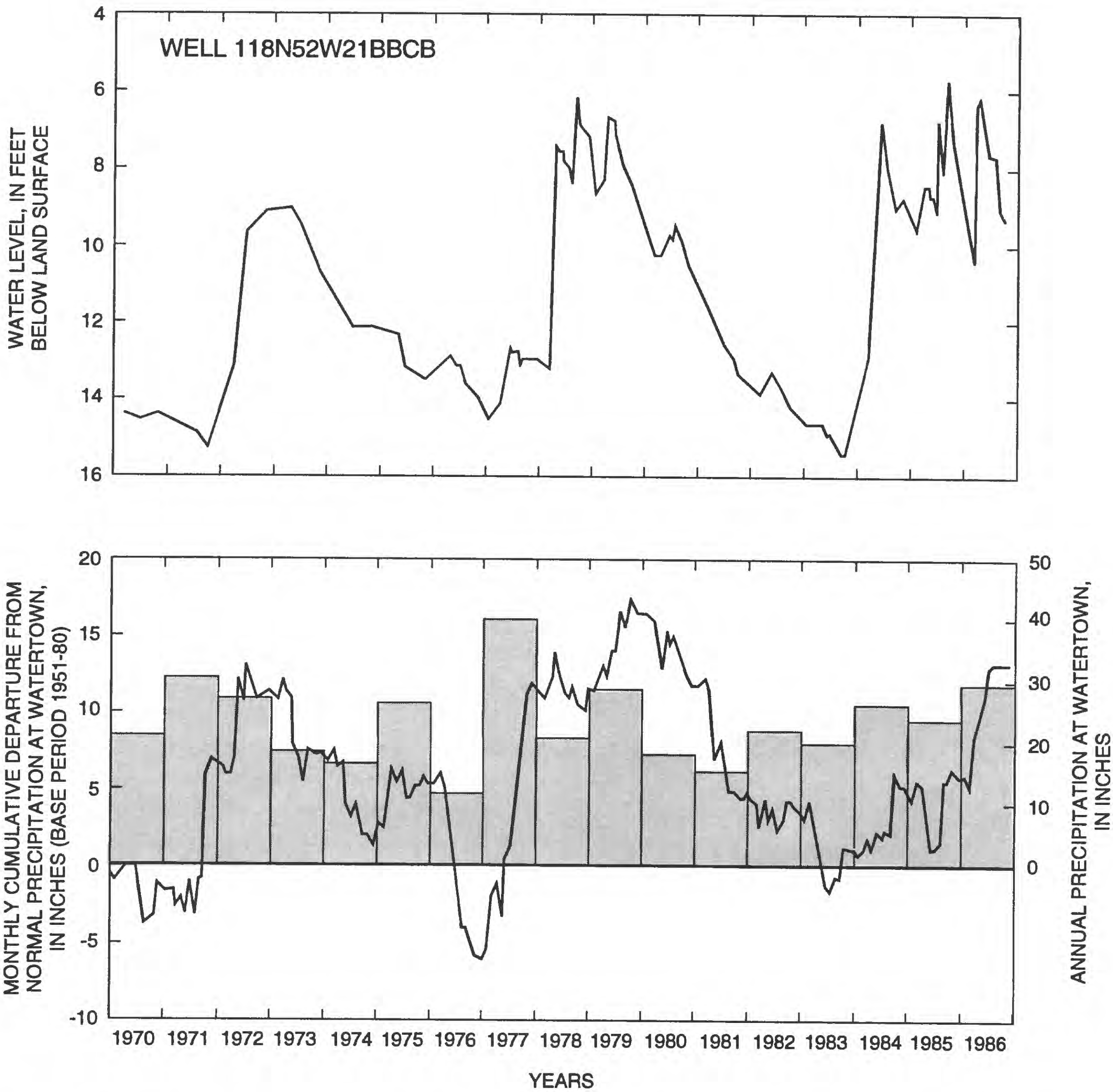


Figure 5. Water-level fluctuations in the Big Sioux aquifer and annual precipitation and monthly cumulative departure from normal precipitation.

approximately one-half mile of the Florence (fig. 6) and Watertown (fig. 7) streamflow-gaging stations, it is evident that the aquifer generally discharges to the river over time. A plot with a shorter period of well record revealed the same relationship near the Castlewood gaging station.

A range of potential recharge was determined by making a correlation to water-level rises in observation wells at least one-quarter mile from the river and believed to be unaffected by pumpage. At this distance, possible errors introduced by bank storage are assumed to be minimized. Nine such wells had an average annual water-level rise of 2.40 ft during 1978 to 1985. This period of time was selected because climatic conditions are believed to approximate long-term average conditions, and adequate water-level data are available. Using a specific yield range of 0.10 to 0.20, the average annual recharge to the aquifer may be expected to be in the approximate range of 2.9 to 5.8 in/yr. Precipitation recorded at the Watertown Airport weather station was used to correlate precipitation to recharge over the study area. The average precipitation at Watertown was 22.14 in/yr for 1961 to 1990 (U.S. Department of Commerce, 1991) and was 22.33 in/yr for 1978 to 1985 (U.S. Department of Commerce, 1978-85). Temporal recharge to the aquifer was estimated considering trends of observation-well hydrographs, precipitation, and the growth cycle of crops and associated variation in the amount of precipitation lost to evapotranspiration.

Examination of hydrographs of wells in the Big Sioux aquifer (figs. 6 and 7) indicates that most of the recharge to the aquifer occurs during the months of March through June. During much of the year, recharge is inversely related to pan evaporation. Recharge is negligible in the winter months when the ground is frozen. As the ground thaws in the spring, and snowmelt and heavy precipitation occur, recharge is at its maximum. Later in spring, warmer weather and developing plant cover increase evapotranspiration and decrease recharge. The main growing season of June through August has the most rapid plant growth and associated evapotranspiration, and thus, the lowest recharge during the growing season. By September and October, the crops generally are mature, no longer actively growing, and a killing frost may have occurred; thus, less moisture is lost through evapotranspiration and recharge increases.

The Big Sioux River has three gaging stations located within the study area—Big Sioux River near Florence (06479215), Big Sioux River near Watertown (06479438), and Big Sioux River near Castlewood (06479525). The location of gaging stations on the Big Sioux River, the drainage-area boundaries (Amundson and others, 1985), and the contributing lakes, rivers, creeks, and wastewater ponds that affect recharge and discharge of the Big Sioux aquifer are shown in figure 8. Lake Kampeska, Pelican Lake, and Still Lake are connected hydraulically to the aquifer and the Big Sioux surface-drainage system. The lakes recharge the aquifer during periods of high runoff when lake water levels are high in relation to the water levels in the aquifer. The aquifer discharges to the lakes when water levels in the lakes are low in relation to water levels in the aquifer. A water-table map of the Lake Kampeska area for June 30, 1970 (when the lake was slightly higher than average), by Barari (1971), shows gradients converging toward Lake Kampeska and Pelican Lake in the range of 3 to 8 ft/mi. The altitude of the water level in Lake Kampeska was 1,717 ft and in Pelican Lake was 1,709 ft. An additional source of recharge to the aquifer is infiltration of municipal wastewater. The city of Watertown uses spreading-basin infiltration-percolation ponds (fig. 8) as the final step for its municipal wastewater treatment. Under normal climatic conditions, the head in the ponds is greater than that in the aquifer, and the aquifer gains water from the wastewater infiltration ponds. Under extreme wet conditions, the head in the aquifer may be greater than that in the ponds, and the aquifer would lose water to the ponds.

Discharge from the Big Sioux aquifer is by evapotranspiration, ground-water discharge to lakes and to the Big Sioux River, and pumping from irrigation, municipal, rural-water-system, industrial, domestic, and stock wells. Pan evaporation was used to estimate the amount of evapotranspiration from the Big Sioux aquifer. The average potential evapotranspiration in the study area is estimated to be about 70 percent of the pan evaporation (Farnsworth and others, 1982), or about 34.7 in/yr. The average 1978-85 monthly pan evaporation and the 1984 and 1985 monthly pan evaporation at Brookings, and the estimated potential evapotranspiration, are shown in table 1. These years were chosen to correlate with detailed water-level records and other data available in 1984 and 1985. The potential maximum evapotranspiration represents the evapotranspiration that would occur when the water level is at the land surface. Greater transpiration occurs from cottonwood trees and other phreatophyte vegetation along creeks and the river.

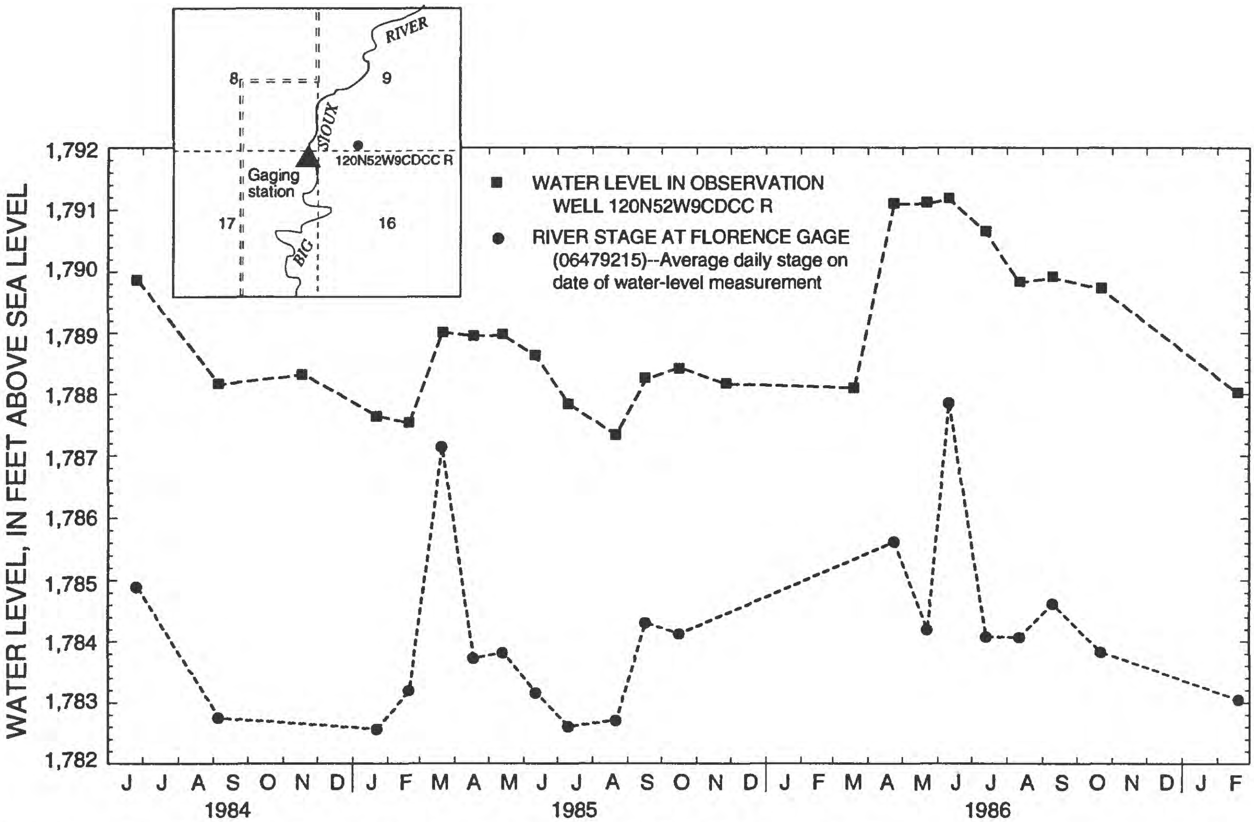


Figure 6. Ground-water level and river stage at Big Sioux River near Florence gaging station.

