A DIGITAL SIMULATION OF THE GLACIAL-AQUIFER SYSTEM IN THE NORTHERN THREE-FOURTHS OF BROWN COUNTY, SOUTH DAKOTA

By Patrick J. Emmons

U.S. GEOLOGICAL SURVEY

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DEPARTMENT OF THE INTERIOR

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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	4,047	square meter
acre-foot per year (acre-ft/yr)	1,234	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
foot per mile (ft/mi)	0.1894	meter per kilometer
inch	25.4	millimeter
inch per year (in/yr)	25.4	millimeter per year
gallon per minute (gal/min)	0.06308	liter per second
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer

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

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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 intensively irrigated areas of the James River basin in eastern South Dakota. These increases in demand for irrigation water from the glacial-aquifer system and concern about the adequacy of the ground-water resources for future development have created a need for a systematic water-management program to avoid overdevelopment of these aquifers in the James River basin.

A complex hydrologic system exists in the glacial drift overlying the bedrock in Brown County. This system has been subdivided in descending order into three aquifers: the Elm (aquifer layer 1), Middle James (aquifer layer 2), and Deep James (aquifer layer 3) aquifers. These sand and gravel outwash aquifers generally are separated from each other by till and other fine-grained sediments. Fine-grained lake deposits, which were deposited on the bed of ancient Lake Dakota, are not an important aquifer but commonly provide recharge to and discharge from the Elm and Middle James aquifers.

These three aquifers were simulated under steady-state and transient conditions using the U.S. Geological Survey's modular, three-dimensional, finite-difference, ground-water flow model program. The simulations were used to help understand flow in the glacial-aquifer system. An equally spaced grid containing 86 rows and 70 columns was used to simulate the glacial-aquifer system. The steady-state simulation represents the system under equilibrium conditions. The maximum available steady-state recharge to the aquifer was 7.0 inches per year and the maximum potential evapotranspiration was 35.4 inches per year. The thickness of the confining bed overlying the uppermost active aquifer controls the actual amount of recharge or evapotranspiration which can occur in each model grid block. The simulated steady-state potentiometic heads were compared to the average water levels from observation wells in aquifer layer 1 and aquifer layer 2 as a means of checking the accuracy of the simulations. The average differences were 0.78 feet for aquifer layer 1 and 3.49 feet for aquifer layer 2. The average absolute difference was 4.59 feet for aquifer layer 1 and 5.10 feet for aquifer layer 2. There were no water-level data available to check the accuracy of aquifer layer 3. The steady-state simulated water budget indicates recharge from precipitation accounts for 94.8 percent of the water entering the system and evapotranspiration accounts for 95.8 percent of the water leaving the system. The sensitivity analysis of the steady-state model indicates that it is most sensitive to reductions in recharge and least sensitive to changes in hydraulic conductivity.

In the monthly transient simulations for 1985, recharge, evapotranspiration, and pumpage were varied monthly. The maximum monthly recharge ranged from zero during the winter months to 2.51 inches in May. The maximum potential evapotranspiration ranged from zero during the winter months to 7.03 inches in July. The average monthly difference between the simulated and observed water levels ranged from -2.54 feet in July to 1.48 feet in January for aquifer layer 1 and from -1.22 feet in April to 4.98 feet in October for aquifer layer 2. The average absolute differences for aquifer layer 1 ranged from 4.16 feet in September to 6.31 feet in February and ranged from 3.96 feet in April to 8.23 feet in August for aquifer layer 2. The water levels for the monthly transient simulations varied considerably as a result of changes in recharge, 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 intensively irrigated 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, written commun., 1973) to 35,422 acre-ft/yr (South Dakota Department of Water and Natural Resources, written commun., 1981), an increase of greater than 600 percent. These increases in demand for irrigation water from the glacial-aquifer system and concern about the adequacy of ground-water resources for future development have created a need for a systematic water-management program to avoid overdevelopment of these aquifers in the James River basin.

In 1983, the South Dakota Department of Water and Natural Resources and the City of Aberdeen entered into a cooperative agreement with the U.S. Geological Survey to define the flow system of the glacial-aquifer system in the northern three-fourths of Brown County (fig. 1). More specifically, the study will improve definition of the glacial-aquifer boundaries; determine the aquifer thicknesses, 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 investigation of the glacial-aquifer system in the northern three-fourths of Brown County using a three-dimensional ground-water flow model and describes the design and calibration of that model.

The scope of this investigation included the collation and synthesis of aquifer-test data, 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 transmissivity, hydraulic conductivity, and storage coefficient of the aquifers. Well and test-hole data for Brown, Marshall, and Day 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. Where existing data were inadequate, the South Dakota Geological Survey drilled 32 additional test holes. Water-level data that 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. During 1984 and 1985, the U.S. Geological Survey measured 47 wells to provide additional

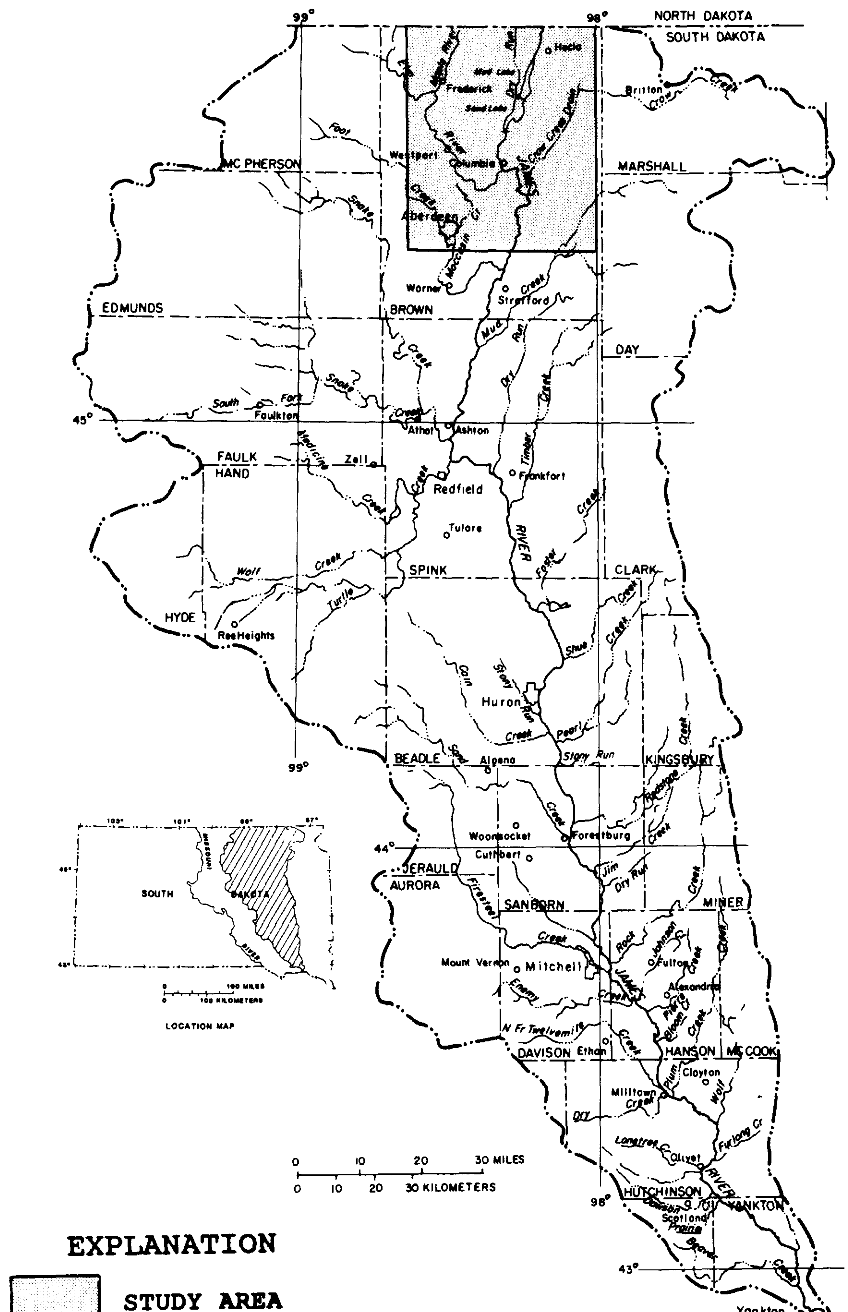


Figure 1.--Location of study area.

water-level data where it was lacking or insufficient. The South Dakota Department of Water and Natural Resources provided pumpage data that were used to determine the magnitude of the stress being applied to the aquifer system.

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 glaciers from the north and east covered eastern South Dakota, depositing a blanket of glacial drift over the eroded preglacial bedrock surface. The glaciers radically altered the topography by partially filling major valleys, entirely obliterating many small valleys, scouring new valleys, and forming massive end moraines. The overall effect of glaciation has been to lower 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 River basin is a lowland of low to moderate relief trending northward between the Coteau du Missouri highlands to the west and the Coteau des Prairies highlands to the east. The basin is 50 to 75 mi wide and about 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).

Most of the surficial deposits in the study area (fig. 1) are the result of glaciation and collectively are called drift, which is any material deposited by or from a glacier. Drift in Brown County can be subdivided into three major types--till, outwash, and lake deposits--that differ greatly in physical and hydraulic characteristics. Till, which was deposited directly from or by glacial ice, is a heterogeneous mixture of clay, silt, sand, and gravel. 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 sand and gravel but may contain silt and clay and interbedded layers of sandy or gravelly silt and clay. Beds of well-sorted sand and gravel are contained in the outwash but generally are small and discontinuous (Howells and Stephens, 1968). Leap (1986) subdivided the glacial outwash into three major types of deposits--surface deposits, intratill deposits, and buried meltwater channel deposits. The buried meltwater channel deposits, primarily of proglacial origin, were further subdivided into three different levels--the lowermost or basal outwash, middle outwash, and the upper buried outwash.

When the glacial ice sheet of Wisconsin age melted back into North Dakota, meltwater flowing from it accumulated in a shallow depression, glacial Lake Dakota which includes about the eastern two-thirds of Brown County. The area is a distinct physiographic unit known as the Lake Dakota plain. The lake deposits consist generally of silt, however, in a large area in northeastern Brown County, the silt is overlaid by lake deposits of fine to medium-grained sand (Leap, 1986). Koch and Bradford (1976) classified the lake deposits as very fine to fine-grained sand.

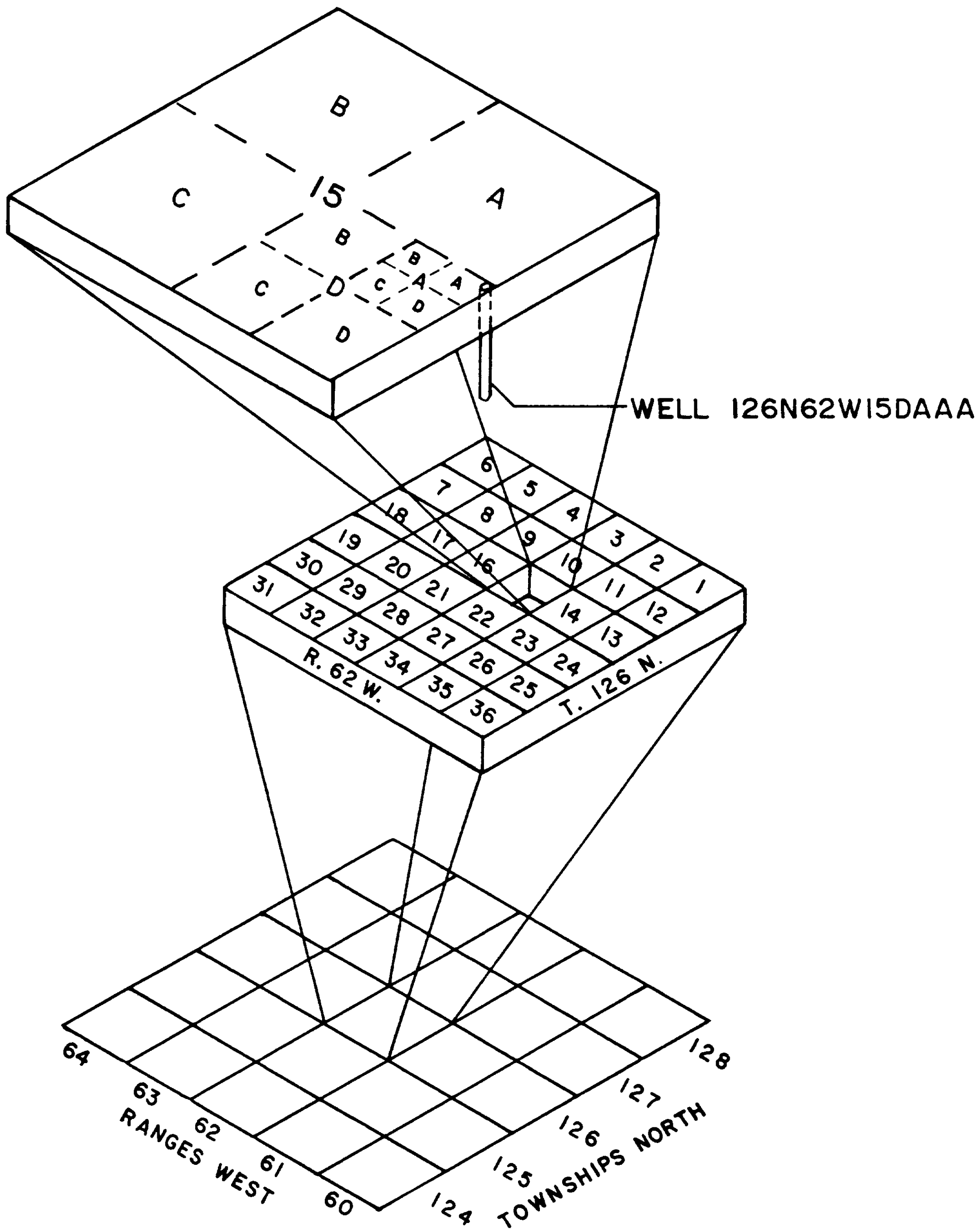


Figure 2.--Site-numbering system. The well number consists of township followed by "N," range followed by "W," and section number, followed by a maximum of four uppercase letters that indicate, respectively, the 160-, 40-, 10-, and 2½-acre tract in which the well is located. These letters are assigned in a counterclockwise direction beginning with "A" in the northeast quarter. Thus, well 126N62W15DAAA is the well recorded in the NE¼ of the NE¼ of the NE¼ of the SE¼ of section 15 in township 126 north and range 62 west of the 5th meridian and baseline system.

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, especially on the Lake Dakota plain. The alluvium, which consists of poorly sorted, poorly stratified, thin, discontinuous layers of material, ranges in size from clay to boulders, but usually has a large silt content and does not contain significant sand and gravel.

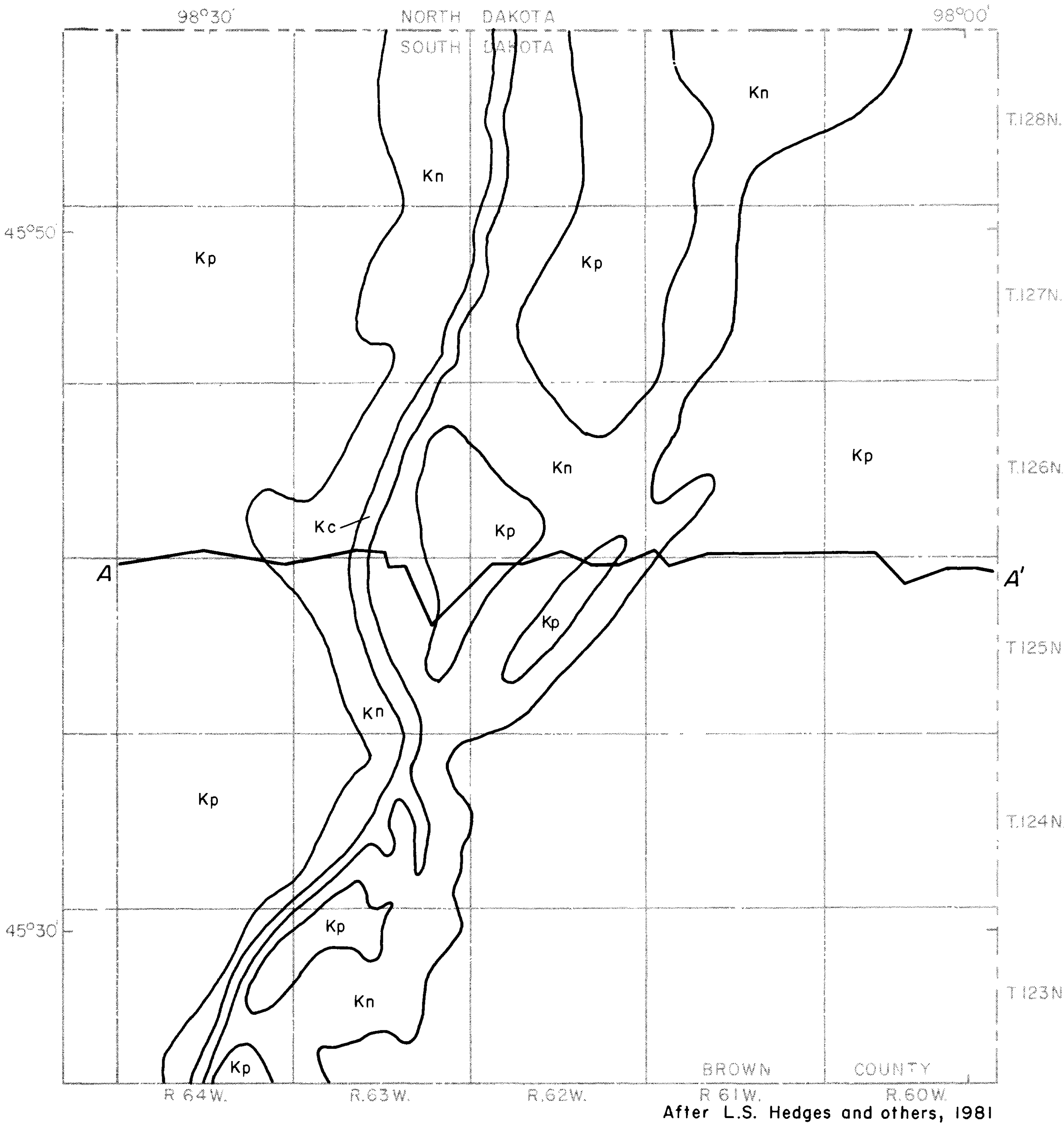
The bedrock directly underlying the drift in the study area, in descending order, consists of the Upper Cretaceous Pierre Shale, Niobrara Formation, and Carlile Shale (fig. 3). The Pierre Shale is predominantly a dark-gray, fissile, bentonitic clay-shale. The Pierre Shale has a maximum thickness of 320 ft in Brown County and is exposed along stream channels in the western part of the county. The Pierre Shale is believed to be conformable with the underlying Niobrara Formation (Leap, 1986). The Niobrara Formation is predominantly a light- to dark-gray speckled marl or calcareous clay with some "chalk" and shaly beds. The marl contains shells of foraminifera that impart the distinctive, white-speckled appearance. The Niobrara is not exposed on the surface but is found in subcrops beneath the drift where the Pierre Shale is absent. According to Hedges and others (1983), the Carlile Shale directly underlies the drift in a narrow band trending roughly north-south through the study area. This band of Carlile Shale is located in the preglacial Grand-Moreau-Cheyenne River channel (Leap, 1986). The Carlile consists mostly of light-gray to black shale containing silty and sandy zones. The maximum thickness of the Carlile Shale in Brown County is 275 ft.

Hydrologic Setting

Ground water is a major source of water for irrigation, municipal, farm, and domestic use in the James River basin. In the unconsolidated surficial deposits, only the more sandy and gravelly glacial-outwash deposits yield substantial quantities 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 natural 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 may confine the outwash aquifers. In the study area, the till, the Pierre Shale, Niobrara Formation, and Carlile Shale generally yield little or no water to wells and are considered to be confining beds.

The units that comprise the complex hydrologic system in the glacial outwash have been subdivided into three aquifers in Brown County by Koch and Bradford (1976). They are the Elm, Middle James, and Deep James aquifers. The Elm, Middle James, and Deep James aquifers equate to Leap's (1986) three levels of buried meltwater channel deposits: the uppermost buried outwash, the middle outwash, and the lowermost or basal outwash, respectively. The topographic and stratigraphic relations of these aquifers are shown in the geohydrologic section in figure 4. Koch and Bradford (1976) defined the Elm, Middle James, and Deep James aquifers based on altitude. The maximum altitude of the top of the Elm aquifer is 1,400 ft and the minimum altitude of the bottom is 1,225 ft. The maximum altitude of the top of the Middle James aquifer is 1,250 ft and the minimum altitude of the bottom is 1,150 ft. The maximum altitude of the top of the Deep James aquifer is 1,175 ft and the minimum altitude of the bottom is 950 ft.



EXPLANATION

Kp PIERRE SHALE
Kn NIOBRARA FORMATION
Kc CARLILE SHALE
A—A' LINE OF GEOLOGIC SECTION

Figure 3.--Bedrock geology.

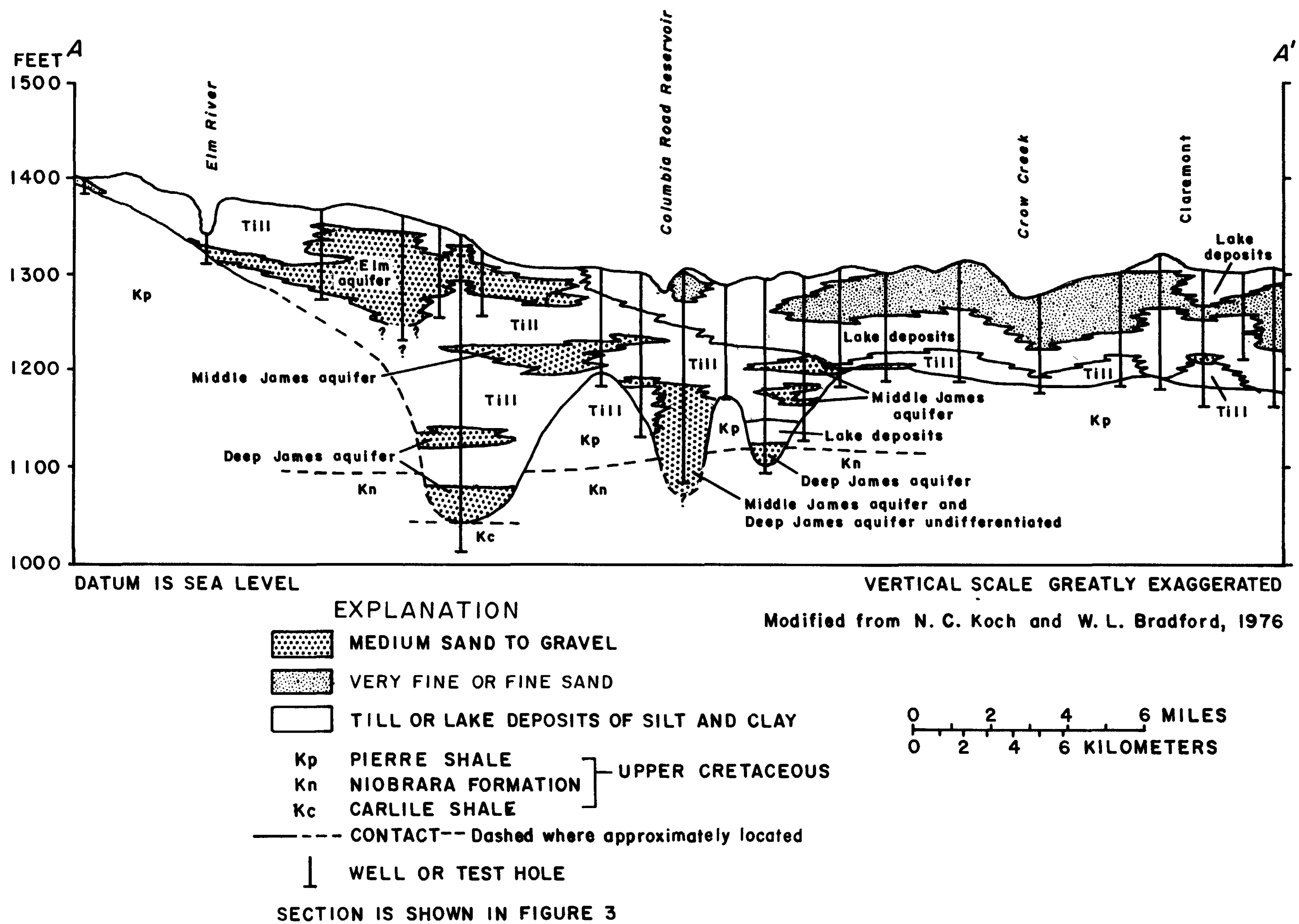


Figure 4.--Typical geohydrologic section showing the relation between the glacial aquifers and confining beds.

The three glacial-outwash aquifers generally are separated from each other by till, as shown in figure 4, and may be internally separated by till and thin clay and silt outwash layers. The till and thin clay and silt outwash layers allow some flow to occur between and within the aquifers.

Sandy Lake Deposits

The Lake Dakota plain covers much of the study area. Glacial meltwaters deposited an average of about 75 ft of fine sand, silt, and clay on the bed of ancient Lake Dakota. Figure 5 shows the average thickness and extent of the sandy lake deposits. Wells completed in the sandy lake deposits may yield 1 to 5 gal/min, but well failure is common because of clogging of the well screens by fine-grained sediments. These sediments commonly pass through the well screen and enter the water system, not only clogging the well but abrading and seriously damaging pumps and other equipment (Koch and Bradford, 1976). Because of the low yield potential of the sandy lake deposits compared to the glacial outwash, the lake deposits are considered a minor aquifer in the study area; however, they commonly provide recharge to and accept discharge from the Elm and Middle James aquifers, as well as provide for interaquifer flow.

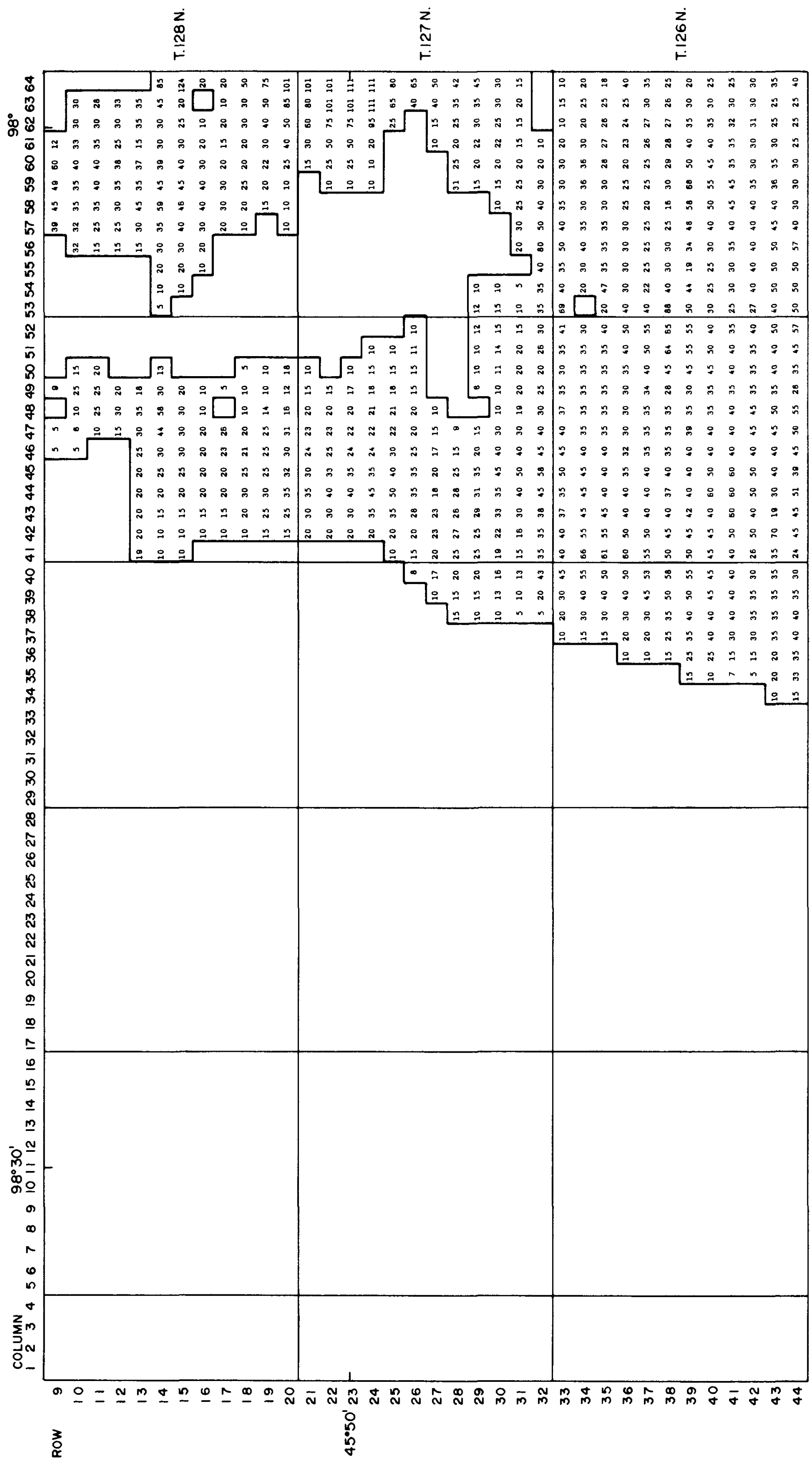
Elm Aquifer

The Elm aquifer is the uppermost and largest sand-and-gravel outwash aquifer in the glacial-aquifer system (figs. 4 and 6) in the study area. Because of the availability of additional well logs, reinterpretation of existing logs, and the averaging of the aquifer thicknesses, the boundaries shown in figure 6 may differ from those of Koch and Bradford (1976). The aquifer underlies about 351 mi^2 of the study area. The thickness ranges from zero at the boundaries to 113 ft in 127N62W18A and averages about 32 ft. The aquifer slopes to the east at about the same gradient as the topographic surface, about 15 ft/mi.

