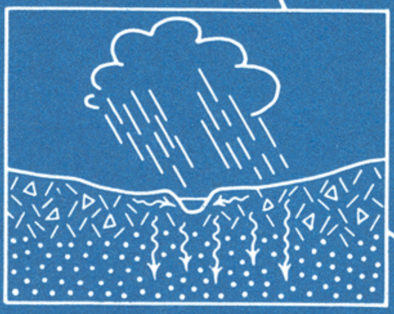
REPORT OF INVESTIGATION 48

STATE OF ILLINOIS

DEPARTMENT OF REGISTRATION AND EDUCATION

Ground Water Recharge and Runoff in Illinois

by WILLIAM C. WALTON



ILLINOIS STATE WATER SURVEY

URBANA

1965

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STATE OF ILLINOIS
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STATE WATER SURVEY DIVISION
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URBANA

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Ground-Water Recharge and Runoff in Illinois

by William C. Walton

ABSTRACT

Recharge conditions in several areas of northeastern Illinois are described, and recharge rates for several aquifers in central and southern Illinois are given. Recharge rates to deeply buried bedrock and sand and gravel aquifers vary from 1300 to 500,000 gallons per day per square mile (gpd/sq mi). The lowest rate is for an area where the Cambrian-Ordovician Aquifer is overlain by the Maquoketa Formation consisting mostly of shale; the highest rate is for an area where a sand and gravel aquifer is overlain by permeable coarse-grained deposits. Ground-water recharge generally is at a maximum during wet spring months; in many years there is little recharge during the 5-month period July through November.

The theoretical aspects of recharge from precipitation are discussed; recharge rates vary with the coefficient of vertical permeability, the vertical head loss associated with recharge, and the saturated thickness of deposits through which vertical leakage of water occurs. Recharge rates are not constant but vary in space and time.

A summary of coefficients of vertical permeability and leakage of deposits overlying aquifers within the state is presented. Coefficients of vertical permeability of glacial deposits range from 1.60 to 0.01 gallons per day per square foot (gpd/sq ft). The average coefficient of vertical permeability of the Maquoketa Formation is 0.00005 gpd/sq ft. Coefficients of leakage of glacial deposits and bedrock confining beds range from 2.3×10^{-1} to 2.5×10^{-7}.

Annual ground-water runoff from 109 drainage basins scattered throughout Illinois is estimated with streamflow hydrograph separation methods and flow-duration curves. The relations between ground-water runoffs during years of near, below, and above normal precipitation and basin characteristics such as geologic environment, topography, and land use were determined by statistical analysis. Ground-water runoff is greatest from glaciated and unglaciated basins having considerable surface sand and gravel and underlain by permeable bedrock. Ground-water runoff is least from glaciated basins with surface lakebed sediments and underlain by impermeable bedrock. Ground-water runoff during a year of near normal precipitation ranges from 0.06 to 0.43 cubic feet per second per square mile (cfs/sq mi). Ground-water runoff is at a maximum during spring and early summer months, and is least in late summer and fall months. Annual ground-water runoff depends upon antecedent moisture conditions as well as the amount and distribution of annual precipitation.

Because many aquifers in Illinois are deeply buried, not all ground-water runoff can be diverted into cones of depression because there is some lateral as well as vertical movement of water in surface deposits. Data on ground-water runoff can be useful in estimating recharge to aquifers and in evaluating the potential yield of ground-water reservoirs. However, studies indicate that no simple relation exists between ground-water runoff and the potential or practical sustained yields of aquifers.

INTRODUCTION

Recharge rates for aquifers must be estimated before ground-water resources can be evaluated and the consequences of the utilization of aquifers can be forecast. A thorough search of the literature reveals very little quantitative data on recharge rates, and even less quantitative data on the influence of geohydrologic factors on recharge rates. During the past several years the Illinois State Water Survey has made intensive studies pertaining to recharge to several aquifers scattered throughout the state. Studies made on 109 drainage basins scattered throughout Illinois indicate that no simple relation exists between ground-water runoff and recharge rates. In most parts of Illinois not all ground-water runoff can be diverted into cones of depression.

The major sources of recharge to aquifers in Illinois are direct precipitation on intake areas and downward percolation of stream runoff (induced infiltration). Recharge from precipitation on intake areas is irregularly distributed in time and place. Most recharge occurs during spring months when evapotranspiration is small and soil moisture is maintained at or above field capacity by frequent rains. During summer and early fall months evapotranspiration and soil-moisture requirements are so great that little precipitation percolates to the water table except during periods of excessive rainfall. Recharge during winter months when the ground is frozen is negligible. Only a small fraction of the annual precipitation percolates downward to the water table. A large proportion of precipitation runs overland to streams or is discharged by the process of evapotranspiration before it reaches aquifers. The amount of precipitation that reaches the zone of saturation depends upon several factors. Among these are the character and thickness of the soil and other deposits above and below the water table; the topography; vegetal cover; land use; soil-moisture content; the depth to the water table; the intensity, duration, and seasonal distribution of rainfall; the occurrence of precipitation as rain or snow; and the air temperature.

Recharge to aquifers by induced infiltration of surface water occurs when the water table is below the water surface of a stream and the streambed is permeable. The rate of induced infiltration depends upon several factors: the surface water temperature, the permeability of the streambed and the aquifer, the thickness of the streambed, the position of the water table, and the depth of water in the stream. Few streambeds remain stable over a long period because of alternate sedimentation and scouring by the stream. During periods of low streamflow fine sediment may settle from the slowly moving water and greatly reduce the permeability of the streambed. At high stages the fine sediments are scoured from the streambed and the permeability is increased.

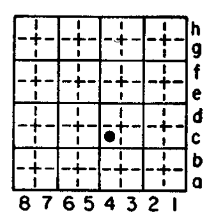
Recharge direct from precipitation and by induced infiltration of surface water involves the vertical movement of water under the influence of vertical head differentials. Thus, recharge is vertical leakage of water through deposits. The quantity of vertical leakage varies from place to place and it is controlled by the vertical permeability and thickness of the deposits through which leakage occurs, the head differential between sources of water and the aquifer, and the area through which leakage occurs.

In parts of northern Illinois, deeply buried sandstone and shallow dolomite aquifers are recharged in part by the vertical leakage of water through glacial drift and relatively impermeable shale or shaly dolomite beds. Large areas in western, south central, and southern Illinois are covered by glacial drift of Illinoian age; the drift cover is relatively thin and seldom exceeds 75 feet in thickness. In the area of the Wisconsinan glacial drift in the east central and northern part of Illinois, drift is thicker. Large deposits of water-yielding sand and gravel occur in drift areas chiefly in existing or buried bedrock valleys and as lenticular and discontinuous layers on bedrock uplands. The sand and gravel aquifers are commonly interbedded and overlain by deposits of till that contain a high percentage of silt and clay and have a low permeability. In many areas, recharge to these aquifers is derived from vertical leakage through the till.

The water level in shallow, 10 to 30 feet deep, dug or bored wells fluctuates through a wide range in response to above or below normal precipitation; in drought years many shallow dug wells go dry. However, water stored in thick deposits of glacial drift is available to deeply buried aquifers so that drought periods have little influence on water levels in these aquifers. Ground-water storage in deposits above aquifers and in aquifers permits pumping for short periods of time at rates greater than recharge. However, many aquifers are greatly limited in areal extent and thickness, and pumping at rates much above recharge rates for extended periods results in the rapid depletion of aquifers.

Well-Numbering System

The well-numbering system used in this report is based on the location of the well, and uses the township, range, and section for identification. The well number consists of five parts: county abbreviation, township, range, section, and coordinate within the section. Sections are divided into rows of 1/8-mile squares; each 1/8-mile square contains 10 acres and corresponds to a quarter of a quarter of a quarter of a section. A normal section of 1 square mile contains eight rows of 1/8-mile squares; an odd-sized section contains more or fewer rows. Rows are numbered from east to west and lettered from south to north as shown below:



St. Clair County

T2N, R10W

Section 23

The number of the well shown in the diagram is: STC 2N10W-23.4c. Where there is more than one well in a 10-acre square they are identified by arabic numbers after the lower case letter in the well number.

The abbreviations for counties discussed are:

Champaign	CHM	Kane	KNE
Cook	COK	Kendall	KEN
DeKalb	DEK	Lake	LKE
DuPage	DUP	Livingston	LIV

Madison	MAD	St. Clair	STC
Mason	MSN	Tazewell	TAZ
McHenry	MCH	Will	WIL
McLean	MCL	Woodford	WDF
Monroe	MON		

Acknowledgments

This study was made under the general supervision of William C. Ackermann, Chief of the State Water Survey, and Harman F. Smith, Head of the Engineering Section. Many former and present members of the State Water Survey and State Geological Survey assisted in the collection of data, wrote earlier reports which have been used as reference material, or aided the writer indirectly in preparing this report. Grateful acknowledgment is made, therefore, to the following engineers and geologists: T. A. Prickett, R. T. Sasman, W. H. Baker, W. H. Walker, G. E. Reitz, R. J. Schicht, R. R. Russell, H. R. Hoover, G. B. Maxey, R. E. Bergstrom, J. E. Hackett, A. J. Zeizel, and G. H. Emrich. J. W. Brother prepared the illustrations for this report.

GROUND-WATER RECHARGE

Cambrian-Ordovician Aquifer in Northeastern Illinois

The Cambrian-Ordovician Aquifer receives recharge from rocks of Silurian age or glacial drift by the vertical leakage of water through the Maquoketa Formation in areas of northeastern Illinois where the Maquoketa Formation (confining bed) overlies the aquifer. In areas where the Galena-Platteville Dolomite, the uppermost unit of the aquifer, directly underlies the glacial drift, recharge is by the vertical leakage of water largely through glacial drift. Silurian rocks and the glacial drift are recharged from precipitation that falls locally.

The Maquoketa Formation overlies the Cambrian-Ordovician Aquifer in large parts of northeastern Illinois, including the Chicago region, and to a great extent confines the water in the aquifer under artesian pressure. As described in a detailed report on the ground-water resources of the Chicago region (Suter et al., 1959), the Cambrian-Ordovician Aquifer is the most highly developed source of large ground-water supplies in northeastern Illinois and consists in downward order of the Galena-Platteville Dolomite, Glenwood-St. Peter Sandstone, and Prairie du Chien Series of Ordovician age; the Trempealeau Dolomite, Franconia Formation, and Ironton-Galesville Sandstone of Cambrian age. The sequence, structure, and general characteristics of these rocks are shown in figure 1A. The Cambrian-Ordovician Aquifer is underlain by shale beds of the Eau Claire Formation which have a very low permeability. Available data indicate that, on a regional basis, the entire sequence of strata from the top of the Galena-Platteville to the top of the shale beds of the Eau Claire Formation essentially behaves hydraulically as one aquifer.

As shown in figure 1B, the Maquoketa Formation has a maximum thickness of about 250 feet and thins to the north and west to less than 50 feet. The formation dips regionally to the east at a uniform rate of about 10 feet per mile. Bergstrom and Emrich (see Suter et al., 1959) divided the Maquoketa Formation into three units —lower, middle, and upper. As described by Bergstrom and Emrich: "The lower unit is normally a brittle, dark brown, occasionally gray or grayish brown, dolomitic shale grading locally to dark brown, argillaceous dolomite. The middle unit is dominantly brown to gray, fine- to coarse-grained, fossiliferous, argillaceous, speckled dolomite and limestone. It is commonly interbedded with a fossiliferous brownish gray to gray, dolomitic shale. The upper unit is a greenish gray, weak, silty, dolomitic shale that grades into very argillaceous, greenish gray to gray dolomite. The lower unit is thicker in Cook and Will Counties where it exceeds 100 feet. It thins to the north and west to less than 50 feet. The middle unit is thicker to the west where it is more than 100 feet locally and thins to the east. The upper unit ranges in thickness from less than 50 feet in the west to more than 100 feet in parts of Cook and Will Counties. The lower dense shale unit is the most impermeable unit. Dolomite beds in the middle unit yield small quantities of ground water."

The piezometric surface of the Cambrian-Ordovician Aquifer in 1864 (figure 2A) indicates that under natural conditions water entering or recharging the aquifer was discharged in areas to the east and south by vertical leakage upward through the Maquoketa Formation and by leakage into the Illinois River Valley. Flow lines were drawn from the ground-water divide in McHenry County toward the northern and southern boundaries of Cook County at right angles to the estimated piezometric surface contours for 1864. The part of the aquifer (area 1) which is enclosed by the flow lines, the ground-water divide, and section B—B' was considered. In 1864 the piezometric surface was below the water table and downward leakage through the Maquoketa Formation into the aquifer was occurring in area 1. Because near steady-state conditions prevailed and there was no appreciable change in storage within the aquifer, leakage was equal to the quantity of water percolating through section B—B'. At section B—B' the hydraulic gradient of the piezometric surface was about 2 feet per mile (ft/mi), and the distance between limiting flow lines was about 25 miles. Based on data given by Suter et al. (1959), the average coefficient of transmissibility of the aquifer at section B—B' is about 19,000 gpd/ft.

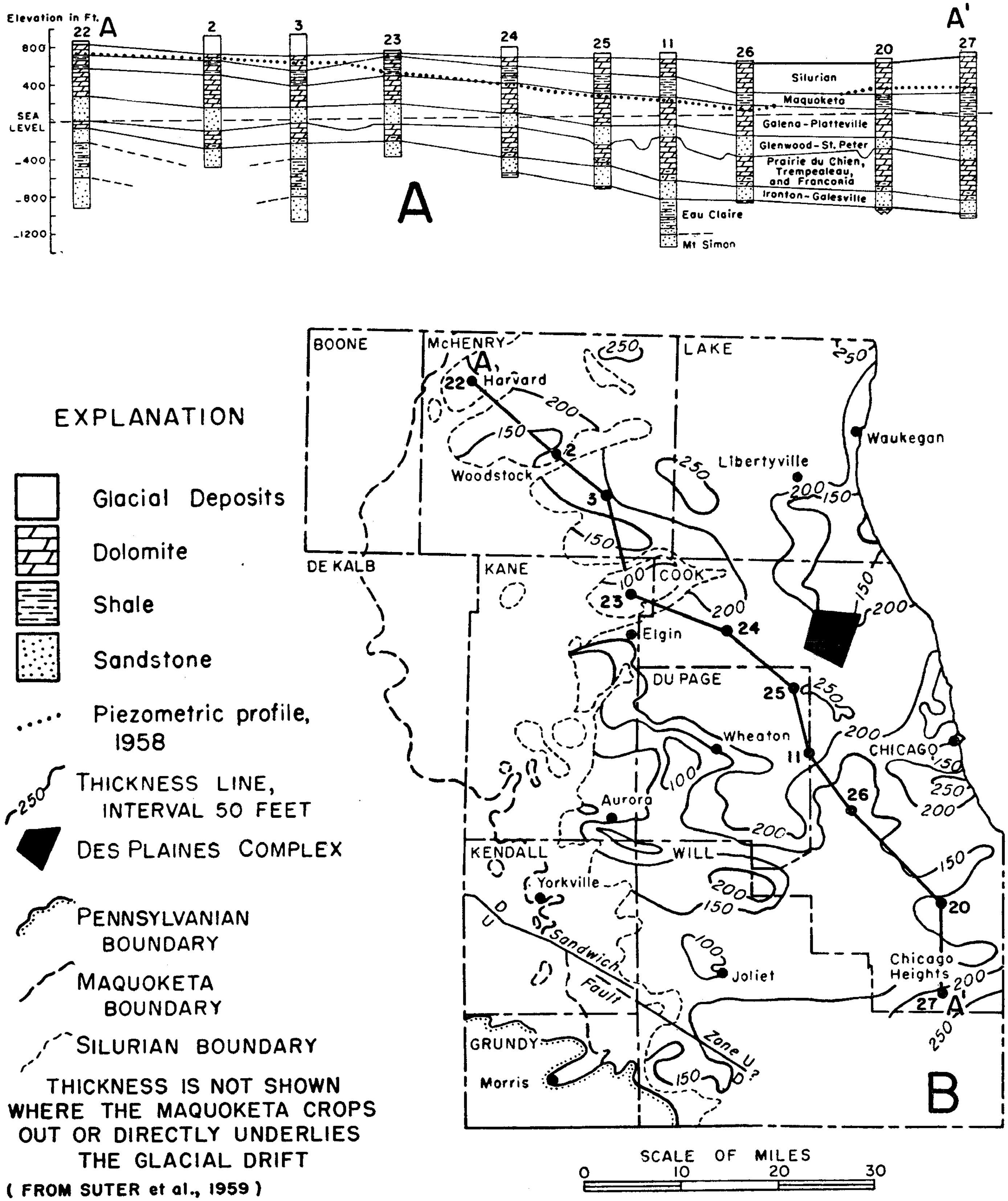


Figure 1. Geohydrologic cross section (A) and thickness of Maquoketa Formation (B) in northeastern Illinois

The quantity of water percolating through a given cross section of an aquifer is proportional to the hydraulic gradient (slope) of the piezometric surface and the coefficient of transmissibility, and can be computed by using the following modified form of the Darcy equation (see Ferris, 1959):

$$Q = TIL \qquad (1)$$

where:

Q = discharge, in gpd
T = coefficient of transmissibility, in gpd/ft
I = hydraulic gradient, in ft/mi
L = width of cross section through which discharge occurs, in mi

Using equation 1, the quantity of water moving southeastward through the aquifer at section B—B' was computed to be about 1 million gallons per day (mgd). Leakage downward through the Maquoketa Formation in area 1 was therefore about 1 mgd in 1864. As measured from figure 2A, area 1 is about 750 square miles. The recharge rate, $Q/area\ 1$, for the Cambrian-Ordovician Aquifer in area 1 was 1330 gpd/sq mi or about 0.9 gallon per minute per square mile (gpm/sq mi) in 1864.

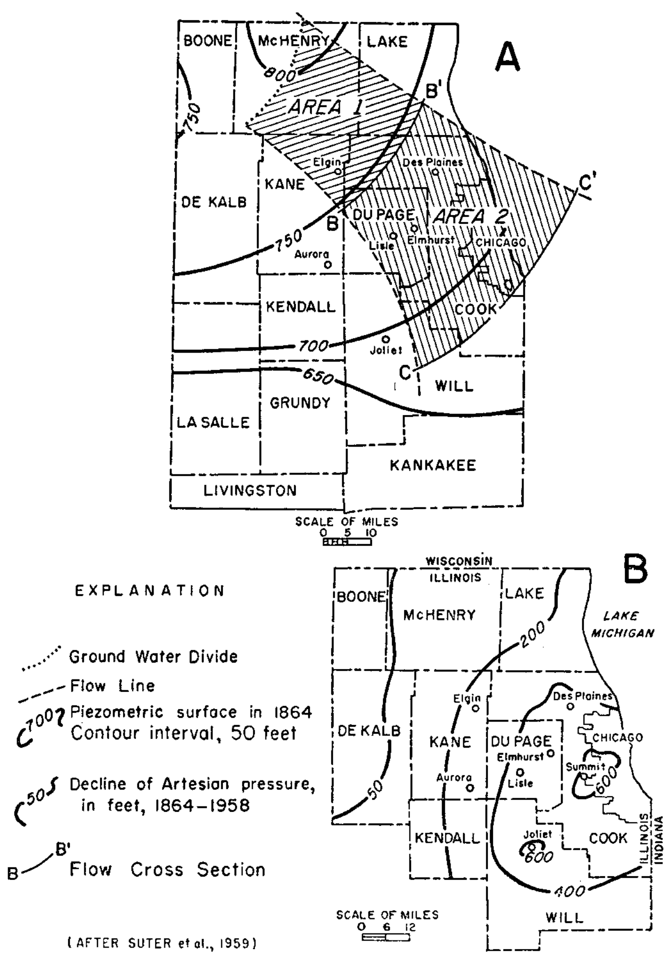


Figure 2. Piezometric surface of Cambrian-Ordovician Aquifer, about 1864 (A) and decline in artesian pressure, 1864-1958 (B) in northeastern Illinois

Flow-net analysis indicates that the upward leakage rate for area 2 was 450 gpd/sq mi or 0.3 gpm/sq mi in 1864 and the leakage rate for the Cambrian-Ordovician Aquifer increases to the north and west. Available geologic information supports this conclusion. The lower unit of the Maquoketa Formation, probably the least permeable of the three units (Bergstrom and Emrich, personal communication), thins to the west. In addition, the Maquoketa Formation is the uppermost bedrock formation below the glacial deposits in a large part of area 1 and locally may be completely removed by erosion.

The changes in artesian pressure produced by pumping since the days of early settlement have been pronounced and widespread. Pumpage from deep sandstone wells increased from 200,000 gpd in 1864 to about 78 mgd in 1958. Figure 2B shows the decline of artesian pressure in the Cambrian-Ordovician Aquifer from 1864 to 1958 as the result of heavy pumping. The greatest declines, more than 600 feet, have occurred in areas of heavy pumpage west of Chicago, at Summit, and Joliet. In 1958, the piezometric surface of the Cambrian-Ordovician Aquifer was several hundred feet below the water table in most of northeastern Illinois. Downward movement of water through the Maquoketa Formation was appreciable under the influence of large differentials in head between shallow deposits and the Cambrian-Ordovician Aquifer.

Even though the recharge rate is very low, leakage in 1958 through the Maquoketa Formation was appreciable. The area of the confining bed within the part of Illinois shown in figure 3A through which leakage occurred (4000 square miles) and the average head differential between the piezometric surface of the Cambrian-Ordovician Aquifer and the water table (300 feet) were great. Computations made by Walton (1960) indicate that leakage through the Maquoketa Formation within the part of Illinois shown in figure 3A was about 8.4 mgd or about 11 percent of the water pumped from deep sandstone wells in 1958. The average recharge rate in 1958 was 2100 gpd/sq mi which is much greater than the recharge rates computed for 1864. The differences between the heads in the aquifer and in the source beds above the Maquoketa Formation were 85 and 70 feet respectively in areas 1 and 2 in 1864, whereas the average head differential was 300 feet in 1958. Thus, recharge rates can change with time and are a function of the head differential between the piezometric surface of the Cambrian-Ordovician Aquifer and the water table of shallow deposits.

The piezometric surface map for 1958 in figure 3A was used to determine the recharge rate for the Cambrian-Ordovician Aquifer in areas where the Maquoketa Formation is missing. Flow lines were drawn from DeKalb County toward the cone of depression at Joliet to describe the flow channel shown in figure 3A. The flow channel is north of the Sandwich Fault Zone (see Suter et al., 1959),

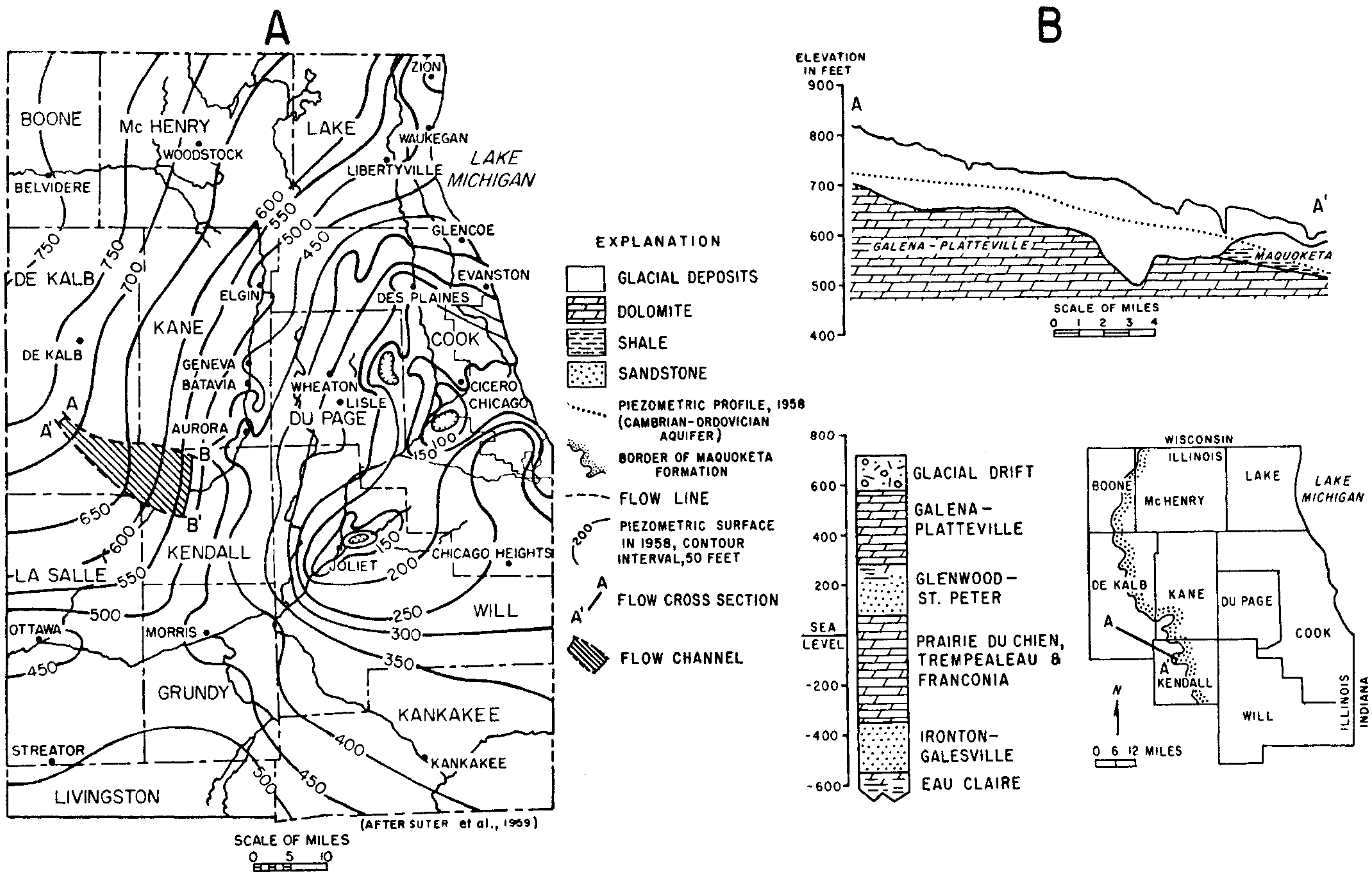


Figure 3. Piezometric surface of Cambrian-Ordovician Aquifer in 1958 (A) and geohydrologic cross section in DeKalb and Kendall Counties (B)

and the sections A—A' and B—B' are both west of the border of the Maquoketa Formation where the Galena-Platteville Dolomite immediately underlies the glacial drift as shown in figure 3B. In 1958 the piezometric surface was below the water table, and downward leakage through glacial deposits and underlying bedrock into the various units of the Cambrian-Ordovician Aquifer was occurring in the flow channel. The recharge rate was computed to be 18,000 gpd/sq mi or 12.5 gpm/sq mi by substituting data in equation 1 and the following equation (see Walton, 1962):

$$Q/A_l = [(Q_1 - Q_2) - h_t S A_l (2.1\text{X}10^8)]/A_l \quad (2)$$

where:

Q/A_l = recharge rate, in gpd/sq mi

Q_1 = quantity of water percolating through flow cross section B—B', in gpd (computed with equation 1)

Q_2 = quantity of water percolating through flow cross section A—A', in gpd (computed with equation 1)

h_t = average rate of water-level decline in flow channel, in feet per day (fpd)

S = coefficient of storage of Cambrian-Ordovician Aquifer, fraction

A_l = area of flow channel between flow lines and flow cross sections A—A' and B—B', in sq mi

At sections A—A' and B—B' the hydraulic gradients of the piezometric surface were about 8.3 and 15.0 ft/mi, respectively. The distances between flow lines at sections A—A' and B—B' were 1.5 and 7 miles, respectively. Based on data given by Suter et al. (1959), the average coefficients of transmissibility and storage of the Cambrian-Ordovician Aquifer within the flow channel are 20,000 gpd/ft and 0.00034, respectively. The area within the flow channel is 100 square miles, and the average water-level decline in 1958 was 2 feet per year (ft/yr) or 0.0055 fpd.

The recharge rate for the Cambrian-Ordovician Aquifer in areas west of the border of the Maquoketa Formation, computed by flow-net analysis of water-level data for 1958, is about 14 times as great as the recharge rate in area 1, computed by flow-net analysis of water-level data for 1864, where the Maquoketa Formation overlies the aquifer. Thus, the Maquoketa Formation greatly retards but does not completely prevent recharge to the Cambrian-Ordovician Aquifer.

Dolomite Aquifer in DuPage County

DuPage County, about six miles west of the corporate limits of Chicago, is underlain at depths averaging about

100 feet by a dolomite aquifer that has yielded large quantities of ground water for more than 70 years. The dolomite aquifer consists mostly of Silurian rocks; rocks of Silurian age in ascending order are the Alexandrian and Niagaran Series. The dolomite aquifer is overlain in most areas by glacial drift; the thickness of unconsolidated deposits ranges from 0 to more than 200 feet (*see figure 9 in* Zeizel et al., 1962). The glacial drift contains a high percentage of silt and clay in many places. Permeable zones within the dolomite aquifer are recharged by the vertical leakage of water through glacial drift and less permeable zones of the dolomite.

For a detailed discussion of the geology of DuPage County, the reader is referred to Suter et al. (1959) and Zeizel et al. (1962). The following geologic description is based largely upon these reports. Except in small areas in the north-central, western, and southwestern parts of the county, the bedrock surface beneath the glacial drift is formed by rocks of the Niagaran Series (*see figure 7 in* Zeizel et al., 1962). The Alexandrian Series occurs immediately below the glacial drift in narrow bands associated with buried bedrock valleys in north-central parts of the county, and in fairly extensive areas in the southwest corner of the county. The Maquoketa Formation underlies Silurian rocks and is the uppermost bedrock beneath the glacial drift only in a narrow band coinciding with the deeper portions of a buried bedrock valley in the north-central part of the county.

The Niagaran Series is composed chiefly of dolomite; however, shaly dolomite beds occur at the base, and reefs and associated strata are present in upper formations. The Niagaran Series has been removed by erosion mostly in the southwest corner of the county and in parts of the north-central portion of the county. The maximum known thickness is 175 feet, and thicknesses of more than 50 feet are common in the eastern two-thirds of the county. Basal shaly dolomite beds have a fairly uniform total thickness of about 30 feet except in the southwestern part of the county where the beds are missing. The Alexandrian Series is composed chiefly of dolomite; shale and argillaceous dolomite beds occur near the base, and relatively pure dolomite is present in upper formations. The Alexandrian Series occurs everywhere in the county except in a narrow band associated with a deep bedrock valley in the north-central part of the county. The thickness of the rocks commonly exceeds 40 feet and reaches 90 feet at some places.

The Silurian rocks increase in thickness from less than 50 feet in the northwestern and southwestern parts of the county to more than 250 feet in the southeastern part of the county (*see figure 16 in* Zeizel et al., 1962). The rocks dip to the southeast at an average rate of about 10 ft/mi; gentle folds pitch towards the southeast. The boundary of the Silurian rocks is about 6 miles west of the county.

The productivity of the dolomite aquifer is inconsistent; specific capacities of wells range from 0.6 to 530 gallons per minute per foot of drawdown (gpm/ft) and average 42 gpm/ft. However, the inconsistency has little effect on the regional response of the aquifer to pumping and areas of influence of production wells extend for considerable distances. On a regional basis the dolomite aquifer has high to moderate coefficients of transmissibility. Mean annual precipitation of 34.2 inches occurs in the county. Mean annual temperature is 49.6 F.

Recharge rates for the dolomite aquifer were estimated with a piezometric surface map and past records of pumpage and water levels. A comparison of hydrographs for dolomite wells and pumpage graphs indicates that water-level declines are directly proportional to pumping rates, and water levels will stabilize within a short time after each increase in pumpage. Thus, in the past, recharge has balanced discharge. In order to determine areas of recharge, a piezometric surface map (*see figure 52 in* Zeizel et al., 1962) was prepared with water-level data collected during August 1960. Total withdrawal from the dolomite aquifer in August 1960 was about 20 mgd. Areas of diversion, A_c, for production wells in the West Chicago area (1), Wheaton-Glen Ellyn-Lombard area (2), Downers Grove-Westmont-Clarendon Hills-Hinsdale area (3), and Argonne National Laboratory area (4), shown in figure 4, were delineated by flow-net analysis of the piezometric surface map. The boundaries of areas of diversion enclose areas within which the general movement of water in the aquifer is towards pumping centers.

