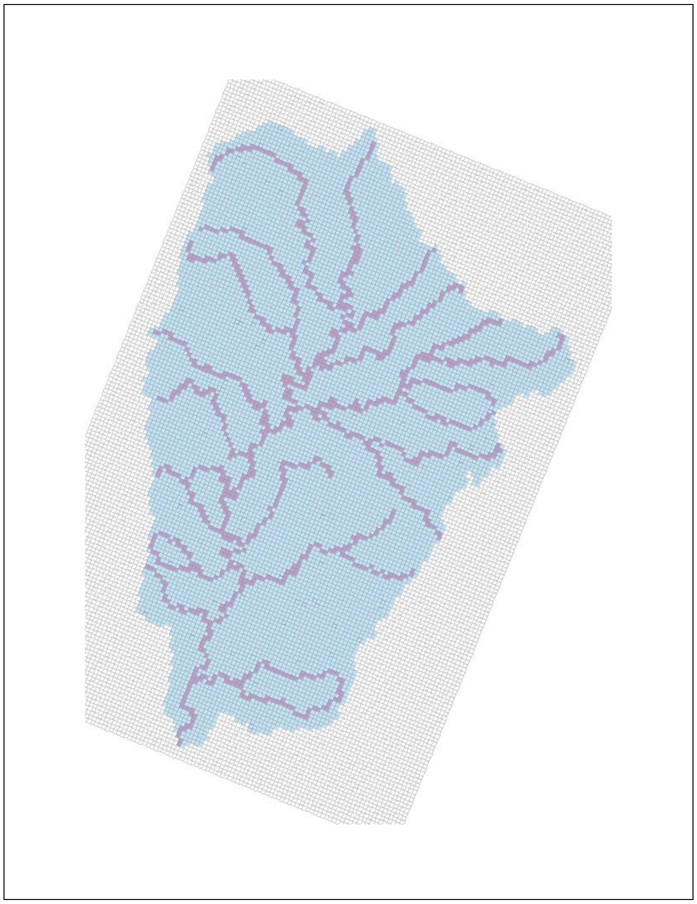
Ground-water/Stream Flow Model of the Monocacy River Basin, Maryland and Pennsylvania
Phase I: Steady-State Model



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Executive Summary

We report on the first phase of development of a regional model of ground-water and stream flow in the Monocacy River basin, a 970 square mile drainage area located in Adams County, Pennsylvania, and Frederick, Carroll and Montgomery Counties in Maryland. Many areas of the Monocacy basin are experiencing rapid population growth due to proximity to the Washington, DC metropolitan area. Residents and industries rely on water drawn both from streams and from wells drilled into the bedrock aquifers which underlie the basin. There is concern that development will strain water resources beyond their capacity to meet increased demand.

Regional ground-water/stream flow models are a tool for understanding the impact of population growth and development on water levels in wells and on streams. They incorporate available geologic, hydrologic and meteorological data and make use of our understanding of the physical processes responsible for the movement of water through the ground to simulate ground-water flow patterns, aquifer water levels, and stream flow rates. Models can provide information on the spatial distribution and availability of ground-water and stream flow under current withdrawal conditions and under future growth scenarios.

The sub-surface of the Monocacy basin consists of a layer of unconsolidated material, or “regolith”, composed of soil, clay, sand, and pieces of weathered bedrock and with a thickness typically in the range of 15 to 40 feet, overlying thousands of feet of bedrock. Ground-water in the basin’s “fractured bedrock aquifers” resides in the pores of the regolith and in the crevices and fractures of the underlying bedrock. The ground-water/stream flow model simulates ground-water levels in these aquifers, the discharge of ground-water from aquifers to basin streams, and the rate of flow of water in streams. The model was calibrated with available data from 1980, a time period which was chosen to represent pre-development conditions in the basin. Mean annual water levels and stream flows for 1980 were simulated with reasonable accuracy. After the calibration was completed, the model’s sensitivity to changes in inputs was investigated, with a focus on stream flow predictions.

The model described in this report simulates “steady-state” conditions, and is capable of simulating average ground-water levels and stream flows. It may be used in future projects to simulate average summertime conditions and to investigate the impact of future demand on summertime water availability. In upcoming phases of model development, the steady-state model will serve as the basis for construction of a transient model, which will incorporate the effects of aquifer storage and be capable of simulating seasonal fluctuations in aquifer water levels and stream flows, providing a more realistic predictive tool for evaluating summertime water availability.

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Ground-water/Stream Flow Model of the Monocacy River Basin, Maryland and Pennsylvania

1. Introduction

The Monocacy River basin is a 970 square mile land area drained by the river and its network of tributaries, covering portions of Adams and Franklin Counties in Pennsylvania and Frederick, Carroll and Montgomery Counties in Maryland. The waters of the Monocacy River flow into the Potomac River just south of Point of Rocks, Maryland. The largest municipalities in the Monocacy basin are the City of Frederick in Maryland and the Borough of Gettysburg in Pennsylvania. These and other basin localities are experiencing rapid population growth due to proximity to the Washington, DC metropolitan area. Residents and industries rely on water drawn both from streams and from wells drilled into the bedrock aquifers which underlie the basin. There is concern that development will strain water resources because of the potential for increased demand, and because the accompanying increases in pavement and other “impervious surfaces” tend to decrease the amount of water able to percolate downward to replenish the basin’s aquifers.

We report on the first phase of development of a regional model of ground-water and stream flow in the Monocacy River basin. Ground-water/stream flow models are an important tool for understanding the impact of population growth and development on water levels in wells and on stream flows. They incorporate available geologic, hydrologic and meteorological data and make use of our understanding of the physical processes responsible for the movement of water through the ground to simulate ground-water flow patterns, aquifer water levels, and stream flow rates. Models can provide information on the spatial distribution and availability of ground-water and stream flow under current withdrawal conditions and under future growth scenarios.

Regional ground-water/stream flow models are constructed to help answer questions at a regional level. Unlike smaller-scale models, which typically address issues such as how ground-water pumping at a single location affects water levels in adjacent wells, regional models can provide information on how significant and widespread increases in ground-water withdrawals may affect small and medium-sized watersheds. A model of the Monocacy basin can potentially help answer the following questions:

- How will ground-water withdrawals associated with proposed growth scenarios affect flows in nearby streams?
- How will significant new ground-water withdrawals in one of the basin’s tributary watersheds affect wells in that watershed, both upstream and downstream of the withdrawals?

- How will new ground-water withdrawals in one of the basin's tributary watersheds affect aquifer levels in adjacent watersheds?

Construction of a model for the Monocacy basin poses a number of challenges. The basin is underlain by bedrock aquifers, and ground-water flow is governed, to a large degree, by the distribution and characteristics of bedrock fracture systems which lie deep below the ground's surface and for which little information is available. The Monocacy basin is an area with fairly high topographic relief, constraining options for the design of a finite difference grid, as used by the USGS's MODFLOW ground-water flow modeling package. Finally, the data set of available well level and stream flow observations is limited in both its spatial and temporal coverage.

The first phase of the Monocacy ground-water/stream flow model, described in this report, is a "steady-state" model of average annual conditions. Steady state models are capable of simulating the average spatial distribution of water levels in basin aquifers and average flows in streams, but are not able to simulate seasonal changes in water availability and the effects of aquifer storage. In subsequent phases of model development, the steady-state model will serve as the basis for construction of a transient model capable of simulating seasonal fluctuations in aquifer water levels and stream flows.

This study was conducted as part of the Potomac River Basin Ground-water Assessment Project, a collaboration between the Interstate Commission on the Potomac River Basin (ICPRB) and the U.S. Geological Survey (USGS). Funding for the project was provided by the USGS under Cooperative Agreement, Award Number 03ERAG0042, by the Maryland Department of Environment (MDE), and by ICPRB.

2. Purpose

This report provides results from a steady-state ground-water/stream flow model of the Monocacy River basin, located in Frederick, Carroll, and Montgomery Counties, Maryland and Adams and Franklin Counties, Pennsylvania. The model was calibrated to available data from the time period, January 1 through December 31, 1980, selected to represent pre-development conditions in the basin. The year, 1980, has a reasonable amount of observation well and stream flow data available, and it is the earliest year in which water use data were collected throughout the state by the Maryland Department of Environment. This report describes data compilation and data analysis, model geometry and initial values for model inputs, adjustments made to model inputs during the calibration process, and comparisons of final calibrated model predictions versus available well level and stream flow data.

The purpose of the steady-state version of the flow model, described below, is threefold. First, the steady-state model can serve as a preliminary tool for the evaluation of the impact of future demand on summertime water availability, by running the model with average summertime recharge inputs. Second, it is anticipated that the steady-state model will serve as the basis for development of a transient model of the basin, which will be able to simulate seasonal changes in water levels and stream flow and incorporate the effects of aquifer storage. Third, the steady-state model of the Monocacy basin can provide information and a framework for constructing more detailed finer-scaled models of selected watershed in the basin.

3. Previous Investigations

The ground-water resources and hydrogeology of the Monocacy River basin have been investigated and summarized in numerous reports covering this region of the country. Among the sources used to provide information for this report are: Ground-water occurrence in the Maryland Piedmont (Nutter and Otton, 1969), Hydrogeology of the carbonate rocks, Frederick and Hagerstown Valleys, Maryland (Nutter, 1973), Geohydrologic reconnaissance of the upper Potomac basin (Trainer and Watkins, 1975), Hydrogeology of the Triassic rocks of Maryland (Nutter, 1975), Ground-water in the Piedmont upland of central Maryland (Richardson, 1982), Ground-water and surface-water data for Frederick County, Maryland (Dine, Tompkins and Duigon,1985), Water Resources of Frederick County, Maryland (Duigon and Dine, 1987), Geohydrology and water quality in the vicinity of the Gettysburg National Military Park and Eisenhower National Historic Site, Pennsylvania (Becher, 1989), Plan of study for the regional aquifer system analysis of the Appalachian Valley and Ridge, Piedmont, and Blue Ridge physiographic provinces of the Eastern and Southeastern United States with a description of study area geology and hydrogeology (Swain and others, 1991), Estimated hydrologic characteristics of shallow aquifer systems in the Valley and Ridge, the Blue Ridge, and the Piedmont physiographic provinces based on analysis of streamflow recession and base flow (Rutledge and Mesko, 1996), Summary of hydrogeologic and ground-water quality data and hydrogeologic framework at selected well sites, Adams County, Pennsylvania (Low and Dugas, 1999), Hydrogeology and simulation of ground-water flow at the Gettysburg Elevator Plant Superfund site, Adams County, Pennsylvania (Low, Goode and Risser, 2000), Geohydrology of Southeastern Pennsylvania (Low, Hippe and Yannacci, 2002), Ground-water availability in part of the Borough of Carroll Valley, Adams County, Pennsylvania and the establishment of a drought-monitor well (Low and Conger, 2002), and Stratigraphy-karst relationships in the Frederick Valley of Maryland (Brezinski and Reger, 2002).

4. Location and Hydrogeologic Setting

4.1. Location

The Monocacy River drains approximately 970 square miles of Frederick, Carroll, and Montgomery Counties in Maryland and Adams and Franklin Counties in Pennsylvania (Figures 1 and 2). It discharges into the Potomac River downstream of Point of Rocks, Md at an average annual rate of approximately 600 million gallons per day (based on the streamflow record from

1930 to 2002 at USGS gage 01643000, Monocacy River at Jug Bridge near Frederick Maryland, (USGS-NWIS, 2006). The climate of the Monocacy basin is moderately humid temperate. Precipitation records of varying record length are available at several stations within and nearby the basin. The average annual precipitation based on records for the period 1985 to 2002 is 42.7 inches per year. The mean annual temperature at Frederick is 53.3°F (Duigon and Dine, 1987). The basin is located in parts of two major physiographic provinces, the Blue Ridge province and the Piedmont province as described in Fenneman (1938). The Piedmont physiographic province in Maryland has been subdivided into the Eastern and Western Piedmont provinces (Bolton, 1996). The surface physiography of the Monocacy River Basin varies from gently rolling hills in the south and central parts of Frederick County, Maryland, and central Adams County, Pennsylvania to the relatively steep mountainous eastern edge of the Blue Ridge Mountains (Stose and Stose, 1946) making up the western boundary of the basin.

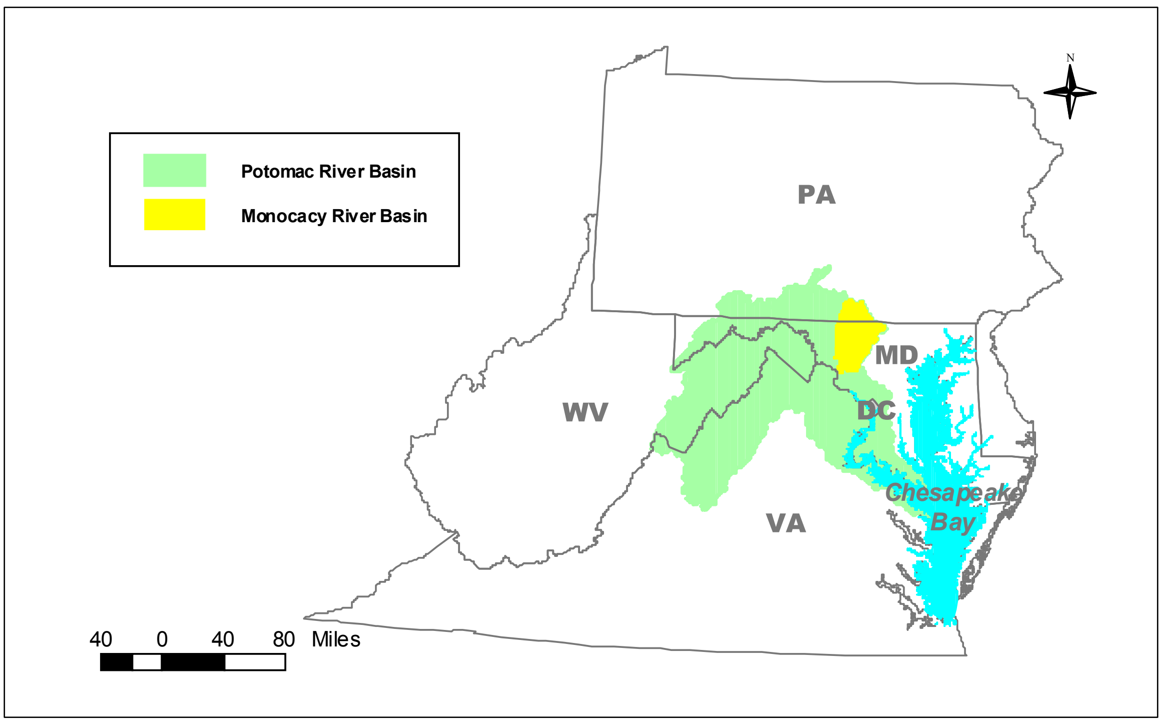

Figure 1. Location of Monocacy River basin, a sub-basin of the Potomac River basin

4.2. Geology

Approximately 47% of the Monocacy River basin is in the Western Piedmont physiographic province, which forms most of the eastern part of the basin (Figure 3). The Western Piedmont province is from 5 to 13 miles wide in the basin and is present mostly in Frederick and Carroll Counties in Maryland with lesser areas in Montgomery County, Maryland,

and Adams County, Pennsylvania. It forms a gently rolling upland with an average elevation of 700 to 800 feet (ft), with relief generally less than 500 ft and is incised by many deep narrow stream valleys (Stose and Stose, 1946). The Western Piedmont within the Monocacy River basin is underlain by Precambrian and Cambrian metamorphic and igneous rocks (approximately 82%) with some imbedded carbonate and quartzite bodies (approx. 2%). The remainder (approx. 16%) of the Western Piedmont province is underlain by Cambrian age carbonates similar in composition and structure to the Great Valley carbonates in western Maryland, Virginia and West Virginia. These carbonates make up the floor of Frederick Valley. At the southwestern corner of the basin is a monadnock, named Sugarloaf Mountain, with a summit of 1,282 ft above sea level composed of Cambrian age quartzite overlying phyllite (Vokes and Edwards, 1974).

