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# Hydrogeology and Water Quality of the Dublin and Midville Aquifer Systems at Waynesboro, Burke County, Georgia, 2011

Scientific Investigations Report 2013–5026

U.S. Department of the Interior
U.S. Geological Survey

**Cover photograph.** Well field just north of City of Waynesboro during the 24-hour aquifer test at the new production well, August 23, 2011. Photo by Michael D. Hamrick, USGS.

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KEN SALAZAR, Secretary

**U.S. Geological Survey**
Suzette M. Kimball, Acting Director

U.S. Geological Survey, Reston, Virginia: 2013

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Amentrout Matheny Thurmond, consulting group, and Rowe Drilling Company designed and constructed the production well for the City of Waynesboro. The author appreciates the support of Charles Armentrout (Armentrout Matheny Thurmond) and Blake Hare (Rowe Drilling Company) who provided technical assistance and guidance throughout the planning and implementation of study objectives.

Michael Hamrick and Gary Holloway with the U.S. Geological Survey (USGS) Georgia Water Science Center performed geophysical logging, wellbore-flowmeter surveys, and conducted groundwater-quality sampling. Mr. Hamrick also provided valuable assistance during the aquifer test and performed late instrument download and maintenance. Greg Cherry (USGS) assisted in the installation of monitoring equipment and collection of manual measurements during the aquifer test. John Clarke (USGS) provided assistance during the aquifer test and in the interpretation of wellbore-flowmeter-survey and water-quality data. Lynn Torak (USGS) checked additional spreadsheets developed for this study. Eve Kuniansky (USGS) contributed a literature review of the development of radial-flow models.

# Contents



# Figures

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# Conversion Factors and Abbreviations

Inch/Pound to SI units

| Multiply | By | To obtain |
|---|---|---|
| | Length | |
| inch | 2.54 | centimeter (cm) |
| foot (ft) | 0.3048 | meter (m) |
| mile (mi) | 1.609 | kilometer (mi) |
| | Flow | |
| foot per day (f/d) | 0.3048 | meter per year (m/yr) |
| gallon per minute (gal/min) | 0.06309 | liter per second (L/s) |
| gallon per day (gal/d) | 0.003785 | cubic meter per day ($m^3$/d) |
| | Transmissivity | |
| foot squared per day ($ft^2$/d) | 0.09290 | meter squared per day ($m^2$/d) |

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 the distance above the vertical datum.

# Abbreviations

RMS root mean square

µg/L micrograms per liter

mg/L milligrams per liter

mm micrometer (micron)

µS/cm microSiemens per centimeter

PEST parameter estimation

USGS U.S. Geological Survey

# Hydrogeology and Water Quality of the Dublin and Midville Aquifer Systems at Waynesboro, Burke County, Georgia, 2011

By Gerard J. Gonthier

## Abstract

The hydrogeology and water quality of the Dublin and Midville aquifer systems were characterized in the City of Waynesboro area in Burke County, Georgia, based on geophysical and drillers' logs, flowmeter surveys, a 24-hour aquifer test, and the collection and chemical analysis of water samples in a newly constructed well. At the test site, the Dublin aquifer system consists of interlayered sands and clays between depths of 396 and 691 feet, and the Midville aquifer system consists of a sandy clay layer overlying a sand and gravel layer between depths of 728 and 936 feet. The new well was constructed with three screened intervals in the Dublin aquifer system and four screened intervals in the Midville aquifer system. Wellbore-flowmeter testing at a pumping rate of 1,000 gallons per minute indicated that 52.2 percent of the total flow was from the shallower Dublin aquifer system with the remaining 47.8 percent from the deeper Midville aquifer system. The lower part of the lower Midville aquifer (900 to 930 feet deep), contributed only 0.1 percent of the total flow.

Hydraulic properties of the two aquifer systems were estimated using data from two wellbore-flowmeter surveys and a 24-hour aquifer test. Estimated values of transmissivity for the Dublin and Midville aquifer systems were 2,000 and 1,000 feet squared per day, respectively. The upper and lower Dublin aquifers have a combined thickness of about 150 feet and the horizontal hydraulic conductivity of the Dublin aquifer system averages 10 feet per day. The upper Midville aquifer, lower Midville confining unit, and lower Midville aquifer have a combined thickness of about 210 feet, and the horizontal hydraulic conductivity of the Midville aquifer system averages 6 feet per day. Storage coefficient of the Dublin aquifer system, computed using the Theis method on water-level data from one observation well, was estimated to be 0.0003. With a thickness of about 150 feet, the specific storage of the Dublin aquifer system averages about $2\times10^{-6}$ per foot.

Water quality of the Dublin and Midville aquifer systems was characterized during the aquifer test on the basis of water samples collected from composite well flow originating from five depths in the completed production well during the aquifer test. Samples were analyzed for total dissolved solids, specific conductance, pH, alkalinity, and major ions. Water-quality results from composite samples, known flow contribution from individual screens, and a mixing equation were used to calculate water-quality values for sample intervals between sample depths or below the bottom sample depth. With the exception of iron and manganese, constituent concentrations of water from each of the sampled intervals and total flow from the well were within U.S. Environmental Protection Agency primary and secondary drinking-water standards. Water from the bottommost sample interval in the lower part of the lower Midville aquifer (900 to 930 feet) contained manganese and iron concentrations of 59.1 and 1,160 micrograms per liter, respectively, which exceeded secondary drinking-water standards. Because this interval contributed only 0.1 percent of the total flow to the well, water quality of this interval had little effect on the composite well water quality. Two other sample intervals from the Midville aquifer system and the total flow from both aquifer systems contained iron concentrations that slightly exceeded the secondary drinking-water standard of 300 micrograms per liter.

## Introduction

The Dublin and Midville aquifer systems are the principal source of groundwater in the northern Coastal Plain of east-central Georgia (fig. 1; Clarke and others, 1985). The City of Waynesboro in Burke County, Ga., relies on groundwater from the Dublin and Midville aquifer systems for most of its water supply (Fanning and Trent, 2009). Waynesboro is located within an area characterized by high dissolved iron concentrations in the Dublin and Midville aquifer systems (Clarke and others, 1985, p. 46). The presence of iron in drinking water is objectionable because of its taste, staining capacity, and encrusting property (U.S. Environmental Protection Agency, 1976; Hem, 1985).

At Waynesboro, groundwater is supplied by two wells open to the Dublin and Midville aquifer systems. Well 29Y009 is located within the city limits, and well 29Y010 is located 1.5 miles (mi) north of downtown in what is referred to as the north well field. Supply well 29Y010 was constructed during

1994 and was gradually taken out of service by 2006 because of lost productivity when iron-oxide clogged the well screens (Reggie Hanton, Waynesboro Water Department, oral commun., September 7, 2012). The city currently is investigating the means of increasing its water supply at the north well field by installing a replacement well (29Y015) located 42 feet (ft) northwest of well 29Y010 (referred to as old production well 29Y010 in this report; fig. 1). The installation and testing of the replacement well provided the opportunity to study the hydraulic properties of the Dublin and Midville aquifer systems.

The U.S. Geological Survey (USGS), in cooperation with the City of Waynesboro, conducted site investigations during April through August 2011 to assess the hydrogeology and water quality of the Dublin and Midville aquifer systems. This assessment provided an improved regional characterization of the two aquifer systems.

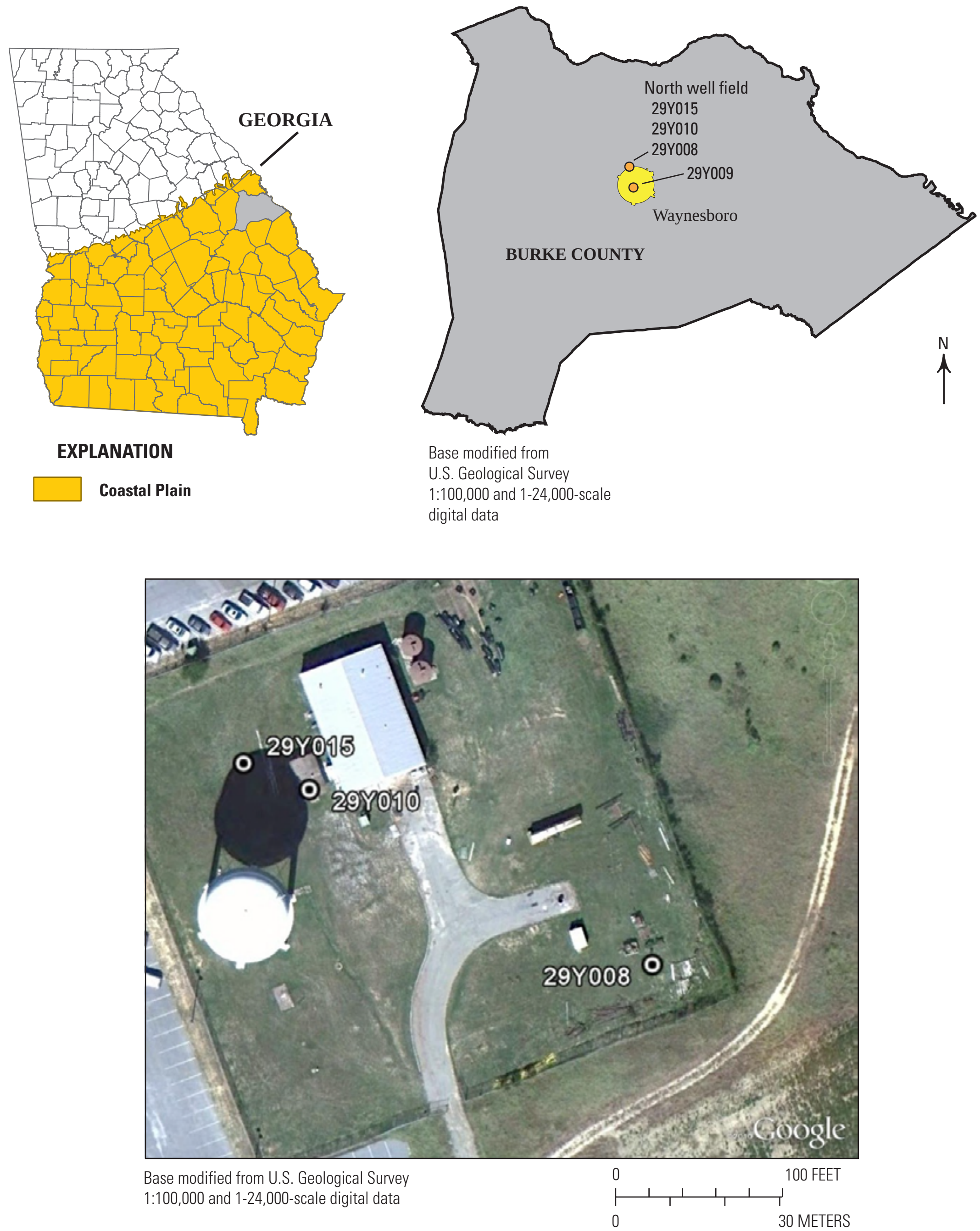



**Figure 1.** Location of wells in the north well field 1.5 miles north of Waynesboro in Burke County, Georgia, 2011.

## Purpose and Scope

This report describes the hydrogeology and water quality of the Dublin and Midville aquifer systems at Waynesboro, Burke County, Ga., based largely on data collected from well 29Y015, constructed during June–August 2011. Description of the hydrogeology included the lithology and thickness of geologic and hydrogeologic units and hydraulic properties of hydrogeologic units. The data from well 29Y015 consist of geophysical and drillers' logs, a wellbore-flowmeter survey, a 24-hour aquifer test, and the collection and chemical analyses of water samples. Similar data from wells 29Y010, 29Y008, and 29Y009 also were used for hydrogeologic characterization in the Waynesboro area. The percent contribution of water from specific screened intervals in old production well 29Y010 and new production well 29Y015 was determined by wellbore-flowmeter surveys performed during April and August 2011, respectively. A 24-hour aquifer test was performed at well 29Y015 during August 2011 in which well 29Y015 was pumped at a discharge rate of 1,000 gallons per minute (gal/min). Water levels were monitored in the pumped well 29Y015 and in wells 29Y010 and 29Y008. Transmissivity and storage coefficient of the Dublin and Midville aquifer systems were estimated using results from the wellbore-flowmeter surveys and the 24-hour aquifer test. Water samples were collected from five depths in well 29Y015 during the aquifer test and analyzed for total dissolved solids, specific conductance, pH, alkalinity, sodium, potassium, magnesium, calcium, manganese, iron, fluoride, chloride, and sulfate.

The results of this study provide hydrogeologic information that can be used to determine a regional assessment of aquifers. Such information supports the USGS Groundwater Resources Program to provide objective scientific information and develop the interdisciplinary understanding necessary to assess and quantify the availability of the Nation's groundwater resources (*http://water.usgs.gov/ogw/gwrp/* accessed December 18, 2012).

## Site Description and Water Use

In 2005, total water use for the City of Waynesboro was 930,000 gallons per day (gal/d). Groundwater accounted for 790,000 gal/d and surface water obtained from Brier Creek accounted for 140,000 gal/d (Fanning and Trent, 2009). Groundwater was supplied by wells at two locations—well 29Y009 is located within the Waynesboro city limits, and well 29Y010 is located 1.5 mi north of downtown in what is referred to as the north well field (fig. 1). This study focused primarily on data from existing wells and a new well (29Y015) constructed at the north well field. At the north well field, old production well 29Y010 and observation well 29Y008 are located 42 and 271 ft southeast of new production well 29Y015, respectively. Topography generally is undulating with land-surface altitudes ranging from 330 ft at hilltops to 230 ft in streambeds. Wells 29Y015 and 29Y008 are located in an upland area with land-surface altitudes of 302 and 292 ft, respectively. Production well 29Y009, located near downtown Waynesboro, has a land-surface altitude of about 305 ft.

### Climate

The study area has a mild climate with warm, humid summers and mild winters. Long-term climatic patterns in the area were derived from records provided by the National Weather Service station 2 mi northeast of Waynesboro, Ga. (climatological station "Waynesboro 2 NE, Georgia [099194]," accessed March 19, 2012, at *http://www.sercc.com/cgi-bin/sercc/cliMAIN.pl?ga9194*). During the climate period 1971–2000, precipitation at station 099194 averaged 48.08 inches per year (in/yr). Monthly rates of precipitation reached maximum levels in late winter with 4.61 inches per month (in/mo) during February and in late summer with 5.09 in/mo during August. Monthly rates of precipitation reached minimum levels in late spring with 3.02 in/mo during May and in late autumn with 2.80 in/mo during November.

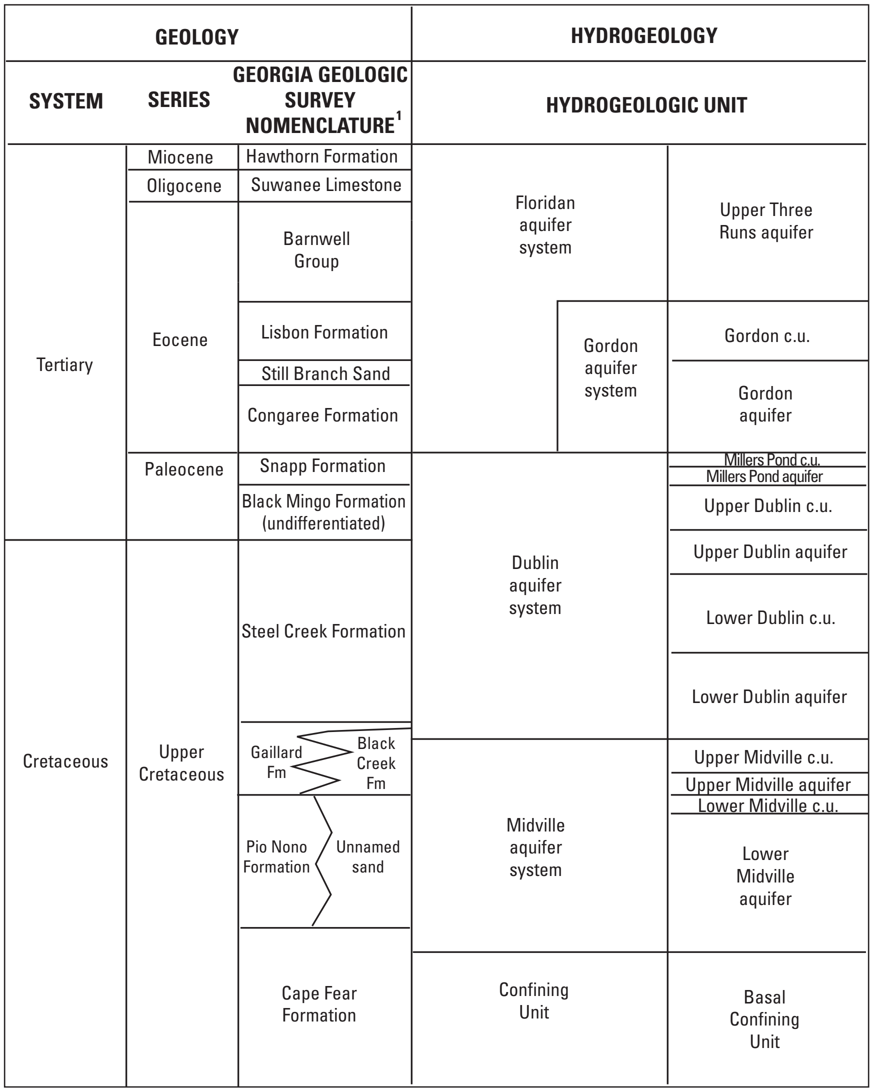
## Hydrogeology

The Waynesboro, Burke County, area is underlain by Coastal Plain strata consisting mostly of unconsolidated layers of upper Cretaceous to lower Miocene sand and clay (fig. 2), and some layers of limestone. The Coastal Plain strata attain a maximum thickness of about 985 ft (Falls and others, 1997). These sediments constitute three major aquifer systems, in order of descending depth: the Floridan aquifer system, Dublin aquifer system, and Midville aquifer system. Clarke and others (1985) reported that hydraulic separation between these aquifer systems decreases in an updip direction and that the Dublin and Midville aquifer systems coalesce into a single aquifer system called the Dublin-Midville aquifer system.

The Floridan aquifer system in Burke County consists of the Upper Three Runs aquifer and Gordon aquifer system (fig. 2). The Upper Three Runs aquifer consists of largely clastic Eocene to Miocene sediments and is hydrogeologically equivalent to the carbonate Upper Floridan aquifer in coastal Georgia (Payne and others, 2005). The Gordon aquifer system, as classified by Brooks and others (1985), consists of Eocene clastic sediments and includes the Gordon confining unit and Gordon aquifer. The Gordon aquifer system is hydrogeologically equivalent to the Lower Floridan confining unit and Lower Floridan aquifer in coastal Georgia (Payne and others, 2005).

The Dublin aquifer system consists of six hydrogeologic units, in order of descending depth: the Millers Pond confining unit, Millers Pond aquifer, upper Dublin confining unit, upper Dublin aquifer, lower Dublin confining unit, and lower Dublin aquifer (Clarke and others, 1985; Falls and others, 1997). The Millers Pond confining unit, Millers Pond aquifer, and upper Dublin confining unit consist of Paleocene clastic sediments. The Millers Pond hydrogeologic units are thin or absent in the vicinity of Waynesboro. The upper Dublin aquifer, lower Dublin confining unit, and lower Dublin aquifer consist of upper Cretaceous clastic sediments.

The Midville aquifer system consists of upper Cretaceous clastic sediments and includes, in order of descending depth: the upper Midville confining unit, upper Midville aquifer, lower Midville confining unit, and lower Midville aquifer. In the Waynesboro area, the aquifer system is underlain by a basal confining unit consisting of clay-matrix-filled sands and saprolite (Falls and others, 1997).




**EXPLANATION**

**c.u. confining unit**

**Fm formation**

[1] Huddlestun and Hetrick, 1991; Summerour and others, 1994; Huddlestun and Summerour, 1996
Modified from Falls and others, 1997
Gordon aquifer system distinction from Brooks and others, 1985

**Figure 2.** Correlation chart of geologic and hydrogeologic units in Waynesboro, Burke County, Georgia.

# Methods of Data Collection and Analysis

Hydrogeologic characteristics of the Dublin and Midville aquifer systems were characterized using geophysical and drillers' logs, wellbore-flowmeter surveys, and results from a 24-hour aquifer test. Water quality was assessed by collecting wellbore grab samples at selected depths during the aquifer test.

