DIGITAL SIMULATION OF GROUND-WATER FLOW IN THE WARWICK AQUIFER, FORT TOTTEN INDIAN RESERVATION, NORTH DAKOTA

By Thomas B. Reed

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DIGITAL SIMULATION OF GROUND-WATER FLOW IN THE WARWICK AQUIFER, FORT TOTTEN INDIAN RESERVATION, NORTH DAKOTA

By Thomas B. Reed

Abstract

The demand for water from the Warwick aquifer, which underlies the Fort Totten Indian Reservation in northeastern North Dakota, has been increasing during recent years. Therefore, the Spirit Lake Sioux Nation is interested in resolving questions about the quantity and quality of water in the aquifer and in developing a water-management plan for future water use. A study was conducted to evaluate the surface-water and ground-water resources of the Fort Totten Indian Reservation and, in particular, the ground-water resources in the area of the Warwick aquifer. A major component of the study, addressed by this report, was to define the ground-water flow system of the aquifer.

The Warwick aquifer consists of outwash deposits of the Warwick outwash plain that are as much as 30 feet thick and buried-valley deposits beneath the outwash plain that are as much as 200 feet thick. The aquifer is bounded on the north and west by end-moraine deposits and Devils Lake, on the south by the Sheyenne River Valley, and on the east by outwash deposits and ravines. The aquifer is underlain by Pierre Shale or by glacial till, clay, or silt. Ground-water gradients generally are small and rarely are more than 3 or 4 feet per mile. From 1982 to 1993, withdrawals from the Devils Lake well field averaged 1.5 cubic feet per second, and withdrawals from irrigation wells averaged 1.29 cubic feet per second. The combined discharge from springs may be about 3 cubic feet per second. During the early 1990s, the Warwick aquifer probably was in a steady-state condition with regard to storage change in the aquifer.

A finite-difference, three-dimensional, ground-water flow model provided a reasonable simulation of ground-water flow in the Warwick aquifer. The aquifer was divided vertically into two layers and horizontally into a grid of 83 by 109 cells, each measuring 656 feet (200 meters) per side. The steady-state simulation was conducted using 1992 pumpage rates and October 1992 water levels. The mean absolute difference between simulated and derived water-level altitudes during final calibration of the model was 1.52 feet. The two transient simulations were conducted for 20 time intervals of 1 year each using both the small and large storage estimates, doubled 1992 pumpage from the Devils Lake well field, 1992 irrigation pumpage, and initial water-level altitudes simulated by the October 1992 steady-state simulation. In the simulation using the small storage estimate and doubled pumpage, model cells in the area of the well field went dry after 13 years.

Assumptions made in the design of the model generally are supported by the digital simulation. Except in the area of Warwick Springs and smaller springs, lateral and basal boundaries of

the aquifer are impermeable. The flow system is dominated by recharge and evapotranspiration. Recharge rates obtained during the calibration process were lower in topographically high areas than in topographically low areas. Hydraulic conductivity in the area of the Devils Lake well field was larger than that in the rest of the aquifer.

INTRODUCTION

The Warwick aquifer (fig. 1) underlies the southeastern corner of the Fort Totten Indian Reservation in northeastern North Dakota and extends southward from just south of Devils Lake to the Sheyenne River Valley. The demand for water from the aquifer has been increasing during recent years. Since 1962, the aquifer has supplied water to the city of Devils Lake, which withdraws water from the Devils Lake well field located in the eastern part of the reservation south of Elbow Lake. The aquifer also supplies water to more than a dozen irrigation wells on the reservation, and the number of center-pivot irrigation systems is increasing. In addition, the Spirit Lake Sioux Nation is developing a municipal and rural water-supply system on the reservation, and production wells for this system are located near production wells used by the city of Devils Lake.

The Spirit Lake Sioux Nation is interested in resolving questions about the quantity and quality of water in the Warwick aquifer, including (1) whether the current or additional demand for water can be met and (2) the effect this demand will have on the quality of water in the aquifer. To develop a water-management plan for future water use, the ground-water flow system of the aquifer and the effects on the aquifer first needed to be evaluated. Therefore, from 1992 to 1995, the U.S. Geological Survey, in cooperation with the Spirit Lake Sioux Nation, conducted a study to evaluate the surface-water and ground-water resources of the Fort Totten Indian Reservation and, in particular, the ground-water resources in the area of the Warwick aquifer. A major component of the study was to define the ground-water flow system of the aquifer. The study included the collection and evaluation of well-log, borehole, aquifer-test, water-level, and land-surface altitude data.

Purpose and Scope

This report describes the ground-water flow system of the Warwick aquifer and presents the results of simulations of ground-water flow in the aquifer with the use of a numerical, three-dimensional, ground-water flow model. The report describes the areal extent, thickness, hydrologic properties, and boundaries of the aquifer and the direction of ground-water flow in the aquifer; identifies areas of ground-water recharge and discharge; and gives estimated rates of the recharge and discharge. Model simulations include steady-state conditions of the aquifer and transient conditions of the well field.

Previous Investigations

Paulson and Akin (1964) provided a specialized report on the Devils Lake area in Benson, Ramsey, and Eddy Counties that included a description of the Warwick aquifer. Bluemle (1965, 1973), Trapp (1966, 1968), Downey (1971, 1973), Randich (1971, 1977), Carlson and Freers (1975), Hutchinson (1977), and Hutchinson and Klausing (1980) provided substantial information about the aquifer in county ground-water studies reports. Pusc (1992a, 1992b) documented results of hydrologic investigations in the area of the aquifer and included basic water-level data, descriptions of the geologic and hydrologic setting, and a conceptual model of ground-water flow. Many of these reports and other specific studies provide information on hydraulic properties of the aquifer.

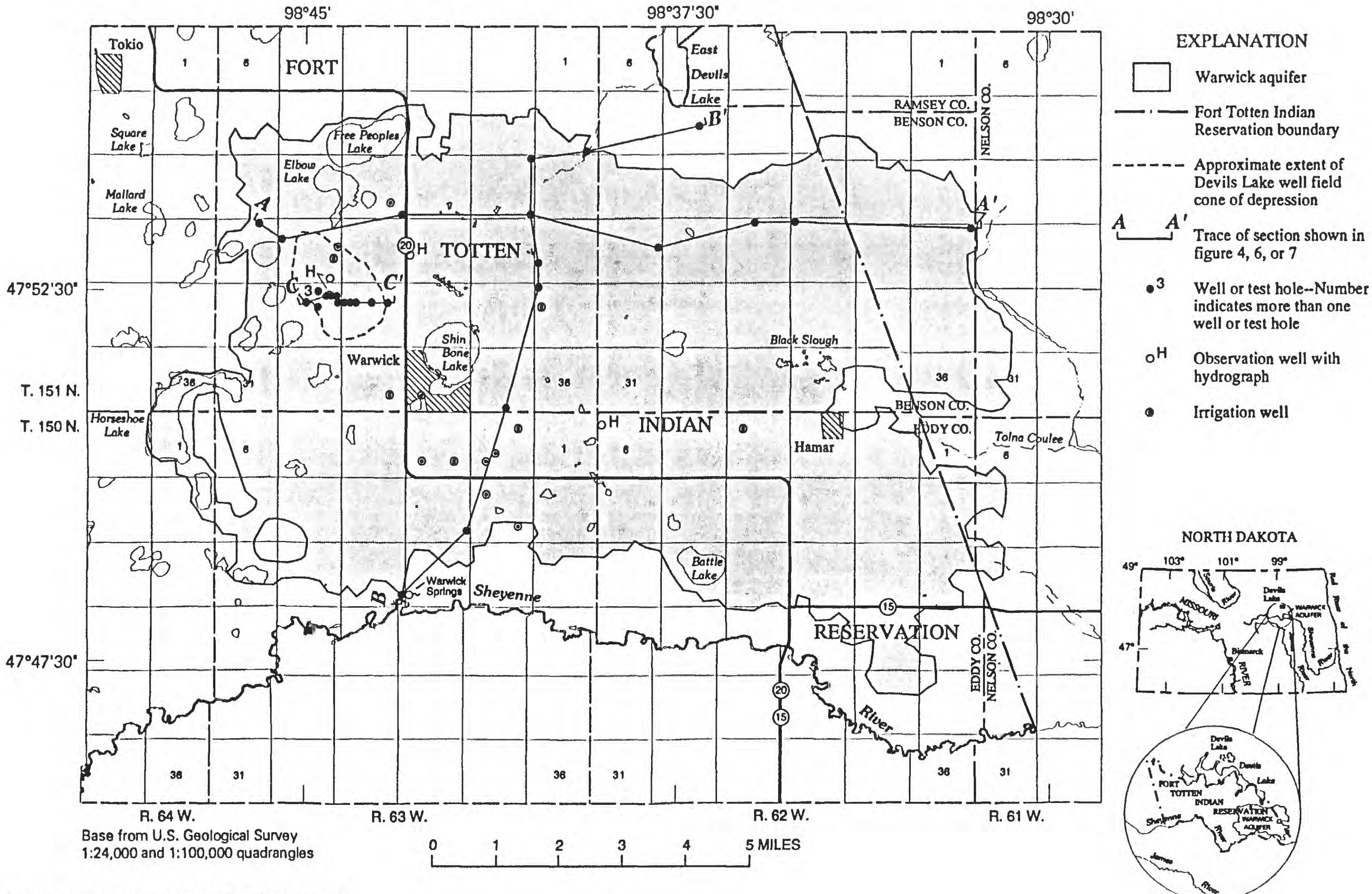


Figure 1. Location of the Warwick aquifer.

Location-Numbering System

The location-numbering system used to identify wells, test holes, and springs in this report (fig. 2) is based on the Federal system of rectangular surveys of the public lands. The first number denotes the township north of a base line, the second number denotes the range west of the fifth principal meridian, and the third number denotes the section in which the well, test hole, or spring is located. The letters A, B, C, and D designate, respectively, the northeast, northwest, southwest, and southeast quarter section, quarter-quarter section, and quarter-quarter-quarter section (10-acre tract); thus, well 150-063-16DAA is in the NE1/4NE1/4SE1/4 sec. 16, T. 150 N., R. 63 W. Consecutive terminal numbers are added if more than one well, test hole, or spring is located in the 10-acre tract.

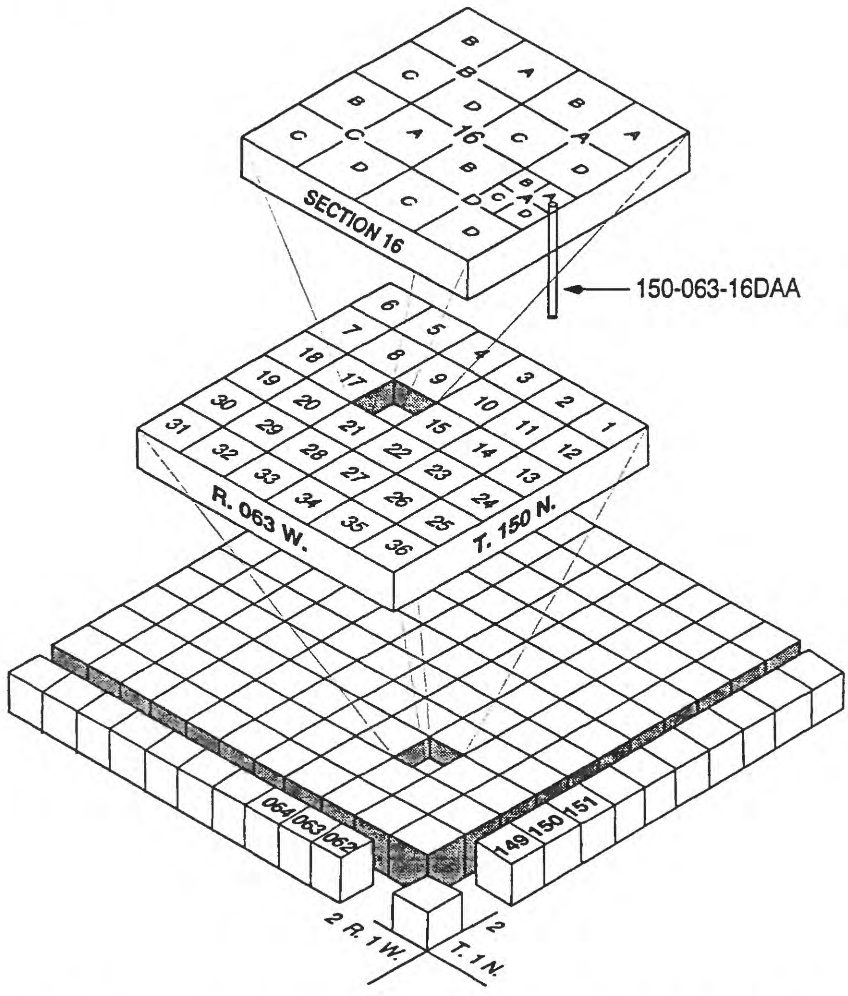


Figure 2. Township-range location-numbering system used to identify wells, test holes, and springs.

Sources of Data

Data on the thickness and extent of the Warwick aquifer were obtained from drillers' logs and test-hole logs published in county ground-water studies reports by Trapp (1966), Downey (1971), Randich (1971), and Hutchinson (1977). These reports were published cooperatively by the U.S. Geological Survey, the North Dakota Geological Survey, and the North Dakota State Water Commission. Data also were obtained from drillers' logs published in reports by Paulson and Akin (1964) and HKM Associates (undated and 1992). Unpublished data exist for about 1,500 boreholes augered by the Bureau of Reclamation in the Warwick and McVille aquifers. The McVille aquifer is located about 6 miles (mi) east of the Warwick aquifer. The boreholes in the Warwick aquifer were augered to a maximum depth of 30 feet (ft) or to the first occurrence of clay or glacial till, and the data were used to delineate much of the aquifer. Ground-water level data for wells were obtained from the U.S. Geological Survey Ground-Water Site Inventory (GWSI) system, and surface-water level data for perennial lakes were obtained from contours on topographic maps. Land-surface altitudes were obtained from topographic maps, unpublished Bureau of Reclamation data, well data stored in the GWSI system, and a 1993 global positioning system survey of altitudes of observation wells.

Acknowledgments

Personnel from the North Dakota State Water Commission, particularly Christopher Bader and Gordon Baesler, helped determine withdrawals from the Devils Lake well field. Arden Mathison and Kurt Webber from the Bureau of Reclamation provided estimates of recharge values used in the digital ground-water flow model.

DESCRIPTION OF STUDY AREA

Topography and Surficial Geology

The Warwick glacial-outwash plain is a generally rectangular area that extends about 8 mi from north to south and 12 mi from east to west. The land-surface altitude of the outwash plain is between about 1,460 and 1,470 ft above sea level. The land surface generally slopes to the southeast and is flat except for a few low hills or knobs of ground moraine. The surface relief is greater in the southeastern corner of the reservation where aeolian processes formed dunes from glacial-outwash sediments (fig. 3).

The Warwick outwash deposit is terminated on the south by the Sheyenne River Valley (figs. 1 and 3), which is incised about 60 ft lower than the outwash deposit. To the north and west, the outwash deposit generally is bordered by topographically higher end-moraine deposits (fig. 3), and, to the east, the outwash deposit is interrupted by deep coulees and ravines.

Surface drainage on the Warwick outwash deposit generally is undeveloped and consists of many wetlands and about 20 lakes. East of the outwash deposit, the surface is deeply dissected by coulees and ravines, some of which are incised about 20 ft into the surrounding material. These coulees and ravines direct surface drainage toward the southeast, eventually into the Sheyenne River Valley.

Hydrogeologic Setting

The Warwick aquifer (fig. 1) is an unconfined glacial aquifer recognized surficially as the approximately 85 square miles (mi^2) of outwash, dune, and alluvial deposits located north of the Sheyenne River on the eastern end of the Reservation (fig. 3). The aquifer typically is 20 to 30 ft thick but thickens where the underlying bedrock is channelized. Cretaceous Pierre Shale underlies the Warwick aquifer and is exposed in parts of the Sheyenne River Valley. The Pierre Shale was deeply incised by various glacial events and, therefore, provides a highly irregular basal surface for the overlying glacial units. Fractures

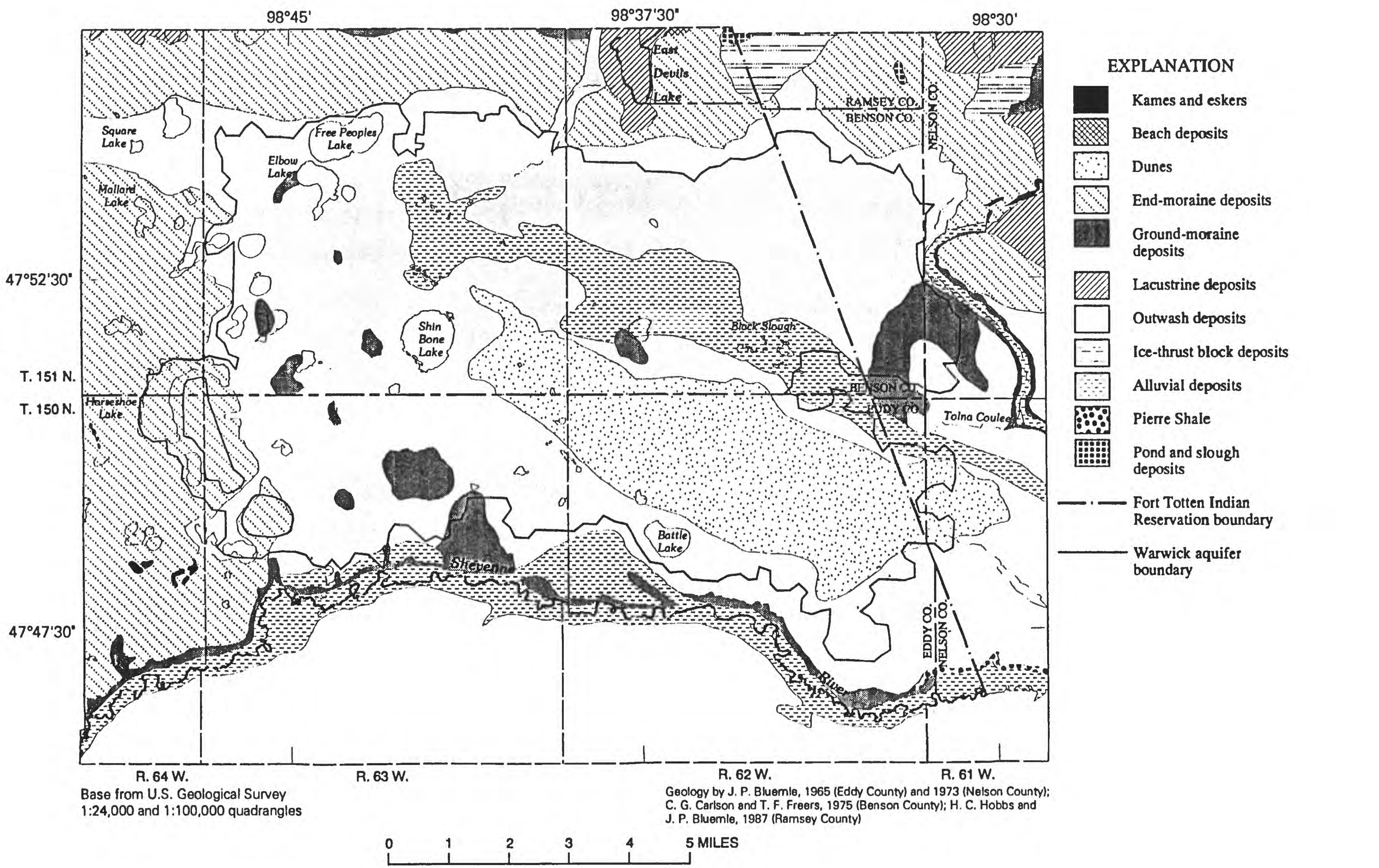


Figure 3. Surficial geology of the Warwick aquifer and surrounding area.

and weathering have made the Pierre Shale slightly permeable in places, and the shale supplies water for domestic use in areas where glacial aquifers are missing.

The geologic history of the Warwick aquifer is complex and consists of several sequential events in the glacial history of North Dakota. Glacial sediments of variable thickness overlie the Pierre Shale and represent numerous stages of glacial advance and retreat. Reconstruction of the glacial sequences is difficult because of the complexity of glacial-sedimentation processes and partial or complete removal of previous deposits by subsequent glacial advances. The primary glacial sediments associated with the Warwick aquifer are end- and ground-moraine deposits, buried-valley deposits, and outwash deposits.

End-moraine deposits generally are ridges of poorly sorted boulders, gravels, sands, silts, and clays (tills). The deposits, which are transverse to glacial-flow directions, represent margins of glacial advance or stagnation during general retreat and generally are low-yield water-bearing units. The end-moraine deposits located along the northern and western boundaries of the Warwick aquifer (fig. 3) consist mainly of local materials with some transported materials. The deposits have low permeability because of the poorly sorted fine materials and, therefore, act as lateral boundaries or confining units for the aquifer. End-moraine deposits that consist of till with local concentrations of sand and gravel probably represent the initial stages in the formation of the aquifer. The deposits, acting as dams, may have retained meltwater from retreating glaciers and prevented the meltwater from entering drainage valleys. This probably explains why glacial-lacustrine sediments are common along end-moraine deposits. The buildup and subsequent release of meltwater probably account for the buried-valley and outwash deposits that compose the Warwick aquifer.

