HYDRAULIC PROPERTIES OF THREE TYPES OF GLACIAL DEPOSITS IN OHIO

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CONVERSION FACTORS, ABBREVIATIONS, AND VERTICAL DATUM

Multiply	*By*	*To obtain*
inch (in.)	25.4	millimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
foot per mile (ft/mi)	0.1894	meter per kilometer
foot per minute (ft/min)	0.3048	meter per minute
foot squared per day (ft^2/d)[1]	0.09290	meter squared per day
cubic foot per day (ft^3/d)	0.02832	cubic meter per day
pound (lb)	453.6	gram
pound per foot (lb/ft)	1.488	kilogram per meter
ton per square foot (ton/ft^2)	95.785	kilopascal

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929--a geodetic datum derived from a general adjustment of the first-order level nets of the United States and Canada, formerly called Sea Level Datum of 1929.

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[1] This is the unit used to express transmissivity, a measure of the capacity of an aquifer to transmit water. Conceptually, transmissivity is cubic feet (of water) per day per square foot (of aquifer area) per foot (of aquifer thickness). In this respect, the unit is reduced to its simplest terms.

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ABSTRACT

The effects of thickness, grain size, fractures, weathering, and atmospheric conditions on vertical ground-water flow in glacial deposits were studied at three sites in Ohio that represent ground moraine, end moraine, and lacustrine depositional environments. Vertical hydraulic conductivities computed from pumped-well tests were 3.24×10^{-1} to 6.47×10^{-1} ft/d (feet per day) at the site representing end moraine and 1.17 ft/d at the site representing lacustrine deposits. Test data for the site representing ground moraine could not be used to calculate hydraulic conductivities, but the data indicate that ground water flows through the total thickness of deposits in response to discharge from a lower gravel unit. Vertical hydraulic conductivities computed from pumped-well tests of nested wells and data from drill-core analyses indicate that fractures affect the migration of ground water downward through the glacial deposits at these sites. Flow through glacial deposits is complex; it is controlled by fractures, grain-size distribution, clay content, thickness, degree of weathering, and is also affected by atmospheric conditions.

INTRODUCTION

The sedimentology and structure of Pleistocene glacial deposits affect the hydraulic characteristics and the amounts of infiltration that recharge glacial and bedrock aquifers in Ohio. Variations in ground-water flow can result from variations in the composition, texture, and structural character of the deposits. These deposit variations can cause differences in hydraulic properties over short distances.

Purpose and Scope

This report identifies some of the characteristics that control the hydraulic properties and affect ground-water flow in ground moraine, end moraine, and lacustrine deposits. Three sites selected to represent the generalized types of till-bearing glacial and lacustrine deposits in Ohio are discussed. Grain size, fractures, and composition were analyzed by use of texture analysis; pumped-well tests, mapping of core stratigraphy, and observation of long- and short-term fluctuations of water levels in response to precipitation, evapotranspiration, and atmospheric-pressure changes were also analyzed.

Previous Investigations

The composition of glacial deposits can be a factor in the rate of ground-water flow. The dissolution of carbonate minerals in glacial deposits, particularly till, produces secondary porosity that increases the permeability. Cartwright and others (1977) observed differences in the hydraulic conductivities in laboratory analysis of leachate movement through various mixtures of sand and clay containing either montmorillonite, kaolinite, or illite. Volume shrinkage caused by the oxidation of minerals can enhance the development of fractures and can modify hydraulic conductivity (Hendry and others, 1986). In contrast, secondary mineralization of the matrix or along fractures can decrease hydraulic conductivity.

In sediments, intrinsic permeability is a function of particle size (Summers and Weber, 1984; Shepherd, 1989) and grain-size distribution (Wenzel, 1942; Masch and Denny, 1966; Beard and Weyl, 1973). In general, the larger the median grain size and greater the degree of uniformity of a sediment, the greater the intrinsic permeability.

Previous investigations of the hydraulic properties of glacial tills in the Midwest and Great Plains of North America commonly have divided tills into two units; brown, weathered till overlying gray, unweathered till (Hendry and others, 1986; Keller and others, 1986; Hendry, 1988). Fractures associated with weathering in tills can result in hydraulic conductivities that are one to three orders of magnitude greater than those in unweathered

tills (Grisak and others, 1976; Prudic, 1986; Keller and others, 1988).

Hydraulically conductive fractures generally are not present in thick, unoxidized tills of the Great Plains (Keller and others, 1988; Keller and others, 1989). Field observations and core analyses in Ohio indicate that fracture occurrence is sparse in the unoxidized tills near the oxidized-unoxidized contact and absent beneath this contact. Keller and others (1986) documented fracture permeability at the base of an unweathered till unit 59 ft thick in western Canada. Comparison between field and laboratory hydraulic conductivities (Norris, 1962) show differences of as much as three orders of magnitude in data from tills in Ohio and South Dakota. The differences between hydraulic conductivities in the laboratory and field tests may be caused by interconnected sand lenses and fractures in the field that are not represented in laboratory samples (Sharp, 1984). Bradbury and Muldoon (1990) found hydraulic conductivities measured in laboratory samples to be as much as two orders of magnitude lower than hydraulic conductivities calculated from field tests on the same materials.

The amount of recharge that passes through glacial deposits also is affected by the degree of evapotranspiration. In glacial deposits in the Great Plains, most of the potential recharge from precipitation that penetrates into weathered tills is lost to evapotranspiration (Cravens and Ruedisilli, 1987; Hendry, 1988). In Ohio, about 70 percent of the annual precipitation is returned to the atmosphere by evapotranspiration (Shindel and others, 1990).

As part of the Midwestern Basins and Arches glacial and carbonate Regional Aquifer-System Analysis (MBA-RASA), the sedimentological and structural characteristics of glacial deposits were analyzed at three sites in Ohio to determine the effects of these characteristics on ground-water flow through till, with emphasis on estimation of quantities of recharge to underlying aquifers.

Geologic Setting

The bedrock units in Ohio that are present as subcrops beneath glacial deposits or crop out

consist of Paleozoic sedimentary rocks that range from Ordovician to Permian age (west to east, respectively) across the State (fig. 1). Ordovician bedrock in Ohio is mostly calcareous shale, argillaceous and (or) cherty limestone, and fine, clastic rock (Bugliosi, 1990). Silurian bedrock in Ohio consists mainly of limestone and dolomite, with evaporites and dolomite that grades into shale in the upper part of the Silurian sequence (Shearrow, 1957; Janssens, 1977). The Lower and Middle Devonian rocks are predominantly limestones, whereas the Upper Devonian consists of the Ohio and Olentangy shales.

The regional structure in Ohio is due mainly to the Cincinnati Arch in southwestern Ohio and the Findley Arch in northwestern Ohio (fig. 1). The dips on either side of the arches are gentle, only 5 to 10 ft/mi in the northwest and approximately 25 ft/mi along a line extending from western to central Ohio (Norris and Fidler, 1973).

Approximately two-thirds of the State is overlain by Pleistocene deposits (fig. 2); most are of Wisconsinan age and represent three major stages: Early, Middle, and Late Wisconsinan. The Early and Late stages of Wisconsinan glaciation represent extensive ice advances into southern Ohio (Dreimanis and Goldthwait, 1973). The Middle Wisconsinan is distinguished by the development of soil such as the Sidney paleosol (Gooding, 1975). Advances by the Late Wisconsinan Laurentide ice sheet removed evidence of prior glaciations in many places by eroding the surface and incorporating earlier deposits with those transported by the glacier. In many places, the ice sheets overrode and deposited sediments on top of earlier deposits. The resulting composite landforms contain sedimentologic sequences from multiple glacial advances.

The composition of till in the surficial deposits mainly depends on glacial-flow paths and local geology. Although most of the mineral composition of a till is representative of local bedrock, about 4 percent of the material in Ohio moraines was transported from the Canadian Shield north of the study area (Strobel and Faure, 1987). This material was transported englacially, whereas most transport and deposition occurs along the base of the ice.

Figure 1. Generalized bedrock geology of Ohio and approximate location of the Cincinnati and Findley Arch Axes (modified from Bownrocker, 1920; Burger and others, 1971; Gray and others, 1972).

Figure 2. Generalized glacial geology of Ohio. (Modified from Goldthwait and others, 1979; Darr, 1990.)

The following labels appear on the map:

84° 83° 82° Cleveland 81°

41°

40°

Columbus

Cincinnati 39°

Base from U.S. Geological Survey
State Base Map of Ohio, 1971, 1973

0   50 MILES

0   50 KILOMETERS
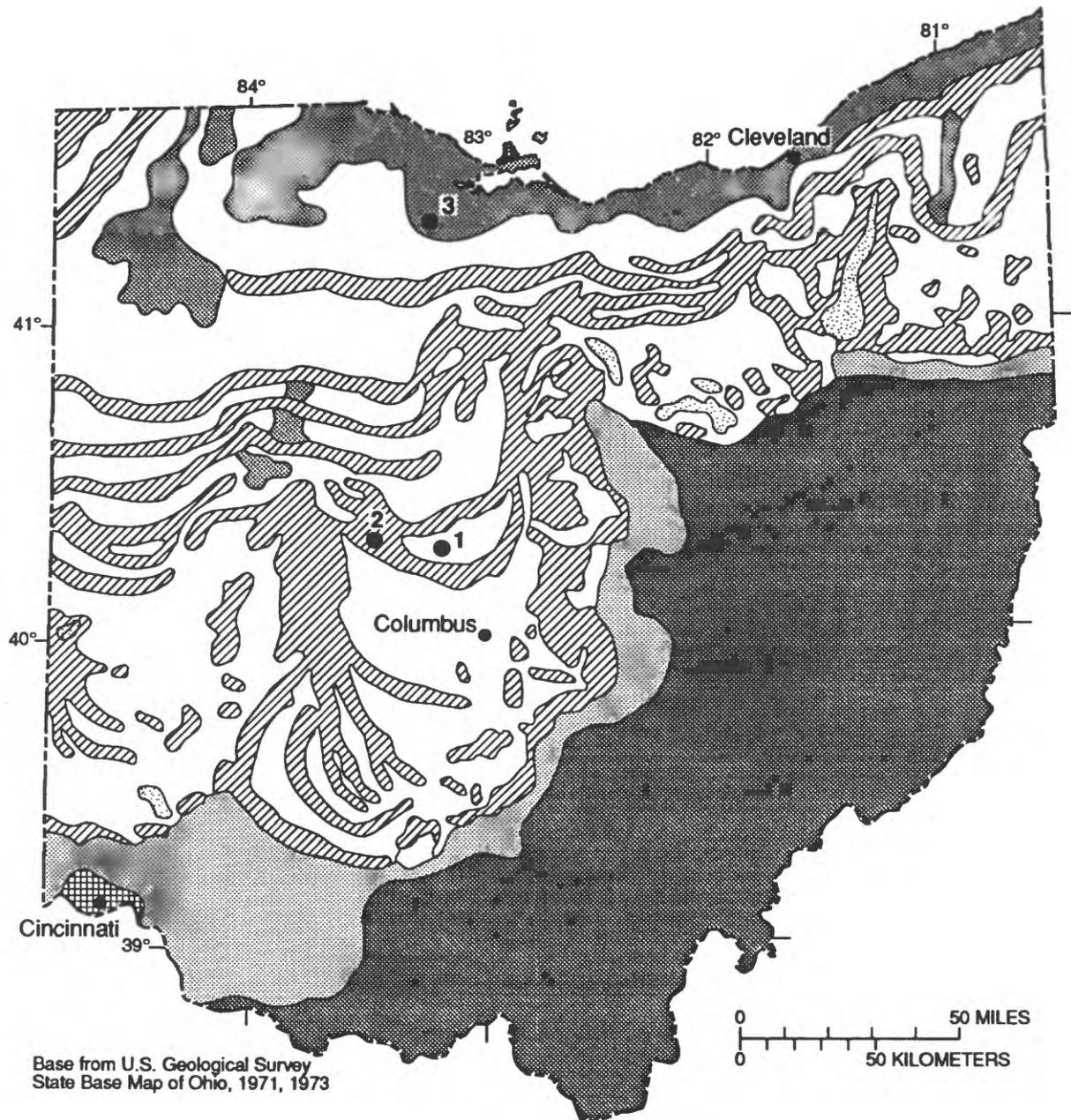
EXPLANATION

WISCONSINAN

End Moraine

Ground Moraine

Kames and Eskers

Lacustrine Deposits

ILLINOIAN

Undifferentiated

KANSAN / NEBRASKAN

Undifferentiated

Unglaciated

●¹ SITES AND IDENTIFIERS

Therefore, composition of ground moraine and end moraines in Ohio should reflect previous local surficial deposits in addition to bedrock geology. Many of the till units in Ohio are clay rich because of the low resistance of local shale and carbonate bedrock to glacial erosion. In the Late Wisconsinan deposits, however, till composition is strongly determined by the incorporation of earlier glacial and interstadial material during ice readvance. Flow paths determined from erosional features (striations and chattermarks) and from depositional features (eskers and drumlins) indicate that ice advances during the Late Wisconsinan occupied and traversed the Lake Erie Basin and other interstadial lakes. This flow of ice across lake areas probably accounts for much of the clay component in some of the tills.

Acknowledgments

Wells used in this study were drilled and installed by the Ohio Department of Natural Resources, Division of Geological Survey (the Division is the repository for the core samples mentioned in this report). The test sites were provided by the Ohio Department of Transportation (District 6), the Ohio Department of Natural Resources (Forestry Division), and the Keckley Rural Life Center. The author appreciates the assistance of the people from these organizations.

STUDY METHODS

Study sites were chosen to represent glacial deposits in three common glacial morphologies in Ohio (figs. 2 and 3). Site 1 is within the ground moraine in Union County, site 2 is within the end moraine in the western part of Union County, and site 3 is within the lacustrine deposits in Sandusky County.

Well Drilling and Installation

Drilling and well installation at the three sites was completed during February 1990. Fourteen continuous cores from depths of about 7 to 75 ft were obtained by drilling with a rotary auger rig equipped with a Shelby tube and (or) a split- spoon sampler. Holes were augered without drilling fluid or water in an attempt to prevent alteration of the samples. Cores initially were examined in the field to determine

depths for well-screen placement, and later, in greater detail in the laboratory. A split-spoon sampling tube was attached to the drill rods and placed inside the hollow-stem auger for the initial hole at each site. Simultaneous augering and sample collection in 5-ft intervals provided a continuous stratigraphic record at the site.

Wells that were placed in a fractured till were installed by augering to a depth above the till interval and then pushing a Shelby tube past the bottom of the auger into the desired material. The Shelby tubes may close fractures, but the amount of disturbance and displacement is minor (M.J. Hendry, National Water Well Association, oral commun., 1989).

Observation wells consisted of either 2-in. or 4-in. (inside diameter) threaded, flush-joint polyvinyl chloride (PVC) pipe. Slotted screens with 0.010-in. openings in 5-ft sections were attached to the bottom of each well casing, with the exception of well 1-3, for which a 2-ft screen was constructed from PVC and cut with a hacksaw at a slot interval and width approximately equivalent to 0.010 in. A small screen size was selected for all wells because of the difficulty in predicting lithologies in these heterogeneous sediments during augering.

Wells 3-1 and 3-2 at the Sandusky County site were screened in a fine-grained (0.005-in. diameter) sand that collapsed around the screens and provided a natural filter pack. Although the mean grain size of this sand was finer than the slot opening in the screen, the low pumping rate, low transmissivity of these sediments, and bridging of the grains allowed only small amounts of sediment to enter the wells during well development and pumped-well tests. The other wells at the three sites were completed by packing a washed, silica quartz filter sand (rated at 90 percent retained at 0.036-in. diameter) in the annulus of the well between the screen and the borehole wall to prevent formation collapse around the screen and to promote unrestricted ground-water flow to the well screen.

All the wells were sealed above the filter sand with 0.25-in. bentonite pellets. The holes were then backfilled with auger cuttings, and the upper 2 to 3 ft were sealed with Portland cement around steel well guards to protect the

SITE 3
Green Springs State Nursery, Sandusky Co.
Setting : Lacustrine deposits

3-4
3-3
3-2
3-1

0   5   10 FEET
0   1   2   3 METERS

N

CR 198

0   1,000   2,000 FEET
0   250   500 METERS

CR 113

Camp Quilter
CCC Camp

To Green Springs
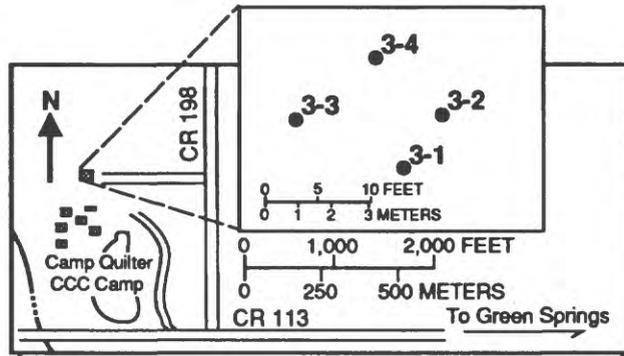
EXPLANATION

----- INTERMITTENT STREAM

3-1 ● WELL AND IDENTIFIER

2-5
2-2
2-1
2-4
2-3

0   5   10 FEET
0   1   2   3 METERS

N

Bear Swamp Road

R.R.