The water in the aquifer is under water-table (unconfined) conditions in some places and under artesian (confined) conditions where the confining bed overlying the aquifer is sufficiently thick. Emmons (1988), in developing a ground-water flow model of the glacial-aquifer system in the Sanborn-Beadle County area (fig. 1), estimated that a confining-bed thickness of 10 ft or greater overlying the aquifer probably is sufficient to confine the aquifer, causing artesian conditions. Even in areas where the aquifer is under water-table conditions, silt and clay layers within the aquifer may confine its lower parts. The thickness of the confining bed overlying the Elm aquifer or sandy lake deposits is shown in figure 7. The thickness of confining bed overlying the uppermost aquifer averages about 20 ft.

Recharge to the Elm aquifer is by infiltration of precipitation and snowmelt, and possibly by leakage from the Elm River and Foot Creek during periods of high flow directly into the aquifer or by percolation through the overlying lake sediments and till. Recharge occurs more rapidly in level areas where the aquifer is at or near land surface or where more permeable sandy lake deposits overlie the aquifer. Recharge occurs more slowly where the aquifer is overlaid by less permeable clayey or silty lake deposits or till.

Hydrographs of water levels from two observation wells completed in the Elm aquifer are shown in figure 8. Examination of the hydrographs indicates that no long-term water-level declines have occurred in the aquifer, although seasonal and year-to-year changes have occurred because of variations in available recharge. Koch and Bradford (1976) determined that the water level in the Elm aquifer varied in direct response to snowmelt, rainfall, or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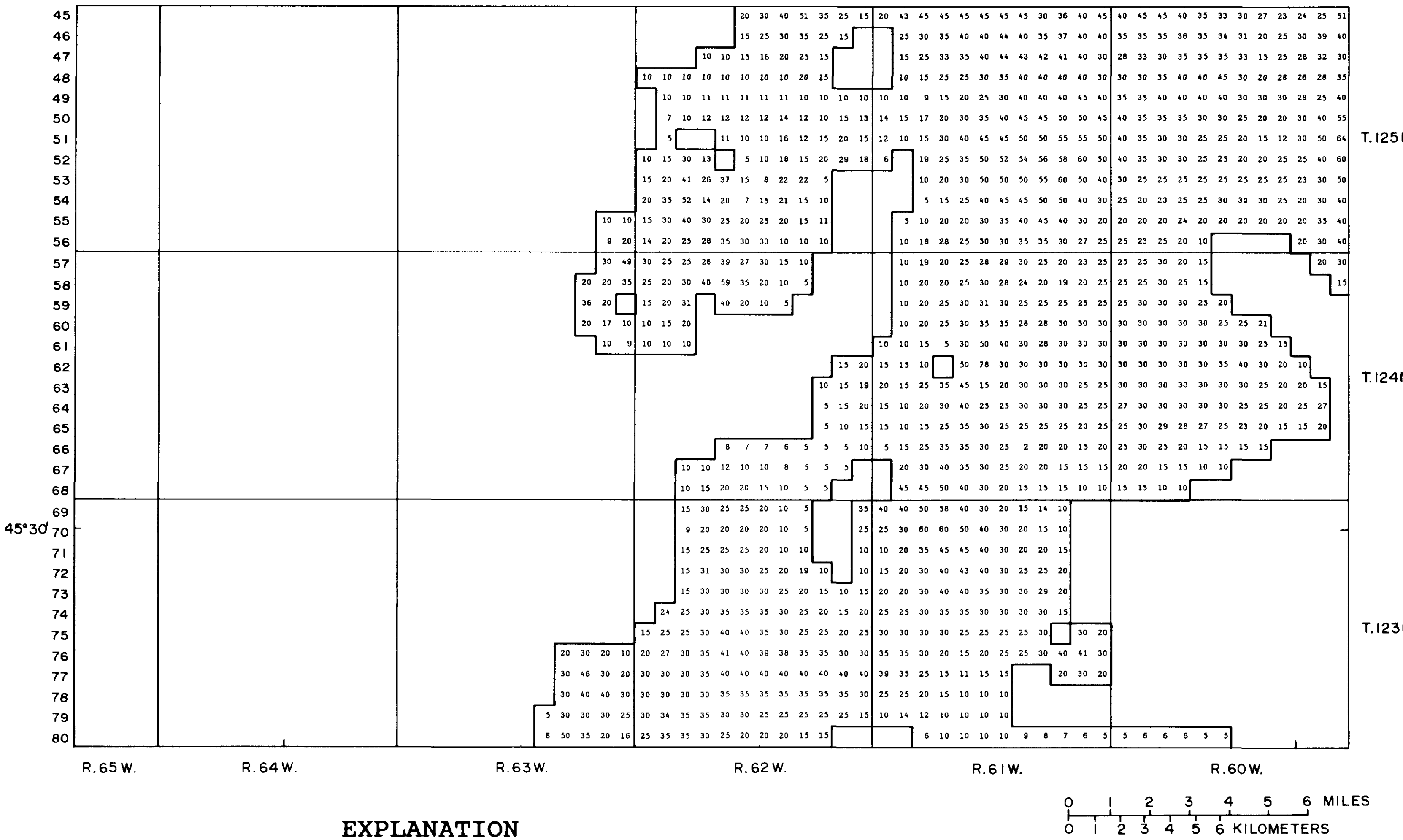
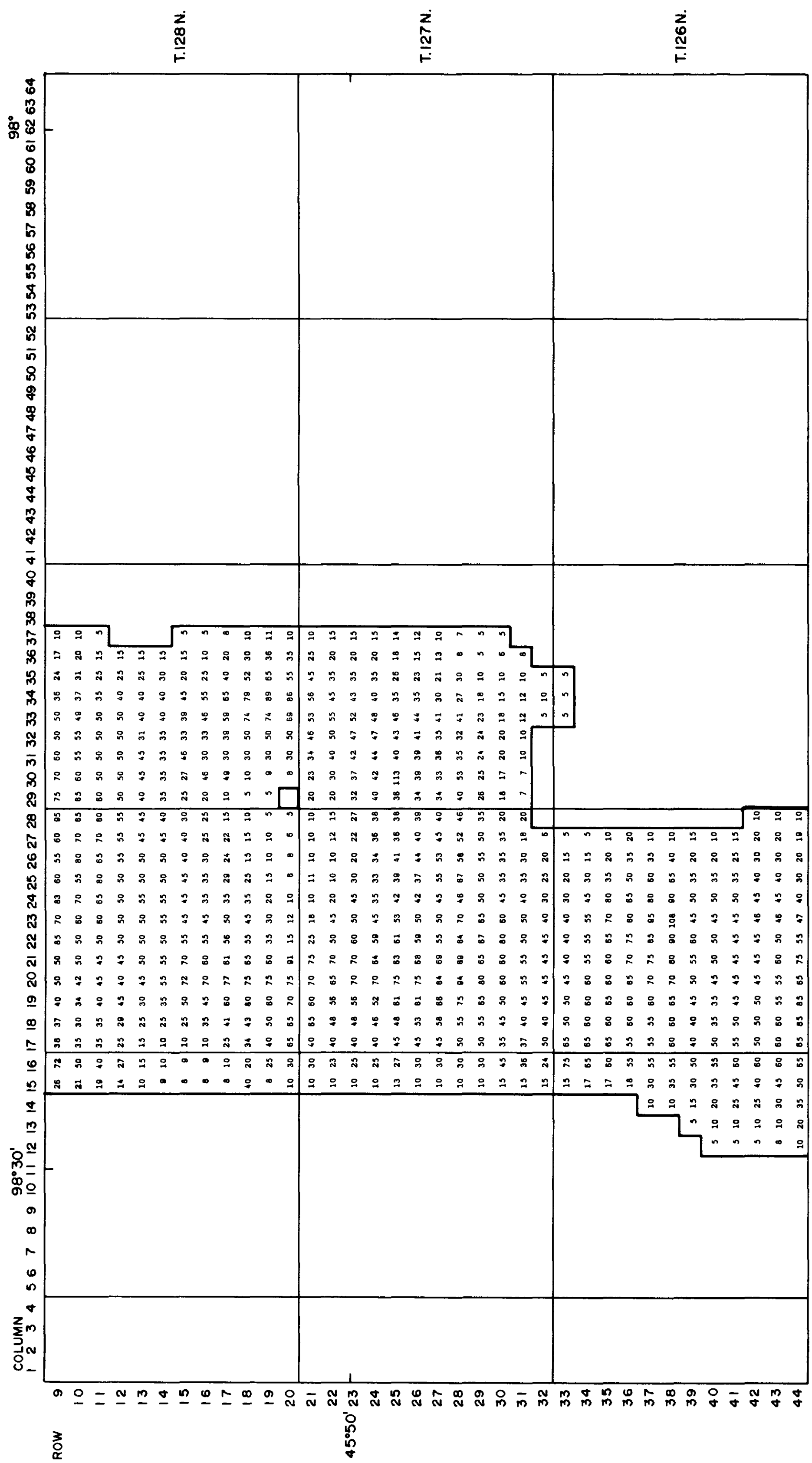


EXPLANATION

25 AQUIFER THICKNESS--Number is the average value for each grid block (0.25 square mile), in feet

—— AQUIFER MODEL BOUNDARY

Figure 5.--Average sand thickness of the sandy lake deposits for each grid block.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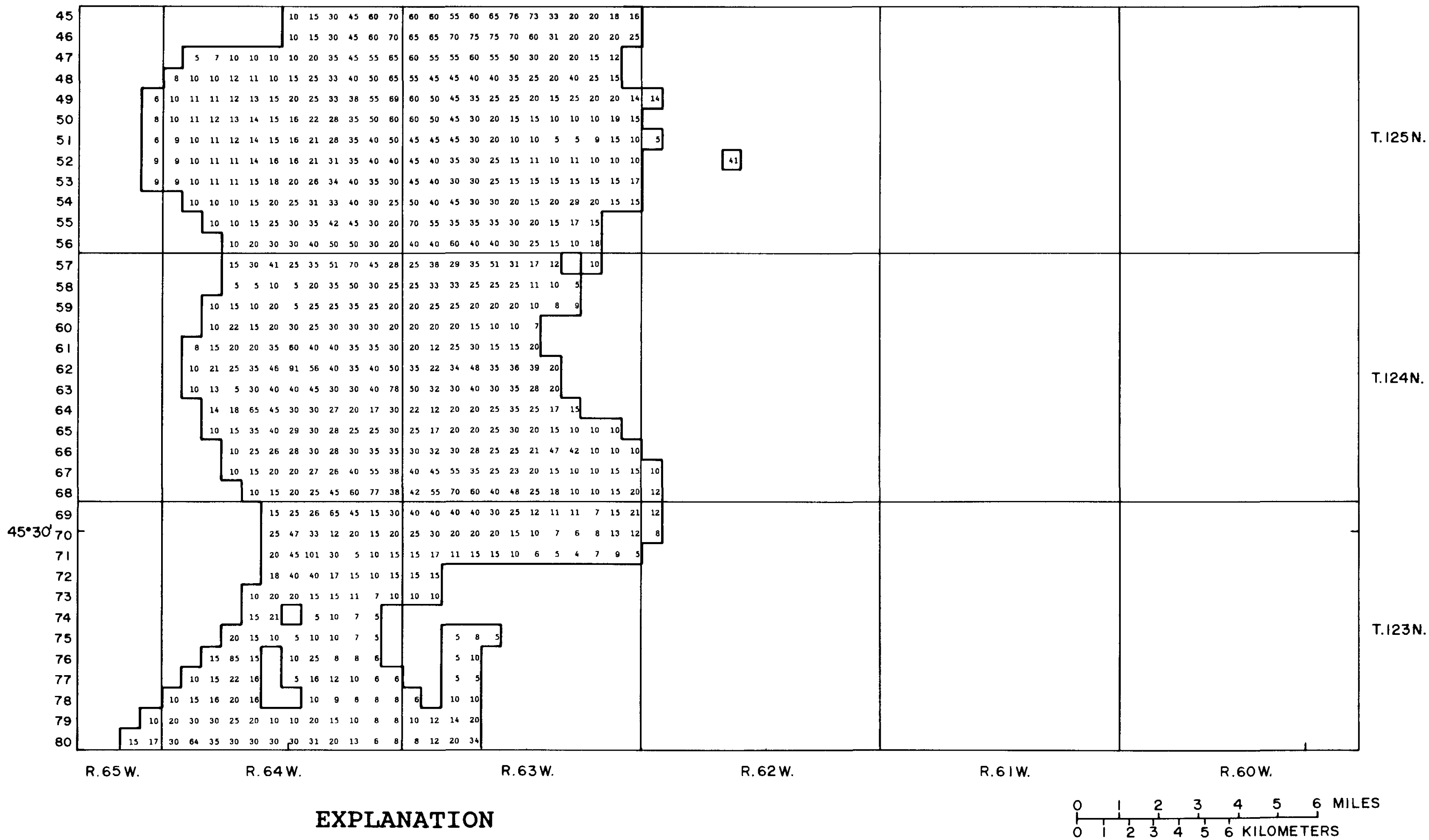
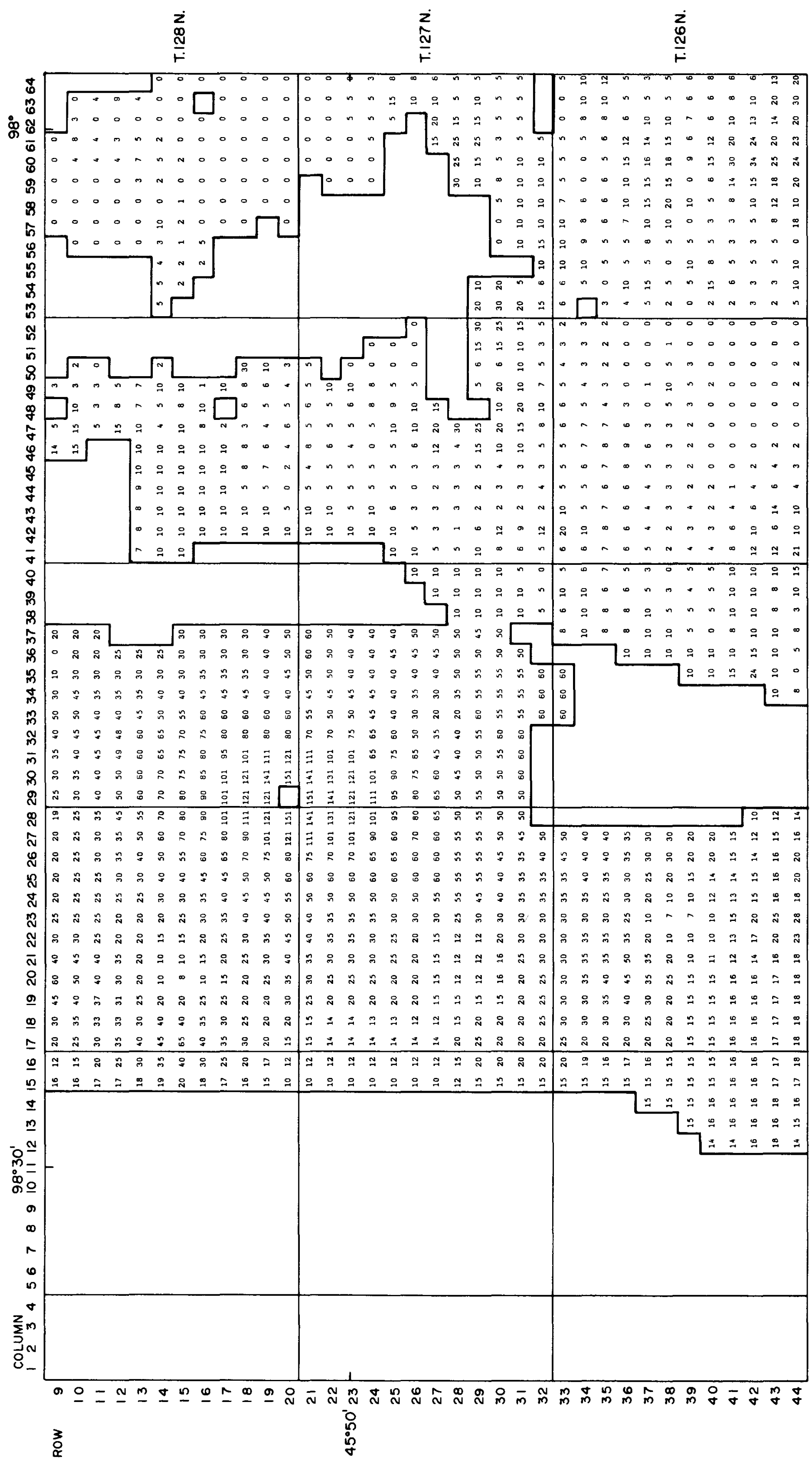


EXPLANATION

25 AQUIFER THICKNESS--Number is the average value for each grid block (0.25 square mile), in feet

—— AQUIFER MODEL BOUNDARY

Figure 6.--Average thickness of the Elm aquifer for each grid block.



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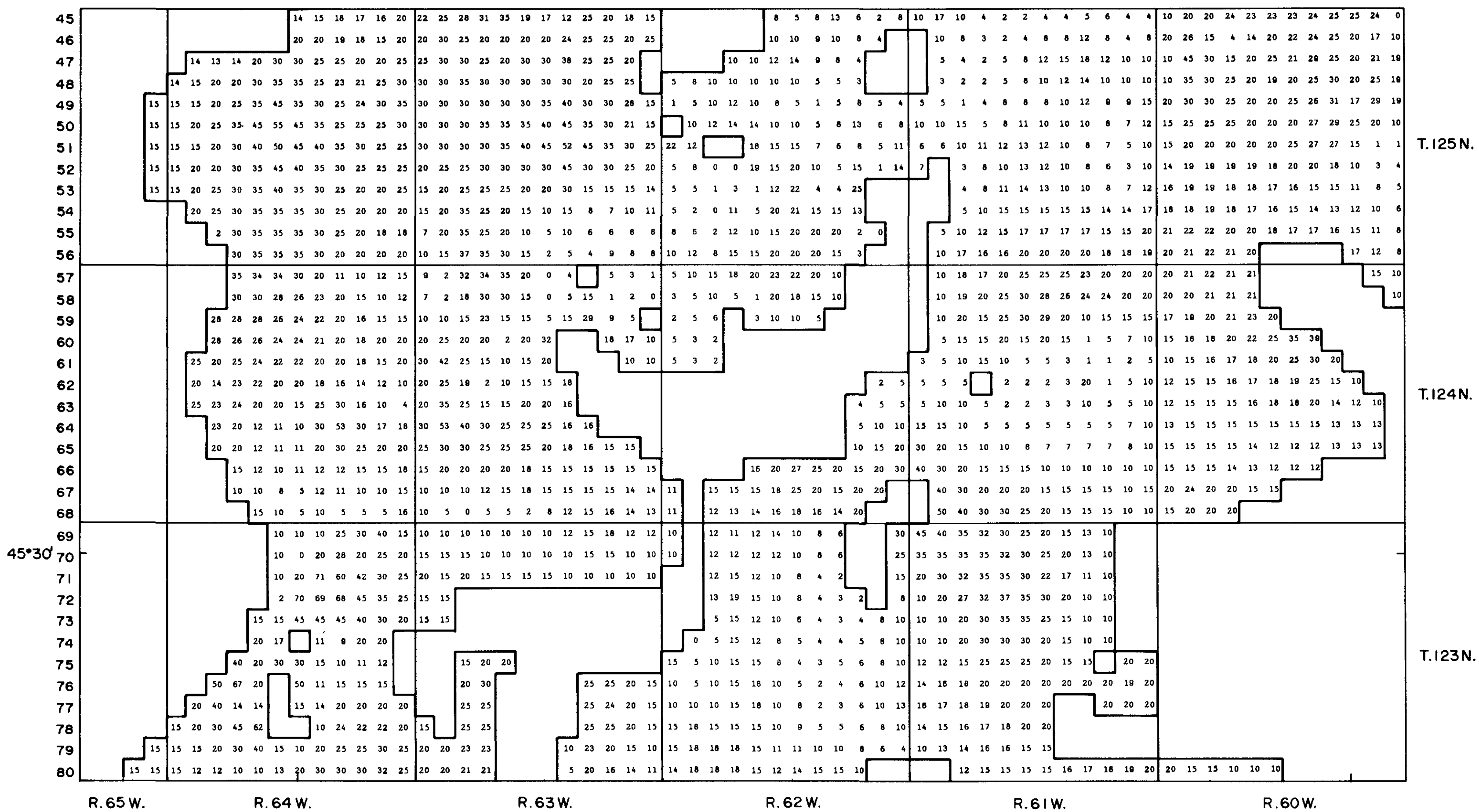


0 1 2 3 4 5 6 MILES
0 1 2 3 4 5 6 KILOMETERS

EXPLANATION

10 THICKNESS OF CONFINING BED--Number is the average value for each grid block (0.25 square mile), in feet

—— AQUIFER OR SANDY LAKE DEPOSIT MODEL BOUNDARY

Figure 7.--Average thickness of confining beds overlying the Elm aquifer and the sandy lake deposits for each grid block.

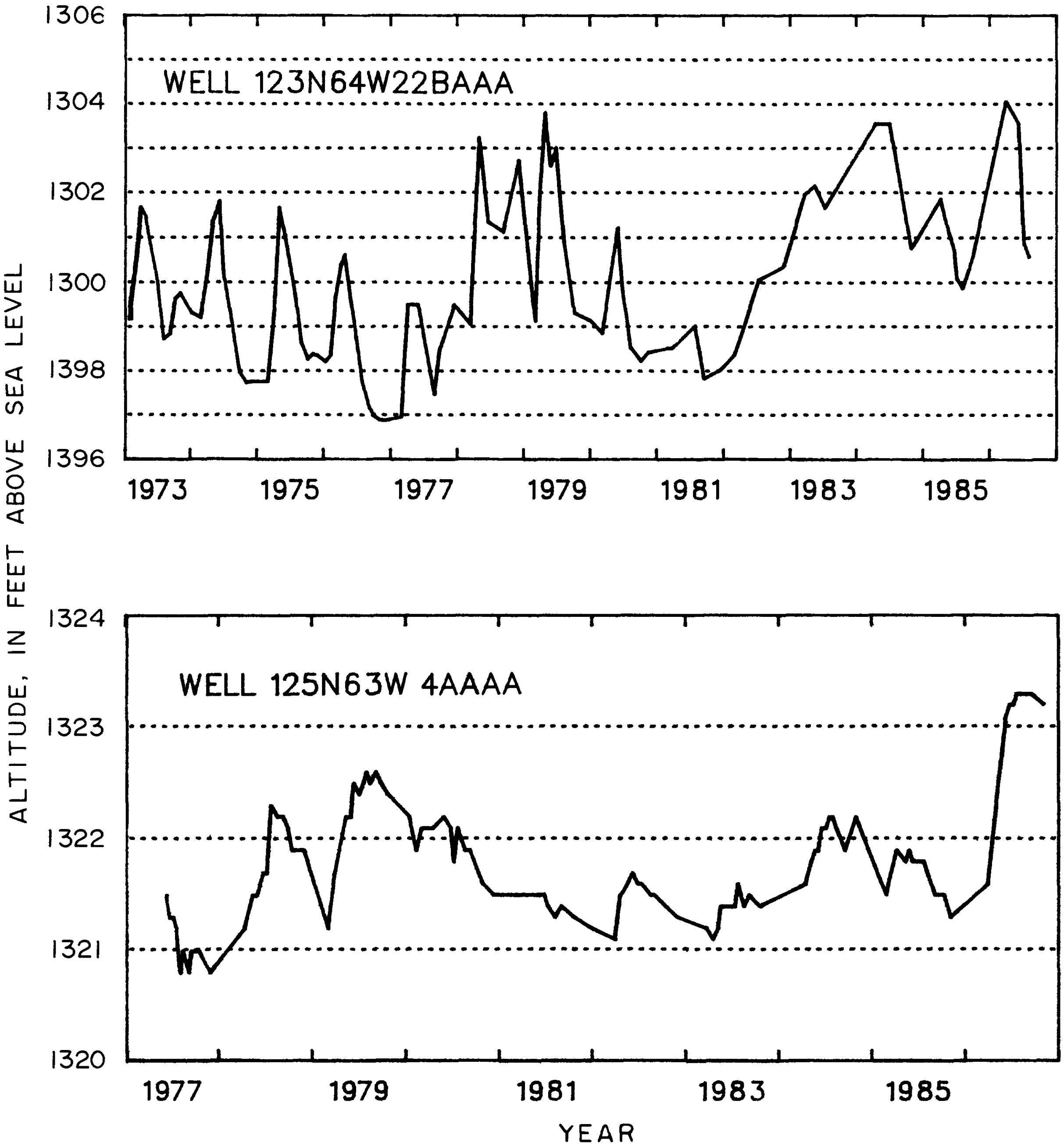


Figure 8.--Water-level hydrographs for selected wells completed in the Elm aquifer.

drought with only a short time lag. They also compared long-term water-level data and cumulative departure of precipitation from normal which showed that even though precipitation was more than 25 inches less than normal from 1950 to 1972, water levels in the Elm aquifer remained about the same.

Natural discharge occurs as evapotranspiration, eastward flow into the sandy lake deposits underlying the Lake Dakota plain, recharge into the Middle James aquifer, and minor seepage into the Elm River and Foot Creek during periods of low flow. The Elm aquifer does not appear to be in hydraulic connection with the James River. According to Koch and Bradford (1976), the general direction of water movement in the aquifer is to the southeast at a gradient of about 10 ft/mi. Discharge from the aquifer also occurs by pumpage from wells and as seepage into gravel pits that penetrate the aquifer. Most of the wells completed in the Elm aquifer are small-yield domestic, stock, and farm wells. In addition, currently (1986) about 120 ft^3/s of irrigation and municipal pumpage is permitted from the glacial-aquifer system from about 55 permits in the study area according to unpublished data provided by the South Dakota Department of Water and Natural Resources. The 55 locations are either single wells, multiple wells on a single permit, or gravel pits. These wells may yield water from the Elm aquifer, the Middle James aquifer, the Deep James aquifer, or any combination. According to Koch and Bradford (1976), wells can be constructed to yield 500 gal/min or more where at least 40 ft of medium-grained sand is present. For coarser material, a lesser thickness is needed.

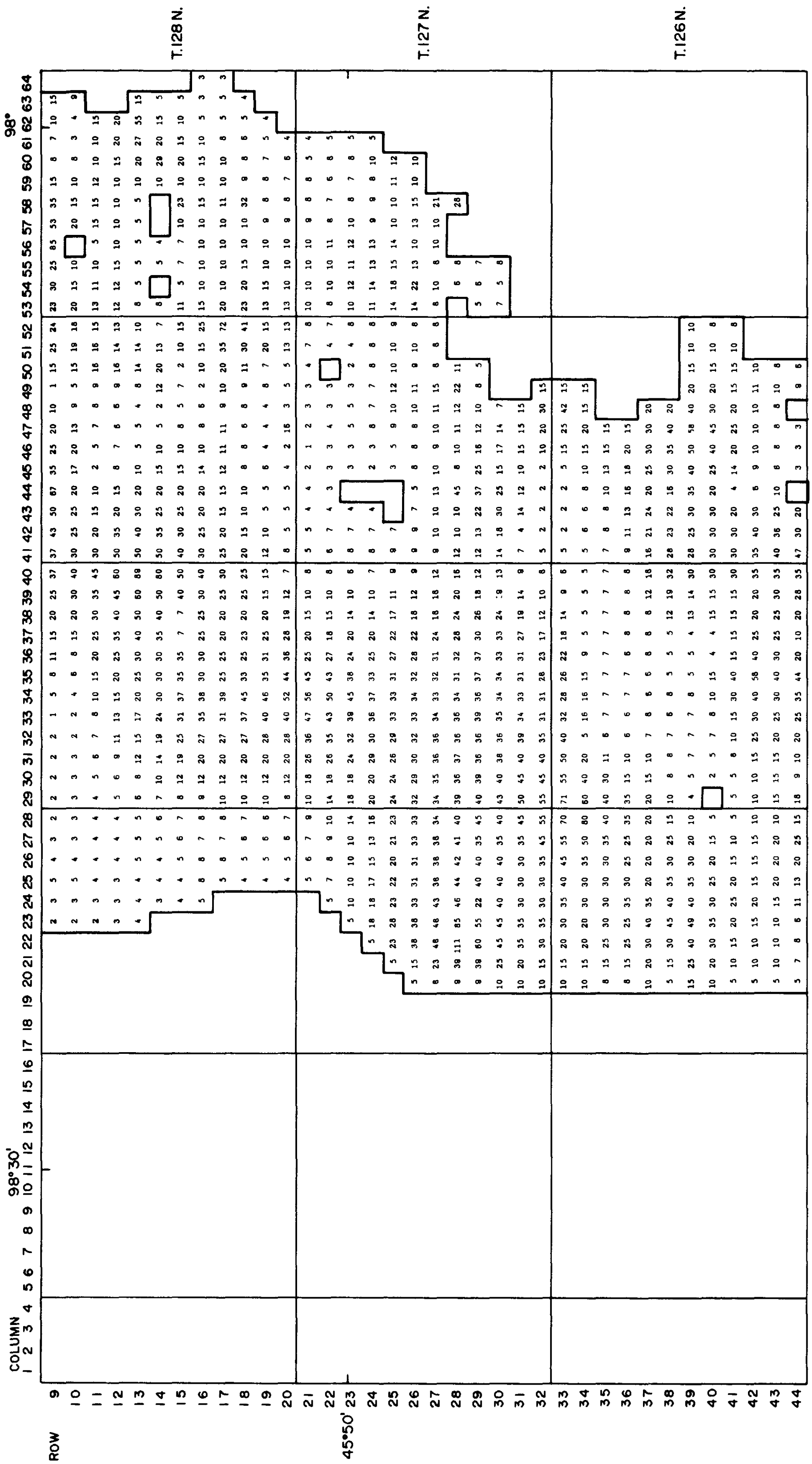
Middle James Aquifer

The Middle James aquifer stratigraphically underlies the Elm aquifer and covers about 500 mi^2 of the study area (fig. 9). The boundaries shown in figure 9 may differ from those of Koch and Bradford (1976) due to the availability of additional well logs, reinterpretation of existing logs, and the averaging of aquifer thicknesses. The thickness ranges from zero at the boundaries to 111 ft in 123N64W21C and 127N63W21D and averages about 17 ft. The aquifer is lenticular and contains many clay and silt layers.

