Hydrogeologic Setting and Ground-Water Flow Simulations of the Great Miami River Basin Regional Study Area, Ohio

By Rodney A. Sheets

Section 7 of

Hydrogeologic Settings and Ground-Water Flow Simulations for Regional Studies of the Transport of Anthropogenic and Natural Contaminants to Public-Supply Wells—Studies Begun in 2001

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Hydrogeologic Setting and Ground-Water Flow Simulations of the Great Miami River Basin Regional Study Area, Ohio

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Abstract

The transport of anthropogenic and natural contaminants to public-supply wells was evaluated for the glacially derived valley-fill aquifer underlying the Great Miami River and its tributaries near Dayton, Ohio, as part of the U.S. Geological Survey National Water-Quality Assessment Program. The glacial valley-fill aquifer in the Great Miami River regional study area is representative of the glacial aquifer system in the North-Central United States. Water needs in the study area for industry, agriculture, and public water supply are met almost entirely by pumping from the glacial aquifer, and the aquifer is susceptible and vulnerable to contamination. An existing three-layer, steady-state ground-water flow model of the study area was modified to include some bedrock islands within the glacial valleys and recalibrated with parameter estimation to represent average ground-water flow conditions for the period from 1997 to 2001. The calibrated model and advective particle-tracking simulations were used to compute areas contributing recharge and traveltimes from recharge areas to 60 public-supply wells. Model results indicate streamflow loss from streams (39 percent of inflow) and precipitation (39 percent of inflow) provide most of the ground-water inflow, while the majority of ground-water discharge is to pumping wells (53 percent of outflow) and gaining stream reaches (40 percent of outflow). Median simulated traveltimes from recharge areas to wells ranged from 21 days to 184 years. Approximately 73 percent of the traveltimes were less than 25 years indicating water quality in the aquifer is susceptible to the affects of overlying land use.

Introduction

The Great Miami River Basin regional study area for the transport of anthropogenic and natural contaminants to public-supply wells (TANC) study is located in the east-central portion of the White River-Great and Little Miami River Basin National Water-Quality Assessment (NAWQA) study unit (fig. 7.1). Nationwide, the glacial aquifer system is the largest in areal extent of any principal aquifer and is an important source of water for public water supply providing public and domestic water for approximately 41 million people in 2000 (Warner and Arnold, 2005).

Purpose and Scope

The purpose of this Professional Paper section is to present the hydrogeologic setting of the Great Miami River Basin regional study area. The section also documents the setup and calibration of a steady-state regional ground-water flow model for the study area. Ground-water flow characteristics, pumping-well information, and water-quality data were compiled from existing data to develop a conceptual understanding of ground-water conditions in the study area. An existing ground-water flow model was modified to include some bedrock islands within the glacial valleys and was recalibrated with parameter estimation to represent average conditions for the period from 1997 to 2001. The 5-year period 1997–2001 was selected for data compilation and modeling exercises for all TANC regional study areas to facilitate future comparisons between study areas. The recalibrated ground-water flow model and associated particle tracking were used to simulate advective ground-water flow paths and to delineate areas contributing recharge to selected public-supply wells. Ground-water traveltimes from recharge to public-supply wells, oxidation-reduction (redox) conditions along flow paths, and the presence of potential contaminant sources in areas contributing recharge were tabulated into a relational database as described in Section 1 of this Professional Paper. This section provides the foundation for future ground-water susceptibility and vulnerability analyses of the study area and comparisons among regional aquifer systems.

Figure 7.1. Location of the Great Miami River Basin regional study area within the glacial aquifer system.

Study Area Description

The study area is centered on the city of Dayton, Ohio (fig. 7.2), and examines the effects of pumping unconfined ground water from the unconsolidated, glacially derived valley-fill aquifer underlying the Great Miami River and its tributaries. The Great Miami River Basin regional study area represents the glacial aquifer system (Warner and Arnold, 2005) in the North-Central United States (table 7.1) with ground water occurring in the valley-fill glacial aquifer underlying the Great Miami River and its tributaries. The study area includes the confluence of the Great Miami River and Mad River at Dayton (1,700 km^2) and several tributaries to the Great Miami River, including the Stillwater River (1,750 km^2). The study area also includes the parts of the upper Little Miami River drainage basin. In the Great Miami River Basin regional study area, water needs for industry, agriculture, and public water supply are met almost entirely by pumping from the glacial aquifer.

The primary aquifer within the study area consists of Quaternary glacial deposits overlying Ordovician and Silurian bedrock (Dumouchelle, 1998). The glacial deposits are primarily a result of glacial meltwater or outwash deposits left by retreating continental glaciers. The most productive aquifers in the White-Miami study unit are those that were deposited in buried-valley settings underlying the Great Miami River (fig. 7.2). The buried-valley aquifer in the study area is heavily used by industry and municipalities with well yields commonly greater than 5,300 m^3/d (Dumouchelle, 1998). These high pumping rates often induce infiltration from nearby rivers or artificial recharge lagoons. The buried-valley aquifer has been designated a Sole-Source Aquifer by the U.S. Environmental Protection Agency (Debrewer and others, 2000). Low permeability Ordovician shale with well yields less than approximately 6 m^3/d underlies the buried-valley aquifer beneath most of the study area; Silurian limestone and dolomite form a thin carbonate aquifer that underlies the buried-valley aquifer in the higher elevations of the study area (Dumouchelle, 1998).

The study area is in the Till Plains section of the Central Lowland Physiographic Province. The topography of the till plains was formed by several continental glaciations resulting in a flat to gently rolling land surface (Fenneman, 1938). Bedrock features formed by pre- and periglacial drainage systems were buried under the glacial deposits (Dumouchelle, 1998). In areas where modern rivers dissect the land surface, topographic relief is low to moderate, with steep-walled valleys formed by the river drainages. Surface drainage through the study area is from north to south, toward the Ohio River. Average annual precipitation is 96.5 cm (Harstine, 1991).

Even though Dayton lies near the center of the study area (fig. 7.2), approximately 44 percent of the land use in the study area (2001) is agricultural (row crops, pasture, and so forth); about 41 percent of the land use in the study area is developed. Primarily, the upper reaches of the subbasins are agricultural, and the areas along the main drainages are suburban to urban. The remaining land uses are forested or grasslands (13 percent) and wetlands, quarries, or open water (2 percent) (Homer and others, 2004).

Water use in the study area consists almost entirely of ground-water supplies for domestic, agricultural, industrial and public-supply users. In 1993, Dayton supplied approximately 285,000 m^3/d of water to the areas in and around the city with two well fields (Dumouchelle, 1998). The Dayton Rohrers' Island well field lies along the Mad River, just south of Wright Patterson Air Force Base, and supplied approximately 190,000 m^3/d. The Miami River well field, the second Dayton well field, is located along the Great Miami River and consists of about 50 wells that pumped ground water at the rate of approximately 95,000 m^3/d in 1993. The remaining water use in the study area, approximately 245,000 m^3/d, is provided by other public ground-water supplies (for example, Wright Patterson Air Force Base, smaller cities, and hospitals), industrial ground-water supplies, and agricultural ground-water supplies (Dumouchelle, 1998).

Conceptual Understanding of the Ground-Water System

Ground-water flow is controlled by the geology and the hydraulic properties of the geologic materials. Areal recharge and discharge to streams in this system are the primary inputs and outputs to this system; the ground-water chemistry seems to be controlled by the residence time (from recharge to discharge) and geology. The generalized geologic section in figure 7.3 illustrates features of the conceptual model for the Dayton study area.

Geology

Surficial geology of the study area is dominated by Wisconsinan glacial deposits. Glacial till (ground moraine) covers most of the bedrock in upland areas. The clay-rich tills are unstratified and poorly sorted; grain size ranges from clay size to boulders. Some thin gravel deposits can be found interspersed within the till cover.

Glacial deposits cover the bedrock in the buried river valleys. Illinoian glacial deposits may underlie the Wisconsinan deposits in the deepest areas of the buried valleys (Dumouchelle, 1998). The glacial deposits range from fine-grained sand to gravel with some glacial till interspersed within as sheets that sometimes extend across the buried valleys (Dumouchelle, 1998).

The buried-valley floor and walls consist of Late Ordovician interbedded shale and limestone and form the base of the buried-valley aquifer. The limestone beds typically are thin—from less than a centimeter to several centimeters thick, and typically account for about 25 percent of the Ordovician rocks (Eberts and George, 2000). In upland areas, the Brassfield Formation (Early Silurian) overlies the Ordovician rocks