The central portion of the Monocacy River basin is underlain by rocks of the Mesozoic Lowland province in a narrow belt extending northward through the basin from the Potomac River to the northern boundary of the basin in Pennsylvania. This belt is, at its narrowest, about 1 mile wide near Frederick and is about 17 miles wide just north of the Maryland - Pennsylvania border. The Mesozoic Lowland underlies approximately 37% of the Monocacy River basin. The rocks in this province include Triassic-age consolidated and compacted sedimentary layers of sandstone, shale, arkose, and conglomerate with numerous igneous intrusions. These igneous intrusions are formed of diabase, a dense, fine-grained rock that is resistant to weathering. These intrusive bodies frequently form low ridges and act as impermeable barriers to ground-water flow (Focazio, et al, 1997). However, there are reports from well drillers that very high-yielding wells have been installed in the altered margins between the diabase intrusions and the country rock.

The Blue Ridge province is represented in the Monocacy River basin in the mountain ridges that make up the western boundary of the basin, Catoctin and South Mountains. These ridges are joined in the north forming a highland about 9 miles wide at its widest extending from the northwestern corner of the basin to just south of the Maryland-Pennsylvania border. Where separate, the ridges run almost parallel south to the Potomac River with the eastern ridge, Catoctin Mountain, forming the western boundary of the basin. In total, the two ridges, and thus the Blue Ridge province, make up approximately 16% of the basin. The highest point in the basin is on South Mountain in Adams County, Pa, at 1,982 ft above sea level. The lowest point in the Blue Ridge province is approximately 374 ft, along the eastern edge of Catoctin Mountain in southern Frederick County, Md. The rocks that form these ridges consist of Precambrian metavolcanic rocks of the Catoctin Formation, and the members of the Cambrian age Chilhowee Group; phyllite of the Loudoun Formation, quartzites of the Weverton Formation, metasiltstone of the Harpers Formation, and metasandstone of the Antietam Formation (Southworth, et al, 2002).

Overlying the fractured bedrock of the basin is a layer of overburden, or regolith, composed of weathered bedrock, soil, alluvium, and colluvium. The length of casing installed in ground-water wells has been used as an indicator of the thickness of the regolith in many studies in the region (e.g. Nutter and Otton, 1969, Richardson, 1982, and Low and others, 2002).

Richardson (1982) in a study in the Piedmont Upland in Maryland reported regolith thickness from 0 to more than 100 ft, generally ranging between 20 and 40 ft. Low and Dugas (1999) reported on well data by hydrogeologic unit and topographic setting in Adams County, Pa. The range of depth to bedrock was 0 ft to 356 ft with the median value for the hydrogeologic units ranging from 14 to 35 ft. Low, Hippe and Yannacii (2002) summarized information on the geohydrologic system in the Southeastern Pennsylvania. Of the six physiographic provinces present in their study area, three are also present within the Monocacy River basin; the Piedmont Upland, Gettysburg-Newark (Mesozoic) Lowland, and the South Mountain section of the Blue Ridge province. The Piedmont Lowland is also present in the eastern corner of the Adams County portion of the basin but makes up less than 1% of the basin area. The median casing length in wells installed in the units within the Monocacy basin ranged from 25 ft to 118 ft. The authors of this report reviewed records of 5,792 wells in Frederick, Carroll, and Montgomery Counties in Maryland and Adams, Franklin and York Counties in Pennsylvania with casing length reported. The casing lengths ranged from 0 to 360 ft with a median casing length of 37 ft. These wells represent all hydrogeologic units and water use types.

Figure 2. Geographic features, counties, and municipalities of the Monocacy basin

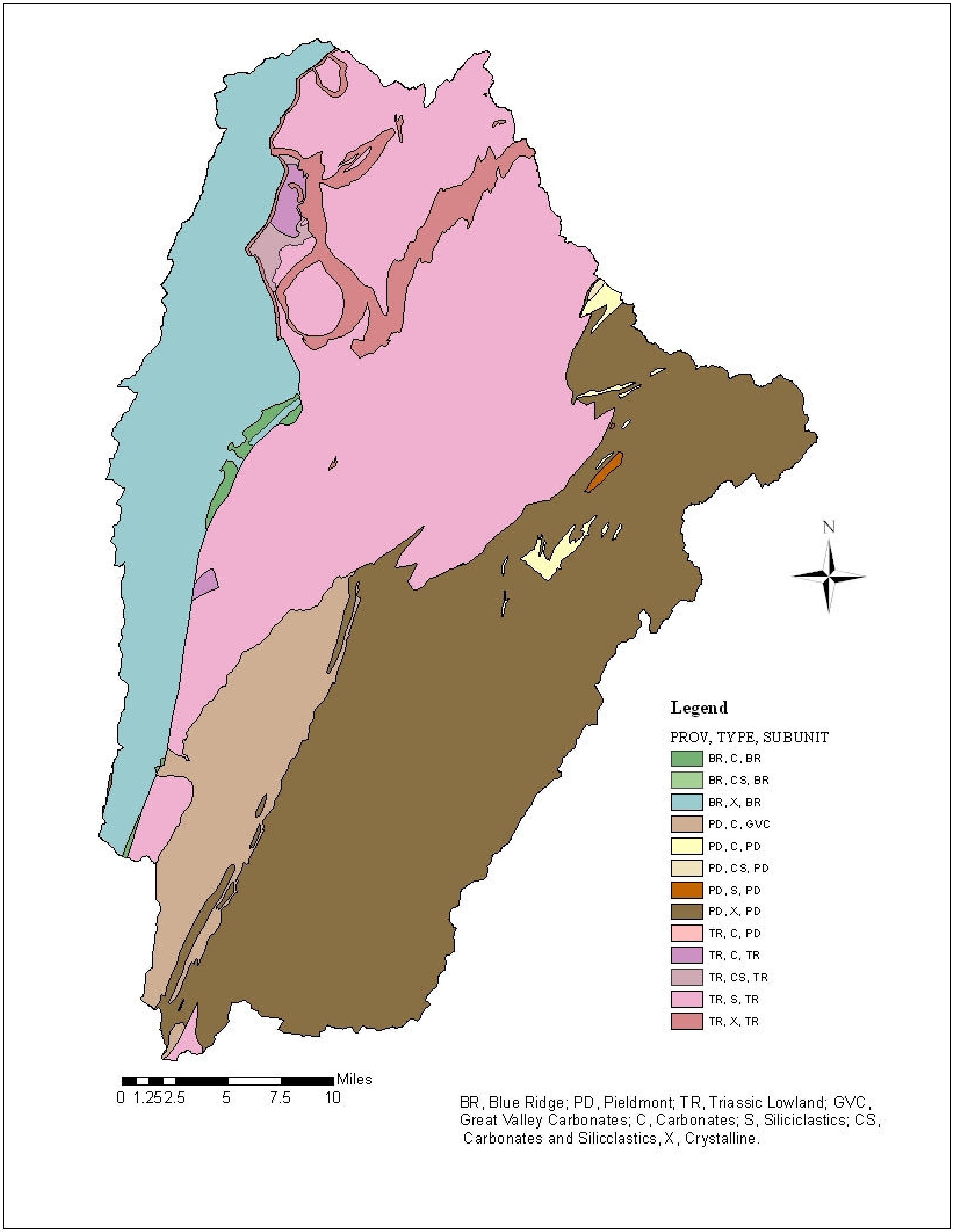

Figure 3. Geologic units in the Monocacy River basin (Derosier and others, 1998).

4.3. Available Data

Only data that could be readily accessed in electronic format were used. Data were obtained from the USGS Ground-water Site Inventory (GWSI) database, USGS 10 m National Elevation Dataset (NED), USGS National Hydrography Dataset (NHD) (USGS-NHD, 2003), USGS National Water Information System (USGS-NWIS, 2006), and the Pennsylvania Ground Water Information System (PaGWIS) database.

4.3.1. Water-Level Measurements

Water level measurement data are available from the USGS National Water Information System (NWIS) (USGS-NWIS, 2006) for wells throughout the nation. Data for wells within the Monocacy River basin were downloaded from this database and reviewed. From this initial dataset, wells were selected which had water level measurements in 1980, the "pre-development" period assumed in the definition of the conceptual model, and which provided a relatively even spatial coverage of the basin. This process resulted in 48 observation wells within the basin selected to provide water level data for model calibration. At many of these wells, only a single water level measurement was recorded in 1980, and observations at the wells were taken at various times throughout the year. Because ground-water levels vary seasonally in the basin, values for mean well levels in the calibration data set may differ considerably from actual 1980 annual means. As such, they give only an approximate representation of average, steady-state ground-water levels in 1980. Risser (2006) used water level measurements from similar sources and estimated that the water levels represented average, steady-state conditions to about +/- 30 ft of the true value. The locations of the observations wells used in model calibration are shown in Figure 4, and their mean 1980 water levels are given in Table A1 in Appendix A.

4.3.2. Discharge Measurements

Historical stream flow data are available at 20 stream gaging stations that have been operated by USGS within the Monocacy River basin. Data from twelve of these gages were used for model calibration (see Figure 4). Of the eight gages not used, six were in stream reaches not represented by the model stream network and had areas of less than five square miles, one, 01642000, had a period of record judged to be too far outside the model simulation period, and one, 01642190, had a period of record of less than three years at the time this report was prepared. Table 1 is a list of the gaging stations used in model calibration, along with their associated drainage areas, periods of record, coordinates, and flow information. Annual (calendar year) mean discharge data for these gages were downloaded from their USGS NWIS web pages. In some cases, where there were little or no records in 1980, an estimate was made using the annual average flow per watershed area of a watershed in the Monocacy basin having the same underlying lithology. The mean annual discharge was then multiplied by a Base Flow Index (BFI) (Stewart et al., 2006; Wahl and Wahl, 1995) to estimate a mean annual base flow at the gage for comparison with the model calculated flows. In Table 1, stream discharge is given in units of cubic feet per second (cfs).

4.3.3. Extraction Wells

One of the major stresses on the ground-water system is the withdrawal of water through extraction wells. Ground-water extraction data was taken from databases provided by USGS of compiled data from the Maryland Department of the Environment and Pennsylvania Ground-water Information System (PaGWIS). Only wells that were in use in 1980 were used, as the initial version of the flow model is a steady state model with the hydrologic conditions in 1980 approximating "pre-development" conditions within the basin. Table A1 in Appendix A lists the extraction wells, and Figure 4 shows the locations of model cells that contain simulated extraction wells in the model. The depths of the wells ranged from 9 meters (m) to 312 m, with a mean depth of 87 m. Twenty-six percent of the wells were 50 m or less in depth, 39% were between 50 m and 100 m, 31 % were between 100 m and 200 m, and only 4% were greater than 200 m in depth.

Table 1. USGS Stream gaging stations used for model calibration

Station Name	Station ID	Watershed Area (sq. mi.)	BFI	Period of Record Begin Date	Period of Record End Date [c]	Lat	Long	Mean CY1980 Flow (cfs)	Mean Annual Flow for Period of Record (cfs)
Monocacy River at Bridgeport, MD	01639000	173.0	0.24	5-1-1942	9-30-2005	39 40 43.8	77 14 04.2	139.2	210.4
Piney Creek near Taneytown, MD	01639140	31.3	0.29	5-1-1990	1-15-2002	39 39 38.7	77 13 15.5	26.7[a]	34.0
Toms Creek at Emittsburg, MD	01639375	41.3	0.42	3-1-1986	9-30-1990	39 42 13	77 20 41	46.2[b]	52.0
Big Pipe Creek at Bruceville, MD	01639500	102.0	0.52	10-1-1947	9-30-2005	39 36 44.5	77 14 14.8	91.3	115.6
Little Pipe Creek at Avondale, MD	01640000	8.1	0.72	8-29-1947	9-30-1956	39 33 40	77 02 38	7.1[a]	9.0
Owens Creek at Lantz, MD	01640500	5.9	0.50	10-1-1931	9-30-1984	39 40 36	77 27 50	7.7	9.4
Hunting Creek near Thurmont, MD	01640975	7.1	0.58	12-31-1981	5-6-1986	39 37 42	77 27 44	9.5[b]	11.7
Hunting Creek at Jimtown, MD	01641000	18.4	0.50	10-1-1949	2-18-1992	39 35 37.0	77 23 47.48	23.0	26.5
Fishing Creek near Lewistown, MD	01641500	7.3	0.65	10-1-1947	9-30-1984	39 31 37.57	77 28 01.27	11.5	11.6
Linganore Creek near Frederick, MD	01642500	82.3	0.55	11-27-1931	9-30-1982	39 24 55	77 20 00	73.1	86.1
Monocacy River at Jug Bridge Near Frederick, MD	01643000	817.0	0.43	10-1-1929	9-30-2005	39 24 10.2	77 21 57.9	750.7	954.4
Bennett Creek at Park Mills, MD	01643500	62.8	0.55	7-29-1948	9-30-2005	39 17 38.9	77 24 25.5	60.1	70.9

[a] Estimated 1980 flow from flow/sq mi of Big Pipe Creek.

