Origin of the middle Pleistocene transition by ice sheet erosion of regolith

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Abstract. The transition in the middle Pleistocene (~0.9 Ma) seen in δ¹⁸O deep-sea-core records from relatively low-amplitude, high-frequency (41 kyr) to high-amplitude, low-frequency (100 kyr) ice volume variations under essentially the same orbital forcing can be attributed to a change from an all soft-bedded to a mixed hard-soft bedded Laurentide ice sheet through glacial erosion of a thick regolith and resulting exposure of unweathered crystalline bedrock. A one-dimensional ice sheet and bedrock model which includes transport of sediment and ice by subglacial sediment deformation demonstrates that a widespread deforming sediment layer maintains thin ice sheets before the transition which respond linearly to the dominant (23 and 41 kyr) orbital forcing. Progressive removal of the sediment layer eventually causes a transition to thicker ice sheets whose dominant timescale of change (100 kyr) reflects nonlinear deglaciation processes. In model simulations over the last 3 Ma initialized with no ice and a uniform 50 m sediment layer the time series of ice volume and extent agree in several important aspects with the observed records.

1. Introduction

Oxygen isotopes (δ¹⁸O) measured in benthic foraminifera from deep-sea sediments reveal that former northern hemisphere ice sheets cyclically waxed and waned at the same periodicities (100, 41, and 23 kyr) as the orbital parameters that control the amount and distribution of solar radiation received by the Earth, suggesting a causal relationship between orbital changes in insolation and ice volume (Milankovitch hypothesis) [Hays et al., 1976; Imbrie et al., 1984]. However, there are two significant problems in explaining global ice volume variations since the onset of northern hemisphere continental glaciation at ~2.8 Ma solely by orbital variations in insolation: (1) a transition, in the middle Pleistocene (~0.9 Ma), from low-amplitude, higher-frequency (41 kyr) to high-amplitude, lower-frequency (100 kyr) ice volume variations under essentially the same orbital forcing and (2) the dominance of the 100 kyr ice volume signal since the middle Pleistocene when insolation forcing at that period (eccentricity) has essentially no power (Figure 1) [Shackleton and Opdyke, 1976; Pisias and Moore, 1981; Start and Prell, 1984; Ruddiman et al., 1989; Imbrie et al., 1993; Mix et al., 1995].

Many hypotheses (summarized by Imbrie et al. [1993]) have been developed to explain these two aspects of the evolution and response of the earth’s climate system to the known orbital forcing. Generally speaking, a large number of models have successfully reproduced the ~23 and 41 kyr cycles in direct response to the orbital Milankovitch forcing, but only a few have reproduced the observed dominant 100 kyr cycles with rapid and complete deglaciations over the last ~0.9 Myr. To achieve the latter, it has been necessary to include additional physical processes (e.g., calving [Pollard, 1983], meltwater discharge feedback [Pollard, 1983; DeBlonde and Peltier, 1991], or snow aging [Gallee et al., 1992]) or to prescribe additional time-dependent forcing (atmospheric dust loading [Peltier and Marshall, 1995] or atmospheric CO₂ variations [DeBlonde et al., 1996]). Furthermore, relatively few attempts have been made to model the middle Pleistocene transition observed in marine δ¹⁸O records, i.e., with relatively little 100 kyr power before ~0.9 Ma, and dominant 100 kyr cycles after that time. In general, these ice sheet-climate models produce a transition as a nonlinear response to a prescribed long-term cooling trend [Oerlemans, 1984; Saltzman and Maasch, 1991; Abe-Ouchi, 1996] or to a suddenly imposed switch in model physics [DeBlonde and Peltier, 1991; Mudelsee and Schütz, 1997]. However, evidence for a general cooling trend or decrease in CO₂ over the last 2 Ma is uncertain [Raymo et al., 1997]. Nevertheless, a gradual global cooling continuing through the last million years would be a viable candidate for the cause of the middle Pleistocene transition.

Previous workers have also suggested that a change from a land-based to a marine-based ice sheet through long-term glacial erosion was responsible for the middle Pleistocene transition [Pisias and Moore, 1981; Berger and Jansen, 1994]. Deep erosion beneath the large Laurentide ice sheet, causing it to become a marine-based ice sheet [Pisias and Moore, 1981], however, does not appear to be a viable hypothesis [Sugden, 1976; Bell and Laine, 1985; Dyke et al., 1989]. A transition to a marine-based condition may have occurred beneath the Barents Sea ice sheet [Berger and Jansen, 1994], but the timing of such a transition is not well constrained [Hjelstuen et al., 1996; Solheim et al., 1996], and the small size of the Barents Sea ice sheet could not account for the spectral change seen in the δ¹⁸O record (Figure 1).
Figure 1. (a) Summer half-year insolation (in W m⁻²) at 55°N for the last 3 Myr computed using orbital elements from Berger and Loutre [1991]. (b) Oxygen isotope records for Ocean Drilling Program Sites 677 and 849 over the last 3 Myr (data and age models are from Mix et al. [1995]).

