Modeling the deglaciation of the Green Bay Lobe of the southern Laurentide Ice Sheet

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We use a time-dependent two-dimensional ice-flow model to explore the development of the Green Bay Lobe, an outlet glacier of the southern Laurentide Ice Sheet, leading up to the time of maximum ice extent and during subsequent deglaciation (*c*. 30 to 8 cal. ka BP). We focus on conditions at the ice-bed interface in order to evaluate their possible impact on glacial landscape evolution. Air temperatures for model input have been reconstructed using the GRIP δ^{18}O record calibrated to speleothem records from Missouri that cover the time periods of *c*. 65 to 30 cal. ka BP and 13.25 to 12.4 cal. ka BP. Using that input, the known ice extents during maximum glaciation and early deglaciation can be reproduced reasonably well. The model fails, however, to reproduce short-term ice margin retreat and readvance events during later stages of deglaciation. Model results indicate that the area exposed after the retreat of the Green Bay Lobe was characterized by permafrost until at least 14 cal. ka BP. The extensive drumlin zones that formed behind the ice margins of the outermost Johnstown phase and the later Green Lake phase are associated with modeled ice margins that were stable for at least 1000 years, high basal shear stresses (*c*. 100 kPa) and permafrost depths of 80–200 m. During deglaciation, basal meltwater and sliding became more important.

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Deposits and former extents of the southern Laurentide Ice Sheet have been studied extensively in the field (e.g. Mickelson *et al.* 1983; Dyke & Prest 1987; Attig *et al.* 1989, and references therein) and glacial landscape assemblages have been compiled in a comprehensive GIS database (Colgan *et al.* in press). Dating of end moraines provides a time frame for the ice sheet's extent at different stages. However, many aspects of the evolution of the southern Laurentide Ice Sheet since the Last Glacial Maximum (LGM) still remain unknown or are controversial: Did the ice retreat early and progressively or late and rapidly? Where and when did permafrost develop during ice retreat? What were the conditions at the ice-bed interface that led to the formation of distinct glacial landscapes that differ within one lobe (from phase to phase) or between neighboring lobes? For example, how did the ice-bed interface conditions of ice-sheet phases associated with drumlin formation differ from 'drumlin-free' phases?

Timing of the ice retreat is, in a broader context, important with regard to the impact of climate forcing on ice-sheet oscillations (e.g. McCabe & Clark 1998). Linking subglacial conditions and ice-sheet behavior to landforms is useful when interpreting the glaciation history of similar locations.

Numerical ice-sheet models are a valuable tool for predicting and quantitatively evaluating subglacial environments. Conditions postulated from glacial landscape interpretations can thus be tested independently. Glacier-bed conditions for the southern Laurentide Ice Sheet have been inferred from landform distributions (Mickelson *et al.* 1983; Attig *et al.* 1989; Johnson & Hansel 1999). Previous model studies focused mainly on extent and volume of the Laurentide Ice Sheet in general around the LGM (e.g. Peltier 1994; Clark *et al.* 1996; Fabre *et al.* 1997; Marshall *et al.* 2000, 2002) and on deglaciation aspects of the northern hemisphere ice sheets with relatively low spatial resolution (e.g. Deblonde *et al.* 1992; Licciardi *et al.* 1998; Marshall & Clarke 1999; Charbit *et al.* 2002). For the southern Laurentide Ice Sheet, the importance of permafrost and the role of calving and morainal-bank evolution have been investigated for the period of ice advance up to the LGM (Cutler *et al.* 2000, 2001). However, in order to examine the questions raised above, it is necessary to carry out high-resolution transient model runs that cover not only the time of advance to the LGM, but also at least part of the subsequent deglaciation.

In this study, we focus on modeling the evolution of the Green Bay Lobe for the time since the LGM in order to test the validity of using a tuned ice core δ^{18}O record from Greenland as a proxy for temperature, to investigate the behavior of the ice sheet under deglaciation conditions, and to explore the possibilities of connect-

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ing model results to specific glacial landscape features. We use the GRIP $\delta^{18}O$ record (Dansgaard *et al.* 1993) adjusted to existing speleothem $\delta^{18}O$ records from Missouri (Dorale *et al.* 1998; Denniston *et al.* 2001) in order to provide us with a temperature input throughout the time of deglaciation. Precipitation is mostly parameterized following modern values and gradients. The only geologic constraints placed on input data and used for model validation are bed materials, topography and ice extents. This independence from using geologic data as input allows us to examine landform distribution and suppositions concerning basal conditions as unconstrained variables that can be compared to model results.

The study area

The extent of the southern Laurentide Ice Sheet has been summarized for the LGM (e.g. Mickelson *et al.* 1983; Attig *et al.* 1985, 1989; Dyke & Prest 1987) and several phases during deglaciation of the Green Bay Lobe (e.g. Clayton & Moran 1982; Mickelson *et al.* 1983; Attig *et al.* 1985; Maher & Mickelson 1996; Colgan 1996, 1999; Colgan & Mickelson 1997). In a recent review paper, Dyke *et al.* (2002) stated that the advance of the Laurentide Ice Sheet to its maximum extent started at 30–27^{14}C ka BP, the ice sheet reached its maximum southern extent at *c.* 23 ^{14}C ka BP, and rapid ice margin recession started around 14 ^{14}C ka BP. Five margin positions of the Green Bay Lobe are shown in Fig. 1. Their dating is partly uncertain; numbers given here represent our best estimate, based on previous studies (Attig *et al.* 1985; Maher & Mickelson 1996; Colgan 1999; Clayton *et al.* 2001). The phases are called Johnstown (maximum extent, before and after LGM), Green Lake (around 14.5 ^{14}C ka BP, $\sim$17 cal. ka BP), Chilton (around 13 ^{14}C ka BP, $\sim$15 cal. ka BP), Two Rivers (starting at *c.* 11.8 ^{14}C ka BP, $\sim$13.8 cal. ka BP), and Marquette (at *c.* 9.9 ^{14}C ka BP, $\sim$10.9 cal. ka BP) (see Table 1). Rates of ice retreat and readvance are controversial (e.g. Maher & Mickelson 1996; Colgan 1999), but in general the Green Bay Lobe seems to have been more stable and shows less ice margin fluctuation than the adjacent lobes, as can be inferred, for example, from the deposition of only one till unit before about 13 ^{14}C ka BP (Clayton & Moran 1982; Johnson & Hansel 1999). Retreat from the Johnstown moraine either occurred relatively early and progressively at rates of about 50 m/yr (Colgan 1999), or it took place late and quite rapidly at rates of 300 to 900 m/yr (Maher & Mickelson 1996). Between the Chilton and the Two Rivers phases, the Two Creeks forest developed, which was then overridden by ice again at *c.* 11.8 ^{14}C ka BP.

