Simulation of Ground-Water Flow in the Glaciofluvial, Saginaw, Parma-Bayport, and Marshall Aquifers, Central Lower Peninsula of Michigan

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By J.R. Hoaglund, G.C. Huffman, and N.G. Grannemann

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CONVERSION FACTORS, VERTICAL DATUM, AND ABBREVATIONS

For use of readers who prefer the International System of Units (SI), the conversion factors for terms used in this report are listed below.

Multiply	By	To obtain
inch (in.)	2.54	centimeter
foot (ft)	0.3048	meter
mile (mi)	1.609	kilometer
square mile (mi^2)	2.59	square kilometer
foot per day (ft/d)	0.3048	meter per day
cubic foot (ft^3)	0.02832	cubic meter (m^3)

Sea level: In this report "sea level" refers to National Geodetic Vertical Datum of 1929 (NGVD of 1929), a geodetic datum derived from a general adjustment of the first-order level nets of both the United States and Canada, called Mean Sea Level of 1929.

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ABSTRACT

A steady-state finite difference model was developed to simulate ground-water flow in four regional aquifers in Michigan's Lower Peninsula. The Glaciofluvial, Saginaw, Parma-Bayport, and Marshall aquifers were simulated as layers 1 through 4, respectively, in the model. Separately calculated vertical conductances input to the model were used to simulate the intervening Till/"Red Beds", Saginaw, and Michigan confining units, respectively. The model domain was laterally bound by a continuous specified-head boundary, formed from Lakes Michigan, Huron, St. Clair, and Erie, together with the St. Clair and Detroit River connecting channels.

The model was developed to quantify regional ground-water flow in the aquifer systems using independently determined recharge estimates. The flow model showed that ground-water heads and flows in the Glaciofluvial aquifer are controlled by local stream stages and discharges, resulting in localized flow cells accounting for 95-percent of the overall model water budget. Simulation of recharge to an unspecified water table also enabled the estimation of ground-water discharge to three Great Lakes.

A computer diskette contains all MODFLOW and MODFLOWP input files, as well as digital model surfaces and several Fortran processing routines used to construct the surfaces. The diskette also provides the data used for calibration and sensitivity analysis.

INTRODUCTION

Mississippian through Jurassic bedrock units within the central Lower Peninsula of Michigan, together with peninsula-wide Pleistocene glacial deposits, form a regional system of aquifers and confining units in the central Lower Peninsula of Michigan. Comprehensive hydrogeological, geochemical, and ground-water recharge investigations of this regional aquifer system were completed as part of the U.S. Geological Survey (USGS) Regional Aquifer-System Analysis (RASA) project for the Michigan Basin, one of 25 RASA hydrogeologic investigations of regional aquifer systems in the United States (Sun and others, 1997). The purposes of the RASA program were to define the regional geohydrology and geochemistry of the major regional ground-water systems of the United States and to establish a framework of background information that can be used for regional assessment of ground-water resources. A purpose of the Michigan Basin RASA study, as defined by Mandle (1986), was to "develop a regional ground-water-flow model to simulate present, and possibly, paleo-ground-water-flow directions [and] through model simulation [monitor] future water use [and] water levels ..."

The Michigan Basin RASA study area included 56,980 km^2 (22,000 mi^2) of Michigan's Lower Peninsula. The Michigan Basin RASA project included study of the geology (Westjohn and others, 1994; Westjohn and Weaver, 1996 a,b,c; Westjohn and Weaver, 1998), aqueous geochemistry (Wahrer and others, 1996; Meissner and others, 1996; Ging and others, 1996), water level (Barton and others, 1996), and recharge (Holtschlag, 1996; 1997) of aquifers that are primary sources of ground-water supply for human needs. The following ground-water modeling analysis completes the major objective of the Michigan Basin RASA study: to better understand the natural ground-water flow system that existed prior to large-scale withdrawal of ground water, brine, gas, and oil, using the information obtained from the studies of the hydrogeologic framework and the geochemistry of ground water, as well as using basic hydrologic principles. An analysis of paleo-ground-water flow directions was provided by Hoaglund (1996).

Background

The Michigan Basin aquifer system is a major source of ground water in the central Lower Peninsula of Michigan. Ground-water flow in the aquifer system is directly linked to areal recharge, stream baseflow, and Great Lakes direct riparian discharge. An initial simulation of the RASA study area assumed a fixed water table and simulated the effects of the water table on the bedrock hydrology (Mandle and Westjohn, 1989). However, this model did not simulate or balance ground-water inputs and outputs other than flows between specified heads. The actual ground-water system balances areal recharge inputs to discharge outputs, including baseflow to the rivers, and a component of direct ground-water discharge to the inland lakes and Great Lakes.

Although the RASA study area was restricted to the 56,980 km^2 (22,000 mi^2) defined by the extent of the bedrock units, the incorporation of information on the peninsula-wide glacial units, including land-surface elevation and glacial thickness information from the Western Michigan University hydrogeologic atlas (1981), extended the modeled area across nearly the entire Lower Peninsula of Michigan, with the Great Lakes as boundaries on three sides of the model. As a result, direct riparian ground-water discharge to the Great Lakes was calculated and the results are presented below.

River discharge provides the "runoff" (i.e. tributary) hydrologic inputs to the inland lakes and Great Lakes. This runoff component to the Great Lakes is significant, including 43 percent of the non-diversion hydrologic inputs to Lake Michigan, with the remaining 57 percent provided by direct precipitation (Croley and Hunter, 1994). Quinn (1992) used estimates of runoff, precipitation, connecting channel flow, and diversion flow to calculate hydraulic residence times for the Great Lakes, but these estimates did not include a component of direct, riparian ground-water discharge (Quinn, oral communication, 1999). The ground-water model in this report uses estimates of ground-water recharge (Holtschlag, 1996; 1997) to a simulated water table, and provides estimates of both the direct (riparian) and indirect (stream baseflow) ground-water discharges to three of the Great Lakes from the Lower Peninsula of Michigan.

This report presents the design, calibration, and sensitivity analysis of the digital model of ground-water flow within the central Lower Peninsula of Michigan, work conducted both as part of the RASA project as well as individual research efforts continued outside the USGS since the RASA program was terminated, and attempts to describe the regional, pre-development ground-water flow system. The report includes sensitivity analysis of calculations and reports limitations for using the model. Parameter estimation of selected hydrologic parameters, and sensitivity to boundary conditions and stresses, including pumping, will be the focus of planned future publications.

Acknowledgments

The authors would like to thank David Westjohn for guidance through complex Michigan Basin geology, and David Holtschlag for interpretation of the many complex signals that must be incorporated into optimum recharge estimation. Both individuals patiently contributed to our understanding through their manuscripts and countless discussions. The authors gratefully acknowledge the following organizations and individual for their support throughout this study: Michigan State University, Department of Geological Sciences; the University of Michigan, Department of Geological Sciences; and Rose Bohn.

HYDROGEOLOGY

Physiographic Setting

The entire Lower Peninsula of Michigan is underlain by layered sedimentary rock and almost entirely overlain by glacial deposits. Three of the Great Lakes and their connecting channels bound the Lower Peninsula (fig. 1). Upland and lowland physiographic provinces were described by Leverett and Taylor (1915). Three lowland areas are shown on figure 2. Land surface altitudes in the Saginaw and Michigan lowlands range from 177 to about 213 meters (580 to about 700 ft); in the Erie lowlands they range from 174 to about 229 meters (572 to about 750 ft). Lowland areas are generally flat to gently sloping toward Saginaw Bay, Lake Michigan, Lake Huron, Lake Erie or the connecting channels. Two upland areas are shown on figure 2, delimited by 244-meter (800 feet) water table contours. Land surface altitudes range from about 260 to 525 meters (850 to 1,725 ft) in the Northern Uplands and from about 250 to 365 meters (825 to 1,200 ft) in the Southern Uplands.

Geologic Setting

The Michigan Basin is an ovate shaped accumulation of sedimentary rocks in the Lower Peninsula of Michigan and parts of Michigan's Upper Peninsula, Wisconsin, Illinois, Indiana, Ohio, and Ontario, Canada. The maximum thickness of Precambrian through Jurassic rocks is about 5,334 meters (17,500 ft) (Lillienthal, 1978). Paleozoic through Jurassic rocks are mantled by glacial deposits that are the result of the Wisconsinan and, possibly, earlier glaciations. Ice from the last glaciation receded from Michigan about 10,000 years ago (Eschman, 1985, p. 164).

Aquifers and Confining Units

Four major aquifers were identified in the Michigan Basin RASA study area: the Glaciofluvial, Saginaw, Parma-Bayport, and Marshall aquifers (fig. 3). The four aquifers are underlain by four confining units: the Till/Red Beds, Saginaw, Michigan, and Coldwater Shale, respectively. These units are described stratigraphically from top to bottom below.

The Glaciofluvial aquifer consists dominantly of thick sequences of glacial and and/or fluvial sand and gravel. In parts of the study area, however, it consists of sand and gravel beds within till or other fine-grained glacial deposits. For this study, the uppermost aquifer is referred to as a single unit, even though it is composed of multiple layers of sand and gravel. More than 120 meters (400 ft) of freshwater-bearing glacial deposits overlie finer grained deposits including “red beds” of Jurassic age and finer grained glacial deposits throughout most of the study area.

Together with the fine-grained glacial deposits, "red beds" form a subregional confining unit (Westjohn and others, 1994). The "red beds" separate the Glaciofluvial aquifer from the Saginaw aquifer in the west-central part of the study area. These "red beds" are primarily composed of red mud, poorly consolidated red shale, gypsum, and minor amounts of sandstone.

Hydrogeological characterization of the Saginaw and Grand River Formations defined the Saginaw aquifer as “the cumulative thickness of sandstones that overlie the Saginaw confining unit...” (Westjohn and Weaver, 1996a, p. 9). This lithology of the Pennsylvanian rock sequence includes sandstone, siltstone, shale, coal and limestone, but consists primarily of shale in the lower part and sandstones in the upper part (Westjohn and Weaver, 1996a). These observations distinguished an upper sandstone

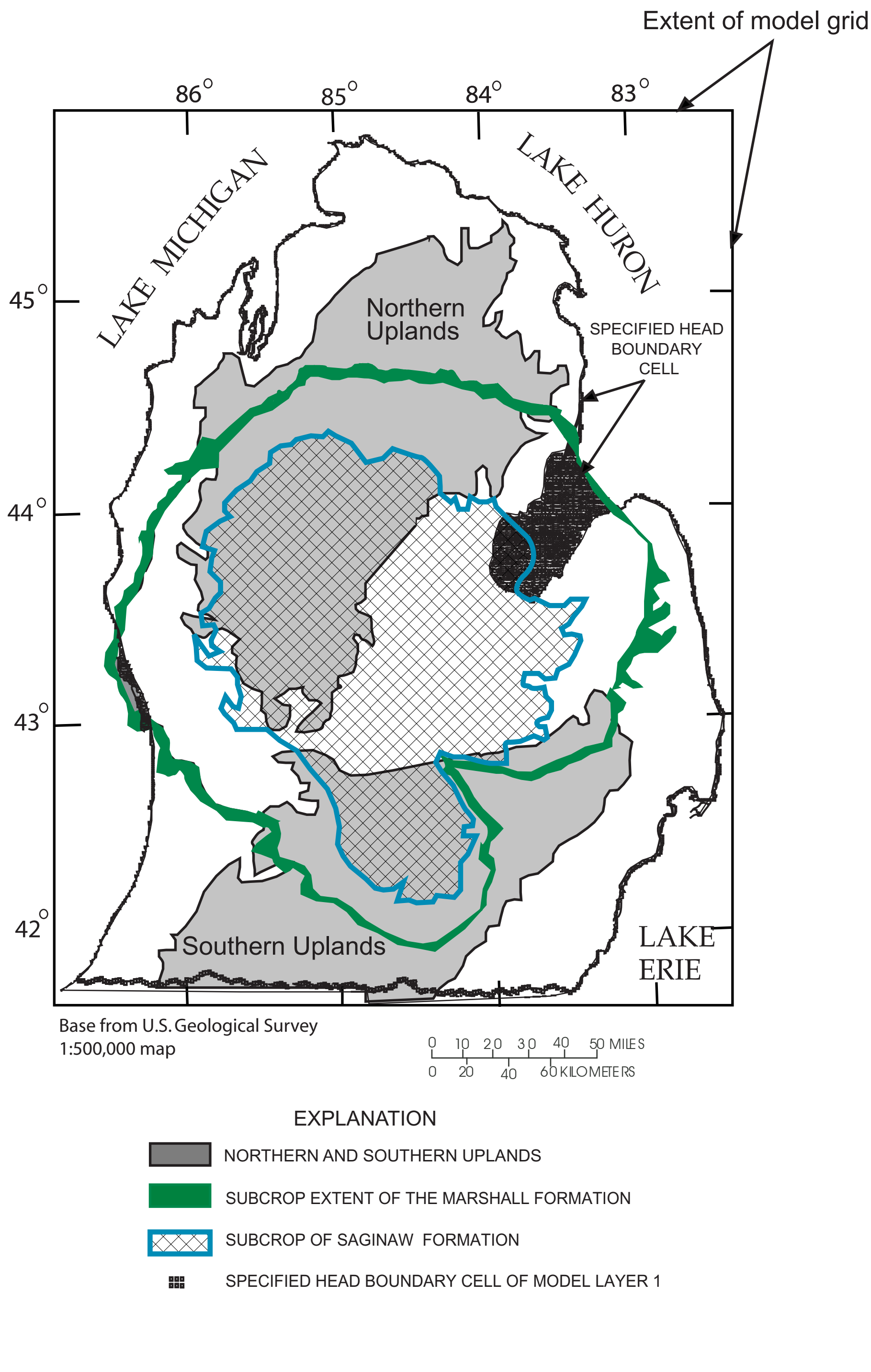


Figure 1. Location of the RASA Study Area in the Lower Peninsula of Michigan.

