HYDROGEOLOGY OF CONFINED-DRIFT AQUIFERS NEAR THE POMME DE TERRE AND CHIPPEWA RIVERS, WESTERN MINNESOTA

By G. N. Delin

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CONVERSION FACTORS AND ABBREVIATIONS

The conversion factors listed below are for the convenience of readers who prefer to use metric (International System) units rather than the inch-pound units used in this report.

Multiply Inch-Pound Unit	By	To obtain Metric Unit
foot (ft)	0.3048	meter (m)
foot per day (ft/d)	.3048	meter per day (m/d)
foot squared per day (ft^2/d)	.09294	meter squared per day (m^2/d)
cubic foot per second (ft^3/s)	.02832	cubic meter per second (m^3/s)
gallon (gal)	3.785	liter (L)
gallon per minute (gal/min)	.06308	liter per second (L/s)
million gallons per year (Mgal/yr)	.00012	cubic meter per second (m^3/s)
inch (in)	25.4	millimeter (mm)
inch per year (in/yr)	25.4	millimeter per year (mm/yr)
mile (mi)	1.609	kilometer (km)
square mile (mi^2)	2.590	square kilometer (km^2)

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ABSTRACT

Confined-drift aquifers in a 1,380-square-mile area of western Minnesota range in thickness from less than 10 feet to 114 feet. Transmissivities range from less than 1,000 square feet per day to over 16,000 square feet per day and theoretical well yields range from less than 100 gallons per minute to more than 1,800 gallons per minute.

Regional ground water flow in the confined-drift aquifers is toward the Minnesota River and locally toward smaller streams, lakes, wetlands, and wells. Water levels near high-capacity pumping wells generally fluctuate 5 to 10 feet annually, compared to annual fluctuations of 2 to 3 feet in the surficial aquifers.

Water from confined-drift aquifers generally is suitable for most uses. The water is hard to very hard and contains locally elevated concentrations of some chemical constituents. Dissolved-solids concentrations ranged from about 400 to 1,800 milligrams per liter.

A ground-water-flow model indicated that increased pumping from two of the confined aquifers simulated, the Appleton and Benson-middle aquifers, would not adversely affect water levels. The addition of 30 hypothetical wells in the Benson-middle aquifer, pumping a total of approximately 792 million gallons per year, resulted in regional water-level declines of as much as 1.4 and 2.7 feet in the surficial and Benson-middle aquifers, respectively. The addition of 28 hypothetical wells in the Appleton aquifer, pumping a total of approximately 756 million gallons per year, lowered water levels as much as 5 feet in the surficial and Appleton aquifers. Simulations of reduced recharge and increased pumping, which could represent a 3-year drought, probably would lower water levels 2 to 6 feet regionally in the surficial and confined aquifers and as much as 11 feet near aquifer boundaries. Ground-water discharge to the Pomme de Terre and Chippewa Rivers in the southern part of the study area probably would be reduced by approximately 15.2 and 7.4 cubic feet per second, respectively, as a result of the simulated drought. Mean discharge of the Pomme de Terre and Chippewa Rivers is 104 and 267 cubic feet per second, respectively.

INTRODUCTION

Ground-water withdrawals from drift aquifers have increased dramatically during the last decade in western Minnesota. The increase is primarily due to increased crop irrigation from wells following the 1976-77 drought. The MDNR (Minnesota Department of Natural Resources) received 38 applications for ground-water permits for irrigation in Swift County before 1976. Conversely, 105 applications were received in 1977 and 278 in 1984. Most ground water is obtained from surficial aquifers, but an increasing amount has been obtained from confined-drift aquifers during the last decade.

The MDNR is concerned about the effects of increased withdrawals from the confined-drift aquifers because of uncertainty about (1) long-term yields of wells open to these aquifers, (2) effects of pumping and drought on water levels and streamflow, and (3) possible interference between nearby wells pumping from the same aquifer. Consequently, the U.S. Geological Survey, in cooperation with the MDNR and the Pomme de Terre and Chippewa Ground-Water Study Steering Committee, conducted a 5-year study (1979-84) to appraise the ground-water resources along these rivers in Chippewa, Grant, Pope, Stevens, and Swift Counties.

Purpose and Scope

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

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

Objectives of the second phase of the study were to appraise the ground-water resources of confined-drift aquifers near the Pomme de Terre and Chippewa Rivers. Results of the second phase of the study are summarized in this report.

Two additional U.S. Geological Survey reports were prepared in conjunction with the second phase of this study: Delin (1984) and modeling results in Swift County, and Delin (1986) provides a detailed description of the three-dimensional ground-water-flow model constructed for this study.

Location and Description of Study Area

The study area is about 150 miles west of Minneapolis and St. Paul and covers approximately 1,380 mi^2 including parts of Chippewa, Grant, Pope, Stevens, and Swift Counties (fig. 1). The area is drained by the Pomme de Terre and Chippewa Rivers, which are tributaries of the Minnesota River. The topography is generally flat or gently rolling. Mean annual precipitation is about 24 inches (Baker and Kuehnast, 1978), with most of it occurring between May and September. Mean potential evapotranspiration is about 24.5 inches and average annual runoff is about 2 inches (Baker and others, 1979).

Previous Investigations

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

Methods of Investigation

Field work for this phase of the study was conducted during 1981-82. Hydrogeologic maps were prepared using reported data from approximately 400 wells and test holes, from files of the Minnesota Geological Survey and the U.S. Geological Survey, and geologic logs from 19 test holes drilled for this phase of the study. Geologic logs for the 19 test holes drilled for this

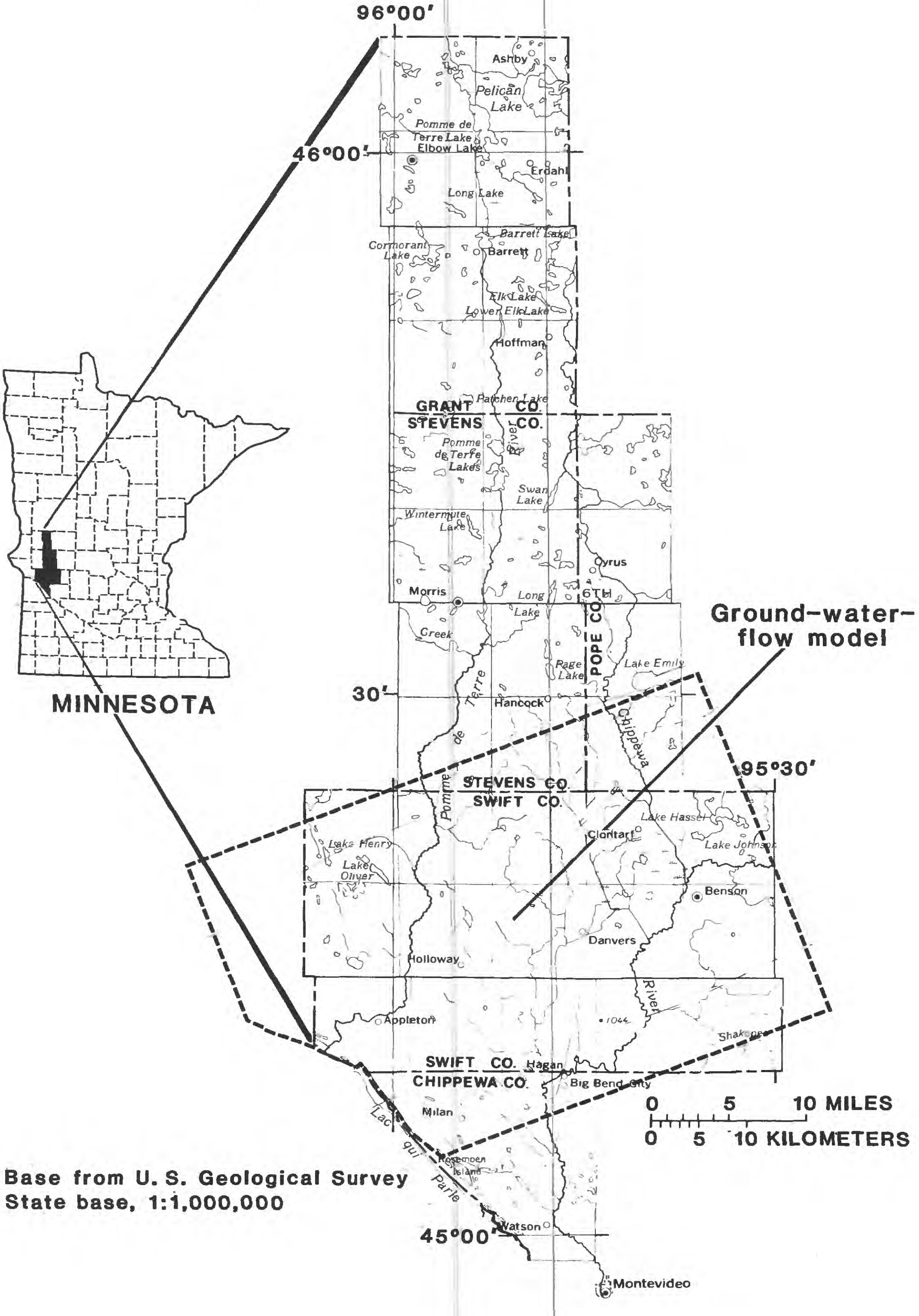


Figure 1.--Location of study area and extent of ground-water-flow model

phase of the study are located in Appendix A. These data were used to determine the areal distribution, thickness, depth of burial, and physical composition of confined-drift aquifers (hereafter called confined aquifers) in the study area. These data also were used to construct the ground-water-flow model.

Sixteen test holes were completed as observation wells to determine changes in water levels in confined aquifers and to collect water samples for chemical analysis. Water levels were measured in a total of 197 domestic, irrigation, and observation wells between November 29 and December 15, 1982. These data were used in constructing the potentiometric-surface maps in this report and in calibrating the ground-water-flow model.

Values of mean hydraulic conductivity were determined primarily through analysis of 22 aquifer tests. Aquifer hydraulic conductivity also was estimated at approximately 200 other locations from specific-capacity data. Transmissivities were determined by multiplying the mean hydraulic conductivity of each aquifer by the aquifer thickness. Transmissivity values obtained from specific-capacity and aquifer tests were used as control points during map construction. Therefore, the transmissivity contours reflect local variations in hydraulic conductivity.

Vertical hydraulic conductivity of the till was calculated through analysis of 12 aquifer tests using a method presented by Cooper (1963). These data were used in constructing the ground-water-flow model.

Water-use data were collected from the Minnesota Water-Use Data System at the MDNR. These data were used in the ground-water-flow model to simulate current (1984) ground-water usage.

A three-dimensional finite-difference model was constructed to simulate ground-water flow. The model was calibrated to steady-state conditions based primarily on hydrologic data collected for this study. Transient calibration also was conducted, based on 3 years of water-level data collected by the Swift County Soil and Water Conservation District. The model was used to estimate the effects of hypothetical pumping and drought conditions and to determine the possible effects on regional ground-water levels and streamflow.

Test-Hole and Well-Numbering System

The system of numbering wells and test holes is based on the U.S. Bureau of Land Management's system of land subdivision (township, range, and section). Figure 2 illustrates the numbering system. The first numeral of a test-hole or well number indicates the township, the second the range, and the third the section in which the point is located. Uppercase letters after the section number indicate the location within the section; the first letter denotes the 160-acre tract, the second the 40-acre tract, and the third the 10-acre tract. The letters A, B, C, and D are assigned in a counterclockwise direction, beginning in the northeast corner of each tract. The number of uppercase letters indicates the accuracy of the location number; if a point can be located within a 10-acre tract, three uppercase letters are shown in the location number. For example, the number 129.41.15ADC indicates a test hole or well located in the SW¼SE¼NE¼, Sec. 15, T.129 N., R.41 W.

Acknowledgments

The author is grateful to well owners, well drillers, and State and local agencies for data used in preparing this report. Special thanks are given to Ken Tosel who permitted an aquifer test using his irrigation well, to land owners who permitted the drilling of test holes and the installation of observation wells, and to well owners who permitted sampling of their wells and measurement of water levels.

GEOLOGY

Drift

Glacial deposits cover the entire study area. These deposits are termed drift, and consist primarily of till and of outwash sand and gravel. The deposits range in thickness from less than 100 ft near the Minnesota River to about 400 ft where they fill bedrock valleys. (fig. 3).

The drift was deposited by various mechanisms during successive glacial advances and retreats during the Wisconsin glaciation and reflects a complex glacial history. During glacial advances, till was deposited at the base of the glaciers. During periods of glacial stagnation, silt and clay was deposited in glacial ponds and lakes. During glacial retreats, melt-water streams deposited sand and gravel in stream channels, outwash plains (commonly referred to as sand plains), kames, eskers, and beach ridges. Some sand and gravel deposits were covered by till during subsequent glacial advances. These deeper sand and gravel units are present throughout the study area and are covered by 3 to 170 ft of till.

Figure 4 shows areas where till and outwash exist at land surface. The topography in till areas is rolling and irregular; in outwash areas, it is nearly flat to gently rolling.

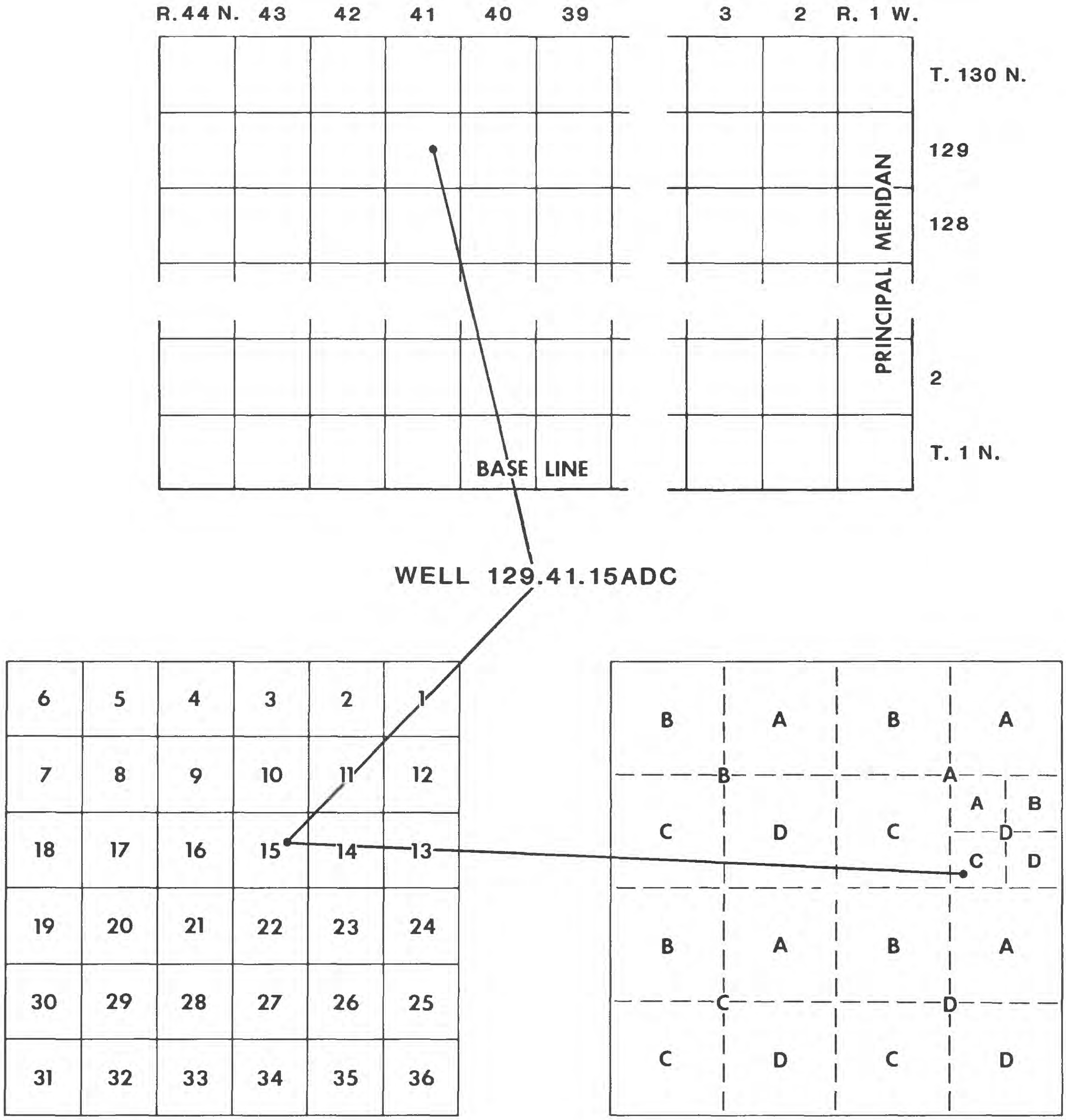
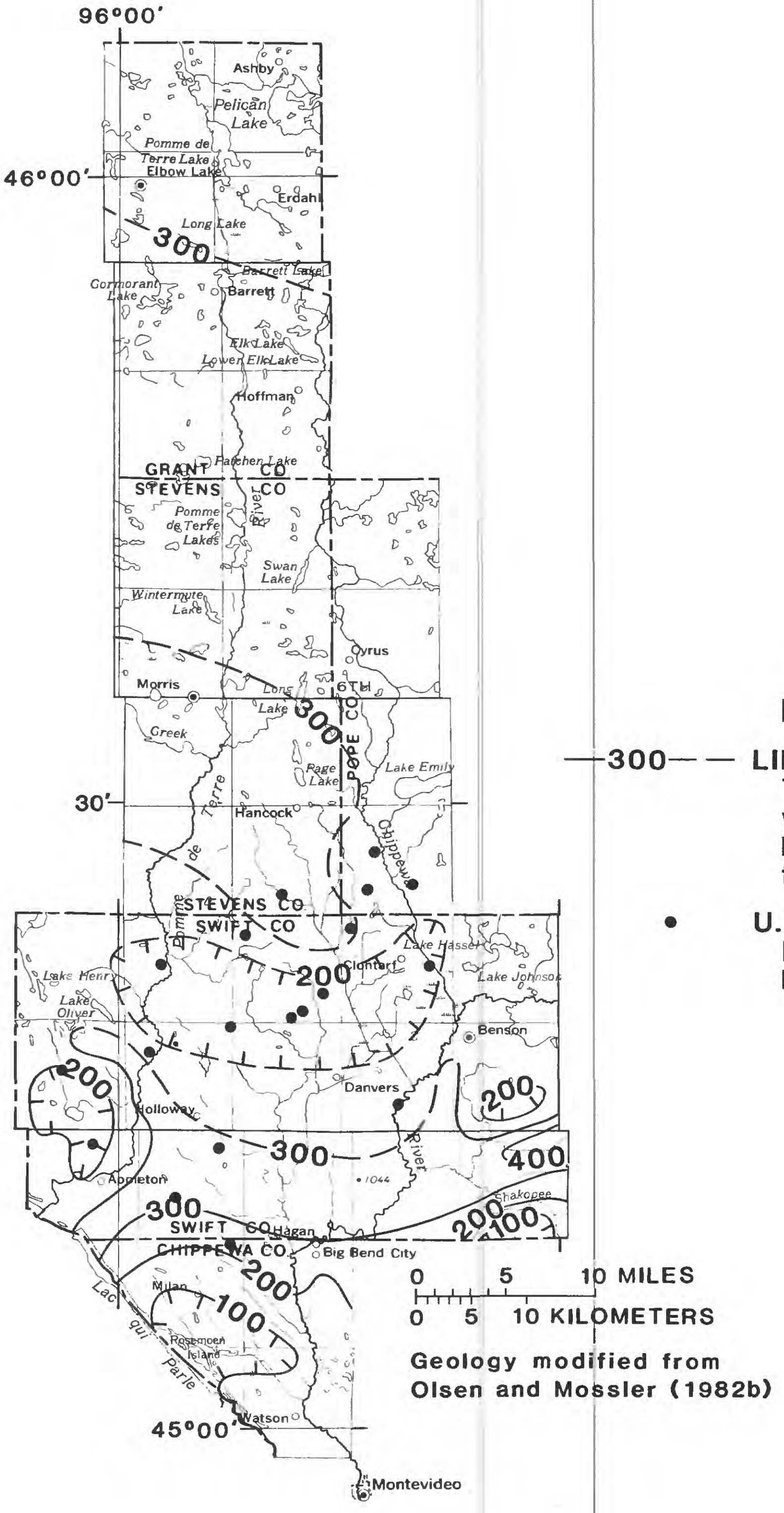


Figure 2.--Test-hole and well-numbering system



EXPLANATION

—300— — LINE OF EQUAL DRIFT THICKNESS--Dashed where approximately located. Interval 100 feet

● U. S. G. S. TEST HOLE DRILLED FOR THIS PHASE OF THE STUDY

Figure 3.--Drift thickness

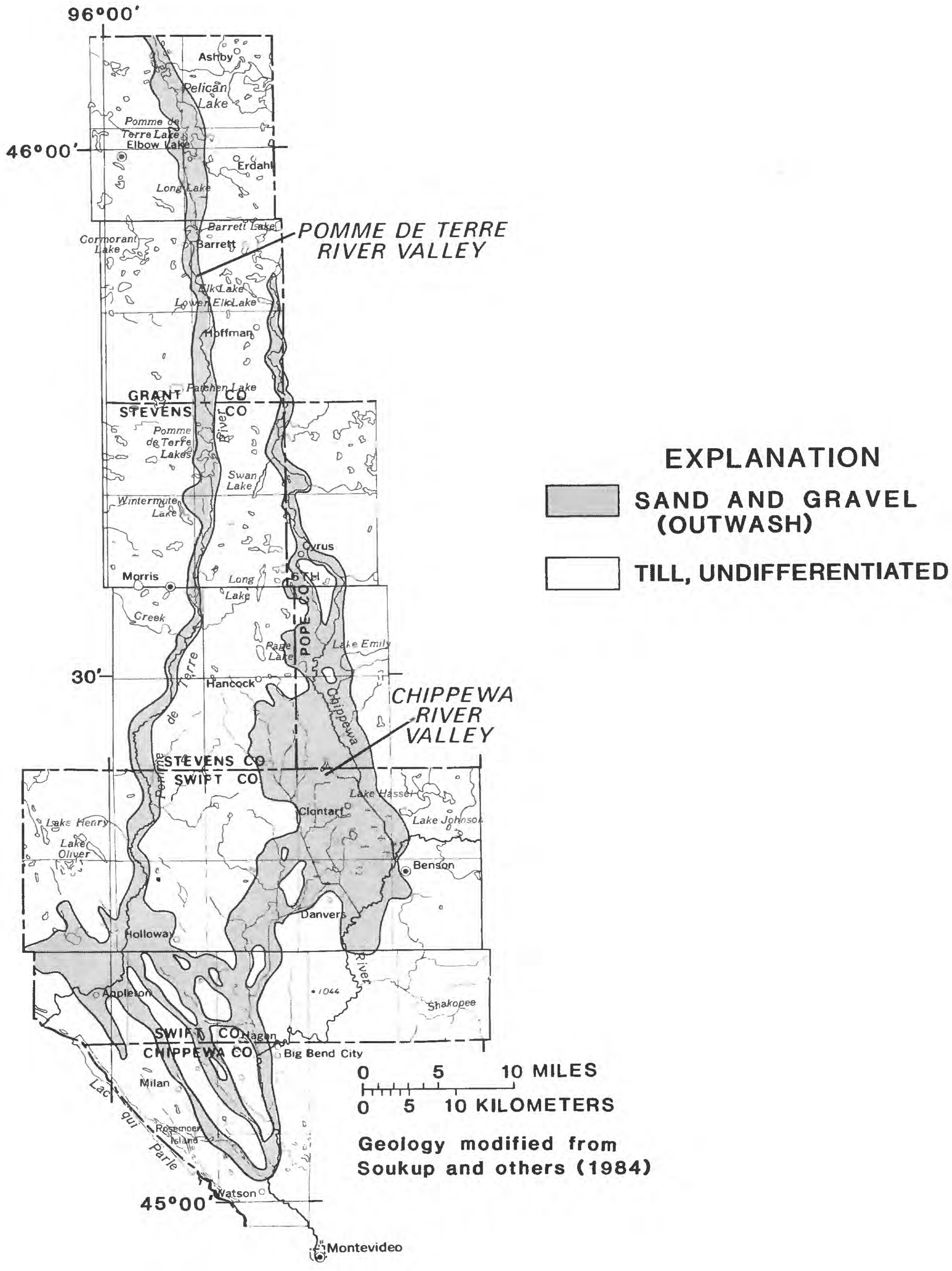


Figure 4.--Surficial geology

Bedrock

Proterozoic (Precambrian) igneous and metamorphic rocks directly underlie the drift throughout most of the study area. The rocks consist primarily of granite, gneiss, and schist. These rock types were largely inferred from gravity and magnetic data (Sims, 1970). Some outcrops are present in the Minnesota River valley in Chippewa County. Water occurs only in fractures and in weathered zones near the top of these rocks, which generally are dense with low porosity and permeability; they are not used for water supplies within the study area.