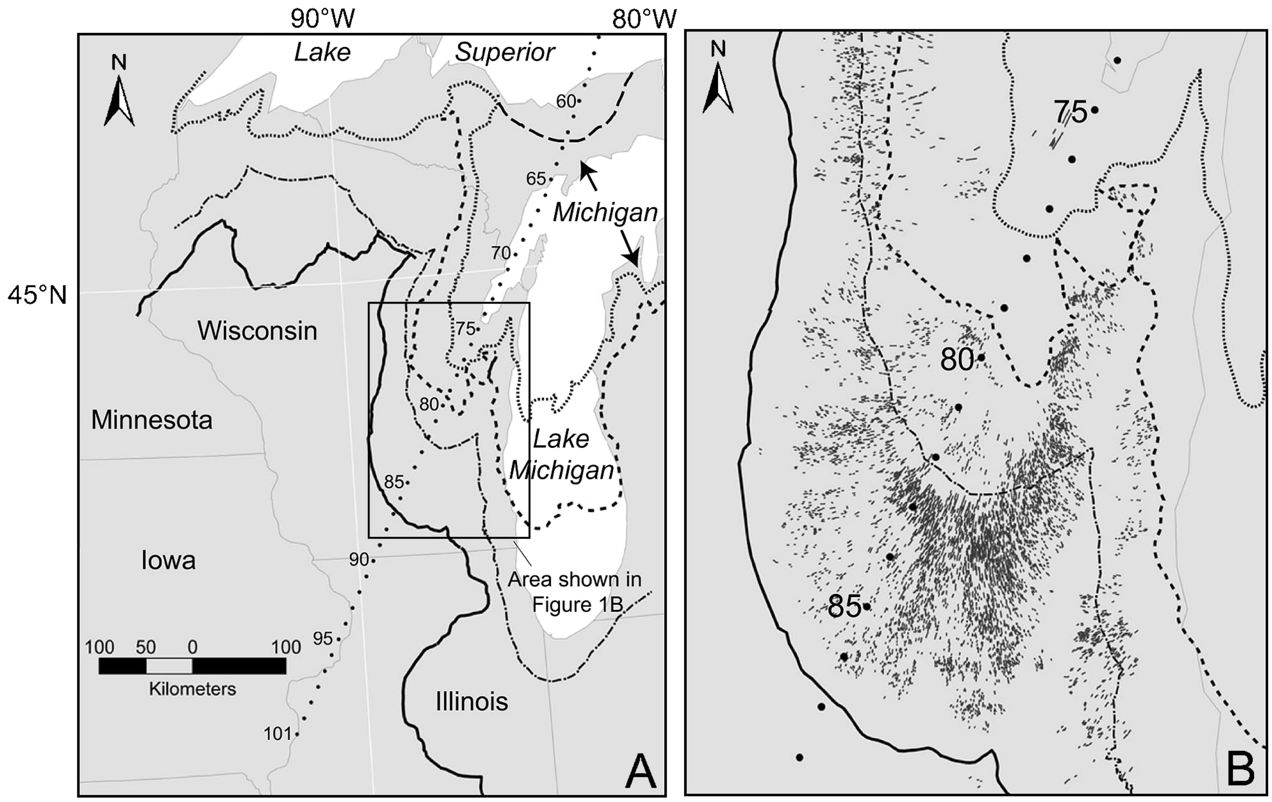
The glacial landscape of the area covered by the Green Bay Lobe is characterized by narrow (<1 km), moderate-to-high-relief (10–40 m) end moraines in the ice-marginal zone, a zone with mainly rolling till plain behind these moraines, and a zone with a surface extensively streamlined by the ice flow farther up-ice (Colgan 1999). Drumlins and megaflutes characterize this subglacial zone. Most drumlins are between 1 and 6 km long, but some superimposed drumlins are smaller. Most sediment in them appears to predate the drumlin-forming phase (Colgan & Mickelson 1997). Sandy diamicton was deposited during the Johnstown and Green Lake phases, and lake-sediment-derived silty-clayey diamicton was deposited during the Chilton and Two Rivers phases (McCartney & Mickelson 1982). Moraine relief is much higher for the Johnstown and the Green Lake phases (15 and 10 m) and the slopes are steeper (*c.* 0.0018) than for the Chilton and the Two Rivers phases (5 and 3 m relief and slope values of 0.001) (Colgan & Mickelson 1997). Permafrost features (e.g. ice-wedge casts and ice-wedge polygons as well as a lack of trees) have been documented in Wisconsin for the period of *c.* 26 to 13 ^{14}C ka BP (Attig *et al.* 1989; Clayton *et al.* 2001). Tunnel channels near the outermost margin of the Green Bay Lobe suggest the release of subglacial meltwater trapped behind an ice-marginal frozen-bed zone (Cutler *et al.* 2002). Eskers are present near the former ice margin and also in the drumlin zone; they probably formed during ice wastage when the ice base was thawed (Attig *et al.* 1989).

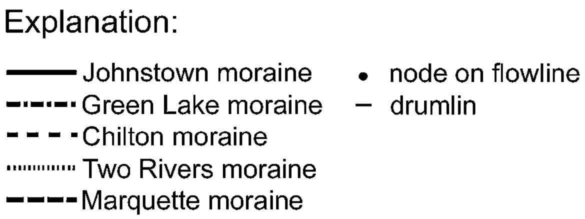
The Green Bay Lobe flowline we use in this study starts northeast of James Bay and ends south of Madison (see Fig. 1 for its southern part). The northernmost *c.* 400 km of the flowline is underlain by Paleozoic carbonate rocks, followed by *c.* 100 km of carbonate-rich till that overlies crystalline rocks. South of this, bedrock is mostly exposed to the southern edge of Lake Superior. Paleozoic sedimentary rock (mainly Ordovician dolomite and limestone) forms the base along the flowline south of Lake Superior (cf. Cutler *et al.* 2000, and references therein).

The model

Model outline

We use a two-dimensional, time-dependent, thermo-mechanically coupled finite-element ice model that includes flow divergence. The model has been used and described in detail by Cutler *et al.* (2000, 2001) and Parizek (2000). Horizontal ice velocity consists of ice deformation and sliding velocity, the latter occurring only if the temperature at the ice base is at the pressure melting point at two or more neighboring nodes. Glen's flow law is used for ice deformation; sliding parameterization follows Payne (1995) and Greve & MacAyeal (1996). Sediment deformation is not treated separately. By choosing appropriate sliding parameters for 'hard', predominantly igneous and 'soft', sedimentary bedrock, the sliding law aims at incorporating the influence of the substrate geology on basal motion, with





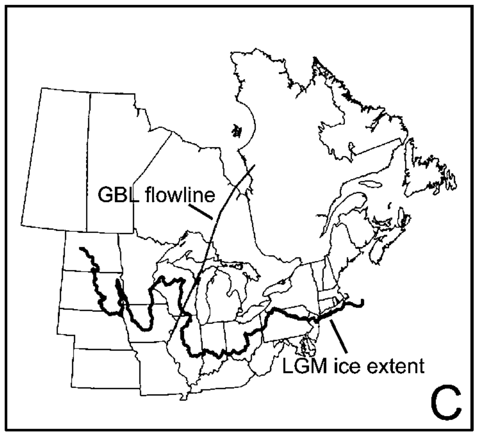


Fig. 1. Maximum extent and several phases during deglaciation of the southern Laurentide Ice Sheet in the Green Bay Lobe area (A). Part of the flowline used here with numbered nodes and the occurrence of drumlin fields (B), as well as the location of the complete flowline used in this study (C), are also shown.

Table 1. Phases of the Green Bay Lobe and their timing derived from geologic field data in comparison with the extents and their durations generated by the ice model (Experiment 01). Also given are the flowline nodes that are considered in the study and correspond to the modeled ice margin and a relevant zone up-ice.

Phase (field data)	Timing derived from field data (cal. ka)	Modeled ice extent	Duration of modeled ice extent (cal. ka)	Flowline nodes
Johnstown	Around 21	Maximum stillstand	22.2–17.2	78–83
Green Lake	Around 17	2nd stillstand	17–16 (or 14.2)	77–82
Chilton	Around 15	Model year 14.1	14.1	77–79
Two Rivers	At 13.8	Model year 13.7	13.7	75–77
Marquette	At 10.9	3rd stillstand	11.6–10.5	61–62

'soft' bed facilitating fast ice flow (Clark *et al.* 1999). The model includes permafrost development and calving into freshwater lakes and accommodates for isostasy. Model values along the Green Bay Lobe flowline are calculated at 101 almost equally spaced nodes (resulting in a horizontal resolution of about 17 km) and for 25 ice and 36 bedrock layers (down to a depth of 1830 m below the surface). The timestep of the runs was set to 10 years. Information on topography, bedrock, sediment thickness and flow divergence (estimated from striations and drumlins) is read into the model from a GIS database (Colgan *et al.* in press). A geothermal flux value of 50 mW m^{-2} is applied at the base of the model domain.