Ground-moraine deposits are till deposits released directly from melting ice. These deposits, which consist of poorly sorted clay, sand, gravel, and silt, generally are low-yield water-bearing units.

Buried-valley deposits, which consist primarily of medium to coarse gravel with till or clay lenses, occur in many places within the Warwick aquifer. The deposits are formed by many processes, including fluvial deposition in preglacial valleys, glacial re-advance over glaciofluvial (esker) sediments, and catastrophic flooding from glacial lakes. Such flooding incises channels into the surrounding glacial sediments or into the underlying bedrock. The process whereby these incised channels fill with sediments of variable permeability is complex and is discussed by Kehew and Boettger (1986), Shaver and Pusc (1992), and others. Pockets of permeable sediments can occur along the sides of the channels from alluvial-fan fill and can be isolated by low-permeability sediments and bedrock. The processes probably account for the thick buried-valley deposits that occur in parts of the Warwick aquifer, such as in the area of the Devils Lake well field. Buried-valley deposits near the well field are as much as 200 ft thick. Other narrow channels have been detected within the aquifer, and many more channels may exist. The Spiritwood aquifer, a very large system of high-permeability sediments that underlie the eastern part of the Warwick aquifer, consists of buried-valley deposits (fig. 4).

The surficial part of the Warwick aquifer consists primarily of outwash deposits that are bordered by alluvial deposits on the south and end-moraine deposits on the north and west. The outwash deposits generally are between 20 and 30 ft thick and are formed as outwash aprons just downslope from a melting glacier, as alluvial fans resulting from the release of water from ice- or moraine-dammed glacial-lacustrine settings, as deposits from braided rivers that drain meltwater from the margin of the glacier, and as sediment in or on stagnant or dead-ice blocks. The deposits consist of fine to coarse sand and gravel and often are interfingered with or capped by till deposits (often shown in figure 3 as ground-moraine deposits). Dune and alluvial deposits also occur in the surficial part of the Warwick aquifer. Buried-valley deposits form the lower part of the aquifer in some areas, as reflected by areas of greater thickness (fig. 5).

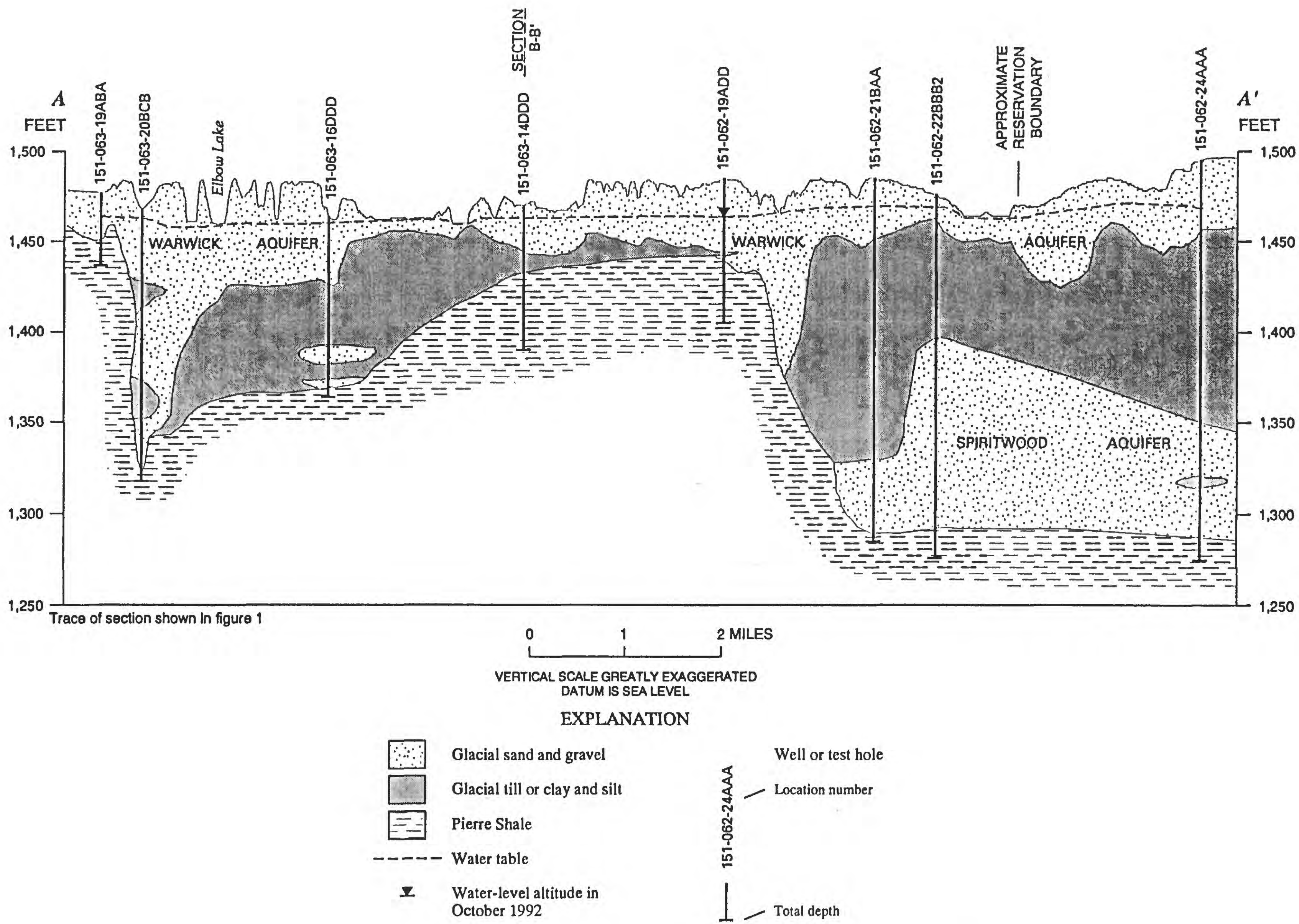


Figure 4. Geohydrologic section A-A' through the Warwick aquifer, the Spiritwood aquifer, and the Pierre Shale.

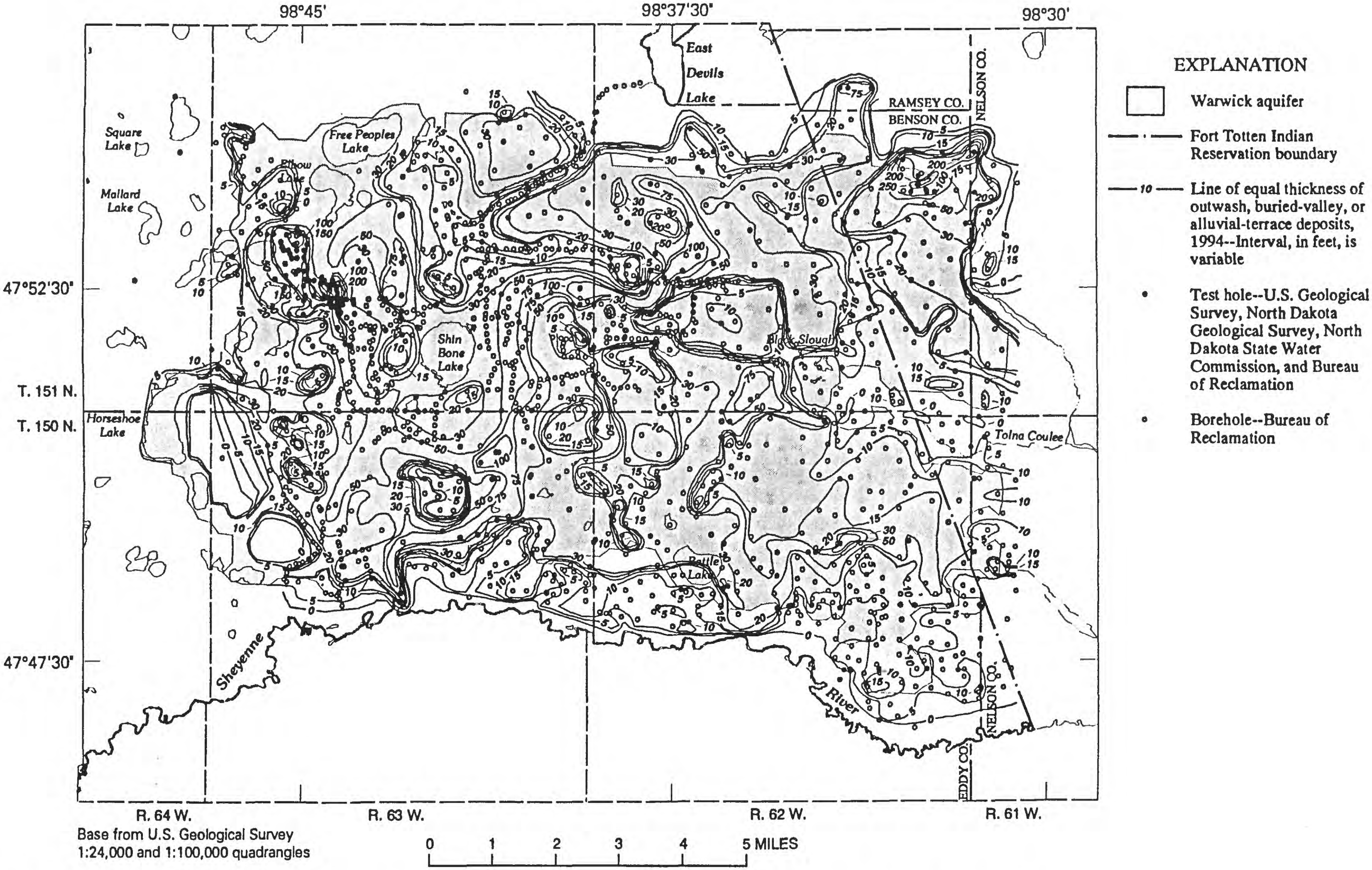


Figure 5. Thickness of outwash, buried-valley, and alluvial-terrace deposits as determined from borehole and test-hole data.

Geohydrologic sections through the Warwick aquifer from west to east and south to north are shown in figures 4 and 6, and a geohydrologic section through the area of the Devils Lake well field is shown in figure 7. Traces of the sections are shown in figure 1. The glacial sand and gravel shown in the figures includes outwash, buried-valley, dune, and alluvial deposits.

CONCEPTUAL HYDROLOGIC MODEL OF THE WARWICK AQUIFER

Aquifer Boundaries

The Warwick aquifer is bounded on the top by the water table, including perennial lakes that probably reflect the position of the water table; on the north and west by topographically higher end-moraine deposits; on the south by the Sheyenne River Valley; and on the east by topographically lower, dissected outwash deposits (fig. 3) and ravines. The aquifer is underlain by Pierre Shale or by glacial till, clay, or silt. The geometry of the aquifer is illustrated by its surficial extent (figs. 1 and 3) and by its thickness (fig. 5).

The water-level altitude in the Warwick aquifer in October 1992 is shown in figure 8. The altitude was calculated from ground-water level data for 54 observation wells measured in October 1992, land-surface altitude and water-level altitude data obtained from U.S. Geological Survey topographic maps, and land-surface altitude data obtained from Bureau of Reclamation land-surface contours. The land-surface altitude data were used to establish upper limits for the water-level altitudes. The water table in the aquifer is relatively shallow. The mean depth to water for the period of record for 87 observation wells is 12.92 ft below land surface. During 1992, precipitation was less than normal, and the mean depth to water for the 54 observation wells was 1.15 ft greater than the mean depth to water for the period of record for each of the 54 wells. The depth to water in the Warwick aquifer in October 1992 (fig. 9) was determined using computer-interpolated values for both land-surface and water-level altitudes. The depth to water was 10 ft or less in 70 percent of the aquifer and 5 ft or less in 41 percent of the aquifer.

Lateral boundaries of an aquifer may be determined on the basis of several criteria. An aquifer may thin out or become truncated by material of much lower permeability; a ground-water divide may represent a plane of no flow in a ground-water flow system; or a recharge or discharge area, such as a stream or lake, may constrain the ground-water altitude to a stable, relatively fixed value. The boundaries of the Warwick aquifer were determined on the basis of a combination of these conditions.

The northern boundary of the Warwick aquifer is in contact with topographically higher end-moraine deposits. These deposits are bordered farther north by Devils Lake, which has a water level of about 35 to 60 ft less than the water table along the northern boundary of the aquifer. The difference probably indicates highly restricted hydraulic connection between the aquifer and the lake. Water-table gradients indicate no regional flow northward from the middle of the aquifer toward the Devils Lake chain of lakes although some local flow occurs at the northern margin of the aquifer, and a general flow occurs southward toward the topographically lower areas in the middle of the aquifer. In addition, drillers' logs and borehole data indicate a thinning of the aquifer along parts of the northern boundary. Therefore, a ground-water divide across which no flow occurs is considered to represent the northern boundary of the aquifer. Because outwash deposits occur north of this ground-water divide, the boundary of the aquifer is south of the northern extent of outwash as shown in figure 3.

The southern boundary of the Warwick aquifer is formed by the Sheyenne River Valley. Water-level altitudes and historic surface-water levels in those parts of the aquifer adjacent to the Sheyenne River

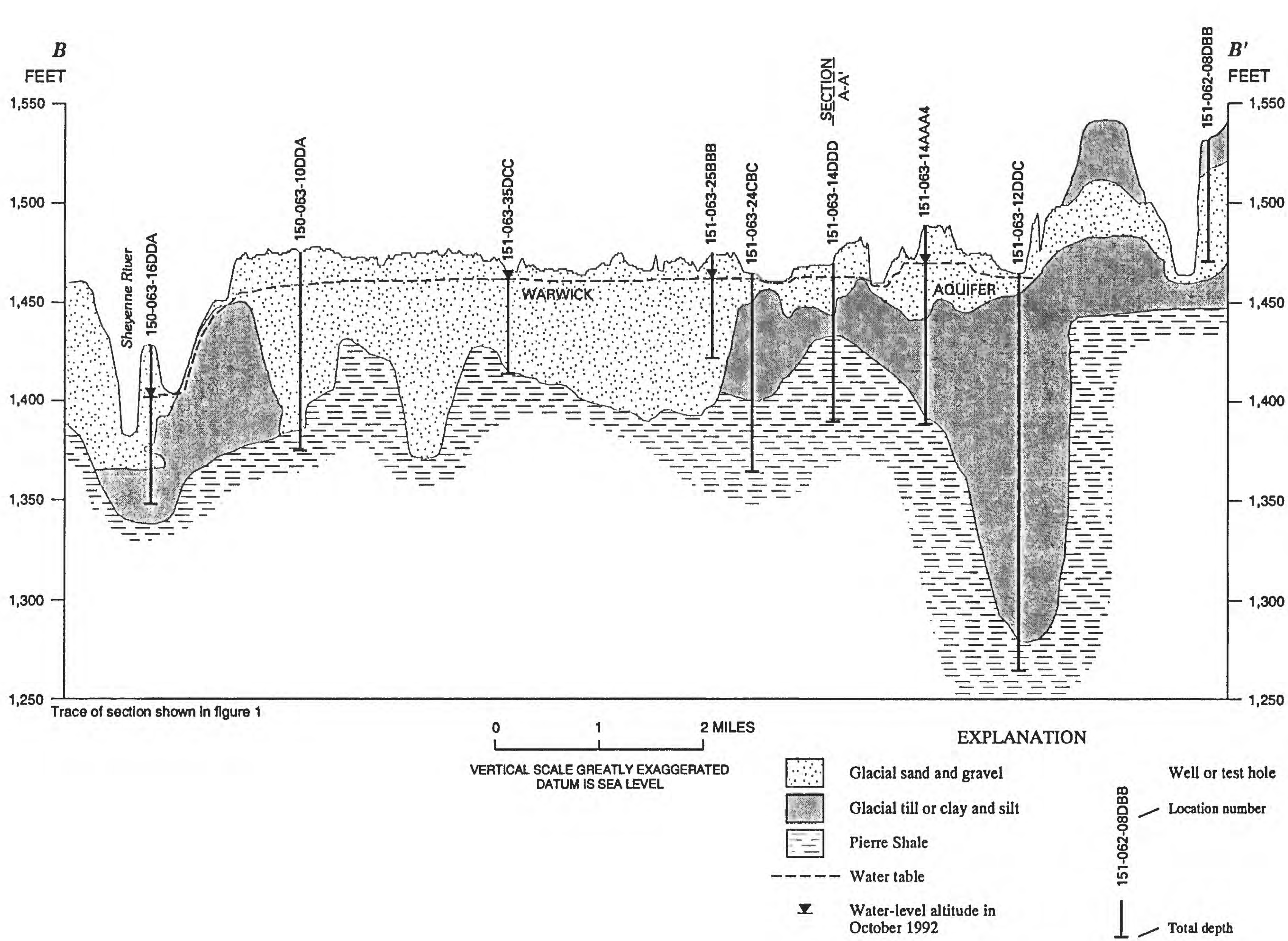


Figure 6. Geohydrologic section B-B' through the Warwick aquifer and the Pierre Shale.

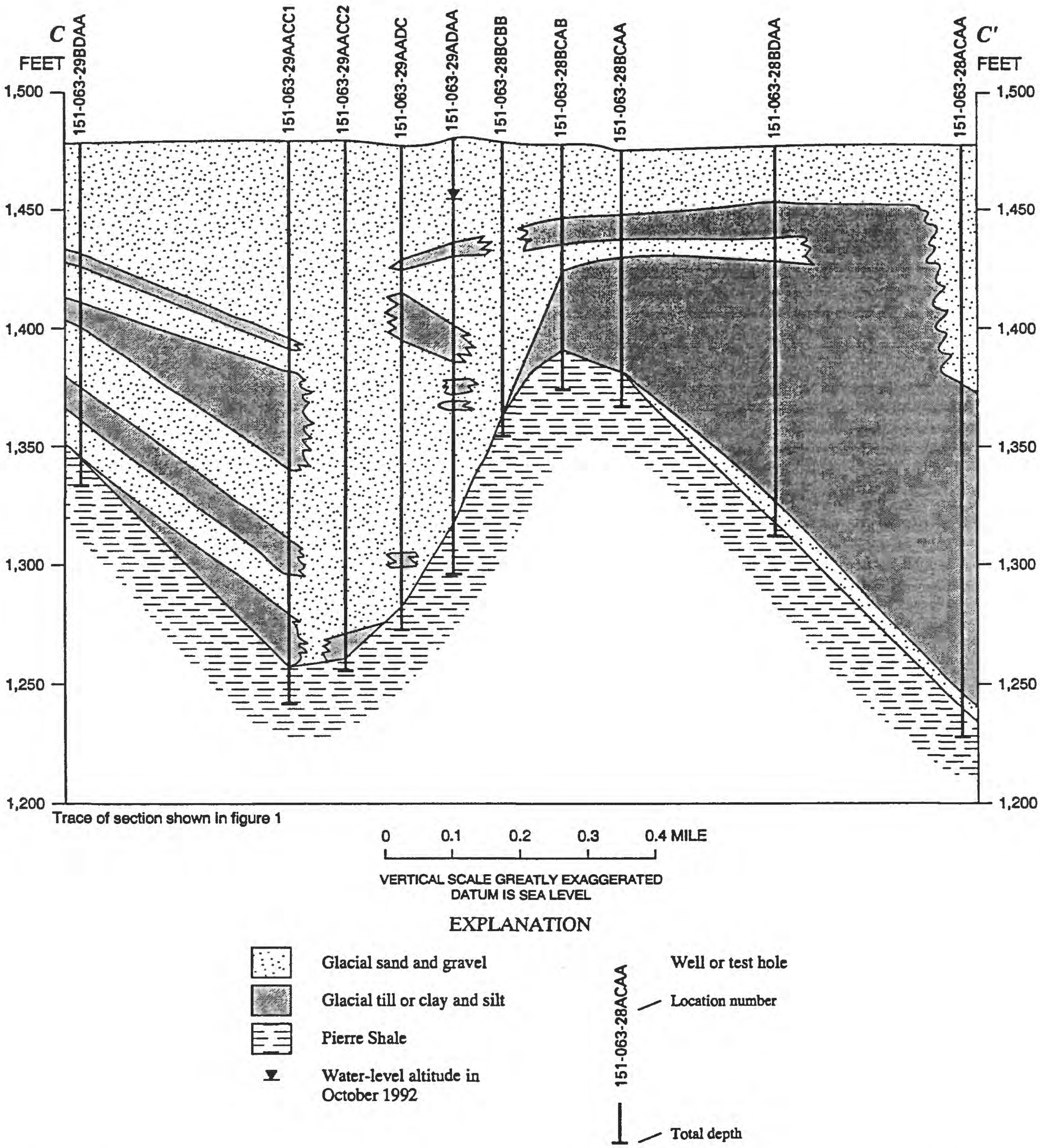


Figure 7. Geohydrologic section C-C' through the Warwick aquifer and the Pierre Shale in the area of the Devils Lake well field.

