Hydrogeology and Simulation of Source Areas of Water to Production Wells in a Colluvium-Mantled Carbonate-Bedrock Aquifer near Shippensburg, Cumberland and Franklin Counties, Pennsylvania

By Bruce D. Lindsey

In cooperation with the
Shippensburg Borough Authority

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Conversion Factors and Datum

Multiply	By	To obtain
	Length	
inch (in.)	25.4	millimeter (mm)
foot (ft)	0.3048	meter (m)
	Area	
acre	4,047	square meter (m^2)
square foot (ft^2)	0.09290	square meter (m^2)
square mile (mi^2)	2.590	square kilometer (km^2)
	Volume	
million gallons (Mgal)	3,785	cubic meter (m^3)
	Density	
pound per cubic foot (lb/ft^3)	16.02	kilogram per cubic meter
	Flow rate	
cubic foot per second (ft^3/s)	0.02832	cubic meter per second (m^3/s)
million gallons per day per square mile [$(Mgal/d)/mi^2$]	1,461	cubic meter per day per square kilometer [$(m^3/d)/km^2$]
	Mass	
pound, avoirdupois (lb)	0.4536	kilogram (kg)

Temperature in degrees Celsius (˚C) may be converted to degrees Fahrenheit (˚F) as follows:

˚F = (1.8 x ˚C) + 32

Temperature in degrees Fahrenheit (˚F) may be converted to degrees Celsius (˚C) as follows:

˚C = (˚F - 32) / 1.8

Vertical coordinate information is referenced to the North American Vertical Datum of 1988 (NAVD 88).

Horizontal coordinate information is referenced to the North American Datum of 1983 (NAD 83).

Altitude, as used in this report, refers to distance above the vertical datum.

Specific conductance is given in microsiemens per centimeter at 25 degrees Celsius (µS/cm at 25˚C).

Concentrations of chemical constituents in water are given either in milligrams per liter (mg/L) or micrograms per liter (µg/L).

Mesh sizes of sampling devices are given in micrometers (µm); 1,000 micrometers equal a millimeter (mm).

Hydrogeology and Simulation of Source Areas of Water to Production Wells in a Colluvium-Mantled Carbonate-Bedrock Aquifer near Shippensburg, Cumberland and Franklin Counties, Pennsylvania

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Abstract

This report presents the results of a study by the U.S. Geological Survey in cooperation with the Shippensburg Borough Authority to evaluate the source areas of water to production wells in a colluvium-mantled carbonate-bedrock aquifer in Cumberland and Franklin Counties, Pa. The areal extent of the zone of contribution was simulated for three production wells near Shippensburg, Pa. by use of a ground-water-flow model. A 111-square-mile area was selected as the model area and includes areas of the South Mountain Section and the Great Valley Section of the Valley and Ridge Physiographic Province. Within the model area, the geologic units in the South Mountain area are predominantly metamorphic rocks and the geologic units in the Great Valley are predominantly carbonate rocks. Hydrologic and geologic information were compiled to establish a conceptual model of ground-water flow. Characteristics of aquifer materials were determined, and streamflow and water levels were measured. Streamflow measurements in November 2003 showed all streams lost water as they flowed from South Mountain over the colluvium-mantled carbonate aquifer into the Great Valley. Some streams lost more than 1 cubic foot per second to the aquifer in this area. The Shippensburg Borough Authority owns three production wells in the model area. Two wells, Cu 969 and Fr 823, are currently (2004) used as production wells and produce 500,000 and 800,000 gallons per day, respectively. Well Cu 970 is intended to be brought on line as a production well in the future. Water levels were measured in 43 wells to use for model calibration. Water-level fluctuations and geophysical logs indicated confined conditions in well Cu 970.

Ground-water flow was simulated with a model that consisted of two vertical layers, with five zones in each layer. The units were hydrostratigraphic units that initially were based on geologic formations, but boundaries were adjusted during model calibration. Model calibration resulted in a root mean square error of 9.8 feet. A parameter-estimation package was used during model calibration to estimate three parameters. The parameter estimation resulted in a value of 233 feet per day for horizontal hydraulic conductivity of the highly fractured carbonate rocks and sandy colluvium in layer 1; 3.97 feet per day for horizontal hydraulic conductivity of the ridge-forming unit in layer 1; and a value of 1.73 for horizontal anisotropy in both layers.

The calibrated model was used to delineate the areal extent of the zone of contribution for wells Cu 969 and Fr 823. Although well Cu 970 is not currently (2004) being used, the areal extent of its zone of contribution also was simulated without additional model calibration. The shape of the areal extent of the zone of contribution was similar for each well and included an area that extended from the well southwest along the Tomstown Formation, and then extended southeast into the metamorphic rocks of South Mountain. The contributing areas from the watersheds of losing streams were also delineated because losing stream reaches bisect the areal extent of the zones of contribution.

Spatial uncertainty of the areal extent of the zone of contribution was illustrated using a Monte-Carlo analysis. The model was run 1,000 times using randomly generated parameter sets that were normally distributed within the confidence interval around the optimal values for the three estimated parameters. The model converged and had a reasonable water budget for 980 of the model runs. For each of those 980 model runs, the recharge area was determined, and the results for all runs were compiled and contoured. The results of the Monte-Carlo analysis were compared to the results of the deterministic model, illustrating that the deterministic model has the greatest certainty in the area closest to each well in the Tomstown Formation. The areas farther from the well, upgradient, and in the metamorphic rocks have a higher degree of uncertainty than those areas closer to the well.

Introduction

Many communities in Pennsylvania rely on ground water as a source for municipal supply. In areas underlain by carbonate bedrock, wells commonly yield sufficient quantities of potable water to supply small communities; however, carbonate-bedrock aquifers typically are more susceptible to contamination than other aquifer types (Lindsey and others, 1997). Understanding the source of the water to wells is essential in protecting this important resource.

The Borough of Shippensburg, a community in Cumberland and Franklin Counties, Pa. (fig. 1), relies largely on ground water from a colluvium-mantled carbonate-bedrock aquifer for municipal supply. Shippensburg Borough Authority supplies water to the Borough of Shippensburg, and to several municipalities in the surrounding area. The borough sold their surface-water supply reservoir in 2004 and is in transition to a system that relies mostly on ground water. Because of reliance on ground water and the inherent susceptibility of the carbonate-bedrock aquifer to contamination, the borough has initiated a wellhead-protection plan. As a part of this effort, the U.S. Geological Survey (USGS), in cooperation with the Shippensburg Borough Authority, conducted a study of the sources of water to the Shippensburg Borough Authority production wells. The results of this study can be applied to similar carbonate-bedrock aquifers in Pennsylvania and other similar areas.

Water is supplied by the Shippensburg Borough Authority (hereafter referred to as the “Authority”) to a population of approximately 15,000 people with about 4,200 connections (William Wolfe, Borough of Shippensburg, written commun., 2004), including connections in Southampton Township, Franklin Co.; Southampton Township, Cumberland Co.; Shippensburg Township; Lurgan Township; Letterkenny Township; and the Borough of Orrstown (fig. 2). The Authority supplies water to the Huckleberry Land and Water Association, and also is a backup supply for Southern Cumberland Water. The current (2004) water usage is approximately 2 Mgal/d; the majority of that water is supplied by wells.

Two wells are used by the Authority for production, and a third well is intended to be brought on line by 2006 to meet future needs. Well 1 (USGS well number Cu 969) is in Cleversburg (fig. 3). It was drilled in 1988 and has a total depth of 590 ft, with 343 ft of casing. This well is pumped intermittently and supplies 350,000-500,000 gal/d. Well 2 (USGS well number Fr 823) is in Mainsville (fig. 3). This well was drilled in 1994 and has a total depth of 400 ft, with 304 ft of casing. Well Fr 823 is the most productive well in use by the Authority and is pumped continuously to supply the system with 800,000 to 1,000,000 gal/d. Well 3 (USGS well number Cu 970) was drilled in 2001 to a depth of 360 ft, with 240 ft of casing. The

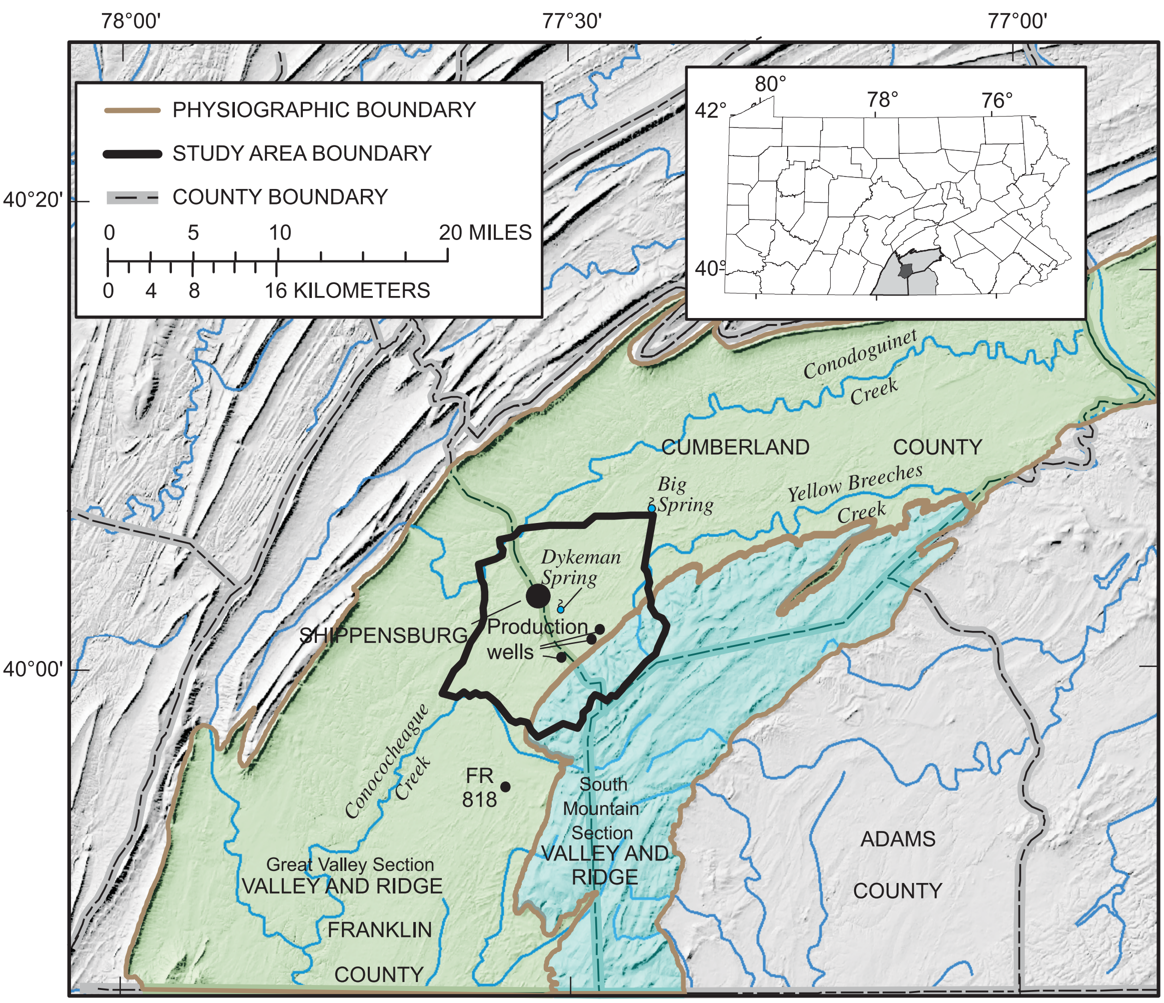


Figure 1. Physiography (Sevon, 2000), major streams, springs, production wells, county observation well, and study-area boundary near Shippensburg, Pa.

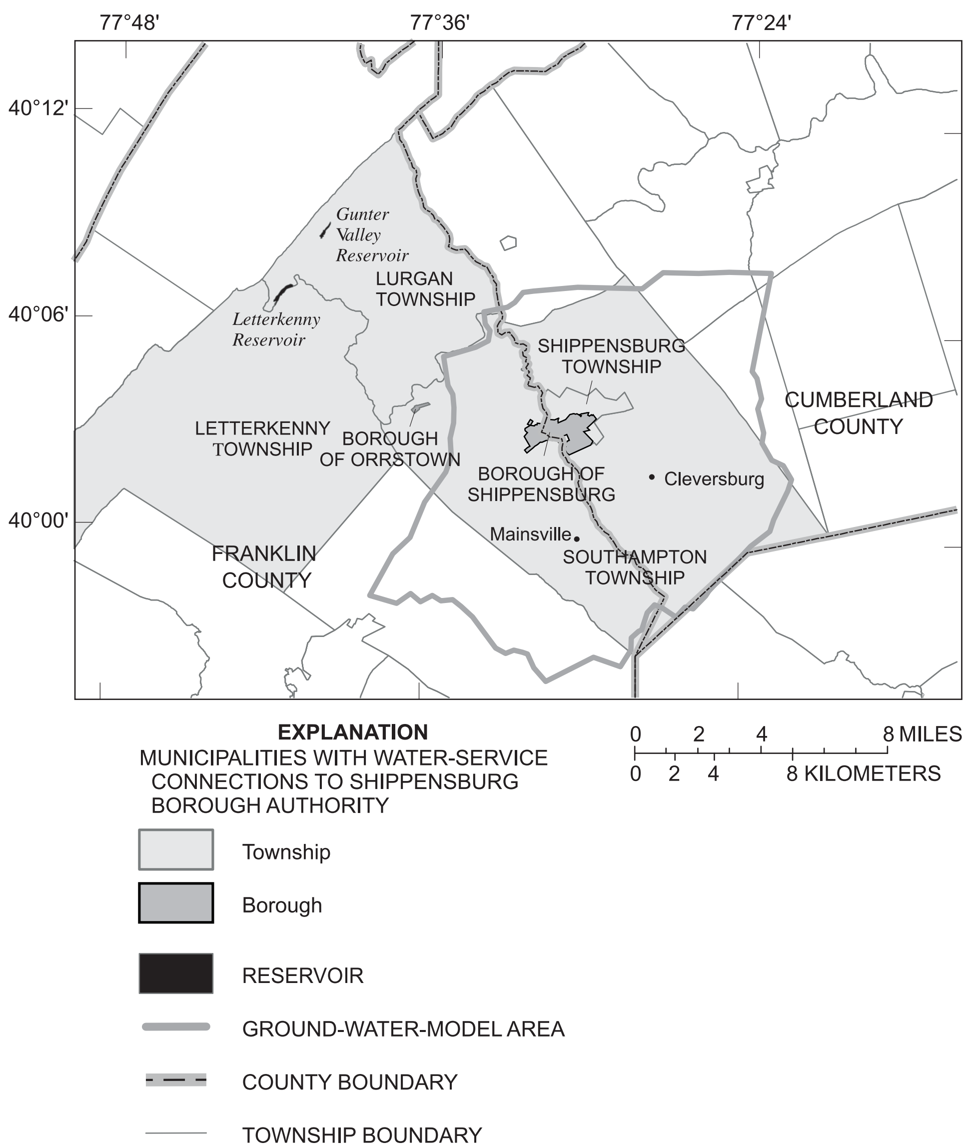


Figure 2. Geographic features and municipalities with water-service connections to Shippensburg Borough Authority water system near Shippensburg, Pa.

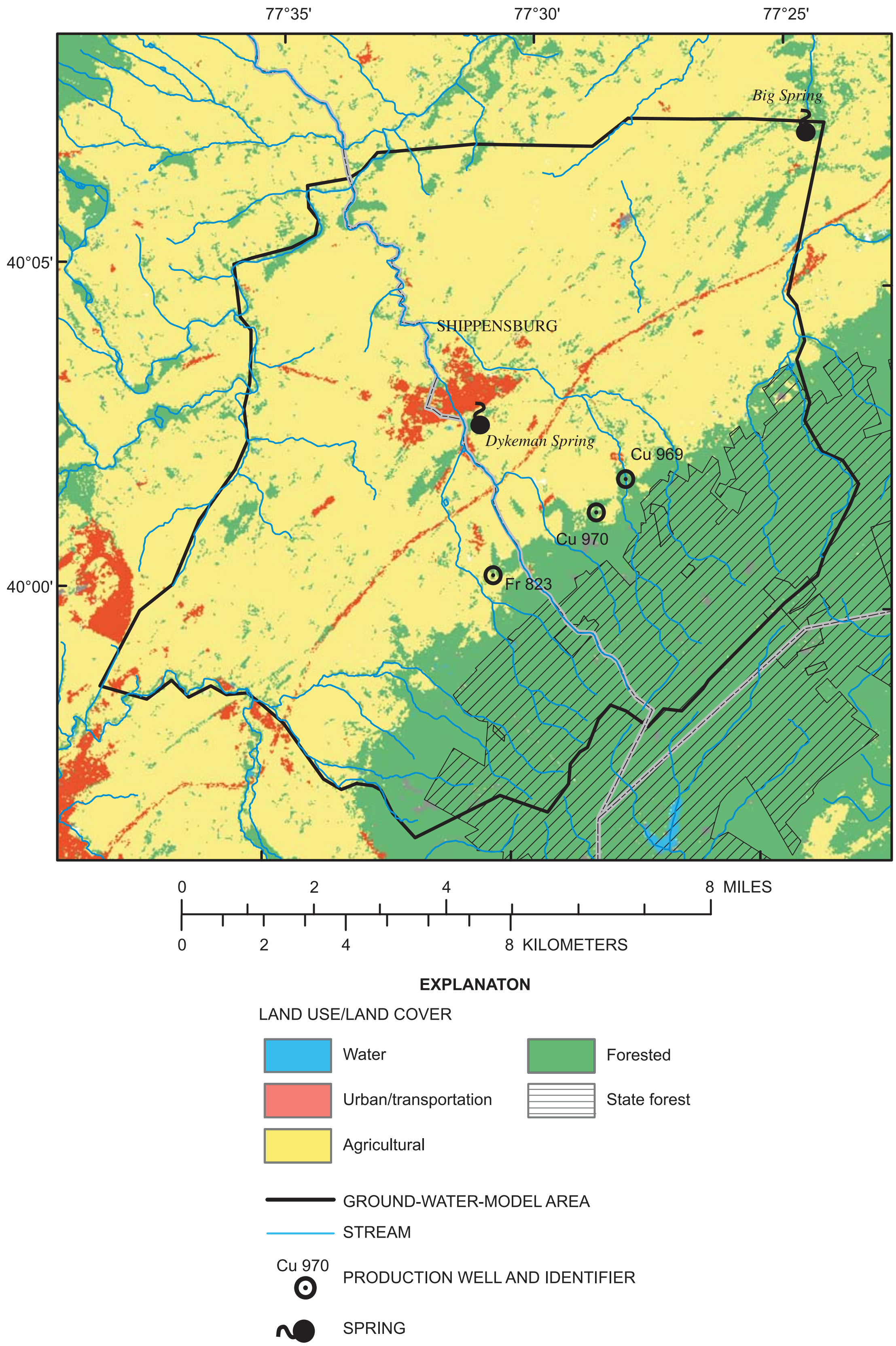


Figure 3. Land use and land cover near Shippensburg, Pa. (Vogelmann and others, 1998a, 1998b)

wells are all completed in the Tomstown Formation and are on the northern flank of South Mountain. The treatment for the ground-water system is chlorination. Storage for wells Cu 969 and Fr 823 is provided by a 500,000-gal tank in Cleversburg and a 250,000-gal tank in Mainsville. A central reservoir provides storage of 1.4 Mgal for both wells and water from the water-treatment plant. About 600,000 gal/d is provided from the Gunter Valley Reservoir (fig. 2), but the reservoir will be abandoned when an interconnect with the Letterkenny Reservoir is established and brought on line.

Dykeman Spring (fig. 3) was formerly used by the Authority as a water-supply source. In 1996, a study was conducted as part of an effort to determine the suitability of springs used for public supply (U.S. Environmental Protection Agency, 1992). This study determined that Dykeman Spring was under the direct influence of surface water, primarily because of the presence of algae in water discharging from the spring. In 2001, the spring was abandoned as a water-supply source and is no longer being considered by the borough as a potential source for future water supply. (Thomas Feeney, Shippensburg Borough Authority, oral commun., 2004).

Municipal water supplies are required to test routinely for selected organic, inorganic, and radiochemical constituents (U.S. Environmental Protection Agency, 2002). Concentrations of these constituents have not exceeded the U.S. Environmental Protection Agency (USEPA) Maximum Contaminant Levels (MCLs) (U.S. Environmental Protection Agency, 2002) in water from the Authority production wells (William Wolfe, Borough of Shippensburg, written commun., 2004). The USGS, however, conducted a study of ground-water quality that included the carbonate-bedrock aquifer in the Shippensburg area (Lindsey and others, 1997) and reported a median nitrate concentration of 9 mg/L as nitrogen, and numerous wells that exceeded the USEPA MCL of 10 mg/L in water from the carbonate-bedrock aquifer. Although elevated concentrations of nitrate and other contaminants have not been detected in Wells Cu 969 and Fr 823, it has been demonstrated that protecting a ground-water supply can be up to 200 times less expensive than mitigating a ground-water contamination problem (U.S. Environmental Protection Agency, 1996). Therefore, the Authority is determining the probable source of water to the production wells and implementing a wellhead-protection plan to prevent or minimize the possibility of contamination.

The land use and land cover in the area around the Shippensburg production wells include forested, agricultural, and low-density residential (fig. 3). Fertilizers, animal manure, and individual private septic systems are potential sources of nitrate that could cause concentrations in ground water to approach or exceed the USEPA MCL. Several roads are in this area, and spills of contaminants along these roadways potentially could affect the quality of water in the production wells.

Purpose and Scope

This report defines the contributing area and sources of water to three production wells used by the Authority, on the basis of field measurements and simulation of ground-water flow. The field measurements and simulation of ground-water flow were in a 111-mi^2 study area in Cumberland and Franklin Counties, Pa., near the Borough of Shippensburg (fig. 1). Field measurements used to determine sources of water to wells include water levels measured in 43 wells from November 2003 to April 2004, streamflow measured in 20 streams during November 2003, and discharge measured from 1 spring during November 2003. Ground-water flow was simulated for steady-state conditions.

Description of Study Area

Shippensburg is in the Great Valley Section of the Valley and Ridge Physiographic Province (Sevon, 2000)(fig. 1). The Great Valley Section is characterized as a broad valley flanked on the south and east by the South Mountain Section of the Valley and Ridge Physiographic Province and by Blue Mountain on the north and west. The valley has a large amount of agricultural land and is an important transportation corridor with urban centers along the major transportation routes. The mountains flanking the valley are sparsely populated and predominantly forested. An area of approximately 111 mi^2 around the Authority production wells was the focus of the study and is referred to herein as the ground-water-model area. This area is bounded on the southeast by the top of the ridge of South Mountain, extends northeast to include the uppermost tributaries of the Yellow Breeches Creek, extends across the valley to the northwest to Big Spring, follows the general trace of the mainstem of Conodoguinet Creek to the southwest, then extends to the uppermost tributaries of Conococheague Creek that flow from South Mountain. The model boundaries are a combination of stream-watershed boundaries or assumed no-flow boundaries parallel to streams (fig. 3). These boundaries are larger than typically would be needed to simulate ground-water flow near the wells. Because of the large volume of water in this area that is transferred among surface-water basins through the ground-water system (underflow), these boundaries allow better simulation of ground-water flow by allowing water to travel to its natural discharge locations.

Land use and land cover in the study area are predominantly agricultural (fig. 3). Agricultural land is about 64 percent of the land use, forested land is about 33 percent, urban land is about 2 percent, and all other land uses are less than 1 percent. The distribution of land use is important in assessing the potential sources of contaminants near the well fields. All three production wells are near the boundary of the forested and agricultural area. The area upgradient from each of these wells is mostly undeveloped forest. A large percentage of the forested land on South Mountain is part of the Pennsylvania State Forest system (fig. 3), making the future land use and development more predictable than if this area was privately owned.

Contributing Area and Related Teminology

Some of the terms used by the Pennsylvania Department of Environmental Protection (PaDEP) and the USGS with regard to contributing areas to wells are described herein. The term "area contributing recharge to a discharging well" is the "surface area of the three-dimensional boundary of the ground-water system that delineates the location of the water entering the ground-water system that eventually flows to the well and discharges" (modified from Reilly and Pollock, 1993). This area is shown in gray on figure 4 and is determined by the sum of all the recharge locations. Only a subset of recharge locations is shown as examples; the gray shaded area would be entirely covered by, and only include, the recharge points if all were shown. Another term is the "areal extent of the zone of contribution to a discharging well," which is the three-dimensional volume of water flowing to a discharging well, projected to the land surface. This is shown with the bold outline on figure 4. This area encompasses the area contributing recharge and also includes projection of ground-water-flow paths to the land surface, which may be important if there are concerns about contaminants entering the zone of contribution. Another term, which was used by Risser and Barton (1995) is the "watershed of losing streams" (fig. 5). If a well induces recharge from a stream and draws that water to the well, the watershed upstream of the location of the losing reach should be considered in wellhead-protection planning. For this report, both the areal extent

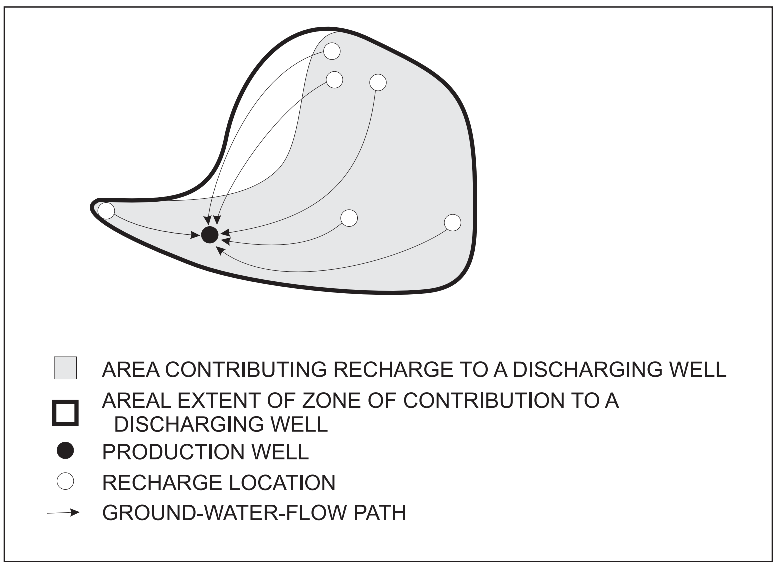


Figure 4. Illustration of terms related to areas contributing recharge to a well.

