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WRC RESEARCH REPORT

GROUND-WATER RECHARGE RATES IN MINNESOTA AS RELATED TO PRECIPITATION

Roman Kanivetsky and Barbara Palen
Minnesota Geological Survey
University of Minnesota

REPORT
Project No. B--153

WATER RESOURCES CENTER

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Abstract

Title: Ground-Water Recharge Rates in Minnesota as Related to Precipitation

OWRT Project Number B-153-Minn
Contract Number: 14-34-0001-0268

Principal Investigators: Matt Walton and Roman Kanivetsky

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Quantitative estimates of recharge rates and storage coefficients in sufficient detail are essential to any meaningful state water-resources management, policies, plans and legislation. A methodology was developed to derive quantitative estimates of ground-water recharge rates and storage coefficients for unconfined aquifers in the various hydrologic regions of Minnesota.

Recharge estimates are based on information from ground-water hydrographs.

Statistical analysis of ground-water level fluctuations and their correlation with precipitation, evaporation, stream discharge were evaluated. Harmonic analysis and autoregression analysis of water levels were done.

Storage coefficients were estimated from pumping test data, soil moisture method and empirical equations.

Ground-water recharge rates were determined by combining data on ground-water level fluctuations and storage coefficient evaluation for unconfined aquifers in Minnesota. This work is fundamental to the ultimate goal of an interactive water resources model of the state.

Key words; Evaporation/ groundwater/ hydrogeology/ recharge/ permeability/ precipitation/ storage coefficient.

OUTLINE OF CONTENTS

FOREWORD

This report includes a compilation of all data relating to the present study with the exception of ground-water level correlations and determination of ground-water recharge rates. The latter determination depends on final synthesis of the ground-water data, which are extremely voluminous and are still in the process of being assembled and analyzed as of the date of this report. The present report, plus the ground-water data and determination of recharge rates, will eventually be reviewed internally and edited by professional staff of the Minnesota Geological Survey and will be published in its Report of Investigations series.

GROUNDWATER RECHARGE RATES IN MINNESOTA AS RELATED TO PRECIPITATION

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GROUND-WATER RECHARGE RATES IN MINNESOTA AS RELATED TO PRECIPITATION

by

Roman Kanivetsky and Barbara Palen

I. INTRODUCTION

1. Overview

The state of Minnesota has an abundance of ground water, but its distribution varies considerably throughout the state. In 1975-76 much of the United States, particularly Minnesota and other midwestern states, was affected by prolonged periods of drought. The drought forced many people, especially in the western and central parts of the state, to develop additional ground-water sources. Currently, more than two thirds of the state's population depends on ground water, a dependence which increases each year.

Another factor affecting ground water is man's activities such as waste disposal, land management, and urbanization. The effect of man's activities on hydrologic systems may be gradual and not recognized until considerable time has passed.

Widely publicized ground-water contamination cases and shortages of water supply increased the public awareness and created concern on the part of governmental officials for more adequate information on ground-water resources.

Two important quantitative parameters of ground-water resources are recharge rates and storage coefficients. These two parameters are essential for quantifying ground-water recharge in specific areas (required for environmental impact statements), assessing potential impact of aquifer development, evaluating contamination problems, and for developing protection and use programs.

No data are availableto determine recharge rates for various hydrogeologic domains. This report provides quantitative results for various unconfined aquifers in Minnesota, which also can be useful for other areas in the midwestern U.S. that have similar hydrogeologic conditions.

2. Purpose and Objectives of the Investigation

The purpose of the project was to develop data that can serve as the foundation for legal, social and economic aspects of water development and protection. Results of the investigation will allow policy makers and government officials to make decisions regarding ground-water problems on the basis of scientific information.

Specific objectives of the project were to:

a. Analyze existing data on precipitation, ground-water levels, soil and air temperatures, summer evaporation, winter depth of snow and stream discharge records;

b. Statistically analyze ground-water levels under water-table conditions;

c. Evaluate storage coefficients; and

d. Appraise recharge rates for unconfined aquifers.

3. Acknowledgments

This project was supported by funds provided by the University of Minnesota Water Resources Research Center from support by the Office of Water Research and Technology, U.S. Department of the Interior (P.L. 88-379). Additional support was provided by the Minnesota Geological Survey.

Special appreciation is due to the U.S. Geological Survey (USGS) District Office in St. Paul for providing access to data on observation wells, pumping tests, ground-water levels, and boundaries of water-table conditions in bedrock aquifers. Special thanks go to Dennis Woodward and Donald Siegel of the USGS. The Minnesota Department of Natural Resources provided observation well water-level data for Minnesota. We would like to thank Donald Slack and Curtis Larson from the Department of Agricultural Engineering, University of Minnesota, for providing data on soil moisture, water levels and temperatures at the site of Andover, Anoka County.

II. METHODOLOGY

Ground-water level rise after rainfall and snowmelt in water-table aquifers approximately reflects the recharge rate in Minnesota. Recharge to the water-table aquifers occurs mainly during the springs and causes water levels to rise significantly. During the summer, evaporation eliminates much of the available precipitation, and recharge to the ground-water system is negligible. In the fall, when evapotranspiration decreases, some recharge may result from precipitation and water levels may rise slightly. During the winter period, water levels decline when precipitation is stored on the land surface as ice and snow and typically water levels reach a low point before the spring.

