Quantifying ice-sheet feedbacks during the last glacial inception

M. Kageyama, S. Charbit, C. Ritz, M. Khodri, Gilles Ramstein

Quantifying ice-sheet feedbacks during the last glacial inception

M. Kageyama,[1] S. Charbit,[1] C. Ritz,[2] M. Khodri,[3,4] and G. Ramstein[1]

Received 25 August 2004; revised 20 October 2004; accepted 4 November 2004; published 18 December 2004.

[1] The last glacial inception (~116 ky ago) has long been used to test the sensitivity of climate models to insolation. From these simulations, atmospheric, oceanic and vegetation feedbacks have been shown to amplify the initial insolation signal into a rapid growth of ice-sheets over the northern hemisphere. However, due to the lack of comprehensive atmosphere-ocean-vegetation-northern hemisphere ice-sheet models, the impact of all these feedbacks acting concurrently has not yet been evaluated. Here we present the results from such a model, which simulates significant ice-sheet growth over North America, but none over Eurasia. Our analyses focus on the different behaviours over these regions, and the quantification of the ice-sheet feedbacks on climate. *INDEX TERMS:* 1827 Hydrology: Glaciology (1863); 3322 Meteorology and Atmospheric Dynamics: Land/atmosphere interactions; 3337 Meteorology and Atmospheric Dynamics: Numerical modeling and data assimilation; 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology. **Citation:** Kageyama, M., S. Charbit, C. Ritz, M. Khodri, and G. Ramstein (2004), Quantifying ice-sheet feedbacks during the last glacial inception, *Geophys. Res. Lett.*, *31*, L24203, doi:10.1029/2004GL021339.

1. Introduction

[2] The simulation of the last glacial inception (~116 ky ago) has long constituted a test for climate models (for a review, see *Vettoretti and Peltier* [2004]). After the failure of most atmospheric general circulation models (AGCMs) to simulate perennial snow cover under 116 or 115 ky BP insolation forcing, vegetation feedbacks —via a replacement of high-latitude taiga by tundra and its impact on snow albedo— have been shown to amplify the climatic response to the insolation forcing [e.g., *DeNoblet et al.*, 1996]. In addition, *Khodri et al.* [2001], using a coupled atmosphere-ocean (AO) GCM, have shown that the 115 ky BP insolation forcing was favourable to cold sea-surface temperatures (SST) and increased sea-ice cover at high northern latitudes. This high latitude cooling results in an amplified meridional temperature gradient between low and high northern latitudes, which yields enhanced moisture fluxes to the Arctic regions and help to maintain a perennial snow cover. *Vettoretti and Peltier* [2003] also obtained amplified moisture fluxes as a response to Arctic perennial snow in an AGCM experiment.

[3] Climate models of intermediate complexity (EMICS) are useful tools for the investigation of climate change on the Milankovitch timescale because they can be run for millenia. For instance, *Gallée et al.* [1992] successfully simulated the last 130 ky climate-ice-sheet evolution using a 2D EMIC forced by insolation and CO_2 only. They showed that variations in ablation are more important in driving the ice-sheet response than variations in accumulation, and that the ice-sheet altitude effect is stronger than the ice-sheet extent (albedo) feedback. Although the representation of the vegetation (tundra-taiga transition) and oceanic (slab-ocean) components are very simple in this model, these simulations indicate that the vegetation and sea-ice feedbacks are necessary for the simulation of the glacial-interglacial cycle. Other studies have focussed on the inception. *Khodri et al.* [2003] use the AO CLIMBER2.3 model [*Petoukhov et al.*, 2000] forced by the 126 to 115 ky BP insolation and a preindustrial 280 ppm CO_2, to show that the insolation forcing favours a cool state (North Altantic SSTs 0.8°C cooler and more extensive Arctic sea-ice) as early as 120 ky BP, in line with the SST cooling (2.2°C) suggested by *Cortijo et al.* [1999]. *Crucifix and Loutre* [2002], using the MoBidiC atmosphere-ocean-vegetation (AOV) model, show how vegetation, snow cover and sea-ice work in synergy to produce inception conditions. *Wang and Mysak* [2002] use an AO EMIC coupled to a 2D ISM to study the 122–110 ky BP period and simulate glacial inception over North America (mainly over Alaska and the eastern Canada) and Fennoscandia. They emphasize the role of the ice-sheet altitude and refreezing of rain and meltwater at the ice-sheet surface. Their total ice volume at 110 ky BP is $\sim 15 \times 10^6$ km^3, i.e., 30 m drop in sea-level, 2/3 of the observed sea-level drop [*Waelbroeck et al.*, 2002]. They also suggest an active role for the thermohaline circulation, which strengthens during ice-sheet growth due to the deficit in fresh water delivered to the ocean.