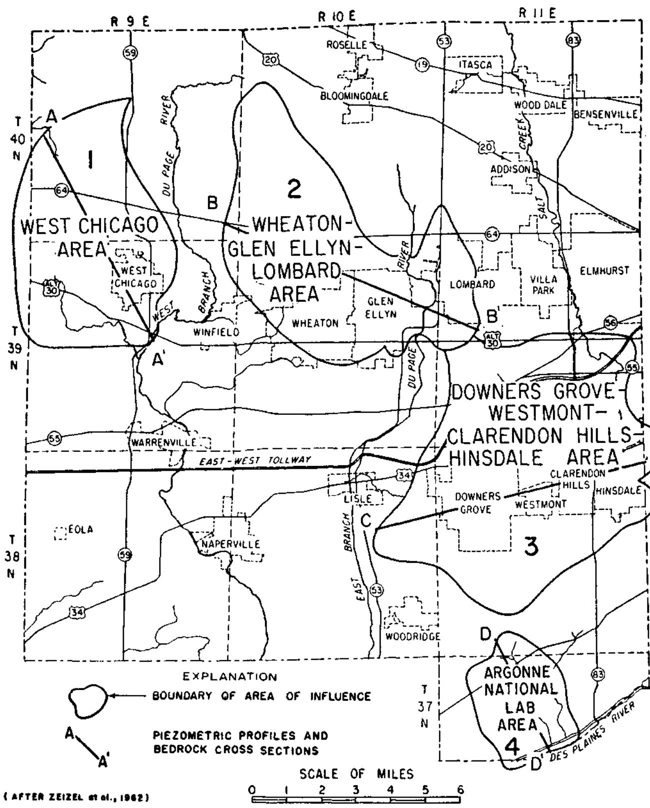


Figure 4. Areas of diversion in DuPage County, August 1960

Table 1. Rates of Recharge for Dolomite Aquifer in DuPage County

Area	Pumpage in 1960 *(mgd)*	Area of diversion *(sq mi)*	Recharge rate *(gpd / sq mi)*
1	1.8	28.0	64,000
2	4.5	32.5	138,000
3	6.3	46.2	136,000
4	1.2	7.6	158,000

(from Zeizel et al., 1962)

Measured areas of diversion, pumpage data, and recharge rates computed as the quotient of pumpage and area (Q/A_c) are given in table 1. Except for area 1, recharge rates for the areas are approximately the same. On a gross basis the glacial drift deposits are similar in character and thickness in the four areas and slightly more permeable in area 1 than in the other three areas. The profiles in figure 5 show that the piezometric surface

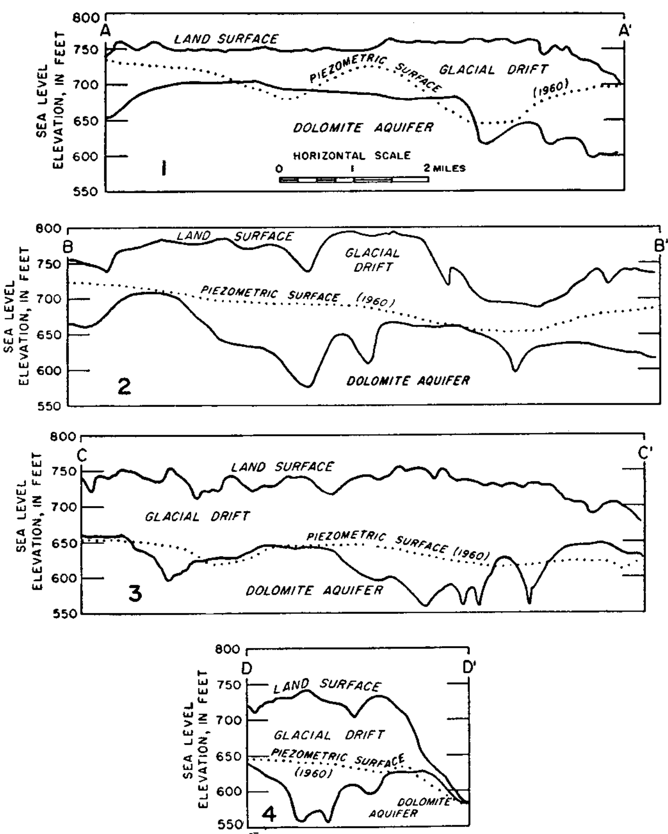


Figure 5. Geohydrologic cross sections in West Chicago area (1), Wheaton-Glen Ellyn-Lombard area (2), Downers Grove-Hinsdale area (3), and Argonne area (4)

is more than 50 feet below the ground surface in most of the areas. Average vertical hydraulic gradients do not differ appreciably from area to area. Thus, the low recharge rate in area 1 cannot be explained by differences in the character and thickness of the glacial drift or average vertical hydraulic gradients. In areas 2, 3, and 4 most of the water pumped is obtained from the thick rocks of the Niagaran Series above the shaly dolomite unit. The rocks of the Niagaran Series are thin in area 1, and much of the water withdrawn from wells is obtained from rocks of the Alexandrian Series below the shaly dolomite unit. The shaly dolomite unit greatly retards the vertical movement of ground water and is responsible for the low recharge rate in area 1.

Areas in DuPage County where recharge will be low and about the same as in the West Chicago area under heavy pumping conditions were delineated by assuming that recharge will be limited east of the Niagaran-Alexandrian contact where rocks of the Niagaran Series overlying the shaly dolomite unit are less than 25 feet thick. In addition, the recharge rate will be low in areas where water is obtained from the Maquoketa Formation because permeable dolomite beds of the formation are interbedded with, and often overlain by, beds of dolomitic shale with a very low permeability. Areas where recharge will be about the same as in areas 2, 3, and 4 were delineated by assuming that recharge will be high west of the Niagaran-Alexandrian contact where rocks of the Alexandrian Series are more than 25 feet thick and east of the Niagaran-Alexandrian contact where rocks of the Niagaran Series overlying the shaly dolomite unit are more than 25 feet thick. A map showing estimated recharge rates under heavy pumping conditions for the dolomite aquifer is shown in figure 6. It is probable that recharge to the dolomite aquifer aver-

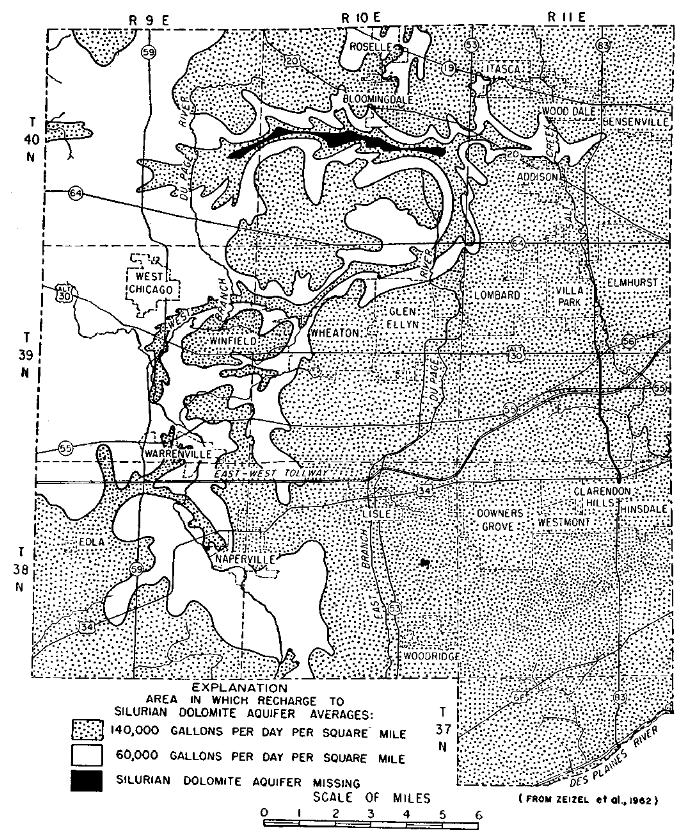


Figure 6. Estimated recharge rates for dolomite aquifer in DuPage County

ages about 140,000 gpd/sq mi in parts of the eastern one-third of the county and averages about 60,000 gpd/sq mi in large areas of the western two-thirds of the county.

Dolomite Aquifer in LaGrange Area

Water for municipal use at the villages of LaGrange and Western Springs, just east of DuPage County in western Cook County, is obtained largely from wells in Silurian rocks (Prickett et al., 1964). The dolomite aquifer averages 250 feet in thickness and is overlain largely by clayey materals of glacial origin. In 1962 the average withdrawal from municipal wells was about 2.4 mgd. Water-level declines resulting from heavy concentrated pumpage averaged about 70 feet in the vicinity of LaGrange.

For a detailed discussion of the geology in the LaGrange area, the reader is referred to Suter et al. (1959) and Horberg (1950). The following section is based largely on these two reports. The LaGrange area is covered largely by glacial drift in all but a few places where bedrock is exposed. The thickness of unconsolidated deposits is variable but averages about 50 feet in the Cook County part and about 100 feet in the DuPage County part of the LaGrange area (*see figure 111 in* Prickett et al., 1964). The basal portion of the glacial drift contains an extensive deposit of sand and gravel which varies from a few feet or less to more than 50 feet in thickness. The remainder of the unconsolidated deposits is composed largely of clayey materials (confining bed) with intercolated lenses and layers of sand and gravel.

The rocks immediately underlying the glacial drift are Silurian in age. A bedrock valley trends eastward across the center of the area (*see figure 112 in* Prickett et al., 1964). The channel of the bedrock valley averages ½ mile in width, has walls of moderate relief, and averages 50 feet in depth.

The Silurian rocks consist of the Alexandrian Series overlain by the Niagaran Series and are underlain by the Maquoketa Formation of Ordovician age. The thickness of the Silurian rocks generally increases from less than 150 feet in the northwestern part to more than 350 feet in the southeastern part of the LaGrange area (*see figure 113 in* Prickett et al., 1964). The rocks of the Niagaran Series range from relatively pure dolomite to silty, argillaceous, and cherty dolomite with some thin shale beds and reefs. The Alexandrian Series is composed chiefly of dolomite that increases in clastic content from the top of the series downward. Shale and argillaceous dolomite beds occur near the base of the series.

Logs of wells in table 2 illustrate the character of the unconsolidated deposits and the bedrock. The glacial drift consists largely of till that contains a high percentage of silt and clay.

Because of heavy concentrated pumpage from rock quarries and wells in the LaGrange area, extensive dewatering of upper portions of the dolomite aquifer has taken place. The coefficient of transmissibility of the aquifer averages about 100,000 gpd/ft in areas where dewatering has not occurred; in areas where dewatering has occurred the coefficient of transmissibility averages only

Table 2. Logs of Selected Wells in LaGrange Area

Well number	Record* number	Formation	Thickness *(ft)*	Depth *(ft)*
DUP 38N11E-				
2.1g	42074	Yellow clay	8	8
		Gray clay	17	25
		Gravel and clay	26	51
		Silty gravel	25	76
		Coarse gravel and clay	4	80
		Gravel and clay	32	112
		Broken dolomite	5	117
		Hard gray dolomite	8	125
		Gray dolomite	40	165
		White dolomite	23	188
		Gray dolomite	35	223
		White dolomite	4	227
		Gray dolomite	45	272
		Green shale	2	274
		Broken dolomite and shale	16	290
		Gray shale	21	311
COK 38N12E-				
11.6c	10746	Pleistocene Series		
		Drift	10	10
		Silurian System		
		Niagaran Series		
		Dolomite, white to light buff	75	85
		Dolomite cherty, glauconitic, buff	45	130
		Dolomite, light gray to light buff	60	190
		Dolomite, light buff, silty	28	218
		Alexandrian Series		
		Kankakee Formation		
		Dolomite, white to buff, fine	47	265
		Dolomite, cherty, glauconitic, buff	22	287
		Dolomite, cherty, buff, brown, silty	23	310
		Edgewood Formation		
		Dolomite, shaly, silty; shale	30	340
		Dolomite, very shaly, gray, silty	20	360
		Ordovician System		
		Maquoketa Formation		
		Shale, light gray, weak, little dolomite	5	365

**State Geological Survey sample set number*

30,000 gpd/ft. During the period 1900 to 1962 more than 100 well-production tests were made on dolomite wells in the vicinity of LaGrange. The productivity of the dolomite aquifer is inconsistent; specific capacities of wells range from 0.12 to 500 gpm/ft and average about 29 gpm/ft.

The mean annual precipitation is 33.13 inches; mean annual temperature is 49.8 F.

Most of the recharge to the dolomite aquifer is from vertical leakage of water through overlying glacial deposits. A comparison of water-level hydrographs and pumpage graphs indicates that, in general, drawdown is proportional to pumpage, and water levels stabilize shortly after each pumpage increase. Thus, in the past recharge has balanced pumpage. The rate of recharge in 1962 was estimated using the piezometric surface map in figure 7. Pronounced cones of depression are centered around LaGrange and the quarries in the southeastern part of the LaGrange area. Other cones of depression are present at Downers Grove, Hinsdale, and Clarendon Hills. Flow lines were drawn at right angles to the piezometric surface contours to define the area of diversion of production wells at LaGrange and Western Springs. As measured from figure 7, the area of diversion is about 18 square miles. Ground-water pumpage within the area of diversion averaged about 2.9 mgd in 1962. Computations show that the quotient of the pumpage and the area of diversion was about 161,000 gpd/sq mi. The recharge rate in 1962 for the LaGrange area was, therefore, about 161,000 gpd/sq mi.

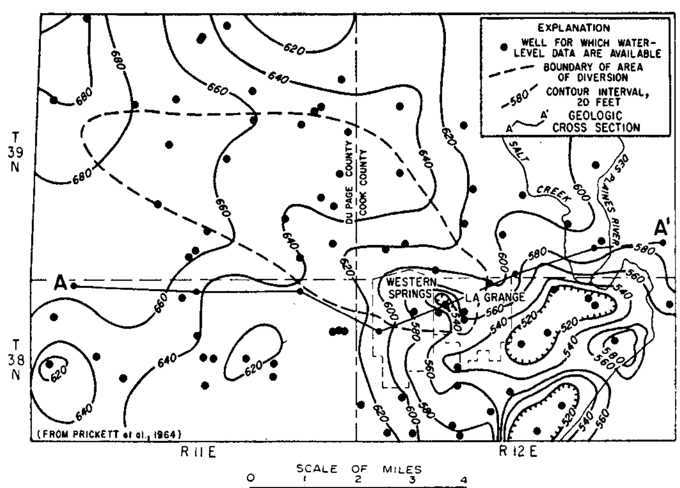


Figure 7. Piezometric surface of dolomite aquifer in LaGrange area, November 1962

Dolomite Aquifer in Chicago Heights Area

The most heavily pumped aquifer in the Chicago Heights area (about 27 miles south of the Loop in Chicago) is a dolomite aquifer of Silurian age which averages 400 feet in thickness (Prickett et al., 1964). The aquifer is overlain by materials of glacial origin (largely till) averaging 75 feet thick. In 1962 the average daily withdrawal from municipal wells at Chicago Heights and Park Forest was 7.84 mgd. Exploitation of ground-water resources caused a local water-level decline averaging about 40 feet within a 3-mile radius of Chicago Heights.

For a detailed discussion of the geology in the Chicago Heights area, the reader is referred to Suter et al. (1959) and Horberg (1950). The following section is based largely upon these two reports. The Chicago Heights area is covered mostly with glacial drift which varies in thickness from a few feet in the north-central part to more than 100 feet in the southern part (*see figure 89 in* Prickett et al., 1964). Numerous bedrock exposures are found northeast of Chicago Heights. Logs of wells given in table 3 illustrate the nature of the unconsolidated deposits above bedrock. The glacial drift

Table 3. Logs of Selected Wells in Chicago Heights Area

Well number	Formation	Thickness (*ft*)	Depth (*ft*)
WIL 34N14E- 8.1a	Sandy clay	12	12
	Fine sand	68	80
	Dolomite	80	160
	Siltstone (dolomitic)	35	195
	Dolomite	—	264
WIL 34N14E-16.8a	Clay	30	30
	Sand	45	75
	Dolomite	—	265
WIL 34N14E-21.6b	Soil and clay	70	70
	Sand and gravel	30	100
	Limestone	—	379
WIL 34Nl4E-33.6h	Soil	30	30
	Sandy clay	55	85
	Sand and gravel	30	115
	Limestone	402	517
	Shale	—	526
COK 35N13E- 1.2c	Pleistocene Series		
	Clay, gravel, and boulders	65	65
	Silurian System		
	Niagaran Series		
	Lime and broken lime	23	88
	Red and mixed shale	36	124
	Lime rock	1	125
	Red and mixed shale	23	148
	Lime rock	12	160
	Dolomite, light gray, pink, green, fine	85	245
	Dolomite, white, fine	40	285
	Dolomite, gray, fine	20	305
	Dolomite, white, very fine	20	325
	Dolomite, silty, fine	30	355
	Alexandrian Series		
	Kankakee Formation		
	Dolomite, light buff, fine	15	370
	Dolomite, buff, fine to medium	20	390
	Dolomite, cherty, buff, fine to medium	20	410

contains a thick and extensive deposit of sand and gravel immediately above the bedrock. The thickness of the sand and gravel deposit, often exceeding 25 feet, tends to increase with increasing drift thickness, thinning generally from the southwest to the northeast. The average thickness of the sand and gravel deposit ranges from about 15 feet in the northeastern part to about 40 feet in the southern part of thé Chicago Heights area. Relatively impermeable deposits (confining bed), consisting of sandy and silty clay and gravel, overlie the basal sand and gravel deposits. The thickness of these clayey materials varies considerably but often exceeds 25 feet.

Rocks of Silurian age form the bedrock surface throughout the Chicago Heights area. They are mainly dolomites, though shaly dolomite beds occur at the base. The Silurian rocks are divided into the Niagaran Series above and the Alexandrian Series below. Based on logs of a few wells that completely penetrate the Silurian rocks, the combined thickness of the Niagaran and Alexandrian Series is fairly uniform and averages about 400 feet. The Niagaran Series is white to light gray, finely to medium crystalline, compact dolomite with varying amounts of shale and argillaceous dolomite beds, and averages about 360 feet in thickness. The Alexandrian Series is relatively thin, averaging about 40 feet in thickness, and is composed chiefly of dolomite; shale and argillaceous dolomite beds occur near the base. The character of the rocks is illustrated by the logs of selected wells in table 3.

In general, the bedrock surface slopes to the northeast toward Lake Michigan at an average rate of about 7 ft/mi (*see figure 91 in* Prickett et al., 1964). Chicago Heights is located on a bedrock upland; the maximum elevation of the bedrock surface reaches about 660 feet.

As a result of the close spacing of wells and well fields and the heavy pumpage at Chicago Heights, extensive dewatering of upper portions of the aquifer has taken place. The coefficient of transmissibility of the dolomite aquifer averages about 65,000 gpd/ft in areas where no dewatering has taken place; the coefficient of transmissibility in areas where extensive dewatering has taken place averages about 22,000 gpd/ft. The gravity yield of the upper portion of the dolomite aquifer is about 0.03. During the period 1900-1962, well-production tests were made on more than 150 dolomite wells in the Chicago Heights area. Specific capacities of the wells range from 0.38 to 3450 gpm/ft and average 54 gpm/ft.

The mean annual precipitation is 33.65 inches; mean annual temperature is 49.1 F.

A comparison of water-level hydrographs and pumpage graphs indicates that in general water-level declines are proportional to the pumpage rates. Although the water levels vary considerably from time to time because of shifts in pumpage in well fields and variations in recharge from precipitation, the hydrographs show no "permanent" decline in water levels that cannot be explained by pumpage increases and short-term dry periods. The relation between water-level decline and pumpage suggests that recharge has balanced discharge in the past.

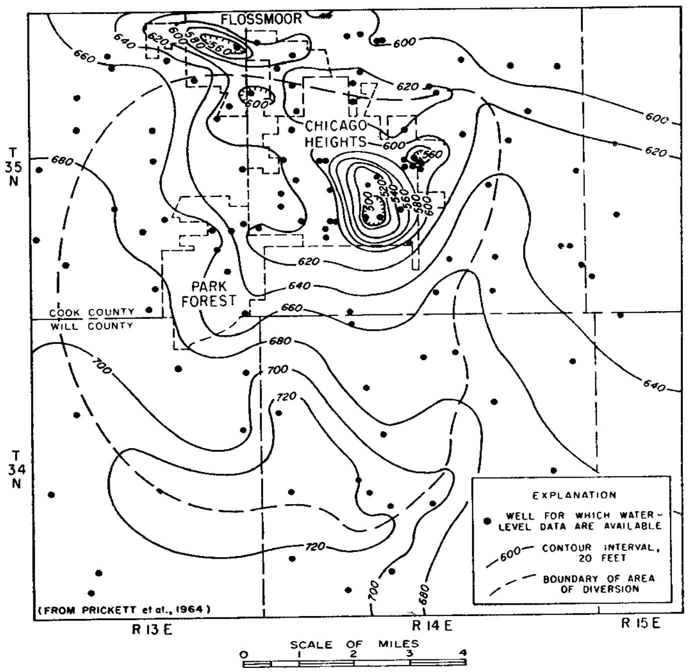


Figure 8. Piezometric surface of dolomite aquifer in Chicago Heights area, August 1962

As shown in figure 8, ground water in the Chicago Heights area moves in all directions from topographic uplands toward streams and well fields. Heavy concentration of pumpage has produced cones of depressions in many parts of the Chicago Heights area. The piezometric surface map shows well-defined cones of depression at Chicago Heights and Flossmoor and in the vicinity of the industrial complex northeast of Chicago Heights.

Flow lines were drawn at right angles to the piezometric surface contours to define the area of diversion of production wells in the vicinity of Chicago Heights. The area of diversion as measured from figure 8 is about 60 square miles. The piezometric surface map of the dolomite aquifer was compared with water-level data for the period prior to development, and water-level changes were computed. The greatest declines in the piezometric surface occurred in the immediate vicinity of Chicago Heights and averaged about 100 feet. The average slope of the piezometric surface in areas unaffected by pumpage is about 15 ft/mi. Gradients are much steeper and exceed 100 ft/mi near and within cones of depression.

Recharge to the dolomite aquifer occurs locally, mostly as vertical leakage of water through unconsolidated deposits, and has precipitation as its source. The rate of recharge to the aquifer was estimated using the piezometric surface map and past records of pumpage and water levels. The area of diversion of pumping was delineated as explained earlier; pumpage within the area of diversion was 13.5 mgd in 1962. Because recharge balanced discharge, the average rate of recharge to the aquifer during 1962 is the quotient of the average pumping rate and the area of diversion. Computations show that the average recharge rate to the dolomite aquifer in the Chicago Heights area was about 225,000 gpd/sq mi in 1962.

The aquifer is not recharged entirely by the direct percolation of precipitation to the water table. Small amounts of recharge to the aquifer by induced infiltration of surface water occurs because the piezometric surface is below stream levels and the streambeds have some permeability. In some cases the streams lie directly on the aquifer where bedrock outcrops east of Chicago Heights.

Dolomite and Sand and Gravel Aquifers in Libertyville Area

Water for municipal use at the villages of Libertyville and Mundelein, about 25 miles north-northwest of Chicago, is obtained locally from wells in deeply buried dolomite and sand and gravel aquifers (Prickett et al. 1964). The dolomite aquifer averages 150 feet thick and is overlain mostly by clayey materials of glacial origin averaging 175 feet thick. The sand and gravel aquifer is thin and occurs above bedrock at the base of the glacial drift. In 1962 the average daily withdrawal from municipal wells at Libertyville and Mundelein was 2.14 mgd. Exploitation of ground-water resources caused a local water-level decline averaging about 70 feet within a 2-mile radius of Libertyville.

For a detailed discussion of the geology in the Libertyville area, the reader is referred to Suter et al. (1959) and Horberg (1950). The following section is based largely upon these two reports. The Libertyville area is covered mostly with glacial drift which exceeds 200 feet in thickness at places. The bedrock immediately underlying the glacial drift is mainly dolomite of the Niagaran Series of Silurian age. In the western part of the Libertyville area the Niagaran Series has been removed by erosion (*see figure 65 in* Prickett et al., 1964) and dolomite of the Alexandrian Series of Silurian age is the uppermost bedrock. The Maquoketa Formation of Ordovician age underlies the Alexandrian Series. The glacial drift contains a thick and fairly extensive deposit of sand and gravel immediately above the bedrock which commonly exceeds 20 feet in thickness. The remainder of the glacial drift is mainly composed of clayey materials (confining bed) and commonly exceeds 150 feet in thickness. Lenses of sand and gravel are intercolated in the confining bed.

A bedrock valley extends northeastward across the center of the Libertyville area (*see figure 66 in* Prickett et al., 1964). The channel of the bedrock valley exceeds a mile in width in most places, has walls of moderate relief, and averages about 50 feet in depth.

The Niagaran Series is composed chiefly of dolomite, however, shaly dolomite beds occur at the base. The Niagaran Series in the Libertyville area is relatively more argillaceous than the Niagaran Series in other parts of northeastern Illinois. The thickness of the Niagaran Series varies but averages about 60 feet and generally increases from the Niagaran-Alexandrian contact in the western part toward the southeastern part. The Alexandrian Series is composed chiefly of dolomite; shale and argillaceous dolomite beds occur near the base. The thickness of the Alexandrian Series commonly exceeds 75 feet and averages about 90 feet. The thickness of the Silurian rocks increases from less than 50 feet in the western part to over 300 feet in the southeastern corner of the Libertyville area (*see figure 67 in* Prickett et al., 1964).

Table 4. Logs of Selected Wells in Libertyville Area

Well number	Formation	Thickness (ft)	Depth (ft)
LKE 44N11E-16.1b2	Fill	6	6
	Clay	2	8
	Gravel and clay	47	55
	Clay	60	115
	Sand	10	125
	Clay and gravel	25	150
	Gravel	40	190
	Dolomite	110	300
LKE 44N11E-30.6c2*	Soil, dark brown; till, yellow; clay, fine	15	15
	Soil, dark brown; little till; gravel, fine	10	25
	Clay, gray; silt, gray	40	65
	Till, gray, sandy	10	75
	Sand, yellow, fine to coarse, clean	10	85
	Clay, gray; some silt	85	170
	Sand, medium to coarse; gravel, fine, clean	20	190

**Record of State Geological Survey sample set 19869*

Logs in table 4 illustrate in general the nature of the unconsolidated deposits above bedrock. The unconsolidated deposits are mainly glacial drift and increase in thickness from less than 100 feet southeast of Libertyville to over 300 feet in the western part of the Libertyville area (*see figure 69 in* Prickett et al., 1964). The glacial drift consists largely of deposits of till that contain a high percentage of silt and clay.

Logs of wells show that permeable sand and gravel deposits are found in numerous zones within the glacial drift. Sand and gravel occur at the base of the glacial drift over most of the Libertyville area. The thickness of this zone is variable but averages 20 feet except in the vicinity of the channel of the buried bedrock valley near the center of the area. Based on logs of a few wells which completely penetrate the glacial drift, the thickness of this basal sand and gravel deposit increases in the vicinity of the bedrock valley and commonly exceeds 40 feet. Geologic data suggest that the materials of the basal sand and gravel deposit may be predominantly fine-grained in the vicinity of the buried bedrock valley. Relatively impermeable deposits consisting of sandy and silty clay and gravel overlie the basal sand and gravel deposits. The thickness of these clayey materials varies considerably but averages about 175 feet. Deposits of permeable sand and gravel of limited areal extent are interbedded in the confining bed.

During the period 1929 to 1961, well-production tests were made by the State Water Survey on more than 80 dolomite wells in and near the Libertyville area. The productivity of the dolomite aquifer is inconsistent; specific capacities of wells range from 0.13 to 358 gpm/ft and average about 7 gpm/ft. Data for municipal and industrial wells obtaining water from glacial drift aquifers in the Libertyville area indicate that the specific capacities of sand and gravel wells range from 1.0 to 47.4 gpm/ft and average about 14 gpm/ft. Flow-net analysis suggests that the average coefficient of transmissibility of the part of the dolomite aquifer within the Libertyville cone of depression is 9500 gpd/ft.

The mean annual precipitation is 32.14 inches; mean annual temperature is 48.7 F.

A comparison of water-level hydrographs and pumpage graphs indicates that water levels stabilize after each pumpage increase, drawdowns are proportional to pumpage, and in the past recharge has balanced withdrawals.

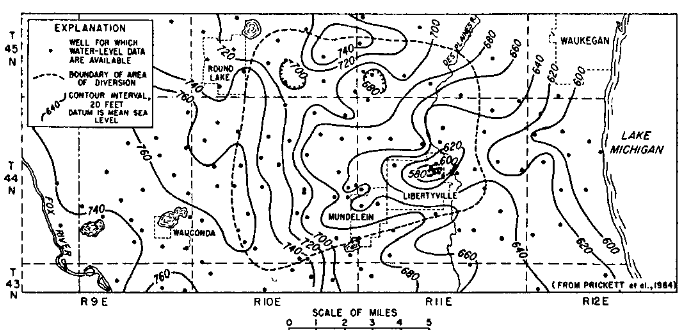


Figure 9. Piezometric surface of dolomite aquifer in Libertyville area, August 1962

The piezometric surface map in figure 9 represents the elevation to which water will rise in a well completed in the dolomite aquifer and does not usually coincide with the position of the water table in shallow sand and gravel aquifers. The map was prepared from water-level measurements made mostly during the months of July and August 1962. A pronounced cane of depression is centered around Libertyville and Mundelein. Other cones of depression are present at the village of Grays Lake and at Wildwood Subdivision in the north-central part of the Libertyville area. Ground-water movement is in all directions toward well fields or topographic lowlands. Flow lines were drawn at right angles to the piezometric surface contours to define the area of diversion of production wells. As measured from figure 9, the area of diversion is about 58 square miles.

The piezometric surface map of the Silurian dolomite aquifer was compared with water-level data for the period prior to development and water-level changes were computed. The greatest declines in the piezometric surface occurred in the immediate vicinity of Libertyville and averaged about 85 feet.

Recharge to aquifers in the Libertyville area occurs locally as vertical leakage of water through clayey materials and has precipitation as its source. The quotient of the quantity of leakage and the area of diversion is the rate of recharge. Pumpage was 3 mgd in 1962 and the area of diversion was 58 square miles, therefore, the recharge rate to the Silurian dolomite aquifer was about 52,000 gpd/sq mi in 1962.

Sand and Gravel Aquifer at Woodstock

Water for municipal use at the city of Woodstock, about 50 miles northwest of Chicago, is obtained from wells in deeply buried sand and gravel aquifers (Prickett et al., 1964). The most heavily pumped aquifer is a layer of sand and gravel of large areal extent which averages about 50 feet thick. In 1962 the average daily withdrawal was about 1.9 mgd.

For a detailed discussion of the geology of the Woodstock area the reader is referred to Suter et al. (1959). The following section is largely based on this report. The Woodstock area is covered with glacial drift which commonly exceeds 200 feet in thickness. The bedrock beneath the glacial drift is mainly dolomite of the Alexandrian Series of Silurian age. In a narrow belt averaging about 1 mile wide south and east of Woodstock, it is largely shale of the Maquoketa Formation of Ordovician age (*see figure 37 in* Prickett et al., 1964). The glacial drift contains thick and extensive deposits of sand and gravel in two zones; near the surface (upper aquifer) and immediately above bedrock (lower aquifer). The upper and lower aquifers exceed 30 feet in thickness at many places and are separated by clayey materials (confining bed) commonly exceeding 75 feet in thickness.

A bedrock valley extends northeastward across the southern part of the Woodstock area (*see figure 36 in* Prickett et al., 1964). The channel of the bedrock valley, roughly delineated by 650-foot contours, is over 1.5 miles wide in most places, has walls of moderate to low relief, and averages about 50 feet in depth. The bedrock surface at Woodstock slopes southeastward toward the channel of the bedrock valley and has an average elevation of 700 feet.

Table 5. Logs of Selected Wells in Woodstock Area

Well number	Formation	Thickness (ft)	Depth (ft)
MCH 44N7E- 5.7d2	Cinders and sand	4	4
	Sand and boulders	50	54
	Clay and boulders	24	78
	Sand	9	87
	Clay and boulders	25	112
	Sand and gravel	28	140
	Clay and boulders	2	142
	Fine sand	12	154
	Clay	3	157
	Coarse sand	5	162
	Clay	4	166
	Fine sand	5	171
	Clay	7	178
	Gravel	13	191
	Clay	6	197
	Gravel and sand	9	206
MCH 45N7E-32.7c*	Top soil	0	1
	Yellow silty clay with streaks of gray sand and gray clay	1	7
	Gray sandy clay	7	31
	Gray fine sand to coarse gravel	31	43
	Reddish sandy clay, gravel embedded	43	118
	Reddish clay with streaks of fine sand	118	123
	Gray, tight, fine to coarse sand	123	130
	Reddish soft sandy clay	130	134
	Gray fine sand	134	138
	Gray soft sandy clay	138	141
	Tight, gray, fine sand to medium gravel; occasional streak of clay	141	152
	Gray clay	152	154
	Tight fine sand to medium gravel	154	158
	Gray tight fine sand to coarse gravel	158	164
	Boulders and limestone chips, at 166 ft lost circulation	164	171
	Solid limestone	172	176

**Record of State Geological Survey sample set 35322*

Logs in table 5 illustrate the nature of the unconsolidated deposits above bedrock. The unconsolidated deposits are mostly glacial drift and increase in thickness from 150 feet in the northwestern part of the Woodstock area to more than 300 feet southeast of Woodstock (*see figure 39 in* Prickett et al., 1964). The entire thickness of the unconsolidated deposits has been penetrated in only 10 wells located mostly at Woodstock. The thickness in other areas is based on regional bedrock surface and topographic maps. The glacial drift consists largely of deposits of till that contain a high percentage of silt and clay. Logs of wells show that widely distributed permeable sand and gravel are found in two major zones within the glacial drift; near the surface and immediately above bedrock. Sand and gravel is encountered at many places at shallow depths. The thickness of this zone, herein termed the "upper aquifer," is variable but averages 30 feet. Data are not sufficient to delineate the boundaries of the upper aquifer, but available information suggest that the upper aquifer has a large areal extent in the Woodstock area. Large supplies of sand and gravel have been mined from several gravel pits in the upper aquifer. Medium- to coarse-grained sand and gravel occur at the base of the glacial drift over most of the Woodstock area. The thickness of this zone, herein termed the "lower aquifer," is variable but averages 50 feet except in the vicinity of the channel of the buried bedrock valley southeast of Woodstock. The lower aquifer is interbedded with lenses of clay at places, and it changes in character from place to place. Based on the regional bedrock-surface map and logs of a few wells which do not completely penetrate the glacial drift, the thickness of the lower aquifer increases from about 50 feet at Woodstock to more than 150 feet in the channel of the buried bedrock valley southeast of Woodstock. Logs of Wells and other geologic data suggest that the materials of the lower aquifer may be predominantly fine-grained in the vicinity of the buried bedrock channel. Mechanical (particle-size) analyses of samples of the materials obtained from test wells show that the lower aquifer is composed mainly of medium to very coarse sand (*see figure 40 in* Prickett et al., 1964). The Upper and lower aquifers are separated by beds of sandy and silty clay and gravel that are relatively impermeable. The thickness of this zone, herein termed the "confining bed," is variable but averages 80 feet. The confining bed is interbedded with permeable sand and gravel deposits (middle aquifer) of limited areal extent. The thickness of the middle aquifer is variable and at places exceeds 10 feet.