Single nodes in areas of special interest along the modeled flowline are monitored through time for certain parameters (e.g. ice velocities, meltwater production, basal shear stress) in order to provide a detailed record of ice-bed interface conditions for the interpretation of landform genesis.

Time-transient climate parameterization

The most important parameters driving the model are temperature and precipitation. It is in general a great challenge to acquire an appropriate climate input for time-transient model runs. Cutler *et al.* (2001) used δ^{18}O variations from a U/Th-dated speleothem record in Missouri, USA (Dorale *et al.* 1998), which is located *c*. 200 km south of the LGM ice margin, for reconstructing the paleo-temperature from 65 to 30 ka BP (the only period of continuous record). The mean summer insolation at 40°N served as a temperature proxy for the period from 30 to 18 ka BP.

In this study, we test the use of the GRIP δ^{18}O record (Dansgaard *et al.* 1993) in conjunction with the local speleothem records in order to provide us with a longer time series of temperature through deglaciation (up to *c*. 8 ka BP). The time scale of the GRIP δ^{18}O record has been revised by Johnsen *et al.* (2001). It is too young for events older than 14.5 ka BP, up to *c*. 2000 years for the time-span we are investigating in this study. These discrepancies are taken into account in the discussion of the results. The GRIP record was interpolated onto the 500-year timesteps of the speleothem record. It was then adjusted to the speleothem record by modifyng the amplitude (by a factor of 0.45) and adding a constant (13.25). The obtained δ^{18}O record was converted into temperature using the relationship that had been found appropriate for the speleothem record (a change of 0.35δ δ^{18}O corresponding to a temperature change of 1°C; see Dorale *et al.* (1998) for details on effects that have been taken into account and a description of possible error sources). Fig. 2A illustrates the mean annual temperature at the latitude and elevation of Madison generated using the speleothem record that had been used previously (55–30 ka BP) and using the adjusted GRIP δ^{18}O record (55–8 ka BP). The modeled temperature shows a drop of *c*. 3.7°C at 13.5–12.0 ka BP, which agrees well with estimates from the Missouri speleothem record described by Denniston *et al.* (2001). Furthermore, the curve is consistent with evidence from Minnesota indicating coldest conditions between 30 and 18 ka BP (Lively 1983) and with field evidence for LGM temperatures at the ice margin of at most −6°C, and probably lower (Attig *et al.* 1989). Additional support for our approach of establishing a paleo-temperature input curve comes from a study by Lowell *et al.* (1999) that suggests a connection between the millennial-scale phasing of the Greenland ice core record and the expansion of the Laurentide Ice Sheet at several well-dated sites.

The input temperature is varied along the flowline using an elevation lapse rate (0.008°C/m; Huybrechts & T'siobbel 1995) and a latitudinal lapse rate (0.8°C/degree latitude). In addition, one run uses a steepening temperature gradient south of the advancing ice sheet that is up to 1.5 times the modern temperature gradient and which decreases again during ice retreat (in accordance with GCM results, e.g. Bartlein *et al.* 1998) in order to test the model's response in terms of ice build-up and the other parameters of particular interest for this study.

Proxy data from ice-free areas in Illinois around the LGM suggest similar precipitation values as today (Curry & Baker 2000); GCM results predict precipitation similar to or higher than today for the area of the southern Laurentide Ice Sheet (e.g. Kutzbach *et al.* 1998; Bartlein *et al.* 1998; Vettoretti *et al.* 2000). Therefore, baseline precipitation in most model simulations is kept at the modern value but changes with latitude (0.03 m/degree latitude, decreasing northwards) and longitude (0.015 m/degree longitude, increasing eastwards), following modern trends. Precipitation decreases with elevation up to a minimum on the ice sheet of 0.3 m/yr at 3000 m (cf. Cutler *et al.* 2001, following Vettoretti *et al.* 2000). Equations treating the variation of temperature and precipitation along the flowline as well as the most important constants used in the model are listed in Cutler *et al.* (2000).

Model sensitivity to some of the climate input parameters was tested; examples are described and discussed later in the text.

Experiments and results

In Experiment 01 we used the speleothem record as a temperature proxy for the time between 65 and 30 ka BP (as this had already produced results in earlier studies that satisfied age and ice extent constraints; Cutler *et al.* 2001) and prolonged it after 30 ka BP with the adjusted GRIP record. This experiment yielded reasonable results in terms of ice extent and duration of ice stillstands (validated by the geologic field record) and was therefore chosen as our standard run (Fig. 2B).

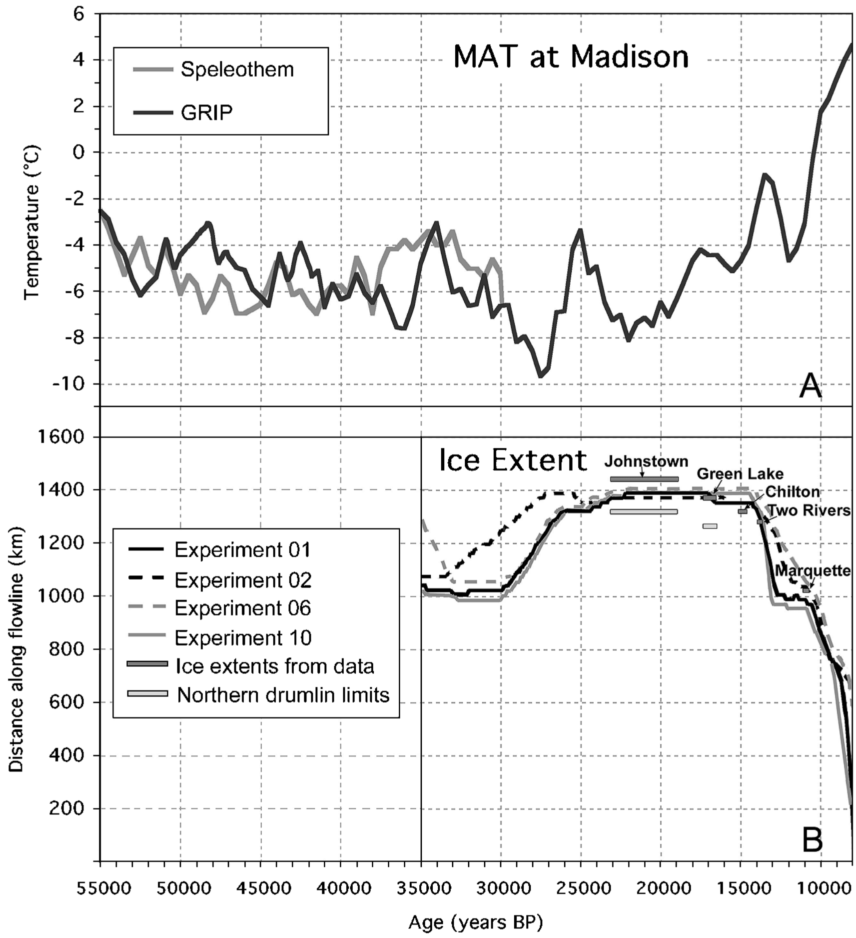


Fig. 2. A. Mean annual temperature at Madison, generated by using $\delta^{18}O$ from a speleothem record in Missouri (55–30 ka BP) and an adjusted GRIP $\delta^{18}O$ record (55–8 ka BP). B. Modeled ice extents for Experiments 01, 02, 06, 10. The ice extents known from field data as well as the northern margin of the drumlin fields associated with the Johnstown and the Green Lake phases are marked.

