

ground water

Michigan Basin Regional Ground Water Flow Discharge to Three Great Lakes

by John Robert Hoaglund III¹, Gary Cecil Huffman², and Norman Guy Grannemann³

Abstract/

Ground water discharge to the Great Lakes around the Lower Peninsula of Michigan is primarily from recharge in riparian basins and proximal upland areas that are especially important to the northern half of the Lake Michigan shoreline. 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 interlayered with the 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, 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. According to the flow model, local stream stages and discharges account for 95% of the overall model water budget; only 5% enters the lakes directly from the ground water system. Direct ground water discharge to the Great Lakes' shorelines was calculated at 36 m³/sec, accounting for 5% of the overall model water budget. Lowland areas contribute far less ground water discharge to the Great Lakes than upland areas. The model indicates that Saginaw Bay receives only ~1.13 m³/sec ground water; the southern half of the Lake Michigan shoreline receives only ~2.83 m³/sec. In contrast, the northern half of the Lake Michigan shoreline receives more than 17 m³/sec from upland areas.

Introduction

The water balances of the Great Lakes are of considerable interest to numerous stakeholders. Airborne and riverine fluxes are more accessible and thus better known than ground water discharges. In their budget for the Great Lakes, Croley and Hunter (1994) neglect ground water and separately itemize interbasin diversions, limiting nondiversion lake hydrologic inputs to "basin runoff" (i.e., tributary river discharge) and direct precipitation, with runoff providing 43% and precipitation 57% of these inputs to Lake Michigan, for example. 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

1999). Unlike precipitation and runoff, no comprehensive monitoring program of ground water fluxes is in place. Grannemann and Weaver (1998) summarized estimates of ground water discharges to the Great Lakes made by others; however, few of these studies involve regional models that can balance recharge inputs to stream and riparian discharge outputs. Nauta (1987) provided ground water discharge estimates from a peninsula within Lake Michigan. Sellinger (1995) provided ground water discharge estimates from a lowland area along Lake Michigan. More recently, Boutt et al. (2001) provided an estimate of ground water flow to an embayment along Lake Michigan. To examine the effects of physiography, geology, and shoreline geometry on discharge, we developed a regional ground water model with internally consistent recharge that provides estimates of ground water discharge from the entire Lower Peninsula of Michigan to the surrounding Great Lakes.

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 ~5334 m (17,500 feet) (Lillienthal 1978). Paleozoic through Jurassic rocks are mantled by glacial deposits that are the result of

¹The Pennsylvania State University, Earth and Mineral Sciences Environment Institute, 2217 Earth-Engineering Sciences, University Park, PA 16802-6813, (814) 865-4792; fax (814) 865-3191; hoaglund@essc.psu.edu

²U.S. Geological Survey, WRD, Lansing, Michigan (retired)

³U.S. Geological Survey, WRD, 6520 Mercantile Way, Ste 5, Lansing, MI 48911; (517) 887-8936; fax (517) 887-8937; nggranne@usgs.gov

Received May 2001, accepted March 2002.

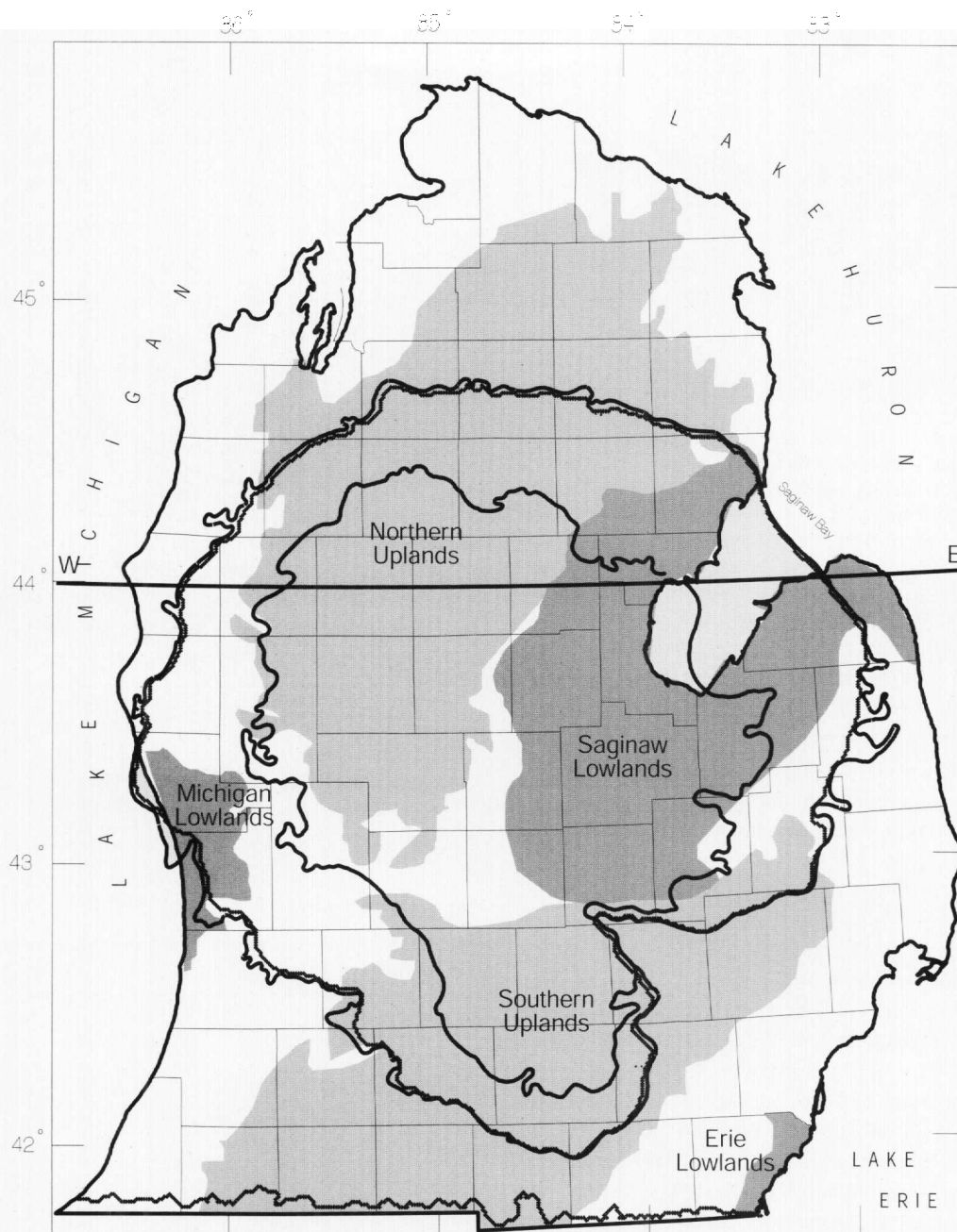


Figure 1. Location of the RASA study area in the Lower Peninsula of Michigan showing upland and lowland areas, and boundaries for model layers 1–4 of the RASA ground water flow model. Latitude and longitude bounding rectangle corresponds to edges of the 361-column \times 470-row RASA model grid. Line W–E denotes cross section of Figure 2.

the Wisconsinan and, possibly, earlier glaciations. Ice from the last glaciation receded from Michigan ~10,000 years ago (Eschman 1985, p. 164).

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 Michigan Basin that are bounded laterally by three of the Great Lakes. 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, providing the framework for a MODFLOW model of the system. The Michigan Basin RASA study area was restricted to a 56,980 km² (22,000

square miles) region of Michigan's Lower Peninsula, defined by the extent of the bedrock units (Figure 1). Geologic data, including thicknesses of glacial deposits and land-surface elevations, were obtained from existing sources to characterize the Glaciofluvial Aquifer across the entire Lower Peninsula of Michigan, with the Great Lakes as boundaries on three sides of the model. As a result, the model can calculate both fluvial and direct riparian components of Great Lakes ground water discharge from the Lower Peninsula of Michigan. Steady-state results are presented later.

The Michigan Basin RASA project included study of the geology (Westjohn et al. 1994; Westjohn and Weaver 1996a, 1996b, 1996c; Westjohn and Weaver 1998), aqueous geochemistry (Wahrer et al. 1996; Meissner et al. 1996;

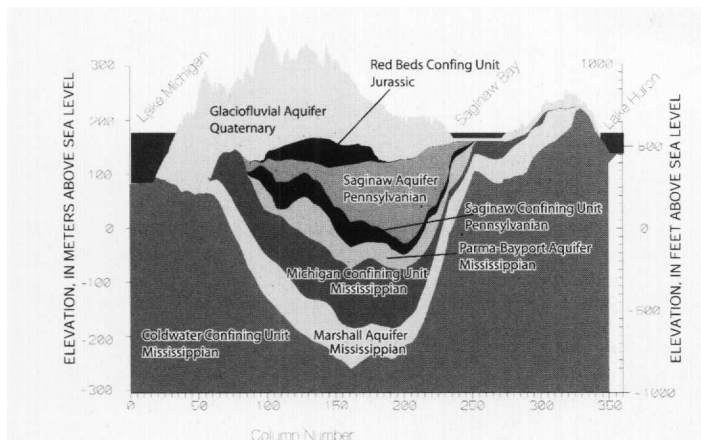


Figure 2. Geologic cross section along model row 240 showing structural relations among aquifer and confining units. Vertical exaggeration is 394 times.

Ging et al. 1996), water level (Barton et al. 1996), and recharge (Holtschlag 1996, 1997) of aquifers that are primary sources of ground water supply for human needs. Hoaglund et al. (2002) present the design, calibration, and sensitivity analysis of the MODFLOW model. Hoaglund et al. (2002) also includes sensitivity analysis of calculations and reports limitations for using the model. The model simulates the regional, predevelopment ground water flow system. An analysis of paleo-ground water flow directions was provided by Hoaglund (1996). This report focusses on the discharge to three Great Lakes and highlights the ground water flow system from which these numbers are based.

