AN ABSTRACT OF THE THESIS OF

Christopher W. Breemer for the degree of Master of Science in Geology presented on May 25, 2000. Title: Subglacial Hydrology of the Lake Michigan Lobe, Laurentide Ice Sheet.

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Peter U. Clark

Fast ice flow and unstable ice sheet behavior were characteristic features of the Lake Michigan Lobe of the southern Laurentide Ice Sheet. Such behavior may result from some combination of subglacial-sediment deformation and decoupled sliding at the ice-bed interface. Both mechanisms depend on high water pressure relative to ice pressure. Using the finite-difference groundwater modeling package MODFLOW we simulate groundwater flow along a 1,040 km flowline, extending from the south shore of Lake Superior to the Mississippi River near Carbondale, Illinois. Model simulations indicate that subglacial aquifers were not capable of evacuating the estimated basal meltwater. A basal drainage system consisting of a distributed film or canal system, similar to systems hypothesized as underlying Ice Stream B, West Antarctic Ice Sheet, would transmit sufficient water to prevent basal water pressure from exceeding the ice overburden pressure. The buried Mahomet bedrock valley system may have drained enough subglacial meltwater to

stop the advance of the Lake Michigan Lobe. Simulations also suggest that groundwater flow directions and velocities were substantially different than modern conditions.

Subglacial Hydrology of the Lake Michigan Lobe, Laurentide Ice Sheet

by

Christopher W. Breemer

A Thesis Submitted

to

Oregon State University

In Partial Fulfillment of
the requirements for the
degree of

Master of Science

Presented May 25, 2000
Commencement June 2001

Master of Science thesis of Christopher W. Breemer presented on May 25, 2000

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Christopher W. Breemer, Author

ACKNOWLEDGEMENTS

This paper is the culmination of three years of research, during which time I have benefited immensely from the suggestions and contributions of numerous people. Most importantly, Peter Clark and Roy Haggerty helped organize the project, provided limitless conceptual guidance, and improved the writing. Staff members of the Illinois State Geological Survey and the Wisconsin State Natural History Survey graciously offered their time and facilities for my use. Most notable among those staff is Ardith Hansel who housed me and guided me through stacks of files and piles of sediment in Illinois. Jack Istok encouraged the pursuit of this project and edited this manuscript.

This project was supported by grants from Sigma Xi and the National Science Foundation

CONTRIBUTION OF AUTHORS

Peter Clark helped with every phase of this project including its design, implementation, and editing. Roy Haggerty helped design the groundwater simulations and he edited every draft of this thesis.

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DEDICATION

This thesis is dedicated to my wife Mary and my parents Jan and Ann. Mary encouraged me to pursue a project that made me happy, an advisor I respect, and a life I enjoy. She also showed limitless patience toward a thesis that, at times, seemed as if it would last forever. My parents gave me the desire, means, and freedom to explore, and that ultimately lead to my affair with ice.

Subglacial Hydrology of the Lake Michigan Lobe, Laurentide Ice Sheet

INTRODUCTION

The Lake Michigan Lobe (LML) was a dynamic feature of the Laurentide Ice Sheet during the last glaciation. The lobe had an extremely low surface profile, suggesting that the ice advanced under a driving stress of approximately 0.9–1.7 kPa (Clark, 1992), yet experienced fast ice flow with a number of significant oscillations (Mickelson et al., 1983; Hansel and Johnson, 1992), suggesting that the lobe flowed over a bed offering very low yield stress. Such stress could be offered by some combination of subglacial sediment deformation and decoupling of the ice from its bed (Alley, 1991). Both conditions 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.

We report on our investigation of the subglacial hydrology of the LML during its last glacial maximum. We use MODFLOW (McDonald and Harbaugh, 1988), a finite-difference groundwater model, to explore the likely groundwater pressure distribution and flow patterns along a profile flowline extending from Lake Superior to the Mississippi River near Carbondale, Illinois (Figure 1). Our flowline includes a realistic description of the major hydrogeologic units that existed beneath the LML. We perform a number of sensitivity tests to assess the capacity of the substrate to transmit estimated glacial meltwater fluxes. We also investigate plausible drainage systems that may have existed at the ice-bed

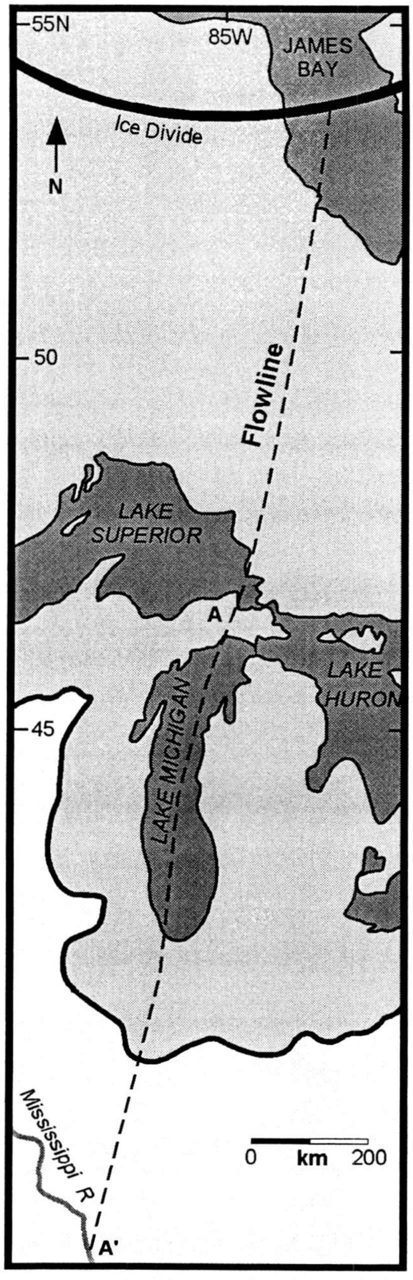


Figure 1: Simulated flowline. Ice limit is shown with ice cover represented by the grey pattern. A-A' show the boundaries of the finite difference grid.

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.

Output from these simulations allows the estimation of water pressure at the ice-bed interface that would have influenced mechanisms of fast ice flow. Our results also include the water pressure in the deeper aquifers, thus enabling us to reconstruct likely ground water flow vectors in subglacial aquifers.

PREVIOUS WORK

Numerical simulations 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; Piotrowski, 1997a,b). Piotrowski (1997a,b) concluded that the aquifers underlying the Fennoscandian Ice Sheet in northwestern Germany were not capable of transmitting the estimated basal meltwater discharge, thus requiring that subglacial meltwater flow through tunnel valleys, probably during spontaneous outburst events. In contrast, Boulton et al. (1995) suggested that subglacial aquifers had the capacity to transmit all of the basal meltwater. These different results likely reflect the fact that Boulton et al. (1995) did not include glacial drift in their model, based on the assumption that the glacial drift had little influence on the transmission of subglacial meltwater.

Research in Iowa, North Dakota, and Michigan also suggests that glaciation profoundly affected regional groundwater flow patterns. Siegel and Mandle (1984) and Siegel (1989, 1991) argued that significantly depleted $\delta^{18}O$ values in groundwater in Iowa was due to a glacial meltwater origin. They concluded that glacial meltwater in Iowa had been forced through confining layers, in a direction opposite that of modern flow, implying that glacial-stage pressure gradients were different from modern gradients. Similar geochemical anomalies in the Fox Hills Aquifer of North Dakota also suggest that modern discharge areas served as recharge zones during one or more glacial periods (Carlson, 1994). Based on a regional groundwater and particle tracking model, Hoaglund (1996) showed that ice loading may have caused groundwater flow reversals in Michigan. His simulations indicate a strong downward component of flow beneath the former ice sheet and an upward component into a large proglacial lake.