Based on aquifer and well-production test data the coefficients of transmissibility, permeability, and storage of the lower aquifer are 57,000 gpd/ft, 1100 gpd/sq ft, and 0.00034, respectively. Specific capacity data indicate that the coefficients of transmissibility and permeability of the middle aquifer are 170,000 gpd/ft and 8500 gpd/sq ft, respectively. The coefficients of transmissibility and permeability of the upper aquifer are 200,000 gpd/ft and 6600 gpd/sq ft based on aquifer-test data.

The mean annual precipitation is 32.12 inches; mean annual temperature is 48.6 F.

A comparison of water-level hydrographs and pumpage graphs indicates that water-level decline is proportional to the rate of pumpage. The consistent relationship between decline and pumpage, and the fact that water levels stabilize after each pumpage increase, indicate that in the past recharge has balanced withdrawals.

Prior to development, the piezometric surfaces of the middle and lower aquifers were near the land surface in the northern half of the Woodstock area and fairly high under the surrounding uplands to the south and east. Data on water levels in wells prior to development suggest that the piezometric surfaces of the middle and lower aquifers were subdued replicas of the topography. The general movement of ground water was from the uplands toward the streams draining the area. Ground-water divides roughly coincided with topographic divides. The approximate piezometric surface of the lower aquifer after development is shown in figure 10. The map

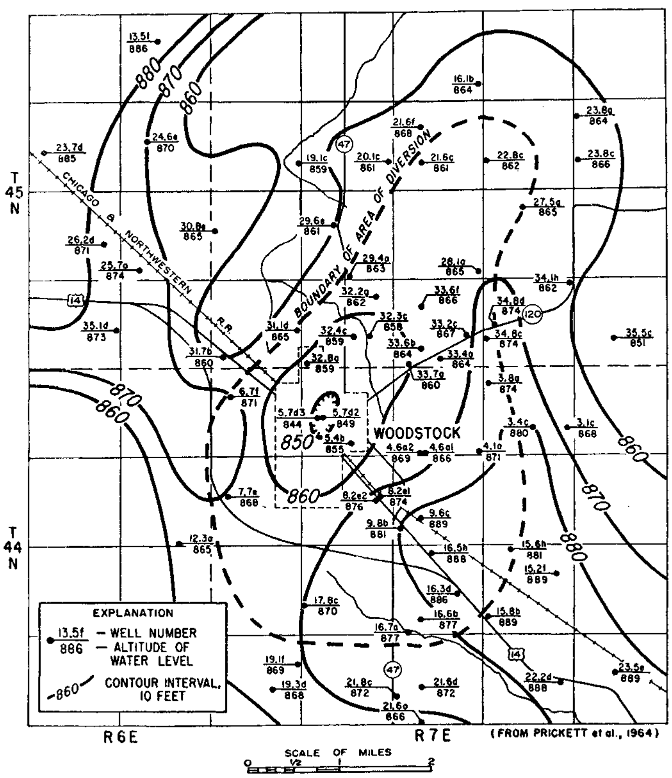


Figure 10. Piezometric surface of sand and gravel aquifer in Woodstock area, September 1962

was prepared from water-level measurements made during the latter part of August and the early part of September 1962. A pronounced cone of depression is centered at the old municipal well field. A ground-water ridge occurs southeast of Woodstock and a ground-water trough exists north of Woodstock. Pumping has reduced natural discharge of ground water to the shallow aquifer north of Woodstock. Contours are warped around the new municipal well field. As shown in figure 10, ground-water movement is in all directions toward well fields or topographic lowlands. Flow lines were drawn at right angles to the piezometric surface contours from production wells up gradient to define the area of diversion of production wells. As measured from figure 10, the area of diversion is 15 square miles.

Recharge to aquifers in the Woodstock area occurs locally as vertical leakage of water through glacial deposits and has precipitation as its source. A large proportion of precipitation runs off to streams or is discharged by evapotranspiration without reaching aquifers. Some precipitation reaches the water table and becomes ground water. Part of the water stored temporarily in the upper aquifer moves downward through the confining bed and into the middle and lower aquifers. Vertical movement is possible because of the large differentials in head between the water table in the upper aquifer and the piezometric surfaces of the middle and lower aquifers. The rate of recharge to the lower aquifer was estimated from the piezometric surface map in figure 10 and the quantity of leakage. The quotient of the quantity of leakage (pumpage) and the area of diversion is the rate of recharge. Pumpage was 1.9 mgd and the area of diversion was about 15 square miles, therefore, the recharge rate to the lower aquifer was about 127,000 gpd/sq mi in 1962.

Sand and Gravel Aquifer near Joliet

Water for municipal use at the city of Joliet, about 25 miles southwest of Chicago, is obtained in part from wells in a shallow sand and gravel aquifer northeast of the city (Prickett et al., 1964). The aquifer is a semi-infinite strip of sand and gravel approximately 2 miles wide and 60 feet thick. Dolomite with some permeability bounds the aquifer on the sides and bottom. The aquifer is overlain by fine-grained materials averaging 30 feet thick. Since 1951 when the aquifer was first tapped by the city of Joliet, the average withdrawal from a 5-well system has been about 3.7 mgd.

For a detailed discussion of the geology in the Joliet area, the reader is referred to Horberg and Emery (1943) and Suter et al. (1959). The following section is based upon these two reports. The area east of Joliet is largely covered with glacial drift which seldom exceeds 100 feet in thickness. The bedrock immediately beneath the glacial drift is mainly dolomite of Silurian age. Silurian rocks commonly exceed 250 feet in thickness, except where they have been deeply eroded as in buried bedrock valleys, and yield moderate amounts of water to wells. Large deposits of water-yielding sand and gravel are scarce in the glacial drift, and they occur chiefly in existing or buried valleys and as lenticular and discontinuous layers. The glacial drift is more than 100 feet thick and contains thick deposits of sand and gravel in a deeply buried valley which extends northeastward from Joliet for a distance of at least 10 miles. Two large bedrock valleys extend northeastward from Joliet and roughly coincide with the present valleys of Spring Creek and Hickory Creek (*see figure 4 in* Prickett et al., 1964). These two bedrock valleys are connected by a third short bedrock valley about 2 miles east of Forest Park. An island-like upland is surrounded by the bedrock valleys and rises 100 feet above the floors of the bedrock valleys. The channels of the bedrock valleys are about 1 mile wide, have relatively steep walls, and average 100 feet in depth. The buried valley beneath Spring Creek and the connecting buried valley are collectively called Hadley Valley (see Horberg and Emery, 1943).

Table 6. Logs of Selected Wells near Joliet

Well number	Formation	Thickness (*ft*)	Depth (*ft*)
WIL 35N11E-5.4g	Soil and clay	10	10
	Sand and gravel, clayey	10	20
	Sand and gravel, cemented	20	40
	Sand and gravel, coarse	20	60
	Fine sand	10	70
	Sand and gravel, clean	13	83
	Muddy clay, soft, till	22	105
	Sand and gravel, some till	7	112
	Sand and gravel, clean	13	125
	Sand and gravel, clay seams	10	135
	Limestone	5	140
WIL 35N11E-6.3h	Soil	20	20
	Till	24	44
	Sand and gravel	31	75
	Sand	25	100
	Till, sandy	5	105
	Gravel	10	115
	Till, gravelly	26	141
	Dolomite	19	160

Logs in table 6 illustrate the nature of the glacial drift east of Joliet. The fill in Hadley Valley contains a large proportion of sand and gravel. At places the lower part of the fill contains finer-grained material than the upper part. The sand and gravel is overlain at places by deposits of till (confining bed) that contain a high percentage of silt and clay. Till deposits are missing in many places in the present valley of Spring Creek and Hickory Creek. The glacial drift which nearly fills Hadley Valley may be Illinoian in age (Horberg and Potter, 1955). The map of the saturated thickness of sand and gravel (*see figure 6 in* Prickett et al., 1964) shows that the aquifer exceeds 100 feet in thickness in the vicinity and east of the municipal well field. Thicknesses exceeding 60 feet occur in a belt averaging ¾ mile in width. The sand and gravel deposits range in width from about ¾ to 3 miles and trend southwest to northeast. The logs of wells in table 6 illustrate the character of the glacial deposits in Hadley Valley.

Based on the results of aquifer and well-production tests, the average coefficients of transmissibility, premeability, and storage of the aquifer are 186,000 gpd/ft, 3100 gpd/sq ft, and 0.0015, respectively.

The mean annual precipitation is 34.25 inches; the mean annual temperature is 49.1 F.

The approximate piezometric surface map for the aquifer is shown in figure 11. The map was prepared

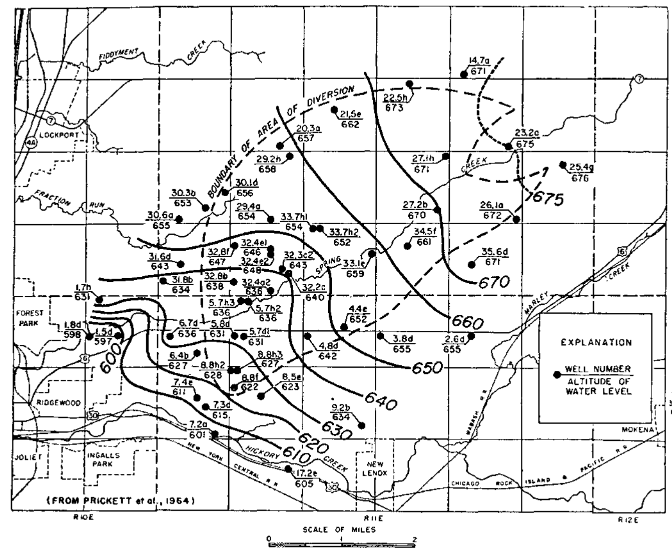


Figure 11. Piezometric surface of sand and gravel aquifer near Joliet, May 1962

from water-level measurements made on May 24 and 25, 1962, when withdrawals averaged 3.3 mgd. There are no well-defined cones of depression in the area. Pumping has considerably reduced natural discharge of ground water to Spring Creek and has warped contours upstream. Most of the ground water which under natural conditions discharged into Spring Creek, in the reach between gaging stations 4 and 6 (figure 12), is now diverted into production wells. Ground water still discharges into Spring Creek above gaging station 6 and below gaging station 3. The 650-foot contour has been displaced by pumping

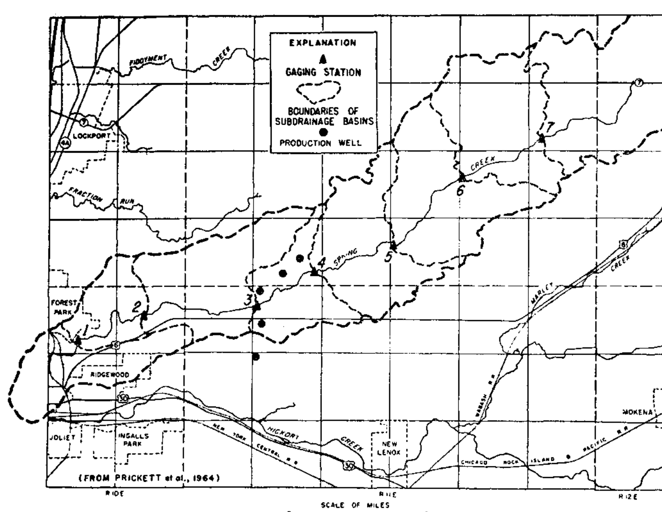


Figure 12. Gaging stations near Joliet

about 1 mile northeast of its original position. Withdrawals by the city have also moved the 640-foot contour approximately 1 mile from its estimated original position. Water levels near Sprng Creek between gaging stations 4 and 6 are probably a few feet below the bed of the creek.

Flow lines were drawn at right angles to the piezometric surface contours from production wells up gradient to define the area of diversion of production wells. As measured from figure 11, the area of diversion is 11 square miles. Ground-water movement outside the area of diversion is toward Spring Creek and other streams and scattered pumping centers beyond the Hadley Valley area.

The piezometric surface map was compared with water-level data for the period prior to development and water-level changes were computed. The greatest declines occurred in the immediate vicinity of the production wells and averaged about 8 feet. Water levels in the area of diversion of pumping and about 1 mile from production wells have declined on the average of about 2 feet. The average decline of water levels within the area of diversion is about 5 feet.

Streamflow records for a gaging station on Spring Creek, at the Benton Street Bridge in Joliet about 4 miles southwest of the municipal well field, are available for the period 1926 through 1933. The drainage area of Spring Creek above the station is 19.7 square miles. Ground-water runoff to Spring Creek was estimated with streamflow hydrograph separation methods outlined by Linsley, Kohler, and Paulhus (1958). Ground-water runoff during a year of near normal precipitation prior to development averaged about 5.78 inches of precipitation over the basin or about 275,000 gpd/sq mi, and was about 69 percent of streamflow and 18 percent of precipitation. Ground-water runoff averaged about 476,000 gpd/sq mi during 1927 when precipitation at Joliet was 47.49 inches and much above normal. On the basis of streamflow records for other drainage basins (see Schicht and Walton, 1961), it is probable that ground-water runoff during a year of much below normal precipitation averages about 135,000 gpd/sq mi.

The rate of recharge to the aquifer was estimated with the piezometric surface map and past records of pumpage and water levels. Comparisons of pumpage and water-level graphs indicate that in general water-level declines are directly proportional to pumping rates (*see figures 17 and 21 in* Prickett et al., 1964). Within a relatively short time after each increase in pumpage the area of diversion expanded in proportion to pumpage and water levels stabilized. Thus, recharge balanced discharge, and the average recharge rate to the aquifer is the quotient of the average pumping rate and the area of diversion. Computations with pumpage and water-level data show that the average recharge rate to the aquifer in 1961 was about 300,000 gpd/sq mi.

The aquifer is not recharged entirely by the percolation of precipitation to the water table. Recharge to the aquifer by induced infiltration of surface water occurs because the piezometric surface is below stream level and the streambed and surficial deposits in the flood plain of Spring Creek have some permeability. The streambed is only a few feet wide and is silted; streamflow during much of the time is low. Very little recharge from Spring Creek occurs during periods of low flow. However, at high flood stream stages the flood plain is inundated, and fairly large amounts of recharge by induced infiltration occur for short periods of time. The average rate of recharge from streamflow was computed as the difference between the average recharge rate to the aquifer and the amount of ground-water runoff diverted into cones of depression.

Ground-water runoff to Spring Creek under natural conditions was estimated to average about 275,000 gpd/sq mi earlier in this report. Not all ground-water runoff can be diverted into cones of depression because even under heavy pumping conditions there is some shallow lateral as well as deeper vertical movement of ground water in the surficial deposits. Precipitation, and therefore recharge, is unevenly distributed throughout the year, and there are periods of time during the wet spring months when recharge temporarily exceeds the rate of vertical movement of water. From studies in DuPage County (Zeizel et al., 1962) it is estimated that about 75 percent of ground-water runoff is diverted into existing cones of depression. The rate of recharge directly from precipitation is therefore about 200,000 gpd/sq mi based on the average ground-water runoff and the 75 percent factor. The average rate of recharge from streamflow was computed to be about 100,000 gpd/sq mi by subtracting the rate of recharge directly from precipitation from the average rate of recharge to the aquifer.

Ground-water runoff was estimated to average about 135,000 gpd/sq mi during years of much below normal precipitation earlier in this report. It is probable that recharge from streamflow will be less than 100,000 gpd/sq mi during dry periods. On the basis of the ratio of ground-water runoff during years of normal precipitation (275,000 gpd/sq mi) and ground-water runoff during dry years (135,000 gpd/sq mi) it is probable that recharge from streamflow may average about 50,000 gpd/sq mi during dry years. It is estimated that the average rate of the recharge to the aquifer during dry periods is about 185,000 gpd/sq mi.

Recharge from the flow in Spring Creek was determined along two reaches of the stream by measuring

Table 7. Streamflow Measurements for Spring Creek near Joliet

Gaging station number	Discharge (*cfs*)	Discharge (*mgd*)	Loss of flow between stations (*mgd*)	Percent of flow infiltrated	Infiltration rate (*gpd/acre*)	Average depth of water in stream (*ft*)
1	5.16	3.59				0.65
2	3.47	2.41				0.70
3	1.30	0.90				0.73
4	1.11	0.77	0.07	8	58,000	1.02
5	1.21	0.84				0.35
6	1.67	1.16	0.32	29	175,000	0.60
7	0.75	0.52				1.05

(*from Prickett et al.*, 1964)

the water lost between successive gaging stations. Discharge measurements were made on April 26, 1962, at the stations listed in table 7 and shown in figure 12. The infiltration rates were measured during a period of low streamflow. They are probably far greater when the flow in Spring Creek is high.

Sand and Gravel Aquifer at Champaign-Urbana

Ground-water supplies at Champaign-Urbana are developed from wells in deeply buried sand and gravel aquifers in the Mahomet buried bedrock valley, which extends across the central part of Illinois from the Indiana border to the Illinois River Valley. The normal annual precipitation at Champaign-Urbana is 36.43 inches; average annual temperature is 51.9 F.

For a detailed discussion of the geology of the Champaign-Urbana area, the reader is referred to Foster and Buhle (1951) and Selkregg and Kempton (1958). The following geologic description is based largely upon these reports. The Mahomet buried bedrock valley averages about 12 miles in the area and is largely filled with glacial drift ranging in thickness from 50 to 440 feet. The glacial drift is composed chiefly of pebbly, silty till and deposits of glaciofluvial sand and gravel that have various areal and cross-sectional patterns as shown in figure 13.

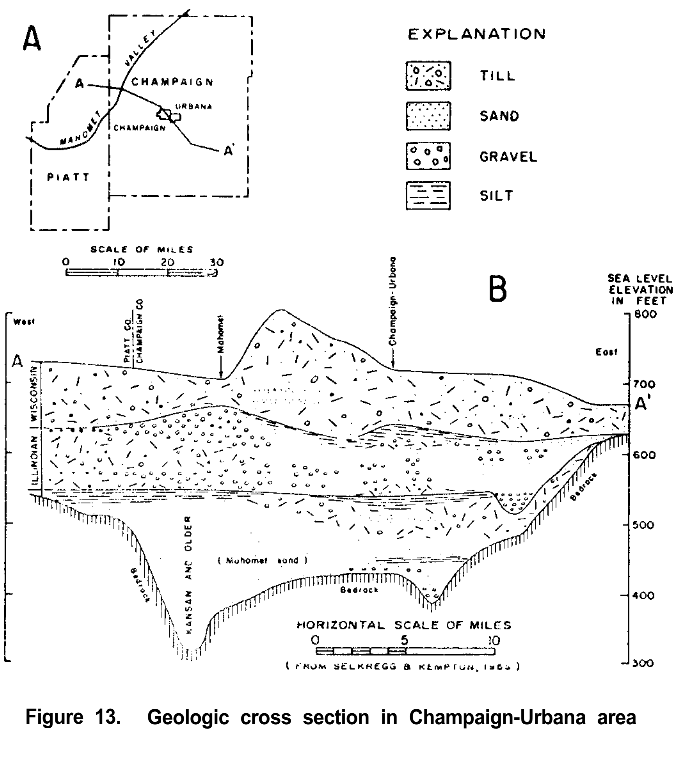


Figure 13. Geologic cross section in Champaign-Urbana area

The bedrock directly underlying the drift is composed mainly of Pennsylvanian shale with thin beds of limestone, sandstone, and coal. The bedrock surface in the area has a maximum relief of over 300 feet as shown in figure 14. The channel of the deeply entrenched Mahomet buried valley lies about 9 miles west of the corporate limits of Champaign. A bedrock tributary valley trends west and lies south of Champaign-Urbana.

Sand and gravel are encountered within the glacial drift at depths between 60 and 120 feet (upper aquifer), 140 and 170 feet (middle aquifer), and below a depth of 200 feet (lower aquifer). The upper Wisconsinan aquifer is thin, discontinuous, scattered, and lenticular in nature, whereas the other two aquifers have fairly large areal extents. The middle Illinoian aquifer ranges in thickness from less than 20 feet to more than 60 feet as shown in figure 15 and has an average thickness of about 43 feet in the immediate vicinity of Champaign-Urbana. The middle aquifer is overlain in most places by a confining bed (upper) consisting largely of clayey silt with varying amounts of sand. The thickness of the confining bed ranges from more than 150 feet to less than 50 feet and averages about 120 feet in the immediate vicinity of Champaign-Urbana as shown in figure 16. The upper aquifer is intercalated in the confining bed at places. The lower Kansan aquifer, partially filling the deep channel of the Mahomet buried bedrock valley, often exceeds 100 feet in thickness west of Champaign-Urbana. Except in local areas basal Illinoian till, typically composed of pebbly silt with a varying amount of clay, separates the middle and lower aquifers. The lower confining bed averages about 30 feet thick. Complex facie changes and interfingering silt is typical of the aquifers. The log in figure 17 illustrates the character of the glacial drift at Champaign-Urbana.

Recharge to aquifers at Champaign-Urbana occurs as vertical leakage of water through overlying confining beds. Quantities of leakage through confining beds vary from place to place, and are primarily controlled by vertical permeabilities and thicknesses of confining beds and by the differences between the heads in aquifers and in shallower deposits.

Prior to 1947, ground-water supplies for municipal, commercial, industrial, and university use at Champaign-Urbana were obtained from wells in the middle aquifer. Pumpage was concentrated in three major pumping centers as shown in figure 15. A new municipal well field was developed west of town in the lower aquifer in 1947 and pumpage from the middle aquifer was considerably reduced. Average daily ground-water withdrawals from the middle aquifer 1900 to 1961 are shown in figure 18. As the result of heavy concentrated pumping, nonpumping water levels declined more than 50 feet at places as shown in figure 19. Pumping levels in the municipal well field declined to critical levels in 1947. As the result of a reduction in pumpage from 6.3 to 2.8 mgd, water levels recovered as much as 25 feet at places during the period 1947 to 1961. A comparison of the pumpage and water-level hydrographs in figures 18 and 19 shows that water-level decline is directly proportional to the pumping rate.

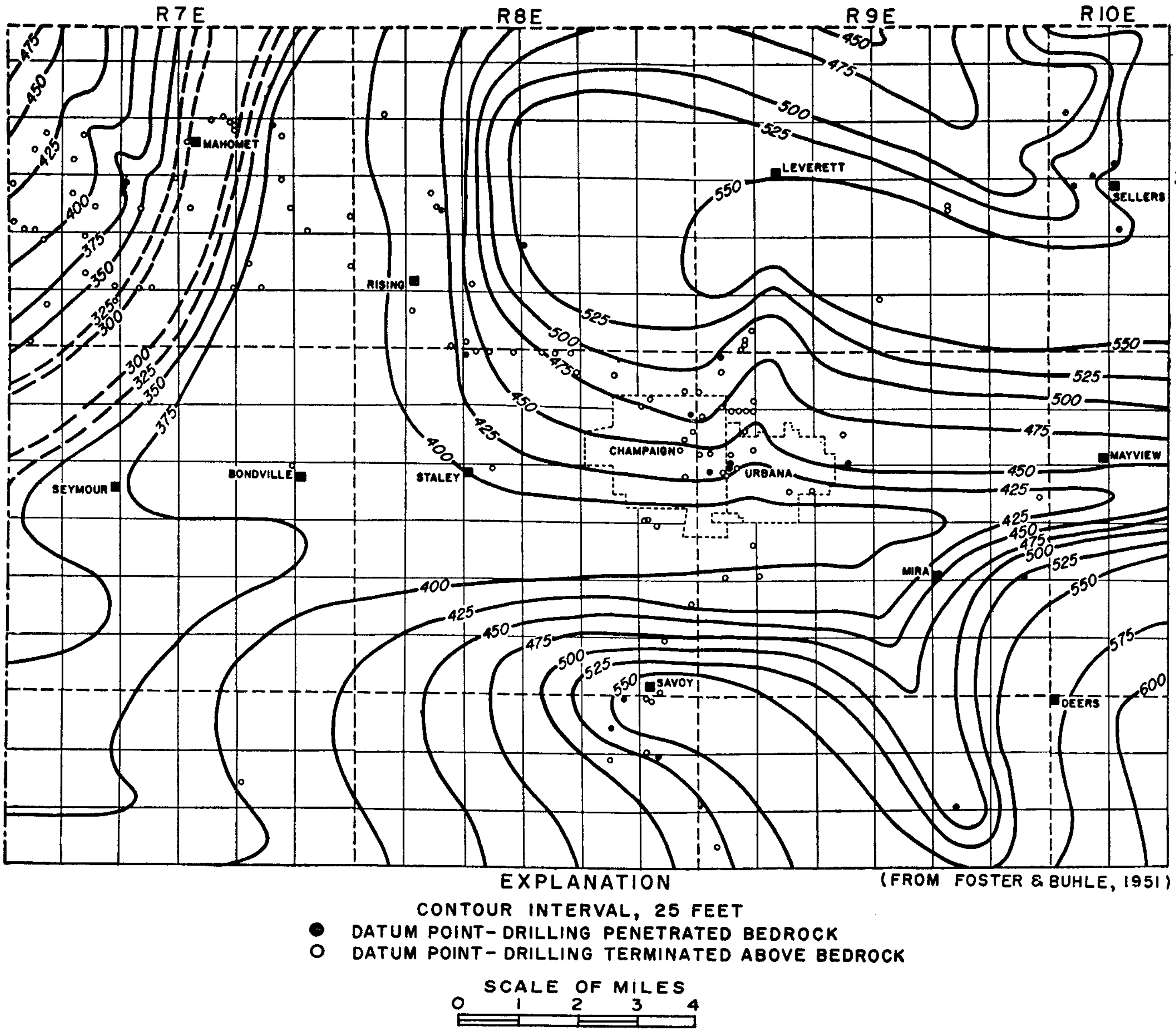


Figure 14. Bedrock topography in Champaign-Urbana area

Within a relatively short time after each increase in pumping rate, leakage through the upper confining bed increased in proportion to pumpage and balanced discharge.

Several aquifer tests were made to determine the hydraulic properties of the middle aquifer. The coefficients of permeability and transmissibility vary from place to place and average 850 gpd/sq ft and 37,000 gpd/ft, respectively. The average coefficient of storage is 0.00024 (Smith, 1950). The coefficient of vertical permeability of the upper confining bed is very low, and the effects of vertical leakage were not measurable during the aquifer tests. The vertical permeability was determined with a model aquifer (see Walton and Walker, 1961) and past records of pumpage and water levels.

The results of geologic and hydrologic studies indicate that it is possible to simulate complex middle aquifer conditions with an idealized model aquifer. The model aquifer is a layer of sand and gravel extending beyond cones of depression, averaging 43 feet thick and overlain by a confining bed averaging 120 feet thick. The coefficients of transmissibility and storage of the model aquifer are 37,000 gpd/ft and 0.00024, respectively. The coefficient of vertical permeability of the confining bed is low, based on well logs and drilling experiences.

The water-level declines in the wells in figure 19 from 1900 to 1947 were computed by using the model aquifer, the computed hydraulic properties, the estimated pumpage data, the steady state leaky artesian equation (Jacob, 1946), and several assumed values of the coefficient of permeability of the confining bed. The computed declines were then compared with actual declines. The coefficient of vertical permeability used to compute declines equal to actual declines was assigned to the aquifer.

Ground-water withdrawals were grouped into three centers of pumping. The centers of pumping and observation wells were located on a map, and the distances between them were scaled from the map. Distance-drawdown graphs were prepared with computed and assumed hydraulic properties of the model aquifer. The effects of production wells on water levels in the observation wells

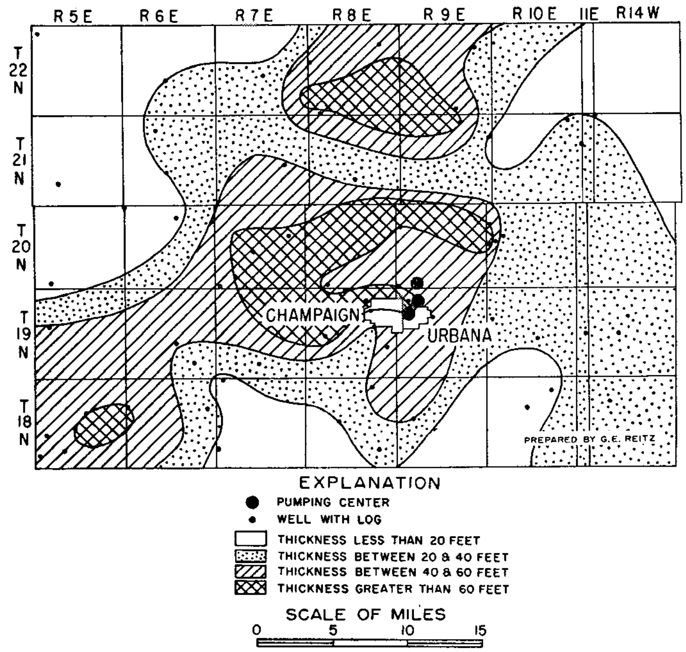


Figure 15. Thickness of middlse sand and gravel aquifer in Champaign-Urbana area

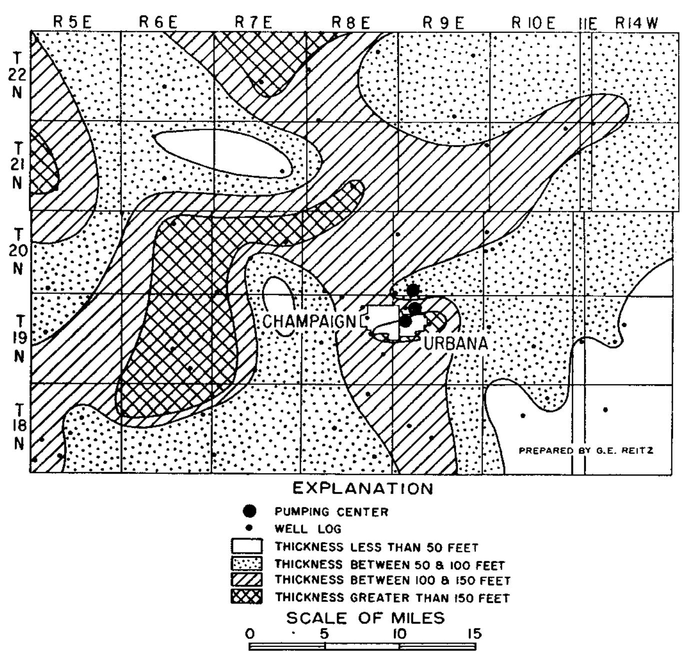


Figure 16. Tickness of upper confining bed in Champaign-Urbana area

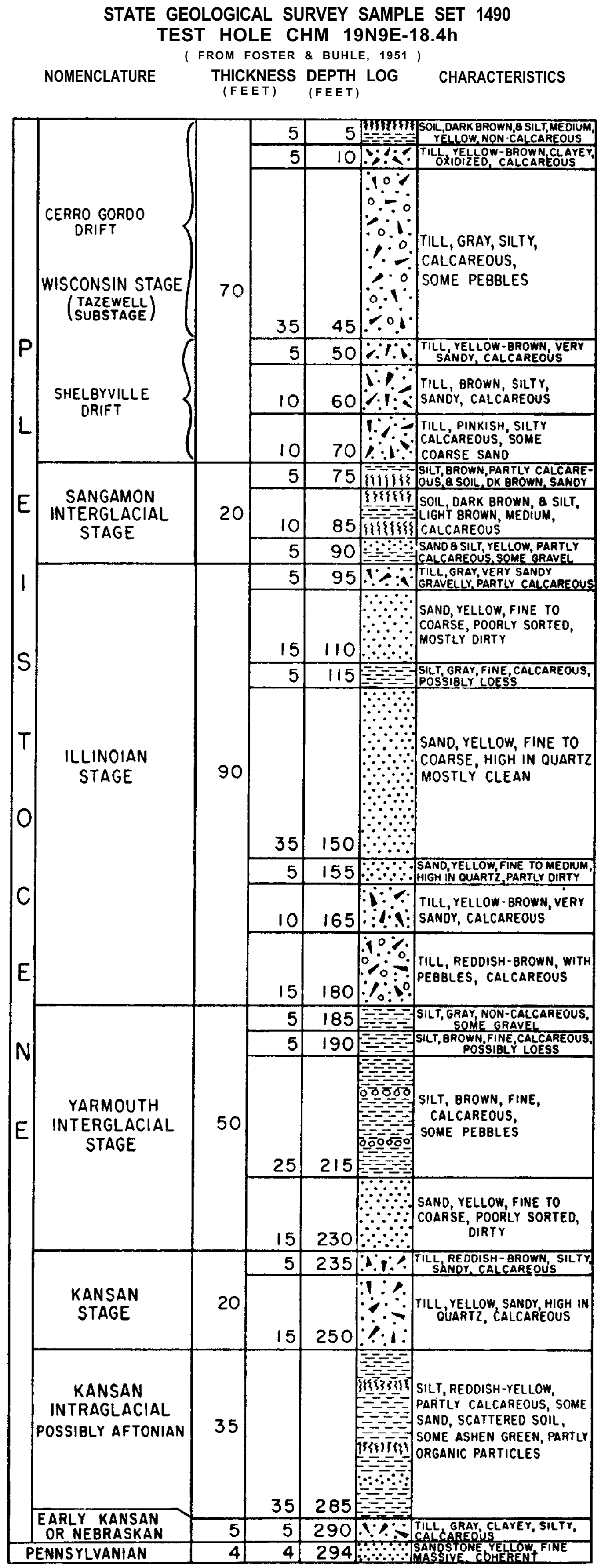


Figure 17. Log of well at champaign-Urbana

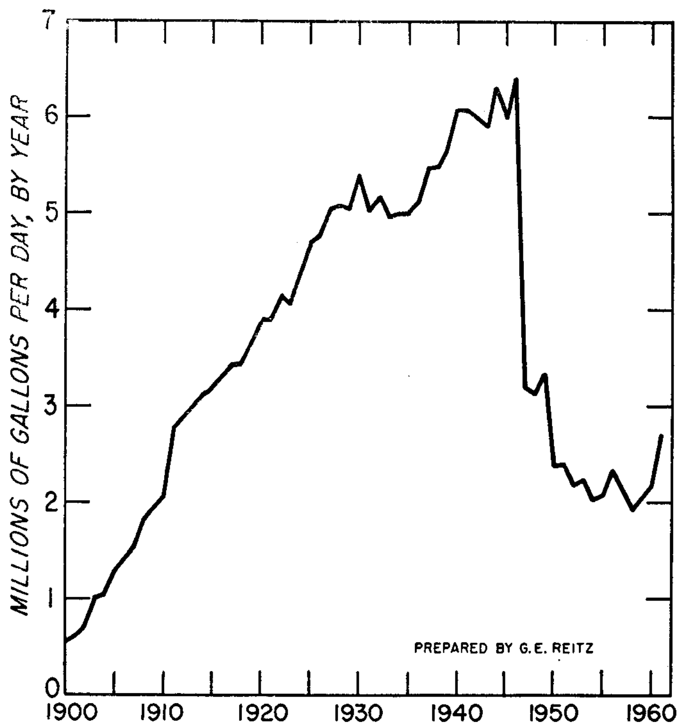


Figure 18. Pumpage from middle sand and gravel aquifer in Champaign-Urbana area, 1900-1961

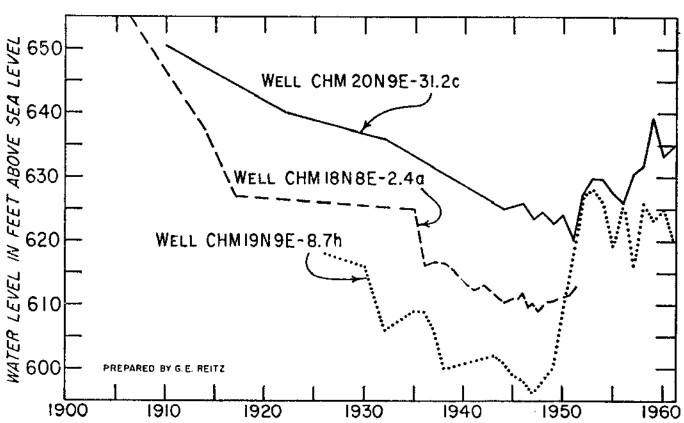


Figure 19. Ground-water levels in middle sand and gravel aquifer in Champaign-Urbana area, 1907-1961

were determined with the distance-drawdown graphs and pumpage data in figure 18. Water-level declines based on a coefficient of vertical permeability of 0.01 gpd/sq ft compare favorably with actual declines. The coefficient of vertical permeability of the confining bed is, therefore, estimated to be 0.01 gpd/sq ft.