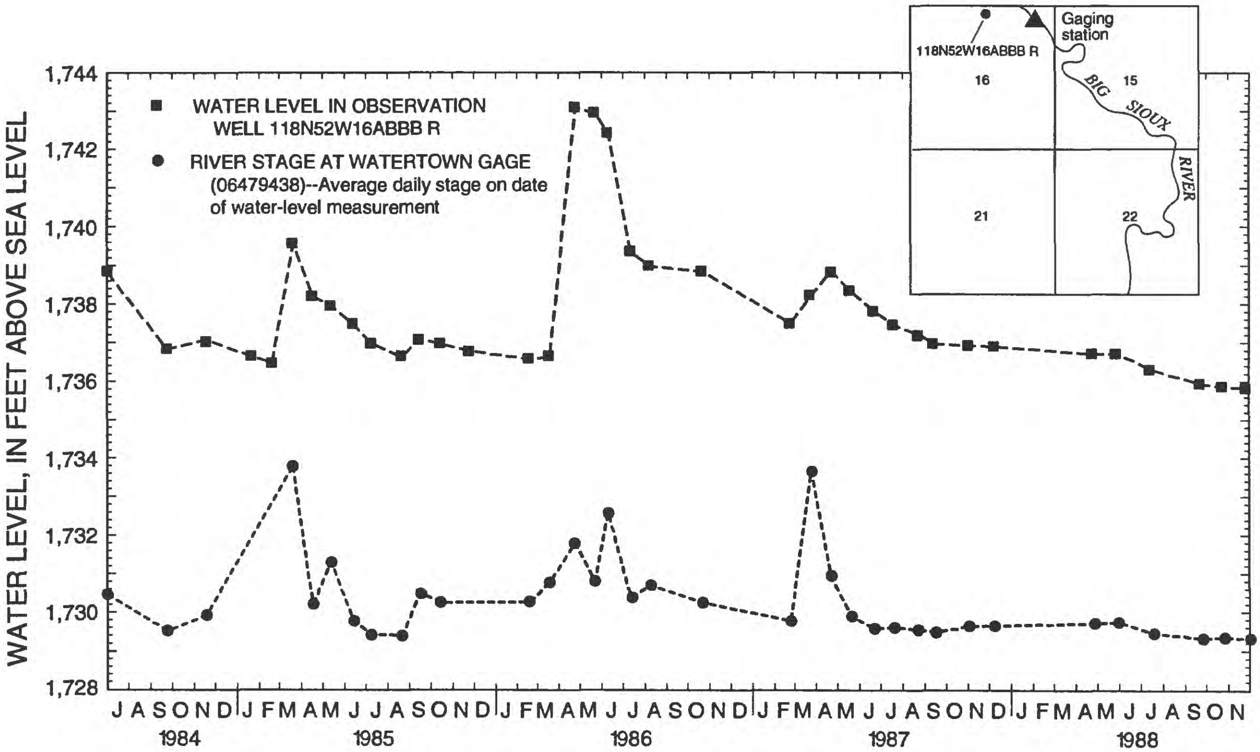


Figure 7. Ground-water level and river stage at Big Sioux River near Watertown gaging station.

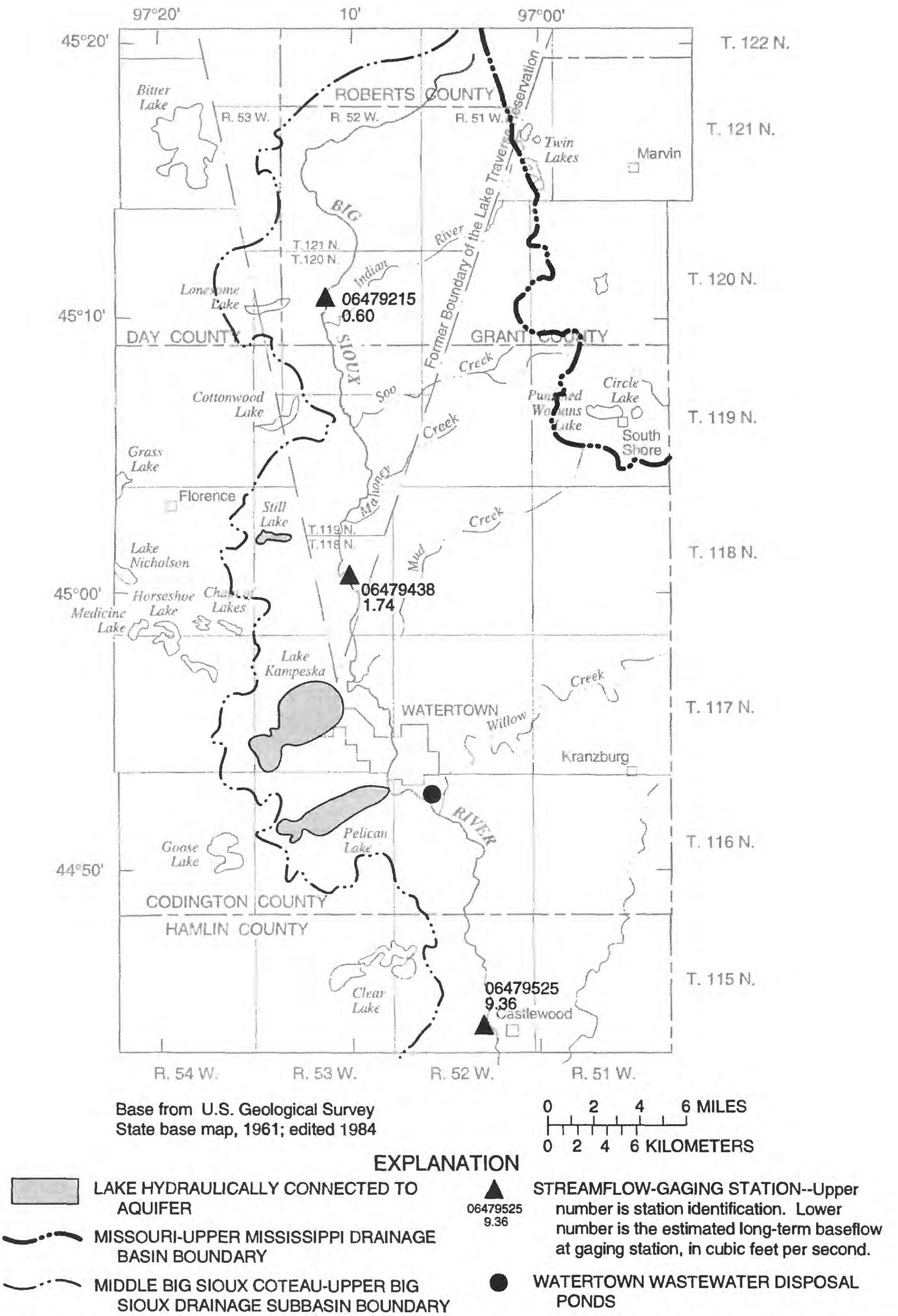


Figure 8. Drainage basins, locations of streamflow-gaging stations, locations of lakes and wastewater disposal ponds that are hydraulically connected to the aquifer, and the estimated baseflow at gaging stations.

Table 1. Pan evaporation for Brookings and estimated potential evapotranspiration

[--, no data or not calculated]

Month	Average for 1978-85: Pan evaporation (inches)	Average for 1978-85: Maximum potential evapotranspiration[1] (inches)	1984: Pan evaporation[2] (inches)	1984: Maximum potential evapotranspiration[1] (inches)	1985: Pan evaporation[2] (inches)	1985: Maximum potential evapotranspiration[1] (inches)
January	--	--	--	--	--	--
February	--	--	--	--	--	--
March	--	--	--	--	--	--
April	5.39	3.77	[3]6.00	[3]4.20	[4]6.17	4.32
May	7.56	5.29	7.75	5.43	8.04	5.63
June	8.65	6.06	[3]8.00	5.60	8.35	5.85
July	9.39	6.57	9.05	6.34	11.01	7.71
August	8.26	5.78	7.56	5.29	7.02	4.91
September	6.07	4.25	5.81	4.07	[4]3.64	2.55
October	4.27	2.99	[3]3.00	2.10	[4]3.04	2.13
November	--	--	--	--	--	--
December	--	--	--	--	--	--
Total	49.59	34.71	47.17	33.02	47.27	33.10

[1]Calculated as 0.7 times the monthly pan evaporation.
[2]Data for Brookings, South Dakota (U.S. Department of Commerce, 1984-85), except as indicated.
[3]Estimated from other years.
[4]Data for Madison, South Dakota (U.S. Department of Commerce, 1985).

Average baseflow from the Big Sioux aquifer to the Big Sioux River was estimated by analysis of streamflow data. December and January streamflow normally are not influenced by runoff events, so these months provide an estimate of baseflow. The long-term average December and January flows in the Big Sioux River at the Florence gage (06479215) are 0.36 and 0.85 ft^3/s, respectively, for the period of record (U.S. Geological Survey, 1994). Averaging these two values to estimate baseflow yields 0.60 ft^3/s, or about 5 percent of the annual mean flow of 11.4 ft^3/s. The long-term average December and January flows in the Big Sioux River at the Watertown gage (06479438) are 2.45 and 1.02 ft^3/s, respectively, for the period of record (U.S. Geological Survey, 1994). The average of these two numbers would result in a baseflow estimate of 1.74 ft^3/s, which is about 6 percent of the annual mean flow of 27.7 ft^3/s. The long-term average December and January flows in the Big Sioux River at the Castlewood gage (06479525) are 6.73 and 12.7 ft^3/s, respectively, for the period of record (U.S. Geological Survey, 1994). The average of these two months would yield a baseflow estimate of 9.72 ft^3/s, or about 15 percent of the annual mean flow of 64.2 ft^3/s. The locations of these three gaging stations and their estimated baseflows are shown in figure 8.

Municipal, rural-water system, and irrigation use are the largest withdrawals from the Big Sioux aquifer. Industrial, farm, and domestic use generally is very small. Discharge by irrigation, rural-water-system, and municipal wells was obtained from annual irrigation reports supplied by the South Dakota Department of Environment and Natural Resources, Water Rights Program, and from pumpage records obtained from rural water systems and municipalities. The average withdrawal from 1978-85 for irrigation use was 1.0 ft^3/s. Municipal use by the city of Watertown averaged 1.1 ft^3/s, and rural-water-system use averaged 0.2 ft^3/s, as reported by the Sioux Rural Water System. A summary of the ground-water withdrawal rates from the aquifer is shown in table 2.

Table 2. Ground-water withdrawal rates for the study area

Period		Ground-water withdrawal rates, in cubic feet per second			
		Irrigation[1]	Municipal use[2]	Rural water[3]	Total
Average (1978-85)		1.0	1.1	0.2	2.3
1984	January	0	.4	.6	1.0
	February	0	2.4	.6	3.0
	March	0	2.1	.5	2.6
	April	0	1.6	.6	2.2
	May	.1	1.5	.5	2.1
	June	.2	.8	.5	1.5
	July	.7	2.4	.5	3.6
	August	.8	2.1	.5	3.4
	September	.3	2.2	.6	3.1
	October	.1	1.0	.5	1.6
	November	0	1.8	.5	2.3
	December	0	1.7	.5	2.2
	Average	.2	1.7	.5	2.4
1985	January	0	1.8	.5	2.3
	February	0	1.9	.6	2.5
	March	0	1.6	.6	2.2
	April	0	1.6	.6	2.2
	May	.1	1.5	.6	2.2
	June	.4	2.6	.5	3.5
	July	.7	3.2	.6	4.5
	August	.5	1.7	.5	2.7
	September	.1	2.2	.6	2.9
	October	0	.5	.4	.9
	November	0	1.2	.5	1.7
	December	0	.7	.5	1.2
	Average	.2	1.7	.5	2.4

[1]Data from South Dakota Department of Environment and Natural Resources, Water Rights Program.

[2]Data from City of Watertown.

[3]Data from Sioux Rural Water System.

DESCRIPTION OF THE DIGITAL MODEL

A digital-computer model, or simply a digital model, is a mathematical model that uses a digital computer to obtain approximate solutions to the partial-differential equations that describe ground-water flow. Continuous derivatives of the partial-differential equations of ground-water flow are replaced by finite-difference approximations at the node (centroid) of cells arranged in a rectangular grid. The digital model used in this study was the U.S. Geological Survey's modular three-dimensional finite-difference ground-water-flow model (MODFLOW), written by McDonald and Harbaugh (1988).