Recharge rates can be expressed through the equation:

$$W = s\frac{\Delta h}{\Delta t};$$

where: W = rate of recharge (in inches/year),

s = storage coefficient or specific yield;

for water-table conditions (dimensionless),

Δh = rise of water level in the well

Δt = time (years).

However, this basic formula does not take into account outflow of ground water toward the discharge areas, such as rivers, lakes, and streams, and calculated values of recharge will be underestimated. Outflow values can be approximated on the basis of a known outflow period, such as a previous winter when there is no recharge due to frozen ground conditions. The equation then could be written (Fig. 1):

$$W = s\frac{\Delta h + \Delta Z}{t};$$

where: ΔZ = water level decline due to outflow (inches).

Water level rise also is a function of the difference between inflow and outflow, and the water-table surface is generally convex and not horizontal. Therefore, the outflow in any given cross section is greater toward the lower part of the ground-water flow than the upper part. Closer to the discharge area (river, lake, stream) the difference between inflow and outflow is greater and the recharge value from precipitation is insignificant. The ground-water regime near streams is closely related to the regime of surface water and generally does not relate to precipitation.

Wells located on slopes show little relationship between ground-water levels and precipitation. Here the water level rise is mostly related to inflow from the recharge area and possibly some recharge from the deep aquifers. Discharge of water-table aquifers on slopes occur through springs or underground to surface water bodies.

Wells located on river terraces usually do not have significant fluctuation because of the close proximity to surface water, or because stabilization of ground-water levels occurs due to shallowness of the underground water body and excessive evaporation.

Wells located on a ground-water divide will provide the best estimates of recharge values because the difference between inflow and outflow approaches zero (the water-table surface is almost horizontal), and therefore recharge depends directly on precipitation. Water-table wells on slopes and locations where the water-table fluctuation is directly related to human activities such as pumping and dewatering have been excluded from the study. In those cases, the recharge parameter is nearly unrelated to precipitation.

Two parameters (Δh and ΔZ) must be evaluated before the recharge value can be estimated. The Δh can be obtained from the water-table well observation and the ΔZ can be estimated.

For the terrace-type ground-water regime, ΔZ was determined from the ground-water hydrograph by extending the line of ground-water decline to the point of the next observed maximum rise in water level. For the divide type regime, ΔZ was determined based on recommendation by Kovalevsky (1968).

Recharge in this case is directly proportional to changes of the aquifer thickness and is a function of it. Determining the saturated thickness of the aquifer before recharge occurs and knowing the average thickness of the aquifer, it is possible to estimate changes in thickness of the flow at any given position of the water level. The ΔZ parameter, therefore, can be determined as follows:

$$\Delta Z = \frac{(h_1 - h_2)\ (h_3 + h_2)\,\Delta t_2}{(h_1 + h_2)\ \Delta t_1}\ ;$$

where: h_1 and h_2 = saturated thicknesses of the aquifer during the time when there is no recharge in inches;

h_3 = saturated thickness of the aquifer corresponding to the maximum rise of water level (maximum thickness), in inches;

Δt_1 = period corresponding to the interval $(h_1 - h_2)$ when there is no recharge in years;

Δt_2 = period corresponding to the interval $(h_3 + h_2)$ when the ground-water reservoir receives recharge, in years.

The rise of water level (Δh) when determined for short periods of observation can be erroneous, random or reflect artificial conditions (pumping, injection, etc.) and so may not represent natural processes. Therefore, it was necessary that the observation well measurements be compared to precipitation and other climatic data for the corresponding time periods to evaluate the reliability of the level record. After a correlation was established between water-level records and precipitation data, the average amplitude of static water-level fluctuation was determined from

The second critical parameter in estimating recharge rates is the storage coefficient. In areas where there were no data on pumping tests or pumping tests were short, data on specific capacity from domestic wells were used. Storage coefficients in these cases were calculated from empirical equations. No data on storage coefficients exist for bedrock aquifers under water-table conditions, and storage coefficients were approximated from literature data.

Statistically processed ground-water levels were used for the evaluation of long-term rise (Δh). This rise with the correction (ΔZ water-level decline due to outflow) was multiplied by the storage coefficient to determine the recharge rate. This methodology gave reasonably valid estimates of recharge rates on a regional scale although they are probably underestimated because of certain unavoidable limitations, such as the following: (1) the water-table can never be ideally horizontal, (2) the observed rise of the water table sometimes will not be a true maximum, (3) ground-water outflow during the spring and early fall periods of water-level rise will always be slightly higher than in late fall and winter due to a corresponding increase of water level and flow gradient.

The resulting long-term recharge values were used for comparison with the results obtained by hydrologic methods (low-flow stream data and ground-water contribution to streamflow in formation) evaluated in previous investigations (Kanivetsky, 1979b).

III. HYDROGEOLOGIC SETTING

The following discussion is mainly about Quaternary deposits in the state, and also in the southeastern part where recharge from precipitation occurs mainly in the bedrock aquifers.

