USGS
science for a changing world

Simulation of Ground-Water Flow in Glaciofluvial Aquifers in the Grand Rapids Area, Minnesota

Scientific Investigation Report 2004-5159

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

Simulation of Ground-Water Flow in Glaciofluvial Aquifers in the Grand Rapids Area, Minnesota

By Perry M. Jones

Prepared in cooperation with the Minnesota Department of Health

Scientific Investigation Report 2004-5159

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

U.S. Department of the Interior
Gale A. Norton, Secretary

U.S. Geological Survey
Charles G. Groat, Director

U.S. Geological Survey, Reston, Virginia: 2004

For sale by U.S. Geological Survey, Information Services
Box 25286, Denver Federal Center
Denver, CO 80225

For more information about the USGS and its products:
Telephone: 1-888-ASK-USGS
World Wide Web: http://www.usgs.gov/

Jones, P.M., 2004, Simulatioin of ground-water flow in glaciofluvial aquifers in the Grand Rapids area, Minnesota: U.S. Geological Survey Scientific Investigations Report 2004-5159, 23 p.

Contents

Figures

Tables

Conversion Factors, Abbreviations, and Vertical and Horizontal Datums,

Multiply	By	To obtain
inch (in.)	2.54	centimeter (cm)
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
square mile (mi^2)	2.590	square kilometer
foot per day (ft/day)	0.3048	meter per day
square foot per day (ft^2/day)	0.09290	square meter per day
cubic foot (ft^3)	0.02832	cubic meter
gallon (gal)	0.003785	cubic meter
cubic foot per second (ft^3/s)	0.02832	cubic meter per second
degrees Fahrenheit	˚C = (°F – 32) / 1.8	degrees Celsius

Vertical coordinate information to the North American Vertical Datum of 1988 (NAVD88): horizontal coordinate informatioin is referred to the North American Datum of 1927 (NAVD27)

Water year: The water year is October 1 through September 30 and is named for the calendar year in which it ends.

Simulation of Ground-Water Flow in Glaciofluvial Aquifers in the Grand Rapids Area, Minnesota

By Perry M. Jones

ABSTRACT

A calibrated steady-state, finite-difference, ground-water-flow model was constructed to simulate ground-water flow in three glaciofluvial aquifers, defined in this report as the upper, middle, and lower aquifers, in an area of about 114 mi^2 surrounding the city of Grand Rapids in north-central Minnesota. The calibrated model will be used by Minnesota Department of Health and communities in the Grand Rapids area in the development of wellhead protection plans for their water supplies. The model was calibrated through comparison of simulated ground-water levels to measured static water levels in 351 wells, and comparison of simulated base-flow rates to estimated base-flow rates for reaches of the Mississippi and Prairie Rivers. Model statistics indicate that the model tends to overestimate ground-water levels. The root mean square errors ranged from +12.83 ft in wells completed in the upper aquifer to +19.10 ft in wells completed in the middle aquifer. Mean absolute differences between simulated and measured water levels ranged from +4.43 ft for wells completed in the upper aquifer to +9.25 ft for wells completed in the middle aquifer. Mean algebraic differences ranged from +9.35 ft for wells completed in the upper aquifer to +14.44 ft for wells completed in the middle aquifer, with the positive differences indicating that the simulated water levels were higher than the measured water levels. Percentage errors between simulated and estimated base-flow rates for the three monitored reaches all were less than 10 percent, indicating good agreement. Simulated ground-water levels were most sensitive to changes in general-head boundary conductance, indicating that this characteristic is the predominant model input variable controlling steady-state water-level conditions. Simulated ground-water flow to stream reaches was most sensitive to changes in horizontal hydraulic conductivity, indicating that this characteristic is the predominant model input variable controlling steady-state flow conditions.

INTRODUCTION

About two-thirds of the population of Minnesota uses ground water for their drinking water. About 2,600 community water-supply wells are present in Minnesota, with about one-half of these wells potentially vulnerable to contamination (Minnesota Department of Health, 2003).

Public water supplies in Minnesota are defined as sources of water to at least 15 service connections, or as sources to 26 or more people per day for at least 60 days per year (Minnesota Department of Health, 2003). Public water supplies are divided into two categories: community and non-community water supplies. Community water supplies serve residents year-round, and include municipalities, mobile home parks, and apartment complexes. Non-community water supplies provide water to people in places other than their homes, such as hotels, restaurants, campgrounds, and schools. Non-community water supplies consist of transient and non-transient systems, where transient systems serve the traveling or transient public, and non-transient systems serve at least 25 non-transient individuals (Minnesota Department of Health, 2003).

Communities in Minnesota and throughout the United States are being required to develop and implement wellhead protection plans to safeguard their public water supplies from contamination. Few tools are currently (2004) available to communities to effectively delineate drinking-water-supply management (or protection) areas surrounding their public supply wells.

The Minnesota Department of Health (MDH) is working to assist communities to develop and implement wellhead protection plans through delineation of drinking-water-supply management (or protection) areas surrounding public supply wells. MDH hydrologists are developing tools and techniques for determining capture zones around public supply wells. One tool is a set of regional ground-water flow models in areas where groups of community water supply wells are located. The Grand Rapids area in north-central Minnesota is one of the areas where a group of public wells are located (fig. 1). Eighteen community supply wells and seven non-community, non-transient wells are present in a 45-mi^2 area around Grand Rapids. Most of the water supplied to these wells is obtained from glaciofluvial aquifers. Glaciofluvial aquifers consist of boulders, gravels, sands, and silts deposited by running waters from melting glaciers. In the Grand Rapids area, these aquifers lie between glacial till units.

The U.S. Geological Survey (USGS), in cooperation with the MDH, conducted a 5-year study to simulate ground-water-flow conditions in aquifers in Minnesota. As part of this 5-year study, a 2-year study (2002-2003) was conducted

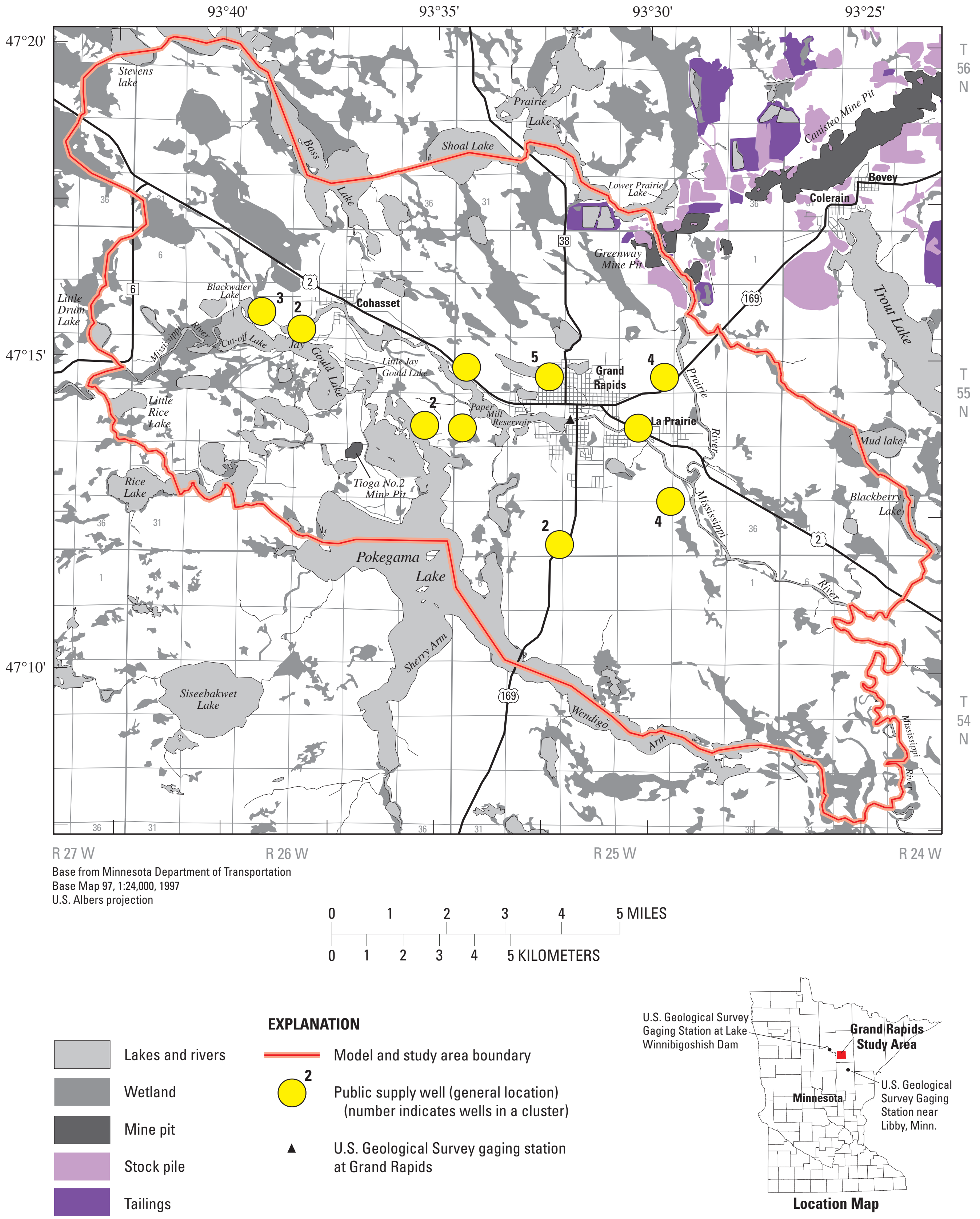


Figure 1. Location of modeled area, mining features, wetlands, public supply wells, gaging stations, lakes, and rivers in the study area.

to simulate ground-water flow conditions in glaciofluvial aquifers in the Grand Rapids area (fig. 1). A steady-state, regional, ground-water-flow model of glaciofluvial aquifers was constructed to simulate ground-water flow in the area. This regional ground-water flow model will be used by MDH and communities in the Grand Rapids area in the development of drinking-water-supply management areas surrounding public supply wells. This report presents the data sets, calibration results, and sensitivity analyses from the ground-water-flow model of the Grand Rapids area. Available geologic and hydrologic data collected from 1975 through 2001 were used for model construction and calibration. Data from scientific literature and other sources also were used.

Physical Setting of the Grand Rapids Area

The study area is about 114 mi^2 surrounding the city of Grand Rapids (population 7,764 in 2000) (U.S. Census Bureau, 2002) in Itasca County, north-central Minnesota (fig. 1). Cohasset and La Prairie are the only other cities in the study area. The watershed of the study area is part of the upper Mississippi River Basin, near the headwaters. Land-surface altitude in the study area ranges from 1,240 ft along the Mississippi River in the southeast portion of the study area to 1,480 ft east of Pokegama Lake and south of Grand Rapids.

Climate in the watershed is continental, with cold winters and hot summers. The mean annual temperature (1961-90) at Grand Rapids, Minnesota is 38.6°F, and the mean annual precipitation is 27.54 in. (Minnesota State Climatologist, 2001). January is the coldest month, and July is the warmest month. February is the driest month and June is the wettest month. Mean January temperature is 3.8°F, and mean July temperature is 67.4°F. Mean February precipitation is 0.54 in., and mean June precipitation is 4.11 in.

Vegetation in the study area consists of a mix of northern hardwood forest and grasslands. Thirty-seven percent of the study area is forest, which consists of 18 percent deciduous forest, 11 percent mixed-wood forest, 5 percent young forest, and 3 percent coniferous forest (Minnesota Department of Natural Resources, 2002a). Grasslands cover 25 percent of the study area (Minnesota Department of Natural Resources, 2002a).

Open water and wetlands cover 29 percent of the study area (Minnesota Department of Natural Resources, 2002a). Farmsteads, cultivated lands, and other rural development cover 3 percent of the study area, whereas 4 percent of the land cover consists of urban and industrial areas, mainly in the cities of Grand Rapids and Cohasset (Minnesota Department of Natural Resources, 2002a).

Two percent of the study area consists of gravel pits, abandoned iron-ore open mines, and lands affected by mining activities (Minnesota Department of Natural Resources, 2002a). Local topography and hydrology south of Lower Prairie Lake and west of Pokegema Lake have been affected by previous iron-ore mining activities at the Tioga No. 2 and Greenway Mines, and gravel mines (fig. 1). Abandoned open mine pits, tailings, stockpiles, and tailings and settling ponds are present in these areas (fig. 1) (Minnesota Department of Natural Resources, 2002a). No active iron-ore mining currently (2004) occurs in the study area. Tailings and stockpiles are as high as 150 ft above the land surface and extend over an area of several square miles. Because no active iron ore mining is present in the study area, the heights and extents of these piles are relatively static. Water levels in iron-ore mine pit lakes are relatively stable.

Surface drainage and ground-water flow through the study area is generally to the south-southeast (Oakes, 1970). Surface drainage north of the Mississippi River flows south to the river. The Mississippi River enters the study area from the west, flows east-southeast through a series of lakes (Cut-off, Blackwater, and Jay Gould Lakes) and the Paper Mill Reservoir, through the city of Grand Rapids, and southeast out of the study area (fig. 1). The Prairie River drains Lower Prairie Lake, located north of Grand Rapids, and flows south into the Mississippi River. Surface drainage south of the Mississippi River flows either north to the Mississippi River or south to Pokegama Lake (fig. 1). Pokegama Lake drains to the northwest through Little Jay Gould and Jay Gould Lakes to the Mississippi River.