The model was designed taking into consideration the hydrogeologic setting, aquifer boundaries, hydraulic properties, recharge, and discharge. These hydrologic aspects, which require simplifying assumptions, were quantified by subdividing the simulated area into a series of finite-difference cells within which aquifer properties were assumed to be constant. A value was then assigned for the aquifer properties that characterize the system at each model cell. Flow in the aquifer was assumed to be lateral and two dimensional. The resulting arrays of aquifer parameters for specified time periods were assembled to portray the aquifer in a form such that computerized numerical-solution techniques could be used. The Strongly Implicit Procedure (SIP) numerical technique was used to solve the series of finite difference equations. This solution sequence was used to test interpretations, calibrate the model, and analyze hypothetical hydrologic situations.

Representation of Physical Geometry

The equally spaced finite-difference grid selected to represent the study area has 172 rows and 60 columns (fig. 9). The model cells are 1,320 ft on a side to remain compatible with the models presented in previous reports of this series. Each model cell shown in figure 9 represents one-sixteenth of a 640-acre section. The model area was extended beyond the study area (the Big Sioux aquifer in Codington and Grant Counties) 3 mi north into Roberts County to minimize the effect of boundary conditions, and 6 mi south into Hamlin County to include the streamflow-gaging station on the Big Sioux River near Castlewood. The modeled aquifer boundary was modified from the actual aquifer boundary in areas where the saturated thickness was insufficient to simulate aquifer stresses.

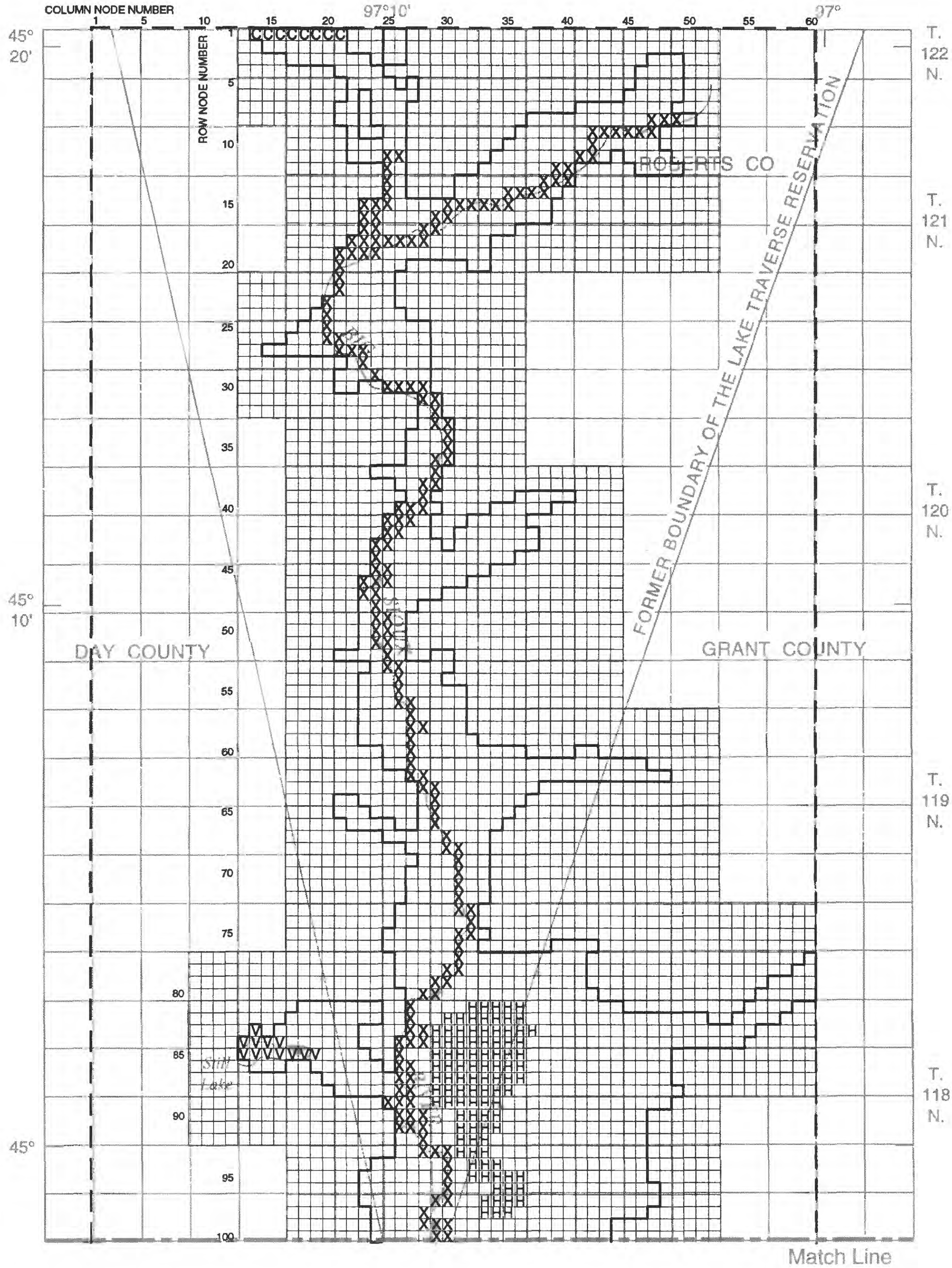


Figure 9. Model area and boundary conditions represented in the model.

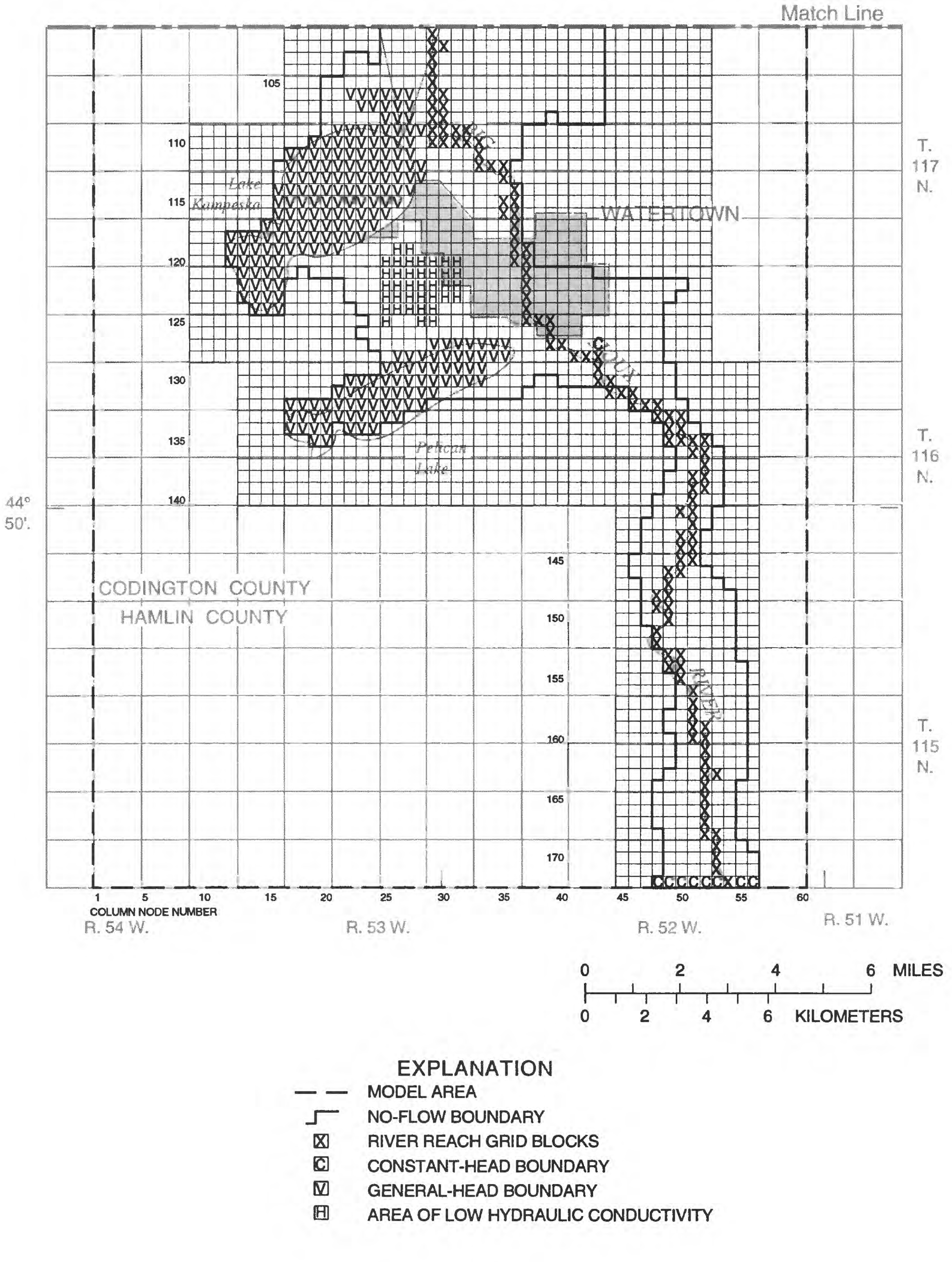


Figure 9. Model area and boundary conditions represented in the model.--Continued

Although the thickness of the alluvium along Willow Creek is as much as 30 ft, the gradient of the bottom of the aquifer was such that cells went dry during periods of low recharge. A zone of inactive model cells surrounds the aquifer boundary to allow for possible future reinterpretations of the aquifer extent. This resulted in a total of 10,320 cells, with 2,341 active cells representing the aquifer.

The boundary conditions in the model were selected to best reflect the field conditions. Where the aquifer is bounded by till, no-flow boundaries were used. General-head boundaries were used for Lake Kampeska, Pelican Lake, and Still Lake to enable simulated lake-level elevations to be varied monthly during transient simulations. The Watertown wastewater infiltration ponds were modeled as a constant-head boundary because the water levels in the ponds remain relatively steady throughout the year. On the north edge of the model, near the headwaters of the Big Sioux River, constant-head cells were used to represent the continuation of the aquifer north beyond the study area. The south edge of the study area was treated in the same way.

Representation of Hydraulic Properties

The assignment of an average aquifer hydraulic conductivity to each model cell is based on the assumption that aquifer materials are uniform in each cell and the test-hole and drillers' logs adequately depict the types of aquifer materials in the area surrounding the test hole. Hydraulic conductivity of the aquifer for each cell in the model was estimated by comparing grain size reported by driller's logs with similar grain sizes where aquifer tests had been performed to determine hydraulic conductivity. The 55 aquifer tests that were referenced were completed in areas of the Big Sioux aquifer outside of this study area. Hydraulic conductivity for model cells with no test-hole data was estimated using known values in nearby model cells. If the estimated hydraulic conductivity values caused simulated water levels to be different from measured water levels, the values were adjusted slightly within a plausible range to resolve the differences. The assigned aquifer hydraulic conductivity to the model cells ranged from 50 to 500 ft/d, with most equaling 350 ft/d. Two zones of sandy till within the aquifer along the east edge of T. 118 N., R. 53 W. and the south edge of T. 117 N., R. 53 W. were assumed to have a significantly lower hydraulic conductivity than the surrounding areas and were assigned a value of 50 ft/d.

A uniform specific yield of 14 percent was used in the transient model simulations based on the range of 10 to 17 percent determined from neutron-probe analysis. This falls within the range of values used in previous modeling efforts on the Big Sioux aquifer.

Transmissivity, the ability of an aquifer to transmit water, is computed for unconfined aquifers as the product of the hydraulic conductivity and the saturated thickness of the aquifer. The model calculates the saturated thickness at each node by subtracting the altitude of the bottom of the aquifer (based on fig. 3) from the altitude of the simulated water level.