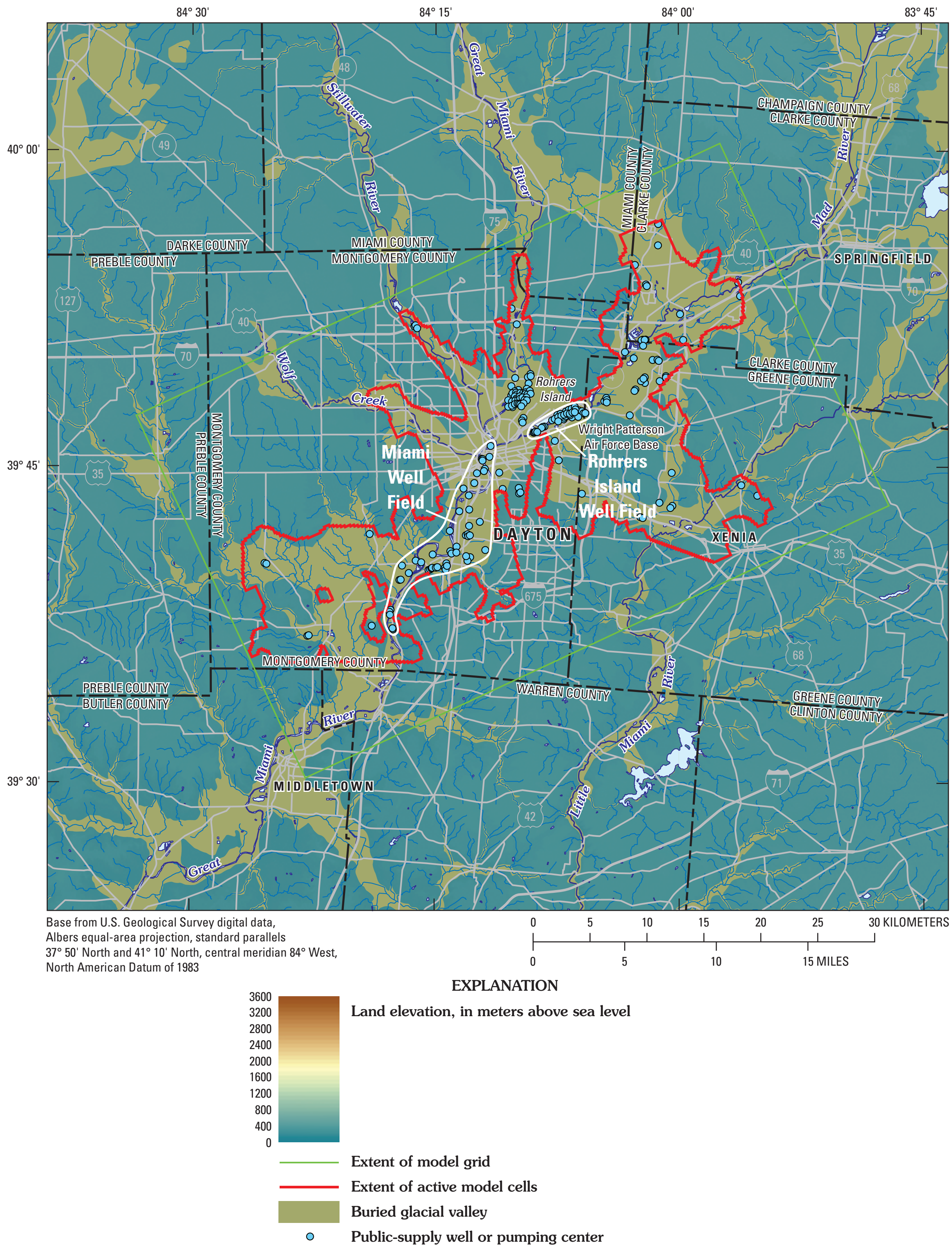

Figure 7.2. Topography, hydrologic features, and locations of public-supply wells, Great Miami River Basin regional study area, Ohio.

(Dumouchelle, 1998). The Brassfield Formation consists of fossiliferous, fine-grained limestone beds that range from massive to irregularly bedded (Sheets and Yost, 1994). Overlying the Brassfield Formation in the study area are thinly bedded Silurian shale and limestone formations.

The study area lies over the axis of the Cincinnati arch, a bedrock high formed between the Appalachian and Illinois Basins (Casey, 1994, 1997). The bedrock at the axis of the arch essentially has zero dip.

Ground-Water Occurrence and Aquifer Properties

The clay-rich tills that cover the upland areas in the study area are generally thin (less than 9 m) and poorly permeable–wells completed in these units yield from 11 to 65 m^3/d (Norris and others, 1948, 1950; Schmidt, 1986, 1991). Vertical hydraulic conductivity of the till ranges from approximately 4 $X10^{-3}$ to 43 m/d (Dumouchelle, 1998). Typically, wells in the upland unconsolidated deposits are completed in sand "stringers." Ground-water flow in the upland unconsolidated deposits is toward small streams or local pumping centers.

The coarse sand and gravels in the glacial deposits are as much as 90 m thick in the center of the valleys and yield as much as 11,500 m^3/d of water to wells (Norris and others, 1950). Hydraulic conductivities range from 860 to $1.1X10^6$ m/d (Dumouchelle, 1998, table 2) based on more than 30 multiple-well aquifer tests. Transmissivities have been reported as much as 65,000 m^2/d but generally range from 280 to 6,500 m^2/d (Dumouchelle, 1998). Where laterally continuous tills are present within the glacial deposits, the outwash deposits are separated into two or more aquifers, and the lower aquifers may be confined or semiconfined by the till layers (Norris and Spieker, 1966). Porosity in the glacial deposits has been estimated from 0.15 to 0.25 (Cunningham and others, 1994; Sheets and others, 1998). In the absence of pumping, ground-water flow in the glacial deposits is lateral and upward toward the major streams in the valleys. Ground-water flow can be reversed and streams can lose water to the aquifer near areas such as Dayton where pumping rates are large.

Generally, the Ordovician bedrock is not considered an aquifer—most wells drilled into these rocks yield less than 6 m^3/day and drawdown is generally very large (Dumouchelle, 1998). Only near the subcrop of the Ordovician units, where weathering has created some secondary permeability in fractures and bedding planes, is the permeability high enough to support localized domestic water supplies (hydraulic conductivity of approximately 430 m/d; Dumouchelle and others, 1993). Sheets and others (1998) showed that excess radiogenic helium found in ^{3}Hydrogen-^{3}Helium samples might indicate some ground-water flow from the Ordovician bedrock through the valley walls and into the glacial deposits (fig. 7.3). Wells in the Brassfield Formation and overlying formations ordinarily yield about 30 to 90 m^3/d, which is generally adequate for domestic water supplies in the uplands (Sheets and Yost, 1994). Springs occur at the base of the Brassfield Formation and at the top of the shale units above the Brassfield (Sheets and Yost, 1994).

Recharge and Discharge

Recharge to the Great Miami River Basin regional study area is primarily through direct precipitation and infiltration, but recharge also occurs from surface-water infiltration and inflow from bedrock (Dumouchelle, 1998). Recharge estimates for the glacial deposits are from water-budget analyses (31.5 cm/yr; Walton and Scudder, 1960), recession-curve analyses (31.5 to 40.0 cm/yr; Dumouchelle and others, 1993) and ground-water flow modeling (8 to 31 cm/yr; Dumouchelle and others, 1993; Dumouchelle, 1998). Recharge estimates for the upland areas (2.5 to 15 cm/yr) are lower than estimates for the glacial deposits because of lower permeability soils and till (Sheets, 1994). Particle-tracking analyses show the majority of recharge to the ground-water system is from direct precipitation and infiltration to the glacial deposits and from the upland areas (Cunningham and others, 1994). Additional sources of recharge are induced infiltration from streams near pumping wells and, to a much lesser extent, inflow from the Ordovician bedrock. Some recharge to the ground-water flow systems occurs at the boundary between the valleys and the bedrock hills, where surface-water runoff flows from the low-permeability tills to much higher permeability glacial deposits (Sheets, 1994; Dumouchelle, 1998). Recharge to the glacial deposits from bedrock is generally considered negligible, relative to the amount of recharge from other sources (Dumouchelle, 1998).

Under natural flow conditions, the major streams and rivers within the glacial deposits are the primary discharge areas for the regional ground-water flow system (Yost, 1995; Dumouchelle, 1998; Sheets, 1994). However, ground water in the study area is heavily pumped for water supply, and this pumping locally alters natural ground-water flow conditions in the glacial deposits. Because most of the large pumping centers are near the major cities and rivers (for example, Dayton), induced infiltration from streams lessens the effect of pumping on regional ground-water flow.

Ground-Water Flow Directions, Depth to Water, and Hydraulic Gradients

Dumouchelle (1998) compiled water-level data from 678 wells to assess ground-water levels and flow directions in the buried-valley aquifer for September 1993. Wells were completed at varying depths within the glacial deposits. Ground-water flow directions in the buried-valley aquifer are generally toward the major rivers and downvalley toward the south/southwest. Horizontal hydraulic-head gradients in the center part of the buried valley and away from the discharge areas (rivers, pumping centers) are approximately 0.0006 m/m; in pumping areas (Miami Well Field, for example) horizon-

Table 7.1. Summary of hydrogeologic and ground-water-quality characteristics for the glacial aquifer system and the Great Miami River Basin regional study area, Ohio.

[m, meters; cm/yr, centimeters per year; m^3/s, cubic meters per second; m^3/d, cubic meters per day; km, kilometers; Kh, horizontal hydraulic conductivity; Kz, vertical hydraulic conductivity; Sy, specific yield; n, porosity; mg/L, milligrams per liter]

Characteristic	Glacial aquifer system	Great Miami River Basin regional study area
	Geography	
Topography	Relief generally less than 300 m. (Randall, 2001).	Relief approximately 200 m.
Climate	Precipitation 91 to 127 cm/yr; evapotranspiration 46 to 58 cm/yr (Randall, 1996).	Precipitation 100 cm/yr; evapotranspiration up to 67 cm/yr (Harstine, 1991; Debrewer and others, 2000).
Surface-water hydrology	Runoff 41 to 76 cm/yr (Randall, 1996); streamflow varies widely with the size of drainage basin. Water-supply reservoirs and former mill ponds are common in upland and valley settings.	Runoff 31 cm/yr (Debrewer and others, 2000); average flow in Great Miami River at Dayton is 96 m^3/s (Shindel, and others, 2002). Some water-supply reservoirs are present in upper reaches.
Land use	Urban, suburban, rural residential, woodlands, farmland.	Urban, suburban, rural residential, woodlands, farmland.
Water use	Potential aquifer yields generally less than 60,500 m^3/d (Kontis and others, 2004).	Pumpage for public supply about 510,000 m^3/d. Aquifer yields range from 300 to as much as 11,000 m^3/d (Dumouchelle, 1998).
	Geology	
Surficial geology	Glacially-derived sand and gravel in valleys that slope away from retreating ice sheets; limited fine-grained deposits; till prevalent in uplands but discontinuous under valley fill (Randall and others, 1988; Randall, 2001).	Mainly glacial alluvial sand and gravel in buried bedrock valleys; till covers uplands and is interspersed in valley fill (Dumouchelle, 1998).
Bedrock geology	Crystalline granitic and metamorphic rocks and sedimentary rocks; limited carbonate rocks (Randall, 2001; Randall and others, 1988).	Uplands underlain by limestone/dolomite; buried valleys underlain by shales and limestones (Dumouchelle, 1998).
	Ground-water hydrology	
Aquifer conditions	Valley-fill aquifers that are generally less than 2.5 km wide and are unconfined; valley fill generally less than 67 m thick; depth to water generally less than 15 m. Streams that cross valley fill from upland areas are commonly sources of recharge; pumping near surface water commonly induces infiltration (Kontis and others, 2004).	A valley-fill aquifer that is generally less than 3 km wide and unconfined; valley fill generally less than 100 m thick; depth to water generally less than 15 m. Areal recharge dominates with some leakage in from bedrock. Induced infiltration from streams where pumping is nearby.