Benton Road

Shirk Road

0   1,000   2,000 FEET
0   250   500 METERS
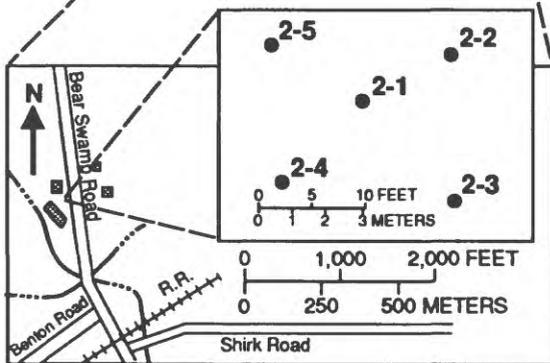
SITE 2
Keckley Rural Life Center, Union Co.
Setting : End moraine

1-5
1-4
1-2
1-1
1-3

0   5   10 FEET
0   1   2   3 METERS

N

S.R. 33

0   1,000   2,000 FEET
0   250   500 METERS
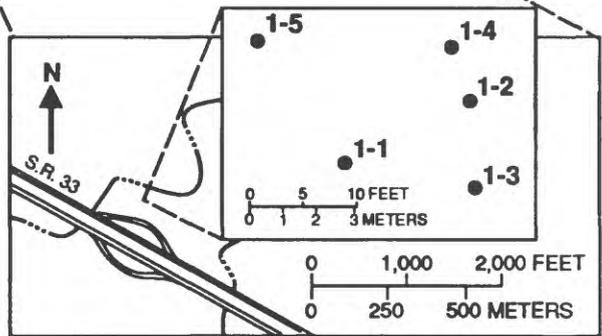
SITE 1
Rt. 33 Roadside Rest Area, Union Co.
Setting : Ground moraine

Figure 3. Locations of study sites. (Well designations and relative positions illustrated in insets.)

plastic casing and prevent surface water from flowing down the annulus (fig. 4).

Well Development and Water-Level Monitoring

After drilling and well installation was completed, the wells were developed to facilitate natural flow through the well screens and to reopen any fractures that were smeared shut during augering. The methods used to develop the wells included surging and bailing by hand in shallow wells, jetting with compressed air in deep wells screened in sediments with small hydraulic conductivities (water levels in these wells require days to recover), and pumping in deep wells screened in sediments with relatively large hydraulic conductivities (water levels in wells recovered in a few minutes).

Water-level measurements were made periodically in all wells from the time of well installation throughout the completion of the study. Measurements were made manually with a steel tape read to 0.01 ft and (or) continuously by use of a counterbalanced float system connected to an incremental encoder and coupled to a Campbell Scientific CR-10[1] data-logger. One float system was placed at each site in wells open to the lower hydrologic units.

Pumped-Well Tests

The rate of vertical ground-water movement was determined by measuring water levels in a set of nested observation wells completed either at the same level as, or above, a well being pumped from a lower hydrologic unit. A cone of depression created around the pumped well affected the water levels in the observation wells. Measurements of the rate of water-level decline in the upper wells allowed analysis of

[1] Use of trade and firm names in this report is for identification purposes only and does not constitute endorsement by the U.S. Geological Survey.
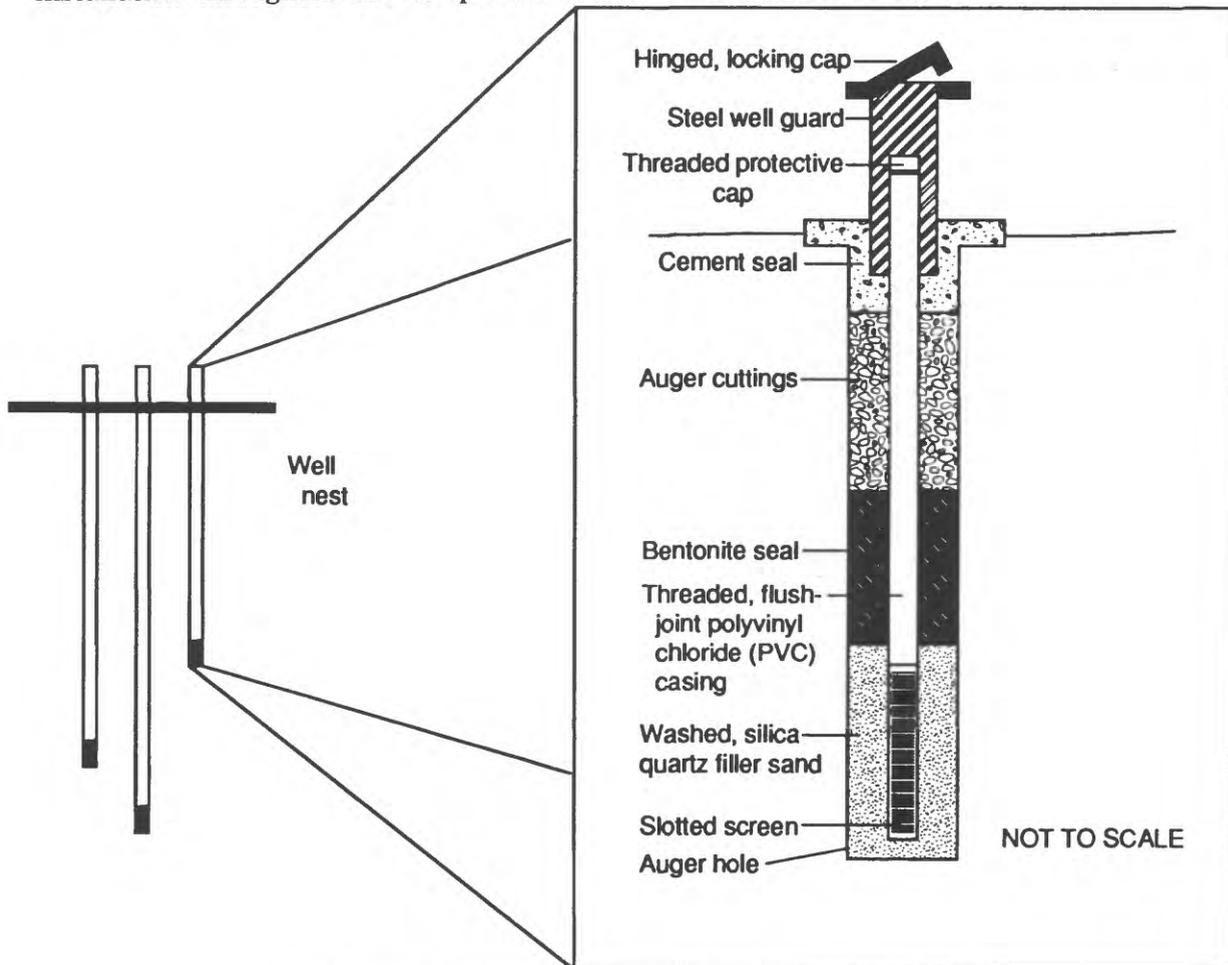
Figure 4. Diagram of typical monitoring-well construction.

the vertical flux of ground water through these units. To be used in geologic settings consisting of thick clay units with low vertical hydraulic conductivity and in hydrologic units of low transmissivity (such as the sites in this study), the pump must be designed for continuous operation over a long period (weeks), must be able to pump at very low rates (fractions of cubic feet per minute), and must have a portable power source.

A positive-displacement bladder pump was selected for several reasons; first, the bladder pump is portable and inexpensive to operate; second, the pump works by gas displacement and there are no mechanical components to fail over prolonged operation; third, silts and fine sands suspended in the water do not affect the performance of the pump; finally, the pumping rate can be adjusted easily.

Water levels during the pumped-well tests were measured at intervals of 1 minute for the first 10 minutes, then at increasing intervals with time. Pressure transducers were placed in four wells at each site to provide supplemental measurements. Water-level fluctuations due to the cyclic action of the bladder pump are reflected in the drawdown curves from the pumped wells (fig. 8).

A few hours before a pumped-well test began, water-level measurements were made in the test well and compared to continuous records from wells equipped with dataloggers to establish water-table trends. The bladder pump and transducers were placed in the wells at least 12 hours before the start of the test to allow water levels to equilibrate. Measurements of barometric pressure and air temperature were started at this time and then measured hourly.

The pumped-well tests required three to four people for the initial day of each test. The first hour of each test required approximately 40-50 water-level measurements in wells in the lower hydrologic units, where the response to pumping was rapid. In the upper wells, measurements were made less often, as the response to pumping required days to be noticed, if it was apparent at all. After the first day of each test, only one person was needed to

measure water levels, pumping rate, and atmospheric conditions.

Each pumped-well test lasted from 1 to 6 weeks, after which time pumping was stopped and recovery levels were measured. Multiple measurements of water levels were made during the first few hours of recovery when the rate of change was greatest, and measurement intervals were decreased during the following few days. When recovery rates became stable, the pump and transducers were removed, and the water levels were measured at 2- to 3-week intervals.

RESULTS OF GRAIN-SIZE DISTRIBUTION ANALYSES

One core recovered from each site during drilling was sampled for grain-size analysis. The sample intervals are indicated in figures 5, 6, and 7, and are representative of the major lithologies or significant sedimentary features in each core.

Samples weighing 150 to 300 grams were soaked in water for at least 12 hours to disperse clumps of fine-grained material and then wet sieved through 16-mm, 8.0-mm, 4.0-mm, 2.0-mm, 1.0-mm, 0.5-mm, 0.25-mm, 0.125-mm, and 0.063-mm (millimeter) sieves. The fraction of the sample that passed through all the sieves represents grain sizes less than 0.063 mm. Grain-size distribution, as a function of percent by weight, was then calculated (table 1). Grain size was plotted against weight (percent finer) and presented to illustrate sample uniformity (figs. 8, 9, and 10). The grain sizes, designated by Phi (ϕ) values for specific fractions, follow the Wentworth classification scheme (Krumbein and Pettijohn, 1938).

In the samples from site 1, representing ground moraine, the grain-size distribution differs little within the tills of the upper confining unit. The brown, weathered tills, represented by samples 1-1 to 1-7, are composed mainly of silts and clays and some sands and gravel. The gray, unweathered tills, represented by samples 1-8 and 1-9, likewise are composed mainly of silts and clays; however, the percentage of sand and gravel fractions in the unweathered till decreases. This decrease may be attributed to movement of clays and

WELLS

DESCRIPTION

1-1 1-2 1-3 1-4 1-5

Feet   Sample
0   1-1 — DISTURBED

1-2 — TILL; SILTY CLAY; LIGHT BROWN ;
        SOFT; WHITE CALICHE; FRACTURED.

5   1-3 — TILL; SILTY CLAY; PEBBLY; DARK
        BROWN; CONTAINS CARBONATE, SHALE
        AND CRYSTALLINE CLASTS; HIGHLY
1-4   FRACTURED , NO APPARENT ORIENTATION
        OF FRACTURES

1-5
10

1-6 — TILL; SILTY CLAY; PEBBLY; GRAY; HARD;
        HIGHLY FRACTURED.

— SAND; LENSES AT 12.5 AND 14 FEET ;
        INTERSECTS FRACTURES.
15
— TILL; SILTY CLAY; LIGHT BROWN AND GRAY;
1-7   FEW PEBBLES; HARD; NO FRACTURES.

20

1-8
— TILL; SILTY CLAY; GRAY; PLASTIC; WET;
        SOFT; FEW PEBBLES; NO FRACTURES.

1-9
25
1-10

— SAND, GRAVEL, SILT; MAINLY CARBONATE
        ROCKS WITH SOME SHALE; WET.

1-11
1-12
30
— TILL; SILTY CLAY; BROWN; SOME PEBBLES;
1-13   HARD; NO FRACTURES.
— BEDROCK

EXPLANATION

Disturbed soil   Sand

Till   Bedrock

Fractures   Screen

Sand, gravel, and silt ..▽.. Water level

   1-2   Well identifier

Figure 5. Stratigraphic section of glacial deposits, well depths, and sampling sequence at site 1.
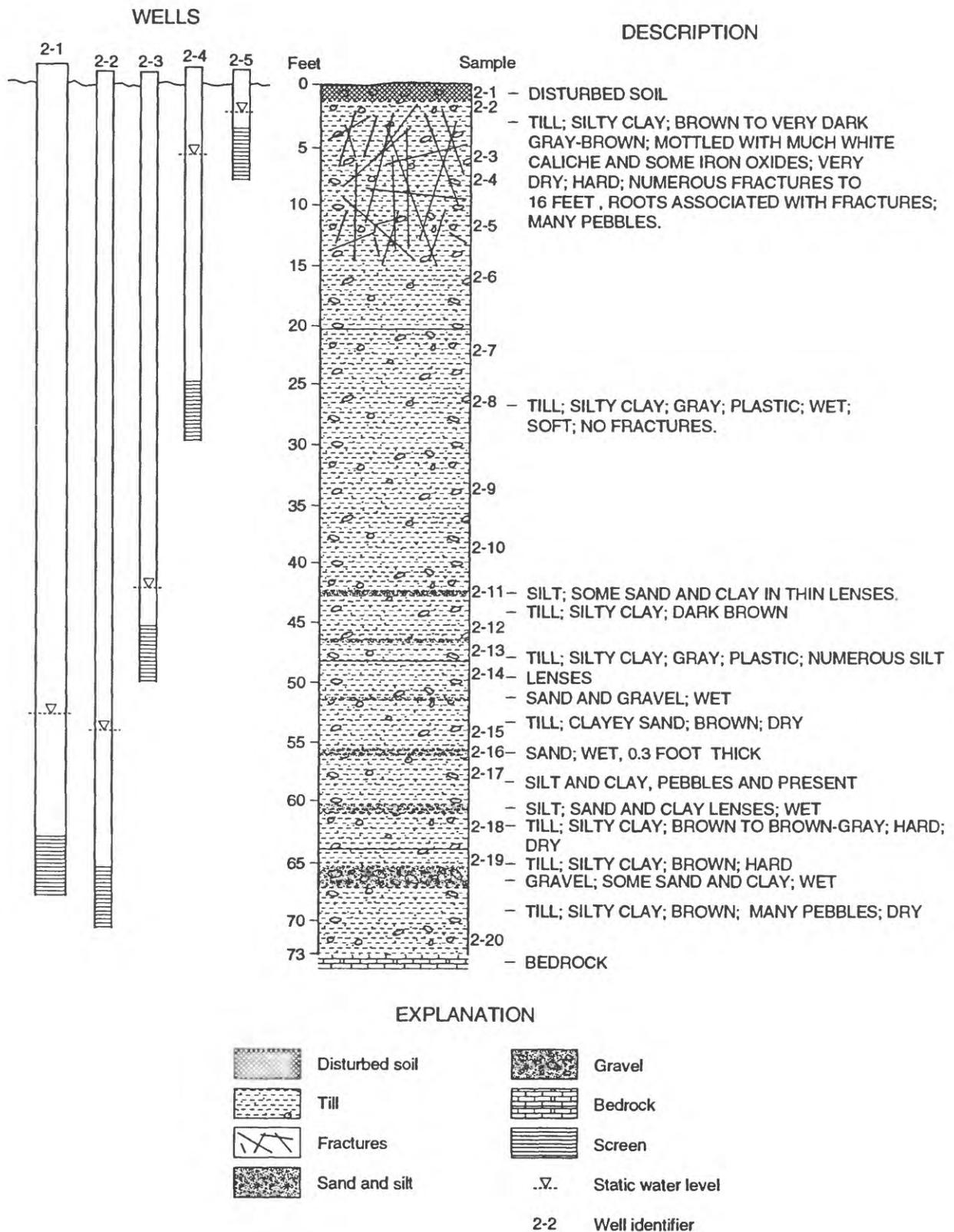
**Figure 6.** Stratigraphic section of glacial deposits, well depths, and sampling sequence at site 2.
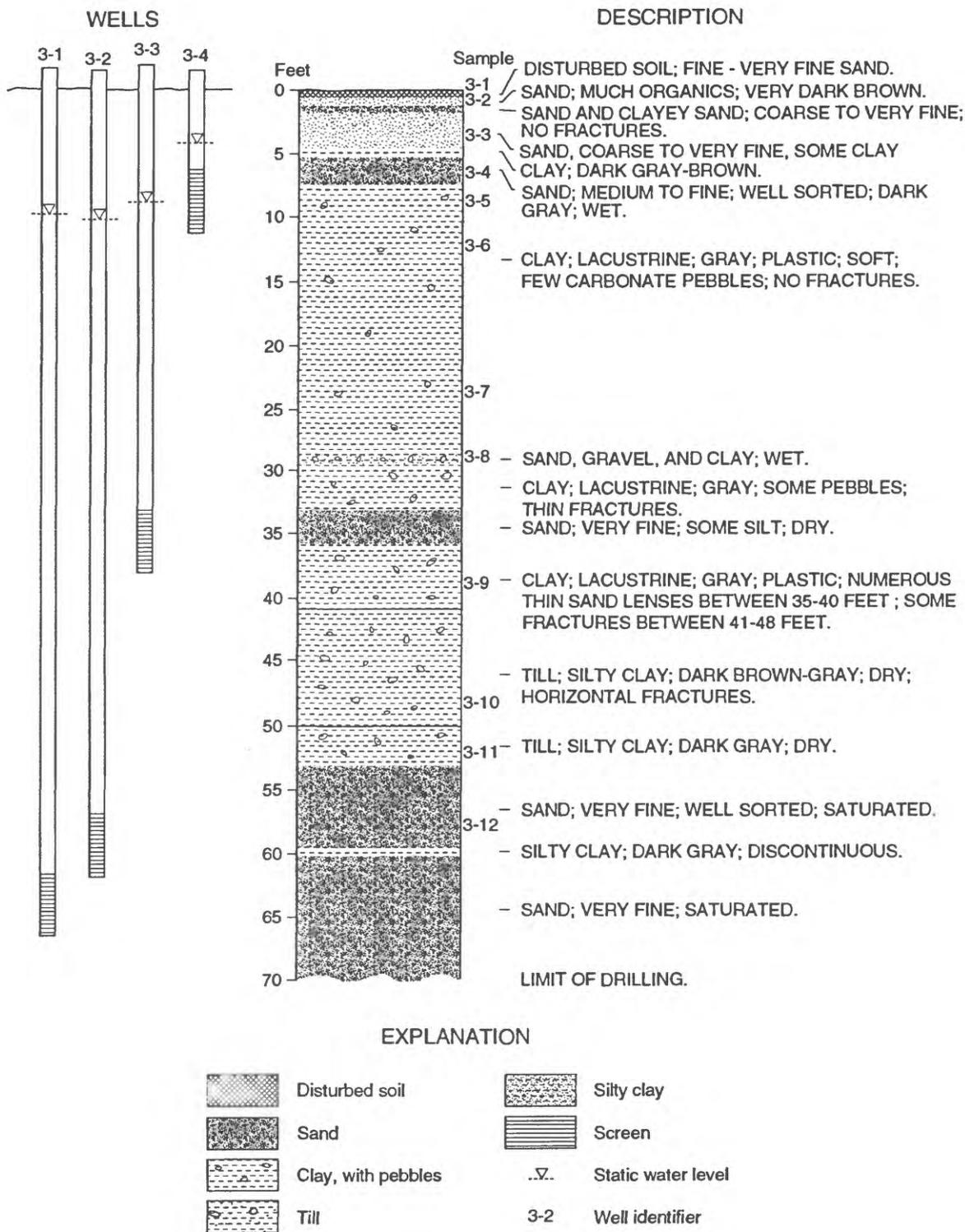
Figure 7. Stratigraphic section of glacial deposits, well depths, and sampling sequence at site 3.

