EVALUATION OF AVAILABILITY OF WATER FROM DRIFT AQUIFERS NEAR THE POMME DE TERRE AND CHIPPEWA RIVERS, WESTERN MINNESOTA

By G. N. Delin

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CONVERSION FACTORS AND ABBREVIATIONS

For the convenience of readers who may prefer to use metric (International System) units rather than the inch-pound units used in this report, values may be converted by using the following factors:

Multiply Inch-Pound Unit	By	To Obtain Metric Unit
foot (ft)	0.3048	meter (m)
foot per day (ft/d)	0.3048	meter per day (m/day)
foot squared per day (ft^2/d)	0.09290	meter squared per day (m^2/day)
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
gallon (gal)	3.785	liter (L)
gallon per minute (gal/min)	0.06308	liter per second (L/s)
million gallons per year (Mgal/yr)	0.00012	cubic meter per second (m^3/s)
inch (in.)	25.4	millimeter (mm)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.590	square kilometer (km^2)

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ABSTRACT

Ground-water flow in the confined- and unconfined-drift aquifers near Appleton and Benson, Minnesota, was simulated with a three-dimensional finite-difference ground-water-flow model. Model results indicate that 98 percent of the total inflow to the modeled area is from precipitation. Of the total outflow, 38 percent is ground-water discharge to the Pomme de Terre and Chippewa Rivers, 36 percent is evapotranspiration, 17 percent is ground-water pumpage, and 8 percent is ground-water discharge to the Minnesota River.

The model was used to simulate the effects of below-normal precipitation (drought) and hypothetical increases in ground-water development. Model results indicate that reduced recharge and increased pumping during a three-year extended drought probably would lower water levels 2 to 6 feet regionally in the surficial aquifer and in the Appleton and Benson-middle aquifers and as much as 11 feet near aquifer boundaries. Ground-water discharge to the Pomme de Terre and Chippewa Rivers in the modeled area probably would be reduced during the simulated drought by 15.2 and 7.4 cubic feet per second, respectively, compared to 1982 conditions. The addition of 30 hypothetical wells in the Benson-middle aquifer near Benson, pumping a total of 810 million gallons per year, resulted in water-level declines of as much as 1.3 and 2.7 feet in the surficial and Benson-middle aquifers, respectively. The addition of 28 hypothetical wells in the Appleton aquifer east and southeast of Appleton, pumping a total of 756 million gallons per year, lowered water levels in the surficial and Appleton confined aquifers as much as 5 feet.

INTRODUCTION

Ground-water withdrawals from drift aquifers have increased dramatically during the last decade in western Minnesota. The increase is due primarily to an increase in the ground water used for irrigation following the 1976-77 drought. The MDNR (Minnesota Department of Natural Resources) received only 38 applications for permits for irrigation from ground water in Swift County prior to 1976. Conversely, 105 applications were received in 1977 alone and 278 in 1984 alone. Most ground water is obtained from surficial aquifers although an increasing amount has been pumped from confined-drift aquifers. The Appleton and Benson areas of Swift County, Minnesota, obtain water supplies almost entirely from drift aquifers. The MDNR is concerned about the effects of increased withdrawals from the confined drift aquifers because of uncertainty about (1) long-term yields of wells open to these aquifers, (2) effects of pumping and drought on water levels and streamflow, and (3) possible interference between nearby wells pumping from the same aquifer. Consequently, the U.S. Geological Survey, in cooperation with the MDNR and the Pomme de Terre and Chippewa Ground-Water Study Steering Committee, conducted a five-year study

(1979-84) to appraise the ground-water resources along these rivers in Chippewa, Grant, Pope, Stevens, and Swift Counties.

The purpose of this study was to describe the occurrence, availability, and quality of ground water near the Pomme de Terre and Chippewa Rivers. Study objectives were to (1) map the areal extent and thickness of the confined- and unconfined-drift aquifers, (2) determine hydraulic characteristics of the aquifers, (3) estimate the potential yield of each aquifer, (4) describe the water quality of each aquifer, and (5) determine the probable effects of future development on the aquifer system through ground-water-flow simulations.

The study was divided into two phases. The purpose of the first phase was to determine the ground-water resources of the surficial aquifers along the Pomme de Terre and Chippewa Rivers. Results from this phase of the study are described by Soukup and others (1984).

Objectives of the second phase of the study were to appraise the ground-water resources of confined-drift aquifers near the Pomme de Terre and Chippewa Rivers. Results of the second phase of the study are summarized in two U.S. Geological Survey reports: Delin (1984) provides a preliminary description of the confined-drift aquifers and modeling results, and Delin (1986) provides a detailed hydrogeologic description of confined-drift aquifers.

Purpose and Scope

The purpose of this report is to evaluate the availability of water from unconfined- and confined-drift aquifers in the Appleton and Benson areas of Swift County. The evaluation was done by use of a ground-water flow model. The model is a practical tool used to estimate the long-term effects of projected pumping and climatic conditions. This report describes the sources and types of data used in constructing the model, model calibration, hypothetical model simulations, and limitations of the model.

Location and Description of Study Area and Modeled Area

The study area is about 150 miles west of Minneapolis and St. Paul and covers approximately 1,380 mi^2 including parts of Chippewa, Grant, Pope, Stevens, and Swift Counties (fig. 1). The southern part of the study area was modeled (fig. 1), encompassing primarily the area in Swift County near the cities of Appleton and Benson. The modeled area covers approximately 780 mi^2 and includes parts of Swift, Stevens, Pope, and Big Stone Counties. Ground-water supplies have been developed more here than elsewhere in the study area. The sandy soils in the area indicate that there is potential for expanded ground-water development. The model grid was oriented in a northeast-southwest direction because of (1) the areal extent of the confined aquifers, (2) the location of major hydrologic boundaries, and (3) the direction of regional flow in the confined aquifers. The area is drained by the Pomme de Terre and Chippewa Rivers, which are tributaries of the Minnesota River. The topography is generally flat or gently rolling. Mean annual precipitation is about 24 inches (Baker and Kuehnast, 1978), with most of it occurring between May and September. Mean potential evapotranspiration is about 24.5 inches and average annual runoff is about 2 inches (Baker and others, 1979).

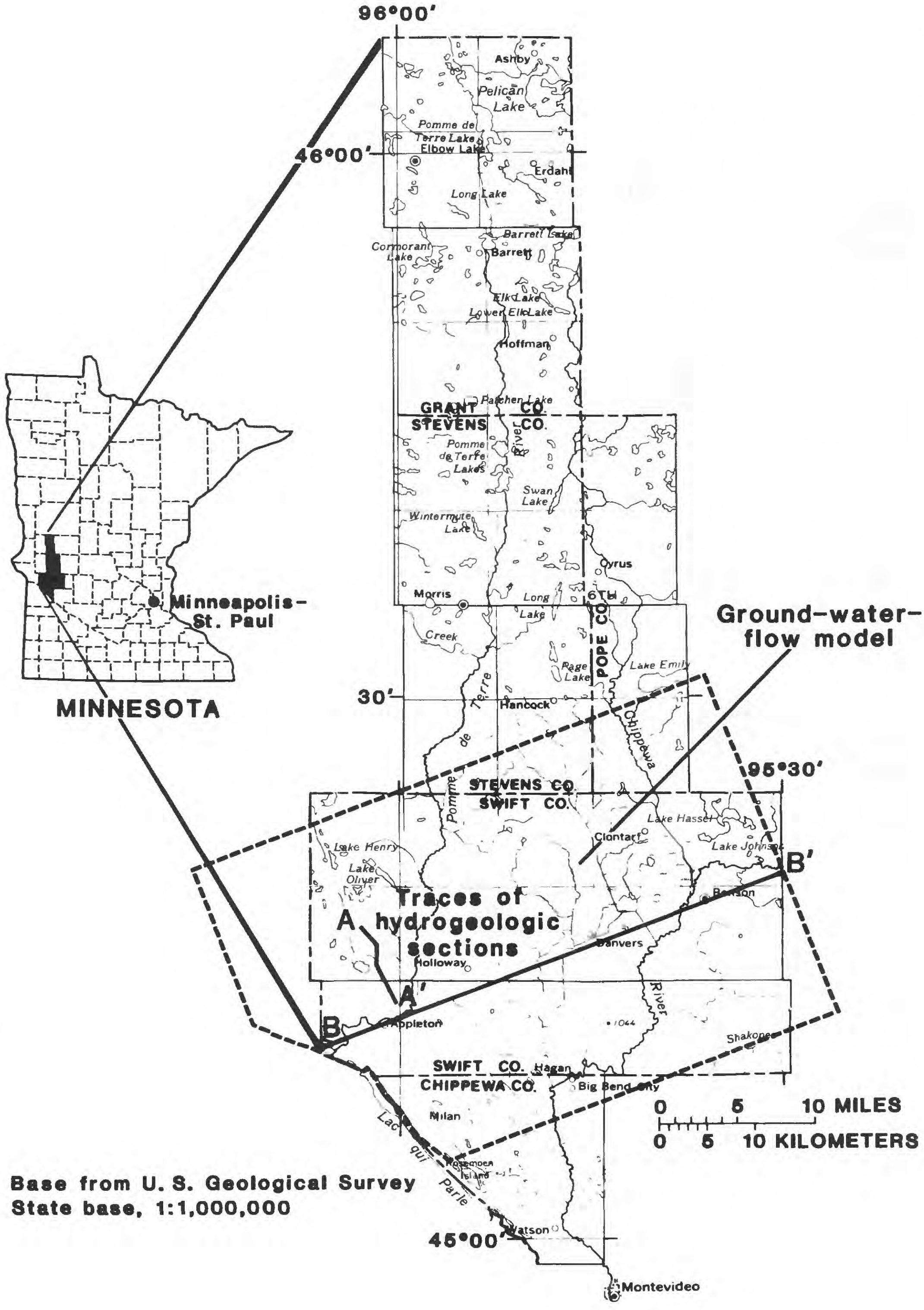


Figure 1.--Location of study area and extent of ground-water-flow model

Previous Investigations

Winchell and Upham (1888) first summarized the geology and natural history of west-central Minnesota. A general description of the glacial geology in the study area is presented by Leverett (1932). The glacial geology was reinterpreted by Wright and Ruhe (1965) and Wright (1972). Outwash deposits in the vicinity of the Pomme de Terre River were described by Sandeson (1919). Glacial Lake Benson and Lake Agassiz outwash deposits are discussed in Matsch and Wright (1967). Hall and others (1911) and Theil (1944) investigated the hydrogeology of southern Minnesota including Swift and Chippewa Counties. Allison (1932) provides a general description of the geology and ground water in Grant, Stevens, and Pope Counties. A general description of ground water in the study area is provided by Lindholm and Norvitch (1976). More detailed hydrologic studies were conducted near Lake Emily by Van Voast (1971) and Wolf (1976). Larson (1976) discussed the ground water available from surficial aquifers near Appleton. Hydrologic reconnaissances of the Pomme de Terre and Chippewa River watersheds were made by Cotter and Bidwell (1966) and Cotter and others (1968), respectively. A preliminary investigation and data summary containing well logs, water levels, and geologic sections for Swift County was completed by Fax and Beissel (1980).

Test-Hole and Well-Numbering System

The system of numbering wells and test holes is based on the U.S. Bureau of Land Management's system of land subdivision (township, range, and section). Figure 2 illustrates the system of numbering data-collection points for location. The first numeral of a location number indicates the township, the second the range, and the third the section in which the point is located. Uppercase letters after the section number indicate the location within the section; the first letter denotes the 160-acre tract, the second the 40-acre tract, and the third the 10-acre tract, and so on. The letters A, B, C, and D are assigned in a counterclockwise direction, beginning in the northeast corner of each tract. The number of uppercase letters indicates the accuracy of the location number; if a point can be located within a 10-acre tract, three uppercase letters are shown in the location number. For example, the number 129.41.15ADC indicates a test hole or well located in the SW1/4, SE1/4, NE1/4, section 15, T.129N., R.41W.

HYDROGEOLOGIC SETTING

Drift Geology

Glacial drift covers the entire modeled area. The drift consists primarily of till and outwash and ranges in thickness from less than 100 ft near the Minnesota River to about 400 ft where it fills bedrock valleys. Drift in the area has been subdivided hydrogeologically into three types (1) sand and gravel deposits at land surface that compose the unconfined (or surficial) aquifers, (2) till that overlies and confines deeper sand and gravel deposits, and (3) deeper sand and gravel deposits that compose the confined aquifers. Delin (1986) provides a detailed description of these deposits; a general description follows.

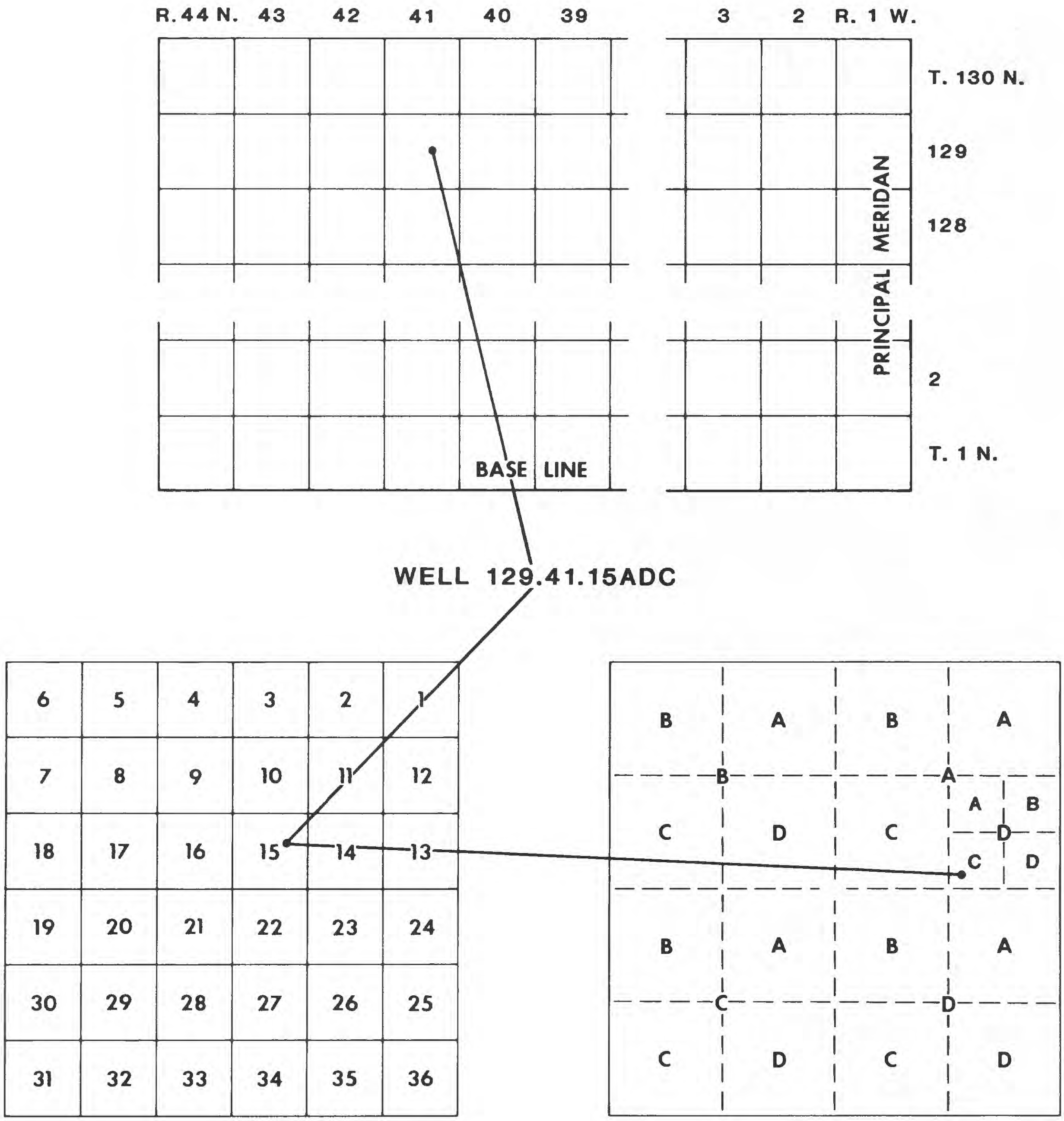


Figure 2.--Test-hole and well-numbering system

The unconfined (surficial) aquifer occupies the saturated zone between the water table and underlying till; it is present in narrow channels along the Pomme de Terre and Chippewa Rivers and as sand plains in the southern part of the study area. The aquifer generally consists of coarse sand and gravel to the north and fine to medium sand to the south that was deposited during the last glacial retreat. Surficial aquifers commonly range in thickness from 10 to 40 ft (Soukup and others, 1984), although the aquifer can be as much as 100 ft thick in the northern part of the Pomme de Terre River valley. The surficial aquifers are described in detail by Soukup and others (1984).

Till is an unsorted mixture of clay, silt, sand, gravel, and boulders. Till confining the deeper aquifers in the drift ranges in thickness from 3 to 170 ft. Hydrogeologic section A-A' north of Appleton (fig. 3) illustrates the relation between the till, several of the prinicipal aquifers, the deposits of Cretaceous age, and bedrock. Hydrogeologic section A-A' also shows that the surficial and confined aquifers coalesce in part of the study area.

Confined-drift aquifers, herein referred to as confined aquifers, consist of saturated sand-and-gravel deposits bounded above and below by lower-permeability till. Using techniques presented by Winter (1975), five confined aquifers have been identified within the modeled area (figs. 1 & 4): (1) the Morris aquifer, present between the Pomme de Terre and Chippewa Rivers north of Holloway and Danvers and west of the Pomme de Terre River north of Appleton; (2) the Benson-upper aquifer, present primarily in the Benson area; (3) the Appleton aquifer, present primarily in the Appleton area; (4) the Benson-middle aquifer, present primarily in the eastern two-thirds of the modeled area; and (5) the Benson-lower aquifer, present in drift-filled bedrock valleys. These aquifers were named and described in detail by Delin (1986).

Bedrock Geology

Igneous and metamorphic rocks of Proterozoic (Late Precambrian) age directly underlie drift throughout most of the model area. The rocks consist primarily of granite, gneiss, and schist. Some outcrops are present along the Minnesota River valley in Chippewa County. Virtually all water is present in fractures and in weathered zones near the top. These rocks generally are dense, having low porosities and permeabilities, and do not yield water to wells in the modeled area.

Deposits of Cretaceous age overlie the Proterozoic rocks in parts of Swift and Chippewa Counties. These discontinuous and generally semiconsolidated shale and sandstone deposits are difficult to differentiate from drift. The maximum known thickness of Cretaceous deposits in the modeled area is 33 ft. Although a few wells are known to yield as much as 50 gal/min from Cretaceous rocks, the deposits are not considered to be a major confined aquifer in the modeled area.