Hydrogeology

Glaciofluvial and bedrock aquifers are present in the study area (table 1). Three glaciofluvial aquifers are defined for the purposes of this report as the upper, middle, and lower aquifers. These aquifers are composed of glaciofluvial sediments that are separated or overlain by glacial tills or discontinuous deposits of glaciofluvial sediments. The primary bedrock aquifer in the study area is the Biwabik Iron Formation, which is composed of chert, shale, and iron minerals (table 1). The geologic units and water-bearing characteristics in the study area are presented in table 1.

Precambrian-age metavolcanic, sedimentary, and igneous bedrock, and Cretaceous-age sedimentary bedrock underlie glacial sediments in the study area (Morey, 1972). The Ely Greenstone of Archean age underlies glacial sediments in the northwestern part of the study area (fig. 2), and the Giants Range Granite of lower Precambrian age lie to the south-southeast of the Ely Greenstone. A series of Precambrian-age and Cretaceous-age bedrock units lie above the Giants Range Granite in the central and southeastern part of the study area, striking to the east-northeast and dipping generally between 5° and 15° to the south-southeast (Oakes, 1970). The Pokegama Quartzite of Precambrian age overlies the Giants Range Granite in the central part of the study area (fig. 2). The Biwabik Iron Formation, composed of chert, shale, and iron minerals, overlies the Pokegama Quartzite and is overlain and bounded to the south by the Virginia Formation, consisting of argillites, siltstones, and graywackes (Morey, 1972) (fig. 2). The Biwabik Iron Formation is mined for iron ore and the main

Table 1. Geologic and hydrogeologic units in the Grand Rapids area, Minnesota and their water-bearing characteristics

[gal/min, gallons per minute]

System	Series	Geologic unit	Lithology	Water-bearing characteristics	Hydrogeologic unit in model	Equivalent layer in the digital ground-water flow model
Quaternary	Pleistocene	Glaciofluvial sediments	fluvial sands, gravels, and boulders	well yields more than 1,000 gal/min in middle and lower aquifers,	upper, middle, and lower aquifers; glaciofluvial sediments separating aquifers	layers 1, 2, 3, 4, 5, and 6
Quaternary	Pleistocene	Glacial tills	clay, sands, silts, pebbles, cobbles, and boulders	limited permeability	glacial tills	layers 1, 2, 3, 4, 5, and 6, basal (no-flow) boundary below lower aquifer
Cretaceous	Upper	Coleraine Formation	sandstones, iron-formation, and shales	limited, occurs in isolated patches above the Virginia Formation	not represented in model	basal (no-flow) boundary
Precambrian	Middle	Virginia Formation	argillite, siltstones, and graywackes	limited permeability, well yields as much as 10 gal/mn	not represented in model	basal (no-flow) boundary
Precambrian	Middle	Biwabik Iron-Formation	ferruginous chert, shale, and iron minerals, also called taconite.	well yields as much as 500 gal/min	not represented in model	basal (no-flow) boundary
Precambrian	Middle	Pokegama Quartzite	quartzite, dense, hard, conglomeratic at base	not used as a source of water	not represented in model	basal (no-flow) boundary
Precambrian	Middle	Giants Range Granite	granite intrusive	well yields are low (less than 10 gal/min), only used for domestic water supplies where glacial drift thicknesses are small	not represented in model	basal (no-flow) boundary
Precambrian	Middle	Ely Greenstone	greenstone, green schist	not used as a source of water	not represented in model	basal (no-flow) boundary

bedrock aquifer in the area. Isolated patches of sandstones, iron-formation, and shales of the Cretaceous-age Coleraine Formation overlie the Virginia Formation in the southeastern part of the study area. Bedrock valleys have been identified in the study area (Bruce A. Bloomgren, Minnesota Geological Survey, oral commun., April 17, 2001), but the extent and depth of these valleys is poorly defined by the small number and distribution of wells and boreholes.

Glacial sediments cover the entire study area, with thicknesses ranging from less than 50 ft in the central and northwestern part of the study area to greater than 400 ft in the southeastern part of the study area (Oakes, 1970; Winter, 1973a; Jirsa, and others, 2002). Three major morainal tills and associated glaciofluvial sediments exist, which were formed during the Wisconsin ice advances from the north and west of the study area (Winter, 1971). All three of the tills are discontinuous in the study area, with the extents and thicknesses varying over short distances, forming areas of hydrologic confinement. The stratigraphically lowest till, the basal till, is a dark-greenish and brownish-gray till that ranges from clayey to sandy in texture, and is calcareous (Winter, 1971). The middle boulder till ranges widely in color from gray to yellow, and consists of sands and silts, with abundant cobbles and boulders (Winter, 1971). This till tends to be thickest in the southeastern part of the study area, and greater than 100 ft thick. The surficial till is brown in color; sandy, silty, and calcareous; and is generally less than 30 ft thick in the study area. Sand, silt, and clay deposits reworked and redeposited by Glacial Lake Aitkin overlie surficial till and glaciofluvial sediments southeast and northwest of Grand Rapids (Winter, 1973b).

Glaciofluvial sediments of the three glaciofluvial aquifers lie stratigraphically above the surficial till, between surficial and boulder tills, and commonly lie between the boulder and basal till or bedrock (Winter, 1973a). These deposits consist largely of sands, gravels, and boulders. Glaciofluvial sediments above the surficial and between the surficial and boulder tills, the upper and middle aquifers, respectively, are the most continuous outwash deposits in the study area, commonly greater than 50 ft thick and sometimes greater than 100 ft thick in portions of buried valleys (Winter, 1973a). These sediments consist of fine-grained sands throughout much of the study area. However, outwash deposits can be highly transmissive where coarse-grained sands, gravels, and boulders were deposited in buried valleys and other locations where the bedrock surface was low. The glaciofluvial sediments below the boulder till, the lower aquifer, are fairly continuous in most of the study area, but are absent in the northwestern part of the study area. The glaciofluvial sediments below the boulder till are poorly sorted and are generally less than 50 ft thick, but are greater than 100 ft thick at some locations in buried bedrock valleys (Winter, 1973a). Some glaciofluvial sediments are in direct contact with glaciofluvial sediments of earlier or later glacial flow events.

Glaciofluvial aquifers and the Biwabik Iron Formation are the main sources of ground-water in the Grand Rapids area. Well yields from glaciofluvial aquifers can be more than 1,000 gal/min where glaciofluvial sediment thickness and the horizontal extent are large (table 1). These high well yields are found in the middle or lower aquifers. The Biwabik Iron Formation ranges in thickness between 350 to 500 ft (Oakes, 1970), producing water mainly within fractures and bedding planes. Wells opened to the Biwabik Iron Formation can produce yields as high as 500 gal/min (table 1). These wells generally are located near the subcrop of the formation in the central portion of the study area (fig. 2). The Virginia Formation and Giants Range Granite are used as a source of water for some domestic water supplies. However, well yields typically are low, generally less than 10 gal/min (table 1).

Ground-water withdrawals mainly are from glaciofluvial aquifers through municipal, small industrial, and domestic wells. In 2001, the city of Grand Rapids withdrew 342.4 million gallons of water from four wells completed in glaciofluvial sands and gravels, 256.6 million gallons of water from a well completed in the Biwabik Iron Formation, and 15.9 million gallons of water from Pokegama Lake (fig. 1) (Minnesota Department of Natural Resources, 2002b). The city of La Prairie (population 605 in 2000) (U.S. Census Bureau, 2002) purchases water from Grand Rapids. In 2001, the city of Cohasset (population 2,481 in 2000) (U.S. Census Bureau, 2002) withdrew 22.5 million gallons of water from two wells completed in buried glaciofluvial sands and gravels to serve 755 people (Minnesota Department of Natural Resources, 2002b). The remaining population of Cohasset withdrew water from private wells (Michael Stejskal, City of Cohasset, Minnesota, oral commun., March 4, 2004). In 2001, Minnesota Power withdrew 357.4 million gallons of water from three wells completed in glaciofluvial sands and gravels for generation of electricity from their coal-fire power plant in Cohasset (Minnesota Department of Natural Resources, 2002b). Within the study area, 14 other public supply wells and 2 irrigation wells annually pumped about 39.6 million gallons of water from wells completed in glaciofluvial aquifers in 2001 (Minnesota Department of Natural Resources, 2002b). Domestic wells in the study area extract water mostly from glaciofluvial aquifers, with a few households using the aquifer in the Biwabik Iron Formation for a source of water. Only withdrawals from wells completed in glaciofluvial aquifers were simulated in the model described in this report.

SIMULATION OF GROUND-WATER FLOW

Ground-water flow in the Grand Rapids area was simulated using a calibrated, steady-state, finite-difference, ground-water-flow model. The model was calibrated using water levels in wells and estimated base flows in reaches of the Mississippi and Prairie Rivers. Once the steady-state simulation was calibrated, a series of sensitivity analyses were run with the model to determine the effect of changes in hydraulic

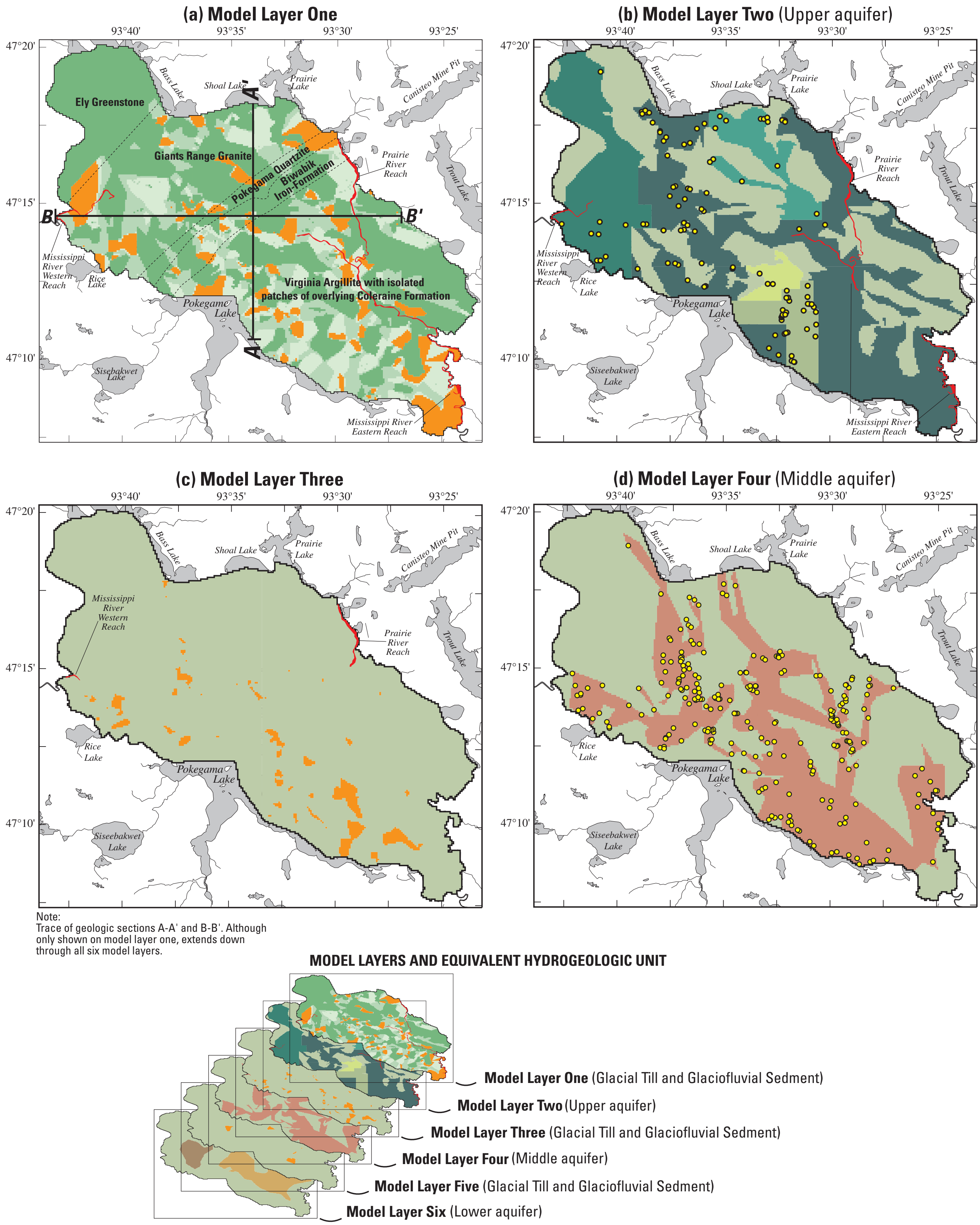
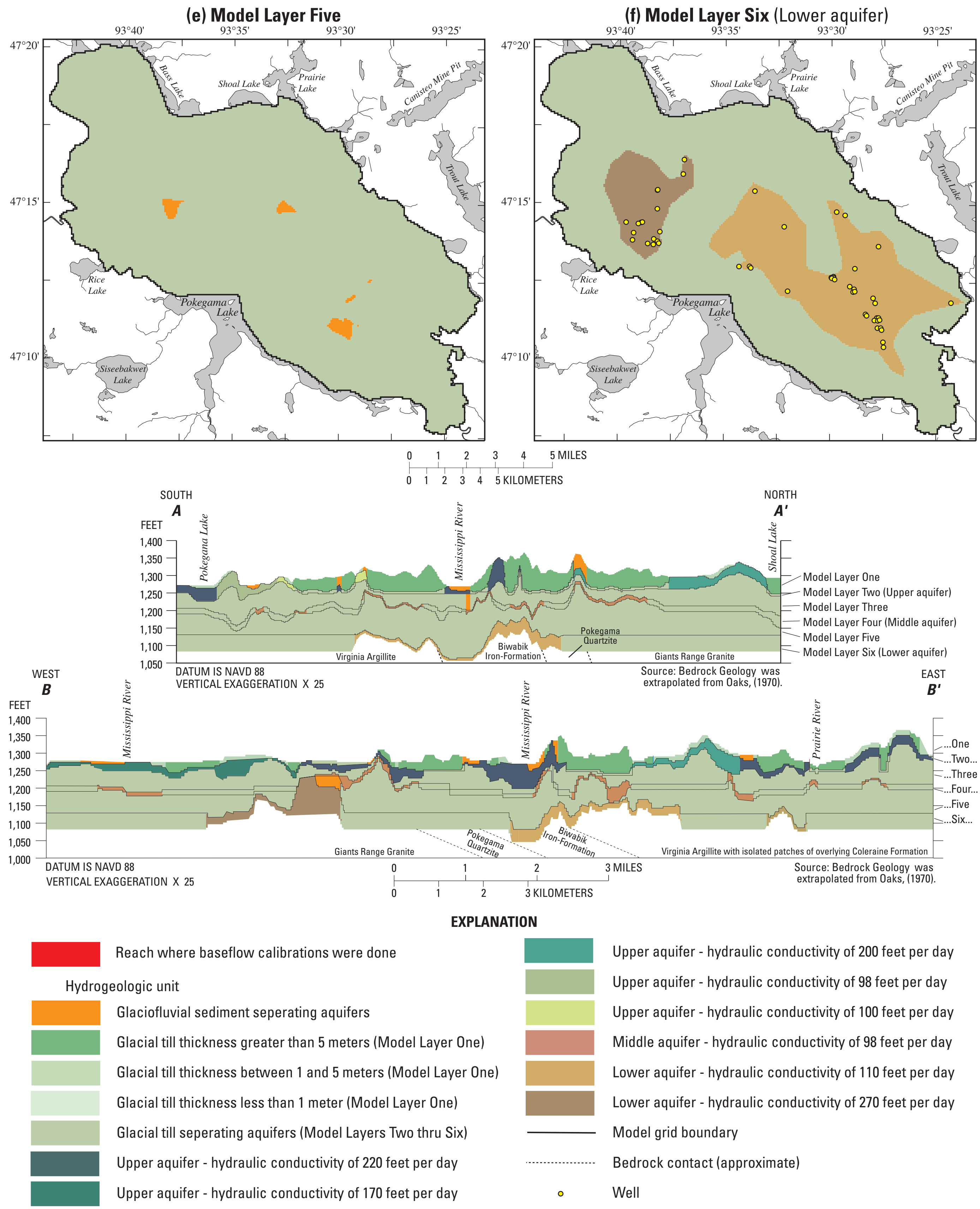