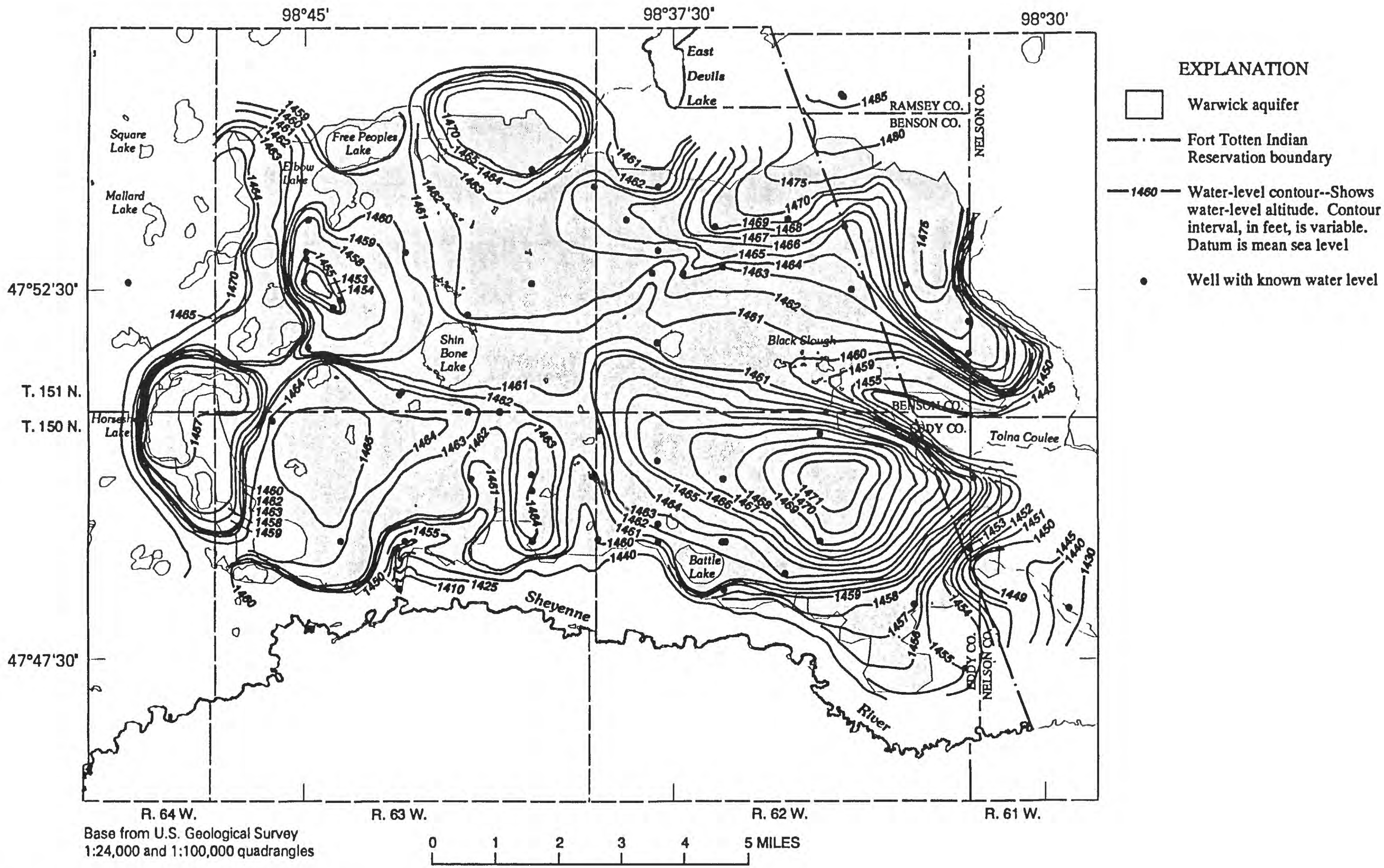


Figure 8. Water-level altitude in the Warwick aquifer in October 1992.

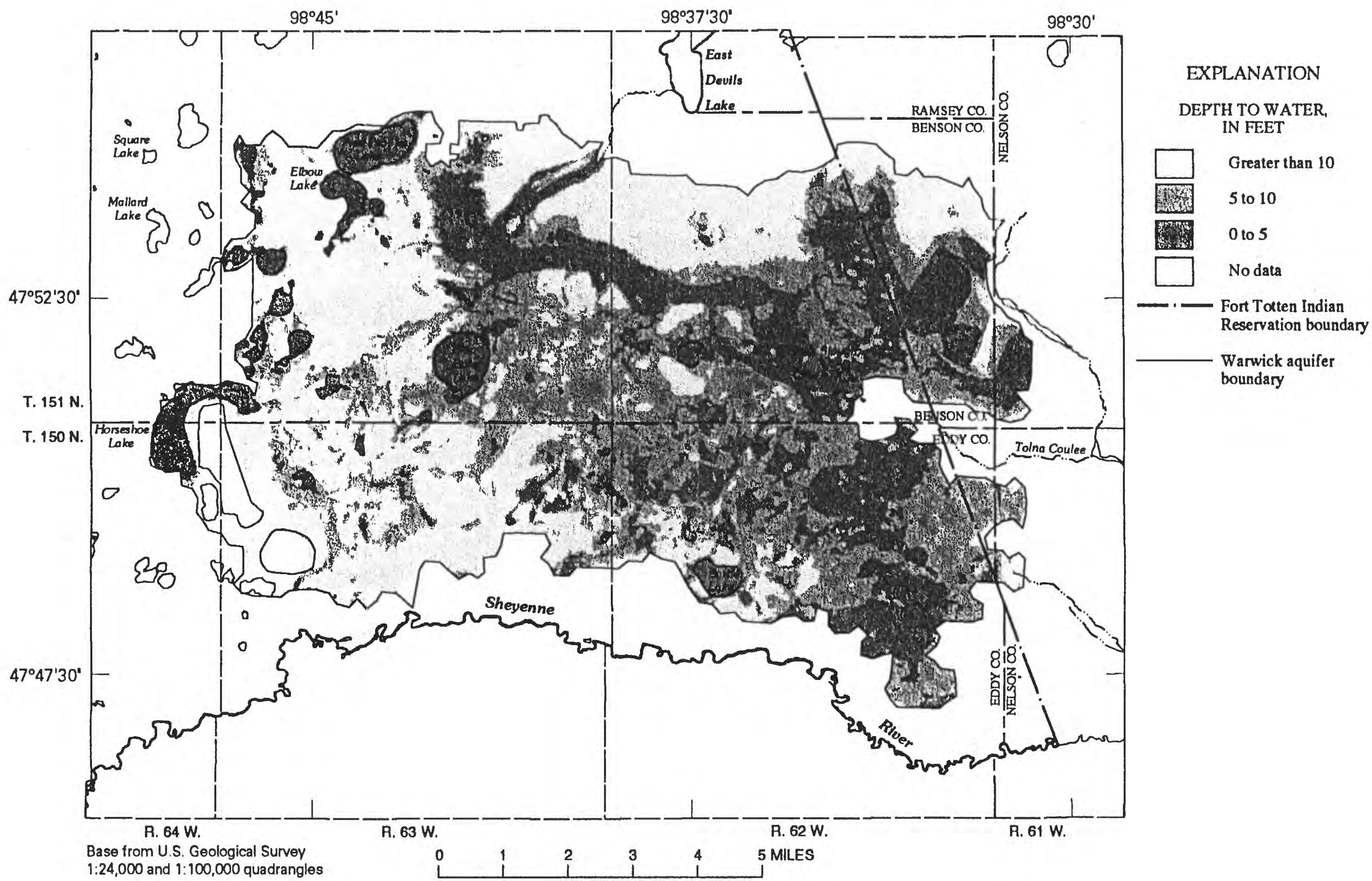


Figure 9. Depth to water in the Warwick aquifer in October 1992.

Valley are about 60 ft higher than the altitude of the valley floor. This difference, along with little ground-water flow from the aquifer southward, reflects the presence of low-permeability material, such as clay or glacial till, that separates the aquifer from the alluvium of the Sheyenne River Valley, in contrast to what is shown in figure 3. Well-log and borehole data indicate that the aquifer thins just north of the Sheyenne River Valley. The saturated thickness of the aquifer also thins farther north of the southern extent of outwash deposits. Therefore, the southern boundary of the aquifer is north of the southern extent of outwash as shown in figure 3. Warwick Springs is the only known location where the aquifer is in direct contact with the Sheyenne River alluvium (fig. 1). Streamflow measurements on the Sheyenne River in October 1986 indicated that about 3.1 cubic feet per second (ft^3/s) of ground-water discharge entered the river within the reach where the river valley is bordered by the aquifer. Warwick Springs is the major source of ground-water discharge from the aquifer to the Sheyenne River Valley. Along the rest of the southern boundary, ground-water flow from the aquifer southward probably is restricted by thin, low-permeability deposits. Therefore, the quantity of ground-water flow to the south is minimal except from Warwick Springs and smaller springs that serve as drains from the aquifer.

The eastern boundary of the Warwick aquifer may be in contact with thin, unsaturated, sand deposits (apparent in some well logs) that are 5 to 10 ft thick. These thin deposits are largely truncated by Tolna Coulee and other deep ravines located just beyond the eastern boundary. The large water-table gradients along the eastern boundary (fig. 8) reflect the large difference between the water-level altitudes in the aquifer and the land-surface altitude to the east. Although numerous small springs probably exist along the eastern boundary and indicate some ground-water flow to the east, the eastern boundary of the aquifer probably transmits minimal flow in most areas.

The western boundary of the Warwick aquifer is not directly known, but borehole data generally indicate a thinning in the subsurface toward the west. Borehole, water-level, and land-surface altitude data indicate that the saturated thickness in the aquifer becomes greatly constricted toward the west of its surficial expression. Immediately west of the surficial expression of the aquifer, the land-surface altitude rises about 60 or 70 ft. Perennial lakes and ponds that occur to the west have surface-water levels that typically are about 60 ft higher than the water-level altitudes in the aquifer. The large hydraulic-head differences between the aquifer and the area immediately west of the aquifer contrast with the much smaller gradients in the central part of the aquifer. This, along with the trend of thinner deposits toward the west, suggests that ground-water flow from surficial deposits to the west is much less than ground-water flow within the aquifer, and little hydraulic connection occurs with deposits to the west.

The Warwick aquifer generally is underlain by Pierre Shale or by glacial till, clay, or silt that overlies the Pierre Shale (fig. 6). As shown in geohydrologic section A-A' (fig. 4), the aquifer is underlain directly by Pierre Shale in the west and by glacial till or clay and silt in the east. The section also shows that, in the east, the Spiritwood aquifer overlies the Pierre Shale and is separated from the Warwick aquifer by the glacial till or clay and silt. Ground-water levels in wells in the Spiritwood aquifer generally are about 10 ft lower than those in nearby wells in the Warwick aquifer. The ground-water levels in wells in the two aquifers are known to be about the same only in and near T. 151 N., R. 62 W., secs. 19 and 20. Also, the water levels in wells in the Warwick aquifer are not lower in the areas of possible contact between the two aquifers, such as near T. 151 N., R. 62 W., secs. 19 to 22, indicating minimal hydraulic connection with the Spiritwood aquifer. The water quality of the Spiritwood aquifer is highly variable (R.M. Lent, oral commun., 1996) and generally is inconclusive with regard to any possible hydraulic connection between the Warwick and Spiritwood aquifers. Therefore, although drillers' logs indicate some localized contact may exist, the contact is poorly defined and will be disregarded in this report. Although the Pierre Shale can act as a low-yield aquifer in places and can provide yields of 1 to 10 gallons per minute (gal/min) (Randich, 1977, p. 16), the shale generally has much lower permeability than the Warwick aquifer, and the bottom of the aquifer probably can be considered a no-flow boundary.

Aquifer Properties

Hydraulic conductivities in the upper 30 ft of the Warwick aquifer were estimated by the Bureau of Reclamation during irrigation drainage studies (Arden Mathison, oral commun., 1994). On-site estimates of horizontal and vertical hydraulic conductivity in specific lithologic units were made at selected sites in the aquifer. Horizontal hydraulic conductivities were obtained from slug tests, and vertical hydraulic conductivities were measured using ring permeameters. The conductivities then were extrapolated to the other boreholes by using lithologic correlation. Horizontal hydraulic conductivities in the upper 30 ft of the aquifer ranged from nearly zero to 100 feet per day (ft/d). Additional data from aquifer tests (Randich, 1977, p. 47) indicate transmissivities that ranged from 6,300 to 20,600 feet squared per day (ft^2/d) in screened intervals of sand and gravel. Given the reported length of the screened intervals, these transmissivities would correspond to hydraulic conductivities that range from 520 to 4,100 ft/d. If ground-water flow from the full saturated thickness is assumed, the hydraulic conductivities from screened intervals would be lower. Actual hydraulic conductivities in screened intervals of sand and gravel probably are at least 1,000 ft/d. Particle-size distributions of the aquifer material indicate a mean porosity of 42 percent and estimated hydraulic conductivities that range from 21 to 220 ft/d (Randich, 1977, p. 45). Randich (1977, p. 45) reported that the specific yield of the Warwick aquifer is high, but no specific value was given. Downey (1973, p. 33) reported an estimated specific yield of 0.15 for the McVille aquifer.

Ground-Water Flow

Ground-water flow patterns in the Warwick aquifer (fig. 8) indicate no clear regional trend. Water-table gradients indicate that ground-water movement generally is from topographically higher areas of the aquifer toward topographically lower areas within and outside the aquifer. Gradients generally are small and rarely are more than 3 or 4 feet per mile (ft/mi). A large cone of depression extends outward to a radius of about 1 mi around the Devils Lake well field. A low area in the water table also occurs within several miles northeast of Warwick Springs and probably is caused by flow from the springs. Except where withdrawals from the aquifer channels are large, such as in the Devils Lake well field, the water-table gradients are similar to the overlying land-surface gradients.

Recharge and Evapotranspiration

Natural recharge to the Warwick aquifer occurs as direct infiltration of precipitation or snowmelt. Irrigation return is likely to add a small component to the recharge because of the porous nature of the aquifer and the generally small depth to water. Seasonal variations occur in evapotranspiration, recharge, and water-level altitudes and also may occur in water-table gradients in the aquifer. Areas that are recharged by spring snowmelt may become areas of discharge during the summer because of large evapotranspiration rates. Water enters the aquifer through downward percolation and exits the aquifer primarily as evapotranspiration from the shallow water table and as evaporation from the numerous lakes and ponds that receive ground-water discharge from the aquifer. The relation between precipitation and recharge is demonstrated in the rise in ground-water levels that occurs during the spring recharge period. Recharge in a particular area is controlled by the topography as well as by the soil characteristics of the area. During recharge events, recharge is greater in topographically low areas than in topographically high areas. In topographically low areas, surface drainage flows into ponds, lakes, and potholes that intersect the water table or overlie the aquifer, and a focused downward percolation exists. In topographically high areas, surface drainage is directed away and, thus, some precipitation becomes a part of overland flow. The Bureau of Reclamation conducted studies on the Warwick aquifer to provide deep-percolation ground-water recharge values for various soil types during given climatological conditions (Bureau of Reclamation, unpub. data, 1976, and written commun., 1994). These studies give an area-weighted average recharge of 4.5 inches per year (in/yr) during average precipitation conditions for soils overlying the Warwick aquifer. The area-weighted values range from zero for a dry year to about 8.3 in/yr for a wet year, such as 1965.

The areal distribution of evapotranspiration from an aquifer is a function of depth to the water table, potential evapotranspiration (meteorological conditions and plant physiology), and maximum depth at which evapotranspiration occurs. The depth to water in the Warwick aquifer is shown in figure 9. Potential evapotranspiration is the amount of water that would be removed from the land surface by evaporation and transpiration if unlimited water was available in the soil. Potential evapotranspiration serves as the upper limit for actual evapotranspiration and is assumed to be equal to the surface evaporation from a small body of water. The maximum actual net evapotranspiration over long time periods can be estimated as evaporation from the water table, when at land surface, minus average precipitation. The estimated average long-term surface evaporation for eastern North Dakota is 34 in/yr (Farnsworth and others, 1982, map 3), and the estimated average normal annual precipitation is 17.71 inches (in.) (U.S. Department of Commerce, National Oceanic and Atmospheric Administration, 1992). Thus, the average maximum potential net loss to evapotranspiration for eastern North Dakota is 16.29 in/yr. The maximum depth at which evapotranspiration occurs is the approximate depth of the base of the rooting zone for the overlying vegetation. The vegetation overlying the Warwick aquifer is primarily native and introduced grass. Maximum rooting depths for the grass *Agropyron smithii* vary from 4.9 to 12 ft in the United States and from 1.9 to 4.9 ft in the moister Saskatchewan prairies (Fitter and Hay, 1987, p. 157).

Spring Discharge and Well Pumpage

Other than evapotranspiration, the major ground-water discharge from the Warwick aquifer is Warwick Springs, a natural spring located southwest of the Warwick outwash plain at 150-063-15CCA. The altitude of Warwick Springs is about 60 ft lower than water-level altitudes about a mile to the north in the Warwick aquifer. Outflows for Warwick Springs were measured at roughly monthly intervals from October 1970 to June 1983. The measured outflows ranged from 0.63 to 2.70 ft^3/s and had a mean of 1.33 ft^3/s. The large variation in discharge indicates that much of the ground-water discharge to Warwick Springs may be augmented by sources other than the Warwick aquifer, such as seasonal recharge to terrace deposits located below the Warwick aquifer near the springs, or by overland surface-water flow during precipitation events. The altitude difference between water levels in the aquifer and water levels in Warwick Springs remains fairly stable and rarely varies by more than a few feet. Streamflow measurements on the Sheyenne River in October 1986 (Harkness and others, 1988, p. 356-357) indicated that, in addition to the flow from Warwick Springs, about 1.5 ft^3/s of ground-water discharge entered the river from areas adjacent to the Warwick aquifer. The ground-water discharge is either from the aquifer or from the south bank of the Sheyenne River. This, along with the occurrence of vegetation along the southern and eastern boundaries of the aquifer, indicates that smaller springs also serve as discharge points from the aquifer. The combined discharge from springs may be about 3 ft^3/s.

Ground water from the Warwick aquifer is used for municipal, irrigation, farm, ranch, and domestic purposes. The aquifer is the source of water for the city of Devils Lake, which withdraws water from the Devils Lake well field, and for at least 15 irrigation wells. Withdrawals from the Devils Lake well field, which represents the area of most of the pumpage from the aquifer, began in 1962 (Randich, 1977, p. 49). Records of pumpage from the well field were obtained from the North Dakota State Water Commission (Christopher Bader, oral commun., 1994), and pumpage for 1976-93 is shown in figure 10. Pumpage varied from about 1.17 ft^3/s in 1976 to about 1.83 ft^3/s in 1993. Records of pumpage from irrigation wells also were obtained from the North Dakota State Water Commission, but withdrawals from the Spirit Lake Sioux Nation irrigation system are not recorded. From 1982 to 1993, withdrawals from the Devils Lake

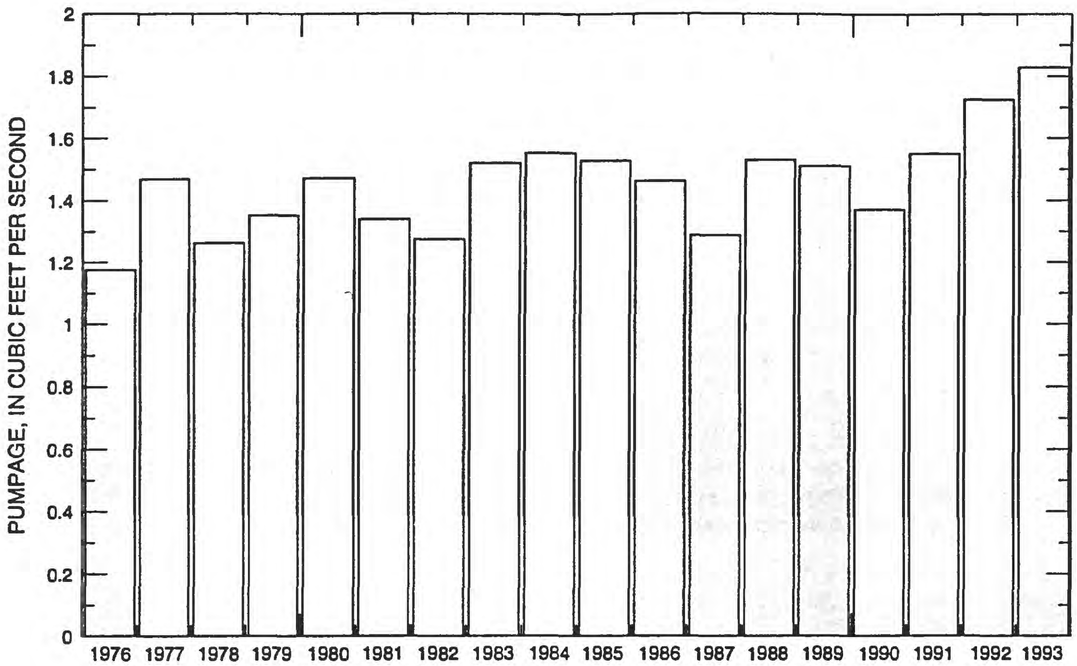


Figure 10. Average pumpage from the Devils Lake well field for 1976-93. (Data from Christopher Bader, North Dakota State Water Commission, oral commun., 1994.)

well field averaged 1,087 acre-feet per year (acre-ft/yr) (1.5 ft^3/s), and withdrawals from the irrigation wells averaged 932 acre-ft/yr (1.29 ft^3/s). Discharge from the irrigation wells was at a maximum of 2.31 ft^3/s in 1992, nearly twice the average rate for 1982-93 (Christopher Bader, North Dakota State Water Commission, written commun., 1994), and at a minimum of 0.31 ft^3/s in 1993. Additional production wells have been installed within one-half mile of the Devils Lake well field by the Spirit Lake Sioux Nation. These wells eventually are intended to withdraw water at a rate of 0.77 ft^3/s (Oliver Gourd, Spirit Lake Sioux Nation, oral commun., 1995). Other than from municipal and irrigation wells, withdrawals from the Warwick aquifer, such as withdrawals from domestic and stock wells, are small.

