U.S. Department of the Interior
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Estimated Hydraulic Properties for the Surficial- and Bedrock-Aquifer System, Meddybemps, Maine

Open-File Report 99-199

Prepared in cooperation with the
U.S. ENVIRONMENTAL PROTECTION AGENCY

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By FOREST P. LYFORD, STEPHEN P. GARABEDIAN, and BRUCE P. HANSEN

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Northborough, Massachusetts
1999

U.S. DEPARTMENT OF THE INTERIOR
BRUCE BABBITT, Secretary

U.S. GEOLOGICAL SURVEY
Charles G. Groat, Director

For additional information write to:
Chief, Massachusetts-Rhode Island District
U.S. Geological Survey
Water Resources Division
10 Bearfoot Rd.
Northborough, MA 01532

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CONVERSION FACTORS

Multiply	By	To obtain
acre	0.4047	hectare
cubic foot per day (ft^3/d)	28.317	liters per day
foot (ft)	0.3048	meter
foot per day (ft/d)	0.3048	meter per day
foot per minute (ft/min)	0.3048	meter per minute
foot squared per day (ft^2/d)	0.09290	meter squared per day
gallons per day (gal/d)	0.003785	cubic meter per day
gallon per minute (gal/min)	0.06309	liter per second
gallon per minute per foot [(gal/min)/ft)]	0.2070	liter per second per meter
inch (in.)	25.4	millimeter
inch per year (in/yr)	25.4	millimeter per year
mile (mi)	1.609	kilometer

VERTICAL DATUM

Sea level: In this report, "sea level" refers to the National Geodetic Vertical Datum of 1929 (NGVD 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.

Estimated Hydraulic Properties for the Surficial- and Bedrock-Aquifer System, Meddybemps, Maine

By Forest P. Lyford, Stephen P. Garabedian, *and* Bruce P. Hansen

Abstract

Analytical and numerical-modeling methods were used to estimate hydraulic properties of the aquifer system underlying the Eastern Surplus Company Superfund Site in Meddybemps, Maine. Estimates of hydraulic properties are needed to evaluate pathways for contaminants in ground water and to support evaluation and selection of remediation measures for contaminated ground water at this site.

The hydraulic conductivity of surficial materials, determined from specific-capacity tests, ranges from 17 to 78 feet per day for wells completed in coarse-grained glaciomarine sediments, and from about 0.1 to 1.0 foot per day for wells completed in till. The transmissivity of fractured bedrock determined from specific-capacity tests and aquifer tests in wells completed in less than 200 feet of bedrock ranges from about 0.09 to 130 feet squared per day. Relatively high values of transmissivity at the south end of the study area appear to be associated with a high-angle fracture or fracture zone that hydraulically connects two wells completed in bedrock. Transmissivities at six low-yielding (less than 0.5 gallon per minute) wells, which appear to lie within a poorly transmissive block of the bedrock, are consistently in a range of about 0.09 to 0.5 foot squared per day.

The estimates of hydraulic conductivity and transmissivity in the southern half of the study area are supported by results of steady-state calibration of a numerical model and simulation of a 24-hour pumping test at a well completed in bedrock. Hydraulic conductivity values for the surficial aquifer used in the model were 30 feet per day for coarse-grained glaciomarine sediments, 0.001 to 0.01 foot per day for fine-grained glaciomarine sediments, and 0.1 to 0.5 foot per day for till. As part of model calibration, a relatively transmissive zone in the surficial aquifer was extended beyond the hypothesized extent of coarse-grained sediments eastward to the Dennys River.

Hydraulic conductivity values used for bedrock in the model ranged from 3×10^{-4} to 1.5 feet per day. The highest values were in the fracture zone that hydraulically connects two wells and apparently extends to the Dennys River. The transmissivity of bedrock used in the model ranged from 0.03 to 150 feet squared per day, with the majority of the bedrock transmissivities set at 0.3 foot squared per day. Numerical modeling results indicated that a very low ratio of vertical hydraulic conductivity to thickness (1×10^{-9} days^{-1}) was required to simulate a persistent cone of depression near a residential well that lies in the previously identified poorly transmissive block of bedrock.

INTRODUCTION

Volatile organic compounds (VOCs) have been detected in ground water in surficial materials and bedrock in two areas near the Eastern Surplus Superfund Site in Meddybemps, Maine (Lyford and others, 1998). The U.S. Environmental Protection Agency (USEPA) and local residents are concerned that contaminants, principally tetrachloroethylene (PCE), can move to existing residential wells and limit future development of ground-water resources in the area. Information on the hydraulic properties of the aquifer system is needed to assess the potential for contaminants in ground water to affect residential wells and to help evaluate remediation approaches. During 1997–98, the U.S. Geological Survey (USGS), in cooperation with USEPA, studied characteristics of the fractured crystalline bedrock and the hydraulic properties of the aquifer system near the Eastern Surplus Superfund Site.

The purpose of this report is to provide estimates of hydraulic properties for the surficial- and bedrock-aquifer system near the Eastern Surplus Superfund Site. Estimates of transmissivity and hydraulic conductivity determined from specific-capacity data and aquifer tests were refined by calibration of a numerical model for steady-state and transient conditions. Calibration of the numerical model also provided estimates of vertical hydraulic conductivity and reinforced the conceptual ground-water-flow model for the study area. The study focused on the area near the southernmost of two contaminant plumes that is closest to existing residential wells. The characteristics of fractures near the site are described in a companion report (Hansen and others, 1999).

Thanks are extended to Edward Hathaway, USEPA Project Manager, for logistical support during the study and to Mona Van Wart and Charlotte Smith for access to their wells during aquifer testing. Also, thanks are extended to Madge Orchard, Terry Lord, Greg Smith, and Harry Smith for access to their property.

DESCRIPTION OF STUDY AREA

The Eastern Surplus Superfund Site is on the western bank of the Dennys River at the outlet from Meddybemps Lake (figs. 1 and 2). A study area of approximately 30 acres encompasses the 4-acre Superfund site. The primary focus of this investigation was on the southern half of the study area. The following description of the hydrogeology of the region and study area is summarized from Lyford and others (1998).

The region that encompasses the study area is underlain mainly by the Meddybemps Granite. A small area centered on the study area is underlain by a gabbro-diorite, which is most likely a detached body of mafic rock within the Meddybemps Granite. Surficial materials include till, generally less than 10 to 20 ft thick, and extensive glaciomarine deposits, including coarse-grained and fine-grained (Presumpscot Formation) sediments deposited during deglaciation of the region. The coarse gravel and sand was deposited in an ancestral sea, probably as a subaqueous fan at the ice margin during retreat of the glacier. Glaciomarine silty clay of the Presumpscot Formation underlies much of the lowland area in the region. A sandy facies in the upper section of the Presumpscot Formation was deposited as the land rose relative to sea level and the shoreline regressed southeastward through the area.