Table 1. *Grain-size distribution data from the study sites*

[Grain size in percent, by weight, for each sieve size (phi (ϕ)); >, greater than]

Ground Moraine

Sample number	Grain size										
	-5	-4	-3	-2	-1	0	1	2	3	4	>4
1-1	0.00	0.00	4.39	1.11	0.46	0.92	1.57	4.00	6.36	5.11	76.07
1-2	.00	.00	.00	2.03	2.65	2.38	2.56	4.02	4.50	4.55	77.31
1-3	.00	.00	2.95	2.37	3.50	2.85	3.34	4.50	4.86	4.67	70.96
1-4	.00	.00	.00	1.98	2.54	2.73	2.87	3.86	4.05	4.61	77.36
1-5	.00	.00	3.57	2.01	2.76	2.76	2.85	3.48	4.06	4.37	74.14
1-6	.00	.00	.00	1.53	2.16	2.16	2.31	3.47	3.95	4.21	80.21
1-7	.00	.00	2.91	2.05	2.09	2.09	2.01	2.77	3.81	3.74	79.53
1-8	.00	.00	.00	1.50	.93	1.13	1.22	1.70	2.79	3.60	87.12
1-9	.00	.00	3.80	.74	1.04	.99	1.09	1.48	2.87	3.06	84.93
1-10	.00	.00	13.73	21.29	12.24	10.64	16.18	10.64	4.52	3.03	7.88
1-11	.00	.00	.00	2.74	4.70	12.60	40.34	24.93	3.85	.98	9.86
1-12	.00	.00	9.53	2.36	5.80	5.13	5.32	5.38	5.77	5.19	55.53
1-13	.00	.00	4.25	5.27	6.11	5.66	5.31	5.97	8.32	6.28	47.17

End Moraine

Sample number	Grain size										
	-5	-4	-3	-2	-1	0	1	2	3	4	>4
2-1	0.00	0.00	0.00	.81	1.00	1.25	2.21	3.87	3.56	1.44	85.96
2-2	.00	.00	.00	1.56	1.73	1.90	1.82	2.77	4.33	4.07	81.82
2-3	.00	.00	1.33	.64	1.63	1.54	2.18	3.51	5.00	2.99	81.18
2-4	.00	.00	2.17	1.53	1.62	1.80	1.80	2.48	3.70	3.16	81.73
2-5	.00	.00	.00	.00	.24	.24	.60	1.39	1.69	.91	94.93
2-6	.00	.00	.00	1.68	1.83	1.68	1.68	2.05	3.65	3.58	83.86
2-7	.00	.00	.00	1.26	1.50	1.50	1.68	2.04	2.95	2.89	86.17
2-8	.00	.00	1.27	1.59	1.69	1.97	1.78	2.30	3.94	3.38	82.08
2-9	.00	.00	3.86	1.64	1.73	1.89	2.05	2.67	3.12	4.68	78.35
2-10	.00	.00	.00	1.37	2.28	1.91	1.69	2.14	3.33	3.52	80.08
2-11	.00	.00	8.42	8.85	5.05	5.77	8.61	11.41	6.02	5.00	27.48
2-12	.00	.00	.00	2.57	2.77	2.91	3.16	4.27	5.44	4.52	66.49
2-13	.00	.00	.00	2.26	2.56	2.34	2.13	2.68	3.62	3.96	80.44
2-14	.00	.00	.00	2.10	3.20	2.60	3.25	3.99	4.24	4.39	56.67
2-15	.00	.00	7.33	7.37	6.45	6.20	5.19	6.24	7.04	6.12	42.19
2-16	.00	.00	6.45	2.90	2.54	2.60	2.78	2.90	2.90	3.73	21.76
2-17	.00	.00	4.42	4.19	4.49	4.35	4.66	4.72	8.22	7.39	40.84
2-18	.00	.00	3.53	6.65	5.70	7.14	7.60	6.65	5.70	9.49	47.55
2-19	.00	.00	10.29	8.80	7.44	7.36	5.70	5.53	5.36	4.68	44.86
2-20	.00	.00	.00	5.65	6.54	6.77	8.05	6.04	5.53	6.23	55.19

Table 1. *Grain-size distribution data from the study sites*--Continued

Lacustrine Deposits

Sample number	Grain size										
	-5	-4	-3	-2	-1	0	1	2	3	4	>4
3-1	0.00	0.00	1.11	0.61	1.27	2.87	6.36	24.54	38.20	10.67	14.37
3-2	.00	.00	.00	.51	1.27	2.04	6.04	42.75	20.42	4.33	22.65
3-3	.00	.00	.45	1.82	6.78	10.15	12.06	18.65	19.93	4.09	26.07
3-4	.00	.00	.91	1.78	5.28	13.10	27.37	31.74	10.36	3.25	6.20
3-5	.00	.00	.00	.62	2.20	2.78	4.57	7.32	10.56	5.82	66.13
3-6	.00	.00	11.25	1.72	2.05	2.55	3.05	4.38	5.65	3.99	65.37
3-7	.00	.00	3.32	1.80	1.77	2.77	3.46	6.16	5.82	3.57	71.33
3-8	.00	.00	.57	4.11	7.70	8.16	6.56	6.79	5.82	4.45	55.85
3-9	.00	.00	.00	.76	.38	.13	.19	.44	.70	1.01	96.39
3-10	.00	.00	.00	.29	.46	.69	.87	1.04	1.67	1.50	93.48
3-11	.00	.00	.00	.68	.42	.42	.53	1.32	6.90	13.59	76.13
3-12	.00	.00	.00	.40	.17	.17	.29	6.14	54.54	18.43	19.86

fine-grained sediments in the upper, weathered tills to the lower, unweathered tills by water flow along fractures. The eluviation of the clay-size fraction from the weathered tills is represented by an increase in the coarser sediments in a grain-size distribution curve. Sample 1-10 is poorly sorted, ranging from a medium gravel to silt and clay. Sixty-five percent of sample 1-11 is a well-sorted, coarse-to-medium sand; the remainder of the sample is distributed between very coarse sand to silts and clays. Samples 1-12 and 1-13 are both silt, clay, and minor amounts of poorly sorted gravels and sands.

The upper sections of the core from site 2, representing end moraine, have grain-size distributions similar to those of the tills at site 1. The entire core is primarily silt and clay and some sand and gravel, except for sample 2-11, which is poorly sorted medium gravel to silt and clay, and sample 2-16, a poorly sorted coarse gravel. In general, the percentage of sand and gravel in the till increases with depth from 43 percent in sample 2-14 to 55 percent in sample 2-19. All samples from 2-14 to 2-19 contain larger percentages of gravel than sample 2-20, the sample representing the lower unit confining the gravel unit that was pumped.

The samples from site 3 represent lacustrine deposits and are much more diverse in grain-size distribution than samples from the other two sites. Samples 3-1 and 3-2 are both 63 percent well-sorted, medium-to-fine sand (sample 3-2 is the coarser). Sample 3-3 is poorly sorted, ranging from granules to silt and clay. Sample 3-4 is well-sorted, coarse-to-medium sand. Lacustrine deposits of silt and clay are found in samples 3-5 to 3-9. Some coarser materials are found in samples 3-5 to 3-8 but are almost nonexistent in 3-9. Sample 3-10 was mapped as till, but its grain-size distribution resembles that of sample 3-9 (lacustrine deposit). Sample 3-11, also a till, is mainly silt and clay that contains some fine to very fine sand. The aquifer, represented by sample 3-12, is a well-sorted, fine to very fine sand.

Estimates of hydraulic conductivity (K) based on grain-size distribution (Norris and Fidler, 1965; Summers and Weber, 1984) indicate a significant relation between K and grain size. These estimates, however, are based on the assumption that the permeability of a sediment is controlled by intergranular flow. In clay-dominated sediments such as tills and lacustrine deposits, secondary porosity attributed to fractures can be the dominant effect on ground-water flow. Because of this factor and because of the inconsistency of glacial

Figure 8. Grain-size distribution at site 1.

deposits, including thickness, texture, and composition both vertically and laterally, quantitative analysis of the relation between grain-size distribution and hydraulic conductivity is difficult.

Bradbury and Muldoon (1990) found laboratory measurements of K to be significantly lower than those measured in field tests and concluded that laboratory measurements of K are of little use. Driscoll (1986) warns that no

Figure 9. Grain-size distribution at site 2.

accurate method of calculating K by grain-size distribution is yet available. For the purposes of this study, grain-size analysis simply offers a means of observing lithologic changes in the geologic column and illustrates the similarities and differences of tills at a particular site.

HYDRAULIC PROPERTIES OF SELECTED GLACIAL AND LACUSTRINE DEPOSITS

The hydraulic properties of the study sites were determined by evaluating the relation of

GRAIN-SIZE ANALYSIS
PERCENT FINER VS GRAIN SIZE
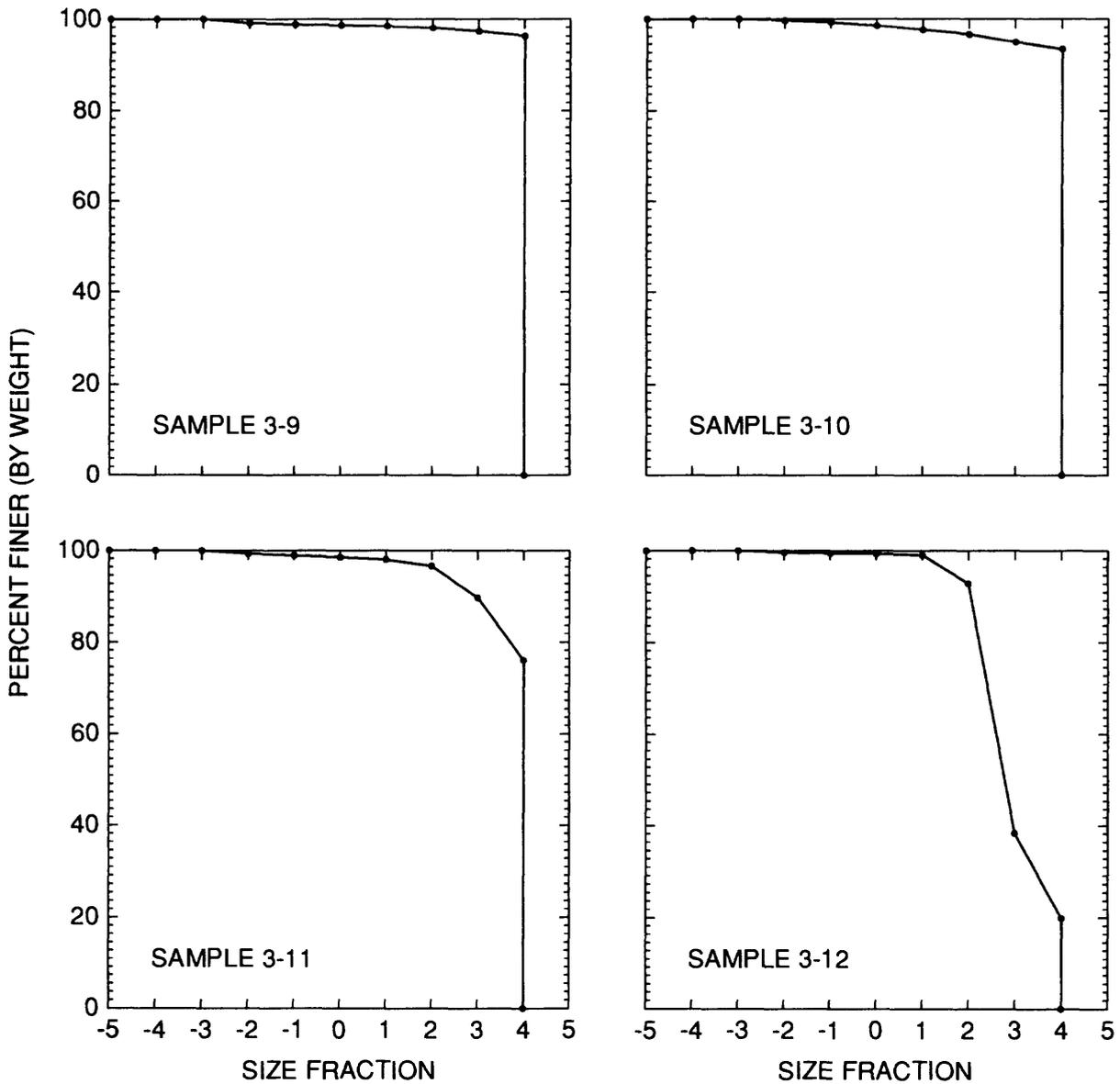
Figure 10. Grain-size distribution at site 3.

geologic setting and textural variations with the results of pumped-well tests. The effects of fractures and atmospheric conditions on ground-water flow were identified.

Results of Pumped-Well Tests

Pumped-well tests were conducted at three sites to determine the hydraulic properties of the glacial and (or) lacustrine deposits at each site. Each site represented a different type of surficial deposit found throughout western Ohio, and included ground moraine, end moraine, and lacustrine deposits.

Site 1--Ground Moraine

Union County is underlain almost entirely by the Bass Islands Group of Upper Silurian age, except in the southeastern part of the county, which is underlain by Columbus Limestone of Middle Devonian age (fig. 11). These rocks are part of the regional carbonate bedrock aquifer in Ohio (Bugliosi, 1990). Site 1 is underlain by the Bass Islands Group.

Surficial deposits in Union County range in thickness from 10 to 255 ft (fig. 12) and consist of ground moraine, end moraine, and buried-valley deposits. The Powell and Broadway Moraines cross the county in a northwest-southeast trend, and the St. Johns Moraine parallels the northern border. Paris (1985) describes the tills south of the Powell Moraine in Union County as being correlated lithostratigraphically to the Darby Till, a sandy till with a mean sand-to-clay ratio of 1.54. The till south of the Powell Moraine was deposited by the Scioto sublobe of the Huron-Erie Lobe (Fullerton, 1986). Erosion of carbonate bedrock and glaciofluvial deposits by the advancing ice incorporated sediments into the basal till being transported, resulting in the deposition of a sandy or silty till.

The readvance of the Ontario-Erie Lobe after the Erie Interstade followed a flow path through the Lake Erie Basin and southwestward into Ohio (Strobel and Faure, 1987). Lacustrine deposits from the basin and other smaller Wisconsinan lakes in the flow path produced clay-rich tills transported by the ice. In addition, the flow paths of this advance crossed the soft Ohio and Olentangy shales, which probably further increased the clay content in these tills. Tills on and north of the Powell Moraine in Union County all are clay rich in composition, with a mean sand-to-clay ratio of 0.66 (Paris, 1985). This difference in glacial flow paths explains the different glacial lithologies north and south of the Powell Moraine in Union County.

Site 1 is north of the Powell Moraine and represents ground-moraine deposits. The glacial deposits are 30 to 35 ft thick in this area. The core log, well-screen placement, and sample depths are illustrated in figure 5.

The pumped-well test at site 1 was started on August 1, 1990. The pump was shut off after 1 week, and recovery measurements were made for 2 weeks. The results of the test are plotted in figure 13.

The bladder pump used for the test pumped water in pulses because of bladder inflation-deflation cycles; a cycle generally took about 15 seconds to complete. Throughout the test, the pump maintained a constant discharge of 46 ft^3/d; however, the curve plotted for the pumped well (well 1-4) shows large fluctuations in water levels because water-level measurements were based on set time intervals regardless of the position in the pumping cycle. Linear regression of the measurements after 10 minutes provides a nearly horizontal best-fit line (slope = -8.5 x 10^{-7} ft/min) for the pumped well (fig. 14). The flat, best-fit line indicates that steady-state conditions were obtained after about 10 minutes of pumping. The large scatter of the data points resulting from the cyclic action of the pump, however, produces a poor correlation coefficient. Nonetheless, the line graphically illustrates the average drawdown in the pumped well. The small number of data points compared to fluctuations resulting from the pumping cycle and the effects of well-bore storage make the fitting of a curve to these data difficult. For these reasons, measured water levels were included on the figure only to demonstrate the general trend of the drawdown.

The results of this test do not provide adequate data for quantitative analysis of the hydraulic properties of the aquifer units. The curves in figure 14 indicate that reaction of the water levels to the pumping stress is noticeable first in the shallow wells (wells 1-3 and 1-5), then in the well in the till that confines the pumped gravel layer from above (well 1-2), and last in the well in the gravel at the same depth as the pumped well (well 1-1) (see fig. 5).

