ESTIMATION OF AQUIFER HYDRAULIC PROPERTIES

Hydraulic conductivity and storage are aquifer properties that may vary spatially because of geologic heterogeneity. Estimation of these properties allows quantitative prediction of the hydraulic response of the aquifer to recharge and pumping. Storage coefficients are important for understanding hydraulic response to transient stresses on aquifers. These properties can be estimated on a local scale by analysis of data from aquifer tests, such as single-well or multiple-well aquifer tests, or on a regional scale by a numerical simulation of ground-water flow by use of a computer-based model. The local scale ranges from tens of feet to hundreds of feet. The regional scale is characterized by lengths of hundreds to thousands of feet. Transmissivity, the hydraulic conductivity multiplied by the saturated thickness of the aquifer, represents a vertical average of hydraulic conductivities that may vary with depth. Most of the analytical techniques used to estimate the hydraulic properties of aquifers were developed for porous media, such as unconsolidated sediments. These techniques may provide reasonable estimates of hydraulic properties in fractured rocks, however, when the hydraulic response of the fractured-rock aquifers approximates porous media at the scale of interest. In this report, the regional-scale flow model assumes steady-state conditions, hence the storage coefficient cannot be estimated from it.

Aquifer Tests

As part of this study, several types of aquifer tests were conducted by the USGS and others in the Lansdale area since 1995. At each of three sites, both a single well, aquifer-interval-isolation test in one borehole and a multiple-well test (single pumping well and multiple observation wells) were done by USGS. At a fourth site (J.W. Rex), single-well, interval-isolation tests in two wells and a multi-well test were done by a private contractor for the property owner (QST Environmental, Inc., 1998). In addition, specific-capacity data are available for wells pumped during ground-water sampling done by the USEPA contractor (Lusheng Yan, Black & Veatch Waste Science, Inc., written commun., 1997). This report presents in detail the tests done by USGS and briefly discusses tests done by others.

In a review of aquifer-test data collected prior to this study (pre-1995), Goode and Senior (1998) summarized the range of estimated transmissivity and storage coefficients. Estimates of transmissivity ranged from 0 to about 5,400 ft^2/d (0 to 500 m^2/d); estimates from most tests ranged from 108 to 1,080 ft^2/d (10 to 100 m^2/d). Estimates of storage coefficients ranged from 0.00001 to 0.26; most estimates ranged from 0.0001 to 0.007.

Single-Well, Interval-Isolation Tests

Water enters open-hole wells through discrete openings or zones in fractured-rock aquifers. Most ground-water flow and contaminant movement at the site is through distinct water-bearing zones consisting of one or more fracture(s), and the hydraulic and chemical characteristics of each water-bearing zone can differ. By isolating these discrete zones with inflatable packers, hydraulic properties of individual zones and the extent of vertical hydraulic connection between zones can be determined. This determination provides data on the vertical distribution of hydraulic properties.

The USGS performed single-well, aquifer-interval-isolation tests in three wells known to yield water containing VOC’s and near known sources of soil contamination. The wells were Mg-80 (at Keystone Hydraulics), Mg-1443 (at Philadelphia Toboggan), and Mg-1444 (at Rogers Mechanical) (pl. 1). The objectives of the single-well, interval-isolation tests were to (1) provide information on hydraulic heads and specific capacities of discrete vertical intervals and the hydraulic connection between intervals, and (2) provide water samples from discrete water-bearing zones to allow the USEPA to characterize the vertical extent of contamination in each well. Similar single-well, aquifer-interval-isolation tests were done in two wells, Mg-624 and Mg-1639, at the J.W. Rex property by QST Environmental, Inc.

Packers were set to isolate selected water-bearing (producing or receiving) zones. The number and depths of intervals to be tested in each open-hole well were based on an analysis of the borehole geophysical logs. A straddle packer was used to isolate three intervals and a single packer was used to isolate two intervals in the open-hole wells. When inflated, the rubber bladder of each packer acts as a plug sealing off 4 ft (1.2 m) of the borehole between two zones. Water levels in each isolated zone were measured before and after packer inflation by use of electric tapes. The reference measuring point for water levels and all logged depths was land surface. When possible, water levels also were measured during pumping by use of pressure transducers; drawdowns were recorded at a specified change in water level [0.1 ft (.03 m)]. Pumping duration was approximately 1 to 2 hours; rates ranged from about 0.2 to 4 gal/min (0.76 to 15 L/min) for each test.

Specific capacity and transmissivity for each isolated zone were calculated. These results are compared to additional data, where available, on specific capacities of the open-hole wells determined from pumping rates and drawdowns during pumping for open-hole tests (Conger, 1999; Black & Veatch Waste Science Inc., 1998). The transmissivity (T) was calculated by use of the Thiem equation (Bear, 1979), assuming steady-state conditions, as follows:

$$T = \frac{Q}{2\pi\Delta h}\ln\frac{R}{r_w}, \qquad (1)$$

where Q is pumping rate,

Δh is change in head,

R is radius of influence of pumping, and

r_w is radius of well.

For analysis of data from single-well, interval-isolation tests at the three wells (Mg-80, Mg-1443, and Mg-1444), R was assumed to equal 328 ft (100 m). This method of estimating transmissivity is similar to that used by Shapiro and Hsieh (1998) for short-term, low-injection-rate, single-well, interval-isolation tests in low-permeability fractured rocks. For the tests by Shapiro and Hsieh (1998), R was assumed to equal 9.8 ft (3 m). The rate and duration of pumping of tests for the present study were greater than in the tests by Shapiro and Hsieh (1998), and it is reasonable to assume that R would be greater than 9.8 ft (3 m).

Single-well, interval-isolation aquifer tests at three wells in Lansdale (Mg-80, Mg-1444, Mg-1443) generally indicate that (1) discrete water-bearing openings are not well connected in the vertical direction and (2) specific capacity and estimated transmissivity ranged over two to three orders of magnitude in the water-bearing zones tested. No relation between depth and specific capacity or estimated transmissivity was noted in the results of tests of isolated zones in the three wells, Evidence for limited vertical hydraulic connection between water-bearing openings includes differences in static potentiometric head up to 15 ft (46 m) over 300 vertical ft (91 m) and typically small drawdown in zones adjacent to the isolated pumped zone.

The chemical and physical properties of borehole discharge were measured at various times during pumping by the USGS by the use of temperature-compensated pH and specific-conductance meters. After physical and chemical properties stabilized or after three test-interval volumes of borehole water were pumped, water samples for measurement of pH, specific conductance, temperature, and dissolved oxygen concentration were collected. Samples for VOC analysis then were collected by the USGS and forwarded to USEPA’s contractor, B&V, for analysis. In single-well, aquifer-interval-isolation tests by QST Environmental, Inc., in wells Mg-624 and Mg-1639, the USGS measured chemical and physical properties and QST Environmental, Inc., collected samples for VOC analysis. The pH and specific conductance were measured by methods outlined in Wood (1976). Dissolved oxygen was measured by use of the azide modification of the Winkler titration method (American Public Health Association and others, 1976).

Well Mg-80

The open-hole well is about 270 ft (82.3 m) in depth with a few feet of soft sediment at the bottom of the well. An 8-in. (0.2-m) diameter casing extends to a depth of 138 ft bls (42.1 m). Geophysical logging (Conger, 1999) indicated water-bearing zones at 144-154 ft bls (43.4-46.9 m) and 253-258 ft bls (77.1-78.6 m) (fig. 24). Under non-pumping conditions, upward flow in the borehole was measured with inflow from fractures at 253-258 ft bls (77.1-78.6 m) and outflow through fractures at 144-154 ft bls (43.4-46.9 m). The flow pattern indicated a difference in hydraulic heads in the well. When the open-hole well was pumped at a rate of about 1 gal/min (3.785 L/min) in summer 1996, the fractures at 144-154 ft bls (43.4-46.9 m) produced most of the fluid.

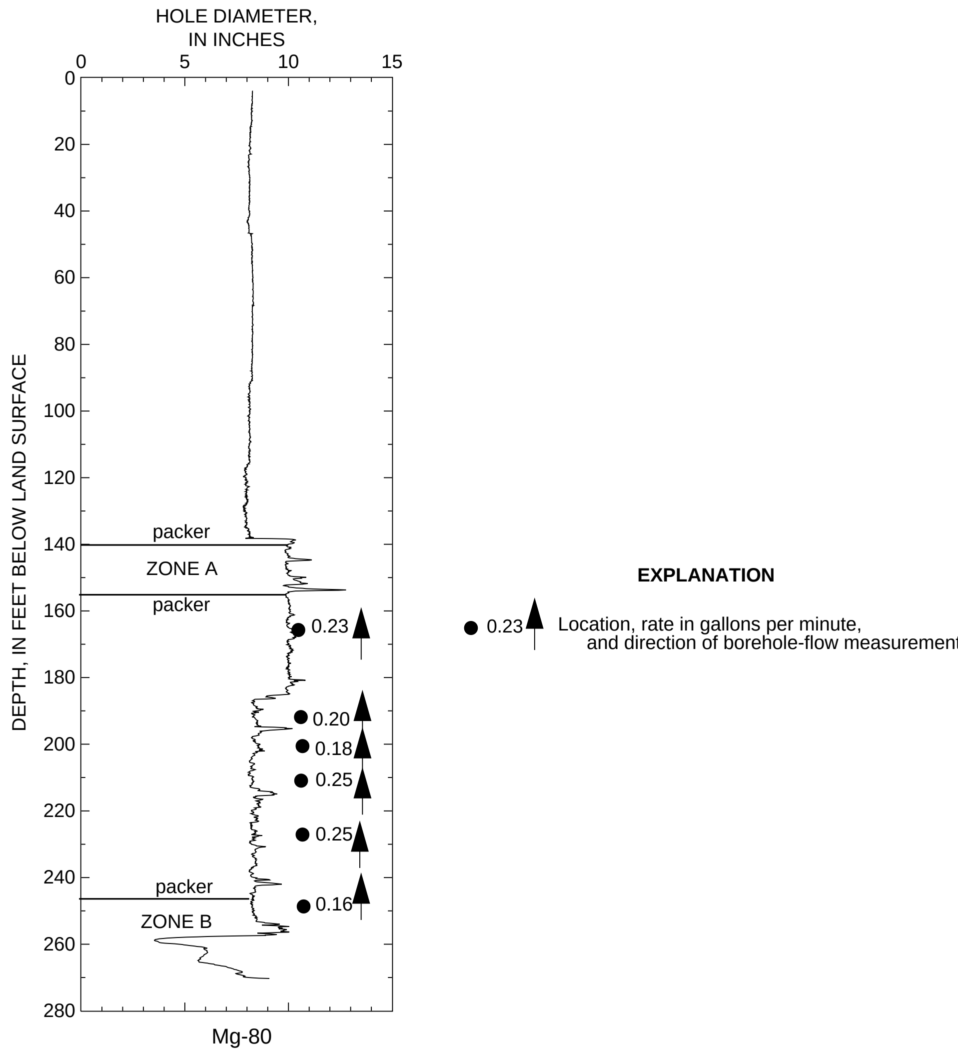

Figure 24. Depth of packers for aquifer-interval-isolation tests and direction of nonpumping flow in well Mg-80 in Lansdale, Pa.

Tests in well Mg-80 were conducted on March 24-27, 1997. Packers isolated two intervals (fig. 24) for testing, including below 246 ft bls (75 m) (zone B) and 142-157 ft bls (43.3-47.8 m) (zone A). Depth to water in the open borehole was 12.43 ft bls (3.79 m). After packer inflation, water levels were measured above, in, and for zone A below the isolated intervals. Water levels in isolated intervals stabilized in about 15 minutes after packer inflation. In test of zone A, the isolated interval was pumped at about 2 gal/min (7.6 L/min), and drawdown was observed in all three intervals (fig. 25, table 7). The observed drawdowns indicate either the packers did not isolate the interval (seal the borehole) effectively or the intervals are connected outside of the well. In the test of zone B, a single packer was placed at 246 ft bls (75 m) and the pump was placed below the packer. Drawdown was observed only in the pumped zone (fig. 26, table 7). These results indicate that the zone below 246 ft bls (75 m) is hydraulically isolated from water-bearing zones above that depth. In the test of zone A, a straddle packer with a 15-ft (4.6-m) spacing between center of packers was used to isolate the interval of 142-157 ft bls (43.3-47.8 m). The water level in the isolated interval was slightly higher than in the upper or lower intervals after packer inflation (table 7).

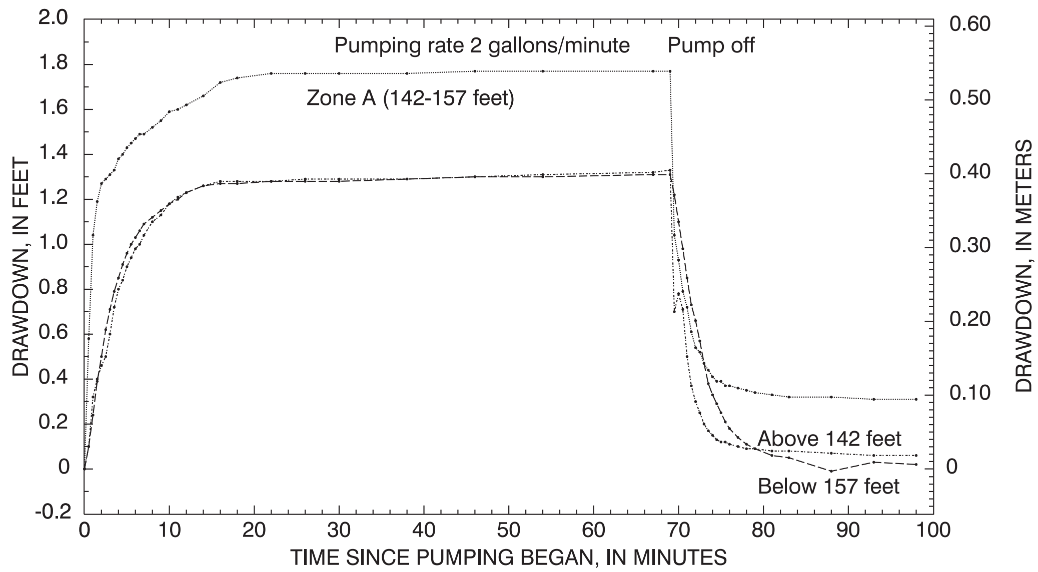

Figure 25. Drawdown as a function of time in aquifer-interval-isolation test of zone A in well Mg-80 in Lansdale, Pa., March 26, 1997.

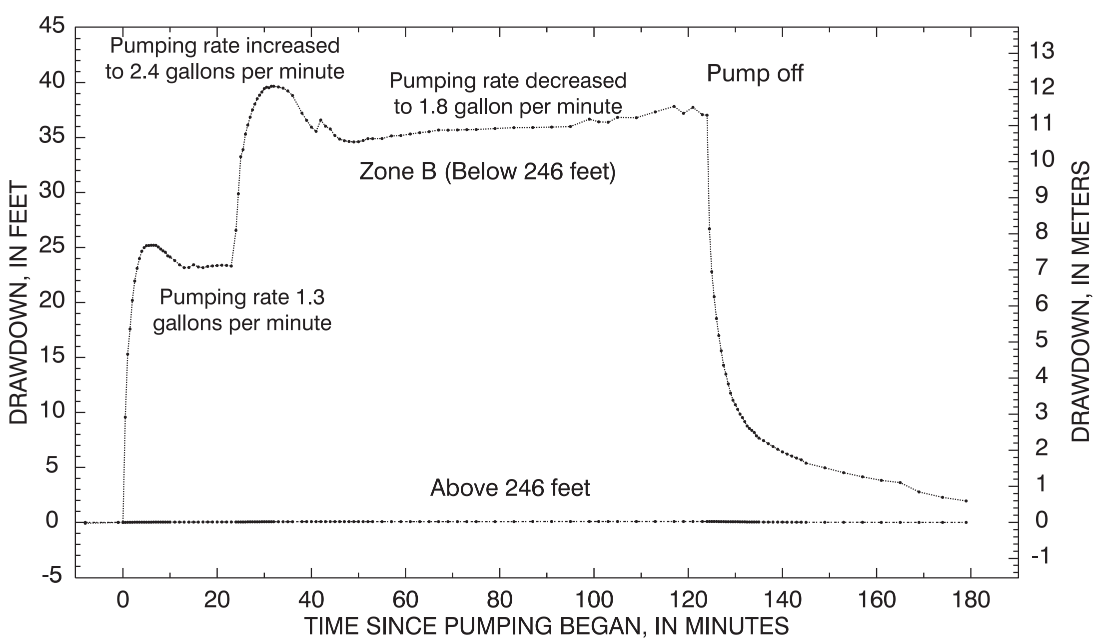

Figure 26. Drawdown as a function of time in aquifer-interval-isolation test of zone B in well Mg-80 in Lansdale, Pa., March 27, 1997.

The interval between 142-157 ft bls (43.3-47.8 m) has a greater specific capacity than the interval below 246 ft bls (75 m). These specific-capacity measurements are consistent with the heatpulse-flowmeter measurements that indicated fractures in the upper zone produced most water when the open well was pumped (Conger, 1999). The calculated specific capacity for the zone A (table 7) in this borehole probably is greater than actual specific capacity for the zone because of contribution from other intervals. The sum of specific capacities determined for isolated zones A and B is similar or somewhat less than the specific capacity determined for the open-hole tests (table 7).

Table 7. *Depths, water levels, specific capacity, and transmissivity of aquifer intervals isolated by packers and of the open hole for well Mg-80 in Lansdale, Pa., March 1997, May 1996, and September 1997*

[ft bls, feet below land surface; ft, feet; gal/min, gallons per minute; min, minutes; (gal/min)/ft, gallons per minute per foot; ft^2/d, square feet per day; NA, not applicable]

Depth of isolated intervals (ft bls)	Date of test	Pre-pumping depth to water in interval[1] (ft bls)	Depth to water in interval at end of test[2] (ft bls)	Drawdown at end of test (ft)	Pumping rate (gal/min)	Pumping duration (min)	Specific capacity [(gal/min)/ft]	Trans-missivity[3] (ft^2/d)
				Zone A (142-157 ft bls)				
Open hole	3-26-97	12.43	NA	NA	NA	NA	NA	NA
Above 142	3-26-97	11.93	13.26	1.33	NA	NA	NA	NA
142-157 (pumped)	3-26-97	11.88	13.65	1.77	2	69	[4] 1.13	[5] 238
Below 157	3-26-97	12.03	13.34	1.31	NA	NA	NA	NA
				Zone B (below 246 ft bls)				
Above 246	3-27-97	12.11	12.19	.08	NA	NA	NA	NA
Below 246 (pumped)	3-27-97	12.07	49.10	37.03	1.8	124	.037	10.2
Sum of specific capacities or transmissivities for intervals tested							1.17	248
				Open-hole tests				
Open hole	5-23-97	13.29	13.8	.51	1	79	1.96	413
Open hole	9-30-97	15.2	25.78	10.58	12	65	1.13	239

[1] Stabilized water levels after packers were inflated.
[2] Depth to water at end of pumping at a constant rate before the pump was shut off.
[3] Calculated using Thiem equation, assuming a radius of influence, r_0, of 328 feet (100 meters).
[4] Measured specific capacity for zone greater than actual specific capacity because of contributions of flow from other intervals.
[5] Calculated transmissivity for zone greater than actual transmissivity because of contributions of flow from other intervals.

Well Mg-1443

The caliper log indicated fractures at 35-41 ft bls (10.7-12.5 m), 104-106 ft bls (31.7-32.3 m), 175-178 ft bls (53.3-54.3 m), and 289-291 ft bls (88.1-88.7 m) in the 339-ft (103.3-m) deep, 8-in.- (0.2 m) diameter borehole (fig. 27). When the open-hole well was pumped at a rate of about 1 gal/min (3.785 L/min) in summer 1996, the fractures at 289-291 ft bls (88.1-88.7 m) appeared to produce most of the water and fractures at 104-106 ft bls (31.7-32.3 m) produced the second greatest amount (Conger, 1999). Under nonpumping conditions in summer 1996, minor upward flow was measured between the depths of 332 ft bls (101.2 m) and 68 ft bls (20.7 m) (Conger, 1999). This flow pattern indicates a difference in hydraulic heads between water-bearing zones in the borehole.

