Influence of the Great Lakes on the dynamics of the southern Laurentide ice sheet: Numerical experiments

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ABSTRACT

A time-dependent, flow-line ice-sheet model is used to explore interactions between last glacial climate, the Laurentide ice sheet, and the Great Lakes. We seek to understand perturbations in the ice-flow field that caused three neighboring lobes to leave different geomorphic and sedimentary records. Driven by reconstructed air-temperature variations from 65 to 18 ka, the simulated lobe dynamics are consistent with constraints from loess and till chronologies, till stratigraphy, and ice-surface profiles. Contrasts in lobe dynamics are best explained by the geometry of lakes along each flow line. As ice entered these lakes, calving influenced glacier mass balance. In particular, deep water in Lake Superior likely delayed ice advance into northern Wisconsin by promoting large calving losses, whereas lobes to the east that encountered shallower water were less affected by calving. The Driftless Area of Wisconsin may thus owe its existence, at least in part, to the presence of Lake Superior. Our results suggest that morainal-bank evolution should be treated in ice-sheet models in lacustrine and shallow-marine settings. Determining the sediment flux to the morainal bank remains a difficult task. For example, ice advance across central Lake Superior was probably sediment-flux limited. Otherwise, the Driftless Area would have been glaciated.

Keywords: Laurentide ice sheet, numerical model, Great Lakes, glacial geology, mass balance.

INTRODUCTION

The Laurentide ice sheet profoundly influenced global climate, sea level, meltwater routing, and landscape evolution during the last glaciation (Clark et al., 1999). For example, fluctuations of the southern ice margin modulated meltwater fluxes to the North Atlantic, influencing the thermohaline circulation (Licciardi et al., 1998). However, little is known of ice-sheet history prior to the Last Glacial Maximum (LGM, ~21 ka) because of limited time constraints and physical evidence, particularly southwest of James Bay (Clark et al., 1993). This paper explores the influence of the Great Lakes on lobes of the Laurentide ice sheet as these lobes advanced. Our numerical modeling experiments suggest sources of perturbation in the ice sheet flow field that led to striking contrasts in lobe dynamics and extent at the LGM.

The importance of permafrost to Laurentide ice sheet behavior and deposits, in particular its impact on subglacial drainage and sliding, is discussed in Cutler et al. (2000). Another major influence on lobes entering the Great Lakes region was the presence of large lakes. We investigate the effects of the lobes advancing over a permafrost-free lake bottom and of the calving that would occur. These effects may help to explain contrasts in LGM extent and sediment-landform assemblages of the adjacent Lake Michigan, Green Bay, and Langlade lobes (Fig. 1A; Mickelson et al., 1983). For example, multiple till sheets with interbedded proglacial deposits of the Lake Michigan lobe attest to a rapidly fluctuating lobe with a thawed bed. In contrast, the Green Bay lobe deposited a single LGM till and formed high-relief moraines that suggest greater stability than the Lake Michigan lobe. Tunnel channels near the margin of the Green Bay lobe, but not the Lake Michigan lobe, hint at the release of subglacial meltwater trapped behind an ice-margin frozen-bed zone (Attig et al., 1989). Tunnel channels are accompanied by ice-walled lake plains in deposits from the Langlade lobe, suggesting stagnating ice on a deeply frozen substrate (Ham and Attig, 1996). Last, in an area of overlap, Langlade lobe till overlies LGM till of the Green Bay lobe (Mickelson, 1986), indicating later arrival of Langlade ice at its maximum extent.