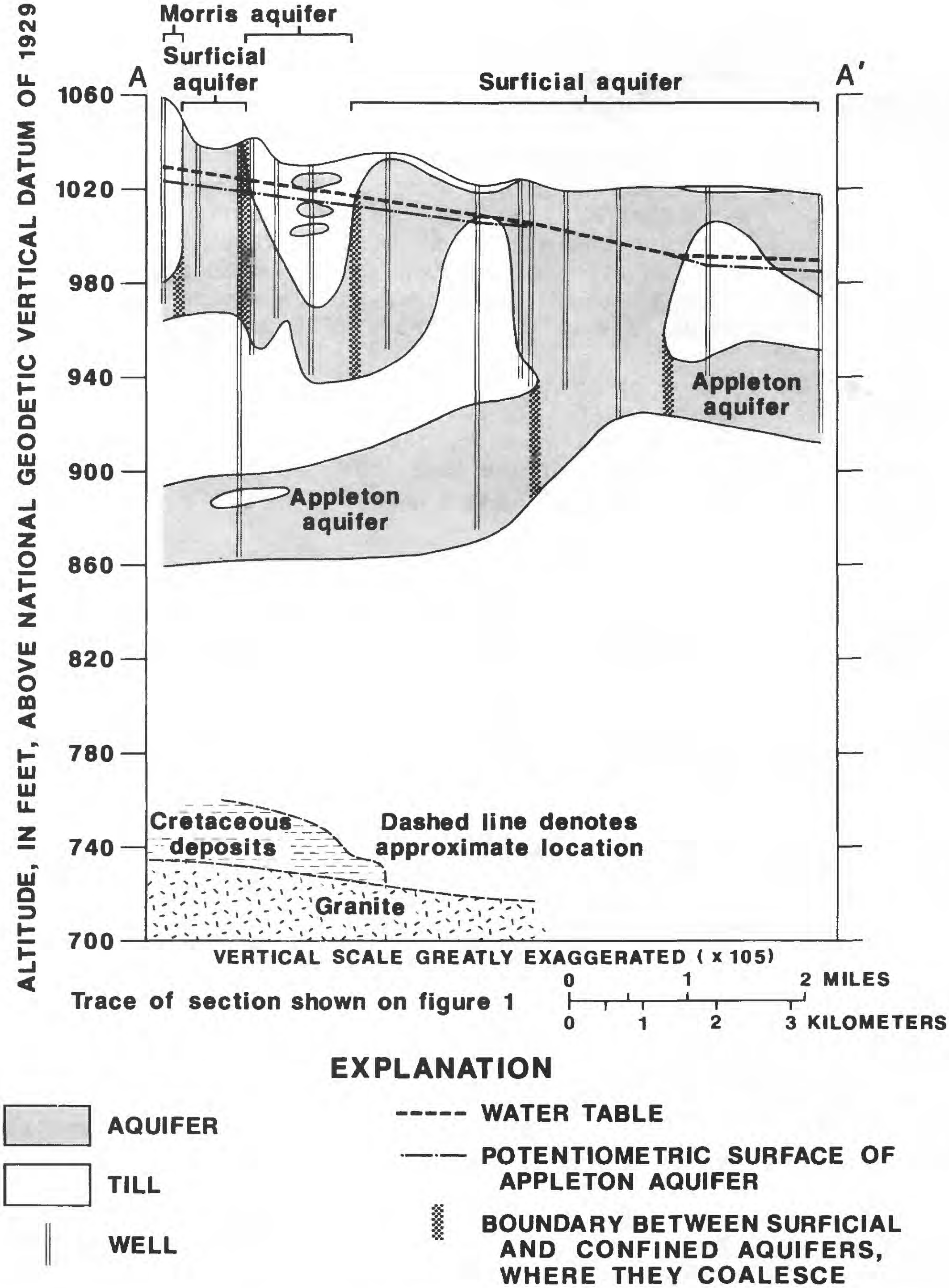


Figure 3.--Hydrogeologic section A-A' north of Appleton showing relationship between till and confined and unconfined aquifers

Hydrology

Aquifers in the modeled area generally are recharged near topographic highs and discharged near topographic lows, such as the Minnesota, Pomme de Terre, and Chippewa Rivers. Flow in aquifers is predominantly horizontal whereas flow in confining beds is predominantly vertical.

Where confined and surficial aquifers coalesce, ground water can flow directly from one aquifer to the other in response to natural or man-made stresses. Ground water generally flows under a natural head gradient from confined aquifers into the surficial aquifer in such areas. The surficial aquifer coalesces with (1) the Appleton aquifer near Appleton, (2) the Morris aquifer north of Appleton, (3) the Benson-middle aquifer northeast of Appleton, and (4) the Benson-upper aquifer northwest of Benson.

The head in each of the confined aquifers in the modeled area generally decreases with depth, indicating downward flow. Near the Minnesota, Pomme de Terre, and Chippewa Rivers, however, the head increases with depth and flow is upward. Water-level data indicate that heads in the Benson-middle aquifer, for example, are 1 to 5 ft higher than heads in the surficial aquifer near these rivers.

The major source of areal recharge to the ground-water system is precipitation. Recharge is greater in areas where the surficial aquifer is present. Areal recharge commonly is greatest in spring, due to snowmelt, spring rain, and little evapotranspiration, which results in rising ground-water levels. Conversely, ground-water levels generally decline in summer because most precipitation is lost by evaporation or is transpired by plants. Areal recharge sometimes occurs in the fall, depending on rainfall, runoff, and evapotranspiration conditions.

Ground water moves into and out of the study area primarily where the drift aquifers extend beyond the modeled area boundaries. The directions of ground-water flow in the modeled area are generally parallel to its boundaries, however. Therefore, natural ground-water flux across the boundaries is negligible, considering the total amount of water in the ground-water system. Flow to or from areas outside the study area could be significant locally, however.

Water levels fluctuate in response to seasonal variations in recharge to and discharge from the ground-water system. Variations in ground-water pumping, evapotranspiration, soil moisture, vegetation type, precipitation, and runoff are the major factors affecting water-level fluctuations.

Water levels in wells completed in the surficial aquifer generally flucuate 2 to 3 ft annually, even within approximately one mile of a high-capacity pumping well. Water levels in wells completed in confined aquifers generally fluctuate 5 to 10 ft annually near high-capacity pumping wells (Delin, 1986).

Water levels in confined and surficial aquifers generally recover to pre-pumping levels following each irrigation season. The net change in water level from 1980-82 in 12 observation wells completed in confined and surficial aquifers ranged from about -2.0 ft to +1.1 ft. These data suggest that, although ground-water levels fluctuate in response to seasonal variations in recharge and discharge, the ground-water system is in dynamic equilibrium. Hydrology of the drift system is described in greater detail by Delin (1986).

EVALUATION OF AVAILABILITY OF WATER FROM DRIFT AQUIFERS

One of the primary objectives of this study was to evaluate the availability of water from the unconfined- and confined-drift aquifers. The ground-water system is too complex to be analyzed by analytical methods alone. Consequently, a ground-water-flow model was constructed as a tool to help evaluate the potential of these aquifers and estimate the long-term effects of projected pumping and climatic conditions.

Model Description

Model objectives were to determine (1) the vertical head gradient between the drift aquifers and (2) the probable effects of future ground-water development and drought on ground-water levels and on streamflow.

The computer code of McDonald and Harbaugh (1984) was used to simulate the ground-water-flow system in three dimensions. The model program uses finite-difference methods to obtain a solution to the partial-differential equation of ground-water flow in three dimensions as given below:

$$\frac{\partial}{\partial x} K_{xx} \frac{\partial h}{\partial x} + \frac{\partial}{\partial y} K_{yy} \frac{\partial h}{\partial y} + \frac{\partial}{\partial z} K_{zz} \frac{\partial h}{\partial z} - W = S_s \frac{\partial h}{\partial t}$$

where,

x, y, and z are cartesian coordinates aligned along major axes of hydraulic conductivity K_{xx}, K_{yy}, K_{zz}, (L/T);

h is the head (L);

W is a volumetric flux per unit volume and represents sources and/or sinks of water (t^{-1});

S_s is the specific storage of the porous material (L^{-1}); and

t is time (t).

Model Assumptions

A conceptual model of ground-water flow in the aquifer system was developed prior to constructing the digital model. The conceptual model consists of simplifying assumptions for the geometry and hydrologic properties used to simulate ground-water flow with the model. These assumptions are necessary because the actual ground-water system is too complex to simulate in detail. Major simplifying assumptions of the conceptual model are as follows:

1. Ground-waterflow in the drift aquifers is primarily horizontal and flow in the confining units separating them is primarily vertical; based on available water-level data;

2. The aquifers and confing units are continuous, homogeneous, and isotropics;

3. The ratio of vertical to horizontal conductivity in both the aquifers and confining units is 1:1;

4. The stage of the Minnesota River does not fluctutate significantly in time and, therefore, may be simulated as a constant-head boundary;

5. Because accurate field data are lacking, streambeds are assumed to be 1 ft thick and composed of permeable material of lower hydraulic conductivity than the aquifers:

6. Minor streams and ditches are insignificant discharge points for the ground-water system and may be ignored:

7. Areal recharge to the water table is from precipitation and occurs primarily from April through June and secondarily from October through December;

8. Where till is present at land surface, vertical leakage through the till is constant and does not flucutate seasonally:

9. The rate of evaportranspiration declines linearly to zero at a depth of 5 ft below land surface; and

10. Ground water withdrawn for irrigation is consumed by evaportranspiration and, therefore, return flow to the aquifer system is neglible.

Layering Scheme and Finite-Difference Grid

The drift was divided into three layers (table 1 and fig. 4): layer one (the top layer) represents the surficial and Morris aquifers; layer two represents the Benson-middle aquifer; and layer three represents the Appleton aquifer. Vertical leakage through till also was simulated by the model.

Table 1.--Relationship between model layers and hydrogeologic units

Model layer	Vertical-leakage between model layers	Hydrogeologic unit
1	---	Surficial aquifer, Morris aquifer, and till (where aquifer is absent)
	1-2	Till confining unit
2	---	Benson-middle aquifer and till (where aquifer is absent)
	2-3	Till confining unit
3	---	Appleton aquifer and till (where aquifer is absent)

The modeled area was subdivided into discrete blocks by a variable grid of 57 rows and 53 columns (fig. 5). The center of each block is referred to as a node. Grid spacings range from 0.25 to 1.25 mi and grid-block areas range from 0.125 to 0.94 mi^2. Smaller blocks were used in the highly irrigated areas near Appleton and Benson and along the Pomme de Terre and Chippewa Rivers.

Simulation of Aquifers

Transmissivity of the confined aquifers in each layer was assigned by overlaying the model grid on the transmissivity map for that aquifer (Delin, 1986, pl. 3) and averaging the transmissivity within each grid block. Transmissivity of the surficial aquifer is calculated by the model as the product of hydraulic conductivity and saturated thickness. Where till is present in each of the layers, as illustrated in figure 4B, a transmissivity of 1 ft^2/d was assigned. This transmissivity was based on a horizontal hydraulic conductivity of till equal to 0.1 ft/d multiplied by a 10-ft thickness of till. The 0.1 ft/d horizontal hydraulic-conductivity value is assumed to represent the average conductivity of till in the study area and is within the range of values for till given by Todd (1959, p. 53) and Freeze and Cherry (1979, p. 29). Where the surficial and Appleton aquifers coalesce (figs. 5 & 6), a transmissivity of 260 ft^2/d was assigned in layer 2. This represents the the transmissivity of a 1-ft thickness of the surficial aquifer.

The Benson-upper aquifer was simulted as a separate layer during the initial stages of model calibration. The aquifer was not simulated as a separate layer in the final model, however, because (1) significant ground-water withdrawals from the aquifer do not occur, and (2) inclusion of the layer was not considered to be essential for accurate simulation of the ground-water system or to meet project objectives. The initial simulations indicated that water-level declines for the aquifer, in response to pumping from the overlying

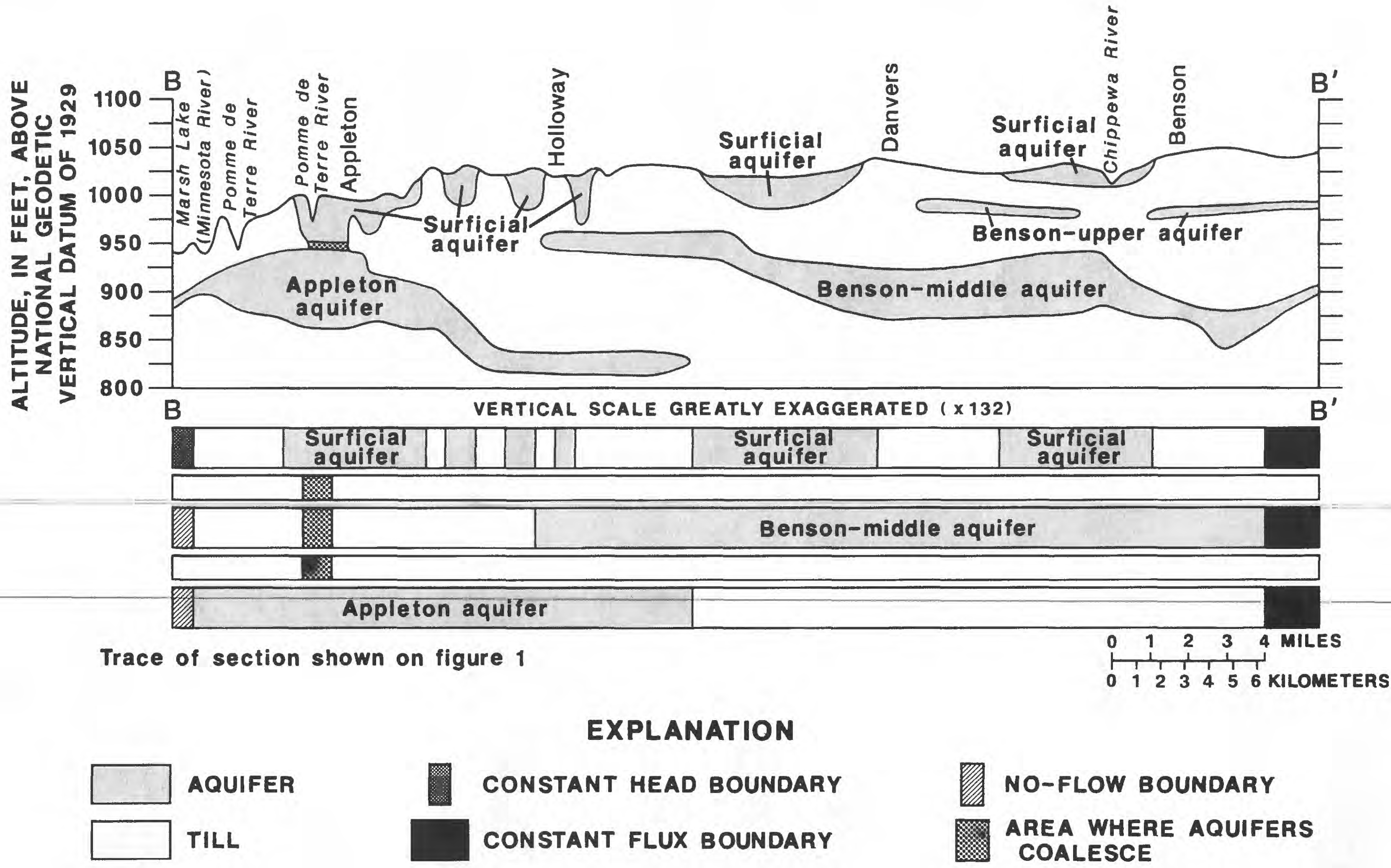


Figure 4.--Hydrogeologic section B-B' showing drift lithology and representative layering scheme for the model

surfical aquifer or from the underlying Benson-middle aquifer, are in the range of model-computed water-level declines for the overlying and underlying aquifers. An approximation of water-level declines in the Benson-upper confined aquifer can be calculated for the model results shown in later sections of this report, however. To obtain this approximate water-level decline at a given point, the reader should add the model-computed water-level declines in the surfical aquifer and the Benson-middle aquifer and divide by two.

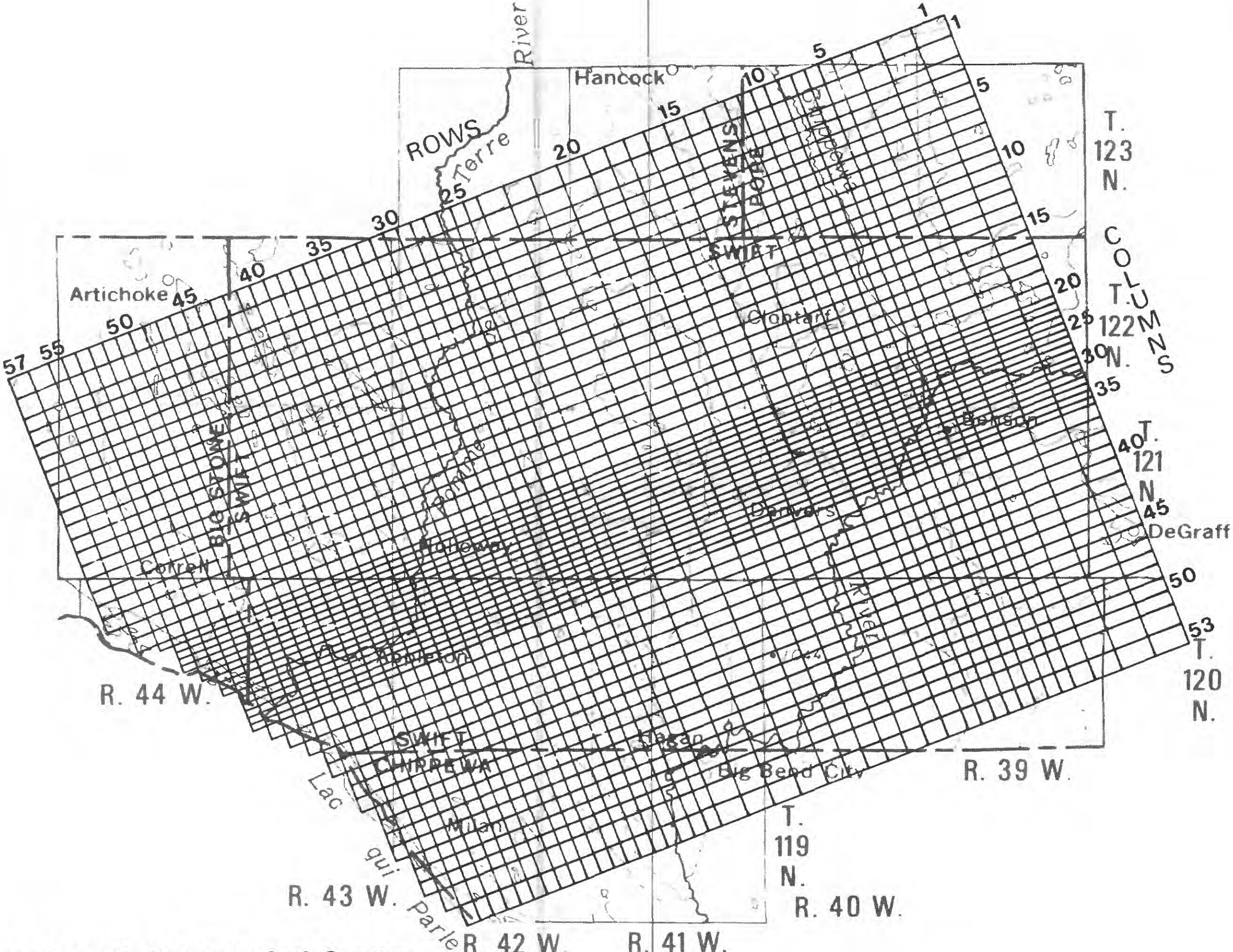
The Benson-upper aquifer extends beyond the model boundaries and, therefore, a finite amount of water flows into or out of the modeled area through the aquifer. In addition, the Benson-upper aquifer coalesces with the surficial aquifer north of Benson (fig. 6), allowing ground water to flow between them. To account for the flux of water into the model and between the aquifers, head-dependent flux-boundary nodes were assigned where the aquifers coalesce (fig. 6). The head-dependent-flux nodes allow leakage into or out of the modeled area based on the hydraulic conductivity and head in the Benson-upper aquifer of 260 ft/d and 1,052 ft, respectively.

The Benson-lower aquifer also was not simulated. Although little hydrogeologic data exist for this aquifer, available data suggest that ground-water flow between the Benson-lower aquifer and overlying aquifers is minimal. Therefore, a no-flow boundary at the base of the Appleton aquifer was deemed justifiable.