Figure 2. Geology, geologic sections, locations of wells, and reaches of Mississippi and Prairie Rivers where base flow calibrations Model Layer five, and (f) Model Layer six—Lower aquifer



were done in (a) Model Layer one, (b) Model Layer two—Upper aquifer, (c) Model Layer three, (d) Model Layer four—Middle aquifer, (e) conductivity, areal recharge, and river-bed conductance on simulated water levels and streamflows in the model.

Ground-Water-Flow Model of the Grand Rapids Study Area

A three-dimensional, finite-difference, ground-water-flow model was developed, incorporating an area of about 114 mi^2 surrounding the city of Grand Rapids (fig. 1). The USGS MODular ground-water-FLOW model, MODFLOW, was used to simulate ground-water flow conditions in the Grand Rapids area. MODFLOW ground-water-flow model that simulates steady-state and transient ground-water flow in multiple aquifers (Harbaugh and McDonald, 1996) and simultaneously solves a series of mathematical equations that describe saturated ground-water flow where Darcy's Law applies. In MODFLOW, ground-water flow can be simulated in unconfined aquifers, confined aquifers, and confining units. A variety of hydrologic features and processes, such as rivers, streams, wells, evapotranspiration, and recharge from precipitation, also can be simulated. The steady-state representation of the study area was done using the BASIC, BCF, General-Head, River, Well, Recharge, and Preconditioned Conjugate Gradient Solver (PCG2) packages of MODFLOW. The MODFLOW simulation was developed, run, and analyzed using the Groundwater Modeling System (GMS) graphic-user interface (U.S. Department of Defense, 1998). Hydrologic and geologic data available in GIS coverages were imported into the model using GMS.

Simulation of ground-water flow conditions in the study area was done using the following four-step approach: (1) compile existing hydrologic and geologic data; (2) discretize the compiled data; (3) calibrate the model through the comparison of simulated and measured ground-water levels and stream base-flow rates; and (4) perform sensitivity analyses on the calibrated model to assess the effects of hydraulic conductivity, areal recharge, and river-bed conductance on simulated ground-water levels and base-flow rates.

Data Used

A variety of data sources were used to represent various hydrologic features in the model (table 2). The model properties required to represent these features include extent, thicknesses, and boundaries of aquifers; areal recharge rates, water-level altitudes, depths of surface-water bodies that affect ground-water flow, and withdrawal rates from wells completed in aquifers. The types of data sources consisted of geologic descriptions, Geographic Information System (GIS) data, water-level altitudes, and municipal well pumping records.

A three-dimensional representation of the glacial hydrogeologic units in the study area was created by MDH hydrologists based on interpolation of borehole data and geologic maps. Geologic logs of existing municipal, domestic, and monitoring wells were used to develop representations of hydrologic units. Logs for wells were obtained from water-well records in the Minnesota Geological Survey's County Well Index System (table 2). Geologic descriptions on the logs were not detailed enough to distinguish between the surficial, middle boulder, and basal till units, so these units were simulated as a single hydrogeologic unit named glacial tills (table 1). The inverse-distance-weighting method was used to interpolate between well logs, using the 12 nearest neighbors, a power of 6, and outlined aquifer boundaries for barriers. The interpolation was refined by referencing geologic maps and publications by Oakes (1970) and Winter (1971, 1973a, and 1973b), and bedrock-depth maps developed by the Minnesota Department of Natural Resources (MNDNR).

A series of GIS coverages were used to identify and represent hydrologic features and processes in the study area (table 2). Data sets for perennial wetlands, lakes, and rivers were obtained from the National Wetlands Inventory data base (U.S. Fish & Wildlife Service, 1994). Included were natural and man-made lakes, such as tailings ponds, settling ponds, and mine pit lakes. Water-level altitude data for the perennial wetlands, rivers, and lakes used in the model were obtained directly from the National Wetlands Inventory data base and from the MNDNR Lake Level data base (Minnesota Department of Natural Resources, 2002c). A 98-ft (30-m) digital elevation model (DEM) based on the hypsography of 1:24,000-scale USGS quadrangle maps was interpolated to develop a 328-ft (100-m) land-surface altitude grid used for the top of the upper layer of the model.

Annual pumping records for 1997-2001 for public supply wells in the study area were obtained from the MNDNR water-appropriation-permit data base (Minnesota Department of Natural Resources, 2002b). The maximum annual withdrawal rates for 1997-2001 were used in the model to represent steady-state pumping rates from public supply wells.

Model Description and Assumptions

A three-dimensional, numerical model of ground-water flow was constructed based on a conceptual model of the hydrogeology in the study area. The conceptual model was created based on knowledge of the hydrogeologic setting, aquifer characteristics, distributions and amounts of ground-water recharge and discharge, and aquifer boundaries. No storage terms were included in the model. Static water levels measured between 1975 and 1999 and the maximum annual withdrawal rates for 1997-2001 were considered an acceptable estimate of the ground-water system at equilibrium. Withdrawals from individual wells vary annually, but the maximum annual withdrawal rate for 1997-2001 approximates the average withdrawals through the 1975-1999 water-level measurement period (James Walsh, Minnesota Department of Health, oral commun., March 7, 2003).

A "true-layer" approach was used to define layering represented by the model, explicitly defining altitudes and aquifer hydraulic properties of cells in each layer based on the three-dimensional geologic model representation. Using this

Table 2. Data sources used in the MODFLOW steady-state simulation of ground-water flow in glaciofluvial aquifers in the Grand Rapids area, Minnesota

Hydrologic feature	Types of data	Sources of data	Types of information	Date compiled
Aquifer and model boundaries and geology	GIS data sets (shape-files) and Digital Elevation Model (DEM)	U.S. Geological Survey, Minnesota Department of Health, which used Minnesota Geological Survey's County Well Index record, Minnesota Department of Natural Resources bedrock-depth maps, and U.S. Geological Survey publications	location, geologic logs, geologic description, stratigraphy, land-surface altitude, bedrock-depth maps	February - July 2002
Wetlands	GIS data sets	U.S. Fish & Wildlife Service, National Wetlands Inventory data base (1994)	location, area, water-level altitude	February 2002
Lakes	GIS data sets	U.S. Fish & Wildlife Service, National Wetlands Inventory data base (1994)	location, area, water-level altitude	February 2002
Rivers (polygons)	GIS data sets	U.S. Fish & Wildlife Service, National Wetlands Inventory data base (1994)	location, area, water-level altitude (stage)	February 2002
Rivers (Arc and Node)	GIS data sets	U.S. Fish & Wildlife Service, National Wetlands Inventory data base (1994)	location, area, water-level altitude (stage)	February 2002
Lake	water-level altitude	U.S. Fish & Wildlife Service, National Wetlands Inventory data base (1994), Minnesota Department of Natural Resources, Division of Waters, Lake Level data base (2002c)	altitude for lakes	February and March 2002
Areal Recharge	GIS data sets	Developed from U.S. Geological Survey data	location, recharge rate	June 2002
Public supply and other large-capacity wells	well pumping records	Minnesota Department of Natural Resources, Division of Waters, Water-Appropriations-Permit data base (2002b)	annual pumping rates	June 2002

approach, MODFLOW geometry-dependent parameters such as transmissivity and Vcont ("leakance", see McDonald and Harbaugh, 1988) are automatically computed when the model files are saved. The study area was discretized into rectangular finite-difference grid cells within which the properties of the hydrogeologic unit represented were assumed to be uniform and isotropic. Ground-water flow was simulated within the aquifers using a block-centered approach, in which flow is calculated between discretized cells based on head conditions at the central nodes of the cells (McDonald and Harbaugh, 1988). Hydrogeologic properties and stresses were applied assuming that the assigned properties and stresses represent average conditions within that cell. Starting hydraulic head values in the model were extrapolated from a water-level surface obtained from interpolation of measured water levels in wells. Those head values were later reset to head values calculated by early iterations of the model to improve model convergence.

The three-dimensional, finite-difference grid used in the model was evenly spaced, consisting of 239 rows and 263 columns. The dimensions of the grid cells were 328.08 ft (100 m) along rows and columns. The model was vertically divided into six layers, based on the hydrogeologic units (fig. 2, table 1). The layers were represented as either confined or unconfined, with their transmissivities varying with saturated thicknesses under unconfined conditions and being constant under confined conditions. Simulation of flow in the model was dependent on the thicknesses and hydraulic conductivities between adjacent cells and layers. A detailed discussion of the simulation of flow in the model can be found in McDonald and Harbaugh (1988).

The three principal glaciofluvial aquifers used in the study area for ground-water supplies and represented in the model are herein defined as the upper, middle, and lower aquifers. In the model, these aquifers are simulated in layer 2 (upper aquifer), layer 4 (middle aquifer), and layer 6 (lower aquifer) (fig. 2). Glaciofluvial sediments and glacial tills in layers 1, 3, and 5 represent glaciofluvial sediments and glacial tills lying above or between, and hydraulically interconnected with the principal aquifers. Glacial clays and tills were present in parts of each model layer, and were simulated in the model as a single hydrogeologic unit; herein termed glacial tills. Hydraulic conductivities were specified for the glaciofluvial sediments in each layer and a single hydraulic conductivity value was specified for the glacial tills (table 3). Initial hydraulic conductivity values for each of the hydrogeologic units were obtained from MODFLOW simulations in the Coleraine/Bovey area east of the Grand Rapids modeled area (Jones, 2002). In GMS, horizontal and vertical hydraulic conductivity values are entered directly into the program. Model layer elevations are used by GMS with the entered vertical hydraulic conductivity values to get the Vcont ("leakance") values used in MODFLOW (McDonald and Harbaugh, 1988). Precambrian and Cretaceous bedrock, and basal till below the lower aquifer, represented the basal boundary of the model (table 1).

Perennial rivers, streams, lakes, and wetlands were specified as head-dependent boundaries (general-head or river) in layers 1-5 of the model (fig. 3). The model area was bounded to the north by Stevens Lake, Bass Lake, Shoal Lake, the Prairie Lake-Lower Prairie Lake System, and a series of wetlands (figs. 1 and 3). The eastern part of the model is bounded by the Prairie River, Mud Lake, Blackberry Lake, Mississippi River, and a series of wetlands and tributaries (fig. 1). Pokegama Lake, Rice Lake, and a series of small lakes, ponds, and wetlands bound the model to the south. The western boundary of the model consists of wetlands and two lakes, Little Drum and Little Rice Lakes. Cells located inside the model boundary were active, and cells outside of the model boundaries were inactive, where no flow into or out of the cells was simulated.