The bedrock surface is irregular, with as much as 100 ft of relief in one mile (fig. 5). An east-west trending bedrock valley in southern Swift County is the predominant feature. Two smaller northwest-southeast trending bedrock valleys form tributaries to the main valley. These bedrock valleys probably reflect the flowpaths of a glacial or preglacial drainage system. Erosion from glacial streams and ice during the Wisconsin glaciation further altered the bedrock surface.

Cretaceous deposits overlie the Proterozoic rocks in parts of Swift and Chippewa Counties (fig. 5). These discontinuous and generally semiconsolidated deposits are difficult to differentiate from drift. The maximum thickness of Cretaceous deposits penetrated during test drilling for this phase of the study is 33 ft. Although isolated wells are known to yield as much as 50 gal/min from Cretaceous formations, the deposits are not considered to be a major confined aquifer in the study area.

EXTENT AND HYDRAULIC PROPERTIES OF DRIFT UNITS

Drift in the study area has been hydrogeologically subdivided into three types: (1) sand and gravel deposits at land surface that compose the unconfined aquifers; (2) till deposits that overlie and confine deeper sand and gravel deposits, and (3) deeper sand and gravel deposits that compose the confined aquifers. The hydraulic properties of these three drift types are distinctly different and are described in the following sections.

Unconfined Aquifers

The unconfined (surficial) aquifers occur in narrow channels along the Pomme de Terre and Chippewa Rivers and as sand plains in the southern part of the study area (fig. 5). The aquifers generally consist of coarse sand and gravel in the north and fine to medium sand in the south, deposited during the last glacial retreat. Surficial aquifers commonly vary in thickness from 10 to 40 ft (Soukup and others, 1984), although the aquifer can be as much as 100 ft thick in the northern part of the Pomme de Terre River valley.

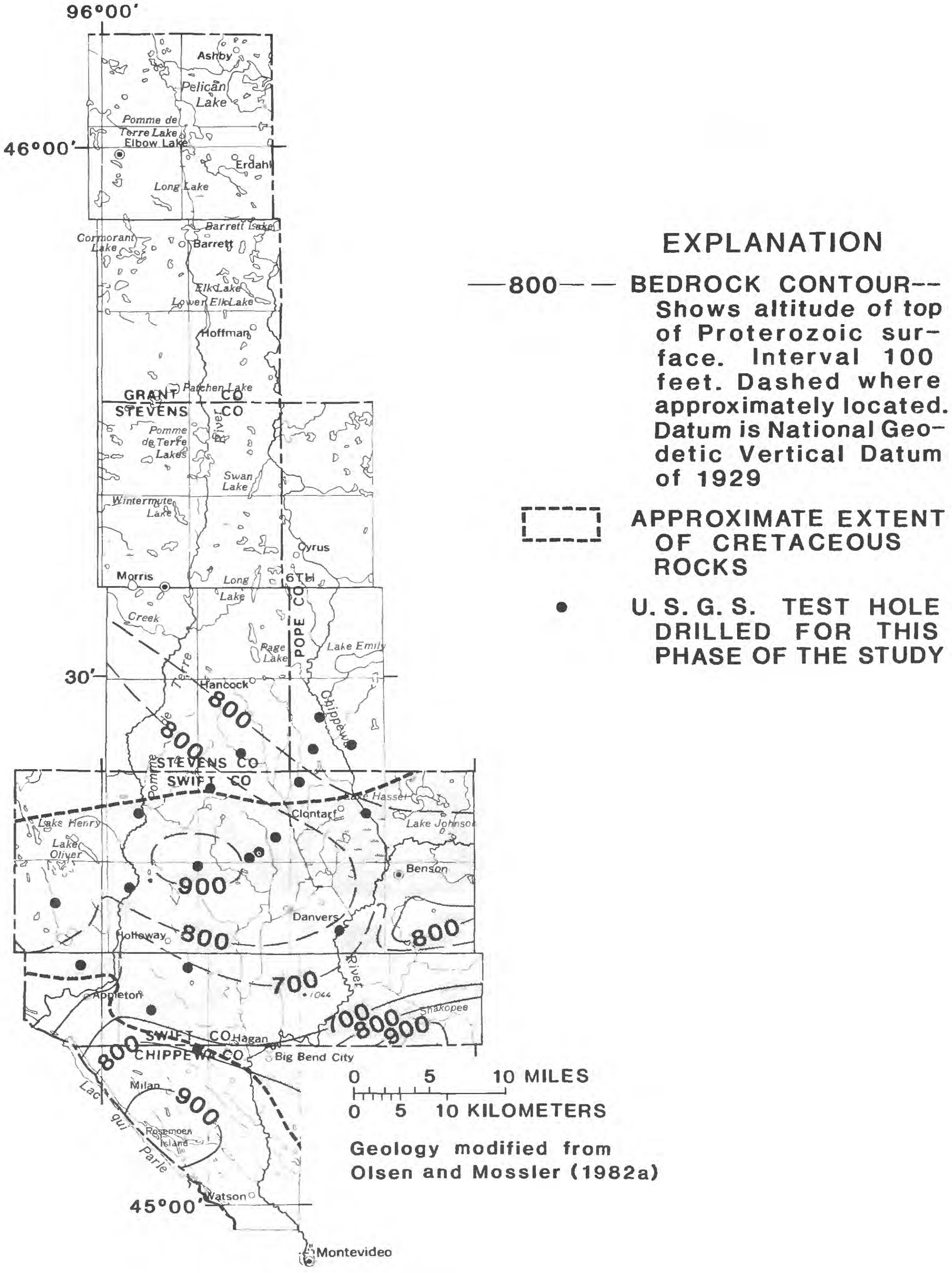


Figure 5.--Bedrock geology

Transmissivities generally range from less than 10,000 ft^2/d in the south to 35,000 ft^2/d in the north (Soukup and others, 1984). Well yields of 500 to 1,500 gal/min are probable (Soukup and others, 1984). Regional ground-water flow is south and southwest toward the Minnesota River. Locally, flow is toward the Pomme de Terre and Chippewa Rivers and high-capacity pumping wells. The reader is referred to Soukup and others (1984) for a detailed description of surficial aquifers in the study area.

Till Confining Beds

Till consists of an unsorted mixture of clay, silt, sand, gravel, and boulders generally deposited beneath stagnated or advancing glaciers. However, some clayey till may have been deposited by proglacial lakes. The gray till in the study area, although sandy, has a matrix consisting primarily of clay and silt.

The hydraulic properties of till have been considered because they control the vertical flow of ground water to and between the drift aquifers. The vertical hydraulic conductivity of till generally is much lower than the hydraulic conductivity of the drift aquifers. Therefore, till is considered to be a confining unit. The mean vertical hydraulic conductivity of till in the study area, based on analysis of 12 aquifer tests, is 0.025 ft/d. This compares favorably with the value of 0.018 ft/d for the Detroit Lakes area of Minnesota (Miller, 1982). These values of vertical hydraulic conductivity are slightly higher than those reported for other parts of the glaciated northern United States and reflect the sandy nature of till in the study area. Permeameter tests conducted by Prudic (1982), for example, indicate that the vertical hydraulic conductivity of till in Cattaraugus County, New York, ranges from 3.1×10^{-5} to 4.3×10^{-4} ft/d.

Although no field tests were made, the horizontal hydraulic conductivity of till in the study area probably is about two orders of magnitude higher than the vertical hydraulic conductivity. A value of 1 ft/d for the horizontal conductivity of alluvial clay was given by Lohman (1972). A value of 1 ft/d is also at the upper end of conductivity values for till given by Heath (1983).

Confined Aquifers

The confined aquifers are composed of saturated sand and gravel that, within the study area, are bounded above and below by lower-permeability till. These aquifers are the main source of ground-water supplies where surficial aquifers are thin or absent.

The areal extent of each confined aquifer was determined based on a modification of methods for delineating confined aquifers described by Winter (1975). Rodis (1961) describes a similar method for delineating confined aquifers. The basic assumptions for this technique are that (1) wells at a given altitude are completed in deposits that form a single aquifer, (2) the sand and gravel deposits forming each aquifer are continuous between known points of occurrence, and (3) thin, areally discontinuous sand and gravel deposits are considered to be distributed randomly, and, therefore, are ignored. These thin, discontinuous deposits could supply water sufficient for domestic purposes but not for long-term high-capacity water supplies.

Based on the above assumptions, the top and bottom elevations of sand and gravel deposits noted on test holes and well logs were plotted on a map. Deposits at common altitudes then were correlated and designated aquifers, with the aid of hydrogeologic sections and fence diagrams. When using plates 1-6, it is important to remember that deposits forming each aquifer are assumed to be continuous between known points of occurrence in each aquifer. The actual presence of an aquifer, however, must be confirmed by test drilling. In other words, the chance that a sand and gravel deposit occurs as shown on the maps is good, but the hydrogeologic maps are intended only as a guide.

Confined aquifers described in this report are named for reference purposes only. These names were based on either their vertical relation to each other (for example; Benson-upper, Benson-middle, and Benson-lower) or their proximity to a city or township (for example; Appleton, Morris, and Erdahl). The aquifer boundaries shown on each hydrogeologic map in this report represent the known areal extent of each aquifer.

The hydrologic properties of each confined aquifer of particular interest are their (1) location and areal extent, (2) composition and origin, (3) thickness and depth below land surface, (4) occurrence in relation to other aquifers, (5) hydraulic properties (hydraulic conductivity, transmissivity, and storage coefficient), and (6) known well yields. Later sections of the report describe the annual reported water use, ground-water movement, and theoretical well yields for each confined aquifer. Some of the hydrologic properties of each aquifer are summarized in table 1.

Short-term well yields depend primarily on local aquifer thickness, hydraulic conductivity, transmissivity, storage, saturated thickness, and on the condition of the well pump and screen. Long-term yields also depend on recharge rates and boundary conditions. Hydraulic conductivity and transmissivity are indicators of an aquifer's ability to yield water to wells. Transmissivity is the product of hydraulic conductivity and

Table 1.--Summary of hydrologic characterisitics for major confined aquifers near the Pomme de Terre and Chippewa Rivers, western Minnesota

Aquifer name	Approximate areal extent (mi^2)	Maximum known thickness (feet)	Average thickness (feet)	Range and average depth below land surface (feet)	Range and average hydraulic conductivity (ft/d)	Typical range in transmissivity (ft^2/d)	Range and average depth to water below land surface (feet)	Range in reported well discharge (gal/min)	Primary use of water (1984) [irrigation (I), domestic and/or stock (DS), municipal (M), or industrial (IN)]
Appleton	220	114	60	32-203 (92)	80-330 (230)	1,400-14,000	0-65 (26)	5-1,500	I,M,DS,IN
Barrett	60	49	19	68-168 (128)	50-140 (100)	1,000-3,000	8-110 (48)	25-200	DS,M
Benson-lower	80	70	45	220-273 (251)	50-480 (100)	1,000-4,000	1-20 (15)	5-500	DS
Benson-middle	520	90	30	50-200 (135)	40-150 (90)	1,000-8,000	0-80 (23)	10-1,600	I,M,DS,IN
Benson-upper	90	92	16	21-174 (73)	20-1,000 (260)	1,000-5,000	6-95 (35)	10-700	DS,I
Elbow Lake	40	37	20	168-248 (196)	6-80 (50)	800-1,600	30-74 (53)	25-85	DS
Erdahl	120	52	20	33-130 (79)	12-1,300 (200)	1,500-6,000	6-98 (52)	10-1,140	DS,I,M
Morris	430	80	16	30-213 (80)	10-1,000 (180)	1,800-10,000	14-115 (48)	8-1300	DS,I
Pomme de Terre	50	16	7	67-134 (102)	133-140 (137)	700-1,400	13-47 (30)	10-30	DS
Sanford	30	42	23	200-260 (231)	140-140 (140)	1,400-2,800	43-104 (65)	15-50	DS

thickness. Transmissivity variations reflect changes in aquifer thickness and composition. Areas of greatest transmissivity generally concide with areas of greatest thickness. Lower values of hydraulic conductivity generally indicate poor sorting of aquifer material and(or) an increase in the percentage of clay-size particles. Conversely, higher hydraulic-conductivity values generally correspond to well-sorted sands with a smaller percentage of clay-size particles. The storage coefficient is an indicator of an aquifer's ability to store or release water and determines when the effects of pumping will stabilize. Consequently, the values of hydraulic conductivity, transmissivity, and storage coefficient are important indicators of a confined aquifer's ability to yield water to wells.

Geologic logs indicate that several of the confined aquifers identified in the study area coalesce with surficial aquifers near the Pomme de Terre and Chippewa Rivers. In addition to the hydraulic connection between surficial and confined aquifers, several confined aquifers probably coalesce with other confined aquifers. Present hydrogeologic data are insufficient to delineate these areas, however.

Ten areally extensive confined aquifers were identified using the well-completion-altitude method described earlier. A description of each confined aquifer, in alphabetical order, is presented in the following sections.

Appleton Aquifer

The Appleton aquifer, located in the southwestern part of the study area near the town of Appleton, covers approximately 220 mi^2 (pl. 1a). Municipal water supplies for the town of Appleton, are obtained, in part, from this aquifer. It consists of fine to coarse sand and gravel, with lenses of silt, clay, or till as much as 20 ft thick interbedded locally in the aquifer north of Appleton. This material probably was deposited as a proglacial sand plain and subsequently covered by drift from later glacial advances. The relative complexity of deposits composing this confined aquifer indicates that it probably was the result of several glacial advances.

Fax and Beissel (1980) identified two confined aquifers in the Appleton area. These sand units are shown in section E-E' (pl. 6). Although separated locally by till, geologic logs indicate that the sand units coalesce into one aquifer near Appleton. Therefore, they were lumped together as one continuous aquifer in constructing the hydrogeologic maps presented in this report. The thicknesses of both sand units were combined in constructing the Appleton aquifer thickness map (pl. 1a). The top of the uppermost unit is shown on plate 2a.

Larson (1976) reported that the surficial aquifer is in direct hydraulic connection (coalesces) with the Appleton aquifer near Appleton. This area of interconnection has been modified slightly in this report, based on additional data, but is essentially the same area identified by Larson. The boundary between the Appleton aquifer and the surficial aquifer are shown on plate 1a and in section F-F' (pl. 6). This interconnection probably is the result of glacial action that eroded through till into the Appleton aquifer followed by deposition of the surficial aquifer over the exposed Appleton aquifer. Since the till confining layer is absent in these areas, only the surficial aquifer is present.

The maximum known thickness of the Appleton aquifer is 114 ft, but it may be more than 120 ft thick southeast of Appleton (pl. 1a). The average thickness is 60 ft. Aquifer thicknesses generally decrease to the east and southeast.

The Appleton aquifer is closest to land surface near Appleton and its depth of burial increases to the east and southeast (pl. 2a and section G-G', pl. 6). Depth below land surface to the top of the aquifer ranges from 32 ft near Appleton to 203 ft southeast of Hoffman and averages 92 ft. The aquifer generally occurs between the altitudes 960 and 820 ft (sections E-E', F-F', and G-G', pl. 6).

The Appleton aquifer is, in general, the only confined aquifer in the area. In some areas east and southeast of Appleton, however, the Benson-middle and Benson-lower aquifers occur above and below the aquifer, respectively. A surficial aquifer also is present locally above the Appleton aquifer (sections E-E', F-F', and G-G', pl. 6). The Morris aquifer also is present locally above the aquifer north of Appleton (section F-F' pl. 6).

Hydraulic properties of the Appleton aquifer were determined primarily from analysis of four aquifer tests, one of which was conducted during this study. Data from 38 specific-capacity tests supplemented the aquifer-test data. The hydraulic conductivity computed from aquifer-test data range from 80 to 330 ft/d and average 230 ft/d. Hydraulic-conductivity values obtained from specific-capacity tests were consistently lower than this average value, however, indicating that the average may be too high. Therefore, a reduced value of 140 ft/d was used in constructing the transmissivity map (pl. 3a). Transmissivities greater than 16,000 ft^2/d occur both northwest and southeast of Appleton. Storage coefficients for the aquifer range from 0.001 to 0.0001 and average 0.0002. Reported well yields for the Appleton aquifer range from 5 to 1,500 gal/min.

Results of the aquifer test conducted for this study (location: 120.43.11BCA) indicate that the aquifer may change from confined to unconfined conditions after pumping begins. Moench and Prickett (1972) provide a detailed discussion of radial flow to a pumping well for this type of pumping condition. The change from confined to unconfined conditions occurs primarily near where the Appleton aquifer and the surficial aquifer coalesce (pl. 1a). After the potentiometric surface drops below the top of the aquifer, aquifer response is similar to that of an unconfined aquifer. Therefore, drawdowns in the vicinity of the pumping well are less than would have occurred under confined conditions. Aquifer hydraulic properties were calculated for this aquifer test using the Boulton (1954) and Stallman (1954) methods for unconfined aquifers with delayed yield from storage. A transmissivity of 6,430 ft^2/d and storage coefficient of 0.2 were calculated from test results.

Barrett Aquifer

The Barrett aquifer is located in the northern part of the study area near the town of Barrett; it covers approximately 60 mi^2 (pl. 1c). Municipal water supplies for the town of Elbow Lake are obtained from the Barrett aquifer. Grain-size information for the aquifer is lacking; however, hydraulic-conductivity values obtained from four specific-capacity tests indicate that the aquifer probably consists of fine sand. Thin lenses of silt, clay, or till probably are interbedded locally in the aquifer. The aquifer material was probably deposited in proglacial melt-water stream channels and subsequently covered by drift from later glacial advances.

The maximum known thickness of the Barrett aquifer is 42 ft; the average thickness is 19 ft (pl. 1c). Depth below land surface to the top of the aquifer ranges from 68 ft near the Pomme de Terre River to 168 ft near Elbow Lake. The average depth to the aquifer is 128 ft. The relatively large range in depth to the top of the aquifer is due primarily to variations in topography. The aquifer generally occurs between the altitudes of 1,070 and 1,130 ft (sections A-A' and B-B', pl. 6). The top of the aquifer generally slopes to the north (pl. 2c).

The Pomme de Terre and Erdahl aquifers and the surficial aquifer are present locally above the Barrett aquifer (sections A-A' and B-B', pl. 6). The Morris, Elbow Lake, and Sanford aquifers are present locally below the aquifer. Geologic logs indicate that the aquifer coalesces with the surficial aquifer along the Pomme de Terre River west of Hoffman (section B-B', pl. 6).

Hydraulic properties of the Barrett aquifer were determined from analysis of four specific-capacity tests. The hydraulic conductivity ranges from 50 to 140 ft/d and averages 100 ft/d. Locally, transmissivities of 3,000 ft^2/d are probable northwest and southeast of Barrett (pl. 3c). Although data were not available to determine a storage coefficient for this aquifer, a storage coefficient of 0.0001, similar to that for other confined aquifers in the area, is probable. Reported well yields for the Barrett aquifer range from 25 to 200 gal/min.

Benson-Lower Aquifer

The Benson-lower aquifer, named after the town of Benson, is located primarily in Swift County and covers about 80 mi^2 (pl. 1b). Hydrogeologic data for this aquifer are lacking. The areal extent of the aquifer was partially inferred from bedrock topography information presented by Olsen and Mossler (1982a). Bedrock topgraphy indicates that the aquifer may extend west to the Minnesota River and north into Stevens County (pl. 1). The aquifer consists primarily of fine sand. Thin lenses of silt, clay, or till probably are interbedded locally in the aquifer. The aquifer material probably was deposited in valleys eroded in Proterozoic bedrock by glacial melt-water streams, and subsequently was covered by drift from later glacial advances.

The maximum known thickness of the Benson-lower aquifer is 70 ft and the average thickness is 45 ft (pl. 1b). Depth below land surface to the top of the aquifer ranges from 220 ft near Holloway to 273 ft near Benson. Depth to the top of the aquifer averages 261 ft. The aquifer generally occurs between the altitudes of 730 and 800 ft (sections H-H' and I-I' pl. 6). The top of the aquifer generally slopes to the east (pl. 2b).

The Appleton, Benson-middle, and Benson-upper aquifers and the surficial aquifer are present locally above the Benson-lower aquifer (sections H-H' and I-I' pl. 6), which is the lowermost confined aquifer in the area.

Hydraulic properties of the Benson-lower aquifer were estimated from analysis of one specific-capacity test and grain-size analyses. Based on these data, aquifer hydraulic conductivity probably is about 100 ft/d. Locally, transmissivities of 5,000 ft^2/d are probable (pl. 3b). Although data were not available to determine a storage coefficient for this aquifer, a storage coefficient of 0.0001, similar to that for other confined aquifers in the area, is probable. Reported well yields for Benson-lower aquifer range from 5 to 500 gal/min.

Benson-Middle Aquifer

The Benson-middle aquifer is located between the between cities of Morris, Holloway, and Benson (pl. 1c). This is the most areally extensive confined aquifer in the study area, covering approximately 520 mi^2 (pl. 1c). Municipal water supplies for the city of Benson are obtained from this aquifer.

The aquifer consists primarily of fine sand to gravel with clay, silt, or till interbedded locally. The aquifer material probably was deposited as a proglacial sand plain, perhaps during several glacial advances, and subsequently was covered by drift from later glacial advances.

The Benson-middle aquifer generally occurs as one continuous sand unit (sections C-C', D-D', H-H', and I-I', pl. 6). Near Danvers, however, the aquifer is split into two interconnected confined sand units separated by 10 to 30 ft of till. The thicknesses of both sand units were combined in constructing the thickness map (pl. 1c), and the configuration of the top of the uppermost sand unit is shown in plate 2c.

The maximum known thickness of the Benson-middle aquifer is 90 ft and the average thickness is 30 ft (pl. 1c). Depth below land surface to the top of the aquifer ranges from 50 ft near Holloway to 200 ft southeast of Benson. Depth to the top of the aquifer averages 135 ft. The aquifer generally occurs between the altitudes of 830 and 960 ft (sections C-C', D-D', H-H', and I-I', pl. 6). The top of the aquifer generally slopes to the west (pl. 2c). Several northwest-southeast trending buried ridges and valleys occur at the top of the aquifer (pl. 2c), which may indicate that parts of the aquifer were deposited within or on glacial ice and that the present structures result from melting of the glacial ice.

The Benson-upper aquifer and the surficial aquifer are present locally above the Benson-middle aquifer (sections C-C', D-D', H-H', and I-I', pl. 6). The Appleton and Benson-lower aquifers are present locally below the aquifer. Geologic logs indicate that the aquifer coalesces with the surficial aquifer along the Pomme de Terre River northwest of Holloway (pl. 1c).

Hydraulic properties of the Benson-middle aquifer were determined from analysis of data of 16 aquifer tests, four specific-capacity tests, and drill cuttings. The hydraulic conductivity, determined from aquifer tests, ranges from 40 to 150 ft/d and averages 90 ft/d. The distribution of hydraulic-conductivity values obtained from specific-capacity tests and drill cuttings, however, indicate that this value probably is too low and is not a good approximation of the average for the aquifer. Therefore, following incorporation of values obtained from specific-capacity tests and drill cuttings, an average hydraulic-conductivity value of 140 ft/d was caculated and used in constructing the transmissivity map (pl. 3c).