Table 7.1. Summary of hydrogeologic and ground-water-quality characteristics for the glacial aquifer system and the Great Miami River Basin regional study area, Ohio.—Continued

[m, meters; cm/yr, centimeters per year; m^3/s, cubic meters per second; m^3/d, cubic meters per day; km, kilometers; Kh, horizontal hydraulic conductivity; Kz, vertical hydraulic conductivity; Sy, specific yield; n, porosity; mg/L, milligrams per liter]

Characteristic	Glacial aquifer system	Great Miami River Basin regional study area
Hydraulic properties	Valley fill: Kh=6 to 150 m/d; Kh/Kz=10:1 (commonly); n=0.3 to 0.4; Sy=0.2 to 0.3 Till: Kh=0.003 to 3 m/d; Kh/Kz=1; n=0.1 to 0.3; Sy=0.04 to 0.28 Bedrock: Kh=0.003 to 0.3 m/d; Kh/Kz (limited information); n=0.005 to 0.02; Sy=0.0001 to 0.005 (Randall and others, 1988; Bradbury and others, 1991; Melvin and others, 1992; Gburek and others, 1999).	Valley fill: Kh=0.1 to 150 m/d; n=0.15 to 0.25; Kh/Kz estimated at10:1. Till: Kh=0.02 to 3 m/d; n=0.15 to 0.25 Bedrock: K=0.001−1.5 m/d. (Dumouchelle, 1998; Sheets and others, 1998).
Ground-water hydrology—Continued		
Ground-water budget	Recharge to valley fill from infiltration of precipitation, 36 to 76 cm/yr. Recharge to valley fill from upland runoff often exceeds recharge from precipitation. (Kontis and others, 2004; Morrissey and others, 1988). Pumpage generally less than 15 percent of water budget; most discharge is to streams (Morrissey, 1983; Tepper and others, 1990; Dickerman and others, 1990; Dickerman and others, 1997; Mullaney and Grady, 1997; Starn and others, 2000; Barlow and Dickerman, 2001; DeSimone and others, 2002.)	Recharge to valley fill from infiltration of precipitation, 8 to 40 cm/yr. Recharge increases along valley walls from upland runoff. Pumpage for public supply greater than 25 percent of water budget; where induced infiltration doesn’t occur, most ground water discharges to streams. (Dumouchelle, 1998).
Ground-water quality		
	Dissolved solids less than 150 mg/L in crystalline-rock terrains and greater than 150 mg/L in sedimentary-rock terrains; pH, 6−8; oxic. Calcium and bicarbonate are the principal ions (Rogers, 1989). Redox conditions not defined regionally.	Calcium and bicarbonate are the principal dissolved ions. Redox conditions are typically oxic in shallow valley fill, suboxic to anoxic in deep valley fill and in bedrock (Rowe and others, 1999; this study).

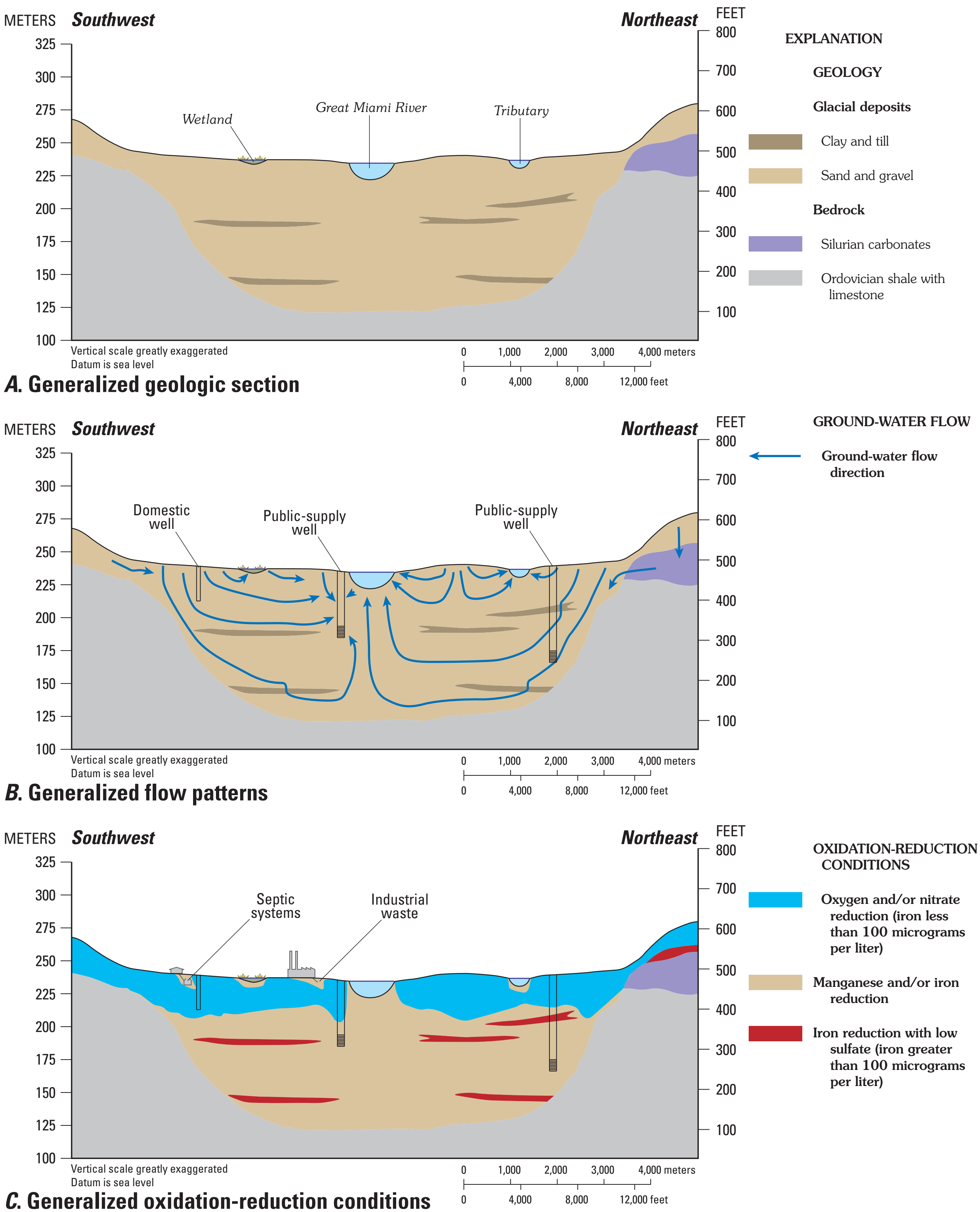

Figure 7.3. (*A*) geology, (*B*) ground-water flow directions, and (*C*) geochemical conditions within the buried valleys around Dayton, Ohio.

tal hydraulic-head gradients are about 0.001 m/m. Near the edge of the buried valley, adjacent to the uplands, horizontal hydraulic-head gradients commonly are greater than 0.0003 m/m.

Water levels for the uplands were not measured by Dumouchelle (1998), but previous work at Wright Patterson Air Force Base indicates the ground-water potentiometric-head surface mimics land-surface topography in the uplands (Schalk, 1992). Ground water flows roughly radial off of upland highs and toward the center of the major river valleys (Dumouchelle, 1998).

Because the ground-water and surface-water systems are generally connected in the study area, depth to ground water often is directly related to land-surface topography. Near the rivers and streams, depth to ground water commonly is just a few feet. Away from the rivers, as elevation increases, depth to ground water generally increases; maximum depth to water away from the river is approximately 15 m, under natural conditions. In areas of heavy pumping, depth to water can be greater than 30 m. The saturated thickness of the buried-valley aquifer also varies in relation to proximity to the axis of the buried valley and in relation to pumping centers. Near the axis of the valley, the maximum depth to bedrock is approximately 90 m; therefore, the maximum buried-valley aquifer saturated thickness is about 80 m. Depth to water in the upland bedrock aquifer is generally between 6 and 9 m (Schalk, 1992). Saturated thickness of the upland bedrock aquifer also varies in relation to topography but generally is about 15 m.

Ground-Water Quality

Ground-water quality in the study area has been divided into two compositional groups—calcium-magnesium-bicarbonate (Ca-Mg-HCO_3) and sodium-chloride (NaCl) type waters (Lloyd and Lyke, 1995; Dumouchelle and others, 1993; Debrewer and others, 2000). Water from the unconsolidated deposits is predominantly Ca-Mg-HCO_3, and water from the bedrock ranges from NaCl to Ca-Mg-HCO_3 type water.

Within the unconsolidated deposits, only subtle variations in major cation concentrations were observed between shallow, intermediate, and deep wells (Dumouchelle and others, 1993). Shallow ground water (less than 18 m below land surface) had higher mean temperatures, specific conductances, and dissolved-solids concentrations than ground water from deep (20 to 70 m below land surface) wells. Median ground-water pH from the unconsolidated deposits was 7.3.

