A DIGITAL SIMULATION OF THE GLACIAL-AQUIFER SYSTEM IN SANBORN AND PARTS OF BEADLE, MINER, HANSON, DAVISON, AND JERAULD COUNTIES, SOUTH DAKOTA

By Patrick J. Emmons

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CONVERSION FACTORS

For readers who may prefer to use metric (International System) units rather than inch-pound units, the conversion factors for the terms in this report are listed below:

Multiply inch-pound unit	By	To obtain metric unit
acre	0.4047	hectare
acre-foot per day (acre-ft/d)	1,233	cubic meter per year
acre-foot per year (acre-ft/yr)	1,233	cubic meter per year
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
gallon per minute (gal/min)	0.06308	liter per second
inch	25.4	millimeter
inch per year (in/yr)	25.4	millimeter per year
mile (mi)	1.609	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 "Mean Sea Level of 1929."

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ABSTRACT

The drought in South Dakota from 1974-76 and the near-drought conditions in 1980-81 have resulted in increased demands on the ground-water resources within many of the irrigated areas of the James River basin in eastern South Dakota. These increases in demand for irrigation water from the glacial-aquifer system and continued requests to the State of South Dakota for additional irrigation well permits have created a need for a systematic water-management program to avoid overdevelopment of this system in the James River basin.

The aquifer system which can be unconfined at shallow depths and confined at greater depths in the same section has a thickness ranging from less than 10 feet to greater than 200 feet and an average hydraulic conductivity of 316 feet per day. Calculated storage coefficients of the aquifer system range from 0.00039 to 0.000017 and specific yield values are as great as 0.28. Calculated recharge rates to the unconfined aquifer range from 0.9 to 3.4 inches per year and to the confined aquifer, 0.24 to 0.72 inch per year. Evapotranspiration, which accounts for most of the natural discharge from the aquifer, was estimated to be as great as 36.2 inches per year in some locations.

An equally spaced grid containing 56 rows and 52 columns was used to simulate the glacial-aquifer system. The steady-state simulation was calibrated using water-level data collected before significant ground-water development (before 1973). The aquifer was also simulated in 11 annual transient stress periods from 1973 through 1983 and in 12 monthly transient stress periods for 1976.

The simulated predevelopment potentiometric heads were compared to average water levels from 32 observation wells to check the accuracy of the simulated potentiometric surface. The average arithmetic difference between the simulated and observed water levels was 1.68 feet and the average absolute difference was 4.38 feet. The nonpumping steady-state simulated water budget indicates that recharge from precipitation accounts for 97.1 percent of the water entering the aquifer and evapotranspiration accounts for 98.2 percent of the water leaving the aquifer. The sensitivity analysis of the steady-state model indicates that the model is most sensitive to reductions in recharge and least sensitive to changes in hydraulic conductivity.

In the annual transient simulation, recharge, evapotranspiration, and pumpage were adjusted annually. The maximum annual recharge varied from

0.10 inch in 1976 to 8.14 inches in 1977. The potential annual evapotranspiration varied from 29.9 inches in 1982 to 48.9 inches in 1976. Withdrawals from the glacial-aquifer system increased 2.6 times between 1975 and 1976. Since 1976, the pumpage has fluctuated annually in both distribution and quantity, however, the maximum annual withdrawals have not increased significantly since 1976. The average annual arithmetic difference between the simulated and observed water levels ranged from -3.88 feet in 1974 to 2.23 feet in 1982; the average absolute difference ranged from 4.70 feet in 1973 to 11.70 feet in 1982. The annual transient simulated water budget varies considerably as a result of changes in recharge and evapotranspiration.

In the 1976 monthly transient simulation, the maximum annual recharge rate of 0.10 inch was distributed over the months of March, April, and September. The potential monthly evapotranspiration rate ranged from 12.50 inches in August to 0.00 inch during the winter when the ground was frozen. The average arithmetic and absolute differences between the simulated and observed potentiometric heads for each of the 12 monthly simulation periods were calculated. The average arithmetic difference ranged from -1.25 feet in November to 2.68 feet in July. The average absolute difference ranged from 3.82 feet in October to 6.88 feet in July. The simulated monthly water budgets varied considerably as a result of changes in the monthly evapotranspiration, storage, and pumpage.

INTRODUCTION

The drought in South Dakota from 1974-76 and the near-drought conditions in 1980-81 have resulted in increased demands on the ground-water resources within many of the irrigated agricultural areas of the James River basin in eastern South Dakota. Between 1972 and 1980, the total quantity of ground-water irrigation from the glacial-aquifer system in the James River basin increased from 4,999 acre-ft/yr (South Dakota Water Resources Commission, 1973) to 35,422 acre-ft/yr (South Dakota Dept. of Water and Natural Resources, 1981), an increase of greater than 600 percent. These increases in demand for irrigation water from the glacial-aquifer system and continued requests to the State of South Dakota for additional irrigation well permits have created a need for a systematic water-management program to avoid overdevelopment of these aquifers in the James River basin.

In 1979, the South Dakota Department of Water and Natural Resources and Sanborn County entered into a cooperative agreement with the U.S. Geological Survey to define the flow system of the glacial-aquifer system in part of the James River basin (fig. 1). The study area has been divided into a northern part and a southern part. An appraisal of the northern part of the aquifer system in Spink and northern Beadle Counties has been completed (Kuiper, 1984).

The purpose of this study was to describe the flow system of the glacial aquifers in the southern part of the James River basin by using a digital flow model. More specifically, the study will better define the glacial aquifer boundaries; determine the aquifer thickness, direction of ground-water movement, and hydrologic properties of the glacial-aquifer system; and identify areas of ground-water recharge and discharge and determine rates of natural recharge and discharge. This report presents the results of the

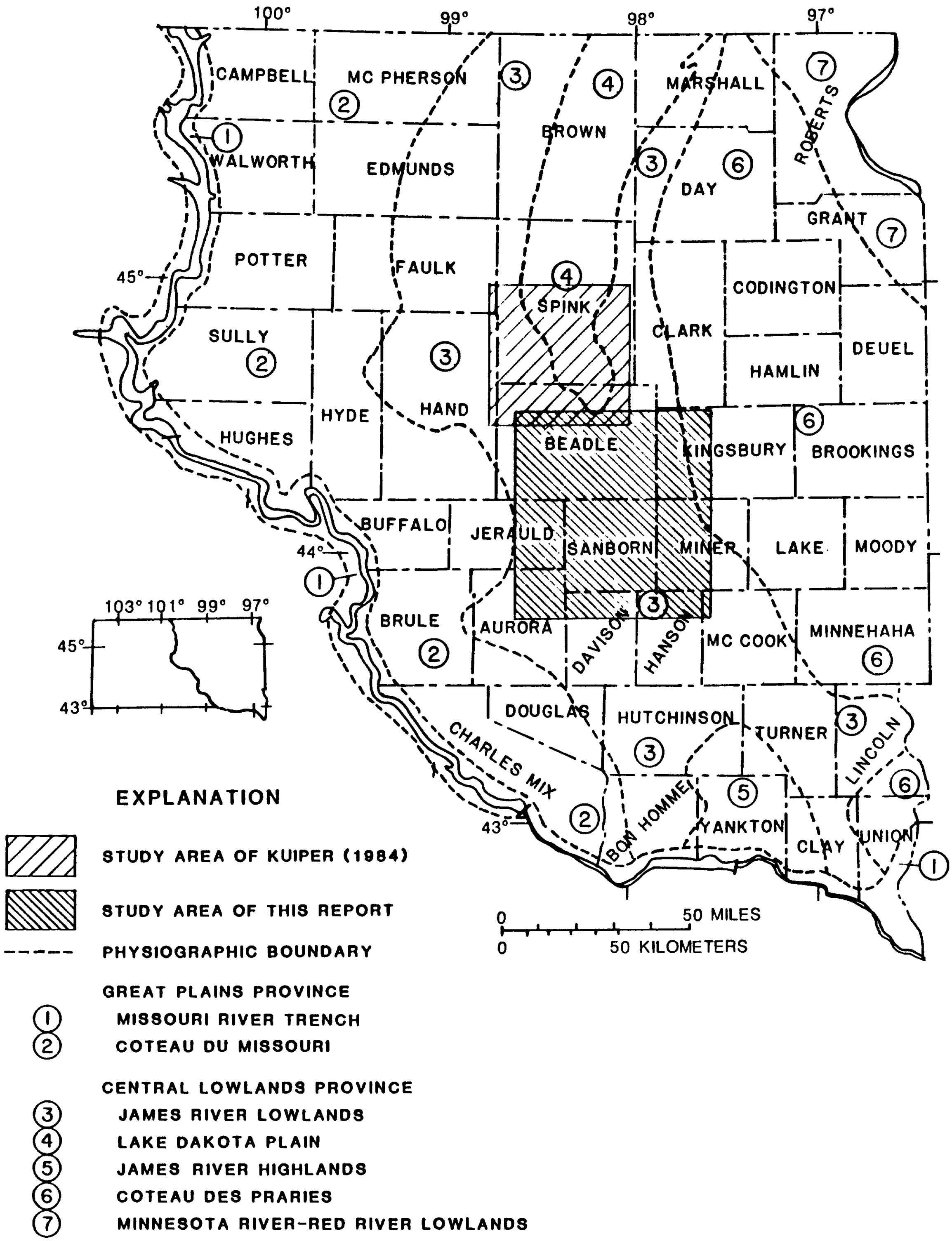


Figure 1.--Location of the study area and major physiographic divisions in eastern South Dakota.

investigation of the glacial aquifer in the southern part of the James River basin using a two-dimensional ground-flow model and describes the design and calibration of that model.

The scope of this investigation included the collation, examination, and synthesis of aquifer-test data, several thousand well and test hole logs, water-level measurements, pumpage data, and other miscellaneous geohydrologic data.

The aquifer-test data provided site-specific information on the aquifer's characteristics, such as hydraulic conductivity and storage coefficient. Koglin, Stack-Goodman, and Ambroson (1981) and Schroeder (1982) have compiled well and test-hole data for Beadle and Miner Counties, respectively. Well and test-hole data for Kingsbury, Mc Cook, Hanson, Davison, Aurora, Jerauld, and Sanborn Counties were obtained from the South Dakota Geological Survey, U.S. Geological Survey, U.S. Bureau of Reclamation, private drillers, and other miscellaneous sources. The well and test-hole data provided detailed information on the extent, thickness, and composition of the aquifers and confining beds. Water-level data which were obtained from the South Dakota Department of Water and Natural Resources provided historical water-level data and allowed for the determination of long-term water-level changes. The South Dakota Department of Water and Natural Resources also provided the pumpage data. This data was used to determine the magnitude of the stress being applied to the aquifer system as a result of pumpage.

Where existing data were inadequate, the South Dakota Geological Survey drilled five additional test wells. All these data were used to develop a digital flow model of the aquifer system. The aquifer system was simulated by using the U.S. Geological Survey's modular, three-dimensional, finite-difference, ground-water flow model program developed by McDonald and Harbaugh (1984).

Wells and test holes used in this report are numbered according to the Federal land-survey system of eastern South Dakota (fig. 2).

GEOLOGIC SETTING

During the Pleistocene Epoch, continental glaciation from the north and east covered eastern South Dakota, depositing a blanket of glacial drift over the eroded preglacial bedrock surface. Glaciation radically altered the topography by partially filling major valleys and entirely obliterating many small valleys, forcing the cutting of new valleys and forming massive end moraines. The overall effect of glaciation has been to reduce the local topographic relief. One of the greatest changes caused by the glaciers was the rearrangement of the surface drainage. Before glaciation, the main streams flowed toward the east. As a result of glaciation, the drainage in eastern South Dakota is now predominately southward (Flint, 1955).

The James basin is a lowland of low to moderate relief that trends north-south between the Coteau du Missouri and the Coteau des Prairies highlands, which are of glacial origin (fig. 1). The basin is 50 to 75 mi wide and approximately 250 mi long in South Dakota. The James River, which occupies the central axis of the basin, drains the basin to the south (Flint, 1955).

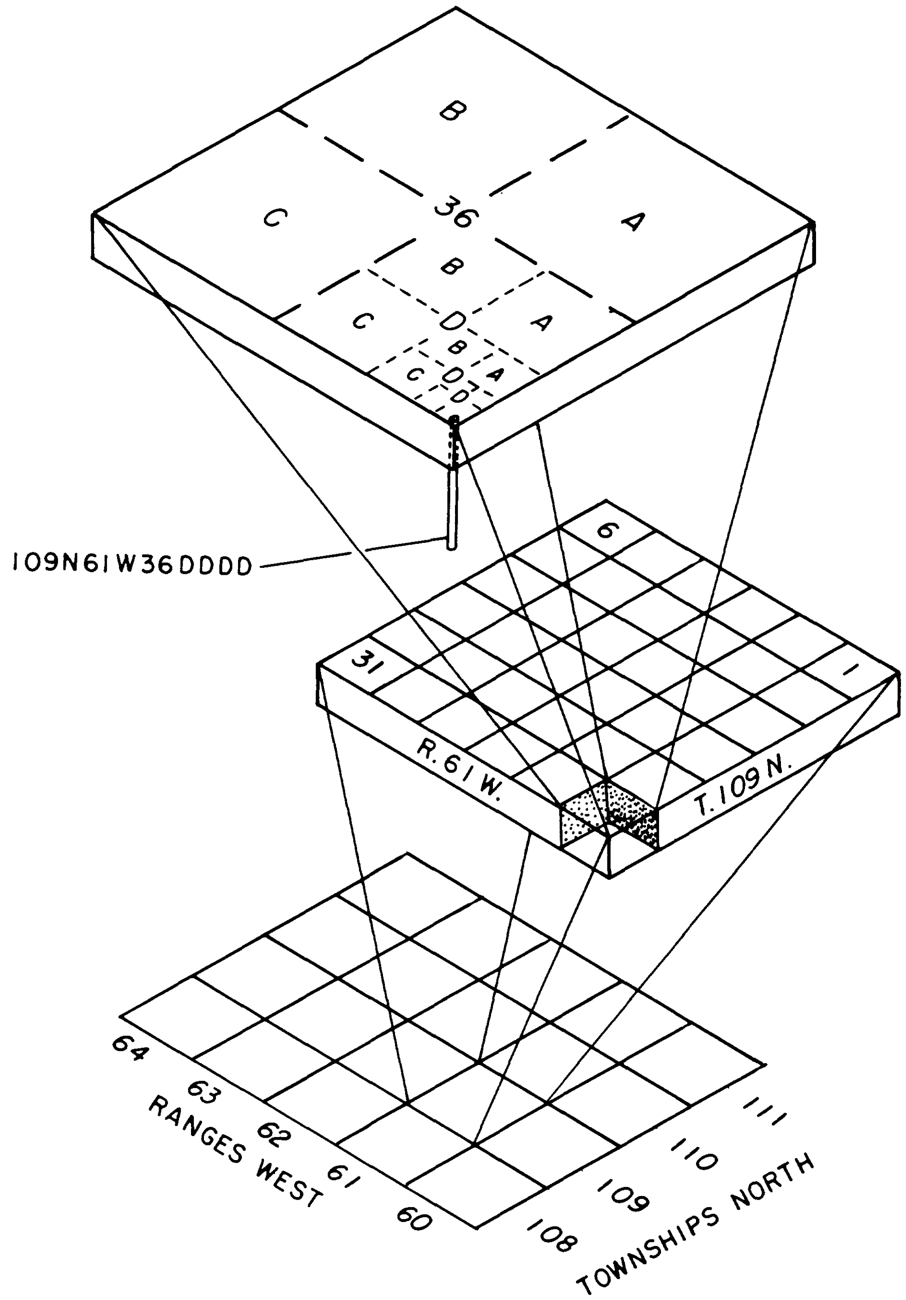


Figure 2.--Site-numbering system. The well number consists of township followed by "N," range followed by "W," and section number, followed by a maximum of four uppercase letters that indicate, respectively, the 160-, 40-, 10-, and $2\frac{1}{2}$-acre tract in which the well is located. These letters are assigned in a counterclockwise direction beginning with "A" in the northeast quarter. A serial number following the last letter is used to distinguish between wells in the same tract. Thus, well 109N61W36DDDD is the well recorded in the $SE\frac{1}{4}$ of the $SE\frac{1}{4}$ of the $SE\frac{1}{4}$ of the $SE\frac{1}{4}$ of section 36 in township 109 north and range 61 west of the 5th meridian and baseline system.

Most of the surficial deposits in the study area are the result of glaciation and are collectively called drift, which is any material deposited by or from a glacier. Drift can be subdivided into two major types, till and outwash, which differ greatly in both physical and hydrologic characteristics. Till, which was deposited directly from or by glacial ice, is a heterogeneous mixture of silt, sand, gravel, and boulders in a clay matrix. Outwash, which was deposited from or by meltwater streams on top of the ice or beyond the margin of the active glacial ice, consists primarily of layers of clayey or silty sand and sandy gravel, interbedded with layers of sandy or gravelly silt or clay. Beds of well-sorted sand and gravel are contained in the outwash but are generally small and discontinuous (Howells and Stephens, 1968).

The drift may be covered by deposits of alluvium along streams and rivers and locally, the drift may be covered by windblown sand and silt. The alluvium consists of poorly sorted, poorly stratified, thin, discontinuous layers of material that ranges in size from clay to boulders. Alluvium underlying the James River flood plain is as much as 25 ft thick and generally contains a much higher proportion of silt than does the alluvium elsewhere in the study area (Howells and Stephens, 1968).

The bedrock units directly underlying the drift in the study area in descending order are the Cretaceous Pierre Shale, Niobrara Formation, and Carlile Shale, and the Precambrian Sioux Quartzite (fig. 3). The Pierre Shale consists of a light to dark-gray fissile bentonitic clay-shale. Hedges (1968) reports that the Pierre Shale in Beadle County contains marly zones and chalky beds. Also, thin limestone beds, concretions, and bentonite stringers may be present. In the study area, the shale ranges from 0 to 600 ft in thickness.

The Niobrara Formation is predominantly a light- to dark-gray speckled marl with some chalk and shaly beds. The marl contains shells of Foraminifera (one-celled organisms) which give the marl a distinctive white speckled appearance. The formation ranges from 0 to 110 ft in thickness.

The Carlile Shale directly underlies the drift only in the southeast and a small area in the south-central part of the study area. The Carlile Shale consists mostly of light-gray to black shale containing silty and sandy zones. The thickness of the shale ranges from 0 to 312 ft. The Codell Sandstone Member is situated at or near the top of the Carlile Shale. A light-blue to black shale zone may separate the Codell Sandstone Member from the overlying Niobrara Formation. The Codell is a brown, fine- to medium-grained, moderately cemented sandstone. Thin shale layers in the Codell are common. The Codell Sandstone Member ranges from 0 to 120 ft in thickness.

The Sioux Quartzite underlies the drift only in the southeastern part of the study area. The Sioux Quartzite is a hard, massive, pink siliceous orthoquartzite which is horizontally bedded, cross-bedded, and jointed. Thickness of the quartzite in the study area is unknown (Hedges, 1968).

