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Modeling the oxygen-isotopic composition of the North American Ice Sheet and its effect on the isotopic composition of the ocean during the last glacial cycle

Adriana Sima,[1,2] André Paul,[3] Michael Schulz,[3] and Johannes Oerlemans[4]

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[1] We used a 2.5-dimensional thermomechanical ice-sheet model including the oxygen-isotope ratio $^{18}O/^{16}O$ as a passive tracer to simulate the isotopic composition ($\delta^{18}O$) of the North American Ice Sheet (NAIS) during the last glacial cycle. This model allowed us to estimate the NAIS contribution to the change of seawater $\delta^{18}O$ (δ_w) between the Last Glacial Maximum (LGM) and the Holocene and to evaluate the effect of nonequilibrium isotopic composition of the NAIS on the relationship between ice-volume variations and the ocean isotopic enrichment. The enrichment due to the NAIS at the LGM was 0.63‰, corresponding to ~60% of the LGM sea-level lowstand and to a mean $\delta^{18}O$ of the NAIS of approximately −31‰. The modeled NAIS volume variations and the induced δ_w changes over the past 120,000 years indicated no significant time lag. The inaccuracy associated with linearly inferring ice-volume variations from δ_w changes was generally less than 10%. **Citation:** Sima, A., A. Paul, M. Schulz, and J. Oerlemans (2006), Modeling the oxygen-isotopic composition of the North American Ice Sheet and its effect on the isotopic composition of the ocean during the last glacial cycle, *Geophys. Res. Lett.*, *33*, L15706, doi:10.1029/2006GL026923.

1. Introduction

[2] The oxygen-isotopic composition ($\delta^{18}O$) of seawater (δ_w) is a key parameter in paleoclimate research. It depends on global ice volume (V_g) and the mean isotopic composition of ice (δ_i), and, together with seawater temperature, it determines the isotopic composition of foraminiferal carbonate (δ_c). The relationship between V_g (or equivalent sea level) and δ_c via δ_w can be applied in both directions: to infer the V_g history from marine $\delta^{18}O$ records using information on seawater temperature [e.g., *Lea et al.*, 2002], or, vice-versa, to reconstruct seawater-temperature changes from δ_c variations in conjunction with independent sea-level data [e.g., *Cutler et al.*, 2003]. The main problem in either approach, is to separate the contributions of δ_w and temperature to the total change in δ_c [*Duplessy et al.*, 2002]. Other difficulties are due to climatically-driven variations of δ_i and to hydrography, two factors which have the potential to decouple V_g and δ_w signals.

[3] In common paleoceanographic practice the effect of changing isotopic composition of ice is considered small [*Chappell and Shackleton*, 1986] and a linear relationship between V_g and δ_w (implying constant δ_i) is assumed in reconstructing sea-level variations from δ_c records [e.g., *Waelbroeck et al.*, 2002]. But the mean $\delta^{18}O$ of ice does change during the course of a glacial-interglacial cycle. For an ice sheet in the growth phase this is mainly because the isotopic composition of snow changes: from heavy values, corresponding to warm, low-elevation accumulation areas at the beginning, toward lighter values (due to increased Rayleigh fractionation of water vapor in the atmosphere), on colder, higher ice-sheet surfaces at later stages. During decay ablation dominates and at each moment ice with different isotopic composition is ablated, so the mean $\delta^{18}O$ of the ice sheet changes through time as well. Also, different ice sheets had different isotopic evolution, depending on their volume and the latitudinal area where they developed. *Mix and Ruddiman* [1984] (hereinafter referred to as *MR*) suggest that, due to changes of the mean $\delta^{18}O$ of ice sheets, δ_w may misrepresent the amplitude of the ice-volume signal by up to 30% and lag ice volume by 1 to 3 thousand years (kyr). If this holds true, linearly relating δ_w to ice volume might introduce a significant error in the stratigraphic and paleoclimatic interpretation of δ_c records. For example, as *Clark and Mix* [2002] note, the nonequilibrium isotopic composition of ice sheets might be partly responsible for the timing differences observed in marine $\delta^{18}O$ records between the last isotopic maximum (LIM, centered at ~18 kyr before present (BP)) and the LGM (which ended at ~19 kyr BP).