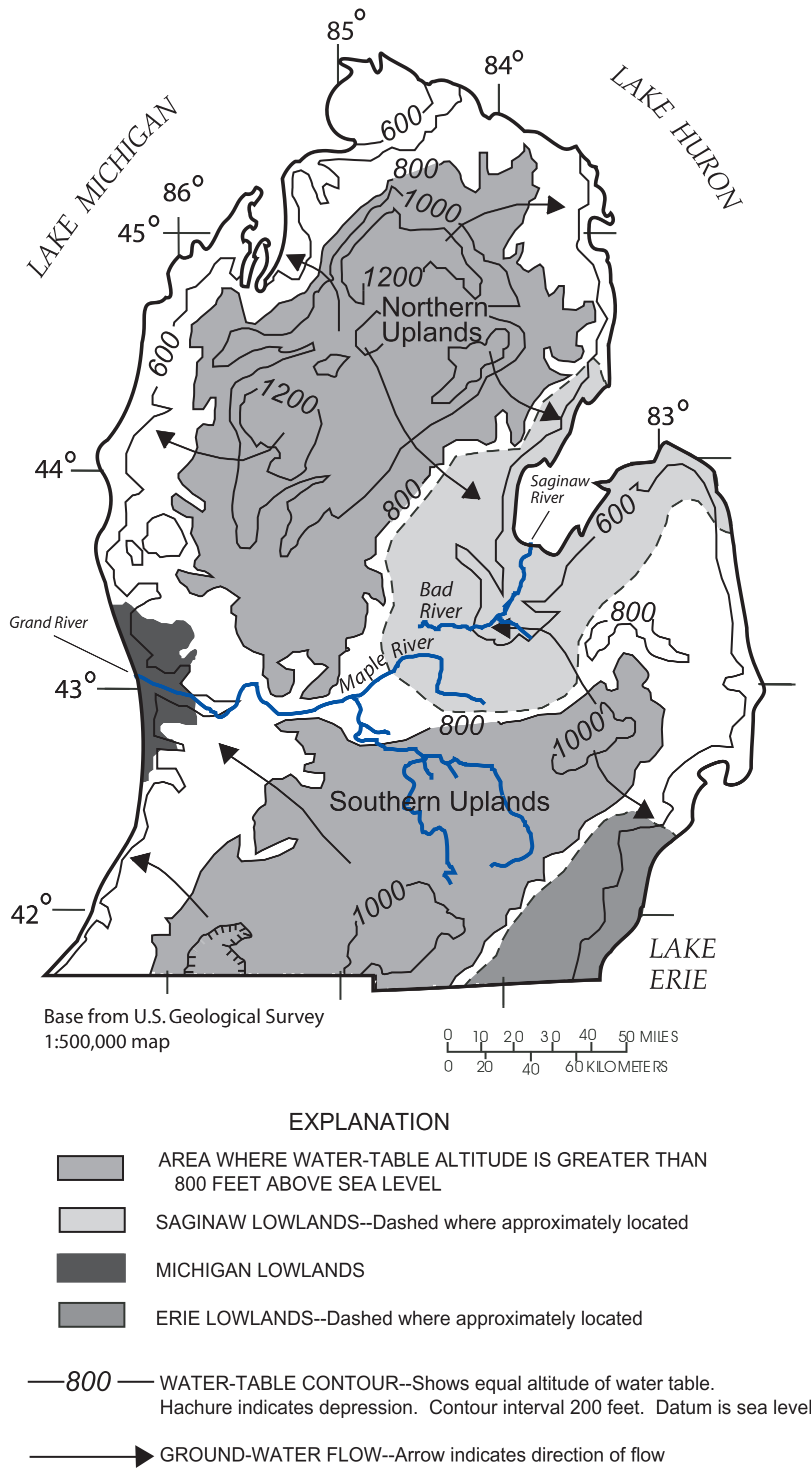


Figure 2. General configuration of the water-table in the Glaciofluvial aquifer.

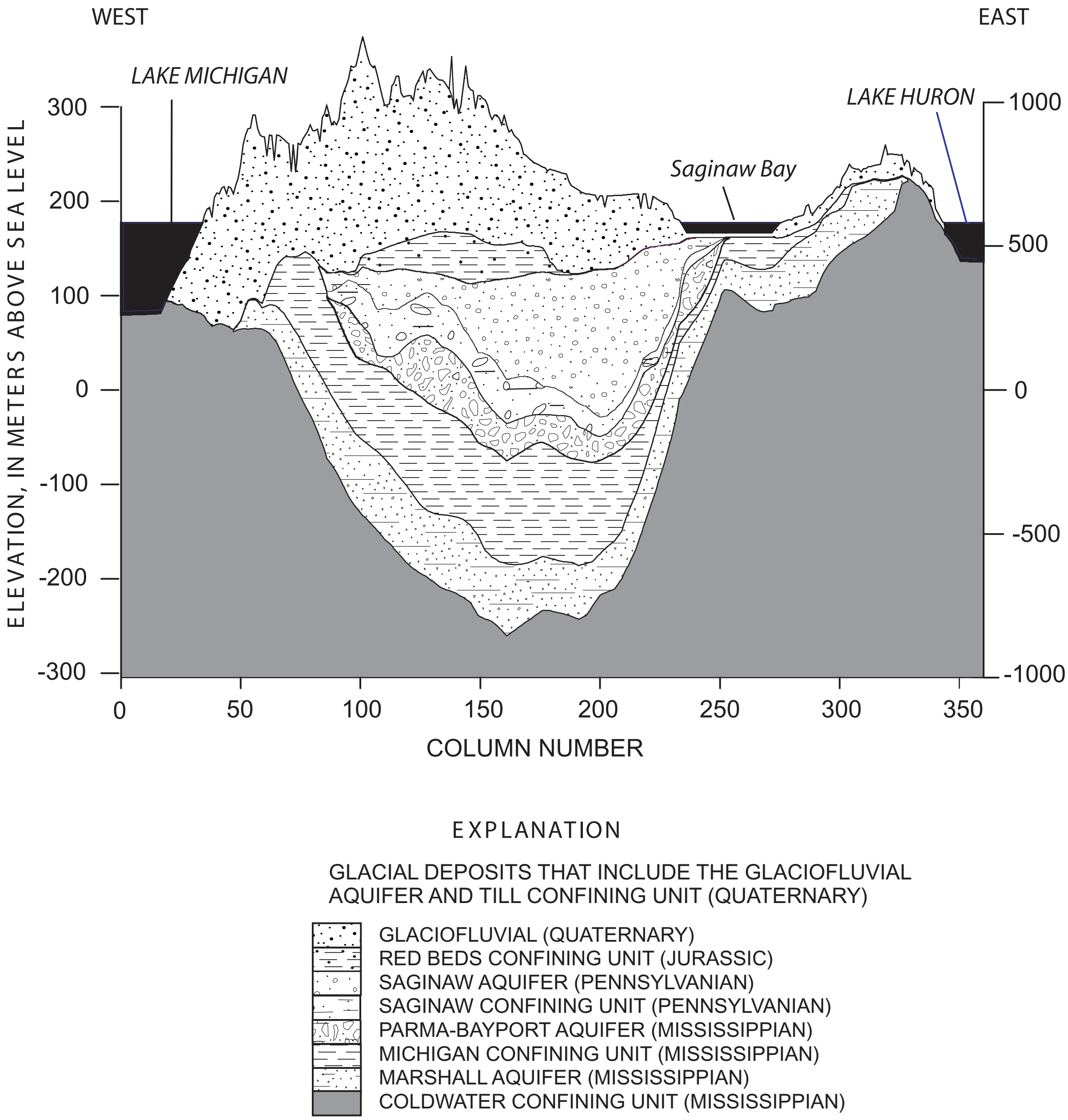


Figure 3. Geologic section showing structural relations among aquifers and confining units.

aquifer from a lower shale confining unit, though both are complexly intercalated sandstone and shale. For the purposes of computer simulation, the upper sandstones are assumed to be hydraulically connected at the scale of the study area. These sandstones are the most productive aquifer material of the unit, but they are generally less than 30 meters (100 ft) thick, except in the east-central part of the basin, where the composite thickness of sandstone ranges from 60 to 113 meters (200 to 370 ft) (Westjohn and Weaver, 1996a).

In most areas of the basin, the lower, predominantly shale portion of the Saginaw Formation underlies the Saginaw aquifer. This shale, which constitutes the Saginaw confining unit, separates the Saginaw aquifer from the underlying Parma-Bayport aquifer, and ranges in thickness from 0 to 91 meters (300 ft) (Westjohn and Weaver, 1996a).

The Parma-Bayport aquifer ranges in thickness from 30 to 45 meters (100 to 150 ft) and consists of the Parma Sandstone and the Bayport Limestone (Westjohn and Weaver, 1996a). The Parma Sandstone contains sandstone, shale, siltstone, and thin lenses of limestone. The Bayport Limestone is predominantly limestone, sandstone, and sandy limestone.

The Michigan confining unit underlies the Parma-Bayport aquifer, and is an intercalated sequence of thin bedded limestone, dolomite, shale, gypsum, anhydrite, and lenses of sandstone. The unit ranges in thickness from 15 meters (50 ft) near the fringes of the subcrop area to about 122 meters (400 ft) over the central part of the study area (Westjohn and Weaver, 1996b).

The Marshall aquifer is the lowermost aquifer defined by the RASA study. It includes the Marshall Sandstone and sandstones that form the lower part of the Michigan Formation. The basal unit of the Marshall Sandstone consists of 15 to 30 meters (50 to 100 ft) of poorly permeable micaceous sandstone or micaceous siltstone that overlies the Coldwater Shale (Westjohn and Weaver, 1996b). Above this basal unit of the Marshall Sandstone is a permeable, fine- to medium-grained sandstone which is generally 15 to 30 meters (50 to 100 ft) in thickness. This unit is commonly referred to as the lower Marshall Sandstone. However, the lithologic relations of strata that overlie the lower Marshall Sandstone are complex. In most of the study area, a sandstone unit known as either the upper Marshall or Napoleon Sandstone Member, ranges in thickness from 15 to 38 meters (50 to 125 ft). It is hydraulically similar to the lower Marshall Sandstone.

A thick sequence of shale, with local occurrences of limestone, dolomite, and sandstone, constitutes the Coldwater confining unit, and is assumed to form the lower boundary of the RASA aquifer systems. Thickness of the Coldwater ranges from 152 meters (500 ft) in the east to 335 meters (1,100 ft) in the west (Westjohn and Weaver, 1996b).

The Coldwater Shale / Marshall Sandstone contact delimited the RASA study area for the purposes of defining the bedrock aquifers. However, geologic data, including thicknesses of glacial deposits and land-surface elevations, were used to characterize the Glaciofluvial aquifer for the Lower Peninsula. The Coldwater Shale underlies these glacial deposits in most of the region between the study area and the Great Lakes shoreline, and it was therefore assumed that the Coldwater Shale formed a basal no-flow boundary for the Glaciofluvial aquifer. This enabled the modeled area to be extended beyond the study area, to the near-border areas of Michigan with Indiana or Ohio, and to the Great Lakes shorelines.

Generalized Ground-Water Flow System

Difference in hydraulic head caused by topographic relief is the most significant driving force for ground-water flow in the aquifer system. Altitude of the land surface ranges from 526 meters (1,725 ft) in the west-central part to 174 meters (572 ft) at Lake Erie in the southeastern part of the Lower Peninsula. Except for a few areas where bedrock is near land surface, the water table is in the Glaciofluvial aquifer. The water table closely follows trends in the land surface. Figure 2 illustrates the relation between land-surface elevation and water-table elevation. The two main areas where the water table is highest coincide with the Northern and Southern uplands. Low water-table altitudes coincide with the Saginaw, Michigan, and Erie lowlands, as well as with an elongated northeast-southwest depression that trends from Saginaw Bay to Lake Michigan. The depression is located in the proglacial Grand River valley, which is the site of the present day Grand, Maple, Bad, and Saginaw Rivers (fig. 2). Generalized ground-water flow directions in the Glaciofluvial aquifer are toward the lowland areas, the Great Lakes, and the proglacial Grand River.

Head Observations

Predevelopment freshwater head maps for the Saginaw (fig. 4) and Marshall aquifers (fig. 5) were prepared using water levels measured and reported in the late 1800's and early 1900's, and some records from early oil and gas exploration in the 1930's (Barton and others, 1996). The configuration of freshwater heads in these two bedrock aquifers is similar to those of the water table, although the magnitude of the heads are less. Vugrinovich (1986) noted a similar configuration of hydraulic heads. He concluded that predevelopment hydraulic head in both aquifers are generally in equilibrium with the present day land-surface elevations.

