Global budget of water isotopes inferred from polar ice sheets

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Received 9 June 2005; revised 13 September 2005; accepted 19 September 2005; published 27 October 2005.

[1] Water isotope ratios in ice cores and marine sediments are a key indicator of past temperature and global ice volume. Quantitative interpretation of these ratios requires understanding of the storage capacity and exchanges among the ocean, atmosphere, and cryosphere. We combine numerical models of ice dynamics and tracer transport to predict bulk ice properties by simulating the fine layering of ice sheets locally validated at ice core sites. The $^{18}O/^{16}O$ content of ice sheets is found to vary between the present and 20 kyr ago from −34‰ to −37‰ for Greenland, from −41‰ to −42.5‰ for West Antarctica, and always remained near −56.5‰ for East Antarctica. Their combined effect on sea-water $^{18}O/^{16}O$ is a 0.08–0.12‰ increase 20 kyr ago, a 1.11‰ decrease if ice sheets were to vanish. We confirm that ice volume changes in Antarctica and Greenland linearly affect ocean composition, though at different rates. **Citation:** Lhomme, N., G. K. C. Clarke, and C. Ritz (2005), Global budget of water isotopes inferred from polar ice sheets, *Geophys. Res. Lett.*, *32*, L20502, doi:10.1029/2005GL023774.

1. Introduction

[2] The isotopic composition of carbonate shells in deep-sea sediments has been widely used for reconstructing climate history [*Zachos et al.*, 2001]. Interpretation of these sedimentary records is complicated by the fact that $\delta^{18}O$ reflects the combined influence of ocean temperature and global sea-water $^{18}O/^{16}O$ composition ($\delta^{18}O_{sw}$) [*Shackleton*, 1974; *Labeyrie et al.*, 1987]. Distinguishing the effects of temperature and $\delta^{18}O_{sw}$ remains a long-standing problem in paleoceanography [*Duplessy et al.*, 2002], largely because of uncertainties concerning the amount of ^{18}O-depleted water stored in continental ice masses. The consequences are especially serious for evaluating temperatures at the Last Glacial Maximum (LGM) [*Dansgaard and Taubert*, 1969; *Waelbroeck et al.*, 2002], when ice volume peaked. Previous attempts to gauge the isotopic content of polar reservoirs have assumed that their bulk composition is proportional to that of single ice cores, from Camp Century for Greenland [*Dansgaard and Taubert*, 1969] and Vostok for Antarctica [*Duplessy et al.*, 2002].

[3] The scaling premise that underlies these estimates ignores the facts that isotopic deposition rates and patterns vary spatially and temporally and that isotopic signatures in snowfall are redistributed by ice flow. In this paper we address these complexities using numerical three-dimensional Ice Sheet Models (ISM) for Greenland [*Marshall and Cuffey*, 2000] and Antarctica [*Ritz et al.*, 2001]. We incorporate physical models of isotopic deposition and tracer transport [*Clarke and Marshall*, 2002; *Lhomme*, 2004; *Clarke et al.*, 2005] to predict the fine-scale layering of ice sheets and identify plausible climatic and dynamic reconstructions that simulate ice sheets with geometric features (volume, areal extent, surface elevation) and ice layering that are in close agreement with present-day observations and major ice core records. These conditions allow us then to calculate the bulk composition of polar ice masses.