Storage Changes

The depth to water in two shallow observation wells located outside the Devils Lake well field cone of depression is shown in figure 11, and the depth to water in two observation wells located in the Devils Lake well field cone of depression is shown in figure 12. Water-level fluctuations represent storage changes in the Warwick aquifer. Well 151-063-29AAC2 is equipped with a continuous recorder and provides a long-term record of water levels in the Devils Lake well field since 1951. Water levels in the well field generally declined throughout the period of record, and a correlation generally exists between changes in pumpage from the well field and changes in water levels in the well field. However, this correlation does not exist from about 1988 to 1992. From 1988 through 1990, pumpage from the well field remained relatively stable, but water levels in the well field declined because of less-than-normal precipitation (fig. 13). In 1991, pumpage was normal and precipitation was greater than normal, but the water levels continued to decline. In the fall of 1993, the water levels rose because of wet climatic conditions and decreased pumpage. The water levels in well 151-063-29AAC2 indicate that net storage change from 1979 to 1985 was minor although storage changed from year to year. Because water levels are measured at discrete intervals throughout the year and precipitation and pumpage are reported as total annual values,

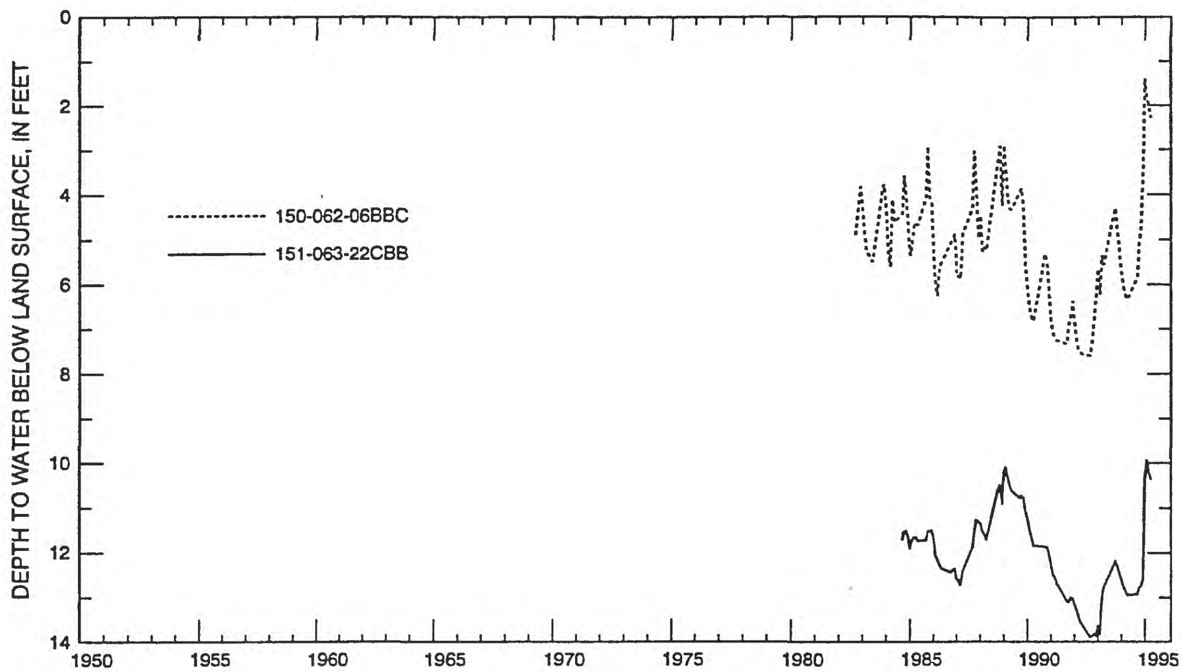


Figure 11. Depth to water in shallow observation wells outside the Devils Lake well field, 1983-95.

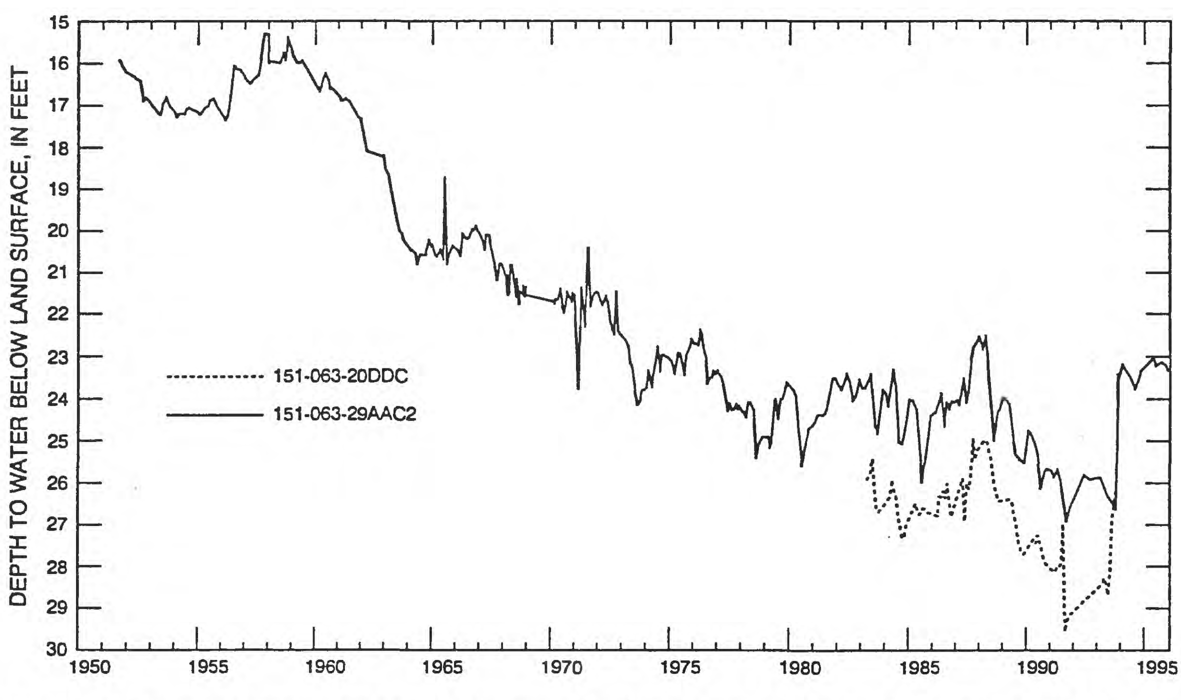


Figure 12. Depth to water in observation wells in the Devils Lake well field cone of depression, 1951-95.

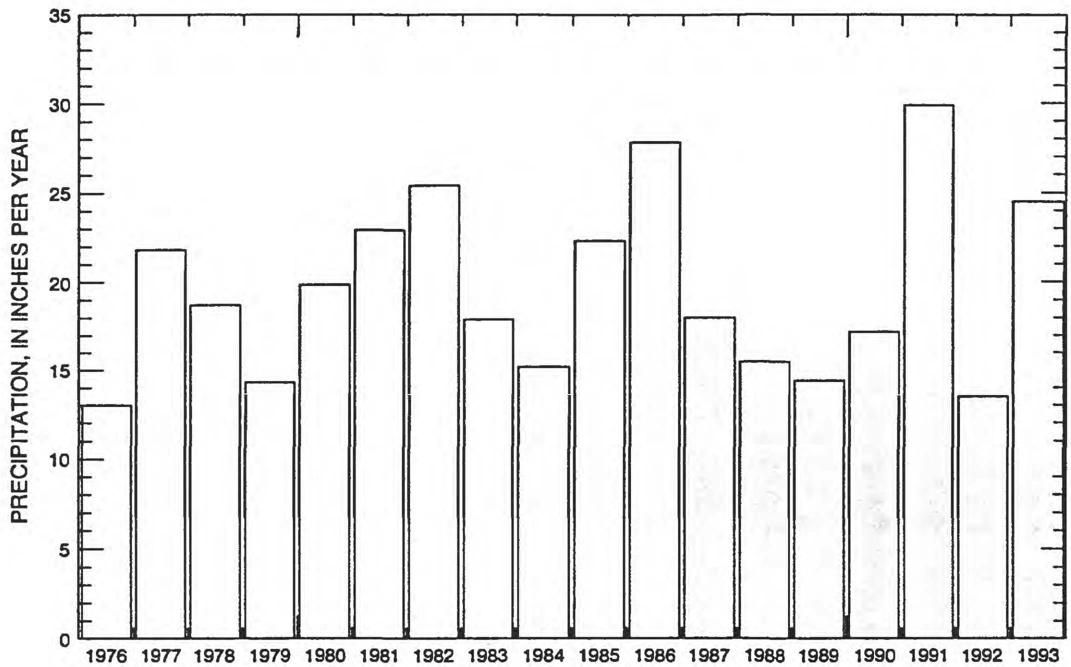


Figure 13. Annual precipitation at the city of Devils Lake for 1976-93. (From U.S. Department of Commerce, National Oceanic and Atmospheric Administration, Environmental Data Service, 1977-94.)

figures 10, 12, and 13 may not always show clear relation (for example, the decline in water levels through 1991). Also, precipitation is measured in the city of Devils Lake, which is about 16 mi away from the Devils Lake well field, and water levels can vary within a month by as much as 3 ft in the well field and by more than 2 ft outside the well field. Thus, the precise causative relation may be difficult to determine without continuous records of water levels, pumpage, and precipitation. Outside of the well field, water levels (and, thus, storage) generally fluctuate (fig. 11) in response to variations in precipitation (fig. 13).

DESIGN OF DIGITAL GROUND-WATER FLOW MODEL

Description of Digital Model

The U.S. Geological Survey finite-difference, three-dimensional, ground-water flow model MODFLOW (McDonald and Harbaugh, 1988) was used to develop and calibrate a ground-water flow model of the Warwick aquifer for steady-state and transient conditions. The calibrated model was used to simulate ground-water flow in the aquifer and to evaluate the range of plausible values for hydrologic characteristics. MODFLOW was used to solve finite-difference, ground-water flow equation approximations for spatial distributions of hydraulic head over time with certain simplifying assumptions.

A map of the Warwick aquifer was overlain by a rectangular grid that discretized the aquifer into cells. Spatial and vertical variations in hydrologic characteristics and in the aquifer framework can be represented in the model, and characteristics that vary continuously in space can be represented by discrete values in model cells. Model cells extend vertically into the aquifer and divide the aquifer into discrete volumes of aquifer material that are assumed to have uniform hydrologic characteristics. The Warwick aquifer was divided vertically into two layers. The method by which the Warwick aquifer was discretized into horizontal and vertical model cells is shown in figure 14.

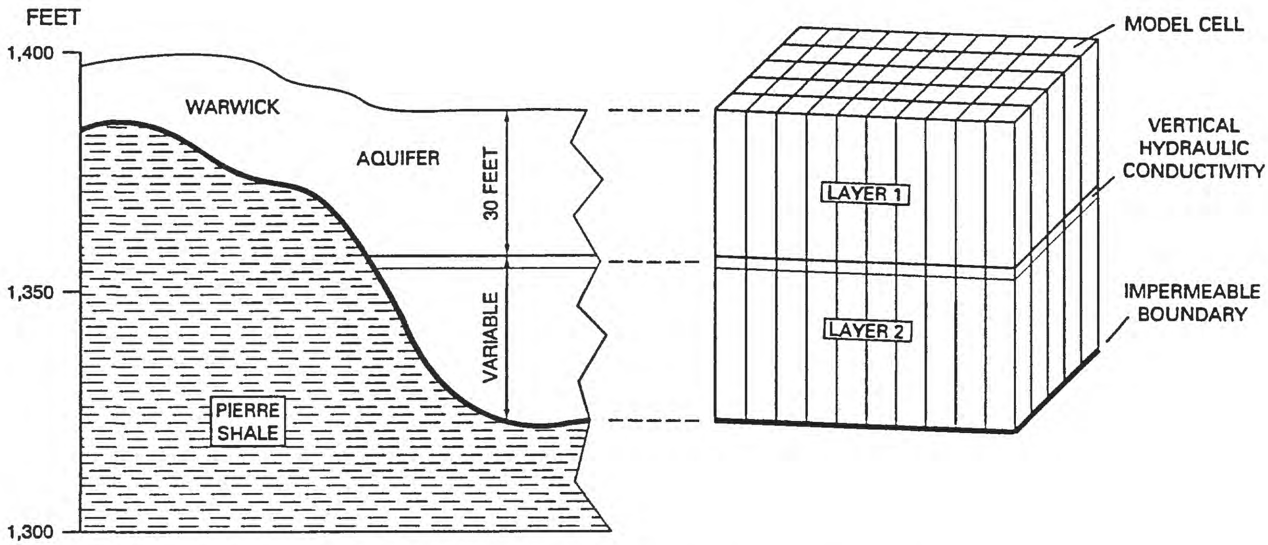


Figure 14. Discretization of the Warwick aquifer into horizontal and vertical model cells.

Simplifying Assumptions

The use of a digital model required that simplifying assumptions be made for the complex and interrelated hydrologic features and processes occurring in the Warwick aquifer. The use of such simplifying assumptions limits the correlation between simulated values and actual conditions. The simplifying assumptions are as follow:

1. The aquifer is represented by two layers. The first layer represents the upper 30 ft of the aquifer and, where the aquifer is thicker than 30 ft, the active model cells of the second layer represent the rest of the aquifer. All flow within the layers is horizontal and all flow between the layers is vertical.

2. The first layer is unconfined, and the second layer is confined.

3. All lateral boundaries are impermeable except in the area of Warwick Springs and smaller springs along the eastern and southern boundaries of the aquifer.

4. The only source of recharge to the aquifer is precipitation. Irrigation return, which possibly contributes a small part of the recharge, was considered negligible.

5. Discharge from the aquifer occurs as evapotranspiration, pumpage from the Devils Lake well field, pumpage from irrigation wells, discharge from Warwick Springs, and discharge from smaller springs. Discharge to lakes and ponds was treated as evapotranspiration from the water table at zero depth to water.

6. Discharge from Warwick Springs is represented by an arc of 14 model cells located about 1,000 ft northeast of the actual location of Warwick Springs at 150-063-15CCA.

7. Discharge from the smaller springs is represented by 255 model cells located along the southern and eastern boundaries of the aquifer.

8. The shale bedrock or till underlying the aquifer is the impermeable lower boundary of the aquifer.

9. During the early 1990s, the aquifer was in a steady-state condition in which inflow to the aquifer balanced outflow and no net change occurred in storage in the aquifer.

Approach Used in Simulations

A total of three simulations were conducted--one of steady-state conditions and two of transient conditions. Steady-state conditions exist when inflow to an aquifer equals outflow from the aquifer. During steady-state conditions, the amount of water stored in the aquifer remains the same. During transient conditions, the amount of water stored in the aquifer may change. A transient simulation typically may be conducted for an aquifer that is not at equilibrium with withdrawals from wells or changes in recharge.

General steady-state conditions that probably existed in the Warwick aquifer in 1992 were assumed for the first steady-state simulation. The hydrograph of well 151-063-29AAC2 in the Devils Lake well field (fig. 12) shows that, from 1979 to 1985, the net change in storage was minor although storage changed from year to year. By 1992, ground-water levels in the aquifer probably had recovered from the effects of increased pumpage and drought conditions. Water-level altitudes for October 1992 were chosen as representative initial conditions for the steady-state simulation. Thus, the steady-state simulation was conducted using 1992 pumpage rates and October 1992 water levels. The steady-state model was calibrated so that simulated water-level altitudes more closely resembled the October 1992 water-level altitudes and the corresponding ground-water flows. Model calibration is the process by which model parameters are adjusted so that simulated water-level altitudes and ground-water flows produced by the model closely resemble derived water-level altitudes and ground-water flows. Selected hydrologic characteristics of the aquifer are varied within reasonable limits during the process. During calibration of the steady-state model, hydraulic-conductivity values for layer 2 and recharge values for layer 1 were varied within reasonable limits as guided by values published in other reports.

To estimate potential effects of future pumpage, two transient simulations were conducted using small and large storage values and doubled pumpage rates. To ensure steady-state initial conditions for these two transient simulations, the simulated water-level altitudes obtained from the steady-state simulation conducted using 1992 pumpage rates were used as the initial altitudes.

Model Specifications

Model Grid and Layers

Digital simulation of ground-water flow in the Warwick aquifer required that aquifer properties be assigned to each model cell. The Warwick aquifer was divided horizontally into model cells by a rectangular grid. For each model cell, the hydrologic characteristics of the aquifer were interpolated so as to provide a single value for each cell. A grid of 83 by 109 cells, each measuring 656 ft (200 meters) per side, was overlain on the Warwick aquifer. The orientation of the rectangular grid was approximately east to west. The horizontal arrangement of the grid of model cells is shown in figure 15.

The Warwick aquifer also was divided vertically into two layers on the basis of availability of data from Bureau of Reclamation studies. Layer 1 consists of that part of the aquifer that is less than or equal to

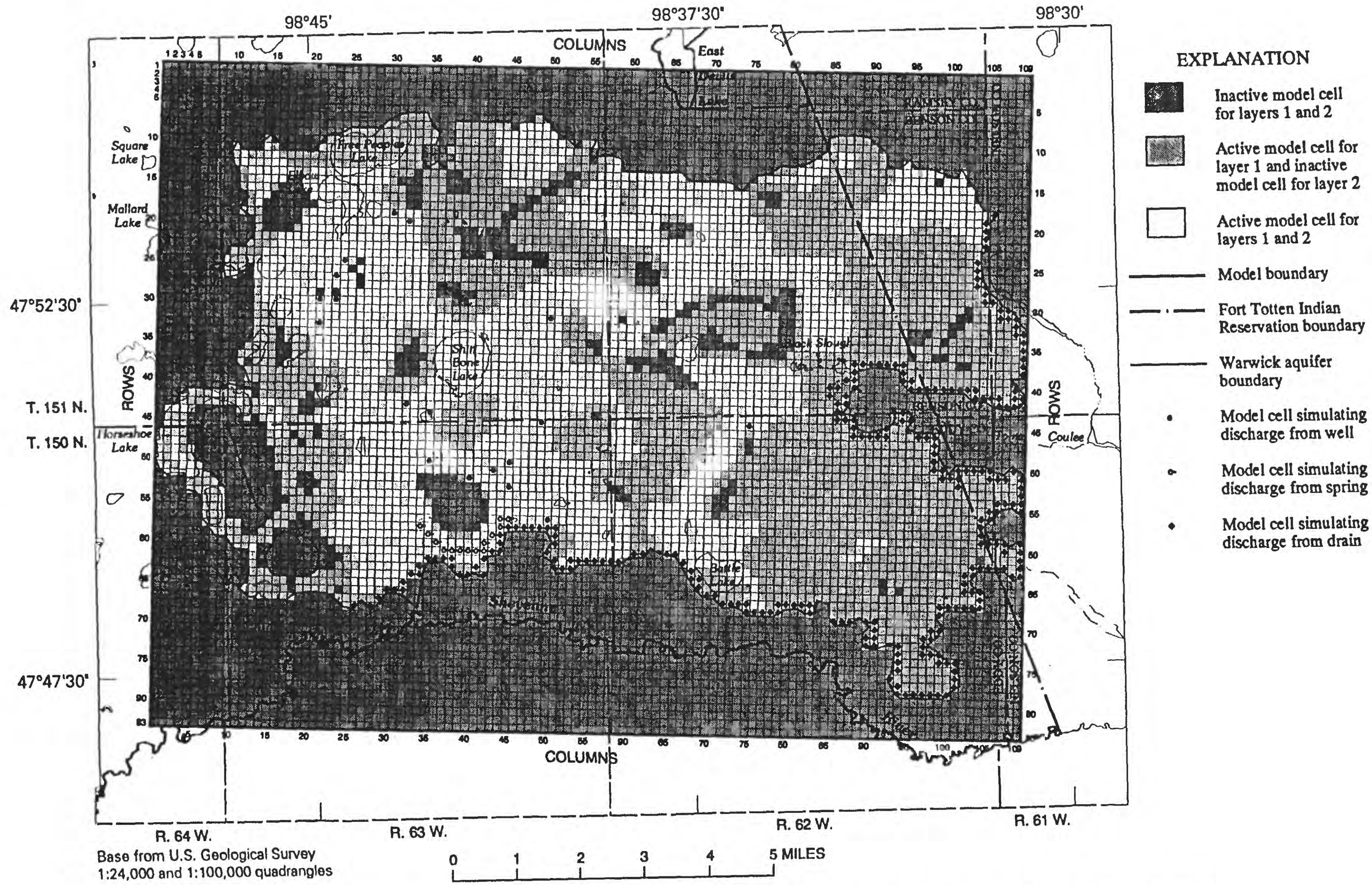


Figure 15. Finite-difference grid of model cells and model boundaries used to define the digital ground-water flow model of the Warwick aquifer.