Representation of Recharge and Discharge

Recharge was assigned uniformly to each active model cell. Based on water-level-rise measurements, the most likely average recharge rate was estimated to be between 2.9 and 5.8 in/yr. A steady-state recharge rate of 5.53 in/yr best reproduced average water levels (1978-85). The temporal distribution of recharge was quantified by the simulated matching of water-level hydrographs as described in the subsequent model-calibration section. An empirical monthly recharge multiplication factor was used to convert precipitation values to aquifer-recharge values. Thus, the monthly recharge factors represent the percentage of total precipitation that infiltrates into the soil and is not lost to evapotranspiration. Recharge rates to the aquifer on a monthly basis (table 3) agree with the conceptual model of recharge from precipitation.

The evapotranspiration computed for each model cell decreased with increasing depth of the water level below the land surface. The altitude of the land surface was determined from U.S. Geological Survey 7.5-minute topographic maps. The model calculates a linear reduction in evapotranspiration between the maximum potential evapotranspiration (table 1) and a depth at which the evapotranspiration is zero. This depth is referred to as the "extinction depth." Following Koch (1982), an extinction depth of 5 ft was used except along the Big Sioux River. Hansen (1988) reported an improved correlation between simulated and observed water levels when the extinction depth along river cells was increased to greater than 5 ft. Therefore, an extinction depth of 10 ft was used along the river.

Table 3. Precipitation for Watertown and estimated aquifer recharge

Month	Recharge multipli-cation factor[1]	Normal (1978-85)		1984		1985	
		Precipitation[2] (inches)	Recharge to the aquifer[3] (inches)	Precipitation[2] (inches)	Recharge to the aquifer[3] (inches)	Precipitation[2] (inches)	Recharge to the aquifer[3] (inches)
January	0.00	0.56	0.00	0.46	0.00	0.26	0.00
February	.00	.70	.00	.37	.00	.14	.00
March	.55	1.21	.67	1.34	.74	2.22	1.22
April	.55	2.20	1.21	3.03	1.67	1.99	1.09
May	.35	3.14	1.10	2.30	.81	1.27	.44
June	.25	3.96	.99	5.28	1.32	1.57	.39
July	.10	2.96	.30	2.38	.24	[4]2.95	.30
August	.10	2.77	.28	3.46	.35	[4]3.62	.36
September	.30	1.60	.48	1.29	.39	[4]8.03	2.41
October	.30	1.66	.50	5.51	1.65	1.68	.50
November	.00	.89	.00	.06	.00	1.42	.00
December	.00	.68	.00	.74	.00	.45	.00
Total	--	22.33	5.53	26.22	7.17	25.60	6.71

[1]The maximum decimal fraction of average precipitation that could potentially recharge the aquifer.
[2]Data for Watertown (U.S. Department of Commerce, 1978-85), except as indicated.
[3]Calculated by multiplying the monthly precipitation by the recharge multiplication factor.
[4]Data for Castlewood (U.S. Department of Commerce, 1978-85).

Recharge to the aquifer from the river and discharge from the aquifer to the river was simulated using the river package within MODFLOW (McDonald and Harbaugh, 1988). Flow between the river and the aquifer is calculated by applying Darcy's law: flow to or from the river in a model cell is equal to the riverbed conductance multiplied by the difference in head between the river and the aquifer in that cell. The riverbed conductance is equal to the hydraulic conductivity of the riverbed material multiplied by the river area in the cell and divided by the thickness of the riverbed material. The riverbed hydraulic conductivity and thickness is constant throughout the steady-state and transient simulation for each node. Riverbed conductance and head in the river (stage) are model inputs. If the water level in the aquifer is below the bottom of the riverbed, the difference in head is equal to the river stage minus the altitude of the bottom of the riverbed material. The altitude of the bottom of the river was determined using records from gaging stations and measurements at bridges. The bottom of the river at reaches between measurements was interpolated using U.S. Geological Survey 7.5-minute topographic maps.

The values of hydraulic conductivity used to estimate riverbed conductance within the model area ranged from 0.05 to 1.0 ft/d. The river was discretized by noting where the river traverses model cells as shown in figure 9. The reach length and the width of the riverbed were determined from U.S. Geological Survey 7.5-minute topographic maps. The width of the riverbed ranged from 50 to 250 ft. The thickness of the riverbed material was assumed to be 1 ft.

The average stage of the river for 1978 through 1985 was estimated from data collected at streamflow-gaging stations (U.S. Geological Survey, 1979-86) located near Florence, Watertown, and Castlewood and from various bridge measurements. The average stage of the river for the monthly transient simulations was interpolated from the mean monthly measured stages at the gages near Florence, Watertown, and Castlewood.

Lake Kampeska, Pelican Lake, and Still Lake were represented by general-head boundary model cells. The steady-state altitude for Lake Kampeska was 1,717 ft, for Pelican Lake was 1,709 ft, and for Still Lake was 1,741 ft. Watertown wastewater infiltration ponds were represented as constant-head cells with an altitude of 1,715 ft. For transient simulations, the water levels in Lake Kampeska were varied according to measurements recorded by the city of Watertown. Water-level records for Pelican and Still Lakes were not available, so changes in their water levels were estimated based on the changes in Lake Kampeska water level for the same time period.

CALIBRATION OF THE DIGITAL MODEL OF THE BIG SIOUX AQUIFER

Model calibration is the process by which model parameters are adjusted within reasonable hydrologic constraints to obtain the optimum match with historical water levels and river baseflows. The model calibration process involved a steady-state simulation, a steady-state sensitivity analysis, and a transient simulation.

Steady-state conditions (1978 through 1985) were simulated by setting the change in storage to zero and using average recharge, evapotranspiration, river stage, and pumpage. Monthly transient simulations in 1984 and 1985, and the transient antecedent simulations leading up to them, included storage and time-dependent recharge, evapotranspiration, river stage, and pumpage. Parameters that were varied included hydraulic conductivity, recharge, evapotranspiration extinction depth, and specific yield. Recharge to the aquifer was varied by adjusting the monthly recharge factors to best approximate water levels during the transient calibration. Following each change in recharge factors, the steady-state and antecedent simulations were run to provide appropriately adjusted antecedent conditions for the transient simulation. The best composite set of average and absolute differences obtained between the simulated and observed water levels for the steady-state simulation and the 1984 and 1985 monthly transient simulations is shown in table 4.

Monthly water-level measurements at 17 wells completed in the aquifer were recorded from 1978 to

Table 4. Comparison between simulated and observed water levels in the aquifer for steady-state and transient simulations

[--, no data]

Model simulation		Average difference between simulated and observed water levels[1] (feet)	Average absolute difference between simulated and observed water levels[2] (feet)	Number of observation wells with observed water levels
Steady-state		0.89	1.20	17
Transient				
1984	April	.85	2.02	17
	May	.79	2.12	30
	June	-.16	1.69	44
	July	.11	2.09	21
	August	-.39	2.21	40
	September	.38	2.04	55
	October	-.35	1.29	3
	November	-.30	2.07	80
	December	--	--	0
1985	January	-.09	1.62	55
	February	.24	1.81	55
	March	-.65	1.78	59
	April	-.22	1.54	62
	May	-.61	1.48	61
	June	-.37	1.50	61
	July	-.12	1.34	62
	August	-.16	1.39	60
	September	-.21	1.57	62
	October	-.15	1.58	62
	November	.51	1.47	30

[1]Summation of the difference between simulated and observed water levels divided by number of observation wells. Positive number indicates simulated water level was higher than the observed water level; negative number indicates simulated water level was lower than the observed water level.

[2]Summation of the absolute values of simulated minus observed water levels divided by number of observation wells.

1985. An additional set of observation wells were measured for calibration of the transient-state time period of April 1984 through November 1985. In order to take into account antecedent conditions, heads from a steady-state simulation were used as initial conditions for a series of yearly transient antecedent simulations of 1978 through 1982. End-of-1982 conditions were used to start seasonal transient simulations of 1983. Beginning with December 1983, monthly transient simulations were run to provide starting heads for the monthly transient simulations used to calibrate the model from April 1984 through November 1985.

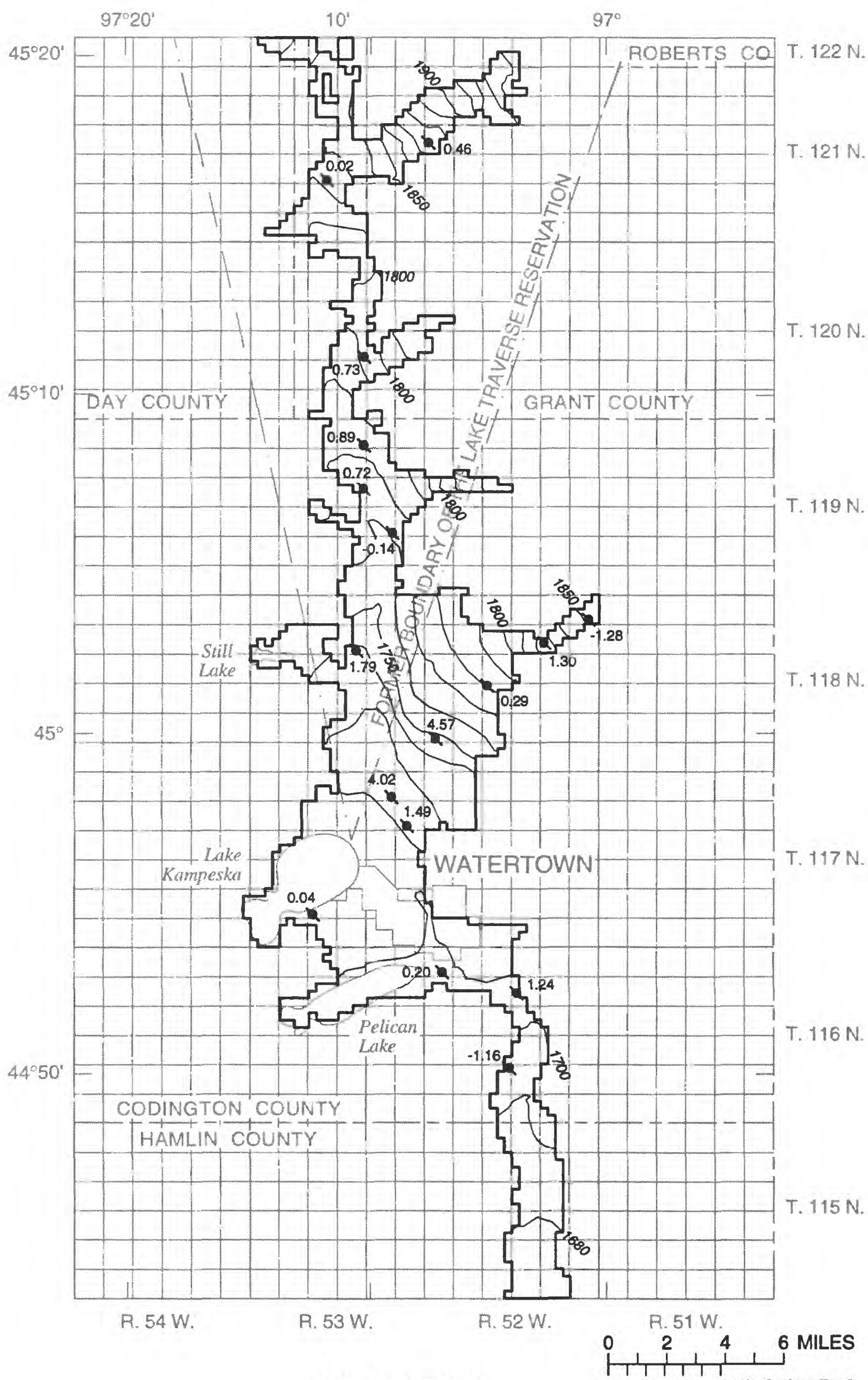
Steady-State Simulation

The steady-state simulation shows the general flow pattern in the aquifer to the south and towards the river (fig. 10). The flatter gradients occur along the Big Sioux River, Lake Kampeska, and Pelican Lake. The steeper gradients occur in the glacial outwash in the tributary valleys. The aquifer water level ranges from 1,940 ft at the south edge of Roberts County to 1,680 ft in Hamlin County, with an average north-south gradient of about 5 ft/mi. The gradient ranges from about 14 ft/mi along the upper reaches of Mud Creek valley to 3 ft/mi near Lake Kampeska. Calibration of the steady-state model was accomplished by comparing a representative observed water level at each well for the 1978-85 period with the simulated water level at the nearest node. Due to the grid size used in the model, the distance between the actual well and the nearest node could be as much as 900 ft, and may account for some of the difference between simulated and observed heads.