1. Hydrogeologic Units

Quaternary deposits represent only a very small part of the state's geologic history. However, glacial drift spread by Pleistocene ice sheets covers most of Minnesota and ranges to hundreds of feet in thickness, so that the Quaternary geology is the major influence on topography, soils, water and land--in short, on the landscape of Minnesota. The deposits now exposed at the surface are of late Wisconsin age except in the southeast and southwest corners of the state where older deposits (possibly of Illinoisan or Kansan age) occur near the surface, covered at most places with loess (Goebel and Walton, 1979). Pre-Wisconsin deposits probably are widespread beneath the younger Wisconsin deposits, especially in buried bedrock valleys. However, the age of the buried Quaternary sediments in them is uncertain. The total thickness of the drift exceeds 200 feet in most of the state and is from 400 to more than 700 feet in the south-central part, and at certain localities in the west exceeds 900 feet (Olsen, 1982). In general, the drift is thicker in the southwestern and thinner in the southeastern parts of the state. Hydrologically (Kanivetsky, 1979a) the Quaternary deposits may be divided into seven units: ground and end moraines, outwash, alluvium, lake deposits, loess, peat and buried outwash. Ground and end moraines are prominent features of the surface deposits. Their principal constituent is till of low permeability, which therefore cannot yield appreciable amounts of water. Till deposited at ice margins or from stagnating masses of melting ice (end moraine) forms irregular, pitted to hilly, topography with many ponds and lakes.

Outwash deposits of admixed and interbedded sand, silt and gravel were carried from glaciers by meltwater and spread over wide areas. The sustained yield ranges from 25 to 500 gallons per minute (Kanivetsky, 1979a) and depends on the thickness of the outwash deposits and their permeability. The outwash aquifers are the main source of water supply in central Minnesota and represent areas where recharge occurs from precipitation.

Alluvial sand and gravel deposits on present floodplains, generally locally interbedded with silt, clay and organic materials, are also an important source of water supply. The sustained yield rating ranges from 25 to more than 500 gallons per minute. The highest yields for alluvial deposits are those along the Minnesota and Mississippi Rivers, where recharge from the stream is possible and therefore the probability of irreversible depletion of a ground-water reservoir is minimal. Major glacial lake deposits are associated with glacial Lake Agassiz in northwestern and north-central Minnesota and glacial Lakes Upham and Aitkin, northwest of Duluth. The sediments formed in these lakes commonly vary from nearly clean sand along the outer edges through sandy silt to fine clay in the central part. The yield rating ranges from very low (<1 gallon per minute--practically impermeable) in clay to 100 gallons per minute in some thicker sections of sandy lake deposits in the Beltrami County area in the northwestern part of the state.

Minnesota has a large area of wetlands, swamps, bogs and marshes which covers about 11 percent of the state (P. Olson, oral communication, 1983). Organic deposits of these wetlands are generally composed of peat. Because the water is absorbed in the fabric of the peat, the permeability is very low. Little is known of the areal distribution of glacial materials below the peat. A few boreholes indicate that some sand and gravel deposits are found below the peat (Morris Eng, oral communication, 1981). Although the extent of these deposits is unknown, where they are present, they may be important aquifers. Loess deposits occur in the southwestern and southeastern corners of Minnesota. The loess is composed mainly of silt and has a low-yield capacity. Very little is known on distribution of buried bodies of sand and gravel. Sporadic bodies of sand and gravel of uncertain extent are found in central and western Minnesota (Kanivetsky, 1979a).

Recharge from precipitation to bedrock aquifers occurs mainly in southeastern Minnesota. Consolidated sedimentary rocks in this part of the state comprise five aquifers (from top to bottom): Cedar Valley-Maquoketa-Dubuque-Galena, St. Peter, Prairie du Chien-Jordan, Franconia-Ironton-Galesville and Mt. Simon-Hinckley-Fond du Lac, each of which is separated by confining beds (Kanivetsky, 1978; Kanivetsky and Walton, 1979).

Cedar Valley-Maquoketa-Dubuque-Galena aquifer comprises the four youngest formations in the sequence of Paleozoic sedimentary strata. The formations composing the aquifer consist mainly of grayish-yellow or olive-gray limestone, dolomite and dolomitic limestone. Beds are a few inches to a few feet in thickness and the rock is vuggy and fractured. Locally thin shale beds in the Galena, Dubuque and Maquoketa Formations somewhat confine the flow of water in separate layers, and differences in static water levels may be found in wells in the same general area. Regionally, however, for all practical purposes, the four formations function as a single aquifer. Although the aquifer is not capped by a Paleozoic regional confining bed, most confinement is provided by unconsolidated Quaternary deposits and locally by patches of Cretaceous shale. Unconfined (water table) conditions exist in the extreme eastern part of the aquifer. The aquifer yields water mainly from solution channels, joints and fissure systems, which tend to be more prevalent and open in the uppermost beds present at any location. Water yields from this aquifer are typically between 200 and 500 gallons per minute.