## Hydrogeologic Data Collection

Four wells provided data used to characterize the hydrogeology and water quality at Waynesboro (figs. 3–6). Wells 29Y015, 29Y010, and 29Y008 are located at the north well field, and well 29Y009 is located near downtown Waynesboro (fig. 1). Well construction and location information for these wells are listed in table 1.

At the north well field, a 1,109-ft test hole (fig. 3) was drilled during early May 2011, using a mud-rotary drill to characterize the lithology of sediments above basement rock. Cuttings were collected every 10 ft or as lithologic changes were observed. The following borehole geophysical logs were collected to characterize geophysical properties of the penetrated sediments and interstitial fluid: caliper; natural gamma; spontaneous potential; lateral, long-normal, 64 inches, and short-normal, 16 inches, fluid resistivity; and temperature. The borehole was developed into well 29Y015, completed to a depth of 940 ft using 12-inch inner-diameter schedule-40 steel casing, and stainless-steel wire-wrapped screen at seven separate intervals. The bottom 169 ft of the test hole was backfilled with bentonite grout to create a hydraulic barrier between the aquifer systems and the underlying basal confining unit. A gravel pack around the screens extends upward to a depth between 365 and 385 ft; an outer seal of bentonite laterally seals the gravel pack down to a depth of 405 ft (fig. 7; Rowe Drilling Company, written commun., 2011).

Flowmeter surveys were conducted using an electromagnetic flowmeter to determine the relative contribution of flow from screens in wells 29Y015 and 29Y010 that were open to the Dublin and Midville aquifer systems. A pump operated above the screens and the depth interval of the flowmeter survey to produce upward flow in the screened sections of the wellbore. A flowmeter survey was performed in well 29Y010 in April 2011 while the well was pumped at a rate of 300 gal/min. In well 29Y015, a flowmeter survey was completed about 8 hours into the August 2011 aquifer test when the well was pumped at a rate of 1,000 gal/min. Duplicate discharge measurements between the screens in well 29Y015 indicate measurement uncertainty to be ±4 percent.

A 24-hour aquifer test was performed in well 29Y015, which included continuous and intermittent groundwater-level measurements at the pumped well and two observation wells, according to USGS methods (Stallman, 1971; Cunningham and Schalk, 2011). Manual intermittent water-level measurements in wells were made for calibration of groundwater-level recorder readings and for direct monitoring of the aquifer test. An electric tape was used to make manual measurements to the nearest 0.01 ft following procedures described in Garber and Koopman (1968) and Cunningham and Schalk (2011). Continuous groundwater-level recorders were installed in the pumped well and two observation wells to monitor water-level changes using submerged, vented pressure transducers. For long-term monitoring, pressure transducers measured water levels that were recorded every 15 minutes; for the aquifer test, pressure transducers measured water levels that were recorded every half second to every 15 minutes.

Discharge during the aquifer test was measured by the well driller who used the orifice method described by Schwab (2007) and Cunningham and Schalk (2011). A 6-inch-diameter orifice partially blocked outflow from an 8-inch-diameter pipe. A manometer was connected to the 8-inch pipe. The pump flow was adjusted so that the water level in the manometer was maintained 41.5 inches above the center of the outflow (1,000 gal/min). The manometer water level was checked regularly to ensure that the correct height, and therefore the correct discharge, was maintained.

Drawdown was estimated by subtracting the water level measured during the aquifer test from the background water level. The background water level was determined based on the groundwater-level trend for up to 3 months prior to the aquifer test. Just prior to the start of the aquifer test, the background water level was set equal to the measured water level. The background water-level was represented as a decreasing linear trend with time that was slightly adjusted for each well to create an appropriately representative drawdown time series.

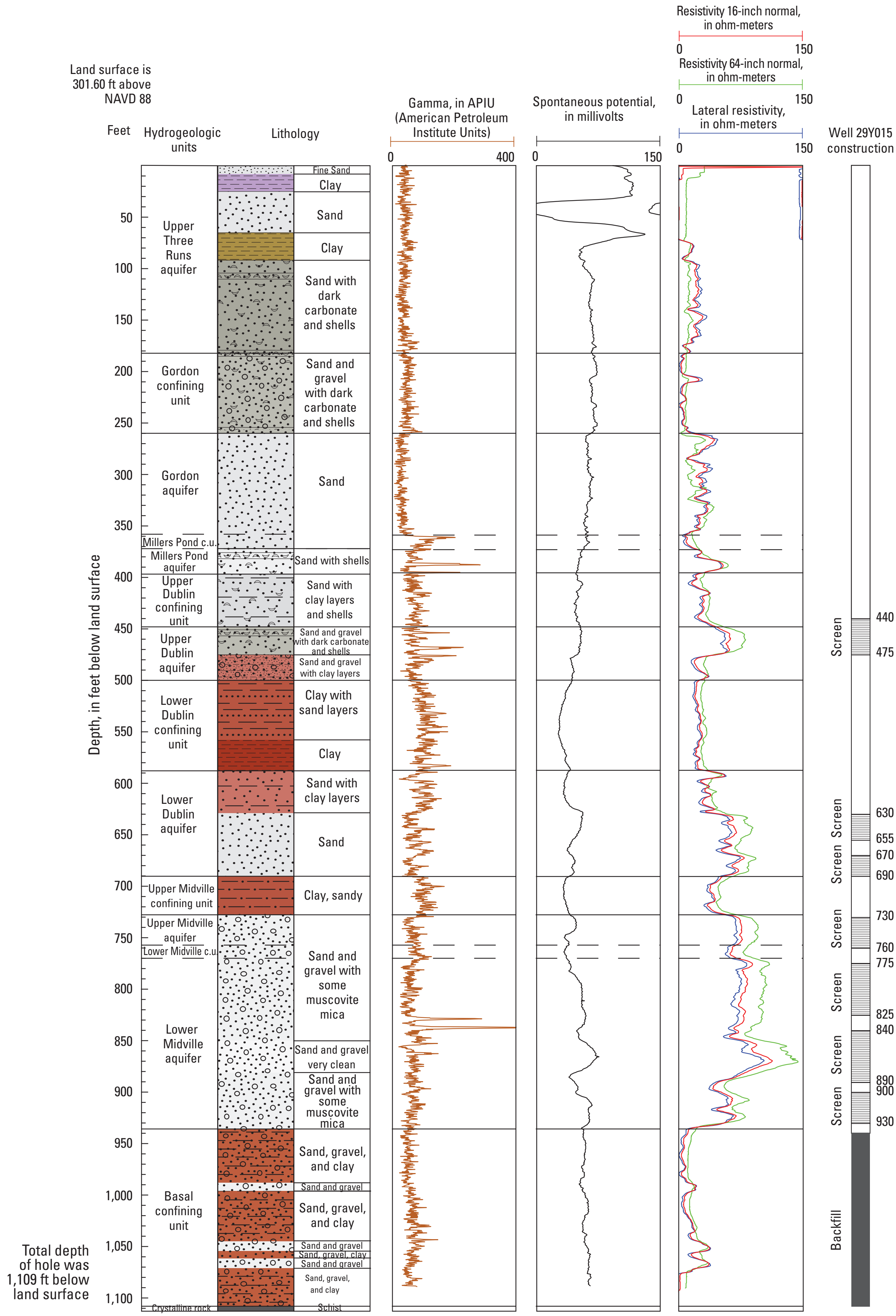



**Figure 3.** Lithology and geophysical properties of well 29Y015 near Waynesboro, Georgia. [c.u., confining unit.]

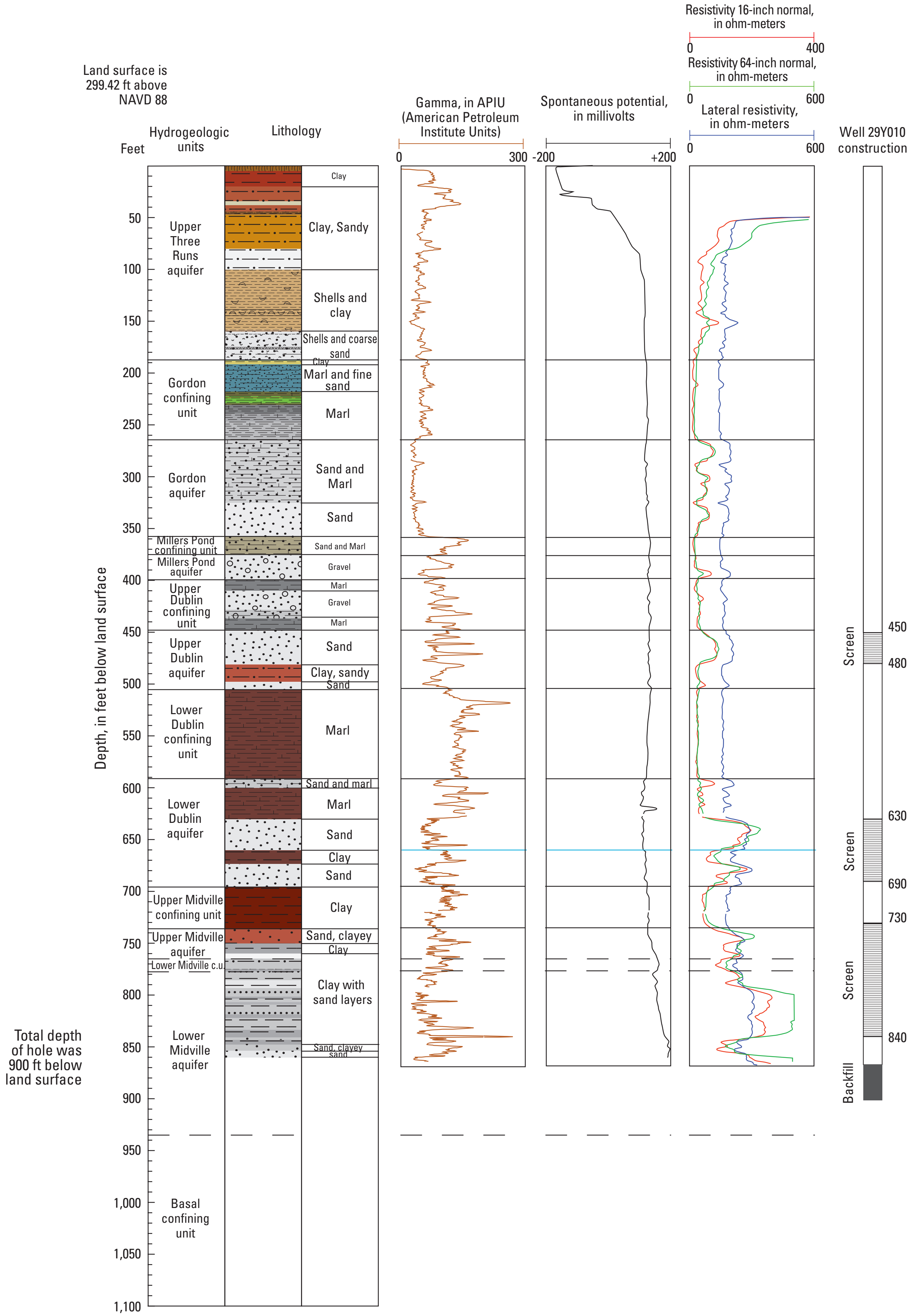



**Figure 4.** Lithology and geophysical properties of well 29Y010 near Waynesboro, Georgia. [c.u., confining unit.]

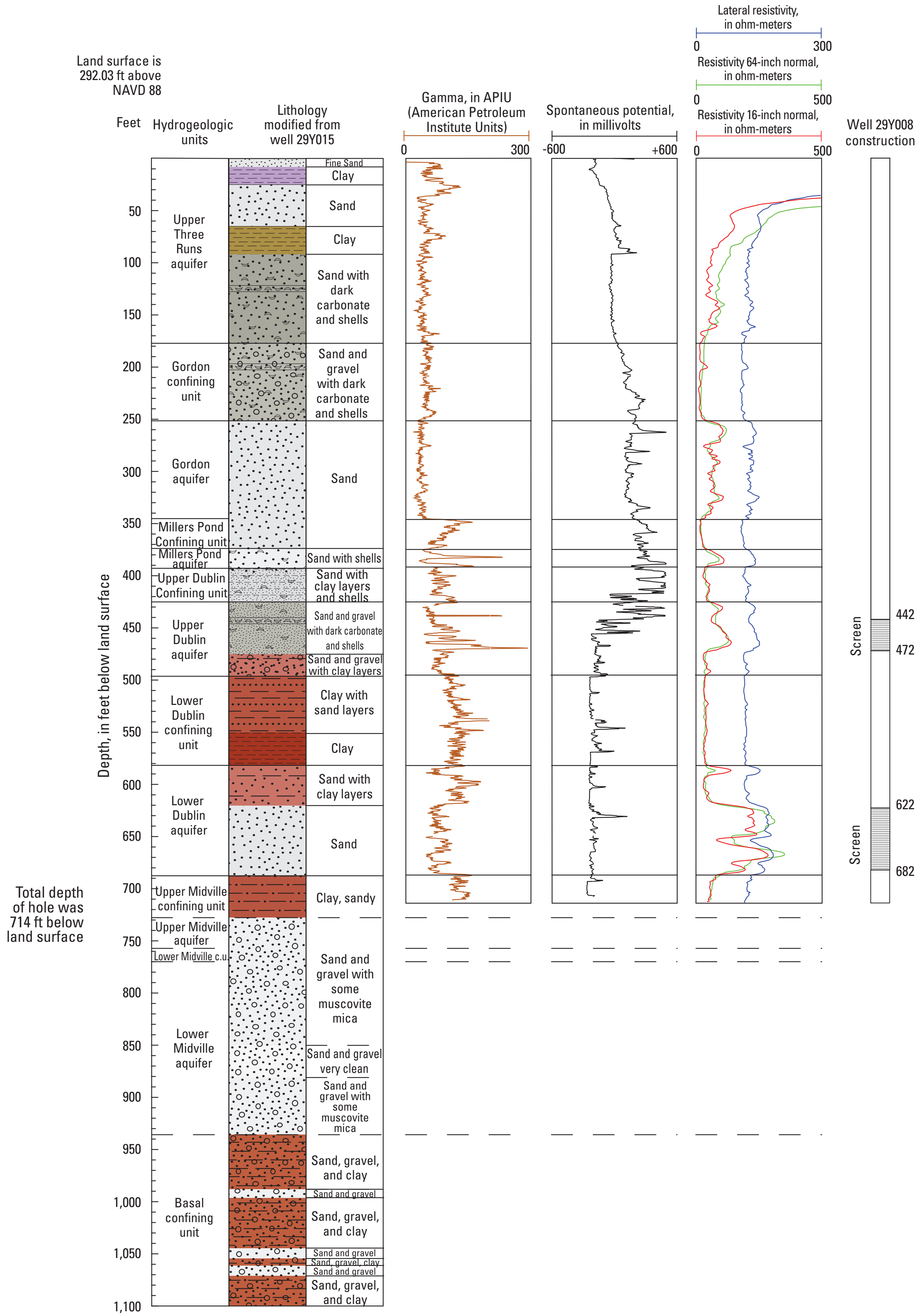



**Figure 5.** Geophysical properties of well 29Y008 near Waynesboro, Georgia. [c.u., confining unit.]

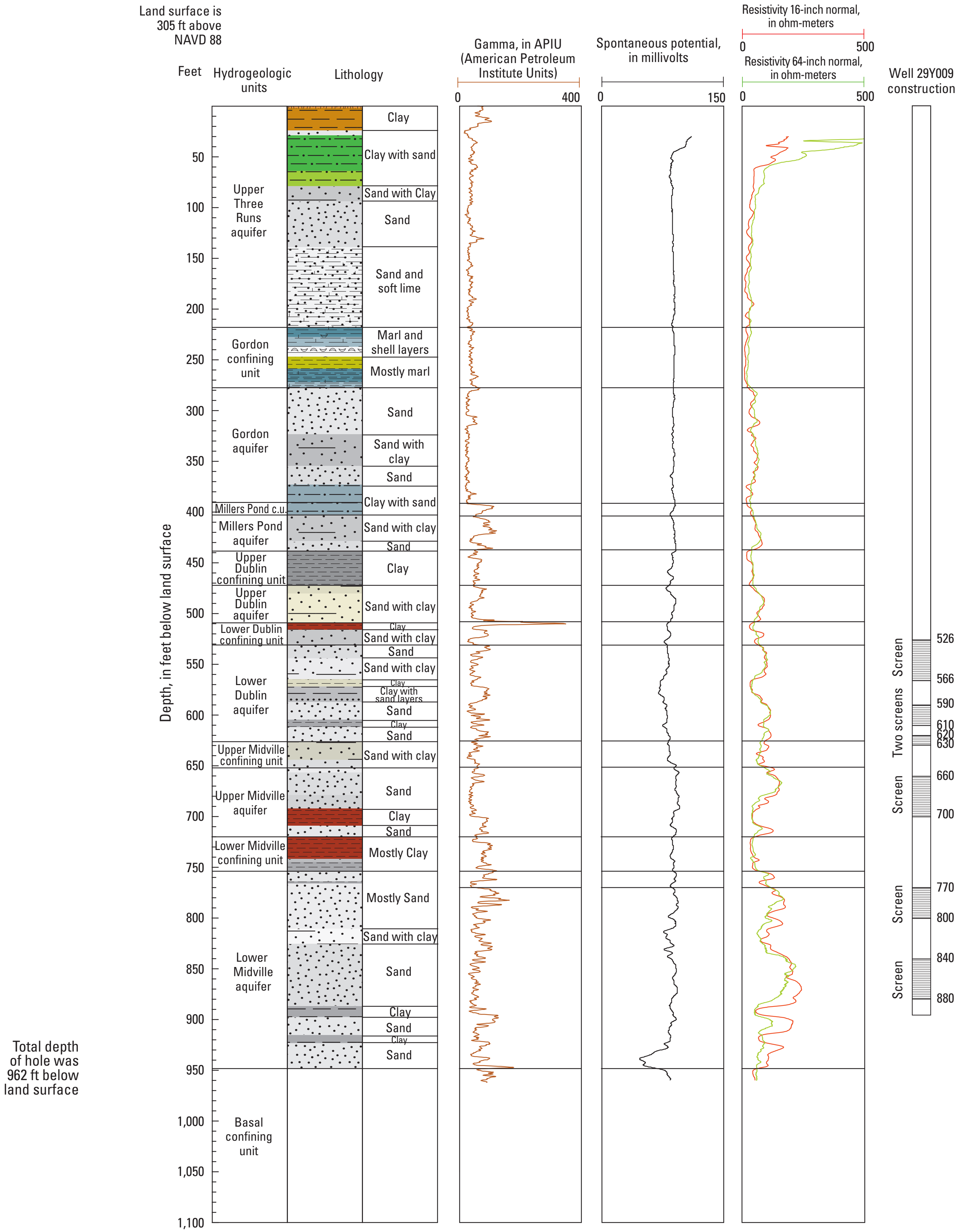



**Figure 6.** Lithology and geophysical properties of well 29Y009 in Waynesboro, Georgia. [c.u., confining unit.]

**Table 1.** Information for wells in and just north of Waynesboro in Burke County, Georgia.

[NAVD 88, North American Vertical Datum of 1988; N/A, not applicable]

| Well name | Site identification | Altitude of land surface, in feet above NAVD 88 | Number of screens or openings | Depth to top of screen, in feet | Depth to bottom of screen, in feet | Radius from pumped well, in feet | Hydrogeologic unit to which screens are aligned | Driller's log | Geophysical logs | Flow-meter survey | Monitored for water levels during the aquifer test? | Data were used in aquifer-test analysis? |
|---|---|---|---|---|---|---|---|---|---|---|---|---|
| 29Y015 | 330638082013601 | 301.60 | 7 | 440 | 475 | 0.5 | Upper Dublin aquifer | Yes | Yes | Yes | Yes | Yes |
| | | | | 630 | 655 | | Lower Dublin aquifer | | | | | |
| | | | | 670 | 690 | | Lower Dublin aquifer | | | | | |
| | | | | 730 | 760 | | Upper Midville aquifer | | | | | |
| | | | | 775 | 825 | | Lower Midville aquifer | | | | | |
| | | | | 840 | 890 | | Lower Midville aquifer | | | | | |
| | | | | 900 | 930 | | Lower Midville aquifer | | | | | |
| 29Y010 | 330604082012802 | 299.42 | 3 | 450 | 480 | 42 | Upper Dublin aquifer | Yes | Yes | Yes | Yes | No |
| | | | | 630 | 690 | | Lower Dublin aquifer | | | | | |
| | | | | 730 | 840 | | Midville aquifer system | | | | | |
| 29Y008 | 330640082012802 | 292.03 | 2 | 442 | 472 | 271 | Upper Dublin aquifer | No | Yes | No | Yes | Yes |
| | | | | 622 | 682 | | Lower Dublin aquifer | | | | | |
| 29Y009 | 33051608211502 | 305 | 6 | 526 | 566 | 8,500 | Lower Dublin aquifer | Yes | Yes | No | No | N/A |
| | | | | 590 | 610 | | Lower Dublin aquifer | | | | | |
| | | | | 620 | 630 | | Lower Dublin aquifer | | | | | |
| | | | | 660 | 700 | | Upper Midville aquifer | | | | | |
| | | | | 770 | 800 | | Lower Midville aquifer | | | | | |
| | | | | 840 | 880 | | Lower Midville aquifer | | | | | |