30 ft thick, and layer 2 consists of that part of the aquifer that is greater than 30 ft thick. Model cells for layer 2 are active only where the aquifer is greater than 30 ft thick. Where the aquifer is absent in the rectangular grid, model cells are inactive and no flow is simulated. The altitude of the top of layer 1 is the land-surface altitude. The land-surface altitude for each model cell was interpolated from 1-ft land-surface contours developed for Bureau of Reclamation studies, well measuring-point altitudes, and land-surface contours shown on U.S. Geological Survey topographic maps. Well measuring points were resurveyed to an accuracy within 1 ft or less using global positioning system techniques. Where necessary, 1:24,000-scale topographic maps were hand digitized to provide additional land-surface altitude data. These data then were integrated to the model grid using geographic information system (ARC/INFO TIN and GRID) functions. The ranges of land-surface altitudes calculated by this process are shown in figure 16.

The aquifer thickness map shown in figure 5 was discretized and used to assign a thickness to each model cell in the grid. The altitude of the bottom of the Warwick aquifer was determined by subtracting the aquifer thickness from the land-surface altitude of each model cell. Where the aquifer is less than or equal to 30 ft thick, the altitude of the bottom of layer 1 is the altitude of the bottom of the aquifer, and layer 2 cells are inactive. Where the aquifer is greater than 30 ft thick, the altitude of the bottom of layer 1 is the land-surface altitude minus 30 ft, and the altitude of the bottom of layer 2 is the altitude of the bottom of the aquifer. Where active model cells are present in layer 2, the altitude of the bottom of layer 1 is the altitude of the top of layer 2. The thicknesses used in the model are shown in figure 17.

Initial Water-Level Altitudes

Initial water-level altitudes were required for each active model cell. The derived water-level altitudes for October 1992 (fig. 8) were used as initial altitudes for the steady-state simulation and were compared to simulated water-level altitudes. The mean depth to water for all observation wells measured in October 1992 was similar to (1.15 ft greater than) the mean depth to water measured over the period of record. Therefore, the October 1992 water-level altitudes probably represent reasonable values for a steady-state simulation. The depth to water in wells 151-063-20DDC and 151-063-29AAC2, which are located in the Devils Lake well field, increased from 1989 to 1992 but stabilized or began to decrease from 1992 to 1994, further justifying the use of October 1992 water levels. The October 1992 water-level altitudes used in the model are shown in figure 18. The initial water-level altitudes were derived from October 1992 water-level altitudes, land-surface altitudes, and lake-stage altitudes from 1:24,000-scale maps.

Boundaries

The upper boundary of the model represents the water table of the Warwick aquifer, and the no-flow lower boundary represents either the glacial till or the Pierre Shale underlying the aquifer. Lateral boundaries of the model generally represent physical limits of the aquifer or ground-water divides in the aquifer. The western no-flow boundary represents the western edge of the surficial expression of the aquifer because, although the aquifer may extend west in the subsurface, data indicate that the saturated thickness becomes very constricted. Model cells on the southern and eastern lateral boundaries were designated as "drains" to enable the simulation of seepage from springs into topographically lower areas to the south and east of the aquifer. The hydraulic conductances and altitudes of the drains were adjusted during model calibration to produce simulated ground-water level altitudes that closely resembled actual conditions.

Horizontal Hydraulic Conductivity

Hydraulic-conductivity values were assigned to each active model cell. Horizontal hydraulic conductivities for the upper 30 ft of the Warwick aquifer (layer 1) were obtained from the Bureau of Reclamation.

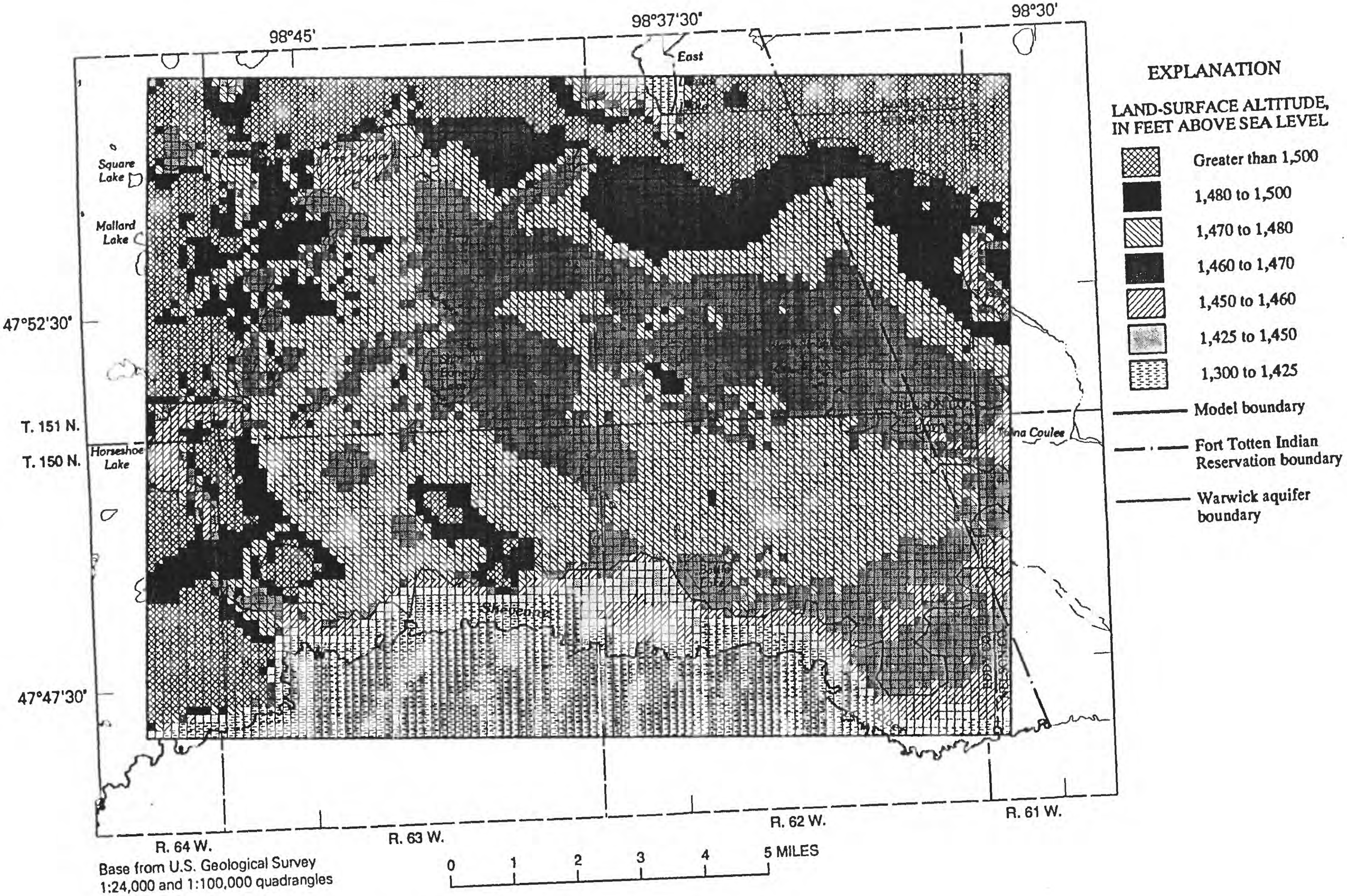


Figure 16. Land-surface altitudes assigned to model cells.

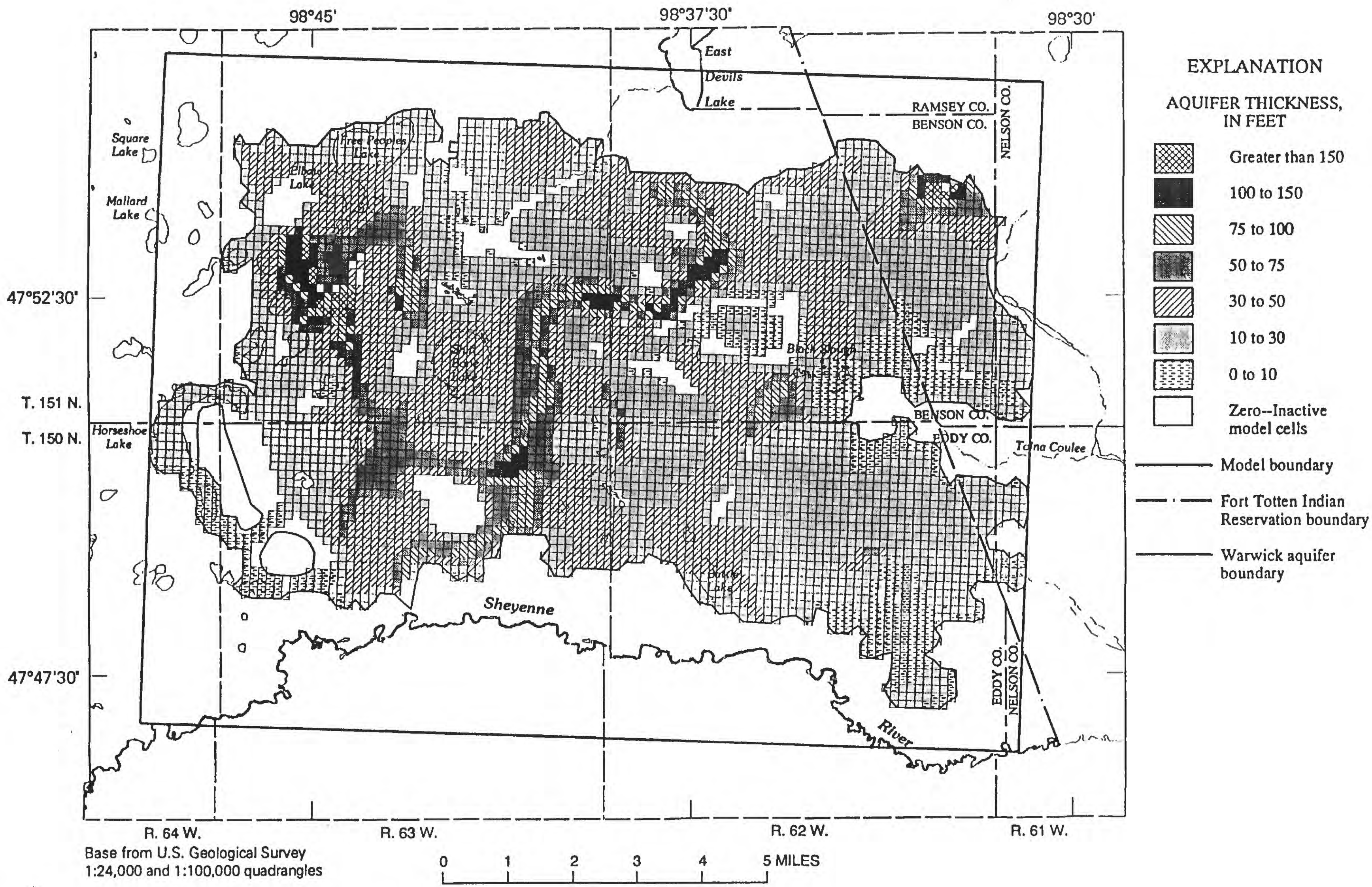


Figure 17. Thickness of the Warwick aquifer assigned to model cells.

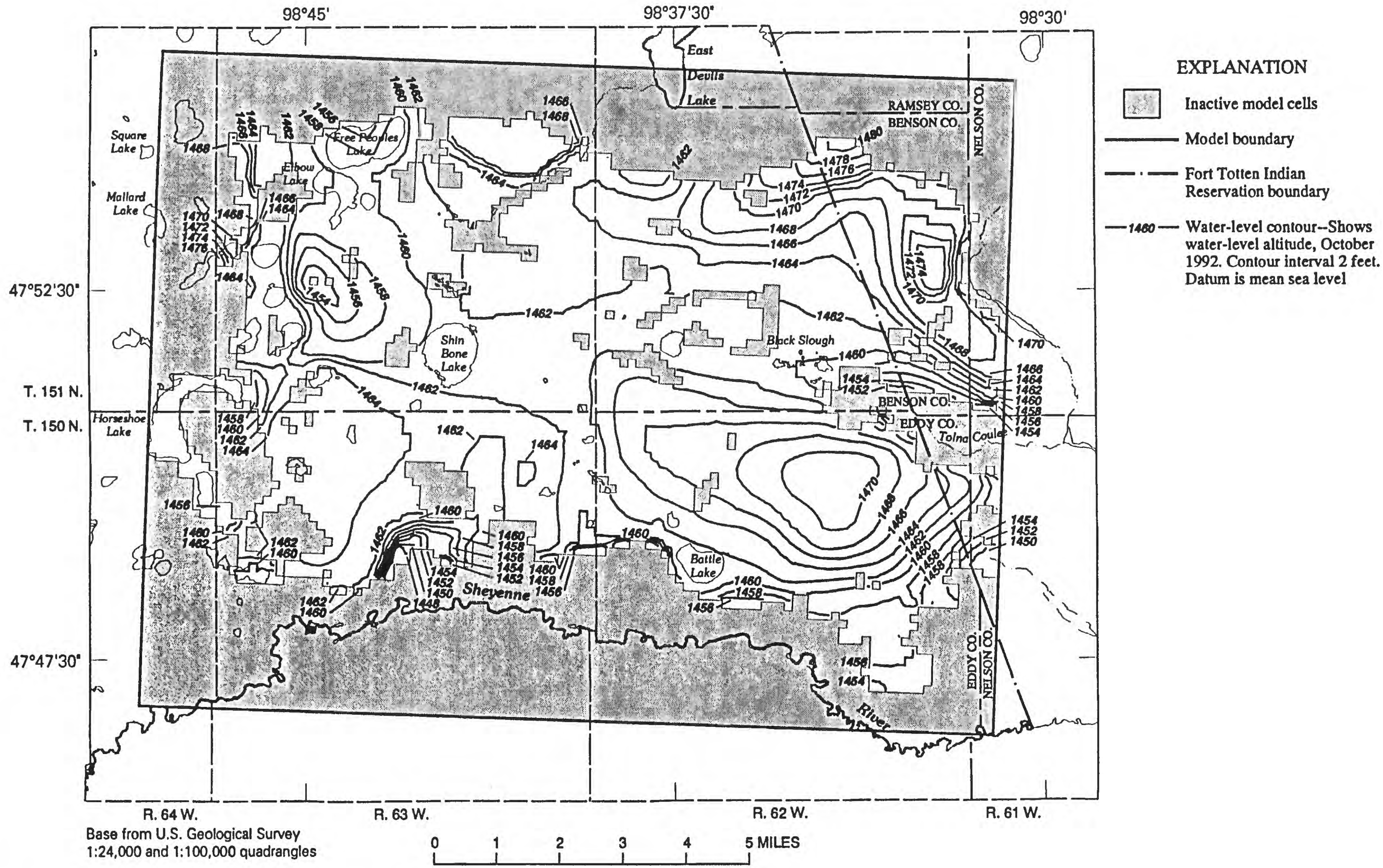


Figure 18. Initial water-level altitudes assigned to model cells.

The values used for model cells in layer 1 are shown in figure 19. Simulated hydraulic-conductivity values assigned to model cells in layer 2 are shown in figure 20. These values were estimated during the calibration process because hydraulic-conductivity data for layer 2 were limited. Thick deposits of coarse gravels may be common at deeper levels of the aquifer in layer 2 and may account for the areas where hydraulic conductivity in layer 2 is larger than in layer 1. Discontinuous clay layers in layer 2 also affect hydraulic conductivities in these areas and account for the smaller hydraulic-conductivity values. Because of the discontinuity of these clay layers, no attempt was made to further discretize the model vertically.

To determine the simulated values of horizontal hydraulic conductivity (fig. 20), the areas where layer 2 was present were divided into three zones on the basis of thickness. In addition to the three zones determined on the basis of thickness, two zones were selected near the Devils Lake well field because of interest in the well field and because of large pumpage stresses placed on this area of the Warwick aquifer. During the calibration process, the hydraulic-conductivity values in all five zones were adjusted separately by multipliers in an attempt to simulate water-level altitudes that most resembled altitudes derived from October 1992 data. Where layer 2 was less than or equal to 15 ft thick, the initial hydraulic-conductivity value assigned was the hydraulic-conductivity value of the overlying model cell in layer 1. Where layer 2 was greater than 15 ft thick, the initial value assigned was 100 ft/d, the maximum hydraulic-conductivity value in layer 1. This value was chosen because sand and gravel deposits often are present in the deep channel deposits in layer 2. Model cells in layer 2 in the two zones located near the Devils Lake well field were assigned an initial hydraulic-conductivity value of 100 ft/d.

Vertical Hydraulic Conductivity

To simulate ground-water flow between layers, MODFLOW requires that a vertical conductance value be applied to all applicable model cells. The value was calculated by the use of equation 51 in McDonald and Harbaugh (1988, p. 5-13), with the relevant horizontal hydraulic conductivities. Vertical hydraulic conductivity is assumed to be 10 percent of horizontal hydraulic conductivity. The model does not simulate the vertical flow of ground water other than between layers 1 and 2.

Recharge

The initial recharge value for each model cell was 4.5 in/yr. During the calibration process, an effort was made to relate areas that had large residuals between derived and simulated water levels to plausible variations in recharge. Active model cells were grouped into zones for this effort. Topographically low areas, especially those where wetlands and water bodies exist, generally have greater recharge rates than topographically high areas because of runoff during precipitation. During the spring when evapotranspiration is low, snowmelt and runoff are major sources of recharge in North Dakota, and wetlands and small bodies of water where depth to the water table is small are important recharge areas. Therefore, topography and wetlands provide a basis for spatial variations in recharge. Recharge to the selected zones of active model cells was adjusted by multipliers during the calibration process to simulate water-level altitudes that most resembled derived water-level altitudes. The altitudes shown in figure 18 indicate that a large ground-water mound exists beneath the sand dune deposits in the southeastern part of the Warwick aquifer. The gently undulating sand dunes have highly permeable soils, classified as the Claire-Lohnes-Hamar soil association, and an estimated permeability of 6 to 10 inches per hour (in/hr) (Wright, 1977, p. 146). This indicates that the sand dune deposits have a relatively large recharge capacity. Therefore, during the calibration process, model cells simulating this area were assigned recharge values that were greater than the initial values.

