Modeling the subglacial hydrology of the late Pleistocene Lake Michigan Lobe, Laurentide Ice Sheet

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ABSTRACT

The fast ice flow characteristic of the low-gradient Lake Michigan Lobe of the southern Laurentide Ice Sheet during the last glaciation likely resulted from some combination of subglacial sediment deformation and sliding at the ice-bed interface. Both mechanisms require high basal water pressure relative to overlying ice pressure. To evaluate the basal water pressures of the Lake Michigan Lobe, we used the groundwater model MODFLOW to simulate groundwater flow along a 1040-km-long flow line corresponding to the flow path taken by the ice lobe. Our simulations show that groundwater flow directions and velocities were substantially different than modern conditions. Simulations also indicate that subglacial aquifers were not capable of evacuating the estimated basal meltwater, but that excess water draining through a basal drainage system such as one that may underlie Ice Stream B, West Antarctic Ice Sheet, would prevent basal water pressure from exceeding the ice-overburden pressure. The buried Mahomet bedrock-valley system near the distal end of the Lake Michigan Lobe could have drained enough subglacial water to substantially lower basal water pressure and thus affect the dynamics of the lobe. By draining basal meltwater away from the Lake Michigan Lobe, the buried Mahomet bedrock-valley system may have attenuated the effect of permafrost on basal water pressure. During the Last Glacial Maximum, groundwater-flow patterns and velocities were altered or reversed relative to present conditions.

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INTRODUCTION

The Lake Michigan Lobe was a dynamic feature of the Laurentide Ice Sheet during the last glaciation. The lobe had an extremely low surface profile (Clark, 1992), yet sustained fast ice flow (Mickelson et al., 1983; Hansel and Johnson, 1992), suggesting that the lobe flowed over a low-friction bed. Such bed conditions most likely occurred through some combination of subglacial sediment deformation and decoupling of the ice from its bed (Alley, 1991). Both mechanisms are dependent on high subglacial water pressure relative to ice pressure and are, therefore, critically dependent on the character and behavior of the subglacial drainage system (Alley, 1989).

Here, we use the groundwater model MODFLOW (McDonald and Harbaugh, 1988) to estimate the water pressure at the base of the Lake Michigan Lobe that would have influenced mechanisms of fast ice flow during the last glaciation. During the Last Glacial Maximum, the Lake Michigan Lobe originated from an ice divide near James Bay and flowed south through the Lake Michigan basin to terminate at ∼40°N (Fig. 1). We first explore the likely groundwater-pressure distribution and flow patterns along a flow line parallel to the central trajectory of the Lake Michigan Lobe that extends from Lake Superior to the Mississippi River, ∼350 km south of the lobe terminus (Fig. 1). Our flow line includes a hydrostratigraphic description of the major hydrogeologic units that existed beneath the Lake Michigan Lobe, and we perform a number of sensitivity tests to assess the capacity of these units to transmit estimated meltwater fluxes. We also investigate plausible drainage systems that may have existed at the ice-bed interface and evaluate the potential impact of permafrost on the subglacial drainage system. Finally, we compare simulated glacial-stage groundwater-flow patterns and velocities to the groundwater conditions that exist today.

PREVIOUS WORK

Previous model results and data demonstrate that former glaciation in the central United States profoundly affected regional groundwater chemistry and groundwater-flow patterns. Geochemical ($\delta^{18}O$) anomalies in the Fox Hills aquifer of North Dakota suggest that modern discharge areas served as recharge zones beneath former ice sheets during one or more glacial periods (Carlson, 1994). On the basis of a regional groundwater and particle-tracking model, Hoaglund (1996) showed that ice loading may have caused groundwater-flow reversals in Michigan in which a strong downward component of flow existed beneath the former ice sheet and an upward component fed into a large proglacial lake. Similar research has not been performed in the region formerly overlain by the Lake Michigan Lobe. Nevertheless, because the extent and timing of glaciation in North Dakota and Michigan were similar to the extent and timing in Illinois, it is reasonable to expect that the Lake Michigan Lobe may have similarly affected regional groundwater-flow patterns.

Numerical simulations also indicate that groundwater velocity and flow directions in northern Europe were altered or, in some cases, reversed under the influence of the Fennoscandian Ice Sheet. Boulton et al. (1995) suggested that subglacial aquifers underlying the southern Fennoscandian Ice Sheet had the capacity to transmit all of the basal meltwater. In contrast, Piotrowski (1997a, 1997b) concluded that the aquifers in northwestern Germany were not capable of transmitting the estimated basal meltwater discharge, thus

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requiring subglacial meltwater to flow through tunnel valleys, probably during spontaneous outburst events. These different results likely reflect the fact that Boulton et al. (1995) did not explicitly include glacial drift in their model because they assumed that the glacial drift had little influence on the transmission of subglacial meltwater.

REGIONAL GEOLOGY

The bedrock geology beneath our modeled flow line includes a southward-thickening sequence of sedimentary units that overlies crystalline basement (Fig. 2). In Lake Michigan, the bedrock dips gently to the east toward the Michigan basin. South of Lake Michigan, bedrock generally dips to the south, ultimately reaching a thickness of >7000 m in the Illinois basin.

Currently, a sequence of glacial drift as thick as 200 m overlies bedrock along the flow line south of Lake Superior. Because we are concerned with the Lake Michigan Lobe at the Last Glacial Maximum, we do not include drift that is younger than the Tiskilwa Till, the oldest member of the late Wisconsinan Wedron Formation (Hansel and Johnson, 1996). Tiskilwa Till and older drift are absent or undocumented along the flow line between Lake Superior and a point ~130 km south of Chicago. It is reasonable to assume, however, that some quantity of drift was present in this region during the Last Glacial Maximum. We thus include 1 m of till over the bedrock from the southern shore of Lake Superior to a point ~130 km south of Chicago (more till may have overlain the bedrock in this area; however, any assumption of thicker till will result in increased subglacial water pressure). On the basis of data maintained by the Illinois State Geological Survey, we model the Tiskilwa Till south of this point as a wedge-shaped deposit with a maximum thickness of 15 m at the lobe terminus. Along the flow line beyond the lobe terminus, we assume a 15 m drift thickness for older Illinoian drift deposits (Soller, 1997).

The Mahomet bedrock valley in central Illinois (Fig. 1) is a complex lowland eroded into Pennsylvanian and older rocks underlying glacial drift and filled with fluvial deposits that are the most extensive and productive sand and gravel aquifers in the southern three quarters of Illinois (Kempton et al., 1991). The Mahomet bedrock valley is between 13 and 22 km wide, the width generally increasing toward the west. It traverses central Illinois roughly in an east-west direction, ultimately terminating in the Mississippi River Valley.