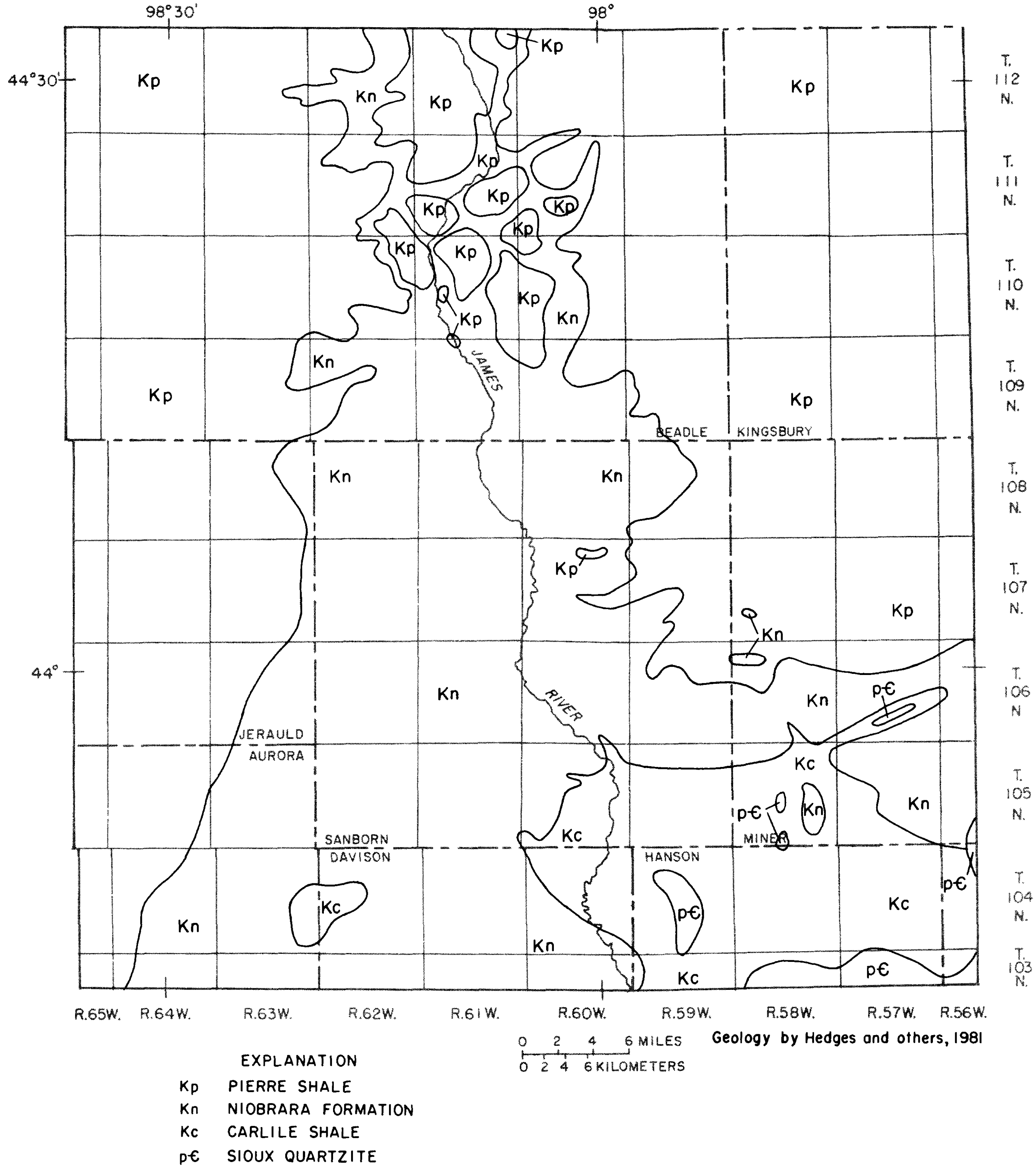


Figure 3.--Bedrock geology.

HYDROLOGIC SETTING

Ground water is a major source of water in the James River basin. In the unconsolidated surficial deposits, only the more sandy and gravelly glacial outwash deposits yield significant quantities of as much as 1,000 gal/min of water to wells. The remaining unconsolidated surficial deposits generally are either too clayey and silty or are too thin to serve as major sources of water except in very localized situations.

The recharge, movement, and discharge of water in the outwash aquifers are controlled by the lithology and stratigraphy of the surficial deposits and the underlying bedrock units. The till and the layers of silt and clay within the outwash deposits act to confine the outwash aquifers. The Niobrara Formation and the Codell Sandstone Member of the Carlile Shale may provide significant quantities of water to wells; however, these bedrock aquifers generally are isolated to some extent from the overlying outwash aquifers by till, clay and silt layers within the outwash deposits or shale units, or both. The Pierre Shale, Carlile Shale, and the Sioux Quartzite generally yield little or no water to wells and are considered to be confining beds.

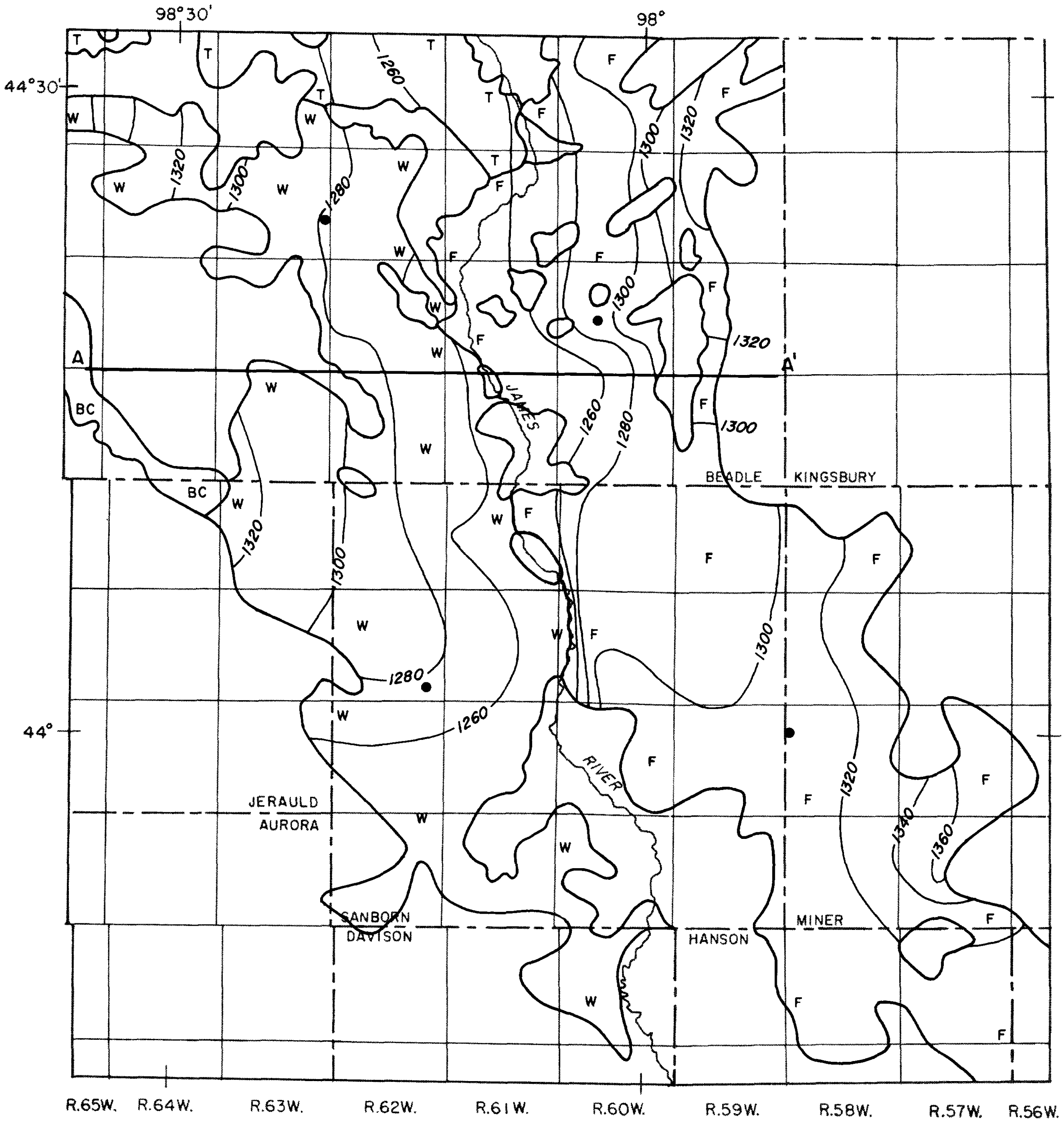
The complex hydrologic system which exists in the glacial outwash has been subdivided into four aquifers in the study area (fig. 4). They are the Floyd, Warren, Tulare, and Bad-Cheyenne aquifers. According to Hedges and others (1981), the aquifer boundaries are based on one or more of the following criteria:

a) A thinning or constriction of the aquifer.
b) A facies change from high to low permeability of the aquifer material.
c) A change from unconfined to confined conditions or vice versa.
d) A ground-water divide.
e) A ground-water discharge point such as a stream or lake.

The study area encompasses most of the Warren and Floyd aquifers and only a small part of the Tulare and Bad-Cheyenne aquifers. Most of the Tulare aquifer is located to the north and west in Spink and Hand Counties, respectively. The Bad-Cheyenne aquifer extends northwest into Hand and Hyde Counties.

The four glacial outwash aquifers generally are separated from each other by till confining beds and may be internally confined by till and thin clay and silt outwash layers (fig. 5). However, the till, and clay and silt outwash layers generally allow some flow to occur between and within aquifers.

Table 1 indicates that there is generally more variation of hydraulic conductivity within the aquifer than among the four aquifers. The hydraulic conductivity of the Warren aquifer ranges from 160 to 670 ft/d with an average of 410 ft/d. Hydraulic conductivity of the Floyd aquifer ranges from 37 to 589 ft/d with an average of 260 ft/d and hydraulic conductivity of the Tulare aquifer ranges from 20 to 1,430 ft/d with an average of 270 ft/d. There is no hydraulic conductivity data available for the Bad-Cheyenne aquifer.



EXPLANATION

F	FLOYD AQUIFER
W	WARREN AQUIFER
T	TULARE AQUIFER
BC	BAD-CHEYENNE AQUIFER
———	APPROXIMATE LIMITS OF AQUIFER
A———A'	LINE OF GEOHYDROLOGIC SECTION SHOWN ON FIGURE 5
—1300—	POTENTIOMETRIC CONTOUR--Shows altitude of potentiometric surface in May, 1980. Contour interval 20 feet. Datum is sea level
●	LOCATION OF HYDROGRAPHS SHOWN ON FIGURE 9

Figure 4.--Location and potentiometric surface of the aquifers in the glacial outwash.

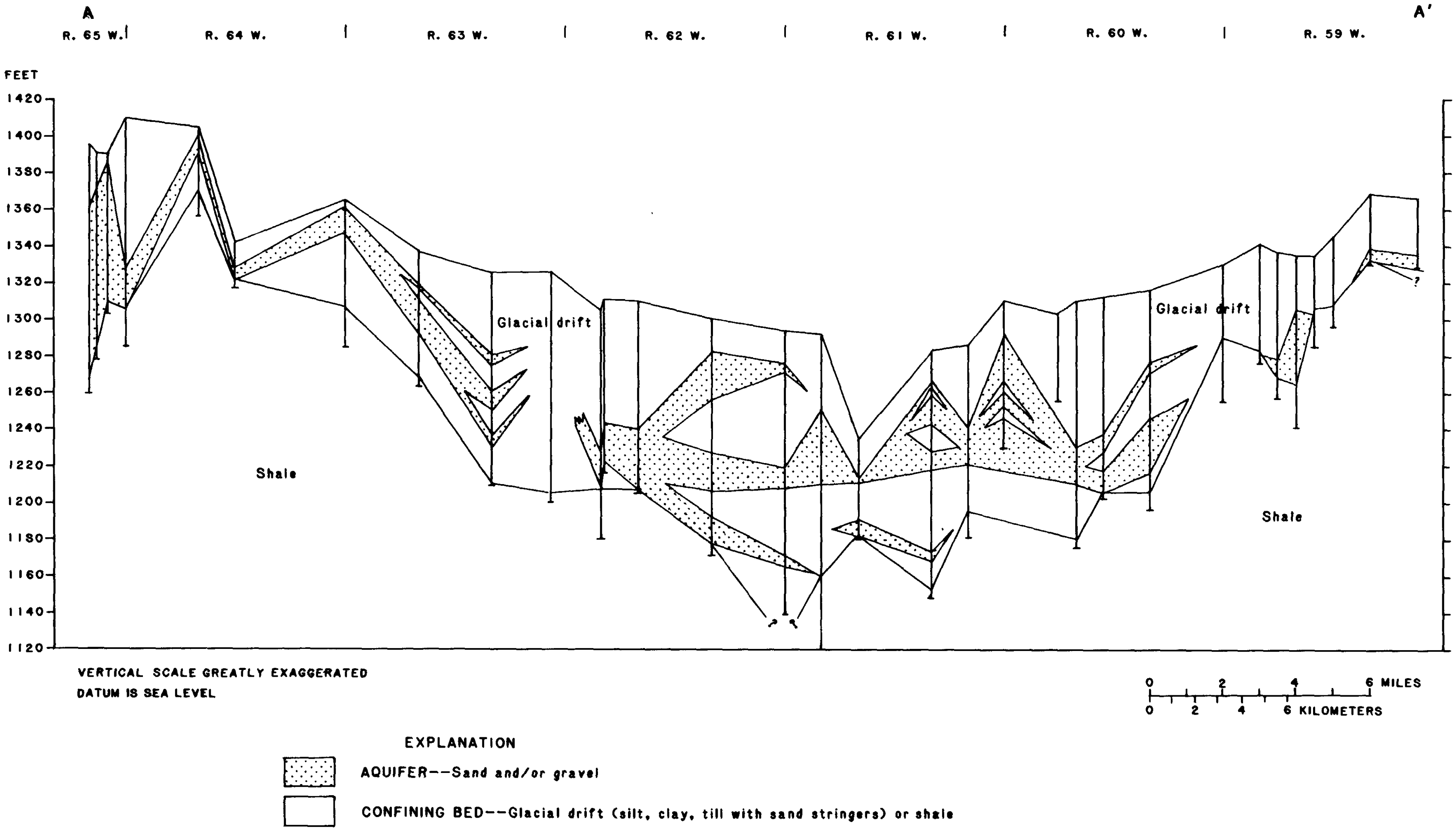


Figure 5.--Geohydrologic section showing the relation between glacial aquifers and confining beds.

Table 1.--Thickness, storage coefficient, and hydraulic conductivity of the glacial aquifers

Location	Aquifer	Aquifer thickness (feet)	Storage coefficient (dimensionless)	Hydraulic conductivity (feet per day)
		[1]Beadle County		
109N63W34ACAB	Warren	85	--	670
110N60W11BBD	Floyd	55	0.04	80
110N61W06ACDD2	Floyd	24	.00035	160
111N59W06BBBB2	Floyd	33	.00017	450
112N59W31CCCD1	Floyd	40	.00017	380
112N59W32DDDD2	Floyd	40	.00039	230
113N62W05DDBB1	Tulare	50	--	470
113N62W18BCAD1	Tulare	42	.00027	90
113N62W22ABBA2	Tulare	43	.00034	260
113N62W22ABBA4	Tulare	43	.00033	260
113N62W34CDCC2	Tulare	49	.00044	140
113N62W36DCDB1	Tulare	30	.00039	210
		[2]Hand County		
115N66218DCCC	Tulare	71	.0135	81
115N66220DABD	Tulare	30	.15	116
115N66W20DADB4	Tulare	30	.28	116
115N66W20DACA	Tulare	80	.14	95
115N66W20DABD3	Tulare	40	.00038	217
115N67W19CABB2	Tulare	53	.00016	139
115N67W19CABB3	Tulare	53	.00016	139
115N68W23BBAB	Tulare	19	.00052	20
		[3]Hanson County		
--	Floyd	4-87 (mean = 38, median = 32)	--	37-589 (mean = 255, median = 245)
		[4]Jerauld County		
108N63W20ABCC	Warren	70	.0001	402
		[1]Sanborn County		
108N61W17AACC	Warren	42	.000017	160
		[1]Spink County		
114N63W24CBAA1	Tulare	65	--	1,430
114N63W26ACAA1	Tulare	60	--	560

[1]From Howells and Stephens, 1968.
[2]From Koch, 1980a.
[3]From Hansen, 1983; Aquifer thickness and hydraulic conductivity were calculated from transmissivity and thickness of the Floyd aquifer at 31 test holes.
[4]From Hamilton, 1985.

Because all of the aquifers are in outwash deposits with similar hydraulic conductivities and are hydraulically connected by zones of material with lower hydraulic conductivity, the aquifers are treated as a single glacial-aquifer system rather than individual aquifers in this report.

A reliable delineation of the glacial-aquifer system is difficult to obtain due to the glacial processes that deposited the glacial outwash. The system is comprised of a series of connected and disconnected lenses, fingers, stringers, and channels of sand and gravel separated by layers of clay and silt outwash and till (fig. 5). The thickness of the sand and gravel layers in the aquifer system as well as other hydrologic characteristics vary greatly over short distances. For example, in the southeastern corner of T. 111 N., R. 63 W., Sec. 22, the aquifer consisted of 13 ft of gravel and 14 ft of sand. In section 23, approximately 0.25 mi east, the aquifer is composed of 63 ft of sand in one test hole and 30 ft of sand and 22 ft of gravel in another test hole. As a result of the extreme variations in thickness and composition, individual aquifer units often can be traced for only short distances or not at all. The composite glacial-aquifer system thickness ranges from less than 10 ft to greater than 200 ft (Howells and Stephens, 1968). The average thickness of the aquifer ranges from 4 to 144 ft and averages 56 ft in thickness (fig. 6). More than 1,000 drillers' logs were used to estimate the thickness. The aquifer top is defined as the uppermost occurrence of sand and gravel below the till where present. If no till is present, the aquifer top is land surface. The bottom of the aquifer is defined as the top of the bedrock or the bottom of the lowermost sand and gravel. The average aquifer thickness also includes all silt and clay layers in the aquifer zone. Lateral boundaries for the aquifer system were placed where the average sand and gravel thickness was less than about 5 ft.

The average thickness of the confining bed ranges from 4 to 170 ft and averages 49 ft (fig. 7). The thickness was estimated from the drillers' logs. The confining bed thickness was calculated as the thickness of all the clay and silt between land surface and the top of the aquifer. The confining bed controls, in part, the quantity of water which can recharge the aquifer and also the quantity of water available for evapotranspiration.

The hydraulic conductivity of the glacial-aquifer system, calculated from aquifer tests in the study area, ranges from 80 to 670 ft/d with an average of 316 ft/d (table 1). If the 31 hydraulic conductivity values estimated from test-hole data from Hanson County south of the study area are included, the average hydraulic conductivity decreases to 267 ft/d.

Water in the glacial-aquifer system occurs under unconfined water-table conditions and confined or artesian conditions. Due to the complexity of the aquifer system, an aquifer can be confined and unconfined in the same area. The value of the storage coefficient derived from an aquifer test is an indication of whether the aquifer is confined or unconfined in the vicinity of the test. The storage coefficient of most confined aquifers ranges from about 0.00001 to 0.001. The storage coefficient in an unconfined aquifer, often referred to as specific yield, generally ranges from 0.1 to 0.3. With one exception, the storage coefficients calculated from aquifer tests in the study area range from 0.00039 to 0.000017, indicating the aquifer system is artesian in these areas. One storage value, 0.04, indicates transitional conditions

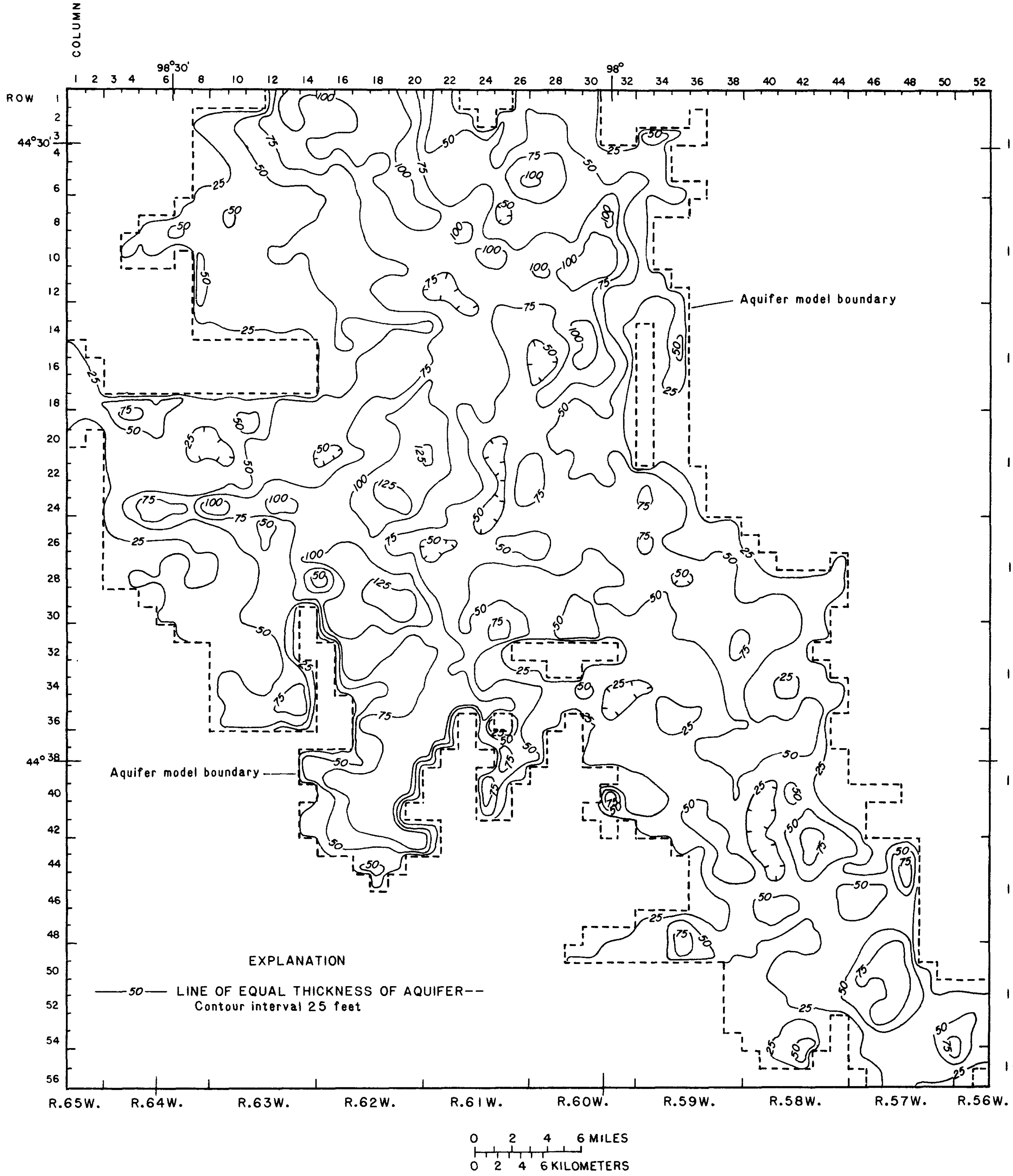


Figure 6.--Thickness of the glacial-aquifer system.