[b] Estimated 1980 flow from flow/sq mi of Hunting Creek at Jimtown.

[c] Period of record end date for daily statistics, at the time this study was conducted.

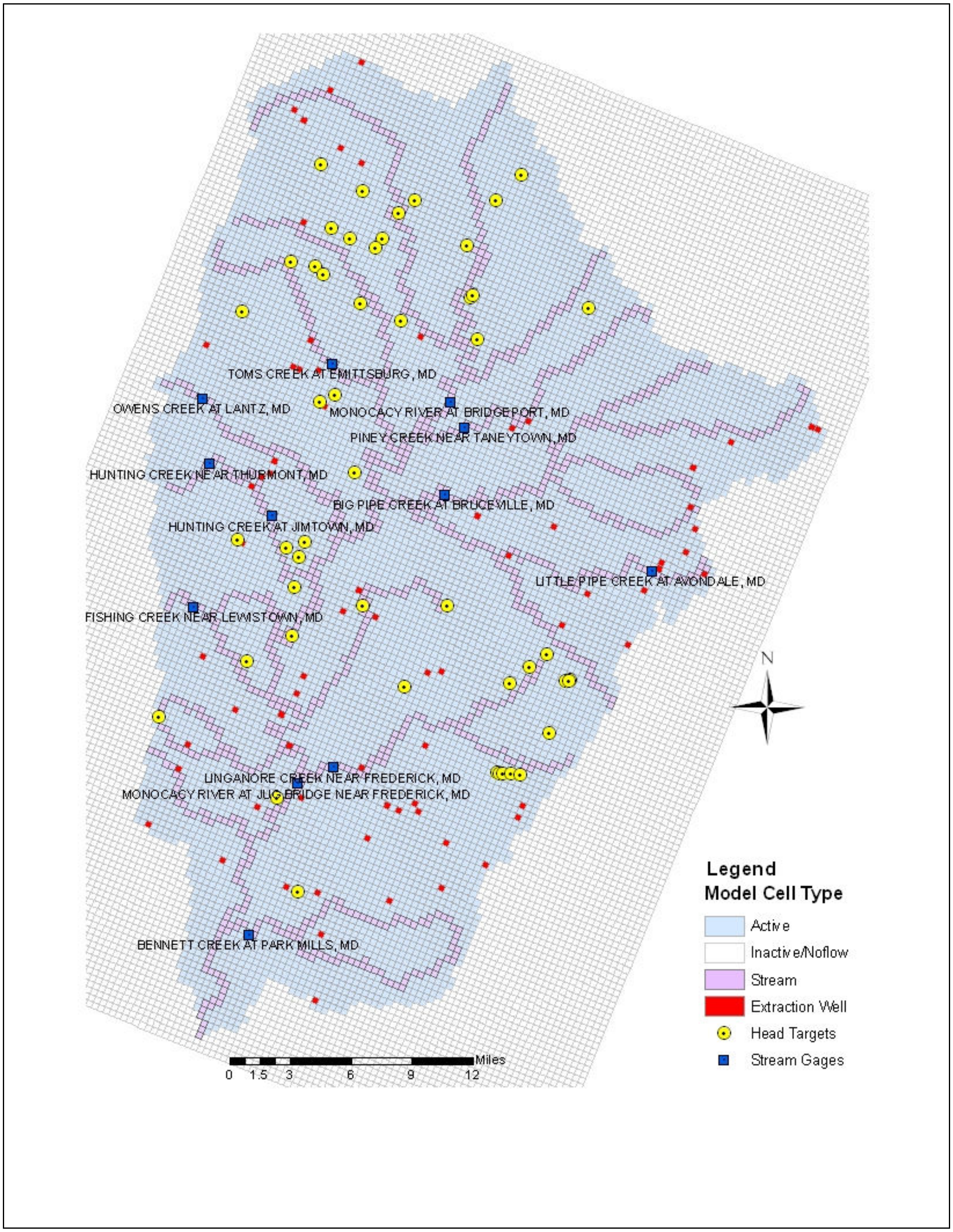

Figure 4. Model grid, with locations of stream and extraction well cells, water level targets, and stream gages.

4.4. Monocacy Basin Water Budget

Information on the annual water budget for the Monocacy basin is given below. In Table 2, ground-water withdrawals, in million gallons per day (MGD), are given for three years of interest. The year 1980 is the simulation year of this study, representing pre-development conditions. The years 2000 and 2030 represent current conditions and projected conditions, respectively (Wolman, 2004). In Table 3, estimates of basin recharge are given, based on statistical analyses of annual stream base flows (Schultz et al., 2005). In this table, recharge for the 2-year return period represents recharge in an average year, and the 10-year represents recharge in a moderately dry year. The 20-year return period represents recharge in a moderate drought year, with a severity likely to occur once in every 20 years.

Table 2. Monocacy basin annual ground-water withdrawals

	1980	2000	2030 - projected
Withdrawal amount (MDG)	15	25	34

Table 3. Monocacy basin annual recharge

	2-year return period	10-year return period	20-year return period
Recharge amount (cfs/MGD)	602/389	418/270	378/244

Total ground-water withdrawals for the Monocacy basin as a percentage of total annual recharge can be computed from Tables 2 and 3. Withdrawals for 1980, 2000, and 2030 (projected) are approximately 3%, 6%, and 9%, respectively, of annual recharge in an average year, but 5%, 9%, and 13% of annual recharge, respectively, in a moderately dry year (10-year return period). In a moderate drought year (20-year return period), ground-water withdrawals are projected to reach 14% of annual recharge by 2030. In this preliminary model for the Monocacy basin, withdrawals from and discharges into streams were not taken into account, because they are expected to have little impact on ground-water levels.

Estimates of annual withdrawals and recharge, such as those given above, provide a perspective for the steady-state ground-water/stream flow model of average annual conditions described in this report. However, it should be noted that these figures are only a starting point for understanding water resources in the Monocacy basin. Aquifer levels and stream flows fall significantly below their annual averages in the months of summer and early autumn, so estimates of seasonal recharge and water demands are necessary to fully evaluate the impact of future demands. A seasonal water budget for three Monocacy tributary watersheds is given in Schultz, et al. (2005). A transient ground-water/stream flow model for the basin, to be completed in the next phase of this modeling effort, will provide more information on the effects of future demands on basin water resources during summertime conditions.

4.5. The Conceptual Model

A conceptual model provides a generalized description and interpretation of the hydrogeologic framework of the ground-water system. This conceptual model is then used to design the model grid, boundary conditions, stresses, and distribution of hydraulic properties within the computer model, providing the inputs that the model computer code uses in its simulations.

The aquifer system in the Monocacy River basin consists of multiple types of fractured bedrock overlain by a mantle of regolith. The thickness of the regolith varies greatly depending on the lithology of the underlying bedrock and the topographic setting. The density and size of the fractures varies with rock type, terrain, and relation to tectonic features such as faults. The primary source of recharge to the aquifer is precipitation infiltrating into the regolith and in turn recharging the network of fractures within the underlying bedrock. Ground-water discharges to the surface streams through streambed seepage and via discharge from springs. The ground-water divides outlining the ground-water basin are assumed to coincide with the surface water boundaries the Monocacy River basin.

An underlying assumption of ground-water flow simulation models is that the aquifer material behaves as a porous permeable media. For the most part, the bedrock in the study area has practically no primary porosity and permeability. Within most, if not all, of the bedrock units essentially all flow is within the secondary porosity; bedding surfaces, joints, fractures, fault zones, and solution enhanced openings (Nutter and Otton, 1969, Low and others, 2002). However, given the scale used in the model (500 m grid spacing) and the secondary porosity demonstrated by specific capacities of wells in these no-primary-porosity bedrock units (Nutter, 1974, Duigon and Dine, 1987, Low and Dugas, 1999), the assumption seems reasonable.

The conceptual model of the ground-water flow system is: a layer of permeable regolith overlying impermeable bedrock with a dense fracture network (the term “fractures” is used to refer to all secondary porosity providing openings, regardless of their source). The regolith receives water in the form of infiltration from precipitation and acts as a source of recharge to the fracture network within the bedrock. The density and size of the fractures due to weathering, and therefore the transmissivity of the fractures, decreases with depth. Where the elevation of the water table is above the elevation of the streambed, water discharges from the aquifer to the streams through the streambed, resulting in a “gaining” stream reach. Where the water level in the aquifer is below the elevation of the streambed, water will flow from the stream, recharging the aquifer through the streambed and resulting in a “losing” stream reach. There are also extraction wells removing water from the aquifer.

The ground-water flow model represents the regolith and the upper portion of the fractured bedrock as a single layer. The upper surface of this layer represents the ground surface. Processes occurring in the unsaturated zone are not represented in the model. The overall size of the modeled area requires large grid cells, 500 meters in both the X and Y dimensions. The horizontal dimensions of the grid cells and the high topographic relief in some portions of the basin results in differences in mean ground surface elevations of up to 100 meter between adjacent grid cells. Therefore, representing the regolith as a separate layer in the model would be computationally unworkable. The lower part of the fractured bedrock is represented in the second, lower layer. This layer has reduced hydraulic conductivities to represent the decease in fracture dimensions and density with depth. The reduction of fracture density and size is reflected in the number of wells completed at depths represented by the layers in the model. Approximately 80% of the wells in the basin are 150 m or less in depth and are represented in the upper model layer.

5. Flow Simulation Model

5.1. Model Design

The ground-water flow model was implemented using the US Geologic Survey’s MODFLOW-2000 (Harbaugh and others, 2000) finite difference model code. (References in this report to MODFLOW refer to the MODFLOW-2000 version of the USGS program.) The MODFLOW code was used because it is a widely used, well-tested and verified model code. It also has a wide variety of modules available to simulate various hydrogeologic conditions for predicting ground-water flow. The program Ground-water Vistas, version 4.2 (GWV), by Environmental Simulations, Inc., was used as a pre- and post-processor for MODFLOW. The parameter estimation program, PEST (Dogherty, 2004), and some of its associated utility programs were also used in the model calibration process.

5.2. Finite Difference Grid

MODFLOW is a computer program that simulates ground-water flow using the finite difference method. The grid used by MODFLOW is rectangular in the horizontal plane while the vertical dimension (thickness) of the layers can be varied spatially (Harbaugh and others, 2000). In the interest of minimizing the computational requirements for the model, grid spacing of 500 m in both horizontal dimensions was used, with the layer thickness selected to allow representation of the change in hydraulic parameters with depth.

The Mesozoic Lowland province within the Monocacy River basin is a half grabben with the western side dropped resulting in sedimentary layers dipping to the west-northwest. The boundary between the western Piedmont and the Blue Ridge provinces is formed by a high-angle Triassic age western border fault (Stose and Stose, 1946). The orientation of this fault and of the Mesozoic sedimentary basin in the Monocacy River basin provides a regional structural orientation to the rocks in these areas. This orientation is recreated in the flow model by the rotation of the model grid N22°E.

The grid is constructed with two layers; the uppermost layer representing the regolith/weathered bedrock part of the fractured bedrock aquifer and the second layer representing the part of the aquifer where the fractures are less numerous and well connected or where they pinch-out altogether with depth. The change in hydraulic characteristics with depth is gradual and therefore there is no transition marking one layer from the next, so the thickness of the layers was chosen somewhat arbitrarily, with an effort to allow for computational efficiency. The elevation of the top surface of the upper layer was taken from the USGS 10-m digital elevation model (DEM). The thickness of the layers was set to arbitrary constants of 150 m for layer 1 and 150 m for layer 2.

5.3. Boundary Conditions

The lateral extent of the modeled area was defined to coincide with the boundary of the watershed of the Monocacy River as defined by the boundary of the 8-digit hydrologic unit 02070009 (Seaber and others, 1987). It was assumed that the ground-water divide bounding the Monocacy basin coincides with the surface water divide of the basin. The active part of the model grid was shaped to reproduce the outline of the basin and was surrounded by inactive areas. The inactive areas within the model grid were defined with no-flow cells. By default, MODFLOW assigns a no-flow boundary to the bottom of the lowest layer.

5.4. Recharge

Recharge to ground-water was simulated as a flux applied to the top of each cell in the top layer (Harbaugh and others, 2000). The zones of recharge used in the model, shown in Figure 5, were created from the estimated mean annual natural ground-water recharge national dataset (Wolock, 2003). The dataset contains the geographic distribution of recharge values in a 1-kilometer resolution raster (grid) dataset and was computed by multiplying a grid of base-flow index (BFI) values by a grid of mean annual runoff values derived from a 1951-80 mean annual runoff contour map. The BFI is the ratio of base flow to total flow, expressed as a percentage. The BFI grid used was interpolated from BFI point values estimated for U.S. Geological Survey stream gages (Wolock, 2003a). The grid of average annual runoff in the conterminous United States, 1951-1980 is a thematic data layer representing average annual runoff, in inches per year, for the conterminous United States. The data reflects the runoff of tributary streams rather than in major rivers in order to represent more accurately the local or small scale variation in runoff with precipitation and other geographical characteristics. The range of recharge values within the Monocacy basin were grouped into 10 categories. Each of the 10 categories is represented by a zone in the model and range from approximately 5 in/yr to 10 in/yr.

Annual stream flows in the simulation year, 1980, are approximately 80% lower than long-term averages of annual flows, as shown in Table 1. To account for 1980 flow conditions, the set of recharge values that were computed from the national recharge dataset, as described above, were multiplied by an overall adjustment factor of 0.79, the ratio of 1980 to long-term average annual flow at the Jug Bridge gage. The resulting total annual recharge used in the model is 279 MGD. Recharge values given in Table 3 indicate that this simulated recharge represents conditions in a moderately dry year, consistent with the values for mean and 1980 stream flows appearing in Table 1.