Four major aquifers were identified in the Michigan Basin RASA study area (Figure 2): the Glaciofluvial (Westjohn et al. 1994), Saginaw (Westjohn and Weaver 1996a), Parma-Bayport (Westjohn and Weaver 1996a), and Marshall aquifers (Westjohn and Weaver 1996b). These four aquifers are underlain by four confining units: the Till/Red Beds (Westjohn et al., 1994), Saginaw (Westjohn and Weaver 1996a), Michigan (Westjohn and Weaver 1996b), and Coldwater (Westjohn and Weaver 1996b) respectively. Significant geological assumptions affecting model construction are addressed in Hoaglund (1996) and Hoaglund et al. (2002).

The Coldwater Shale/Marshall Sandstone contact delimited the RASA study area for the purposes of defining the bedrock aquifers. The modeled area was extended across the entire Lower Peninsula of Michigan by incorporating peninsula-wide information on the glacial units, including landsurface elevation and glacial thickness information from the Western Michigan University hydrogeologic atlas (1981). The Coldwater Shale underlies these glacial deposits for most of the region between the study area and the Great Lakes shoreline, except for the Grand Traverse area where Devonian limestones, stratigraphically lower than the Coldwater, form the bedrock floor. It was assumed that the top of bedrock, including the Coldwater Shale and the Devonian limestones, formed a basal no-flow boundary for the Glaciofluvial Aquifer in this region.

Ground water flow in the aquifer system balances areal recharge inputs with indirect stream baseflow and direct

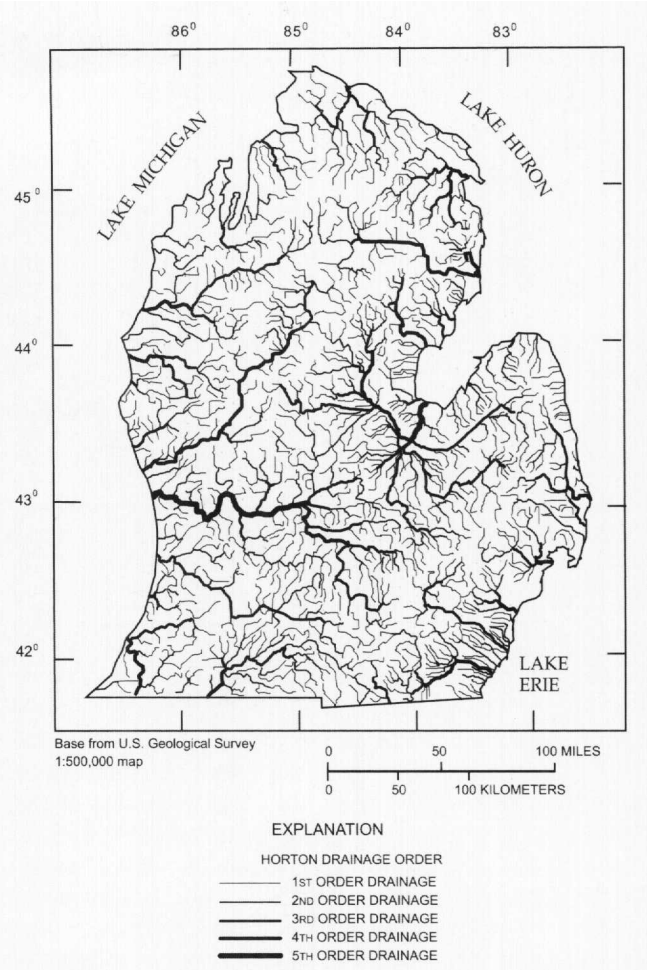


Figure 3. Location of nodes used to simulate streams and lakes in the ground water flow model.

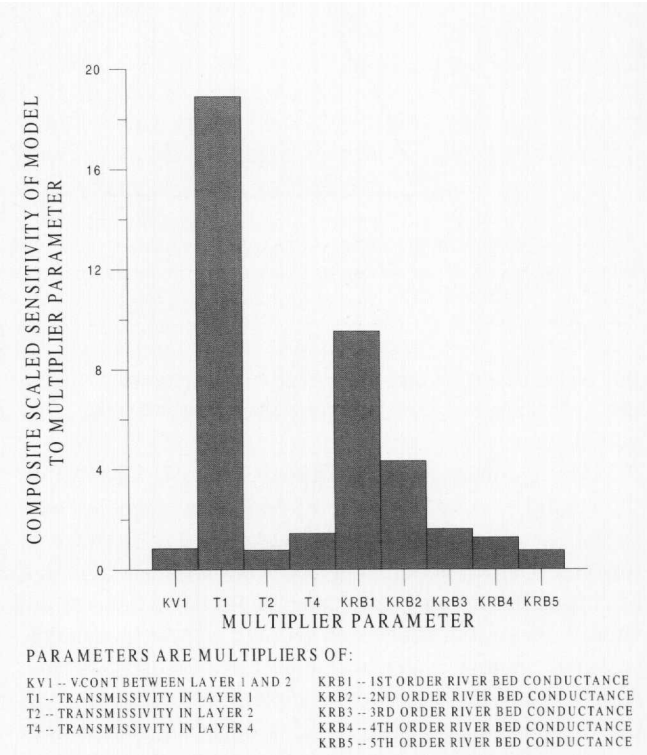


Figure 4. Composite scaled sensitivities for the unit-parameter simulation.

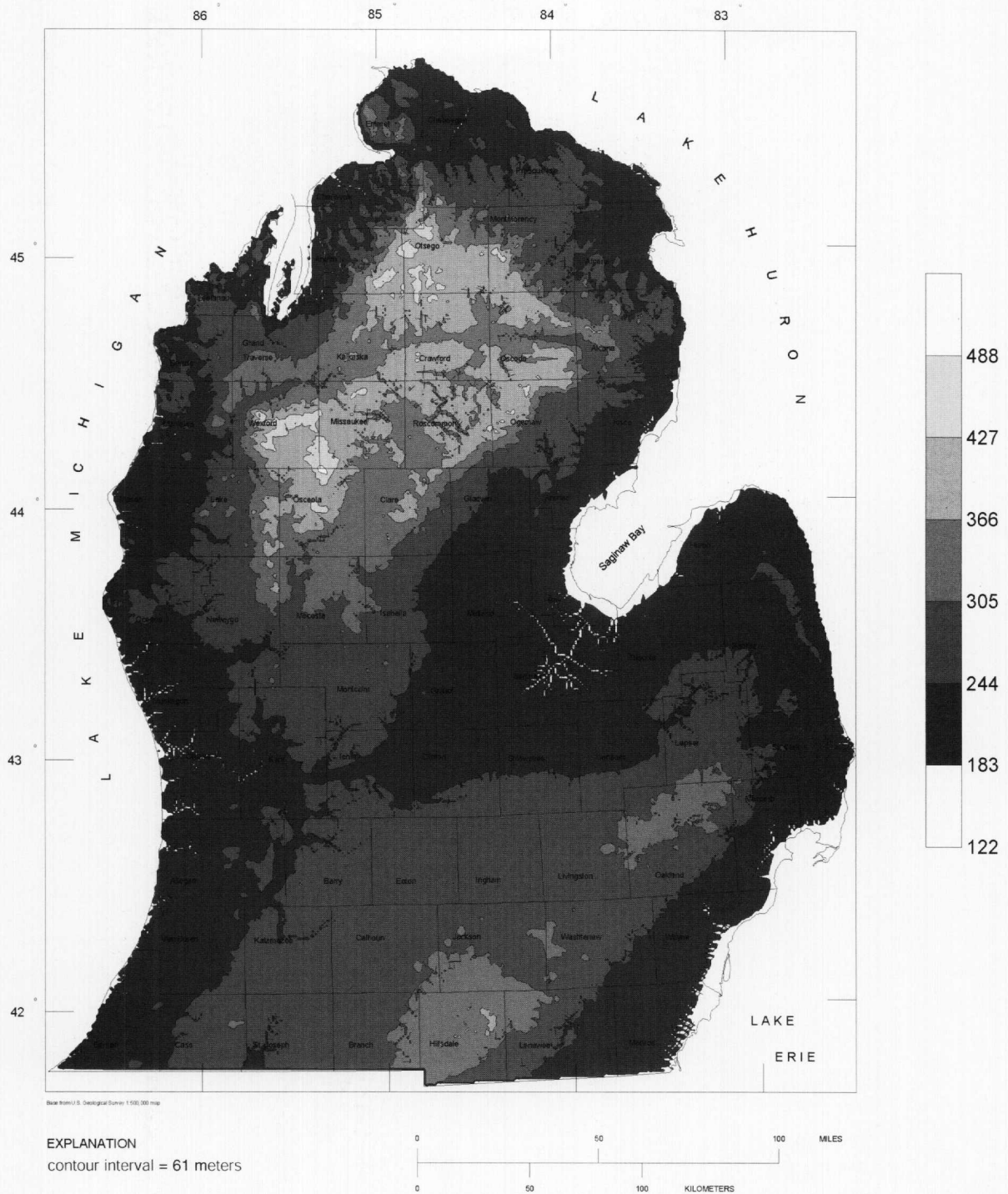


Figure 5. Simulated hydraulic head (Dupuit water table) for the glacial aquifer, model layer 1.

riparian discharge outputs to the Great Lakes. 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 input and output other than flows between specified heads. 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.

Holtschlag (1996, 1997) regressed precipitation and average annual basin baseflow rates from 114 unregulated drainage basins to determine steady-state normal baseflow for a 30-year period (1951 through 1980). The normal basin baseflow calculation most accurately reflects predevelopment, steady-state baseflow, and was assumed to equal steady-state recharge for the basin. Recharge was then mapped statewide from a separate regression relating recharge to basin characteristics.