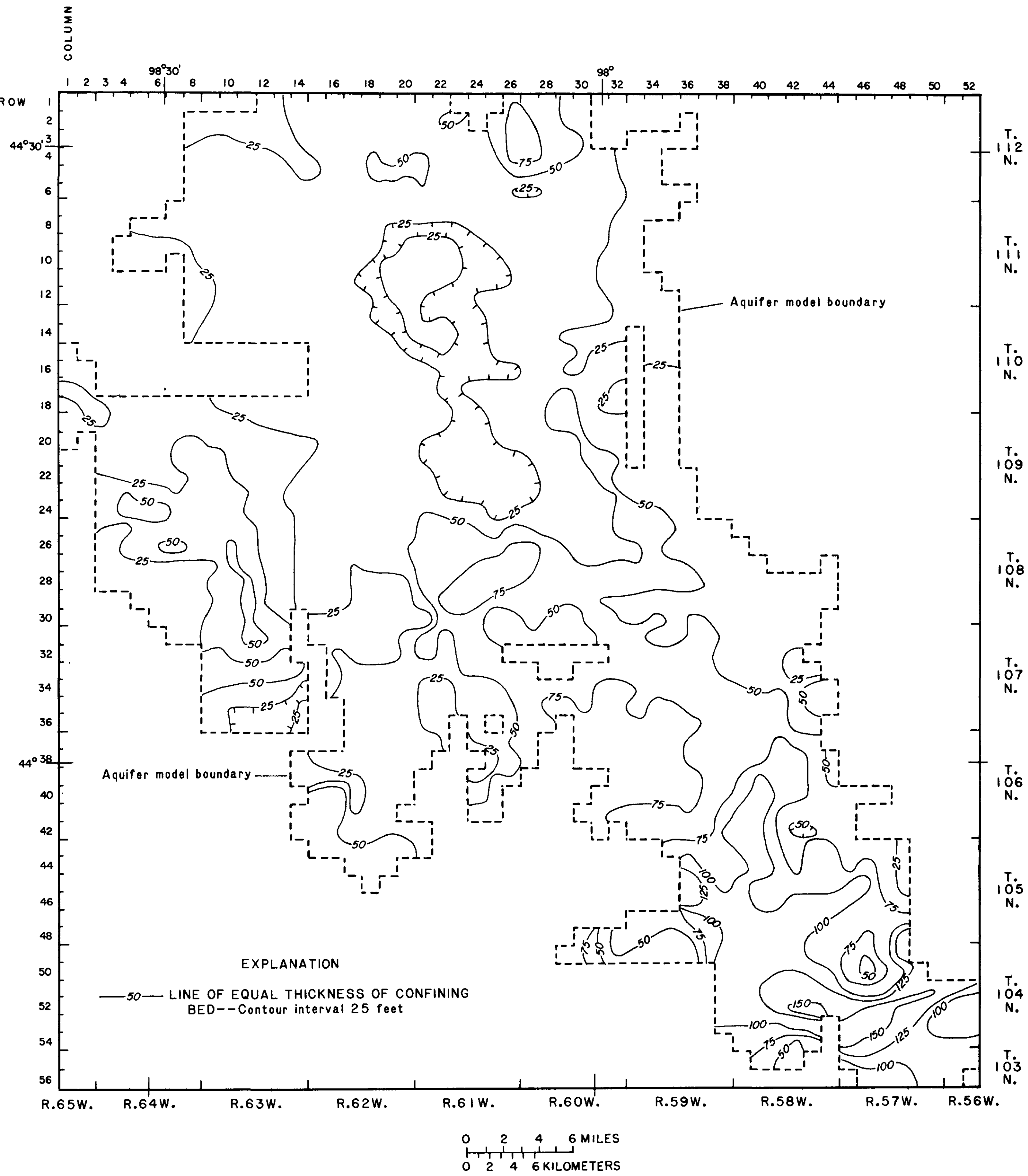


Figure 7.--Thickness of confining bed overlying the glacial-aquifer system.

between confined and unconfined. No specific yield values indicating unconfined conditions were calculated in the study area. Specific yield values as large as 0.28, however, were calculated from aquifer tests in Hand County located west of the study area.

Recharge to the glacial-aquifer system occurs as infiltration of precipitation and snowmelt directly into the aquifer or through the overlying confining bed. The thickness of the clay and silt in the overlying till and alluvium (fig. 7) controls the rate at which the underlying aquifer system can be recharged. Recharge occurs rapidly where there are permeable sediments overlying the aquifer. When the clay and silt are sufficiently thick (generally greater than 40 ft), there is probably little or no recharge by infiltration to the underlying aquifer. To a lesser extent, the aquifer in the study area also receives water as underflow from the west and as leakage from the lateral till boundaries.

The relationship between precipitation and recharge can be observed by comparing precipitation data and hydrographs for selected wells. For example, between 1968 and 1973, precipitation was generally at or above normal (fig. 8). Examination of the hydrographs (fig. 9) indicates a general water-level rise over the same period of time. Hedges and others (1983) calculated recharge rates to the Floyd and Warren aquifers from observation-well data and by flow net analysis. They also report results of recharge rates estimated from computer model analyses of the Tulare aquifer (table 2). The recharge rates to the unconfined parts of the glacial-aquifer system range from 0.9 to 3.4 in/yr and in the confined parts of the aquifer range from 0.24 to 0.72 in/yr.

Ground water flows downgradient perpendicular to the potentiometric contours as shown in figure 4. Although some long-term water-level declines have occurred as indicated by the hydrographs (fig. 9), the general direction of ground-water movement has remained the same. The direction of ground-water movement in the glacial-aquifer system in the study area is generally eastward or southeastward, west of the James River and westward or northwestward east of the James River. However, there is no or a very poor hydraulic connection between the aquifer system and the James River. Benson (1983, p. 47) states, "...there is probably no significant interchange between the James River and the underlying aquifers in Beadle County." According to Steece and Howells (1965), little natural surface discharge occurs from the glacial-aquifer system in Sanborn County. There is some discharge from the aquifer to Dry Run Creek and to the James River in Hanson and Davison Counties (Benson, 1983; Hansen, 1983); however, most of this discharge is south of the study area. A small amount of water may leak from the aquifer in the lateral low hydraulic conductivity till layers and into the underlying bedrock.

Evapotranspiration accounts for most of the natural discharge from the aquifer system. The potential evapotranspiration in the study area is estimated to be 72 to 74 percent of the pan evaporation or about 36.2 in/yr (Farnsworth and others, 1982). The potential evapotranspiration of an area can be estimated from pan-evaporation data (table 3). According to Farnsworth and Thompson (1982), the monthly estimated pan evaporation at Huron computed from meteorological measurements between January 1956 and December 1970 using a form of the Penman Equation are as follows:

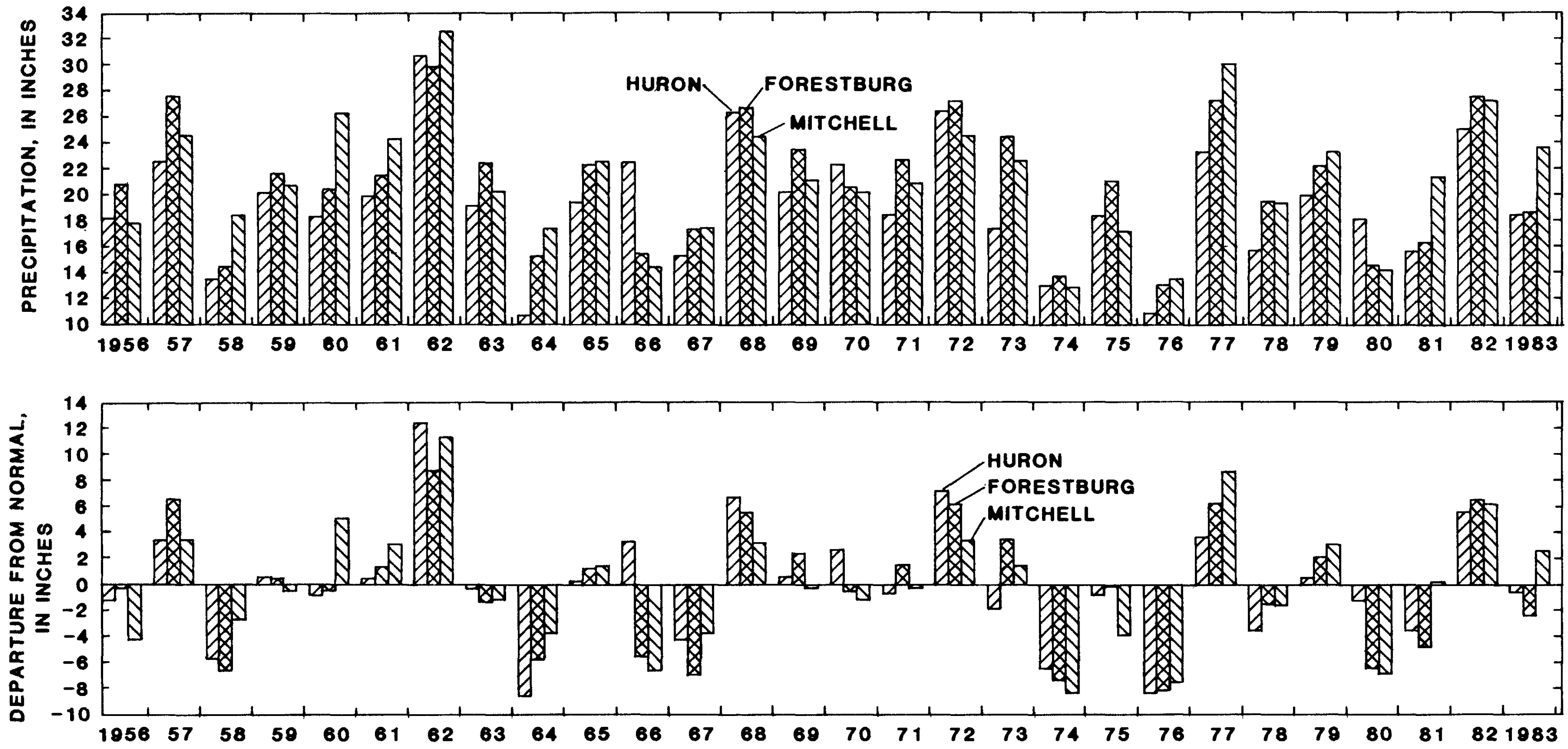


Figure 8.--Annual precipitation and departure from normal at Huron, Forestburg, and Mitchell, 1956-83.

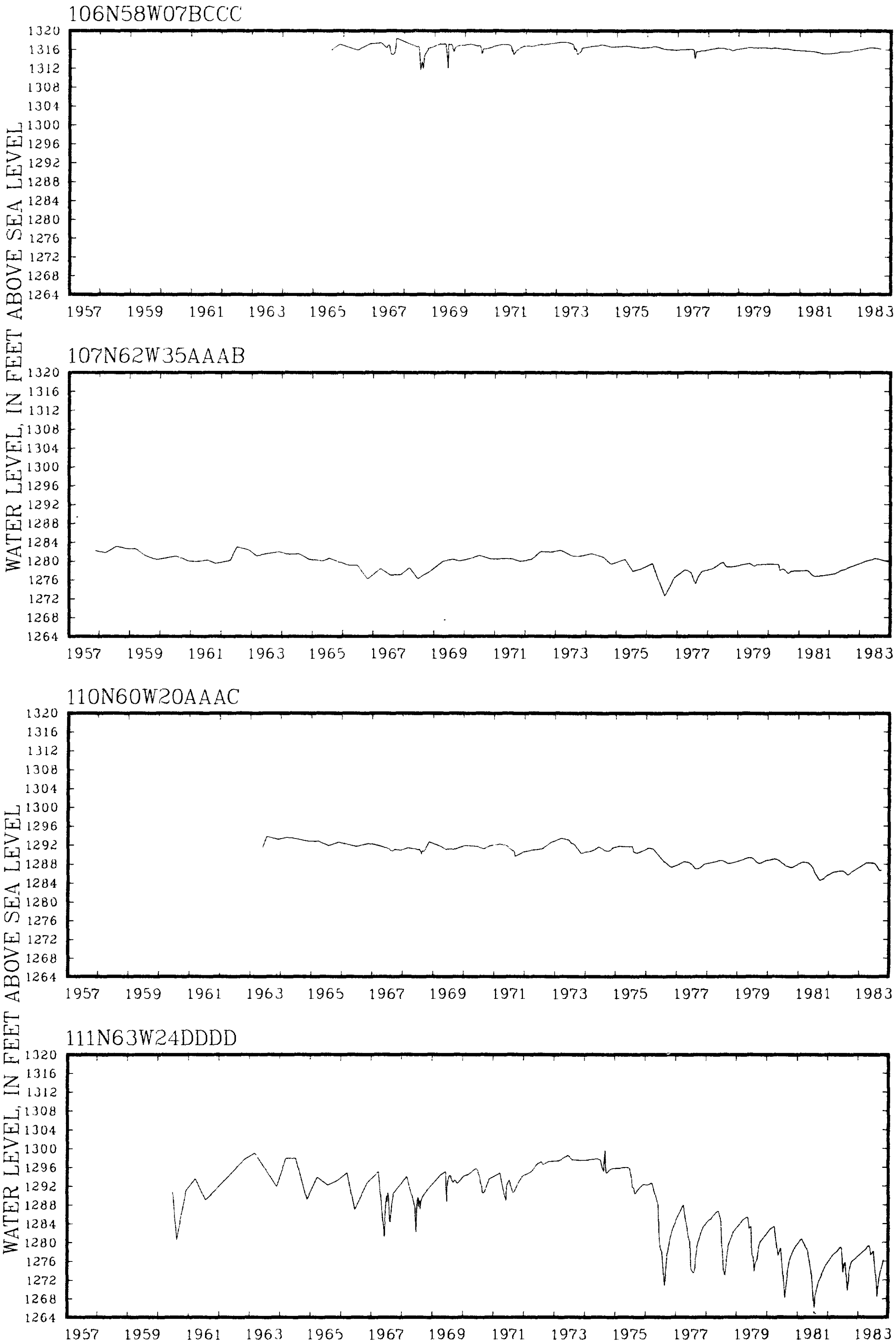


Figure 9.--Water-level changes in wells completed in the glacial-aquifer system.

Table 2.--Recharge to the glacial aquifers

[From: Hedges and others, 1983]

Aquifer	Units in inches per year: From computer model analysis	From observation well data	From flow net analysis
Floyd aquifer:			
East James			
Buried-unconfined (part)	--	3.4	--
Buried-confined	--	--	0.30
Pearl Creek:			
Buried-confined (all)	--	.9	--
Tulare aquifer:			
East James	0.38 to 1.52 (best fit) 0.76		
Buried-unconfined (part)	--	2.5	.72
Hitchcock	0.41 to 1.66 (best fit) 0.83		
Buried-unconfined (part)	--	2.0	.24
Western Spink	0.41 to 1.66 (best fit) 0.83	--	--
Hand			
Buried-unconfined (part)	--	3.4	--
Warren aquifer:			
Morris Creek			
Buried-confined	--	--	.35
Warren aquifer: (continued)			
West James			
Buried-unconfined (part)	--	3.0	--
Buried-confined	--	--	.35

Table 3.--Monthly pan evaporation and estimated potential evapotranspiration

Month	Mean monthly "pan evaporation" (inches)	Mean monthly potential evapotranspiration[1] (inches)
January	0.69	0.50
February	.83	.61
March	2.15	1.57
April	4.45	3.25
May	6.26	4.57
June	7.68	5.61
July	8.89	6.49
August	7.68	5.61
September	4.96	3.62
October	3.52	2.57
November	1.60	1.17
December	.84	.61
Annual	49.55	36.18

[1]The mean monthly pan evaporation multiplied by 0.73.

The estimated potential evapotranspiration is from wet soil or other moist natural surfaces. The potential evapotranspiration from the glacial-aquifer system is reduced by the depth to water in the aquifer and by the confining bed overlying the aquifer. As the depth to water in the aquifer increases, the amount of water for evapotranspiration decreases. There is probably very little water removed by evapotranspiration where the aquifer potentiometric surface is greater than 10 to 15 ft below land surface. A till confining bed overlying the aquifer can decrease the quantity of water available for potential evapotranspiration from the aquifer even though the potentiometric surface may be less than 15 ft below land surface.

Ground water is used for irrigation, municipal, industrial, farm, ranch, and domestic use. However, most of the water withdrawn from the glacial-aquifer system is used for irrigation. Withdrawals other than irrigation generally have little effect on the aquifer system. Permitted ground-water withdrawal rates in the study area have increased from 136 ft^3/s in 1973 to 499 ft^3/s in 1981, a 267 percent increase in 8 years (table 4). The actual ground-water withdrawal rate for irrigation has increased from 2.66 to 12.45 ft^3/s. The effects of the continued increase in the withdrawal of ground water are shown on the hydrographs (fig. 9). The hydrographs indicate that possible long-term water-level declines are beginning to occur in some areas as the result of the continuing increase in ground-water withdrawals.

Table 4.--Ground-water use in the James River basin and the study area

Year	Projected acreage that was irrigated with ground water in the James River basin[1] (acres)	Average depth of ground-water applied in the James River basin[1] (inches)	Permitted ground water acreage in the study area (acres)	Permitted ground-water withdrawal in the study area (cubic feet per second)	Ground-water withdrawal in the study area (cubic feet per second)
1970	14,463.87	12.5	--	--	--
1971	--	--	--	--	--
1972	9,181.32	8.3	--	--	--
1973	13,479.39	13.2	12,287.6	135.64	2.66
1974	13,661.64	13.0	9,997.1	109.96	3.37
1975	16,793.85	12.7	10,222.2	119.02	5.33
1976	34,459.84	16.0	15,819.8	190.60	14.10
1977	49,237.02	12.17	25,562.8	285.92	15.64
1978	53,044.99	10.15	28,304.2	311.48	11.73
1979	60,045.60	7.76	36,042.2	402.54	8.47
1980	69,240.91	10.51	39,559.9	445.65	9.79
1981	71,370.36	12.52	43,767.9	499.21	16.52
1982	74,991.95	8.78	Not calcualted	Not calculated	[2]9.52
1983	60,569.11	9.18	Not calculated	Not calculated	12.45

[1]From South Dakota Department of Water and Natural Resources irrigation questionnaire information (written commun., 1970-83).
[2]Withdrawal data not available for 1982. Withdrawal was estimated using 1983 pumpage and comparing precipitation data for 1981-83.

SIMULATION OF FLOW IN THE GROUND-WATER SYSTEM

Simplifying Assumptions

Ground-water flow within an aquifer system is governed by a complex series of interrelated hydrologic processes. A number of simplifying assumptions make it possible to describe these hydrologic processes and allow the aquifer system to be represented mathematically. The simplifying assumptions may not exactly represent the hydrologic processes, but should include the basic assumptions and logic governing these processes.

The simplifying assumptions for simulation of the glacial-aquifer system are:

1. The aquifer consists of one layer. The top of the aquifer is defined as the first occurrence of sand or gravel below the till confining bed or land surface if the confining bed is not present. The bottom of the aquifer is defined as the bottom of the lowest sand or gravel layer or the top of the bedrock. The aquifer includes all of the deposits between these vertical limits.

2. The overlying confining bed allows recharge to infiltrate downward to the aquifer and ground water to migrate upward through the till when the confining bed is less than about 45 ft thick.

3. The clay and silt or bedrock below the aquifer is an impermeable lower boundary of the aquifer system.

4. All lateral boundaries of the aquifer system are impermeable or no-flow boundaries. At 12 internal locations, the potentiometric heads are held constant or are specified-head boundaries.

5. The James River is hydraulically isolated from the aquifer system by the confining bed except at 2 river-head boundary locations (fig. 10). At these 2 locations, water discharges from the aquifer to the river.

6. All flow in the aquifer is horizontal and in the overlying confining bed, vertical. Storage occurs only in the aquifer. The confining bed yields no water to wells.

7. The principal source of recharge to the aquifer system is precipitation. The thickness of the overlying confining bed (fig. 7) controls the rate at which the aquifer system can be recharged. The greater the confining bed thickness, the lower the recharge rate.

