SIMULATION OF GROUND-WATER FLOW AT THE LOCAL SCALE

Aquifer tests provide information about aquifer properties and local-scale ground-water flow. Drawdown and changes in ground-water flow on a local scale near a pumping well can be highly variable because of aquifer heterogeneity. Results of aquifer tests in Lansdale indicate that transmissivity differs in both vertical and horizontal directions in the fractured sedimentary rocks that underlie the area (Goode and Senior, 1998; Senior and Goode, 1999). The extent of hydraulic connection between water-bearing fractures is not necessarily related to distance between the fractures but may be related to geologic structure. For example, wells with water-bearing zones located in the projected bed of the pumped interval responded to pumping in aquifer tests (Senior and Goode, 1999). In the Triassic-age sedimentary rocks of the Brunswick Group and the Lockatong Formation, cones of depression caused by pumping have been observed to extend preferentially along strike of bedding planes or in the direction of fracture orientation (Longwill and Wood, 1965).

Local ground-water flow is defined for this report as flow to a single pumping well from a distance of about 1,000 ft (300 m). In this report, data from three multiple-well aquifer tests done in 1997 (QST, Inc. 1998; Senior and Goode, 1999) are used to simulate local-scale ground-water flow for two areas in Lansdale. Detailed local ground-water flow within a borehole between sets of water-bearing fractures is beyond the scope of the simulations in this report.

Approach

A three-dimensional finite-difference numerical model, MODFLOW (Harbaugh and McDonald, 1996), is used to simulate local flow. The model is 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, is used to calculate and display ground-water-flow pathlines from the output of the flow model. This general approach is the same as that used by Senior and Goode (1999) for a regional-scale model of flow in the Lansdale area.

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 using 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 if it were distributed throughout all parts of the formations. Local-scale ground-water flow in fractures and fracture zones is modeled as occurring in stratigraphic beds of high hydraulic conductivity. This approach captures the dominant effect of zones of high hydraulic conductivity on local-scale flow. Detailed characteristics of flow within fractures or within individual beds at scales of a few feet or less are not accurately simulated by the models.

The entire thickness of rock represented by 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).

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 significantly reduces the number of model parameters and improves the reliability of parameter estimates. Zones are delineated on the basis of hydrogeologic information.

Limitations and Uncertainties in Predictive Simulations

The contributing areas for pumping wells in the Lansdale area are approximated by the predictive simulations in this report. Although the calibrated models match many of the measured water-level changes during pumping and the regional model reasonably matches overall regional water-level trends, the measurements are not precisely reproduced by the models. Furthermore, steady-state flow under alternative pumping conditions cannot be measured to compare to the model simulations. The actual ground-water flowpaths are likely to be more complex than those shown here because of the highly heterogeneous characteristics of the fractured-rock aquifers, and the flowpaths are likely to change in time because of changing recharge and pumping conditions. The results here can be used to compare the potential effects of alternative ground-water management methods and to indicate general characteristics of contributing areas for these wells. The uncertainties in the predictive simulations could be reduced by more detailed field studies and longer-term aquifer and tracer tests, which are beyond the scope of this report.

North-Central Lansdale Model

A separate local-scale model is constructed for aquifer-test analysis and particle tracking in north-central Lansdale (fig. 4). This area does not contain any streams draining the ground-water system. Hence, the components of the water budget include only recharge, pumping within the area, and fluxes to and from the parts of the aquifer outside the local-scale model area. Furthermore, the aquifer-test results provide information only about a single high-permeability bed (Senior and Goode, 1999). These features allow a small local-scale model to be used that incorporates boundary fluxes determined from the regional-scale model. This approach is computationally efficient but cannot be efficiently used in locations where streams are present within the local-scale model area, or where a regional hydrogeologic structure is simulated beyond the boundaries of the local-scale model.

Aquifer-Test Results

One aquifer test was done at the John Evans and Sons property (Evans) on November 21, 1997 (Senior and Goode, 1999). 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. 6) with pressure transducers and electric tapes. Barometric pressure at a nearby site also was recorded with a transducer. The configuration of wells 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-152) 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. 7 and 8). Bedding strikes about N45°E and dips about 12° NW in the vicinity of the site (Conger, 1999). Pumping for industrial use occurred intermittently during the aquifer test in well Mg-153 near well Mg-618 (fig. 6).

Positive drawdown during the aquifer test was measured in the pumped well and in 7 of the 10 observation wells (fig. 6). Negative drawdown, probably due in part to barometric effects, was measured in observation wells Mg-618, Mg-1608, 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. 7 and 8); Mg-152, the next-closest observation well 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 downdip of the pumped well but open to the same beds (fig. 7 and 8). 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 more than 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. 7).