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

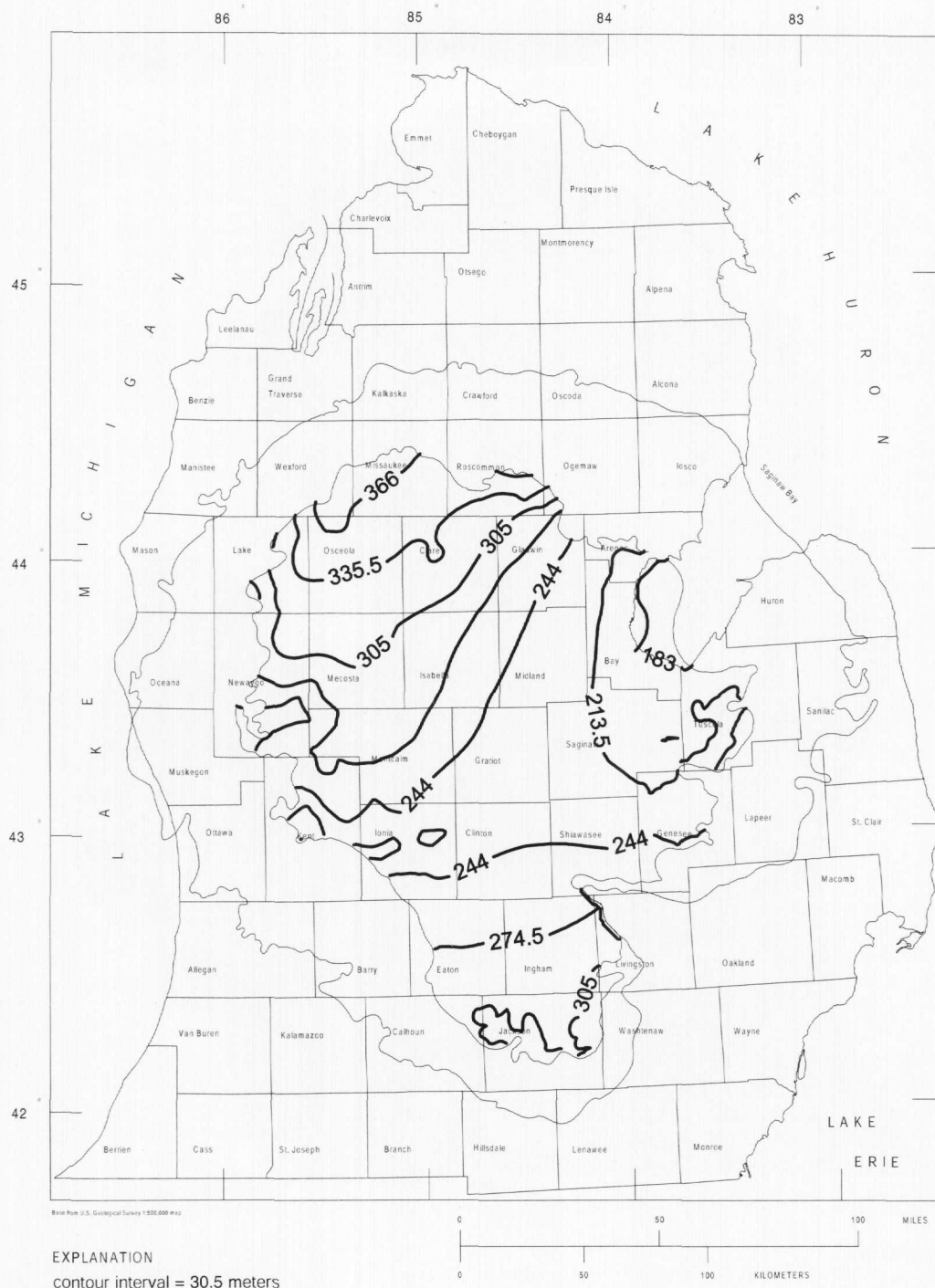


Figure 6. Simulated hydraulic head for the Saginaw Aquifer, model layer 2.

ranges from 526 m (1725 feet) above mean sea level in the west-central part to 174 m (572 feet) above mean sea level 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. High water table altitudes exist in the Northern and Southern uplands (Figure 1 of Fenneman 1938). Low water table altitudes coincide with the Saginaw, Michigan, and Erie lowlands (Figure 1 of Fenneman 1938), as well as along the paleo-, proglacial Grand River valley, which is the site of the present day Grand, Maple, Bad, and Saginaw rivers (Figure 3). Generalized ground water flow directions in the

Glaciofluvial Aquifer are toward the lowland areas, the Great Lakes, and the proglacial Grand River.

A predevelopment water table map was prepared by Mandel and Westjohn (1989) and reproduced as Figure 5 of Barton et al. (1996). Predevelopment fresh water head maps for the Saginaw and Marshall aquifers were prepared using data from collected water-level records measured and reported in the late 1800s and early 1900s, and some records from early oil and gas exploration in the 1930s (Figures 6 and 8, respectively, of Barton et al. 1996). The distribution of fresh water head in these two bedrock aquifers is similar to that of the water table, although the

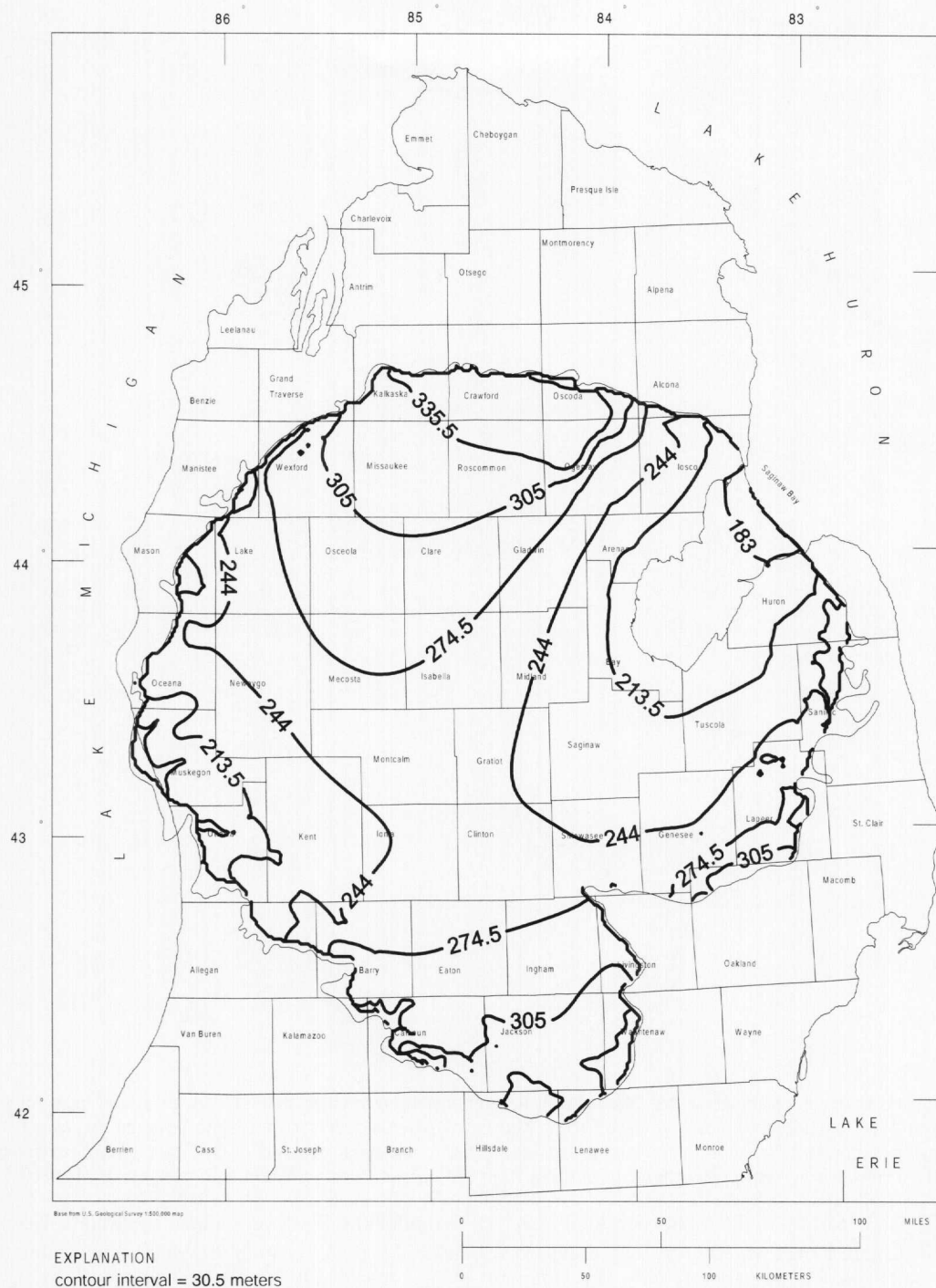


Figure 7. Simulated hydraulic head for the Marshall Aquifer, model layer 4.

magnitude of the heads in both aquifers is dampened relative to the water table. Vugrinovich (1986) noted a similar distribution of hydraulic heads. He concluded that predevelopment hydraulic head in both aquifers were generally in equilibrium with the present day land-surface elevations.

A major hydrologic feature of the Michigan Basin aquifer system is the presence of saline water near surface in the lowland areas (Wahrer et al. 1996), and saline water and brine down dip in the Parma-Bayport and Marshall aquifers (Westjohn and Weaver 1996c; Ging et al. 1996). Saline water also occurs in the Saginaw Aquifer in the west-central part of the study area, and with brine in the

Saginaw Lowlands (Westjohn and Weaver 1996c; Meissner et al. 1996).