5.5. Streams

The stream network in the Monocacy River basin is extensive. The National Hydrologic Dataset for 2003 (USGS-NHD, 2003) contains almost 2,200 miles of streams within the Monocacy River basin. Many of these streams are small first-order streams with very little flow and in times of drought, some may have no flow at all. The data burden to represent all these streams in a flow model is prohibitive. Therefore, the main stem of the Monocacy River and its major tributaries, with a total stream length of approximately 419 miles, are represented in the flow model. A total of 27 streams are included in the model using the MODFLOW Stream (STR) package (Prudic, 1989). Streambed elevation data required for the STR package were estimated using USGS topographic quadrangle maps. Stream width was also estimated from the USGS topographic maps and from reconnaissance observations of several of the streams in the basin. Each stream was subdivided into segments for which constant stream characteristics such as stream width, streambed thickness, slope, etc. could be reasonably assigned. This process resulted in 112 stream segments being defined for the model. During several reconnaissance visits to streams within the basin, it was observed that most of the streams had little sediment covering the streambeds, with gravel or larger clast sizes dominant. In almost all streams visited, bedrock was visible in a large percentage of the streambed. As a result of this observation, a streambed thickness of 0.5 m was assumed for the majority of the tributaries and 1.0 m for the mainstem of the Monocacy River. The depth of water in the streams was also observed at the visited streams and these observations were used to estimate typical stream depths by comparing the dimensions and the flows of these observed streams. Similarities of stream dimensions, general topography, drainage area, and underlying geology were used to estimate stream depths of the remaining stream segments in the model.

5.6. Extraction Wells

Extraction wells are represented in the model using the MODFLOW well boundary condition. The well data was taken from water use databases from the USGS of compiled data from the Maryland Department of the Environment and Pennsylvania Ground-water Information System. Individual well records were selected that were active during 1980. Some wells with active permits but showing zero withdrawal amounts in 1980 were eliminated from the list. In the process of locating the withdrawal wells in the flow model grid, there were instances where nests of multiple wells on a permit were located in the same model grid cell but at multiple depth intervals. In these cases the withdrawal amounts were totaled by depth to combine withdrawals within each model layer. In these cases, only one well is then represented in each appropriate model layer. The sum of withdrawals per day represented in the model is about 15 MGD (approximately 55,700 m^3), or about 4% of the daily recharge applied to the model. This represents the 1980 annual average daily withdrawal from ground-water for all use types.

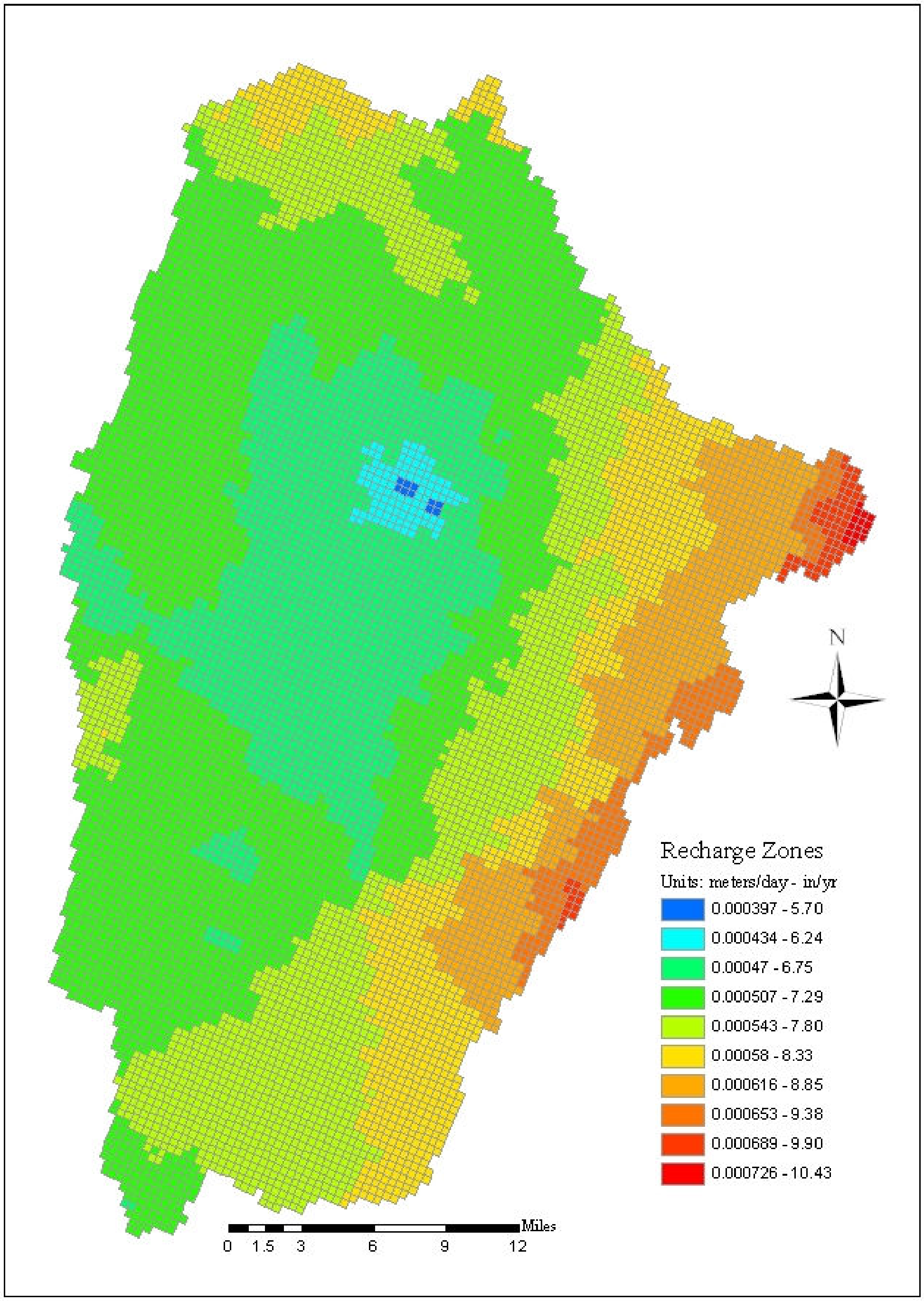

Figure 5. Model recharge zones.

5.7. Hydraulic properties

The distribution of hydraulic conductivity assigned to the model finite difference cells was based on the spatial distribution of a combination of physiographic province, physiographic province sub-unit, and lithology type as defined in the USGS National Water-Quality Assessment (NAWQA) Program's study of the Potomac River basin (Derosier and others, 1998). This dataset provided thirteen combinations of province, type, and sub-unit within the Monocacy River basin, as depicted in Figure 3. Each of these combinations could represent a hydrogeologic unit with unique hydraulic characteristics. Model calibration inputs were based on a grouping of these combinations into five zones, listed in Table 4 and shown in Figure 6. A discussion of these groupings is given below.

5.7.1. Results from analysis of available well data

In order to determine to what extent the thirteen hydrogeologic units shown in Figure 3 have significantly different in hydraulic conductivities, statistical analyses were done using available well data. Construction and specific capacity information pertaining to approximately 2,900 wells within the Monocacy basin was downloaded from the USGS GWSI database and analyzed. An estimate of transmissivity and hydraulic conductivity was calculated for each well record having sufficient data using the method described by Theis and others (1963) (Appendix B, Table B1). An Analysis of Variance (ANOVA) was performed on the resulting set of hydraulic conductivity values (Appendix C). The resulting hydraulic conductivity values were grouped by geologic formation included in each well record. Separate analyses were performed on the groups of formations making up the Blue Ridge, Piedmont Crystalline, Mesozoic Lowland, and Frederic Valley provinces. In most cases these analyses showed that the difference between the hydraulic conductivities of the geologic formations within the provinces was not statistically significant. The exception was in the Piedmont Crystalline province where the difference between the hydraulic conductivity values calculated for the Wakefiled Marble and several of the other formations differed significantly. However, the relative area of the study area underlain by Wakefield Marble (PD-C-PD in Figure 3) is small and it is completely surrounded by Piedmont Crystalline phyllites and schists. Given this, the small carbonate and metavolcanic units within the Piedmont Crystalline could be combined and represented in the model by a single hydraulic conductivity zone. Applying this method to the hydrogeologic units as identified in the NAWQA geology dataset, the thirteen formations were grouped into five zones used to represent the distribution of hydraulic conductivity in the model. These zones represent, broadly, the Mesozoic Lowland (zone 1), Blue Ridge (zone 2), Diabase intrusions within the Mesozoic Lowland (zone 3), Piedmont Lowland (zone 4), and Frederick Valley (zone 5) (see Figure 6). Summary statistics for hydraulic conductivities derived from Monocacy basin well data, grouped by zone, is given in Table 4.

5.7.2. Results from previous studies

In their summary of the geohydrology of southeastern Pennsylvania, Low, Hippe, and Yannacci (2002) provided tables of hydraulic conductivity values calculated from 1 hour or longer, single-well aquifer tests or specific capacity tests by use of a modified Theis formula (Theis and others, 1963) for 51 geohydrologic units, that include most of the units present in the Monocacy River basin. For units that make up the Mesozoic Lowland (model zone 1) the hydraulic conductivities reported range from 0 to 37 m/d with a median value of 0.21 m/d from a total of 282 well records. For units that make up the Blue Ridge (model zone 2) they report values from 0 to 8.5 with a median of 0.063 m/d from 52 well records. They reported hydraulic conductivity values for the Diabase (model zone 3) of from 0 to 51.8 m/d and 0.07 m/d for the median of values from 44 well records. Of the units Low and others (2002) described that make up the Piedmont Upland and Piedmont Lowland as named in Pa. the Marburg Schist also makes up a large portion of the Western Piedmont represented in the model. They reported hydraulic conductivities for the Marburg Schist from 0.003 to 2.9 m/d and a median of 0.076 m/d from records of 35 wells.

Gerhart and Lazorchick have developed ground-water models of parts of Lancaster and Berks Counties, Pa. (Gerhart and Lazorchick, 1984) and the lower Susquehanna River Basin in Pennsylvania and Maryland (Gerhart and Lazochick, 1988) wherein they represented some of the units present in the Monocacy River basin. In their model of the lower Susquehanna River Basin they used values of hydraulic conductivity ranging from 0.81 to 13.24 m/day for carbonate units, 0.04 to 1.02 m/d for Triassic sedimentary units, and 0.20 to 0.41 m/d for crystalline units.

In the Monocacy basin model described in this report, each hydraulic conductivity zone was assigned an initial value based on typical literature values for the lithology and from the estimated conductivities from the well data, as given in Table 4. These values were used as the starting conductivity values for the model calibration process.

Table 4. Hydraulic conductivity values (m/d) by model zone.

Model Zone	Zone Description	Range of values from previous studies	Range of values from analysis of available well data							Number of well records
			min	10^{th} %	25^{th} %	50^{th} %	75^{th} %	90^{th} %	max	
1	Mesozoic Lowland	0 to 37	0.0000	0.0029	0.0104	0.029	0.111	0.46	48	347
2	Blue Ridge	0 to 8.5	0.0000	0.0005	0.0031	0.027	0.202	0.67	41	238
3	Diabase	0 to 52	0.0001	0.0006	0.0055	0.011	0.050	0.54	0.82	11
4	Piedmont Crystalline	0.003 to 2.9	0.0000	0.0006	0.0032	0.019	0.111	0.40	13.1	607
5	Frederick Valley Carbonates	- -	0.0000	0.0012	0.0094	0.104	0.88	3.2	162	175

6. Model Calibration and Sensitivity Tests

6.1. Calibration Results

The calibration of the model was performed by comparing model predictions of aquifer levels, or "heads", and stream flows to values derived from available data for the simulation period, January 1 to December 31, 1980. Model input parameters representing aquifer and streambed conductivities were defined at a scale judged to be appropriate for a regional model. These parameters were adjusted to provide the best fit to available data by using the PEST parameter estimation program developed by John Doherty of Watermark Computing in native DOS mode and as implemented in Ground-water Vistas, and by trial-and-error methods.

The model's steady-state simulated ground-water levels were compared to mean 1980 water levels computed from data for 48 wells listed in Table A1 in Appendix A. These values represent approximate average, steady-state ground-water levels in the Monocacy basin in 1980. In addition, simulated flows in selected stream reaches were compared to mean annual base flows, computed using data from USGS stream gages. Base flow at each gage was computed as the average of the annual flow for 1980 multiplied by the BFI for that gage. The MODFLOW STR stream package reports the stream inflows to the reach, the flux between the stream and the aquifer (base flow if positive or recharge if negative), and the flows out of the reach to the next downstream reach. Comparison was made between the observed base flow at the twelve stream gages listed in Table 1 and the reported outflows of the stream reach in the model cell corresponding to the location of the stream gage.

Aquifer hydraulic conductivities were defined for each of the two model layers for each of the five hydrogeologic zones listed in Table 4 and shown in Figure 6. In order to reduce the number of degrees of freedom of the model, horizontal and vertical hydraulic conductivities for each layer were assumed to be related by a proportionality constant that was uniform throughout the model domain. Similarly, horizontal anisotropy was assumed to be uniform throughout each zone of hydraulic conductivity, being 1.0 in all zones except zones 1 and 5, representing the Mesozoic Lowland and Frederick Valley. The Mesozoic Lowland was assumed to have a horizontal anisotropy caused by its tilted sedimentary structure, the strike of which is in the Y-model direction through the rotation of the model grid by 22 degrees East of North. The anisotropy of the Frederick Valley hydraulic conductivity zone represents the solution-enhanced features present in this geologic unit. In this first phase of the Monocacy model, the values of horizontal anisotropy in these two zones were assumed to be equal.

Because no data were available on stream leakage rates, streambed conductivities for tributary stream segments were initially assigned to be equal to the assumed hydraulic conductivity of the underlying lithology. However, two separate streambed conductivities were defined for the river's main channel, one for the portion of the main channel located in the Mesozoic Lowland and one for the portion of the main channel located in the Frederick Valley. Streambed conductances were computed by MODFLOW from parameters representing streambed conductivities combined with a set of model inputs constructed from estimated values of stream reach lengths, widths, and bed thicknesses.