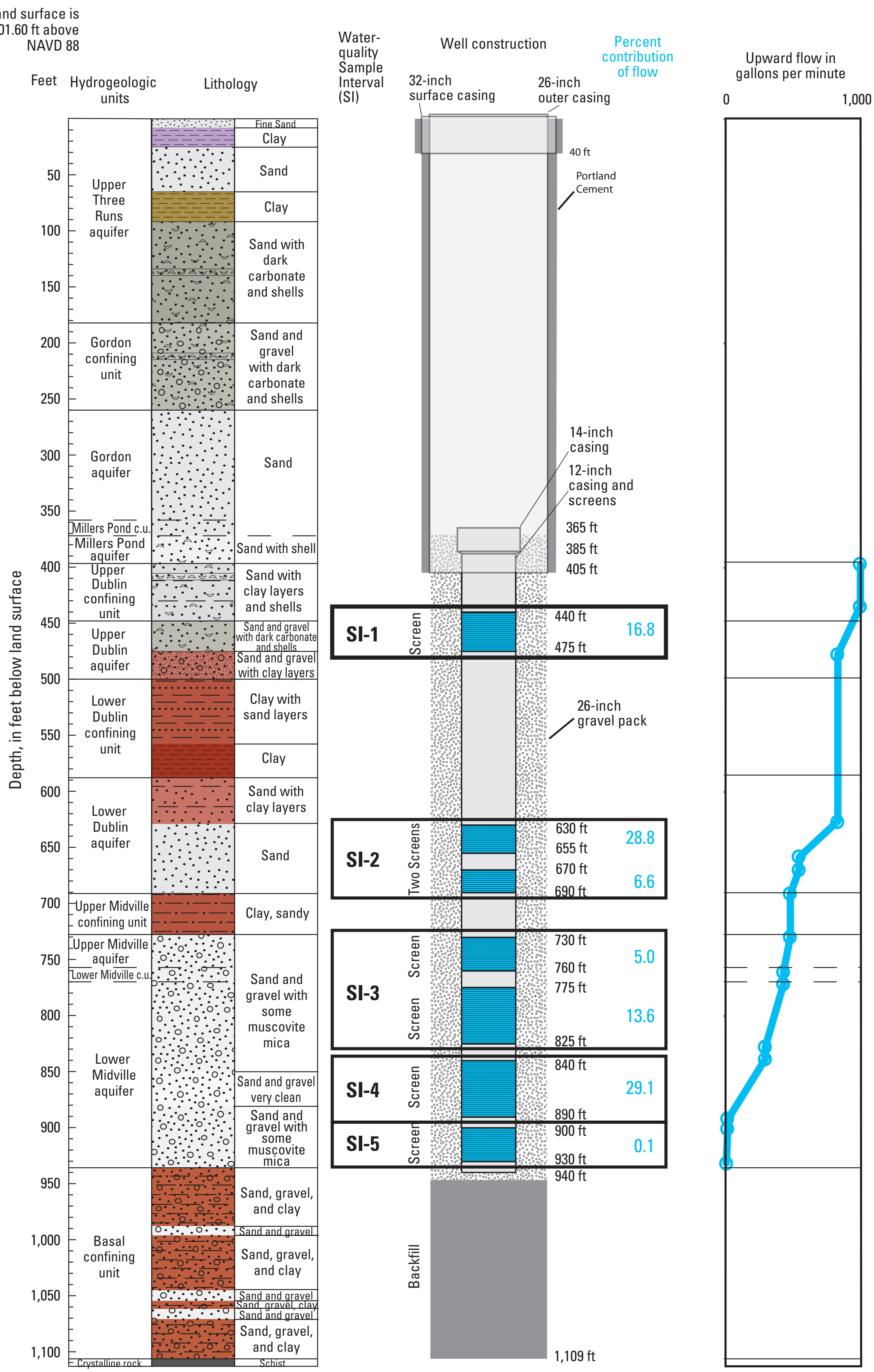



**Figure 7.** Construction, flowmeter survey results, and water-quality sample intervals (SI) for well 29Y015 near Waynesboro, Georgia, August 2011. [c.u., confining unit.]

## Aquifer-Test Analysis

Hydraulic properties of the Dublin and Midville aquifer systems were estimated using AnalyzeHOLE (Halford, 2009) and analytical solutions (Theis, 1935; Cooper and Jacob, 1946; Papadopulos and Cooper, 1967). AnalyzeHOLE is an integrated analysis tool for simulating flow and transport in a pumped well that uses observations from the flowmeter survey performed in the pumped well and drawdown from the pumped well and nearby observation wells. These data are input into the two-dimensional, axisymmetric radial, transient, finite-difference model, MODFLOW (Harbaugh and others, 2000), which uses the parameter-estimation program PEST (Doherty, 2005) to estimate the horizontal hydraulic conductivity of hydrogeologic units and the gravel pack. Results from the AnalyzeHOLE simulation provided estimates of transmissivity for the Dublin and Midville aquifer systems. For model simulation and parameter estimation, initial single values of specific storage, porosity, and vertical anisotropy were entered for all hydrogeologic units. An initial value of composite (or target) transmissivity was entered for the entire hydrogeologic column. Full descriptions of the derivation of two-dimensional radial models using a single layer or multiple layers are provided in the following references: Rutledge (1991); Reily and Harbaugh (1993); Langevin (2008); and Halford (2009). The method of computation of flow observations for parameter estimation with radial models is described in Clemo (2002). The composite transmissivity is estimated from analytical solution methods, such as those published by Theis (1935) or Cooper and Jacob (1946) and is required input for the use of the program AnalyzeHOLE. An advantage of using AnalyzeHOLE is that flowmeter-survey data and drawdown data from pumped well 29Y015 and observation well 29Y008 are integrated into the estimation process; thus, more detailed estimates of the hydraulic conductivity and transmissivity of individual hydrogeologic units or subunits can be estimated. However, the total transmissivity is constrained by the target composite transmissivity determined from the analytical solutions.

AnalyzeHOLE has two limitations or requirements for additional data that required the use of analytical solution methods, such as Theis (1935); Cooper and Jacob (1946); and Papadopulos and Cooper (1967) methods, prior to using the analysis tool. AnalyzeHOLE does not estimate storage properties and only uses a single diameter for the entire wellbore. Because well 29Y015 was constructed with different casing diameters in the upper and lower parts of the well, wellbore storage effects, as described by Papadopulos and Cooper (1967), are not adequately addressed. The Theis method was used on water-level data from observation well 29Y008 to estimate a value of specific storage that was entered into AnalyzeHOLE. Papadopulos-Cooper type curves were used to determine the amount of time for wellbore storage effects to abate; this time was used as the starting time for inputting drawdown data to the axisymmetric model and for using the PEST program. Additionally, the composite transmissivity is required in order to constrain the parameters estimated with AnalyzeHOLE (Halford, 2009).

Spreadsheets for the Cooper-Jacob method are fully documented in Halford and Kuniansky (2002). Additional spreadsheets were developed for this study and are described in the following paragraphs.

The Theis method and Papadopulos-Cooper type curves were iteratively used to simultaneously determine the hydraulic properties of the aquifer systems and wellbore storage effects. The Theis method was used to determine the transmissivity and storage coefficient of the Dublin aquifer system by using water-level data from observation well 29Y008, and the combined transmissivity of the Dublin and Midville aquifer systems by using water-level data from well 29Y015. The Papadopulos-Cooper method was used only to determine when wellbore storage effects became small following a change in discharge in the pumped well.

The Theis and Cooper-Jacob methods are based on the same analytical solution for the one-dimensional radial coordinate system partial differential equation for flow to a well in a confined aquifer and assume the following: the aquifer has infinite extent, has uniform thickness, is horizontal, homogeneous and isotropic; the aquifer is fully confined and discharge is derived exclusively from storage in the aquifer; the potentiometric surface is horizontal initially; the well fully penetrates the confined aquifer resulting in horizontal flow to the well and flow is laminar; the well has infinitesimally small diameter (no wellbore storage); and well discharge is at a constant rate.

Papadopulos-Cooper published type curves (used to determine drawdown as a function of time) follow some of the same assumptions for a confined aquifer as the Theis and Cooper-Jacob methods. But unlike the assumption that wellbore storage is negligible, the Papadopulos-Cooper type curves consider the effects of wellbore storage. As time elapses following the start of an aquifer test, the wellbore storage effects become small and the Papadopulos-Cooper type curves converge with the Theis curve. In addition to the argument, $u$, of the Theis function $W(u)$, the Papadopulos-Cooper function $F(u,\alpha,\rho)$ contains two additional variables that affect drawdown, $\alpha$ and $\rho$, which take into account the radius of the screened interval and the radius of the wellbore where water level is changing:

$$\alpha = \frac{r_w^2 S}{r_c^2}, \rho = \frac{r}{r_w} , \qquad \text{(1 and 2)}$$

where

$r_w$ is the radius of the screened interval of the wellbore, in feet;
$r_c$ is the radius of the wellbore where the water level is changing, in feet;
$S$ is the aquifer storage coefficient; and
$r$ is the distance from the center of pumping, in feet.

From Papadopulos and Cooper (1967), drawdown inside the well is simulated from the following equation:

$$s(t) = \frac{Q}{4\pi T} F(u, \alpha, \rho) \tag{3}$$

where

$u$ is the variable in the Theis function $W(u)$ (Theis, 1935) and is $\frac{r_w^2 S}{4Tt}$, dimensionless;

$Q$ is the discharge pumped from the well, in cubic feet per day;

$T$ is transmissivity of the aquifer, in feet squared per day;

$t$ is for time after the start of the aquifer test, in days; and

$\rho$ equals 1.

Although the equation from Papadopulos and Cooper allows for the estimation of storage coefficient from a single-well test, as with other methods, estimates of storage coefficient from single well tests are not reliable (Halford and Kuniansky, 2002; Halford and others, 2006).

The combined transmissivity for the Dublin and Midville aquifer systems also was estimated using the Cooper-Jacob straight-line method, which is based on the same equations as the Theis method, and thus has the same assumptions (Cooper and Jacob, 1946; Halford and Kuniansky, 2002). The Cooper-Jacob straight-line method provides a good estimate of transmissivity from the slope of a straight-line fit to an arithmetic scale for drawdown on the y-axis versus time plotted with a logarithmic scale on the x-axis; where transmissivity is estimated from the change in drawdown per log cycle of a line fit to the data after wellbore storage effects are gone (Halford and Kuniansky, 2002). The transmissivity estimates from the Cooper-Jacob straight-line method have been found to be fairly robust even with well losses and partial-penetration (Halford and others, 2006). The storage coefficient could be estimated from the intercept of the straight line after the transmissivity is calculated (Cooper and Jacob, 1946); however, well losses affect the intercept without affecting the slope of the straight line. Thus, it is ill-advised to estimate storage from a single well test and this value is not calculated for users of the spreadsheet published in Halford and Kuniansky (2002).

A goodness of fit or “match” between Theis method simulated and observed drawdown and recovery was first determined from a graph where drawdown and recovery are on the y-axis and log time after a change in discharge is on the x-axis. Wellbore-storage effects occur in early time on the graph as a deviation in observed drawdown and recovery from the drawdown and recovery that are simulated using the Theis method. Wellbore storage effects abate with time. The point on the x-axis of the graph where the observed drawdown and recovery and the Papadopulos-Cooper type curves converge with the simulated drawdown and recovery from the Theis method is the time when wellbore storage effects became small. Hydraulic properties then were adjusted to match simulated drawdown and recovery with observed drawdown and recovery after the wellbore storage effects became small. Goodness of fit was then estimated with the root-mean-square (RMS) of the difference between simulated and observed drawdown and recovery; the smaller the RMS the better the match. Radius or distance from the pumping source, discharge, and the time of pumping are known values. The transmissivity and, for observation well 29Y008, the storage coefficient were adjusted until a minimum value of RMS was reached. The transmissivity and storage coefficient used to minimize the RMS was considered the estimate of those hydraulic properties. Details of estimating hydraulic properties of hydrogeologic units are discussed in the section Aquifer-Test Analysis under the section Aquifer Test.

## Water-Quality Data Collection and Analysis

Water samples were collected using a wireline grab sampler just above five of the seven screens in pumped well 29Y015. Sample collection began after the completion of a wellbore-flowmeter survey about 9 hours into the aquifer test. Water was transferred from the grab sampler to sample bottles using a peristaltic pump. Samples were analyzed for total dissolved solids, specific conductance, pH, and alkalinity, reported as calcium carbonate; and for major ions including sodium, potassium, magnesium, calcium, manganese, iron, fluoride, chloride and sulfate. Water collected for major ions was filtered using a capsule filter with a 0.45-micrometer (µm) pore medium. Samples for cations were preserved with nitric acid. Samples were analyzed at TestAmerica Laboratories, Inc., Savannah, Ga. Cations were analyzed using induced coupled plasma; anions were analyzed using ion chromatography. Bicarbonate concentrations were calculated from values of alkalinity. Water type was determined from the percentage of equivalents of sodium plus potassium, calcium, magnesium, chloride plus fluoride, sulfate, and carbonate plus bicarbonate that were plotted on a trilinear diagram (Piper, 1944).

Each sample is a composite of flow contributed from all screens below its sample-collection point. A simple mixing equation and the known flow contribution from screens from the wellbore-flowmeter survey were used to convert composite water-sample concentrations into concentrations of individual sample intervals between sample-collection points. Water was assumed to be flowing through screens from adjacent hydrogeologic units and completely mixing before reaching the collection point. The mixing equation from Kendall and Caldwell (1998, p. 80) was applied to sample intervals in the well as follows:

$$Q_{T,n} C_{T,n} = Q_{T,n-1} C_{T,n-1} + Q_{I,n} C_{I,n} \tag{4}$$

where

$Q_{T,n}$ is the composite discharge at sample-collection point $n$, contributed to or flowing up the wellbore from all screened intervals below sample-collection point $n$, in gallons per minute;

$C_{T,n}$ is the concentration of a specific conservative constituent in discharge water $Q_{T,n}$, expressed in a linear-unit value that varies with constituent, but represents the mass of the constituent per volume of water;

$Q_{T,n-1}$ is the composite discharge at sample-collection point $n-1$, contributed to or flowing up the wellbore from all screened intervals below sample-collection point $n-1$, in gallons per minute;

$C_{T,n-1}$ is the concentration of a specific conservative constituent in discharge water $Q_{T,n-1}$, expressed in a linear-unit value that varies with constituent, but represents the mass of the constituent per volume of water;

$Q_{I,n}$ is the discharge entering the well from the interval between sample-collection points $n$ and $n-1$, in gallons per minute; and

$C_{I,n}$ is the concentration of a specific conservative constituent in discharge water $Q_{I,n}$, expressed in a linear-unit value that varies with constituent, but represents the mass of the constituent per volume of water.

Composite water-sample concentrations at sample-collection points are known from sampling and analysis, and discharge rates are known from the wellbore-flowmeter survey. Therefore, equation 4 can be rearranged to solve for the concentration, $C_{I,n}$, of the specific conservative constituent occurring in discharge water entering the well between the two sample-collection points $n$ and $n-1$ ($Q_{I,n}$):

$$C_{I,n} = \frac{Q_{T,n}C_{T,n} - Q_{T,n-1}C_{T,n-1}}{Q_{I,n}} . \qquad (5)$$

# Hydrogeology and Water Quality

To assess the hydrogeology of the Dublin and Midville aquifer systems, detailed site investigations were performed during 2011 in the vicinity of Waynesboro, Burke County, Ga. The study included delineation of hydrogeologic units at three wells located at the north well field—newly constructed well 29Y015, old production well 29Y010, and observation well 29Y008—and at the production well (29Y009) located in downtown Waynesboro. The hydrogeologic assessment included a test boring to a depth of 1,109 ft; construction of well 29Y015 in the borehole by installing seven screens from 440 to 930 ft deep; geophysical testing and examination of geophysical logs from the four wells; drillers' logs from three of the wells; flowmeter surveys in two wells; and a 24-hour constant-discharge aquifer test at newly constructed well 29Y015.

## Hydrogeologic Units

The Coastal Plain sediments beneath Waynesboro consist mostly of sands and clays with some gravel and minor amounts of marl that range in age from upper Cretaceous through Miocene (figs. 3–6). At well 29Y015, these sediments are 1,109 ft thick and overlie crystalline basement rock. Water-bearing units that were penetrated in well 29Y015 consist of the Upper Three Runs aquifer, Gordon aquifer system, Dublin aquifer system, and Midville aquifer system. The Upper Three Runs aquifer was present from land surface to a depth of 182 ft in well 29Y015. The base of the Upper Three Runs aquifer is identified with slightly elevated formation resistivities relative to the underlying Gordon confining unit (fig. 3). The slightly elevated resistivities extended from 74 to about 182 ft deep in well 29Y015. The upper half of the Upper Three Runs aquifer is mostly medium- to coarse-grained sand and clay, whereas the lower half contains very fine gravel-sized clasts of a dark carbonate rock, shells, and sand. The dark carbonate rock could be calcareous mudstone.

The Gordon confining unit underlies the Upper Three Runs aquifer at 182 to 260 ft deep in well 29Y015. The unit is composed of a 78-ft-thick layer with a lithology similar to that of the lower half of the Upper Three Runs aquifer, but has lower resistivity values than that of the Upper Three Runs aquifer, possibly indicating higher clay content.

The Gordon aquifer, Millers Pond confining unit, and Millers Pond aquifer are lithologically similar in well 29Y015 and constitute a 136-ft-thick layer of predominantly medium- to coarse-grained sand having variable clay content. The Gordon aquifer is juxtaposed in the upper 98 ft of sand at 260 to 358 ft deep. The Millers Pond confining unit is lithologically indistinguishable from the Gordon aquifer in well 29Y015 and is associated with sand and clay or marl in wells 29Y010 and 29Y009 (figs. 4 and 6). In well 29Y015, the Millers Pond confining unit is interpreted to be 14 ft thick. At the top of the Millers Pond confining unit, natural-gamma radiation sharply increases with depth, then gradually decreases. The natural-gamma signature of this confining unit is evident in the geophysical logs collected from the four wells (figs. 3–6). The Coastal Plain sediments beneath this depth generally have higher natural-gamma radiation than units at shallower depths (figs. 3–6). In well 29Y015, the Millers Pond aquifer is associated with a shelly sand layer in the bottom 24 ft of the 136-ft-thick sand layer at 372 to 396 ft deep (fig. 3).

The upper Dublin confining unit and aquifer are associated with a lithologic transition with depth consisting of a decrease in shells and marl and an increase in red clay. The upper Dublin confining unit is present at 396 to 448 ft deep and consists of sand with clay layers and shells in well 29Y015 and sand and marl in well 29Y010. The upper Dublin aquifer consists of sand and gravel present at 448 to 500 ft deep in well 29Y015 and is associated with relatively higher resistivity values than the adjacent confining units. In addition to sand and gravel, the upper Dublin aquifer in well 29Y015 contains dark carbonate rock grains and shells at 448 to

475 ft deep, and contains thin layers of red clay at 475 to 500 ft deep (fig. 3).The upper Dublin confining unit and upper Dublin aquifer each are 52 ft thick.