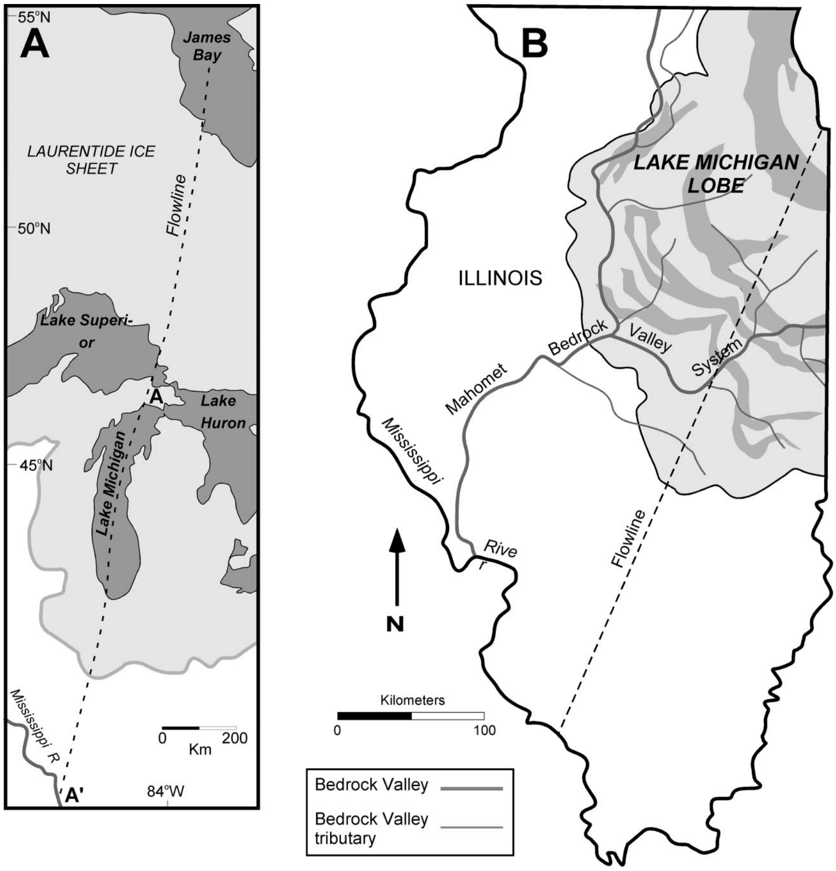

Figure 1. (A) Simulated flow line (dashed line). The ice limit is shown with ice cover represented by the gray pattern. A and A′ show the boundaries of the finite-difference grid. (B) Location of the Lake Michigan Lobe and buried bedrock valleys in Illinois.

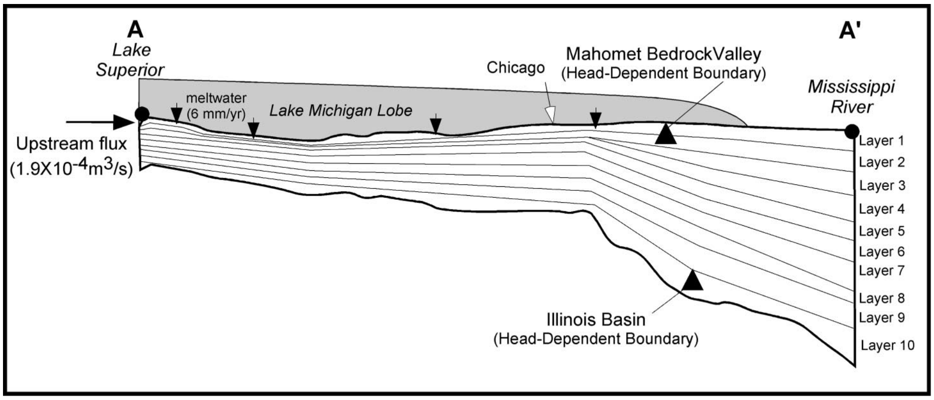

Figure 2. Generalized model geometry and boundary conditions along flow line A–A′ (see Fig. 1A). Horizontal length is 1040 km. Vertical exaggeration is 100×. Illustrated stratigraphy shows 10 modeled hydrostratigraphic units (layers; not to scale) (see Table 1 for description).

HYDROSTRATIGRAPHY

We grouped the geologic formations underlying the Lake Michigan Lobe flow line into six regional aquifer units and four regional confining units on the basis of their hydraulic characteristics (Fig. 2; Table 1). Each unit is represented in the model by 1040 cells, and each cell represents a region that is 1000 m long in the direction parallel to the flow line and a variable vertical distance, defined by the thickness of the hydrostratigraphic unit. The drift is modeled as two distinct layers to increase the model resolution directly beneath the ice: The top drift layer is defined as 0.5 m thick, and the lower drift layer thickness is equal to the total drift thickness minus 0.5 m. We estimated the extent and thickness of the bedrock hydrostratigraphic units by examination of borehole logs, stratigraphic cross sections, and literature published by the Illinois State Geological Survey, the Wisconsin Geological and Natural History Survey, the Michigan Department of Natural Resources, and the U.S. Geological Survey.

We assigned hydraulic conductivity (K) values to bedrock aquifers and confining units on the basis of the estimates of Mandle and Kontis (1992) (Table 2). Because the hydraulic conductivity values applied to bedrock layers in this model are based on a regional-scale model of the Northern Midwest aquifer system (Mandle and Kontis, 1992), the conductivity in our simulations should not be subject to scaling errors. Each unit is characterized as isotropic.

The K value assigned to the drift layers (3.7×10^{-7} m/s) is based on a statewide average for Illinois (Walton, 1965). The K value for drift may have been lower than 3.7×10^{-7} m/s; however, a lower assumed K would yield qualitatively identical results. Glacial drift is composed of cobble- through clay-sized fractions, the hydraulic conductivity of which may vary by several orders of magnitude (Davis, 1969). Drift may be well sorted to unsorted and may be laterally continuous or discontinuous. These characteristics make the hydraulic conductivity of drift extremely difficult to quantify.

Our aquifer characterizations are primarily based on data gathered from hydraulic tests performed under current interglacial field conditions. Some hydraulic characteristics, however, may have been different during the glacial maximum. Hydraulic conductivity, for example, is inversely related to the viscosity of the fluid. A reduction of water temperature, therefore, leads to a reduction of hydraulic conductivity. Today, in central Illinois, the mean annual air temperature is ~15 °C. Near-surface groundwater typically has a temperature approximately equal to the mean annual air temperature. Subglacial meltwater, however, would be close to 0 °C, resulting in a viscosity that is ~60% greater than that of water at 15 °C. During the Last Glacial Maximum, therefore, K values near the ground surface were probably reduced relative to modern values.

TABLE 1. MODEL HYDROSTRATIGRAPHY

Model layer	Hydrostratigraphic unit
1	Glacial drift; 0.5 m
2	Glacial drift; variable thickness
3	Pennsylvanian–Middle Devonian limestone, shale, sandstone, siltstone, and coal
4	Silurian and Devonian carbonate rocks
5	Maquoketa Shale; Galena Dolomite; and limestone, dolomite, sandstone, and shale of the Decorah, Platteville, and Glenwood Formations
6	St. Peter Sandstone; limestone, dolomite, and discontinuous layers of sandstone, siltstone, and shale of the Prairie du Chien Group; and Jordan Sandstone
7	Shaly sandstone, limestone, and dolomite of the St. Lawrence and Franconia Formations
8	Ironton and Galesville Sandstones
9	Eau Claire Formation (mainly sandstone)
10	Mount Simon Sandstone and Elmhurst Sandstone