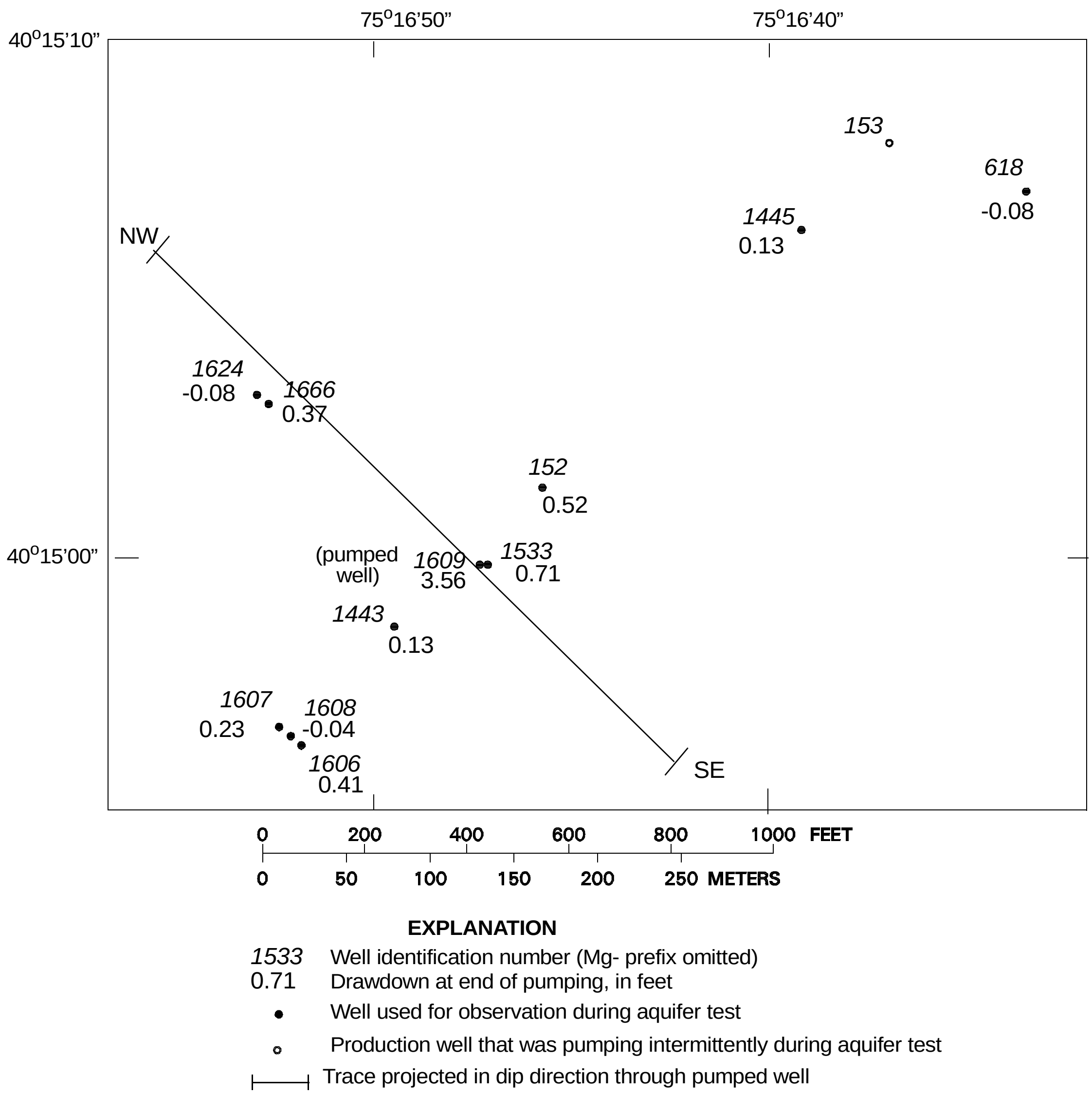

Figure 6.-- Well locations and drawdown at end of pumping well Mg-1609 at the John Evans and Sons property in north-central Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours (from Senior and Goode, 1999).

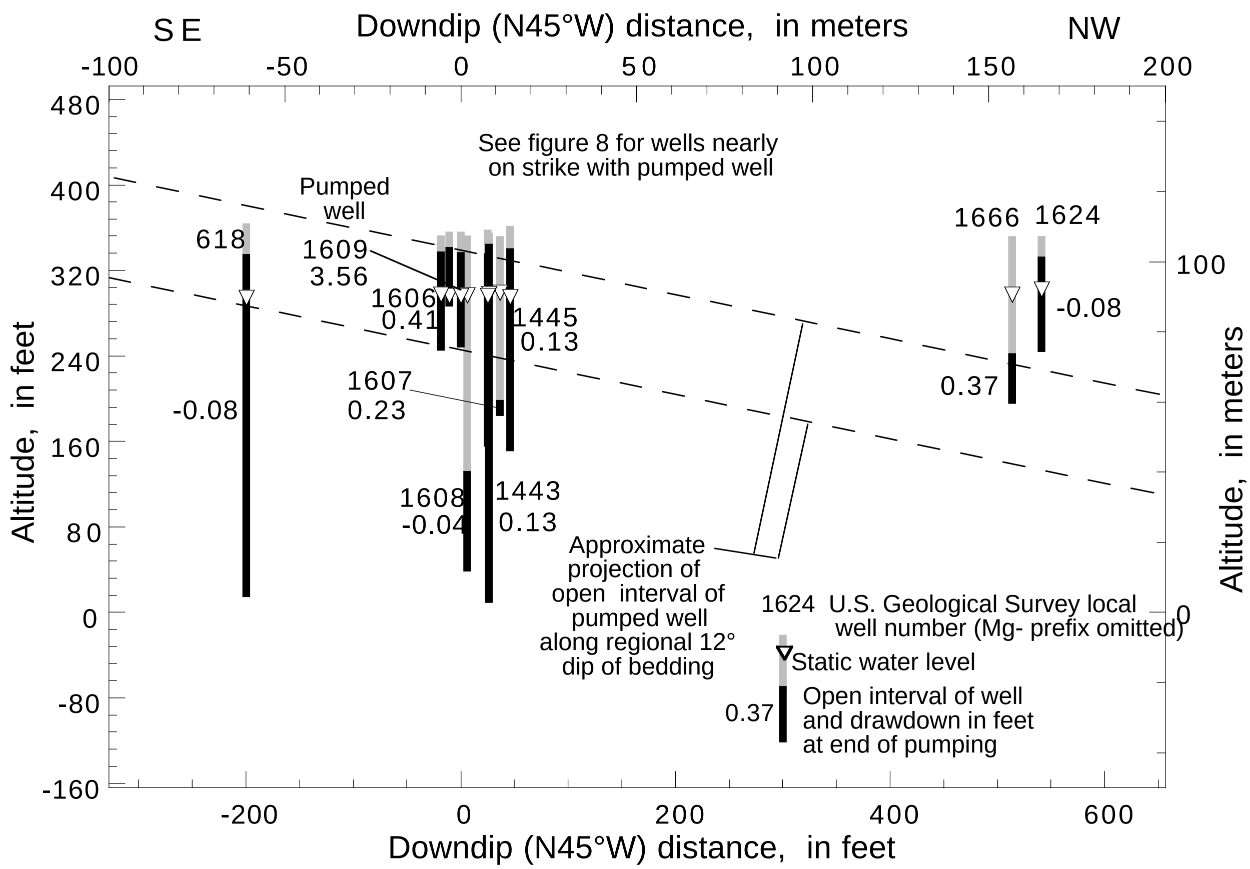

Figure 7.-- Cross-section of open intervals of wells, static depth to water, and drawdown at end of pumping at the John Evans and Sons property in north-central Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours. All wells are projected onto a vertical plane parallel to the dip direction (from Senior and Goode, 1999).

