REGIONAL EVALUATION OF HYDROLOGIC FACTORS AND EFFECTS OF PUMPING, ST. PETER-JORDAN AQUIFER, IOWA

By Michael R. Burkart and Robert C. Buchmiller

U.S. GEOLOGICAL SURVEY

Water-Resources Investigations Report 90-4009

Iowa City, Iowa
1990

U.S. DEPARTMENT OF THE INTERIOR

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CONVERSION FACTORS

For readers who prefer to use International System units rather than the inch-pound units used in this report, conversion factors are listed below:

Multiply inch-pound unit	By	To obtain SI unit
foot (ft)	0.3048	meter(m)
mile (mi)	0.6214	kilometer (km)
gallon (gal)	0.0038	cubic meter (m^3)
foot per second (ft/s)	0.3048	meter per second (m/s)
foot per day (ft/d)	0.3048	meter per day (m/d)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
gallon per minute (gal/min)	0.06308	cubic meter per minute (m^3/min)
gallon per day (gal/d)	0.003785	cubic meter per day (m^3/d)
gallon per minute per foot [(gal/min)/ft}	0.0270	square meter per minute (m^2/min)
foot per day (ft/d)	0.3048	meter per day (m/d)
square foot per day (ft^2/d)	0.09290	square meter per day (m^2/d)

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 St. Peter-Jordan aquifer includes the Cambrian Jordan Sandstone and the overlying Ordovician Prairie du Chien Group and St. Peter Sandstone. The aquifer is present throughout Iowa and is confined beneath other aquifers in much of the State. Information on the aquifer available from drillers and contractors, provided estimates of aquifer transmissivity values ranging from about 500 to about 3,000 square feet per day. The largest transmissivity values are for dolomite and dolomite-cemented sandstone, indicating that permeability in much of the aquifer is due to secondary fractures. The aquifer is vertically bounded by an upper leaky confining unit with a vertical hydraulic conductivity of 10^{-10} feet per second.

The aquifer was simulated using a two-layer finite-difference ground-water flow model. The upper layer simulated a source bed in aquifers composed of Silurian and Devonian rocks overlying the St. Peter-Jordan aquifer. The lower layer simulated flow in the St. Peter-Jordan aquifer. Lateral boundaries assigned in the model include constant heads in northeastern Iowa, where the aquifer is in contact with the Mississippi River or is unconfined, and no-flow boundaries in western and northwestern Iowa, where the rocks are insufficiently permeable to form an aquifer. The aquifer boundaries to the north, east, and south of Iowa were determined by geohydrologic conditions and the relation of the St. Peter-Jordan aquifer with the lateral extent of adjacent aquifers.

An assumption that the largest part of recharge to the aquifer is from outcrop areas in northeastern Iowa and from Minnesota is not supported by the results of this study. Vertical leakage from overlying rocks accounted for most of the recharge to the aquifer in northwestern Iowa. Discharge is mostly through lateral boundaries and to rivers.

Pumping has caused changes in the flow system that include regional declines in the potentiometric surface of the aquifer. Simulation indicates that pumping through 1980 increased net vertical leakage into the aquifer to about double the predevelopment rate. Discharge across lateral boundaries has been substantially reduced or reversed by pumping. Aquifer storage provided about one-third of the water required to supply pumping in the 1970's. Simulation of future conditions, assuming no increase in pumping rates, indicates that the rate of decline in water levels will decrease by the year 2020. As equilibrium with pumping is approached in 2020, 75 percent of the pumpage will be balanced by vertical leakage, eight percent by water released from aquifer storage, and 17 percent by increases in boundary recharge or decreases in boundary discharge. Future pumping at an increasing rate of about 10 percent per decade of the average pumping rate in 1975 will require about one and one-half times the vertical leakage of the 1971-1980 period and about five-times the net inflow from lateral boundaries; however, the rate of water released from aquifer storage will be about half the 1970's rate. Under these conditions, the head in the aquifer will continue to decline at an almost constant rate until 2020.

INTRODUCTION

This report was prepared as part of the U.S. Geological Survey's Northern Midwest Regional Aquifer System Analysis (RASA) project (Steinhilber and Young, 1979), which focused on the regional aquifer system formed by Cambrian and Ordovician rocks in six Midwestern States (fig. 1). Aquifers composed of these rocks are commonly referred to as the Cambrian-Ordovician aquifer system and supply a major part of ground water needs of the states in the northern Midwest, including Iowa. The aquifer system contains three aquifers in Iowa-Dresbach, St. Peter-Jordan, and Galena aquifers, which are separated by rocks that act as confining units.

Base from U. S. Geological Survey
U. S. base map, 1:2,500,000

Geology modified from P. B. King and H. M. Beikman (1974). Structure from P. B. King (1969)

EXPLANATION

- Mississippian and younger formations
- Silurian and Devonian formations *Mostly dolomite*
- Ordovician Maquoketa Shale
- Ordovician and Cambrian formations *Mostly sandstone and dolomite*
- Precambrian sandstone
- Precambrian crystalline rock
- Northern Midwest Regional Aquifer-System Analysis Boundary
- — 500 — STRUCTURE CONTOUR *Shows altitude of top of Precambrian crystaline basement. Contour interval 500 meters. Datum is sea level*
- A —— A' Line of section - - section shown in figure 2

Figure 1. - - Location and general bedrock geology of the Northern Midwest Regional Aquifer-System Analysis area

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In Iowa, Illinois, Minnesota, and Wisconsin, major metropolitan areas depend on the Cambrian-Ordovician aquifer system for at least part of their water supplies. In addition, many major industries, small cities, villages and communities depend on these aquifers as a source of water.

Water levels in the regional aquifer system have declined more than 150 ft (feet) in parts of Iowa (Horick and Steinhilber, 1978) and more than 900 ft in the Chicago-Milwaukee area (Young, in press) since deep well development began in the late 19th century. If the demand for water continues at the present rate or increases, water levels in the aquifer system will continue to decline at present (1987) or greater rates. Additionally, in some areas of Iowa or locally in other states of the region, the Cambrian-Ordovician aquifer system and overlying aquifers contain water with undesirable concentrations of dissolved minerals. Pumping from an aquifer near these areas may change the natural ground-water flow direction and cause poorer-quality water to move into the pumped areas.

A regional study of the Cambrian-Ordovician aquifer system was needed for a comprehensive understanding of the aquifer system and its response to further development (Steinhilber and Young, 1979). In addition to this regional-scale evaluation, a study of the aquifers within Iowa was conducted to further improve the understanding of the aquifer system in Iowa.

Purpose and Scope

This study includes that part of the Cambrian-Ordovician aquifer system in Iowa and parts of the adjoining states that contain the St. Peter-Jordan aquifer. This report defines the range of values of hydrologic characteristics that affect ground-water movement in the St. Peter-Jordan aquifer and provides estimates of the effects of continued withdrawal on water levels in the aquifer. This purpose was accomplished by developing a ground-water flow model to quantitatively determine the source, and the movement and discharge of water under predevelopment and stressed conditions. The St. Peter-Jordan aquifer includes the Cambrian Jordan Sandstone and the overlying Ordovician Prairie du Chien Group and St. Peter Sandstone.

To evaluate the vertical leakage from overlying materials, the Ordovician rocks, except St. Peter Sandstone and Prairie du chien Group (table 1), are considered to be the confining unit above this aquifer system. The rocks from Silurian through Quaternary age above the confining unit are considered the source of vertical recharge to the aquifer system. The aquifers beneath the St. Peter-Jordan aquifer, although very important in other parts of the northern Midwest, are used only in very limited parts of eastern Iowa. Including the deeper aquifers in this analysis would have required hydrologic data for these aquifers that are not available in most of Iowa. For the purpose of this study, the degree of hydrologic connection between these deeper aquifers and the St. Peter-Jordan is believed to be minor, therefore the omission of these deep aquifers probably yields insignificant errors in the study.

The data used in this study include well-sample logs, geophysical logs, reported specific-capacity tests, historical water-level measurements, and reported measurements and water-use information collected from water-supply facilities in Iowa. Hydrologic and geologic interpretations of conditions in the St. Peter-Jordan aquifer in adjacent states were provided by U.S. Geological Survey offices participating in the Northern Midwest RASA.

Acknowledgments

Much of the information compiled for this report was obtained from the Iowa Department of Natural Resources and many municipalities throughout Iowa. Without the efforts, patience, and cooperation of the people in these governmental units, this study would not have been possible.

HYDROGEOLOGY

Geologic Framework

The Cambrian and Ordovician rocks in Iowa are predominantly sandstone, limestone, and dolomite with a few thin beds of shale. The sandstone may consist of beds interlayered with dolomite or shale, or, as is the case with the St. Peter Sandstone in some areas, consist only of loosely cemented sandstone. The dolomite has sufficient permeability, probably secondary, such as solution openings, to yield up to several hundred gal/min (gallons per minute) of water to wells. The

Table 1.--*Stratigraphic and hydrogeologic units and equivalent layers used in the digital model*

STRATIGRAPHIC UNITS		HYDROGEOLOGIC UNITS	DIGITAL MODEL LAYERS
SYSTEM	GROUP/FORMATION		
Quaternary	Undifferentiated deposits	Local aquifers	Aquifer layer 1--Source layer
Cretaceous	Dakota Formation	Dakota aquifer	
Pennsylvanian	Pennsylvanian rocks	Confining unit	
Mississippian	Mississippian rocks	Aquifers composed of Mississippian rocks	
Devonian	Devonian and Silurian rocks	Aquifers composed of Devonian and Silurian rocks	
Silurian			
Ordovician	Maquoketa Shale	Confining unit	Confining unit--controls vertical leakage between aquifer layers
	Galena Dolomite	Galena aquifer	
	Decorah Formation	Confining unit	
	Platteville Formation		
	Glenwood Shale		
	St. Peter Sandstone	St. Peter - Jordan aquifer	Aquifer layer 2--active layer
	Prairie du Chien Group		
Cambrian	Jordan Sandstone		
	St. Lawrence Dolomite	Confining unit	Impermeable boundary
	Franconia Sandstone		
	Dresbach Group	Dresbach aquifer	
Precambrian	Sandstone and crystalline rocks	Confining unit	

limestone is generally much less permeable than dolomite and does not produce appreciable amounts of water.

The Cambrian-Ordovician aquifer system in Iowa consists of three major aquifers in rocks of Cambrian and Ordovician age. These are, in ascending order, the Dresbach, St. Peter-Jordan, and Galena aquifers (table 1).

The Dresbach aquifer (Horick and Steinhilber, 1978) underlies the St. Peter-Jordan aquifer and includes in ascending order the Mount Simon, EauClaire, and Galesville Sandstones of the Dresbach Group of Cambrian age. It is overlain and confined by the glauconitic shale and sandstone of the Franconia Sandstone and by the coarsely crystalline St. Lawrence Dolomite.

The St. Peter-Jordan aquifer includes the Cambrian Jordan Sandstone and the Ordovician Prairie du Chien Group and St. Peter Sandstone (table 1). The St. Peter-Jordan aquifer is divisible into two permeable zones in some locations, the Prairie du Chien-Jordan and St. Peter, separated by a dolomite confining unit in the upper part of the Prairie du Chien Group (Horick and Steinhilber, 1978). However, a difference in hydraulic head between the two permeable zones can be measured at only a few places in Iowa and many wells are open to both units; therefore, the two permeable zones are treated as a single aquifer for this study. The St. Peter-Jordan aquifer is overlain and confined by the Ordovician Glenwood Shale, and by the Platteville and Decorah Formations (mostly limestone with interbedded shale).

The uppermost aquifer in the Cambrian and Ordovician aquifer system is the Galena aquifer. Rocks of the Ordovician Galena Dolomite form the Galena aquifer in Iowa only where these rocks immediately underlie glacial drift or are at land surface. Such conditions occur only in northern and northeastern Iowa. Throughout most of Iowa, the Galena Dolomite can be combined with rocks of the overlying Maquoketa Shale and the underlying Decorah Formation, Platteville Formation and Glenwood Shale, all of Ordovician age, to form a confining unit overlying the St. Peter-Jordan aquifer. Therefore, the Galena aquifer was not included in this study as an important aquifer in Iowa.

Cambrian and Ordovician rocks are overlain by consolidated deposits throughout most of Iowa. In most areas the rocks immediately above those of Ordovician age are of Silurian and Devonian age. The Silurian and Devonian rocks form the bedrock surface beneath unconsolidated deposits in much of eastern Iowa. Mississippian and younger rocks are the uppermost bedrock in much of the rest of the State, except where Cretaceous rocks overlie the older rocks, including the Cambrian and Ordovician rocks, in northwestern Iowa (Burkart, 1984). Cambrian and Ordovician rocks form the bedrock surface in only a small part of the State, in extreme northern and northeastern Iowa.

The configuration of the top of the St. Peter-Jordan aquifer in Iowa generally follows the configuration of the Precambrian basement as shown in figure 1. The aquifer is at land surface in extreme northeastern Iowa and immediately underlies the Cretaceous Dakota Formation in northwest Iowa. Depths to the aquifer are commonly greater than 2,000 ft in southern Iowa. A detailed structure contour map of the top of the Jordan Sandstone (Horick and Steinhilber, 1978) illustrates more precisely the structural framework of the St. Peter-Jordan aquifer. The vertical structural relationships among the geologic units are shown in the geologic section in figure 2.

Hydrology

The hydrogeologic factors controlling ground-water movement in the St. Peter-Jordan aquifer were defined and were used to simulate ground-water flow in the aquifer using a digital ground-water flow model. The conditions that affect the rate of ground-water movement are the distribution of hydraulic head, transmissivity, and storage coefficient of the aquifer. Other factors considered in the flow model are the boundary conditions affecting lateral flow into the area and the location and rate of recharge to and discharge from the aquifer including leakage through confining units.

Potentiometric Surfaces

Methods of obtaining data on head in the St. Peter-Jordan aquifer in Iowa, are limited to water-supply wells for communities and industries. Consequently, measurements of the undisturbed static head in the aquifer can only be made during

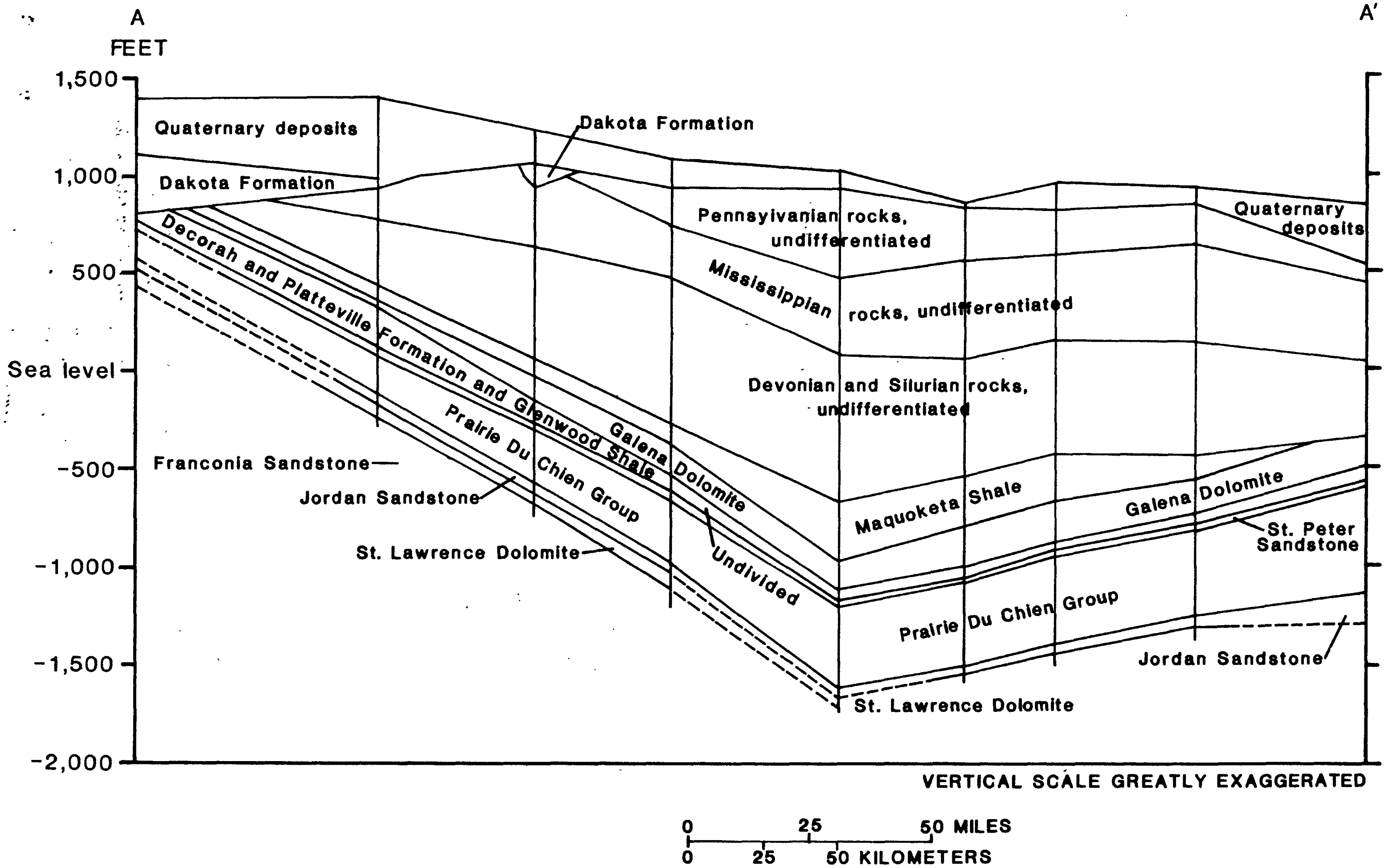


Figure 2.--Generalized geologic section from northwestern to southeastern Iowa.

a short time interval either after drilling or before start of pumping. Once a well has been pumped, only an approximate static water level can be measured, and then only if the well has not been pumped for a period of several hours to several days. Measurements of head may also be affected by pumping in nearby wells and well construction. Historical information on water level is available from drillers, contractors, water plant operators or other municipal officials, and water levels reported and measured by the U.S. Geological Survey.