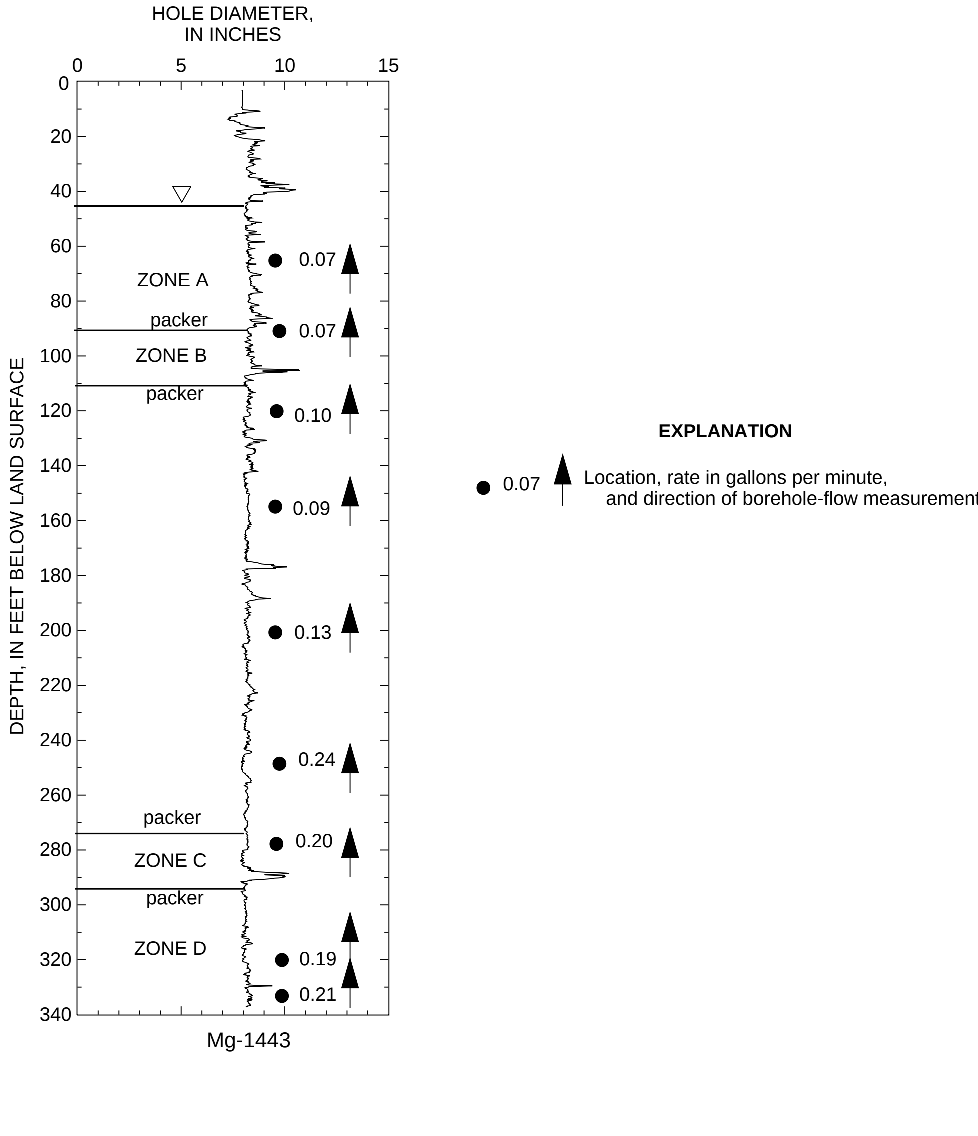

Figure 27. Depth of packers for aquifer-interval-isolation tests and direction of nonpumping flow in well Mg-1443 in Lansdale, Pa.

Tests in well Mg-1443 were conducted on April 9-11, 1997. On the basis of results of geophysical logging, four intervals were selected for testing (fig. 27) including below 296 ft bls (90.2 m) (zone D); 276-296 ft bls (84.1-90.2 m) (zone C); 90.5-110.5 ft bls (27.6-33.7 m) (zone B); and above 90.5 ft bls (27.6 m) (zone A).

In the test of zone A, the pre-pumping level in the pumped zone was about 2.4 ft (0.73 m) higher than the level in the interval immediately below (90.5-110.5 ft), indicating a downward vertical gradient between these intervals. The pre-pumping level in zone A was about 1 ft (0.3 m) lower than the interval below 110.5 ft, indicating an upward gradient between these intervals. Because testing of zone A was done soon after testing of zone B, water levels may not have fully recovered from the test of zone B. When zone A was pumped, drawdown was measured in the interval between 90.5 and 110.5 ft (27.6-33.7 m) but not in the interval below 110.5 ft (33.7 m) (fig. 28).

In the test of zone B, the pre-pumping water level in the isolated interval was almost equal to the level in the overlying interval and 0.52 ft (0.16 m) lower than the level in the underlying interval zone; the latter head difference was similar to the head difference [0.36 ft (0.11 m)] between the isolated zone C and the interval above zone C (table 8). When zone B was pumped, no drawdown was measured in the underlying interval, and about 1 ft (0.3 m) of drawdown was measured in the overlying interval (fig. 29), indicating some hydraulic connection between zone B and the interval above zone B.

In the test of zone C, the water level in the isolated interval before pumping was 4.79 ft (1.46 m) lower than the level in the underlying interval and 0.56 ft (0.17 m) higher than the level in the overlying interval, also indicating an upward vertical gradient. When pumped, small but measurable drawdown in intervals above and below zone C were observed (fig. 30), suggesting an incomplete seal by packers or hydraulic connection outside the borehole.

In the test of zone D, the water level in the isolated interval before pumping was 9.07 ft (2.76 m) higher than in the interval above 296 ft bls (90.2 m), indicating an upward vertical gradient. When zone D was pumped at a rate of about 0.2 gal/min (0.76 L/min), a large drawdown was observed in the pumped interval and very little drawdown was observed in the overlying interval (fig. 31). Zone D appeared to be hydraulically isolated from other intervals and to produce little water. Thus, water-bearing zones near the bottom of the well appear hydraulically isolated from the water-bearing zones near the top of the well.

The calculated specific capacities for zones A and C are lower than the specific capacity of zone B (table 8), which is consistent with the relative yields of these zones determined by heatpulse-flowmeter measurements while pumping (Conger, 1999). The specific capacity of zone D determined from the isolated-interval tests is probably higher than the actual specific capacity. In addition to the apparent hydraulic connection between zone D and adjacent intervals, the short duration of pumping and variable pumping rates may have affected the test. Specific capacity commonly tends to decrease with increases in pumping time. The sum of specific capacities of individual isolated zones is greater than the specific capacity determined for the open borehole in summer 1996 (Conger, 1999), possibly because of the over-estimated specific capacity of zone D (table 8).

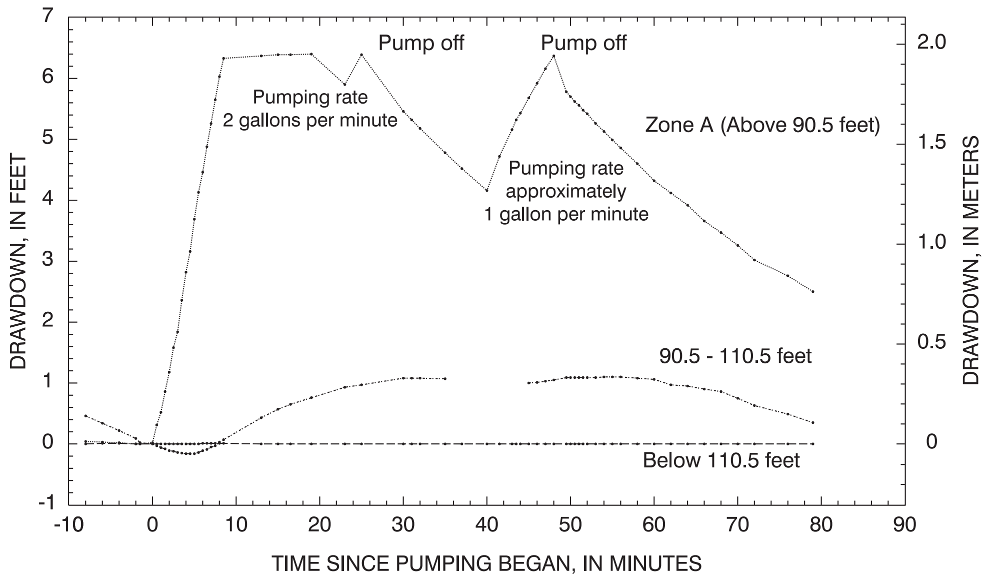

Figure 28. Drawdown as a function of time in aquifer-interval-isolation test of zone A of borehole Mg-1443 in Lansdale, Pa., April 11, 1997.

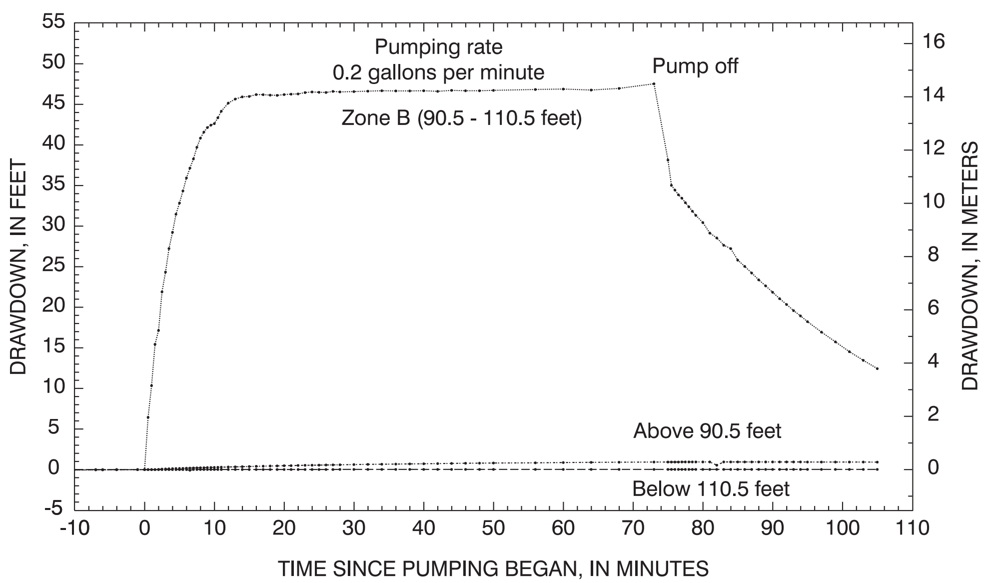

Figure 29. Drawdown as a function of time in aquifer interval-isolation test of zone B of borehole Mg-1443 in Lansdale, Pa., April 11, 1997.

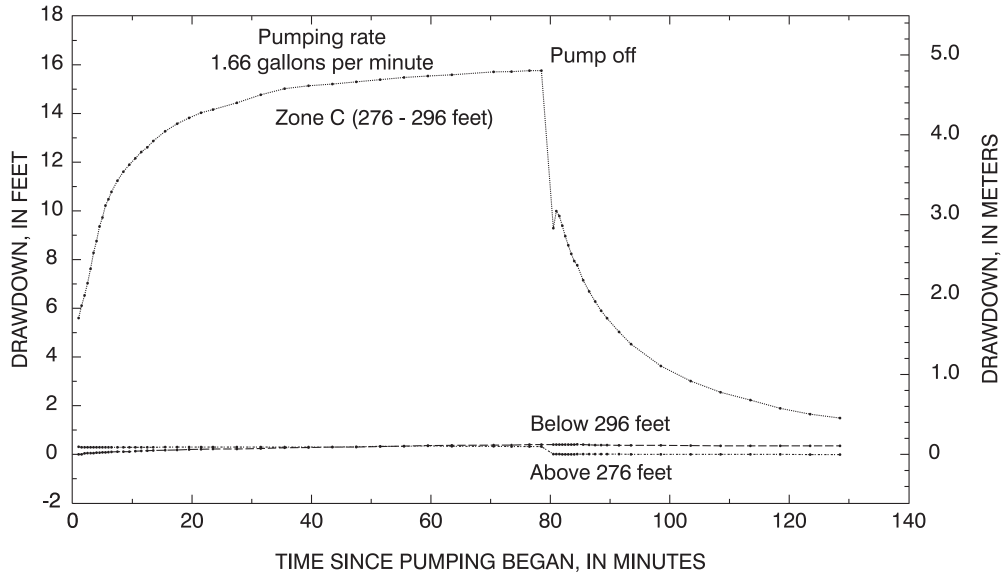

Figure 30. Drawdown as a function of time in aquifer-interval-isolation test of zone C of borehole Mg-1443 in Lansdale, Pa., April 10, 1997.

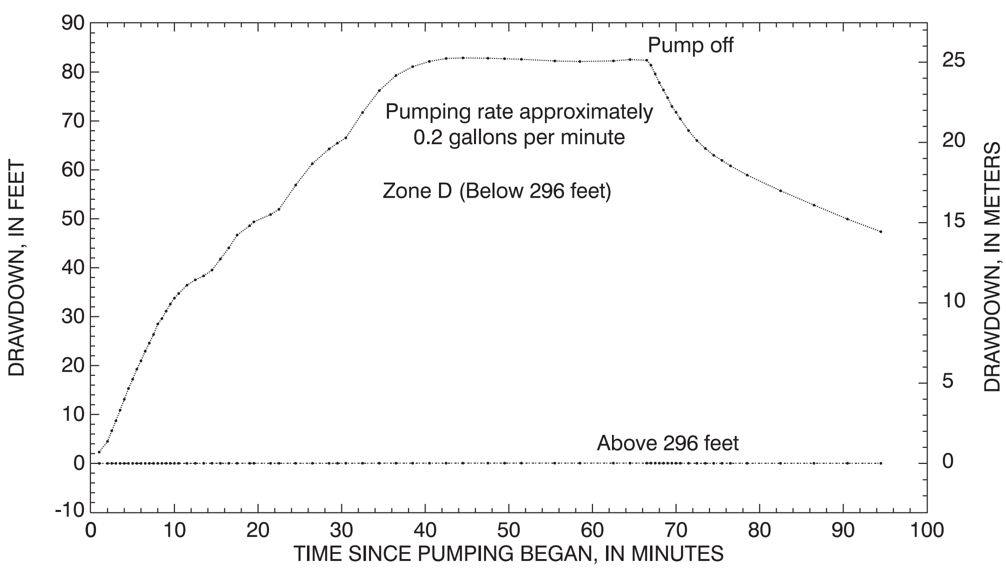

Figure 31. Drawdown as a function of time in aquifer-interval-isolation test of zone D of borehole Mg-1443 in Lansdale, Pa., April 9, 1997.

Table 8. *Depths, water levels, specific capacity, and transmissivity of aquifer intervals isolated by packers and of the open hole for well Mg-1443 in Lansdale, Pa., April 1997, May 1996, and October 1997*

[ft bls, feet below land surface; ft, feet; gal/min, gallons per minute; min, minutes; (gal/min)/ft, gallons per minute per foot; NA, not applicable]

Depth of isolated interval (ft bls)	Date of test	Pre-pumping depth to water in zone[1] (ft bls)	Depth to water in zone at end of test[2] (ft bls)	Drawdown at end of test (ft)	Pumping rate (gal/min)	Pumping duration (min)	Specific capacity [(gal/min/ft]	Trans-missivity[3] (ft^2/d)
			Zone A (above 90.5 ft bls)					
Above 90.5 (pumped)	4-11-97	42.90	49.27	6.37	[4] 1	21	[5] 0.16	[6] 34.4
90.5 - 110.5	4-11-97	45.29	46.34	1.05	NA	NA	NA	NA
Below 110.5	4-11-97	41.91	41.91	0	NA	NA	NA	NA
			Zone B (90.5-110.5 ft bls)					
Above 90.5	4-11-97	42.39	43.32	.93	NA	NA	NA	NA
90.5 - 110.5 (pumped)	4-11-97	42.41	89.95	47.54	.2	73	.004	.86
Below 110.5	4-11-97	41.89	41.91	.02	NA	NA	NA	NA
			Zone C (276-296 ft bls)					
Above 276	4-10-97	42.40	42.72	.32	NA	NA	NA	NA
276 - 296 (pumped)	4-10-97	42.04	57.80	15.76	1.7	78.5	.108	22.6
Below 296	4-10-97	37.25	37.65	.40	NA	NA	NA	NA
			Zone D (below 296 ft bls)					
Above 296	4-9-97	41.95	42.00	.05	NA	NA	NA	NA
Below 296 (pumped)	4-9-97	32.88	115.43	82.55	.2	65	.002	.54
Sum of specific capacities or transmissivities for zones tested							.274	58.4
			Open hole tests					
Open hole	5-22-97	42.09	47.35	[7]5.26	1	98	.19	39.8
Open hole	10-23-97	51.61	94.2	42.59	5.5	150	.13	26.9

[1] Stabilized water levels after packers were inflated.
[2] Depth to water at end of pumping at a constant rate before pump was shut off.
[3] Calculated using Thiem equation, assuming radius of influence, r_0, is 328 feet (100 meters).
[4] Estimated time-weighted average of variable pumping rates ranging from 0.18 to 2.2 gallons/minute.
[5] Calculated specific capacity for zone greater than actual specific capacity because of contributions of flow from other intervals, short duration of pumping, and variable pumping rates.
[6] Calculated transmissivity for zone greater than actual transmissivity because of contributions of flow from other intervals, short duration of pumping, and variable pumping rates.
[7] Drawdown did not stabilize during this test.

Well Mg-1444

Logging of well Mg-1444 identified producing fractures and vertical hydraulic head differences (Conger, 1999). The caliper log indicated major fractures at 70-72 ft bls (21.3-21.9 m), 138-141 ft bls (42.1-43 m), 153 ft bls (46.6 m), 260-265 ft bls (79.2-80.8 m) and numerous minor fractures along the open interval of the 294-ft (89.6-m) deep, 6-in.- (0.15 m) diameter borehole (fig. 32). During heatpulse-flowmeter measurements of the borehole under nonpumping conditions in summer 1996, upward borehole flow of about 1 gal/min (3.785 L/min) was measured, with inflow through fractures below 270 ft bls (82.3 m), at 260-265 ft bls (79.2-80.8 m), and possibly at 138-141 ft bls (42.1-43 m), and outflow through fractures at 70-72 ft bls (21.3-21.9 m). The observed upward flow indicated a difference in hydraulic heads in the borehole.

Tests in well Mg-1444 were conducted on April 3-7, 1997. On the basis of results of geophysical logging, five intervals were selected for testing (fig. 32) including below 268 ft bls (81.7 m) (zone E); 248-269 ft bls (75.6-82 m) (zone D); 136.5-157.5 ft bls (41.6-48 m) (zone C); 64-85 ft bls (19.5-25.9 m) (zone B); and above 64 ft bls (19.5 m) (zone A).

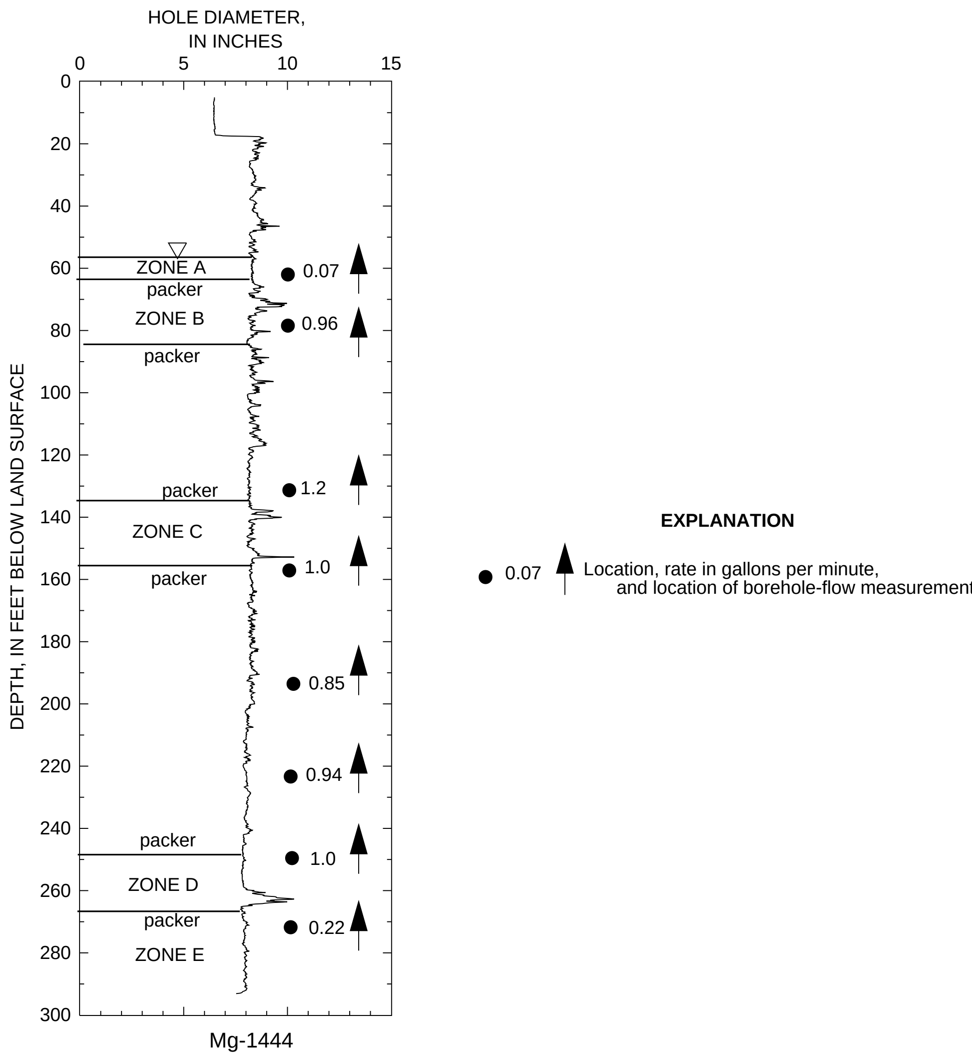

Figure 32. Depth of packers for aquifer-interval-isolation tests and direction of nonpumping flow in well Mg-1444 in Lansdale, Pa.

In the test of zone A, the pre-pumping water level in zone A was 0.28 ft (0.9 m) above the level in the interval between 64-85 ft bls (19.5-25.9 m) and 14.1 ft bls (4.30 m) lower than in the interval below 85 ft bls (25.9 m), similar to head differences measured in the test of zone B. Pumping of zone E was short in duration and at small, variable rates because the zone produced little water and dewatered rapidly. Little drawdown was measured in the interval immediately underlying zone E, and no drawdown was measured in the interval below 85 ft bls (25.9 m) **(fig. 33).**

In the test of zone B, the pre-pumping water level in zone B was 1.01 ft (0.31 m) lower than the level in the overlying interval and 12.12 ft (3.69 m) lower than the level in the underlying interval; these head differences indicate a downward vertical gradient from above and upward vertical gradient from below the isolated interval. Geophysical logging indicated fractures at 70-72 ft bls (21.3-21.9 m) were receiving, consistent with the lower heads measured in zone B compared to adjacent intervals. When zone B was pumped, gradual drawdown of up to 3 ft (0.91 m) in the interval above zone B and minor drawdown in the interval below zone B were measured **(fig. 34).** These results indicate leakage around packers or hydraulic connection outside the borehole between the zone B and the overlying interval and near hydraulic isolation between zone B and the underlying interval.

