Artificial Recharge Through a Well in Fissured Carbonate Rock, West St. Paul, Minnesota

GEOLOGICAL SURVEY WATER-SUPPLY PAPER 2004

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By HAROLD O. REEDER, WARREN W. WOOD, GARRY G. EHRLICH, *and* REN JEN SUN

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Report describes an experiment and related tests in which treated water was injected into a fissured carbonate rock aquifer overlying a sandstone aquifer

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METRIC-ENGLISH EQUIVALENTS

Factors for converting English units to metric units are shown to four significant figures. However, in the text the metric equivalents are shown only to the number of significant figures consistent with the values for the English units.

To convert English units	*Multiply by*	*To get Metric units*
in. (inches)	2.540×10^{1}	mm (millimetres).
ft (feet)	3.048×10^{-1}	m (metres).
mi (miles	1.609	km (kilometres).
mi^2 (square miles)	2.590	km^2 (square kilometres).
gal (gallons)	3.785×10^{-3}	m^3 (cubic metres).
ft^3 (cubic feet)	2.832×10^{-2}	m^3 (cubic metres).
ft/s (feet per second)	3.048×10^{-1}	m/s (metres per second).
ft/d (feet per day)	3.048×10^{-1}	m/d (metres per day).
gal/min (gallons per minute)	6.309×10^{-2}	l/s (litres per second).
ft^3/s (cubic feet per second)	2.832×10^{1}	l/s (litres per second).
ft^3/min (cubic feet per minute)	2.832×10^{-2}	m^3/min (cubic metres per minute).
ft^3/d (cubic feet per day)	2.832×10^{-2}	m^3/d (cubic metres per day).
Mgal/d (million gallons per day)	4.381×10^{-2}	m^3/s (cubic metres per second).
(gal/min)/ft (gallons per minute per foot)	2.070×10^{-1}	(l/s)/m (litres per second per metre).
(gal/d)/ft (gallons per day per foot)	1.242×10^{-2}	(m^3/d)/m (cubic metres per day per metre).
(ft^3/d)/ft (cubic feet per day per foot)	9.291×10^{-2}	(m^3/d)/m (cubic metres per day per metre).

ARTIFICIAL RECHARGE THROUGH A WELL IN FISSURED CARBONATE ROCK, WEST ST. PAUL, MINNESOTA

By Harold O. Reeder

ABSTRACT

The Prairie du Chien Group was injected with 2,754,000 gallons (368,200 cubic feet), or 10,430 cubic metres, of municipally treated water at about 100 gallons per minute (13.4 cubic feet per minute), or 6.3 litres per second, for 20 days. The injection-pipe system was designed to utilize pipe friction rather than a remote-controlled valve in the well to maintain positive pressure and eliminate air entrainment in the injection water and the escape of dissolved gasses from the water. During the 20-day injection period the temperature of the injection water declined gradually from 15.0° to 11.2°C, and the flow rate decreased from 108 to 90 gallons per minute (6.8 to 5.7 litres per second).

Analyses of test data were, in some instances, based upon hydrologic judgement as well as observations. Results of aquifer tests before and after injections indicated that the transmissivity had decreased 18 percent during the intervening injection periods; however, the specific capacity remained the same, indicating no change in transmissivity during pumping. Analysis of water-level changes in observation wells during injection indicated a reduction in transmissivity of more than 50 percent; however, the specific capacity of the injection well decreased only about 5 percent during injection.

A comparison of water-level changes with the discharge or recharge rates of the three tests showed that the water-level changes in the two observation wells tapping the Prairie du Chien Group during the injection test were greater than those projected from the two aquifer pumping tests. The deviations in the water-level changes and in the analysis of aquifer-test data indicate that the methods used to analyze data from these wells may not be wholly applicable, inasmuch as anisotropic and nonhomogeneous conditions prevail in at least the Prairie du Chien part of the aquifer.

The native water and the injected water averaged 0.8 and 25 milligrams per litre chloride, respectively. The chloride, utilized as a tracer, showed that the injected water was detected only in the lower part of the nearest observation well, 99 feet (30.2 metres) from the injection well. The chemistry of the water and the rock formation showed little likelihood of plugging of the recharge well by chemical precipitation. Microbiological phenomena apparently did not become a significant factor in the recharge test.

The hydraulic gradient of the aquifer in October and December 1971 (before and after injection) was estimated to be N. 36° E., 0.0013, and N. 39° E., 0.0012, respectively, on the basis of measurements of water levels in the three wells in the Prairie du Chien Group. The single-well tracer-dilution method of calculation showed a hydraulic gradient of 0.0016. A longitudinal dispersivity of 280 feet (85 metres) was calculated. Such a value of dispersivity is typical of fractured reservoirs and shows that the Prairie du Chien Group is a heterogeneous aquifer.

The injection test demonstrated that it is hydrologically feasible to recharge the Prairie du Chien Group and the Jordan Sandstone artificially through wells completed in the Prairie du Chien Group. The fissures in the Prairie du Chien Group act as conduits through which water spreads. The water passes into the Jordan Sandstone from the Prairie du Chien over a larger area than it would if it were injected directly into the Jordan.

INTRODUCTION

Most of the research and experimentation in artificial recharge has been done in aquifers composed of sand, sand and gravel, or sandstone. This project was initiated by the U.S. Geological Survey to investigate artificial recharge of fissured carbonate-rock terrane. Southeastern Minnesota was selected for these experiments of artificial recharge through wells for two reasons: (1) Its geologic setting is similar to that of several other parts of the Nation, so the results would be applicable to other areas; and (2) it is an area where problems of local overdevelopment exist and, thus, would derive direct benefit from the research.

PURPOSE

The general purpose of the project was to test and demonstrate the hydrologic feasibility of artificially recharging fissured carbonate rocks. The purpose of injection into the carbonate rocks was to determine whether the fractures and solution channels in the dolomite would act as conduits through which water would spread out over a large area in a short time and subsequently move downward into the underlying Jordan Sandstone, the major aquifer in the area. Because of the potential rapid spreading along fissures, fewer recharge wells would be required in the fissured rocks for the underground spreading of water than would be required to inject directly into the sandstone. Also, plugging materials near the well face would be less of a problem in fissured carbonate rocks than in sandstone; hence, less extensive redevelopment work would be required for injection in dolomite than in sandstone.

Ground water is being used in metropolitan areas in the United States at an increasing rate. As a result, pumping levels in many areas of use are declining. In the Minneapolis–St. Paul metropolitan area, ground-water pumping averaged about 155 Mgal/d (6.8 m^3/s) in 1965 and increased to 195 Mgal/d (8.5 m^3/s) in 1970. Generally, the ground water is pumped out, used once through a system, and dumped as waste into a sewer or nearby stream. The large ground-water withdrawals have caused potentiometric levels to decline. The potentiometric surface of the Jordan Sandstone had declined as much as 90 feet (27 m) by 1965 since the beginning of pumping in the downtown area of Minneapolis and St. Paul. Because of the increasing use of ground water, there is an increasing need to: (1) Recycle or reuse water within a system, for such use as air conditioning, (2) replenish the ground-water supply by artificial means, and (3) develop surface-water sources to replace ground-water pumpage. Several schemes for developing future water supplies from surface-water sources are being considered for the metropolitan area that would involve much engineering and construction work. Probably, increasing the ground-water supplies by artificial recharge would involve less work.

SCOPE

Southeastern Minnesota and, particularly, the Minneapolis–St. Paul area is well suited to several methods of artificial recharge. These might include ponding, spreading, lake-level manipulation, induced recharge, and recharge through wells. The present demonstration and research experiment, however, is limited to injection of water through wells into the fissured carbonate rocks. Problems of principal concern in well-recharge operations include plugging of the material near the well face by suspended particulate matter, by entrained air, by chemical reactions between the injected water and the native ground water or the aquifer materials, or by bacterial activity. These problems can be minimized by proper well design and by various methods of water treatment.

To make and analyze the injection test, the project included construction of the injection well and six observation wells, determination of the geologic and hydrologic characteristics of the rock units, instrumentation of the wells, design of an injection system, and investigation of chemical and biological aspects.

SELECTION OF SITE

The area selected is well suited geologically because of the presence of carbonate rocks in contact with a heavily pumped sandstone aquifer. These rocks are continuous into adjacent States, and they are similar to carbonate rocks and sandstones in many other parts of the Nation.

The experiment site is on land owned by the city of West St. Paul, Minn., near the intersection of Charlton Street and Marie Avenue. This site, in addition to providing considerable transfer to other parts of the Nation, was selected because (1) the land is publicly owned; (2) an adequate supply of city-treated water is available at the site for recharge; (3) electric power is available; (4) the site is near the local problem area of heavy pumping; thus, the geology and hydrology of the test site are similar to those of an area of potential application of the results; and (5) interference from the large-scale pumping centers is small. (The Mississippi River is incised into the Prairie du Chien aquifer between the site and a heavy pumping center at St. Paul.)

In the study the effects of water injected into the Prairie du Chien Group were observed in the Prairie du Chien Group, in the underlying Jordan Sandstone, and in the overlying St. Peter Sandstone.

ACKNOWLEDGMENTS

The authors thank the officials of the city of West St. Paul for permitting the use of the city park land for construction of the wells and instrument shelters and for access to the structures for the experimental work.

GEOHYDROLOGY

The Minneapolis–St. Paul area is underlain by consolidated sedimentary bedrock units, which are overlain by unconsolidated glacial drift. The bedrock consists of layers of sandstone, shale, and carbonate rock that structurally have the configuration of a relatively shallow basin. The area of the entire sedimentary rock basin—that is, the area of the extent of the Hinckley Sandstone—is about 6,000 mi^2 (15,500 km^2). Each successively higher rock unit in the stratigraphic section covers a smaller area. A generalized stratigraphic section of the area is shown in table 1.

The bedrock has an erosional surface of channels and valleys, which have been filled and buried with glacial drift. The glacial drift ranges in thickness from a few feet to as much as 500 feet (152 m). The major channels are as much as 1.5 miles (2.4 km) wide and 400 feet (122 m) deep. Consequently, the glacial drift is in contact in places with formations as far down the geologic section as the St. Lawrence Formation (table 1). This is an important hydrologic factor because water may move more easily from one aquifer to another through the glacial drift that fills the channels than it does through the consolidated materials between the aquifers. The nearest erosional channel to the test site that penetrates the Prairie du Chien Group is adjacent to the Mississippi River, which is 2 miles (3.2 km) northwest and 2.5 miles (4.0 km) northeast of the test site (Norvitch and Brietkrietz, 1971; and Payne, 1965).

Ground water in the Twin Cities area is pumped from eight geologic units. The units (aquifers) are interconnected to some extent but are separated in most areas by confining layers of less permeable material, which restricts the flow of water vertically. Some of the units, such as the St. Peter Sandstone, the Prairie du Chien Group, and the Jordan Sandstone, are permeable and yield large quantities of water to wells.

The geologic formations of primary concern at the recharge experiment site are the Prairie du Chien Group, in which water was injected, and the underlying Jordan Sandstone, which is the major aquifer in the area. The Prairie du Chien Group is primarily dolomite, ranging in composition from 60 to 95 percent dolomite and from 2 to 30 percent quartz in samples analyzed by the X-ray-diffraction method (table 2). The Prairie du Chien is highly fractured and contains solution openings (figs. 1, 2). Both types of fissures were noted during drilling and caused difficulty in core-sample recovery. Although data are not sufficient to show details, there is indication that the fissures may have selective orientation, and they are not uniformly distributed vertically in the formation. Thus, the Prairie du Chien Group is apparently anisotropic and nonhomogeneous.

The Jordan Sandstone is primarily quartz, ranging in composition from 50 to 95 percent quartz and from 2 to 25 percent dolomite in the samples analyzed (table 2). The Jordan Sandstone is more uniform than

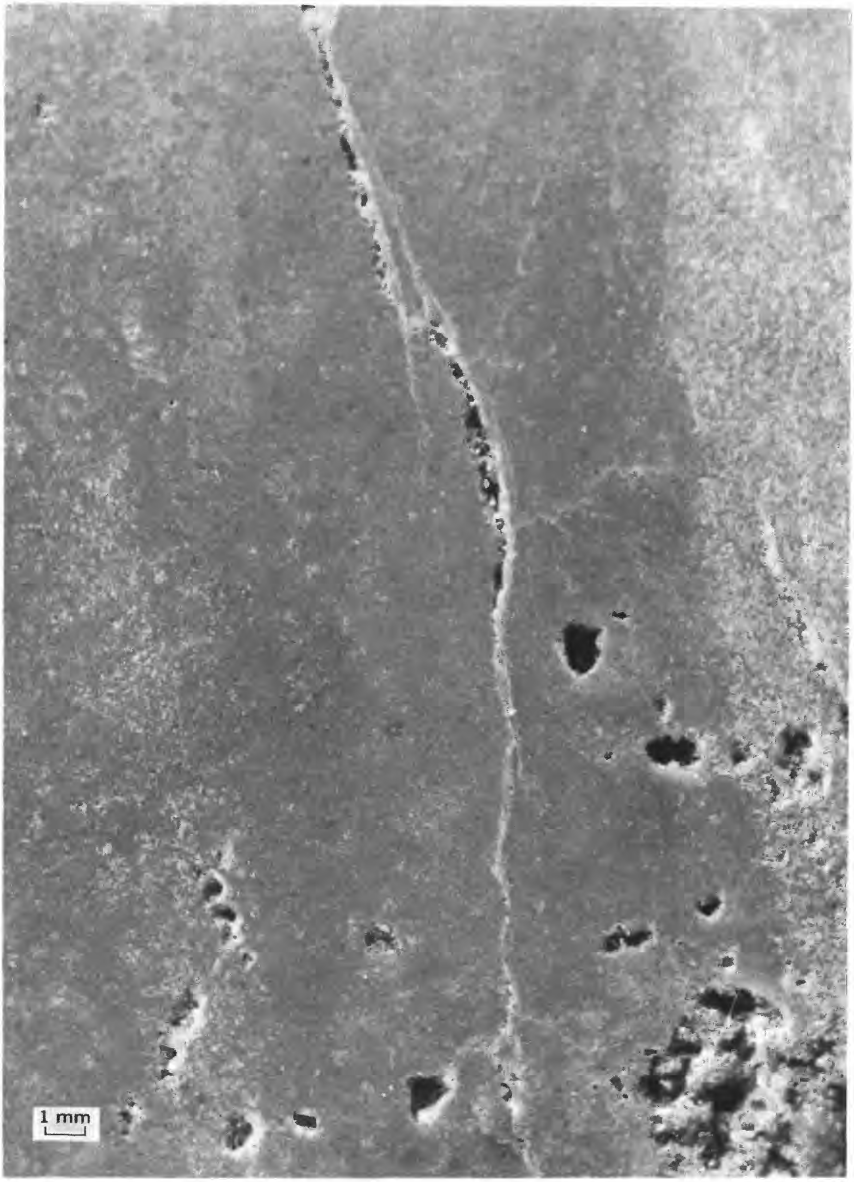

FIGURE 1.—Photograph of a core of rock from the Prairie du Chien Group at a depth of 442 feet (135 m) in the injection well, 1-S, showing fracture and solution openings in horizontal section (× 5).

the Prairie du Chien Group and is considered to be homogeneous, although small fractures and variations in material are recognized. (See table 1.) The grain size is fine to coarse, and the grains are loosely to well

TABLE 1.—*Geologic units and their water-bearing characteristics*

[After Norvitch and others, 1973]

System	Geologic unit	Approx. range in thickness (feet)	Description	Water-bearing characteristics
Quaternary	Undifferentiated glacial drift	0–400+	Glacial till, outwash sand and gravel, valley-train sand and gravel, lake deposits, and alluvium of several ages and several provenances; vertical and horizontal distribution of units is complex.	Distribution of aquifers and relatively impermeable confining beds is poorly known, especially in subsurface. Where saturated, stratified well-sorted deposits of sand and gravel (alluvium, valley train, outwash, some lake and ice-contact deposits) yield moderate to large supplies of water to wells. Records of 24 large-diameter wells completed in sand and gravel show yields ranging from 240 to 2,000 gal/min with from 2 to 69 feet of drawdown. Des Moines Lobe till is non water bearing; Superior Lobe till is sandy and may yield small supplies suitable for domestic or farm use.
Ordovician	Decorah Shale	0–95	Shale, bluish-green to bluish-gray; blocky; thin, discontinuous beds of fossiliferous limestone throughout formation.	Confining bed.
Ordovician	Platteville Limestone	0–35	Dolomitic limestone and dolomite, dark-gray, hard, thin-bedded to medium-bedded; some shale partings; can be divided into five members.	Where saturated, fractures and solution cavities in rock generally yield small supplies to wells. Records of 23 wells show an average yield of 23 gal/min. Water is generally under artesian pressure where overlain by Decorah Shale. Not considered to be an important source of water in area of study.
Ordovician	Glenwood Shale	0–18	Shale, bluish-gray to bluish-green; generally soft but becomes dolomitic and harder to the east.	Confining bed; locally, some springs issue from the Glenwood–Platteville contact in the river bluffs.
Ordovician	St. Peter Sandstone	0–150+	Sandstone, white, fine- to medium-grained, well-sorted, quartzose; locally iron stained and well cemented; rounding and frosting of grains is common; 5–50 feet of siltstone and shale near bottom of formation.	Not fully saturated throughout area. Most wells completed in the sandstone are of small diameter and used for domestic supply. They yield 9 to 100 gal/min with 1 to 21 feet of drawdown. Two wells known to be used for public supply have been pumped at 600 and 1,250 gal/min. Water occurs under both confined and unconfined conditions. Confining bed near bottom of formation seems extensive and hydraulically separates sandstone from underlying Prairie du Chien–Jordan aquifer. Not considered to be an important source for public supplies, but is suitable source for domestic supplies.
Ordovician	Prairie du Chien Group: Shakopee Dolomite		Dolomite, light-brown to buff, thinly to thickly bedded, cherty; shale partings: commonly sandy and oolitic.	About 2,000 mi² in extent in metropolitan area. Together, the Prairie du Chien Group and the Jordan Sandstone constitute the major aquifer unit. The two are hydraulically connected throughout most of the area, but locally some small head differences may exist owing to intervening low-permeable confining beds of limited extent. *Prairie du Chien:* Permeability is due to fractures, joints and solution cavities in the rock. Yields small to large supplies of water to wells. Pumping rates of up to 1,800 gal-min have been obtained. *Prairie du Chien-Jordan aquifer:* Supplies about 75 percent of ground water pumped in the metropolitan area. Yields of 115 wells (8- to 24- inch-diameter casings), open to both rocks, ranged from 85 to 2,765 gal/min with 3 to 133 feet of drawdown. Higher
Ordovician	Prairie du Chien Group: New Richmond Sandstone	0–250+	Sandstone and sandy dolomite, buff; often missing.	
Ordovician	Prairie du Chien Group: Oneota Dolomite		Dolomite, light-brownish-gray to buff; thinly to thickly bedded, vuggy.	

Cambrian	Jordan Sandstone	0–100+	Sandstone, white to yellowish, *fine-to* coarse-grained, massive to bedded, cross bedded in places, quartzose; commonly iron stained; loosely to well cemented.	obtainable yields seem to reflect closeness to the Mississippi and Minnesota Rivers or to places where the aquifer is overlain directly by glacial deposits, particularly where drift-filled valleys penetrate. *Jordan:* Permeability is mostly intergranular but may be due to joint partings in cemented parts. Main source of water for public supply in metropolitan area. Almost all wells completed in the sandstone are of large diameter. Recorded yields are from 86 to over 2,400 gal/min with 2 to 155 feet of drawdown.
	St. Lawrence Formation	0–65	Dolomitic siltstone and fine-grained dolomitic sandstone; glauconitic, in part.	Confining bed. No wells are known to obtain water from this formation.
	Franconia Formation	0–200+	Sandstone, very fine grained; moderately to highly glauconitic; worm bored in places. Interbedded very fine grained sandstone and shale; mica flakes common. Glauconitic fine-grained sandstone and orange to buff silty fine-grained sandstone (often worm bored).	Small amounts of water may be obtainable from the medium- to coarse-grained members of the formation; very little water from the fine-grained members. Not considered to be an important water source in the area. Records of wells completed only in the Franconia Formation are lacking.
	Ironton Sandstone	0–80+	Sandstone, white, medium- to fine-grained, poorly sorted and silty.	An important aquifer beyond the limits of the Prairie du Chien–Jordan aquifer. Yields of wells range from 40 to 400 gal/min with 4 to 110 feet of drawdown.
	Galesville Sandstone		Sandstone, yellow to white, medium- to coarse-grained, poorly cemented.	
	Eau Claire Sandstone	0–150	Sandstone, siltstone, and shale, gray to reddish-brown, fossiliferous.	Confining bed. Sandstone beds may yield small quantities of water to wells for domestic use. Shale of very low permeability and apparent large areal extent constitutes the main confining bed for water in the underlying aquifer.
	Mt. Simon Sandstone	As much as 200	Sandstone, gray to pink, medium- to coarse-grained. Some pebble zones and thin shaley beds.	Secondary major aquifer, supplies about 15 percent of ground water pumped in the metropolitan area. Recorded yields of municipal and industrial wells ranged from 125 to 2,000 gal/min with 20 to 209 feet of drawdown.
Precambrian	Hinckley Sandstone	As much as 200	Sandstone, buff to red, medium- to coarse-grained, well sorted, well cemented.	
	Red clastics	As much as 4,000	Silty feldspathic sandstone and lithic sandstone, fine-grained; probably includes red shale.	Data are lacking in metropolitan area.
	Volcanic rocks	As much as 20,000	Mostly mafic lava flows, but includes thin interlayers of tuff and breccia.	Rock is at and near the surface at Taylors Falls, north of metropolitan area. Deeply buried in metropolitan area, and no data available.

FIGURE 2.—Photograph of a core of rock from the Prairie du Chien Group at a depth of 483 feet (147 m) in the injection well, 1–S, showing fracture and solution openings in vertical section (× 5).

cemented. The Prairie du Chien Group and the Jordan Sandstone are hydraulically interconnected, to some degree, at the test site, but in some other areas a layer of fine-grained material occurs at the contact. The St.

TABLE 2.—*Semiquantitative X-ray-diffraction analysis, in percent, of the cores from the St. Peter Sandstone, Prairie du Chien Group, Jordan Sandstone, and St. Lawrence Formation*

[Analyses by Barbara J. Anderson, U.S. Geological Survey Hydrologic Laboratory, Denver, Colo. ≈, approximately; <, less than; leaders (....) indicate not present]

Geologic unit	Depth (ft) From	Depth (ft) To	Quartz	Dolomite	Calcite	Potash feldspar	Plagioclase feldspar	Other minerals
Core samples, well 1-S								
Prairie du Chien Group	442	443	≈5	90-95	....	....	1-2	
Do	469	469.7	2	≈95	....	....	1-2	
Do	482.5	483	15	75-80	....	....	1-2	Pyrite, <5
Prairie du Chien Group (new Richmond Sandstone).	484	....	25-30	60-65	≈1	1-2	....	Anhydrite(?), <2.
Prairie du Chien Group	497	....	6	85-90	....	....	....	Pyrite, 5-10.
Drill-cutting samples, well 1-S								
St. Peter Sandstone	265	270	≈60	7	6	1-2	....	Pyrite, <5.
Do	280	285	≈95	2	2	≈1	....	
Do	360	365	≈90	....	≈1	5-8	....	
Do	390	395	85-90	1	1-2	≈10	....	
Do	405	410	70-75	2	1-2	≈15	...	Pyrite, <3.
Prairie du Chien Group	418	419	13	75-80	1-2	≈5	....	Anhydrite(?), <3.
Do	420	424	20-25	70-75	....	3-5	....	
Do	435	440	15	80-85	....	2-3	....	
Do	440	445	15	≈80	2	1-2	1-2	
Do	460	465	19	75-80	....	2-3	≈1	
Do	475	480	8	≈90	....	....	≈1	
Do	490	495	30	60-65	....	2-3	....	Pyrite, <3; anhydrite(?), <2.
Do	515	520	12	85-90	....	....	1-2	
Do	520	525	5	≈90	....	≈1	2-3	
Do	525	530	7	≈90	....	1-2	1-2	
Do	535	540	5	≈90	....	≈1	2-3	Anhydrite(?), <2.
Do	540	545	6	≈90	....	1-2	1-2	Anhydrite(?), <2.
Do	545	550	5	≈90	....	....	2-3	
Drill-cutting samples, well 1-N								
Jordan Sandstone	561	565	≈95	4	≈1	....	....	
Do	575	580	≈90	4	....	≈4	....	
Do	580	585	60-65	≈25	1-2	≈10	....	
Do	590	595	≈95	2	....	....	....	
Do	615	620	55-60	10-15	1-2	≈20	....	
Do	625	630	≈75	≈15	1-2	≈10	....	
St. Lawrence Formation	645	650	25-30	≈20	2-3	10-15	....	Illite, 20-30; pyrite, <3

Peter Sandstone, overlying the dolomite of the Prairie du Chien Group, is a more homogeneous aquifer, but it varies in cementation. The basal part of the St. Peter Sandstone at the experiment site is an effective confining layer. The St. Lawrence Formation underlying the Jordan Sandstone is a confining bed. The two confining layers create an artesian system in the Prairie du Chien-Jordan aquifer.

The hydraulic conductivity of the dolomite-rock matrix of the Prairie du Chien is insignificant compared with that resulting from the fissures. For this reason, no attempt was made to determine the hydraulic conductivity of the dolomite. The horizontal hydraulic conductivity of core samples are 4.6 ft/d (1.4 m/d) for the Jordan Sandstone and 0.16 ft/d (0.05 m/d) for the base of the St. Peter Sandstone, as determined in the laboratory. The total porosities of two samples of the Jordan were 19.9 and 22.9 percent. The effective porosities of the same two samples were 19.1 and 22.6 percent, respectively. These porosity values are probably higher than average or than is representative of the Jordan Sandstone at this location. The neutron log of well 1-N shows that the top of the Jordan is slightly more porous than the rest of the formation.

ALTITUDES OF WATER LEVELS IN AQUIFERS

Each aquifer has its own potentiometric surface (the altitude to which water will rise in a well completed in an aquifer), which is lower in each successively deeper aquifer at the test site. The water level in the drift is 1,011 to 1,025 feet (308 to 312 m) above mean sea level. Although the Prairie du Chien and the underlying Jordan Sandstone are hydraulically connected, the water levels in the Prairie du Chien wells are at an altitude of 724 feet (221 m) and in the Jordan well at an altitude of 722 feet (220 m). (See table 4.)