Ground water from the bedrock deposits in the study area varies greatly in its composition. Ordovician shale produces exclusively NaCl-type water with highly variable calcium, magnesium, and bicarbonate concentrations. Median pH of bedrock ground water from Ordovician shale was 7.4. Ground-water quality in the Brassfield limestone, however, is similar to ground-water quality in the glacial deposits consisting of Ca-Mg-HCO_3 type water with a similar pH range of 7.1 to 7.6 (Dumouchelle and others, 1993).

Dissolved oxygen is typically present only in ground water from the shallow glacial deposits (Dumouchelle and others, 1993). Bedrock wells rarely yield water with any measurable dissolved oxygen (greater than 0.1 mg/L). The dissolved oxygen in the shallow aquifers tends to react quickly with organic material in glacial sediments and causes reducing conditions that promote dissolution of iron and manganese oxyhydroxide minerals. Under these low-oxygen and circumneutral pH conditions that are typical of the buried-valley sediments, iron or manganese concentrations can increase in excess of several milligrams per liter (Debrewer and others, 2000). Ground-water concentrations of nitrate greater than a few milligrams per liter are present in shallow, oxygenated parts of the buried-valley aquifers (Debrewer and others, 2000; fig. 7.3), which also are characterized by relatively short ground-water residence times (less than 5 to 10 years) (Rowe and others, 1999). Nitrate in the shallow parts of aquifers in the study area can be reduced to nitrogen gas by denitrification processes as ground-water flows from the oxygenated recharge areas to iron-reducing parts of the buried-valley aquifer (Rowe and others, 1999; fig. 7.3).

Ground-Water Flow Simulations

A numerical ground-water flow model was developed for the glacial deposits in the Great Miami River Basin regional study area by Dumouchelle (1998) (fig. 7.4). Revisions to the steady-state model were made during this study. The following is a brief description of the previously developed model including the original modeled area, boundary conditions, aquifer properties, model stresses, modifications to the original model, and rationale for the changes.

The original model was developed using MODFLOW88 (McDonald and Harbaugh, 1988), MODFLOWARC (Orzol and McGrath, 1992), and ARC/INFO (Environmental Systems Research Institute, 1987). For this study, the model input parameters were translated into files compatible with MODFLOW-2000 (Harbaugh and others, 2000; Hill and others, 2000) using MODFLOW-GUI (Shapiro and others, 1997; Winston, 2000), a graphical user interface for MODFLOW-2000 (Argus Interware, 1997).

The original conceptual model was largely based on previous hydrogeologic investigations and analysis of modeling performed at Wright Patterson Air Force Base (Dumouchelle, 1998). The original model simulated ground-water flow only in the buried-valley aquifer because the underlying Ordovician bedrock and bedrock uplands were not considered significant sources of water to the buried-valley aquifer. The buried-valley aquifer was simulated using three model layers of varying thickness, and the till layers within the aquifer were represented implicitly by reducing horizontal and vertical hydraulic conductivity in locations where tills are present (Dumouchelle, 1998).

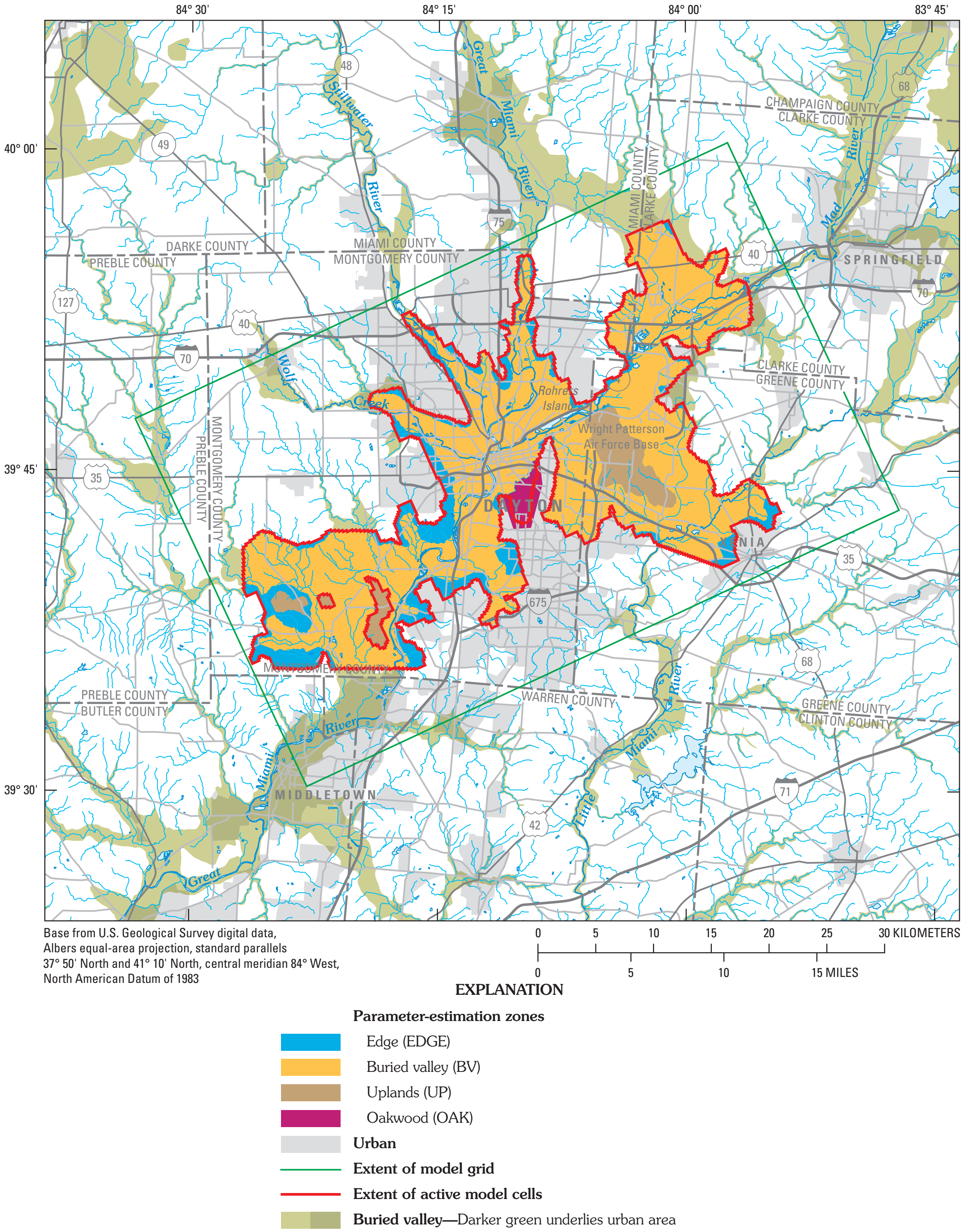

Figure 7.4. Location of ground-water flow model grid and parameter-estimation zones, Great Miami River Basin regional study area, Ohio.

The original model extended laterally to the 800-ft topographic contour around the valley because this contour roughly approximates the knick point between the land surface of the river valleys and the uplands. In some areas, the model was extended into the uplands to account for substantial (greater than 15 m) thickness of sand and gravel deposits (Dumouchelle, 1998). Bedrock islands within the active area of the model (see "Uplands" parameter-estimation zones, fig. 7.4) were not represented in the original model.

The original model simulates steady-state conditions, as observed from water-level and streamflow gain/loss data for late 1993. The revised TANC model simulates these same conditions, as well as those from 1997 through 2001, using water-level data taken from various sources for 1997–2001. The revised model includes two steady-state stress periods with the same model stresses (recharge, river stage and conductance, and so forth) as the original model, except that pumping rates were updated for 1997–2001.

Modeled Area and Spatial Discretization

The original active modeled area covers approximately 620 km^2 and is shown in figure 7.4. The original finite-difference model grid was spatially discretized into 230 rows and 370 columns, corresponding to cells 152.4 m on each side. The model grid was oriented 25 degrees north of east. The original model grid spacing and orientation were not changed by this study.

The ground-water flow system was simulated using three model layers. Layer 1 is simulated as unconfined, with the top representing the ground-water surface and bottom set to the altitude of the uppermost clay layer, an equivalent altitude to the clay layer, or the bedrock surface. Layers 2 and 3 were simulated as confined. The bottom of layer 2 was set to approximately 46 m below the estimated water level or the bedrock altitude. The bedrock valley floor defined the bottom of layer 3 (Dumouchelle, 1998). The vertical connection between model layers was simulated by a vertical leakance parameter.

Boundary Conditions and Model Stresses

No-flow boundaries were used to simulate the Ordovician bedrock-valley walls and floor (Dumouchelle, 1998). The steepness of the valley walls created some numerical instabilities in the original (and revised) model, as narrow or very thin model layers were created during the automatic population of the model grid using MODFLOWARC. Numerical instabilities occurred at the junction between the buried-valley aquifer and uplands or near the four upland "islands" within the modeled area (see "Uplands" parameter estimation zones in fig. 7.4). Similar problems occurred after transferring the model to Argus (Argus Interware, 1997). To decrease numerical instability, the four bedrock islands were simulated using hydraulic conductivities and recharge rates comparable to those given by Sheets and others (1998), who demonstrated that changing the conceptualization of bedrock hills in the Dayton area can improve calibration of numerical ground-water flow models.

Numerical instability in the original model also occurred in the vicinity of the city of Oakwood (fig. 7.4). The Oakwood wells are located in a "hanging valley"—a glacially formed valley that is higher in altitude than the primary buried valley. This area was originally simulated as two layers, with pumping primarily in the second (lower) layer. In the revised model, the Oakwood area was modified to include only one active model layer, and hydraulic conductivity of the layer was assigned as a composite of the original two layers.

At several locations along the edge of the active model layers, specified-head boundaries were used to represent downvalley flow into or out of the modeled area. Specified-head boundaries were typically located in narrow sections of bedrock valleys and far away from the pumping areas, to minimize the effect of boundaries on the simulations (Dumouchelle, 1998). The specified-head boundaries remained unchanged in the revised model.