Evapotranspiration may account for much of the water loss in weathered tills. For example, results of studies in South Dakota indicate that very little recharge passes through the weathered tills into the underlying bedrock aquifers because the annual evapotranspiration rate is nearly equal to the annual average precipitation rate (Barari and Hedges, 1985; Cravens and Ruedisilli, 1987). Hendry (1988)
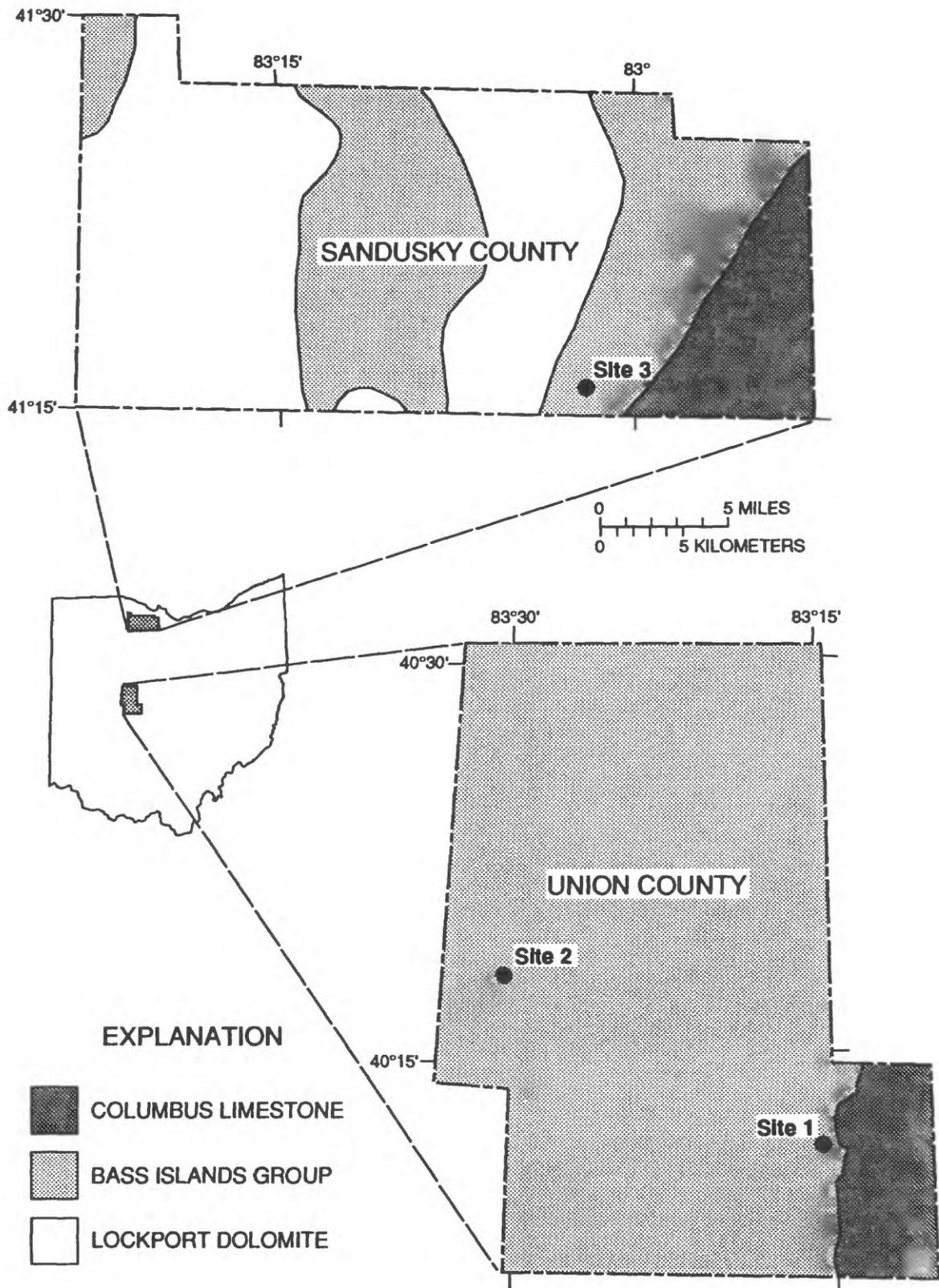
Figure 11. Bedrock geology of Union and Sandusky Counties, Ohio (modified from Bownocker, 1920).
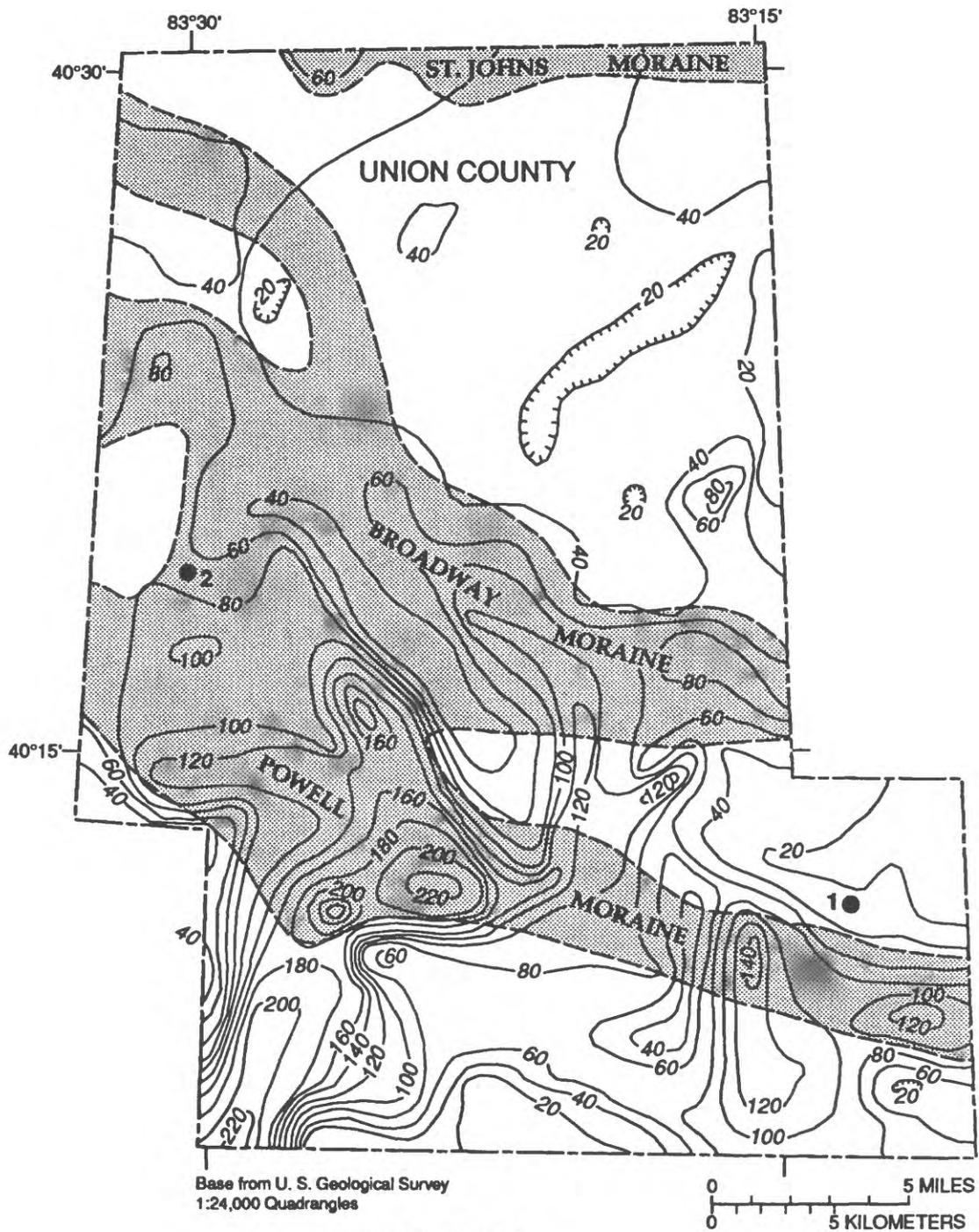
Figure 12. Thickness of surficial deposits in Union County, Ohio, with outlines of major end moraines and locations of study sites. (Data from Schmidt, 1978; Goldthwait and others, 1979; well logs from the Ohio Department of Natural Resources.)
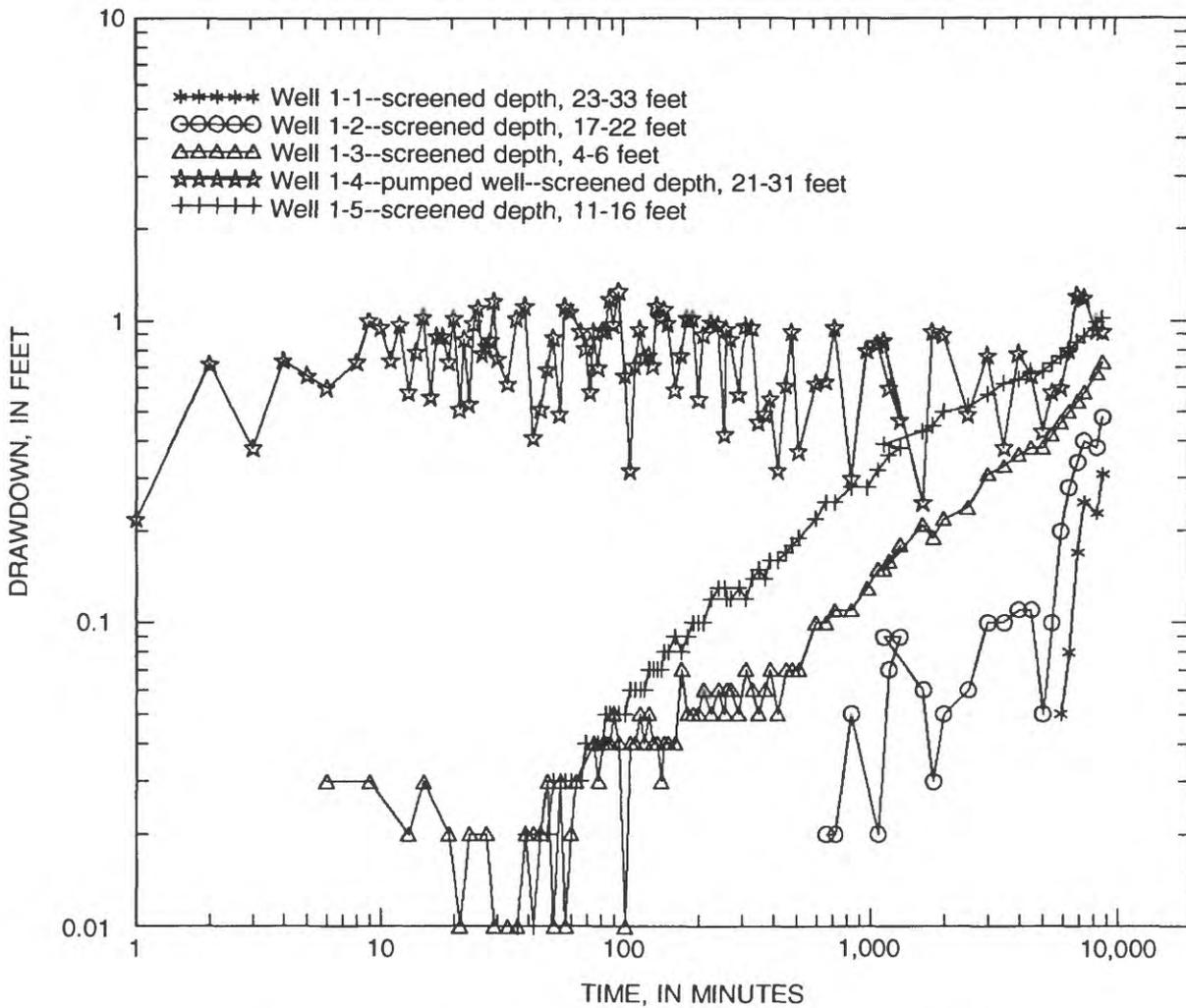
Figure 13. Drawdown curves from pumped-well test at site 1, August 1-7, 1990.

found that potential evapotranspiration exceeded annual precipitation in southern Alberta, Canada, resulting in slightly upward hydraulic gradients in the weathered tills during the summer months. He suggests that fractures in the weathered tills increase the vertical hydraulic conductivity.

Wells 1-3 and 1-5 are screened in the weathered and fractured till in the upper 14 ft of ground moraine. Measurements made from February through October 1990 indicate that declines in water levels can be attributed to seasonal trends (fig. 15). The sharp decline in water levels during the pumped-well test, however, indicates that these wells are affected by lowering the water levels in the gravel below (figs. 15 and 16). Because it is difficult to separate the effects of evapotranspiration from

the effects of drawdown related to pumping and because measurements of evapotranspiration were not recorded during the pumped-well test, detailed analysis of the drawdown data for the shallow wells is not conclusive.

The two deep wells (wells 1-1 and 1-2) demonstrate similar lag in response to the pumping stress (fig. 14). Well 1-1 is screened in the pumped hydrologic unit and, theoretically, should react to pumping before the wells screened in the overlying, less conductive tills. Well 1-1, however, reacts after drawdown is observed in well 1-2. This delayed response could be attributed to incomplete well development in well 1-1 and probably does not reflect the hydraulic characteristics of the hydrologic unit. Inadequate well development that results in clay clogging of screens can also be an
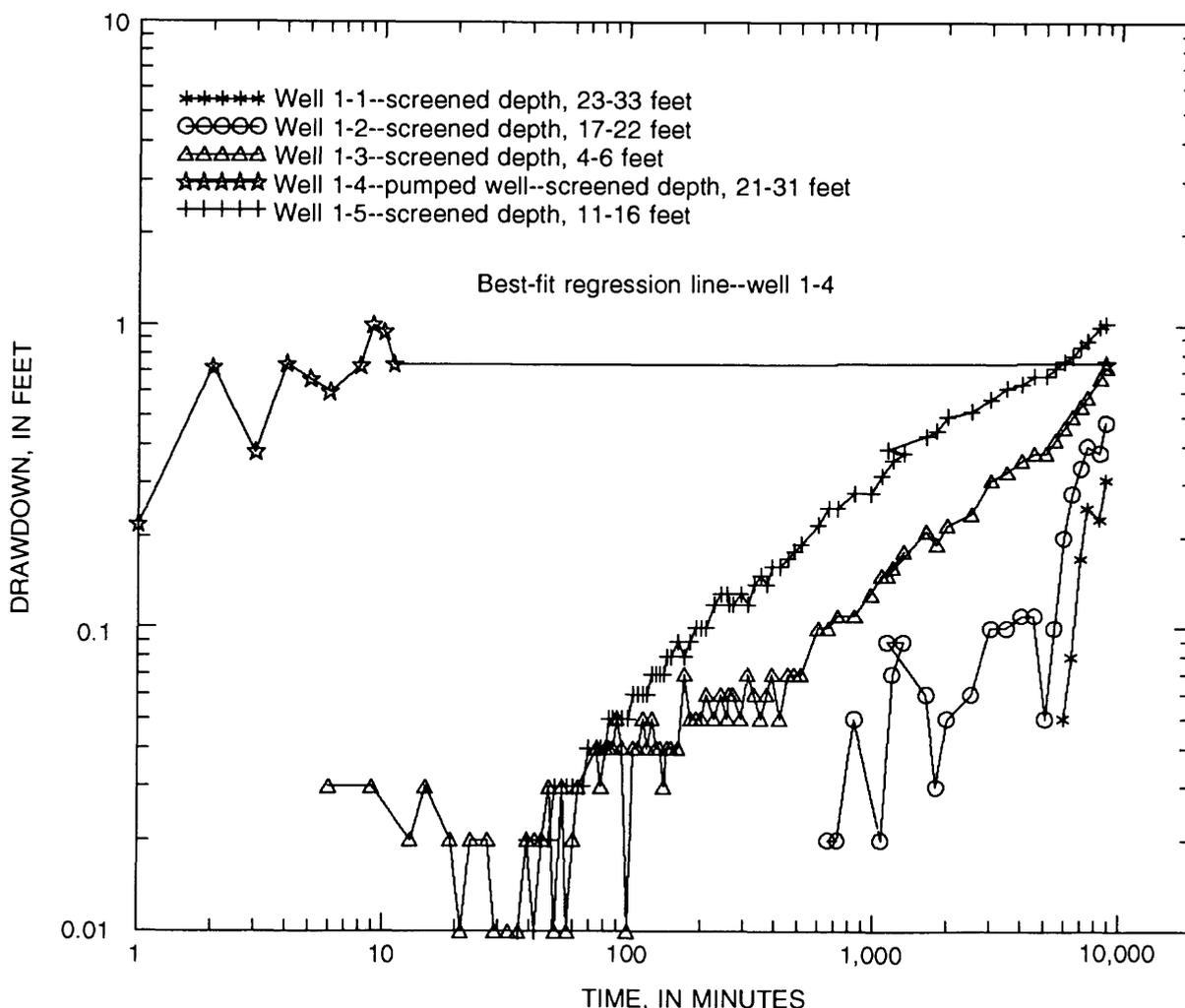
Figure 14. Drawdown curves from pumped-well test at site 1 and best fit for well 1-4, August 1-7, 1990.

important factor in analysis of pumped-well tests. Moench (1985) describes well-screen or formation plugging as a thin "skin" of low permeability that can strongly affect the drawdown response of a well. The lag in response, however, could also be the result of a poor hydraulic connection with the hydrologic unit being pumped from well 1-4 even though wells 1-1 and 1-4 are screened at equal depths.

Well 1-2 is screened partly in the brown, weathered till and partly in the gray, unweathered till. The bottom of the screen is about 3 ft above the top of a gravel unit. As indicated in figure 14, drawdown is first noticeable after about 600 minutes (0.4 day) after the start of the pumping and increases during the course of the test. Water levels are erratic until about 5,000 minutes (3.5 days)

after the start of the pumping, when the drawdown becomes steady. Well 1-1 reflects this steady decrease, but it is delayed by 1 day.