The lower Dublin confining unit is associated with an 87-ft-thick layer of mostly red clay in well 29Y015 at 500 to 587 ft deep that has lower resistivity than adjacent aquifer units (fig. 3). The driller's log for well 29Y010 describes this layer as red marl (fig. 4). The lower Dublin aquifer is associated with a transition with depth from red clay to medium to coarse sand and increasing resistivity values. In well 29Y015, the aquifer occurs at 587 to 691 ft deep for a total thickness of 104 ft.

The upper Midville confining unit is associated with a 37-ft-thick layer of red clay, and relatively low-resistivity values compared to adjacent units. At well 29Y015, the confining unit is at 691 to 728 ft deep.

At well 29Y015, the upper Midville aquifer, lower Midville confining unit, and lower Midville aquifer have similar lithology and geophysical properties consisting largely of sand and gravel with muscovite mica at 728 to 936 ft deep for a total thickness of 208 ft. The lower Midville confining unit is not lithologically distinct at well 29Y015 but is interpreted to be near the upper part of the continuous sand and gravel layer in association with a minor decrease in resistivity at 757 to 770 ft deep (fig. 3). The placement of the lower Midville confining unit makes the upper Midville aquifer 29 ft thick, the lower Midville confining unit 13 ft thick, and the lower Midville aquifer 166 ft thick. The most productive part of the lower Midville aquifer is a 31-ft-thick layer with high-resistivity values associated with sand and gravel with a low concentration of clay material at 850 to 881 ft deep in well 29Y015. The lower Midville confining unit in well 29Y009 (downtown Waynesboro) is represented by a 30-ft-thick layer of clay.

The Midville aquifer system is in contact with a basal confining unit at a depth of 936 ft in well 29Y015. The basal confining unit is composed of numerous thin red clay layers interbedded with sand and gravel. Three of the sand-and-gravel layers are thick enough to be discerned from the rest of the confining unit with drill cuttings and high-resistivity values. In well 29Y015, these three sand-and-gravel layers are at depths of 987–996, 1,044–1,054, and 1,060–1,070 ft. The bottom of the basal confining unit is at a depth of 1,109 ft in well 29Y015. Beneath the basal confining unit lies crystalline rock, likely schist, as evidenced in drill cuttings.

## Well Construction and Wellbore-Flowmeter Surveys

Flowmeter surveys were used to determine the relative contribution of water from hydrogeologic units at wells 29Y015 and 29Y010 (figs. 7 and 8; table 1). Each wellbore-flowmeter survey was followed by grab-water sampling at selected depths in an attempt to identify zones contributing high concentrations of iron to the well.

The first flowmeter survey was completed in the old production well (29Y010) at the north well field during April 2011. Well 29Y010 was constructed with 16-inch-diameter casing to a depth of 350 ft, where it transitions to a 12-inch-diameter casing and screen line from 387 ft to a depth of 610 ft (fig. 8). An 8-inch-diameter casing and screen line extends from 610 ft to a total depth of 867 ft. One screened interval in the 12-inch-diameter section is open to the upper Dublin aquifer at 450 to 480 ft deep. Two screened intervals in the 8-inch-diameter section exist, one open to the lower Dublin aquifer at 630 to 690 ft deep the other open to the Midville aquifer system at 730 to 840 ft deep. Wellbore-flowmeter-survey results indicate that nearly three quarters of the flow to the old production well was derived from the Midville hydrolgeologic units open to the bottom screen below a depth of 730 ft (fig. 8). Of the flow to the well, 9.1 percent was derived from the bottom of the Gordon aquifer at a depth of about 350 ft (figs. 4 and 8) near the base of the 16-inch-diameter casing. The upper Dublin confining unit contributed 7.5 percent of the flow to the well through the 12-inch-diameter casing at 400 to 450 ft deep. The lower Dublin aquifer contributed 6.5 percent of the flow near the transition from the 12-inch-diameter casing to the 8-inch-diameter casing at a depth of 610 ft. No measurable flow came from the screened interval open to the upper Dublin aquifer at 450 to 480 ft deep, and 3.5 percent of the flow came from the screen interval installed at 630 to 690 ft deep, open to the lower Dublin aquifer.

The second flowmeter survey was completed in well 29Y015 during August 23, 2011 (fig. 7). Well 29Y015 was constructed using 26-inch-diameter casing to a depth of 405 ft that telescopes to a 12-inch-diameter casing and screen line from 385 ft to a total depth of 940 ft. A 14-inch-diameter casing extends from 365 to 386 ft deep and overlaps the 26-inch-diameter casing. Seven screened intervals were installed in the 12-inch-diameter section: three in the Dublin aquifer system at depths of 440–475, 630–655, and 670–690 ft; and four in the Midville aquifer system at depths of 730–760, 775–825, 840–890, and 900–930 ft (figs. 3 and 7). Wellbore-flowmeter-survey results indicate that 52.2 percent of the well discharge came from the three screens open to the Dublin aquifer system; the remaining 47.8 percent came from the four screens open to the Midville aquifer system (fig. 7). The deepest screened interval at 900–930 ft depth contributed less than 0.1 percent of the total flow of this well.

Well 29Y008 is 682 ft deep and contains two screens, one open to the upper Dublin aquifer at 442 to 472 ft deep, and the other open to the lower Dublin aquifer at 622 to 682 ft deep (figs. 5 and 9). Casing and screen diameters are constant with depth at 4 inches (fig. 9). A flowmeter survey was not performed in this well.

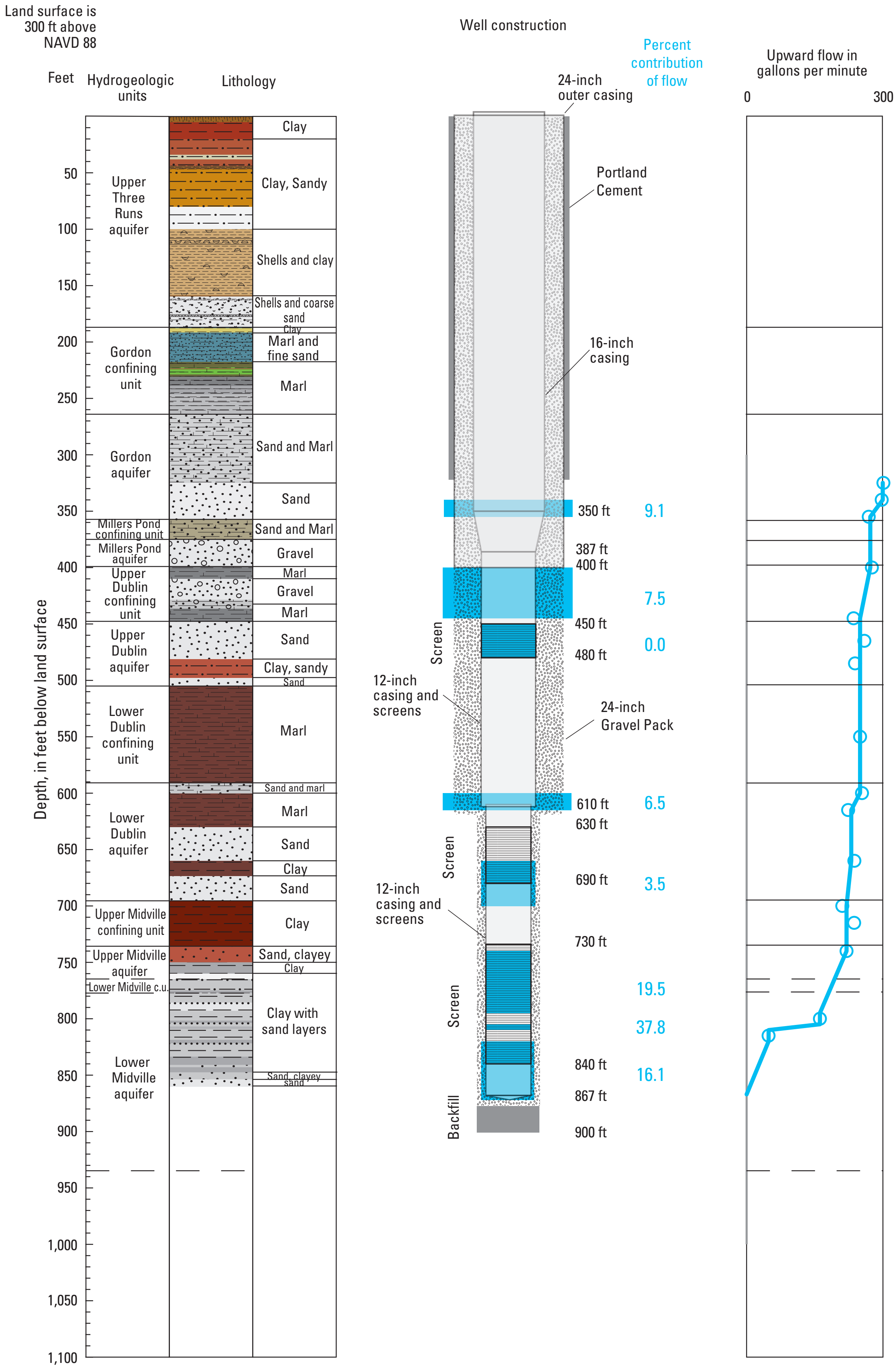



**Figure 8.** Construction and flowmeter survey results for well 29Y010 near Waynesboro, Georgia, August 2011. [c.u., confining unit.]

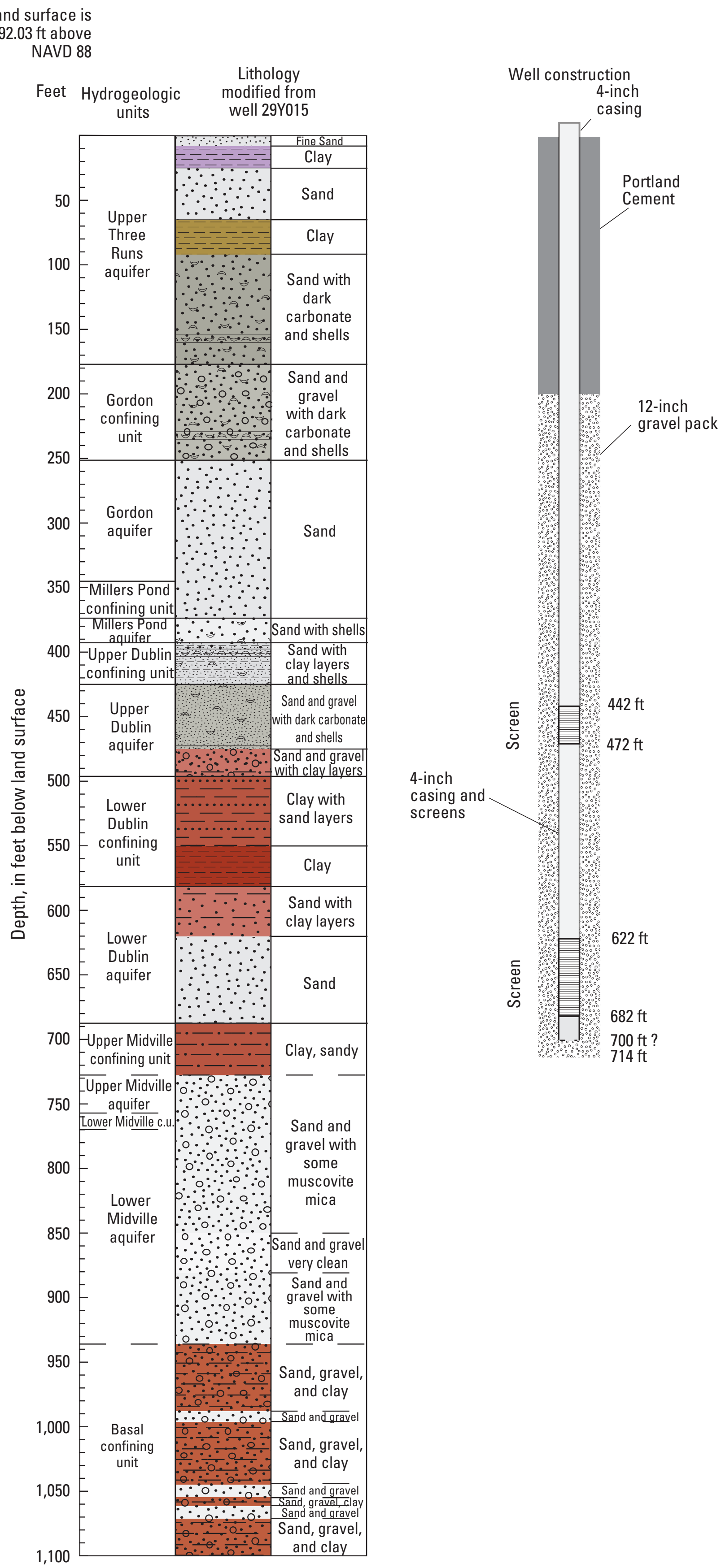



**Figure 9.** Construction for well 29Y008 near Waynesboro, Georgia, August 2011. [c.u., confining unit.]

## Aquifer Test

To determine the hydraulic properties of the Dublin and Midville aquifer systems, well 29Y015 was pumped at an average rate of 1,000 gal/min beginning at 6:45:15 a.m. on August 23, 2011, and continued for about 24 hours and 15 minutes until 7:00:00 a.m. on August 24. Water levels were monitored in the pumped well and in wells 29Y008 and 29Y010 until August 29. Water-level data from wells 29Y015 and 29Y008 were analyzed to determine hydraulic properties. Data from well 29Y010 were not analyzed, because the wellbore-flowmeter survey indicated leakage of water through the well casing, which could adversely affect test analysis. The aquifer-test-site layout and the position of screened intervals in the production well and observation wells is shown in figure 10. Factors affecting well response to the aquifer test are the wellbore storage of pumped well 29Y015 and the multiple screens open in different aquifers. The open interval for observation well 29Y008 is a subset of the open interval for pumped well 29Y015, which has screens open to the upper and lower Dublin aquifers as well as the upper and lower Midville aquifers. Observation well 29Y008 has one screen open to the upper Dublin aquifer and one screen open to the lower Dublin aquifer (figs. 5, 9, and 10). To simplify the analysis when using analytical methods, and because units are highly interconnected at the wellbore, the upper and lower Dublin aquifers were grouped into a single unit referred to as the Dublin aquifer system. Similarly, the upper and lower Midville aquifers and intervening confining unit were grouped into the Midville aquifer system. The lower Midville confining unit is not distinct as a confining unit at the aquifer-test site. The four lower screens in pumped well 29Y015 pervade all three units of the Midville aquifer system. Based on lithology and screen placement, the lower Midville confining unit is considered to contribute water to screens open to the Midville aquifer system, compared to the lower Dublin confining unit, which is not considered to contribute water to screens open to the Dublin aquifer system.

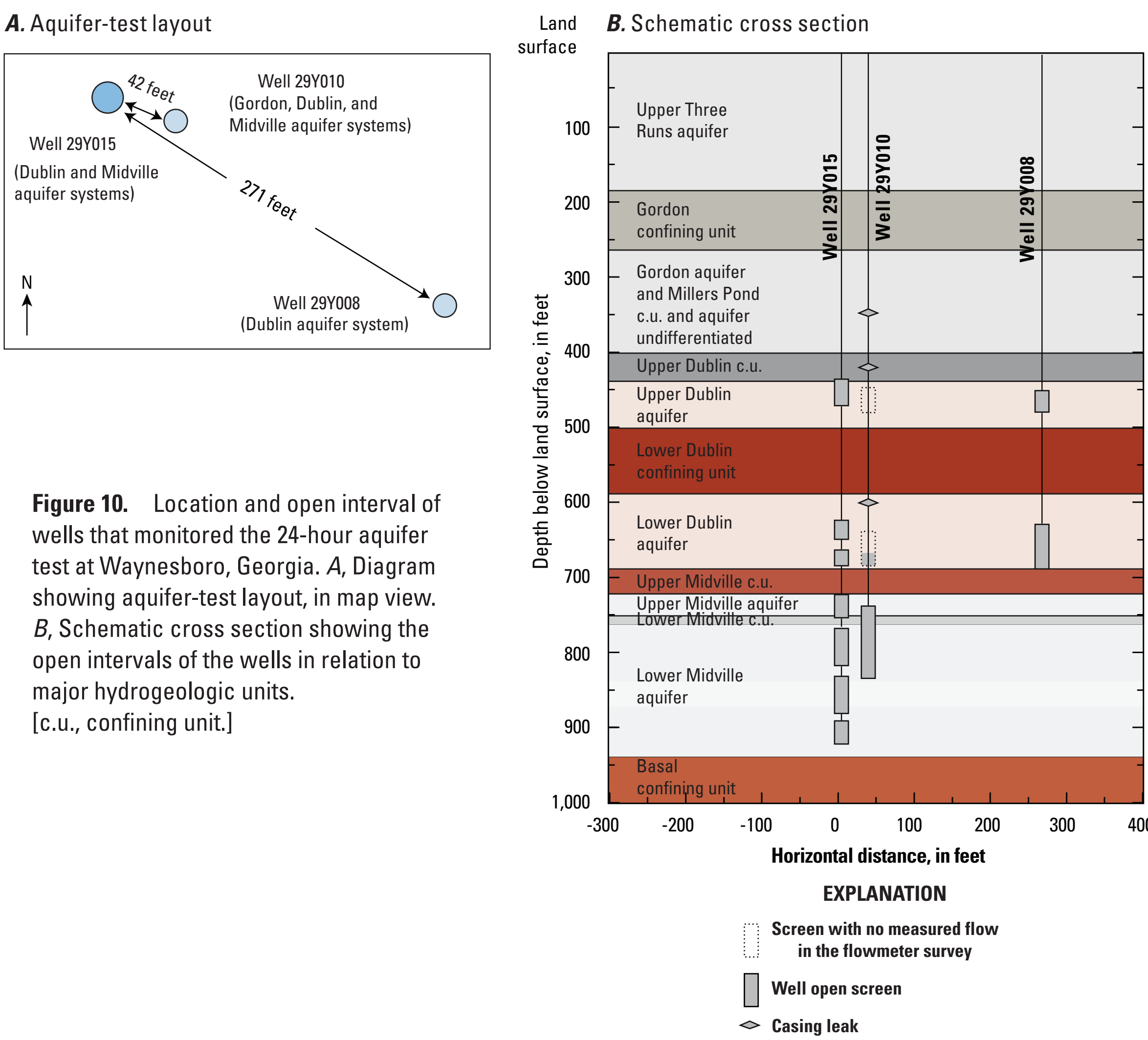



**Figure 10.** Location and open interval of wells that monitored the 24-hour aquifer test at Waynesboro, Georgia. *A*, Diagram showing aquifer-test layout, in map view. *B*, Schematic cross section showing the open intervals of the wells in relation to major hydrogeologic units. [c.u., confining unit.]

## Drawdown Estimations

Just prior to the start of the aquifer test, water levels were observed to be 134.75 and 125.72 ft deep in pumped well 29Y015 and observation well 29Y008, respectively (table 2). Drawdown in response to 24 hours of pumping ranged from 22.00 to 93.35 ft, in observation well 29Y008 and pumped well 29Y015, respectively. Background water-level trend was based on water-level data from well 29Y008 during May through July 2011. Water levels at 29Y008 were steadily declining during this period at a rate of about 0.16 feet per day (ft/d) before being affected by well-construction and development activities. The regional water-level decline caused water levels in well 29Y008 to drop below the depth of the transducer that was monitoring water level at the time (123 ft), resulting in a consistent loss of record during August 11–23, 2011 (fig. 11). Background water-level trends during the aquifer test and recovery period for wells 29Y015 and 29Y008, from which drawdown was estimated, were determined as -0.216 and -0.183 ft/d, respectively (table 2; fig. 12). The decreasing water-level trends were removed to estimate the drawdown at both wells resulting from the aquifer-test pumping.