Transmissivities of 10,000 ft^2/d occur in a narrow band located north of Clontarf and transmissivities greater than 8,000 ft^2/d are common near Danvers (pl. 3c). The storage coefficient ranges from 0.00003 to 0.0004 and averages about 0.0002. Reported well yields from the Benson-middle aquifer range from 10 to 1,600 gal/min.

Benson-Upper Aquifer

The Benson-upper aquifer is located in the central part of the study area extending from near Morris to south of Benson, covering approximately 90 mi^2 (pl. 1a). The aquifer consists of very fine to coarse sand and gravel interbedded locally with thin lenses of silt, clay, or till. The aquifer material probably was deposited in proglacial melt-water stream channels and subsequently was covered by drift from later glacial advances.

The maximum known thickness of the Benson-upper aquifer is 92 ft, northeast of Clontarf, and the average thickness is 16 ft (pl. 1a). Depth below land surface to the top of the aquifer ranges from 21 ft near Benson to 174 ft in the north; it averages 73 ft. The aquifer generally occurs between the altitudes of 980 and 1,020 ft (sections C-C', D-D, I-I, and H-H', pl. 6). The top of the aquifer is irregular (pl. 2a).

The Morris aquifer and the surficial aquifer are present locally above the Benson-upper aquifer (sections C-C' and D-D', pl. 6). The Benson-middle and Benson-lower aquifers are present locally below the aquifer. Geologic logs indicate that the aquifer coalesces with the surficial aquifer along the Pomme de Terre River east of Clontarf (pl. 1a).

Hydraulic properties of the Benson-upper aquifer were determined from analysis of data from one aquifer test and two specific-capacity tests. The hydraulic conductivity ranges from 20 to 1,000 ft/d and averages 260 ft/d. Transmissivities generally are less than 5,200 ft^2/d, but transmissivities greater than 10,000 ft^2/d are probable east of Hancock (pl. 3a). Because of the variability in hydraulic conductivity of this aquifer, transmissivities shown on plate 3a may differ greatly from actual transmissivites. Although data were not available to determine a storage coefficient for this aquifer, a storage coefficient of 0.0001, similar to that for other confined aquifers in the area, is probable. Reported well yields from the Benson-upper aquifer range from 10 to 700 gal/min.

Elbow Lake Aquifer

The Elbow Lake aquifer, located in the northern part of the study area south of the town of Elbow Lake, covers about 40 mi^2 (pl. 1a). Grain-size information for the aquifer is lacking; however, hydraulic-conductivity values obtained from four specific-capacity tests indicate that the aquifer probably consists of very fine to fine sand. Thin lenses of silt, clay, or till probably are interbedded locally in the aquifer. The aquifer material probably was deposited in proglacial melt-water stream channels and subsequently covered by drift from later glacial advances.

The maximum known thickness of the Elbow Lake aquifer is 37 ft and the average thickness is 20 ft (pl. 1a). Depth below land surface to the top of the aquifer ranges from 168 ft east of Hoffman to 248 ft near Elbow Lake. Depth to the top of the aquifer averages 196 ft. The aquifer generally occurs between the altitudes of 960 and 1,010 ft (sections A-A' and B-B', pl. 6). The top of the aquifer generally slopes to the north (pl. 2c).

The Pomme de Terre, Barrett, and Morris aquifers locally are present above the Elbow Lake aquifer (sections A-A' and B-B', pl. 6). The Sanford aquifer is present locally below the aquifer.

Hydraulic properties of the Elbow Lake aquifer were estimated from analysis of four specific-capacity tests. The hydraulic conductivity ranges from 6 to 80 ft/d and averages 50 ft/d. Transmissivities of 1,600 ft^2/d are probable locally north of Elbow Lake. Although data were not available to determine a storage coefficient for this aquifer, a storage coefficient of 0.0001, similar to that for other confined aquifers in the area, is probable. Reported well yields from the Elbow Lake aquifer are less than 85 gal/min.

Erdahl Aquifer

The Erdahl aquifer, in the northern part of the study area near the towns of Erdahl and Hoffman, covers approximately 120 mi^2 (pl. 1a). Water supplies for the town of Ashby and Erdahl Township are obtained from this aquifer. Grain-size information for the aquifer is lacking; however, hydraulic-conductivity values obtained from 10 specific-capacity tests indicate that the aquifer probably consists of very fine to coarse sand and gravel. Thin lenses of silt, clay, or till probably are interbedded locally in the aquifer. The aquifer material probably was deposited in a proglacial sand plain and in melt-water stream channels and was covered subsequently by drift from later glacial advances.

The maximum known thickness of the Erdahl aquifer is 52 ft and the average thickness is 20 ft (pl. 1a). Depth below land surface to the top of the aquifer ranges from 33 ft near the Pomme de Terre River to 130 ft east of Erdahl. Depth to the top of the aquifer averages 52 ft. The aquifer generally occurs between the altitudes 1,140 and 1,200 ft (sections A-A' B-B', and C-C', pl. 6). The top of the aquifer generally slopes to the west (pl. 2a).

The surficial aquifer is present locally above the Erdahl aquifer along the Chippewa River near Hoffman (section C-C', pl. 6). The Morris, Pomme de Terre, and Barrett aquifers are present locally below the aquifer (sections A-A', B-B', and C-C', pl. 6.) Geologic logs indicate that the aquifer coalesces with the surficial aquifer along the Pomme de Terre River north of Barrett and along the Chippewa River north of Cyrus (section A-A' and C-C', pl. 6).

Hydraulic properties of the Erdahl aquifer were estimated from analysis of 10 specific-capacity tests. The hydraulic conductivity ranges from 12 to 1,300 ft/d and averages approximately 200 ft/d. Transmissivities of 6,000 ft^2/d are common in Erdahl Township and northeast of Cyrus, and transmissivities of 20,000 ft^2/d are probable in areas of high hydraulic conductivity. Although data were not available to determine a storage coefficient for this aquifer, a storage coefficient of 0.0001, similar to that for other aquifers in the area, is probable. Reported well yields from the Erdahl aquifer range from 10 to 1,140 gal/min.

Morris Aquifer

The Morris aquifer, named after the town of Morris, located in the central part of the study area extending from Elbow Lake in Grant County to parts of Swift County north of the town of Holloway, covers about 430 mi^2 (pl. 1c). The aquifer also is present in isolated areas north of Appleton and north of Danvers. The aquifer consists of very fine to coarse sand and gravel interbedded locally with thin lenses of silt, clay, or till. The aquifer material probably was deposited as a sand plain and in proglacial melt-water stream channels, and subsequently was covered by drift from later glacial advances.

The Morris aquifer generally occurs as one continuous sand unit (sections A-A', B-B', C-C', and D-D', pl. 6). However, north of Morris the aquifer splits into several interconnected confined sand and gravel units separated by 5 to 30 ft of till. The thickness and top of this aquifer were constructed using the same methodology as described for interconnected sand units in the Benson-middle aquifer.

The maximum known thickness of the Morris aquifer is 80 ft and the average thickness is 16 ft (pl. 1b). Depth below land surface to the top of the aquifer ranges from about 20 ft near the Pomme de Terre and Chippewa Rivers to 180 ft near Elbow Lake. Depth to the top of the aquifer averages 80 ft. The aquifer generally occurs between altitudes of 1,020 and 1,100 ft (sections A-A', B-B', C-C', and D-D', pl. 6). The top of the aquifer generally slopes to the west and south from a structural high approximately 8 miles northeast of Morris (pl. 2b).

The Barrett, Pomme de Terre, and Erdahl aquifers locally are present above the Morris aquifer (sections A-A', B-B', C-C', and D-D', pl. 6). The surficial aquifer also is present locally above the aquifer near the Pomme de Terre River north of Perkins Lake and along the Chippewa River north of Lake Emily. The Appleton, Benson-upper, Benson-middle, Elbow Lake, and Sanford aquifers are present locally below the aquifer. Geologic logs indicate that the aquifer coalesces with the surficial aquifer along the Pomme de Terre River from just north of Appleton to north of Morris and along the Chippewa River north of Danvers and south of Hancock (pl. 1b).

Hydraulic properties of the Morris aquifer were estimated from analysis of 23 specific-capacity tests. The hydraulic conductivity range from 10 to 1,000 ft/d. A hydraulic conductivity of 180 ft/d was used in constructing the transmissivity map (pl. 3b). Locally, transmissivities of 12,000 ft^2/d are probable east and northeast of Morris. Although data were not available to determine a storage coefficient for this aquifer, a storage coefficient of 0.0001, similar to that for other confined aquifers in the area, is probable. Reported well yields from the Morris aquifer range from 8 to 1,300 gal/min.

Pomme de Terre Aquifer

The Pomme de Terre aquifer is located in the northern part of the study area near Pomme de Terre Lake; it covers 50 mi^2 (fig. 6). Grain-size information for the aquifer is lacking; however, hydraulic-conductivity values obtained from two specific-capacity tests indicate that the aquifer probably consists of medium sand. Thin lenses of silt, clay, or till probably are interbedded locally in the aquifer. The aquifer material probably was deposited in a proglacial sand plain and subsequently covered by drift from later glacial advances.

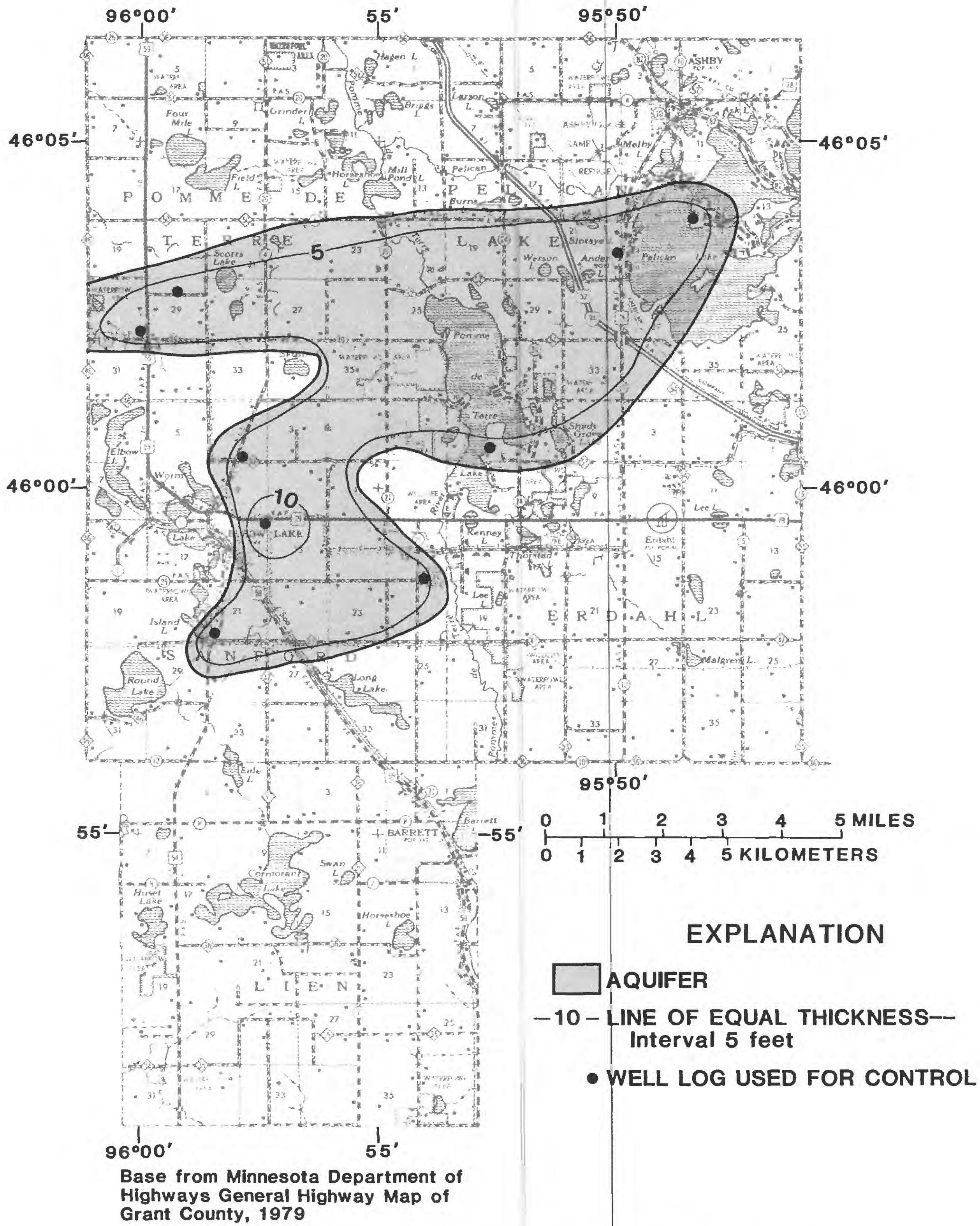


Figure 6.--Thickness of the Pomme de Terre aquifer

The maximum known thickness of the Pomme de Terre aquifer is 16 ft and the average thickness is 7 ft (fig. 6). Depth below land surface to the top of the aquifer ranges from 67 ft north of Elbow Lake to 134 ft near Pelican Lake. The average depth to the aquifer is 102 ft. The aquifer generally occurs between the altitudes of 1,100 and 1,130 ft (section A-A' pl. 6). The top of the aquifer generally slopes toward Pomme de Terre Lake (fig. 7).

The Erdahl aquifer and the surficial aquifer are present locally above the Pomme de Terre aquifer (sections A-A', pl. 6). The Barrett, Morris, Elbow Lake, and Sanford aquifers are present locally below the aquifer.

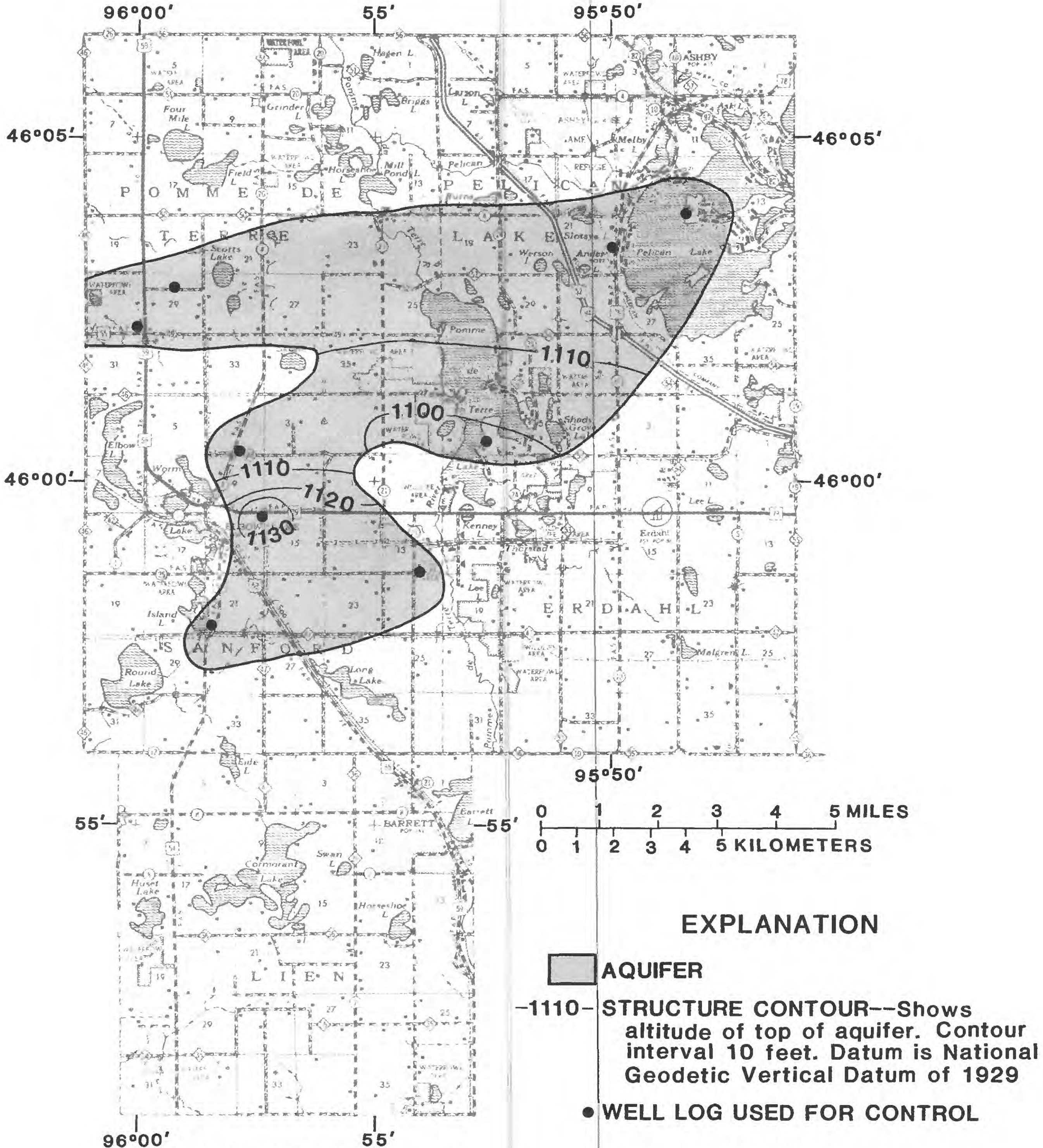
Hydraulic properties of the Pomme de Terre aquifer were estimated from analysis of two specific-capacity tests. The hydraulic-conductivity values were 133 to 140 ft/d. Transmissivities generally range between 700 and 800 ft^2/d, but transmissivities of 1,400 ft^2/d are probable where the aquifer is thickest east of Elbow Lake (fig. 8). Although data were not available to determine a storage coefficient for this aquifer, a storage coefficient of 0.0001, similar to that for other confined aquifers in the area, is probable. Reported well yields from the Pomme de Terre aquifer are less than 30 gal/min.

Sanford Confined Aquifer

The Sanford aquifer, named after the township of Sanford, located in the northern part of the study area generally south of Elbow Lake, covers approximately 30 mi^2 (fig. 9). Grain-size information for the aquifer is lacking; however, the hydraulic-conductivity value derived from a specific-capacity test indicates that the aquifer probably consists of medium sand. Thin lenses of silt, clay, or till probably are interbedded locally in the aquifer. The aquifer material probably was deposited in proglacial melt-water stream channels and subsequently was covered by drift from later glacial advances.

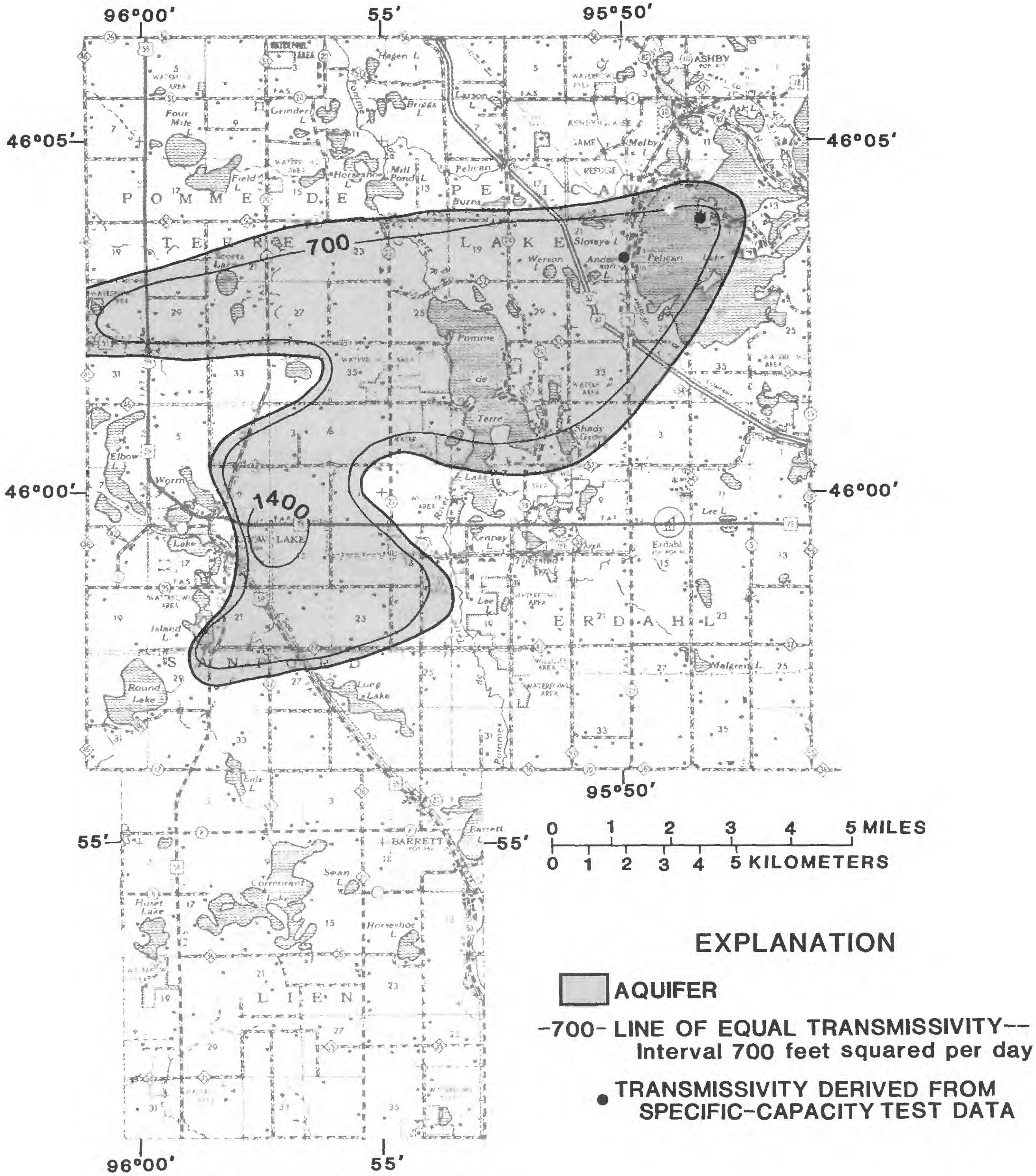
The maximum known thickness of the Sanford aquifer is 42 ft and the average thickness is 23 ft (fig. 9). Depth below land surface to the top of the aquifer ranges from 200 ft north of Elbow Lake to 260 ft near Round Lake. Depth to the top of the aquifer averages 231 ft. The aquifer generally occurs between the altitudes of 900 and 950 ft (sections A-A', pl. 6). The top of the aquifer generally slopes to the west (fig. 10).

The Pomme de Terre, Elbow Lake, Morris, and Barrett aquifers and the surficial aquifer are present locally above the Sanford aquifer (sections A-A' and B-B', pl. 6). The Sanford aquifer is the lowermost confined aquifer in the area.