Regionally, the water level in the St. Peter Sandstone is at an altitude of about 735 feet (224 m). However, the water level in well 3-N, which is finished in the St. Peter Sandstone, was at an altitude of 962 feet (293 m), indicating that well 3-N is open also to the shallower Platteville Limestone. The 238-foot (73-m) difference in head between the water level in the St. Peter well and the water level in the Prairie du Chien well 55 feet away indicates that the base of the St. Peter acts as a very effective confining bed at the test site.

MOVEMENT OF WATER

The successively lower water levels in successively deeper aquifers indicate that the site is in an area of natural recharge. That is, the availability of water in the upper aquifers and the increase in the head differential with depth form the head-potential conditions by which water moves downward from one aquifer to another. The rate of movement and the amount of water moving downward in a unit cross-sectional area, however, is small compared with the lateral movement through aquifers, such as the dolomite of the Prairie du Chien Group or the Jordan Sandstone.

Movement of water through an aquifer is governed by the hydraulic characteristics of the aquifer and by the head difference from place to place. The experiment site is near the center, or divide, from which the ground water moves westward, northward, and eastward toward the Mississippi River (Reeder, 1966; Norvitch and Brietkrietz, 1971). The altitude of water levels in three wells (1-S, 2-S, and 2-N) in the dolomite of the Prairie du Chien Group infers that the water moves northeastward. From April 7, 1971, through February 2, 1972, the direction of flow ranged from about N. 20° E. to N. 60° E., and averaged N. 40° E. The hydraulic gradient ranged from 0.00092 to 0.0013. The average gradient was 0.0012. The variations in direction of flow and in the gradient were probably due to the effects of seasonal pumping from supply wells 2 to 3 miles (3.2 to 4.8 km) from the test site. The gradient and the direction of flow immediately before injection were nearly the same as the average for the year. Table 3 lists water levels in the three wells from which the gradient and direction of flow, also listed, were computed. Figure 25 shows the flow direction in relation to the well locations.

WATER-LEVEL FLUCTUATIONS

The potentiometric surface fluctuates, rising with recharge of water to the ground-water reservoir and declining with discharge. Natural fluctuations are commonly small compared with fluctuations resulting from pumping. The cause of natural fluctuations, such as changes in atmospheric pressure, must be taken into account when computing the aquifer characteristics from pumping-test data.

Changes in atmospheric pressure cause changes in water level in wells under certain conditions (Ferris and others, 1962, p. 83-85). Normally, under artesian conditions, as are found in dolomite of the Prairie du Chien Group and in the Jordan Sandstone at the test site, as atmospheric pressure increases, the water level in a well declines, and, as the atmospheric pressures decreases, the water level in a well rises. Water levels in the Prairie du Chien and Jordan wells fluctuated as much as 1.2 feet (0.37 m) in response to atmospheric-pressure changes.

TABLE 3.—*Static water levels in wells tapping the injection aquifer (fractured dolomite of the Prairie du Chien Group)*

Date	Altitude of water level, in feet above mean sea level, observed in wells			Estimated flow direction and hydraulic gradient
	Well 1-S	Well 2-S	Well 2-N	
4- 9-71	724.10	724.16	723.73	N. 53° E., 0.0009
5- 7-71	723.04	723.12	722.63	N. 59° E., 0.0011
7-30-71	720.47	720.49	719.92	N. 24° E., 0.0011
8-29-71	717.54	717.58	716.97	N. 33° E., 0.0012
10-20-71	721.47	721.52	720.85	N. 36° E., 0.0013
11-29-71 (5 days after stop of injection)	724.33	724.37	723.75	N. 32° E., 0.0013
12-15-71	724.22	724.27	723.68	N. 39° E., 0.0012
2- 2-72	723.46	723.47	723.04	N. 58° E., 0.0010

The barometric efficiency of the Prairie du Chien and Jordan wells at the site is about 80 percent. That is, an atmospheric-pressure change of 1.00 foot (0.3 m) of water (0.883 in. or 22.4 mm of mercury) will cause water levels in wells to change 0.80 foot (0.24 m). As water levels in the glacial drift and St. Peter wells were unaffected by pumping at the site, no attempt was made to use the data in the aquifer-test analysis nor to determine atmospheric-pressure effects on the water levels in those aquifers.

In addition to the effects of pumping at the test site, water levels in the Prairie du Chien and Jordan wells fluctuate in response to regional pumping. Two patterns of fluctuations—one daily and one weekly—that are the result of regional pumping are noted during the winter, when pumping is limited to year-round uses. The weekly fluctuation occurs in response to the pumping in the South St. Paul area, about 3 miles (5 km) to the east. At that time water levels gradually decline generally from Monday until Friday and rise from Friday until Monday. (See fig. 12.) At the test site the weekly water-level change ranged from 0.2 foot to 1.2 feet (0.06 to 0.37 m). The daily fluctuation results from daily changes in pumping in the South St. Paul area. At about 0900 hours (9 a.m.) water levels in the Prairie du Chien observation wells decline 0.1 to 0.2 foot (0.03 to 0.06 m) in about 45 minutes and then gradually rise to their former levels.

In the summer the water-level fluctuations due to regional pumping have no definite pattern with time, but the water levels fluctuate as much as 4 feet (1.2 m) in a few hours. Aquifer tests of more than a few hours' duration in the summer would not be easily analyzed because of the effects of regional pumping.

INJECTION AND PUMPING TESTS

Aquifer tests by both injection and pumping were made at the site. A series of aquifer pumping tests were preformed before injection to determine the hydraulic characteristics of the injection well and the hydraulic properties of the aquifer, and to provide samples for determining chemical and bacterial quality of the native ground water. An injection test was made to determine the hydraulic behavior of the well and aquifer during injection and to determine chemical and bacterial changes induced by recharging treated surface water. After injection, additional aquifer pumping tests were run to determine, by comparison with the pre-injection aquifer pumping tests, changes in the hydraulic characteristics of the injection well and in the hydraulic properties of the aquifer that were induced by injection. Also, this pumping allowed for additional sampling of the chemical and bacterial effects of recharge.

Problems common to recharge operations and for which experiments were incorporated into the injection test can be categorized in the fields of hydraulics, chemistry, and microbiology. The multipurpose nature of the injection test multiplied problems of instrumentation and of data collection. The site could be instrumented and observations made with relative ease for one facet of the project, but the several fields of interest are interrelated, and none could be omitted. Basic knowledge of the system is needed initially in all fields of activity before a specialized study of a single approach can be made. In practice, however, some loss of data for one field is necessary in order to gain information in another. For example, some water-level data and some degree of accuracy are sacrificed so that water samples can be collected. Conversely, the best method of collecting water samples may have to be somewhat compromised in order to measure water levels.

WELL CONSTRUCTION

Seven wells, the minimum deemed necessary, were installed at the site to obtain drill samples and geophysical logs, to determine aquifer characteristics, to observe water levels, to inject water and observe its effects, to determine the chemical quality of the ground water, and to observe chemical and biological aspects of recharge. The layout of the seven wells at the test site is shown in figure 3. The injection well, 1–S, and two observation wells, 2–S and 2–N, were finished in the Prairie du Chien Group. Observation well 1–N was finished in the underlying Jordan Sandstone, and observation well 3–N was finished in the overlying St. Peter Sandstone. Two shallow piezometers, 3–S and 4–N, were installed in the upper part of the glacial drift. Figure 4 is a photograph of the south wells and instrument shelters.

Drilling was begun on September 3, 1970, and was completed on January 14, 1971. Four 6-inch (150-mm)-diameter observation wells were drilled by the hydraulic-rotary method, and the 24-inch (610-mm)-diameter injection well was drilled by the cable-tool method. Samples of drill cuttings were collected at 5-foot (1.5-m) intervals and at formation changes.

Observation wells 1–N, 2–S, and 2–N (fig. 3) were drilled to the top of the Prairie du Chien Group with a bentonite drilling mud and to completion with a biodegradable drilling fluid. Well 3–N was drilled with bentonite to the top of the St. Peter Sandstone and with biodegradable mud to completion. Use of the self-destroying drilling fluid minimized plugging of the formation; therefore, less extensive washing and developing of the wells were needed. Little or no drilling mud was needed in the cable-tool drilling; however, biodegradable drilling fluid was used in the pilot hole, and coring was done with the rotary drill in the Prairie du Chien part of the injection well, 1–S, before that part of the hole was enlarged with the cable-tool drill. The purpose of the 5-inch (130-mm) pilot hole in the injection well was to collect core samples at several intervals and to provide a small hole for geophysical logging. Details of well construction are given in table 4 and figure 5. Geophysical logging was done in the injection well and the four observation wells. The logs made include resistivity, spontaneous potential, gamma, neutron, gamma-gamma, and caliper.

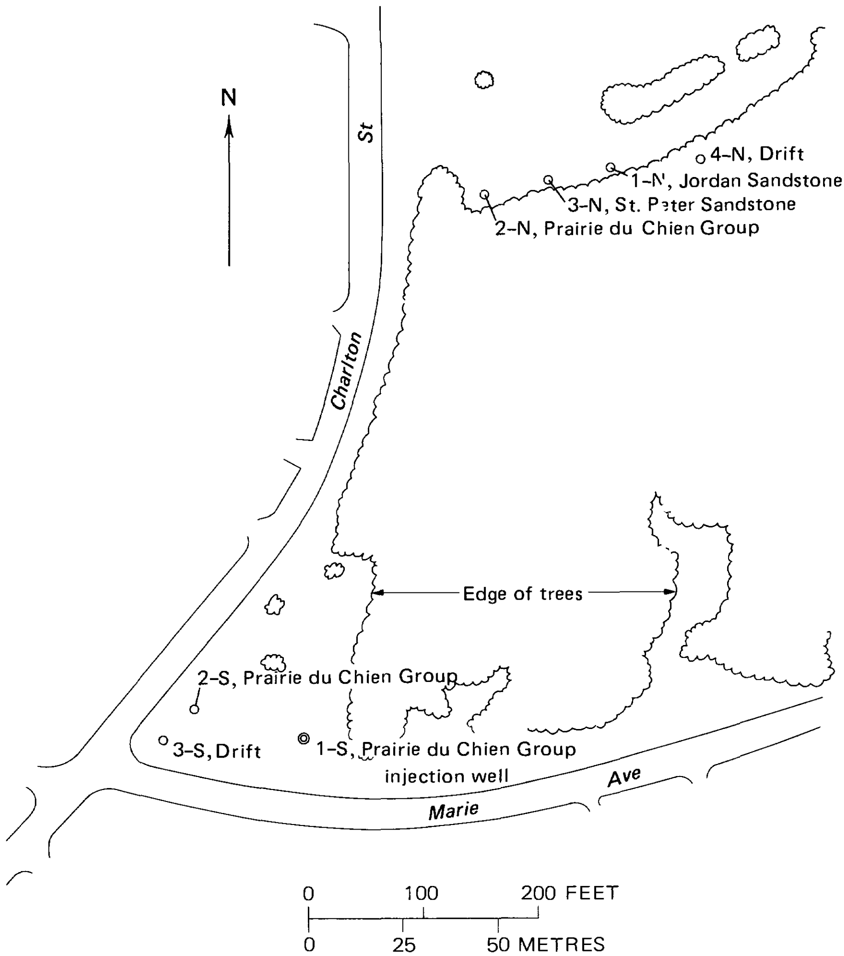

FIGURE 3.—Location of wells at the artificial-recharge site.

Two shallow piezometers were placed in glacial drift with a power auger. Holes 4–N and 3–S were augered 76 feet (23 m) and 57 feet (17.4 m) deep, respectively, into glacial drift and 1¼-inch (32-mm)-diameter sandpoints and pipe were placed at a depth of 55 feet (17 m) in each hole.

INSTRUMENTATION AND FIELD PROCEDURES

PUMPING AND INJECTION FACILITIES

The initial aquifer tests in January and February 1971 included a preliminary production test and a constant-rate aquifer test. An electrically driven deep-well turbine pump was used for these aquifer tests. After the injection test, a deep-well turbine pump was placed in well 1–S and powered with a diesel engine. (Pumping with the diesel engine was per-

FIGURE 4.—The recharge site, 1–S, injection well in Prairie du Chien Group; 2–S, observation well in Prairie du Chien Group; and 3–S, observation piezometer in glacial drift.

mitted only between 0600 hours (6 a.m.) and 2200 hours (10 p.m.) because of a local prohibitive ordinance.)

The water injected was city-treated water derived from the Mississippi River and several lakes. A 4-inch (100-mm) water-service connection was installed from the 12-inch (300-mm) water main under Marie Avenue to a point above ground level, 2.5 feet (0.76 m) from the well head. At this point, the pipe size was reduced to 3 inches (77 mm). The pipe system included valves for control of flow rate and devices for measuring flow rate and volume, for monitoring pressure, and for sampling and monitoring chemical and bacterial quality. Figure 6 is a diagram of pipe and fittings installed above ground. Figure 7 shows the pipes and equipment installed inside the small house.

A major problem in some past experiments of recharge through wells has been plugging of the formation or wells with air bubbles. The source of the problem has been air entrainment in the water and release of dissolved gasses from the water under decreased pressure caused by dropping the water from near land surface to the water level in the well and out an open-end pipe. In some experiments a control valve was placed below the water level in the injection well to maintain a full pipe and positive

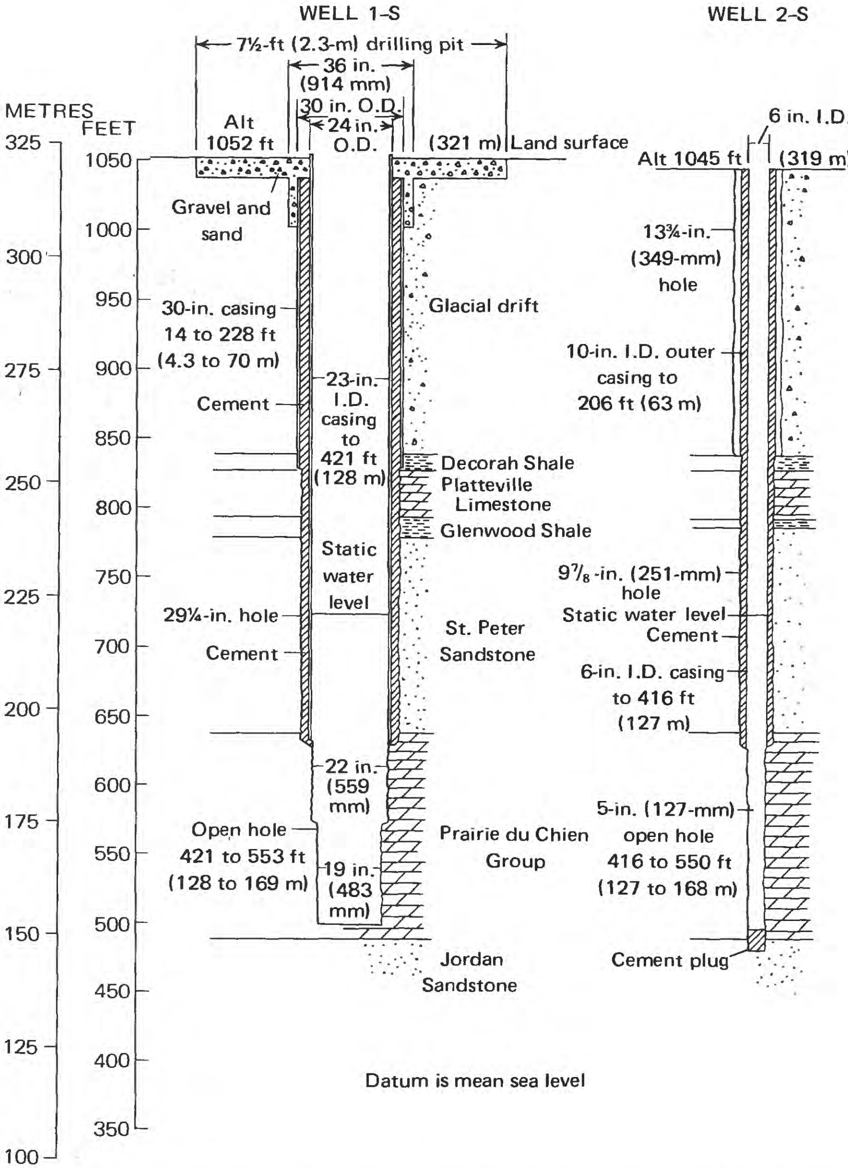

FIGURE 5 (above and next two pages).—Diagrams of wells constructed for recharge experiment.

pressure to prevent air entrainment and gas release. Placing the valve downhole, however, requires a large-diameter well and a means for controlling the valve from the ground surface.

In the present study, the injection system was designed to use pipe friction instead of a valve to maintain positive pressure in the pipe. The pipe

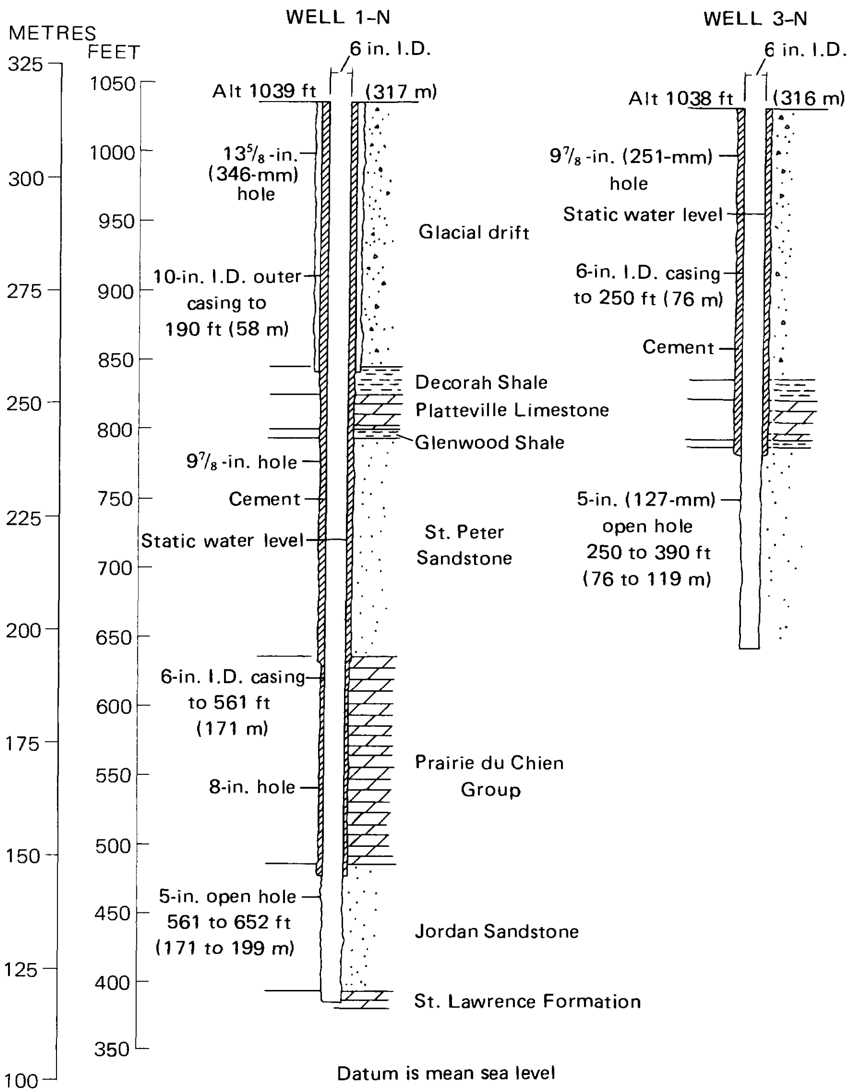

FIGURE 5.—Continued.

size was selected for the desired injection rate on the basis of the maximum flow that a pipe can carry when friction causes unit head loss for each unit length of pipe.

The following form of the Hazen-Williams formula (King, 1939, p. 176) was used in the computation:

$$v=1.318Cr^{0.63}s^{0.54},$$

in which v is the mean velocity of the water, in feet per second; C is the Hazen-Williams roughness coefficient of the pipe; r is the mean hydrau-

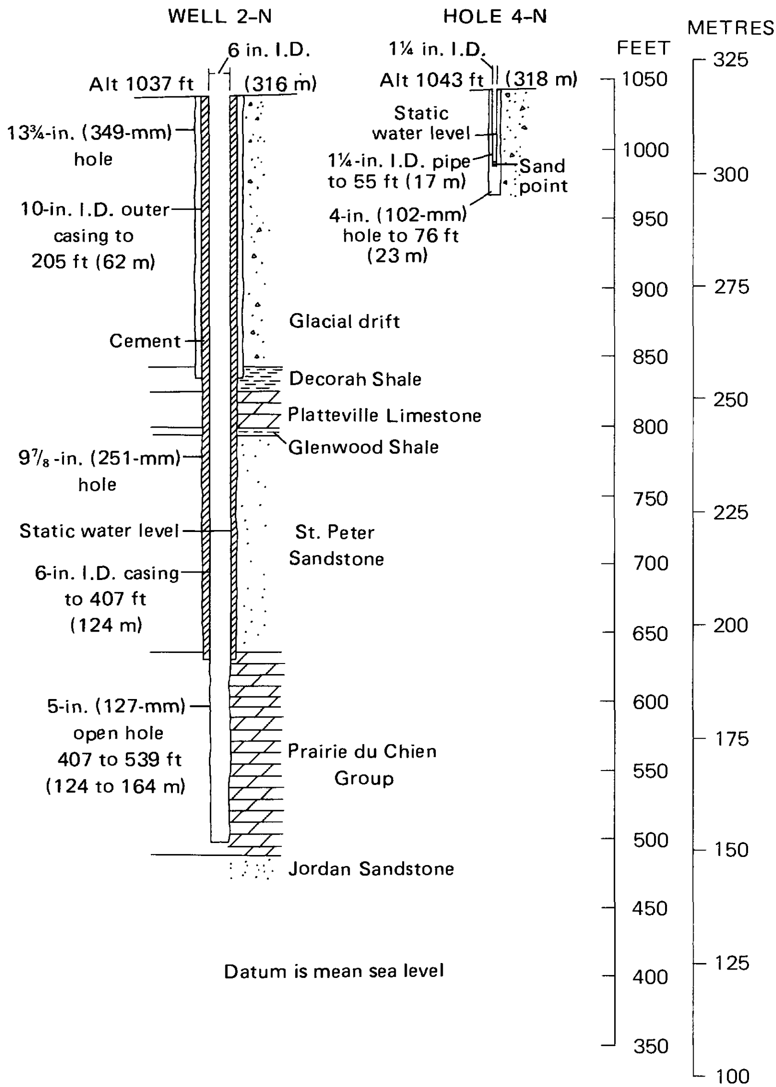

FIGURE 5.—Continued

lic radius = $d/4$; s is the head loss, in feet per foot length of pipe, the mean slope of the hydraulic gradient; and d is the inside diameter of the pipe, in feet. The form of the equation can be modified to solve for the rate of flow as $Q=vA$; in which A is the cross sectional area of the pipe. The formula becomes:

$$Q=1.318C\frac{\pi d^{2.63}}{4^{1.63}}s^{0.54}.$$

TABLE 4.—*Summary of well data for wells constructed at the recharge-experiment site*

Well	Altitude of land surface (ft above mean sea level)	Depth of well (ft)	Casing: Diameter (nominal size) (in.)	Casing: Depth (ft)	Open hole: Diameter (in.)	Open hole: Depth (ft) below land surface	Formation tapped by well[1]	Water level: Depth to water below land surface (ft)	Water level: Date	Water level: Altitude of water (ft above mean sea level)	Outer casing: Diameter (nominal size) (in.)	Outer casing: Depth (ft)
1-S.........	1,051.59	553	24	421	22	421–480	P. du C................	327.49	4-9-71	724.10	30	228
					20	480–553	P. du C................					
2-S.........	1,045.32	550	6	416	5	416–550	P. du C................	321.16	4-9-71	724.16	10	206
3-S.........	1,043.45	55	1¼	53	Sandpoint........		Drift......................	18.55	4-9-71	1,024.90	None.	
1-N.........	1,039.01	652	6	561	5	561–652	Jordan..................	317.15	4-9-71	721.86	10	195
2-N.........	1,036.85	539	6	407	5	407–539	P. du C................	313.12	4-9-71	723.73	10	205
3-N.........	1,038.09	390	6	250	5	250–390	St. Peter................	75.77	4-9-71	[2]962.32	None.	
4-N.........	1,043.18	55	1¼	53	Sandpoint........		Drift......................	32.18	4-9-71	1,011.00	None.	

[1]Formation: P. du C., Prairie du Chien Group; Jordan, Jordan Sandstone; St. Peter, St. Peter Sandstone; Drift, upper part of glacial drift.

[2]Well 3-N apparently is open to the *Platteville Limestone* in addition to the St. Peter Sandstone. The regional altitude of the water level in the St. Peter only should be about 735 feet above mean sea level, on the basis of other evidence.

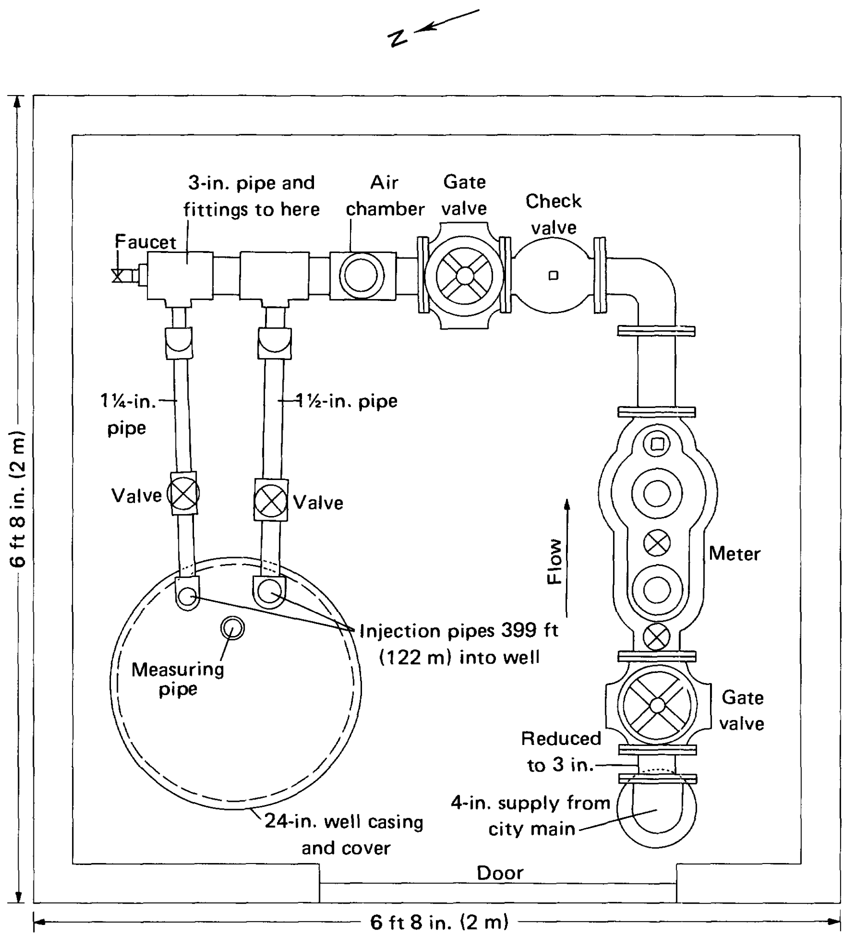

FIGURE 6.—Injection-pipe layout and arrangement at the injection well, 1-S, at the recharge test site.