Similar research has not been performed in the region formerly overlain by the LML. Nevertheless, since the extent and timing of glaciation in Iowa, North Dakota, and Michigan was similar to that of Illinois, it is reasonable to expect that the LML may have similarly affected regional groundwater flow patterns.

REGIONAL GEOLOGY

The LML probably originated from an ice divide near James Bay. During the last glacial maximum it flowed south through the Lake Michigan basin, ultimately terminating near Shelbyville, Illinois. A 400 m thick layer of Paleozoic carbonate bedrock, overlying Precambrian crystalline basement rock and mantled by 0-7.5 m of drift, formed the bed of the LML for the first 400 km of the flowline (Marshall et al., 1996). Between the Paleozoic carbonates and the southern shore of Lake Superior Precambrian basement rock is exposed at the surface, occasionally covered by a thin cover of glacial drift.

South of Lake Superior, a southward thickening sequence of sedimentary units overlies the crystalline basement. 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 greater than 7,000 m in the Illinois Basin. These sedimentary units primarily consist of carbonates, sandstone, siltstone, shale, and coal.

A sequence of glacial drift between 0-200 m thick overlies the bedrock units along the flowline south of Lake Superior. Because we are concerned with the bedrock and drift affected by the glacial maximum LML, our model is limited to the late Wisconsinan Tiskilwa Till, the oldest member of the Wedron Formation (Hansel and Johnson, 1996), and earlier drift units. Tiskilwa Till and older drift is absent or undocumented along the flowline between Lake Superior and a point about 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 assume that 1 m of Tiskilwa Till existed from the southern shore of Lake Superior to a point approximately 130 km south of Chicago. South of that point, the extent and thickness of Tiskilwa Till is constrained by published maps, literature, and borehole records maintained by the Illinois State Geological Survey. Based on these data we model the Tiskilwa Till as a wedge-shaped deposit, with the thickness increasing to a maximum of 15 m at the lobe terminus near Shelbyville, Illinois. Along the flowline south of Shelbyville, Tiskilwa Till is absent, but earlier Illinoian drift deposits are widespread. These deposits are between 0 and 30 m thick (Soller, 1997), and we assume a 15 m drift thickness for the entire cross section south of Shelbyville.

Drift and bedrock stratigraphy is interrupted in central Illinois by the buried Mahomet Bedrock Valley system (Figure 2). The Mahomet Bedrock Valley is a complex lowland eroded into Pennsylvanian and older rocks underlying glacial drift. It traverses central Illinois, roughly in an east-west direction, ultimately terminating in the Mississippi River Valley. The Mahomet Bedrock Valley is between 13 and 22 km wide, with width generally increasing toward the west. Fluvial deposits that fill the Mahomet Bedrock Valley system are the most extensive and highly productive sand and gravel aquifers in the southern three quarters of Illinois (Kempton, 1991).

DESCRIPTION OF THE GROUNDWATER MODEL

We have used a 2-D groundwater model to simulate flow underneath the LML along a flowline that extends from the ice divide at James Bay, Canada, to the terminus of the LML at Shelbyville, Illinois. Beyond the ice terminus, the model and flowline continues to the Mississippi River near Carbondale, Illinois. Our finite-difference grid is limited to the region south of Lake Superior where aquifers with significant transmissivity occur.

Inherent to the usage of a 2-D model is the assumption that all flow occurs parallel to the transect (Anderson and Woessner, 1992). Subglacial head is largely defined by ice thickness (Paterson, 1994). Therefore, since most geologic material has hydraulic conductivity that is approximately isotropic in the horizontal plane, the horizontal component of groundwater flowlines runs approximately parallel to that of ice flowlines. We thus assume that the groundwater flow paths are approximately parallel to ice flowlines, and therefore that a 2-D model adequately describes the hydrogeologic system.

The assumption of 2-D flow will be violated where large-scale, high conductivity geologic units strike perpendicular or oblique to ice flowlines. An example of such a violation would be an east-west trending alluvial unit, such as the Mahomet bedrock valley, in a location where ice flowlines run north-south. A small number of such cases can be dealt with using source-sink terms, as will be explained shortly.

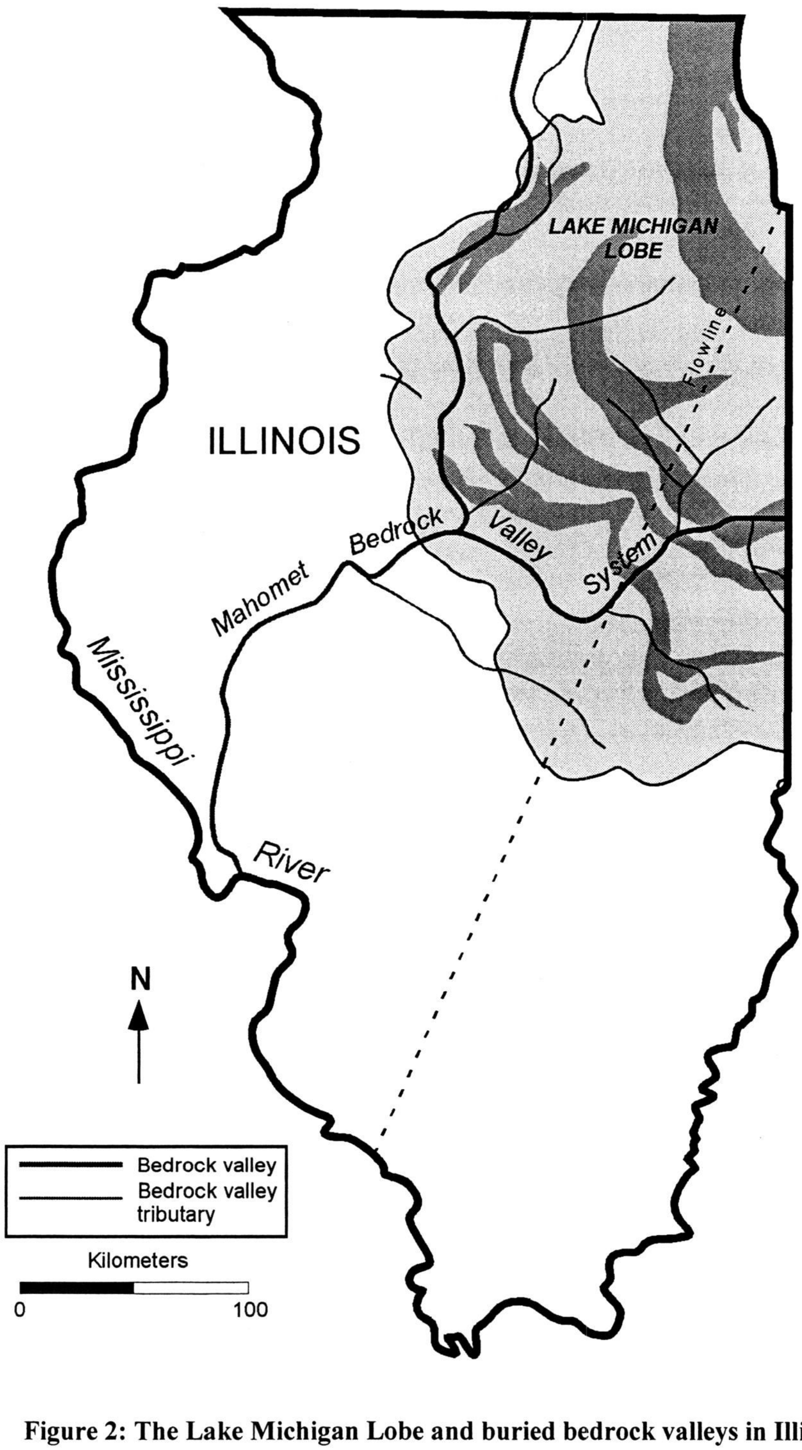


Figure 2: The Lake Michigan Lobe and buried bedrock valleys in Illinois

Some amount of water may have flowed in directions oblique to our flowline, but that flow should not have a significant effect on our results. Oblique flow would probably be limited to the area south of modern Lake Michigan, where the LML extended beyond the main mass of the Laurentide Ice Sheet. Beyond the ice sheet, potential gradients perpendicular or oblique to the central flowline may have existed. Those potential gradients would force water toward the lateral margins of the lobe. Due to the roughly circular shape of the LML, the distance from the model flowline to the southern terminus near Shelbyville is generally similar to the distance to other lobe margins. Illinois hydrostratigraphy does not change substantially in the regions paralleling the flowline, suggesting that water flowing oblique to the modeled flowline would encounter an aquifer system similar to that along the flowline. Oblique flow systems would thus require a similar head gradient to evacuate the meltwater, and thus will not significantly change our results.