TABLE 2. HYDRAULIC CONDUCTIVITY VALUES IN ILLINOIS

Layer	K (m/s) assigned to the validation simulation	Geometric mean (m/s)	-1σ (m/s)	$+1\sigma$ (m/s)	Total observations
Quaternary drift 1	3.7×10^{-7}		N.A.		
Quaternary drift 2	3.7×10^{-7}		N.A.		
Pennsylvanian through Middle Devonian rocks	7.6×10^{-11}		N.A.		
Silurian–Devonian carbonates	1.4×10^{-5}	9.0×10^{-5}	4.6×10^{-6}	1.6×10^{-4}	1816
Maquoketa Shale, Galena Dolomite, and the Decorah, Platteville, and Glenwood Formations	1.8×10^{-10}		N.A.		
St. Peter Sandstone, Prairie du Chien Group, and the Jordan Sandstone	1.5×10^{-5}	1.0×10^{-5}	1.6×10^{-6}	6.4×10^{-5}	539
St. Peter Sandstone, Prairie du Chien Group, and the Jordan Sandstone (southern third of the flow line)	6.1×10^{-5}	1.0×10^{-5}	1.6×10^{-6}	6.4×10^{-5}	539
St. Lawrence and Franconia Formations	2.4×10^{-9}		N.A.		
St. Lawrence and Franconia Formations (southern third of the flow line)	3.0×10^{-11}		N.A.		
Ironton-Galesville Sandstones	2.1×10^{-4}	4.3×10^{-5}	1.2×10^{-5}	1.5×10^{-4}	58
Ironton and Galesville Sandstones (southern quarter of the flow line)	1.5×10^{-4}	4.3×10^{-5}	1.2×10^{-5}	1.5×10^{-4}	58
Eau Claire Formation	3.0×10^{-10}		N.A.		
Eau Claire Formation (southern third of the flow line)	2.4×10^{-10}		N.A.		
Mount Simon Sandstone and Elmhurst Sandstone	1.1×10^{-4}	1.8×10^{-5}	2.4×10^{-6}	1.4×10^{-4}	99

Note: All K values except those of Quaternary drift are based on the best estimates of the Northern Midwest Regional Aquifer System Assessment (Mandle and Kontis, 1992). Standard deviation data are available only for the bedrock aquifer units. K values assigned to the confining units are close to the values estimated by Walton (1960) and Young (1976). Quaternary drift K is one order of magnitude less than Walton's (1965) estimate for diverse drift deposits in Illinois.

THE GROUNDWATER MODEL

Groundwater flow was simulated by using MODFLOW, a three-dimensional finite-difference code describing steady-state or transient groundwater flow (McDonald and Harbaugh, 1988). Three-dimensional groundwater flow is described by the following equation, combining Darcy's law with mass conservation (Domenico and Schwartz, 1990):

$$\nabla \cdot (K\nabla h) - w = S_s \frac{\partial h}{\partial t} \quad (1)$$

where K [L/T] is the hydraulic conductivity, h [L] is hydraulic head, w [1/T] is a source-sink term, and S_s [1/L] is specific storage. In the case of steady-state flow (the first simulations described here), $\partial h/\partial t = 0$. Hydraulic conductivity is defined by

$$K = \frac{k\rho g}{\mu} \quad (2)$$

where k [L^2] is permeability, ρ [M/L^3] is the density of water, g [L/T^2] is the acceleration of gravity, and μ [M/L/T] is the viscosity of water. Equation 1 is solved by a finite-difference method, which is the basis of our numerical simulations.

We applied the model to a 1-m-wide flow line that is parallel to ice flow, assuming that all water flow occurs parallel to the transect. Most geologic material is characterized by hydraulic conductivity that is approximately isotropic in the horizontal plane. Because the subglacial head gradient is largely defined by the ice-surface gradient (Paterson, 1994), the horizontal component of groundwater flow runs approximately parallel to the direction of ice flow. Some amount of water may have flowed in directions oblique to our flow line, but in general, that flow should not have a significant impact on our results. Oblique flow would probably be limited to the area south of Lake Michigan, where the Lake Michigan Lobe extended beyond the main mass of the Laurentide Ice Sheet. Owing to the roughly circular shape of the Lake Michigan Lobe terminus, the distance from the model flow line to the southern terminus is generally similar to the distance to other lobe margins. Illinois hydrostratigraphy does not change substantially in the regions paralleling the flow line, suggesting that water flowing oblique to the modeled flow line would encounter an aquifer system similar to that along the flow line. Oblique flow systems would thus require a similar head gradient to evacuate the meltwater and thus would not significantly affect our results.

For those locations where groundwater may flow in directions oblique to the simulated flow line of our model, we apply a gradient-dependent source-sink term to estimate transverse flow:

$$w = \frac{K}{d}\frac{\partial h}{\partial n} \tag{3}$$

where d [L] is the aquifer thickness and n [L] is a coordinate oblique to the flow line. This source-sink term, which is included in equation 1, allows the calculation of flow into or out of a specified cell, on the basis of the difference between the computed head in the specified cell and a constant-head value defined for an outside source or sink.

We utilize a gradient-dependent source-sink term in two locations, representing the deep aquifers of the Illinois basin and the buried Mahomet bedrock-valley system (Fig. 2). Generally, the deep aquifers in the Illinois basin are recharged at the basin margins in Missouri, Indiana, and Kentucky. Deep groundwater discharges by upward leakage in the basin, possibly along faults. The recharge areas for the Illinois basin do not occur along our modeled flow line, but discharge does occur in the southern part of our flow line. To account for recharge to the deep basin aquifers, we have assigned a gradient-dependent source-sink term to the Mount Simon aquifer in the Illinois basin (layer 10 in Fig. 2). The Mount Simon aquifer crops out in northwest Missouri ~90 km west of the modeled flow line at altitudes up to 500 m, but more generally at ~300 m. We assume that 300 m adequately describes the head in the Mount Simon in its recharge zone. The thickness of the aquifer is ~300 m, and the hydraulic conductivity is $\sim 6 \times 10^{-5}$ m/s. We base our gradient-dependent source-sink term on these values.

Much of the Mahomet bedrock-valley system underlies the area formerly overlain by the Lake Michigan Lobe (Fig. 1). However, the discharge area for the valley system lies to the west of the ice margin, suggesting that the Mahomet bedrock valley may have functioned as a subglacial meltwater conduit that discharged to the west in nonglaciated areas. To investigate the influence of the Mahomet bedrock valley on subglacial hydrology, we assigned a gradient-dependent source-sink term to those model cells in the layer 2 drift aquifer that encounter the Mahomet bedrock-valley system (Fig. 2). The modeled flow line intersects the Mahomet bedrock valley ~100 km east of the western boundary of the Lake Michigan Lobe; thus the length (n) term applied to equation 3 is 100 km. Currently, the potentiometric surface in the Mahomet Sand aquifer roughly parallels the surface of the Mahomet Sand. We thus assume that the head in the Mahomet aquifer at the ice-sheet boundary was equal to the modern ground surface altitude (~200 m). The transmissivity of the Mahomet Sand and other aquifers in the Mahomet bedrock valley is $\sim 7 \times 10^{-3}$ m^2/s (Kempton et al., 1991). These aquifers predate the Last Glacial Maximum. We assume a constant 14 km width for the bedrock valley. Accordingly, a gradient-dependent source-sink term is applied to each of the 14 cells that represent the layer 2 drift aquifer overlying the thalweg of the Mahomet bedrock valley. By limiting the valley width to 14 km, we intended to avoid overestimating the capacity of the bedrock-valley aquifers to transmit meltwater.

BOUNDARY CONDITIONS

For our upstream boundary condition, we assume that all of the meltwater generated along the flow line north of Lake Superior arrived at the ice-bed interface. This simplification is justified because the ice sheet rested directly on crystalline rock over and north of Lake Superior, forcing groundwater driven by an ice-pressure gradient to flow at the ice-bed interface and/or through a very thin layer of drift.