With the design head loss gradient due to friction of s=1, then $s^{0.54}$=1, the formula becomes

$$Q=\frac{1.318\ \pi C d^{2.63}}{4^{1.63}}=0.432 C d^{2.63},$$

where Q is rate of flow, in cubic feet per second.

The Hazen-Williams formula is most applicable for pipes of 2 inches (50 mm) or larger and velocities less than 10 ft/s (3 m/s) (Chow, 1964, p. 7-18). However, results from the formula in this design were valid. An advantage of this formula is that all problems have direct solutions.

Two pipes, 1¼-inch (32-mm) and 1½-inch (38-mm) diameter heavy duty wrought iron black pipe, were installed from the supply pipe, with a static pressure of 125 feet (38 m), to 75 feet (23 m) below the water level in

FIGURE 7.—Equipment installation during injection at well 1-S.

the well. The pressure head in the injection pipe at the top of the well remained fairly uniform at about 97 feet (30 m) after the valve was fully opened during injection. The injection head loss (difference in entrance and exit pressure heads) remained about the same after the first day of injection. The maximum (uncontrolled by valve) injection rates, with a friction loss gradient of 1 in this system are: 70 gal/min (4.4 l/s) through the 1¼-inch pipe, 106 gal/min (6.7 l/s) through the 1½-inch pipe, and 176 gal/min (11.1 l/s) through both pipes together (based on a *C* value of 130 for new steel or cast-iron pipe). A graph of flow rates for small pipes is shown in figure 8. The range of values in the graph generally are not shown in published pipe-friction tables because the flow rates and head losses exceed those normally considered in pipeline design. Olson (1966, p. 248) presents a nomograph of pipe flow for the Hazen-Williams formula, with C=100, in the range of rates and head losses used in pipe-

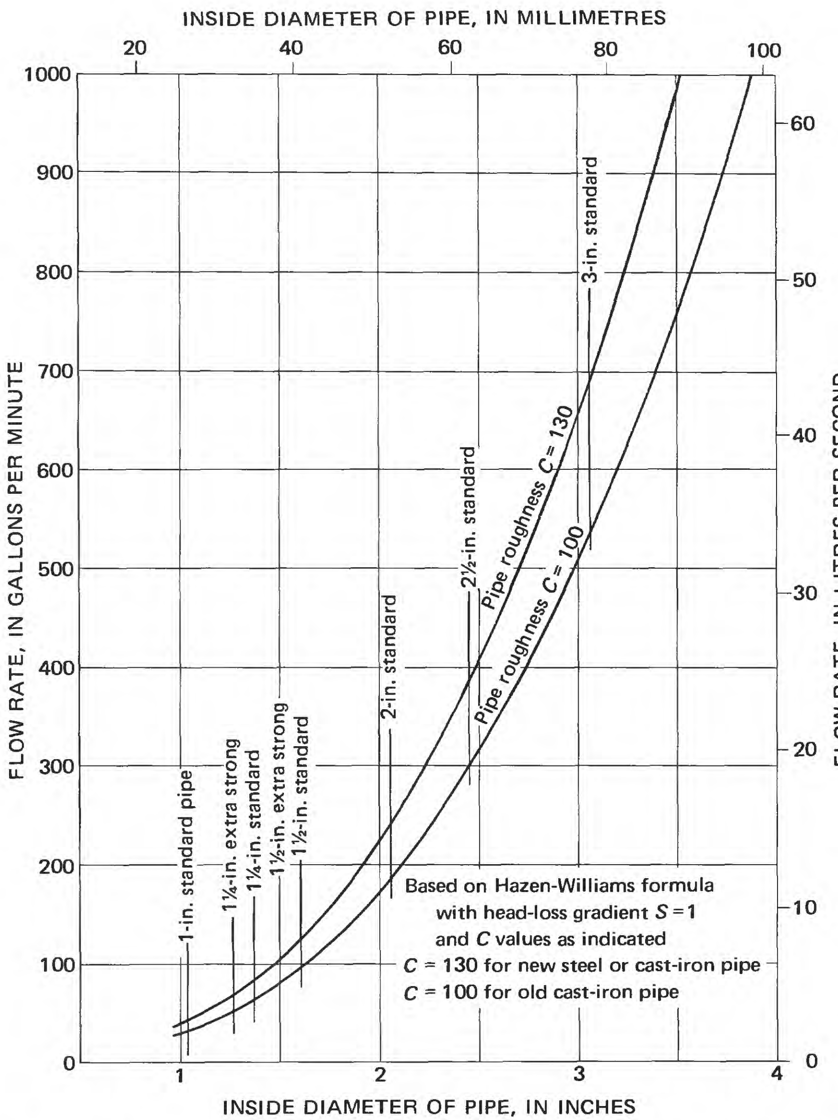

FIGURE 8.—Flow rates in small pipes, with unit head loss per unit length of pipe.

line designs. By extending the head-loss scale to 1,000 feet (305 m) of head loss per 1,000 feet (305 m) of pipe, approximate values could be determined comparable to those shown in figure 8.

During the injection test, maximum flow rates varied. The variation was probably due in part to the change in water temperature, suggesting that a viscosity correction factor is needed in the preceding equations, at least for the range of pipe sizes and friction head loss used in the computations. Part of the decrease in flow rate could possibly be due to a change in pipe roughness.

MEASURING AND SAMPLING PROCEDURES

Recording gages were installed on all the observation wells used for all testing. Water-level changes in the pumped well during the initial aquifer pumping tests were measured with an electric cable. During the injection test a recording gage coupled with an electronic water-surface seeking device were used on the injection well to record water levels continuously. During the second pumping series (pumping after injection), an electric tape was used to measure water-level changes in the large well. A microbarograph was in operation for the duration of the project to provide data for correcting water levels in the Prairie du Chien and Jordan wells for barometric effects. Discharge rates were measured with an orifice meter at the well head for all the pumping tests.

Water samples collected at the well head of the pumped well during the initial aquifer tests were analyzed for chemical content and for total bacteria count. In addition to standard laboratory analyses, field determinations were made of time-critical constituents and properties. A small pump was placed in each of the four deep observation wells in June and October 1971 to collect representative water samples from the aquifers for laboratory and field analyses. Wells 3–N in the St. Peter Sandstone and 1–N in the Jordan Sandstone were pumped for about 7 hours each at nearly 10 gal/min (0.63 l/s) in June. Wells 1–N, 2–N, and 2–S were pumped in October for 64 hours, 22 hours, and 42 hours, respectively, and water samples were obtained.

During the injection test, a chemical-feed injector (tank and pump) was used to add sodium chloride (canning salt) to the injection water and thus magnify the difference in chloride concentration between the injection water and the native ground water. The salt solution was fed into the pipe downstream from a tap used to collect samples of water directly from the city water main before any treatment at the site. The chloride concentration of the water, as delivered to the site, was 19 mg/l before addition of the salt solution. The chloride concentration of the injection water was increased to an average of 25 mg/l. The native ground water contained about 1 mg/l chloride.

An automatic water sampler (fig. 9) was used to collect 250-ml water samples at 2-hour intervals throughout the injection test. The sampler removed water from the pipe downstream from the chemical-feed-injector pump. Specific conductance and turbidity were determined for all samples. One or two samples were selected each day, and chloride, pH, and bicarbonate were determined. Additionally, a sample of the city water without added chloride was collected daily, and determinations were made for chloride, specific conductance, pH, bicarbonate, and tempera-

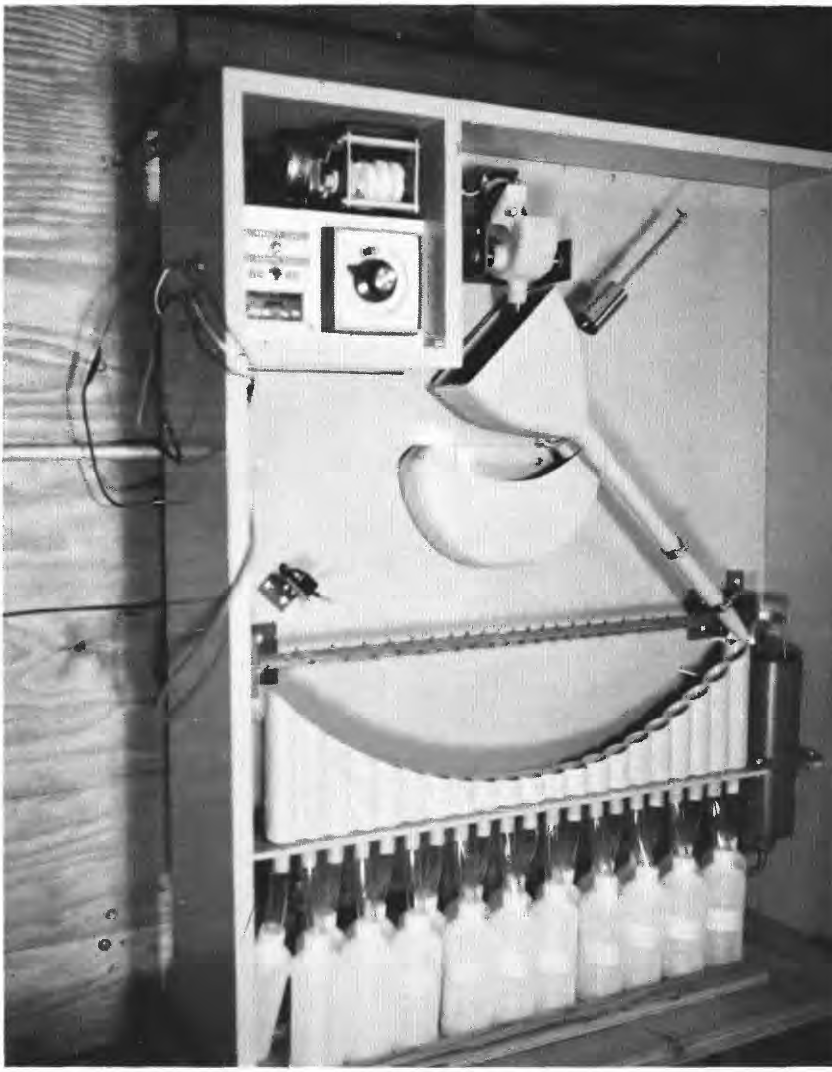

FIGURE 9.—Automatic water sampler used during injection and repumping operations at well 1–S.

ture. The analyses of samples of water before and after addition of salt were compared as a check on the amount of salt added and chloride injected into the well. (See sections entitled "Geochemical Aspects of Artificial Recharge" and "Hydrodynamic Dispersion and Movement of Injected Water.")

Observation wells in the Prairie du Chien Group, 2–S and 2–N, and in the Jordan Sandstone, 1–N, were sampled daily for chloride concentration, specific conductance, and pH determinations. Three water samples were collected daily from each well—from the top, middle, and bottom of the formation open to the well. A ball-valve sampler was used to bail samples from the observation wells on all days but one. On November 19, a plunger-valve sampler was used. The ball-valve sampler was also used to take water samples from the injection well and the observation wells periodically between the end of injection and the start of pumping.

During the second set of pumping operations, samples of the pumped water were collected for standard complete laboratory analysis. Additional samples for chloride and specific-conductance analyses were collected with an automatic water sampler at 2-hour intervals. Samples were prepared at the site for total bacteria count and for sulfate-reducing bacteria.

DESCRIPTION OF TESTS

TIMES AND DISCHARGE OR RECHARGE RATES DURING TESTS

Series of pumping tests were run both before and after the injection test. The hydraulics part of the pumping tests consisted of pumping one well at a uniform rate and observing effects in the pumped well and observation wells. The injection test consisted of running water into the injection well within a closed-pipe system and observing the effects in the injection well and in observation wells.

The two principal hydraulic properties of aquifers that influence the yield of wells and the accompanying lowering of water levels are transmissivity (the hydraulic conductivity multiplied by the thickness of the aquifer) and storage coefficient. A prediction of the rate of change of water levels and the effects of artificial recharge depends largely on accurate determinations of these parameters. Aquifer tests measuring magnitude and rate of decline of water levels in response to pumping at a known rate commonly yield information from which aquifer characteristics can be computed. The extent to which an aquifer departs from an ideal homogeneous aquifer and the effectiveness of well construction and development mainly govern the accuracy of aquifer test results.

Pumping tests were made in January and February 1971 to determine aquifer and well characteristics before injection of water and to obtain data for the design of the injection well system. A summary of this pumping is listed in table 5. The aquifer test on January 26 and 27, 1971, had to be stopped after only 16 hours of pumping, and data from the initial part of the test were used in comparison with step-drawdown tests only.

The injection test (table 6) was run in November because the injected water should have nearly the same temperature as the native ground water at this time. Effects of temperature differences on the test results would thus be avoided. For example, the viscosity of water changes with temperature which, in turn, affects the specific capacity of the pumped well. However, this assumption did not hold true, and the temperature of the injection water ranged from 15.0°C at the beginning of the test, to 11.2°C at the end, as compared with a ground-water temperature of 10°C. In some past years, the average of the municipal water in November was lower—8° to 9°C.

The rate of injection, using only the 1½-inch (38-mm) pipe, was started at 108 gal/min (6.8 l/s) but gradually decreased to 90 gal/min (5.7 l/s). Figure 10 shows graphs of injection rate and water temperature. In addition to the gradual decrease in flow rate, also noted was a daily fluctuation in flow rate due to daytime decreased pipeline pressure caused by higher daytime water use. Momentary line-pressure fluctuations also were noted at the time of momentary fluctuations of a few hundredths of a foot in water level in the injection well. Table 6 lists accumulative injected volumes, daily flow rates, and daily temperature of the injected water. Total injection was 368,200 ft^3 (10,430 m^3) at an average rate of 18,410 ft^3/d (521.4 m^3/d).

After the injection test, the second series of pumping tests was made to pump out the injection water, to determine whether the injection caused a change in aquifer and well characteristics, and to make chemical-quality and bacterial analyses. Two million gallons (7,570 m^3) of water was pumped from February 25 to March 8, 1972. Table 7 shows the pumping times and rates and the amount pumped each day. The first day, February 25, was used primarily to sample water for chemical and bacterial analysis. The second day, February 28, a controlled aquifer test of drawdown (10 hours) and recovery (10 hours) was performed. The first 4½ hours

TABLE 5.—*Pumping intervals, rates, and volumes during the initial pumping operations, January-February 1971, well 1-S*

Pumping began		Pumping stoped		Rate of pumping (gal/min)	Volume of water pumped (gal)	Remarks
Date	Time	Date	Time			
1-20-71	1030	1-20-71	1615	400–1,000	([1])	Development and preliminary production test.
1-22-71	1230	1-22-71	1401	250	22,860	
1-25-71	0830	1-25-71	1000	508	45,720	
1-25-71	1300	1-25-71	1430	752	67,680	
1-26-71	0845	1-27-71	0045	556	533,769	Aquifer test.
2- 3-71	1240	2- 5-71	1240	458	1,319,000	Aquifer test.

[1]Not measured.

TABLE 6.—*Cumulative injected volume, daily injection rates, and temperature of injection water during injection test, November 1971.*

NOTE.—Additional readings were made during the test.

Day in Nov. 1971	Time	Cumulative injected volume (ft³)	Cumulative injected volume Thousands of gallons	Rate of Flow: At time measured (gal/min)	Rate of Flow: Average for 24-hour period (gal/min)	Water temperature (°C)
Start. 4	0930	0	0	0	107.0	14.5
5	0745	19,138	143.1	105.5	105.8	15.0
6	0755	39,615	296.3	104.5	103.5	14.5
7	0815	59,821	447.4	103.0	101.0	14.5
8	0745	78,877	590.0	100.5	100.2	14.2
9	0745	98,182	734.4	100.0	98.9	14.0
10	0750	117,262	877.1	98.8	97.4	13.8
11	0745	135,953	1,016.9	100.0	97.3	13.5
12	0740	154,438	1,155.2	98.6	95.2	13.0
13	0750	172,874	1,293.1	94.4	94.3	13.0
14	0805	191,166	1,429.9	94.1	93.2	13.0
15	0830	209,459	1,566.7	93.4	93.0	12.8
16	0740	226,722	1,685.9	93.5	92.4	12.5
17	0735	244,443	1,828.4	96.6	91.7	12.0
18	0740	262,142	1,960.8	96.8	91.7	12.0
19	0745	279,850	2,093.3	93.4	90.9	12.0
20	0835	297,937	2,228.6	92.8	91.0	12.0
21	0805	315,075	2,356.7	90.6	90.5	11.8
22	1125	334,908	2,505.1	91.6	90.3	11.5
23	1410	354,247	2,649.8	89.4	90.1	11.4
24 End.	0930	368,200	2,754.1	89.9	....	11.2

of pumping on February 29 was used primarily for a step-drawdown test. The remainder of the pumping was primarily for water sampling.

WATER-LEVEL FLUCTUATIONS DURING TESTS

AQUIFER TESTS BY PUMPING

Water levels in the Prairie du Chien observation wells, 2-S and 2-N, responded immediately to pumping from the Prairie du Chien well (1-S), whereas, at the beginning of pumping, response in the Jordan well (1-N) lagged 3 to 7 minutes behind that in the Prairie du Chien wells (fig. 11). Also, the early water-level change in the Jordan well was not so abrupt and rapid as it was in the Prairie du Chien well. These effects are primarily due to the lower vertical hydraulic conductivity at the contact between the two rock units and to the difference in the hydraulic characteristics of the two rock types. A minor part of these effects is due to the greater distance to the Jordan well from the pumped well. The Jordan well, 1-N, is 560 feet (171 m) from the pumped well, as compared with 495

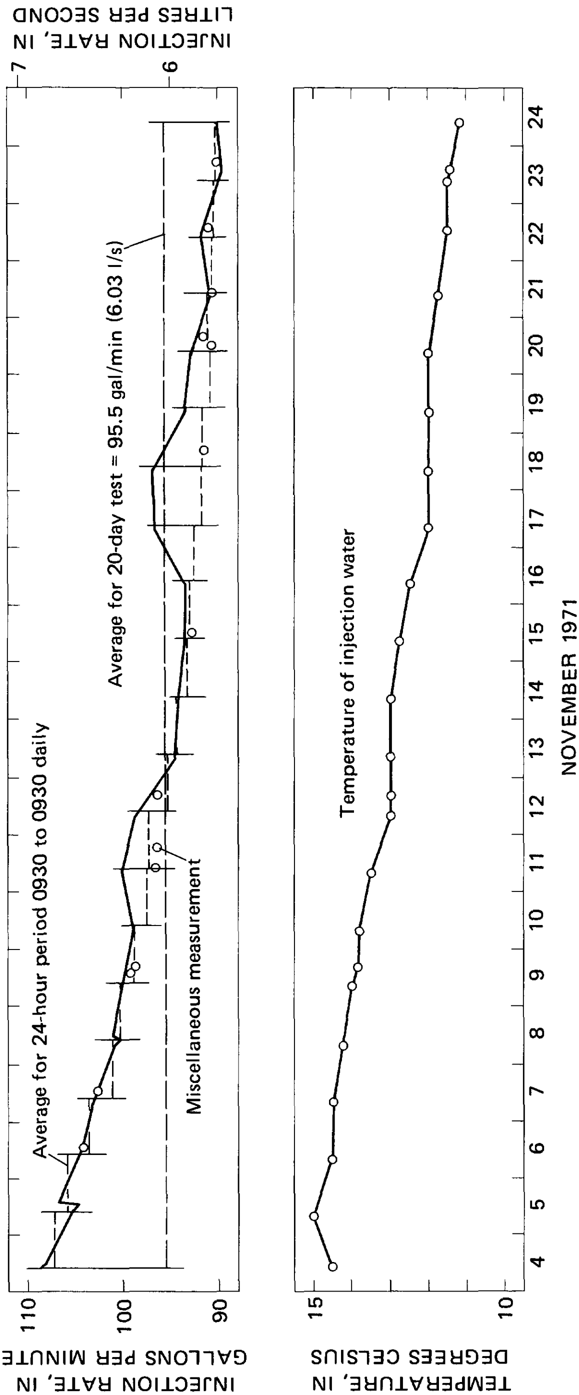

FIGURE 10.—Injection rate and temperature of injection water during injection test.

TABLE 7.—*Pumping intervals, rates, and volumes during second pumping operations (after injection), February-March 1972*

[Leaders (....) indicate no data available]

Date	Time of pumping	Rate of pumping (gal/min)	Volume pumped for the day (gal)	Remarks
2-25-72	1036-1600	375	122,000	Sampling.
2-28-72	0811-1800	485	266,700	Aquifer drawdown-recovery test.
2-29-72	0804-0932	242	....	Step-drawdown test.
Do	0932-1102	334	....	
Do	1102-1232	455	....	
Do	1232-1800	455	233,100	
3- 1-72	0800-1800±	395	239,000	
3- 2-72	0800-1805	400	242,400	
3- 3-72	0757-1001	285	....	
	1001-1605	365	169,500	
3- 6-72	0755-1810	355	217,000	
3- 7-72	0800±-1800±	365	214,000	
3- 8-72	0800±-1817	....	296,600	

feet (151 m) from the pumped well for the Prairie du Chien well 2-N. The total drawdown in the Jordan well several hours after pumping from the Prairie du Chien began was not as great as it was in the Prairie du Chien well. Water levels in the glacial drift and St. Peter Sandstone observation wells did not respond to pumping from the Prairie du Chien, indicating that the degree of hydraulic connection between the Prairie du Chien and the upper aquifers at the test site is small.

AQUIFER TEST BY INJECTION

Injection caused water-level changes similar to but opposite those caused by pumping. The injection of water caused water-level rises in the wells in the first part of the test, but after the third day of injection, the effects of atmospheric-pressure changes and of distant pumping obscured the water-level changes due to recharge. Consequently, only data from the first 3 days of the injection test were analyzed by well-hydraulics equations. Injection was continued, however, in order to complete the quality-of-water and chloride-tracer experiments. After injection was stopped, water levels declined rapidly to their preinjection levels. Figure 12 is a graph of water levels in the wells during the 20 days of injection and the 20 days after injection. The more pronounced fluctuations shown are the water-level changes due to injection, barometric fluctuations, and a weekly cycle of fluctuations due to distant pumping.

SPECIFIC CAPACITY AND WELL-LOSS TESTS

SPECIFIC CAPACITIES

The well loss, or loss due to turbulent flow in the vicinity of the pumped well, is 15.7 feet (4.79 m) compared with the aquifer loss of 22.1 feet (6.74 m) at the pumping rate of 458 gal/min (28.9 l/s) for 24 hours.

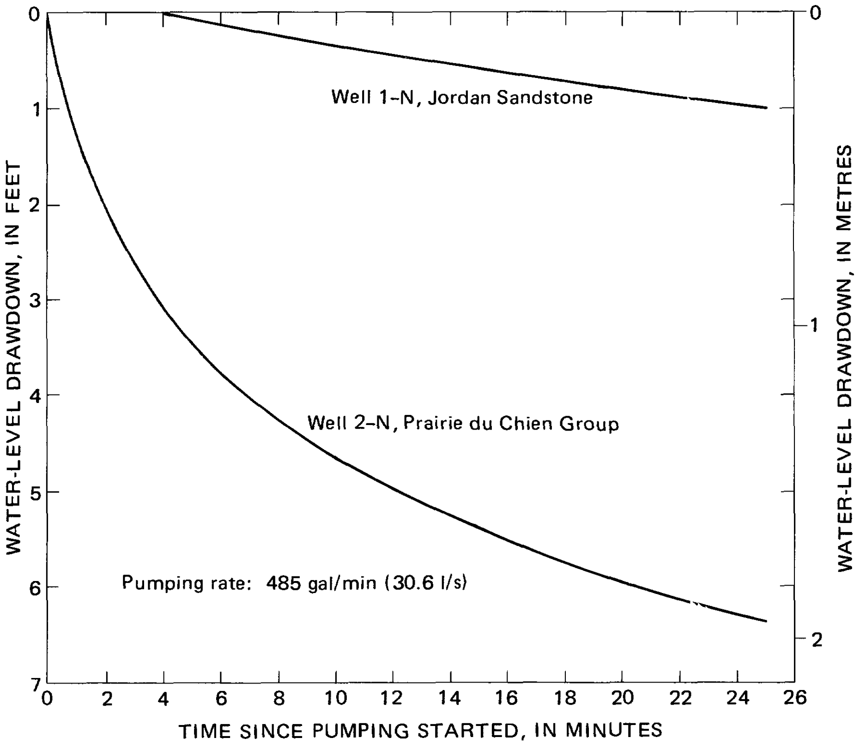

FIGURE 11.—Water-level changes in observation wells 1-N and 2-N at the beginning of pumping from well 1-S in the Prairie du Chien at the test site.

The indicated specific capacity calculated from the measured drawdown during the initial pumping test seems small—12.1 (gal/min)/ft, or 2.50 (l/s)/m, However, when correction is made for well loss, the specific capacity is 20.7 (gal/min)/ft, or 4.28 (l/s)/m.

The indicated specific capacity during the injection test was 17.2 (gal/min)/ft, or 3.56 (l/s)/m, computed from the measured drawdown. The specific capacity is 19.7 (gal/min)/ft, or 4.08 (l/s)/m, allowing for well loss. A correction for the difference in temperature of the injected water and the native ground water (Sniegocki, 1963, p. 11; Wenzel, 1942, p. 62) reduced the indicated specific capacities to 15.3 (gal/min)/ft, or 3.17 (l/s)/m, as calculated from the measured drawdown, and 17.5 (gal/min)/ft, or 3.62 (l/s)/m, corrected for well loss. Vecchioli and Ku (1972, p. A6–A7), however, stated: "In applying the temperature adjustments, * * * the true specific capacity probably lies somewhere between the observed value and the adjusted value."