Simulation of Confining Units

Vertical flow in the ground-water system was simulated by allowing leakage through till between model layers (table 1 and fig. 4). A vertical-leakage coefficient was used to control the rate of flow between the aquifers. A coefficient was calculated for each model node by first estimating the till thickness between model layers. The till thickness was then divided into the mean vertical hydraulic conductivity for till in the study area, which is 0.025 ft/d, (Delin, 1986) to obtain the vertical-leakage coefficient. Although thin discontinuous sand units occur in the till, their presence does not affect calculation of the vertical leakage coefficient. In areas where the surficial and confined aquifers coalesce (fig. 6) a vertical hydraulic conductivity of 260 ft/d was used in computing vertical leakage. This value represents the approximate vertical hydraulic conductivity for the surficial aquifer (Soukup and others, 1984). Vertical leakage through till to the uppermost confined aquifer also was simulated in the model where the surficial aquifers are absent. In these areas, recharge rates to the water table were adjusted to reflect net leakage to the underlying confined aquifer.

Till confining units were not simulated as separate layers in the model because vertical flow is dominant in these units. Horizontal flow in till was simulated within each model layer, however, where drift aquifers are absent. Horizontal flow in till overlying the Morris confined aquifer was not simulated, however. Only vertical leakage to the Morris confined aquifer was simulated. The simulation of horizontal flow through till was deemed necessary largely due to the occurrence of seepage faces along the Minnesota River. These seepage faces indicate that ground-water flow through till to the river



Base from U. S. Geological Survey
State base map, 1:500,000, 1965

Figure 5.--Finite-difference grid for the ground-water-flow model

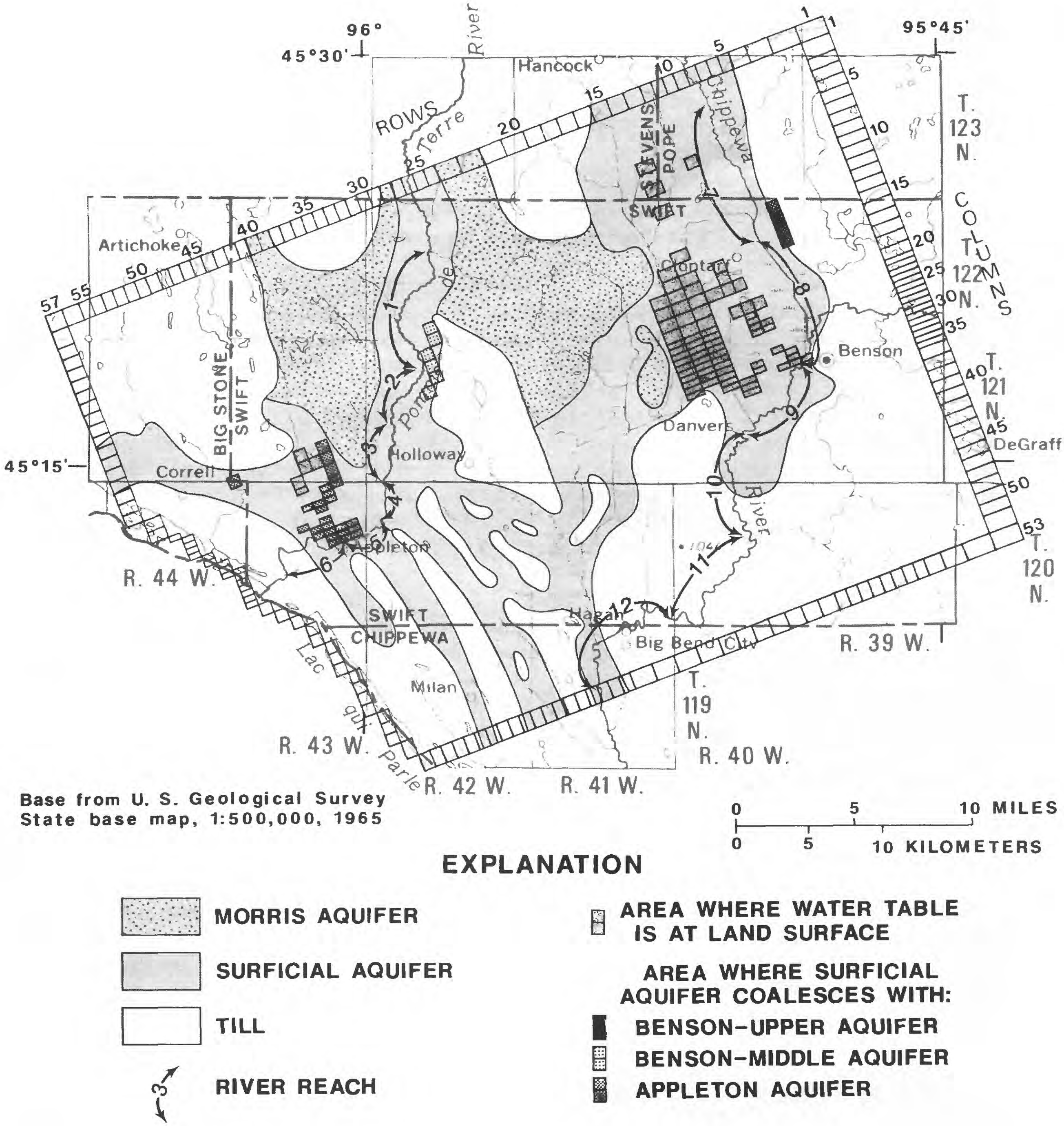


Figure 6.--Areas of surficial aquifer where water table is at land surface, river reaches for which ground-water discharge rates were calculated, and areas where surficial and confined aquifers coalesce

is significant. Minnesota River base-flow measurements (Lindskov, 1977) also indicate that as much as 7 ft^3/s may discharge to the Minnesota River valley from the ground-water system in the modeled area. Delin (1986) indicates that till in the model area is sandy and capable there of transmitting more water than in other parts of the State.

Simulation of Streams, Ground-Water Withdrawals, Areal Recharge, and Evapotranspiration

Ground-water discharge to the Pomme de Terre and Chippewa Rivers is significant. These rivers were simulated as head-dependent-flux nodes in layer 1 of the model (fig. 7). These nodes allow leakage between the aquifer and river nodes based on a specified head in the river and a streambed leakage coefficient. A streambed leakage coefficient was calculated for each river node and is equal to the vertical hydraulic conductivity of the streambed (K') divided by its thickness (m) multiplied by the streambed area in that node. An initial value of 0.1 (ft/d)/ft was used for the value of K'/m. This value is similar to that used in previous investigations by Larson (1976), Lindholm (1980), and Soukup and others (1984). Simulation of the Minnesota River is discussed in the model boundary section.

Because ground-water withdrawals for irrigation and municipal use are a significant part of the total water budget, they were included in the steady-state simulation. Ground-water pumpage was based on MDNR records of reported pumpage from high-capacity wells for 1982. Additional pumpage not reported to the MDNR was estimated using a technique developed by Horn (1984). Total pumpage of 4,954 Mgal/yr was divided among 206 pumping centers in the three model layers. Locations of the pumping centers simulated during steady-state simulations are shown in figures 7 through 9.

Areal recharge, as used in the model, represents the rate of water reaching the water table. The value of areal recharge for areas where the surficial aquifer is present was calculated by analysis of hydrographs using a standard method of hydrograph analysis described by Rassmussen and Andreason (1959). Application of this method for this study is described by Delin (1986). Areal recharge rates were calculated in areas where the water table is generally greater than 5 ft below land surface. Hydrograph analyses indicate that recharge to the surficial aquifer ranges between approximately 3.5 and 11 in/yr and averages about 6 in/yr.

Areal recharge to the water table also was applied in areas where till comprises the uppermost unit. Recharge rates in these areas can be estimated, as leakage to a confined aquifer, using the following form of Darcy's Law:

$$Q_c = \frac{K'}{m'} \Delta h \, A_c$$

where:

Q_c = leakage through confining bed to confined aquifers, in ft^3/d;

K' = vertical hydraulic conductivity of confining bed, in ft/d;

m' = confining bed thickness, in feet;

Δh = difference between head in confined aquifer and in source bed above confining bed, through which leakage occurs, in feet; and

A_c = area of confining bed through which leakage occurs, in ft^2.

Leakage rates to confined aquifers of 0.4 to 3.4 in/yr were calculated for five locations in the modeled area. An areal recharge rate of 1.0 in/yr was applied initially in areas where till is the uppemost unit.

Evapotranspiration (ET) of ground water is an important part of the water budget in areas where the water table is at or near land surface. ET occurs where the water table is within the root zone of vegetation. Based on values of root-zone depth of Thornthwaite and Mather (1957), a depth of 5 ft is considered applicable for the modeled area. The water table is within 5 ft of land surface over a large part of the surficial aquifer in the modeled area, particularly near the city of Benson. A maximum rate of 9 in/yr was specified where the surficial aquifer is present and when the water table is at land surface. The maximum ET rate was calculated using the following formula:

$$ET(max) = ET(pot) - (Precip. - Rech. - RO)$$

where:

ET(max) = maximum ET rate used in the model;

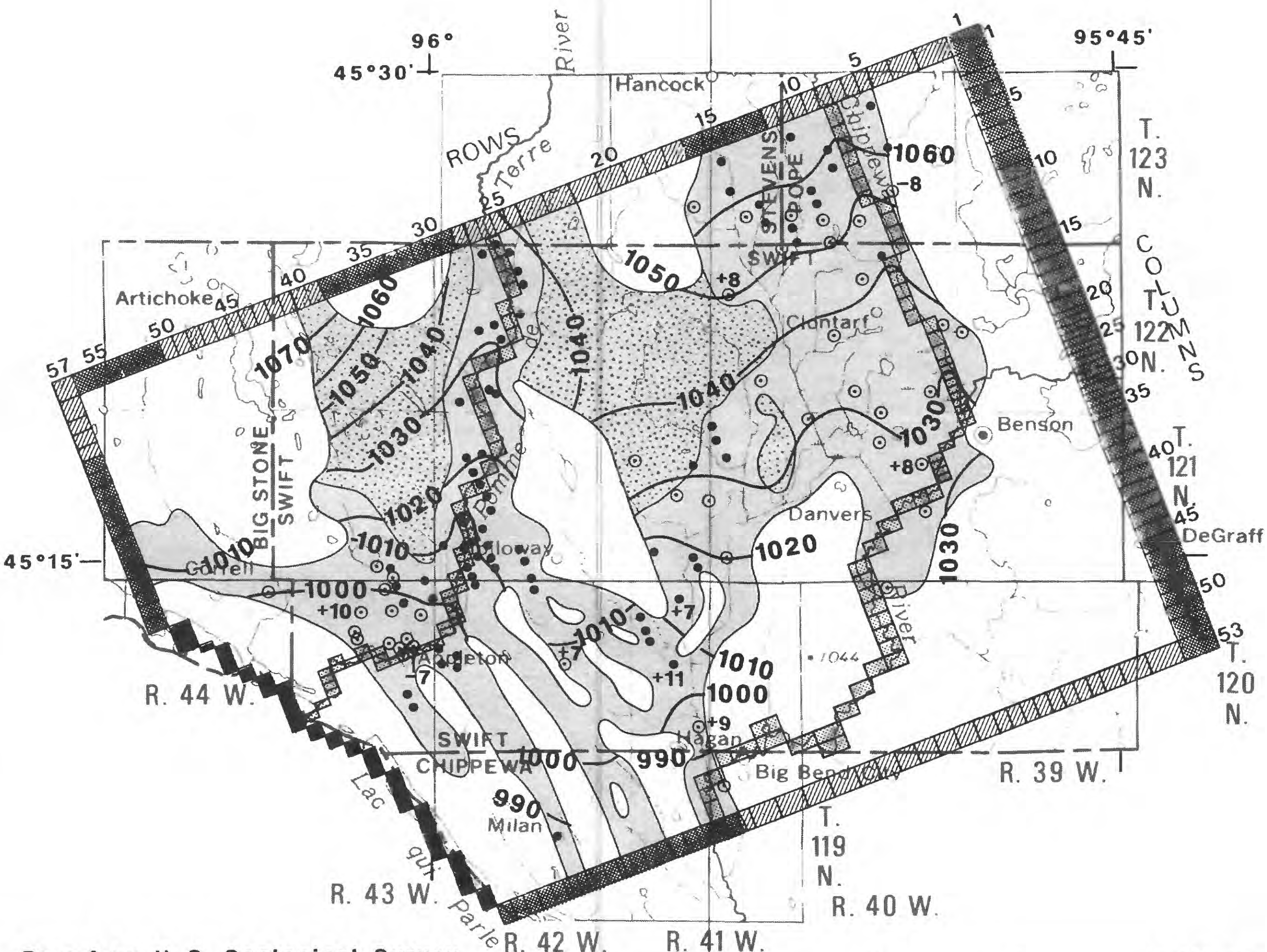
ET(pot) = mean potential ET rate in the model area of 25 in/yr, calculated by the Thornthwaite and Mather (1957) method (Baker and others, 1979);

Precip = mean annual precipitation in the model area of approximately 24 in/yr (Baker and others, 1979);

Rech = Average annual recharge to the ground-water system of 6 in/yr, calculated by hydrograph analysis; and

RO = average annual runoff in the model area of 2 in/yr (Baker and others, 1979).

The quantity (Precip - Rech - RO) is equivalent to that part of the mean annual precipitation subject to ET after losses to recharge and runoff.



Base from U. S. Geological Survey
State base map, 1:500,000, 1965

0 5 10 MILES
0 5 10 KILOMETERS

EXPLANATION

MORRIS AQUIFER

SURFICIAL AQUIFER

TILL

—1010— MODEL GENERATED POTENTIOMETRIC CONTOUR--Interval 10 feet. Datum is National Geodetic Vertical Datum of 1929

RIVER NODE

CONSTANT HEAD BOUNDARY

CONSTANT FLUX BOUNDARY

NO-FLOW BOUNDARY

DATA POINT--Number denotes deviation from measured water level if greater than +6 or less than -6 feet:

+11 1982 PUMPING WELL

o-8 DOMESTIC OR OBSERVATION WELL

Figure 7.--Boundary conditions, river nodes, and difference between model-computed and measured water levels in the surficial and Morris aquifers for the steady-state simulation

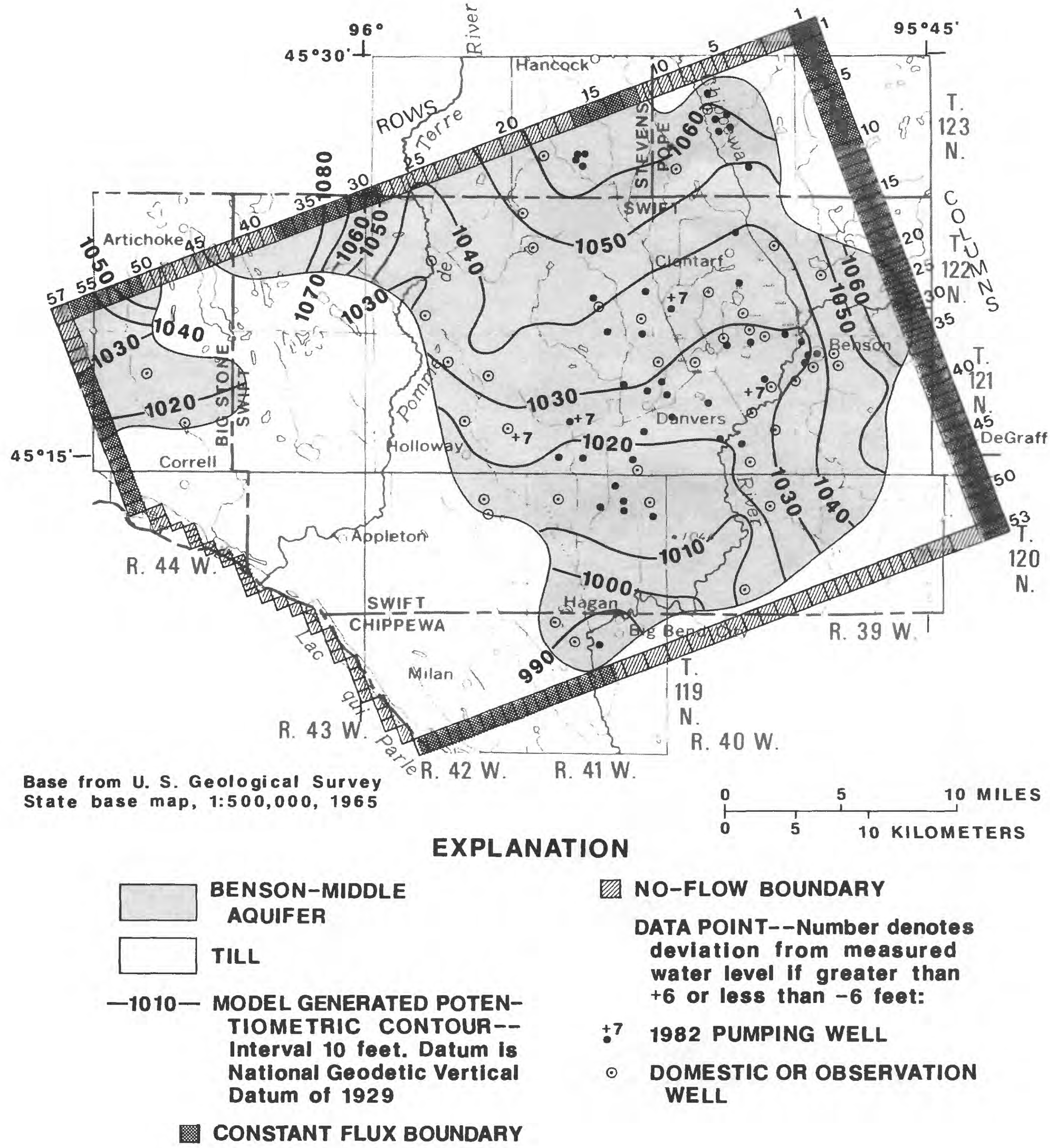
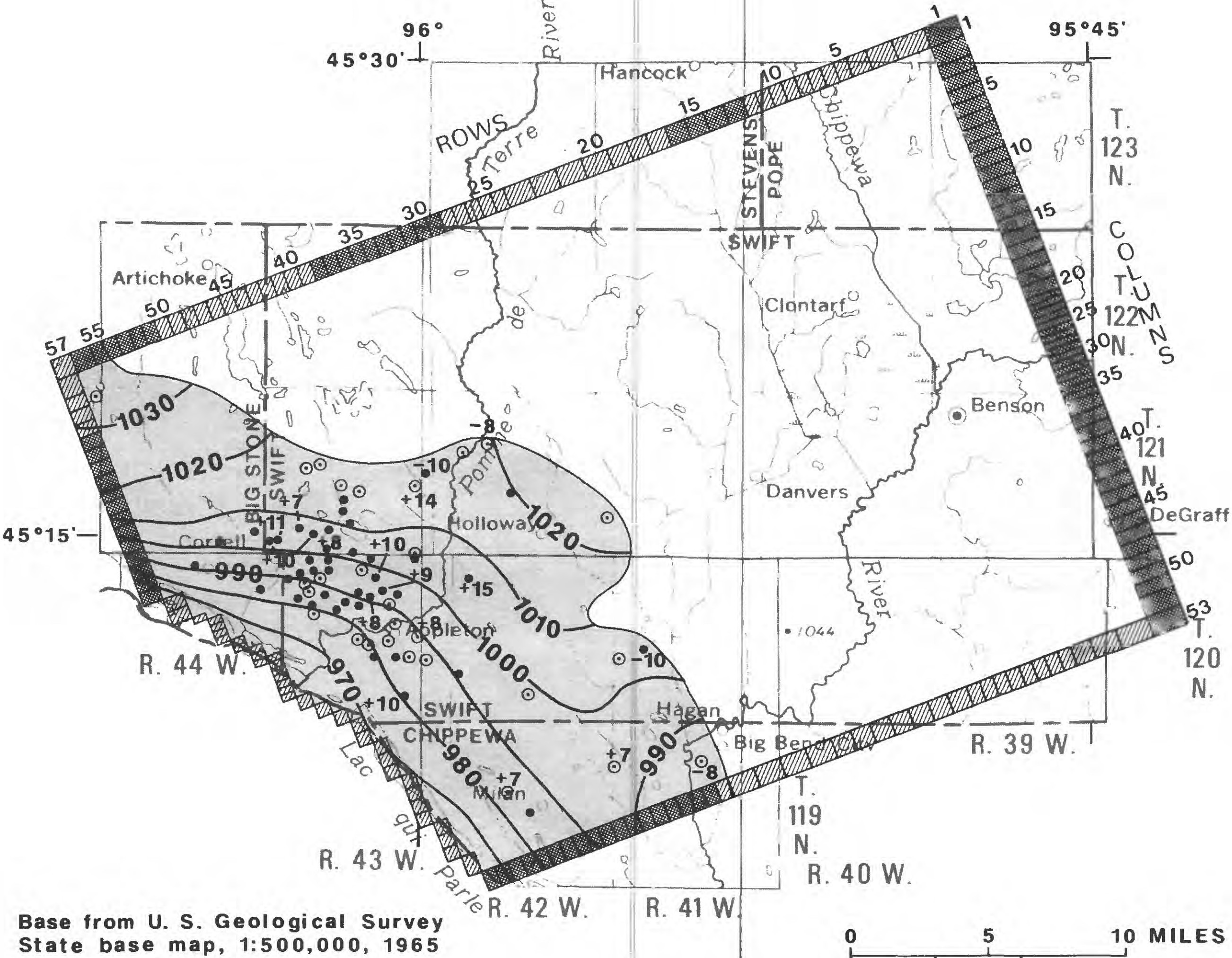


Figure 8.--Boundary conditions and difference between model-computed and measured water levels in the Benson-middle aquifer for the steady-state simulation



EXPLANATION

APPLETON AQUIFER

TILL

—990— MODEL GENERATED POTENTIOMETRIC CONTOUR--Interval 10 feet. Datum is National Geodetic Vertical Datum of 1929

CONSTANT FLUX BOUNDARY

NO-FLOW BOUNDARY

DATA POINT--Number denotes deviation from measured water level if greater than +6 or less than -6 feet:

-10 1982 PUMPING WELL

+7 DOMESTIC OR OBSERVATION WELL

Figure 9.--Boundary conditions and difference between model-computed and measured water levels in the Appleton aquifer for the steady-state simulation

Boundary Conditions

The specification of appropriate boundary conditions is an essential part of accurate simulation of the ground-water-flow system. The natural (physical) hydrologic boundaries of the ground-water system were selected as model boundaries where possible. Selection of the boundary conditions, however, involves considerable simplification of actual hydrogeologic conditions.