The average head loss associated with vertical leakage was estimated to be about 50 feet, based on the distance-drawdown graph for the model aquifer and confining bed and on data of water levels in shallow and deep deposits measured prior to development. The estimated average head loss, computed coefficient of vertical permeability of the confining bed, saturated thickness of the confining bed, and estimated pumpage data were substituted in the following equation to determine the area of diversion associated with recharge:

$$A_c = Q_c\ m'/(P'\ \ h) \qquad (3)$$

where:

A_c = area of confining bed through which recharge occurs, in sq ft
Q_c = leakage through confining bed which is equal to pumpage, in gpd
m' = thickness of confining bed through which leakage occurs, in ft
P' = coefficient of vertical permeability, in gpd/sq ft
h = average differential between the head in the aquifer and in the source bed above the confining bed, in ft

The area of diversion in 1947 was about 55 square miles. The recharge rate for the middle aquifer, computed as the quotient of pumpage and area of diversion, was about 115,000 gpd/sq mi in 1947.

Sand and Gravel Aquifer in Havana Region

The Havana region in west-central Illinois covers about 720 square miles mostly in Mason County. The area is bounded on the west by the Illinois River; on the east by ground-water divides which roughly trend east-northeast in the vicinity of Mason City, San Jose, and Delavan; on the north by Pekin; and on the south by the Sangamon River. The area lies between 40°00' and 40°35' north latitude and 89°30' and 90°30' west longitude. Principal municipalities within the area are Delavan, Havana, Mason City, Manito, Kilbourne, San Jose, Easton, Forest City, Green Valley, and South Pekin. The area is primarily a wide, low rolling sandy plain east of the Illinois River, bordered to the east by glaciated uplands. The normal annual precipitation is 35.18 inches; average annual temperature is 51 F.

For a detailed discussion of the geology of the area, the reader is referred to Selkregg and Kempton (1958) and to Walker et al. (1965). The following geologic description is based largely upon these two reports. The area is a wide bedrock lowland at the confluence of the ancient Mississippi and Mahomet Rivers now buried beneath a thick mantle of glacial drift (*see figure 10 in* Walker et al., 1965). Glacial deposits, mainly sand and gravel, include ancient stream fills and outwash and exceed 100 feet in thickness in most of the area. Mississippian and Pennsylvanian bedrock formations, consisting mostly of shale and limestone, underlie the glacial deposits.

The bedrock channel is generally below an elevation of about 400 feet, whereas the adjoining bedrock upland has elevations ranging from 500 to more than 600 feet. The glacial deposits range in thickness from a few feet to more than 400 feet. The deposits are commonly from 125 to 150 feet thick above bedrock benches and exceed 200 feet thick above bedrock channels. Throughout much of the Havana region the upper part of the deposits is composed of sand and gravel and the lower part is mainly sand. In upland areas, such as at Mason City, the sand and gravel deposits are overlain by glacial till. The sequences of deposits described by logs in table 8 are considered typical of the glacial deposits in lowland and upland areas.

Table 8. Logs of Selected Wells in Havana Region

Formation	Thickness *(ft)*	Depth *(ft)*
Well located 1.5 *miles south of Forest City in sec.* 19, *T22N, R6W (representative of the glacial deposits in lowland areas)*		
Sand, fine to medium, some coarse sand at base	50	50
Sand, fine to medium	40	40
Sand, fine to medium, silty	30	120
Sand, fine to medium, some granular gravel, very silty	10	130
Sand, fine to medium	10	140
Shale, dark gray	10	150
Well located at Mason City in sec. 8, *T20N, R5W (representative of the glacial deposits in upland areas)*		
Soil	5	5
Sand, fine	35	40
Silt	5	45
Till, silty, brown	5	50
Sand, medium to coarse	5	55
Sand, very coarse, some gravel, dirty	10	65
Till, yellowish, brown	20	85
Sand, fine to medium	5	90
Sample missing	105	195
Sand, very fine to fine	4	199
Sand, very coarse, some gravel	11	210
Sand, medium to coarse	10	220

(from Selkregg and Kempton, 1958 *)*

Pre-Illinoian Sankoty Sand, ranging from fine sand to very coarse sand with granule gravel, overlies the bedrock throughout much of the area. Its thickness is generally less than 125 feet. Illinoian deposits, consisting of till, silt, and sand and gravel, overlie the Sankoty Sand in upland areas to the east. In much of the area the Sankoty Sand is overlain by Wisconsinan outwash up to 100 feet thick. The outwash deposits consist largely of sandy gravel and are the coarsest and most permeable sediments of the area. Wisconsinan dune sand mantles the outwash in extensive areas and commonly exceeds 20 feet thick. Upland areas are mantled with Wisconsinan loess and silt generally more than 5 feet thick and often more than 30 feet thick. Wisconsinan drift mantled by loess cover Illinoian and older deposits in upland areas to the east. Here, the Wisconsinan drift is mainly composed of clayey till, with little sand and gravel, and is more than 100 feet thick at places. Floodplains in the area are generally floored with clay, silt, and sand deposited by flood waters and by slope wash from loess covered uplands.

Recharge conditions are most favorable in lowland areas where the Sankoty Sand is covered by Wisconsinan outwash and dune sands. Recharge conditions are less favorable in upland areas where Wisconsinan or Illinoian tills, or both, are present. Till and silt beds of low permeability retard the vertical movement of water.

Based on aquifer and well-production tests and flow-net analyses, the coefficient of permeability of the aquifer ranges between 7500 and 15,000 gpd/sq ft in the northern part of the area, and between 1600 and 2200 gpd/sq ft in the eastern part of the area. The coefficient of transmissibility ranges between 240,000 and 500,000 gpd/ft in the north to about 300,000 gpd/ft in the east. In areas of thick Wisconsinan outwash where the coefficient of permeability ranges from 4000 to 7500 gpd/sq ft, the coefficient of transmissibility ranges from 200,000 to 700,000 gpd/ft.

In lowland areas municipal and irrigation wells are finished at depths ranging from 40 to 120 feet; in upland areas such as at Mason City wells are finished below a depth of 200 feet. Total withdrawal from the sand and gravel aquifer in 1960 for municipal, industrial, rural, and irrigation water supplies was about 3.2 mgd. The potential yield of the aquifer greatly exceeds present pumpage; the Havana region probably contains the largest undeveloped aquifer in Illinois.

A water-table map for the area is shown in figure 20. The map was prepared from water-level measurements made in 103 observation wells. Movement of water is generally in northwesterly and southwesterly directions toward the Illinois and Sangamon Rivers and other small streams and ditches.

Table 9. Log of a Well in East St. Louis Area

Formation	Thickness *(ft)*	Depth *(ft)*
Pleistocene Series		
Recent and older alluvium		
Soil, clay, and silt, dark gray	10	10
Sand, fine to coarse, subangular grains, abundant feldspar, tiny calcareous spicules, coal	30	40
Sand, medium, with granule gravel, as above, mollusk shell fragments	10	50
Sand, fine, with granule gravel, poor sorting, calcareous spicules, abundant dark grains of igneous rocks, ferromagnesium minerals, and coal	10	60
Gravel, granule size, with coarse sand, granules, mainly igneous rocks and feldspar	10	70
No samples	10	80
Sand, medium to fine, calcareous spicules, subangular grains, coal	10	90
No samples	5	95
Sand, very coarse to coarse, with granule gravel, pinkish cast, abundant pinkstained quartz grains, subangular to subrounded grains	15	110
Sand, medium, well sorted, pink, subrounded to subangular grains, abundant pink feldspar	5	115

(from Bergstrom and Walker, 1956)

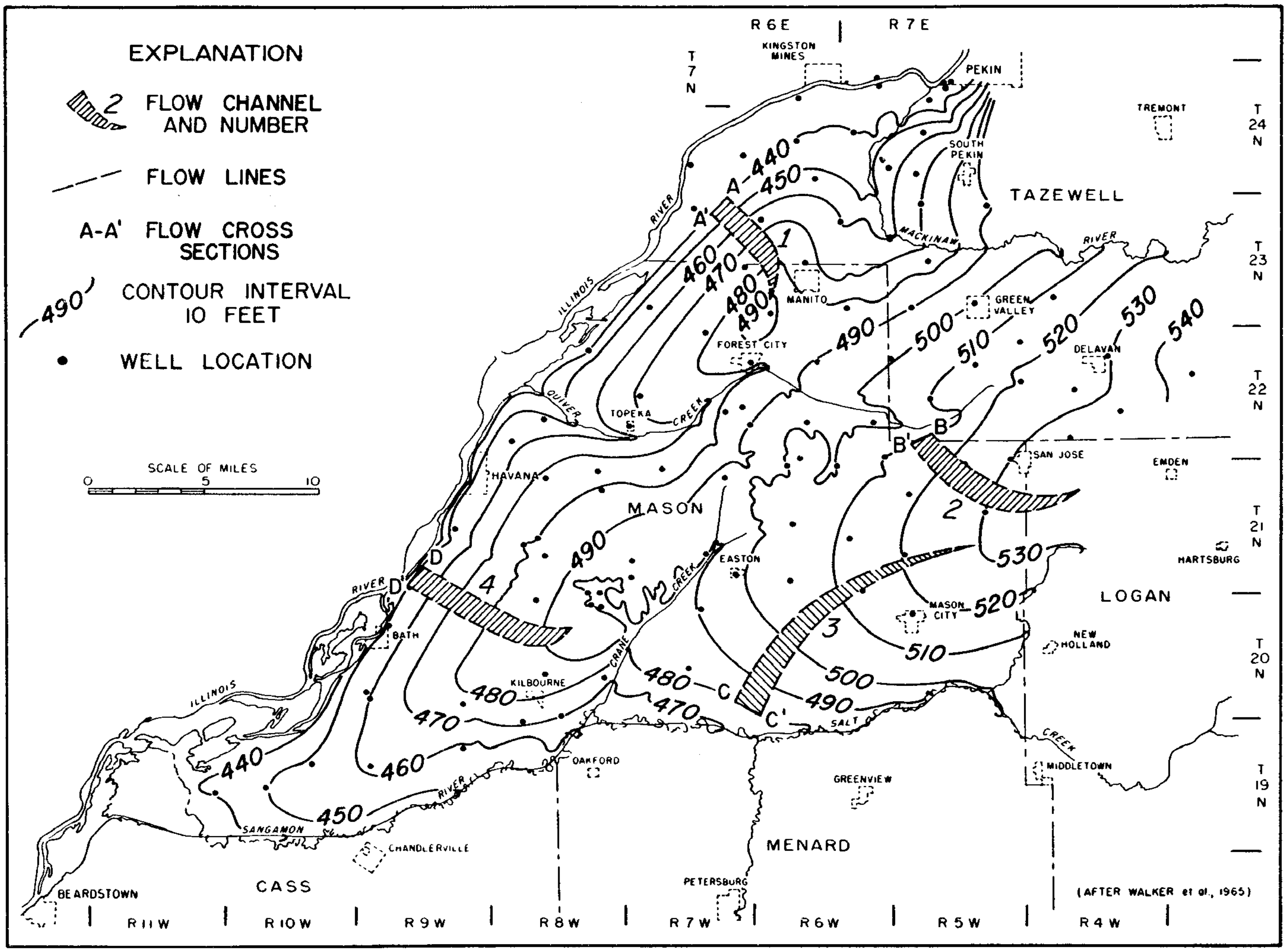


Figure 20. Water table of sand and gravel aquifer in Havana region, September 1960, and locations of flow channels

Flow lines were drawn at right angles to water-table contours down gradient from ground-water mounds and ridges to delineate the four flow channels shown in figure 20. Equation 1, coefficients of transmissibility, and the water-table map were used to estimate the amounts of water percolating through cross sections A—A', B—B', C—C', and D—D'. The flow through these sections is 1.70, 2.15, 3.20, and 2.04 mgd, respectively. Recharge rates, computed as the quotients of flow through sections and areas of flow channels, average about 270,000 gpd/sq mi in flow channels 1 and 2 and about 490,000 gpd/sq mi in flow channels 3 and 4. Flow channels 1 and 2 lie in areas where layers of till overlie the aquifer and retard the vertical movement of water. Flow channels 3 and 4 lie in areas where fairly coarse grained sand and gravel deposits occur from the surface down to bedrock.

Sand and Gravel Aquifer in East St. Louis Area

The East St. Louis area is in southwestern Illinois and includes portions of Madison, St. Clair, and Monroe Counties (Schicht and Jones, 1962). It encompasses the major cities of East St. Louis, Granite City, and Wood River, and extends along the valley lowlands of the Mississippi River from the city of Alton south to the village of Dupo. The area covers about 175 square miles and is approximately 30 miles long and 11 miles wide at the widest point. The normal annual precipitation is about 38 inches; average annual temperature is 56.4 F.

For a detailed discussion of the geology of the area, the reader is referred to Bergstrom and Walker (1956). The following geologic description is based largely upon this report. Unconsolidated valley fill in the area is composed of recent alluvium and glacial valley-train material and is underlain by Mississippian and Pennsylvanian rocks consisting of limestone and dolomite with subordinate amounts of sandstone and shale. The valley fill has an average thickness of 120 feet and ranges in thickness from a feathers edge, near the bluff boundaries of the area and along the reach of the Mississippi River known as the "Chain of Rocks," to more than 170 feet near the city of Wood River (*see figure* 2 *in* Schicht and

Jones, 1962). The thickness of the valley fill exceeds 120 feet and is generally greatest in places near the center of a buried bedrock valley that bisects the area.

Recent alluvium comprises the major portion of the valley fill in most of the area. The alluvium is composed of fine-grained materials with a low permeability; the grain size increases from the surface down. Recent alluvium rests on older deposits including in many places glacial valley-train materials. The valley-train deposits are predominantly medium-to-coarse sand and gravel and increase in grain size with depth. The coarsest deposits are commonly encountered near bedrock and often average 30 to 40 feet in thickness. Logs of wells (*see figure 3 in* Schicht and Jones, 1962) show that the valley fill grades from clay to silt to sand and gravel interbedded with layers of silt and clay with increasing depth. The log in table 9 for a well at Granite City located in T3N, R10W is typical of many wells in the East St. Louis area.

Ground water in the valley fill occurs under leaky artesian and water-table conditions. Leaky artesian conditions exist at places where fine-grained alluvium overlies valley-train deposits and water in the valley-train deposits is under artesian pressure. Water-table conditions prevail where alluvium is missing and the upper surface of the zone of saturation is in valley-train deposits, and at places within deep cones of depression where water is unconfined. Water occurs most commonly under leaky artesian conditions, and the surface to which water rises in wells is called the piezometric surface.

Recharge is from precipitation within the area, induced infiltration of surface water of the Mississippi River and small streams and ditches traversing the area, and subsurface flow from the bluffs bordering the area.

A map showing estimated coefficients of transmissibility of the aquifer in the East St. Louis area is shown in figure 21. The map is based on aquifer and well-production test data, flow-net analyses, and data on the saturated thickness of the aquifer. The coefficient of transmissibility commonly exceeds 150,000 gpd/ft and exceeds 200,000 gpd/ft in the Monsanto, East St. Louis, National City, and Granite City areas. Coefficients of permeability range from 3000 to 1000 gpd/sq ft and decrease rapidly near the bluffs bordering the area.

Large quantities of ground water are withdrawn from sand and gravel wells concentrated in five major pumping centers: the Alton, Wood River, Granite City, National City, and Monsanto areas. Pumpage mostly for industrial use increased from 2.1 mgd in 1900 to 111.0 in 1956 and was 93.0 mgd in 1960 (Schicht and Jones, 1962). As a result of heavy pumping, water levels declined about 50 feet in the Monsanto area and more than 10 feet in other pumping centers.

A piezometric surface map for the East St. Louis area is shown in figure 22. The map was prepared from water-level measurements made in 225 wells in November 1961. The general pattern of flow of water in 1961

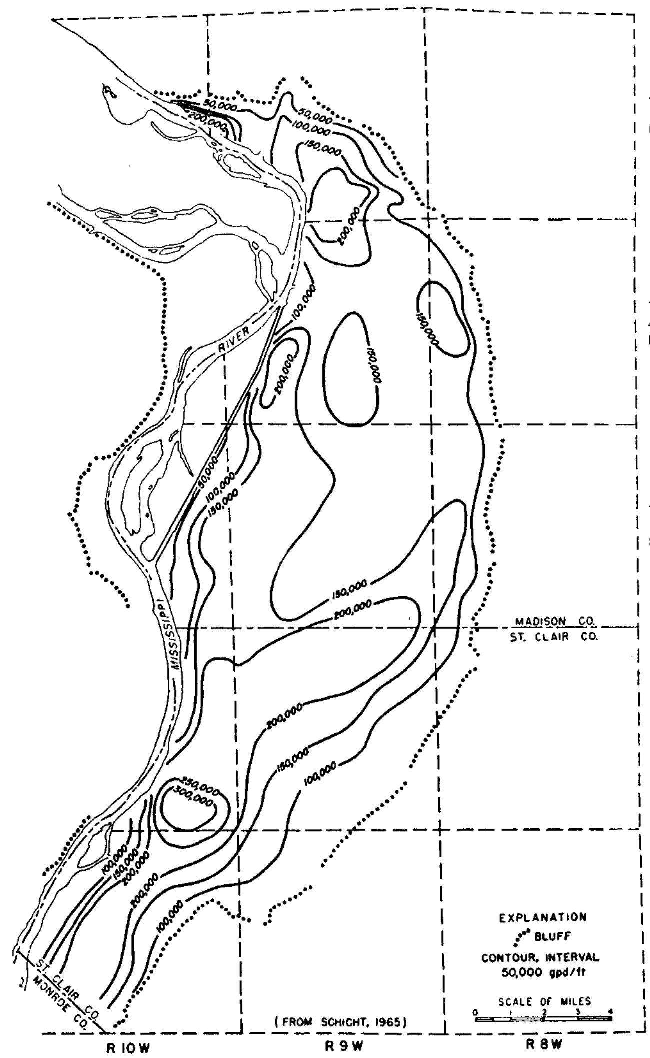


Figure 21. Coefficient of transmissibility of sand and gravel aquifer in East St. Louis area

was slow movement from all directions toward cones of depression in pumping centers or streams and lakes. Pumping of wells and draining of lowlands have considerably reduced ground-water discharge to the Mississippi River, but has not reversed at all places the natural slope of the piezometric surface toward that stream. Ground-water levels were below the river at places, and appreciable quantities of water were diverted from the river into the aquifer by the process of induced infiltration.

Flow lines were drawn at right angles to the piezometric surface contours from pumping centers up the hydraulic gradient on the land sides of pumping centers at Wood River, Granite City, National City, and Monsanto to define flow channels 1 through 4 on figure 22. Equation 2 and the data in table 10 were used to compute recharge rates. The quantity ($Q_2 - Q_1$) is the difference in the amount of water entering and leaving flow channels. Coefficients of transmissibilities were obtained from figure 21, hydraulic gradients were computed with the

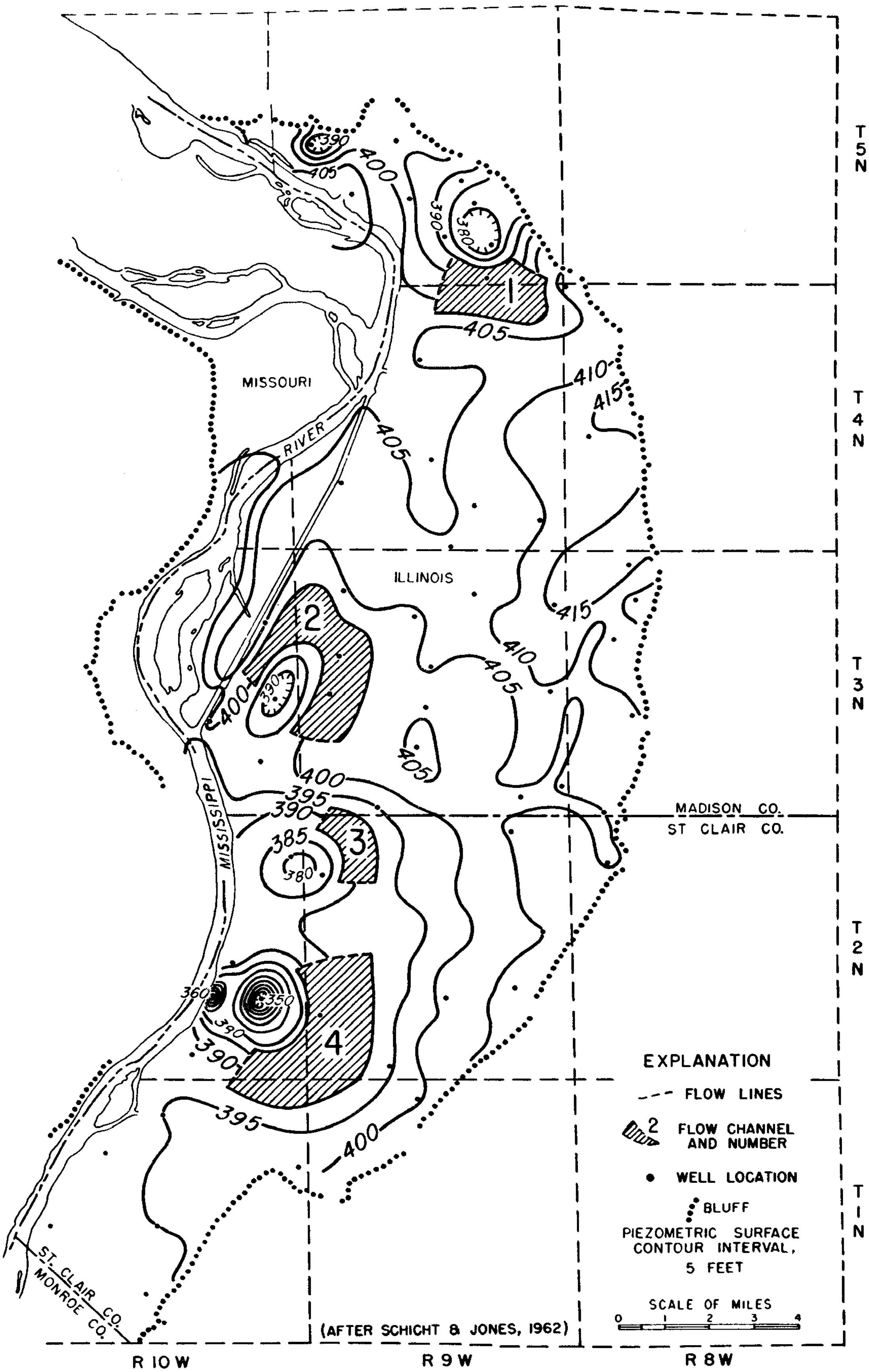


Figure 22. Piezometric surface of sand and gravel aquifer in East St. Louis area, November 1961, and locations of flow channels

Table 10. Flow-Net Analysis Data for East St. Louis Area

Flow channel	Q_1 (gpd)	Q_2 (gpd)	h_t (ft /day)	A_l (sq mi)	Coefficient of storage (fraction)	Recharge rate (gpd /sq mi)
1	3.46×10^6	4.18×10^6	+.011	3.34	0.15	347,000
2	4.74×10^6	6.25×10^6	neg*	4.40	—	343,000
3	2.18×10^6	2.40×10^6	+.0056	1.20	0.10	299,000
4	5.30×10^6	9.10×10^6	—.015	5.70	0.10	370,000

*negligible

(after Schicht, 1965*)*

piezometric surface map, and water-level declines were estimated from hydrographs of observation wells in the area. As shown in table 10, recharge rates do not vary greatly from flow channel to flow channel and average about 340,000 gpd/sq mi. Recharge from the subsurface flow of water through bluffs into the valley fill was computed to average about 329,000 gpd/mi of bluff, from study of the movement of water through four flow channels near the bluffs.

Three Small Watersheds in Central Illinois

To determine how ground-water recharge varies from year to year, month to month, and basin to basin, hydrologic and ground-water budgets were prepared for three small watersheds in central Illinois (Schicht and Walton, 1961). The *hydrologic budget* is a quantitative statement of the balance between total water gains and losses of a basin; the *ground-water budget* is a quantitative statement of the balance between water gains and losses of the ground-water reservoir.

During study periods ranging from 2.5 to 8 years the State Water Survey and cooperating state and federal agencies measured precipitation on, stream discharge from, and ground-water levels in parts of the drainage basins of Panther, Hadley, and Goose Creeks. The study areas are in north-central west-southwestern, and east-central Illinois, respectively. Information pertaining to Panther Creek Basin is presented in detail to illustrate the influence of climatic conditions on ground-water recharge. Data for the other two basins are summarized and compared with data for Panther Creek Basin to show the influence of basin characteristics on ground-water recharge.

Location and general features. The Panther Creek Basin is about 30 miles east of Peoria and about 20 miles north of Bloomington. The part of Panther Creek drainage basin considered, hereafter referred to as "the basin," is approximately between 88°52´ and 89°07´ west longitude and between 40°44´ and 40°54´ north latitude. The basin covers 95 square miles mostly in Woodford County, although small parts are in Livingston and McLean Counties, and is in T26N to T28N and R1E to R3E (*see figure 2 in* Schicht and Walton, 1961). The basin is above a stream gaging station about 4 miles northwest of the city of El Paso.

The basin lies in the Till Plains Section of the Central Lowland Physiographic Province (Fenneman, 1914). The topography consists mostly of gently undulating uplands. Rolling topography is found in belts on moraines along the west, northeast, and east edges of the basin. The uplands are eroded in the immediate vicinity of Panther Creek in the extreme southwest corner of the area where the topography is more diversified. The elevation of the land surface of the basin declines from 770 feet near Benson and Gridley to 660 feet at the stream gaging station northwest of El Paso. Except in the southwest corner adjacent to Panther Creek where the elevation of the land surface declines about 50 feet in a distance of ¼ mile, the relief seldom exceeds 20 ft/mi.

Panther Creek is the principal stream, and flows in a generally southwestward course. A small tributary, East Branch Panther Creek, drains the southern quarter of the basin and flows westward to a confluence with Panther Creek 4 miles northwest of El Paso. The average gradients of Panther Creek and East Branch are 4.7 and 5.0 ft/mi, respectively. The water table was very near the surface and shallow ponds, swamps, and poorly drained areas were widespread prior to settlement. Extensive surface and subsurface drainage was necessary to permit agricultural development.

The population of the basin is chiefly rural and, according to the U. S. Bureau of the Census, had a density of about 37 persons per square mile in 1950. The populations of incorporated municipalities within or bordering the basin are as follows: Benson, 387; El Paso, 1818; Gridley, 817; Minonk, 1955; and Panola, 52.

At the time of this study about 80 percent of the basin was cleared and cultivated; the remainder was pasture, woodland, and farm lots. The cleared land was devoted to three major crops, field corn, oats, and soybeans for beans, and to other crops such as alfalfa, clover and timothy hay, winter wheat, rye, and sweet corn.

Climate. The basin lies in the north temperate zone. Its climate is characterized by warm summers and moderately cold winters. The mean length of the growing season is about 170 days. Based on records collected by the U. S. Weather Bureau at Minonk, the mean annual temperature is 51 F. June, July, and August are the hottest months with mean temperatures of 71, 76, and 73 F, respectively; January is the coldest month with a mean temperature of 25 F. Mean monthly temperatures during December, January, and February are below 32 F. Normal annual precipitation is 33.6 inches, based on 1900 to 1944 U. S. Weather Bureau records at Minonk and Gridley. The months of greatest precipitation are April, May, June, August, and September, each having an average of more than 3 inches. December, January, and February are the months of least precipitation, each having an average of less than 2 inches. Monthly and annual precipitation, 1950-1958, is given in table 11. Precipitation was above, near, and below normal in 1951, 1952, and 1956, respectively. The annual maximum precipitation

Table 11. Monthly and Annual Precipitation in Inches, 1950-1958, Panther Creek Basin

Month	1950	1951	1952	1953	1954	1955	1956	1957	1958
Jan	4.90	1.41	1.01	1.36	1.23	1.92	0.14	1.51	1.02
Feb	2.71	2.88	1.19	1.19	2.11	1.50	1.45	1.16	0.45
Mar	1.13	3.58	2.73	4.38	3.95	1.55	0.73	1.64	0.33
Apr	5.99	4.20	4.66	1.94	4.46	4.28	2.39	7.47	2.56
May	1.07	2.93	3.36	2.06	4.58	3.53	3.24	4.42	2.57
Jun	6.91	7.16	7.07	3.52	2.58	2.81	0.89	4.64	5.67
Jul	6.42	8.40	2.18	6.29	4.42	3.12	3.22	2.28	6.05
Aug	0.62	4.11	4.47	1.22	5.18	4.33	3.23	1.96	4.24
Sep	3.83	2.34	1.43	2.32	0.81	1.86	1.08	1.31	1.82
Oct	0.90	2.99	0.64	0.71	3.42	3.71	0.40	5.14	0.64
Nov	1.81	2.70	2.31	0.72	1.75	0.83	1.54	2.08	2.62
Dec	0.78	1.54	1.57	2.53	1.61	0.35	1.18	2.75	0.49
Annual	37.07	44.24	32.62	28.24	36.10	29.79	19.49	36.36	28.46

(from Schicht and Walton, 1961*)*

amounts occurring on an average of once in 5 and once in 50 years are 39 and 50 inches, respectively; annual minimum amounts expected for the same intervals are 30 and 24 inches, respectively. Amounts are based on data given in the Atlas of Illinois Resources, Section 1 (1958). The mean annual snowfall is 24 inches. On the average more than 28 days have 1 inch or more, and more than 13 days have 3 inches or more, of ground snow cover in a year. The average depth of maximum frost penetration is 26 inches.