[4] Ice-sheet models incorporating oxygen-isotope transport represent a valuable tool for addressing these intricacies, providing information about δ_w independent of foraminiferal $\delta^{18}O$, as well as the possibility to analyze the relationship between the different quantities involved. Such models,, which are able to realistically simulate the time-evolving three-dimensional isotopic stratigraphy of the Greenland [e.g., *Clarke et al.*, 2005] and Antarctic [*Lhomme*, 2004] ice sheets, suggest contributions to the change of δ_w ($\Delta\delta_w$) at the LGM of 0.01‰ for Greenland, 0.11–0.16‰ for West Antarctica and −0.04‰ for East Antarctica [*Lhomme et al.*, 2005]. We developed a new numerical model for the North American Ice Sheet (NAIS) to investigate the isotopic composition of this ice sheet and its impact on the ocean isotopic enrichment during the last glacial cycle.

[5] With respect to the relationship between ice volume and seawater $\delta^{18}O$, volume variations of the polar ice sheets

[1]Department of Geosciences, University of Bremen, Bremen, Germany.
[2]Now at Institut des Sciences de l'Evolution, Universite Montpellier II, Montpellier, France.
[3]Department of Geosciences and Research Center Ocean Margins, University of Bremen, Bremen, Germany.
[4]Institute for Marine and Atmospheric Research Utrecht, Utrecht University, Utrecht, Netherlands.

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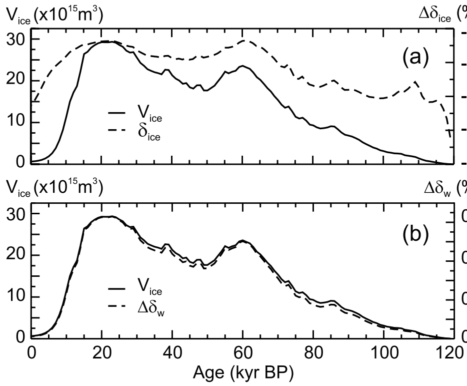

Figure 1. (a) Simulated volume history (solid) and mean ice $\delta^{18}O$ variations (dashed) for NAIS over the last glacial-interglacial cycle. (b) Simulated NAIS volume history (solid) and induced seawater isotopic enrichment $\Delta\delta_w$ (dashed).

seem indeed to have linearly affected δ_w [*Lhomme et al.*, 2005]. But at the LGM they accounted for only ~15% of the global ice-volume excess [cf. *Peltier*, 2004], and ~10% of the total ocean enrichment (approximately 1‰) [e.g., *Schrag et al.*, 2002]. We re-examined in detail the possible error induced by the assumption of a linear relationship between ice volume and $\Delta\delta_w$ from the point of view of the NAIS, the largest contributor to ice-volume variations during the late Pleistocene [cf. *Peltier*, 2004].

2. Model Description

[6] The 2.5-dimensional thermomechanical ice-sheet model [after *Gallée et al.*, 1992; *Payne*, 1995; *Crucifix and Berger*, 2002] simulates the evolution of the NAIS during the last glacial cycle (the past 120 kyr), at millennial to orbital (10^3 to 10^4 year) timescales. Non-linear ice flow is computed along a north-south flow line, whereas in the zonal direction a perfectly-plastic profile symmetric to the ice-sheet crest is assumed (see auxiliary material[1] for a detailed model description).

[7] Climate forcing is parameterized by the snow-line concept [*Oerlemans*, 1982] for mass balance and by the glacial index method [*Marshall et al.*, 2002] for surface temperature. Following this method, the Northern Hemisphere climatic state at every moment is described as a weighting between the LGM and present-day states, provided by a comprehensive atmospheric general circulation model (AGCM) [*Romanova et al.*, 2004], using the NGRIP $\delta^{18}O$ record [*North Greenland Ice Core Project Members*, 2004] as a weight function.