Water in the Middle James aquifer generally is under artesian conditions except where the overlying confining bed is less than about 10 ft thick. Even in areas where the upper part of the aquifer is under water-table conditions, the intervening clay and silt layers commonly confine the lower parts of the aquifer.

Recharge to the Middle James aquifer is from the Elm aquifer and from percolation of snowmelt and precipitation through the overlying lake deposits and till. Probably the largest source of recharge water is from the overlying Elm aquifer where it is in contact with the Middle James aquifer (Koch and Bradford, 1976).

Examination of hydrographs of the water levels measured in two observation wells completed in the Middle James aquifer (fig. 10) indicates no long-term water-level declines have developed. Seasonal and annual fluctuations occur because of variations in recharge, evapotranspiration, and ground-water withdrawal due to pumping. Koch and Bradford (1976) determined that water levels in the Middle James aquifer have remained about the same from 1950 to 1972 even though precipitation totaled 25 inches less than normal for that period.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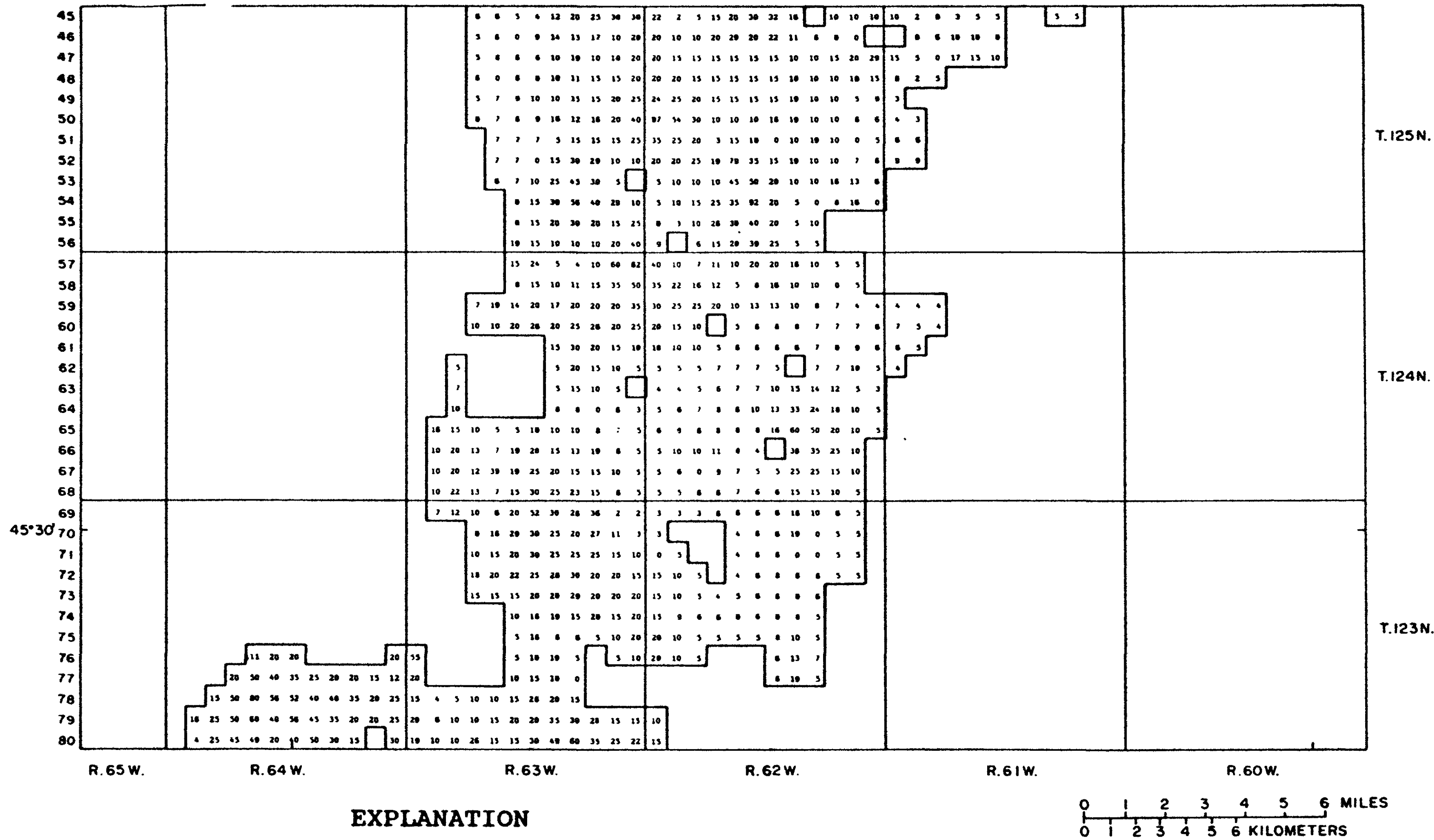


EXPLANATION

25 AQUIFER THICKNESS--Number is the average value for each grid block (0.25 square mile), in feet

—— AQUIFER MODEL BOUNDARY

Figure 9.--Average thickness of the Middle James aquifer for each grid block.

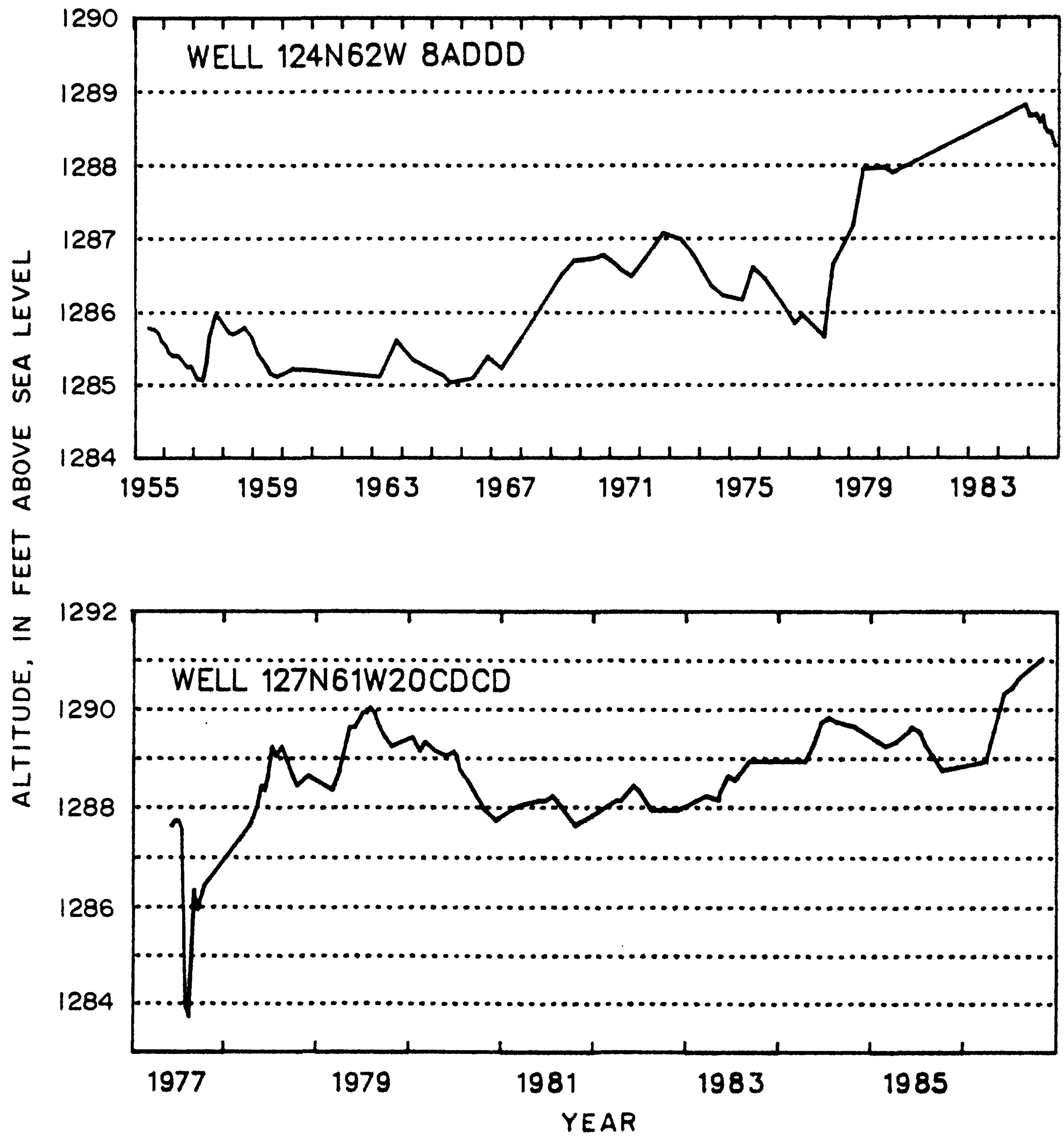


Figure 10.--Water-level hydrographs for selected wells completed in the Middle James aquifer.

Natural discharge from the aquifer occurs as percolation into the Deep James aquifer, which underlies parts of the Middle James, and as eastward flow into the lake deposits and till. The Middle James aquifer does not appear to be in hydraulic connection with the James River. Apparent gains to and losses from the James River in the study area are the result of bank storage, evapotranspiration, and flooding (Koch, 1970). According to Koch and Bradford (1976), the general direction of water movement in the aquifer is to the east. Discharge from the aquifer also occurs by pumpage from wells. Like the Elm aquifer, most of the wells that penetrate the Middle James are small-yield domestic, farm, and stock wells.

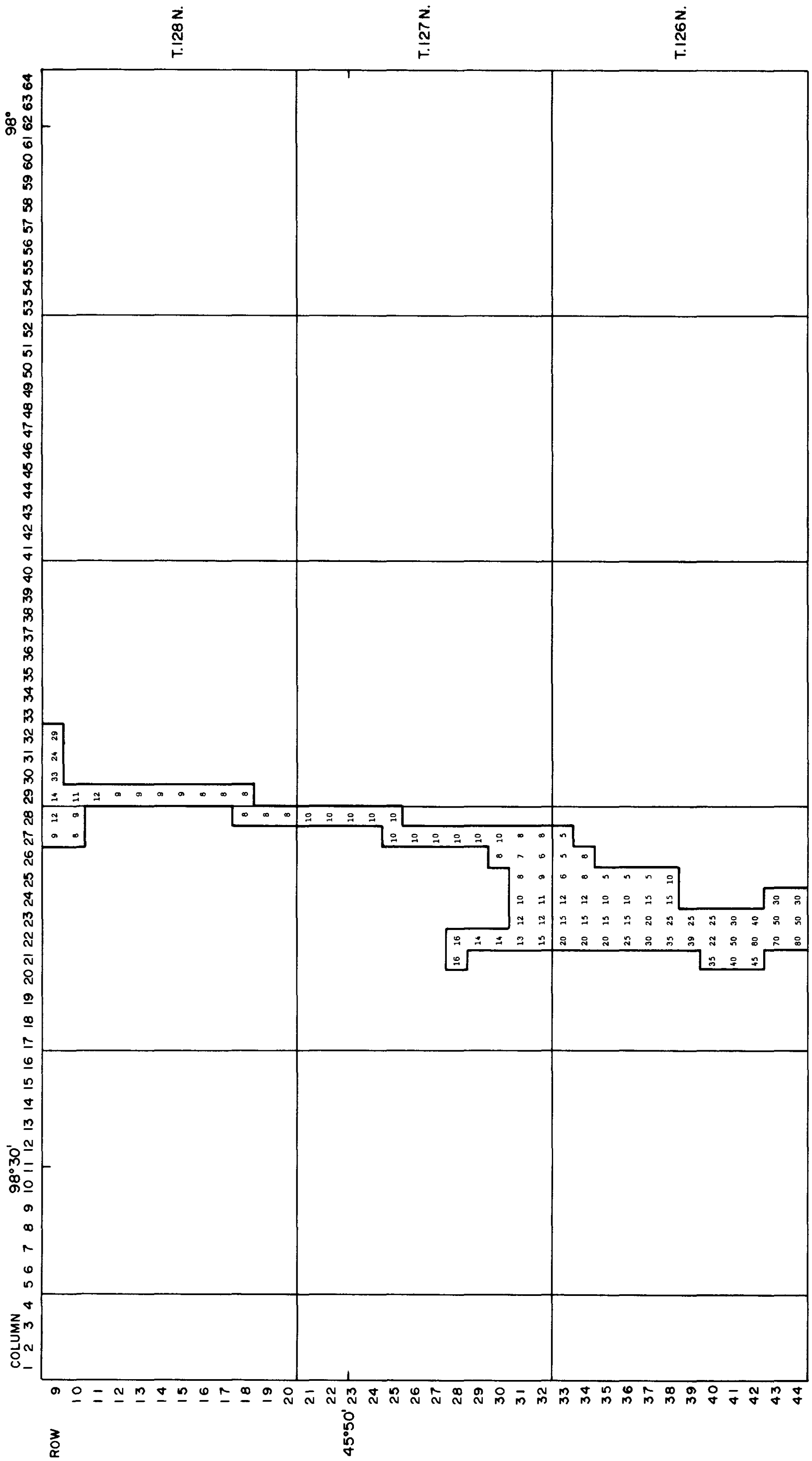
Deep James Aquifer

The artesian Deep James aquifer, which underlies the Elm and Middle James aquifers (figs. 4 and 11), is a buried interconnected system of ancient river channels containing outwash and alluvium. The boundaries shown in figure 11 differ from those of Koch and Bradford (1976) due to the availability of additional well logs, reinterpretation of existing logs, and the averaging of the aquifer thicknesses. The aquifer underlies about 52 mi^2 of the study area. The thickness ranges from zero at the boundaries to 132 ft in 125N62W18C and averages about 16 ft.

Recharge to the aquifer is from the overlying aquifers and to a lesser extent by percolation through the overlying lake deposits, outwash, and till. Recharge also may occur as underflow from Spink and Marshall Counties (Koch and Bradford, 1976).

There are no long-term water-level data available for the Deep James aquifer in the study area. However, Koch and Bradford (1976) reported that "there appears to be no relationship of water-level change in the Deep James aquifer to seasonal or long-term changes in precipitation."

Natural discharge from the aquifer occurs as subsurface outflow into North Dakota and locally by leakage into the overlying till (Koch and Bradford, 1976). Discharge from the aquifer also occurs by pumpage from low-yield domestic and farm wells. No large-capacity wells are reported to have been completed in the Deep James aquifer although the Deep James can be affected by pumpage from the overlying aquifers, as indicated in the hydrograph from observation well 121N65W34CCCC (fig. 12). According to Koch and Bradford (1976), the general direction of ground-water movement in the aquifer is to the north.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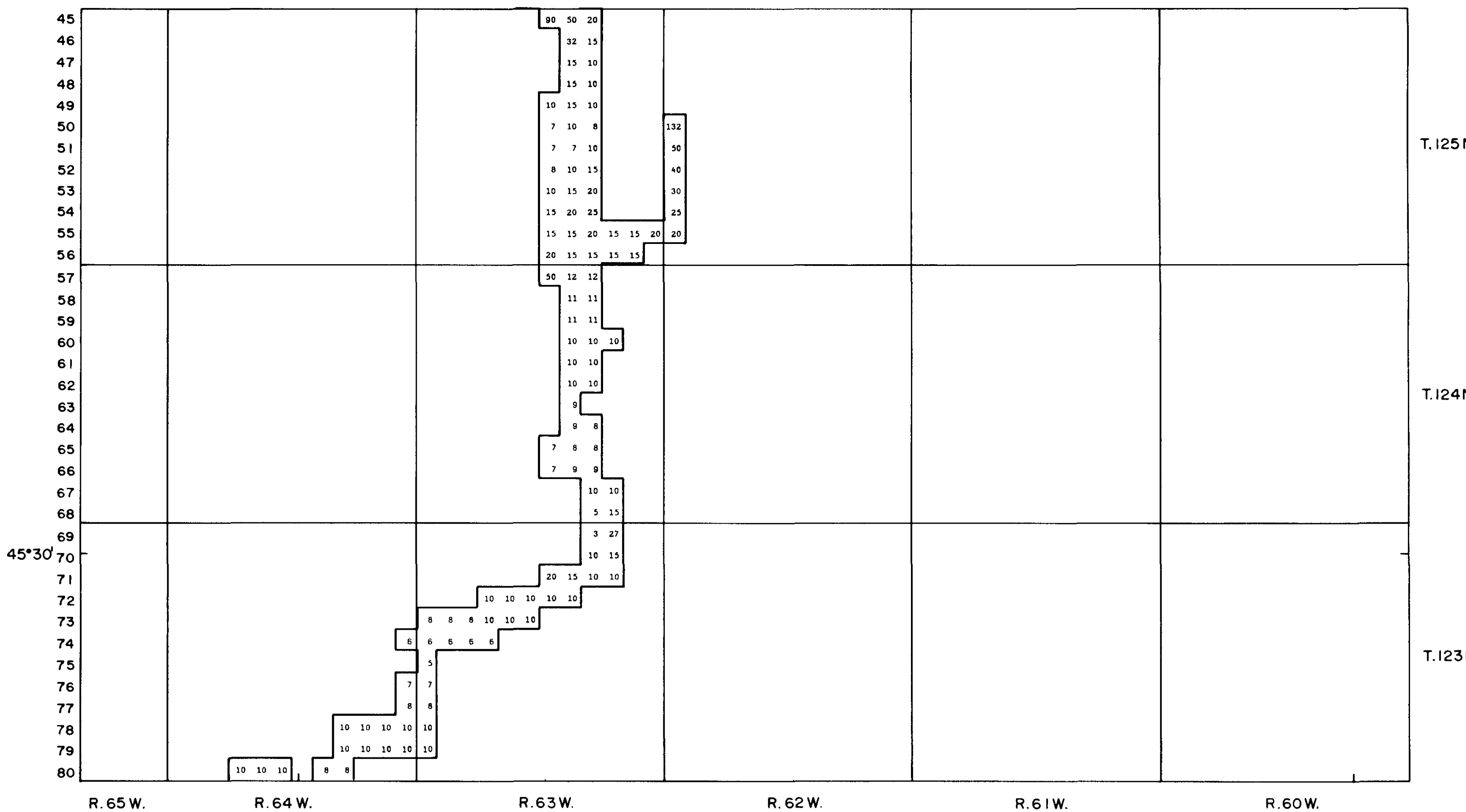
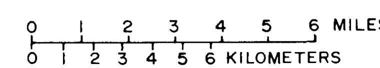
COLUMN
ROW





EXPLANATION

25 AQUIFER THICKNESS--Number is the average value for each grid block (0.25 square mile), in feet

—— AQUIFER MODEL BOUNDARY

Figure 11.--Average thickness of the Deep James aquifer for each grid block.

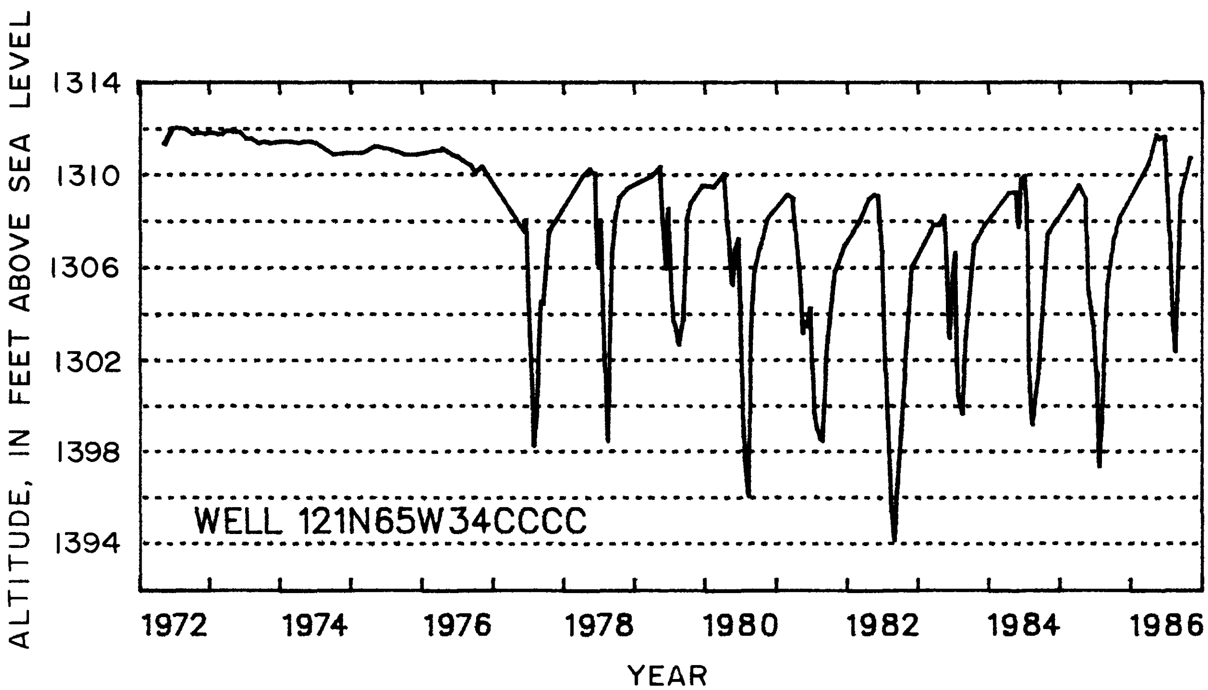


Figure 12.--Water-level hydrograph for a selected well completed in the Deep James aquifer.

DESIGN OF THE GLACIAL-AQUIFER SYSTEM FLOW MODEL

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 applied to the glacial-aquifer system are:

1. The system consists of three aquifer layers (fig. 13). The upper aquifer (aquifer layer 1) represents the Elm aquifer and the sandy lake deposits. The middle aquifer (aquifer layer 2) represents the Middle James aquifer, and the bottom aquifer (aquifer layer 3) represents the Deep James aquifer.

2. The aquifer layers are overlaid by confining bed layers. The uppermost confining layer which extends from land surface to the top of aquifer layer 1 restricts the downward infiltration of recharge water to the aquifer and the upward migration of water from the aquifer. The confining beds between aquifer layers 1 and 2 and aquifer layers 2 and 3 act to restrict the vertical flow between these aquifers.

3. The bedrock is an impermeable lower boundary of the system.

4. All lateral boundaries of the aquifers are impermeable (no-flow boundaries). Along the northern and eastern boundaries, internal potentiometric heads are held constant (specified-head boundaries).

5. The James River, Maple River, Foot Creek, and Moccasin Creek are hydraulically isolated from the glacial-aquifer system. The Elm River is hydraulically connected to the Elm aquifer along a reach from about 125N64W3B to 124N63W15B (fig. 14).

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

7. The principal source of recharge to the aquifer system is precipitation. The uppermost active aquifer is recharged directly by infiltration of precipitation, however, the thickness of the confining bed overlying the aquifer controls the rate at which the recharge can occur. The greater the confining bed thickness, the less the recharge rate. Recharge to the lower aquifers occurs as infiltration through the overlying deposits. Recharge can also occur at river-head and specified-head boundaries.

8. The primary method of discharge from the aquifer system is evapotranspiration. Upward leakage of water from the uppermost active 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 less the rate at which evapotranspiration can occur. Discharge from the aquifer system also can occur as pumpage and at river-head and specified-head boundaries.

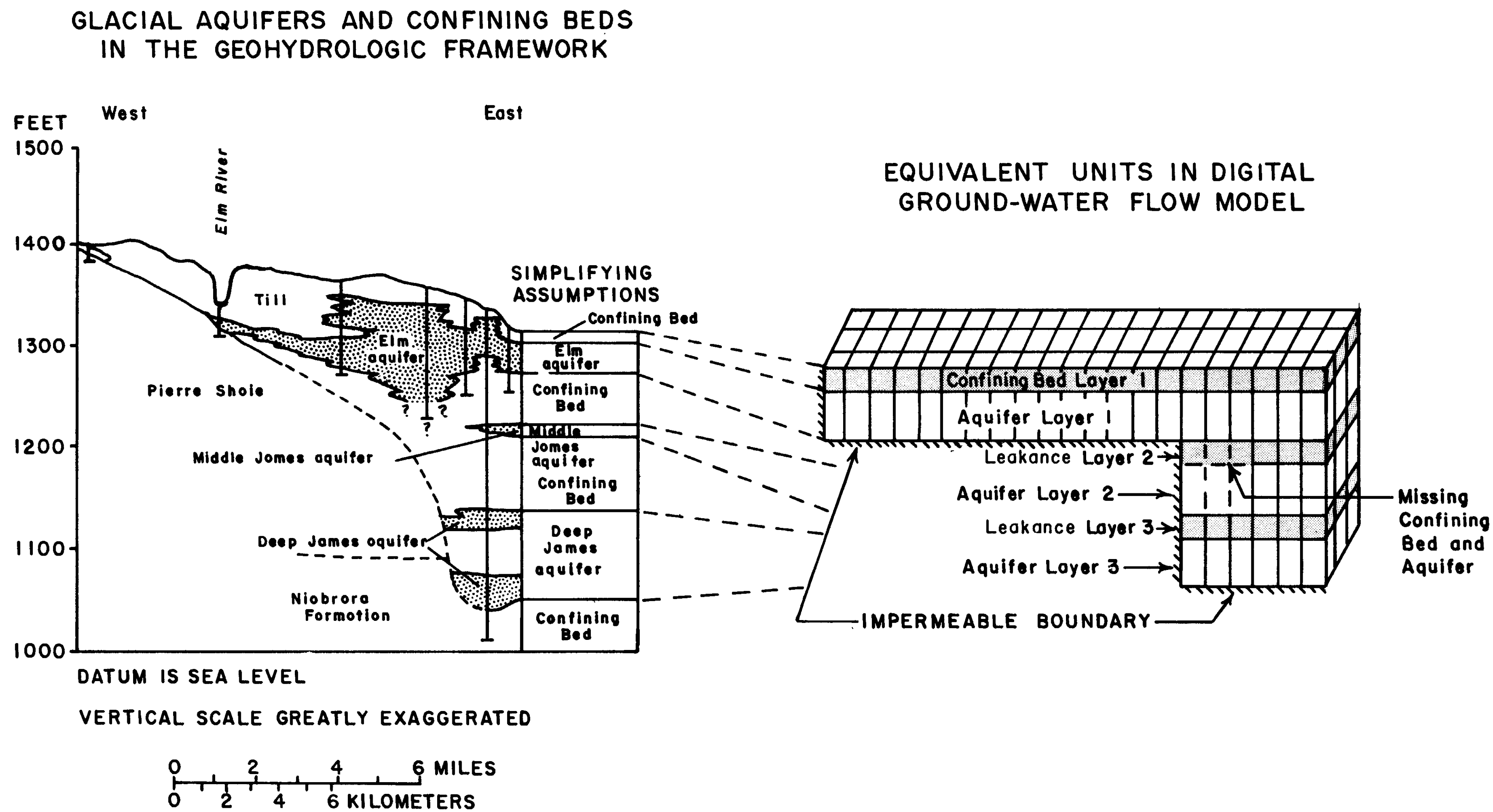
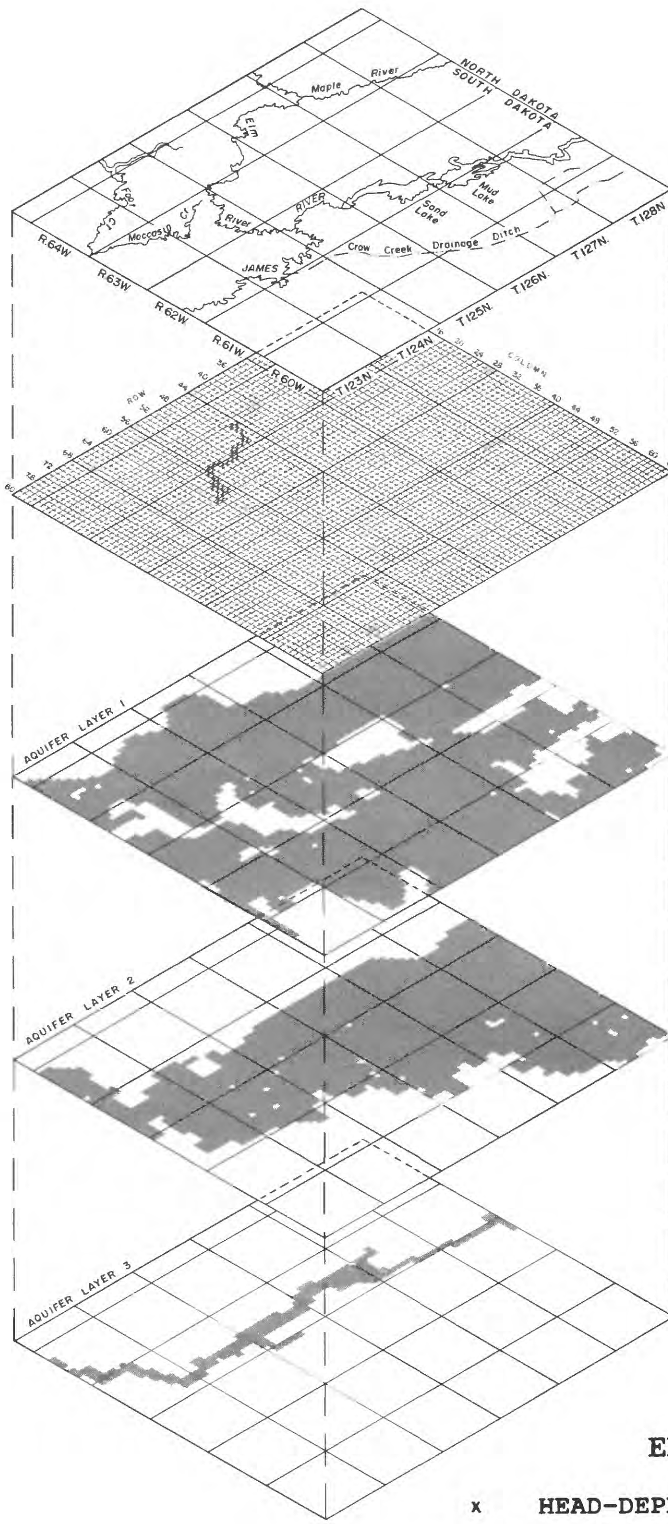


Figure 13.--Simplifying assumptions used to develop the digital ground-water flow model.



