Modeling North American Freshwater Runoff through the Last Glacial Cycle

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The Northern Hemisphere ice sheets decayed rapidly during deglacial phases of the ice-age cycle, producing meltwater fluxes that may have been of sufficient magnitude to perturb oceanic circulation. The continental record of ice-sheet history is more obscured during the growth and advance of the last great ice sheets, ca. 120,000–20,000 yr B.P., but ice cores tell of high-amplitude, millennial-scale climate fluctuations that prevailed throughout this period. These climatic excursions would have provoked significant fluctuation of ice-sheet margins and runoff variability whenever ice sheets extended to mid-latitudes, giving a complex pattern of freshwater delivery to the oceans. A model of continental surface hydrology is coupled with an ice-dynamics model simulating the last glacial cycle in North America. Meltwater discharged from ice sheets is either channeled down continental drainage pathways or stored temporarily in large systems of proglacial lakes that border the retreating ice-sheet margin. The coupled treatment provides quantitative estimates of the spatial and temporal patterns of freshwater flux to the continental margins. Results imply an intensified surface hydrological environment when ice sheets are present, despite a net decrease in precipitation during glacial periods. Diminished continental evaporation and high levels of meltwater production combine to give mid-latitude runoff values that are highly variable through the glacial cycle, but are two to three times in excess of modern river fluxes; drainage to the North Atlantic via the St. Lawrence, Hudson, and Mississippi River catchments averages 0.356 Sv for the period 60,000–10,000 yr B.P., compared to 0.122 Sv for the past 10,000 yr. High-amplitude meltwater pulses to the Gulf of Mexico, North Atlantic, and North Pacific occur throughout the glacial period, with ice-sheet geometry controlling intricate patterns of freshwater routing variability. Runoff from North America is staged in the final deglaciation, with a stepped sequence of pulses through the Mississippi, St. Lawrence, Arctic, and Hudson Strait drainages.

Key Words: Laurentide; proglacial lake; meltwater pulse; thermohaline; Dansgaard-Oeschger.

INTRODUCTION

The present-day hydrological cycle is essentially in balance over annual time scales, with evaporation over the ocean replenished by marine precipitation and continental runoff. Atmosphere–ocean models that consider the hydrological cycle can neglect continental storage of water in glaciers, lakes, or the groundwater system to good approximation, and assume that total annual rainfall over the continents is returned to the ocean via river systems. Continents can be divided into topographic catchment basins that feed the dominant drainage rivers, coarsely providing the correct geographic distribution of freshwater flux from the continents. Figure 1a, adapted from Weaver and Hughes (1996), illustrates such catchment basins for the principal drainage systems of North America. In oceanic general circulation model (OGCM) studies by Weaver and co-workers, rain that falls in a given catchment is reinjected in the ocean in a small number of OGCM cells on the continental shelf (Weaver and Hughes, 1996; Fanning and Weaver, 1997).

The relative stability of the Holocene sea level suggests that the hydrological cycle is indeed in near-balance, although for long time-scale Holocene investigations one may need to consider the small deviations in freshwater runoff (hydrological disequilibrium) introduced by fluctuations of polar ice caps and mountain glaciers. A different treatment is required for Pleistocene studies. The Northern Hemisphere ice sheets changed both the hydrological balance and the topographic disposition of the continents. Water was stored on the continents in the ice sheet and proglacial lake systems, and runoff patterns changed dramatically (e.g., Teller, 1990). The late-glacial proglacial lakes are known to have been immense, and they were likely to have had important regional climate impacts (Hostetler et al., 1994). The lakes may also have exacted a dynamical influence on ice sheets through calving of wide regions of the southern ice-sheet margins (e.g., Pollard, 1983).

We introduce a model of ice-sheet and continental surface hydrology to investigate meltwater runoff patterns during the last glacial cycle in North America. The model describes time-dependent meltwater storage and routing on the surface of the ice sheet and continent, including a simple representation of continental water storage in topographically controlled proglacial lakes. Evolution of the drainage system is a function of meltwater generation, ice-sheet surface topography, continental topography, and the pattern of ice-sheet retreat. Teller...
Licciardi et al. (1987, 1995) and Licciardi et al. (in press) have mapped the proglacial lake distribution and drainage history of North America in tremendous detail through the last deglaciation, offering a means of model evaluation.

This work is motivated by arguments that the strength of the North Atlantic thermohaline circulation has varied dramatically in the last glacial period, presumably in response to variations in the surface freshwater budget (e.g., Boyle and Keigwin, 1982, 1987; Broecker et al., 1985, 1990; Keigwin et al., 1991). Ocean model studies suggest that North Atlantic deep water (NADW) formation can be extremely sensitive to glacial meltwater runoff from the continents (Stocker and Wright, 1991; Weaver et al., 1993; Manabe and Stouffer, 1995, 1997; Rahmstorf, 1995). Severe weakening or shutdown of NADW formation is expected from large meltwater injections such as the deglacial meltwater pulses (mwp) described by Fairbanks (1989). OGCM simulations by Manabe and Stouffer (1997) and Fanning and Weaver (1997) support the notion of ocean “pre-conditioning” by ice-sheet meltwater, increasing the sensitivity of the glacial North Atlantic to freshwater pulses. The amplified high-latitude freshwater runoff which is expected in glacial periods weakens the thermohaline circulation, leaving it highly susceptible to collapse.

The collective observational, conceptual, and modeling evidence implies that the rapid, recurrent climatic fluctuations of the glacial period, as recorded in Greenland ice cores (Dansgaard et al., 1984, 1993; Oeschger et al., 1984), may be associated with freshwater-driven oscillations of the thermohaline circulation. The atmospheric and oceanic signal of large-scale ocean circulation changes can be expected to be rapid and widespread, so the climatic impact of NADW fluctuations would not be limited to the North Atlantic region (e.g., Fawcett et al., 1997; Hostetler et al., 1999). Stadial–interstadial variations in climate, dubbed Dansgaard–Oeschger (D–O) events, may therefore have their origin in ice sheet–ocean interactions, via the global hydrological cycle. This paper introduces a continental ice-sheet/water balance model suitable for paleoclimate studies, as a step toward explicit modeling of the global hydrological budget.