## Aquifer-Test Analysis

Hydraulic properties of the Dublin and Midville aquifer systems were estimated using the Theis method (Theis, 1935) and an analysis tool with parameter estimation, AnalyzeHOLE (Halford, 2009). Storage coefficient could not be estimated for the Midville aquifer system because observation wells did not adequately monitor water levels in this aquifer. Analytical methods applied to pumped well 29Y015 could determine only the combined transmissivity of the Dublin and Midville aquifer systems, owing to the multiple screen intervals in all units. The proportion of total well flow contributed by specific hydrogeologic units of the Dublin aquifer system was determined from the flowmeter survey. Therefore, water-level data from observation well 29Y008 were used to estimate the transmissivity and storage coefficient of the Dublin aquifer system.

The first step of the aquifer-test analysis estimated the storage coefficient of the Dublin aquifer system using the Theis method (Theis, 1935) and water-level data from observation well 29Y008. Of the total discharge of 1,000 gal/min that was pumped from well 29Y015, borehole-flowmeter-survey data indicated that an estimated 522 gal/min was pumped from the Dublin aquifer system during the aquifer

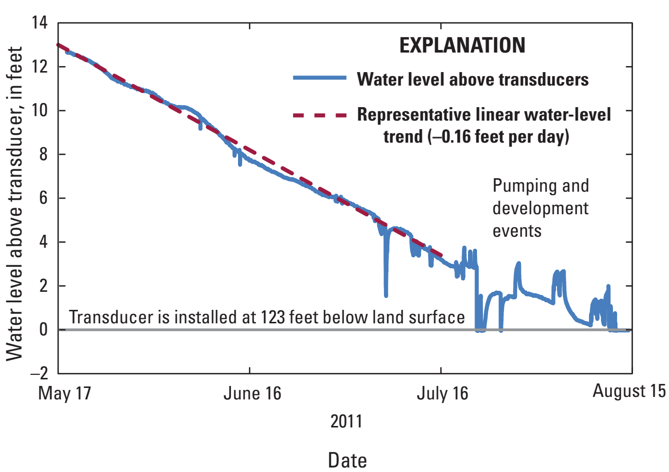



**Figure 11.** Water level above transducer and representative linear water-level trend in well 29Y008 3 months prior to the aquifer test, Waynesboro, Georgia, May–August, 2011.

**Table 2.** Well response to aquifer test in well 29Y015, just north of Waynesboro in Burke County, Georgia, August 23–24, 2011

| Well name | Linear background water-level trend, in feet per day | Static water level, in feet below land surface | 24-hour drawdown |
|---|---|---|---|
| 29Y015 | –0.216 | 134.75 | 93.35 |
| 29Y010 | –0.174 | 131.70 | 30.78 |
| 29Y008 | –0.183 | 125.72 | 22.00 |

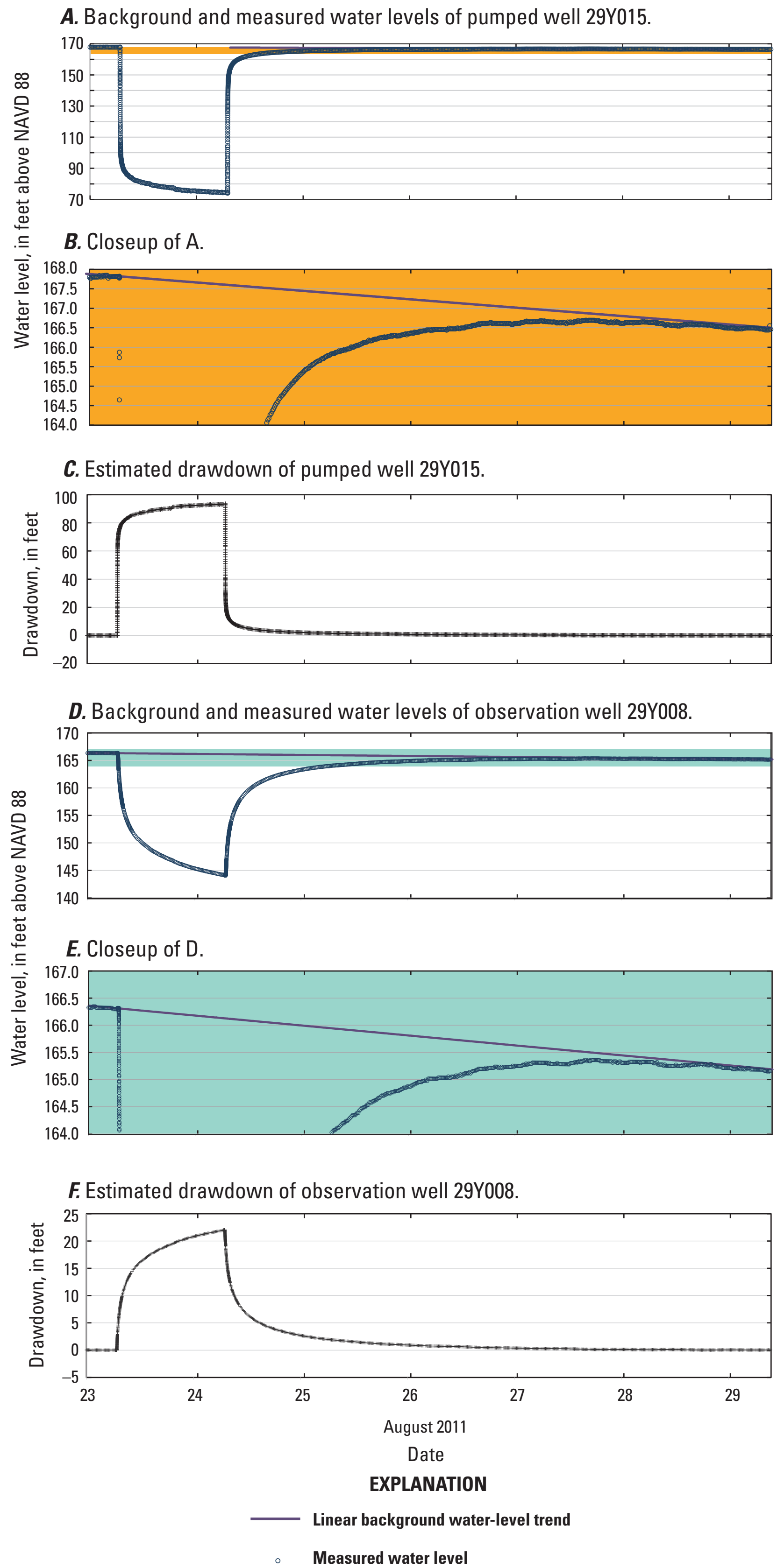



**Figure 12.** Background and measured water levels and estimated drawdown for wells 29Y015 and 29Y008, Waynesboro, Georgia, August 23–29, 2011. *A*, Representative background and measured water level for aquifer-test pumped well 29Y015; *B*, close up of *A*; *C*, estimated drawdown for pumped well 29Y015; *D*, Representative background and measured water level for observation well 29Y008; *E*, close up of *D*; and *F*, estimated drawdown for observation well 29Y008.

test (top three screens in fig. 7). Using a discharge rate of 522 gal/min, a radius of 271 ft, and the pumping schedule, a transmissivity of 1,900 feet squared per day ($ft^2/d$) and storage coefficient of 0.00032 showed the best match to field observations of drawdown and recovery at observation well 29Y008 (fig. 13; table 3). The Cooper-Jacob method (Cooper and Jacob, 1946) provided similar estimates of transmissivity and storage coefficient for the Dublin aquifer system of 1,900 $ft^2/d$ and 0.00029, respectively. With a total aquifer thickness of about 150 ft, the specific storage of the Dublin aquifer system was estimated as $2.13 \times 10^{-6}$ per day ($ft^{-1}$).

A second step of the aquifer-test analysis used the Theis method (Theis, 1935) and Papadopulos-Cooper type curves (Papadopulos and Cooper, 1967) to determine when wellbore storage effects became negligible for pumped well 29Y015. Concurrently, the combined transmissivity of the Dublin and Midville aquifer systems also was estimated using the Theis method. Storage coefficient of the combined Dublin and Midville aquifer systems, analyzed for pumped well 29Y015, was estimated as 0.00077, based on the estimated specific storage at well 29Y008 for the Dublin aquifer system ($2.13\times10^{-6}$ $ft^{-1}$), and the combined thickness of about 360 ft for both aquifer systems. Using the estimated storage coefficient of 0.00077, a discharge rate of 1,000 gal/min, radius of 0.5 ft, and the pumping schedule, the transmissivity value that yielded the best match to observed drawdown and recovery data in late time with the Theis method was 2,800 $ft^2/d$ (fig. 14). Based on the graph, wellbore storage effects become very small by 0.01 days (0.01 days is equivalent to 14 minutes 24 seconds) after a change in discharge rate. Constraining the storage coefficient to 0.00077 did not allow for the best match of simulated drawdown and recovery to observed drawdown and recovery. An unrealistically low value of storage coefficient ($5\times10^{-5}$), which converts to a specific storage of $1.45\times10^{-7}$ $ft^{-1}$, led to a better match than the match attained using an estimated storage coefficient of 0.00077. Note: A specific storage of $1.45\times10^{-7}$ $ft^{-1}$ is less than the lowest reported values of specific storage for sand or gravel and is less than the specific storage that is derived from the compressibility of water (Domenico, 1972; and Mercer and others, 1982). The Cooper-Jacob method (Cooper and Jacob, 1946) provided approximately the same estimate as the Theis method of combined transmissivity for the Dublin and Midville aquifer systems of 3,300 $ft^2/d$ (table 3).

Transmissivity values of the Dublin and Midville hydrogeologic subunits were estimated using the software AnalyzeHOLE (Halford, 2009). Horizontal hydraulic conductivity was assumed to be homogenous within each hydrogeologic subunit (representing a lithologically homogenous layer); specific

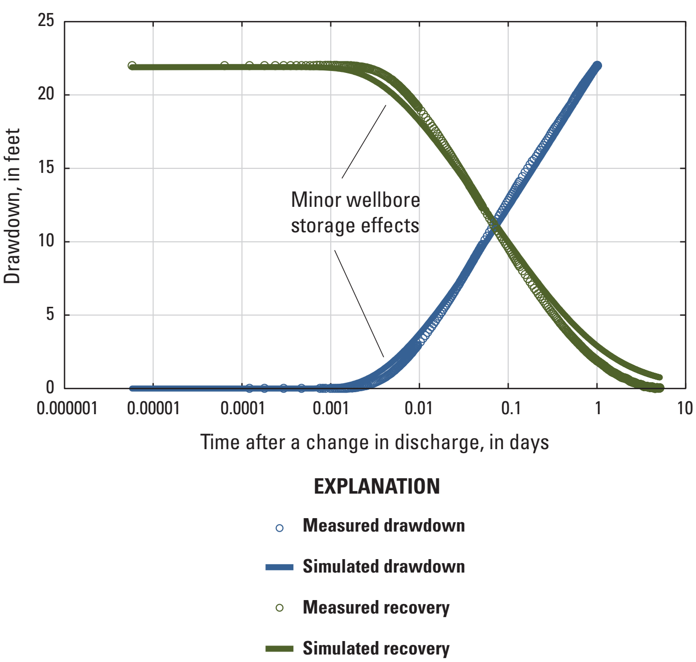



**Figure 13.** Simulated and measured drawdown in well 29Y008 in response to the 24-hour aquifer test at well 29Y015 near Waynesboro, Georgia, August 23–29, 2011. Drawdown during the aquifer test and recovery after the aquifer test are superimposed on the graph. The Theis method (Theis, 1935) and superposition were used for the simulation. Discharge was 522 gallons per minute; transmissivity was 1,906 feet squared per day; storativity was 0.00032; distance from pumped well 29Y015 was 271 feet.

**Table 3.** Methods and final determined hydraulic characteristics for the Dublin and Midville aquifer systems.

[Methods used were the Theis method (1935), Cooper-Jacob method (1946), and AnalyzeHOLE analysis tool (Halford, 2009); $ft^2/d$, feet squared per day; ft/d, feet per day]

| Method | Hydraulic characteristic | Aquifer system: Dublin | Aquifer system: Midville | Aquifer system: Dublin and Midville |
|---|---|---|---|---|
| | Thickness (feet) | 150 | 208 | 358 |
| SUMTheis | Transmissivity ($ft^2/d$) | 1,900 | 900[a] | 2,800[b] |
| | Storativity | $3.2 \times 10^{-4}$ | – | – |
| Cooper-Jacob | Transmissivity ($ft^2/d$) | 1900 | 1,300[a] | 3,300 |
| | Storativity | $2.9 \times 10^{-4}$ | – | – |
| AnalyzeHOLE | Transmissivity ($ft^2/d$) | 1,600 | 1,300 | 2,800[c] |
| Final results | Transmissivity ($ft^2/d$) | 2,000 | 1,000 | 3,000 |
| | Horizontal hydraulic conductivity (ft/d) | 10 | 6 | 8 |
| | Storativity | $3.0 \times 10^{-4}$ | – | – |
| | Specific storage ($ft^{-1}$) | $2.0 \times 10^{-4}$ | – | – |

[a]The transmissivity of both aquifers minus the transmissivity of the Dublin aquifer.

[b]Uses a storativity of 0.00077.

[c]The sum of the transmissivity of the Dublin and Midville aquifer systems.

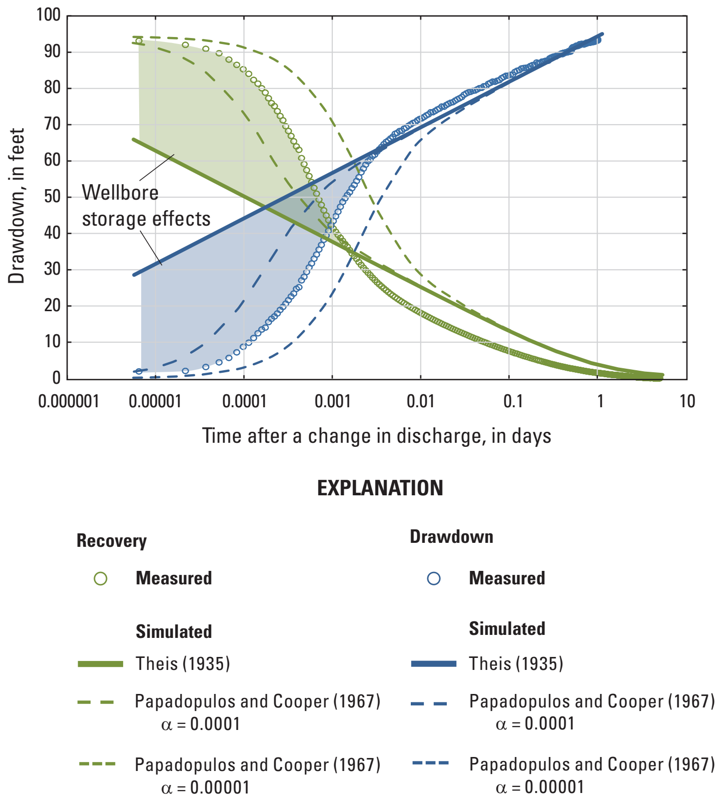



**Figure 14.** Simulated and measured drawdown in well 29Y015 in response to the 24-hour aquifer test at well 29Y015 near Waynesboro, Georgia, August 23–29, 2011. Drawdown during the aquifer test andrecovery after the aquifer test are superimposed on the graph. The Theis (1935) and Papadopulos and Cooper (1967) methods and superposition were used for the simulation. Discharge was 1,000 gallons per minute; transmissivity was 2,810 feet squared per day; storativity was 0.00077; screen radius 0.5 feet. Wellbore storage effects abate by 0.01 days. Value for Papadopulos and Cooper (1967) variable ρ was 1.

storage and vertical anisotropy ratio (vertical hydraulic conductivity divided by the horizontal hydraulic conductivity) of $2.13\times10^{-6}$ ft$^{-1}$ and 0.1581, respectively, were assumed for the entire hydrogeologic section. The vertical anisotropy ratio of 0.1581 was used because it was found, during sensitivity analysis, to yield the best match of model results with the data (see app. 1).

The axisymmetric model domain was discretized into 120 rows that represented the vertical dimension and various hydrogeologic-unit boundaries within the subsurface depth interval between 133 and 937 ft below land surface; 57 columns represented the radial distance (200,000 ft) from pumped well 29Y015 to the external model boundary (fig. 15). Radial grid spacing (column width) increased from 0.5 ft at pumped well 29Y015 to 37,037 ft at the edge of the model. Each row height represented a vertical thickness of 7.8083 ft for the simulated aquifers and intervening confining units.

Hydrogeologic units represented in the model are the saturated part of the Upper Three Runs aquifer and the Gordon confining unit undifferentiated, Gordon aquifer, upper Dublin confining unit, upper Dublin aquifer, lower Dublin confining unit, lower Dublin aquifer, upper Midville confining unit, upper Midville aquifer, lower Midville confining unit, lower Midville aquifer, and the basal confining unit (fig. 15). The 120 model rows (layers) were grouped into 30 subunits of equal lithology. Each model layer and subunit exists within a hydrogeologic unit. An initial horizontal hydraulic conductivity was assigned to each lithology. The initial horizontal hydraulic conductivities for the lithologies ranged from about 0.1 to 15 ft/d. The AnalyzeHOLE parameter “K-Lithology Bound” was set at 1,000, meaning that the final estimated horizontal hydraulic conductivity could be between 0.001 and 1,000 times that of the initially assigned value.

The radial edge distal to the well, center of the well, and edges adjacent to the base and top (first and last row) of the model were simulated as no-flow boundaries; the upper boundary (first row), represents the water table. The distal edge being 200,000 ft from the pumping well is beyond the radius of the influence of the pumping well. The aquifer-test stress period of 1.01 days was approximately represented using 60 timesteps. Timesteps during the pumping stress period ranged from 0.31 second to 4 hours 2 minutes 24 seconds, with each successive timestep increasing by a factor of 1.2.

Horizontal hydraulic conductivity was estimated for subunits and the simulated versus observed data with the final estimated values are shown in figure 16. The observation data used for model calibration were from 0.01 (about 14 minutes) to 1 day following the start of the aquifer test, after wellbore storage effects were no longer a factor. The weighting used for AnalyzeHOLE was subjectively determined, with observed flows having the highest weight, then drawdown and the target composite transmissivity the least weight: 1, 0.25, and 0.0032, respectively (table 4). The difference between simulated drawdown to observed drawdown ranged from 0.06 to 0.87 ft following the start of the aquifer test after wellbore storage effects were no longer a factor. The difference between simulated and measured wellbore flows ranged from 0.017 to 3.3 gal/min. Table 4 provides information for the data and parameters that were used to estimate the horizontal hydraulic conductivity with AnalyzeHOLE (Halford, 2009).

Results of the simulation and parameter estimation indicate a good fit between simulated and observed values of wellbore flow and drawdown (fig. 16). Table 5 shows the estimated horizontal hydraulic conductivity of the subunits within the Dublin and Midville systems and the calculated transmissivity (thickness multiplied by hydraulic conductivity) for each subunit and the total for each aquifer. Estimated transmissivities of the Dublin and Midville aquifer systems were approximately 1,600 and 1,300 ft$^2$/d, respectively, with a total transmissivity of approximately 3,000 ft$^2$/d. Based on a thickness for the Dublin aquifers of about 150 ft, the horizontal hydraulic conductivity of the Dublin aquifers is estimated to be about 10 ft/d; for a model thickness of the Midville aquifer system of about 210 ft, the horizontal hydraulic conductivity of the Midville aquifer system is estimated to be about 6 ft/d (table 5). Confining subunits within the Dublin and Midville aquifer units had estimated horizontal hydraulic conductivity of 0.01 to 0.5 ft/d.