The boundary conditions of each model layer are similar and were simulated initially as either constant head or no flow. Potentiometric-surface data (Delin, 1986) indicate that flow lines parallel to parts of the model boundary are steady, that it, they do not change significantly with time. The flow lines represent horizontal flow to the Pomme de Terre, Chippewa, and Minnesota Rivers in each layer. Because no flow crosses a flow line, no-flow boundary nodes were specified coincident and parallel to the flow lines (figs. 7-9). Since ground-water flow is vertical beneath the Minnesota River, no-flow nodes were also specified in layers 2 and 3 beneath the river. Potentiometric-surface data indicate that flow beneath the Appleton aquifer (layer 3) is predominantly horizontal. Consequently, the base of the model was simulated as a no-flow boundary.

A constant-head-boundary condition was specified initially in each layer where flow lines cross model boundaries (figs. 7-9). The value of head for each constant-head node was based on heads measured in each aquifer. Where till exists at the model boundary in layer 1, heads were estimated from lake, stream, and land-surface elevations recorded on U.S. Geological Survey 7-1/2-minute quadrangle maps. Using the heads specified at the constant-head-boundary nodes, the model computed ground-water flow into or out of the modeled area. The Minnesota River also was simulated as a constant-head boundary because control structures on the river prevent significant fluctuations in river stage. These constant-head nodes were assigned heads equivalent to the average river stage.

During the later stages of steady-state calibration, the constant-head nodes were changed to constant flux, except those representing the Minnesota River. The specified flux was set equal to the model-computed amount of water flowing into or out of each node when it was simulated as a constant head. These nodes were changed because the constant-flux boundary provides the most accurate representation of the ground-water-flow system when the model is stressed. If a model-simulated stress reaches a constant-head boundary, erroneously large quantities of water will be induced across model boundaries. A no-flow boundary condition also is inappropriate because some flow, albeit small, does cross the boundary. The combination of constant flux and no-flow boundary conditions selected simulate the ground-water system during hypothetical stress most accurately. These boundary conditions also maximize water-level declines and streamflow depletion computed by the model. Consequently, the model computed the probable "worst-case" results of the stresses applied to the system. These results are of greatest value to the water-resource manager in making ground-water management decisions. A detailed description of the properties of each type of boundary condition specified in the model is presented by Franke and others (1984).

Steady-State Calibration and Sensitivity Analysis

The model was calibrated to assure that the hydrologic properties and boundaries selected were reasonable for simulation of flow in the ground-water system. The model was calibrated for steady-state conditions by comparing measured ground-water levels and calculated ground-water discharge to rivers with corresponding values computed by the model. Calibration of the model was achieved when model-computed water levels and ground-water discharge rates acceptably matched corresponding measured values.

Model-computed water levels were compared to heads measured in the field during November and December 1982. These heads approximate equilibrium conditions in the ground-water system. Hydrographs near Appleton, however, indicate that water levels in this area still were recovering from pumping during the summer at the time they were measured. Therefore, model-computed heads were expected locally to be as much as 1 to 2 ft higher than measured values, particularly in the Appleton area. In addition, land-surface elevations for all data points were estimated to within 5 ft. Thus, model-computed water levels were considered to be acceptable if within 7 ft of observed water levels.

Model-computed leakage to or from the Pomme de Terre and Chippewa Rivers for various stream reaches (fig. 6) was compared to leakage calculated from stream discharge measured May 23 and November 6, 1980. Flow duration on the Pomme de Terre River at Appleton was approximately 40 percent on May 23, and 55 percent on November 6, 1980 (Soukup and others, 1984).

The calibration procedure consisted of successively adjusting hydrologic-input values until model-computed water levels and ground-water-discharge rates matched field measurements. The transmissivity and hydraulic conductivity distribution were well defined for the aquifers and till. The streambed-leakage coefficient and areal recharge to the ground-water system, although based on field data, were less reliable. Therefore, adjustments during steady-state calibration centered largely on the streambed-leakage coefficient and the areal-recharge rate.

In the initial estimate of recharge to the ground-water system, a 'lumped' recharge value was applied, representing the net amount of water reaching the ground-water system after losses to ET. This approach generated unacceptable results in the Benson area, and the model was particularly sensitive to changes in recharge rates near Benson. Because the water table is at or near land surface over a large part of the Benson area (fig. 5), ground-water loss to ET was simulated separately from areal recharge in all subsequent model simulations. Following this adjustment, a more acceptable match between model-computed and measured heads was achieved. Figure 10 shows where and what rate water was removed from the ground-water system by ET. Figure 10 also clearly illustrates that ground-water losses are significant near Benson and negligible near Appleton. The areas where the model indicates that ground water is removed by ET correspond favorably with areas where the water table is known to be less than 5 ft below land surface. The ET rates shown in figure 10 may differ, however, from actual field conditions due to the inaccuracies inherent to simulation of this complex drift system.

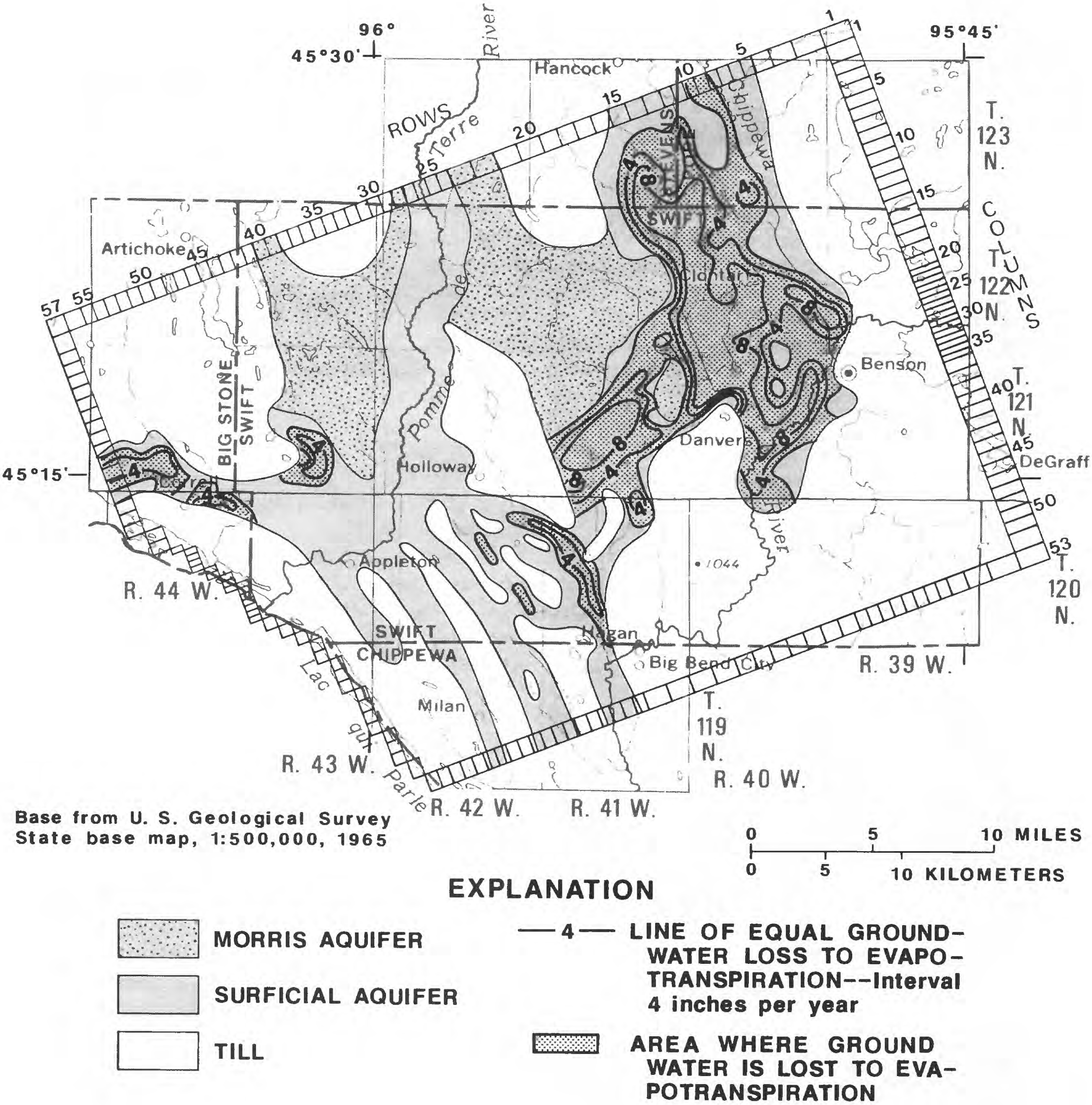


Figure 10.--Location and model-computed rate of ground-water loss to evapotranspiration for steady-state conditions

During calibration, the areal recharge and maximum ET rates to the surficial aquifer were varied uniformly between 4 and 8 in/yr and 6 and 12 in/yr, respectively. A recharge rate of 6 in/yr and an ET rate of 9 in/yr, produced the best model simulation of heads. The maximum ET rate corresponds to the value calculated using the formula described earlier.

The areal recharge rate applied in areas where till is the uppermost unit was set initially to 1 in/yr. This rate was varied areally throughout the model on successive simulations until model-computed water levels were acceptable. This adjustment process is justifiable because recharge rates to the till probably vary areally due to changes in the vertical hydraulic conductivity of till. The areal-recharge rate to till at land surface for the final steady-state model run varied between 0.1 and 3.4 in/yr. These rates agree favorably with the leakage rates to confined aquifers estimated from field data.

Adjustment of streambed-leakage coefficients of one order of magnitude produced differences in model-computed water levels of as much as 7 ft near the Pomme de Terre and Chippewa Rivers, which demonstrates that the model is sensitive to changes in leakage. Adjustments during steady-state calibration resulted in leakage coefficients ranging from 0.001 to 10.0 (ft/d)/ft. Model-computed and calculated ground-water discharge to the Pomme de Terre and Chippewa Rivers are shown in table 2. River reaches for which ground-water discharge rates were calculated are shown on figure 6. Most model-computed rates compare favorably with field data and generally are within the range of calculated values or are within about 20 percent of the calculated values. The lack of match in some river reaches probably reflects a lack of knowledge of the ground-water system near those reaches. However, model-computed values represent steady-state, or long-term average, conditions, whereas, the calculated values represent a specific time. Consequently, the calculated values may not be at steady state, depending on the hydrologic conditions, and cannot be expected to match model-computed values perfectly.

Model sensitivity to the lateral boundaries was tested during steady-state calibration by specifying various boundary conditions. Constant-head, no-flow, constant-flux, and head-dependent-flux boundaries were simulated alternately. The test produced 1- to 3-ft differences in model-computed water levels within two or three nodes of the boundary, and insignificant effects toward the areas of intensive ground-water pumpage near Appleton and Benson. No-flow and constant-flux boundaries were specified for the final steady-state calibration because these conditions probably simulate the ground-water system during hypothetical stress most accurately.

Model sensitivity to the vertical hydraulic conductivity of till was tested by varying the value uniformly between 2.5×10^{-3} and 2.5×10^{1} ft/d. Model-computed water levels varied less than 2 ft when the value was increased or decreased one order of magnitude. A uniform value of 2.5×10^{-2} ft/d was used for the final steady-state calibration model run. This value corresponds favorably with the vertical hydraulic conductivity values calculated from aquifer-test data (Delin, 1986).

Table 2.--Model-computed and calculated ground-water discharge to the Pomme de Terre and Chippewa Rivers
(Discharge values are in cubic feet per second)

River	River reach number[1]	Ground-water discharge to river: Calculated[2], Date 5-23-80	Calculated[2], Date 11-6-80	Model computed
Pomme de Terre	1	5.4	0.9	9.6
Do.	2	-2.3	8.0	2.4
Do.	3	5.4	-6.3	4.4
Do.	4	-0.6	1.3	1.6
Do.	5	-2.0	-0.5	0.3
Do.	6	12.5	3.8	6.5
Chippewa	7	7.0	4.6	7.5
Do.	8	14.7	3.6	3.0
Do.	9	-8.0	-2.0	0.1
Do.	10	0.0	1.0	0.9
Do.	11	22.5	2.9	1.3
Do.	12	11.5	11.5	2.5

[1] See figure 6 for location of river reaches.
[2] Negative number indicates loss of flow to the ground-water system.

Model sensitivity to differences in the horizontal hydraulic conductivity of till was tested by varying the value uniformly throughout the model in the range from 0.01 to 100 ft/d. Increases or decreases of one order of magnitude resulted in model-computed water-level differences of less than 1 ft. A value of 1.0 ft/d produced the best water-level match and was used for the final steady-state model run. This value is within the range of values given by Todd (1959) and Freeze and Cherry (1979).

Model sensitivity to simulation of horizontal flow in till was tested by making till nodes inactive (no flow) in each model layer. As a result, model-simulated water levels were several feet higher and ground-water discharge to the Minnesota River was reduced by approximately 20 percent, compared to simulations which account for horizontal flow in till. This suggests that simulation of horizontal flow in till is necessary to accurately simulate the ground-water-flow system.

The horizontal hydraulic conductivity of each aquifer was varied between 50 and 400 ft/d. A variation in conductivity of 50 ft/d generally produced water-level differences of less than 1 ft in each aquifer. Conductivity values of 260 ft/d near Appleton and 350 ft/d near Benson were assigned to the surficial aquifer and conductivities of 260, 140, and 100 ft/d were assigned to the Morris, Appleton, and Benson-middle aquifers, respectively, for the final steady-state model run. These values agree closely with values calculated from aquifer tests conducted for this study (Soukup and others, 1984, and Delin, 1986).

The model-computed steady-state potentiometric surface for each model layer is shown in figures 7-9. The general flow pattern shown in these figures agrees favorably with potentiometric maps published by Delin (1986, pl. 4). A positive number at well locations indicates a model-computed water level higher than the measured value. Model-computed water levels generally matched measured water levels within 7 ft, but about 10 percent of the model-computed values differ by more than 7 ft. Most of the larger differences between computed and observed water levels in the surficial and Appleton aquifers near Appleton probably result from measurement of water levels that had not recovered from pumping during the summer irrigation season.

A water budget is an accounting of inflow to, outflow from, and storage in the ground-water system. For steady-state conditions, the inflow (sources) to the system, equal the outflow (discharges) from the system. A general equation of the steady-state water budget in the modeled area can be written as:

Recharge from precipitation + ground-water flow into the modeled area = evapotranspiration + ground-water discharge to rivers + ground-water pumpage

The steady-state water budget for the calibrated model (table 3) shows that recharge from precipitation accounts for the major inflow to the system. The table also shows that evapotranspiration and discharge to the principal streams account for most of the discharge from the system.

The model was used to determine the amount of ground-water flow between the confined and surficial aquifers. Ground-water flow from the surficial aquifer through till to the Benson-middle confined aquifer is about 1,900 Mgal/yr. Approximately 1,200 Mgal/yr flows from overlying aquifers to the Appleton confined aquifer. About 500 Mgal/yr of this total flows directly from the surficial aquifer to the Appleton aquifer where they are connected just north of Appleton (fig. 5). Approximately 50 Mgal/yr flows from the surficial aquifer to the Benson-middle aquifer where they are connected north of Appleton (fig. 5). The steady-state water budget (table 3) indicates that 35 Mgal/yr flows from the Benson-upper aquifer to the surficial aquifer, which occurs where the aquifers merge north of Benson (fig. 5). Model results indicate that, for steady-state conditions, heads in the drift aquifers decrease with depth, except near rivers, and are generally within 1 ft of each other. On a seasonal basis, however, head differences between the aquifers is greater because of seasonal pumpage effects. The model suggests that short-term head differences of as much as 13 ft could be expected between the surficial aquifer and the Benson-middle aquifer in the vicinity of high-capacity pumping centers near Benson.

Table 3.--Steady-state water budget for the calibrated model

Sources	Rate (Mgal/yr)	Percent
Recharge from precipitation.............................	28,501	98
Ground-water flow across model boundaries into the modeled area (constant flux)..........................	490	2
Leakage from the Benson-upper aquifer to the surficial aquifer..................................	35	0
Total inflow......	29,026	100

Discharges	Rate (Mgal/yr)	Percent
Evapotranspiration.....................................	11,204	39
Ground-water discharge to the Pomme de Terre and Chippewa Rivers....................................	10,509	36
Ground-water pumpage...................................	4,954	17
Ground-water discharge to the Minnesota River............	2,359	8
Total outflow.....	29,026	100

Transient Simulations

In order to establish that the model can accurately simulate changes in ground-water flow with time, the model was used to simulate the effects of historical pumping and climatic stresses on the ground-water system. Model results were compared to water-level fluctuations measured in observation wells during the period 1980-82.

Heads from the 1982 steady-state solution were used as starting heads for the transient simulations. At most observation points in the modeled area, water levels in the winter of 1980 and 1982 compared within 2 to 3 ft. Considering the margin of error of 5 ft allowed in estimating land-surface elevations at each observation point, the 2- to 3-ft difference between 1980 and 1982 water levels is insignificant.

Areal recharge rates applied for each year of the transient simulations were calculated using the hydrograph analysis method of Rasmussen and Andreason (1959). The average areal recharge rate for the surficial aquifer is listed in table 4. Maximum ET rates were the same for each year simulated and were calculated using the equation described on page 39. While the seasonal breakdown of ET was based on studies of seasonal fluctuations in panevaporation rates conducted by Baker and others (1979), pumping rates were based predominantly on total annual pumpage reported to the MDNR. Aquifer and confining-bed input values were identical to final steady-state values. Storage coefficients of 0.2 (Soukup and others, 1984) and 2.0×10^{-4} (Delin, 1986) were specified for the surficial and confined aquifers, respectively. The no-flow and constant-flux boundary conditions specified for the final steady-state simulation were used for the transient simulations.