St. Peter aquifer is capped regionally by the Decorah-Platteville-Glenwood confining bed. Where the confining bed has been removed by erosion, the St. Peter lies directly beneath glacial drift or crops out at the surface. In the eastern part, along the Mississippi River where glacial drift is thin or absent, the aquifer is under water-table conditions. The aquifer consists of sandstone that is light yellow or white, massive, quartzose, fine to medium grained, well sorted and friable. Thin beds of siltstone and shale near the base of the St. Peter Sandstone serve as a lower confining layer. Typical yield to individual wells ranges from 100 to 250 gallons per minute.

Prairie du Chien-Jordan aquifer is a major regional aquifer underlying the St. Peter Sandstone and forming the uppermost bedrock aquifer over large areas beyond the limits of the St. Peter Sandstone. Water-table conditions exist in the eastern part of the aquifer as with the other bedrock aquifers . However, the largest area under water-table conditions in southeastern MInnesota is occupied by the Prairie du Chien-Jordan aquifer (Woodward, 1981).

The Prairie du Chien Group comprises two principal formations, the Shakopee Formation and the Oneota Dolomite, both of which are predominantly light brownish-gray or buff, sandy, thin- to thick-bedded dolomite, which is vuggy and fractured and contains some thin layers of interbedded grayish-green shale. The underlying Jordan Sandstone consists of a white to yellow quartzose, fine- to coarse-grained sandstone, which is massive and thick to thin bedded and varies from friable to well cemented.

Despite their differing lithologies, the Prairie du Chien Group and the Jordan Sandstone function as one aquifer because there is no regional confining bed between them. Locally, however, small head differences may exist due to the presence of some relatively impermeable beds of limited extent. Typical yields to wells range from 500 to 1000 gallons per minute.

Franconia-Ironton-Galesville aquifer is separated from the Prairie du Chien-Jordan aquifer by the St. Lawrence confining bed. The Franconia Formation consists of a gray to greenish, glauconitic very fine- to coarse-grained quartz sandstone with some interbedded greenish-gray micaceous shale and dolomitic and nonglauconitic sandstone layers. The Ironton Sandstone is a white, medium-grained, moderately well to poorly sorted quartz arenite that contains a significant amount of admixed silt-size material. The Galesville Sandstone consists of a white to light-gray, slightly glauconitic, well- to moderately well-sorted, mostly medium-grained quartzose sandstone. Water-table conditions exist in the eastern part in a very narrow strip along the Mississippi River and tributaries.

Typical yield to individual wells ranges between 250 and 500 gallons per minute.

Mt. Simon-Hinckley-Fond du Lac aquifer is separated from the upper Franconia-Ironton-Galesville aquifer by the Eau Claire confining bed and dips southward and eastward below the confining bed at the northern and western margins of the Twin Cities basin and is continuous beneath younger rocks southward.

The Mt. Simon Sandstone of Late Cambrian age is fine to coarse grained, contains some thin beds of shale, and is commonly white, gray or pink in color, locally yellow. The Upper Precambrian Hinckley Sandstone is medium to very thin bedded, fine to coarse grained, and pale red to light pinkish or brownish gray in color. The Fond du Lac Formation, also Upper Precambrian, is characterized by lenticular beds of arkosic sandstone and interbedded mudstone, and is dark red to pink in color. Ground water of the aquifer is generally under confined (artesian) conditions.

Typical yields for individual wells range from 400 to 700 gallons per minute.

2. Ground-Water Movement

Water from precipitation infiltrates into the ground and then under the influence of gravity moves toward the water table, which is the upper surface of the zone of saturation. Ground-water movement within the system is controlled by distribution of hydraulic potential. Hydraulic potential, in turn, is related to recharge-discharge relationships, hydraulic conductivity of geologic materials and topography. Recharge is generally concentrated in uplands and discharge in lowlands. The water-table tends to be closer to the surface in relatively impermeable materials and in lowlands (discharge areas). It is deeper in relatively permeable materials and beneath topographic elevations (uplands, recharge areas). The water table usually resembles a flattened form of the surface relief with elevations under hills and depressions below valleys.

Several continental divides within the state boundaries separate water flowing northward into the Red River of the North from water flowing southward into the Minnesota and Mississippi Rivers, and eastward to Lake Superior. Regional ground-water movement in the Quaternary deposits is influenced by these divides (Kanivetsky, 1979a). Although ground water has the above regional movement patterns, most recharge water moves only within local systems. The distance that the ground water travels from a recharge to a discharge area depends on the size of the basin and is generally only a few miles. Thus, total discharge to lakes, wetlands, and streams is greater than discharge directly into major rivers. Further, in areas where Quaternary deposits are underlain by Precambrian igneous and metamorphic rocks and topographic relief is significant, the local flow systems extend to the full thickness of the ground-water reservoir, and regional flow is negligible, particularly in the central part of the state. In the southeast the flow is more complex than in most other places because of the basinal structure of the underlying Paleozoic and Proterozoic rocks. These sedimentary bedrock aquifers underlie the drift and ground water moves into the bedrock, becoming part of the regional flow. Local relief has little effect on ground-water movement in southeastern Minnesota.