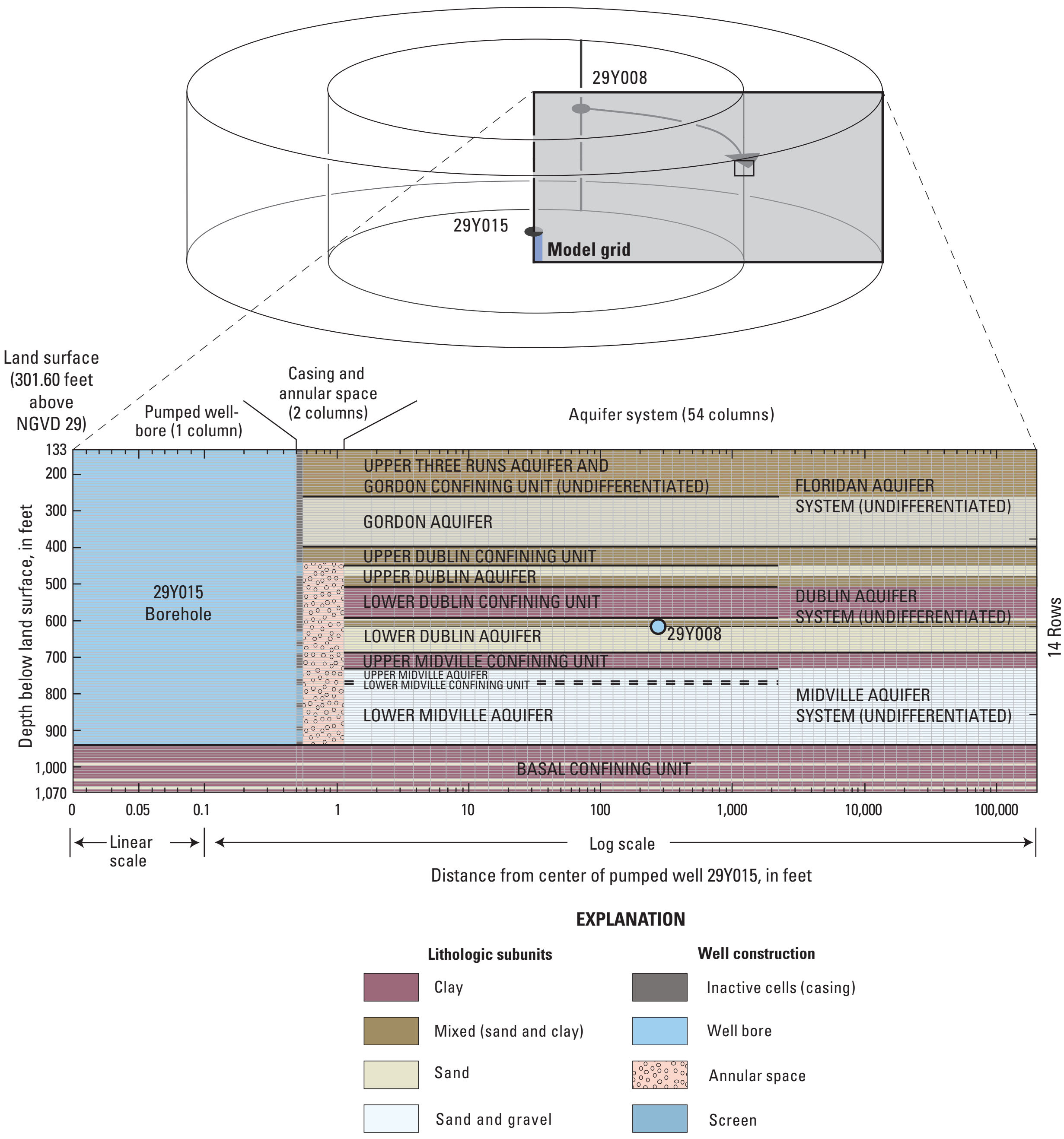



**Figure 15.** Axisymmetric model for 24-hour aquifer test at pumped well 29Y015 just north of Waynesboro, Georgia, August 23–24, 2011. Observation well 29Y008 and pumped well 29Y015 are included on the top diagram. Curved arrows in the top diagram represent transforming three-dimensional location of well opening to the two-dimensional model grid.

**Table 4.** Selected input data and parameters used for the AnalyzeHOLE run on the aquifer test at well 29Y015, just north of Waynesboro in Burke County, Georgia, August 23–24, 2011.

[ft/d, feet per day; ft²/d, feet squared per day; ft bls, feet below land surface]

| Parameter | Entry | Units |
|---|---|---|
| Number of columns | 57 | radial |
| Number of rows | 120 | vertical |
| Stress period | 1.01 | days |
| Time steps | 60 | |
| Time multiplier | 1.2 | |
| Well radius | 0.5 | feet |
| Annulus radius | 1.083 | feet |
| Model extent radius | 200,000 | feet |
| Wellbore hydaulic conductivity | 3,889,124 | ft/d |
| Specific yield for top layer | 0.1 | |
| Specific storage, all except the top layer | $2.13 \times 10^{-6}$ | 1/ft |
| Vertical anisoptropy | 0.158 | |
| Prepumping water table | 133 | ft bls |
| Target transmissivity | 5,000 | ft²/d |
| Calibration weight on drawdown | 0.25 | |
| Calibration weight on discharge | 1 | |
| Calibration weight on target transmissivity | 0.003 | |
| 29Y015 depth to top of top screen | 440 | ft bls |
| 29Y015 depth to base of bottom screen | 940 | ft bls |
| 29Y015 radius | 0.5 | feet |
| 29Y008 depth to top of top screen | 590 | ft bls |
| 29Y008 depth to base of bottom screen | 640 | ft bls |
| 29Y008 radius | 271.3 | feet |

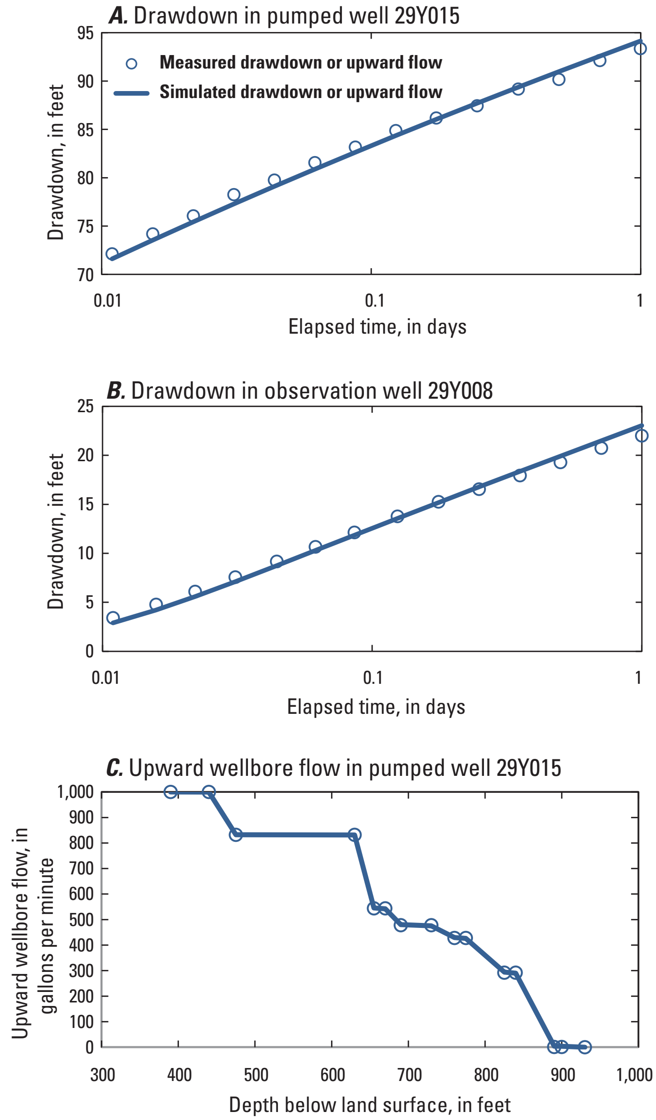



**Figure 16.** Simulated and measured drawdown and wellbore flow during an aquifer test on well 29Y015 near Waynesboro, Georgia, August 23–29, 2011: *A*, drawdown in pumped well 29Y015; *B*, drawdown in observation well 29Y008; *C*, upward wellbore flow in pumped well 29Y015. Parameter estimation (PEST) in AnalyzeHOLE analysis tool was used to match simulated values to measured values.

**Table 5.** AnalyzeHOLE model output of transmissivity for the aquifer test at well 29Y015 just north of Waynesboro in Burke County, Georgia, August 23–24, 2011.

[Input settings are listed in table 4. $ft^2/d$, feet squared per day; –, not applicable]

| Hydrogeologic unit | Depth to top of hydrogeologic unit, in feet below land surface | Thickness, in feet | Transmissivity model output, in $ft^2/d$ | Hydraulic conductivity, in ft/d | Specific storage, in $ft^{-1}$ | Storage coefficient |
|---|---|---|---|---|---|---|
| Gordon aquifer | 260 | 140 | 1,971.51 | 14.08 | $2.13x10^{-6}$ | $2.98x10^{-4}$ |
| Upper Dublin confining unit | 400 | 50 | 25.04 | 0.50 | $2.13x10^{-6}$ | $1.07x10^{-4}$ |
| Upper Dublin aquifer | 450 | 50 | 420.01 | 8.40 | $2.13x10^{-6}$ | $1.07x10^{-4}$ |
| Lower Dublin confining unit | 500 | 90 | 0.51 | 0.01 | $2.13x10^{-6}$ | $1.92x10^{-4}$ |
| Lower Dublin aquifer | 590 | 100 | 1,147.49 | 11.47 | $2.13x10^{-6}$ | $2.13x10^{-4}$ |
| Upper Midville confining unit | 690 | 40 | 0.22 | 0.01 | $2.13x10^{-6}$ | $8.52x10^{-5}$ |
| Upper Midville aquifer | 730 | 30 | 96.83 | 3.23 | $2.13x10^{-6}$ | $6.39x10^{-5}$ |
| Lower Midville confining unit | 760 | 10 | 34.29 | 3.43 | $2.13x10^{-6}$ | $2.13x10^{-5}$ |
| Lower Midville aquifer | 770 | 170 | 1,133.69 | 6.67 | $2.13x10^{-6}$ | $3.62x10^{-4}$ |
| Basal confining unit | 940 | 130 | 393.27 | 3.03 | $2.13x10^{-6}$ | $2.77x10^{-4}$ |
| Base | 1,070 | – | – | – | – | – |
| Dublin aquifer system | 400 | 150 | 1,567.50 | 10.45 | $2.13x10^{-6}$ | $3.20x10^{-4}$ |
| Midville aquifer system | 690 | 210 | 1,264.81 | 6.02 | $2.13x10^{-6}$ | $4.47x10^{-4}$ |
| Total | 400 | 360 | 2,832.31 | 7.87 | $2.13x10^{-6}$ | $7.67x10^{-4}$ |

## Water Quality

Wellbore water samples were collected from well 29Y015 while pumping at a rate of 1,000 gal/min (fig. 7). Five sample-collection points were located just above the screened intervals at depths of 440–475, 630–655, 730–760, 840–890, and 900–930 ft (table 6). Water type borders on calcium bicarbonate and sodium bicarbonate based on water from the five sample intervals and the total flow from the well (fig. 17). Water collected at each sample-collection point is a composite of all water entering the well below the sample-collection point. Water-quality results from composite samples with known flow contribution from screens from the flowmeter survey were used in the mixing equation (eq. 5) to calculate constituent concentrations for sample intervals between sample-collection points or below the bottom sample-collection point (fig. 18; table 7). The five sample intervals represent the upper Dublin aquifer (sample interval one or SI-1), lower Dublin aquifer (SI-2), upper Midville aquifer and the upper part of the lower Midville aquifer (SI-3), the clean sand-and-gravel interval in the lower Midville aquifer (SI-4), and the lower part of the lower Midville aquifer (SI-5).

**Table 6.** Sample intervals in pumped well 29Y015 during a 24-hour aquifer test just north of Waynesboro in Burke County, Georgia, August 23–24, 2011.

| Sample interval number | Number of screens | Depth to top of top screen, in feet below land surface | Depth to base of bottom screen, in feet below land surface | Hydrogeologic unit(s) to which the sample interval is open | Percent contribution of flow |
|---|---|---|---|---|---|
| SI-1 | 1 | 440 | 475 | Upper Dublin aquifer | 16.8 |
| SI-2 | 2 | 630 | 655 | Lower Dublin aquifer | 28.8 |
| | | 670 | 690 | Lower Dublin aquifer | 6.6 |
| SI-3 | 2 | 730 | 760 | Upper Midville aquifer/Lower Midville confining unit strata[a] | 5.0 |
| | | 775 | 825 | Upper part of Lower Midville aquifer | 13.6 |
| SI-4 | 1 | 840 | 890 | Clean[b] within Lower Midville aquifer | 29.1 |
| SI-5 | 1 | 900 | 930 | Lower part of Lower Midville aquifer | 0.1 |

[a]Does not act like a confining unit at pumped well 29Y015.

[b]Sand-and-gravel layer devoid of clay material.

**Table 7.** Water-quality results for composite samples and sample intervals during a 24-hour aquifer test in pumped well 29Y015 just north of Waynesboro in Burke County, Georgia, August 23, 2011.

[Range of fluoride concentrations are due to censored values from cumulative samples. Composite-sample results and flow contribution from screens were used to calculate water quality in sample intervals. –, sample-interval values for pH were not calculated; mg/L, milligrams per liter; µS/cm, microsiemens per centimeter; µg/L, micrograms per liter; SU, pH standard units; <, less than]

| Constituent parameter | Units | Composite sample with depth range, in feet below land surface | | | | | Individual Sample Intervals (SI) with depth range, in feet below land surface | | | | | Water-quality criteria | Criteria type[j] |
|---|---|---|---|---|---|---|---|---|---|---|---|---|---|
| | | 930–900[a] | 930–840[b] | 930–730[c] | 930–630[d] | 930–440[e] | 930–900[a] | 890–840[f] | 825–730[g] | 690–630[h] | 475–440[i] | | |
| | | Well discharge | | | | | SI-5 | SI-4 | SI-3 | SI-2 | SI-1 | | |
| Total dissolved solids | mg/L | 100 | 65 | 92 | 84 | 108 | 100 | 65 | 135 | 73 | 227 | 500 | SDWR |
| Specific conductance | µS/cm | 130 | 91.9 | 142 | 151 | 165 | 130 | 91.8 | 221 | 163 | 235 | – | – |
| pH | SU | 7.24 | 6.63 | 7.06 | 7.16 | 6.97 | – | – | – | – | – | 6.5 | SDWR |
| Alkalinity | mg/L | 53.2 | 27.6 | 52.2 | 57.7 | 68.2 | 53.2 | 27.5 | 91.0 | 65.1 | 120 | – | – |
| Sodium | mg/L | 13.3 | 8.34 | 14.9 | 14.9 | 16.2 | 13.3 | 8.32 | 25.2 | 14.9 | 22.7 | 20 | HBVs |
| Potassium | mg/L | 3.93 | 1.75 | 1.37 | 1.03 | 1.02 | 3.93 | 1.74 | 0.77 | 0.57 | 0.97 | – | – |
| Magnesium | mg/L | 2.06 | 1.02 | 1.16 | 0.958 | 0.954 | 2.06 | 1.02 | 1.38 | 0.685 | 0.934 | – | – |
| Calcium | mg/L | 10.3 | 9.39 | 13.4 | 15.2 | 16.6 | 10.3 | 9.39 | 19.7 | 17.6 | 23.6 | – | – |
| Manganese | µg/L | 59.1 | 32.0 | 26.1 | 19.6 | 17.4 | 59.1 | 31.9 | 16.8 | 10.8 | 6.5 | 50 | SDWR |
| Iron | µg/L | 1,160 | 493 | 419 | 351 | 314 | 1,160 | 491 | 302 | 259 | 130 | 300 | SDWR |
| Fluoride | mg/L | <0.2 | <0.2 | 0.322 | 0.289 | 0.284 | <0.2 | 0.000–0.201 | 0.514–0.829 | 0.244 | 0.259 | 2 | SDWR |
| Chloride | mg/L | 2.81 | 3.05 | 3.97 | 3.86 | 4.03 | 2.81 | 3.05 | 5.42 | 3.71 | 4.87 | 250 | SDWR |
| Sulfate | mg/L | 11.5 | 12.5 | 16.2 | 15.6 | 15.7 | 11.5 | 12.5 | 22.0 | 14.8 | 16.2 | 500 | HBV |
| Bicarbonate | mg/L | 64.9 | 33.7 | 63.6 | 70.4 | 83.2 | 64.9 | 33.5 | 110.9 | 79.4 | 146.7 | – | – |
| Contribution of flow | Percent | 0.1 | 29.2 | 47.8 | 83.2 | 100.0 | 0.1 | 29.1 | 18.6 | 35.4 | 16.8 | | |

[a] Lower part of the lower Midville aquifer.

[b] Lower and clean part of the lower Midville aquifer.

[c] The Midville aquifer system.

[d] The Midville aquifer system and the lower Dublin aquifer.

[e] The Midville and Dublin aquifer systems.

[f] Clean part of the lower Midville aquifer.

[g] Upper Midville aquifer, lower Midville confining unit, and the upper part of the lower Midville aquifer.

[h] Lower Dublin aquifer.

[i] Upper Dublin aquifer.

[j] SDWR Secondary Drinking Water Regulation.

HBV Health-based value.

HBVs Health-based value for individuals on a 500 millgrams per day restricted sodium diet.

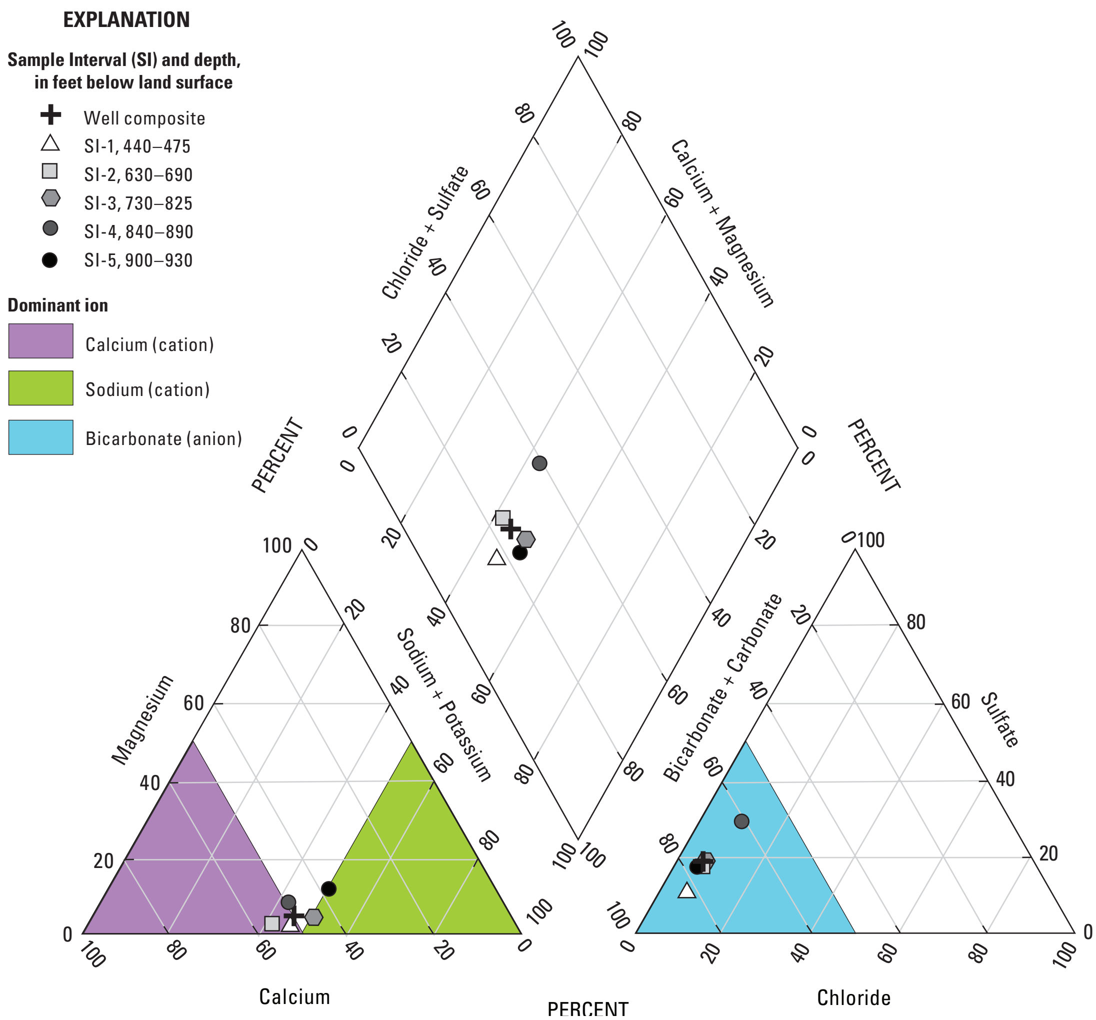



**Figure 17.** Piper diagram of water quality from sample intervals from pumped well 29Y015 during an aquifer test near Waynesboro, Georgia, August 23, 2011. Composite water sample results and flow contribution from screens were used to calculate water quality from sample intervals. Numbers are in percent of equivalents.