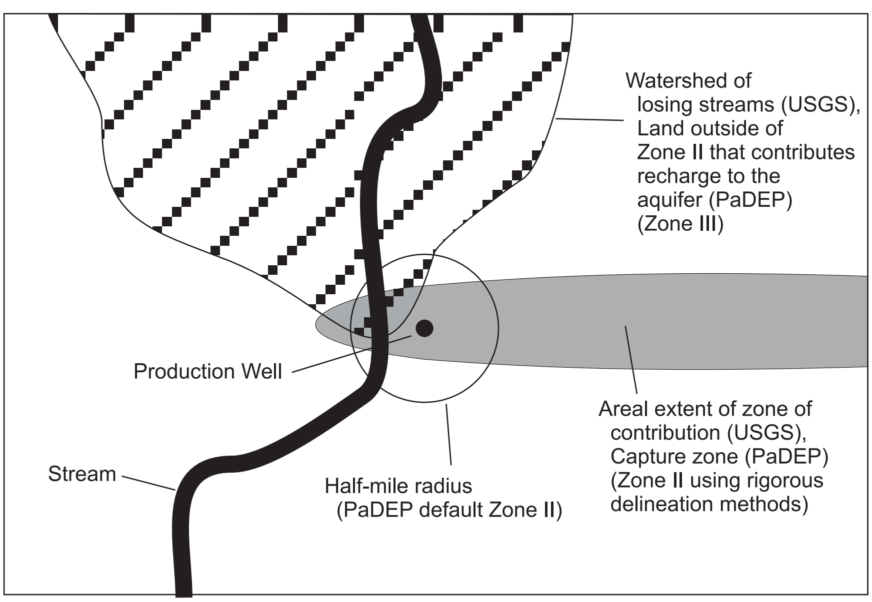


Figure 5. Illustration of terms used by the U.S. Geological Survey and the Pennsylvania Department of Environmental Protection related to wellhead protection. (Modified from Risser and Barton, 1995) [USGS, U.S. Geological Survey; PaDEP, Pennsylvania Department of Environmental Protection].

of the zone of contribution and the watershed of losing streams are considered to be source areas.

Some terms used by the PaDEP for wellhead-protection purposes include “zones I, II, and III,” “capture zone”, and “contributing area of a well”. The following excerpt from the PaDEP Source Water and Assessment Program (Pennsylvania Department of Environmental Protection, 2000) describes these terms: “The first (Zone I) is a 100 to 400 feet radius based on site specific source and aquifer characteristics. The second (Zone II) is the capture zone of the source which is a half-mile radius unless a more rigorous delineation is conducted. The third (Zone III) area is the land area beyond Zone II that contributes recharge to the aquifer within the first two areas via surface water or groundwater. Collectively, Zones II and III constitute the “contributing area of a well.” The relations between the USGS terms and the PaDEP terms are illustrated in figure 5. For the purposes of this report, the areal extent of the zone of contribution will be equivalent to PaDEP Zone II. The watershed of losing streams upstream of the losing reach will be considered to be equivalent to PaDEP Zone III.

Hydrogeology

Information from previous studies of geology, hydrology, and aquifer characteristics are combined with water level, streamflow, and geophysical data from the current study to establish a conceptual model of the hydrogeology that will be used in the simulation of ground-water flow.

Geology

The geology of the Great Valley is characterized by carbonate bedrock in the southeastern part of the valley and siliciclastic bedrock (shale) in the northwestern part of the valley. The Great Valley is bounded on the south by South Mountain and on the north by Blue Mountain (fig. 6). South Mountain is predominantly underlain by metamorphic bedrock (quartzite) that is resistant to weathering and erosion. The bedrock of Blue Mountain is siliciclastic (sandstone), which also is resistant to weathering and erosion (fig. 6). The stratigraphic nomenclature used in this report is that of the Pennsylvania Geological Survey and is based on work by Berg and others (1980) and Root (1968).

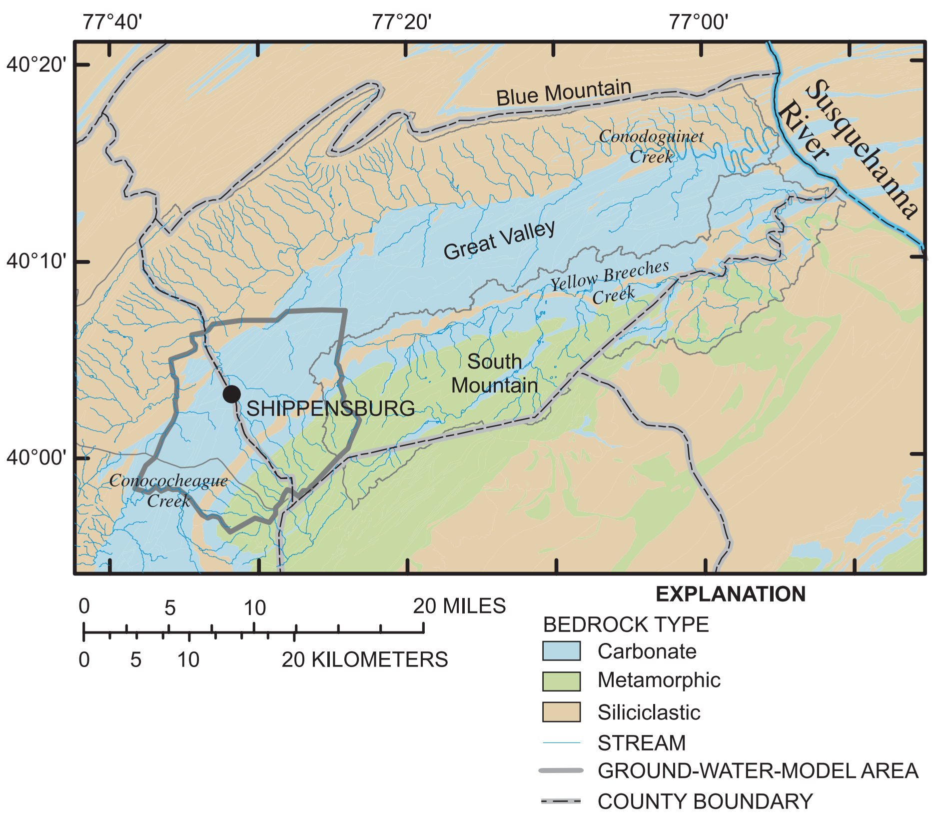


Figure 6. Generalized geology in the Conodoguinet and Yellow Breeches Creek watersheds, Great Valley and South Mountain, Pa. Modified from Berg and others (1980).

The ground-water-model area includes parts of South Mountain underlain by metamorphic rocks such as quartzite and schist and parts of the Great Valley underlain by carbonate rocks such as limestone and dolomite (fig. 7). The overall geologic structure of South Mountain is that of an anticline, and Blue Mountain is a synclinal ridge. The general structure in the valley is a sequence of formations striking about N. 50° E.; the older formations are to the southeast, and the younger formations to the northwest. Dip angles in the ground-water-model area are typically in the range of 50 to 70°; however, bedrock outcrops are rare in the vicinity of the production wells. In some areas, bedding is near-horizontal. Minor folds are superimposed within the broader regional structure (fig. 7); therefore, dip angles are not uniform. For example, the surface location of well Fr 823 is in the Waynesboro Formation in map view (fig. 7); however, it is completed in the Tomstown Formation because of the relatively shallow dip angles locally. The Shippensburg Fault is a major structural feature within the ground-

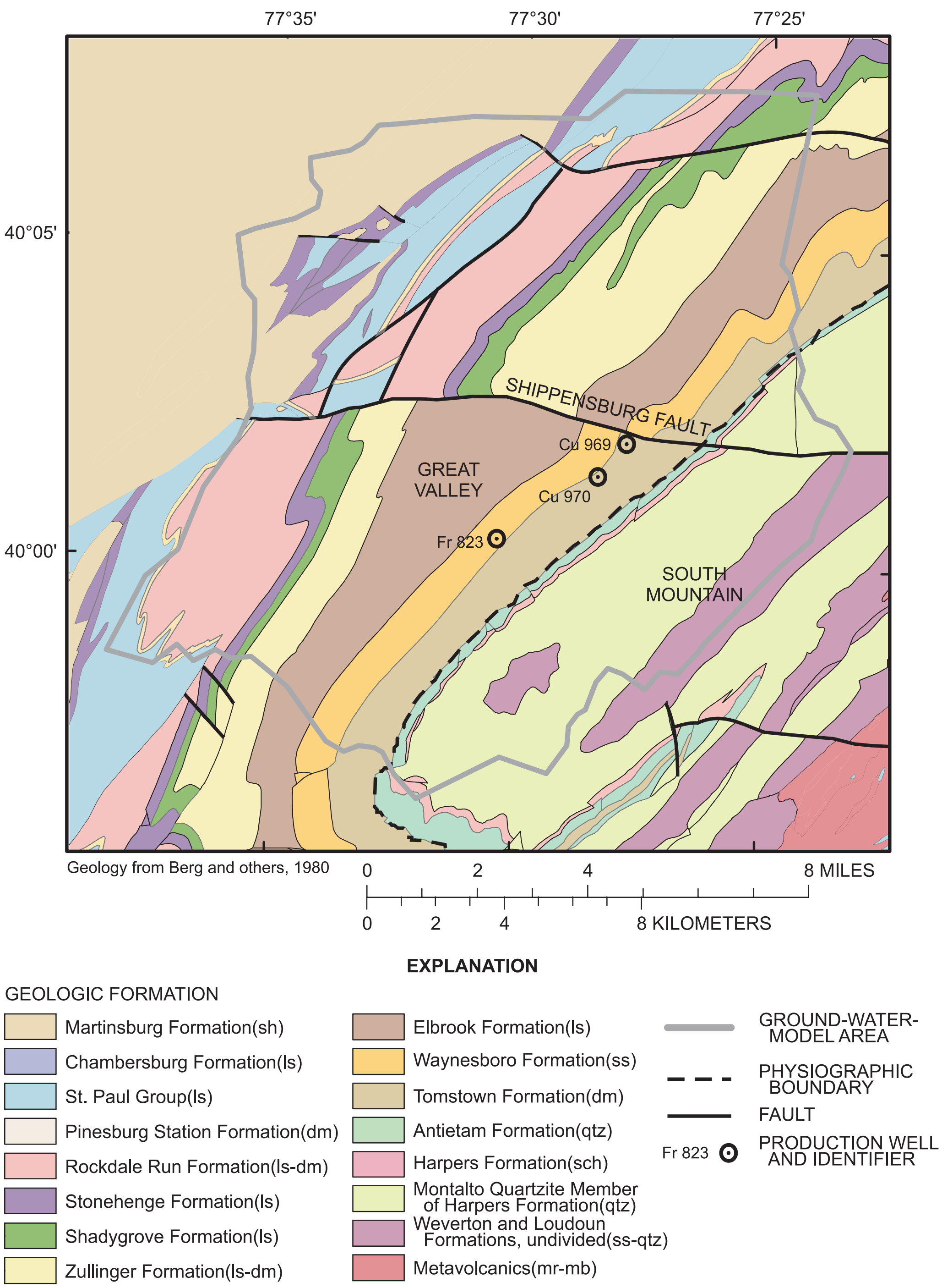


Figure 7. Bedrock geology in the Great Valley and South Mountain area near Shippensburg, Pa. (Fm., formation; Mbr., member; Gp, group; ls, limestone; dm, dolomite; sh, shale; ss, sandstone; qtz, quartzite, sch, schist; mr, metarhyolite; mb, metabasalt.)

water-model area. Descriptions of the geologic units in the study area are given in table 1.

An important geologic unit in this area is the colluvium. The carbonate formations near South Mountain are covered with a mantle of colluvium (Becher and Root, 1981). The colluvium has formed at the base of South Mountain from erosion and mass-wasting processes such as landslides from upslope formations that have been deposited over the carbonate formations. Therefore, the composition of the colluvium includes quartz sand, quartzite boulders, and clays. Because of the extensive weathering of the carbonate formations near the mountain front, the carbonate land surface has been lowered a great deal by solution of carbonate rocks, and colluvium from the upslope area fills the voids left in the carbonate rock. Therefore, the unconsolidated deposits (colluvium and residuum) extend much further below the land surface near the carbonate-metamorphic bedrock interface than the soils on the upper slopes and further out in the valley (fig. 8). Additionally, residual materials such as clays that are left behind when the carbonate rock dissolves are found beneath the colluvium (fig. 8). Although the characteristics of colluvium and residuum may be very different, the interface between these two materials is not well delineated. The thickness of the unconsolidated materials illustrated in figure 9 is the combined thickness of the colluvium and residuum.

The karst topography caused by the dissolution of carbonate bedrock in this area is an important factor affecting ground-water flow. Karst topography includes features such as sinkholes, caves, and closed depressions. These features have been mapped for this area by Kochanov (1989a, 1989b). The ground-water-model area contains many karst features, primarily closed depressions (fig. 10). Sinkholes as defined by Kochanov have an open orifice into the bedrock. Closed depressions in areas that have a thin mantle of soil also may have a nearly direct conduit into the bedrock aquifer (area A on fig. 8). In the area covered by colluvium, closed depressions may be the surface expression of a collapse feature hundreds of feet below the land surface (area B on fig. 8). Several caves also are located in the ground-water-model area (fig. 10), but most of the caves are above the water table. Although the Waynesboro Formation is predominantly sandstone, it includes some carbonate rocks, and some karst features are underlain by the Waynesboro Formation.

Hydrology

Information on the hydrology of the study area can be determined by analysis of historical records of streamflow and precipitation. Hydrologic characteristics of the aquifers can be determined from previous studies. Water levels and streamflow measurements from the current study can be incorporated to supplement the hydrologic data available from previous studies.

Water Budget

A good understanding of the water budget is essential for simulation of ground-water flow. The water budget has several components, some of which can be measured physically and others that are inferred or calculated. The water budget is represented by equation 1.

$$P = SRO + GWD + ET, \qquad (1)$$

where P is precipitation, SRO is surface runoff, GWD is ground-water discharge, and ET is evapotranspiration. Precipitation and streamflow are measured. Evapotranspiration, the sum of evaporation and transpiration, is not measured directly, and, therefore, includes potential errors in the other terms.

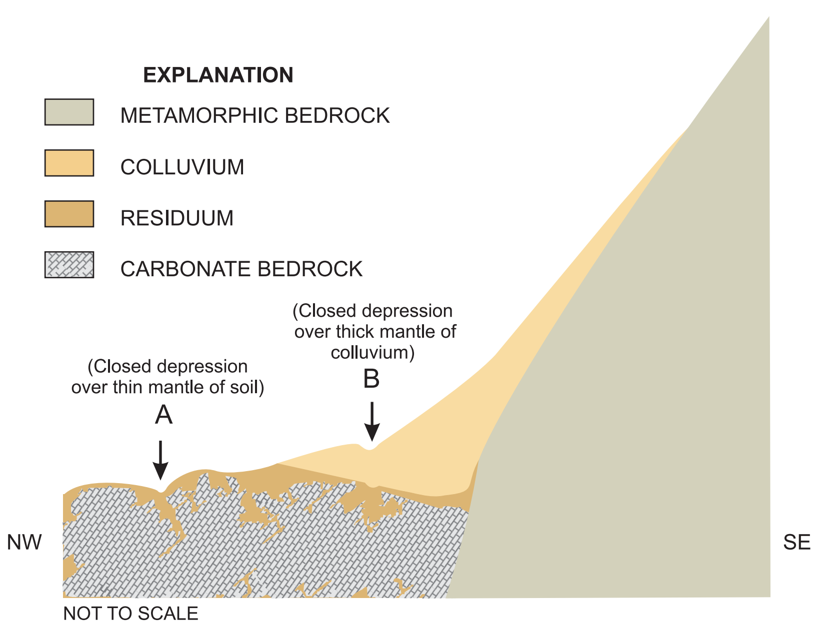


Figure 8. Generalized cross section illustrating relations among carbonate bedrock, metamorphic bedrock, and unconsolidated materials on the flank of South Mountain near Shippensburg, Pa. Modified from Nutter (1974). Locations A and B are described in text.

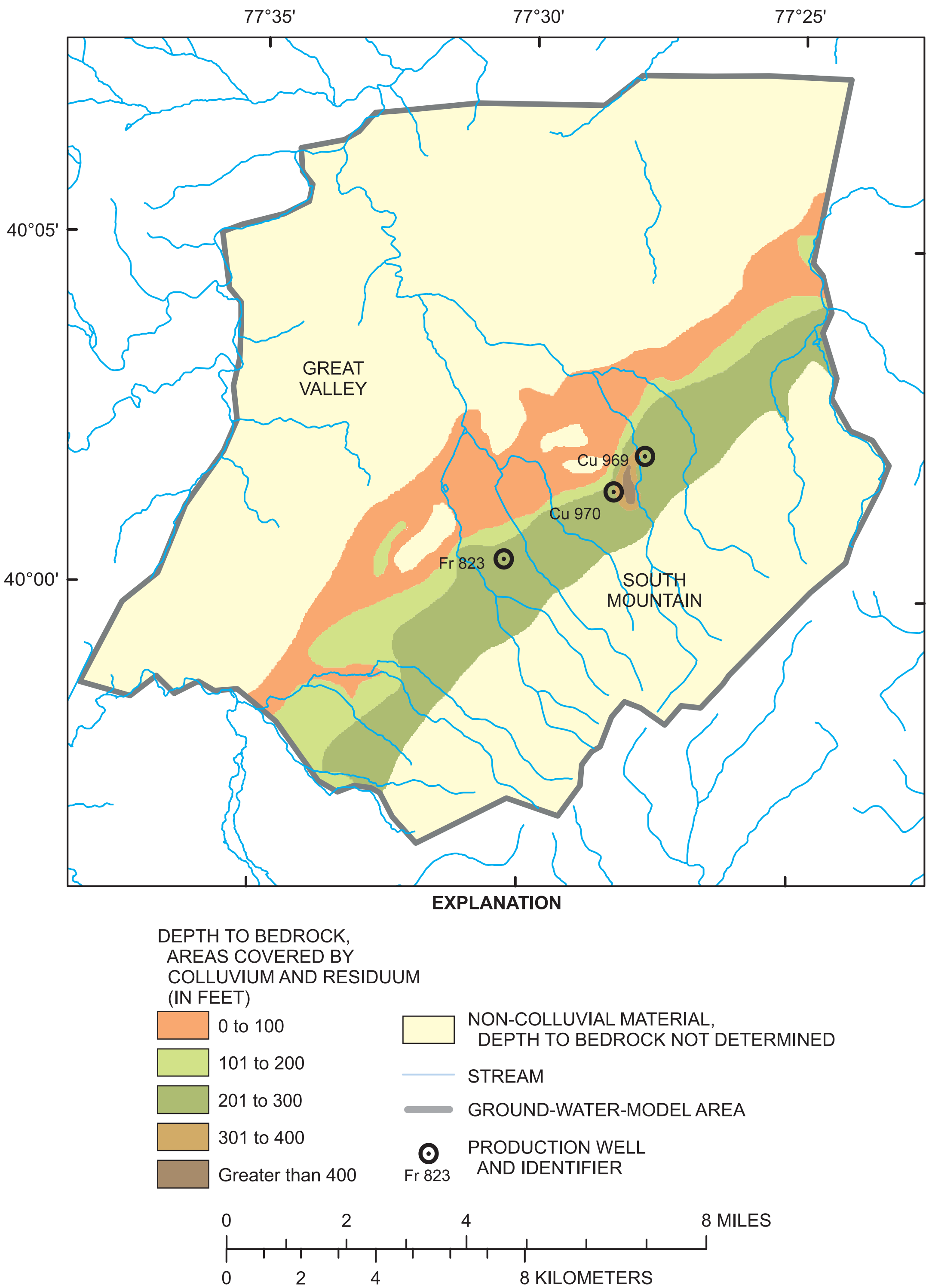


Figure 9. Depth to bedrock or thickness of unconsolidated materials on the flank of South Mountain near Shippensburg, Pa. (from Sevon, 2001).

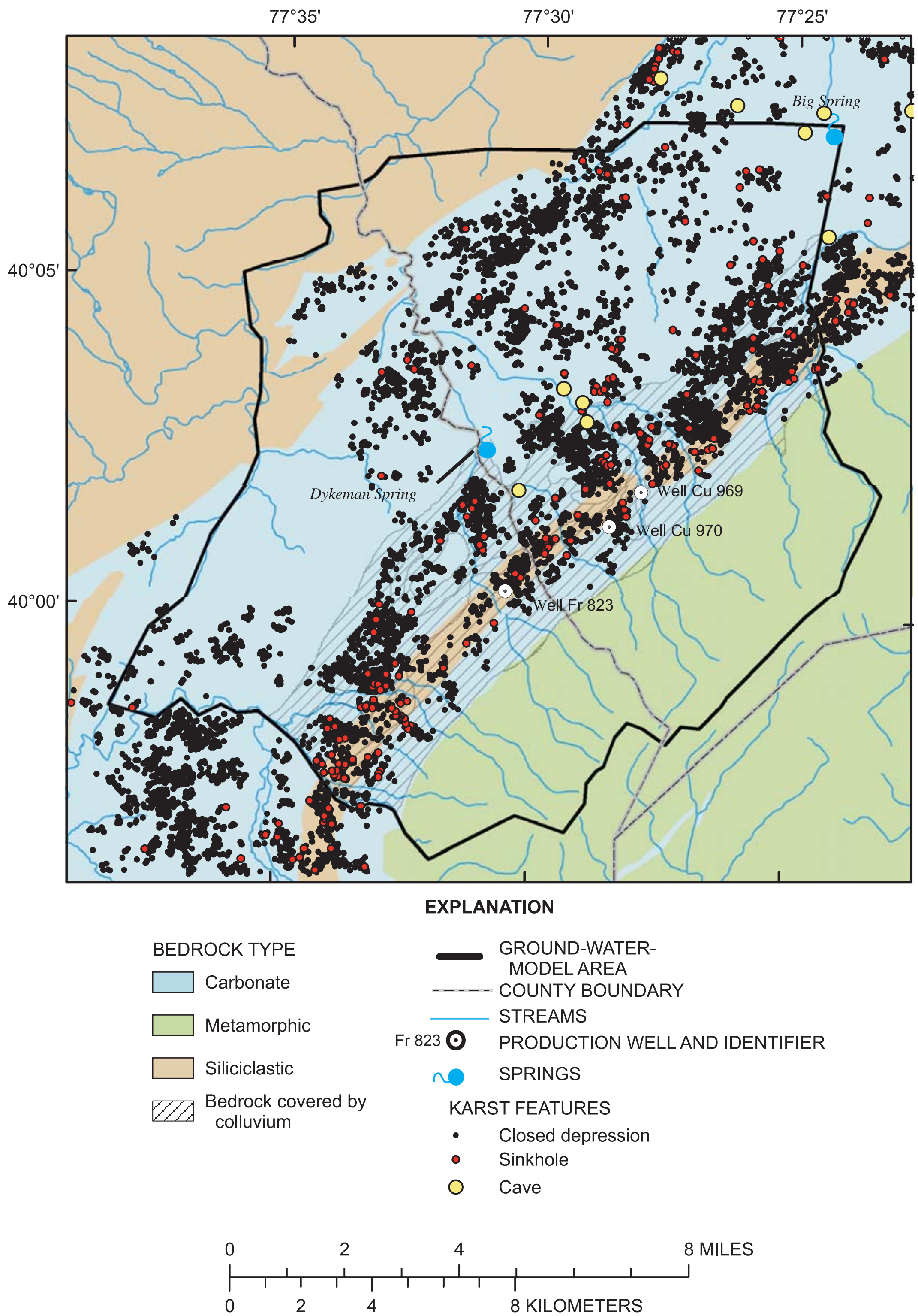


Figure 10. Bedrock type, colluvium, and locations of karst features near Shippensburg, Pa. Karst features from Kochanov (1989a, 1989b).

Table 1. Geologic-stratigraphic column near Shippensburg, Cumberland and Franklin Counties, Pa.