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 for the Glaciofluvial Aquifer near the southern border of Michigan, separating flow in Indiana and extreme southern Michigan from flow in the study area. Although this ground water divide is not a real boundary and could shift, the boundary is sufficiently removed so as to not significantly affect calculation of the budget of the Lower Peninsula. Flows in the southern

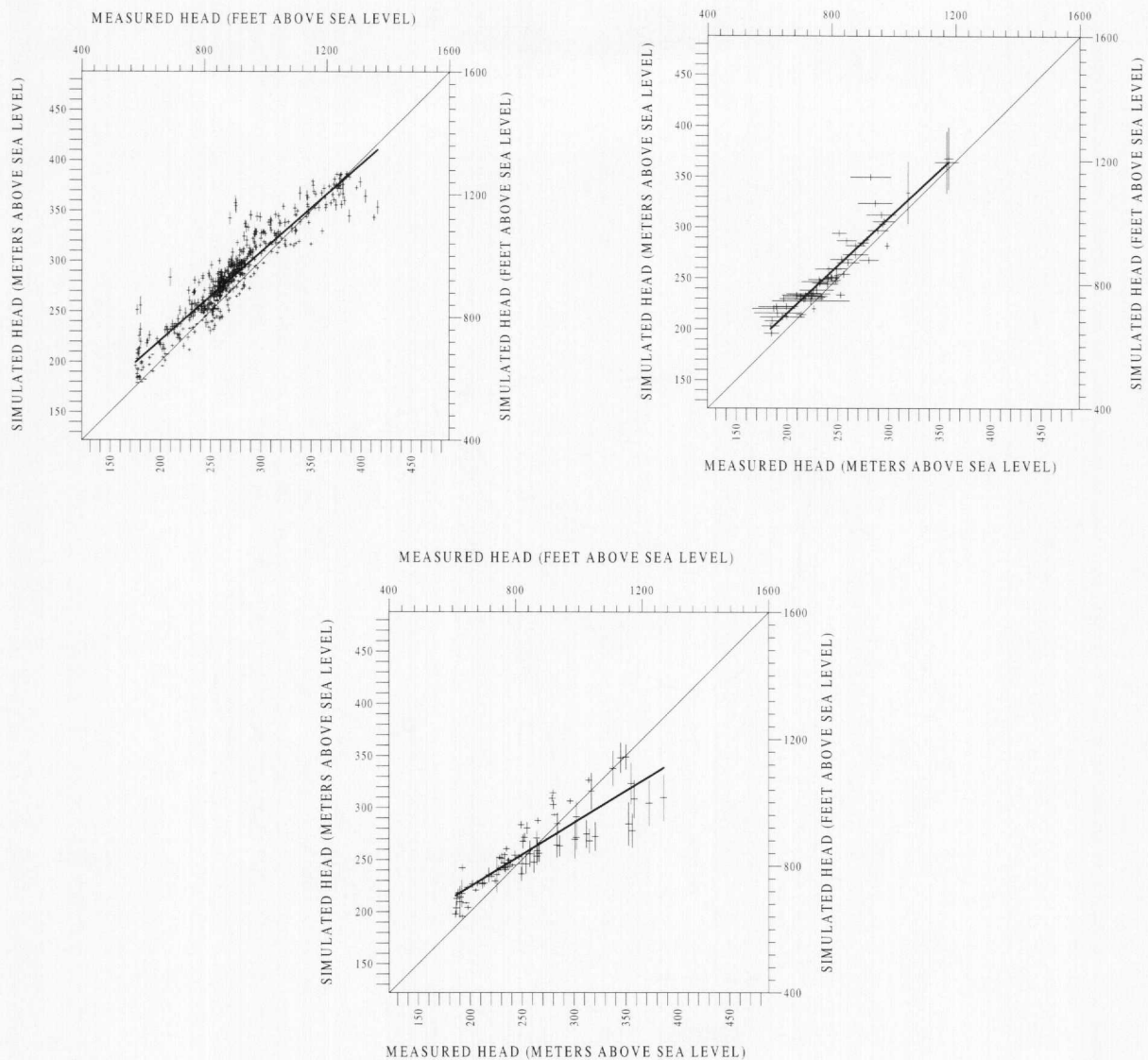


Figure 8. (a) Simulated vs. measured head for the Glaciofluvial Aquifer, model layer 1; (b) simulated vs. measured head for the Saginaw Aquifer, model layer 2; (c) simulated vs. measured head for the Marshall Aquifer, model layer 4. Diagonal line on graphs is line of perfect agreement. Vertical bars represent "sensitivity intervals" calculated from the unit-parameter MODFLOW simulation. Horizontal bars represent assumed, 95% confidence, measurement error bars, taken from the variance used in determining weights in MODFLOW.

Lower Peninsula are generally divided between the Great Lakes and the St. Joseph River, which ultimately drains to Lake Michigan. Subcrop-extent pinchouts form lateral no-flow boundaries for all bedrock aquifers. 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.

Methods

MODFLOW (McDonald and Harbaugh 1988) was used to simulate ground water flow in the aquifer system using the boundary conditions discussed above; assigning values for the hydraulic characteristics of the aquifers and confining units; and using independently estimated recharge rates for the ground water system. The model was calibrated using data that duplicates predevelopment measured heads and flow conditions as closely as possible,

assuming the predevelopment conditions were in steady state. At present, pumping and other alterations have changed the aquifer system so that the steady-state assumptions no longer apply. The program MODFLOW (Hill 1992) was used to conduct sensitivity analysis on an initial trial-and-error calibration, and to calibrate a refined set of parameters for an alternative simulation, both presented later.

MODFLOW is designed to simulate flow of water with constant density, and the results presented herein assume that the effect of observed variable density is negligible. This assumption is most reasonable for the Glaciofluvial Aquifer, where the observed variability of the density is negligible, and most unreasonable for the Marshall Aquifer, where brine exists below the Michigan confining unit. Most of the results presented herein stem from the heads and flows simulated in the Glaciofluvial Aquifer.

The Michigan RASA model grid comprises square finite-difference cells that are 1 km (3281 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 km (CRS1km) data set developed by Michigan State University's Center for Remote Sensing (CRS) to cover the state of Michigan (Lusch and Enslin 1984). The centroid of the upper left cell (Row 1, Column 1) corresponds to an easting of 290568 m and a northing of 557128 m.

Figure 2 shows a cross section of the geologic surfaces used in the model, reconstructed as model layer surfaces along a west to east transect depicted on Figure 1, model row 240 with an approximate latitude of 43°45' N. 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 et al. 1994; Westjohn and Weaver 1996a; Westjohn and Weaver 1996b), using interpolation methods described in Hoaglund (1996) and Hoaglund et al. (2002).

The top of bedrock represents a major angular unconformity (Figure 2) to the hydrogeologic flow system, the base of the Glaciofluvial Aquifer. Hoaglund (1996) and Hoaglund et al. (2002) describe adjustments to model layers that were required to handle pinch-outs and the angular unconformity where model layer 1 must communicate directly with model layer 4 in the region between the Saginaw and Marshall subcrops.

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 (Figure 1). 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 to simulate vertical leakage between the Glaciofluvial Aquifer and underlying bedrock aquifers (Figure 1). The southern boundary of the Glaciofluvial Aquifer consisted of drainage divides forming a continuous no-flow boundary (Figure 1). It was assumed that the surface drainage divides coincide with the ground water divide in this area.

Internal boundaries that represent the major streams in the modeled area were simulated using the RIVER module of MODFLOW (Figure 3). River reaches in the model were identified from the CRS1km data set (Lusch and Enslin 1984). The data set contains pixelated drainages corresponding to the Michigan Hydrologic Unit map (U.S. Geological Survey 1974). River stages were set by gridding a water table from 1220 river-crossing and 609 lake-level observations and assigning the values of the grid to the RIVER stages at river reach locations, and interpolating stages from headland to confluence to mouth in downstream order. Streambed thickness, reach length, and vertical hydraulic conductivity was set to 0.3048 m (1 foot), 1 km, and 1×10^{-4} cm/sec, respectively, for all river nodes, whereas the remaining component of stream conductance, river width, was set to 1, 2, 20, 60, and 80 m corresponding to 1 through 5 Horton (1945) stream orders, respectively.

Measured head values from the Glaciofluvial, Saginaw, and Marshall aquifers were compiled for calibration. Data for calibration were not available for the Parma-Bayport Aquifer. The measured head values for the Glacioflu-

vial Aquifer correspond to 499 of the 609 lake elevations digitized from the 1:500,000 hydrologic basemap. The remaining 110 of the 609 lake data points were located in a model river cell and thus were eliminated to reduce internal boundary bias in the calibration. The measured head values also include 57 well data points for the Saginaw Aquifer and 75 well data for the Marshall Aquifer, used to prepare Figures 6 and 8 of Barton et al. (1996, data not reported), respectively.

Hydraulic conductivity estimates (Table 1) for the aquifer (coarse-textured) and nonaquifer (fine-textured) portions of each model aquifer layer were used in conjunction with their respective fractions to construct effective, heterogeneous, model layer conductivity arrays, both horizontally and vertically (Hoaglund 1996; Hoaglund et al. 2002), as input. The bedrock aquifer estimates, summarized in Table 1, were based on a study by Westjohn et al. (1990) for the RASA project. The Glaciofluvial conductivity estimates, also summarized in Table 1, were based on a range of aquifer and confining unit pump test and modeling studies (results discussed in Mandel and Westjohn [1989]). Aquifer and nonaquifer portions of model aquifer layers were determined from separate aquifer isopach maps from the reports of Westjohn and Weaver (1996a, 1996c) and Westjohn et al. (1994). The vertical hydraulic conductivity estimates, additional estimates of the vertical hydraulic conductivities of intervening confining units, and structural contour top and bottom grids were used to construct effective VCONTS by implementing Equation 52 of the MODFLOW documentation (McDonald and Harbaugh 1988, pp. 5-16). The hydraulic conductivity input is thus scaled to the entire thickness of the layer and may be smaller than aquifer test values determined at the scale of a given field method. Further subdivision of the model layers may better conform model layers with field testable units, but may also limit model layer continuity at this regional scale.

Ground water recharge rates were estimated in an independent study for the RASA modeled area by Holschlag (1996 and 1997, discussed previously). The estimates were incorporated into the recharge module of MODFLOW as a steady-state grid. The recharge rates are heterogeneous even within basins, vary from 5 to 566 mm/year (0.19 to 22.3 in./year), and average 214 mm/year (8.41 in./year) (Holschlag 1996, 1997).