Stresses on the aquifer system consisted of recharge, rivers, drains, and pumping/recharging wells. The original and revised models are nearly identical in terms of how recharge (except for the bedrock islands), rivers, and drains are simulated. After calibration, recharge in the original model ranged from 15 to 31.0 cm/yr. During the transfer from ARC/INFO to Argus in the revised model, recharge was increased by a small percentage (less than 10 percent) in some model cells near the valley walls to increase numerical stability.

Rivers and larger streams were simulated with the MODFLOW River package (McDonald and Harbaugh, 1988). River stage and river-bottom altitude were obtained by surveying, supplied survey data, U.S. Geological Survey (USGS) topographic maps, or estimates from Geographic Information Systems (GIS) (Dumouchelle, 1998) and were unchanged in the revised model. Values of riverbed hydraulic conductivity for the rivers and streams in the model ranged from 0.09 to 4.3 m/d (Dumouchelle, 1998, p. 20). Drains were used to simulate many streams in the study area, specifically those marked as intermittent on USGS topographic maps. Drains can only remove water from the system, using a drain conductance value (McDonald and Harbaugh, 1988). Drain conductance was not altered from the original model and ranged from 0.0007 to 21.5 m^2/d (most values were 0.003 m^2/d).

In 1993, pumping data were collected for 281 nonresidential pumping wells and 3 recharge wells. As a result of adjustments for model grid size, area, and layering, 187 pumping wells were simulated in the original model at 138 model-cell locations. The total pumping for the original model was approximately 500,000 m^3/d.

Ground-water pumping data for wells in the study area were collected for 1997–2001 from State of Ohio files. Most of these wells were the same as wells simulated in the original model, and well locations, screen lengths, and model grid cells in the revised model were the same as in the original model. Some of the wells in the 1997–2001 data set were previously unused wells, and the locations and screen lengths were taken from either USGS water-use files (on file at USGS Ohio Water Science Center, Columbus) or Ohio Department of Natural Resource-Division of Water well-log files. In a few cases where inadequate information was obtained from the water-use files, location and screen-length information was obtained from the well owner or a State or local agency. Ground-water pumping data for well fields (for example, Rohrer's Island Well Field) was reported as a single value; the pumping rate for an individual well in the well field was apportioned pumping on a weighted basis of flow, relative to the specific capacity reported. To obtain the pumping rate for a particular well, the total volume reported for the well field was multiplied by the specific capacity of the well divided by the sum of specific capacity for all wells in the well field.

Pumping-rate data for all the wells for 1997–2001 were averaged to obtain average pumping rates representative of the time period. In the revised model, 309 pumping centers were simulated (fig. 7.2); the number of pumping nodes simulated in model layers 1, 2, and 3 were 131, 153, and 25, respectively. The total pumping for the second steady-state stress period of the revised model was approximately 500,000 m^3/d, which is the same as the pumping rate used for the first steady-state stress period, although the spatial distribution of pumping was slightly different for the two different stress periods.

Aquifer Hydraulic Properties

Layer 1 in the original model was simulated as an unconfined layer, and horizontal hydraulic-conductivity values in the original calibrated model ranged from 1.6X10^{-3} to 140 m/d (Dumouchelle, 1998). The lowest values were near the edge of the buried valley near the bedrock subcrop, and the highest values were concentrated in the main valleys (Dumouchelle, 1998). The original model uses the Block-Centered Flow (BCF) package in MODFLOW-96 (Harbaugh and McDonald, 1996). The revised model uses the Layer Property Flow (LPF) in MODFLOW-2000 (Harbaugh and others, 2000). The LPF package requires that an unconfined layer be simulated as convertible between unconfined and confined flow conditions. This does not make an effectual change in the internal solutions to the ground-water flow equation, as the top of layer 1 in the revised model was assigned to the land-surface elevation, and water levels in layer 1 did not rise above land surface.

In the original model, layers 2 and 3 were simulated as confined layers using the BCF package. Assigned transmissivities ranged from 0.09 m^2/d to approximately 3,900 m^2/d in layer 2 and from 0.09 m^2/d to 2,800 m^2/d in layer 3. These transmissivity values represent hydraulic-conductivity values ranging from approximately 1.2 X 10^{-3} to 1.6 X 10^{-2} m/d (Dumouchelle, 1998). In the revised model, the LPF package requires that hydraulic conductivity and thickness, not transmissivity, are assigned to grid cells. The original GIS layers of hydraulic conductivity and thickness (tops and bottoms of model layers) were available and were used to assign the appropriate values in the revised model.

A vertical conductance value was used to simulate hydraulic connection between model layers in the original model; the discontinuity of the clay layers did not allow for direct simulation of confining units. In areas where clay layers were present, vertical hydraulic conductivity was set to 3.5 X 10^{-5} m/d (Dumouchelle, 1998). In areas where no clay layer was present, vertical conductance was calculated using one-tenth of the horizontal hydraulic conductivity, weighted for the thickness of grid cells above and below. In the revised model, the LPF package allowed use of a vertical hydraulic-conductivity value, and vertical hydraulic conductivity was set to one-tenth of the horizontal hydraulic conductivity.

Model Calibration

Calibration of the revised model consisted of several steps including evaluation and reevaluation of input values of hydraulic parameters (as previously discussed), comparisons of original model to revised model output heads and flows, sensitivity analysis, and parameter estimation. With each step of the calibration process, hypotheses about the hydrologic system were tested. Some of these hypotheses tested were local scale (well field) and some were regional scale (buried valley), with the goal of calibrating the model to represent average conditions for 1997–2001.

Comparison of Models

Upon reconstruction of the original model, including modifying the isolated bedrock hills and changing the flow package to accommodate MODFLOW-2000, the model was recalibrated to 1993 conditions. The model calibration to 1993 conditions was evaluated by comparing model-computed and measured hydraulic heads for each layer and comparing model-computed ground-water discharge to gain-loss data collected from selected river reaches. The root-mean-squared error (RMSE) and the mean-absolute difference (MAD), which are statistical measures of the variance and bias, respectively, were used to evaluate the model calibration. The RMSE was calculated as:

$$RMSE = \sqrt{\frac{\sum_{i=1}^{N} (h_{meas} - h_{sim})i^2}{N}}$$

where h_{meas} is the measured hydraulic head, h_{sim} is the model-computed (simulated) hydraulic head, $(h_{meas} - h_{sim})$ is the head residual, and N is the number of wells used in the computation.

The MAD was calculated using:

$$MAD = \frac{\sum_{i=1}^{N} abs(h_{meas} - h_{sim})i}{N}$$

where *abs* indicates the absolute value of the expression. The heads calculated by the ground-water flow model were interpolated to their relative position within the grid cell.

Table 7.2 shows the total number of measured hydraulic-head values changed in all three layers during reconstruction of the original model. In an attempt to isolate the numerical instability in the original model, slight changes were made to the bottom of layer 1, especially in the uplands, which changed the layer in which some head observations and pumping wells were located. In a few locations, observations and wells originally in layer 2 were reassigned to layer 1. In one case, a layer 3 well was reassigned to layer 2. Four pumping wells in the city of Oakwood were changed from layer 2 to layer 1.

The RMSE and MAD values for the revised model are higher for layer 1, nearly the same for layer 2, and somewhat lower for layer 3 than for the original model, and the overall (average) RMSE and MAD are slightly higher for the revised model than for the original model. Differences between model-computed and measured hydraulic heads in the vicinity of the Oakwood pumping wells affected the MAD and RMSE for layer 1 and the entire model. In the original model, the Oakwood measured hydraulic heads were in layer 2, but the measured hydraulic heads were in layer 1 of the revised model; inclusion of these points in layer 1 of the revised model produces a higher RMSE and MAD than the original model. There also was a large difference between model-computed and measured hydraulic heads for the Oakwood well field in the original model, which Dumouchelle (1998) attributed to (1) error in spatial distribution of pumping, (2) large grid size (measured hydraulic heads and pumping well within same cell), (3) measured hydraulic-head location near the pumping wells, or (4) a combination of these and other factors. The same factors likely affected model-computed hydraulic heads in the revised model.

Table 7.3 presents a comparison of model-computed and measured streamflows for the two versions of the model for the first steady-state period. The revised parameter-estimation model-computed streamflow gains and losses were roughly equivalent to those computed by the original model, although the model-computed flow from the original model more closely matched the measured streamflow than the revised model for the Great Miami River from Needmore Road to Railroad and for Little Beaver Creek.

Model Zonation, Sensitivity Analysis, and Parameter Estimation

Because the revised model was constructed with MODFLOW-2000, parameter estimation (Hill and others, 2000) was used to assess and improve performance of the revised model. An additional 34 measured hydraulic-head values were available for 1997–2001, so a second steady-state stress period was added to the model for that period. Some water-level measurements were averaged over the period and some were single measurements. All water-level measurements from both stress periods were weighted according to the range and accuracy of water levels and confidence in measuring average conditions. Some wells were within the immediate influence of drawdown from pumping wells, and water levels from these wells were weighted lower than wells distal from pumping because the distal wells were more representative of regional conditions. Streamflow measurements from 1993 were weighted with a coefficient of variation according to the error in consecutive downstream measurements (based on subjective estimates by the measurement taker; for example, Excellent, 2 percent;

Table 7.2. Comparison of root-mean-squared error and mean-absolute difference of model-computed and measured hydraulic heads between the original ground-water flow model (Dumouchelle, 1998) and the revised parameter-estimation model, Great Miami River basin regional study area, Ohio.

[N, number of observations; RMSE, root-mean-squared error in meters; MAD, mean absolute difference in meters; —, not applicable]

Layer	N (original)	N (modified)	RMSE (original)	RMSE (revised)	MAD (original)	MAD (revised)
1	303	311	2.23	4.88	1.37	2.07
2	259	252	3.08	3.01	1.98	2.10
3	17	16	2.69	2.03	2.07	1.46
Total[1]	579	579	—	—	—	—
Weighted* averages	—	—	2.62	3.99	1.66	2.07

*Total RMSE and MAD values are weighted averages.