Final values of aquifer and stream bed conductivities for the calibrated model are given in Table 5 and shown in Figure 7. During the calibration process, a variety of sets of conductivity values were found which simulated observed aquifer water levels reasonably well. Of these solutions, sets of parameters with the lowest horizontal hydraulic conductivity values for zone 2 and zone 4 produced the best match to flow observations in Blue Ridge and Piedmont Province headwater streams. In some of the upland portions of the basin, especially in the Blue Ridge, the model tends to over-predict heads and under-predict stream flows, and, during model calibration, conductivities were adjusted to provide a reasonable tradeoff between these two effects. These inaccuracies are probably due to model grid size, which is relatively coarse compared with the size of the small catchments in the upland areas, and the fact that many of the smallest first and second order streams are not represented in the model, likely causing ground-water which discharges to unrepresented upland reaches to discharge, in the simulation, to reaches further downstream.

During calibration runs using PEST, the parameter representing horizontal anisotropy in zones 1 and 5 was found to be correlated with the value of hydraulic conductivity for the Frederick Valley carbonates, with high y to x anisotropies corresponding to high values for the carbonate hydraulic conductivity. Stream flows were also sensitive to the horizontal anisotropy parameter. In the final calibration the horizontal anisotropy was set at 3.3 for model zones 1 and 5, and at 1.0 elsewhere. The ratio of horizontal to vertical hydraulic conductivity was set at 10 throughout the model domain, and the final ratio of hydraulic conductivities in layer 1 to conductivities in layer 2 was set at 10.

The comparison of observed versus simulated heads resulting from the calibration of the model are shown in Figure 8 and in Table 6 and Table 7 below. The model simulated the heads within the Monocacy basin reasonably well, considering the regional nature of the model. The mean residual was –1.03 m, the standard deviation of residuals was 17.10 m and the correlation was 0.919. Of the 48 simulated heads 22 matched the observed heads within the estimated accuracy of the water level measurements (+/- 10 m). On the basis of lithologic units, as represented by zone of hydraulic conductivity in the model, the largest mean residual was in model zone 2, representing the Blue Ridge, with a mean residual of -17.5 m, while the smallest residual was in zone 1, the Mesozoic Lowland, with a 1.0 m mean residual.

Observed versus simulated stream base flows are compared in Table 8 and in Figure 9. From these results, it is evident that the calibrated model does a fairly good job in simulating 1980 flows. Stream flows for the larger drainage areas were found to be most sensitive to changes in recharge, so model errors in simulation of these flows reflect inaccuracies in recharge input values. Recharge in the model is distributed based on recharge zones constructed from the national natural ground-water recharge dataset, as discussed in Section 5.4. If the recharge in a particular watershed is anomalous due to non-represented characteristics such as land use, this deviation from the estimated recharge would affect the simulated results compared to the observed flows. Flow for the largest drainage area, upstream of the gage at Jug Bridge, is over-predicted by about 7%, indicating that overall model recharge inputs are slightly high. Base flow predictions for mid-sized streams are good except for the drainage area above the gage at Bridgeport, where flow is significantly over-predicted. This suggests that model recharge values over-estimate recharge in the Mesozoic Lowland, consistent with results from the recent study on annual and seasonal water budgets in the Monocacy basin (Schultz et al., 2005), which found that recharge in the Mesozoic Lowland was significantly lower than recharge in other portions of the basin.

The model tends to under-simulate base flows for stream reaches associated with small drainage areas. This under-prediction of flows in the headwater streams may be due to model cell size, which is relatively coarse compared with the size of the small catchments in the upland portions of the basin. It also may be due to the fact that many of the smallest first and second order streams were not represented in the model, in an effort to keep model complexity to a manageable level. Thus, ground-water which discharges to upland stream reaches which are not represented in the model may, in the simulation, be discharging to reaches further downstream.

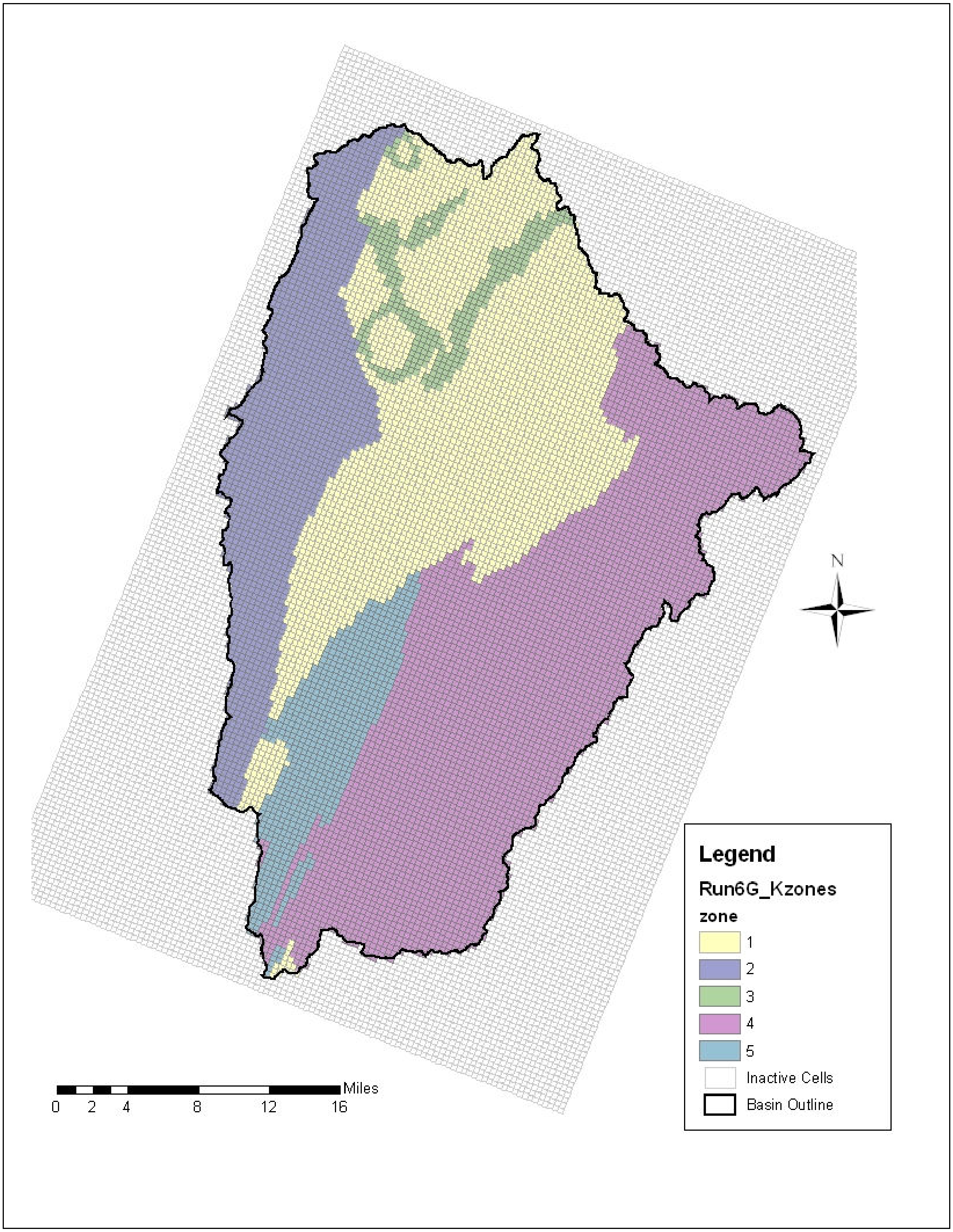

Figure 6. Model hydraulic conductivity zones.

Table 5. Final values of aquifer and streambed conductivities fpr the calibrated model.

Zone	Layer	Aquifer Conductivities			Streambed Conductivities	Unit
		Kx (m/day)	Ky (m/day)	Kz (m/day)	Tribs/Main Channel (m/day)	
1	1	0.89	2.9	0.089	0.19/0.001	Mesozoic Lowland Siliciclastics and Carbonates
2	1	0.038	0.038	0.0038	0.21	Blue Ridge Metavolcanics and Quartzites
3	1	0.10	0.10	0.010	0.18	Jurassic Diabase
4	1	0.57	0.57	0.057	0.10	Piedmont Crystalline
5	1	6.8	22.4	0.68	0.001/0.0044	Frederick Valley Carbonates
6	2	0.089	0.29	0.0089	NA	Mesozoic Lowland Siliciclastics and Carbonates
7	2	0.0038	0.0038	0.00038	NA	Blue Ridge Metavolcanics and Quartzites
8	2	0.010	0.010	0.0010	NA	Jurassic Diabase
9	2	0.057	0.057	0.0057	NA	Piedmont Crystalline
10	2	0.68	2.2	0.068	NA	Frederick Valley Carbonates

Note: Kx = hydraulic conductivity in the x-direction, i.e. along model rows, Ky = hydraulic conductivity in the y-direction, along model column, and Kz = hydraulic conductivity in the vertical direction.

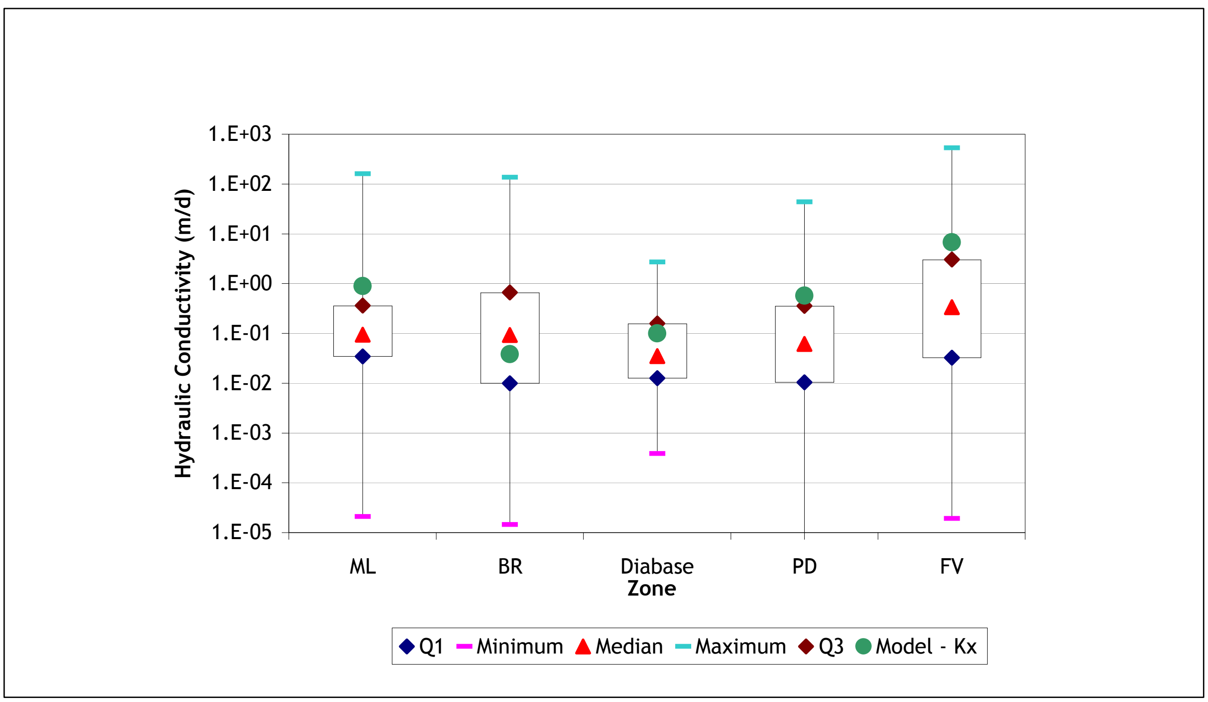

Figure 7. Comparisons of model hydraulic conductivities with values derived from well data

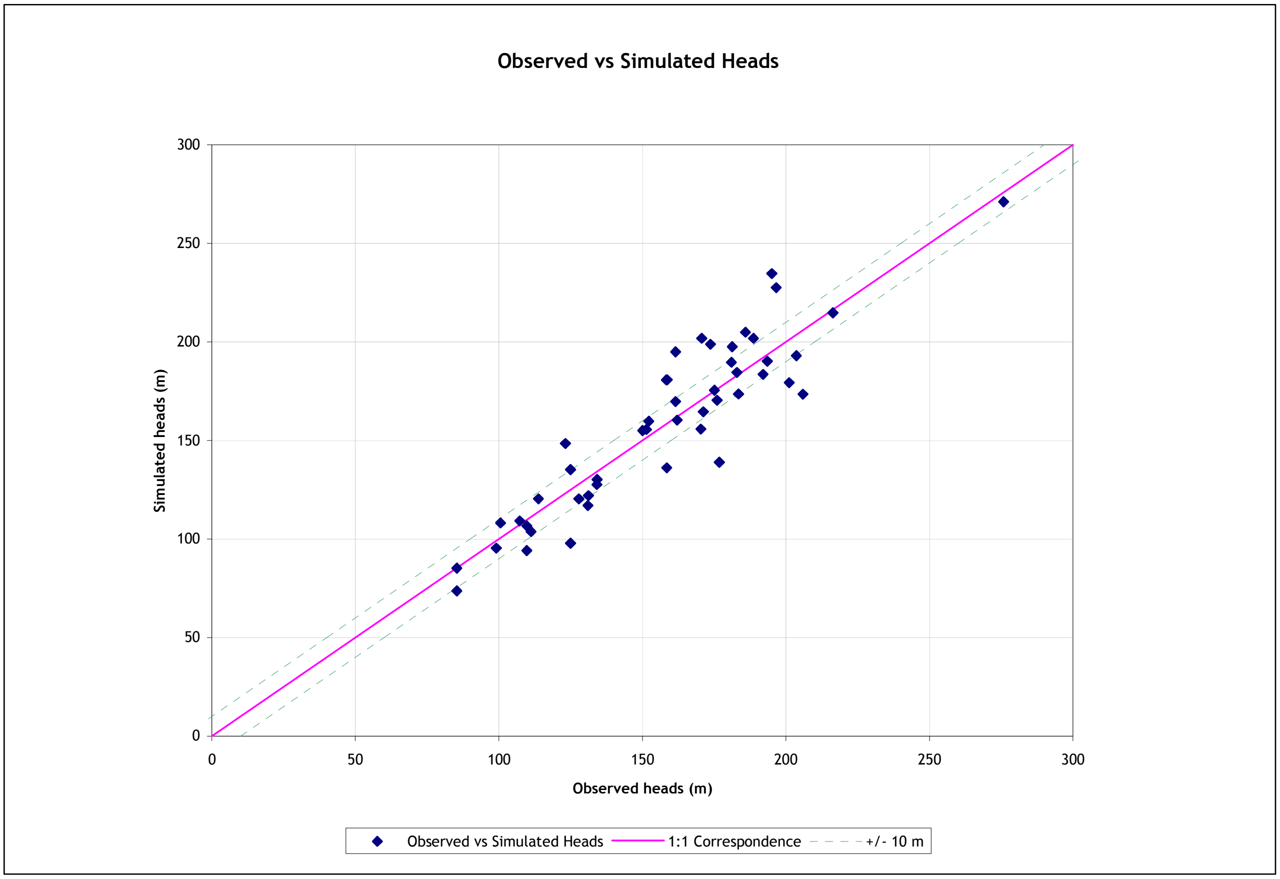

Figure 8. Observed versus simulated heads

Table 6. Mean residuals in meters of head targets by conductivity zone.