[--, undetermined; Descriptions from Root, 1968]

Geologic period	Geologic formation and abbreviation (units as mapped locally)	Geologic description Modified from Root (1968)	Thickness (feet)
Quaternary	Colluvium	Mixture of clay, silt, sand, pebbles, cobbles and boulders overlying a thick residual clay layer.	[1]0-400
Ordovician	Martinsburg - Om	Lower member: buff-weathering, dark-gray shale and thin interbeds of siltstone, metabentonite, and fine-grained graywacke. Middle member, thick to massive brown-weathering medium-grained graywacke containing shale and siltstone interbeds.	1,500-3,000
	Chambersburg- Oc	Dark-gray, thin to medium-bedded, nodular limestone and minor units of thin, even-bedded, argillaceous limestone; thin bands of metabentonite.	650
	St. Paul Group - Osp	Light-gray, thick-bedded, high-calcium, micritic limestone with prominent beds and medial zone of medium-gray, granular black chert-bearing limestone, dolomite, and skeletal-detrital limestone.	600-900
	Pinesburg Station - Ops	Thick-bedded, light- to medium-gray laminated to banded dolomite that contains sparse black chert nodules and white quartz; interbeds of blue-gray limestone.	175-300
	Rockdale Run - Orr	Upper two thirds is light-gray, medium to thick-bedded, detrital to detrital-skeletal and micrograined limestone. Abundant dolomite laminae and sparse dolomite beds, white quartz beds near top. Lower third is medium-bedded, finely laminated to homogeneous, chert-bearing micritic limestone and stromatolitic limestone.	2,000-2,500
	Stonehenge - Osh	Light- to medium-gray micrograined to micritic limestone containing zones and beds that are detrital; some pinkish, chert-bearing limestone beds.	500
Cambrian	Shady Grove - Csg	Light-gray to pinkish-gray, micritic limestone; abundant nodules of brown chert, a few sandstone beds, and a few beds of laminated dolomite.	[2]800-1,000
	Zullinger - Cz	Medium-gray sand-to pebble-sized detrital limestone, stromatolitic limestone, and banded limestone containing siliceous seams; some thick beds of dolomite and calcareous sandstone.	2,500
	Elbrook - Ce	Interbedded calcareous shale, argillaceous limestone, and limestone in beds; local calcareous sandstone and siltstone beds.	3,500
	Waynesboro - Cwb	Thick-bedded, laminated, fine to coarse-grained, well-sorted, quartzitic sandstone containing thick interbeds of medium to dark-gray silty mudstone; probably includes some interbeds of carbonate rocks.	1,000-1,500
	Tomstown - Ct	Some thick massive dolomites present in middle of unit; limestone, siltstone, claystone, in lower and upper part.	1,000 -2,000
	Antietam - Ca	Light-gray, buff-weathering quartzite and quartz schist. It contains some ferruginous quartzite. It is fine-grained. It is moderately well to well bedded, with thick beds.	300
	Harpers - Ch	Dark-banded, hackly schist to slate.	2,750
	Montalto Member of Harpers Formation - Chm	Prominent middle member of massive hard, white quartzite that thickens to the north.	(included in Ch)
	Weverton and Loudon Formations, undivided - Cwl	Gray feldspathic sandstone, coarse grained. Some white quartzites. Conglomerate at base.	1,250
Pre-Cambrian	Metavolcanics - mv	Includes metarhyolite and metabasalt	--

[1]Thickness of colluvium is equivalent to depth below land surface. Other thicknesses are relative to stratigraphic column and do not represent depth.
[2]Thickness is from Becher and Taylor (1982).

Two water budget components used in simulating ground-water flow are recharge and stream base flow. Both can be estimated by analysis of long-term streamflow records or instantaneous streamflow measurements. Total streamflow is the sum of surface runoff and ground-water discharge to streams. Hydrograph separation using HYSEP (Sloto and Crouse, 1996) and data from continuous streamflow records was used to divide total streamflow into components of surface runoff and stream base flow (ground-water discharge) for streams in or near the ground-water-model area. In areas that are far enough downstream so that losses from the stream to the aquifer are not significant, the ground-water discharge component of streamflow is equivalent to recharge for long time periods. Recharge to ground water (as calculated by determining stream base flow) is one component of the water budget determined from long-term flow records and used for simulation of ground-water flow. Instantaneous measurements of stream base flow are another component of the budget used in simulation of ground-water flow. Although continuous streamflow measurements may not be available within the model area, instantaneous streamflow can be measured wherever data are needed. If instantaneous stream base-flow measurements are made during conditions representing long-term average conditions, assuming no withdrawals of water or underflow is occurring, stream base flow should be equivalent to recharge.

The average annual precipitation (1934 to 2004) at Shippensburg Weather Station is 39.4 in/yr (100.4 cm/yr). The temporal distribution of rainfall is relatively even throughout the year; however, annual rainfall is variable from year to year (Pennsylvania State Climatologist, 2004).

The ground-water model area is drained by the Conodoguinet Creek and the Yellow Breeches Creek, which both flow from west to east to the Susquehanna River, and Conococheague Creek, which flows south and west to the Potomac River (fig. 6). Conodoguinet Creek drains a 470-mi^2 watershed on the north side of the Cumberland Valley, with the exception of Middle Spring Creek, a tributary of Conodoguinet Creek that originates on South Mountain and flows northward across the valley. Between 1911 and 2004, there are 72 years of streamflow records for the Conodoguinet Creek streamflow-gaging station at Hogestown, Pa. Yellow Breeches Creek drains a 216-mi^2 watershed in the southern part of the Cumberland Valley. Between 1909 and 2004, there are 58 years of streamflow records for the Yellow Breeches streamflow-gaging station at New Cumberland, Pa. The years of common record between the Yellow Breeches and Conodoguinet Creeks (water years 1912-1916, 1955-1958, and 1968-2003) were used for calculations of base flow. The average total streamflow for Conodoguinet Creek was approximately 16.7 in. (580 ft^3/s) for that period (U.S. Geological Survey, 2004). Hydrograph separation indicated a median of 9.5 in. (or 57 percent) of the total streamflow was base flow (ground-water discharge) during that period. The average discharge for the Yellow Breeches Creek was about 19.4 in. (309 ft^3/s) for that period. Hydrograph separation indicated that 14.8 in. (or 76 percent) of the total streamflow was base flow. A summary of the water budget is given in table 2.

Evapotranspiration (ET) and ground-water evapotranspiration (GW-ET) are important processes. In the water budget, ET is calculated as the difference between precipitation and total runoff. GW-ET (evapotranspiration directly from the ground water reservoir) is the difference between recharge and ground-water discharge. Therefore, values for recharge calculated from stream base flow are actually the difference between true recharge and GW-ET. This approximation allows recharge and ground-water discharge to be in balance.

Although the streams near the well fields actually flow into Conodoguinet Creek (fig. 6), hydrologic characteristics of these streams are similar to those of Yellow Breeches Creek, based on geology and hydrology. Many of the tributaries to Conodoguinet Creek originate on Blue Mountain. These tributaries originate in the ridge-forming sandstone formations and flow across the Martinsburg Formation, which is predominantly shale. The stream network in the northern tributaries of Conodoguinet Creek watershed are very dense and have very different runoff

Table 2. Summary of water budget for Conodoguinet and Yellow Breeches Creeks. Data for base flow and surface runoff from water years 1912-1916, 1955-1958, 1968-2003.

	Conodoguinet		Yellow Breeches	
	Inches per year	Percent of budget	Inches per year	Percent of budget
Precipitation[1]	39.4	100	39.4	100
Base-flow component of streamflow	9.5	24	14.8	38
Surface-runoff component of streamflow	7.2	18	4.6	11
Evapotranspiration	22.7	58	20.0	51

[1]Average annual precipitation at Shippensburg, Pennsylvania, 1934-2004 (Pennsylvania State Climatologist, 2004).

characteristics than those of streams originating on South Mountain. A few tributaries to Conodoguinet Creek originate from the south as spring-fed streams that drain the central part of the valley, which is underlain by carbonate bedrock. Yellow Breeches Creek is unique in that most of its tributaries originate from the south and flow northerly from South Mountain into the main stem, making this area hydrologically similar to the area in which the Authority production wells are located.

Aquifer Characteristics

The storage and movement of water through an aquifer are controlled by the properties of the rock or unconsolidated material through which it is moving. These properties include hydraulic conductivity, transmissivity, specific capacity, porosity, specific yield, and degree of confinement. Definitions of these terms are from Lohman and others (1972). Hydraulic conductivity is defined as the volume of water that will move in a unit time under a unit hydraulic gradient through an area measured at right angles to the direction of flow. Transmissivity is the rate at which water will be transmitted through a unit width of the aquifer under a unit hydraulic gradient. Specific capacity is the rate of discharge of water from a well divided by the drawdown in the well. Porosity is the ratio of the volume of the voids to the total volume. Specific yield is the ratio of the amount of water that a volume of aquifer material will yield under gravity drainage to the volume of the aquifer material.

Hydraulic Conductivity and Transmissivity

The characteristics of the aquifer that affect ground-water flow include hydraulic conductivity and a related characteristic called transmissivity. The relation between transmissivity and hydraulic conductivity is given by equation 2:

$$T = K \times b, \qquad (2)$$

where T is transmissivity, K is hydraulic conductivity, and b is the aquifer thickness.

Values for hydraulic conductivity and transmissivity can be determined by the use of information from several sources. (1) An aquifer test conducted on well Cu 969, which is completed in the Tomstown Formation, was used to determine transmissivity. Transmissivity from the aquifer test was calculated as 20,000 ft^2/d (Stuart Reese, Pennsylvania Department of Environmental Protection, written commun., 1990). Using an estimated aquifer thickness of 200 ft, the hydraulic conductivity for this formation is 100 ft/d. The thickness of the aquifer was not known, but the distance between the average water level and the bottom of a well is about 200 ft. (2) A pump test conducted on well Cu 970 (Higgins, 2001) indicated a transmissivity of 13,100 ft^2/d and, using an aquifer thickness of 160 ft, a hydraulic conductivity of 82 ft/d. (3) Other estimates of hydraulic conductivity were obtained or calculated from previous studies and are summarized in table 3. Values used by Chichester (1996) in a ground-water-flow model of the Cumberland Valley and values from aquifer tests compiled by Low and others (2002) are included in table 3. Some of the values of hydraulic conductivity in table 3 are calculated from specific-capacity tests using an iterative method developed by Thomas and others (1999) shown in equation 3:

$$T = (Q/4\pi s)\ \ln\ (2.25 T_e t/r^2 S), \qquad (3)$$

where

T is transmissivity, in feet squared per day;

Q is the pumping rate, in cubic feet per day;

s is drawdown, in feet;

T_e is an initial estimate of T (the specific capacity times 100 was used for the initial estimate);

t is duration of the test, in days;

r is the radius of the well, in feet;

and

S is the formation storage coefficient (dimensionless).

T_e is replaced with T iteratively until T_e is approximately equal to T. Values for K were calculated by dividing the resultant "T" by the open interval of the well.

Specific-capacity test data from Fleeger and others (2004) were used to make the calculations. Values for the 25th and 75th percentile of data are included to show potential variability in hydraulic conductivity. The values for hydraulic conductivity from Chichester (1996) or Low and others (2002) were used as initial estimates for modeling and also used to evaluate calibrated values of hydraulic conductivity.

Porosity and Specific Yield

The amount of water that can be stored in an aquifer is related to the porosity of the aquifer material. In unconsolidated aquifers, the water fills the voids or open spaces between the particles. In fractured bedrock, the water fills the fractures in the bedrock and, to some degree, penetrates the rock itself. The voids in the rock are referred to as matrix porosity and are generally a very small percentage of the total porosity of the rock. The fractures in the rock are referred to as secondary or fracture porosity. Porosity is difficult to measure in the field without extensive testing; however, values of porosity have been estimated from numerous previous studies. A characteristic closely related to porosity is specific yield, which is the ratio of the volume of water that is released from the pore space when the aquifer is drained by gravity to the aquifer volume. Some materials such as clay may have a large value of porosity yet retain much of that water, whereas other materials such as sand have a lower value of porosity but retain very little water when drained. Because the purpose of this model is to determine the shape of the areal extent of the zone of contribution to the production wells, the model is a steady-state model and values of porosity and specific yield have no effect on the outcome of the modeling. Porosity would be important in determining traveltimes,

Table 3. Reported hydraulic conductivity values for geologic units in the Great Valley and values calculated from specific-capacity tests.

	Hydraulic conductivity, in feet per day				
			Value calculated from specific capacity[3]		
	Chichester[1]	Low[2]	25th percentile	Median	75th Percentile
Colluvium	75	--			
Martinsburg Formation	2.5		0.72	1.8	4.3
Chambersburg Formation	25		.12	.32	3.2
St. Paul Group	55		.14	2.1	18
Pinesburg Station Formation	31	--	.22	2.6	9.0
Rockdale Run Formation	56	--	1.8	22	83
Stonehenge Formation	26	[4]52	.44	16	56
Shady Grove Formation	10	--	.10	3.6	11
Zullinger Formation	.94	--	.40	2.2	20
Elbrook Formation	.83	[4].58	.50	12	44
Waynesboro Formation	5.7	--	1.4	7.0	21
Tomstown Formation	28	--	8.8	71	240
Antietam Formation	--	[4].68/.47			
Harpers Formation	--	[4].52/.16			
Montalto Member of Harpers Formation	--	.41			
Weverton Formation	--	.01			

[1]Data from calibrated model of Chichester (1996).
[2]Data from Low and others (2002).
[3]Data from Fleeger and others (2004) using the method of Thomas and others (1999).
[4]Data from Low and others (2002) in Piedmont Physiographic Province.

and specific yield would be important in determining the aquifer response to changes in conditions, such as a drought or adding additional wells.

Water Levels and Streamflow

Data documenting hydrologic conditions during the study were collected from June 2003 through March 2004. These data included continuous measurements of water levels in one well (Cu 970), a single water-level measurement in wells throughout the ground-water model area, a single streamflow measurement in streams throughout the area, and periodic measurements of stream discharge from a major spring in the area. A ground-water-flow model by Chichester (1996) simulated water levels and generated a water-level contour map that provides background information on conditions in a part of the current ground-water-model area (fig. 11). Chichester's model did not extend into the metamorphic- bedrock aquifer or into the Conococheague Creek watershed.

Well Cu 970, a well owned by the Authority, is intended to be used as a production well; however, this well was unused at the start of this study. The location of this well is shown on figure 12. A submersible pressure transducer was placed in the well to record fluctuations in the water level at 15 minute intervals from June 2003 to April 2004. Daily mean water levels from this well are illustrated in figure 13. Measurements of flow in this well indicated a strong upward flow of water within the borehole under static conditions (fig. 14). Water levels observed in well Cu 970 show a somewhat regular oscillation that could be related to earth tides or pumping of well Cu 969, a nearby production well. Earth tides are related to aquifer response to the diurnal lunar/solar cycle and are indicated by a rise and fall in water levels that corresponds to this cycle. Potential vertical displacement due to earth tides (University of Bern Astronomical Institute, 2005) was compared to the water levels recorded in well Cu 970 and pumping cycles at well Cu 969 (fig. 15). The fluctuations in well Cu 970 did not correlate to the pumping cycles in well Cu 969 or the earth-tide cycles. Water levels in well Cu 970 also were observed to respond to changes in barometric pressure. Therefore, the daily oscillations seen in well Cu 970 are likely to be a combination of earth tides, pumping of well Cu 969, barometric response, and recharge. The nearest USGS observation well, Fr 818 (in Franklin County about 15 mi to the south) (fig. 13), is in a similar geologic for-

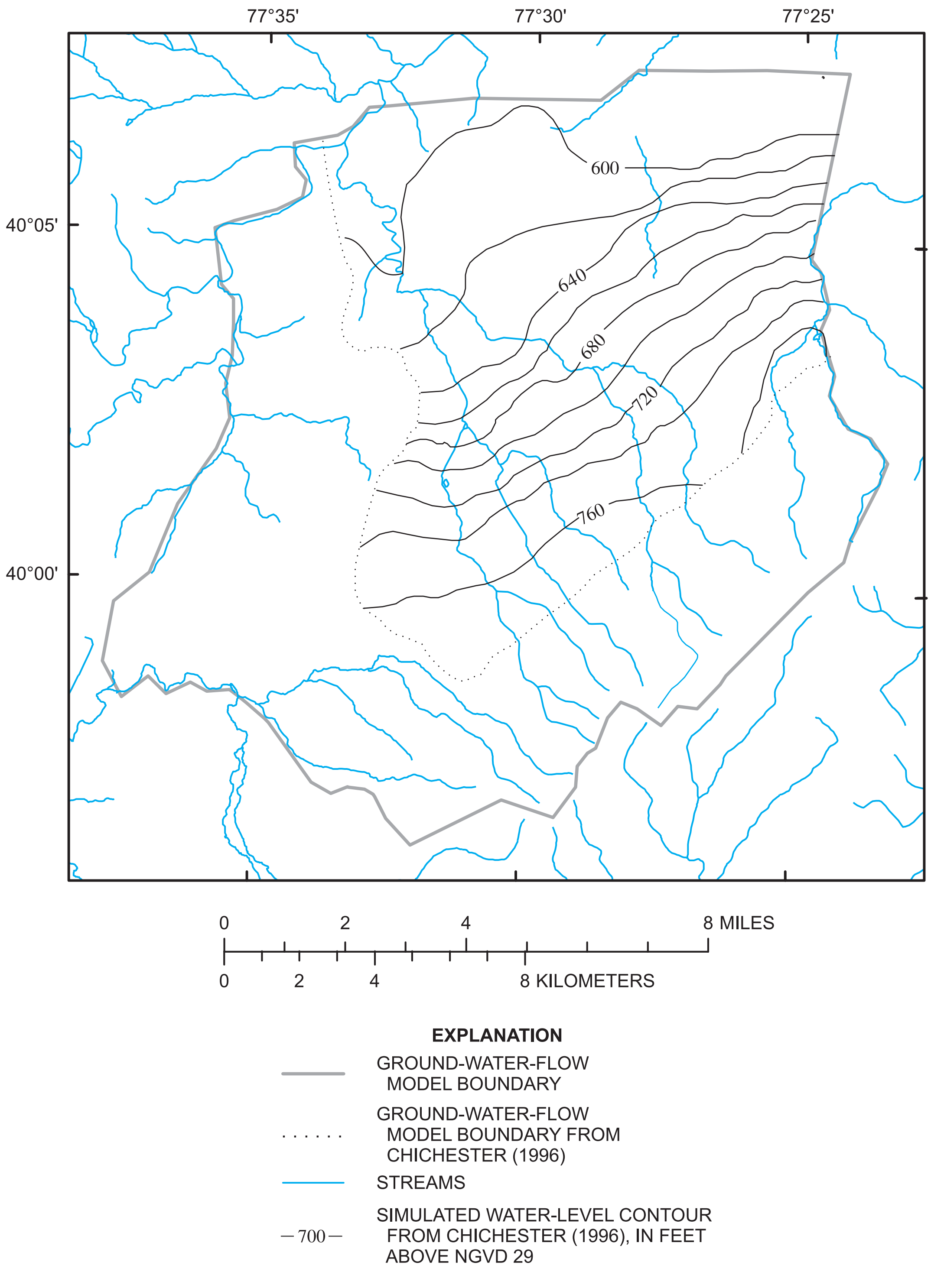


Figure 11. Simulated water-level contours from a ground-water-flow model near Shippensburg, Pa. (modified from Chichester, 1996).

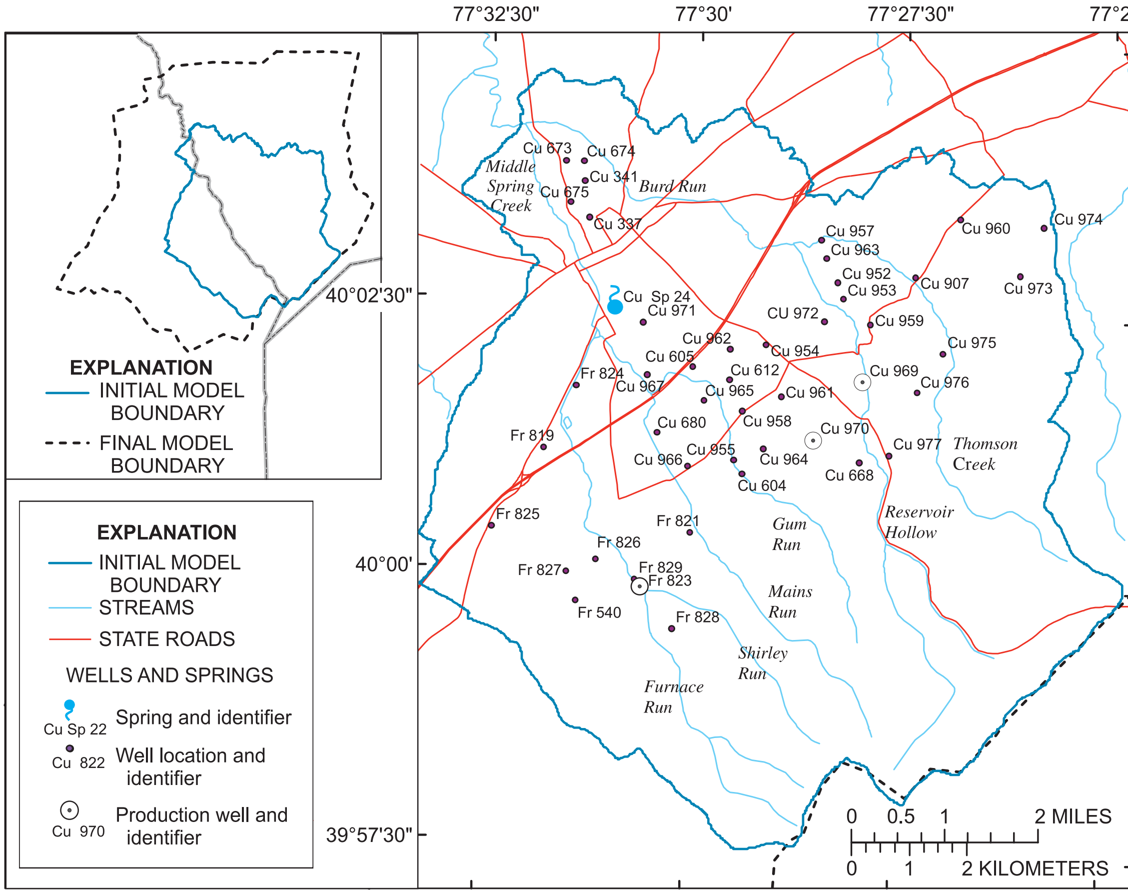


Figure 12. Locations of springs, wells, and production wells measured near Shippensburg, Pa.