Flow Observations

Holtschlag (1996, 1997) determined annual basin discharge rates for 114 unregulated drainage basins, regressing average annual discharge to average annual precipitation and the previous year's average annual discharge. Then, using a 30-year mean precipitation in a steady-state form of the regression relation, he calculated a steady state, normal basin discharge for the 30-year period (1951-1980). The normal basin discharge calculation most accurately reflects pre-development, steady-state discharge.

Saline Ground Water and Brine

A major hydrologic feature of the Michigan Basin aquifer system is the presence of saline water near surface in the lowland areas (Wahrer and others, 1996), and saline water and brine down dip in the Parma-Bayport and Marshall aquifers (Westjohn and Weaver, 1996c; Ging and others, 1996). Saline water also occurs in the Saginaw aquifer in the west-central part of the study area, and saline water and brine occur in the Saginaw Lowlands (Westjohn and Weaver, 1996c; Meissner and others, 1996). Brine and saline water tends to accumulate down dip relative to fresh water, due to the density differences related to higher dissolved-solids concentrations of these fluids. Also, where ground-water density varies, the flow directions are not necessarily perpendicular to head contours because a driving force, caused by water density and the geometry of the aquifer, may alter ground-water flow directions.

Lateral, Upper, and Basal Boundaries

The Great Lakes form lateral specified head boundaries for ground-water flow on three sides of the Lower Peninsula. Surface water and coincident ground-water divides form a no-flow boundary near the southern border of Michigan, separating flow in Indiana and extreme southern Michigan from flow in the study area. Vertically, the water table is the upper flow boundary and the Coldwater confining unit is the lower boundary. The thick shale sequence of the Coldwater Shale is assumed to completely restrict vertical flow. Most ground-water flow in the Glaciofluvial, Saginaw, Parma-Bayport, and Marshall aquifers is horizontal. Flow in the confining units that separate the aquifers is mostly vertical.

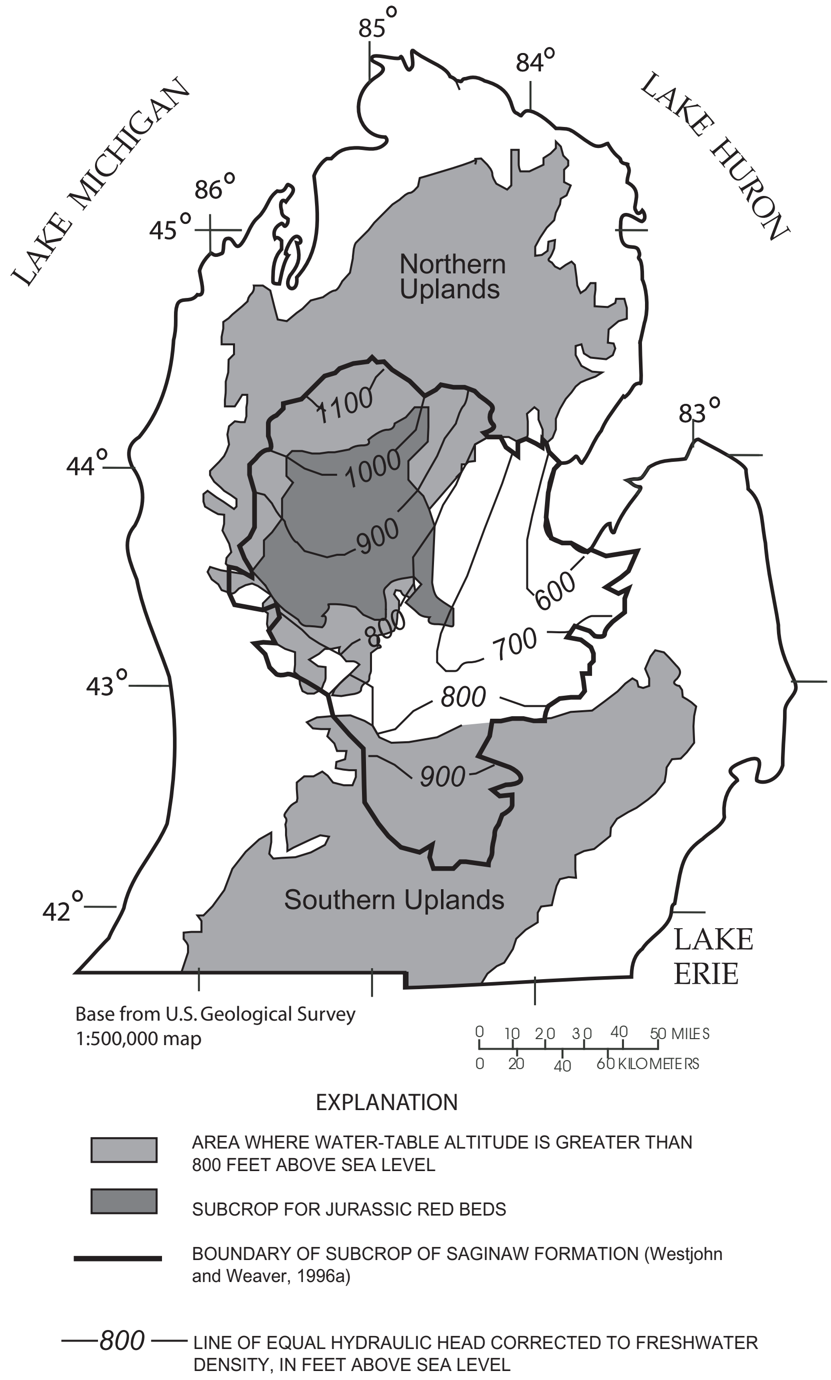


Figure 4. Pre-development freshwater heads in the Saginaw aquifer.

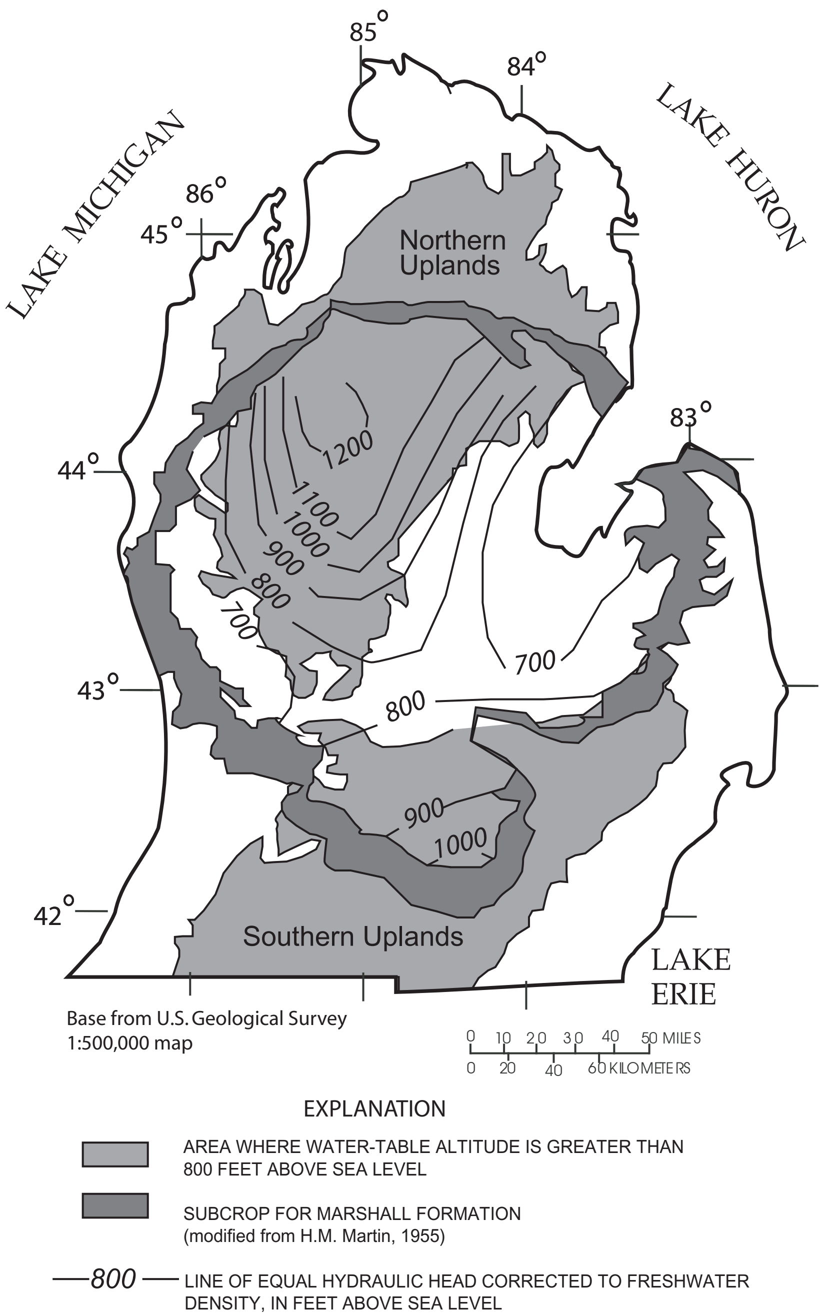


Figure 5. Pre-development freshwater heads in the Marshall aquifer.

SIMULATION OF GROUND-WATER FLOW

The USGS multilayer finite-difference model, commonly known as MODFLOW (McDonald and Harbaugh, 1988), was used to simulate ground-water flow in the aquifer system. The simulation was designed by defining boundary conditions and spatial discretization for the finite-difference calculations; by assigning values for the hydraulic characteristics of the aquifers and confining units; and by estimating recharge rates for the ground-water system. The sources of data were evaluated during simulation design and analysis. The program MODFLOWP (Hill, 1992) was used to conduct sensitivity analysis, a procedure that assesses the effects of changes in assigned hydraulic characteristics on model behavior.

Simplifying Assumptions

Prior to stresses caused by human activities–such as ground-water pumping, natural gas removal, brine withdrawal, coal mining, and agricultural drainage--the aquifer system may have been in a state of equilibrium. At present, pumping and other alterations have changed the aquifer system so that the steady-state assumption may not apply. Therefore, data that duplicate predevelopment measured heads and flow conditions as closely as possible were used to calibrate the model.

MODFLOW is designed to simulate flow of water with constant density, and the results presented herein assume the effect of observed variable density is negligible. This assumption is most reasonable for the Glaciofluvial aquifer, where the observed variable density itself is negligible. Simulation of the Glaciofluvial aquifer comprises most of the results presented herein.

Finite-Difference Grid

The Michigan RASA model grid comprises square finite-difference cells that are 1 kilometer (3,281 feet) on a side. There are 361 columns and 470 rows that correspond to a subset of the 633 column by 733 row Center for Remote Sensing 1-kilometer (CRS1km) data set developed by Michigan State University's Center for Remote Sensing (CRS) to cover the State of Michigan. The bounding rectangle, showing latitudes and longitudes on figure 1, corresponds to edges of the 361 column by 470 row RASA model grid. The Lower Peninsula subset extends to the Great Lakes, with RASA array index (1, 1), the northwestern most model cell, corresponding to CRS1km array index (273, 264). The 361 column by 470 row grid is flush with the ERDAS lower right corner, and shares its transverse mercator (TM) metric coordinate system (Lusch and Enslin, 1984; table 1).

Digital Representation of Model Layers

Figure 3 shows a cross section of the geologic surfaces used in the model, reconstructed as model layer surfaces along a west to east transect (model row 240). Geological structure contour maps and isopach maps were digitally reproduced from the original hydrogeological reports for the Glaciofluvial, Saginaw, Parma-Bayport, and Marshall aquifers, and the confining units that separate them (Westjohn and others, 1994; Westjohn and Weaver, 1996a; and Westjohn and Weaver, 1996b). The surfaces were reproduced using the computer program WANGRID.FOR (see diskette), which uses the model grid and Gauss Seidel iteration to solve a finite-difference approximation of the Laplace equation between specified-value (Dirichlet) and specified-rate (Neuman) boundaries (Wang and Anderson, 1982). Structure contour maps were reproduced by specifying the edges (boundary) of the subcrop as Neuman boundaries (specified rate = 0), and the digitized contour lines as Dirichlet boundaries

Table 1. Grid coordinates, basis for coordinates, and digitizer origin of the RASA model relative to those for the Michigan State University Center for Remote Sensing 1 kilometer transverse Mercator (CRS1k) coordinate system

RASA Model Grid Coordinates, Basis, and Digitizer Origin	CRS1k[a] Easting (meters)	CRS1k[a] Northing (meters)	Latitude (degrees)	Longitude (degrees)
CRS1k[a] Basis	359987	344917	44.0000	86.0000
Centroid Row 1, Column 1	290568	557128	45.9068	86.8950
Upper Left Corner, Row 1, Column 1	290068	557628	45.9113	86.9015
Upper Right Corner, Row 1 Column 361	651068	557628	45.8533	82.2508
Lower Left Corner, Row 470, Column 1	290068	087628	41.6799	86.8400
Lower Right Corner, Row 470, Column 361	651068	087628	41.6298	82.5058
Digitizer Origin	276662	079939	41.6094	87.0000

[a] Center for Remote Sensing, Michigan State University, transverse mercator, 1 kilometer

(specified value = contour value), meeting the Neuman boundaries at right angles. Isopach maps were digitially reproduced specifying Dirichlet boundaries for the subcrop (specified as a zero thickness boundary), and at the contour lines (specified value = contour value). The finite-difference approximation to the LaPlace equation thus provides a linear interpolation between the contour lines. The linear interpolation was greatly improved by specifying cells with original data in addition to the digitized contour lines and subcrops. Specified values of hand-drawn contour lines and data points were assigned to nearest model grid nodes. Unassigned nodes were thus linearly interpolated between assigned nodes.