The large fluctuations in the drawdown in well 1-2 during the first 5,000 minutes of the test is probably the result of diurnal variations in evapotranspiration affecting the water levels. Because the screen in well 1-2 is partly within the fractured weathered till, there could be a hydraulic connection to the land surface that results in changes in water levels with changes in evapotranspiration and atmospheric pressure. Well 1-1, which is screened below the fractured till, does not exhibit similar water-level fluctuations to well 1-2 during the first 5,000 minutes of the test. The water levels in both wells decline steadily after 5,000 minutes of the test, indicating that
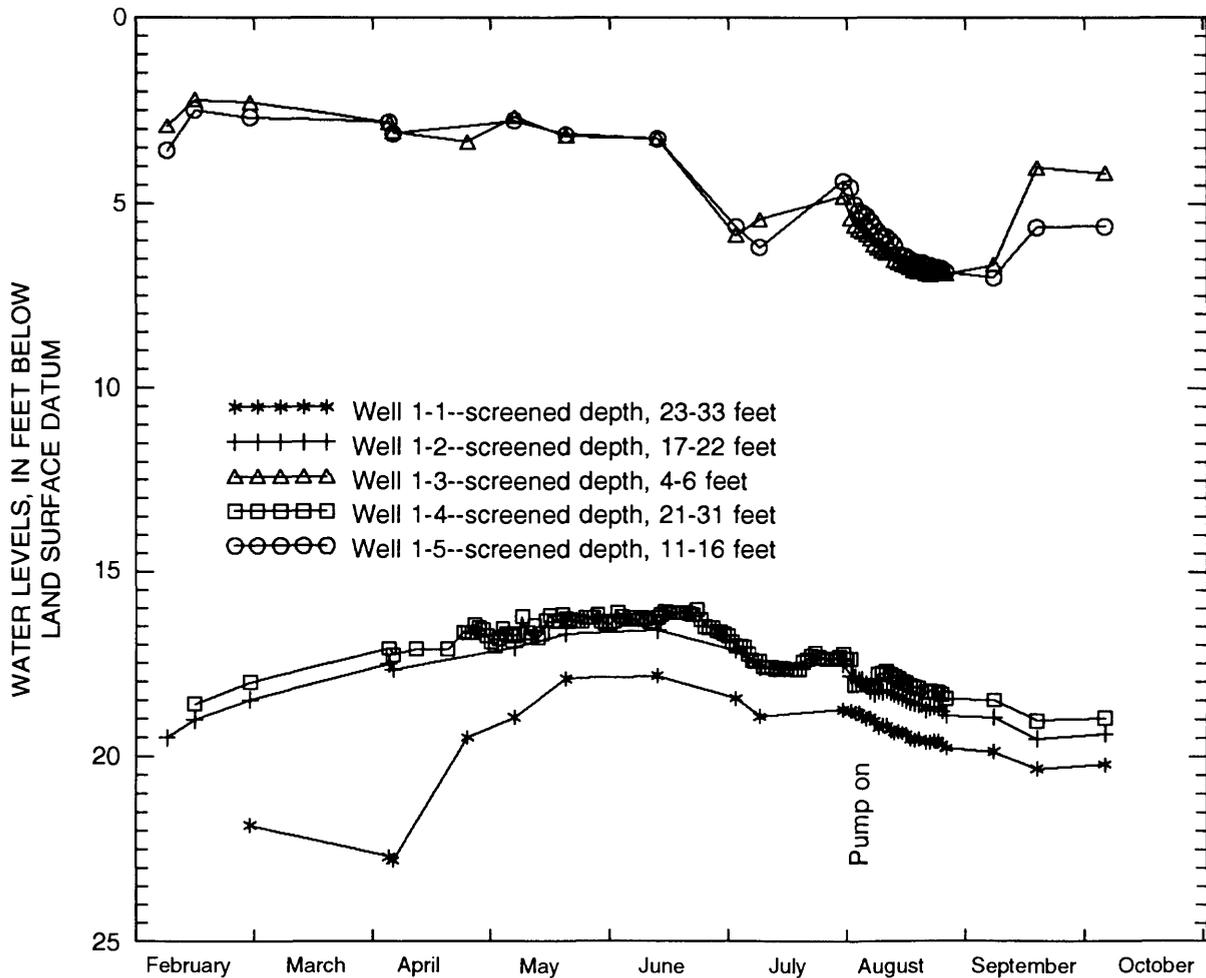
Figure 15. Water-level trends in wells at site 1, February 9-October 5, 1990.

drawdown resulting from the pumped-well test has a dominant effect following that time.

Recovery measurements provided additional information for interpretation of the hydrogeologic setting at site 1 (fig. 16). Water levels in well 1-4 (pumped well) remained constant when the pump was shut off, indicating no water-level recovery in the gravel. Water levels in wells 1-1 and 1-2 continued to drop during recovery, probably in delayed response to the pumping stress, but the rate of water-level decline decreased until levels became constant after 14 days. This period of constant levels was followed by a steady decrease in water levels in all three of the deep wells, continuing the trend indicated before the pumping stress. Wells 1-3 and 1-5 continued a similar decrease as before

the pumping stress, equilibrating after about 1 month. Evapotranspiration is probably the major cause of these water-level declines.

Similar trends in water levels in all five wells indicate a hydraulic connection between the weathered and unweathered tills and the gravel. Drawdown in the gravel during pumping, followed by only minor recovery, indicates that the gravel has limited lateral extent and that pumping from the gravel effectively reduced the quantity of water in storage. The rate of recharge to the gravel through the overlying tills depends on the amounts of evapotranspiration and precipitation and is controlled by the vertical hydraulic gradient across the till.
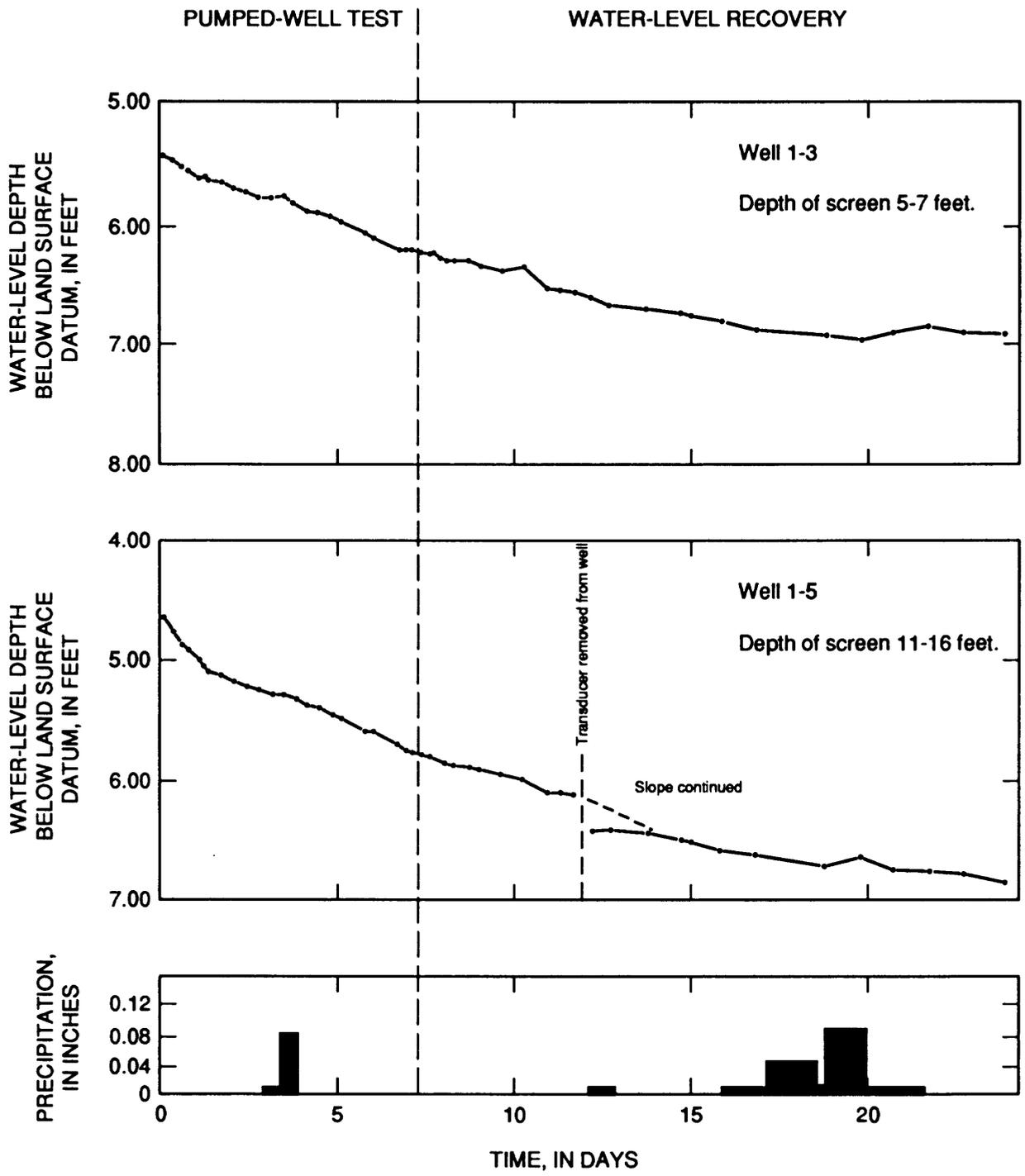
Figure 16. Water-level trends in wells 1-3 and 1-5 and precipitation at site 1 during the pumped-well test and recovery measurements, August 1-22, 1990.

Site 2--End Moraine

The bedrock geology at site 2 is the same as described for site 1. Site 2 is underlain by the Bass Islands Group (fig. 11).

Site 2 lies on the Powell Moraine in the western part of Union County, where the glacial deposits are about 73 ft thick (fig. 12). The Powell Moraine at site 2 is composed of a silty-clay till underlying a clayey till. The core log and well-screen and sample depths are illustrated in figure 6.

The pumped-well test at site 2, at the Keckley Rural Life Center, lasted from September 17 through October 26; recovery measurements were made through October 29. Discharge from the pump remained constant throughout the test at 27.0 ft³/d.

Wells 2-1 (pumped well) and 2-2, which are both screened the full thickness of the lower gravel unit (fig. 6), exhibited drawdown in response to pumping (fig. 17). Water levels in the three wells screened in the overlying tills changed during the test; however, the data indicate that these changes were independent of the pumping stress.

Trends in water levels for all wells from April 4 through October 30, 1990, are shown in figure 18. These water-level trends indicate that the steady decline in well 2-3 during the
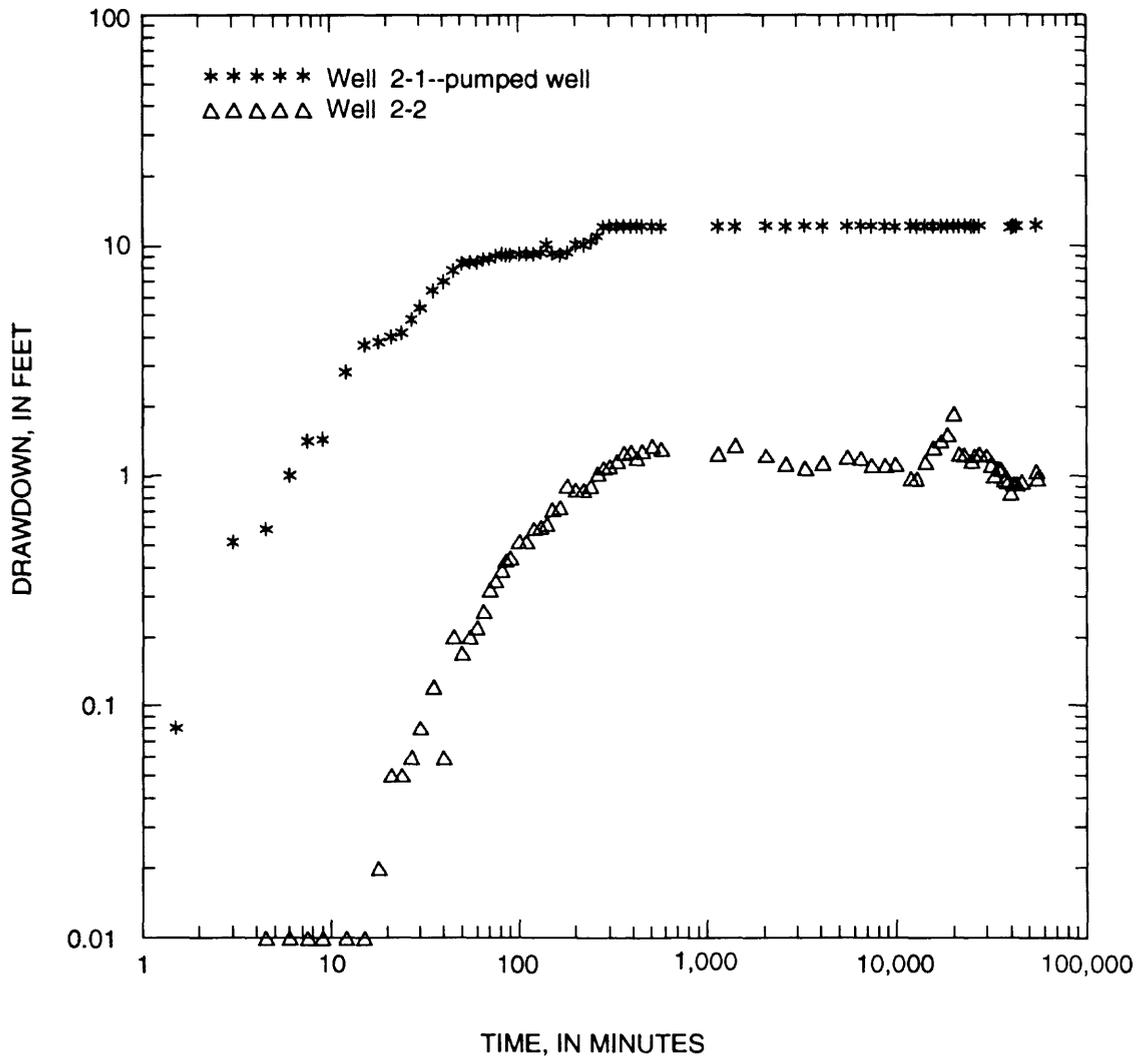
Figure 17. Drawdown curves for wells 2-1 and 2-2 during the pumped-well test at site 2, September 17-October 26, 1990.
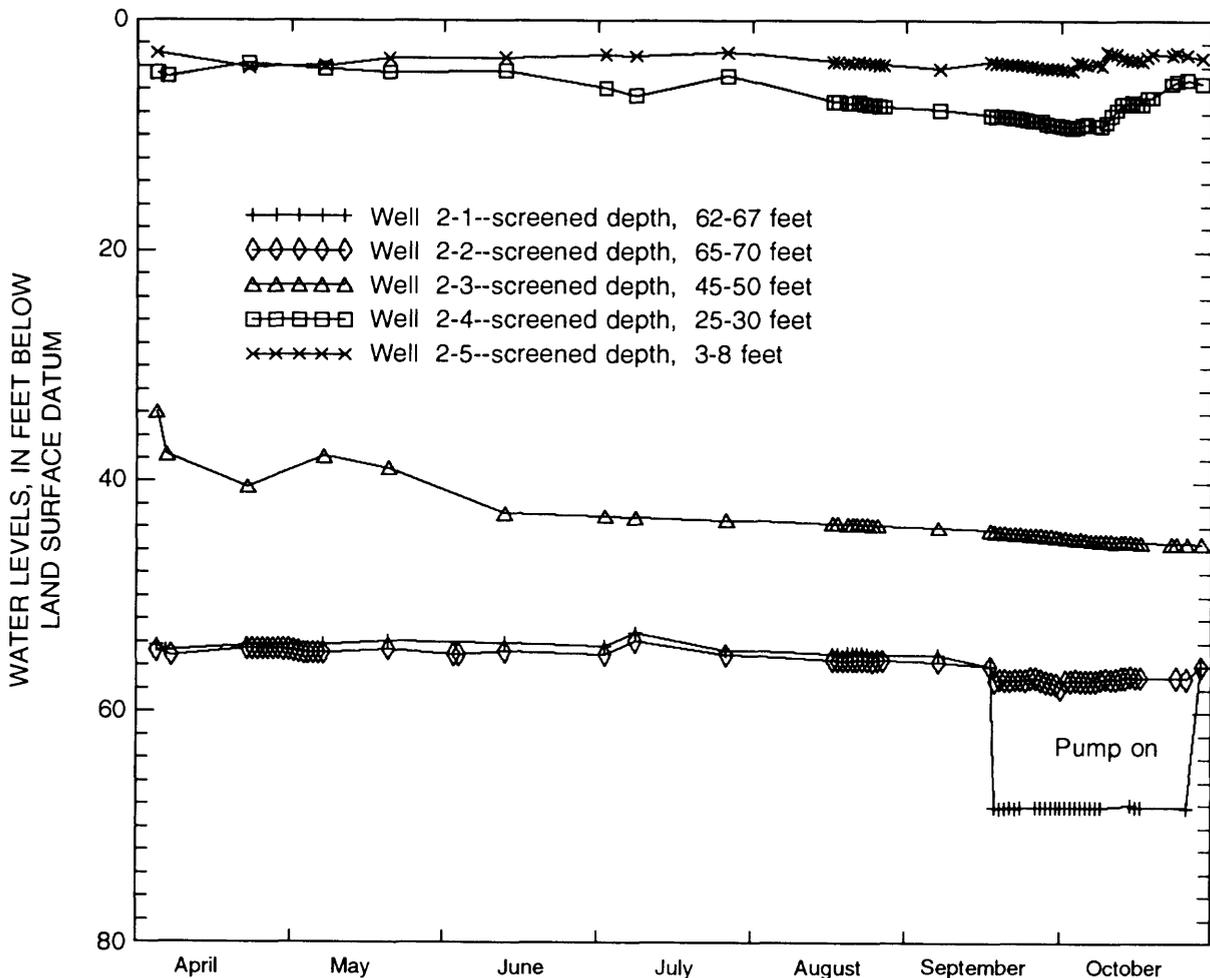
Figure 18. Water-level trends in wells at site 2, April 5-October 29, 1990.

pumped-well test was not an effect of pumping but, rather, was part of a seasonal decline in water levels. Wells 2-4 and 2-5 are shallow, and water levels in these wells indicate no correlation with pumping stress; thus, variations in water levels in these wells are probably the result of atmospheric influences.