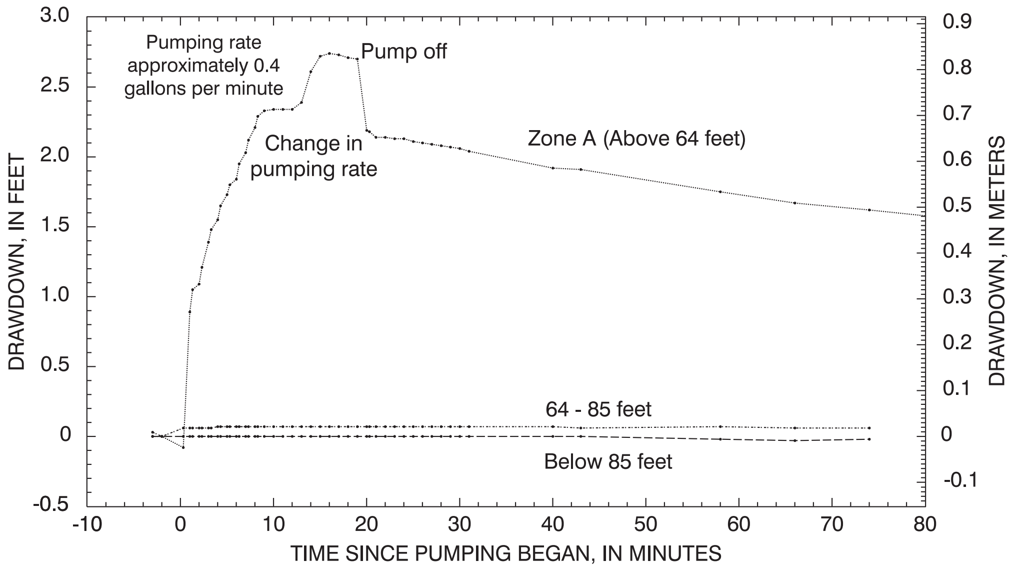

Figure 33. Drawdown as a function of time in aquifer-interval isolation test of zone A of borehole Mg-1444 in Lansdale, Pa., April 7, 1997.

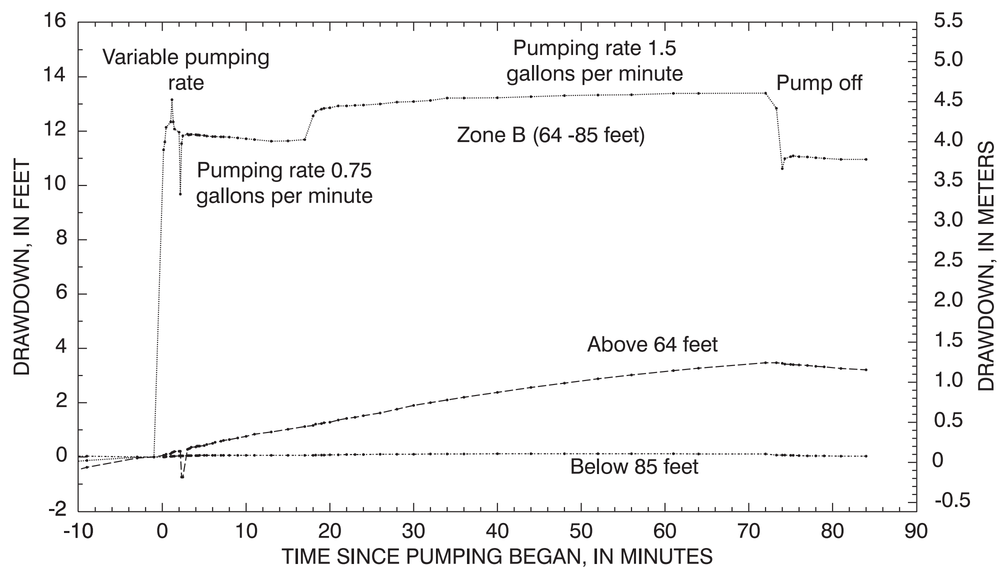

Figure 34. Drawdown as a function of time in aquifer-interval isolation test of zone B of borehole Mg-1444 in Lansdale, Pa., April 4, 1997.

In the test of zone C, the pre-pumping water level in zone C was 16.71 ft (5.09 m) higher than the level in the overlying interval and 1.06 ft (0.32 m) lower than the level in the underlying interval. These head differences are consistent with the upward flow measured with the heatpulse-flowmeter at 160 ft bls (48.8 m) and 130 ft bls (39.6 m) in summer 1996 (Conger, 1999). When zone C was pumped, very little drawdown was measured in the interval above zone C and virtually no drawdown was measured in the interval below zone C (fig. 35), suggesting hydraulic isolation between these intervals.

In the test of zone D, the pre-pumping water level in the isolated interval was 15.35 ft (4.68 m) higher than in the level in the overlying interval and 0.88 ft (0.27 m) higher than the level in the underlying interval. These head differences indicate upward and downward vertical gradients between zone D and adjacent intervals. The upward vertical gradient is consistent with the upward flow measured earlier with the heatpulse flowmeter at and above 256 ft bls (78 m) (Conger, 1999). Drawdown of more than 2 ft (0.61 m) was measured in the interval below zone D when zone D was pumped (fig. 36). These results suggest leakage around packers or a hydraulic connection outside the borehole between the isolated zone D and the underlying interval. In the test of zone D, little drawdown measured in the overlying interval indicates that zone D and the overlying interval were hydraulically isolated.

In the test of zone E, the pre-pumping water level in zone E was 6.45 ft (1.97 m) lower than the level in the overlying interval. Although upward flow was observed during heatpulse-flowmeter measurements in summer 1996, the observed head differences for zone E in April 1997 indicate a downward vertical gradient between the isolated interval and the overlying interval. Drawdown of less than 1 ft was measured in the interval above zone E during pumping of zone E (fig. 37, table 9), suggesting either leakage around packers or a hydraulic connection outside the borehole similar to the test results of zone D.

The total specific capacity of 0.89 (gal/min)/ft [11.1 (L/min)/m] determined from the interval-isolation tests was less than the specific capacity of 1.56 (gal/min)/ft [19.4 (L/min)/m] determined from an open-hole test (table 9). Results of heatpulse-flowmeter measurements in summer 1996 suggest that the zone between 248-269 ft bls (75.6-82 m) is the most productive (Conger, 1999), which is consistent with the results of the interval-isolation tests.

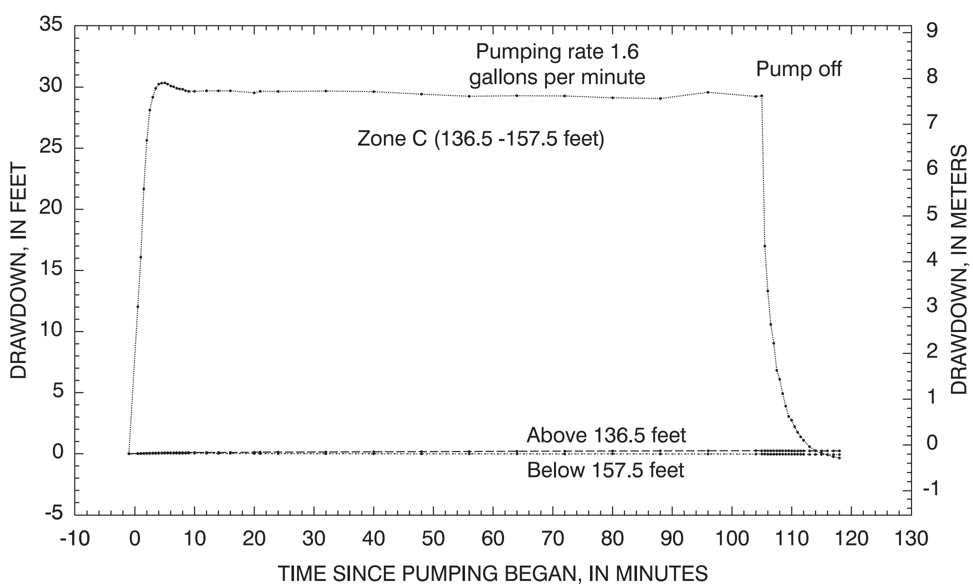

Figure 35. Drawdown as a function of time in aquifer-interval-isolation test of zone C of borehole Mg-1444 in Lansdale, Pa., April 4, 1997.

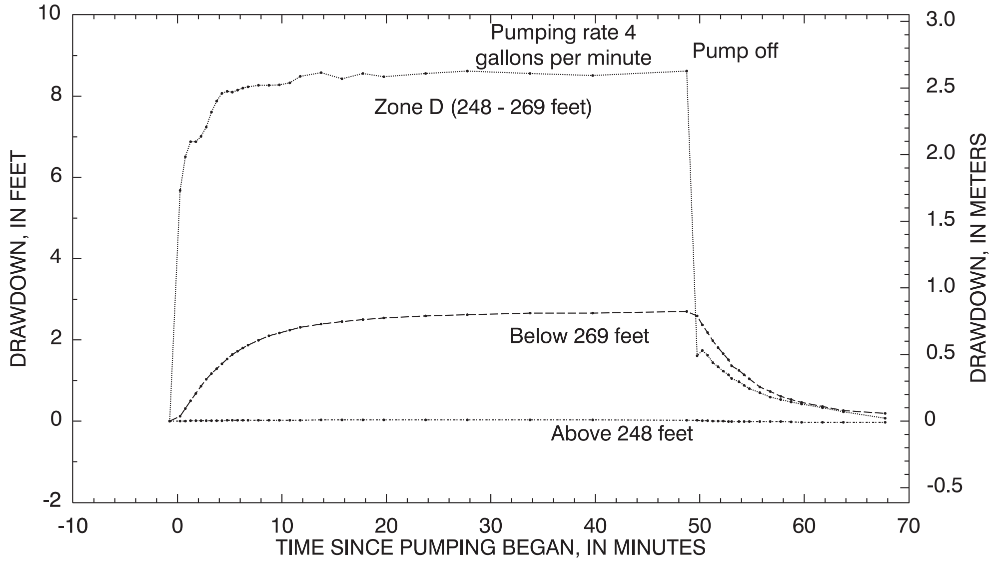

Figure 36. Drawdown as a function of time in aquifer-interval-isolation test of zone D of borehole Mg-1444 in Lansdale, Pa., April 3, 1997.

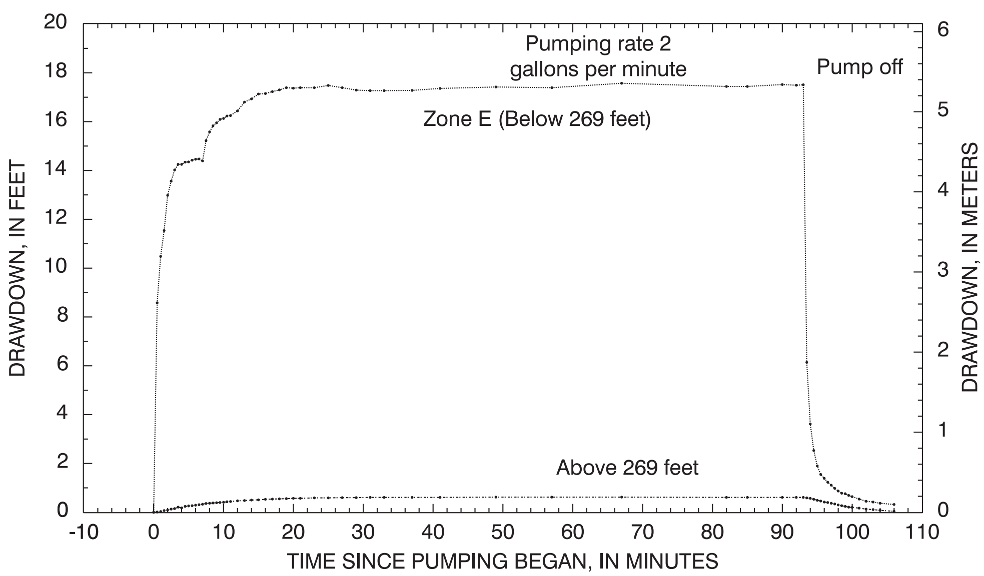

Figure 37. Drawdown as a function of time in aquifer-interval-isolation test of zone E of borehole Mg-1444 in Lansdale, Pa., April 3, 1997.

Table 9. *Depths, water levels, specific capacity, and transmissivity of aquifer intervals isolated by packers and of the open hole for well Mg-1444 in Lansdale, Pa., April 1997 and October 1997*

[ft bls, feet below land surface; ft, feet; gal/min, gallons per minute; min, minutes; (gal/min)/ft, gallons per minute per foot; ft^2/d, square feet per day; NA, not applicable]

Depth of isolated zone in borehole (ft bls)	Date of test	Pre-pumping depth to water in interval[1] (ft bls)	Depth to water in interval at end of test[2] (ft bls)	Drawdown (ft)	Pumping rate (gal/min)	Pumping duration (min)	Specific capacity [(gal/min)/ft]	Trans-missivity[3] (ft^2/d)
				Zone A (above 64 ft bls)				
Above 64 (pumped)	4-7-97	56.34	59.04	2.7	0.4	19	0.15	32.5
64-85	4-7-97	56.62	57.32	.7	NA	NA	NA	NA
Below 64	4-7-97	42.52	42.52	0	NA	NA	NA	NA
				Zone B (64-85 ft bls)				
Above 64	4-4-97	54.31	57.78	3.47	NA	NA	NA	NA
64-85 (pumped)	4-4-97	55.32	68.72	13.40	1.5	72	[4] .11	[5] 24.1
Below 85	4-4-97	43.20	43.31	.11	NA	NA	NA	NA
				Zone C (136.5-157.5 ft bls)				
Above 136.5	4-4-97	58.15	58.38	.24	NA	NA	NA	NA
136.5-157.5 (pumped)	4-4-97	41.44	70.73	29.29	1.67	105	.057	12.5
Below 157.5	4-4-97	40.38	40.36	-.02	NA	NA	NA	NA
				Zone D (248-269 ft bls)				
Above 248	4-3-97	54.58	54.60	.02	NA	NA	NA	NA
248 - 269 (pumped)	4-3-97	39.23	47.85	8.62	4	49	.46	102
Below 269	4-3-97	40.11	42.81	2.7	NA	n	NA	NA
				Zone E (below 268 ft bls)				
Above 268	4-3-97	41.54	42.12	.61	NA	NA	NA	NA
Below 268 (pumped)	4-3-97	47.99	65.50	17.51	2	93	.11	25.1
Sum of specific capacities or transmissivities for zones tested							.89	196
				Open-hole tests				
Open hole	10-1-97	58.8	65.85	7.05	11	130	1.56	342

[1] Stabilized water levels after packers were inflated.

[2] Depth to water at end of pumping at a constant rate before pump was shut off.

[3] Calculated using Thiem equation, assuming radius of influence, r_0, is 328 feet (100 meters).

[4] Calculated specific capacity for zone greater than actual specific capacity because of contributions of flow from other intervals.

[5] Calculated transmissivity for zone greater than actual transmissivity because of contributions of flow from other intervals.

Wells Mg-624 and MG-1639

Aquifer-isolation tests were done in wells Mg-624 and Mg-1639 on the J.W. Rex property in Lansdale by QST Environmental, Inc., during late August and early September 1997. Well Mg-624 is about 633 ft (193 m) deep and well Mg-1639 was about 150 ft (46 m) deep at the time of testing. Intervals for testing were selected on the basis of a review of geophysical logs done by USGS. Three intervals in well Mg-624 and four intervals in well Mg-1639 were tested.

The aquifer-interval-isolation tests in well Mg-624 indicated the tested intervals had relatively low permeability (table 10). The sum of transmissivities for tested zones was about 9.2 ft^2/d, similar to a value of about 6 ft^2/d reported for an earlier aquifer test of the well (Goode and Senior, 1998). In the test of zones A and B [116-146 ft (35.3-44.5 m) and 185-215 ft (56.3-65.5 m)], water levels in isolated intervals indicated a downward vertical gradient. In the test of zone C [290-320 ft (88.4-97.5 m)], water levels in the isolated intervals indicated a small

Table 10. *Summary of aquifer-isolation tests of wells Mg-624 and Mg-1639, Lansdale, Pa., August and September 1997. Data from QST Environmental, Inc. (1998)*

[ft bls, feet below land surface; ft/d, feet per day; ft^2/d, square feet per day; --, no data]

U.S. Geological Survey local well number Mg-	Pumped interval	Depths of isolated intervals (ft bls)	Date of test	Pre-pumping depth to water (ft bls)	Pumping depth to water (ft bls)	Hydraulic conductivity[1] (ft/d)	Transmissivity[2] (ft^2/d)
		Above 116	9-2-97	12.34	36.61		
624	Zone A	116-146	9-2-97	12.98	111.68	0.13	3.9
		Below 146	9-2-97	13.06	28.20		
		Above 185	9-2-97	15.65	55.21		
624	Zone B	185-215	9-2-97	15.85	197.27	.006	.18
		Below 215	9-2-97	17.34	21.54		
		Above 290	9-3-97	12.7	30.7		
624	Zone C	290-320	9-3-97	12.6	150.7	.17	5.1
		Below 320	9-3-97	12.2	17.6		
624	Sum of zones tested					.306	9.2
1639	Zone A	0-40	8-28-97	24.78	32.39	--	--
		Below 40	8-28-97	26.00	34.00		
		Above 40	8-28-97	25.30	25.96		
1639	Zone B	40-60	8-28-97	24.89	28.62	--	--
		Below 60	8-28-97	29.60	30.90		
		Above 80	8-28-97	23.98	24.51		
1639	Zone C	80-100	8-28-97	26.61	59.74	.03	.6
		Below 100	8-28-97	26.47	51.47		
		Above 100	8-29-97	24.29	24.98		
1639	Zone D	100-120	8-29-97	27.51	103.27	.001	.02
		Below 120	8-29-97	27.28	96.30		

[1] Determined from analysis of slug tests (QST Environmental, Inc. 1998).
[2] Calculated by multiplying thickness of isolated interval (20 or 30 feet) by hydraulic conductivity for interval.

upward vertical gradient. The upward vertical gradient is consistent with measurements of upward flow at very low rates (less than 0.01 gal/min) at depths of 66, 100, 214, and 322 ft (20, 30, 65, and 98 m) in well Mg-624 during logging in September 1995 (Conger, 1999). Pumping in well Mg-624 produced little to no drawdown in the nearby shallow [50 ft (15 m)] well Mg-1641, indicating the tested zones are hydraulically isolated from the shallow interval open to well Mg-1641. However, pumping in the three tested zones resulted in drawdown in adjacent intervals in well Mg-624, indicating leakage around packers or hydraulic connection outside of the borehole.

The aquifer-interval-isolation tests in well Mg-1639 indicated the deeper two tested intervals had relatively low permeability (table 10). The upper two intervals tested recovered too quickly from the slug test for analysis and no estimates of permeability were made (QST Environmental, Inc., 1998). Pumping in the upper intervals tested [above 40 ft (12 m) and 40-60 ft (12-18m)] resulted in drawdown in the adjacent shallow monitoring well (Mg-1640) and in the adjacent shallow zone but little to no drawdown in the deeper intervals. Thus, the shallow intervals of well Mg-1639 appear to be hydraulically isolated from the deep zones of the well. Water levels in the isolated intervals of well Mg-1639 indicate a downward vertical gradient. Downward flow was measured during geophysical logging of the well (Conger, 1999).

Chemical and physical properties of water

Chemical and physical properties and selected results of VOC analysis of water samples collected at the end of pumping of each isolated zone of wells Mg-80, Mg-1443, and Mg-1444 are summarized in table 11. Selected water-quality data from the aquifer-interval-isolation tests done by QST Environmental, Inc., in wells Mg-624 and Mg-1639 also are presented in table 11. For each pumped zone, the chemical and physical properties stabilized after about 20 to 40 minutes of pumping.

Comparison of the data for each isolated zone indicates the chemical and physical properties of the water differed slightly within the boreholes and differed to a greater extent between boreholes. Minor differences in chemical and properties of water from isolated zones may indicate hydraulic connection between zones, either by vertical flow between water-bearing zones in the borehole or through fracture networks outside the borehole. Upward flow was observed in wells Mg-1443 and Mg-1444, downward borehole flow was observed in well Mg-80 and Mg-1639, and upward and downward vertical gradients were measured in well Mg-624 (Conger, 1999; QST Environmental, Inc., 1998). Differences in properties measured in water from upper (shallow) and lower (deep) zones of wells Mg-1443, Mg-1444, and Mg-1639 included (1) the water temperature in the upper zones tended to be higher than that in the lower zones; (2) water temperature in zones with relatively high productivity generally was lower than in other zones of the borehole; and (3) the dissolved oxygen concentration commonly was higher, the pH lower, and the specific conductance was lower in water from the uppermost zone than in water from the lower zones. The water in upper zones commonly was more oxygenated and more dilute at the time of sampling compared to water from lower zones, suggesting that water in the upper zone had greater or more recent contact or exchange with the atmosphere and less contact time with aquifer materials than water in the lower zones. Wells Mg-80 and Mg-624 have deeper casing than the other wells and, thus, lack an upper zone open to shallow ground water. As such, water temperature, pH, specific conductance, and dissolved oxygen concentration varied little with depth in tested zones of these wells.

In comparison of chemical and physical properties of water from wells Mg-80, Mg-1443, Mg-1444, Mg-624 and Mg-1639, the water from wells Mg-1444 and Mg-624 had the highest pH (was the most basic), the water from Mg-80 contained the lowest concentration of dissolved oxygen (less than 0.1 mg/L), and the water from wells Mg-1443 and Mg-624 had the lowest specific conductance (table 11). Differences in the specific conductance of water from the five wells also were indicated by the fluid-resistivity logs (Conger, 1999). Fluid resistivity is the inverse of fluid conductivity. The differences in properties of water from these wells may be related to the residence time of ground water in the vicinity of the wells, differences in aquifer mineralogy, and (or) differences in compounds introduced by human activities in recharge areas of the wells. Ground water with a short residence time generally is more similar chemically to recharge water (dilute, oxygenated, and acidic) than ground water with a long residence time.