We assume that basal melting occurred along the entire length of the flow line that was overlain by ice. We assign a basal melt rate of 6 mm/yr, a value that is typical of modern ice sheets (Drewry, 1986) and is within reasonable bounds given by the regional geothermal gradient and the latent heat of fusion for ice. With this basal melt rate and a 1000 km flow-line length upstream of Lake Superior, we estimate an upstream meltwater flow rate of 1.9×10^{-4} m^3·m^{-1}·s^{-1}. Precipitation recharge is not added to the model because we assume that the ice is impermeable. The flux of basal meltwater may have been significantly >6 mm/yr owing to surface melting, precipitation, and frictional melting; however, by assuming a 6 mm/yr basal melt rate, our simulations offer a conservative estimate of the capacity of the basal aquifers to transmit meltwater.

Because the location and surface altitude of the Mississippi River was similar to its current position during the Last Glacial Maximum, we assume that the surface altitude of the modern Mississippi River adequately describes head in the drift and the uppermost bedrock layer at the southern boundary of our flow line (Fig. 2). The entire length of the modeled flow line is underlain by Precambrian crystalline basement rock. We define the top of this basement as an impermeable boundary. At the surface, boundary conditions depend on the presence or absence of ice. On the part of the flow line covered by the ice sheet, the ice is treated as an impermeable layer. South of the lobe terminus, the top of the model is treated as an unconfined aquifer.

MODEL VALIDATION

By using modern nonglacial boundary conditions, we simulate steady-state head values and surface-discharge volumes that are close to measured modern values, suggesting that the model functions as a reasonable approximation of the hydrogeologic system along the modeled flow line. Modern boundary conditions were applied as follows. (1) A constant head was applied to the drift layers and the Silurian–Devonian aquifer underlying Lake Michigan. The head was set equal to the modern surface altitude of the lake (176 m). (2) A constant head was applied to the northern boundary drift layers and in the underlying Silurian–Devonian layer, representing Lake Superior (184 m). (3) A constant head was applied to the southernmost boundary of the glacial drift and the underlying Pennsylvanian–Devonian confining layer representing the surface altitude of the Mississippi River (110 m). (4) A constant recharge rate of 38 cm·yr^{-1} was applied to the top layer of the model to simulate precipitation recharge. The value 38 cm·yr^{-1} is based on the average soil-moisture surplus for the period 1967–1988 in east-central Illinois (Cravens et al., 1990). The stratigraphy and hydraulic characteristics remain unchanged.

Glacial drift younger than the Tiskilwa Till was not added to the validation-model stratigraphy because (1) the younger till occupies a very small percentage of the stratigraphic cross section and, therefore, is likely to have relatively little effect on regional groundwater-flow patterns, (2) data describing the vertical and horizontal extent of younger till are limited, and (3) the number and quality of data describing the hydraulic properties of the till are limited.

The steady-state solution computed for the validation model was compared to published maps of field-measured hydraulic head (Mandle and Kontis, 1992). Where field-measured data are not available or data are very sparse, such as underneath Lake Michigan and in the very deep aquifers of the Illinois basin, simulation results were compared to values generated by the regional aquifer system analysis (RASA) model for the Northern Midwest aquifer system (Mandle and Kontis, 1992).

The validation simulation generates a head solution within 65 m of the measured and simulated head values (Mandle and Kontis, 1992) (Fig. 3). More generally, validation-model results are within 25 m of the RASA values. The greatest discrepancies occur in the Mount Simon Sandstone and St. Peter Sandstone–Prairie du Chien Group. In the Mount Simon Sandstone, relatively large differences occur in the Illinois basin at the southern end of the flow line. Head measurements in the Mount Simon Sandstone in the Illinois basin are extremely sparse. To compare our results to field-measured data, we rely on relatively large extrapolations between field-measured points. This approach may cause some of the divergence between the field data and our results. This numerical divergence is most prevalent in the deepest aquifer, south of the ice terminus, and thus should have an insignificant effect on our simulations.

Simulation head values are up to 65 m greater than measured head values in the St. Peter–Prairie du Chien aquifer. The maximum divergence occurs ~800 km south of the flowline starting point. This difference suggests that our model fails to account for discharge mechanisms that are currently operating. Regardless of the exact causes of the discrepancy, the head difference is relatively small given the scale of our simulations and the level of detail that we are concerned with.

Simulated head shows far greater variability than measured head in the Silurian–Devonian aquifer (Fig. 3), which is an artifact of the scales of measurement. Simulation results are calculated at 1 km intervals, whereas RASA results are calculated at 25 km intervals. Field data are reported at random intervals.

The model can be further validated by comparing the quantity of simulated surface discharge to measured modern surface-runoff values in Illinois. The Vermillion River drains 3341 km^2 in central Illinois. For the years 1978–1998, U.S. Geological Survey stream gage data indicate that the Vermillion River had an average discharge of 37 m^3·s^{-1}, or 1.1 $\times$ 10^{-8} m^3·m^{-2}·s^{-1} (cubic meters of water per meter squared of land surface per second). Our simulated surface-discharge value for this area is 1.2 $\times$ 10^{-8} m^3·m^{-2}·s^{-1}.

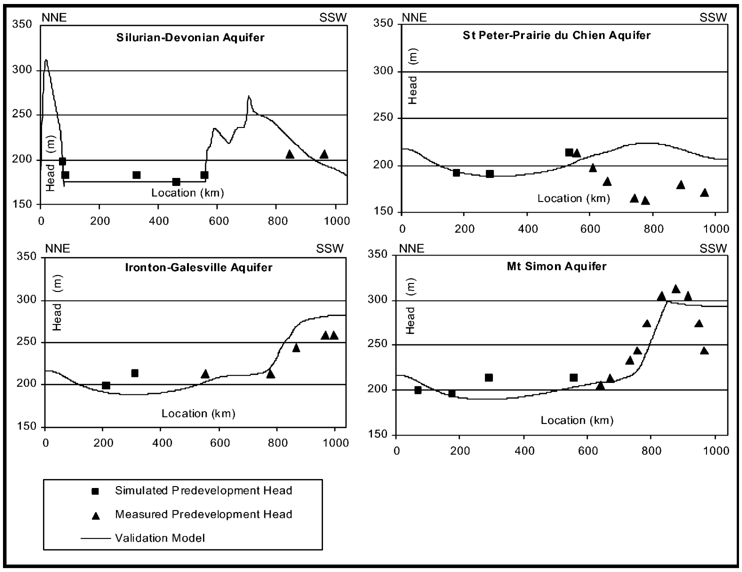

Figure 3. Comparison of validation-model groundwater head to measured and simulated head for four different aquifers. (A) Silurian–Devonian aquifer. (B) St. Peter–Prairie du Chien aquifer. (C) Ironton-Galesville aquifer. (D) Mount Simon aquifer.

RESULTS

Our primary goal in examining water pressure beneath the Lake Michigan Lobe is to examine those conditions by which the effective pressure (the difference between ice pressure and water pressure) at the ice-bed interface remains greater than zero and thus prevents ice flotation. If effective pressure reached zero, the ice would become decoupled from its bed, leading to catastrophic collapse of the lobe. On the basis of numerical reconstructions of the Lake Michigan Lobe, we consider the likely ice thickness at Lake Superior to have ranged from 1200 to 2000 m (Licciardi et al., 1998); we refer to these endpoints later as ''thin-ice'' and ''thick-ice'' conditions, respectively. We then (1) evaluate the capacity of the subglacial aquifers to transmit the estimated subglacial meltwater of the Lake Michigan Lobe, (2) estimate the water pressure at the ice-bed interface under likely geologic and hydrologic conditions, (3) investigate the potential effect of permafrost on subglacial hydrologic processes, and (4) compare modern versus glacial-stage hydrogeologic flow patterns and magnitudes.