The variety of water-level measurements reported at each site, including the methods of measurement, the means by which the altitude of the measuring point was determined, and the possible effects of pumping on individual measurements, were examined. It was determined from this analysis that the regional estimates of the potentiometric surface in the St. Peter-Jordan aquifer obtained from production wells could have an error range of ± 25 ft. This error is included in the estimated predevelopment potentiometric surface shown in figure 3. This predevelopment potentiometric surface is assumed to be the potentiometric surface prior to 1900. The surface was derived from water-level measurements made in newly drilled wells before pumping began. Some of these measurements were made late in the 19th and early 20th century, but most are from wells drilled between 1920 and 1950. The newer wells were included to provide an adequate number of data points for approximation of the local predevelopment head. The decision whether to include the initial water level in each well was based on an estimate of the drawdown at the location of the well that has resulted from pumping in nearby wells. If the estimated cumulative drawdown exceeded five ft, the water level from the new well was not included in the predevelopment water-level interpretation. This selection process indicated that after 1950 the density and magnitude of pumping throughout Iowa was sufficient to cause most new well sites to be influenced by pumping. Consequently, no water levels from wells drilled after 1950 were used for the predevelopment potentiometric surface.

Another criterion for selecting water-level measurements was that wells be open only to those rock units included in the St. Peter-Jordan aquifer. Many wells in Iowa, particularly older wells, are completed in part or all rocks of Devonian through Cambrian age. Some of these include industrial and municipal wells which produce large quantities of water from combinations of aquifers. Any tests or water levels from these wells represent the combined effect of all aquifers in the open section and are not suitable for this analysis of the St.Peter-Jordan aquifer.

Comparison of the predevelopment (fig. 3) and 1980 (fig. 4) potentiometric surfaces shows the changes in the head and in the direction of potential ground-water movement that have occurred during the last 80 years. Both maps indicate that the general movement of water is from the north and northwest toward the east and southeast. Head declines due to pumping are generally less than 50 ft in the northwest and as much as 100 to 150 ft in eastern and central Iowa.

The effect of pumping on the potentiometric surface is evident. Large depressions, represented by closed contours and bulges in contours shown in figure 4, are associated with major pumping centers. An up-gradient shift in all contours represents areal decline in the altitude of the potentiometric surface. In northeastern and north-central Iowa, flow direction has shifted from an easterly to a more southerly direction. In southern Iowa, the southeastward flow direction shifted to a greater component of flow eastward.

Recharge and Discharge

The St. Peter-Jordan aquifer receives recharge in areas where the aquifer directly underlies another aquifer that has a higher head, or where the St. Peter-Jordan aquifer crops out. Direct discharge from the St. Peter-Jordan aquifer occurs in areas where the head in a directly overlying aquifer is lower or where the St. Peter-Jordan aquifer is in direct contact with a stream or stream alluvium. Indirect recharge and discharge (leakage) occur where the aquifer is separated from an adjacent aquifer by a confining unit and there is a head difference between the two aquifers.

In northeastern Iowa, direct recharge and discharge occurs throughout the aquifer outcrop (fig. 5). Stream valleys deeply incised into the Cambrian and Ordovician rocks in the outcrop area are discharge areas for the aquifer. The upland areas lacking substantial glacial drift are areas of direct recharge to the aquifer. These areas of direct recharge and discharge form localized flow systems throughout northeastern Iowa.

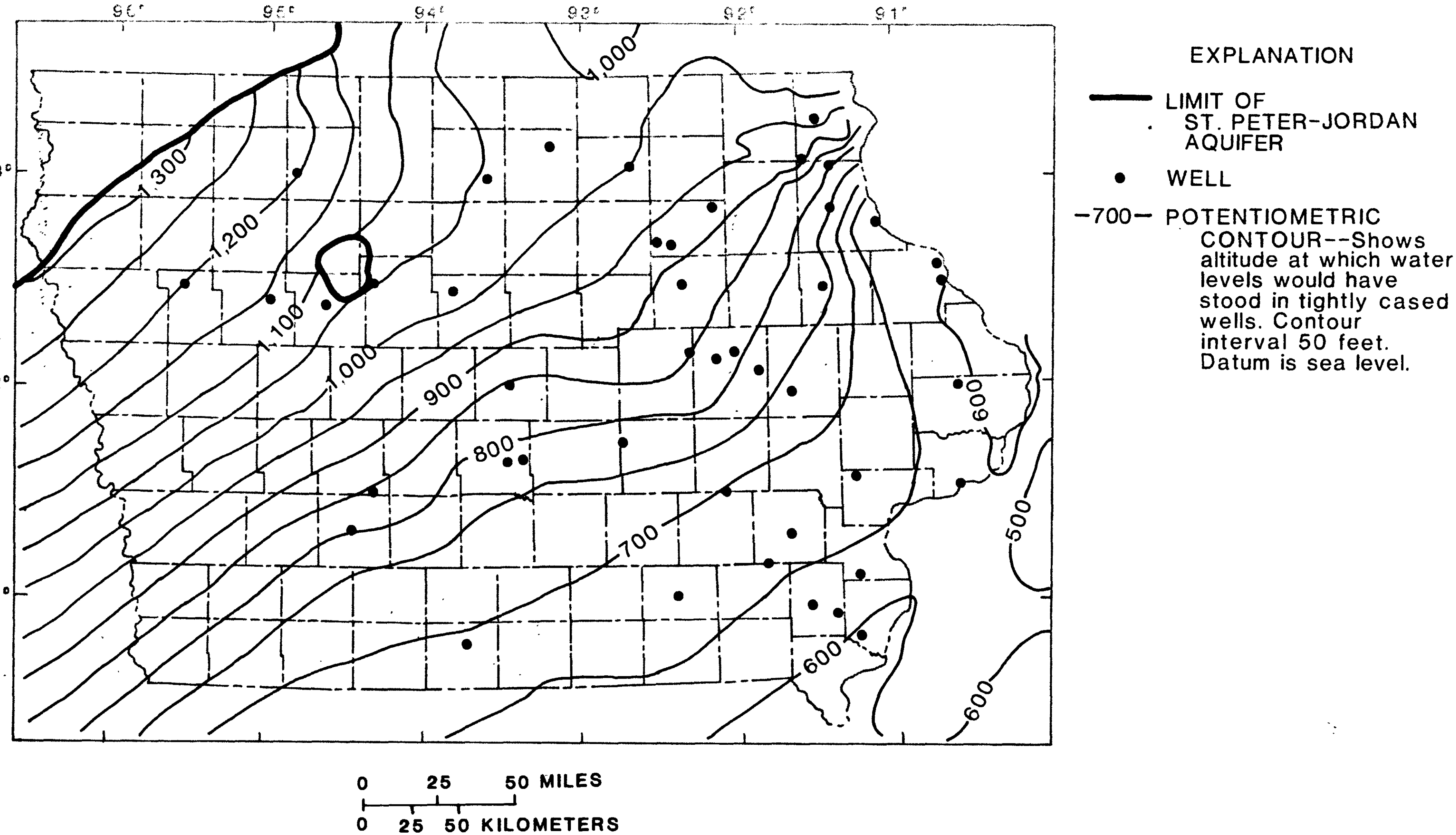
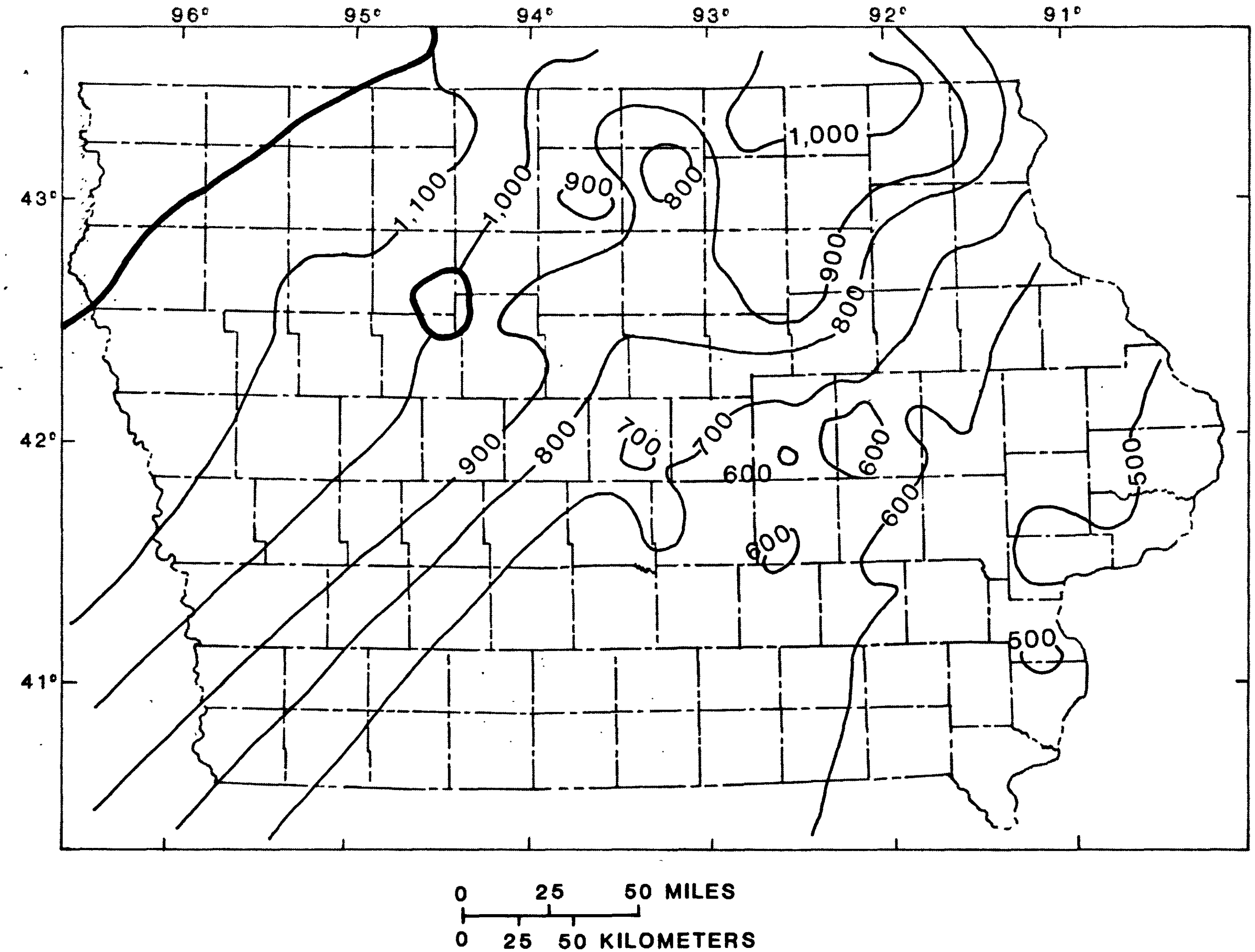


Figure 3.--Estimated predevelopment potentiometric surface for the St. Peter-Jordan aquifer.



EXPLANATION

LIMIT OF ST. PETER-JORDAN AQUIFER

—700— POTENTIOMETRIC CONTOUR--Shows altitude at which water levels would have stood in tightly cased wells. Contour interval 100 feet. Datum is sea level.

Figure 4.--Potentiometric surface for the St. Peter-Jordan aquifer, 1980.

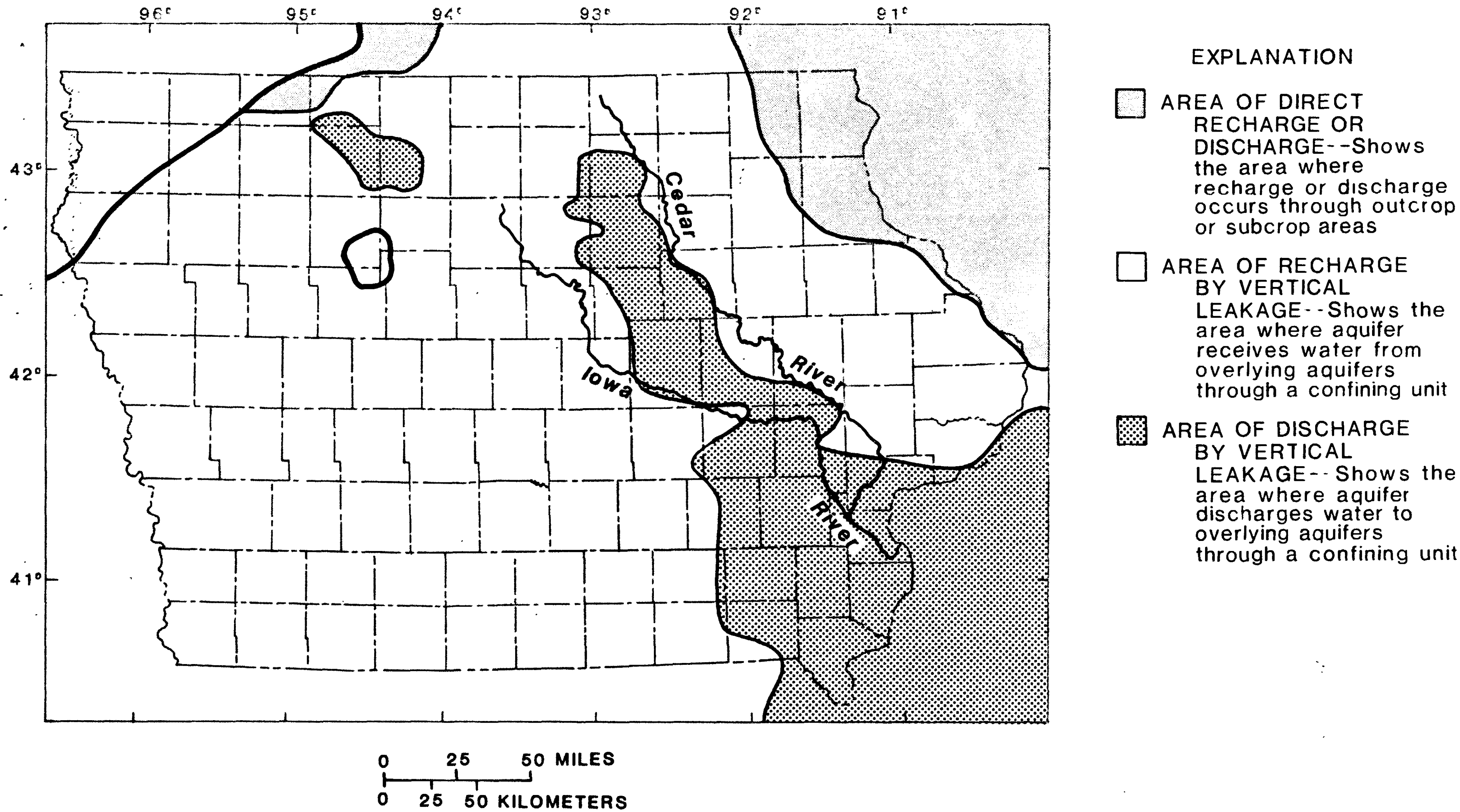


Figure 5.--Distribution of predevelopment recharge and discharge in the St. Peter-Jordan aquifer.

The St. Peter-Jordan aquifer is directly recharged also in northwestern Iowa where it directly underlies the Dakota aquifer (Burkart, 1984). Field measurements indicate that under predevelopment conditions, the head in the Dakota aquifer is higher than the underlying aquifers, including the St.Peter-Jordan. The heads in the St. Peter-Jordan aquifer in this area are the highest in Iowa (figs. 3 and 4). However, the hydraulic properties of the Dakota and St. Peter-Jordan aquifers in this area are poorly known.

Throughout the rest of Iowa, recharge to and discharge from the St. Peter-Jordan aquifer are assumed to be vertical leakage through confining units separating the overlying aquifers, comprised of Silurian and Devonian rocks, from the St. Peter-Jordan aquifer, or from flow across lateral boundaries. Leakage may also occur between the St. Peter-Jordan aquifer and underlying aquifers, however no head data from the underlying aquifers are available except in the extreme eastern part of Iowa; therefore, interpretation of the rate or direction of such leakage cannot be made.

The areas of predevelopment recharge and discharge shown in figure 5 were determined by comparing the potentiometric surface of the St. Peter-Jordan aquifer (fig. 3) with that of the overlying aquifers comprised of Silurian and Devonian rocks (fig. 6). Where the head in Silurian and Devonian rocks is higher, the St. Peter-Jordan aquifer is recharged and where the head differential is reversed, ground water in the St. Peter-Jordan aquifer discharges upward to the overlying aquifers.

The largest area of vertical leakage from the St. Peter-Jordan aquifer is a corridor in eastern and southeastern Iowa. This area is between the Iowa and Cedar Rivers (fig. 5) and is east of and parallel to the western edge of the outcrop of Silurian and Devonian rocks (fig. 1). The northern part of the discharge area coincides with a potentiometric ridge in the predevelopment St. Peter-Jordan head (fig. 3). Remnants of the potentiometric ridge remain in 1980, even though the 1980 potentiometric surface (fig. 4) is strongly affected by pumping. The Cedar and Iowa River valleys are incised into the Silurian and Devonian rocks and are recharge and discharge areas for aquifers comprised of these rocks. This local discharge area also seems to affect the flow in the St. Peter-Jordan aquifer.

The 1980 head differences between the aquifers separated by the confining unit can be calculated by comparing figures 4 and 6. The thickness of the confining unit (fig. 7) is the composite thickness of the Maquoketa Shale, Galena Dolomite, Decorah Formation, Platteville Formation, and Glenwood Shale.

Hydraulic Properties of the St. Peter-Jordan Aquifer

The data used to estimate transmissivity are from 238 pumping tests in 148 wells that penetrate the St. Peter-Jordan aquifer. These data are from single-well pumping tests conducted by well drillers or contractors for the purpose of determining well yield and drawdown. Tests selected were for wells that are open to the entire thickness of the Jordan Sandstone and Prairie du Chien Group; many also are open to part of the St. Peter Sandstone. Data used to determine transmissivity include the well diameter, prepumping (static) water level, pumping rate, water level at the end of a pumping period, test duration, and water-level measurements during recovery. For tests that did not include pumping duration, 12 hours was assumed.