The concentrations of VOC’s also differed between the isolated intervals of the boreholes (table 11). For intervals isolated in well Mg-80, the highest concentrations of PCE and TCE were measured in the upper (shallow) zone and the highest concentrations of VC was measured in the lower (deep) zone. Because PCE and TCE are the

Table 11. *Physical properties and concentrations of selected volatile organic compounds in samples collected from isolated intervals at the end of pumping in wells Mg-80, Mg-1443, and Mg-1444 in Lansdale, Pa., March 26-April 11, 1997, and in wells Mg-624 and Mg-1639, August 28 - September 2, 1997*

[ft bls, feet below land surface; °C, degrees Celsius; µS/cm, microsiemens per centimeter; mg/L, milligrams per liter; µg/L, micrograms per liter; PCE, tetrachloroethylene; TCE, trichloroethylene; DCE, dichloroethylene; VC, vinyl chloride; <, less than; ND, not detected; --, no data]

Well and interval sampled	Depth of interval sampled (ft bls)	Temperature (°C)	Specific conductance (µS/cm)	pH (units)	Dissolved oxygen (mg/L)	Compound concentration[1] (µg/L)				
						PCE	TCE	1,2-cis-DCE	VC	Toluene
Mg-80 - Zone A	142-157	12.7	600	6.86	<0.1	10.5	19.6	24.0	10.5	0.6
Mg-80 - Zone B	Below 246	12.7	[2] 600	6.79	<.1	ND	ND	5.2	57.2	ND
Mg-1443 - Zone A	Above 90.5	19.0	372	5.97	[3]10.3	408	3,550	512	2.6	8.9
Mg-1443 - Zone B	90.5-110.5	19.9	405	6.10	5.7	510	3,680	524	ND	14.4
Mg-1443 - Zone C	276-296	14.8	427	7.03	1.6	199	1,670	167	ND	.6
Mg-1443 - Zone D	Below 296	16.4	386	6.74	1.4	208	3,350	265	ND	13.6
Mg-1444 - Zone A	Above 64	19.9	445	7.35	6.3	11.4	1,220	.4	ND	33.7
Mg-1444 - Zone B	64-85	14.8	586	7.61	--	1.7	141	ND	ND	1.8
Mg-1444 - Zone C	136.5-157.5	15.4	625	7.58	1.1	.1	.6	ND	ND	.7
Mg-1444 - Zone D	248-269	14.0	600	7.58	2.0	ND	.5	ND	ND	.4
Mg-1444 - Zone E	Below 268	14.9	590	7.57	1.2	ND	1.2	ND	ND	.6
Mg-624 - Zone A	116-146	15.3	429	8.03	2.5	<1	7	4	<1	--
Mg-624 - Zone B	185-215	15.6	431	8.07	3.4	<1	2	4	<1	--
Mg-624 - Zone C	290-320	13.6	426	7.39	<.1	<1	<1	1	<1	--
Mg-1639 - Zone A	0-40	14.7	941	6.80	1.8	350	660	620	20	--
Mg-1639 - Zone B	40-60	15.3	950	6.80	.7	350	700	630	20	--
Mg-1639 - Zone C	80-100	15.2	976	6.91	.25	500	890	650	20	--
Mg-1639 - Zone D	100-120	14.2	1,017	6.88	.4	420	780	660	20	--

[1] VOC analytical results for Mg-80, Mg-1443, Mg-1444 from Black & Veatch Waste Science, Inc. (1998) and for Mg-624, Mg-1639 from QST Environmental, Inc. (1998).

[2] At beginning of pumping; probe fouled at end of pumping.

[3] Sample aerated during pumping; reported concentration in sample probably higher than in unaerated ground water from zone.

primary contaminants at the site, these results suggest the upper zone draws water that may be close to contaminant sources near land surface as contaminated water moves deeper into the aquifer under flow or density gradients. VC at this site was likely formed during the chemical breakdown of PCE and TCE.

In samples from well Mg-1443, concentrations of PCE and cis-1,2-DCE were higher in the shallow zones than in the deep zones (table 8), suggesting the upper zones may be closest to contaminant sources near land surface. TCE and toluene concentrations in three of the four zones sampled in well Mg-1443 were similar, perhaps indicating greater areal and depth extent of contamination than that of the other VOC’s detected. Zone D (276-296 ft bls) was the least contaminated but most productive zone of the well. This relation indicates that although fractures may provide preferential pathways for contaminants, increased flow through fractures may result in dilution of contaminants. Upward flow and upward vertical flow gradients were measured at all but the shallowest depths tested. A small downward vertical flow gradient was measured between the shallowest zone tested (zone A) and the underlying zone (zone B), indicating potential for transport of contamination from the shallowest zone to receiving fractures at 104-106 ft bls (31.7-32.3 m).

In samples from well Mg-1444, concentrations of PCE, TCE, and toluene were much greater in the shallowest zone than in the other zones sampled, indicating proximity to a contaminant source near the surface. Because a downward vertical flow gradient was measured from the shallowest zone tested (zone E) to the underlying zone (zone D), movement of contamination in the borehole from above 64 ft (19.5 m) to receiving water-bearing fractures at 70-72 ft bls (21.3-21.9 m) is possible. However, upward flow was measured in well Mg-1444 at depths below 72 ft (21.9 m) (Conger, 1999), indicating that under nonpumping conditions, the contaminants near the upper zones of the well are not moving to depths below 72 ft (21.9 m) in the borehole.

In samples from well Mg-624, only low concentrations of VOC's were detected. The concentrations of TCE and its breakdown product, cis-1,2-DCE, were greater in samples from the upper two zones than the deepest zone tested, suggesting a contaminant source near the surface. An upward vertical flow gradient was observed from the lower zone to the intermediate zone tested, indicating little potential for downward migration in the borehole over that interval.

In samples from well Mg-1639, relatively high concentrations of PCE, TCE, and cis-1,2-DCE were measured in water from all four zones tested. Concentrations of these compounds were higher in the lower two zones than in the upper two zones. VC was present in the same concentrations in all zones. Cross-contamination between zones in and outside of the borehole may explain the similar concentrations of contaminants in the four zones. Downward vertical flow gradients were noted between zones in this well.

Multiple-Well Tests

Aquifer tests involving 1 pumped well and 6 to 10 observation wells were done by USGS at 3 sites in November 1997. The pumped wells were Mg-1610 at Keystone Hydraulics property, Mg-1609 at John Evans Co. property, and Mg-1600 at Rogers Mechanical property (pl. 1). Information about the pumped and observation wells and the aquifer tests is summarized in table 12. Another aquifer test was done at the J.W. Rex property by QST Environmental, Inc., during which well Mg-625 was pumped. The tests were done in areas of known soil and ground-water contamination. The observation wells were oriented at various screened-depth intervals in both dip and strike directions from the pumped well and include open-hole wells and wells constructed in 1997. Wells constructed in 1997 generally have about 20-ft (6.1-m) of screen open to one water-production zone. The tests were done, in part, to determine the relation between transmissivity and aquifer-bed orientation. Information about vertical and horizontal transmissivity also was obtained.

At each site, one well of a nest was pumped. The pumped wells ranged in depth from about 100-150 ft (30.5-45.7 m) and were deeper than the companion monitor wells in the nests. New monitor-well nests were installed during summer 1997. Each well in a nest was constructed to be open to one water-bearing zone. Water levels in other monitor wells and in unused, deep, open-hole wells at or near the sites were measured before, during, and after the test by use of pressure transducers or floats and digital shaft encoders. Water levels were checked periodically by use of an electric tape to verify transducer and float readings. Barometric pressure was measured by use of a transducer during tests at Rogers Mechanical (pumped well Mg-1600) and John Evans (pumped well Mg-1609) at a nearby site in Warminster Township, Bucks County, Pa., about 10 mi (16.1 km) of Lansdale. Wells were pumped by use of a 0.5 horse-power submersible pump at rates of 8 - 10 gal/min (30.3 - 37.9 L/min) for about 8 hours. All pumped water was passed through granulated activated carbon to remove contaminants and then discharged to sanitary sewers. Pumping rates during the first 10-60 minutes tended to be variable and higher than later, stable pumping rates because of adjustments required to avoid exceeding the flow capacity of the carbon-filtration tanks. Drawdown of water levels during pumping and recovery after pumping ceased was measured at each site. For tests at two sites, Rogers Mechanical (pumped well Mg-1600) and John Evans (pumped well Mg-1609), recovery coincided with periods of rainfall that affected the cause and rate of rise in water levels.

Table 12. *Well characteristics and locations and pumping data for aquifer tests done in Lansdale, Pa., November 1997*

[P, pumped well; O, observation well; ft, feet; in., inches; ft bls, feet below land surface; ft asl, feet above sea level; gal/min, gallons per minute]

U.S. Geological Survey local well number Mg-	Site well name[1]	Well status	Well depth (ft)	Casing depth (ft)	Well diameter (in.)	Depth to water before test (ft bls)	Drawdown at end of test (ft)	Altitude of land surface (ft asl)	Location relative to pumped well	
									Radial distance (ft)	Direction[2] (degrees)
Rogers Mechanical site - date: 11-13-97, start time: 12:21, duration: 6.15 hours, stable pumping rate: 8.1 gal/min										
1600	Rog 3I	P	150	130	6	50.8	3.94	365.7	--	--
1605	Rog 1S	O	95	15	6	66.5	-.03	380.5	410.55	116.83
1604	Rog 1D	O	221	210	6	67.5	-.07	380.7	417.83	117.02
1603	Rog 2S	O	98	15	6	74.8	-.10	376.0	307.49	158.04
1602	Rog2I	O	131	110	6	61.7	.71	376.0	313.17	159.64
1601	Rog 3S	O	100	18	6	65.1	-.06	365.5	12.01	41.83
1444	TA-1	O	294	17	6	58.6	-.06	367.0	210.75	195.32
Keystone Hydraulics site - date: 11-18-97, start time: 10:55, duration: 8.05 hours, stable pumping rate: 10.0 gal/min										
1610	Key 1I	P	122	100	6	16.6	2.31	326.7	--	--
1611	Key 1S	O	88	15	6	16.4	.44	326.6	25.15	126.41
80	KH1	O	270	138	8	15.9	.46	326.0	153.01	122.69
1620	Key 2S	O	101	20	6	14.3	.32	324.4	364.26	115.27
1619	Key 2I	O	190	150	6	14.2	.13	324.1	354.69	111.89
67	L-8	O	294	19	8	15.3	.14	325.4	516.12	102.42
163	RY2	O	318	22	8	29.2	.14	339.2	759.97	233.81
164	RY2	O	385	23	8	31.0	-.04	340.1	1,013.51	201.17
John Evans site - date: 11-21-97, start time:10:10, duration: 7.93 hours, stable pumping rate: 9.1 gal/min										
1609	Ev2S	P	101	19	10	55.6	3.56	352.4	--	--
1533	JE-1	O	63	14	6	55.5	.71	352.3	15.76	86.67
1624	Ev 1S	O	101	19	6	45.5	-.08	347.9	549.25	305.56
1666	Ev 1I	O	150	110	6	51.3	.37	348.3	520.69	305.61
1606	PhT 1S	O	101	15.5	6	51.2	.41	349.0	496.90	222.91
1607	PhT 1I	O	161	153	6	49.4	-.04	348.0	506.05	229.21
1608	PhT 1D	O	307	220	6	51.3	.23	348.5	501.64	225.77
1443	PTC	O	339	10	8	52.6	.13	351.1	206.17	232.52
152	PW1	O	196	22	12	57.8	.52	354.1	194.75	37.42
1445	AOmw1	O	204	21	5	61.8	.13	357.9	910.35	42.07
618	NPP	O	343	29	6	66.0	-.08	360.4	1,298.65	53.83

[1] Name given by Black & Veatch Waste Science, Inc.
[2] Due north is 0 degrees, due east is 90 degrees, due south is 180 degrees, due west is 270 degrees.

Method of aquifer-test analysis

The general approach for analyzing the aquifer-test data for this study was to match the measured drawdown with simulated drawdown using analytical models. These simple models treat the aquifer system as homogeneous and of infinite horizontal extent. Three models were used, including: (1) an isotropic single-aquifer model (Theis, 1935); (2) an anisotropic single-aquifer model (Papadopulos, 1965); and (3) an isotropic two-aquifer model (Neuman and Witherspoon, 1969). The fit between measured and simulated drawdown was judged by visual inspection of the log-log graph of drawdown as a function of time since the start of pumping. The model parameters are adjusted such that an optimum fit was achieved.

The Theis (1935) model assumes all wells fully penetrate a confined aquifer in which the transmissivity (T) is independent of direction. The model parameters are transmissivity and storage coefficient (S). Simulated drawdown depends on the radial distance from the pumped well (r) and time elapsed since pumping began (t). The aquifer in the conceptual model corresponds to the network of fractures that are the most permeable and provide most of the flow to the pumped well. Low-permeability parts of the aquifer system, for example, large blocks of unfractured rock, are not explicitly included in the model. Furthermore, wells that are isolated from the pumped aquifer by low-permeability barriers to flow are not included in the model. For example, a well may be open to productive fractures, but those fractures may be isolated from the pumped aquifer by intervening beds of relatively unfractured low-permeability beds. The response of such a well cannot be simulated by use of the simple Theis single-aquifer model.

The anisotropic model (Papadopulos, 1965) is similar to the Theis model, except that transmissivity depends on direction. Directional transmissivity is an ellipse characterized by three parameters. The three parameters are the transmissivity in the direction of maximum transmissivity (T_{max}), the transmissivity in the direction of minimum transmissivity (T_{min}), and the direction of maximum transmissivity, which is specified by the angle between north and the direction of maximum transmissivity (θ_{max}). In addition to depending on r and t, drawdown at an observation well depends on the angle of the line joining the pumped and observation well (θ_{obs}). This model can approximate the apparent large-scale anisotropy often observed in dipping Triassic formations (Vecchioli, 1967; Carleton and others, 1999). As a single-aquifer model, however, this model cannot simulate drawdown in wells in low-permeability blocks or isolated wells, as described in the previous paragraph.

The isotropic two-aquifer model (Neuman and Witherspoon, 1969) assumes two semi-confined isotropic infinite aquifers separated by a confining unit. Only horizontal flow is considered in each of the aquifers, whereas only vertical flow is considered in the confining unit. The pumping well penetrates only one of the aquifers, and each observation well is assumed to fully penetrate either the pumped or unpumped aquifer. For the case considered here, no observation wells are located in the aquitard. The parameters for the isotropic two-aquifer model are transmissivity (T_1) and storage coefficient (S_1) in the pumped aquifer; transmissivity (T_2) and storage coefficient (S_2) in the unpumped aquifer; and thickness (B), vertical hydraulic conductivity (K_v), and specific storage (S_s) of the aquitard. This model can approximate water levels in wells that penetrate (1) a network of fractures hydraulically connected with the pumped well and (2) a second network of high-permeability fractures that are separated from the pumped aquifer by intervening low-permeability parts of the formation. As with the Theis and anisotropic models, wells that are completed in low-permeability parts of the formation (other than the intervening aquitard) cannot be simulated by use of this model.

The models used to simulate drawdown are simplifications of natural conditions. The models do not incorporate several known and unknown complexities that affect measured drawdown. Ground-water flow in the fractured rocks is through a complex network of interconnected fractures. The models are used to approximate the response of the system in the relatively well-connected network of fractures that most-readily contributes water to the pumped well. The typically small drawdown measured in wells that are not well-connected to the primary water-producing fracture network cannot be simulated by use of these models, except to the extent that a connection can be approximated by an infinite, homogeneous confining unit in the case of the two-aquifer model. The uniform parameters (T and S) determined from aquifer tests are effective values at the scale of the well field. Using these effective values, the simulated drawdown most closely matches measured drawdown during the test.

The approach of estimating a few large-scale effective parameters is consistent with the goal of developing a model of regional flow in the formations underlying Lansdale. Regional models, however, cannot fully describe local details of flow in heterogeneous formations. More complex models could be used to more closely simulate the aquifer-test data and describe local flow characteristics. For example, transmissivity could vary in space, having different values in separate zones. In this case, each transmissivity value in each zone would be a separate model parameter. Although such a model may provide a better match between measured and simulated drawdowns, the reliability of each parameter (each zone's transmissivity in the example) decreases sharply as the number of parameters increases. Furthermore, the parameters become non-unique; several different combinations of parameters yield virtually the same match between measured and simulated drawdown. The parameters estimated from these simple models are relatively well-constrained by the measured drawdown, but the field situation is considerably more complex than these conceptual models imply.

Effects of heterogeneity and limited vertical hydraulic conductivity were observed in all three tests. At the Rogers Mechanical site, the water levels in only one observation well responded to pumping. Water levels in other wells, closer to the pumping well but open to parts of the formation above the pumped beds, did not respond, indicating limited hydraulic connection across beds. An anisotropic flow model in a single confined aquifer is used to analyze the drawdown at the Keystone Hydraulics site. However, this analysis included only the four observation wells with largest drawdown. Lower drawdown at several other observation wells does not match this model. Conceptually, these observation wells are located outside the high-permeability pumped beds and their response is muted by the limited cross-bed hydraulic conductivity. Finally, the test at the John Evans site is analyzed by use of a two-aquifer model. In this case, one observation well was located in a relatively moderate-permeability 'aquifer,' but the drawdown was significantly reduced because of intervening low-permeability parts of the formation. Here again, no drawdown because of pumping was measured at several observation wells indicating low hydraulic conductivity connections between these wells and the pumping well. The variability of the extent of response to pumping at all three sites underscores the heterogeneity of three-dimensional hydraulic conductivity in these fractured-rock formations. These results are consistent with a multi-aquifer conceptual model of the ground-water system in which flow is primarily in zones oriented parallel to bedding.

Rogers Mechanical site

One aquifer test was done at this site on November 13, 1997. Well Mg-1600 was pumped for 6.15 hours at rates that ranged from 7.9 to 14.7 gal/min (0.5 to 0.93 L/sec) during the early part of the test. The pumping rate was stable at about 8.1 gal/min (0.51 L/sec) from 7 minutes after pumping started to the end of pumping. Water levels were measured in seven wells (fig. 38) by use of pressure transducers and electric tapes. Barometric pressure at a nearby site also was recorded with a transducer. The configuration of wells included the shallow [less than 100-ft (30-m) deep] wells, Mg-1601, Mg-1603, and Mg-1605; the intermediate-depth [about 150-ft (46-m) deep] wells, Mg-1600 (pumped well) and Mg-1602; deep [222 ft (67.7 m)] well Mg-1604; and an open-hole well [open from 18 to 294 ft (5.5 to 89.6-m)], Mg-1444 (fig. 39; table 12).

Positive drawdown during the aquifer test was measured in the pumped well (Mg-1600) and observation well Mg-1602 (figs. 38 and 39). The effect of variable pumping rate during the early part of the aquifer test also is reflected in the hydrographs of water levels in the pumped well and the observation well Mg-1602. Drawdown in the remaining wells was negative, indicating the water level in those wells rose during the aquifer test. The intermediate-depth observation well (Mg-1602) that responded to pumping is open to a slightly shallower depth in the formation as the pumped well (table 11). Because the local dip is relatively shallow (10°), the open interval of Mg-1602 is open to the same beds as the open interval of the pumped well (fig. 39). Several wells that did not respond to pumping are located closer to the pumped well than well Mg-1602. Measured water levels during the aquifer test illustrate the lack of apparent hydraulic connection between the pumped well and all but one of the observation wells (fig. 40). Because only one observation well had positive drawdown during the aquifer test, the Theis model is used for data analysis.

Drawdown during the later part of the aquifer test in the pumped well (Mg-1600) and in the single observation well (Mg-1602) that responded to pumping can be matched by use of the single-aquifer isotropic model of Theis (1935) (fig. 41). Measured drawdown during the early part of the test is not matched because the pumping rate was elevated for about the first 5 minutes of pumping. The estimated hydraulic properties from this match are $T = 600$ ft^2/d (56 m^2/d) and $S = 3 \times 10$-5 (table 13).

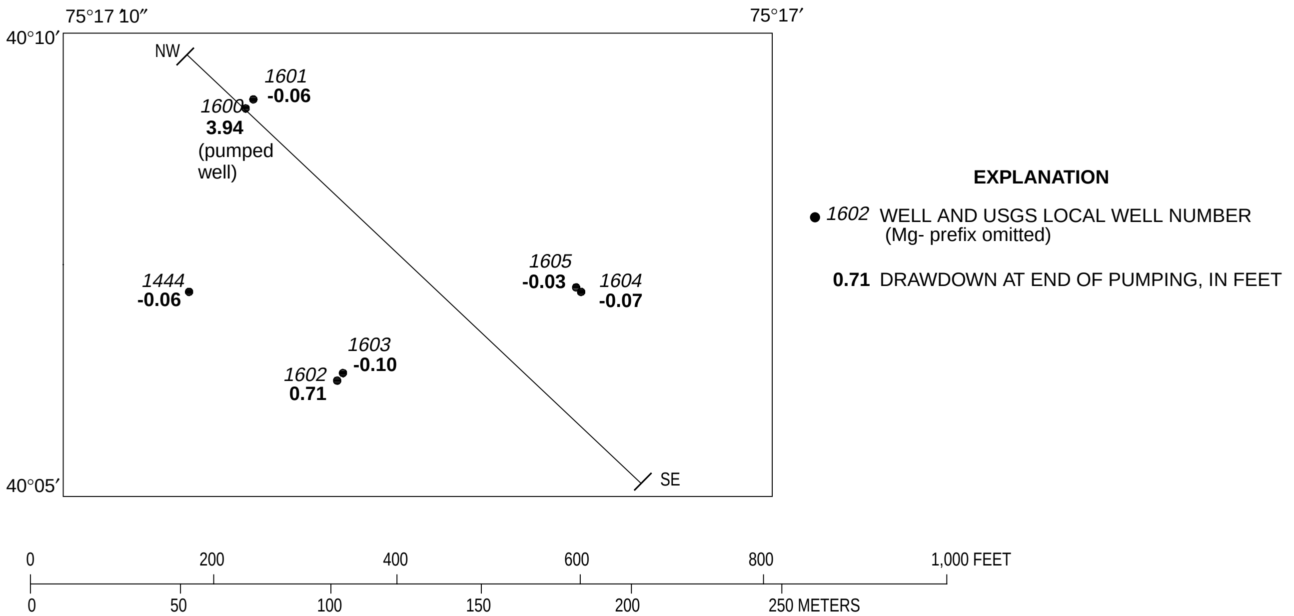

Figure 38. Well locations and drawdown at end of pumping well Mg-1600 at the Rogers Mechanical site in Lansdale, Pa., November 13, 1997. Well Mg-1600 was pumped at a rate of about 8.1 gallons per minute for 6.15 hours.