MODEL DESCRIPTION

We combine models of ice-sheet dynamics, mass balance, bed isostatic adjustment, and surface water transport. The physical treatments of ice thermomechanics, ice-sheet mass balance, and isostatic response are standard and are the central ingredients for models of the ice-age cycle (e.g., Huybrechts and T'Siobbel, 1995; Tarasov and Peltier, 1997; Marshall et al., in press). Our paleoclimate treatment is based on a forcing derived from the Greenland Ice Core Project (GRIP) δ¹⁸O record (Dansgaard et al., 1993), in conjunction with atmospheric general circulation model (AGCM) climate reconstructions for the present day and the last glacial maximum (LGM), from the CCC model of the Canadian Centre for Climate Modelling and Analysis (Vettoretti et al., in press). This paleoclimate forcing, hereafter denoted the CCC/GRIP climatology, is described in detail in Marshall et al. (in press) and is summarized below.

These basic components of models of the ice-age cycle also embody the essential physics needed for analysis of continental
meltwater routing and proglacial lake evolution, ice-sheet position and thickness, patterns and rates of precipitation and glacial melt, and bed topography. We add to this a simple description of ice-sheet and continental surface hydrology to account for transport and ponding of rainfall and meltwater. The theory is outlined below in a spherical coordinate system with longitude \(\lambda\), colatitude \(\theta\), and Earth radius \(R_E\). Three-dimensional fields are a function of \((\lambda, \theta, z, t)\), where \(z\) is positive upward and the \(z = 0\) datum is set to present-day mean sea level. The numerical grid for the simulations has a resolution of \(1^\circ\) longitude by \(0.5^\circ\) latitude, giving cell sizes of \(19–85\ \text{km} \times 55\ \text{km}\) for the latitude range of the ice sheets in North America.

**Ice-Sheet Dynamics**

The ice-sheet model we employ is a three-dimensional, time-dependent, finite difference model in the style of Huybrechts (1990a, 1990b), with ice dynamics based on Mahaffy (1976) and ice thermodynamics solved after Jenssen (1977). We present the ice dynamics model in detail in Marshall and Clarke (1997a). Three-dimensional velocity and temperature distributions are solved to give predictions of ice-sheet area and thickness evolution. Ice temperature is important because the effective viscosity of ice changes by a factor of 1400 over the range of temperatures found in ice sheets (0 to \(-50^\circ\text{C}\)). Thermomechanical models include this influence on ice rheology. The temperature solution requires a prescription of surface air temperature and geothermal heat flux at the ice-sheet base.

**Bed Isostatic Adjustment**

Long integrations of ice-sheet history require a treatment of bed isostatic adjustment under transient loading of the crust. Define the bed elevation \(h^b(\lambda, \theta, t)\), with an equilibrium (unloaded) bed topography \(h^b_0\). For an ice load of thickness \(H^I\), we assume a local, damped return to isostatic equilibrium (Deblonde and Peltier, 1993),

\[
\frac{\partial h^b}{\partial t} = -\left[\frac{h^b_0 - h^b}{\tau} + \frac{\rho^I H^I + \rho^W \delta H^W}{\rho^W \tau}\right].
\]  

(1)

Here, \(\rho^W\), \(\rho^I\), and \(\rho^b\) are the densities of water, ice, and the underlying bedrock, and \(\delta H^W\) accounts for changes in water-layer thickness associated with varying lake depths and with sea-level fluctuations at marine points. The adjustment time scale \(\tau\) (\(e\)-folding time) applied is 5000 yr, a relatively long response time appropriate to the Archean crust underlying the core of the Laurentide ice sheet (Tarasov and Peltier, 1997).

**CCC/GRIP Paleoclimatology**

The climate fields required by the ice-sheet model are the surface air temperature, \(T^\lambda(\lambda, \theta, t)\), and precipitation rate, \(P(\lambda, \theta, t)\). We construct paleoclimatologies based on perturbations to present-day monthly mean values of \(T^\lambda\) and \(P\). Perturbation fields are derived from AGCM output and the paleoclimatic history archived in the GRIP ice core from Summit, central Greenland. We treat the pattern of variability in the GRIP \(\delta^{18}\text{O}\) record (Dansgaard et al., 1993) as diagnostic of the Northern Hemisphere climate state (Marshall et al., in press). A “glacial index” is derived from the \(\delta^{18}\text{O}\) record by assigning climate severities of \(I = 0\) to present day (1950) and \(I = 1\) to LGM \(\delta^{18}\text{O}\) values in the ice core. Glacial index values for all other times in the ice core, \(I(t)\), are linearly interpolated from these reference values. This is essentially a measure of the “glacialness” of past climate.

The magnitude and sign of paleoclimate perturbations are adopted from the present day and LGM simulations of Vettoretti et al. (in press), using v2.0 of the CCC AGCM. In order to compare available moisture in modern and glacial climates, we consider precipitation minus evaporation fields, \(P - E\). The LGM simulations were performed with orbital and ice-sheet configurations corresponding to 21,000 yr B.P., with ice sheets from the ICE-4G reconstruction (Peltier, 1994). We bilinearly interpolate the climate fields onto the \(1^\circ\) by \(0.5^\circ\) model grid. Temperature-field perturbations through a glacial cycle are approximated from the glacial index value and a linear combination of the end-member CCC present-day and LGM model predictions,

\[
\Delta T(\lambda, \theta, t) = I(t)[T_{\text{ccc}}(\lambda, \theta, 21) - T_{\text{ccc}}(\lambda, \theta, 0)].
\]  

(2)

Air temperatures over the continent as a function of position and time are then calculated from the observational climatology, the perturbation field, and an atmospheric lapse rate, \(\beta = 0.0075^\circ\text{C m}^{-1}\),

\[
T^\lambda(\lambda, \theta, t) = T_{\text{obs}}(\lambda, \theta) + \Delta T(\lambda, \theta, t) - \beta[h^I(\lambda, \theta, t) - h^I(\lambda, \theta, 0)],
\]  

(3)

where \(h^I(\lambda, \theta, t)\) is the surface elevation, \(h^I = h^b + H^I\).