THE NUMERICAL MODEL

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

$$\nabla \bullet (K \nabla h) - w = 0 \tag{1}$$

where K [L/T] is the hydraulic conductivity, h [L] is hydraulic head, and w [1/T] is a source-sink term. Hydraulic conductivity is defined by:

$$K = \frac{-k\rho g}{\mu} \tag{2}$$

where k [L^2] is permeability, ρ [M/L^3] is the density of the fluid, g [L/T^2]is the acceleration of gravity, and μ [M/L/T] is the viscosity of the fluid. Equation (1) can be transformed to a finite-difference form and is the basis of our numerical simulations.

The complete model consists of 10,400 cells. Each cell represents a region 1,000 m in the direction parallel to the flowline and a variable vertical distance, defined by the thickness of the each hydrostratigraphic unit. The model is divided into 10 layers (Figure 3), each of which consists of 1,040 cells. 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.

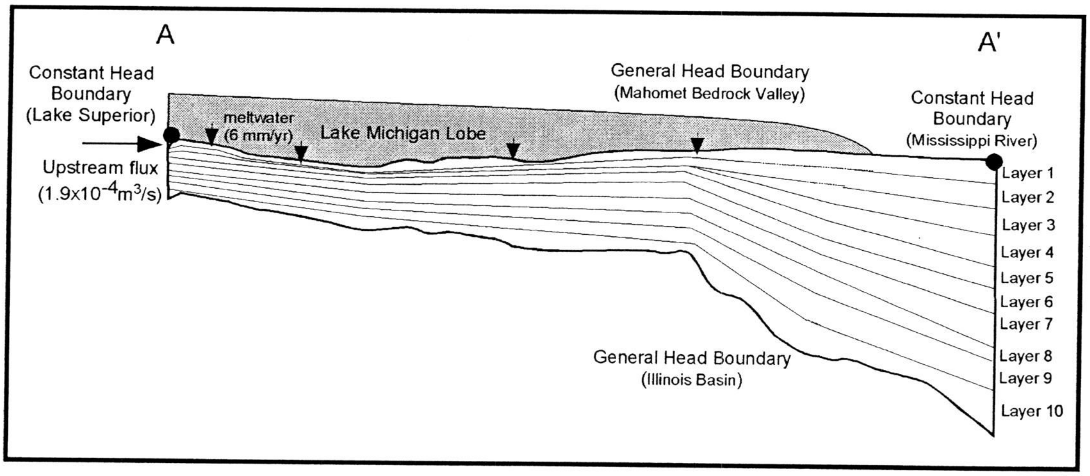


Figure 3: Generalized model geometry and boundary conditions. Vertical exaggeration is 100X. Model stratigraphy is not to scale.

HYDROSTRATIGRAPHY

We grouped geologic formations into six regional aquifer layers and four regional confining layers based on their hydraulic characteristics (Table 1). Hydrostratigraphic groupings generally follow those outlined by Mandle and Kontis (1992) in their simulation of the Northern Midwest Regional Aquifer System.

Table 1: Model stratigraphy

Model Layer	Hydrostratigraphic Unit
1	Glacial drift. Layer 1 is 0.5 m. Layer 2 is variable thickness
2	
3	Pennsylvanian through Middle Devonian rocks. Consist of limestone, shale, sandstone, siltstone, and coal
4	Silurian and Devonian carbonates
5	Maquoketa Shale, Galena Dolomite, and the Decorah, Platteville, and Glenwood Formations. The Decorah, Glenwood, and Platteville Formations consist of limestone, dolomite, sandstone, and shale.
6	St. Peter Sandstone, Prairie du Chien Group, and the Jordan Sandstone. Prairie du Chien Group consists of limestone, dolomite, and discontinuous layers of sandstone, siltstone, and shale.
7	St. Lawrence and Franconia Formations. Consist of shaly sandstone, limestone, and dolomite.
8	Ironton and Galesville Sandstones
9	Eau Claire Formation. Consists mainly of sandstone.
10	Mt. Simon Sandstone and Elmhurst Sandstone

We estimated the extent and thickness of the bedrock hydrostratigraphic units based on borehole logs, stratigraphic cross sections, and literature published by the Illinois State Geological Survey, the Wisconsin Geologic 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 based on the estimates of Mandel and Kontis (1992) (Table 2). The hydraulic conductivity assigned to the drift aquifer is one order of magnitude greater than Walton's (1965) estimate for vertical conductivity in Illinois glacial sediment. Drift conductivity was increased relative to Walton's estimate so that the validation model head solution would more closely resemble the modern head distribution. Aquifers and confining beds are all characterized as isotropic. Since 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.

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. The *K* value assigned to the drift layers is based on a statewide average for Illinois and therefore should account for heterogeneities that are often responsible for estimation errors.

Table 2: Hydraulic conductivity values in Illinois. 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.

Layer	***K* (m/s) Assigned to the validation simulation**	**Geometric Mean (m/s)**	**-1σ (m/s)**	**+1σ (m/s)**	**Total Obs.**
Quaternary drift	3.7×10^{-7}	N/A			
Quaternary drift					
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}	1,816
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 flowline)	6.1×10^{-5}				
St. Lawrence and Franconia Formations	2.4×10^{-9}	N/A			
St. Lawrence and Franconia Formations (southern third of the flowline)	3.0×10^{-11}				
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 flowline)	1.5×10^{-4}				
Eau Claire Formation	3.0×10^{-10}	N/A			
Eau Claire Formation (southern third of the flowline)	2.4×10^{-10}				
Mt. Simon Sandstone and Elmhurst Sandstone	1.1×10^{-4}	1.8×10^{-5}	2.4×10^{-6}	1.4×10^{-4}	99

Our aquifer characterizations are primarily based on data gathered from hydraulic tests performed under current interglacial field conditions. An assumption implicit to this methodology is that modern hydraulic characteristics adequately describe the glacial aquifer system. Some hydraulic characteristics, however, may have been different during the glacial maximum. Hydraulic conductivity is a function of both the medium through which a fluid is flowing and the fluid itself. Hydraulic conductivity 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 about 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 approximately 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.

BOUNDARY CONDITIONS

To simplify the model, we assume that all of the meltwater that was generated along the flowline between the ice divide and the south shore of Lake Superior flowed at the ice-bed interface (Figure 4). This simplification is justified because north of Lake Superior the meltwater is confined by a relatively shallow boundary of crystalline basement rock. Near the north shore of Lake Superior, where the crystalline rock subcrops and frequently crops out, groundwater driven by an ice-pressure gradient would be forced to flow at the ice-bed interface and/or through a very thin layer of surficial drift.