These lobes bordered the Driftless Area in Wisconsin, an area that remained unglaciated at the LGM. The northern part of this area was glaciated in pre-Wisconsin episodes, as were areas south of LGM deposits in Illinois (Fig. 1A). Debate on the cause of the Driftless Area

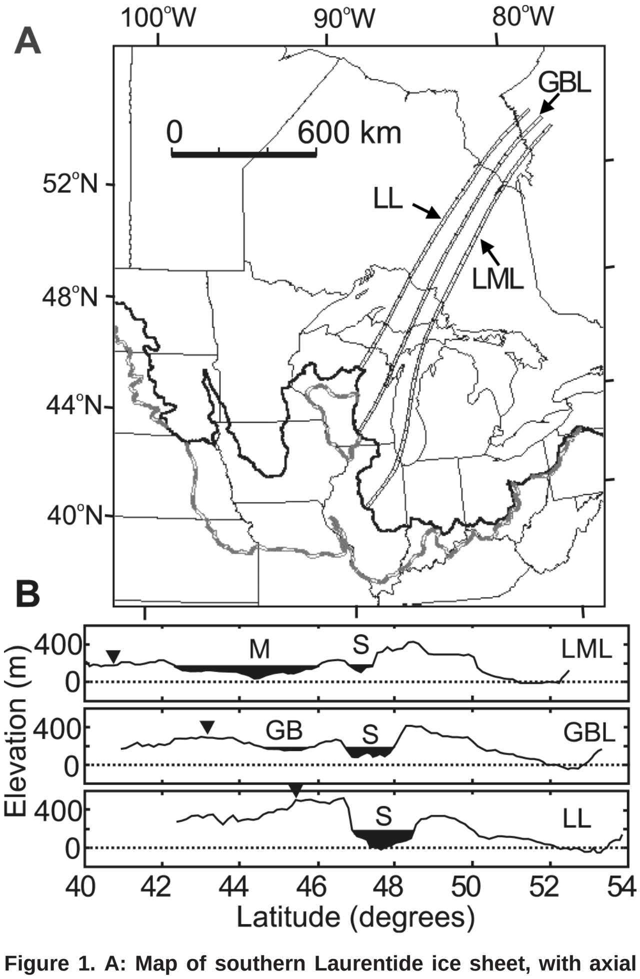

Figure 1. A: Map of southern Laurentide ice sheet, with axial flow lines of Green Bay lobe (GBL), Lake Michigan lobe (LML), and Langlade lobe (LL). Thick black line—LGM (Last Glacial Maximum) extent, gray dashed line—pre-Wisconsin maximum extent. B: Cross-section profiles along flow lines from A. Dark areas are lake basins, M—Lake Michigan, S—Lake Superior, GB—Green Bay. Triangles indicate LGM ice extent.

focuses on the nature of the bed, the up-ice-flow presence of high terrain, and the basins of Lake Michigan and western Lake Superior down which ice was funneled (cf. Hobbs, 1999). Here, we consider the possibility that deep water in Lake Superior induced calving rates during advance that were high enough to prevent LGM ice incursion into the Driftless Area.

METHODS

Outline of Model

A finite-element ice-sheet model is used to explore interactions among the atmosphere, cryosphere, and lithosphere. The model is two-dimensional, time-dependent, and thermomechanically coupled (Cutler et al., 2000; Parizek, 2000), with a domain that includes the upper 1 km of the lithosphere and accommodates isostasy. The horizontal node spacing is about 17 km, and runs are performed with time steps of 10 yr. A sliding law is implemented in some runs (Cutler et al., 2000). Within this law, a sliding parameter allows differentiation between hard, mostly igneous bedrock and soft, sedimentary bedrock (Clark et al., 1996). Parameter values are selected such that sliding speeds are 20–200 m/yr with a basal shear stress (τ) of 10 kPa over soft rocks and 1–10 m/yr for $\tau = 50$ kPa over hard rocks.

Mass Balance

The mass-balance formulation follows Greve and MacAyeal (1996) and accommodates flow divergence. Mass input from precipitation is assumed to follow modern spatial trends (Legates and Wilmott, 1990), but it declines with elevation on the ice sheet to a minimum of 0.3 m/yr at about 3000 m (cf. Vettoretti et al., 2000). The fraction of precipitation falling as snow is estimated following Marshall et al. (2000).