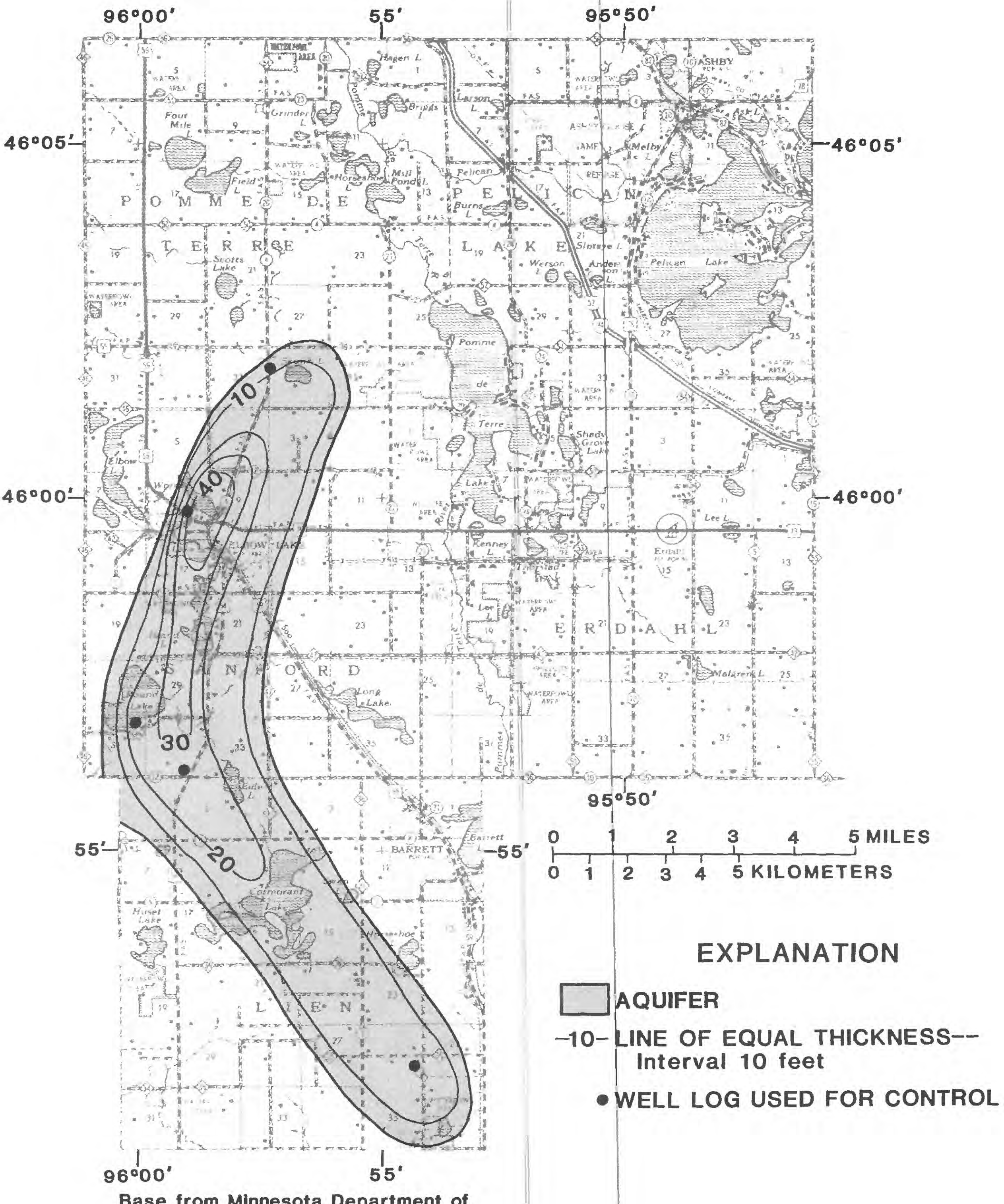
Base from Minnesota Department of Highways General Highway Map of Grant County, 1979

Figure 7.--Configuration of top of Pomme de Terre aquifer



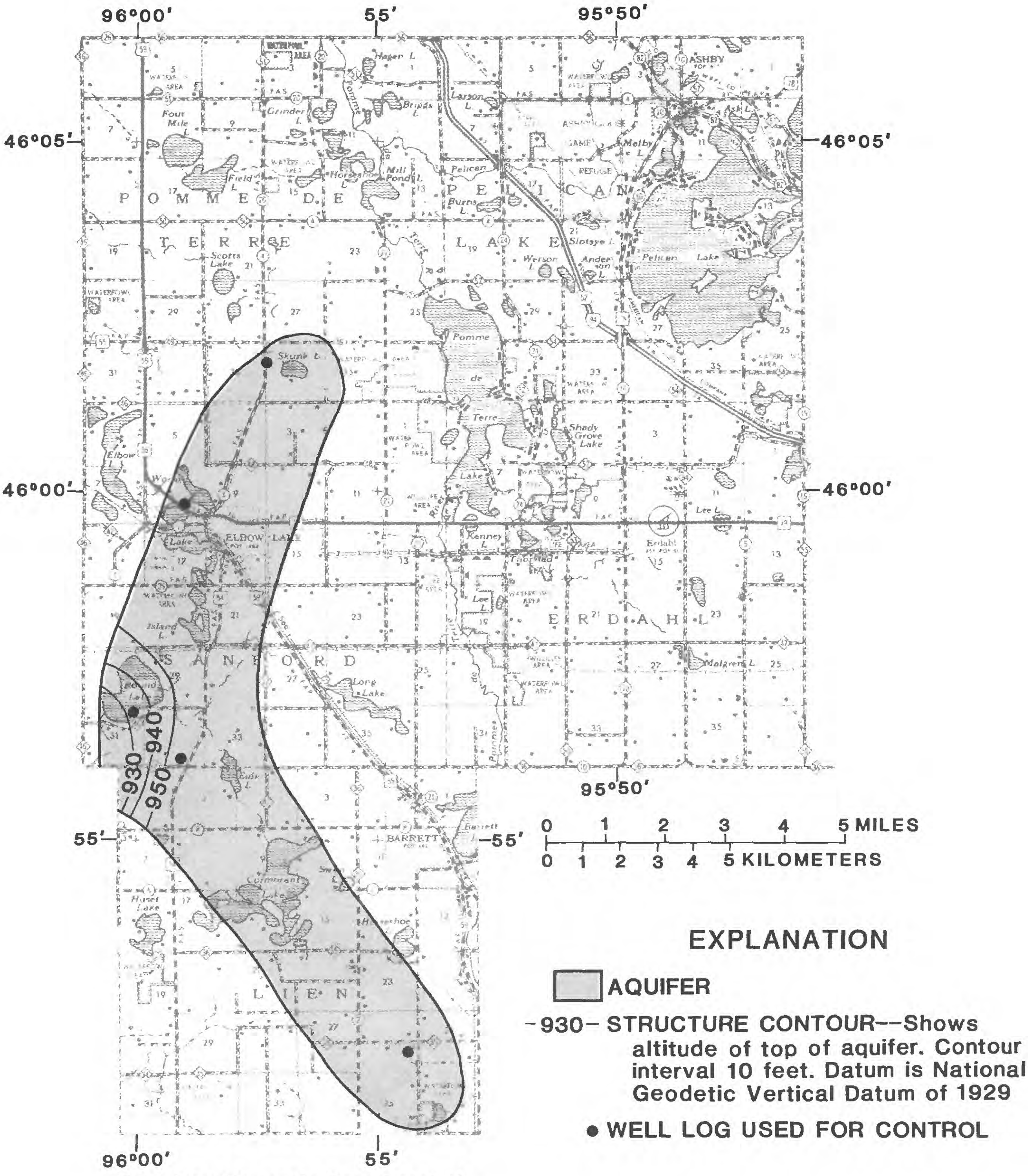
Base from Minnesota Department of Highways General Highway Map of Grant County, 1979

Figure 8.--Transmissivity of the Pomme de Terre aquifer



Base from Minnesota Department of Highways General Highway Map of Grant County, 1979

Figure 9.--Thickness of the Sanford aquifer



Base from Minnesota Department of Highways General Highway Map of Grant County, 1979

Figure 10.--Configuration of top of Sanford aquifer

Hydraulic properties of the Sanford aquifer were estimated from analysis of one specific-capacity test. A hydraulic conductivity of 140 ft/d was estimated. Based on this value of hydraulic conductivity, transmissivities generally range between 1,400 and 2,800 ft^2/d (fig. 11). Although data were not available to determine a storage coefficient for this aquifer, a storage coefficient of 0.0001, similar to that for other confined aquifers in the area, is probable. Reported well yields from the Sanford confined aquifer are less than 50 gal/min.

GROUND-WATER HYDROLOGY OF CONFINED AQUIFERS

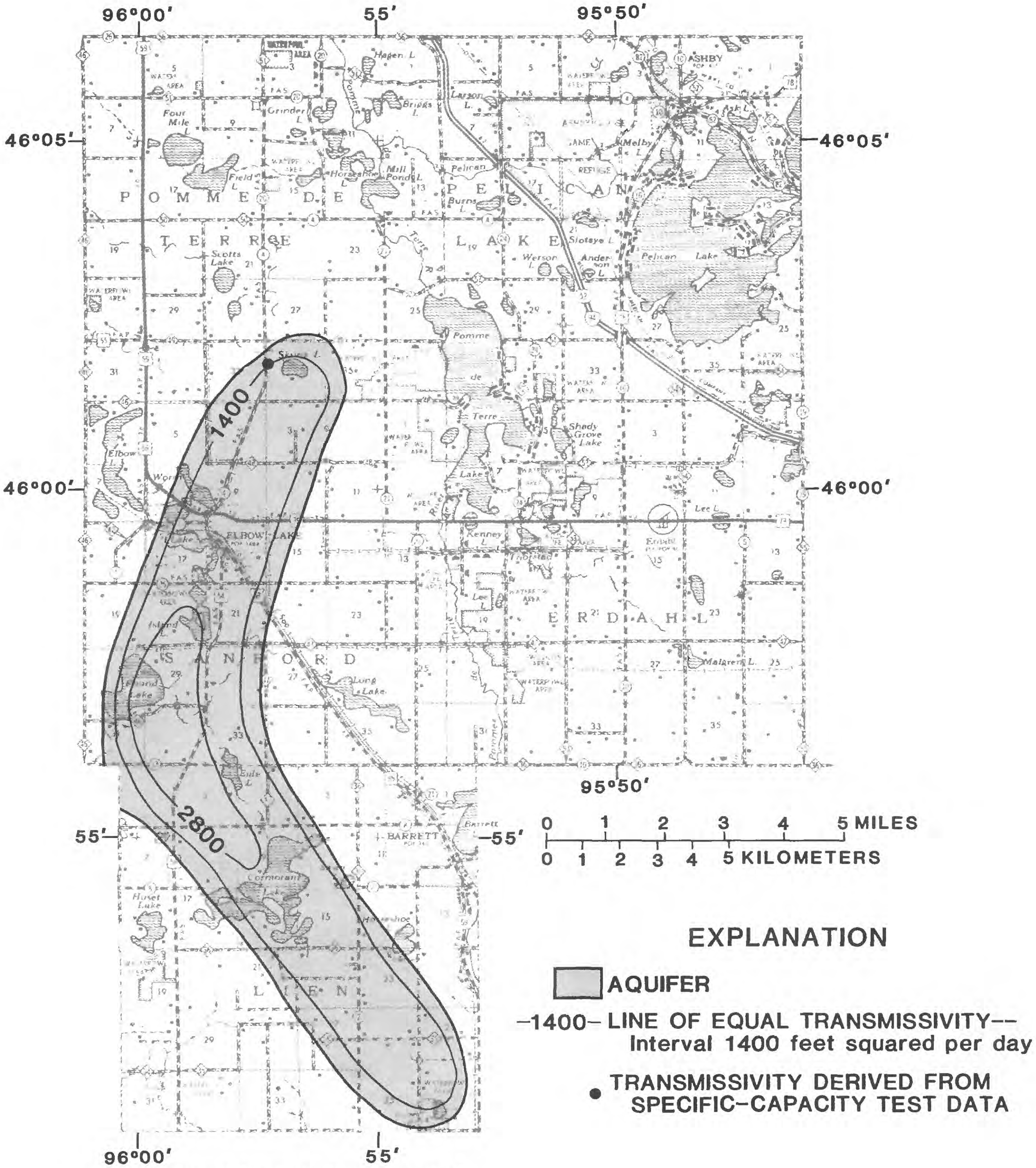
Water moves through the system of aquifers and confining beds described above according to a dynamic set of hydrologic processes. The movement and quality of water in the ground-water system is described in the following sections.

Ground-Water Movement

Ground-water moves under the force of gravity in the direction of decreasing head. The direction and rate of movement is indirectly related to recharge and discharge to the ground-water system and directly related to the hydraulic conductivity of drift material and to the hydraulic gradient. Aquifers are generally recharged near topographic highs and discharge near topographic lows. Ground water flows not only through aquifers, but also across confining beds. Because the hydraulic conductivity of aquifers is much greater than for confining beds, aquifers offer the least resistance to flow. Consequently, flow in aquifers is predominantly horizontal whereas flow in confining beds is predominantly vertical (fig. 12).

The general direction of horizontal ground-water movement in each confined aquifer is shown in plate 4 and figures 13 and 14. The average and range in depth to water below land surface for each confined aquifer is listed in table 1. Ground water generally flows southwest toward the Minnesota River and west toward regional discharge areas outside the study area. Ground water also discharges locally to lakes, wetlands, wells, and smaller streams. The potentiometric surfaces of the confined aquifers in a given location generally are similar to the water table and to each other. Potentiometric gradients generally are low, but steepen near recharge and discharge areas.

Where a confined aquifer coalesces with the surficial aquifer, ground water can flow directly between the aquifers in response to natural or pumping stresses. Ground water generally flows under a natural head gradient from confined aquifers into the surficial aquifer in these areas.



Base from Minnesota Department of Highways General Highway Map of Grant County, 1979

Figure 11.--Transmissivity of the Sanford aquifer

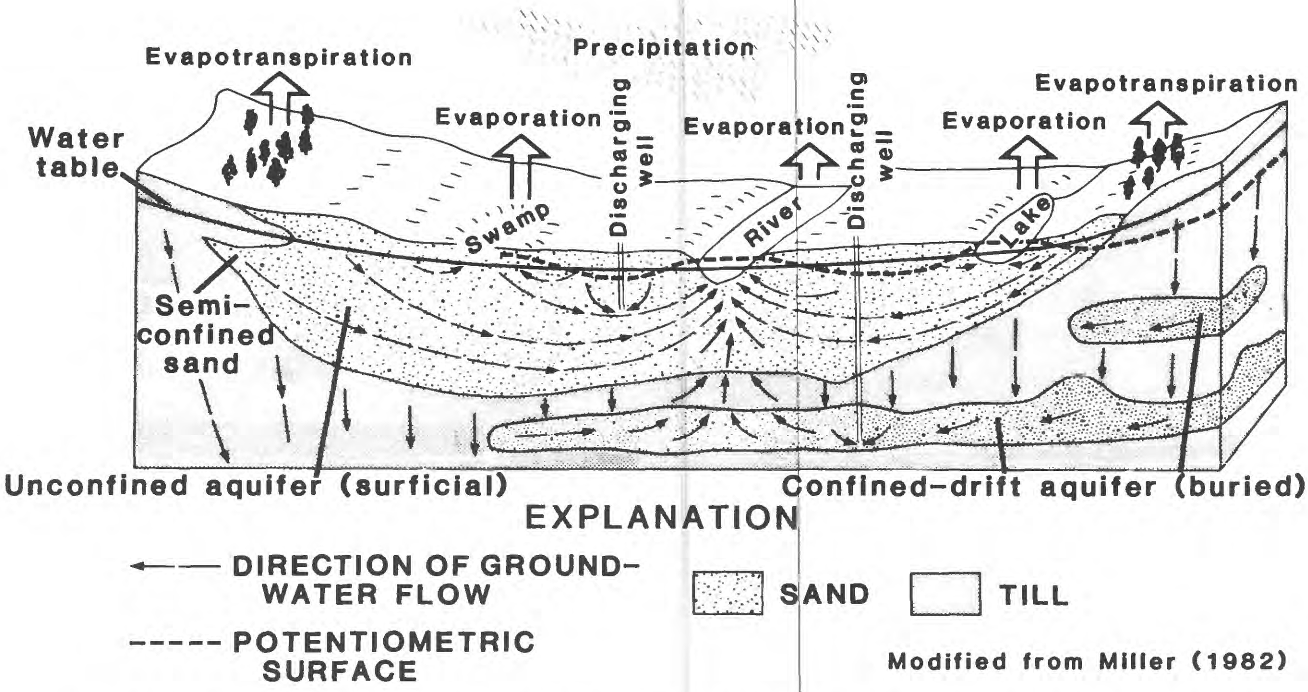


Figure 12.--Generalized ground-water-flow system showing source and discharge areas for ground water

Ground-water flow in the Appleton aquifer generally is southwest toward the Minnesota River (pl. 4a), to which it discharges. Ground-water also discharges locally to the Pomme de Terre River and, to a lesser extent, the Chippewa River in the south. The irregular potentiometric contours northwest and southeast of Appleton in fall 1982 probably reflect the residual effects of summer pumping from the aquifer. Inflections of the 980- and 990-ft contours north of Appleton may indicate ground-water discharge from the Appleton aquifer to the surficial aquifer, where the aquifers coalesce (pl. 4a). The lateral hydraulic gradient in the aquifer generally is about 8 ft/mi, but steepens near the Minnesota River. The potentiometric surface of the aquifer ranges from approximately 1,020 ft north of Hoffman to approximately 940 ft near the Minnesota River. Depth to water below land surface ranges from zero to 65 ft and averages 26 ft.

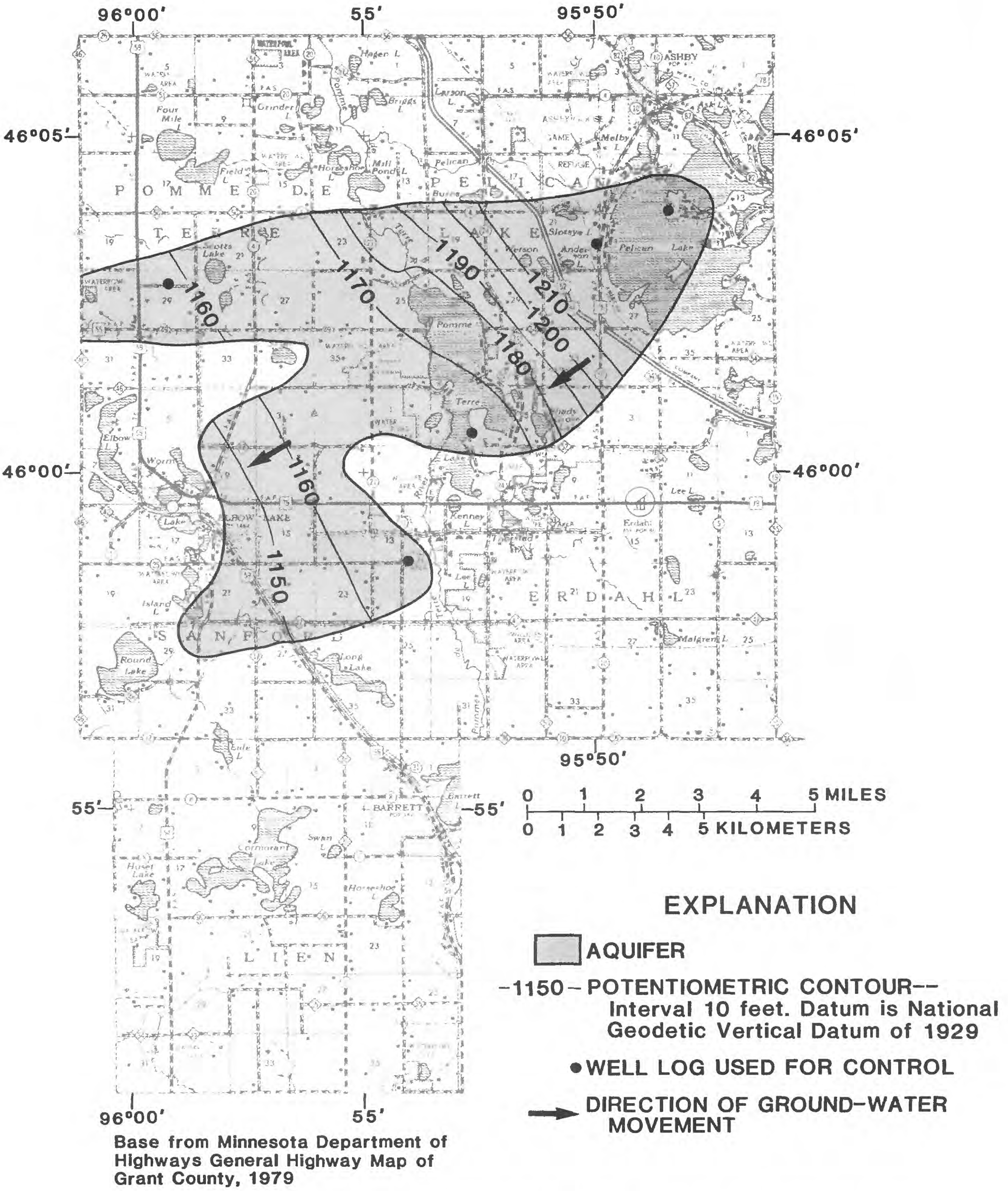


Figure 13.--Potentiometric surface of the Pomme de Terre aquifer

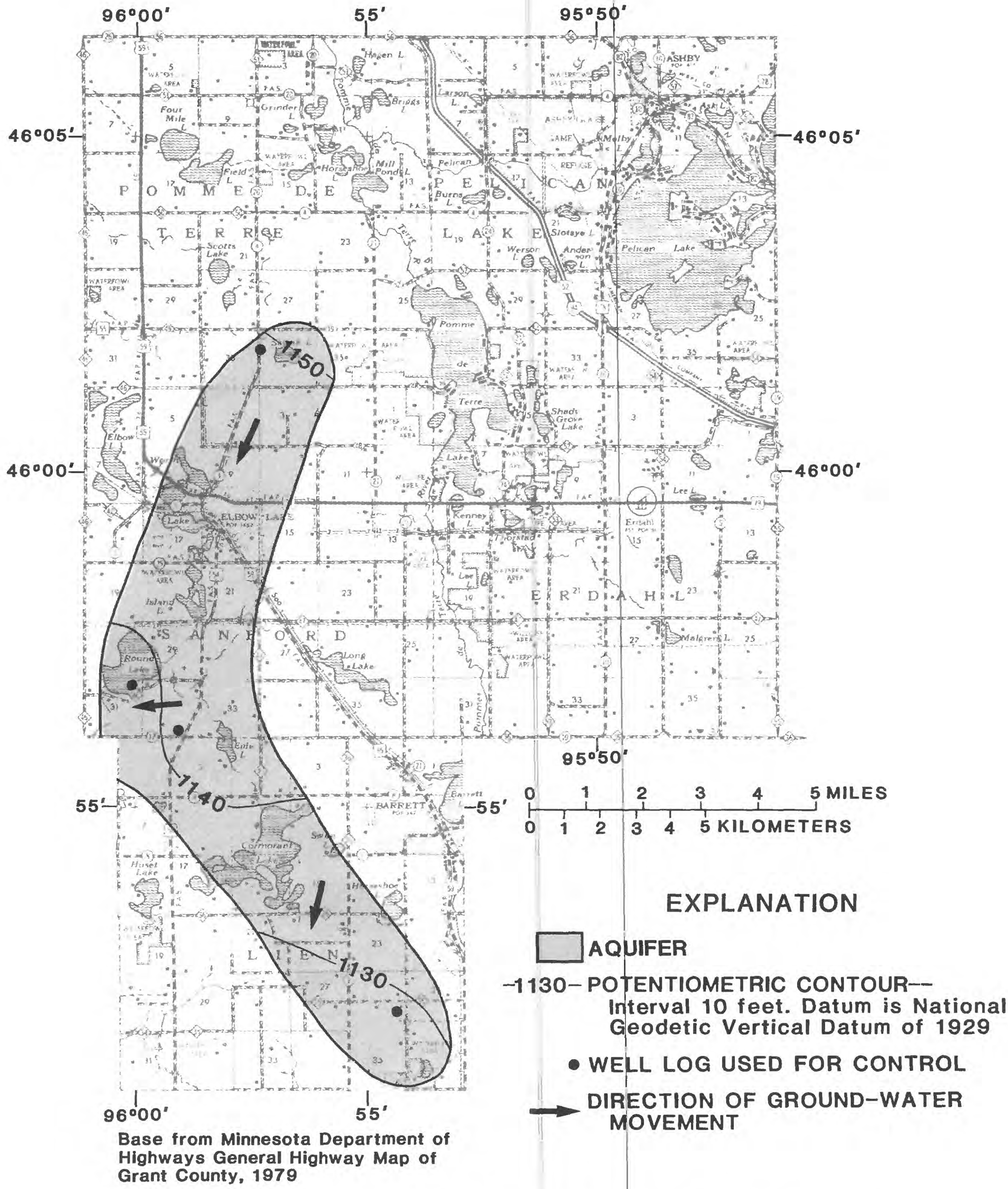


Figure 14.--Potentiometric surface of the Sanford aquifer

Ground-water flow in the Benson-middle aquifer generally is toward the south (pl. 4c), but ground-water also discharges locally to the Pomme de Terre and Chippewa Rivers. Extensive pumping from high-capacity municipal and industrial wells near Benson has created a cone of depression in the potentiometric surface and regional ground-water flow is diverted to this cone. The lateral hydraulic gradient generally is about 4 to 5 ft/mi. The head in the aquifer ranges from approximately 1,090 ft near Morris to approximately 990 ft near Big Bend City in Chippewa County. Depth to water below land surface ranges from zero to 80 ft and averages 23 ft.