8. Discharge from the aquifer occurs as evapotranspiration, pumpage, and at the specified-head boundary. The primary method of discharge is evapotranspiration. Upward leakage of water from the aquifer to the zone where evapotranspiration can occur is controlled by the thickness of the overlying confining bed. The greater the thickness of the confining bed, the lower the rate at which evapotranspiration can occur.

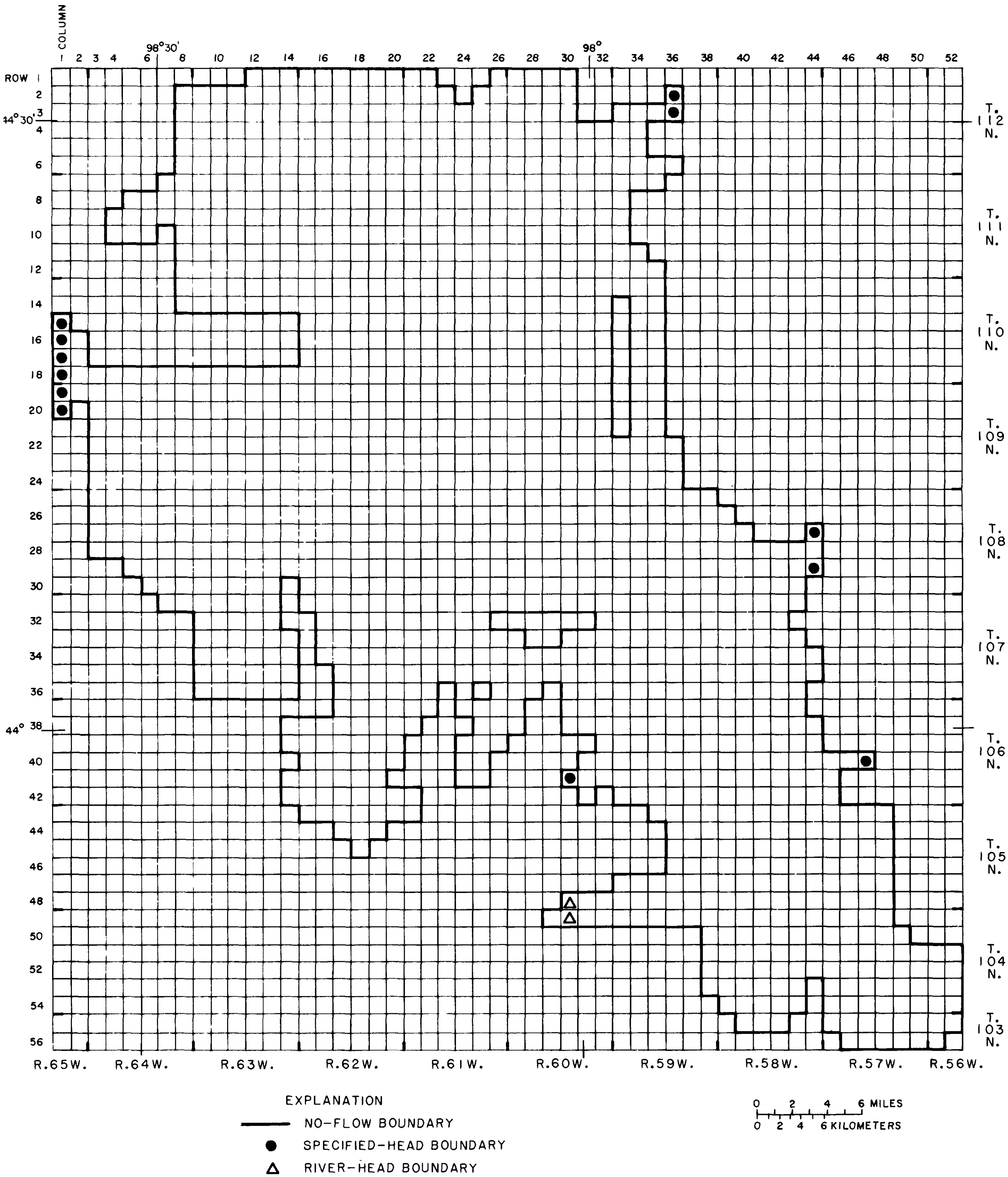


Figure 10.--Finite-difference grid blocks and model boundaries used to define the glacial-aquifer system.

The Digital Model

A mathematical model of an aquifer system is the application of mathematical equations describing ground-water flow and certain simplifying assumptions to a concept of the flow system. 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 of ground-water flow. The digital model used in this study is the U.S. Geological Survey modular three-dimensional finite-difference ground-water flow model of McDonald and Harbaugh (1984).

The model uses finite-difference methods to obtain approximate solutions to partial-differential equations of ground-water flow. The modeling area was subdivided into a series of finite-difference grid blocks in which the aquifer properties are assumed to be constant (fig. 10). The continuous derivatives of the partial differential equation of ground-water flow are replaced by the finite-difference approximations at the center (node) of each of the grid blocks. The result is a series of finite-difference equations that were solved with the slice-successive overrelaxation (SSOR) numerical technique.

Model Data

A ground-water flow model is constructed by entering a value for the hydrologic components that define the system at each finite-difference node. The value assigned to the node is considered to be representative of the entire grid block. The following is a list of components used in the model of the glacial-aquifer system:

1. Dimensions of the finite-difference grid.
2. Altitude of the top of the aquifer.
3. Altitude of the bottom of the aquifer.
4. Hydraulic conductivity of the aquifer.
5. Storage of the aquifer.
6. Recharge to the aquifer.
7. Evapotranspiration from the aquifer.
8. Pumpage from the aquifer.
9. River leakage from the aquifer.

Dimensions of the Finite-Difference Grid

A finite-difference grid is required so the geohydrologic data can be put in a form to be entered and manipulated by the computer-model program. The equally spaced finite-difference grid selected to represent the model area has 56 rows and 52 columns. The grid blocks are 1 mi or 5,280 ft on a side. Each grid block, shown in figure 10, overlies a 640-acre section.

Altitude of the Top of the Aquifer

The altitude of the top of the aquifer (fig. 11) is the top of the first sand or gravel layer below the overlying till confining bed. Where the till is not present, the altitude of the aquifer top is land surface.

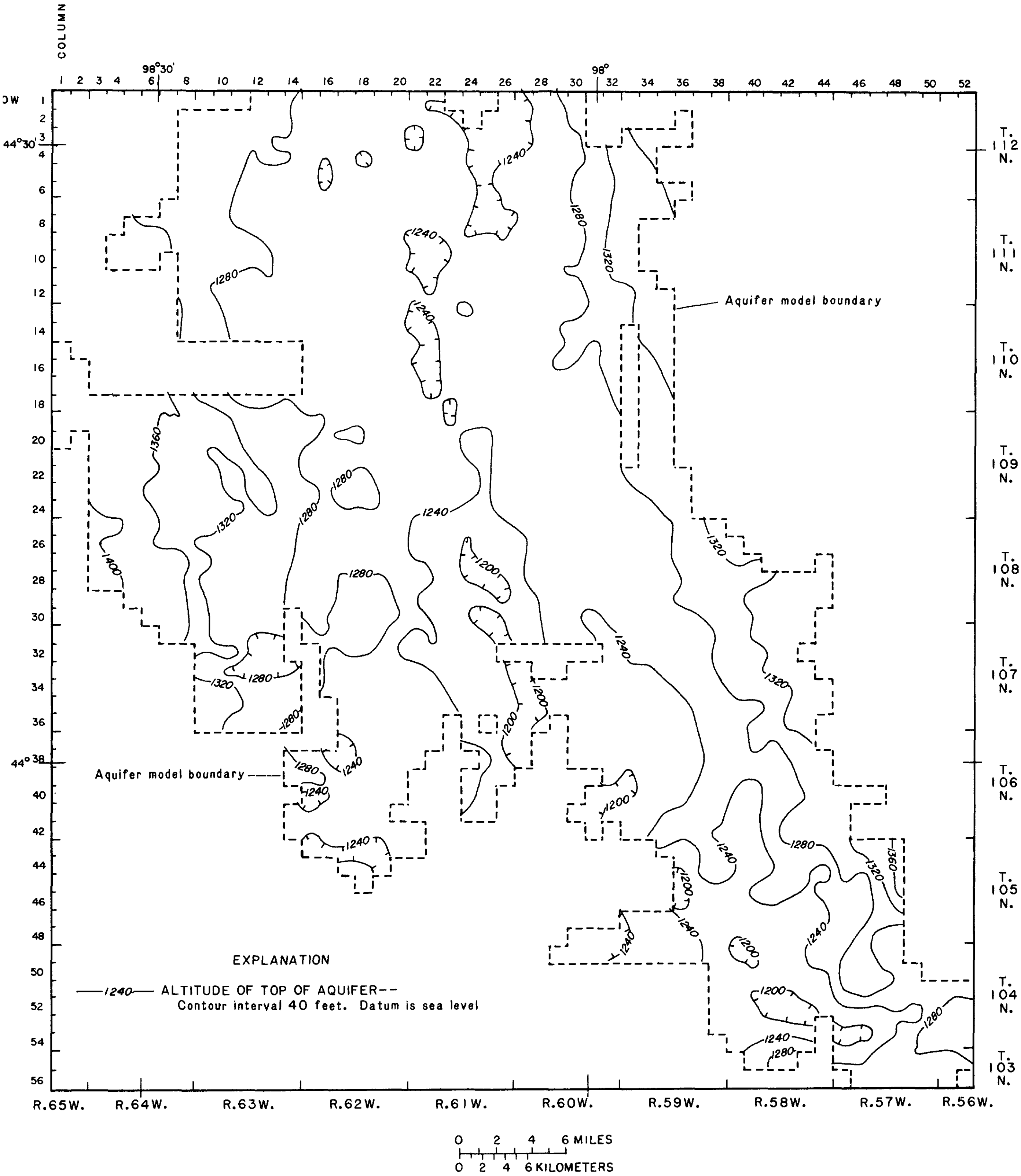


Figure 11.--Altitude of the top of the glacial-aquifer system.

Altitude of the Bottom of the Aquifer

The altitude of the bottom of the aquifer is the bottom of the lowermost sand and gravel layer (fig. 12).

Hydraulic Conductivity of the Aquifer

The hydraulic conductivity, a measure of the ability of the aquifer to transmit water, varies greatly over short distances due to the variability of the glacial deposits. All of the test-hole and drillers' logs were examined and an average composite log for each section was developed. Using the average composite logs for each section of land and the hydraulic conductivity values (table 5), an average composite aquifer hydraulic conductivity was calculated for each section of land (fig. 13). The relationship between hydraulic conductivity and grain size was varied to achieve the best overall model results. The assignment of the average composite hydraulic conductivity for each section is based on the assumption that aquifer materials are uniformly variable and the test-hole and drillers' logs adequately depict the range of the types and thicknesses of aquifer materials in each section.

The hydraulic conductivity of the aquifer system is generally much less than assigned by Koch (1980b) to the alluvial-mantled outwash deposits of the Big Sioux aquifer, located east of the James River basin in South Dakota (table 5). The hydraulic conductivities of the outwash deposits in the glacial-aquifer system are less because they contain much more silt and clay.

The average composite hydraulic conductivities for each section of land range from 11 to 320 ft/d (fig. 13). These average composite hydraulic conductivities are less than the hydraulic conductivities calculated from aquifer tests (table 1). This is expected as the aquifer tests are site specific and generally are conducted in areas where the aquifer has greater hydraulic conductivity and thickness.

As a result of the averaging process for hydraulic conductivity, the ground-water-flow model will approximate the glacial-aquifer system on a regional scale, but locally, deviations may occur.

Storage in the Aquifer

With one exception, storage coefficients calculated from aquifer tests in the study area range from 0.00039 to 0.000017 (table 1), indicating artesian conditions exist at these locations in the glacial-aquifer system. The exception, a storage value of 0.04, most likely indicates a transition between confined and unconfined conditions. Specific yield values as high as 0.28 were calculated from aquifer tests in the glacial-aquifer system west of the study area.

A storage coefficient of 0.0003 was used in the ground-water flow model to represent the glacial-aquifer system in a grid block where the average potentiometric head was higher than the average altitude of the top of the aquifer (artesian conditions). A specific yield of 0.15 was assigned when the average potentiometric head in the section was lower than the average altitude of the top of the aquifer for the same section (water-table conditions).

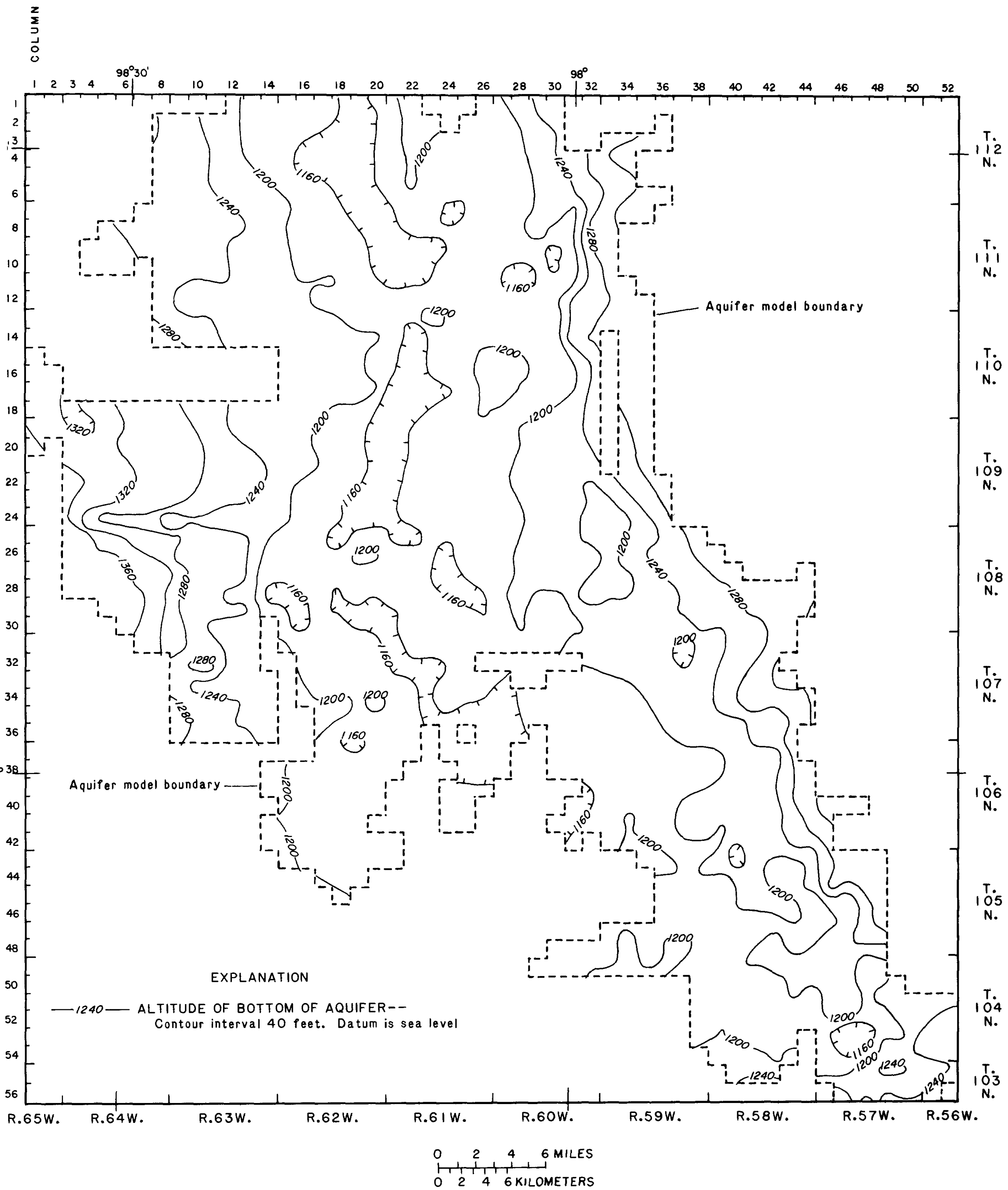
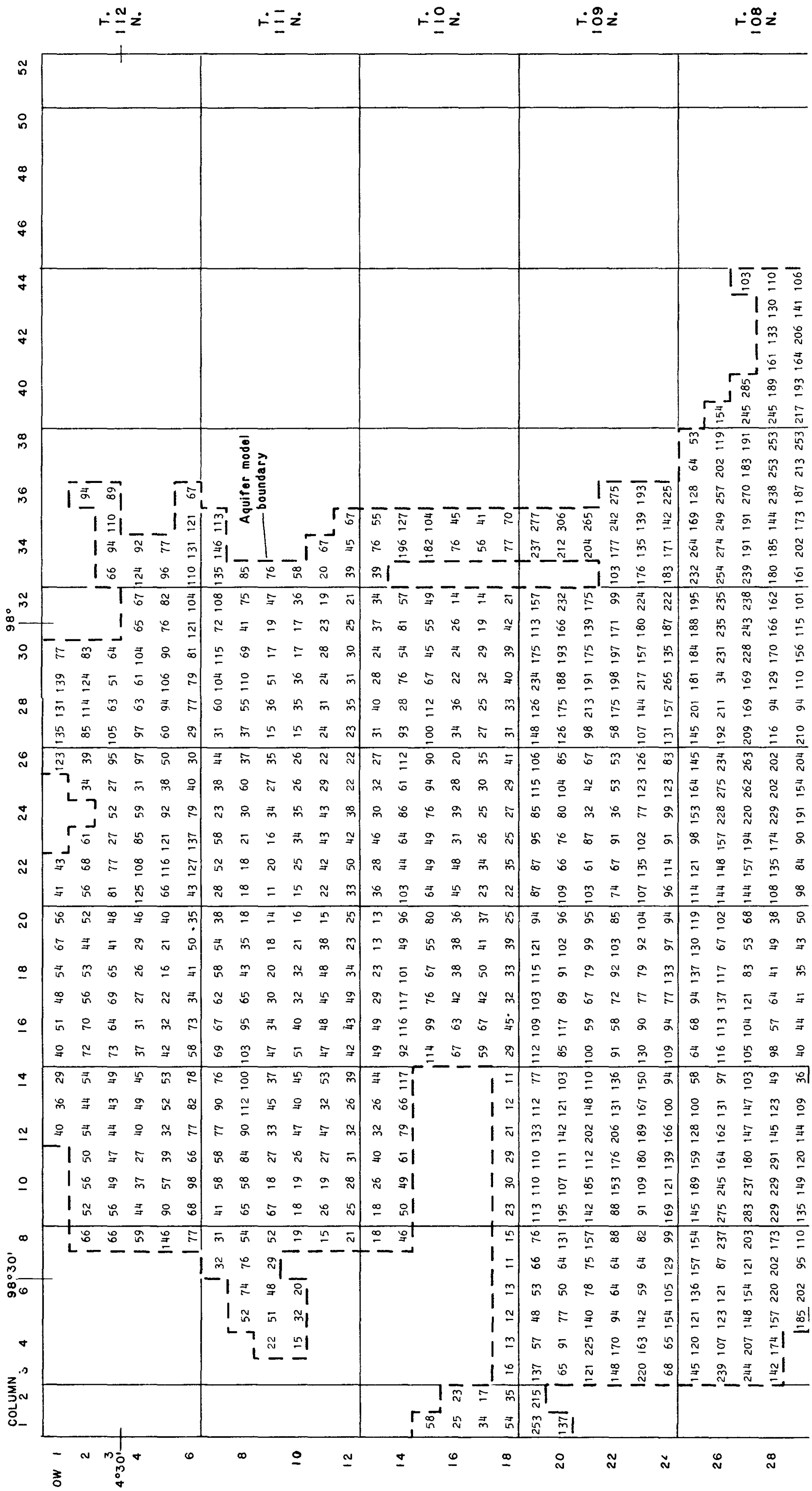


Figure 12.--Altitude of the bottom of the glacial-aquifer system.

Table 5.--Relation between grain size and hydraulic conductivity

Grain size	Range of hydraulic conductivities in glacial drift[1] (feet per day)	Hydraulic conductivity assigned to Big Sioux aquifer[1] (feet per day)	Hydraulic conductivities used in this model (feet per day)				
			Model row 1-8	Model row 9-13	Model row 14-15	Model row 16-18	Model row 19-56
Clay or silt	<20	10	12	6	15	6	22
Sand, very fine	10-80	40	--	--	--	--	--
Sand, fine	70-140	70	--	--	--	--	--
Sand, fine to medium	70-400	200	--	--	--	--	--
Sand, medium	130-400	270	--	--	--	--	--
Sand, fine to coarse	70-600	300	150	75	188	75	281
Sand, medium to coarse	130-800	400	--	--	--	--	--
Sand, coarse	400-1,000	540	--	--	--	--	--
Sand and gravel	400-1,200	600	170	85	212	85	319
Sand, coarse, and gravel	400-1,400	670	--	--	--	--	--
Gravel	800-2,000	800	190	95	238	95	356

[1]From Koch, 1980b.