The recharge rates used in the model were extrapolated from selected stream base flow 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, therefore, may be correspondingly underestimated or overestimated. Thus, the deep seepage (a.k.a. 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 because of unaccounted deep seepage losses, and overestimated in regional discharge areas because of unaccounted deep seepage gains. Deep seepage is not measurable, except as interbasin transfer discernible by comparing water budgets between basins or between a riparian basin

Table 1
Summary of Horizontal and Vertical Hydraulic Conductivities for the Aquifers and Confining Units Used to Construct Model Conductivities and VCONTs

	Texture	Horizontal Hydraulic Conductivity (ft/day)	Vertical Hydraulic Conductivity (ft/day)	Horizontal Hydraulic Conductivity (cm/sec)	Vertical Hydraulic Conductivity (cm/sec)
Glaciofluvial Aquifer	Coarse grained	50	5	1.8×10^{-2}	1.8×10^{-3}
Fine-grained till or lacustrine deposits in Glaciofluvial Aquifer	Fine grained	2.9×10^{-4}	2.8×10^{-4}	1×10^{-2}	1×10^{-7}
Glacial till—J _R "red beds" confining unit	—	—	2.8×10^{-4}	—	1×10^{-7}
Saginaw Aquifer	Coarse grained	2.8	2.8	1×10^{-3}	1×10^{-3}
Intercalated fine-grained rock in Saginaw Aquifer	Fine grained	2.8×10^{-4}	2.8×10^{-6}	1×10^{-7}	1×10^{-9}
Saginaw confining unit	—	—	2.8×10^{-4}	—	1×10^{-7}
Parma-Bayport Aquifer	—	7.1	1.1×10^{-3}	2.5×10^{-3}	4×10^{-7}
Michigan confining unit	—	—	2.8×10^{-7}	—	1×10^{-10}
Marshall Aquifer	Coarse grained	1.4×10^{-1}	1.4×10^{-1}	5×10^{-5}	5×10^{-5}
Intercalated fine-grained rock in Marshall Aquifer	Fine grained	1.4×10^{-3}	1.4×10^{-3}	5×10^{-7}	5×10^{-7}

and its associated discharge to the lake's specified head boundary.

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 (1) calculate composite scaled sensitivities from the sensitivity matrix (Hill 1992, pp. 90–94) for the set of trial-and-error parameters to evaluate relative model parameter sensitivity; (2) calculate "sensitivity intervals" from the covariance matrix on the trial-and-error parameters, using the formulas for individual confidence intervals reserved for optimized parameters; and (3) estimate an alternative set of parameters using Gauss-Newton optimization. The nine input multipliers chosen as parameters include a multiplier for each of the following: the transmissivity arrays of the Glaciofluvial (T1), Saginaw (T2), and Marshall (T4) aquifers; the VCONT array between the Glaciofluvial and Saginaw aquifers (KVI); and the stream conductances for each "n" of five Horton (1945) stream orders (KRBn). The "sensitivity intervals" for simulated values and error bars for measured values are shown on plots of simulated versus measured heads and flows in Figures 7 and 8.

Results

Hydraulic conductivity and river conductance model input was adjusted until a trial-and-error calibration was achieved, balancing a match of heads and flows under conditions of fixed recharge. This final model input includes the hydraulic conductivities of Table 1 and the river conductances summarized in the "Methods" section. The model was then unit-parameterized, as discussed previously, and although the fit to observed head data was improved with parameter estimation (discussed later), the regression did not converge and thus parameters were not optimized. Composite scaled sensitivities from the unit-parameterization were then calculated with MODFLOWP, revealing the greatest sensitivity to layer 1 transmissivity and river conductances (Figure 4).

The unit-parameter solutions of hydraulic head for the Glaciofluvial, Saginaw, and Marshall aquifers are shown in Figures 5, 6, and 7, respectively. The hydraulic head solution of the Glaciofluvial Aquifer (layer 1) is the simulated water table elevation by the unconfined Dupuit assumptions used by the model. The output shown in Figure 5 compares well with the regional distribution of head, coincident with the water table map of Mandle and Westjohn (1989), which is based on stream crossing and lake elevation data, reproduced in Barton et al. (1996, Figure 5). The

heads are highest in the northern and southern upland areas with steep gradients, and lowest in the Saginaw, Michigan, and Erie lowlands with flat gradients. The Saginaw output (layer 2) shown in Figure 6 compares well with the regional distribution of head shown in Barton et al. (1996, Figure 6) based on well data. The solution of hydraulic head for the Parma-Bayport Aquifer (layer 3) is shown in Hoaglund (1996). The regions of high and low head in the Saginaw and Parma-Bayport aquifers are coincident with those in the Glaciofluvial Aquifer. The heads are only moderately dampened (lowered) by the overlying confining units, suggesting the Till "red beds" and Saginaw confining units leak. The solution of hydraulic head for the Marshall Aquifer (layer 4) shown in Figure 7 compares reasonably well with the regional distribution of head based on well data, with the exception of an area of high heads (>1100 feet) in the northwest on the observed map (Barton et al. 1996, Figure 8) that are not reproduced on the simulated map (Figure 7). The Marshall solution (layer 4) is considerably dampened relative to the water table because of 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. High and low areas still correspond to the northern and southern upland areas where the Michigan confining unit pinches out and the Marshall Aquifer (layer 4) has more communication with the water table in the Glaciofluvial Aquifer (layer 1) as seen in Figure 2.

Figures 8a through 8c show that the unit-parameter simulated heads are in good agreement with measured heads for the Glaciofluvial Aquifer, layer 1 (Mandle and Westjohn 1989; data not reported) and Saginaw Aquifer, layer 2 (Barton et al. 1996), and are in general agreement with the Marshall Aquifer, layer 4 (Barton et al. 1996). The simulated heads generally fall along the respective lines of agreement from low to high head, although 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. Marshall Aquifer simulation does not match some of the higher (>1100 feet) observed head values in the Marshall for which the simulated values are too low (Figure 8c). The confined Saginaw Aquifer that pinches out in this region shows lower head values than both the overlying unconfined Glaciofluvial Aquifer and the underlying confined Marshall Aquifer, suggesting that the higher Marshall heads must be transmitted laterally from a Marshall subcrop source, up-dip 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 these high head values in the Marshall Aquifer would require simulation of variable density, beyond the capabilities of MODFLOW. On all three calibration plots, slopes less than 1.0, together with positive intercepts, cause the best-fit lines to cross the agreement lines. The deviations from perfect agreement are not random, indicating some bias is present (i.e., residuals are not random with zero mean).

Concurrently with head calibration, the model was also calibrated against stream discharge measurements. Adjacent, downstream basins within the 114 discharge data set of Holschlag (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 ZONE-BUDGET (Harbaugh 1990) to compare modeled versus measured discharge. The 72 measured stream discharge values plot along a line of agreement on a graph (Figure 9a) of simulated versus measured stream discharge for river basins ranging from < 0.28 m³/sec (10 cfs) up to 28 m³/sec (1000 cfs). For the purpose of calibrating to the regional scale of the model and calculating confidence intervals on the flows, 44 basin zones were delineated, 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 plot along a line of agreement on a graph (Figure 9b) of simulated versus measured stream discharge.

For the calibration plots of Figures 8a through 8c and 9b, "sensitivity intervals" shown as vertical bars were calculated using formulas for individual confidence intervals from YCINT.FOR (Hill 1994), but with nonoptimized parameters from the nine-parameter MODFLOWP simulation. Assumed measurement errors are shown as 2σ (95% confidence) horizontal bars about each point, and the respective variances were used to set weights in MODFLOWP. The "sensitivity intervals" for the higher Marshall heads (Figure 8c) are notably larger than those for lower values because of the larger residuals translated into a larger parameter covariance for these uncalibrated observations. The 22 "sensitivity intervals" of Figure 9b averaged 38% of their respective flows.

The volumetric budget for the unit-parameter model is shown in Table 2 (model run A). In addition, a sum-of-squares for heads and a sum-of-squares for flows are provided in Table 2, both calculated by MODFLOWP to compare relative fit with other model runs. Ground water inflow is derived almost entirely from the ground water recharge, with a relatively small contribution derived from inflow from rivers. Ground water outflow is divided between flow to rivers (95%) and flow to the Great Lakes specified heads (5%). Ground water inflow and outflow balances within -0.01%.

Discussion

The low ground water outflow to specified heads relative to flow to rivers indicates that flow in the Glaciofluvial Aquifer (layer 1) is predominantly in local flow cells, directing flow to rivers as opposed to regional flow cells that direct flow to the Great Lakes. The overall pattern of the water table solution is affected by the internal river boundaries that display a dendritic pattern shown in Figure 3 and divide the flow system into localized flow cells. As a result, most shallow ground water flow discharges to the rivers rather than to the Great Lakes.

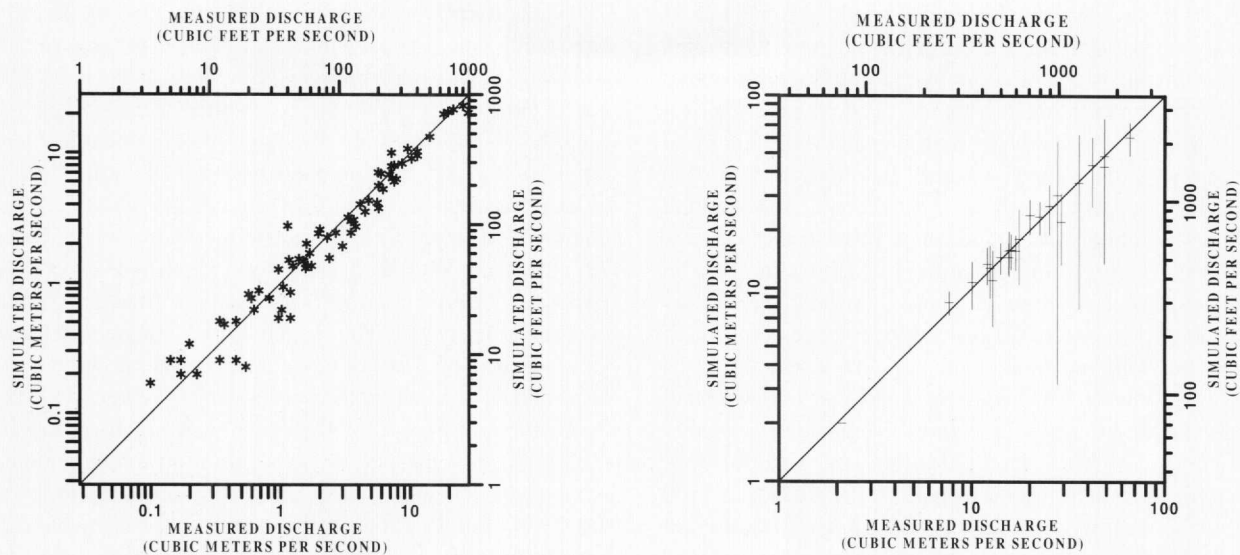


Figure 9. (a) Simulated vs. measured stream discharge for 72 gauging stations. Diagonal line on graph is line of perfect agreement; (b) simulated vs. measured stream discharge for 22 large basin gauging stations. Diagonal line on graph is line of perfect agreement. Vertical bars represent "sensitivity intervals" calculated from the unit-parameter MODFLOWP simulation. Horizontal bars represent assumed, 95% confidence, measurement error (5% of flow) bars, taken from the variance used in determining weights in MODFLOWP.