Mass is lost by surface and basal melting and by calving. The surface-melting rate is calculated by using the positive degree-day (PDD) method (Reeh, 1991), requiring an estimate of air temperature at each node. The latitudinal temperature gradient is 0.8 K/degree and the elevation lapse rate is 0.008 K/m. We adopt an empirical calving relationship for grounded glaciers terminating in fresh water in which calving speed scales linearly with water depth and exponentially with ice temperature (Warren et al., 1995; Marshall et al., 2000). A related model component treats morainal-bank evolution: we use methods and parameter values of Alley (1991). As with tidewater glaciers, morainal banks can assist ice across deeper basins by locally reducing the water depth. In the model, sediment is delivered to the moraine either by melt out of debris-rich basal ice at a rate governed by calving speed or by till advection in a thin, deforming layer if the bed is thawed and sliding is implemented.

Input Data

Ice flowed into the Great Lakes from a spreading center east of James Bay at the LGM (Fig. 1A; Veillette et al., 1999). The bedrock distribution along each flow line is differentiated as either hard or soft following Licciardi et al. (1998). Pre-Wisconsin topography is unknown, although only meters to tens of meters of erosion probably occurred on the Canadian Shield during the Quaternary (Shilts et al., 1987). Flow-line topography is thus taken from the U.S. Geological Survey ETOP05 data set. Spatial trends in striations and drumlins are used to estimate flow divergence in the Great Lakes region, and a uniform value of 0.15%/km is assumed over the Canadian Shield. A geothermal flux of 50 mW/m^2 is applied at the base of the model domain.

Paleo–lake level is needed to estimate water depth in the calving calculations. Modern levels are assigned unless the relevant outlet falls below this level due to ice loading. The elevation of former Lake Superior outlets increases westward along the southern shore (Lineback et al., 1979), whereas Lake Michigan drained through the Chicago

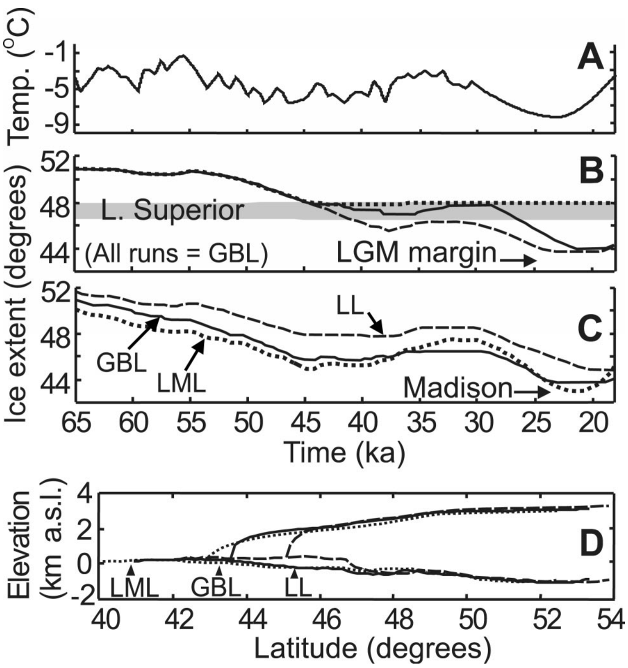

Figure 2. A: Estimated air temperature at Madison, Wisconsin, from 65 to 18 ka (see text for details). B: Simulated extent of Green Bay lobe from 65 to 18 ka, driven by temperatures shown in A. No sliding is allowed. Runs feature no calving or morainal-bank growth (dashed line), calving only (dotted line), and calving with morainal-bank growth (solid line). Gray shading shows extent of Lake Superior. C: Simulated extent of Green Bay lobe (GBL, solid line), Lake Michigan lobe (LML, dotted line), and Langlade lobe (LL, dashed line) from 65 to 18 ka, with sliding. D: Profiles of Green Bay lobe (solid line), Lake Michigan lobe (dotted line), and Langlade lobe (dashed line) at their maximum extent. Triangles show position of actual Last Glacial Maximum (LGM) moraines.

outlet whenever ice blocked the northern outlet (Hansel and Mickelson, 1988). There was potential for significantly deeper water in Lake Superior than in Lake Michigan during ice advance because ice sequentially blocked higher outlets along the southern shore of the lake.