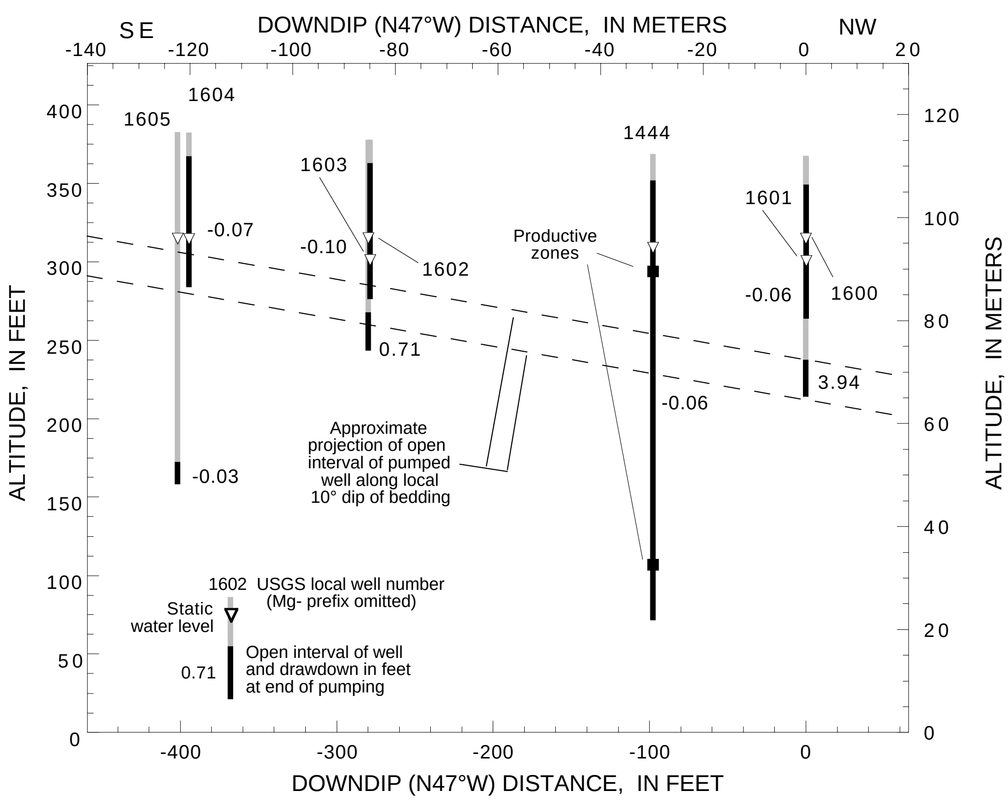

Figure 39. Open intervals of wells, static water level, and drawdown at end of pumping at the Rogers Mechanical site in Lansdale, Pa., November 13, 1997. Well Mg-1600 was pumped at a rate of about 8.1 gallons per minute for 6.15 hours. All wells are projected onto a vertical plane parallel to the dip direction.

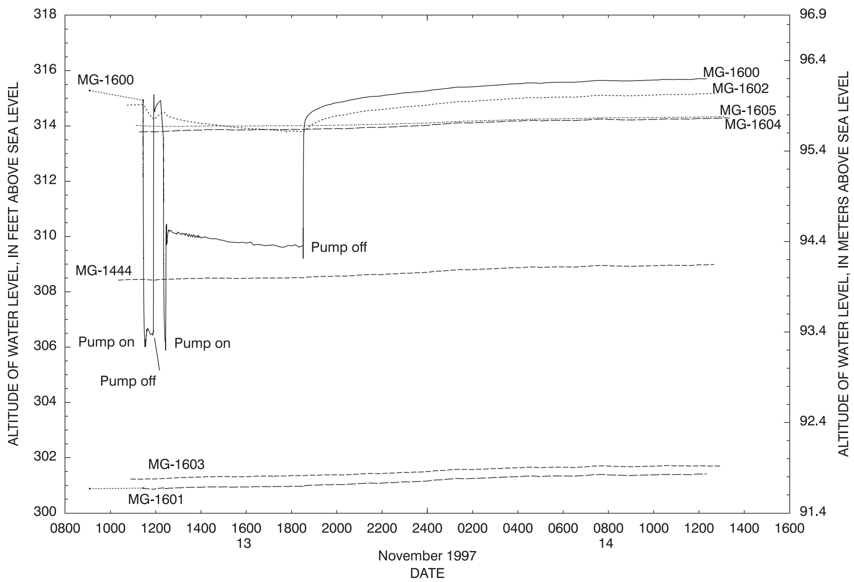

Figure 40. Measured water levels at the Rogers Mechanical site in Lansdale, Pa., November 13-14, 1997. Well Mg-1600 was pumped at a rate of about 8.1 gallons per minute for 6.15 hours on November 13.

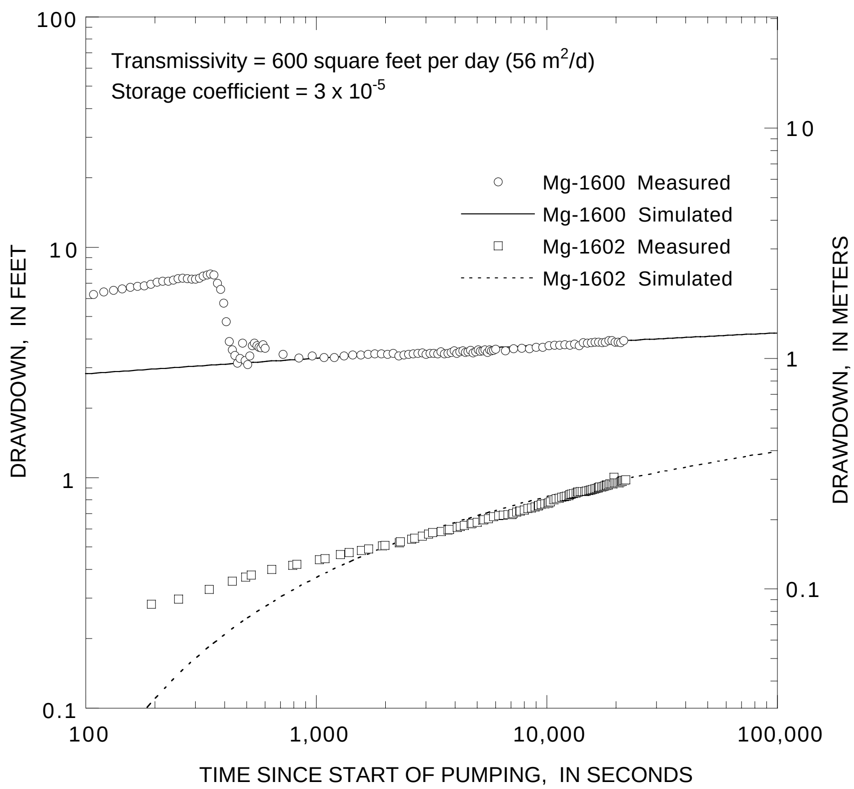

Figure 41. Measured and simulated drawdown in wells Mg-1600 and Mg-1602 at the Rogers Mechanical site in Lansdale, Pa., November 13, 1997. Well Mg-1600 was pumped at a rate of about 8.1 gallons per minute for 6.15 hours. Simulated drawdown is from the isotropic single-aquifer model of Theis (1935) using hydraulic properties of T = 600 ft^2/d (56 m^2/d) and S = 3 x 10^{-5}.

Table 13. *Summary of estimated hydraulic properties determined from analyses of multiple-well aquifer tests in Lansdale, Pa.*

[T, transmissivity; K, hydraulic conductivity; ft^2/d, square feet per day; ft/d, feet per day; S, storage; S_s, specific storage; /ft, per foot; θ, angle]

Site	Conceptual model	Estimated hydraulic properties: Transmissivity (ft^2/d) or vertical hydraulic conductivity (ft/d)		Estimated hydraulic properties: Storage (dimensionless) or Specific storage (/ft)	
Rogers Mechanical	isotropic aquifer	T	600	S	3×10^{-5}
Keystone Hydraulics	anisotropic aquifer	T_{max} (θ_{max} = N. 51° W.) T_{min} (θ_{min} = N. 39° E.) $(T_{max}T_{min})^{1/2}$	10,700 520 2,300	S	3×10^{-5}
John Evans	isotropic two-aquifer	T_1 (lower pumped aquifer) T_2 (upper aquifer) K_v (aquitard)	1,300 15 0.044	S_1 S_2 S_s	8×10^{-5} 8×10^{-5} 1×10^{-6}
J. W. Rex	isotropic aquifer	T	[1]160 - 665	S	[2]2×10^{-5} - 1×10^{-3}

[1] Range of transmissivity values determined by QST Environmental, Inc. (1998).
[2] Range of storage values determined by QST Environmental, Inc. (1998).

Keystone Hydraulics site

One aquifer test was done at this site on November 18, 1987. Well Mg-1610 was pumped for 8.05 hours at rates that ranged from 8.1 to 15 gal/min (0.51 to 0.95 L/sec) during the early part of the test. The pumping rate was stable at about 10 gal/min (0.63 L/sec) from 42 minutes after pumping started until the end of pumping. Water levels were measured in eight wells (fig. 42) by use of pressure transducers and electric tape. The configuration of wells included shallow [less than 100 ft (30 m)] wells Mg-1611 and Mg-1620; intermediate-depth wells [up to 190 ft (60 m)] wells Mg-1610 (pumped well) and Mg-1619; and several deep [more than 270 ft (82 m)] open-hole wells (Mg-67, Mg-80, Mg-163, and Mg-164) (fig. 43). The observation wells were updip and along strike from the pumped well. Bedding at the Keystone Hydraulics site strikes about N. 57° E. and dips about 8° to the northwest (Conger, 1999).

Positive drawdown during the aquifer test was measured in all wells but Mg-164 (fig. 42). Drawdown exceeded 0.3 ft (0.09 m) in three observation wells that were among the closest to and updip of the pumped well (table 12) including Mg-1611, a shallow well within 25 ft (7.6 m) of the intermediate depth pumped well; Mg-80, an open-hole deep well with 138 ft (42 m) of casing and within 153 ft (46.6 m) of the pumped well; and Mg-1620, a shallow well within 365 ft (111 m) of the pumped well. Well Mg-1611 is not open to the projected pumped interval. Although the primary water-bearing zone in well Mg-80 is about 30 ft (9.1 m) below the projected dip of bedding through the pumped zone, aquifer interval-isolation testing indicated this water-bearing zone in well Mg-80 may be hydraulically connected to shallower zones outside the borehole. Shallow well Mg-1620 intersects the projected dip of bedding through the pumped zone (fig. 43). Well Mg-1619 is at a similar distance from the pumped well as well Mg-1620 and is within 25 ft (7.6 m) of well Mg-1620, yet drawdown in well Mg-1619 is only 0.14 ft (0.04 m). Well Mg-1619 is open to beds that are projected to be below the pumped bed (fig. 43). Water levels in well Mg-163, approximately along strike with the pumped well, were drawn down by over 0.18 ft (0.05 m), whereas water levels in well Mg-164, at a similar radial distance but more updip, were not affected by pumping.

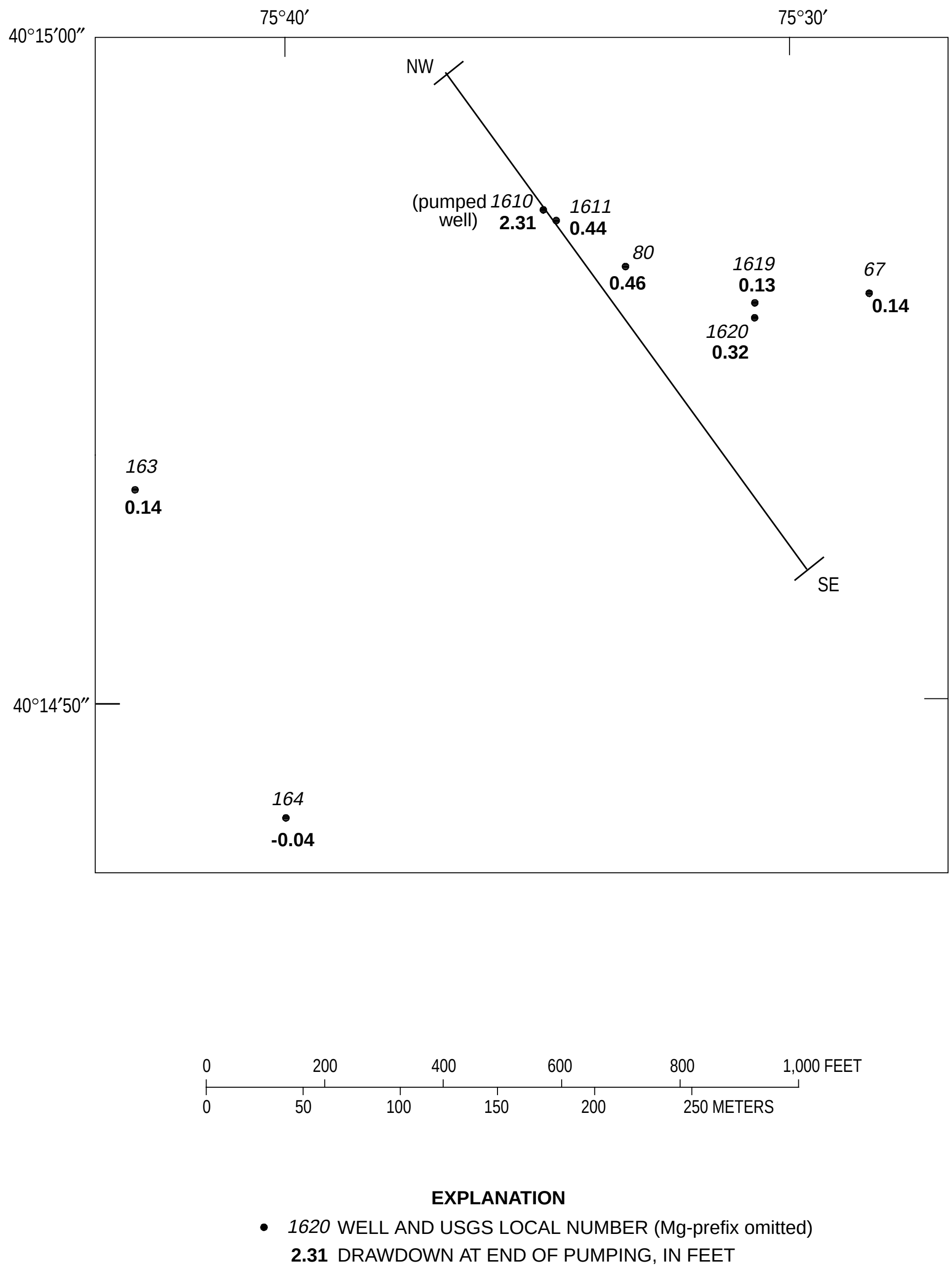

Figure 42. Well locations and drawdown at end of pumping well Mg-1610 at the Keystone Hydraulics site in Lansdale, Pa., November 18, 1997. Well Mg-1610 was pumped at a rate of 10 gallons per minute for 8.05 hours.

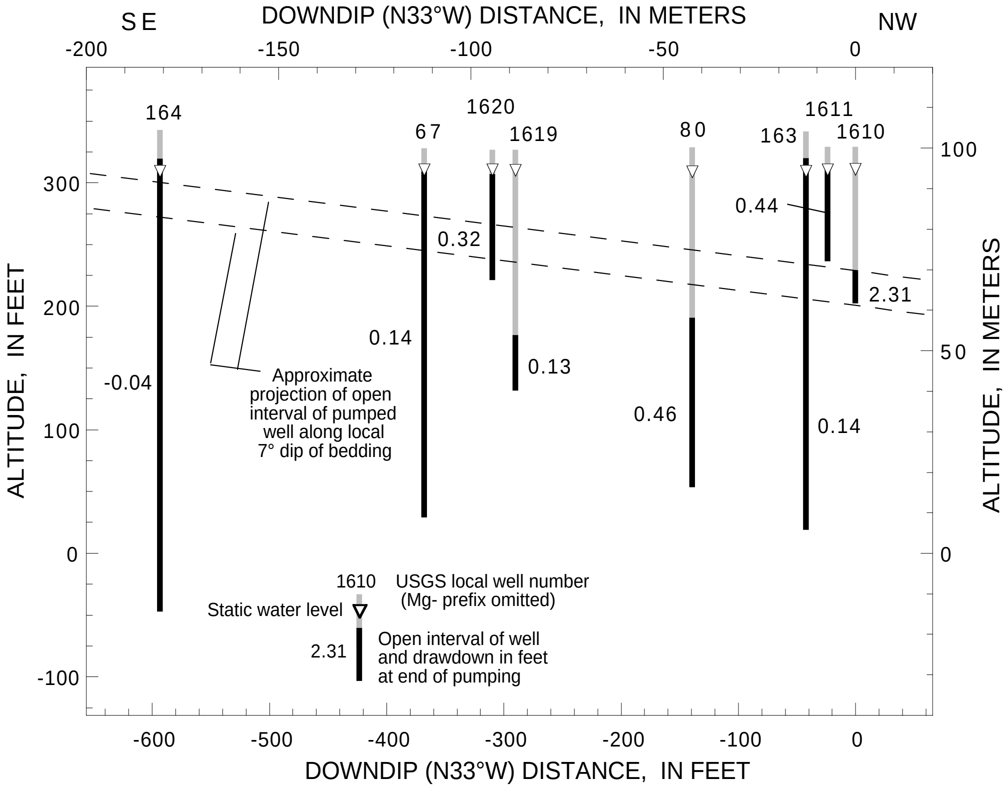

Figure 43. Open intervals of wells, static depth to water, and drawdown at end of pumping at the Keystone Hydraulics site in Lansdale, Pa., November 18, 1997. Well Mg-1610 was pumped at a rate of 10 gallons per minute for 8.05 hours. All wells are projected onto a vertical plane parallel to the dip direction.

Measured water levels during the aquifer test illustrate the effect of pumping, including variable pumping rates at the beginning of the test and fluctuations associated with regional water-level trends (fig. 44). Decreases in barometric pressure resulted in corresponding increases in water levels in wells during the aquifer test. Because the drawdowns resulting from pumping were small, the effect of the barometric-pressure changes was removed prior to analysis of drawdown by use of analytical aquifer-test models. By matching water-level trends in each observation well before and after pumping with the trends in a well unaffected by pumping (well Mg-164), a linear estimation can be made of water levels in the observation wells had pumping not occurred. Drawdown is computed as the difference between this predicted ‘nonpumping’ water level and the measured water level. This correction removes the effects of barometric-pressure fluctuations and other regional trends from the measured drawdown to the extent that those trends at each observation well are the same as the trends at the unaffected well (Mg-164).

Drawdown in four observation wells is selected for analysis by use of the single-aquifer anisotropic model of Papadopulos (1965). Of the six observation wells with positive drawdown, two wells are not matched. Well Mg-1611 is very close to the pumping well but was drawndown less than more distant wells, and the well is not open to the projected pumped bed (fig. 43). Well Mg-1619 was drawdown less than half as much as the nearby well Mg-1620, and it also is not open to the projected pumped bed. Drawdown in these wells cannot be matched by a single-aquifer model because in such a model all observation wells are assumed to be located in the pumped aquifer. These wells are not included in the analysis here in order to use the directional variability of drawdown in the pumped bed to estimate large-scale anisotropy. Well Mg-80 is included in the analysis even though it also is open outside the projected pumped interval. The measured drawdown and aquifer-isolation test results suggest it is hydraulically connected to the pumped interval, as discussed above.

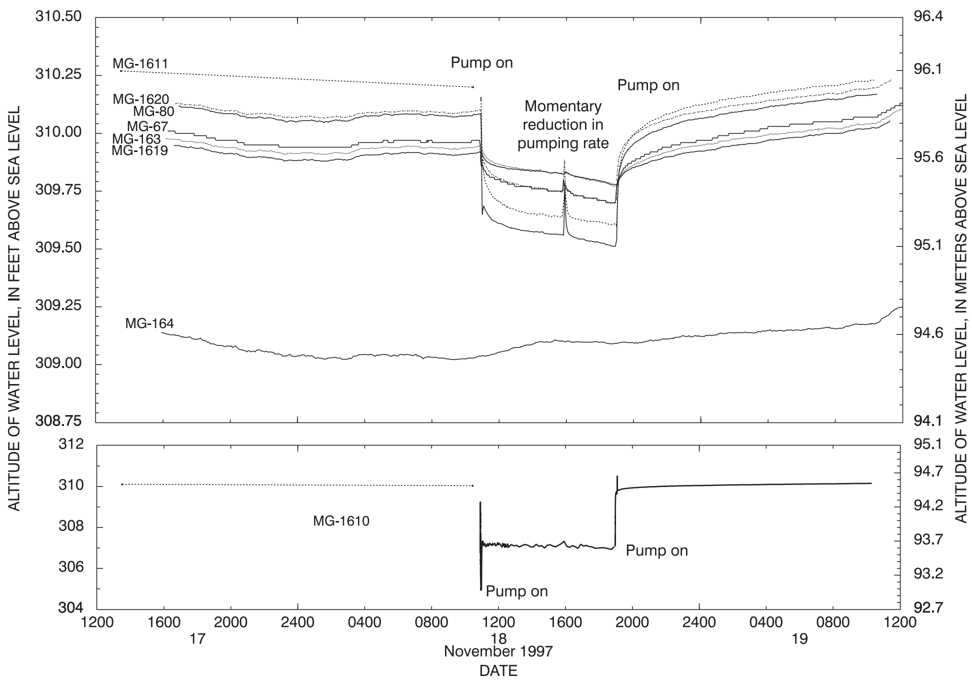

Figure 44. Measured water levels at the Keystone Hydraulics site in Lansdale, Pa., November 17-19, 1997. Well Mg-1610 was pumped at a rate of 10 gallons per minute for 8.05 hours on November 18.