The top of bedrock represents a major angular unconformity (fig. 3) to the hydrogeologic flow system, the base of the Glaciofluvial aquifer. This surface was defined for the entire Lower Peninsula of Michigan by merging separate bedrock top maps for the region within RASA and the region beyond the RASA study area. Within the RASA study area, the altitude of bedrock top for each model cell was linearly interpolated using the map and data from Westjohn and others (1994), corresponding locally to the top of the Saginaw aquifer, or the top of Jurassic "red beds" confining unit. Beyond the RASA study area, bedrock elevations, corresponding to a basal no-flow boundary, were constructed by subtracting glacial thicknesses from land-surface elevations. Both the glacial thickness and land-surface elevation contour sets were provided from the CRS1km database (Lusch and Enslin, 1984) . Outside the RASA study area, the bedrock top map is highly correlated to the land elevation map from which drift thickness was subtracted. No alternative bedrock top map was available at the time of model construction for the region outside of the RASA study area. The combined bedrock top map forms the base of the Glaciofluvial aquifer, the base of model layer 1 (diskette: GLACBOT.DAT).

The top of the Saginaw aquifer (top of model layer 2; diskette: PENNBEDT.DAT), the top of the Parma-Bayport aquifer (top of model layer 3; diskette: PARMBEDT.DAT), and the thickness of the Saginaw confining unit (diskette: SGCFTHCK.WNG) were interpolated by WANGRID.FOR using data and contours from Westjohn and Weaver (1996a). The Saginaw confining unit thicknesses were added to the top of the Parma-Bayport aquifer in the subcrop region of the Saginaw aquifer to form the base of model layer 2 (diskette: PRSGBEDT.DAT).

The top of the Michigan confining unit (diskette: MICHBEDT.DAT), the top of the Marshall aquifer (diskette: MARSBEDT.DAT), and the top of the Coldwater confining unit (diskette: COLDBEDT.DAT) were interpolated by WANGRID.FOR using data and contours from Westjohn and Weaver (1996b). The top of the Michigan confining unit forms the base of the Parma-Bayport aquifer, which is the base of model layer 3. The top of the Marshall aquifer forms the top of model layer 4 in the sub-region of the Michigan confining unit. In the Marshall subcrop area, the bedrock surface was used as the top of model layer 4, because the bedrock surface is the top of the Marshall aquifer in this region. The top of the Coldwater confining unit forms the base of the Marshall aquifer in the sub-region of the Marshall aquifer (the RASA study area), which is the base of model layer 4.

Digital Representation of Model Layers under Saginaw Bay

To complete the aquifer model under Saginaw Bay, where no data were available, the layers were completed by 1) interpreting contacts for each aquifer and confining unit drawn across the bay; 2) interpolating isopachs for each aquifer and confining unit under the bay, between the existing top maps exterior to Saginaw Bay, and the zero-thickness contacts interpreted across the bay; and 3) reconstructing the surface of each layer by subtracting the isopach from the overlying reconstructed surface, starting with 162 meters (530 ft) above sea level for the reconstructed base of glacial deposits beneath the bay. The reconstructions were therefore accomplished by linear interpolation and linear projection, following a conservation of mass argument and an assumed base of drift angular unconformity of 162 meters (530 ft) MSL below Saginaw Bay. An anticlinal structure on the Marshall aquifer resulted from the reconstruction (fig. 3). Though arguably a by-product of the method chosen, anticlinal structures under Saginaw Bay are also interpreted on the oil and gas chart of Cohee et. al. (1951), and Lane and Hubbard (1895, pl. 68, nos. 1 and 2).

Adjustments to Interpolated Layers

Because each top and bottom surface was contoured and interpolated independently, the hydrogeologic units were ill-defined in some regions where the bottom of a layer was defined higher than its top. In these regions, the aquifer was assumed to pinch out, and the bottom was redefined 1 foot lower than the overlying top using TOPDOWN.FOR (see diskette). TOPDOWN.FOR starts from the topmost layer and works through the model downward, thus propagating any consecutive errors downward.

The Saginaw and Parma-Bayport aquifers, which correspond to model layers 2 and 3 in the model, pinch out interior to the Marshall aquifer subcrop (fig. 3). Therefore, model layer 1 must communicate directly with model layer 4 in the region between the Saginaw and Marshall subcrops. A program was developed, ANGUNCON.FOR (see diskette), to define layers 2 and 3 as infinitesimal layers between layers 1 and 4 in this region. The infinitesimal layers in the subcrop region were given the same hydraulic properties, both vertically and horizontally, as the overlying glacial deposits to simulate the direct hydraulic connection between model layers 1 and 4.

Model Hydrologic Boundary Conditions

Lateral boundaries for the model include specified heads and no-flow conditions for the Glaciofluvial aquifer (model layer 1), and no-flow conditions for the bedrock aquifers (model layers 2, 3, and 4) (fig. 1). The Glaciofluvial aquifer (layer 1) is bounded by Lake Michigan to the west and northwest, Lake Huron to the east and northeast, and Lake Erie to the southeast. Together with the connecting channels of the St. Clair and Detroit Rivers and Lake St. Clair, which bound part of the model area to the east, the surface-water bodies form a continuous specified head boundary. The boundary was assigned specified heads of 176.8 meters (580 ft) for lakes Michigan and Huron, linearly interpolated heads between 176.8 and 175.3 meters (580 and 575 ft) for the St. Clair River, specified heads of 175.3 meters (575 ft) for Lake St. Clair, linearly interpolated heads between 175.3 and 174.3 meters (575 and 572 ft) for the Detroit River, and specified heads of 174.3 meters (572 ft) for Lake Erie. In the Saginaw Bay area and for a small portion of Lake Michigan that is underlain by the Marshall aquifer, the Glaciofluvial aquifer was simulated as part of the specified-head boundary in order to simulate vertical leakance between the Glaciofluvial aquifer and the underlying bedrock aquifers. The southern boundary of the Glaciofluvial aquifer consisted of drainage divides forming a continuous no-flow boundary. It was assumed that the surface drainage divides coincide with the ground-water divide in this area.

Layers 2, 3, and 4 are bound by the Coldwater-confining-unit/Marshall-aquifer contact (Marshall subcrop), which forms an aquifer pinchout, no-flow boundary for layer 4 (fig. 1). Model layers 2 and 3 also extend to the Marshall subcrop, but are effectively bound by the Michigan confining unit/Saginaw aquifer contact (Pennsylvanian subcrop), which forms an aquifer pinchout, no flow boundary shared by layers 2 and 3. Between the Pennsylvanian and

Marshall subcrops, the thickness of layers 2 and 3 are infinitesimal.

Stream and Lake Elevations

Internal boundaries that represent the major streams and natural lakes in the modeled area were simulated using the RIVER module of MODFLOW (fig. 6). River reaches in the model were identified from the Center for Remote Sensing's ERDAS data set (Lusch and Enslin, 1984). The data set contains pixelated drainages corresponding to the Michigan Hydrologic Unit map (U.S. Geological Survey, 1974). The river data set was modified to minimize the number of pixels needed to define the continuous drainage channel, mainly by reducing the number of pixels used to define stream confluences. Some pixels were eliminated at stream headlands to separate stream reaches.

River stages were set by gridding the water-table surface and taking the values of the grid at river reach locations. The water table was gridded by linearly interpolating the hand-drawn water table contours (Mandle and Westjohn, 1989), specified heads at boundaries, and river and lake elevations using WANGRID.FOR (diskette: GLACHEAD.DAT). Data for the water table interpolation included 1,220 river crossings (diskette: RIVERELV.OUT) and 609 lake-level observations (diskette: LAKEELVS.OUT).

River widths were set by applying the program CANOE.FOR (see diskette), which orders the stream segments on the basis of Horton stream ordering (Horton, 1945) and puts the segments and reaches in downstream order. River stages were reset by linearly interpolating the stages from headland to confluence, confluence to confluence, and confluence to mouth. Uphill or flat segments were corrected to within 1.5 meters (5 ft) by reading the correct stages from 7.5- and 15- minute topographic maps and reprocessing the river file with CANOE.FOR. Up to five stream orders resulted from the processing. The Grand River, a fifth order stream, was followed along its length to determine river widths corresponding to the different stream orders. River widths at order-starting confluences along the Grand River were obtained from gaging station discharge records, and were used to set river widths for each respective stream order for the Lower Peninsula. The widths are 1 m (3.28 ft), 2 m (6.56 ft), 20 m (65.6 ft), 60 m (197 ft), and 80 m (262 ft) for first through fifth order streams, respectively. It was assumed that all streambeds had a thickness of 0.3048 m (1 ft) and a vertical hydraulic conductivity of 1×10^{-4} cm/s (0.28 ft/d). The river data constitute the major part of the river package input file, GLACRIV2.DAT (see diskette).

Horizontal and Vertical Hydraulic Conductivities

Horizontal and vertical hydraulic conductivities for the aquifer (coarse-textured) and non-aquifer (fine-textured) parts of each layer were used in conjunction with the percentages of each part to total layer thickness of aquifer and non-aquifer fractions to construct effective model layer conductivities. The effective vertical hydraulic conductivities in each model layer, and additional estimates of the vertical hydraulic conductivities of intervening confining units, were used to construct VCONTS (see below). Aquifer and non-aquifer parts of model layers were determined from separate aquifer isopach maps from the reports of Westjohn and Weaver (1996a,b) and Westjohn and others (1994), as described below. The effective hydraulic conductivities used in the model thus represent values for the entire thickness of the layer, and will generally be smaller than values determined for specific parts of aquifers by field methods such as aquifer tests. Equation 52 of the MODFLOW documentation (McDonald and Harbaugh, 1988, p. 5-16) was implemented to construct effective VCONTS, using vertical

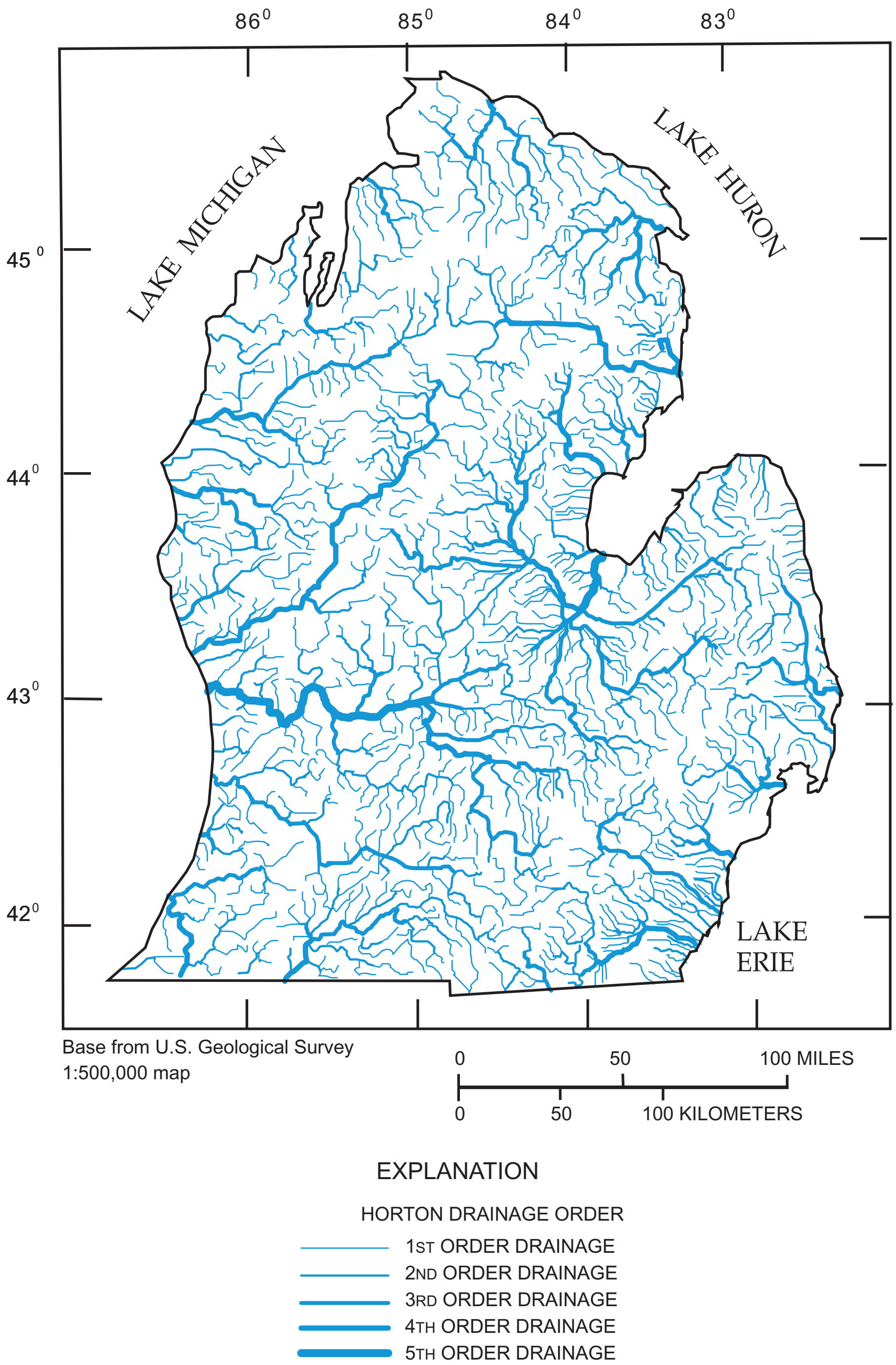


Figure 6. Location of nodes used to simulate streams and lakes.

hydraulic conductivity estimates for the coarse- and fine-textured parts of the aquifer layers; additional estimates of the vertical hydraulic conductivities of intervening confining units; and structural contour top and bottom grids. The bedrock aquifer estimates, summarized in table 2, were based on a study by Westjohn and others (1990) for the RASA project. The Glaciofluvial conductivity estimates, summarized in table 2, were based on a range of aquifer and confining unit pump test and modelling studies, the results of which are discussed in Mandle and Westjohn (1989).