After injection, the indicated specific capacity of the pumped well was 12.7 (gal/min)/ft, or 2.63 (l/s)/m, computed from the measured drawdown after pumping 10 hours. The turbulent-flow well loss was 17.6 feet

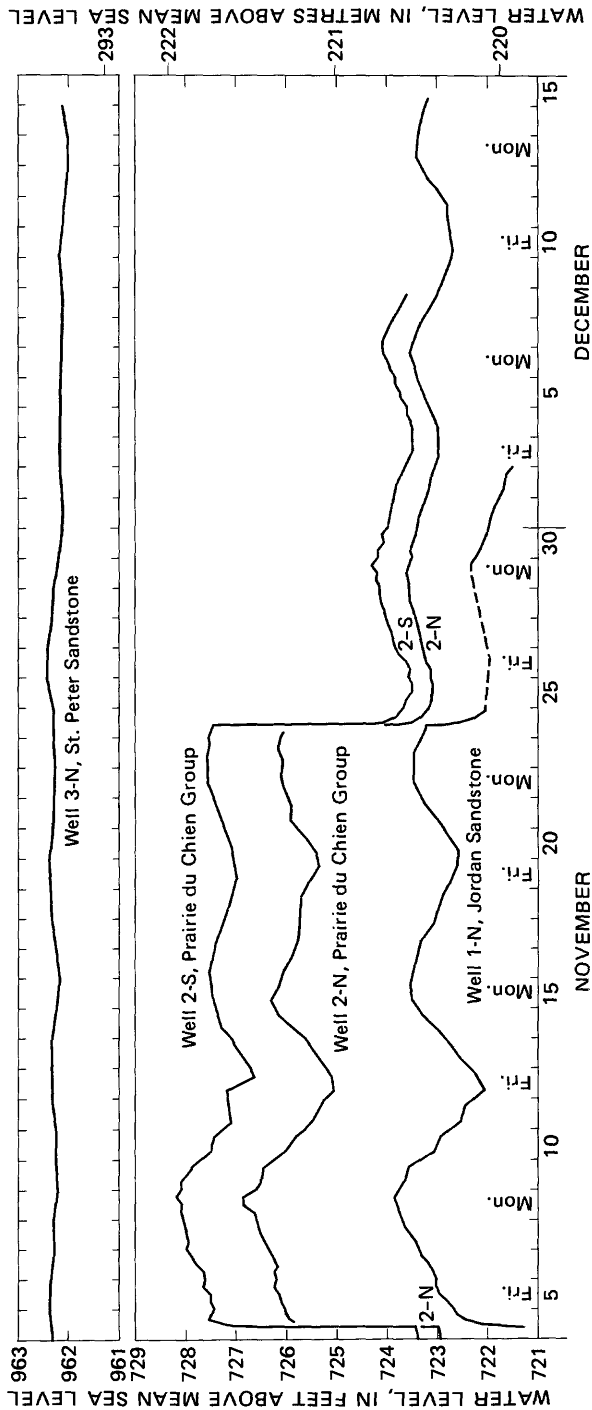

FIGURE 12.—Water levels in four observation wells during and after 20-day injection period.

(5.36 m), compared with the aquifer loss of 20.4 feet (6.22 m), at the pumping rate of 485 gal/min (30.6 l/s) for 10 hours. The specific capacity allowing for well loss gives a specific capacity of 23.8 (gal/min)/ft, or 4.93 (l/s)/m, after 10 hours of pumping. The specific capacity after 24 hours of pumping was less than it was after 10 hours. In order to obtain comparable values, the ratio of the reduction in specific capacity from 10 to 24 hours of pumping determined from the initial aquifer pumping test was applied to the 10-hour test. The 24-hour comparative specific capacities are 11.8 (gal/min)/ft, or 2.44 (l/s)/m, as indicated from the measured drawdown, and 21.0 (gal/min)/ft, or 4.35 (l/s)/m, which excludes well loss. These values are near the specific capacities obtained from the initial pumping test.

STEP-DRAWDOWN TESTS

Step-drawdown data are generally analyzed by plotting the ratio of drawdown in the pumped well to discharge (s_w/Q) at a specified time after pumping began against discharge. In this series of pumping and injection tests, the time interval was 90 minutes. Two step-drawdown tests were made, one before and one after the injection test, to determine whether plugging of materials near the well face occurred because of injection. The step-drawdown tests also provided a means to determine well loss in the pumped well and to separate well loss from aquifer-head loss. The term "well loss" as used here is the head loss, or water-level change, in the well due to turbulent flow in the vicinity of the pumped well. As the well was finished as an open hole, there is no entrance loss due to well-screen constrictions or other construction features. Some wells in the Twin Cities area that are finished in the dolomite of the Prairie du Chien Group and also in the Jordan Sandstone have been blasted to fracture the rock and enlarge the fissures and the effective well size in order to decrease well losses and increase yields. In this experiment, conditions were maintained as nearly natural as possible in order to determine aquifer characteristics with sufficient accuracy. Data from both step tests and, also, the comparative s_w/Q values after 90 minutes of pumping or injection in other tests are given in table 8.

Results of the step-drawdown test were computed by the Jacob (1944) equation, which separates the components of drawdown in a well as follows:

$$s_w = BQ + CQ^2,$$

in which s_w is the drawdown in the pumped well, Q is the discharge of the pumped well, B is the aquifer constant, and C is the well-loss constant. Subsequent work by Rorabaugh (1953) has stated that the exponent for turbulent flow (CQ term) is variable, and the equation is written:

$$s_w = BQ + CQ^n.$$

For the range of well discharge, Q, in this test, the value of n–2 gave satis-

TABLE 8.—*Values of* s_w/Q *after 90 minutes of pumping and injection for various rates*

Date of test	Pumping rate Q (gal/min)	Incremental drawdown in 90 minutes (ft)	Drawdown s_w at pumping rate Q, for 90 minutes (ft)	s_w/Q [ft/(gal/min)]
1971				
Jan. 22	250	16.68	16.68	0.0667
25	508	42.32	42.32	.0834
25	752	95.9	95.9	.127
26	556	51.5	51.5	.0925
Feb. 3	458	34.4	34.4	.0752
Nov. 4	108	5.49	5 49	.0509
1972				
Feb. 28	485	36.5	36.5	.0753
29	242	13.65	13.65	.0564
29	334	7.40	21.05	.0632
29	455	11.90	32.95	.0724
Mar. 1	395	26.2	26.2	.0664
2	400	26.9	26.9	.0673
3	285	15.9	15.9	.0559
3	365	6.8	22.7	.0622
6	355	21.6	21.6	.0609

factory results. Values for *B* and *C* were obtained by the graphical method described by Bruin and Hudson (1961, p. 29–35).

Data from the second step test, on February 29, 1972, are used in the following discussion. The incremental drawdowns of each pumping rate were determined graphically and were used in the computations and in the graphical solution of the equation. The values of s_w/Q are plotted against the pumping rates, *Q*, and shown in figure 13. The s_w/Q value at the intercept of the line with the zero pumping rate is the aquifer constant, *B*, and is 0.0383. (See fig. 13.) The slope of the line through the points is the well-loss constant, *C*, and is computed as 7.5×10^{-5}. The values of *B* and *C* were then used to prepare figure 14, which graphically shows the solution of the equation $s_w = BQ + CQ^2$ and the separation of aquifer loss and well loss. The well-loss values, as used in analysis of the pumping tests of the aquifer, were determined from figure 14. For example, the step test gave a drawdown for well losses at 500 gal/min (31.5 l/s) of 19.1 feet (5.82 m) and an aquifer drawdown of 18.8 feet (5.73 m).

A graphical analysis similar to that shown in figure 13 of the s_w/Q values in table 8, of all pumping and injection tests, shows a gradual development of the well with additional pumping. Also, the s_w/Q for the injection test, as compared with the values from pumping, indicates that no plugging of the well occurred during injection.

In addition to separating well and aquifer losses, the results of the step-drawdown test can be used to estimate a discharge or recharge rate for a given head change (drawdown or recovery) in the well. For example, assuming that the exponent n=2 for turbulent flow (CQ term) remains con-

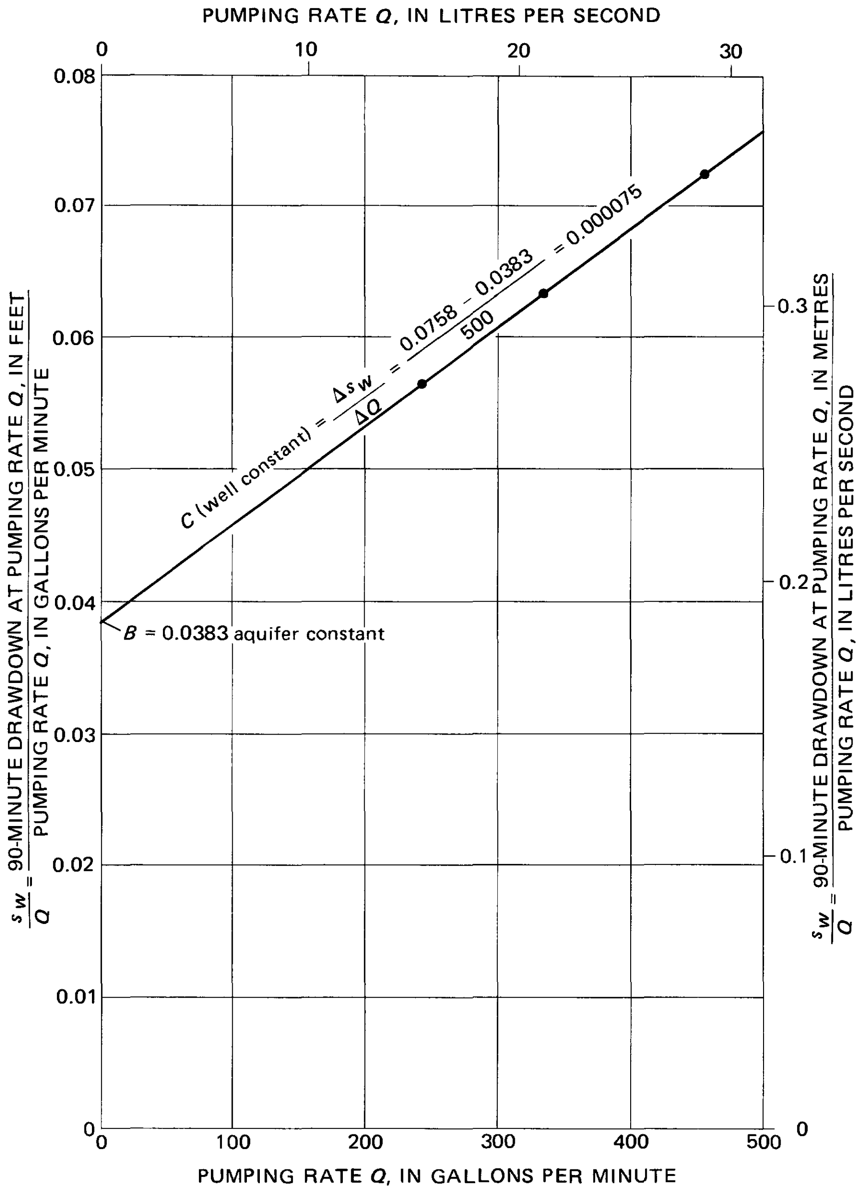

FIGURE 13.—s_w/Q against Q, to solve for the values of B and C of the step-drawdown test.

stant, water can be injected at the test site at a theoretical rate of 1,700 gal/min (107 l/s) with the available head buildup of 300 feet (91.4 m).

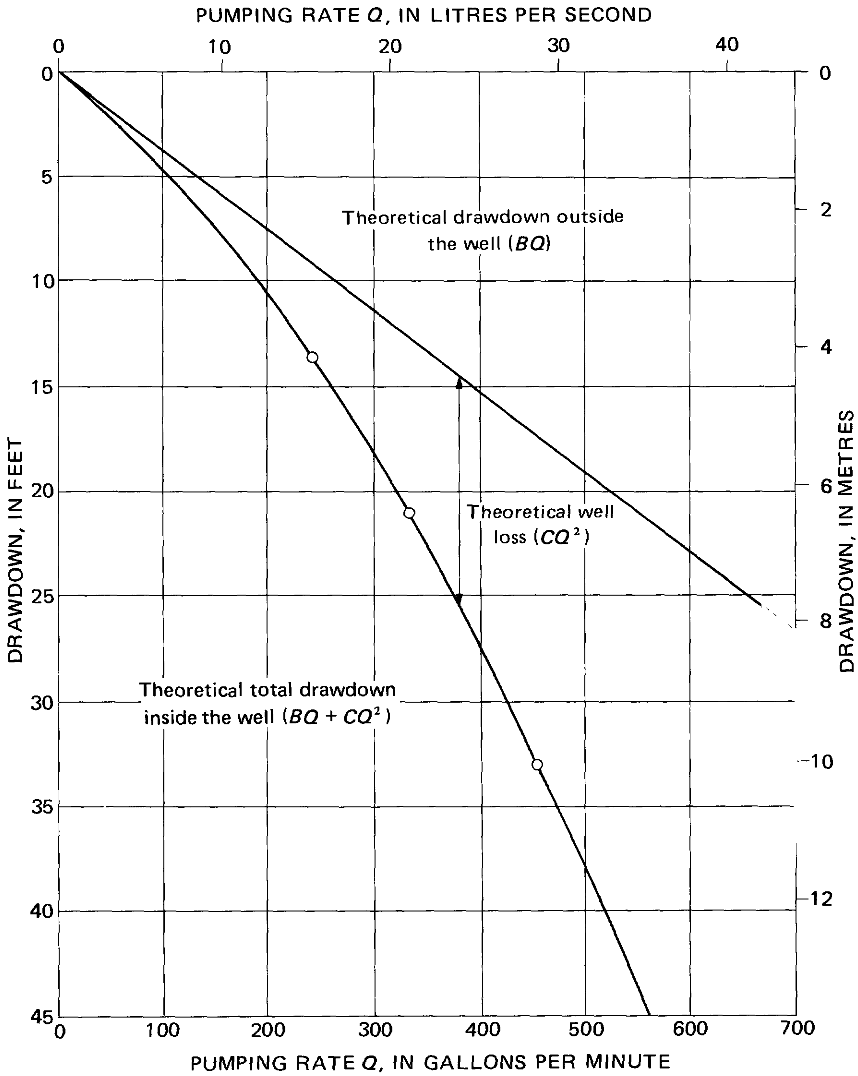

FIGURE 14.—Drawdown-yield graph for the injection well, 1-S, at the recharge test site.

AQUIFER TEST ANALYSES

METHODS OF ANALYSIS

Drawdown and recovery of water levels during and after pumping, and of head buildup and decline during and following injection, were analyzed to determine the transmissivity and storage coefficient of the aquifer, on the basis of analytical equations derived for ideal aquifers.

Because no analytical equation was available for analyzing the two-layer aquifer formed by the combined thickness of the dolomite of the Prairie du Chien Group and the Jordan Sandstone, two different idealizations were made of the aquifer system. For one such idealization the Prairie du Chien dolomite was considered to be a leaky aquifer, and it was assumed that leakage was transmissed from a constant-head source through a semiconfining bed that does not release water from storage. Theoretical drawdowns in such an aquifer due to withdrawals at a constant rate are described by Hantush and Jacob (1955). For the other set of analyses, the Prairie du Chien dolomite and Jordan Sandstone were assumed to consist of a single aquifer bounded above and below by impermeable beds. For these assumptions, the analyses were made using the modified nonequilibrium method described by Cooper and Jacob (1946) and the recovery method described by Theis (1935).

The methods of analyses are based on several assumptions concerning the aquifer being tested and the pumped well. The basic assumptions common to the several methods used in the analyses of data from the test site include: The aquifer is homogeneous, isotropic, and has infinite areal extent; the discharge or recharge well penetrates and receives water from the entire thickness of the aquifer; and the transmissivity is constant at all times and at all places (Ferris and others, 1962, p. 91–122). The hydrologic conditions at the test site do not completely meet the stated assumptions. As described in the section on hydrology, the aquifer tested forms a two-layered (nonhomogeneous) leaky artesian system, and the upper layer is anisotropic because of preferred orientation of fractures in the dolomite. The pumped well, 1–S, and two observation wells, 2–S and 2–N, are open only through the upper layer, the dolomite of the Prairie du Chien Group. One observation well is open only through the underlying Jordan Sandstone. As the effects of pumping move outward from the pumped well and include a larger part of the aquifer, the hydrologic variations are assumed to become less significant, and the aquifer responds as a homogeneous system.

As the dolomite and sandstone are interconnected to some degree, as is shown by changes in the water levels during pumping, the pumping-test data were then treated as applying to a single aquifer in which the wells are not fully penetrating. In this application, of course, the system fails to meet the analytical assumption that the aquifer is homogeneous and isotropic. For the analysis, however, this assumption must be included. In the single-aquifer analysis, a partial penetration correction was made using method 2 described by Weeks (1969, p. 207–208). The partial penetration effects are variable at the beginning of the pumping interval and become constant with length of pumping time. In the method by Weeks (1969), the deviations in drawdowns due to partial penetration of the wells were determined from the field data by graphical analysis. The deviations were used to determine the storage coefficient and the ratio of horizontal to vertical permeability. The analysis assumed that the partial penetration correction was constant and, thus, that transmissivity values determined by the modified nonequilibrium method would not be affected by partial penetration.

Transmissivity is given in this report in gallons per day per foot but can be converted to cubic feet per day per foot and cubic metres per day per metre by multiplying the given values by 0.134 and 0.0124, respectively. The data from the injection test were analyzed in the same ways as the pumping-test data in order to arrive at the transmissivity and the storage coefficient of the aquifer. Table 9 lists the transmissivity values from the pumping and the injection tests.

TRANSMISSIVITY DETERMINATIONS

AQUIFER-TEST ANALYSIS

The analysis of data from the two Prairie du Chien observation wells, 2–S and 2–N, during the initial pumping test gave a transmissivity of about 38,000 (gal/d)/ft [472 (m^3/d)/m], as determined by the leaky-aquifer type-curve solution (Hantush and Jacob, 1955). This method of analysis gave a transmissivity for these wells of 17,000 (gal/d) ft [211 (m^3/d)/m] for the injection test and 29,000 (gal/d)/ft [360 (m^3/d)/m] for the pumping test after injection. Data from the Jordan observation well (1–N) analyzed by the leaky-aquifer type-curve solution (Hantush and Jacob, 1955) gave a transmissivity of about 33,000 (gal/d)/ft [410 (m^3/d)/m] during the initial test, 30,000 (gal/d)/ft [373 (m^3/d)/m] for the injection test, and 28,000 (gal/d)/ft [347 (m^3/d)/m] for the later test. (See table 9 for a summary of transmissivity values.) No adequate explanation has been found to date for the low indicated transmissivity during injection. The leaky-aquifer type-curve solution seems to be the more reliable method of analysis because of the apparent leaky artesian conditions or possibly because of one or more recharging boundaries.

TABLE 9.—*Transmissivity from pumping and injection tests*

Pumped well: 1-S, in Prairie du Chien Group.
Observation wells: 2-S and 2-N, in Prairie du Chien Group; and 1-N, in Jordan Sandstone.

Wells	Transmissivity			Methods of analysis	Remarks
	Initial test Feb. 1971 [(gal d)/ft]	Injection test Nov. 1971 [(gal/d)/ft]	Test after injection Feb. 1972 [(gal d) ft]		
2-S, 2-N	38,000	17,000	29,000	(1)	
2-S, 2-N, 1-S	38,000	17,000	31,000	(2)	Early data.
2-S, 2-N	56,000	56,000	....	(2)	Later data.
1-S	55,000	44,000	45,000	(2)	Later data.
1-S	42,000	....	31,000	(3)	
1-S	44,600	42,500	45,400	(4)	
1-N	33,000	30,000	28,000	(1)	
1-N	38,000	....	....	(2)	

[1]Leaky-artesian type curve (Hantush and Jacob, 1955).
[2]Modified nonequilibrium, semilog (Cooper and Jacob, 1946).
[3]Simplified nonequilibrium recovery, semilog (Theis, 1935).
[4]Specific capacity (Theis and others, 1963).

A range of values is obtained from the data on wells 2–S and 2–N by the straight-line semilog solution (Cooper and Jacob, 1946). Water-level data throughout the periods of the test do not fall on a continuous straight line. Data from the early part of the test periods yield a transmissivity value of about 30,000 (gal/d)/ft [472 (m^3/d)/m]. The data from the middle part of the test show an apparent transmissivity of about 56,000 (gal/d)/ft [696 (m^{3}d)/m]. The later part of the data curve that seems to be the result of leakage also could be due to hydraulic effects of the underlying sandstone aquifer, which would become more significant with duration of pumping. In this case, the transmissivity of 38,000 (gal/d)/ft [472 (m^3/d)/m] would be representative of the Prairie du Chien aquifer as a unit by itself. The transmissivity of 56,000 (gal/d)/ft [696 (m^3/d)/m] may be representative of the Prairie du Chien–Jordan aquifer as a combined unit. The injection test data for wells 2–S and 2–N analyzed by this method gave a transmissivity of about 17,000 (gal/d)/ft [211 (m^3/d)/m] for the first part of the test and 56,000 (gal/d)/ft [696 (m^3/d)/m] for the later part of the test. This method gave a transmissivity of about 31,000 (gal/d)/ft [385 (m^3/d)/m] for the same two wells for the pumping test after injection. As this was a 10-hour test, no value was obtained that can be compared with the later value from the initial 48-hour pumping test and the injection test.

The recovery data of the pumped well, 1–S, for the initial pumping test, computed by the modified nonequilibrium method (Cooper and Jacob, 1946), yielded transmissivity values of 38,000 (gal/d)/ft [472 (m^3/d)/m] for the early part of the test and 55,000 (gal/d)/ft [683 (m^3/d)/m] for the middle part of the test. Data from the injection-test well, 1–S, did not fall on a straight line in the analysis by the modified nonequilibrium method. Three segments of the data, however, gave transmissivity values of 5,000, 16,000, and 44,000, (gal/d)/ft [62, 199, 546 (m^3/d)/m], respectively, from early pumping time to later pumping time. The continuous change in the data curve or increasing values of transmissivity suggest that initially the transmissivity is of the dolomite aquifer, and the effect of the sandstone aquifer becomes increasingly significant with duration of injection. The recovery data of the pumping-test well, 1–S, after injection yielded transmissivity values of 31,000 (gal/d)/ft [385 (m^3/d)/m] for the early part of the test and 45,000 (gal/d)/ft [558 (m^3/d)/m] for the last part of the test, as computed by the modified nonequilibrium method.

The simplified nonequilibrium recovery method (Theis, 1935) for well 1–S gave a transmissivity of 42,000 (gal/d)/ft [522 (m^3/d)/m] for the initial pumping test and 31,000 (gal/d)/ft [385 (m^3/d)/m] for the pumping test after the injection test. No comparable value was obtained from the injection test because of the long period (20 days) of injection and relatively short time (about 3 days) of usable drawdown data.

SPECIFIC-CAPACITY ANALYSIS

Transmissivity can also be estimated from the specific capacity of the pumped well (Theis and others, 1963). One of the requirements is that the well is 100-percent efficient. Using the specific capacity of 20.7 (gal/min)/ft [4.28 (l/s)/m], which excludes well loss, to simulate a 100-percent efficient well, the effective transmissivity is 44,600 (gal/d)/ft [554 (m^3/d)/m] during the initial pumping test. The specific capacity of 19.7 (gal/min)/ft [4.08 (l/s)/m], with well-loss correction, gives an estimated effective transmissivity of 42,500 (gal/min)/ft [528 (m^3/d)/m] during the injection test. The specific capacity of 17.5 (gal/min)/ft [3.62 (l/s)/m], which also includes temperature correction (Sniegocki, 1963), gives an estimated effective transmissivity of 37,800 (gal/d)/ft [469 (m^3/d)/m]. As stated earlier, the true specific capacity is probably between 19.7 and 17.5 (gal/min)/ft [4.08 and 3.62 (l/s)/m]; thus, the transmissivity is probably between 42,500 and 37,800 (gal/d)/ft [528 and 469 (m^3/d)/m] as estimated from the injection test. The specific capacities during the aquifer test after injection are near those determined from the initial aquifer test and indicate no change in the well or transmissivity of the formation.

STORAGE-COEFFICIENT DETERMINATIONS

The storage coefficient is the volume of water an aquifer releases from or takes into storage per unit surface area of the aquifer per unit change in head. In a confined water body the water derived from storage with decline in head comes from expansion of the water and compression of the aquifer; similarly, water added to storage with a rise in head is accommodated partly by compression of the water and partly by expansion of the aquifer (Lohman and others, 1973, p. 13). In this study the storage coefficient was computed or estimated from the pumping and injection tests, from the porosity of the aquifer, and from the barometric efficiency of the aquifer.

The storage coefficient obtained from the Prairie du Chien Group observation wells, 2-S and 2-N, is in the range of 2×10^{-4} to 2×10^{-5} and the Jordan well, 1-N, yielded a value of 4×10^{-4}. The storage coefficient 3×10^{-4} computed from the injection-test data was somewhat higher than that computed from the pumping tests.

The storage coefficient as a function of the elasticity of an artesian aquifer can be computed from the following formula (Ferris and others, 1962, p. 88):

$$S=\gamma_0\,\theta m(\beta+\frac{\alpha}{\theta}),$$

in which γ_0 is the specific weight of water, θ is the porosity of the aquifer, m is the aquifer thickness, β is the bulk modulus of compression of water (reciprocal of the bulk modulus of elasticity), and α is the bulk modulus of compression of the solid skeleton of the aquifer. The bulk modulus of elasticity of the aquifer is not known, but the porosity was determined from diffusion tests. (See later section in this report.) Thus, considering only that part resulting from the expansion of water in the aquifer, the partial storage coefficient value is about 1.4×10^{-5}, a useful value for determining a lower limit of the storage coefficient.

A slightly higher storage coefficient (related to the barometric efficiency of the aquifer) was obtained from the formula (Ferris and others, 1962, p. 90):

$$S = \gamma_0 \theta m \beta \frac{1}{BE},$$

in which, γ_0, θ, m, and β are as defined above, and BE is the barometric efficiency, which has been determined as 80 percent. This formula yields a storage coefficient of 1.8×10^{-5}. Although the storage coefficient varied widely, as determined for the data from the several wells and methods of analysis, it did not change significantly from the initial pumping test to the test after injection.