Hydrogeologic units in the study area include till, coarse-grained glaciomarine deposits, fine-grained glaciomarine deposits, and bedrock. The vertical and lateral distribution of hydrogeologic units is shown in sections on figure 3. Till thickness ranges from less than 5 ft on the western side of the Dennys River to about 40 ft on the eastern side. The coarse-grained glaciomarine deposits are present at or near the surface in the western part of the study area; thickness ranges from 0 to more than 30 ft. The thick (more than 10 ft) coarse-grained sections are largely above the water table. Fine-grained glaciomarine deposits (Presumpscot Formation) are present in the central and southern parts of the study area where thickness ranges from 0 to about 20 ft. The silt-clay facies of this unit is poorly permeable and serves as a confining layer for ground water in underlying till and coarse-grained glaciomarine deposits. Ground water in bedrock occurs principally in fractures. The occurrence of water-yielding fractures ranges widely; in some wells only one or two fractures supply measurable quantities of water (more than 0.02 gal/min).

Ground-water levels in bedrock wells on the north side of the study area respond rapidly to rainfall. Responses to precipitation in surficial materials and bedrock are subdued or are not apparent where silts and clays of the Presumpscot Formation are present. The annual recharge may approach a potential rate of 24 to 26 in. where coarse-grained materials are present at the surface, but is probably less where till, silts, and clays are at the surface.

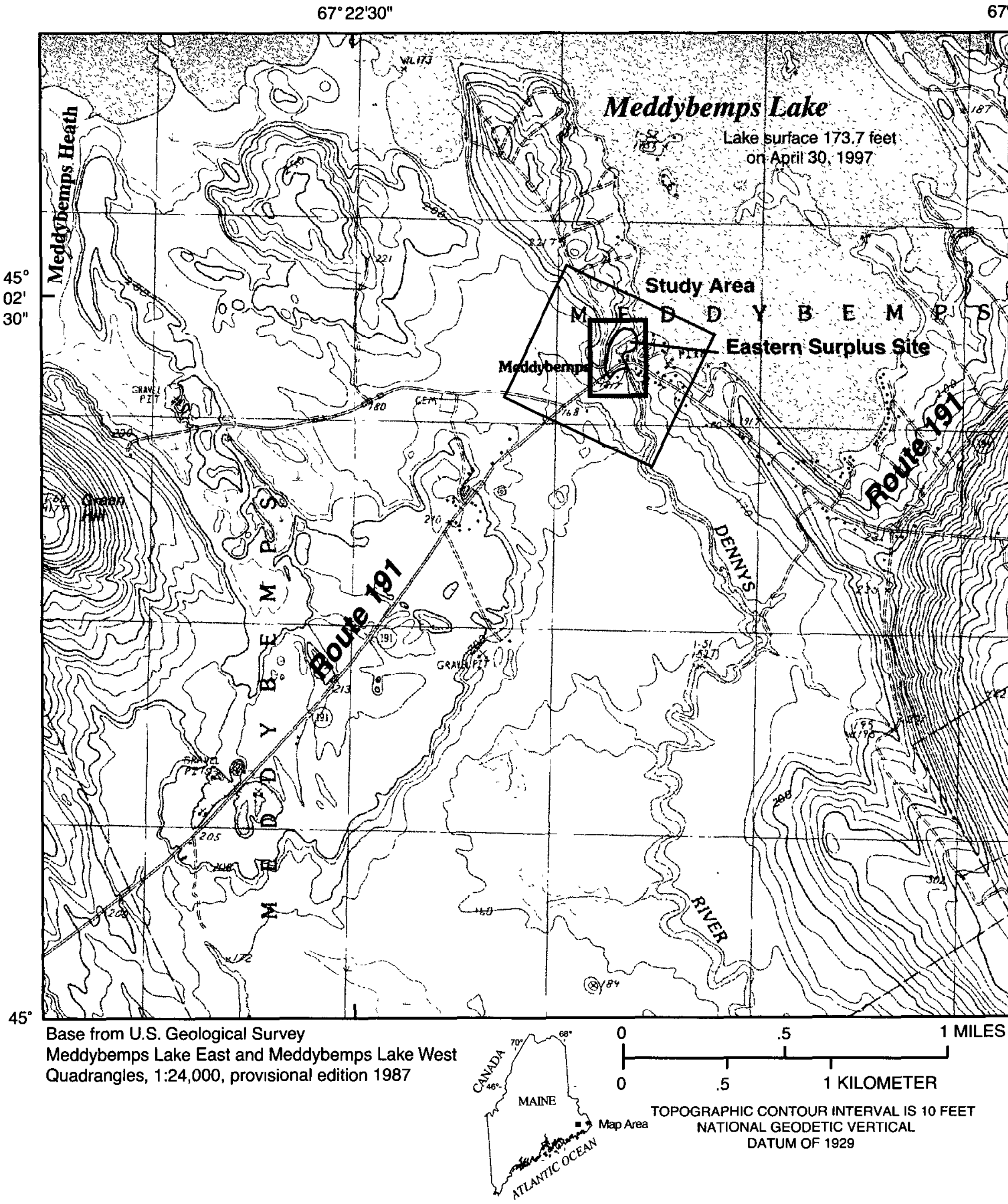


Figure 1. Location of the Eastern Surplus Superfund Site, study area, and numerical model area, Meddybemps, Maine. (Modified from Lyford and others, 1998, fig. 1.)