Hydraulic conductivities for each simulated layer were summarized into model grids for the four-layer model by 1) using or determining percents (x%) of aquifer thickness (coarse fraction composite thicknesses) to total thickness (structural thicknesses that include coarse and fine fraction thicknesses); 2) selecting corresponding representative hydraulic conductivity estimates for each aquifer for both the coarse- and fine-textured fraction in both the horizontal and vertical direction (table 2); and 3) calculating effective vertical (in series) and horizontal (in parallel) hydraulic conductivities for the layer, using equations from Freeze and Cherry (1979). The percent aquifer thickness to total thickness (x%) for the Glaciofluvial aquifer was interpolated (diskette: GLCPRCNT.WNG) using WANGRID.FOR from a map provided by Westjohn and Weaver (1994, fig. 5, p. 8), with contours extended beyond the RASA study area boundary of the map. Geologic interpretation, based on major glacial ice-marginal positions, was used to extend the contours of percent aquifer (percent coarse fraction) in order to cover the entire Lower Peninsula model area. For the bedrock aquifers, the percent aquifer thickness to total thickness (x%) was determined by 1) subtracting aquifer bottom from aquifer top for total thickness; 2) digitally interpolating separate contours of aquifer (coarse-textured) thickness for each model node (diskette: PSNDWELL.WNG, PARMTHCK.WNG, MSSAQUIF.WNG) from maps provided by Westjohn and Weaver (1996a, fig. 6, p. 13 and fig. 8, p. 17;1996b, fig. 7, p. 14) respectively; and 3) calculating the coarse fraction's percent of total thickness. Effective layer hydraulic conductivities were calculated in the program PRMPRCNT.FOR (see diskette) with equations modified from Freeze and Cherry (1979), using the aquifer hydraulic conductivity estimates and respective thickness estimates for both the coarse (x%) and the fine (100% – x%) fraction. The diskette provides data files of the resulting effective horizontal (diskette: GLACPERM.KH, PENNPERM.DAT, PARMPERM.DAT, MARSPERM.DAT) and effective vertical (diskette: GLACPERM.KV, PENNPERM.KV, PARMPERM.KV, MARSPERM.KV) hydraulic conductivities, varying at each node primarily due to the varying percentages of aquifer thickness to total thickness. Thus, the hydraulic characteristics used in the model for each layer represent values for the entire thickness of the layer, and will generally be smaller than values determined for specific parts of aquifers by field methods such as aquifer tests.

Vertical Leakance

MODFLOW input requires calculation of vertical leakances outside of the model. The calculation requires structure contour grids of the top and bottom of adjacent layers as well as vertical hydraulic conductivities of aquifer and intervening confining units. On the basis of vertical hydraulic conductivities for the aquifers, calculated as discussed above, and additional estimates for the confining units (table 2), vertical leakances between layers were calculated with the program VCONT.FOR (see diskette). The program VCONT.FOR calculates a leakance value for each active node by implementing equation 52 of the MODFLOW documentation (McDonald and Harbaugh, 1988, p. 5-16). The diskette contains data files for the VCONTs between each layer (diskette: VCONT12.VCT, VCONT23.VCT, VCONT34.VCT).

Table 2. Summary of horizontal and vertical hydraulic conductivities for the aquifers and confining units used to construct model conductivities and VCONTs

Aquifer or Confining unit	Texture	Horizontal hydraulic conductivity, in feet per day	Vertical hydraulic conductivity, in feet per day	Horizontal hydraulic conductivity, in centimeters per second	Vertical hydraulic conductivity, in centimeters per second
Glaciofluvial aquifer	Coarse grained	50	5	1.75×10^{-2}	1.75×10^{-3}
Fine-grained till or lacustrine deposits in Glaciofluvial aquifer	Fine grained	2.83×10^{-4}	2.83×10^{-4}	1×10^{-7}	1×10^{-7}
Glacial till-- J_R "red beds" confining unit	--	--	2.83×10^{-4}	--	1×10^{-7}
Saginaw aquifer	Coarse grained	2.83	2.83	1×10^{-3}	1×10^{-3}
Intercalated fine-grained rock in Saginaw aquifer	Fine grained	2.83×10^{-4}	2.83×10^{-6}	1×10^{-7}	1×10^{-9}
Saginaw confining unit	--	--	2.83×10^{-4}	--	1×10^{-7}
Parma-Bayport aquifer	--	7.09	1.13×10^{-3}	2.5×10^{-3}	4×10^{-7}
Michigan confining unit	--	--	2.83×10^{-7}	--	1×10^{-10}
Marshall aquifer	Coarse grained	1.42×10^{-1}	1.42×10^{-1}	5×10^{-5}	5×10^{-5}
Intercalated fine-grained rock in Marshall aquifer	Fine grained	1.42×10^{-3}	1.42×10^{-3}	5×10^{-7}	5×10^{-7}

Ground-Water Recharge

Ground-water recharge rates were estimated in an independent study for the RASA modeled area by Holtschlag (1996; 1997). The estimates were made by analysis of streamflow data from 114 basins throughout the Lower Peninsula of Michigan.

Ground-water runoff (baseflow) was determined from hydrograph separation of stream discharge measurements, and annual ground-water discharge (equal to annual basin recharge) for each basin was calculated with QPART (Rutledge, 1993). Steady state, normal basin recharge was then related to 30-year normal precipitation by a set of basin specific regression equations. To interpolate the normal basin recharge estimates statewide, a second regression relation was developed, relating the normal basin recharge estimates to 1) latitude and longitude (as a proxy for climate), 2) fine- and coarse-textured surficial geologic material, and 3) deciduous versus coniferous forest cover in each basin. The relation was used to determine a recharge rate for each active node of model layer 1. The recharge rates range from 5 to 566 mm/yr (0.19 to 22.3 in./yr) and average 214 mm/yr (8.41 in./yr) (Holtschlag, 1996; 1997). The recharge values were entered into a file, GLACRCH7.DAT, called from the recharge input file, RASA1A18.RCH (see diskette).

The recharge rates used in the model were extrapolated from selected stream baseflow measurements related to basin characteristics and precipitation. This method assumes that all recharge water of the selected basins discharges at the stream as baseflow without losses to, or gains from, deep seepage. The extrapolated recharge rates of the model may be similarly underestimated or overestimated. Therefore, the deep seepage [also referred to as underflow] component of flow in the model may not be properly accounted for in the model budget. Generally, model recharge would be underestimated in the regional recharge areas due to unaccounted for deep seepage losses, and overestimated in regional discharge areas due to unaccounted for deep seepage gains.

Model Parameterization

For the purpose of estimating model sensitivity, nine input multipliers were used as unit parameters in a separate MODFLOWP model constructed to exactly reproduce the MODFLOW model. The sensitivity equation method was used to calculate the sensitivity matrix (Hill, 1992, p. 90 - 94), which can be used to 1) estimate parameters using Gauss-Newton optimization, 2) calculate composite scaled sensitivities for a set of parameters to evaluate relative model parameter sensitivity, and/or 3) calculate individual confidence intervals from the covariance matrix on the parameters. The nine parameters chosen include unit multipliers for the transmissivity arrays of the Glaciofluvial (T1), Saginaw (T2), and Marshall (T4) aquifers, the VCONT array between the Glaciofluvial and Saginaw aquifers (KV1), and the stream conductances for each “n” of five Horton stream orders (KRBn). The MODFLOWP model was used to calculate individual confidence intervals for simulated values, which are compared to error bars for observations; both are shown on plots of simulated versus measured heads and flows in figures 11 and 12.

RESULTS OF SIMULATIONS

Potentiometric Surfaces and Hydraulic Head Calibration and Sensitivity

Simulated regional flow is most affected by, and therefore most readily summarized by, the solution of hydraulic head for the Glaciofluvial aquifer (layer 1) shown in figure 7, and provided on the diskette as file GLACHEAD.DAT. The hydraulic head solution simulates the water table elevation, given the unconfined Dupuit assumptions used by the model, and is highest in the northern and southern upland areas coincident with the water table map of Mandle and Westjohn (1989). The output compares well with the regional distribution of head shown in figure 2 (Mandle and Westjohn, 1989; Barton and others, 1996), which is based on stream crossing and lake elevation data. Figure 10a shows that simulated heads are in general agreement with measured heads for the Glaciofluvial aquifer, layer 1 (Mandle and Westjohn, 1989; data not originally reported, but provided on this report's diskette: see below). The measured head values correspond to 499 (see diskette: HEADCONF.XLS) of 609 (see diskette: LAKEELVS.OUT) lake elevations digitized from the 1:500,000 hydrologic basemap. The remaining 110 of 609 lake data points, as well as the 1,220 river data points, were located in a model river cell, and thus were eliminated to reduce internal boundary bias in the calibration. Data used to construct a water table grid, which in turn was used to set the stages in river cells (the internal boundaries), included the complete 609 lake and 1,220 river data points. Individual confidence intervals calculated from YCINT.FOR (Hill, 1994) from the 9-parameter MODFLOWP simulation are shown as 2σ (95-percent confidence) vertical bars about each point. Assumed measurement errors are shown as 2σ (95-percent confidence) horizontal bars about each point, taken from the variance used in determining weights in MODFLOWP.

The solution of hydraulic head for the Glaciofluvial aquifer (layer 1; fig. 7) is strongly imprinted on the solution of hydraulic head for the Saginaw aquifer (layer 2) shown in figure 8, and provided on the diskette as file PENNHEAD.DAT. The regions of high and low head in the Saginaw aquifer are coincident with those in the Glaciofluvial aquifer, and head is only moderately dampened (lowered) by the overlying glacial-till / Jurassic-"red-beds" confining unit. The output compares well with the regional distribution of head shown in figure 4 (Barton and others, 1996) based on well data. Figure 10b shows that simulated heads are in general agreement with measured heads for layer 2, the Saginaw aquifer. The measured head values correspond to 57 data points from Barton and others (1996; data not originally reported, but provided on this report's diskette: PENNHEAD.OUT). Individual confidence intervals calculated from YCINT.FOR (Hill, 1994) from the 9-parameter MODFLOWP simulation are shown as 2σ (95-percent confidence) vertical bars. Assumed measurement errors are shown as 2σ (95-percent confidence) horizontal bars about each point, taken from the variance used in determining weights in MODFLOWP.

The solution of hydraulic head for the Saginaw aquifer (layer 2; fig. 8) is strongly imprinted on the solution of hydraulic head for the Parma–Bayport aquifer (layer 3; see diskette: PARMHEAD.DAT). The regions of high and low head in the Parma-Bayport aquifer are coincident with those in the Saginaw aquifer, and head is only moderately dampened (lowered) by the overlying Saginaw confining unit. Data for calibration was not available for the Parma–Bayport aquifer.