Good, 5 percent; Fair, 8 percent; Poor, more than 8 percent [Rantz and others (1982)].

The modeled area was divided into four zones representing different hydrologic conditions to evaluate each area's sensitivity to model input parameters. The four parameter-estimation zones in the revised model are (1) the main buried valley (BV); (2) the edge of the main buried valley, next to the upland areas (EDGE); (3) the newly modified upland areas (UP); and (4) the Oakwood area (OAK) (fig. 7.4). Model sensitivity to the input parameters of horizontal hydraulic conductivity, recharge, river conductance, and vertical hydraulic conductivity are compared among the four parameter-estimation zones.

After reconstructing the original model in MODFLOW-2000 as described in the "Initial Calibration" section, the model input parameters of horizontal and vertical hydraulic conductivity, recharge, and riverbed conductance were parameterized within the MODFLOW-2000 framework. MODFLOW-2000 multiplier arrays were created for each parameter in each layer for each of the two stress periods. The original model output was recreated by using a multiplier value of 1.0 for horizontal hydraulic conductivity in layers 1, 2, and 3 and for recharge in all parameter-estimation zones. A multiplier value of 0.1 was used for vertical hydraulic conductivity to represent vertical hydraulic conductivity as one-tenth of horizontal hydraulic conductivity. The riverbed hydraulic conductivity was also parameterized using the same conductance values as in the original model. A list of parameters used for each parameter-estimation zone is in table 7.4.

Table 7.3. Comparison of model-computed and measured streamflow gains or losses between the original ground-water flow model (Dumouchelle, 1998) and the revised parameter-estimation model, Great Miami River Basin regional study area, Ohio.

[units in m^3/d, cubic meters per day; WPAFB, Wright Patterson Air Force Base]

Stream reach	Streamflow gain or loss (1993)		
	Measured	Model computed (original)	Model computed (revised)
Great Miami River, Taylorsville Dam to Needmore Road	+0.481	–0.139	–0.033
Great Miami River, Needmore Road to Railroad	–0.983	–0.949	+0.0006
Mad River, Huffman Dam to Harshman Road	–2.24	–1.55	–1.45
Little Miami River from Dayton-Xenia Road to Narrows Park	+0.439	+0.425	–0.405
Hebble Creek, WPAFB to Mad River	–0.008	–0.006	–0.006
Little Beaver Creek, Research Road to Factory Road	–0.062	+0.042	+0.031

Table 7.4. Parameters used in construction of the revised parameter-estimation ground-water flow model, Great Miami River Basin regional study area, Ohio.

Parameter	Layer	Main buried valley	Edge of main buried valley	Uplands	Oakwood
Horizontal hydraulic conductivity	1	HK1_BV	HK1_EDGE	HK1_UP	HK1_OAK
	2	HK2	HK2	HK2	HK2
	3	HK3	HK3	HK3	HK3
Vertical hydraulic conductivity	1	VK1	VK1	VK1	VK1
	2	VK2	VK2	VK2	VK2
	3	VK3	VK3	VK3	VK3
Recharge*		RCH_BV	RCH_EDGE	RCH_UP	RCH_OAK
River conductance		RIV_BV		RIV_UP	

*Recharge also was parametized separately for stress period 2 (1997–2001); parameters have a suffix of 2 (for example, RCH_UP2).

The revised model was run in sensitivity mode to determine the relative effect a parameter might have on the weighted residuals (difference between model-computed and measured heads and flows). Figure 7.5 shows the composite scaled sensitivities of all parameters for the two steady-state stress periods. Composite scaled sensitivities are used to evaluate whether the available observations provide enough information for parameter estimation (Hill and others, 2000). The most sensitive parameters in the modified model are shown in red on figure 7.5 and include the horizontal hydraulic conductivity in layer 1 of the main part of the buried valley (HK1_BV), recharge to the buried valley (RCH_BV) and the edges of the buried valley (RCH_EDGE), and the river conductance of that part of the river which overlies the buried valley (RIV_BV). The range of composite scaled sensitivities indicates that parameter estimation should be able to estimate all of the parameters. If the composite scaled sensitivities ranged over an order of magnitude or more, the optimization procedure may have difficulty estimating values for the least sensitive parameters.

All parameters were input to the parameter-estimation mode of MODFLOW-2000 to determine which parameters of the nonlinear regression are highly correlated. Highly correlated parameter pairs are problematic in parameter estimation because of the difficulty in determining unique values for these parameters (Hill and others, 2000). Preliminary results of the nonlinear regression indicate that a few of the values chosen for regression are highly correlated to each other (r^2>0.85). The highly correlated pairs are as follows: HK1_EDGE and RCH_EDGE, and RCH_OAK and HK1_OAK. HK1_EDGE was fixed using an interim value obtained during preliminary model runs; this value was used in the final parameter estimation and was not allowed to vary with the parameter estimation, whereas RCH_EDGE was estimated because it had a somewhat higher composite scaled sensitivity than HK1_EDGE. RCH_OAK and HK1_OAK had relatively low composite scaled sensitivities (fig. 7.5) and were fixed using values obtained during preliminary model runs.

Based on relative sensitivity and correlation of parameters, six parameters (HK1_BV, HK2, RCH_BV, RCH_UP, RCH_EDGE, and RIV_BV) were chosen for final parameter estimation, where the least-squares objective function (derived from the sum of squared, weighted residuals between model-computed and measured values) is minimized using

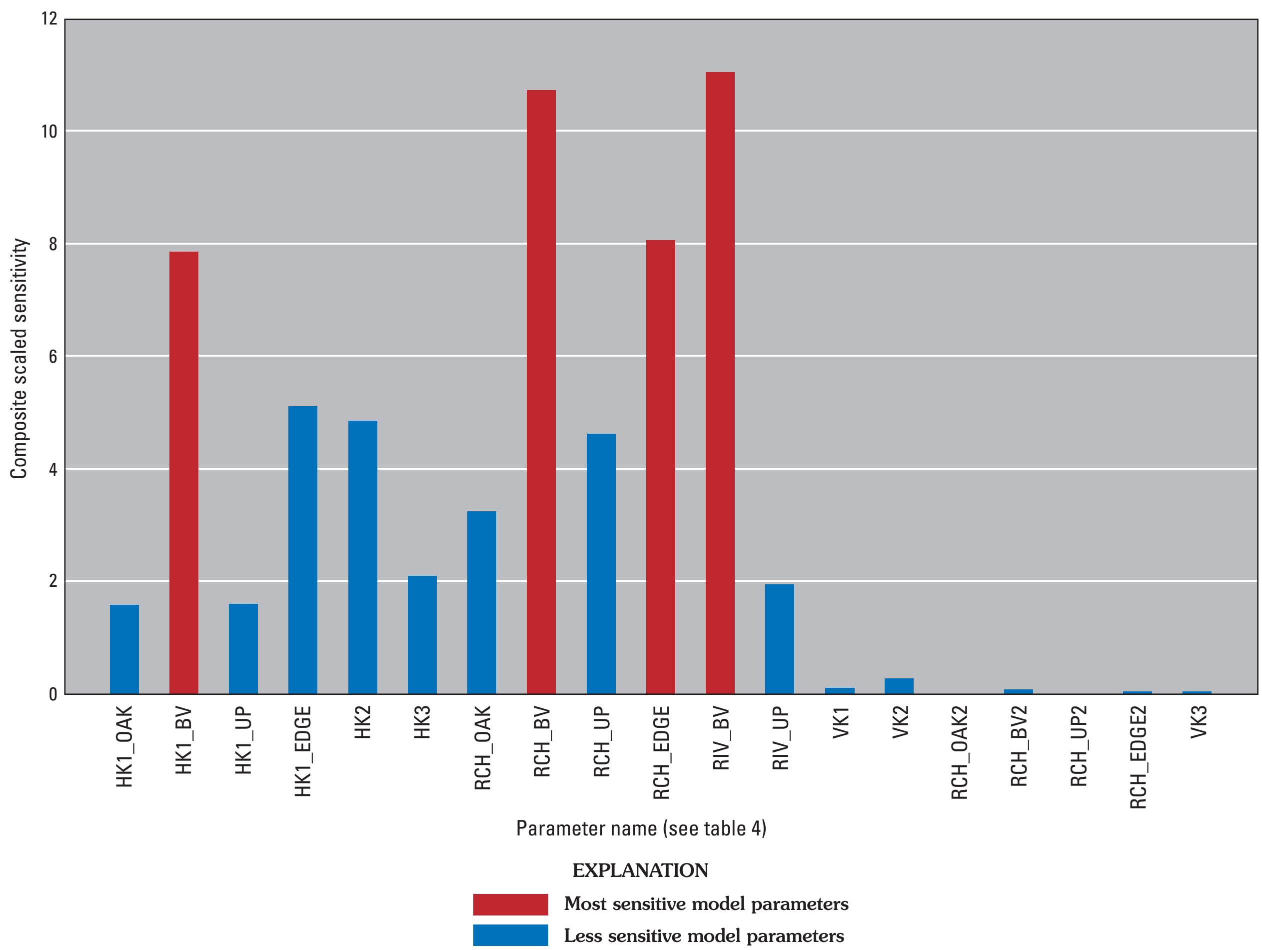

Figure 7.5. Composite scaled sensitivity of ground-water flow model parameters, Great Miami River Basin regional study area, Ohio.

the modified Gauss-Newton method (Hill, 1998). During each parameter-estimation model simulation, each parameter value is modified by MODFLOW-2000 by using their sensitivities, and new parameter values are input to the model to calculate a new objective function. No prior information was used in the nonlinear regression. Parameter estimation was run with a low primary convergence criterion (0.01) and a low convergence criterion (0.001).