Water in the outwash, alluvial and sandy lake deposits is under unconfined water-table conditions except locally where the deposits are fully saturated and are overlain by a relatively impermeable layer of silt or clay. Water in buried outwash deposits is under confined (artesian) conditions. In bedrock aquifers water is generally under artesian conditions except in southeastern Minnesota along the Mississippi River and tributaries ( ).

IV. DETERMINATION OF STORAGE COEFFICIENTS

1. Factors Affecting the Storage Coefficient

The storage coefficient or specific yield of the water-table aquifers is the ratio of the volume of water which the rock or soil, after being saturated, will yield by gravity to the volume of the rock or soil. In the natural environment, the storage coefficient is generally observed as the change that occurs in the amount of water in storage per unit area of a water-table aquifer as the result of a unit change in the head. Such a change in storage is produced by the drainage or filling of pore space and is therefore dependent on particle size, rate of change of water table, time, temperature and viscosity of water and various physical characteristics of the host material. The quantity of water yielded to wells from a body of saturated rock or soil depends on the storage coefficient of the rocks or soil and not on its porosity. Certain deposits, such as clay or peat, have high porosity and thus, are capable of holding large quantities of water but yield only a small part of it under gravity drainage, even after a long period of time. Consequently these materials do not transmit water and are not aquifers. The volume of water drained by the rapid lowering of the water table in the immediate vicinity of a well during a pumping test is much less than the total that would be yielded by long-term drainage. In contrast the drainage that accompanies the annual fluctuation of the water table is virtually complete. Williams and Lohman (1949, p. 213, 220) stated that the true value of specific yield is obtained only after the saturated material has been drained for a long period of time.

Prill, Johnson and Morris (1965) present quantitative information for sand size materials that indicates that a period of two months to more than one year would be required for the drainage to reach equilibrium and, thus, give the valid storage coefficient.

All methods to determine the storage coefficient of rock or soil have some limitations. Laboratory samples are very small and often do not represent all variabilities of rock or soil within the area. The measurement and control of variables in the field is difficult. Pumping tests and other field methods are labor intensive and expensive. In addition, pumping test data should be used with caution when the period of pumping from the well is not sufficient.

2. Literature Review

A number of laboratory and field methods have been developed for the determination of storage coefficients, and during the past decades a vast amount of information on storage coefficients has been collected by many researchers in the U.S. and abroad. The present authors have tried to present a uniform table for different types of material based on laboratory, field and other data.

Johnson (1967) has published the following average values for storage coefficients.

Material	Storage coefficient
clay	0.02
silt	0.08
sandy clay	0.07
fine sand	0.21

medium sand	0.26
coarse sand	0.27
gravelly sand	0.25
fine gravel	0.25
medium gravel	0.23
coarse gravel	0.22

However, he stated that no particular values are recommended. In addition, the number of determinations of the storage coefficient was insufficient for general application of these results.

Lebedev (1976) reports ranges of storage coefficients that he derived from Maslov's work and compared data with his research and other investigators. He stated that there was good agreement between the storage coefficient determined by Maslov, himself and other researchers. The ranges of storage coefficients are as follows:

Material	Storage coefficient
peat (upland)	0.02-0.1
peat (lowland)	0.08-0.14
fine sand	0.01-0.2
clayey sand	0.06-0.15
sandy clay	0.01-0.06
clay	0.00 x n*-0.05
gravel	0.25-0.35

*n is any positive number from 1 to 10.

Volftsun (1972) suggests approximate values of storage coefficients for material under water-table conditions based on a relationship between the storage coefficient and aquifer depth, with an additional review of literature. The author recommends the use of these data only for coarse approximation and not for site specific analysis. Values of the storage coefficients are as follows:

Material	Storage coefficient
silty clay	0.005-0.02
sandy clay	0.02-0.04
silt	0.04-0.06
silty sand	0.06-0.08
fine sand	0.08-0.12
medium sand	0.12-0.2
coarse sand	0.2-0.3
gravelly sand	0.25-0.35
fractured limestone	0.01-0.1

Todd (1970) in the Water Encyclopedia gives a wide range for storage coefficients using driller's terms for various materials. A summary of this information is as follows:

Material	Storage coefficient
crystalline bedrock (fresh)	0.0
clay and related material	0.03
clay and gravel, sandy clay and materials	0.05

fine sand, compacted sand, compacted gravel and similar materials	0.1
gravel, sand, sand and gravel and similar materials	0.25

Some researchers (Bochever et al., 1965) suggest determnination of storage coefficents based on grain sizes for clean, sandy material:

Grain size	Storage coefficient
0.1 mm	0.18
0.2 mm	0.25
0.5 mm	0.28

For silty sand and sandy clay or silt, they recommend approximate values respectively of 0.1-0.15 and 0.01-0.1.

Lohman (1972) states that the storage coefficient generally ranges between 0.1 and 0.3 and that a long period of pumping may be required to drain water-bearing material. Thus, in the absence of any field or laboratory determination in a rapid reconnaissance, it would not be very far off in assuming that the storage coefficient for a water-table aquifer composed of sand and gravel is about 0.2--the average value between the general limits indicated.