Geology. The soils of the basin were divided into four groups by Smith et al. (1927): upland prairie, upland timber, swamp and bottomland, and terrace soils. Except for small areas adjacent to Panther Creek and East Branch, upland prairie soils predominate, and these are largely very dark gray to dark brown silt loams formed under prairie vegetation from thin loess (Wascher et al., 1950). The surface layer is a very dark gray to dark brown silt loam, 6 to 8 inches thick, which is medium in organic matter and slightly to medium acid. The subsurface is a light silty clay loam, very dark grayish brown and 6 to 8 inches thick. The subsoil beginning at a depth of 12 to 16 inches is a brown to dark grayish yellow silty clay. In a small area in the north-central part of the basin, the surface layer is a brown to dark brown heavy silt loam 8 to 10 inches thick, or a granular black clay loam to silty clay loam 8 to 10 inches thick (Wascher et al., 1949). These materials are high in organic matter and nitrogen and slightly acid to neutral. The subsurface layer is a brown or pale yellowish-brown silt loam, a very dark gray or grayish-black clay loam, or silt clay loam. The subsoil layer which begins at a depth of 14 to 18 inches is a silty clay loam, ranging from yellowish brown to dark gray.

The upland prairie soils occur on 1 to 6 percent slopes. Surface drainage is moderate, and artificial drainage is often required for agricultural development. The permeability is moderately slow, but underdrainage by tiles is satisfactory under proper farm management. The materials beneath the subsoils to depths of 40 to 60 inches are compact calcareous or plastic calcareous glacial tills, except in a small area in the north-central part of the basin where there is stratified silt and sand or stratified clay, silt, and sand. The permeability of the materials beneath the subsoils is moderate to slow.

Thick deposits of glacial drift chiefly of Wisconsinan age cover the bedrock and constitute the main features of the present land surface. The deposits are composed predominantly of unstratified clayey materials called glacial till, but include some stratified beds of silt, sand, and gravel as shown by logs of wells in figure 23. The

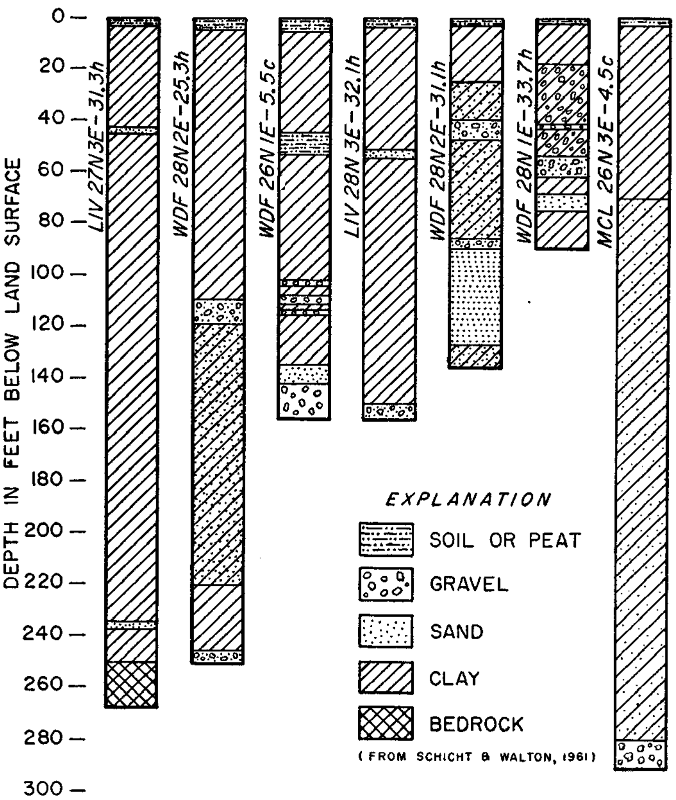


Figure 23. Logs of selected wells in Panther Creek Basin

average thickness of the glacial drift on the bedrock uplands is about 100 feet. Along the eastern edge of the basin, in Danvers bedrock valley, it may reach a thickness of more than 290 feet (see log of well MCL 26N3E-4.5c in figure 23). Sand and gravel, ranging in thickness from a few inches to more than 40 feet, occurs as irregular lenses or layers in the till. These deposits are discontinuous and are limited greatly in areal extent (Buhle, 1943). In general, because of the complex glacial history, the character of the drift varies greatly both vertically and horizontally. However, for the basin as a whole, the character of the drift in relation to the occurrence and movement of ground water is fairly uniform.

There are great variations in the water-bearing openings of till. At places where clayey materials predominate, the till is nearly impervious and yields very little water; a sandy till is somewhat more porous and permeable. Most dug wells in till have small yields and obtain water from the lenses or layers of sand and gravel that are interbedded in the compact clayey materials. The porosity and specific yield of till are not great because the sorting of material is poor and small sediments occupy pore spaces between larger fragments of rock (Dapples, 1959).

The surficial glacial deposits are immediately underlain by bedrock formations of Pennsylvanian age consisting predominantly of shale with alternating thin beds of limestone, sandstone, siltstone, fire clay, and coal. These formations are situated structurally on the northwest flank of the Illinois basin, and dip regionally south-southeastward at uniform rates less than 15 ft/mi. At Minonk the thickness of the Pennsylvanian rocks is about 515 feet. The Pennsylvanian rocks generally have low porosities and permeabilities and yield small amounts of water to wells from interconnected cracks, fractures, crevices, joints, and bedding planes. Water-bearing openings are variable from place to place and are best developed near the surface and in the thin limestones and sandstones. Practically speaking, the rocks are important because they act as a barrier to deep percolation. The bedrock surface topography is relatively flat except for the Danvers bedrock valley. The elevation of the bedrock upland averages 625 feet according to Horberg (1957). The bedrock surface slopes eastward along the eastern edge toward Danvers bedrock valley, and its elevation declines from about 600 to 450 feet in a distance of 4 miles.

Streamflow. Daily mean streamflow at gaging station 1 was plotted for 1951, 1952, and 1956 (*see figures 5, 6, and 7 in* Schicht and Walton, 1961). In general, streamflow is high in winter and spring and low in the summer and fall. In the winter melting of accumulated snow often produces disproportionately high streamflow for short periods of time. Daily mean streamflow exceeded 6000 cfs in July 1951, and was less than 0.1 cfs during parts of August and September 1956. Monthly and annual streamflow during 1951, 1952, and 1956 expressed in inches of water over the basin, are given in table 12.

Table 12. Monthly and Annual Streamflow in Inches, 1951, 1952, and 1956, Panther Creek Basin

	1951			1952			1956		
Month	R_s	R_g	R	R_s	R_g	R	R_s	R_g	R
Jan	0.61	0.16	0.77	0.39	0.77	1.16	neg	0.01	0.01
Feb	2.85	0.15	3.00	0.08	057	0.65	0.14	0.08	0.22
Mar	0.97	0.30	1.27	0.43	1.57	2.00	0.01	0.04	0.05
Apr	1.08	1.44	2.52	0.65	1.94	2.59	0.03	0.03	0.06
May	0.12	0.82	0.94	0.06	0.82	0.88	0.34	0.08	0.42
Jun	1.80	0.56	2.36	1.03	1.10	2.13	0.04	0.07	0.11
Jul	3.63	1.13	4.76	neg	0.27	0.27	0.04	0.03	0.07
Aug	0.16	0.22	0.38	0.01	0.04	0.05	0.01	0.01	0.02
Sep	0.03	0.10	0.13	neg	0.02	0.02	neg	neg	neg
Oct	0.07	0.22	0.29	neg	0.01	0.01	neg	neg	neg
Nov	0.97	0.55	1.52	neg	0.02	0.02	neg	0.01	0.01
Dec	0.05	0.35	0.40	0.01	0.03	0.04	neg	0.01	0.01
Annual	12.34	6.00	18.34	2.66	7.16	9.82	0.61	0.37	0.98

(*from Schicht and Walton,* 1961)

Streamflow was greatest in 1951 largely as a result of above normal precipitation during that year, and was least in 1956 when precipitation was much below normal. Several conditions were responsible for the low streamflow in 1956. Precipitation was below normal during most of 1955; consequently, the mean ground-water stage was low at the beginning of 1956. Precipitation during 1956 was only slightly in excess of evapotranspiration and soil-moisture requirements. Very little precipitation reached the water table and the mean ground-water stage and ground-water runoff were abnormally low throughout the year.

Streamflow consists of surface runoff, R_s, and ground-water runoff, R_g. Surface runoff is precipitation that finds its way into the stream channel without infiltrating into the soil. Ground-water runoff is precipitation that infiltrates into the soil or to the water table and then percolates into the stream channel. Surface runoff reaches streams rapidly and is discharged from the basins within a few days. Ground water percolates slowly towards and reaches streams gradually.

Ground-water runoff. Rating curves were prepared to determine the relationship between mean ground-water stage and ground-water runoff. Fluctuations of the water table in the basin were shown by hydrographs of selected wells (*see figure 8 in* Schicht and Walton, 1961). Daily averages of ground-water levels in wells in the basin were computed for selected dates when streamflow consisted entirely of ground-water runoff. Three to five days after precipitation ceases, there is no surface runoff and streamflow is derived entirely from ground-water runoff. Mean ground-water stages were plotted against ground-water runoff on corresponding dates as shown in figure 24. Closed circles represent sets of data for dates November to April, when evapotranspiration is at a minimum; open circles represent sets of data for dates April through October, when evapotranspiration is great. Daily mean ground-water stages during 1951, 1952, and 1956 were plotted as yearly hydrographs (*see figure 11, 12, and 13 in* Schicht and Walton, 1961). Ground-water runoff corresponding to each mean ground-water stage was read directly from the rating curves in figure 24. During protracted

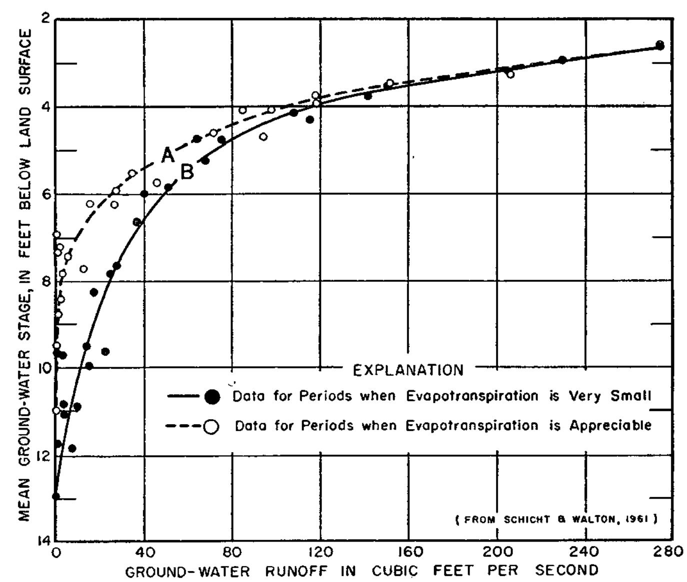


Figure 24. Rating curves of mean ground-water stage versus ground-water runoff for gaging station in Panther Creek Basin

rainless periods actual streamflow is ground-water runoff. Monthly and annual ground-water and surface runoff, expressed in inches of water over the basin, were computed by Schicht and Walton (1961) and are given in table 12. These data indicate that ground-water runoff is at a maximum during spring and early summer months and is least in late summer and fall months. More than half of annual ground-water runoff occurs during the first six months of the year. Annual ground-water runoff depends upon antecedent ground-water stage conditions as well as the amount and distribution of annual precipitation. Ground-water runoff was less in 1951 than in 1952 although precipitation was much greater in 1951 than in 1952. Ground-water stages, and consequently ground-water runoff, during the first six months of 1952 were higher than those for the same period in 1951 because of excessive precipitation during the summer months of 1951 and near normal precipitation in 1950. During extended dry periods ground-water runoff is reduced greatly. Ground-water runoff was very small during 1956 because precipitation was much below normal in 1955 and 1956. Ground-water runoff amounted to 33, 73, and 38 percent of streamflow in 1951, 1952, and 1956, respectively.

Hydrologic budget. The basin is contiguous to headwater reaches of Panther Creek, and except in the vicinity of the stream gaging station, the boundaries of the basin are reasonably congruous with ground-water and topographic divides. There is no surface or subsurface flow into or out of the basin except subsurface underflow from the basin in the vicinity of the stream gaging station. Water stored on the surface of the basin in ponds is very small, and withdrawals from wells is mostly for domestic and livestock use and is not significant. Thus, for this basin several items of the general hydrologic budget can be eliminated because they do not measurably affect the balance between water gains and losses.

Precipitation, including rain and snow, is the source of water entering the basin and is the only water gain considered in the hydrologic budget. Water leaving the basin includes streamflow, evapotranspiration, and subsurface underflow. Water is stored beneath the surface in soils and in the ground-water reservoir. Changes in storage of water in the soil are reflected in changes in soil moisture, and changes in water levels in wells indicate changes in storage of water in the ground-water reservoir. Stated as an equation, the hydrologic budget is:

$$P = R + ET + U \pm S_s \pm S_g \quad (4)$$

where:

P = precipitation
R = streamflow
ET = evapotranspiration
U = subsurface underflow
S_s = change in soil moisture
S_g = change in ground-water storage

Evapotranspiration can be determined by balancing equation 4. Soil moisture, one of the hydrologic factors, was not measured during these investigations; therefore, daily, weekly, and monthly evapotranspiration cannot be appraised. However, soil moisture is near field capacity during January of most years, and annual change in soil moisture is very small. Equation 4 can be rewritten for an annual inventory period as follows:

$$ET = P - R - U \pm S_g \quad (5)$$

Application of equation 5 requires that the annual change in soil moisture is not significant. Annual values of evapotranspiration in 1951, 1952, and 1956 estimated by balancing equation 5 are given in table 13. Methods used

Table 13. Monthly and Annual Evapotranspiration in Inches, 1951, 1952 and 1956, Panther Creek Basin

	1951			1952			1956		
Month	ET_s	ET_g	ET	ET_s	ET_g	ET	ET_s	ET_g	ET
Jan		neg			neg			neg	
Feb		neg			neg			neg	
Mar		neg			neg			neg	
Apr		0.08			0.13			0.06	
May		0.27			0.43			0.11	
Jun		0.18			0.18			0.12	
Jul		0.05			0.47			0.13	
Aug		0.34			0.33			0.14	
Sep		0.23			0.28			0.12	
Oct		0.04			0.19			0.06	
Nov		neg			neg			neg	
Dec		neg			neg			neg	
Annual	23.52	1.19	24.71	21.93	2.01	23.94	18.01	0.74	18.75

(from Schicht and Walton, 1961)

to determine underflow and change in ground-water storage are described later in this section. The range in annual evapotranspiration is much less than the range in annual precipitation. Evapotranspiration was less in 1952, a year of near normal precipitation, than in 1951, a year of above normal precipitation, even though the average temperature during the growing season of 1951 was below normal and the average temperature during the growing season of 1952 was above normal. The ratio of evapotranspiration and precipitation was 56 percent in 1951, 73 percent in 1952, and 96 percent in 1956. Evapotranspiration may be subdivided into two parts according to the source of the water discharged into the atmosphere as follows: 1) surface and soil evapotranspiration, ET_s, and 2) ground-water evapotranspiration, ET_g. The part of evapotranspiration derived from soil moisture and by evaporation from the surfaces of water, vegetation, buildings, and other objects is surface and soil evapotranspiration; the part derived from the water table is ground-water evapotranspiration.

Ground water continuously percolates toward streams; however, the roots of plants and soil capillaries intercept and discharge into the atmosphere some of the water which otherwise would become ground-water runoff. *Ground-water evapotranspiration* can be estimated from rating curves of mean ground-water stage versus ground-water runoff. Ground-water runoff corresponding to a ground-water stage is read from rating curves prepared for dates April through October, when ground-water evapotranspiration is great, and for dates November through March, when ground-water evapotranspiration is very small. The difference in ground-water runoff between the two curves is the approximate ground-water evapotranspiration. Estimates of daily ground-water evapotranspiration during 1951, 1952, and 1956 were computed from mean ground-water stages and the ground-water stage runoff rating curves in figure 24. Monthly and annual ground-water evapotranspiration are given in table 13. These data indicate that monthly ground-water evapotranspiration is greatest generally during July and August and that annual ground-water evapotranspiration is least during dry years. The ratio of ground-water evapotranspiration to total evapotranspiration was 5, 8, and 4 percent in 1951, 1952, and 1956, respectively.

Subsurface underflow out of the basin occurs in the vicinity of the stream gaging station. Underflow can be estimated from equation 1. The width of the lowlands adjacent to Panther Creek through which underflow occurs is about 500 feet. Based on the bedrock surface and topographic maps, the thickness of the glacial drift is estimated to be less than 25 feet and the hydraulic gradient of the water table in the vicinity of stream gaging station 1 is estimated to be less than 50 ft/mi. The coefficient of transmissibility of the deposits through which underflow occurs is low and is probably in the magnitude of 500 gpd/ft. By substituting the above data in equation 1 underflow was computed to be about 0.01 cfs and is so small that it was omitted from budget computations.

The change in mean ground-water stage during an inventory period, H, multiplied by the gravity yield, Y_g, of the deposits within the zone of ground-water fluctuation is equal to the *change in ground-water storage,* S_g. Stated as an equation:

$$S_g = H(Y_g) \qquad (6)$$

Gravity yield (Rasmussen and Andreasen, 1959) may be defined as the ratio of the volume of water that deposits will yield by gravity drainage to the total volume of deposits drained during a given period of ground-water decline. The gravity drainage of deposits is not immediate, and as a result the gravity yield is not constant but increases at a diminishing rate with the time of drainage, gradually approaching the specific yield. The specific yield is the ratio of the volume of water that deposits will yield by complete gravity drainage to the total volume of deposits.

The gravity yield of the deposits beneath the basin can be determined from the hydrologic budget. Equation 6 contains two factors, evapotranspiration and change in soil moisture, which were not measured during the study described here. However, during winter and early spring months (December, January, February, and early March) evapotranspiration and soil-moisture change are very small (Thornthwaite et al., 1958). A reasonable estimate of evapotranspiration for periods during winter and early spring months is 0.3 inch per month. Soil-moisture change can be eliminated and evapotranspiration estimated to average 0.3 inch per month without introducing serious error in the hydrologic budget. Equation 6 may be rewritten for inventory periods during winter and early spring months when the water table is rising as follows:

$$Y_g = (P - R - ET - U) / H \qquad (7)$$

Equation 7 is valid for periods when the soil-moisture change is not significant.

Computations of gravity yield were made using equation 7 and data for nine inventory periods during winter and early spring months, 1951-1958. Values of Y_g were plotted against the average time of drainage preceding the inventory periods (*see figure 14 in* Schicht and Walton, 1961). These data indicate that the average gravity yield of the glacial deposits increases at a diminishing rate from about 1 percent for a drainage period of 10 days to about 8 percent for a drainage period of 140 days. Extrapolation of the data suggests that the average specific yield of glacial deposits beneath the basin is about 12 percent. Monthly increases or decreases in ground-water storage during 1951, 1952, and 1956 were estimated by multiplying mean ground-water stage changes by appropriate values of Y_g. The data on changes in ground-water storage appear in table 14.

Table 14. Monthly and Annual Ground-Water Recharge in Inches, 1951, 1952 and 1956, Panther Creek Basin

	1951		1952		1956	
Month	P_g	S_g	P_g	S_g	P_g	S_g
Jan	0.44	+0.28	0.69	—0.08	neg	—0.01
Feb	0.20	+0.05	0.57	neg	0.29	+0.21
Mar	1.16	+0.86	1.71	+0.14	neg	—0.04
Apr	2.20	+0.68	1.92	—0.15	0.11	+0.02
May	0.89	—0.20	1.11	—0.14	0.20	+0.01
Jun	0.79	+0.05	1.36	+0.08	0.09	—0.10
Jul	1.03	—0.15	0.15	—0.59	0.06	—0.10
Aug	0.41	—0.15	0.18	—0.19	0.06	—0.09
Sep	0.12	—0.21	0.03	—0.27	0.02	—0.10
Oct	0.03	—0.23	0.02	—0.18	0.02	—0.04
Nov	0.88	+0.33	0.02	neg	0.01	neg
Dec	0.23	—0.12	0.27	+0.24	0.01	neg
Annual	8.38	+1.19	8.03	—1.14	0.87	—0.24

(from Schicht and Walton, 1961*)*

Ground-water budget. With a few possible exceptions the water table rose, or declined less than was necessary to balance ground-water runoff and evapotranspiration, during portions of every month of 1951, 1952, and 1956. There was, therefore, some ground-water recharge in most months of these years. Monthly and annual ground-water recharge was estimated by balancing the following equation (ground-water budget) and is given in table 14:

$$P_g = R_g + ET_g + U \pm S_g \qquad (8)$$

where:

P_g = ground-water recharge
R_g = ground-water runoff
ET_g = ground-water evapotranspiration
U = subsurface underflow
S_g = change in ground-water storage

Ground-water recharge during the three years ranged from 8.38 inches (400,000 gpd/sq mi) in 1951 to 0.87 inch (41,000 gpd/sq mi) in 1956, and 8.03 inches (380,000 gpd/sq mi) in 1952. Ground-water recharge was 19 percent of precipitation during a year of above normal precipitation, 4.5 percent of precipitation during a year of below normal precipitation, and 25 percent of precipitation during a year of near normal precipitaton.

Data in table 14 show the pronounced adverse effects of extended dry periods on ground-water recharge. Monthly ground-water recharge is largest in spring months of heavy rainfall and least in summer and fall months. Most ordinary summer rains have little or no effect on the water table. However, occasionally there was appreciable ground-water recharge when summer rainfall was in excess of evapotranspiration and soil-moisture requirements. In February, March, and the early part of April 1951, precipitation was above normal; however, ground-water recharge was only moderate. Temperatures during part of November and December of 1950 and March and the early part of April of 1951 were below normal. As a result, there was a snow cover over frozen ground much of February and March which impeded the infiltration of precipitation to the water table. Thus, most precipitation in February and March was discharged from the basin by surface runoff.

Data in figures 25 and 26 indicate that ground-water recharge generally is at a maximum during April and most recharge occurs prior to July. In many years there is very little recharge during the 5-month period July through November. Thus, water must be taken from storage within shallow aquifers for periods of at least 5 months to balance discharge.

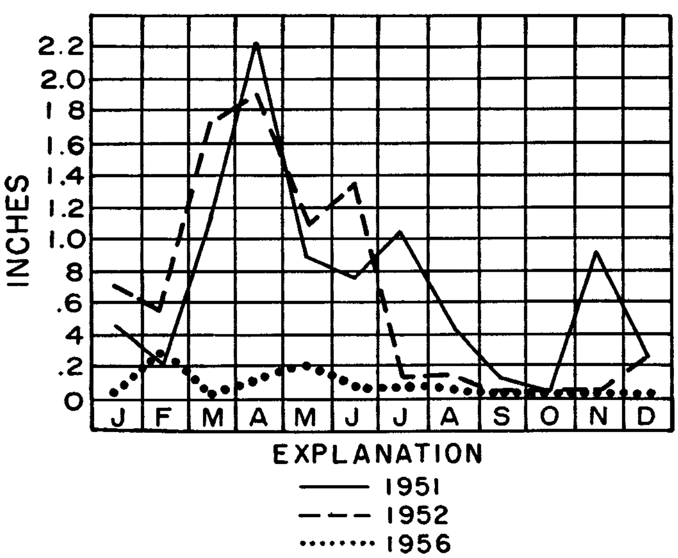


Figure 25. Monthly ground-water recharge, Panther Creek Basin

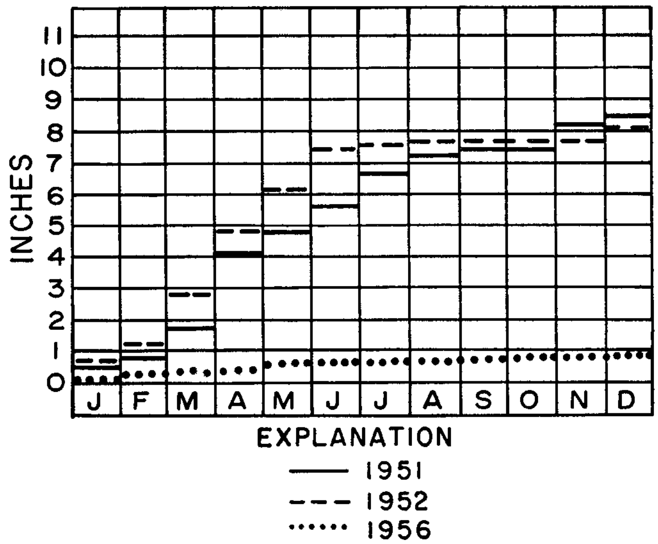


Figure 26. Cumulative monthly ground-water recharge, Panther Creek Basin

Comparison of three basins. Comparative results of annual hydrologic and ground-water budget factors for Panther, Hadley, and Goose Creek Basins are given in table 15. Data for years during which precipitation was near normal are presented. A comparison of the characteristics of the basins is given in table 16. Ground-water recharge is much greater in Panther and Goose Creek Basins than in Hadley Creek Basin. The lower ground-water recharge in Hadley Creek Basin is probably due to rugged upland topography and thin unconsolidated deposits.

Table 15. Comparison of Budget Factors for Basins in Central Illinois

Budget factor	Panther Creek 1952	Goose Creek 1957	Hadley Creek 1957
		(inches)	
Precipitation	32.62	37.18	39.73
Streamflow	9.82	9.48	13.93
Surface runoff	2.66	5.68	12.04
Ground-water runoff	7.16	3.80	1.89
Evapotranspiration	23.94	24.30	24.68
Surface and soil evapotranspiration	21.93	21.10	23.80
Ground-water evapotranspiration	2.01	3.20	0.88
Ground-water recharge	8.03	10.40	3.89
Change in ground-water storage	–1.14	+3.40	+1.05
Underflow	neg	neg	0.07

(from Schicht and Walton, 1961*)*

Summary of Recharge Rates

Data concerning recharge rates are summarized in table 17. Recharge rates vary from 1330 to 500,000 gpd/sq mi. The lowest recharge rate is for an area where the

Table 16. Comparison of Characteristics of Basins in Central Illinois

Characteristics	Panther Creek	Goose Creek	Hadley Creek
Topography	Gently undulating uplands	Level uplands	Rugged uplands
Average stream gradient	4.7 ft/mi	3.9 ft/mi	16.6 ft/mi
Vegetal cover	80% corn, oats, and soybeans; 20% pasture, woodland, and farm lots	86% corn, oats, soybeans, alfalfa, hay, wheat, rye; 14% pasture, woodland, and farm lots	40% row crops, small grain, and hay; 60% pasture, woodland, and farm lots
Soil	Upland prairie silt loams	Drummer silty clay loam and Flanagan silt loam	Upland prairie and timber silt loams
Unconsolidated deposits	100 feet of glacial till	175 feet of glacial till	25 feet of loess and 50 feet of glacial till
Bedrock formations	Shale of Pennsylvanian age	Shale of Pennsylvanian age	Shale of Mississippian and Pennsylvanian age
Average depth to water table	7 feet (below land surface)	8 feet (below land surface)	20 feet (below land surface)
North latitude	40°44'-40°54'	40°05'-40°13'	39°41'-39°50'
Mean annual temperature	51 F	53 F	55 F
Mean annual precipitation	33.6 inches	37.0 inches	36.0 inches

(from Schicht and Walton, 1961)

Table 17. Summary of Recharge Rates

Location	Flow channel discharge or pumpage *(mgd)*	Area of flow channel or diversion *(sq mi)*	Recharge rate *(gpd /sq mi)*	Lithology of deposits above aquifer or permeable zones	Aquifer
Northeastern Illinois	1.00	750	1,330	Maquoketa Formation, largely shale	Cambrian-Ordovician
DeKalb and Kendall Counties	1.80	100	18,000*	Glacial drift and units of Cambrian-Ordovician Aquifer	Cambrian-Ordovician
DuPage County	1.80	28	64,000	Glacial drift, largely till, and shaly dolomite	Dolomite of Silurian age
	4.50	32.5	138,000	Glacial drift, largely till, and dolomite	
	6.30	46.2	136,000	Glacial drift, largely till, and dolomite	
	1.20	7.6	158,000	Glacial drift, largely till, and dolomite	
LaGrange, Cook County	2.90	18	161,000	Glacial drift, largely till, and dolomite	Dolomite of Silurian age
Chicago Heights, Cook County	13.50	60	225,000	Glacial drift, largely till, and dolomite	Dolomite of Silurian age
Libertyville, Lake County	3.00	58	52,000	Glacial drift, largely till, and shaly dolomite	Glacial sand and gravel
Woodstock, McHenry County	1.90	15	127,000	Glacial drift, largely till	Glacial sand and gravel
Near Joliet, Will County	2.20	11	200,000	Glacial drift, largely silt and and sand	Glacial sand and gravel
Champaign-Urbana, Champaign County	6.30	55	115,000	Glacial drift, largely till	Glacial sand and gravel
Havana region, Mason and Tazewell Counties	1.70	6.6	258,000	Glacial drift, largely till	Glacial sand and gravel
	2.15	7.7	279,000	Glacial drift, largely till	
	3.20	6.4	500,000	Glacial drift, largely sand and gravel	
	2.04	4.2	486,000	Glacial drift, largely sand and gravel	
East St. Louis, Madison and St. Clair Counties	0.72	3.3	347,000*	Glacial drift and alluvium, largely sand and gravel	Glacial sand and gravel
	1.51	4.4	343,000*	Glacial drift and alluvium, largely sand and gravel	
	0.22	1.2	299,000*	Glacial drift and alluvium, largely sand and gravel	
	3.80	5.7	370,000*	Glacial drift and alluvium, largely sand and gravel	
Panther Creek Basin, Woodford, Livingston, and McLean Counties	—	—	380,000	Glacial drift, largely till	Glacial drift

**Changes in storage of water within aquifer taken into consideration in computing recharge rates*

Cambrian-Ordovician Aquifer is overlain by a thick layer of shale bedrock (Maquoketa Formation); the second lowest recharge rate (18,000 gpd/sq mi) is for an area where the Cambrian-Ordovician Aquifer is overlain by glacial drift and shaly bedrock retards the vertical movement of water from the glacial drift to permeable bedrock units of the multiunit aquifer.

Recharge rates for dolomite aquifers of Silurian age, overlain largely with till, range from 52,000 to 225,000 gpd/sq mi. Low rates are computed for areas where shaly dolomite beds overlie permeable zones within the dolomite aquifers. In areas where permeable zones within the dolomite aquifers are overlain by permeable dolomite beds and thick glacial drift consisting largely of till, the recharge rate averages about 150,000 gpd/sq mi.

Recharge rates for glacial sand and gravel aquifers range from 115,000 to 500,000 gpd/sq mi. The lowest rate is for an area where the sand and gravel aquifer is overlain by thick glacial drift consisting largely of till. In areas where sand and gravel deposits occur from the surface to bedrock, recharge rates for sand and gravel aquifers commonly exceed 300,000 gpd/sq mi.

Theoretical Aspects

The rate of recharge may be expressed mathematically by the following form of Darcy's law:

$$Q_c / A_c = 2.8 \text{ X } 10^7 \, (P'/m') \quad h \qquad (9)$$

where:

- Q_c / A_c = recharge rate, in gpd/sq mi
- Q_c = leakage (recharge) through deposits, in gpd
- A_c = area of diversion, in sq mi
- P' = coefficient of vertical permeability of deposits, in gpd/sq ft
- m' = saturated thickness of deposits, in ft
- h = difference between the head in the aquifer and in the source bed above deposits through which leakage occurs, in ft

As shown in equation 9, the recharge rate varies with the vertical head loss associated with leakage of water through deposits. The recharge rate per unit area, being dependent upon vertical head loss, is not constant but varies in space and time. The recharge rate is generally greatest in the deepest parts of cones of depression and decreases with distance from a pumping center. The recharge rate increases as the piezometric surface declines and the vertical head loss increases. The recharge rate per unit area is at a maximum when the piezometric surface of the aquifer is at the base of the deposits through which leakage occurs, provided the head in the source bed above the deposits remains fairly constant.

The recharge rate per unit area is valid for one and only one average vertical head loss. On the other hand, the recharge rate per unit area per foot of head loss remains constant as long as the saturated thickness and vertical permeability of the deposits through which leakage occurs does not change and the piezometric surface does not decline below the base of the deposits. Thus, the recharge rate per unit area per foot of head loss is much more meaningful than the recharge rate per unit area. The recharge rate per unit area per foot of head loss (P'/m') is the leakage coefficient (Hantush, 1956) and is given by the following equation:

$$P'/m' = Q_c / [A_c \quad h \; (2.8 \times 10^7)] \qquad (10)$$

Differences in recharge rates per unit area or leakage coefficients from place to place cannot be attributed only to differences in the hydraulic conductivities of deposits. Recharge rates per unit area and leakage coefficients can vary from place to place because of variations in vertical head loss and/or saturated thickness of deposits, as well as variations in hydraulic conductivities. The hydraulic conductivities of deposits through which leakage occurs are expressed as coefficients of vertical permeability (P′) and can be computed by multiplying leakage coefficients by the saturated thickness of deposits.