We performed seven sensitivity tests (Table 3) to identify those conditions under which the water pressure at the base of the Lake Michigan Lobe would remain equal to or less than the ice-overburden pressure. Simulations 1–4 are based on the assumption that basal meltwater does not flow along the ice-bed interface; the meltwater only flows through subglacial aquifers. Simulations 5–7 include a basal drainage system. Under boundary conditions of the Last Glacial Maximum with hydrogeologic properties set as for the validation model, the water-pressure head beneath the Lake Michigan Lobe ranges from 50 240 m at the northern boundary of the model to 2292 m at the southern terminus (Fig. 4), indicating that simulated water pressure at the ice-bed interface would have far exceeded the ice-overburden pressure for any reasonable Lake Michigan Lobe thickness. These results suggest either that the hydraulic conductivity values applied to the glacial drift and the bedrock aquifers are too low or that the transmissivity of subglacial aquifers beneath the Lake Michigan Lobe was insufficient to transmit estimated basal meltwater.

TABLE 3. SENSITIVITY TESTS

Simulation	Boundary conditions	Results
1	All *K* values remain unchanged from the values applied to the validation model. Drift *K* is 3.7 × 10^{-7}m/s. Permafrost is not modeled.	Maximum head is 50 240 m in the northernmost cell in model layer 1. Head in layer 1 is 2292 m at the ice margin.
2	Bedrock *K* remains unchanged from the validation model. Drift *K* is increased to 1 × 10^{-6} m/s. Permafrost is not modeled.	Maximum head is 34 602 m in the northernmost cell in model layer 1. Head in layer 1 is 1093 m at the ice margin.
3	Bedrock *K* remains unchanged. Drift *K* is increased to 0.037 m/s. Permafrost is not modeled.	Maximum head is 3498 m in the northernmost cell in model layer 1. Head in layer 1 at the ice margin is 177 m.
4	Bedrock *K* is increased 1–2 orders of magnitude relative to validation model values. Drift *K* is 3.7 × 10^{-7} m/s. Permafrost is not modeled.	When *K* is increased one order of magnitude, maximum head is 4707 m in the northernmost cell in model layer 1. Head in layer 1 is 702 m at the ice margin. When *K* is increased two orders of magnitude, maximum head is 816 m and 359 m.
5	Layer 1: T = 0.16–0.23 m²/s. Layer 2 drift and bedrock *K* values remain unchanged from the values applied to the validation model. Permafrost is not modeled.	Maximum head is 1447–1099 m in the northernmost cell in model layer 1.
6	This simulation is based on simulation 5 parameters, with T = 0.16 and 0.23 m²/s. A 100-km-wide band of impermeable permafrost is assumed at the ice margin.	Maximum head is 1736 m and 1418 m. Ninety-nine percent of the meltwater is removed through the porous aquifers in the Mahomet Valley.
7	*K* and *T* remain identical to simulation 6. Permafrost is increased to 50 km in the up-ice direction to cover the Mahomet Valley.	Maximum head is >50 000 m.

To test the sensitivity of the model to the value of hydraulic conductivity applied to the drift aquifer, we increased the hydraulic conductivity of the drift incrementally. By increasing the hydraulic conductivity of the glacial drift to 3.7 × 10^{-2} m/s, which is a typical value of gravel, the maximum simulated subglacial head was reduced to 3498 m (simulation 3; Table 3), a value close to, but still greater than, the likely ice-overburden pressure. Glacial till in Illinois, however, normally has *K* values that are three to nine orders of magnitude smaller (Soller and Berg, 1992), suggesting that the subglacial aquifers could not evacuate the meltwater while maintaining subglacial water pressure at a level less than the ice-overburden pressure.

To test the sensitivity of the model to bedrock *K*, we increased the hydraulic conductivity of all bedrock units by one to two orders of magnitude (Simulation 4; Table 3). Drift *K* was unchanged from the validation model. When bedrock *K* is increased one order of magnitude, the model generates 4707 m of head at the northernmost boundary of the drift aquifer, or two to four times the likely ice thickness at this location (Licciardi et al., 1998). When bedrock *K* is increased two orders of magnitude, subglacial head reaches a maximum of 816 m at the northern upper boundary and 359 m at the ice margin.

Given the conservative assumptions applied to the model, these results indicate that neither the bedrock nor the glacial drift had the capacity to transmit the basal meltwater while maintaining a subglacial head less than the ice-overburden pressure. Therefore, we suggest that some type of drainage system would have developed at the ice-bed interface in response to the high basal water pressure. Alley (1989) and Engelhardt and Kamb (1997) reached a similar conclusion for Ice Stream B of the West Antarctic Ice Sheet. Like the Lake Michigan Lobe, Ice Stream B has a low ice-surface gradient, exhibits fast flow, and rests on a bed of unconsolidated sediment overlying sedimentary bedrock units that are not capable of transmitting all of the subglacial melt (Lingle and Brown, 1987; Alley, 1989; Engelhardt and Kamb, 1997).

The specific drainage system underlying Ice Stream B remains unknown, but likely consists of some combination of a water film and a network of canals, but not R-channels (Rothlisberger channels) (Alley, 1989; Walder and Fowler, 1994; Engelhardt and Kamb, 1997). The absence of eskers in the area of the Lake Michigan Lobe similarly indicates that R-channels did not develop under the Lake Michigan Lobe (Clark and Walder, 1994). Weertman (1970) suggested that a distributed meltwater film with a minimum thickness of 0.14 mm and occupying 100% of the bed might exist at the base of Ice Stream B. Alley et al. (1989) suggested that Ice Stream B might have an ice-bed drainage system consisting of a distributed water film between 0 and 7 mm thick, occupying an unspecified part of the glacier sole. Engelhardt and Kamb (1997) found that the existence of such a film beneath the ice stream was not supported by their borehole observations, and they inferred that a system of canals spaced 50–300 m apart, and with dimensions of ~0.1 m deep and >>0.1 m wide (as described by Walder and Fowler, 1994), exists beneath the ice stream.

The transmissivity of a film at the ice-bed interface can be quantified on the basis of an equation describing flow in a fracture. A laminar fluid film flowing between two smooth parallel surfaces has a transmissivity (T) equal to (Romm, 1966)

$$T = \frac{\rho g b^3}{12\mu} \tag{4}$$

where b [L] is the fracture aperture and μ is the viscosity of water (1.798 × 10^{-3} Pa·s at 0 °C). To incorporate the concept of film in our model, we redefined layers 1 and 2 of the drift aquifer as a single layer. Above the newly redefined drift aquifer, we simulated a subglacial film with transmissivity determined by equation 4. All other model parameters remain unchanged.

Our simulations indicate that a 7 mm film reduces maximum subglacial head to 1447 m (Fig. 4) and an 8 mm film reduces head to 1099 m, suggesting that a water film with a thickness between 7 and 8 mm is sufficient to reduce subglacial head to values close to or less than the ice-flotation level for thin and thick-ice conditions (Fig. 5). When a 7 mm film is included in the simulation, 64% of the meltwater flows at the ice-bed interface, while the remaining 36% of the subglacial meltwater is directed downward into aquifers.