The basal melt rate of the ice is difficult to estimate without knowing the ice velocity, accumulation rate, geothermal gradient, and paleoclimate conditions. Given these limitations, we assume 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 1,000 km flowline length upstream of Lake Superior, we estimate an upstream meltwater flow rate of 1.9×10^{-4} $m^3/m/s$. Precipitation recharge is not added to the model since we assume the ice is impermeable.

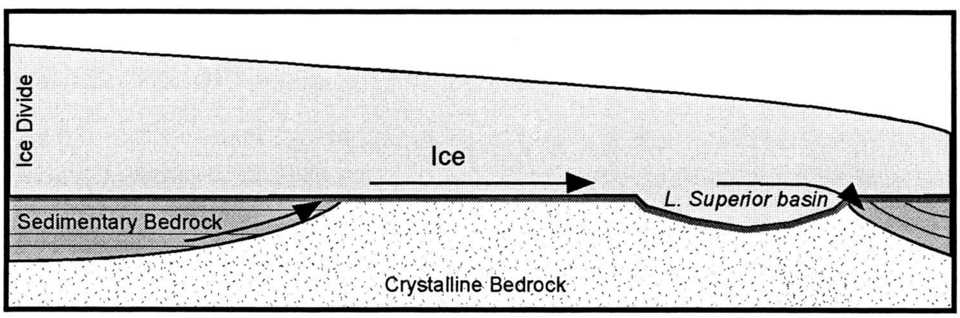


Figure 4: Conceptual model of groundwater flow north of Lake Superior.

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 the model. The entire length of the modeled flowline 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 whether ice was or was not present. On the part of the flowline covered by the glacier, the ice is treated as an impermeable layer. South of the lobe terminus, the top of the model is treated as an unconfined aquifer.

Two-dimensional groundwater models are limited by the assumption that all flow is parallel to the plane of the model. In some cases, however, groundwater may flow in directions oblique to the simulated 2-D plane. Under these conditions a gradient-dependent source-sink term can be utilized to estimate transverse flow. The term is:

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

where K is hydraulic conductivity, d is the aquifer thickness, h is hydraulic head, and n is the distance oblique to the flowline. This term is included in equation (1) and calculates flow into or out of a specified cell, based on the difference between the computed head in the specified cell and a constant head value defined for an outside source or sink. We apply a gradient-dependent source-sink term to our model in two locations, representing the deep aquifers of the Illinois Basin and the buried Mahomet Bedrock Valley system.

Despite the work of numerous researchers (Bredehoeft, 1963, Cartwright, 1970, Bond, 1972), groundwater flow in the Illinois Basin is poorly understood. Generally, the deep aquifers in the Illinois Basin are recharged at the basin margins at outcrops and subcrops in Missouri, Indiana, and Kentucky. Deep groundwater discharges by upward leakage in the basin, possibly through faults. The recharge areas for the Illinois Basin do not occur along our modeled flowline, but discharge does occur in the southern portion of our flowline.

To account for recharge to the deep basin aquifers, we have assigned a gradient-dependent source-sink term to the Mt. Simon aquifer in the Illinois Basin. The Mt. Simon aquifer subcrops in southern Indiana and crops out in northwest Missouri. The outcrops in Missouri occur in hills about 90 km west of the modeled flowline at altitudes up to 500 m, but more generally at about 300 m. We assume that 300 m adequately describes the head in the Mt. Simon in its recharge zone. The thickness of the aquifer is approximately 300 meters and the hydraulic conductivity is about 6×10^{-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 LML. However, the discharge area for the valley system lies beyond the ice margin, suggesting that the Mahomet Bedrock Valley may have functioned as a subglacial meltwater conduit. Subglacial water could have entered the bedrock valley by vertical leakage through overlying till and other drift, as occurs today, drained through the Mahomet valley system, and discharged to the west in non-glaciated areas. If the Mahomet Bedrock Valley behaved as suggested above, it is likely that subglacial water pressure in the vicinity of the bedrock valley was substantially reduced relative to water pressure north of the bedrock valley.

To investigate the influence of the Mahomet Bedrock Valley on subglacial hydrology we assigned a gradient-dependent source-sink term to the model cells in the layer 2 drift aquifer that represent the Mahomet Bedrock Valley system. The modeled flowline intersects the Mahomet Valley about 100 km east of the western boundary of the LML; thus the length (l) 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. Water levels in the overlying Glasford Formation are nearly always 1.5 to 9 m above those of the Mahomet Sand (Kempton, 1991). Artesian wells drilled in the Mahomet Sand are common in parts of central Illinois, indicating that the potentiometric surface is equal to or greater than the land surface. For these reasons, we assume that 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 the Glasford aquifers in the Mahomet Bedrock Valley is about 7×10^{-3} m^2/s (Kempton, 1991). We assume a constant 14 km width for the bedrock valley. Accordingly, a gradient dependent source-sink term is applied to each of the fourteen 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 should avoid overestimating the capacity of the bedrock valley aquifers to transmit meltwater.

Numerous tributary bedrock valleys join the Mahomet Bedrock Valley system, but they are frequently short, relatively shallow, and narrow. Commonly, the tributary valleys are filled with silt and clay (Kempton, 1991). For these reasons, we assume that the tributary valleys play an insignificant role in the regional hydrogeology, and therefore we do not include them in our simulations.

MODEL VALIDATION

We apply modern non-glacial boundary conditions to the model for validation. The stratigraphy and hydraulic characteristics remain unchanged. The 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 Northern Midwest Regional Aquifer System Analysis (RASA) model (Mandle and Kontis, 1992). 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 was applied to the top layer of the model to simulate precipitation recharge. The value 38 cm/yr is based on average soil-moisture surplus for the period 1967-1988 in east-central Illinois (Cravens, 1990).

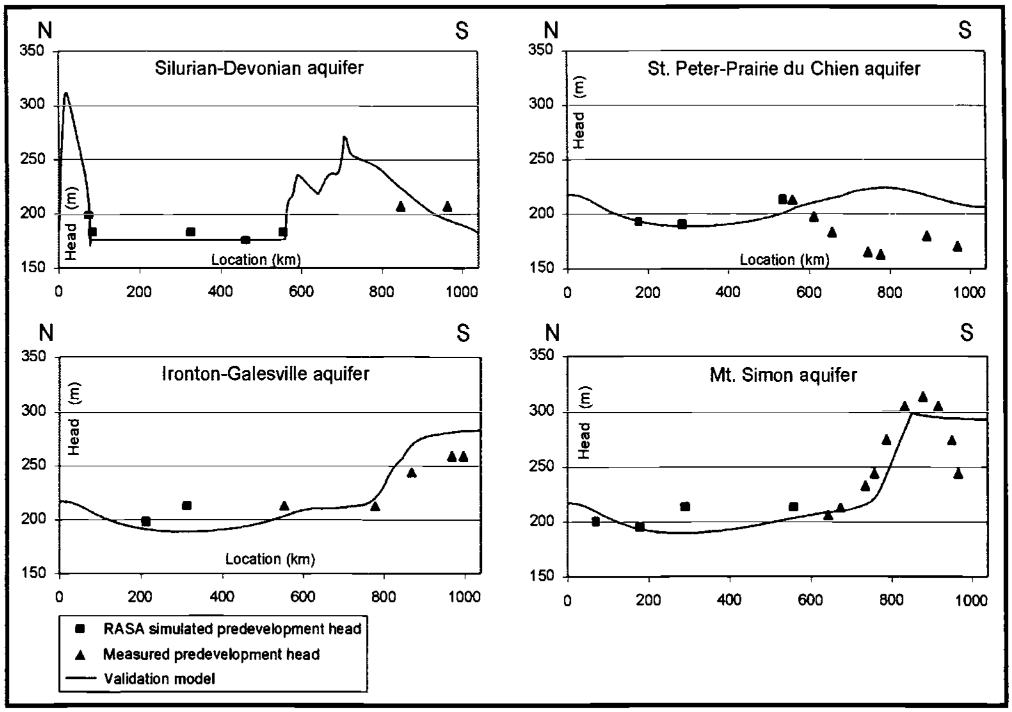


Figure 5: Comparison of validation model head to RASA measured and simulated head.