The solution of hydraulic head for the Marshall aquifer (layer 4), shown in figure 9 and provided on the diskette as file MARSHEAD.DAT, is considerably dampened, though the high and low areas are still controlled by the solution in the Glaciofluvial aquifer (layer 1). The dampening of the solution of hydraulic head in the Marshall aquifer (layer 4) is due to the low vertical hydraulic conductivity of the overlying Michigan confining unit (low VCONT between layers 3 and 4). The Michigan confining unit compartmentalizes the Michigan basin in the study area (Westjohn and Weaver, 1996b), and its effects can be seen in the solution for layer 4. The output compares well with the regional distribution of head shown in figure 5 (Barton and others, 1996) based on well data, with the exception of an area of high heads (>1,100 ft) in the northwest on the observed map (fig. 5) that

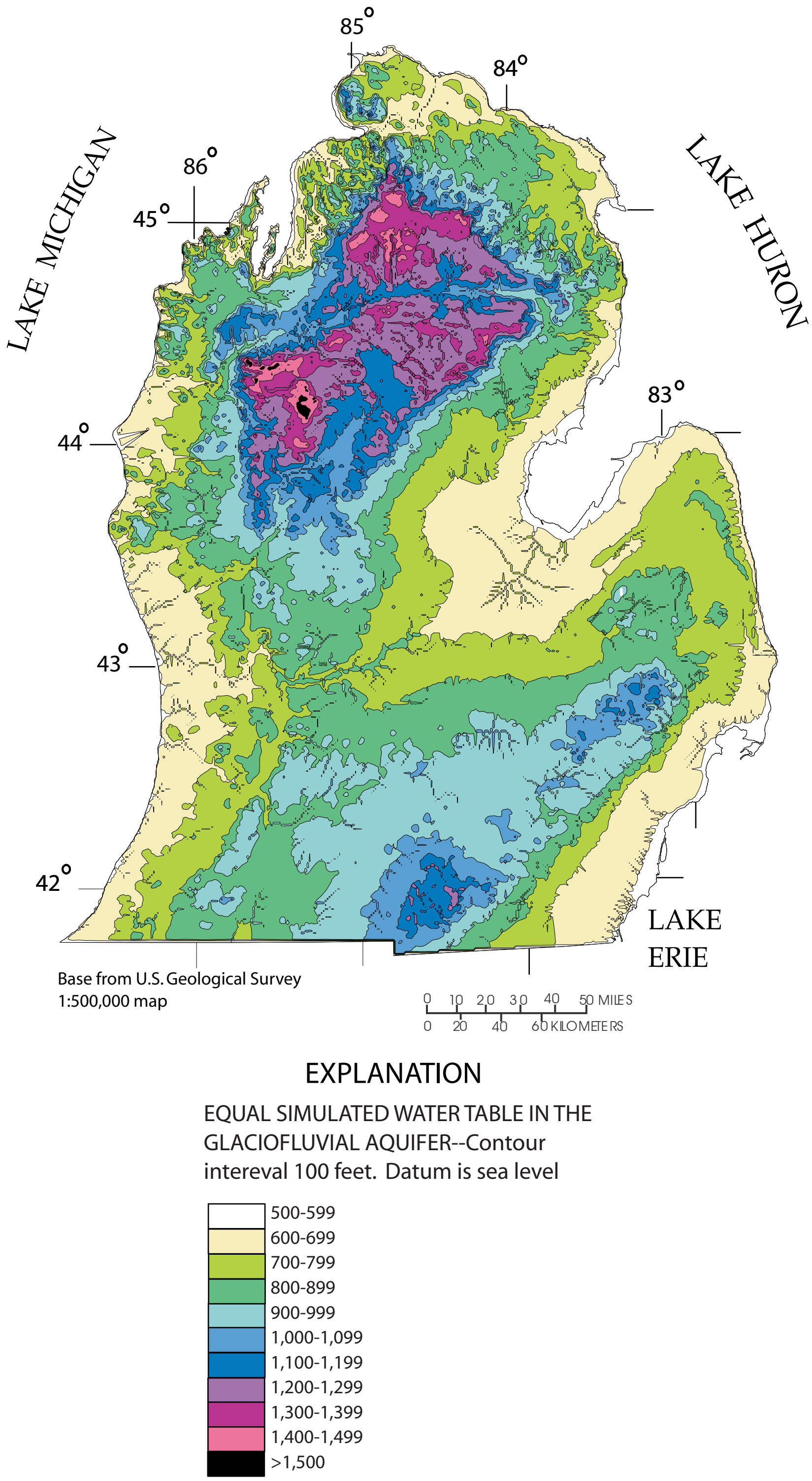


Figure 7. Simulated hydraulic head (Dupuit water table) for the glacial aquifer, model layer 1.

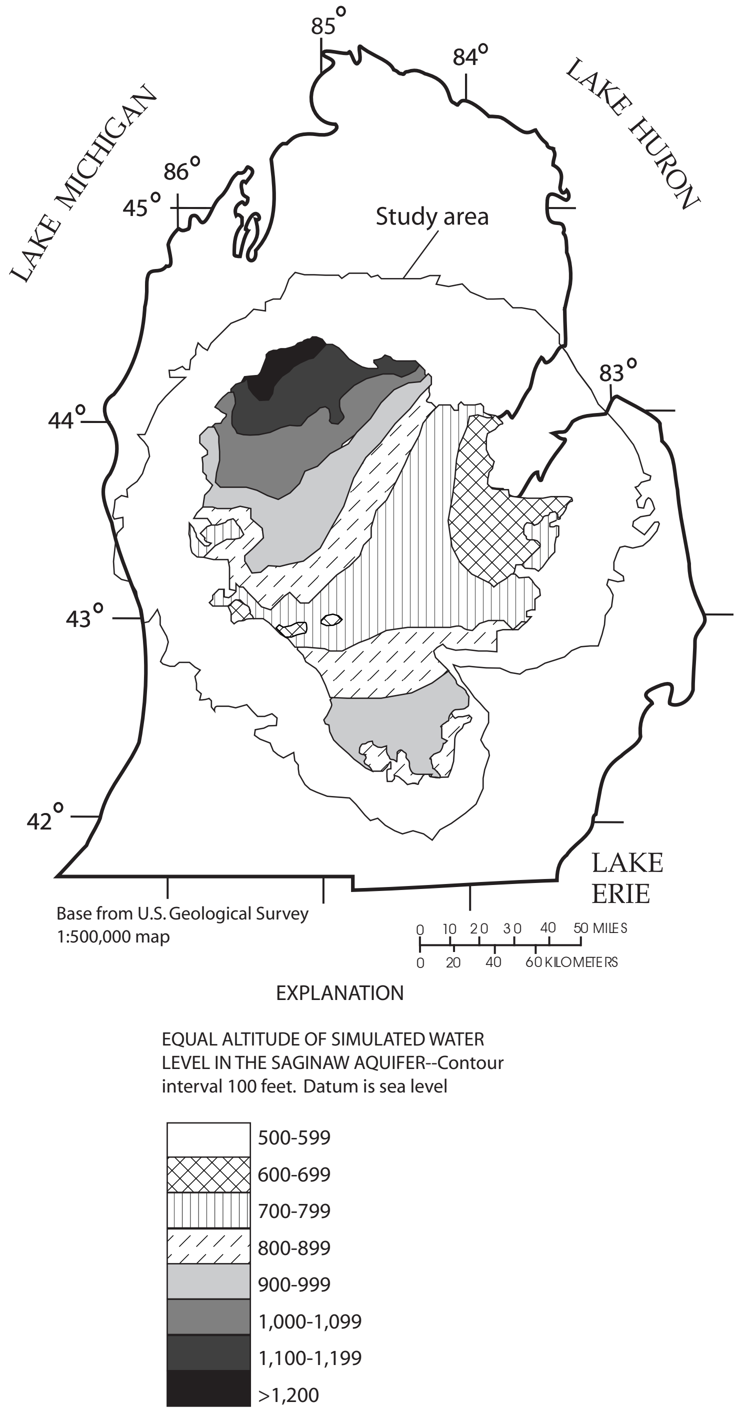


Figure 8. Simulated hydraulic head for the Saginaw aquifer, model layer 2.

are not reproduced on the simulated map (fig. 9). Figure 10 shows that simulated heads are generally in good agreement with 75 measured heads for layer 4, the Marshall aquifer (Barton and others, 1996; data not originally reported, but provided on this report's diskette: MARSHEAD.OUT), except for some of the higher (>1,100 ft) observed head values for which the simulated values are too low. The region of higher head values corresponds to the high water table in the unconfined Glaciofluvial aquifer of the northern uplands. The confined Saginaw aquifer shows lower head values, however, than both the overlying unconfined Glaciofluvial aquifer and the underlying confined Marshall aquifer, suggesting that the higher heads in the Marshall aquifer must be transmitted laterally from a Marshall subcrop source, updip and in direct hydraulic connection with the water table. The lateral transmission of heads up- or down-dip is significantly influenced by brine in the Marshall. Proper simulation of the high head values in the Marshall aquifer, which would result in a better calibration, would require simulation of variable density, which is beyond the capabilities of MODFLOW and the project scope.

Individual confidence intervals calculated from YCINT.FOR (Hill, 1994) from the 9-parameter MODFLOWP simulation are shown as 2σ (95-percent confidence) vertical bars. The confidence intervals for the higher Marshall heads are notably larger than those for lower values due to the larger residuals translated into a larger parameter covariance for these uncalibrated observations. Assumed measurement errors are shown as 2σ (95-percent confidence) horizontal bars about each point, taken from the variance used in determining weights in MODFLOWP.

The simulated heads on all three calibration plots (figs. 10 a,b,c) generally fall along the respective lines of agreement from low to high head. A deviation from perfect agreement (slope = 1.0) can be observed in linear best-fit lines for the Glaciofluvial (slope = 0.88), Saginaw (slope = 0.94), and Marshall (slope = 0.61) aquifers. These slopes less than 1.0, together with positive intercepts, on all three calibration plots cause the best-fit lines to cross the agreement lines. This may be related to the distribution of ground-water recharge (see the earlier discussion on Ground-water Recharge). An underestimation of recharge in the regional recharge areas, due to unaccounted for deep seepage (underflow) losses, together with an overestimation of recharge in regional discharge areas, due to unaccounted for deep seepage gains, would result in the observed trends where low heads are too high and high heads are too low.

Stream Baseflow Calibration and Sensitivity

In addition to head calibration, the model was also calibrated against stream discharge measurements. Adjacent, downstream basins within the 114 discharge data set of Holtschlag (1996) were assumed to be cumulative such that the downstream discharge measurement site was representative of the combined basins. Combining basins reduced the data set to 72 downstream measurement sites that were assumed to represent regional discharge. These 72 combined, unique basins were identified in ZONEBUDGET (Harbaugh, 1990) to compare modeled versus measured discharge. The 72 measured stream discharge values (diskette: BASORT.XLS) plot along a line of agreement on a graph (fig. 11) of simulated versus measured stream discharge for river basins ranging from less than 0.28 m^3/s (10 ft^3/s) up to 28 m^3/s (1000 ft^3/s). For the purpose of calibrating to the regional scale of the model and calculating confidence intervals on the flows, 44 basin zones were delineated (diskette: BASIN361.BOL) from which 22 large basins were identified and the discharge data and corresponding collective river cells entered into the MODFLOWP flow observation data (set 7). The 22 measured stream discharge data (diskette: FLOWCONF2.XLS) plot along a line of agreement on a graph (fig. 12) of simulated

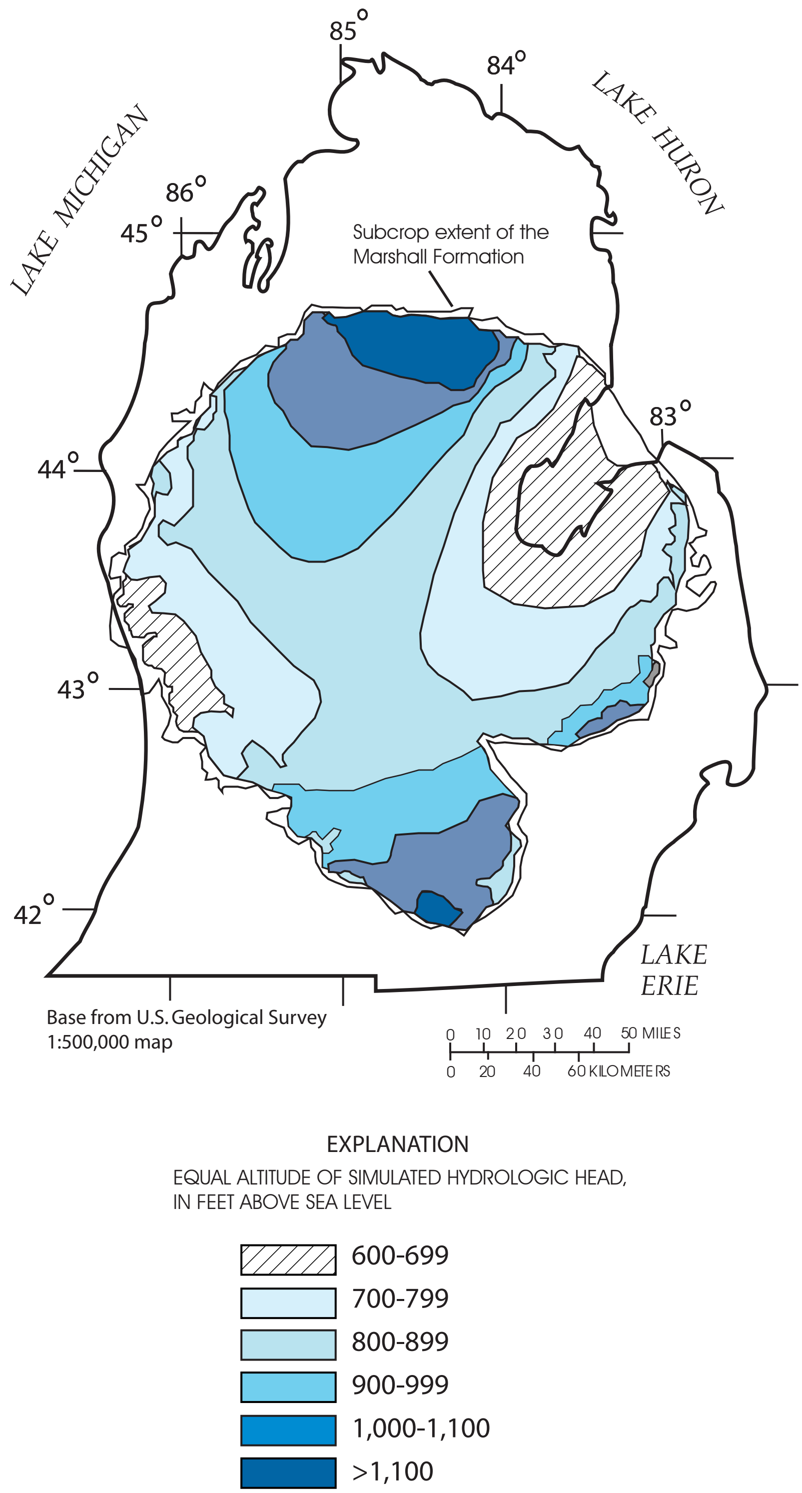


Figure 9. Simulated hydraulic head for the Marshall aquifer, model layer 4.