Data from tests that represent the hydraulic properties of the aquifer are needed to estimate transmissivity on a regional scale. A common practice in Iowa is to attempt to increase the capacity of St. Peter-Jordan aquifer wells by treating them with hydrochloric acid or shooting them with explosives. These treatments have been successful in increasing the specific capacity by factors of 10 to 20. Thus, tests conducted after treatment do not provide specific capacity data that are representative of the natural conditions in the aquifer.

Transmissivity was estimated from specific capacity data using an equation from Lohman (1972, p. 52). A storage coefficient of 0.00025 (Horick and Steinhilber, 1978), was used for these estimates.

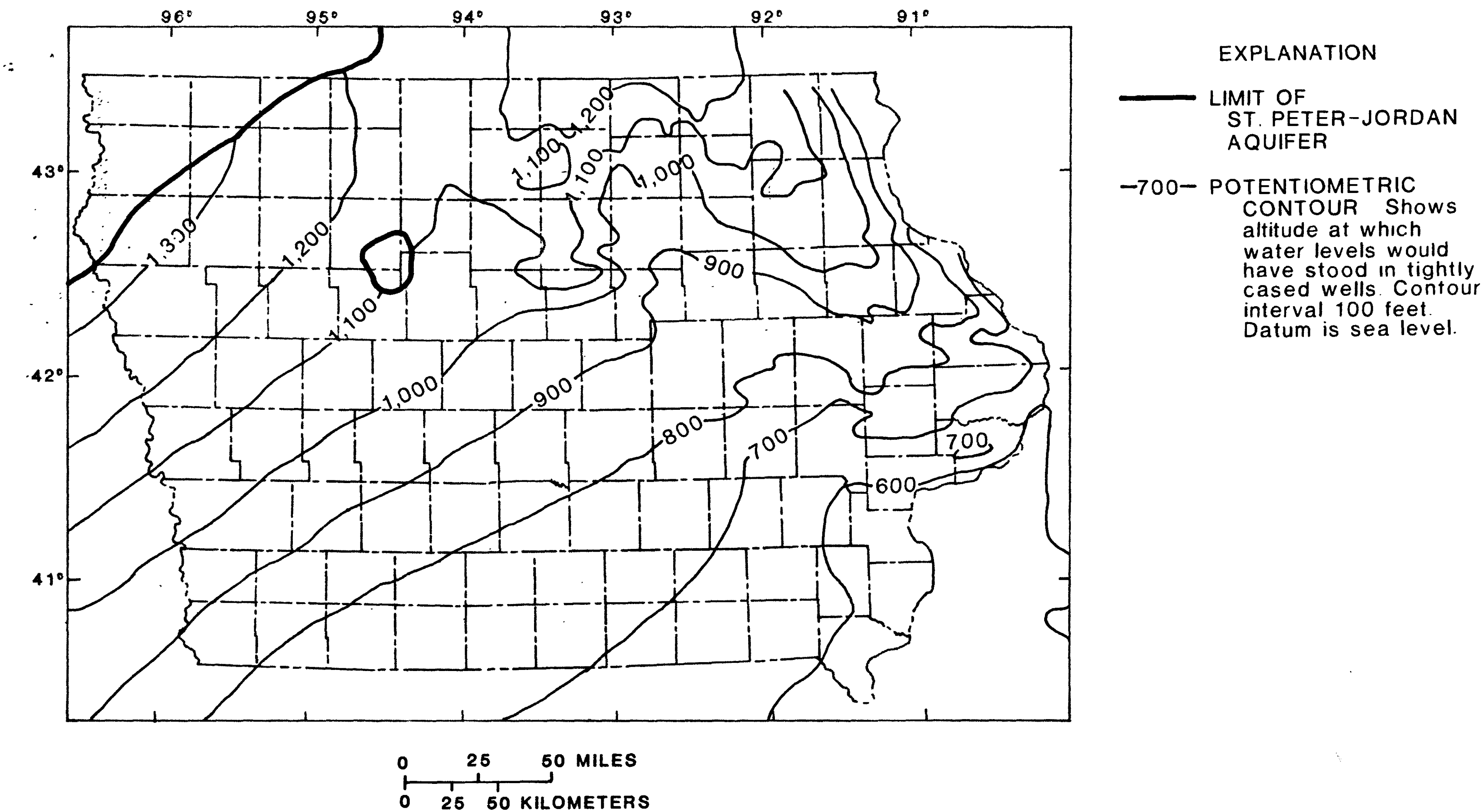


Figure 6.--Potentiometric surface of aquifers comprised of Silurian and Devonian rocks, 1980.

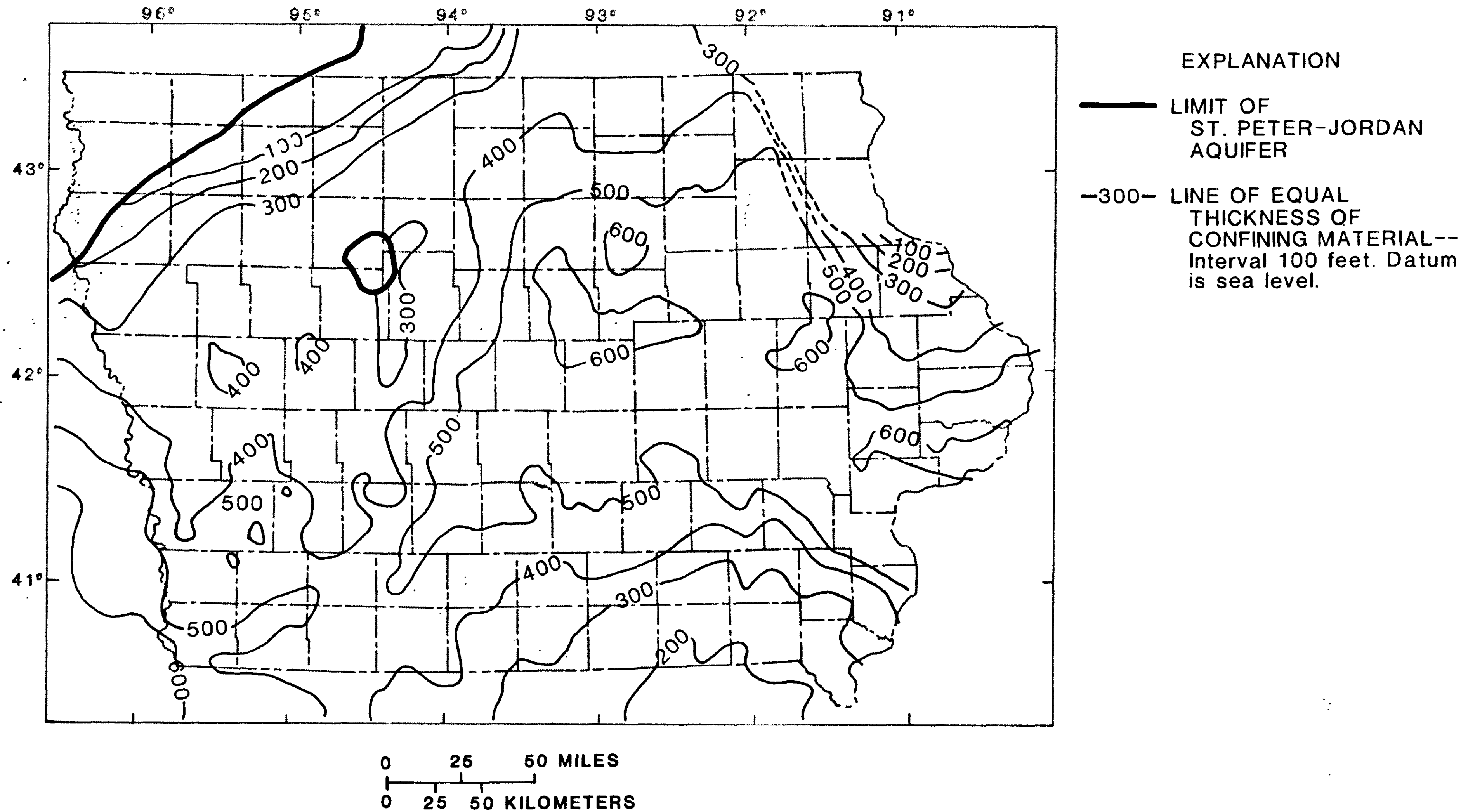


Figure 7.--Thickness of confining unit overlying the St. Peter-Jordan aquifer.

$$T = \frac{Q/s\ 2.30\log_{10} 2.25Tt/r^2S}{4\pi}$$

where: T = transmissivity (L^2T^{-1})
Q/s = specific capacity (L^2T^{-1})
t = duration of pumping (T)
r = radius of well (L)
S = storage coefficient (Dimensionless)

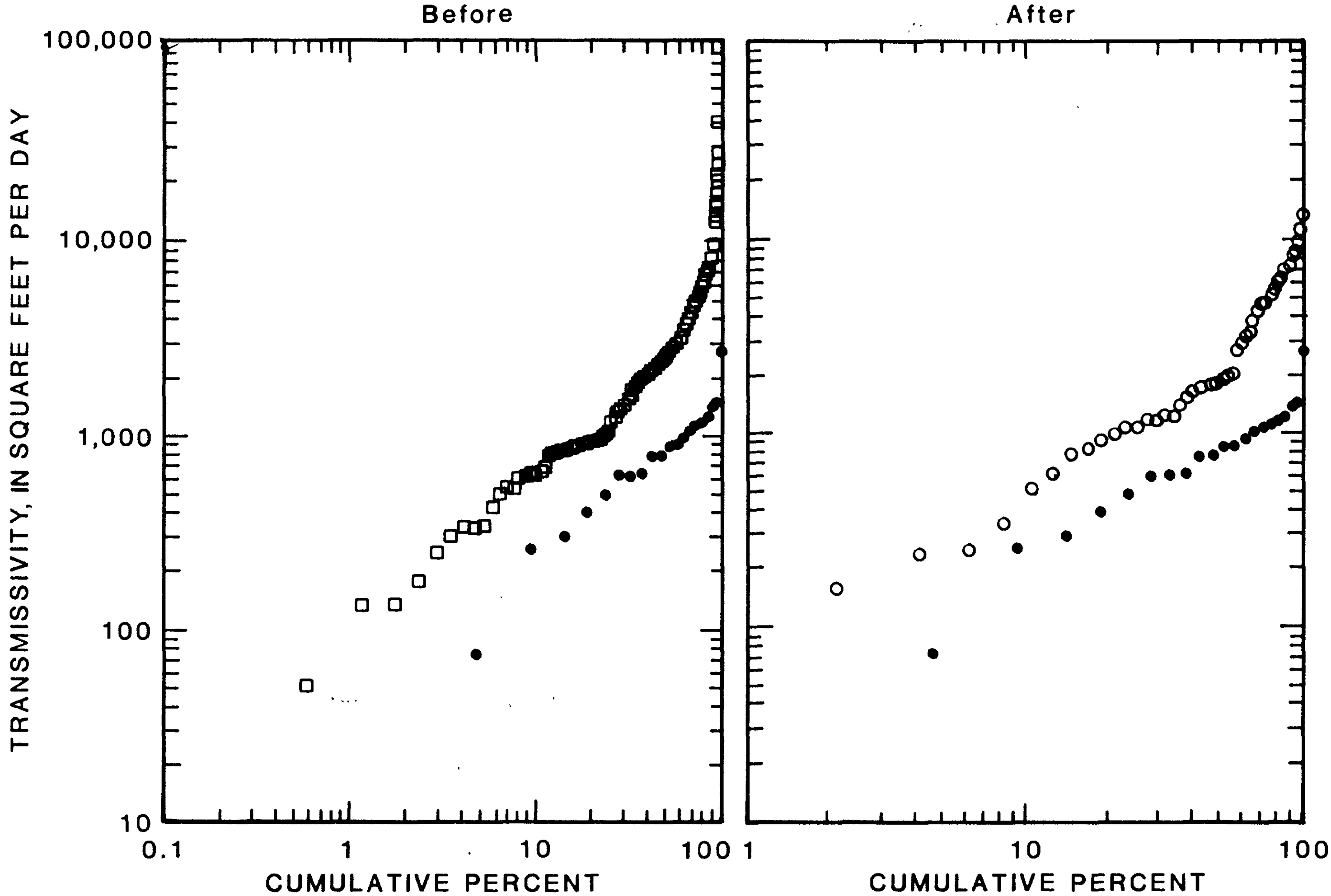
The equation is based on the following assumptions: (1) the aquifer is homogeneous and isotropic and (2) has infinite areal extent; (3) the well penetrates the entire aquifer; and (4) the water removed from storage is discharged instantaneously with decline in head. These assumptions do not accurately describe the aquifer or conditions of all tests, but the approximation is useful because of the limited available data.

The effect of well development on transmissivity values is shown in figure 8. Transmissivity calculated from tests prior to well development is lower than that from tests known to have been made after development (fig. 8b). The relationship is similar for transmissivity from tests whose treatment history is unknown and that from tests made prior to treatment (fig. 8a). Thus, if test data were without descriptions of treatment, the tests were assumed to be from untreated wells, the distribution of transmissivity values would be significantly different. It was concluded, therefore, that many of the tests whose treatment history is unknown were made after treatment. Therefore, a subjective evaluation of individual transmissivities from unknown tests was made to determine whether each value fit the estimated regional pattern of transmissivity derived from tests conducted before treatment. Unknown tests that fit into the estimated regional patterns were accepted for use in the interpretation of the hydraulic characteristics of the St. Peter-Jordan aquifer.

The range of transmissivity values estimated from specific capacity of production wells is from 50 to more than 38,000 ft^2/d (square feet per day). However, this range may include tests in poorly developed wells and wells in which production had been successfully increased by multiple acidization or explosive treatment or both. The large variation in transmissivity values over short distances allows the interpretation of a number of patterns of transmissivity, thus, it is difficult to select representative regional transmissivity values. In all interpretations of the data, however, the assumption must be made that the aquifer is isotropic.

One result of the transmissivity analysis conducted for this study is interpretation of the nature of the permeability of the rocks making up the aquifer. The commonly held hypothesis for flow through the aquifer has been that the sandstones have a greater capability of transmitting and storing water than do the dolomites (Horick and Steinhilber, 1978). The transmissivity, as determined for this study, is greatest in northeastern and southeastern Iowa and smallest in the west (fig. 9). Many of the largest values of transmissivity are in areas where sandstone is a small part of the total aquifer thickness. In these areas the Jordan Sandstone is either a well cemented sandstone with interbedded dolomite, or a sandy dolomite. Therefore, it is likely that the largest transmissivities result from fracture or solution-opening permeability in the dolomite of the Prairie du Chien and Jordan. In Missouri, wells locally yield as much as 1,000 gal/min through well developed solution channels in dolomites equivalent to the Prairie du Chien Group and Jordan Sandstone (Imes, 1985). Fracture or solution-opening permeability is more compatible with the observed large variability in specific capacity or transmissivity over short distances than is the interstitial permeability. An additional factor contributing to large increases toward the south is the increase in the amount of more permeable sandstone in the Prairie du Chien Group or stratigraphically equivalent units. The Gunter Sandstone Member of the Gasconade Dolomite, equivalent to part of the lower Prairie du Chien Group in Missouri, is reported to be a valuable source of water (Imes, 1985).

The storage coefficient can not be accurately estimated from the available data. Consequently, the value of 0.00025, as reported in Horick and Steinhilber (1978), was taken to make initial estimates of transmissivity from specific-capacity data. This value of storage coefficient was applied uniformly throughout the study area. The value reported in Horick and Steinhilber (1978) was "...based on a few aquifer tests in northern and eastern parts of the state...". Uniform variation of storage coefficient was tested during flow simulations which will be described later.



EXPLANATION

● TRANSMISSIVITY DETERMINED FROM PUMPING TEST PRIOR TO TREATMENT FOR WELL DEVELOPMENT

○ TRANSMISSIVITY DETERMINED FROM PUMPING TEST AFTER TREATMENT FOR WELL DEVELOPMENT

□ TRANSMISSIVITY DETERMINED FROM PUMPING TEST WHERE DETAILS OF WELL TREATMENT IS UNKNOWN

Figure 8.--Transmissivity of the St. Peter-Jordan aquifer determined from specific-capacity data.