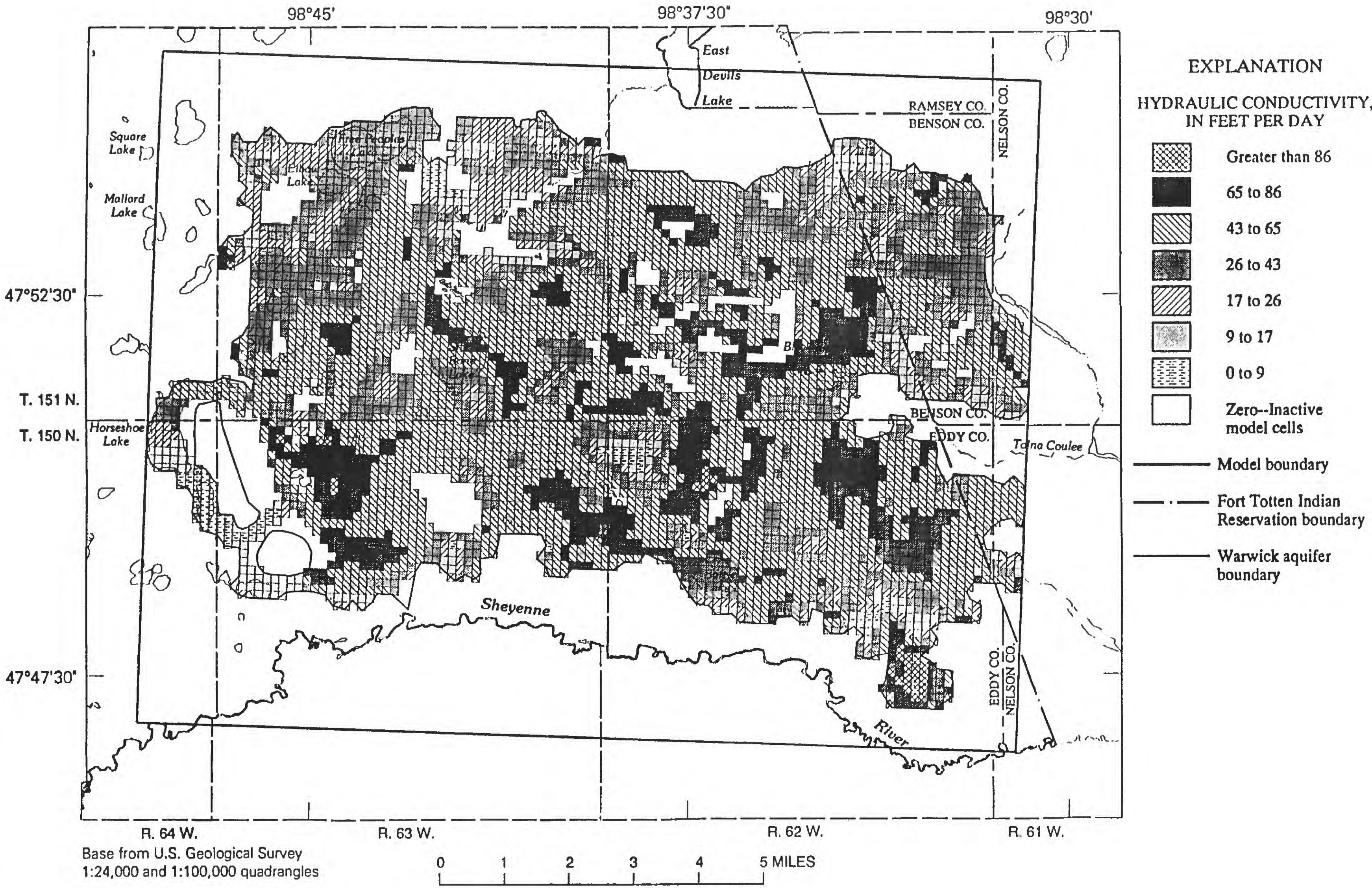
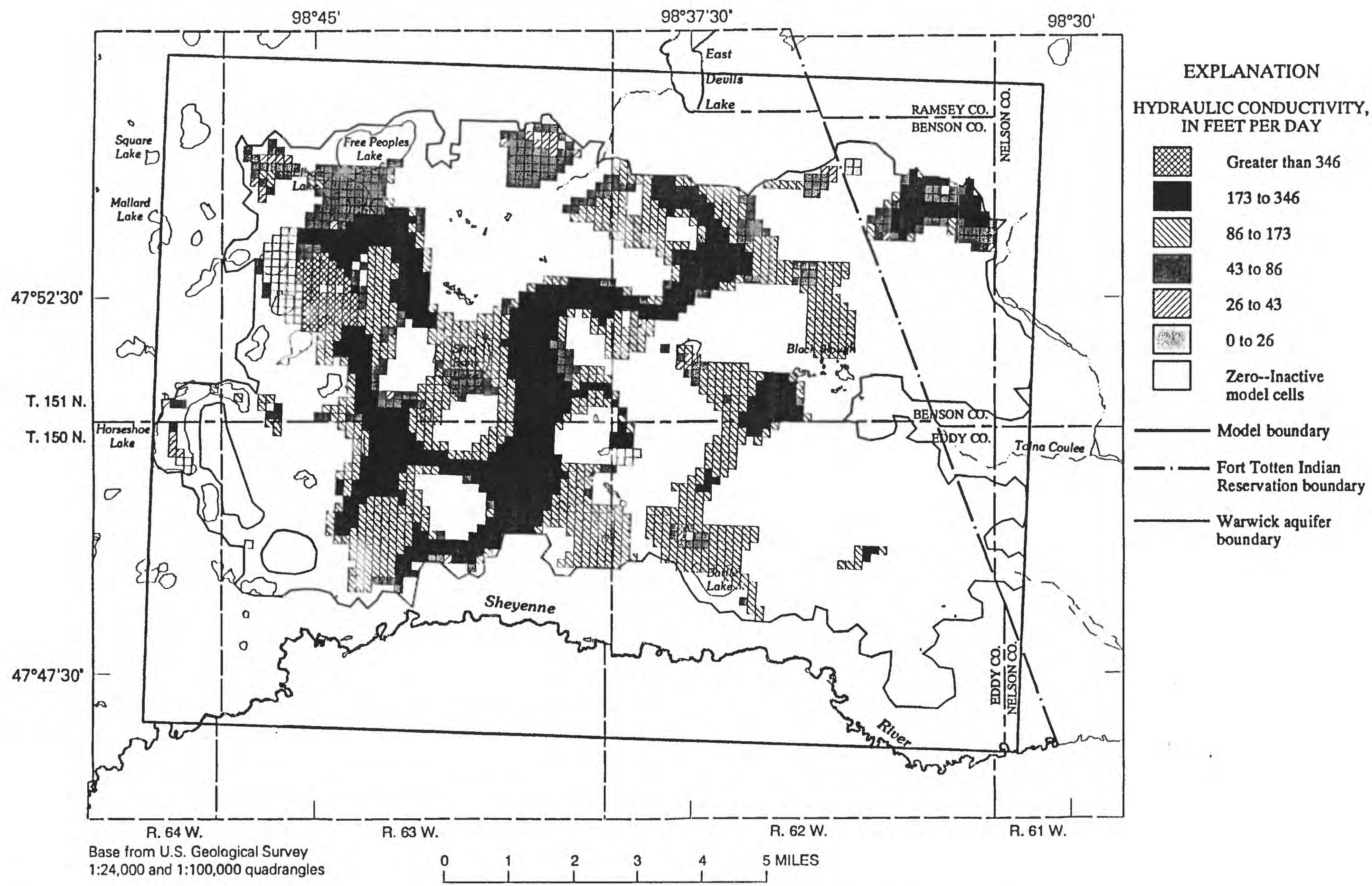


Figure 19. Hydraulic-conductivity values assigned to model cells in layer 1.



Base from U.S. Geological Survey
1:24,000 and 1:100,000 quadrangles

Figure 20. Simulated hydraulic-conductivity values estimated during the calibration process for model cells in layer 2.

Evapotranspiration

The areal distribution of evapotranspiration from an aquifer is a function of depth to the water table, potential evapotranspiration, and maximum depth at which evapotranspiration occurs. For each model cell, the potential evapotranspiration rate assigned was the evaporation from a small pond minus average precipitation. Using data from Farnsworth and others (1982, map 3) and the U.S. Department of Commerce, National Oceanic and Atmospheric Administration (1992), a value of 16.29 in/yr was calculated for net evapotranspiration loss. The maximum depth at which evapotranspiration occurs was set at 10 ft. The model simulates evapotranspiration as a linear relation where the evapotranspiration rate is maximum at ground surface and decreases linearly to zero at the maximum depth at which evapotranspiration occurs. Evapotranspiration values were not varied during the calibration process.

Spring Discharge and Well Pumpage

Discharge from Warwick Springs was represented by an arc of 14 model cells (table 1) located about 1,000 ft northeast of the actual location of Warwick Springs. The combined discharge from the cells was 1.33 ft^3/s. This approach was used because of numerical difficulties with such a large discharge too close to the model boundary. Discharge from smaller springs along the southern and eastern boundaries of the Warwick aquifer was simulated by 255 drain cells in layer 1. The initial hydraulic conductances for these cells were set arbitrarily at 1 foot squared per second (ft^2/s) and adjusted during calibration to reflect the derived water-level altitudes along the southern and eastern boundaries of the aquifer. The locations of these drain cells and the altitudes and hydraulic conductances determined during the final calibration process are given in table 2.

Records of pumpage from the Devils Lake well field, the largest area of pumpage from the aquifer, before 1976 are not available, but pumpage from the well field for 1976-93 is shown in figure 10. Other than from irrigation wells and the Devils lake well field, withdrawals from the aquifer, such as withdrawals from domestic and stock wells, probably are small. Pumpage data for 1992 were obtained from the North Dakota State Water Commission for the Devils Lake well field and for irrigation wells for which data are available. Pumpage rates for 1992 were used as the basis for all simulations.

All screened intervals in pumped wells in the Warwick aquifer are deeper than 30 ft. Therefore, pumpage from the wells was simulated by model cells in layer 2. The locations of model cells simulating discharge from the Devils Lake well field, irrigation wells, and Warwick Springs are given in table 1.

Storage

Specific-yield values were not reported for aquifer tests conducted by Randich (1977, p. 47). Lohman (1972, p. 8) reported values that ranged from about 0.1 to 0.3 for most unconfined aquifers. Williamson and others (1989, p. D-29) reported elastic specific-storage values that ranged from $7x10^{-7}$ to $1x10^{-6}$ per ft for coarse-grained sediments in confined aquifers in the Central Valley of California. They also reported elastic specific-storage values of $3x10^{-6}$ per ft for deposits that are half coarse grained and half fine grained. These values represent a probable range for specific storage for the Warwick aquifer. Specific-yield values (applicable to unconfined conditions) were entered for layer 1, and storage coefficients (applicable to confined conditions) were determined for layer 2 by multiplying specific-storage values by the thickness of layer 2. Given that particle-size distributions of the aquifer material indicate a mean porosity of 42 percent (Randich, 1977, p. 45), the actual specific-yield and specific-storage values probably are at the higher end of the reported ranges. Rather than conducting a single, calibrated, transient simulation, two separate transient simulations were conducted using small and large storage estimates. The large

Table 1. Model cells in layer 2 that represent discharge from the Devils Lake well field, irrigation wells, and Warwick Springs

Model cell[1] Row	Column	Location	Discharge (cubic feet per second)	Type of discharge
19	30	151-063-16D	0.063495	Irrigation well
20	32	151-063-16D	0	Irrigation well
25	24	151-063-20ADD	.137210	Irrigation well
27	23	151-063-20DA	.086271	Irrigation well
30	21	151-063-29AB1	.517210	Devils Lake well field
30	23	151-063-29AB2	.517070	Devils Lake well field
30	23	151-063-29AB3	.344810	Devils Lake well field
32	50	151-063-25BC	.167710	Irrigation well
33	21	151-063-29AC	.344810	Devils Lake well field
43	32	151-063-33D	.220850	Irrigation well
44	35	151-063-34C	.004141	Irrigation well
45	49	150-063-02A	.217130	Irrigation well
45	75	150-062-04B	.105870	Irrigation well
50	35	150-063-03C	.063495	Irrigation well
50	45	150-063-02CA	0	Irrigation well
51	43	150-063-02C	.538330	Irrigation well
52	40	150-063-03D	0	Irrigation well
53	45	150-063-11B	.706040	Irrigation well
57	44	150-063-15CCA[2]	.095000	Warwick Springs
57	45	150-063-15CCA[2]	.095000	Warwick Springs
57	50	150-063-11D[2]	0	Irrigation well
58	34	150-063-15CCA[2]	.095000	Warwick Springs
58	44	150-063-15CCA[2]	.095000	Warwick Springs
59	35	150-063-15CCA[2]	.095000	Warwick Springs
59	43	150-063-15CCA[2]	.095000	Warwick Springs
60	36	150-063-15CCA[2]	.095000	Warwick Springs
60	42	150-063-15CCA[2]	.095000	Warwick Springs
61	37	150-063-15CCA[2]	.095000	Warwick Springs
61	38	150-063-15CCA[2]	.095000	Warwick Springs
61	39	150-063-15CCA[2]	.095000	Warwick Springs
61	40	150-063-15CCA[2]	.095000	Warwick Springs
61	41	150-063-15CCA[2]	.095000	Warwick Springs
61	42	150-063-15CCA[2]	.095000	Warwick Springs

[1]Two wells can be in the same model cell.

[2]Actual location of Warwick Springs; differs from location of cells used in model simulation.

specific-yield value was 0.30, and the large specific-storage value was 3×10^{-6} per ft. The small specific-yield value was 0.10, and the small specific-storage value was 9×10^{-7} per ft.

DIGITAL SIMULATION OF GROUND-WATER FLOW IN THE WARWICK AQUIFER

Steady-State Simulation

During steady-state conditions, the amount of water stored in an aquifer remains the same. All ground-water flow is assumed to represent recharge or discharge, and none represents changes in ground-

Table 2. Model cells in layer 1 that represent discharge from small springs along the southern and eastern boundaries of the Warwick aquifer and altitudes and hydraulic conductances determined during final calibration of the steady-state model

Row	Column	Altitude (feet)	Hydraulic conductance (feet squared per second)
18	105	1,441.844	0.00232511
19	104	1,460.442	.00180317
20	104	1,459.910	.00152288
21	104	1,457.319	.00142010
22	104	1,447.008	.00184965
24	103	1,458.924	.00267030
25	103	1,459.067	.00266388
26	103	1,461.247	.00200257
27	103	1,462.285	.00176497
27	104	1,450.222	.00274983
28	104	1,462.913	.00145184
29	104	1,464.330	.00141768
29	105	1,460.520	.00225504
30	106	1,469.173	1.00000000
30	107	1,449.737	.00363702
31	106	1,479.818	1.00000000
31	107	1,464.135	.00128961
31	108	1,463.054	.00133719
32	108	1,463.491	.00186301
33	108	1,465.061	.00142770
33	109	1,463.220	.00206033
34	109	1,463.497	.00203789
35	109	1,477.936	1.00000000
36	109	1,477.137	1.00000000
37	88	1,461.000	1.00000000
37	89	1,463.602	1.00000000
37	90	1,465.557	1.00000000
37	91	1,465.725	1.00000000
37	92	1,465.730	1.00000000
37	93	1,465.000	1.00000000
37	94	1,464.276	1.00000000
37	109	1,476.445	1.00000000
38	88	1,461.000	1.00000000
38	94	1,465.732	1.00000000
38	109	1,475.753	1.00000000
39	87	1,461.000	1.00000000
39	88	1,461.000	1.00000000
39	94	1,449.139	.00093902
39	95	1,452.245	.00020832
40	85	1,461.816	1.00000000
40	86	1,463.000	1.00000000
40	88	1,461.880	1.00000000
40	95	1,442.067	.00063369
40	96	1,447.399	.00052830
40	97	1,446.834	.00176630
40	98	1,448.994	.00166965
40	99	1,449.803	.00136092
40	100	1,449.667	.00140285
40	101	1,450.415	.00099114
40	102	1,453.317	.00035616

Table 2. Model cells in layer 1 that represent discharge from small springs along the southern and eastern boundaries of the Warwick aquifer and altitudes and hydraulic conductances determined during final calibration of the steady-state model—Continued

Row	Column	Altitude (feet)	Hydraulic conductance (feet squared per second)
40	103	1,454.214	0.00023711
40	108	1,461.658	1.00000000
41	85	1,463.112	1.00000000
41	87	1,463.158	1.00000000
41	103	1,447.767	.00100841
41	104	1,448.729	.00090716
41	105	1,463.820	1.00000000
41	106	1,462.934	1.00000000
41	108	1,460.590	1.00000000
41	109	1,459.454	1.00000000
42	86	1,466.000	1.00000000
42	87	1,464.752	1.00000000
42	106	1,462.028	1.00000000
42	107	1,461.191	1.00000000
42	108	1,460.274	1.00000000
43	86	1,469.393	1.00000000
43	93	1,460.558	1.00000000
43	94	1,460.000	1.00000000
44	86	1,472.896	1.00000000
44	87	1,470.000	1.00000000
44	88	1,470.000	1.00000000
44	94	1,461.848	1.00000000
44	95	1,460.000	1.00000000
45	88	1,472.147	1.00000000
45	92	1,468.892	1.00000000
46	88	1,471.232	1.00000000
46	89	1,471.000	1.00000000
46	90	1,469.000	1.00000000
46	91	1,470.000	1.00000000
46	92	1,470.278	1.00000000
46	96	1,468.592	1.00000000
46	97	1,466.066	1.00000000
46	98	1,465.053	1.00000000
47	98	1,469.452	1.00000000
48	98	1,471.531	1.00000000
49	98	1,471.271	1.00000000
50	98	1,473.957	1.00000000
50	99	1,472.000	1.00000000
50	101	1,470.000	1.00000000
50	102	1,471.733	1.00000000
50	103	1,458.305	.00127340
50	104	1,457.538	.00137613
50	105	1,457.192	.00146208
50	108	1,462.098	1.00000000
51	99	1,471.884	1.00000000
51	100	1,472.000	1.00000000
51	101	1,470.934	1.00000000
51	105	1,469.000	1.00000000
51	106	1,469.497	1.00000000
51	107	1,465.737	1.00000000

Table 2. Model cells in layer 1 that represent discharge from small springs along the southern and eastern boundaries of the Warwick aquifer and altitudes and hydraulic conductances determined during final calibration of the steady-state model—Continued

Row	Column	Altitude (feet)	Hydraulic conductance (feet squared per second)
51	109	1,463.184	1.00000000
52	101	1,471.709	1.00000000
52	109	1,464.241	1.00000000
53	109	1,464.445	1.00000000
54	106	1,469.650	1.00000000
54	107	1,467.983	1.00000000
54	108	1,466.316	1.00000000
54	109	1,465.286	1.00000000
55	105	1,458.240	.00102782
55	106	1,454.755	.00212524
55	109	1,464.376	1.00000000
56	104	1,457.199	.00151650
56	105	1,454.556	.00196880
57	104	1,454.726	.00164002
58	44	1,480.000	1.00000000
58	45	1,480.000	1.00000000
58	46	1,487.097	1.00000000
58	47	1,489.347	1.00000000
58	48	1,479.781	1.00000000
58	49	1,478.888	1.00000000
58	50	1,477.785	1.00000000
58	51	1,476.645	1.00000000
58	104	1,454.679	.00123760
59	44	1,480.000	1.00000000
59	51	1,478.258	1.00000000
59	104	1,460.000	1.00000000
59	105	1,460.000	1.00000000
59	106	1,460.598	1.00000000
59	107	1,458.274	1.00000000
59	109	1,457.091	1.00000000
60	44	1,481.286	1.00000000
60	51	1,477.660	1.00000000
60	107	1,453.845	1.00000000
60	109	1,451.645	1.00000000
61	42	1,476.960	1.00000000
61	43	1,479.238	1.00000000
61	44	1,485.363	1.00000000
61	50	1,477.912	1.00000000
61	61	1,447.038	.00389820
61	62	1,444.798	.00392763
61	65	1,443.543	.00532802
61	66	1,446.865	.00553459
61	109	1,450.000	1.00000000
62	35	1,441.000	.00443318
62	36	1,443.344	.00342447
62	37	1,449.000	.00161063
62	38	1,446.000	.00140723
62	42	1,477.328	1.00000000
62	51	1,473.601	1.00000000
62	53	1,470.997	1.00000000

Table 2. Model cells in layer 1 that represent discharge from small springs along the southern and eastern boundaries of the Warwick aquifer and altitudes and hydraulic conductances determined during final calibration of the steady-state model—Continued

Row	Column	Altitude (feet)	Hydraulic conductance (feet squared per second)
62	54	1,468.656	1.00000000
62	55	1,453.943	.00244897
62	56	1,444.824	.00544942
62	57	1,443.624	.00534038
62	58	1,443.399	.00481736
62	59	1,441.090	.00471020
62	67	1,444.153	.00575072
62	103	1,460.074	1.00000000
62	104	1,458.141	1.00000000
62	105	1,453.000	1.00000000
62	106	1,450.621	1.00000000
62	107	1,450.000	1.00000000
62	109	1,450.000	1.00000000
63	34	1,443.443	.00221249
63	38	1,451.000	.00058694
63	41	1,468.780	1.00000000
63	51	1,466.470	1.00000000
63	52	1,465.025	1.00000000
63	53	1,463.969	1.00000000
63	68	1,460.473	1.00000000
63	102	1,460.755	1.00000000
63	103	1,461.770	1.00000000
63	108	1,451.025	1.00000000
63	109	1,450.000	1.00000000
64	33	1,432.315	.00264994
64	34	1,427.866	.00196824
64	39	1,443.802	.00096552
64	40	1,459.330	1.00000000
64	42	1,465.325	1.00000000
64	68	1,460.103	1.00000000
64	69	1,460.661	1.00000000
64	102	1,462.809	1.00000000
65	32	1,437.137	.00311301
65	69	1,460.292	1.00000000
65	102	1,460.145	1.00000000
65	103	1,460.000	1.00000000
65	104	1,458.000	1.00000000
66	30	1,470.149	1.00000000
66	31	1,452.111	1.00000000
66	32	1,431.106	.00349544
66	70	1,460.269	1.00000000
66	71	1,460.700	1.00000000
66	104	1,460.000	1.00000000
67	29	1,466.764	1.00000000
67	72	1,460.364	1.00000000
67	73	1,460.374	1.00000000
67	74	1,463.328	1.00000000
67	75	1,471.088	1.00000000
67	79	1,465.905	1.00000000
67	80	1,462.190	1.00000000

Table 2. Model cells in layer 1 that represent discharge from small springs along the southern and eastern boundaries of the Warwick aquifer and altitudes and hydraulic conductances determined during final calibration of the steady-state model—Continued

Row	Column	Altitude (feet)	Hydraulic conductance (feet squared per second)
67	81	1,456.245	1.00000000
67	82	1,460.096	1.00000000
67	83	1,464.510	1.00000000
67	99	1,460.491	1.00000000
67	100	1,461.786	1.00000000
67	101	1,460.751	1.00000000
67	102	1,460.000	1.00000000
67	103	1,461.000	1.00000000
67	104	1,462.000	1.00000000
68	26	1,466.857	1.00000000
68	28	1,464.466	1.00000000
68	75	1,438.646	.00148992
68	76	1,439.556	.00334046
68	77	1,442.027	.00415797
68	78	1,441.376	.00439015
68	79	1,439.713	.00552681
68	83	1,459.817	1.00000000
68	99	1,458.275	1.00000000
69	86	1,459.095	1.00000000
69	87	1,461.377	1.00000000
69	88	1,461.981	1.00000000
69	90	1,462.000	1.00000000
69	91	1,462.411	1.00000000
70	91	1,461.938	1.00000000
70	98	1,461.000	1.00000000
70	99	1,461.952	1.00000000
71	90	1,461.000	1.00000000
71	91	1,460.568	1.00000000
71	97	1,460.224	1.00000000
71	98	1,462.896	1.00000000
72	90	1,459.069	1.00000000
72	91	1,460.000	1.00000000
72	94	1,460.834	1.00000000
72	97	1,461.000	1.00000000
73	94	1,462.010	1.00000000
73	97	1,462.000	1.00000000
73	98	1,462.000	1.00000000
73	99	1,462.000	1.00000000
74	94	1,463.648	1.00000000
74	99	1,462.857	1.00000000
74	100	1,462.051	1.00000000
74	101	1,462.218	1.00000000
75	94	1,463.000	1.00000000
75	101	1,462.639	1.00000000
76	93	1,462.753	1.00000000
76	101	1,462.101	1.00000000
77	94	1,453.097	.00176817
77	101	1,460.407	1.00000000
78	94	1,440.812	.00523129
78	95	1,449.271	.00230312

Table 2. *Model cells in layer 1 that represent discharge from small springs along the southern and eastern boundaries of the Warwick aquifer and altitudes and hydraulic conductances determined during final calibration of the steady-state model*—Continued

Row	Column	Altitude (feet)	Hydraulic conductance (feet squared per second)
78	96	1,461.000	1.00000000
78	97	1,460.922	1.00000000
78	98	1,459.956	1.00000000
78	99	1,459.231	1.00000000
78	100	1,459.895	1.00000000

water storage. In the steady-state simulation, the model was calibrated to simulate October 1992 derived water-level altitudes. During calibration, hydraulic-conductivity values for layer 2 and recharge values for layer 1 were varied within reasonable limits to better simulate the altitudes.