RELIABILITY OF AQUIFER-TEST RESULTS

The specific capacity of the pumped well before and after injection was nearly the same—20.7 and 21.0 (gal/min)/ft [4.28 and 4.?5 (l/s)/m] of drawdown, respectively. The transmissivity estimated from the specific capacity is about 44,600 and 45,400 (gal/d)/ft [554 and 564 (m^3/d)/m], respectively, for the two tests.

The transmissivity was 38,000 (gal/d)/ft [472 (m^3/d)/m] computed from the early part of the data from the initial pumping test and 31,000 (gal/d)/ft [385 (m^3/d)/m] computed from the test after injection. The apparent reduction in transmissivity probably is not significant because of the variables involved in the tests, as discussed earlier and based on the specific capacity of the well, which did not change significantly. Data from the later part of the test indicated an effective transmissivity of 56,000 (gal/d)/ft [696 (m^3/d)/m] for the Prairie du Chien–Jordan aquifer. The earliest part of the pumping-test data are more representative of the dolomite of the Prairie du Chien Group and the late part of the pumping test is representative of the Prairie du Chien Group--Jordan Sandstone aquifer combined.

The coefficient of storage determined from the pumping and injection tests seem to agree satisfactorily with values obtained on the basis of porosity and barometric efficiency of the aquifer. The storage coefficient is in the range of 2×10^{-4} to 2×10^{-5}.

The aquifer-test data were analyzed by methods based on homogeneous and isotropic aquifer conditions. As nonhomogeneous and anisotropic conditions prevail at the test site, this analysis yields only apparent values of the aquifer characteristics. Comparative values are obtained by this analysis for particular wells in repeated tests of the same pumped well, but the values derived from data of one observation well may not be comparable with those derived from data of another observation well used in the same test.

A graph of water-level changes against discharge or recharge rate of the three tests is shown in figure 15. The straight-line projections show theoretical water-level changes that are assumed to be due to aquifer resistance only and that might be expected to occur under pumping or recharging conditions. During the injection test, the changes in wells 1–S

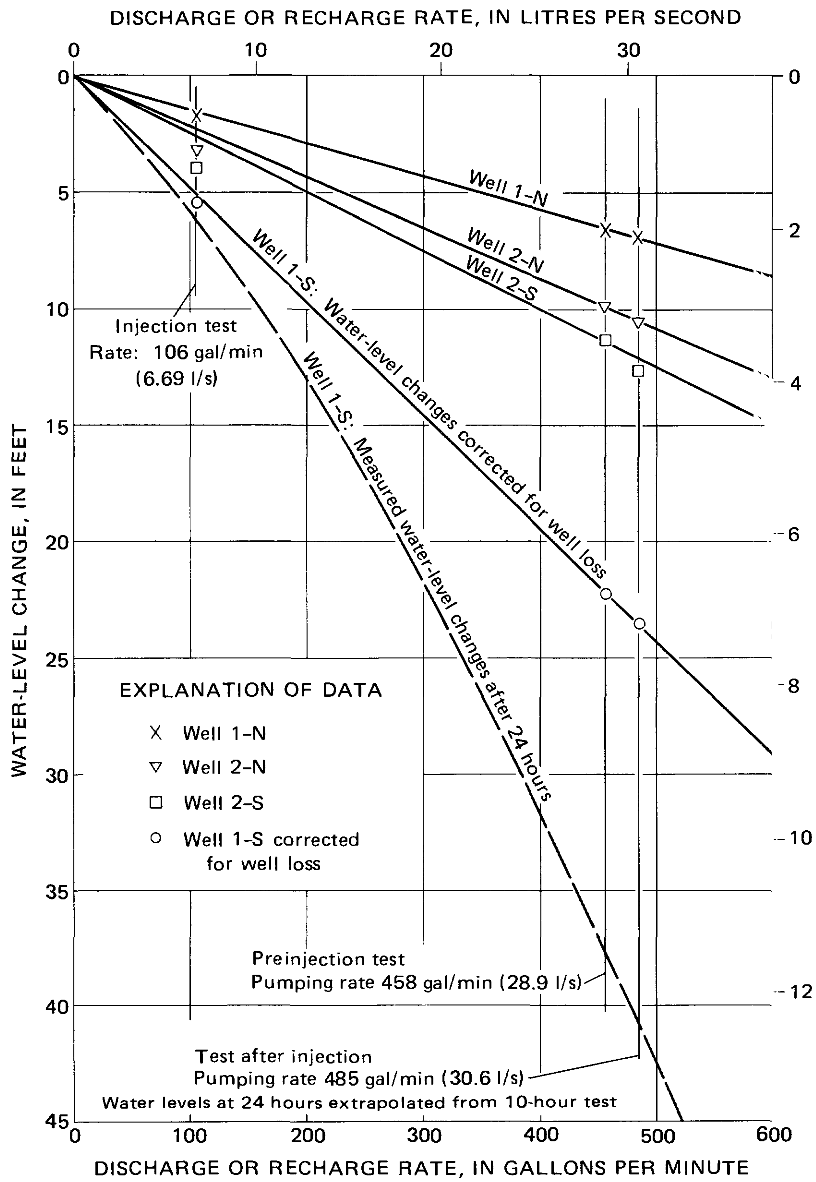

FIGURE 15.—Drawdown-yield graph for pumping and injection tests.

(pumped well) and 1-N (finished in the Jordan Sandstone), measured during injection, nearly fall on the theoretical change line, thus showing that these two wells correspond to the theoretical change values. Therefore, it is assumed that the available analytical methods of pumping test analyses may be applied with some confidence to data obtained from these wells. However, the injection-test data for observation wells 2-S and 2-N (finished in the Prairie du Chien) deviated 1.4 and 0.9 feet (0.43 and 0.27 m), respectively, from the theoretical change values shown, indicating that the available analytical methods may not be wholly applicable. Although no satisfactory explanation has been made, the deviation suggests that anisotropic and nonhomogeneous conditions prevail in at least the Prairie du Chien part of the aquifer.

Although considerable effort was made toward determining values for transmissivity and for storage coefficient, the primary concern should remain whether water injected into a well in the dolomite recharges the underlying sandstone in addition to recharging the dolomite. The water-level changes observed in the observation wells and in the pumped well show that head (water-level) changes are sufficiently large and rapid to cause transfer of water from one aquifer to the other. This is shown also by the results of the injection test.

GEOCHEMICAL ASPECTS OF ARTIFICIAL RECHARGE

By WARREN W. WOOD

INTRODUCTION

Water-quality problems associated with the operation of an injection well include aquifer and well-bore plugging, with the consequent loss of transmissivity or specific capacity. This plugging may be caused by one of several different factors or by a combination of factors. Accumulations of particulate material in the well, bacterial growth at the well-aquifer interface, and chemical reaction of the injected water with the aquifer or native water, forming a mineral precipitate, have been observed to cause a loss of transmissivity.

INORGANIC SUSPENDED MATERIAL IN THE INJECTED WATER

The treated city water used for injection contained an average of approximately 3 mg/l suspended sediment, a relatively low value. X-ray analysis of this suspended sediment indicates that it is amorphous, and no mineral identification could be made. The ranges in turbidity that were observed in the injected water are shown in figure 16. Turbidity values can be converted to milligrams per litre suspended solids in this water by multiplying the turbidity value (in Jackson turbidity units, JTU) by 4. The value of 4 was obtained by averaging several samples in which turbidity and suspended sediments were determined. On the basis of sediment load computed from this factor and the amount of injected water, it was calculated that approximately 36 kg (kilograms) of suspended sediment was injected into the well during recharge. Because of the small amount of suspended material in the recharge water and because the carbonate aquifer was cavernous, there was no apparent plugging by suspended sediment. Chemical reaction is a greater potential problem in this environment.

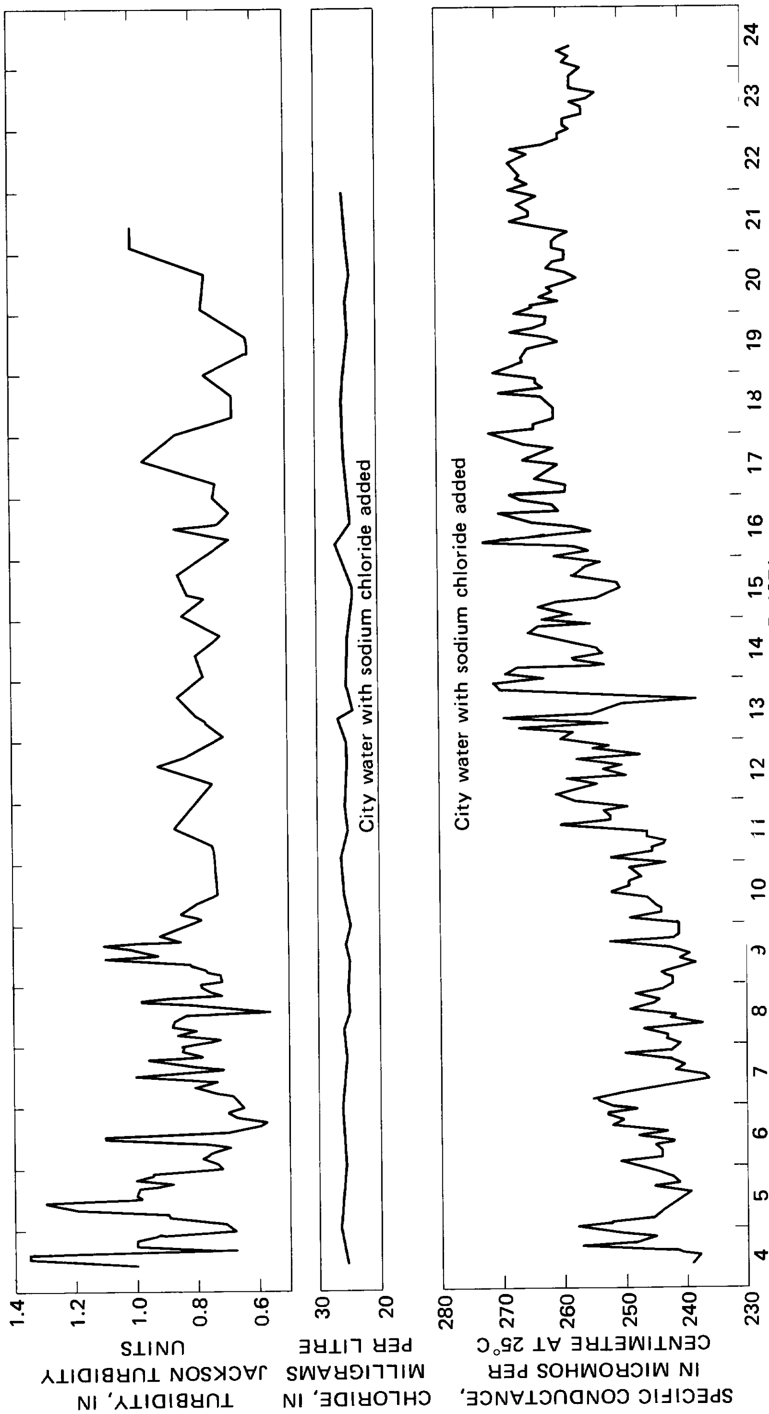

FIGURE 16.—Specific conductance, chloride, and turbidity of the injection water.

GEOCHEMICAL EVALUATION OF THE WEST ST. PAUL RECHARGE EXPERIMENT

The Prairie du Chien Group is composed of dolomite and limestone (table 2). The injected water is a calcium-magnesium bicarbonate type. Therefore, this analysis is confined primarily to evaluation of the stability of the common carbonate minerals in the native and injected water.

One of the major objectives of a geochemical evaluation of artificial recharge is to determine the chemical equilibrium of the water with respect to the aquifer—that is, will the water react with the aquifer to precipitate minerals, which would reduce transmissivity, or will the water dissolve minerals from the aquifer, which would increase transmissivity?

The goal of equilibrium calculations consists of comparing the ion-activity product (K_{iap}) calculated from the chemical analysis of the water with the known equilibrium value (K_{eq}), to determine if the water is in equilibrium with a given mineral. Before the K_{iap} can be calculated, the chemical analyses must be adjusted to activity values by the methods suggested in Garrels and Christ (1965).

From the chemical analyses (table 10) calculations for activity were made and adjusted for ionic strength, temperature, and ion complexing. A programmable desk calculator was used in making these adjustments and in determining the degree of saturation of the various waters with the different minerals (table 11). If the ratio K_{iap}/K_{eq} is equal to 1, then the water is in equilibrium with the mineral; if it is less than 1, the water is undersaturated and, if present, that mineral will dissolve; if the ratio is greater than 1, the water is supersaturated with respect to the given mineral, and mineral precipitation is a distinct possibility. However, because of analytical uncertainties and uncertainties as to the exact values of equilibrium constants (K_{eq}), the table should be viewed as only an approximation, and values that fall between 0.90 and 1.10 probably represent saturation conditions.

The data in table 11 show that the native water from both the Prairie du Chien and the Jordan is generally undersaturated with all the common minerals except calcite. The injected water is greatly undersaturated with all the minerals evaluated and can be expected to dissolve these minerals if they are present in the aquifer. However, the data from table 11 indicate that, after injection, the injected water is going toward calcite saturation but remains undersaturated with all the common minerals found in limestone.

In order to detect the presence of injected water and determine the amount of mixing with native water in the aquifer, a tracer was added to the injected water. Because of anticipated sampling difficulties, the tracer study was designed only to detect the arrival of the injected water at an observation well, not for more sophisticated analyses. Sampling was accomplished by using a ball-valve sampler. This grab-sampling technique causes some mixing, and it is difficult to sample from the same spot each time. Temperature difference between injected and native water, which ordinarily is an excellent tracer in artificial recharge (Schneider and others, 1971), could not be used because of the attempt to match the injected-water temperature with the native-water temperature. After consideration of cost, safety, ease of analysis, and effect on the hydrologic

TABLE 10.—*Chemical analyses of native and recharge water*

[Constituents in milligrams per litre unless otherwise stated]

Constituent	Well 2-N Prairie du Chien Group collected Oct. 6, 1971	Well 2-S Prairie du Chien Group collected Oct. 8, 1971	Well 1-N Jordan Sandstone collected Oct. 4, 1971	City water[1] (injection water) collected Nov. 9, 1971	Water from well 2-S during injection collected Nov. 10, 1971
Silica (SiO_2)	14	15	15	5.9	8.8
Iron (Fe) (μg/l)	550	550	460	40	600
Calcium (Ca)	62	63	61	23	41
Magnesium (Mg)	27	27	24	7.6	21
Sodium (Na)	6.0	5.9	3.9	8.4	8.2
Potassium (K)	1.7	1.7	2.1	2.8	2.2
Bicarbonate (field) (HCO_3)	346	342	317	73	195
Carbonate (CO_3)	0	0	0	0	0
Sulfate (SO_4)	5.8	4.3	7.8	26	17
Chloride (Cl)	1.1	1.2	0.7	19	13
Fluoride (F)	.4	.4	.4	1.2	.9
Nitrate (NO_3)	.01	.01	.01	.96	.36
Phosphorus (PO_4)	....	.07	....	.05	.05
Dissolved solids at 180°C	304	306	320	192	242
Hardness as $CaCO_3$	270	270	250	89	190
Specific conductance (field) (micromhos at 25°C)	508	499	470	236	371
pH (field	7.49	7.48	7.50	8.18	7.78
Temperature (°C)	10.0	10.0	10.3	10	10
Color	3	3	3	5	5
Boron (B) (μg/l)	40	50	40	60	50
Manganese (Mn) (μg/l)	210	120	90	10	120
Cobalt (Co) (μg/l)	0	0	0	....	....
Lead (Pb) (μg/l)	0	0	0	....	....
Cadmium (Cd) (μg/l)	0	0	0	....	....
Molybdenum (Mo) (μg/l)	2	2	8	....	....
Arsenic (As) (μg/l)	1	0	4	....	....
Chromium hexovalent (Cr) (μg/l)	0	0	0	....	....
Copper (Cu) (μg/l)	18	38	0	....	....
Lithium (Li) (μg/l)	10	10	10	....	....
Nickel (Ni) (μg/l)	2	0	4	....	....
Strontium (Sr) (μg/l)	190	260	300	....	....
Zinc (Zn) (μg/l)	30	30	20	....	....
Selenium (Se) (μg/l)	2	4	7	....	....
Eh (millivolts)	+120	+170	+185	....	....
Dissolved oxygen	0	....	0	....	....

[1]Analyses were performed before NaCl was added to the water.

TABLE 11.—*Values of K_{iap}/K_{eq} for common minerals found in limestone*

Mineral	Chemical formula	Injection water from city supply	Well 2-N Prairie du Chien Group	Well 2-S Prairie du Chien Group	Well 1-N Jordan Sandstone	Well 2-S after city water was injected
Calcite	$CaCO_3$	0.49	0.94	0.92	0.89	0.76
Dolomite	$CaMg(CO_3)_2$	.13	.67	.64	55	.50
Aragonite	$CaCO_3$	.35	.67	.65	.64	.54
Siderite	$FeCO_3$	.0077	.081	.078	.066	.021
Magnesite	$MgCO_3$	.14	.38	.36	.32	.34
Strontianite	$SrCO_3$	....	.032	.042	.048	....
Smithsonite	$ZnCO_3$	....	.0075	.0072	.0048	....
Rhodochrosite	$MnCO_3$	.011	.17	.094	.070	.080
Fluorite	CaF_2	.0018	.00046	.00047	.00046	.0017
Gypsum	$CaSO_4 \cdot 2H_2O$	.0048	.0014	.0011	.0020	.0036
Anhydrite	$CaSO_4$	.0023	.00069	.00052	.00099	.0017

system, sodium chloride was chosen as the tracer. Chloride is harmless, easily detected, already present in the system, and a mobile ion that does not enter into chemical reaction in this ground-water environment. Sodium chloride was added to the injected water, by means of a constant-rate injection pump, to increase the chloride concentration of the injected water from 19 to 25 mg/l. Because the native water is low in chloride, 1.1 mg/l (table 10), the contrast is large, and, therefore, makes a good tracer of the injected water in this situation.

The injected water, indicated by the chloride concentration, arrived at the lower zone in well 2–S, 99 feet (30.2 m) away, in less than 1 day. The chloride concentration rose from a background value of 1.1 mg/l to a high of 16 mg/l during the period of the test and decreased rapidly to 3.5 mg/l after the injection was stopped (fig. 24). There was no increase in chloride concentration in the upper zone in well 2–S or in any zone in well 2–N, 495 feet (150.9 m) away.

The large variations in chloride concentration in the different wells and within well 2–S indicate flow in fractures and solution openings. Photographic evidence (figs. 1, 2), supports the cavernous anisotropic nature of the material. Flow in fractures and solution openings results in mixing of injected water with the native water rather than piston volume-for-volume displacement of native water within the aquifer.

If the injection water and native water (table 10) are normalized on the basis of chloride concentration, the result of simple mixing can be seen. Chloride and sodium will be 25 mg/l and 12.3 mg/l, instead of 19 mg/l and 8.4 mg/l, respectively, because of the amount added by chemical injection. Solving two simultaneous equations,

$$1.1x+25y=13,$$

and

$$x+y=100 \text{ percent},$$

where 1.1 mg/l is the background chloride concentration, 25 mg/l is the chloride concentration of the injected water and 13 mg/l is chloride concentration in the sample collected from well 2-S (table 10). The total chloride (100 percent) must come only from two sources—the native water, x, and the injected water, y. Solution of the above equation yields a value of 49.8 percent for y—that is, approximately 50 percent of the water from this zone is injection water. By using this mixing ratio, the concentrations of the other major ions were calculated and are given in table 12. If reaction of the water with the aquifer had occurred, the observed value of the constituent would be considerably different than the calculated value. Table 12 indicates mixing of the two types of water with little chemical reaction. The arrival of the injected water at the observation well (fig. 24) may have been so rapid that there was insufficient time for chemical reactions to occur. However, the lack of reaction probably results from the formation of weathered-like reaction rims on the minerals that prevent contact of the water with the fresh mineral faces along the permeable fractures in the aquifer.

Because of uncertainties in the sampling procedure and normal analytical variations, iron apparently is the only element that underwent a significant chemical reaction at this time and distance from the injection well. The stability of the common iron minerals can be shown by means of an Eh-pH diagram (Garrels and Christ, 1965). From figure 17, which takes sulfur and carbonate species into consideration, it is possible to determine which mineral controls the water-mineral reaction and if the water is in equilibrium with the environment. Eh was determined in the field by a flow cell similar to that used by Clark and Barnes (1969). The diagram shows that $Fe(OH)_3$ is the controlling compound of the water naturally present in the formation. The dashed lines in figure 17 are lines of equal iron concentration that are in equilibrium with the pH and Eh at any given point. For example, the water in well 2-S should have a concentration of 10^{-7} M iron [5.5 μg/l (micrograms per litre)] at the observed pH and Eh. However, it contains 600 μg/l, or about 100 times greater than

TABLE 12—*Comparison of observed water in well 2-S and calculated value normalized on chloride*

[All values in milligrams per litre unless otherwise indicated]

Constituent	Observed concentration	Calculated concentration	Constituent	Observed concentration	Calculated concentration
Silica	8.8	10.5	Sulfate	17	15
Iron (μ g/l)	600	300	Chloride	13	13
Calcium	41	43	Fluoride	.9	.8
Magnesium	21	17	Nitrate	.36	.48
Sodium	8.2	9.1	Phosphorus	.05	.06
Potassium	2.2	2.2	Dissolved solids at 180°C	242	249
Bicarbonate	195	210			

equilibrium values. This supersaturation is also observed in wells 2-N and 1-N. Supersaturation of water with a mineral is not uncommon and may be controlled by the lack of critical activation energy with which to initiate the precipitation or some other kinetically controlled process.

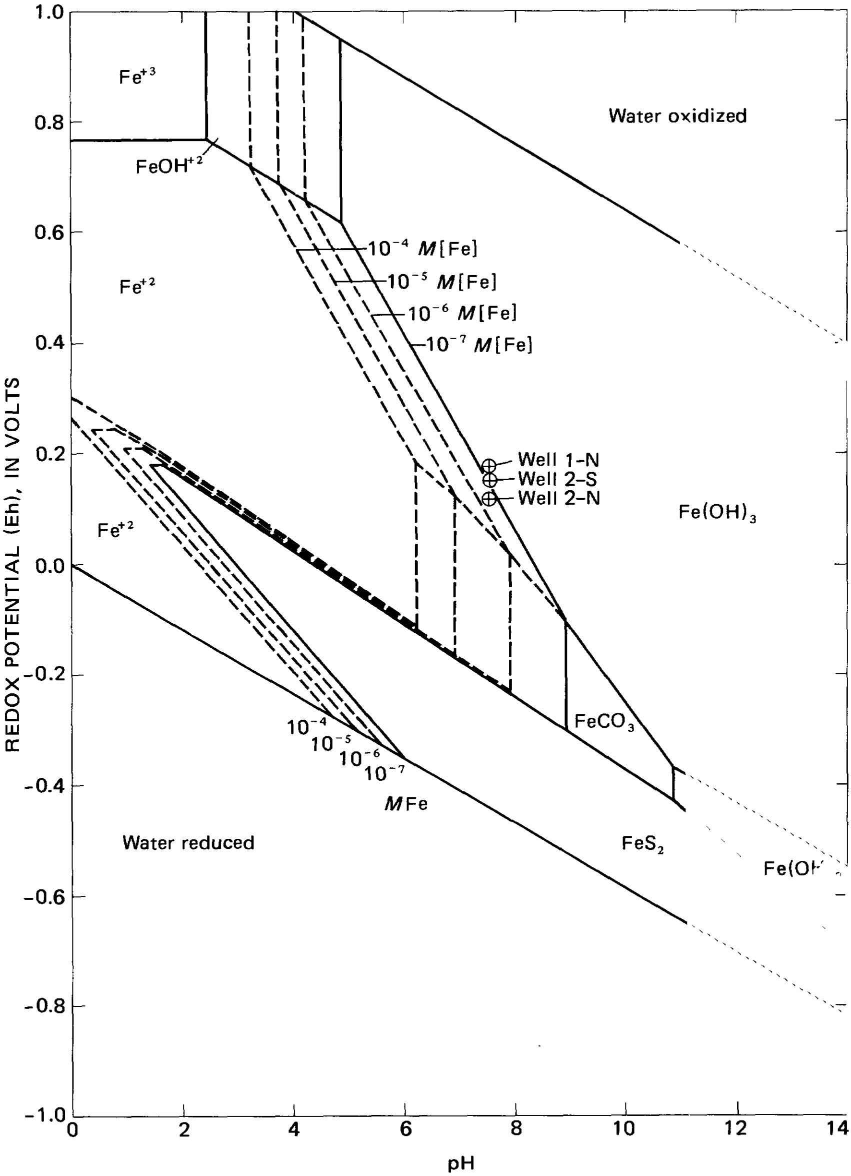

FIGURE 17.—Solubility of iron in relation to pH and Eh at 10°C and 1 atmosphere.

Iron is present in greater concentrations in water from well 2-S than is calculated by a simple mixing model. Dissolved oxygen in the injected water probably coupled with pyrite in the formation to form ferric hydroxide by the following reaction:

$$2FeS_2+3H_2O+\frac{19}{2}O_2=2Fe(OH)_3+4SO_4^{-2}.$$

The addition of oxygenated recharge water would raise the Eh, pushing the equilibrium further into the $Fe(OH)_3$ field. At the same time more iron would be dissolved from the pyrite, and, ultimately, $Fe(OH)_3$ would precipitate. The longer the delay in precipitation of $Fe(OH)_3$, the greater the expected life of a recharge well because the water containing the iron will be spread over a larger area; therefore, the same amount of iron will precipitate over a greater area.

From the above formula it can be shown that for every 19/2 molecules of O_2 added, 2 molecules of $Fe(OH)_3$ will form. This is probably the major oxygen-consuming reaction in this system, as very little organic material is present.