The observed water levels at each calibration well are listed in table 5 and compared to the simulated water level at the nearest node. The average difference between simulated and observed water levels was 0.89 ft, and the average absolute difference was 1.20 ft. Much of the average difference between simulated and observed heads is due to a large difference in a small

Table 5. Difference between simulated and observed water levels for steady-state simulation

Well Location	Location of nearest model cell: Row	Location of nearest model cell: Column	Observed water levels (feet above sea level)	Simulated water levels (feet above sea level)	Arithmetic difference between simulated and observed water levels (feet)
121N52W1CBBBR	15	37	1,874.34	1,874.80	0.46
121N52W8DCCCR	20	23	1,823.00	1,823.02	.02
120N52W9DDDDR	44	28	1,791.05	1,791.78	.73
120N52W28DDDDR	56	28	1,772.84	1,773.73	.89
119N52W4ADDDR	62	28	1,763.68	1,764.40	.72
119N52W10DDDDR	68	32	1,761.71	1,761.57	-.14
118N52W1DCDC	80	59	1,853.29	1,852.03	-1.26
118N52W11CBBC	83	53	1,824.30	1,825.60	1.30
119N52W33DCDCR	84	27	1,736.77	1,738.56	1.79
118N52W21BBCB	89	45	1,780.11	1,780.40	.29
118N52W30CDCD	96	38	1,743.17	1,747.74	4.57
117N53W2DDDC	104	32	1,718.76	1,722.78	4.02
117N53W12CDDD	108	34	1,720.53	1,722.02	1.49
117N53W28CCBB	120	21	1,717.89	1,717.93	.04
116N52W6DCCC	128	39	1,708.95	1,709.15	.20
116N52W10CBBC	131	49	1,705.11	1,706.35	1.24
116N52W28AAAA	141	48	1,695.88	1,694.72	-1.16
	Average				0.89



EXPLANATION

—*1800*— SIMULATED WATER-TABLE CONTOUR--Shows altitude of simulated water table based on average hydrologic conditions, 1978-85. Contour interval, 10 feet. Datum is sea level.

1.24 OBSERVATION WELL--Number indicates the difference between simulated and observed water level, in feet. Positive number indicates simulated water level was higher than observed water level.

Figure 10. Water-table contours and differences between simulated and observed water levels, steady-state conditions.

portion of the model near observation wells 118N52W30CDCD and 117N53W2DDDC (located in row 96, column 38 and row 104, column 32, respectively). The average arithmetic difference for all other wells outside this area is 0.44 ft. A possible explanation for the relatively large deviation of these two wells may be associated with the proximity of the wells to a 6-mi^2 area described previously and within the boundaries of the aquifer where test holes indicate sandy till. An intermittent stream, Mud Creek, runs along the eastern edge of this till area and may have influenced the measured water levels at the affected observation wells. Efforts to improve the fit of the water levels in this small area had a negative effect on the agreement between simulated and measured water levels in the surrounding area.

The simulated steady-state water budget is shown in table 6. Approximately 76 percent of the inflow was recharge from precipitation and 54 percent of the outflow was evapotranspiration. Evapotranspiration occurred in 399 of the 2,341 active cells in the model. Many of the cells with evapotranspiration are simulated as, or are located adjacent to, river cells. Some of these cells represent scattered, isolated areas where the river is losing water to the aquifer. This water is gained by the aquifer and is in turn lost to the atmosphere through evapotranspiration.

Table 6. Simulated water budget for model area for steady-state conditions

Budget component	Flow rates, in cubic feet per second	Percent
INFLOW		
Recharge to the aquifer from precipitation	49.5	75.9
Recharge to the aquifer from the river	11.7	17.9
Inflow at constant-head boundaries	3.7	5.7
Recharge to aquifer from lakes	.3	.5
Total inflow	65.2	100.0
OUTFLOW		
Evapotranspiration from the aquifer	35.4	54.3
Pumpage	2.3	3.5
Discharge from the aquifer to the river	23.6	36.2
Outflow at constant-head boundaries	1.0	1.5
Discharge from the aquifer to lakes	2.9	4.5
Total outflow	65.2	100.0

By dividing the model into zones whose boundaries were at gaging stations, the model-generated baseflow could be compared to that measured at gaging stations. For the period of record at Florence, the average January and December flows were 0.36 and 0.85 ft^3/s, respectively (U.S. Geological Survey, 1994). Averaging these two values to estimate baseflow yields 0.60 ft^3/s. For the period of record at Watertown, the average January and December flows were 1.02 and 2.45 ft^3/s, respectively (U.S. Geological Survey, 1994). Averaging these two values to estimate baseflow yields 1.74 ft^3/s. The period of record at Castlewood had average January and December flows of 6.73 and 12.7 ft^3/s, respectively (U.S. Geological Survey, 1994). Averaging these two values to estimate baseflow yields 9.72 ft^3/s.

Starting with the uppermost reach of the river (above the Florence gaging station), the modeled baseflow gain was 0.86 ft^3/s while the estimated baseflow gain was 0.60 ft^3/s. For the river reach between the Florence and Watertown gaging stations, the modeled baseflow gain was 1.62 ft^3/s. Estimated baseflow gain was 1.74 minus 0.60 ft^3/s, or 1.14 ft^3/s. Finally, for the river reach between the Watertown and Castlewood gaging stations, the modeled baseflow gain was 9.36 ft^3/s. Estimated baseflow gain for this reach was 9.72 minus 1.74 ft^3/s, or 7.98 ft^3/s. The slightly higher baseflow gains generated by the model were representative of average aquifer levels, while the baseflow gains estimated from gaging-station data were from the lower aquifer levels of winter.

The average saturated thickness of the aquifer, as indicated by steady-state simulation, was 20 ft. The largest area with an average saturated thickness less than 10 ft was in the north one-third of Codington County. The areas with less saturated thickness would limit the amount of pumping that could be supported.

Steady-State Sensitivity Analysis

A sensitivity analysis of the steady-state model was made by comparing the simulated water levels with those simulated using a modified model parameter. The change in the simulated water levels in relation to the steady-state simulated water levels provides a description of the relative effect of changes in model parameters on the results of the steady-state model. The sensitivity of the steady-state condition to changes in recharge, maximum evapotranspiration rate, evapotranspiration extinction depth, aquifer hydraulic conductivity, and riverbed hydraulic conductivity is shown in table 7.

Table 7. Model sensitivity to changes in recharge, maximum evapotranspiration rate, evapotranspiration extinction depth, aquifer hydraulic conductivity, and riverbed hydraulic conductivity

Steady-state model simulation	Average arithmetic difference between simulated and observed water levels[1] (feet)	Average absolute difference between simulated and observed water level[2] (feet)	Number of observation wells with observed water levels
Calibrated model	0.89	1.20	17
Recharge reduced 20 percent	.37	1.21	17
Recharge increased 20 percent	1.33	1.41	17
Maximum evapotranspiration rate decreased 20 percent	1.04	1.30	17
Maximum evapotranspiration rate increased 20 percent	.77	1.12	17
Evapotranspiration extinction depth reduced 20 percent	1.20	1.42	17
Evapotranspiration extinction depth increased 20 percent	.58	1.03	17
Aquifer hydraulic conductivity reduced 20 percent	1.20	1.32	17
Aquifer hydraulic conductivity increased 20 percent	.63	1.27	17
Riverbed hydraulic conductivity decreased 20 percent	.93	1.23	17
Riverbed hydraulic conductivity increased 20 percent	.87	1.17	17

[1]Summation of simulated minus observed water levels in corresponding model cells divided by number of observation wells with observed water levels. Positive number indicates simulated water level was higher than the observed water level; negative number indicates simulated water level was lower than the observed water level.

[2]Summation of the absolute values of simulated minus observed water levels in corresponding model cells divided by number of observation wells with observed water levels.

The steady-state simulation was the most sensitive to changes in recharge. A 20-percent reduction in the recharge rate resulted in a 0.52-ft decrease in the average arithmetic difference and 0.01-ft increase in the average absolute difference. A 20-percent increase in the recharge rate resulted in a 0.44-ft increase in the average arithmetic difference and 0.21-ft increase in the average absolute difference.

The effects of a similar change in the evapotranspiration extinction depth produced less change in water level. A 20-percent reduction in the extinction depth resulted in a 0.31-ft increase in the average arithmetic difference and 0.22-ft increase in the average absolute difference. A 20-percent increase in the extinction depth resulted in a 0.31-ft decrease in the average arithmetic difference and 0.17-ft decrease in the average absolute difference. Changing the evapotranspiration rate had a similar, but smaller effect.

The steady-state simulation of changes in aquifer hydraulic conductivity indicated less sensitivity than changes in the recharge rate or extinction depth. A 20-percent reduction in hydraulic conductivity resulted in a 0.31-ft increase in the average arithmetic difference and a 0.12-ft increase in the average absolute difference. A 20-percent increase in the hydraulic conductivity resulted in a 0.26-ft decrease in the average arithmetic difference and 0.07-ft increase in the average absolute difference.

This sensitivity analysis indicated that the accuracy of the recharge rate was most important, followed by evapotranspiration extinction depth and aquifer hydraulic conductivity. The steady-state model was least sensitive to changes in the hydraulic conductivity of the riverbed.

The sensitivity analysis on the steady-state simulation reveals some conditions that appear to improve the agreement between simulated and observed water levels. However, the apparent improvement is, in fact, mainly a result of an improvement of the two wells near the region of low conductivity in the aquifer. The improvements that may occur in the remaining wells in the steady-state simulation do not reflect improvements in transient simulations. The conditions that appear to improve the arithmetic and absolute differences of the steady-state model also cause a decrease in the agreement between modeled and estimated baseflow in the Big Sioux River, and result in the upper reaches of the river changing from gaining to losing water to the aquifer in the simulation. Thus, the steady-state simulation that provided the best antecedent conditions for transient calibration was chosen.

Transient Simulation

Transient simulations allow changes in storage with time. Monthly water-level observations were available at as many as 80 locations during 1984 and 1985. Comparison of hydrographs of observed water levels and simulated water levels for the corresponding model cells in which the observation wells are located was a means of assessing the model's ability to simulate water-level changes in the aquifer. Analyses of these hydrographs provided a basis for adjusting the monthly recharge factors. Three selected hydrographs representing the simulated versus observed water levels are shown in figure 11. The simulated water-level trends match the observed trends of rising water levels in the spring, declines during the summer months, and rises in response to above-normal precipitation in October 1984 and September 1985.

The average of the monthly arithmetic difference between simulated and observed water levels was -0.05 ft, with a range of 0.85 ft in April 1984 to -0.65 ft in March 1985. The average arithmetic difference was within +0.5 to -0.5 ft on 14 of the 19 monthly simulations. The average absolute difference between simulated and observed water levels (table 4) was 1.72 ft. The average absolute difference ranged from 1.29 ft in October 1984 to 2.21 ft in August 1984. The deviations of the nearest-node simulated water levels for the July 1985 simulation from the observed water levels at 62 observation wells with observed water levels are shown in figure 12. The largest deviations occurred where observation wells were located in or adjacent to a cell simulated as a river cell. These differences are related to the grid size and could be minimized if a smaller grid were used to better discretize the river. A deviation also occurred near the previously discussed area along the east edge of T. 118 N., R. 53 W., where test holes indicated sandy till instead of aquifer material. The simulated water level in a model cell east of this area was 3.04 ft higher than the observed water level, and the simulated water level in a model cell west of the area was 5.58 ft lower than the observed water level. A possible explanation for this could be heterogeneity in the aquifer material. Gravel lenses may occur in the till that were not identified in the driller's logs.