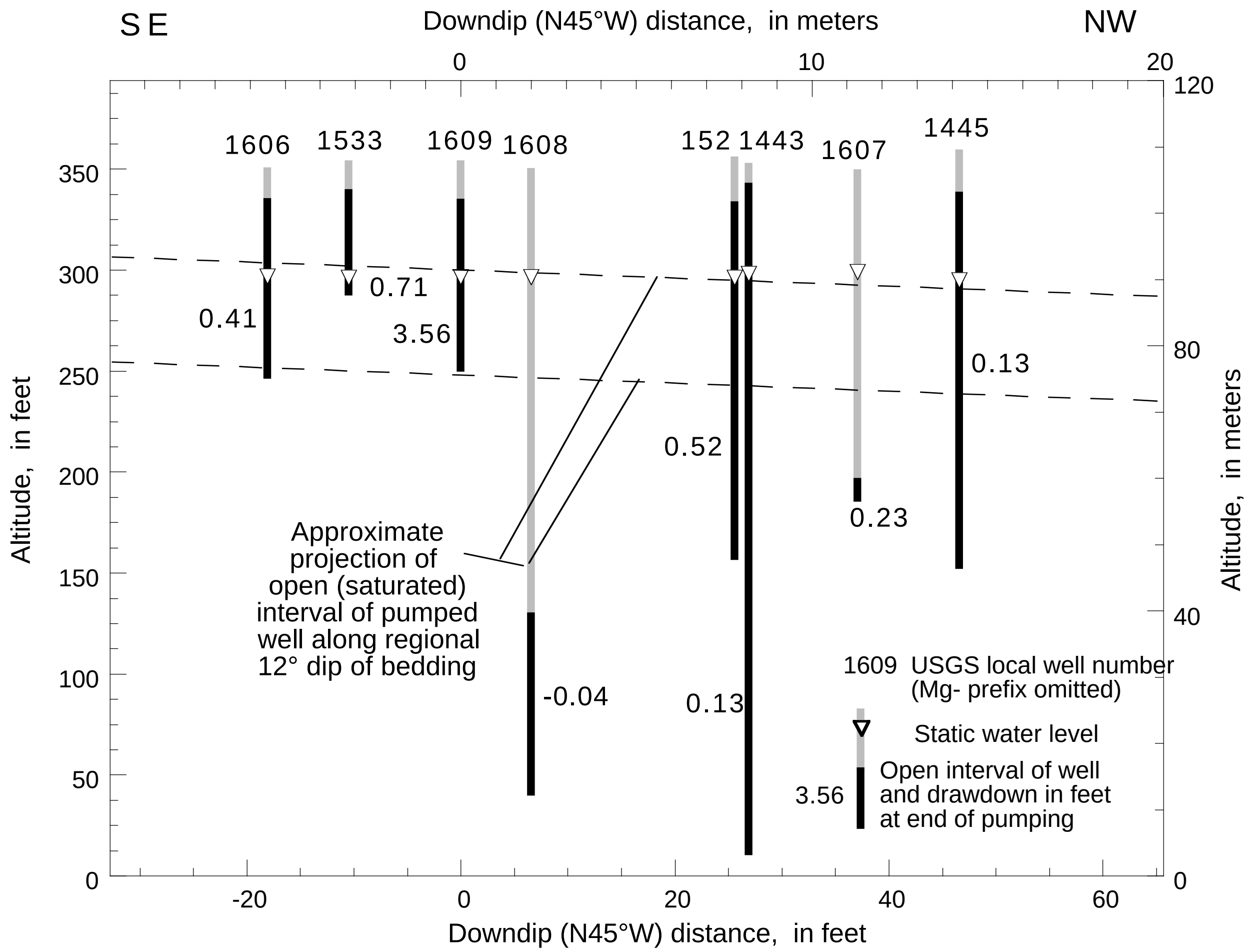

Figure 8.-- Cross-section of open intervals of wells nearly on strike with the pumped well, static depth to water, and drawdown at end of pumping at the John Evans and Sons property in north-central Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours. All wells are projected onto a vertical plane parallel to the dip direction (from Senior and Goode, 1999).

Measured water levels during the aquifer test illustrate the effect of pumping, including variable rates of pumping at the beginning of the test and fluctuations associated with regional water-level trends (fig. 9). 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-1608 (figs. 8 and 9) 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-1608 did respond to changes in barometric pressure (fig. 9) 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 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.