Several sensitivity experiments were carried out in order to explore the effects of varying climate parameters within a plausible range on the model results, mainly with regard to ice-bed interface conditions during the time of deglaciation (Table 2). Experiment 02 used the adjusted GRIP record for the whole run instead of the speleothem-GRIP combination. In Experiment 03, the effect of a changing temperature gradient south of the ice margin (see above) was tested. Experiment 04 explored the effect of a precipitation increase (precipitation factor) by 10% during the time of maximum glaciation (25–19 ka BP). Shifts of temperature input by +1 and −1°C were tested in Experiments 05 and 06. Experiments 07 and 08 dealt with changes in the difference between mean annual temperature and summer temperature (Tsummer_diff) and in the standard deviation of daily temperature (sigma) for the time of deglaciation (19–11 ka BP). In Experiment 09, the GRIP record (after 30 ka BP) was sampled at higher resolution (100 instead of 500 years). Lastly, we explored the effect of enhanced sliding over soft bed (by increasing the sliding parameter) in Experiment 10.

We let the runs start at 65 ka BP in order to allow the ice to build up to its maximum extent. The initial 10 000 model years, however, are treated as 'spin-up' time to allow for possible errors in the set of initial ice cover and initial ice and ground temperature (see Cutler *et al.* 2001). In the following discussion, we focus on the results for the time of 30 to 8 ka BP. The flowline nodes that cover the ice margin and a marginal zone of 17 to 89 km up-ice are 78–83 (1298–1387 km from the ice divide) for the maximum extent generated by the model (corresponding to the Johnstown phase) and 77–82 (1281–1369 km from the ice divide) for the second model-ice stillstand (corresponding to the Green Lake phase). Nodes 77–79 (1281–1316 km from the ice divide) are relevant for the modeled ice extent corresponding to the Chilton phase, 75–77 (1246–1281 km from the ice divide) for the model ice margin corresponding to the Two Rivers phase, and 61–62

Table 2. List of experiments. Experiment 01 was treated as standard. Changes in Experiments 2–9 compared to Experiment 01 are marked by italics. For explanation of parameters see text.

Experiment	Temperature input	Parameters
01 (standard)	Speleothem (until 30 ka BP) and GRIP record (after 30 ka BP)	Precipitation factor = 1.0 Tsummer_diff = 15°C Sigma = 5.0 Sliding parameter = 2.0×10^{-3} m yr^{-1} Pa^{-1}
02	*GRIP only*	Like 01
03	Like 01, *but steepening temperature gradient in front of advancing ice*	Like 01
04	Like 01	*Precipitation factor = 1.1* *(for 25–19 ka BP)*
05	*Warmer by 1°C*	Like 01
06	*Colder by 1°C*	Like 01
07	Like 01	*Tsummer_diff = 17°C*
08	Like 01	*Sigma = 6.0*
09	Like 01, *but higher data frequencies used for GRIP (every 100 instead of 500 years)*	Like 01
10	Like 01	*Sliding parameter = 5.0×10^{-3} m yr^{-1} Pa^{-1}*

(1002–1020 km from the ice divide) for the third modeled stillstand, corresponding to the Marquette phase (see Fig. 1 and Table 1). Although we realize that the occurrences of stillstands in the model results do not necessarily equate with real moraine positions, for ease of discussion we use the names of the phases in quotes when referring to the model results.

Ice extent

The ice extent generated by the standard model run is generally in good agreement with the known and dated extents from field evidence (Fig. 2B). The duration of the first ('Johnstown') stillstand period in Experiment 01 is *c*. 5000 years. Stillstands corresponding to the extents of the Green Lake and the Marquette phase are also generated by the model. The second stillstand ('Green Lake' phase) lasted *c*. 1000–2800 years, the third stillstand ('Marquette' phase) *c*. 1100 years. Modeled ice extents representing the Chilton and the Two Rivers phases are reached at model years 14.1 and 13.7 ka BP, respectively, but do not show stillstands (Table 1). When also using the GRIP record for the time of ice build-up (Experiment 02), the maximum extent is reached *c*. 4500 years earlier and interrupted by a minor retreat; the retreating extent curve looks similar to the standard run. Experiments 03, 04 and 06 all display a slightly farther maximum ice advance (1 node or *c*. 17 km), but the maximum stillstand lasts very long, deglaciation starts too late and then occurs quite rapidly and uniformly, without further stillstands to create later moraines. This scenario seems less probable than the results from Experiments 01 or 02. In Experiment 05, the ice advance stops more than 100 km short of its target, and in Experiment 07 deglaciation starts very early and none of the later phases are represented by the modeled ice extent. Experiments 08 and 10 yield similar results as Experiment 01, except that the 'Marquette' extent is not quite reached. The use of a GRIP input curve with a higher sampling rate in Experiment 09 lets the ice advance slightly farther to the maximum extent, but during the 'Marquette' phase, the modeled ice extent lies more than 100 km south of the mapped extent.

None of the experiments produced an ice retreat and readvance between the 'Johnstown' and 'Green Lake' phases and between the 'Chilton' and 'Two Rivers' phases (representing the Two Creeks interval), as suggested by the field data. Experiments 05 and 07 clearly fail to provide reasonable ice extents through time and are therefore not considered further. Because ice extents of Experiment 01 are in best agreement with the known ice extents, the following discussion is mainly based on the results of this experiment.

Ice profile and basal shear stress

Figure 3 illustrates ice thickness and profile development of the Green Bay Lobe through time. The ice thickness developed in Experiment 01 is very large. Approximately 17 km back from the margin, the ice is about 1000 m thick during the maximum extent and increases to 1800–2000 m *c*. 140 km from the margin (near the location of Green Bay). For the 'Green Lake' stillstand phase, the thickness is slightly less (*c*. 800 m at the first node behind the margin and *c*. 1800 m 140 km up-ice). Ice surface profiles are flatter and the ice thinner for the next two phases (decreasing from *c*. 500 m close to the margin to *c*. 1000 m at 100 km up-ice). Surface slope during the 'Johnstown' phase decreases from *c*. 0.02–0.03 (corresponding to 1.1–1.7°) close to the margin to *c*. 0.002 (equivalent to 0.1°) at *c*. 100 km from the margin and remains more or less consistent for a long distance up-ice (Table 3). Slope values are similar for the 'Green Lake' stillstand phase, whereas during the following two phases the slope is less steep at the margin (*c*. 0.01, corresponding to 0.6°) and then decreases in a similar way as during the two