Table 4.--Summary of seasonally variable input data used in transient simulations

[Areal recharge in inches per year, maximum evapotranspiration rate (ET) inches per year, and pumping rate in millions of gallons per year]

Stress period (annually)	Duration (days)	Hydrologic property	1980	1981	1982
1 January through March	90	Areal recharge	0.0	0.0	0.0
		Maximum ET rate	0.0	0.0	0.0
		Pumping rate	0.0	0.0	0.0
2 April through June	90	Areal recharge	5.8	4.8	5.4
		Maximum ET rate	4.05	4.05	4.05
		Pumping rate	0.0	0.0	0.0
3 July	30	Areal recharge	0.0	0.0	0.0
		Maximum ET rate	2.25	2.25	2.25
		Pumping rate	5,640	4,899	4,954
		Number of wells	194	184	219
4 August through September	60	Areal recharge	0.0	0.0	0.0
		Maximum ET rate	2.7	2.7	2.7
		Pumping rate	0.0	0.0	0.0
5 October through December	90	Areal recharge	0.6	0.5	0.6
		Maximum ET rate	0.0	0.0	0.0
		Pumping rate	0.0	0.0	0.0

Each year of a transient simulation was subdivided into five stress periods. Seasonally variable input data used in the transient simulations are shown in table 4. The areal recharge, maximum ET, and pumping rates were varied during each stress period to approximate actual field conditions. During April through June, 90 percent of the total annual recharge and 45 percent of the total annual ET was applied; during July, 25 percent of the ET and 100 percent of the pumping was simulated; during August through September, 30 percent of the ET was simulated; and during October through December, 10 percent of the recharge was simulated. In contrast to areal recharge, leakage to the confined aquifers is continuous throughout the year with only the rate varying seasonally. Simulation of this leakage in the model was simplified, however, by assuming that leakage to confined aquifers, where till is present at land surface, fluctuates in response to precipitation. During the initial transient runs, pumping was simulated during a stress period covering June through August.

Unfortunately, drawdowns associated with a June to August pumping scheme were not closely duplicated. This was not surprising because (1) pumping was averaged over a three-month period rather than when it actually occurred, (2) model-computed heads were averaged over each grid block and assigned to the appropriate node--therefore, if an observation point used in water-level comparisons for the transient simulations is not located at the node, a discrepancy can be expected, and (3) the observation points for confined aquifers, used in water-level comparisons, are located within approximately 1,000 ft of high-capacity irrigation wells. Consequently, seasonal water-level fluctuations at these observation points are greater than the more regionalized fluctuations computed by the model. To compensate for these expected inaccuracies and, in the process, more closely simulate the summer water-level declines, 100 percent of the pumping was simulated during a July stress period.

A comparison of model-computed and measured hydrographs was made at 10 locations. The hydrographs in figures 11 through 13 show model-computed and measured water-level fluctuations in selected wells in each aquifer for 1980-82. Generally, there is good agreement between model-computed and observed water-level fluctuations. The model also generally simulated water-level declines from 1980-82 to within 15 percent of measured declines, over the 3-year period. Therefore, the model generally simulates regional trends in aquifer response to stress. Several peaks and valleys in the model-computed hydrographs are displaced somewhat in time compared to observed hydrographs (fig. 13, for example), because spring recharge always was simulated as occurring in May and 100 percent of the annual pumpage was simulated in July to facilitate modeling. Based on analysis of the transient-simulation results, further adjustment of model input values was not considered to be necessary.

The model-computed water budget for July 1980 (stress period 3) is shown in table 5. These values reflect the maximum effect of irrigation pumping on the ground-water system and indicate the sources of water necessary to sustain pumping. Model results indicate that the vast majority of the water pumped during the summer irrigation season comes from storage, because areal recharge is assumed to be zero in July. The table also shows that ground-water pumpage accounts for most of the discharge from the system during July. In addition, ground-water pumpage during July decreases evapotranspiration losses, ground-

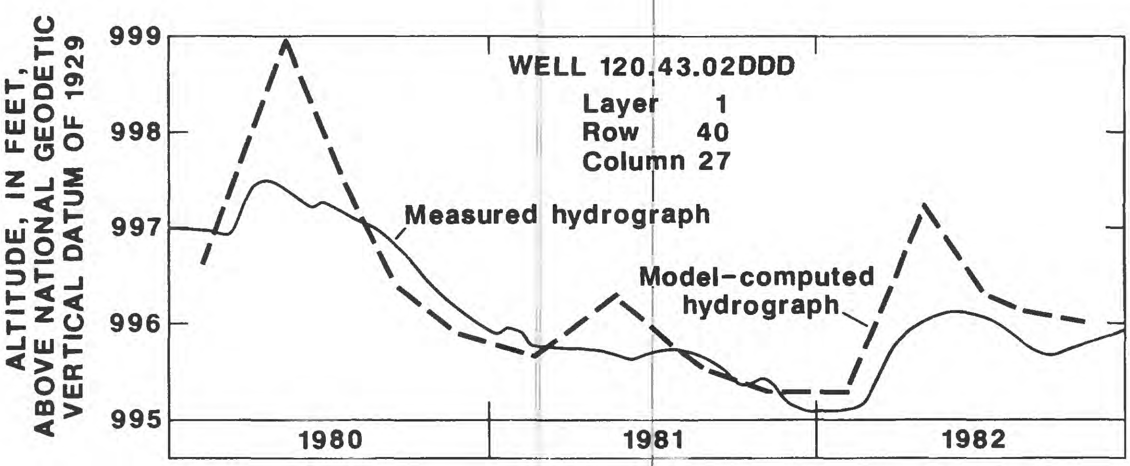


Figure 11.--Hydrographs showing model-computed and measured water levels for the surficial aquifer, 1980-82

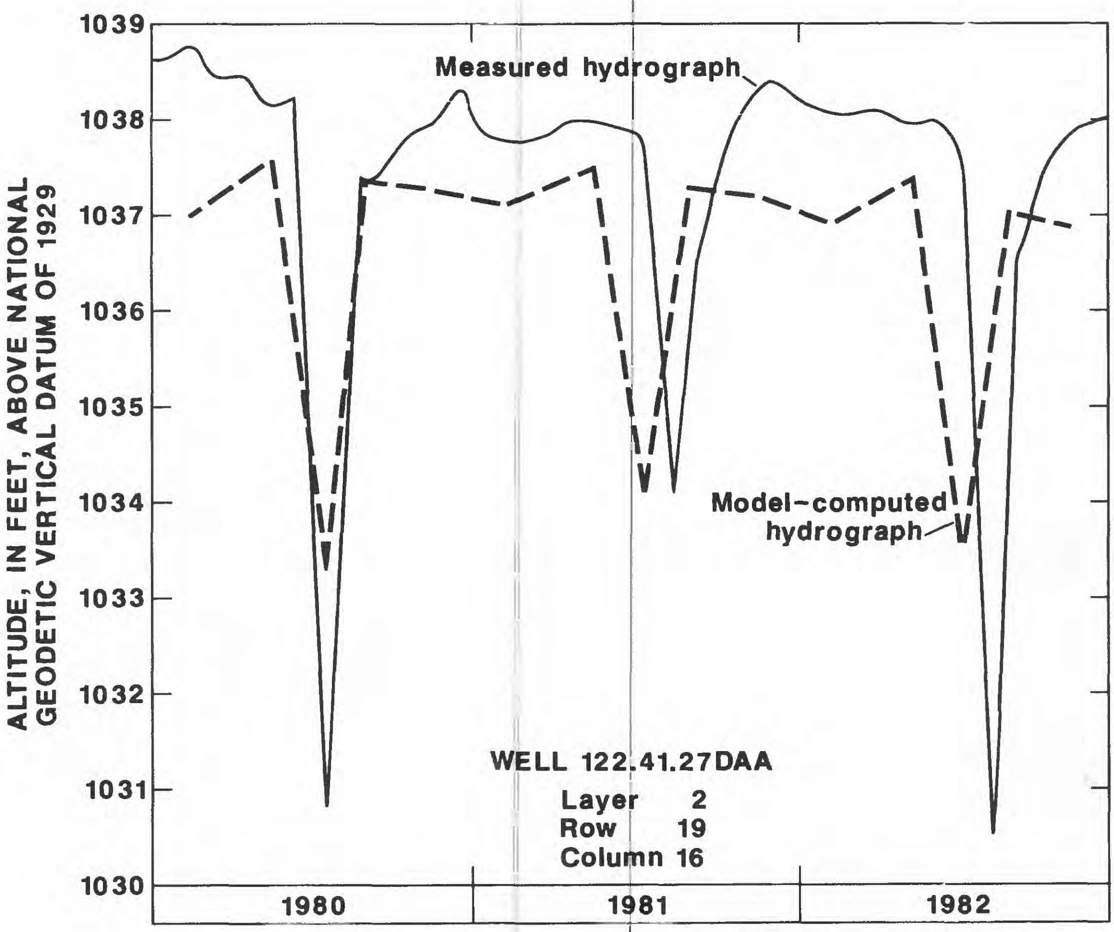


Figure 12.--Hydrographs showing model-computed and measured water levels for the Benson-middle aquifer, 1980-82

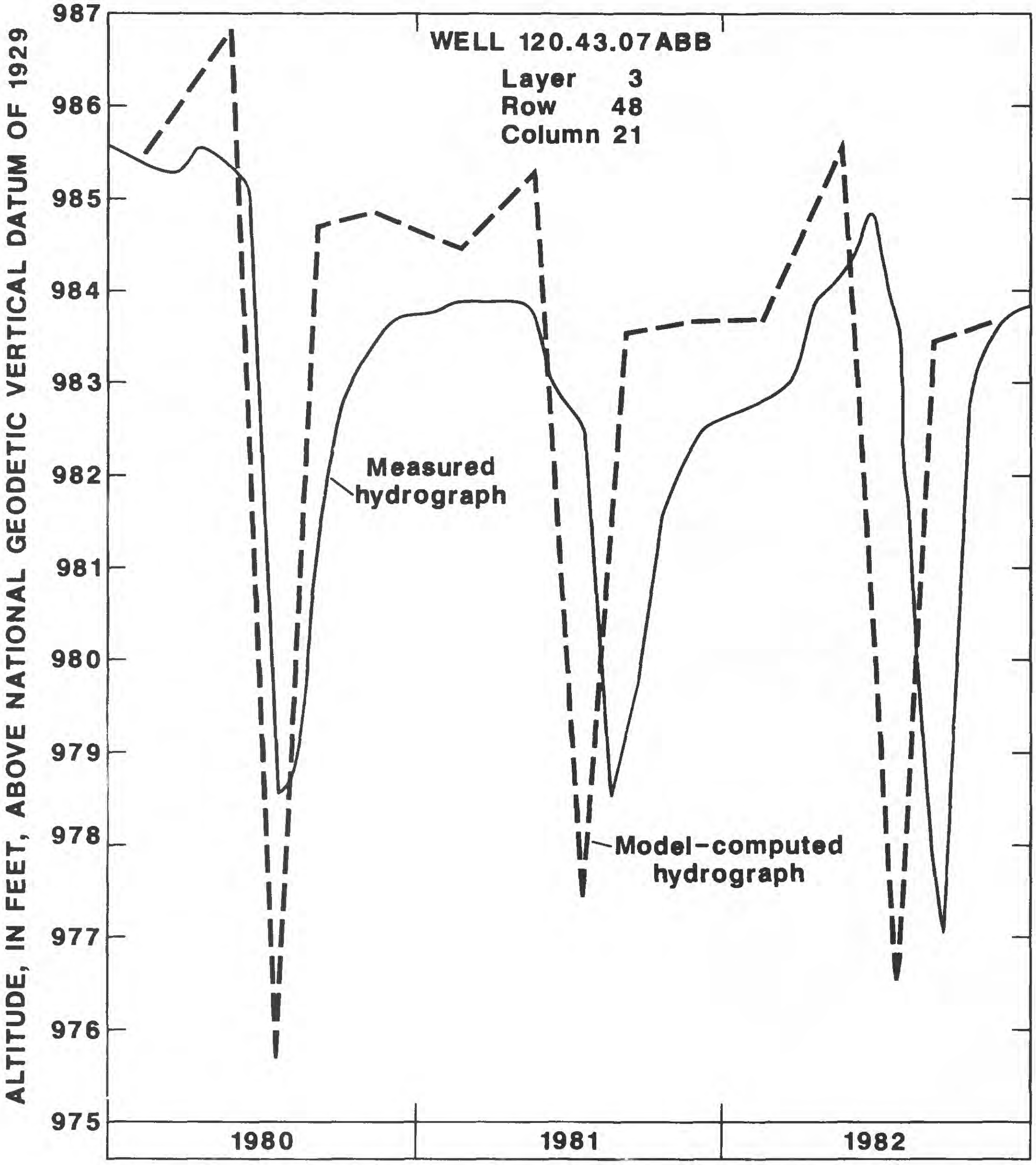


Figure 13.--Hydrograph showing model-computed and measured water levels for the Appleton aquifer, 1980-82

water discharge to the Minnesota River, and ground-water discharge to the Pomme de Terre and Chippewa Rivers compared to steady-state conditions (table 3).

Results of Model Analysis of Ground-Water Availability

The calibrated steady-state model was used to evaluate ground-water availability by assessing the potential effects of hypothetical conditions on ground-water levels and streamflow in the area. The hypothetical simulations investigate (1) the effects of historical and 1982 pumping, (2) the effects of an extended drought, and (3) the potential for additional ground-water development from the Appleton and Benson-middle aquifers and from the surficial aquifer near Benson.

The constant-flux and no-flow boundary conditions specified for the final steady-state calibration simulation were used in all the simulations discussed in this section. The 1982 heads were used as starting heads for each steady-state simulation. Thus, model results indicate the long-term effects of the hypothetical conditions simulated because, unlike transient simulations, no water is derived from storage. Table 6 is a summary of the hypothetical model simulations and corresponding aquifer responses. Table 7 summarizes the water budget for each simulation.

Table 5.--Model-computed water budget for July 1980 (stress period 3) for the transient simulation

Sources	Rate (Mgal/yr)	Percent
Change in storage	6,305	95
Leakage from the Pomme de Terre and Chippewa Rivers	133	2
Ground-water flow across model boundaries into the modeled area (constant flux)	118	2
Leakage from the Benson-upper aquifer	79	1
Total inflow	6,635	100
Discharges	**Rate (Mgal/yr)**	**Percent**
Ground-water pumpage	5,640	85
Evapotranspiration	664	10
Discharge to the Minnesota River	331	5
Total outflow	6,635	100

Table 6.--Summary of results of hypothetical model simulations A, B, C, and D

Simulation	Conditions of simulation	Model results: Layer 1[a]	Model results: Layer 2[b]	Model results: Layer 3[c]
A	Predevelopment: 1982 pumping removed to determine effects of historical pumpage Average areal recharge	Water levels have declined 2 and 1 ft regionally in the Appleton and Benson areas, respectively, and as much as 4 ft locally near Appleton; ground-water discharge to rivers has decreased 18 percent since predevelopment.	Water levels have declined 1 ft regionally and as much as 13 ft locally near the Benson city wells.	Water levels have declined 1 to 2 ft regionally and as much as 4 ft north of Appleton and near Holloway.
B	Present well development (206 wells) Pumping stress: actual (1982) X 1.5 Drought: 30-percent less recharge for 3-year duration	Water levels decline 4 and 2 ft regionally in the Appleton and Benson areas, respectively, and as much as 10 ft locally; ground-water discharge to river is 49 percent less than steady state; Pomme de Terre River discharge is reduced by 15.2 ft^3/s.	Water levels decline 3 to 4 ft regionally and as much as 11 ft east of Benson and northwest of Lake Oliver.	Water levels decline 3 to 6 ft regionally and as much as 11 ft north of Appleton.
C	Present + hypothetical well development: 14 in layer 1[a] 16 in layer 2[b] 28 in layer 3[c] (264 wells) Pumping stress: actual + estimated Average areal recharge	Water levels decline 1 and 0.5 ft regionally and as much as 5 and 1 ft locally in the Appleton and Benson areas, respectively; ground-water discharge to rivers is 13-percent less than for steady state.	Water levels decline 0.5 ft regionally and as much as 1 ft locally near Benson. Declines of as much as 2 ft occur near Holloway as a result of hypothetical pumping in layer 3.	Water levels decline 1 to 3 ft regionally and as much as 5 ft in some areas.
D	Present + hypothetical well development: 14 in layer 1[a] 16 in layer 2[b] 28 in layer 3[c] Pumping stress: (actual + estimated) X 1.5 Drought: 30-percent less recharge for 3-year duration	Water levels decline 5 and 3 ft regionally in the Appleton and Benson areas, respectively, and as much as 11 ft south of Appleton. Some channels are dewatered. Ground-water discharge to rivers is 57 percent less than for steady state. Pomme de Terre River flow is reduced by 21.7 ft 3/s.	Water levels decline 3 to 5 ft regionally and as much as 11 ft east of Benson and northwest of Lake Oliver.	Water levels decline 5 to 9 ft regionally and as much as 13 ft southeast of Appleton.