[8] The parameterization of oxygen-isotopic composition of snow combines a relationship between mean annual surface temperature and precipitation $\delta^{18}O$ for present-day Greenland [*Johnsen and White*, 1989] and results of AGCM simulations [*Jouzel et al.*, 1994] for the NAIS domain. The $\delta^{18}O$ of ice is treated as a passive tracer and, as well as temperature, is advected using a first-order upwind scheme. The mean ice $\delta^{18}O$ computed along the flow line is considered to represent the mean for the entire ice sheet. We assume an ocean with $\delta_w = 0$ at present day and calculate the changes of mean δ_w induced by NAIS volume variations as:

$$\Delta\delta_w = -\frac{S_i}{d_0 - S_g}\delta_i \tag{1}$$

where d_o is the present average depth of the ocean, S_i is the NAIS volume-equivalent sea level and S_g the global ice-equivalent sea level. NAIS volume V_{ice} and S_i are related by $S_i = \rho_i V_{ice}/(\rho_w A_o)$, with A_o the present ocean area and ρ_w and ρ_i the average densities of water and ice ($A_o = 360.5 \cdot 10^6$ km^2, $d_o = 3800$ m, $\rho_w = 1000$ kg/m^3 and $\rho_w = 910$ kg/m^3). No correction is made for variations in ocean area between the LGM and present. The global sea-level change is computed as $S_g = 1.7\, S_i$, by assuming a ~60% contribution of the NAIS to the total ice-volume variations [*Peltier*, 2004]. We also calculate the "isotopic volume" [cf. *MR*], V_{iso}, which is the ice volume linearly inferred from $\Delta\delta_w$, that is, by assuming a constant reference value δ_{ref} for the mean $\delta^{18}O$ of ice. V_{iso} is well approximated by:

$$V_{iso} = \frac{\delta_i}{\delta_{ref}} V_{ice} \tag{2}$$

and its departure from V_{ice} illustrates the misrepresentation of the real ice volume by $\Delta\delta_w$. In order to fit V_{ice} and V_{iso} values at the LGM (21 kyr BP), the computed LGM value of δ_i is chosen for δ_{ref}.

3. Results

[9] The simulated history of NAIS volume over the last glacial-interglacial cycle is shown in Figure 1a. The ice sheet approached its maximum volume at 25 kyr BP and then exhibited only small variations until 19 kyr BP, when the deglaciation started. At the LGM the volume was $29.3 \cdot 10^{15}$ m^3, corresponding to ~74 m of eustatic sea-level change. The northern ice margin remained always at 72°N; the southern margin reached 43.5°N at the LGM. The maximum ice thickness was ~3500 m, of which ~3000 m lay above the present-day sea level. In our simulation the deglaciation was not complete and an ice volume equivalent to 3 m of sea level was still left at present day.

[10] At the LGM most of the ice had a temperature between −33° and −15°C (Figure 2a). Values above −15°C were found at the base and at the southern margin of the ice sheet. About one third of the ice-sheet base was at the pressure melting point. The values for ice $\delta^{18}O$ ranged approximately between −35 and −27‰ (Figure 2b). Most of the basal ice had $\delta^{18}O$ values close to −30‰.

[11] The time evolution of δ_i (Figure 1a) showed a rapid decrease during glacial inception (from about −24‰ to −29‰ within 10 kyr), due to decreasing snow-$\delta^{18}O$ related to cooling climate and, more important, to increasing altitude of accumulation areas. After that, the ice sheet continued to grow, but less in height and more by extending toward lower latitudes, incorporating warmer, isotopically less-negative snow. Therefore, until the LGM, variations of only up to 3‰ in amplitude (approximately between −28‰ and −31‰) could be seen, modulated by changing ice volume. The lightest values (−31.3‰) were reached at

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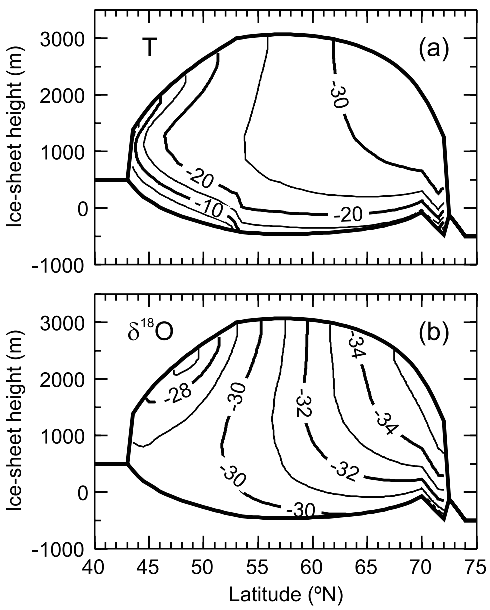

Figure 2. Distributions of (a) ice temperature (°C) and (b) δ^{18}O (‰) of the modeled NAIS along the flow line at the LGM.