The bottom of the lowest layer in the model was simulated as a no-flow boundary. The volume of water moving downward across the bottom of the lowest layer in the model was assumed to be small relative to the amount of horizontal flow. With this assumption, ground-water withdrawals from the Biwabik Iron Formation were assumed to have minimal effect on steady-state water levels in the glaciofluvial aquifers.

A total of 14,314 general-head boundary cells, representing lakes and perennial wetlands, were specified in the model. Flow into and out of these cells was calculated by the model based on the difference between the head in the cell and the assigned general head, multiplied by a conductance term (McDonald and Harbaugh, 1988). In each of the general-head boundary cells, the conductance was defined as the hydraulic conductivity of the lake or wetland bed material divided by the vertical thickness of that material, multiplied by the area of the lake or wetland in that cell. Both hydraulic conductivity and vertical thickness of lake and wetland bed material can be highly variable for lakes and wetlands. The selected hydraulic conductivity value of 0.066 ft/day (0.021 m/day) represents a small value typical of the range of hydraulic conductivity values for glacial tills obtained from single-well hydraulic conductivity tests done by Jones (2002) and for hydraulic conductivity values determined for lake-bed material of nearby Shingobee Lake (Kishel and Gerla, 2002). Many of the lakes and wetlands in the study area lie above glacial till, so a hydraulic conductivity value for the glacial till is relevant for the lake-bed material. An assumed value of 3.28 ft (1.0 m) was used to represent an average thickness for the bed material for lakes and wetlands in the area, which was used by Jones (2002) for general-head boundaries. Lake and wetland areas and altitudes from the National Wetland Inventory GIS coverages were used for the areas and head values, respectively, for the general-head boundaries representing the lake and perennial wetland segments.

Streams and rivers were simulated using the river package of MODFLOW (fig. 3). The river package was used to simulate flow between the surface-water features and ground-water systems (McDonald and Harbaugh, 1988). Streams and rivers were segmented into reaches based on underlying geology and altitudes of the streams and rivers for assigning hydrologic variables to model cells. These reaches were segmented into cells (fig. 3). A total of 1,840 river cells were simulated in the model. Flow between the reaches and the ground-water flow systems was calculated for each cell based on the head difference between the river and the aquifer, multiplied by a conductance term (McDonald and Harbaugh, 1988). In the river package, the conductance term is defined as the hydraulic conductivity of the river-bed materials divided by the vertical thickness of the river-bed materials, multiplied by the surficial area of the river bed in the cell (McDonald and Harbaugh, 1988). Hydraulic conductivities of 0.066 ft/day (0.021 m/day) and 13 ft/day (4.0 m/day) were used for the conductance terms for river segments where the underlying geology was till or glaciofluvial sediments, respectively. A vertical thickness of 3.28 ft (1.0 m) was used for the thickness of each river segment. Jones (2002) also used these hydraulic conductivity and vertical thickness values in simulation of ground-water flow surrounding the nearby Canisteo Mine (fig. 1). Surficial areas and altitudes for the river segments were obtained from the National Wetland Inventory GIS coverages. The altitude of the river bed was assumed to be 6.56 ft (2.0 m) below the altitudes for the river segments. Jones (2002) also used this assumption.

A specified-flux boundary was used to represent areal recharge to the upper layer (layer 1) of the model using the recharge package in MODFLOW. Areal recharge was represented as the net difference between precipitation and evapotranspiration losses occurring above the water table. Initial recharge rates were proportioned based on results from stream hydrograph analyses from Jones (2002) and the geology of the upper two layers of the model. The largest areal recharge rate (1.1×10^{-3} ft/day (3.4×10^{-4} m/day)) was assumed to occur where glaciofluvial sediments (sands, gravels and boulders) were present in layer 1 of the model and where glacial tills in model layer 1 were less than 3.28 ft (1.0 m) thick and overlied glaciofluvial sediments in layer 2 (fig. 2; table 3). The smallest recharge rate (4.6×10^{-6} ft/day (1.4×10^{-6} m/day)) occurred

Table 3. Ground-water recharge and hydraulic conductivity values for the best-fit calibration of the MODFLOW steady-state simulation of ground-water flow in glaciofluvial aquifers in the Grand Rapids area, Minnesota

[All values are in feet per day: ---, no value]

Geologic unit, model layer, and hydrogeologic units	Ground-water recharge		Horizontal hydraulic conductivity		Vertical hydraulic conductivity	
	Initial	Final	Initial	Final	Initial	Final
Glacial tills - Layer 1, where till thickness is equal to or greater than 5 meters	4.6×10^{-6}	4.6×10^{-6}	6.6×10^{-2}	2.0×10^{-2}	6.6×10^{-2}	2.0×10^{-3}
Glacial tills - Layer 1, where till thickness is equal to or greater than 1 meter and less than 5 meters	2.3×10^{-4}	2.3×10^{-4}	6.6×10^{-2}	2.0×10^{-2}	6.6×10^{-2}	2.0×10^{-3}
Glacial tills - Layer 1, where till thickness is less than 1 meter and glaciofluvial sediments are present in Layer 2	1.1×10^{-3}	1.1×10^{-3}	6.6×10^{-2}	2.0×10^{-2}	6.6×10^{-2}	2.0×10^{-3}
Glaciofluvial sediments - Layer 1	1.1×10^{-3}	1.1×10^{-3}	13	160	13	16
Glacial tills - Layers 2-6	----	----	6.6×10^{-2}	2.0×10^{-2}	6.6×10^{-2}	2.0×10^{-3}
Glaciofluvial sediments - Layer 2 (upper aquifer)	----	----	13	98, 100, 170, 200, 220	13	9.8, 10, 17, 20, 22
Glaciofluvial sediments - Layer 3	----	----	13	98	13	9.8
Glaciofluvial sediments - Layer 4 (middle aquifer)	----	----	13	98	13	9.8
Glaciofluvial sediments - Layer 5	----	----	13	98	13	9.8
Glaciofluvial sediments - Layer 6 (lower aquifer)	----	----	13	110, 270	13	11, 27

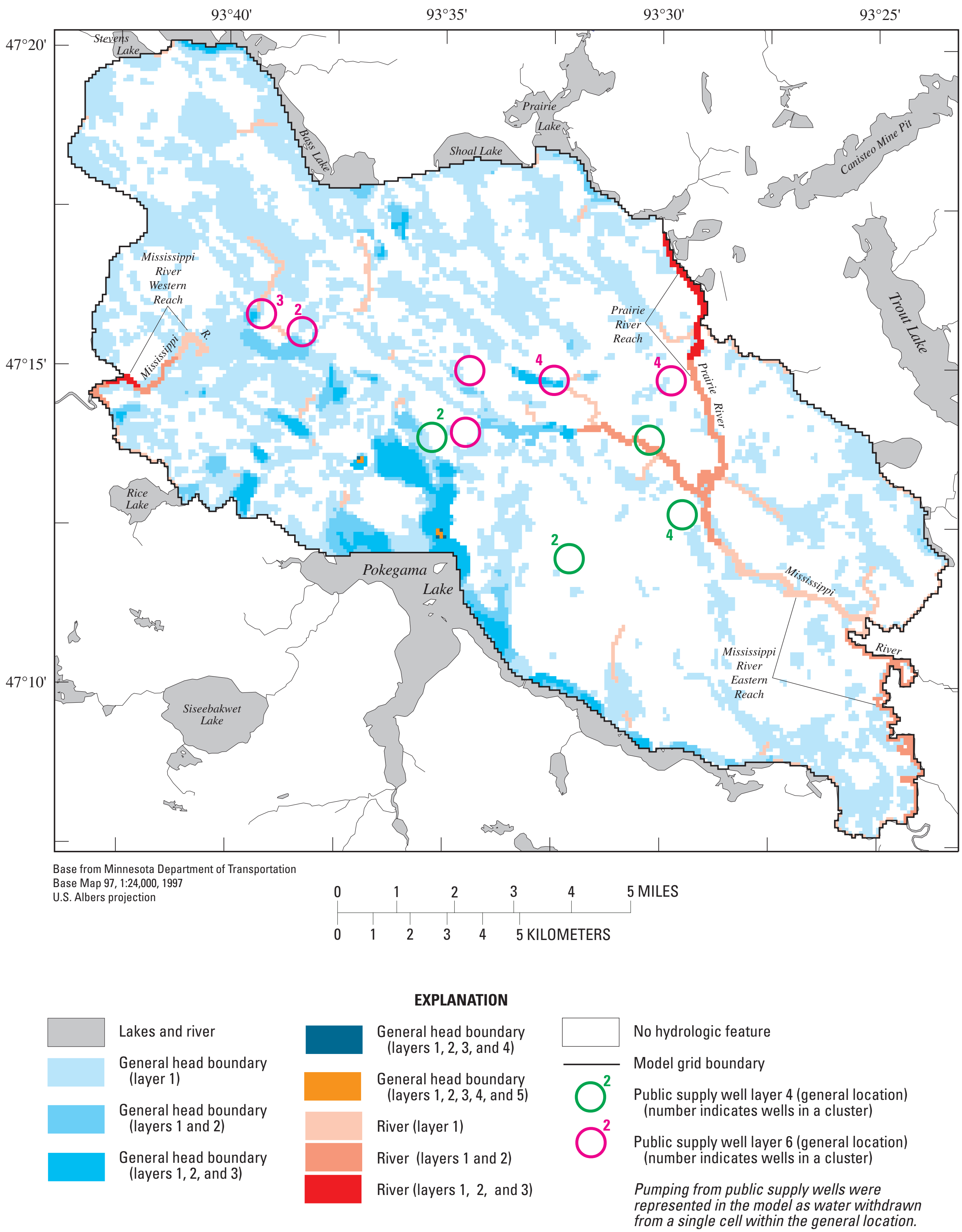


Figure 3. Model segmentation, showing model boundaries, and locations of ground-water withdrawals, for the MODFLOW simulation of ground-water flow in glaciofluvial aquifers in the Grand Rapids area, Minnesota

where glacial tills were present in layer 1 of the model and were at least 16.4 ft (5.0 m) thick (fig. 2; table 3). Recharge rates were 2.3 x 10^{-4} ft/day (7.0 x 10^{-5} m/day) for glacial tills present in model layer 1 that were greater than 3.28 ft (1.0 m) and less than 16.4 ft (5.0 m) thick (fig. 2; table 3).

Ground-water withdrawals from the 23 public supply wells and 2 irrigation wells were simulated in the model using the well package in MODFLOW (fig. 3). These withdrawals included withdrawals from four municipal wells for the city of Grand Rapids and two municipal wells for the city of Cohasset. In the model, water was withdrawn from either the middle or lower aquifers at a specified rate, where the rate was independent of the cell area and head (McDonald and Harbaugh, 1988). Little water is used from the upper aquifer for public supply or irrigation use in the study area, and therefore was not simulated in the model. Pumping rates for the wells were based on the highest annual pumping rate between 1997 and 2001. Annual pumping rates for the municipal wells were obtained from the MNDNR Water Appropriations Permit Records (Minnesota Department of Natural Resources, 2002b). A total of 383,500 ft^3/day of pumping was simulated from the lower aquifer, whereas a total of 3,700 ft^3/day of pumping was simulated from the middle aquifer. Withdrawals from the Biwabik Iron Formation were not simulated in the model because it was assumed that the volume of water moving downward across the bottom of the lowest layer in the model was small relative to the amount of horizontal flow. The actual recharge could be greater than the simulated recharge in localized areas of the model where water is leaving the model through the bottom of the lowest model layer, recharging the Biwabik Iron Formation.

The Preconditioned Conjugate Gradient (PCG2) solver package was used with the modified incomplete Cholesky preconditioning option (relaxation parameter = 1.0) to solve the matrix equations produced by the model (Hill, 1990). Convergence criteria of 0.03 ft (0.01 m) of both head change and residual criterion were used during model calibration. Within the model boundaries, no-flow model cells were not allowed to rewet.

Model Calibration

Model calibration was done by adjusting initial estimates of aquifer properties and boundary conditions within a plausible range of values until simulated water levels and ground-water discharge to streams acceptably match measured water levels and estimated base flows, respectively. Recharge rates and hydraulic conductivities were varied during calibration because values for these variables were uncertain and initial model runs were most sensitive to changes in these variables.

Water levels in 351 wells were used to calibrate the model. The 351 wells were completed in the upper (60 wells), middle (240 wells), and lower (51 wells) aquifers (fig. 2). In general, static water levels for these wells were reported by well drillers following installation, and were obtained from water well records in the Minnesota Geological Survey's County Well Index System. Measurement dates for these static water levels ranged from 1975 to 1999. A water-table contour map published by Oakes (1970) also was used as a reference for assessing the simulated water-table contours.

Base-flow rates into two reaches of the Mississippi River (eastern and western) and a reach of the Prairie River (fig. 2) were used in model calibration. The base-flow rate for the western Mississippi River reach was determined by multiplying the length of the reach by the base flow per river mile value determined by Payne (1995) for the subreach between the USGS gaging station at Lake Winnibigoshish Dam (67 miles upstream from Grand Rapids) and the USGS gaging station at Grand Rapids (fig. 1). In a similar manner, base-flow rates for the eastern Mississippi River reach and the Prairie River reach were determined by multiplying the length of the reaches by the base flow per river mile value determined by Payne (1995) for the subreach between the USGS gaging station at Grand Rapids and the USGS gaging station near Libby, Minnesota (75 miles downstream from Grand Rapids) (fig.1).