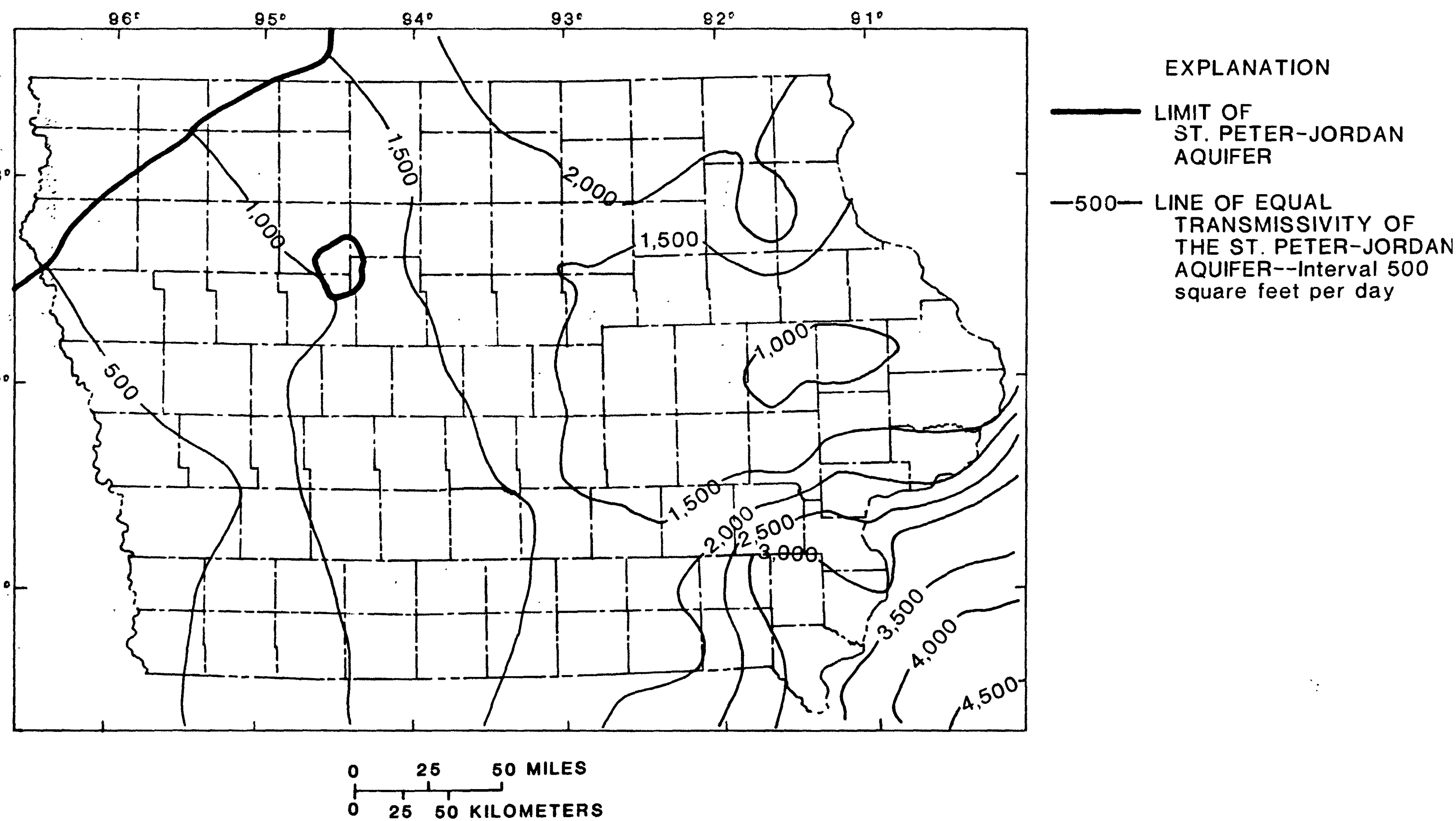


Figure 9.--Transmissivity of the St. Peter-Jordan aquifer in Iowa.

SIMULATION OF FLOW IN THE ST. PETER-JORDAN AQUIFER

A three-dimensional ground-water flow model was developed and calibrated to simulate the flow in the St. Peter-Jordan aquifer under steady-state and transient conditions. The calibrated models were used to evaluate flow in the St. Peter-Jordan aquifer, determine the applicable range of aquifer hydraulic characteristics, and predict the effects of future pumping on the aquifer.

Description of Digital Model

The computer code used to simulate the ground-water flow in the St. Peter-Jordan aquifer is a finite-difference, three-dimensional, flow model developed by McDonald and Harbaugh (1984). The model solves the ground-water flow equation with respect to time and space.

A rectangular two-dimensional grid is superimposed on the defined aquifer system. The discretized hydraulic values of the aquifer system of each cell are assigned to the center of a cell within the grid system. This discretization requires that assumptions be made about hydrologic conditions and that the direct correlation of simulated results to observed conditions be limited. For example, hydraulic head, vertical leakance, and transmissivity distributions must be represented by a single interpolated value for each grid cell, which eliminates local variations. Values computed at the center of the cell cannot be compared directly to observations made near the edges of the cell.

Ground-water flow in the St. Peter-Jordan was simulated for predevelopment (steady-state) and 1980 pumping (transient) conditions. The variables common to both phases were calibrated using predetermined calibration criteria within the limits of known geohydrologic factors. The calibrated steady-state model was used to estimate hydraulic conditions in the aquifer and ground-water flow related to hydrologic boundary assumptions. The calibrated model consists of those values of factors that produce a head distribution within the predetermined calibration criterion of $\pm$ 25 ft. In order to test which values of transmissivity and leakance, from the range obtained in the steady-state model, were the most reasonable values, various combinations were used to simulate head with the transient model for the period 1901-1980. The computed head changes were then compared to observed head changes during the 80-year period. A discussion of transmissivity, vertical leakance, boundaries, and storage coefficients based on model calibration is presented below, along with a sensitivity analysis of variables.

The horizontal dimensions of the grid system used to simulate flow in the St. Peter-Jordan aquifer in Iowa are 340 mi (miles) from west to east and 235 mi from north to south. This area is divided into 3,196 cells, each five mi^2 (square mile) (fig. 10). Five-mile spacing approximates the smallest distance between pumping centers. The hydrologic conditions for the entire cell are represented by a single value for each variable at a central point of a cell as discussed before.

The types of cells used in the model are no-flow, variable-head or active, and constant-head cells. No-flow, or inactive cells have zero transmissivity values. Active cells are those in which the head may vary throughout the simulation, although conditions may be specified to restrict flux and head in the cells. Within an active cell, a variety of options are available to simulate various types and combinations of hydrologic conditions. Constant-head cells are those in which the heads remain equal to the initial head throughout the simulation.

The model contains two aquifer layers. The uppermost represents the overlying aquifer comprised of Silurian and Devonian rocks (layer 1) which act as a source of recharge to or discharge sink from the St. Peter-Jordan aquifer. The lower of the two represents the St. Peter-Jordan aquifer (layer 2) itself. These two aquifers are separated by a confining unit which allows vertical leakage to or from the St. Peter-Jordan aquifer. The head at all cells within the overlying (source) aquifer was assumed to remain constant with time. No other hydraulic interpretations are needed for the overlying aquifer because flow within this layer was not simulated. The relation between hydrogeologic units and equivalent layers in the digital model is shown in table 1.

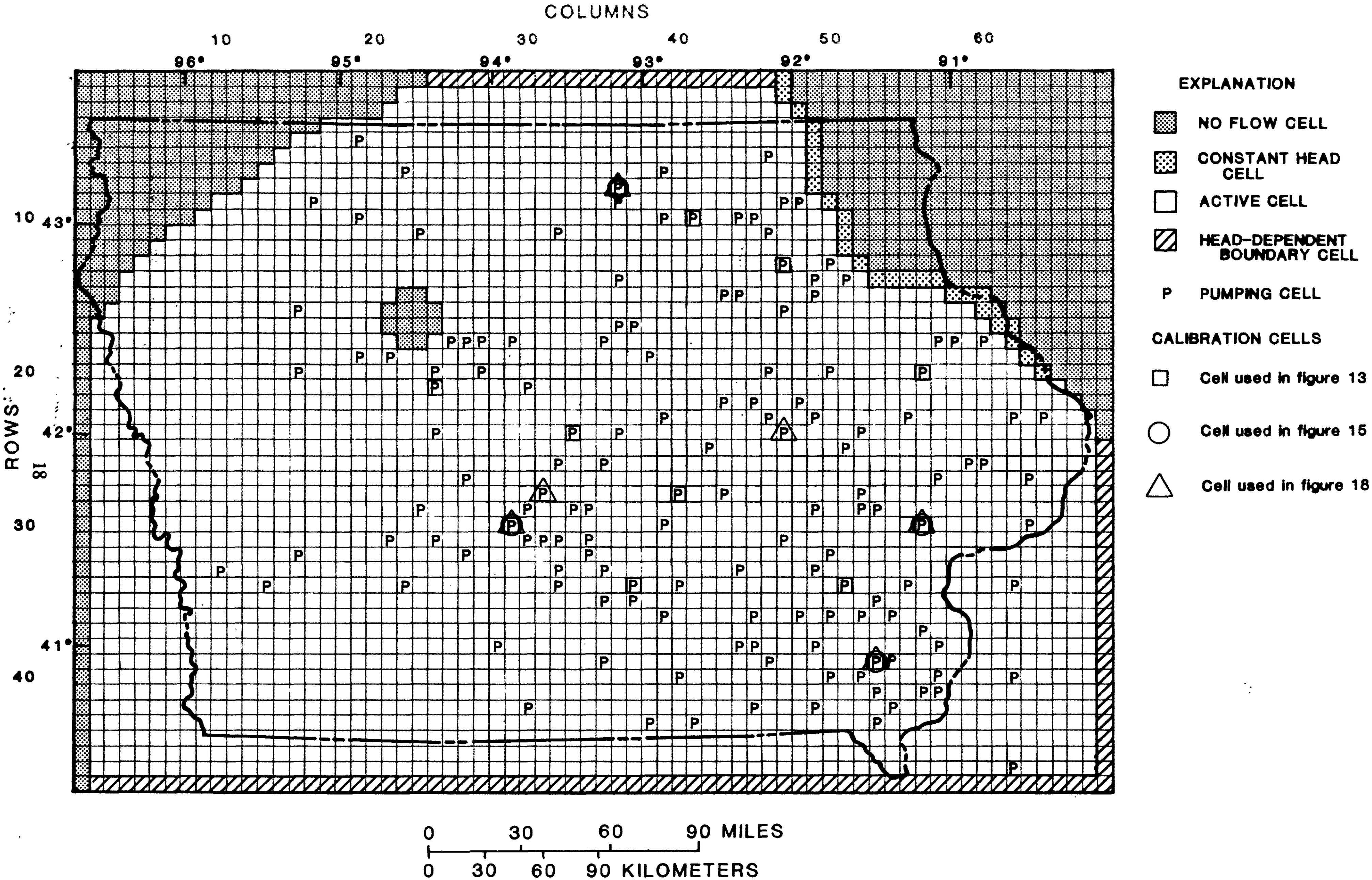


Figure 10.--Model grid and location of boundary, pumping and active cells used in the simulation of the St. Peter-Jordan aquifer.

Model Requirements

Boundaries

The dimensions of the simulated area were extended more than 15 mi to the south into Missouri and to the north into Minnesota, and sufficiently far to the west to include all of Iowa in the active cells of the model (fig. 10). The eastern limit was extended to include the part of Illinois necessary to simulate flow beneath the Mississippi River.

The boundary cells used in the final simulation are either active or no-flow. A small number of constant-head cells are used in the area where the aquifer is in direct contact with alluvium underlying the Mississippi River or the area of water-table conditions in northeastern Iowa (fig. 10). The head values used in these constant-head cells are the predevelopment heads shown in figure 3. It is assumed that heads in the aquifer along the Mississippi River segment are controlled by the stage in the river and that the stage has changed less than ±25 ft in the last 80 years. In northeastern Iowa, constant-head cells are used to represent the edge of the outcrop area of the rocks that comprise the aquifer. Much of this boundary parallels eastern Iowa streams, which are cut into and receive ground-water discharge from local aquifers comprised of carbonate rocks overlying the St. Peter-Jordan aquifer. These rocks are greatly fractured or have solution openings and are assumed to provide a path of discharge for the St. Peter-Jordan aquifer.

Interpretations from a regional scale (Young and others, 1986) show that the St. Peter-Jordan aquifer beneath the Mississippi River in southeast Iowa allows ground-water to flow into and out of Illinois. Therefore, this part of the eastern boundary of the model is represented by active cells. The area of active cells included in western Illinois is believed to be sufficient to allow simulation of flow in the aquifer beneath the Mississippi River. No attempt was made to accurately simulate hydrologic conditions in the part of western Illinois included in this report. Head and transmissivity data used in the simulation of western Illinois cells are from Young (1987).

Inactive cells are used to represent a no-flow boundary at the western edge of the model where the aquifer is interpreted to pinch out. In northwestern Iowa the St. Peter-Jordan aquifer directly underlies the Dakota aquifer in the area shown on figure 10. This area was simulated by incorporating the head for the Dakota aquifer in layer 1 and setting the confining unit thickness to zero ft.

The northern and southern boundaries are simulated by active, head-dependent, variable-flux cells (fig. 10). The heads in these boundary cells were allowed to vary and the flux from each boundary cell outside the model was calculated by using the river module of the digital model (McDonald and Harbaugh, 1984).

Use of this type of active boundary incorporates the interpretation that the aquifer exists, hydrologically, beyond the defined boundary and that flow to or from the boundary is dependent on the head in the adjacent areas. This allows the head at the model boundary to vary with pumping stresses, but requires the assumption that heads in the adjacent areas outside the model be stable and have little effect due to stresses to the St. Peter-Jordan aquifer. North of the modeled area, constant heads were assumed at the points where the aquifer subcrops beneath drift as shown in Delin and Woodward (1984). Along the southern boundary, heads were assigned at a distance of 10 mi from the last active cell. This distance is appropriate because there is no significant use of the St. Peter-Jordan aquifer in Missouri (Imes, 1985). A similar analysis of the aquifer in Missouri used no-flow boundaries at the Iowa border following the assumption that this area is sufficiently far from areas of pumping in Missouri so that boundary conditions will not affect pumping (Imes,1985).

In addition to the boundaries described above, which yielded the best results, simulations were made using variable-flux boundaries in combination with no-flow and constant head boundaries. Attempts to simulate boundary conditions with areas of constant head different than those shown in figure 10 produced a calibrated steady-state model. However, results of transient simulations diverged from the measured head, particularly in boundary areas. When experimenting with variable-head cells, injection and withdrawal wells were used to simulate estimated constant-flux boundary conditions. Little progress was made toward a calibrated steady-state simulation using all or some constant-flux boundaries, possibly because little independent basis for estimation of boundary flux was available. Additionally, no-flow boundaries were simulated along the north boundary under the assumption that, conceptually, little flow crossed this boundary. Constant head boundaries were used extensively in some simulations, particularly along the southern boundary, because there is little independent information about the head or other data which would allow an interpretation of the flux across this boundary. These attempts resulted in steady-state simulations that met calibration criteria. However, the results of transient simulations were poor and, as will be shown later, the boundaries exchange a significant amount of water.

Leakage

Leakage to and from the St. Peter-Jordan aquifer is simulated by specifying an overlying aquifer (layer 1) that represents the aquifers comprised of Silurian and Devonian rocks or other aquifers overlying the St. Peter-Jordan such as the Dakota aquifer. A leakance array for the confining unit between layer 1 and the St. Peter-Jordan aquifer (layer 2) was estimated and specifically defined. The head in layer 1 is simulated using a constant head value for each cell. The head values were obtained by overlaying the grid system shown in figure 10 on the potentiometric surface map defined for layer 1 in 1980 (fig. 6) and selecting a representative head value within each cell by interpolation. It is assumed that the changes in head in the overlying aquifer have been small. Few data are available to test this assumption, however in one of the areas where the aquifers comprised of Silurian and Devonian rocks was most intensively pumped, there had been a measured decline in head of only about 20 ft (Horick, 1984).

Leakance defines the rate of movement of water per unit area per unit thickness of the confining unit between vertically adjacent cells (McDonald and Harbaugh,1984). Leakance is then defined as the vertical hydraulic conductivity of a confining unit divided by its thickness. The program multiplies the leakance by the area of the cell and the head difference between layer 1 and layer 2 to obtain the vertical flux rate and direction across the adjacent cell faces. To obtain an array of leakance values, the thickness of the confining unit between the aquifers comprised of Silurian and Devonian rocks and the St. Peter-Jordan aquifer (fig. 7) at each cell was calculated. This thickness was then divided into each of several values of vertical hydraulic conductivity, within the range of estimated values, to produce several leakance arrays spanning the range of likely values of leakance. Vertical conductivity values between 10^{-12} to 10^{-6} ft/s (foot per second) (approximately 10^{-7} to 0.1 feet per day) were tested. Simulation with non-uniform vertical hydraulic conductivity values yielded no significantly different results compared to results with uniform values. Consequently, the leakance array has been calculated using a constant vertical conductivity of 10^{-10} ft/s and the variable thickness of confining unit shown in figure 7. The final values of leakance range from 1.43×10^{-13} to 5×10^{-12} ft/s.

In northwestern Iowa, where the St. Peter-Jordan aquifer is in direct contact with the overlying Dakota aquifer (fig. 10), it is assumed that the vertical hydraulic conductivity of the Dakota aquifer is equal to the vertical hydraulic conductivity of the St. Peter-Jordan and that both aquifers are isotropic.The leakance between the Dakota and the St. Peter-Jordan aquifers in these areas was calculated according to the procedure described by McDonald and Harbaugh (1984, p. 142). This procedure results in a leakance between the two aquifers that is one-half of the vertical hydraulic conductivity of the aquifer divided by the thickness of the confining unit between the cells of the two aquifers.

Transmissivity

Steady-state simulations were used to evaluate distribution of the transmissivity estimated from specific-capacity data. The transmissivities that satisfied steady-state calibration criteria were used in transient simulations. The attempts to match transient data with the transmissivities used in the steady-state model help narrow the acceptable range and distribution of transmissivity values. Several iterations between steady-state and transient simulations were necessary to arrive at an optimum result.

In addition to varying transmissivity values areally, calibration attempts were made using uniform values of transmissivity for the entire aquifer. Generally, the results demonstrated that under steady-state conditions a uniform transmissivity of 1,500 to 3,000 ft^2/d produced a head distribution within the calibration criteria. The uniform value that produced the best fit was 1,800 ft^2/d. However, calibration to transient data was difficult because measured drawdown in individual cells matched the simulated drawdown only where the uniform transmissivity was within the relatively narrow range known to occur at that location. The distribution of transmissivity values finally used in the simulations is the distribution derived from specific-capacity data (fig. 9). The steady-state and transient simulations using this distribution provided the best results. A sensitivity analysis, presented in a later section, demonstrates some of the effects of varying transmissivity values.

Using a geostatistical approach, Hoeksema and Kitanidis (1984) independently estimated the spatial distribution of transmissivity values in the St. Peter-Jordan aquifer under steady-state conditions. They used head and transmissivity data similar to those used in this report. The distribution in figure 9 is within the limits of the 95 percent confidence interval of transmissivities discussed by Hoeksema and Kitanidis.

Storage Coefficient

Storage coefficients that ranged from 0.01 to 0.000001 were tested in order to determine an optimum value that could be uniformly applied to the St. Peter-Jordan aquifer. A spatially uniform coefficient was used because no data are available to adequately determine the coefficient at any point. The value that provided the most reasonable fit was 0.00025, which is the value suggested by Horick and Steinhilber (1978).

Pumpage

Withdrawals from wells in the St. Peter-Jordan aquifer were simulated by summing the known pumpage within each model cell for 10-year periods from 1901-1980. However, some well discharge may also come from wells that tap other aquifers. This is a potential source of error for pumpage estimates of the St. Peter-Jordan aquifer. Approximately 20 percent of the wells are open to other aquifers, including about 7 percent that are open to the aquifers comprised of Silurian and Devonian rocks.