The validation simulation generates a head solution within 65 m of the measured and simulated head values published by Mandle and Kontis (1992) (Figure 5). More generally, validation model results are within 25 m of the RASA values. The greatest discrepancies occur in the Mt. Simon Sandstone and St. Peter Sandstone-Prairie du Chien Group. In the Mt. Simon Sandstone relatively large differences occur in the Illinois Basin at the southern end of the flowline. Head measurements in the Mt. Simon Sandstone in the Illinois Basin are extremely sparse. To compare our results to field measured data, we rely on relatively large extrapolations, which may explain 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 in the St. Peter-Prairie du Chien aquifer. The maximum divergence occurs about 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 (Figure 5). This 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 runoff to measured modern surface runoff values in Illinois. The Vermillion River drains 3,341 km^2 in central Illinois. During the years 1978-1998, United States Geological Survey stream gage data indicate that the Vermillion River had an average discharge of 37 m^3/sec, or 1.1×10^{-8} m^3/m^2/sec. Our simulated runoff value for this area is 1.2×10^{-8} m^3/m^2/sec, or 91% of modern values. Both simulated head values and runoff volumes are thus close to measured modern values, suggesting that the model functions as a reasonable approximation of the hydrogeologic system along the modeled flowline.

RESULTS

Model simulations were designed 1) to evaluate the capacity of the subglacial aquifers to transmit the estimated subglacial meltwater of the LML, 2) to estimate the water pressure at the ice-bed interface under likely geological and hydrologic conditions, 3) to investigate the potential effect of permafrost on subglacial hydrologic processes, and 4) to offer a comparison of modern versus glacial-stage hydrogeologic flow patterns and magnitudes. Table 3 provides a summary of the hydraulic conditions applied to each of our simulations and the results of those simulations.

Under boundary conditions of the last glacial maximum with hydrogeologic properties set as for the validation model, maximum water pressure occurs at the northern boundary of the model where it is 50,240 m at the ice-bed interface, while at the southern terminus of the ice, the simulated water pressure is 2,292 m (Figure 6). Numerical reconstructions of the LML suggest that ice was between 1,200 m and 2,000 m thick at Lake Superior (Licciardi et al., 1998). Hereafter these ice reconstructions will be referred to as "thin ice" and "thick ice conditions". Under either of these conditions, water pressure at the ice-bed interface exceeded the ice overburden pressure, and the glacier would be decoupled from its bed. These results suggest either that the hydraulic conductivity values applied to the glacial drift and the bedrock aquifers are too low or that subglacial aquifers beneath the LML were insufficient to transmit estimated basal meltwater, thus requiring some other type of drainage system at the ice-bed interface to drain the excess water. We next evaluate these influences on our model.

Table 3: Summary of simulations

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 2,292 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 1,093 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 3,498 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 1 order of magnitude maximum head is 4,707 m in the northernmost cell in model layer 1. Head in Layer 1 is 702 m at the ice margin. When K is increased 2 orders of magnitude maximum head is 816 m and 359 m.
5	Layer 1 T=0.16 - 0.23 m^2/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 1,447-1,099 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 m^2/s. A 100 km wide band of impermeable permafrost is assumed at the ice margin.	Maximum head is 1,736 m and 1,418 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 width is increased to 50 km in the up-ice direction to cover the Mahomet Valley.	Maximum head is >50,000 m.

MODEL SENSITIVITY TO DRIFT CONDUCTIVITY

During geophysical investigations in Martinsville, Illinois, numerous field tests were performed to evaluate the hydraulic conductivity of the glacial drift. Thirty-four field tests were performed in fractured Vandalia Till, a member of the Illinoian Stage Glasford Formation (Battelle Memorial Institute and Hanson Engineers, Inc, 1990). All of the reported values were based on the hydrostratigraphic unit within the test section of the borehole estimated to have the highest *K*. The geometric mean of the tests, $K = 1\times10^{-6}$ m/s, should therefore be considered a high estimate of the conductivity. The *K* of the sand facies interbedded with the Vandalia Till is typically an order of magnitude greater than the till conductivity. A simulation based on drift *K* equal to 1×10^{-6} m/s, with bedrock *K* unchanged from the validation simulation (Simulation 2; Table 3), generates a maximum of 34,602 m head at the upper northern boundary of the model and a simulated head of 1,093 m at the ice margin.

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, five orders of magnitude greater the value applied to the validation model, simulated subglacial head was reduced to 3,498 m (Simulation 3; Table 3), a value close to, but still greater than, the ice overburden pressure under both thin and thick ice conditions. This simulation is based on a *K* value typical of gravel. Glacial till in Illinois, however, normally has *K* values that are three to nine orders of magnitude smaller (Soller and Berg, 1992, Curry, 1994), suggesting that the drift could not evacuate the meltwater while maintaining subglacial water pressure at a level less than the ice-overburden pressure.

We base our simulations on drift layers that are effectively homogenous at the scale of the model. While glacial drift in Illinois has ubiquitous sand lenses, gravel deposits, and other highly permeable inclusions, these inclusions are normally vertically and horizontally discontinuous (Kempton and Morse, 1982; Wickham et al., 1988). Therefore they lack hydraulic connections to other permeable units, and do not significantly affect regional groundwater patterns. The Mahomet Bedrock Valley is an exception to this characterization.

MODEL SENSITIVITY TO BEDROCK CONDUCTIVITY

In order to test the sensitivity of the model to bedrock *K*, we increased the hydraulic conductivity of all bedrock units by 1-2 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 4,707 m of head at the northernmost boundary of the drift aquifer. At the ice terminus, the model generates 707 m of head. 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.

EFFECT OF AN ICE BED DRAINAGE SYSTEM

Neither the bedrock nor the glacial drift had the capacity to transmit the basal meltwater while maintaining subglacial head less than the ice overburden pressure. These results suggest that some type of drainage system must have existed at the ice-bed interface. The subglacial drainage system of the LML may have been analogous to present drainage of Ice Stream B in the West Antarctic Ice Sheet. Like Ice Stream B, the LML exhibited fast flow and it rested on a bed of unconsolidated sediment overlying sedimentary bedrock units that are not capable of transmitting all of the subglacial melt (Alley, 1989, Lingle and Brown, 1987, Engelhardt and Kamb, 1997). This drainage system may have been a thin film of water at the ice-bed interface (Weertman, 1970), a canal system of the type envisioned by Walder and Fowler (1994), a distributed film similar to that theorized by Alley (1989), or some combination of these drainage systems.

Based on borehole observations, Weertman (1970) suggested that a distributed meltwater film might exist at the base of Ice Stream B. Weertman calculated a thickness of 0.14 mm for a film occupying 100% of the bed, although he suggested that this value might underestimate the actual film thickness because of experimental complications. Alley (1989) calculated that Ice Stream B might have an ice-bed drainage system consisting of a distributed water film between 0-7 mm thick, occupying an unspecified portion of the glacier sole. Engelhardt and Kamb (1997) calculated a thickness of 1.4-4.3 mm for a film occupying 50% of the bed.