Simulation Results

The final calibration values for hydraulic conductivity and ground-water recharge rates are listed in table 3. The match between simulated and measured water levels was improved by (1) decreasing the horizontal hydraulic conductivity for the glacial tills from 6.6 x 10^{-2} to 2.0 x 10^{-2} ft/day, (2) increasing the horizontal hydraulic conductivity for the glaciofluvial sediments from 13 to values ranging from 98 to 270 ft/day, (3) decreasing the vertical hydraulic conductivity for the glacial tills from 6.6 x 10^{-2} to 2.0 x 10^{-3} ft/day, (4) increasing the vertical hydraulic conductivity for glaciofluvial sediments in layers 1, 2, and 6 from 13 to values ranging from 16 to 27 ft/day, and (5) decreasing the vertical hydraulic conductivity for glaciofluvial sediments in layers 2–6 from 13 to values ranging from 9.8 to 11 ft/day. The above changes are within the range of values calculated or measured in existing aquifer tests or reported in previous studies cited in this report.

Plots of measured and best-fit simulated water levels and model statistics indicate that the model tends to overestimate ground-water levels. Plots of measured and best-fit simulated water levels for the wells completed in all glaciofluvial aquifers (fig. 4a), the upper (fig. 4b), the middle (fig. 4c), and the lower (fig. 4d) glaciofluvial aquifers are shown with the 1:1 linear relation between measured and simulated water levels in figure 4. Root mean square error, mean absolute differences, and mean algebraic differences for the simulated and measured water levels indicate that the simulated water levels are biased high compared to the measured water levels (table 4). The root mean square errors ranged from +12.83 ft in wells completed in the upper aquifer to +19.10 ft in wells completed in the middle aquifer. The mean absolute difference between simulated and measured water levels, computed as the sum of the absolute values of the differences divided by the number of

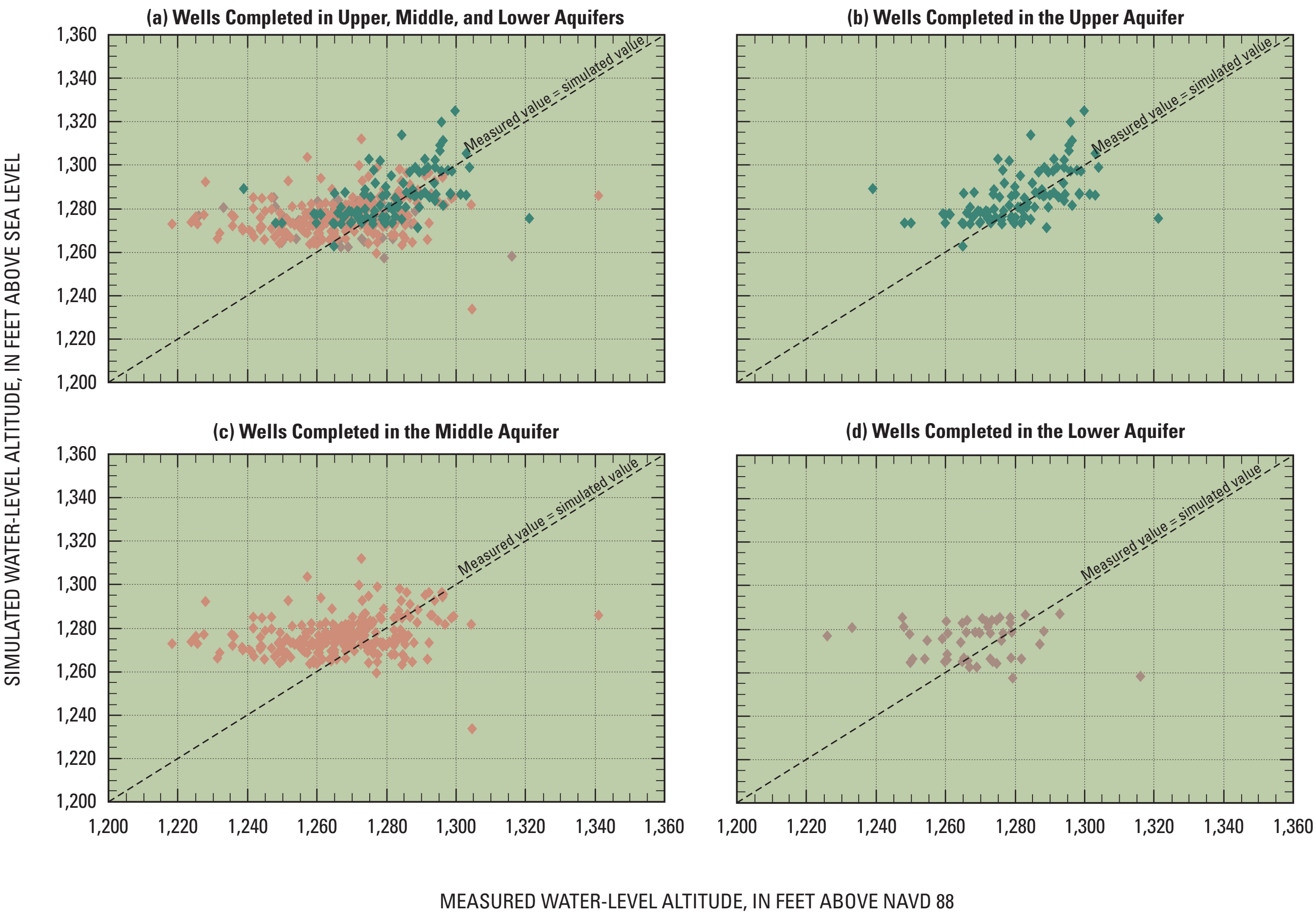


Figure 4. Measured and simulated water-level altitude in wells (a) completed in upper, middle, and lower aquifers, (b) completed in the upper aquifer, (c) completed in the middle aquifer, and (d) completed in the lower aquifer for the calibrated MODFLOW simulation of ground-water flow in glaciofluvial aquifers in the Grand Rapids area, Minnesota

Table 4. Regression statistics for measured and simulated water levels in wells and simulated and estimated base flow to reaches of the Mississippi and Prairie Rivers for the best-fit calibration of the MODFLOW steady-state simulation of ground-water flow in glaciofluvial aquifers in the Grand Rapids area, Minnesota

[ft³/day, cubic feet per day; ft, foot; values in parentheses are percentage error]

Groups of wells	Well measured versus simulated water-level regression statistics (ft)		
	Root mean squared error	Mean absolute difference	Mean algebraic difference
All wells	+17.55	+7.61	+12.99
Wells completed in upper aquifer	+12.83	+4.43	+9.35
Wells completed in middle aquifer	+19.10	+9.25	+14.44
Wells completed in lower aquifer	+18.18	+6.17	+13.35
Monitored reaches	**Base flow to reach (ft³/day)**		
	Simulated	**Estimated**	**Simulated - estimated (ft³/day)**
Eastern reach of Mississippi River	700,000	730,000	- 30,000 (-4.1)
Western reach of Mississippi River	132,000	133,000	- 1,000 (-0.8)
Reach of Prairie River	324,000	297,000	+ 27,000 (+9.1)
Total reach	1,160,000	1,160,000	0 (0.0)

wells, ranged from +4.43 ft for wells completed in the upper aquifer to +9.25 ft in wells completed in the middle aquifer. These mean absolute differences are less than 10 percent of the differences between maximum and minimum ground-water altitudes in each aquifer. The mean algebraic difference between simulated and measured water levels, computed as the algebraic sum of the differences divided by the number of wells, ranged from +9.35 ft in wells completed in the upper aquifer to +14.44 ft in wells completed in the middle aquifer, indicating the positive differences were not balanced by the negative differences.

Differences between simulated and measured water levels at the well locations ranged between –74.25 to +56.10 ft, with a standard deviation of 15.87 ft for all of the wells. The largest positive differences occurred in isolated wells at relatively low land-surface altitudes near the Mississippi River or Pokegama Lake, or at relatively high land-surface altitudes in the central part of study area (fig. 5). The largest negative differences occurred in the middle and lower aquifers in the vicinity of the Grand Rapids well field (fig. 5). One possible explanation for these large differences is extreme weather conditions (droughts or periods of high precipitation) during the time of water-level measurements, which would make the assumption of steady-state conditions inaccurate. Extreme weather conditions are unlikely factors because no correlation was found between these large differences and local precipitation records or the Palmer Hydrological Drought Index for Central Minnesota. Other possible explanations for the large differences include substantial differences between simulated land-surface altitude for the model grid and actual land-surface altitude at the wells and local areas of high or low hydraulic conductivity. Differences between simulated land-surface altitude for the model grid and actual land-surface altitude at some wells were as great as 20 ft. Positive differences at some wells in the central part of the study area may result from ground-water recharge to the Biwabik Iron Formation where the formation subcrops in the modeled area (figs. 2 and 5). Recharge to the formation from the overlying glaciofluvial sediments results from withdrawals from the Biwabik Iron Formation aquifer.

Percentage errors between simulated and estimated base-flow rates for the three monitored reaches all were less than 10 percent (table 4), indicating good agreement. Percentage errors between simulated base-flow rates and estimated base-flow rates were –4.1 for the eastern reach of the Mississippi River, –0.8 for the western reach of the Mississippi River, and +9.1 for the reach of the Prairie River (table 4). The percentage error was 0.0 between total simulated and total estimated base-flow rates for the three reaches (table 4).

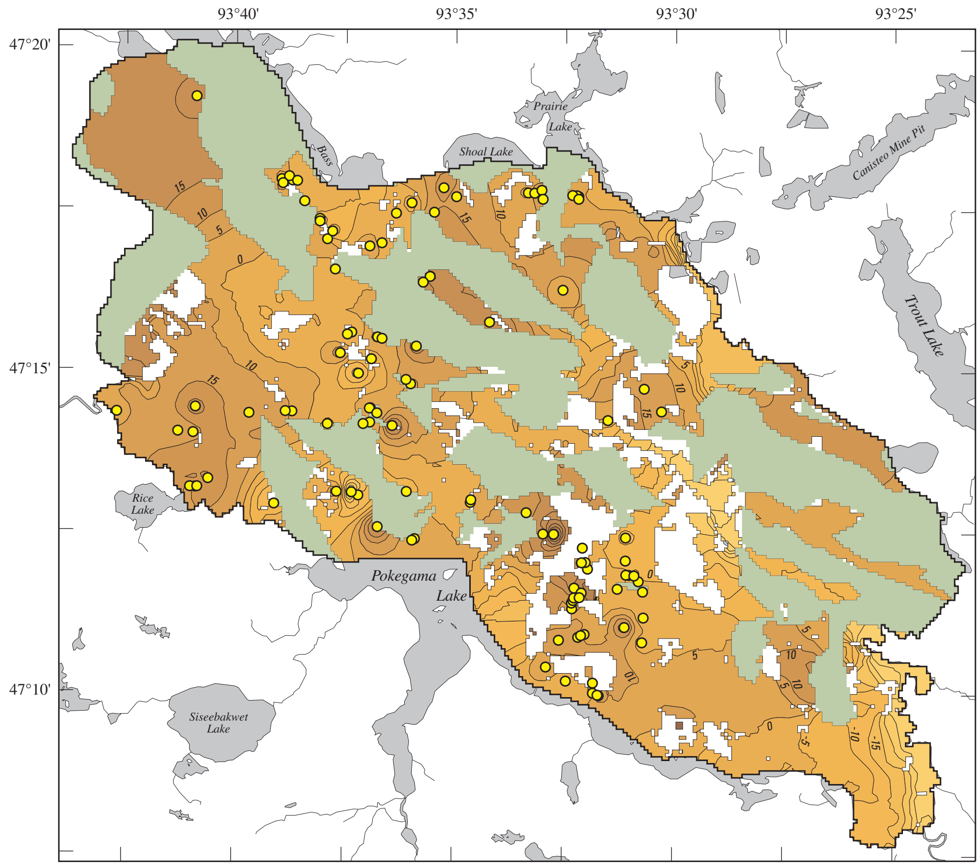
Simulated ground-water flow in the upper aquifer is discontinuous due to the discontinuous areal extent of the aquifer and the presence of dry cells that are mostly found in the higher altitudes of the aquifer (fig. 6). In general, ground water in this aquifer flows toward the Mississippi River and the chain of lakes connected with the river in the central part of the study area (fig. 6a). A local cone of induced drawdown averaging about 8 ft is present beneath the city of Grand Rapids well field, resulting from pumping from the lower aquifer. Because steady-state conditions are assumed for the simulation, this cone of induced drawdown is only an estimate of long-term average conditions. The highest altitudes in the upper aquifer are found in the south-central part of the study area east of Pokegama Lake, an area of low hydraulic conductivity, and the lowest altitudes are in the southeastern part of the study area near the Mississippi River (fig. 6a).

Pumping from the lower aquifer by the city of Grand Rapids public-supply wells lowers the potentiometric surface of both the middle and lower aquifers by about 60 ft (fig. 6b and 6c). A radial cone of depression of greater than 1 mile in diameter exists in both aquifers beneath the well field. The highest altitudes in the middle aquifer are found in the south-central part of the study area east of Pokegama Lake, in the north-central part of the study area near Shoal Lake, and in the eastern part of the study area near the Prairie River (fig. 6b). The lowest altitudes are present beneath the city of Grand Rapids well field and in the southeastern part of the study area (fig. 6b). In the lower aquifer, the highest altitudes are found in the southeastern part of the aquifer and the lowest altitudes are at the city of Grand Rapids well field (fig. 6c).