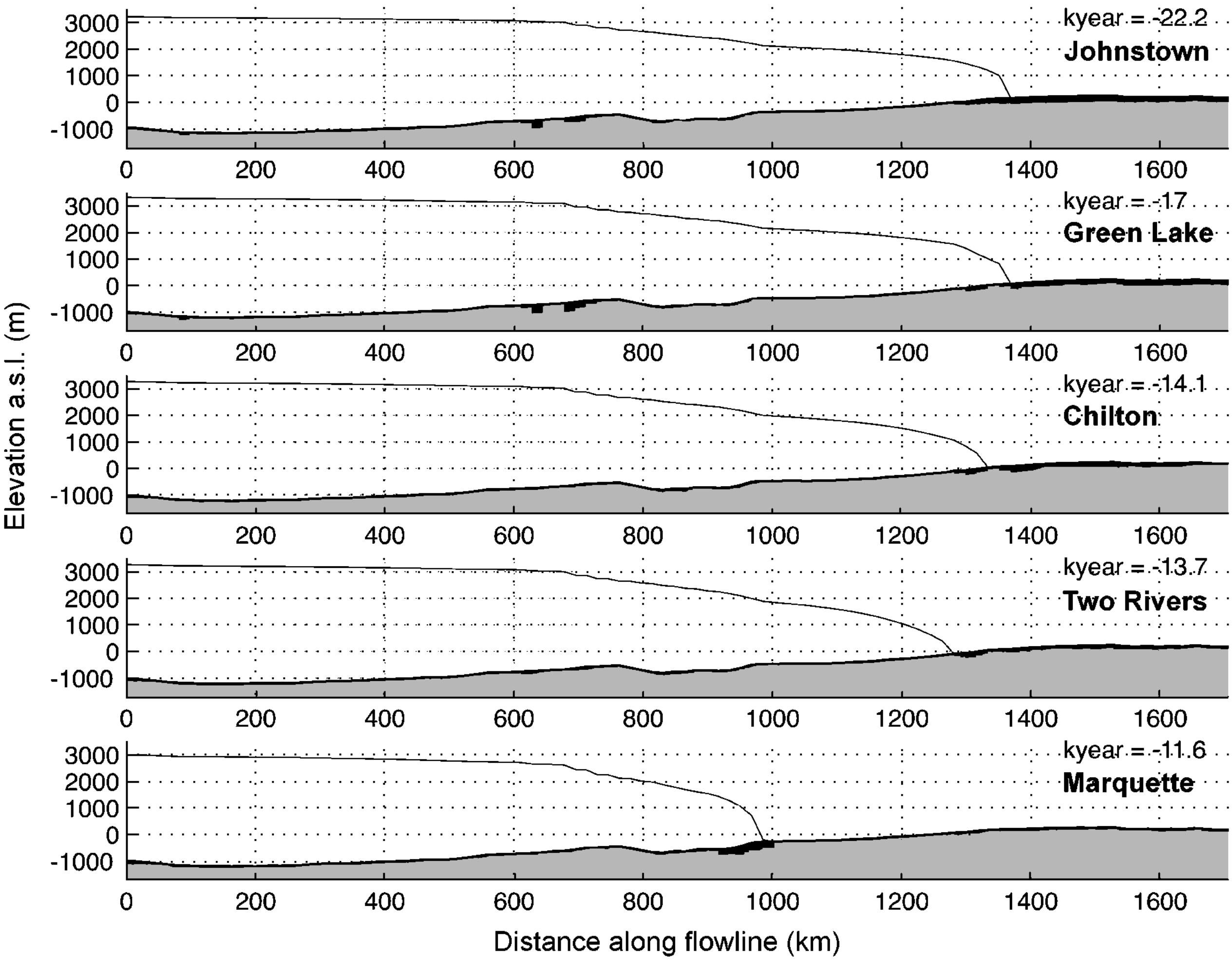


Fig. 3. Ice thickness profiles and permafrost extent (black) during maximum extent and at different times during deglaciation for Experiment 01.

earlier phases. Thickness and slope values for the 'Marquette' stillstand phase are similar to the 'Johnstown' and 'Green Lake' phases.

Basal shear stress rises to 100 kPa close to the ice margin of the maximum extent phase. It decreases to 50 kPa at *c.* 150 km behind the margin and to 30 for the next *c.* 250 km up-ice (Fig. 4 and Table 3). During the 'Green Lake' stillstand phase, basal shear stress values are very similar. For the 'Chilton' and 'Two Rivers' phases, maximum basal shear stress is similar to the two earlier phases, but it decreases more evenly behind the margin. Basal shear stress is higher for the 'Marquette' phase, probably because the ice moves up a steep slope. It decreases to *c.* 100 kPa over 200 km behind the ice margin.

Ice-flow velocity

Ice-flow velocities of our standard experiment (01) are shown in Fig. 5 and listed in Table 3. Total ice-flow velocities (combined sliding and average ice deformation velocities) up to the end of the 'Johnstown' stillstand are quite uniform in the area examined, up to *c.* 150 km up-ice. They range between 60 and 100 m/yr. Ice-flow velocities are similar for the 'Green Lake' stillstand phase (around 100–150 m/yr). For the modeled ice cover during the 'Chilton' and 'Two Rivers' phases, ice-flow velocities are *c.* 100 m/yr near the ice margin and up to 250 m/yr *c.* 70 km up-ice. A velocity peak of *c.* 150 m/yr occurs close to the ice margin of the 'Marquette' stillstand. In all experiments, deformation velocity is relatively high close to the ice margin due to the steep ice surface slope during the 'Johnstown' and 'Green Lake' stillstand phases and is more evenly distributed during the later phases.

During the 'Johnstown' stillstand, sliding (of *c.* 80 m/yr) only takes place far up-ice, *c.* 150–200 km from the ice margin. Sliding velocities are similar or slightly higher for the 'Green Lake' stillstand, but sliding is also limited to an area of at least 100 km (and at least 70 km

Table 3. Results for ice-flow velocity, sliding velocity, slope values, basal shear stress and the occurrence of permafrost, derived from standard Experiment 01, listed for the times and extents that correspond to the phases derived from field evidence.

Phases corresponding to modeled ice extents	Total ice-flow velocity (m/yr)	Sliding velocity (m/yr)	Slope value (at 5–70–100 km from margin)	Basal shear stress (kPa)	Permafrost occurrence
Johnstown	60–100 (at margin and up to 150 km up-ice)	80 (at 150–200 km up-ice)	0.02 to 0.03–0.006–0.002	100 close to margin and up to *c.* 100 km up-ice, 30–50 up to *c.* 400 km up-ice	Width *c.* 105 km, thickness 100–200 m
Green Lake	100–150 (at margin and up to 150 km up-ice)	80 (at 100 and more km up-ice)	0.02 to 0.03–0.006–0.002	100–120 close to margin and up to *c.* 100 km up-ice, 30–50 up to *c.* 400 km up-ice	Width 70–80 km, thickness 80–180 m
Chilton	100 at margin, up to 250 *c.* 70 km up-ice	180–200, peaks up to 400 (up to 70 km from the margin)	0.01–0.006–0.002	100 close to margin, decreasing to 40 at 300 km up-ice	Width *c.* 25–50 km, thickness up to 50 m
Two Rivers	100 at margin, up to 250 *c.* 70 km up-ice	180–200, peaks up to 400 (up to 35 km from the margin)	0.01–0.006–0.002	100 close to margin, decreasing to 50 at 280 km up-ice	Width *c.* 0–25 km, thickness up to 30 m
Marquette	150 at margin	Up to 400 (up to 35 km from the margin)	0.02 to 0.03–0.006–0.002	200 close to margin, 100 at 200 km up-ice	Width *c.* 80 km, thickness up to 200 m

in all other experiments) behind the ice margin. For the ‘Chilton’ and ‘Two Rivers’ phases, sliding velocities are higher (180–200 m/yr, single peaks up to 400 m/yr) and sliding occurs up to 70 and 35 km from the ice margin (similar for all experiments).

Permafrost development

During ice advance to its maximum extent, permafrost occurs *c.* 1260 km from the ice divide and southward to the end of the flowline (Fig. 3). Permafrost begins to thin when covered by thick ice (*c.* 1260 to 1350 km from the ice divide). When the ‘Johnstown’ stillstand is reached, permafrost has disappeared at *c.* 1260 km from the ice divide (*c.* 120 km from the ice margin) and discontinuous permafrost occurs around 1280 km from the ice divide (*c.* 100 km from the margin). The glacier bed at the margin and at the first node up-ice (1369–1387 km from the ice divide) remains frozen throughout the first stillstand phase to a depth of about 200 m, whereas in the zone between the deeply frozen margin and the discontinuous permafrost, maximum permafrost depth decreases throughout the ‘Johnstown’ phase from about 200 to about 100 m. During the ‘Green Lake’ stillstand phase, the bed at the margin (1369 km from the ice divide) is frozen to a depth of 200 m (and -150 at 17 km up-ice), and permafrost thickness is 80–180 m in the zone up to *c.* 85 km back from the margin.