Paleo-precipitation perturbations are calculated from a scaling based on the harmonic combination of LGM and present-day CCC fields, designated \((P - E)_h(\lambda, \theta, t)\),

\[
P(\lambda, \theta, t) = \left[\frac{(P - E)_h(\lambda, \theta, t)}{(P - E)_{\text{ccc}}(\lambda, \theta, 0)}\right] P_{\text{obs}}(\lambda, \theta)
\]  

(4)

We adopt this scaling relationship because there is considerable uncertainty in the magnitude of simulated AGCM precipitation, even in modern control runs. Fractionally scaling the
observed precipitation fields makes the perturbation less sensitive to AGCM magnitudes and more sensitive to the glacial/interglacial contrast. Harmonic averaging was selected from similar reasoning, as a linear combination of CCC \( P - E \) magnitudes gives excessive weight to high (wet) modern values on occasions when the glacial aridity gives \( P - E \) values close to 0. Note that the CCC simulations contain the ice-sheet surface orography, with its resultant effects on precipitation patterns. The altitude-desert effect which occurs in the high, interior regions of ice sheets is therefore embedded in the precipitation perturbation in (4).

This temperature and precipitation scaling gives climate conditions that are generally intermediate between the LGM and modern states, although intervals of the last interglacial and early Holocene periods are warmer than present \((I < 0)\). Climate forcing is updated every 100 yr in the model, introducing significant high-frequency variability as evident in the GRIP core.

We emphasize that the glacial index forcing does not necessarily give a climate representation that is internally consistent with ice-sheet topography and albedo fields at a particular time. It is a forcing, not a dynamical climate system model. The GRIP-based chronology essentially drives the ice sheet toward the LGM configuration during cold climates, while climates similar to present day drive the ice sheet toward its deglaciated Holocene state. Because the CCC LGM simulation has the ICE-4G ice sheets built in, with the distribution of ice based on the geological reconstructions of Dyke and Prest (1987a, 1987b), simulation of reasonable LGM ice-sheet coverage is “predestined” to some degree in our simulations. The glacial-index approach is therefore very different from a coupled ice-sheet/climate model, which permits internal feedbacks. The important advantage of the CCC/GRIP methodology for our purposes is the representation of high-frequency and high-amplitude climate variability. The GRIP record of climate oscillations on 100–10,000 yr time scales is much larger than the variability expressed in climate models. This has important implications for meltwater runoff fluctuations.

Ice-Sheet Mass Balance

Temperatures and precipitation rates must be translated to an estimate of annual mass balance,

\[
b(\lambda, \theta, t) = \dot{a}(\lambda, \theta, t) - \dot{m}(\lambda, \theta, t), \tag{5}
\]

where \( \dot{a} \) and \( \dot{m} \) are ice-equivalent annual accumulation and melt rates. The ablation budget, \( \dot{m} \), includes ice loss from surface melt, basal melt, and iceberg calving into proglacial lakes or the ocean. We calculate surface melt and the fraction of precipitation to fall as snow (\( \dot{a} \)) from the degree-day method (Braithwaite, 1984, 1995; Reeh, 1991). Air temperatures are assumed to be normally distributed about the monthly mean values, with standard deviation \( \sigma = 5^\circ C \). Iceberg calving on marine margins is calculated as a function of water depth, ice thickness, and ice stiffness (temperature), as described in Marshall et al. (in press).

Surface Hydrology

We introduce a simplistic surface hydrology model to calculate water redistribution and storage on the continent. Consider a water layer of thickness \( H^w(\lambda, \theta, t) \) on the surface of the ice sheet and continent. This layer can of course vanish, where there is no standing water. Locations where \( H^w > 0 \) correspond to supraglacial and proglacial (continental) lakes. The elevation (head) of the water sheet or lake surface is \( h^w = h^1 + H^w \). For incompressible fluid, the transport equation which governs the surface water balance is

\[
\frac{\partial H^w}{\partial t} = \nabla \cdot \mathbf{q}^w + \phi^w, \tag{6}
\]

where \( \mathbf{q}^w \) is the horizontal water flux vector and \( \phi^w(\lambda, \theta, t) \) is a source/sink term that includes rainfall, surface melt, basal water outflow from the ice sheet, and losses from the ice-sheet surface to the subglacial water system, via drainage into moulins and crevasses. We do not account for groundwater drainage in this version of the model. Other than differences in the source term, Eq. (6) applies equally for the ice-sheet surface and for subaerial regions.

We have experimented with different formulations of the water flux, including equations representative of open-channel laminar flow and nonlinear diffusive flow. For the coarse physics that we wish to capture here though—downslope drainage without regard for the form or rate of flow—simple linear diffusion of surface water has proven to be the most expedient means of expediting water transport. Hence

\[
\mathbf{q}^w = -\kappa \nabla h^w, \tag{7}
\]

where \( \kappa \) is the hydraulic diffusivity, and the water balance follows,

\[
\frac{\partial H^w}{\partial t} = -\kappa \nabla^2 h^w + \phi^w. \tag{8}
\]

Surface water is diffusively redistributed; water essentially flows downhill, filling up isostatic bowls, and then overflowing to run off at the continental margins. This is the simplest possible treatment of what Nature accomplishes in complicated fashion, through intricate river networks. We intend it only to describe the first-order influences of changing continental topography and meltwater supply on freshwater routing.

Numerical Model

We apply finite difference discretizations on a spherical North American grid, with grid cells 1° in longitude by 0.5° in
latitude. Ice dynamics, thermal evolution, and surface hydrology are solved asynchronously, with model time steps of 4–20 yr for the implicit dynamical solution, 10–50 yr for thermodynamic updates, and 100 yr for the surface hydrology solution. Actual time steps for modeling surface water flow are 1800 s, although this depends on the arbitrary choice of \( \kappa \). We essentially allow water to “pile up” over 100 yr, from the past century of rainfall and meltwater, then invoke the runoff model and run it out with 1800 s time steps until the water from the past century has completely run off or come to rest in topographic basins. This is handled by monitoring river fluxes and integrating Eq. (8) forward until there is no further flow. The surface hydrology acts “instantaneously” as far as the ice dynamics model is concerned. This is a simple but effective way to approximate the prevailing runoff patterns of the previous century, and it is valid as long as ice geometry and precipitation patterns have not changed significantly in that period. We found negligible difference with experiments in which the climate and runoff models were invoked every 50 yr.

**Input data sets.** Bed topography is derived from the TerrainBase DEM (Row and Hastings, 1994). We interpolate mean monthly air temperatures from 14-yr averages (1982–1995) of the NOAA NCEP-NCAR CDAS-1 2-m temperature reanalysis data set (Kalnay et al., 1996). Precipitation rates are from Legates and Willmott (1990), and geothermal heat maps for North America are digitized from the maps of Blackwell and Steele (1992).