Ground-water flow in the Morris aquifer generally is from northeast to southwest (pl. 4b) with some discharge locally to the Pomme de Terre and Chippewa Rivers. The lateral hydraulic gradient generally is about 5 ft/mi, but steepens to about 23 ft/mi northeast of Cyrus. The head in the aquifer ranges from approximately 1,200 ft northeast of Cyrus to approximately 1,020 ft north of Appleton. Depth to water below land surface ranges from 14 to 115 ft and averages 48 ft.

Ground-water flow in the remaining confined aquifers generally is from east to west. Lateral hydraulic gradients generally are 4 to 8 ft/mi; however, hydraulic gradients for the Barrett and Erdahl aquifers steepen near recharge areas to the east.

The head in each confined aquifer generally decrease with depth, indicating downward flow. Near the Pomme de Terre and Chippewa Rivers, however, the head increases with depth and flow is upward. Water-level data indicate that heads in the Benson-middle aquifer, for example, are 1 to 5 ft higher than in the surficial aquifer near these rivers. Hydrogeologic section A-A' (fig. 15) illustrates ground-water flow and head relationships near the Pomme de Terre River in Grant County.

Areal Recharge

The major source of recharge to the ground-water system is precipitation. Recharge is greatest in areas where the surficial aquifer is present (fig. 5). Precipitation, referred to as areal recharge, usually is greatest in spring due to snowmelt, spring rain, and little evapotranspiration, which results in rising ground-water levels. Conversely, ground-water levels generally decline in summer because most precipitation is lost as evaporation or as transpiration by plants. Areal recharge sometimes occurs in the fall, depending on rainfall, runoff, and evapotranspiration conditions.

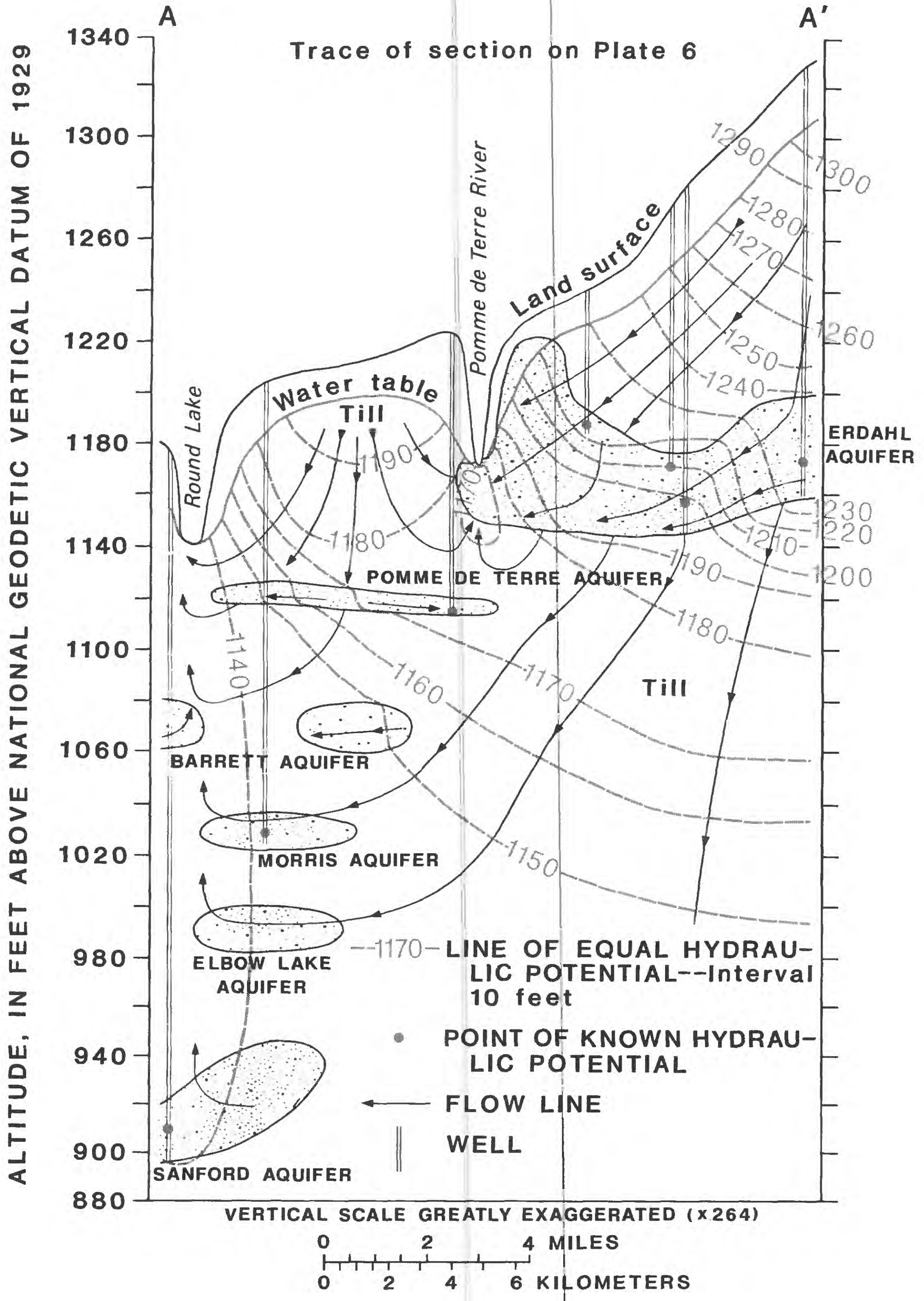


Figure 15.--Hydrogeologic section A-A' showing ground-water flow near the Pomme de Terre River

The rate at which water reaches the water table, where the surficial aquifer is present, can be estimated using a method of hydrograph analysis (Rasmussen and Andreason, 1959). The method assumes that (1) all water-level rises in the surficial aquifer result from areal recharge and (2) the rate of areal recharge per year nearly equals the sum of individual water-level rises within the year multiplied by the specific yield of the surficial aquifer. The water-level rise thus calculated, however, falls short of the true water-level rise by the amount of water-level decline that would have occurred if recharge had not taken place. To account for this part of areal recharge, the hydrograph, prior to the rise, is projected to the date on which the peak occurred (fig. 16). The corrected areal recharge rate, therefore, equals the difference between the peak stage and the projected water-level decline, on the day of the peak, multiplied by the specific yield of the surficial aquifer in the study area of 0.2 (fig. 16). Annual recharge was computed for 1980-82 using hydrographs from 12 observation wells near Appleton and Benson. Areal recharge ranged from 1.2 to 15.1 in/yr and averaged 6.0 in/yr.

Although areal recharge to the ground-water system is greatest where the surficial aquifer is present, areal recharge also occurs where till is present at land surface. Leakage to confined aquifers in these areas depends on (1) the head difference between the water table in the overlying till confining bed and the water level in the confined aquifer, (2) the vertical hydraulic conductivity of the till confining bed, and (3) the thickness of the till confining bed. Leakage rates to confined aquifers, in the areas where the surficial aquifer is absent, were estimated using the following form of Darcy's Law:

$$Q_C = \frac{K' \Delta h A_C}{m'}$$

where:

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

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

m' = confining bed thickness, in feet;

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

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

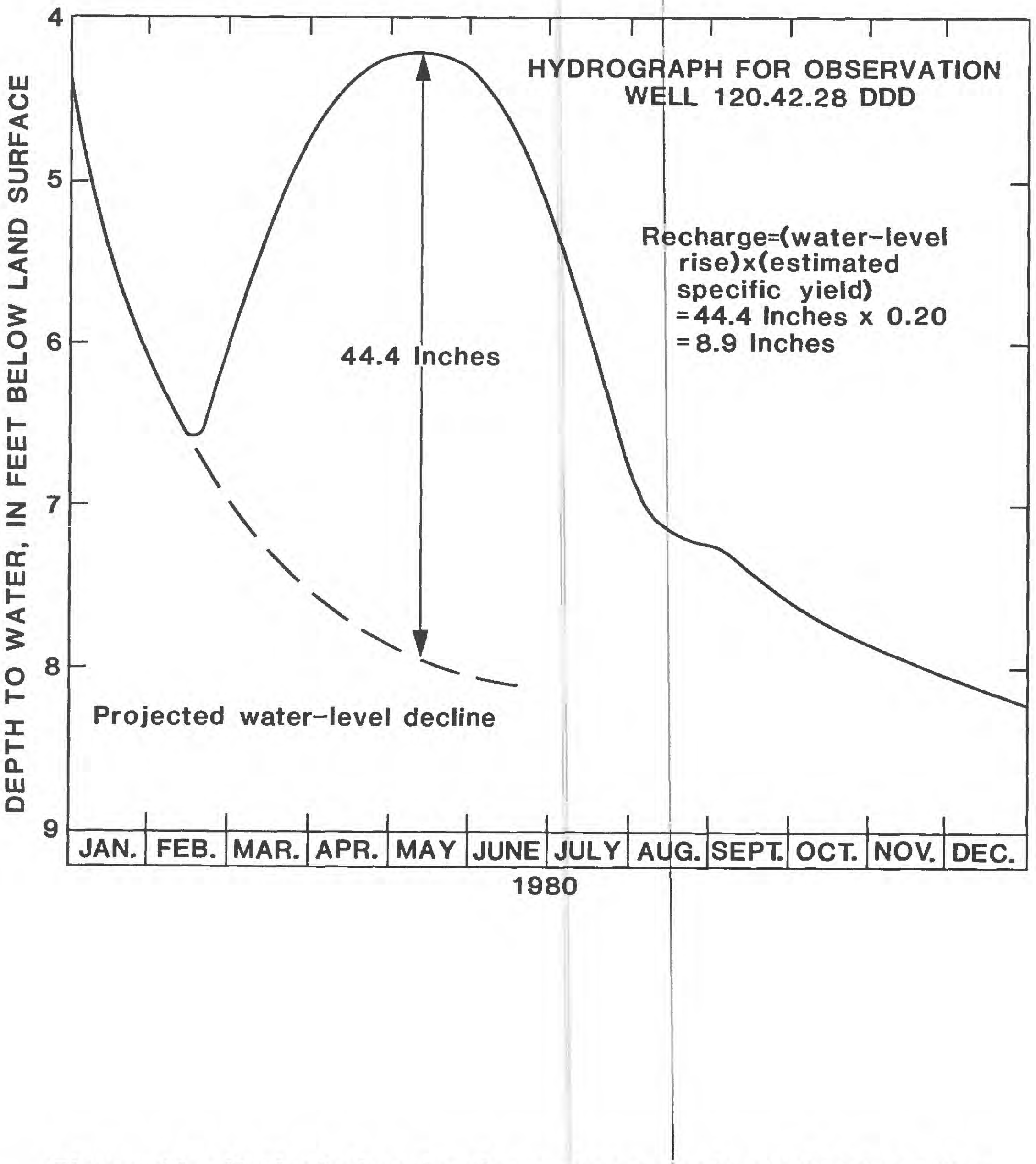


Figure 16.--Method of estimating recharge to the surficial aquifer

Leakage rates to confined aquifers of 0.4 to 3.4 in/yr were calculated for five sites using Darcy's Law.

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

Discharge

Discharge from the ground-water system occurs naturally and artificially. Ground water discharges naturally to streams, lakes, and swamps and by evapotranspiration (fig. 12). Artificial discharge is by wells.

Ground-Water Discharge to Streams

A significant part of discharge from the ground-water system is to streams. The amount of this discharge was estimated for the Pomme de Terre and Chippewa Rivers from base-flow measurements made May 23 and November 11, 1980. Streamflow measurements indicate a wide range in base-flow conditions. Total gain in streamflow to the Pomme de Terre River between Pomme de Terre Lake in Grant County and the town of Appleton in Swift County (pl. 1) was measured at approximately 45 and 23 ft^3/s during May and November 1980, respectively (U.S. Geological Survey, 1981, p. 303-304; U.S. Geological Survey, 1982, p. 336-337). Total gain in streamflow to the Chippewa River between Ellingson Lake in Grant County and the town of Hagen in Chippewa County (pl. 1) was approximately 186 and 98 ft^3/s during May and November 1980, respectively (U.S. Geological Survey, 1981, p. 304-306; U.S. Geological Survey, 1982, p. 337-339). Mean discharge of the Pomme de Terre and Chippewa Rivers in the study area, for the period of record, is 104 and 267 ft^3/s, respectively (U.S. Geological Survey, 1983). Discharge to or from the rivers depends on (1) thickness of the riverbed material, (2) vertical hydraulic conductivity of the riverbed material, and (3) head differences between the aquifer and river. In general, ground water discharge to rivers is greater than leakage from rivers into the ground-water system. Ground-water discharge to and from the Pomme de Terre and Chippewa Rivers is discussed in greater detail by Soukup and others (1984).

Evapotranspiration

Evapotranspiration is a combination of direct evaporation of surface water and soil moisture and transpiration of water by plants. The amount of ground-water loss to evapotranspiration depends on (1) water availability (depth to the water table below land surface), (2) the solar energy supplied, (3) air temperature, and (4) humidity of the air. The rate of evapotranspiration is assumed to be maximum (25 in/yr; Baker and others, 1979) where water levels are at land surface. The rate also is assumed to decrease to zero at the root-zone depth. The approximate root-zone depth for vegetation in the study area is assumed to be 5 ft.

Large quantities of water are discharged from the ground-water system through evapotranspiration during the summer. These losses decrease rapidly in the fall and are near zero in the winter. This variation in ground-water loss to evapotranspiration with time is approximately the same from year to year, provided the vegetation cover does not change significantly. Ground-water loss to evapotranspiration is greatest near the Chippewa River west of Benson. This area is characterized by wetlands or a depth to the water table of less than 5 ft. Ground-water-flow model simulations (Delin, 1986) indicate that ground-water loss to evapotranspiration during May through September exceeds total annual recharge in these areas.

Ground-Water Pumpage

Ground-water pumpage is a significant part of the total water budget locally in the study area. Surficial aquifers provided the majority of the ground water used prior to about 1975, but increasing amounts of water have been withdrawn from confined aquifers since.

The primary use of ground water from each confined aquifer in 1984, as shown in table 1, was for irrigation. Significant amounts of water are pumped also for municipal and industrial purposes. The Appleton, Erdahl, Benson-middle, Morris, and Benson-upper aquifers currently (1984) are used for irrigation, municipal, and(or) industrial purposes. The Benson-lower aquifer probably could yield sufficient water for these purposes also. All the confined aquifers are used for domestic and(or) stock purposes.

Most municipalities obtain their water supplies from surficial aquifers. Although the city of Appleton has wells completed in the surficial aquifer, the Appleton aquifer is directly connected to the surficial aquifer in the area. Therefore, some of the city's water supplies are diverted from the Appleton confined aquifer. The municipal supply for the city of Benson is obtained from the Benson-middle aquifer. Ashby and Erdahl Townships obtain ground-water supplies from the Erdahl aquifer. The city of Elbow Lake has developed a municipal supply from the Barrett aquifer.

Pumpage shown in table 2 represents annual ground-water pumpage reported to the Minnesota Department of Natural Resources (MDNR) during 1980-82 by permitted high-capacity ground-water users (irrigators, municipalities, and industrial users). These data represent most of the ground water used in the study area. Ground water used for domestic and stock purposes is insignificant compared to the large-scale uses. It is beyond the scope of this study to account for all water withdrawn from the ground-water system. Pumpage shown in table 2 is totaled by county and by aquifer, and includes a comparison of pumpage from confined and surficial aquifers. The five aquifers shown are the only aquifers for which high-capacity wells were reported to the MDNR. Pumpage included in the unidentified-aquifer category probably represents pumpage from one or more of the five confined aquifers listed.

Data in table 2 show that ground-water pumpage is greatest in Swift County and that the Appleton and Benson-middle aquifers were the most heavily used from 1980-82. Areas where these confined aquifers are present generally coincide with areas of sandy soils; thus, a greater amount of crop irrigation is necessary. A comparison of data for confined and surficial aquifers shows that total pumpage from surficial aquifers exceeds total pumpage from confined aquifers during 1980-82. With the exception of Swift County, pumpage from surficial aquifers generally exceed pumpage from confined aquifers in 1980-82 for all counties. Pumpage from confined aquifers generally decreased from 1980-82; however, pumpage from the Erdahl, Benson-middle, and Benson-upper aquifers increased slightly from 1981-82. By comparison, pumpage from surficial aquifers generally decreased from 1980-82 in each county. The decline in total pumpage from 1980-82 probably is due to climatic changes. Precipitation measured during 1981-82 was slightly greater than that measured during the previous 4-year period. Consequently, the need for crop irrigation was reduced during 1981-82.

Water-Level Fluctuations

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

Table 2.—Annual pumpage from confined and surficial aquifers near the Pomme de Terre and Chippewa Rivers, western Minnesota, 1980-82

(Pumpage in millions of gallons)

Confined aquifer	Year	Chippewa County	Grant County	Pope County	Stevens County	Swift County	Ground-water pumpage totals (by aquifer)
Appleton	1980	0.0	—	—	—	1,212.7	1,212.7
	1981	51.6	—	—	—	1,172.1	1,223.7
	1982	90.9	—	—	—	954.3	1,045.2
Erdahl	1980	—	40.6	—	—	—	40.6
	1981	—	43.6	—	—	—	43.6
	1982	—	63.1	—	—	—	63.1
Benson-middle	1980	—	—	246.9	134.4	1,203.8	1,585.1
	1981	—	—	184.1	65.2	1,117.1	1,366.4
	1982	—	—	187.1	112.1	1,218.4	1,517.6
Morris	1980	—	—	62.2	61.1	—	123.3
	1981	—	—	65.1	59.8	—	124.9
	1982	—	—	62.6	53.2	—	115.8
Benson-upper	1980	—	—	240.5	173.2	31.8	445.5
	1981	—	—	49.4	103.8	14.3	167.5
	1982	—	—	68.7	101.2	5.5	175.4
Unidentified	1980	—	77.9	28.2	142.2	—	248.3
	1981	—	43.5	32.5	106.8	—	182.8
	1982	—	44.8	27.6	135.6	—	208.0
Confined aquifer pumpage totals	1980	0.0	118.5	577.8	510.9	2,448.3	3,655.5
	1981	51.6	87.1	331.1	335.6	2,303.5	3,108.9
	1982	90.9	107.9	346.0	402.1	2,178.2	3,125.1
Surficial aquifer pumpage totals	1980	62.5	483.0	818.6	972.9	1,849.3	4,186.3
	1981	56.2	403.3	565.6	989.5	1,657.5	3,672.1
	1982	59.8	333.4	665.6	848.8	1,276.5	3,184.1

Water levels in wells completed in the surficial aquifer generally fluctuate 2 to 3 ft annually, even within approximately 1 mile of a high-capacity pumping well (fig. 17A). Water levels in wells completed in confined aquifers generally fluctuate 5 to 10 ft annually near high-capacity pumping wells (figs. 17B and 17C). Water-level fluctuations are greater for confined aquifers compared to surficial aquifers because of their lower ability to release water from storage in response to pumping. The deep troughs in the hydrographs are caused primarily by ground-water withdrawals from nearby high-capacity irrigation wells.

Water levels in confined and surficial aquifers in the study area generally recover to prepumping levels following each irrigation season. The net change in water level from 1980 to 1982 in 12 observation wells completed in confined and surficial aquifers in the study area ranged from about -2.0 ft to +1.1 ft. These data suggest that, although ground-water levels fluctuate in response to seasonal variations in recharge and discharge, the ground-water system is in dynamic equilibrium. In other words, the ground-water levels fluctuate around mean water levels that remain relatively constant in time. If the system were not in dynamic equilibrium, the general trend of ground-water levels would be rising or falling. A period of falling water levels throughout a region would indicate that recharge to the ground-water system was less than discharge from it.

Theoretical Maximum Yield of Wells in Confined Aquifers

The theoretical maximum yields of wells in confined aquifers were estimated using a chart developed by Meyer (1963) that relates well diameter, specific capacity, and the coefficients of transmissivity and storage. The relation shows that for confined aquifers (storage coefficients less than about 0.005), large differences in storage coefficient correspond to relatively small differences in specific capacity. Therefore, inaccurate estimation of aquifer storage is not a serious limiting factor in estimating theoretical well yields. The relation shows that for transmissivities between approximately 270 and 13,000 ft^2/d, the ratio of transmissivity to specific capacity is about 320 to 1. The ratio is larger for greater transmissivities. Therefore, for confined aquifers with transmissivities of 13,000 ft^2/d or less, the specific capacity can be approximated by dividing the transmissivity by 320. The theoretical maximum well yield at a specific site can then be estimated by multiplying the specific capacity by an arbitrarily selected drawdown, such as 30 ft. The estimates of theoretical maximum well yield included in this report were based on the following assumptions:

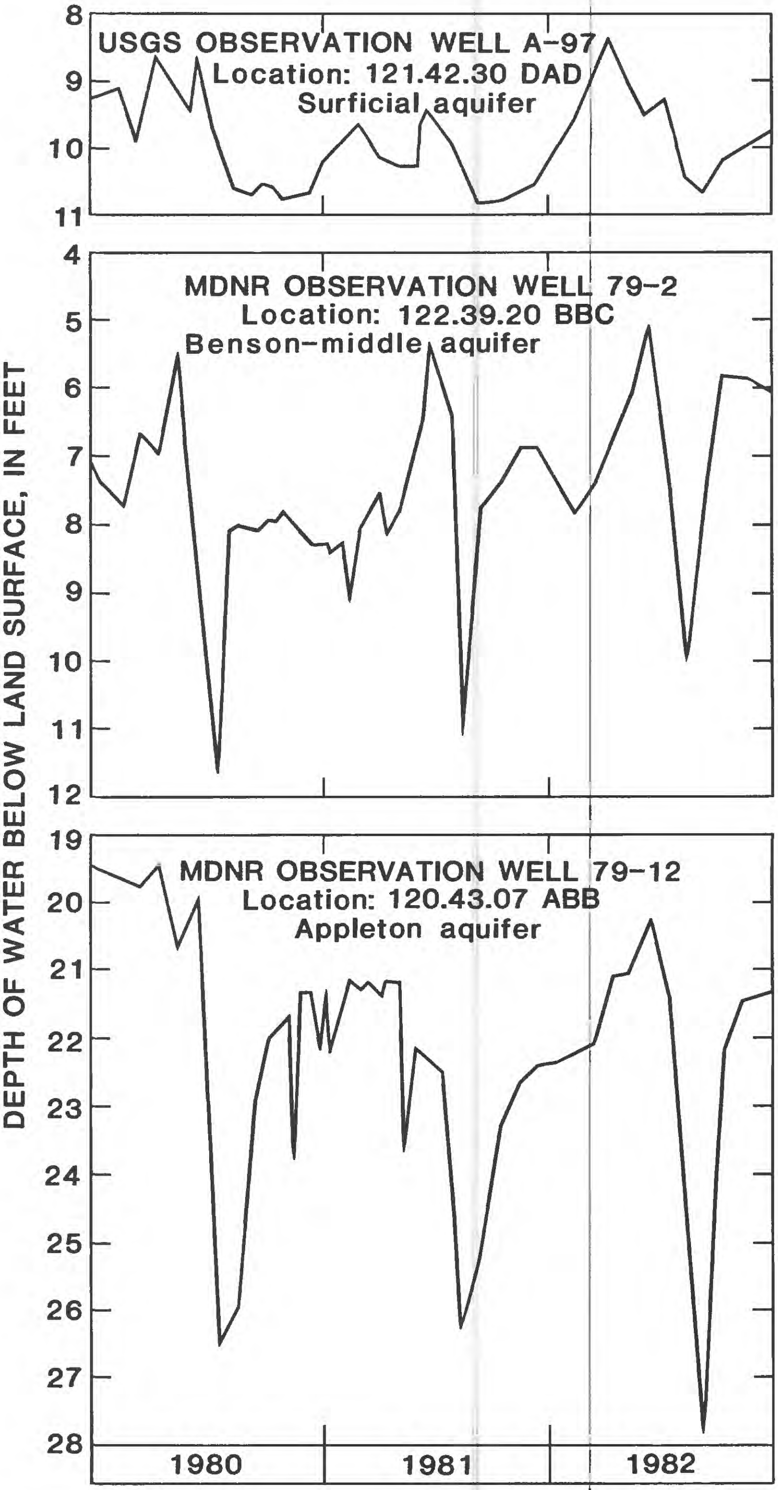


Figure 17.--Ground-water levels in confined and surficial aquifers, 1980-82

1. The aquifer is homogeneous, isotropic, and infinite in areal extent.
2. The well is screened through the entire thickness of the aquifer, is 100 percent efficient, and has a diameter of 2 inches.
3. The well is pumped continuously for 24 hours.
4. Drawdown is 30 ft.
5. Effects of recharge, hydrologic boundaries, and other pumping wells are negligible.