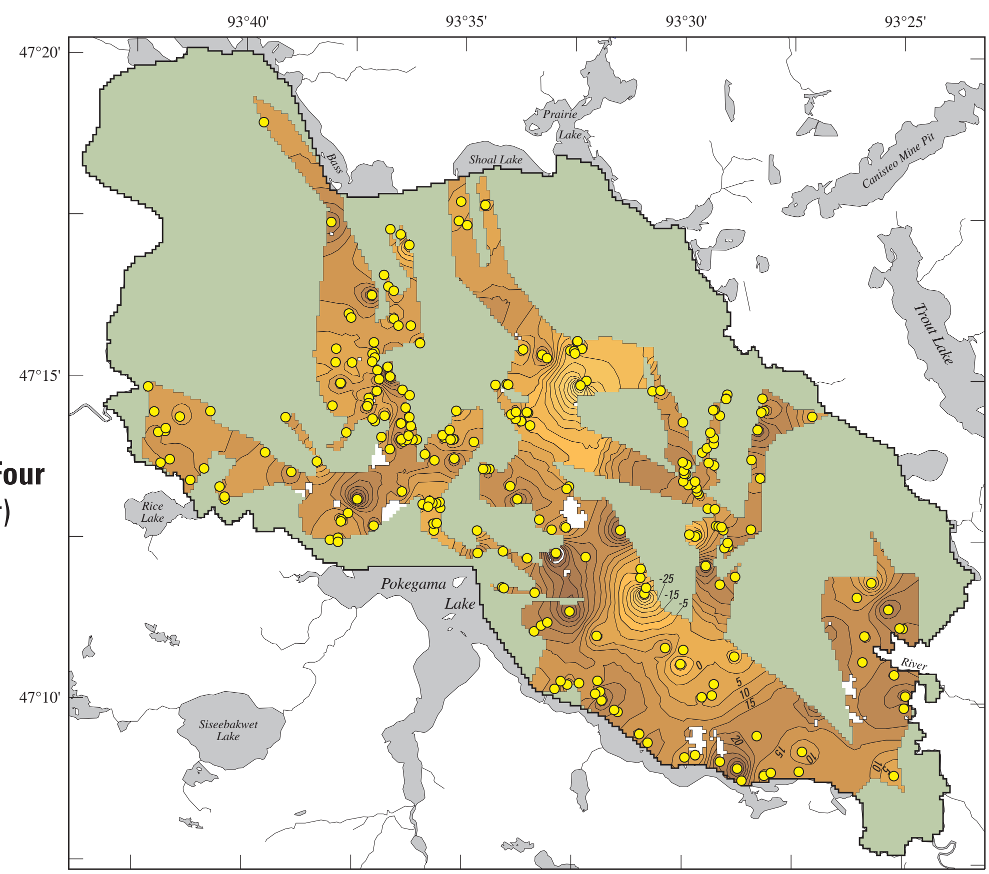
A water budget computed by a steady-state model is an accounting of inflow to, and outflow from, the model. Inflow (sources of water) to the model should equal outflow (discharges). In the calibrated model, total inflow was 51.38 ft^3/s, whereas total outflow was 51.37 ft^3/s. Of the sources of water, inflow from wetlands and lakes accounted for 63.0 percent (32.35 ft^3/s), simulated recharge from the river boundaries accounted for 19.3 percent (9.93 ft^3/s), and simulated areal recharge (precipitation – evaporation) accounted for 17.7 percent of the total recharge to the model (9.10 ft^3/s). Discharge from rivers accounted for 45.7 percent (23.46 ft^3/s) and discharge from wetlands and lakes accounted for 45.6 percent (23.42 ft^3/s) of the total discharges from the model. Pumping wells accounted for the remaining 8.7 percent (4.48 ft^3/s) of the total simulated discharge.

Model Sensitivity to Hydraulic Conductivity, Areal Recharge, River-Bed Conductance, and General-Head-Boundary Conductance

Sensitivity analyses were conducted to determine the effect of changes in areal recharge, hydraulic conductivity, river-bed conductance, and general-head-boundary conductivity on simulated water levels and ground-water discharge to streams. Sensitivity analyses provides an understanding of the importance of these changes on simulated results, and identifies areas for possible model improvement. Thirty-nine simulations were run to assess the sensitivity of the model to changes in areal recharge rates, aquifer hydraulic conductivity, river-bed conductance, and general-head-boundary conductance. Eight of the 39 simulations were conducted varying the recharge rates by factors of 0.6, 0.7, 0.8, 0.9, 1.1, 1.2, 1.3, and 1.4 times the calibrated rates (table 5). Twenty-four simula-



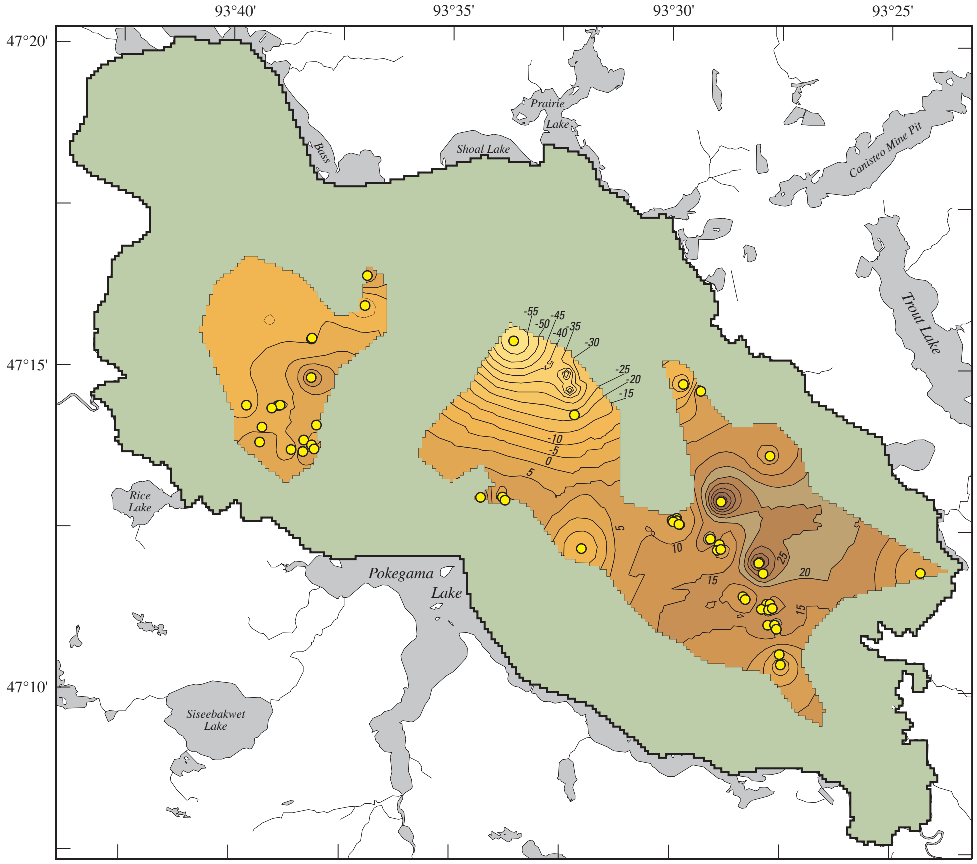
(a) Model Layer Two
(Upper aquifer)

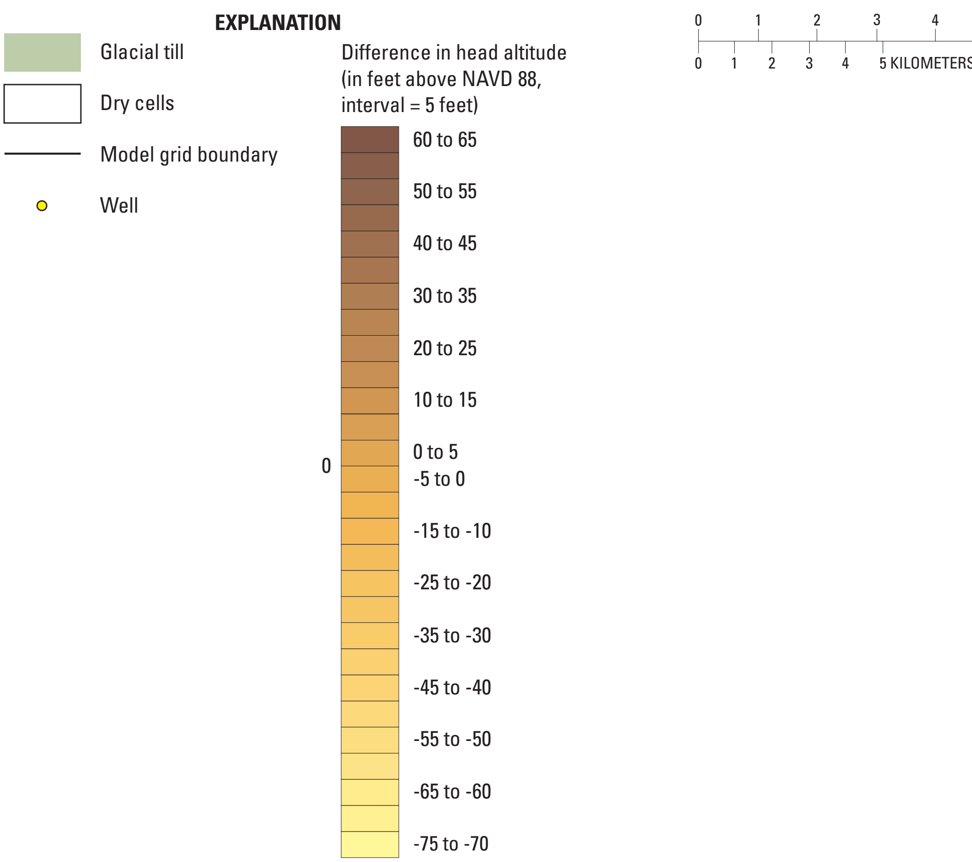


(b) Model Layer Four
(Middle aquifer)

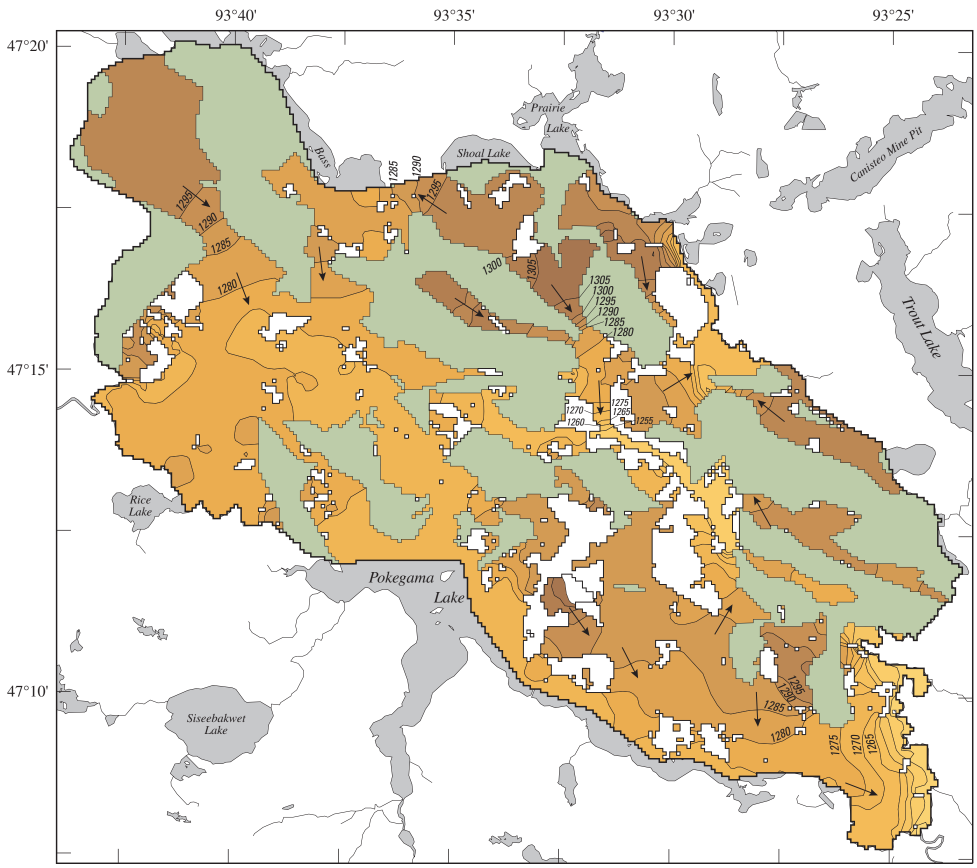
Figure 5. Difference between simulated and measured water-level contours for ground-water flow for the (a) Model Layer two—

(c) Model Layer Six
(Lower Aquifer)