The pumpage data were acquired with the cooperation of municipal and industrial officials. This information was either reported as estimates or was obtained directly from pumping records maintained by water-supply operators. Where neither records nor estimates could be obtained, no withdrawal was assumed, even though there may be evidence that wells were used. Consequently, there are sites where head data are available, but pumping data are not. The underestimation of pumpage may not create significant errors in simulations because the communities for which pumping information was not available are generally the smaller ones, or the missing records are in the periods when pumpage was generally the smallest.

For each of the simulated periods of the 10-year pumping interval, the total pumpage within each cell is expressed as a constant rate. Pumpage is simulated as coming from the entire cell, thus drawdown in pumping wells would be greater than that for the simulated value in the respective cell.

The simulation period, pumping rate, and location of pumping wells in cells are shown in table 2. The increase in pumpage with time is shown in figure 11a. This increase suggests that pumpage increased about threefold between the thirties and forties and has more than doubled the value again from the fifties to the seventies. An examination of the changes in average pumping rate per pumping cell (fig. 11b) indicates a bell-shape trend. This fact indicates that a larger number of pumping cells show pumpage from the St. Peter-Jordan aquifer during the sixties and seventies. This correlates well with the fact that more wells were drilled to the St. Peter-Jordan during recent years. Data error may also be reduced due to improvement of methods of data collection and record-keeping. That is, data from the seventies and eighties are more complete than that from the twenties and thirties.

Steady-State Simulation

Under steady-state conditions, inflow is equal to outflow and no change in storage should occur. Before the aquifer was developed, the aquifer was assumed to be under steady-state conditions and was simulated without pumping.

The predevelopment potentiometric surface shown in figure 3 was interpolated for the head in each cell which was used as the initial head for the simulations. The simulation program calculates new head values and the difference between the calculated and initial head in each cell. The calculated new head will replace the original initial head as the initial head value in the simulation program which calculates new head values and

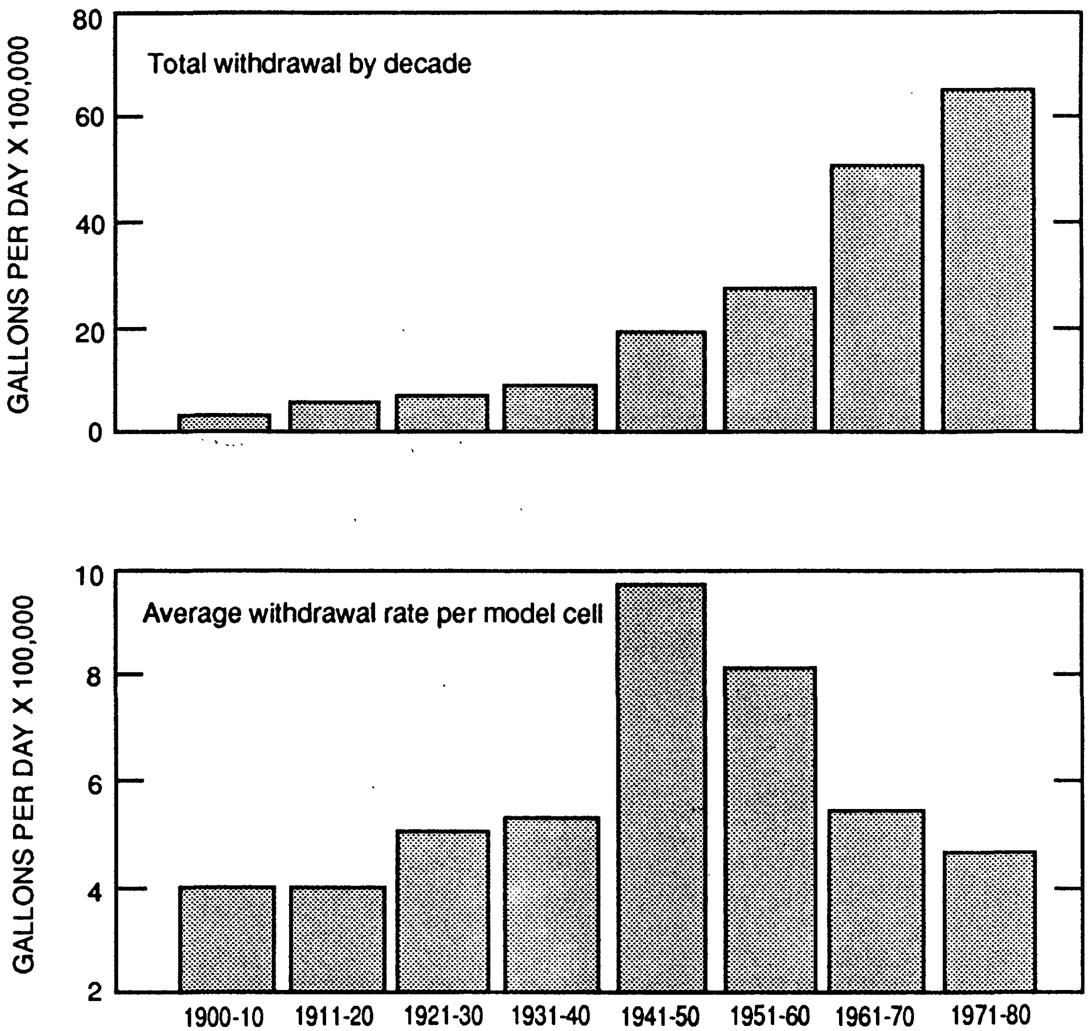


Figure 11.--Pumpage from the St. Peter-Jordan aquifer, 1901-1980.

Table 2.--*Historical pumpage from the St. Peter-Jordan aquifer in Iowa*

[Units of million gallons per day; --, no data]

Model (row)	Cells (column)	Simulation Period 1971-1980	1961-1970	1951-1960	1941-1950	1931-1940	1921-1930	1911-1920	1901-1910
5	19	2.19	1.97	--	--	--	--	--	--
6	46	.33	.28	--	--	--	--	--	--
7	22	.05	--	--	--	--	--	--	--
7	39	.42	.35	0.27	0.22	--	--	--	--
8	36	1.17	1.75	1.75	1.75	--	--	--	--
9	16	1.44	2.27	--	--	--	--	--	--
9	36	7.12	5.94	5.19	4.96	1.80	0.75	0.52	0.15
9	47	.01	.01	--	--	--	--	--	--
9	48	.02	.01	--	--	--	--	--	--
10	19	.03	--	--	--	--	--	--	--
10	39	.04	.04	--	--	--	--	--	--
10	41	.18	.42	.62	.44	.68	.49	.29	--
10	44	.05	.24	--	--	--	--	--	--
10	45	.58	--	--	--	--	--	--	--
11	23	.16	.03	--	--	--	--	--	--
11	32	.05	--	--	--	--	--	--	--
11	46	.24	.24	--	--	--	--	--	--
13	47	.20	.15	.15	--	--	--	--	--
13	50	.14	.04	--	--	--	--	--	--
14	36	--	.15	.25	.12	.09	--	--	-
14	49	.05	--	--	--	--	--	--	--
14	51	.02	.02	--	--	--	--	--	--
15	43	.58	.58	--	--	--	--	--	--
15	44	.04	--	--	--	--	--	--	--
15	49	.83	.77	--	--	--	--	--	--
16	15	.15	.10	--	--	--	--	--	--
16	47	.06	.03	--	--	--	--	--	--
17	36	.23	.24	--	--	--	--	--	--
17	37	.28	.26	--	--	--	--	--	--
18	25	.58	--	--	--	--	--	--	--
18	26	3.37	2.79	1.11	--	1.17	1.01	.74	--
18	27	1.22	.61	--	--	--	--	--	--
18	29	1.18	1.02	.88	.62	--	--	--	--
18	35	.04	.14	.14	.11	--	--	--	--
18	57	.01	--	--	--	--	--	--	--

Table 2.--*Historical pumpage from the St. Peter-Jordan aquifer in Iowa--Continued*

Model (row)	Cells (column)	Simulation Period 1971-1980	1961-1970	1951-1960	1941-1950	1931-1940	1921-1930	1911-1920	1901-1910
18	58	0.05	--	--	--	--	--	--	--
18	60	.02	--	--	--	--	--	--	--
19	19	.06	0.06	0.03	--	--	--	--	--
19	21	.30	.25	--	--	--	--	--	--
19	38	.07	.05	--	--	--	--	--	--
20	15	.14	--	--	--	--	--	--	--
20	24	.01	--	--	--	--	--	--	--
20	27	.02	.02	.02	--	--	--	--	--
20	46	.20	--	--	--	--	--	--	--
20	50	.04	.01	--	--	--	--	--	--
20	56	.11	.04	--	--	--	--	--	--
20	64	.22	.22	--	--	--	--	--	--
21	24	.15	.10	.07	--	--	--	--	--
21	30	--	.03	--	--	--	--	--	--
22	43	.16	.08	--	--	--	--	--	--
22	45	.12	.09	--	--	--	--	--	--
22	48	.32	--	--	--	--	--	--	--
23	39	.32	.04	--	--	--	--	--	--
23	46	.03	.03	.01	--	--	--	--	--
23	49	.05	--	--	--	--	--	--	--
23	55	.33	--	--	--	--	--	--	--
23	62	.58	.44	--	--	--	--	--	--
23	64	1.64	.86	--	--	--	--	--	--
23	67	1.01	1.03	.57	0.49	0.59	0.44	0.21	0.18
24	24	.06	--	--	--	--	--	--	--
24	33	.48	.40	--	--	--	--	--	--
24	36	.19	--	--	--	--	--	--	--
24	47	.07	.06	--	--	--	--	--	--
24	52	.86	.62	.13	--	--	--	--	--
25	42	1.91	1.46	--	--	--	--	--	--
25	51	4.72	2.54	2.00	1.08	.16	1.36	1.54	1.15
26	32	.04	--	--	--	--	--	--	--
26	35	.01	--	--	--	--	--	--	--
26	59	.06	.06	.06	.06	--	--	--	--
26	60	.06	.04	--	--	--	--	--	--

Table 2.--*Historical pumpage from the St. Peter-Jordan aquifer in Iowa--Continued*

Model (row)	Cells (column)	Simulation Period 1971-1980	1961-1970	1951-1960	1941-1950	1931-1940	1921-1930	1911-1920	1901-1910
26	67	1.69	1.64	1.24	1.14	1.05	--	--	--
27	26	.58	.03	--	--	--	--	--	--
27	57	--	--	.08	.17	.13	0.10	0.06	--
27	63	.21	--	--	--	--	--	--	--
27	66	4.61	3.77	1.50	1.25	.50	.50	.49	0.48
28	31	1.12	.36	--	--	--	--	--	--
28	40	1.18	.98	.68	--	--	--	--	--
28	43	.12	.08	--	--	--	--	--	--
28	52	.16	.12	--	--	--	--	--	--
29	23	.12	--	--	--	--	--	--	--
29	30	.24	.08	--	--	--	--	--	--
29	33	.52	.21	--	--	--	--	--	--
29	34	--	.03	.04	--	--	--	--	--
29	49	.02	--	--	--	--	--	--	--
29	52	.44	.20	--	--	--	--	--	--
29	53	.10	.08	--	--	--	--	--	--
30	29	.35	.08	--	--	--	--	--	--
30	39	.06	.06	--	--	--	--	--	--
30	56	.86	.19	.16	.14	.11	.08	.06	--
30	63	.01	.01	.01	--	--	--	--	--
31	21	.10	--	--	--	--	--	--	--
31	24	.16	.10	--	--	--	--	--	--
31	30	1.59	.43	--	--	--	--	--	--
31	31	.08	.02	--	--	--	--	--	--
31	32	.05	.06	--	--	--	--	--	--
31	34	.03	.01	--	--	--	--	--	--
31	47	.08	--	--	--	--	--	--	--
31	63	2.68	2.00	.53	.49	.05	.03	.02	.01
32	15	.09	--	--	--	--	--	--	--
32	26	.12	.04	--	--	--	--	--	--
32	34	.02	--	--	--	--	--	--	--
32	50	.14	--	--	--	--	--	--	--
33	10	.01	--	--	--	--	--	--	--
33	32	.70	.18	--	--	--	--	--	--
33	35	.18	--	--	--	--	--	--	--

Table 2.--*Historical pumpage from the St. Peter-Jordan aquifer in Iowa--Continued*

Model (row)	Cells (column)	Simulation Period 1971-1980	1961-1970	1951-1960	1941-1950	1931-1940	1921-1930	1911-1920	1901-1910
33	44	0.09	--	--	--	--	--	--	--
33	49	.10	--	--	--	--	--	--	--
34	13	.35	0.02	--	--	--	--	--	--
34	22	.07	.04	0.04	0.04	0.04	--	--	--
34	32	.16	.42	--	--	--	--	--	--
34	37	1.02	.18	--	--	--	--	--	--
34	40	.01	--	--	--	--	--	--	--
34	51	.87	.61	.53	.46	.38	0.30	0.23	--
34	55	.37	.45	--	--	--	--	--	--
35	35	.08	--	--	--	--	--	--	--
35	37	.03	--	--	--	--	--	--	--
35	53	.02	.02	.01	--	--	--	--	--
36	39	.04	--	--	--	--	--	--	--
36	45	.06	--	--	--	--	--	--	--
36	48	.06	--	--	--	--	--	--	-
36	50	.06	--	--	--	--	--	--	--
36	52	.05	--	--	--	--	--	--	--
36	54	.07	--	--	--	--	--	--	--
37	56	.09	.05	.06	--	--	--	--	--
38	28	.03	.01	--	--	--	--	--	--
38	44	2.00	6.00	6.00	4.00	--	--	--	--
38	45	.04	--	--	--	--	--	--	--
38	49	.46	.44	--	--	--	--	--	--
38	57	.15	--	--	--	--	--	--	--
39	35	.06	--	--	--	--	--	--	--
39	46	.10	.10	--	--	--	--	--	--
39	53	.96	.82	.70	.58	.46	.33	.21	--
39	54	.17	.12	--	--	--	--	--	--
40	40	.03	--	--	--	--	--	--	--
40	50	.02	--	--	--	--	--	--	--
40	52	.02	--	--	--	--	--	--	--
40	57	.26	.11	--	--	--	--	--	--
41	53	.01	--	--	--	--	--	--	--
41	56	.02	--	--	--	--	--	--	--
41	57	.13	.06	--	--	--	--	--	--

Table 2.--*Historical pumpage from the St. Peter-Jordan aquifer in Iowa--Continued*

Model (row)	Cells (column)	Simulation Period 1971-1980	1961-1970	1951-1960	1941-1950	1931-1940	1921-1930	1911-1920	1901-1910
42	30	0.16	0.12	--	--	--	--	--	--
42	45	.02	.01	--	--	--	--	--	--
42	49	.13	--	--	--	--	--	--	--
42	54	.02	--	--	--	--	--	--	--
43	38	.02	--	--	--	--	--	--	--
43	41	.08	.08	--	--	--	--	--	--
43	53	.10	.10	--	--	--	--	--	--
Total		64.48	49.50	24.83	18.12	7.21	5.39	4.37	1.97
Number of cells		143	94	31	19	14	11	11	5

head differences. These repeated calculations of head are the iteration process of the simulation. Until there is no head difference or the head difference in each cell is equal to or smaller than a predefined, very small value, then the calculation process is terminated, and a solution of the simulation is obtained.

Two methods of comparing the simulated and predevelopment head were used. The first was a visual comparison of the simulated potentiometric surface with the potentiometric surface constructed from measured heads (fig. 3). The second method used the arithmetic average of the absolute value of head differences in all active cells.

The goal of using visual comparison of potentiometric surfaces was to simulate a head array in which all head contours were no farther from the contour of the same value than from the next contour. This criterion permitted a maximum deviation of 50 ft. The other criterion for calibration was that the arithmetic average of the absolute value of head difference in each cell is equal to or less than 25 ft.

The comparison of predevelopment potentiometric surfaces constructed from measured heads and simulated heads is shown in figure 12. Note that the simulated potentiometric surface has smoother contour lines and the overall position of the simulated potentiometric surface is generally higher than the potentiometric surface constructed from measured heads except in north-central and parts of east-central Iowa.

The average of the absolute value of head differences in all active cells is 19.81 ft. The number of active cells in which the head differences exceed 50 ft is 174, or about 6 percent of all active cells. The number of active cells in which the head differences exceed 25 ft is 778, or about 30 percent of the total active cells. Head differences greater than 25 ft generally occur in areas where head data for construction of the predevelopment surface shown in figure 3 are uncertain or the hydraulic gradients in areas are very large between cells, such as in northeastern Iowa.