EXPLANATION

x HEAD-DEPENDENT RIVER GRID BLOCK

AQUIFER LAYER

Figure 14.--Finite-difference grid blocks and aquifer layers 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 about 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 simulated area was subdivided into a series of finite-difference grid blocks in which the aquifer properties are assumed to be constant (fig. 14). 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 properties 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 properties used in the model of the glacial-aquifer system:

1. Dimensions of the finite-difference grid.
2. Altitude of the top and bottom of the aquifers.
3. Hydraulic conductivity of the aquifers.
4. Leakance of the confining beds.
5. Storage in the aquifers.
6. Recharge to the aquifers.
7. Evapotranspiration from the aquifers.
8. Altitude of land surface.
9. Pumpage from the aquifers.
10. Hydraulic connection between the rivers and aquifers.

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 digital model. The equally spaced finite-difference grid selected to represent the model area has 86 rows and 70 columns. The grid blocks are one-half mile or 2,640 ft on a side. Each grid block, shown in figure 14, represents one-quarter of a 640-acre section. The model grid was extended beyond the study area, 4 mi to the north and 3 mi to the east and south to minimize boundary effects in the study area.

Altitude of the Top and Bottom of the Aquifers

Koch and Bradford (1976) defined the Elm, Middle James, and Deep James aquifers based on altitude. The maximum altitude of the top of the Elm aquifer was defined as 1,400 ft and the minimum altitude of the bottom was 1,225 ft. The maximum altitude of the top of the Middle James aquifer was 1,250 ft and the minimum altitude of the bottom was 1,150 ft. The maximum altitude of the top of the Deep James aquifer was 1,175 ft, and the minimum altitude of the bottom was 950 ft.

The method used to define the tops and bottoms of the aquifers for the model is a modification of the method used by Koch and Bradford (1976). Aquifer layer 1 is the Elm aquifer and the sandy lake deposits. The altitude of the top of aquifer layer 1 is the top of the first sand or gravel layer encountered below the overlying till or other fine-grained sediment. Where the overlying confining bed is not present, the altitude of the aquifer top is land surface.

The bottom of aquifer layer 1 is the bottom of the lowermost sand or gravel layer above 1,250 ft above sea level. If no confining bed exists between aquifer layers 1 and 2, the bottom of aquifer layer 1 is 1,250 ft above sea level.

Aquifer layer 2 is the Middle James aquifer. The top of aquifer layer 2 is the first occurrence of a sand or gravel layer below the altitude of 1,250 ft. If no confining bed exists between aquifer layers 1 and 2, the top of aquifer layer 2 is 1,250 ft above sea level.

The bottom of aquifer layer 2 is the bottom of the lowermost sand or gravel layer above 1,150 ft above sea level or if no confining bed exists between aquifer layers 2 and 3, the bottom of aquifer layer 2 is 1,150 ft above sea level.

Aquifer layer 3 is the Deep James aquifer. The top of aquifer layer 3 is the first occurrence of a sand or gravel layer below the altitude of 1,150 ft. If no confining bed exists between aquifer layers 2 and 3, the top of aquifer layer 3 is 1,150 ft above sea level. The bottom of aquifer layer 3 is the lowermost sand or gravel layer below 1,150 ft above sea level. Since aquifer layer 3 was simulated as a confined aquifer, no top or bottom arrays were required.

Hydraulic Conductivity of the Aquifers

The hydraulic conductivity or the ability of the aquifer to transmit water varies greatly over very 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 grid block was developed. Using the average composite logs for each grid block and the hydraulic conductivity values shown in table 1, an average composite aquifer hydraulic conductivity was calculated for each grid block for each aquifer layer as shown in figures 15-17. The assignment of the average composite hydraulic conductivity for each grid block 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 grid block.

ROW \ COLUMN	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16
9															273	272
10															193	202
11															187	220
12															218	232
13															250	250
14															250	250
15															250	250
16															250	250
17															250	300
18															337	325
19															312	330
20															350	350
21															350	350
22															330	337
45°50' 23															310	310
24															300	290
25															304	287
26															280	283
27															270	283
28															250	250
29															250	250
30															350	257
31															183	189
32															183	67
33															183	100
34															191	118
35															191	163
36															228	217
37														350	293	257
38														350	301	288
39													350	350	350	306
40												350	350	350	336	301
41												350	350	330	317	317
42												350	300	310	300	300
43												350	300	283	283	283
44												300	300	279	270	273

98°30' (above columns 10–11)

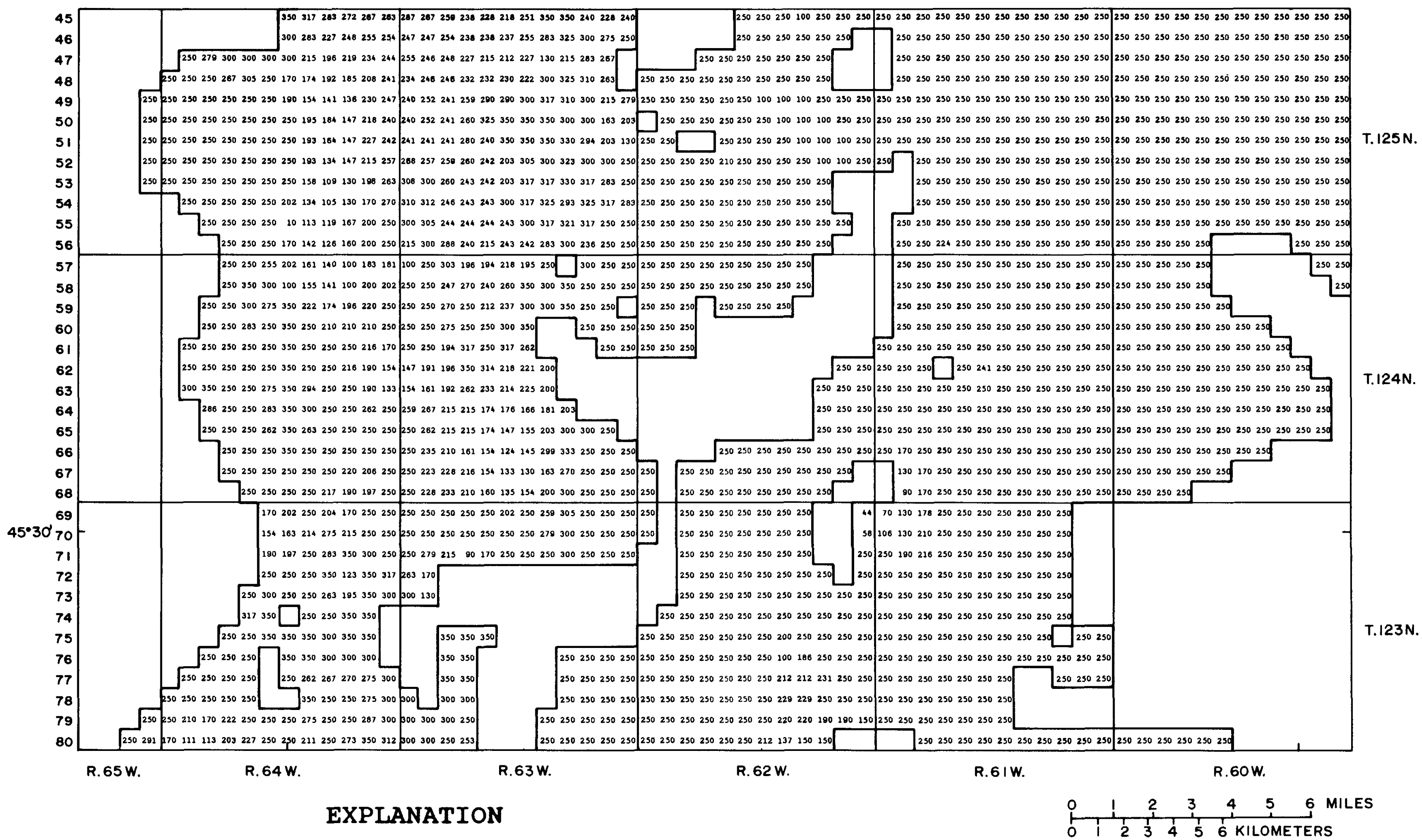
ROW \ COLUMN	17	18	19	20	21	22	23	24	25	26	27	28
9	195	126	70	58	92	152	204	223	237	195	118	109
10	181	187	121	83	116	146	180	204	228	163	127	115
11	216	196	185	139	168	160	163	212	172	144	120	121
12	202	181	203	197	203	184	184	194	180	137	128	128
13	250	222	243	241	218	218	218	199	184	150	128	128
14	250	242	259	252	230	218	199	199	184	184	166	185
15	250	242	242	241	224	230	241	203	203	185	185	147
16	250	273	139	231	220	239	252	224	224	203	174	174
17	282	252	248	224	220	223	228	224	202	175	184	203
18	288	264	228	223	212	217	226	239	262	203	203	300
19	350	282	237	207	180	204	220	240	317	300	300	250
20	350	298	229	191	156	350	350	350	325	300	267	250
21	350	298	237	180	169	282	293	350	341	330	320	300
22	325	262	218	185	173	204	226	240	300	300	292	283
23	312	287	239	190	190	192	204	283	277	240	241	241
24	300	274	254	207	195	203	188	239	246	244	234	226
25	294	267	232	210	213	180	235	241	237	221	211	194
26	283	265	224	228	231	244	282	281	226	205	190	158
27	272	264	226	233	247	270	308	312	245	188	166	130
28	270	265	242	238	270	297	301	315	187	172	135	146
29	250	265	252	227	275	292	298	282	230	200	140	196
30	250	246	256	243	260	277	293	274	250	200	161	215
31	223	232	246	234	243	242	282	240	224	203	200	70
32	164	202	219	219	230	230	227	203	242	215	267	
33	128	164	248	219	215	215	215	243	300	283	250	
34	138	147	167	187	203	203	225	223	283	283	250	
35	170	175	181	187	207	202	208	210	216	250	250	
36	208	203	195	192	189	186	190	195	202	216	250	
37	248	230	223	220	215	199	199	205	210	216	250	
38	277	260	252	244	224	208	194	197	195	190	250	
39	295	282	288	258	285	300	350	262	273	300	283	
40	296	321	307	306	310	310	350	296	273	300	300	
41	314	310	305	306	306	306	350	306	336	350	350	
42	300	300	300	294	294	294	350	350	350	350	350	350
43	283	283	277	268	267	290	350	350	350	350	350	350
44	273	265	258	250	250	288	350	350	350	350	350	350

ROW \ COLUMN	29	30	31	32	33	34	35	36	37	38	39	40
9	129	144	155	170	173	208	279	350	350			
10	128	127	146	155	177	212	240	350	350			
11	138	150	160	170	180	204	214	237	350			
12	140	150	160	170	170	210	194	203				
13	142	166	177	98	197	197	194	203				
14	161	161	161	161	185	185	220	203				
15	174	180	229	171	189	192	215	203	250			
16	215	277	187	178	203	212	222	130	250			
17	300	299	209	208	227	225	215	215	267			
18	250	232	243	242	245	245	256	243	300			
19	250	283	243	242	245	256	258	258	295			
20		287	243	242	245	255	274	259	300			
21	240	307	255	251	261	199	208	202	250			
22	300	300	270	276	252	230	196	190	170			
23	241	241	241	241	233	229	210	215	170			
24	219	215	211	210	211	227	210	215	170			
25	192	300	197	197	198	210	215	200	181			
26	170	193	184	190	185	210	220	186	190			250
27	144	154	158	172	157	164	191	195	202		250	250
28	160	210	173	181	140	169	260	180	181	250	250	250
29	222	231	240	240	250	250	250	250	250	250	250	250
30	210	222	250	250	250	250	250	250	250	250	250	250
31	250	250	250	250	250	250	250	250		250	250	250
32					250	250	250			250	250	250
33					250	250	250		250	250	250	250
34									250	250	250	250
35									250	250	250	250
36								250	250	250	250	250
37								250	250	250	250	250
38								250	250	250	250	250
39							250	250	250	250	250	250
40							250	250	250	250	250	250
41							250	250	250	250	250	250
42							250	250	250	250	250	250
43						250	250	250	250	250	250	250
44						250	250	250	250	250	250	250

ROW \ COLUMN	41	42	43	44	45	46	47	48	49	50	51	52
9						250	250		250			
10						250	250	250	250	250		
11							250	250	250	250		
12							250	250	250			
13	250	250	250	250	250	250	250	250	250			
14	250	250	250	250	250	250	250	250	250	213		
15	250	250	250	250	250	250	250	250	250			
16		250	250	250	250	250	250	250	250			
17		250	250	250	250	250	250		250			
18		250	250	250	250	250	250	250	250	250		
19		250	250	250	250	250	250	250	250	250		
20		250	250	250	250	250	250	250	250	143		
21		250	250	250	250	250	250	250	250	250		
22		250	250	250	250	250	250	250	250			
23		250	250	250	250	250	250	250	250	250		
24		250	250	250	250	220	250	250	250	250	250	
25	250	250	250	250	250	250	250	250	250	250	250	
26	250	250	250	250	250	250	250	250	250	250	200	250
27	250	250	250	250	250	250	250	250				
28	250	250	250	224	250	250	250					
29	250	250	250	250	175	250	250		250	250	250	250
30	250	250	250	250	250	250	250	250	250	250	250	250
31	250	250	250	250	250	250	250	250	250	250	250	250
32	250	250	250	250	242	250	250	250	250	250	204	250
33	250	250	250	250	250	250	250	250	250	250	250	250
34	250	250	250	250	250	250	250	250	250	250	250	250
35	250	250	250	250	250	250	250	250	250	250	250	250
36	250	250	250	250	250	250	250	250	250	250	250	250
37	250	250	250	250	250	250	250	250	250	250	250	250
38	250	250	250	250	250	250	250	250	250	250	250	250
39	250	250	250	250	250	250	250	250	250	250	250	250
40	250	250	250	250	250	250	250	250	250	250	250	250
41	250	250	250	250	250	250	250	250	250	250	250	250
42	250	250	250	250	250	250	250	250	250	250	250	250
43	250	250	250	250	250	250	250	250	250	250	250	250
44	250	250	250	250	250	250	250	250	250	250	250	250

98° (above column 62)

ROW \ COLUMN	53	54	55	56	57	58	59	60	61	62	63	64	
9					250	250	250	250	250				
10				250	250	250	250	250	250	250	250		
11				250	250	250	250	250	250	250	250		
12				250	250	250	250	250	250	250	250		
13				250	250	250	250	250	250	250	250		
14	250	250	250	250	250	250	250	250	250	250	250	250	
15		250	250	250	250	250	250	250	250	250	250	244	T.128 N.
16			250	250	250	250	250	250	250	250		250	
17					250	250	250	250	250	250	250	250	
18					250	250	250	250	250	250	250	250	
19						250	250	250	250	250	250	250	
20					250	250	250	250	250	250	250	250	
21								250	250	250	250	250	
22							250	250	250	250	250	250	
23							250	250	250	250	250	250	
24							250	250	250	250	250	250	
25										250	250	250	
26											250	250	
27									250	250	250	250	T.127 N.
28							250	250	250	250	250	250	
29	250	250					250	250	250	250	250	250	
30	250	250		250	250	250	250	250	174	250	250	250	
31	250	250		250	250	250	250	250	250	250	250	250	
32	250	250	250	250	250	250	250	250	250				
33	250	250	250	250	250	250	250	250	250	250	250	250	
34		250	250	250	250	250	250	250	250	250	250	250	
35	250	250	250	250	250	250	250	250	250	250	250	250	
36	250	250	250	250	250	250	250	250	250	250	250	250	
37	250	250	250	250	250	250	250	250	250	250	250	250	
38	250	250	250	250	250	250	250	250	250	250	250	250	
39	250	174	250	250	250	250	250	250	250	250	250	250	T.126 N.
40	250	250	250	250	250	250	250	250	250	250	250	250	
41	250	250	250	250	250	250	250	250	250	250	250	250	
42	206	250	250	250	250	250	250	250	250	250	250	250	
43	250	250	250	250	250	250	250	250	250	250	250	250	
44	250	250	250	250	250	250	250	250	250	250	250	250	



EXPLANATION

250 **HYDRAULIC CONDUCTIVITY--Number is average value for each grid block (0.25 square mile), in feet per day**

—— **AQUIFER MODEL BOUNDARY**

Figure 15.--Average hydraulic conductivity of aquifer layer 1 for each grid block.

ROW	20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38	39	40
9				350	350	350	350	350	350	350	350	300	250	250	250	250	250	283	350	350	322
10				350	350	350	350	350	350	350	317	317	300	300	275	250	250	283	350	293	307
11				350	350	350	350	350	350	350	350	333	321	312	300	283	275	280	350	350	312
12				350	350	350	350	350	350	350	333	326	323	319	317	325	330	301	307	312	293
13				350	350	350	350	350	350	350	325	317	317	321	325	330	333	307	282	265	271
14					350	350	350	350	350	350	330	321	316	317	317	317	333	301	307	316	265
15					350	350	350	350	350	350	333	324	322	321	305	321	321	279	350	307	282
16					350	350	350	350	350	350	333	325	324	315	297	333	333	350	350	350	307
17						350	350	350	350	350	333	325	324	299	289	330	330	350	350	350	350
18						350	350	350	350	350	333	325	324	305	290	328	330	337	350	350	350
19						350	350	350	350	350	333	325	321	325	289	321	331	330	340	350	350
20						350	350	350	350	350	333	325	321	325	294	296	299	304	329	333	350
21						350	350	350	350	350	333	327	325	310	296	290	330	350	350	350	350
22					350	350	350	350	350	301	333	327	321	320	297	279	281	350	350	350	350
23				350	350	350	350	350	287	252	279	312	312	314	266	234	262	301	336	330	317
24			350	350	350	310	305	298	252	226	252	269	284	299	269	255	256	279	304	320	321
25		350	276	289	276	273	265	243	229	212	232	259	285	294	259	245	254	261	275	314	308
26	350	350	261	261	247	240	234	235	229	219	222	242	264	284	259	249	245	241	241	260	308
27	350	350	279	208	192	216	211	207	215	216	221	236	249	273	256	238	219	232	241	270	280
28	350	350	350	270	187	191	195	203	211	220	229	239	249	249	242	238	212	217	232	252	282
29	350	350	350	266	173	252	240	176	203	210	224	211	242	224	205	161	168	187	188	217	300
30	350	350	312	312	295	240	227	210	172	201	215	195	233	219	201	186	177	177	175	218	273
31	350	350	301	301	277	203	187	147	170	188	181	172	233	215	199	193	193	198	206	259	294
32	350	350	293	301	277	260	196	170	185	185	170	172	210	206	193	200	201	215	243	290	283
33	350	350	350	350	273	215	206	185	161	186	185	178	172	191	163	196	199	228	241	306	300
34	350	350	350	325	350	317	244	222	198	190	190	215	350	222	222	237	199	350	350	350	350
35	350	350	350	350	350	293	277	273	257	215	243	305	350	350	350	350	350	350	350	350	350
36	350	350	350	262	350	350	350	330	273	273	317	350	350	350	350	350	350	350	350	350	350
37	350	350	350	350	350	350	350	350	350	350	350	350	350	350	350	350	350	350	265	192	139
38	350	350	350	350	350	350	350	350	350	350	350	350	350	350	350	350	350	350	192	150	134
39	350	350	350	350	350	350	350	350	350	350	350	350	350	350	350	350	350	350	219	200	163
40	350	350	350	350	350	350	350	350	350		350	350	350	312	300	203	350	350	237	203	163
41	350	350	350	350	350	350	350	350	350	350	350	312	300	283	187	142	203	237	237	237	180
42	310	320	350	350	350	350	317	317	350	350	350	317	242	167	130	109	142	194	265	265	204
43	250	300	330	350	350	325	300	275	300	317	317	317	300	310	260	240	203	262	262	237	204
44	250	279	312	350	305	273	250	270	283	294	294	350	350	350	350	350	325	130	160	168	156

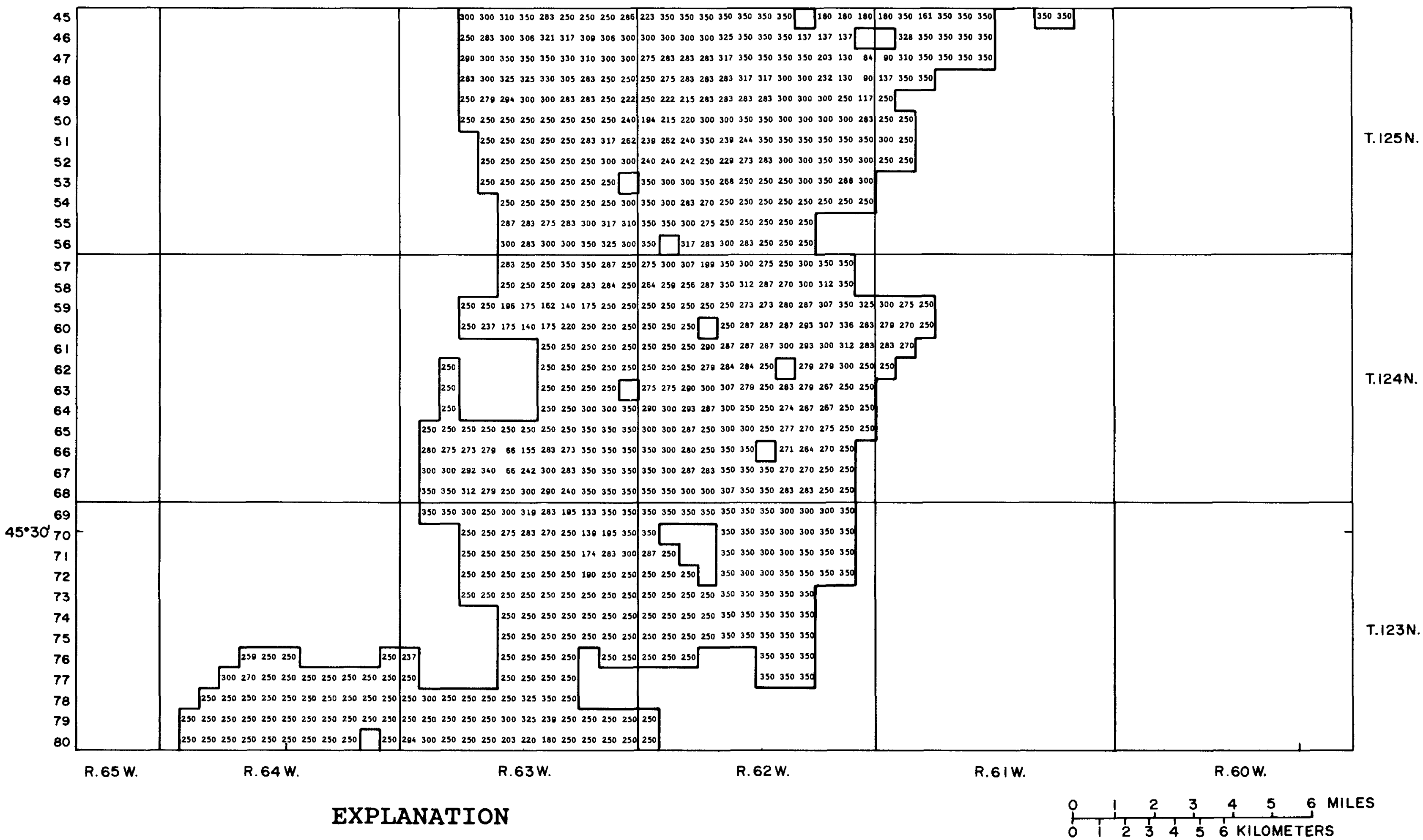
ROW	41	42	43	44	45	46	47	48	49	50	51	52	53	54	55	56	57	58	59	60	61	62	63	64
9	341	350	350	350	350	330	325	300	250	317	350	350	350	350	350	350	92	113	170	250	250	300	317	
10	350	330	310	300	262	300	312	294	250	310	250	294	300	350	350		130	90	130	250	250	250	306	
11	350	350	350	300	250	250	279	267	294	308	287	303	304	305	300	250	170	170	210	250	250	283		
12	282	301	350	350	317	293	283	300	294	275	250	250	250	250	250	250	250	250	250	250	250	275		
13	282	307	350	350	350	350	310	300	275	279	264	250	287	250	250	250	250	250	250	250	250	242	283	
14	282	301	350	350	350	350	350	350	267	265	258	250	300		250	300			250	225	250	283	250	
15	307	350	350	350	350	350	350	350	279	350	300	317	350	350	350	350	350	350	250	250	283	300	250	
16	350	350	350	350	350	350	350	300	350	270	317	330	283	300	300	300	300	317	250	250	250	250	250	250
17	350	350	350	350	350	323	295	283	330	265	253	222	250	350	300	300	300	286	250	250	250	250	250	250
18	350	350	350	350	350	300	300	300	328	350	237	211	253	130	170	280	280	328	250	250	250	250	250	
19	350	350	350	350	350	300	300	300	325	350	265	350	350	237	300	250	250	250	250	250	250	250		
20	350	350	350	350	350	250	344	317	330	350	350	350	250	250	250	250	250	250	250	250	250			
21	350	350	350	350	350	250	300	317	350	350	350	250	250	250	250	250	250	250	250	250	250			
22	350	321	350	350	350	317	325	317	317		350	307	275	270	270	268	275	264	250	250	250			
23	317	307	350		350	350	330	330	317	300	300	300	290	283	268	292	290	275	264	250	250			
24	263	279	300		350	350	333	336	321	317	317	300	295	293	296	312	294	250	250	250	250			
25	272	250			350	350	350	340	325	320	320	306	307	268	294	304	300	250	250	250				
26	294	283	279	250	300	300	325	330	323	317	310	300	259	252	288	300	288	250	250	250				
27	306	300	270	250	300	283	310	314	276	300	300	300	300	300	275	250	250	255						
28	333	350	300	350	350	250	305	317	257	257				317	300			245						
29	308	312	282	277	350	350	350	300	350	350			350	317	307									
30	271	275	273	270	263	309	300	250					350	350	300									
31	279	275	279	267	250	283	263	250																
32	270	250	250	250	250	250	250	250	250															
33	290	250	250	250	250	250	250	252	250															
34	350	300	250	250	250	250	250	250	250															
35	350	300	300	290	268	283	283																	
36	350	314	312	312	306	300	283																	
37	212	236	250	300	310	317	317	350																
38	143	172	195	250	317	336	350	350																
39	150	194	169	253	350	350	350	350	350	350	350	350												
40	163	220	180	265	350	350	350	350	300	317	350	350												
41	180	237	180	350	350	350	350	300	283	263	300	300												
42	204	265	237	350	350	350	350	300	250	250														
43	180	350	282	350	350	350	350	300	300	300														
44	147	180	95		350	350	350		350	350														

COLUMN 1–64; 98°30' (at column 11); 98° (at column 62); 45°50' (at row 23)

T.128 N.

T.127 N.

T.126 N.