Water flowing through these films may not be laminar (an assumption of Darcy's law), in which case our film thicknesses may be underestimated. Our simulations are also based on the assumption that a water film occupies 100% of the ice-bed interface. Research by Walder (1994) and Engelhardt and Kamb (1997) suggests that such a film drainage system is unstable. Instead, drainage may have occurred through some type of canal system. A similar system may have existed underneath the Lake Michigan Lobe. To test this idea, we simulate the effect that drainage at the ice-bed interface has on simulated subglacial water pressure. Because the transmissivity of a laminar film is proportional to the cube of the thickness of the film (equation 4), only a very small increase in film thickness is needed to decrease the subglacial drainage area. This effect suggests that water did not drain through a film occupying 100% of the bed, but instead, may have drained through a more organized canal-type system. Because the film thickness suggested by our simulations is similar to but slightly greater than the estimates for Ice Stream B, a drainage system similar to that of

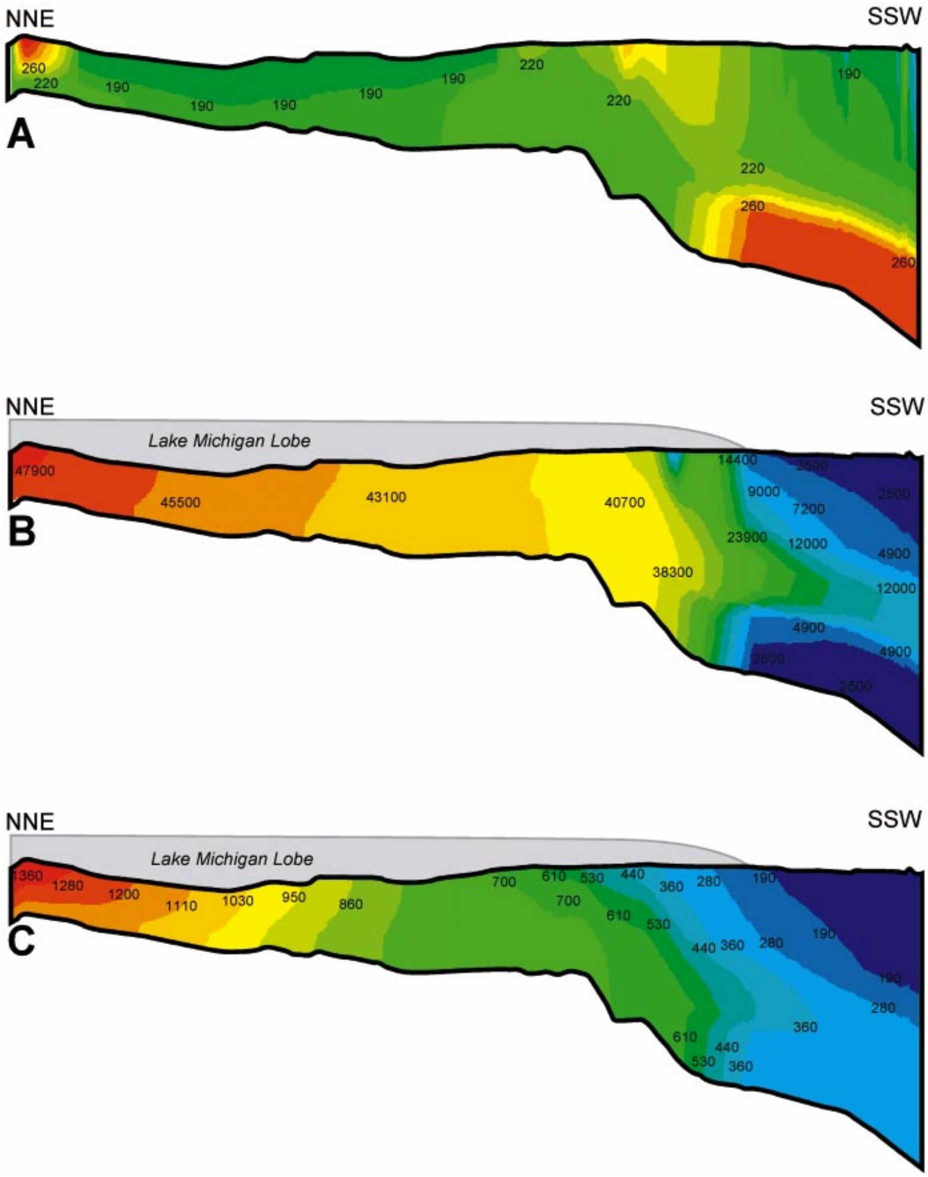

Figure 4. Simulated groundwater head (in meters) for three different conditions along modeled flow line A–A′. (A) Modern state. (B) Simulation based on modern *K* values with Lake Michigan Lobe present but no drainage system at the ice-bed interface. (C) Simulation based on modern *K* values with Lake Michigan Lobe present and a 7-mm-thick water film at the ice-bed interface.

Ice Stream B may have operated underneath the Lake Michigan Lobe.

Effect of Permafrost

Permafrost may have been widespread near and beneath the Lake Michigan Lobe margin during its last maximum extent (Johnson, 1990). Permafrost may influence subglacial hydrology by reducing the hydraulic conductivity of the frozen unit (Burt and Williams, 1976), leading to larger subglacial water pressure. Furthermore, if permafrost exists at an ice-sheet margin, then a drainage system at the ice-bed interface cannot discharge through the frozen terminal areas of the margin.

To investigate the possible hydrologic effects of a band of permafrost near the ice-sheet margin, we altered the boundary conditions describing the previous simulation by defining all of the layer 1 water film and layer 2 drift cells within 50 km of either side of the ice margin as impermeable to water (simulation 6; Table 3). The transmissivity of the film layer was set to 0.16 $m^2 \cdot s^{-1}$, the equivalent of a 7 mm layer of water. All of the other model parameters remained unchanged. Simulations based on these conditions generate a maximum head of 1736 m. When the film thickness is increased to 8 mm, the simulation generates a maximum head of 1418 m. The corresponding effective pressure would be negative for thin-ice conditions but positive for thick-ice conditions (Fig. 5).

Why does permafrost cause only a slight increase in subglacial water pressure? This result reflects the influence of the Mahomet bedrock-valley system on the subglacial hydrology. The Mahomet bedrock-valley system lies >50 km upstream from the lobe terminus (Fig. 1) and is thus beyond the range of our assumed permafrost extent. Because the Mahomet bedrock-valley system is filled with thick layers of highly permeable sediment and extends far beyond the ice margin, it had the capacity to divert water away from the permafrost, thus preventing subglacial water pressure from increasing substantially. The gradient-dependent source-sink term that represents discharge through the Mahomet bedrock valley accounts for 99% of the estimated subglacial meltwater, indicating that the Mahomet bedrock valley effectively drained subglacial water away from the lobe terminus and prevented the development of high subglacial heads.

When the width of simulated permafrost is increased by an additional 50 km in the up-ice direction, it completely covers and fills the simulated Mahomet bedrock valley. Under these conditions, the model generates subglacial heads of >50 000 m, further illustrating the importance of an unfrozen Mahomet valley in maintaining low subglacial water pressure. If the valley was frozen, the mechanism that could reduce the subglacial water pressure is unclear. Cutler et al. (2000) suggested that permafrost underneath the neighboring Green Bay Lobe may have led to the development of subglacial lakes, and they speculated that those lakes may have been released occasionally during sudden jökulhlaup-type flood events. The lack of large-scale flood features in Illinois, however, suggests that such floods did not occur.