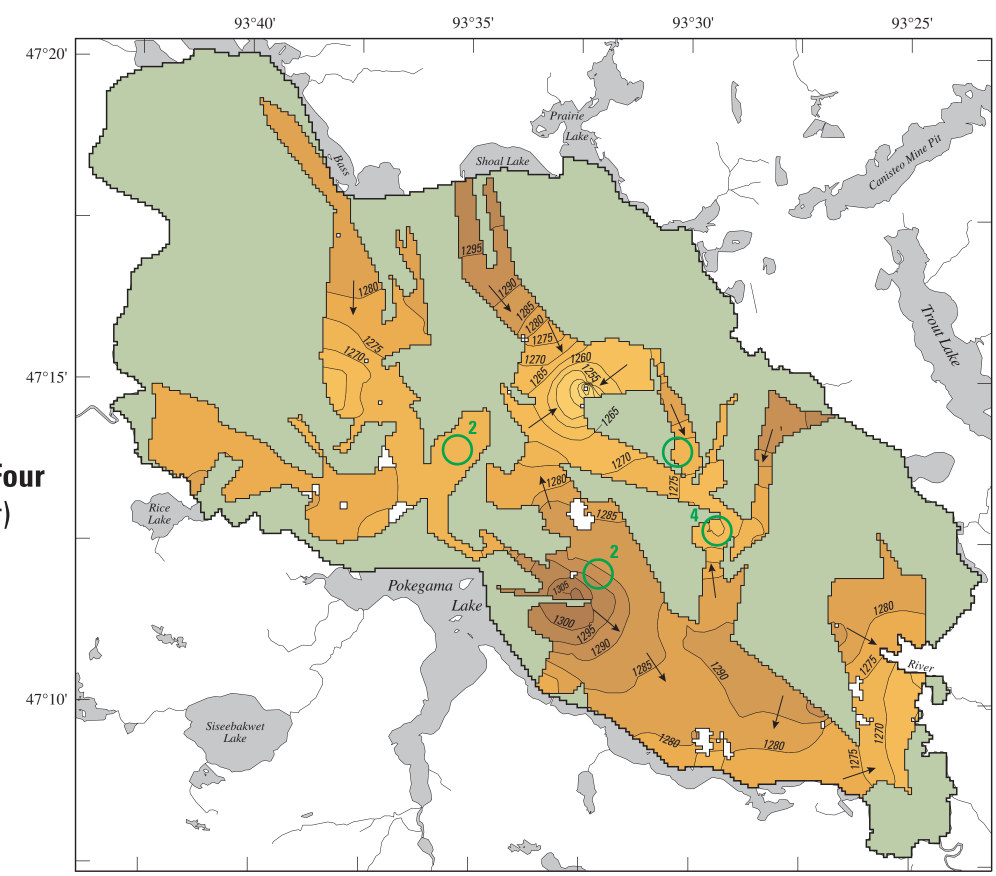




Upper aquifer, (b) Model Layer four—Middle aquifer, and (c) Model Layer six—Lower aquifer, in the Grand Rapids area, Minnesota



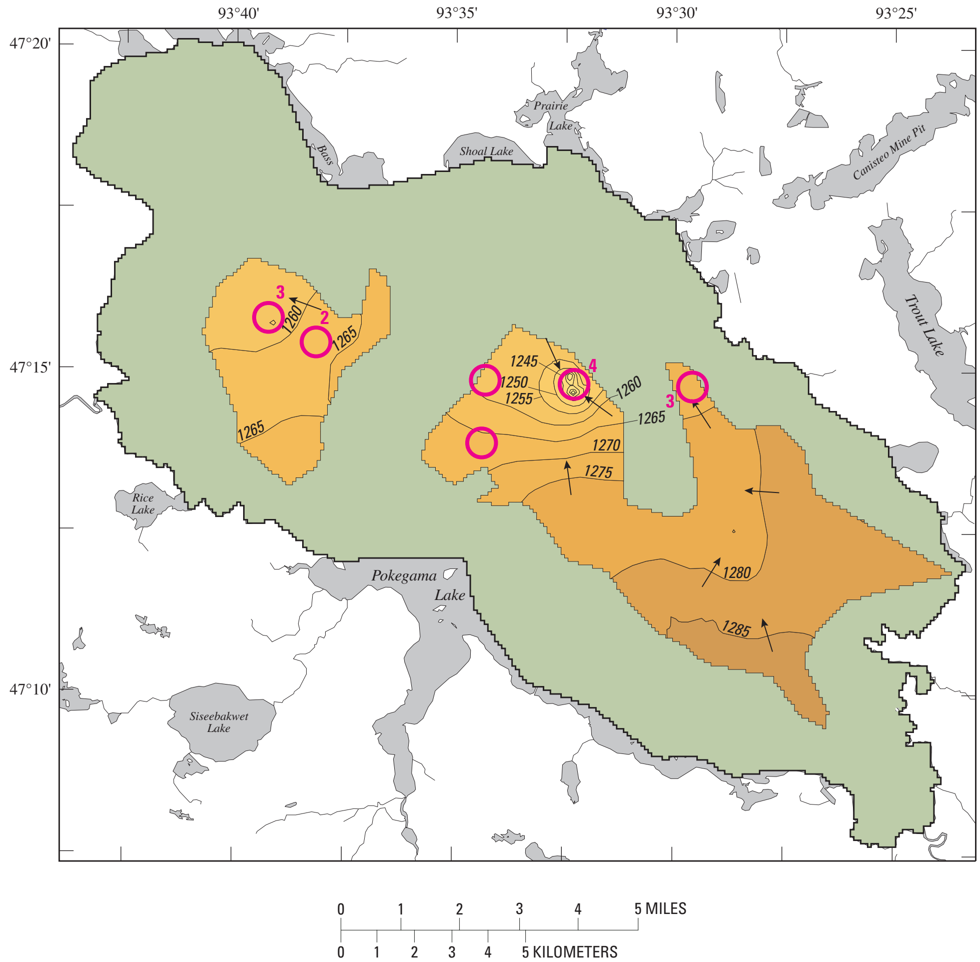
(a) Model Layer Two
(Upper aquifer)



(b) Model Layer Four
(Middle aquifer)

Figure 6. Simulated potentiometric surfaces for ground-water flow and ground-water flow direction for the (a) Model Layer two—

(c) Model Layer Six
(Lower Aquifer)



EXPLANATION

Glacial till

Dry cells

Model grid boundary

Public supply well layer 4 (general location) (number indicates wells in a cluster)

Public supply well layer 6 (general location) (number indicates wells in a cluster)

Pumping from public supply wells were represented in the model as water withdrawn from a single cell within the general location.

Generalized direction of ground-water flow

Potentiometric surface (in feet above NAVD 88, interval = 5 feet)

1,320 to 1,325
1,310 to 1,315
1,300 to 1,305
1,290 to 1,295
1,280 to 1,285
1,270 to 1,275
1,260 to 1,265
1,250 to 1,255
1,240 to 1,245
1,230 to 1,235
1,220 to 1,225

Upper aquifer, (b) Model Layer four—Middle aquifer, and (c) Model Layer six—Lower aquifer, in the Grand Rapids area, Minnesota

Table 5. Relation among model variables used in sensitivity analysis for the calibrated MODFLOW simulation of ground-water flow in the upper, middle, and lower aquifers in the Grand Rapids area, Minnesota

[---, no value]

Groups of wells	Multiplication factor	Root mean square error in ground-water levels (ft)				
		Areal recharge	Horizontal hydraulic conductivity	Vertical hydraulic conductivity	River-bed conductance	General-head boundary conductance
Wells completed in upper aquifer	0.6	12.70	14.01	12.99	12.83	12.21
	0.7	12.73	13.65	12.89	12.83	12.30
	0.8	12.76	13.35	12.83	12.83	12.50
	0.9	12.80	13.06	12.83	12.83	12.66
	1.0	12.83	12.83	12.83	12.83	12.83
	1.1	12.86	12.66	12.80	12.83	12.96
	1.2	12.89	12.50	12.80	12.83	13.12
	1.3	12.89	12.57	12.80	12.83	13.26
	1.4	12.93	---	12.80	12.83	13.39
Wells completed in middle aquifer	0.6	19.03	19.00	18.83	19.10	18.18
	0.7	18.96	18.93	18.73	19.10	18.28
	0.8	19.00	18.83	18.67	19.10	18.93
	0.9	19.03	18.64	18.64	19.10	19.03
	1.0	19.10	19.10	19.10	19.10	19.10
	1.1	19.13	18.96	19.03	19.10	19.16
	1.2	19.16	18.87	18.96	19.10	19.19
	1.3	19.16	18.83	18.93	19.10	19.26
	1.4	19.03	---	18.90	19.10	19.33
Wells completed in lower aquifer	0.6	18.05	17.95	18.18	18.21	17.88
	0.7	18.05	17.91	18.57	18.21	17.98
	0.8	18.08	17.68	18.37	18.18	18.08
	0.9	18.14	17.88	18.28	18.18	18.14
	1.0	18.18	18.18	18.18	18.18	18.18
	1.1	18.24	18.01	18.11	18.18	18.24
	1.2	18.28	17.88	18.08	18.18	18.31
	1.3	18.28	17.75	18.01	18.18	18.31
	1.4	18.24	---	17.98	18.18	18.41
Monitored reaches	**Multiplication factor**	**Percentage error in base flow**				
		Areal Recharge	**Horizontal hydraulic conductivity**	**Vertical hydraulic conductivity**	**River-bed conductance**	**General-head boundary conductance**
Eastern reach of Mississippi River	0.6	-9.1	-27.4	-9.6	-5.0	-15.0
	0.7	-7.0	-20.1	-8.4	-4.7	-12.0
	0.8	-6.1	-14.8	-6.9	-4.5	-9.5
	0.9	-5.2	-9.7	-5.6	-4.4	-6.4
	1.0	-4.1	-4.1	-4.1	-4.1	-4.1
	1.1	-3.4	1.0	-3.2	-4.3	-2.4
	1.2	-2.5	5.3	-2.1	-4.2	-0.7
	1.3	-1.0	8.0	-1.1	-4.0	-1.1
	1.4	-0.9	---	-0.1	-4.0	-2.7
Western reach of Mississippi River	0.6	-5.3	-24.1	-5.0	-0.5	-12.8
	0.7	-3.9	-17.3	-4.1	-0.4	-9.1
	0.8	-2.4	-11.1	-2.2	-0.2	-5.8
	0.9	-1.6	-6.0	-1.1	-0.1	-2.8
	1.0	-0.8	-0.8	-0.8	-0.1	-0.8
	1.1	0.7	4.9	0.9	-0.1	1.7
	1.2	2.1	9.5	1.9	-0.1	4.8
	1.3	3.2	13.7	2.8	0.0	7.0
	1.4	4.2	---	3.7	-0.1	8.3
Reach of Prairie River	0.6	4.8	-5.6	-7.1	8.6	-1.6
	0.7	5.8	1.2	-3.1	8.8	1.7
	0.8	5.5	-1.8	0.1	8.9	4.7
	0.9	8.0	3.8	6.0	9.0	7.0
	1.0	9.1	9.1	9.1	9.1	9.1
	1.1	10.0	13.7	12.0	9.0	10.8
	1.2	10.9	15.0	14.8	9.2	12.2
	1.3	11.7	16.6	20.8	9.2	13.6
	1.4	10.0	---	23.5	9.2	14.8

tions were run using various vertical hydraulic conductivity, river-bed conductance, and general-head-boundary values, varying the values by the same multiplication factors as in the areal recharge simulations. Horizontal hydraulic conductivity values were varied by factors of 0.6, 0.7, 0.8, 0.9, 1.1, 1.2, and 1.3 times the calibrated values in seven simulations. Sensitivity simulations were compared to the calibrated model simulations by observing changes in the root mean square error in water levels in the wells and percentage error in the base-flow rates to the Mississippi and Prairie River reaches.

Simulated water levels were most sensitive to changes in general-head-boundary conductance (fig. 7 and table 5), indicating that it is the predominant variable controlling steady-state water levels. Simulated base-flow for the model was most sensitive to changes in horizontal hydraulic conductivity (fig. 7, table 5), indicating that it is the predominant variable controlling steady-state flow conditions. Compared to general-head-boundary conductance and horizontal hydraulic conductivity, the model was less sensitive to changes in vertical hydraulic conductivity and areal recharge values (table 5).

The model was least sensitive to the changes in river-bed conductance (table 5 and fig. 7). This low sensitivity may be due in part to the smaller values for river-bed conductance relative to the horizontal hydraulic conductivity values used in the model and the smaller areal extent of the river beds relative to the extent of the aquifers, general-head boundaries, and areal recharge.

The model sensitivity varied among the three aquifers and among the different reaches. Simulated ground-water levels in the upper aquifer were most sensitive to changes in horizontal hydraulic conductivity (table 5), whereas ground-water levels in the middle aquifer were most sensitive to changes in general-head boundary conductance. Simulated ground-water levels in the lower aquifer were equally sensitive to changes in horizontal hydraulic conductivity, vertical hydraulic conductivity, and general-head boundary conductance (table 5). Base flow to the simulated reaches for the Mississippi River were most sensitive to changes in horizontal hydraulic conductivity, whereas base flow to the Prairie River reach was most sensitive to changes in vertical hydraulic conductivity of the aquifer (table 5).