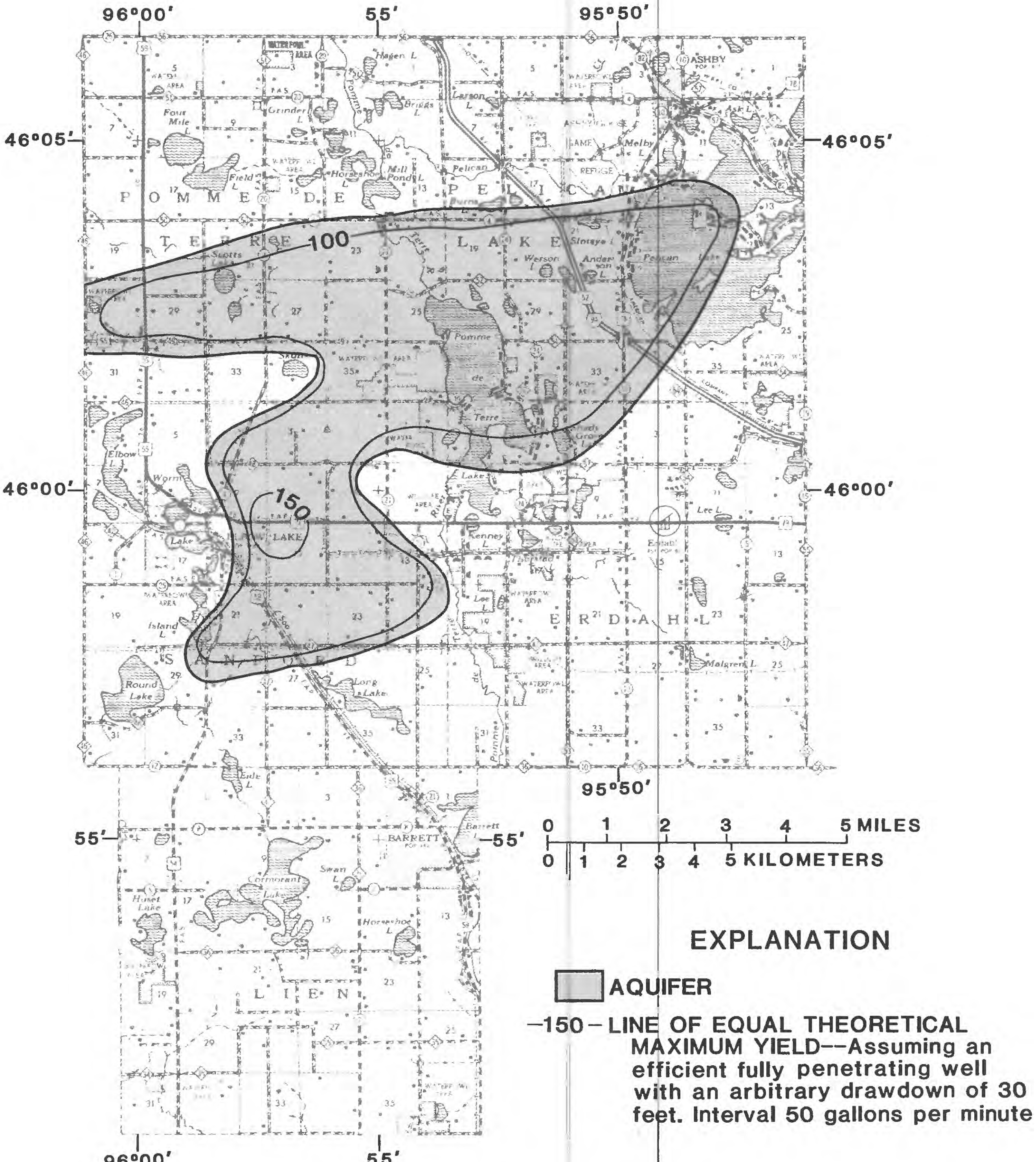
The reader should keep in mind that no aquifer or well fully satisfies the above assumptions. Local variations in aquifer hydraulic properties, recharge, proximity of the well to other pumping wells, effects of hydrologic boundaries (for example, rivers or the edge of the aquifer), well diameter and efficiency, and duration of pumping will cause local exceptions to the values shown on plate 5 and figures 18 and 19. The theoretical maximum well yields for each confined aquifer are intended to show only general conditions and relative differences in water-yielding capability. The maps cannot be used for accurate projection of well yields at a given location.

The areas of greatest theoretical maximum yield coincide with areas of greatest transmissivity. High-capacity wells generally are located in these areas. Theoretical maximum well yields range from less than 100 gal/min to about 1,800 gal/min locally in the Appleton confined aquifer.

Well Interference

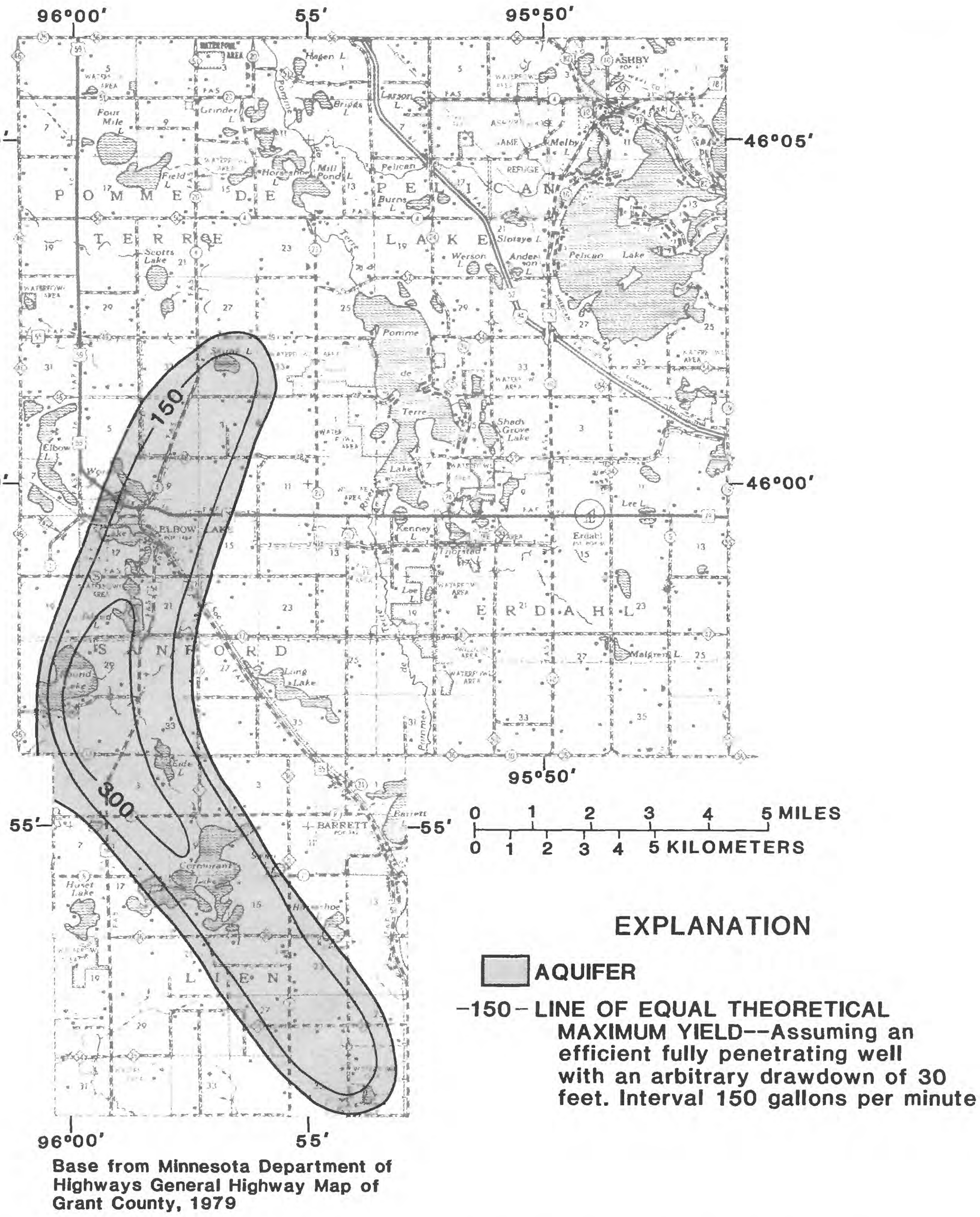
Pumping a well causes drawdown of nearby ground-water levels. This drawdown forms a depression in the water table or potentiometric surface that commonly is referred to as a cone of depression (fig. 20a). Where pumping wells are spaced relatively close together, pumping one well causes drawdown in the others. Total drawdown in a pumping well is equal to its own drawdown plus the drawdown, at that point, caused by other pumping wells (fig. 20b). This additional drawdown is referred to as well interference.

Well interference is of greatest concern to owners of wells completed in confined aquifers. Figure 21 illustrates the limit of the cone of depression for wells, pumping at the same rate, completed in surficial and confined aquifers. From this figure it is clear that the cone of depression around a well completed in a confined aquifer extends much further from the well than the cone of depression around a well completed in a surficial aquifer. Withdrawals from surficial aquifers (Fig. 21a) result in drainage of water from the sand and gravel through which the water table declines. The storage coefficient of a surficial aquifer virtually equals the specific yield of the aquifer material. Therefore, the cone of depression expands slowly



Base from Minnesota Department of Highways General Highway Map of Grant County, 1979

Figure 18.--Theoretical maximum yield of wells in the Pomme de Terre aquifer



Base from Minnesota Department of Highways General Highway Map of Grant County, 1979

Figure 19.--Theoretical maximum yield of wells in the Sanford aquifer

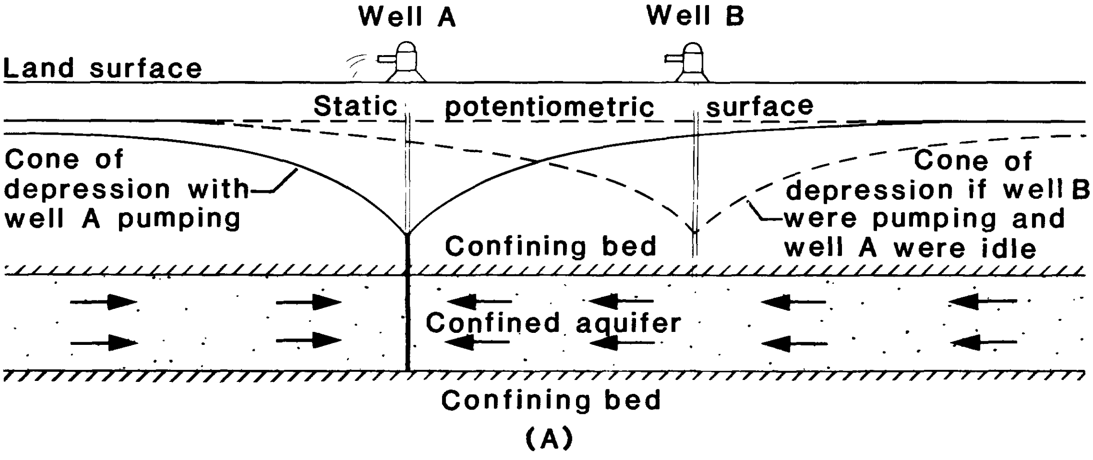
because water of sufficient quantity to sustain pumping is available in the immediate vicinity of the well. Conversely, withdrawals from confined aquifers (Fig. 21b) cause a drawdown in the potentiometric surface but normally do not cause dewatering of the aquifer. Confined aquifers have very small storage coefficients. Therefore, the cone of depression for a confined aquifer expands very rapidly since water is derived from expansion of water and compression of the rock skeleton of the aquifer.

Increased drawdown in a well, due to well interference, reduces the amount of available drawdown. A lowered pumping level also will result in increased pumping costs and decreased maximum well yield. The most serious problem related to well interference is the dewatering of a nearby domestic well by withdrawals from a high-capacity well. This problem can usually be avoided by screening domestic wells near the bottom of a confined aquifer as opposed to the top. The potential for well interference always should be considered prior to drilling a well.

Well interference also could occur near where confined and surficial aquifers coalesce. Where confined and surficial aquifers are separated by till (fig. 22a), water levels in the surficial aquifer are relatively unaffected by pumping from the confined aquifer. However, where confined and surficial aquifers coalesce, ground water can flow freely from one aquifer to the other. Therefore, if the aquifers are connected, pumping from a confined aquifer can cause drawdown in a nearby well completed in a surficial aquifer (fig. 22b). Because confined and surficial aquifers are connected in many places throughout the study area (pl. 1), interference between wells is a potential problem.

The location of a well near a physical boundary can affect drawdown also. Close proximity of a well to a sand-till boundary, for example, will increase drawdown in the well. Conversely, close proximity to lakes, streams, and swamps may induce infiltration of water to the aquifer, causing less drawdown.

Other problems relating to the use of confined aquifers is their relatively low rate of recharge, compared to surficial aquifers, and their areal discontinuity. Because of these factors, confined aquifers may initially yield sufficient quantities of water for irrigation, but may not be able to sustain these yields for an entire irrigation season.



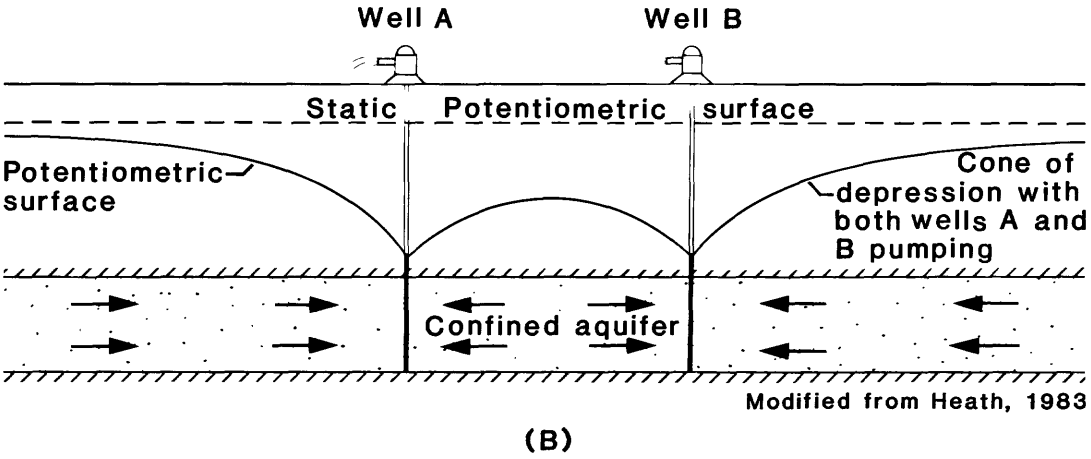


Figure 20.--Well interference in a confined aquifer

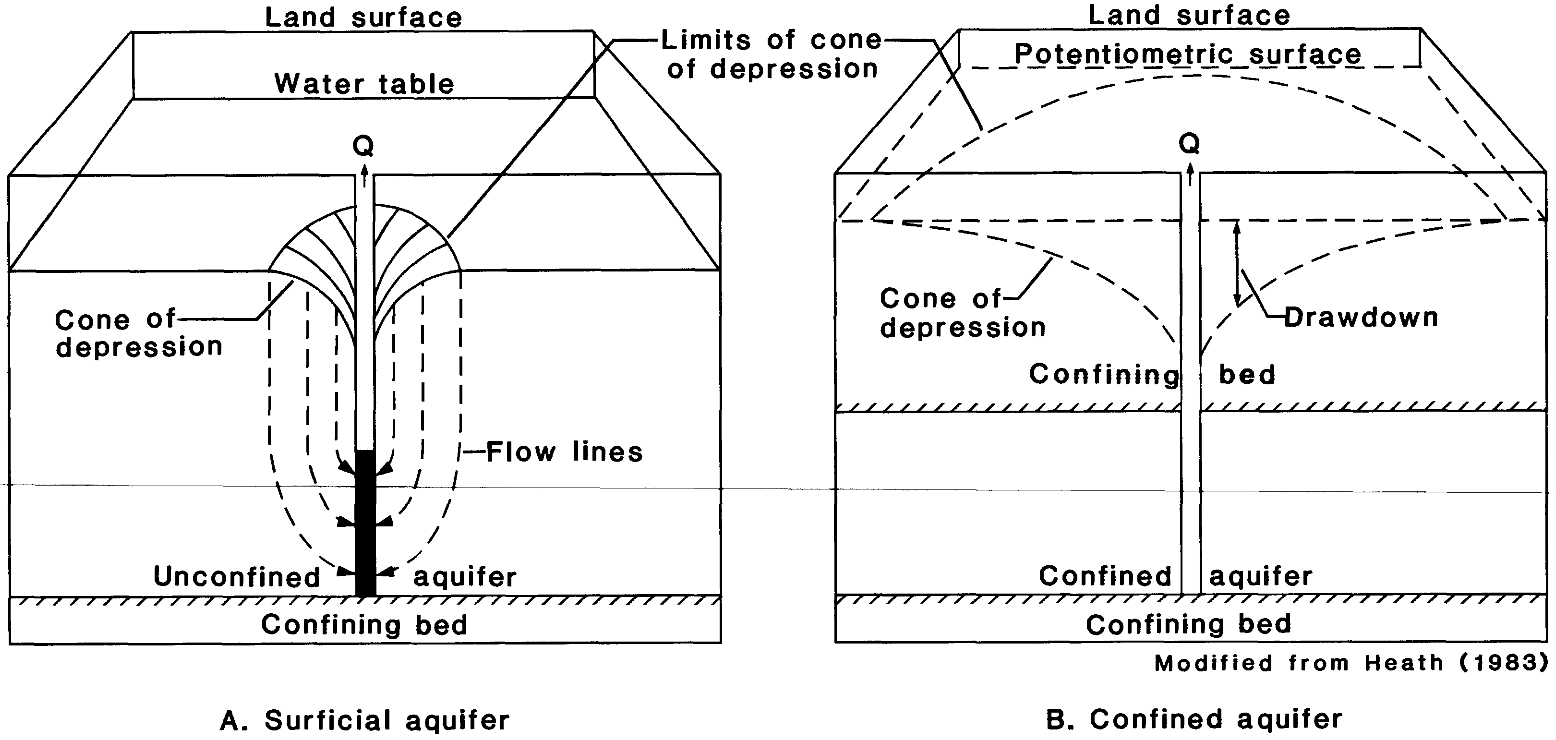


Figure 21.--Cones of depression for wells, pumping at the same rate, completed in surficial and confined aquifers

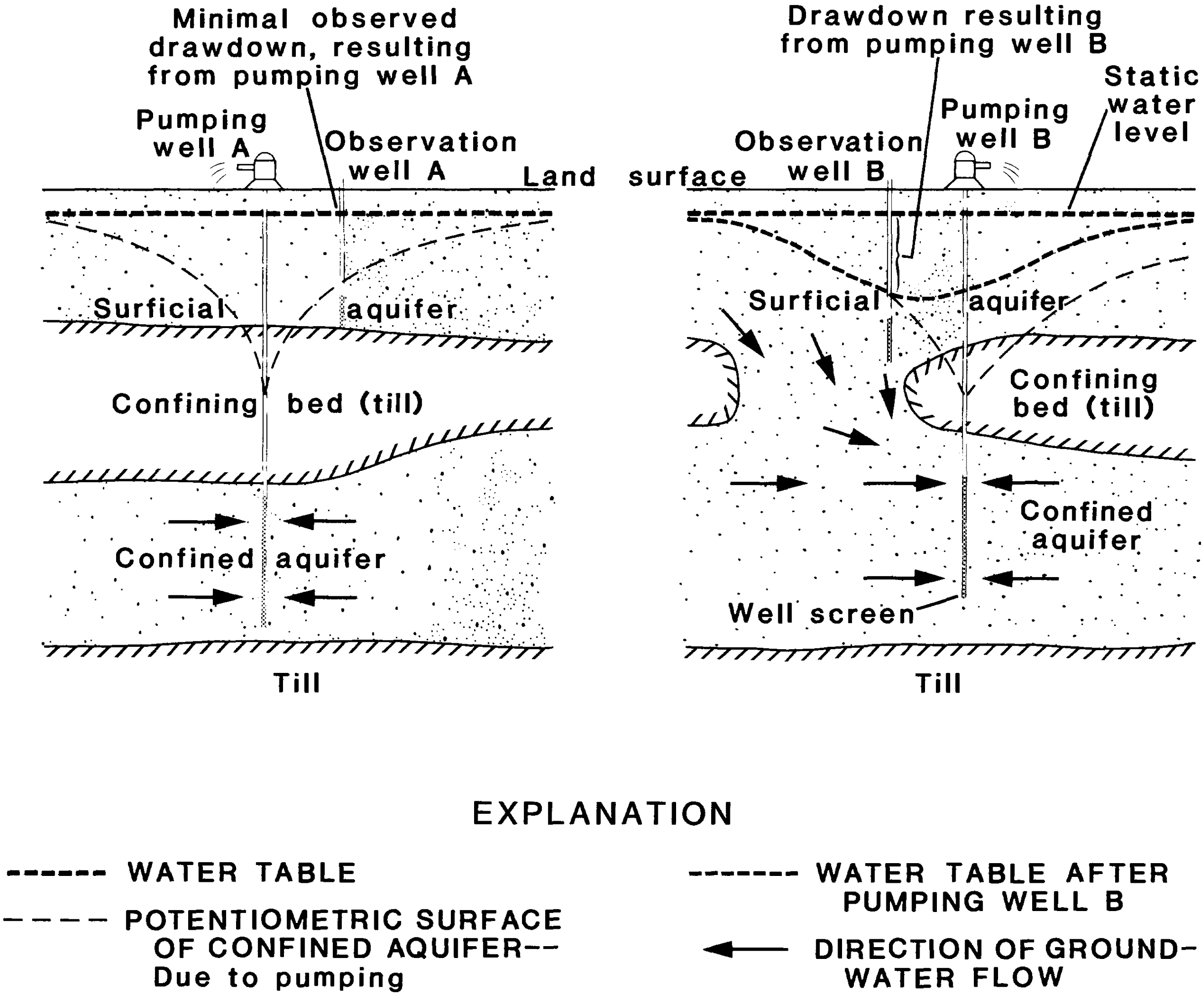


Figure 22.--Potential well interference near where confined and surficial aquifers coalesce

WATER QUALITY

Chemical constituents dissolved in ground water are derived mainly from the materials (soil, drift, etc.) through which the water moves. Ground-water quality varies in response to changes in residence time, length of flow path, temperature, precipitation, water chemistry, land use, and chemical reactions with minerals and aquifer materials.