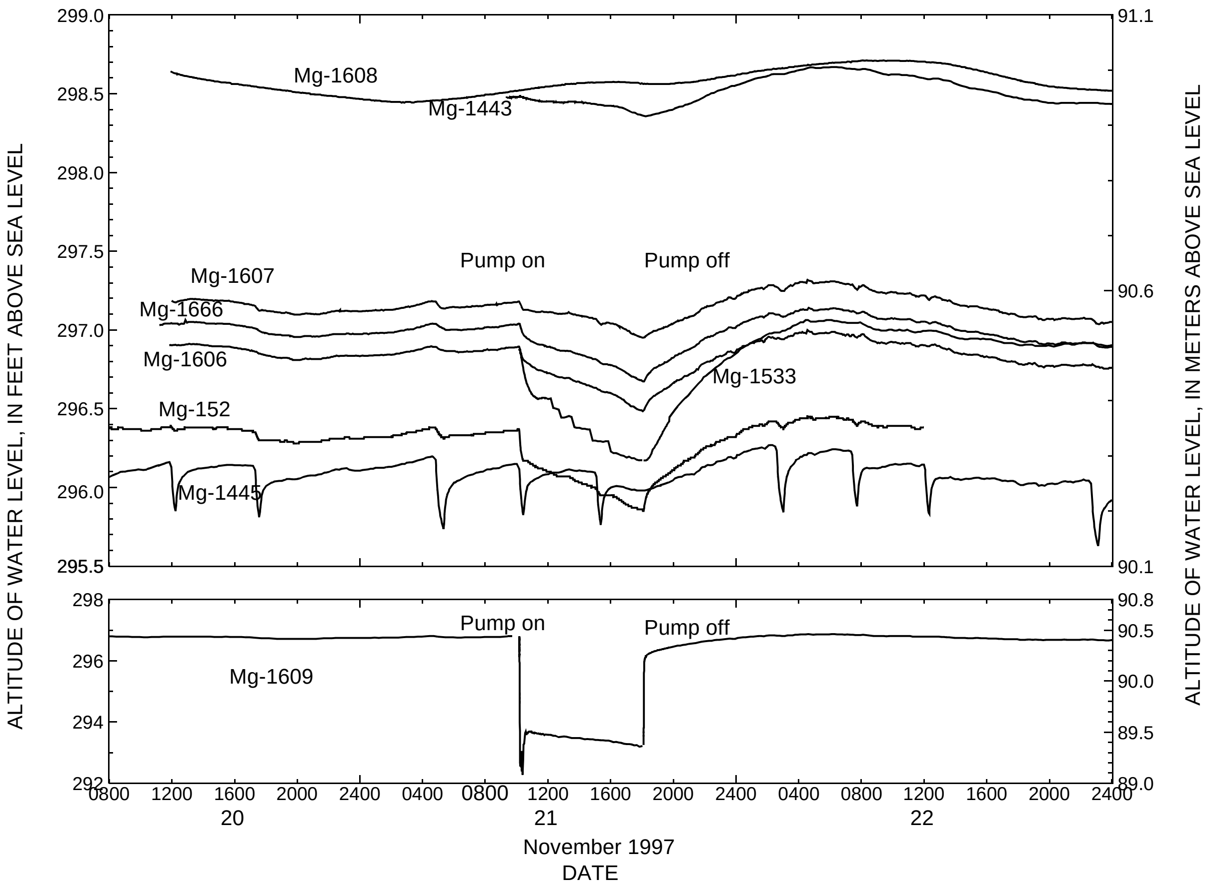

Figure 9.-- Measured water levels at the John Evans and Sons property in north-central Lansdale, Pa., November 20-22, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours on November 21 (from Senior and Goode, 1999).

Senior and Goode (1999) matched drawdown in four observation wells using the two-aquifer analytical model of Neuman and Witherspoon (1969) to estimate transmissivity (T), storage coefficient (S), hydraulic conductivity (K), and specific storage (S_s) (fig. 10). 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 using 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 x 10^{-5} for the pumped 'aquifer' or network of fractures; T_2 = 15 ft^2 / d (1.4 m^2 / d), S_2 = 8 x 10^{-5} for the unpumped 'aquifer'; and K_v = 0.044 ft / d (0.013 m / d), and S_s = 1 x 10^{-6} / ft (3 x 10^{-6} / m) for the low-permeability unit separating the two aquifers. These results are consistent with the results of aquifer interval-isolation tests (Senior and Goode, 1999) in that the vertical hydraulic conductivity is very low for bedrock between high-permeability zones oriented along bedding.

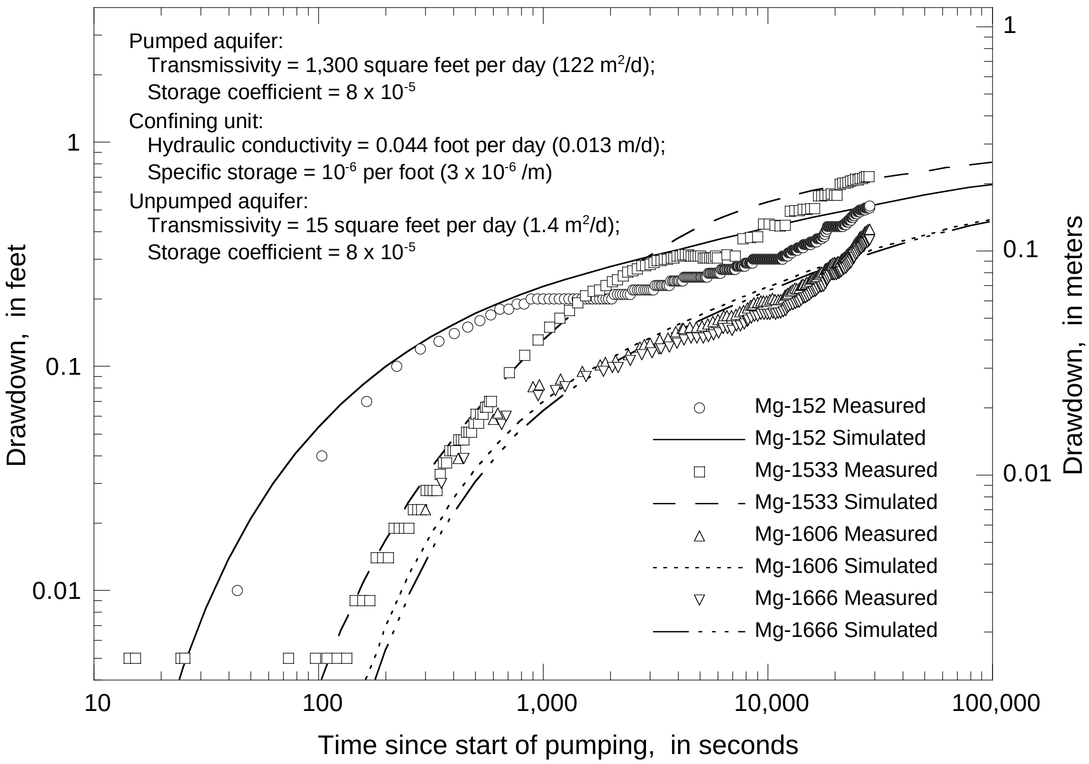

Figure 10.-- Measured and simulated drawdown, using two-aquifer model of Neuman and Witherspoon (1969), in wells Mg-67, Mg-80, Mg-163, and Mg-1666 at the John Evans and Sons property in north-central Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours (from Senior and Goode, 1999).