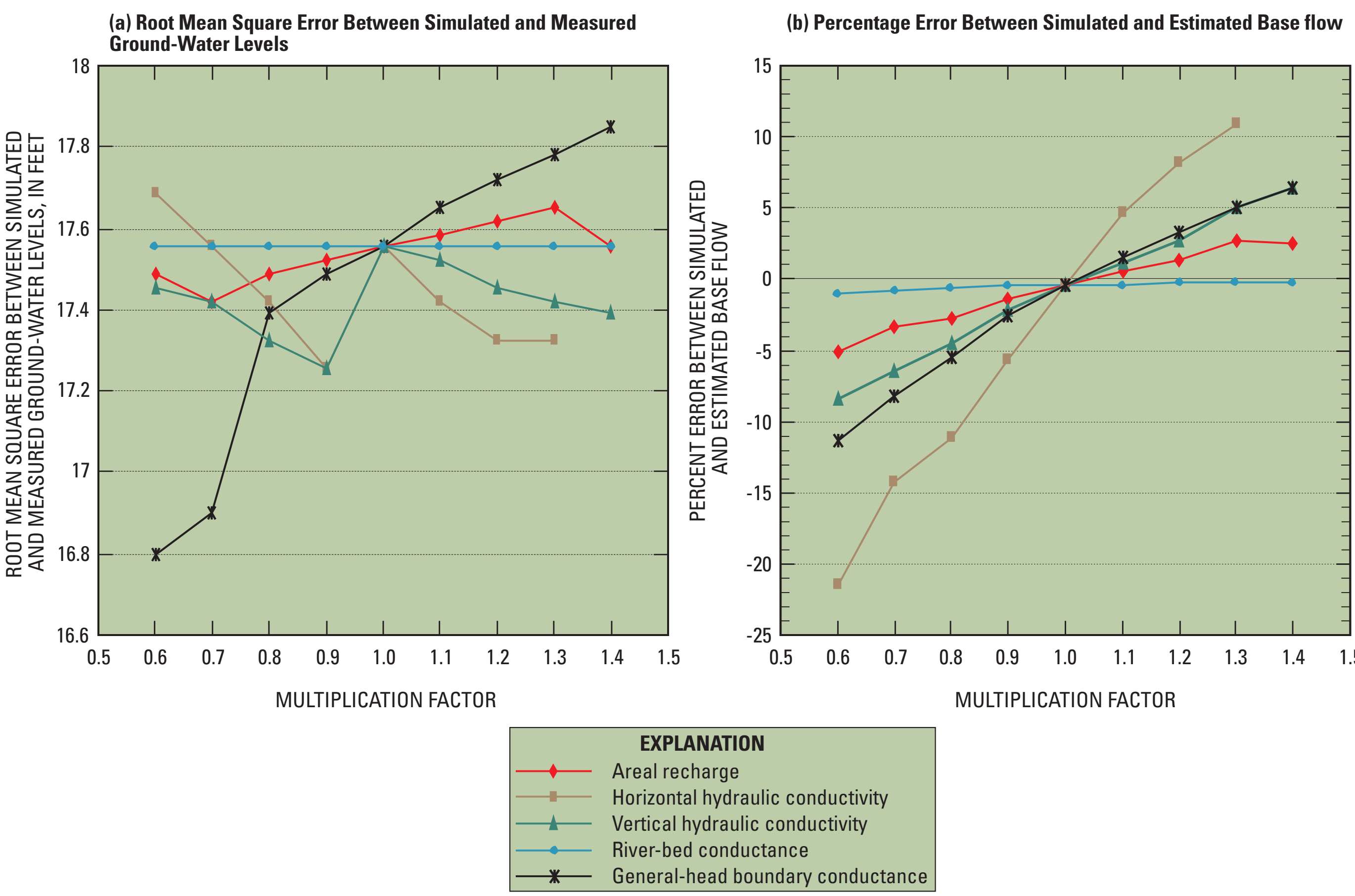


Figure 7. Model sensitivity to selected input variables based on (a) root mean square error between simulated and measured ground-water levels, and (b) percentage error between simulated and estimated base flow for the calibrated MODFLOW simulation of ground-water flow in glaciofluvial aquifers in the Grand Rapids area, Minnesota

Model Limitations and Accuracy

The numerical flow model is a simplification of a complex glaciated-terrain, ground-water flow system. The accuracy of the simulation is limited to the accuracy, amount, and distribution of the data used to describe the hydrologic variables of the flow system and the accuracy of the assumption of steady-state. These variables include the hydrologic properties of the aquifers and confining units, areal recharge rates, and hydrologic boundary conditions. Variables determined for the model during calibration may not be unique. Different combinations of variables for the model could produce similar results from the model. The model grid was designed to simulate hydrologic conditions at the scale of the study area. For example, the model can be modified to include storage terms and used with particle-tracking software (Pollock, 1994) to determine approximate areal extents of the zones of contribution for discharging wells in the study area on a regional scale. However, the model grid could be refined to address hydrologic issues on a smaller, more accurate scale, such as the estimation of the areal extent of the zone of contribution for individual public supply wells.

The accuracy of a model depends upon the accuracy of the measured data used to calibrate a model. The model was calibrated using water-level data recorded, for the most part, within a week following well installation; therefore, it is unlikely that static conditions were present for all of the wells following installation. These wells were installed between 1975 and 1999, when hydrologic conditions varied between periods of droughts to periods of above-normal precipitation. Thus, these water levels do not necessarily represent steady-state conditions. Also, the base-flow rates used for calibration were estimated from values obtained by Payne (1995) over large reaches of the Mississippi River. Any local variations in the base-flow values were not taken into account in the estimated values. Therefore, the model may not accurately represent small-scale flow conditions where local variations from more regional patterns may exist.

The calibrated, steady-state model is a tool for water-resources management based on the assumption that future hydrologic conditions will be similar to historical, or assumed steady-state, conditions. Furthermore, it would be assumed that variations in annual recharge and discharge for a hypothetical simulation using the calibrated model are similar to hydrologic conditions of the calibrated model. Accuracy of the hypothetical simulation becomes more uncertain if the variation in annual recharge or discharge exceeds the range used in the calibrated model. Because the numerical model was calibrated under the assumption of steady-state flow conditions, the model will most accurately reflect the effects of long-term, annual or multiple-year stresses.

SUMMARY

A calibrated steady-state, finite-difference, ground-water flow model was constructed to simulate a system of three glaciofluvial aquifers, defined as the upper, middle, and lower aquifers, in an area of about 114 mi^2 surrounding the city of Grand Rapids in north-central Minnesota. This model was constructed as part of a 5-year cooperative study between the U.S. Geological Survey (USGS) and the Minnesota Department of Health (MDH) to simulate ground-water-flow conditions in aquifers. The calibrated model will be used by MDH and communities in the Grand Rapids area in the development of wellhead protection plans for their water supplies.

The USGS MODular ground-water-FLOW model, MODFLOW, was used to simulate ground-water flow conditions in the Grand Rapids area. Available geologic and hydrologic data collected from 1975 through 2001 were used for model construction and calibration. Data from scientific literature and other sources also were used. A series of GIS coverages were used to identify and represent hydrologic features and processes in the model. Data sets for perennial wetlands, lakes, and rivers were obtained from the National Wetlands Inventory data base. A “true-layer” approach was used to define layering represented by the model, explicitly defining altitudes and aquifer hydraulic properties of cells in each model layer based on the three-dimensional geologic model representation. Perennial rivers, streams, lakes, and wetlands were specified as head-dependent boundaries in five of the six layers in the model. Lakes and wetlands were simulated as general-head boundaries, and streams and were simulated using the river package of MODFLOW. A specified-flux boundary was used to represent areal recharge to the upper layer (layer 1) of the model using the recharge package in MODFLOW. Ground-water withdrawals from the 23 public supply wells and 2 irrigation wells were simulated in the model using the well package in MODFLOW.

The model was calibrated through comparison of simulated ground-water levels to measured static water levels in 351 wells, and comparison of simulated base-flow rates to estimated base-flow rates for two reaches of the Mississippi River (eastern and western) and a reach of the Prairie River. Measurement dates for the static water levels in the wells ranged from 1975 to 1999. Model statistics indicate that the model tends to overestimate ground-water levels. The root mean square errors ranged from +12.83 ft in wells completed in the upper aquifer to +19.10 ft in wells completed in the middle aquifer. Mean absolute differences between simulated and measured water levels ranged from +4.43 ft for wells completed in the upper aquifer to +9.25 ft for wells completed in the middle aquifer. Mean algebraic differences ranged from +9.35 ft for wells completed in the upper aquifer to +14.44 ft for wells completed in the middle aquifer, with the positive differences indicating that positive water levels were not balanced by the negative differences. Percentage errors between simulated and estimated base-flow rates for the three moni-

tored reaches were less that 10 percent, indicating good agreement. Simulated ground-water levels were most sensitive to changes in general-head boundary conductance, indicating that this characteristic is the predominant model input variable controlling steady-state water-level conditions. Simulated ground-water flow to stream reaches was most sensitive to changes in horizontal hydraulic conductivity, indicating that this characteristic is the predominant model input variable controlling steady-state flow conditions.

The accuracy of the model is limited to the accuracy, amount, and distribution of the data used to describe the hydrologic variables of the flow system and the accuracy of the assumption of steady-state. The model grid was designed to simulate hydrologic conditions at the scale of the study area. The model may not accurately represent small-scale flow conditions where local variations from more regional patterns may exist.

ACKNOWLEDGMENTS

The author thanks the Minnesota Department of Health, in particular Mr. James Walsh and Mr. Bruce Olsen, for providing assistance needed to complete this study. The author thanks the Minnesota Department of Natural Resources, in particular Mr. Sean Hunt and Mr. Kevin Zytkovicz, for providing annual pumping records for 1997-2001 for public supply wells and bathymetric data for lakes in the study area. The author thanks Minnesota Power, in particular Lowell Neudahl, for providing bathymetric data for Blackwater Lake.

REFERENCES

Harbaugh, A.W., and McDonald, M.G., 1996, User's documentation for MODFLOW-96, an update to the U.S. Geological Survey Modular Finite-Difference Ground-Water Flow Model: U.S. Geological Survey Open-File Report 96-485, 63 p.

Hill, M.C., 1990, Preconditioned conjugate-gradient 2 (PCG2)—A computer program for solving ground-water flow equations: U.S. Geological Survey Water-Resources Investigations Report 90-4048, 43 p.

Jirsa, M.A., Setterholm, D.R., Bloomgren, B.A., Bauer, E.J., and Lively, R.S., 2002, Bedrock topographic and depth to bedrock maps of the western half of the Mesabi Iron Range, Northern Minnesota: Minnesota Geological Survey, Miscellaneous Map Series, Map M-126.

Jones, P.M., 2002, Characterization of ground-water flow between the Canisteo mine pit and surrounding aquifers, Mesabi Iron Range, Minnesota: U.S. Geological Survey Water-Resources Investigations Report 02-4198, 30 p.

Kishel, H.F., and Gerla, P.J., 2002, Effects of small-scale hydraulic variation on ground-water discharge to Shingobee Lake, Minnesota: The University of North Dakota, Department of Geology web page, accessed April 23, 2002, at URL http://www.geology.und.nodak.edu/gerla/home/cur_res.html

McDonald, M.G., and Harbaugh, A.W., 1988, A modular three-dimensional finite-difference ground-water flow model: U.S. Geological Survey Techniques of Water-Resources Investigations, book 6, chap A-1, p. 2-1 – 2-35.

Minnesota Department of Health, 2003, Community Public Water Supply General Information: accessed March 18, 2003, at URL http://www.health.state.mn.us/divs/eh/water/index.htm

Minnesota Department of Natural Resources, 2002a, LandSat-based land use-land cover (Raster), 1995 (Minnesota): Minnesota Department of Natural Resources online metadata, accessed February 11, 2002, at URL http://deli.dnr.state.mn.us/metadata/full/lusatra3.html

Minnesota Department of Natural Resources, 2002b, Minnesota Water Use Information (Raster), 1996 (Minnesota): Minnesota Department of Natural Resources online metadata, accessed February 14, 2002, at URL http://www.dnr.state.mn.us/waters/programs/water_mgt_section/appropriations/wateruse.html

Minnesota Department of Natural Resources, 2002c, Lake finder, 1996: accessed March 25, 2002, at URL http://www.dnr.state.mn.us/lakefind/index.html

Minnesota State Climatologist, 2001, Normal precipitation and temperature data: accessed July 11, 2001, at URL http://climate.umn.edu/doc/historical/normals.htm

Morey, G.B., 1972, Mesabi Range, *in* Sims, P.K., and Morey, G.B., eds., Geology of Minnesota—A Centennial Volume: Minnesota Geological Survey, p. 204-217.

Oakes, E.L., 1970, Geology and ground-water resources of the Grand Rapids area, north-central Minnesota: U.S. Geological Survey Hydrologic Investigations Atlas, HA-322, scale 1:48,000.

Payne, G.A., 1995, Ground-water base flow to the Upper Mississippi River upstream of the Minneapolis-St. Paul Area, Minnesota during July 1988: U.S. Geological Survey Open-File Report, 94-478, 28 p.

Pollock, D.W., 1994, User's guide for MODPATH/MODPATH-PLOT, Version 3—A particle tracking post-processing package for MODFLOW, the U. S. Geological Survey finite-difference ground-water flow model: U.S. Geological Survey Open-File Report 94-464, 249 p.

Winter, T.C., 1971, Sequence of glaciation in the Mesabi-Vermilion Iron Range area, northeastern Minnesota: U.S. Geological Survey Professional Paper 750-C, p. C82-C88.

Winter, T.C., 1973a, Hydrogeology of glacial drift, Mesabi Iron Range, northeastern Minnesota: U.S. Geological Survey Water-Supply Paper 2029-A, 23 p.

Winter, T.C., 1973b, Petrography and stratigraphy of glacial drift, Mesabi-Vermilion Iron Range area, northeastern Minnesota: U.S. Geological Survey Bulletin 1331-C, 41 p.

U.S. Census Bureau, 2002, Population data: accessed February 6, 2002, at URL http://factfinder.census.gov/servlet/BasicFactsServlet

U.S. Department of Defense, 1998, Groundwater modeling system, user's manual: Brigham Young University, 620 p.

U.S. Fish & Wildlife Service, 1994, National wetlands inventory digital data, October 1991 to 1994: accessed February 11, 2002, at URL http://www.nwi.fws.gov