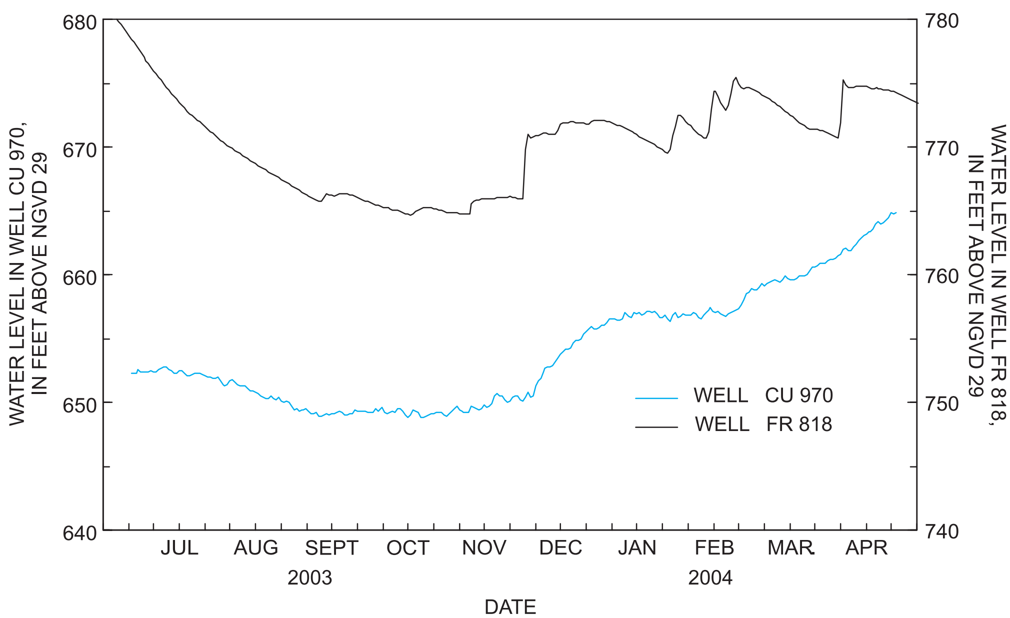
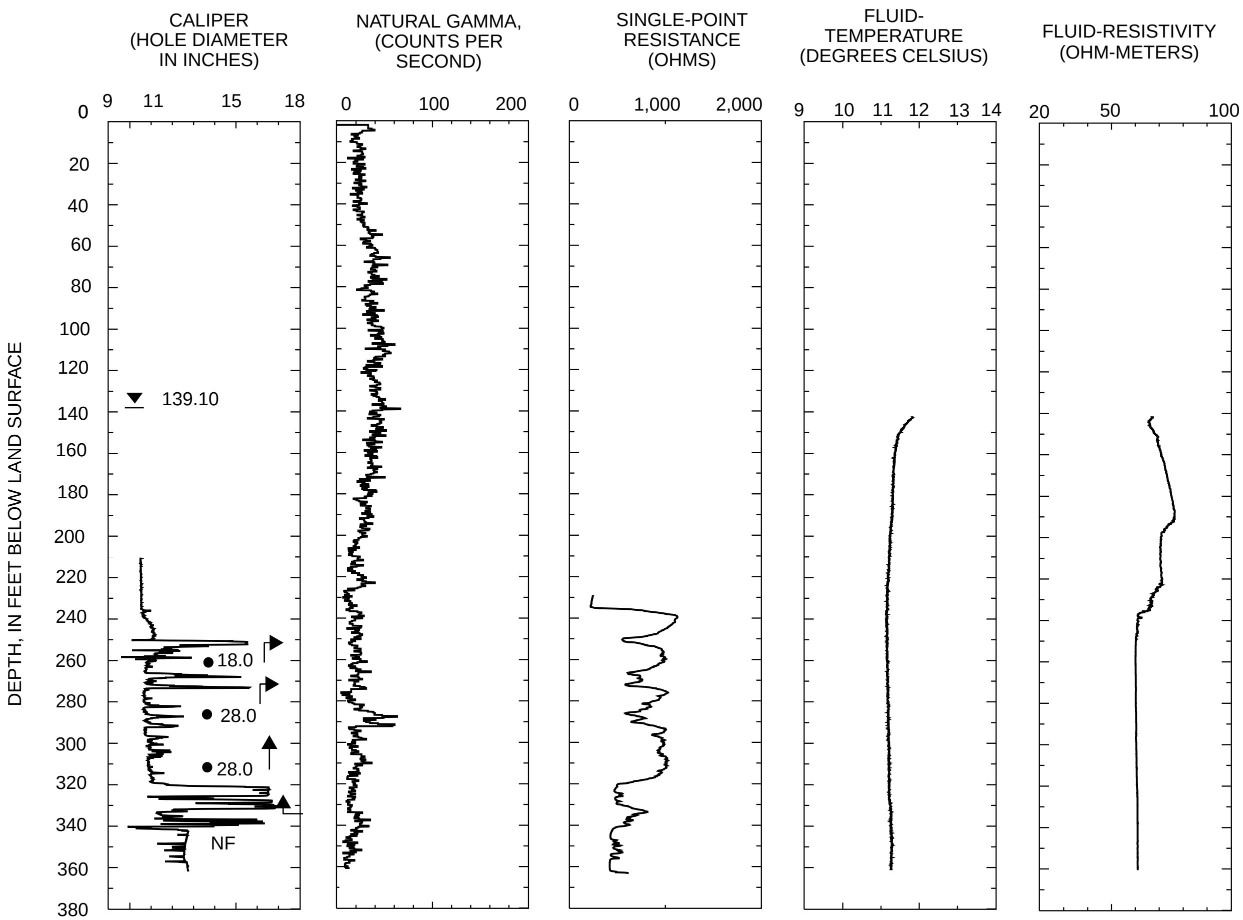


Figure 13. Water levels in well Cu 970 near Shippensburg, Pa., and the U.S. Geological Survey Franklin County observation well (Fr 818), near Chambersburg, Pennsylvania, June 2003 to October 2004.



EXPLANATION

139.10 STATIC WATER LEVEL—Measured in well at the time of geophysical logging.

18.0 BOREHOLE-FLOW MEASUREMENT UNDER NONPUMPING CONDITIONS— Circle at depth of flow measurement. Number is estimated flow in gallons per minute.

DIRECTION OF BOREHOLE FLOW UNDER NONPUMPING CONDITIONS

WATER-PRODUCING ZONE UNDER NONPUMPING CONDITIONS (Arrow indicates direction of flow into or out of the borehole)

WATER-RECEIVING ZONE UNDER NONPUMPING CONDITIONS

NF NO FLOW

Figure 14. Borehole geophysical logs and direction of flow within borehole Cu 970 (Shippensburg Well #3), near Shippensburg, Pa. (Borehole geophysical logs collected on April 6, 2004).

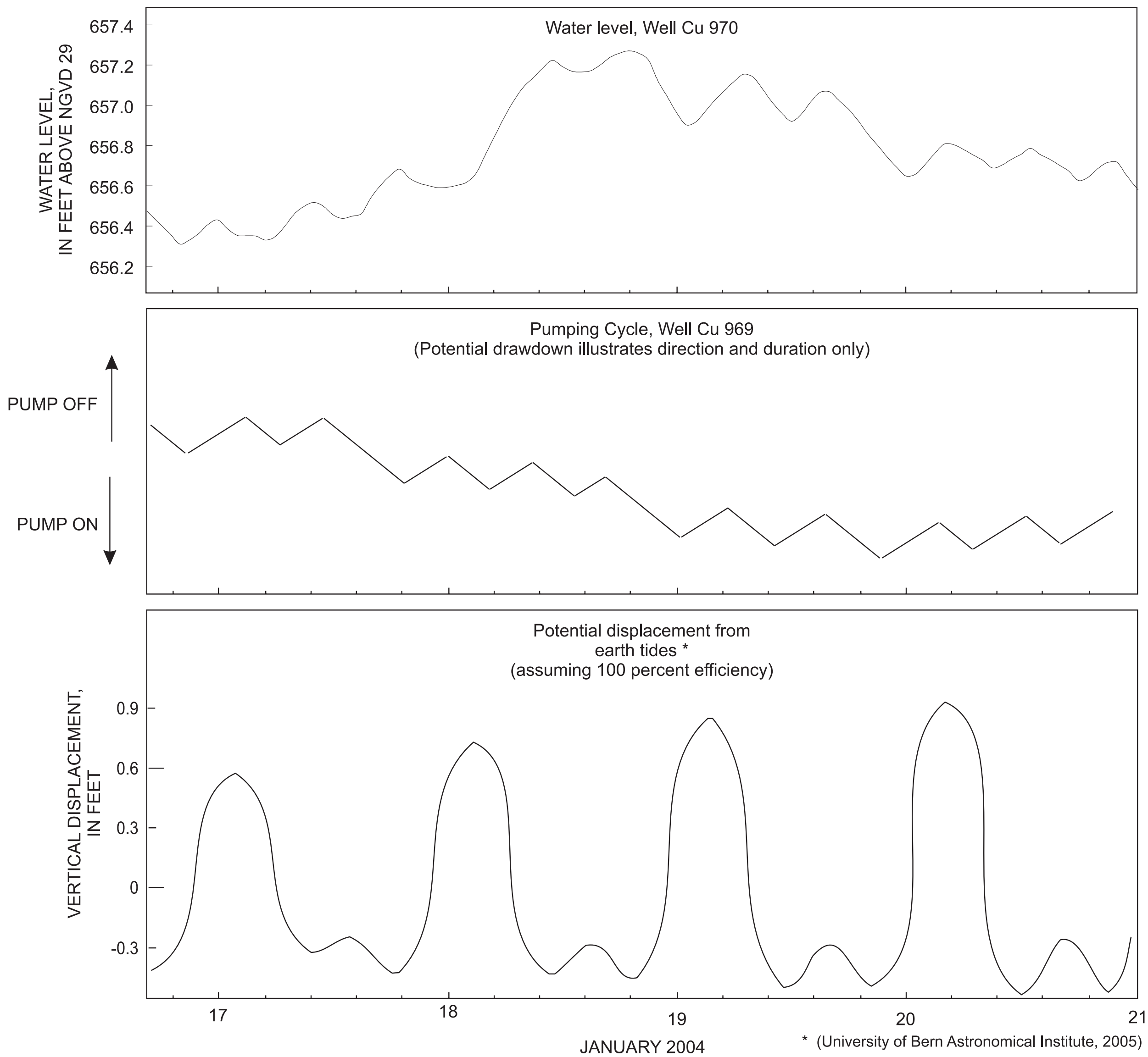


Figure 15. Fluctuations in water levels in well Cu 970, pumping cycle in well Cu 969, and potential displacement due to earth tides near Shippensburg, Pa., January 2003.

mation to Cu 970, although it is shallower and not in an area with a deep colluvium cover. A comparison of water levels in well Cu 970 to USGS observation well Fr 818 illustrates that well Cu 970 has a much lower response to hydrologic events than does well Fr 818. The muted response to hydrologic events, upward flow in the borehole, barometric response, and earth tides indicate well Cu 970 probably is completed in a confined aquifer.

Wells were used as locations to measure static water levels or to measure water pumped from the aquifer. Static water levels were measured once in 43 wells (fig. 12) throughout the ground-water-model area during November 2003 through April 2004 and used to calibrate water levels in the model. Most water levels were from domestic wells and were measured during November 2003–January 2004. Water levels and locations of these wells are given in table 4. Water-level measurements were made inside the ground-water-model boundary used for the initial modeling. The ground-water-model boundary was modified to its final location during the modeling process and, therefore, no water levels were measured in the area outside the initial ground-water-model boundary. Measurements of discharge from production wells were used to simulate pumping in the ground-water model (fig. 12). Water levels from production wells were not used to calibrate water levels in the model.

Streamflow was measured at 20 sites in the initial ground-water-model area. These measurements were made along stream segments to document gaining and losing reaches of the stream. Measurements were made within a 2-day period to allow comparisons of measurements made at approximately the same time, in a type of study called a seepage run. Two hydro-

Table 4. Wells and springs used for model design and/or calibration near Shippensburg, Pa.

[ft, feet; n/a, not applicable; --, no data; NAD 83, North American Datum of 1983; NGVD 29, National Geodetic Vertical Datum of 1929; *, water level not used for model calibration]

County well or spring number (Locations shown on figure 12)	Aquifer	Latitude (degrees, minutes, seconds, NAD 83)	Longitude (degrees, minutes, seconds, NAD 83)	Elevation of land surface (ft above NGVD 29)	Well depth (ft)	Casing length (ft)	Water level below land surface (ft)	Model Layer	Date of water-level measure-ment
Cu 337	Rockdale Run	400322	773117	655	142	39	14.9	2	1/6/2004
Cu 339	Rockdale Run	400342	773121	655	105	20	22.6	2	1/6/2004
Cu 604	Tomstown	400102	772922	840	235	225	106.4	2	11/21/2003
Cu 605	Elbrook	400201	772959	770	110	80	26.1	2	11/24/2003
Cu 612	Elbrook	400154	772933	775	125	51	35.4	2	11/21/2003
Cu 668	Tomstown	400112	772758	935	460	450	199	2	3/01/1972
Cu 673	Rockdale Run	400353	773135	645	144	52	34	2	1/6/2004
Cu 674	Rockdale Run	400353	773122	625	60	47	9	2	1/6/2004
Cu 675	Rockdale Run	400330	773131	655	150	24	33	2	1/23/2004
Cu 680	Elbrook	400124	773024	748	97	50	9.3	1	12/5/2003
Cu 907	Elbrook	400253	772720	790	129	120	48.95	1,2	11/17/2003
Cu 952	Elbrook	400249	772816	770	232	80	60.8	1,2	11/24/2003
Cu 953	Elbrook	400240	772812	775	285	40	57.8	1,2	11/24/2003
Cu 954	Elbrook	400214	772907	803	198	147	79.1	2	11/21/2003
Cu 955	Waynesboro	400110	772928	815	273	200	80.3	2	11/21/2003
Cu 957	Zullinger	400313	772829	705	180	105	10.9	2	11/21/2003
Cu 958	Elbrook	400137	772923	775	73	63	22.9	1	11/24/2003
Cu 959	Elbrook	400226	772752	850	252	242	115.4	2	11/24/2003
Cu 960	Elbrook	400326	772649	790	250	60	87.9	1,2	11/24/2003
Cu 961	Waynesboro	400145	772855	880	246	241	149	2	11/24/2003
Cu 962	Elbrook	400211	772933	830	248	100	66.3	*	12/03/2004
Cu 963	Zullinger	400303	772825	730	248	191	36.4	2	11/21/2003
Cu 964	Waynesboro	400016	772906	835	248	222	129	*	11/17/2003
Cu 965	Elbrook	400142	772951	770	181	118	21.3	2	12/5/2003
Cu 966	Waynesboro	400106	773001	835	180	161	81.4	1	12/5/2003
Cu 967	Elbrook	400156	773032	720	120	97	21	2	12/5/2003
Cu 969	Tomstown	400155	772755	835	590	343	--	2	12/16/2003
Cu 970	Tomstown	400122	772828	870	360	240	149.65	2	12/16/2003
Cu 971	Elbrook	400225	773036	705	148	84	25.20	2	2/23/2004
Cu 972	Elbrook	400228	772825	780	170	147	57.2	2	3/10/2004
Cu 973	Waynesboro	400256	772604	840	273	152	145.4	2	3/10/2004
Cu 974	Waynesboro	400323	772548	870	280	240	156.7	2	3/10/2004
Cu 975	Tomstown	400211	772659	880	448	209	157.1	2	2/23/2004
Cu 976	Tomstown	400150	772717	960	475	400	264.1	2	2/25/2004
Cu 977	Tomstown	400114	772736	985	480	371	245.2	2	4/6/2004
Cu Sp 22	Shady Grove	400746	772427	510	n/a	n/a	n/a	1	n/a
Cu Sp 24	Elbrook	400231	773054	670	n/a	n/a	n/a	1	n/a
Fr 540	Waynesboro	395954	773121	830	200	123	60	*	3/10/2004
Fr 819	Elbrook	400114	773146	725	68	40	40.5	1,2	12/5/2003
Fr 821	Tomstown	400029	772958	905	300	260	118.9	2	12/5/2003

Table 4. Wells and springs used for model design and/or calibration near Shippensburg, Pa.

[ft, feet; n/a, not applicable; --, no data; NAD 83, North American Datum of 1983; NGVD 29, National Geodetic Vertical Datum of 1929; *, water level not used for model calibration]

County well or spring number (Locations shown on figure 12)	Aquifer	Latitude (degrees, minutes, seconds, NAD 83)	Longitude (degrees, minutes, seconds, NAD 83)	Elevation of land surface (ft above NGVD 29)	Well depth (ft)	Casing length (ft)	Water level below land surface (ft)	Model Layer	Date of water-level measure-ment
Fr 823	Tomstown	400001	773031	852	400	304	92.7	2	12/16/2003
Fr 824	Elbrook	400149	773123	710	62	40	32.3	1	3/10/2004
Fr 825	Elbrook	400030	773222	760	140	60	75.5	2	3/10/2004
Fr 826	Waynesboro	400017	773106	810	180	130	19.9	*	2/23/2004
Fr 827	Elbrook	400010	773126	800	140	80	9.0	*	3/10/2004
Fr 828	Tomstown	395935	773010	980	600	560	185.5	2	2/23/2004
Fr 829	Waynesboro	400002	773038	850	310	229	90	2	4/6/2004

logic technicians made all streamflow measurements, wading the stream and using a pygmy current meter as described in Rantz (1982). Each individual making measurements focused on a single tributary at a time, and measurements were made sequentially from upstream to downstream on each tributary. The results of this seepage run are illustrated on figure 16. The study showed that all the streams flowing from South Mountain into the Great Valley lose water as they cross the contact between metamorphic and carbonate rocks and begin to flow across the colluvium, with some streams losing more than one cubic foot per second. The volume of water loss varies, but all the streams showed some decrease in flow through this area during November, 2003. Given a potential streamflow-measurement error of 10 percent, the determination of streamflow gain or loss illustrated on figure 16 would not change for most reaches. Because streams are flowing in the uppermost reaches measured and also are near the stream headwaters, streams are assumed to be gaining upstream from the uppermost measurement site, the actual location where a stream begins gaining is undetermined. Under drought conditions, some of these streams flow from the mountain, but lose all of their water as they enter the valley (Thomas Feeney, Shippensburg Borough Authority, oral commun., 2004).

Borehole Geophysical Logs

In addition to the hydrologic data collected from wells and streams, geophysical logs were collected in wells Cu 970 and Fr 829. Well Cu 970 is a future supply well, and well Fr 829 is near current supply well Fr 823. Geophysical logs provide information on aquifer characteristics, locations of open intervals, borehole flow, changes in lithology, and orientation of fractures. Findings from the borehole geophysical logs were used to help build the conceptual model used to simulate ground-water flow.

The caliper log for well Cu 970 indicated the casing extended to 240 ft below land surface, and the total depth of the well was 360 ft. The water level at the time of logging was 139.10 ft below land surface. The caliper log indicated fractures or large fracture zones at 250, 267, 274, 320, 330, and 337 ft (fig. 14). Some of the fractures appeared to be voids. The existence of voids was verified by downhole video logs; most voids were near horizontal and extended beyond the viewing field of the camera. The gamma log indicated the lithology was either sand or limestone throughout the length of the borehole with a low clay content in the sandy material, and little shale in the limestone. Flowmetering was conducted under static conditions; however, the heatpulse flowmeter measured a large volume of water flowing upward through the borehole (fig. 14). Water was produced from the bottom fracture at 337 ft, and upward flow was sustained at an estimated rate of about 28 gal/min through most of the open interval of the borehole. About 40 percent of the upward flow exited the borehole through fractures 260–280 ft below land surface, and the remainder exited at the fracture below the casing at approximately 250 ft below land surface. Acoustic televiewer logs could not be conducted on this well because the diameter of the borehole was too large.

The caliper log for well Fr 829 indicated the 6-in. casing extended to 229 ft, and the total depth was 310 ft below land surface (fig. 17). The bottom of the borehole appeared to be collapsed, and the logging could not determine the depth of the original borehole. The water level at the time of logging was 90.40 ft below land surface. The caliper log shows fractures or fracture zones at 232, 246, 263, 274, 281, and 293 ft below land surface. The gamma log indicates the lithology was either sand or limestone throughout the length of the borehole with a low clay content in the unconsolidated material and little or no shale in the bedrock. Also, the gamma log shows an increase in counts from 18 to 119 ft below land surface that may be because

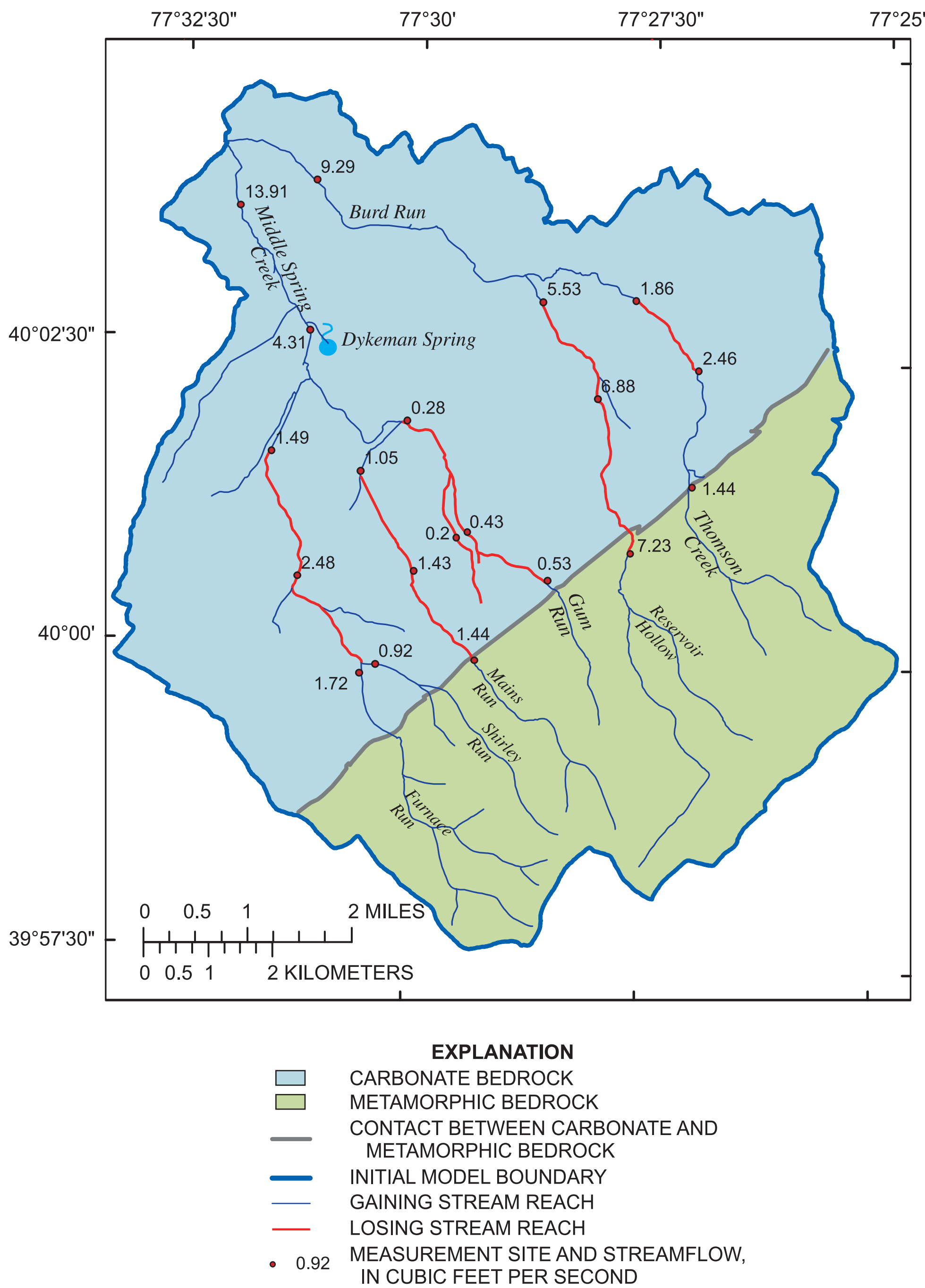
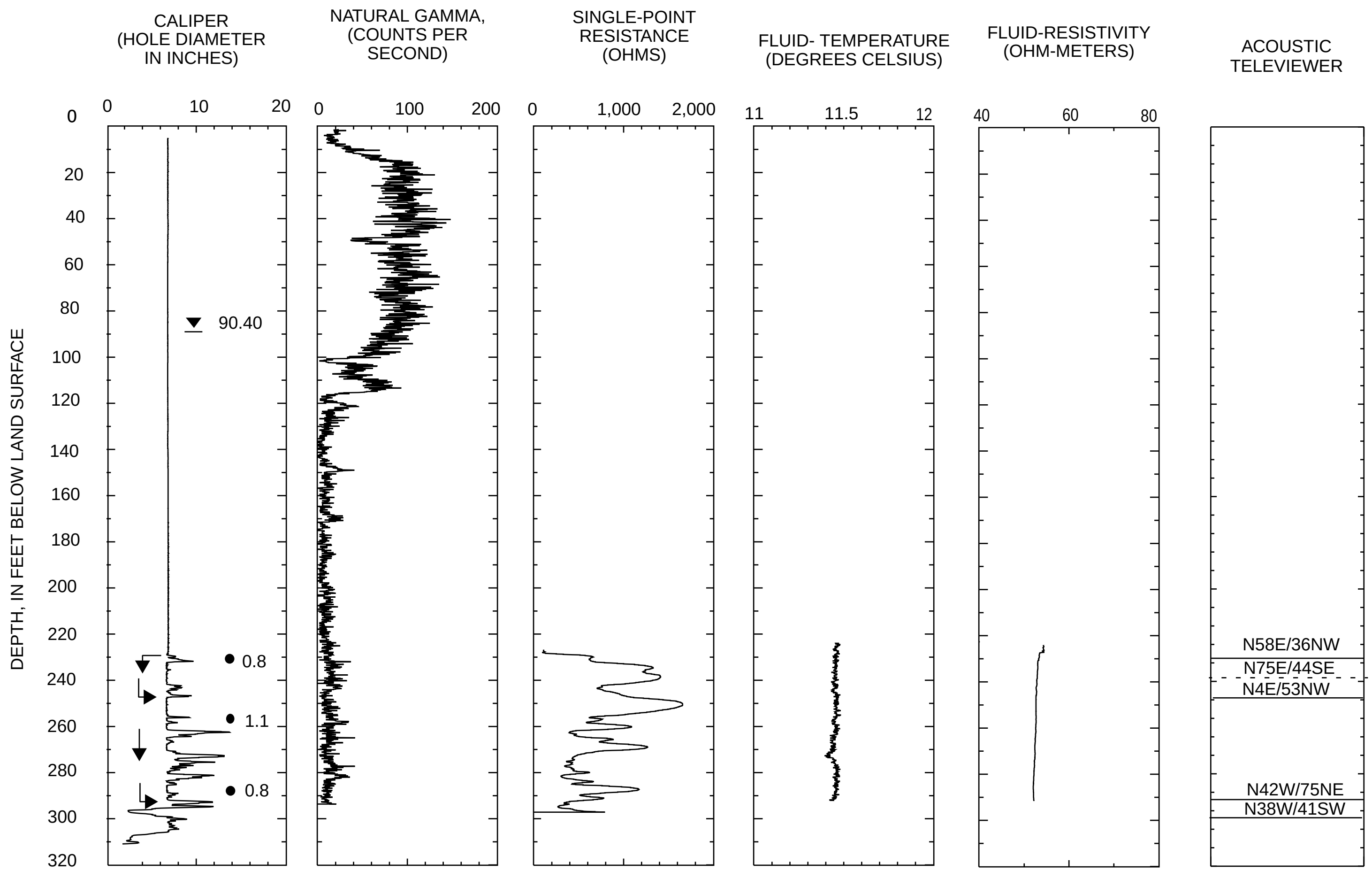


Figure 16. Streamflow-gaging sites and losing stream reaches near Shippensburg, Pa., November 2003.