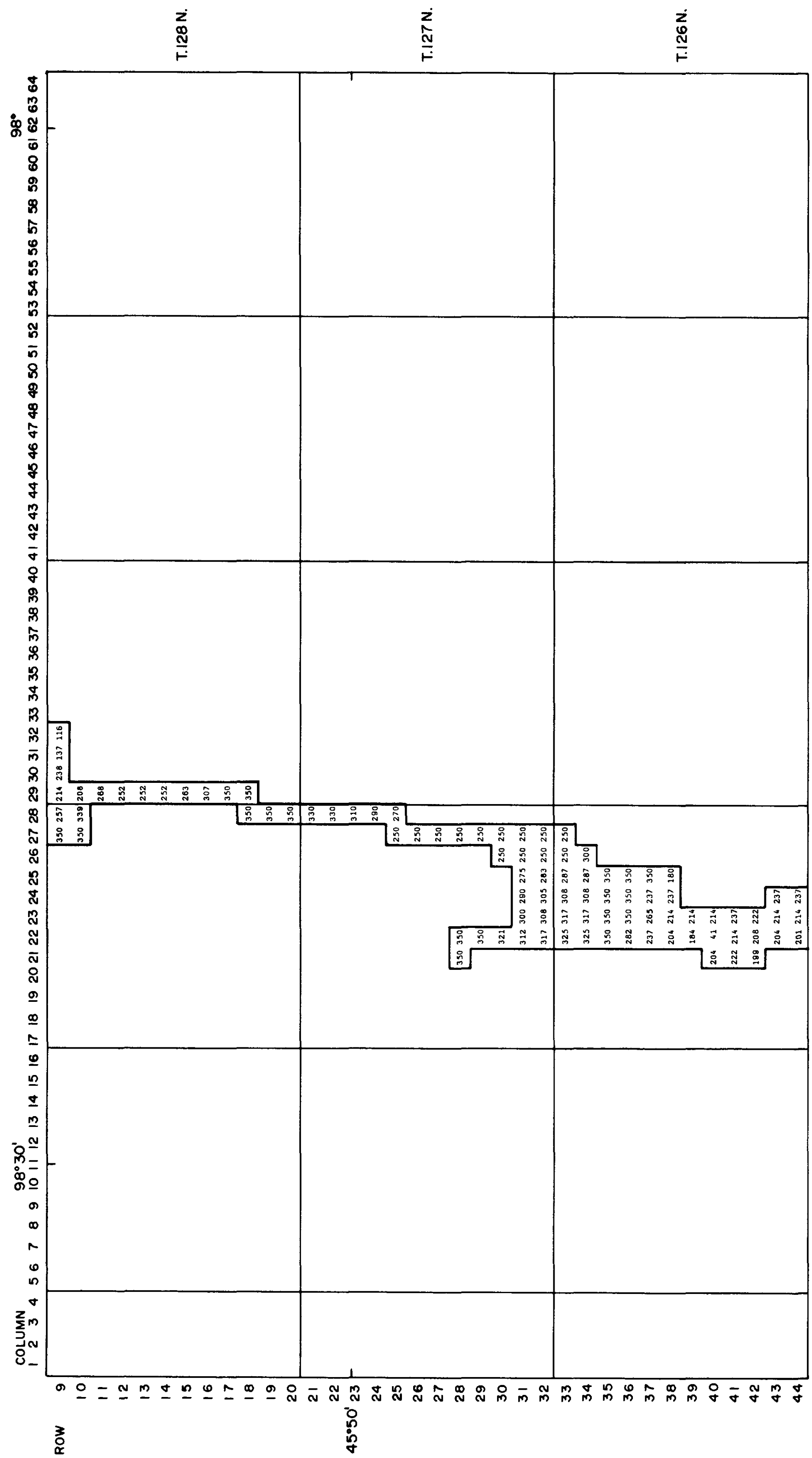


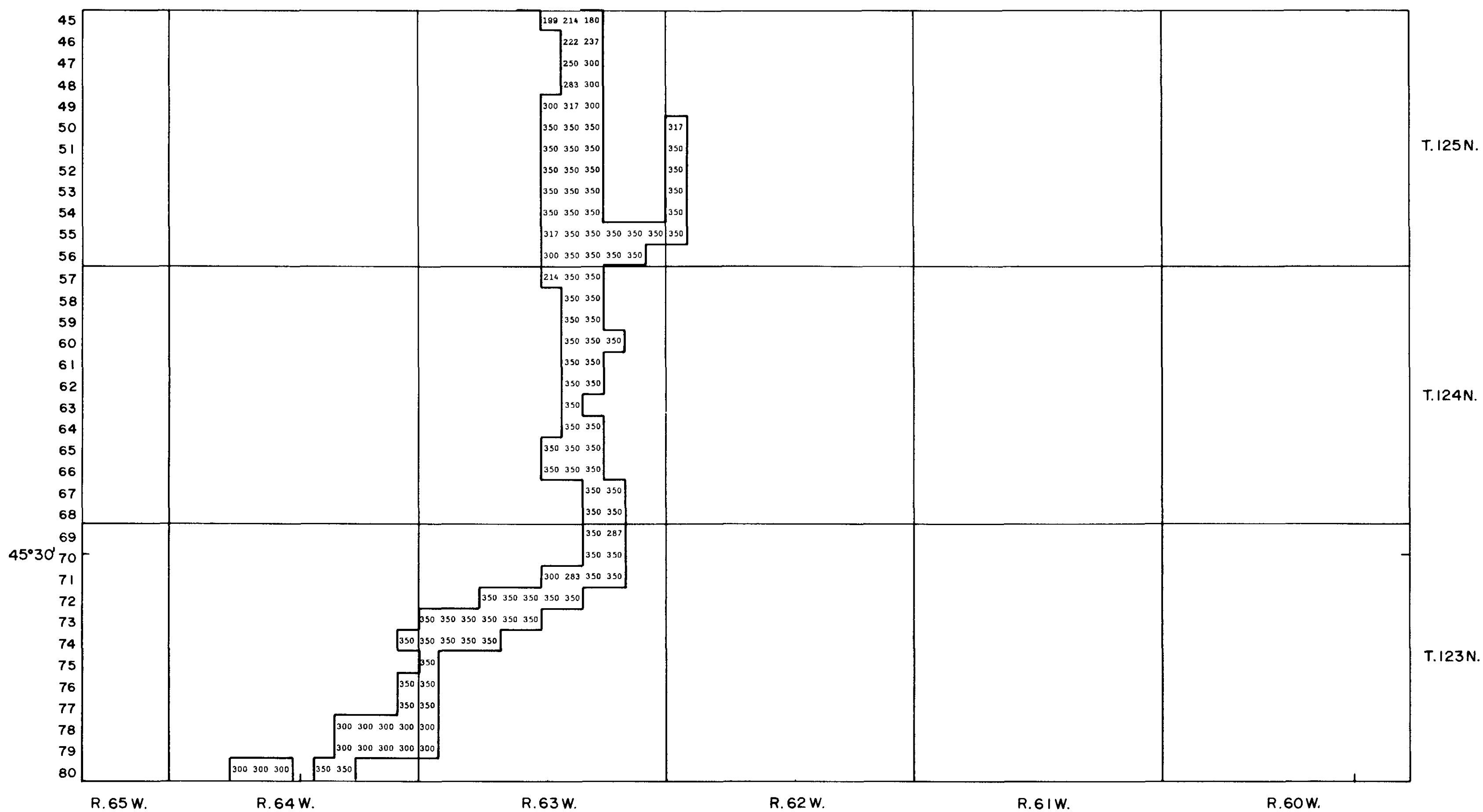
EXPLANATION

250 HYDRAULIC CONDUCTIVITY--Number is average value for each grid block (0.25 square mile), in feet per day

—— AQUIFER MODEL BOUNDARY

Figure 16.--Average hydraulic conductivity of aquifer layer 2 for each grid block.


T.128 N.
T.127 N.
T.126 N.
98°
98°30'
45°50'
COLUMN
ROW



EXPLANATION

250 HYDRAULIC CONDUCTIVITY--Number is average value for each grid block (0.25 square mile), in feet per day

—— AQUIFER MODEL BOUNDARY

Figure 17.--Average hydraulic conductivity of aquifer layer 3 for each grid block.

Table 1.--Relation between grain size and hydraulic conductivity

[<, less than; --, no data]

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 conductivity assigned to glacial-aquifer system in Sanborn-Beadle County[2] (feet per day)	Hydraulic conductivity used in this model (feet per day)
Clay or silt	<20	10	6-22	10
Clay or silt, sandy or gravelly	--	--	--	20
Sand, clayey or silty	--	--	--	150
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	75-281	250
Sand, medium to coarse	130-800	400	--	--
Sand, coarse	400-1,000	540	--	--
Sand, gravelly	--	--	--	350
Sand and gravel	400-1,200	600	85-319	350
Sand, coarse, and gravel	400-1,400	670	--	--
Gravel, clayey or silty	--	--	--	200
Gravel	800-2,000	800	95-356	350

[1]From Koch, 1980.
[2]From Emmons, 1988.

No aquifer-test data are known to be available for the Elm, Middle James, or Deep James aquifers in or around the area of study. The hydraulic conductivity calculated from aquifer-test data from the Sanborn-Beadle County area (fig. 1) ranges from 20 to 1,430 ft/d with an average of about 300 ft/d (Emmons, 1988). Because the composition of the glacial drift in Brown, Sanborn, and Beadle Counties are similar, it is assumed that the aquifer characteristics also are similar.

Examination of table 1 indicates that the hydraulic conductivity of the aquifer system generally is much smaller than assigned by Koch (1980) to the alluvium-mantled outwash deposits of the Big Sioux aquifer, located east of the James River basin in South Dakota. The hydraulic conductivities of the outwash deposits in the glacial-aquifer system in the James River basin are smaller because they contain a higher percentage of silt and clay.

The average composite hydraulic conductivity at each node ranges from 10 to 350 ft/d with an average of 245 ft/d in aquifer layer 1. In aquifer layer 2, the average composite hydraulic conductivity ranged from 59 to 350 ft/d with an average of 288 ft/d, and in aquifer layer 3 the hydraulic conductivity ranged from 116 to 350 ft/d with an average of 305 ft/d. Since aquifer layer 3 was simulated as a confined aquifer, transmissivity rather than hydraulic conductivity is required in the model. The transmissivity of a confined aquifer is equal to the hydraulic conductivity of the aquifer (fig. 17) multiplied by the thickness of the aquifer (fig. 11). These average nodal composite hydraulic conductivities generally are smaller than the hydraulic conductivities calculated from aquifer tests in Sanborn and Beadle Counties. This is expected because the aquifer tests are site specific and are generally conducted in areas where the aquifer has greater hydraulic conductivity and thickness. As a result of the averaging process for the hydraulic conductivities the ground-water flow model will approximate the glacial-aquifer system on a county scale, but may differ locally.

Leakance of the Confining Beds

The leakance or leakage coefficient is the volumetric rate at which water will flow vertically from one aquifer to another through an intervening confining bed per unit area per foot of head loss between the aquifers.

The leakance arrays are calculated as the vertical hydraulic conductivity divided by the thickness of each confining bed at each node. There are no data available for the vertical hydraulic conductivities of the confining beds in the study area. In northeastern Illinois, the vertical hydraulic conductivity of the drift calculated from seven aquifer tests ranged from 0.01 to 0.08 ft/d (Walton, 1960). The vertical hydraulic conductivity of the till calculated from an aquifer test which included six well clusters completed in the till in northwestern Beadle County (fig. 1) ranged from 0.000025 to 0.10 ft/d (James M. Montgomery, Consulting Engineers, Inc., 1986).

The vertical hydraulic conductivity of the confining beds overlying aquifer layers 2 and 3 were calculated as 0.1 times their horizontal hydraulic conductivity from table 1. The vertical hydraulic conductivity of confining bed layer 2 ranged from 1.0 to 35 ft/d with an average of 1.5 ft/d. For confining bed layer 3, the vertical hydraulic conductivity ranged from 1.0 to 10.0 ft/d with an average of 1.1 ft/d. Although these vertical hydraulic conductivities are larger than those from Illinois or Beadle County, they were found to give acceptable model results.

COLUMN

98°30' (columns 10–11) 98° (columns 61–62)

	ROW	1	2	3	4	5	6	7	8	9	10	11	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32
	9															68	76	60	40	10	1	20	40	50	60	60	60	80	62	50	40	30	20
	10															68	70	50	30	20	5	30	50	55	50	50	50	50	50	40	30	20	10
	11															66	60	40	34	26	20	40	65	60	50	50	40	40	30	20	20	10	10
	12															66	50	30	34	38	40	50	65	65	50	40	30	30	10	1	1	2	4
	13															64	40	20	40	50	80	60	60	60	50	40	20	1	1	1	1	1	1
	14															62	30	10	20	80	80	80	70	60	40	20	1	1	1	1	1	1	1
	15															60	20	1	20	60	88	80	70	50	40	20	1	1	1	1	1	1	1
	16															84	40	20	30	50	80	70	60	40	30	10	1	1	1	1	1	1	1
	17															66	50	30	40	50	70	60	50	30	20	10	1	1	1	1	1	1	1
	18															88	60	40	50	60	60	50	40	20	10	1	1	1	1	1	1	1	1
	19															70	66	60	60	60	50	40	30	20	10	1	1	1	1	1	1	1	1
	20															80	76	70	60	40	30	20	10	1	1	1	1	1	1	42	1	1	1
	21															80	76	70	70	50	40	30	20	20	1	1	1	1	1	1	1	1	1
	22															80	76	72	72	60	50	40	30	30	1	1	1	1	1	1	1	1	1
45°50'	23															80	76	72	72	60	50	40	30	30	1	1	1	1	1	1	1	1	1
	24															80	76	72	74	60	50	40	40	30	1	1	1	1	1	1	1	1	1
	25															80	76	72	74	60	60	50	50	40	1	1	1	1	1	1	1	1	1
	26															80	76	72	76	60	60	60	60	40	1	1	1	1	1	1	1	1	1
	27															80	78	72	76	70	70	70	70	40	1	1	1	1	1	1	1	10	30
	28															76	70	60	70	70	76	76	76	50	1	1	1	1	1	1	10	20	20
	29															70	60	50	60	70	76	76	76	40	10	1	1	1	1	1	1	1	1
	30															70	80	80	60	70	88	84	60	40	20	20	10	1	1	1	1	1	1
	31															70	80	60	60	60	60	50	40	40	30	30	30	10	1	1	1	1	1
	32															70	60	60	50	50	50	40	40	40	30	30	20	1	1	1	1	1	1
	33															70	60	50	40	40	40	40	40	40	30	30	10	1	1	1	1	0	0
	34															70	62	60	40	40	30	30	30	25	20	15	20	20	1	1	1	0	0
	35															70	64	60	40	30	20	30	30	25	25	20	20	20	1	1	1	0	0
	36															70	66	60	40	20	10	1	30	30	25	25	30	30	1	1	1	1	1
	37														70	70	68	60	50	40	30	30	80	80	60	50	40	40	1	1	1	1	1
	38														70	70	70	60	60	60	50	80	80	86	80	60	40	40	1	1	1	1	1
	39													70	70	70	70	70	70	70	70	80	80	86	80	70	60	60	1	1	1	1	1
	40												72	68	68	70	70	70	70	70	70	78	80	80	76	72	60	60	1		1	1	1
	41												72	68	68	68	88	68	68	68	68	76	74	70	74	72	70	70	1	1	1	1	1
	42												68	68	68	68	68	68	68	68	68	72	66	60	70	70	72	76	80	1	1	1	1
	43												84	88	88	66	66	66	66	66	68	88	60	50	68	68	68	70	76	1	1	1	1
	44												72	70	68	66	84	64	64	84	84	84	54	44	64	80	80	45	50	0	1	1	1

ROW	33	34	35	36	37	38	39	40	41	42	43	44	45	46	47	48	49	50	51	52	53	54	55	56	57	58	59	60	61	62	63	64	
9	1	40	80	100	60	1	1	1	1	1	1	1	1	72	90	10	94	0	0	0	1	1	1	1	100	100	100	100	100	1	1		
10	1	10	40	60	60	1	1	1	1	1	1	1	1	70	70	80	94	75	0	0	1	1	1	100	100	100	100	92	84	94	100		
11	20	30	40	80	80	1	1	1	1	1	1	1	1	10	90	94	94	75	0	0	1	1	1	100	100	100	100	92	92	100	92		
12	20	30	40	50	1	1	1	1	1	1	1	1	10	20	70	84	90	0	0	0	1	1	1	100	100	100	100	92	94	100	82		
13	10	30	40	50	25	25	25	1	86	84	84	82	80	80	80	86	86	1	1	1	1	1	1	100	100	100	94	86	90	100	92		
14	1	20	40	50	25	25	25	1	80	80	80	80	80	80	92	90	80	96	1	0	45	45	92	94	80	100	96	90	96	100	100	100	
15	1	20	40	40	40	25	25	1	80	80	80	80	80	80	80	42	40	0	1	0	0	48	96	98	98	98	100	96	100	100	100	100	T.128 N.
16	1	10	30	40	40	1	1	1	1	80	80	80	80	80	84	40	49	0	1	0	0	0	96	90	100	100	100	100	100	100	1	100	
17	1	10	30	40	40	1	1	1	1	80	80	80	80	80	96	0	40	0	1	1	1	1	1	1	100	100	100	100	100	100	100	100	
18	1	10	30	40	40	1	1	1	1	80	80	90	84	84	94	88	84	40	1	1	1	1	1	1	100	100	100	100	100	100	100	100	
19	1	20	20	20	20	1	1	1	1	80	80	90	86	88	92	90	29	26	0	1	1	1	1	1	1	100	100	100	100	100	100	100	
20	1	20	10	1	1	1	1	1	1	80	90	100	96	92	88	90	30	30	0	1	1	1	1	1	100	100	100	100	100	100	100	100	
21	1	10	1	1	1	1	1	1	1	80	80	90	92	84	90	88	30	30	0	1	1	1	1	1	1	1	1	100	100	100	100	100	
22	1	10	1	1	1	1	1	1	1	80	80	90	90	84	90	90	80		1	1	1	1	1	1	1	1	100	100	100	100	100	100	
23	1	10	20	20	20	1	1	1	1	80	90	90	90	92	90	84	80	100	1	1	1	1	1	1	1	1	100	100	100	90	90	100	
24	10	10	20	20	20	1	1	1	1	80	80	90	90	100	90	84	84	100	100	1	1	1	1	1	1	1	100	100	90	90	90	94	
25	20	20	20	10	20	1	1	1	80	80	84	90	90	90	80	82	90	100	100	1	1	1	1	1	1	1	1	1		90	70	84	
26	40	30	20	10	1	1	1	80	80	90	94	100	94	84	80	80	90	100	100	100	1	1	1	1	1	1	1	1			80	84	
27	60	40	20	10	1	1	80	80	90	94	94	96	94	78	60	70	1	1	1	1	1	1	1	1	1	1			70	60	80	88	T.127 N.
28	60	30	1	1	1	80	80	80	90	98	94	94	94	92	40	1	1	1				1	1			1	40	50	50	70	90	90	
29	1	1	1	1	10	80	80	80	80	88	98	98	90	70	50	1	90	88	70	40	60	80	1				80	70	50	70	80	90	
30	1	1	1	1	1	80	80	80	84	76	96	94	92	80	60	80	80	88	70	50	40	60	1	0	0	90	84	90	94	90	90	90	
31	1	1	1	1	1	80	80	90	88	82	96	94	94	80	70	60	80	80	80	70	60	90		80	80	80	80	80	90	90	90	90	
32	1	1	1	1	1	90	90	100	90	78	98	92	94	90	84	80	86	90	94	90	70	88	80	70	80	80	80	80	90				
33	0	1	1	1	84	84	80	90	88	80	80	90	90	90	88	88	90	92	94	96	88	88	80	80	80	86	90	90	90	100	100	90	
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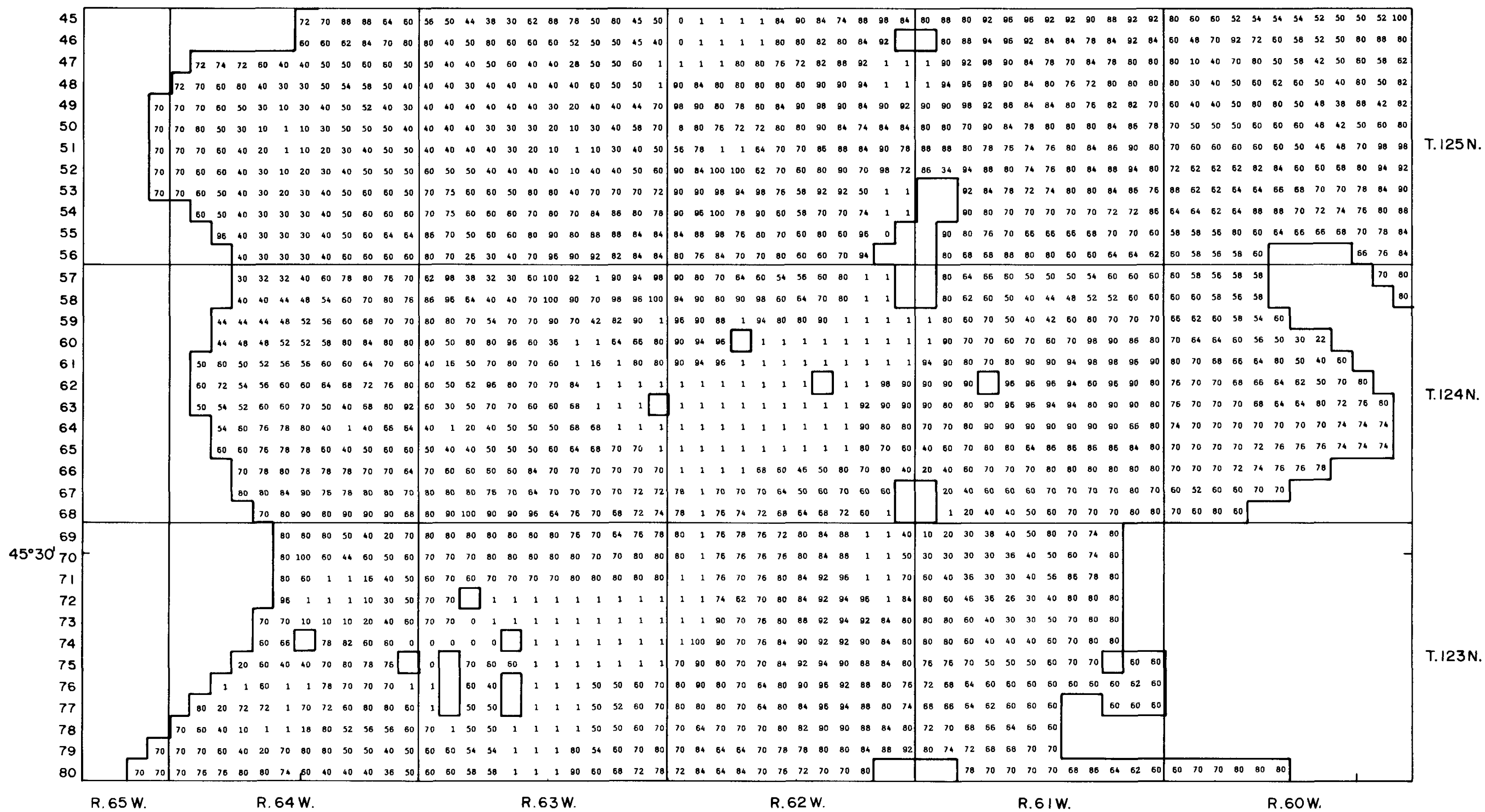
38	1	1	1	80	80	90	94	100	96	98	94	94	94	94	94	90	80	90	98	100	98	90	100	90	80	60	70	64	70	80	90	90	
39	1	1	80	80	90	90	92	90	92	94	92	98	98	98	96	94	90	94	100	100	100	90	80	90	100	80	100	82	84	86	88	88	T.126 N.
40	1	1	80	80	100	90	90	90	92	94	96	96	100	100	100	100	96	100	100	100	96	70	84	90	94	90	88	70	76	84	84	84	
41	1	1	70	80	84	80	80	80	84	88	92	98	100	100	100	100	100	100	100	100	96	84	90	94	94	84	72	40	60	80	84	88	
42	1	1	52	70	80	80	80	80	78	80	88	92	98	100	100	100	100	100	100	100	94	94	90	94	90	80	70	32	52	74	80	88	
43	1	80	80	80	80	84	84	80	78	88	72	88	92	98	100	100	100	100	100	100	96	94	90	90	84	78	64	50	60	72	80	74	
44	1	84	100	90	84	94	80	70	58	80	80	92	94	96	100	100	96	98	100	100	90	80	80	100	64	80	80	52	54	60	40	60	



0 1 2 3 4 5 6 MILES
0 1 2 3 4 5 6 KILOMETERS

EXPLANATION

70 **PERCENT RECHARGE**--Number is average value for each grid block, in percentage of available recharge which reaches the aquifer

—— **AQUIFER MODEL BOUNDARY**

Figure 18.--Average available recharge to the uppermost active aquifer for each grid block.

Storage in the Aquifers

With one exception, storage coefficients calculated from aquifer tests in Sanborn and Beadle Counties range from 0.00039 to 0.000017, 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 large as 0.28 were calculated from aquifer tests in the glacial-aquifer system in Hand County (fig. 1).

A storage coefficient of 0.0003 was used in the ground-water flow model to represent aquifer layers 1, 2, and 3 in the grid blocks where the average potentiometric head was higher than the average altitude of the top of the aquifer (artesian conditions). A specific yield of 0.20 was assigned to represent aquifer layers 1 and 2 when the average potentiometric head in the grid block was lower than the average altitude of the top of the aquifer for the same grid block (water-table conditions).

Recharge to the Aquifers

The principal source of recharge to the aquifer system is precipitation. The average annual precipitation is about 18 inches in the model area. The areal distribution of recharge to the aquifer was based on analyses of precipitation data and on the thickness of the confining bed overlying the uppermost aquifer (fig. 7). The thickness of this confining bed controls the rate at which the underlying aquifer can be recharged. Recharge occurs rapidly where there are permeable sediments overlying the uppermost aquifer. When the clay and silt are sufficiently thick (generally greater than 50 ft), there probably is little or no recharge by infiltration.

Hedges and others (1983) calculated recharge rates to the Elm aquifer by flow-net analysis. Recharge ranged from 0.06 in/yr in northern Brown County to 0.42 in/yr in southern Brown County. The recharge rates estimated from computer-model analyses of the glacial-aquifer system in the Sanborn-Beadle County area ranged from 0 to 7.0 in/yr with an average of 0.97 in/yr (Emmons, 1988). The glacial-aquifer system in the Sanborn-Beadle County area is similar in character to the system in Brown County.

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 average recharge to the aquifer was 7.0 in/yr and occurred only where aquifer material was at land surface. When the average confining bed thickness was greater than zero, the rate of recharge to the underlying aquifer decreased linearly to 0.0 in/yr at 50 ft below average land surface. No recharge occurs when the average confining bed thickness exceeds 50 ft. Figure 18 shows the percentage of available recharge that reaches the aquifer. Because an empirical relation was developed between recharge and thickness of the confining bed overlying the aquifer, values should not be considered absolute. The values are hydrologically reasonable and provide the best overall model results.

Using the percentage of potential recharge which can reach the uppermost aquifer and potential steady-state recharge of 7.0 inches, the areally distributed, average steady-state recharge to the aquifer system is 4.3 in/yr. This is in poor agreement with Hedges and others (1983), calculated recharge rates of 0.06 in/yr for northern Brown County and 0.42 in/yr for southern Brown County. The reason for the poor agreement is the result of including the sandy lake deposits as part of aquifer layer 1. The confining bed overlying the lake deposits tends to be thinner than those overlying the Elm aquifer, thereby skewing the average steady-state recharge.

Monthly potential recharge to the aquifers (table 2) was estimated using the monthly average precipitation data (U.S. Department of Commerce, published annually) and hydrograph analysis. Examination of the hydrographs of wells completed in the Elm and Middle James aquifers indicates that most of the recharge to the aquifers occurs in the spring months (figs. 8 and 10). The multiplication factors to convert average monthly precipitation data to monthly potential recharge was adjusted as part of the transient model calibration process to obtain the best fit between the observed and calculated potentiometric heads in the three aquifers.

Evapotranspiration from the Aquifers

The areal distribution of evapotranspiration from the aquifer is controlled by the potential evapotranspiration, the thickness of the confining bed overlying the aquifer, and the depth of the water below land surface. The potential evapotranspiration in the study area is estimated to be 77 percent of the pan evaporation (Farnsworth and others, 1982). The average class A pan evaporation at Aberdeen for 1946-55 is about 46 in/yr (Spuhler and others, 1971). The calculated average potential evapotranspiration in the study area is about 35.4 in/yr.

The areal distribution of the potential evapotranspiration from the aquifers was tested and refined as part of the steady-state simulation process. The potential evapotranspiration rate can occur only where no confining bed is present above the uppermost aquifer. Even though the potentiometric head in the aquifers may be close to land surface, it is assumed that the confining beds overlying the aquifers will restrict upward movement of water and reduce the potential evapotranspiration rate. When the average confining bed thickness is between zero and 50 ft, the potential evapotranspiration rate decreases linearly from 35.4 to 0.0 in/yr. The potential evapotranspiration is 0.0 in/yr for confining bed thicknesses greater than 50 ft. Figure 19 shows the percentage of the potential evapotranspiration available from the uppermost aquifer. The uppermost aquifer layer is generally aquifer layer 1, however, when not present, the percentage of the potential evapotranspiration is calculated for the next deeper aquifer layer.

The evapotranspiration from the aquifers is controlled also by the depth of the potentiometric surface for the uppermost aquifer below land surface. When the potentiometric head is at or greater than 9.0 ft below land surface, evapotranspiration is no longer simulated.

Monthly evapotranspiration was estimated for 1983-85 using U.S. Department of Commerce data (1983-85) and the relationship between pan evaporation and evapotranspiration described by Farnsworth and others (1982). No pan evaporation is available in the study area for 1983-85. Based on pan evaporation data at Redfield, South Dakota, about 40 mi south of Aberdeen and at Pickstown, 220 mi south of Aberdeen, evaporation was estimated and evapotranspiration was calculated for the study area (table 3). The calculated maximum potential evapotranspiration rate was 38.45 inches in 1983, 34.69 inches in 1984, and 32.14 inches in 1985. The minimum potential evapotranspiration was 0.0 inch in the winter months when the ground is frozen.

Because an empirical relation 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 and provide the best overall model results.