Transient Simulation

All of the simulations that we have outlined indicate that groundwater-flow patterns beneath the Lake Michigan Lobe were significantly altered relative to modern conditions. Here we examine the time required to change from steady-state nonglacial to glacial conditions by performing a series of transient simulations based on an aquifer system that includes a 7 mm film at the ice-bed interface

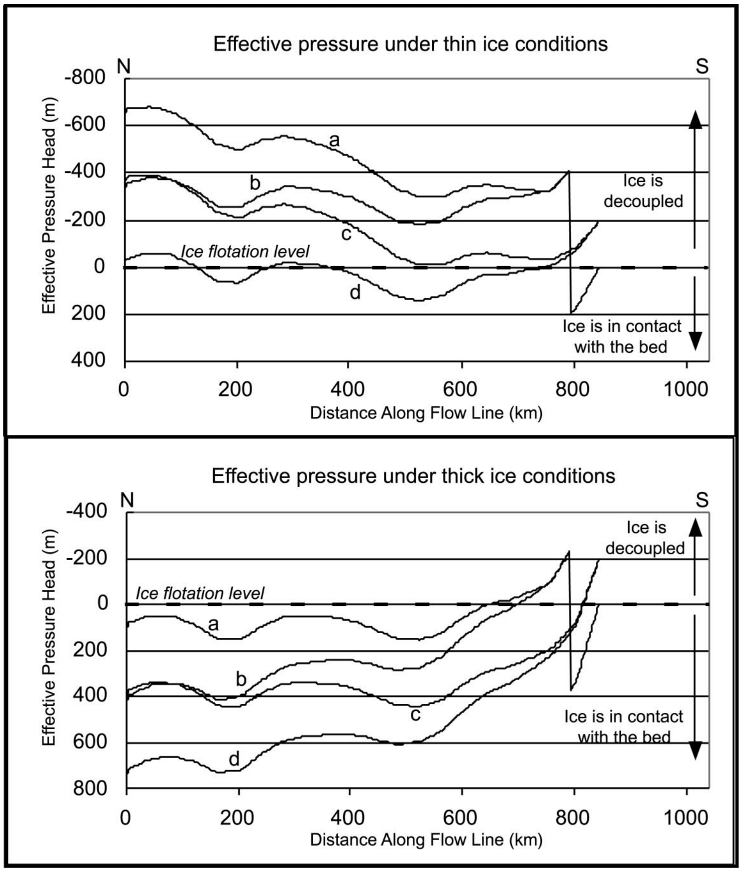

Figure 5. (A) Effective pressure (in meters) at the ice-bed interface for thin-ice condition (ice thickness is 1200 m at 0 km distance). Ice-flotation level occurs where effective pressure is <0 m. Curves are as follows: (a) Simulation with 100 km of permafrost bracketing the ice margin and water flowing through a 7 mm film. (b) Simulation with 100 km of permafrost bracketing the ice margin and water flowing through an 8 mm film. (c) Simulation with no permafrost and water flowing through a 7 mm water film. (d) Simulation with no permafrost and water flowing through an 8 mm water film. (B) Same as in A but for thick-ice condition (ice thickness is 2000 m at 0 km distance).

(simulation 5 with a 7 mm film). All model layers were assigned a specific storage (S_s) of 4.6×10^{-6} m^{-1}, on the basis of a coefficient of vertical compressibility (β_p) equal to 3.3×10^{-10} m^2/N, a value typical of solid rock (Domenico and Mifflin, 1965; Johnson et al., 1968). Aquifer porosity was assumed to be 0.3. The coefficient β_p is typically one to four orders of magnitude greater for unconsolidated sediment than for solid rock. Along our simulated flow line, however, bedrock generally occupies >95% of the entire profile thickness, suggesting that $S_s = 4.6 \times 10^{-6}$ m^{-1} is a reasonable generalization of the aquifer system. All initial head values were defined as equal to those generated by the validation simulation.

Under transient conditions, subglacial head reaches 97% of the simulated maximum steady-state head value in ~1000 yr, and within 1900 yr it reaches the steady-state maximum of 1447 m (Fig. 6). When S_s is increased to 8.2×10^{-6} m^{-1}, a value typical of fractured rock, the subglacial head reaches 95% of the maximum value after 1000 yr and reaches the steady-state maximum after 2900 yr (Fig. 6). These results suggest that water pressure underneath the Lake Michigan Lobe could have equilibrated to steady-state values while it remained at its maximum extent between 22 and 19 ^{14}C ka (Hansel and Johnson, 1992). These transient simulations, however, predict the response of subglacial aquifers to an instantaneous application of the maximum Lake Michigan Lobe. Because the lobe required some amount of time to advance to this maximum position, it is likely that subglacial aquifers were significantly perturbed prior to the period examined in this simulation. For this reason, our transient simulation probably overestimates the amount of time necessary for subglacial water pressure to reach a steady-state maximum.

Comparison of Groundwater-Flow Patterns During Glacial and Nonglacial Periods

Under nonglacial conditions, groundwater is typically recharged in topographically high areas and discharged from topographic lows. Our simulations indicate, however, that the low-gradient Lake Michigan Lobe that preferentially flowed through the topographically low Lake Michigan basin, altered or reversed topographically driven pressure gradients, resulting in flow patterns and velocities that differed substantially from those of nonglacial periods. To discuss the possible rearrangement of groundwater-flow patterns, we use results from the validation model, the glacial-stage simulation based on validation-model *K* values, and the glacial-stage simulation that includes a 7 mm film and no permafrost as three end-member scenarios.

The validation model shows a flow pattern consisting of recharge in the high-elevation areas north and south of Lake Michigan and upward discharge through aquifers underlying the lake. South of Lake Michigan, groundwater velocity is substantially reduced, and flow directions are variable (Fig. 7). This transition occurs because a thick Pennsylvanian and Mississippian confining bed that forms a subcrop below the southern half of Illinois severely restricts precipitation recharge. Furthermore, the Mahomet bedrock valley probably

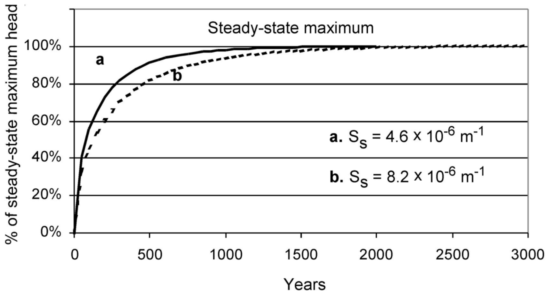

Figure 6. Transient aquifer response to glaciation for two cases of specific storage (S_s).

intercepts much of the groundwater flowing through the upper aquifers, causing the water to flow toward the southwest, ultimately discharging to the Mississippi River.

Under glacial conditions, when a subglacial water film is not present, groundwater velocities are high, and flow is consistently directed toward the southern end of the flow line (Fig. 7). Along the northern two-thirds of the flow line, groundwater has a relatively strong downward component in the drift aquifer. Along the southern one-third of the flow line, flow in the drift aquifer is negligible, owing to flow through the Mahomet bedrock valley and discharge at the ground surface. Deeper groundwater, in the St. Peter–Prairie du Chien, Ironton-Galesville, and Mount Simon aquifers, is dominated by horizontal flow. South of the ice margin, groundwater in the St. Peter–Prairie du Chien and Silurian aquifers is directed upward. Groundwater in the Mount Simon aquifer in the Illinois basin discharges to regions that currently serve as recharge zones. When a 7 mm film is included in the simulation, subglacial meltwater that is directed downward into aquifers exhibits flow in a pattern similar to, but at lower velocity than, that described for conditions in which a film does not exist at the ice-bed interface (Fig. 7).