Hydraulic-conductivity values for layer 2 were varied according to zones. Of the five zones, three were determined on the basis of thickness as previously described, and two were selected from near the Devils Lake well field because of interest in the well field and because of the large pumpage stresses in the area. A total of 55 model cells were within the well field and 46 model cells were in an adjacent wetland northwest of the well field. The hydraulic conductivities were varied by applying a multiplier to 100 ft/d, the maximum hydraulic-conductivity value in layer 1. In the final calibration, the multipliers were 2 for thicknesses less than 15 ft and thicknesses from 15 to 100 ft, 0.6 for thicknesses greater than 100 ft, 3.6 for the 55 model cells within the well field, and 0.2 for the 46 model cells in the wetland. Simulated hydraulic-conductivity values obtained during the calibration process are shown in figure 20 for model cells in layer 2.

Recharge to selected zones in layer 1 was varied during the calibration process by multipliers that ranged from zero to 3.5. As previously stated, topographically high areas are likely to have smaller recharge rates than topographically low areas because of runoff during precipitation. Simulated recharge values estimated during the calibration process are shown in figure 21 for model cells in layer 1. The amount of recharge was small (0.05 to 0.20 times the base value) for some model cells in the area of the well field but large (2.5 times the base value) for model cells simulating wetlands located to the west of the well field.

The altitudes and hydraulic conductances for simulated small springs (drain cells) along the southern and eastern boundaries were varied during the calibration process. The final values for these properties are shown in table 2.

Calibration of the model continued until final simulation of steady-state conditions, in which 39 active model cells went dry, closely resembled derived water-level altitudes. Water-level altitudes (hydraulic heads in model layer 1) obtained during the final simulation of steady-state conditions, using a closure criteria of 0.02 ft, are shown in figure 22. Except in the area southeast of the Devils Lake well field, the simulated water-level altitudes are reasonably close to the initial altitudes shown in figure 18. However, altitudes in the area of the well field, especially near Shin Bone Lake, were difficult to match. Derived altitudes typically were a few feet higher than simulated altitudes.

The cone of depression around the Devils Lake well field, as represented by the 1,460-ft water-level altitude contour, does not extend as far spatially in simulated conditions (fig. 22) as in actual conditions (fig. 18) and also is not as deep as in actual conditions. These differences probably can be attributed to the

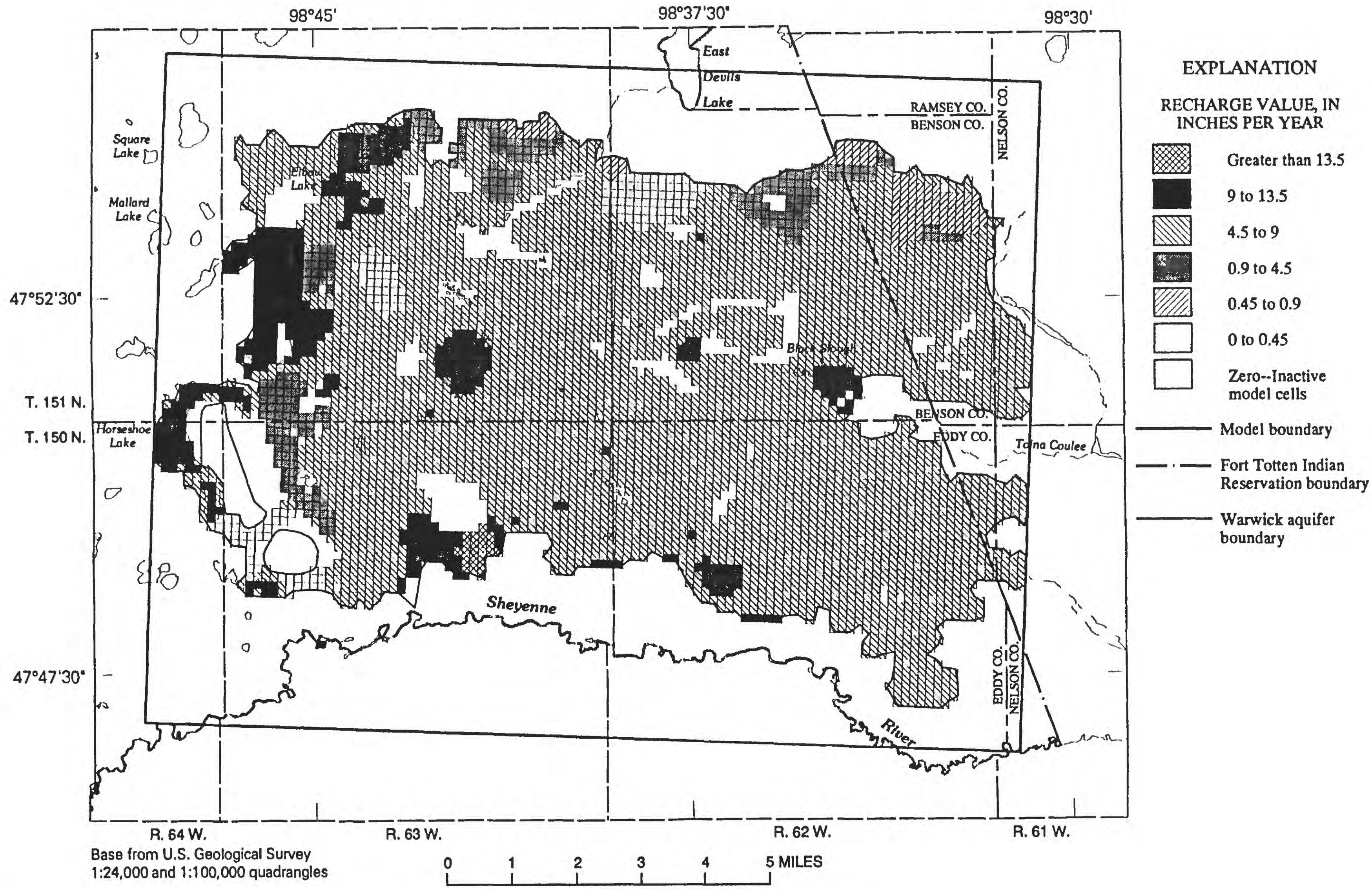


Figure 21. Simulated recharge values estimated during the calibration process for model cells in layer 1.

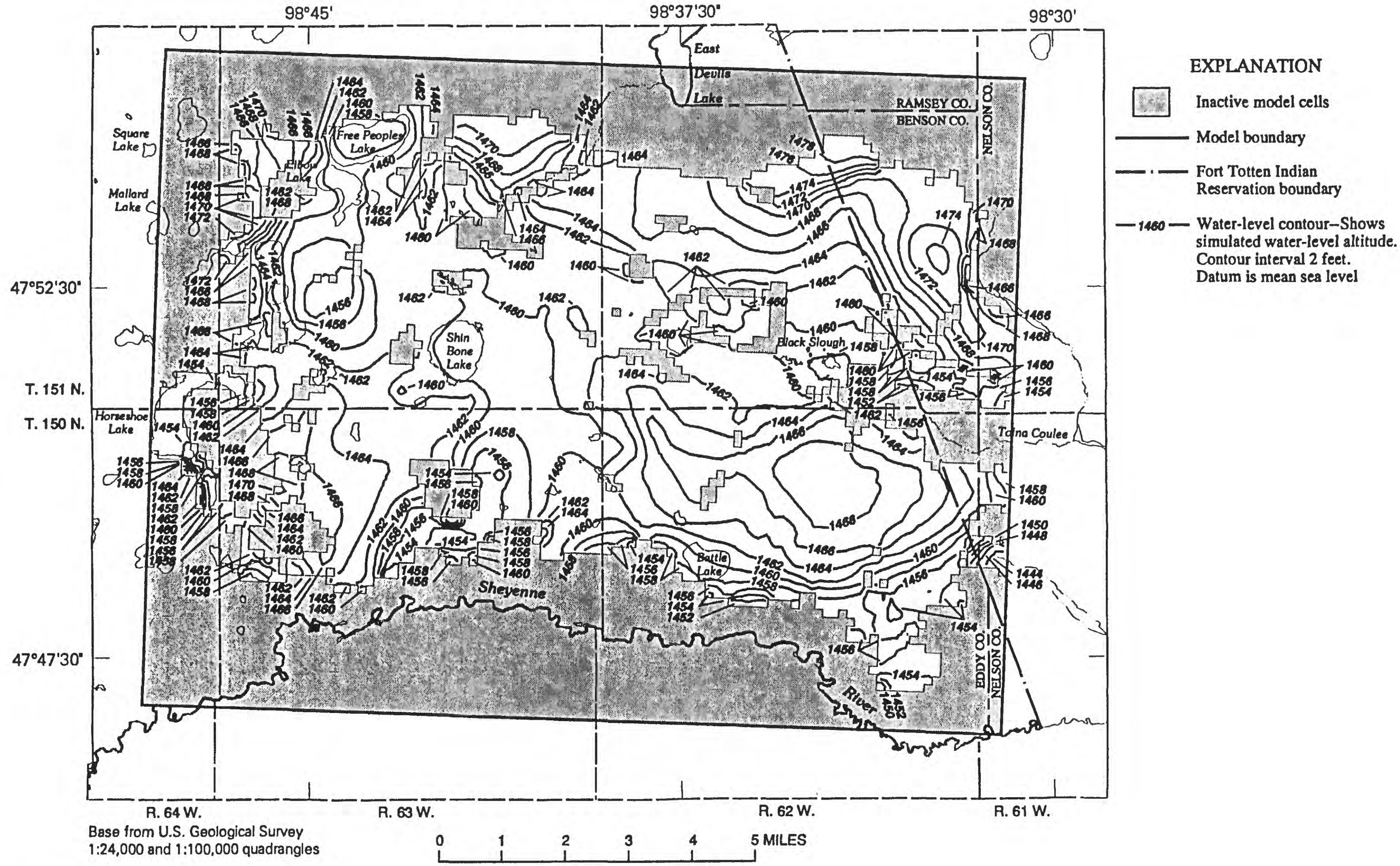


Figure 22. Simulated water-level altitudes obtained from the steady-state model.

use of only two model layers to simulate the thicker deposits in the area of the Devils Lake well field. Because the glacial deposits are deeply channelized in the area of the well field (figs. 7 and 17), the model cannot accurately represent vertical variations in hydraulic conductivity and thickness and, therefore, water-level altitudes in the area of the well field.

The simulated water-level altitudes generally were lower than the derived altitudes in the area southeast of the Devils Lake well field, and small cones of depression are shown around irrigation wells in simulated conditions of the area. Measured water-level data are not dense enough to confirm the presence of these small cones of depression that are indicated by the model simulation.

The minimum, maximum, and mean differences, in feet, between simulated and derived water-level altitudes for all active model cells are shown in table 3. The mean absolute difference for all active model cells was 1.52 ft, and the mean absolute difference for the 55 model cells in the area of the Devils Lake well field was 2.17 ft (table 3). The mean absolute difference for the 54 observation wells was 1.89 ft.

Table 3. Absolute differences, in feet, between simulated and derived water-level altitudes for the steady-state model

	Number of cells	Minimum difference (feet)	Maximum difference (feet)	Mean difference (feet)
All active model cells	4,742	0	17.18	1.52
All model cells in the area of the Devils Lake well field	55	.15	6.59	2.17

The simulated mass-balance budget obtained during final calibration of the steady-state model is shown in table 4. The approximately 27.78 ft^3/s of recharge combined with the approximate model error of 1.00 ft^3/s was balanced by about 21.83 ft^3/s of discharge from evapotranspiration and about 6.95 ft^3/s of discharge from wells and springs. The combined simulated outflow from wells, Warwick Springs, and smaller springs was about 24 percent of the total outflow. The flow system in the Warwick aquifer, as simulated by the digital model, is dominated by recharge and evapotranspiration; the simulated discharge from wells comes from water that would otherwise be lost to evapotranspiration. Other stresses on the flow system are relatively minor. Water discharged from the Devils Lake well field probably comes from a distance of 2 or 3 mi as shown by the cone of depression in figure 22. The rate of leakage at model cells simulating springs was about 1.59 ft^3/s. This rate of leakage is reasonable for leakage from the aquifer into the topographically lower regions to the south and east, excluding Warwick Springs.

The steady-state model is numerically unstable and has a mass-balance error of 3.47 percent because of the large variations in saturated thickness within short lateral distances. A smaller mass-balance error would have been accompanied by more model cells going dry. The simulation required a balance between the mass-balance error and the number of model cells going dry. The only altitude-dependent fluxes in the steady-state model are from model cells that simulate evapotranspiration and small springs. The simulated fluxes from small springs generally were small, about 1.6 ft^3/s, and changed little between different simulations. Therefore, these fluxes are unlikely to be the source of mass-balance error in the steady-state simulation. The main source of error probably is evapotranspiration flux, the remaining altitude-dependent flux in the simulation. The mass-balance error is about 1.0 ft^3/s, which is about 5 percent of the evapotranspiration flux. Because evapotranspiration varies linearly throughout 10 ft of depth to water, the error represents a variation in water-level altitude of about 0.5 ft, a value that is unlikely to affect any of the major interpretations of ground-water flow that are made in this report.

Table 4. Simulated mass-balance budget obtained during final calibration of the steady-state model

[Values shown are computer generated]

Inflow to the aquifer, in cubic feet per second
    Recharge = 27.781
    Total inflow = 27.781

Outflow from the aquifer, in cubic feet per second
    Discharge from Devils Lake well field = 1.7239
    Discharge from irrigation wells = 2.31054
    Discharge from Warwick Springs = 1.33
    Discharge from small springs = 1.5887
    Discharge from evapotranspiration = 21.826
    Total outflow = 28.779

Model error (inflow minus outflow, in cubic feet per second) = -0.99724

Percent discrepancy = -3.47

Sensitivity Analysis of Steady-State Model

A limited amount of data was available for digital simulation of ground-water flow in the Warwick aquifer. Also, hydrologic-characteristic values estimated during the calibration process do not represent a unique set of values that alone can produce the simulated results. Therefore, steady-state simulations were conducted to determine the sensitivity of the model to variations in selected hydrologic characteristics of the aquifer.

Because most of the well pumpage is from the deeper part of the aquifer, a sensitivity analysis was conducted by varying hydraulic-conductivity values for layer 2. The final calibrated values for layer 2 were multiplied by 0.25, 0.5, 1, 2, 3, 4, and 8, and steady-state simulations were conducted for each set of values. The mean absolute differences between simulated water-level altitudes obtained during the sensitivity analyses and derived water-level altitudes are shown in figures 23 through 25. The mean differences are shown for all model cells as a whole (fig. 23) and for the 55 model cells in the area of the Devils Lake well field (fig. 24). For the cells representing the locations of the 54 observation wells, the mean absolute differences are shown in figure 25. The model does not appear to be sensitive to small changes in hydraulic conductivity in layer 2.

The steady-state model was tested further by varying the rate of recharge. When the final calibrated values for recharge were used in the model, the mean absolute difference between simulated and derived water-level altitudes was 1.52 ft. When recharge was multiplied by a factor of 0.75, the mean absolute difference was 2.03 ft, and when recharge was multiplied by a factor of 1.25, the mean absolute difference was 1.44 ft. The model is sensitive to small changes in recharge.

The mean absolute difference between simulated and derived water-level altitudes for the cells representing the locations of the 54 observation wells was 1.89 ft. The mean values for all active model cells probably represent a better measure of the model simulation because land-surface altitude is considered as well as the measured values of water-level altitude.

Transient Simulations Using Doubled 1992 Pumpage from the Devils Lake Well Field

To test the response of the Warwick aquifer to additional pumpage, two transient simulations were conducted for 20 time intervals of 1 year each. In these simulations, the water-level altitudes, discharges,

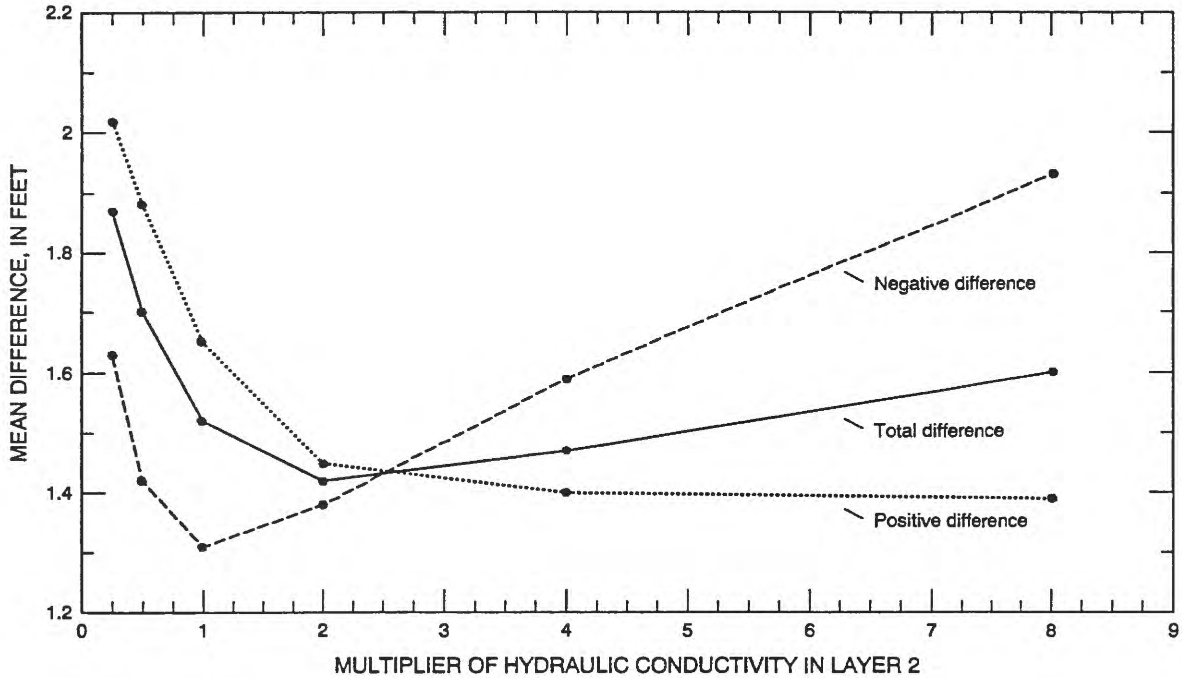


Figure 23. Differences, in feet, between simulated and derived water-level altitudes for various simulated values of hydraulic conductivity.