Transient Simulation

Transient conditions were simulated by incorporating historical pumpage, coefficient of storage, and hydrologic variables used in the steady-state simulation. The transient simulation used the head obtained by the steady-state simulation as the initial head array. This simulated predevelopment head represents the head in the year 1900. The steady-state simulated head was used rather than the head interpreted from the

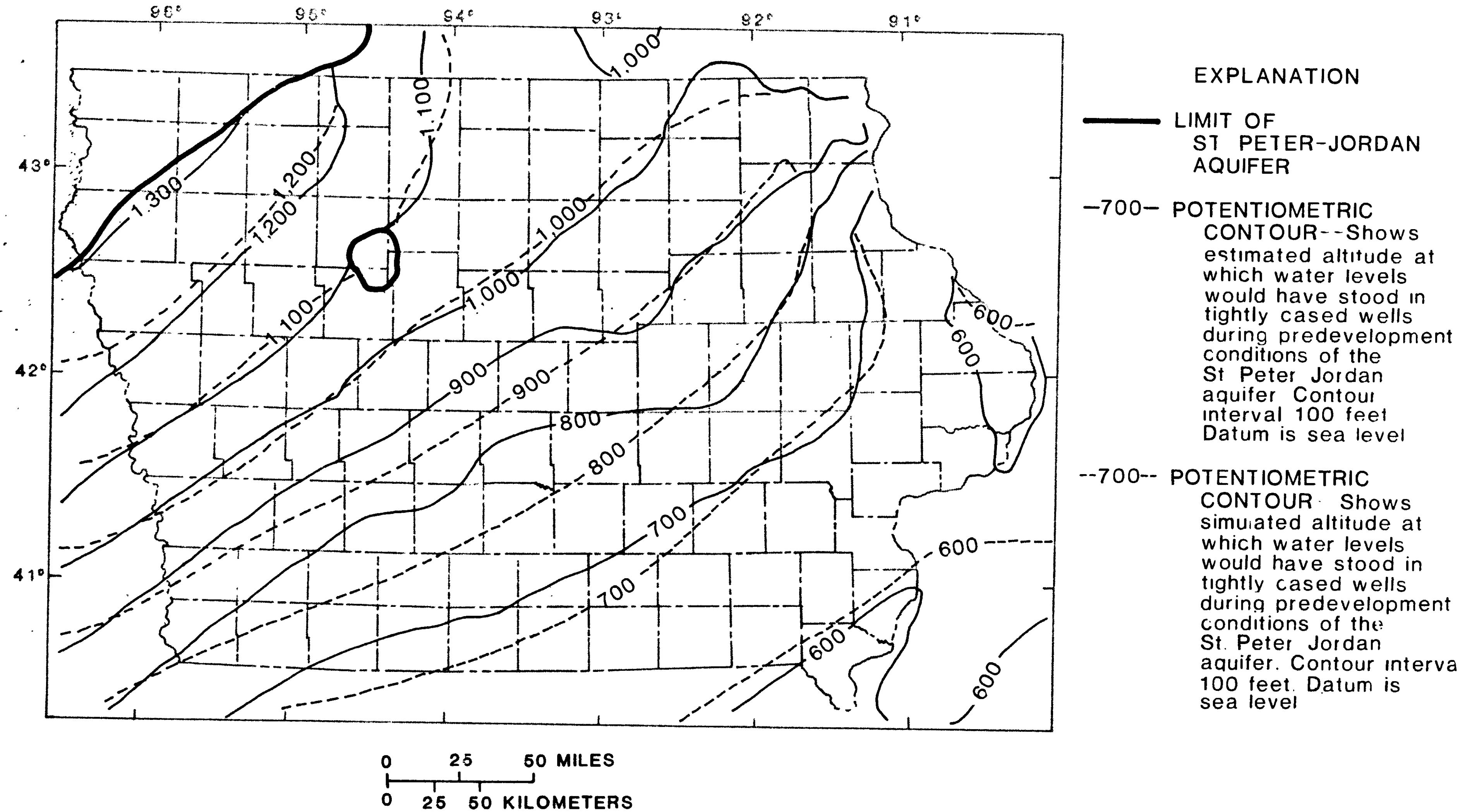


Figure 12.--Estimated predevelopment and simulated steady-state predevelopment potentiometric surfaces for the St. Peter-Jordan aquifer.

potentiometric surface constructed from measured heads. This is to eliminate the errors introduced during the interpretation. Eight pumping periods of ten years per period were used to simulate head changes resulting from pumping from 1901 to 1980. The head calculated at the end of each pumping period is used as the initial head array for the next period.

The cells used for transient calibration (fig. 10) contained wells with reliable water-level measurements for at least the last 20 years. The lack of widely distributed transient head data, in both time and location, requires that the model calibration be based on those few areas where accurate comparisons can be made. Some areas with a large number of wells, such as cell 25-51, have some anomalous heads that were not considered in the analysis. In some areas, only two recent head measurements may be available, but these observations show the effects on the aquifer resulting from intensive pumping. In addition, this recent period represents the most critical time for the model calibration because the head decline is an ultimate result of the largest pumping and the quality of data is generally good during this period.

The transient simulation was calibrated by comparing the model-calculated heads with the measured heads from predevelopment to 1980 in selected cells and comparing the overall head declines. Hydrographs were drawn for wells located in specific cells. Some of the hydrographs are a composite of head measurements in several wells. The model results for the specific cells were plotted and compared to the well hydrographs. Because a model-calculated head represents the head at the center of a cell and is not a head at a well located at the cell center, inaccuracies can be introduced when comparing the calculated head to a measured head from various locations within the cell. These inaccuracies are greatest in areas of steepest ground-water gradient and where nearby pumping wells may greatly influence the water level measured at a site.

The results of the transient simulation were used to describe net head changes. The shapes of the hydrographs of simulated head in figure 13 are similar to those of hydrographs of measured head. Although there are variations between measured and simulated head, generally the hydrograph slopes are similar and, in some cases, nearly identical. Further improvement of the transient simulation may be achieved by adjusting the hydrologic variables of the aquifer, however, the distribution of known information is not sufficient to support such refinement of the simulation.

Sensitivity Analysis

Steady-State Simulation

The sensitivity of the calibrated steady-state simulation to variation of the hydrologic parameters was tested by separately changing the transmissivity of layer 2 and the leakance between layers 1 and 2. The sensitivity to each parameter was tested by the same procedure used in the calibration process, namely, comparisons of contoured potentiometric surfaces and of average head differences.

The sensitivity of the steady-state simulation to changes in transmissivity was tested by multiplying each value within the calibrated transmissivity array by 0.1, 0.5, 2.0 and 10.0. Each of these transmissivity arrays was used to simulate new head distributions with other parameters unchanged from the calibrated simulation. The graph in figure 14 compares the average head differences from the calibrated steady-state simulation (19.81 ft) and those from the simulations using other transmissivity values.

The effect of increasing or decreasing the transmissivity array, as shown in figure 14, is to increase the average head difference per active cell. The effects of increasing transmissivity by a factor of two is to increase the average head difference by over 17 percent to 23.19 ft and increasing the transmissivity by a factor of 10 more than doubles the average head difference to 43.41 ft. The effect of reducing the transmissivity array from the calibrated array is similar. Reduction by a factor of two increases the average head difference by about 19 percent to 23.56 ft and a reduction of the transmissivity by a factor of 10 increases the average head difference to 38.86 ft, double the average head difference determined from the calibrated simulation.

Increases in transmissivity values produce a potentiometric surface generally having lower altitudes than that from the calibrated simulation. The more the transmissivity is increased, the more the potentiometric-surface altitude and hydraulic

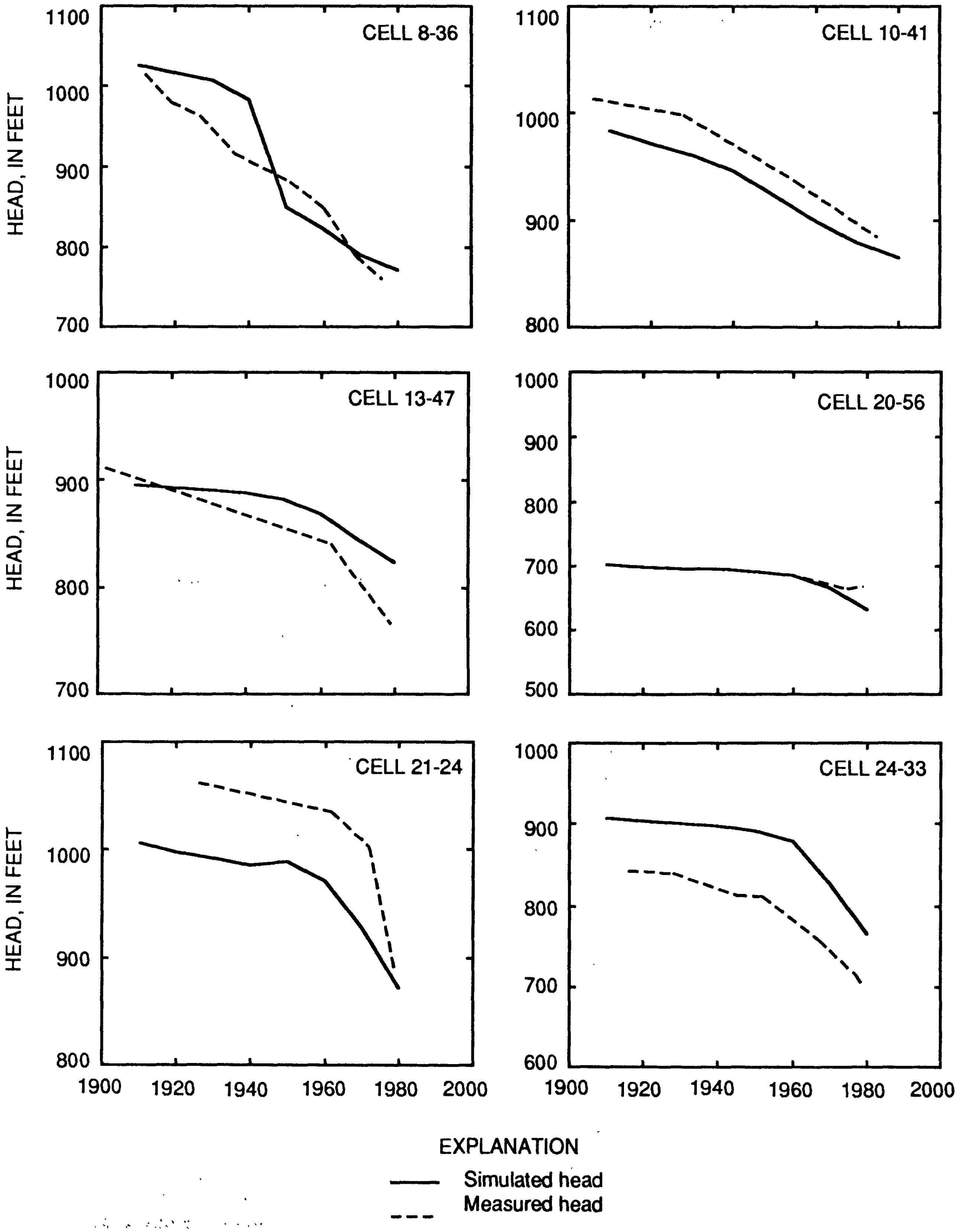


Figure 13.--Simulated and measured head at selected model cells.

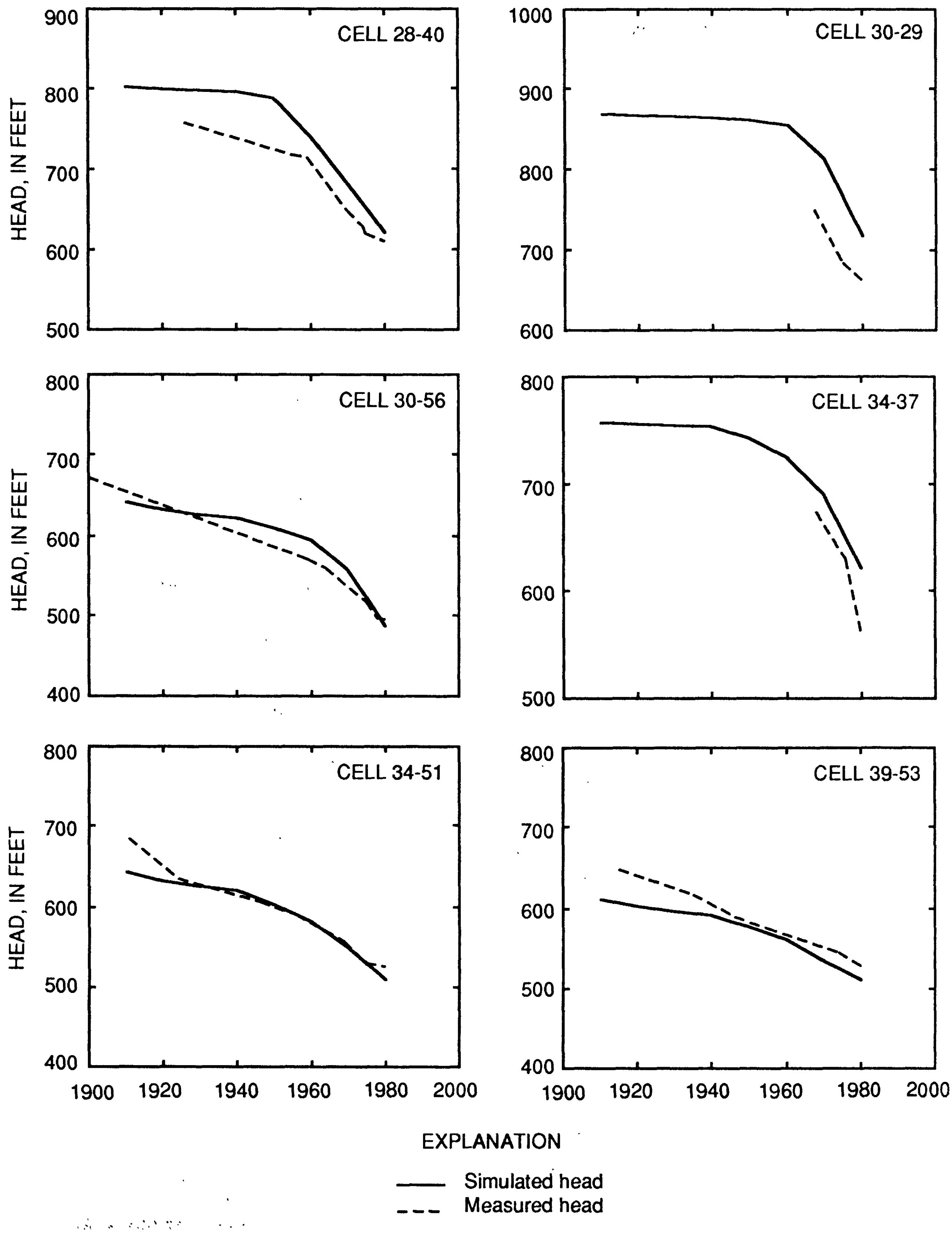


Figure 13.--Simulated and measured head at selected model cells--Continued.

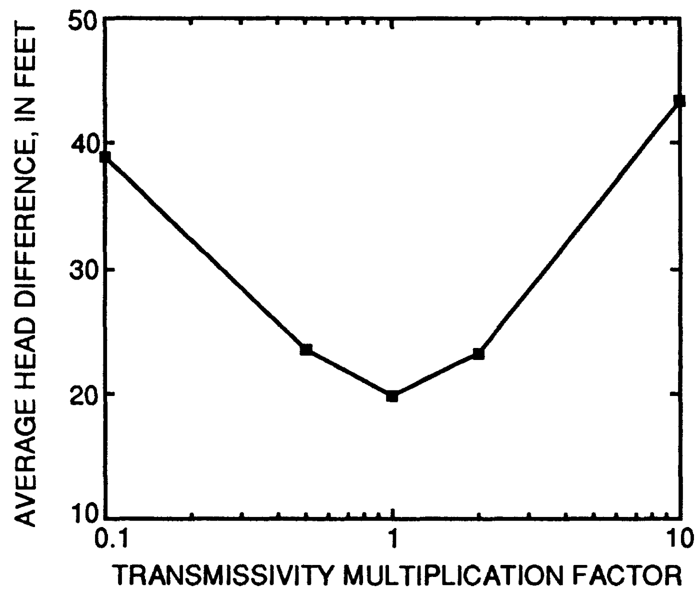


Figure 14.--Effect of variation in transmissivity on steady-stead head.

gradient decrease, except near lateral boundaries. As would be expected, a decrease in transmissivity values has the opposite effect on head, although the change is greater.

The sensitivity of the steady-state simulation to changes in leakance can be described as essentially the inverse of the sensitivity to change in transmissivity values. When the leakance was increased, the potentiometric surface rose and when the leakance was decreased, the surface declined. The magnitude of these shifts is essentially identical to those caused by changes in transmissivity values, except that they are reversed.

The results of the steady-state sensitivity analysis indicate that the model is quite sensitive to model-wide changes in transmissivity and leakance values of an order of magnitude. Doubling or halving the values of these two parameters seems to produce small changes that are within the calibration criteria. Transmissivity and leakance have an opposite effect on the model head, indicating that many combinations of these two parameters could produce acceptable results in the model. A relatively wide range of values can be substituted for these hydrologic variables and still produce an acceptable steady-state simulation of the St. Peter-Jordan aquifer in Iowa.

Transient Simulation

The sensitivity of the calibrated transient simulation was tested by separately varying the transmissivity and storage coefficient in layer 2 and varying the leakance between layers 1 and 2. Each variable was adjusted by multiplying the calibrated value by 2.0 and 0.5, except storage, which was adjusted by an order of magnitude. Hydrographs of the calculated head produced by the sensitivity trials were compared to those produced by the calibrated simulation (figs. 15, 16, and 17).