Model Structure and Boundary Conditions

The local-scale model of ground-water flow in the north-central part of Lansdale (fig. 4) is based on the regional-scale model. The thickness of the entire aquifer is divided into three major layers and includes a dipping high-permeability bed within the bedrock (fig. 11). The hydrogeologic layers are divided into 12 layers for model computations. The soil zone is represented by model layer 1 and is uniformly 16.4 ft (5 m) thick. The upper weathered part of the bedrock is represented by model layer 2 and is uniformly 16.4 ft (5 m) thick. The unweathered fractured bedrock is represented by model layers 3 through 12; the thickness of the layers increases progressively with depth from 16.4 to 82 ft (5 to 25 m). The bedrock layers in this model correspond to layers 2 and 3 of the regional-scale model of Senior and Goode (1999). The pumped well is open to a dipping, high-permeability bed that extends throughout the area of the local model. The vertical position of this bed depends on the local dip and strike. Hence, this pumped bed is represented by a stair-step configuration of high-permeability cells occurring in all model layers 2-12 (fig. 11).

Areally, model rows are aligned with the strike of the local stratigraphy (fig. 4). The horizontal dimensions of model grid cells range from 9.8 by 9.8 ft (3 by 3 m) at the pumping well to 65.6 by 65.6 ft (20 by 20 m) for cells more than about 164 ft (50 m) from the pumping well.

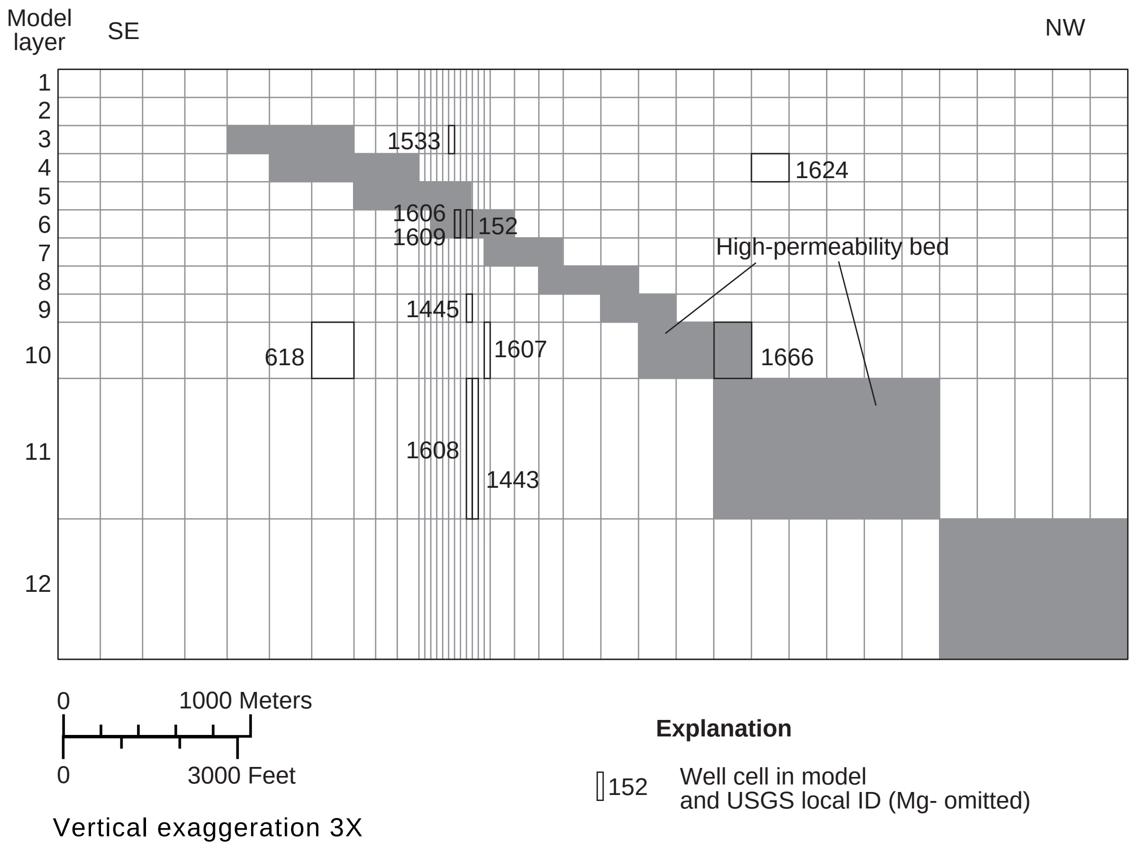

Figure 11.-- Cross section of local-scale model of ground-water flow at the John Evans and Sons property in north-central Lansdale, Pa. Well locations in the model are indicated; wells 1606 and 1609 are collocated in the cross section but are in different model columns. See fig. 4 for location of local-scale model.

Two different model configurations (fig. 12) are used: (a) a local-scale model for calibration to aquifer-test results; and (b) a local-scale model with overlap areas beyond the boundary of the local-scale model for ground-water-flowpath simulation. The local-scale model is calibrated to the aquifer-test results assuming that there is insignificant interaction between the local model and regional flow. For model calibration to the aquifer-test results, initial heads are specified as zero throughout the local model domain, and no-flow conditions are applied along all model boundaries.