We describe geologic data and modeling results which suggest that a change in the basal boundary condition of the Laurentide ice sheet was responsible for the middle Pleistocene transition. In particular, our hypothesis invokes only internal ice sheet mechanisms with no prescribed external trends except for Milankovitch orbital forcing. Our proposed mechanism accounts for the otherwise perplexing evidence that the late Pleistocene-early Pleistocene Laurentide ice sheet was as extensive as but volumetrically significantly smaller (Figure 1) than the later, similarly extensive but volumetrically larger Laurentide ice sheet. Furthermore, modeling results identify how this change in ice sheet size, in turn, resulted in a different response of the ice sheet to the same orbital forcing, giving rise to the middle Pleistocene spectral change.

2. Geologic Records for Change in Basal Boundary Condition

The Laurentide and Fennoscandian ice sheets both developed at 2.6-2.8 Ma [Raymo, 1994; Clemens et al., 1996], but the Fennoscandian ice sheet has remained much smaller than the Laurentide. (We note that our proposed mechanism for the middle Pleistocene transition may equally well apply to the Fennoscandian ice sheet). The East Antarctic Ice Sheet has remained stable since the middle Miocene [Marchant et al., 1996], with orbital-scale changes in the ice sheet being "paced" by global sea level changes driven from the northern hemisphere [Denton et al., 1986]. We thus assume that the Laurentide ice sheet has dominated the global δ¹⁸O signal during the last ~2.8 Myr.

Geologic records suggest that the Laurentide ice sheet was at times as or more extensive in the late Pliocene and early Pleistocene than it was in the late Pleistocene. Several tills deposited by the Laurentide ice sheet in Iowa and Nebraska, or south of the ice margin during the last glacial maximum (~21 ka), are overlain by the Huckleberry Ridge Ash [Boeblstorff, 1973, 1978], which was derived from the eruption of the Yellowstone caldera at 1.97 Ma [Sarna-Wojcicki and Davis, 1991]. Similarly, a δ¹⁸O record from the Gulf of Mexico suggests that the southern margin of the Laurentide ice sheet repeatedly entered the Mississippi River drainage after ~2.3 Ma and discharged light δ¹⁸O meltwaters to the Gulf of Mexico [Joyce et al., 1993].

Because the global δ¹⁸O record suggests that late Pliocene-early Pleistocene ice sheets were volumetrically between one half to two thirds as large as late Pleistocene ice sheets (Figure 1), the existence of a Laurentide ice sheet during the late Pliocene and early Pleistocene that was as extensive as during the late Pleistocene must reflect a change from a thinner to a thicker ice sheet. Ice sheet thickness is controlled primarily by mechanisms at the base of the ice sheet which control ice movement. Ice sheets which move entirely by internal deformation of ice will generally be thicker than ice sheets which move by some combination of internal ice deformation, basal sliding, and subglacial sediment deformation (soft beds) [Paterson, 1994]. A change from a thinner to a thicker ice sheet during the middle Pleistocene thus implicates a change in the basal boundary condition which governs ice thickness.

Several characteristic landform and sediment assemblages suggest that soft beds underlay a significant fraction (of the order of 50%) of former northern hemisphere ice sheets during the last glacial maximum [Alley, 1991; Hicock and Dreimanis, 1992; Clark and Walder, 1994; Clark, 1997]. In the areas covered by the Laurentide ice sheet, easily deformable finely grained till nearly completely mantles areas of sedimentary bedrock, so exposed bedrock is largely absent. In contrast, hard-bedded areas, which are characterized by discontinuous, coarsely grained till with a large fraction of exposed bedrock, are typical of crystalline bedrock areas underlying the Laurentide ice sheet such as the Canadian shield [Marshall et al., 1996]. A numerical reconstruction of the Laurentide ice sheet with this prescribed distribution of hard and soft beds [Clark et al., 1996] agrees with the ICE-4G reconstruction [Peltier, 1994] both in surface topography and ice volume. This agreement suggests that the distribution of soft beds provides a glaciological explanation for the shape and volume of the Laurentide ice sheet during the last glacial maximum that is most consistent with observations of relative sea level and other geodynamic considerations [Clark et al., 1996].