Equation 10 may be rewritten in terms of P' as follows:

$$P' = Q_c \, m' / [\quad h \, A_c \; (2.8 \times 10^7)] \qquad (11)$$

Available data on recharge rates, vertical head losses, and thickness of deposits for areas described earlier in this report can be substituted into equation 11 to compute coefficients of vertical permeability.

Your attention is directed to data pertaining to recharge to the Cambrian-Ordovician Aquifer in northeastern Illinois. The recharge rate per unit area that was computed for area 1 where the Maquoketa Formation overlies the aquifer and with piezometric surface data for 1864 is much less than the recharge rate per unit area computed for the entire Chicago region with data for 1958. Thus, if the recharge rate per unit area for 1864 was used to estimate the amount of recharge in 1958, resulting computations would be in error. The recharge rate per unit area for 1958 is much greater than the recharge rate per unit area for 1864 largely because the vertical head loss greatly increased with declines in the piezometric surface during the period 1864 to 1958. It is apparent that recharge under heavy pumping conditions cannot be estimated unless vertical head loss is considered.

The recharge rate per unit area for the Cambrian-Ordovician Aquifer in 1958 in areas where the Galena-Platteville Dolomite is the uppermost bedrock is about 8.6 times the recharge rate per unit area for the Cambrian-Ordovician Aquifer in areas where the Maquoketa Formation overlies the aquifer, if vertical losses are not considered. Taking into consideration vertical head losses, the recharge rate per unit area per foot of head loss in areas where the Maquoketa Formation is missing is about 64 times the recharge rate per unit area per foot of head loss in areas where the Maquoketa Formation overlies the aquifer. It is apparent that comparisons of recharge rates are meaningless unless vertical head losses are considered; the saturated thickness of deposits through which leakage occurs must also be taken into consideration.

Coefficients of Leakage and Vertical Permeability

Several controlled aquifer tests, involving one or more observation wells, were made under leaky artesian conditions (Walton, 1960). Test data were analyzed with the leaky artesian aquifer equation (Hantush and Jacob, 1955) to determine the coefficients of leakage and vertical permeability of confining beds overlying glacial drift aquifers. Coefficients of vertical permeability computed from aquifer-test, flow-net, and geohydrologic system analyses described earlier are given in table 18 and are summarized in table 19.

Values of P´ for drift deposits consisting largely of sand and gravel exceed 1.0 gpd/sq ft and average 1.31 gpd/sq ft. As the clay content of the drift increases, values of *P´* decrease and average about 0.25 gpd/sq ft when considerable sand and gravel is present and about 0.03 gpd/sq ft when little sand and gravel is present. It is apparent that for the purpose of estimating vertical permeability or recharge rates it is not sufficient to classify deposits as drift because the range in values of P´ for drift is great.

A comparison of the values of P´ for drift and for shale (Maquoketa Formation) indicates that the least permeable drift is about 200 times as permeable as the Maquoketa Formation.

In places where shaly dolomite beds overlie permeable zones within the dolomite aquifers the vertical permeability is considerably lower than in places where shaly dolomite beds are absent. The leakage coefficient for the Maquoketa Formation is about 1/300 of the lowest value of *P´/m´* for drift.

Recharge rates under heavy pumping conditions can be estimated by substituting in equation 9 data on the coefficient of vertical permeability (table 18), saturated thickness of deposits, and available vertical head losses.

Very few data on the coefficient of vertical permeability of till have been published. Norris (1962) gave several values of the coefficient of vertical permeability based on laboratory and aquifer-test data for tills in Ohio, Illinois,

Table 18. Coefficients of Leakage and Vertical Permeability

Location	*P´* (gpd /sq ft)	*h* (ft)	*m´* (ft)	*P´/m´* (gpd /cu ft)	Lithology
Beecher City, Effingham County	0.25		12	2.1×10^{-2}	Drift, clay with considerable sand and gravel
Dieterich, Effingham County	0.10		14	7.1×10^{-3}	Drift, clay with considerable sand and gravel
Cowden, Shelby County	1.60		7	2.3×10^{-1}	Drift, sand and gravel with some clay
Assumption, Christian County	0.19		8	2.4×10^{-2}	Drift, clay with considerable sand and gravel
Mattoon, Coles County	0.63		12	5.2×10^{-2}	Drift, clay with considerable sand and gravel
Barry, Pike County	0.15		16	9.4×10^{-3}	Drift, clay with considerable sand and gravel
Winchester, Scott County	0.08		16	5.0×10^{-3}	Drift, clay with some sand and gravel
Arcola, Douglas County	0.04		70	5.7×10^{-4}	Drift, clay with some sand and gravel
Northeastern Illinois	0.00005	300	200	2.5×10^{-7}	Dolomitic shale
West Chicago, DuPage County	0.0046	45	90	5.1×10^{-5}	Drift, clay with some sand and gravel and shaly dolomite
Downers Grove, DuPage County	0.0068	65	90	7.6×10^{-5}	Drift, clay with some sand and gravel and dolomite
Woodstock, McHenry County	0.012	30	80	1.5×10^{-4}	Drift, clay with some sand and gravel
Champaign-Urbana, Champaign County	0.01	50	120	8.3×10^{-5}	Drift, clay with some sand and gravel
Near Joliet, Will County	1.02		30	3.4×10^{-2}	Drift, sand and gravel with some clay
Champaign-Urbana, Champaign County	0.21		35	6.1×10^{-3}	Drift, clay with considerable sand and gravel
LaGrange, Cook County	0.008	30	40	2.0×10^{-4}	Drift, clay with some sand and gravel and dolomite
Libertyville, Lake County	0.009	40	200	4.5×10^{-5}	Drift, clay with some sand and gravel and dolomite
Chicago Heights, Cook County	0.011	25	35	3.2×10^{-4}	Drift, clay with some sand and gravel and dolomite

Table 19. Summary of Coefficients of Leakage and Vertical Permeability

Lithology	P'/m' (gpd /cu ft) range	P' (gpd /sq ft) range	average
Drift, sand and gravel, some clay and silt	3.4×10^{-2} - 2.3×10^{-1}	1.02 - 1.60	1.31
Drift, clay and silt with considerable sand and gravel	6.1×10^{-3} - 5.2×10^{-2}	0.10 - 0.63	0.25
Drift, clay and silt with some sand and gravel	8.3×10^{-5} - 5.0×10^{-3}	0.01 - 0.08	0.03
Drift, clay and silt with some sand and gravel and dolomite	4.5×10^{-5} - 3.2×10^{-4}	0.005 - 0.011	0.008
Drift, clay and silt with some sand and gravel and shaly dolomite	5.1×10^{-5}	0.005	0.005
Dolomite shale	2.5×10^{-7}	0.00005	0.00005

and South Dakota. As shown in figure 27 permeability values range from 0.0003 to 0.9 gpd/sq ft but commonly exceed 0.01 gpd/sq ft. The permeability values for till in Illinois are in about the same general range as that of the tills in Ohio and South Dakota.

As stated by Norris (1962) there is a widespread misconception that till is a more or less random accumulation of drift, ranging widely in its physical properties from place to place and lacking continuity in many of the common lithofacies characteristics upon which correlation of sediments is based. Evidence shows that till in widely separated areas actually is reasonably uniform in permeability and, by inference, in other related properties. With sound professional judgment, permeability values can be extrapolated over fairly large distances and applied with reasonable confidence in estimating recharge rates.

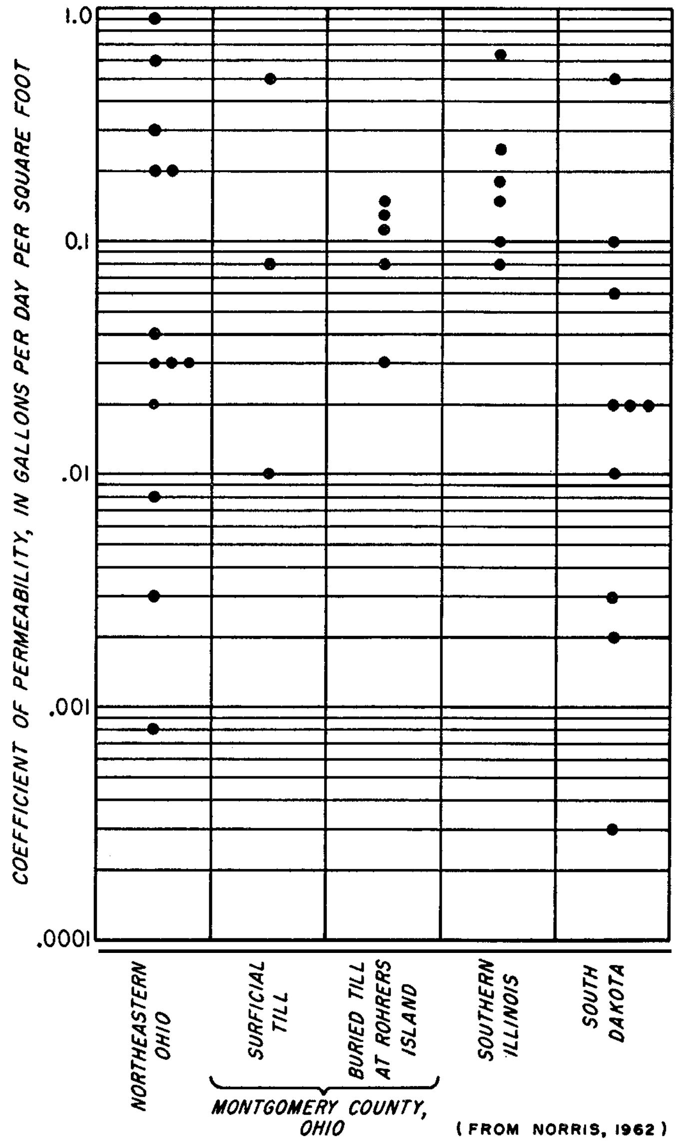


Figure 27. Chart showing range of till vertical permeability in Ohio, Illinois, and South Dakota

GROUND-WATER RUNOFF

Streamflow consists of surface runoff and ground-water runoff. Surface runoff is here defined as precipitation that finds its way into the stream channel without infiltrating into the soil. Ground-water runoff is precipitation that infiltrates into the soil or to the water table and then percolates into the stream channel. Ground-water runoff includes bank storage.

Estimating Ground-Water Runoff

Streamflow data in the Water-Supply Papers published by the U. S. Geological Survey and flow-duration studies by Mitchell (1957) were used to determine annual ground-water runoff from the 109 drainage basins within Illinois shown in figure 28. Data for the Fox and Illinois Rivers were not processed because of the complicating effects of discharge of sewage into these streams, diversion of water from Lake Michigan into the Illinois River, and flow control structures.

Streamflow data for years of near (1948), below (1953 or 1956), and above (1942 or 1951) normal precipitation were investigated. Based on existing geologic, topographic, and land use maps for Illinois, 21 widely scattered basins having contrasting characteristics and size were selected for detailed analysis. Basin areas varied from 10.1 to 8700 square miles; basin characteristics varied from glaciated, impermeable bedrock, thick drift, gentle topographic relief, gentle stream gradient, little

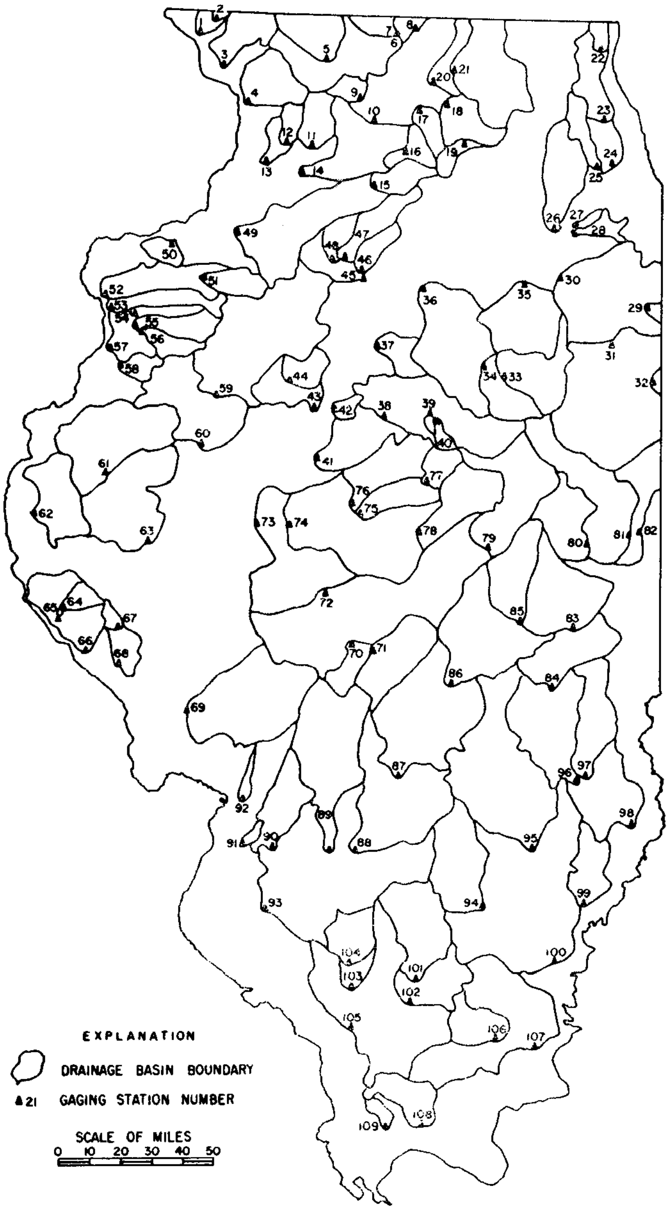


Figure 28. Location of drainage basins

forest and woodland to unglaciated, permeable bedrock, thin unconsolidated deposits, rugged topography, steep stream gradient, and considerable forest and woodland.

Daily mean streamflow at the 21 selected gaging stations during years of near, below, and above normal precipitation were plotted on semilogarithmic hydrograph paper. Hydrographs were divided into two components, surface runoff and ground-water runoff, with streamflow hydrograph separation methods outlined by Linsley, Kohler, and Paulhus (1958).

Principles of separating the streamflow hydrograph into its two major components are not well developed. Ground-water runoff several days after precipitation ceases is readily determined; however, ground-water runoff under flood hydrographs is the subject of much discussion. Wisler and Brater (1959) made the following pertinent remarks concerning ground-water runoff under flood hydrographs.

"Except for the smallest stream rises, the rise in the stage of the river occurs more quickly and is much greater in magnitude than the corresponding rise of the water table . . . Consequently as quickly as the water surface in the stream rises higher than the adjacent water table, thus creating at any given elevation a greater hydrostatic pressure in the stream than in the banks, ground-water inflow into the stream channel ceases temporarily and the direction of flow reverses, creating bank storage . . . The volume of this bank storage continues to increase as long as the water level in the stream is higher than the water table . . . or until after the stream has passed its peak stage. As soon as the stage starts to fall, the direction of flow again reverses, and for a time, because of the accumulated bank storage, the ground-water contribution to the stream is considerably increased. As soon as the bank storage is drained out, the ground-water flow again follows the normal depletion curve."

Even though ground-water runoff into the stream channel ceases temporarily during periods of flood, ground water continues to percolate towards the stream creating ground-water storage in the lowlands adjacent to the stream channel. As soon as the stream stage starts to fall, ground-water runoff is considerably increased not only because of the accumulated bank storage but also because of the accumulated ground-water storage. When bank and ground-water storage is drained out, ground-water runoff will generally be greater than before precipitation occurred because during most flood periods precipitation infiltrating into the ground-water reservoir causes the water table to rise and the hydraulic gradient toward the stream to increase.

Ground-water runoff under flood hydrographs was estimated in the following manner: A straight line was drawn from the point of rise to the hydrograph N days after the peak; N, defined as the time after the peak of the streamflow hydrograph at which surface runoff terminates, was approximated by $N = A^{0.2}$ where A is the drainage basin area in square miles. Another line was constructed by projecting the recession of the streamflow after the storm back under the hydrograph to a point under the inflection point of the falling limb; an arbitrary rising limb was sketched from the point of rise of the hydrograph to connect with the projected streamflow recession. Ground-water runoff was estimated as the average of these two lines. On the basis of studies made by Schicht and Walton (1961) it is believed that the procedures described above give reasonably accurate estimates of ground-water runoff under flood hydrographs even though the lines do not describe the time sequence of events occurring during periods of flood. Annual ground-water runoff during years of near, below, and above normal precipitation for the 21 drainage basins is given in table 20.

Table 20. Annual Ground-Water Runoff and Frequencies of Occurrence of Streamflows

Basin number	Basin, north latitude *(degrees)*	Ground-water runoff based on hydrograph separation *(cfs /sq mi)*			Frequencies of occurrence of streamflows *(percent)*			Ground-water runoff based on flow duration curves *(cfs /sq mi)*		
		near*	below*	above*	near	below	above	near	below	above
4	42°06'50"	0.23	0.12	0.59	58	80	29	0.26	0.15	0.49
17	42°05'55"	0.22	0.08	0.45	43	75	24	0.16	0.08	0.30
21	42°15'20"	0.30	0.15	0.62	52	76	23	0.24	0.16	0.43
23	42°04'55"	0.15	0.12	0.50	53	56	37	0.14	0.05	0.60
25	41°31'10"	0.35	0.17	0.63	43	60	32	0.22	0.09	0.62
26	41°31'20"	0.40	0.21	0.69	46	66	33	0.32	0.19	0.61
28	41°31'10"	0.28	0.17	0.44	46	57	30	0.18	0.11	0.38
35	41°17'10"	0.15	0.01	0.29	45	62	35	0.07	0.01	0.29
38	40°37'25"	0.23	0.05	0.45	48	66	34	0.24	0.05	0.46
49	41°29'20"	0.23	0.11	0.63	58	82	24	0.28	0.16	0.43
70	39°34'40"	0.22	0.07	0.47	52	70	38	0.30	0.12	0.58
73	40°07'25"	0.30	0.16	0.55	49	61	37	0.34	0.15	0.55
76	40°13'10"	0.23	0.09	0.40	50	62	41	0.32	0.08	0.50
81	40°06'09"	0.39	0.26	0.41	42	51	39	0.32	0.14	0.56
93	38°19'22"	0.33	0.15	0.71	40	59	26	0.28	0.17	0.44
95	38°38'05"	0.24	0.12	0.27	36	51	33	0.17	0.08	0.23
96	38°56'10"	0.44	0.27	0.51	36	50	32	0.30	0.18	0.42
99	38°22'50"	0.11	0.08	0.53	37	40	18	0.06	0.02	0.12
100	38°03'40"	0.28	0.18	0.39	40	46	35	0.19	0.10	0.39
103	37°58'00"	0.14	0.09	0.29	45	52	25	0.14	0.08	0.18
108	37°21'00"	0.74	0.23	0.79	32	47	30	0.35	0.18	0.58

****Words indicate data are for years of near, below, and above normal precipitation***

Annual ground-water runoffs for the 21 basins were compared with the standard-period flow-duration curves for the basins given by Mitchell (1957). The flow-duration curve is an accumulative frequency curve of a continuous time series of mean daily discharges displaying the relative duration of various magnitudes of streamflow. The frequencies of occurrence of streamflows corresponding to ground-water runoffs (table 20) were obtained from the flow-duration curves and were plotted against the latitudes of the basins as shown in figure 29.

Three straight lines were fitted to data to describe the relations between latitudes of basins and frequencies of occurrence of streamflows corresponding to ground-water runoffs during years of near, below, and above normal precipitation. The data are scattered in large part due to the fact that amounts and distribution of annual precipitation and antecedent moisture conditions preceding study periods vary from basin to basin. Annual ground-water runoffs from three basins in northeastern Illinois, during three different years of near normal precipitation, were estimated with streamflow hydrograph separation methods described earlier. The annual ground-water runoffs ranged about in the same manner as the data for a year of near normal precipitation in figure 29. The three straight lines, therefore, represent ground-water runoff conditions averaged over a number of years of near, below, and above normal precipitation and cannot be expected to precisely describe ground-water runoff during any one given year.

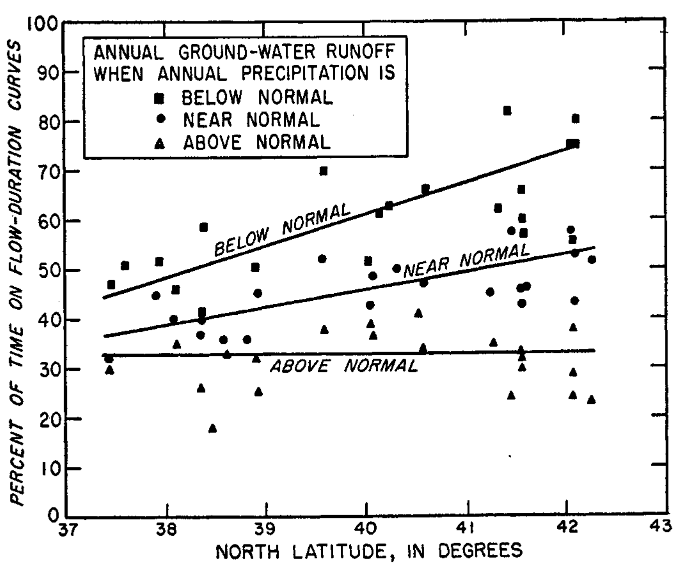


Figure 29. Relation between flow-duration curves and annual ground-water runoff

Ground-water runoffs from the 21 selected basins estimated with the straight-line graphs in figure 29 and flow-duration curves were compared with ground-water runoffs estimated with streamflow hydrograph separation methods. Differences between ground-water runoffs estimated with the two methods averaged about 23 percent of ground-water runoffs estimated with streamflow hydrograph separation methods. It is concluded that ground-water runoff in Illinois can be rapidly estimated without excessive error with standard-period flow-duration curves and figure 29. Annual ground-water runoffs (table 21) from the 109 basins in figure 28 were estimated by the following procedure. According to the latitude of the particular basin and precipitation conditions, figure 29 was used to determine the frequency of occurrence of streamflow corresponding to ground-water runoff. The

Table 21. Gaging Station Locations and Annual Ground-Water Runoff

Basin number	Gaging station location	Annual ground-water runoff (*cfs/sq mi*) near*	below*	above*	Ratio $(Q_{25}/Q_{75})^{1/2}$	Basin area (*sq mi*)	Basin north latitude (*degrees*)
1	Galena River at Galena	0.36	0.27	0.51	1.52	192	42°24'50"
2	East Fork, Galena River at Council Hill	0.30	0.18	0.40	1.67	20.1	42°28'05"
3	Apple River near Hanover	0.30	0.19	0.50	1.85	244	42°15'05"
4	Plum River below Carroll Creek	0.26	0.15	0.49	2.04	231	42°06'50"
5	Pecatonica River at Freeport	0.41	0.28	0.58	1.53	1330	42°18'13"
6	Pecatonica River at Shirland	0.38	0.29	0.59	1.53	2540	42°26'10"
7	Sugar River at Shirland	0.40	0.32	0.53	1.37	757	42°26'10"
8	Rock River at Rockton	0.39	0.28	0.59	1.70	6290	42°27'05"
9	Leaf River at Leaf	0.31	0.20	0.47	1.70	102	42°07'35"
10	Rock River at Oregon	0.36	0.27	0.56	1.64	8120	42°01'00"
11	Elkhorn Creek near Penrose	0.29	0.20	0.41	1.61	153	41°54'10"
12	Rock Creek near Coleta	0.27	0.17	0.39	1.77	81.6	41°55'00"
13	Rock Creek near Morrison	0.25	0.18	0.36	1.61	143	41°49'50"
14	Rock River at Como	0.39	0.28	0.58	1.64	8700	41°47'00"
15	Green River at Amboy	0.20	0.10	0.35	2.24	199	41°42'35"
16	Kyte River near Flag Center	0.13	0.08	0.31	2.48	125	41°56'00"
17	Killbuck Creek near Monroe Center	0.16	0.08	0.30	2.31	114	42°05'55"
18	South Branch, Kishwaukee River near Fairdale	0.19	0.07	0.40	2.98	386	42°06'40"
19	South Branch, Kishwaukee River at DeKalb	0.17	0.03	0.50	6.29	70	41°55'50"
20	Kishwaukee River at Perryville	0.24	0.13	0.50	2.24	1090	42°11'45"
21	Kishwaukee River at Belvidere	0.24	0.16	0.43	1.97	525	42°15'20"
22	DesPlaines River near Gurnee	0.04	0.01	0.38	8.84	215	42°20'40"
23	DesPlaines River near DesPlaines	0.14	0.05	0.60	4.90	374	42°04'55"
24	DesPlaines River at Riverside	0.12	0.05	0.48	3.95	635	41°49'20"
25	Salt Creek at Western Springs	0.22	0.09	0.62	3.16	122	41°49'35"
26	DuPage River at Troy	0.32	0.19	0.61	2.30	325	41°31'20"
27	Spring Creek at Joliet	0.29	0.18	0.48	1.91	19.7	41°31'45"
28	Hickory Creek at Joliet	0.18	0.11	0.38	2.30	107	41°31'10"
29	Kankakee River at Momence	0.55	0.40	0.72	1.60	2340	41°09'36"
30	Kankakee River near Wilmington	0.37	0.24	0.58	1.91	5250	41°20'48"
31	Iroquois River near Chebanse	0.24	0.10	0.52	3.65	2120	41°00'29"
32	Iroquois River at Iroquois	0.22	0.13	0.50	2.91	682	40°49'25"
33	North Fork, Vermilion River	0.09	0.02	0.28	7.75	184	40°50'08"
34	Vermilion River at Pontiac	0.10	0.03	0.27	5.40	568	40°52'40"
35	Mazon River near Coal City	0.07	0.01	0.29	30.6	470	41°17'10"
36	Vermilion River at Lowell	0.15	0.05	0.37	4.46	1230	41°15'18"
37	Crow Creek near Washburn	0.20	0.06	0.40	7.75	123	40°57'15"
38	Mackinaw River near Congerville	0.24	0.05	0.46	5.79	764	40°37'25"
39	Hickory Creek above Lake Bloomington	0.30	0.01	0.59		10.1	40°38'15"
40	Money Creek above Lake Bloomington	0.28	0.03	0.56	9.50	51.9	40°37'13"
41	Mackinaw River near Green Valley	0.26	0.10	0.45	3.26	1100	40°26'43"
42	Farm Creek at East Peoria	0.11	0.06	0.26	3.10	60.9	40°40'05"
43	Kickapoo Creek near Peoria	0.16	0.07	0.36	3.10	296	40°40'55"
44	Kickapoo Creek near Kickapoo	0.18	0.07	0.36	3.70	120	40°48'00"
45	Bureau Creek at Bureau	0.25	0.17	0.45	2.16	481	41°16'40"
46	East Bureau Creek near Bureau	0.12	0.04	0.32	5.20	101	41°20'06"
47	Bureau Creek at Princeton	0.20	0.08	0.48	4.19	186	41°21'55"
48	West Bureau Creek at Wyanet	0.16	0.07	0.39	4.70	83.3	41°21'54"
49	Green River near Geneseo	0.28	0.16	0.43	2.03	958	41°29'20"
50	Mill Creek at Milan	0.16	0.07	0.34	3.22	62.5	41°26'35"
51	Edwards River near Orion	0.19	0.09	0.38	2.83	163	41°16'20"
52	Edwards River near New Boston	0.23	0.14	0.42	2.57	434	41°11'15"
53	Pope Creek near Keithsburg	0.20	0.10	0.37	2.73	171	41°07'45"
54	North Henderson Creek near Seaton	0.18	0.08	0.40	3.21	66.4	41°05'25"
55	Henderson Creek near Little York	0.17	0.08	0.34	3.16	151	41°02'35"
56	Cedar Creek at Little York	0.25	0.13	0.46	2.82	128	41°00'50"
57	Henderson Creek near Oquawka	0.19	0.12	0.38	2.40	428	41°00'05"
58	South Henderson Creek at Biggsville	0.32	0.13	0.62	3.65	81.4	40°51'25"
59	Spoon River at London Mills	0.22	0.08	0.40	3.29	1070	40°42'51"
60	Spoon River at Seville	0.28	0.14	0.45	3.16	1600	40°29'10"
61	LaMoine River at Colmar	0.29	0.14	0.46	3.05	655	40°19'45"
62	Bear Creek near Marcelline	0.10	0.05	0.15	3.60	348	40°08'34"

Table 21 (Continued)

Basin number	Gaging station location	Annual ground-water runoff (*cfs/sq mi*) near*	below*	above*	Ratio $(Q_{25}/Q_{75})^{1/2}$	Basin area (*sq mi*)	Basin north latitude (*degrees*)
63	LaMoine River at Ripley	0.22	0.09	0.38	3.46	1310	40°01'31"
64	Hadley Creek at Kinderhook	0.22	0.08	0.32	3.00	72.7	39°41'35"
65	Hadley Creek near Shinn	0.24	0.13	0.35	2.73	73.6	39°39'55"
66	The Sny River at Atlas	0.32	0.17	0.52	2.58	451	39°30'20"
67	Bay Creek at Pittsfield	0.15	0.05	0.22	4.12	39.6	39°37'30"
68	Bay Creek at Nebo	0.26	0.12	0.38	3.84	162	39°26'35"
69	Macoupin Creek near Kane	0.14	0.07	0.19	3.46	875	39°14'00"
70	South Fork, Sangamon River at Kincaid	0.30	0.12	0.58	4.25	510	39°34'40"
71	South Fork, Sangamon River at Taylorville	0.36	0.17	0.60	3.00	427	39°30'25"
72	Sangamon River at Riverton	0.30	0.13	0.44	3.94	2560	39°50'34"
73	Sangamon River near Oakford	0.34	0.15	0.55	3.16	5120	40°07'25"
74	Salt Creek near Greenview	0.38	0.17	0.59	2.83	1800	40°08'01"
75	Kickapoo Creek near Lincoln	0.28	0.10	0.46	3.86	306	40°11'30"
76	Sugar Creek near Hartsburg	0.32	0.08	0.50	3.75	335	40°13'10"
77	Kickapoo Creek near Heyworth	0.19	0.08	0.35	4.32	71.4	40°21'00"
78	Salt Creek near Rowell	0.36	0.13	0.58	3.34	334	40°07'00"
79	Sangamon River at Monticello	0.33	0.14	0.58	3.84	550	40°01'40"
80	Salt Fork, Vermilion River near Homer	0.39	0.14	0.58	3.06	344	40°03'20"
81	Vermilion River near Catlin	0.32	0.14	0.56	3.58	959	40°06'09"
82	Vermilion River near Danville	0.35	0.14	0.60	4.08	1280	40°05'53"
83	Embarras River near Oakland	0.28	0.10	0.48	4.32	535	39°40'50"
84	Embarras River near Diona	0.34	0.15	0.60	4.25	903	39°20'40"
85	Kaskaskia River near Arcola	0.36	0.12	0.52	4.95	390	39°40'50"
86	Kaskaskia River at Shelbyville	0.33	0.09	0.58	5.50	1030	39°24'20"
87	Kaskaskia River at Vandalia	0.31	0.18	0.48	3.46	1980	38°57'35"
88	Kaskaskia River at Carlyle	0.32	0.18	0.55	4.25	2680	38°36'42"
89	Shoal Creek near Breese	0.17	0.10	0.28	3.16	760	38°36'35"
90	Silver Creek near Lebanon	0.13	0.08	0.18	3.56	335	38°35'40"
91	Canteen Creek at Caseyville	0.28	0.18	0.40	2.92	22.5	38°38'35"
92	Indian Creek at Wanda	0.09	0.04	0.15	5.30	37.0	38°50'30"
93	Kaskaskia River at New Athens	0.28	0.17	0.44	3.38	5220	38°19'22"
94	Skillet Fork at Wayne City	0.08	0.04	0.18	5.65	475	38°21'25"
95	Little Wabash River at Wilcox	0.17	0.08	0.23	4.07	1130	38°38'05"
96	Embarras River at Ste. Marie	0.30	0.18	0.42	3.26	1540	38°56'10"
97	North Fork, Embarras River near Oblong	0.16	0.09	0.22	3.60	304	39°00'35"
98	Embarras River at Lawrenceville	0.40	0.28	0.62	3.33	2260	38°43'25"
99	Bonpas Creek at Browns	0.06	0.02	0.12	18.6	235	38°22'50"
100	Little Wabash River at Carmi	0.19	0.10	0.39	5.76	3090	38°03'40"
101	Big Muddy River near Benton	0.10	0.05	0.20	5.89	498	37°59'40"
102	Big Muddy River at Plumfield	0.15	0.06	0.28	7.06	753	37°54'05"
103	Beaucoup Creek near Matthews	0.14	0.08	0.18	3.87	291	37°58'00"
104	Beaucoup Creek near Pinckneyville	0.07	0.03	0.18	6.11	227	38°03'40"
105	Big Muddy River at Murphysboro	0.25	0.12	0.50	7.06	2170	37°44'55"
106	Middle Fork, Saline River near Harrisburg	0.08	0.05	0.14	5.82	198	37°44'25"
107	Saline River near Junction	0.17	0.10	0.30	7.75	1040	37°41'52"
108	Cache River at Forman	0.35	0.18	0.58	7.41	242	37°21'00"
109	Big Creek near Wetang	0.22	0.17	0.35	3.32	32.2	37°19'00"

Words indicate data are for years of near, below, and above normal precipitation

streamflow corresponding to this frequency on the flow-duration curve for the basin was selected as the ground-water runoff. Distribution of ground-water runoff is shown in figures 30, 31, and 32.