The climate driving ice-sheet evolution is distilled to temperature and precipitation (Cutler et al., 2000). No consensus exists on the precipitation regime in the midwestern United States during glacial time, and a simple Arrhenius-type relationship between air temperature and moisture fails to capture the influence of changes in atmospheric circulation on precipitation (Vettoretti et al., 2000). A contemporary precipitation field is therefore retained in unglaciated areas, leaving air-temperature variation as the sole forcing function for ice-sheet evolution. Figure 2A illustrates the adopted temperature regime at the latitude and elevation of Madison, Wisconsin. This reconstruction is based on δ^{18}O variations from a Missouri speleothem record (Dorale et al., 1998) that begins at 65 ka and remains reliable until ca. 30 ka. After 30 ka, this record is replaced by a trend that follows summer insolation at 40°N. The reconstruction is consistent with (1) evidence from Minnesota indicating coldest conditions at 30–18 ka (Lively, 1983) and (2) evidence for LGM mean annual temperature (MAT) of at most −6 °C (Attig et al., 1989) and, perhaps, −10 °C near Madison (Vettoretti et al., 2000).

Initial Conditions and Model Sensitivity

No direct evidence exists for Laurentide ice sheet extent southwest of James Bay at 65 ka. Nonetheless, trends in global sea level (Clark et al., 1999), chronology in the James Bay Lowlands (Clark et al., 1993), onset of loess deposition in the Upper Mississippi Valley (Dorale et al., 1998), and a ^{10}Be inventory from Illinois (Curry and Pavich, 1996) point to ice extending to somewhere between James Bay and Lake Superior at 65 ka. Fortunately, model runs with initial ice positions from 50 to 400 km north of Lake Superior converge after 10 k.y., as do runs with initial ice and ground temperature fields that differ by up to 10 °C at each node. Initial ice cover is thus arbitrarily set at the northern 25% (~400 km) of each flow line, with an initial ice-surface profile that yields $\tau = 50$ kPa. The initial 10 k.y. of any model run is treated as model ''spin-up'' time. Ice extent is most sensitive to air temperature and precipitation. For example, perturbations of +10% in precipitation or −1 K in temperature increase ice extent by up to 250 km. In contrast, a 10% perturbation in flow divergence, calving rate, or sliding speed causes a maximum difference of a single node spacing (17 km).

RESULTS

In light of uncertainties in climatic forcing, we emphasize *qualitative*, but nonetheless significant, differences in lobe behavior. First, we examine the impact of calving on ice advance across Lake Superior (Fig. 2B). The ice is coupled to its bed in these runs. With no calving (dashed line), ice advance is controlled by climatic influences on mass balance. The impact of calving is severe (dotted line): though calving speed for a given water depth is roughly 10 times less than for tidewater glaciers (Warren et al., 1995), ice fails to traverse Lake Superior. Marshall et al. (2000), encountering a similar model result in Hudson Bay, noted that clogging of embayments by sea ice would diminish calving rates. The equivalent impact of lake ice may not apply in the Great Lakes region, because energy-balance calculations (per unit width) for the unglaciated part of a lake, using a conservative degree-day constant of 0.003 m per PDD, consistently yield enough energy to melt all icebergs and at least 2 m of lake ice per year. Such a thickness of lake ice fails to form today in the Brooks Range, Alaska, where the MAT is about −10 °C and precipitation is below Great Lakes values.