Application of a particle tracking algorithm (Hoaglund 1996) indicates that local flow cells drastically reduce the ground water residence time within layer 1 (Hoaglund 1996). Some regional ground water flow is evident from the particle tracking (Hoaglund 1996), particularly along flowpaths directed downward to the bedrock aquifers from the upland areas and which on average are directed toward the Great Lakes. 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, located beyond the study area. Bedrock aquifer conductivities are from 10 to 100 times lower than the respective (coarse versus fine) Glaciofluvial Aquifer conductivities, resulting in regional ground water flowpaths of long residence time within the bedrock (Hoaglund 1996), except possibly through faults. In addition, the bedrock is commonly tens to hundreds of meters below the water table, and interior bedrock aquifers pinch-out before reaching the Great Lakes (Figure 2). The result is that the Glaciofluvial Aquifer transmits the most water, either to rivers or the Great Lakes, through shallower flowpaths with 10 to 100 times shorter residence times (Hoaglund 1996).

Two sources of model bias may be evident from the head calibration of the Glaciofluvial and Saginaw aquifers, and thus may affect the discharge results: (1) heads for the aquifers are simulated too high, indicating that more water needs to be released to the rivers and/or transmitted through the aquifer layers; and (2) low heads are simulated too high and high heads are simulated too low, indicating that the recharge-equals-baseflow assumption underestimates recharge in the regional recharge areas because of unaccounted for deep seepage (underflow) losses, and overestimates recharge in regional discharge areas because of unaccounted for deep seepage gains.

Heads for the Glaciofluvial and Saginaw aquifers may be simulated too high because conservative values of horizontal conductivities were used in constructing model input. Given the high composite scaled sensitivity (Figure 4), increasing the Glaciofluvial (layer 1) transmissivity would have the largest effect here. The net effect would be to lower the heads and increase the model's calculation of direct discharge to the Great Lakes. Alternatively, heads for these aquifers could be lowered by releasing more water to the rivers. The net effect would be to lower heads and decrease the model's calculation of direct discharge to the Great Lakes. To assess the effects of these alternatives on the discharge calculation, two model runs are presented in Table 2. In model run B, the river conductances were increased by a factor of 10, improving the head calibration but worsening the flow calibration. The calculation of direct discharge to the Great Lakes lowered to 29.5 m³/sec (1041 cfs), reducing its budget to 4%. In model run C, the Glaciofluvial transmissivity was increased by a factor of 2, and leakance to the Saginaw and the Saginaw transmissivity were increased by a factor of 10, improving the head calibration but worsening the flow calibration. The calculation of direct discharge to the Great Lakes raised to 48.8 m³/sec (1725 cfs), increasing its budget to 7%.

MODFLOWP produced the best head calibration after 20 iterations, summarized by model run D. Together with model runs B and C, it is evident that better head calibration is obtained at the expense of the flow calibration, indicating that there is a disconnect between the head and flow data that might be responsible for the non-convergence. We believe this is a limitation of regional ground water studies generally, where in our example, 30-year steady-state baseflows were mixed with scale- and time-dependent head data.

High heads may be simulated too low and low heads may be simulated too high in the Glaciofluvial and Saginaw aquifers because the model is calibrated with a

Table 2
Summary of the Hydrologic Budget Estimated by the Model

Model Run	Sum of Squares (heads) ($\times 1000$)	Sum of Squares (flows)	Hydrologic Source/Sink	Inflow from:			Outflow to:		
				m ³ /sec	cfs	%	m ³ /sec	cfs	%
A	411	311	Recharge	711	25,124	99.8	0	0	0
			Rivers	1.4	49	0.2	677	23,902	95
			Great Lakes	0	0	0	36.0	1272	5
B	124	492	Recharge	711	25,124	97	0	0	0
			Rivers	22.7	802	3	705	24,885	96
			Great Lakes	0	0	0	29.5	1041	4
C	309	761	Recharge	711	25,123	99.5	0	0	0
			Rivers	3.5	122	0.5	666	23,521	93
			Great Lakes	0	0	0	48.8	1725	7
D	230	1142	Recharge	711	25,121	82	0	0	0
			Rivers	159	5618	18	835	29,503	96
			Great Lakes	0	0	0	34.4	1216	4
Run	KV1	T1	T2	T4	KRB1	KRB2	KRB3	KRB4	KRB5
A	1	1	1	1	1	1	1	1	1
B	1	1	1	1	10	10	10	10	10
C	10	2	10	1	1	1	1	1	1
D	0.059	1.43	0.0056	0.039	34710	4.53	0.16	0.94	0.012

recharge-equals-baseflow assumption. The deviations from perfect agreement in the Glaciofluvial and Saginaw aquifers may be related to the distribution of ground water recharge. An underestimation of recharge in the regional recharge areas because of unaccounted for deep seepage (underflow) losses, together with an overestimation of recharge in regional discharge areas because of unaccounted for deep seepage gains, would result in the observed trends where low heads are too high and high heads are too low. It is unclear whether the same is true for the Marshall Aquifer where most of the deviating trend may be related to the region of higher-than-simulated, observed head values that correspond to the high water table in the unconfined Glaciofluvial Aquifer of the northern uplands. For the Glaciofluvial Aquifer, areas of recharge underestimation (recharge areas) are of greater areal extent than areas of recharge overestimation (discharge areas), with the net effect that recharge is generally underestimated in the model. Recharge underestimation, caused by of unaccounted for deep seepage, would imply that the direct ground water discharge to the Great Lakes is generally underestimated.

Increased recharge as a correction to the model would increase underflow with a resultant increase in direct discharge to the Great Lakes. In their study of a small mountain watershed, Mau and Winter (1997) concluded that a recharge-equals-baseflow assumption resulted in a 25% recharge underestimation, which they attributed to deep seepage losses and losses from transpiration directly from

ground water. In the Michigan basin model, recharge constitutes 100% of the model input budget while the model output budget is split between two components: stream discharge and direct discharge. The model does not simulate evapotranspiration, which is instead incorporated in the recharge numbers. The percentage of model recharge correction would have to be entirely accommodated by underflows to lateral specified heads for the recalibrated model to match stream discharges. As a result, a fractional change in recharge, f , would change the fraction of direct discharge, a , to the value $(a+f)/(1+f)$ while changing the fraction of stream discharge, b , to the value $b/(1+f)$. Using the current model's direct discharge ($a = 5\%$) and stream discharge ($b = 95\%$) fractions and a 10% increase in recharge ($f = 10\%$), the new model output budget would be 14% direct discharge and 86% stream discharge. The conservative 10% increase in recharge would account for transpiration directly from ground water and scaling from a small mountain watershed with high relief to the Michigan basin's regional watersheds with low relief. Net recharge budget changes would be reflected only in the flows to specified heads component of the total regional water budget; flows to individual specified heads would likely show a wider range of changes. Furthermore, recharge should be increased in recharge areas but decreased in discharge areas to reflect regional underflow losses and gains respectively, modifications that are not incorporated in net recharge changes.