GEOCHEMICAL CONSIDERATIONS FOR RECHARGE IN THE TWIN CITIES

Water that is free of sediment and bacteria and that is of the same general chemical type as used in this study could be injected into the fractured limestone and dolomite with few apparent geochemical problems. The injected water should be undersaturated with the major carbonate minerals and should contain as little dissolved oxygen as possible. The temperature of the injected water is important in considering the degree of saturation. The equilibrium constant (K_{eq}) is sensitive to temperature as a power function, whereas the chemical constituents are linearly additive. Therefore, the temperature of the injected water should be chosen carefully. Use of cooler water is preferable, so as not to cause precipitation of minerals in native water. However, this cooler water could increase the dissolved-oxygen concentration, thereby increasing the potential for pyrite solution. Consideration might be given to adding iron-stabilizing compounds to the water prior to injection, if iron precipitation is a potential plugging problem.

Because the rock is fractured and the minerals form reaction rims along the fractures, chemical reactions of all types would take place over a large areal distance; consequently, a relatively long well life can be expected before plugging occurs and the well needs to be abandoned or rehabilitated.

MICROBIOLOGICAL ASPECTS OF ARTIFICIAL RECHARGE OF FISSURED-ROCK TERRANE

By Garry G. Ehrlich

Few detailed studies of the occurrence and activities of biota in ground water have been made, but it is possible to give a few generalizations.

Absence of oxygen, as is typical of most deep ground water, precludes the existence of air-breathing animals, and lack of sunlight excludes plants from ground water. Although many species of bacteria flourish under anoxic conditions, the organic content of most ground water is insufficient to sustain a bacterial population. Undisturbed ground water is therefore usually considered to be sterile. In a few places, however, bacteria have been isolated from spring water, well water, and lithologic cores. Their existence in certain types of organic-rich ground water, such as oil-field brines, is well documented. If conditions for bacterial growth are satisfactory, the presence of these organisms is to be expected. This could occur, for example, in the vicinity of an artificial-recharge well. Therefore, a microbiological-surveillance program was made as a part of the West St. Paul recharge study.

Two major areas of concern can be defined relative to the presence of microorganisms in and near artificial-recharge wells. The first of these involves the possible introduction into the aquifer of pathogenic bacteria and viruses, which would render the water unsafe for domestic consumption. The results of several studies on the movement of bacteria in soils and in nonsaline ground water showed that the travel of bacteria to fine- to medium-grained unconsolidated materials was generally restricted to distances less than 150 feet (45 m) from the point of injection (Romero, 1970). Bacterial survival times under these conditions may be 100 days or longer. Under most conditions the restricted travel distances result more from filtering action of the porous formation material than from organism die-off. These results suggest that little or no danger exists when wastes are injected into formations of low to moderate permeability because of the limited travel distance from the point of injection.

If bacterially contaminated water is injected into fissured carbonate rocks, the bacteria may travel for long distances. The travel distance may be limited only by the life spans of the bacteria in the formation. Unfortunately, there is very little data concerning the longevity of bacteria in fissured carbonate rocks. Romero (1970) reported that some species of bacteria have been observed to survive in the ground for as much as 5 years under favorable temperature and nutrient conditions. However, the results of most studies indicate that maximum life spans generally range from 60 to 100 days. Thus, the injection of heavily contaminated water into fractured or otherwise highly porous formations seems inadvisable.

The second area of concern involves the effects of biological growths upon the operation of the recharge well. Hydraulic conductivity of the aquifer may be reduced by growth of organisms in the formation, leading to clogging of recharge wells. Effects of biological clogging of sand and sandstone aquifers near recharge wells have been reported by the California State Water Pollution Control Board, (1954), Rebhun and Schwarz (1968), and Ehrlich, Ehlke, and Vecchioli (1972). Biological clogging of wells recharging highly permeable formations is uncommon. Rebhun and Schwarz (1968), for example, found that wells recharging lake water into sandstone gradually clogged, whereas the same type of lake water could be injected into limestone for equally long time periods with no significant effects on specific capacity of the wells during either recharge or discharge.

Samples of ground water from the Jordan Sandstone and Prairie du Chien Group were collected before and during the injection test. Samples of injection water were collected daily during injection. Water samples were collected during the pump-out test, and a single water sample was taken 10 days after the completion of the pump-out test. Total aerobic bacterial counts were done essentially according to the membrane-filter method described in Standard Methods (American Public Health Association, 1971, p. 679–683). Exposed membrane filters were incubated on Plate Count Agar at room temperature for 24 to 72 hours before the bacteria count was made. After incubation, the filters with developed colonies were placed for 15 minutes on an absorbent pad saturated with 1 percent methylene blue-dye solution before the count. Bacterial colonies were stained to a darker color than the background. Results of these tests are given in tables 13 and 14.

TABLE 13.—*Bacteria count in water from the Prairie du Chien and Jordan wells at the West St. Paul recharge site*

Date	Well	Geologic unit	Total bacteria count (colonies/100 ml of sample)
		Preinjection sampling	
10- 4-71	1-N	Jordan Sandstone	$>10^4$
10- 5-71	2-N	Prairie du Chien Group	$>10^4$
10- 7-71	2-S	......do.	$>10^4$
		During injection test	
11-18-71	2-S	Prairie du Chien Group(top)	$>10^4$
Do	2-S	.....do. (middle)	$>10^4$
Do	2-S	.....do. (bottom)	$>10^4$
Do	2-N	.....do. (top)	$>10^4$
Do	2-N	.....do. (middle)	$>10^4$
Do	2-N	.....do. (bottom)	$>10^4$
Do	1-N	Jordan Sandstone (top)	$>10^4$
Do	1-N	.....do. (middle)	$>10^4$
Do	1-N	.....do. (bottom)	$>10^4$
		During pump-out test	
2-25-72	1-S	Prairie du Chien Group	$>10^4$
2-28-72	1-S	.....do.	$>10^4$
2-29-72	1-S	.....do.	$>10^4$
3- 1-72	1-S	.....do.	$>10^4$
3- 3-72	1-S	.....do.	$>10^4$
3- 6-72	1-S	.....do.	$>10^4$
3- 7-72	1-S	.....do.	$>10^4$
3- 8-72	1-S	.....do.	$>10^4$
		Post-pump-out sampling	
3-24-72	1-S	Prairie du Chien Group	$>10^4$

The test results were inconclusive. In general, the biological phenomena seemed to have no impact upon the recharge of water to the Prairie du Chien aquifer during this test.

TABLE 14.—*Total bacteria count and residual chlorine content of injected water*
[n.d., not determined]

Sample collection date	Total bacteria count (colonies 100 ml of sample)	Residual chlorine (mg/l)	Sample collection date	Total bacteria count (colonies 100 ml of sample)	Residual chlorine (mg/l)
11- 4-71	$>10^3$	n.d.	11-14-71	250	<.01
11- 5-71	$>10^3$	< 0.01	11-15-71	100	<.01
11- 6-71	$>10^3$	.10	11-16-71	10	.06
11- 7-71	$>10^3$	<.01	11-17-71	n.d.	.05
11- 8-71	$>10^3$	<.01	11-18-71	40	.04
11- 9-71	$>10^3$	<.01	11-19-71	300	.07
11-10-71	$>10^3$	.08	11-20-71	400	.10
11-11-71	$>10^3$	.08	11-21-71	$>10^3$	.04
11-12-71	600	.02	11-22-71	$>10^3$	.02
11-13-71	450	.16	11-23-71	$>10^3$	<.01

HYDRODYNAMIC DISPERSION AND MOVEMENT OF INJECTED WATER

By REN JEN SUN

INTRODUCTION

Quality-control problems arise when the recharged waters are different from the native water or when wastes are injected into an aquifer for storage. Forecasting the degree of mixing and the movement of the injected water are major tasks for recharge and subsurface waste-storage operations. The aim is to control the movement of the injected fluid in order to assure adequate mixing ratios that will dilute the injected wastes below harmful concentrations or, if the recharged waters are of a quality not acceptable for use, to assure adequate dilution by mixing with fresh water in the aquifer before it is pumped out. In this study the quality of the injected water differed only slightly from that in the formation; however, the data obtained are useful for testing the applicability of some analytical methods for forecasting mixing and movement.

Theories of hydrodynamic dispersion and flow through isotropic homogeneous porous media of infinite extent have been well developed. On the basis of these theories, a discussion is given below of methods for predicting the degree of mixing and the movement of the injected water, as well as the use of a single-well tracer-dilution method to estimate the natural flow velocity in the aquifer.

MOVEMENT OF INJECTED WATER

A well penetrating a theoretically static aquifer, injecting or pumping at some rate, Q, produces flow so that streamlines representing the flow pattern radially diverge from or converge toward the well, respectively.

But, if natural flow exists in the aquifer, the streamlines are affected by the direction of the natural flow. During recharge of an aquifer, the injected water spreads at a higher rate in the downstream direction and at a slower rate in the upstream direction. During pumping, conditions are reversed.

On the basis of potential theory, the shape of the injected volume at time t after the start of injection can be defined mathematically. For simplicity and without using numerical integration, only steady-state and two-dimensional flow are considered. Also, the differences in density and viscosity between the displacing and displaced fluids are assumed to be small and may be neglected. Following is the mathematical derivation of the equation defining the shape of the injected water body at time t.

Consider a well which fully penetrates a confined homogeneous isotropic infinite aquifer and recharges at a constant rate, Q_i. The aquifer has a uniform thickness, b, and a constant hydraulic conductivity, K, and uniform natural flow takes place at an average flow rate, v_0, in the positive direction along the x axis (fig. 18).

The complex potential $f(z)$ for the combined natural flow and the radial flow in the (x, y) plane resulting from the injection (Jacob, 1950; Bear and Jacobs, 1965) is

$$f(z) = \varphi + i\psi = -v_0 z - (Q_i/2\pi b) \ln(z);\ z = x + iy, \tag{1}$$

where φ is the velocity potential and ψ the stream function. Therefore,

$$\varphi = -v_0 x - (Q_i/4\pi b) \ln[(x^2+y^2)/r_w^2], \text{ and} \tag{2}$$

$$\psi = -v_0 y - (Q_i/2\pi b) \arctan(y/x), \tag{3}$$

where r_w is the radius of the well.

Average seepage velocity is defined (Ferris, 1958; De Wiest, 1965; and Bear and others, 1968) as

$$v_a = v/n, \tag{4}$$

where v_a denotes average seepage velocity; v, Darcian velocity; and n, the effective porosity of the aquifer. By Cauchy-Riemann equations

$$v_x = -\partial\varphi/\partial x = -\partial\psi/\partial y, \tag{5}$$

and

$$v_y = -\partial\varphi/\partial y = \partial\psi/\partial x, \tag{6}$$

where v_x and v_y are Darcian velocity components along the x axis and y axis, respectively, the (x, y) components of the average seepage velocity can be written in the following expressions:

$$v_{ax} = dx/dt = -(1/n)(\partial\varphi/\partial x) = (v_0/n) + Q_i x/[2\pi nb(x^2+y^2)], \tag{7}$$

$$v_{ay} = dy/dt = -(1/n)(\partial\varphi/\partial y) = Q_i y/[2\pi nb(x^2+y^2)]. \tag{8}$$

From equation 3, along a given streamline with ψ being constant,

$$x = y \cot[(-2\pi b/Q_i)(\psi + v_0 y)]. \tag{9}$$

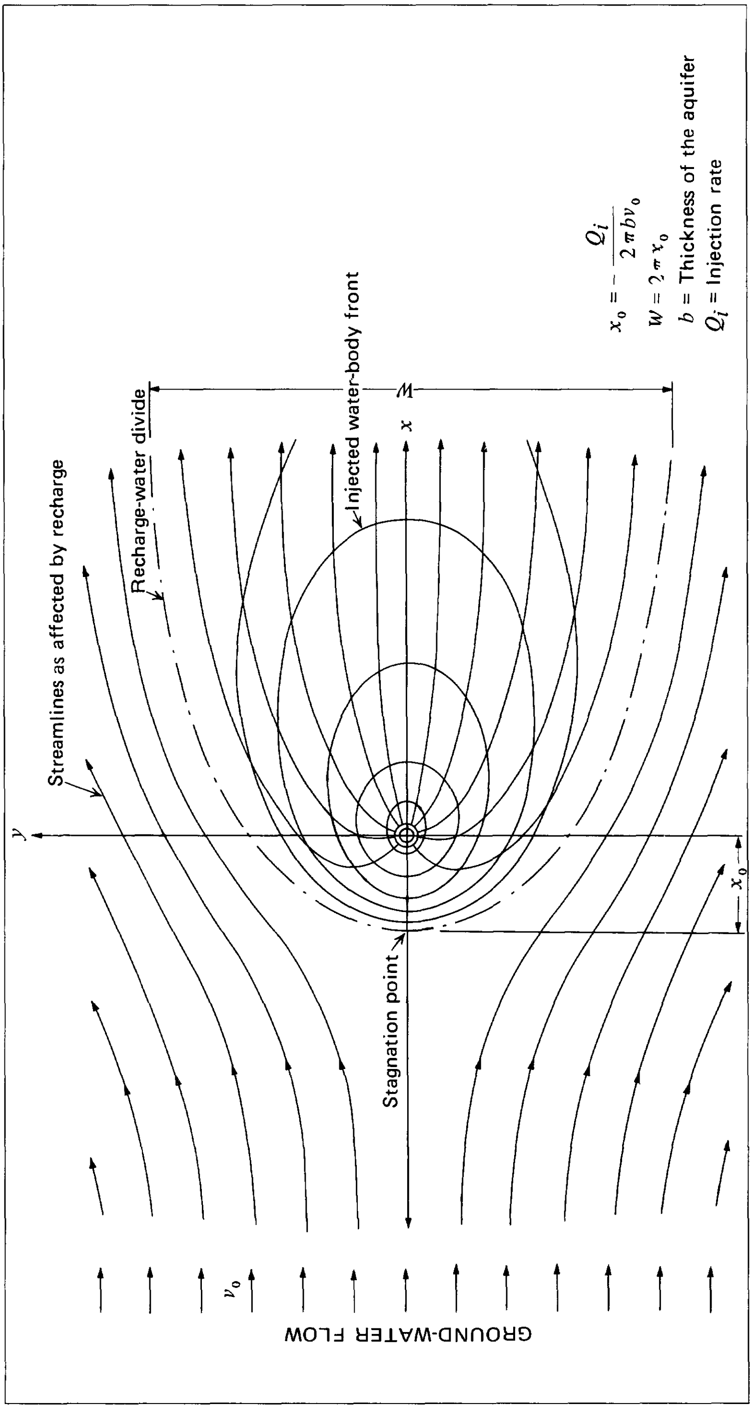

FIGURE 18.—Streamlines and position of injected water body.

Substituting equation 9 into equation 8, then

$$dt = (2\pi nby/Q_i)\csc^2[(2\pi b/Q_i)(\psi + v_0 y)]dy. \tag{10}$$

Integrating equation 10,

$$t = (-ny/v_0)\cot[(2\pi b/Q_i)(\psi+v_0y)] + (nQ_i/2\pi bv_0^2)\ln\{\sin[(2\pi b/Q_i)(\psi+v_0y)]\}+C, \tag{11}$$

where C is an integration constant.

Let water particles leave the origin x=0, y=0 at t=0. Then

$$C = (-nQ_i/2\pi bv_0^2)\ln[\sin(2\pi b\psi/Q_i)] \tag{12}$$

and

$$t = (-ny/v_0)\cot[(2\pi b/Q_i)(\psi+v_0y)] + (nQ_i/2\pi bv_0^2)\ln[\sin(2\pi b/Q_i)(\psi+v_0y)/\sin(2\pi b\psi/Q_i)]. \tag{13}$$

Let

$$\theta = \arctan(y/x), \tag{14}$$

and from equation 3,

$$\psi = -v_0y - (Q_i/2\pi b)\theta, \tag{15}$$

substituting equation 15 into equation 13,

$$\begin{aligned} t &= (-ny/v_0)\cot\{(2\pi b/Q_i)[-v_0y - (Q_i/2\pi b)\theta + v_0y]\} \\ &\quad + (nQ_i/2\pi bv_0^2)\ln\{\sin(2\pi b/Q_i)[-v_0y-(Q_i/2\pi b)\theta + v_0y] \\ &\quad /[\sin(2\pi b/Q_i)[-v_0y-(Q_i/2\pi b)\theta]]\} \\ &= (ny/v_0)\cot\theta + (nQ_i/2\pi bv_0^2)\ln\{\sin\theta/\sin[(2\pi bv_0y/Q_i)+\theta]\} \\ &= (nx/v_0) + (nQ_i/2\pi bv_0^2)\ln\{\sin\theta/\sin[(2\pi bv_0y/Q_i)+\theta]\}, \end{aligned} \tag{16}$$

which may be written in terms of dimensionless parameters,

$$\begin{aligned} \bar{t} &= \bar{x} - \ln[\sin(\bar{y}+\theta)/\sin\theta] \\ &= \bar{x} - \ln(\cot\theta\sin\bar{y} + \cos\bar{y}), \end{aligned}$$

or

$$\exp(\bar{x} - \bar{t}) = \cos\bar{y} + (\bar{x}/\bar{y})\sin\bar{y}, \tag{17}$$

where

$$\bar{x} = (2\pi bv_0/Q_i)x = (2\pi TI/Q_i)x;\ v_0 = KI, \text{ and } T = Kb; \tag{18}$$

$$\bar{y} = (2\pi bv_0/Q_i)y = (2\pi TI/Q_i)y; \text{ and} \tag{19}$$

$$\bar{t} = (2\pi bv_0^2/nQ_i)t = [2\pi(TI)^2/nbQ_i]t. \tag{20}$$

Equation 17 describes the shape of the injected water front at time t, which is symmetric with respect to the x axis (fig. 18).

If the well recharges indefinitely, $t \rightarrow \infty$, equation 17 will reduce to

$$\bar{x} = -\bar{y}\cot\bar{y}. \tag{21}$$

This equation represents a theorectical water divide. Mathematically, injected water will never penetrate this water divide, or no water will cross this divide to move toward a pumping well, even after infinite pumping time.

With a recharge well, a point of stagnation, at which Darcian velocity is zero, theoretically will be located upstream from the recharge well along the x axis; with a pumping well, this point theoretically will be located along the x axis, downstream from the pumping well. The position of a stagnation point can be obtained through equations 2 and 5. The distance of the stagnation point, x_0, along the x axis from the recharge well is found by the following expressions:

$$v_x = -\partial\varphi / \partial x = -v_0 - (Q_i x) / [2\pi b(x^2 + y^2)] = 0, \text{ with } y = 0;$$

whence

$$x = x_0 = -Q_i/(2\pi b v_0). \tag{22}$$

Equation 22 also applies to pumping when Q_i is replaced by $(-Q_p)$, the pumping rate.

Theoretically, observation wells constructed for a recharge or pumping study should be located within the flow area bounded by the water divide and the point of stagnation, which can be calculated by equations 21 and 22.

In a homogeneous and isotropic porous medium after injection ceases, the injected water will move downgradient with the natural flow (Bear and Jacobs, 1965). The extent of the injected water body at a time, t_d, after the end of injection, is given by

$$\exp(\bar{x} - \bar{t}) = \cos \bar{y} + [(\bar{x} - \bar{t}_d)/\bar{y}]\sin \bar{y}, \tag{23}$$

where $\bar{t} = \bar{t}_d + \bar{t}_i$. $\bar{t}_d$ and $\bar{t}_i$ are defined by equation 20; t_i is the total injection time; $\bar{x}$ and $\bar{y}$ are defined by equations 18 and 19.

HYDRODYNAMIC DISPERSION AND MIXING

When a fluid is injected into an aquifer and the injected fluid differs from the native water in the aquifer, two kinds of physical mixing processes take place. First, the injected fluid mixes with the native water, provided the two fluids are miscible. Immiscible fluids, such as water and oil, remain physically separated, forming a sharp interface. However, miscible fluids, such as fresh water and salt water, have no such distinct interface; instead, a transition zone forms between them, increasing in extent with time. This phenomenon is known as hydrodynamic dispersion.

The second mixing process occurs as a result of pumping. If injected fluid is within the "water divide" produced by pumping, part or all of the injected liquid moves into the pumping well. Under such a condition, a pumping well draws from the aquifer some native water and some injected fluid.

Harpaz and Bear (1963) proposed an approximate solution for the degree of mixing at such a pumping installation. On the basis of the assumptions that the flow toward the pumping well is radial and that the injected water is a cylindrical body, they derived an equation for calculating the concentration of a selected constituent in the pumped water. The solution is approximate and is justified if the transition zone between the fronts of the two miscible fluids is narrow compared with the size of the injected water body and if the injection time is short.

The following is a discussion of an approximate solution of mixing proposed by Harpaz and Bear (1963).

Consider a center of an injected water body located at a distance, d, from a pumping well (fig. 19). At time t, all fluid particles on the circumference of a cylinder with radius r and height b will reach the circumference of the pumping well at the same time. At that time the injected water will be entering the well through arc ABC, while native water will be entering through arc ADC.

If C is the concentration in mass of solute per unit volume of fluid in the pumped water; C_g, the concentration in the native water; and C_0, the concentration in the injected water, then the relative concentration at any time during the pumping period is defined as (fig. 19):

$$(C-C_g)/(C_0-C_g) = \theta/\pi. \tag{24}$$

By using a trigonometrical relation, θ is obtained as

$$\theta = \arccos\left\{(r^2+d^2-R^2)/(2rd)\right\}. \tag{25}$$

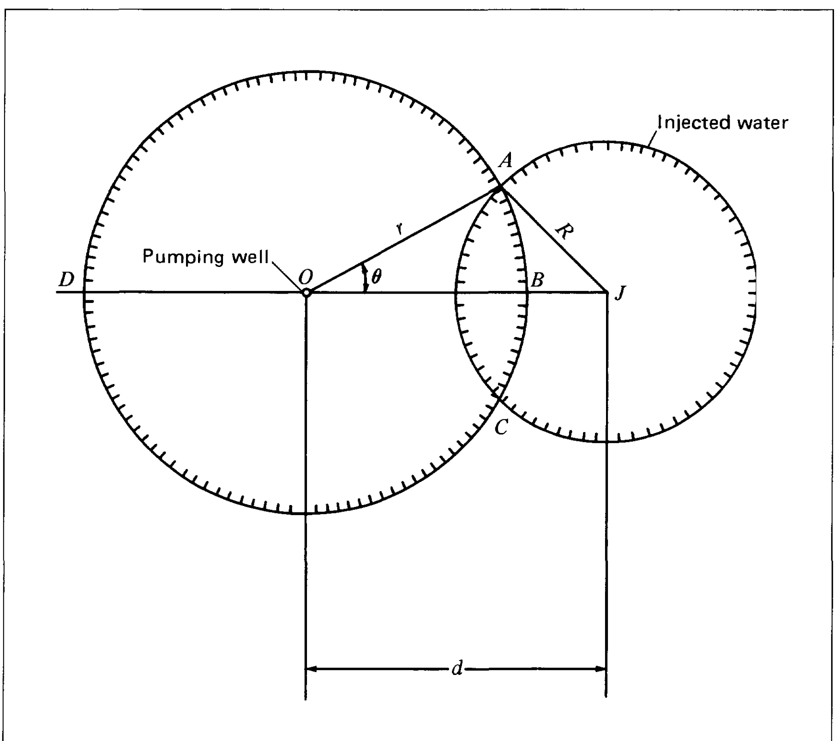

FIGURE 19.—Pumping near a cylindrical body of injected water.

Let

V_i = volume of injected water = $\pi n b R^2$;
V_p = volume of pumped water at time t = $\pi n b r^2$;
V_f = a reference volume = $\pi n b d^2$;
n = effective porosity of the aquifer; and
b = thickness of the aquifer.

Equations 24 and 25 can be combined as

$$(C-C_g)/(C_0-C_g)=(1/\pi) \arccos \left\{[(V_p/V_f)-(V_i/V_f)+1]/2(V_p/V_f)^{1/2}\right\}. \quad (26)$$

The distance, d, which is used to calculate V_f shown in equation 26, can be obtained from the following expression:

$$d=(v/n)t_d=[TI/(nb)]t_d, \quad (27)$$

where v is Darcian velocity; n, effective porosity of the formation; T, transmissivity; I, hydraulic gradient along the natural flow direction; b, effective thickness of the aquifer; and t_d, time since cessation of injection. If the injection lasts so long that the injected water is no longer a cylindrical body, or if a wide transition zone exists, then the degree of mixing at a pumping well can be determined with the aid of equation 23. From equation 23, the extent of the injected water body and the limits of the cone of depression at time t can be determined. Using a planimeter the percentage of the injected water entering the pumping well can be calculated. The above discussion concerns only the conditions of mixing of injected water and native water drawn by a pumping well in the vicinity of a cylindrical body of injected water, but it does not consider the effects of hydrodynamic dispersion.

Hydrodynamic dispersion is present because of variations in local velocity, both in magnitude and direction, along the tortuous paths. Two basic elements involved in hydrodynamic dispersion are the flow and the existence of the pore system through which flow takes place (Bear and others, 1968; Ogata, 1970; Fried and Combarnous, 1971).

During the last 10 to 15 years, extensive research work has been done on problems of hydrodynamic dispersion through porous media. This work has resulted in the general acceptance of a differential equation that describes distribution of an injected dissolved substance in fluids flowing through uniform porous media. The following assumptions must be made to derive the differential equation shown by equation 28: (1) Darcy's law governs the flow, (2) no chemical reaction occurs between the solid and the liquid phases, (3) no loss or addition of substance takes place in the flow system, (4) no vertical flow occurs, and (5) the concentration distribution is symmetrical with respect to polar angle θ.

Under these assumptions, the general differential equation describing the spatial and temporal distribution of a dissolved substance in a radial flow pattern from a source or to a sink (Lau and others, 1959; Bachmat and Bear, 1964; Hoopes and Harleman, 1967; Ogata, 1970) is

$$\partial C/\partial t \pm \bar{v}\,(\partial C/\partial r) = D(\partial^2 C/\partial r^2), \quad (28)$$

in which C is concentration in mass of solute per unit volume of fluid; $\bar{v}$, the mean seepage velocity between the well and the radial distance, r; D, coefficient of hydrodynamic dispersion; and t, time. Positive sign before $\partial C/\partial r$ is for divergent flow (for example, injection), and negative sign is for convergent flow (for example, pumping).

The coefficient of hydrodynamic dispersion, D, depends on velocity. By experiments, two limiting cases have been found (Scheidegger, 1960; Bear and others, 1968):

$$D = D_m \bar{v}^2, \tag{29}$$

and

$$D = D_m \bar{v}, \tag{30}$$

in which D_m is the medium dispersivity. Theoretically and experimentally, the value of D_m is related to the pore geometry of the medium.