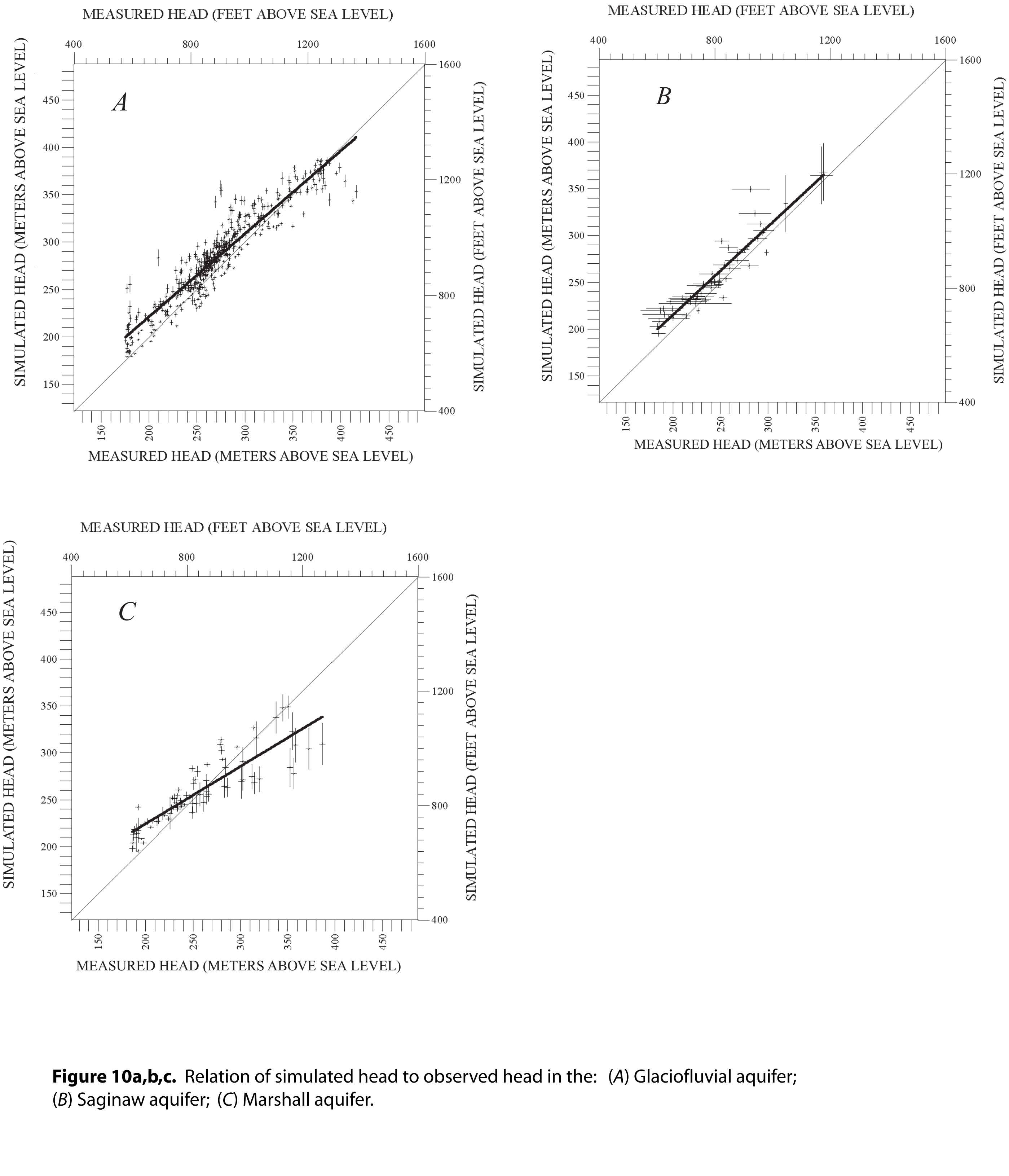


Figure 10a,b,c. Relation of simulated head to observed head in the: (*A*) Glaciofluvial aquifer; (*B*) Saginaw aquifer; (*C*) Marshall aquifer.

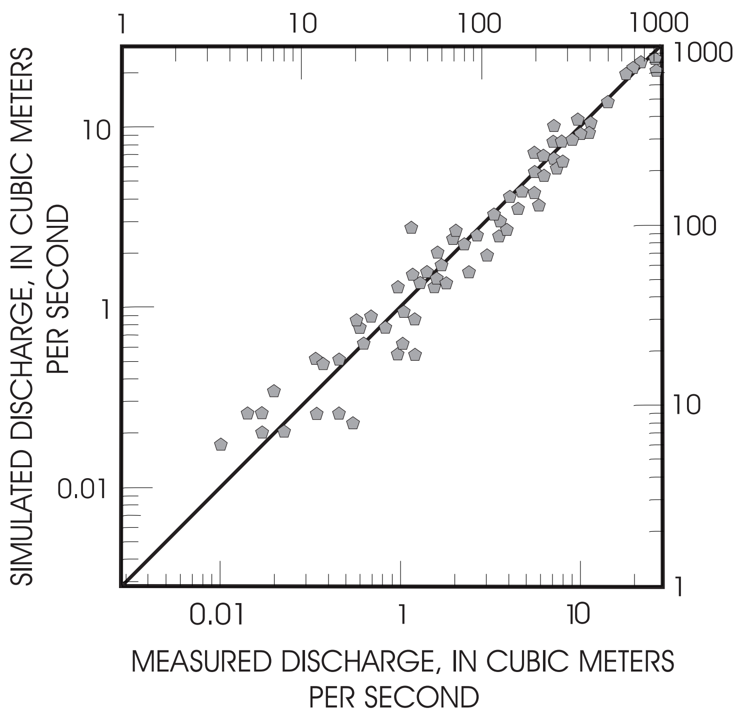


Figure 11. Relation of simulated stream discharge to observed stream discharge for 72 gaging stations.

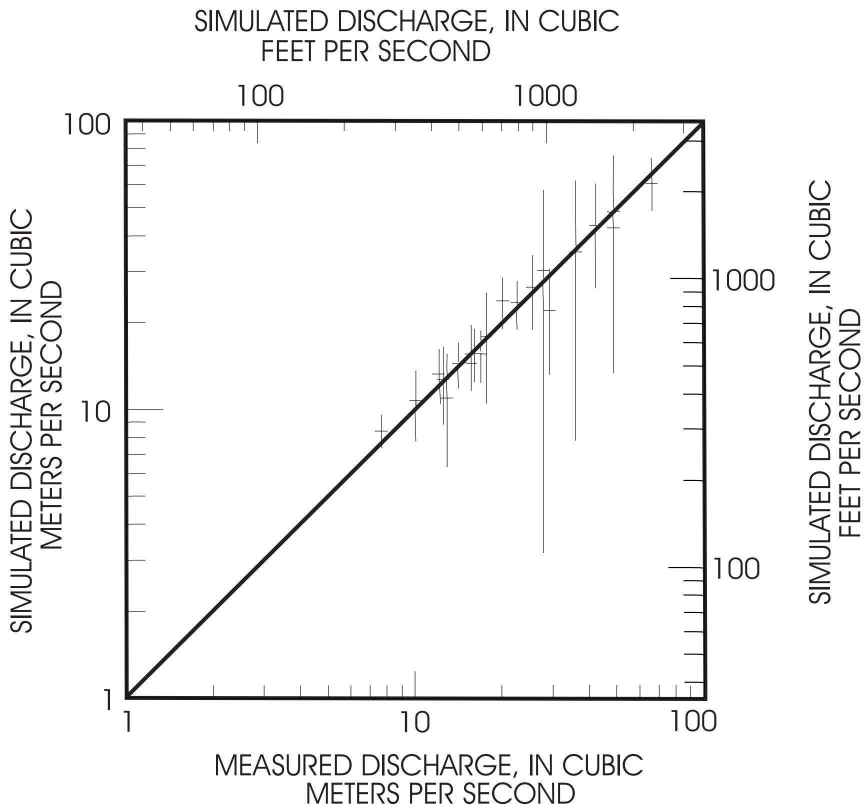


Figure 12. Relation of simulated stream discharge for 22 large basin gaging stations. Diagonal line on graph is line of perfect agreement. Vertical bars represent 95 percent individual confidence intervals calculated from the 9 parameter MODFLOWP simulation. Horizontal bars represent assumed, 95 percent confidence, measurement error (5 percent of flow) bars, taken from the variance used in determining weights in MODFLOWP.

verses measured stream discharge. Individual confidence intervals calculated from YCINT.FOR (Hill, 1994) from the 9-parameter MODFLOWP simulation are shown as 2σ (95 percent confidence) vertical bars. The twenty-two 95-percent confidence intervals averaged 38 percent of their respective flows. Assumed measurement errors are shown as 2σ (95-percent confidence) horizontal bars about each point, taken from the variance used in determining weights in MODFLOWP.

Relative Model Parameter Sensitivity

The composite scaled sensitivity matrix calculated by MODFLOWP yields a dimensionless and scaled measure of model sensitivity for each parameter. Composite scaled sensitivities can thus be used to compare relative model sensitivity between parameters. Figure 13 is a bar graph of composite scaled sensitivities for each of the nine parameters chosen for the sensitivity analysis. Though model sensitivity is at least partly controlled by data quantity, quality, and distribution, some conclusions about the effect of model setup and parameterization on sensitivity can be drawn. The parameter with the largest sensitivity coefficient is the multiplier of layer 1 transmissivity, T1. followed by the multiplier of the Horton first-order stream cells' river bed conductivities, KRB1. The layer 1 transmissivity controls the bulk of horizontal ground-water flow, while the KRB1 controls most of the outflow in the model; therefore this result was expected. The multipliers of the transmissivities in the bedrock, T2 and T4, as well as the multiplier of KV1, which controls communication between bedrock and the overlying drift, have the smallest sensitivity coefficients, indicating that model output is relatively insensitive to these parameters. The insensitivity of model output to bedrock parameters may reflect the position of boundary conditions in the model: inflows and outflows are primarily bounded in layer 1, the Glaciofluvial aquifer. The remaining parameters, multipliers of river bed conductances KRB2 through KRB5, have decreasing sensitivity coefficients corresponding to increasing stream order. The decreasing model sensitivity to changes in parameters of increasing stream order is due to the effect on the model solution of predominantly localized ground-water flow cells, which correspond to the lower Horton-order drainages (see Hoaglund, 1996).

Regional and Local Ground-Water Budgets

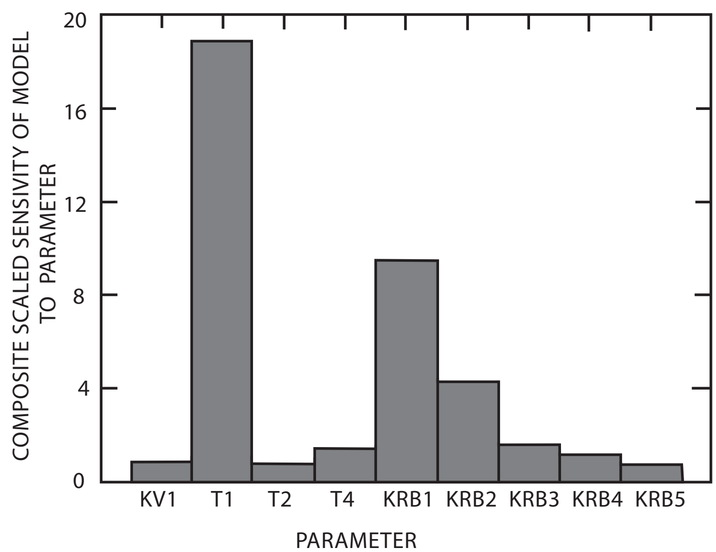
The volumetric budget for the entire model is shown in table 3. Ground-water inflow is derived almost entirely from ground-water recharge, with a relatively small contribution derived from inflow from rivers. Ground-water outflow is divided between flow to rivers and flow to the Great Lakes specified heads. Ground-water inflow and outflow balances within -0.01 percent. The low ground-water outflow to specified heads relative to flow to rivers indicates that local flow cells, which direct flow to rivers, predominate flow in the Glaciofluvial aquifer (layer 1). The model is calibrated with a recharge-equals-baseflow assumption, however, and recharge under-estimation, due to unaccounted for deep seepage (see earlier section on Ground-water Recharge), would imply that this direct ground-water discharge to the Great Lakes is generally underestimated. Areas of recharge under-estimation (recharge areas) are of greater areal extent than areas of recharge over-estimation (discharge areas), with the net effect that recharge is generally underestimated in the model.