Ground-water flowpaths under steady-state conditions are simulated with overlap areas around the local model to accommodate the specified flux boundary conditions from the regional-scale model. The outer five columns on the SW and NE ends of the model and the outer three rows on the NW and SE edges of the model are overlap areas beyond the boundaries of the local-model domain (fig. 12). In the regional model, the bedrock is homogeneous; hence, flow is nearly evenly distributed vertically. In the local-scale model, however, some of the cells along the boundary between the local and regional models have high permeability, and others have low permeability. Rather than setting a similar specified flux in cells with different properties, the specified flux is applied on the outside of the model overlap region (fig. 12). In this overlap region, properties are uniform. The specified boundary flux from the regional-scale model is applied on the outside of the overlap area. The same overall flux occurs at the local / regional boundary, but most of the flow occurs where high-permeability cells are located. The specified flux values are determined by regional-scale model simulation using pumping rates of the “1997” simulation of Senior and Goode (1999) with the additional specified pumping at well Mg-1609.

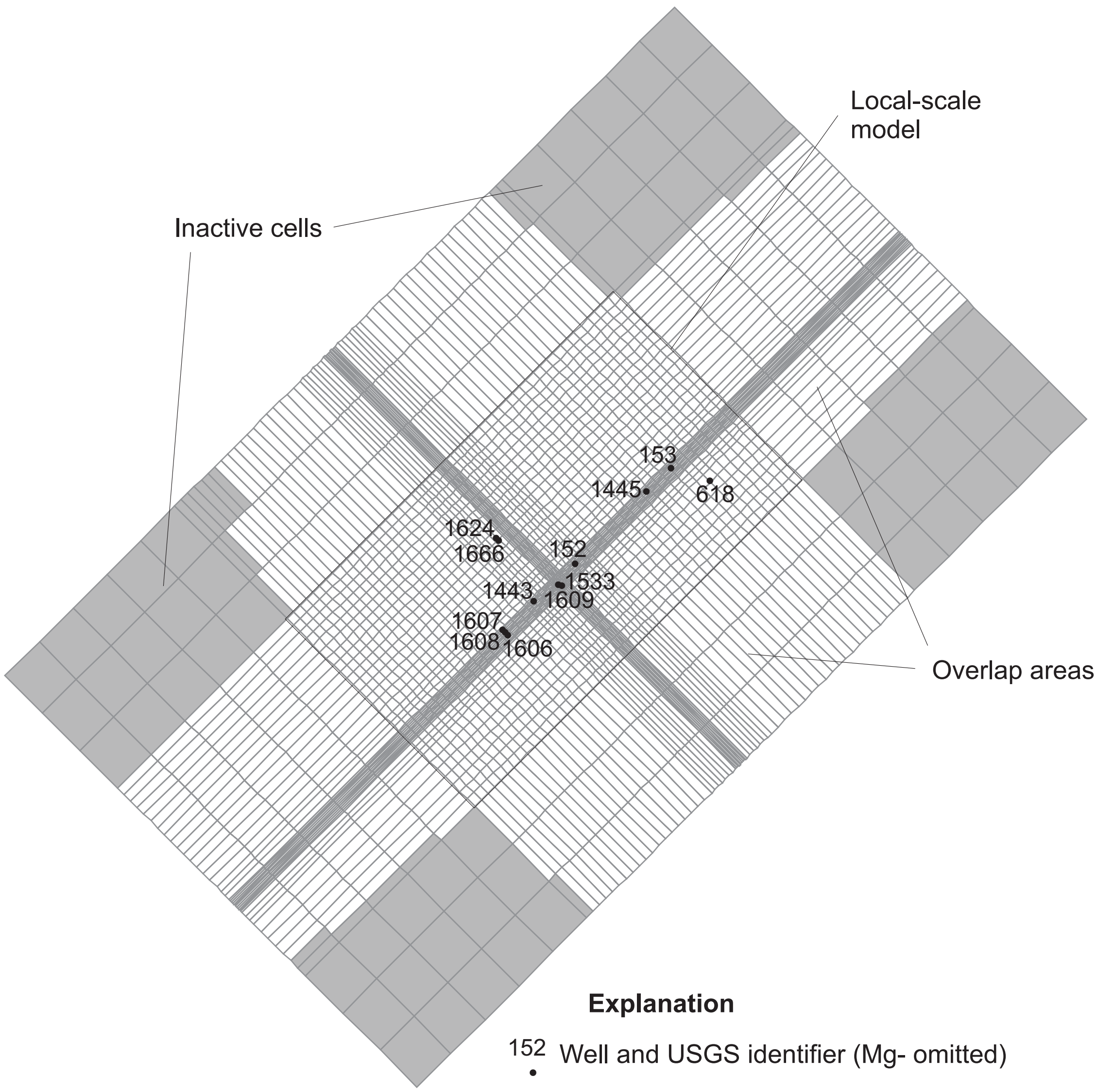

Figure 12.-- Well location and horizontal grid configuration for simulation of local-scale ground-water flow at the John Evans and Sons property in north-central Lansdale, Pa. showing local-scale grid and overlap areas for incorporation of regional-model flux boundary conditions.

Aquifer-Test Simulation

The local-scale model is calibrated by simulation of drawdown during the aquifer test of November 1997. The model is calibrated to drawdown measured in the same four wells used for the analytical model analysis summarized in the Aquifer-Test Results section. Three of the wells used in the analysis are in high-permeability model cells, representing the pumped bed, and the fourth well is in a cell that is within the low-permeability part of the aquifer, not in the pumped bed.

MODFLOWP automatically determines the optimum values of model parameters (hydraulic conductivity and storage coefficient) that yield the minimum sum of squared errors (table 3). Model errors are the difference between simulated and measured drawdown. This procedure is similar to that of matching analytical models to the measured drawdown (e.g. Senior and Goode, 1999, p. 55), except that a numerical model of flow is used here. The results of simulations obtained here with a three-dimensional numerical model are similar to the results obtained by Senior and Goode (1999, p. 67) using a two-aquifer analytical model. The transmissivity of the pumped bed estimated here, 1,900 ft^2/d, is about 45 percent higher than the transmissivity of the pumped aquifer reported by Senior and Goode (1999), 1,300 ft^2/d. The hydraulic conductivity of the rest of the bedrock, 0.05 ft / d, is only slightly larger than the hydraulic conductivity of the low-permeability unit separating the aquifers in the analytical model (Senior and Goode, 1999), 0.044 ft / d. The storage parameters agree within a factor of 2. The analytical model included a second unpumped aquifer that does not have a corresponding part in the numerical model used here.