Table 2.--Average monthly precipitation and estimated recharge

Month	Recharge multiplication factor[1]	Normal: Average precipitation for Aberdeen and Columbia[2] (inches)	Normal: Maximum recharge to the glacial-aquifer system[3] (inches)	1983: Average precipitation for Aberdeen and Columbia[2] (inches)	1983: Maximum recharge to the glacial-aquifer system[3] (inches)	1984: Average precipitation for Aberdeen and Columbia[2] (inches)	1984: Maximum recharge to the glacial-aquifer system[3] (inches)	1985: Average precipitation for Aberdeen and Columbia[2] (inches)	1985: Maximum recharge to the glacial-aquifer system[3] (inches)
January	0.00	0.49	0.00	0.24	0.00	0.48	0.00	0.30	0.00
February	.00	.60	.00	.16	.00	.83	.00	.08	.00
March	.70	.99	.69	2.82	1.13	1.71	.68	1.48	.59
April	.90	2.06	1.85	.68	.61	2.66	2.39	.54	.49
May	.80	2.60	2.08	1.66	1.16	1.16	.81	3.58	2.51
June	.40	3.38	1.35	3.68	1.10	5.94	1.78	1.94	.58
July	.05	2.34	.12	5.78	.29	2.10	.10	2.34	.12
August	.05	2.00	.10	2.60	.13	2.24	.11	2.75	.14
September	.25	1.55	.39	1.20	.24	.84	.17	2.50	.50
October	.40	1.06	.42	1.37	.55	3.34	1.34	.70	.28
November	.00	.62	.00	.64	.00	.06	.00	1.54	.00
December	.00	.50	.00	.48	.00	.59	.00	.45	.00
Total	--	18.19	7.00	21.31	5.21	21.95	7.38	18.20	5.21

[1]The maximum decimal fraction of average precipitation which could potentially recharge the glacial-aquifer system.
[2]Data from the U.S. Department of Commerce, 1983-85.
[3]Calculated by multiplying the average monthly precipitation by the recharge multiplication factor.

Table 3.--Estimated monthly pan evaporation and estimated potential evapotranspiration

[--, no data or not calculated]

Month	1983		1984		1985	
	Pan evaporation[1] (inches)	Potential evapo-transpiration[2] (inches)	Pan evaporation[1] (inches)	Potential evapo-transpiration[2] (inches)	Pan evaporation[1] (inches)	Potential evapo-transpiration[2] (inches)
January	--	--	--	--	--	--
February	--	--	--	--	--	--
March	--	--	--	--	--	--
April	4.19	3.23	6.51	5.01	7.27	5.60
May	11.72	9.02	7.22	5.56	8.06	6.21
June	6.54	5.04	7.54	5.81	7.38	5.68
July	9.08	6.99	8.36	6.44	9.13	7.03
August	9.35	7.20	8.12	6.25	6.06	4.67
September	6.05	4.66	4.35	3.35	3.83	2.95
October	3.00	2.31	2.95	2.27	2.60	2.00
November	--	--	--	--	--	--
December	--	--	--	--	--	--
Total	49.93	38.45	45.05	34.69	44.33	32.14

[1]Estimated using pan evaporation data for Redfield and Pickstown, South Dakota; U.S. Department of Commerce, 1983-85.
[2]Calculated as 0.77 times the monthly evaporation.

ROW	12	13	14	15	16	17	18	19	20	21	22	23	24	25	26	27	28	29	30	31	32	33	34	35	36	37	38	39	40
9				78	84	87	44	11	0	22	44	56	87	67	87	87	89	58	44	33	22	0	44	69	100	67	0	0	0
10				76	78	56	33	22	0	10	30	40	56	56	56	56	56	44	33	22	11	0	11	44	67	67	0	0	0
11				73	87	44	38	29	22	15	25	40	56	56	44	44	33	22	22	11	11	22	33	44	87	87	0	0	0
12				73	56	33	38	42	44	25	45	50	58	44	33	33	11	0	0	2	4	22	33	44	56	0	0	0	0
13				71	44	22	44	58	67	67	67	67	56	44	22	0	0	0	0	0	0	11	33	44	56	0	0	0	0
14				89	33	11	22	67	89	89	78	67	44	22	0	0	0	0	0	0	0	0	22	44	56	0	0	0	0
15				67	22	0	22	67	98	89	78	58	44	22	0	0	0	0	0	0	0	0	22	44	44	35	0	0	0
16				71	44	22	33	56	89	78	67	44	33	11	0	0	0	0	0	0	0	0	11	33	44	44	0	0	0
17				73	56	33	44	56	78	67	56	33	22	11	0	0	0	0	0	0	0	0	11	33	44	44	0	0	0
18				76	67	44	56	67	67	56	44	22	11	0	0	0	0	0	0	0	0	0	11	33	44	44	0	0	0
19				78	73	67	67	67	56	44	33	22	11	0	0	0	0	0	0	0	0	0	22	22	22	22	0	0	0
20				89	84	78	67	44	33	22	11	0	0	0	0	0	0	47	0	0	0	0	22	11	0	0	0	0	0
21				89	84	78	78	58	44	33	22	22	0	0	0	0	0	0	0	0	0	0	11	0	0	0	0	0	0
22				89	84	80	80	67	56	44	33	33	0	0	0	0	0	0	0	0	0	0	11	0	0	0	0	0	0
45°50' 23				89	84	80	80	67	56	44	33	33	0	0	0	0	0	0	0	0	0	0	11	22	22	22	0	0	0
24				89	84	80	82	67	56	44	44	33	0	0	0	0	0	0	0	0	0	11	11	22	22	22	0	0	0
25				89	84	60	82	67	67	56	56	44	0	0	0	0	0	0	0	0	0	22	22	22	11	22	0	0	0
26				89	84	80	84	67	67	67	67	44	0	0	0	0	0	0	0	0	0	44	33	22	11	0	0	0	89
27				89	84	80	84	78	78	78	78	44	0	0	0	0	0	0	0	11	33	67	44	22	11	0	0	89	89
28				84	78	67	78	78	84	84	84	56	0	0	0	0	0	0	11	22	22	67	33	0	0	0	89	89	89
29				78	67	56	67	78	84	84	84	44	11	0	0	0	0	0	0	0	0	0	0	0	0	11	89	89	89
30				78	67	67	67	78	76	71	67	44	22	22	11	0	0	0	0	0	0	0	0	0	0	0	89	89	89
31				78	67	67	67	67	67	56	44	44	33	33	33	11	0	0	0	0	0	0	0	0	0	0	89	89	100
32				78	67	67	56	56	56	44	44	44	33	33	22	0	0	0	0	0	0	0	0	0	0	0	100	100	100
33				78	67	56	44	44	44	44	44	44	33	33	11	0	0	0	0	25	25	25	0	0	0	93	93	89	100
34				78	69	67	44	44	33	33	33	55	45	35	22	22	0	0	0	25	25	25	0	0	0	89	89	89	98
35				78	71	67	44	33	22	33	33	55	65	50	22	22	0	0	0	25	25	25	0	0	0	93	93	93	96
36				78	73	67	44	22	11	5	33	65	55	50	33	33	0	0	0	0	0	0	0	0	89	93	93	93	100
37			78	78	76	67	56	44	33	33	67	89	67	56	44	44	0	0	0	0	0	0	0	0	89	69	89	100	100
38			78	78	78	67	67	67	56	67	89	96	89	67	44	44	0	0	0	0	0	0	0	0	89	89	100	100	100
39		78	78	78	78	76	78	78	78	89	89	96	89	78	87	87	0	0	0	0	0	0	0	89	89	100	100	100	100
40	80	76	76	76	76	78	78	78	78	87	89	89	84	80	67	67	0		0	0	0	0	0	69	89	100	100	100	100
41	80	76	78	78	76	76	76	76	76	84	82	78	82	80	78	78	0	0	0	0	0	0	0	78	89	93	89	89	89
42	76	76	76	78	76	76	76	76	76	80	73	67	78	78	80	64	89	0	0	0	0	0	0	58	78	89	89	89	89
43	71	76	76	73	73	73	73	73	73	76	67	56	76	76	76	78	84	0	0	0	0	0	89	69	89	89	93	93	89
44	80	78	76	73	71	71	71	71	71	71	60	49	71	67	67	80	85	25	0	0	0	0	93	100	100	93	100	89	76

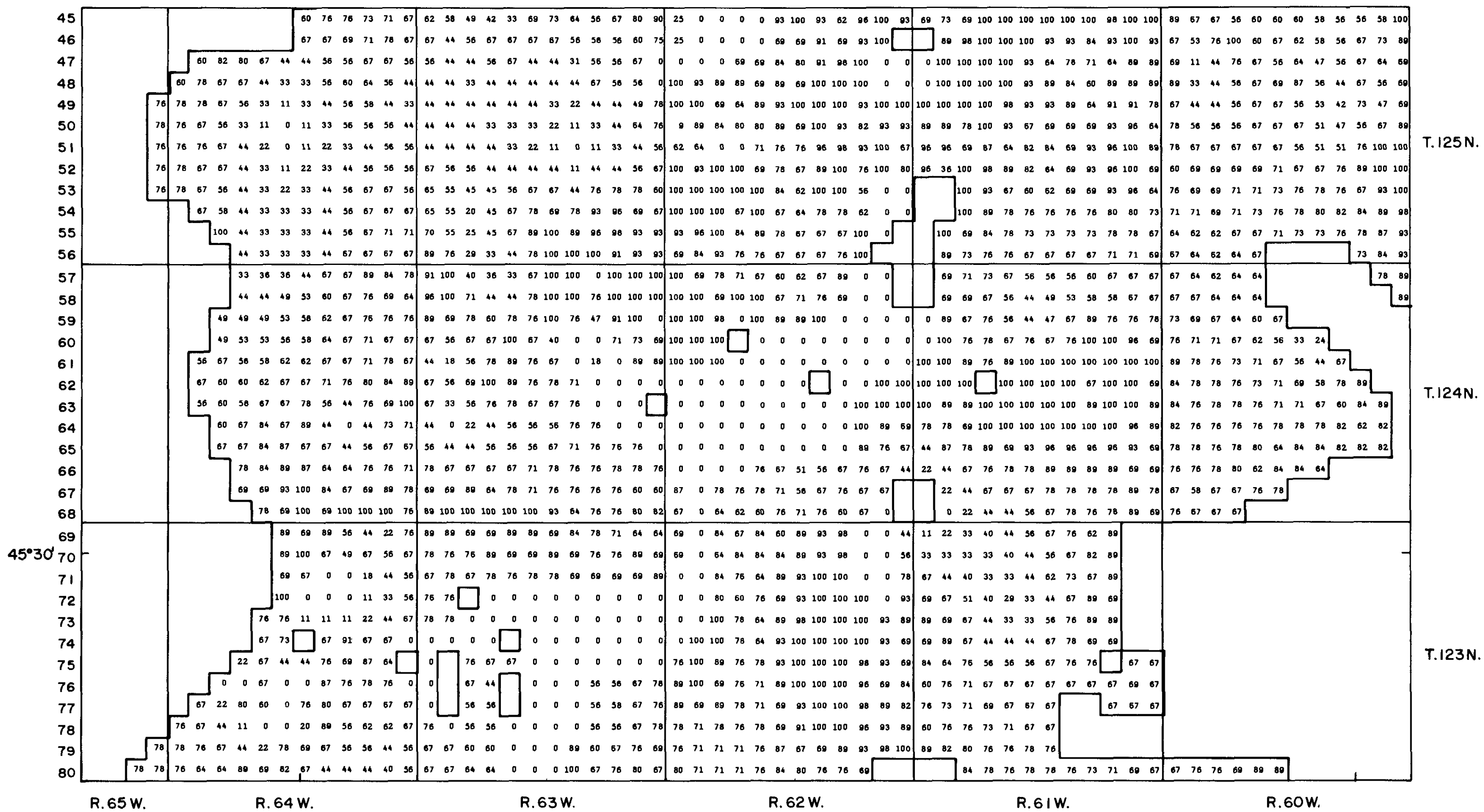
ROW	41	42	43	44	45	46	47	48	49	50	51	52	53	54	55	56	57	58	59	60	61	62	63	64
9	0	0	0	0	0	80	100	11	100	30	20	20	0	0	0	0	100	100	100	100	100	0	0	
10	0	0	0	0	0	78	78	89	100	100	20	20	0	0	0	100	100	100	100	100	93	100	100	
11	0	0	0	0	0	11	100	100	100	100	20	20	0	0	0	100	100	100	100	100	100	100	100	
12	0	0	0	0	11	22	78	93	100	20	20	20	0	0	0	100	100	100	100	100	100	100	91	
13	96	93	93	91	89	89	89	96	96	0	0	0	0	0	0	100	100	100	100	98	100	100	100	
14	89	89	89	89	89	89	100	100	89	100	50	50	100	100	100	100	89	100	100	100	100	100	100	100
15	89	89	89	69	89	89	89	93	89	20	25	50	50	100	100	100	100	100	100	100	100	100	100	100
16	0	89	69	89	89	89	93	89	100	20	25	50	50	50	100	100	100	100	100	100	100	100	0	100
17	0	89	89	89	89	89	100	25	89	25	0	0	0	0	0	0	100	100	100	100	100	100	100	100
18	0	89	89	100	93	93	100	98	93	44	0	0	0	0	0	0	100	100	100	100	100	100	100	100
19	0	89	89	100	96	98	100	100	98	89	25	0	0	0	0	0	0	100	100	100	100	100	100	100
20	0	89	100	100	100	100	98	100	100	100	25	0	0	0	0	0	100	100	100	100	100	100	100	100
21	0	89	89	100	100	93	100	98	100	100	25	0	0	0	0	0	0	0	0	100	100	100	100	100
22	0	89	89	100	100	93	100	100	89		0	0	0	0	0	0	0	0	100	100	100	100	100	100
45°50' 23	0	89	100	100	100	100	100	93	89	100	0	0	0	0	0	0	0	0	100	100	100	100	100	100
24	0	89	89	100	100	100	100	93	93	100	100	0	0	0	0	0	0	0	100	100	100	100	100	100
25	89	89	93	100	100	100	89	91	100	100	100	0	0	0	0	0	0	0	0	0		100	78	93
26	89	100	100	100	100	93	89	89	100	100	100	100	0	0	0	0	0	0	0	0			89	93
27	100	100	100	100	100	84	67	78	0	0	0	0	0	0	0	0	0	0			78	67	89	98
28	100	100	100	100	100	100	44	0	0	0				0	0			0	44	56	56	78	100	100
29	89	98	100	100	100	78	56	0	100	98	78	44	67	89	0				89	78	56	78	89	100
30	93	84	100	100	100	89	67	89	67	98	78	56	44	67	0	0	0	100	93	100	100	100	100	100
31	98	91	100	100	100	89	78	67	89	89	89	78	67	100		89	89	89	89	89	100	100	100	100
32	100	84	100	100	100	100	93	89	96	100	100	100	78	98	89	78	89	89	89	89	100			
33	98	67	89	100	100	100	98	98	100	100	100	100	98	98	89	89	89	96	100	100	100	100	100	100
34	98	89	100	100	98	96	96	100	100	100	100	100		100	89	91	93	98	100	100	100	93	89	89
35	96	93	96	98	96	93	96	100	100	100	100	100	100	100	100	100	98	98	98	100	98	93	89	84
36	98	98	98	98	93	91	98	100	100	100	100	100	100	69	100	100	96	89	89	78	84	93	100	100
37	100	100	100	100	100	98	100	100	100	100	100	100	100	78	100	93	69	78	78	76	80	69	100	100
38	100	100	100	100	100	100	100	100	89	100	100	100	100	100	100	100	89	67	78	71	78	89	100	100
39	100	100	100	100	100	100	100	100	100	100	100	100	100	100	89	100	100	89	100	91	93	96	98	98
40	100	100	100	100	100	100	100	100	100	100	100	100	100	78	93	100	100	100	98	78	84	93	93	93
41	93	98	100	100	100	100	100	100	100	100	100	100	100	93	100	100	100	93	80	44	67	89	93	98
42	84	89	98	100	100	100	100	100	100	100	100	100	100	100	100	100	100	89	78	36	58	82	69	98
43	84	98	80	96	100	100	100	100	100	100	100	100	100	100	100	100	93	84	71	56	67	80	67	82
44	64	69	89	100	100	100	100	100	100	100	100	100	100	89	89	100	71	89	67	56	60	67	44	67

COLUMN 1–64; 98°30' (at columns 10–11); 98° (at column 62)

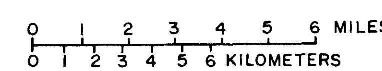
T.128 N. (rows 9–20)

T.127 N. (rows 21–32)

T.126 N. (rows 33–44)



EXPLANATION



78 **PERCENT EVAPOTRANSPIRATION--Number is value for each grid block, in percentage of available evapotranspiration which can occur from the uppermost active aquifer**

——— **AQUIFER MODEL BOUNDARY**

Figure 19.--Average available evapotranspiration from the uppermost active aquifer for each grid block.

Pumpage from the Aquifers

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.

Ground-water withdrawal data are required to simulate the glacial-aquifer system. Pumpage data were collected for 1973 through 1985. Before 1973, little pumping occurred from the aquifers. The aquifers are in approximately 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 (figs. 8 and 10). The average 1973 through 1982 pumpage from aquifer layer 1 used in the simulation of the steady-state model was about 1.66 ft^3/s. No pumpage was reported for aquifer layers 2 and 3 for 1973 through 1982.

Pumpage for 1983 through 1985 was used to simulate the monthly pumping from the aquifer system (table 4). Pumpage is large during the summer months because irrigation is the major use of water from the aquifers.

Hydraulic Connection Between the Rivers and Aquifers

Where the glacial-aquifer system is hydraulically connected to a river, stream, or surface-water body, they may contribute water to the aquifer or drain water from the aquifer, depending on the head gradient. According to Koch and Bradford (1976), aquifer layer 1 discharges into Foot Creek and the Elm River although no data on streamflow gains are available. Considering the extent and thickness of aquifer layer 1 and thickness of the overlying confining bed, the hydraulic connection between Foot Creek and aquifer layer 1 is poor. The Elm River is hydraulically connected to aquifer layer 1 in 32 grid blocks (fig. 14).

SIMULATION OF GROUND-WATER FLOW IN THE GLACIAL-AQUIFER SYSTEM

Calibration of the Ground-Water Flow Model

Model calibration is the process by which model parameters are adjusted so the model will adequately simulate historical potentiometric heads and flows. The initial equilibrium conditions were simulated by entering average river stage, recharge, evapotranspiration, and pumpage and by setting the storage in the aquifers 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-1983 potentiometric heads to assess the accuracy of the steady-state simulation. The monthly transient simulation includes storage and time-dependent river stage, recharge, evapotranspiration, and pumpage. Simulated monthly transient potentiometric heads were compared to observed monthly potentiometric heads.

Calibration involves varying the values of hydraulic conductivity, recharge, evapotranspiration, leakance, 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 in each aquifer was obtained.

Table 4.--Monthly ground-water withdrawal rates from the study area, 1983-85

[--, assumed to be zero]

Year	Month	Ground-water withdrawal rates (cubic feet per second) Aquifer layer 1	Aquifer layer 2	Aquifer layer 3
1983	January	--	--	--
	February	--	--	--
	March	--	--	--
	April	0.120	0.008	--
	May	.291	.575	0.054
	June	2.487	3.013	.440
	July	3.930	6.287	.565
	August	4.220	5.288	.652
	September	1.120	2.950	.080
	October	.038	.022	--
	November	--	--	--
	December	--	--	--
1984	January	--	--	--
	February	--	--	--
	March	--	--	--
	April	--	--	--
	May	.306	.202	.042
	June	.349	.666	.105
	July	5.596	6.766	.578
	August	10.492	12.693	.645
	September	3.289	6.007	.214
	October	--	.260	--
	November	--	--	--
	December	--	--	--
1985	January	--	--	--
	February	--	--	--
	March	--	--	--
	April	.031	.019	--
	May	.376	.706	.028
	June	1.030	2.796	.144
	July	3.900	10.307	.563
	August	4.548	11.391	.331
	September	1.318	4.682	.120
	October	.062	.038	--
	November	--	--	--
	December	--	--	--

Tables 5 and 6 give 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 scattered 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 aquifers and their stratigraphic relations can result in seemingly unusual water levels. In addition, nearby pumping can result in observed water levels which do not reflect conditions throughout the area. Inaccurate measurement or error in 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 potentiometric heads and the observed potentiometric heads. Tables 5 and 6 reflect the best composite set of average and absolute differences obtained between the simulated and observed potentiometric heads for the steady-state simulation and the 1985 monthly 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 are included in the simulation.

As indicated by the hydrographs (figs. 8 and 10), pumpage for irrigation from the aquifers (table 4) have not produced significant water-level declines in the aquifers. The aquifers generally are in equilibrium; water levels usually recover during the nonirrigation fall, winter, and spring seasons.

The simulated steady-state potentiometric surfaces are shown in figures 20-22. There are water-level data from 22 observation wells completed in aquifer layer 1 for the period before 1983 and from 13 observation wells completed in aquifer layer 2. There are no water-level data available for aquifer layer 3, however, observation well 121N65W34CCCC located south of the study area was used as an aid for estimating water levels in the study area. The maximum positive head difference between the simulated and observed water levels in aquifer layer 1 was 8.41 ft in observation well 126N63W10AAAA located in grid block row 35, column 24 and the maximum negative difference was 11.78 ft in observation well 125N63W29CCCC in grid block row 54, column 19. The maximum positive head difference between the simulated and observed water levels in aquifer layer 2 was 17.17 ft in observation well 128N60W19BBCC located in grid block row 15, column 53 and the maximum negative difference was 5.43 ft in observation well 123N64W34BBBB in grid block row 79, column 11.

Table 5.--Comparison between simulated and observed potentiometric heads for aquifer layer 1

[--, no data]

Model simulation	Average 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	0.78	4.59	8.41	11.78	22
Transient, 1985					
January	1.48	4.96	10.59	11.47	9
February	1.44	6.31	11.89	16.08	18
March	--	--	--	--	--
April	-.76	4.68	7.76	96.08	21
May	.59	5.09	11.79	10.79	21
June	-.86	4.63	8.05	13.62	22
July	-2.54	4.66	6.24	17.31	22
August	-2.53	5.00	5.79	17.16	19
September	-1.12	4.16	5.51	11.89	17
October	-2.31	4.18	5.34	16.34	10
November	-.98	4.93	8.71	16.17	19
December	--	--	--	--	--

[1]Summation of simulated minus observed potentiometric heads in corresponding grid blocks divided by number of observation wells with observed potentiometric heads. Positive number indicates simulated head was higher than the observed head; negative number indicates simulated head was lower than the observed head.

[2]Summation of the absolute values of simulated minus observed potentiometric heads in corresponding grid blocks divided by number of observation wells with observed potentiometric heads. 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.

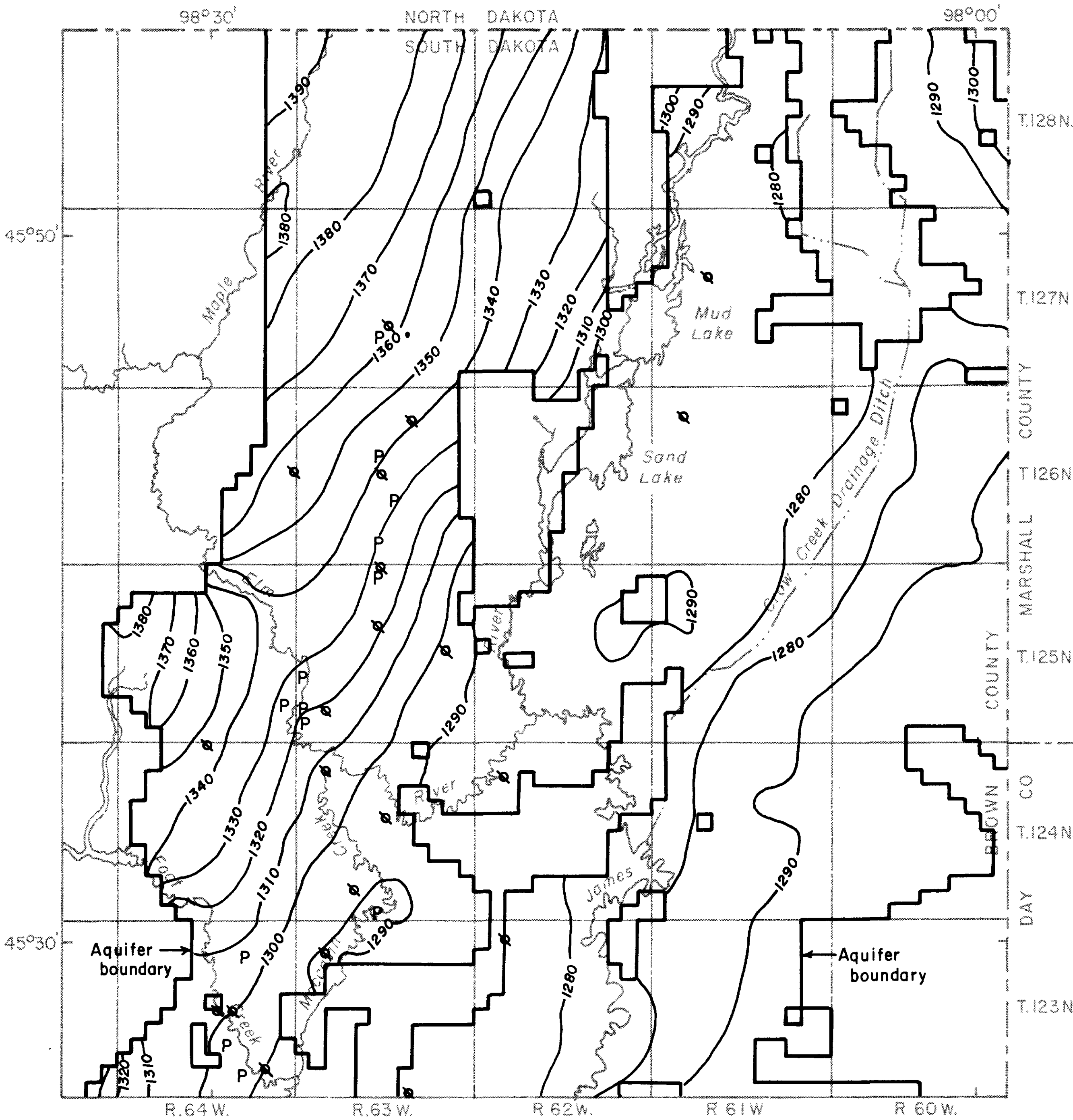
[4]Negative difference when simulated head is less than observed water level.

Table 6.--Comparison between simulated and observed potentiometric heads for aquifer layer 2

[--, no data]

Model simulation	Average 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	3.49	5.10	17.17	5.43	13
Transient, 1985					
January	1.67	6.90	25.76	19.11	13
February	2.15	5.43	27.01	18.49	20
March	--	--	--	--	--
April	-1.22	3.96	9.48	18.45	23
May	.93	5.28	32.67	17.11	23
June	-.49	5.15	27.59	18.77	23
July	2.21	7.37	46.35	21.48	23
August	3.21	8.23	42.26	21.87	23
September	2.99	7.21	29.17	21.88	21
October	4.98	7.44	15.92	6.45	9
November	2.75	6.97	17.46	21.30	22
December	--	--	--	--	--

[1]Summation of simulated minus observed potentiometric heads in corresponding grid blocks divided by number of observation wells with observed potentiometric heads. Positive number indicates simulated head was higher than the observed head; negative number indicates simulated head was lower than the observed head.
[2]Summation of the absolute values of simulated minus observed potentiometric heads in corresponding grid blocks divided by number of observation wells with observed potentiometric heads. 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.



EXPLANATION

—1300— SIMULATED POTENTIOMETRIC-SURFACE CONTOUR-- Shows altitude of simulated potentiometric surface based on average hydrologic conditions, 1972-82. Contour interval is 10 feet. Datum is sea level

ø OBSERVATION WELL

P PUMPING NODE

0 2 4 6 MILES
0 2 4 6 KILOMETERS

Figure 20.--Simulated steady-state potentiometric surface of aquifer layer 1.