Water-quality data for the confined aquifers sampled in this study are insufficient to determine chemical-constituent variations within each aquifer. In addition to the 11 wells sampled for this study during July through October 1982, 18 water-quality analyses of samples from confined aquifers, collected during 1964-65, were used in this report. The location and aquifer designation, if known, for the analyses are shown on plate 1. The median, standard deviation, and range in chemical-constituent concentrations for confined aquifers are given in table 3. Concentrations are in milligrams per liter (mg/L) and micrograms per liter (ug/L). Specific-conductance units are micromhos per centimeter at 25°C (umhos).

Calcium, bicarbonate, and sulfate are the predominant ions in ground water from confined aquifers (fig. 23). Calcium and bicarbonate are derived primarily from soil and rock weathering (Hem, 1970). Sulfate is contributed primarily by precipitation, organic material in sediments, and sulfide minerals in rocks.

Water from confined aquifers in the study area is hard to very hard, but generally is suitable for domestic consumption, crop irrigation, and other uses. However, concentrations of sulfate, iron, total dissolved solids locally exceed limits recommended by the Minnesota Pollution Control Agency (MPCA) (1978) for domestic consumption. Boron and specific conductance locally exceed the MPCA (1978) limit for agricultural and wildlife use. Table 4 lists the recommended limits for domestic consumption and table 5 lists the recommended limits for use by agriculture and wildlife. Also included in the tables are the percentage of wells sampled which exceeded the recommended limits.

The suitability of water for irrigation commonly is determined by relating conductivity of the water to the sodium-adsorption ratio (fig. 24), which can be used to classify the water in terms of its sodium and salinity hazards. This classification system was developed by the U.S. Salinity Laboratory (1954). The sodium-adsorption ratio is a measure of the amount of sodium with respect to calcium and magnesium. High values of the sodium-adsorption ratio can be an indication of tendency for ground water to destroy soil structure and thereby reduce permeability. High salinity concentrations endanger plants by reducing the amount of water absorbed by roots.

Table 3.—Comparison of water quality in confined and surficial aquifers near the Pomme de Terre and Chippewa Rivers, western Minnesota

Chemical constituent or property	Confined aquifers				Surficial aquifers			
	Number of analyses	Median	Range	Standard deviation	Number of analyses	Median	Range	Standard deviation
Specific conductance (lab) (umhos)	11	1,010	580-2,250	506	6	819	649-1,030	142
pH (standard units)	28	7.6	6.8-8.3	.4	19	7.5	7.2-8.2	.2
Temperature (Degrees C)	16	9.9	8.3-13	1.5	17	9.0	7.8-10	.8
Hardness, (mg/L as $CaCO_3$)	29	590	120-1,400	288	21	380	290-800	129
Hardness noncarbonate (mg/L as $CaCO_3$)	29	304	0-1,030	296	13	119	78-351	89
Calcium, dissolved (mg/L as Ca)	29	132	24-360	73	19	100	53-180	32
Magnesium, dissolved (mg/L as Mg)	29	57	14-137	29.5	18	36.5	25-64	11
Sodium, dissolved (mg/L as Na)	20	38.5	8.5-141	40.7	20	12.5	2.3-40	9
Potassium, dissolved (mg/L as K)	20	5.2	2.7-9.6	2.1	20	3.9	1.7-6.6	1.2
Alkalinity (lab) (mg/L as $CaCO_3$)	11	329	214-469	79	6	255	250-310	23
Sulfate, dissolved (mg/L as SO_4)	29	270	1-1,080	320	21	150	37-374	80
Chloride, dissolved (mg/L as Cl)	29	4.0	1.4-80	14.3	17	5.7	.5-46	14.8

Table 3.—Comparison of water quality in confined and surficial aquifers near the Pomme de Terre and Chippewa Rivers, western Minnesota—Continued

Chemical constituent	Confined aquifers				Surficial aquifers			
	Number of analyses	Median	Range	Standard deviation	Number of analyses	Median	Range	Standard deviation
Fluoride, dissolved (mg/L as F)	18	.2	.2-0.6	.1	21	.2	.1-0.3	.1
Silica, dissolved (mg/L as SiO_2)	18	27	12-33	5.5	18	26.5	23-29	1.5
Solids, residue at 180°C, dissolved (mg/L)	29	770	388-1,960	467	15	510	366-970	142
Solids, sum of constituents, dissolved (mg/L)	17	700	380-1,800	468	13	520	340-880	146
Nitrogen, NO_2+NO_3, dissolved (mg/L as N)	11	.1	0.1-1	.3	17	.5	.0-20	6
Phosphorus, ortho, dissolved (mg/L as P)	11	.02	.01-0.08	.03	17	.02	.00-0.05	.01
Boron, dissolved (ug/L as B)	18	210	.2-1,600	407	18	105	.2-240	59.8
Iron, dissolved (ug/L as Fe)	19	1,800	70-11,000	2,577	21	1,100	10-6,400	1,893
Manganese, dissolved (ug/L as Mn)	20	175	0-720	161	21	250	10-580	165
Carbon, organic, total (mg/L as C)	11	3.1	2.3-7	1.4	---	---	---	---

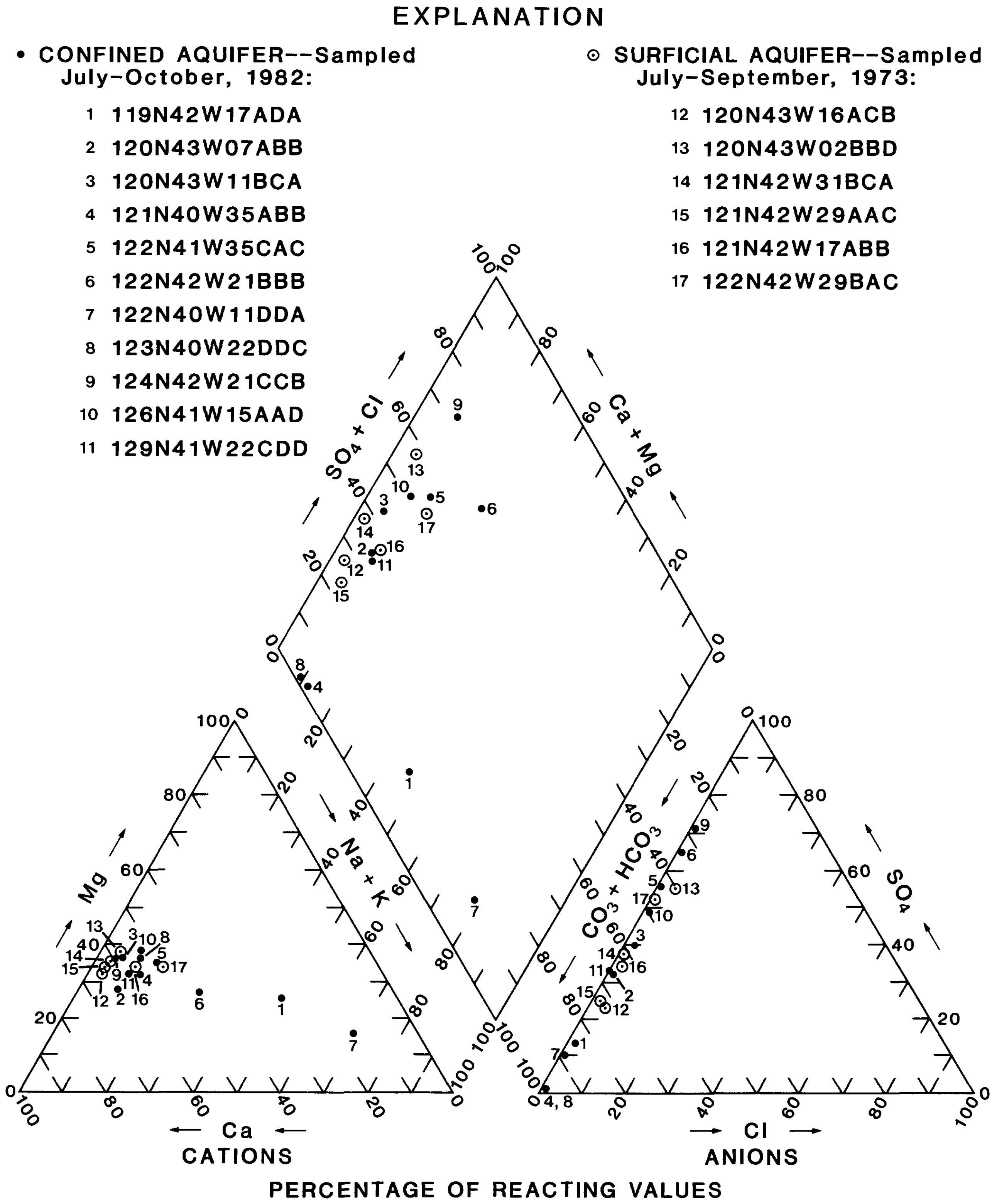


Figure 23.--Trilinear diagram showing chemical character of water in the confined and surficial aquifers

Salinity is directly related to the specific conductance of water. Water from the confined aquifers generally has a low sodium hazard and a medium to high salinity hazard (fig. 24).

Boron is essential to plant growth, but is toxic if present in concentrations much above recommended limits. Boron concentrations in water from confined aquifers were generally below the limit of 500 ug/L recommended by MPCA (1978) for agricultural and wildlife use (table 3). However, the boron concentration in water for one well was 1,600 ug/L. This concentration of boron may be toxic to semitolerant plants such as corn, sunflowers, wheat, barley, oats, and potatoes.

Dissolved iron and manganese are essential to plants and animals, but, in high concentrations, may cause objectionable taste, odors, and staining of plumbing fixtures. Concentrations of dissolved iron and managanese in water from confined aquifers generally exceed limits recommended by MPCA (1978) for domestic use. The concentrations (table 3) should not adversely affect plants, but treatment of the water may be necessary prior to domestic use.

High concentrations of sulfate in drinking water commonly result in an objectional taste and may have a laxative effect. Sulfate concentrations in water from confined aquifers locally are above the recommended limit for domestic use. The high sulfate concentrations (table 3) may be the result of mixing with ground water from deposits of Cretaceous age that are present locally. Sulfate concentrations in Cretaceous deposits generally are higher than in drift aquifers (Soukup, 1980).

High concentrations of dissolved solids in ground water can cause well-screen incrustation and reduced well yield. Dissolved-solids concentrations in water from confined aquifers (table 3) generally exceed the recommended limit for domestic use.

The quality of water from the confined aquifers generally is similar to the quality of water from the surficial aquifers. This relationship is illustrated in figure 23, which shows most of the results of chemical analyses plotted in the same general area of the trilinear diagram. This grouping of data indicates that major cations and anions in samples from confined and surficial aquifers are present in similar concentrations and that mixing of water between the aquifers probably occurs. Mixing of water from confined and surficial aquifers is highly probable where the aquifers coalesce (pl. 1).

The similarities in quality of water from the confined and surficial aquifers are shown also in figure 24. Water from both aquifers generally has a low sodium hazard and medium to high salinity hazard to soils. A comparison of chemical-constituent concentrations in water from the confined and surficial aquifers is shown also in table 3.

There are several differences in the quality of water from the confined and surficial aquifers. Water from surficial aquifers generally has higher concentrations of nitrate (NO_2+NO_3 as N) and chloride compared to water from the confined aquifers. The higher nitrate concentrations probably result from infiltration of runofff from livestock feedlots, domestic septic systems, and(or) fertilizers. Confined aquifers generally are less affected by these nitrogen sources, primarily because the overlying till confining beds prevent rapid leakage of nitrogen-rich water to the confined aquifers. In addition, studies conducted by Myette (1984) near Staples, Minnesota, indicate that concentrations of nitrate and chloride generally are greatest in samples from the shallowest part of the surficial aquifer, near the water table. This indicates that water containing elevated levels of nitrate and chloride moves vertically to the water table and then laterally, discharging primarily to streams and lakes, rather than moving deeper into the ground-water system. Only a minor amount of mixing occurs within the saturated part of the surficial aquifer.

Concentrations of dissolved solids, hardness, magnesium, sodium, potassium, sulfate, boron, and iron are, in general, slightly higher in water from confined aquifers than from surficial aquifers (table 3). These higher concentrations may be the result of mixing with water from deposits of Cretaceous age. Data from drift aquifers in Big Stone County, Minnesota, (Soukup, 1980) indicate that the pH, sodium-adsorption ratio, specific conductance, and concentrations of boron, iron, sodium, sulfate, chloride, fluoride, and dissolved solids generally increase with depth and are highest in water from the Cretaceous deposits.

Higher concentrations of chemical constituents in water from confined aquifers are directly related to longer residence times, in comparison to surficial aquifers. Longer residence times are primarily the result of (1) the discontinuity of confined aquifers, (2) lower ground-water-flow velocities due to that discontinuity, and (3) greater depth of burial, which results in ground-water interaction with intermediate and regional flow systems in addition to the local flow systems associated with surficial aquifers. The longer time for minerals to dissolve in the ground water results in higher concentrations of chemical constituents.

One of the most significant advantages of developing water supplies from confined aquifers, rather than surficial aquifers, is their lower susceptibility to ground-water contamination. Till confining units greatly impede the migration of contaminants from or near land surface to confined aquifers. Conversely, surficial aquifers are vulnerable to contamination from a variety

Table 4.--Minnesota Pollution Control Agency (1978) recommended limits for domestic consumption of selected chemical constituents

Chemical constituent	Recommended limit	Percent of wells exceeding limit
Chloride	205 mg/l	0
Fluoride	1.5 mg/l	0
Iron	300 ug/l	95
Nitrate (NO_2+NO_3)	10 mg/l	0
Sulfate	250 mg/l	59
Total dissolved solids	500 mg/l	83

Table 5.--Minnesota Pollution Control Agency (1978) recommendations for agriculture and wildlife use of selected chemical constituents.

Chemical constituent	Recommended limit	Percent of wells exceeding limit
Boron	500 mg/l	11
pH	6.0-8.5 units	0
Specific conductance	1,000 umhos	72

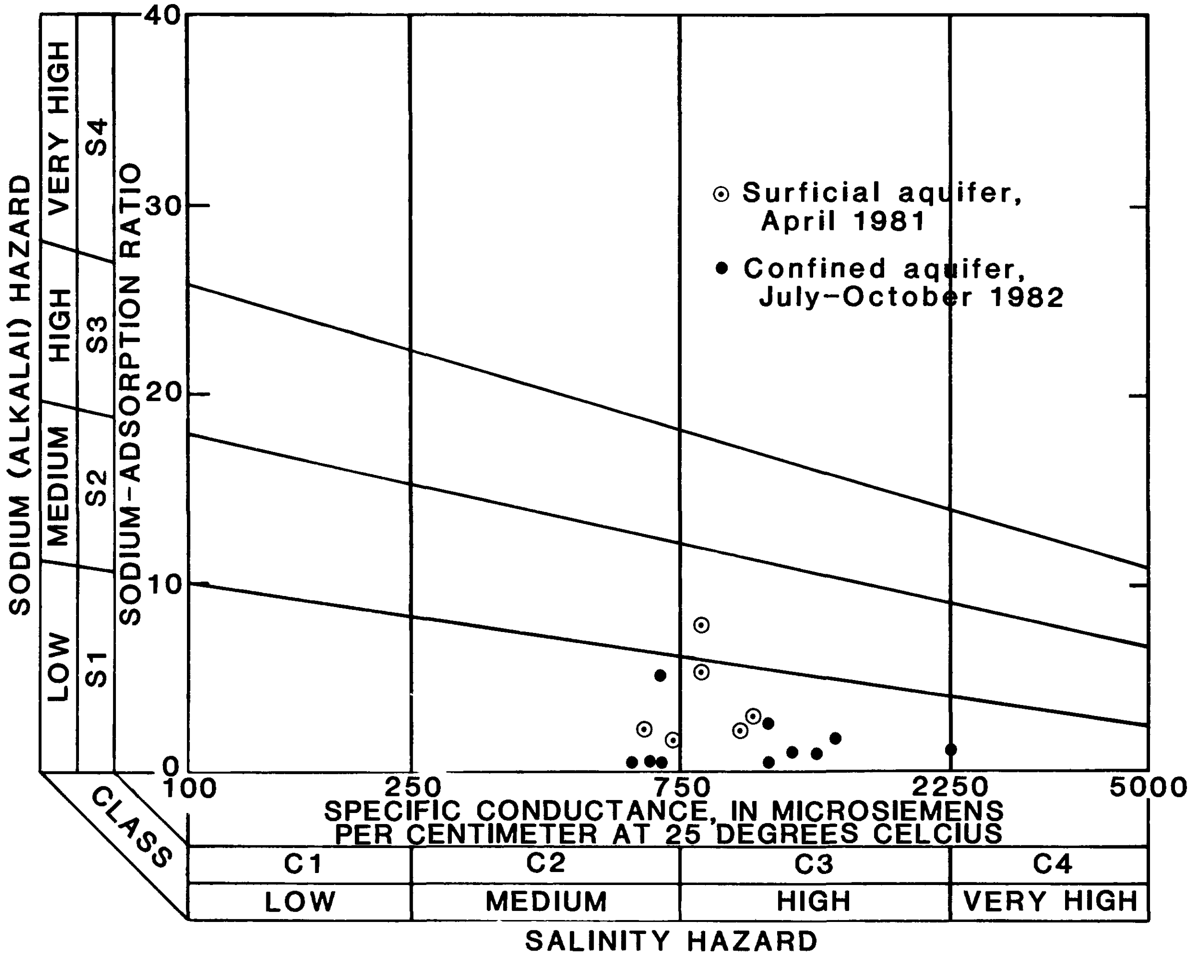


Figure 24.--Suitability of water from confined and surficial aquifers for irrigation in terms of sodium and salinity hazards

of sources, including excessive fertilizer applications and drainage from livestock feedlots and septic systems.

A dramatic future increase in pumpage from confined aquifers could have adverse effects on both ground-water quantity and quality. Overall ground-water quality of the confined aquifers, for example, could be decreased locally as a result of increased pumping, which could induce migration of poorer-quality water from underlying or overlying deposits.

GROUND-WATER-FLOW MODEL

Ground-water-flow models are useful tools for management of the ground-water system. A model was constructed for the southern part of the study area (fig. 1). The model area covers approximately 780 mi^2 and includes parts of Swift, Stevens, Pope, and Big Stone Counties. The modeled area is located in Swift County primarily because the ground-water resources have been developed more there than elsewhere in the study area. Hydrogeologic data also indicate that this area has the greatest potential for additional ground-water development. Model objectives were to (1) determine the vertical head gradient between the drift aquifers and (2) determine the probable effects of future ground-water development on water levels and storage of water in the aquifers.

The computer code of McDonald and Harbaugh (1984) was used to simulate ground-water flow in three-dimensions. Several simplifying assumptions were made in constructing the model. The assumptions are:

1. Ground-water flow in the drift aquifers is primarily horizontal and flow in the till-confining units separating them is primarily vertical, based on available water-level data;
2. The aquifers and confining units simulated are continuous, homogenous, and isotropic;
3. The ratio of vertical to horizontal conductivity in both the aquifers and confining units is 1 to 1;
4. The stage of the Minnesota River does not fluctuate significantly in time and, therefore, is simulated as a constant-head boundary;
5. Due to lack of accurate field data, streambeds are assumed to be 1 ft thick and composed of permeable material of lower hydraulic conductivity than the aquifers;
6. Minor streams and ditches are insignificant discharge points for the ground-water system and are ignored;
7. Areal recharge to the water table is from precipitation and occurs primarily from April to June and secondarily from October to December;

8. Where till is present at land surface, vertical leakage through till is constant and does not fluctuate seasonally;
9. The rate of evapotranspiration declines linearly to zero at a depth of 5 ft below land surface; and
10. Ground water used for irrigation is consumed by evapotranspiration and, therefore, return flow to the aquifer system is negligible.

The drift system was divided into three model layers: layer 1 (the top layer) represents the Morris aquifer and the surficial aquifer; layer 2 represents the Benson-middle aquifer; and layer 3 represents the Appleton aquifer. Horizontal ground-water flow was simulated in each aquifer. Vertical flow in the ground-water system was simulated by allowing leakage between model layers. A detailed description of the model, including steady-state and transient calibrations, is provided by Delin (1986).

The model was calibrated to assure that the hydrologic properties and boundaries selected were reasonable for the simulation of flow in the ground-water system. The model was calibrated for steady-state conditions by comparing measured water levels and calculated ground-water discharge to rivers with corresponding values computed by the model. Calibration of the model was achieved by successively adjusting hydrologic input values until model-computed water levels and ground-water discharge rates acceptably matched corresponding measured values. Transient calibration of the model was performed also to establish that the model can reasonably simulate changes in ground-water flow and water level in time. Transient calibration was accomplished by simulating water-level fluctuations during 1980-82.

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

Precipitation + ground-water flow into the modeled area = evapotranspiration + ground-water discharge to rivers + ground-water pumpage.

The steady-state water budget for the calibrated model (table 6) shows that recharge from precipitation accounts for the major inflow to the system. The table also shows that evapotranspiration and discharge to the principal streams account for most of the discharge from the system.

Table 6.--Steady-state water budget for the calibrated ground-water-flow model

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

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

Following calibration, the model was used to simulate the effects of pumping in 1982, potential effects of hypothetical increases in ground-water development, and potential effects of below-normal precipitation (drought). Results of these simulations can be used to estimate regional aquifer response to future stress. However, caution should be used in making ground-water management decisions based on the model simulations. Model-computed water-level declines reflect simplified assumptions and should be considered only in assessing regional water-level changes. The projected declines represent average declines over model grid blocks that are as large as 0.94 mi^2.

values, and declines in or near individual high-capacity wells generally will be greater.

The effects on the ground-water system of historical and 1982 pumping were evaluated using the model. Model results indicate that pumping has lowered water levels between 1 and 2 ft regionally in all aquifers and as much as 13 ft locally near Benson in the Benson-middle aquifer. Ground-water discharge to the Pomme de Terre and Chippewa Rivers has been reduced by approximately 18 percent compared to predevelopment conditions. Ground-water loss to evapotranspiration has decreased by approximately 20 percent because pumping has lowered the water table.

Simulations of hypothetical development indicate that the Appleton and Benson-middle aquifers and the surficial aquifer are capable of supporting additional pumping. Hypotetical wells were located in two areas with sandy soils, near the towns of Appleton and Benson, where there is little irrigation of crops but where irrigation could expand in the future. The hypothetical wells were spaced throughout these areas to minimize well-interference problems. The average pumping rate for irrigation wells in the modeled area, 27 Mgal/yr, was simulated for each hypothetical well. All model-computed water-level declines mentioned in the remainder of this section are in addition to the historical declines which ocurred prior to 1982. Model results indicate that the addition of 30 hypothetical high-capacity wells near Benson, pumping a total of 810 Mgal/yr, would lower water levels about 1 ft regionally in the Benson-middle aquifer and the surficial aquifer. Hypothetical pumping from only the Benson-middle aquifer resulted in a maximum water-level decline of 2.7 ft in the aquifer compared to a maximum decline of 1.3 ft if the hypothetical pumping was simulated only in the surficial aquifer. The addition of 28 hypothetical wells in the Appleton aquifer east and southeast of Appleton, pumping a total of 756 Mgal/yr, would lower water levels in the Appleton aquifer and the surficial aquifer 5 ft regionally.

The model was used to simulate the potential effects of a hypothetical drought of 30 percent less recharge for 3 years, accompanied by a 50 percent increase in pumpage. Model results indicate that increased pumping during the hypothetical drought probably would lower water levels 3 to 7 ft regionally in each aquifer and as much as 11 ft locally near aquifer-till boundaries. Ground-water discharge to the Pomme de Terre and Chippewa Rivers in the modeled area would be reduced by 15.2 and 7.4 ft^3/s, respectively, during the simulated drought compared to steady-state conditions. A detailed description of all model simulations is provided by Delin (1986).