COLUMN
ROW
98°30'
98°
44°30'
Aquifer model boundary
T. 112 N.
T. 111 N.
T. 110 N.
T. 109 N.
T. 108 N.

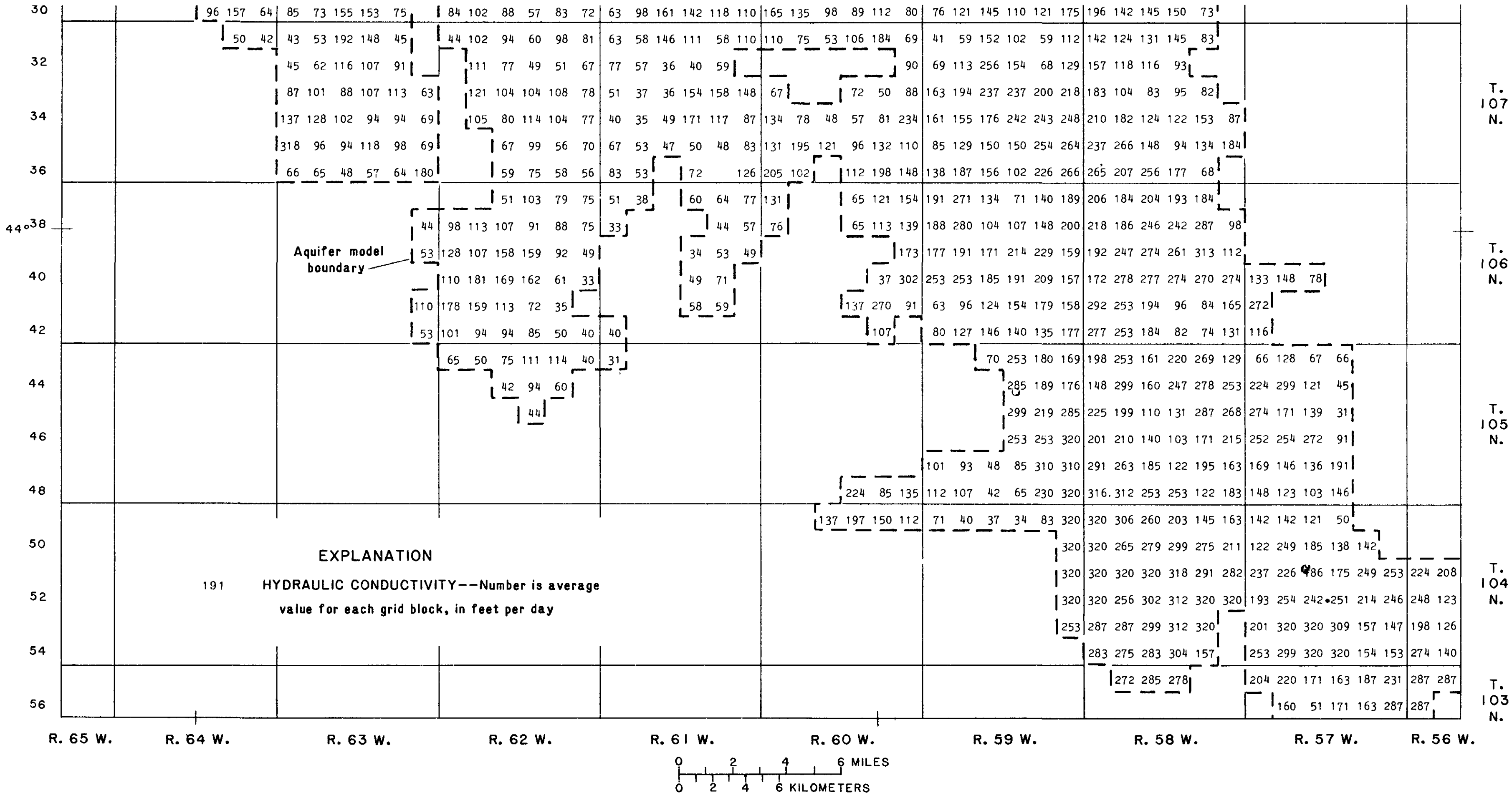


Figure 13.--Average hydraulic conductivity for each grid block.

Recharge to the Aquifer

The areal distribution of recharge to the aquifer was based on analyses of precipitation data (fig. 8) and on the thickness of the confining bed overlying the aquifer (fig. 7). The areal distribution of recharge to the aquifer was tested and refined as part of the steady-state simulation process. It was determined that the maximum available recharge to the aquifer was 7.0 in/yr and occurred only where the average thickness of the confining bed in the section overlying the aquifer was less than 10 ft. With average confining bed thicknesses between 10 and 45 ft, the rate of recharge to the aquifer decreased linearly to 0.0 in/yr. No recharge occurs when the average confining bed thickness in the section exceeds 45 ft. Figure 14 shows the percentage of available recharge that reaches the aquifer. Because an empirical relationship was developed between recharge and thickness of the confining bed overlying the aquifer, the values should not be considered absolute. The values are hydrologically reasonable and provide the best overall model results.

Evapotranspiration from the Aquifer

The areal distribution of evapotranspiration from the aquifer is controlled by the potential evapotranspiration (table 3), the thickness of the confining bed overlying the aquifer (fig. 7), and the depth of the potentiometric head below land surface. The areal distribution of potential evapotranspiration from the aquifer in each section was tested and refined as part of the steady-state simulation process. The best overall model results were obtained when the potential steady-state evapotranspiration rate was 36.0 in/yr. The potential evapotranspiration rate can occur only where no confining bed is present above the aquifer. Even though the potentiometric head in the aquifer may be close to land surface, the confining bed will restrict upward movement of water and reduce the potential evapotranspiration rate. Therefore, when the average confining bed thickness is between 0 and 45 ft, the potential evapotranspiration rate decreases linearly from 36.0 to 0.15 in/yr. The potential evapotranspiration remains constant at 0.15 in/yr for confining bed thicknesses greater than 45 ft. Figure 15 shows the percentage of the potential evapotranspiration available from the aquifer.

The evapotranspiration from the aquifer is controlled also by the depth of the aquifers' potentiometric surface below land surface. Evapotranspiration did not occur when the potentiometric heads are greater than 15 ft below land surface. Because an empirical relationship was developed between potential evapotranspiration rate from the aquifer, thickness of the confining bed overlying the aquifer, and depth of the potentiometric head below land surface, does not mean these values should be considered absolute, but only that they are reasonable hydrologically and provide the best overall model results.

Pumpage from the Aquifer

Ground-water withdrawal data are required to simulate the glacial-aquifer system. Pumpage data were collected for the period 1973 through 1983. Before 1973, the aquifer was in steady-state or equilibrium conditions. That is, although the water levels in the aquifer system may have declined during the summer months due to reduced recharge, increased evapotranspiration or pumpage, the water levels generally recovered to approximately the same or equilibrium levels during the winter or early spring months. Because the aquifer system was in equilibrium before 1973, pumpage was not included in the pre-1973 steady-state simulation.

Beginning in about 1973, ground-water withdrawals (table 4) had increased in some parts of the aquifer to the point that the aquifer did not fully recover before the next pumping season. To simulate the period from 1973 through 1983, the annual pumpage by section was included in the annual transient simulation. Examination of precipitation data (fig. 8) indicates that 1974 through 1976 was a period of below normal precipitation with 1976 being the driest. To simulate 1976, monthly pumpage by section was included in the 1976 monthly transient simulation (table 6).

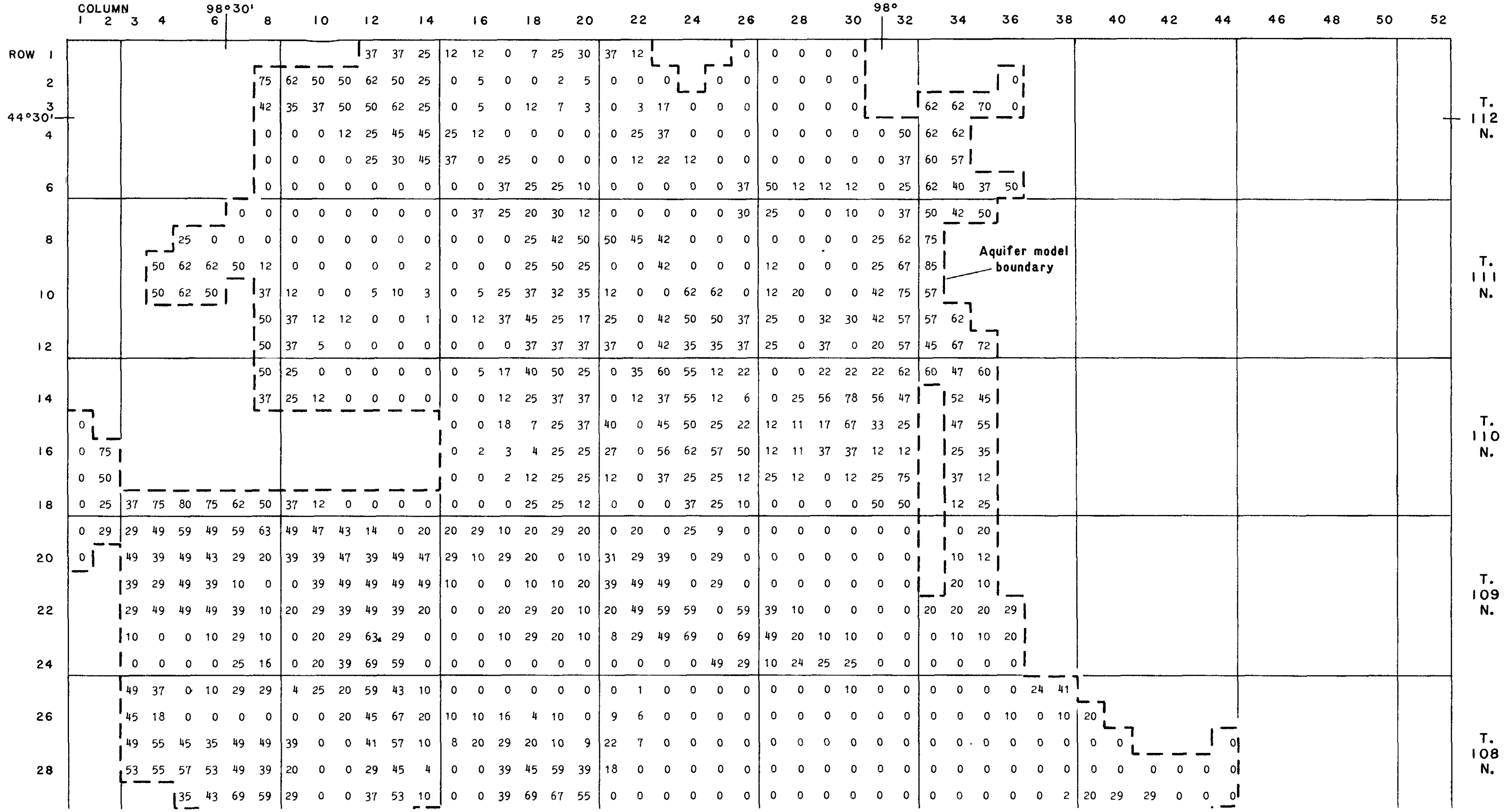
Table 6.--1976 monthly ground-water withdrawal from the study area

Month	Pumpage[1] (cubic feet per second)
January	1.59
February	1.59
March	1.59
April	.00
May	19.85
June	40.50
July	48.99
August	45.64
September	15.34
October	.00
November	1.59
December	1.59

[1]South Dakota Department of Water and Natural Resources (written commun.).

River Leakage from the Aquifer

Where the glacial-aquifer system is hydraulically connected to the James River, the river may contribute water to or drain water from the aquifer, depending on the head gradient. Hansen (1983) indicates that the river is hydraulically connected to the aquifer in a small part of the study area in Davison County; the locations of these river grid blocks are shown in figure 10 as river-head boundaries.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T. 111 N.
T. 110 N.
T. 109 N.
T. 108 N.
Aquifer model boundary

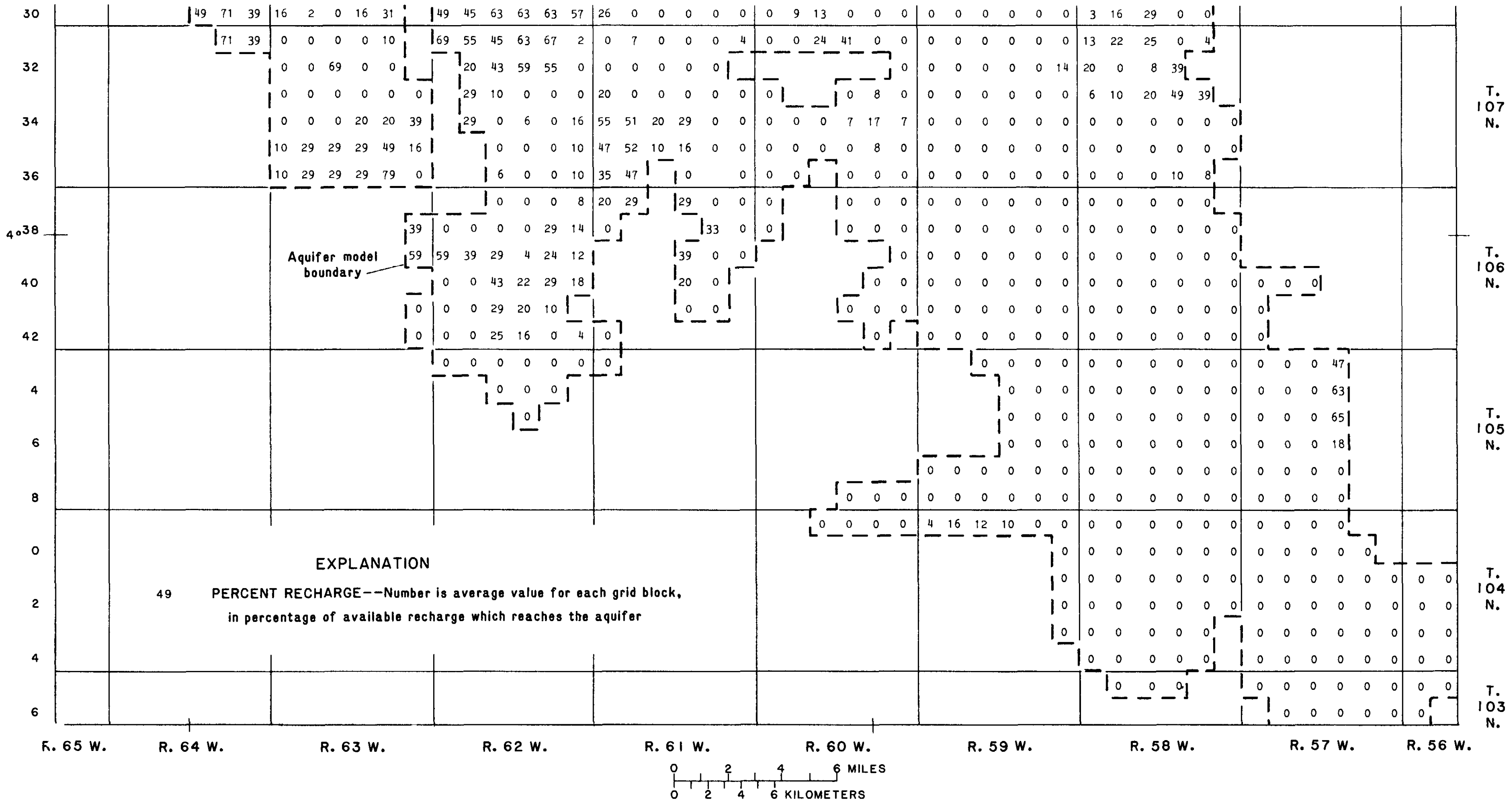
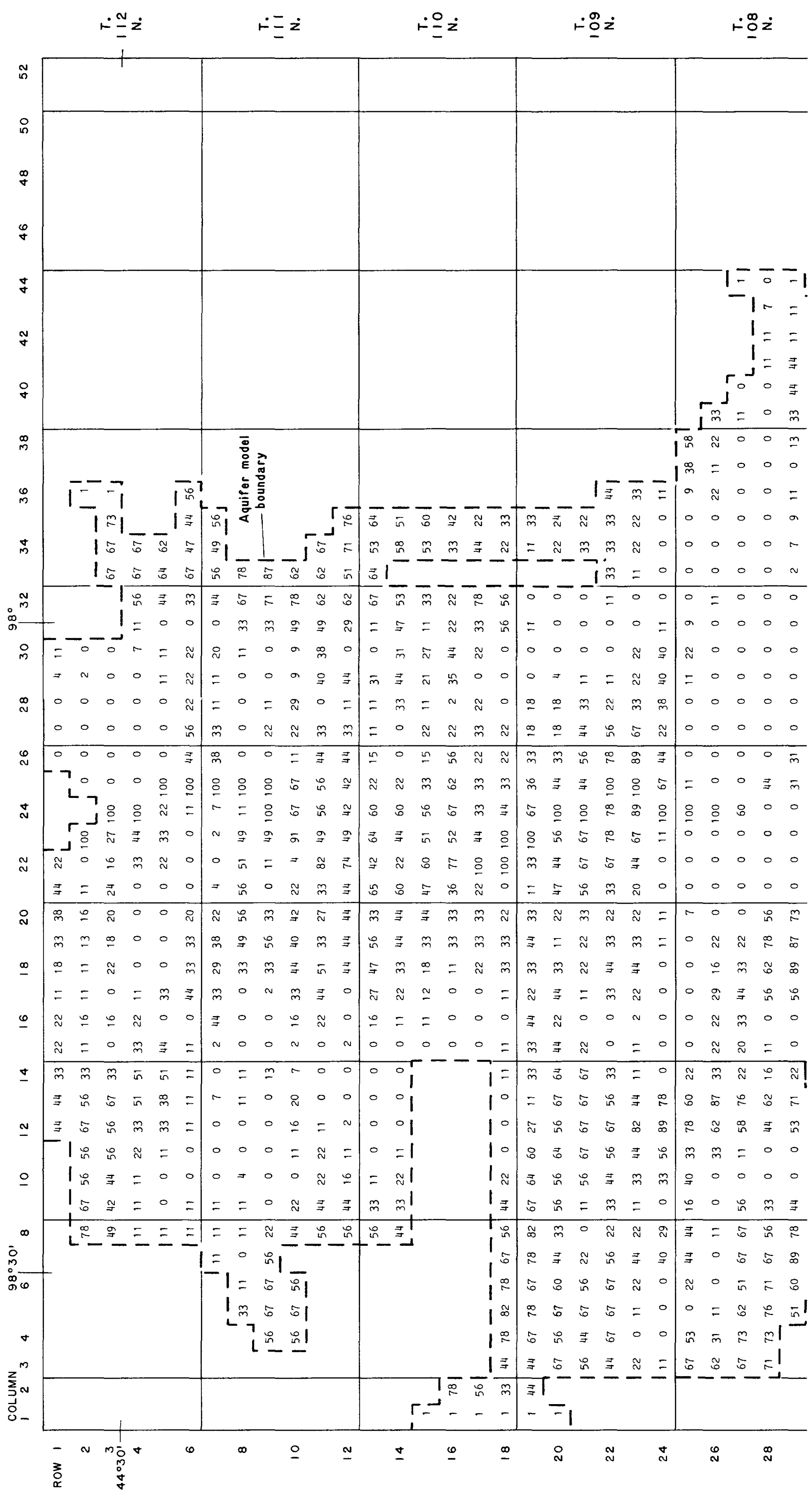


Figure 14.--Average recharge for each grid block.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Aquifer model boundary
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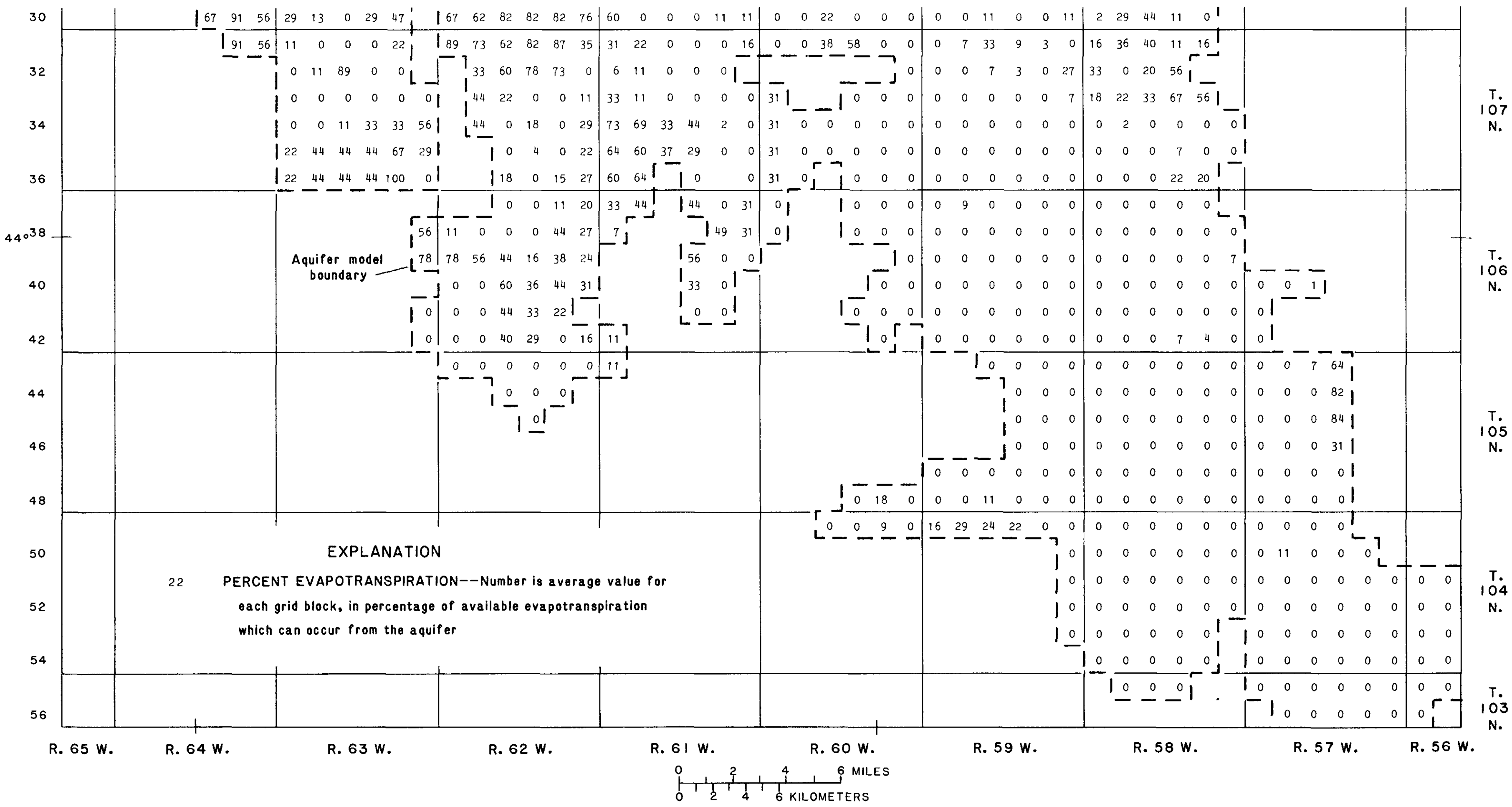

Figure 15.--Average evapotranspiration for each grid block.