EXPLANATION

90.40 STATIC WATER LEVEL—Measured in well at the time of geophysical logging.

● 1.0 BOREHOLE-FLOW MEASUREMENT UNDER NONPUMPING CONDITIONS— Circle at depth of flow measurement. Number is measured flow in gallons per minute.

DIRECTION OF BOREHOLE FLOW UNDER NONPUMPING CONDITIONS

N58E/36NW WATER PRODUCING/RECEIVING FRACTURE—Showing strike and dip in degrees.

N75E/44SE BEDDING PLANE—Showing strike and dip in degrees.

WATER-PRODUCING ZONE UNDER NONPUMPING CONDITIONS

WATER-RECEIVING ZONE UNDER NONPUMPING CONDITIONS

Figure 17. Borehole geophysical logs and direction of flow within borehole Fr 829 (Mainsville Monitor Well), near Shippensburg, Pa. (Borehole geophysical logs collected on April 7, 2004.)

of the clay content of the construction grout. This increase is typical of deflections that indicate grout rather than lithologic changes. The heatpulse flowmeter measured downward flow throughout the open borehole, with a flow rate of about 0.8 gal/min produced from the fracture at 232 ft below land surface, and flow rates ranging from 0.8 to 1.1 gal/min flowing downward and possibly exiting the borehole through fractures at 293 ft below land surface. The downward flow of about 0.8 gal/min was exited the borehole at some depth greater than 293 ft below land surface. Acoustic televiewer logs indicated two predominant fracture sets in the borehole. One of the fracture sets is oriented NE-SW and the other is oriented NW-SE. Only those fractures producing or receiving water and an example of fractures representing bedding are illustrated in the acoustic televiewer log on figure 17. The set of fractures that probably represents bedding is oriented with a strike of N. 75° E. and a dip of about 44° SE. The dip is not in accordance with the dip of bedding that would be expected in this area (50 to 70° NW.) and may indicate the well is completed in a unmapped local fold in this strata.

Analysis of the geophysical logs illustrates several important hydrologic factors: 1) the unconsolidated materials appear to be more sand than clay on both logs, 2) voids exist in the boreholes, indicating ground-water flow is likely to be dominated by conduit flow in some areas, and 3) vertical gradients (both upward and downward) exist in this area. The large upward flow rates in well Cu 970 indicate the deeper aquifer is confined, and water recharged upgradient from the well is moving through the deeper aquifer and is locally isolated from the upper aquifer. The downward flow in well Fr 829 is noteworthy also. The well is within about 20 ft of Furnace Run, but the water level is 90.4 ft below the land surface. This indicates that Furnace Run, which is a losing reach near well Fr 829, is not directly connected to the bedrock aquifer in this area. In addition, the downward flow in the borehole indicates a downward hydraulic gradient in the aquifer at this location. Well Fr 829 is within several hundred feet of well Fr 823, which was pumping during geophysical logging, and the downward gradient may be because of drawdown from the pumping well. Finally, the acoustic televiewer logs indicate the orientation of bedding in this area is not in accordance with what would be expected on the basis of geologic maps. The logs indicate minor folds in these geologic formations that are not apparent on geologic maps.

Simulation of Source Areas of Water to Production Wells

Ground-water flow was simulated for the area of South Mountain and the Great Valley referred to as the ground-water-model area using MODFLOW 2000 (Harbaugh and others, 2000; Hill and others, 2000). The model was developed to determine the areal extent of the zone of contribution to the Authority wells. The areal extent of the zone of contribution, assuming no infiltration from streams, is the aquifer volume through which water is drawn to a well, projected to the land surface (Risser and Barton, 1995). A rigorous approach to determine the areal extent of the zones of contribution to wells commonly includes ground-water-flow modeling. In areas of fractured bedrock and, in particular, areas underlain by carbonate bedrock, predictions of ground-water-flow direction using ground-water-flow models can be uncertain. This uncertainty results because the model used was designed for porous media in a homogeneous, isotropic aquifer. Fractured rock is not a porous medium; however, on a large scale, the model can simulate flow through fractured rock as an equivalent to a porous medium. Issues related to anisotropy and heterogeneity of aquifer materials can be simulated in the model. Because of these issues, estimates of the uncertainty of the simulation are included. Because measurements of streamflow indicate recharge from the stream to the aquifer, the watersheds of these streams also are evaluated as potential contributing areas.

Conceptual Model

A conceptual model is a simplified description of how the ground-water-flow system functions. The conceptual model uses measurable physical properties of the hydrologic system, such as topography and streamflow, coupled with basic principles of hydrology to infer a logical starting point for more detailed numerical modeling. Some components of the conceptual model include recharge mechanisms, interbasin transfer of ground water, relative permeability of aquifers, and the effect of geologic structure on ground-water flow.

Aquifer recharge in the study area has two potential sources. The first is areal recharge from precipitation. The amount of areal recharge from precipitation can be estimated from hydrograph separation, using the assumption that long-term base flow is equal to recharge. The distribution of recharge can vary spatially, depending on topography and the characteristics of surficial materials; however, because the streamflow-gaging stations are on large streams, information about streamflow to distinguish spatial variations in recharge on a local scale is insufficient. The second potential source of aquifer recharge is from streams that lose water to the aquifer. The rate of recharge from the losing streams was measured in the seepage run in November 2003. On the scale of the entire model area, water in the streams available for recharge originated from ground water discharged from the aquifer to the stream further upgradient, so with respect to water moving from streams to the aquifer, there is no net gain of water in the overall water budget as there would be if the stream was flowing in from outside of the model area. Thus, the two components of recharge are areally distributed recharge from precipitation and recharge from streams to the aquifer; however, only the recharge from precipitation is a gain to the system as a whole.

Another component of the conceptual model is interbasin transfer of water (underflow). Becher and Root (1981) determined that ground-water basins did not coincide with surface-water basins in the Cumberland Valley. Their findings indicated ground water flowed from the Yellow Breeches Creek watershed northward across the valley to the Conodoguinet Creek watershed beneath the surface-water divide. Chichester (1996) illustrated with a digital flow model that water discharging from Big Spring originated on the South Mountain side of Yellow Breeches Creek. Similarly, the seepage run conducted for this study indicated streams flowing towards the north from South Mountain not only lost water as they traveled across the colluvium, but they had normalized flows (flow per unit area) that were well below the values those streams would have if the streams were transmitting all the water recharged in that area. This indicates a large percentage of the water recharged in the headwaters areas is moving downgradient in the ground-water system and is likely discharging to a larger regional stream such as Conodoguinet Creek or Yellow Breeches Creek.

Anisotropy is a characteristic of an aquifer that allows ground water to flow easier in one direction than in another direction. The surface drainage system is perpendicular to the geologic structure in this area. In some areas, streams intersect a resistant topographic high and change their course until they can bypass the obstacle. Similarly, it is possible that in the fractured-bedrock aquifer, ground water may be encountering geologic units of higher and lower permeability as it flows downgradient. In addition, fractures may be oriented along planar surfaces, such as bedding, and therefore allow preferential flow in the direction of these fractures. Because the hydraulic gradient is perpendicular to the strike of the beds, flow in the downgradient direction would be through fracture sets developed along cleavage or other cross-fractures, which are likely to be less abundant than those developed parallel to the strike of beds. These factors would impart anisotropy into the aquifer in that it may be more difficult for water to move perpendicular to strike than it is to move parallel to strike of the geologic formations. Another potential source of anisotropy can be from the orientation of major fractures. Fracture traces measured by Becher and Root (1981) and Becher and Taylor (1982) are summarized in a rose diagram (fig. 18). This illustration shows the fracture traces have a preferential orientation, which is approximately east-west. Because these fracture traces were identified from land-surface features, this set may overrepresent vertical or high-angle fractures. Both conceptual models of anisotropy (fracture orientation and geologic structure) were evaluated during modeling.

There are four general aquifer types in the ground-water-model area; three are based on general rock type, and the fourth

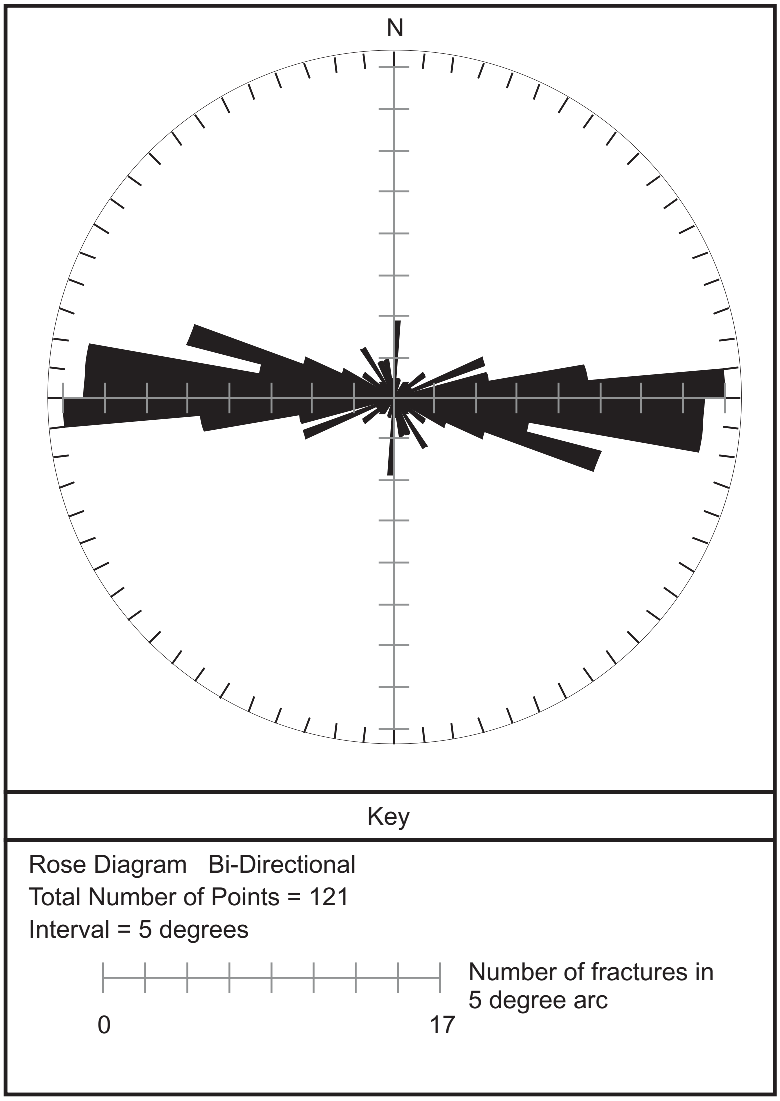


Figure 18. Rose diagram illustrating orientation of fracture traces in the study area (from Becher and Root, 1981; Becher and Taylor, 1982).

is composed of unconsolidated materials. The three general rock types are carbonate, metamorphic, and siliciclastic. The carbonate rocks include limestones and dolomites; many of the carbonate formations have a mixture of both lithologies. Generally, the carbonate-bedrock aquifers have solutionally enlarged fractures, yield large quantities of water, and are a reliable source of water for supply wells. The degree of solutional enlargement of the carbonate rocks has a large effect on the permeability of the carbonate-bedrock aquifer. Metamorphic rocks in the area are predominantly quartzite and schist. In general, the metamorphic rocks are resistant to weathering and are poorly fractured. Because metamorphic rocks are not as susceptible to chemical weathering as the carbonate rocks, those fractures that do exist have not been enlarged by dissolution; therefore, the metamorphic-rock aquifers transmit smaller volumes of water in comparison to the carbonate-bedrock aquifers. The siliciclastic rocks include shales and sandstones. Typically, siliciclastic-bedrock aquifers are less transmissive than the carbonate-bedrock aquifers but more transmissive than the metamorphic-rock aquifers. The fourth aquifer type consists of unconsolidated materials. The most important unconsolidated material is colluvium, which includes rubble, sands, and clays. The permeability of the colluvium depends on the relative percentage of sand and clay. The higher the percentage of sand, the higher the permeability. The areal extent and thickness of the colluvium is illustrated in figure 9. The other unconsolidated materials in this area are soils that overlie the bedrock aquifers where colluvium does not exist and residuum that exists between the colluvium and bedrock (fig 8). Typically, the soils at the land surface have a high clay content, but they can have macropore structures (root holes, animal burrows, and desiccation or cooling cracks) that are highly permeable. Other materials also overlie the bedrock, such as highly weathered remnants of the original rock. The permeability of these residual materials is highly variable, but they typically are not a barrier to the vertical infiltration of water into the deeper bedrock aquifer.

Model Design

The ground-water model is designed by taking the precepts of the conceptual model and assigning numerical values to represent the characteristics of ground-water flow. The first step in establishing the ground-water model is to define the boundaries. The external boundaries in the model are either stream cells, constant-head cells, or no-flow boundaries. The no-flow boundaries are either hydrologic divides or boundaries assumed to be parallel to ground-water-flow directions. Initially, the ground-water-model boundary was established using the watersheds of Burd Run and Middle Spring Creek Run (fig. 16). However, in developing the model it was determined that because of the significant amount of ground water moving downgradient under the surface-stream network, it was necessary to move the model boundaries out to the natural discharge points for ground water. The boundaries of the model include a no-flow boundary at the top of South Mountain, a surface-water divide that is likely also a ground-water divide (location “A”, fig. 19). The eastern model boundary is Yellow Breeches Creek and is represented by stream cells (location “B”, fig. 19) and a no-flow boundary (assumed surface-water divide-parallel to ground-water-flow direction) between Yellow Breeches Creek and Big Spring, using the assumption that ground-water flow is generally parallel to the no-flow boundary in this area. Big Spring (location “C” in figure 19) is simulated as a constant-head node. Big Spring is a large regional drain, and the assumption is that much of the water in the northern part of the model area is moving towards this location. The no-flow boundary along the northern model boundary essentially forces ground water in that area to move northeast toward Big Spring. The remainder of the northwestern boundary of the model represents tributaries to Conodoguinet Creek (location “D”, fig. 19) and are simulated as stream cells. A small no-flow boundary (assumed surface-water divide) is between the Conodoguinet Creek and the Conococheague Creek (location “E”, fig. 19), and the remaining southwest model boundary is stream cells representing Conococheague Creek (location “F”, fig. 19). The model boundaries are placed far enough away from the municipal wells so that inaccuracies in boundary locations will not have a large effect on the delineation of the contributing area.

In addition to defining the external boundaries, locations were designated for known inflows and outflows from the aquifer. Recharge was assigned to the top model layer, simulating areal recharge from precipitation. Municipal-supply wells are assigned a value that correlates to the average discharge rate. Well Cu 969 was assigned an average daily pumping rate of 500,000 gal/d, and well Fr 823 was assigned an average daily pumping rate of 800,000 gal/d. The model was calibrated with no pumping from Cu 970, but after calibration was completed, the areal extent of the zone of contribution was simulated for well Cu 970 by assigning a pumping rate of 500,000 gal/d. Dykeman Spring was simulated using a constant flux (a pumping well node) of 3,000,000 gal/d to simulate a known rate of discharge from the aquifer. Big Spring was simulated using a constant head, which allows water to enter or leave a cell depending on the water level in the adjacent cells. The other outflow simulated was the stream network. Stream cells allow water to enter the stream when the hydraulic head in an adjacent cell is sufficient to move water through the streambed material (stream package, Prudic, 1989). The model also allows the water that enters a stream cell to transfer water to the adjacent downstream cell so that if the water table drops below the bottom of the stream, the stream can provide recharge to the aquifer, but the amount of water that the stream loses is limited to the amount of water that the stream has gained in upstream reaches.

The model was established by placing a uniformly spaced grid of 500 by 500 ft cells with 167 columns and 120 rows over the external model boundaries. Grid orientation allows the grid to be aligned with features that may be the source of anisotropy. Two hypothetical grid orientations were tested during model calibration, one aligned with the strike of the geologic formations and another aligned with the predominant orientation of fracture traces. The alignment with the predominant orientation of strike of the geologic formations yielded the best model fit during calibration; therefore, the grid was aligned so that the model rows were parallel to the predominant direction of the strike of the geologic formations, which was approximately N. 50° E. (fig. 7).

The model was designed such that the simplest model that could accurately describe the system would be used. Therefore, the minimum number of layers and zones was used. It was assumed there were two important layers for simulation of ground-water flow. Layer 1 represents colluvium, residuum, and highly weathered bedrock. Layer 2 represents fractured-bedrock aquifers. Both layers were simulated as confined aquifers.

The average elevation of each grid cell for the top of layer 1 was calculated from a 10-m (32.8 ft) digital elevation model (DEM) from the Shippensburg, Caledonia Park, Walnut Bottom, and Scotland, Pa., quadrangles. The bottom of layer 1, and the top of layer 2, was established by subtracting the colluvium thickness (fig. 9) from the top of layer 1. For those areas of South Mountain where colluvium is not present, a value of 180 ft was assigned for the thickness of layer 1, and for the areas of the valley where colluvium is not present, a value of 160 ft was assigned. In areas where colluvium is not present, layer 1 represents soil, unconsolidated materials, and highly fractured bedrock. The thickness of layer 2 was assumed to be uniform and was assigned a value of 200 ft.

The layer property flow (LPF) package (Harbaugh and others, 2000) was used for simulation of ground-water flow. This package allows aquifer characteristics to be represented vertically by layers and horizontal differences within each layer to be specified by zones. Initially, the zones for layer 1 were

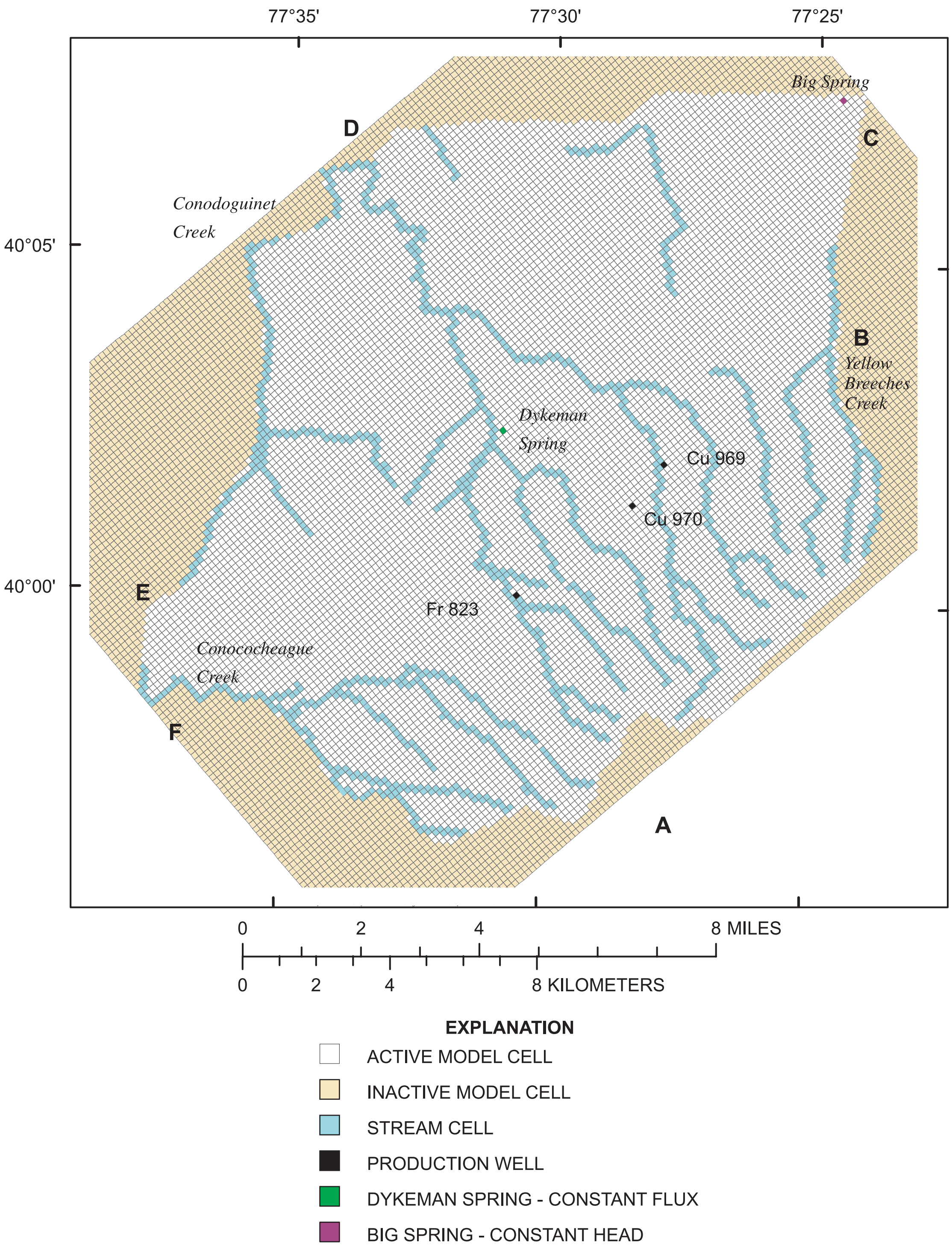


Figure 19. Model grid, stresses, and boundaries used to simulate ground-water flow near Shippensburg, Pa. Locations A-F are described in text.

assigned assuming three basic types of materials: 1) highly fractured carbonate-bedrock aquifer and residual carbonate soils, 2) highly fractured carbonate-bedrock aquifer and sandy colluvium, and 3) highly fractured metamorphic rocks and residual sandy soils. Zones for layer 2 initially were assigned to represent each geologic formation.

Early model runs indicated the characteristics of layer 1 as assigned by these geologic units did not appear to match the hydraulic characteristics indicated by the model. The Tomstown Formation appeared to have appreciably different characteristics from the other carbonate units; therefore, a zone was designated to represent this area and was referred to as highly fractured dolomite bedrock and sandy colluvium for layer 1. The area delineated as the Waynesboro Formation (fig. 7) had different characteristics on either side of the Shippensburg Fault during calibration. The hydraulic head in clusters of wells varied in a specific pattern, indicating areas with higher and lower transmissivity. To the south of the fault, the Waynesboro Formation appeared to have low transmissivity; however, to the north of the fault, the transmissivity of the formation appeared to be similar to that of the other carbonate formations. In addition, a hilly area just to the south of the Shippensburg Fault that is mapped as the Elbrook Formation appeared to have low transmissivity similar to the Waynesboro Formation south of the fault. Because the mapping of these units is limited by a lack of outcrops (they are covered by the colluvium), the geologic contacts are difficult to locate precisely. Because of this uncertainty and the hydrologic information provided by the modeling, both layers were divided into hydrostratigraphic units rather than specific formations. The result of this was designation of a unit called the Waynesboro Formation, ridge-forming units and sandy colluvium, which includes the Waynesboro Formation south of the Shippensburg Fault and the hill part of the Elbrook Formation. Those areas mapped as the Waynesboro Formation north of the Shippensburg Fault were included with the rest of the carbonate formations. A small part of the area simulated as highly fractured carbonate rocks and carbonate soils is actually underlain by a shale formation; however, because the characteristics of this layer were similar to the carbonate unit and it is in an area that was not near the municipal wells, it was not considered important enough to differentiate in layer 1. Descriptions of the zones used for modeling and the geologic formations they are based on are given in table 5. Thicknesses of newly defined zones were unchanged. The geographic extent of zones used in layer 1 differ slightly from the geographic extent of the geologic formations and are shown in figure 20.

The initial assumption for zones in layer 2 was to assign individual zones for each geologic formation. In early simulations, model sensitivity indicated many of the carbonate formations could be combined into a single zone representing carbonate bedrock. Zones also were assigned for areas underlain by shale, metamorphic bedrock, and dolomite. The unit previously described as the Waynesboro Formation and ridge-forming units was used for layer 2 as well. The term hydrostratigraphic units is used to describe the zones used for modeling. Although many of these units are based on geologic formations, the boundaries have been modified to group hydrologically similar units. The geographic extent of the zones used for layer 2 are shown on figure 21.

Model Calibration and Sensitivity

Model calibration consisted of comparing results of simulations to field measurements to obtain information about how well the model matches observed conditions. The field measurements available for calibration included water levels from wells and streamflow measurements made at springs and streams within the study area. The final model area was much larger than initially anticipated because of the difficulty in simulating ground-water flow out of the smaller area. The calibration points are largely within the initial model area (fig. 12). The results in the area outside the initial model area are less likely to simulate the actual conditions; however, this area is far enough away from the wells so that potential errors in simulation for that area are unlikely to have an appreciable effect on the outcome of the simulation for the wells.

A steady-state calibration simulated average conditions in the aquifer. Water levels from 43 wells were used for calibration (fig. 12). Streamflow measurements at 20 locations were used in early model calibration but not used during final model calibration (fig. 16). Initial calibration was conducted using manual calibration and the parameter-estimation process in MODFLOW 2000. Both streamflow and water levels in wells were used in the early part of the calibration process.