These simulations indicate that groundwater vectors were significantly altered underneath the Lake Michigan Lobe compared to modern conditions. The two glacial end-member drainage-system scenarios (no film vs. 7-mm thick film) indicate that groundwater-flow patterns were fundamentally different under the influence of the Lake Michigan Lobe, indicating that regardless of which end member scenario is more realistic, groundwater-flow patterns were rearranged under the influence of the lobe. Under glacial conditions, all simulations show groundwater consistently flowing downward along the northern half of the flow line, which is opposite to the vectors shown by the validation simulation and opposite to flow patterns currently observed. In general, these results agree with work elsewhere in the glaciated Midwest by showing that some areas that currently serve as groundwater discharge zones were recharge zones during glacial periods (Carlson, 1994; Hoaglund, 1996).

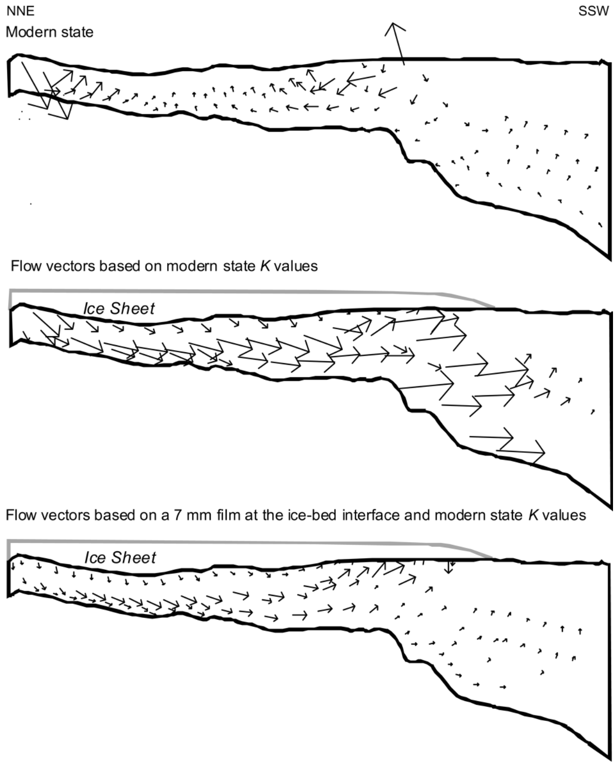

Figure 7. Groundwater vectors for three different conditions along modeled flow line A–A′ (see Fig. 4) (the vector scale varies between illustrations). (A) Modern state. (B) Simulation based on modern *K* values with Lake Michigan Lobe present but no drainage system at the ice-bed interface. (C) Simulation based on modern *K* values with Lake Michigan Lobe present and a 7-mm-thick water film at the ice-bed interface.

DISCUSSION AND CONCLUSIONS

Our simulations indicate that during the Last Glacial Maximum, the Lake Michigan Lobe rested on a substrate with a transmissivity less than that needed to drain the estimated basal discharge. The hydraulic conductivity of the underlying aquifers must be increased to unrealistically high values to reduce simulated water pressure to levels less than the ice-overburden pressure and allow the lobe to achieve a steady-state condition. Because flow through porous subglacial aquifers is clearly an inadequate mechanism for transporting meltwater, an additional drainage system at the ice-bed interface would have developed in response to the high basal water pressure.

Our simulations are based on conservatively estimated values of glacial-drift *K,* drift thickness, and subglacial meltwater flux. Actual values may have been greater than the estimates applied to these simulations; however, simulations based on increased values of drift *K*, drift thickness, and basal meltwater flux would yield qualitatively identical results and strengthen our conclusion that the aquifers underneath the Lake Michigan Lobe were insufficient to evacuate the basal meltwater.

Our simulations show that a drainage system at the ice-bed interface—possibly equivalent to a water film on the order of 7–8 mm thick and covering much of the glacier bed—would have been adequate to maintain subglacial water pressure at levels less than or equal to the ice-overburden pressure. A small increase in film thickness can accommodate a large decrease in the fraction of the bed that is covered by water; thus, a film- or canal-type drainage system occupying only a limited part of the glacier bed would have been sufficient to drain the subglacial meltwater. Similar drainage systems have been hypothesized as underlying Ice Stream B, West Antarctica (Lingle and Brown; 1987, Alley et al., 1989; Engelhardt and Kamb, 1997).

Fast ice flow, such as that exhibited by the Lake Michigan Lobe, requires a bed offering low friction, which in turn is dependent on high subglacial water pressure. Extremely permeable sediment in the buried Mahomet bedrock-valley system had the capacity to transmit virtually all of the basal meltwater delivered to it from upstream to regions beyond the boundaries of the Lake Michigan Lobe. Such a diversion would reduce subglacial pore pressure and/or drain any film at the ice-bed interface, significantly increasing the resistance to ice flow at the bed and slowing the advance of the ice. Anandakrishan and Alley (1997) described a comparable situation for Ice Stream C in West Antarctica, wherein subglacial diversion of basal water away from the Ice Stream C catchment may have caused the ice stream to slow dramatically. Although the cause of the diversion is related to a change in the orientation of the head gradient rather than a buried bedrock valley, the effect on downstream basal water pressure is the same. Given that the southern and western boundaries of the maximum Lake Michigan Lobe are roughly parallel to the main tributaries of the buried bedrock-valley system (Fig. 1), we suggest that the high transmissivity of the tributaries of the buried bedrock valley in Illinois, and perhaps Indiana, played a role in determining the maximum extent of the lobe.

Apparently, permafrost had very little effect on subglacial hydrology as long as the permafrost was limited to areas south of the Mahomet bedrock valley at the time of the Lake Michigan Lobe's maximum extent. Cutler et al. (2000) calculated that the bed of the Green Bay Lobe was frozen for 60–200 km upstream of the ice-sheet margin. If $>$50 km of permafrost were underlying the Lake Michigan Lobe, then the Mahomet bedrock valley would be occupied by permafrost and therefore be ineffectual at draining the subglacial meltwater. Although permafrost existed near the margin of the Lake Michigan Lobe during the Last Glacial Maximum, it apparently formed only in areas near the ice margins (Johnson, 1990) and thus had little effect on ice behavior and subglacial hydrology.

In conclusion, our numerical simulations suggest the following. (1) Aquifers underneath the Last Glacial Maximum Lake Michigan Lobe of the southern Laurentide Ice Sheet were incapable of draining the estimated basal meltwater. (2) Meltwater in excess of that transmitted through aquifers may have been discharged through some type of water film at the ice-bed interface. A basal drainage system equivalent to a 7–8-mm-thick water film, occupying 100% of the bed surface, could have removed enough meltwater to maintain water pressure at a level less than that of the overlying ice. (3) The Mahomet bedrock-valley system drained much of the basal meltwater, leading to substantially reduced subglacial pore pressure; thus the mechanism of fast ice flow was removed, possibly resulting in ice stagnation. (4) Permafrost probably did not have a significant effect on the subglacial hydrology. Meltwater diversion through the buried Mahomet bedrock-valley system probably minimized the possible effects of permafrost. (5) During the Last Glacial Maximum, groundwater-flow patterns and velocities were altered or reversed relative to those in nonglacial periods.

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