In equation 29, the combined effects of a velocity distribution across a flow channel and transverse molecular diffusion are considered (Bear and others, 1968). Taylor (1953) found this relation by experiments of flow through a tube. In equation 30, only the main motion in a flow channel is considered, while mixing occurs at junctions connecting different flow channels (Bear and others, 1968). Experiments run in porous media seem to indicate that equation 30 corresponds to physical reality (Scheidegger, 1961). Therefore, equation 28 is written as

$$\partial C/\partial t \pm \bar{v}(\partial C/\partial r) = D_m \bar{v}(\partial^2 C/\partial r^2). \tag{31}$$

The boundary conditions for a recharge well with a radius r_w fully penetrating a confined homogeneous and isotropic aquifer of infinite areal extent are

$$\begin{aligned} C(r_w, t) &= C_0, \text{ when } t \geqslant 0, \\ C(r, 0) &= 0, \text{ when } r > r_w, \\ C(\infty, t) &= 0, \text{ when } t \geqslant 0. \end{aligned} \tag{32}$$

Attempts to solve equations 31 and 32 analytically have failed to date. Ogata (1970) has obtained an integral form solution. The integral solution is too complicated to evaluate because of the appearance of a Bessel function in both numerator and denominator. Numerical integration would be required.

If an approximate analytical solution of acceptable accuracy for practical problems could be found, it would have advantages over the numerical scheme. First, its time of calculation would be shorter. Second, the concentration at any point or time could be calculated directly without having to carry the solution all the way from the initial conditions to the desired point.

Based on two assumptions, Lau, Kaufman, and Todd (1959) found an approximate dispersion solution for divergent radial flow by superimposing the radial-flow geometry on the one-dimensional longitudinal dispersion solution solved from equation 31. Their first assumption is that the breakthrough curves in radial flow have a nearly normal distribution as breakthrough curves for one-dimensional flow. The second assumption is that the solution can be separated into the effects of radial distance and divergence.

Based on the above assumptions, the approximate solution of equations 31 and 32 is

$$C/C_0 = \tfrac{1}{2}\, erfc\left\{(r-\bar{r})/[(4/3)(D_m\ \bar{r})]^{1/2}\right\}, \tag{33}$$

where C_0 is the initial concentration in the injected water; $\bar{r}$, the average radius of the injected water body; C, the concentration at time t; D_m, the medium dispersivity; and $erfc$ (x), the complimentary error function. Results obtained from this approximate solution are within acceptable error limits for solving practical problems (Hoopes and Harleman, 1967; Schamir and Harleman, 1967).

If V_0 represents the required volume for the circular front of the injected water body reaching the radial distance, r, then

$$r = [V_0/(\pi nb)]^{1/2},$$

and

$$\bar{r} = [V/(\pi nb)]^{1/2}, \tag{34}$$

where V is the injected volume at any time t.

Substituting equation 34 into equation 33, then

$$C/C_0 = \tfrac{1}{2}\, erfc[\{1/[(4D_m/3V_0)^{1/2}(\pi nb)^{1/4}]\}\, U];\ U = [1-(V/V_0)^{1/2}]/(V)^{1/4}, \tag{35}$$

which is the approximate dispersion solution for a radial and divergent flow.

SINGLE-WELL TRACER-DILUTION METHOD

If ground-water flow is governed by Darcy's law, then flow velocity can be calculated by Darcy's equation, when hydraulic gradient and hydraulic conductivity are known. Hydraulic conductivity can be determined from a pumping test, even if there is only one well in the area, but hydraulic gradient can not be determined under this condition. However, natural flow velocity can be estimated by the use of the single-well tracer-dilution method.

In the single-well tracer-dilution method, the concentration decrease of a tracer solution in a well is measured over a period of time. Consider a well fully penetrating a confined homogeneous isotropic aquifer with a uniform natural flow. A tracer solution at a concentration C_0 is introduced into and evenly dispersed in the well. If the rate of decay of the tracer substance and the extent of molecular diffusion are small, they may be neglected. Also, it is assumed that there is no vertical flow in the well. The change in concentration of tracer in the well is then caused by dilution of the horizontal flow that passes through the well.

Let

C_0= initial tracer concentration of water in the well;
C = tracer concentration of water in the well at time t;
C_g= tracer concentration of native water which enters the well;

V = volume of water in the well;
L = length of effective perforation of the well;
l = length of column of water in the well;
D = diameter of the well;
b = thickness of the aquifer;
K = hydraulic conductivity of the aquifer;
T = transmissivity of the aquifer;
I = hydraulic gradient of the aquifer at the well site;
v = flow velocity = KI;
A = width of flow field intercepted by the well (fig. 20); and
t = time.

According to the law of conservation of mass, the change in tracer mass in the well is equal to the tracer mass leaving the well. Therefore,

$$V(dC/dt) = -ALv\ (C - C_g). \qquad (36)$$

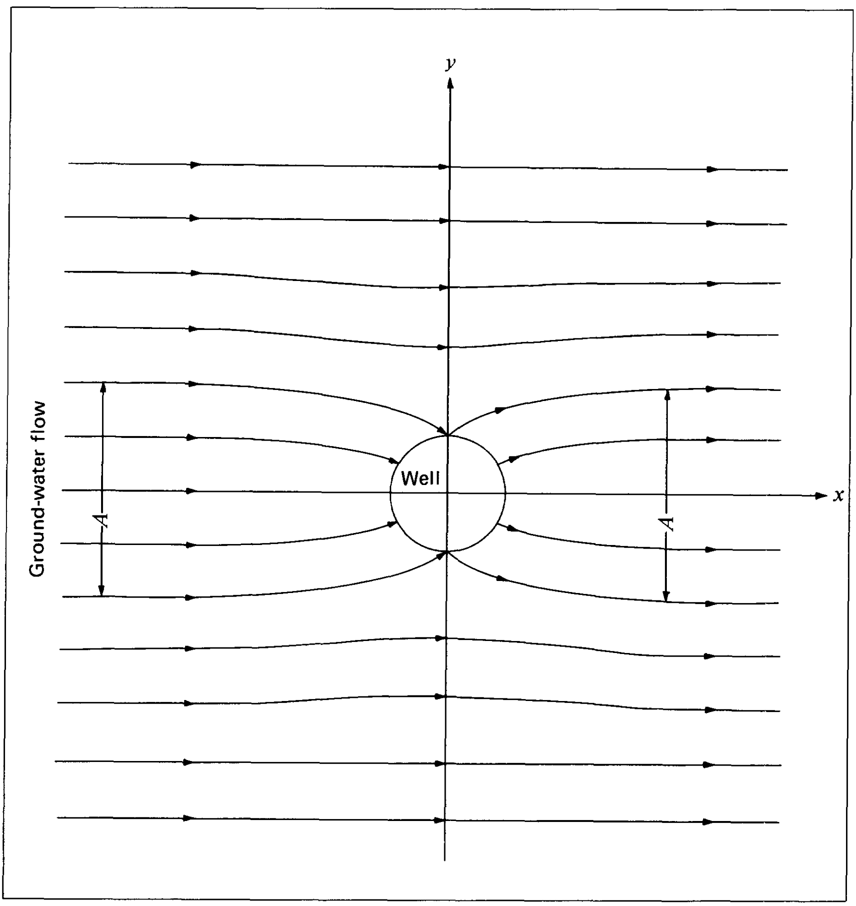

FIGURE 20.—Flow field intercepted by a well.

After integration and substituting initial conditions, $t = 0$, $C = C_0$, into the integrated solution, the following equation is obtained.

$$(C - C_g)/(C_0 - C_g) = \exp\,(-ALvt/V). \tag{37}$$

Because

$$V = (\pi/4)D^2 l\,,$$

and

$$v = TI/b,$$

then equation 37 can be written as

$$(C - C_g)/(C_0 - C_g) = \exp(-4ATILt/\pi D^2 lb). \tag{38}$$

The value of A can be determined from the potential theory. Consider a well to be constructed with perforated casings and gravel packing, then three concentric zones with different hydraulic conductivities are formed around the well (fig. 21). The equation to calculate the value of A is given (Halevy and others, 1967; Drost and others, 1968; and Saleem, 1971) by

$$A = 16\,r_1 \Big/ \left(1+\frac{K_3}{K_2}\right)\left\{\left[1+\left(\frac{r_1}{r_2}\right)^2\right]+\frac{K_2}{K_1}\left[1-\left(\frac{r_1}{r_2}\right)^2\right]\right\}+\left(1-\frac{K_3}{K_2}\right)$$
$$\left\{\left[\left(\frac{r_1}{r_3}\right)^2+\left(\frac{r_2}{r_3}\right)^2\right]+\frac{K_2}{K_1}\left[\left(\frac{r_1}{r_3}\right)^2-\left(\frac{r_2}{r_3}\right)^2\right]\right\}, \tag{39}$$

where K and r are the hydraulic conductivity and radius of each zone shown in figure 21, respectively.

If a well is constructed without gravel packing, then the three concentric zones will be reduced to two. Thus, $K_2 = K_3$, equation 39 can be written as

$$A = 8r_1 \Big/ \left\{[1+(r_1/r_2)^2]+(K_2/K_1)[1-(r_1/r_2)^2]\right\}. \tag{40}$$

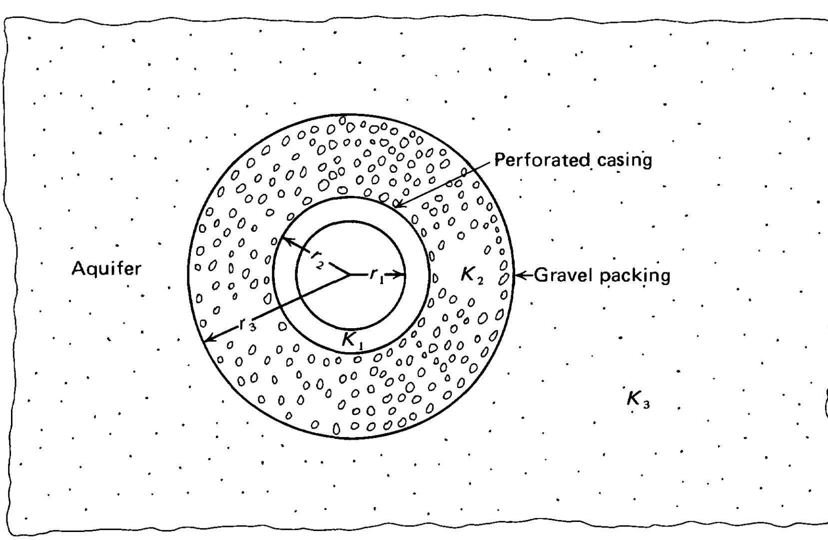

FIGURE 21.—Well with perforated casing and gravel packing in a homogeneous aquifer.

Also, if a well is an uncased open hole, then $r_1=r_2$ and $K_1=K_3$, and the value of A becomes

$$A = 4r_1 = 2D, \tag{41}$$

where D is the diameter of the well.

INTERPRETATION OF INJECTION-TEST DATA

The injection experiment in West St. Paul, Minn., was designed to study the problems of artificial recharge of fissured carbonate rocks, without specific concern for studying the hydrodynamic dispersion and mixing processes involved. Nevertheless, an attempt was made to interpret the nature of dispersion and mixing on the basis of the available data.

The water was injected into the Prairie du Chien Group through well 1-S for 20 days, from November 4 (Thursday) to November 24 (Wednesday), 1971. Chloride was used as the tracer element. Kaufman and Orlob (1956) found that chloride is a good tracer where density effects can be avoided. Total injection was 368,200 ft³ (10,430 m³) at an average rate of 18,410 ft³/d (521.4 m³/d). Pumping was started on February 25, 1972, 93 days after the injection was ended, and continued through March 8, 1972. Interruptions due to intermittent pumping made the interpretation of mixing of the injected water during the entire pumping period impractical. Total pumpage was 267,433 ft³ (7,574 m³) at an average rate of 53.5 ft³/min (1.52 m³/min).

Water samples were taken from the observation wells 1-N, 2-N, and 2-S for chloride determination. The observed data are shown in figures 22 through 24, respectively. The chloride concentrations observed in the pumped water during the pumping period after the injection are shown in table 15. The average chloride concentration in the injection water was determined to be 25.3 mg/l. Chloride concentrations observed in the injection well between the end of injection and start of pumping are shown in table 16.

Table 3 shows the records of water levels observed at the injection well (1-S) and the two observation wells, 2-S and 2-N, from April 9, 1971, through February 2, 1972, that were used to determine the direction of flow and the hydraulic gradient in the Prairie du Chien Group. The estimated flow direction ranged from N. 25°E. to N. 40°E. from July through December 1971 and ranged from N. 50°E. to N. 60°E. from April through May 1971 and in February 1972. The average flow direction was about N. 40°E. (See fig. 25.) The natural flow direction and hydraulic gradient in October and December 1971, were estimated to be N. 36°E and 0.0013; N. 39°E. and 0.0012; respectively (table 3). Five days after the end of injection (November 29, 1971), the flow direction was estimated to be N. 32°E., and the hydraulic gradient to be 0.0013 (table 3). Therefore, the influence of the injection on head gradients and flow rates was concluded to have diminished 5 days after injection stopped. The decrease of chloride concentration in the observation well 2-S since that time probably was caused mainly by the natural flow.

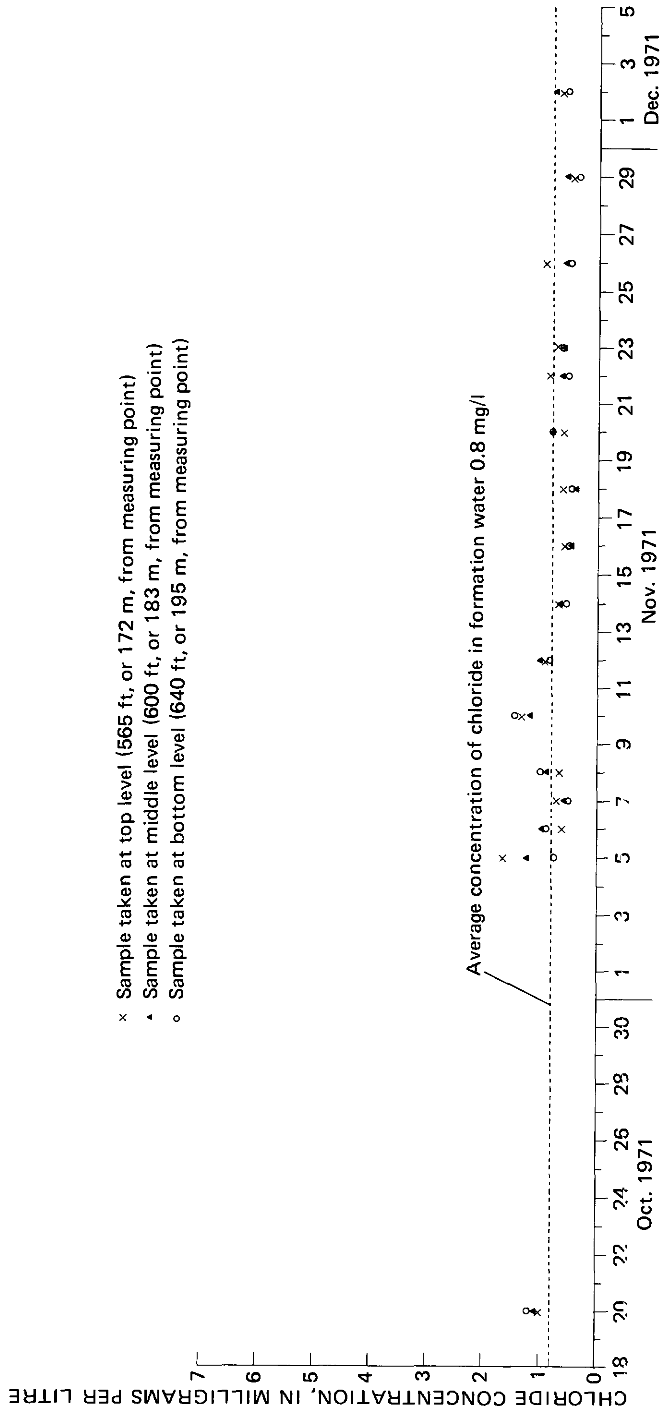

Figure 22.—Concentration of chloride observed in well 1-N.

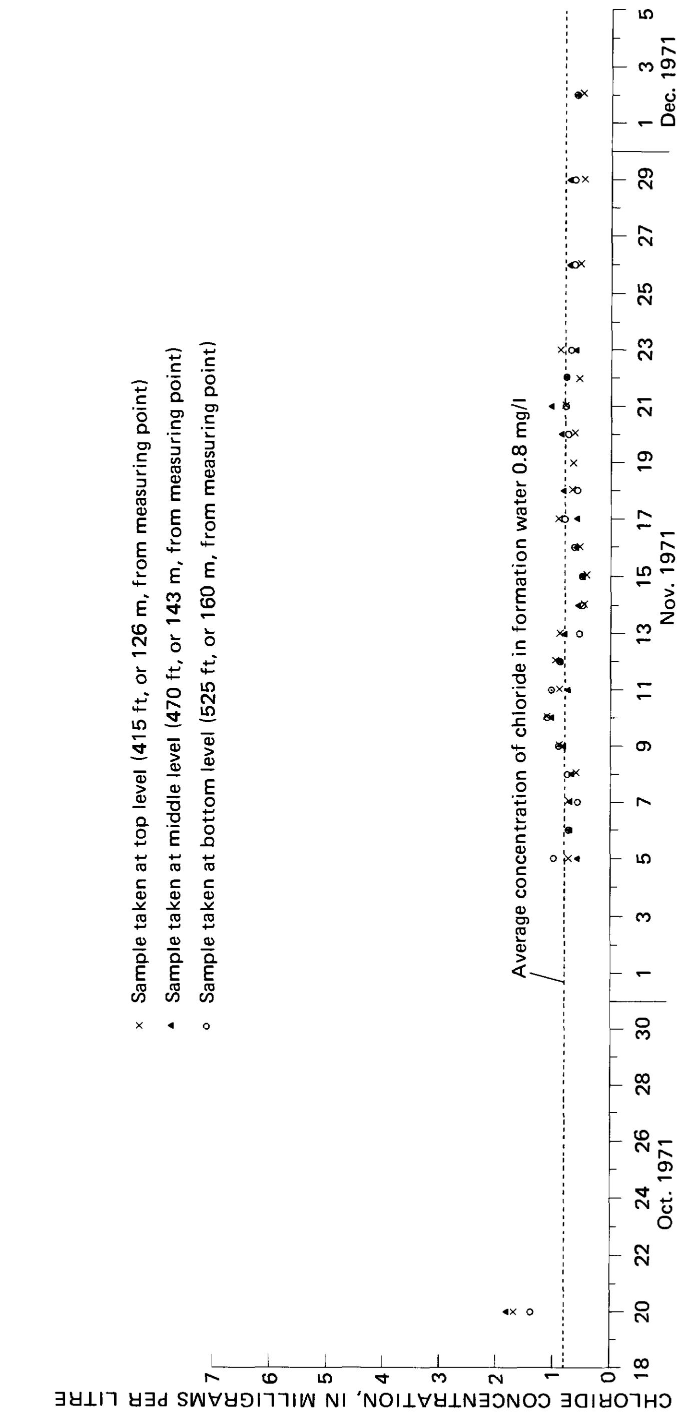

FIGURE 23.—Concentration of chloride observed in well 2–N.

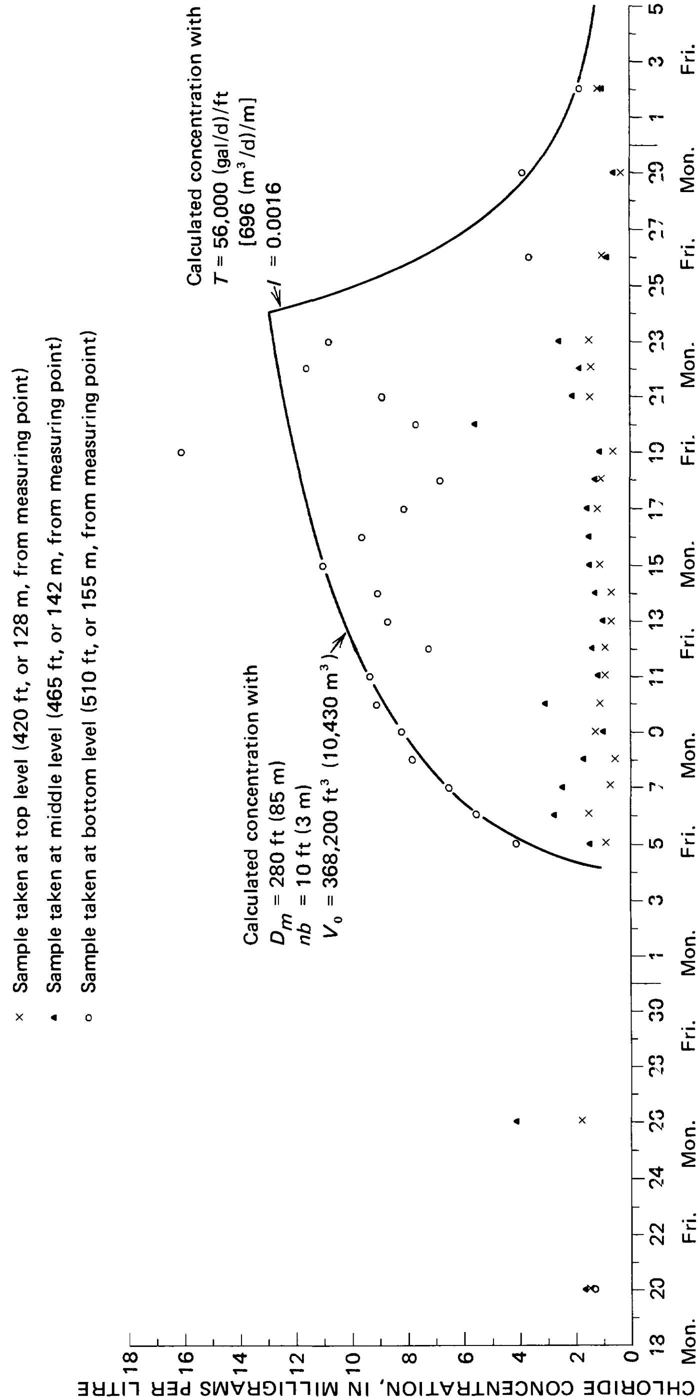

FIGURE 24.—Concentration of chloride observed in well 2–S compared with the calculated concentrations.

TABLE 15.—*Chloride concentrations observed in pumped water*
[Water samples were taken at well 1-S]

Date	Time (hr and min)	Volume of water pumped daily at the time chloride concentration was observed (ft^3)	Volume of water pumped during the day (ft^3)	Chloride concentration (mg/l)
2-25-72	1115	1,951.00	16,288.14	2.85
	1600	16,288.14		4.10
2-28-72	0840	1,751.71	35,577.78	2.70
	1110	10,812.26		3.10
	1605	28,631.35		2.75
2-29-72	1115	7,748.54	31,095.54	2.90
	1730	30,509.91		5.50
3- 1-72	0900	23,898.00	31,882.60	3.90
	1800	31,882.60		3.30
3- 2-72	1000	6,413.78	32,336.16	3.45
3- 3-72	1000	4,676.34	22,344.50	3.05
	1600	22,194.43		3.35
3- 6-72	1100	8,707.87	28,947.80	2.60
	1700	25,652.53		3.00
3- 7-72	0900	2,854.76	28,547.60	2.90
3- 8-72	0900	3,839.84	39,486.40	2.80
	1445	25,918.91		2.90

TABLE 16.—*Chloride concentrations observed in the injection well (1-S) between end of injection and start of pumping*

Depth sample taken from (ft)	425	485	545
Date	Chloride concentration (mg/l)		
12- 2-71	8.0	24.5	25.0
12- 8-71	24.2	10.2	13.2
12-10-71	15.0	9.6	12.0
1- 5-72	5.8	7.8	9.1
1-21-72	23.5	14.5	19.5
2- 2-72	23.0	16.6	7.6
2-14-72	17.8	11.3	9.4

A test of the single-well tracer-dilution method (eq 38) was made by comparing measured values of chloride concentration with those computed from the equation. The boundary conditions existing following the end of the injection departed from those assumed for derivation of the equation. However, because the injected fluid just had reached well 2-S, it was assumed that the method might be applicable. Well 2-S is constructed as an open hole, so that the width of flow field intercepted by this well is twice as wide as the well diameter (eq 41), which is 5 inches (127 mm). The length of water column in the well containing the chloride is the same as the length of the effective perforation of the well; therefore, L

is equal to l in equation 38. By using the value of transmissivity of 56,000 (gal/d)/ft [695 (m^3/d)/m] (as indicated from the hydraulic tests to be representative of the Prairie du Chien and Jordan together as an aquifer), the thickness of the aquifer (the Prairie du Chien Group and the Jordan Sandstone) of 242 feet (74 m), the best value of hydraulic gradient to fit the chloride concentrations (fig. 24) by the single-well tracer-dilution method

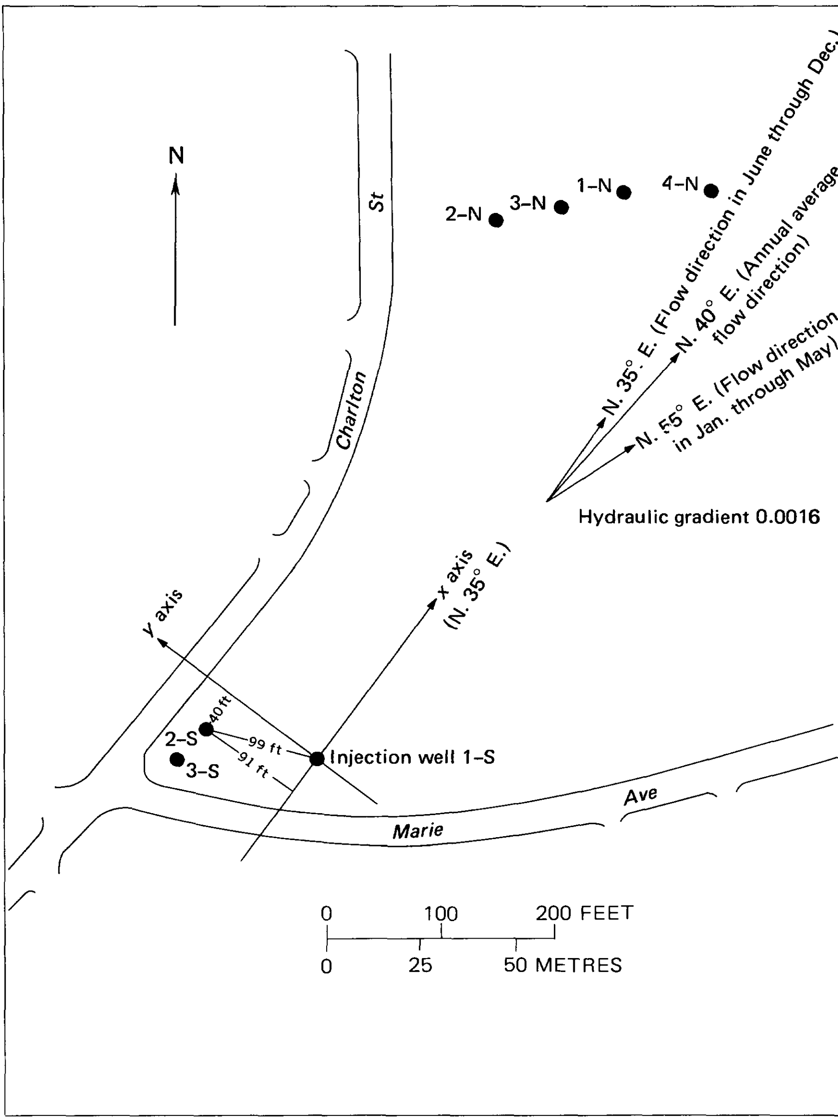

FIGURE 25.—Natural flow direction in the Prairie du Chien at the West St. Paul site, Minn.

is determined to be 0.0016. This value is close to the hydraulic gradient of 0.0013 estimated from the water-level measurements. However, there are only three measured values of concentration for the post-injection period, and only two points fall on the calculated curve. Therefore, it is impossible to judge the applicability of the method.