The influence of morainal-bank construction on ice advance is illustrated in Figure 2B (solid line). With no sliding in this run, sediment delivery to the moraine occurs solely by melt out of basal debris. The sediment flux is small, not exceeding 10 $m^3 \cdot m^{-1} \cdot yr^{-1}$ (cf. Alley, 1991), but perhaps reasonable for ice flowing over shield lithologies that were deeply frozen prior to ice occupation. Even this flux is sufficient to aid ice across Lake Superior, however. The thickness of the morainal bank reaches 120 m, less than the 300 m bank in front of the Columbia Glacier (B. Molnia, 2000, personal commun.). Ice attains its maximum extent later than in the noncalving case (dashed line), although with a factor-of-five increase in sediment flux, the calving and noncalving cases become almost identical.

Figure 2C contrasts the results of driving ice along the three flow lines in Figure 1A; the same temperature forcing is used in each case (Fig. 2A). Differences arise primarily from the lake geometry along each flow line. Sliding constants are nonzero, yielding sliding speeds at the lower end of the range already discussed. The Green Bay and Lake Michigan lobes reach the south shore of Lake Superior at about 50 ka, within the onset window for loess deposition in the Upper Mississippi Valley (Dorale et al., 1998). The Lake Michigan lobe remains north of 42.5°N prior to 25 ka (cf. Curry and Pavich, 1996) and advances and then retreats more rapidly than the Green Bay lobe because of its passage through the thawed Lake Michigan basin rather than across frozen terrain. This is consistent with the glacial record. The surface profile of the Lake Michigan lobe near its margin is shallower than that of the Green Bay lobe (Fig. 2D), also a result of the thawed versus frozen bed conditions, and supported by independent reconstructions (Colgan, 1996).

The Langlade lobe reaches its maximum extent later than the other two lobes (Fig. 2C), consistent with till stratigraphy (Mickelson, 1986) and geomorphology (e.g., Ham and Attig, 1996). In the model, the delay is caused by calving losses while ice traverses the deepest part of Lake Superior (Fig. 1B).

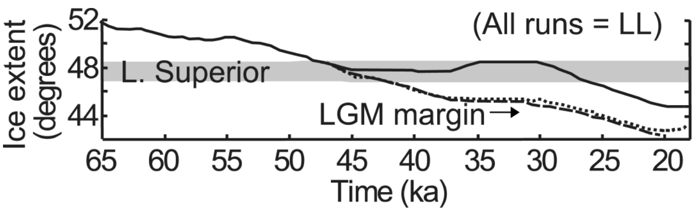

Figure 3. Impact of halving initial depth of Lake Superior on extent of Langlade lobe (dotted line), compared to baseline run with modern bathymetry (solid line). Also contrasted: runs with basal-ice sediment concentrations of 0.1 m^3/m (dashed line) and 0.01 m^3/m (solid line). Gray shading indicates extent of Lake Superior. LL—Langlade lobe; LGM—Last Glacial Maximum.

The Green Bay and Langlade lobe simulations terminate within tens of kilometers of their respective targets, whereas the Lake Michigan lobe simulation is far short of its goal (Fig. 2D). Clearly, reduced or negative flow divergence along sections of the flow line, increased precipitation, or surging could have driven the lobe farther south. Runs in which baseline precipitation was increased by 10%, or basal sliding on soft beds was 200 m/yr for $\tau = 10$ kPa instead of 20 m/yr, terminated an additional degree south but still short of the target by 1°. Given the weak constraints on the precipitation and divergence fields, and no treatment of surge mechanisms, little is gained by pushing the model further.

In light of the unknown bathymetry of the pre-Wisconsin Great Lakes, we explored the impact on lobe dynamics of halving the initial lake depth in Lake Superior and Lake Michigan. This extreme change affects the Langlade lobe (Fig. 3) but has little effect on the Lake Michigan lobe (not shown), because modern Lake Michigan is shallower than Lake Superior (Fig. 1B) and the Chicago outlet is tens to more than 100 m below Superior outlets. This result argues against a shallower pre-Wisconsin Lake Michigan basin as a major influence on earlier, more extensive glaciations.