MODFLOW 2000 includes a feature called parameter estimation. A parameter, as used with respect to this program, is the assignment of a name to a variable used by the ground-water-flow equation at one or more model cells. A parameter is assigned for a range of cells that are assumed to have similar characteristics. For example, a parameter could be assigned to the carbonate-bedrock aquifer in layer 1. The parameter can be used to define the variables used in the model or it can be used for processes such as model sensitivity (how much does a change in the parameter value affect the model fit) and parameter estimation (when a parameter is iteratively adjusted to improve model fit). Parameters were assigned for layer-property flow characteristics and recharge. Eleven parameters were defined for layer-property flow, including 10 for horizontal hydraulic conductivity and 1 for horizontal anisotropy. Vertical anisotropy for hydraulic conductivity was introduced as a parameter but was found to be insensitive and was removed early in the modeling process, and vertical anisotropy was set equal to one thereafter. Therefore, the value of horizontal hydraulic conductivity along rows (K_x) for a given unit is the same value used for vertical hydraulic conductivity (K_v), with the exception of the highly fractured carbonate bedrock and residual carbonate soils in layer 1, which was assigned a value of 15 ft/d. Five parameters were assigned to represent hydraulic conductivity for zones in layer 1 (fig. 20), and five parameters were assigned to represent hydraulic conductivity for zones

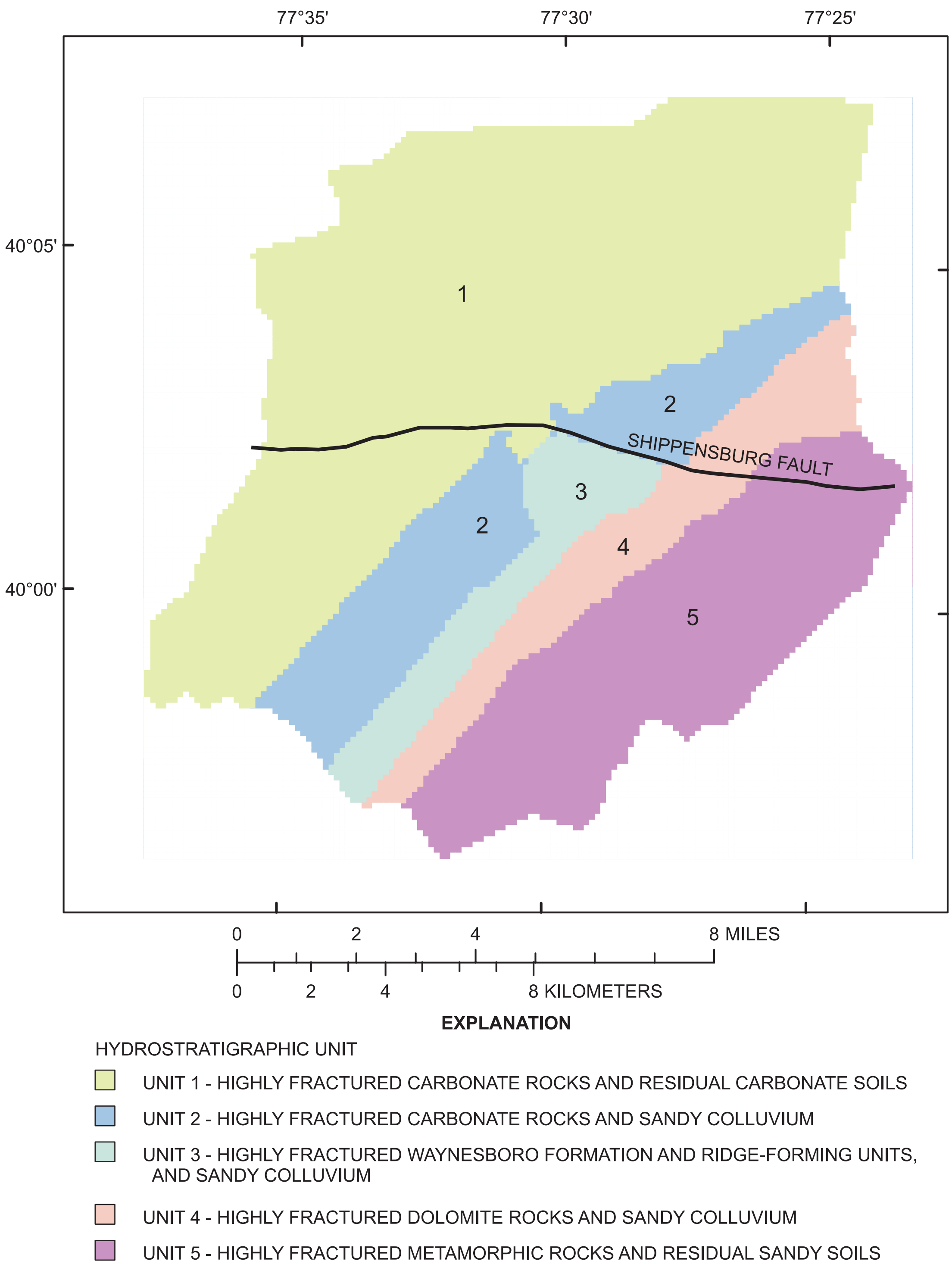


Figure 20. Hydrostratigraphic units used to simulate ground-water flow in model layer 1 near Shippensburg, Pa.

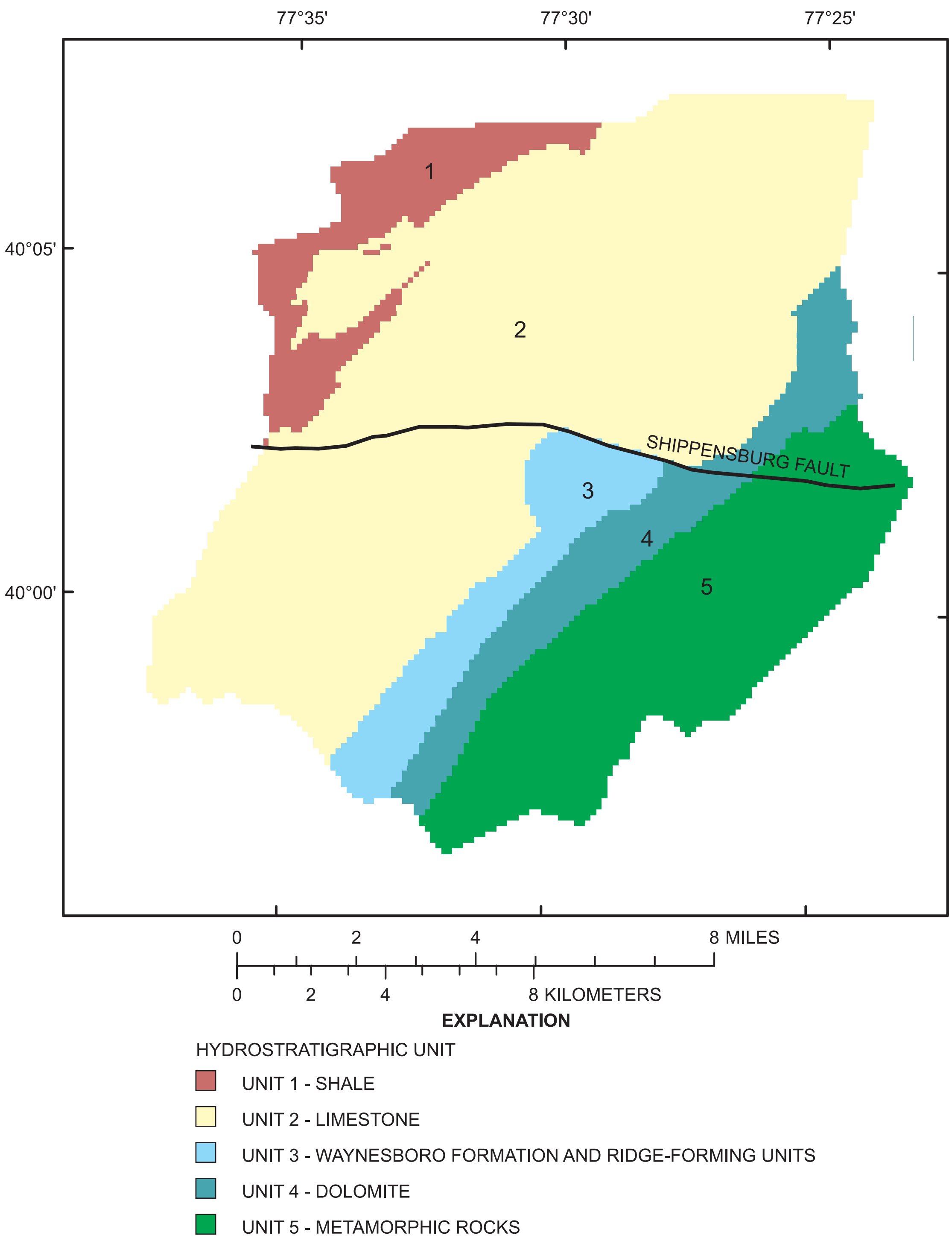


Figure 21. Hydrostratigraphic units used to simulate ground-water flow in model layer 2 near Shippensburg, Pa.

Table 5. Description of hydrostratigraphic units used for simulation of ground-water flow near Shippensburg, Pa.

Hydrostratigraphic unit description	Model layer	Zone number	Geologic formations as mapped by Berg and others (1980)
Highly fractured carbonate rocks and residual carbonate (or shale) soils	1	1	Martinsburg; Chambersburg; St. Paul Group; Pinesburg Station; Rockdale Run; Stonehenge; Shady Grove, Zullinger, (where no colluvium cover present); and Elbrook (where no colluvium cover present)
Highly fractured carbonate rocks and sandy colluvium	1	2	Zullinger and Elbrook (where covered by colluvium); Waynesboro, north of Shippensburg Fault
Highly fractured Waynesboro Formation and ridge-forming units and sandy colluvium	1	3	Waynesboro, south of Shippensburg Fault; Elbrook, ridge-forming area
Highly fractured dolomite rocks and sandy colluvium	1	4	Tomstown
Highly fractured metamorphic rocks and residual sandy soils	1	5	Antietam; Harpers; Montalto Member of Harpers Formation; Weverton and Loudon Formations, undivided
Shale	2	1	Martinsburg
Limestone	2	2	Chambersburg; St. Paul Group; Pinesburg Station; Rockdale Run; Stonehenge; Shady Grove; Zullinger; Elbrook; Waynesboro (north of Shippensburg Fault)
Waynesboro and ridge-forming units	2	3	Waynesboro (south of Shippensburg Fault); Elbrook, ridge-forming area
Dolomite	2	4	Tomstown
Metamorphic rocks	2	5	Antietam; Harpers; Montalto Member of Harpers Formation; Weverton and Loudon Formations, undivided

in layer 2 (fig. 21). One parameter was set to represent horizontal anisotropy within the entire model area in both layer 1 and layer 2. Recharge parameters were assigned for three areas within the model. These three areas were the urban area around Shippensburg, the steep terrain on South Mountain, and the relatively flat valley floor.

Because there were more parameters (14) than could be estimated simultaneously, parameters were first evaluated for sensitivity, which indicates whether or not enough information is available to justify modifying the values of the parameters. Those parameters that are insensitive are set to a commonly accepted value (typically taken from previous studies) and not modified further. The values used for parameters not estimated are given in table 6. The most sensitive parameters are used in the parameter-estimation process, in which parameter values are altered in successive model runs in an attempt to minimize the objective function. The objective function is the sum of squared-weighted residuals obtained by comparing water levels simulated by the model to water levels measured in the field. The theory behind calibration is that a model that accurately matches measured data, such as streamflow and water levels, will more accurately portray processes that cannot be directly measured, such as direction of ground-water flow.

During calibration, several important issues were explored. One issue was the validity of assigning a unique value of horizontal hydraulic conductivity to each geologic formation. The characteristics of limestone units, the variability in the colluvium, and the similarity of the Elbrook and western part of the Waynesboro Formations led to the use of hydrostratigraphic units (table 5) rather than strictly geologic units as previously discussed. Also, the potential effect of the Shippensburg Fault on ground-water flow was evaluated. The fault was simulated both as a highly transmissive zone and a barrier, but neither simulation resulted in improvement to the model.

Another issue explored during calibration was determining values of streambed hydraulic conductivity to correctly simulate the gaining and losing reaches of the streams. Using the parameter-estimation process to estimate streambed hydraulic conductivity and calibrate the model to streamflow led to simulations in which streams either lost all of their flow or simulations where no streamflow was lost. In this environment, streambed conductivity actually changes on a scale much smaller than the model can simulate. Variation in clay percentage in the streambed material is typically the controlling variable. Because parameter estimation could not accurately simulate streamflow, values for streambed hydraulic conductivity were not estimated by the model; rather, the value for each stream reach was set individually so that gaining and losing reaches of the stream generally were accurate. Stream hydraulic conductivity was set segment by segment from the top of each reach downstream until streamflow in each reach had approximately the correct flow and gain or loss. In the final simulations, streams gained and lost flow in the correct locations, but the volume of gain or loss was less than measured during the

Table 6. Values of hydraulic conductivity assigned for parameters not estimated (fixed) in simulation of ground-water flow near Shippensburg, Pa.

Hydrostratigraphic unit	Model layer	Zone number	[1]Final value of hydraulic conductivity (feet per day)
Highly fractured carbonate bedrock and residual carbonate soils	1	1	[2]0.1
Highly fractured dolomite bedrock and sandy colluvium	1	4	98.4
Highly fractured metamorphic rocks and residual sandy soils	1	5	.98
Shale	2	1	8.2
Limestone	2	2	37.1
Waynesboro and ridge-forming unit	2	3	.02
Dolomite	2	4	94.2
Metamorphic rocks	2	5	.06

[1]Value given is for both vertical and horizontal hydraulic conductivity.
[2]Parameter values were adjusted from initial values early in calibration process but not modified during final model calibration because of low sensitivity.

seepage run. This result is not unexpected because streamflow during the time of the seepage run was slightly above average base flow. Therefore, although streambed hydraulic conductivity is a manually calibrated parameter, it is not included in the calculations of model error. Streamflow was not used for calibration in final model runs. Values assigned for streambed hydraulic conductivity ranged from 0.3 to 1,500 ft/d. The majority of the streams were given values in the range of 1.5 to 3 ft/d. The thickness of the streambed was set to 3.2 ft for all stream segments, the widths of the stream reaches varied from 3.2 ft for headwater streams to 9 ft for larger streams and were based on measurements from the seepage run.

Recharge parameters were assigned for the urban area around Shippensburg, the steep terrain on South Mountain, and the relatively flat valley floor. Although the model was set up to allow different values for recharge in the valley and South Mountain, the final model used a value of 14.8 in/yr for recharge in both areas (the median annual base flow for Yellow Breeches Creek). Using the assumption that runoff is greater in the urban parts of the model area, a value of 9.8 in/yr (the median value for Conodoguinet Creek) was assigned for recharge in that area. These values were assigned early in model calibration to obtain reasonable water levels in the urban area around Shippensburg and reasonable volumes of stream base flow in tributaries flowing from South Mountain.

Final calibration of the model was conducted after the model parameters were close to simulating observed water levels. This final calibration consisted of selecting the most sensitive parameters. Sensitive parameters are those that cause a relatively large change in the model solution for an incremental change in the parameter value. Composite sensitivity is calculated as follows: the sensitivities are scaled by multiplying them by the product of the parameter value and the square root of the weight of the observation to obtain dimensionless values. The scaled sensitivities for each parameter are then squared, and the sum of these values is divided by the number of observations. The composite-scaled sensitivities equal the square root of these values. Composite-scaled sensitivity is indicated in figure 22. The parameters with the largest composite-scaled sensitivity also are the parameters most likely to be most accurately estimated by the parameter-estimation process.

The parameters of recharge in the valley, horizontal anisotropy, hydraulic conductivity of limestone bedrock in layer 2, hydraulic conductivity of the Waynesboro and ridge-forming hydrostratigraphic unit and sandy colluvium in layer 1, and hydraulic conductivity of highly fractured carbonate bedrock and sandy colluvium in layer 1 had the highest composite-scaled sensitivities of all parameters in the model. In initial model calibration, all five of these parameters were adjusted; however, sufficient data were not available for parameter estimation to converge using all five parameters. The value of recharge was not estimated because it was established on the basis of base-flow data and allowing recharge to change during calibration resulted in unreasonable water budgets. The value for horizontal hydraulic conductivity of limestone bedrock in layer 2 was highly correlated with horizontal anisotropy, and therefore, both could not be estimated independently. For final model calibration, hydraulic conductivity of the highly fractured carbonate rocks and sandy colluvium in layer 1, hydraulic conductivity of the Waynesboro and ridge-forming hydrostratigraphic unit and sandy colluvium in layer 1, and horizontal anisotropy were selected for estimation.

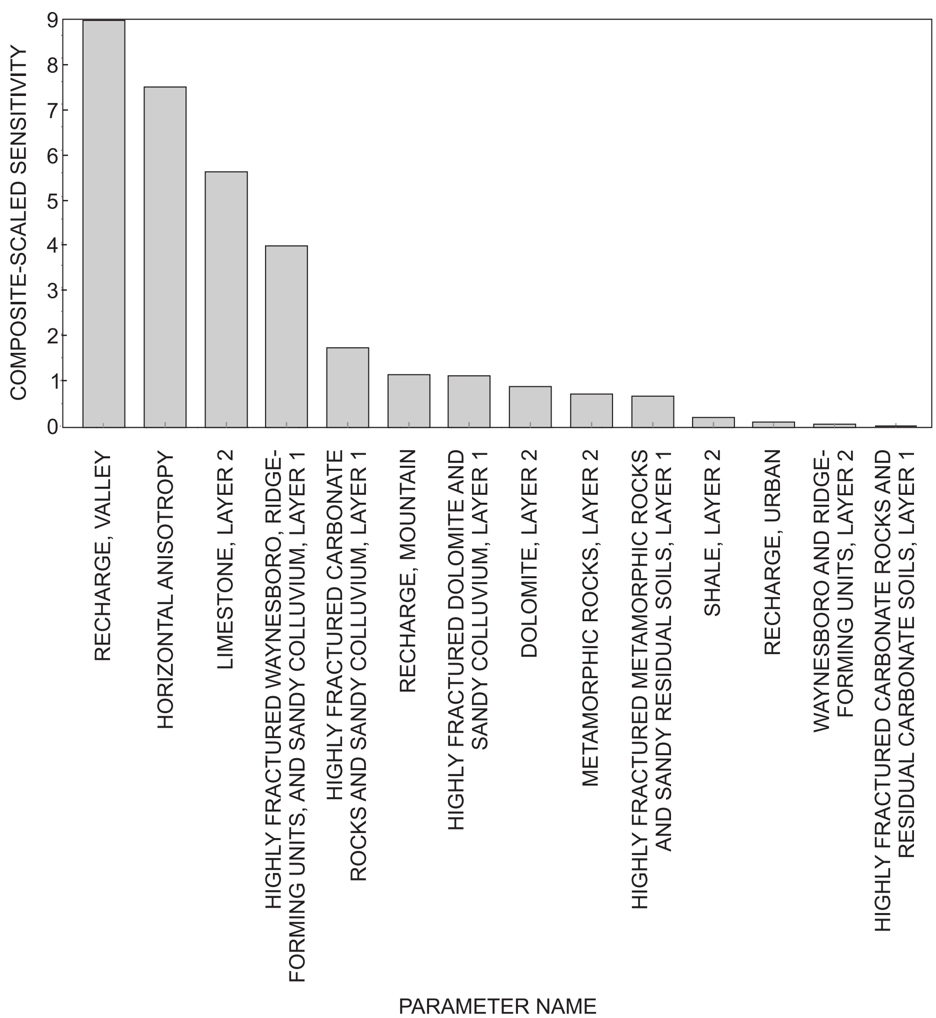


Figure 22. Model sensitivity for parameters used to simulate ground-water flow near Shippensburg, Pa.

Model Results and Numerical Uncertainty

The model calibration resulted in a sum of squared, weighted residuals of 1,036 and a root-mean-square error of 9.8 ft. The largest positive residual was 23.5 ft, and the largest negative residual was -17.22 ft (fig. 23). The water budget had a discrepancy of 0.15 percent. Correlation between parameters was not significant at a 95-percent confidence interval. The final values from parameter estimation are given on table 7. The value for the horizontal hydraulic conductivity of highly fractured Waynesboro Formation and ridge-forming units and sandy colluvium is very close to that determined by previous studies. The value for horizontal hydraulic conductivity of highly fractured carbonate bedrock and sandy colluvium is higher than that given in previous studies in this area, but it is within the range of general values given for sand and gravel aquifers (Todd, 1980). The value of 1.73 for horizontal anisotropy indicates horizontal hydraulic conductivity along columns is 1.73 times greater than horizontal hydraulic conductivity along rows. This finding is somewhat unique in that it indicates greater permeability across the strike of the geologic formations than along strike. The reason for this is undetermined but may represent greater development of solution fractures across strike (along the gradient) than along strike. The parameters estimated and shown on table 7 include estimates for confidence intervals for the optimal parameter value. The confidence interval for horizontal hydraulic conductivity of highly fractured carbonate bedrock and sandy colluvium is quite large, indicating a lower level of confidence in that value. These confidence intervals are based on linear theory and are only accurate if the model is linear. One measure of model linearity is the correlation between weighted residuals and normal order statistics, which for this model was 0.974. This value is greater than the threshold of 0.945, indicating the model is linear. Another measure of linearity is Beale's measure (Hill and others, 2000). The value of Beale's measure for this model is 0.24, which is lower than the threshold of 0.35; values above the threshold would be considered non-linear. Therefore, the confidence intervals based on linear theory are considered to be accurate.

The final simulation of ground-water flow in the model area produced a simulated water-level surface as indicated in figure 24. The simulated water-level surface does not match those generated by Becher and Root (1981), who generated contours on the basis of water-level measurements; however, a model by Chichester (1996) produced results similar to that of the present study (fig. 11). The water-level contours are not parallel to the topographic contours. This is partially caused by the juxtaposition of the high transmissivity dolomite and colluvium with the resistant Waynesboro and ridge-forming hydrostratigraphic units, which causes the ground-water gradient to be oriented in a direction that is not parallel to the topographic gradient. The contours in figure 24 illustrate details of the simulated water-level surface that are not evident on figure 11 because of the higher resolution of the current model.

The RESAN program (Hill and others, 2000) was used to evaluate the effect of individual water levels on the model results. Cook's D and DFBeta statistics indicated water levels from wells Cu 976, Cu 612, Cu 680, Cu 971, and Cu 959 had

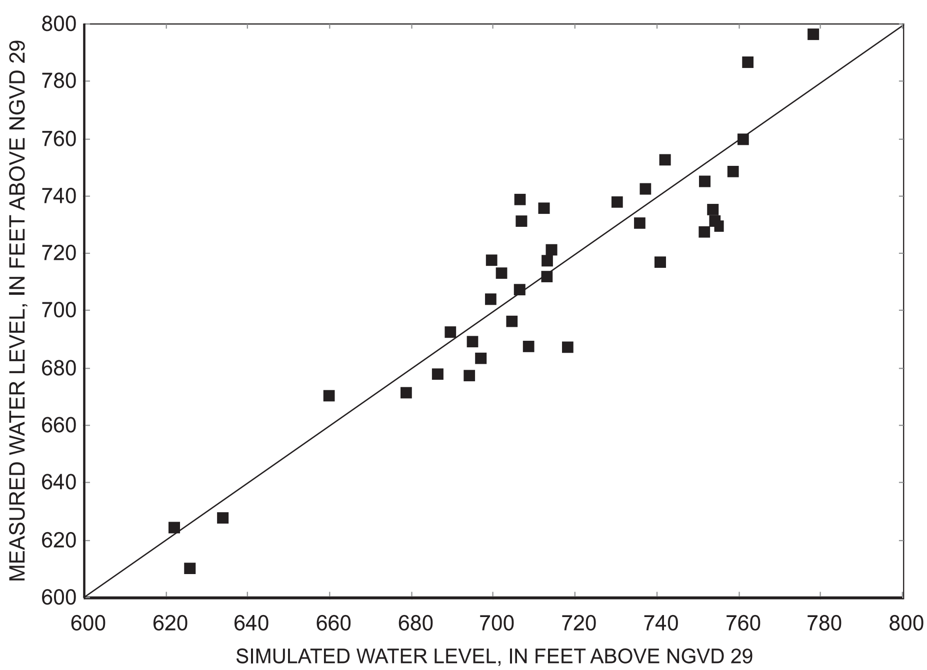


Figure 23. Relation between measured and simulated water levels used to calibrate the ground-water-flow model near Shippensburg, Pa.