The calibrated model can approximately simulate measured drawdowns measured during the aquifer test (fig. 13). Compared to the analytical match using the two-aquifer model by Senior and Goode (1999, fig. 10), this model does not match early drawdown as well but more closely matches the rate of drawdown increase at late time (fig. 13). The best fit of the model is obtained with high K in the pumped bed and low K in the lower permeability rock (table 3). The weathered rock and soil also have low K.

Table 3.-- Optimum and approximate, individual, 95-percent confidence-interval values for hydraulic conductivity and specific storage for calibrated simulation of ground-water flow at the John Evans and Sons property in north-central Lansdale, Pa.

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

Parameter	Units	Optimum value	Approximate, individual, 95-percent confidence interval	
			Lower value	Upper value
Pumped bed transmissivity	ft^2/d	1,900	500	2,300
Bulk rock hydraulic conductivity	ft/d	0.05	.033	.077
Pumped bed storage coefficient	-	4.0×10^{-5}	2.3×10^{-5}	7.0×10^{-5}
Bulk rock specific storage	per foot	1.5×10^{-6}	9.9×10^{-7}	2.3×10^{-6}

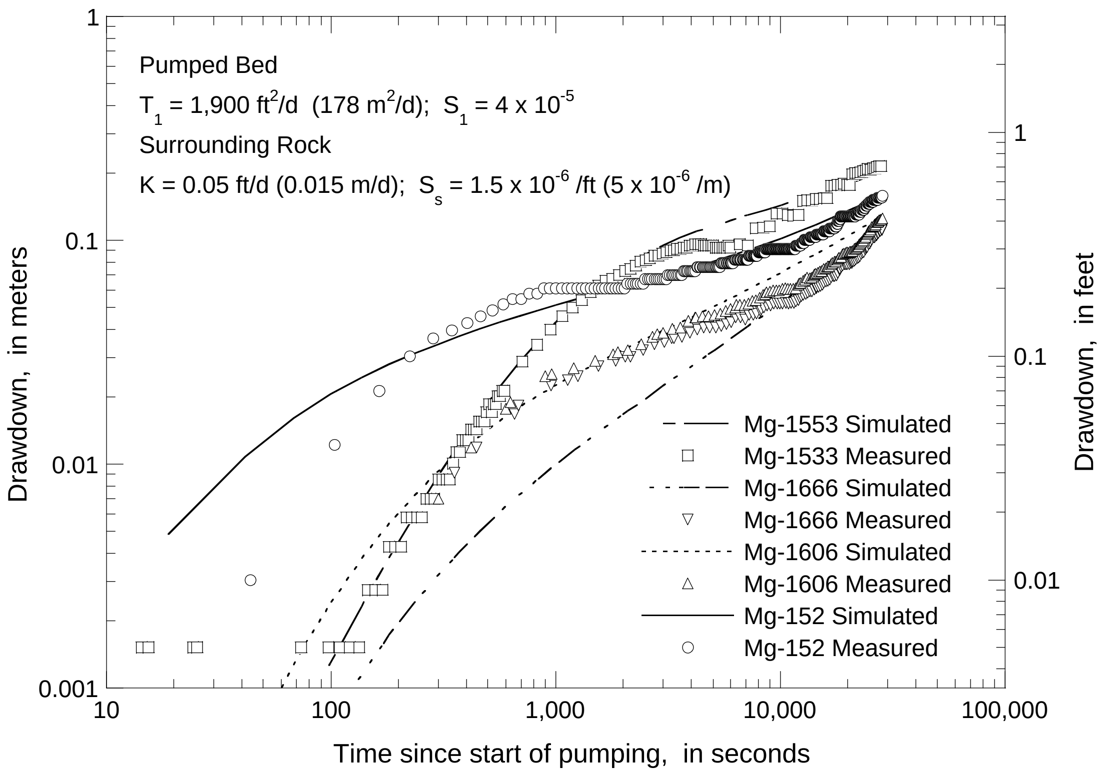

Figure 13.-- Measured and simulated drawdown, using local-scale flow model, in wells Mg-67, Mg-80, Mg-163, and Mg-1666 at the John Evans and Sons property in north-central Lansdale, Pa., November 21, 1997. Well Mg-1609 was pumped at a rate of 9.1 gallons per minute for 7.93 hours (measured drawdown from Senior and Goode, 1999).

The spatial pattern of drawdown within a horizontal model layer is characteristic of anisotropy (fig. 14), although the model is configured to represent a heterogeneous isotropic aquifer. Within a model layer, only some of the cells are located in the high-hydraulic-conductivity pumped bed. Head gradients within these cells are small because the hydraulic conductivity is high. The configuration of cells with high hydraulic conductivity along rows leads to the elongated drawdown contours, similar to results in anisotropic homogeneous aquifers.