Drawdown in four observation wells can be matched by use of the single-aquifer anisotropic model of Papadopulos (1965) (fig. 45). The response of anisotropic aquifers to aquifer tests include larger drawdowns in one direction than in another for similar distances from the pumped well. The early-time part of the measured drawdown is not matched because the pumping rate was variable for about the first 42 minutes of pumping. The estimated hydraulic properties from this match are: T_{max} = 10,700 ft^2/d (990 m^2/d); T_{min} = 520 ft^2/d (48 m^2/d); θ_{max} = N. 51° W.; and $S = 3 \times 10^{-5}$ (table 13). The non-directional geometric-mean transmissivity is 2,300 ft^2/d (220 m^2/d). These aquifer-test results from this match represent a preferred flow direction within the pumped bed that is oriented in the dip direction (about N. 33° W.). Previous aquifer test results in similar formations (Morin and others, 1997; Welty and Carleton, 1996) present a preferred flow direction oriented in the strike direction.

The difference between the isotropic and anisotropic model match is illustrated by comparing figure 45 to a similar plot using the isotropic Theis model with the nondirectional geometric-mean transmissivity (fig. 46). The isotropic model does not simulate the observed directional dependence of drawdown. Drawdowns at the observation wells estimated by the isotropic model are a function of distance from the pumped well only and more similar in magnitude than those estimated by the anisotropic model. Drawdown simulated by the anisotropic model in two wells (Mg-80 and Mg-1620) updip of the pumped well is greater than drawdown simulated by the isotropic model. Conversely for a well (Mg-163) along strike of the pumped well, drawdown simulated by the anisotropic model is less than drawdown simulated by the isotropic model. Differences in drawdown simulated by the two models are relatively small for well Mg-67, which is oriented between the strike and dip directions.

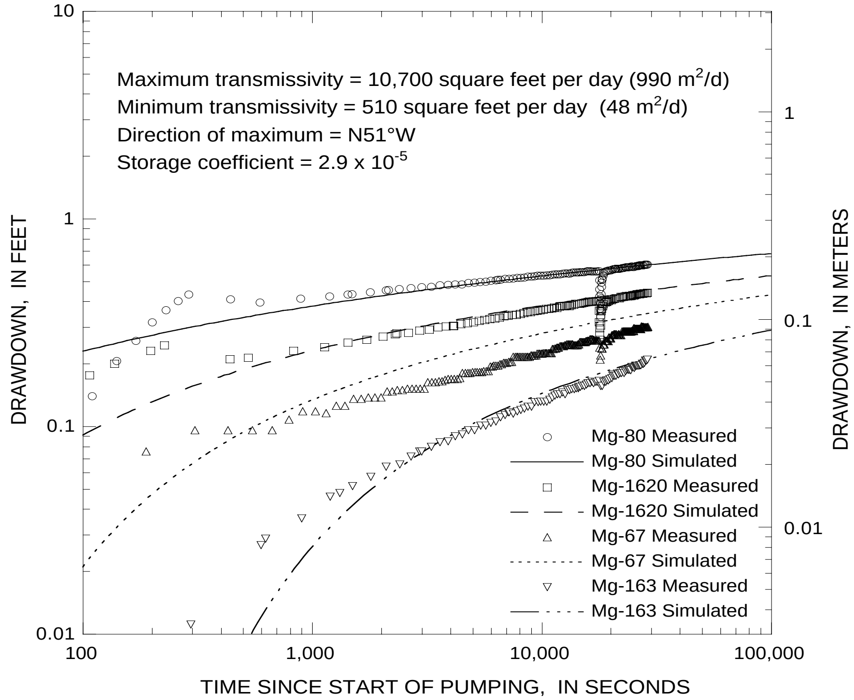

Figure 45. Measured and simulated drawdown, using anisotropic model of Papadopulos (1965), in wells Mg-67, Mg-80, Mg-163, and Mg-1620 at the Keystone Hydraulics site in Lansdale, Pa., November 18, 1997. Well Mg-1610 was pumped at a rate of 10 gallons per minute for 8.05 hours.

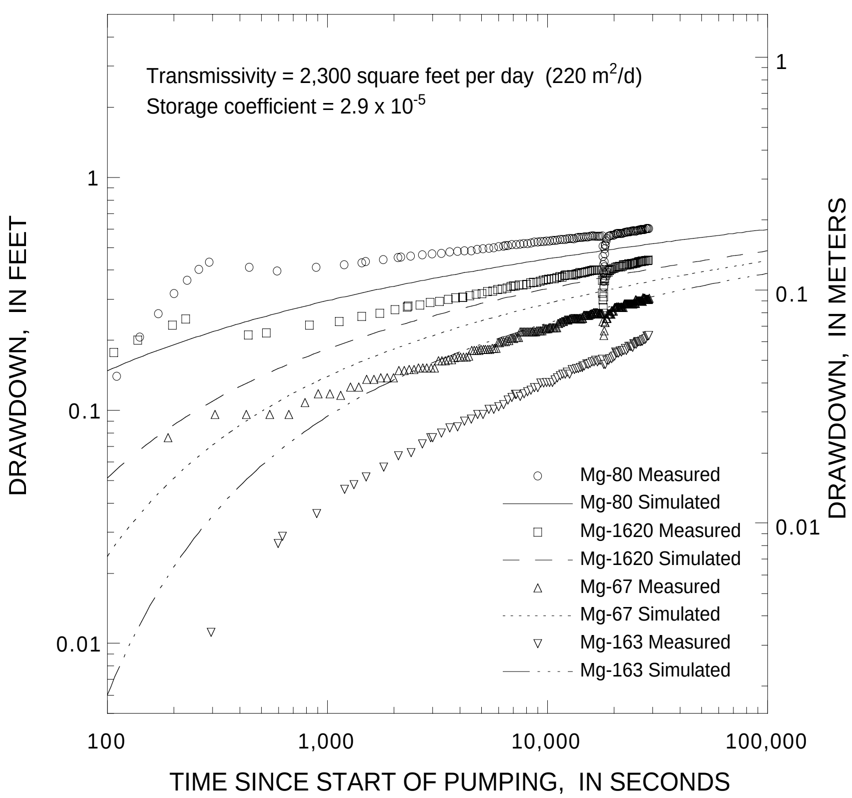

Figure 46. Measured and simulated drawdown, using isotropic model of Theis (1935), in wells Mg-67, Mg-80, Mg-163, and Mg-1620 at the Keystone Hydraulics site in Lansdale, Pa., November 18, 1997. Well Mg-1610 was pumped at a rate of 10 gallons per minute for 8.05 hours.

John Evans site

One aquifer test was done at this site on November 21, 1997. Well Mg-1609 was pumped for 7.93 hours at rates that ranged from 6 to 10 gal/min (0.38 to 0.63 L/sec) during the early part of the test. The pumping rate was stable at about 9.1 gal/min (0.57 L/sec) from 35 minutes after pumping started until the end of pumping. Water levels were measured in 11 wells (fig. 47) by use of pressure transducers and electric tapes. Barometric pressure at a nearby site also was recorded with a transducer. The well configuration included shallow [about 100 ft (30 m) or less in depth] wells Mg-1533, Mg-1606, Mg-1609 (pumped well), and Mg-1624; an open-hole well (Mg-142) with intermediate [less than about 200 ft (61 m)] and shallow water-bearing zones; intermediate wells Mg-1607, Mg-1666, and Mg-1445; deep [about 300 ft (91 m)] well Mg-1608; and two deep open-hole wells, Mg-618 and Mg-1443, open to a large part of the formation (figs. 48 and 49; table 12). Bedding strikes about N. 45° E. and dips about 12° NW. in the vicinity of the site (Conger, 1999).

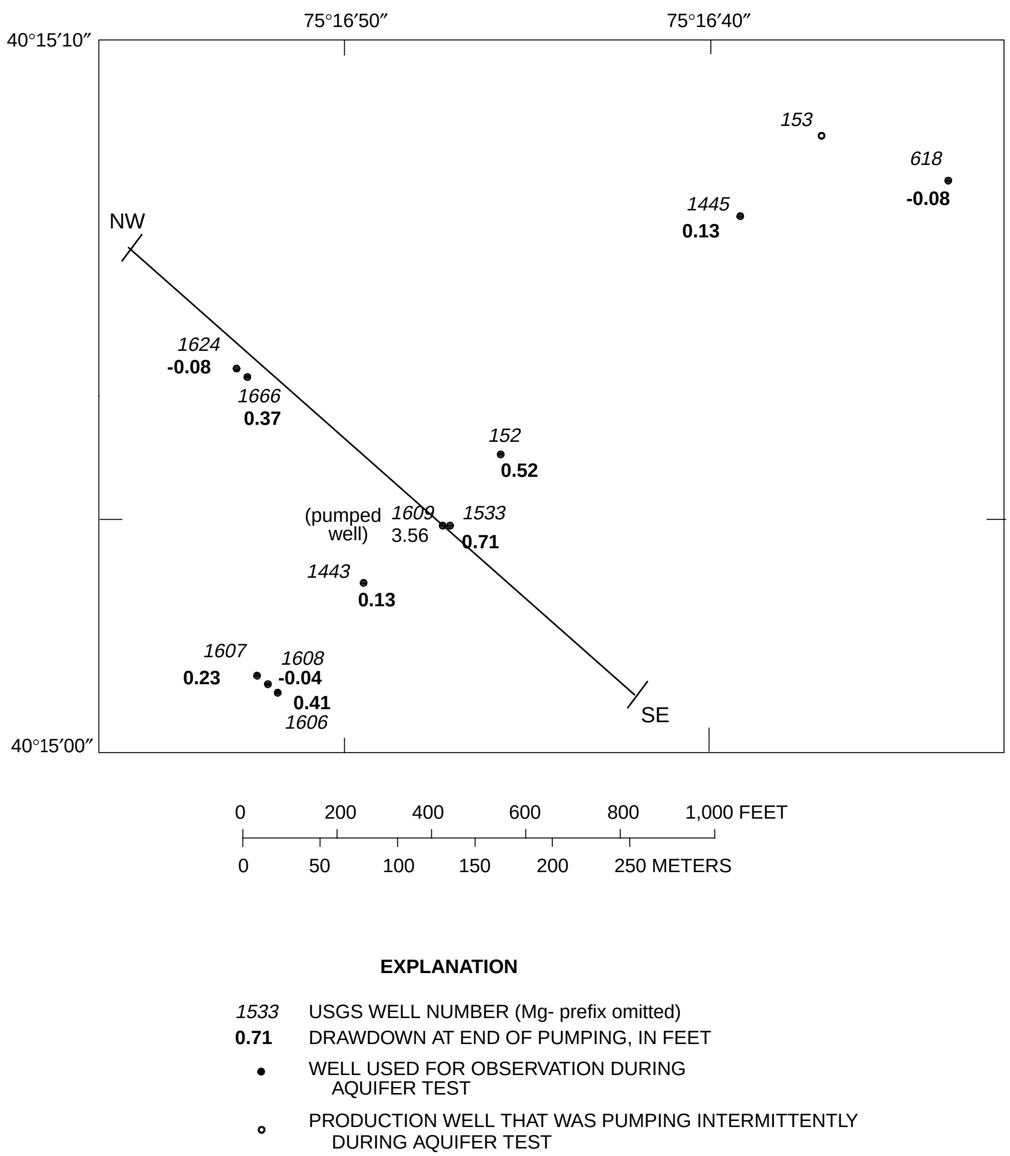

Figure 47. Well locations and drawdown at end of pumping well Mg-1609 at the John Evans site in Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours.

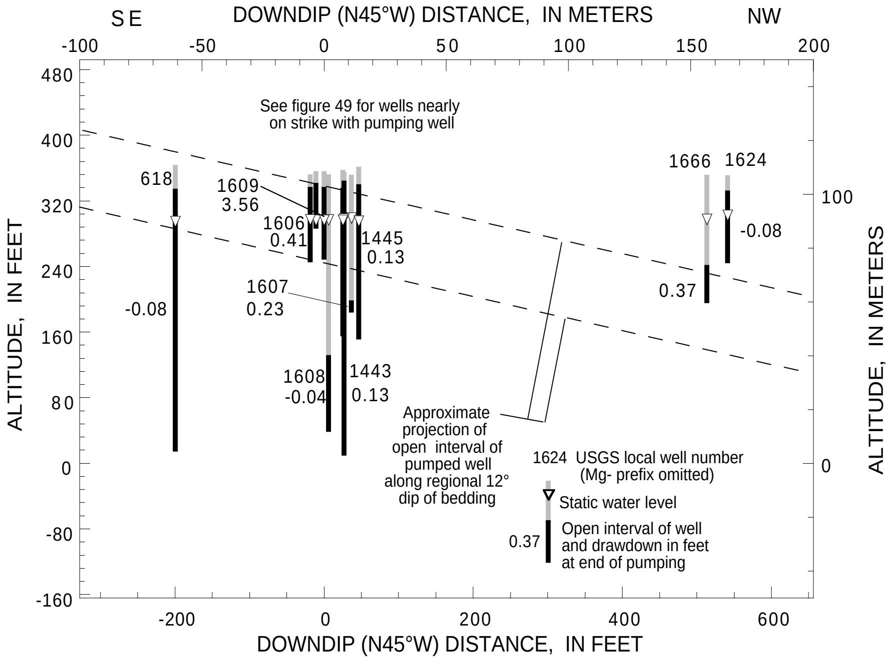

Figure 48. Open intervals of wells, static depth to water, and drawdown at end of pumping at the John Evans site in Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours. All wells are projected onto a vertical plane parallel to the dip direction.

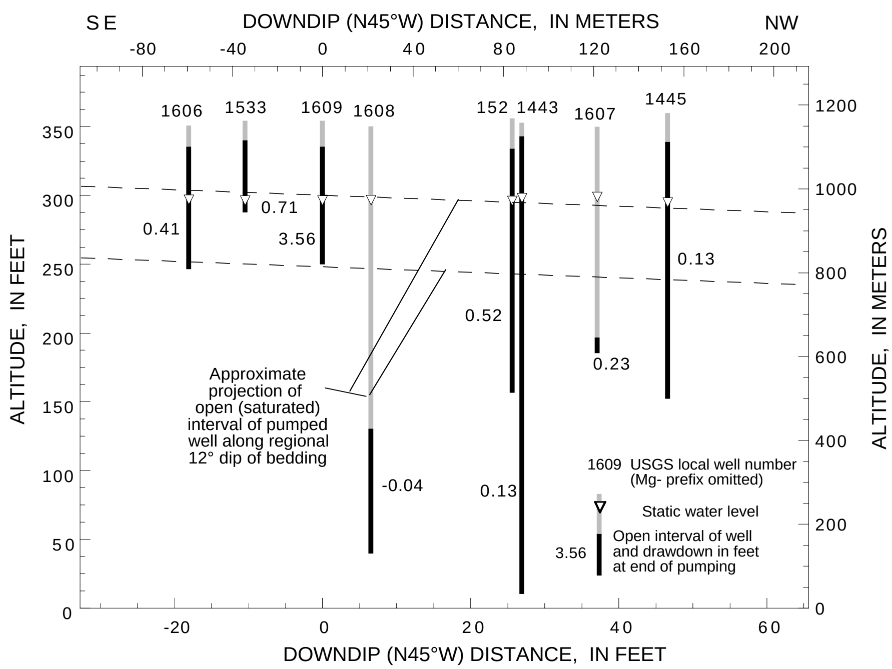

Figure 49. Open intervals of wells nearly on strike with the pumped well, static depth to water, and drawdown at end of pumping at the John Evans site in Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours. All wells are projected onto a vertical plane parallel to the dip direction.

Positive drawdown during the aquifer test was measured in the pumped well and in 7 of the 10 observation wells (fig. 47). Negative drawdown was measured in observation wells Mg-618, Mg-1607, and Mg-1624. Drawdown exceeded 0.3 ft (0.1 m) in four observation wells: Mg-1533, a shallow well adjacent to the shallow pumping well (fig. 49); Mg-152, the next closest observation well that is open to shallow and intermediate depths; Mg-1606, a shallow well relatively far from the pumping well but along strike; and Mg-1666, an intermediate depth well that is downdip of the pumped well but open to the same beds (fig. 48). Well Mg-1443 is about the same distance from the pumped well as well Mg-152, in the opposite direction along strike, and is open to a large part of the formation. Measured drawdown in well Mg-1443 was less than 0.16 ft (0.05 m), which is less than one-third the drawdown at Mg-152. Drawdown in shallow well Mg-1624 was negative, whereas drawdown in the adjacent intermediate well Mg-1666 was over 0.3 ft (0.1 m). These differences in drawdown are consistent with the projection of the pumped beds through the open interval of well Mg-1666 but below that of well Mg-1624 (fig. 48).

Measured water levels during the aquifer test illustrate the effect of pumping, including variable pumping rates at the beginning of the test and fluctuations associated with regional water-level trends (fig. 50). The initial pumping rate was up to about 1 gal/min (0.06 L/sec) greater than the long-term average rate, as evidenced by greater drawdown in the pumped well during the first 15 minutes of the test. The water levels in well Mg-1607 (figs. 49 and 50) are representative of the other two observation wells (Mg-618 and Mg-1624) that did not respond to pumping. The water level in well Mg-1607 did respond to changes in barometric pressure (fig. 18) and rose about 0.04 ft (0.01 m) over the pumping period of the test. Water levels in well Mg-1445 apparently responded to pumping in well Mg-1609 but also

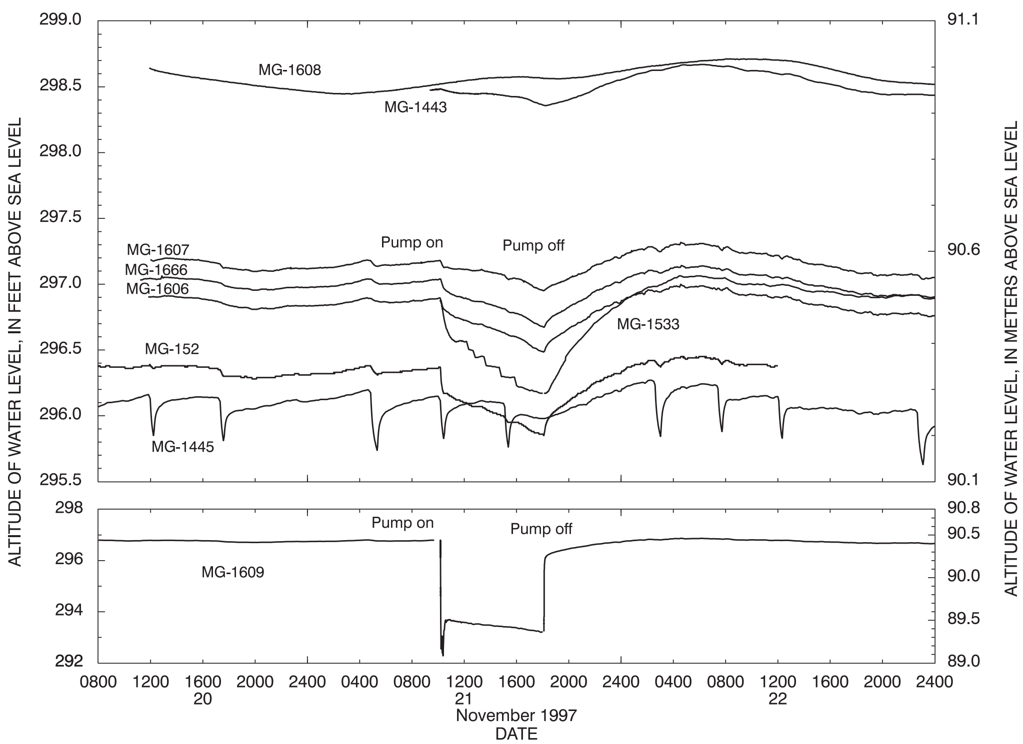

Figure 50. Measured water levels at the John Evans site in Lansdale, Pa., November 20-22, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours on November 21.

responded strongly to other pumping in the area. Other pumping also resulted in minor water-level changes in the other observation wells. For wells included in the aquifer-test analysis, drawdown was not corrected for the apparently small effects of barometric-pressure decrease or other pumping wells. The recovery of water levels in the pumped well is similar to that reported for many pumping tests in the Lansdale area (Goode and Senior, 1998). A very rapid recovery of more than 75 percent of the drawdown at the end of pumping was followed by a much more gradual recovery to the static water level.

Drawdown in four observation wells was matched by use of the two-aquifer model of Neuman and Witherspoon (1969) to estimate hydraulic properties (fig. 51). These four wells had the largest measured drawdowns. The two-aquifer model matches the measured drawdown in these four wells better than either the isotropic Theis model or the anisotropic single-aquifer model (Papadopulos, 1965). Smaller drawdown at several other observation wells could not be matched by use of this conceptual model. The estimated hydraulic properties from this match are T_1 = 1,300 ft^2/d (122 m^2/d), $S_1 = 8 \times 10^{-5}$ for the pumped 'aquifer' or network of fractures; T_2 = 15 ft^2/d (1.4 m^2/d), $S_2 = 8 \times 10^{-5}$ for the unpumped 'aquifer'; and K_v = 0.044 ft/d (0.013 m/d), and $S_s = 1 \times 10^{-6}$ /ft (3×10^{-6} /m) for the 'aquitard' (table 13). These results are consistent with the results of aquifer interval-isolation tests in that the vertical hydraulic conductivity is very low for bedrock between high-permeability zones oriented along bedding.