The transmissivity of a film at the ice-bed interface can be quantified based on 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):

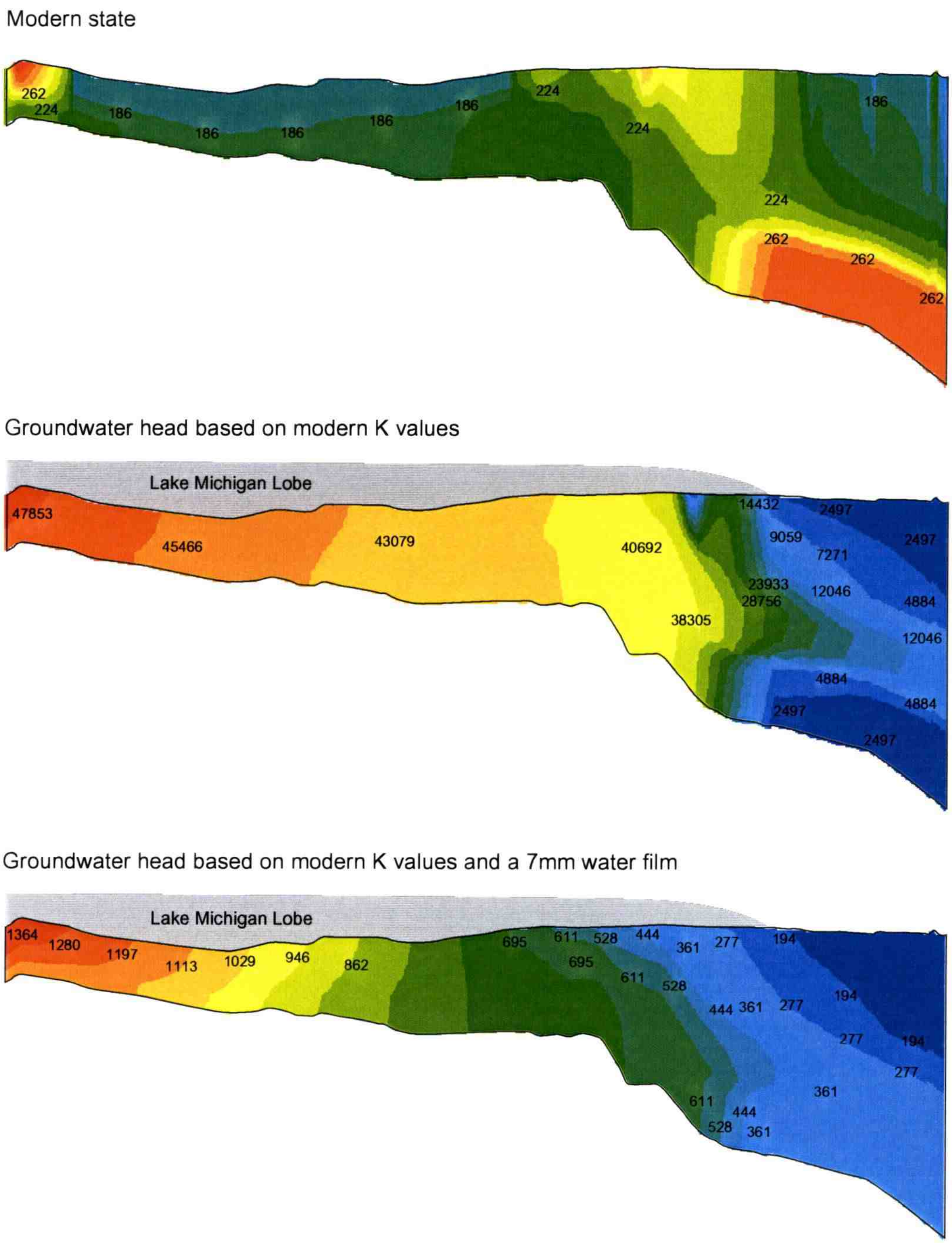


Figure 6: Simulated groundwater head (m)

$$T = \frac{\rho_w g d^3}{12\mu} \tag{4}$$

where ρ_w is the density of water, g is the gravitational constant, d is the distance between the plates, 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 5 mm layer of water at the ice-bed interface would reduce subglacial head to a maximum of 3,145 m, or about one and a half times more than the thick-ice overburden pressure, while a 7 mm film reduces maximum subglacial head to 1,447 m and an 8 mm film reduces head to 1,099 m (Figure 6). These results indicate that a water film with a thickness between 7-8 mm is sufficient to reduce subglacial head to values less than the ice flotation level under both thick-ice and thin-ice scenarios. Water flowing through these films may, however, exceed the Reynold's number describing laminar flow, in which case our film thicknesses may be underestimated. The film thickness suggested by our numerical simulations is similar to but slightly greater than the estimates that other researchers have calculated for Ice Stream B, suggesting that a drainage system similar to that of Ice Stream B may have operated underneath the Lake Michigan Lobe.

EFFECT OF PERMAFROST

Johnson (1990) suggested that permafrost may have been widespread near the LML margin during the last glacial maximum. 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 distributed film, canal system, or any other type of subglacial drainage system cannot discharge through the frozen terminal areas of the margin.

To investigate the possible hydrological 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 (100 total per layer) as impermeable to water (Simulation 6; Table 3). The transmissivity of the film layer was set to 0.16 m^2/s, the equivalent of a 7 mm layer of water. All of the other model parameters remained unchanged. Simulations based on the above conditions generate maximum head of 1,736 m. When the film thickness is increased to 8 mm, the simulation generates a maximum of 1,418 m head.

These results indicate that permafrost leads to slightly increased subglacial water pressure. This simulation is based on conditions in which permafrost is limited to an area south of the buried Mahomet Bedrock Valley system. Since the Mahomet 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 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 exceeding 50,000 m, again illustrating the importance of an unfrozen Mahomet valley in maintaining low subglacial water pressure. If the valley was frozen, the mechanism or mechanisms that could reduce the subglacial water pressure are unclear. Cutler et al. (in press) suggest that permafrost may have lead to the development of subglacial lakes underneath the neighboring Green Bay Lobe, and they speculate that those lakes may have been released occasionally during sudden Jokelhaup-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 outlined above suggest that groundwater flow patterns were significantly altered beneath the LML relative to modern conditions. Here we examine the time lag that may have existed between steady-state non-glacial and glacial conditions. To explore this question we performed a series of transient simulations based on an aquifer system that includes a 7mm film at the ice-bed interface. All model layers were assigned a specific storage (S_s) of 4.6×10^{-6} m^{-1}, based on 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 1- 4 orders of magnitude greater for unconsolidated sediment than for solid rock. Along our simulated flowline, however, bedrock generally occupies more than 95% of the entire profile thickness, suggesting that $S_s = 4.6\times10^{-6} m^{-1}$ 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 maximum value in about 1,000 years and within 1,900 years it reaches the steady-state maximum of 1,445 m. When S_s is increased to 8.2×10^{-6} m^{-1}, a value typical of fractured rock, subglacial head reaches 95% of the maximum value after 1000 years and the steady-state maximum after 2900 years (Figure 7). These results suggest that water pressure underneath the LML would have equilibrated to steady-state values while at its maximum extent between 22 to 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 LML. 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 time period examined in this simulation. For this reason, our transient simulation probably thus overestimates the amount of time necessary for subglacial water pressure to reach a steady-state maximum.