It is of interest to note that table 21 and figure 29 indicate that differences between ground-water runoffs from basins during years of near, below, and above normal precipitation decrease from north to south in Illinois. There is little difference between ground-water runoff during years of near and above normal precipitation in many parts of southern Illinois. Average annual precipitation increases from north to south as shown in figure 33. The annual maximum and minimum precipitation amounts expected on an average of once in 5 and once in 50 years are shown in figure 34.

Characteristics of Basins

The general characteristics of the 109 basins were determined from the information given by Thornburn

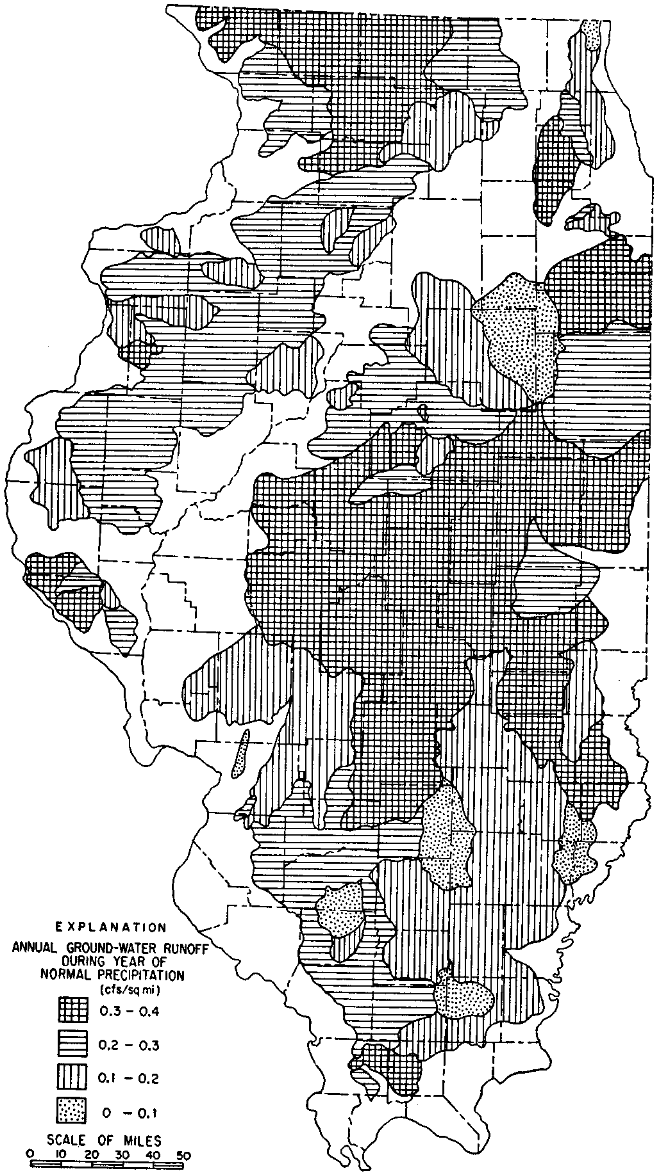


Figure 30. Distribution of annual ground-water runoff during a year of near normal precipitation

(1960); Horberg (1950); and Atlas of Illinois Resources (Section 2, Mineral Resources; and Section 3, Forest, Wildlife, and Recreational Resources). The f ollowing sections on geology and topography were abstracted largely from Thornburn (1960).

Table 22 gives an abbreviated geologic column for Illinois and indicates the sequence and general characteristics of rocks. The most common bedrock strata are those of Paleozoic age; only in extreme southern Illinois are rocks of Cretaceous or early Cenozoic age found, as shown in figure 35. Exposures of bedrock are limited to a small percentage of Illinois; unconsolidated deposits are absent or very thin chiefly in areas in the extreme western and southern parts of the state. The bedrock is scored by numerous stream valleys and their tributaries. Figure 36 shows the location of the major bedrock valleys. Horberg's (1957) Map of Bedrock Topography (see figure 37) in combination with topographic maps of the state was used to make estimates of the thickness of unconsolidated deposits overlying bedrock within basins.

Most of the bedrock exposed in Illinois is of sedimentary origin and can be classified as limestone, sandstone, or shale. In Hardin and Pope Counties in southeastern Illinois occur the state's only surface exposure of igneous rocks. The sedimentary rocks, originally deposited as relatively flat-lying beds or strata, were subsequently down-folded into a large spoon-shaped basin, the deepest part of which lies in southeastern Illinois. In the central part of the state, the strata are relatively flat lying; however, at the borders of the state, particularly in the north, south, and southwest, the strata dip up and rocks of

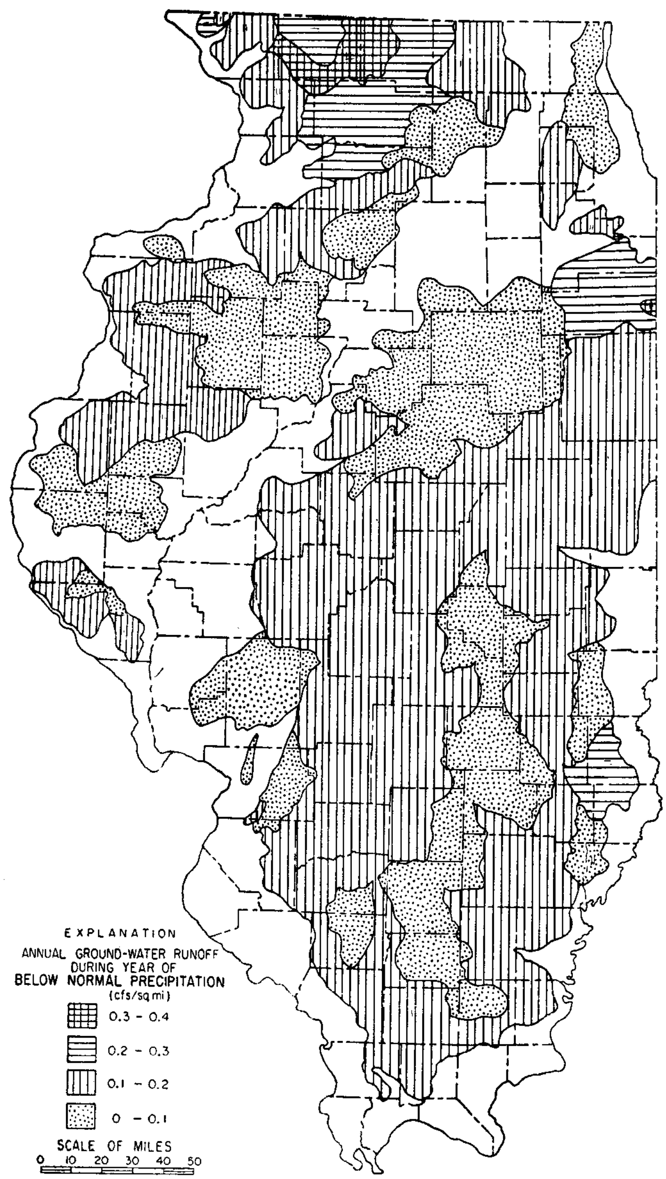


Figure 31. Distribution of annual ground-water runoff during a year of below normal precipitation

increasingly greater age are exposed outward from the deepest part of the troughlike structure.

Only small scattered supplies of ground water are available from Pennsylvanian, Mississippian, and Devonian rocks. These bedrock formations are classified as relatively impermeable and cover all but the northern third of Illinois. Tertiary, Cretaceous, Silurian, Ordovician, and Cambrian bedrock formations are favorable aquifers and yield large quantities of water to wells. These rocks are classified as permeable and cover the northern third and extreme southern tip of the state.

Since most streams in Illinois are cut into materials laid down during the glacial epoch, referred to geologically as the Pleistocene, the character of glacial deposits is an important characteristic of basins. Most of the state is mantled by unconsolidated deposits left by the

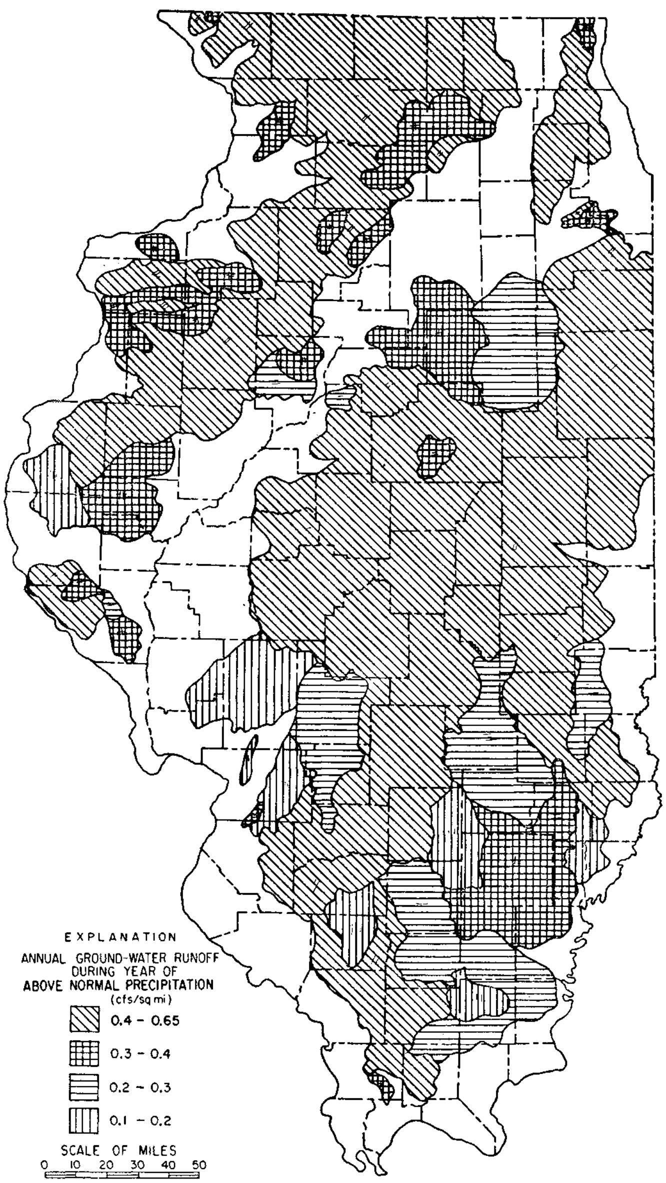


Figure 32. **Distribution of annual ground-water runoff during a year of above normal precipitation**

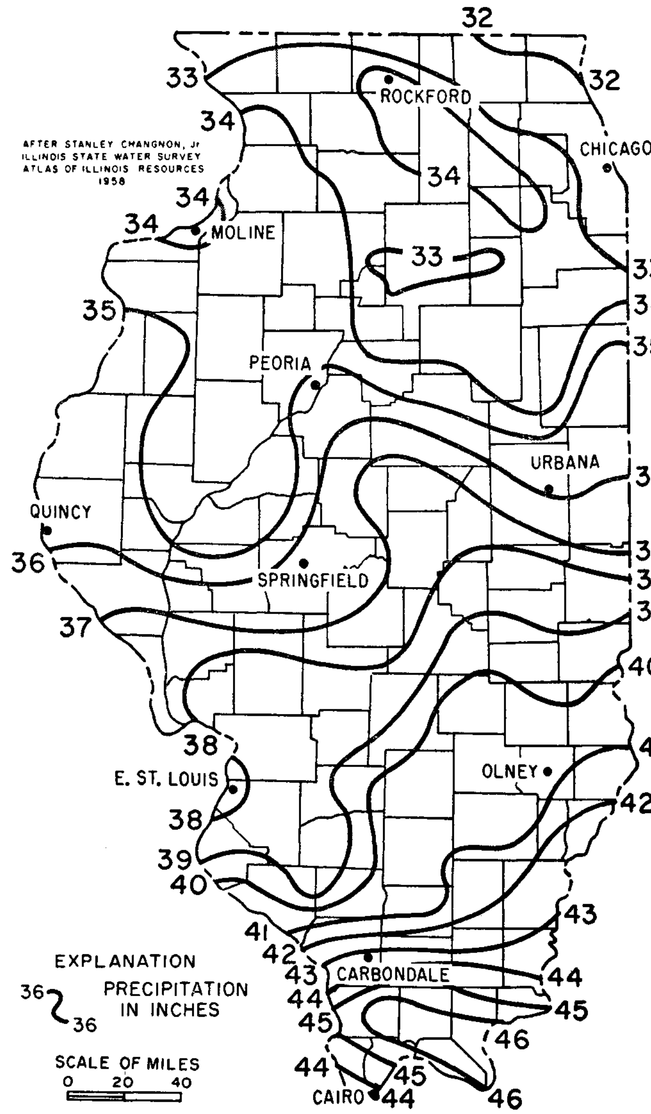


Figure 33. **Average annual precipitation in Illinois**

glaciers; only small areas in the extreme northwest corner, southwest border, and southern tip of the state lie entirely outside the glacial boundary as shown in figure 38A. Wisconsinan deposits cover only the northeastern third of the state, whereas Illinoian deposits are exposed over most of the remaining two-thirds. Kansan deposits are limited to small areas along the southwestern border of the state. Large areas in western, south-central, and southern Illinois are covered by glacial drift of Illinoian age. The drift cover is relatively thin and seldom exceeds 75 feet in thickness. In the area of the Wisconsinan glacial drift in the east-central and northern parts of Illinois, drift is thicker. The glacial drift is several hundred feet thick in deeply buried bedrock valleys such as the Mahomet Valley in east-central Illinois. Permeable outwash sand and gravel deposits partly fill bedrock valleys and exceed 100 feet thick at places. Permeable glacial deposits also occur on bedrock uplands and are commonly interbedded and overlain by till. Possibili-

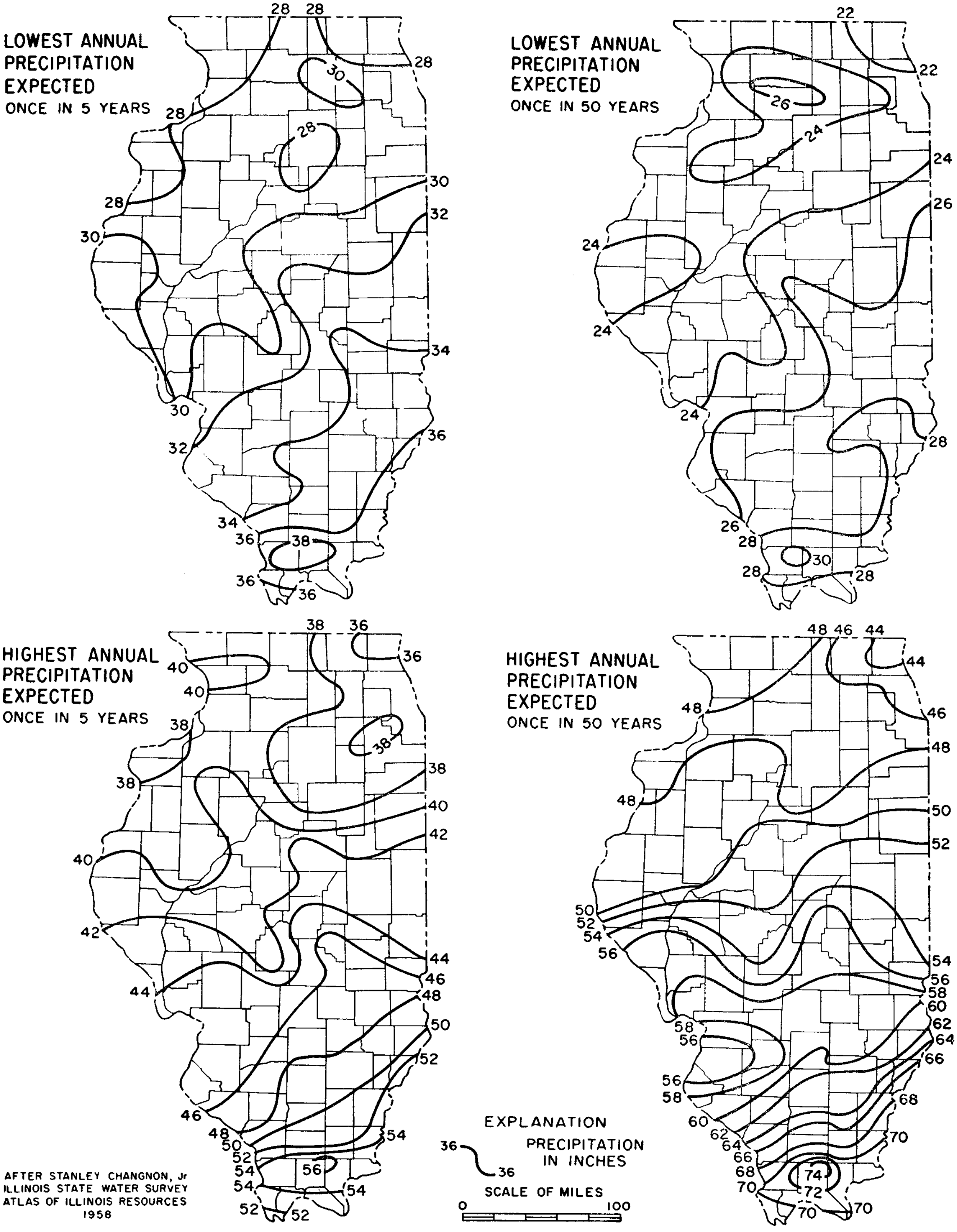


Figure 34. Frequency of annual maximum and minimum precipitation in Illinois

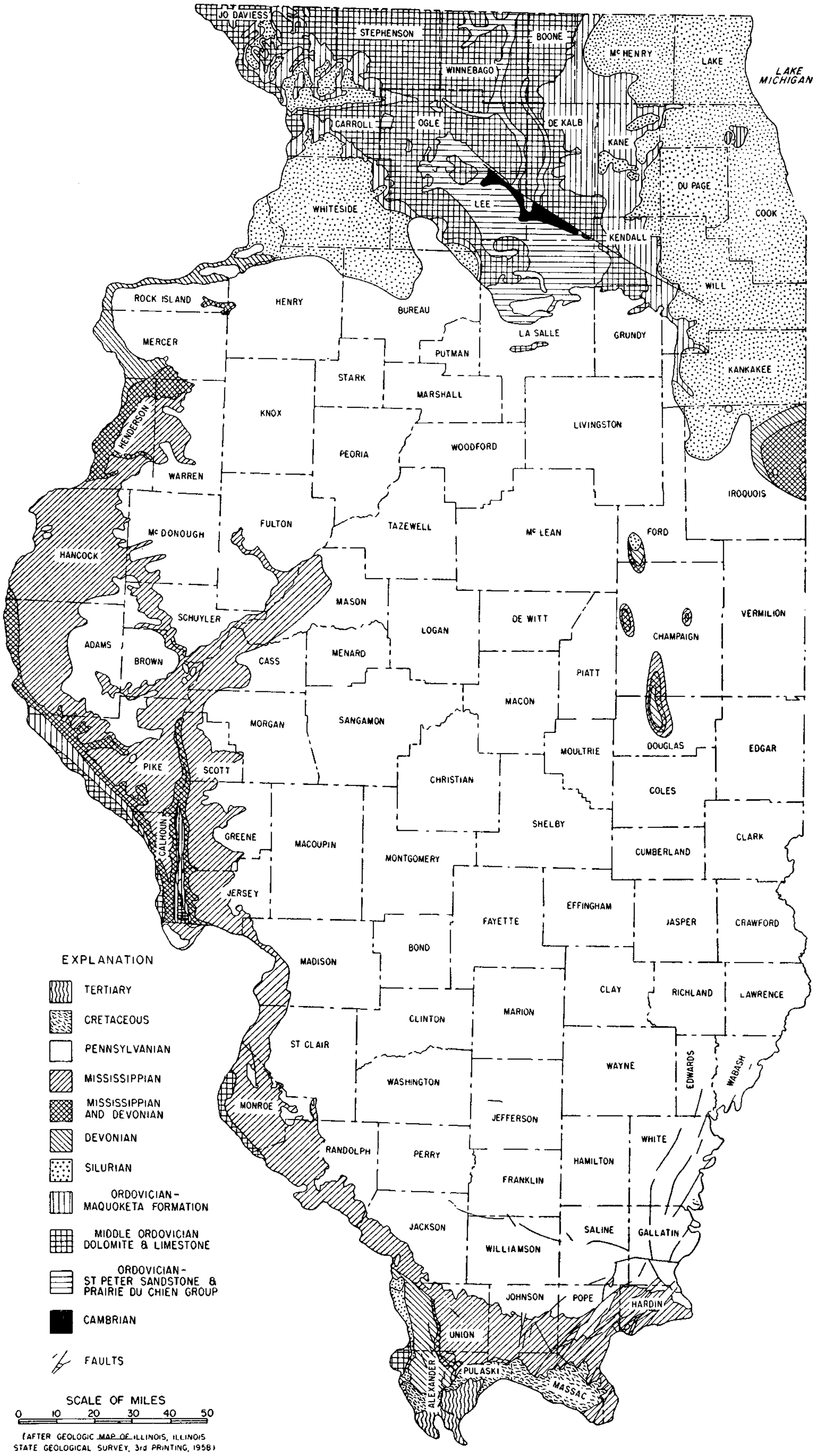


Figure 35. Generalized bedrock geology of Illinois

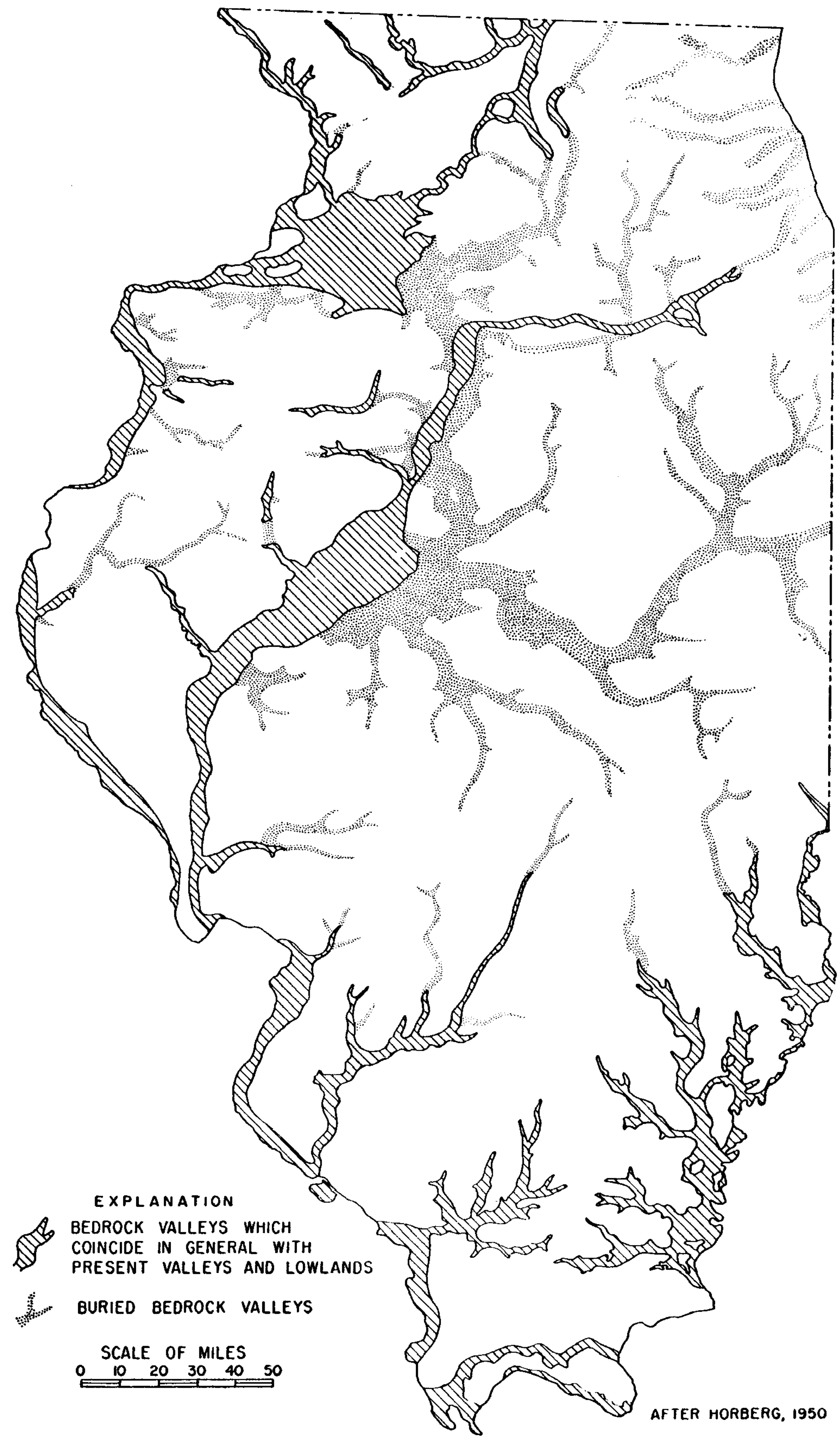


Figure 36. Major bedrock valleys in Illinois

Figure 37. Generalized bedrock topography of Illinois

ties for occurrence of sand and gravel within the glacial drift are shown in figure 39. Best possibilities are confined mostly to major bedrock valley areas; poor possibilities cover large areas in southern and western Illinois.

Figure 38 summarizes: 1) the principal glacial deposits of Wisconsinan age, including morainic ridges, lakebed areas, outwash areas, and ground moraine ridges; 2) locations of the principal morainic ridges of Illinoian age; 3) locations of the principal alluviated valleys including those which contain primarily outwash deposits and those which were impounded and contain lakebed type deposits; 4) unglaciated areas; 5) glacial boundaries; and 6) depth of the loess on uneroded topography throughout the state whether in glaciated or unglaciated regions.

Ground moraine is the predominating surface deposit of Illinois; glacial till is the principal constituent of the deposit. Till is typically a heterogeneous unsorted mixture of particles ranging in size from boulders to fine clay. At most places the till contains a high percentage of silt and clay and has a low permeability and gravity yield. In central and northeastern Illinois the accumulative thickness of several till sheets of Wisconsinan and Illinoian age often exceeds 100 feet; in southern Illinois, where the till is solely of Illinoian age, the total thickness is often 20 feet or less.

Ridges called morainic ridges or end moraines are prominent features of surface deposits in the Wisconsinan drift area in northeastern Illinois; in the Illinoian drift areas in southern Illinois they are not so prominent. The principal constituent of the morainic ridges is till; however, interbedded sand and gravel is more common in morainic ridges than in ground moraine.

Outwash plain deposits of glaciofluvial character are generally associated with moraines and lie in front of them. Normally, the coarsest-textured sediments of sand

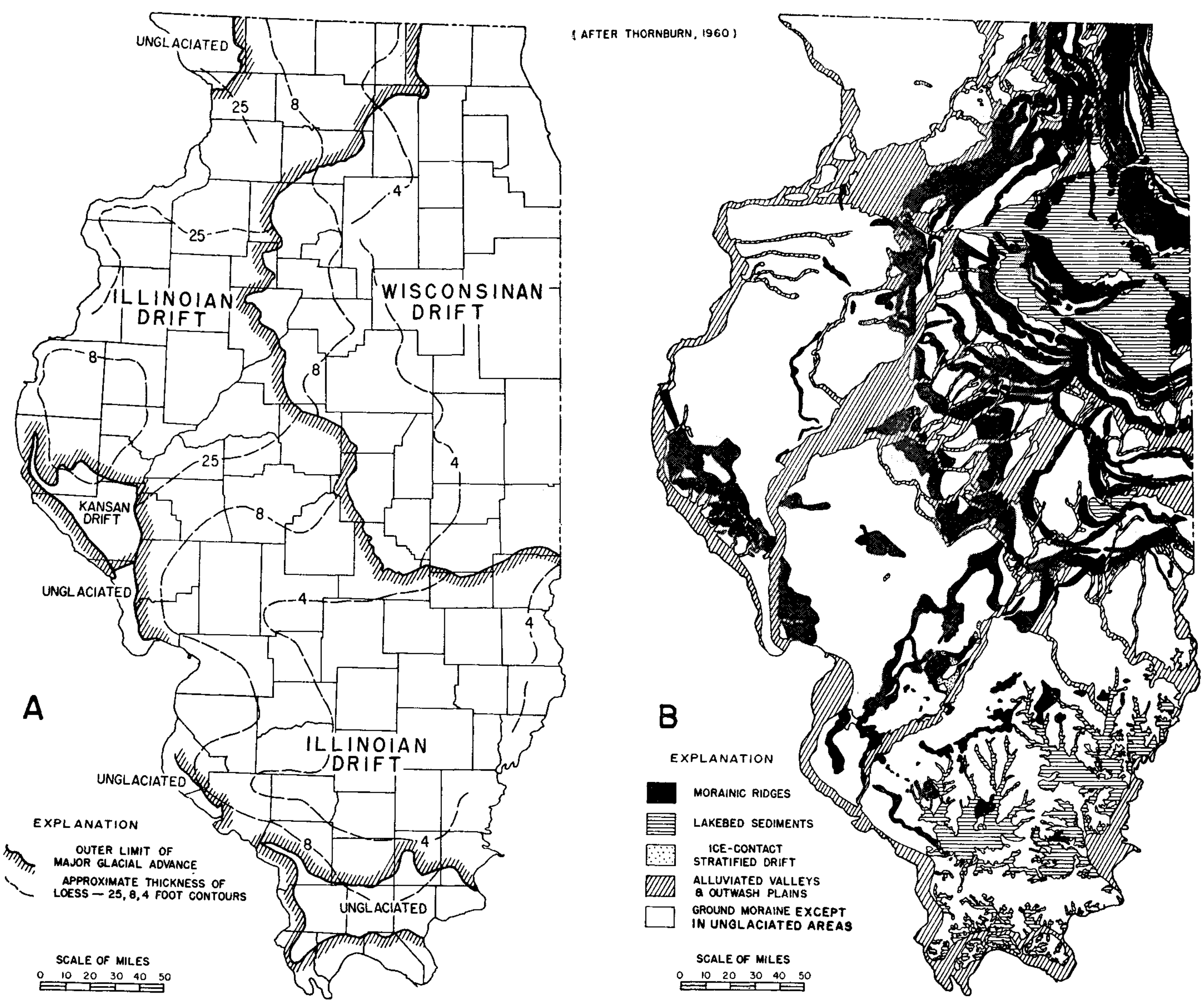


Figure 38. Outer limits of major glacial advances and thickness of loess (A), and surficial deposits (B), in Illinois

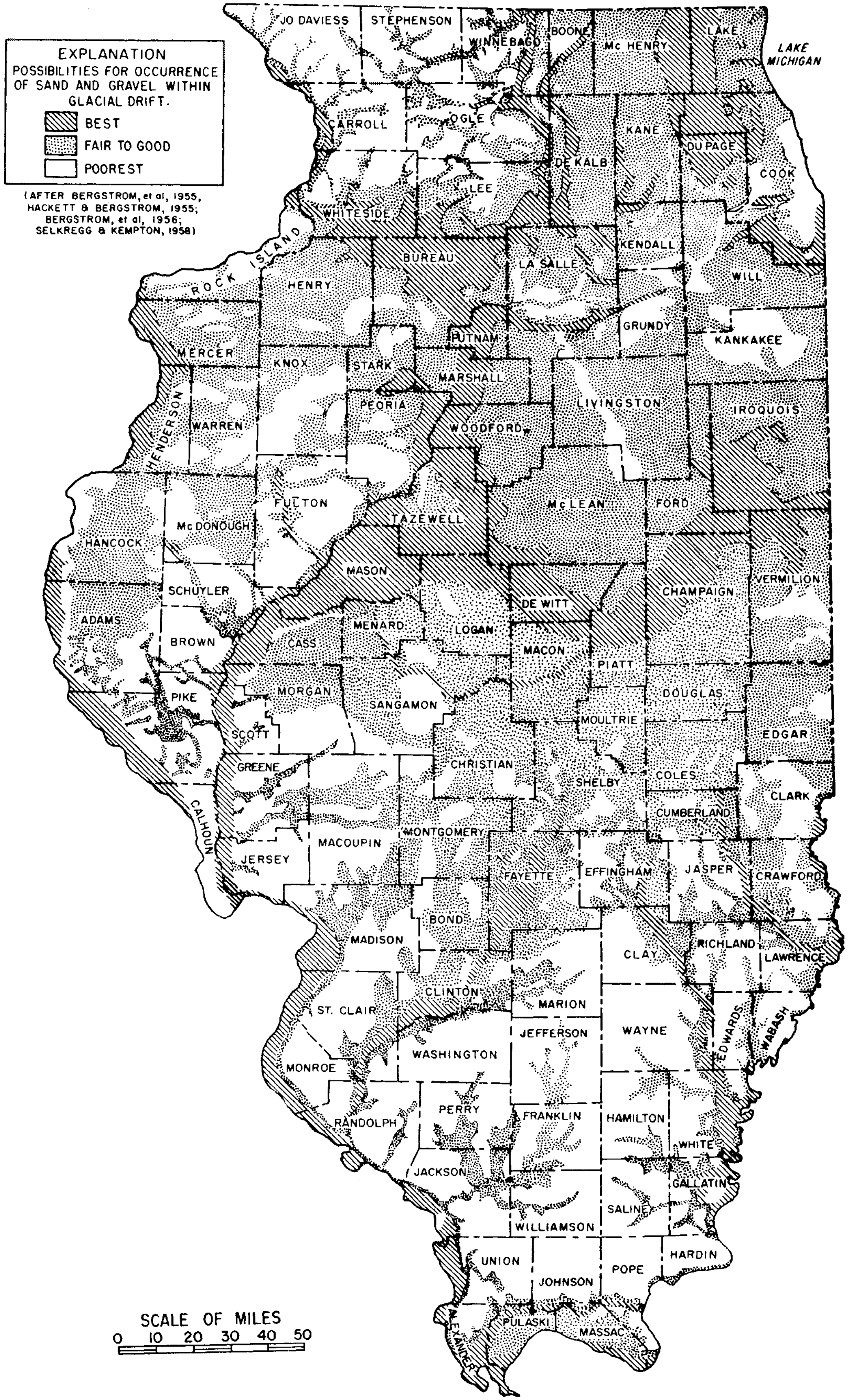


Figure 39. Possibilities for occurrence of sand and gravel aquifers in Illinois

Table 22. Generalized Geologic Column for Illinois

Era	Period	Series	Nature of deposit
	Quaternary	Pleistocene	Till, gravel, sand, silt, and clay
Cenozoic	Tertiary	Pliocene Miocene Oligocene Eocene Paleocene	Gravel, sand, and clay
Mesozoic	Cretaceous		Sand and clay
	Jurassic		Absent
	Triassic		Absent
Paleozoic	Permian		Absent
	Pennsylvanian		Shale, sandstone, siltstone, coal, clay, and limestone
	Mississippian		Limestone, sandstone, siltstone, and shale
	Devonian		Limestone, dolomite, shale, and sandstone
	Silurian		Dolomite and limestone
	Ordovician		Dolomite, limestone, sandstone, and shale
	Cambrian		Dolomite, sandstone, shale, siltstone, and limestone
Proterozoic	Keweenawan Huronian		Crystalline rocks
Archeozoic	Keewatin		

(after Thornburn, 1960)

and gravel and the thickest deposits occur immediately in front of morainic ridges; grain-size and thickness usually gradually decrease at successively greater distances from the ridges. Closely associated with the outwash plains are deposits of water-sorted material in major stream valleys. Figure 38B shows the location of the major alluviated valleys, but it does not indicate areas where the alluvium is composed chiefly of granular outwash materials. Unpublished road material maps prepared by G. E. Ekblaw, Illinois State Geological Survey, were used with figure 38 to delineate areas alluviated with permeable sand and gravel.