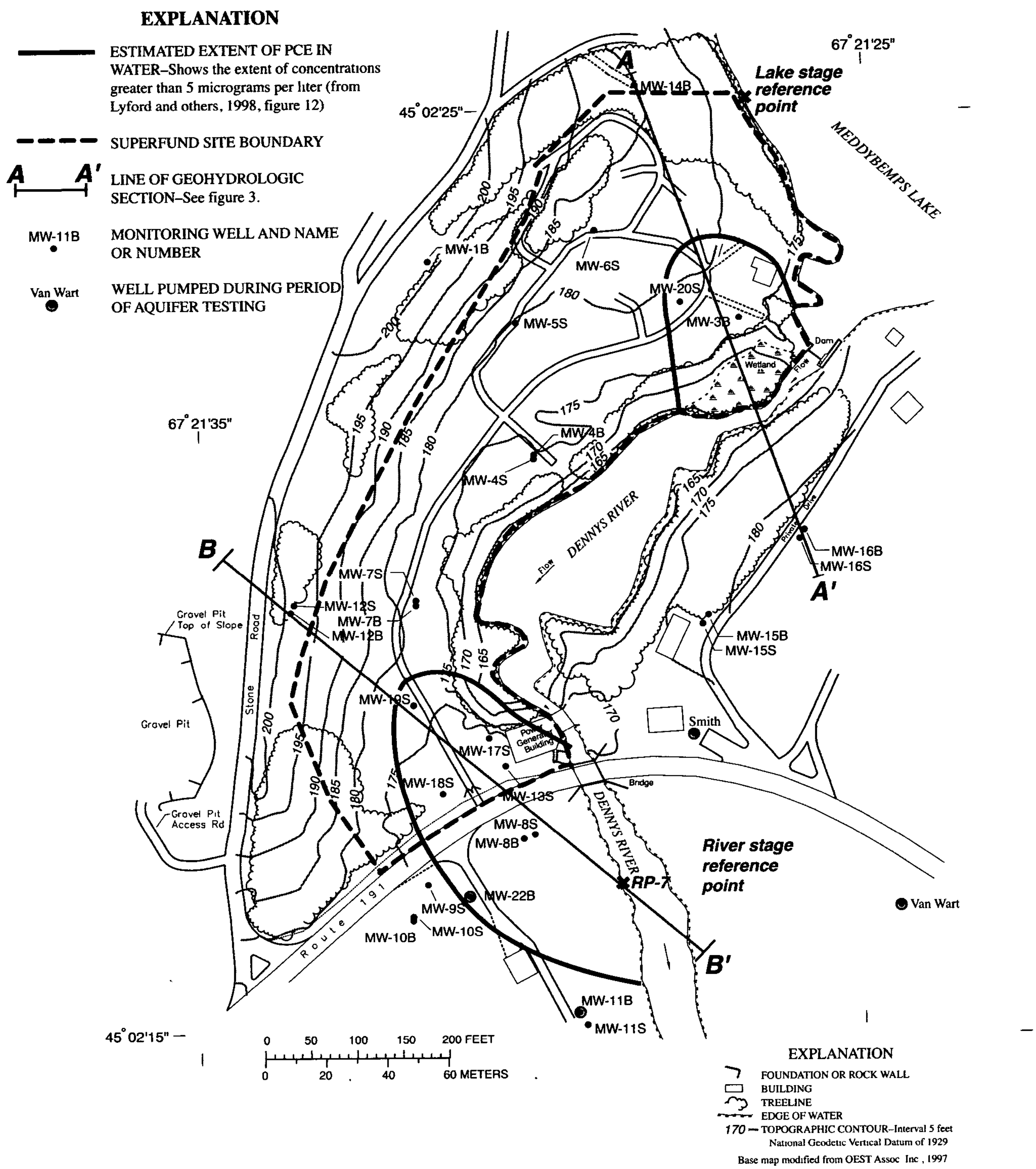


Figure 2. Location of study area, extent of tetrachloroethylene (PCE) in ground water and locations of wells, Meddybemps, Maine.

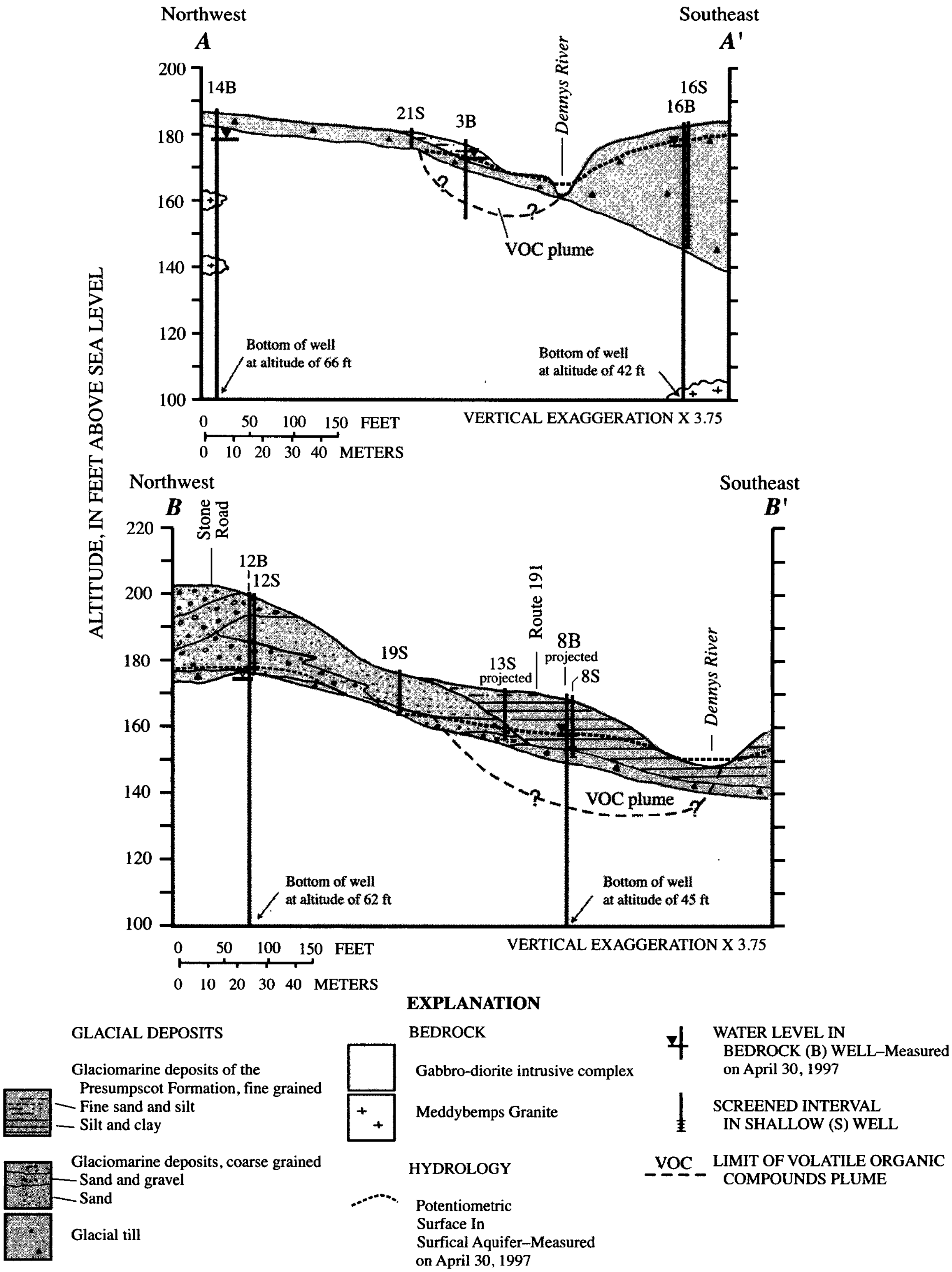


Figure 3. Geohydrologic sections A-A′ and B-B′, Meddybemps, Maine.

Ground water in surficial materials generally flows towards the Dennys River. The saturated thickness of surficial materials under the study area is generally less than 10 ft, and several monitoring wells screened in surficial materials "go dry" during extended periods of little or no recharge. South of Highway 191, ground water in coarse-grained sediments is confined below silts and clays of the Presumpscot Formation, but flow also is toward the Dennys River.

Ground water in bedrock flows towards the Dennys River from the eastern and western sides of the study area. In addition, hydraulic gradients are generally downward from the surficial aquifer to bedrock. Water-level data also indicate a potential for flow under the river from the western side to a cone of depression near a residential well on the eastern side. The cone of depression may extend laterally several hundred feet and affect water levels within a block of the aquifer characterized by low-yielding wells (less than 0.5 gal/min). In contrast, it is likely that the higher-yielding bedrock wells outside the low-yielding zone intersect high-angle fractures that extend to the overlying surficial aquifer.