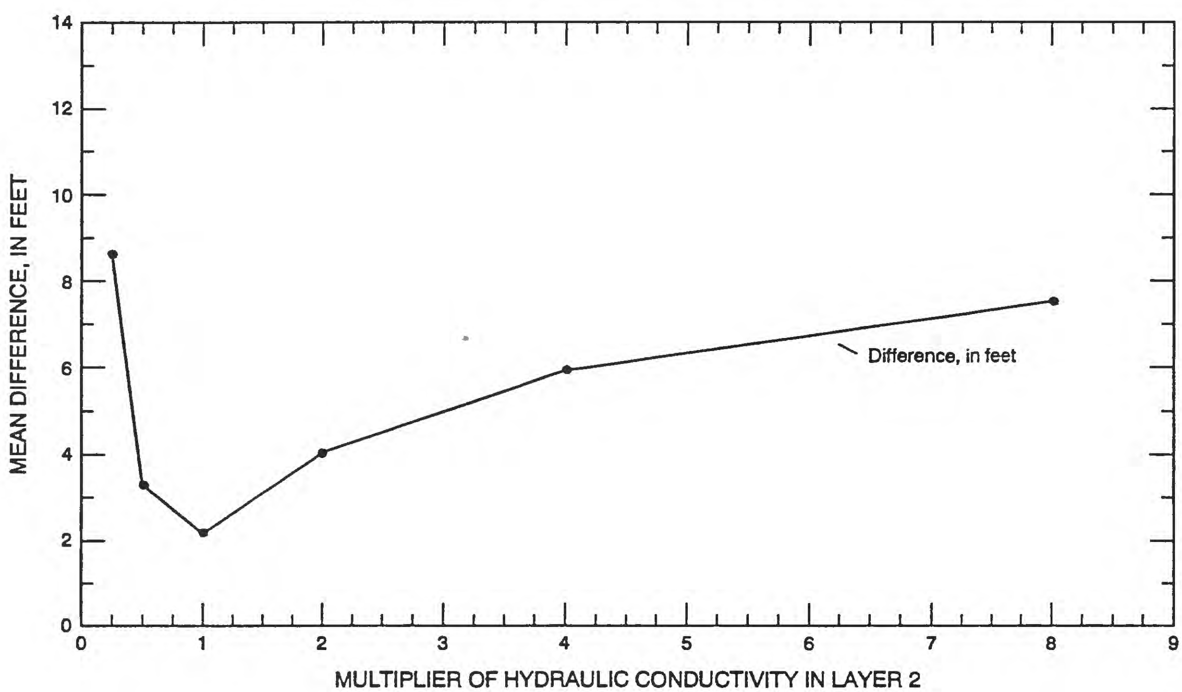


Figure 24. Differences between simulated and derived water-level altitudes for 55 model cells in the area of the Devils Lake well field.

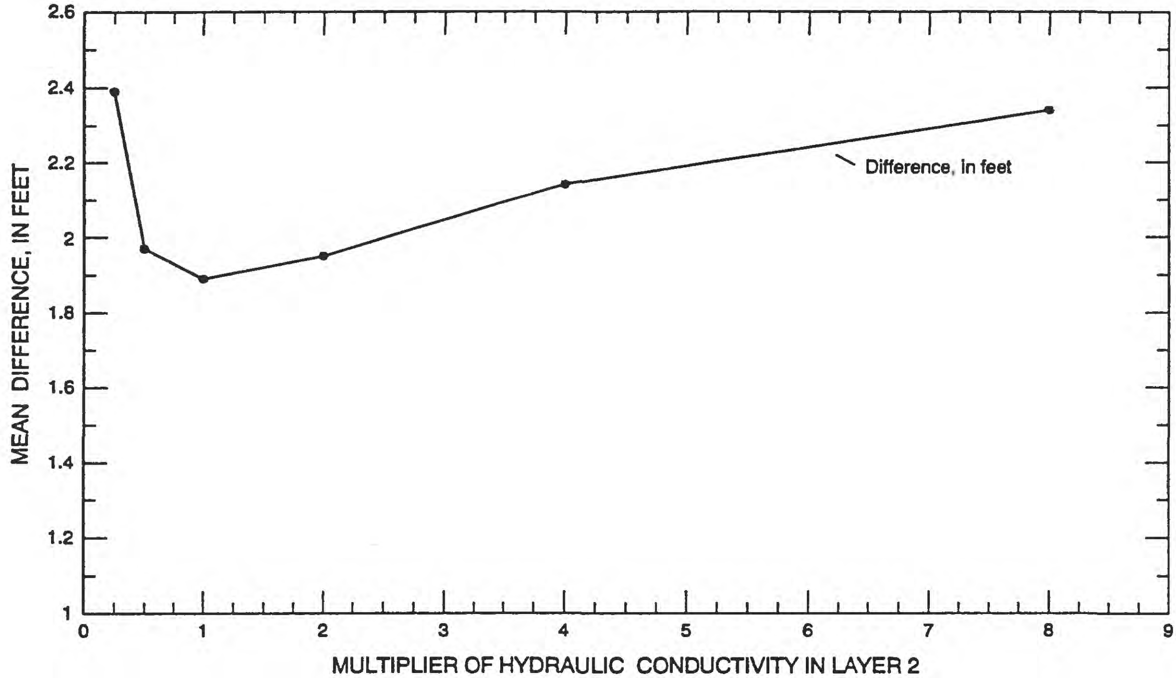


Figure 25. Differences between simulated and derived water-level altitudes for model cells representing the locations of the 54 observation wells in the Warwick aquifer.

recharge rates, irrigation well pumpage, and evapotranspiration values remained the same, but 1992 pumpage was doubled. The simulations were conducted using both the small and large storage estimates, doubled pumpage from the Devils Lake well field, 1992 irrigation pumpage, and initial water-level altitudes identical to those simulated by the October 1992 steady-state simulation (fig. 22). By using the simulated steady-state altitudes rather than the derived October 1992 altitudes, the potential effects from simulated unsteady flow were removed from the doubled-pumpage simulations.

The two simulations produced significant differences in the aquifer's simulated response to pumpage. In the simulation using the small storage estimate and doubled pumpage, model cells simulating discharge from wells in the area of the well field went dry after 13 years, and one model cell simulating irrigation pumpage went dry at 13 years. In the simulation using the large storage estimate and doubled pumpage, these model cells remained saturated for 20 years. Nevertheless, the simulated water-level altitude contours shown in figures 26 and 27 indicate that drawdowns around the well field were substantially larger in these simulations than in the steady-state simulations. Also, the mass-balance budgets indicate net change in storage obtained using the small storage estimate and doubled pumpage was about 0.70 ft^3/s (lost from the aquifer) (table 5), and net change in storage obtained using the large storage estimate and doubled pumpage was about 0.30 ft^3/s (lost from the aquifer) (table 6). These inflows, which are equal to about 20 and 9 percent, respectively, of the well field pumpage, indicate steady-state conditions have not been reached in the simulations using doubled pumpage, and drawdown probably will continue to increase until some or all of the model cells simulating discharge from wells go dry. Increases in well discharge reduce saturated thickness in the area near the wells and, thus, progressively reduce the quantity of ground-water flow toward the well field.

Interpretation of the results obtained from the simulations using doubled pumpage should consider the sensitivity of the model to recharge. More accurate determinations of the variation of recharge throughout the aquifer would help confirm whether increased pumpage would affect the aquifer to the extent indicated

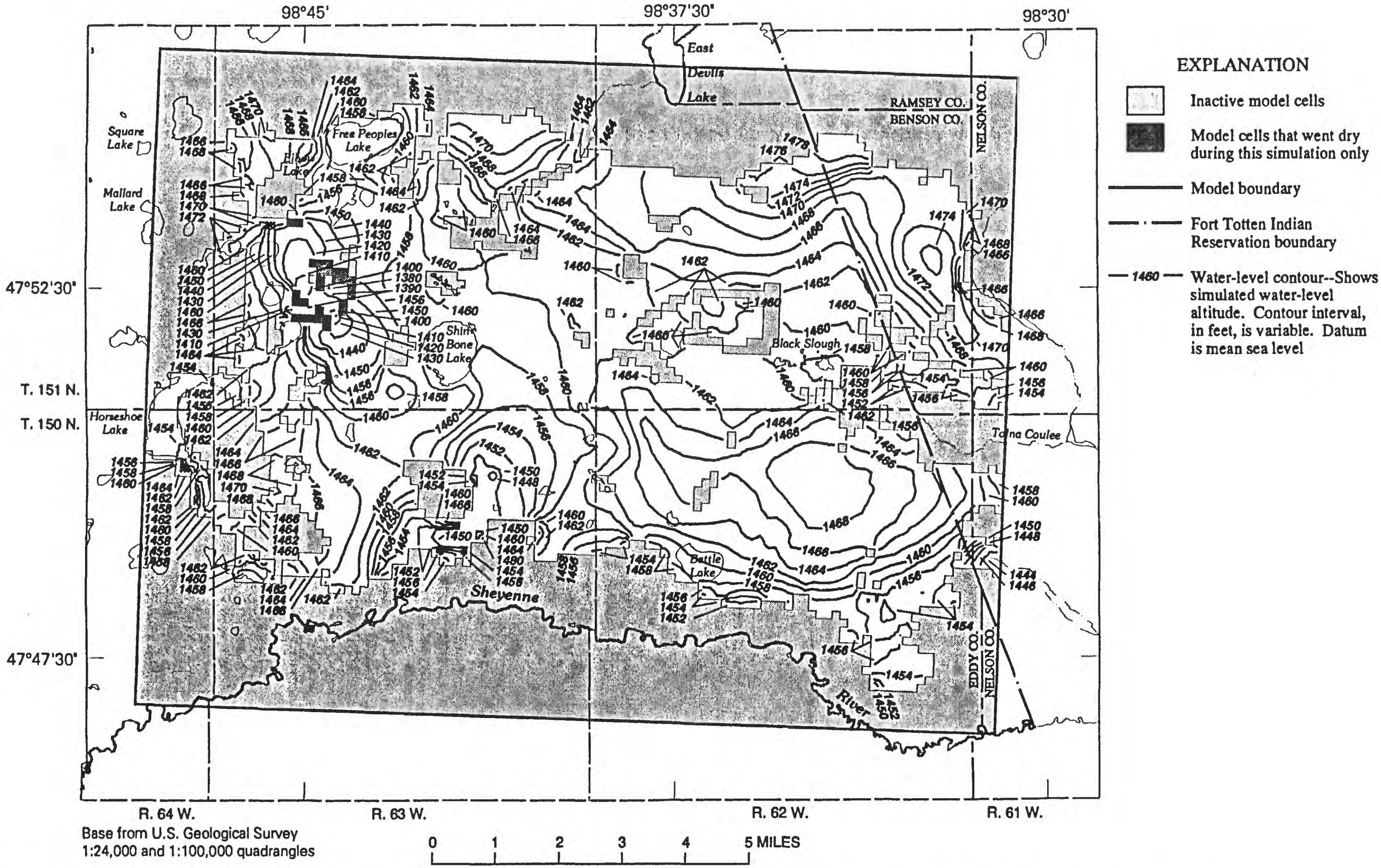


Figure 26. Simulated water-level altitudes (after 13 years) obtained from the transient model using the small storage estimate and doubled pumpage.

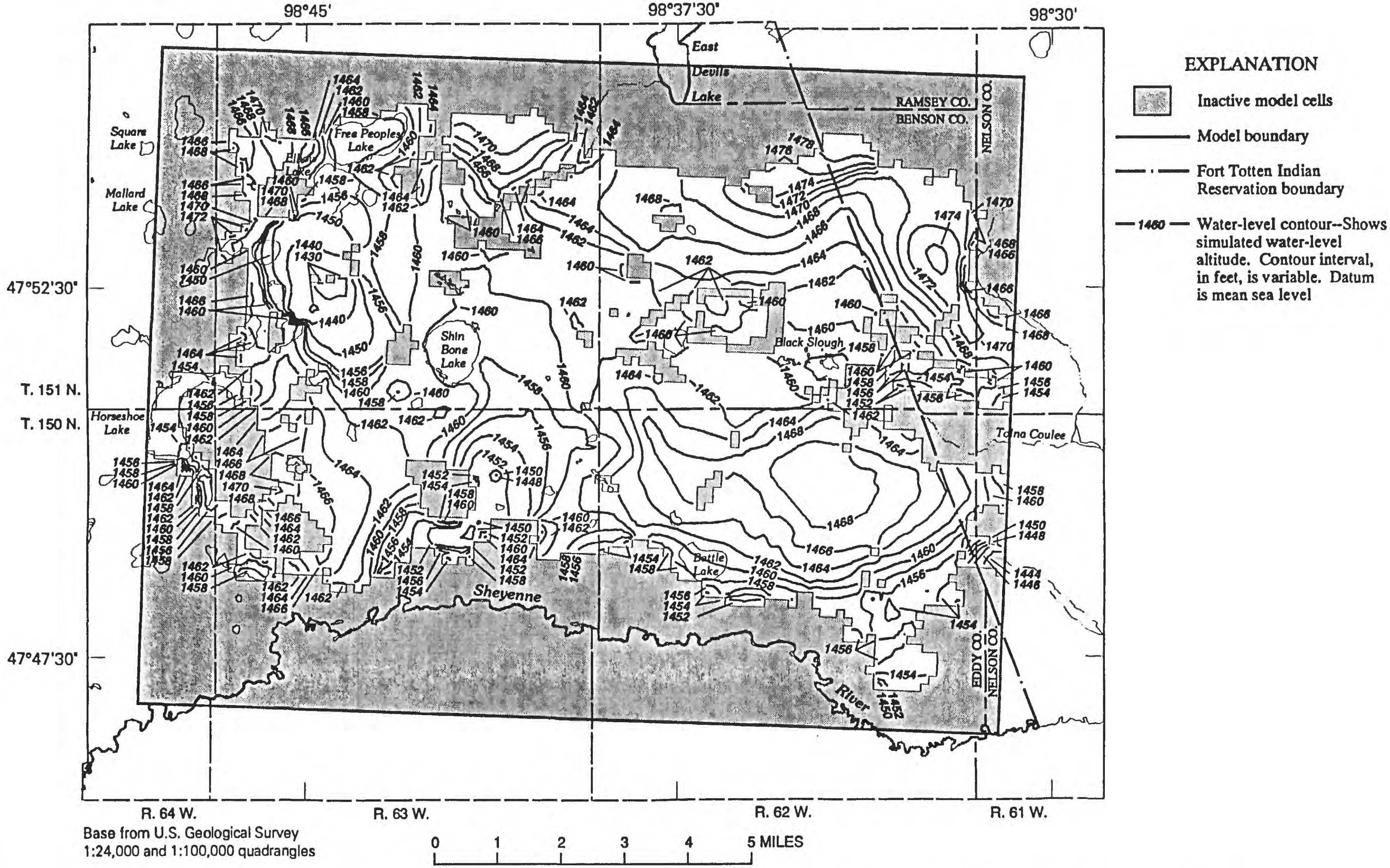


Figure 27. Simulated water-level altitudes (after 20 years) obtained from the transient model using the large storage estimate and doubled pumpage.

Table 5. Simulated mass-balance budget (after 13 years) obtained from the transient model using the small storage estimate and doubled 1992 pumpage

[Values shown are computer generated]

Inflow to the aquifer, in cubic feet per second
- Recharge = 27.582
- Storage = 0.72124
- Total inflow = 28.303

Outflow from the aquifer, in cubic feet per second
- Discharge from Devils Lake well field = 3.4478
- Discharge from irrigation wells = 2.2242
- Discharge from Warwick Springs = 1.33
- Discharge from small springs = 1.521
- Discharge from evapotranspiration = 19.817
- Storage = 0.032596
- Total outflow = 28.373

Model error (inflow minus outflow, in cubic feet per second) = -0.069414

Percent discrepancy = -0.24

Table 6. Simulated mass-balance budget (after 20 years) obtained from the transient model using the large storage estimate and doubled 1992 pumpage

[Values shown are computer generated]

Inflow to the aquifer, in cubic feet per second
- Recharge = 27.778
- Storage = 0.34598
- Total inflow = 28.124

Outflow from the aquifer, in cubic feet per second
- Discharge from Devils Lake well field = 3.4478
- Discharge from irrigation wells = 2.31054
- Discharge from Warwick Springs = 1.33
- Discharge from small springs = 1.5512
- Discharge from evapotranspiration = 20.287
- Storage = 0.022537
- Total outflow = 28.949

Model error (inflow minus outflow, in cubic feet per second) = -0.82494

Percent discrepancy = -2.89

by these simulations. Also, vertical ground-water flow may be important in the vicinity of the well field, and a finer vertical discretization may more adequately account for this vertical flow.

SUMMARY

The demand for water from the Warwick aquifer, which underlies the Fort Totten Indian Reservation in northeastern North Dakota, has been increasing during recent years. Therefore, the Spirit Lake Sioux Nation is interested in resolving questions about the quantity and quality of water in the aquifer and in developing a water-management plan for future water use. A study was conducted to evaluate the surface-water and ground-water resources of the Fort Totten Indian Reservation and, in particular, the ground-water resources in the area of the Warwick aquifer. A major component of the study, addressed by this report, was to define the ground-water flow system of the aquifer.

The Warwick aquifer consists of outwash deposits of the Warwick outwash plain that are as much as 30 feet thick and buried-valley deposits beneath the outwash plain that are as much as 200 feet thick. The aquifer is bounded on the north and west by end-moraine deposits and Devils Lake, on the south by the Sheyenne River Valley, and on the east by outwash deposits and ravines. The aquifer is underlain by Pierre Shale or by glacial till, clay, or silt. Ground-water gradients generally are small and rarely are more than 3 or 4 feet per mile. From 1982 to 1993, withdrawals from the Devils Lake well field averaged 1.5 cubic feet per second, and withdrawals from irrigation wells averaged 1.29 cubic feet per second. The combined discharge from springs may be about 3 cubic feet per second. During the early 1990s, the Warwick aquifer probably was in a steady-state condition with regard to storage change in the aquifer.

A finite-difference, three-dimensional, ground-water flow model provided a reasonable simulation of ground-water flow in the Warwick aquifer. The aquifer was divided vertically into two layers and horizontally into a grid of 83 by 109 cells, each measuring 656 feet (200 meters) per side. The steady-state simulation was conducted using 1992 pumpage rates and October 1992 water levels. The lateral boundaries of the model were assumed to be impermeable except in the area of Warwick Springs and smaller springs along the eastern and southern boundaries of the aquifer. Hydraulic connection between the Warwick aquifer and the underlying Spiritwood aquifer is very limited and has little effect on the ground-water flow system. Therefore, the bottom of the aquifer also was assumed to be impermeable.

Hydraulic-conductivity values for layer 2 and recharge rates obtained during the calibration process generally are supported by the steady-state simulation. The calibration process indicates that hydraulic conductivity in the area of the Devils Lake well field is larger than that in the rest of the aquifer. This agrees well with the large values, as much as 1,000 feet per day, indicated by previous aquifer tests. The calibration process also indicates that recharge rates are highly variable over the surface of the aquifer. Recharge rates obtained during the calibration process were lower in topographically high areas than in topographically low areas, such as ponds and lakes. The recharge rates were somewhat greater in dune deposits in the southeastern part of the aquifer than elsewhere in the aquifer.

The combined discharge of about 1.33 cubic feet per second from Warwick Springs was represented by an arc of 14 model cells, and additional discharge of about 1.59 cubic feet per second from the smaller springs was represented by 255 model cells. Pumpage rates for 1992 were used to simulate discharges from the Devils Lake well field and the irrigation wells. Simulated discharge from irrigation wells was 2.31 cubic feet per second in 1992, nearly twice the average rate for 1982-93.

Well hydrographs indicate that steady-state conditions generally existed in the Warwick aquifer in the early 1990s with regard to discharge from the Devils Lake well field and with regard to other stresses on the aquifer. Although storage in the aquifer changed from year to year, the net change in storage was minor.

The mean absolute difference between simulated and derived water-level altitudes during final calibration of the model was 1.52 feet. The assumptions made in the design of the model, including those of boundary conditions and steady-state conditions, generally are supported by this result. Simulated recharge for the steady-state model was about 27.78 cubic feet per second. Simulated discharge from evapotranspiration was about 21.83 cubic feet per second. The combined simulated outflow from wells, Warwick Springs, and smaller springs was about 6.95 cubic feet per second. The flow system in the Warwick aquifer, as simulated by the digital model, is dominated by recharge and evapotranspiration. Water is supplied to wells in the well field from a distance of 2 or 3 miles. Other stresses on the flow system are relatively minor.

Simulations of increased pumpage indicate limits to the well field's capacity. The two transient simulations were conducted for 20 time intervals of 1 year each using both the small and large storage estimates, doubled 1992 pumpage from the Devils Lake well field, 1992 irrigation pumpage, and initial water-level altitudes simulated by the October 1992 steady-state simulation. In the simulation using the small storage estimate and doubled pumpage, model cells in the area of the well field went dry after 13 years. Because an increase in discharge causes a decrease in saturated thickness, the quantity of ground-water flow toward the well field is limited.

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