[a] Surficial and Morris aquifers
[b] Benson-middle aquifer
[c] Appleton aquifer

Table 7.--Model-computed water budget for the steady-state calibration and hypothetical model simulations A, B, C, C1, C2, and D

[Budget figures are in million gallons per year]

Simulation	Well development (number of wells)	Pumping stress	Areal recharge conditions	Inflow					Outflow				
				Areal recharge	Head dependent boundaries	Constant heads (MN Riv.)[d]	River leakage	Constant flux	Evapotranspiration	River leakage	Ground-water pumpage	Constant heads (MN Riv.)[d]	Constant flux
Steady-state (1982)	1982 (206)	1982	Average	28,495	24	24	71	710	10,591	11,205	4,954	2,359	121
A	Predevelopment (Pumping removed)	None	Average	28,495	24	24	24	710	13,328	13,233	0	2,500	121
B	1982 (206)	1982 X 1.5	Drought (0.7 X average)	19,965	47	47	637	710	6,180	5,496	7,454	2,076	212
C	1982 + hypothetical: 14 in layer 1[a] 16 in layer 2[b] 28 in layer 3[c] (264)	1982 + estimated	Average	28,495	47	24	94	710	10,261	9,695	6,912	2,288	212
C1	1982 + hypothetical: 30 in layer 1[a] 28 in layer 3[c] (264)	1982 + estimated	Average	28,495	47	24	94	710	10,261	9,695	6,912	2,288	212
C2	1982 + hypothetical: 30 in layer 2[b] 28 in layer 3[c] (264)	1982 + estimated	Average	28,495	47	24	94	710	10,261	9,695	6,912	2,288	212
D	1982 + hypothetical: 14 in layer 1[a] 16 in layer 2[b] 28 in layer 3[c] (264)	1982 + estimated X 1.5	Drought (0.7 X average)	19,956	47	165	1,651	710	4,836	5,284	10,379	1,816	212

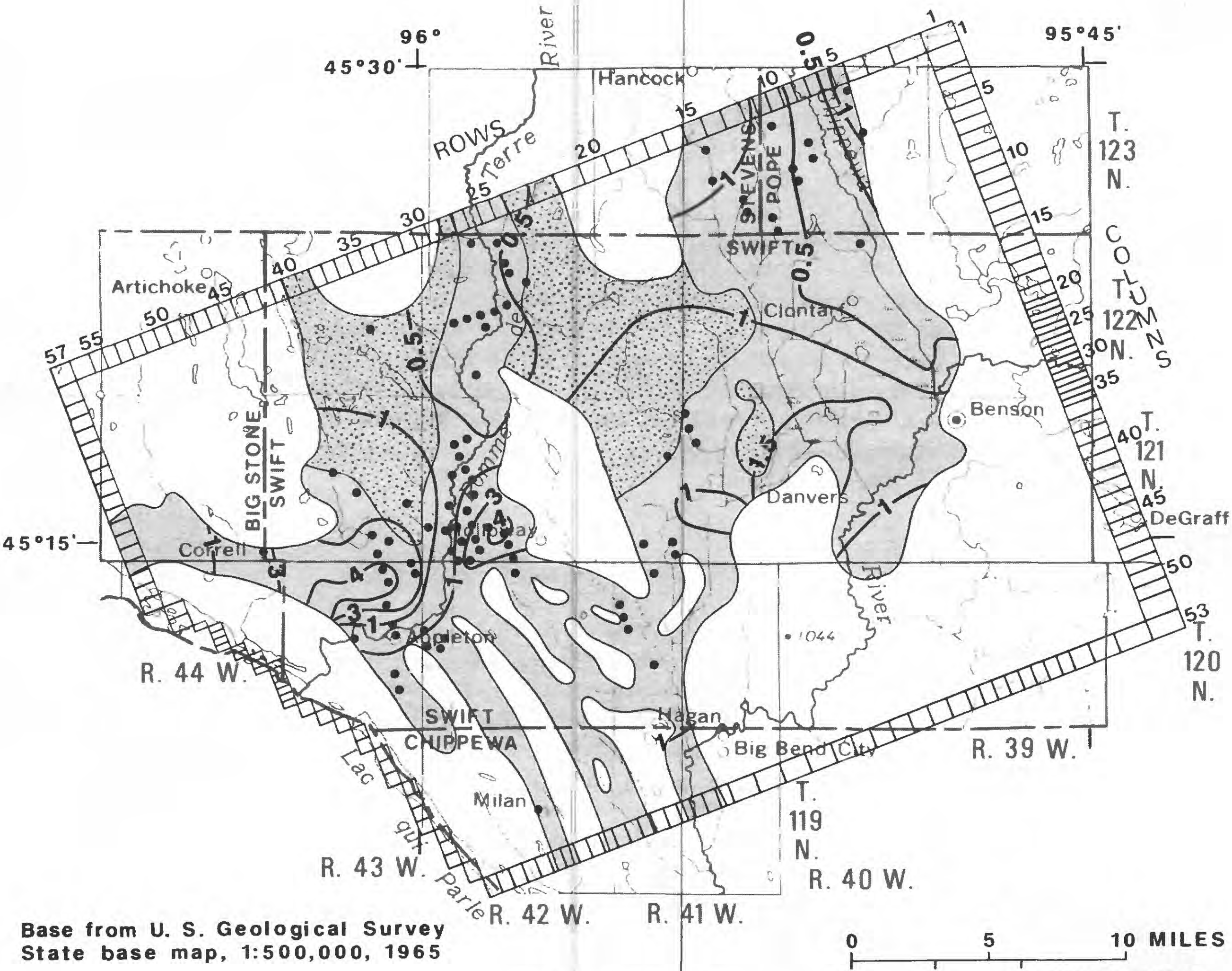
[a] Surficial and Morris aquifers
[b] Benson-middle aquifer
[c] Appleton aquifer
[d] MN Riv. denotes Minnesota River

Historical Pumpage

Simulation A was designed to evaluate the effects on the ground-water system of historical and 1982 pumping. This was achieved by removing pumping from the steady-state model and simulating average recharge conditions. Model results, therefore, are an estimate of predevelopment equilibrium conditions. By comparing results of simulation A with the steady-state (1982) calibration, effects of historical pumping can be estimated. Model results indicate that pumping has lowered water levels between 1 and 2 ft regionally in all layers and as much as 13 ft locally near Benson in the Benson-middle aquifer [layer 2 (figs. 14-16)]. Ground-water discharge to the Pomme de Terre and Chippewa Rivers has been reduced by 18 percent compared to predevelopment conditions and ground-water discharge to the Minnesota River has been reduced by 5 percent. Ground-water loss to ET has decreased 20 percent because pumping has lowered the water table.

Drought

Simulation B was designed to investigate the effects of a 3-year drought similar in severity to the drought of 1976-77. The experiment was a steady-state simulation in which areal recharge was reduced by 30 percent throughout the model and pumping was increased by 50 percent. A separate transient simulation, using the storage coefficients listed on page 44, indicated that it would take the ground-water system 3 years to reach steady-state under these conditions. Thus, the steady-state simulation represents a drought of 3 years duration. Model results indicate that water levels may decline 3 to 7 ft regionally in each aquifer and as much as 11 ft locally near aquifer boundaries as a result of the simulated drought (figs. 17-19). All model-computed water-level declines mentioned in this and following sections are in addition to the historical declines which occurred prior to 1982. Several water-budget terms summarized in table 7 are significantly affected by the simulated drought, compared to 1982 conditions. The model indicates that ground-water loss to ET would decrease 42 percent and ground-water discharge to the Minnesota River would be 13 percent less than for 1982 conditions. In addition, discharge to the Pomme de Terre and Chippewa Rivers would be reduced by 50 percent and leakage from streams increased to 900 percent of 1982 conditions. Flow of the Pomme de Terre and Chippewa Rivers in the modeled area would be reduced by 15.2 and 7.4 ft^3/s, respectively, compared to 1982 conditions. Soukup and others (1984, p. 31-34) show that depletion of the Pomme de Terre River north of the modeled area also would be significant because of irrigation pumping in that area. Depletion of flow in the Chippewa River north of the modeled area also is probable. Therefore, total streamflow depletions for the Pomme de Terre and Chippewa Rivers probably would exceed the model-computed streamflow depletions.



EXPLANATION

MORRIS AQUIFER

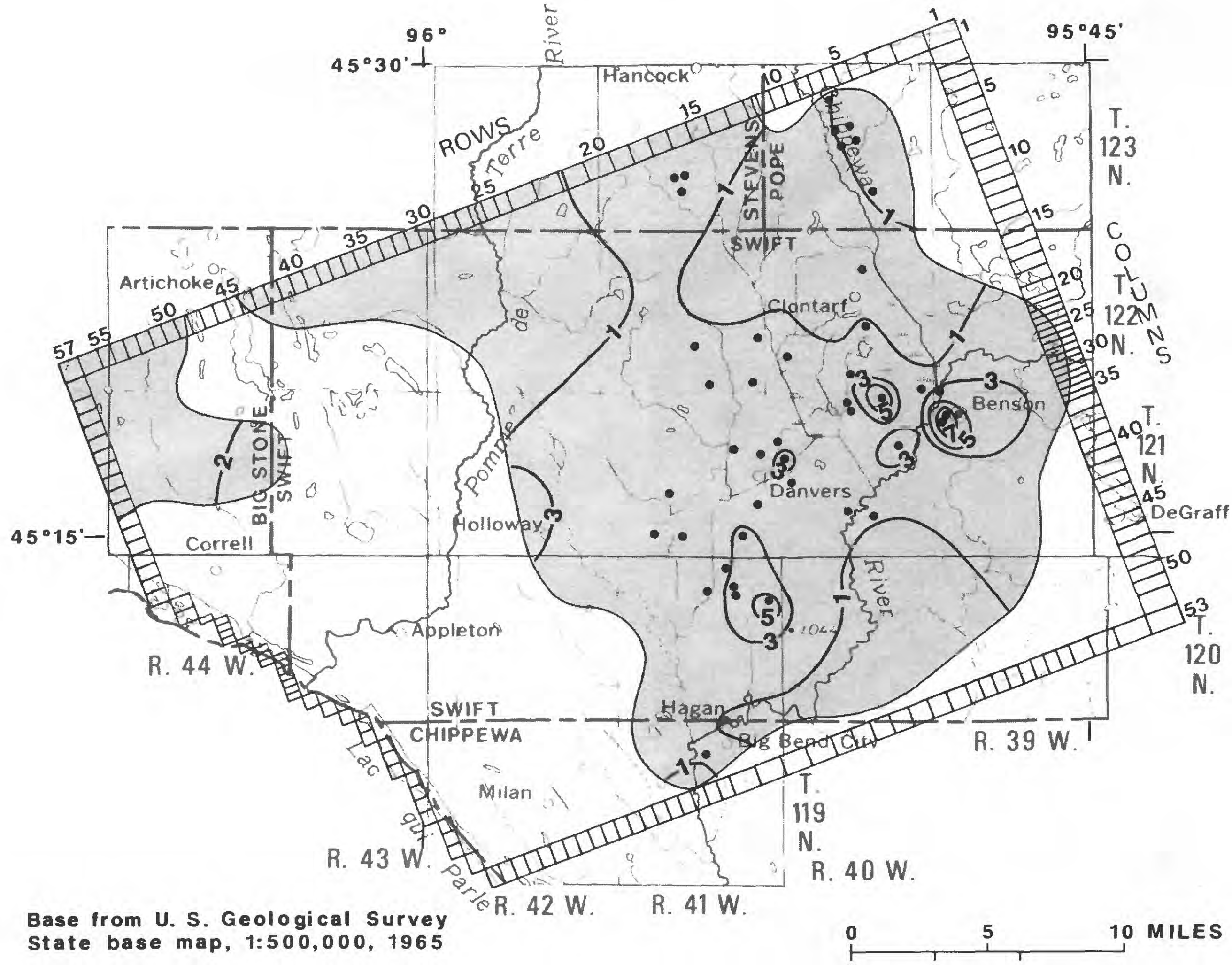
SURFICIAL AQUIFER

TILL

—1— LINE OF EQUAL WATER-LEVEL DECLINE--Interval 0.5, 1, and 2 feet

• 1982 PUMPING WELL

Figure 14.--Model-computed water-level declines in the surficial and Morris aquifers that have resulted from 1982 pumping under steady-state conditions (simulation A)



EXPLANATION

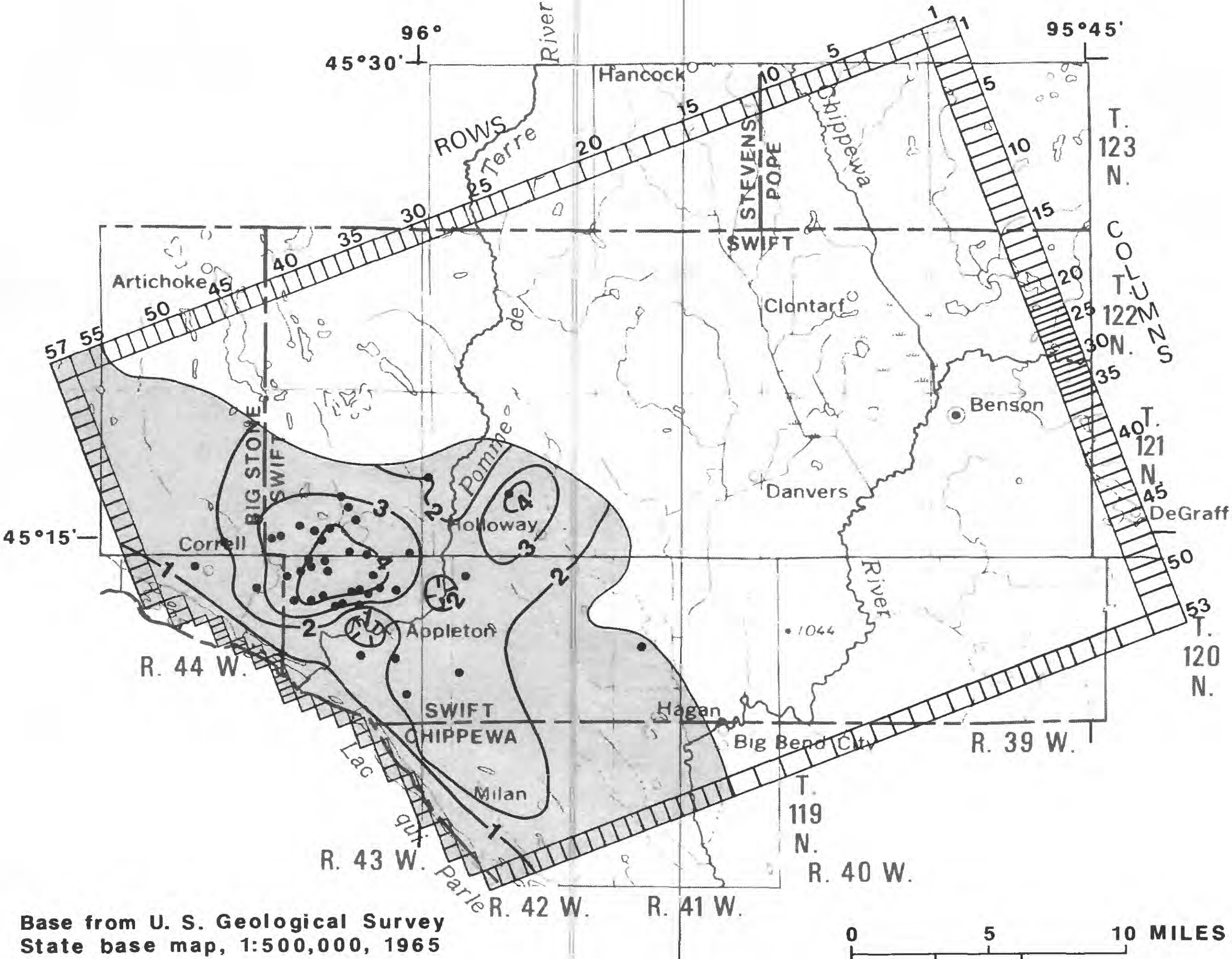
BENSON-MIDDLE AQUIFER

TILL

—5— LINE OF EQUAL WATER-LEVEL DECLINE--Interval 2 feet

• 1982 PUMPING WELL

Figure 15.--Model-computed water-level declines in the Benson-middle aquifer that have resulted from 1982 pumping under steady-state conditions (simulation A)



EXPLANATION

APPLETON AQUIFER	—2— LINE OF EQUAL WATER-LEVEL DECLINE--Interval 1 foot
TILL	• 1982 PUMPING WELL

Figure 16.--Model-computed water-level declines in the Appleton aquifer that have resulted from 1982 pumping under steady-state conditions (simulation A)

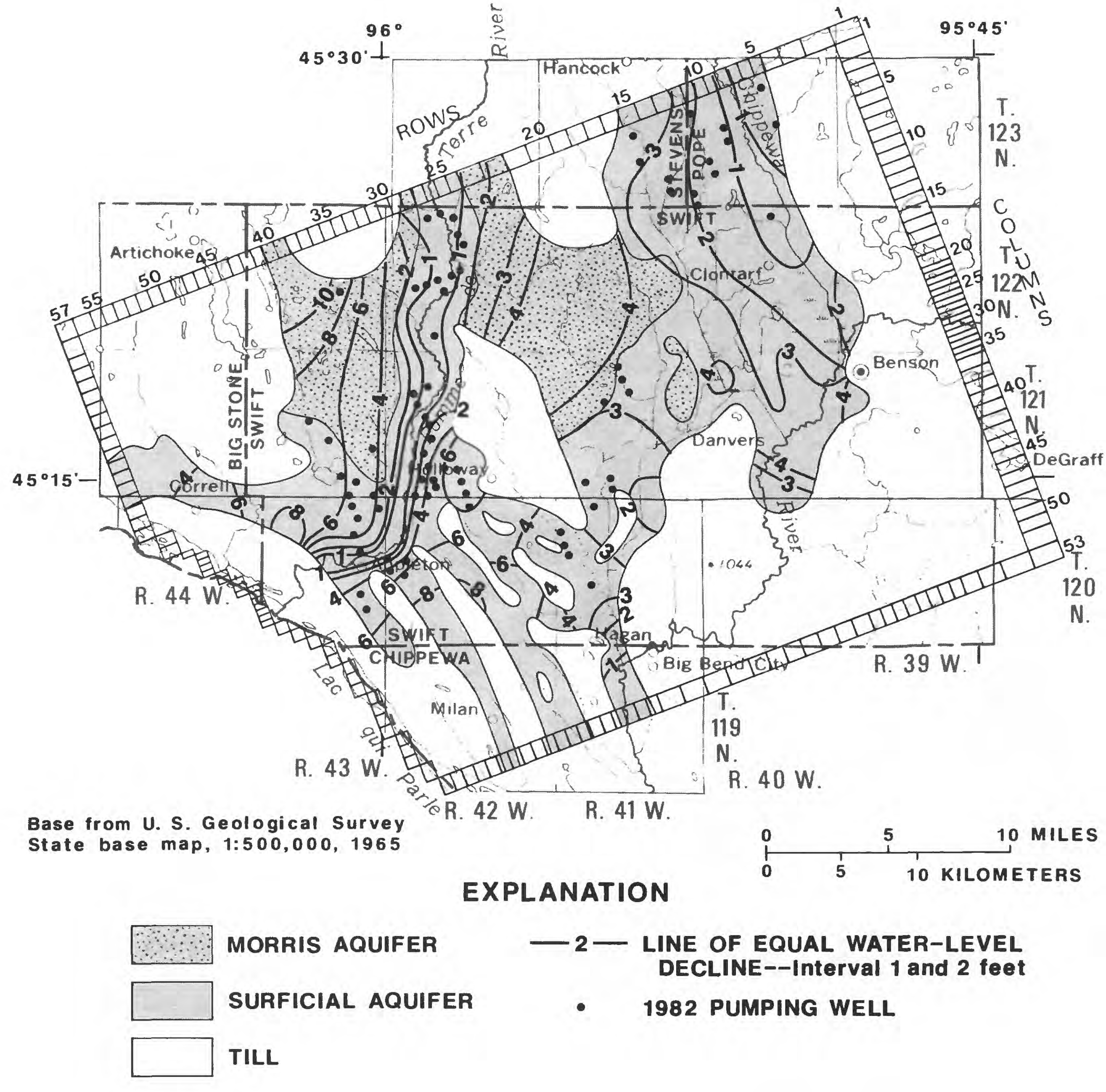
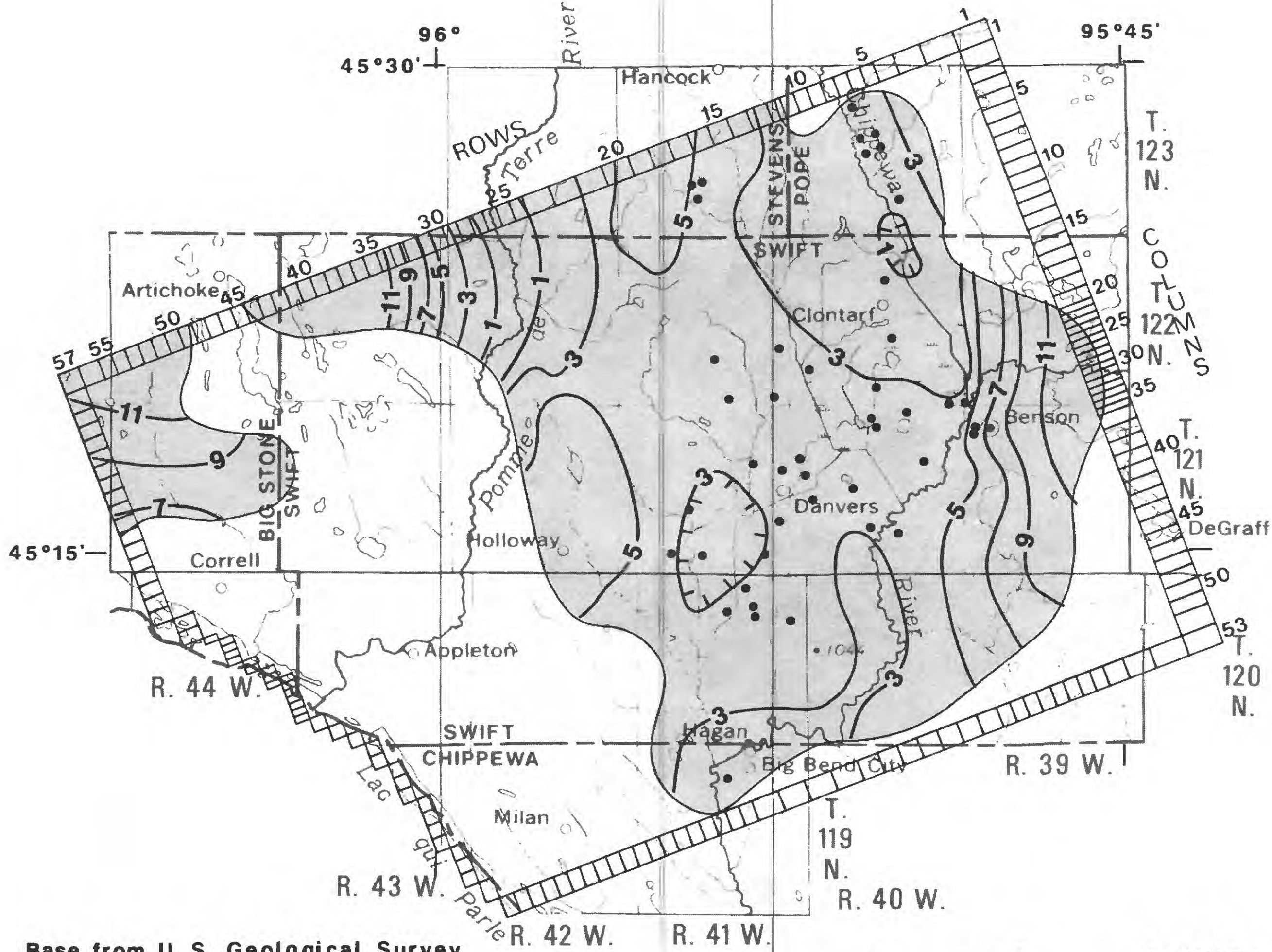