Plumes of VOCs, including PCE and trichloroethylene (TCE), have been detected in ground water in two areas. VOCs in both plumes move through surficial materials and shallow bedrock towards the Dennys River. Contaminants in the southern plume could potentially move through fractures in bedrock to the local cone of depression east of the Dennys River.

ESTIMATION OF HYDRAULIC PROPERTIES USING ANALYTICAL METHODS

Specific-capacity data and drawdown data from aquifer tests conducted in wells in the study area were used to estimate aquifer transmissivity and hydraulic conductivity by applying conventional analytical methods. For tests in bedrock, an equivalent porous medium and radial flow are assumed. This section describes results of these analyses.

Specific-Capacity Measurements

The specific capacity of a well is the ratio of pumping rate of the well to drawdown at the well. The following formula presented by Cooper and Jacob (1946) is used here to estimate aquifer transmissivity from specific capacity data. Application of the method has been described by Fetter (1994).

$$T = \frac{2.30Q}{4\pi s}\log\frac{2.25Tt}{r^2S}, \qquad (1)$$

where:

T = transmissivity,
s = drawdown,
Q = pumping rate,
S = storage coefficient,
t = time since pumping began, and
r = well radius.

Application of equation 1 requires an estimate of the storage coefficient and an initial estimate of transmissivity. The final transmissivity is then determined through a series of iterations that uses the previously estimated or calculated value of transmissivity. An initial estimate of transmissivity, in feet squared per day, can be derived by multiplying the specific capacity of the well, in gallons per minute per foot of drawdown, times 200. The transmissivity determined by the specific-capacity method is, however, subject to considerable uncertainty. In addition to using an estimate of storage coefficient, which could be in error by a factor of 10 or more, the method requires the following assumptions: (1) well-entry losses are negligible, (2) the pumping period is sufficiently long to satisfy the requirements for applying the Cooper-Jacob formula ($r^2S/4Tt < 0.01$), and (3) wellbore storage effects are negligible. At a minimum, transmissivity values determined from specific-capacity tests provide relative values that can be used as indices for some types of hydrologic analyses. Specific-capacity data were collected during water-quality sampling (Roy F. Weston, Inc., 1997; 1998a), during borehole flowmeter tests, during recovery measurements in bedrock wells after drilling, and during aquifer tests. Well-construction data for the wells tested are summarized in table 1.

Table 1. Well construction data, water levels and stage on September 9–10, 1997, and wells equipped with water-level recorders during aquifer testing, Meddybemps, Maine

[All depths in feet below land surface except water level and stage, which are in feet below the measuring point. ft, feet; No., number; --, no data; na, not applicable]

Well or reference point No. or name	Date drilled	Altitude of land surface (ft)	Altitude of measuring point (ft)	Total depth of well (ft)	Depth to bedrock or refusal (r) (ft)	Screened (s) or open-hole (o) interval (ft)	Water level or stage below measuring point (ft)	Water-level altitude (ft)	Equipped with recorder during aquifer testing
MW-1B	4/17/88	201.60	204.18	57.8	34.6	s 38–53	36.62	167.56	no
MW-3B	4/17/88	177.37	179.89	23.3	9	s 13.3–23.3	8.96	170.93	no
MW-4S	4/15/88	174.84	177.60	18	19.5	s 13.0–18.0	13.82	163.78	no
MW-4B	4/14/88	174.75	177.43	39.7	19.5	s 24.7–39.7	15.48	161.95	no
MW-5S	10/23/96	179.86	182.06	13	r 13	s 10–13	13.78	168.28	no
MW-6S	10/23/96	182.34	184.71	7.0	r 7.0	o 4.5–7.0	dry	--	no
MW-7S	10/28/96	177.79	180.09	17.2	17	s 12–17	18.22	161.57	no
MW-7B	10/28/96	177.81	178.75	117.8	18	o 21–117.8	22.12	156.63	yes
MW-8S	10/25/96	167.30	169.14	16.5	r 16.5	s 14–16.5	12.66	156.48	yes
MW-8B	11/04/96	169.04	169.35	124	20.5	o 25.7–124	13.45	155.90	yes
MW-9S	10/25/96	174.03	175.52	16.5	r 16.5	s 14–16.5	dry	--	no
MW-10S	11/06/96	174.42	176.13	23	22	s 18–23	18.58	157.55	yes
MW-10B	11/4/96	174.24	175.64	120	20	o 26.4–120	18.97	156.67	yes
MW-11S	10/26/96	169.34	170.70	26	r 26	s 21–26	16.55	154.15	yes
MW-11B	11/04/96	169.69	170.63	129	29	o 35.1–129	15.83	154.80	yes
MW-12S	10/26/96	199.11	200.21	22	r 22	s 19–21.5	dry	--	no
MW-12B	11/4/96	200.13	201.34	138	22.5	o 27.7–138	26.72	174.62	yes
MW-13S	10/29/96	171.36	174.14	14	r 14	s 11–13.5	dry	--	no
MW-14B	11/05/96	185.70	187.33	120	3.5	o 9.4–120	14.17	173.16	no
MW-15S	11/06/96	178.46	179.32	38	36	s 26–36	17.08	162.24	no
MW-15B	11/05/96	178.97	180.11	240	39	o 46.9–240	25.76	154.35	yes
MW-16S	11/06/96	182.88	183.48	38	36	s 28–38	12.60	170.88	yes
MW-16B	11/05/96	182.18	183.91	138	38	o 42.3–140	13.39	170.52	yes
MW-17S	4/22/97	172.42	174.34	23	18.0	s 15–17.5	15.98	158.47	yes
MW-18S	4/23/97	172.90	174.82	19.5	18.0	s 16–18.5	17.10	157.85	yes
MW-19S	4/23/97	177.08	178.60	13.5	11.8	s 9.3–11.8	dry	--	no
MW-20S	4/24/97	178.57	180.33	8.0	6.0	s 3.5–6.0	dry	--	no
MW-22B	1950s	172.35	174.28	49	18	o 25.5–49	17.47	156.81	yes
VanWart	--	171.78	174.13	142	29	o 39–142	9.15	164.98	yes
Smith	--	173.35	174.55	420	--	--	110.10	[1]64.45	yes
Meddybemps Lake	na	na	174.09	na	na	na	2.33	171.76	no
Dennys River at RP-7	na	na	156.41	na	na	na	3.85	152.56	no

[1]Water level in the Smith well was recovering slowly at the time of measurement.

During sampling for water quality, wells were pumped at nearly constant rates that resulted in minimal and nearly constant drawdown. Pumping continued until water-quality parameters measured in the field had stabilized. Data are available for sampling periods in December 1996, June 1997, and October 1997. The pumping rate differed with well yield and was typically less than 0.4 gal/min. Yields for wells MW-11S, MW-4B, MW-8B, MW-12B, MW-15B, and MW-16B (fig. 1) were too low to sustain a constant pumping rate. These wells were sampled after purging and allowing the water level to recover (Roy F. Weston, Inc., 1998a). Several wells were dry or nearly dry in October 1997 after an extended period of low precipitation and could not be sampled. Specific-capacity data also were collected during borehole-flowmeter logging at the Van Wart well, during a brief (30 min) aquifer test at well MW-3B, and during aquifer tests at wells MW-11B and MW-22B.