The most important and extensive areas of lakebed sediments are in northeastern and southern Illinois. Fine-grained plastic sediments are the principal constituent of the lakebed areas in southern Illinois. The character of the lakebed sediments in northeastern Illinois is often variable, ranging from almost clean sands along the outer edges of the lakes through sandy silts to fine textured clays in the central part of the lakebed areas. No deep lacustrine clay deposits are found.

Associated with both the Wisconsinan ground moraine and morainic ridges are deposits of ice-contact stratified drift composed of water-sorted glacial materials. These deposits are most common in the northeastern part of the state; some stratified drift is found in the Illinoian drift region.

Accumulations of wind-blown silt-size material called loess are associated with glacial deposits. Deposits of loess cover much of the western half of Illinois to a depth of 6 feet or more; much of the eastern half of the state is covered with loess varying from 2 to 4 feet in thickness. Figure 38 indicates the depth of loess on uneroded topography at contour intervals of 4, 8, and 25 feet.

There is a close relation between the map of surficial deposits and the maps of physiographic divisions (see figure 40) and of bedrock. The state of Illinois lies mostly within the Central Lowland Physiographic Province and is essentially a prairie plain. The relief over most of the state is moderate to slight; large scale relief features are generally absent. The Wisconsin Driftless Section in northwestern Illinois and much of the Ozark Plateaus, the Interior Low Plateaus, and the Coastal Plain in extreme southern and southwestern Illinois lie outside of the glacial boundary. Normally local topography of ground moraine areas is of the order of 30 feet or less; in southern Illinois the topography is controlled largely by the underlying bedrock. The local relief in morainic areas seldom exceeds 50 feet although in some locations differences in elevation may be as much as 100 feet. Topographically the outwash plains are nearly level and the lakebeds have slight to moderate local relief.

The Great Lake Section of the Central Lowland Province is an area in which bold moraines encircle the Lake Michigan Basin and distinguish it from the nearly level to gently undulating till plains to the south and west. The Chicago Lake Plain is characterized by a nearly flat surface sloping gently toward Lake Michigan. The flatness of the plain is interrupted only by the presence of low sandy beach ridges and a few morainic remnants. The Wheaton Morainal Country is an area in which young rolling morainal topography is best developed. The topography is the roughest found in any part of the state covered by Wisconsinan drift.

The Till Plains Section, covering about four-fifths of the state, is characterized by broad till plains which are uneroded or in the youthful stage of erosion. The Kankakee Plain is generally described as a lake plain because of its nearly flat topography. In the Kankakee River Valley dolomitic bedrock occurs at or very near the surface. Most of the region is poorly drained by low gradient streams. Throughout the Bloomington Ridged Plain are

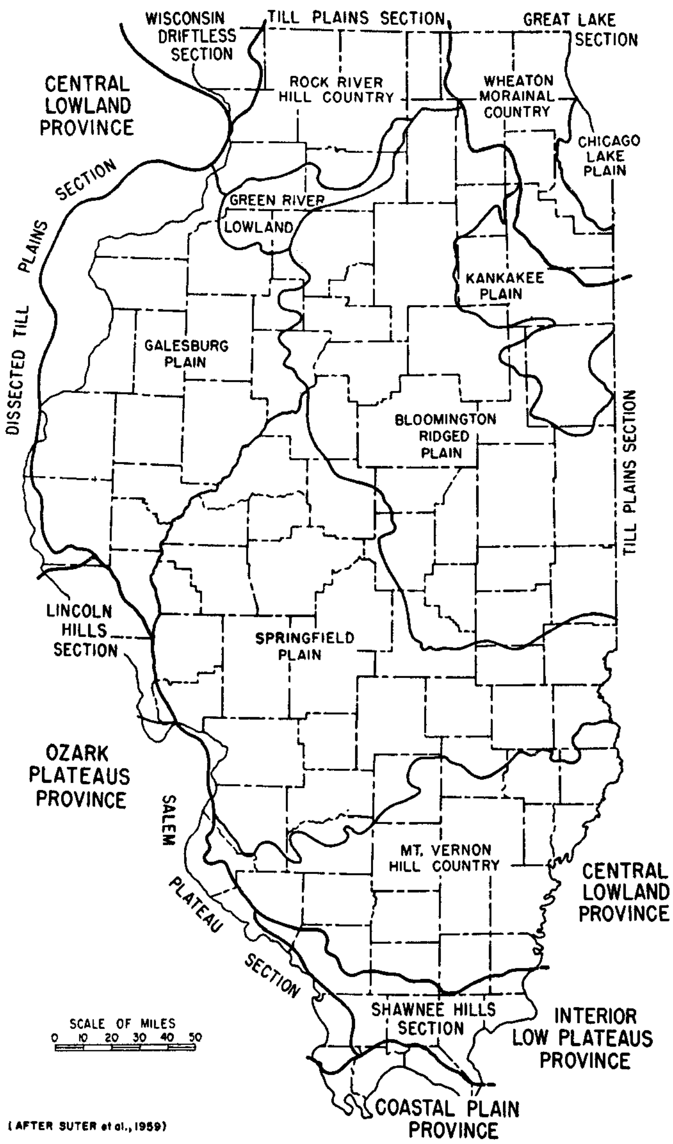


Figure 40. Physiographic divisions of Illinois

found low broad morainic ridges alternating with intervening wide stretches of relatively flat or gently undulating till plains. Broad outwash plains are associated with the Shelbyville, Champaign, and Bloomington moraines. The Rock River Hill Country is characterized by rather gently rolling topography. The Green River Lowland is described as a broad alluviated valley and is a modified outwash plain related to the Bloomington moraine. Throughout much of the Galesburg Plain the topography is relatively level to gently undulating except where dissection has proceeded along the major river valleys. The Springfield Plain differs from the Galesburg Plain in that the drainage systems, while well developed, are not as deeply entrenched. Within the Mt. Vernon Hill Country the topography is controlled chiefly by the character of the underlying bedrock. Most streams have broad valleys with low gradients.

The topography of the Dissected Till Plains Section of the Central Lowland Province is controlled mostly by the ruggedness of the underlying bedrock. Thus, the physiographic characteristics of the region are similar in many respects to the southern part of the Mt. Vernon Hill Country. The topography of the Wisconsin Driftless Section of the Central Lowland Province is controlled by the bedrock which has been maturely dissected. The Mississippi River and its tributaries have carved valleys to depths of 150 to 300 feet below the general upland surface. The principal physiographic feature of the Lincoln Hills Section of the Ozark Plateaus Province is the mature ridge which forms the watershed between the Illinois and Mississippi Rivers. The northern part of the Salem Plateau Section of the Ozark Plateaus Province has been glaciated and is covered by a thin layer of drift, whereas the southern part lies south of the glaciated area. Sinkhole topography has developed on the limestone strata underlying the northern part; the southern part has rugged topography resulting from a well developed drainage pattern. The rugged topography of the Shawnee Hills Section of the Interior Low Plateaus Province is largely the result of variations in composition of the bedrock. The Coastal Plain Province comprises predominantly two areas; the alluviated valleys of the Cache, Ohio, and Mississippi Rivers, and the upland areas surrounded by them. The upland area topography is not as rugged as the Shawnee Hills Section.

In the prairie areas, woodlands are widely scattered and are generally limited to those bottomlands along the major streams. Woodland areas generally constitute but a few percent of acreage in Illinois except in the extreme southern and southwestern parts where woodlands occupy more than 20 percent of many areas.

Relation Between Ground-Water Runoff and Basin Characteristics

Flow-duration curves can be used in making comparisons of the flow characteristics of streams (Cross and Hedges, 1959). The shape of the flow-duration curve is governed in part by the water-yielding properties and areal extent of the unconsolidated and consolidated deposits within a drainage basin. The more nearly horizontal the curve, the greater are the values of the water-yielding properties and/or the areal extent of deposits. Thus, the shape of the flow-duration curve is an index of the effects of geology of a basin on streamflow.

Grain-size frequency-distribution curves (see Dapples, 1959) are somewhat analogous to flow-duration curves in that their shapes are indicative of the water-yielding properties of deposits (personal communication, J. E. Hackett, Illinois State Geological Survey). A measure of the degree to which all the grains approach one size, and therefore the slope of the grain-size frequency-distribution curve, is the sorting. One parameter of sorting is obtained by the ratio (Pettijohn, 1949) $(D_{25}/D_{75})^{1/2}$, where D_{25} is the size that has 25 percent larger and 75 percent smaller grains in the distribution and D_{75} is the size that has 75 percent larger and 25 percent smaller grains in the distribution.

Because geology and therefore grain-size frequency distribution affects streamflow to a great degree, the parameter selected to describe the slope of the flow-duration curve is the ratio $(Q_{25}/Q_{75})^{1/2}$, where Q_{25} is the streamflow equalled or exceeded 25 percent of the time and Q_{75} is the streamflow equalled or exceeded 75 percent of the time. Streamflow data (for most basins) over the range Q_{25} to Q_{75} describe straight lines. Ratios for the 109 drainage basins based on flow-duration curves are given in table 21.

The characteristics of the 109 drainage basins were determined by superimposing figure 28 over maps such as those shown in figures 35 to 40. The relations between ground-water runoff during years of near, below, and above normal precipitation, the ratios $(Q_{25}/Q_{75})^{1/2}$, and the basin characteristics listed in table 23 were studied by a quantitative method of statistical analysis.

Basins were segregated into categories depending upon the characteristics of the basins. It was found that most basins could be classified into one of the 16 multiple basin characteristic categories listed in table 24. Ground-water runoffs and the ratios for wells in each of the 16 categories were tabulated in order of magnitude, and frequencies were computed by the Kimball (1946) method. Values of ground-water runoffs and ratios were then plotted against percent of basins on logarithmic probability paper as illustrated in figures 41 and 42. The range (10 and 90 percent frequencies) and medium (50 percent frequency) of ground-water runoff and the ratio $(Q_{25}/Q_{75})^{1/2}$ were determined for each category from the frequency graphs and are given in table 25.

Ground-water runoff is greatest in categories 3, 6, and 8; ground-water runoff is least in category 7. Drainage basin characteristics common to categories 3 and 6 are: glaciated, impermeable bedrock, thick drift, bedrock valleys, fair or poor possibilities for occurrence of sand and gravel within drift, considerable surface sand and gravel of limited areal extent, ground moraine or morainic ridges, slight to moderate topographic relief, slight to moderate stream gradient, and little forest and woodland. Category 8 drainage basin characteristics are: unglaciated, permeable bedrock, thin unconsolidated deposit, bedrock valleys, some surface sand and gravel of limited areal extent, rugged topography, steep stream gradient, and considerable forest and woodland. The differences in ground-water runoff between categories 3 and 6 and 8 are small and may not be significant; however, ground-water runoff appears to be greatest in category 8. Category 7, in which ground-water runoff is least, has the same drainage basin characteristics as categories 3 and 6

Table 23. Basin Characteristics

1. Surface deposits are predominantly ground moraine
2. Surface deposits are predominantly morainic ridges
3. Surface deposits are predominantly lakebed sediments
4. Alluviated valley or outwash plain surface deposits are present
5. Considerable surface sand and gravel are present
6. Relatively impermeable bedrock underlies unconsolidated deposits
7. Permeable bedrock underlies unconsolidated deposits
8. Basin is glaciated
9. Basin is unglaciated
10. Unconsolidated deposits are thick (commonly exceed 50 feet)
11. Unconsolidated deposits are thin
12. Possibility for occurrence of sand and gravel within glacial drift is good
13. Possibility for occurrence of sand and gravel within glacial drift is fair
14. Possibility for occurrence of sand and gravel within glacial drift is poor
15. Major bedrock valleys are present
16. No major bedrock valleys are present
17. Age of exposed glacial deposits is Wisconsinan
18. Age of exposed glacial deposits is Illinoian
19. Age of exposed glacial deposits is Kansan
20. Topographic relief is slight to moderate
21. Topographic relief is rugged
22. Stream gradient is slight to moderate
23. Stream gradient is steep
24. Forest and woodland area is small
25. Forest and woodland area is considerable
26. Normal precipitation is less than 38 inches
27. Normal precipitation is greater than 38 inches
28. Growing season less than 175 days
29. Growing season greater than 175 days

Table 24. Selected Basin Categories

Category	Basin characteristics (see table 23)
1	1, 4, 6, 8, 10, 13 or 14, 15, 17, 18 or 19, 20, 22, 24, 26 or 27
2	1, 4, 7, 8, 10, 13 or 14, 15, 17, 18 or 19, 20, 22, 24, 26 or 27
3	1, 5, 6, 8, 10, 13 or 14, 15, 17, 18 or 19, 20, 22, 24, 26 or 27
4	2, 4, 6, 8, 10, 13 or 14, 15, 17, 18 or 19, 20, 22, 24, 26 or 27
5	2, 4, 7, 8, 10, 13 or 14, 15, 17, 18 or 19, 20, 22, 24, 26 or 27
6	2, 5, 6, 8, 10, 13 or 14, 15, 17, 18 or 19, 20, 22, 24, 26 or 27
7	3, 4, 6, 8, 10, 13 or 14, 15, 17, 18 or 19, 20, 22, 24, 26 or 27
8	4, 7, 9, 11, 15, 21, 23, 25, 26 or 27
9	4, 6, 9, 11, 15, 21, 23, 25, 26 or 27
10	1 or 2, 4, 6, 8, 10, 12, 15, 17 or 18, 20, 22, 24, 26 or 27
11	1 or 2, 4, 6, 8, 10, 13 or 14, 16, 17 or 18, 20, 22, 24, 26 or 27
12	1 or 2, 4, 6, 8, 10, 13 or 14, 15 or 16, 18, 20, 22, 24, 26 or 27
13	1 or 2, 4, 6, 8, 10, 13 or 14, 15 or 16, 19, 21, 23, 25, 27
14	1 or 2, 4, 6, 8, 10, 13 or 14, 15 or 16, 17, 20, 22, 24, 26 or 27
15	1 or 2, 4, 6, 8, 10, 13 or 14, 15 or 16, 17, 18 or 19, 20, 22, 24, 26, 28
16	1 or 2, 4, 6, 8, 10, 13 or 14, 15 or 16, 17, 18 or 19, 20, 22, 24, 27, 29

except less surface sand and gravel and lakebed sediments. There seems to be little difference in ground-water runoff between categories 3 and 6; differences in ground-water runoff between ground moraine and morainic ridges are small and, as shown in figure 41, morainic ridges may yield slightly more water to streams than ground moraine.

Ground-water runoff is significantly greater in similar basins when the bedrock is permeable than it is when the bedrock is relatively impermeable as shown in figure 42. Ground-water runoff increases appreciably as the amount of surface sand and gravel increases; in fact, surface sand and gravel deposits control ground-water runoff to a great extent.

The ratio $(Q_{25}/Q_{75})^{1/2}$ is controlled largely by the areal

The greater the ratio, the less the areal extent and thickness and/or hydraulic properties. The least ratios are for basins underlain by permeable bedrock having large areal

Table 25. Annual Ground-Water Runoff and Basin Characteristics

	Annual ground-water runoff *(cfs /sq mi)*						Ratio $(Q_{25}/Q_{75})^{1/2}$		
	near*		below*		above*				
Category	range	median	range	median	range	median	range	median	Basin Characteristics
1	0.13 to 0.36	0.22	0.05 to 0.18	0.09	0.19 to 0.57	0.33	2.7 to 5.2	3.7	Glaciated, relatively impermeable bedrock, thick drift (commonly exceeding 50 feet), bedrock valleys; fair or poor possibility for occurrence of sand and gravel within drift, some surface sand and gravel of limited areal extent, ground moraine, slight to moderate stream gradient; little forest and woodland
2	0.24 to 0.38	0.28	0.17 to 0.22	0.19	0.32 to 0.59	0.40	1.5 to 1.9	1.7	Same as category 1 except permeable bedrock
3	0.20 to 0.38	0.29	0.09 to 0.18	0.14	0.35 to 0.56	0.44	2.8 to 3.8	3.2	Same as category 1 except considerable surface sand and gravel
4	0.15 to 0.38	0.23	0.04 to 0.17	0.08	0.29 to 0.62	0.43	2.7 to 6.2	4.0	Same as category 1 except morainic ridges
5	0.17 to 0.39	0.25	0.05 to 0.18	0.10	0.38 to 0.70	0.52	1.7 to 4.6	2.8	Same as category 1 except morainic ridges and permeable bedrock
6	0.17 to 0.43	0.27	0.07 to 0.22	0.14	0.38 to 0.72	0.52	2.0 to 5.0	3.1	Same as category 1 except morainic ridges and considerable surface sand and gravel
7	0.06 to 0.20	0.11	0.01 to 0.12	0.04	0.12 to 0.49	0.24	3.2 to 13	6.5	Same as category 1 except lakebed sediments
8	0.25 to 0.43	0.32	0.13 to 0.31	0.19	0.41 to 0.58	0.48	1.4 to 2.4	1.7	Unglaciated, permeable bedrock, thin unconsolidated deposits, bedrock valleys, some surface sand and gravel of limited areal extent, rugged topography, steep stream gradient, considerable forest and woodland
9	0.14 to 0.42	0.27	0.11 to 0.22	0.17	0.20 to 0.54	0.42	1.5 to 6.0	4.7	Same as category 8 except impermeable bedrock
10		0.30		0.10		0.49		3.9	Same as category 1 or 4 except major buried bedrock valleys
11		0.29		0.07		0.48		4.2	Same as category 1 or 4 except no major buried bedrock valleys
12		0.24		0.14		0.41		2.8	Same as category 1 or 4 except drift of Illinoian age
13		0.23		0.11		0.35		3.2	Same as category 1 or 4 except drift of Kansan age, rugged topography, steep stream gradient, and considerable forest and woodland
14		0.23		0.09		0.46		3.7	Same as category 1 or 4 except drift of Wisconsinan age
15		0.20		0.08		0.41		3.2	Same as category 1 or 4 except normal precipitation less than 38 inches and growing season less than 175 days
16		0.30		0.13		0.45		3.4	Same as category 1 or 4 except normal precipitation greater than 38 inches and growing season greater than 175 days

****Words indicate data are for years of near, below, and above normal precipitation***

extent. The presence of surface sand and gravel deposits tends to reduce ratios; however, the limited areal extent and thickness of most sand and gravel deposits results in small reductions. The ratio is highest in basins containing lakebed sediments with little or no surface sand and gravel. Thus, the size and hydraulic properties of the ground-water reservoir in connection with streams determines in large part ground-water runoff during extended dry periods. The greater the size and hydraulic properties of deposits the greater low flows in streams, and the less the difference between ground-water runoff during wet and dry periods.

A comparison of ground-water runoffs and ratios for categories 10 and 11 indicates that the presence of major buried valleys with accompanying greater drift thicknesses tends to slightly increase ground-water runoff only during years of below normal precipitation. Thick layers of till of low permeability in buried bedrock valley areas greatly retard the vertical movement of water toward streams only partially penetrating the upper part of the drift. Vertical movement of water toward streams is greatest during dry periods when the difference between the water table and the piezometric surface of deeply buried deposits is greatest. Data for categories 12, 13, and 14 suggest that the age of exposed glacial drift has little influence on ground-water runoff. Ground-water runoff may be slightly greater in areas where Illinoian Drift is exposed than in areas where Wisconsinan Drift is exposed.

A comparison of ground-water runoffs for categories 15 and 16 indicates that ground-water runoff increases as precipitation increases. As shown in figure 33, precipitation increases from north to south in Illinois. The growing season, and therefore frost-free period, and the mean annual temperature also increase from north to south. Thus, ground-water runoff increases with increases in precipitation from north to south despite the fact that evapotranspiration also increases from north to south.

There are probably some important relations between ground-water runoff and basin characteristics other than those discovered during the present study and discussed above. The other relations may be too subtle and complex to permit detection. The profound influence of geology on ground-water runoff is apparent.

Annual ground-water runoff from ungaged basins may be estimated with reasonable accuracy with data in table 25 and information on the characteristics of the ungaged basins, as follows: 1) The generalized characteristics of the ungaged basin are listed and one of the general categories 1 through 9 in table 24 is assigned to the ungaged basin. 2) The range and median ground-water runoffs for the selected category are noted. 3) Annual ground-water runoff is then estimated largely on the basis of median ground-water runoff, taking into consideration the range of ground-water runoff, detailed basin characteristics, and data for categories 10 through 16 pertaining to the influence of detailed basin characteristics on ground-water runoff.

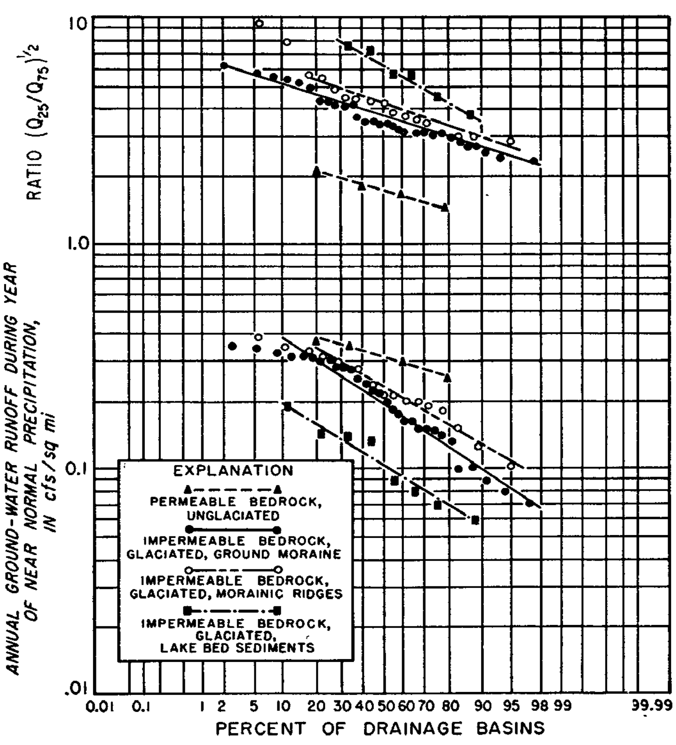


Figure 41. Relation between annual ground-water runoff and ratios and character of surface deposits

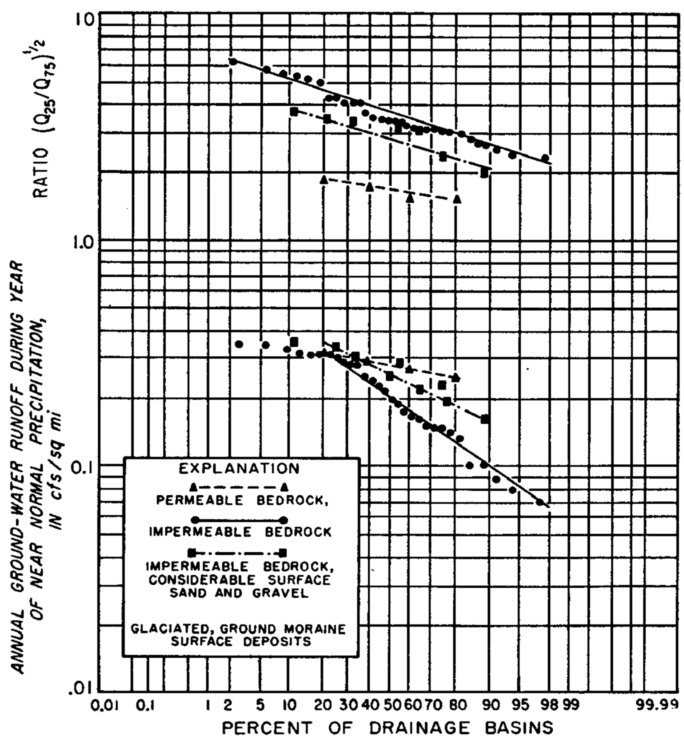


Figure 42. Relation between annual ground-water runoff and ratios and character of bedrock

Panther Creek Basin

Hydrologic and ground-water budgets were prepared as described earlier for the small drainage basin of Panther Creek in north-central Illinois (Schicht and Walton, 1961) to determine how ground-water runoff varies throughout the year and is affected by evapotranspiration losses. The general characteristics of Panther Creek Basin are those of categories 1 and 4 in table 24.

Data in figures 43 and 44 indicate that ground-water runoff generally is at a maximum during spring and early summer months, and is least in late summer and fall months. More than half of annual ground-water runoff occurs during the first six months of the year. Annual ground-water runoff depends upon antecedent ground-water stage conditions as well as the amount and distribution of annual precipitation. Ground-water runoff is greatly reduced during extended dry periods. Ground-water runoff amounted to 33, 73, and 38 percent of streamflow in 1951, 1952, and 1956, respectively.

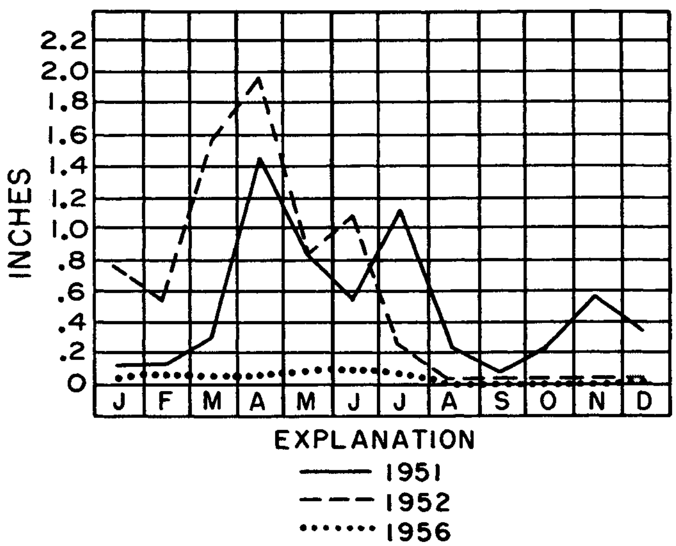


Figure 43. Monthly ground-water runoff, Panther Creek Basin

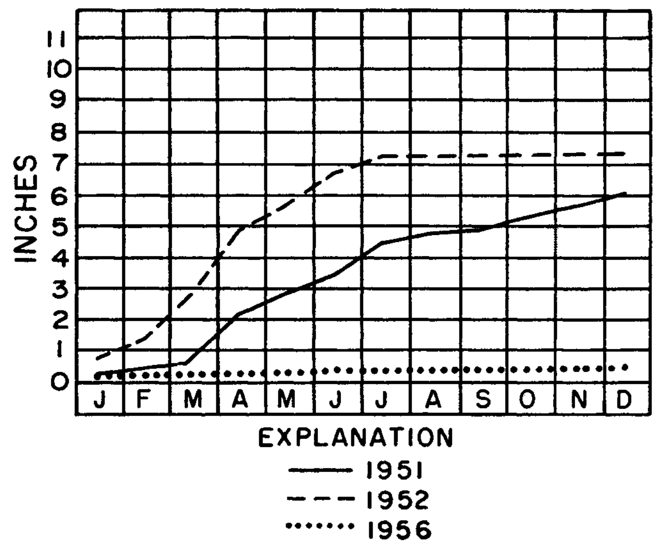


Figure 44. Cumulative monthly ground-water runoff, Panther Creek Basin

RELATION BETWEEN RECHARGE RATES AND GROUND-WATER RUNOFF

Large areas in Illinois are covered by glacial drift commonly exceeding 50 feet in thickness. Sand and gravel and bedrock aquifers are often deeply buried, and in many areas recharge to the aquifers is derived from vertical leakage through thick layers of till having a low permeability. Vertical leakage with maximum vertical hydraulic gradients is often much less than recharge to surface deposits. Because many aquifers in Illinois are deeply buried, not all ground-water runoff can be diverted into cones of depression. Even under heavy pumping conditions there is some lateral as well as vertical movement of water in surface deposits. Also, some ground-water runoff is bank storage which cannot be entirely diverted into cones of depression at some distance from streams.

Precipitation and, therefore, ground-water runoff and recharge are unevenly distributed throughout the year. There are periods of time during the wet spring months when the rate of recharge to surface deposits greatly exceeds, at least temporarily, maximum rates of vertical leakage of water to cones of depression in deeply buried aquifers.

The amount of recharge to many aquifers depends upon the coefficient of vertical permeability of till, the saturated thickness of till, the area of diversion of the well or well field, and the difference between the head in the aquifer and that in surface deposits above the till. The area of diversion depends in part upon the hydraulic properties and areal extent of the aquifer.

The amounts of recharge from vertical leakage of water through till to deeply buried dolomite aquifers in DuPage County (see section "Dolomite Aquifer in DuPage County") were compared with ground-water runoffs under natural conditions from basins having characteristics similar to those of the areas within cones of depression.

The average ratio between recharge to cones of depression in deeply buried aquifers and ground-water runoff was about 60 percent.

RELATION BETWEEN GROUND-WATER RUNOFF AND POTENTIAL OR PRACTICAL SUSTAINED YIELDS OF AQUIFERS

Before ground-water resources can be evaluated and the consequences of the utilization of aquifers can be forecast, recharge to aquifers must be appraised. Data on ground-water runoff can be useful in estimating recharge to aquifers; however, studies indicate that no simple relation exists between ground-water runoff and the potential recharge or practical sustained yields of aquifers. In most parts of Illinois not all ground-water runoff can be diverted into cones of depression in deeply buried aquifers.

Cursory consideration may suggest that a map showing the distribution of ground-water runoff could be used directly to evaluate the potential yield of ground-water reservoirs. However, areas of high or low ground-water runoffs do not necessarily have to be areas of high or low potential yield. Ground-water runoff can be high largely because of the presence of considerable surface sand and gravel. But because of their greatly limited areal extent and thickness, the surface sand and gravel deposits may be capable of yielding only small amounts of water to wells and well fields. Ground-water runoff can be low because of the lack of surface sand and gravel and the presence of surface lakebed sediments. However, deeply buried sand and gravel deposits, extending beyond basin boundaries, may be capable of yielding moderate amounts of water to wells and well fields. Thus, a map showing the distribution of ground-water runoff can be misleading if used directly to delineate favorable areas for development of ground-water resources.

It is apparent that no simple relation exists between the amount of ground-water runoff and the potential or practical sustained yields of aquifers. However, with sound professional judgment, rough estimates of recharge to aquifers which will be useful in making preliminary evaluations of ground-water resources can be made from ground-water runoff data and hydrogeologic data. If recharge by induced infiltration of surface water is excluded, recharge to most deeply buried aquifers will be less than ground-water runoff under natural conditions. If decreases in evapotranspiration due to lowering of water levels under heavy pumping conditions and recharge by induced infiltration of surface water are excluded, the maps of ground-water runoff can be useful in setting upper limits on the potential yields of ground-water reservoirs in Illinois.

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