2. Fine Stratigraphy of Polar Ice Sheets

[4] The climate forcing applied to the ISMs prescribes surface temperature, accumulation and ablation rates as functions of time and space. We use standard parameterizations for all three components [*Ritz et al.*, 2001; *Clarke and Marshall*, 2002; *Huybrechts*, 2002]. Specifically, past surface temperature $T(x, y, t)$ and precipitation $P(x, y, t)$ are obtained by scaling their present distributions (t = 0) according to a history of temperature change ΔT_c inferred from an ice core record representative of each ice sheet. For Greenland, the methodology with the GRIP–Vostok $\delta^{18}O$ splice adopted by *Cuffey and Marshall* [2000] and subsequent studies is applied for 320 kyr of climate history, extracting ΔT_c from

$$\Delta\delta^{18}O(t) = \alpha_c \Delta T_c(t) + \beta_\delta \Delta Z(t) \qquad (1)$$

with α_c the isotopic sensitivity or "temporal slope" [*Jouzel et al.*, 1997] set for a glacial–interglacial change of 22–24°C [*Cuffey et al.*, 1995; *Dahl-Jensen et al.*, 1998], β_δ the isotopic lapse rate (β_δ = −6.2‰ km^{-1} in central Greenland) and $\Delta Z(t)$ the elevation change at the site. For Antarctica we take 700 kyr of temperature history inferred from EPICA-Dome C (EDC) [*EPICA Community Members*, 2004] with a glacial–interglacial amplitude of 10–12°C. Past surface temperature is then redistributed with an ice-sheet-specific temperature lapse rate β_T: $\Delta T(x, y, t) = \Delta T_c + \beta_T \Delta Z(x, y, t)$. Saturation vapour pressure governs precipitation through

$$P(x, y, t) = P(x, y, 0)\exp[D_{acc}(t)\Delta T(x, y, t)] \qquad (2)$$

with a precipitation sensitivity $D_{acc}(t)$ inferred from ice cores (0.068–0.085 °C^{-1}). Ablation is derived from the

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0094-8276/05/2005GL023774

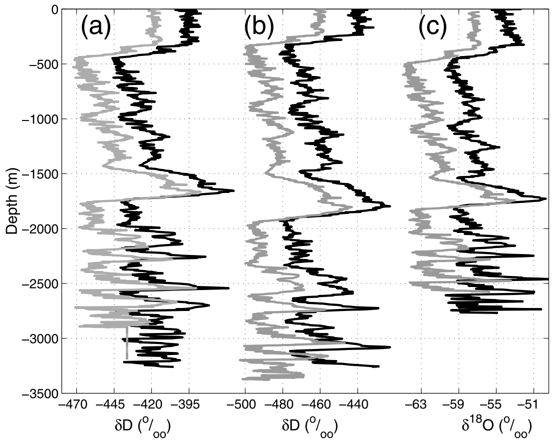

Figure 1. Comparison of simulated (dashed, left) and recorded (solid, right) ice cores in Antarctica, with offset for clarity of display. (a) EPICA-Dome C. (b) Vostok (longer modeled δ^{18}O-core than the actual record because the influence of Lake Vostok is not included in our model). (c) Dome Fuji.

standard Positive-Degree-Day method. Grounding line migration is controlled by ice flow and changes in sea level, here a prescribed external forcing due to the predominant influence of Northern Hemisphere ice sheets [*Ritz et al.*, 2001].