Hydraulic Conductivity zone	Number of targets	Mean residual
1 (Mesozoic Lowland)	21	0.991
2 (Blue Ridge)	2	-17.478
3 (Diabase)	7	1.842
4 (Piedmont Crystalline)	15	-4.240
5 (Frederick Valley)	3	5.175

Table 7. Calibration statistics of head targets (meters).

Residual Mean	-1.028
Residual Standard Deviation	17.096
Sum of Squared Residuals	13787
Absolute Residual Mean	13.165
Minimum Residual	-39.635
Maximum Residual	37.846
Range of Simulated Heads	197.5
Stdev/Range	0.087
Correlation	0.919
R^2	0.845

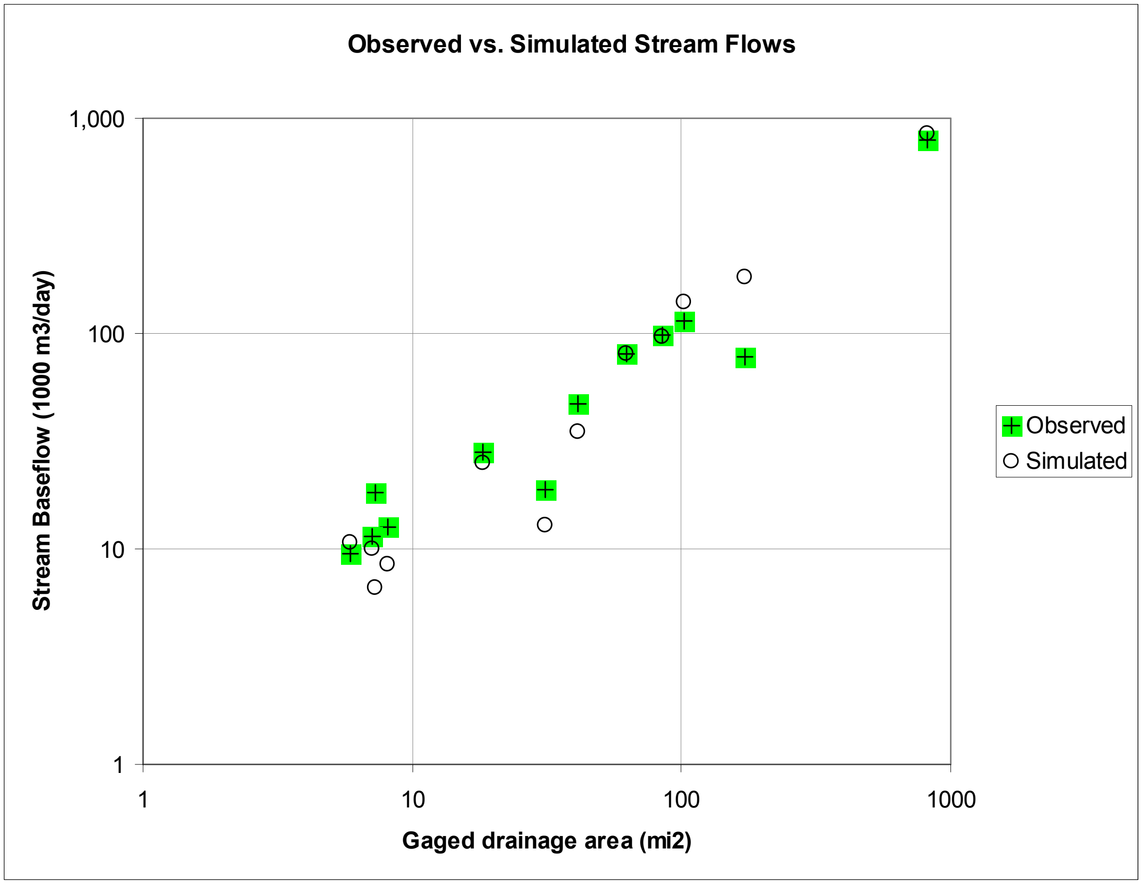

Figure 9. Observed versus simulated stream base flows for the year, 1980

Table 8. Observed versus simulated stream base flows (1000 m^3/day)

Station Name	Station ID	Drainage Area (mi^2)	Observed	Calibrated Model	Sensitivity Test 1	Sensitivity Test 2	Sensitivity Test 3
Monocacy River at Bridgeport, MD	01639000	173.0	78	182	182	163	182
Piney Creek near Taneytown, MD	01639140	31.3	18.7	12.8	12.8	9.8	12.8
Toms Creek at Emittsburg, MD	01639375	41.3	47	35	35	32	35
Big Pipe Creek at Bruceville, MD	01639500	102.0	115	139	138	124	137
Little Pipe Creek at Avondale, MD	01640000	8.1	12.6	8.4	8.9	7.3	10.8
Owens Creek at Lantz, MD	01640500	5.9	9.5	10.6	10.6	9.5	10.6
Hunting Creek near Thurmont, MD	01640975	7.1	11.5	10.0	10.0	8.8	10.0
Hunting Creek at Jimtown, MD	01641000	18.4	28	25	25	22	25
Fishing Creek near Lewistown, MD	01641500	7.3	18.3	6.6	6.6	5.7	6.6
Linganore Creek near Frederick, MD	01642500	82.3	98	97	97	87	97
Monocacy River at Jug Bridge Near Frederick, MD	01643000	817.0	790	847	847	757	847
Bennett Creek at Park Mills, MD	01643500	62.8	81	80	80	72	80

6.2. Sensitivity Tests

One of the potential uses of a regional ground-water/stream flow model is to estimate the impact of increased ground-water withdrawals on stream flows. Simulated flows in large streams are primarily sensitive to model recharge inputs, so simulated changes in net recharge are reasonably accurately reflected in changes in predicted base flows in the large streams. But flows in reaches of the smaller streams are less predictable when model inputs are changed. Several tests were conducted to improve understanding of the sensitivity of predicted flows in small streams to changes in model inputs.

The first test was done to help evaluate how uncertainty in stream bed elevations may affect predicted flows in small head water streams. Values for stream bed elevations used in the model, estimated from 7.5 minute, 1:24,000-scale USGS topographic quad maps, are likely to only be accurate to within 10 feet (3.05 m). In sensitivity test 1, stream bed elevations, and corresponding stages, for all reaches above the gages for Little Pipe Creek, Hunting Creek near Thurmont, and Fishing Creek were decreased by 4 meters. In particular, it was thought that results from this test might help explain the under-prediction of flows in small headwater streams. Resulting changes in predicted flows (see Table 8) were found to be insignificant in Hunting Creek and Fishing Creek, where simulated heads were already well above the bottom of the streambeds. In Little Pipe Creek simulated heads were close to or below the streambed in some reaches, and lowering the streambed did increase flow by 6%. However, this change was not significant compared to the discrepancy between predicted and observed flows in Little Pipe.

A second sensitivity test was done to evaluate how a potential increase in ground-water withdrawals would affect basin stream flows, and in particular, flows in small headwater streams. A hypothetical basin-wide increase in withdrawals of 10% was simulated by decreasing net recharge in all model cells by 10%. This change corresponds, roughly, to the increase in total basin withdrawals expected to occur between 1980 and 2030, as a percentage of annual recharge in a moderately dry year (see Table 2 and Table 3). Results of this test appear in the second from the last column of Table 8. Flows in most streams decreased by approximately 10%, but flows in several streams decreased by a somewhat greater amount. The greatest change occurred in Piney Creek, where predicted flows decreased by approximately 24%. Flows in Little Pipe Creek, Hunting Creek near Thurmont, and Fishing Creek decreased by approximately 13%, 12%, and 13%, respectively. The results of sensitivity test 2 are very preliminary, and only represent potential changes in mean annual flow conditions. Changes in steam flows due to potential increases in ground-water withdrawals must be evaluated using a transient flow model in order to simulate decreases in flow in summer months, when withdrawal rates represent a much greater fraction of seasonal recharge. However, results from the steady-state model give a preliminary indication that increased water demand due to population growth and development in the basin may have a significant impact on some headwater streams.

A third test was done to evaluate the effects of changes in streambed conductivities on flows in headwater stream segments. In the calibrated model, stream segment streambed conductivities were uniform throughout model conductivity zones (Figure 6), except that the main channel of the Monocacy River, which runs through zones 1 and 5, was assigned separate streambed conductivity values for each of these two zones. In sensitivity test 3, stream bed conductivities for all reaches above the gages for Little Pipe Creek, Hunting Creek near Thurmont, and Fishing Creek were increased by a factor of 10. This change increased simulated stream flow at the Little Pipe gage by almost 30%, but did not significantly change flows at Hunting or Fishing Creeks. Results of this test indicate that fine-tuning of headwater streambed conductivities can improve stream flow predictions in some headwater streams. Increased streambed conductivities in upper stream segments would also be consistent with typical physical streambed conditions, since bed material in upper stream reaches tends to be coarser than material present in lower reaches.

7. Conclusions

This report describes a steady-state ground-water/stream flow model for the Monocacy River basin, a 970 square mile drainage area underlain by fractured bedrock aquifers. The model was calibrated to mean annual well levels and stream base flows for the year, 1980, chosen to represent pre-development conditions in the basin. Ground-water flow is simulated using the USGS's MODFLOW-2000 finite difference model, using a two-layer grid with cell size of 150 meters in the vertical direction and 500 meters in both horizontal directions. Aquifer and streambed hydraulic conductivities are defined for each of five hydrogeologic zones, representing the predominant lithology of the Mesozoic Lowlands, Blue Ridge, diabase, Piedmont crystalline rock, and Frederick Valley carbonates.

7.1. Model performance

Predicted versus observed head and stream flow values match well in the calibrated model, considering the large-scale, regional nature of the model. In some of the upland portions of the basin, especially in the Blue Ridge, the model tends to over-predict heads and under-predict stream flows, and, during model calibration, conductivities were adjusted to provide a reasonable tradeoff between these two effects. These inaccuracies are probably due to model grid size, which is relatively coarse compared with the size of the small catchments in the upland areas. Also, since many of the smallest first and second order streams are not represented in the model, in an effort to keep model complexity to a manageable level, ground-water which discharges to unrepresented upland reaches may, in the simulation, be discharging to reaches further downstream.

Though flows in headwater streams with small drainage areas were quite sensitive to changes in conductivities, flows in the larger basin streams were not. Flows in all streams were sensitive to changes in recharge, and inaccuracies in flows in the large basin streams could be attributed to inaccuracies in model recharge inputs.

One of the potential uses for a regional ground-water/stream flow model is evaluation of the effects of future water demand scenarios on stream flows. In particular, headwater streams are thought to be most vulnerable to significant reductions in flow due to increased withdrawals. For this reason, sensitivity tests were conducted to assess the model's ability to accurately simulate stream base flows and its sensitivity to model inputs affecting base flow in headwater streams. In the first test, it was found that uncertainties in stream bed elevations did not have a significant effect on predicted flows in small head water streams. In the second test it was found that reductions in net recharge, simulating a hypothetical basin-wide increase in withdrawals of 10%, reduced stream flows throughout the basin by approximately 10%, and flow decreases in some but not all small headwater streams were somewhat greater than 10 %. Though flows in small headwater streams only appeared to be slightly more sensitive to changes in net recharge than flows in large streams in the steady-state model, it is likely that this would not be the case in a more realistic transient model. The third sensitivity test indicated that increasing streambed conductivities in upland stream reaches will likely improve model predictions of flows in some headwater streams. An increase of this nature would also be consistent with typical physical streambed conditions, since bed material in upper stream reaches tends to be coarser than material present in lower reaches.

7.2. Future needs

In this first phase of development of a predictive tool for water resources in the Monocacy basin, a number of approximations and simplification have been made that limit the model's predictive abilities. These limitations are summarized below, along with potential refinements that can be made to address them in future phases of model development.

- The current model is steady-state, and only capable of simulating average ground-water level and stream flow conditions. The steady-state model could be run using average summertime recharge inputs to investigate the impact of future demand on summertime water availability. However, a transient model, which could simulate seasonal changes in water levels and stream flow and incorporate the effects of aquifer storage, would provide a more realistic predictive tool.
- The current model does not represent many of the basin's small headwater streams. This approximation is appropriate for some uses of a regional-scale model. However, small headwater streams are thought to be particularly vulnerable to reductions in flow due to increased water demands. For use in studies with the objective of evaluating the impact of development on headwater streams, the model should be reconfigured with a denser stream network.
- The current model simulates the vertical component of flow using only two layers. In this configuration, the model is unable to accurately simulate deep ground-water

flow paths. The two-layer model is capable of simulating the effects of withdrawals in one of the basin's tributary watersheds on ground-water levels in that watershed, both upstream and downstream of the withdrawals, and on stream flows. But in order to accurately simulate the effects of ground-water withdrawals in one of the basin's tributary watersheds on aquifer levels in adjacent watersheds, the model must be reconfigured with more layers.

Computer simulation models rely on data to provide accurate model inputs. The Monocacy basin is fortunate to have relatively rich sets of historical data in comparison to other areas in the Potomac basin, and the Monocacy has adequate ground-water level, stream flow data, and withdrawal data to support development of a steady state model.

A transient ground-water flow model for the Monocacy basin will require data that provide a record of the seasonal variations of aquifer conditions throughout the calibration time period. The upcoming transient model of the Monocacy basin will likely use stream flow data as its primary calibration data set, and well-level data as its secondary calibration data set. Unfortunately, the number of locations where ground-water levels have been recorded on a monthly or seasonal basis is very limited. Also, little stream flow data exist for the Pennsylvania tributaries, Marsh, Rock, Alloway, and Middle Creek. Lastly, only four stream flow gages are currently operating in the Monocacy basin, which limits the possibility of verifying model predictions of the impact of current withdrawals.