We argue that prior to the middle Pleistocene transition the Laurentide ice sheet was thinner because it was underlain everywhere by a soft, easily deformable substrate and that the transition to a thick ice sheet during the middle Pleistocene resulted from a change to an ice sheet with an increasing distribution of hard beds. Geologic data and modeling results discussed below suggest that the change from all soft-bedded to a mixed-bedded ice sheet occurred as thick, clayey soils (regoliths), which had developed by chemical weathering of crystalline bedrock, were progressively eroded by successive glaciations,
eventually exposing crystalline bedrock to the ice sheet as a hard bed.

The development of such a regolith is expected given the long time (of the order of 10^7-10^9 years) that the crystalline bedrock was exposed to weathering prior to the inception of northern hemisphere glaciation. Direct evidence for this regolith comes from glaciated central and western Minnesota, where a deep saprolite developed in Precambrian crystalline bedrock is preserved [Goldich, 1938; Parham, 1970; Setterholm and Morey, 1995]. Stratigraphic relations suggest that the saprolite formed during the Mesozoic and possibly the early Tertiary, when global climate was significantly warmer than present [Crowley and North, 1991]. The weathering profile averages 30 m in thickness, but locally is as thick as 60 m [Parham, 1970; Setterholm and Morey, 1995]. Similar, though less extensive, pre-Quaternary saprolites preserved on glaciated crystalline bedrock also occur in eastern Canada [Wang et al., 1981, 1982; McKeague et al., 1983], and well-developed saprolites are common in the Appalachian Mountains south of the glacial border [Mills and Delcourt, 1991].

Chemical weathering involves the removal of the more reactive mineral species, resulting in a composition that reflects the more resistant minerals as well as the products of chemical weathering of the more reactive species. During chemical weathering, amphiboles alter first, followed by biotite, then the feldspars, and lastly quartz [Goldich, 1938; Setterholm and Morey, 1995]. Plagioclase feldspars are more reactive than K-feldspar, and quartz persists in all but the most intensely weathered saprolite. Dissolution of hornblende and biotite releases ferrous iron, which migrates upward to a level where oxidizing conditions fix it as either goethite or hematite. Chemical weathering also involves the dissolution of feldspars and the concurrent removal of Ca, Na, and K. Al is concentrated in the remaining residue and is thus available to form kaolinite.

These relations suggest that erosion and transport of a regolith will yield mineralogies that differ from those of the source rocks but resemble those found in the weathering profile mantling the source rocks. Sediments derived from fresh bedrock by glacial and fluvial abrasion which have undergone minimal chemical weathering have essentially the same bulk mineralogy as the bedrock source [Nesbitt and Young, 1996]. In contrast, the mineralogy of sediments derived from a heavily weathered crystalline bedrock reflect the composition of the regolith, not the bedrock [Nesbitt et al., 1996].

Existing geologic data are consistent in identifying a change in sediment source from a regolith to fresh bedrock around the time of the middle Pleistocene transition. The oldest (unweathered) tills in the glaciated midcontinent (>2 Ma) are relatively enriched in stable minerals and weathered products (iron oxides) and depleted in unstable minerals (hornblende, biotite, and apatite) compared to younger tills (Table 1) [Boellstorff, 1978]. The oldest tills also contain allochthonous fragments of bauxite (a product of prolonged chemical weathering under tropical to subtropical climates) [Gravenor, 1975]. Finally, the older tills are enriched in sedimentary rock clasts (~90%) whereas the younger tills are enriched in crystalline lithologies [Boellstorff, 1978], which may reflect the change in source from erosion of highly weathered to fresh crystalline bedrock.

Additional mineralogical evidence for a change in basal boundary conditions during the middle Pleistocene comes from sediments drilled by Ocean Drilling Program (ODP) site 645 in Baffin Bay (~71°N), which record a direct influence from the adjacent Laurentide and Greenland ice sheets beginning with the first ice rafting of pebbles in the Pliocene [Thiebault et al., 1989]. In particular, there is a significant change in mineralogy at ~0.95 Ma; sediments older than 0.95 Ma are characterized by clay minerals originating from weathered continental deposits (irregular layered clay layers and "smectite"), whereas sediments younger than 0.95 Ma show an increase in clay minerals derived from crystalline bedrock (chlorite and illite) [Thiebault et al., 1989; Andrews, 1993]. Coincident with this transition is a significant increase in feldspar [Andrews, 1993], which is consistent with a source change from a regolith, in which feldspars have weathered, to crystalline bedrock.