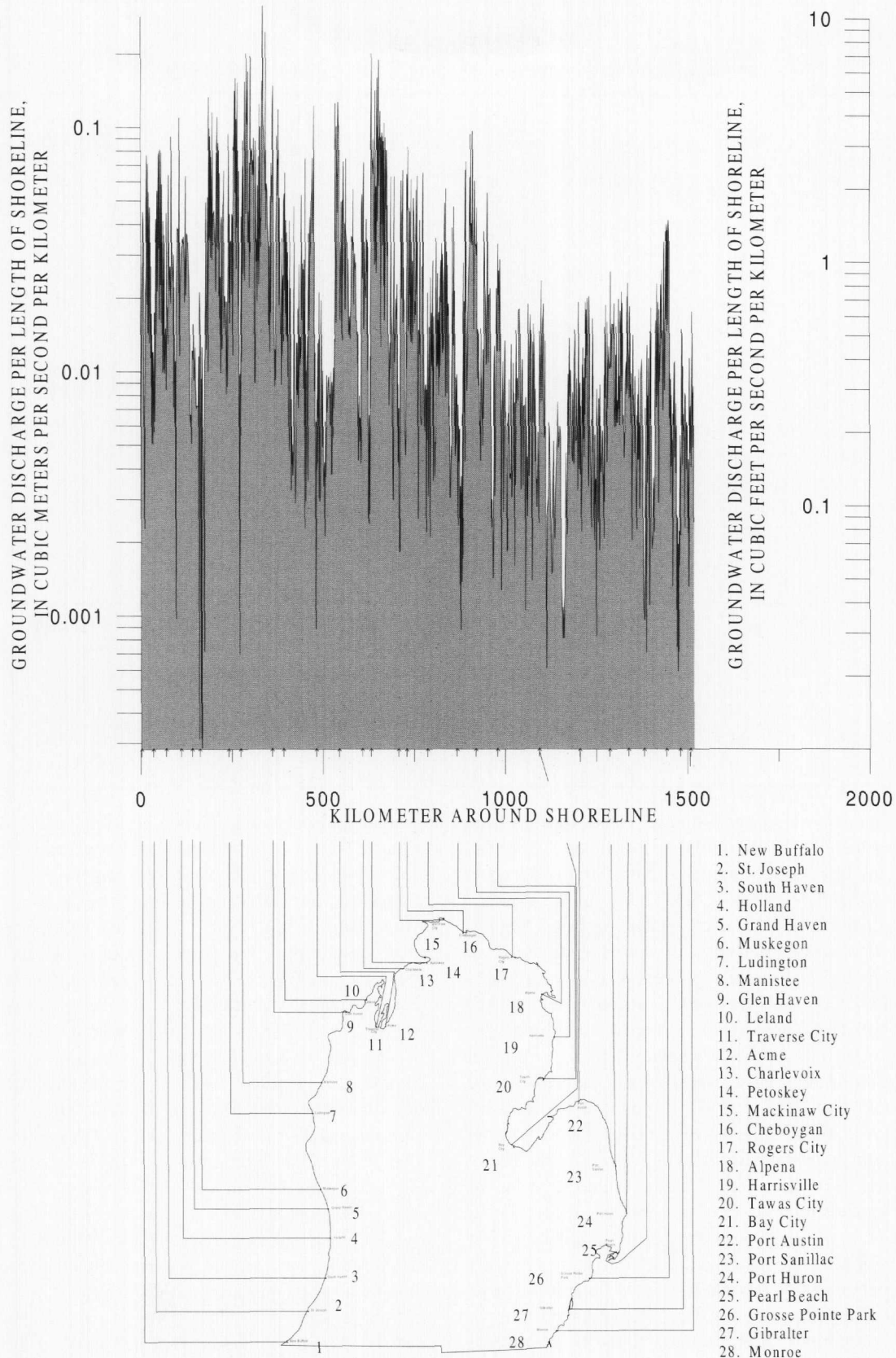


Figure 10. Direct ground water discharge per kilometer of shoreline to the Great Lakes from the Glaciofluvial Aquifer vs. numbered kilometer of shoreline.

To examine the effects of physiography, geology, and shoreline geometry on the discharge to the Great Lakes, flow from the Glaciofluvial Aquifer (layer 1) was calculated for each specified head cell and plotted against the

shoreline length measured from southwest Michigan clockwise to southeast Michigan (Figure 10). Twenty-eight 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.

The calculated discharge is lowest in the Saginaw and Michigan lowlands because of low model conductivity and thickness (low effective transmissivity) and lower gradients for the lowland regions. The model calculates that Saginaw Bay, measured from Tawas Point to Port Crescent State Park, receives only $\sim 1.13 \text{ m}^3/\text{sec}$ (40 cfs) direct discharge ground water from the Saginaw Lowlands with an average discharge of $5.1 \times 10^{-3} \text{ m}^3/\text{sec}$ (0.18 cfs) per km (Figure 11a). Discharge to the inner Saginaw Bay, from Point Au Gres to Sand Point, is $0.623 \text{ m}^3/\text{sec}$ (22 cfs), about half of the bay's total discharge, delivered at an average discharge of $4.5 \times 10^{-3} \text{ m}^3/\text{sec}$ (0.16 cfs) per km. The model calculates that the Michigan lowlands shoreline, measured from South Haven to Benona (Little Sable Point), receives only $\sim 3.11 \text{ m}^3/\text{sec}$ (110 cfs) direct discharge ground water with an average discharge of $2.38 \times 10^{-2} \text{ m}^3/\text{sec}$ (0.84 cfs) per km (Figure 11b).

The calculated discharge is highest along the northern Lake Michigan shoreline because of high model conductivity and thickness (high effective transmissivity) and steep gradients from the northern uplands. The model calculates that the northern Lake Michigan shoreline, measured from Big Sable Point to Mackinaw City, receives $18 \text{ m}^3/\text{sec}$ (637 cfs) direct discharge ground water with an average discharge of $3.96 \times 10^{-2} \text{ m}^3/\text{sec}$ (1.4 cfs) per km (Figure 11c). The highest discharge of $0.311 \text{ m}^3/\text{sec}$ (11 cfs) per km occurs near Sleeping Bear Dunes National Lakeshore, most likely because of locally steep ground water gradients and high hydraulic conductivity associated with a well sorted sand aquifer.

Comparing the discharge numbers to other studies requires complete understanding of the study's objectives and methods, and in the case of shoreline estimates, knowledge of the total contributing shoreline length. From a hydrologic budget analysis of the combined lakes Michigan and Huron, Bergstrom and Hanson (1962) calculated $22.7 \text{ m}^3/\text{sec}$ (800 cfs) ground water "discharge along the lake shores and underflow." Normalizing this value to 5630 km (3500 miles) of shoreline, they obtained an average discharge of $4 \times 10^{-3} \text{ m}^3/\text{s}$ (0.14 cfs) per km, which is equivalent to our lowest discharge average, that of Saginaw Bay. In addition, our discharges are only from the Glaciofluvial Aquifer along the lakeshore and do not include "underflow." Looking exclusively at hydraulic potential of lake bottom sediments, Cartwright et al. (1979) obtained discharges of $181 \text{ m}^3/\text{sec}$ (6380 cfs) through "near-shore sands" and $9 \text{ m}^3/\text{sec}$ (325 cfs) through "fine-grained soft deep-lake sediment" for Lake Michigan. The "near shore" estimate is somewhat analogous to our total shoreline discharge, but exceeds our value of $36 \text{ m}^3/\text{sec}$ (1272 cfs). The "deep lake" estimate and the "near shore" excess can be attributed to underflow. In other studies using MODFLOW, Nauta (1987) obtained discharges of $0.24 \text{ m}^3/\text{sec}$ (8.4 cfs) to Lake Michigan and $0.18 \text{ m}^3/\text{sec}$ (6.4 cfs) to Green Bay from the Door Peninsula of Wisconsin; Sellinger (1995)

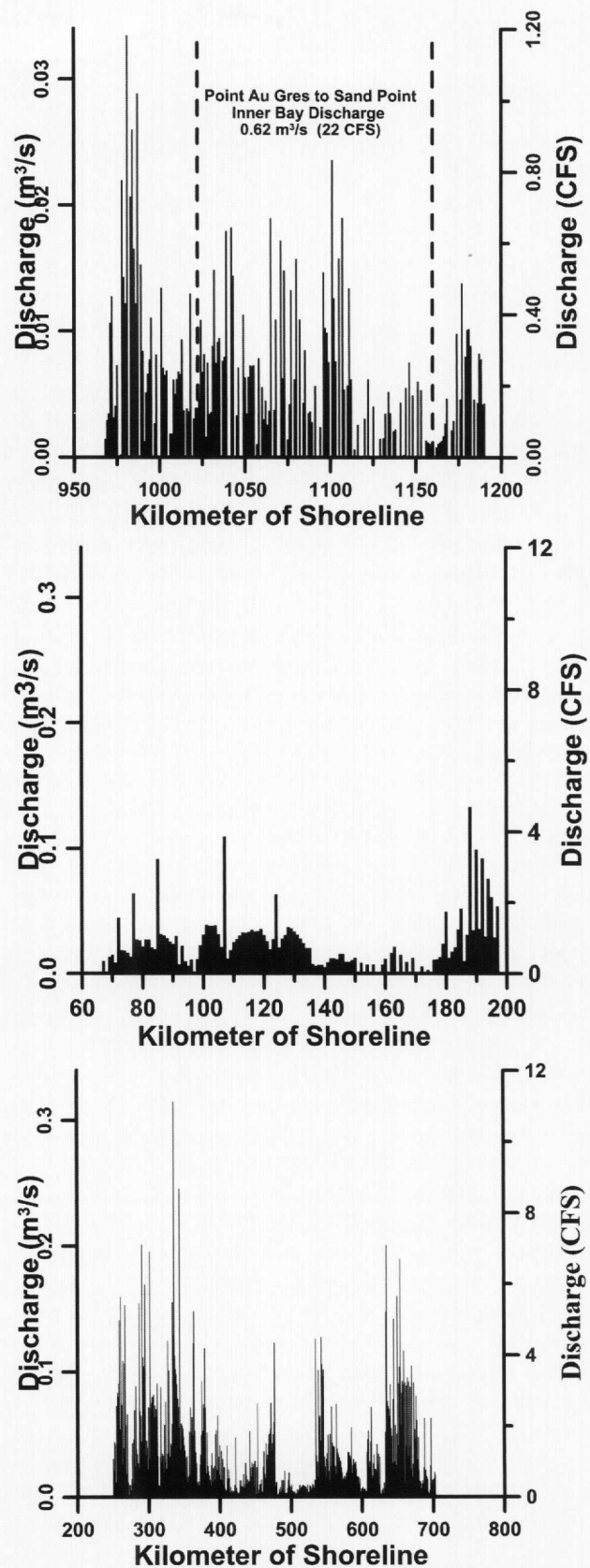


Figure 11. Direct ground water discharge from the Glaciofluvial Aquifer per kilometer of shoreline vs. numbered kilometer of shoreline: (a) total discharge to Saginaw Bay from the Saginaw Lowlands is $1.16 \text{ m}^3/\text{sec}$ (41 CFS) averaging $5.1 \times 10^{-3} \text{ m}^3/\text{sec}/\text{km}$ (0.18 CFS/km); (b) total discharge to Lake Michigan from the Michigan Lowlands is $3.11 \text{ m}^3/\text{sec}$ (110 CFS) averaging $0.024 \text{ m}^3/\text{sec}/\text{km}$ (0.84 CFS/km); (c) total discharge to northern Lake Michigan is $18 \text{ m}^3/\text{sec}$ (640 CFS) averaging $0.040 \text{ m}^3/\text{sec}/\text{km}$ (1.40 CFS/km).

obtained a value of $6.4 \times 10^{-3} \text{ m}^3/\text{sec}$ (0.23 cfs) per km for the Michigan lowlands, approximately 25% of our value for the same region, and compared her results to values of Cherkauer and Hensel (1986) for the Wisconsin shoreline, ranging from 6.7 to $10.2 \times 10^{-3} \text{ m}^3/\text{s}$ per km; and Boutt et al. (2001) obtained a value of 0.23 m^3/s (8 cfs) for an unstated length of shoreline ranging between 100 and 200 km in Grand Traverse Bay. In summary, our estimate of ground water discharge from the shoreline to the Great Lakes exceeds most previous estimates.