SUMMARY AND CONCLUSIONS

Ground-water pumpage from confined aquifers in western Minnesota has increased during the last decade. These aquifers are the main sources of ground-water supplies where the surficial aquifers are absent. A study of confined aquifers in an area near the Pomme de Terre and Chippewa Rivers was conducted to determine the areal distribution, thickness, hydraulic properties, well-yield capabilities, and water quality of the aquifers.

Ten areally extensive confined aquifers were identified with thicknesses ranging from about 10 to 114 ft. Depth below land surface to the top of the aquifers ranges from about 20 to 250 ft. Aquifer transmissivities range from about 1,000 to 16,000 ft^2/d. Theoretical maximum well yields range from about 100 to 1,800 gal/min.

Ground water in confined aquifers generally flows from recharge areas in the north and west to discharge along the Minnesota River. Locally, ground water flows to smaller streams, lakes, wetlands, and pumping wells. Head in each confined aquifer generally is higher than in the underlying aquifer(s), indicating downward flow. However, the head increases with depth near rivers and flow is upward. Areal recharge from precipitation averages 6 in/yr where the surficial aquifer is present, but generally is less than 2 in/yr where till is present at land surface.

The Appleton and Benson-middle aquifers have been most intensely developed for water supplies. Pumpage from the confined aquifers generally decreased from 1980-82. Pumpage from surficial aquifers exceeded total pumpage from all confined aquifers for each of these years.

Water levels in observation wells completed in confined aquifers generally fluctuate 5 to 10 ft annually near high-capacity pumping wells, compared to annual fluctuations of 2 to 3 ft in surficial-aquifer wells. Water levels generally recover to prepumping levels following each irrigation season.

Well interference may occur when wells completed in confined aquifers are spaced relatively close together. Well interference may result in increased pumping costs and decreased well yield.

Water from confined aquifers is hard to very hard, but generally is suitable for domestic consumption and crop irrigation. However, locally elevated concentrations of some chemical constituents may require treatment of the water. Dissolved-solids concentrations range from about 400 to 1,800 mg/L. Concentrations of several chemical constituents are slightly higher in water from the confined aquifers than in water from surficial aquifers. These higher concentrations probably result from longer residence times for confined ground water and mixing with water from Cretaceous deposits.

Results from a ground-water-flow model indicate that historical and 1982 pumping has lowered water levels between 1 to 2 ft regionally and as much as 13 ft locally near Benson in the Benson-middle aquifer. The model also indicates that the Appleton and Benson-middle aquifers are capable of supporting additional pumping. Model results indicate that an extended drought may lower water levels between 3 and 7 ft regionally in each aquifer and as much as 11 ft locally near aquifer-till boundaries.

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GLOSSARY

The geologic and hydrologic terms pertinent to this report are defined as follows:

Aquifer--a formation, group of formations, or part of a formation that contains sufficient saturated permeable material to yield significant quantities of water to wells or springs.

Base flow--sustained streamflow, consists mainly of ground-water discharge.

Beach-ridge deposit--sand and gravel deposited by wave action on the shores of former large glacial lakes.

Cone of Depression--a depression in the potentiometric surface of an aquifer. Has the shape of a cone around a well from which water is being with drawn.

Confined aquifer--an aquifer bounded above and below by confining beds. An aquifer containing confined ground water. Synonymous with buried aquifer.

Confined ground water--ground water under pressure significantly greater than atmospheric and whose upper surface is the bottom of a confining bed.

Confining bed--a body of material with low vertical permeability stratigraphically adjacent to one or more aquifers. Replaces the terms "aquiclude," "aquitard," and "aquifuge."

Drawdown--the vertical distance between the static (nonpumping) water level and the water level caused by pumping.

Drift--a general term applied to all material (clay, sand, gravel, and boulders) transported and deposited by glacial ice or melt water issuing therefrom.

Equipotential line--line connecting points of equal static head. (Head is a measure of the potential.)

Esker--a long narrow ice-contact ridge composed of stratified drift. The drift was deposited in glacial streams flowing over glacial ice masses.

Evapotranspiration--water discharge to the atmosphere by evaporation from water surfaces and moist soil and by transpiration by plants.

Flow line--the idealized path followed by particles of water.

Ground water--that part of subsurface water that is in the saturated zone.

Head, static--the height above a standard datum of the surface of a column of water that can be supported by the static pressure at a given point.

Hydraulic conductivity--capacity of porous material to transmit water under pressure. It is the rate of flow of water passing through a unit section of area under a unit hydraulic gradient.

Hydraulic gradient--the rate of change of pressure head per unit distance of flow at a given point and in a given direction. Synonymous with potentiometric gradient.

Kame--mound-like hill of ice-contact stratified drift, of any size.

Outwash--washed, sorted, and stratified drift deposited beyond the melting glacial ice front by melt-water streams.

Permeability--a measure of the relative ease with which a porous medium can transmit a fluid under a potential gradient.

Potentiometric--surface that represents the static head of water in an aquifer; it is defined by the levels to which water will rise in tightly cased wells from a given point in an aquifer.

Saturated zone--zone in which all voids are ideally filled with water. The water table is the upper limit of this zone, and the water in it is under pressure equal to or greater than atmospheric.

Specific yield--the ratio of the volume of water that a saturated aquifer will yield by gravity drainage to the volume of the aquifer material.

Steady-state flow--flow at any point in a flow field when the magnitude of the flow velocity and the hydraulic head are constant with time.

Storage coefficient--the volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head. In an unconfined aquifer, it is virtually equal to the specific yield.

Surficial aquifer--the saturated zone between the water table and the first lower confining bed; synonymous with unconfined aquifer.

Till--unsorted, unstratified drift deposited directly by glacial ice.

Transmissivity--the rate at which water of the prevailing kinematic viscosity is transmitted through a unit width of an aquifer under a unit hydraulic gradient.

Unconfined aquifer--an aquifer that has a water table; the saturated zone between the water table and the first lower confining bed; synonymous with surficial aquifer.

Water table--that surface in a ground-water body at which the water pressure is atmospheric. Generally, this is the upper surface of the zone of saturation.

Appendix—Geologic logs of test holes

Test hole number: CP1 Location: 123.40.17AACCBA County: Pope

Township: Hoff Land surface altitude: 1,070

Geologic log	Depth (feet)	Thickness (feet)
Soil	0-5	5
Sand, fine to coarse, and fine to medium gravel	5-33	28
Till, sandy, gray	33-60	27
Clay, sandy	60-76	16
Sand, very fine	76-84	8
Clay, with some sand, gray	84-114	30
Sand and gravel, with clay	114-116	2
Sand, fine	124-125	1
Till, gray	125-134	9
Sand, medium, with clay layers	134-137	3
Till, sandy, clayey, gray	137-218	81
Sand, medium, and gravel, fine	218-224	6
Till, gray	224-236	2
Sand, medium, and medium to coarse gravel	236-238	2
Clay, sandy, whitish-gray, possibly Cretaceous age	238-247	11

Test hole number: CP2 Location: 123.40.22DDCCCC County: Pope

Township: Hoff Land surface altitude: 1,065

Geologic log	Depth (feet)	Thickness (feet)
Soil, dark brown, with sandy clay	0-2	2
Sand, coarse, brown	2-5	3
Sand, medium to coarse, and fine to coarse gravel	5-20	15
Till, clayey, gray	20-77	57
Till, clayey, soft, olive	77-87	10
Till, clayey, light gray	87-97	10
Till, clayey, dark gray	97-119	22
Sand, fine to medium, with fine to medium gravel	119-148	29
Cobble	148-149	1
Clay, sandy	149-153	4
Sand, very fine	153-163	10
Till, gray-brown	163-176½	13½
Sand	176½-178½	2
Till, gray-brown	178½-202	23½
Sand	202-203	1
Clay, greenish-white	203-234	31

Appendix—Geologic logs of test holes--Continued

Geologic log	Depth (feet)	Thickness (feet)
Test hole number: CP3 Location: 122.40.11DDAADD County: Swift		
Township: Clontarf Land surface altitude: 1,045		
Soil, dark	0-2	2
Sand, and gravel	2-6	4
Till, gray-brown	6-64	58
Sand, fine to medium, gray	64-65½	1½
Till, gray	65½-78½	13
Sand	78½-79½	1
Till, gray	79½-93½	14
Sand	93½-94	½
Till, with sand and cobbles	94-102½	8½
Sand, medium to coarse	102½-104	1½
Till, gray	104-159½	55½
Sand, with clay lens	149½-165	5½
Till, gray	165-187	22
Sand	187-189	2
Till, with clay layer	189-191	2
Sand	191-192	1
Shale, light blue-gray	192-207	15
Rock, weathered, crystalline, whitish	207-212	5

Geologic log	Depth (feet)	Thickness (feet)
Test hole number: CP4 Location: 123.40.30DADADD County: Pope		
Township: Hoff Land surface altitude: 1,063		
Soil, dark	0-2	2
Sand, medium to coarse, clean, gray	2-29	27
Till, clayey, gray	29-71	42
Sand, fine, gray	71-76	5
Sand, and interbedded clay	76-79	3
Till	79-106	27
Sand, and interbedded clay	106-107	1
Sand, fine to medium, gray	107-111	4
Till	111-117	6
Sand, and interbedded clay	117-118	1
Sand	118-120	2
Till	120-123	3
Sand	123-128	5
Clay	128-131	3
Sand, and gravel with clay layers	131-145	14
Till, clayey, gray	145-154	9
Sand, with some gravel	154-165	11
Till	165-169	4
Sand	169-170	1
Till, greenish-brown	170-182	12
Shale, bluish-green	182-190	8

Geologic log	Depth (feet)	Thickness (feet)
Test hole number: CP5 Location: 122.40.6CDDDDD County: Swift		
Township: Clontarf Land surface altitude: 1,052		
Soil	0-2	1
Sand, medium to coarse, brown	2-25	23
Till, clayey, gray	25-44	19
Sand, and interbedded gray till	44-78	34
Sand, with some clay layers	78-81	3
Till, gray	81-112	31
Sand	112-112½	½
Till, gray	112½-139	26½
Till, reddish-brown	139-159	20
Till, reddish-gray	159-176	17
Sand	176-177	1
Clay, gray	177-189½	12½
Till, gray	189½-210	20½
Sand, medium to coarse	210-218	8
Till, gray	218-243½	25½
Sand	243½-256	12½
Clay, sandy	256-263	7
Sand, fine, with clay lenses	263-273½	10½
Sand, fine	273½-340½	67
Cobbles	340½-343	2½
Rock, weathered, crystalline, light green	343-362	19

Geologic log	Depth (feet)	Thickness (feet)
Test hole number: CP6 Location: 122.41.25CDDD County: Swift		
Township: Tara Land surface altitude: 1,040		
Soil, dark brown	0-2	2
Sand, medium to coarse	2-27	25
Sand and gravel	27-31	4
Till, gray	31-36	5
Sand	36-38	2
Till, gray	38-72½	34½
Sand	72½-74	1½
Till, with clay lens	74-75	1
Sand	75-76	1
Till	76-131½	55½
Sand and gravel	131½-134	2½
Till	134-136	2
Sand, coarse	263-273½ 10½	
Sand, fine	273½-340½	67
Cobbles	134-136	2
Sand, coarse	263-273½ 185-187	2
Clay, light bluish gray	187-192	5
Clay, yellowish green to tan	192-207	17
Rock, highly weathered, crystalline, whitish-red	209-230	21

Geologic log	Depth (feet)	Thickness (feet)
Test hole number: CP7 Location: 122.41.6CDDDCC County: Swift		
Township: Tara Land surface altitude: 1,121		
Soil	0-1	1
Clay, brown	1-21	20
Till, gray	21-33	12
Sand, and fine to coarse gravel	33-39	6
Clay, sandy	39-40	1
Sand	40-41	1
Clay	41-41½	½
Sand and gravel	41½-44	2½
Till, gray	44-80	36
Sand, with clay layers	80-102	22
Till, gray	102-117	15
Sand, with clay layers	117-121	4
Till	121-127½	6½
Sand	127½-128	½
Till	128-147	19
Sand, with clay layers	147-157	10
Till, gray	157-175	18
Sand, coarse at top, fine at bottom, gray	175-281½	106½
Granite, decomposed, white	281½-297	15½

Geologic log	Depth (feet)	Thickness (feet)
Test hole number: CP8 Location: 121.42.17ABBBBB County: Swift		
Township: Moyer Land surface altitude: 1,033		
Soil, dark	0-½	½
Sand, medium to coarse, and fine to coarse gravel, brown	½-50	49½
Till, gray	50-107	57
Sand	107-110	3
Clay, sandy	110-113½	3½
Sand	113½-114	½
Till, light greenish-gray	114-133	19
Sand, with clay layers	133-169	36
Sand, medium to coarse, and gravel with cobbles	169-188	19
Clay, greenish-gray to turquoise	188-206	18
Shale, greenish-gray	206-238	32

Appendix--Geologic logs of test holes--Continued

Geologic log	Depth (feet)	Thickness (feet)
Test hole number: CP9 Location: 122.41.31CCC County: Swift		
Township: Tara Land surface altitude: 1,100		
Soil, dark	0-1	1
Till, weathered brown	1-16	15
Sand and gravel	16-16½	½
Till, clayey, brown	16½-26	9½
Till, clayey, gray	26-63	37
Sand	63-64	1
Till	64-70	6
Sand, very fine, gray	70-76	6
Sand, clayey	76-81	5
Clay, soft, gray	81-108	27
Granite, boulder	108-109½	1½
Till	109½-114	4½
Boulder	114-114½	½
Till, gray	114½-119	4½
Sand	119-120	1
Till, clayey, gray	120-137	17
Sand	137-139	2
Till, clayey, gray	139-143	4
Sand	143-144	1
Till, clayey, gray	144-163	19
Till, greenish-brown	163-165	2
Sand	165-167	2
Shale, greenish brown to green	167-192	25
Granite, decomposed, white	192-212	20

Appendix--Geologic logs of test holes--Continued

Geologic log	Depth (feet)	Thickness (feet)
Test hole number: CP10 Location: 122.42.21BBBBBB County: Swift		
Township: Fairfield Land surface altitude: 1,048		
Soil, dark	0-1	1
Sand, fine to coarse, and very fine to medium gravel	1-69	68
Till, clayey, gray	69-80½	11½
Sand	80½-81½	1
Till, gray	81½-93	11½
Till, clayey matrix, olive-brown	93-101	8
Boulder	101-101½	½
Clay, gray	101½-112	10½
Till, gray-brown	112-118	6
Till, light gray	118-124	6
Sand	124-125	1
Till, light gray	125-137	12
Sand, with clay layers	137-140	3
Clay	140-141	1
Sand	141-146	5
Clay, dark	146-157	11
Clay, dark, peaty	157-159	2
Granite, decomposed, with some shale at top, light blue	159-175	16

Geologic log	Depth (feet)	Thickness (feet)
Test hole number: CP11 Location: 120.43.2DDDDDD County: Swift		
Township: Appleton Land surface altitude: 1,021		
Soil, dark brown	0-1	1
Sand, fine to very coarse, and coarse gravel	1-38	37
Till, clayey, gray	38-40	2
Sand	40-41	1
Till, clayey, gray	41-62	21
Sand	62-63	1
Clay	63-64	1
Sand	64-67	3
Till, clayey, greenish-gray	67-77	10
Till, gray	77-88	11
Clay, gray	88-131	43
Clay, dark gray	131-142	11
Clay, some organic material, very dark gray-brown	142-150½	8½
Sand	150½-151	½
Clay, light blue-gray, becomes darker with depth	151-195	44
Dolomite boulder	195-196	1
Clay, gray	196-227	31
Clay, dark blue	227-228	1
Clay, gray	228-250	22
Sand	250-251½	1½
Shale, highly weathered, white	251½-263	11½
Shale, multicolored	263-278	15
Granite, decomposed, green	278-289	11

Geologic log	Depth	Thickness
Test hole number: CP12 Location: 121.43.21AAAAAA County: Swift		
Township: Shible Land surface altitude: 1,039		
Soil, dark brown	0-1	1
Sand, very fine to medium, brown	1-71½	70½
Boulder	71½-72	½
Till, clayey, gray	72-109	37
Sand	109-110	1
Till	110-113	3
Sand	113-114	1
Till	114-136	19
Sand	136-138½	2½
Till	138½-140	1½
Sand	140-143	3
Till	143-147½	4½
Cobbles	147½-148	½
Sand, with clay layers	148-150	2
Till, gray	150-153	3
Sand, fine	153-165	12
Cobbles, and sand with clay layers	165-167	2
Sand, with clay layers	167-178	10
Granite, decomposed, white	178-183	5
Granite, decomposed, with quartz	183-198	15
Granite, grayish-white	198-202	4

Test hole number: CP13 Location: 120.42.28DDDDDD County: Swift

Township: Edison Land surface altitude: 1,024

Geologic log	Depth (feet)	Thickness (feet)
Soil, dark	0-1	1
Clay, light gray-brown	1-10	9
Sand, medium to very coarse, with some fine to coarse gravel	10-11½	1½
Till, clayey, gray	11½-28	16½
Sand, and gravel, coarse	28-31½	3½
Till, reddish-brown	31½-43	11½
Till, gray	43-50½	7½
Sand	50½-51	½
Till, gray	51-55	4
Sand, medium to very coarse	55-56	1
Till, clayey, gray	56-83	27
Sand, gravel, and cobbles	83-85	2
Till, gray	85-92	7
Sand	92½-92½	½
Till, gray	92½-106	13½
Sand	106-107	1
Till, gray	107-118	11
Sand	118-119	1
Till, gray-brown	119-139	20
Till, sandy, light-gray	139-151	12
Sand, fine, multicolored	151-164	13
Till, light greenish-gray above to darker green below	164-173	9
Sand, fine to coarse, mostly medium	173-190	17
Till, gray	190-196	6
Sand, fine to medium	196-207	11
Sand, with clay layers	207-213	6
Till, reddish gray-brown	213-235	22
Clay, greenish-white	235-244	9
Sandstone, coarse, white	244-254	10
Granite, decomposed, whitish-green	254-303	49

Geologic log	Depth (feet)	Thickness (feet)
Test hole number: CP14 Location: 119.41.7BBBBBB County: Chippewa		
Township: Big Bend Land surface altitude: 1,042		
Soil, dark brown	0-1	1
Sand, fine, brown	1-8	7
Sand	8-16	8
Till, brown	16-22	6
Till, gray	22-100	78
Sand	100-101	1
Till, gray	101-113	12
Sand	113-115	2
Till	115-127	12
Sand, medium to coarse, with some clay layers	127-174	47
Sand, clean	174-213	39
Granite, decomposed, with some greenish-white clay	213-228	15

Test hole number: CP15 Location: 120.42.01DDCDCD County: Swift

Township: Edison Land surface altitude: 1,018

Geologic log	Depth (feet)	Thickness (feet)
Soil, dark brown	0-2	2
Sand, fine to coarse, and fine to medium gravel , gray	2-23	21
Till, clayey, gray	23-38	15
Sand	38-39	1
Till, gray	39-41	2
Sand	41-42	1
Till, clayey, gray	42-50	8
Sand	50-51	1
Till, clayey, reddish-brown	51-63	12
Sand, medium to coarse, multicolored	63-69	6
Till, light brown above, greenish-gray below	69-104	35
Sand	104-108	4
Till, gray	108-136	28
Till, dark gray	136-153	17
Clay, with sand layers	153-168	15
Clay, gray	168-184	16
Sand, fine, with some clay	184-198	14
Till, gray	198-220	22
Clay, soft, with lenses of fine sand	220-226	6
Sand	226-228	2
Till, dark gray	228-243	15
Till, gray	243-258	15
Clay, sandy	258-275	17
Shale, dark gray	275-292	17
Shale, white to gray and rust colored	292-313	21
Granite, decomposed, light green	313-333	20

Geologic log	Depth (feet)	Thickness (feet)
Test hole number: CP16 Location: 122.41.35CACCCD County: Swift		
Township: Tara Land surface altitude: 1,048		
Soil, dark	0-1	1
Gravel and sand	1-2	1
Sand, very fine to medium, with some fine gravel	2-22	20
Sand, fine to coarse, mostly medium, gray	22-47	25
Cobbles, and gravel	47-49	2
Till, clayey, gray	49-57	8
Sand	57-58	1
Till, gray	58-64	6
Sand	64-64½	½
Till, light greenish-white	64½-72	7½
Sand	72-72½	½
Clay	72½-73	½
Sand	73-77	4
Till, gray	77-101	24
Sand	101-101½	½
Till, gray	101½-115	13½
Boulder	115-116	1
Till, gray	116-132	16
Sand, fine to medium, gray	132-162	30
Till, gray	162-168	5
Boulder	167-168	1
Granite, decomposed, white	168-185	17

Geologic log	Depth (feet)	Thickness (feet)
Test hole number: CP17 Location: 122.41.35CCCC County: Swift		
Township: Tara Land surface altitude: 1,047		
Soil, dark brown	0-1	1
Gravel	1-2	1
Clay, brown	2-11	9
Sand, fine to medium, brown, with some fine to medium gravel	11-22	11
Till, clayey, gray	22-32	11
Till, sandy, gray	32-42	10
Sand	42-48	6
Boulder	48-48½	½
Till, gray	48½-64	15½
Sand	64-64½	½
Till, clayey, gray	64½-92	27½
Sand	92-93	1
Till, gray	93-106	13
Sand	106-107	1
Till, light gray, with very small sand lens	107-115	8
Sand	115-117	2
Till, light gray	117-140	23
Till, brown	140-144	4
Boulder	144-145	1
Till, brown above to gray-brown below	145-152	7
Granite, decomposed, bluish-green	152-182	30

Appendix—Geologic logs of test holes--Continued

Test hole number: CP18 Location: 121.40.35ABBBBB County: Swift

Township: Six Mile Grove Land surface altitude: 1,040

Geologic log	Depth (feet)	Thickness (feet)
Soil, dark brown	0-1	1
Clay, subsoil	1-3	2
Sand, fine, gray	3-12	9
Till, clayey, gray	12-21	9
Sand, medium to coarse, and fine gravel	21-22	1
Till, clayey, gray	22-59	37
Sand and gravel, multicolored	57-82	23
Till, gray	82-84	2
Sand	84-88	4
Till, gray	88-91	3
Sand	91-94	3
Sand, with clay lenses	94-110	16
Cobbles	110-113	3
Till, sandy, clayey, light gray	113-118	5
Sand, clayey, with clay lens	118-140	22
Sand, fine to medium	140-179	39
Till, gray, dark	179-241	62
Sand	241-243	2
Rock, decomposed, crystalline, white	243-259	16
Sand	259-267	8
Clay, white	267-277	10

Geologic log	Depth (feet)	Thickness (feet)
Test hole number: CP19 Location: 123.41.27CBBBBB County: Stevens		
Township: Moore Land surface altitude: 1,077		
Soil, dark brown	0-1	1
Sand, fine to coarse, and fine to medium gravel, brown above, gray below	1-40	39
Till, clayey, gray	40-50	10
Till, with interbedded sand	50-62	12
Sand and gravel, clayey, gray	62-65	3
Till, with interbedded sand layers	65-73	8
Sand and gravel	73-77	4
Till, gray	77-111	34
Till, with interbedded sand layers	111-116	5
Till, softer, slightly lighter gray, darkens with depth	116-133	17
Sand	133-134	1
Till, greenish-brown	134-137	3
Till, gray	137-161	24
Sand, fine to medium	161-187	26
Sand, clayey, with interbedded clay	187-195	8
Sand	195-206	11
Till, brick-red	206-222	16
Sand and gravel, coarse, multicolored	222-226	4
Till, greenish-gray	226-257	31
Clay, reddish-brown	257-263	6
Clay, gray	263-286	23
Shale, with fine sand layers, greenish-blue	286-313	27
Granite, decomposed, green	313-332	19

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