Closer examination of the water levels measured during the pumped-well test in the shallow wells (2-4 and 2-5, screened from 25 to 30 ft and 3 to 8 ft, respectively) (fig. 6) and precipitation data indicate that the change in water levels depends on time and on depth (fig. 19). Measurable precipitation during the pumped-well test resulted in an initial water-level rise in well 2-5, followed by a delayed rise in well 2-4. Declines in water levels in these two wells in response to pumping may result from downward leakage of water through the lower tills, but the effect of

evapotranspiration, observed at site 1, needs to be considered in wells 2-4 and 2-5.

The response in well 2-5 to precipitation indicates how fractured till affects the vertical movement of ground water. The water level in well 2-5 (screened in the highly fractured till in the upper 15 ft (fig. 6) and constitutes an unconfined aquifer), responds instantaneously to precipitation then declines steadily as water moves downward through the underlying till or out of the ground-water system through evapotranspiration.

The response to precipitation in well 2-4 (screened within the gray, unfractured till) indicates that water flows into this unit; however, the small amount of time required for the water-level rise in relation to precipitation indicates purely advective ground-water flow is improbable in this situation. The delayed rise in

water levels in well 2-4 may be attributed to fractures extending as deep as 25 to 30 ft in the till but undetected during core analysis. The fractures would provide conduits for vertical ground-water flow and result in response in water levels to precipitation. In addition, the water-level rise in well 2-4 could partly result from increases in formation pressure caused by trapped air below an advancing wetting front caused by precipitation.

Fractures were not apparent in the gray, unweathered till in cores from site 2, as is typical for most locations in Wisconsinan till cited in the literature. Field mapping completed in Union County as part of this study, however, showed that the gray till is fractured near the unoxidized/oxidized contact at nearby stream-cut exposures. The fractures are generally random, and highly fractured and unfractured zones are found within the same till unit.

Although fractures were not observed in cores containing the gray till at site 2, the rise in water levels in response to precipitation in well 2-4 indicates that some extent of fracture permeability may affect vertical ground-water flow throughout at least a part of the unweathered till in this area. Keller and others (1988) observed fracture permeability in unweathered tills in laboratory and field tests in a similar glacial setting in Saskatchewan, Canada. Their study, as well as field observations in Union County for this study, indicate that fractures are present in the unweathered till but may not always be evident in cores.

Fractures in the unweathered till at site 2 may be subject to increased formation pressure with depth, which may result in decreases in the fracture apertures and, in turn, decreases in the hydraulic conductivity of the tills with depth. This decrease in fracture aperture, or secondary porosity, may explain the delayed and larger changes in water levels in well 2-4 after precipitation compared to water levels in well 2-5 for the same periods.

Effects from pressure increases due to trapping of air below an advancing wetting front also may play a role in explaining the water-level trends in wells 2-4 and 2-5. Todd (1980, p. 239-240) expresses the relation between a rise in water level and infiltration from precipitation by

$$\Delta h = \frac{m}{H-m}(10)m, \qquad (1)$$

where
m is the thickness of the saturated zone resulting from infiltrating precipitation, in feet;

H is the thickness of the zone containing interconnected air-filled pores (zone of aeration), in feet; and

Δh is the change in water level, in feet.

According to this equation, the rise attributed to increased pressure can be large. For a shallow water table, the rise in water levels can be an order of magnitude larger than the depth of the infiltrating precipitation because of trapped air (Todd, 1980). This effect dissipates rapidly, however, as air escapes upward through the wetting front.

The minor changes in water levels in wells 2-4 and 2-5 with respect to precipitation during the early part of the pumped-well test (fig. 19) may indicate a moisture content in the tills of less than or equal to the specific retention. As a result, most of the water infiltrating the tills was held initially by capillary tension. The moisture content exceeded the specific retention after about 21 days of the pumping stress because of the effect of precipitation (fig. 19), which resulted in water-level responses to even minor amounts of precipitation after this period. The slow, steady increase in the water level in well 2-4 after 21 days (fig. 19) may indicate the advance of a wetting front downward through the till; however, the expected rapid dissipation of this effect described by Todd (1980) is not apparent. Whether the rise in water levels in well 2-4 is due to decreases in porosity with depth as fracture apertures decrease is uncertain.

Water levels in well 2-3 decrease seasonally (fig. 18). Multiple silt, sand, and gravel horizons intersect the screened interval (45-50 ft) and are probably hydraulically connected to a lower gravel unit (65-67 ft) indicated by the rate of
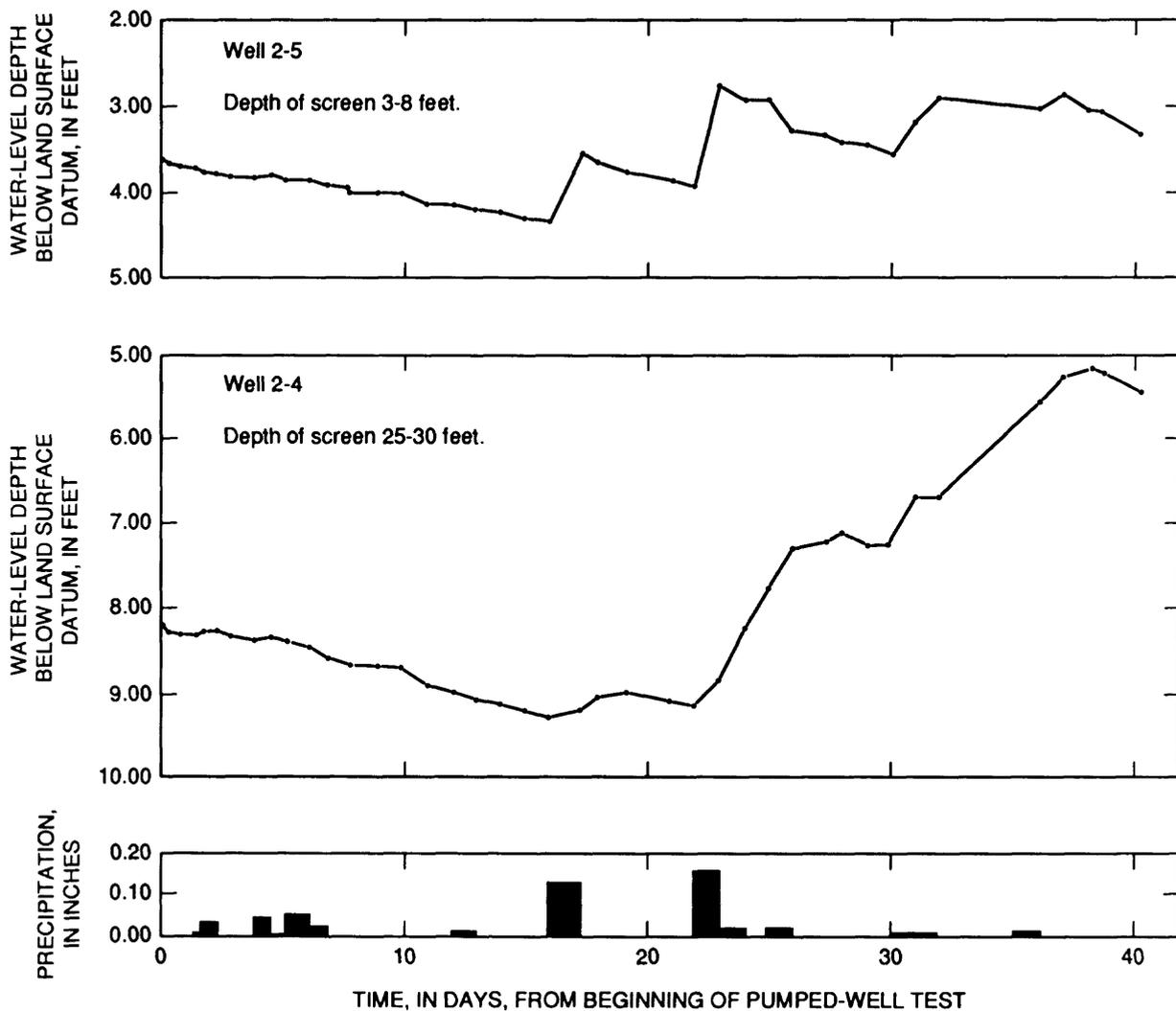
Figure 19. Water-level trends in wells 2-4 and 2-5 and precipitation during the pumped-well test and recovery measurements, September 17-October 29, 1990.

water-level decline in well 2-3 compared to the declines observed in wells 2-1 and 2-2. The higher hydraulic head in well 2-3 indicates a downward gradient compared to heads in wells 2-1 and 2-2. The lack of response in well 2-3 to the pumping stress, however, indicates that although ground water may flow vertically through the lower till, the hydraulic connection of the till to the lower gravel is complex or possibly nonexistent. The decline in the water levels in well 2-3 may illustrate the effect of the reduction in available recharge due to evapotranspiration from overlying layers.

Wells 2-1 and 2-2 are screened in gravel that is present at 65 to 67 ft below land surface

(fig. 6). Well 2-1 extends to the bottom of the gravel unit (67 ft) and has no bentonite seal on the bottom of the casing. Well 2-2 (73 ft) is sealed from the bedrock with bentonite. At least 3 ft of dry, silty clay separates the gravel unit from the underlying bedrock.

Steady-state conditions observed in drawdown curves for the pumped-well test indicate leakage through either the upper, lower, or both confining units to the gravel (fig. 17). Physical indications of downward leakage include numerous saturated sand and silt layers overlying the gravel, dry clay underlying the gravel, slightly higher percentages of silts and clays in the layer

underlying the gravel than in the nearest layers overlying the gravel, and a downward hydraulic gradient evident in the water-level measurements in the wells all indicate downward leakage. Quantitative analysis of these factors is not possible because of the lack of hydrologic information on the units underlying the gravel. On the basis of the stratigraphic and hydraulic-gradient information available, however, downward leakage is assumed.

Hantush and Jacob (1955) considered leakage from a single confining unit as applied to results of aquifer tests. Their method does not account for water released from storage in the compressible confining units and, therefore, may overestimate the drawdown at early time (Moench, 1985). In this study, however, the small amount of stress applied to the gravel and confining units during the pumped-well test probably results in a minor contribution from storage in the confining units. Therefore, the Hantush and Jacob (1955) method was selected for application.

The Hantush and Jacob (1955) method is based on the following assumptions: (1) The aquifer is homogeneous and isotropic; (2) the aquifer has an infinite areal extent; (3) the pumped well is fully penetrating, and a constant pumping rate is maintained; (4) the well has an infinitesimally small diameter; (5) the aquifer is overlain, or underlain, everywhere by a confining unit having a uniform hydraulic conductivity and thickness; (6) the confining unit is overlain, or underlain, by a unit with infinite areal extent and constant head; (7) there is no release from storage in the confining unit; and (8) flow in the aquifer is horizontal, whereas flow in the confining unit is vertical.

For site 2, assumptions 1, 2, 5, 6, and 8 are considered valid for the zone of influence of the pumped well. The pumping rate (assumption 3) varied throughout the pumped-well test because of variations in the gas pressure to the pump and changes in the water pressure in the well as water levels declined. The pumping rate was adjusted throughout the test, however, to keep errors to a minimum. The diameter of the pumped well (4 in.), coupled with the low pumping rate, may make assumption 4 a source of error in the analysis. This error can be overlooked if only data collected about

60 minutes after the test was started (the time required to drain well-bore storage with simultaneous recharge to the well at a pumping rate of 27 ft/d) are used in the analysis. Assumption 7 may not be entirely met at site 2, but release from confining-unit storage probably contributes only minor amounts of water to the gravel.

For site 2, the hydraulic gradient produced from lowering the head in the gravel affects the upper confining till unit, causing leakage through the upper confining unit into the gravel. Equilibrium in the system is reached when the discharge from the pumped gravel unit equals the leakage across the confining unit to the gravel. This conceptualization is valid for site 2 because the discharge rate was constant throughout the test, and drawdown stabilized in the pumped well after the initial drawdown. Water levels are transient in well 2-2 during much of the test (fig. 20), due in part, perhaps, to partial dewatering of the gravel unit and to pulses sent through the formation from cycling of the bladder pump.

The equation for transient radial flow in a leaky confined aquifer of infinite areal extent, defined by Hantush and Jacob (1955, p. 98) and later described by Walton (1962) and Wolansky and Corral (1985), is given as:

$$s = \frac{Q}{4\pi T} \int_{\mu}^{\infty} \frac{1}{y} exp\left(-y - \frac{r^2}{4B^2 y}\right) dy, \qquad (2)$$

where s is   drawdown   of   the potentiometric surface, in feet;

Q is the discharge, in ft^3/d;

T is the transmissivity, in ft^2/d;

B is the leakage factor, in feet;

r is the radius from the pumped well, in feet; and

y is the variable of integration.

The integral can be expressed as $W(u,r/B)$ (the leaky well function of u), and the drawdown in the aquifer expressed as
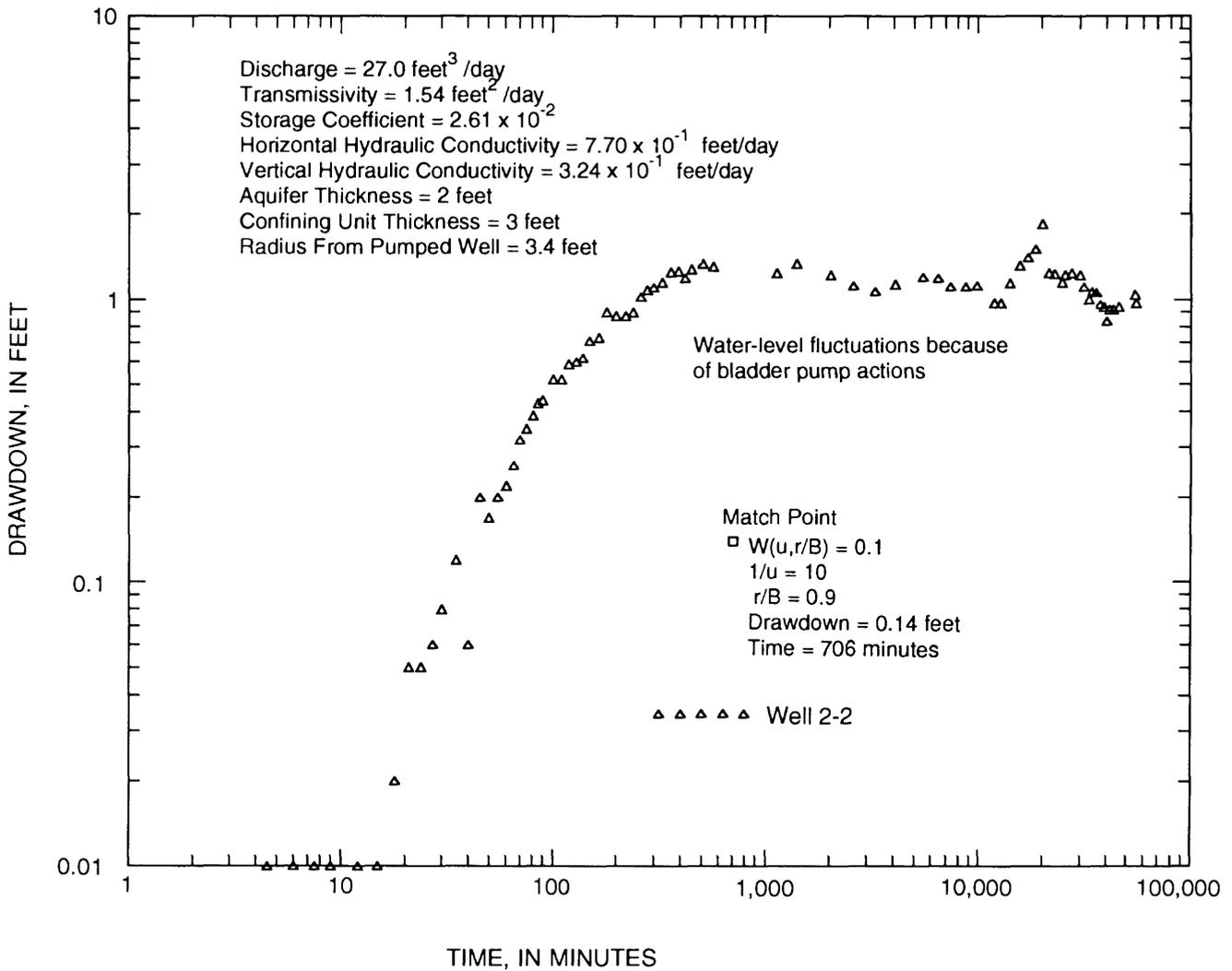
Figure 20. Drawdown in well 2-2 resulting from discharge during the pumped-well test, September 17-October 26, 1990.

$$s = \frac{Q}{4\pi T} W(u, \frac{r}{B}), \tag{3}$$

The quantity u is defined by

$$u = \frac{r^2 S}{4Tt}, \tag{4}$$

and r/B is given by

$$\frac{r}{B} = \frac{r}{(T + \frac{K'}{b'})^{1/2}}, \tag{5}$$

where S is the storage coefficient;

t is time, in days;

K' is the vertical hydraulic conductivity in the overlying confining unit, in ft/d; and

b' is the thickness of the overlying confining unit, in feet.

The method for obtaining values for the function $W(u,r/B)$ was simplified by Hantush (1956) and later by Walton (1960) and Reed (1980) by providing tables and type curves of $W(u,r/B)$ plotted against $1/u$ for various values of r/B. In addition, the horizontal hydraulic conductivity (K) of the pumped aquifer can be expressed as

$$K = \frac{T}{b} \quad , \tag{6}$$

$$\frac{K}{K'} > 100 \times \frac{b}{b'}. \tag{7}$$

where T is the transmissivity, in ft²/d; and

b is the aquifer thickness, in ft.