### Table 1. Heavy Mineral Percentages in Tills from Nebraska

<table>
<thead>
<tr>
<th>Till Unit</th>
<th>Hornblende</th>
<th>Biotite</th>
<th>Apatite</th>
<th>Limonite</th>
<th>Hematite</th>
<th>Opaques</th>
</tr>
</thead>
<tbody>
<tr>
<td>&quot;A&quot; tills</td>
<td>30.7</td>
<td>4.3</td>
<td>0.8</td>
<td>6.4</td>
<td>2.5</td>
<td>21.8</td>
</tr>
<tr>
<td>&quot;B&quot; tills</td>
<td>21.1</td>
<td>3.3</td>
<td>0.7</td>
<td>14.3</td>
<td>3.5</td>
<td>35.9</td>
</tr>
<tr>
<td>&quot;C&quot; tills</td>
<td>8.6</td>
<td>1.6</td>
<td>0.1</td>
<td>31.4</td>
<td>6.8</td>
<td>61.3</td>
</tr>
</tbody>
</table>

Data from tills exposed at City Wide Rock Quarry section, Nebraska [Boellstorff, 1973, 1978; Gravenor, 1975]. The "C" tills are overlain by volcanic ash from the Yellowstone caldera dated at 1.97 Ma, the "B" tills overlay a Yellowstone ash dated at 1.27 Ma, and a Yellowstone ash dated at 0.62 Ma is interbedded with the "A" tills [Boellstorff, 1973, 1978; Sarna-Wojcicki and Davis, 1991].

3. Modeling the Transition

We tested our hypothesis using a one-dimensional ice sheet and bedrock model, in which ice thickness and bedrock elevation are predicted as functions of latitude and time, representing a roughly north-south profile along a typical flow line through the Laurentide ice sheet. Such models have been used extensively to model long-term ice sheet cycles [Oerlemans, 1980; Pollard, 1983; Birchfield and Grumbine, 1985; DeBlonde and Pelletier, 1991]. Ice flow is described by a vertically integrated continuity equation for ice thickness, assuming the ice sheet deforms mainly by shear at or near its base under its own weight:

$$
\frac{\partial h_I}{\partial t} = \frac{\partial}{\partial x} \left[ A h_I^{n} \left( \frac{\partial (h_I + h_s + h_b)}{\partial x} \right)^{\beta} \frac{\partial (h_I + h_s + h_b)}{\partial x} \right] - \frac{\partial}{\partial x} [u_s(0) h_I] + B - C
$$

(1)
Table 2. Symbols Used in the Model

<table>
<thead>
<tr>
<th>Symbol</th>
<th>Meaning</th>
<th>Value</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>x</td>
<td>north-south distance</td>
<td>independent variable</td>
<td>m</td>
</tr>
<tr>
<td>t</td>
<td>time</td>
<td>independent variable</td>
<td>yr</td>
</tr>
<tr>
<td>(h_i(x,t))</td>
<td>ice sheet thickness</td>
<td>prognostic variable</td>
<td>m</td>
</tr>
<tr>
<td>(h_s(x,t))</td>
<td>sediment layer thickness</td>
<td>prognostic variable</td>
<td>m</td>
</tr>
<tr>
<td>(h_b(x,t))</td>
<td>bedrock elevation above sea level</td>
<td>prognostic variable</td>
<td>m</td>
</tr>
<tr>
<td>(h_{eq})</td>
<td>equilibrium bedrock elevation above sea level</td>
<td>prescribed versus (x) (see text)</td>
<td>m</td>
</tr>
<tr>
<td>(B)</td>
<td>mass balance on ice surface</td>
<td>prescribed pattern (see text)</td>
<td>m yr(^{-1})</td>
</tr>
<tr>
<td>(C)</td>
<td>removal of ice by calving</td>
<td>20 if &quot;floating&quot; (see text)</td>
<td>m yr(^{-1})</td>
</tr>
<tr>
<td>(\alpha)</td>
<td>ice rheology exponent</td>
<td>5</td>
<td></td>
</tr>
<tr>
<td>(\beta)</td>
<td>ice rheology exponent</td>
<td>2</td>
<td></td>
</tr>
<tr>
<td>(\tau)</td>
<td>timescale of bedrock relaxation</td>
<td>5000</td>
<td>yr</td>
</tr>
<tr>
<td>(\rho_s)</td>
<td>ice density</td>
<td>910</td>
<td>kg m(^{-3})</td>
</tr>
<tr>
<td>(\rho_{bs})</td>
<td>sediment density (saturated bulk)</td>
<td>2390</td>
<td>kg m(^{-3})</td>
</tr>
<tr>
<td>(\rho_b)</td>
<td>bedrock density</td>
<td>3370</td>
<td>kg m(^{-3})</td>
</tr>
<tr>
<td>(z)</td>
<td>vertical coordinate in sediment layer</td>
<td>independent variable</td>
<td>m</td>
</tr>
<tr>
<td>(u_s)</td>
<td>sediment horizontal velocity</td>
<td>diagnostic versus (z) (see text)</td>
<td>m yr(^{-1})</td>
</tr>
<tr>
<td>(D_{fs})</td>
<td>sediment reference deformation rate</td>
<td>(7.9 	imes 10^{-4})</td>
<td>yr(^{-1})</td>
</tr>
<tr>
<td>(\phi)</td>
<td>sediment reference viscosity</td>
<td>(3 \times 10^6) (Figure 2a), (10^9) (Figure 2b), (3 \times 10^8) (Figure 2c), (6 \times 10^7) (Figure 2d)</td>
<td>Pa s</td>
</tr>
<tr>
<td>(n)</td>
<td>sediment rheology exponent</td>
<td>1.25 (except 3 in Figure 2d)</td>
<td></td>
</tr>
</tbody>
</table>