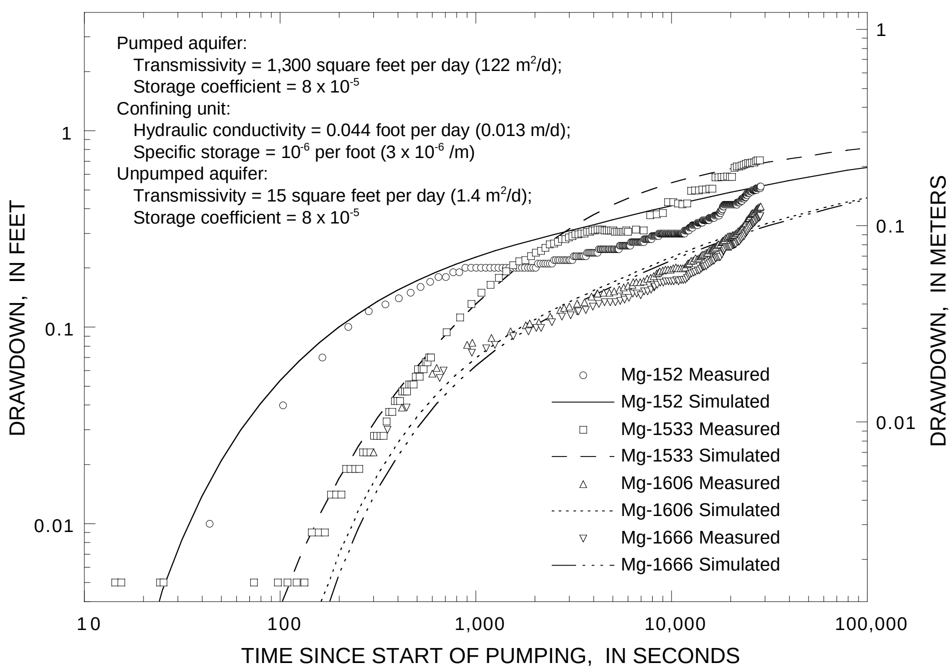

Figure 51. Measured and simulated drawdown, using two-aquifer model of Neuman and Witherspoon (1969), in wells Mg-67, Mg-80, Mg-163 and Mg-1666 at the John Evans site in Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours.

J.W. Rex site

An aquifer test at the J.W. Rex property was done by QST Environmental, Inc. (1998). Production well Mg-625 was pumped at a rate of about 40 gal/min for about 56 hours from October 24-27, 1997. Water levels in the pumped well and 10 other wells, including Mg-82, Mg-157, Mg-1441, Mg-624, Mg-1639, Mg-1640, Mg-1641, Mg-1615, Mg-1617, and Mg-1665 (pl. 1), were measured during the test. Drawdown was observed in all wells. Drawdown was greatest [11.4 ft (3.5 m)] in obervation well Mg-1639. Well Mg-1639 is the closest to the pumped well. Well Mg-1640 is within 10 ft (3 m) of well Mg-1639 but is shallower than well Mg-1639 and had much less drawdown [2.4 ft (0.7m). The downward vertical flow observed during geophysical logging prior to the aquifer tests indicates well Mg-1639 is directly influenced by pumping in production well Mg-625. Estimates of hydraulic properties were determined from analysis of drawdown data assuming an isotropic aquifer. Transmissivity ranged from 160 to 665 ft^2/d (14.5 - 61.8 m^2/d) and storage ranged from about 2×10^{-5} to 4×10^{-3} (QST Environmental Inc., 1998) (table 13). The transmissivities from this test are similar to a transmissivity of 330 ft^2/d (31 m^2/d) estimated from an earlier test (Goode and Senior, 1998).

Chemical measurements during aquifer tests

Water samples were collected during the aquifer tests to determine chemical and physical properties and the concentration of VOC’s at various times while pumping. Field measurements, including temperature, pH, specific conductance, and dissolved oxygen, were made by the USGS. Samples for VOC analysis were collected by the USGS and sent by B&V to a USEPA laboratory.

The measurements of pH, dissolved oxygen, and specific conductance and concentrations of VOC’s generally remained relatively stable during the aquifer tests of the three wells (table 14). PCE, TCE, and cis-1,2-DCE concentrations increased slightly in samples collected during the test of well Mg-1610 (table 14), suggesting that increasingly contaminated water from elsewhere on the site may have been drawn toward the pumped well. The dissolved oxygen concentration in the last sample collected during the test of well Mg-1610 was more than 3 mg/L lower than the earlier samples from the well. Slight increases in PCE, TCE, 1,1-DCE, and cis-1,2-DCE concentrations also were measured in samples from the test of well Mg-1609 at John Evans site.

Numerical Simulation of Regional Ground-Water Flow

A three-dimensional finite-difference numerical model, MODFLOW (McDonald and Harbaugh, 1988), was used to simulate regional steady-state flow. The model was calibrated using an automatic, nonlinear optimization program, MODFLOWP (Hill, 1992), that minimizes the differences between measured and simulated hydraulic heads and streamflow. MODPATH (Pollock, 1994), a particle-tracking module linked to MODFLOW, was used to calculate and display ground-water-flow pathlines from the output of the flow model.

Model and Model Assumptions

The model structure is based on a simplified conceptualization of the ground-water flow system. The weathered and fractured-rock formations were modeled as equivalent porous media, such as unconsolidated granular deposits. Thus, it is assumed that ground-water flow can be described by use of a three-dimensional flow equation based on Darcy’s Law. In this approach, the hydraulic conductivities used in the model represent the bulk properties of the fractured-rock formations. Water flux, which may pass through only a small fraction of the rock mass occupied by fractures, is simulated as distributed throughout the formations. The model cannot simulate localized ground-water flow controlled by a few, discrete permeable fractures or fractures zones. The model is assumed to approximately represent regional-flow conditions that are controlled by a large number of fractures or fracture zones distributed throughout the region.

Table 14. *Field measurements of physical and chemical properties and concentrations of selected volatile organic compounds in water samples collected during aquifer tests of wells Mg-1600, Mg-1610, and Mg-1609 in Lansdale, Pa., November 13-21, 1997*

[µg/L, micrograms per liter; °C, degrees Celsius; µS/cm, microsiemens per centimeter; mg/L, milligrams per liter; DCE, dichloroethylene; PCE, tetrachloroethylene; TCA, trichloroethane; TCE, trichloroethylene; --, not detected]

Cumulative pumped volume (gallons)	Time of sample	Physical or chemical property				Volatile organic compounds (µg/L)									
		Temperature (°C)	Specific conductance (µS/cm)	Dissolved oxygen (mg/L)	pH	Acetone	Carbon disulfide	Carbon tetrachloride	Chloroform	1,1-DCE	cis-1,2-DCE	trans-1,2-DCE	PCE	1,1,1-TCA	TCE
						Test of well Mg-1600, Rogers Mechanical site[1]									
845	14:00	12.4	627	1.0	7.09	2.2	0.1	0.06	0.1	--	0.08	--	0.2	--	1.4
925	14:10	12.5	625	1.0	7.20	--	--	.07	.1	--	.2	--	.2	0.2	2.2
1,330	15:00	12.0	628	1.0	7.40	2.7	.5	.06	.1	--	.08	--	.2	--	1.5
2,060	16:30	12.0	625	1.2	7.52	2.3	.08	.06	.1	--	.1	0.1	.2	--	1.8
2,790	18:00	12.3	627	1.4	7.51	5.0	.1	.07	.1	--	.3	.06	.2	--	4.0
						Test of well Mg-1610, Keystone Hydraulics site[2]									
920	12:20	12.9	687	4.7	6.94	5.0	--	--	.4	0.9	67.0	.4	37	.6	63.6
1,620	13:30	13.0	686	4.8	6.87	2.1	--	--	.4	.8	68.8	.3	40.7	.6	64.0
2,820	15:30	12.6	689	4.9	6.70	3.2	.04	--	.4	.8	75.1	.5	46	.6	69.7
4,670	18:35	11.5	694	1.4	6.80	1.3	100	--	.4	.8	77.2	.4	51.6	.6	72.6
						Test of well Mg-1609, John Evans site[3]									
820	11:40	15.0	673	2.4	7.04	4.7	.4	11.8	1.0	3.4	35.6	.5	78	.9	418
1,370	12:40	14.5	674	2.1	7.16	1.2	--	13.1	1.1	4.7	40.0	.5	90	1.2	526
3,000	15:40	14.5	683	2.1	7.51	3.7	.2	12.6	1.2	5.2	48	.4	100	1.4	544
4,190	17:50	14.5	686	2.4	7.14	3.4	.06	11.7	1.1	5.8	46	.8	97	1.3	530

[1] November 13, 1997.
[2] November 18, 1997.
[3] November 21, 1997.

The model grid is aligned parallel to the regional strike of the dipping sedimentary beds (45° NE.) and corresponds to the assumed major axis of anisotropy of horizontal hydraulic conductivity (fig. 52). The assumed minor axis of anisotropy, therefore, is oriented in the dip direction. Cell dimensions of the horizontal model grid were 328-ft × 328-ft (100-m × 100-m). Lateral boundaries of the model were defined as zero-flux (no flow) cells that include streams (discharge boundaries) and topographic divides that were assumed to be ground-water divides (fig. 52). Definition of the lateral boundaries was based in part on a map of water levels in the area (Senior and others, 1998). The bottom layer of the model also was defined as a no-flow boundary. The top layer of the model was defined as a constant flux boundary, where the flux equals the recharge rate.

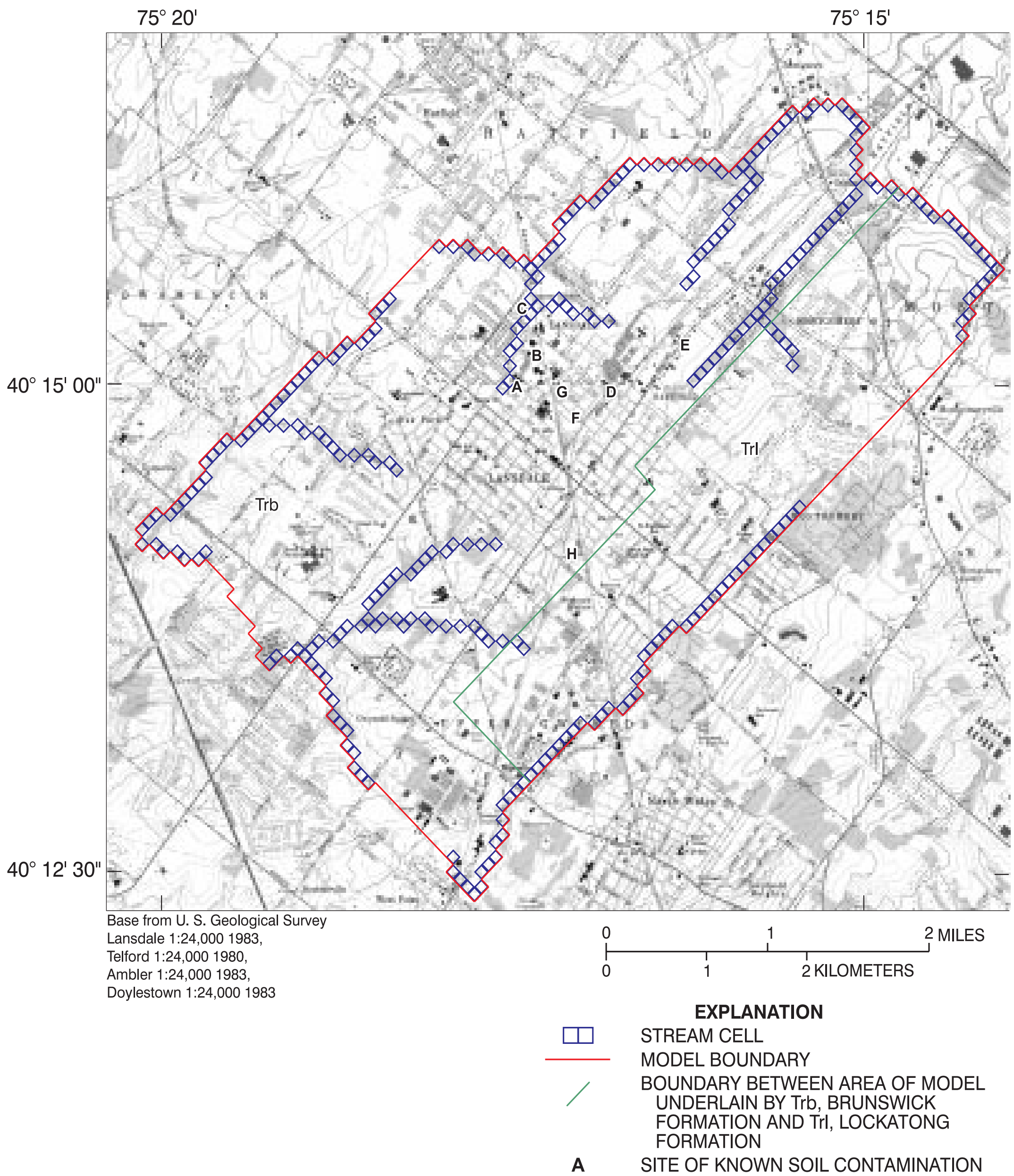

Figure 52. Boundaries and stream cells of model grid and selected areas of soil contamination in and near Lansdale, Pa.

Three model layers represent the shallow [0-40 ft (0-12 m)], intermediate [40-367 ft (12-112 m)], and deep [367-696 ft (112-212 m)] parts of the aquifer (fig. 53). The 40-ft (12-m) thick top layer (1) represents the shallow-flow system, and the 367-ft (100-m) thick second (2) and third (3) model layers represent the deep-flow system (fig. 53). The altitude of the top surface of the model was derived from digital-elevation-model data with 100-ft (30-m) grid spacing. Pumping wells fully penetrate the intermediate layer of the simulated aquifer (fig. 53).

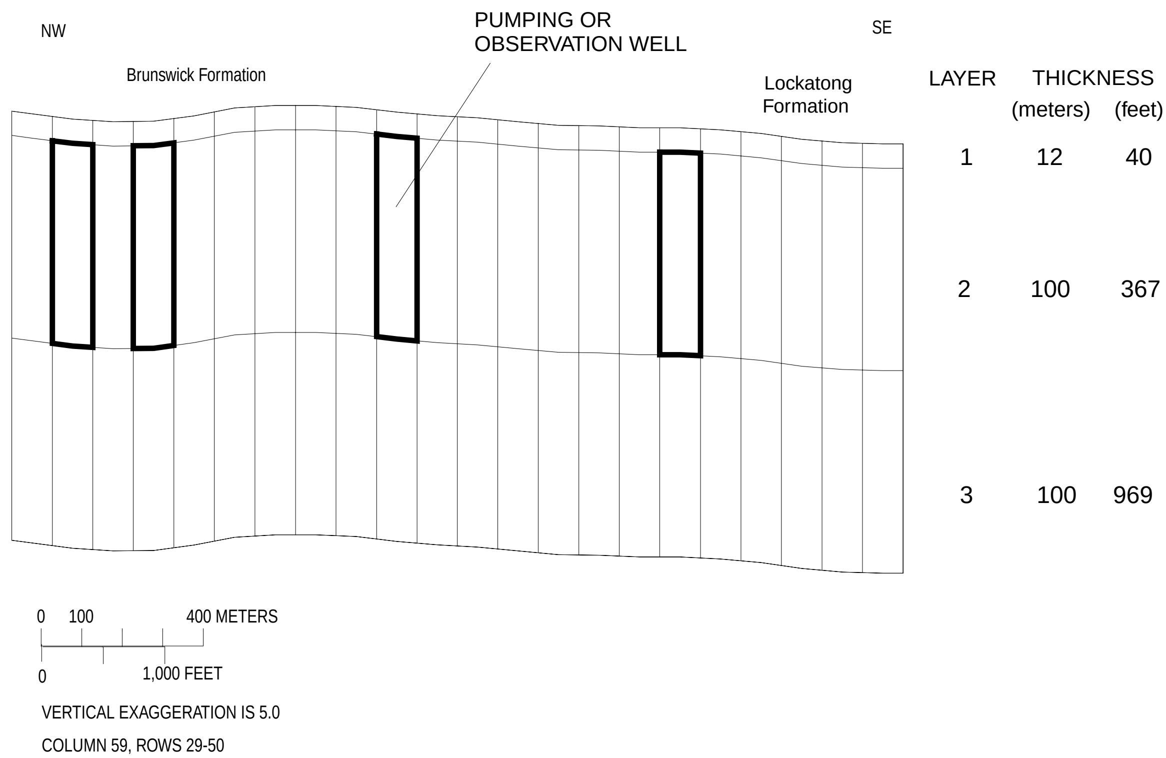

Figure 53. Model structure showing thickness of three layers and location of pumping or observation well in middle layer for simulation of ground-water flow in and near Lansdale, Pa.

The entire thickness of each model layer is assumed to be saturated. This approximation means that the transmissivity (T) of the top model layer is assumed to be independent of the computed hydraulic head. The calibration model MODFLOWP requires this approximation. The model results are relatively insensitive to minor changes in the transmissivity of the top layer because most flow is in the deeper parts of the ground-water system. Where not affected by pumping, the depth to water in the study area commonly is less than 50 ft (15 m) and was less than 30 ft (9 m) in about half of the wells measured in August 1996 (Senior and others, 1998).

Initial transmissivity estimates were determined from analyses of aquifer tests in and near Lansdale (this report; Goode and Senior, 1998). Analysis of some aquifer tests provided estimates of hydraulic conductivity (K), which can be multiplied by saturated thickness to obtain T. Because most tested wells are completed at depths within the intermediate layer [from 40 to 367 ft (12 to 112 m) below land surface], transmissivity estimates from aquifer tests pertain to this layer. Most pumping also is within this layer. The aquifer system also initially was assigned anisotropic properties on the basis of other earlier work (Longwill and Wood, 1965; Goode and others, 1997). The deep layer is assigned the same transmissivity as the intermediate layer. The hydraulic conductivity is assumed to be zero below the bottom of deep model layer, based on a review of data indicating that most water-bearing zones are at depths less than 700 ft (210 m) and because hydraulic conductivity is thought to decrease with depth (Lewis-Brown and Jacobsen, 1995). Areas underlain by the Lockatong Formation were differentiated from areas underlain by the Brunswick Group, in accordance with relatively low transmissivity of the Lockatong Formation (Longwill and Wood, 1965). This zonation of hydraulic properties is described in more detail in the section, "Calibration of Numerical Model."

The vertical hydraulic conductivity is assumed to be equal to the horizontal hydraulic conductivity. Aquifer-interval-isolation tests suggest substantial vertical anisotropy at the borehole scale with the horizontal hydraulic conductivity much higher than the vertical hydraulic conductivity. However, model calibration tests indicate that the observed heads in the intermediate model layer and the observed streamflow are insensitive to the vertical hydraulic conductivity. Furthermore, if the vertical anisotropy is assumed to be uniform throughout the aquifer system, calibration tests indicate that minimum model error is obtained with very high vertical hydraulic conductivity. Vertical fractures may not be located near some of the tested wells but may serve to connect beds at the regional scale. Open boreholes also act as high-permeability connections across bedding. The regional-scale model cannot simulate local-scale vertical flow controlled by a local network of fractures and fracture zones.

The components of the water balance for the saturated zone that are included in the model are (1) uniform recharge to the water table, (2) discharge to pumping wells, and (3) discharge to and infiltration from streams. The steady-state assumption implies that these fluxes are in equilibrium and that hydraulic head is not changing in time. In reality, these fluxes, particularly pumping rates and recharge, are changing in time, and hydraulic head changes in response to these fluctuations. The steady-state model corresponds to the average flow conditions for the month of interest and approximates the average fluxes and hydraulic head during that period. Thus, the steady-state model cannot simulate instantaneous flow conditions.

Recharge to the saturated zone is assumed to be spatially uniform because detailed spatial information on factors affecting infiltration are not available for the area of Lansdale. On average, recharge to the water table is precipitation minus surface runoff and evapotranspiration. Areal recharge enters through the top model layer, and the magnitude of recharge is determined from calibration.

The pumping rates used in the model represent annual-average rates (Pennsylvania Department of Environmental Protection, State Water Plan Division, written commun., 1995), except for some NPWA wells (table 15). NPWA wells are assigned the average pumping rate for the month of interest, if monthly data are available.

Streams are in the shallow top layer of the model, and the aquifer discharges to the stream if the hydraulic head in a model cell is higher than the hydraulic head of the stream in that cell. Streamflow can enter the aquifer if the stream's hydraulic head is higher than the head in the aquifer, provided the stream is flowing. Stream hydraulic heads are estimated from topographic information.

Calibration of Numerical Model

The numerical model is calibrated by use of MODFLOWP (Hill, 1992), a parameter-estimation program that minimizes model error. Model error is defined as the sum of squared, weighted residuals, where residuals are the differences between measured and simulated hydraulic head and streamflow. Values for aquifer discharge to streams are derived from five measurements of base flow made at five locations from May 1995 through November 1996 (table 4). Eighty-seven model cells contain observation wells in which water levels were measured in August 1996. Because few data are available for comparison of measured to simulated heads in the shallow and deep layers, the calibration of the model is relatively insensitive to changes in hydraulic conductivity in these layers.

For model calibration, average pumping rates in August 1996 are assigned to NWPA wells and annual pumping rates in 1995 are assigned to the remaining wells (table 15). On the basis of available information, pumping rates in 1996 were similar to those in 1995.