To support continued development of a predictive tool to evaluate the impacts of future population growth on basin water resources, we recommend that the following two data collection efforts take place:

- Synoptic measurements should be made of ground-water levels in wells throughout the basin at a monthly time interval for a period at least one year. Ideally, measurements should be made at one upstream and one downstream location in each of the 15 significant tributaries of the basin.
- Stream flow gage stations should be re-established at locations where long-term historical flow records are available, such as Toms, Owens, Hunting, Fishing, Little Pipe, Big Pipe, and Piney Creeks. Gage stations should be established on Adams County, Pennsylvania tributaries, such as Marsh, Rock, Middle, and Alloway Creeks.

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Appendix A

Table A1. Well water level data for wells used as head targets.

Well Name	Period of Record	Number of measurements	Mode l Row	Model Column	Land Elevation (m)	Well depth (m)	Mean 1980 Water Level (btc m)	Water Level Elevation (m)
AD_443	05/01/80 - 05/01/80	1	45	51	140.62	47.85	9.46	131.16
AD_444	05/01/80 - 05/01/80	1	47	31	170.5	34.14	0.15	170.34
AD_445	05/01/79 - 05/01/80	2	36	31	161.35	59.44	10.7	151.43
AD_473	09/01/79 - 06/25/80	2	39	47	127.82	89	15.51	113.81
AD_474	02/01/80 - 02/01/80	1	39	47	140.01	60.96	12.19	127.82
AD_477	06/25/80 - 06/25/80	1	32	44	158.29	91.44	24.08	134.21
AD_479	06/25/80 - 06/25/80	1	23	45	164.39	29.57	2.25	162.13
AD_489	06/27/80 - 06/27/80	1	34	65	181.16	38.1	5.15	176.01
AD_491	06/24/80 - 06/24/80	1	28	33	159.21	38.1	9.14	150.07
AD_493	06/24/80 - 06/24/80	1	31	32	143.06	68.58	19.89	123.17
AD_495	06/24/80 - 06/24/80	1	38	30	164.4	47.85	12.18	152.22
AD_496	06/24/80 - 06/24/80	1	38	26	219.27	129.54	13.34	205.93
AD_497	12/01/79 - 12/01/79	1	30	25	164.1	38.1	5.49	158.61
AD_499	06/24/80 - 06/24/80	1	29	17	201.9	30.48	5.28	196.62
AD_505	06/25/80 - 06/25/80	1	37	22	182.69	70.1	1.72	180.97
AD_512	06/25/80 - 06/25/80	1	44	22	176.6	60.96	18.26	158.34
AD_514	06/25/80 - 06/25/80	1	45	24	194.88	84.12	11.38	183.5
AD_515	09/01/79 - 06/26/80	2	47	38	167.75	75.29	7.63	158.45
AD_518	08/01/79 - 06/26/80	2	45	19	194.28	24.99	13.72	182.98
AD_587	06/27/80 - 06/27/80	1	18	47	179.63	21.34	4.51	175.12
FR_Ae_56	12/03/80 - 12/03/80	1	64	31	213.36	213.36	18.29	195.07
FR_Ae_60	12/09/80 - 12/09/80	1	62	33	146.3	144.78	12.19	134.11
FR_Bf_29	07/11/80 - 07/11/80	1	72	41	140.21	38.1	9.14	131.06
FR_Cd_43	08/29/80 - 08/29/80	1	89	27	131.06	99.06	6.1	124.97
FR_Ce_71	09/02/80 - 09/02/80	1	100	41	94.49	122.83	9.14	85.34
FR_Ce_80	10/06/80 - 10/06/80	1	92	39	143.26	91.44	18.29	124.97
FR_Ce_83	07/10/80 - 07/10/80	1	88	37	112.78	68.58	1.52	111.25
FR_Ce_85	06/23/80 - 06/23/80	1	85	37	124.97	68.58	17.68	107.29
FR_Ce_90	10/20/80 - 10/20/80	1	87	35	118.87	63.09	9.14	109.73
FR_Cf_48	10/20/80 - 11/03/83	16	91	50	103.63	21.34	4.57	99.06
FR_Cg_1	06/28/46 - 02/26/04	746	86	62	182.88	13.11	11.63	171.25
FR_Dd_207	12/11/80 - 12/11/80	1	120	26	289.56	44.2	13.72	275.84
FR_De_106	09/08/80 - 09/08/80	1	106	36	118.87	48.77	9.14	109.73
FR_Df_30	05/22/80 - 05/22/80	1	101	61	182.88	45.72	6.1	176.78
FR_Dg_27	09/16/80 - 09/16/80	1	94	76	167.64	38.1	6.1	161.54

Well Name	Period of Record	Number of measurements	Mode l Row	Model Column	Land Elevation (m)	Well depth (m)	Water Level (btc m)	Water Level Elevation (m)
FR_Dh_42	03/25/80 - 03/25/80	1	99	85	234.7	91.44	18.29	216.41
FR_Dh_52	03/25/80 - 03/25/80	1	90	85	216.41	138.68	30.48	185.93
FR_Dh_53	01/17/80 - 01/17/80	1	90	84	207.26	44.2	18.59	188.67
FR_Dh_9	06/06/80 - 06/06/80	1	90	78	201.17	36.58	9.14	192.02
FR_Ee_120	05/06/80 - 05/06/80	1	125	49	88.39	68.58	3.05	85.34
FR_Eg_20	07/17/80 - 07/17/80	1	108	79	179.83	44.2	18.29	161.54
FR_Eg_21	11/07/80 - 11/07/80	1	108	80	188.98	60.96	7.62	181.36
FR_Eg_23	01/30/80 - 01/30/80	1	107	80	204.22	42.67	10.67	193.55
FR_Eg_27	11/11/80 - 11/11/80	1	107	81	219.46	50.29	15.85	203.61
FR_Eh_17	09/30/80 - 09/30/80	1	107	83	185.93	45.72	12.19	173.74
FR_Fd_87	03/11/80 - 03/11/80	1	137	57	112.78	85.34	12.19	100.58

Note: Water levels are depth to water below the top of the well casing (btc) in meters.

Appendix B

Table B1. Calculated hydraulic conductivities (in ft/dy) from specific capacity test data using the equation of Theis and others, 1963.

Formation Code	n	Min	P10	P25	Median	P75	P90	Max	Formation Name
(none)	70	1.6E-04	0.0035	0.0081	0.03	0.36	2.4	20	none provided
ANTM	7	0.019	---	0.082	2.17	2.6	---	7.1	Antietam Formation
BCMV	2	0.0026	0.14	0.36	0.71	1.1	1.3	1.4	Bachman Valley Formation
BLDR	1	3.2	---	---	3.2	---	---	---	Bolder Gneiss of the Wissahickon Formation
BLMR	250	1.0E-04	0.0020	0.010	0.04	0.21	1.0	13	Baltimore Gneiss
CCKV	17	0.0022	0.0046	0.029	0.24	1.0	2.6	3.2	Cockeysville Marble
CTCN	120	1.4E-05	0.0030	0.012	0.12	0.73	2.2	30	Catoctin Metabasalt
DIBS	10	3.9E-04	0.0019	0.015	0.03	0.13	1.9	2.7	Diabase Dikes and Sills
ELCC	1	7.4E-04	---	---	7.4E-04	---	---	---	Ellicott City Granodiorite
FDCK	117	1.7E-05	0.0015	0.013	0.19	2.0	6.4	153	Frederick Limestone
GBRG	169	2.1E-05	0.0081	0.024	0.07	0.23	1.1	159	Gettysburg Shale
GLFD	34	3.9E-04	0.0016	0.0066	0.03	0.22	0.54	8.5	Guildford Quartz Monzonite
GLLS	106	1.3E-05	3.1E-04	0.0011	0.01	0.09	0.40	21	Gillis Formation
GLRM	4	0.0023	---	---	0.04	---	---	0.1	Glenarm Series
GROV	73	2.0E-05	0.0021	0.041	0.38	3.7	18	533	Grove Limestone
HRPR	46	1.1E-04	0.0006	0.0017	0.03	0.19	1.3	7.6	Harpers Formation
IJMV	168	2.0E-05	0.0020	0.0063	0.04	0.33	1.1	21	Ijamsville Formation
LBRN	5	0.0058	---	0.014	0.04	0.11	---	3.9	Libertytown Metarhyolite
LCRV	378	1.0E-05	7.5E-04	0.0032	0.015	0.07	0.48	53	Lock Raven Schist
LPLC	1	3.1	---	---	3.1	---	---	---	Lower Pelitic Schist of the Wissahickon Formation
LUDN	16	0.0026	0.0036	0.0055	0.02	0.15	0.50	3.9	Loudoun Formation
MNWS	10	0.0203	0.403	0.92	1.62	3.58	7.77	17	Alluvial cones of mountain wash
MRBG	53	0.0000	0.0073	0.028	0.10	0.61	2.7	43	Marburg Schist
MRGR	284	7.7E-05	0.0030	0.012	0.06	0.22	0.81	11	Morgan Run Formation
MTRL	25	0.035	0.094	0.157	0.35	1.3	2.3	8.2	Metarhyolite and assoc. pyroclastic sediments
NOXF	138	2.3E-04	0.012	0.043	0.11	0.36	0.92	14	New Oxford Formation
NOXFB	13	0.24	0.29	0.58	1.32	2.7	7.4	21	New Oxford Basal Conglomerate
OELL	53	4.6E-04	0.0045	0.012	0.04	0.45	1.2	11	Oella Formation
PCMB	1	1.1	---	---	1.1	---	---	---	Undifferentiated Precambrian rocks
PLGV	33	0.016	0.057	0.087	0.22	0.89	4.0	9.1	Pleasant Grove Schist
PNRN	14	0.0011	0.067	0.089	0.23	0.93	1.5	8.4	Piney Run Formation
PRTB	185	7.5E-04	0.0051	0.017	0.08	0.46	1.3	7.0	Prettyboy Schist
ROCK	27	0.0044	0.019	0.070	0.22	0.51	3.3	10	Bedrock

Formation Code	n	Min	P10	P25	Median	P75	P90	Max	Formation Name
SKVLS	24	0.0024	0.0048	0.012	0.04	0.16	0.32	5.7	Sykesville Formation-Gneiss
SMCK	13	0.0110	0.022	0.046	0.10	0.20	0.54	0.7	Sams Creek Metabasalt
STRS	42	4.0E-04	0.0053	0.010	0.03	0.22	0.55	2.2	Setters Formation
TMSN	2	0.028	---	---	0.55	---	---	1.1	Tomstown dolomite
UMFC	37	0.0021	0.0039	0.012	0.05	0.14	0.51	1.2	Ultramafic Rocks
UPPC	17	2.2E-04	0.014	0.077	0.37	0.75	2.5	8.1	Upper Pelitic Schist of the Wissahickon Formation
URBN	26	4.0E-04	0.0069	0.017	0.13	0.48	2.4	5.1	Urbana Formation
WDCK	3	0.0037	---	---	0.04	---	---	1.1	Woodstock Quartz Monzonite
WKFD	5	0.0021	---	0.95	0.96	12	---	15	Wakefield Marble
WVRN	20	1.7E-04	3.8E-04	0.0055	0.04	0.11	0.28	136	Weverton Formation

Note: The units shown in the table are feet per day. A value of 0.01 was used for S (specific capacity) in all calculations using the Theis equation.

Appendix C

The following is a discussion of the ANOVA statistical tests of similarity performed on calculated hydraulic conductivities from specific capacity test data.

The equation to calculate transmissivity from specific capacity of Theis and others (1963), was used to calculate transmissivity (ft^2/dy) and hydraulic conductivity (ft/dy) from the specific capacity test data on approximately 2,900 well records in the GWIS dataset. A storage coefficient value of 0.01 was used for all calculations. This data was then tested to determine if the calculated hydraulic conductivities were sufficiently similar to allow combining the representation of the geologic units in the model. The ANOVA analysis tool in Microsoft Excel was used to compare the calculated conductivity values for each formation as identified in the GWIS data. The constituent formations were tested in the major lithologic units represented in the model; the Piedmont Crystalline, Mesozoic Lowland, Blue Ridge, and the Frederick Valley. First the conductivities for the formations were tested using the null hypothesis that the differences in the mean conductivities were zero. If the null hypothesis was rejected, individual pairs of formations were compared.

Zone 2 Anova: Single Factor

alpha = 0.05 for all ANOVA analyses

H_o: mean Ks (ft/dy) of formations making up Zone 2 are the same

SUMMARY

Groups	*Count*	*Sum*	*Average*	*Variance*
ANTM	7	14.59696	2.085279	6.323015
CTCN	120	129.851	1.082092	11.09
HRPR	46	20.21682	0.439496	1.542479
LUDN	16	5.443287	0.340205	0.933615
MTRL	25	28.22901	1.12916	3.598178
WVRN	20	138.5465	6.927325	928.0059

ANOVA

Source of Variation	*SS*	*df*	*MS*	*F*	*P-value*	*F crit*
Between Groups	688.3159	5	137.6632	1.638203	0.150812	2.25364296
Within Groups	19159.53	228	84.03304			
Total	19847.85	233				

Do not reject Ho: calculated $F < F_{crit}$

Geologic Unit Codes: ANTM = Antietam, CTCN = Catoctin, HRPR = Harpers, LUDN = Loudoun, MTRL = Metarhyolite, and WVRN = Weverton Formations.

Zone 5 Anova: Single Factor

H_o: mean Ks (ft/dy) of formations making up Zone 5 are the same

SUMMARY

Groups	*Count*	*Sum*	*Average*	*Variance*
FDCK	117	432.8504	3.699576	237.1363
GROV	73	1127.763	15.44881	4423.244

ANOVA

Source of Variation	*SS*	*df*	*MS*	*F*	*P-value*	*F crit*
Between Groups	6205.464	1	6205.464	3.371936	0.067895	3.89139793
Within Groups	345981.4	188	1840.327			
Total	352186.9	189				

Do not reject H_o: calculated $F < F_{crit}$

Geologic Unit Codes: FDCK = Frederick Limestone, GROV = Grove Limestone.