**MODEL EXPERIMENTS**

Ice-sheet physics and the mass-balance parameterization are not tuned specifically for these tests; rather, free parameters are chosen in accord with calibrations to the Greenland ice sheet (Ritz et al., 1997). This contributes to a modeled Laurentide ice sheet that has a number of discrepancies with the geological record, although no simple tuning would be likely to alleviate this problem (Marshall et al., in press). We consider the paleoclimatic reconstruction to be the principal uncertainty in the simulation, although all components of the physical system are necessarily simplified. In particular, we approximate the ice sheet to be well coupled with the bed in all simulations presented here, with no basal flow (surge lobes or ice streams). This condition is unrealistic and leads to important weaknesses in the ice-sheet reconstruction (Clark et al., 1996a). However, the time-dependent controls of surge behavior and basal flows remain poorly understood and are yet to be satisfactorily quantified for modeling purposes. Surge-lobe physics represent an important area for future development and refinement of the simulations presented here, and we return to this point below.

**Glacial Cycle Simulation**

Time series of North American ice area and volume history through a 122,000-yr integration of the last glacial cycle are presented in Figure 2. The GRIP \( \delta^{18}O \) forcing is evident in the temperature curve of Figure 2a, which plots average air temperature over the model grid. This forcing gives considerable high-frequency response in ice-sheet area (Fig. 2d), but this is very much smoothed in the volume evolution (Fig. 2b). Volumetrically, the ice sheet integrates and essentially filters out the high-frequency variability in the GRIP record. One is left with an ice-volume history that contains high variability on the Milankovitch time scales of 23,000 and 41,000 yr. North American deglaciation is almost complete by 8000 yr B.P., with residual ice on Ellesmere Island and in the St. Elias Mountains.

The high degree of ice-sheet variability through the glacial buildup is an interesting feature of the simulation. Ice-covered area varies in step with climate forcing oscillations, with several advance–retreat sequences in the Keewatin and Labrador–Quebec sectors during ice-sheet buildup. The high variability is a result of climate fluctuations that bracket the stability conditions for mid-latitude ice sheets. Figure 3 plots mean mass balance over the ice sheet through the glacial cycle, with a value of zero indicating ice-sheet equilibrium. The Cordilleran and Laurentide complexes never reach an equilibrium, although there are approaches at the two periods of high ice volume, 60,000 and 17,000 yr B.P.

The well-known asymmetry of ice-sheet growth and retreat is manifest in this plot, as slow ice-volume expansion (small positive net balance) is interrupted by high ablation rates and rapid ice-sheet meltback at numerous intervals in the glacial cycle. Snapshots of LGM ice-sheet thickness and surface topography are plotted in Figure 4. Proglacial and supraglacial lakes are superimposed on the contours in Figure 4a and are discussed below.

The reasonableness of the modeled glacial cycle is unknown. The ice sheet is almost certainly too thick at LGM, with maximum North American ice volume about 35% in excess of estimates based on the ICE-4G geophysical reconstruction of Peltier (1994). This is a well-known problem in Laurentide ice-sheet reconstructions (e.g., Huybrechts and T'Siobbel, 1995; Tarasov and Peltier, 1999); conventional ice-sheet rheology, which works well for modeling of the Greenland and Antarctic ice sheets, invariably leads to overthickened LGM ice sheets in North America (Hughes, 1998; Marshall, 1998). Simulated LGM ice volume for North America (Fig. 4a) is \( 35.6 \times 10^{15} \) m\(^3\), equivalent to 87 m of global sea level. Seventeen percent of this ice is contained in the Cordilleran ice mass, with the bulk divided between the Laurentide and Arctic ice domes. The lack of basal flows contributes to our excess ice; subglacial sliding and sediment deformation were likely to have been important throughout much of the glaciation (e.g., Boulton et al., 1985; Fisher et al., 1985; Alley, 1991; MacAyeal, 1993) and can lead to significantly more gaunt ice sheets, as manifest on the Siple Coast of contemporary West Antarctica. Taking widespread subglacial sediment deformation into consideration in steady-state Laurentide flow-
line reconstructions, Clark and co-workers have modeled a greatly thinned, multidomed LGM ice sheet, comparable to the ICE-4G reconstruction (Jenson et al., 1996; Clark et al., 1996a).

With reference to the geological reconstructions of Dyke and Prest (1987a, 1987b), many details of modeled margin positions during deglaciation are also poor, although the overall timing and patterns of ice retreat are good. Part of the difficulty with matching the geological record can be attributed to the lack of modeled surge lobes on the southern margin. Laurentide ice-sheet deglaciation was punctuated by dramatic advance/retreat sequences in a number of locations on the southern margin (e.g., Mickelson et al., 1983; Clayton et al., 1985; Clark, 1994), which had important ramifications for meltwater routing (Teller, 1990; Clark et al., 1996b; Licciardi et al., in press). Inaccuracies in the glacial cycle simulation impose important caveats on interpretation of the modeled meltwater routing. We are nevertheless confident that the patterns of runoff evolution described below may be representative of those that prevailed in the glacial period.

**Proglacial Lakes**

Proglacial lake evolution is a complex and fascinating aspect of the meltwater runoff history. Proglacial lakes are present through much of the glacial period in our simulation, occupying a shallow, narrow band around the southern margin during ice advance phases (e.g., Fig. 4a), and filling a much larger area following intervals of retreat.

Figure 5 plots ice surface contours at 60,000, 15,000, 13,000, and 9000 yr B.P. in the model simulation, with pro-
FIG. 4. Last glacial maximum (20,000 yr B.P.) ice sheet configuration. (a) Ice sheet thickness, proglacial lake, and supraglacial lake distribution. (b) Surface topography.

FIG. 5. Time slices of ice sheet surface topography and lake history during the glacial cycle: (a) 60,000 yr B.P., (b) 15,000 yr B.P., (c) 13,000 yr B.P., (d) 9000 yr B.P.
glacial and supraglacial lakes superimposed. The Hudson Bay
lowlands host a vast lake system following major ice retreats
ca. 84,000–80,000 yr B.P. and 14,000–10,000 yr B.P. These
lakes occupy an area generally conformable with glacial lakes
Agassiz, Barlow, and Ojibway, although we do not resolve
them as distinct water bodies during early stages of ice retreat/
lake formation.