Little research has been done in determining the storage coefficient for bedrock aquifers under water-table conditions.

Walton (1970) gives storage coefficient values for selected materials including bedrock:

Material	Storage coefficient
clay	0.01-0.1
sand	0.1-0.3
gravel	0.15-0.3
sand and gravel	0.15-0.25
sandstone	0.05-0.15
shale	0.005-0.05
limestone	0.005-0.05

The upper limits for storage coefficients mentioned in the table for clay and sandstone (0.1-0.15) seem to be too high. Those values are more representative of sandy and gravelly materials.

Based on literature review and regional studies, Bochever (1968) obtained storage coefficient values of 0.005-0.1 and 0.001-0.03. for water-table aquifers composed of limestone and sandstone respectively. Storage coefficients for water-table carbonate rocks in Kentucky ranged between 0.0018 and 0.0082 (Walker, 1956). This parameter has been calculated based on data obtained from nearby stream flow and observation wells. In the report it is stated that the observation wells are very close to farm houses and possibly some effect of pumping remains in the calculation. Therefore, the values of storage coefficients are very low. Detailed studies conducted during the past 30 years in the northwestern part of the Soviet Union in formations similar to the Prairie du Chien-Jordan in Minnesota showed that the storage coefficient for the fractured limestone and dolomite under water-table conditions generally lies within the range of 0.03-0.05 (Zaltsberg, 1980).

No data are available on storage coefficients for carbonate rocks in Minnesota and neighboring states. However, in the southeastern part of Minnesota bedrock formations, mainly of the Prairie du Chien-Jordan aquifer, are under water-table conditions. It was not feasible to make an extensive field investigation to determine storage coefficients of bedrock formations, which itself comprises a separate study. Therefore, on the basis of available literature data, the storage coefficient was estimated to be within the range of 0.03-0.05.

Site specific storage coefficients were obtained using methods described below.

3. Soil-Moisture Method

This method is based on the difference between the moisture content of saturated material and the moisture content of the same material after it has been drained. Moisture content was measured using a neutron probe, as recommended by Meyer (1962).

For the analysis of data separate neutron logs were selected near the beginning and end of periods during which water levels persistently declined or rose. The change in moisture storage for the time span between logs was determined by integrating the difference in moisture content over the depth interval through which moisture changes (due to water level changes) occurred. The change in moisture storage was divided by the changes in water level during the period of water decline or rise to obtain an estimate of the storage coefficient.

This method was used to determine the storage coefficient of the sandy deposits near Andover, Anoka County. The Department of A ricultural Engineering, University of Minnesota, drilled three holes to a depth of 61, 16 and 12 feet, which were converted to wells with the depth of 12 (main well), 11 (west well), and 11 (south well) feet and screen intervals of 9.5-12, 8.5-11 and 8.5-11 feet, respectively. The water level is generally from 5 to 7 feet below the surface.

Continuous water level monitoring and moisture measurement with a neutron probe were conducted in 1980 and 1981. Seven cycles of rising or declining water levels were shown on ground-water hydrographs. The decline or rise of water levels is in the same direction in all three wells. However, the rate of fluctuations in the west well is significantly different from the main and south wells. This lower efficiency of the west well may be due to poor well development. Calculated storage coefficient between the main well (0.12-0.25) and the south well (0.12-0.28) are in close agreement; data from the west well showed broad variation of storage coefficient (S = 0.13-0.61). The upper value in the west well is significantly over the representative limit of any clastic sediments and suggests that the west well does not respond to natural changes of water levels. Fourteen values of storage coefficient determined at the Andover site from main and south wells range from 0.12 to 0.28 with an average of 0.18.

4. Pumping Test Method

This method is based on measuring the rate of water level decline (or rise) after pumping (or recovery) a measured volume of water and then determining the volume of sediments drained. The storage coefficient is obtained as the ratio of the volume of water pumped (recovered) to the volume of sediments drained. A pumping well and at least one observation well are needed to determine the storage coefficient. Pumping water from a well lowers the water table surface around the well and creates a cone of depression. Data are obtained on rate of discharge, water levels and the duration of pumping period. The drawdown data are then used to calculate the storage coefficient. The converse of the pumping data are the recovery data. Analysis of pumping tests is based on the classical Theis solution (1935) with modifications by Boulton (1963), Streltsova (1973) and Neuman (1979). These modifications are needed to take into account the existence of an additional source of water which is released from storage with a delay in time, a so-called "delayed yield." The following assumptions are generally used in formulas: (1) the aquifer has infinite areal extent, (2) the well diameter is infinitesimal and (3) water removed from storage is discharged instantaneously with the decline of head.

Many investigators (Johnson, 1967; Stallman, 1967; Price, 1977, 1980) have stated that storage coefficient values determined from pumping tests are smaller than would be expected because the standard pumping tests usually are of short duration. Johnson (1967), referring to the study by Wenzel, reported that saturated water-bearing material when allowed to drain may yield water rather slowly. Samples of saturated material may continue to drain for several years, although most of the water in it may drain in a much shorter time. The slow drainage of water-bearing material in the vicinity of a pumped well causes the water table to decline rapidly at first and then more slowly as draining proceeds.