During ice retreat, thawing of permafrost takes place over *c.* 5 ka from a thickness of 200 m to zero. Melting at the ‘Johnstown’ and ‘Green Lake’ stillstand ice margins starts at *c.* 14 ka BP and slightly later at the margins of the ‘Chilton’ and ‘Two Rivers’ phases. Shortly before and during the ‘Marquette’ stillstand phase, some discontinuous permafrost reforms around the ice margin. The width of the frozen zone below the ice remains stable at *c.* 105 km back from the margin for the duration of the ‘Johnstown’ phase, decreases to 80–70 km for the ‘Green Lake’ phase, decreases rapidly from 50 to 0 km during the ‘Chilton’ and ‘Two Rivers’ phases and increases once more to *c.* 80 km around the ‘Marquette’ phase, before permafrost disappears completely. Permafrost distribution for several times during deglaciation is shown together with the ice surface profiles in Fig. 3 and permafrost values are listed in Table 3.

Experiments with a farther ice advance during the maximum (03, 04, and 06) show almost the same permafrost distribution as Experiment 01, only shifted southwards by one node. The timing of permafrost melting also remains the same. All other experiments (including 05 and 07, which use higher input temperature) behave similarly as Experiment 01 in terms of permafrost distribution, extent and duration.

Basal melting

Figure 6 shows the evolution of the average basal meltwater production along the whole flowline through time. Significant basal meltwater production in the southern part of the flowline (*c.* 1210 km from its beginning and southwards) starts at *c.* 14.3 ka BP and reaches a peak around 13.5 ka BP. Almost no meltwater occurs here prior to this time. After reaching the maximum, meltwater production decreases rapidly and is very low during the ‘Marquette’ stillstand.

Discussion

While the locations of the maximum and subsequent extents of the southern Laurentide Ice Sheet in the Green Bay Lobe are known from moraines, the exact timing of ice advance, duration of ice stillstands, beginning of retreat and extent of readvances are controversial. Our model reproduces ice extent through time reasonably well, except that maximum ice extent is missed by the model results by *c*. 50 km. The model run with the best-fitting ice extent results (Experiment 01), on which the description and interpretation of all other results is mainly based, yields an ice advance starting at 30 ka BP, stillstand durations of 5 kyr for the ‘Johnstown’ phase (22.2–17.2 ka BP) and of 1–2.8 kyr (17–16 or 14.2 ka BP) for the ‘Green Lake’ phase, and then a progressive retreat, interrupted only by a stillstand of 1.1 kyr during the ‘Marquette’ phase. The model years of the ‘Johnstown’ and ‘Green Lake’ phases might be up to 2000 years too young and the durations slightly underestimated due to the use of the old GRIP time scale. Older modeled phases would even be in better agreement with the dated geologic record (cf. Fig. 2B). The fact that ice extents can be reproduced by using the adjusted GRIP δ^{18}O curve for paleotemperature input might reflect the feedback processes described by Clark *et al.* (2001). The authors proposed that the routing of freshwater flow from the southern margin of the Laurentide Ice Sheet influenced North Atlantic circulation, thus linking ice margin fluctuations to North Atlantic climate. Our model results agree with suggestions by Colgan (1999) of a relatively stable and progressively and slowly retreating ice margin of the Green Bay Lobe. The model fails, however, to reproduce what has been interpreted as a major ice-margin retreat between deposition of the sandy Holy Hill member of the Johnstown and Green Lake advances and deposition of the red clayey till of the

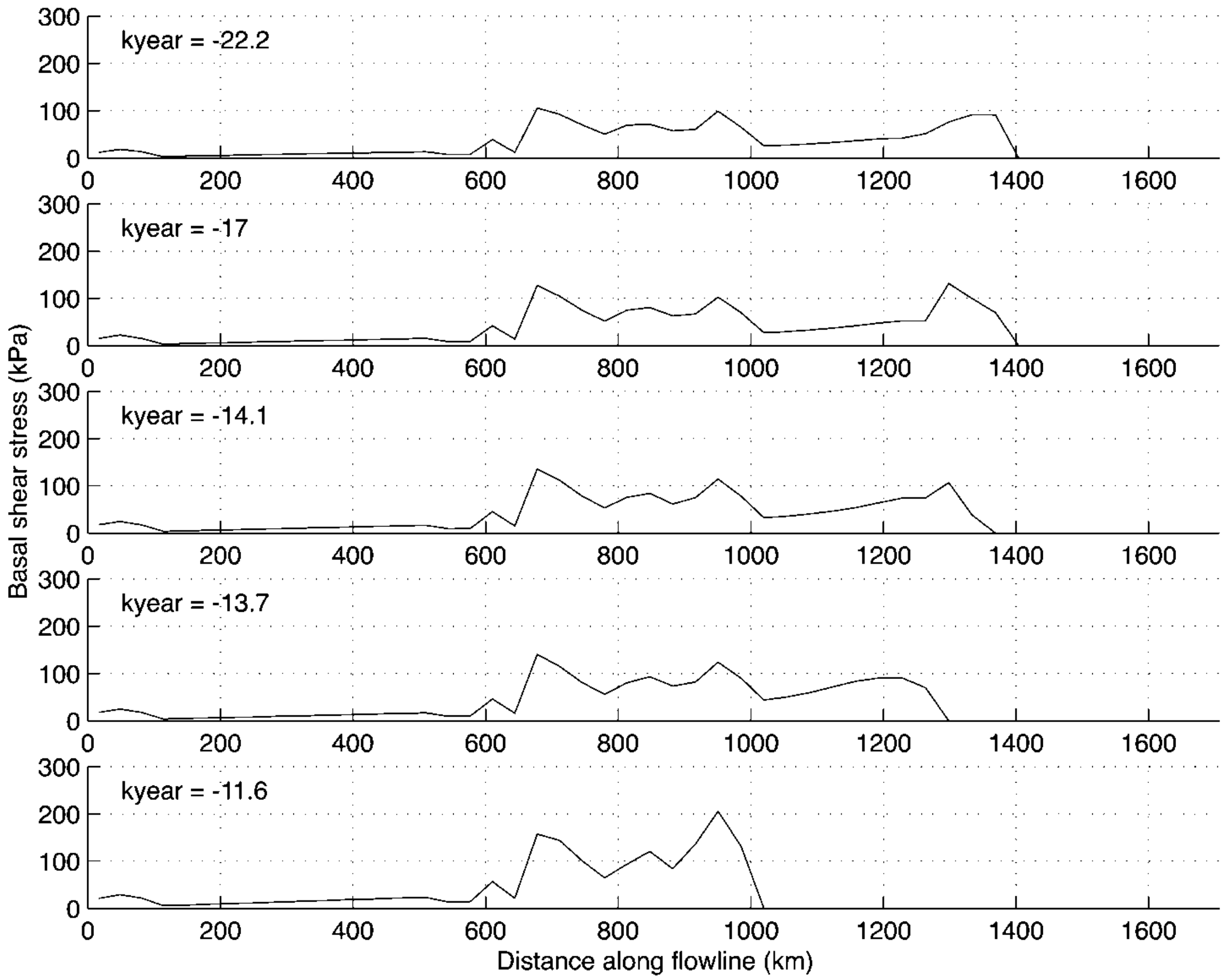


Fig. 4. Basal shear stress distribution along flowline for Experiment 01 at different times during deglaciation.