During the final deglaciation, proglacial lakes form a moat
along the southern and western edges of the Laurentide ice
sheet, occupying basins that are topographically depressed
from ice occupation ca. 30,000–17,000 yr B.P. The moat
expands northward as the ice sheet retreats, with lake depths
approaching 700 m in the central “Lake Agassiz” basin. The
lake system drains and becomes more intricate as isostatic
recovery proceeds and runoff routes open to the Atlantic and
Arctic. Areal distribution of lakes during these late deglacial
periods (Fig. 5d) resembles an overfilling of present-day basins
that ring the Canadian Shield.

Shallow lakes are commonly evident on the ice-sheet sur-
facing. These supraglacial lakes can be areally extensive and are
typically ephemeral features that form when the ice sheet is in
a state of transition. Diffusive relaxation of ice surface topog-
raphy causes changes in margin and divide position can take several
centuries, giving rise to topographic basins on the ice surface.
The saddle dividing the Laurentide and Cordilleran ice sheets
commonly ponds a great deal of water in our simulations.
Hydrologically and energetically, it is questionable whether
these lakes would endure in Nature. Seasonal surface lakes that
have been observed in the accumulation area of the Greenland
ice sheet typically refreeze to form superimposed surface ice,
or drain into the englacial or subglacial water system. Supra-
glacial lakes in Figure 5 form high in the accumulation area of
the ice sheet and would be expected to refreeze and be buried
by winter snowfall before developing to the dimensions that we
model. It is also important to point out that we have not
modeled fast basal flow in these tests; hydrologically lubricated
basal flows, as well as longitudinal stresses in the ice, would
result in more rapid surface relaxations, suppressing the extent
of supraglacial lake basins.

Teller (1987, 1990) has mapped out proglacial lake history
on the southern Laurentide margin through the deglaciation,
providing a detailed observational record to judge the model
by. Like the ice-sheet reconstruction, the general pattern of
proglacial lake distribution is reasonable, although details at
any particular point and time are unrealistic. Good present-day
reconstruction of features like the Great Lakes and the Hudson
Bay shoreline gives some assurance that our simple treatment
is acceptable. Indeed, Figure 5c reasonably approximates the
distribution of proglacial lake sediments deposited in North
America during the last deglaciation (Teller, 1987, Fig. 2). Not
all of this area was likely to have hosted proglacial lakes
simultaneously; however; nor was it one connected water mass.

Figure 6a plots time series of simulated lake volume through
the glacial cycle, expressed in meters of sea-level equivalent
(msl; 1 msl = 3.62 \times 10^{14} \text{m}^3). Millennial-scale variability is
evident in the lake volume, largely associated with stadial/
interstadial ice-sheet area fluctuations, while 10,000–100,000
yr trends reflect the long-term ice volume evolution. The
timing and rate of change of modeled lake volume are consis-
tent with the geological record (Teller, 1987, Fig. 22).

Figure 6b compares the modeled pattern of continental water
storage to the geological reconstruction of Teller (1987) for the
interval 17,000–9000 yr B.P. The age scale of ice-marginal
lake volume from Teller has been converted from $^{14}$C yr B.P.
to model time using CALIB 4.0 (Stuiver et al., 1998). The
period of extensive lake volume has a duration of roughly 2000
yr in both the model and Teller’s reconstruction, but the
maximum modeled water volume occurs at 10,200 yr B.P.,
lagging the reconstructed maximum by about 800 yr.

This comparison is not direct or simply interpretable, as we
have tabulated total lake volume on the continent, including
supraglacial lakes and basins that do not border the ice margin.
The geological reconstruction is restricted to ice-marginal lake
volume. Proglacial lakes represent the main contribution to
time-varying lake storage in the model, however, so the pat-
terns illustrated in Fig. 6b largely reflect the expansion and
drainage of ice-marginal water bodies.

Present-day North America contains about 1 msl in the
model; some water gained during the glacial cycle never finds
its way back to the ocean. We begin the integration with no
water on the continents, so this “residual” volume is largely
contained in the Great Lakes and other contemporary basins.
Relative to the reconstructions of Teller (1987), the modeled
glacial/interglacial lake volume contrast indicates an excess of
water impounded on the continents. This can be attributed to
the coarse resolution in our study and the unrealistic lake
geometry and geography which results. Any grid cell with
ponded water is considered to be entirely occupied by this
water layer, of uniform thickness $H^N$. Lake volume is simply
calculated from this water layer thickness (often several 100 m)
times the cell area, of order 2500 km$^2$.

The hypsometry of genuine lake beds would invariably lead
to lesser water volumes. Similarly, our drainage model is too
simple to veraciously simulate subgrid drainage paths (river
networks) which prevail on the continent. At ~50-km resolu-
tion, we do not expect to portray exact shoreline positions and
transgressions; isolation of marginal lakes in the reconstruc-
tions of Teller (1987) is controlled by topographical features,
theice margin structure (e.g., surge lobes), and drainage routes
that we do not capture. In recent AGCM simulations of modern
continental surface hydrology, Coe (1998) concludes that even
10-km resolution is too coarse to resolve continental lake
basins properly. Coe finds a similar result of surplus ponded
water.

These simplifications in our proglacial lake model chart a
course for further development. Subgrid topographic treat-
ments and the high-resolution digital elevation models avail-
bable today should facilitate refined modeling of lake geometry
and drainage. In this first attempt at explicit modeling of proglacial lakes, however, we believe that we describe many of the qualitative effects of lake evolution on the freshwater runoff history from the continent. We focus the remainder of this paper on details of the modeled runoff patterns, which are surprisingly rich.

**Freshwater Runoff Patterns**

The principal output of the surface hydrology scheme is the prediction of spatial and temporal patterns of freshwater runoff through the glacial cycle. Figure 7 plots time evolution of hydrological budget terms for the glacial cycle simulation pictured above. Runoff includes contributions from all river basins, and rainfall is the total wet precipitation (that which does not fall as snow). Annual rainfall and glacial meltwater are areally integrated over continental regions in the model grid to give a measure of this total source term in Sv (10^6 m^3 s^-1). This is a convenient unit for constructing the hydrological balance, as river runoff is conventionally quantified in Sv or m^3 s^-1.