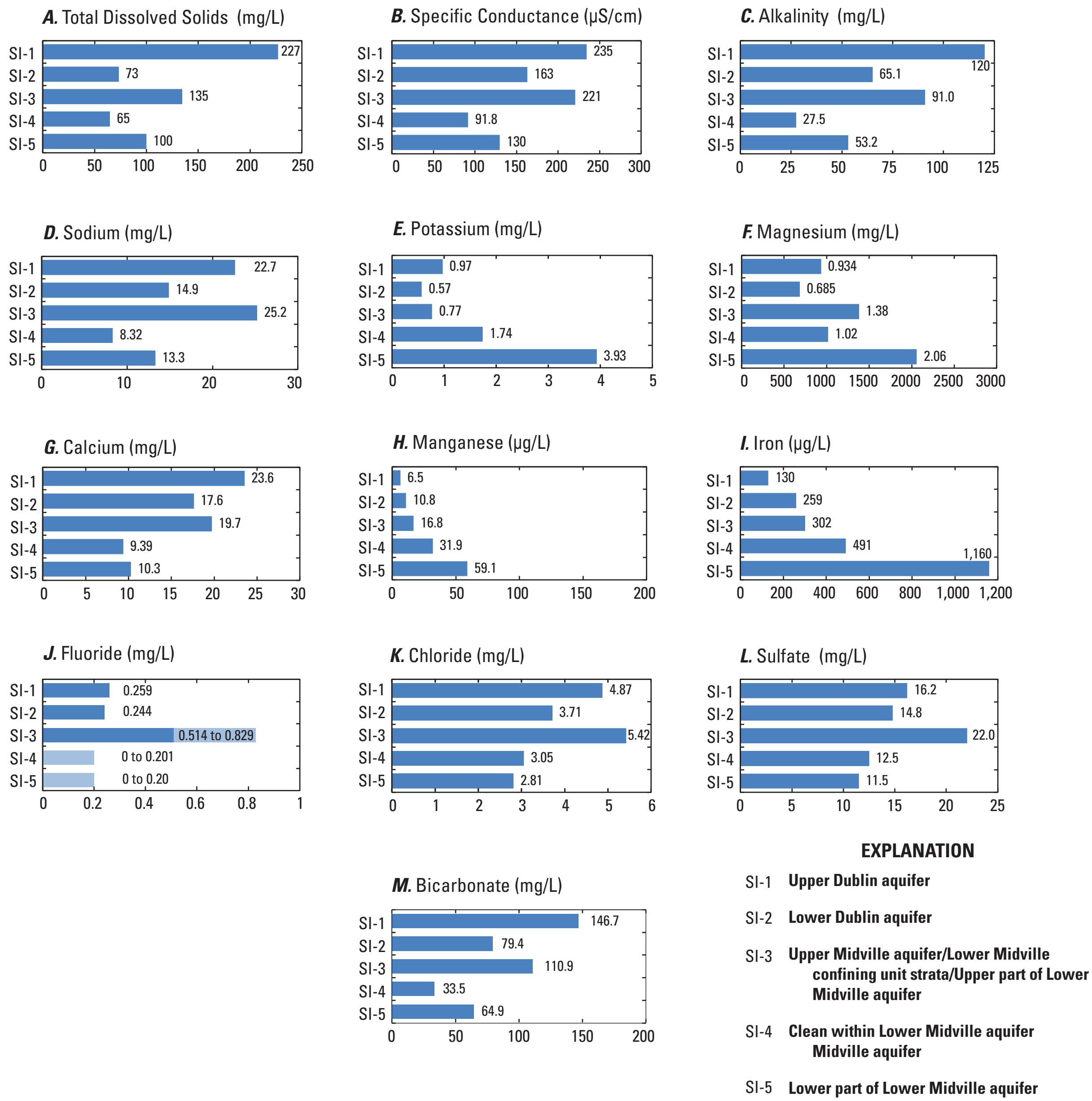



**Figure 18.** Water-quality values by sampleinterval (SI). Sample interval locations are shown in figure 7. Ranges for fluoride insample intervals are due to censored values in cumulative samples. Sample-interval specifications are listed in table 6.

Manganese and iron concentrations increased with depth. Therefore, manganese and iron concentrations were greater in water sampled from the Midville aquifer system (26.1 and 419 micrograms per liter (µg/L), respectively) than the Dublin aquifer system (9.4 and 218 µg/L, respectively). Within sample intervals manganese concentrations increased from 6.5 to 59.1 µg/L (fig. 18*H*); iron concentrations increased from 130 to 1,160 µg/L (fig. 18*I*). Manganese and iron concentrations from the lower part of the lower Midville aquifer (900 to 930 ft deep, SI-5; 59.1 and 1,160 µg/L, respectively) exceeded secondary water-quality standards of 50 and 300 µg/L, respectively (U.S. Environmental Protection Agency, 2011). Concentrations of iron in sample intervals SI-4 and SI-3 (491 and 302 mg/L, respectively), which represent almost all of the water from the Midville aquifer system, and total flow from both aquifer systems exceeded the secondary water-quality standard of 300 µg/L. Because SI-5 contributed only 0.1 percent of the total flow to the well, the total load of manganese and iron coming from this sample interval had little effect on the total concentration coming from both aquifer systems. Based on the mixing equation, excluding the contribution of SI-4 would bring the iron concentration in the total flow from the well below the secondary water-quality standards. Other major ions varied less than iron and manganese and did not exceed general water-quality standards for drinking water (table 7).

The upper Dublin aquifer (SI-1) generally contained the highest concentrations of dissolved constituents, including total dissolved solids (227 milligrams per liter (mg/L)), alkalinity (120 mg/L as calcium carbonate), and calcium (23.6 mg/L). The lower Dublin aquifer (SI-2) contained the lowest concentration of potassium (0.57 mg/L) and magnesium (0.685 mg/L). The upper Midville aquifer and the upper part of the lower Midville aquifer (SI-3) had the highest concentration of sodium (25.2 mg/L), fluoride (between 0.514 and 0.829 mg/L), chloride (5.42 mg/L), and sulfate (22.0 mg/L). The clean sand and gravel interval within the lower Midville aquifer (SI-4) had the lowest concentration of total dissolved solids (65 mg/L), alkalinity (27.5 mg/L as calcium carbonate), sodium (8.32 mg/L), and calcium (9.39 mg/L). The lower part of the lower Midville aquifer (SI-5) had the lowest concentration of chloride (2.81 mg/L) and sulfate (11.5 mg/L).

## Summary

The City of Waynesboro in Burke County, Georgia, has been supplied by two wells completed in the Dublin and Midville aquifer systems. One supply well (29Y010) north of the town lost productivity because iron oxide clogged the well screens and was taken out of service by 2006. The U.S. Geological Survey (USGS), in cooperation with the City of Waynesboro, used a replacement well (29Y015) located 42 feet (ft) northwest of the old production well 29Y010 to conduct site investigations during April through August 2011 to assess the hydrogeology and water quality of the Dublin and Midville aquifer systems. This assessment provided an improved regional characterization of the two aquifer systems.

Hydrogeology and water-quality results are based largely on data collected from new production well 29Y015, consisting of geophysical and drillers' logs, wellbore-flowmeter survey, 24-hour aquifer test, and the collection and chemical analysis of water samples. Data from three other wells also were used for the hydrogeologic characterization, namely, borehole geophysical data from all three wells, drillers' logs from two of the three wells, and a wellbore-flowmeter survey from one of the three wells.

The Coastal Plain sediments beneath Waynesboro consist mostly of sands and clays with some gravel and minor amounts of marl. Water-bearing units that were penetrated in well 29Y015 consist of the Upper Three Runs aquifer, Gordon aquifer system, Dublin aquifer system, and Midville aquifer system. The Millers Pond confining unit and aquifer hydrogeologic units commonly located at the top of the Dublin aquifer system are thin or absent in the area. The top of the upper Dublin confining unit occurs at a depth of 396 ft. The upper Dublin confining unit and upper Dublin aquifer each are 52 ft thick. The lower Dublin confining unit, mostly clay, is 87 ft thick. The lower Dublin aquifer is 104 ft thick. The upper Midville confining unit is 37 ft thick. At the aquifer-test site, the upper Midville aquifer, lower Midville confining unit, and lower Midville aquifer are composed of a 208-ft-thick continuous sand and gravel layer containing muscovite mica. The lower Midville confining unit is not lithologically distinct from the adjacent sand units. The top of the basal confining unit is at a depth of 936 ft in well 29Y015 and extends to basement crystalline rock at 1,109 ft.

Well 29Y015 was constructed with 26-inch-diameter casing to a depth of 405 ft and telescopes to a 12-inch-diameter casing and screen line from 385 ft to a total depth of 940 ft. Seven screened intervals were installed in the 12-inch-diameter section: three in the Dublin aquifer system at depths of 440–475, 630–655, and 670–690 ft; and four in the Midville aquifer system at depths of 730–760, 775–825, 840–890, and 900–930 ft. Wellbore-flowmeter-survey results during the aquifer test indicated that 52.2 percent of the well discharge came from the three screens open to the Dublin aquifer system; the remaining 47.8 percent came from the four screens open to the Midville aquifer system. The lower part of the lower Midville aquifer between depths of 900 and 930 ft contributed 0.1 percent of the total flow of this well.

To determine the hydraulic properties of the Dublin and Midville aquifer systems, well 29Y015 was pumped at an average rate of 1,000 gallons per minute (gal/min) beginning at 6:45:15 a.m. on August 23, 2011, and continued for about 24 hours and 15 minutes until 7:00:00 a.m. on August 24. Water-level data from wells 29Y015 and 29Y008 were analyzed to determine hydraulic properties. Factors affecting well response to the aquifer test are the wellbore storage of pumped well 29Y015, and the multiple screens open in different aquifers. The open interval for observation well 29Y008 has screens open to the Dublin aquifer system, which

is a subset of the open interval for pumped well 29Y015. Just prior to the start of the aquifer test, water levels were 134.75 and 125.72 ft deep in pumped well 29Y015 and for observation well 29Y008, respectively. Drawdown in response to 24 hours of pumping ranged from 22.00 to 93.35 ft for observation well 29Y008 and pumped well 29Y015, respectively.

The hydraulic properties of the Dublin and Midville aquifer systems were estimated using a traditional analytical solution, the Theis method; and calibration of a radial numerical model with parameter estimation, AnalyzeHOLE. Storage coefficient of the Dublin aquifer system was estimated using the Theis method with water-level data from observation well 29Y008. Of the total discharge of 1,000 gal/min pumped from well 29Y015, wellbore-flowmeter-survey data indicate that an estimated 522 gal/min was pumped from the Dublin aquifer system during the aquifer test. Transmissivity (1,900 feet squared per day ($ft^2/d$)) and storage-coefficient (0.00032) values showed the best match to field observations of drawdown and recovery at observation well 29Y008. With a total aquifer thickness of about 150 ft, the specific storage of the Dublin aquifer system was estimated as $2.13 \times 10^{-6}$ per foot ($ft^{-1}$). The Theis method and Papadopulos and Cooper type curves then were used with water-level data from pumped well 29Y015 to determine when wellbore storage effects became negligible for pumped well 29Y015. After a change in discharge rate, wellbore storage effects became negligible in 0.01 days (about 14 minutes).

Transmissivity values of the Dublin and Midville aquifer systems were estimated using the software AnalyzeHOLE and calibrated to the flowmeter-survey data of pumped well 29Y015 and observed drawdown data from pumped well 29Y015 and observation well 29Y008. Horizontal hydraulic conductivity was assumed to be homogenous within each hydrogeologic subunit (representing a lithologically homogenous layer); specific storage and vertical anisotropy ratio (vertical hydraulic conductivity divided by the horizontal hydraulic conductivity) of $2.13 \times 10^{-6}$ $ft^{-1}$ and 0.1581, respectively, were assumed for the entire hydrogeologic section. The aquifer system was simulated using a two-dimensional, axisymmetric radial, transient groundwater-flow model. The observation data used for model calibration was from 0.01 (about 14 minutes) to 1 day following the start of the aquifer test, after wellbore storage effects were no longer a factor.

Results of the simulation and parameter estimation indicate a good fit between simulated and observed values of wellbore flow and drawdown. Rounded to one significant figure, transmissivity estimates of the Dublin and Midville aquifer systems are 2,000 and 1,000 $ft^2/d$, respectively, with a total transmissivity of 3,000 $ft^2/d$. Based on a thickness for the Dublin aquifers of about 150 ft, the horizontal hydraulic conductivity of the Dublin aquifers is estimated to be about 10 ft/d; for a thickness of the Midville aquifer system of about 210 ft, the horizontal hydraulic conductivity of the Midville aquifer system is estimated to be about 6 ft/d. Confining subunits within the Dublin and Midville aquifer units had estimated horizontal hydraulic conductivity of 0.01 to 0.5 ft/d.

Wellbore water samples were collected from well 29Y015 while pumping at a rate of 1,000 gal/min. Five sample-collection points were located just above screens at depths of 440–475, 630–655, 730–760, 840–890, and 900–930 ft. Water type borders on calcium bicarbonate and sodium bicarbonate based on water from the five sample intervals and the total flow from the well. Water-quality results from composite samples, known flow contribution from screens from the flowmeter survey were used in a mixing equation to calculate constituent concentrations for sample intervals between sample-collection points or below the bottom sample-collection point. The five sample intervals represent the upper Dublin aquifer (sample interval one or SI-1), lower Dublin aquifer (SI-2), upper Midville aquifer and the upper part of the lower Midville aquifer (SI-3), the clean sand-and-gravel interval in the lower Midville aquifer (SI-4), and the lower part of the lower Midville aquifer (SI-5).

Manganese and iron concentrations increased with depth. Therefore, manganese and iron concentrations were greater in water sampled from the Midville aquifer system (26.1 and 419 micrograms per liter (µg/L), respectively) than the Dublin aquifer system (9.4 and 218 µg/L, respectively). Manganese and iron concentrations from the lower part of the lower Midville aquifer (900 to 930 ft deep, SI-5; 59.1 and 1,160 µg/L, respectively) exceeded secondary water-quality standards of 50 and 300 µg/L, respectively. Concentrations of iron in sample intervals SI-4 and SI-3 (419 and 302 µg/L, respectively), which represents almost all of the water from the Midville aquifer system, and total flow from both aquifer systems (314 µg/L) exceeded the secondary water-quality standard of 300 µg/L. Because SI-5 contributed only 0.1 percent of the total flow to the well, the total load of manganese and iron coming from this sample interval had little effect on the total concentration coming from both aquifer systems. Based on the mixing equation, excluding the contribution of SI-4 would bring the iron concentration in the total flow from the well below the secondary water-quality standards. Other major ions varied less than iron and manganese and did not exceed general water-quality standards for drinking water.

## Selected References

Brooks, Rebekah, Clarke, J.S., and Faye, R.E., 1985, Hydrogeology of the Gordon aquifer system of east-central Georgia: Georgia Geologic Survey Information Circular 75, 41 p.

Clarke, J.S., Brooks, Rebekah, and Faye, R.E., 1985, Hydrogeology of the Dublin and Midville aquifer systems of east-central Gerogia: U.S. Geological Survey Information Circular 74, 62 p.

Clemo, Tom, 2002, MODFLOW-2000 for cylindrical geometry with internal flow observations and improved water table simulation: Technical Report BSU CGISS 02-01, Boise State University, Boise, Idaho, 29 p.

Cooper, H.H., and Jacob, C.E., 1946, A generalized graphical method for evaluating formation constants and summarizing well-field history: Transactions, American Geophysical Union, v. 27, no. IV, p. 526–534.

Cunningham, W.L., and Schalk, C.W., comps., 2011, Groundwater technical procedures of the U.S. Geological Survey: U.S. Geological Survey Techniques and Methods 1–A1, 151 p. (Available only online at *http://pubs.usgs.gov/tm/1a1/*)

Doherty, John, 2005, Manual for Version 10 of PEST: Brisbane, Australia, Watermark Numerical Computing.

Domenico, P.A., 1972, Concepts and models in groundwater hydrology: McGraw-Hill, New York, 405 p.

Falls, W.F., Baum, J.S., Harrelson, L.G., Brown, L.H., and Jerden, J.L., Jr., 1997, Geology and hydrogeology of Cretaceous and Tertiary strata, and confinement in the vicinity of the U.S. Department of Energy Savannah River site, South Carolina and Georgia: U.S. Geological Survey, Water-Resources Investigations Report 97–4245, 125 p.

Fanning, J.L., and Trent, V.P., 2009, Water use in Georgia by county for 2005; and water-use trends, 1980–2005: U.S. Geological Survey Scientific Investigations Report 2009–5002, 186 p.

Garber, M.S., and Koopman, F.C., 1968, Methods of measuring water levels in deep wells: U.S. Geological Survey Techniques of Water-Resources Investigations, book 8, chap. A1, 23 p.

Halford, K.J., 2009, AnalyzeHOLE—An integrated wellbore flow analysis tool: U.S. Geological Survey, Techniques and Methods 4–F2, 46 p.

Halford, K.J., and Kuniansky, E.L., 2002, Documentation of spreadsheets for the analysis of aquifer-test and slug-test data: U.S. Geological Survey Open-File Report 02–197, 51 p.

Halford, K.J., Weight, W.D., and Schreiber, R.P., 2006, Interpretation of transmissivity estimates from single-well pumping aquifer tests: Ground Water, v. 44, no. 3, p. 467–471.

Harbaugh, A.W., Banta, E.R., Hill, M.C., and McDonald, M.G., 2000, MODFLOW-2000, the U.S. Geological Survey modular ground-water flow model—User guide to modularization concepts and the ground-water flow process: U.S. Geological Survey Open-File Report 00–92, 121 p.

Hem, J.D., 1985, Study and interpretation of the chemical characteristics of natural water: 3d ed., U.S. Geological Survey Water-Supply Paper 2254, 263 p.

Huddlestun, P.F., and Hetrick, J.H., 1991, The stratigraphic framework of the Fort Valley Plateau and the central Georgia kaolin district: Georgia Geological Society, Guidebook for the 26th Annual Field Trip, v. 11, no. 1, 119 p.

Huddlestun, P.F., and Summerour, J.H., 1996, The lithostratigraphic framework of the uppermost Cretaceous and lower Tertiary of eastern Burke County, Georgia: Georgia Geologic Survey Bulletin 127, 94 p.

Kendall, Carol, and Caldwell, E.A., 1998, Fundamentals of isotope geochemistry, *in* Kendall, Carol, and McDonnell, J.J., eds, 1998, Isotope tracers in catchment hydrology: Amsterdam, The Netherlands, Elsevier Science, 839 p.

Langevin, C.D., 2008, Modeling axisymmetric flow and transport: Ground Water, v. 46, no. 4, p. 579–590.

Mercer, J.W., Thomas, S.D., and Ross, Benjamin, 1982, Parameters and variables appearing in repository siting models: U.S. Nuclear Regulatory Commission, NUREG/CR-3066, Washington, D.C., 244 p.

Papadopulos, I.S., and Cooper, H.H., Jr., 1967, Drawdown in a well of large diameter: Water Resources Research, v. 3, no. 1, p. 241–244.

Payne, F.P., Rumman, M.A., and Clarke, J.S., 2005, Simulation of ground-water flow in coastal Georgia and adjacent parts of South Caroline and Florida—Predevelopment, 1980, and 2000: U.S. Geological Survey Scientific Investigations Report 2005–5089, 82 p.

Piper, A.M., 1944, A graphic procedure in the geochemical interpretation of water analyses: American Geophysical Union Transactions, v. 25, p. 914–923.