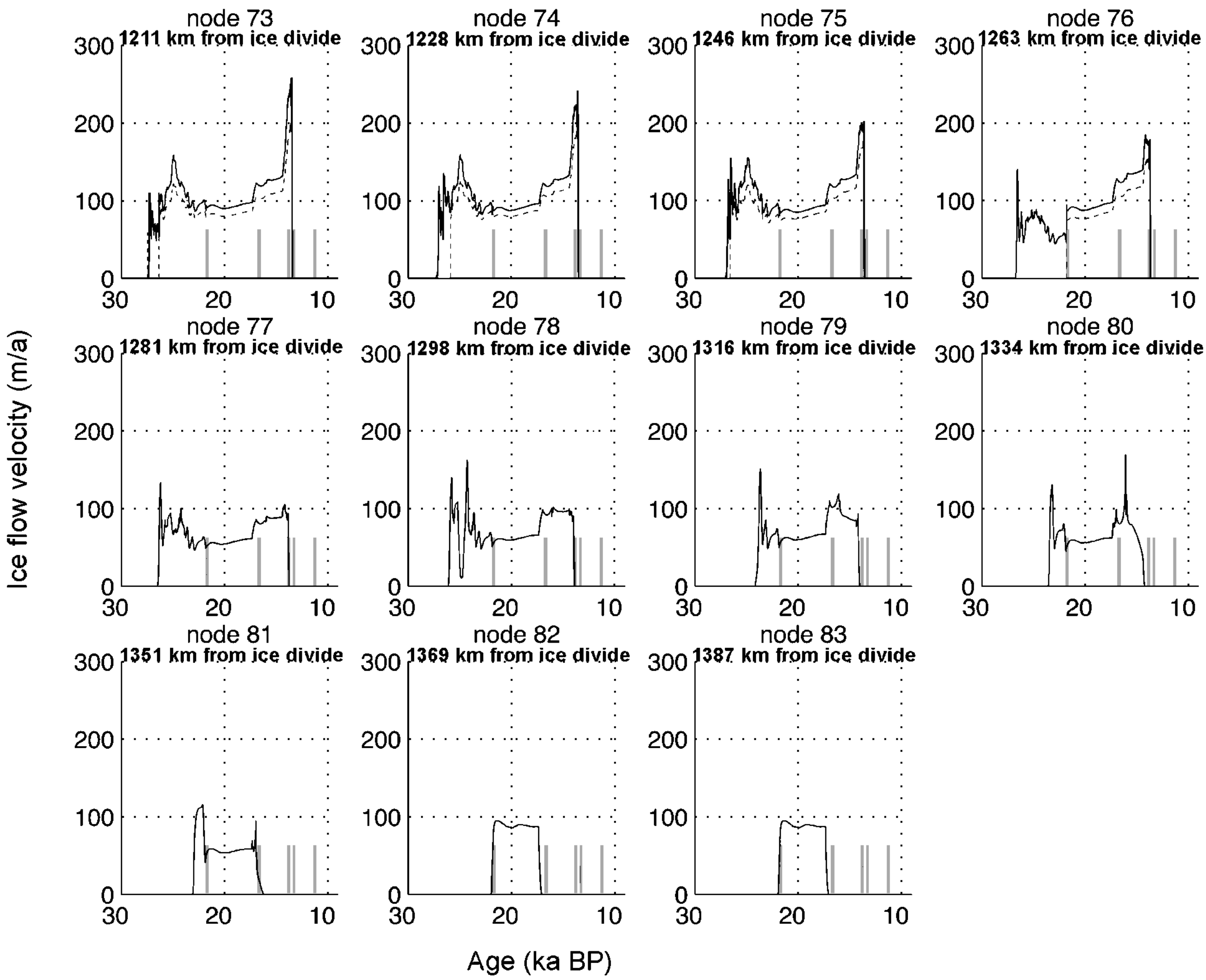


Fig. 5. Ice velocities (total solid line, sliding dashed line) for Experiment 01 at nodes 73 to 83. Marked in grey are the times of the modeled ice extents corresponding to the Johnstown, Green Lake, Chilton and Two Rivers phases.

Kewaunee Formation (McCartney & Mickelson 1982). This retreat is thought to be documented by the transport of red clay from the Lake Superior basin into Green Bay and Lake Michigan. However, this red clay could have been transported subglacially, making this argument for extensive retreat invalid. The model also fails to reproduce the margin retreat that allowed growth of the Two Creeks forest which was then drowned by rising lake level and finally covered by the readvancing ice margin (Maher & Mickelson 1996; Colgan 1999; Socha *et al.* 1999). Even when using higher frequency and higher amplitude temperature input or larger differences between mean annual and summer temperature, the model does not produce rapid oscillations in ice-margin positions. Using a GRIP-based climate forcing, Marshall & Clarke (1999) observed a good agreement of deglaciation timing and pattern of their modeled Laurentide Ice Sheet with the geologic record, but also a lack of advance/retreat sequences. One possible explanation for the differences between model results and field evidence might be that ice surging played an important role during later stages of the Green Bay Lobe. This is not incorporated in the current model, as its mechanism is poorly understood.

The maximum balance velocity of the glacier generated by the model for the two first stillstands is 150 m/yr, which lies at the lower end of the velocity range suggested by Colgan (1999). The ice-flow velocity during the unstable retreat phase is up to 100 m/yr higher. Modeled ice-flow velocities depend on the flow law employed. As ice surging is not incorporated in the model, values calculated here have to be regarded as minimum velocities. During retreat, sliding at the ice base contributes to the higher modeled ice velocities and probably becomes more important in the formation process of drumlins, whereas during the stable ice phases, sliding does not play an important role within a zone of at least 100 km behind the ice margin.

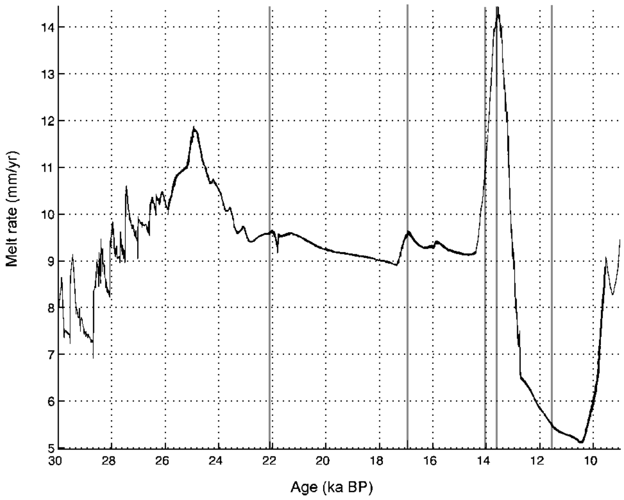


Fig. 6. Average basal meltwater production through time. The beginnings of the Johnstown, Green Lake, Chilton, Two Rivers and Marquette phases are marked.