A storage coefficient of 0.1 was assumed for most wells completed in the surficial aquifer. This value is within the range commonly assumed for unconfined aquifers (Lohman, 1972, p. 8) and is considered to be a reasonable estimate for the short-term tests that were typically 45 to 90 min in duration. Exceptions were wells MW-8S and MW-10S, where ground water is confined by clays of the Presumpscot Formation, and wells MW-15S and MW-16S, where the well screen is considerably below the water table. A storage coefficient of 1×10^{-4}, which is within the range commonly assumed for confined aquifers (Lohman, 1972, p. 8), was assumed for these wells. A storage coefficient of 1×10^{-4} was also assumed for all wells completed in bedrock. This value is at the high end of a range from 5×10^{-7} to 1×10^{-4} reported for fractured-rock aquifers (Earl Greene, U.S. Geological Survey, written commun., 1997). The importance of the storage-coefficient estimates will be discussed later in this section of the report.

An alternative approach to pumping was used to determine the specific capacity of bedrock wells MW-8B, MW-12B, MW-15B, and MW-16B. Water levels in these wells recovered slowly and at a nearly constant rate, indicating a constant inflow rate, for several hours to several days after they were drilled (fig. 4). Conceptually, when the water level in a well declines below the level of a bedrock fracture, drawdown at the well bore is effectively constant and flow to the well would be expected to gradually decline (Jacob and Lohman, 1952). A steady inflow rate while water-bearing fractures are above the water level in the well indicates either that a constant head source, such as leakage from a surficial aquifer, was controlling the inflow rate, or that sufficient time had elapsed so changes in discharge were very slow and not discernible in the recovery record. Computation of changes in flow with time using an equation presented by Jacob and Lohman (1952), and estimates of hydraulic properties at these low-yield wells, indicated that changes in flow should have been discernible during the recovery period. For this reason, leakage from a nearby source, causing steady flow to a well after a relatively short time, seems to be the more likely cause of the steady inflow rate. For this analysis, a time of 100 min for flow to stabilize was assumed.

A transmissivity value was computed for major fractures or fracture zones in these four bedrock wells using equation 1. Equation 1 was derived for constant-flow conditions, but it also applies to constant-drawdown conditions after very small values of time (Lohman, 1972, p. 23). The well yield, in cubic feet per day (ft^3/d), was computed by multiplying the rate of water-level rise by the cross-sectional area of the well. The yield for each fracture was assumed to be the percentage of total yield determined from borehole-flowmeter tests (Hansen and others, 1999). The drawdown for each fracture or fracture zone was the depth to the fracture below a static water level that was measured about 2 weeks after the well was drilled. The depths to water-bearing fractures and percentage of flow reported by Hansen and others (1999) are shown in figure 4. The transmissivity values reported in table 2 are the sum of transmissivity values computed for all water-bearing fractures or fracture zones in a well.

The estimates of transmissivity for all the bedrock wells that were made on the basis of specific-capacity data range widely—from 0.09 ft^2/d in well MW-15B to 130 ft^2/d in well MW-3B. The transmissivity range of 280 to 550 ft^2/d for well MW-22B shown in table 2 probably reflects the transmissivity of the bedrock and surficial aquifers combined, as discussed later in this report. For this reason, the estimates of transmissivity for well MW-22B are not included in this range. The transmissivity values for the low water-yielding wells MW-7B, MW-8B, MW-12B, MW-15B, MW-16B, and possibly MW-4B, are 0.5 ft^2/d or less. For comparison, Lyford and others (1998) report a transmissivity of 0.6 ft^2/d for the Smith Well, another low-yielding well.

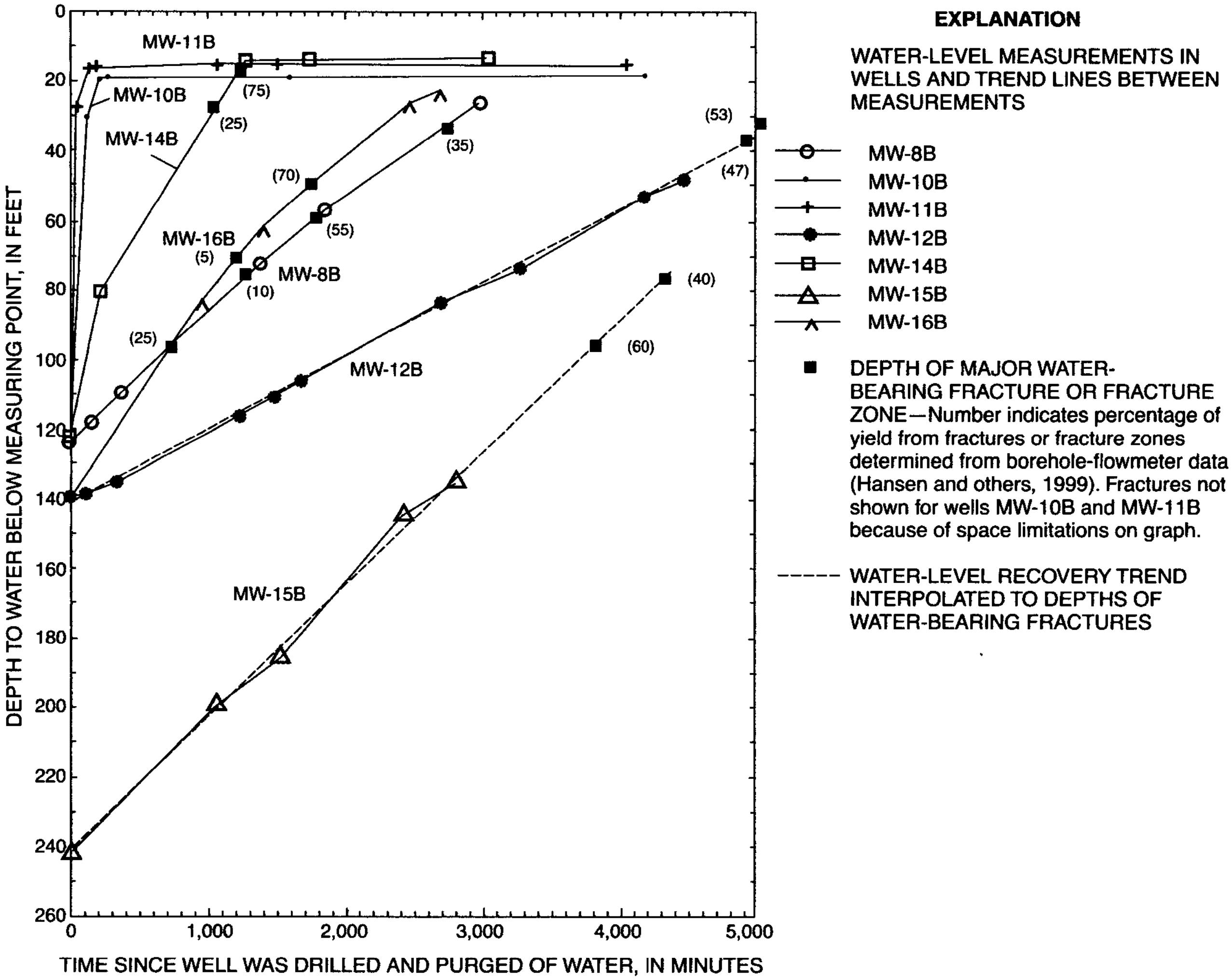


Figure 4. Depth to water during recovery of water levels in selected bedrock wells after drilling and depths of water-bearing fractures or fracture zones.