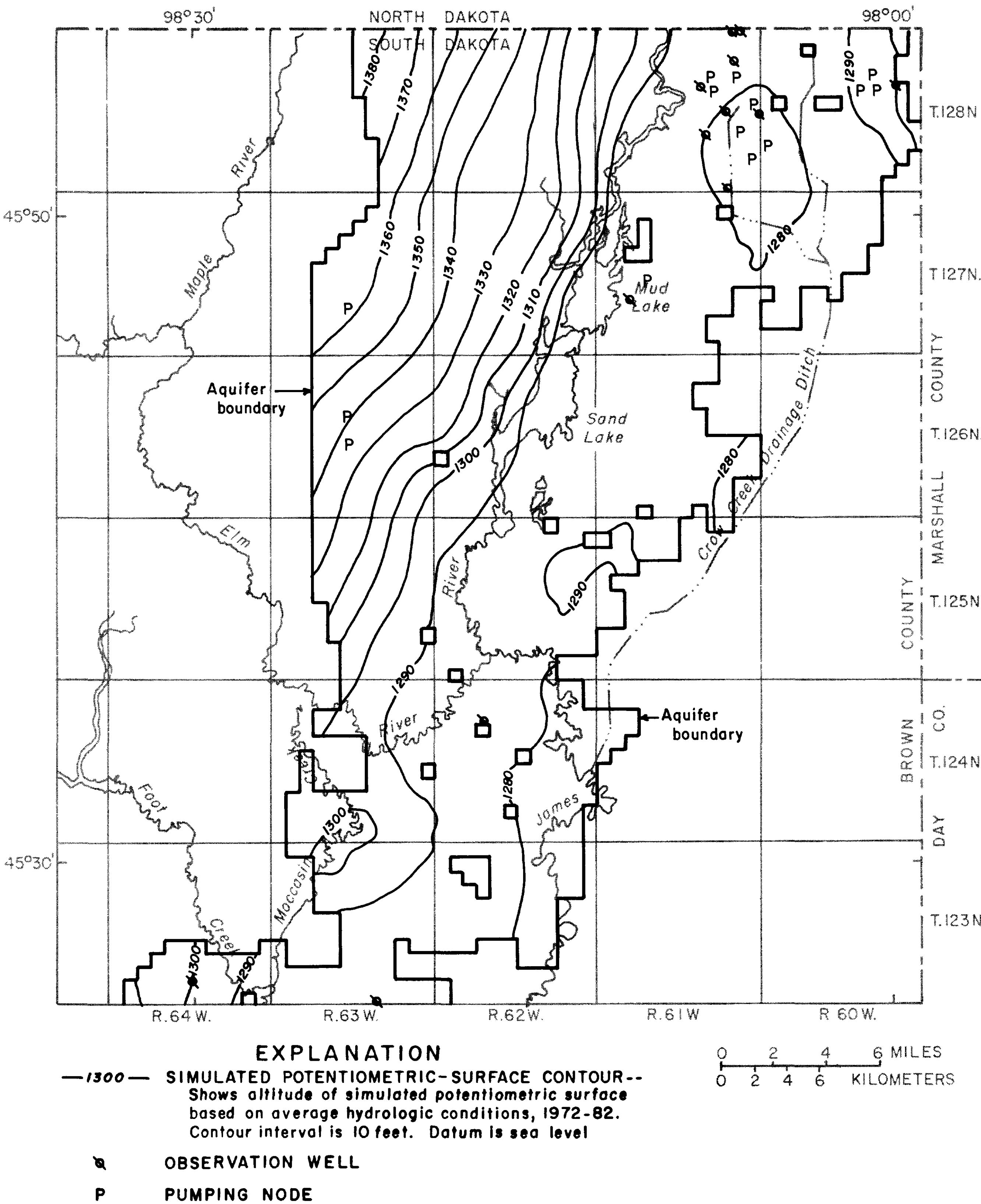
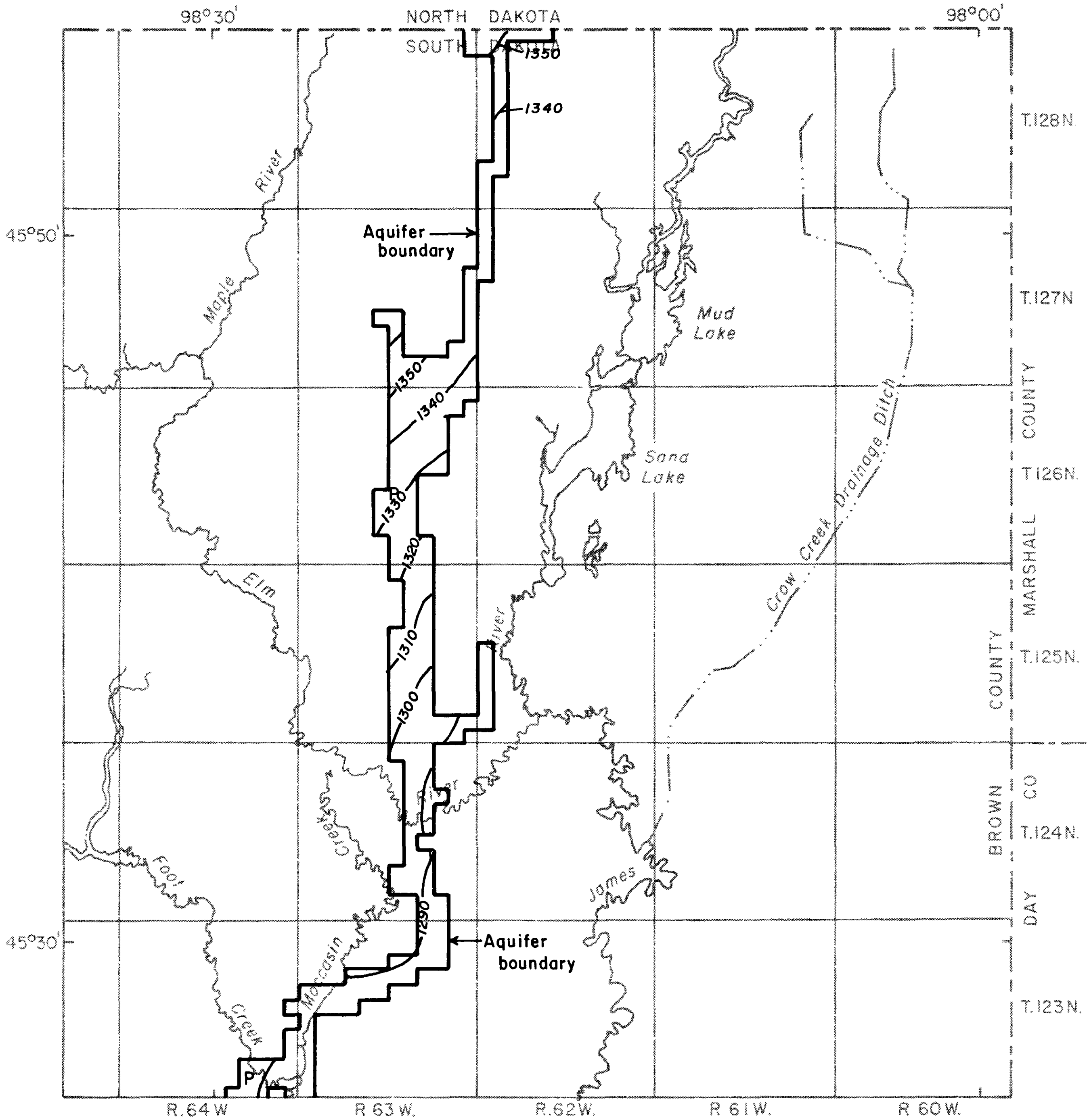


Figure 21.--Simulated steady-state potentiometric surface of aquifer layer 2.



EXPLANATION

—*1300*— SIMULATED POTENTIOMETRIC-SURFACE CONTOUR-- Shows altitude of simulated potentiometric surface based on average hydrologic conditions, 1972-82. Contour interval is 10 feet. Datum is sea level

P PUMPING NODE

0 2 4 6 MILES
0 2 4 6 KILOMETERS

Figure 22.--Simulated steady-state potentiometric surface of aquifer layer 3.

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 of that aquifer. Also due to the simplifying assumptions in the model and the size of the finite-difference grid, the simulated steady-state potentiometric heads will contain inaccuracies. However, the model is one of the best means of improving and evaluating our understanding of the glacial-aquifer system.

The highest potentiometric heads in aquifer layer 1 are located on the western boundary of the study area and the lowest heads are located near the James River which is topographically the lowest area (fig. 20). The potentiometric contours in aquifer layer 1 approximate the topographic contours. Ground water flows from higher to lower potentiometric head and perpendicular to the potentiometric contours. The flow west of the James River generally is eastward toward the river, and east of the James River the flow generally is westward.

Previous studies have indicated that the James River gains little or no water from the underlying glacial-aquifer system in the study area. The steady-state simulation shows that evapotranspiration can reasonably remove enough water from the glacial-aquifer system before it reaches the river to approximate the steady-state potentiometric surface.

The potentiometric heads and flow within aquifer layer 2 (fig. 21) are similar to those of aquifer layer 1. The highest potentiometric heads are located on the western boundary of the study area and the lowest heads are located in the topographically low area near the James River. Flow in aquifer layer 2 generally is to the east, west of the James River and flow generally is to the west, east of the James River.

The steady-state simulation indicates that the potentiometric heads in aquifer layer 3 are highest at the north end of the study area and lowest at the south end (fig. 22). Simulated flow in aquifer layer 3 is from the northwest to the southeast. Based on a couple of measuring points, Koch and Bradford (1976) stated that the direction of ground-water movement is to the north. There are no water-level data available to check the validity of the aquifer layer 3 steady-state simulation.

One of the least understood aspects of the glacial-aquifer system is the potentiometric head relations and flow between the aquifers. The model provides a means of making head comparisons between the aquifers over the study area. The steady-state head differences between aquifer layers 1 and 2 (figs. 20 and 21) were all less than 1.0 ft and most were 0.1 ft or less.

Recharge from precipitation was 94.8 percent of the steady-state inflow to the glacial-aquifer system (table 7). The average annual recharge distributed over the active model area was 4.3 inches. Evapotranspiration accounts for 95.8 percent of the outflow from the predevelopment steady-state glacial-aquifer system. The average evapotranspiration distributed over the active model area was 4.6 in/yr.

Table 7.--Simulated water budget under steady-state conditions

Budget component	Flow rates in cubic feet per second	Percent
INFLOW		
Recharge to the glacial-aquifer system from precipitation	327	94.8
Recharge to the glacial-aquifer system from the stream	.33	.1
Inflow at specified-head boundaries	17.6	5.1
Total inflow (rounded)	345	100.0
OUTFLOW		
Evapotranspiration from the glacial-aquifer system	333	95.8
Pumpage	1.61	.4
Discharge from the glacial-aquifer system to the stream	1.54	.5
Outflow at specified-head boundaries	11.5	3.3
Total outflow (rounded)	348	100.0

Transient Simulation

The transient simulation includes changes in storage in the aquifers. The hydrographs of water levels in the 3 aquifers (figs. 8, 10, and 12) indicate that the aquifers generally are in equilibrium. Water levels decline during the summer months due to increased evapotranspiration and pumpage and rise to near nonpumping water levels during the late winter and spring due to increased recharge.

Since no long-term water-level declines have occurred in the aquifer system, the aquifers were simulated in monthly pumping periods to simulate seasonal changes in the aquifers' water levels. Twelve monthly pumping periods for the year 1985 are simulated. The year 1985 was selected because it has the most water-level data available to check the accuracy of the monthly simulation. To ensure adequate starting heads for the January 1985 simulation, monthly simulations for 1983 and 1984 were also run. The potentiometric heads calculated by the December 1984 simulation were used as the starting heads for January 1985. Water levels at the beginning of 1985 (figs. 8, 10, and 12) were higher than average due to above-normal precipitation (table 2) and subsequent recharge to the aquifers. Subsequent 1985 monthly starting heads used potentiometric heads generated by the preceding monthly simulation.

The average difference between the simulated and observed water levels for the 12 monthly simulation periods (tables 5 and 6) ranged from -2.54 ft in July to 1.48 ft in January for aquifer layer 1, from -1.22 ft in April to 4.98 ft in October for aquifer layer 2. The average absolute difference ranged from 4.16 ft in September to 6.31 ft in February for aquifer layer 1, from 3.96 ft in April to 8.23 ft in August for aquifer layer 2. Insufficient data exists for aquifer layer 3 to calculate differences between the simulated and observed water levels. The number of observation wells is a major factor in the average and absolute monthly differences. The larger the number of observation wells, the better the average differences indicate how well the model simulates the aquifers.

Comparison of hydrographs of observed water levels and simulated potentiometric heads for the corresponding grid blocks in which the observation wells are located are a means of determining the ability of the model to simulate the glacial-aquifer system (figs. 23-25). The hydrographs indicate that there are areas where the model is capable of simulating the aquifers accurately and there are areas where the model does not. The reason for the poor correlation between the simulated and observed hydrographs for May and June possibly is a result of the distribution of precipitation in April and May (table 2). The average precipitation for April was only 0.54 inch or about 26 percent of normal which probably could result in a soil moisture deficiency. Although the average May precipitation was 3.58 inches or about 138 percent of normal, a potential soil moisture deficiency would result in a lower than anticipated recharge to the aquifer. This variability indicates the complexity of the glacial-aquifer system and the apparent need for accurate recharge data.

The hydrographs indicate that the water levels in the aquifers generally are highest in April and lowest in August. The simulated potentiometric surfaces of aquifer layers 1, 2, and 3 for April 1985 are shown in figures 26-28. In April, the average difference between the simulated and observed water levels was -0.76 ft and the average absolute difference was 4.68 ft for aquifer layer 1 (table 5). The maximum positive difference between the simulated and observed water levels was 7.76 ft at observation well 125N63W4AAAA in grid block row 45, column 22, and the maximum negative difference was 16.08 ft at observation well 125N63W29CCCC in grid block row 54, column 19. The maximum positive difference for aquifer layer 2 (table 6) was 9.48 ft at observation well 125N63W1AAAA in grid block row 45, column 28, and the maximum negative difference was 18.45 ft at observation well 128N62W14DDDD in grid block row 14, column 38. There is only one water-level measurement in aquifer layer 3 in April 1985. The simulated potentiometric head in grid block row 27, column 28 was 35.56 ft higher than the observed water level in observation well 127N63W24AAAA. The average and absolute differences between the simulated and measured water levels (tables 5 and 6) and the comparison of selected hydrographs of simulated and measured hydrographs demonstrate the ability of the 1985 monthly transient simulations to approximately replicate the aquifer's response to monthly changes in pumping, recharge, and evapotranspiration.

The configuration of the April 1985 potentiometric surfaces for aquifer layers 1, 2, and 3 (figs. 26-28) are very similar to the corresponding steady-state potentiometric surfaces (figs. 20-22) except that the April potentiometric heads generally are higher. The April 1985 potentiometric heads in aquifer layer 1 range from greater than 1,390 ft along the western boundary of the study area to less than 1,280 ft near the James River. The direction of ground-water movement in aquifer layer 1 generally is eastward, west of the James River and westward, east of the James River. The April 1985 potentiometric heads and direction of flow in aquifer layers 2 and 3 are similar to those of aquifer layer 1.

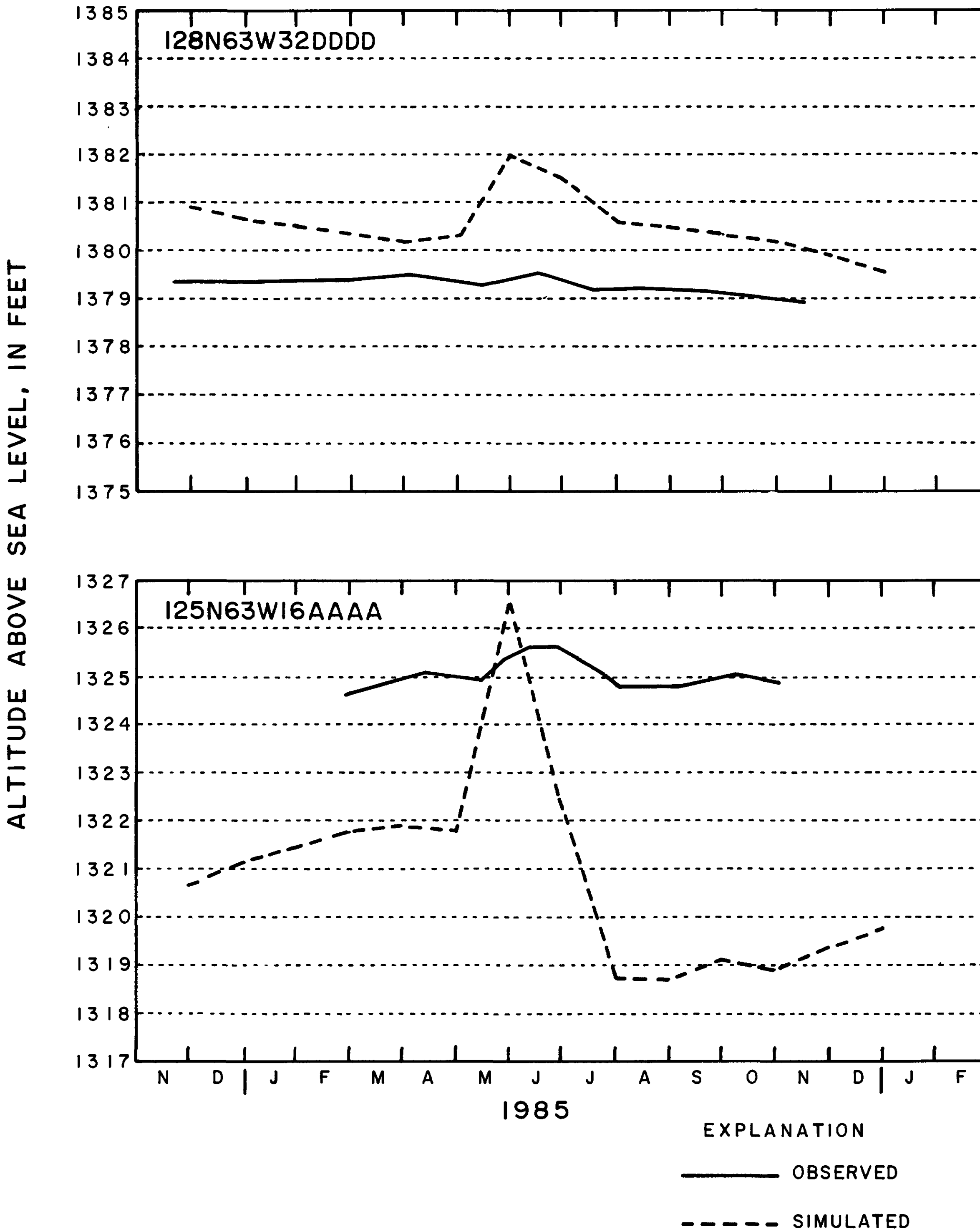


Figure 23.--Hydrographs showing comparison of simulated and observed monthly potentiometric heads in the Elm aquifer part of aquifer layer 1, 1985.

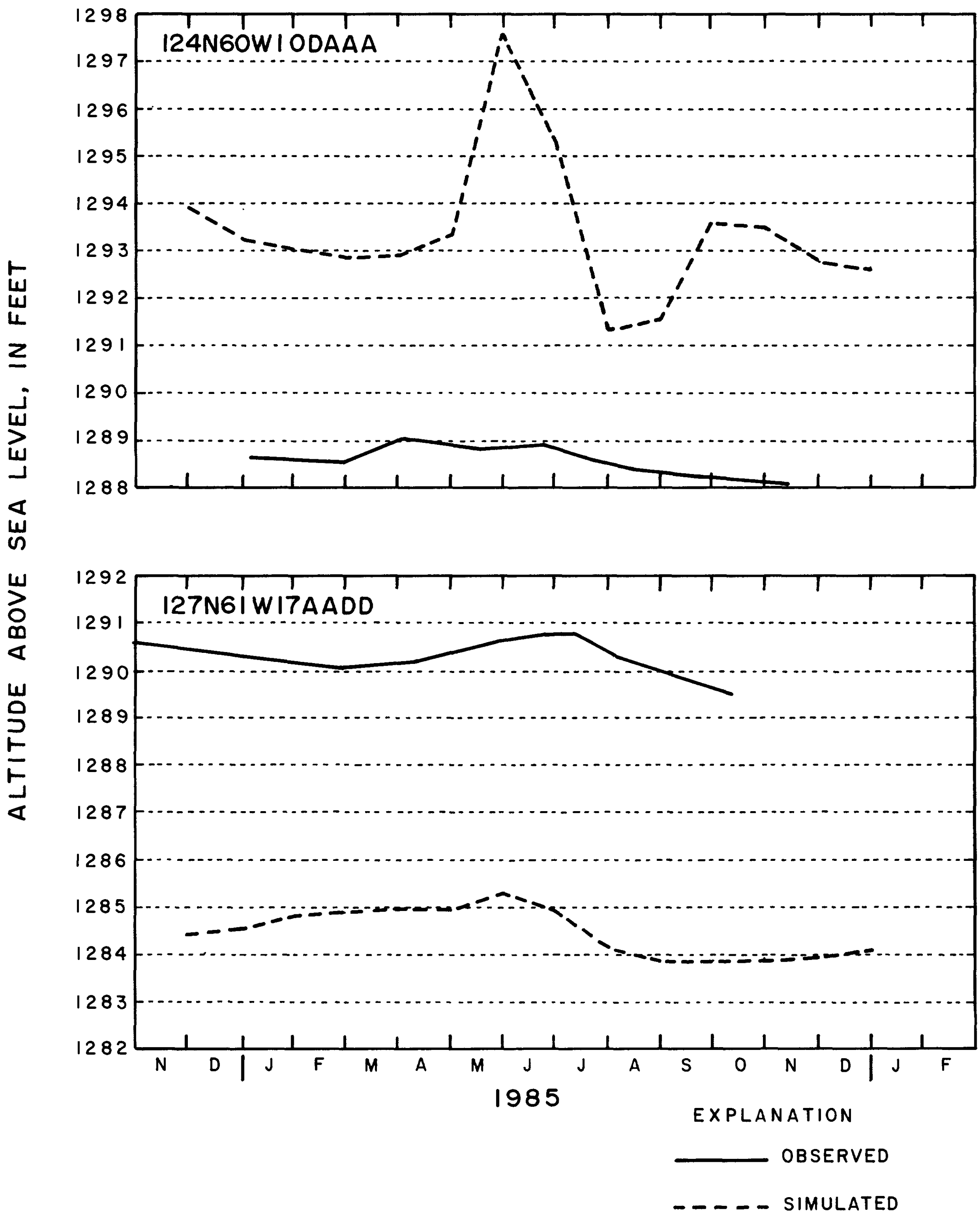


Figure 24.--Hydrographs showing comparison of simulated and observed monthly potentiometric heads in the sandy lake deposits part of aquifer layer 1, 1985.

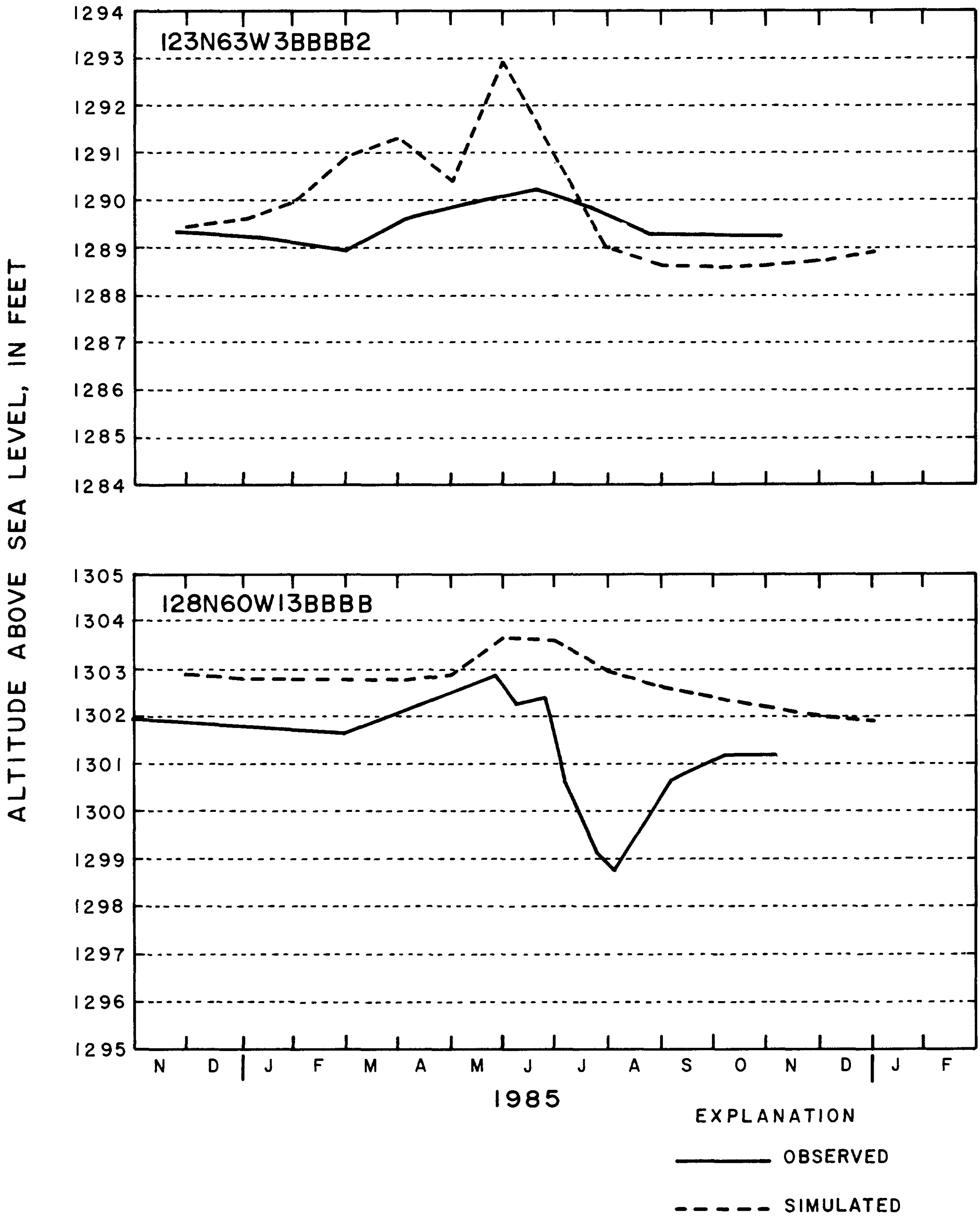


Figure 25.--Hydrographs showing comparison of simulated and observed monthly potentiometric heads in aquifer layer 2, 1985.

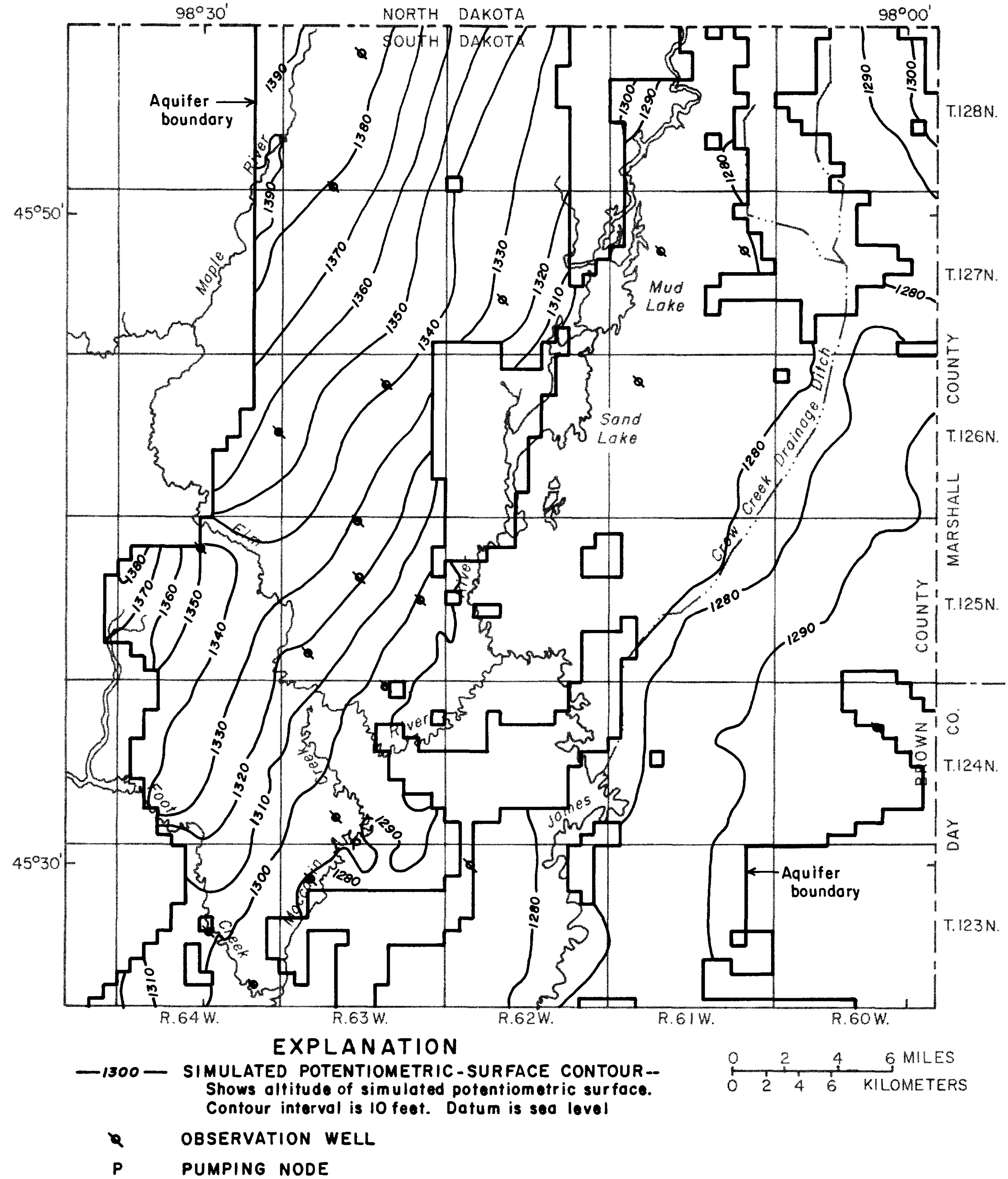


Figure 26.--Simulated potentiometric surface of aquifer layer 1, April 1985.