In the shallow-lake case for the Langlade lobe (Fig. 3, dotted line), ice readily traverses Lake Superior and ultimately terminates far south of the known margin. Without enhanced calving in a deep Lake Superior, the Driftless Area could have been inundated by ice at the LGM (cf. Marshall et al., 2000). A similar result is caused by increasing sediment supply to the morainal bank to about 50 $m^3 \cdot m^{-1} \cdot yr^{-1}$ (Fig. 3, dashed line), while retaining modern lake bathymetry. Within the limitations of the simplistic treatment of calving and morainal-bank evolution, this result hints that the maximum extent of the Langlade lobe and other lobes in north-central Wisconsin was sediment-flux limited when ice crossed Lake Superior.

DISCUSSION

Our results illustrate how a flow-line model can test explanations of lobe behavior in the Great Lakes region. Differences in calving rates and morainal-bank growth are shown to provide considerable perturbations to the flow field as ice advanced into major lake basins. Such perturbations were likely magnified over time as lobes like the Lake Michigan lobe captured ice from their neighbors.

An untreated factor in our model is ice-shelf development. Four arguments reduce its likely importance, however. First, ice shelves are absent where mean air temperature in the warmest month of the year exceeds 0 °C (Mercer, 1968). The reconstructed temperature regime yields values of at least this magnitude whenever ice is near Lake Superior. Second, ice shelves, once established, are primarily nourished by precipitation. Mass-balance calculations indicate strong net melting that would cause shelf disappearance in a few hundred years. Third, thick ice entering the lake on hard shield rocks favors grounding. Flotation is not achieved during advance within the range of basal motion discussed earlier, although sub–grid-scale flotation is possible at the margin. Fourth, if ice were able to traverse Lake Superior aided by a large morainal bank, a shallower lake or a shelf, then the climate was probably favorable for a more extensive advance than actually occurred (Fig. 3). We therefore suspect that shelf development was not influential in ice-sheet expansion across Lake Superior.

The precipitation regime is a key unresolved factor in Laurentide ice sheet simulations. A LGM climate simulation by Vettoretti et al. (2000) shows focused moisture transport to the southern Great Lakes. This effect is certainly favorable for inducing the observed trend in ice extent, but proxy records of paleoprecipitation provided by fossil ostracodes and pollen in Illinois do not concur with this result (Curry and Baker, 2000).

CONCLUSIONS

Our time-dependent, flow-line ice-sheet model provides insight into the causes of contrasting glacial records of the Lake Michigan, Green Bay, and Langlade lobes of the Laurentide ice sheet. Driven by reconstructed temperature variations from 65 to 18 ka, the simulated dynamics of each of these lobes are in reasonable agreement with loess and till chronology, till stratigraphy, and ice-surface profiles. Differences in timing and extent of lobes were enhanced by the geometry of lake basins and differences in calving losses therein. Deep water in Lake Superior potentially delayed ice advance, whereas lobes to the east encountered shallower water and were less inhibited by calving during ice advance. The Driftless Area of Wisconsin may owe its existence, at least in part, to the presence of a deep Lake Superior basin.

Morainal-bank evolution should be treated in ice-sheet models applied to lacustrine and shallow-marine settings, although determining the sediment flux to the bank remains a difficult task. For example, ice advance across Lake Superior was probably sediment limited; if it had not been, the climate appears favorable for ice occupation of the Driftless Area. Other remaining challenges include estimating pre-Wisconsin bathymetry of the Great Lakes, developing a physically based calving treatment, and determining the precipitation and ice-flow divergence fields.

ACKNOWLEDGMENTS

Research was funded by National Science Foundation grants EAR 96-27798 and 98-14371. We thank J. Dorale and B. Curry for advice and data, and S. Hostetler and J. Licciardi for their constructive reviews.

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Manuscript received February 20, 2001
Revised manuscript received July 11, 2001
Manuscript accepted July 27, 2001

Printed in USA