Characteristics of Regional Flow and Base Flow

The overall pattern of the water table solution is affected by the internal river boundaries that display a dendritic pattern (fig. 6). The rivers divide the flow system into local flow

Table 3. Summary of the hydrologic budget components estimated by the model
[M^3/s – cubic meters per second; ft^3/s – cubic feet per second;% – percent of total budget]

Hydrologic Source /Sink	Inflow from			Outflow to		
	M^3/s	ft^3/s	Percent	M^3/s	ft^3/s	Percent
Recharge	711	25124	99.8	0	0	0
Rivers	1.39	49	.2	677	23902	95
Great Lakes	0	0	0	36	1272	5



EXPLANATION

KV1, Vertical conductance between layer 1 and layer 2
T1, Transmissivity of Layer 1
T2, Transmissivity of Layer 2
T4, Transmissivity of Layer 4

KRB1, River bed conductance for first order streams
KRB2, River bed conductance for second order streams
KRB3, River bed conductance for third order streams
KRB4, River bed conductance for fourth order streams
KRB5, River bed conductance for fifth order streams

Figure 13. Composite scaled sensitivities for the 9-parameter MODFLOWP simulation.

cells. As a result, most shallow ground-water flow discharges to the rivers rather than to the Great Lakes (see earlier section on ground-water budgets). Application of a particle tracking algorithm (Hoaglund, 1996) indicates that local flow cells exist, drastically reducing the ground-water residence time within layer 1 (Hoaglund, 1996).

Some regional ground-water flow is evident from the results of particle tracking (Hoaglund, 1996), particularly along flowpaths directed downward to the bedrock aquifers from the upland areas, which on average flow toward the Great Lakes perpendicular to the contour lines on the water table and head maps. Ground-water flow is regionally downward into the bedrock aquifers, except in the Saginaw and Michigan lowlands, where flow is upward. Ground-water flow may be upward in the Erie lowlands, but bedrock aquifers were not simulated in this region beyond the study area. Bedrock aquifer conductivities are considerably lower than Glaciofluvial aquifer conductivities, resulting in regional ground-water flow paths of long residence time within the bedrock (Hoaglund, 1996). However, the volume of water incorporated into these flow paths is considerably less than that in the Glaciofluvial aquifer. The Glaciofluvial aquifer transmits the most water, and with considerably shorter residence times.

Ground-Water Discharge to the Great Lakes

Though direct ground-water flow to the Great Lakes is only 5 percent of the model volumetric outflow budget, it is of great concern to management and study of the lakes. Figure 14 shows direct ground-water discharge from the Glaciofluvial aquifer (layer 1), calculated for each specified head cell and plotted against the shoreline length measured from southwest Michigan clockwise to southeast Michigan. Twenty-eight (28) coastal cities are tagged to the plot from a map of Michigan. The calculated discharge is variable cell to cell, higher in bays where flow is focused into embayments (Cherkauer and McKereghan, 1991) and lower near rivers where flow is diverted into rivers. Confidence intervals are not shown on the plot because direct ground-water discharge data were not available for model calibration. However, ground-water flow to lateral specified head boundaries in riparian drainage basins is similar to ground-water flow to a river from a symmetric half of an interior basin. As a result, the accuracy of model prediction would be similar to the accuracy implied by the individual confidence intervals shown for the 22 large basins, expressed as a percentage of total flow. The average 95-percent confidence interval for these basins was 38 percent.

LIMITATIONS OF THE MODEL

Several limitations to the modeling approach must be considered before the results of the simulations are accepted and applied. These include the degrees to which the assumptions of recharge, Darcy flow, steady state, scale, and internal boundary conditions are met. These limitations are in addition to the input parameter assumptions and external boundary conditions reviewed above.

The recharge rates used in the model were extrapolated from selected stream baseflow measurements and related to basin characteristics and precipitation. This method assumes that all recharge water in the selected basins discharges at the stream as baseflow without losses to, or gains from, deep seepage. The extrapolated recharge rates of the model may be similarly underestimated or overestimated. Therefore, the deep seepage (underflow) component of flow in the model may not be properly accounted for in the model budget. Generally, model recharge would be underestimated in the regional recharge areas due to unaccounted for deep seepage losses, and overestimated in regional discharge areas due to unaccounted for deep seepage gains.

The MODFLOW model solves a governing

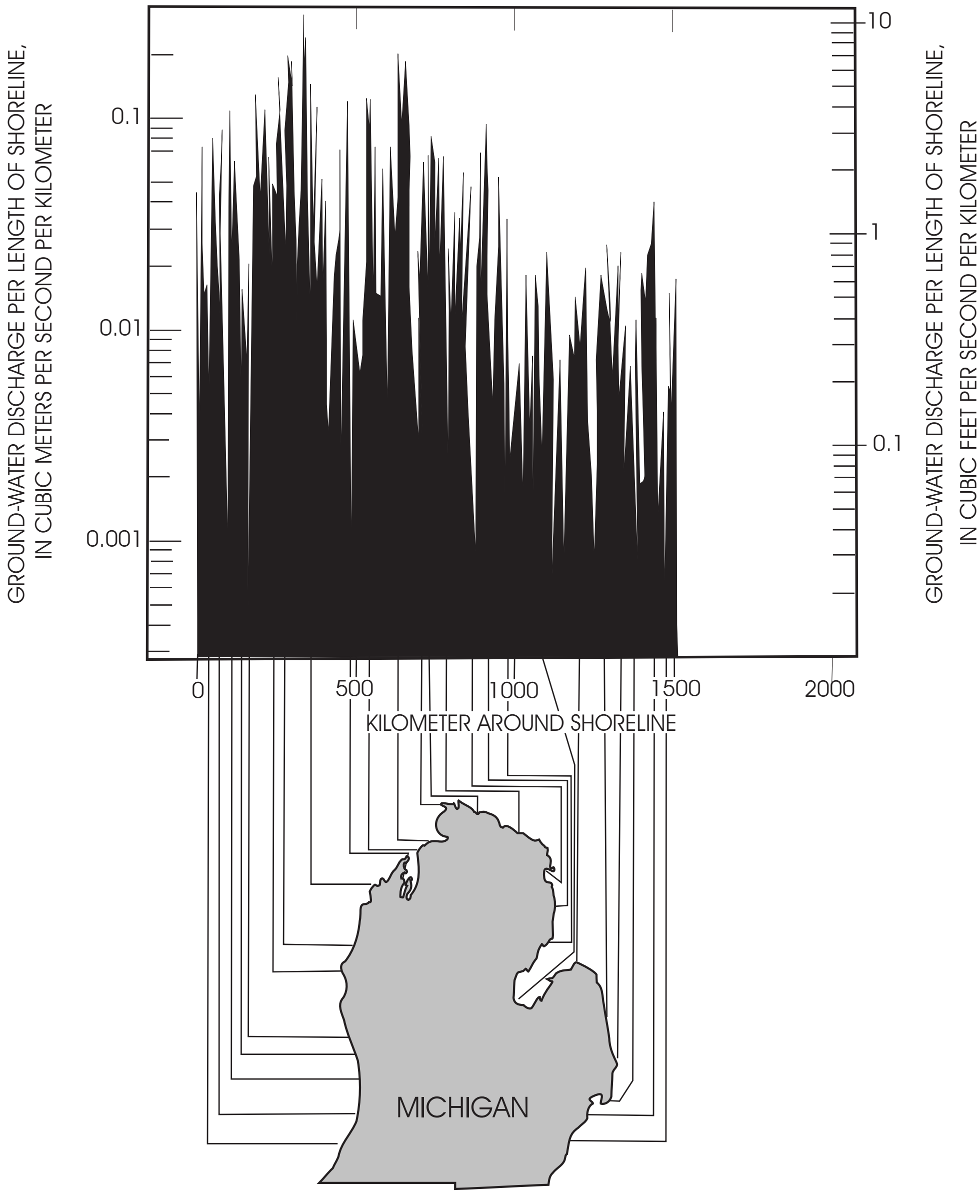


Figure 14. Direct ground-water discharge to the Great Lakes from the Glaciofluvial aquifer.

equation derived from Darcy's law and the principle of continuity. Furthermore, construction of the Michigan RASA model was based on the assumption of heterogeneous and horizontally isotropic flow conditions. The model, therefore, would not be valid in karst regions or under other non-Darcy flow conditions, nor in anisotropic flow conditions such as fracture-dominated flow.

The Michigan RASA model assumes a steady-state condition, which implies that the Michigan Basin has reached equilibrium. Head and flow measurements used to calibrate the model were assumed to be time invariant. The flow budget for the model, including recharge, was also assumed to be time invariant. Because the model is a pre-development simulation, the effects of municipal withdrawal were not modeled.

All model parameters and boundary conditions are scale dependent. As a result, interpretations from the model should be limited to the scale of the investigation.

Approximately 10 percent of layer 1 nodes are stream boundary nodes that form a dendritic pattern over the entire solution area. As a result, boundary influences must be considered when evaluating the heads, or the effect of stresses on heads, in layer 1. For example, data on the elevations of stream crossings were eliminated from the head calibration data set because this same data was used to set the stage boundaries in the river package of MODFLOW.

SUGGESTIONS FOR FUTURE WORK

A modeling effort results in a compilation of a geologic database as much as in an analytical tool for describing ground-water flow. An online digital archive of this data would make it available to other studies, both hydrologic and geologic.

Calibration of a transient solution to a transient database of regional water levels and stream discharges would be the most critical step toward converting the existing model into a regional ground-water management tool. The requirement of pre-development heads and discharges for calibrating a steady-state solution, together with the regional scale of the model, severely limited what little usable regional data existed at the time of model construction. The current, steady-state model does not accurately characterize the effect of municipal water withdrawals, or the effect of stream and/or lake infiltrations that may be induced near regional pumping centers. Corrections of model input parameters are required for the model to sustain current municipal withdrawal rates in a steady-state simulation. Although this suggests that the model may need further stress calibration, there is no indication that current withdrawal rates are, in fact, in steady state. The model in its current configuration may be correct in predicting that current withdrawal rates cannot be sustained, and/or achieve, a steady state.

Simulation of the effect of the variable density of ground water on both the head solution and ground-water advection may be required to characterize the effect of Michigan Basin brines on shallow ground-water quality (Wahrer and others, 1996; Meissner and others, 1996; Ging and others, 1996) and the effect of ground-water discharge on water quality in the Great Lakes (Kolak and others, 1999). Dynamic variable density could be simulated by application of a fully three-dimensional, variable density ground-water code. The current freshwater solution shows little hydrodynamic interaction between shallow ground water and deep bedrock aquifers, except in discharge areas. Given the shape of the confining unit surfaces, variable density would result in even less communication between dilute shallow ground waters and deeper bedrock brines.

The Glaciofluvial aquifer needs to be subdivided into multiple units and modelled with multiple layers because the Glaciofluvial aquifer system critically controls both local and regional recharge and discharge. A statewide glacial geologic mapping effort, part of the Central Great Lakes Geologic Mapping Coalition, is currently in the planning stages.

SUMMARY

A ground-water flow model simulating steady state conditions in glacial deposits and bedrock within the Lower Peninsula of Michigan quantified water levels within four regional aquifers, and discharges to rivers and the shoreline of the Great Lakes bounding the flow system. The recharge source of water into the ground-water flow system was estimated in a separate study relating stream baseflow (assumed equal to recharge) to climate, forest cover, and geologic basin characteristics. Ground-water heads in the Glaciofluvial aquifer were simulated under Dupuit water table conditions. Ground-water heads in the Saginaw, Parma-Bayport, and Marshall aquifers were simulated under confined conditions. The Saginaw and Parma-Bayport aquifer heads were coincident with the Glaciofluvial heads suggesting only moderate dampening (lowering) of heads due to confining units. Marshall aquifer heads are considerably dampened, though still influenced by the Glaciofluvial heads. The Marshall aquifer is affected by brine, and requires variable density simulation for calibration (beyond the scope of this report).

Model output is most sensitive to changes in the Glaciofluvial aquifer conductivity, and the river bed conductivities that regulate discharge from the Glaciofluvial aquifer to streams. Furthermore, Glaciofluvial aquifer heads are strongly imprinted on the head solutions of the deeper bedrock aquifers. Glaciofluvial heads, therefore, most represent the average ground-water flow system of the model, and indicate topographically driven regional ground-water flow from upland to lowland areas. Local flow cells predominate, however, as evidenced by both the model hydrologic budget and a separate particle tracking, ground-water travel-time study. The local flow cells drastically reduce ground-water travel-time and the significance of regional ground-water flow. Recharge estimates used in the model may need to be modified to account for deep seepage. The model recharge estimates resulted in a direct ground-water discharge to the shores of the Great Lakes of approximately 36 M^3/ s (1,272 ft^3/s), or about 5 percent of the model budget.

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