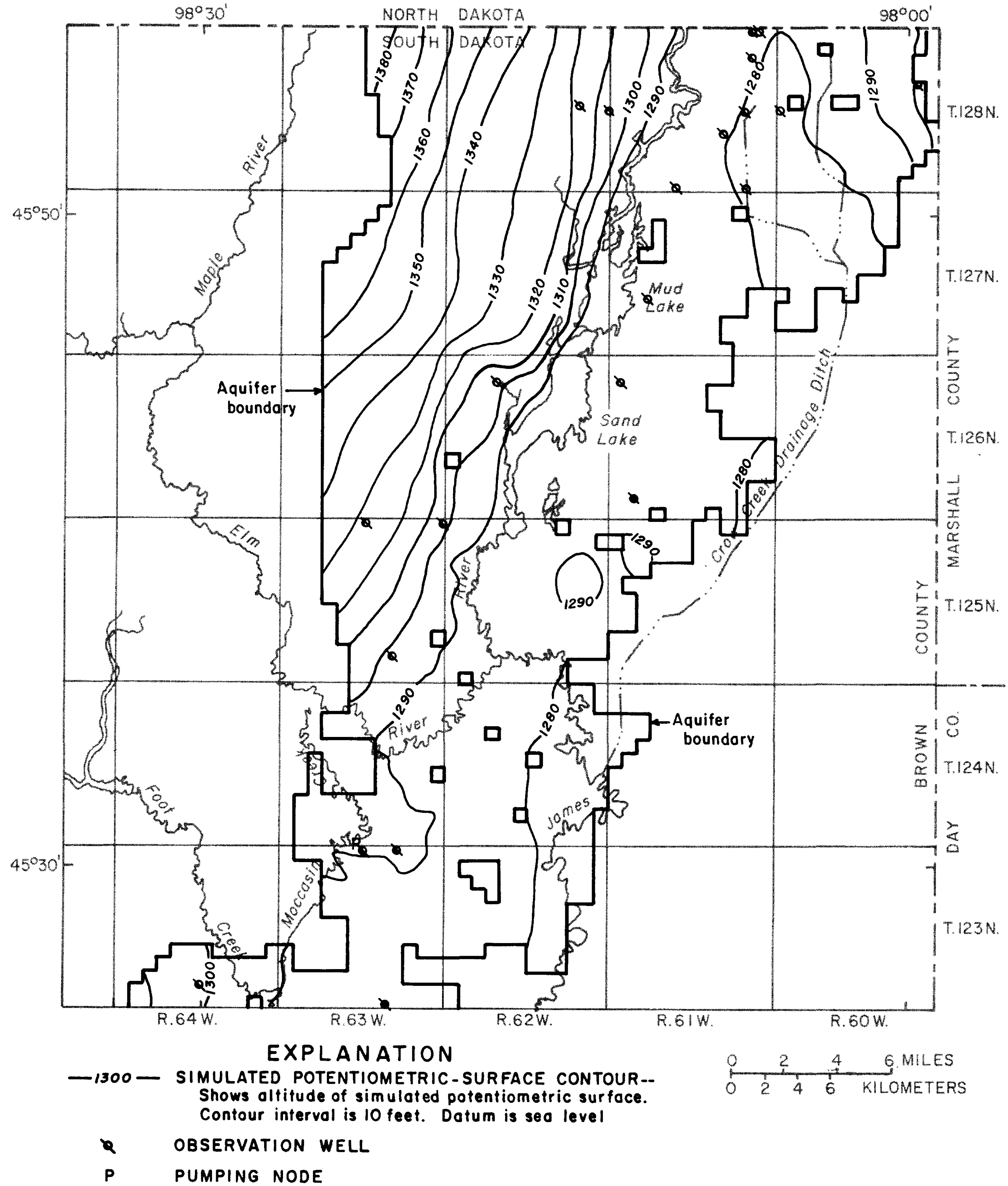
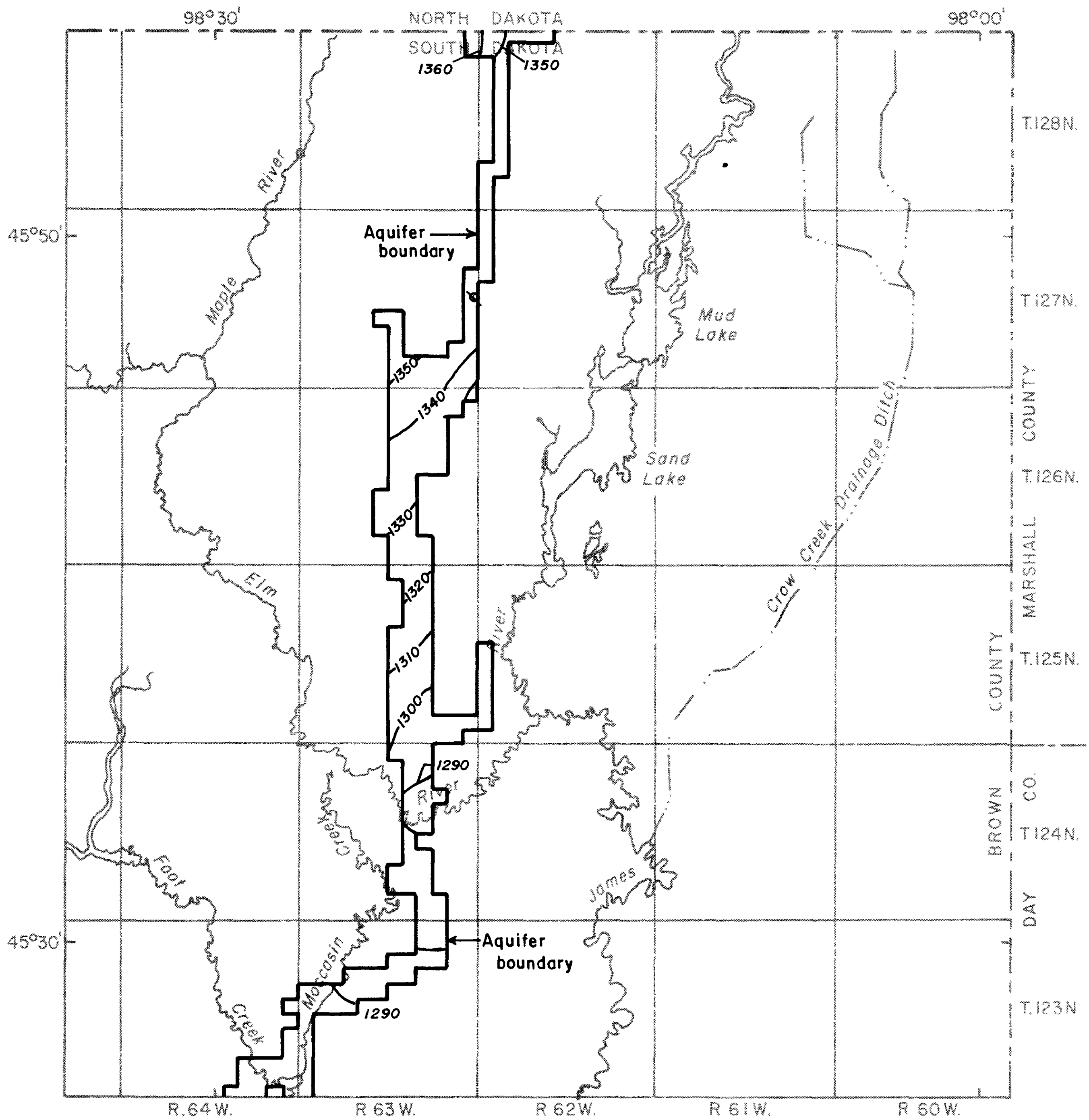


Figure 27.--Simulated potentiometric surface of aquifer layer 2, April 1985.



EXPLANATION

—1300— SIMULATED POTENTIOMETRIC-SURFACE CONTOUR--Shows altitude of simulated potentiometric surface. Contour interval is 10 feet. Datum is sea level

OBSERVATION WELL

Figure 28.--Simulated potentiometric surface of aquifer layer 3, April 1985.

The configuration of the August 1985 potentiometric surfaces for aquifer layers 1, 2, and 3 (figs. 29-31) are lower than those in April due to increased evapotranspiration and pumpage and a decrease in recharge. The August 1985 potentiometric heads in aquifer layer 1 ranged from greater than 1,390 ft above sea level along the western edge of the model area to less than 1,270 ft near the James River (fig. 29). The direction of ground-water movement in aquifer layer 1 generally is eastward, west of the James River and westward, east of the James River. The August 1985 potentiometric heads and direction of flow in aquifer layers 2 and 3 (figs. 30-31) are similar to the heads in aquifer layer 1.

The August potentiometric heads for aquifer layers 1 and 2 generally are lower than the April potentiometric heads. Between April and August 1985, the model-simulated aquifer layer 1 potentiometric heads declined from 0.0 to 15.8 ft with an average decline of 2.4 ft. The potentiometric heads declined from 0.0 to 15.8 ft with an average of 3.3 ft in aquifer layer 2. The simulated potentiometric heads in aquifer layer 3 declined from 0.0 to 13.5 ft with an average decline of 4.8 ft between April and August 1985. The largest head declines generally occurred in areas where pumping from the aquifers is greatest. There are few data available on the potentiometric heads in aquifer layer 3, therefore the accuracy or significances of these head changes between April and August are not known.

The nodal data used to contour the potentiometric head data for aquifer layers 1 and 2 (figs. 26, 27, 29, and 30) for April and August were compared to determine the vertical head differences between the aquifers. The vertical head differences between aquifer layers 1 and 2 in both April and August do not exceed 1.0 ft.

A simulated water budget equating monthly sources and discharges for the model for 1985 is shown in table 8. The water budgets for the 12 monthly simulation periods vary considerably as a result of changes in the monthly evapotranspiration, storage, and pumpage. The maximum error in the monthly mass balances (differences between sources and consumption of water) is about 4 percent.

The primary source of water in 1985 was recharge from precipitation and snowmelt which occurred in the spring and early summer. During this period, leakage across the model boundaries represented by specified head nodes and leakage from the Elm River supplied less than 0.1 percent of the water. During the remainder of the year, the primary source of water was from storage. The major losses of water were pumpage and evapotranspiration during the months of May through August. The amount of water discharged from the aquifer during the summer is recharged to the aquifer in the fall, winter, and spring months.

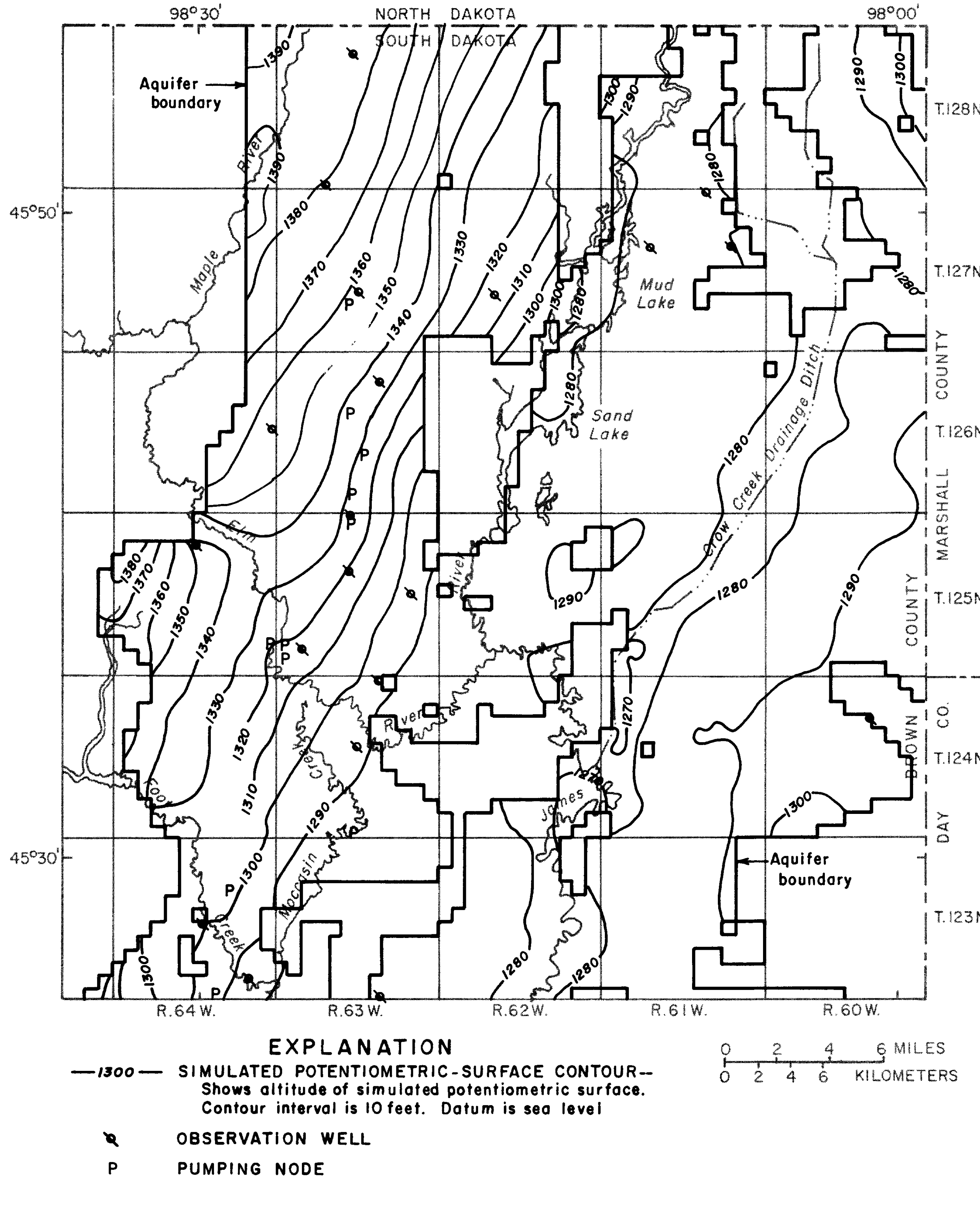


Figure 29.--Simulated potentiometric surface of aquifer layer 1, August 1985.

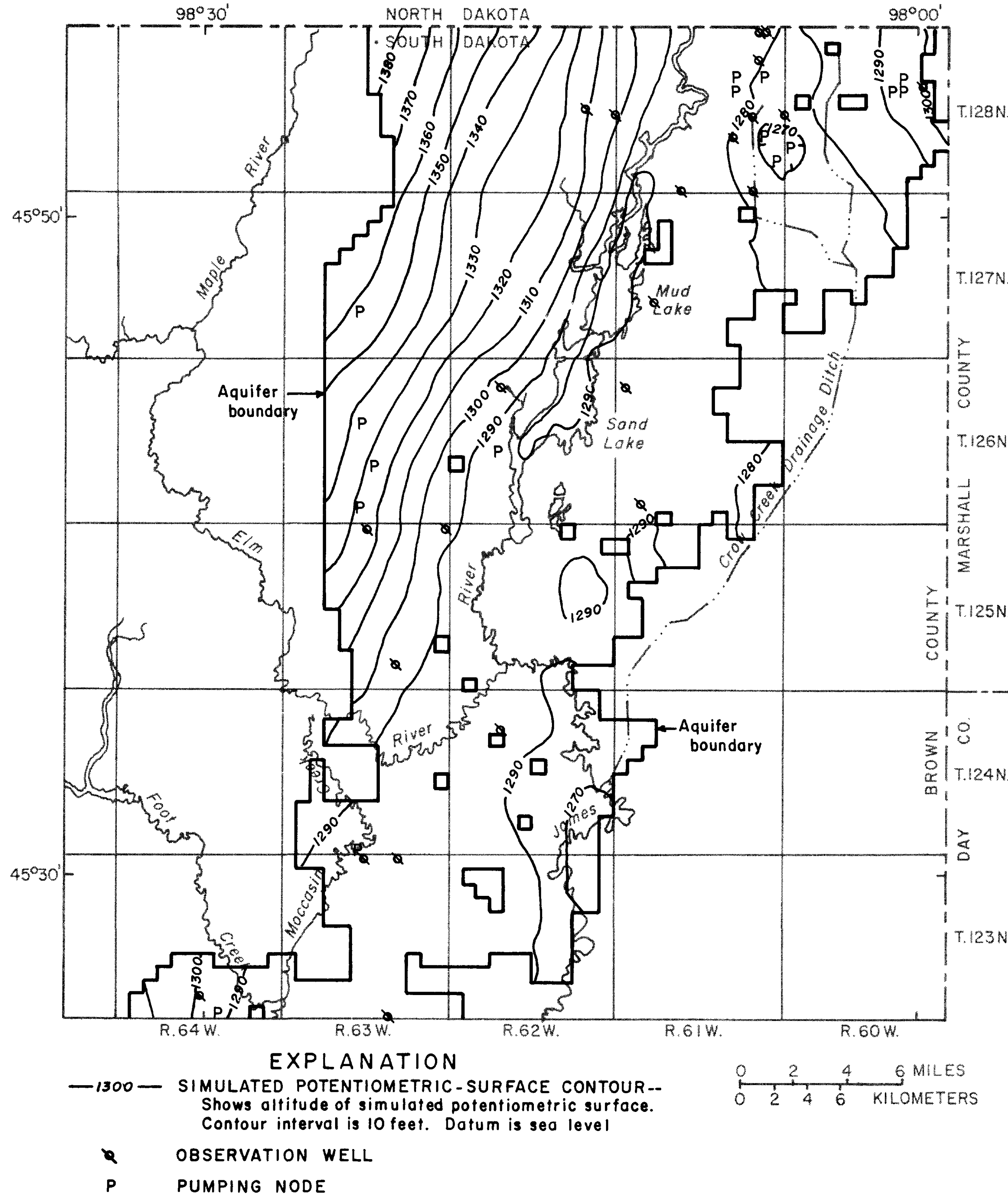


Figure 30.--Simulated potentiometric surface of aquifer layer 2, August 1985.

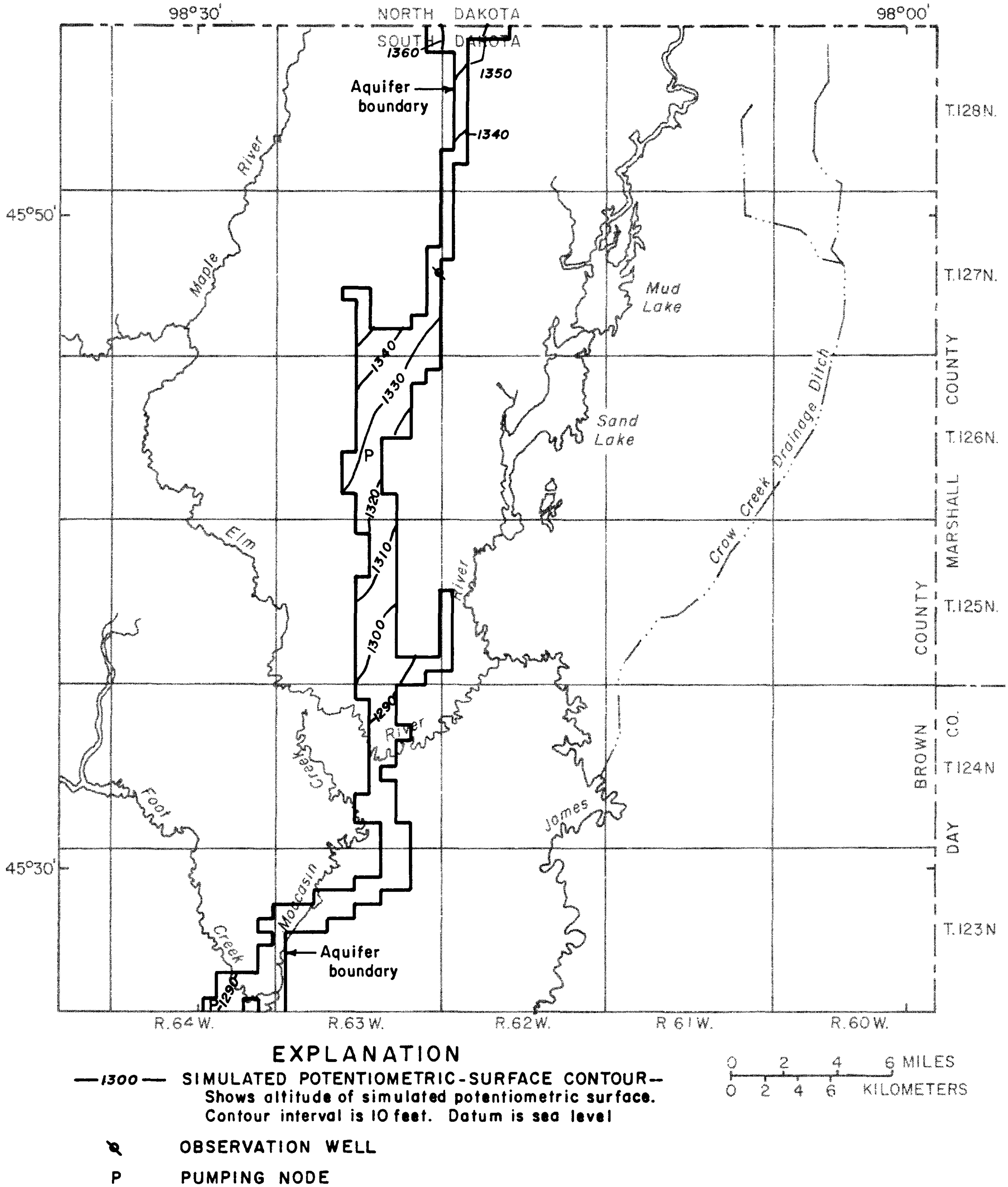


Figure 31.--Simulated potentiometric surface of aquifer layer 3, August 1985.

Table 8.--Monthly simulated water budgets, 1985

	January	February	March	April	May	June	July	August	September	October	November	December	Total
					Sources of water (accretions) (cubic feet per second, rounded)								
Recharge from precipitation	0.00	0.00	437	299	1,690	708	11.8	17.7	153	82.7	0.00	0.00	3,400
Recharge from the stream to the glacial-aquifer system	.10	.09	.00	.39	.11	.27	.68	.67	.43	.30	.15	.13	3.32
Recharge from specified-head boundaries	17.2	17.0	15.5	17.6	16.4	17.0	18.3	19.5	19.4	19.1	19.0	18.9	215
Discharge from storage	73.2	65.8	1.44	248	.92	92.3	463	221	90.6	81.4	79.1	78.3	1,500
Total	90.5	82.9	454	565	1,710	818	494	259	263	183	98.2	97.3	5,120
					Consumption of water (depletions) (cubic feet per second, rounded)								
Evapotranspiration from aquifer	0.00	0.00	0.00	557	1,240	767	471	238	204	119	0.00	0.00	3,600
Pumpage	.00	.00	.00	.05	1.11	3.97	14.8	16.3	6.12	.10	.00	.00	42.4
Discharge from the glacial-aquifer system to the stream	2.33	2.38	4.49	1.55	2.47	1.87	1.40	1.30	1.43	1.71	1.92	2.14	25.0
Discharge from specified-head boundaries	10.8	10.5	13.4	10.5	15.8	14.0	11.0	8.54	8.95	10.1	10.4	10.5	134
Recharge to storage	74.3	66.9	435	1.74	446	34.7	.00	.53	48.9	51.2	89.0	84.1	1,330
Total	87.4	79.8	453	571	1,710	822	498	265	269	182	101	96.7	5,090

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.

The 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).

The sensitivity of the steady-state simulation to changes in recharge, evapotranspiration, and hydraulic conductivity for aquifer layer 1 is shown in table 9 and aquifer layer 2 is shown in table 10. There are no potentiometric-head data available for aquifer layer 3 on which to test the model's sensitivity. It is assumed, however, that the sensitivity of aquifer layer 3 would be similar to that of aquifer layers 1 and 2. The areal distribution of the percentage of maximum recharge and 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 difference decreasing 0.12 ft for aquifer layer 1 and 0.59 ft for aquifer layer 2. The average absolute differences decreased 0.37 ft for aquifer layer 1 and 0.11 ft for aquifer layer 2. Also the maximum positive differences changed 2.01 ft for aquifer layer 1 and 0.21 ft for aquifer layer 2. The maximum negative differences changed 0.64 ft for aquifer layer 1 and 1.07 ft for aquifer layer 2. Increasing the recharge rate 25 percent from 7.00 to 8.75 in/yr produced larger changes in the average difference and in the maximum positive and negative differences for both aquifers layers 1 and 2. The average absolute differences were slightly larger.

The effects of decreasing the potential steady-state evapotranspiration rate from 35.4 to 26.4 in/yr produced a 1.61-ft increase in the average difference for aquifer layer 1 and a 0.93-ft increase for aquifer layer 2. The decreased potential evapotranspiration also resulted in an increase of 0.18 ft for aquifer layer 1 and a 0.48-ft increase for aquifer layer 2 in the average absolute difference from the standard steady-state simulation. An increase in the evapotranspiration rate to 44.25 in/yr resulted in the average difference decreasing 0.62 ft for aquifer layer 1 and 0.57 ft for aquifer layer 2. The average absolute difference decreased 0.17 ft for aquifer layer 1 and 0.28 ft for aquifer layer 2.

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 and absolute differences for aquifer layers 1 and 2 than 25-percent changes 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.

Table 9.--Model sensitivity to changes in recharge, evapotranspiration, and hydraulic conductivity for aquifer layer 1

Model simulation	Average 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
Standard steady-state model	0.78	4.59	8.41	11.78	22
Steady-state model with maximum recharge reduced 25 percent	.66	4.22	10.42	12.42	22
Steady-state model with maximum recharge increased 25 percent	2.35	4.67	11.39	9.73	22
Steady-state model with maximum evapotranspiration reduced 25 percent	2.39	4.77	12.69	10.07	22
Steady-state model with maximum evapotranspiration increased 25 percent	.16	4.42	7.24	12.49	22
Steady-state model with hydraulic conductivity reduced 50 percent	2.21	4.64	10.06	9.81	22
Steady-state model with hydraulic conductivity increased 50 percent	2.00	4.50	14.15	10.43	22

[1]Summation of simulated minus observed potentiometric heads in corresponding grid blocks divided by number of observation wells with observed potentiometric heads. Positive number indicates simulated head was higher than the observed head; negative number indicates simulated head was lower than the observed head.

[2]Summation of the absolute values of simulated minus observed potentiometric heads in corresponding grid blocks divided by number of observation wells with observed potentiometric heads. 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.

Table 10.--Model sensitivity to changes in recharge, evapotranspiration, and hydraulic conductivity for aquifer layer 2

Model simulation	Average difference between simualted 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
Standard steady-state model	3.49	5.10	17.17	5.43	13
Steady-state model with maximum recharge reduced 25 percent	2.90	4.99	16.96	6.50	13
Steady-state model with maximum recharge increased 25 percent	4.07	5.27	17.39	3.85	13
Steady-state model with maximum evapotranspiration reduced 25 percent	4.42	5.58	17.79	4.08	13
Steady-state model with maximum evapotranspiration increased 25 percent	2.92	4.82	16.84	6.12	13
Steady-state model with hydraulic conductivity reduced 50 percent	3.31	4.58	16.44	3.77	13
Steady-state model with hydraulic conductivity increased 50 percent	3.92	5.44	17.83	5.36	13

[1]Summation of simulated minus observed potentiometric heads in corresponding grid blocks divided by number of observation wells with observed potentiometric heads. Positive number indicates simulated head was higher than the observed head; negative number indicates simulated head was lower than the observed head.

[2]Summation of the absolute values of simulated minus observed potentiometric heads in corresponding grid blocks divided by number of observation wells with observed potentiometric heads. 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.

SUMMARY AND CONCLUSIONS

During the Pleistocene Epoch, continental glaciers from the north and east covered eastern South Dakota, depositing a blanket of glacial drift over the preglacial bedrock surface. The drift can be subdivided into two major types, till and outwash, that differ greatly in both physical and hydrologic characteristics. Only the more sandy and gravelly glacial-outwash deposits yield significant quantities of water to wells. The bedrock directly underlying the drift generally yields little or no water to wells. The natural 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 units that comprise the complex hydrologic system in the glacial outwash have been subdivided into three aquifers in the study area: the Elm, Middle James, and Deep James aquifers. These aquifers generally are separated from each other by till or other fine-grained sediments. The Elm aquifer is the uppermost and largest of the aquifers and underlies about 351 mi^2 of the study area. The Elm includes all of the coarser-grained outwash deposits above the altitude of 1,250 ft above sea level. The average thickness of the Elm aquifer ranges from zero to 113 ft. The Middle James aquifer underlies about 500 mi^2 of the study area. The maximum altitude of the top of the aquifer is 1,250 ft and the minimum altitude of the bottom of the aquifer is 1,150 ft. The average thickness ranges from zero to 111 ft. The Deep James aquifer underlies about 52 mi^2. The Deep James includes all of the outwash deposits below the altitude of 1,150 ft above sea level.

Glacial meltwaters deposited an average of about 75 ft of fine sand, silt, and clay on the bed of ancient Lake Dakota. These lake deposits are not an important aquifer but commonly control recharge to and discharge from the Elm and Middle James aquifers.

To improve understanding of the flow in the glacial-aquifer system, a three-dimensional ground-water flow model was developed. 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 system consists of three layers, (2) the aquifers are overlaid by confining beds, (3) the bedrock is an impermeable lower boundary, (4) all lateral boundaries are impermeable except along the northern and eastern boundaries, which are specified head, (5) the James River, Maple River, Foot Creek, and Moccasin Creek are hydraulically isolated from the aquifer system, the Elm River is hydraulically connected at 32 river nodes, (6) all flow in the aquifers is horizontal and in the confining beds vertical, (7) the principal source of recharge is precipitation, which is controlled by the thickness of the confining beds overlying the uppermost aquifer, and (8) the primary method of discharge is evapotranspiration, which is controlled by the thickness of the confining bed overlying the uppermost active layer.

A grid that contains 86 rows and 70 columns of equally spaced blocks, each 0.5-mi wide and 0.5-mi long, was used to simulate the glacial-aquifer system. The aquifer system was simulated under steady-state conditions and under 12 monthly pumping periods for 1985.

The steady-state simulation represents the glacial-aquifer system under equilibrium conditions; that is, water levels recovered to near-prepumping levels during the nonirrigation season. The maximum available recharge to the aquifer was 7.0 in/yr and occurred only where the confining bed overlying the uppermost aquifer was not present. With an average confining bed thickness greater than 0.0 ft and less than 50 ft, the recharge rate declined linearly to 0.0 in/yr. The maximum potential evapotranspiration rate was

35.4 in/yr and can occur only where no confining bed is present above the uppermost aquifer. When the average confining bed thickness is greater than zero and less than 50 ft, the maximum potential evapotranspiration decreases linearly from 35.4 to 0.0 in/yr. The steady-state simulated water budget indicates that recharge from precipitation accounts for 94.8 percent of the water that enters the aquifer or 4.3-in/yr average for each active grid block. Evapotranspiration accounts for 95.8 percent of the water that leaves the aquifer or 4.6-in/yr average for each grid block.

Thirty-six consecutive monthly pumping periods from 1983 through 1985 were simulated. Recharge, evapotranspiration, and pumpage were adjusted monthly. Only the 12 monthly pumping periods for 1985 are presented in this report. In 1985, the maximum monthly recharge varied from 0.00 inch in January, February, November, and December to 1.85 inches in May. The maximum evapotranspiration varied from 0.00 inch in January, February, November, and December to 1.35 inches in May. The simulated monthly water budgets varied considerably as a result of changes in recharge, evapotranspiration, storage, and pumpage.

Because 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 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. Because 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.

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