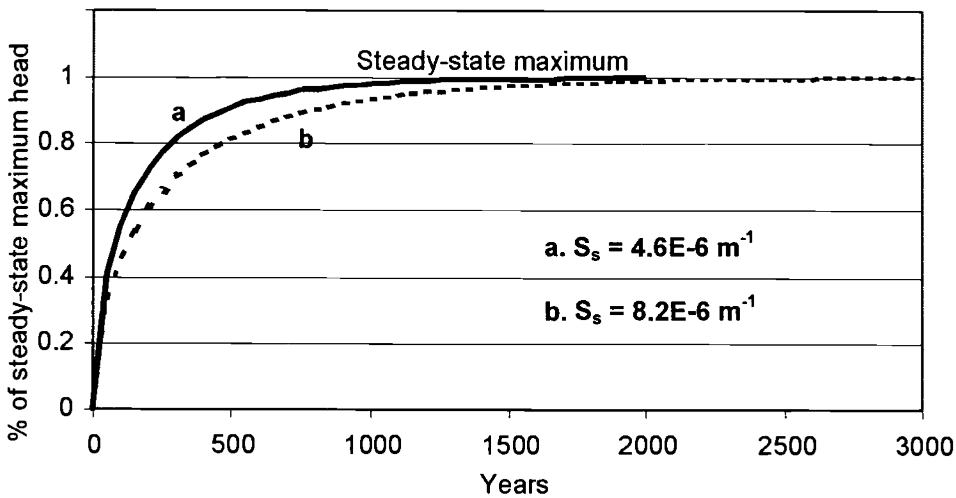


Figure 7: Transient aquifer response to glaciation

COMPARISON OF GROUNDWATER FLOW PATTERNS DURING GLACIAL AND NON-GLACIAL PERIODS

Under non-glacial conditions groundwater is typically recharged in topographically high areas and discharged from topographic lows. Glaciers, however, preferentially flow through topographically low regions. The pressure exerted by the ice, therefore, may alter or reverse topographically driven pressure gradients, resulting in flow patterns and velocities that vary substantially between glacial and non-glacial periods. Our simulations indicate that glacial-stage groundwater flow patterns may have been significantly altered relative to modern conditions. The validation model, the glacial-stage simulation based on validation model *K* values, and the glacial-stage simulation that includes a 7mm film and no permafrost offer three end member scenarios that we use to discuss possible rearrangement of groundwater flow patterns.

The validation model shows a flow pattern consisting of recharge in the high elevation areas bordering Lake Michigan and upward discharge through aquifers underlying the lake. South of Lake Michigan, groundwater velocity is substantially reduced and flow directions are variable (Figure 8). This transition occurs because a thick Pennsylvanian and Mississippian confining bed that subcrops over the southern half of Illinois severely restricts precipitation recharge. Furthermore, the Mahomet Bedrock Valley probably 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 flowline. Along the northern two-thirds of the flowline, groundwater has a relatively strong downward component in the drift aquifer. Along the southern one-third of the flowline, flow in the drift aquifer is negligible, due 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 Mt. 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 Mt. 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, 64% of the meltwater flows at the ice-bed interface, while the remaining 36% of the subglacial meltwater is directed downward into aquifers where it exhibits a flow pattern similar to that described for conditions in which a film does not exist at the ice-bed interface.

These simulations indicate that groundwater vectors were significantly altered underneath the LML. When flow is simulated in a system characterized by K values identical to those applied to the validation model, subglacial water pressure approaches 50,000 m, groundwater velocity is high, and flow patterns are completely rearranged in the deepest aquifers. Subglacial head values generated under these conditions are unrealistically high and therefore flow velocities are exaggerated. Nonetheless, the simulation offers a qualitative description of groundwater flow patterns that may have existed underneath the LML in the absence of a significant drainage system at the ice-bed interface.

When a 7mm film is exists at the ice-bed interface and aquifer K values are equivalent to those applied to the validation model the simulation shows extremely high groundwater flux at the glacier bed and very little flow in the lower aquifers. In both simulations nearly all groundwater vectors trend toward the southern end of the flowline.

We draw several conclusions from these results. First, the two glacial end member drainage system scenarios described above indicate that groundwater flow patterns were fundamentally different under the influence of the LML. Regardlesss of which end member scenario is more realistic, it is clear that groundwater flow patterns were rearranged under the influence of the lobe. Under glacial conditions

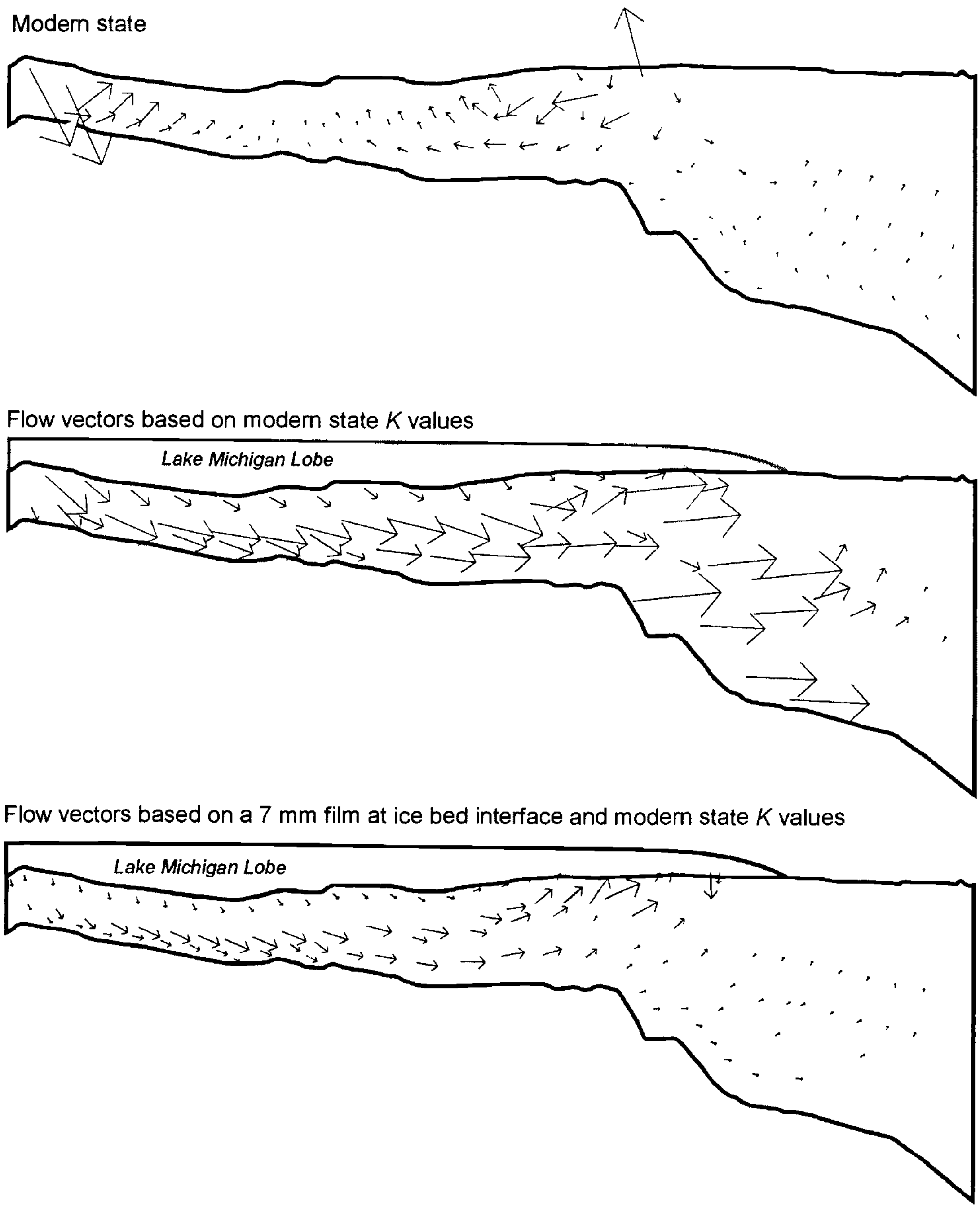


Figure 8: Groundwater vectors. Vector scale varies between simulations.

all simulations show groundwater consistently flowing downward along the northern half of the flowline. This is opposite to the vectors shown by the validation simulation and opposite to flow patterns currently observed. Currently, groundwater in northern Illinois flows north to discharge into Lake Michigan. Our simulations indicate that during the glacial maximum that flow pattern was reversed. Today, deep aquifers in the Illinois Basin are recharged by groundwater originating in recharge areas on the margin of the basin. Under glacial conditions, groundwater discharges through those basin margin areas. In general, these results agree with the conclusions of Siegel and Mandle (1984), Siegel (1989, 1991), Carlson (1994), and Hoaglund (1996), all of whom concluded that some areas that currently serve as groundwater discharge zones were recharge zones during glacial periods.