Conclusions

Direct ground water flow to the Great Lakes is between 4% and 7% of the model outflow budget. The corresponding 29.5 m^3/sec (1041 cfs) to 48.8 m^3/s (1725 cfs) is from 33% to 54% of the legally allowed Chicago diversion of 90.6 m^3/sec (3200 cfs) from Lake Michigan, which underscores the great importance of ground water discharge to the management and study of the lakes. A 10% increase in recharge to account for underflow could increase the direct discharge to 14% of the model outflow budget. Topographic lowland areas correspond to regions of low effective transmissivity and low gradients and contribute far less ground water discharge to the Great Lakes than upland areas.

Suggestions for Future Work

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 predevelopment 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 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 in, 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 impact of Michigan Basin brines on shallow ground water quality (Wahrer et al. 1996; Meissner et al. 1996; Ging et al. 1996) and the effect of ground water discharge on water quality in the Great Lakes (Kolak et al. 1999). Dynamic variable density could be simulated by application of a fully three-dimensional, variable density ground water code. The current fresh water 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 water and deeper bedrock brines.

The Glaciofluvial Aquifer needs to be subdivided into multiple units and modeled 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.

A regional 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.

Acknowledgments

We would like to thank Theo Olsthoorn and our anonymous reviewers who greatly improved the manuscript with their comments and suggestions. We would like to thank David Westjohn for invaluable guidance through complex Michigan Basin geology, and David Holtschlag for priceless 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. John Hoaglund would like to thank Rose Bohn, the U.S. Geological Survey, The Michigan State University, The University of Michigan, The Pennsylvania State University, and NSF Grant ATM-9972956 for various forms of support during this undertaking.

References

- Barton, G.C., R.J. Mandle, and M.A. Baltusis. 1996. Predevelopment freshwater heads in the Glaciofluvial, Saginaw, and Marshall aquifers in the Michigan Basin. USGS Open-File Report 95-319.
- Bergstrom, R.E., and C.F. Hanson. 1962. Ground water supplies in Wisconsin and Illinois adjacent to Lake Michigan. In *Symposium on Great Lakes Basin*, ed. H.J. Pincus. Publication No. 71. Washington, D.C.: American Association for the Advancement of Science.
- Boutt, D.F., D.W. Hyndman, B.C. Pijanowski, and D.T. Long. 2001. Identifying potential land use-derived solute sources to stream baseflow using ground water models and GIS. *Ground Water* 39, no. 1: 24-34.
- Cartwright, K., C.S. Hunt, G.M. Hughes, and R.D. Brower. 1979. Hydraulic potential in Lake Michigan bottom sediments. *Journal of Hydrology* 43, 67-78.
- Cherkauer, D.S., and B.R. Hensel. 1986. Ground water flow into Lake Michigan from Wisconsin. *Journal of Hydrology* 84, nos. 3/4: 261-271.
- Cherkauer, D.S., and P.F. McKereghan. 1991. Ground-water discharge to lakes: Focusing in embayments. *Ground Water* 29, no. 1: 72-80.
- Croley, T.E., II, and T.S. Hunter. 1994. Great Lakes monthly hydrologic data. NOAA Technical Memorandum GLERL-83. Ann Arbor, Michigan: Great Lakes Environmental Research Laboratory.
- Eschman, D.J. 1985. Summary of the Quaternary history of Michigan, Ohio, and Indiana. *Journal of Geological Education* 33, 161-167.
- Fenneman, N.M. 1938. *Physiography of Eastern United States*. New York: McGraw-Hill.

- Ging, P.B., D.T. Long, and R.W. Lee. 1996. Selected geochemical characteristics of ground water from the Marshall Aquifer in the central Lower Peninsula of Michigan. USGS Water-Resources Investigations Report 94-4220.
- Grannemann, N.G., and T.L. Weaver. 1998. An annotated bibliography of selected references on the estimated rates of direct ground-water discharge to the Great Lakes. USGS Water-Resources Investigations Report 98-4039.
- Harbaugh, A.W. 1990. A computer program for calculating sub-regional water budgets using results from the U.S. Geological Survey modular three-dimensional finite-difference ground-water flow model. USGS Open File Report 90-0392.
- Hill, M.C. 1992. A computer program (MODFLOWP) for estimating parameters of a transient, three-dimensional, ground water flow model using non-linear regression. USGS Open-File Report 91-489.
- Hill, M.C. 1994. Five computer programs for testing weighted residuals and calculating linear confidence and prediction intervals on results from the ground water parameter estimation program MODFLOWP. USGS Open File Report 93-481.
- Hoaglund, J.R. 1996. Recharge to discharge ground water travel times in the Michigan Basin and the effect of glacial ice loading. Ph.D. diss., Department of Geological Sciences, Michigan State University-East Lansing.
- Hoaglund, J.R., G.C. Huffmann, and N.G. Grannemann. 2002. Simulation of ground water flow in the Glaciofluvial, Saginaw, Parma-Bayport, and Marshall aquifers, central Lower Peninsula of Michigan. USGS Open-File Report 00-504, in press.
- Holtschlag, D.J. 1996. A generalized estimate of ground-water recharge rates in the Lower Peninsula of Michigan. USGS Open File Report 96-593.
- Holtschlag, D.J. 1997. A generalized estimate of ground-water recharge rates in the Lower Peninsula of Michigan. USGS Water Supply Paper 2437.
- Horton, R.E. 1945. Erosional development of streams and their drainage basins, hydrophysical approach to quantitative morphology. *Geological Society of America Bulletin* 56, no. 3: 275-370.
- Kolak, J.J., D.T. Long, J.M. Matty, G.J. Larson, D.F. Sibley, and T.B. Councell. 1999. Ground-water, large-lake interactions in Saginaw Bay, Lake Huron: A geochemical and isotopic approach. *Geological Society of America Bulletin* 111, no. 2: 177-188.
- Lillienthal, R.T. 1978. Stratigraphic cross-sections of the Michigan Basin. Michigan Department of Natural Resources, Geological Survey Division, Report of Investigation 19.
- Lusch, D.P., and W. Enslin. 1984. Microcomputer based statewide digital land surface information. In *Proceedings of the Pecora IX Symposium on Spatial Information Technology for Remote Sensing, Today and Tomorrow*, Sioux Falls, South Dakota, 40-43. Bethesda, Maryland: American Society of Photogrammetry.
- Mandle, R.J., and D.B. Westjohn. 1989. Geohydrologic framework and ground water flow in the Michigan Basin. *American Water Resources Association Monograph Series* 13, 83-110.
- Mau, D.P., and T.C. Winter. 1997. Estimating ground water recharge from streamflow hydrographs for a small mountain watershed in a temperate humid climate, New Hampshire, USA. *Ground Water* 35, no. 2: 291-304.
- McDonald, M.G., and A.W. Harbaugh. 1988. A modular three-dimensional finite-difference ground water-flow model. USGS Techniques of Water-Resources Investigations, Book 6, Chapter A1.
- Meissner, B.D., D.T. Long, and R.W. Lee. 1996. Selected geochemical characteristics of water from the Saginaw Aquifer in the Michigan Basin. USGS Water-Resources Investigations Report 93-4220.
- Nauta, R. 1987. A three-dimensional ground-water flow model of the Silurian Dolomite Aquifer of Door County, Wisconsin. M.S. thesis, Department of Geology and Geophysics, University of Wisconsin-Madison.
- Quinn, F.H. 1992. Hydraulic residence times for the Laurentian Great Lakes. *Journal of Great Lakes Research* 18, no. 1: 22-28.
- Quinn, F.H. 1999. Personal communication.
- Sellinger, C.E. 1995. Ground water flux into a portion of eastern Lake Michigan. *Journal of Great Lakes Research* 21, 53-63.
- U.S. Geological Survey. 1974. Hydrologic unit map: 1974, State of Michigan. Reston, Virginia: USGS.
- Vugrinovich, R.G. 1986. Patterns of regional subsurface fluid movement in the Michigan Basin. Michigan Geological Survey Open File Report 86-6.
- Wahrer, M.A., D.T. Long, and R.W. Lee. 1996. Selected geochemical characteristics of ground water from the Glaciofluvial Aquifer in the central Lower Peninsula of Michigan. USGS Water-Resources Investigations Report 94-4017.
- Western Michigan University. 1981. Hydrogeologic atlas of Michigan: U.S. Environmental Protection Agency Underground Injection Control Program Report. Department of Geology, College of Arts and Sciences, Western Michigan University-Kalamazoo.
- Westjohn, D.B., H.W. Olsen, and A.T. Willden. 1990. Matrix-controlled hydraulic properties of Mississippian and Pennsylvanian Sandstones from the Michigan Basin. USGS Water-Resources Investigations Report 93-4152.
- Westjohn, D.B., T.L. Weaver, and K.F. Zacharias. 1994. Hydrogeology of Pleistocene glacial deposits and Jurassic "red beds" in the central Lower Peninsula of Michigan. USGS Water Resources Investigation Report 93-4152.
- Westjohn, D.B., and T.L. Weaver. 1996a. Hydrogeology of Pennsylvanian and Late Mississippian rocks in the central Lower Peninsula of Michigan. USGS Water-Resources Investigations Report 94-4107.
- Westjohn, D.B., and T.L. Weaver. 1996b. Hydrogeologic framework of Mississippian rocks in the central Lower Peninsula of Michigan. USGS Water-Resources Investigations Report 94-4246.
- Westjohn, D.B., and T.L. Weaver. 1996c. Configuration of freshwater/saline-water interface and geologic controls on distribution of freshwater in a regional aquifer system, central Lower Peninsula of Michigan. USGS Water-Resources Investigations Report 94-4242.
- Westjohn, D.B., and T.L. Weaver. 1998. Hydrogeologic framework of the Michigan Basin regional aquifer system. USGS Professional Paper 1418.