The simulated monthly water budgets for 1984 and 1985 are shown in table 8. Recharge to the aquifer from precipitation occurred at an average rate of 63.0 ft^3/s, 13.5 ft^3/s greater than steady-state conditions. Evapotranspiration from the aquifer occurred at an average rate of 36.9 ft^3/s, 1.5 ft^3/s greater than steady-state conditions. Net discharge to the river was 4.6 ft^3/s more than steady-state conditions. The monthly transient water budgets for 1984 show that most of the water added to storage occurs during the spring months. The months of October 1984 and September 1985 had above-normal precipitation, which contributed to storage.

APPRAISAL OF THE BIG SIOUX AQUIFER USING THE DIGITAL MODEL

The model of the Big Sioux aquifer was used as a tool to evaluate the effects of various environmental stresses on the water table in Codington and Grant Counties. Stresses important to the hydrologic system include municipal, rural-water-system, and irrigation pumpage; changing river stage and lake levels; evapotranspiration by plant cover; and amount of rainfall. Once the model has been calibrated to a set of observed conditions assumed to represent steady state and used to adjust the monthly recharge factors under transient conditions, it may be used to estimate the response of the hydrologic system to a set of hypothetical stresses. In this study, the model was used to evaluate the effects on the water table of present-day public-supply pumping rates and maximum permitted irrigation pumping. The results of this simulation may be used to evaluate management practices and to aid in prudent utilization of the resources of the Big Sioux aquifer in Codington and Grant Counties.

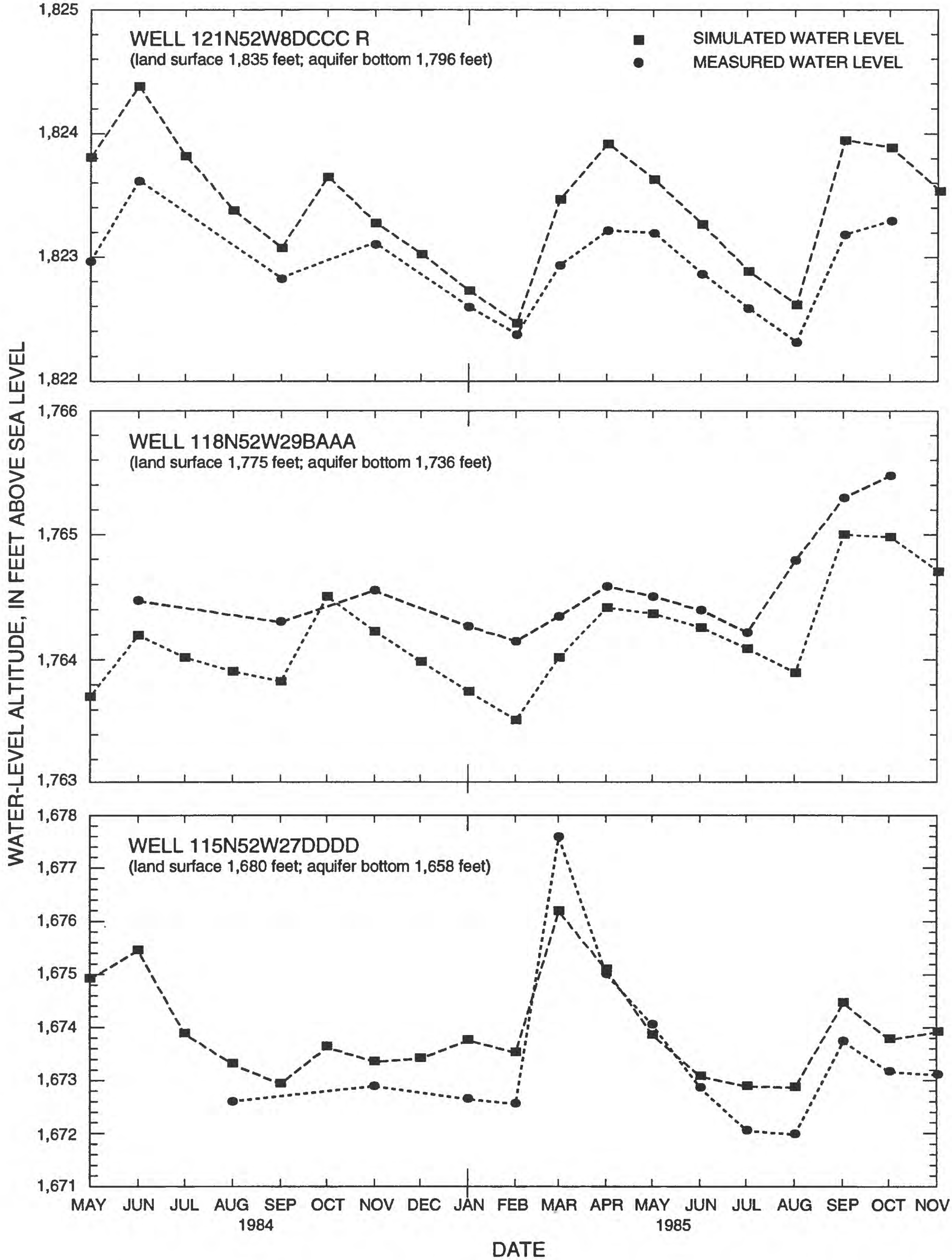
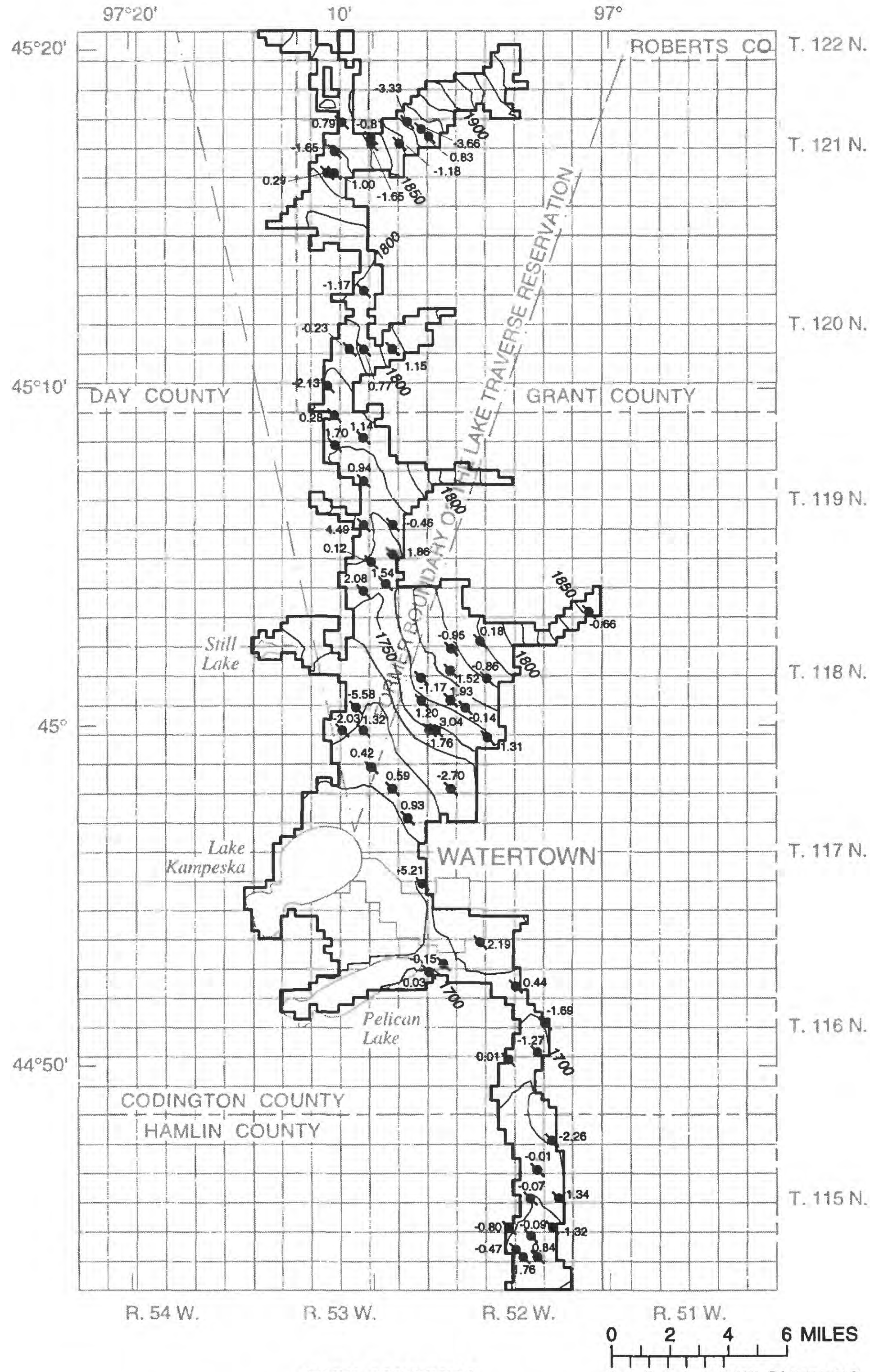


Figure 11. Simulated versus measured water levels during transient simulation.



EXPLANATION

—1800— SIMULATED WATER-TABLE CONTOUR--Shows altitude of simulated potentiometric surface. Contour interval, 10 feet. Datum is sea level.

1.47 OBSERVATION WELL--Number indicates the difference between simulated and observed water level, in feet. Positive number indicates simulated water level was higher than observed water level.

Figure 12. Water-table contours and differences between simulated and observed water levels at the end of July 1985.

Table 8. Simulated monthly water budgets, 1984-85

	January	February	March	April	May	June	July	August	September	October	November	December	Average
							1984						
							Recharge (inflow), in cubic feet per second						
Recharge from precipitation	0.0	0.0	80.3	181.7	87.8	143.9	25.9	37.7	42.2	180.2	0.0	0.0	65.0
Recharge from the river to the aquifer	7.9	11.9	33.8	13.2	16.1	41.7	10.2	11.9	10.3	6.8	7.5	6.4	14.8
Inflow at constant-head boundaries	4.2	4.1	3.6	3.1	3.3	3.0	3.4	3.8	3.9	3.7	3.5	3.7	3.6
Recharge from lakes	.1	3.8	32.1	47.9	13.0	18.1	6.7	4.9	3.5	6.2	5.5	.0	11.8
Total	12.2	19.8	149.8	245.9	120.2	206.7	46.2	58.3	59.9	196.9	16.5	10.1	95.2
							Discharge (outflow), in cubic feet per second						
Evapotranspiration from aquifer	0.0	0.0	0.0	67.6	80.6	91.6	85.7	60.6	45.4	28.0	0.0	0.0	38.3
Pumpage	1.0	3.0	2.6	2.2	2.0	.9	3.4	3.3	3.0	1.5	2.2	2.2	2.3
Discharge from the aquifer to the river	29.8	19.7	8.0	24.0	29.8	8.6	50.1	34.3	33.5	45.0	35.3	35.8	29.5
Outflow at constant-head boundaries	.9	.9	1.1	1.4	1.4	1.5	1.3	1.1	1.0	1.4	1.4	1.2	1.2
Discharge to lakes	3.4	1.5	.4	.0	.4	.6	1.8	16.9	15.5	3.1	2.2	66.2	9.3
Total	35.1	25.1	12.1	95.2	114.2	103.2	142.3	116.2	98.4	79.0	41.1	105.4	80.6
Change in storage increase (+), decrease (-)	-22.9	-5.3	137.7	150.7	6.0	103.5	-96.1	-57.9	-38.5	117.9	-24.6	-95.3	14.6