The relatively high transmissivity values determined for wells MW-10B, MW-11B, MW-14B, and the Van Wart well, can be attributed to the presence of one or more water-yielding fractures or fracture zones (Hansen and others, 1999). The high transmissivity measured in well MW-3B may be attributable to fractures in shallow bedrock that provide a hydraulic connection to the surficial aquifer and to Meddybemps Lake, but fracture data were not available for this well.

Estimates of transmissivity and hydraulic conductivity for surficial materials that were made on the basis of specific-capacity data are summarized in table 3. The hydraulic conductivity values reported in table 3 were determined by dividing the transmissivity of each well by the saturated thickness of the aquifer at the time of the tests. This approach for estimating hydraulic conductivity requires the assumption that the full saturated thickness of the aquifer contributed water to the well during pumping. This assumption is reasonable if the saturated thickness is approximately the same as the length of the well screen, and for wells completed in coarse-grained materials. The assumption may yield values of hydraulic conductivity that are somewhat lower than actual values in fine-grained materials if vertical hydraulic conductivities are relatively low, and if the screen length is considerably less than the saturated thickness. The hydraulic conductivity estimates in table 3 for wells MW-8S, MW-15S, and MW-16S may be lower than actual values for this reason.

Table 2. Estimates of transmissivity for bedrock using specific-capacity data for wells, Meddybemps, Maine

[o, length of open borehole; s, length of screen; ft^2/d, feet squared per day; ft, feet]

Well name	Transmissivity range (ft^2/d)	Number of tests	Length of screen or open borehole (ft)
MW-1B	1.3–2.3	2	s 15
MW-3B	60–130	4	s 10
MW-4B	<1.4	1	o 15
MW-7B	0.1–0.5	2	o 97
MW-8B	0.2	1	o 98
MW-10B	8.0–15	3	o 94
MW-11B	70–110	4	o 94
MW-12B	0.3	1	o 110
MW-14B	20–32	3	o 111
MW-15B	0.09	1	o 193
MW-16B	0.2	1	o 98
MW-22B	280–550	2	o 23
Van Wart	12	1	o 103

Table 3. Estimates of transmissivity and hydraulic conductivity for surficial materials using specific-capacity data for wells, Meddybemps, Maine

[ft, foot; ft/d, feet per day; ft^2/d, feet squared per day]

Well name	Transmissivity range (ft^2/d)	Number of tests	Screen length (ft)	Saturated thickness range (ft)	Hydraulic conductivity range (ft/d)
MW-4S	1.5–5.2	3	5.0	5.5–9.2	0.3–0.6
MW-5S	150–200	2	3.0	6.2–8.6	17–32
MW-7S	230–290	2	5.0	3.7–3.7	63–78
MW-8S	3.9	1	2.5	7.4	0.5
MW-10S	210–220	2	5	5.0–7.2	29–45
MW-13S	50	1	2.5	2.7	19
MW-15S	7.6–19	3	10	18.4–23.7	0.3–1.0
MW-16S	2.9–8.0	3	10	23.1–30.6	0.1–0.3
MW-17S	11–18	2	2.5	3.6–6.7	2.7–3.2
MW-18S	120	1	2.5	5.7	20

Relatively high values of transmissivity and hydraulic conductivity for surficial materials were measured in wells MW-5S, MW-7S, MW-10S, and MW-18S. Estimates of hydraulic conductivity that range from 17 to 78 ft/d in these wells reflect the likely range of hydraulic properties for coarse-grained glaciomarine sediments and is characteristic of silty sand and clean sand (Freeze and Cherry, 1979, table 2.2). Hydraulic conductivity values of 0.1 to 1 ft/d in wells MW-4S, MW-8S, MW-15S, and MW-16S reflect the likely range of hydraulic conductivity for till. This range is within the range of values for tills derived from crystalline rocks in southern New England and northern New Hampshire (Torak, 1979: Pietras, 1981; Melvin and others, 1992, table 3; Tiedeman and others, 1997, p.8).

Specific-capacity data are not available for wells MW-9S, MW-12S, MW-19S, and MW-20S because they were dry or nearly dry and were not sampled. Specific-capacity data are not available for well MW-11S because the yield was too low to sustain pumping during sampling.

The major uncertainty associated with the specific-capacity approach for estimating aquifer transmissivity is the storage coefficient. The computed values of transmissivity, however, are relatively insensitive to the storage coefficient because the coefficient appears in the log term of equation 1. For example, for well MW-12B completed in bedrock, a reduction of the storage coefficient from 0.0001 to 0.00001 increases the transmissivity from 0.16 to 0.21 ft^2/d. For well MW-7S in the surficial aquifer, increasing the storage coefficient from 0.1 to 0.2 decreases the transmissivity from 230 ft^2/d to 207 ft^2/d, and decreasing the storage coefficient from 0.1 to 0.05 increases the transmissivity to 250 ft^2/d. For the low-yielding wells, the time required for the inflow rate to stabilize also is uncertain, but, as with the storage coefficient, the transmissivity estimates are not particularly sensitive to time because time also appears in the log term of equation 1. For example, for well MW-12B, a reduction of the time from 100 min to 10 min in equation 1 reduces the computed transmissivity of the highest-yielding fracture from 0.16 ft^2/d to 0.11 ft^2/d.

Aquifer Tests

Aquifer tests were conducted at wells MW-22B and MW-11B during September 10–14, 1997, to refine estimates of hydraulic properties for the ground-water system. Water-level responses to pumping were observed in these two wells and the Smith and Van Wart wells, which were pumped intermittently for domestic purposes during the aquifer-test period (fig. 5). Water levels in wells that responded to pumping are shown in figure 6. Water levels in the Dennys River, Meddybemps Lake, and wells prior to aquifer testing are summarized in table 1.