CALIBRATION AND APPLICATION OF THE GROUND-WATER-FLOW MODEL

Model calibration is the process by which input data are adjusted so the model will adequately simulate historical potentiometric heads and flows. The initial equilibrium conditions were simulated by entering average recharge and evapotranspiration, and by setting the storage in the aquifer to zero. This is referred to as the steady-state or equilibrium simulation. The simulated steady-state potentiometric heads were compared to the observed annual average pre-1973 potentiometric heads to assess the accuracy of the steady-state simulation. The transient simulation includes storage and time-dependent recharge, evapotranspiration, and pumpage. Again, the simulated transient potentiometric heads were compared to observed potentiometric heads.

Calibration involves varying the values of hydraulic conductivity, recharge, evapotranspiration, and storage to bring simulated potentiometric heads closer to the observed potentiometric heads. The parameters were varied within reasonable hydrologic limits. Calibration was completed when a "best fit" between the simulated and observed potentiometric heads was obtained.

Table 7 gives an indication of how well the model duplicated observed potentiometric heads. The smaller the average difference between the simulated and observed potentiometric heads, the better the model represents the glacial-aquifer system. However, because of the uneven areal distribution of the data, the degree to which the model duplicates observed potentiometric heads can only be assessed where sufficient water-level data exist.

There are several means by which errors can be introduced into the analysis. The complexity of the aquifer can result in seemingly unusual water levels. In addition, nearby pumping can result in observed water levels which do not reflect natural conditions. Inaccurate measurement or recording of water levels can result in additional errors. Errors in the model formulation, estimation of the hydrologic parameters, and the lateral differences between well location and node center in the model will also produce differences between the simulated and the observed potentiometric heads. The table reflects the best composite set of average arithmetic and absolute differences obtained between the simulated and observed potentiometric heads for the steady-state simulation, the 1976 monthly, and 1973-83 transient simulations.

Steady-State Simulation

The steady-state simulation provides information on the hydrologic conditions in the glacial-aquifer system before significant ground-water development; no storage terms or pumpage are included in the simulation.

The ground-water withdrawals in the study area before 1976 were much less than the withdrawals after 1976 (table 4). Before 1973 the aquifer generally was in equilibrium with water levels nearly recovering to prepumping levels during the nonirrigated fall, winter, and spring seasons (fig. 9). Precipitation in the study area was significantly less than normal from 1974 through 1976 (fig. 8). As a result of the drought, large and continued increases in ground-water withdrawals began in 1976. Therefore, the steady-state simulation represents average conditions for the glacial-aquifer system before 1973.

Table 7.--Comparison between simulated and observed potentiometric heads

Model simulation	Average arithmetic difference between simulated and observed potentiometric heads[1] (feet)	Average absolute difference between simulated and observed potentiometric heads[2] (feet)	Maximum positive difference between simulated and observed potentiometric heads[3] (feet)	Maximum negative difference between simulated and observed potentiometric heads[4] (feet)	Number of observation wells with observed potentiometric heads
Steady-state	1.68	4.38	16.22	8.32	32
Transient					
1973	-1.27	4.57	14.39	10.35	32
1974	-3.88	6.65	12.91	15.47	32
1975	-1.52	4.70	13.89	11.57	32
1976	-3.10	5.98	17.65	18.02	50
1977	-.49	5.26	15.58	27.47	60
1978	-.99	6.32	24.32	34.45	74
1979	.75	9.18	43.80	45.38	110
1980	-.21	9.86	44.46	45.64	114
1981	.75	10.39	46.75	45.83	114
1982	2.23	11.70	44.20	46.00	122
1983	.54	11.46	41.91	47.18	120
January 1976	--	--	--	--	0
February 1976	-.93	5.36	14.89	13.84	17
March 1976	-.63	5.28	14.62	16.05	32
April 1976	--	--	--	--	0
May 1976	--	--	--	--	0
June 1976	.17	5.74	9.53	14.55	15
July 1976	2.68	6.88	14.12	14.62	16
August 1976	1.45	5.85	17.62	14.76	35
September 1976	1.42	5.74	9.34	14.88	13
October 1976	.79	3.82	10.59	15.05	24
November 1976	-1.24	4.29	8.73	14.64	14
December 1976	1.55	4.11	17.68	11.98	31

[1]Positive number indicates simulated head was higher than the observed head; negative number indicates simulated head was lower than the observed head.
[2]The absolute value of a number is the number without its associated sign. For example, the absolute value of 2 and -2 are the same.
[3]Positive difference when simulated head is greater than observed water level.
[4]Negative difference when simulated head is less than observed water level.

There are water-level data from 32 observation wells completed in the aquifer system for the period before 1973 (fig. 16). The observed potentiometric heads are used to check the accuracy of the simulated potentiometric surface. The average arithmetic difference between the predevelopment simulated and observed water levels was 1.68 ft and the average absolute difference was 4.38 ft (table 7). The difference between simulated and observed heads was more than 10 ft at three locations. The simulated head was 15.28 ft higher than the average observed water level in the observation well located in grid block: row 5, column 22, 16.22 ft higher in row 20, column 27, and 13.37 ft in row 30, column 35 (fig. 16).

The reasons for the discrepancies are unknown but may be due to the complexity of the glacial-aquifer system. These observation wells may be partly isolated from the surrounding aquifer by till or clay and silt outwash, and therefore, water levels from these wells may not represent the regional potentiometric surface. Also due to the simplifying assumptions in the model and the size of the finite-difference grid, the simulated steady-state heads will contain inaccuracies. However, the model is one of the best means of improving and evaluating our understanding of the aquifer system and of testing the sensitivity of the model to changes in selected aquifer properties.

The highest potentiometric heads are located on the eastern and western boundaries of the model area and the lowest heads are near the James River which runs approximately through the center of the area simulated (fig. 16). Ground water flows from higher to lower potentiometric head and perpendicular to the potentiometric contours. The flow west of the James River is eastward toward the river and east of the James River the flow is westward or northwestward. Previous studies have indicated that the James River gains little water from the underlying glacial-aquifer system. The steady-state simulation shows that evapotranspiration can reasonably remove enough water from the aquifer to approximate the predevelopment potentiometric surface.

Recharge from precipitation was 97.1 percent of the predevelopment inflow to the aquifer (table 8). The average annual recharge to each active model grid block was 0.96 inch. This value is in the range of aquifer recharge calculated from observation-well data, flow-net analyses, and values used in other computer models (table 2). Evapotranspiration accounts for 98.2 percent of the outflow from the predevelopment steady-state aquifer. The average evapotranspiration from each active model grid block was 0.97 in/yr.

Transient Simulation

The transient simulation includes pumpage from and storage in the aquifer system. The transient or pumping simulation includes 11 consecutive pumping periods between 1973 and 1983 (table 4) and 12 monthly pumping periods in 1976 (table 6). The starting potentiometric heads in the 1973 transient simulation are the heads generated by the steady-state simulation. Subsequent annual simulations used the potentiometric heads generated by the preceeding simulation. The starting potentiometric heads in the 1976 monthly transient simulation were those generated by the 1975 annual transient simulation. Subsequent monthly simulations used potentiometric heads generated by the preceeding monthly simulation.

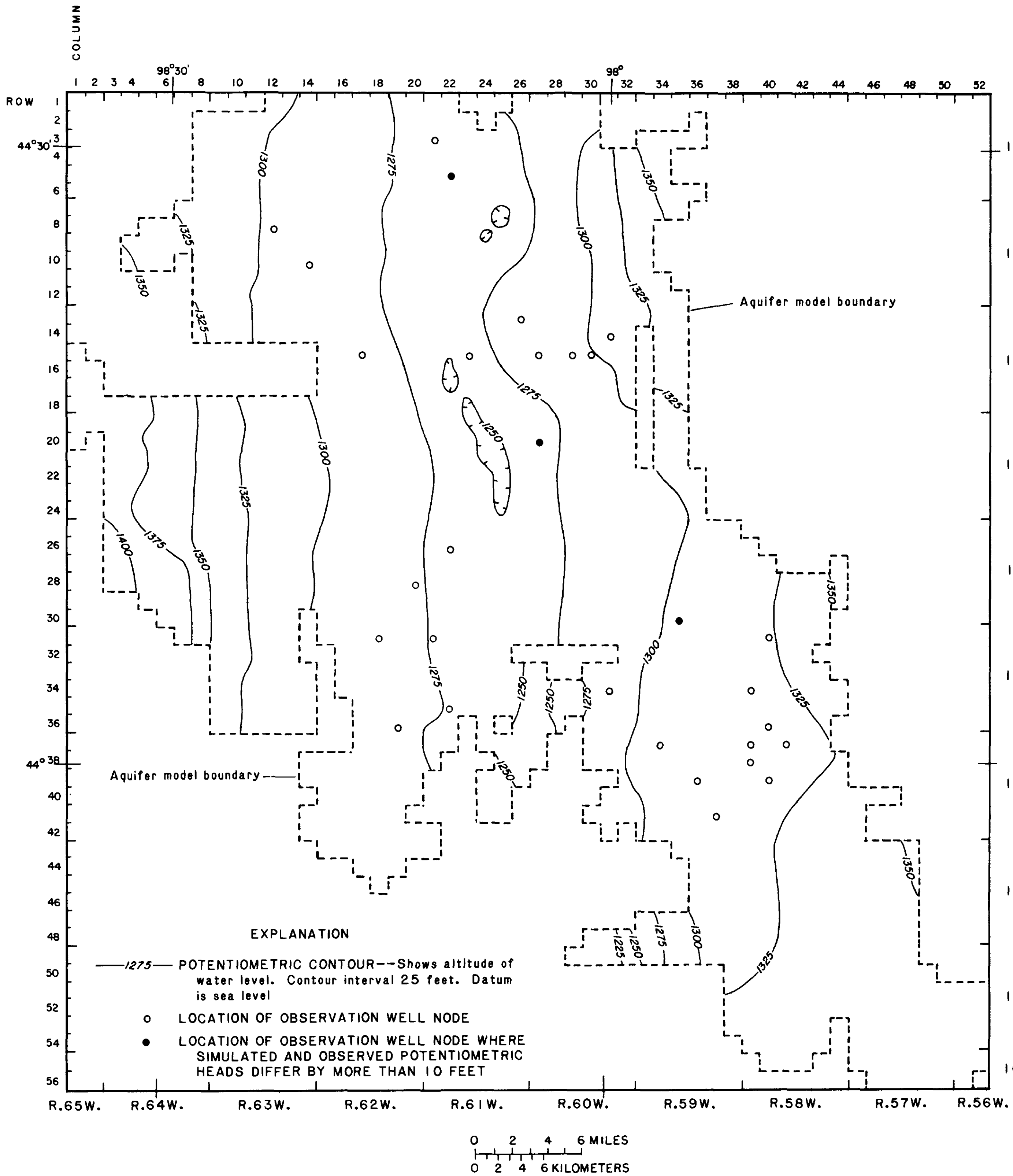

Figure 16.--Simulated prepumping potentiometric surface.

Table 8.--Simulated water budget under steady-state conditions

Budget component	Flow rates in cubic feet per second	Percent
INFLOW		
Recharge to the aquifer from precipitation	101.42	97.1
Inflow at specified-head boundaries	3.06	2.9
Total inflow	104.48	100.0
OUTFLOW		
Evapotranspiration from the aquifer	102.85	98.2
Discharge from the aquifer to the stream	1.36	1.3
Outflow at specified-head boundaries	.51	.5
Total outflow	104.73	100.0

The following sections briefly compare the simulations with the known hydrology at the end of 11 annual and 12 monthly periods. The most recent (1983) annual simulation and the August 1976 monthly simulation will be compared in more detail.

1973-83 Simulation Period

Water levels in the aquifer system were at or near their highest levels at the end of 1972 and the beginning of 1973 (fig. 9). In the spring of 1973, the water levels began to decline due primarily to a substantial decrease in precipitation in the study area (fig. 8). This drought which lasted through 1976 created a large increase in the demand for ground water for irrigation (tables 4 and 6).

Recharge to and evapotranspiration from the glacial-aquifer system are difficult to estimate accurately. They are controlled by a number of complex, interrelated variables which are not fully understood and cannot be accurately measured. The recharge and evapotranspiration rates used in the transient simulations were estimated based on only three of these variables: precipitation, pan evaporation, and thickness of the confining bed overlying the aquifer system.

Maximum annual recharge to the aquifer system was estimated from the average annual precipitation recorded at Huron, Forestburg, and Mitchell (table 9). The maximum annual recharge ranged from 0.10 inch in 1976 to 8.14 inches in 1977 which correspond to the driest and wettest years. The percentage of the maximum available annual recharge which actually infiltrates is controlled by the thickness of the confining bed overlying the glacial-aquifer system (fig. 14).

Table 9.--Average annual precipitation and estimated recharge

Year	Average annual precipitation for Huron, Forestburg, and Mitchell[1] (inches)	Average annual departure[2] (inches)	Maximum annual recharge to the glacial aquifer system[3] (inches)
1973	21.53	-1.50	6.62
1974	13.30	-9.73	.45
1975	18.91	-4.12	4.66
1976	12.55	-10.48	.10
1977	26.82	3.79	8.14
1978	18.35	-4.68	4.24
1979	22.47	-.56	7.00
1980	15.74	-7.29	2.28
1981	17.79	-5.24	3.82
1982	26.69	3.66	8.10
1983	20.39	-2.64	5.77

[1]Data from the U.S. Department of Commerce, 1973-83.
[2]Departure from 23.03 inches per year, the average annual precipitation for Huron, Forestburg, and Mitchell, from 1968 through 1972. Data from the U.S. Department of Commerce, 1968-72.
[3]Calculated from the following table:

Departure (inches)	Recharge calculation
0.0	7.00 inches + (0.30 x departure)
0.0 to -1.00	7.00 inches
<-1.00 to -10.00	7.00 inches - [(departure - 1.00 inch) x 0.75]
<-10.00	0.10 inches

Table 10.--Estimated annual pan evaporation and estimated potential evapotranspiration

Year	Annual pan evaporation[1] (inches)	Annual potential evapotranspiration[2] (inches)
1973	51	37.2
1974	48	35.0
1975	49	35.8
1976	67	48.9
1977	52	38.0
1978	49	35.0
1979	42	30.7
1980	50	36.5
1981	51	37.2
1982	41	29.9
1983	46	33.6
Average	49.6	36.2

[1]Estimated using data from the U.S. Department of Commerce, 1973-83.
[2]Calculated as 0.73 times the annual evaporation.

The evapotranspiration rate from the aquifer also is difficult to estimate. The estimate of the annual potential evapotranspiration is calculated as 73 percent of the estimated total annual pan evaporation (table 10). The estimated total annual pan evaporation varied from 41 inches in 1982 to 67 inches in 1976 with an average of 49.6 inches. Annual potential evapotranspiration varied from 29.9 inches in 1982 to 48.9 inches in 1976 with an average of 36.2 inches. Figure 15 shows the percentage of the potential available evapotranspiration which can be withdrawn from the aquifer system.

The acreage permitted to be irrigated with ground water in the James River basin before 1973 was small (table 4). However, as a result of the 1974-76 drought, the quantity of permitted acreage increased dramatically after 1975. Ground-water withdrawals in the study area prior to 1976 were small. In 1976, withdrawals were 2.6 times the withdrawal in 1975. Since 1976, the ground-water withdrawals in the study area have fluctuated annually in both distribution and quantity; however, the maximum annual withdrawals have not increased significantly since the large increase in 1976. The location of each grid block in which pumping occurred in 1983 is shown in figure 17.

The average arithmetic difference between the simulated and observed water levels for the eleven annual calibration periods ranged from -3.88 ft in 1974 to 2.23 ft in 1982 (table 7). The average absolute difference ranged from 4.57 ft in 1973 to 11.70 ft in 1982. The average arithmetic difference between the simulated and observed water levels for 1983 was 0.54 ft and the average absolute difference was 11.46 ft. The maximum positive difference

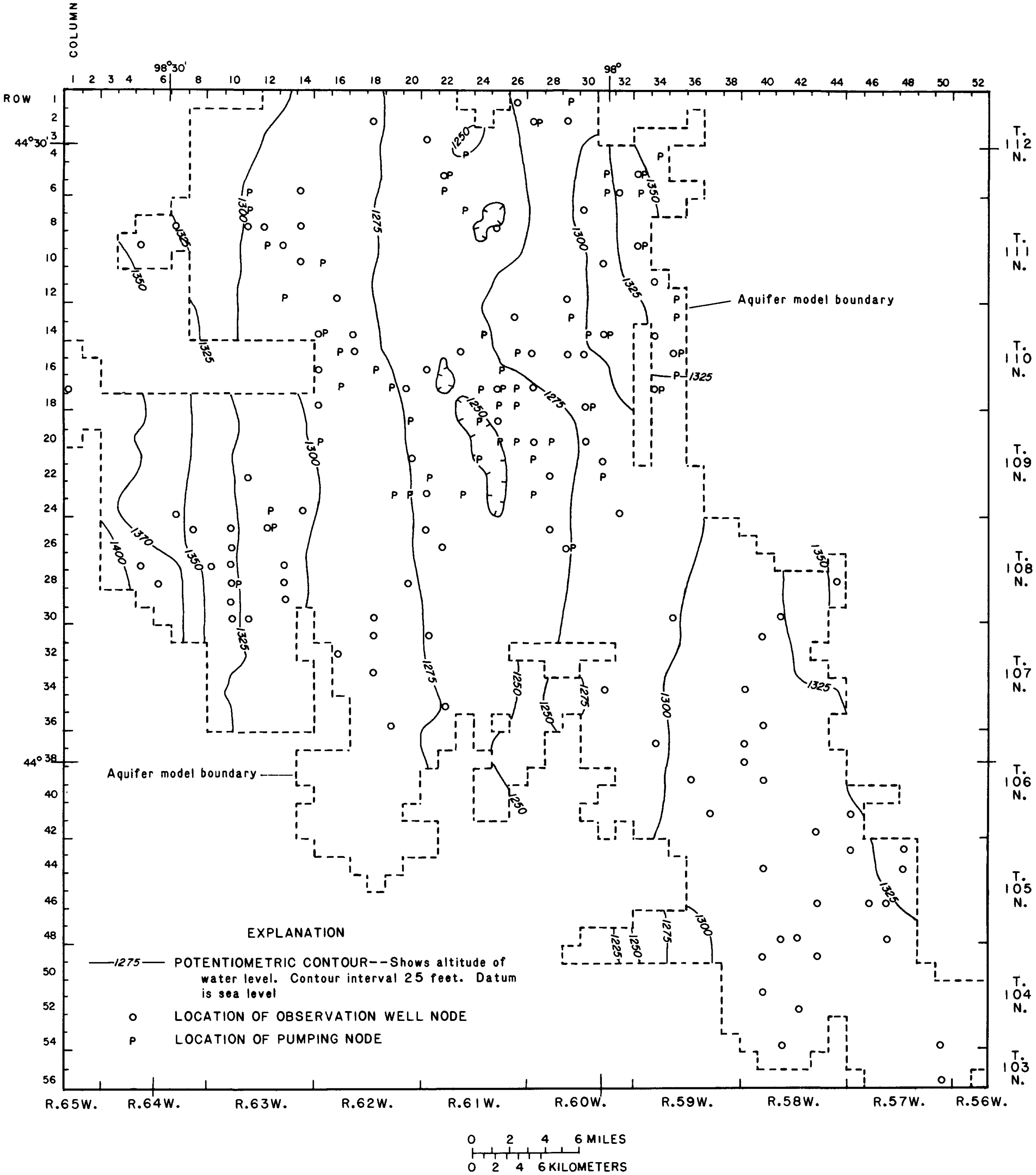

Figure 17.--Simulated 1983 potentiometric surface.

between the simulated and observed water levels was 41.91 ft in the observation well located in grid block row 52, column 42, and the maximum negative difference was 47.18 ft in the observation well located in grid block row 54, column 50 (fig. 17). The relatively large 1983 maximum positive and negative differences are probably a result of the complexity of the glacial-aquifer system and the model's inability to simulate the aquifer on a scale small enough to adequately represent this complexity. The model does, however, adequately simulate the glacial-aquifer system on a one-square-mile scale.