Storage coefficients from pumping tests for this study were obtained from USGS reports (Lindholm, 1971, 1980; Reeder, 1972; Helgesen, 1973, 1977;

Larson, 1976; Wolf, 1981) and unpublished data (Myette, 1981; Soukup, 1981), and from the Department of Natural Resources reports (Johnson et al., 1980) and unpublished data. The length of pumping tests was generally between 12 and 72 hours. In addition, some data provided by the Department of Natural Resources on storage coefficients were unreliable, i.e., storage coefficient for water-table aquifers was in the ranges of artesian conditions. Regional distribution of pumping tests was limited to north-central and east-central Minnesota). Therefore, reasonable geographic distribution of storage coefficients based on pumping tests was not possible.

5. Empirical Equations

In order to obtain a reasonable geographic distribution of storage coefficents for water-table aquifers it was necessary to estimate the storage coefficient using empirical equations. Experimental studies were made by many researchers to determine different hydraulic parameters and their relationship to the field data.

Lebedev (1976) referring to studies conducted by K.P. Lundin and separately by L. Daml and R. Vinger suggests two empirical equations:

$$S = 0.13 + 0.074 \lg k \quad \text{(K.P. Lundin)}$$

$$S = 0.1254 + 0.1152 \lg k \quad \text{(L. Daml \& R. Vinger)}$$

Shestakov (1973) suggests the following equation on the basis of an experimental study conducted by P.A. Betsinsky:

$$S = 0.117\sqrt[7]{K}$$

where: K is hydraulic conductivity of material in m/day.

The basic principle of these equations is that the storage coefficient is a function of hydraulic conductivity.

These formulas are applicable only for clastic materials and cannot be used for rocks that have secondary porosities (limestone, dolomite) or porous bedrock (sandstone).

The transmissivity was determined from specific capacity using a graph method and formulas (Csallany, 1966; Johnson Drillers Journal, 1968; Theis, 1963). The thickness of the aquifer was approximated from well logs and a hydrogeologic map (Kanivetsky, 1979a). Hydraulic conductivity was estimated by dividing transmissivity over the aquifer thickness. In determining the transmissivity of the aquifer the following assumptions were used:

a) the well is 100% efficient

b) well loss is negligible

c) the aquifer is fully penetrated

d) the aquifer is infinite in areal extent

In the empirical equations cited above, the storage coefficient is a function either of a logarithm or of the seventh root of hydraulic conductivity, and therefore, changes of hydraulic conductivity even by a considerable percentage do not significantly alter the value of the storage coefficient.

Although the storage coefficients determined from empirical equations are not precise, they are representative and especially useful for areas where data are absent or scarce.

About 650 determinations of the storage coefficients for water-table aquifers have been made throughout Minnesota. The data distribution is not even, however, with the main concentration in central Minnesota.

6. Review and Comparison of Results

Storage coefficients for different types of sediments constituting the water-table aquifers were determined from (1) soil-moisture measurements; (2) pumping tests; (3) empirical equations; and (4) literature data.

The most reliable data were obtained from soil-moisture measurements at Andover, Anoka County. Because of the unpredictable changes in grain size and degree of sorting of clastic materials the data from the Andover site cannot be extrapolated to other areas.

Storage coefficients determined from pumping tests are concentrated in north-central and east-central Minnesota. In addition to the limited distribution of data, the pumping tests are usually of short duration and therefore in most cases storage coefficients calculated from pumping test data are slightly underestimated.

Analysis of data on storage coefficients calculated from empirical equations shows that values determined from formulas of K.P. Lundin and L. Dami & R. Vinger (Lebedev, 1976) are overestimated. Comparison of storage coefficients calculated from empirical equations of P.A. Betsinsky (Shestakov, 1973) with the storage coefficients obtained from reliable pumping tests (Helgesen, 1977) and literature data (Lebedev, 1976; Volftsun, 1972; Johnson, 1967; Johnson et al., 1980) show good agreement. In most cases, however, the storage coefficients determined from pumping tests were lower than those obtained from Betsinsky's formula. This discrepancy is attributed mostly to the slow drainage of water from sediments during the pumping tests, but also may be the result of interpretation of pumping test data. Freeze and Cherry (1979) describing scientific disadvantages of pumping tests, refer to the nonuniqueness of pumping test interpretation. They stated that "unless there is very clear geologic evidence to direct the ground-water hydrologists in their interpretation, there will be difficulties in providing a unique prediction of the effects of any proposed pumping scheme. The fact that a theoretical curve can be matched by pumping test data in no way proves that the aquifer fits the assumption on which the curve is based."

Storage coefficient values determined from different methods were plotted on a Quaternary hydrogeologic map (Kanivetsky, 1979a). For the purpose of estimating regional distribution of storage coefficients the area covered by the water-table aquifer was divided into units. Areal size and ranges of each unit were based on frequency and distribution of reliable determined storage coefficients, thickness of aquifer, grain size and sorting of the material composing the aquifer. Storage coefficients for areas that lacked sufficient data were estimated by extrapolation from information on neighboring units based on analogy of hydrogeological data derived from Kanivetsky (1979b).