where \(x\) is north-south distance, \(h_i(x,t)\) is the ice thickness, \(h_s(x,t)\) is the sediment thickness, \(h_b(x,t)\) is the bedrock elevation above sea level, \(A = 5.77 \times 10^4\) m yr\(^{-1}\), \(\alpha = 5\), and \(\beta = 2\) (Table 2). The first term on the right-hand side represents internal ice shear and/or basal sliding at the ice-bed interface, as in the earlier models. The second term, involving \(u_s(0)\), is due to horizontal shear within the sediment layer below the ice, as described below. \(B\) is the net annual mass balance on the ice surface, representing the climatic distribution of precipitation and ice melt, and is prescribed as a function of latitude and elevation that is shifted vertically in proportion (35 m W\(^{-1}\) m\(^{-3}\)) to long-term orbital variations in the summer half-year insolation at 55°N [De Jager, 1980; Berger and Loutre, 1991] (see Figure 1).

The last term \(C\) represents calving by proglacial lakes and/or marine incursions [Pollard, 1983]. A large body of water is assumed to form adjacent to the ice sheet whenever the bedrock falls below sea level at the margin. The term \(C\) is set to 20 m yr\(^{-1}\) at any ice sheet grid point if its base is below sea level \((h_i + h_s < 0)\) and if it or one of its neighboring points has floating ice \((\rho_i > \rho_f(h_i + h_s))\). The second condition protects the thick interior ice and limits calving, generally, to the outermost 50-150 km of the ice sheet. In practice, the southern tip is attacked in this way only during deglaciations and is responsible for the model's realistic simulation of the dominant 100 kyr cycles during the last ~1 Myr [Pollard, 1983].

The depression or rebound of the underlying bedrock is simply a local relaxation to isostatic equilibrium, lagged by 5000 years [Peltier and Marshall, 1995]:

\[
\frac{\partial h_b}{\partial t} = \frac{1}{\tau} \left[ h_{eq}^b - h_b - \frac{(\rho_i h_i + \rho_s h_s)}{\rho_b} \right] (2)
\]

where \(h_{eq}^b\) is the ice-free equilibrium bedrock elevation above sea level, taken to be 500 m above sea level southward of 70°N, ramping linearly down to -500 m at 74°N to represent the Arctic Ocean (which limits the northern extent of the ice sheet via calving). Also, \(\tau = 5000\) years, \(\rho_b = 3370\) kg m\(^{-3}\) is the bedrock density, \(\rho_i = 910\) kg m\(^{-3}\) is the ice density, and \(\rho_s = 2390\) kg m\(^{-3}\) is the sediment (saturated bulk) density.

Equations (1) and (2) are standard except for calving and the extra ice motion due to deformation (second term on the right-hand side of (1)). The latter is modeled by a sediment layer of thickness \(h_i(x,t)\) resting on bedrock. Deformation within the sediment layer follows Jenson et al. [1995, 1996], in which the sediment shears according to a slightly nonlinear flow law in response to horizontal shear stress applied by the ice sheet at the top of the sediment. The sediment is assumed to be saturated with liquid water, which transmits the weight of the ice through the upper sediment to the undeforming layers below, so that the effective confining pressure in the shear zone is small and due only to the buoyant weight of the sediment itself. By solving their equation (2) the resulting horizontal velocity profile within the sediment is

\[
u_s(z) = \frac{2D_0 a}{(n+1) b} \left( \frac{a}{2D_0 \mu_0} \right)^n \left[ \left( 1 - \frac{z}{a} \right)^{n+1} - \left[ 1 - \frac{b}{a} \min \left( h_i + h_s, h_b \right) \right]^{n+1} \right]
\]