At site 2, the well 2-2 was in the same unit as the pumped well (2-1) and was, therefore, selected for analysis by means of the Hantush-Jacob method. Because drawdown in the pumped well was strongly affected by well-bore storage and cycling of the bladder pump and thus there was a large potential for error in matching the curves, well 2-1 was omitted from analysis. The type curve of $r/B = 0.9$ provided the best match of the data from well 2-2 (fig. 20). Match points from the curves were

$$
\begin{aligned}
W(u,r/B) &= 0.1, \\
1/u &= 10, \\
s &= 0.14 \text{ ft, and} \\
t &= 0.49 \text{ day.}
\end{aligned}
$$

With $Q = 27.0$ ft /d, $r = 3.4$ ft, b (the effective thickness of the aquifer, although the screen is 5 ft) = 2 ft, and b' (the approximate thickness of the clayey till overlying the gravel and underlying the nearest wet sand layer) = 3 ft, the hydraulic characteristics obtained for site 2 were

$$
\begin{aligned}
T &= 1.54 \text{ ft}^2 /d, \\
S &= 2.61 \times 10^{-2}, \\
K' &= 3.24 \times 10^{-1} \text{ ft/d, and} \\
K &= 7.70 \times 10^{-1} \text{ ft/d.}
\end{aligned}
$$

These calculations indicate that the vertical hydraulic conductivity is large in comparison to textbook ranges for till, possibly because of flow along fractures through the upper confining unit down to the gravel and (or) the possible addition of water to the gravel by leakage from the underlying confining unit, although the previously listed arguments and the apparent lack of fractures in the till sampled below the gravel do not support this later argument. The conditions of assumption 8, that flow is vertical in the confining unit and horizontal in the aquifer, was quantified by Hantush (1967) by the equation

The assumption of vertical flow through the confining unit may be valid at site 2 because of the strong effect of vertical fractures on the flow.

Sensitivity of the results to variations in geologic characteristics were tested by setting b equal to 5 ft in equation 7. In this case, flow across the entire screened interval is considered and equation 7 applied. This consideration affects only K, which becomes 3.08×10^{-1} ft/d, or roughly equivalent to K'. By doubling the value of b' to 6 ft, the computed value of K' is 6.47×10^{-1} ft/d. Therefore, changing the thickness of the aquifer or confining unit by more than double the original estimates at this site would result in values within the same order of magnitude and would have only minor effects on the overall interpretation of the test results.

Site 3--Lacustrine Deposits

The geologic setting of Sandusky County is similar to that in Union County. The Silurian Lockport Dolomite and the overlying Bass Islands Group underlie most of Sandusky County, and the Devonian Columbus Limestone underlies the southeastern corner of the county (fig. 11). These rocks are all part of the carbonate-bedrock aquifer of Ohio (Bugliosi, 1990). Site 3, in Sandusky County, was selected as characteristic of northern Ohio (fig. 3) for several reasons: (1) The surficial lacustrine deposits cover a large part of the county, (2) surficial deposits are thick in southern Sandusky County (about 120 ft to bedrock) but are much thinner in the north, and (3) a number of Late Wisconsinan shoreline deposits (Lake Warren, Lake Arkona, Lake Whittlesey, and stages of Lake Maumee) cross the county.

The glacial geology of site 3 reflects the various lake-level changes during the Pleistocene (fig. 7). The general stratigraphic sequence for the area is alternating sand (shore or nearshore deposits) and lacustrine clay units. The clay units commonly are about 50 feet thick in northern Sandusky County (Palombo, 1974).

In the Green Springs area (fig. 3), lacustrine clays ranging from a few feet to more than 60 ft thick are underlain by continuous fine-grained sand deposits that extend 60 to 70 ft to bedrock.

The pumped-well test at site 3, Green Springs State Nursery, Sandusky County, lasted from June 5 through June 20; recovery measurements were made from June 20 through June 26. Variations in the rate of pumping as a result of changes in the gas pressure to the pump and in hydraulic head presented difficulties in maintaining a constant discharge for the duration of the test. The most rapid change in discharge occurred during the initial drawdown in the well; consequently, the data from the first 15 minutes of the test are considered invalid for quantitative analysis.

After about 45 minutes of pumping, fluctuations in discharge decreased, and a mean discharge of 28.5 ft^3/d was calculated for the period from 45 to 600 minutes after the start of the pumping. From 600 minutes (0.42 day) after the start of the pumping through the end of the test at 19,690 minutes (13.7 days), a mean discharge of 27.0 ft^3/d was calculated. Water levels declined to the top of the pump at 5,480 minutes (3.8 days) and remained near this position for the remainder of the test. This level is approximately at or below the bottom of the confining unit, so dewatering of the aquifer may have affected the water-level measurements; however, as at site 2, the cyclic action of the bladder pump permitted water levels to fluctuate above the bottom of the confining unit during the time between pumping cycles.

Fluctuations in discharge decrease the period of time that dewatering affects the pumped-well test by allowing some recovery in the aquifer. Nevertheless, for the purpose of accuracy, emphasis is placed on the drawdown data before 5,480 minutes (3.8 days). The pumped-well test was continued after the drawdown in the pumped well reached the top of the pump because the cone of depression and the effects of leakage from the confining units could still be examined.

The screens on wells 3-1 and 3-2 partly penetrate the upper part of the confined aquifer (fig. 7). Well 3-3 is screened within the confining unit from 32 to 37 ft below land surface and fully penetrates a 2-ft thick sand unit. Well 3-4 is screened from 6 to 11 ft below land surface in an unconfined sand unit. Water levels in well 3-4 represent the water table.

Hydraulic head declined in well 3-2 as a result of discharge from the pumped well (well 3-1), whereas hydraulic heads in wells 3-3 and 3-4 were unaffected by the pumping stress (fig. 21). Closer examination of water-table fluctuations represented in well 3-4 indicates a correlation with atmospheric pressure (fig. 22). This was the only observation in which atmospheric pressure significantly affected water levels at any of the study sites. As figure 23 shows, increases in atmospheric pressure result in a proportional decrease in water levels in confined aquifers (U.S. Bureau of Reclamation, 1977, p. 156). Todd (1980, p. 235-238) attributes changes in water levels in wells penetrating confined aquifers to changes in the compressive stress on the aquifer due to changes in atmospheric pressure. This relation is expressed as

$$B = \frac{\gamma \Delta h}{\Delta p} \quad , \quad (8)$$

where B is the barometric efficiency;

γ is the specific weight of water, in lb;

h is the change in piezometric level, in ft; and

p is the change in atmospheric pressure, in lb/ft.

In unconfined aquifers, such as represented by water levels in well 3-4, an increase in atmospheric pressure compresses air trapped below the water table. This compression results in increased availability of pore space for soil moisture and consequently, a fall in the water table (Peck, 1960; Freeze and Cherry, 1979).

The relation between water levels and changes in atmospheric pressure in well 3-4 is complicated by the effects of recharge and evapotranspiration. Although trends in the water-level data during the pumped-well test indicate that changes in the water table are proportional to changes in atmospheric
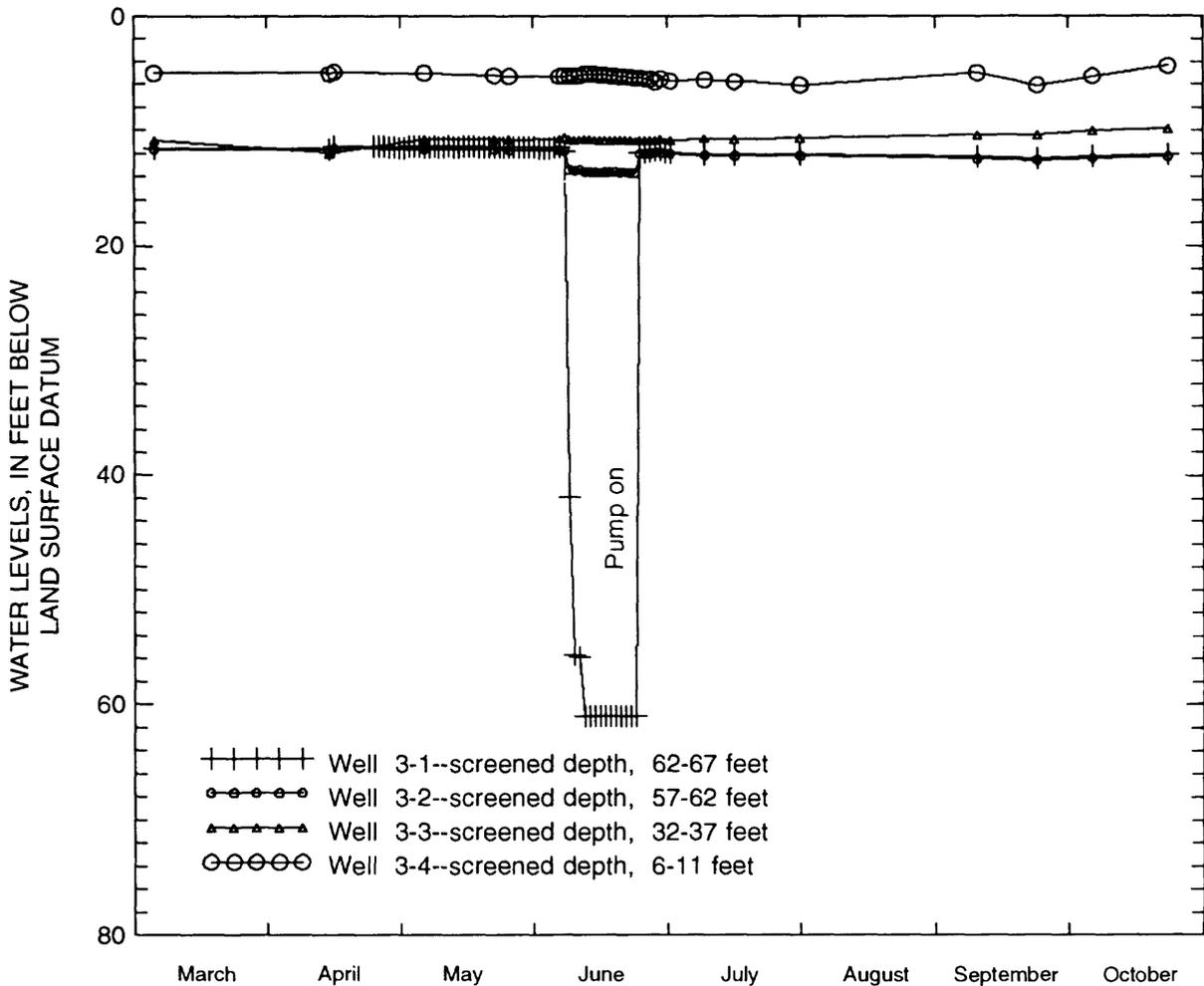
Figure 21. Water-level trends in wells at site 3, March 5-October 5, 1990.

pressure, precipitation can produce a rise in the water table during barometric lows. The steady decline in the water table after precipitation is similar to that observed in the shallow wells at sites 1 and 2 (figs. 17 and 20), where evapotranspiration is believed to affect water-table declines. The sharp fluctuations in water levels and atmospheric pressure late in the pumped-well test, however, support the hypothesis that atmospheric pressure contributes to water-level changes in well 3-4.

The sandy texture of the materials at site 3 allows the transmission of the changes in barometric pressure to hydraulic heads more rapidly at this site than at the other two study sites. The sandy soil and underlying sand units that extend to about 7 ft below land surface and make up a shallow, unconfined aquifer at site 3

are generally more conductive to water and atmospheric influences than the clay-rich soils at sites 1 and 2.

Only minor fluctuations were observed in water levels in well 3-3 during the pumped-well test (fig. 22), and no correlation was found between change in hydraulic head and pumping, precipitation, or changes in atmospheric pressure. For these reasons, there is no indication of vertical water flow downward through the confining till units in response to pumping from the aquifer.

Hydraulic-head declines in wells 3-1 and 3-2, which partly penetrate about 11 percent of the aquifer, are in response to discharge from the aquifer. Because partial penetration creates vertical flow gradients near the pumped well, the effects of partial penetration must be
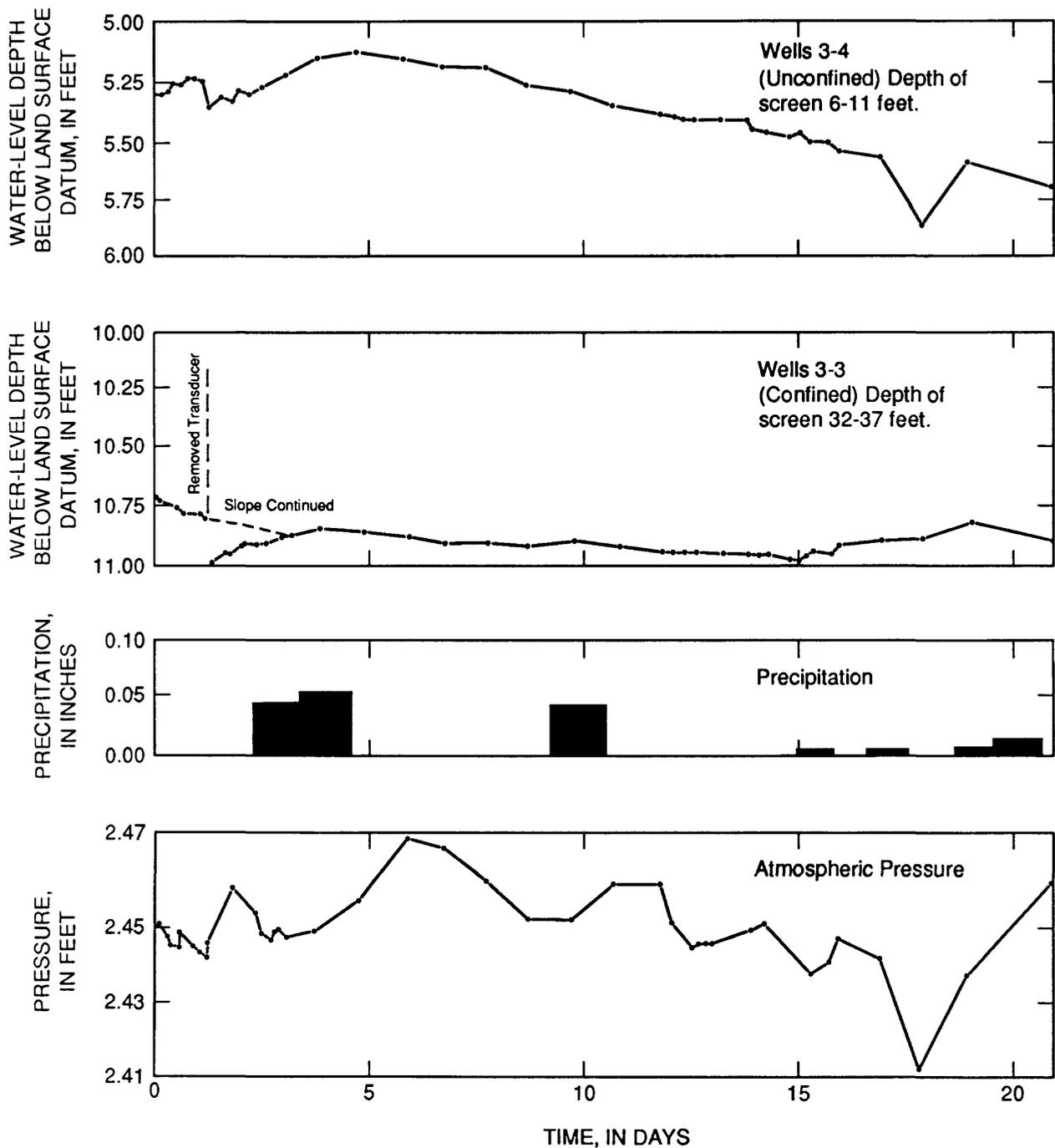
Figure 22. Water-level trends in wells 3-3 and 3-4, precipitation, and atmospheric pressure during the pumped-well test and recovery measurements, June 5-26, 1990.

accounted for in the analysis of the test data. Hantush (1964) concluded that the effects of partial penetration are insignificant in the observation well when

$$r > 1.5b \left(\frac{Kz}{Kr}\right)^{1/2} , \qquad (9)$$

where Kz/Kr is the ratio of the vertical hydraulic conductivity to the radial hydraulic conductivity in

Figure 23. Relation between water levels and barometric pressure (modified from U.S. Bureau of Reclamation, 1977, p. 156).
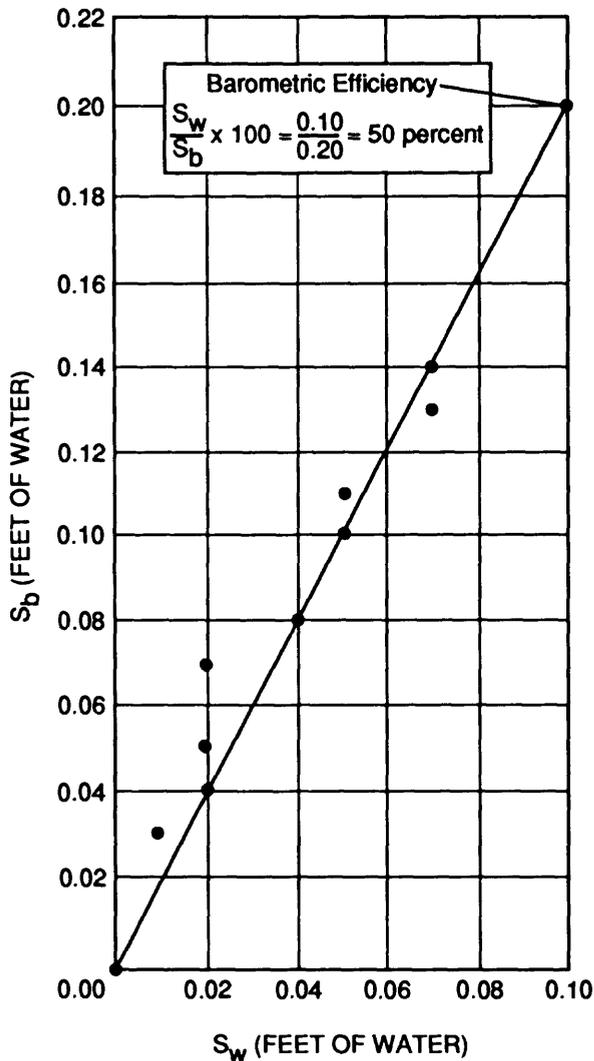
EXPLANATION

S_w ----- Absolute water-level change

S_b ----- Absolute barometric-pressure change

the aquifer; and

b   is the thickness of the aquifer, in ft.