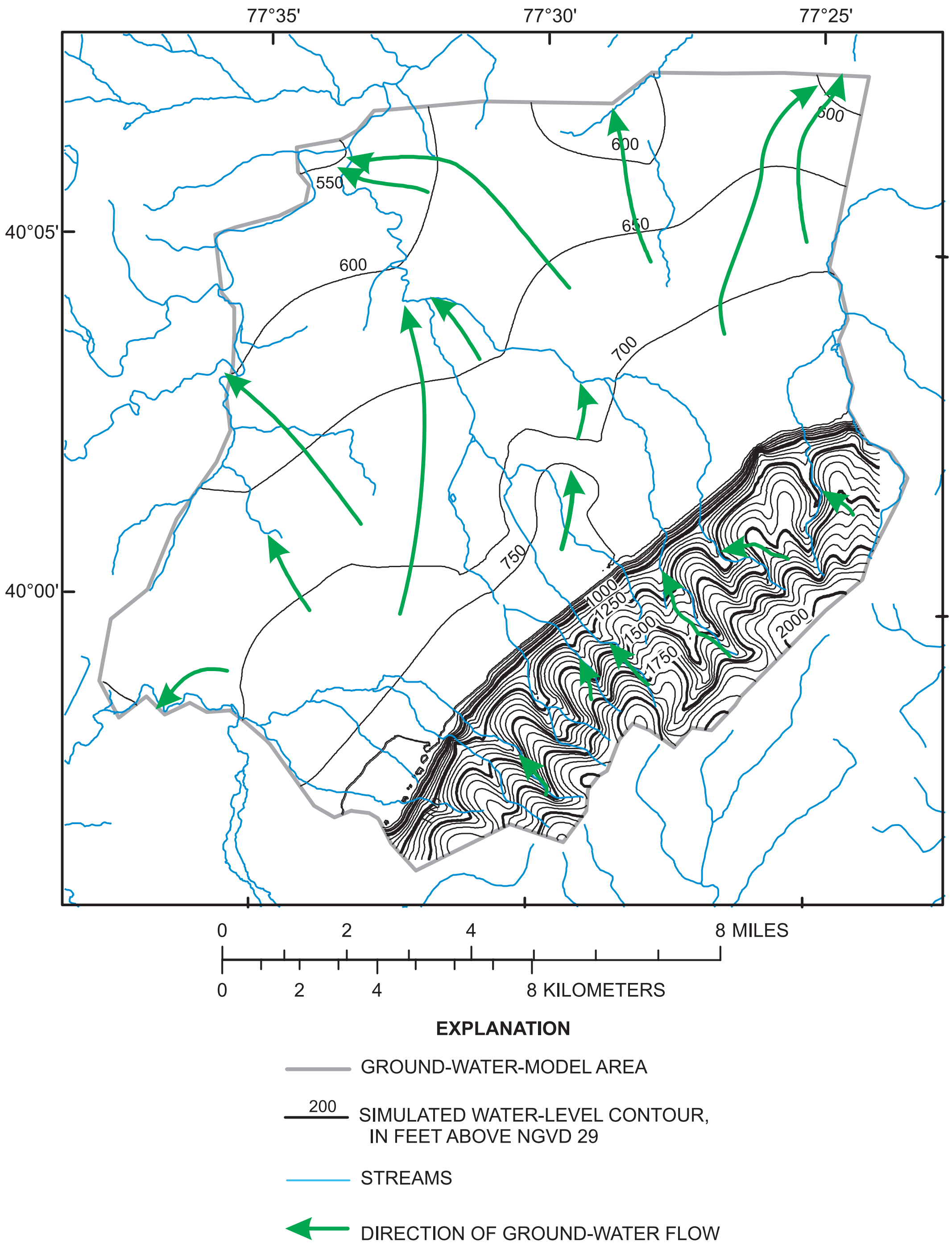


Figure 24. Simulated water-level contours and directions of ground-water flow near Shippensburg, Pa.

Table 7. Final values for aquifer properties determined by parameter estimation near Shippensburg, Pa.

[ft/d, feet per day]

	Layer	Hydro-stratigraphic Unit	Final value	95-percent confidence interval values
Horizontal hydraulic conductivity of highly fractured carbonate bedrock and sandy colluvium	1	2 (fig. 20)	233. ft/d	52.5 - 1,220 ft/d
Horizontal hydraulic conductivity of highly fractured Waynesboro Formation and ridge-forming units and sandy colluvium	1	3 (fig. 20)	3.97 ft/d	2.46 - 6.23 ft/d
Horizontal anisotropy	1 and 2	All	1.73	1.1 - 2.7

the most influence on the estimated parameter values. The observations where Cook's D had a value greater than 0.11 (4/number of observations) are considered to be significant outliers. Five wells had water levels that were significant outliers, and the observations were believed to be hydrologic anomalies (such as a perched water table or possibly affected by pumping). These observations were not included during calibration (table 4). The remaining observations with high values of Cook's D were evaluated for accuracy and spatial patterns, but no discernible spatial pattern was observed.

The water budget determined by simulation of ground-water flow is given in table 8. The water accounted for as input to the model as streamflow is not a net gain because the water that enters the aquifer from streams is the same water that was discharged from the aquifer upstream, which ultimately came from recharge. Therefore, for the total budget calculations, recharge is considered to be the total input. A key consideration in examining the water budget is comparing the pumping from the production wells to the recharge in the entire area upgradient of the wells. The recharge to that part of the aquifer is about 13 Mgal/d, as compared to 1.3 Mgal/d being pumped by production wells in 2004, and about 2 Mgal/d that will be pumped when well Cu 970 is brought on line. The volume of water that makes up the difference between recharge and the pumped volume is not necessarily available for development by production wells in the future. The recharge that is not pumped out by wells sustains stream base flow, and at some point, increased pumping would begin to cause streams to go dry. Determining the amount of pumping that this area could sustain is beyond the scope of this report; however, long-term records of streamflow from tributaries draining this area would be helpful in determining the balance between pumping and streamflow.

The simulated locations where water flows in and out of streams are illustrated in figure 25. The model shows the majority of water lost to the aquifer is in areas where the stream is crossing over the colluvial material. Some water is lost to the aquifer upstream of the area underlain by colluvium and in a few locations in the valley where the water table drops below the streambed. The gaining and losing reaches simulated in the model (fig. 25) compare favorably with the gaining and losing reaches measured during the seepage run (fig. 16). The flow from the constant-head node simulating Big Spring has a flow of 6.2 ft^3/s, which is less than half of the discharge reported by Becher and Root (1981) for that location; however, the construction of this model does not necessarily include the entire contributing area for Big Spring.

Zone of Contribution and Spatial Uncertainty

The areal extent of the zone of contribution to the production wells was determined by using MODPATH (Pollock, 1994) as implemented in MODFLOW-2000. This program simulates movement of water between model cells, and the particle-tracking component of this can be used to simulate the movement of a water particle from its recharge point to the location where it reaches a stream, pumping well, or constant head cell. The output files store all starting and finishing locations so the recharge location of any particle that ends up in the pumping well cell can be determined and plotted. The areal extent of the zone of contribution to the production wells (fig. 26) shows a similar pattern for wells Cu 969, Cu 970, and Fr 823. The points on figure 26 are locations on the land surface where a particle of water recharged in that cell would end up in the pumping well and indicate the center of a 500 by 500 ft model cell (recharge area). The outlined areas include any recharge location plus any flow path of water flowing toward the well, projected to the land surface (areal extent of the zone of contribution). The patterns all follow the strike of the geologic formations, specifically the Tomstown Formation (the dolomite unit), then turn abruptly up the topographic gradient into the Antietam and Harpers Formations (metamorphic units). The reason is that the relatively high transmissivity of the Tomstown Formation allows water to flow easily towards the well, whereas the relatively low transmissivity of the Antietam and Harpers Formations does not allow water to flow directly toward the pumping wells. The water-bearing properties of these ridge-forming units are not sufficient to store or transmit the large volumes of water pumped from the wells.

The areal extent of the zone of contribution illustrated on figure 26 is equivalent to the PaDEP's 'Zone II' for wellhead-protection plans (Pennsylvania Department of Environmental Protection, 2000). The half-mile radius, which is the arbitrary

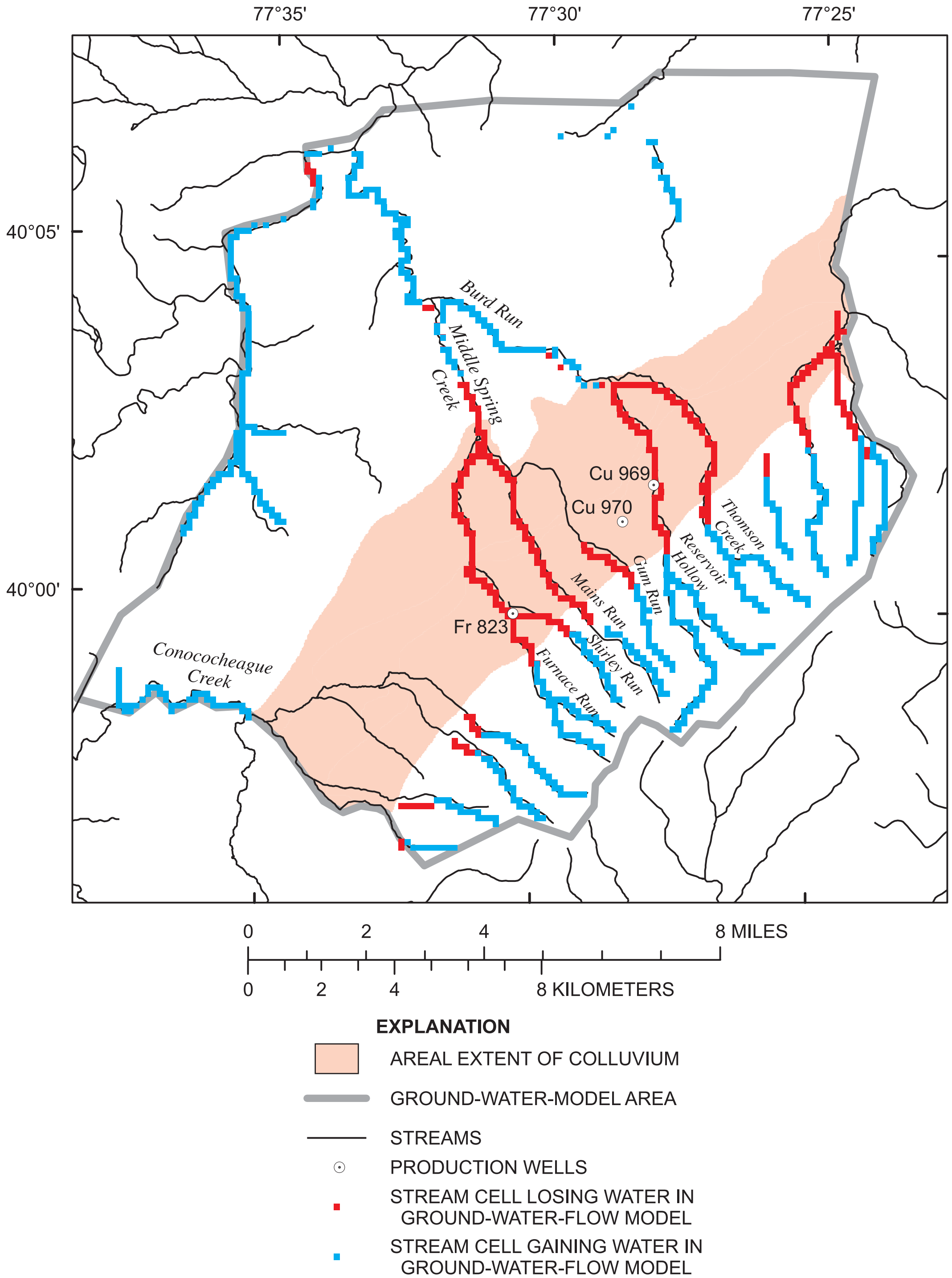


Figure 25. Simulated gaining and losing reaches of streams near Shippensburg, Pa.

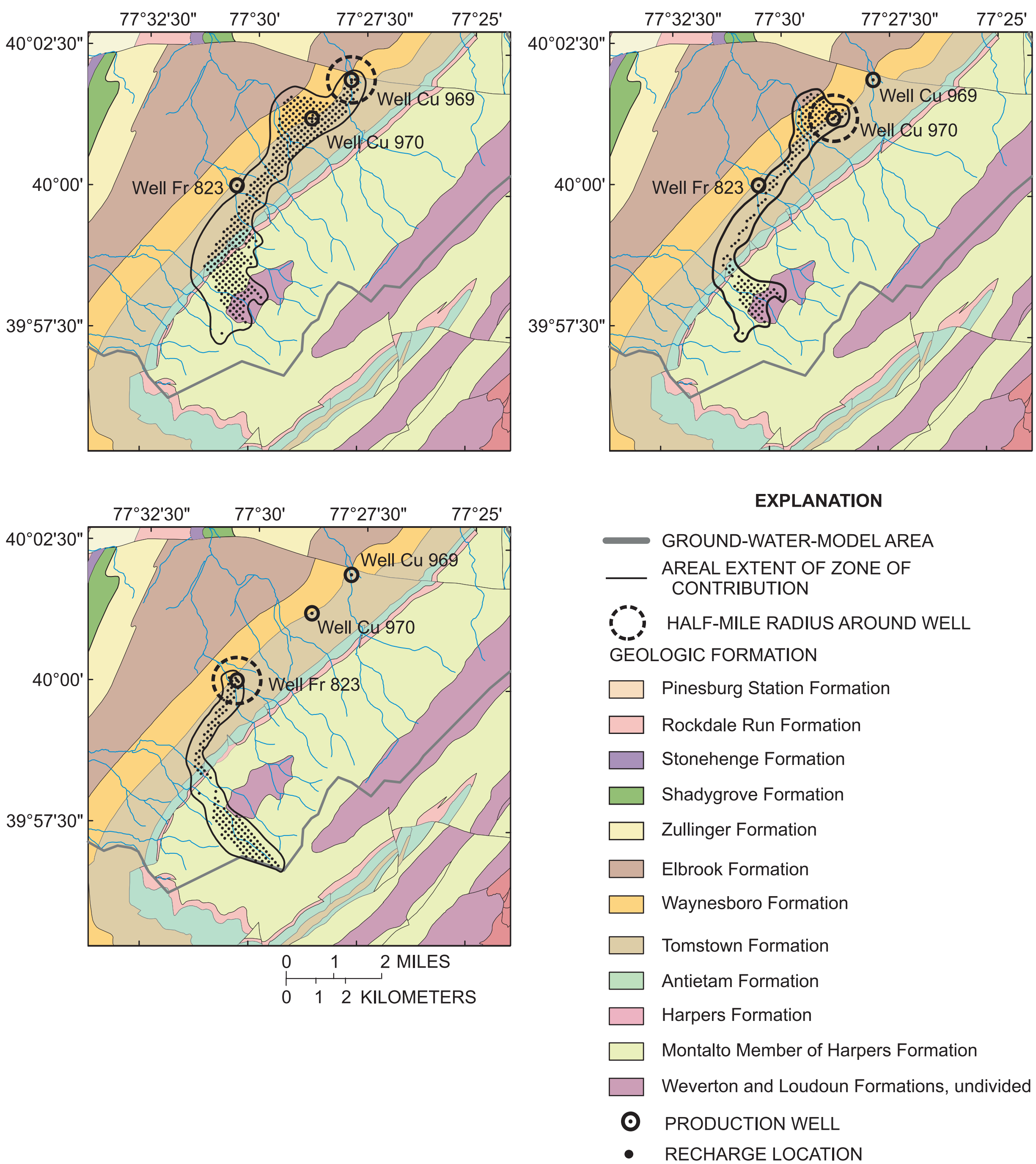


Figure 26. Geologic units, stream network, and contributing area to production wells near Shippensburg, Pa.

Table 8. Water budget from simulation of ground-water flow near Shippensburg, Pa.

[Mgal/d, million gallons per day; ft^3/s, cubic feet per second; in., inches]

	Volume (Mgal/d)	Volume (ft^3/s)	Volume (in.)
Source:		Inflow	
Recharge	75.0	116	14.1
Stream to aquifer	8.7	13.5	1.6
TOTAL INFLOW[1]	75.0	116	14.1
		Outflow	
Big Spring	4.0	6.2	.8
Dykeman Spring	3.1	4.8	.6
Wells (Cu 769 and Fr 823 only)	1.3	2.0	.2
Aquifer to stream	66.5	103	12.5
TOTAL OUTFLOW	74.9	116	14.1

[1]Inflow totals do not count ‘stream to aquifer’ amounts because that water is originally from recharge. Total outflows are adjusted downward by the volume of ‘stream to aquifer’ inflows because that volume of water would be accounted for twice.

area Zone II if a rigorous delineation is not done, also is illustrated on figure 26 for comparison purposes. Well Fr 823, which has the highest pumping rate, has the smallest recharge area delineated. A mass-balance calculation can be used to determine the amount of water recharged in this area. That surface area of the recharge locations, multiplied by the recharge in that area, yields a value of about 425,000 gal/d. Well Fr 823 pumps 800,000 gal/d. This simulation indicates induced recharge from the streams to the aquifer is required to supply sufficient water (375,000 gal/d) to well Fr 823. Wells Cu 969 and Cu 970 have sufficiently large recharge areas to produce the water pumped from these wells. This does not preclude induced infiltration from streams in those situations; it just means it is not required in this model simulation. Although the simulation did not show significant volumes of induced streamflow recharge going to wells Cu 969 and Cu 970, the seepage run documented that water was being lost in the areal extent of the zone of contribution to those two wells. Therefore, it is assumed some induced recharge from the stream to the aquifer enters both of those wells also.

Alternate scenarios were simulated using very high values for streambed hydraulic conductivity in stream segments near the wells, and in these scenarios, almost all the water moving to the production wells came from a few cells immediately under the streams. The results of this simulation produced a poor fit to observed water levels and streamflows, making this an unlikely scenario. However, this represents an example where the source of the water is very different from that situation where the majority of the water is produced from areal recharge to the aquifer. Field measurements of streamflow in the vicinity of the wells indicated all the streams lost some water in that area. These field measurements, in conjunction with model results that showed a similar result (greater for well Fr 823 but likely for all three wells), indicate a high probability the streams flowing from South Mountain are a source of at least some of the water produced by the pumping wells. This loss of water from stream to aquifer in or near the areal extent of the zone of contribution indicates the total contributing area may be larger than that area delineated as the areal extent of the zone of contribution from ground water. This is discussed in Risser and Barton (1995) and is called the contributing area from watershed of losing streams. This area is equivalent to the PaDEP definition of Zone III for wellhead-protection purposes (Pennsylvania Department of Environmental Protection, 2000). The contributing area from the watershed of losing streams is indicated on figure 27. Although no streamflow measurements were made for streams in the headwaters of Conococheague Creek to document losing reaches, this was documented for all the other streams flowing from South Mountain and, therefore, is assumed to be the case for those streams also. It is likely that, as hydrologic conditions change, the sources of water vary as well. In situations where more water comes from induced recharge from streams, the areal extent of the zone of contribution from ground water is smaller, and in situations where less water comes from induced recharge from streams, the areal extent of the zone of contribution from ground water is larger.

One of the implications of the stream-aquifer interaction is that traveltimes of contaminants may be short-circuited by water traveling that route. Although traveltimes were not determined for this study, a previous study of ground-water residence time indicated that Dykeman Spring had an average residence time of 9 years (Lindsey and others, 2003). The study by Lindsey and others was for the spring and not the production wells; however, a contaminant introduced into the aquifer in the upper part of the recharge area may be expected to have several years traveltime before entering the well. That same contami-

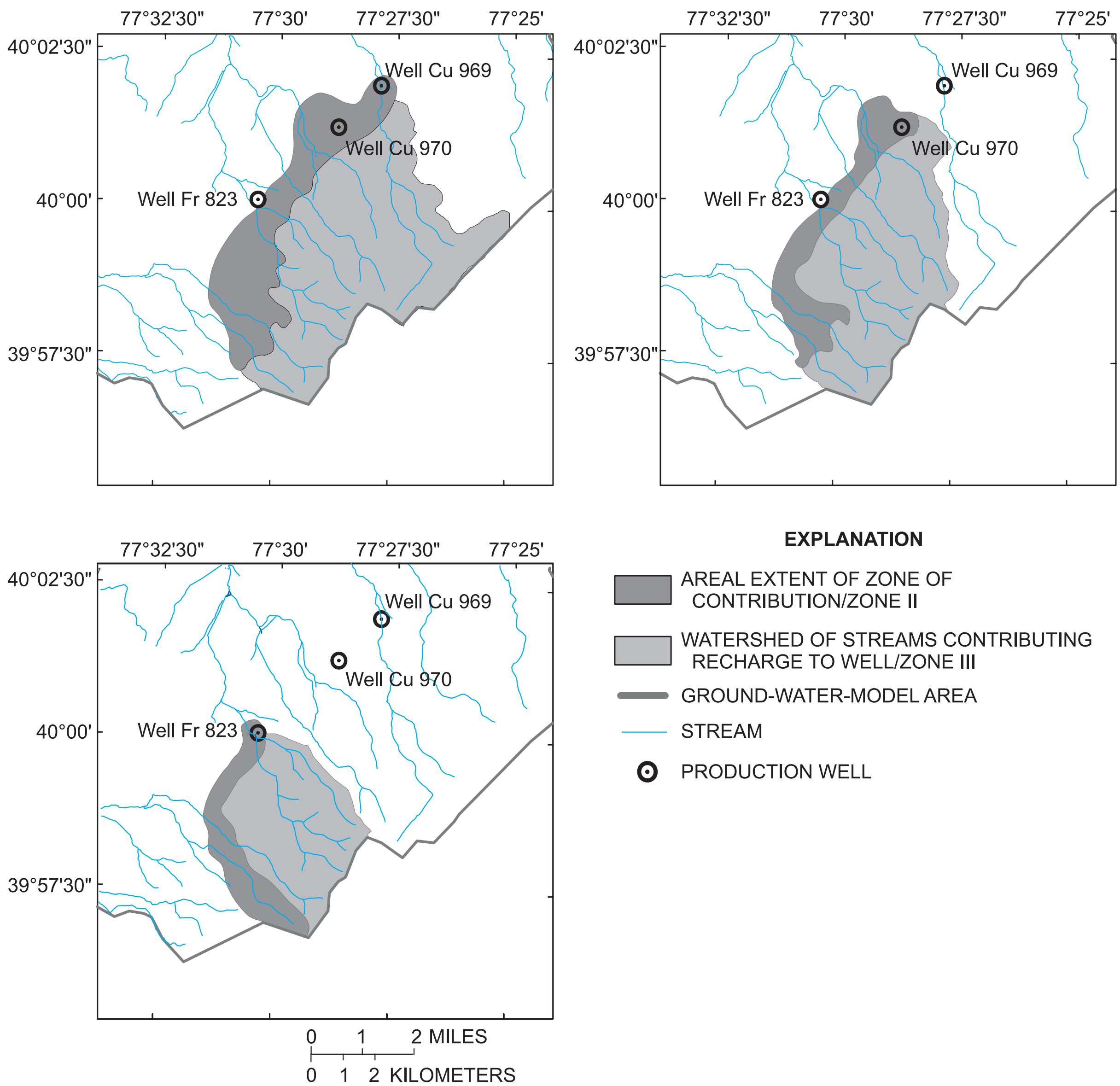


Figure 27. Contributing area of watersheds of streams with losing reaches that may provide recharge to production wells near Shippensburg, Pa.

nant flowing in a stream may move rapidly to a stream reach near the production wells where induced recharge would bring the contaminant into the aquifer very close to the well and into the well in a much shorter time period.

The model was calibrated for steady-state conditions using long-term average values for recharge. The calibrated model also was used to simulate the areal extent of the zone of contribution under drought conditions. The recharge value used to simulate drought conditions was 10 in/yr, which is the 10^{th} percentile of annual base flow for Yellow Breeches Creek. The value for recharge in the urban area was 6.4 in/yr which is the 10th percentile of annual base flow for Conodoguinet Creek. Using these reduced recharge values, the areal extent of the zone of contribution to the production wells was larger than when simulated using the long-term average recharge values. The shape of the area also changed, typically extending further upgradient toward the top of South Mountain. This extended area, however, was entirely within the area of watersheds of streams contributing recharge to the wells as designated in figure 27. Therefore, even under drought conditions, the areal extent of the zone of contribution to the production wells will be in areas already designated as being within the original areal extent of the zone of contribution, or the area of watershed of streams contributing recharge to the well.

The areal extent of the zone of contribution to wells as shown in figure 26 is a result of a deterministic model and is the best representation of the shape of that area on the basis of available information. However, delineation of the areal extent of the zone of contribution has a degree of uncertainty. This uncertainty has been expressed numerically in the section discussing model results, but a probabilistic approach can also be used to express the uncertainty spatially. This is done by use of a Monte Carlo analysis, a probabilistic method in which the model is run under multiple likely scenarios and the recharge locations are compiled from each scenario. The difference between a deterministic model and a probabilistic model is that a deterministic model is trying to answer the question "What specific areas on the ground provide recharge to the production well?" (one "best" answer) and the probabilistic model is answering the question "What is the probability that a given point on the ground is providing recharge to the production well?" (a range of likely answers). The Monte-Carlo method is described in detail in Starn and others (2000).

To provide the input data sets for the Monte-Carlo analysis that represent reasonable possible alternate solutions, the output of the deterministic model is used to constrain those data sets. The parameter-estimation package within MODFLOW 2000 preserves the variance-covariance matrix. The values in this matrix are derived from the regression and include the uncertainty of the parameter values and the correlation among parameter values. The set of random normal variables that are created for the alternate model realizations are created using the following equation from Starn and others (2000):

$$b = z\sigma + \mu, \qquad (4)$$

where

b is a vector of the model parameter values,

z is a vector of normally distributed random numbers,

σ is the square root of the variance/covariance matrix, and

μ is a vector of optimal parameter values.