Reily, T.E., and Harbaugh, A.W., 1993, Computer note: Simulation of cylindrical flow to a well using the U.S. Geological Survey modular finite-difference ground-water flow model: Ground Water, v. 31, no. 3, p. 489–494.

Rutledge, A.T., 1991, An axisymmetric finite-difference flow model to simulate drawdown in and around a pumped well: U.S. Geological Survey Water-Resources Investigations Report 90–4098, 33 p.

Schwab, Delbert, 2007, Irrigation water measurement: Oklahoma State University, Oklahoma Cooperative Extension Service, BAE-1502, 4 p.

Stallman, R.W., 1971, Aquifer-test design, observation, and data analysis: U.S. Geological Survey Techniques of Water-Resources Investigations, book 3, chap. B1, 26 p.

Summerour, J.H., Shapiro, E.A., Lineback, J.A., Huddlestun, P.F., and Hughes, A.C., 1994, An investigation of tritium in the Gordon and other aquifers in Burke County, Georgia: Georgia Geologic Survey Information Circular 95, 93 p.

Theis, C.V., 1935, The relation between the lowering of the piezometric surface and the rate and duration of discharge of a well using groundwater storage: American Geophysical Union Transactions, v. 16, p. 519–524.

U.S. Environmental Protection Agency, 1976, Quality criteria for water: Washington, D.C., EPA PB-263 943, 501 p.

U.S. Environmental Protection Agency, 2011, 2011 Edition of the drinking water standards and health advisories: Washington, D.C., EPA 820-R-11-002, 12 p.

# Appendix 1. Sensitivity Analysis

AnalyzeHOLE does not estimate specific storage or anisotropy (the vertical hydraulic conductivity divided by the horizontal hydraulic conductivity). Therefore, a sensitivity analysis was performed to check or estimate these parameters by changing the specific-storage value and the anisotropy value in all layers of the model. The initial or reference values of specific storage and anisotropy are $2.13\times10^{-6}$ per foot ($ft^{-1}$) and 0.1, respectively. The sensitivity analyses assessed how changes in specific storage or anisotropy input into the model affected the root-mean-square (RMS) of the difference between the simulated and observed drawdown values, the RMS of the difference between the simulated and measured flow contribution from screen intervals, and the model estimated values of transmsissivity for the Dublin and Midville aquifer systems. The RMS of the drawdown values for the wells that is expressed in the discussion and figures was normalized by dividing it by the maximum drawdown for each well.

Specific-storage values were changed from a factor of 0.4 to 2.5 ($8.52\times10^{-7}$ to $5.32\times10^{-6}$ $ft^{-1}$). The RMS of the difference between simulated and observed drawdown values increased with increasing input of specific-storage values into the model. The RMS divided by the maximum drawdown for both wells increased by 75 percent with specific-storage values increasing from $8.52\times10^{-7}$ to $2.13\times10^{-6}$ $ft^{-1}$. With specific-storage values increasing from $2.13\times10^{-6}$ to $5.32\times10^{-6}$ $ft^{-1}$, RMS divided by the maximum drawdown for both wells increased by about 370 percent (table 1-1 and fig. 1-1).

**Table 1-1.** The root-mean-square of the difference between simulated and estimated drawdown and discharge for different values of specific-storage input into the model.

[Vertical anisotropy and target total transmissivity were 0.1 and 5,000 feet squared per day ($ft^2/d$), respectively. Parameter estimation PEST and AnalyzeHOLE analysis tool were used. RMS, root mean square; MAX DD, the maximum drawdown in the well, in feet, after 24 hours of the aquifer test pumping 1,000 gallons per minute (gal/min); ft, feet]

| | | RMS | | | | RMS/MAX DD | | | Transmissivity | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Specific storage, in $ft^{-1}$ | Factor | 29Y015, in ft | 29Y008, in ft | Both wells, in ft | RMS discharge, in gal/min | 29Y015 | 29Y008 | Both wells | Dublin aquifer system, in $ft^2/d$ | Midville aquifer system, in $ft^2/d$ | Both aquifer systems, in $ft^2/d$ |
| 8.52E-7 | 0.40 | 0.37 | 0.38 | 0.37 | 1.98 | 0.004 | 0.017 | 0.012 | 1,751 | 1,355 | 3,106 |
| 1.35E-6 | 0.63 | 0.44 | 0.58 | 0.51 | 1.92 | 0.005 | 0.026 | 0.019 | 1,687 | 1,308 | 2,994 |
| 2.13E-6 | 1.00 | 0.50 | 0.67 | 0.59 | 1.93 | 0.005 | 0.030 | 0.022 | 1,602 | 1,269 | 2,871 |
| 3.37E-6 | 1.58 | 0.83 | 1.72 | 1.35 | 1.89 | 0.009 | 0.078 | 0.056 | 1,555 | 1,226 | 2,781 |
| 5.33E-6 | 2.50 | 1.15 | 3.17 | 2.38 | 1.77 | 0.012 | 0.144 | 0.102 | 1,512 | 1,185 | 2,696 |

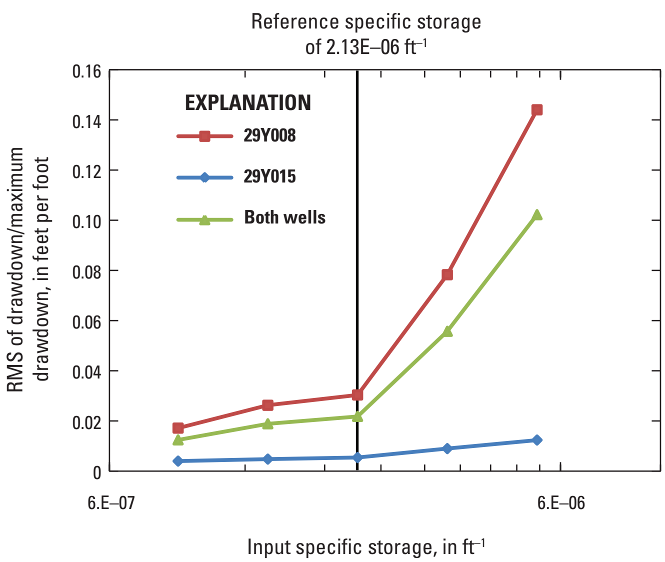



**Figure 1-1.** Root-mean-square (RMS) of the difference between simulated and estimated drawdown divided by maximum drawdown for pumped well 29Y015 and observation well 29Y008 as a function of specific storage value input into model of a 24-hour aquifer test, Waynesboro, Georgia, August 23–24, 2011. Parameter estimation (PEST) and AnalyzeHOLE analysis tool were used. Vertical anisotropy and target total transmissivity were 0.1 and 5,000 feet squared per day, respectively.

The RMS of the difference between simulated and measured contributions of flow from screens to well 29Y015 was not sensitive to changes in specific-storage values input into the model. With specific storage values increasing from $8.52\times10^{-7}$ to $5.32\times10^{-6}$ $ft^{-1}$, the RMS decreased by 10 percent (fig. 1-2 and table 1-1). This general lack of sensitivity to specific storage for aquifer tests is the reason the software AnalyzeHOLE does not estimate this parameter (Halford, 2009).

Transmissivity values were not sensitive to changes in specific-storage values input into the model. With specific-storage values increasing from $8.52\times10^{-7}$ to $5.32\times10^{-6}$ $ft^{-1}$, model estimated transmissivity values for the Dublin and Midville aquifer systems decreased by 239 and 170 feet squared per day ($ft^2/d$), respectively (fig. 1-3 and table 1-1). Rounded to the nearest single significant digit, the determined transmissivity values for the Dublin and Midville aquifer systems, 2,000 and 1,000 $ft^2/d$, respectively, do not change with a change in specific storage from $8.52\times10^{-7}$ to $5.32\times10^{-6}$ $ft^{-1}$.

Anisotropy values were changed from a factor of 0.4 to about 4 (0.04 to 0.395). The RMS of the difference between simulated and observed drawdown values divided by the maximum drawdown, for both wells, was at a minimum when the anisotropy input into the model was 0.158 (fig. 1-4 and table 1-2). Based on the sensitivity analysis, the anisotropy that was used in the final model results was 0.158.

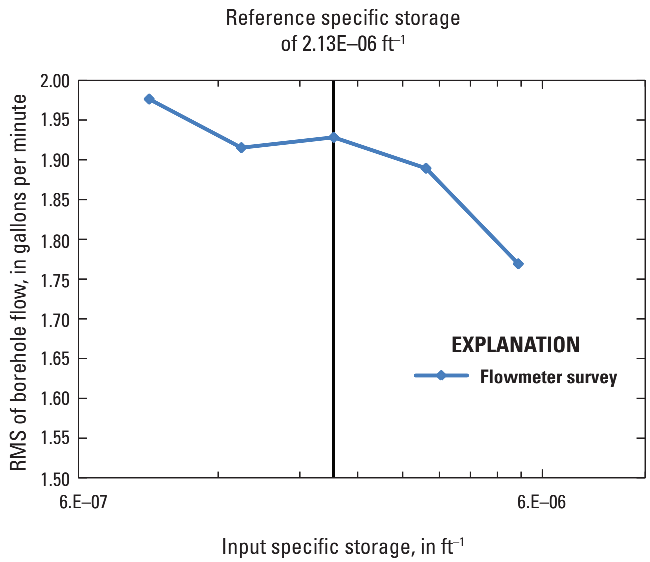



**Figure 1-2.** Root-mean-square (RMS) of the difference between simulated and measured upward borehole flow as a function of specific storage value input into model of a 24-hour aquifer test at pumped well 29Y015, Waynesboro, Georgia, August 23–24, 2011. Parameter estimation (PEST) and AnalyzeHOLE were used. Vertical anisotropy and target total transmissivity were 0.1 and 5,000 feet squared per day, respectively.

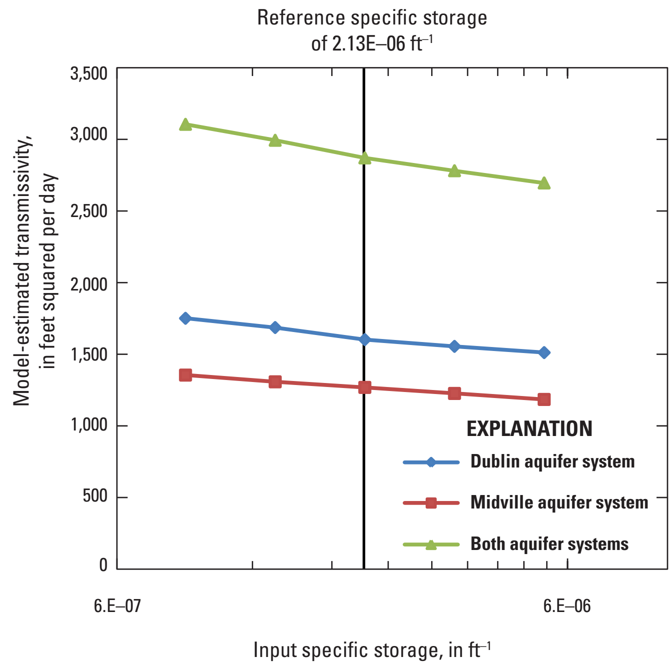



**Figure 1-3.** Model estimated transmissivity of the Dublin and Midville aquifer systems as a function of specific storage value input into model of a 24-hour aquifer test at pumped well 29Y015 and observation well 29Y008, Waynesboro, Georgia, August 23–24, 2011. Vertical anisotropy is the vertical hydraulic conductivity divided by the horizontal hydraulic conductivity. Parameter estimation (PEST) and AnalyzeHOLE analysis tool were used. Vertical anisotropy and target total transmissivity were 0.1 and 5,000 feet squared per day, respectively.

**Table 1-2.** The root-mean-square of the difference between simulated and estimated drawdown and discharge for different values of anisotropy input into the model.

[Specific storage and target total transmissivity were 2.13x10$^{-6}$ ft$^{-1}$ and 5,000 feet squared per day (ft$^2$/d), respectively. Parameter estimation PEST and AnalyzeHOLE analysis tool were used. RMS, root mean square; MAX DD, the maximum drawdown in the well, in feet, after 24 hours of the aquifer test pumping 1,000 gallons per minute (gal/min); ft, feet]

| | | RMS | | | | RMS/MAX DD | | | Transmissivity | | |
|---|---|---|---|---|---|---|---|---|---|---|---|
| Anisotropy | Factor | 29Y015, in ft | 29Y008, in ft | Both wells, in ft | RMS discharge, in gal/min | 29Y015 | 29Y008 | Both wells | Dublin aquifer system, in ft$^2$/d | Midville aquifer system, in ft$^2$/d | Both aquifer systems, in ft$^2$/d |
| 0.0400 | 0.40 | 0.47 | 1.31 | 0.98 | 1.89 | 0.0051 | 0.0593 | 0.0421 | 1,689 | 1,291, | 2,980 |
| 0.0632 | 0.63 | 0.48 | 0.86 | 0.69 | 1.93 | 0.0051 | 0.0390 | 0.0278 | 1,646 | 1,279 | 2,925 |
| 0.1000 | 1.00 | 0.50 | 0.67 | 0.59 | 1.93 | 0.0054 | 0.0303 | 0.0218 | 1,602 | 1,269 | 2,871 |
| 0.1581 | 1.58 | 0.60 | 0.52 | 0.56 | 1.91 | 0.0064 | 0.0236 | 0.0173 | 1,568 | 1,265 | 2,833 |
| 0.2500 | 2.50 | 0.66 | 0.66 | 0.66 | 1.92 | 0.0070 | 0.0300 | 0.0218 | 1,530 | 1,259 | 2,789 |
| 0.3953 | 3.95 | 0.71 | 0.91 | 0.82 | 1.91 | 0.0076 | 0.0413 | 0.0297 | 1,490 | 1,255 | 2,745 |

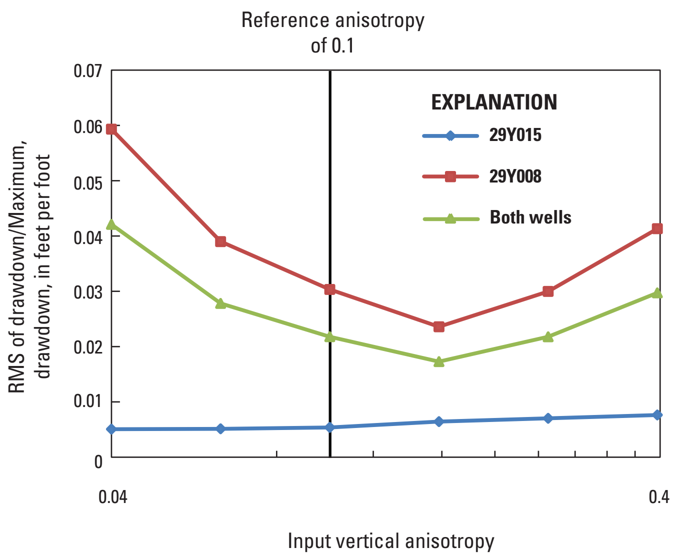



**Figure 1-4.** Root-mean-square (RMS) of the difference between simulated and estimated drawdown divided by maximum drawdown for pumped well 29Y015 and observation well 29Y008, as a function of vertical anisotropy value input into model of a 24-hour aquifer test, Waynesboro, Georgia, August 23–24, 2011. Vertical anisotropy is the vertical hydraulic conductivity divided by the horizontal hydraulic conductivity. Parameter estimation (PEST) and AnalyzeHOLE were used. Specific storage and target total transmissivity were 2.13 X 10$^{-6}$ ft and 5,000 feet squared per day, respectively.

The RMS of the difference between simulated and measured contributions of flow from screens to well 29Y015 was not sensitive to changes in anisotropy values input into the model. The RMS was within a 2-percent range for all values of anisotropy input into the model (fig. 1-5 and table 1-2).

Transmissivity values were not sensitive to changes in anisotropy values input into the model. With anisotropy values increasing from 0.04 to 0.395, model estimated transmissivity values for the Dublin and Midville aquifer systems decreased by 199 and 36 ft$^2$/d, respectively (fig. 1-6 and table 1-2). Rounded to the nearest significant digit, the model estimated transmissivity values for the Dublin and Midville aquifer systems did not change with change in anisotropy values with the exception of the transmissivity of the Dublin aquifer system using the input maximum anisotropy value of 0.395. An anisotropy value of 0.395 input into the model led to a model estimated transmissivity for the Dublin aquifer system of just under 1,500 ft$^2$/d. This general lack of sensitivity to vertical anisotropy for aquifer tests is the reason the software AnalyzeHOLE does not estimate this parameter (Halford, 2009).

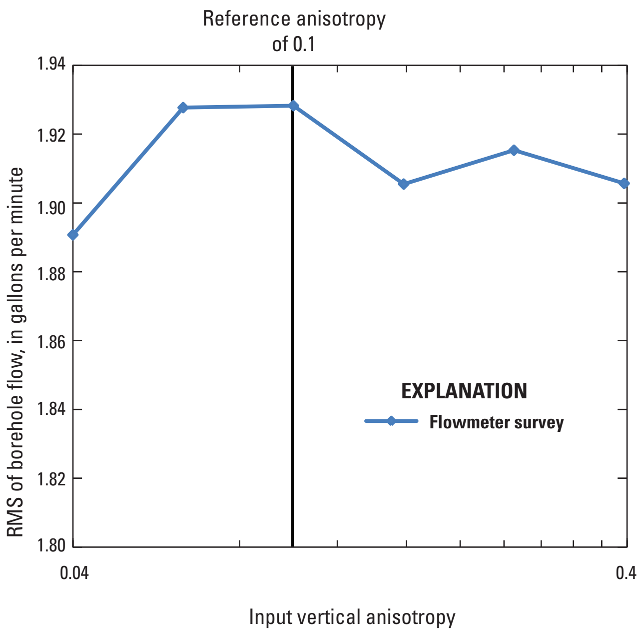



**Figure 1-5.** Root-mean-square (RMS) of the difference between simulated and measured upward borehole flow as a function of vertical anisotropy value input into model of a 24-hour aquifer test at pumped well 29Y015, Waynesboro, Georgia, August 23–24, 2011. Vertical anisotropy is the vertical hydraulic conductivity divided by the horizontal hydraulic conductivity. Parameter estimation (PEST) and AnalyzeHOLE analysis tool were used. Specific storage and target total transmissivity were 2.13x10$^{-6}$ ft$^{-1}$ and 5,000 feet squared per day, respectively.

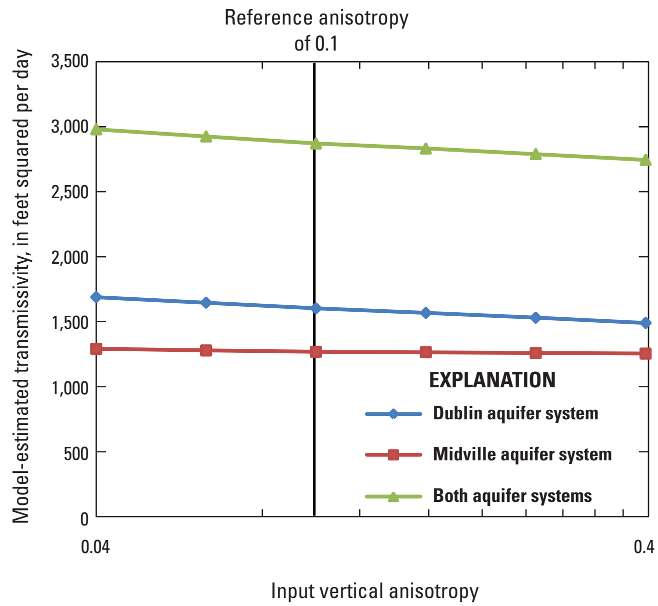



**Figure 1-6.** Model estimated transmissivity of the Dublin and Midville aquifer systems as a function of vertical anisotropy value input into model of a 24-hour aquifer test at pumped well 29Y015 and observation well 29Y008, Waynesboro, Georgia, August 23–24, 2011. Vertical anisotropy is the vertical hydraulic conductivity divided by the horizontal hydraulic conductivity. Parameter estimation (PEST) and AnalyzeHOLE analysis toolwere used. Specific storage and target total transmissivity were 2.13x10$^{-6}$ ft$^{-1}$ and 5,000 feet squared per day, respectively.

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