Kz/Kr is considered equal to 1.0 (isotropic conditions) for the conditions at site 3 because there is insufficient evidence from analysis of the cores to justify an assumption of anisotropy

in this well-sorted sand aquifer. Drillers' logs for site 3 indicate a depth to bedrock of approximately 90 ft, resulting in an aquifer thickness (*b*) of 45 ft. The radius (*r*) of observation well 3-2 from the center of the pumped well is 3.4 ft. As a result, the effects of partial penetration are significant in the aquifer analysis.

Effects of the pumping on water levels were not observed in well 3-3, which is completed in a very fine sand and silt lens within the clay units overlying the pumped aquifer (fig. 7). This absence of effects indicates that the clay deposits surrounding the screen in well 3-3 are effectively impermeable. Therefore, water entering the aquifer during the pumped-well test may come from the underlying bedrock, from release from storage in the confining unit overlying the aquifer, or from fracture flow in the geologic units directly overlying the aquifer.

Analyses of the geologic cores of the units overlying the aquifer obtained during well installation indicate that till deposits from numerous glacial advances are present at depth intervals of 48 to 53 ft at this site (fig. 7), and that hydraulic properties resulting from buried weathered surfaces are variable with depth. Horizontal fractures lined with caliche are present in lacustrine clays at depth intervals of 30 to 35 ft and 41 to 50 ft. These fractures are the result of either exposure to weathering and desiccation in the geologic past or pressure release resulting from isostatic rebound after retreat of the glacial ice or drainage of the glacial lake. The presence of caliche indicates water flow through the fractures. It is uncertain whether the fractures still function as conduits to ground-water flow. Water flow through the fractures, however, is considered in the analysis of the pumped-well test.

Two factors were considered in the analysis of the pumped-well-test data: (1) the pumped well partly penetrated the aquifer and thus affected the drawdown in the nearby observation well; and (2) the drawdown data indicated leakage through the confining unit. The procedures for analysis of a pumped-well test in a leaky confined aquifer with constant discharge from a partly penetrating well are outlined by Hantush (1964). The assumptions of the analysis are: (1) constant discharge is

maintained, (2) the well partly penetrates the aquifer and is of an infinitesimal diameter, (3) vertical hydraulic conductivity (Kz) does not equal the radial hydraulic conductivity (Kr) in the aquifer, (4) the aquifer is overlain or underlain by a unit with infinite areal extent and uniform thickness (b') and vertical hydraulic conductivity (K') throughout, (5) the confining unit is overlain or underlain by a constant-head source bed of infinite areal extent, (6) there is no release from storage in the confining unit, (7) flow in the confining unit is vertical, and (8) leakage from the confining unit into the aquifer has no independent component of vertical flow.

Many of these assumptions were not completely met by the pumped-well test at site 3. Difficulties in maintaining a constant discharge during most of the test and inadequate evidence to support anisotropy in the aquifer may cause deviations from the assumptions outlined by Hantush (1964). It is uncertain how important water released from storage from the confining unit is to the analysis, but it is assumed to be minor because of the negligible stress applied to the aquifer by the small pumping rate. Because the deviations from the assumptions are probably minor, the Hantush (1964) method adequately provides general information about the hydraulic properties of the aquifer and the confining unit. Adjustments to equations 2 and 3 by Hantush (1964, p. 353) to compensate for partial penetration are given by

$$s = \frac{Q}{4\pi T}\left[W(u,\beta) + \bar{f}\left(u, \frac{ar}{B}, \beta, \frac{d}{b}, \frac{l}{b}, \frac{d'}{b}, \frac{l'}{b}\right)\right],$$

in which

$$\bar{f}\left(u, \frac{ar}{B}, \beta, \frac{d}{b}, \frac{l}{b}, \frac{d'}{b}, \frac{l'}{b}\right) = \frac{2b^2}{\pi^2}(l-d)(l'-d') \bullet$$

$$\sum_{n=1}^{\infty} \frac{1}{n^2}\left(\sin\frac{nl}{b} - \sin n\pi\frac{d}{b}\right) \cdot \left(\sin n\pi\frac{l'}{b} - \sin n\pi\frac{d'}{b}\right) \bullet$$

$$W\left(u, (\beta^2 + (n\pi a\frac{r}{b})^2)^{1/2}\right), \tag{10}$$

where a is $(Kz/Kr)^{1/2}$

β is a variable of integration;

B is the leakage factor;

b is the aquifer thickness, in ft;

n is a variable of integration;

l is the depth below the top of the aquifer to the bottom of the pumped-well screen, in ft;

d is the depth, below the top of the aquifer at the top of the pumped-well screen, in ft;

l' is the depth, below the top of the aquifer to the bottom of the observation well screen, in ft; and

d' is the depth, below the top of the aquifer to the top of the observation well screen, in ft

(Reed, 1980; Wolansky and Corral, 1985).

Practical application of the Hantush (1964) method is extremely difficult because of the large number of variables possible in an aquifer setting. Reed (1980) formulated a Fortran program that produces type curves for analyzing aquifer data for specific variables, thereby greatly simplifying the application of the Hantush method. A personal-computer version of Reed's program was used to analyze the data at site 3.

A wide range of $1/u$ values were input in each case, thereby producing adequate representation of the drawdown data along the length of the curve. The r/B values each produce specific curves of $W(u,\beta) + f$ plotted against $1/u$. Matching the appropriate r/B curve to the actual drawdown data for the observation well (well 3-2) was done by trial and error, as was the interpretation of the drawdown data that best met the original assumptions of the method. Constant Q was generally maintained between 45 and 600 minutes into the aquifer test; it then

decreased slightly from 600 (0.42 day) through 5,480 minutes (3.8 days). For this reason, emphasis was placed on the drawdown data between 45 and 600 minutes after the start of the pumping for curve matching, with less emphasis on the data after 600 minutes because the assumption of a constant pumping rate was not maintained.

A curve of r/B equal to 1.5 best fit the drawdown data for well 3-2 (fig. 24). The match point of $1/u$ equal to 10 and $W(u,\beta) + f$ equal to 0.1 produced values for s of 0.11 ft and t of 300 minutes, or 0.21 day. Rearranging equations 3, 4, and 6, the following values of T, S, and K were obtained from the data for site 3:

$$T = 2.07 \text{ ft}^2/\text{d},$$

$$S = 1.50 \times 10^{-2}, \text{ and}$$

$$K = 4.60 \times 10^{-2} \text{ ft/d}.$$

A value for K' was calculated for the unfractured gray till unit directly above the aquifer. The fractured till overlying this gray till was considered to be a source bed to the confining unit of constant hydraulic head. A thickness for b' of 3.0 ft applied to equation 5 gives $K' = 1.17$ ft/d for the confining unit.

If the values computed for T, S, K, and K' are assumed to be valid, the component of vertical flow in the confining unit is greater than the component of lateral flow in the aquifer. This violates the conditions of assumption 7 of Hantush (1964) (defined in equation 7) that states that the horizontal hydraulic conductivity in the aquifer must be significantly greater than the vertical hydraulic conductivity in the confining unit. The large K' for the confining unit similar to that observed at site 2 indicates that leakage alone does not account for the flow across this till layer and that another factor affecting flow must be accounted for. Although vertical fractures were not identified in the cores from the confining unit, they may be present and could produce a larger K' than predicted for unfractured till. If this assumption is true, then flow across the confining unit can exceed flow measured in the aquifer without violating the test assumptions.

The drawdown data for well 3-2 indicates that a flow boundary was encountered around

5,480 minutes (3.8 days) into the test, as indicated in the sudden decline in water levels with no change in Q (fig. 24). This could be explained by a no-flow boundary at some distance from the pumped well; however, the geology of the area at site 3 precludes a no-flow boundary because there is no evidence of any significant changes in stratigraphy within a mile of the test site. Another possible explanation for the sudden water-level decline is from a decrease in vertical flow into the aquifer as water retained in fractures in the confining unit is depleted. The depletion of water in the fractured till of the confining unit could result in a sharp decline in the hydraulic head for wells within the cone of depression because water would not be available. These assumptions, if true, may account for the large K' value calculated from data obtained 45 to 600 minutes after the start of the test.

SUMMARY AND CONCLUSIONS

Glacial deposits are complex and can result from single or multiple erosional and depositional events. Analysis of cores and grain-size distributions shows that a changing depositional environment at the study sites produced variable stratigraphic sequences in these glacial deposits.

Postdepositional change due to weathering of the glacial deposits--including fracturing resulting from desiccation, biological action such as root growth, oxidation of minerals in the sediments, or isostatic rebound after the retreat of the ice sheet--may enhance ground-water flow. The intersection of fractures with one another and with sand and gravel lenses in the till can produce conduits for ground-water flow and K values orders of magnitude larger than those in units commonly considered restrictive to ground-water flow, such as till.

The geology at the three study sites differs greatly. Vertical ground-water flow through the glacial deposits at site 1 is probably enhanced by fractures within the till. Pumped-well-test analysis indicates hydraulic connection between the shallow wells within the till and the lower gravel unit; however, the analysis of the hydraulic effects of the pumped-well test were complicated because of the loss of water in the weathered tills by evapotranspiration. Site 2 is
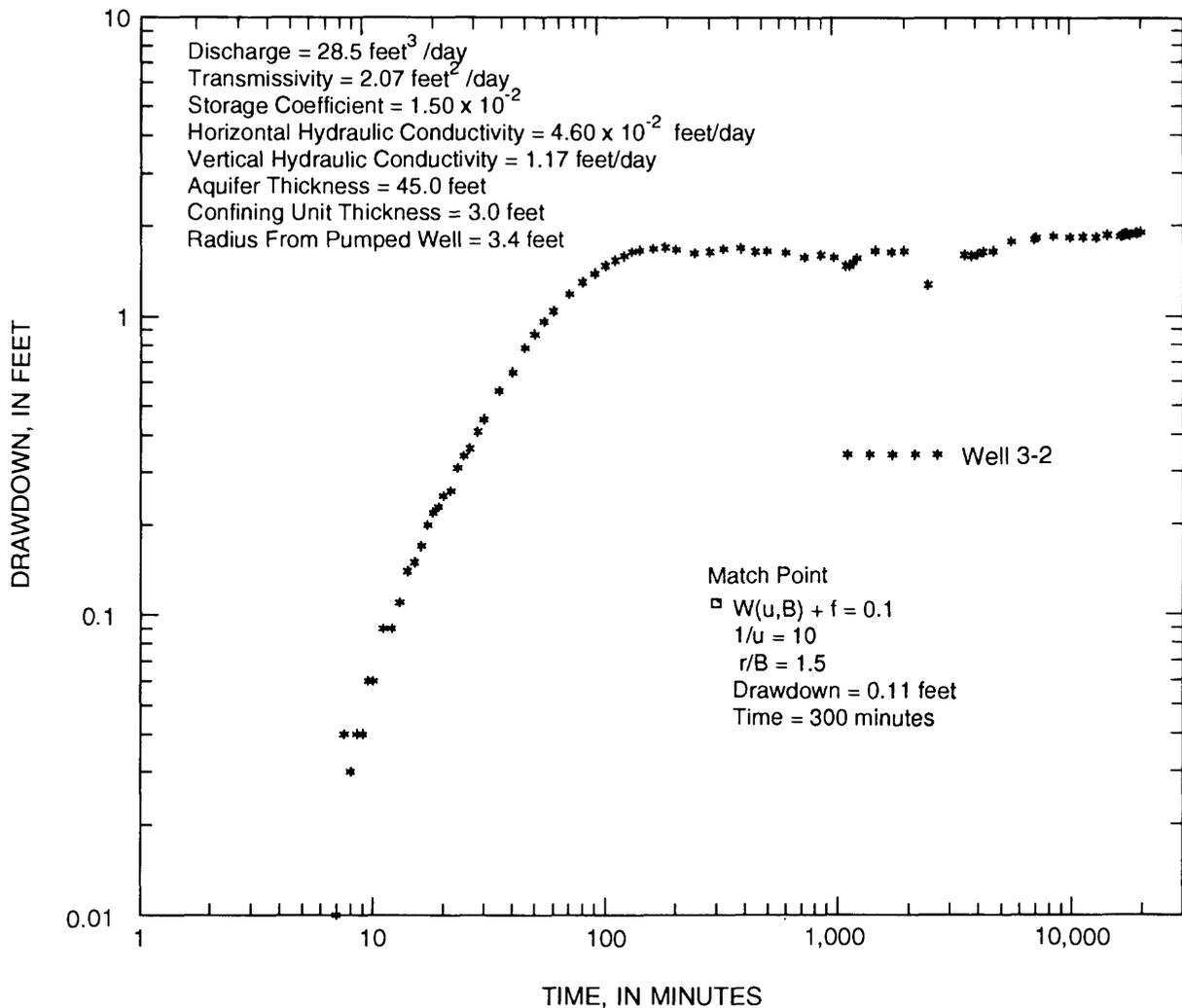
Figure 24. Drawdown in well 3-2 resulting from discharge during the pumped-well test, June 5-20, 1990.

similar to site 1 because fractures were observed to depths of 15 ft and transport of water through the weathered tills was indicated; however, there was no apparent reaction in the shallow wells to drawdown in the lower aquifer.

The greater thickness of the unweathered deposits at site 2, as compared to site 1, may account for the apparent absence of vertical flow at site 2. At site 3, water levels in observation wells in the lacustrine deposits did not change during the pumped-well test. Water in the unconfined sand unit reacted to atmospheric-pressure changes, whereas water levels within the unit in the center of the clay sequence and within the till units were unchanged by pumping.

The pumped-well tests at each site provided information about the hydraulic properties of the aquifers and the units that confine the aquifers (table 2). Although not quantified, the test results for site 1 indicate that the lower gravel unit transmits water in sufficient quantities to exceed the capability of the bladder pump (a pumping capacity of about 48 ft³/d). Recharge to the lower gravel unit is from flow through the overlying tills and is partially controlled by evapotranspiration.

The lower gravel unit at site 2 has a transmissivity of 1.54 ft^2/d, a storage coefficient of 2.61 x 10^{-2}, and a horizontal hydraulic conductivity of 7.70 x 10^{-1} ft/d. These values fall within the range of silty sand (U.S. Bureau of Reclamation, 1977; Freeze and Cherry, 1979), which reflects the poor sorting of these

Table 2. *Computed hydraulic properties from the pumped-well tests at the study sites*

[ft^2/d, feet squared per day; ft/d, feet per day; n/a, not applicable]

Site	Transmissivity (ft^2/d)	Storage coefficient	Horizontal hydraulic conductivity (ft/d)	Vertical hydraulic conductivity (ft/d)
[1]1	n/a	n/a	n/a	n/a
2	1.5	2.6×10^{-2}	7.7×10^{-1}	3.2×10^{-1}
3	2.1	1.5×10^{-2}	4.6×10^{-2}	1.2

[1] Difficulties with equipment at site 1 prevented collection of data suitable for quantitative analyses of hydraulic properties.

sediments and the strong influence of the fine-grained component on the hydraulic properties. The vertical hydraulic conductivity in the confining unit ranges from 3.24×10^{-1} to 6.47×10^{-1} ft/d, which is within the same order of magnitude as K and probably reflects fracture flow through the confining tills.

Data from the pumped-well test at site 3 indicates that the fine to very fine sand in the aquifer has a transmissivity of 2.07 ft^2 /d and a storage coefficient of 1.50×10^{-2}, which are in the same range as those computed for the gravel unit at site 2. The horizontal hydraulic conductivity, however, is much lower (4.6×10^{-2} ft/d) at site 3. The horizontal hydraulic conductivity is below the range outlined by Freeze and Cherry (1979) for clean sand; however, considering the small median grain size of the sand, which probably limits the hydraulic conductivity of the aquifer, the horizontal hydraulic conductivity of 4.6×10^{-2} ft/d appears reasonable.

The vertical hydraulic conductivity of the confining unit above the aquifer at site 3 is 1.17 ft/d. This comparatively high value may indicate flow through fractures in the till. Fractures were not observed in the core analysis at this depth; however, the presence of fracture permeability at the base of a nonweathered till has been documented by Keller and others (1986) for a till in Saskatchewan, Canada. At site 3, the multiple glacial advances may have buried weathered till units below later glacial and lacustrine deposits. Although not observed in the cores, buried weathered tills may be present, and if so, could produce fracture flow at the base of unweathered, unfractured till units. The estimated vertical hydraulic conductivity may be questionable, however, if the analysis is affected by water released from storage in the confining unit or by significant anisotropy in the aquifer that results in a large upward component of flow.

The grain-size analyses provide information on the lithologies of the glacial deposits; however, these analyses do not account for other characteristics that affect ground-water flow, such as fractures and secondary mineralization. The horizontal hydraulic conductivities calculated from the grain-size analysis are not considered valid for these sites because of the large clay content of the tills and lacustrine deposits.

This investigation outlines hydrogeologic characteristics at three sites in Ohio. The quantitative analysis and evaluation provide one possible interpretation of the mechanisms controlling ground-water flow through surficial deposits. Additional studies of the hydraulic properties of sediments with low hydraulic conductivities and increased development of analytical methodology for evaluating various hydrologic settings would be useful for improving an understanding of the mechanisms that affect ground-water flow in glacial deposits.

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