Figure 18 shows four hydrographs of measured water levels from observation wells in the study area and the simulated potentiometric heads for the corresponding grid block in which the observation wells are located. The four hydrographs indicate that there is generally good agreement between the simulated and observed potentiometric heads and demonstrates the ability of the transient simulation to simulate the aquifers' responses to changes in annual recharge, evapotranspiration, and pumping.

The highest simulated 1983 potentiometric heads are located along the eastern and western boundaries of the model area and the lowest heads are found along the James River (fig. 17). The potentiometric heads range from greater than 1,400 ft above sea level along the west boundary of the model to less than 1,250 ft above sea level along the James River. The direction of ground-water flow generally is eastward, west of the James River and westward or northwestward, east of the James River. Configuration of the potentiometric surface and direction of ground-water flow are similar to those of the predevelopment conditions (fig. 16).

Comparison of the simulated predevelopment steady-state and 1983 potentiometric heads indicates that the heads have declined more than 25 ft (fig. 19). Although drawdowns of more than 25 ft have occurred, the conversion from confined to unconfined conditions will result in a significant reduction in this drawdown rate. When the head declines in a confined aquifer, water in storage is released from the expansion of water and from the compression of the aquifer. Where the aquifer is unconfined, the predominant source of water is from gravity drainage of the sediments through which the decline in the water table occurs. In an unconfined aquifer, the volume of water derived from expansion of water and compression of the aquifer is negligible (Heath, 1983).

A hydrologic budget equates accretions to the water supply to depletions of the water supply. A budget equation states that inflow minus outflow equals change in storage. A general equation of the hydrologic budget for the glacial-aquifer system may be written:

precipitation + inflow at boundaries + discharge from ground-water storage = evapotranspiration + outflow at boundaries + recharge to ground-water storage + pumpage.

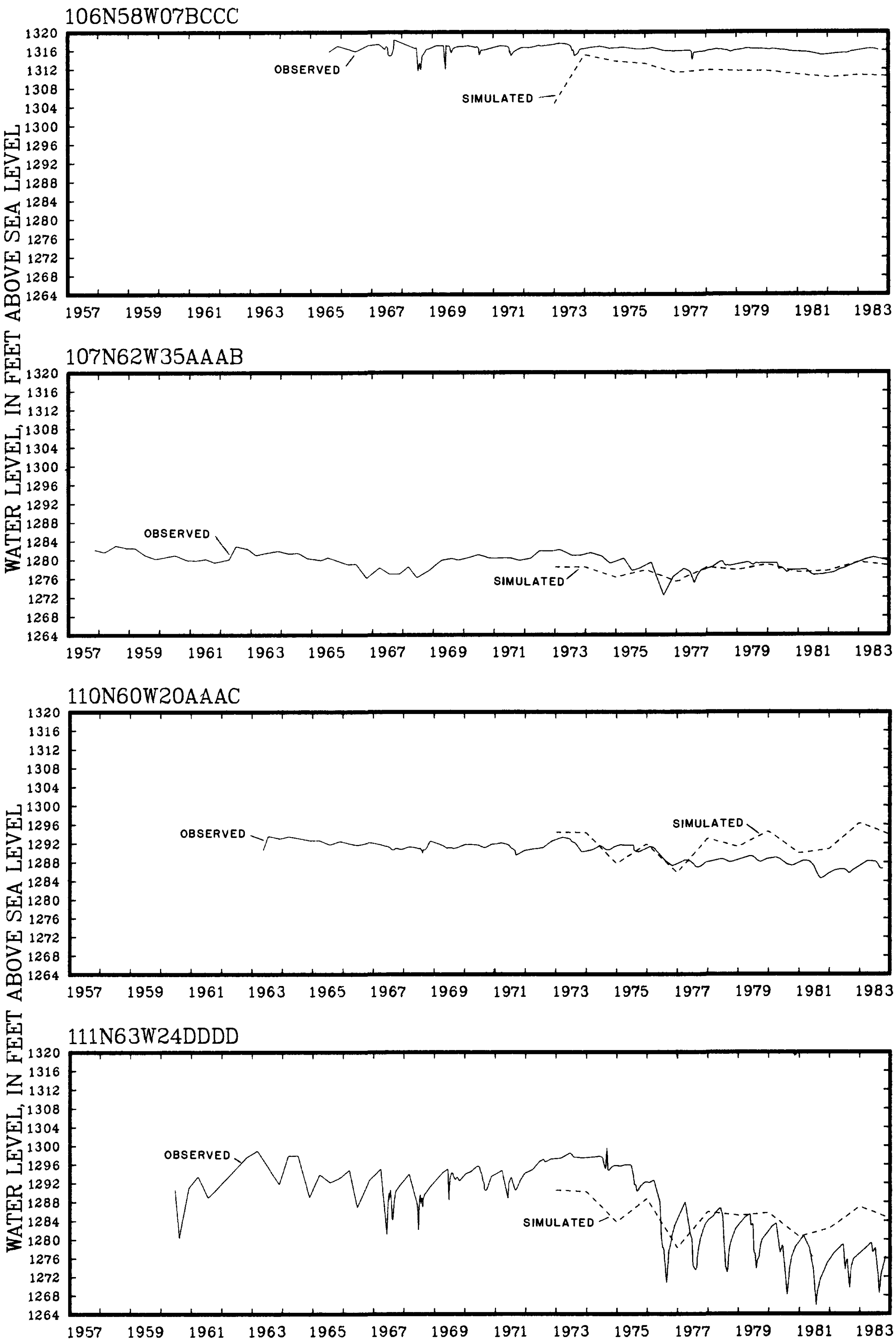

Figure 18.--Comparison of simulated and observed annual potentiometric heads.

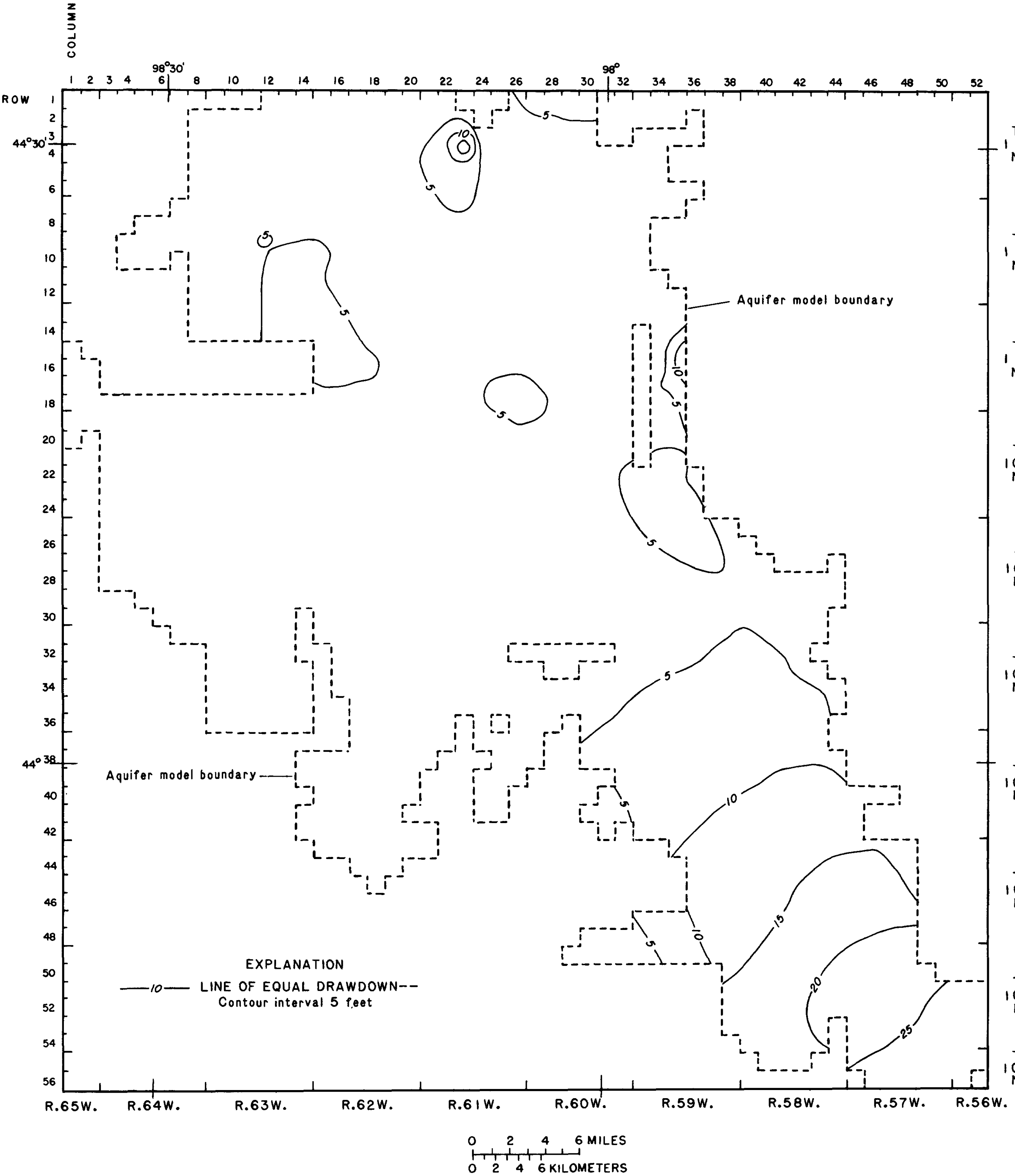

Figure 19.--Drawdown between the simulated predevelopment steady-state potentiometric surface and the simulated 1983 transient potentiometric surface.

Table 11.--Annual simulated water budgets, 1973-83

	1973	1974	1975	1976	1977	1978	1979	1980	1981	1982	1983
Sources of water (accretions) (cubic feet per second)											
Recharge from precipitation	106.09	12.74	88.58	3.62	132.14	75.98	110.72	50.66	75.99	131.42	101.31
Recharge from specified-head boundaries	2.82	3.07	2.93	3.21	2.89	3.00	2.87	3.08	3.04	2.83	2.94
Discharge from storage	3.09	42.33	4.12	55.60	2.43	5.44	1.14	19.88	8.88	1.84	3.86
Total	112.00	58.14	95.63	62.43	137.46	84.42	114.73	73.62	87.91	136.09	108.11
Consumption of water (depletions) (cubic feet per second)											
Evapotranspiration from aquifer	106.61	52.91	81.37	46.55	91.48	69.73	88.24	61.74	69.02	99.80	89.48
Pumpage	2.66	3.37	5.33	14.10	15.64	11.73	8.47	9.79	16.06	9.39	12.32
Discharge from the aquifer to the stream	1.71	1.69	1.47	1.64	1.56	1.30	1.47	1.66	1.59	1.38	1.44
Discharge from specified-head boundaries	.62	.60	.60	.55	.58	.58	.59	.57	.56	.58	.57
Recharge to storage	.58	.00	7.03	.00	28.12	1.36	15.85	.33	1.00	24.70	4.36
Total	112.18	58.57	95.80	62.84	137.38	84.70	114.62	74.09	88.23	135.85	108.17

The water budgets for the 11 transient simulation periods vary considerably as a result of changes in the annual recharge from precipitation and discharge from evapotranspiration (table 11). The recharge rate from precipitation was very small in 1974 and 1976 compared with the other annual simulations. The reduced recharge resulted in decreased potentiometric heads in the aquifer, thereby causing reduced evapotranspiration rates and an increase in the discharge of water from storage.

The 1983 simulated water budget (table 11) indicates recharge from precipitation is the major budget component, contributing approximately 94 percent of the inflow. Specified-head boundaries account for 2.7 percent of the inflow. Quantitatively, recharge from precipitation and inflow from specified-head boundaries are about the same as the predevelopment steady-state water budget (table 8). The major difference between the predevelopment and 1983 simulated water inflows is storage. In 1983 inflow from storage accounted for 3.86 ft^3/s or 3.6 percent of the accretions. In 1983 evapotranspiration accounted for 89.48 ft^3/s and the specified-head boundaries accounted for 0.57 ft^3/s of the depletions from the aquifer compared to 105.82 and 0.70 ft^3/s, respectively in the predevelopment steady-state simulation (table 8). Pumpage accounted for 12.32 ft^3/s and recharge to storage 4.36 ft^3/s in the 1983 depletions.

1976 Monthly Simulation Period

The last and driest year in the 1974-76 drought was 1976. Monthly model runs for 1976 were chosen to determine how well the transient simulation could simulate a year when estimated maximum recharge was very small and potential evapotranspiration was calculated to be the greatest of the 11 annual simulation periods.

The maximum recharge rates to the aquifer for each month in 1976 are shown in table 12. The average annual maximum recharge rate of 0.10 inch (table 9) was distributed in March and April which accounts for recharge due to snowmelt in the spring, and in September when the average annual departure was 0.20 inch above normal due to rain storms. The percentage of the maximum available monthly recharge which can actually reach the aquifer is controlled by the average thickness of the confining bed in each grid block overlying the glacial-aquifer system (fig. 14).

The monthly evapotranspiration for 1976 (table 13) was estimated using Department of Commerce data (1976) and the relationships presented by Farnsworth and Thompson (1982) and Farnsworth, Thompson, and Peck (1982) for estimating evapotranspiration from pan evaporation data. The calculated potential evapotranspiration rate ranges from 12.5 inches in August to 0.0 inch in the winter months when the ground is frozen. Figure 15 shows the percentage of the potential monthly evapotranspiration which can be withdrawn from the glacial-aquifer system.

Figure 20 shows the simulated potentiometric surface at the end of August 1976 and the location of each grid block in which ground-water withdrawal occurred. August was selected because it was the month when the potentiometric heads were the lowest, as indicated by the hydrographs in figure 21. The summer months of June, July, and August had the greatest ground-water withdrawal rates (table 6).

Table 12.--Average monthly precipitation and recharge for 1976

Month	Average monthly precipitation for Huron, Forestburg, and Mitchell[1] (inches)	Average monthly departure[1] (inches)	Maximum monthly recharge to the glacial aquifer system (inches)
January	0.44	+0.02	--
February	.60	-.09	--
March	.86	-.29	0.05
April	1.79	-.43	.02
May	1.01	-1.97	--
June	2.32	-1.48	--
July	1.57	-1.00	--
August	.46	-1.98	--
September	2.17	+.20	.03
October	.98	-.66	--
November	.03	-.73	--
December	.33	-.23	--

[1]Data from the U.S. Department of Commerce, 1976.

Table 13.--Estimated monthly evapotranspiration for 1976

Month	Maximum evapotranspiration (inches)
January	0.0
February	.0
March	2.5
April	4.5
May	7.0
June	8.5
July	12.0
August	12.5
September	2.5
October	.5
November	.0
December	.0
Total	50.0

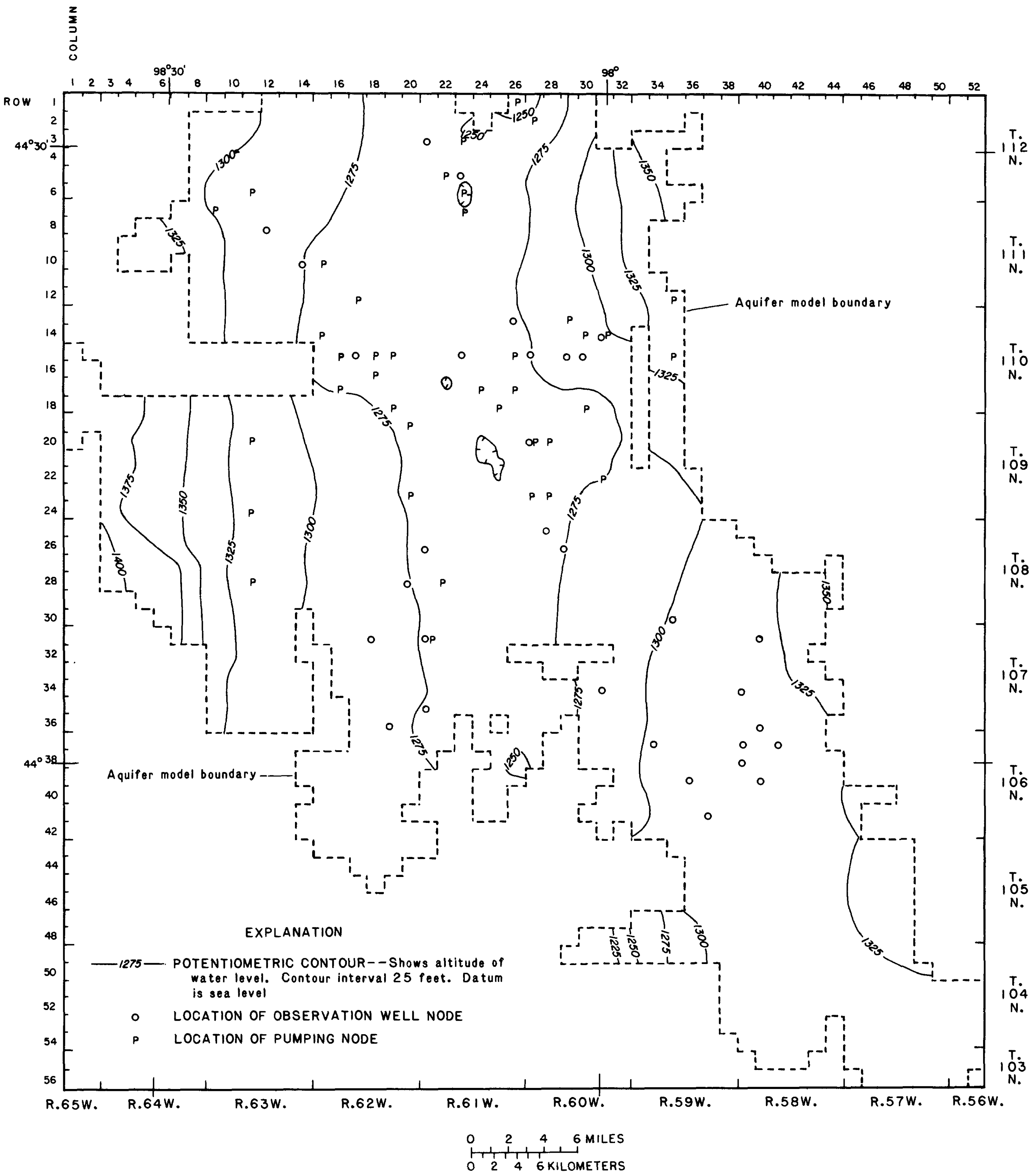

Figure 20.--Simulated potentiometric surface, August 1976.