[5] The three-dimensional time-evolving layering of ice in Greenland, East Antarctica and West Antarctica is obtained by using the numerical ISMs to track ice particles having δ^{18}O and depositional age labels. Applying a specific adjustment for the surface mass balance history of an ice sheet [*Lhomme*, 2004], the method has shown great success at predicting the fine stratigraphy of the Greenland Ice Sheet (GIS) and its ice core records (GRIP, GISP2, Dye 3, Camp Century), proving that the climate derived from the single GRIP record encapsulates most of Greenland variability and validating our method [*Clarke et al.*, 2005; *Lhomme et al.*, 2005]. Similar performance is here achieved for Antarctica, as evidenced in Figure 1 where the observed and predicted isotopic variations and age–depth profiles at the East Antarctic Ice Sheet (EAIS) sites of EDC, Vostok [*Petit et al.*, 1999] and Dome Fuji [*Watanabe et al.*, 2003] fit remarkably. These good fits are obtained for both the 10 and 12°C-climate forcings, which we take as lower and upper bounds on climate history in subsequent calculations. This success with a simple climate parameterization and a full ice sheet model confirms the "homogeneous climate variability across East Antarctica" noted by *Watanabe et al.* [2003] and validates our working hypothesis and previous ice sheet studies relying on a simple climate forcing for all of Antarctica [*Ritz et al.*, 2001; *Huybrechts*, 2002]. Comparison with ice core records from West Antarctica is not attempted here because the ISM does not adequately resolve the particular topography and ice flow at the core sites. Nonetheless, to estimate the bulk composition of the Antarctic Ice Sheet, such limitations play a negligible role because (1) the climate over the depositional areas and the most glaciated parts of Antarctica is confirmed by the observed similarities in the isotopic records of EDC–Dome Fuji with the inland West Antarctica site of Byrd [*Watanabe et al.*, 2003], (2) the accuracy of the ice evolution and tracking schemes has been demonstrated [*Ritz et al.*, 2001; *Lhomme*, 2004], and (3) relative to EAIS, WAIS makes a small contribution to the ice volume (~15% at present) and bulk composition of the entire ice sheet.

3. Bulk Water Isotopic Composition

[6] Estimates of the water-isotope composition of polar ice sheets rely on a priori knowledge of the past surface deposition of ^{18}O/^{16}O. Because this information is only available at ice core sites, we developed a physical model of spatial and temporal isotopic deposition. For Antarctica, atmospheric models that include isotopic fractionation processes [*Delaygue et al.*, 2000] suggest that the present relationship between surface temperature and δ^{18}O remained valid in the past; thus we assume that the temporal slope α_c is constant and spatially uniform. In this manner, past surface δ^{18}O is derived from its present spatial distribution [*Giovinetto and Zwally*, 1997] after correction for changes in global temperature ΔT_c (modulated by α_c) and surface elevation ΔZ, as in (1). The same principle applies to Greenland, though with more complexity. As evidenced by borehole thermometer measurements at GRIP [*Dahl-Jensen et al.*, 1998] and GISP2 [*Cuffey et al.*, 1995] and confirmed by atmospheric models [*Jouzel et al.*, 2000], changes in the seasonality and origin of precipitation during glacial periods modify the relationship between surface temperature and δ^{18}O. Thus α_c varies spatially and temporally. To assess the uncertainty associated with both phenomena, past surface δ^{18}O is obtained from the present distribution [*Zwally and Giovinetto*, 1997] by using either a spatially uniform α_c derived from GRIP/GISP2 or regional values of α_c predicted by the GISS model [*Jouzel et al.*, 2000]. Thus we apply different climate forcings and isotopic sensitivities to compute an average volume and bulk isotopic composition of polar ice sheets over the past 160 kyr.

[7] East Antarctica lost up to 5% of its ice volume during glacial periods (Figure 2a) because of reduced precipitation (down by 50%), implying the equivalent of a 3 m rise of sea level while decreasing the average $\delta^{18}O_{sw}$ composition of sea water by at most 0.04‰ (Figure 2d). Negligible variations of EAIS volume imply that its average isotopic composition remained almost constant near an average of −56.5‰. Conversely, ice volume varied significantly for WAIS, which benefited from the glacial lowering of sea level to over-ride its shallow continental shelf. This further pulled down sea level by an additional 8–10 m at 12 kyr BP (Figure 3), a result on the lower side of the accepted range [*Ritz et al.*, 2001; *Huybrechts*, 2002]. The isotopic composition reflects, with a short lag, the variation of volume and temperature: from −41.1 ± 0.7‰ at present to a minimum of −42.6 ± 1.0‰ at 12 kyr BP, when it would have increased $\delta^{18}O_{sw}$ by 0.11–0.16‰. The uncertainty for WAIS could be diminished with finer treatment of climate (regional features) and dynamics (spatial resolution, hydrology, bedrock topography, geothermal activity, ice streams). In total, the isotopic composition of Antarctic ice remained near −52‰, well within the conservative −40 to −60‰ range from previous estimates [*Duplessy et al.*, 2002]. We note that EAIS and WAIS reacted to the last glacial