In model simulations, runoff is essentially in balance with the sum of rainfall and meltwater, with water storage and release in continental lake systems representing a minor term. (Note that we have not included a description of catastrophic lake-drainage events.) Meltwater pulses in our simulation correspond to interstadials and reflect the shifts in ice area evident in Figure 2d. Runoff is much more stepped than the climate forcing, however. This is because the early stages of a warm interval result in a rapid attack of vulnerable (southward-extending) ice. The ice sheet that survives this initial assault is more stable, and the warm climate may persist but meltwater generation is more modest.

Total river runoff can be broken down into a small number of major river systems, although the catchment basins drawn in Figure 1a are of course no longer germane in the Pleistocene. We calculate drainage in individual river systems by summing the net efflux off the continent in model cells rimming the continental margin. This is essentially equivalent to setting numerical gauging stations on the entire continental margin. Drainage in the model is diffuse rather than channelized, following Eq. (7), so the array of model cells through which a particular “river system” drains is broad. Our classifications are plotted in Figure 1b. The Arctic basin in this classification differs due to the complex island geography; rather than monitoring coastal cells, we keep track of all runoff to reach the ocean in the area blocked off in Figure 1b.

Figure 8a graphs runoff in the principal drainages, the Mississippi River, St. Lawrence River, and Hudson Strait. Runoff in secondary drainage basins is plotted in Figure 8b. There is a great deal of interesting structure in the results. Large melt episodes (the warmest interstades) are seen in all river catchments but the Columbia, which is rarely fed by glacial meltwater in our simulation. Weak interstades or those that occur during long, cold spells are not seen or are only weakly manifest in the northern drainages. From 30,000–10,000 yr B.P., for instance, there is no freshwater runoff in the Arctic catchment and low variability in the Mackenzie and Baffin Bay systems, indicative of very little meltwater generation and immunity from the interstadials of this period.

An intricate pattern of asynchronous local drainage-system variability is superimposed on the widely felt melt episodes. Structure largely reflects ice-sheet geometry, with margin fluctuations creating a complex sequence of river diversions. Drainage through Hudson Strait is heightened early in the glaciation, a result of runoff from the Canadian Arctic Archipelago and Labrador. Beginning about 96,000 yr B.P., efflux from Hudson Strait weakens for several thousand years, shutting down at 87,000 yr B.P. This reflects growing ice cover over Hudson Bay and blocking of the Strait. Hudson Strait remains plugged until about 8000 yr B.P. in this simulation. On
the southern margins, drainage via the Mississippi and St. Lawrence systems is commonly out of phase, a result of closing and opening of the Great Lakes/St. Lawrence drainage route by southward advance of ice (Clark et al., 1996b; Licciardi et al., in press).

Meltwater pulses to all ocean basins are the norm, not the exception, during the last glacial period. Runoff patterns and rates are only stable during the Holocene, although southern runoff basins have near-Holocene conditions through most of marine isotope stage 5. Hudson Strait and Arctic basins exhibit significant variability throughout isotope substages 5e–5b, as ice waxes and wanes over the Canadian Arctic islands, northern Quebec–Labrador, and the Keewatin highlands west of Hudson Bay. The onset of earnest southern expansion of ice, in marine isotope substage 5a, is accompanied by increased runoff to the Mississippi, North Atlantic, and North Pacific basins. Substage 5a and stage 4 are characterized by very intense meltwater events. These represent the largest pulses of the last glacial period, with peak freshwater fluxes in excess of the major deglacial runoff events in the Mississippi and St. Lawrence.

Figure 9 focuses on the final 40,000 yr of the model simulation, illustrating the antiphasing of runoff in the Mississippi and St. Lawrence systems. This image also depicts the staged response in peak runoff to the various ocean basins during deglaciation. From 40,000–20,000 yr B.P. there is a progressive reduction in St. Lawrence runoff, interrupted by quasi-periodic interstadial drainage pulses. Increases in St. Lawrence runoff are almost always at the expense of Mississippian outflow. Freshwater flux to the North Atlantic is at its lowest level at LGM, with most of the interior of the continent draining to the Gulf of Mexico. When deglaciation begins in earnest ca. 14,000 yr B.P., the Laurentide complex is still large and susceptible to climatic attack. The result is a profusive meltwater pulse (mwp), coincident with the well-known “mwp-IA” in the Barbados sea-level record (Fairbanks, 1989). Peak modeled discharge to all ocean basins is almost 1 Sv during this period, with an average runoff of 0.675 Sv for the 1000-yr interval centered on 14,500 yr B.P. This discharge is divided between all river basins save those in the northeastern Arctic and Hudson Strait. The Gulf of Mexico and North Atlantic receive the bulk of this runoff, an average flux of 0.492 Sv for the 1000-yr interval centered on 14,500 yr B.P. Substantial meltwater (0.126 Sv) is also delivered to the North Pacific from the Cordilleran Ice Sheet at this time.
Runoff to the Mississippi plummets to Holocene values shortly after mwp-IA, remaining at this level except for a modest resumption of southward-routed drainage associated with the Younger Dryas ice readvance, ca. 13,000–11,500 yr B.P. The North Atlantic continues to receive very high runoff (0.2–0.4 Sv) in the interval 14,000–10,000 yr B.P. Peak deglacial St. Lawrence discharge occurs from 11,000–10,000 yr B.P., a runoff maximum corresponding to meltwater pulse IB (mwp-IB; Fairbanks, 1989). This drainage episode is less of a pulse in our model than the first interval of enhanced runoff, with high freshwater fluxes for several centuries of intense ice-sheet meltback. Total runoff at this time is 0.667 Sv for the 1000-yr interval centered on 10,500 yr B.P. Freshwater injections are evident in all basins in this period, although runoff to the Gulf of Mexico and North Atlantic once again dominates, averaging 0.484 Sv for the 1000-yr interval centered on 10,500 yr B.P. Continued ice-sheet meltback and isostatic recovery give progressive diversion of drainage routes to the Arctic and Hudson Strait outlets. Arctic and Mackenzie River basins are activated just after 10,000 yr B.P., pirating much of the St. Lawrence outflow. Peak 1000-yr average discharge to the Arctic and Mackenzie systems is 0.156 Sv, in the interval centered on 9500 yr B.P. Arctic runoff in turn gives way to freshwater export via Hudson Strait at 7600 yr B.P., immediately following opening of the Strait.