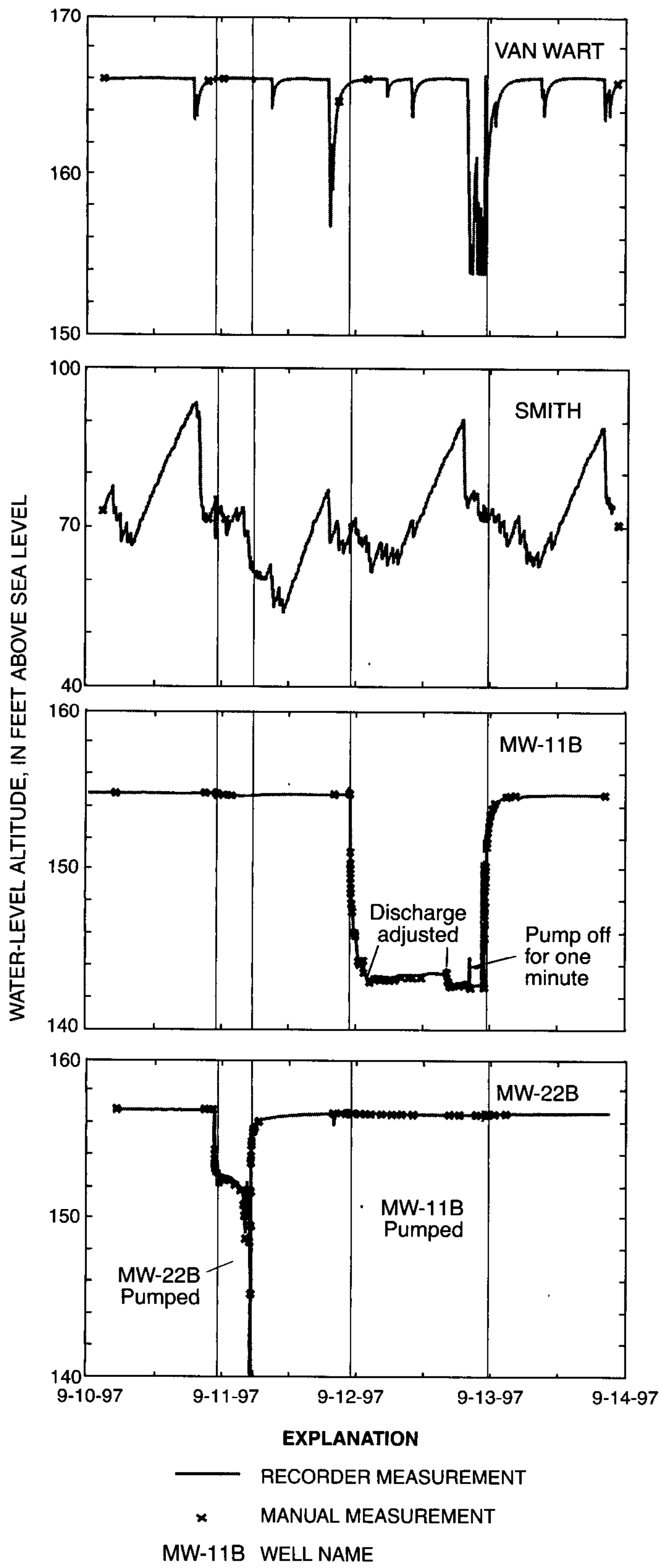


Figure 5. Water-level hydrographs for wells pumped during aquifer testing, Meddybemps, Maine.

The original plan for aquifer testing was to pump well MW-22B at a constant rate for at least 24 hours. After about 6 hours of pumping at a rate of 9 gal/min, however, drawdown suddenly increased and the water level quickly fell to the pump intake. During part of the pumping period, turbidity in the water caused by fine sand and silt indicated that the well may not be fully cased through surficial materials or that fractures in bedrock provided a short connection to the surficial aquifer. The test was terminated after 325 minutes, and water levels were allowed to recover for about 18 hours before a second test was started in well MW-11B.

Water-level data for the first 150 minutes of the test in well MW-22B were used to estimate aquifer transmissivity by applying the straight-line method of Cooper and Jacob (1946). Drawdown during this part of the test followed an approximate straight-line trend on a semi-logarithmic plot, as shown in figure 7, before the well started producing sand and silt and the water level started declining at a greater rate. A transmissivity value of 450 ft^2/d estimated from the time and drawdown data is consistent with estimates using specific-capacity data (table 2). This estimate of transmissivity may be high relative to other wells because it may represent a combination of the surficial and bedrock aquifers.

For the second test, well MW-11B was pumped at a rate of 4.5 gal/min for 24 hours. A uniform pumping rate was difficult to maintain, and the small variations in pumping rate were apparent in the water-level record. Water levels in all wells monitored south of Route 191, except well MW-8B, responded to pumping from well MW-11B (fig. 5). A rise of water level in well MW-8B shortly after pumping started resulted from an estimated 0.1 to 0.2 in. of precipitation that entered the uncovered well during a rain shower.

Water-level data for the first 200 min of the test in well MW-11B (fig. 7) were used to estimate transmissivity by applying the straight-line method of Cooper and Jacob (1946). During this period, the effects of leakage and well-bore storage were assumed to be negligible; however, both factors could have affected the analysis. A transmissivity of 38 ft^2/d calculated by this method is lower than estimates using specific-capacity data (70–110 ft^2/d in table 2), possibly because well-bore storage affected the rate of drawdown during the first hour of pumping. The estimates of transmissivity for this well were refined using numerical methods discussed in the next section.