The MODFLOWP program calculates optimum values of model parameters, such as recharge rate and hydraulic conductivity, for a particular model structure. The model structure includes all quantitative information that establishes the functional relation between model parameters and predicted heads and streamflow. Although properties of model cells can be specified individually, the approach is to group cells with similar properties into zones with uniform parameters. This approach (using zones) significantly reduces the number of model parameters

Table 15. *Annual average pumping rates for wells in and near Lansdale, Pa., during model-calibration period (1996), 1994, and 1997*

[--, not numbered; gal/min, gallons per minute]

U.S. Geological Survey local well number Mg-	Owner	Owner well number	Model cell [1]		Pumping rate (gal/min)		
			Row	Column	Calibration period	1994	1997
498	North Penn Water Authority	L-23	28	39	0	18.7	25.0
143	North Penn Water Authority	L-21	30	31	0	37.4	0
593	North Penn Water Authority	L-25	30	36	0	32.7	34.1
625	J.W. Rex Co.	1	31	52	49.7	49.7	49.7
704	North Penn Water Authority	L-26	32	57	0	29.9	0
67	North Penn Water Authority	L-8	36	45	0	60.0	0
621	American Olean Tile Co.	4	38	58	0	10.0	0
69	North Penn Water Authority	L-10	39	35	36.1	63.3	68.1
1045	American Olean Tile Co.	5	41	59	0	13.4	0
620	American Olean Tile Co.	3	42	61	0	10.6	0
914	North Penn Water Authority	NP-12	42	84	59.6	60.6	54.9
566	Lehigh Valley Dairy	5	43	23	64.4	64.4	64.4
153	American Olean Tile Co.	2	42	56	0	19.8	0
59	Lehigh Valley Dairy	3	44	26	44.4	44.4	44.4
1418	Ziegler	--	44	62	4.4	4.4	4.4
140	Lehigh Valley Dairy	4	45	24	92.5	92.5	92.5
924	North Penn Water Authority	NP-21	45	85	0	65.3	0
1125	North Penn Water Authority	NP-61	47	59	125.3	125.3	125.3
875	North Wales Water Authority	NW-17	48	70	71.0	71.0	71.0
1051	North Wales Water Authority	NW-22	48	77	136.3	136.3	136.3
1198	Merck & Co.	PW9	52	11	26.1	26.1	26.1
125	Merck & Co.	PW2	58	10	[2] 94.1	[2] 94.1	[2] 94.1
130	Merck & Co.	PW7	60	17	91.0	91.0	91.0
171	Precision Tube	1	60	30	6.4	6.4	6.4
204	Precision Tube	2	60	31	6.4	6.4	6.4
126	Merck & Co.	PW3	62	13	96.7	96.7	96.7
169	Leeds & Northrup Co.	1	63	20	0	10.6	0
223	Leeds & Northrup Co.	2	63	24	0	11.6	0
77	North Penn Water Authority	L-18	63	42	70.5	71.0	67.2
75	North Penn Water Authority	L-16	63	54	43.7	32.7	43.5
124	Merck & Co.	PW1	64	10	48.4	48.4	48.4
202	North Penn Water Authority	L-22	64	34	42.9	34.1	37.6
76	North Penn Water Authority	L-17	64	36	41.8	25.1	40.4
73	North Penn Water Authority	L-14	64	47	40.6	27.9	38.5
78	North Penn Water Authority	L-19	64	51	37.3	35.0	31.9

[1] All pumping wells are simulated as fully penetrating the middle layer (40 to 367 feet below land surface) of the model.
[2] Pumping rate at cell is (rate at PW2) + [(rate at PW8)/2].

and improves the reliability of parameter estimates. Zones are determined on the basis of hydrogeologic information. Model parameters are calibrated for several different structures, and the results of these calibrations are compared to identify a calibrated model appropriate for predictive simulation.

Two hydrogeologic zones are delineated from regional geologic mapping. Zone B represents the northwestern area of the model underlain by the Brunswick Group (Trb) (fig. 52). Zone L represents the southeastern area of the model underlain by the Lockatong Formation (Trl). Model parameters for the hydraulic conductivity of the Brunswick and Lockatong zones are designated KB and KL, respectively. Homogeneous hydraulic conductivity is specified by assigning one parameter with the same value of hydraulic conductivity for both of these zones (KB = KL). In some cases, model layer 1, representing saprolite and weathered bedrock, is assigned a value of hydraulic conductivity that differs from that assigned to model layers 2 and 3. In these cases, the model parameter corresponding to the uniform isotropic hydraulic conductivity of layer 1 is designated KW.

Anisotropy of hydraulic conductivity is included in some model structures. Anisotropy refers to a dependence of hydraulic conductivity on direction. Preliminary model evaluation indicated the simulated water levels at the observation well locations, and simulated streamflow, are relatively insensitive to vertical anisotropy. Hence, only horizontal anisotropy is included. The top layer of the model is assumed to be isotropic in all cases because extensive fracture features are less likely to be important in highly weathered rock and saprolite and because preliminary model evaluation indicated the simulated water levels in layer 2, the layer with the most observed data, are not sensitive to the horizontal anisotropy of model layer 1. The model parameter describing the horizontal anisotropy of model layers 2 and 3 is designated ANI23. The parameter is the hydraulic conductivity in the dip direction (y direction in model) divided by the hydraulic conductivity in the strike direction (x direction in model) (or ANI23 = K_y/K_x). In anisotropic cases, the hydraulic conductivity parameters KB and KL are the hydraulic conductivities in the strike direction and KB = KB_x and KL = KL_x. The hydraulic conductivity in the dip direction is the value in the strike direction multiplied by ANI23. Another model parameter estimated by calibration is the uniform recharge rate, designated R.

Several alternative model structures for hydraulic-conductivity parameters were considered to evaluate the relation between model structure and calibration error (table 16). The structures varied by including one effective layer (cases 1t and 2t) or three effective layers (case 3t), one horizontal zone (case 1t) or two horizonal zones (cases 2t and 3t), and isotropy (cases with 1t.iso, 2t.iso, 3t.iso) or anisotropy (1t.ani, 2t.ani, 3t.ani) (table 16). In case 1t.iso, the hydraulic conductivity is assumed to be isotropic and uniform throughout the entire model domain. In case 1t.ani, the

Table 16. *Hydraulic conductivity, anisotropic ratios of hydraulic conductivity, recharge rates, and calibration errors for calibrated cases of different model structures used for simulation of ground-water flow in and near Lansdale, Pa.*

[KB, hydraulic conductivity of Brunswick zone; KL, hydraulic conductivity of Lockatong zone; KW, hydraulic conductivity of model layer 1 representing saprolite and weathered bedrock; ANI23, anisotropy ratio of model layers 2 and 3; R, recharge; SSR, sum of squared, weighted residuals; ft/d, feet per day; ft^2, square feet]

Case	Model parameter					Calibration error
	KB (ft/d)	KL (ft/d)	KW (ft/d)	ANI23	R (ft/d)	SSR (ft^2)
1t.iso	2.53	=KB[1]	=KB	ONE[2]	0.0019	91,280
1t.ani	3.31	=KB	=KB	0.04	.0018	25,190
2t.iso	4.56	0.22	=KB	ONE	.0020	86,110
2t.ani	4.69	1.05	=KB	.041	.0019	16,360
3t.iso	11.4	.19	0.013	ONE	.0017	50,910
3t.ani	5.35	1.12	.161	.090	.0019	16,360

[1] =KB not estimated; set equal to KB.
[2] ONE not estimated; set equal to 1.0.

hydraulic conductivity also is assumed to be uniform throughout the entire model domain, but horizontal anisotropy is included to allow the optimal hydraulic conductivity in the dip direction to differ from the optimal hydraulic conductivity in the strike direction. In case 2t.iso, different hydraulic conductivities are assigned to the Brunswick and Lockatong zones. Hydraulic conductivities in both zones are assumed to be uniform with depth and isotropic. In case 2t.ani, different hydraulic conductivities are assigned to the Brunswick and Lockatong zones and horizontal anisotropy is included for model layers 2 and 3, which represent unweathered bedrock. Because of limitations in the input structure of MODFLOW, the anisotropy ratios of the Brunswick and Lockatong zones are assumed to be identical. In cases 3t.iso and 3t.ani, a separate model parameter represents the uniform isotropic hydraulic conductivity of model layer 1, which represents saprolite and weathered bedrock. Case 3t.iso assumes that hydraulic conductivity of layer 1 and the Brunswick and Lockatong zones are isotropic, whereas case 3t.ani includes one parameter for the horizontal anisotropy of both the Brunswick and Lockatong zones in model layers 2 and 3.

The calibrated model parameters for several alternative model structures are listed in table 16. These optimum values yield simulated hydraulic head and streamflow for each model structure that best match the measured water levels and streamflow. Changes in the model structure, for example, changing which cells represent the Brunswick Group and which represent the Lockatong Formation, would result in different optimum model parameter values. The model error excludes the contribution from the computed streamflow that corresponds to the measurement at SW-13. In the model, the stream is dry or virtually dry in all simulations.

The overall model error (sum of squared, weighted residuals, SSR) decreases as the number of model parameters is increased. From these results, the incorporation of regional horizontal anisotropy is judged to be an important model feature. Separation of the model zones corresponding to the Brunswick Group and the Lockatong Formation also substantially reduces model error and yields different hydraulic conductivities for these zones. Separation of the hydraulic conductivity of the saprolite and weathered zone (model layer 1) yields no appreciable decrease in the model error (difference between cases 2t.ani and 3t.ani, table 16). However, the optimum hydraulic conductivity for model layer 1 is significantly lower than the hydraulic conductivities of the underlying unweathered rock, in agreement with previous observations of relative hydraulic conductivities in these Triassic rocks (Longwill and Wood, 1965). Therefore, the model structure “3t.ani” is chosen for further evaluation and predictive simulation. Because the shared model parameters of structures “3t.ani” and “2t.ani” are similar, simulated water levels and ground-water fluxes should be similar with either set of estimated parameters.

All the high-permeability bed-oriented features contributing to aquifer transmissivity are included into model layers 2 and 3. The actual aquifers may contain many more permeable zones in the top 656 ft (200 m) of unweathered rock, but that level of detail is not included in this regional-flow model. The two-aquifer model used to analyze the aquifer test of well Mg-1609 at the John Evans site identified two aquifers differing in permeability, the pumped aquifer and an overlying unpumped aquifer, separated by a low-permeability bed. Both low-permeability and high-permeability parts of the formation are included within the unweathered bedrock of model layers 2 and 3. In the analysis of the aquifer test at the John Evans site, the transmissivity of the overlying unpumped aquifer is less than that of the pumped aquifer. Although the shallow observation well in the test at the John Evans site is deeper than the thickness of model layer 1, the relation of low-permeability aquifer materials above high-permeability aquifer materials is similar to the relation between model layer 1 and the underlying model layers 2 and 3. The top model layer corresponds to the saprolite and weathered zone lying above the upper aquifer at the John Evans site.

Calibration Errors

The calibrated flow model describes the regional-scale average flow conditions during August 1996 (fig. 54). The contour map of hydraulic head in the intermediate model layer (2) is similar to the contour map of observed water levels in bedrock wells (fig. 19). These similarities include steep head gradients in the Lockatong Formation, a “flat” potentiometric surface underlying the borough of Lansdale, and flow generally away from Lansdale towards regional stream-discharge areas. Pumping has a strong influence on water levels, particularly in the southern part of the modeled area, where public supply and industrial pumping rates are high.

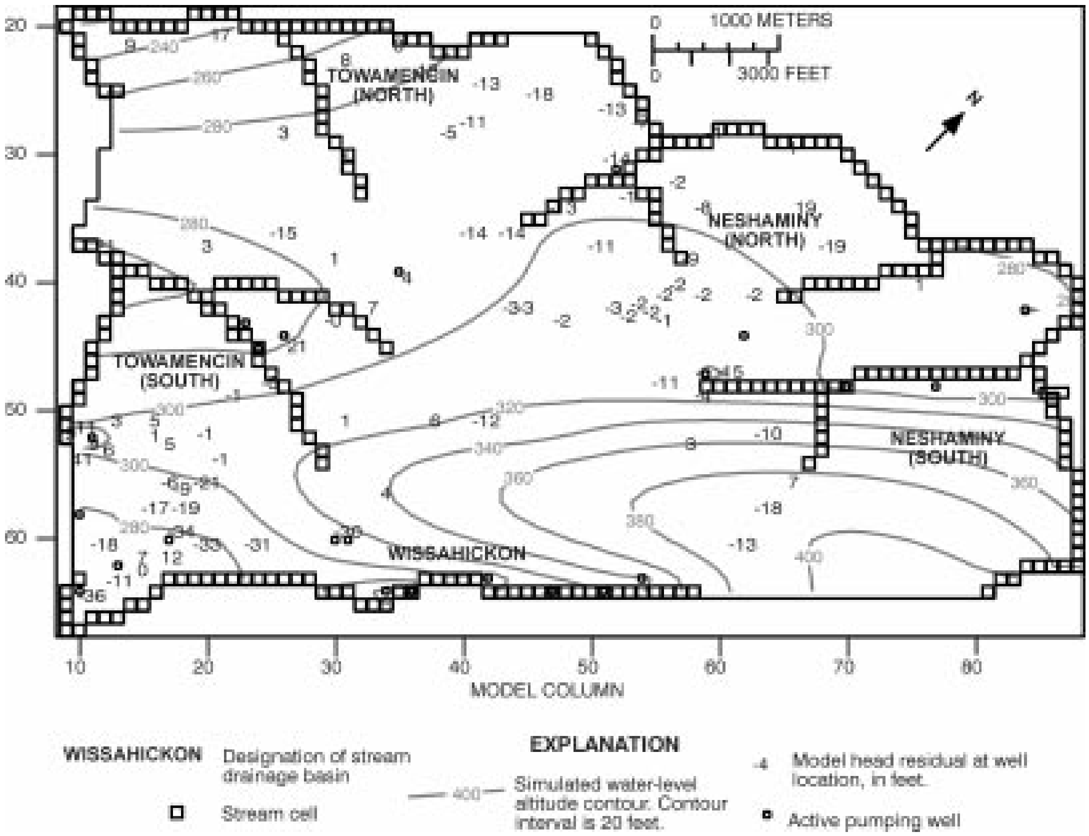

Figure 54. Simulated hydraulic head in model layer 2 representing the upper 328 feet of unweathered, fractured bedrock in and near Lansdale, Pa., and model head residual. The model head residual is the simulated hydraulic head minus the observed hydraulic head.

The root mean square residual for hydraulic head is 13 ft (4.0 m); ground-water-level differences are from -14 to +7 ft (-4 to +2 m) near the center of the model in the area of Lansdale. Maximum head residuals of -36 ft and +41 ft (-11 to +12 m) occur near the southern boundary (bottom left boundary, fig. 54), an area of intense industrial pumping that is outside the main area of interest for this study. These larger residuals may represent the inaccuracy of the regional-scale model in simulating local-scale effects of large pumping wells in this area.

Another feature of the measured water levels that is not reproduced by the model is the local water-level high in the area of the Keystone site (potential source location A), located in the central part of the model where the residuals are about -14 ft (-4 m) at three locations. A uniform transmissivity for the Brunswick Group is used in the model, but these relatively high water levels may be the result of lower permeability at this location or nearby than elsewhere in the modeled area underlain by the Brunswick Group. However, aquifer tests done in 1997 for this study and done prior to 1995 (Goode and Senior, 1998) indicate transmissivity at the Keystone Hydraulics site is higher than at several other locations in the modeled area.

Simulated streamflow agrees reasonably well with four of the observed values, but the optimized model does not include any net streamflow for the stream segments corresponding to the measurement at the site SW-13, Wissahickon Creek near Hancock Street (table 17). All the model structures tested simulated near-zero streamflow for the model stream cells corresponding to site SW-13. This stream is along the southeastern boundary of the modeled area and has been known to go dry during periods of low rainfall. The streamflow residuals are multiplied by a constant weight of 465 ft /(ft^3/s) [0.058 m/(m^3/d)] to account for the difference in units and measurement errors between head and streamflow (see Hill, 1992, p. 38). The chosen weight value yields weighted residuals for

Table 17. *Measured and simulated streamflow for calibrated numerical model of ground-water flow in and near Lansdale, Pa.*

[ft^3/s, cubic feet per second; ft, feet]

Site number	Model cell[1]		Streamflow		Calculated residual (ft^3/s)	Weighted residual[3] (ft)
	Row	Column	Simulated (ft^3/s)	Measured[2] (ft^3/s)		
SW-21, tributary to Towamencin Creek at Troxell Rd.	19	22	0.459	0.411	0.048	22
SW-3, tributary to W. Branch Neshaminy Creek at Cowpath Rd near Kulp Rd.	29	66	.126	.098	.028	13
SW-10, tributary to W.Branch Neshaminy Creek near Line & Cowpath Rd.	48	69	0	.022	-.022	-10
SW-13, Wissahickon Creek at Hancock St. (and at Wissahickon Ave.)	64	29	[4] 0	.170		
SW-17, Towamencin Creek at Sumneytown Pike	39	15	.807	.762	.044	20

[1] All stream cells are in the top layer (1) of the model.

[2] Measured streamflow estimated from five base-flow measurements May 1995 through November 1996; flow was weighted at SW-21 by 70 percent and at both SW-3 and SW-13 by 50 percent to account for reduced amount of contributing areas in these streams at the boundaries of the model.

[3] Weight is 465 feet per cubic feet per second for all flux measurements.

[4] The measurement was not used in the model calibration procedure because all cells of the stream were dry during parameter-estimation iterations (Hill, 1992).

streamflow that are in the same range as head residuals. A smaller weight value would reduce the weighted residuals and the importance of the streamflow measurements in the overall parameter estimation, whereas a larger weight value would increase the importance of streamflow measurements relative to head measurements.

The accuracy of the nonlinear regression methods used here for estimating model parameters is based, in part, on the assumption of normally distributed, independent residuals. Hill (1992) proposes a hypothesis test of normality and independence of weighted residuals. This test compares the correlation coefficient between the ordered weighted residuals and order statistics from the normal distribution. For case “3t.ani,” this correlation coefficient is 0.978. This value is slightly greater than the critical value (0.977) for the 0.10 significance level, indicating the residuals are nearly normally distributed and independent. This suggests the optimum parameters for this model are accurately identified by use of these procedures.

Estimated Large-Scale Hydraulic Conductivity and Recharge

The calibrated model parameters are estimates of the large-scale hydraulic properties controlling ground-water flow in and near Lansdale. Calibrated parameters and estimated confidence intervals are shown in table 18. The confidence intervals correspond to plus and minus two standard deviations from the estimated value. These confidence intervals are based on the assumption that the optimization model is linear near the calibrated parameters. Furthermore, these confidence intervals represent only the uncertainty in the parameter in question under the condition that all other model parameters are held constant. The modified Beale’s measure is computed to examine nonlinearity in the optimization model (Cooley and Naff, 1990). For case 3t.ani, this measure is 19.1, which indicates the model is highly nonlinear. The model is nonlinear if the modified Beale’s measure is greater than 0.43, and it is effectively linear if the measure is less than 0.04. Examination of the output of program BEALEP (Hill, 1994) indicates parameter KW, the hydraulic conductivity of the top model layer, contributes most to the nonlinearity. To test the effect of this parameter on the model nonlinearity, parameter KW is set to its optimal value, 0.16 ft/d (0.049 m/d), and removed from the parameter estimation. For this test case without estimation of parameter KW, the modified Beale’s measure is 0.04 and indicates the model is effectively linear. This implies the linear confidence intervals on the other four parameters may be meaningful, even though the measure indicates the model is highly nonlinear with all parameters included.

Table 18. *Optimum and approximate, individual, 95-percent confidence-interval values for hydraulic conductivity, anisotropic ratio, and recharge or calibrated simulation of ground-water flow in and near Lansdale, Pa.*

[KB, hydraulic conductivity of Brunswick zone; KL, hydraulic conductivity of Lockatong zone; KW, hydraulic conductivity of model layer 1 representing saprolite and weathered bedrock; ANI23, anisotropy ratio of model layers 2 and 3; R, recharge; ft/d, feet per day; -, dimensionless; in/yr, inches per year]

Parameter	Units	Optimum value	Approximate, individual, 95-percent confidence interval	
			Lower value	Upper value
KB	ft/d	5.35	4.04	7.05
KL	ft/d	1.12	.89	1.40
KW	ft/d	.16	.01	2.00
ANI23	-	.090	.060	.119
R	in/yr	8.3	7.9	8.8

The approximate, individual, 95-percent confidence intervals show the hydraulic conductivities of the Brunswick and Lockatong zones are relatively tightly constrained in the optimum model but the hydraulic conductivity of the weathered zone is poorly described. This poor description is probably the result of a lack of water-level data in the top layer of the model. Only two measurements are assigned to that layer. Recharge also is tightly constrained, because streamflow observations are used in the calibration and the specified pumping constitutes a large percentage of the water balance.

The transmissivity of the weathered zone (layer 1) is estimated as 0.16 ft/d (hydraulic conductivity) $\times$ 40 ft (layer thickness) = 6.4 ft^2/d (0.59 m^2/d). The transmissivity of the underlying Brunswick Group (layers 2 and 3) in the strike direction is estimated as 5.35 ft/d $\times$ 656 ft = 3,510 ft^2/d (326 m^2/d). The transmissivity of the Brunswick Group in the dip direction is estimated as 3,510 ft^2/d $\times$ 0.090 = 316 ft^2/d (29 m^2/d). The geometric mean (square root of the product) of the directional transmissivities corresponds to the “effective” isotropic transmissivity controlling drawdown because of pumping (Kruseman and de Ridder, 1990, p. 134). For the Brunswick Group, the geometric mean transmissivity is about 1,050 ft^2/d (97 m^2/d). The transmissivity of the unweathered part of the Lockatong Formation is similarly estimated as 732 ft^2/d (68 m^2/d) in the strike direction and 64 ft^2/d (6 m^2/d) in the dip direction, with a geometric mean of 215 ft^2/d (20 m^2/d). Most water moving horizontally through the model does so in layers 2 and 3, representing unweathered fractured rock. The transmissivity of the zone representing the Brunswick Group is higher than that of the Lockatong Formation zone.

The calibrated recharge rate is 8.3 in/yr (212 mm/yr). This value is somewhat higher than regional estimates of recharge from long-term-average base flow to streams overlying the Brunswick Group and Lockatong Formation (White and Sloto, 1990). The streamflow measurements and assumed pumping rates strongly control the estimated recharge rate. Lower estimated recharge would be obtained by use of lower pumping rates and lower streamflow measurements. Lower streamflow or pumping rates used for calibration also would lead to lower estimated hydraulic conductivity and transmissivity. It is not known how the observed streamflow compares to long-term streamflow because long-term measurements are not available for these streams.