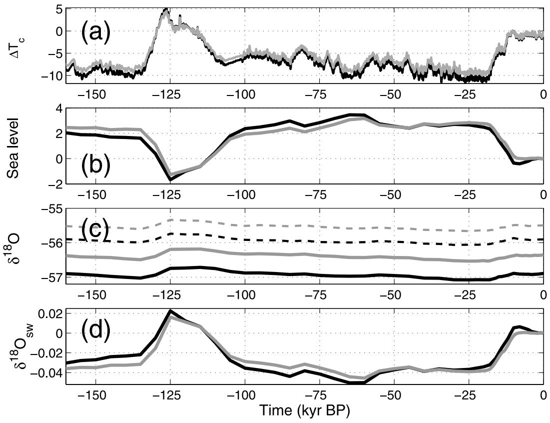

Figure 2. East Antarctic climate history and bulk properties. Tests with glacial–interglacial amplitude of 10°C (gray) or 12°C (black), isotopic sensitivity α_c = 0.80‰ °C^{-1} (solid) [*Petit et al.*, 1999] or α_c = 0.60‰ °C^{-1} (dash), $\beta_\delta = -11.2$‰ km^{-1}. (a) Climate forcing ΔT_c (°C). (b) Effect on sea level (m), assuming that 1 Mkm^3 of ice corresponds to ~2.5 m of sea level [*Huybrechts*, 2002]. (c) Bulk δ^{18}O composition of ice sheet (‰). (d) Effect of the ice sheet on $\delta^{18}O_{sw}$ (‰), assuming that present ocean mass is 1.37 × 10^{21} kg and that ocean area remained constant at 3.61 × 10^8 km^2. Equivalent δD can be obtained via the water meteoric relationship $\delta D = 8\delta^{18}O + 10$.

termination on different time-scales because of the high sensitivity of WAIS to sea level change, controlled by the slow Northern Hemisphere deglaciation.

[8] In Greenland (Figure 4), large changes in temperature during the last glacial cycle only had a moderate effect on ice volume. The average δ^{18}O composition varied between a minimum of −37.1‰ at LGM (±0.8‰ error from α_c) and the present −34.2 ± 0.3‰. Previous estimates suggest −41.5‰ and −36‰ at these respective times [*Duplessy et al.*, 2002; *Dansgaard and Taubert*, 1969]. Small changes in ice volume imply a negligible effect on $\delta^{18}O_{sw}$: at most +0.01‰ at LGM and possibly as low as −0.03‰ during the last interglacial. For the latter period, we refrain from adding up the Antarctic and Greenland contribution to sea level because the climatic forcings for the two ice sheets are not synchronized yet and we lack constraints to properly model WAIS for that time.

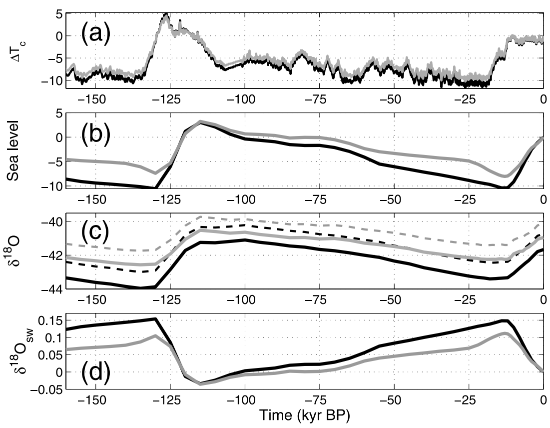

Figure 3. West Antarctic climatic history and bulk properties. (a)–(d) Graphs as in Figure 2.