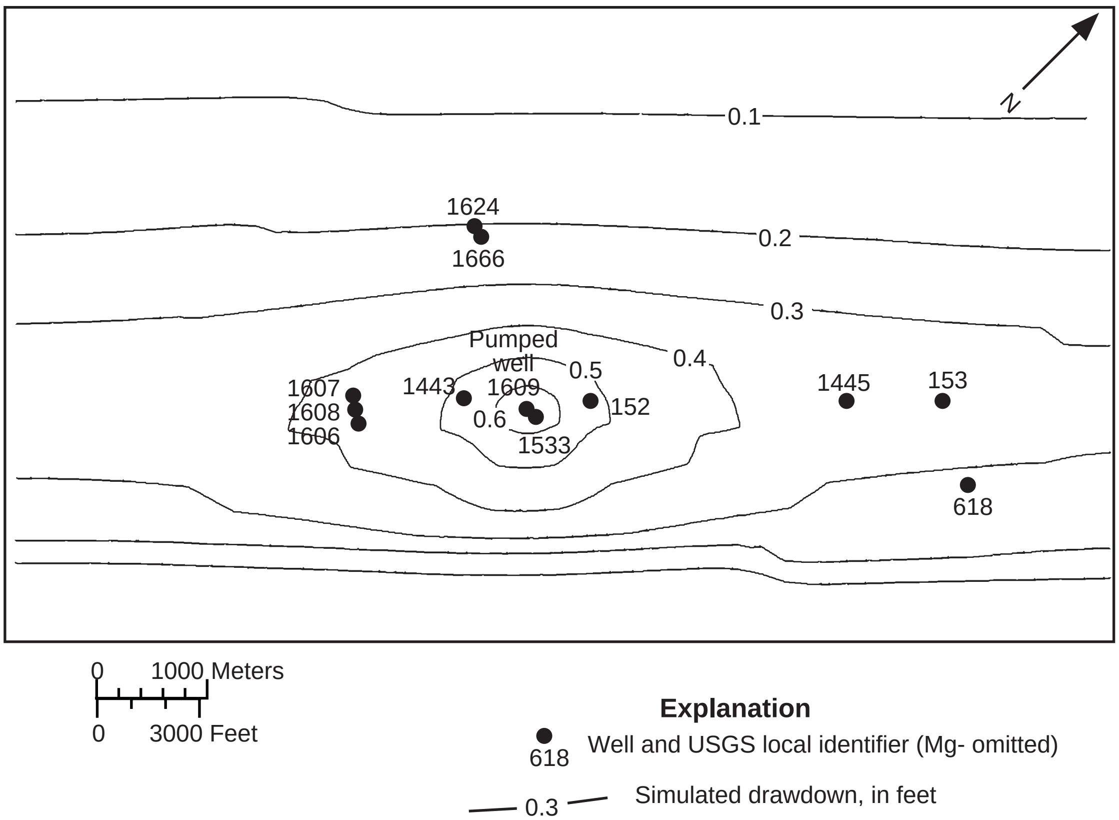

Figure 14.-- Well locations and simulated drawdown in model layer 6 (containing the pumped well) after 7.93 hours of pumping well Mg-1609 at a rate of 9.1 gallons per minute at the John Evans and Sons property in north-central Lansdale, Pa.

Effect of Pumping on Ground-Water Flowpaths

A steady-state-flow field is simulated in the local-scale model using boundary fluxes from the regional-scale model with the specified pumping in well Mg-1609. Regionally, flow into the local-scale model occurs along the SE boundary; most flow out crosses the SW and NE boundaries. When pumping at 10 gal / min, about 12 percent of the inflow to the local model discharges to the pumping well.

The source areas for water pumped from well Mg-1609 are illustrated by results of a particle tracking simulation. Particle paths are backtracked from the pumping well to the local-scale model boundaries using MODPATH (Pollock, 1994). About 56 percent of the particles originate at the water table in the local-model domain (fig. 15). The remaining particles cross the inflow boundary along the SE edge of the model and originate outside the local-model domain. Deeper particles on this edge of the model probably originate at the water table far from the pumped well. Any recharge more than about 200 ft from the pumping well in the downgradient (NW) direction is not captured by the pumping well.

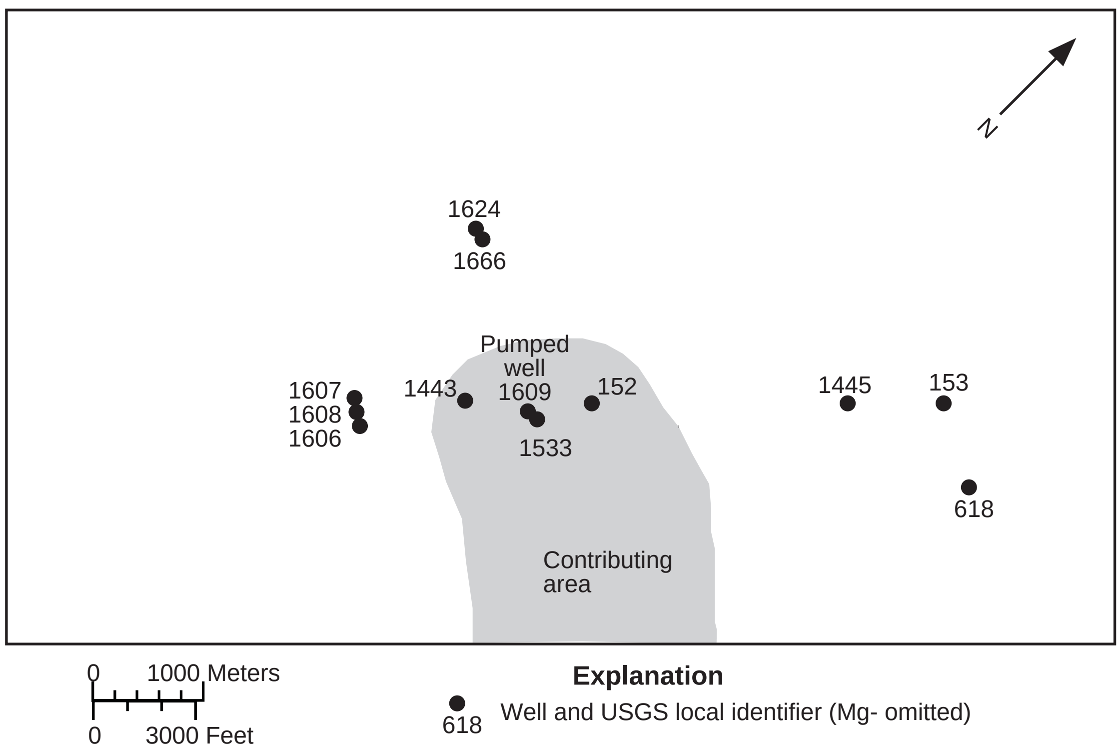

Figure 15.-- Simulated contributing area for well Mg-1609 pumping at a rate of 10 gallons per minute at the John Evans and Sons property in north-central Lansdale, Pa. See figure 4 for location of the local-scale model.