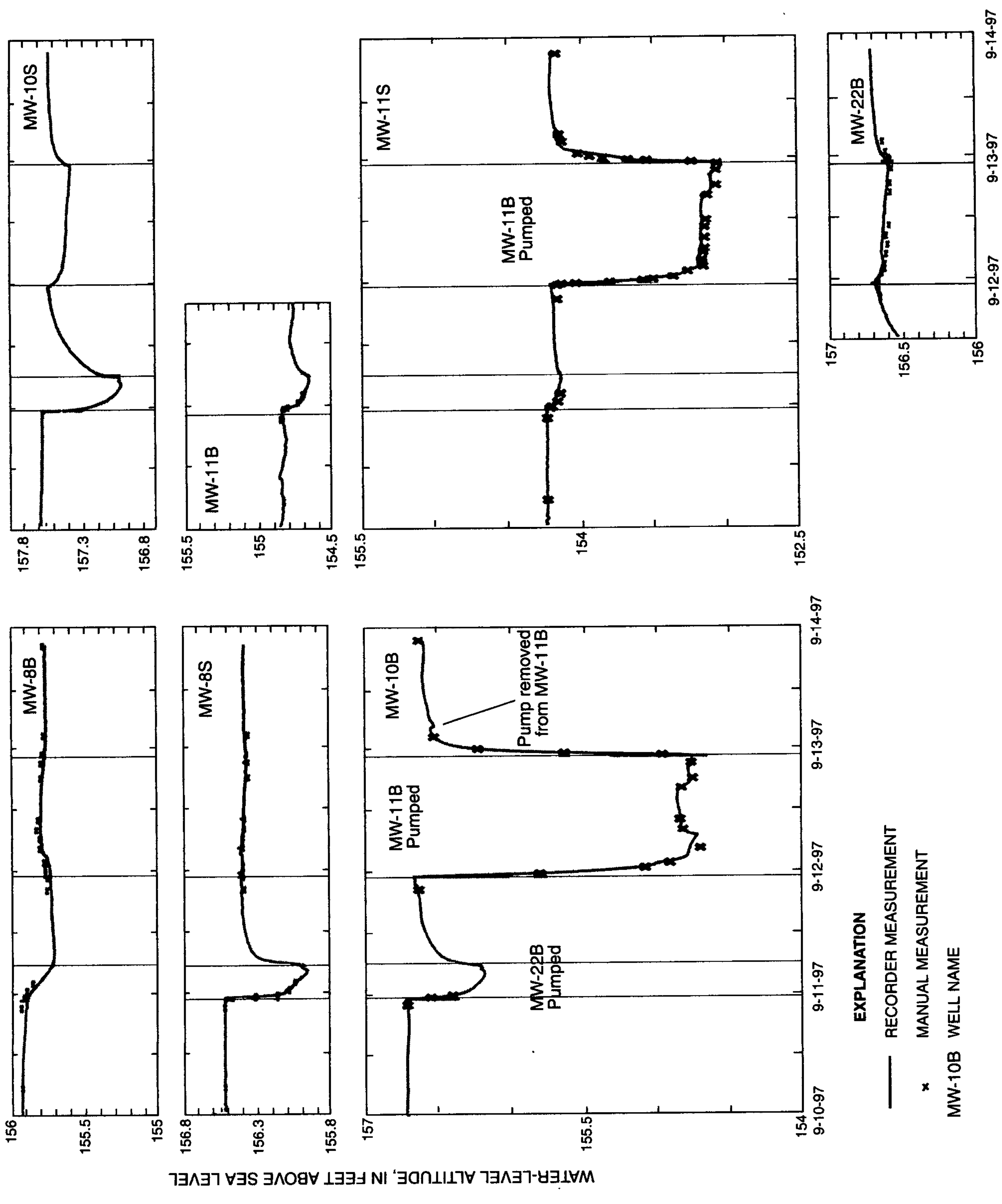
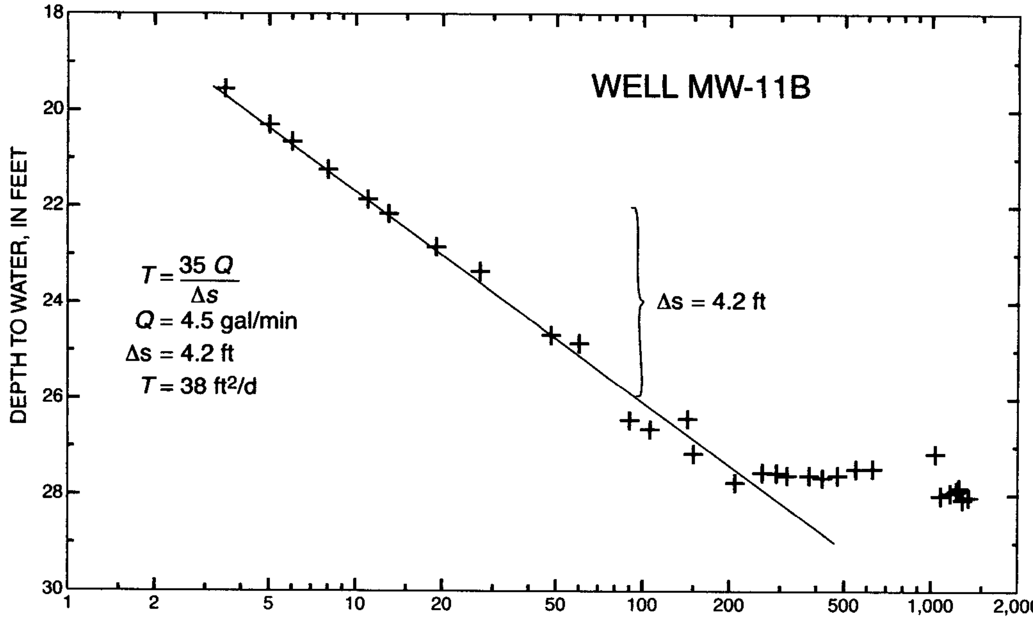
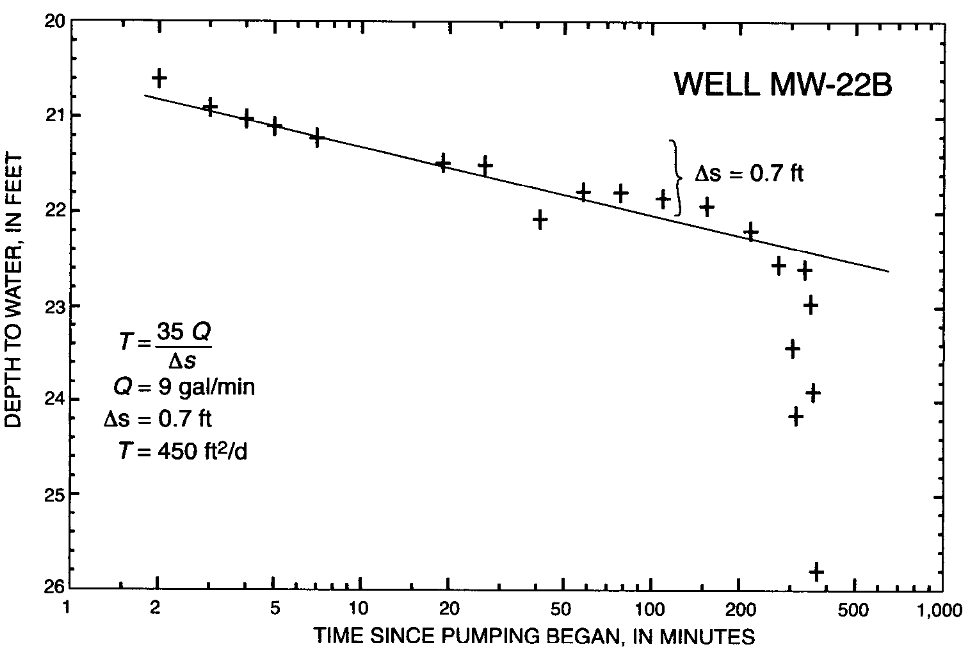


Figure 6. Water-level hydrographs for observation wells that responded to pumping during aquifer testing, Meddybemps, Maine.





EXPLANATION

+ MANUAL WATER-LEVEL MEASUREMENT

——— LINE USED FOR COMPUTATION OF TRANSMISSIVITY

Q PUMPING RATE, IN GALLONS PER MINUTE (gal/min)

Δs CHANGE IN DRAWDOWN, IN FEET (ft) OVER ONE LOG CYCLE

T TRANSMISSIVITY, IN FEET SQUARED PER DAY (ft^2/d)

Figure 7. Depth to water in wells MW-11B and MW-22B during pumping and calculations of transmissivity, Meddybemps, Maine.