In southeastern Minnesota where bedrock aquifers, mainly Prairie du Chien-Jordan, are under water-table conditions storage coefficient values were assigned to the range of 0.03-0.05 (see discussion 2 of this chapter).

V. GROUND-WATER LEVEL CORRELATIONS

1. Sources of Data

Ground-water level records for 200 water-table condition observation wells in Minnesota were obtained from the Minnesota Department of Natural Resources Division of Waters. Ground-water level records for 50 long-term water-table condition observation wells were obtained from U.S. Geological Survey records published in Water-Supply Papers and Water Data Reports for Wisconsin, Iowa, North Dakota and South Dakota.

Climatological data were obtained from U.S. Environmental Data Service published records for Minnesota and the surrounding states.

2A. Regression of Static Water-Level Rise Versus Effective Precipitation--a Method of Estimation of Static Water-Level Rise During Recharge

Static water-level rise is estimated from the graph of the static water-level record, called a hydrograph. A recharge event is indicated by a rise in the static water level in the aquifer. When no recharge from the land surface is occurring, the water level recedes at a rate dependent on the lithology of the aquifer. To estimate static water-level rise during a recharge event, the recession curve is extended graphically from the low point previous to the beginning of recharge to a point directly below the high point of the recharge event. The vertical distance between the low point on the recession curve and the high point of the recharge event is the estimated static water-level rise.

Some water-level fluctuations do not reflect recharge. Pumping effects show up as sharp convex upward water-level declines and recovery in, contrast to the linear to concave upward natural recession curve. Observation wells with water levels less than five feet from the land surface show sharp but temporary water-level fluctuations in summer due to evapotranspiration and water-level rises from storms exaggerated by entrapment of air in water-saturated soil. As a rule, if the natural recession curve is not displaced forward in time on the hydrograph, these fluctuations do not represent recharge and are ignored.

2B. Characteristics of Minnesota's Climate in Relation to Ground-Water Level Fluctuations

During winter precipitation falls mainly as snow. Recharge is delayed until air temperature rises above freezing. Recharge from the melting snow pack begins before the soil is fully thawed. Infiltration can occur because ice in most soils does not form continuous sheets.

From April through October, evapotranspiration occurs, reducing the amount of precipitation available for infiltration. Evapotranspiration subtracted from precipitation equals the effective precipitation available for recharge.

During fall, after killing frosts, evapotranspiration is not significant. During this period recharge occurs most efficiently from precipitation unaffected by evapotranspiration or time lags due to extended periods of below-freezing air temperatures.

2C. Regression of Static Water-Level Rise Versus Precipitation

Due to the delay of recharge from winter precipitation and the effects of evapotranspiration from April through October, a close correlation between water-level fluctuations and precipitation, measured week by week or month by month, is not found in Minnesota.

In addition to the lag between precipitation and recharge due to climate, recharge occurs at varying time lags due to antecedent soil moisture and variation in depth to water table.

Consequently there are two problems to be solved to obtain a good regression equation for water-level rise versus precipitation: accommodation of varying time lags and calculation of effective precipitation.

The first problem is solved by using the interval between peaks in the hydrograph as the time basis for the regression, rather than a set interval of a week, month, or year. The dependent variable, static water-level rise, is regressed against the independent variable, the sum of effective precipitation taken over the interval of the static water-level rise.

There are two benefits to this procedure. Recharge events can last weeks, months, or even years. The reliability of the regression equation is improved by the inclusion of small and large recharge events at the extremes of the ranges of the variables. In addition, most observation wells are taped at greater than monthly intervals, making month by month correlation impractical. The use of the recharge event as a time interval rather than correlation on an annual basis increases the total number of separate events in the sample space.

The calculation of effective precipitation for the months of April through October is based on monthly summations of precipitation (P) at the weather station closest to the well and monthly summations of pan evaporation (E) at five stations in or near the border of Minnesota.

It was assumed that monthly evapotranspiration could be approximated by taking a fraction of the monthly pan evaporation of the nearest station. In general, pan evaporation multiplied by a coefficient of 0.7 aproximates evaporation from open water bodies. Therefore, several trials were run correlating this sum of monthly effective precipitation with static water-level rise where monthly effective precipitation was given a range of (P-0.7E) to (P-0.3E). In most cases monthly sums of (P-0.5E) resulted in the best correlation. If 0.5E is greater than P, the summation is taken as zero.

For the remaining months, November through March, the unmodified precipitation totals were used.

When evaporation data were missing, an estimated value based on the correlation between nonevaporation and monthly average maximum air temperature was used.

If the recharge peak occurs in the middle of the month, daily values of precipitation and evaporation are summed for the relevant fraction of the month.

2D. Results of Regression of Static Water Levels Versus Precipitation

From the linear regression equation, the average annual static water-level rise is determined, given the average annual effective precipitation for the years since pan evaporation measurements were begun in Minnesota.

For each well record, the number of recharge events used, the regression equation, coefficient of determination and standard error are listed. The average water-level rise given average annual effective precipitation for the closest weather station is determined in order to calculate average annual recharge at the well site.

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