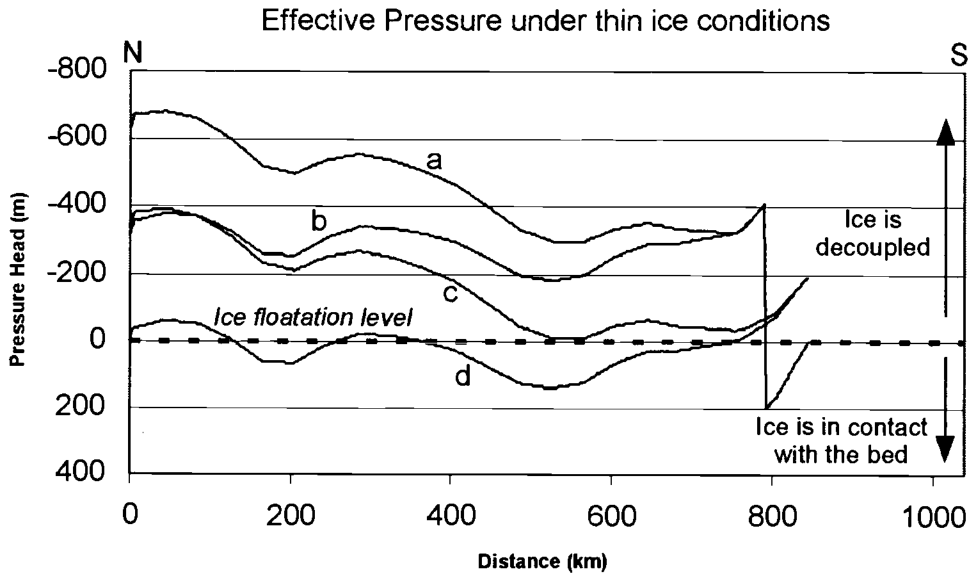
DISCUSSION AND CONCLUSIONS

Our simulations indicate that during its last glacial maximum, the LML 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 (Figure 9). Since flow through porous subglacial aquifers is clearly an inadequate mechanism for transporting meltwater, some additional drainage system must have existed.

A drainage system at the ice-bed interface, possibly consisting of a water film on the order of 7-8 mm thick and covering the entire glacier bed, would have been adequate to maintain subglacial water pressure at levels less than or equal to the ice-overburden pressure. Similar film-type drainage systems have been hypothesized as underlying Ice Stream B, Antarctica (Alley, 1989; Lingle and Brown; 1987, Engelhardt and Kamb, 1997).

Fast ice flow, such as that exhibited by the LML, requires a bed offering low yield stress, 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 to regions beyond the boundaries of the LML. Such a diversion would reduce subglacial pore pressure and/or drain any film at the ice-bed interface, significantly increasing the yield stress of the glacier bed. This should dramatically slow the advance of the ice.

Anandakrishan and Alley (1997) describe a comparable situation for ice Stream C in west Antarctica. They show that Ice Stream C slowed dramatically following the subglacial divergence of meltwater from Ice Stream C to the basin occupied by



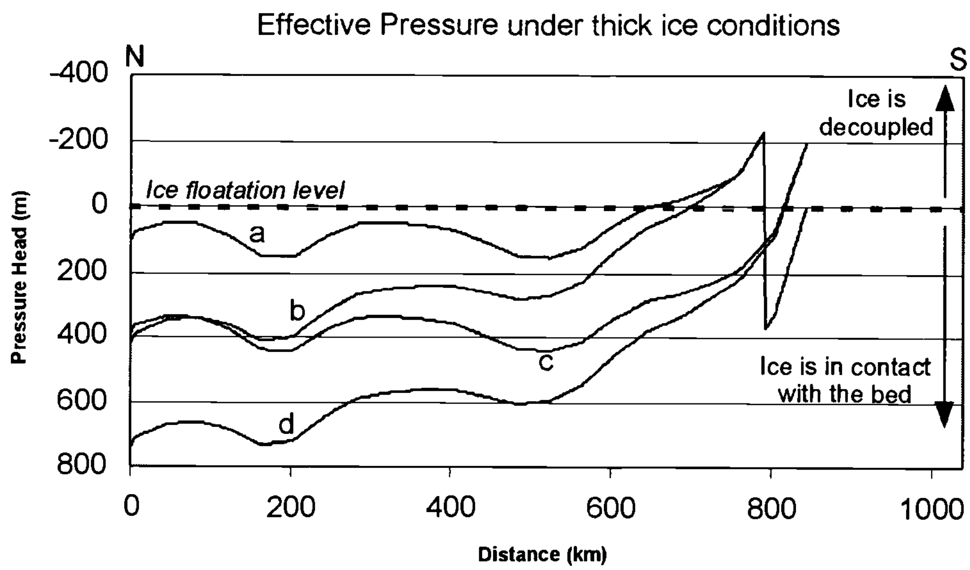


Figure 9: Effective pressure at the ice-bed interface. a) 100 km of permafrost brackets the ice margin and water flows through a 7 mm film. b) 100 km of permafrost and flow through an 8 mm film. c) No permafrost and a 7mm film. d) No permafrost and an 8mm film.

Ice Stream B. Perhaps it is not coincidental that the southern and western boundaries of the maximum LML are roughly parallel to the buried bedrock valleys (Figure 3). The transmissivity of the buried bedrock valleys in Illinois, and perhaps Indiana as well, may have been sufficient to drain water film and/or reduce pore pressure underneath the LML, and thus help define the margins 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. When permafrost forms in the Mahomet Bedrock Valley, our simulations show a dramatic increase of subglacial water pressure. Cutler et al. (in press) calculate that the bed of the Green Bay Lobe was frozen for 60-200 km upstream of the ice sheet margin. If more than 100 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. Research in Illinois indicates that permafrost existed near the margin of the Lake Michigan Lobe during the last glacial maximum. Based on sediment stratigraphy, however, Johnson (1990) concluded that the permafrost formed only in ice marginal areas. If Johnson is correct then we can conclude that permafrost did not block the Mahomet Bedrock Valley, and thus had little effect on ice behavior and subglacial hydrology.

The diversion of water into the Mahomet Bedrock Valley has implications beyond simply minimizing the possible effects of permafrost. The Mahomet Valley apparently had the capacity to divert nearly all of the subglacial meltwater to areas beyond the ice margin.

Our numerical simulations suggest the following. 1) Aquifers underneath the last glacial maximum LML of the southern Laurentide Ice Sheet were incapable of draining the estimated basal meltwater. 2) Meltwater in excess of that transmitted through porous or fractured aquifers may have been discharged through some type of water film at the ice-bed interface. A film 7-8 mm thick, 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 virtually all of the basal meltwater, leading to substantially reduced subglacial pore pressure, and thus removing the mechanisms of fast ice flow and 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 glacial maximum, groundwater flow patterns and velocities were altered or reversed relative to non-glacial periods.

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