Table 8. Simulated monthly water budgets, 1984-85—Continued

	January	February	March	April	May	June	July	August	September	October	November	December	Average
							1985						
							Recharge (inflow), in cubic feet per second						
Recharge from precipitation	0.0	0.0	133.1	119.3	48.5	42.8	32.2	39.5	262.6	54.9	0.0	0.0	61.1
Recharge from the river to the aquifer	7.1	13.7	44.5	8.1	11.3	11.8	22.0	15.3	22.1	6.2	6.3	5.2	14.5
Inflow at constant-head boundaries	3.8	3.6	2.8	3.0	3.4	3.8	3.9	4.0	3.3	3.2	3.4	3.6	3.5
Recharge from lakes	.0	.0	26.8	6.0	.7	.8	1.1	.9	4.9	3.4	.1	6.0	4.2
Total	10.9	17.3	207.2	136.4	63.9	59.2	59.2	59.7	238.9	67.7	9.8	14.8	83.2
							Discharge (outflow), in cubic feet per second						
Evapotranspiration from aquifer	0.0	0.0	0.0	79.8	83.0	71.6	76.7	46.5	36.5	31.2	0.0	0.0	35.4
Pumpage	2.3	2.5	2.2	2.2	2.1	3.3	4.4	2.5	2.8	1.0	1.6	1.3	2.4
Discharge from the aquifer to the river	34.4	22.2	7.1	50.8	37.5	38.2	23.1	26.6	23.9	46.9	40.3	42.1	32.8
Outflow at constant-head boundaries	1.1	1.0	1.4	1.6	1.3	1.1	.9	.9	1.5	1.6	1.4	1.2	1.3
Discharge to lakes	7.8	6.1	7.7	4.2	29.1	16.8	8.5	6.7	6.4	6.6	8.0	6.3	9.5
Total	45.6	31.8	18.4	138.6	153.0	131.0	113.6	83.2	71.1	87.3	51.3	50.9	81.3
Change in storage increase (+), decrease (-)	-34.7	-14.5	188.8	-2.2	-89.1	-71.8	-54.4	-23.5	167.8	-19.6	-41.5	-36.1	1.9

A simulation using 1993 municipal and rural-water-system pumping rates and maximum permitted irrigation pumping rates under dry conditions was made to analyze the effects of increased aquifer depletions. Return flow to the aquifer from irrigation is assumed to be negligible during this simulation. The recharge rates and evapotranspiration rates were chosen based on the two consecutive dry years of 1980 and 1981. The 37 permitted irrigation wells had a combined appropriation of 18,330 acre-ft of water to be withdrawn during the 1993 growing season. For each irrigation well, the appropriation was proportioned monthly as follows: 10 percent in May and September, 25 percent in June and August, 30 percent in July. The stresses for the hypothetical increased-pumping and dry-condition simulation are shown in table 9. Although a previous simulation of water levels at the end of 1979 was available, these heads were not used as starting conditions because of the higher than normal rainfall experienced that year. Instead, the steady-state heads were used as starting conditions for the 2-year simulation of increased pumping and dry conditions. River stages for both years were based on gaging-station records for 1981. Average monthly lake levels for the simulation were based on the lowest lake levels on record, which occurred during 1976 when the Lake Kampeska levels ranged from an elevation of 1,716 to 1,714.5 ft.

During the simulation, many cells containing irrigation wells went dry. Two cells containing irrigation wells went dry during May 1980, which was the first month in which irrigation was simulated. During June 1980, 13 additional cells went dry, followed by six more in July 1980. By that time, most of the cells containing irrigation wells with high simulated pumping rates or locations in areas with a limited saturated thickness had already gone dry. One additional cell went dry in August 1981. By the end of the 2-year dry simulation, 22 of the 37 cells containing irrigation wells had gone dry.

Another simulation was run in which the irrigation rates were one-half the appropriated amount. In this simulation, 15 of the 37 cells still went dry. While the maximum permitted irrigation pumpage of 2 ft/acre (or even one-half that amount) is unlikely to be utilized, the large grid size of the model gives a conservative indication of when wells will go dry. These simulations do show, however, that the aquifer could probably not support extensive irrigation during dry periods such as those experienced in 1980 and 1981, or in 1976. The monthly simulated water budgets are listed in table 10.

Table 9. Assumed monthly hydrologic stresses for increased pumping and dry conditions

Period		Maximum evapotranspiration rate (inches)	Recharge rate (inches)	Pumping rate[1], all wells as of 1993 (cubic feet per second)
1980	January	0	0	1.5
	February	0	0	1.4
	March	0	.45	2.0
	April	4.2	.24	2.1
	May	5.5	.55	28.3
	June	6.4	1.69	35.4
	July	6.7	.11	31.1
	August	5.9	.24	22.8
	September	4.0	.15	11.2
	October	2.1	.26	2.3
	November	0	0	2.7
	December	0	0	2.5
1981	January	0	0	1.5
	February	0	0	1.4
	March	0	.72	2.0
	April	4.9	.56	2.1
	May	5.4	.30	11.2
	June	5.8	.54	22.8
	July	5.7	.30	31.1
	August	5.0	.06	21.4
	September	4.6	.09	11.2
	October	1.9	.50	2.3
	November	0	0	2.7
	December	0	0	2.5

[1]Pumping rates were adjusted as wells went dry during the simulation.

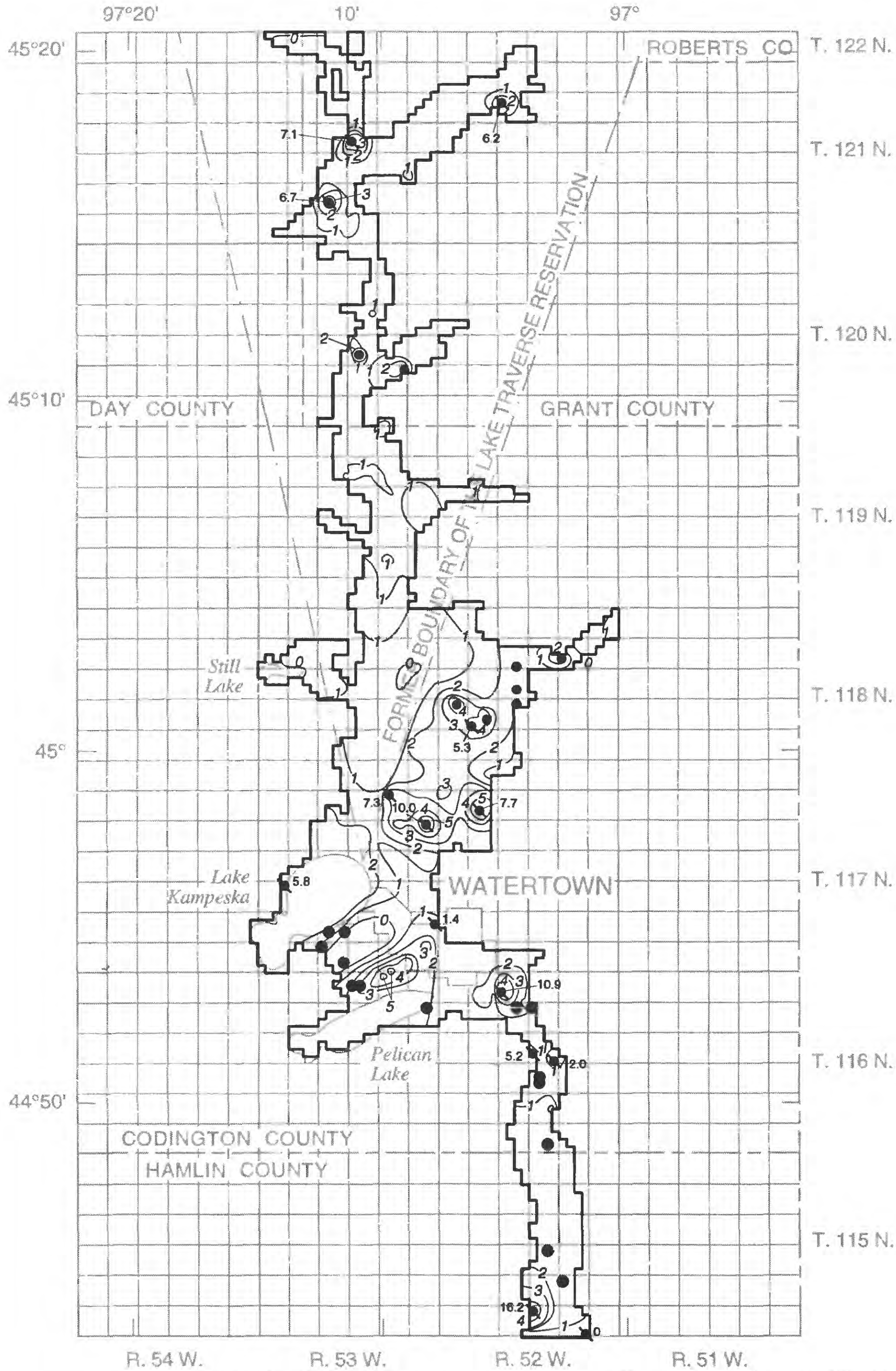
After monthly simulation of dry-condition stress, the average change in water level at the end of August 1981 was approximately 1.5 ft. Within the remaining active model cells, the maximum drawdown of 16.2 ft occurred in a cell that contained a simulated irrigation well. The simulated drawdown after two irrigation seasons of dry-condition stress and maximum permitted irrigation pumping during the growing season is shown in figure 13. The drawdown represents the average drawdown by model cell. Drawdown was largest in the model cells containing irrigation wells, 22 of which went dry during the simulation.

Table 10. Monthly simulated water budgets: hypothetical dry conditions with increased withdrawals

	January	February	March	April	May	June	July	August	September	October	November	December	Average
1980													
Recharge (inflow), in cubic feet per second													
Recharge from precipitation	0	0	48.6	25.8	59.4	182.5	12.2	26.2	16.2	27.8	0	0	33.2
Recharge from the river to the aquifer	8.0	11.6	5.6	12.4	15.7	17.8	22.6	22.3	18.5	12.1	9.4	7.9	13.7
Inflow at constant-head boundaries	3.9	3.8	3.9	4.2	4.4	4.3	5.2	5.5	5.5	5.2	5.0	5.0	4.7
Recharge from lakes	0	0	0	0.1	0.9	0.7	0.6	0.6	0.4	0.1	0	0.3	0.3
Total	11.9	15.4	58.1	42.5	80.4	205.3	40.6	54.6	40.6	45.2	14.4	13.2	51.9
Discharge (outflow), in cubic feet per second													
Evapotranspiration from aquifer	0	0	0	43.0	49.5	66.4	52.6	43.2	30.2	17.2	0	0	25.2
Pumpage	1.5	1.4	2.0	2.1	28.3	35.4	31.1	22.8	11.2	2.3	2.7	2.5	11.9
Discharge from the aquifer to the river	30.3	19.5	39.6	23.8	23.5	26.9	20.5	18.4	17.0	20.7	21.9	25.0	23.9
Outflow at constant-head boundaries	0.9	0.9	1.0	0.9	0.9	1.3	0.9	0.8	0.8	0.8	0.8	0.8	0.9
Discharge to lakes	6.8	5.5	5.4	4.2	3.4	5.7	4.7	3.9	3.8	4.2	4.2	4.0	4.7
Total	39.5	27.3	48.0	74.0	105.6	135.7	109.8	89.1	63.0	45.2	29.6	32.3	66.6
Change in storage increase (+), decrease (-)	-27.6	-11.9	10.1	-31.5	-25.2	69.6	-69.2	-34.5	-22.4	0.0	-15.2	-19.1	-14.7

Table 10. Monthly simulated water budgets: hypothetical dry conditions with increased withdrawals—Continued

	January	February	March	April	May	June	July	August	September	October	November	December	Average
1981													
Recharge (inflow), in cubic feet per second													
Recharge from precipitation	0	0	77.7	60.5	32.8	58.0	32.0	6.6	10.0	54.3	0	0	27.7
Recharge from the river to the aquifer	10.4	14.5	5.9	13.4	17.0	24.0	21.8	23.1	22.1	11.7	4.1	13.0	15.1
Inflow at constant-head boundaries	4.8	4.5	4.3	4.4	4.7	4.9	5.4	5.8	5.7	5.2	5.5	5.1	5.0
Recharge from lakes	0	0	0	0	0.2	0.4	0.5	0.6	0.4	0.1	0	0	0.2
Total	15.2	19.0	87.9	78.3	54.7	87.3	59.7	36.1	38.2	71.3	9.6	18.1	48.0
Discharge (outflow), in cubic feet per second													
Evapotranspiration from aquifer	0	0	0	45.4	43.0	45.6	39.6	32.7	29.8	14.4	0	0	20.9
Pumpage	1.5	1.4	2.0	2.1	11.2	22.8	31.1	21.4	10.6	2.3	2.7	2.5	9.3
Discharge from the aquifer to the river	18.6	12.7	36.6	22.8	19.9	15.1	17.8	14.7	13.1	20.1	40.5	13.8	20.5
Outflow at constant-head boundaries	0.8	0.8	1.0	0.9	0.8	0.8	0.7	0.7	0.7	0.8	0.8	0.8	0.8
Discharge to lakes	4.2	4.5	5.7	4.7	4.4	5.0	5.2	3.9	4.0	4.8	3.8	4.2	4.5
Total	25.1	19.4	45.3	75.9	79.3	89.3	94.4	73.4	58.2	42.4	47.8	21.3	56.0
Change in storage increase (+), decrease (-)	-9.9	-0.4	42.6	2.4	-24.6	-2.0	-34.7	-37.3	-20.0	28.9	38.2	-3.2	-8.0



EXPLANATION

— 4 — LINE OF EQUAL DRAWDOWN--Shows drawdown of simulated water table. Contour interval, in feet, is variable.