Ice-surface profile reconstructions for the Green Bay Lobe have so far mainly been based on morphologic field evidence (Clark 1992; Colgan & Mickelson 1997; Colgan 1999; Socha *et al.* 1999). Highest moraine elevations are assumed to reflect ice surface elevations at the time of moraine formation, and lines of equal elevation are perpendicular to flowlines (that are derived from ice-flow direction indicators). These reconstructions provide information on the ice-margin zone of the Johnstown, Green Lake and Chilton phases and will be compared to our model results in that order. For the Johnstown phase, the very steep margin slope of 0.02–0.03 and the surface slope of 0.002 at *c.* 100 km from the ice margin generated by our model experiments are in good agreement with the slope value of *c.* 0.0018 within 150 km of the terminus inferred from moraine elevations by Colgan & Mickelson (1997). They deduced an ice thickness of at least 100 m within a few hundred meters of the margin and 200 to 600 m (vs. 1000 to 1800 m in the model) for the associated drumlin field during the Johnstown phase. Over the location of Green Bay, the model yields an ice thickness of *c.* 1800 m during maximum extent, whereas Colgan (1999) reconstructed a thickness of *c.* 1000 m. For the Green Lake phase, Colgan & Mickelson (1997) postulated a thickness of at least 100–150 m within 1 km of the terminus and *c.* 200 to 350 m (vs. 800 to 1800 m in the model) over the drumlin area. For the Chilton phase, Socha *et al.* (1999) obtained average surface slope values of 0.002 and an ice thickness of less than 200 m within 35 km of the terminus. These values are also lower than the ones produced by the model for that time (*c.* 0.006 and 500 m). The ice thickness and slope values deduced from field data are, however, minimum values, since they probably reflect times of ice wasting, and not times of maximum ice thickness and slope.

The driving stresses produced by the model are also two to four times higher than the ones inferred from geomorphic data by Clark (1992) and Colgan (1999) for the maximum extent of the Green Bay Lobe and an order of magnitude higher than the estimates by Colgan (1999) and Socha *et al.* (1999) for later phases of the Green Bay Lobe. Geomorphic investigations provide minimum values (see Clark 1992), but the great discrepancies between model results and field evidence during later stages of the Green Bay Lobe might be due to ice surging acting at the ice margin at this time. Because only the presence or absence of water and not the thickness of a water layer or the basal water pressure are implemented into the ice flow and sliding laws used, the model is only able to predict relatively high basal shear stresses, but not low basal shear stresses generated by ice surges. However, little basal water is predicted by the model except for the time of late deglaciation, thus probably limiting the possibility of rapid ice flow. With the presence of more water late during deglaciation, glacier flow might have been much faster. The mechanics of ice flow in the Green Bay Lobe at this time are still poorly understood.

The presence of permafrost probably played an important role in determining the ice lobe form by drastically reducing basal motion, and also in controlling the subglacial hydrology of the ice lobe. The impact of permafrost on subglacial water flow, possibly leading to blockage of the drainage system, has been investigated in groundwater modeling studies by, for example, Piotrowski (1997) and Breemer *et al.* (2002). Permafrost and a frozen bed near the margin may be among the most important factors for shaping the glacial landscape, especially during times of ice retreat. Several glacial landforms have been interpreted as being associated with cold-ice conditions: tunnel channels (e.g. Attig *et al.* 1989), drumlins (e.g. Mickelson *et al.* 1983), ice-walled lake plains (e.g. Ham & Attig 1996), and ice-thrust features (e.g. Attig *et al.* 1989; Mooers 1989). Field evidence such as ice-wedge casts, polygons and the lack of datable wood suggests that the late Wisconsin Green Bay Lobe advanced over permafrost and that permafrost had melted by *c.* 14 ^{14}C ka in southern Wisconsin and by *c.* 13 ^{14}C ka in northern Wisconsin (Attig *et al.* 1989; Clayton *et al.* 2001). The model results show permafrost in the glacier forefield and beneath the ice during advance. The permafrost penetrates deeply, is stable throughout the first two stillstands and still partly present, but less extensive, afterwards. The presence of eskers younger than the tunnel channels, occurring within 10 km of the outermost margin, argues for warming of the glacier while it was still at the outermost moraine. Hence, permafrost seems to be overestimated by the model.

From the comparison of phases associated with drumlins with the phases that do not show evidence of these features, it appears likely that the formation of drumlins is linked to an ice stillstand of at least 1000 years in combination with extensive frozen-bed conditions and subsequent thawing, ice-flow velocities of 50–150 m/yr and high basal shear stress. A longer duration of these conditions may favor more pronounced drumlins, which is suggested by the comparison of the larger Johnstown phase drumlins with the smaller Green Lake phase drumlins. Drumlins are also associated with the Marquette phase, which resembles the Johnstown and Green Lake phases with regard to modeled ice thickness, shear stress and formation of permafrost. Our findings are in agreement with Mitchell (1994), who attributed larger drumlins to areas with high basal driving stress and smaller drumlins to areas with lower basal driving stress. In the Johnstown phase drumlin field, the length-to-width ratio of drumlins decreases towards the margin. This might be related to decreasing ice-flow velocity towards the margin (as suggested, for example, by Stokes & Clark (2002) for a paleo-icestream record, where the down-ice variations in elongation ratio reflect exactly the expected velocity field). However, similar velocity patterns are simulated for all phases by the model, thus suggesting that the duration of ice stillstand and its prevailing ice-flow conditions plays a major role in enabling drumlin formation.

The experiments of Cutler *et al.* (2000) showed that the presence of permafrost probably hindered significant meltwater drainage and thus led to meltwater accumulation behind the frozen zone. When the ice margin stabilized and the frozen-zone width declined, hydro-fracturing occurred, leading to outburst events (Cutler *et al.* 2002) and generating landforms such as the tunnel channels that are observed around the LGM margin of the Green Bay Lobe (e.g. Attig *et al.* 1989). Cutler *et al.* (2002) suggested that perma frost could not build up again during retreat once it had melted. Our deglaciation experiments and field evidence indicate, however, that permafrost remained or reformed as ice retreated until *c*. 14 ka BP. The lack of tunnel channels at younger ice margins suggests that the frozen-bed zone was probably not thick enough or of long enough duration to allow large build-ups of basal water. Basal meltwater production is greatest at about 13.5 ka BP, but at this time thawing of permafrost had probably progressed enough to allow most basal meltwater to escape directly through the aquifer or channels.

Conclusions

Our numerical experiments investigate the behavior of the Green Bay Lobe during the last deglaciation and the model's sensitivity to varying climate input within a reasonable range (based on GCM results and local proxy data). Despite possible model shortcomings with regard to basal ice motion parameterization, as it influences ice surface profile, basal shear stress and ice-flow velocity results, and despite limited data on the paleoclimate during the time of maximum ice extent and deglaciation in our study area, the following conclusions seem justified:

- North Atlantic changes in climate, reflected in the GRIP δ^{18}O record on which the temperature input curve for the model was based, appear able to explain first-order behavior of the southern Laurentide Ice Sheet in the Green Bay Lobe area. Using that input, known ice extents through time (for maximum extent and early deglaciation) can be reproduced reasonably well.
- The model fails to reproduce short-term ice retreats and readvances (second-order behavior) during later stages of deglaciation, perhaps because ice surging played an important role at this time.
- In the model, permafrost is present in the Green Bay Lobe area until *c*. 14 ka BP, agreeing with independent geologic evidence. Stable ice margins leading to steep moraines and a regime of high basal shear stress occur between *c*. 22 and 17 and between *c*. 17 and 14 ka BP. The overall modeled retreat rate from the maximum position to the second stillstand is very low, around 10 m/yr. Ice retreat after the first two stillstand phases takes place at relatively steady rates of 160–230 m/yr until *c*. 9 ka BP; then it occurs quite rapidly (at a rate of *c*. 660 m/yr).
- Associated with the drumlin-producing ice extent phases are stable ice margins with relatively thick ice, steep and stable profiles and high and stable basal shear stresses for a duration of at least 1000 years, together with subglacial permafrost up to about 100 km up-ice from the margin and its subsequent thawing. As long as the frozen ground prevails, no significant basal sliding takes place and meltwater is present only in very limited amounts. During deglaciation, basal meltwater and sliding become more important at the ice base. Geologic evidence suggests that this warming took place as the ice began to retreat from the maximum extent and may have contributed significantly to the streamlining of drumlins.

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