Figure 17.--Model-computed water-level declines in the surficial and Morris aquifers following an extended drought (simulation B)



Base from U. S. Geological Survey
State base map, 1:500,000, 1965

0 5 10 MILES

0 5 10 KILOMETERS

EXPLANATION

BENSON-MIDDLE AQUIFER	—3— LINE OF EQUAL WATER-LEVEL DECLINE--Interval 2 feet
TILL	• 1982 PUMPING WELL

Figure 18.--Model-computed water-level declines in the Benson-middle aquifer following an extended drought (simulation B)

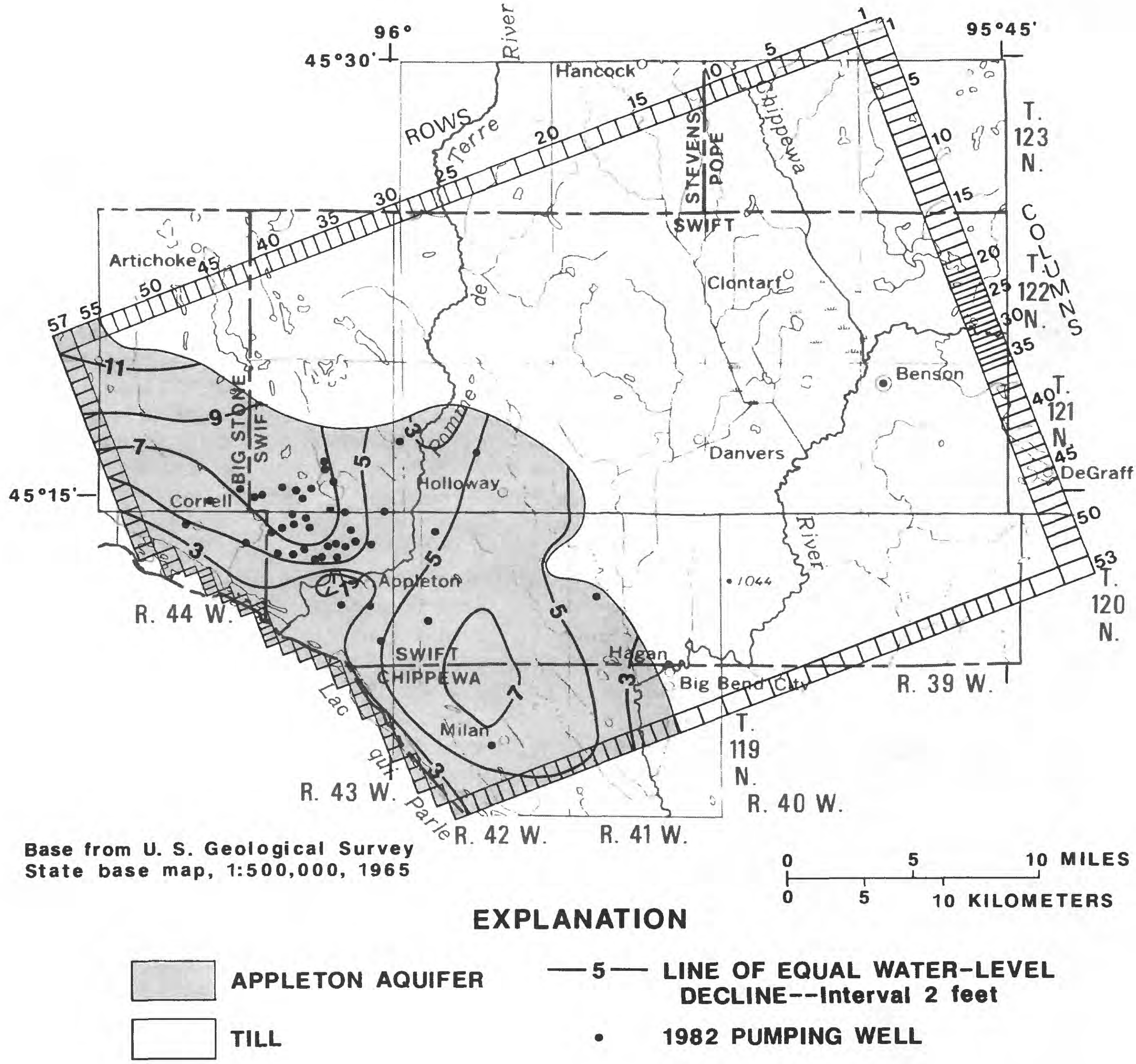


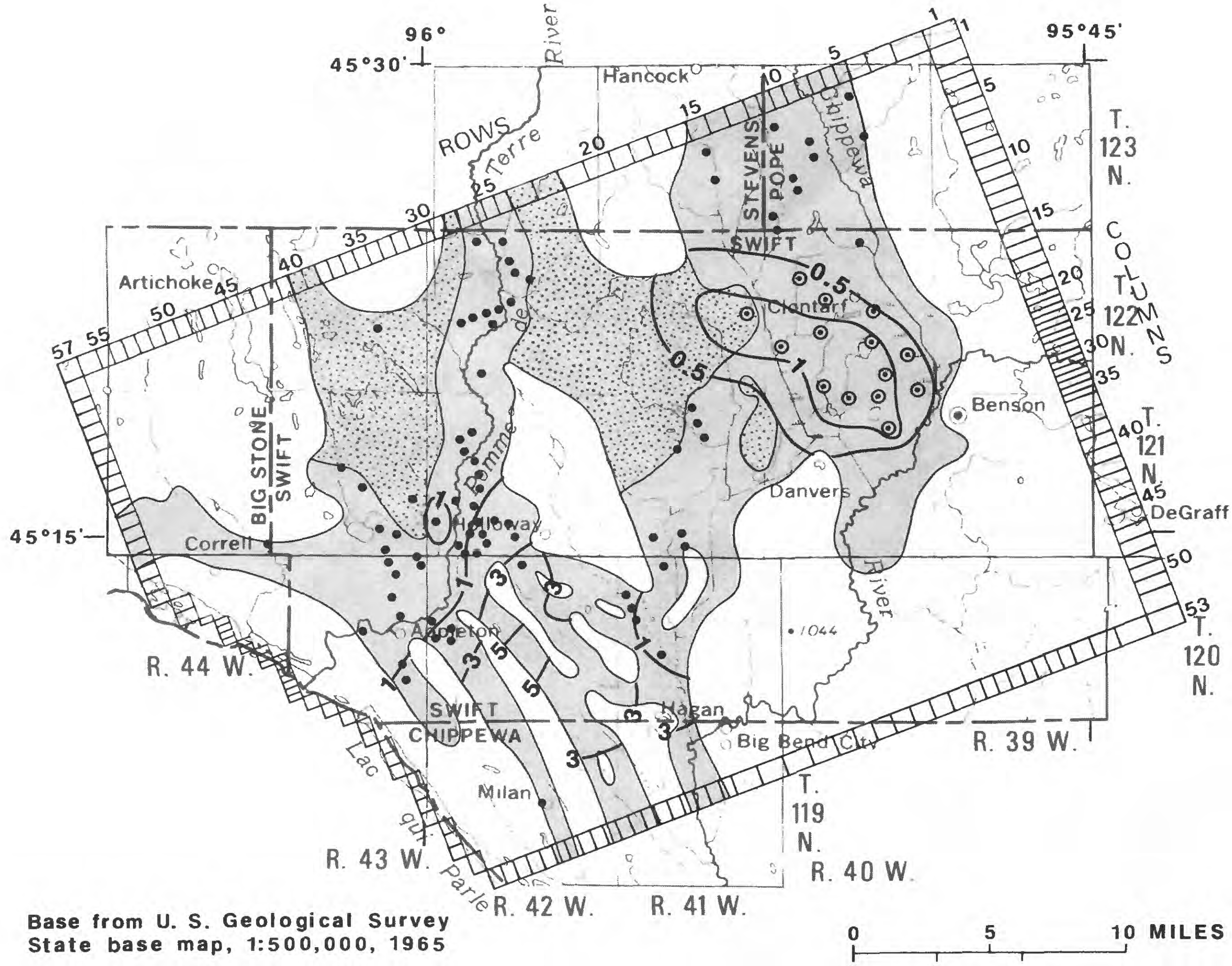
Figure 19.--Model-computed water-level declines in the Appleton aquifer following an extended drought (simulation B)

Expanded Development

Simulation C was designed to simulate the steady-state effects of a hypothetical increase in the number of pumping centers. The MDNR identified two areas with sandy soils near the towns of Appleton and Benson where there is little irrigation of crops but where irrigation could expand in the future. A total of 58 hypothetical wells were simulated. The wells were spaced throughout the two areas to minimize well-interference problems. The average pumping rate for irrigation wells in the modeled area, 27 Mgal/yr, was specified as the pumping rate for each hypothetical well. Withdrawals from 14 hypothetical wells, pumping a total of 378 Mgal/yr, completed in the surficial and Morris aquifers (layer 1) and 16 hypothetical wells, pumping a total of 432 Mgal/yr, completed in the Benson-middle aquifer (layer 2) were simulated west of Benson (figs. 20 and 21). Withdrawals from 28 hypothetical wells, pumping a total of 765 Mgal/yr, were simulated in the Appleton aquifer east and southeast of Appleton (fig. 22). The saturated thickness of the surficial aquifer is between zero and 15 ft in this area near Appleton and is insufficient to sustain additional pumping for an extended period. Therefore, hypothetical wells were not simulated in layer 1 near Appleton.

Simulation C results indicate that water levels may decline between 0.5 and 1 ft regionally in the Morris and Benson-middle aquifers and in the surficial aquifer, and as much as 5 ft in the Appleton aquifer near Appleton due to the hypothetical withdrawals (figs. 20-22). Water-level declines as great as 5 ft in the surficial aquifer and 2 ft in the Benson-middle confined aquifer also may occur in response to hypothetical withdrawals from the Appleton aquifer. Pumping of the hypothetical wells would reduce ground-water discharge to the Pomme de Terre and Chippewa Rivers by an amount equal to 78 percent of the pumpage.

Simulations C1 and C2 were modifications of Simulation C designed to determine the effects of hypothetical pumping from either the surficial aquifer alone or the Benson-middle aquifer alone in the Benson area. In simulation C1, 30 hypothetical wells, pumping a total of 810 Mgal/yr, were simulated in the surficial aquifer west of Benson. In simulation C2, 30 hypothetical wells, pumping the same rates in the same locations as in simulation C1, were simulated in the Benson-middle aquifer. As expected, water-level declines were greater when pumping only from the Benson-middle aquifer. Model results indicate, however, that water-level declines for simulations C1 and C2 would be similar for steady-state conditions. Hypothetical pumping from the surficial aquifer (simulation C1) resulted in a maximum water-level decline of 1.4 ft in the surficial aquifer and 1.3 ft in the Benson-middle aquifer. Hypothetical pumping from the Benson-middle aquifer (simulation C2) resulted in a maximum water-level decline of 1.4 ft in the surficial aquifer and 2.7 ft in the Benson-middle aquifer. Water-level declines in the Benson-middle aquifer in simulation C1 are the result of induced upward leakage through till to the surficial aquifer. Conversely, water-level declines in the surficial aquifer in simulation C2 are the result of induced downward leakage through till to the Benson-middle aquifer. Thus, water-level declines in the Benson-middle aquifer were approximately twice as great when simulating hypothetical pumping only in the Benson-middle aquifer compared to pumping only from the surficial aquifer.



EXPLANATION

MORRIS AQUIFER

SURFICIAL AQUIFER

TILL

—3— LINE OF EQUAL WATER-LEVEL DECLINE--Interval 0.5 and 2 feet

• 1982 PUMPING WELL

⊙ HYPOTHETICAL PUMPING WELL

Figure 20.--Model-computed water-level declines in the surficial and Morris aquifers following expanded development (simulation C)

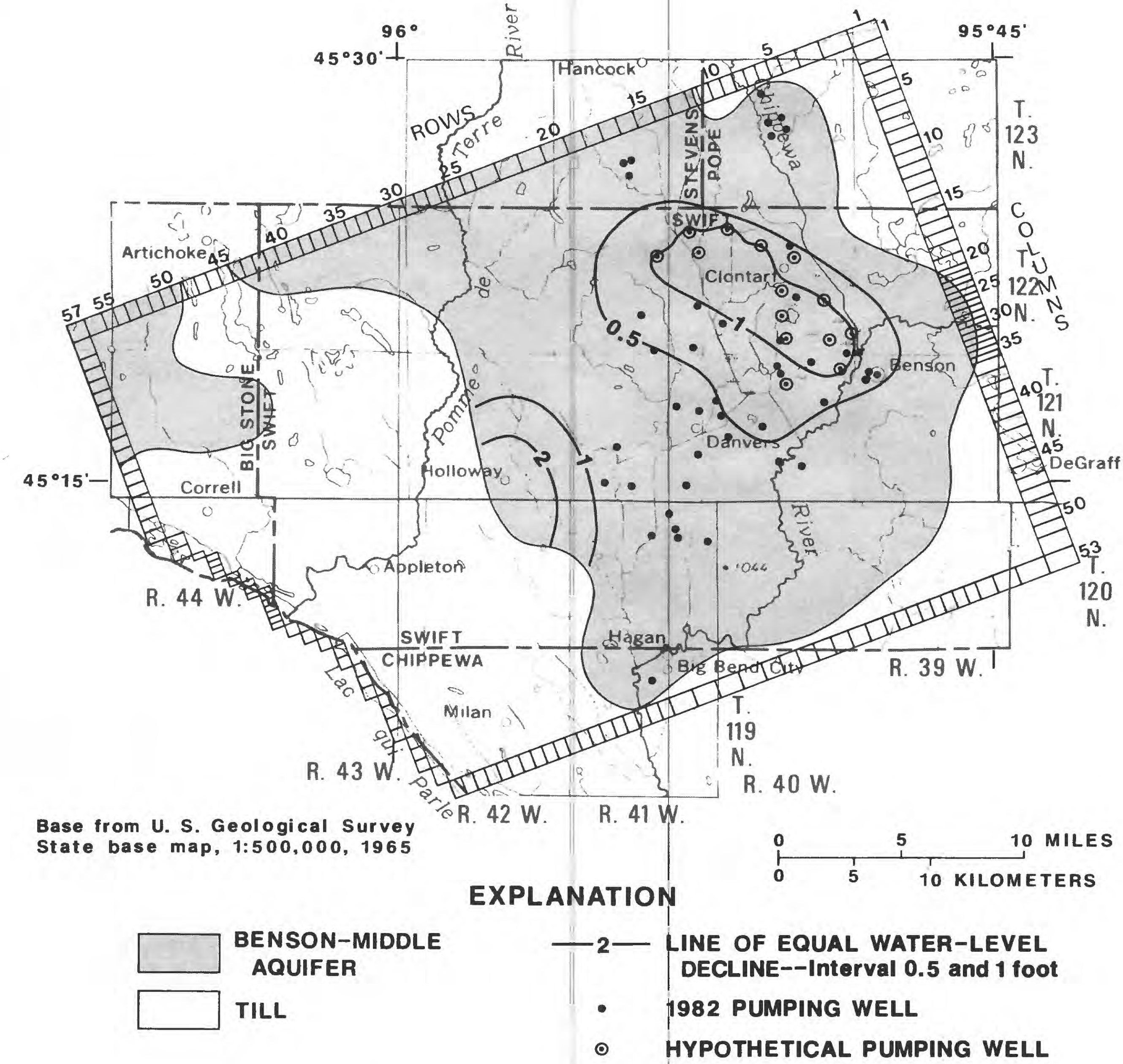
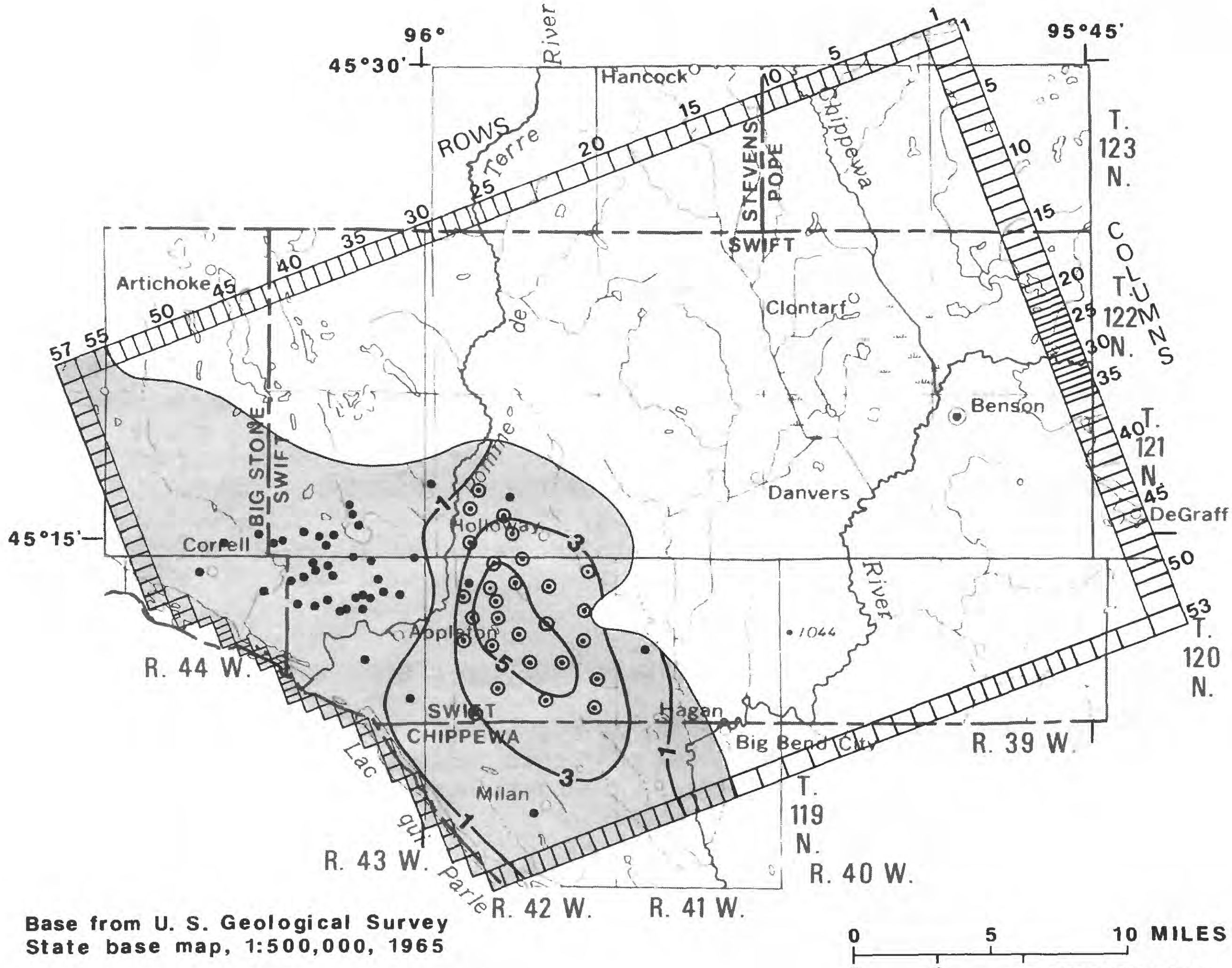


Figure 21.--Model-computed water-level declines in the Benson-middle aquifer following expanded development (simulation C)



EXPLANATION

(shaded)	APPLETON AQUIFER
(unshaded)	TILL
—3—	LINE OF EQUAL WATER-LEVEL DECLINE--Interval 2 feet
•	1982 PUMPING WELL
⊙	HYPOTHETICAL PUMPING WELL

Figure 22.--Model-computed water-level declines in the Appleton aquifer following expanded development (simulation C)

It must be emphasized that the hypothetical simulations were allowed to reach equilibrium conditions, whereas the ground-water system probably would not reach equilibrium during a given pumping season. Therefore, head differences between the surficial and Benson-middle aquifers during a given pumping season probably would be greater than the simulated differences.