where \(z\) is the vertical coordinate increasing downward from zero at the ice-sediment interface, \(a = \rho_i g h_i \left| \frac{\partial (h_i + h_s + h_b)}{\partial x} \right|\) is the shear stress at the ice-sediment interface, and \(b = (\rho_b - \rho_i) g \tan(\phi)\) is the rate of increase of shear strength with depth. The \(u_s(z)\) is taken to be zero at and below the bottom of the shear zone at \(z = a/b\) or at the bottom of the sediment layer itself at \(z = h_i\) if \(h_i < a/b\). The shear-zone thickness \(a/b\) depends on the ice sheet slope and thickness but is typically 1-10 m in our model. We use Jenson et al.'s [1995, 1996] nominal values for the sediment rheology, \(\phi = 22^\circ\) for the angle of internal
friction, $D_o = 7.9 \times 10^{-7} \text{s}^4$ for the reference deformation rate, $\mu_o = 3 \times 10^5 \text{ Pa s}$ for the reference viscosity, $n = 1.25$ for the exponent, and sediment cohesion is neglected. Also, $\rho_s = 1000 \text{ kg m}^{-3}$ is the liquid water density and $g = 9.80616 \text{ m s}^{-2}$ is the gravitational acceleration. Equation (3) assumes that the ice sheet base is at the melting point, allowing the till to be saturated. This assumption can be relaxed by including ice thermodynamics in the model, but as described below, we find that this makes very little difference to the results, at least with a relatively simple treatment of ice temperatures.

All the runs below were initialized with a uniform 50 m thick layer of sediment at 3 Ma ($h_i = 50 \text{ m}$ everywhere) as might be expected from the geologic evidence of preserved saprolites mentioned above. Thereafter the sediment thickness varies according to

$$\frac{\partial h_i}{\partial t} = -\frac{\partial}{\partial z} \left[ \frac{\min(h_i, a/b)}{k_{l'}} \mu_s(z) dz \right]$$

(4)

The condition $h_i \leq 100 \text{ m}$ is imposed for areas that are ice free ($h_i = 0$) and represents rapid erosion of excess sediment in an ice-free environment. Without some removal of sediment during ice-free periods the sediment in our model accumulates unrealistically to values of ~500 m around and under the southern margins by the late Pleistocene, which, in reality, must have been avoided by relatively rapid erosion, probably by river and eolian transport. The imposed condition includes this process as simply as possible with a minimum of free parameters; the value of 100 m is based roughly on the maximum thicknesses of observed Pleistocene tills [Mickelson et al., 1983; Klassen, 1989]. The results shown below are quite insensitive to this value, at least in the range 50-100 m, and also change little when this condition is replaced by horizontal diffusion of sediment in ice-free areas (not shown).

The horizontal velocity $u_s(0)$ at the ice-sediment interface given by (3) causes additional ice sheet transport via the second term in (1), which generally dominates the internal flow/basal sliding (first term in (1)) unless the sediment is very thin. Note that any basal sliding at the ice-sediment interface is not included in the sediment model but is considered to contribute to the first term in (1). The system of equations (1)-(4) are solved numerically on a $0.5^\circ$ latitudinal grid with a time step of 1 year.

4. Model Results

A 3 Myr model run with the nominal parameter values given above (Table 2) shows ice sheet "volume" (cross-sectional area of ice in the latitude versus height plane) and extent (distance from northern to southern tip) (Figure 2a). The only external forcing is the vertical shift of the mass balance pattern $B$ in (1) due to orbital insolation variations. From 3 Ma to ~2 Ma the ice transport is dominated by sediment deformation, which restricts the basal shear stresses to small values and constrains the ice sheet to remain relatively thin (~1.5 km) with gentle slopes near the margins (Figure 3). This results in wide southern ablation zones and prevents the ice sheet from achieving large volumes; however, the maximum linear extents (dashed curve) are close to those of the late Pleistocene in agreement with early observed ice limits as mentioned above.

During the first ~1 Myr of this run the sediment layer is advected rapidly toward the northern and southern margins of the ice sheet (Plate 1). By ~2 Ma most sediment has been removed poleward of ~62° N but remains thick under the southern flanks of the ice sheet, whose maximum extends to ~54°N around this time. Thus the southern slopes and total ice volumes are still inhibited by the sediment. It is not until ~1.5 Ma, when the marginal sediment has been gradually cleared several degrees latitude further south (Figure 3 and Plate 1), that the central zone of crystalline bedrock is sufficiently wide to support ice volumes close to maximum Pleistocene values. After 1.5 Ma the central bedrock depressions are deep enough (Figure 3) to allow significant calving, and the first complete deglaciation occurs at 1.45 Ma, with dominant 100 kyr cycles from then on as given by Pollard [1983]. Although the long-term erosion of model sediment and deepening of maximum bedrock depressions are gradual, the transition to 100 kyr cycles is sudden because the onset of the model's calving events is highly nonlinear and depends mainly on whether the southern tip drops below sea level as the ice sheet recedes. The transition in Figure 2a from relatively thin ice sheets and ~20-40 kyr cycles before 1.5 Ma to thicker ice sheets and dominant ~100 kyr cycles after that time (e.g., Figure 3) is in basic agreement with the observed middle Pleistocene transition in long-term δ18O records, whereby the observed transition begins at ~1.2 Ma and is complete by ~0.9 Ma (Figure 1). (It is still uncertain how abrupt the transition was [Maasch, 1988; Ruddiman et al., 1989; Park and Maasch, 1993].)