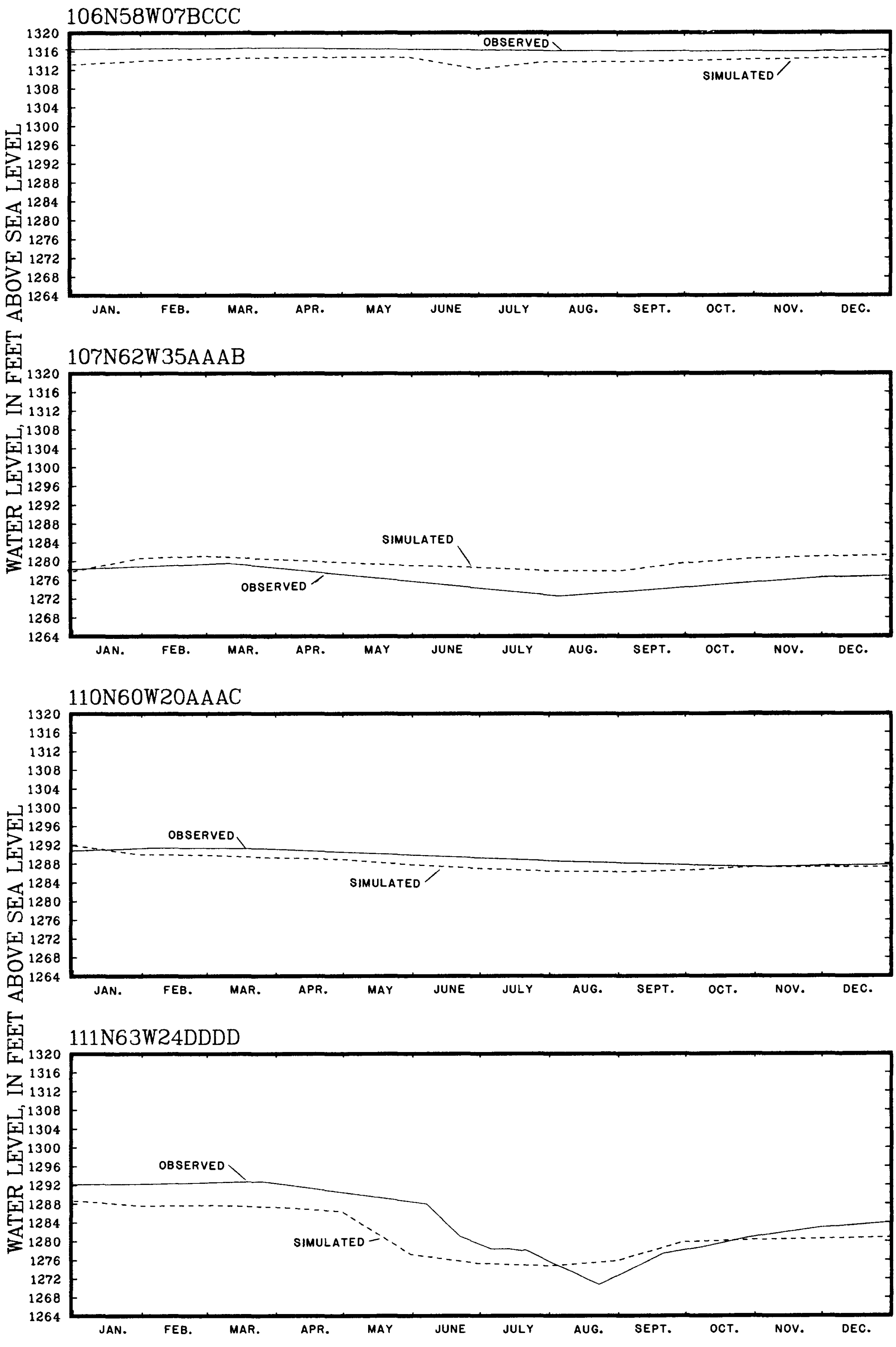

Figure 21.--Comparison of simulated and observed monthly potentiometric heads, 1976.

The average arithmetic difference between the simulated and observed water levels for the 12 monthly simulation periods ranged from -1.24 ft in November to 2.68 ft in July. The average absolute difference ranged from 3.82 ft in October to 6.88 ft in July (table 7). In August, the average arithmetic difference between the computed and measured water levels was 1.45 ft and the average absolute difference was 5.85 ft. The maximum positive difference between the simulated and observed water levels in August was 17.62 ft in the observation well located in grid block row 35, column 21, and the maximum negative difference was 14.76 ft in the observation well located in grid block row 15, column 30 (fig. 20). The average arithmetic and absolute differences between the computed and measured water levels (table 7) and the hydrographs (fig. 21) demonstrate the ability of the monthly transient simulations to very approximately replicate the aquifers' response in 1976 to monthly changes in pumping, recharge, and evapotranspiration.

The configuration of the August 1976 potentiometric surface (fig. 20) is similar to the predevelopment potentiometric surface (fig. 16) and the 1983 simulated potentiometric surface (fig. 17) except that the heads are, in general, lower. The August potentiometric heads range from greater than 1,400 ft above sea level along the west boundary of the model to less than 1,250 ft above sea level along the James River. The direction of ground-water flow generally is eastward, west of the James River and westward or northwestward east of the James.

A simulated water budget equating monthly accretions and depletions of the water supply for 1976 is shown in table 14. The water budget for the 12 monthly simulation periods varies considerably as a result of changes in the monthly evapotranspiration, storage, and pumpage. The primary source of water in 1976 was from storage. Recharge from precipitation and leakage across the model boundaries, represented by specified-head nodes, generally supplied less than 10 percent of the water. The major losses of water were pumpage and evapotranspiration during the months of May through August. In the fall and winter months, most of the water discharged from storage went back into storage elsewhere in the model.

MODEL SENSITIVITY

The confidence in the model's response needs to be based on a subjective appraisal of the analogy between the glacial-aquifer system and the model. A significant part of this analogy is the assumption that the aquifer characteristics have the same or similar characteristics assumed in the model. Because the aquifer characteristics are not known with certainty, the sensitivity of the model to each of several selected characteristics was tested.

The sensitivity of the model was tested by changing the values assigned for recharge, evapotranspiration, and hydraulic conductivity. The extent to which these variations affect the simulated response is a qualitative measure of the sensitivity of the model to uncertainty in that aquifer characteristic. Thus, if the variation produces a minor change in the predicted response, the model is not sensitive to that aquifer characteristic.

Table 14.--Monthly simulated water budgets, 1976

	January	February	March	April	May	June	July	August	September	October	November	December
Sources of water (accretions) (cubic feet per second)												
Recharge from precipitation	0.00	0.00	0.72	0.29	0.00	0.00	0.00	0.00	0.43	0.00	0.00	0.00
Recharge from specified-head boundaries	2.57	2.57	2.67	2.74	2.81	2.85	2.92	2.93	2.70	2.61	2.59	2.54
Discharge from storage	26.30	24.12	25.22	26.08	44.28	64.90	77.10	73.87	38.06	25.98	24.17	22.69
Total	28.87	26.69	28.61	29.11	47.09	67.75	80.02	76.80	41.19	28.59	26.76	25.23
Consumption of water (depletions) (cubic feet per second)												
Evapotranspiration from aquifer	.00	.00	10.49	16.80	21.88	23.76	28.62	29.06	8.44	1.99	.00	.00
Pumpage	1.57	1.57	1.57	.00	19.49	40.03	48.45	44.74	15.15	.00	1.57	1.57
Discharge from the aquifer to the stream	1.93	1.86	1.72	1.51	1.54	1.76	2.00	2.13	2.04	1.97	1.91	1.86
Discharge from specified-head boundaries	.66	.68	.69	.69	.69	.66	.65	.65	.67	.68	.69	.70
Recharge to storage	24.39	22.19	14.10	10.20	3.57	1.73	.34	.20	14.67	23.57	22.20	20.85
Total	28.55	26.30	28.57	29.20	47.17	67.94	80.06	76.78	40.97	28.21	26.37	24.98

Sensitivity of the simulated steady-state condition is described by comparing the standard steady-state simulation (the one described thus far in the report) with an alternative simulation (one in which an aquifer characteristic had an alternative value).

Table 15 shows the sensitivity of the steady-state simulation to changes in recharge, evapotranspiration, and hydraulic conductivity. The areal distribution of the percentage of maximum recharge and potential evapotranspiration in each grid block was not changed. Also the areal distribution of the hydraulic conductivity was not changed.

The steady-state simulation is most sensitive to changes in recharge. A 25-percent reduction in the maximum recharge rate from 7.00 to 5.25 in/yr resulted in the average arithmetic difference decreasing 2.99 ft. Also the maximum positive and negative differences changed 1.77 and 3.57 ft, respectively. Increasing the recharge rate 25 percent from 7.00 to 8.75 in/yr produced somewhat smaller changes in the average arithmetic difference and in the maximum positive and negative differences. The average absolute difference was slightly higher.

The effects of decreasing the potential steady-state evapotranspiration rate from 36.0 to 27 in/yr produced a 1.01-ft increase in the average arithmetic difference and a 0.31-ft difference in the average absolute difference from the standard steady-state simulation. An increase in the evapotranspiration rate to 45 in/yr resulted in the average arithmetic difference decreasing 1.96 ft and the average absolute difference decreasing 0.69 ft.

The steady-state simulation is relatively insensitive to changes in hydraulic conductivity. In general, a 50-percent change in hydraulic conductivity produced less change in the average arithmetic and absolute differences than a 25-percent change in recharge or evapotranspiration. This sensitivity analysis indicates that the accuracy of the recharge and evapotranspiration used in the model is more important than the accuracy of the hydraulic conductivity.

SUMMARY AND CONCLUSIONS

During the Pleistocene Epoch, continental glaciation from the north and east covered eastern South Dakota and deposited a blanket of glacial drift (till and outwash) over the preglacial bedrock surface. The more sandy and gravelly outwash deposits can yield significant quantities of water to wells. The till and other unconsolidated surficial deposits serve primarily as confining beds. The bedrock directly underlying the drift generally yields little or no water to wells or is hydraulically separated from the overlying outwash by till or other silt and clay deposits.

The complex hydrologic system which exists in the glacial outwash has been subdivided into four aquifers in the study area. However, the aquifers have similar hydraulic conductivities and are hydraulically connected by zones of lower hydraulic conductivity; therefore, the aquifers have been treated as one glacial-aquifer system, rather than individual aquifers in this report.

Table 15.--Model sensitivity to changes in recharge, evapotranspiration, and hydraulic conductivity

Model simulation	Average arithmetic difference between simulated and observed water levels[1] (feet)	Average absolute difference between simulated and observed water levels[2] (feet)	Maximum positive difference between simulated and observed water levels[3] (feet)	Maximum negative difference between simulated and observed water levels[4] (feet)	Number of observation wells with observed water levels
Standard steady-state model	1.68	4.38	16.22	8.32	32
Steady-state model with maximum recharge reduced 25 percent	-1.31	4.37	14.45	11.89	32
Steady-state model with maximum recharge increased 25 percent	2.56	4.44	17.45	6.89	32
Steady-state model with maximum evapotranspiration reduced 25 percent	2.69	4.69	17.95	5.89	32
Steady-state model with maximum evapotranspiration increased 25 percent	-.28	3.69	15.45	10.89	32
Steady-state model with hydraulic conductivity reduced 50 percent	2.09	4.16	17.45	7.89	32
Steady-state model with hydraulic conductivity increased 50 percent	-.03	4.05	15.45	8.89	32

[1]Positive numbers indicate simulated head was higher than the observed head; negative numbers indicate simulated head was lower than the observed head.
[2]The absolute value of a number is the number without its associated sign. For example, the absolute value of 2 and -2 are the same.
[3]Positive difference when simulated head is greater than observed water level.
[4]Negative difference when simulated head is less than observed water level.

Thickness of the glacial-aquifer system ranges from less than 10 feet to greater than 200 feet. The lateral boundaries were placed where the aquifer thickness was less than about 5 feet. In the study area, the hydraulic conductivity ranged from 80 to 670 feet per day with an average of 316 feet per day. Because of the complexity of the glacial-aquifer system, the aquifer can be both confined and unconfined. The storage coefficients calculated from aquifer tests in the study area generally ranged from 0.00039 to 0.00017. Specific yield values as large as 0.28 were calculated for the aquifer west of the study area. Recharge to the glacial-aquifer system occurs as infiltration of precipitation and snowmelt directly into the aquifer or through the overlying till confining bed. Calculated recharge rates to the unconfined aquifer ranged from 0.9 to 3.4 inches per year and to the confined parts ranged from 0.24 to 0.72 inch per year. Evapotranspiration accounts for most of the natural discharge from the glacial-aquifer system. The average potential evapotranspiration estimated from pan evaporation is 36.2 inches per year. Little natural discharge from the aquifer system to the James River occurs in the study area. The direction of water movement in the glacial-aquifer system generally is eastward or southeastward, west of the James River and westward or northwestward, east of the James River.

In order to simulate ground-water flow within an aquifer system, a number of simplifying assumptions must be made. The simplifying assumptions for the glacial-aquifer system are: (1) The aquifer consists of one layer, (2) the overlying confining bed controls recharge to and discharge from the aquifer, (3) the silt and clay or bedrock is the lower impermeable boundary of the aquifer, (4) all lateral boundaries exhibit no flow except for 12 specified-head nodes, (5) the James River is hydraulically isolated from the aquifer except at 2 river nodes, (6) most recharge is by precipitation, and (7) most discharge is by evapotranspiration.

A grid that contains 56 rows and 52 columns of equally spaced blocks, each 1 mile wide and 1 mile long, was used to simulate the glacial-aquifer system. The aquifer was simulated prior to significant ground-water development (pre-1973) under steady-state conditions. The aquifer also was simulated in 11 annual pumping periods from 1973 through 1983 and in 12 monthly pumping periods for 1976.

The steady-state simulation represents the glacial-aquifer system prior to 1973, when the aquifer system generally was in equilibrium; that is, water levels recovered to near-prepumping levels during the nonirrigation season. The maximum available recharge to the aquifer was 7.0 inches per year and occurred only where the average thickness of the confining bed was less than 10 feet. With an average confining bed thickness between 10 and 45 feet, the rate of recharge to the aquifer decreased linearly to 0.0 inch per year. The potential evapotranspiration rate was 36.0 inches per year and can occur only where no confining bed is present above the aquifer. When the average confining bed thickness is between 0 and 45 feet, the potential evapotranspiration rate decreases linearly from 36.0 to 0.15 inches per year. The steady-state simulated water budget indicates that recharge from precipitation accounts for 97.1 percent of the water that enters the aquifer or 0.96 inch per year per active grid block, and evapotranspiration accounts for 98.2 percent of the water that leaves the aquifer or 0.97 inch per year per active grid block.

Eleven consecutive annual pumping periods from 1973 through 1983 were simulated. In 1973, water levels began to decline because of a decrease in recharge, which lasted through 1976 and because of an increase in pumping of ground water for irrigation. Recharge, evapotranspiration, and pumpage were adjusted annually. The maximum annual recharge varied from 0.10 inch in 1976 to 8.14 inches in 1977. The potential annual evapotranspiration varied from 29.9 inches in 1982 to 48.9 inches in 1976. Withdrawals from the glacial-aquifer system in 1976 were 2.6 times those in 1975. Since 1976, the pumpage has fluctuated annually in both distribution and quantity, however, the maximum annual withdrawals have not increased significantly. The annual water budget from the transient simulations varies considerably as a result of changes in recharge and evapotranspiration.

For 1976, 12 consecutive monthly pumping periods were simulated. The maximum annual recharge rate of 0.10 inch was distributed over March, April, and September. The potential monthly evapotranspiration rate ranged from 12.50 inches in August to 0.00 inch during the winter months when the ground was frozen. The simulated monthly water budgets varied considerably as a result of changes in evapotranspiration, storage, and pumpage.

Since the model is based on a number of simplifying assumptions, it cannot represent exactly the hydrologic processes in the aquifer system. The confidence in the model's response needs to be based on an objective appraisal of the analogy between the glacial-aquifer system and the model. Because the aquifer characteristics are not known with certainty, the sensitivity of the steady-state simulation to changes in recharge, evapotranspiration, and hydraulic conductivity were tested. The sensitivity analysis indicates that the model is most sensitive to reductions in recharge and least sensitive to changes in hydraulic conductivity. Since the model was insensitive to hydraulic conductivity, and recharge and discharge were widely distributed, a large range of combinations of recharge and evapotranspiration could give an equally good fit to the measured water levels. However, the values of recharge and evapotranspiration used in the model are considered to be reasonable estimates. The model is one of the best means of evaluating and improving our understanding of the aquifer system and of testing the sensitivity of various aquifer properties in the study area.

SELECTED REFERENCES

Bardwell, Lawrence, 1984, Water quality of South Dakota's glacial aquifers: South Dakota Department of Water and Natural Resources, 52 p.

Benson, R.D., 1983, A preliminary assessment of the hydrologic characteristics of the James River in South Dakota: U.S. Geological Survey Water-Resources Investigations Report 83-4077, 114 p.

Driessen, J.L., 1980, Soil survey of Sanborn County: U.S. Department of Agriculture Soil Conservation Service, 191 p.

Dyman, T.S., and Barari, Assad, 1976, Ground-water investigation for the city of Wolsey: South Dakota Geological Survey Special Report 63, 19 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.

Hamilton, L.J., 1980, Major aquifers in Aurora and Jerauld Counties, South Dakota: South Dakota Geological Survey Information Pamphlet No. 23, 5 p.

----- 1985, Water resources of Aurora and Jerauld Counties, South Dakota: U.S. Geological Survey Water-Resources Investigations Report 84-4030, 58 p.

Hansen, D.S., 1983, Water resources of Hanson and Davison Counties, South Dakota: U.S. Geological Survey Water-Resources Investigations Report 83-4108, 55 p.

Heath, R.C., 1983, Basic ground-water hydrology: U.S. Geological Survey Water-Supply Paper 2220, 84 p.

Hedges, L.S., 1968, Geology and water resources of Beadle County, South Dakota, part 1, geology: South Dakota Geological Survey Bulletin 18, 66 p.

Hedges, L.S., Allen, Johnette, and Holly, D.E., 1983, Evaluation of ground-water resources, eastern South Dakota and the upper Big Sioux River South Dakota and Iowa; Task 7, ground water recharge: South Dakota Geological Survey, 56 p.

Hedges, L.S., Burch, S.L., Iles, D.L., Barari, R.A., and Schoon, R.A., 1981, Evaluation of ground-water resources eastern South Dakota and upper Big Sioux River, South Dakota and Iowa: South Dakota Geological Survey.

Heil, D.M., 1979, Soil survey of Beadle County, South Dakota: U.S. Department of Agriculture Soil Conservation Service, 169 p.

Howells, L.W., and Stephens, J.C., 1968, Geology and water resources of Beadle County South Dakota, part II, water resources: South Dakota Geological Survey Bulletin 18, 63 p.

Iles, D.L., 1979a, Ground-water study for the city of Huron: South Dakota Geological Survey Open-File Report No. 24-UR, 37 p.

----- 1979b, Sanitary landfill investigations for the city of Huron: South Dakota Geological Survey Open-File Report No. 26-UR, 30 p.

Koch, N.C., 1980a, Geology and water resources of Hand and Hyde Counties, South Dakota, part II, water resources: South Dakota Geological Survey Bulletin 28, 46 p.

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

Koch, N.C., and McGarvie, S.D., 1988, Water resources of Miner County, South Dakota: U.S. Geological Survey Water-Resources Investigations Report 86-4035, 37 p.

Koglin, E.N., Stack-Goodman, Kathleen, and Ambroson, J.R., 1981, Test hole logs and well logs of Beadle County, volumes 1 and 2: South Dakota Geological Survey Open-File Report No. 4-CS, 1,217 p.

Kuiper, L.K., 1984, Appraisal of the water resources of the eastern part of the Tulare aquifer, Beadle, Hand, and Spink Counties, South Dakota: U.S. Geological Survey Water-Resources Investigations Report 84-4078, 52 p.

McDonald, M.G., and Harbaugh, A.W., 1984, A modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Open-File Report 83-875, 528 p.

McGarvie, S.D., 1983, Major aquifers in Miner County, South Dakota: South Dakota Geological Survey Information Pamphlet no. 20, 10 p.

Norris, S.E., 1962, Permeability of glacial till: U.S. Geological Survey Research 1962, p. E150-E150.

Schoon, R.A., 1971, Geology and hydrology of the Dakota Formation in South Dakota: South Dakota Geological Survey Report of Investigations No. 104, 55 p.

Schroeder, Wayne, 1982, Test hole logs and well logs of Miner County: South Dakota Geological Survey Open-File Report No. 5-CS, 719 p.

Spuhler, Walter, Lytle, W.F., and Moe, Dennis, 1971, Climate of South Dakota: South Dakota State University Agricultural Experiment Station Bulletin 582, 5 p.

Steece, F.V., and Howells, L.W., 1965, Geology and ground water supplies in Sanborn County, South Dakota: South Dakota Geological Survey Bulletin 17, 182 p.

----- 1969, Geology and ground water supplies in Sanborn County South Dakota, part II, ground water basic data: South Dakota Geological Survey Bulletin 17, 200 p.

Walker, I.R., 1961, Shallow outwash deposits in the Huron-Wolsey area Beadle County, South Dakota: South Dakota Geological Survey Report of Investigations No. 91, 44 p.

U.S. Army Corps of Engineers, 1951, Time lag and soil permeability in ground-water observations: U.S. Army Corps of Engineers Bulletin No. 36, 50 p.

U.S. Department of Commerce, 1956-83, Climatological data for South Dakota (issued annually).