Simulation D was designed as a worst-case scenario to simulate the effects of a drought plus the addition of hypothetical pumping centers as in simulation C. Model results indicate that water levels may decline between 3 and 5 ft regionally in the aquifers present in the Benson area. Water-level declines of as much as 11 ft are probable in the surficial aquifer and 13 ft in the Appleton aquifer southeast of Appleton (figs. 23-25). Some channels in which the surficial aquifer occurs southeast of Appleton may be dewatered as a result of the hypothetical stress on the ground-water system (fig. 23). Ground-water discharge to the Pomme de Terre and Chippewa Rivers would be 43 percent of 1982 conditions. Flow of the Pomme de Terre and Chippewa Rivers in the modeled area would be reduced by 21.7 and 7.9 ft^3/s, respectively, compared to 1982 conditions. Discharge to the Minnesota River would be 77 percent of 1982 conditions. Ground-water loss to ET would be 46 percent of 1982 conditions.

Model Application and Limitations

The ground-water-flow model is a practical tool for simulating response of the ground-water system to projected climatic conditions and proposed development schemes. However, the model necessarily is a simplification of a complex flow system. The accuracy of model results is limited by the accuracy of the data that describe aquifer and confining bed properties, areal recharge rates, ET rates, and boundary conditions. In addition, a different combination of input data could achieve the same results. Quantitative field data for river-bed leakage coefficient, amount of irrigation water return to the ground-water system, ground-water loss to ET, and additional geologic data would enhance the model. As additional data become available, the model should be modified and recalibrated to improve its accuracy.

Caution should be used in making ground-water management decisions based on the steady-state model simulations described in this report. Water-level declines computed for hypothetical simulations A through D represent average declines over grid blocks as large as 0.94 mi^2. Actual water-level declines in wells will differ from computed values, and declines in or near individual high-capacity pumping wells generally will be greater. Because simulations A through D are steady-state simulations, results do not reflect the seasonal effects of climatic and pumping stresses. Rather, the results represent the long-term effects of the stresses applied. Steady-state simulations do not consider water from storage, which may appreciably affect short-term water levels.

Model-calibration and transient-simulation results demonstrate that the model reasonably approximates operation of the ground-water system. The model can simulate regional and seasonal changes in ground-water flow and water levels with time. Therefore, the model can be used by state and local water-resource planners to evaluate the regional and seasonal effects of hypothetical

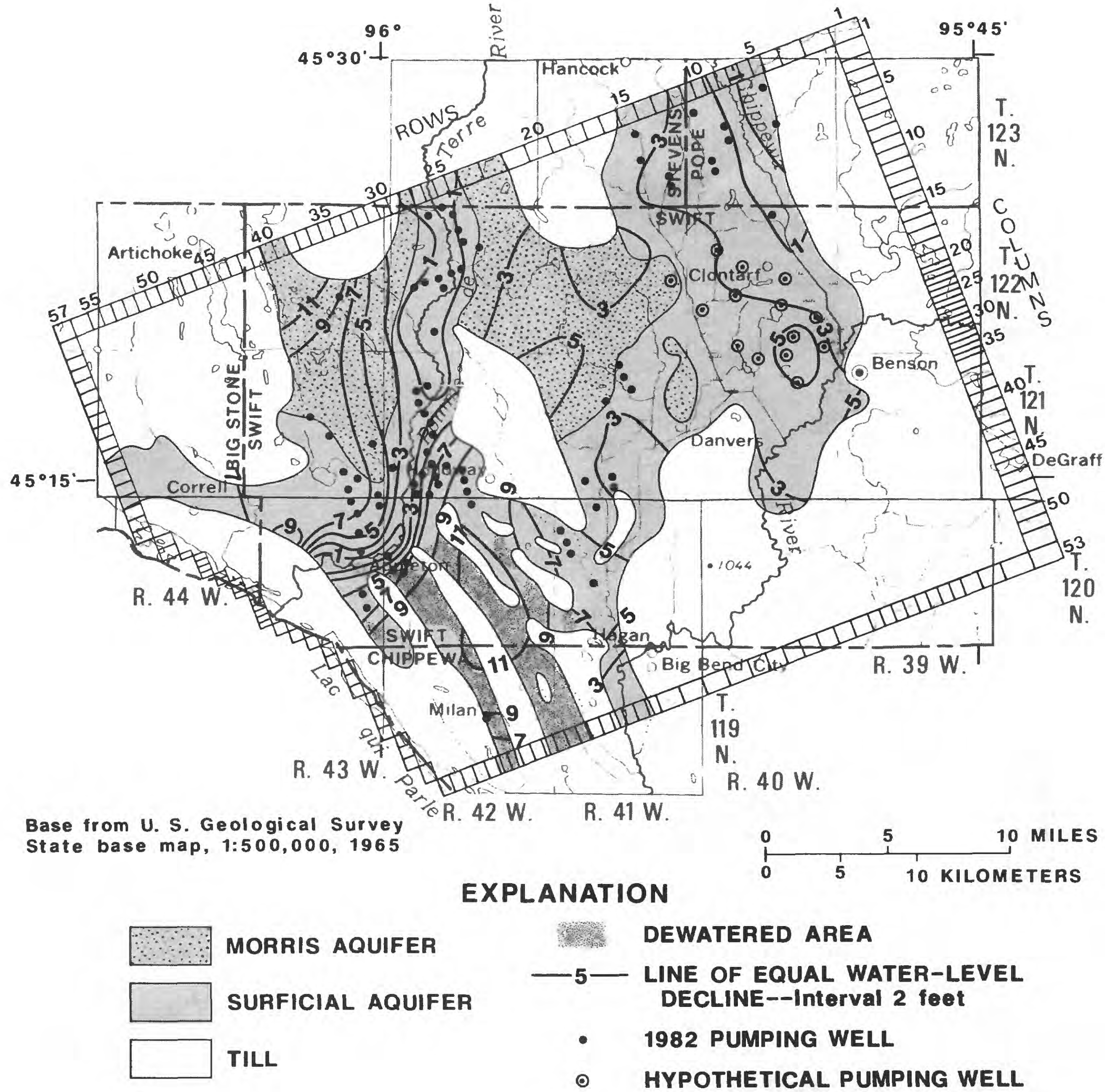
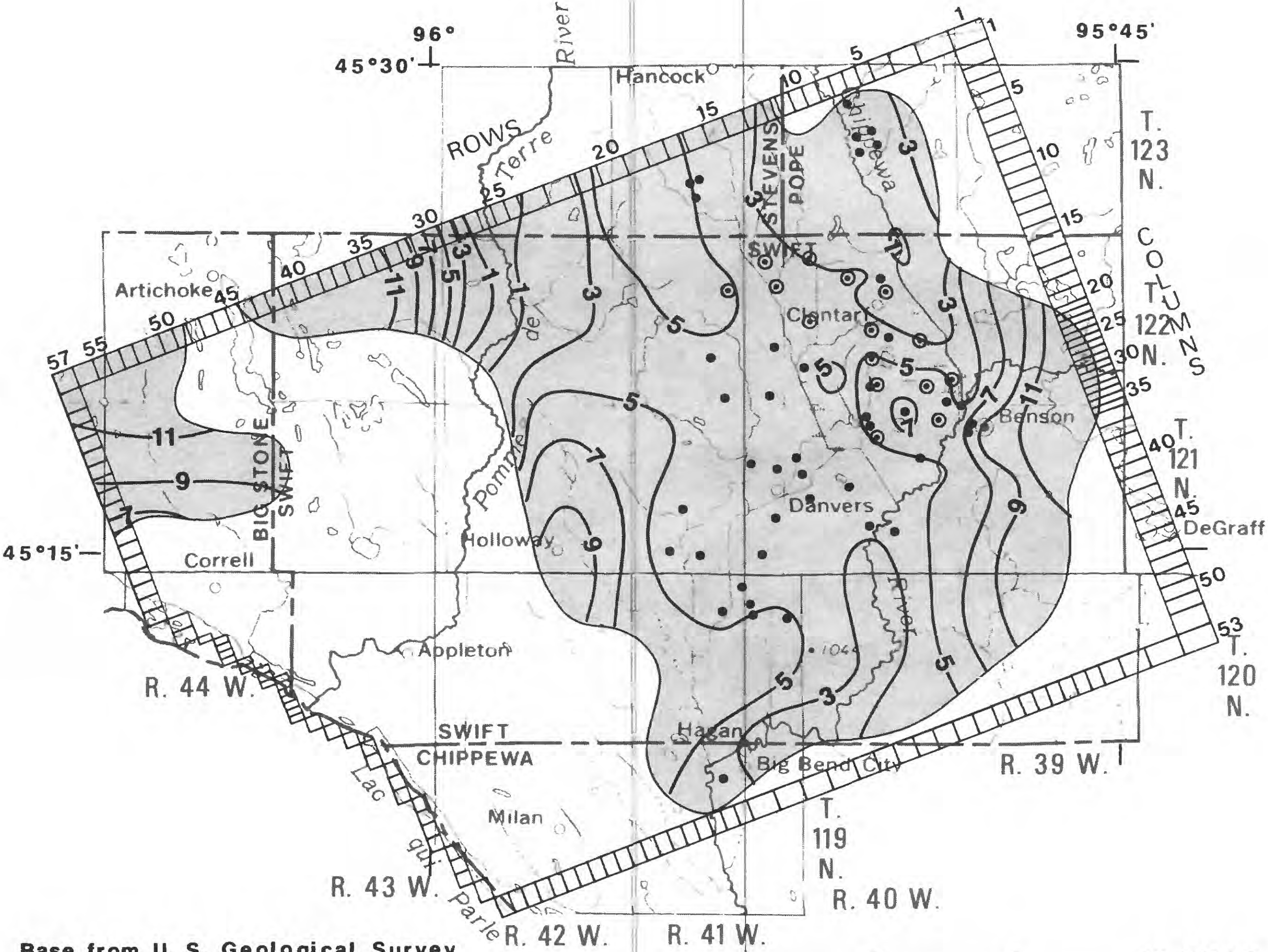


Figure 23.--Model-computed water-level declines in the surficial and Morris aquifers with expanded development following an extended drought (simulation D)



Base from U. S. Geological Survey
State base map, 1:500,000, 1965

0 5 10 MILES
0 5 10 KILOMETERS

EXPLANATION

BENSON-MIDDLE AQUIFER

TILL

—5— LINE OF EQUAL WATER-LEVEL DECLINE--Interval 2 feet

• 1982 PUMPING WELL

⊙ HYPOTHETICAL PUMPING WELL

Figure 24.--Model-computed water-level declines in the Benson-middle aquifer with expanded development following an extended drought (simulation D)

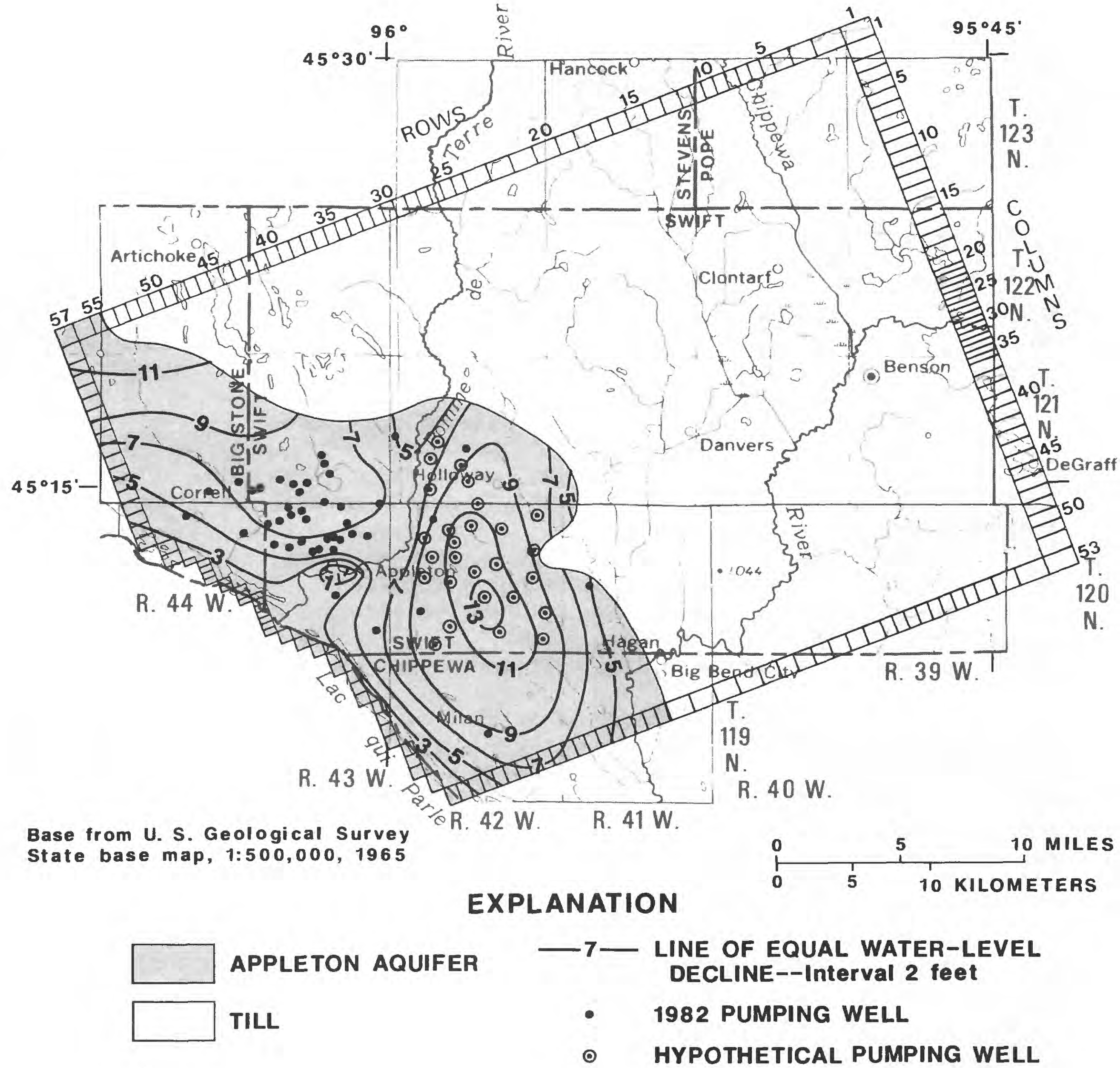


Figure 25.--Model-computed water-level declines in the Appleton aquifer with expanded development following an extended drought (simulation D)

changes in ground-water development and recharge. However, the scale of the model precludes ground-water analyses for a local area. A more detailed model would be necessary for site-specific analyses.

SUMMARY AND CONCLUSIONS

Ground-water withdrawals from drift aquifers have increased dramatically during the last decade in western Minnesota. The increase is due primarily to an increase in the ground water used for irrigation following the 1976-77 drought. The Appleton and Benson areas obtain their water supplies almost entirely from confined- and unconfined-drift aquifers. Management of this ground-water resource requires an understanding of the behavior of this complex aquifer system. Consequently, a ground-water-flow model was constructed to provide information on the regional behavior of the system.

Steady-state calibration of the model produced simulated water levels generally within 7 ft of measured values. Model-computed steady-state ground-water discharge to the Pomme de Terre and Chippewa Rivers matched observed rates generally within 20 percent. The model results indicate that 98 percent of the inflow to the model area is recharge from precipitation. Of the total outflow, 36 percent is evapotranspiration, 38 percent is ground-water seepage to the Pomme de Terre and Chippewa Rivers, 17 percent is ground-water pumpage, and 8 percent is ground-water discharge to the Minnesota River.

The model indicates that ground-water flow between the drift aquifers is considerable. The model computed steady-state ground-water flow of approximately 1,900 Mgal/yr from the surficial aquifer to the Benson-middle aquifer and 1,200 Mgal/yr from overlying drift aquifers to the Appleton aquifer. Approximately 500 Mgal/yr flows directly from the surficial aquifer to the Appleton aquifer where they coalesce north of Appleton.

Transient simulations demonstrated that the model accurately simulates seasonal changes in the ground-water-flow system. The model simulated water-level declines to within 15 percent of measured values during 1980-82. Model results indicate that 95 percent of the water pumped during the summer irrigation season is derived from storage in the aquifer.

Model results indicate that historical pumping has lowered water levels between 1 and 2 ft regionally in all confined and unconfined aquifers, and as much as 13 ft locally near Benson in the Benson-middle aquifer. Reduced recharge and increased pumping during a 3-year hypothetical drought may lower water levels an additional 2 to 6 ft regionally in each aquifer, compared to 1982 water levels, and as much as 11 ft locally near aquifer boundaries. Ground-water discharge to the Pomme de Terre and Chippewa Rivers in the modeled area would be reduced by 15.2 and 7.4 ft^3/s, respectively, during the simulated drought, compared to 1982 conditions.

Model simulations of hypothetical ground-water development indicate that the surficial aquifer and the Appleton and Benson-middle aquifers are capable of supporting additional pumping. Model results indicate that the addition of 30 hypothetical high-capacity wells near Benson, pumping a total of 810

Mgal/yr, would lower water levels about 1 ft regionally in the surficial and Benson-middle aquifers. Hypothetical pumping from only the Benson-middle aquifer would result in a maximum water-level decline of 2.7 ft in the aquifer compared to a maximum decline of 1.3 ft if the hypothetical pumping was simulated only in the surficial aquifer. The addition of 28 hypothetical wells in the Appleton aquifer east and southeast of Appleton, pumping a total of 756 Mgal/yr, would lower water levels in the surficial and Appleton aquifers as much as 5 ft.

Model results indicate that the addition of 58 hypothetical wells, mentioned above, plus the effects of a 3-year drought would result in regional water-level declines of 3 to 5 ft in the aquifers present in the Benson area. Water-level declines in the Appleton area would be greater, with as much as 13 ft of decline in the surficial and Appleton aquifers southeast of Appleton. The model-computed water-level declines are in addition to the historical declines that occurred prior to 1982.

The ground-water-flow model is a practical tool for use in simulating the ground-water system. However, the accuracy of model results is limited by the accuracy of the data that describe aquifer and confining-bed properties, recharge rates, ET rates, and boundary conditions. Caution should be used in making ground-water management decisions based on the steady-state simulations. Actual water-level declines in wells will differ from model-computed values, and declines in or near individual high-capacity pumping wells generally will be greater. As additional data become available, the model should be modified and recalibrated to improve its accuracy. The scale of the model precludes its use for ground-water analyses in local areas. More detailed models of local areas are needed for site-specific analyses.

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