The transient model is considerably more sensitive to changes in transmissivity than is the steady-state model. This interpretation may result from the fact that transient sensitivity was evaluated

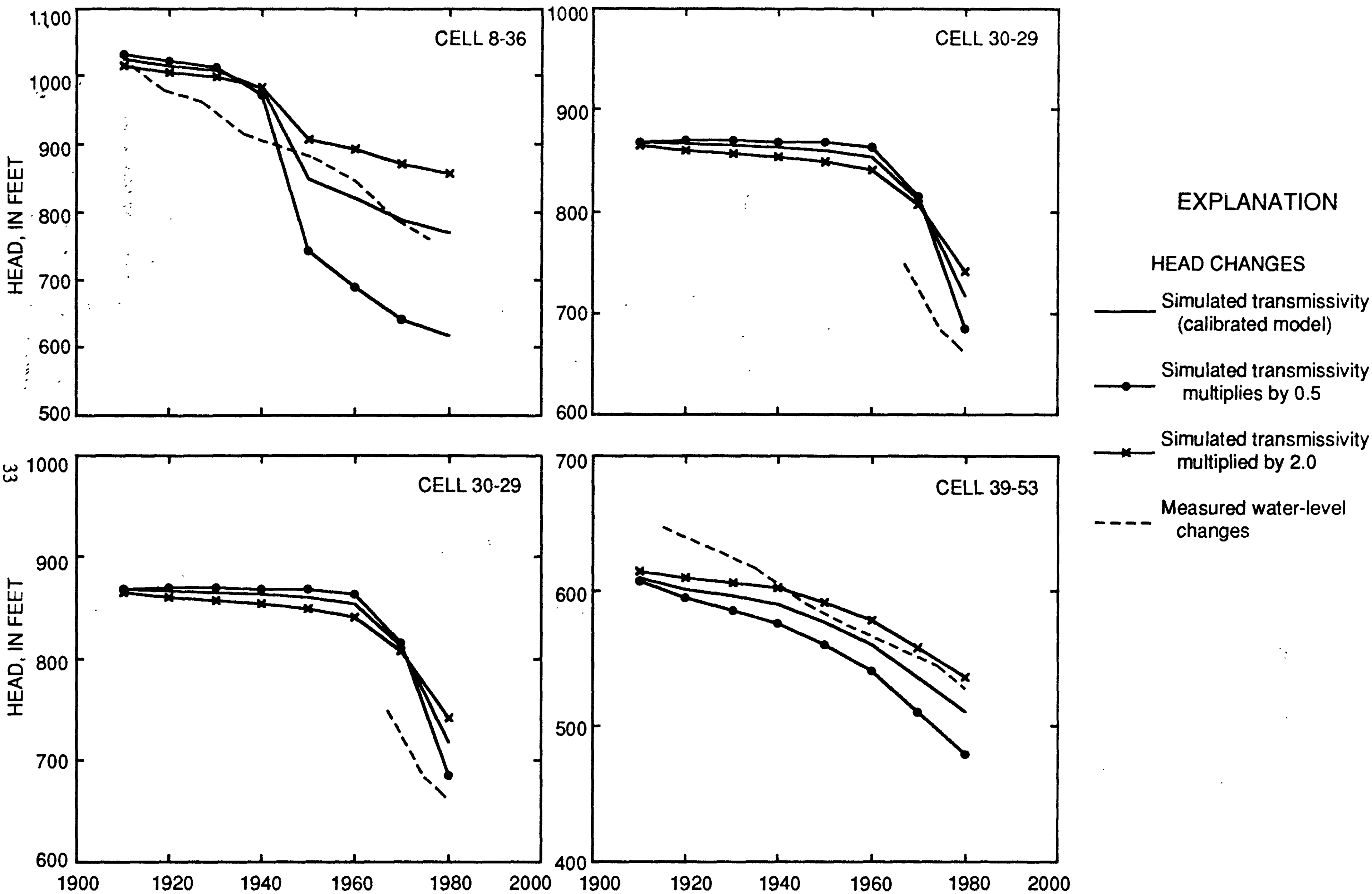


Figure 15.--Effect of variation in transmissivity on simulated head trends in selected model cells.

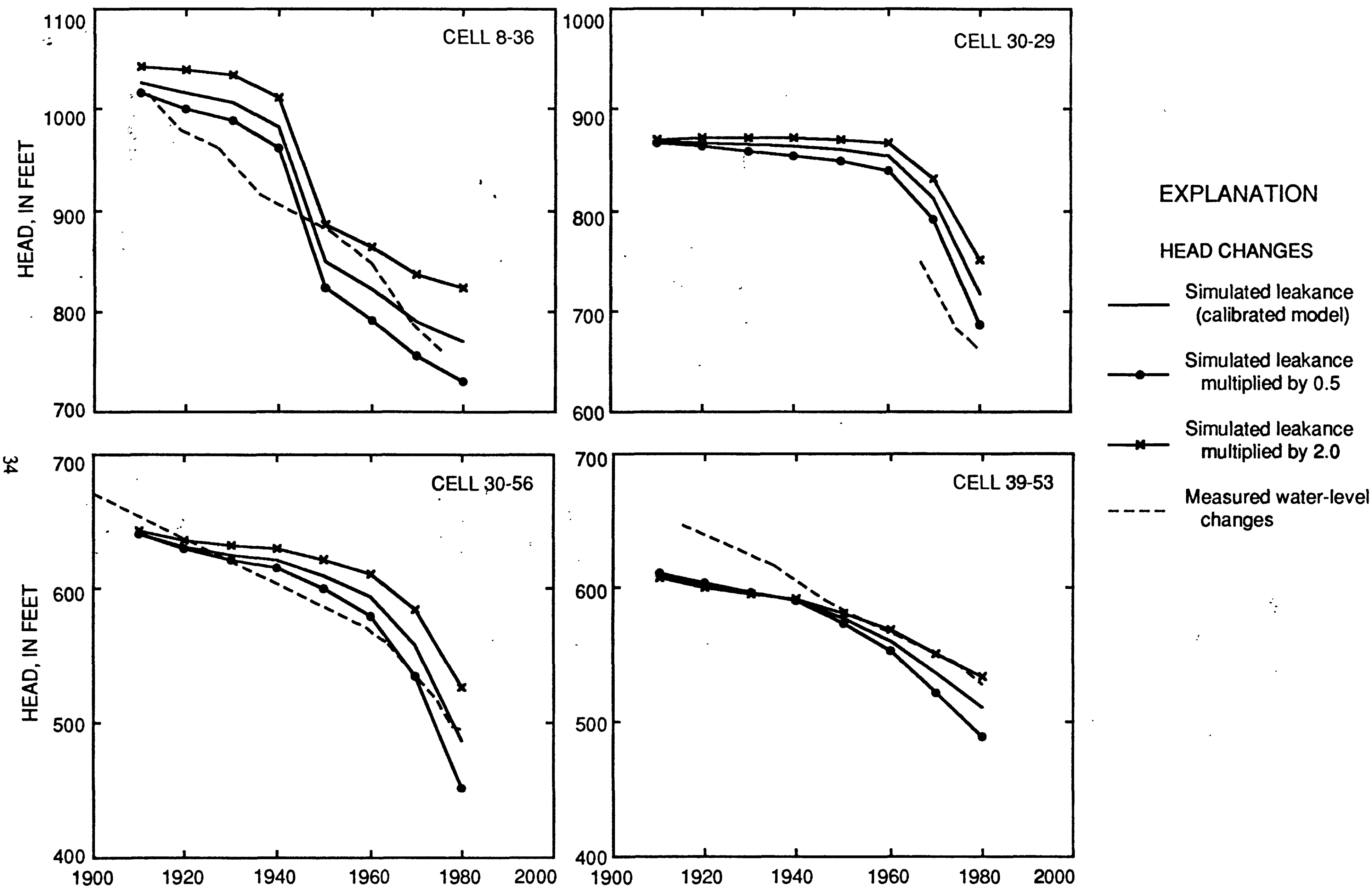


Figure 16.--Effect of variation in leakance on simulated head trends in selected model cells.

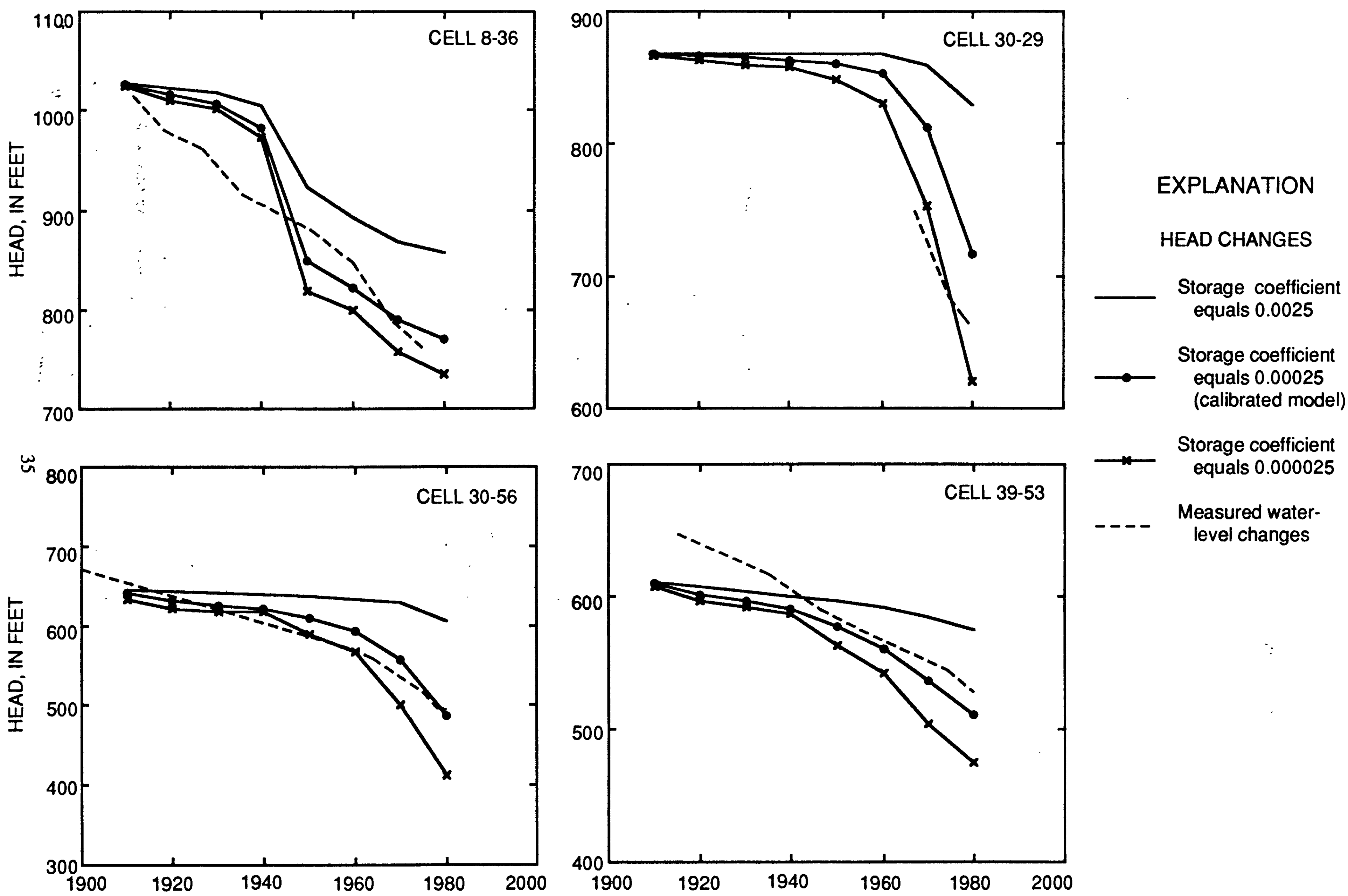


Figure 17.--Effect of variation in storage coefficient on simulated head trends in selected model cells.

at individual cells and steady-state sensitivity was evaluated over the entire aquifer. Multiplying the transmissivity by 2.0 in the transient model produces a head difference for 1980 of more than 85 ft in cell 8-36, from the calibrated head, and multiplying transmissivity by 0.5 produces a head difference of 152 ft in the same cell. The average head difference for the selected cells shown in figure 15 is approximately 42 ft for the simulation with transmissivity multiplied by 2.0 and 62 ft when the transmissivity values were multiplied by 0.5. When transient conditions were simulated using transmissivity arrays multiplied or divided by 10.0, the differences were much greater than those shown here.

The sensitivity of the calibrated transient model to variations in leakance (fig. 16) was tested by multiplying the values of the leakance array by 2.0 and 0.5. This exercise essentially tests the sensitivity of the model to the vertical hydraulic conductivity of the confining unit because the areal variation of the thickness is known with confidence and was not altered in the sensitivity tests. The variations in leakance produce a range of head differences from the simulated 1980 head of 22 to 54 ft with an average of about 34 ft. The effects of varying leakance are similar, although slightly less, to those caused by change in transmissivity values. Consequently, the transient model seems to be generally sensitive to relatively small variations in leakance.

The sensitivity of the transient model to variations in storage coefficient is shown in figure 17. The head differences between the calibrated model, using a storage coefficient of 0.00025, and a simulation using 0.000025 are from 36 to 97 ft and average about 61 ft. Differences in head when a storage coefficient of 0.0025 was used range from 64 to 118 ft with an average of about 85 ft. This indicates that the transient model is sensitive to variations in storage coefficient within the range of estimation reasonably possible with the data available. Further, it seems that the calibrated storage coefficient cannot be increased without introducing fairly large errors in the estimation of head.

A relatively large range of values for transmissivity and leakance are satisfactory under steady-state conditions. However, the transient sensitivity analysis indicates that this large range of values is not acceptable in the transient simulation.

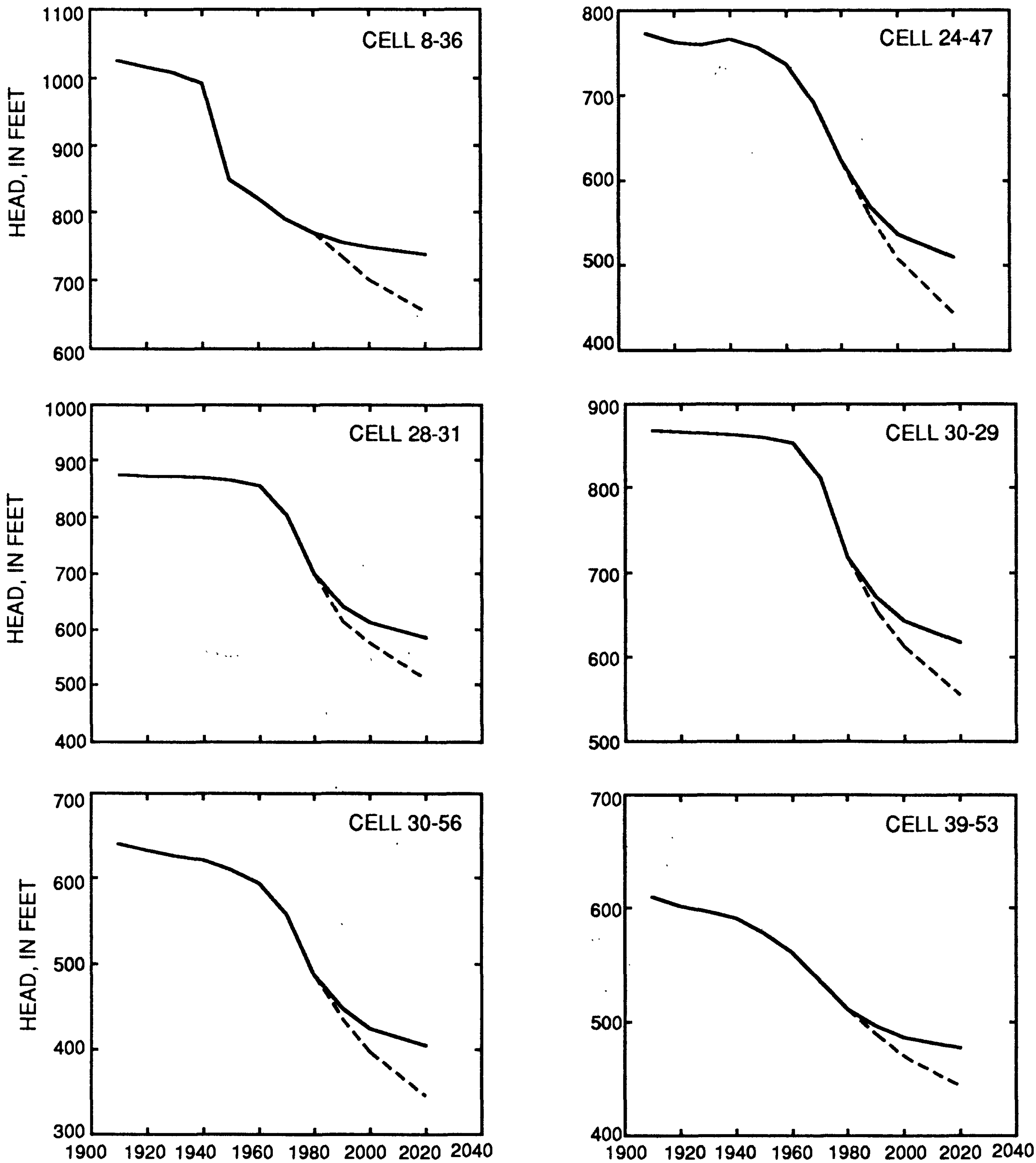
Estimates of Effects of Future Pumping

The objective in estimating future pumping effects is to determine, on a regional scale, the probable magnitude and timing of further drawdown and changes in rates and sources of recharge and discharge. The transient model was used to estimate these factors and to make some estimates of drawdown, particularly near the nodes used for calibration. Predictions of drawdown at a particular point may be inaccurate, however, because of incorrect assumptions about the location and rate of future pumping, errors in the estimates of hydrologic parameter values, proximity to model boundaries, location of nodes with respect to actual wells, or other factors associated with the assumptions necessary to develop the model.

Two sets of future pumping rates were simulated. The first uses the average pumping rate from 1975 to 1980, the most recent and complete pumping data, at each node as the constant rate for the period 1981 to 2020. The second begins with the 1975 to 1980 pumping rates and increases these rates by 10 percent for each successive pumping period. The pumping periods simulated are 1981 to 1990,1991 to 2000, and 2001 to 2020. Other hydraulic factors are identical to those used in the calibrated transient model. The hydrographs that represent estimated future changes in water levels at calibration nodes (fig. 18) are drawn as continuations of the simulated water levels calculated by the transient model.

Simulation results show that continued pumping, at the average pumping rate determined from the 1975 to 1980 period, will produce decreasing rates of water-level decline through the year 2020. The average annual rate of drawdown in the 6 nodes selected for analysis will be about 3.8 ft from 1980 to 1990, 2.2 ft from 1991 to 2000, and 1.0 foot between 2001 and 2020. Because the confining unit above the St. Peter-Jordan aquifer was not simulated as an aquifer layer, storage in the confining unit could not be considered a source of water. The actual drawdown may be smaller if water were available from storage in the confining unit.

Based on the calibrated transient model, continued pumping at the average rates of 1975-1980 will produce additional head declines from the 1980 water level of more than 57 ft by 1990 (node 28-31) and an average of 38 ft in all nodes evaluated. By 2020 the maximum decline from



EXPLANATION

—— Pumping at 1975-1980 rates until 2020

– – – Pumping at 1975-1980 rates increased by 10 percent during each simulation period--1981-1990, 1991-2000, and 2001-2020

Figure 18.--Predicted head in selected model cells.

1980 is predicted to be more than 115 ft (node 24-47) with an average of about 79 ft. The average difference in head from 2001 to 2020 is about 20 ft, or one foot per year, which is about one-fourth of the rate of head change predicted from 1981 to 1990.