60 kyr BP and at the LGM, concurrent with maximum values of ice volume. After the LGM, the ice mean δ^{18}O started to increase, initially rather slowly (by only 0.3‰ until 15 kyr BP), and more rapidly afterward. The ice left at the end of the simulation had a mean δ^{18}O of ~ -27‰.

[12] Changes of δ_w induced by NAIS-volume variations were as large as 0.63‰ at the LGM (Figure 1b). When scaled to fit the maximum and minimum values, $\Delta\delta_w$ and V_{ice} were in very good agreement: only small differences in magnitude were visible, mainly during ice build-up phases. As a consequence, the modeled ice volume V_{ice} and isotopic volume V_{iso} were also in very good agreement (Figure 3a). Again only a slight underestimation of V_{ice} by V_{iso} could be seen, especially during ice-sheet growth intervals. The differences between V_{ice} and V_{iso} were up to $0.8\times \cdot 10^{15}$ m^3 ($\sim$2 m equivalent sea level) with the largest values in the intervals 100 to 70 kyr BP and 50 to 30 kyr BP. These differences represent generally less than 10% of V_{ice} (Figure 3b). Somewhat larger errors, of 10–20%, occurred during the first 10 kyr of the glacial cycle (glacial inception) and in the second part of the deglaciation. During ice build-up the apparent (isotopic) ice volume lagged behind the true ice volume by up to 300–400 years, whereas during decay there was an even smaller lead of V_{iso} over V_{ice} by up to 100–200 years. The steeper the transition, the smaller the lead or lag. The deglaciation was steep enough to practically eliminate the phase difference. The local minima and maxima of V_{ice} and V_{iso} were reached at almost the same time (within less than 100 years).

4. Discussion

[13] Model parameters were chosen to match the NAIS contribution (S_i = 74 m) to the LGM eustatic sea-level drop (S_g = 126 m) of *Peltier* [2004]. The thickness and southern extent of the NAIS at the LGM, as known from reconstructions [e.g., *Dyke et al.*, 2002], were reasonably reproduced. Due to only diagnostic computation in the zonal direction, no distinction was made between the Laurentide and Cordilleran ice sheets, and the presence of ice in Alaska was implied. Since the northern margin of the simulated ice sheet remained at 72°N, the Innuitian ice sheet was not represented. However, as we were mainly interested in the total volume of the NAIS, the model served our purpose very well, and its shortcomings were compensated by computational efficiency.

[14] The modeled history of NAIS volume over the last glacial-interglacial cycle (Figure 1a) closely resembles the simulation with a three-dimensional ice-sheet model by *Marshall et al.* [2002]. The amount of ice left at present day is small compared to glacial-to-interglacial ice-volume change and therefore does not affect our conclusions. We note that the nonlinear physics were not really necessary in producing the 100-kyr cycle, which was contained in the climate forcing (based on the NGRIP δ^{18}O record), but were indispensable in realistically simulating the ice flow and, hence, the tracer transport.

[15] In order to infer the mean δ^{18}O for the ice sheet starting from the profile at the flow line, suppositions on the latitudinal distribution of ice δ^{18}O were needed. Based on the fact that snow falling at ice-sheet crest has the longest residence time as ice, we chose to attribute the mean along the flow line to the mean for the entire ice sheet. Actually, the latter would be less negative, because on every latitude, accumulated snow is isotopically more depleted at the ice divide than at any other point. This may counterbalance the possible underestimation of S_i at the LGM in the $\Delta\delta_w$ calculation (equation 1) by choosing the *Peltier* [2004] value of 74 m [e.g., *Marshall et al.*, 2002]. Using a different value for S_g, for example, 130 m [*Lambeck et al.*, 2002], would change $\Delta\delta_w$ by only 0.001‰.