As discussed by Jenson et al. [1995, 1996], the most appropriate values for the sediment viscosity $\mu_s$ and the exponent $n$ in (4) are uncertain. Several diverse modeling and experimental studies on deformable tills suggest that $\mu_s$ might range from $10^8$ to $10^{11}$ Pa s and that $n$ might lie between 1.25 and ~3.

![Figure 2](image-url)
Figure 3. Cross sections of ice sheet, sediment (solid), and bedrock (stippled) at two different times from the run in Figure 2a. Although the linear north-south extents are the same at the two times, there is still sediment under much of the ice sheet during the earlier time so the ice sheet is relatively thin and the bedrock depression is shallow. In contrast, during the later time, there is no sediment remaining under the ice sheet so the ice sheet is thicker and the bedrock depression is deeper.

Plate 1. Changes in sediment layer thicknesses ($h_1$) (in meters) from the nominal run (Figure 2a). Blank areas indicate negligible sediment ($h_1 < 0.1$ m), and the "50-50" contour interval shows the initial 50 m thickness value. The dashed lines show the latitudes of the ice sheet's northern and southern tips.
[Boulton and Hindmarsh, 1987; Alley, 1989; Paterson, 1994; Vela et al., 1996], although much higher values of n have been suggested [Kamb, 1991]. We examined the effects of different sediment viscosities, with n fixed at 1.25. With much larger viscosity (Figure 2b), sediment deformation and the associated ice transport become negligible, and the model essentially reverts to the no-sediment model of Pollard [1983] with thick ice sheets, calving, and dominant ~100 kyr cycles throughout the run. With much smaller viscosity (Figure 2c), basal shear stresses at the ice-sediment interface are very low and severely limit the ice sheet volumes. Also, the sediment shear zone a/b is much thinner, and the vertically integrated sediment transport in (4) is smaller [Jenson et al., 1995], so the initial 50 m sediment layer takes longer to be cleared away. Thus ice volumes remain small over the entire run (Figure 2c), and the 100 kyr transition does not occur until ~0.2 Ma. Although we might claim from these results (Figures 2a-2c) that 3 x 10^5 Pa s is the most realistic value of sediment viscosity, this is really only an order-of-magnitude constraint since the timing of the transition also depends on our somewhat arbitrary choices of initial sediment thickness (50 m) and start date (3 Ma). By varying those parameters we could probably adjust the timing of the transition in the current model to agree more closely with the observed transition at ~0.9 Ma (Figure 1); however, we feel that sort of fine tuning will be more meaningful at a later stage after some of the model developments mentioned below. At that stage, objective statistical tests will probably be needed to compare the model and observed curves, especially since the exact timing of the transition is somewhat subjective and has been quoted as early as 1.2 Ma [Clemens et al., 1996].

Making the sediment rheology more nonlinear has relatively little effect on the results (Figure 2d). The value of n we used (= 3) is at the high end of the range suggested by a number of studies [Boulton and Hindmarsh, 1987; Alley, 1989; Paterson, 1994; Vela et al., 1996]. If n were increased by 1 or 2 orders of magnitude [Kamb, 1991], one would expect more dramatic effects since the sediment would become essentially plastic, and the whole deforming bed mechanism could become unstable if feedback from basal melting is considered [Kamb, 1991]. However, such large values of n are beyond the numerical design of our model.

5. Discussion

The model described here is obviously very simple compared to reality and contains a number of important assumptions. We regard it solely as a first quantitative test of the hypothesis that deformable beds are important for the middle Pleistocene transition. One assumption is that the calving mechanism is important for the 100 kyr cycles. Clearly, any model addressing the middle Pleistocene transition (with no 100 kyr cycles before and dominant cycles afterward) must be capable of simulating them at all. In our model a calving mechanism is necessary to augment the basic bedrock-depression/lapse-rate feedback to produce these cycles, but other models have suggested other mechanisms [Oerlemans, 1984; DeBlonde and Peltier, 1991; Saltzman and Maasch, 1991; Galley et al., 1992; Peltier and Marshall, 1995; DeBlonde et al., 1996; Abe-Ouchi, 1996; Muddelsee and Schulz, 1997], and their relative importance for the real 100 kyr cycles is uncertain. Nevertheless, several of these other models rely on the same basic behavior as here: that at roughly 100 kyr intervals the ice sheet must become very large and thick with a steep bedrock depression, and thus susceptible to complete and rapid deglaciation in the next warm-orbit interval. A deforming sediment layer prevents these conditions before ~1 Ma and allows them after it has been eroded, so the sediment model proposed here might be expected to produce the same middle Pleistocene transition independent of the exact mechanism for the 100 kyr cycles.