The result of this is a set of random variables for each parameter that has a normal distribution around the optimal value. The distribution is based on the confidence interval of the parameter with each parameter set taking into account the correlation among the parameters. The three parameters estimated for the deterministic model were hydraulic conductivity in highly fractured carbonate rocks and sandy colluvium in layer 1, hydraulic conductivity of the Waynesboro and ridge-forming hydrostratigraphic unit in layer 1, and horizontal anisotropy. The distribution of the values used for each of these parameters is illustrated in figure 28. Each of these parameters was log-transformed in the model; therefore, the distribution of the input parameters is log-normal. The alternate data sets used as input for the Monte-Carlo analysis represent a range of likely values for those parameters, on the basis of the uncertainty of each parameter. This Monte-Carlo analysis only simulates scenarios for three of the parameters used in this model, and potential for variability exists in many of the other parameters. However, because the three parameters used in the Monte-Carlo analysis were among the most sensitive parameters in the model, it is likely this analysis will illustrate more of the potential variability than if other parameters were used for this analysis.

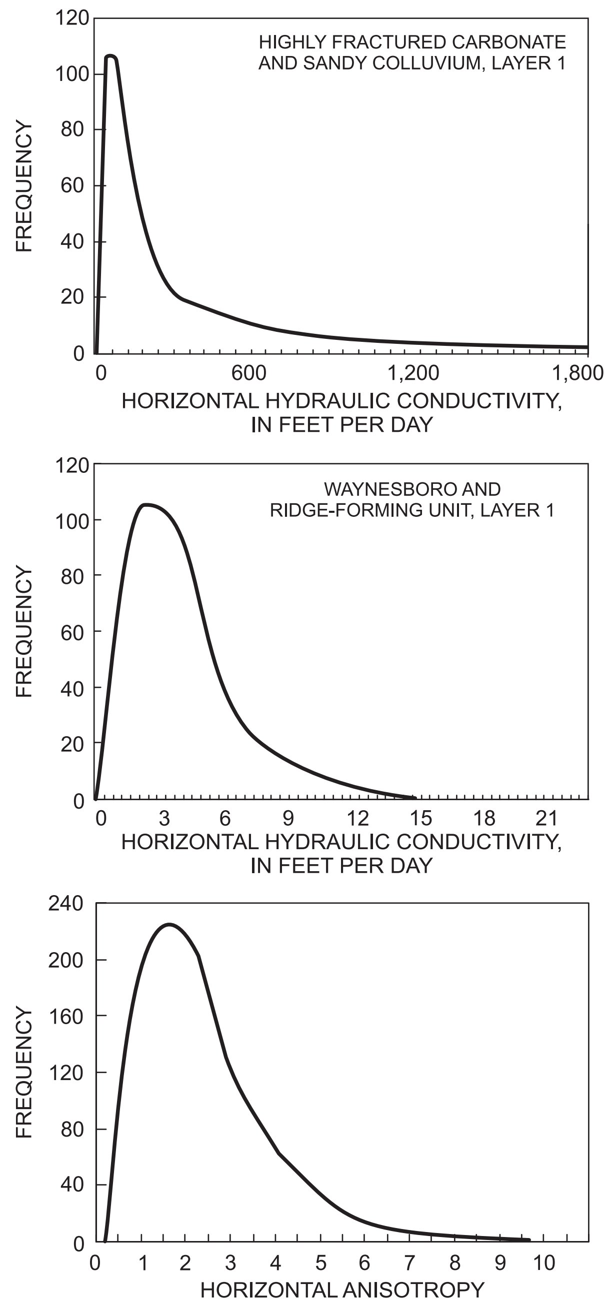


Figure 28. Distribution of parameters used to simulate probabilistic contributing areas to production wells near Shippensburg, Pa.

The Monte-Carlo analysis was run with 1,000 sets of randomly generated parameters. Because this method determines recharge locations and not pathlines, the results are given for recharge locations only rather than the areal extent of the zone of contribution. The particle tracking step is only conducted for those model runs where the model converges and the water budget is within an acceptable limit (2 percent used for this scenario). Of the 1,000 model runs, 980 simulations met that criteria and had particle-tracking analysis done to determine the recharge locations for that set of parameters. Dividing the number of times that a given cell is providing recharge to the production well by the total number of model runs allows the results to be expressed as a percentage for each cell. These percentages can then be contoured to show patterns in the results of these alternate models. The results of these 980 simulations were compiled, and the results are illustrated in figures 29, 31, and 30. The probabilistic method is more useful in illustrating the strengths and weaknesses of the model than the shape of the recharge area. That is, those areas where a high percentage of the Monte-Carlo runs coincide with the deterministic results indicate a high probability that the model is correct for that area, whereas those areas where a low percentage of the Monte-Carlo runs coincide with the deterministic results indicate the model is less likely to be accurate in that area. All three of these figures show a similar pattern. The areas with the highest percentage from the Monte-Carlo analysis coincide with the recharge area from the deterministic model, particularly that area closest to the well. In those areas further from the well, the percentage of particles from the Monte-Carlo analysis that coincide with the results of the deterministic model is lower. Each of the probabilistic model areas also shows a pattern where the spread of the low percentage of particles increases with the distance from the pumping wells. These results are expected for this type of analysis and indicate the probability of the deterministic contributing area coinciding with the true recharge area is greatest near the well and decreases with distance. One exception to this pattern is for well Cu 970, where some of the 26 to 50 percent probability area is in the headwaters and separated from the area closer to the well. This is likely an indicator of the importance of stream recharge in determining the contributing area.

Model Limitations

The accuracy of the model is limited by the accuracy of the data and assumptions that were used in modeling. The major assumption is that, on a sufficiently large scale, flow through fractured bedrock is equivalent to flow through a porous medium that can be simulated by the MODFLOW model. Measurement error for water levels is one source of potential error, but that is likely to be negligible in this situation. Conversion of water-level measurement to water-table elevation is limited to the resolution of the maps used to plot the data, which had a contour interval of 10 ft. Errors in these measurements are not likely to be systematic (that is, always negative in a certain aquifer) but are likely to impart inaccuracy to the model. The more likely sources of error are in cases where a well was recently pumped or was completed in a perched water table. These wells were likely to have been detected during model calibration and removed from consideration. Water levels also were collected over a period of several months where conditions in the aquifer were changing slightly. The most significant potential source of error is the lack of water levels in the South Mountain area of the model. The lack of wells prohibited collection of information that could be used to calibrate water levels in this area. No wells are in the downgradient areas of the model because the model area was expanded late in the process. The lack of data in this area is not likely to affect model predictions because it is downgradient from the area of interest; however, water levels and fluxes in this area are less likely to reflect actual conditions than would be the case in the area where there are more wells for calibration.

Streamflow in and out of the aquifer were simulated in this model; however, these flows were difficult to constrain and are only approximated in the model. Flow measurements were made under a set of conditions that may not be representative of a steady-state condition. Accurate simulation of these flows would require more data about streamflow under various hydrologic conditions. A transient model could be used to illustrate these conditions. Also, the values used for recharge affect the overall water budget, and there is potential that actual recharge is not the same as that used in the model.

Ground-water traveltimes are not simulated in this model. The values of porosity affect the velocity of ground water; however, porosity values are not used for steady-state simulation. Tracer tests and analysis of aquifer materials could help determine porosity and age-dating of ground-water samples could be used to verify these values. The extreme variation of flow velocities may make it difficult to simulate average traveltimes. Flow rates are likely to be very slow in clays, moderate in sands and fractured bedrock, and very fast in karst conduits. For the purposes of wellhead protection, the fastest possible routes from the land surface to the well are the ones that are of greatest concern.

The discretization of the model is another source of potential error. A finer grid could be used to determine some of the minor variations in the water table that could not be determined at the scale used, if additional data are available to support the refinement. Two layers were used to simulate ground-water flow in this model. Additional data on geologic and hydrologic features could indicate a need to include more model layers. Geophysical studies that could determine the depth and characteristics of the various geologic units could help further refine the model as well. Data on the depth and characteristics (such as percent sand or clay) of the colluvium would be particularly helpful in improving the understanding of this system.

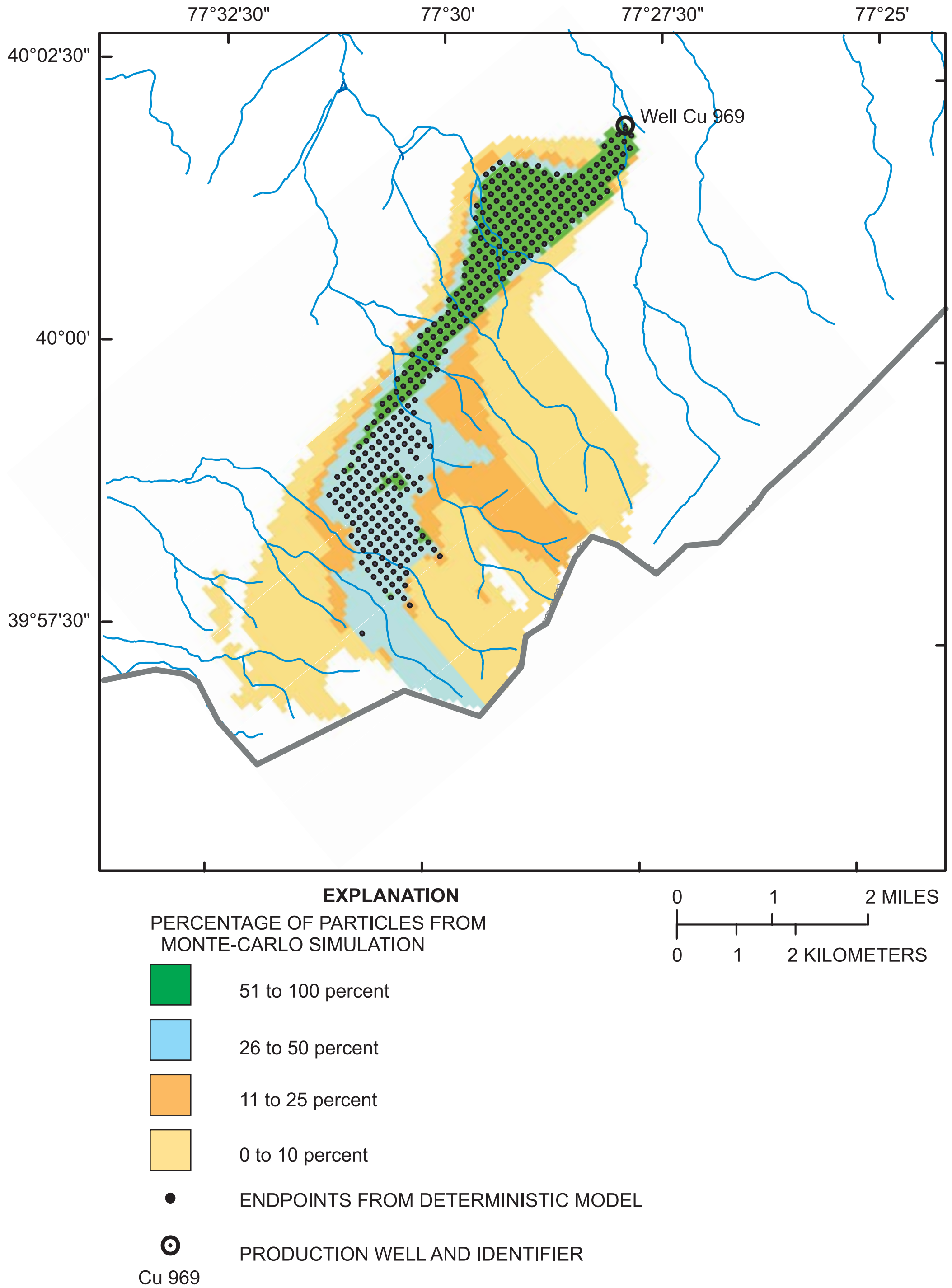


Figure 29. Deterministic recharge locations superimposed on probabilistic recharge locations for production well Cu 969 near Shippensburg, Pa.

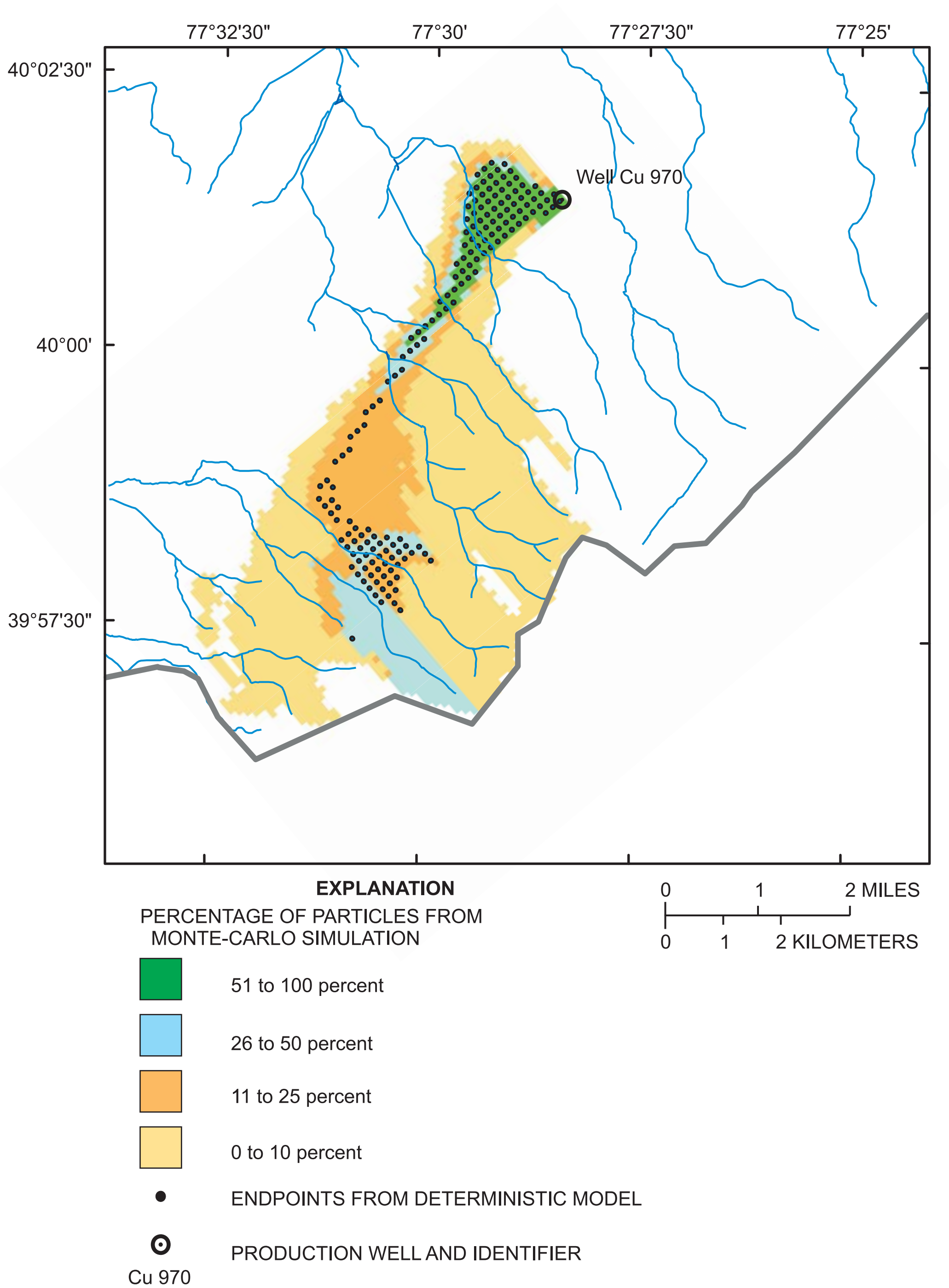


Figure 30. Deterministic recharge locations superimposed on probabilistic recharge locations for production well Cu 970 near Shippensburg, Pa.

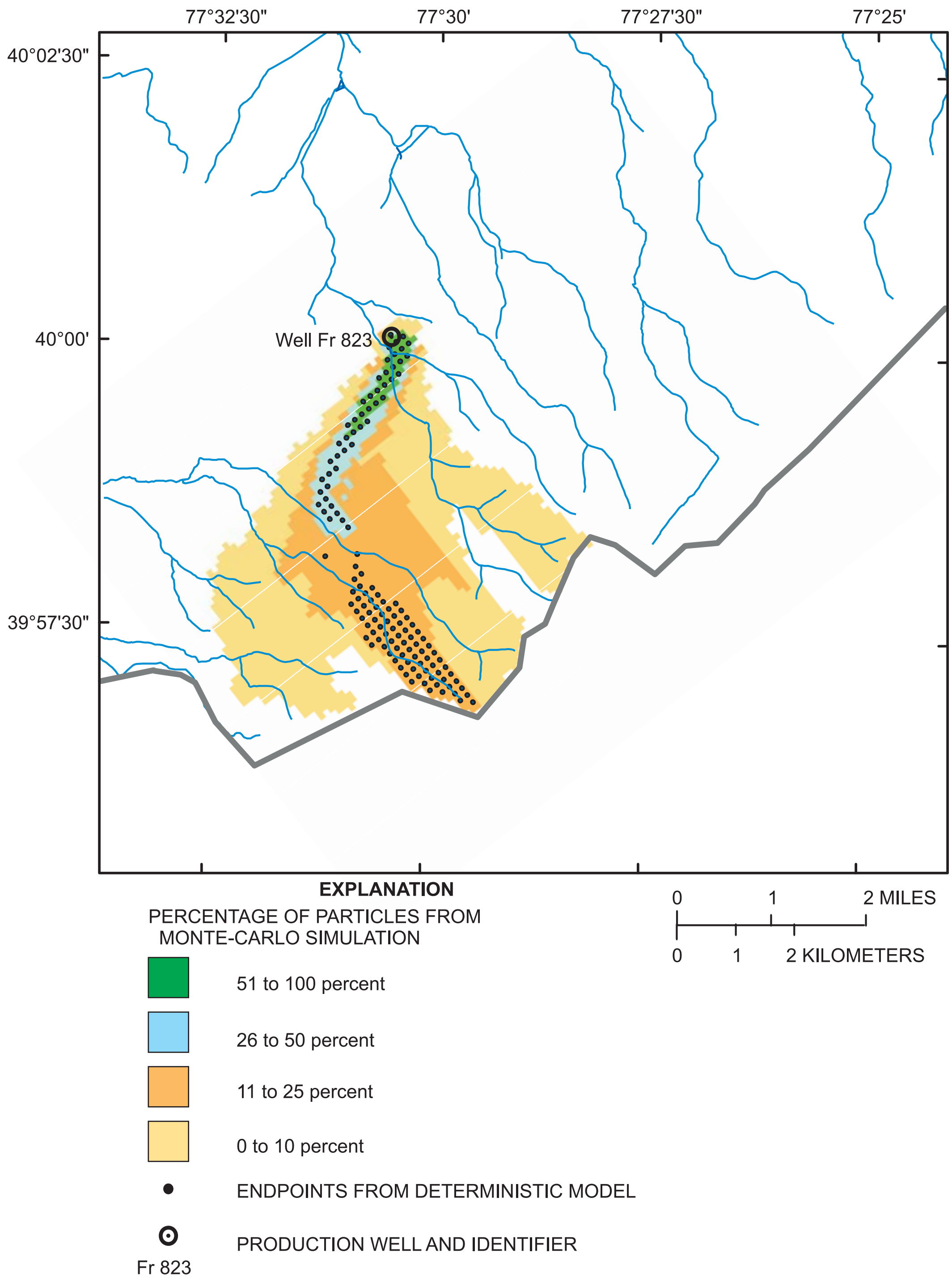


Figure 31. Deterministic recharge locations superimposed on probabilistic recharge locations for production well Fr 823 near Shippensburg, Pa.

Summary and Conclusions

The hydrogeology of South Mountain and the Great Valley near Shippensburg, Pa., was analyzed to determine the areal extent of the zone of contribution to the three production wells owned by the Shippensburg Borough Authority. The results of this study can be applied to similar carbonate-bedrock aquifers in Pennsylvania and other similar areas. The project was conducted in cooperation with the Shippensburg Borough Authority. The geology of this area is complex, and the hydraulic characteristics of the geologic units in the area are highly variable. The metamorphic rocks of South Mountain typically have low transmissivity, and the carbonate-bedrock aquifer of the Great Valley typically has relatively high transmissivity. Some of the siliciclastic aquifers in the Great Valley have relatively low transmissivity. The carbonate-bedrock aquifer adjacent to South Mountain is covered by colluvium. The colluvium is an important part of the hydrogeology of this area, but the exact depth and characteristics of it are unknown. Colluvium is typically deepest near the contact between the metamorphic and carbonate rocks. In these areas, chemical weathering of the carbonate rocks has caused the surface of the carbonate bedrock to be lowered, and mass-wasting processes have moved colluvium from South Mountain downslope to fill in where the carbonate rock was removed. These areas also contain an unknown amount of residuum from the dissolution of the carbonate rock, which is below the colluvium. Numerous karst features, particularly sinkholes and closed depressions, are in the carbonate areas of the Great Valley. In some cases, water can flow into the sinkholes and closed depressions to provide direct recharge to the aquifer. In areas where the closed depressions are covered by colluvium, several hundred feet of unconsolidated material can exist between the surface of the feature and the bedrock aquifer. There are several caves are in the area. Several wells in the area are known to penetrate voids that are likely to be part of a conduit-flow system. Water-level fluctuations and borehole geophysical logs indicate that the deeper carbonate-bedrock aquifer in the area where these wells are completed is confined.

The streams originating in South Mountain are typically perennial streams with normalized flows that are in the same range as the Yellow Breeches or Conodoguinet Creek. As these streams flow from South Mountain into the Great Valley and over the colluvium on the flank of South Mountain, water is lost from the streams to the aquifer. The volume of water lost is variable, depending on streamflow conditions, but some streams lose all their flow to the aquifer under low-flow conditions. Streams flowing over the colluvium, and even further out in the valley, transmit volumes of water that are much less than the values that would be expected given the precipitation and recharge calculated for this area. Therefore, much of the water recharged in this area must travel downgradient through the valley as ground-water flow. The discharge points of these ground-water-flow paths are large streams and springs further downgradient.

Ground-water flow was simulated to determine the areal extent of the zone of contribution to the municipal wells. Simulation of ground-water flow in this area required modeling a 111-mi^2 area to allow ground water to flow to natural discharge points. The model was bounded on the upgradient side by the top of the ridge on South Mountain. The tributaries of Yellow Breeches Creek, Conococheague Creek, and Conodoguinet Creek formed lateral and downgradient model boundaries. Tributaries of Middle Spring Creek and Burd Run were simulated inside the model. Dykeman Spring was simulated as a constant flux, and Big Spring was simulated as a constant head. Aquifer properties were simulated using two layers. The upper layer represented soils, colluvium, and highly fractured bedrock. The lower layer represented the fractured-bedrock aquifer and was simulated as a confined aquifer. Each layer was divided into five zones. The zones initially were based on geologic formations but later were modified into hydrostratigraphic units during model calibration. Initially, recharge was assigned as a single value for the model area, but during calibration, separate values of recharge were assigned to the metamorphic rocks, the carbonate-bedrock aquifer of the valley, and the urban area.

Model calibration initially was done using streamflow and head measurements; however, final model calibration included only head measurements. The model was unstable when calibrating to stream flow; therefore, these measurements were not used for calibration. The root mean square error of the model was 9.8 ft. Post-processing programs indicated the regression was linear, which indicated the confidence intervals based on linear theory were likely to be accurate. Analysis of the Cooke’s D and DFBeta statistics indicated observations with large influence on the regression were accurate and had no spatial pattern. Three parameters were estimated by the parameter-estimation process. The value of horizontal hydraulic conductivity of the highly fractured carbonate-bedrock aquifer and sandy colluvium (layer 1) was 233 ft/d; the value of the horizontal hydraulic conductivity of the Waynesboro and ridge-forming hydrostratigraphic unit (layer 1) was 3.97 ft/d; and the value of horizontal anisotropy was 1.73 (layer 1 and layer 2).

The calibrated model was used to determine the areal extent of the zone of contribution to production wells Cu 969 and Fr 823. Simulation of the areal extent of the zone of contribution to well Cu 970, which is scheduled to be brought on line in 2006, was conducted by adding that well as a pumping well, but without additional model calibration. The areal extent of the zone of contribution to each well showed a similar pattern. The patterns had an area that extended to the southwest from the well, following the Tomstown Formation, then extending to the southeast into the metamorphic rocks of South Mountain. All wells also had part of the areal extent of the zone of contribution bisected by a stream that was losing water in that area. Therefore, the watershed of the streams potentially contributing recharge to the aquifer was also delineated and was considered as the watershed of losing streams providing recharge to the wells.

A probabilistic representation of model uncertainty was determined by a Monte-Carlo analysis. A set of 1,000 combinations of parameters were created to simulate alternate possible model solutions. The model was conditioned to run the MODPATH program only for parameter sets that converged and had a reasonable water budget. The areal extent of the zone of contribution was determined for 980 solutions. The results of these solutions were compiled and contoured. Comparison of the deterministic results to the probabilistic results indicated the model results matched most closely in the area of the Tomstown Formation closest to each of the wells. Areas further from each well and upgradient in the metamorphic rocks of South Mountain had a higher degree of uncertainty.

Acknowledgments

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