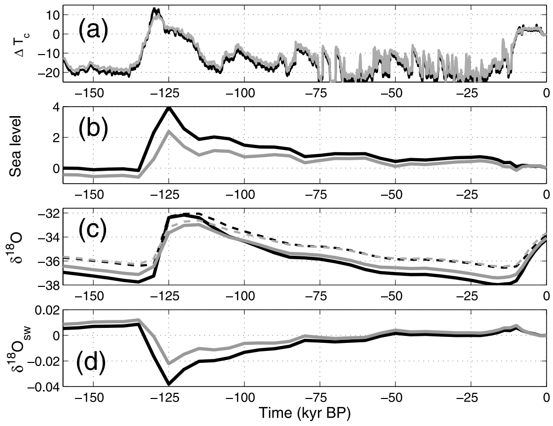

Figure 4. Greenland climatic history and bulk properties. Tests with 2 m (gray) and 4 m (black) effect on sea level at Eemian [*Lhomme et al.*, 2005], α_c from GRIP/GISP2 (α_c (glacial) ≈ 0.33‰ °C^{-1}, dash) or spatially varying (α_c (glacial) = 0.27–0.50‰ °C^{-1}, solid). **a**, **b**, **c**, **d** as in Figure 2.

[9] For the interpretation of ice and marine records, the relationship between ice volume and $\delta^{18}O_{sw}$ is generally assumed to be linear, with rates of 0.085–0.011‰ per metre change in sea level estimated from corals and deep-sea sediments [*Fairbanks*, 1989; *Schrag et al.*, 1996; *Waelbroeck et al.*, 2002]. We find that, when separately considered, there is a high degree of correlation between $\delta^{18}O_{sw}$ and sea-level change for each ice sheet over the past 160 kyr: 0.016, 0.015 and 0.010‰ m^{-1} for EAIS, WAIS and GIS, respectively. Complete melting of these ice sheets would decrease $\delta^{18}O_{sw}$ by 0.91, 0.13 and 0.07‰, respectively, yielding a total effect that is consistent with the ~1‰-step increase in δ^{18}O recorded in deep-sea sediments when, 34 million years ago [*Zachos et al.*, 2001], a full-scale Antarctic Ice Sheet developed and persisted. Using the standard water meteoric relationship, all the above results can be extended to yield estimates of D/H. The joint contribution of polar ice sheets of 1.11 ± 0.03‰ δ^{18}O and −8.88 ± 0.24‰ δD closes the global water-isotope budget and gives the ocean composition of an ice-free world.

4. Conclusion

[10] These estimates of polar ice isotopic composition are the first to use a framework that is consistent with most ice core records and the state-of-the-art in ice sheet modeling. Predictions of Greenland glacial composition (−37‰) establish an absolute lower bound on the past composition of the Laurentide and Eurasian ice sheets, because they formed at low elevation and thus were richer in ^{18}O; detailed studies are in progress. Finally, our method can also provide the present average age of ice in the contemporary polar ice masses: 41 ± 2 kyr for that of Greenland, 125 ± 5 kyr for East Antarctica and 44 ± 3 kyr for West Antarctica.

[11] **Acknowledgments.** This is a contribution to the European Project for Ice Coring in Antarctica (EPICA), a joint European Science Foundation/European Commission (EC) scientific programme, funded by the EC and contributions from Belgium, Denmark, France, Germany, Italy,

the Netherlands, Norway, Sweden, Switzerland and the United Kingdom. This is EPICA publication no. 130. The numerical model was run on the computer facilities of MIRAGE at Université Joseph Fourier in Grenoble, France. Ice core data were obtained from the WDC-A for Paleoclimatology. N.L. acknowledges support by the UBC Graduate Fellowship, the French program ECLIPSE and Région Rhône-Alpes' program EURODOC. G.C. acknowledges support from the Natural Sciences and Engineering Research Council of Canada. The authors thank F. Parrenin for early access to the EDC2 climate forcing and Eric Steig for his review.

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