A value for the medium dispersivity was computed from the data of concentration against time for well 2–S, using equations 17 through 20 and equation 35. For the calculations, 0.0013 and 56,000 (gal/d)/ft [695 (m^3/d)/m] were used as the natural hydraulic gradient and the transmissivity of the aquifer, respectively. The dispersion equation (eq 35) shows that the front of the injected water body arrives at an observation well when the relative concentration C/C_0 reaches 50 percent. The arrival time at an observation well can be determined by plotting observed C/C_0 against time, t, on log-probability paper with t on the log-coordinate. All data should fall nearly on a straight line.

The relation of t against C/C_0 observed at well 2–S was plotted on log-probability paper (fig. 26), and a straight line was drawn to pass through most of the observed data, but also considering the weekly pumping effects at the South St. Paul area (that is, water-level declines from Monday through Friday and rises from Friday through Monday in the wintertime). This plot shows that 20 days after the start of injection, the injected water would reach well 2–S.

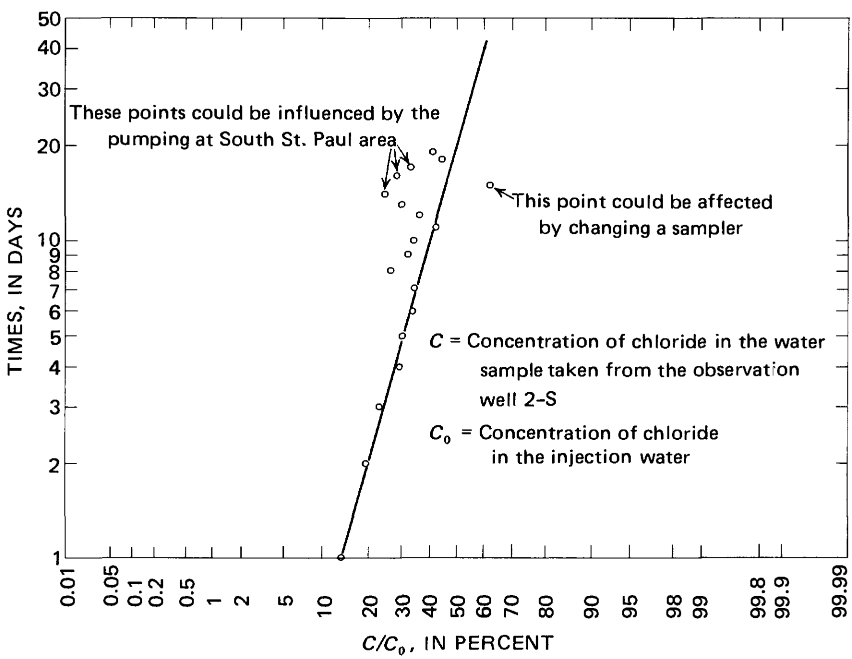

FIGURE 26.—Time and observed C/C_0 at well 2–S plotted on log-probability scale.

Before determining the medium dispersivity D_m, it is necessary to determine the effective thickness of the aquifer, nb. From figures 25 and 28 and equations 17 through 20 with t=20 days, the value of nb is obtained as follows.

$$TI = [56{,}000\ (\text{gal/d})/\text{ft}]\ (0.13368\ \text{ft}^3/\text{gal})\ (0.0013) = 9.73\ (\text{ft}^3/\text{d})/\text{ft};$$

$$\bar{X} = (2\pi TI/18410)\ (-40) = -0.1328;$$

$$\bar{y} = (2\pi TI/18410)\ (91) = 0.3022;$$

$$\bar{t} = \left\{2\pi (TI)^2/[(18410)\ (nb)]\right\}(20) = 0.6462/nb;$$

$$0.6462/nb = -0.1328 - \ln\left[\cos 0.3022 - \left(\frac{0.1328}{0.3022}\right) \sin 0.3022\right]; \text{ and}$$

$$nb = 10.6 \text{ feet } (3.23 \text{ m}).$$

If the hydraulic gradient of 0.0016 estimated from the single-well tracer-dilution method and the transmissivity of 56,000 (gal/d)/ft [695 (m^3/d)/m] are to be used in the calculation, then the value of the effective thickness of the aquifer, nb, is determined to be 10.3 feet (3.14 m). Therefore, it can be concluded that the effective thickness of the aquifer is about 10 feet (3 m).

According to characteristics of the normal distribution of the dispersion equation (eq 35), plotting U versus C/C_0 on arithmetic-probability paper, a straight line will connect all points and will pass through the point of (U=0, C/C_0= 50 percent). From this relation and by using the chloride concentrations observed at well 2–S, D_m can be found. By using equation 35 and C/C_0= 20 percent and 80 percent, as shown in figure 27, D_m is found.

$$D_m = (3/4)\ (368{,}200)\left\{0.09/[(1.2)\ (10\pi)^{1/4}]\right\}^2$$
$$= 277 \text{ feet, say } 280 \text{ feet } (85 \text{ m}).$$

The large value of medium dispersivity (D_m) that resulted from field tests is typical of a fractured reservoir. Webster, Proctor, and Marine (1970) found that D_m is equal to 440 feet (134 m) in the fractured crystalline rock formation at the Savannah River Plant near Aiken, S.C. Grove and Beetem (1971) found D_m to be 125 feet (38 m) for a fractured-carbonate-rock aquifer in Eddy County, N. Mex.

The chloride concentrations at well 2–S have been caluclated by using the value of nb and D_m as determined above. The correlation between the calculated and observed concentration is shown in figure 24. Except for the highest concentration point, which occurred on November 19, 1971, all the low concentration points deviated from the calculated concentration curve (fig. 24), following the weekly water-level fluctuation pattern induced by the pumping at the South St. Paul area (which is east of the experiment site). The deviation of the point on November 19, 1971, is probably due to the change of the sampler; a ball-type sampler was used throughout the test except on November 19, 1971, when a plunger sampler was used.

Theoretically, chloride concentration in well 2–N, which is 495 feet (151 m) away from the injection well, should be 0.9 mg/l at the end of the

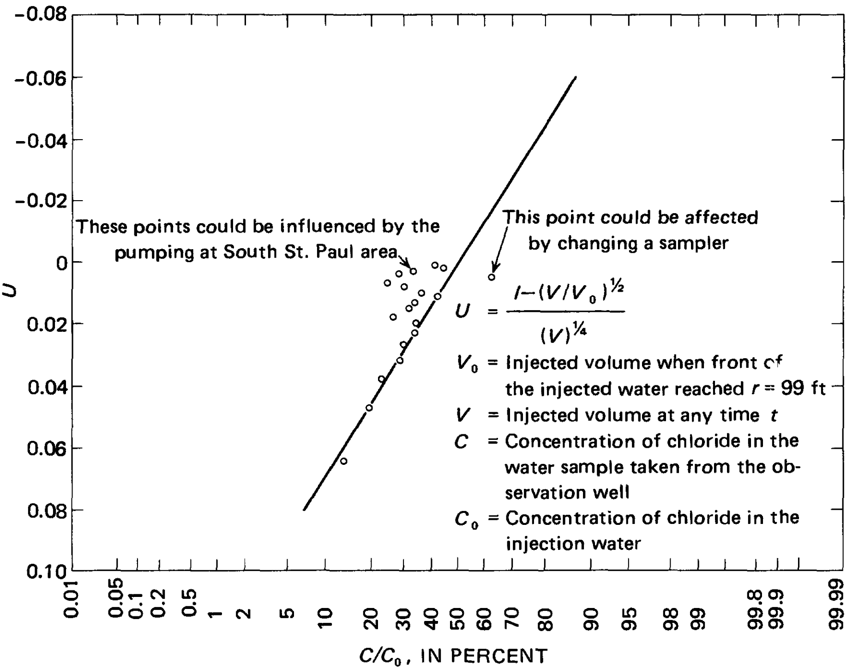

FIGURE 27.—U and observed C/C_0 at well 2–S plotted on arithmetic-probability scale.

injection. This value is nearly that of the chloride concentration of the native water. Therefore, no evidence of increased chloride concentration at well 2–N was observed, nor would it have been theoretically expected (fig. 23).

Based on the determined values of nb and TI, the natural flow rate ranges between 0.97 and 1.20 ft/d (0.30 and 0.37 m/d). Therefore the average flow rate is probably 1.0 ft/d (0.30 m/d). Possibly, fluid in some fractures may have a flow rate as high as a few feet per day, while in other fractures flow may be at a low rate, perhaps only a few inches per day.

An attempt was made to compare the calculated chloride concentrations in the pumped water with the observed values. Because of the intermittent pumping, a steady-state condition was not reached, and the data became too complicated for interpretation. Nevertheless, a correlation between the observed chloride concentrations in the pumped water during the first pumping day and the chloride concentration predicted by the mixing model and the estimated hydraulic parameters of the aquifer was still made.

The position of the two fronts of the injected water body along the x axis at the end of the injection can be found from equation 17 by setting $\bar{y}$=0.

$$\bar{t}=\bar{x}-\ln(1\pm\bar{x}),$$

where $\ln(1-\bar{x})$ is for upstream front and $\ln(1+\bar{x})$ is for downstream front.

By using the hydraulic gradient of 0.0016 estimated by the single-well tracer-dilution method and the transmissivity of 56,000 (gal/d)/ft [695 (m^3/d)/m], the calculated location of the upstream front along the x axis is found at x=–93 feet (28 m) and the downstream front at x=125 feet (38 m). If the injected water body is assumed to be in a cylindrical form, then its center lies at x=16 feet (4.9 m), and its radius is 109 feet (33 m) (fig. 28).

Between the end of the injection and the beginning of the pumping period, the center of the injected water body moved downgradient with the natural flow. The travel distance, d, can be found either by equation 23 or by equation 27, and the distance is found to be 111 feet (34 m). Therefore, the center of the injected water body before repumping is at x=127 feet (39 m).

The radius of the extent of the pumping during the first day of pumping is

$$r=(16{,}288/10\pi)^{1/2}=23 \text{ feet (7 m)}.$$

According to figure 28 and as derived by equation 26, the calculated chloride concentration in the pumped water at the end of the first day of pumping is 4.5 mg/l. The observed concentrations are 2.85 and 4.10 mg/l. The correlation between the predicted value and the observed data is good. However, if the hydraulic gradient of 0.0013 estimated from the water-level measurements and the determined transmissivity of 56,000 (gal/d)/ft [695 (m^3/d)/m] are used in the same calculation, then the calculated chloride concentration in the pumped water at the end of first day of pumping is determined to be 14 mg/l, which is much higher than the observed concentrations. Therefore, the hydraulic gradient at the test area is concluded to be about 0.0016.

DISCUSSIONS AND CONCLUSION

By using the previously determined value of transmissivity of 38,000 (gal/d)/ft [472 (m^3/d)/m], the thickness of the Prairie du Chien Group of 150 feet (46 m) and the hydraulic gradients of 0.0013 and 0.0016, respectively; the chloride concentrations calculated by the single-well tracer-dilution method for the postinjection period at well 2–S roughly correlate with the observed data. The movement of the injected water under these conditions, however, indicates that the water pumped from the aquifer during the first pumping day after the injection test should be the injected water that had a chloride concentration of 25.3 mg/l. The observed chloride concentrations were 2.85 and 4.10 mg/l. Therefore, it can be concluded that the transmissivity of the aquifer receiving the injection is about 56,000 (gal/d)/ft [695 (m^3/d)/m] and that the Prairie du Chien Group is hydraulically connected with the Jordan Sandstone.

The determined ground-water flow rate at the site is about 1.0 ft/d (0.3 m/d), with a natural (not affected by injection) hydraulic gradient of 0.0016. The average flow direction is N. 40°E., with average seasonal variations of N. 35° E., June through December and N. 55° E., January through May (table 3).

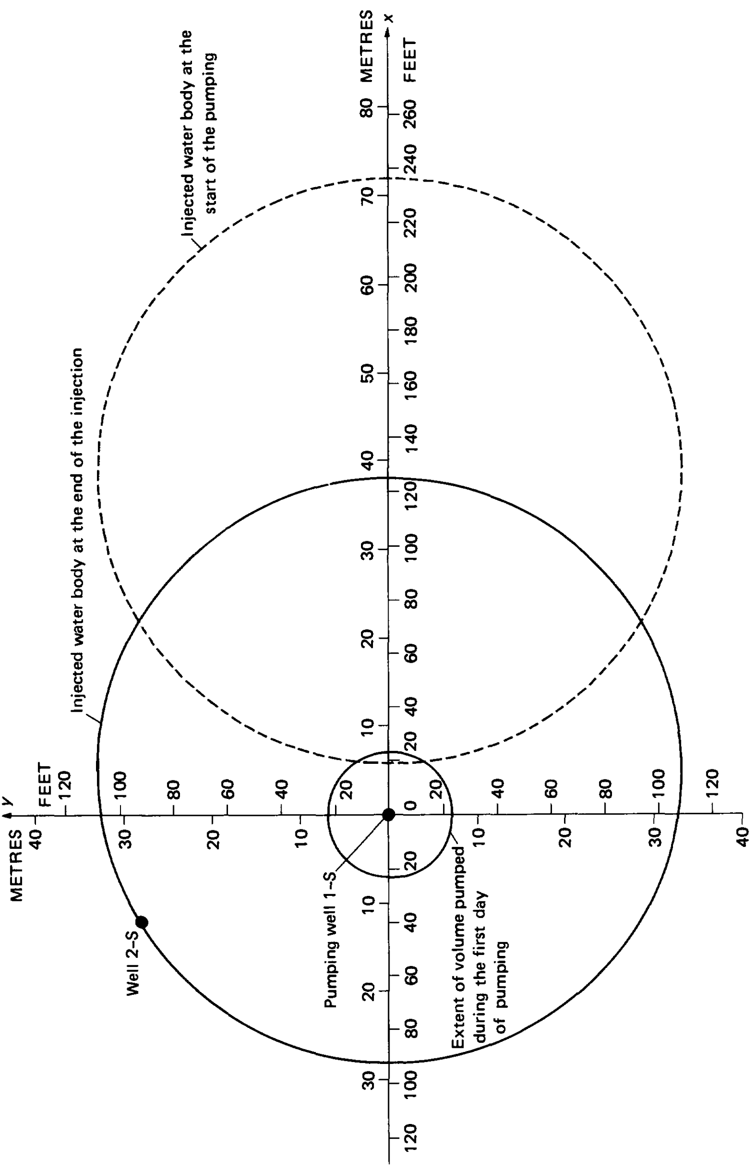

FIGURE 28.—Position of injected water body and the extent of the volume pumped during the first day of pumping.

The effective thickness of the aquifer, *nb*, is determined to be 10 feet (3 m). Therefore, the effective porosity is about 4 percent.

The medium dispersivity D_m is 280 feet (85 m), which is determined from the breakthrough curve of chloride concentration observed at well 2-S. Orlob and Radhakrishna (1958) stated that the shape of a breakthrough curve must be related to the geometry of the intricate passages through which the flow must pass. Hence, the shape of the curve reflects the degree of tortuosity in the medium, the lack of perfect isotropy, and variations in size of passages. As the degree of heterogeneity increases, the degree of dispersion also increases, and, therefore, the value of medium dispersivity becomes larger. They also found that inclusion of a small amount of air could decrease the uniformity of pore spaces, thus increasing dispersion.

In laboratories, flow models are commonly built by using either glass beads or highly uniform sand. The reported medium dispersivity, D_m, determined in laboratories is only a few inches (Ogata and Banks, 1961; Rifai and others, 1956; Hoopes and Harleman, 1967). In the field, the degree of uniformity of aquifers decreases and the determined values of dispersivity are as much as three orders of magnitude higher than those reported from laboratories. Tests run in sand and sandstone aquifers show that values of dispersivity are only a few feet (Mercado, 1966a, b), but in fractured dolomite or limestone aquifers, the reported values of D_m are higher. They range from several tens of feet to several hundreds of feet, depending on the degree of heterogeneity of the aquifer (Schwarz, 1963; Webster and others, 1970; Grove and Beetem, 1971).

The value of the medium dispersivity determined from this test clearly corroborates the fact that the fractured dolomite of the Prairie du Chien Group at the test site is a heterogeneous aquifer. Presumably, there are a few fractures directly connecting the injection well and observation well 2-S; other interconnecting fractures give rise to longer flow paths.

All dispersion and hydrodynamic theories discussed above are based on the assumptions that the flow system is isotropic and homogeneous, and that the flow is laminar, two dimensional, and governed by Darcy's law. In fractured reservoirs, water may flow so fast in some fractures that the flow may not be laminar or in two dimensions. Thus, the assumptions are violated. When a flow system is different from the assumed theoretical model, the theory is inadequate to describe the flow. Unfortunately, at the present time, no adequate theories have yet been developed for flow through a fractured aquifer. Hopefully, the existing theories can be applied advantageously, at least from a coarse macroscopic point of view.

If tests are planned properly, the results should theoretically reflect some average characteristics of the aquifer. Because of the heterogeneity of a fractured aquifer, it is necessary to test the aquifer from a large scale point of view. This would require a long period of injection. It would also require a larger network of observation wells distributed around the injection well. A few observation wells may only represent the extreme conditions in the tested area. For example, field tests run at a fractured dolomite aquifer at the Hof Hacarmel well field in Israel showed that values of medium dispersivity determined from two observation wells differed by a factor of three (D_m=21 ft (6 m) versus D_m=74 ft (22 m); Schwarz, 1963). More observation wells would have given better average values.

SUMMARY AND CONCLUSIONS

Tests were made near St. Paul, Minn., to determine the feasibility of artificially recharging a consolidated aquifer by well injection. The formations of primary concern at the recharge experiment site are the Prairie du Chien Group, into which the water was injected, and the underlying Jordan Sandstone, which is the major aquifer in the area. The Prairie du Chien is made up primarily of well-fractured dolomite, which contains solution openings. The Jordan Sandstone is more uniform and is considered to be homogeneous on a large scale, although small fractures and variations in material are recognized locally. The two layers are hydraulically interconnected and are artesian. A head difference of 2 feet (0.61 m) is noted between water levels in a well open only in the dolomite and in a nearby well open only in the sandstone.

For the tests, a production well was installed in the Prairie du Chien Group, and observation wells were installed nearby, both in the Prairie du Chien and in the Jordan Sandstone. Pumping tests were performed before and after injection, and the injection test itself involved recharge of 2,754,000 gallons (10,424 m^3), occurring at a rate of about 100 gal/min (6.3 l/s) for 20 days. Air entrainment during injection was avoided by use of a small-diameter pipe, in which pipe friction was sufficient to maintain positive pressure throughout the column.

Water-level changes in the injection well and observation wells resulting from injection were analyzed in the same ways as were the data from pumping tests of the aquifer before and after injection in order to obtain the comparative values of transmissivity and the storage coefficient of the aquifer. The following table lists the results of the three tests:

Test	Transmissivity from early part of test [(gal/d)/ft]	Transmissivity from early part of test [(gal/d)/ft]	Specific capacity after well loss and partial penetration corrections [(gal/min)/ft]	Transmissivity from specific capacity [(gal/d)/ft]
Initial	38,000	56,000	20.7	44,600
Injection	17,000	56,000	19.7	42,500
After injection	31,000	....	21.0	45,400

The transmissivity value determined for the early part of the injection test by aquifer-test methods was less than half that before injection. The pumping-test data after injection indicates that the transmissivity was about 18 percent less than the original value. The apparent 18 percent reduction in transmissivity, from 38,000 (gal/d)/ft [472 (m^3/d)/m] before the injection to 31,000 (gal/d)/ft [385 (m^3/d)/m] after the injection, is probably not significant because of other variables in the test and analysis. Also, the transmissivity determined from the specific capacities shows no change, and the chemical analyses indicate no chemical plugging of the formation. No adequate explanation has been found to explain the low value of 17,000 (gal/d)/ft [211 (m^3/d)/m] for transmissivity during injection.

Data from the later part of the initial pumping test indicate a transmissivity of 56,000 (gal/d)/ft [696 (m^3/d)/m]. This is probably because the initial part of the test (perhaps the first few minutes) gives the aquifer characteristics of the Prairie du Chien Group, and the later part of the test data shows the characteristics of both the Prairie du Chien and the Jordan Sandstone. That is, the effects of the lower aquifer become more significant with longer duration of pumping. The storage coefficient in the Prairie du Chien Group is in the range of $2\text{x}10^{-4}$ to $2\text{x}10^{-5}$. The storage coefficient is $4\text{x}10^{-4}$, computed from data from the Jordan Sandstone well.

As the aquifer-test data were analyzed by methods that assume homogeneous and isotropic aquifer conditions, the analyses yields only apparent aquifer characteristic values. A comparison of water-level changes against discharge or recharge rates of the three tests showed that the water-level changes in the two observation wells tapping the Prairie du Chien Group, during the injection tests, were substantially greater than those projected from the two pumping tests. The deviations in water-level changes indicate that the methods used to analyze data from these wells may not be wholly applicable, although no satisfactory explanation has been made of the deviations.

The point of primary concern is whether or not water injected into a well in the Prairie du Chien Group also recharges the underlying Jordan Sandstone. Water-level changes observed in the observation wells and in the injection and pumped well show that head (water-level) changes are sufficiently large and rapid to cause transfer of water from one aquifer to the other.

The injection water was detected only in the bottom part of the nearest observation well (2–S), a lateral distance of 99 feet (30.2 m). Chloride utilized as a tracer was detected at that point within the first day of injection.The ground water averaged 0.8 mg/l chloride, and the injected water averaged 25 mg/l. Comparison of measured water quality in well 2–S with calculated values normalized on chloride indicates mixing of the two types of water with little chemical reaction. The lack of chemical reaction may have resulted from both the rapid arrival of the injected water at the observation well and from the formation of weathered reaction rims on the minerals that prevents contact of the water with the fresh faces along the fractures in the aquifer. Chemical analyses of the injection water indicate that the water is undersaturated with the major minerals in the Prairie du Chien. Therefore, the water used in this experiment can be expected to dissolve these minerals, rather than to precipitate the minerals and plug the formation.

Iron seems to be the only element involved in a significant chemical reaction in the experiment. Supersaturation of iron is noted in water from observation wells in the Prairie du Chien and the Jordan. The addition of oxygenated recharge water would dissolve iron from pyrite in the formation, and, ultimatley, $Fe(OH)_3$ would precipitate. The longer the delay in this precipitation, the larger the area through which the water will spread; consequently, the precipitate per unit area will be less. Temperature of the injected water affects the degree of saturation. Cooler water would be preferred, so as not to cause precipitation of minerals; however, it could increase the dissolved oxygen concentration, increasing the potential for pyrite solution. Chemical reactions of all types should occur over a large area and, consequently, give a relatively long life to a recharge well before it needs to be abandoned or rehabilitated. The expected longevity of the well is due to the fissured nature of the rock, the apparent armoring of the minerals along the fractures, and the unsaturated nature of the recharge water.

The results of the microbiological surveillance program were inconclusive. In general, it did not seem that the biological phenomena had any impact on the recharge of the water to the Prairie du Chien Group during this test.

The hydraulic gradients of the aquifer in October and December 1971 (before and after injection) were estimated to be N. 36° E. and 0.0013, and N. 39° E. and 0.0012, respectively, on the basis of measurements of water levels observed in the three wells in the Prairie du Chien Group. The single-well tracer-dilution method of calculation, using the observed chloride concentration and a transmissivity of 56,000 (gal/d)/ft [69? (m^3/d)/m], shows a hydraulic gradient of 0.0016, which is close to the hydraulic gradient of 0.0013 estimated from water-level measurements. However, on the basis of the correlation between the calculated chloride concentration (the mixing model) and the observed chloride concentrations in the pumped water at the end of the first day of pumping, the hydraulic gradient is probably 0.0016.

From analysis of the breakthrough curve for the chloride concentration observed at well 2–S, the effective aquifer thickness—that is, effective porosity times aquifer thickness—is determined to be 10 feet (3 m), as compared with the actual thickness of 242 feet (74 m).

The longitudinal dispersivity of the aquifer is 280 feet (85 m), which is within the range typical of fractured reservoirs, and shows that the aquifer is heterogeneous at the test site. Presumably, a few fractures directly connect the injection well and the observation well 99 feet (30.2 m) away; other interconnecting fractures provide more circuitous paths for transfer of the water.

The injection test demonstrated that it is hydrologically feasible to recharge the Prairie du Chien Group and the Jordan Sandstone through wells in the Prairie du Chien. The fissures in the Prairie du Chien act as conduits through which the water spreads. The water passes into the Jordan from the Prairie du Chien over a larger area than it would if injected directly into the Jordan. Air entrainment and release of dissolved gasses from the water that accompanies a head drop can be avoided by using a small-diameter pipe, a scheme that also reduces the cost of the injection-pipe system. The expense of an injection system can be further reduced by using pumped wells also as injection wells. Many production wells are large enough to accommodate one or more small pipes beside the pump column and could be used alternately for production or for injection.

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