[16] The parameterization of δ^{18}O of precipitation is critical for the computation of δ_i, and is difficult to evaluate, because little direct evidence exists for the isotopic composition of paleo-precipitations and of the now-vanished NAIS. Our parameterization resulted in LGM values for δ^{18}O of basal ice (-30‰) and for snow accumulating at the southern ice-sheet margin (-14‰ to -18‰) consistent with those obtained for the Laurentide ice from subglacial calcite and early diagenetic concretions from glacio-lacustrine deposits [*Hillaire-Marcel et al.*, 1979; *Hillaire-Marcel and Causse*, 1989]. The range of values for the computed mean isotopic composition of the NAIS (-24 to -31‰) is very close to that of *Bintanja et al.* [2005] (-25 to -32‰)

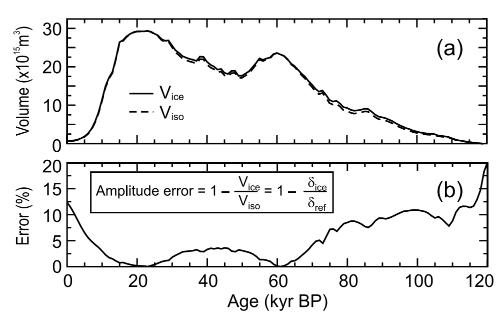

Figure 3. (a) Modeled ice volume V_{ice} (solid) and isotopic volume V_{iso} (dashed). (b) Amplitude misrepresentation of V_{ice} by V_{iso} (relative error).

and the LGM value (−31.3‰) is in agreement with the −31 ± 3‰ estimation of *Duplessy et al.* [2002]. Using a constant slope in the dependence of precipitation δ^{18}O on temperature, in the range between the present-day and the LGM slopes, instead of a function of the glacial index (see auxiliary material), brings relatively little change in our LGM δ_i (values between −30 and −31.4‰) and the corresponding $\Delta\delta_w$ (0.60 to 0.63‰).

[17] The simulated distributions of temperature and δ^{18}O (Figure 2) are artificially smoothed because the first-order upwind scheme that we used for advection in the ice-sheet model introduces numerical diffusion. The few other studies involving oxygen-isotope distribution in ice, aiming at a detailed reconstruction of δ^{18}O distribution of the ice sheets still present, Greenland [e.g., *Tarasov and Peltier*, 2003; *Clarke et al.*, 2005] and Antarctica [*Lhomme et al.*, 2005], employ three-dimensional thermomechanical models combined with high-resolution semi-Lagrangean tracer transport schemes, and are validated by comparing modeled and observed profiles at ice-coring sites. In contrast, little information is available on the isotopic composition of the NAIS, and the results of our ice-sheet model are in reasonable agreement with it. Tracer transport in our model can be improved by increasing the vertical resolution (Figure S1), but this does not significantly affect the result for the mean δ^{18}O of the ice sheet (Figure S2).

[18] As *MR* have shown using simple zero-dimensional models, the nonequilibrium isotopic composition of the ice sheet has theoretically the potential to decouple real and apparent (isotopic) ice-volume signals. If δ_{ref} is chosen to be equal to the LGM value of δ_i, then V_{iso} fits V_{ice} at the LGM, but underestimates it at any other moment of the glacial cycle. The misrepresentation of V_{ice} by V_{iso} is determined by the departure of δ_i from δ_{ref} (cf. equation 2), which explains why the largest errors (here 10–20%) occurred in the beginning and at the end of the glacial period (Figure 3b). This results in apparently slower ice build-up and faster decay, in agreement with some of the *MR* simulations. The phase difference decreases with increasing slope of the transitions, practically vanishing during deglaciation. While in MR's "best cases" V_{iso} still lags V_{ice} by 500–1000 years, in our simulation the phase differences were even smaller. Also, V_{iso} and V_{ice} reached all their minimum and maximum values almost simultaneously, which is important for the interpretation of δ_c records with respect to ice volume.