Figure 7.6 shows graphs comparing model-computed and measured hydraulic heads for both stress periods, by layer, for the results of the parameter-estimation simulations. The 1:1 line shows where the results would plot if the model-computed heads exactly matched the measured heads for the stress periods. The correlation coefficient between model-computed and measured hydraulic heads is 0.865, which indicates a reasonable model fit. Figure 7.7 presents the weighted residuals compared to the weighted simulated equivalents for both head and streamflow observations and shows that the two variables are independent and that the weighted residuals are scattered evenly about 0.0 (Hill, 1998).

The calculated error variance and standard error for the final calibrated model are 152.6 and 12.4, respectively. A standard deviation of 0.22 m (variance = 0.15 m) was used to calculate weights for the majority of head observations, so the fitted standard error is 2.7 m and represents the overall fit achieved for these hydraulic heads. A calculation of the RMSE and MAD for each layer for stress period 1 (1993) and stress period 2 (1997–2001) is shown in table 7.5. When the results of stress period 1 are compared to the original and revised model (from table 7.2), a slight decrease in the overall MAD and RMSE can be seen, indicating the parameter-estimation resulted in a slightly improved model with respect to hydraulic heads. A coefficient of variation of 0.10 (10 percent) was used to calculate the weights associated with streamflow measurements (1993), so the fitted coefficient of variation of 1.24 (124 percent) represents the overall fit with streamflow measurements. Table 7.6 shows the model-computed and measured streamflow from the calibrated parameter-estimation model. The coefficient of variation and direct comparisons with measurements indicate streamflow is not well represented by this model, and the representation of streamflow is not as good as in the original Dayton-area model (Dumouchelle, 1998). Low weights were applied to streamflow data in the parameter-estimation simulations, which indicates confidence in these measurements is low, and likely resulted in the poorer representation of streamflow in the revised model.

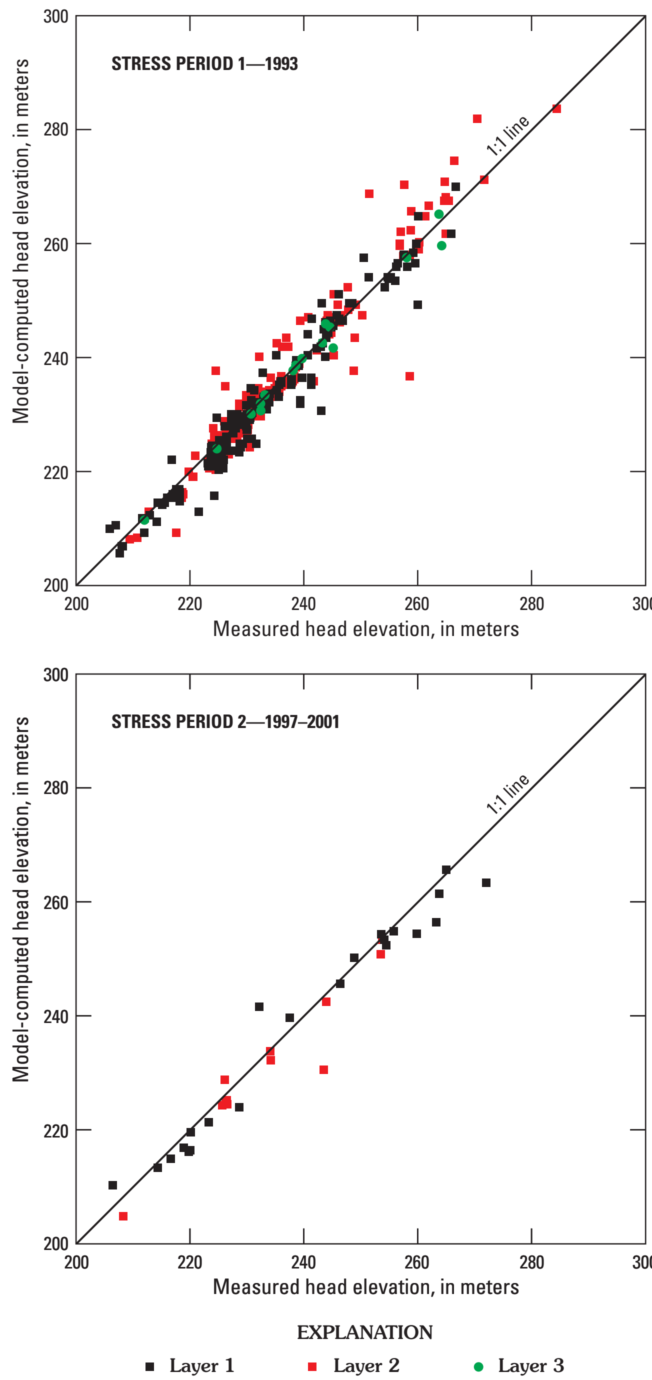

Figure 7.6. Relation between model-computed and measured hydraulic heads, Great Miami River Basin regional study area, Ohio.

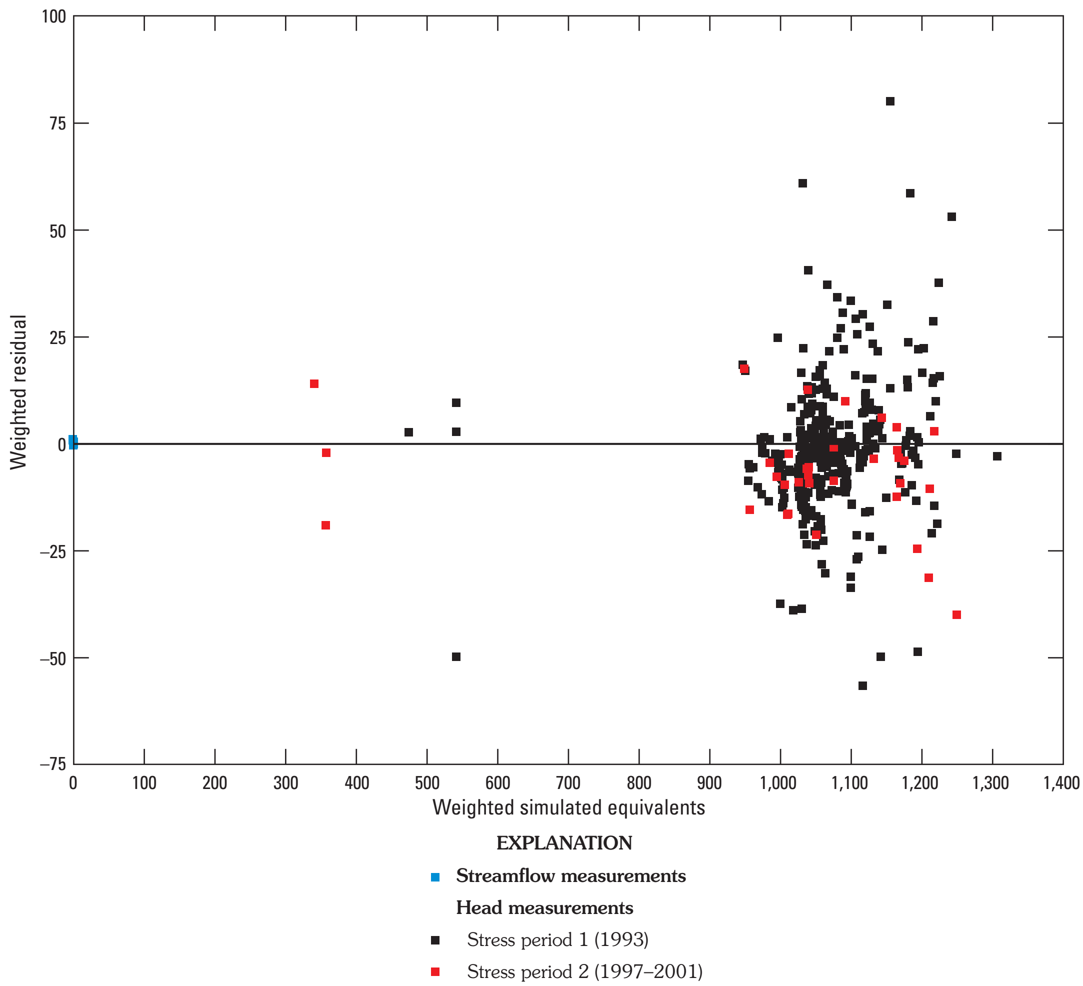

Figure 7.7. Weighted residuals plotted against weighted simulated equivalents, Great Miami River Basin regional study area, Ohio.

Table 7.5. Root-mean-squared error and mean-absolute difference between model-computed and measured hydraulic heads in the revised parameter-estimation ground-water flow model, Great Miami River Basin regional study area, Ohio.

[N, number of observations; RMSE, root-mean-squared error in meters; MAD, mean absolute difference in meters; SP1—Stress period 1 (1993 measurements); SP2—Stress period 2 (1997–2001 measurements); —, not applicable]

Layer	N (SP1)	N (SP2)	RMSE (SP1)	RMSE (SP2)	MAD (SP1)	MAD SP2)
1	311	22	3.04	3.90	1.63	2.96
2	252	12	2.58	4.18	1.73	2.62
3	16	—	1.67	—	1.14	—
Total	579	34	—	—	—	—
Weighted* averages	—	—	2.80	4.00	1.66	2.84

*Total RMSE and MAD values are weighted averages.

Table 7.6. Comparison of model-computed and measured streamflow gains or losses between the original ground-water flow model (Dumouchelle, 1998) and the revised parameter-estimation model, Great Miami River Basin regional study area, Ohio.