The second simulation of future effects on heads is based on a conservative estimate of increases in pumping rates of 10 percent in each of the three successive pumping periods. Historical increases in pumpage have been much larger than 10 percent per decade, but the rate of increase in pumpage has declined in recent years. The pumping rates shown in figure 11 include a 95-percent increase from the 1950's to the 1960's and a 31-percent increase from the 1960's to the 1970's. If the rate of increase of pumping continues to decline, a 10-percent increase in total pumpage is a reasonable estimate. The10-percent increase in pumping rate is within the 5 mi^2 cell and is not necessarily limited to uniform increases in existing wells.

The simulations indicate that if the pumping rate increases by 10 percent for each pumping period, the heads in the nodes used for calibration will continue to decline at an almost constant rate throughout the period of simulation. By 1990 the maximum head change from the 1980 water level will be more than 83 ft (node 28-31) and the average head change for the calibrated nodes will be about 53 ft. The maximum change from 1981 to 2020 is predicted to be 188 ft (node 28-31) and the average to be about 143 ft for calibrated nodes.The average annual head decline is predicted to decrease over the next 40 years, but that would not reach a new equilibrium condition. The average annual change in head is predicted to be 5.3 ft in the 1980's, 3.8 ft in the 1990's, and 2.6 ft from 2001 to 2020.

The estimates of future changes in head in the St. Peter-Jordan aquifer under the two sets of pumping conditions indicate that significant declines in head may occur with increases in pumping rates that are much less than historical trends. Even with no increase in pumping rate, head declines will continue for at least 20 years.

CONCLUSIONS ABOUT REGIONAL GROUND-WATER FLOW BASED ON SIMULATION

Analysis of the St. Peter-Jordan aquifer (Cambrian Jordan Sandstone and overlying Prairie du Chien Group and St. Peter Sandstone) using a digital flow model provided several observations and conclusions regarding the nature of flow in the aquifer and possible effects of future pumping.

Simulated values of transmissivity approximately agree with values estimated from pumping-test results as reported by drillers and contractors. The local variability in specific capacity of wells can be explained by lithologic and structural variations of the rocks comprising the aquifer as well as variations in well treatments.

The largest transmissivity values that satisfy the calibration criteria occur where the aquifer is predominantly dolomite and dolomite-cemented sandstone. Therefore, the permeability of the aquifer is due more to fractures and solution openings in the dolomite than to interstitial sandstone permeability, and the regional distribution of transmissivity cannot be reliably estimated by using only the sandstone thickness. These fractures and solution openings complicate attempts to interpret local flows because variations in transmissivity and anisotropy, even over short distances, cannot be estimated accurately.

In eastern and southeastern Iowa, the aquifer discharges water upward to the aquifers comprised of Silurian and Devonian rocks under simulated predevelopment conditions. This occurs between the Iowa and Cedar Rivers, whose valleys have incised the Silurian and Devonian rocks; thus, apparently influencing discharge from the St. Peter-Jordan aquifer.

Vertical leakage from overlying rocks is the largest source of recharge to the aquifer in the predevelopment simulation. Recharge and discharge components of flow in the St. Peter-Jordan simulated by the model are shown in figure 19. Simulation of predevelopment conditions

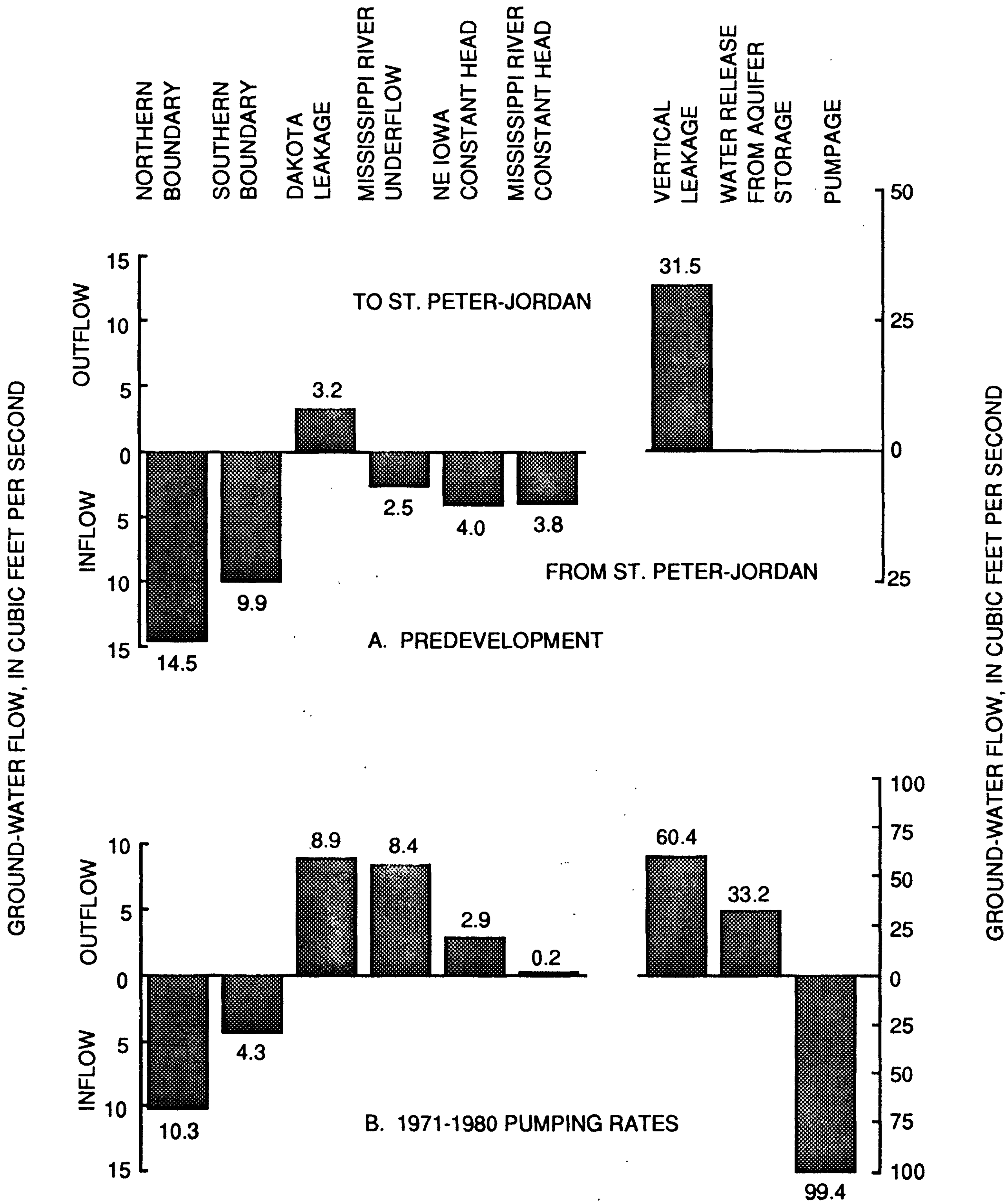


Figure 19.--Simulated components and rates of flow for the St. Peter-Jordan aquifer.

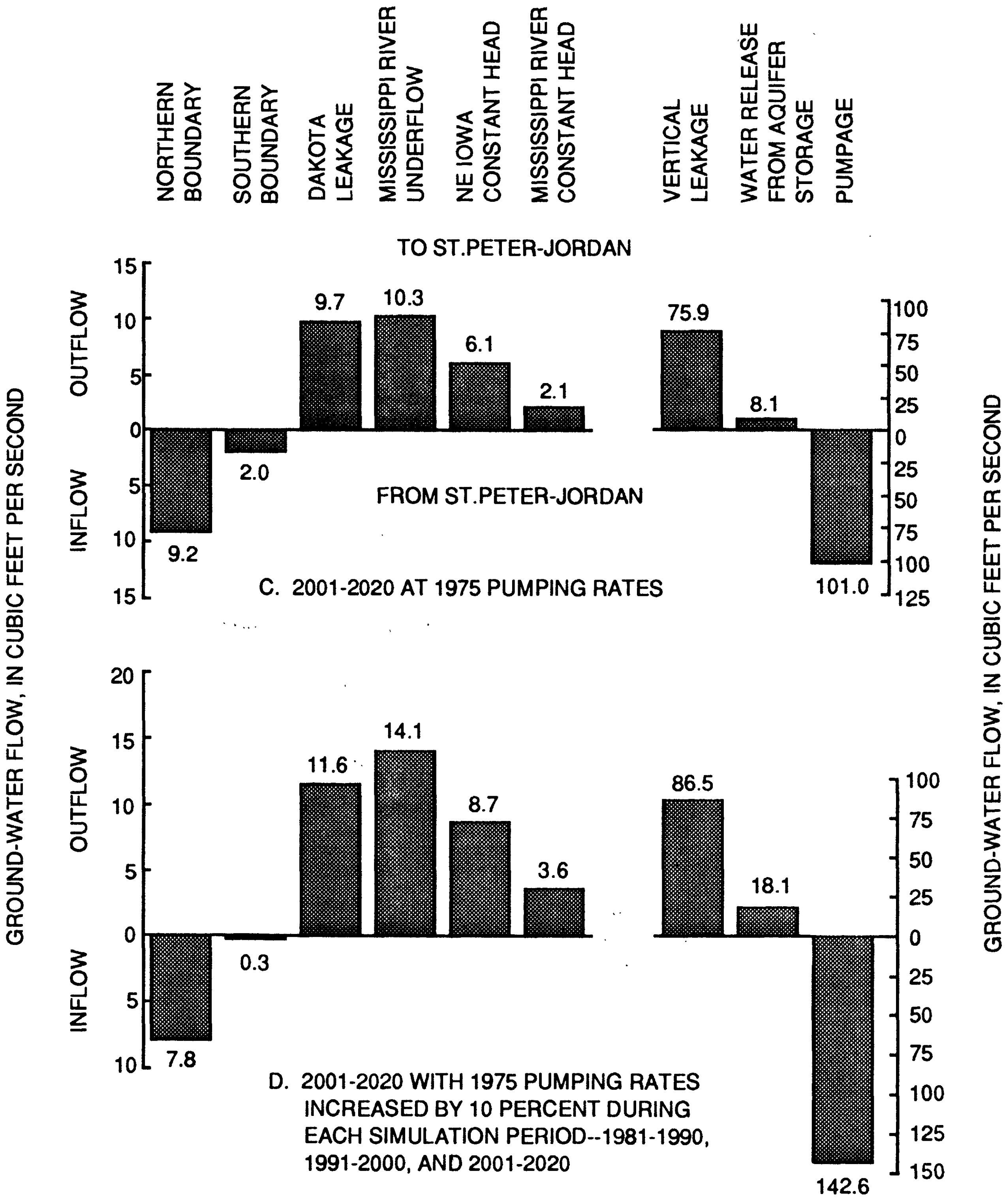


Figure 19.--Simulated components and rates of flow for the St. Peter-Jordan aquifer--Continued.

indicates that vertical leakage into the aquifer, at a rate of 31.5 ft^3/s (cubic feet per second), constituted more than 90 percent of the total recharge. The other significant source of recharge was downward leakage from the Dakota aquifer in northwestern Iowa. Lateral boundary flux provided discharge from the aquifer beneath the Mississippi River and from the outcrop area in northeastern Iowa.

The hypothesis suggested by Steinhilber and Horick (1978) that a large part of the recharge to the aquifer occurred within the outcrop areas in northeastern Iowa and from Minnesota, to the north, is not supported by the simulation. The northern boundary had a net predevelopment discharge rate of 14.5 ft^3/s and the area of constant-head cells in northeastern Iowa had a net discharge of 4.0 ft^3/s. As discussed previously, this latter area parallels the streams. The simulated rate of discharge from the aquifer is very small when compared to the discharge in any single stream in the area. Another area with a large rate of discharge, 9.9 ft^3/s, is along the southern (Missouri) boundary. Discharge from the aquifer in contact with the Mississippi River was at a rate of 3.8 ft^3/s. No independent analysis of flux into Missouri is available to confirm the simulated discharge rate. Discharge to the Mississippi River is extremely small in relation to the flow in the river and therefore cannot be confirmed or denied by flow measurements.

The large proportion of total simulated recharge derived from leakage is supported by water-quality data. Dissolved-solids concentrations increase toward the southwest (Horick and Steinhilber, 1978 and Siegel and Mandle,1984), yet the flow direction in the aquifer is toward the southeast. If, as can be inferred from the simulation, vertical leakage is a significant source of recharge, the water quality in the aquifer may be influenced more by the rock composition through which leakage occurs rather than the materials comprising the aquifer. Sulfate is the dominant ion dissolved in the water of the aquifer beneath that part of Iowa where Mississippian age rocks occur (fig. 1). The Mississippian rocks contain sources of sulfate in the form of minerals such as gypsum and anhydrite (Horick and Steinhilber, 1973). The sulfate concentration in the Silurian-Devonian aquifer (layer 1 in this simulation) immediately beneath the Mississippian rocks generally exceeds 1,000 mg/L (milligrams per liter) (Horick, 1984). The zone of increase from 100 to 1,000 mg/L in the Silurian-Devonian aquifer is only a few mi wide and coincides with the northeastern limit of Mississippian rocks. An anomalously fresh plume of water in the St. Peter-Jordan aquifer extending south beneath the Mississippian rocks has been recently explained by Siegel and Mandle (1984). These researchers used isotopic ratios of hydrogen and oxygen in water to hypothesize that Pleistocene glacial meltwater leaked into the St. Peter-Jordan aquifer beneath the area of north-central Iowa overlain by the most recent glacial deposits. The freshwater plume mapped by Siegel and Mandle conforms to the general shape of the glacial deposits and very likely represents downward leakage into the St. Peter-Jordan aquifer beneath the most recent ice lobe. Because this area is overlain by rocks simulated by layer 1, the likely recharge mechanism was through this layer.

Results of the transient simulation indicate that pumping has lowered the head in the aquifer such that flow across boundaries is changed significantly (fig. 19). Simulated flow through the aquifer more than quadrupled between predevelopment time and 1980. The net predevelopment discharge from the aquifer to the Mississippi River, where they are in direct hydraulic connection, ceased and the river contributed recharge to the aquifer at a rate of 0.2 ft^3/s during the 1970's. Also, net flow across the Minnesota border area was reduced from a discharge rate of 14.5 ft^3/s, under predevelopment conditions, to a discharge rate of 10.3 ft^3/s. Net vertical leakage into the aquifer is still the predominant source of recharge in 1980. The rate of leakage almost doubled from 31.5 ft^3/s, the predevelopment simulation rate, to 60.4 ft^3/s, which represents 53 percent of all recharge. This rate of leakage is very small when compared to the potential rate of replenishment to the source layer through rainfall at a long-term average rate of more than 130,000 ft^3/s. Discharge across the southern boundary is about one-half that of predevelopment conditions and recharge from the Dakota aquifer almost tripled to 8.9 ft^3/s.

Pumping has caused regional declines in the potentiometric surface of the aquifer. Because the only head measurements were from pumping wells, it was not initially clear whether significant declines were localized or whether the head declines also occurred in areas between these wells. However, the transient simulation indicates that head has declined regionally. The simulations imply that these head declines caused changes in flux across boundaries and beneath the Mississippi River between predevelopment and 1980.

The total pumping rate of 99.4 ft^3/s represented about 87 percent of the net discharge from the aquifer in the 1970's (fig. 19). During this period, 33.2 ft^3/s of water was removed from aquifer storage and vertical leakage provided a net recharge of 60.4 ft^3/s. Additional recharge is 20.4 ft^3/s and consists of a combination of: recharge from the Dakota aquifer (8.9 ft^3/s); from the subcrop area in northeastern Iowa (2.9 ft^3/s); direct recharge from the Mississippi River (0.2 ft^3/s); and from its underflow (8.4 ft^3/s). A discharge rate of 4.3 ft^3/s is across the southern boundary and 10.3 ft^3/s across the northern boundary. Therefore, the net boundary flux from the simulation was 5.8 ft^3/s of recharge.

Boundary inflow to the aquifer provides much more water under pumping than under predevelopment conditions. Although pumping produced an increase in the net vertical leakage, other inflow sources increased as well. The total in flow, excluding vertical leakage, increased to more than 18 percent of the total recharge as compared to 9 percent under predevelopment conditions. The net boundary flux is about two-thirds the rate of loss from storage. Because many of these boundaries are directly related to surface or water-table conditions, the water quality in the aquifer near the boundaries may be affected by changes in land or water use in those areas.

Simulation of future pumping at 1975 average rates suggests that the rate at which water is removed from aquifer storage will be significantly reduced by 2020 and the rate of vertical leakage will be increased. Seventy-five percent of the pumpage will be balanced by the vertical leakage to the aquifer, eight percent by water released from aquifer storage and 17 percent by net recharge through boundaries. The rate of storage loss is estimated to be 8.1 ft^3/s (fig. 19), one-fourth the rate of the 1970's and the vertical leakage rate will increase to 75.9 ft^3/s, 25 percent more than the 1970's rate. The net lateral boundary recharge rate will nearly triple that of the 1970's. The increase in boundary recharge is particularly significant in northeastern Iowa and along the Mississippi River, although a net recharge increase also results from decreases in discharge through the north and south boundaries. Recharge from the overlying Dakota aquifer will not increase significantly, possibly because of smaller pumping demand in northwestern Iowa and the smaller values of transmissivity in the western part of the State.

A similar simulation was conducted using 1975 average pumping rates plus an increase of 10 percent in each pumping period. This simulation shows that net recharge through vertical leakage will increase to about 150 percent of the 1970's rate and net recharge from lateral boundaries will increase to about five times that of the 1970's (fig. 19). Discharge through the southern boundary area will be reduced to an insignificant amount and water released from aquifer storage, 18.1 ft^3/s, will be slightly more than half the 1970's rate.

These simulations indicate that vertical leakage will be the dominant source of pumped water. Consequently, more information regarding the overlying and underlying aquifers and their relation to the St. Peter-Jordan aquifer is important to fully understand the sources of recharge. An understanding of the sources of recharge is important, particularly in areas where the underlying aquifers are used or could be the source of water of unsatisfactory quality.

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