7.0 PUMPING IRRIGATION WELL--Number indicates the depth of drawdown cone at the model cell containing irrigation well, in feet

● IRRIGATION WELL IN MODEL CELL THAT HAS GONE DRY DURING SIMULATION

Figure 13. Simulated drawdown of the water table at the end of two irrigation seasons under dry conditions with maximum permitted irrigation pumpage.

In an effort to determine which stress factor had the greatest impact on the water budget, three additional scenarios were completed. These three scenarios were each compared to the main scenario, which simulated dry conditions and maximum permitted irrigation pumpage. In scenario 1, the recharge and evapotranspiration were under dry conditions while the lake and river stages and irrigation pumping were at average conditions. In scenario 2, river and lake stages were under dry conditions while recharge, evapotranspiration, and irrigation pumpage were at average conditions. In scenario 3, irrigation pumpage was at a maximum while recharge, evapotranspiration, and lake and river stages were at average conditions.

The decrease in storage in the aquifer during the main simulation occurred at an average rate of 14.6 ft^3/s during 1980 and 7.9 ft^3/s during 1981. In scenario 1, the decrease in storage occurred at an average rate of 10.1 ft^3/s during 1980 and 5.8 ft^3/s during 1981. Scenario 2 followed the same trend, with an average rate of decrease in storage of 13.9 ft^3/s during 1980 and 2.9 ft^3/s during 1981. Scenario 3 had the lowest average storage depletion rates, 7.5 ft^3/s during 1980 and 1.7 ft^3/s during 1981. Because of the complex interactions of the hydrologic system, the sum of the rates of loss in the three scenarios do not equal the rate of loss in the main hypothetical simulation. Scenario 2, with the lower lake and river stages, produces more of the storage depletion in the aquifer.

Water-loss rates due to evapotranspiration were 25.2 ft^3/s during 1980 and 20.9 ft^3/s during 1981 in the main simulation. The rates of loss in scenario 1 were 28.0 ft^3/s during 1980 and 23.8 ft^3/s during 1981. Scenario 2 produced loss rates of 21.8 ft^3/s during 1980 and 20.1 ft^3/s during 1981, while scenario 3 had losses of 29.6 ft^3/s during 1980 and 28.2 ft^3/s during 1981. Because evapotranspiration losses are greatest when the water is closest to the land surface, the effects of varying evapotranspiration rates are masked by the interactions of the other stress factors and their effects on the water table. It is possible to determine, however, that the lowest evapotranspiration losses occurred during scenario 2, when the water table was the lowest of the three scenarios.

Discharge from the aquifer to the river during the main simulation occurred at a rate of 10.3 ft^3/s during 1980 and 6.2 ft^3/s during 1981. Discharges to the river during scenario 1 averaged 11.5 ft^3/s during 1980 and 6.3 ft^3/s during 1981. Discharges in scenario 2 averaged 38.9 ft^3/s during 1980 and 29.7 ft^3/s during 1981, and in scenario 3 averaged 15.1 ft^3/s during 1980 and 12.2 ft^3/s during 1981. Water flowing from the aquifer to the lakes also was greatest in scenario 2, although flow rates for all three scenarios and for the main simulation were between 4.4 and 5.2 ft^3/s.

Low river stages cause the most rapid depletion of aquifer storage during simulation of drought conditions. The smaller storage depletions simulated for 1981 may indicate a system tending towards a new steady state that would occur under continued dry conditions.

As with any model, it is important to recognize that this model is a simplified representation of the Big Sioux aquifer. Some of the input parameters are difficult to measure in the field. Recharge, evapotranspiration extinction depth, and other parameters are therefore based on estimates. The model output should not be viewed as a prediction, but rather an estimated response of the system to certain stresses.

SUMMARY

This report is part of a series on the hydrology of the Big Sioux River Basin. The series includes county-by-county investigations of the water resources within the basin.

The Big Sioux aquifer in Codington and Grant Counties consists of poorly to well-sorted surficial glacial outwash ranging from medium sand to medium gravel. The unconfined aquifer is a 150-mi^2 area located primarily in the flood plain of the Big Sioux River in Codington and Grant Counties. The Big Sioux aquifer is hydraulically connected to the Big Sioux River, Lake Kampeska, Pelican Lake, and Still Lake. The aquifer material has an average thickness of 24 ft, with a range from 2 to 54 ft, and is underlain by glacial till.

A digital computer model was developed and calibrated under steady-state and transient conditions. A grid that contains 172 rows and 60 columns of equally spaced model cells, each 1,320 ft on a side, was used to simulate the aquifer. The hydraulic conductivities assigned to the model cells ranged from 50 to 500 ft/d and averaged 350 ft/d. A uniform specific yield of 0.14 and steady-state recharge of 5.53 in/yr were used. The riverbed hydraulic conductivity for the Big Sioux River was modeled at 0.05 to 1.0 ft/d, and the steady-state maximum potential evapotranspiration rate was 34.71 in/yr. The evapotranspiration rate decreased linearly to zero at the evapotranspiration extinction depth, which was 5 ft in most areas and 10 ft in cells located in areas likely to contain phreatophytes.

Average hydrologic conditions for the period 1978 through 1985 were used for steady-state simulation. The average absolute difference between simulated and observed water levels at 17 observation wells was 1.20 ft. Transient simulation was done on a monthly basis using hydrologic data from 1984 and 1985. The average arithmetic difference between simulated and observed water levels ranged from -0.65 ft in March 1985 to 0.85 ft in April 1984. The average arithmetic difference for 14 of the 19 months was between +0.5 and -0.5 ft. The average absolute difference between simulated and observed water levels ranged from 2.21 to 1.29 ft, and averaged 1.72 feet.

The steady-state recharge rate to the aquifer from precipitation was 49.5 ft^3/s. A net inflow of 2.7 ft^3/s recharges the aquifer from constant-head cells. Evapotranspiration to the atmosphere was 35.4 ft^3/s. Net outflow from the ground-water system to the river was 11.9 ft^3/s, and net outflow to lakes was 2.6 ft^3/s. Pumpage from the aquifer was 2.3 ft^3/s.

Sensitivity analysis of the steady-state model showed simulated water levels were most sensitive to recharge rate. A 20-percent reduction in the recharge rate of 5.53 in/yr resulted in a 0.52-ft decrease in the average difference between the observed and simulated water levels. A 20-percent increase in the recharge rate of 5.53 in/yr resulted in a 0.44-ft increase in the average difference. The effects of a similar change in the potential evapotranspiration rate produced less change in the simulated water levels. The steady-state simulation of changes in evapotranspiration extinction depth had a slightly greater effect than evapotranspiration rate. Hydraulic conductivity indicated less sensitivity than changes in recharge rate or evapotranspiration extinction depth, but a slightly greater sensitivity than changes in evapotranspiration rate or the riverbed hydraulic conductivity.

The calibrated model was used to study hypothetical scenarios of increased pumping from the aquifer during dry conditions. The municipal and rural-water-system pumping demands were updated to reflect 1993 usage. Irrigation pumpage was simulated as the maximum permitted usage for each well, distributed throughout the growing season. During the 2-year simulation of increased pumping and dry conditions similar to 1980-81, 22 of the 37 irrigation nodes went dry. The average drawdown in the aquifer at the end of August 1981 was about 1.5 ft; maximum drawdown was 16.2 ft. A second dry-condition simulation was run in which the irrigation pumpage was one-half the permitted amount. Under this condition, 15 of the 37 irrigation nodes went dry. These simulations indicate that the Big Sioux Aquifer probably is unable to support extensive irrigation during dry periods such as those that occurred during 1980 and 1981, or in 1976.

SELECTED REFERENCES

Amundson, F.D., Bradford, Wendell, and Koch, N.C., 1985, Drainage areas in the Big Sioux River basin in eastern South Dakota: U.S. Geological Survey Open-File Report 85-348, 1 sheet.

Barari, Assad, 1971, Hydrology of Lake Kampeska: South Dakota Geological Survey Report of Investigations No. 103, 84 p.

Ellis, M.J., and Adolphson, D.G., 1969, Basic hydrologic data, for a part of the Big Sioux drainage basin, eastern South Dakota: South Dakota Geological Survey and South Dakota Water Resources Commission, Water Resources Report 5, 124 p.

Farnsworth, R.K., and Thompson, E.S., 1982, Mean monthly, seasonal, and annual pan evaporation for the United States: National Oceanic and Atmospheric Administration Technical Report NWS 34, 82 p.

Farnsworth, R.K., Thompson, E.S., and Peck, E.L., 1982, Evaporation atlas for the contiguous 48 United States: National Oceanic and Atmospheric Administration Technical Report NWS 33, 26 p.

Flint, R.F., 1955, Pleistocene geology of eastern South Dakota: U.S. Geological Survey Professional Paper 262, 173 p.

Freeze, R.A., and Cherry J.A., 1979, Groundwater: Prentice-Hall, Inc., New Jersey, 604 p.

Hansen, D.S., 1988, Appraisal of the water resources of the Big Sioux aquifer, Moody County, South Dakota: U.S. Geological Survey Water-Resources Investigations Report 87-4057, 38 p.

———1990, Water resources of Codington and Grant Counties, South Dakota: U.S. Geological Survey Water-Resources Investigations Report 89-4147, 47 p.

———1994, Major aquifers in Codington and Grant Counties: South Dakota Geological Survey Information Pamphlet No. 47, 16 p.

Jones, O.R., and Schneider, A.D., 1969, Determining specific yield of the Ogallala aquifer by the neutron method: Water Resources Research, v. 5, no. 6, p. 1267-1272.

Jorgensen, D.G., and Ackroyd, E.A., 1973, Water resources of the Big Sioux River valley near Sioux Falls, South Dakota: U.S. Geological Survey Water-Supply Paper 2024, 50 p.

Koch, N.C., 1980, Appraisal of the water resources of the Big Sioux aquifer, Brookings, Deuel, and Hamlin Counties, South Dakota: U.S. Geological Survey Water-Resources Investigations Report 80-100, 46 p.

———1982, A digital-computer model of the Big Sioux aquifer in Minnehaha County, South Dakota: U.S. Geological Survey Water-Resources Investigations 82-4064, 49 p.

Kume, Jack, 1985, Water resources of Deuel and Hamlin Counties, South Dakota: U.S. Geological Survey Water-Resources Investigations Report 84-4069, 53 p.

Larimer, O.J., 1970, A proposed streamflow data program for South Dakota: U.S. Geological Survey Open-File Report 70-194, 46 p.

McDonald, M.G., and Harbaugh, A.W., 1988, A modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Techniques of Water-Resources Investigations, book 6, chap. A1, 600 p.

Meyer, W.L., 1962, Use of a neutron moisture probe to determine the storage coefficient of an unconfined aquifer: U.S. Geological Survey Professional Paper 450-E, p. E174-E176.

U.S. Department of Commerce, 1978-85, Climatological data for South Dakota (issued annually).

———1991, Climatological data for South Dakota.

U.S. Geological Survey, 1977-94, Water resources data for South Dakota, water years 1978-93: U.S. Geological Survey Water-Data Reports SD-78-1 to SD-93-1 (published annually).