[units in m^3/d, cubic meters per day]

Stream reach	Streamflow gain or loss (1993)		
	Measured	Simulated (original)	Simulated (parameter estimation)
Great Miami River, Taylorsville Dam to Needmore Road	+0.481	–0.139	–0.191
Great Miami River, Needmore Road to Railroad	–0.983	–0.949	–0.752
Mad River, Huffman Dam to Harshman Road	–2.24	–1.55	–1.31
Little Miami River from Dayton-Xenia Road to Narrows Park	+0.439	+0.425	–0.489
Hebble Creek to Mad River	–0.008	–0.006	–0.011
Little Beaver Creek, Research Road to Factory Road	–0.062	+0.042	+0.040

The modified Beale measure is used to test linearity of a model, especially if the model's linear confidence and prediction intervals are to be used. The modified Beale's measure of this model (using the F-statistic of 2.1689) is 0.146, indicating the model is somewhat nonlinear with respect to the parameters. If Beale's measure is greater than 0.46, the model is nonlinear; if Beale's is less than 0.041, the model is effectively linear. These results are particularly applicable if this model will be used in predictive simulations where linear confidence intervals on predictions are to be used.

Model-Computed Water Budget

The water budget simulated by the calibrated parameter-estimation model (table 7.7) indicates less than one percent error in the steady-state-model water balance. Recharge to the modeled area is from losing stream reaches (38.9 percent of inflow), precipitation (38.8 percent of inflow), and downvalley ground-water underflow (22.3 percent of inflow). Discharge from the modeled area is to pumping wells (53.5 percent of outflow), gaining stream reaches (40.1 percent of outflow), and downvalley ground-water underflow (6.4 percent of outflow). The model-computed water balance is a reasonable approximation of the conceptual-model water balance.

Table 7.7. Water budget computed by the revised parameter estimation model for 1997–2001 average conditions, Great Miami River Basin regional study area, Ohio.

[m^3/d, cubic meters per day]

Water-budget component	Flow (m^3/d)	Percentage of inflow or outflow
Model inflow		
Downvalley flow into the modeled area	224,000	22.3
Rivers	391,000	38.9
Precipitation	390,000	38.8
TOTAL	**1,005,000**	**100**
Model outflow		
Downvalley flow out of the modeled area	64,100	6.4
Wells	538,000	53.5
Rivers	403,000	40.1
TOTAL	**1,005,100**	**100**

Simulation of Areas Contributing Recharge to Public-Supply Wells

The calibrated revised ground-water flow model was used to simulate areas contributing recharge for 60 public-supply wells and traveltimes from recharge areas to wells. The particle-tracking software MODPATH (Pollock, 1994) was used in conjunction with flux output from the revised flow model and a constant assumed effective porosity value of 0.2 to calculate flow paths and traveltimes. An effective porosity value of 0.2 is reasonable for the study area on the basis of previous work (Cunningham and others, 1994). A nested rediscretization method for particle tracking near wells (Spitz, 2001) was used for the analysis of contributing areas to identify flow paths for individual wells. A grid spacing of approximately 50 m was used for the rediscretization. Properties from the calibrated model were used for the finer grid. The model-computed areas contributing recharge represent advective ground-water flow and do not account for mechanical dispersion. Advection-dispersion transport simulations would likely yield larger areas contributing recharge than advective particle-tracking simulations because the effects of dispersion caused by aquifer heterogeneity would be included.

The areas contributing recharge are presented in figure 7.8, and several features of contributing areas are noteworthy.

- Areas contributing recharge can extend great distances from some wells.
- The areas contributing recharge and zone of contribution for wells with higher pumping rates are generally larger than those with lower pumping rates.
- Losing stream reaches coincide with several of the areas contributing recharge and likely affect the sizes of contributing areas.
- Zones of contribution often are similar to the areas contributing recharge, indicating that the primary source of water to those wells is water recharged at the water table.

Median simulated traveltimes from recharge areas to wells ranged from 21 days to 184 years; approximately 73.2 percent of the traveltimes were less than 25 years. Generally, traveltimes are shorter for shallow well screens and longer for wells screened in the lower part of the aquifer. For deeper wells, pathlines are longer and zones of contribution and areas contributing recharge are more distal from the well than those for shallower wells. Zones of contribution and areas contributing recharge mostly originate upgradient and upvalley from the wellhead. In some areas around Dayton, where multiple wells are completed in close proximity to each other, nearly all the water recharged to the valley is withdrawn by production wells.

Model Limitations and Uncertainties

The purpose of the original ground-water flow model for the Great Miami River Basin regional study area was to provide a better understanding of the flow system in the valley-fill deposits. The purpose of the revised model was to aid in determining contributing areas to public-supply wells and ground-water traveltimes. Both models simulate steady-state conditions. Water-level data and computed water budgets indicate the Great Miami River valley-fill aquifer in the study area was generally in steady-state equilibrium for 1997–2001. However, further calibration for transient conditions may be needed to accurately represent temporal changes in the system. Also, there may be areas where much more detailed geologic information may be necessary to define local flow paths and traveltimes. Because of the regional nature of the modeling effort, some generalizations, including how surface-water bodies (streams) were treated in the model, may adversely affect traveltimes, especially from wells near streams.

Computed areas contributing recharge and traveltimes through zones of contribution are based on the calibrated revised model and a constant assumed effective porosity value of 0.2 for all three model layers. In a steady-state model, changes to input porosity values do not change the area contributing recharge to a given well. Changes to input porosity values will change computed traveltimes from recharge to discharge areas in direct proportion to changes of effective porosity because there is an inverse linear relation between ground-water flow velocity and effective porosity and a direct linear relation between the traveltime and effective porosity. For example, a one-percent decrease in porosity will result in a one-percent increase in velocity and a one-percent decrease in particle traveltime. A detailed sensitivity analysis of porosity distributions was beyond the scope of this study, although future work could compare simulated ground-water traveltimes to ground-water ages to more thoroughly evaluate effective porosity values.

The Great Miami River Basin regional ground-water flow model uses justifiable aquifer properties and boundary conditions and provides a reasonable representation of ground-water flow conditions in the study area for 1997–2001 average conditions. The model is suitable for evaluating regional water budgets and ground-water flow paths in the study area for the time period of interest but may not be suitable for long-term predictive simulations. This regional model provides a useful tool to evaluate aquifer vulnerability at a regional scale, to facilitate comparisons of ground-water traveltime between regional aquifer systems, and to guide future detailed investigations in the study area.

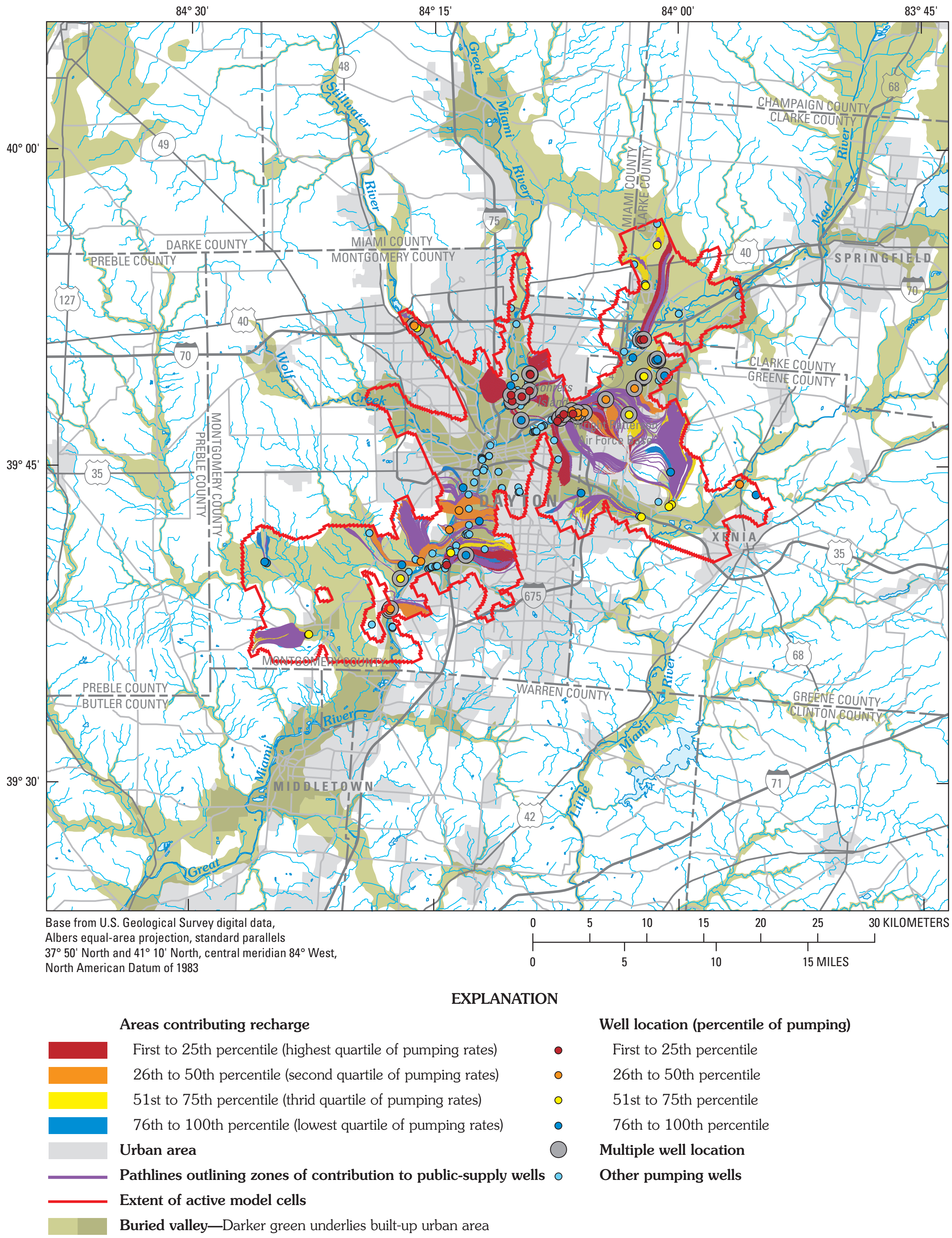

Figure 7.8. Model-computed areas contributing recharge for 60 public-supply wells, Great Miami River Basin regional study area, Ohio.

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