Another assumption is that the sediment is always saturated with just enough liquid water to transmit the vertical weight of the ice sheet. This requires the ice sheet base to be at the melting point; otherwise, the till is frozen, and no sediment deformation can occur. We have tested this assumption by predicting ice temperatures in the model, using a standard equation of vertical heat transfer involving conduction and vertical advection, with lower boundary condition from basal frictional heating and the geothermal heat flux, upper boundary condition from specified annual mean surface ice temperatures, and neglecting horizontal heat advection and thermal inertia [Paterson, 1981]. The annual surface ice temperature is estimated consistently with the mass-balance parameterization for B in (1) [Oerlemans, 1980] by assuming that air temperatures at the equilibrium-line altitude have an annual mean of -10°C, a seasonal amplitude of 25°C, and an atmospheric lapse rate of 7°C km^{-1}. This equation has an analytic solution [Paterson, 1981] that gives the steady state basal temperature at each grid point and time step in the model. Whenever the basal temperature is below the pressure melting point, sediment deformation from (3) and its associated ice transport is reset to zero. The resulting ice volume curves and sediment thickness distributions for the nominal model are virtually the same as our results (Figure 2a and Plate 1). Although the base is usually frozen in the central regions when the ice sheet is large, basal temperatures nearer the flanks are often at the melting point where most sediment transport occurs, so sediment thicknesses and ice volumes evolve over time in essentially the same way as before. The neglect of horizontal heat advection may be significant near the ice sheet flanks, and we plan to include that in future work [Ritz et al., 1997] and to investigate possible freeze-thaw cycles at the bed [MacAyeal, 1992, 1993] and other variations in basal hydrology [Kamb, 1991; Fowler and Johnson, 1995]. However, the first-order inclusion of ice temperatures to date shows that basal temperatures near the Laurentide flanks were at the melting point often enough to allow the sediment transport mechanism proposed here to remain viable. Other possible future model developments are a more realistic nonlocal model for the lithospheric and asthenospheric response [Birchfield and Grumbine, 1985; DeBlonde and Peltier, 1991], the extension to two horizontal dimensions and the use of two-dimensional (2-D) or 3-D climate models [Peltier and Marshall, 1995; DeBlonde et al., 1996], more explicit models of calving water bodies and discharging ice shelves [Pollard, 1983; van der Veen, 1996], refinement of the sediment removal process in (4b), and remoulding of the sediment under the ice causing long-term changes in sediment rheology.

As Birchfield and Ghil [1993] and others have noted, there is relatively little ~23 kyr spectral power in 518O records prior
to the middle Pleistocene transition as compared to after the transition. If the early ice sheets did extend as far south as in the late Pleistocene, as suggested by continental records, one might expect substantial 23 kyr power since variations of orbital precession at those periods dominate the insolation signal more at low latitudes. There may be two reasons why this signal is less before the transition. First, ice volume responds to changes in accumulation and ablation over the whole ice sheet, which, in turn, are affected by large-scale changes in climate and circulation that may differ for thin versus thick ice sheets and do not just follow insolation variations at the southern margin. In our simple model the climate forcing is tied to insolation variations at an arbitrary latitude (55°N), and global climate models will be needed to fully explore this possibility. Second, as suggested by Birchfield and Ghil [1993], there may simply be more downcore loss of signal at shorter periods (~23 kyr compared to ~41 kyr) because of postdepositional diffusive processes such as bioturbation.

In conclusion, we have used a simple numerical model to test the hypothesis that a change from soft-bedded to hard-bedded conditions due to glacial erosion of an initially thick and uniform regolith and progressive exposure of unweathered crystalline bedrock caused the middle Pleistocene transition in 8³¹O records. This first quantitative test shows that with reasonable values of sediment rheology the northern hemisphere ice sheets could have eroded several tens of meters of regolith within the right time frame of 1-2 Myr. Before the transition the model's deforming sediment maintains relatively thin, low-volume ice sheets that nevertheless, extend nearly as far south as the later ice sheets, in agreement with several lines of evidence discussed above. The exposure of undeforming bedrock below the regolith allows ice sheet thicknesses and bedrock depressions to become large enough to allow marine incursions and calving that are responsible for complete model deglaciations and the introduction of the dominant nonlinear ~100 kyr response in the middle Pleistocene.

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