To summarize, the sequence of deglacial meltwater staging is: (1) pulses to the Gulf of Mexico, North Atlantic, and North Pacific, ca. 14,300 yr B.P.; (2) dominantly St. Lawrence drainage from 14,000–10,000 yr B.P., with intermittent periods of water piracy by the Mississippi catchment; (3) a brief pulse to the Mackenzie River, ca. 9900 yr B.P.; (4) a shift to high rates of drainage in the Arctic Archipelago, ca. 9700–7700 yr B.P.; and (5) opening of Hudson Strait drainage at 7600 yr B.P. Shifts from Mississippian to St. Lawrence drainage are oversimplified here due to the generally monotonic meltback in these tests (i.e., no surge lobes on the southern margin). Otherwise this stepped sequence of freshwater diversion is broadly in keeping with the geological record (Teller and Kehew, 1994).

The major meltwater events that coincide with mwp-IA and mwp-IB in the sea level record are a result of two important conditions: (1) large and southward-extending “vulnerable” ice sheets and (2) intense warm intervals as recorded in the GRIP ice-core record (the Bølling warming at ~14,500 yr B.P., and the abrupt warming that follows the Younger Dryas). These climatic shifts are dialed into our simulation via the glacial-index forcing. The response to these well-established warming events is suggestive of plausible Laurentide ice-sheet contributions to rapid sea-level rise during the deglaciation. As discussed in Clark et al. (1996b), however, there is no clear evidence that Fairbanks’ meltwater pulses come from the Laurentide. Furthermore, Bard et al. (1997) question even the existence of a second meltwater pulse in the marine record.

**RANGE OF MODEL PREDICTIONS**

The general sequence of runoff patterns is encouraging, given our simplistic representation of the physical system. The modeled ice sheet is incorrect in a number of respects, containing excess ice and regional discrepancies in the deglacial history. The lack of basal-flow (surge lobe) physics precludes portrayal of the southern margin advance–retreat history, which is known to have been important to Mississippi/St. Lawrence runoff switches. The climate forcing and hydrolog-
ical drainage models are also simple, and AGCM precipitation fields are difficult to validate for paleoclimates. Given these compounding uncertainties and approximations in the various model ingredients, it is important to assess what aspects of the results are significant.

To gain a feeling for features of the model which are robust, we have experimented with several different climate and ice sheet configurations. Air temperature and precipitation fields from the NCAR CCM1 AGCM (Kutzbach et al., 1998) were used in several studies, in perturbative mode as described for the CCC experiments. The ice sheet resulting from the CCM1/GRIP climatology has regional differences from that of the CCC/GRIP climatology, giving subtle variations in the meltwater history. However, both CCM1 and CCC ice sheets contain excess volume at LGM, relative to the ICE-4G reconstruction of Peltier (1994). The overall proglacial lake and meltwater runoff patterns do not differ greatly between CCC and CCM1 climatologies.

We found greater variations in experiments with observed (present-day) precipitation patterns rather than AGCM $P - E$ fields. In the CCC model, $P - E$ at LGM is actually greater than that of present day over many regions of North America, due in part to low evaporative fluxes in the cold climate. While this was likely the case along the southern margins, a more dry glacial climate is expected in the interior of the ice sheet. The impact of an arid climate was tested by using the CCC temperature-field predictions and scaling present-day precipitation rates as a function of the temperature perturbation, after Tarasov and Peltier (1997),

$$P(\lambda, \theta, t) = P_{\text{obs}}(\lambda, \theta)(1.03)^{\Delta T(\lambda, \theta)}.$$  (9)

Figure 10 plots hydrological cycle diagnostics for the deglacial period from a selection of experiments. The CCC/GRIP reference model (bold line) contains 87 msl of ice at 20,000 yr B.P. (Fig. 10a). The other three models presented in Figure 10 use an approximation of the ICE-4G LGM ice sheet (Peltier, 1994) as an initial condition, containing an ice volume of 58 msl. ICE-4G ice sheet margins are geologically constrained, and we believe this to be the most realistic initial condition possible for simulating the deglaciation. In Figure 10, case CCC/ICE-4G uses the CCC/GRIP climatology, identical to that of the reference model. Cases CCC/pconst and CCM/pconst use CCC and CCM1 temperature fields with precipitation calculated from Eq. (9).

The main difference in the ice-volume time series of Figure 10a is the initial condition. With the CCC/GRIP forcing and the CCM temperature fields, the ICE-4G ice sheet grows steadily until 14,000 yr B.P. After this point all models behave similarly, with full deglaciation in the period 14,000–7000 yr B.P. and two major meltwater events. The first freshwater pulse has almost identical magnitudes in all cases (Fig. 10c), despite the significant differences in ice volume; this is because the areal cover of ice is similar in the experiments, giving the same amount of vulnerable ice. Contrastingly, meltwater pulse IB is weaker in low-ice volume cases, as this runoff represents wastage of the bulk of the Laurentide and Cordilleran ice.

The other very evident difference in total runoff is attributable to the low precipitation rates in cases CCC/pconst and CCM/pconst. When glacial meltwater is not dominating the hydrological budget (e.g., from 20,000–14,000 yr B.P.), total runoff in these two cases is about 50% of that with the CCC $P - E$ fields. Models converge in the Holocene as all cases relax to present-day precipitation rates. The lake volume record of Figure 10b shows considerably greater, and less predictable, variability between models. The isostatic basin created by the CCC/GRIP ice sheets is deeper, and this leads to greater lake volumes. All simulations predict maximum water volumes of 3–4 msl impounded on the continent, however, with two to three peaks in lake volume and maximum water storage occurring close to 10,000 yr B.P.

Figure 11 plots hydrological budget terms in these four tests, clearly illustrating the differences in net continental rainfall between cases. In the arid cases, rainfall and glacial melt play almost equal roles during cold climates, with surface melt swamping precipitation during deglaciation. The CCC clima-
tology gives a more vigorous hydrological cycle during cold periods, and even this extremely stable period from 20,000–14,000 yr B.P. has greater runoff variability than that predicted for the Holocene. This is seen in all models, suggesting that differences in ice-sheet thickness and geometry are secondary for net continental hydrological budget and runoff patterns. Overall, precipitation rates account for most of the differences in the water balance in our simulations.