[19] One reason for the difference between MR's results and ours is the higher variability of V_{ice}, induced by the NGRIP forcing and amplified by the thermomechanical coupling. In MR's simulations V_{ice} is prescribed to be constant during a long period (10 kyr), while V_{iso} continues to change due to varying δ_i only, which leads to significant decoupling between the two signals. In our case there were no such long intervals of even nearly constant ice volume. Another reason is our parameterization for the isotopic composition of snow, which resulted in a contrast of only ~20‰ between δ_i at minimum and maximum volume and, therefore, in a smaller underestimation of V_{ice} by V_{iso}. We consider the good agreement between V_{ice} and V_{iso}, better then suggested by *MR*, as a valid result, because it is based on (a) a realistic ice-volume history instead of prescribed idealized sinusoidal transitions and (b) a parameterization for snow δ^{18}O consistent with the current knowledge on the isotopic composition of the Laurentide ice sheet. Based on a different (inverse) modeling setup, *Bintanja et al.* [2005] suggested that ice-volume variations dominate the δ_w component of the marine isotope signal. Our findings on the NAIS, when combined with those on Antarctic and Greenland ice sheets [*Lhomme et al.*, 2005], agree with that inference. Our results also indicate that the changing ice δ^{18}O is not among the main factors responsible for the observed difference between the last isotopic and ice-volume maxima.

[20] The relationship between ice-volume variations and the mean isotopic enrichment of the ocean is only one – proven to be small – uncertainty in relating ice volume to calcite δ^{18}O (δ_c) in marine sediments, via the seawater δ^{18}O. The biggest challenge in interpreting particular marine records remains to distinguish between the local contributions of temperature and δ_w. Therefore, further studies need to take into account the ocean circulation, which may induce phase differences between a freshwater signal due to ice-volume variations and the recording of that signal at different locations and depths in the ocean [*Sima*, 2005]. Also, at the interface between ice-sheets and the ocean, the routing of meltwater input must be considered, because, at least in the case of the NAIS, it changes dramatically [e.g., *Licciardi et al.*, 1999].

5. Conclusions

[21] Our 2.5-dimensional thermomechanical ice-sheet model provided essential insight in the oxygen-isotopic composition of the North American Ice Sheet (NAIS). It allowed us to make an estimation of the NAIS contribution to the ocean isotopic enrichment between the Last Glacial Maximum (LGM) and the Holocene, and reassess the time-evolving relationship between ice-volume variations and changes in the mean oxygen-isotope composition of seawater ($\Delta\delta_w$), first investigated by *Mix and Ruddiman* [1984]. The mean ice isotopic composition of the NAIS was found to vary between approximately −24 and −31‰. The simulated LGM value of −31.3‰ corresponded to a change of the mean seawater isotopic composition of 0.63‰. A comparison between NAIS-volume variations and the corresponding $\Delta\delta_w$ showed practically no time lag between the two signals and only small differences in magnitude, mainly during times of ice-sheet growth. As NAIS has been the largest contributor to $\Delta\delta_w$ during the late Pleistocene, we suggest that the nonequilibrium isotopic composition of ice sheets induces no significant phase difference between global ice volume and the mean isotopic enrichment of the ocean. Linearly translating $\Delta\delta_w$ to ice volume results in an underestimation of true volume by 10–20% during glacial inception and the second part of the deglaciation, and generally below 10% in the rest of the glacial cycle.

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J. Oerlemans, Institute for Marine and Atmospheric research Utrecht, Utrecht University, Princetonplein 5, NL-3508 TA Utrecht, Netherlands. (j.oerlemans@phys.uu.nl)

A. Paul and M. Schulz, Department of Geosciences and Research Center Ocean Margins, University of Bremen, Postfach 330440, D-28334 Bremen, Germany. (apau@palmod.uni-bremen.de; mschulz@palmod.uni-bremen.de)

A. Sima, Institut des Sciences de l'Evolution, Universite Montpellier II, Pl. E Bataillon, Cc 61, F-34095 Montpellier cedex 5, France. (adriana.sima@gmail.com)