Zone 4 Anova: Single Factor

H_o: mean Ks (ft/dy) of formations making up Zone 4 are the same

SUMMARY

Groups	*Count*	*Sum*	*Average*	*Variance*
WKFD	5	28.63967	5.727933	49.77256
BCMV	2	1.42148	0.71074	1.002967
GLLS	106	46.13425	0.435229	4.727696
GLRM	4	0.206095	0.051524	0.002672
IJMV	168	84.63611	0.503786	3.673576
LBRN	5	4.06461	0.812922	2.965583
MRBG	53	88.4512	1.668891	39.06815
PLGV	33	39.099	1.184818	5.10135
PRTB	184	86.23332	0.468659	0.894704
SMCK	13	2.443683	0.187976	0.048378
URBN	26	17.71062	0.681178	1.563235

ANOVA

Source of Variation	*SS*	*df*	*MS*	*F*	*P-value*	*F crit*
Between Groups	212.1355	10	21.21355	3.353075	0.000291	1.84679544
Within Groups	3720.038	588	6.326596			
Total	3932.174	598				

reject H_o: calculated $F > F_{crit}$

Geologic Unit Codes: WKFD = Wakefield Marble, BCWV = Bachman Valley, GLLS = Gillis Formation, GLRM = Glenarm Series, IJMV = Ijamsville, LBRN = Libertytown Metarhyolite, MRBG = Marburg Schist, PLGV = Pleasant Grove Schist, PRTB = Prettyboy Schist, SMCK = Sams Creek Metabasalt, and URBN = Urbana.

As the null hypothesis was rejected, comparisons were performed on individual pairs of formations.

	WKFD	BCMV	GLLS	GLRM	IJMV	LBRN	MRBG	PLGV	PRTB	SMCK	URBN
WKFD		0	1	0	1	0	0	0	1	0	0
BCMV	0		0	0	0	0	0	0	0	0	0
GLLS	1	0		0	0	0	0	0	0	0	0
GLRM	0	0	0		0	0	0	0	0	0	0
IJMV	1	0	0	0		0	0	0	0	0	0
LBRN	0	0	0	0	0		0	0	0	0	0
MRBG	0	0	0	0	0	0		0	0	0	0
PLGV	0	0	0	0	0	0	0		0	0	0
PRTB	1	0	0	0	0	0	0	0		0	0
SMCK	0	0	0	0	0	0	0	0	0		0
URBN	0	0	0	0	0	0	0	0	0	0	

In the table above, a “1” indicates there is a difference in the mean hydraulic conductivities of the units being compared and a “0” indicates no difference in mean hydraulic conductivities.
Conclusion: there is a statistical difference between the mean hydraulic conductivities of WKFD and GLLS, IJMV, and PRTB but there is no between the mean hydraulic conductivities of WKFD and the other formations. Also, there is no statistical difference between the mean hydraulic conductivities of the formations other than WKFD.

Zone 1 Anova: Single Factor

H_o: mean Ks (ft/dy) of formations making up Zone 1 are the same

SUMMARY

Groups	*Count*	*Sum*	*Average*	*Variance*
GBRG	169	281.0106	1.662785	157.056
MNWS	10	35.92137	3.592137	26.5689
NOXF	138	80.18116	0.581023	2.621227
NOXFB	13	42.5852	3.275784	31.5465
DIBS	10	4.770768	0.477077	0.907585

ANOVA

Source of Variation	*SS*	*df*	*MS*	*F*	*P-value*	*F crit*
Between Groups	203.6201	4	50.90502	0.623053	0.646359	2.39860605
Within Groups	27370.37	335	81.70259			
Total	27573.99	339				

Do not reject H_o: calculated $F < F_{crit}$

Geologic Unit Codes: GBRG = Gettysburg Shale, MNWS = Mountain Wash, NOXF = New Oxford Formation, NOXFB = Basal Conglomerate of the New Oxford, DIBS = Diabase.

Although this analysis showed there is no statistical difference in the mean hydraulic conductivities of the Diabase and the other units of the Mesozoic Lowland, due to the size of the area represented in the model, the Diabase was represented separately in the model.

Appendix D

Table D1. List of Monocacy basin extraction wells with withdrawals in 1980.

Well Number	Well Depth (m)	1980 Withdrawal (m^3/d)	Row	Column	Land Surface Elev (m)	Model Layer	Latitude	Longitude
4	52.27	8.998	128	68	161.24	1	39.34861	-77.28639
17	25.91	52.882	104	40	84.68	1	39.48833	-77.36111
20	121.92	543.191	122	32	154.79	1	39.43028	-77.47056
22	138.99	36.162	54	63	152.20	1	39.66167	-77.17028
23	150.88	49.468	68	73	160.80	2	39.59028	-77.13472
24	121.31	49.468	68	73	160.80	1	39.59028	-77.13472
25	121.92	49.468	68	73	160.80	1	39.59028	-77.13472
32	152.40	79.547	55	26	139.94	2	39.71889	-77.36583
39	60.96	163.435	89	48	124.38	1	39.53250	-77.31528
40	90.53	163.435	89	48	124.38	1	39.53250	-77.31528
41	182.88	163.435	89	48	124.38	2	39.53250	-77.31528
42	144.78	163.435	89	48	124.38	1	39.53250	-77.31528
43	182.88	163.435	89	48	124.38	2	39.53250	-77.31528
44	38.10	163.435	89	48	124.38	1	39.53250	-77.31528
45	144.78	163.435	89	48	124.38	1	39.53250	-77.31528
46	274.32	163.435	89	48	124.38	2	39.53250	-77.31528
47	18.90	163.435	89	48	124.38	1	39.53250	-77.31528
48	205.74	163.435	89	48	124.38	2	39.53250	-77.31528
49	152.40	163.435	89	48	124.38	2	39.53250	-77.31528
50	188.98	163.435	89	48	124.38	2	39.53250	-77.31528
51	180.14	163.435	89	48	124.38	2	39.53250	-77.31528
53	23.16	5.303	120	26	289.79	1	39.36444	-77.49500
54	137.16	5.303	120	26	289.79	1	39.36444	-77.49500
55	44.20	5.303	120	26	289.79	1	39.36444	-77.49500
56	9.14	5.303	120	26	289.79	1	39.36444	-77.49500
65	49.07	21.410	60	26	208.99	1	39.69694	-77.39417
66	76.20	21.410	60	26	208.99	1	39.69694	-77.39417
67	96.01	21.410	60	26	208.99	1	39.69694	-77.39417
68	111.25	21.410	60	26	208.99	1	39.69694	-77.39417
69	17.68	21.410	60	26	208.99	1	39.69694	-77.39417
70	29.87	21.410	60	26	208.99	1	39.69694	-77.39417
71	182.88	21.410	60	26	208.99	2	39.69694	-77.39417
73	19.51	1.745	86	55	163.03	1	39.54611	-77.27278
74	47.55	1486.914	137	44	90.01	1	39.34806	-77.43833
75	26.52	1486.914	137	44	90.01	1	39.34806	-77.43833
76	50.60	1486.914	137	44	90.01	1	39.34806	-77.43833
83	121.92	15.168	119	67	165.58	1	39.38417	-77.27583

Well Number	Well Depth (m)	1980 Withdrawal (m^3/d)	Row	Column	Land Surface Elev (m)	Model Layer	Latitude	Longitude
84	94.49	15.168	119	67	165.58	1	39.38417	-77.27583
85	94.49	15.168	119	67	165.58	1	39.38417	-77.27583
86	76.20	15.168	119	67	165.58	1	39.38417	-77.27583
87	91.44	15.168	119	67	165.58	1	39.38417	-77.27583
88	91.44	15.168	119	67	165.58	1	39.38417	-77.27583
95	23.77	573.100	132	45	78.37	1	39.36472	-77.42083
96	41.15	573.100	132	45	78.37	1	39.36472	-77.42083
97	25.60	573.100	132	45	78.37	1	39.36472	-77.42083
137	91.44	102.168	117	55	135.51	1	39.40611	-77.29361
138	92.05	102.168	117	55	135.51	1	39.40611	-77.29361
139	45.72	102.168	117	55	135.51	1	39.40611	-77.29361
140	91.44	102.168	117	55	135.51	1	39.40611	-77.29361
141	91.44	102.168	117	55	135.51	1	39.40611	-77.29361
169	65.23	48.381	108	29	146.84	1	39.49083	-77.45694
175	53.34	37.854	114	37	116.40	1	39.45528	-77.40722
176	76.20	37.854	114	37	116.40	1	39.45528	-77.40722
177	76.20	37.854	114	37	116.40	1	39.45528	-77.40722
184	60.96	1764.168	72	90	167.94	1	39.54389	-77.04944
185	60.96	1764.168	72	90	167.94	1	39.54389	-77.04944
186	60.96	1764.168	72	90	167.94	1	39.54389	-77.04944
187	54.86	32.888	128	43	85.15	1	39.39778	-77.39611
188	19.81	32.888	128	43	85.15	1	39.39778	-77.39611
205	51.46	90.055	88	47	129.17	1	39.54611	-77.31889
210	106.07	3.774	92	52	144.55	1	39.52139	-77.30833
212	75.41	591.145	96	52	114.49	1	39.50778	-77.30472
238	68.35	3.831	116	85	229.51	1	39.36806	-77.16972
250	79.25	816.740	113	85	197.25	1	39.38167	-77.16611
251	54.86	816.740	113	85	197.25	1	39.38167	-77.16611
252	45.72	816.740	113	85	197.25	1	39.38167	-77.16611
253	42.67	816.740	113	85	197.25	1	39.38167	-77.16611
254	55.47	816.740	113	85	197.25	1	39.38167	-77.16611
255	26.52	816.740	113	85	197.25	1	39.38167	-77.16611
256	83.82	816.740	113	85	197.25	1	39.38167	-77.16611
257	91.44	355.208	64	32	165.12	1	39.67778	-77.35167
258	213.36	355.208	64	32	165.12	2	39.67778	-77.35167
259	129.54	355.208	64	32	165.12	1	39.67778	-77.35167
268	106.68	168.901	81	91	228.29	1	39.51083	-77.06361
269	106.68	168.901	81	91	228.29	1	39.51083	-77.06361

Well Number	Well Depth (m)	1980 Withdrawal (m^3/d)	Row	Column	Land Surface Elev (m)	Model Layer	Latitude	Longitude
270	106.68	168.901	81	91	228.29	1	39.51083	-77.06361
309	152.40	150.417	108	45	91.15	2	39.46639	-77.38944
310	178.31	150.417	108	45	91.15	2	39.46639	-77.38944
311	91.44	150.417	108	45	91.15	1	39.46639	-77.38944
324	74.37	1243.970	56	61	155.97	1	39.65889	-77.17750
325	68.58	1243.970	56	61	155.97	1	39.65889	-77.17750
326	119.79	1243.970	56	61	155.97	1	39.65889	-77.17750
327	182.88	1243.970	56	61	155.97	2	39.65889	-77.17750
328	190.50	1243.970	56	61	155.97	2	39.65889	-77.17750
329	179.83	1243.970	56	61	155.97	2	39.65889	-77.17750
330	176.78	1243.970	56	61	155.97	2	39.65889	-77.17750
331	187.45	1243.970	56	61	155.97	2	39.65889	-77.17750
332	132.59	1243.970	56	61	155.97	1	39.65889	-77.17750
333	144.78	1243.970	56	61	155.97	1	39.65889	-77.17750
334	120.70	1243.970	56	61	155.97	1	39.65889	-77.17750
336	60.05	53.382	67	96	188.09	1	39.55750	-77.01056
337	48.77	53.382	67	96	188.09	1	39.55750	-77.01056
340	137.16	499.754	80	26	141.37	1	39.61444	-77.41167
341	32.00	499.754	80	26	141.37	1	39.61444	-77.41167
342	31.09	499.754	80	26	141.37	1	39.61444	-77.41167
343	89.61	499.754	80	26	141.37	1	39.61444	-77.41167
344	32.00	499.754	80	26	141.37	1	39.61444	-77.41167
345	30.48	499.754	80	26	141.37	1	39.61444	-77.41167
346	121.92	499.754	80	26	141.37	1	39.61444	-77.41167
347	91.44	499.754	80	26	141.37	1	39.61444	-77.41167
348	91.44	499.754	80	26	141.37	1	39.61444	-77.41167
349	25.91	499.754	80	26	141.37	1	39.61444	-77.41167
350	21.95	499.754	80	26	141.37	1	39.61444	-77.41167
351	38.10	499.754	80	26	141.37	1	39.61444	-77.41167
352	29.57	499.754	80	26	141.37	1	39.61444	-77.41167
353	47.24	499.754	80	26	141.37	1	39.61444	-77.41167
354	67.06	499.754	80	26	141.37	1	39.61444	-77.41167
368	131.67	99.439	78	19	284.78	1	39.64722	-77.47917
369	109.73	99.439	78	19	284.78	1	39.64722	-77.47917
370	83.82	99.439	78	19	284.78	1	39.64722	-77.47917
371	94.49	99.439	78	19	284.78	1	39.64722	-77.47917
372	243.84	99.439	78	19	284.78	2	39.64722	-77.47917
373	152.40	99.439	78	19	284.78	2	39.64722	-77.47917

Well Number	Well Depth (m)	1980 Withdrawal (m^3/d)	Row	Column	Land Surface Elev (m)	Model Layer	Latitude	Longitude
375	114.30	256.007	74	69	127.29	1	39.56833	-77.18083
376	92.05	256.007	74	69	127.29	1	39.56833	-77.18083
377	73.76	256.007	74	69	127.29	1	39.56833	-77.18083
378	312.42	256.007	74	69	127.29	2	39.56833	-77.18083
379	51.82	256.007	74	69	127.29	1	39.56833	-77.18083
388	50.29	1678.100	105	45	89.38	1	39.48000	-77.36111
389	21.95	1678.100	105	45	89.38	1	39.48000	-77.36111
390	91.44	1678.100	105	45	89.38	1	39.48000	-77.36111
391	36.58	1678.100	105	45	89.38	1	39.48000	-77.36111
392	76.20	1678.100	105	45	89.38	1	39.48000	-77.36111
401	93.57	25.567	67	91	170.17	1	39.56583	-77.03528
402	24.38	25.567	67	91	170.17	1	39.56583	-77.03528
8743	73.15	21.655	40	18	188.99	1	39.79583	-77.37917
8808	80.77	125.945	25	19	192.66	1	39.85944	-77.34167
8838	121.92	25.305	17	14	231.02	1	39.90083	-77.35056