River runoff histories in the four main drainage basins are presented in Figure 12 for each model treatment. Interesting differences in the drainage system evolution are evident here. Mississippian freshwater flux (Fig. 12a) primarily sees the impact of differing rainfall south of the ice sheet; this is the region that becomes significantly more wet in the CCC model (Marshall et al., in press). In all cases the Mississippi becomes a minor player after the first meltwater pulse, while drainage to the North Atlantic steps up until ca. 10,500 yr B.P., with a distinct double-peak structure. Atmospheric moisture is again the main source of discrepancy between models. The story differs in the high-latitude basins, which do not see significant runoff or rainfall during full glacial conditions.

The roles of ice-sheet geometry and volume become important for regional deglaciation and water pathways in parts of the continent draining to the Arctic and Hudson Strait. Figure 12c illustrates the earlier onset of Arctic freshwater runoff in cases with the ICE-4G initial condition. More rapid isostatic recovery in these cases may also contribute to the early and amplified freshwater diversion to the Arctic. The combination of CCC temperature fields and low precipitation rates appears to have important impacts on the regional deglaciation here, opening up Arctic drainage routes as early as 13,400 yr B.P. The timing of runoff diversion to Hudson Strait in Figure 12d is similarly displaced in different simulations, although in this case the CCM1 climatology gives the greatest variation in regional deglaciation, which controls the opening of Hudson Strait.

Despite these important timing discrepancies, all four tests predict a staged deglaciation following the switch to St. Lawrence drainage at 14,000 yr B.P. We conclude from these tests that the general patterns and behavioral modes of the coupled ice/bed/hydrology evolution are robust, although the precise timing and magnitude of freshwater fluxes are uncertain in the model. Deglacial meltwater pulses are extreme in all simulations, however, of order 0.3–0.4 Sv in the Mississippi and St. Lawrence (North Atlantic) drainages. This is certainly in the range of freshwater perturbation that can be expected to disrupt North Atlantic deepwater formation.

**FIG. 11.** Time series of hydrological budget terms in the deglaciation experiments, Sv. (Heavy line) Net river runoff. (Fine line) Glacial meltwater. (Dashed line) Rainfall.

**FIG. 12.** Time series of river basin runoff in the deglaciation experiments, Sv. (Heavy line) CCC/GRIP (reference) model. (Fine line) CCC/ICE-4G model. (Heavy dashed line) CCM/pconst model, ICE-4G initial ice. (Light dash-dot line) CCC/pconst model, ICE-4G initial ice.
DISCUSSION

Our simple representation of the hydrological system describes a great deal of the structure expected in the natural system. Despite uncertainties and approximations in the model ingredients, results show promise for rough estimation of proglacial lake and drainage system evolution. This offers the exciting possibility of allowing one to close the global balance in coupled ocean/atmosphere/land-surface/ice-sheet models of the paleohydrological cycle.

The modeled hydrological cycle is greatly enhanced from present day throughout the glacial period, with runoff from ice-sheet melt up to four times greater than that associated with rainfall. In addition to amplification of the hydrological cycle, a large degree of temporal and spatial runoff variability is exhibited in the presence of mid-latitude ice sheets. This is a characteristic of most of the last glacial cycle. Severe interstadial periods impact most of the runoff basins, although freshwater delivery to the Gulf of Mexico, North Atlantic, and North Pacific is the most susceptible to millennial-scale variability. During deglaciation, responses in different basins are asynchronous. We capture drainage shifts from the Gulf of Mexico to the St. Lawrence, Mackenzie, and Arctic catchments, although the chronology and magnitude of fluxes in our model must be accepted with caution. The timing of ice-sheet retreat is critical to details of the lake and runoff history.

Many trends in modeled drainage evolution do appear robust. In particular, even the minimum predictions of North Atlantic freshwater forcing suggest runoff rates of sufficient magnitude to perturb thermohaline circulation. Long and sustained periods of runoff variability, particularly given the tremendously amplified seasonal cycle, may keep the North Atlantic in a state of marginal thermohaline stability throughout the glacial interval (Fanning and Weaver, 1997). Millennial-scale runoff variability is intense in our experiments, but it is important to stress that the simulations do not explain the time scale or forcing mechanism of D–O events in any way; ice-sheet volume changes are forced from the GRIP climatology, which has the interstadial/stadial D–O events built in. The millennial-scale freshwater pulse cycles are a result of this forcing and we are therefore left with a classical chicken-and-egg problem. Continental runoff should, nevertheless, be recognized as a critical component of the glacial North Atlantic freshwater budget, supplementing the freshening associated with iceberg melt (Blunier et al., 1998; Stocker, 1998; Weaver, in press). Runoff fluxes are considerably in excess of the iceberg flux associated with Heinrich events (Dowdeswell et al., 1995; Marshall and Clarke, 1997b; Clarke et al., in press), but the geographic distribution of the freshwater perturbation differs importantly (see Weaver, in press). Both the location and amplitude of freshwater input play roles in modulating NADW production, so we anticipate that freshwater runoff and iceberg fluxes each play a role in climate instability during glacial periods.

Future Work

The lack of river physics and the necessarily coarse resolution of the study limit significance with regard to the details of modeled lake extent and timing. At this stage in the modeling, we do not expect to replicate individual shorelines or the intricate pattern of connections between proglacial lakes (cf. Teller, 1987). Improvements require a more sophisticated model of lake drainage. Enhanced model resolution would also improve the representation of lakes; continent-scale modeling by Coe (1998) suggests that grid cells of less than 10 km are needed to adequately resolve modern lake basins. Alternatively (or additionally), lake geometry might be refined and outlet channels identified through clever application of subgrid topographic information such as bed hypsometry. Advanced continent-scale surface hydrology schemes being developed for AGCM studies may prove very applicable to this problem. In addition, a full spherical visco-elastic model of isostatic adjustment is likely warranted to address meltwater routing properly. The present treatment does not give a good representation of marginal fore-bulges, which are important to proglacial lake history.

Given the simplifications and uncertainties associated with the ice-sheet, isostatic, and climate models, however, we are encouraged by the qualitative resemblance to proglacial lake history in North America. The geological record of ice-sheet and lake evolution through the deglaciation is of high quality and should be embraced to foster improvements in this sort of modeling. If the simulated freshwater runoff history can be represented in a reasonable fashion through the deglaciation, taking the Licciardi et al. (in press) reconstructions as an observational guide, we may be able to assess the veracity of simulations from earlier periods, which lack observational constraint. Preliminary results presented here suggest a great deal of structure in the freshwater drainage history throughout the last glacial period, with interesting implications for North Atlantic thermohaline circulation.

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