[4] To date, all the important components (i.e., atmosphere, full ocean, vegetation, ice-sheets) thought to play a role in the inception have not been studied within one single model. Here we present results from an atmosphere-ocean-vegetation-northern hemisphere ice-sheet model (AOV-NHIS), developed by coupling the AOV model CLIMBER2.3 and the 3D ISM GREMLINS (GRenoble Model for Land Ice in the Northern hemiSphere) [*Ritz et al.*, 1997]. Our analyses focus on the processes which augment or prevent ice-sheet growth and on the feedbacks of these ice-sheets on the AOV system.

2. Model Description

[5] The model consists of the CLIMBER2.3 AOV model coupled to the GREMLINS ISM. CLIMBER includes

[1]Laboratoire des Sciences du Climat et l'Environnement, Institut Pierre Simon Laplace, Centre National de la Recherche Scientifique, Gif-sur-Yvette, France.

[2]Laboratoire de Glaciologie et Géophysique de l'Environnement, Saint-Martin-d'Hères, France.

[3]Lamont-Doherty Earth Observatory, Columbia University, New York, USA.

[4]Now at Geophysical Fluid Dynamics Laboratory, U.S. Department of Commerce, National Oceanic and Atmospheric Administration, Princeton, New Jersey, USA.

0094-8276/04/2004GL021339$05.00

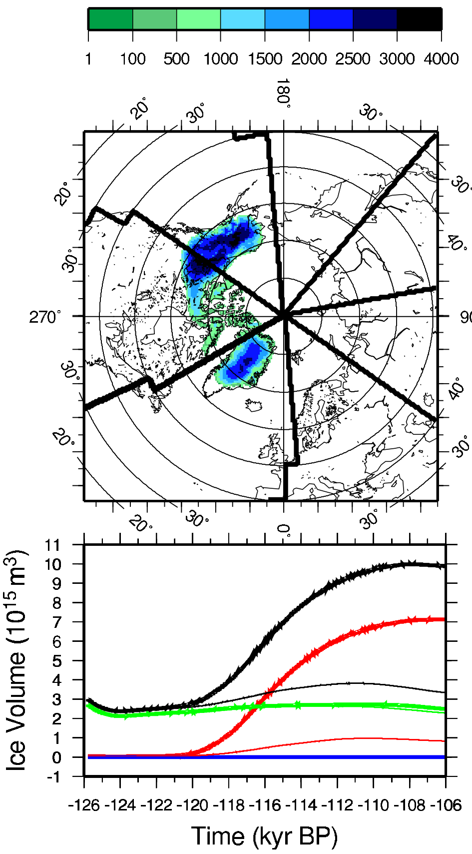

Figure 1. a. Final ice-sheet thickness (in m) in the ice-sheet model on the ISM grid (231 × 241 points) and limits of the CLIMBER sectors (thick lines in longitude, thin parallels for latitudinal boundaries); b. evolution of the volume (in Pm^3) of the main NH ice-sheets simulated by the model forced by the insolation and CO_2 of the period 126–106 ky BP: total volume (black), Greenland (green), Laurentide (red), Fennoscandia (blue); thin lines are for experiment with vegetation fixed at its initial value.

simplified representations of the atmosphere and its hydrological cycle (resolution: ~51° in longitude, 10° in latitude), the ocean and the vegetation. GREMLINS is a 3D thermo-mechanical ISM (resolution: 45 × 45 km^2).

[6] The models exchange information every 200 years. Mean annual temperature (Tann), summer temperature (Tjja) and annual snowfall (Snw) are passed from CLIMBER to GREMLINS. Snw directly yields the accumulation rate over the ice-sheet. Tann and Tjja are used to compute an upper boundary condition for the temperature inside the ice, and an ablation rate via the positive-degree-day method [*Ritz et al.*, 1997]. This is similar to previous studies in which GREMLINS was forced by AGCMs [*Charbit et al.*, 2002], except that we use absolute values for temperature and snowfall, rather than anomalies or ratios w.r.t. a modern situation. The challenge in coupling the two models resides in the downscaling of variables computed on CLIMBER's coarse atmospheric grid onto GREMLINS's fine grid (Figure 1a): within one given CLIMBER gridbox, the hundreds of GREMLINS grid points can lie at altitudes between sea-level and a few thousand meters. We have developed a scheme that computes the surface radiative budget and associated variables, such as the snow fraction, for each CLIMBER grid box at both the grid box altitude and every 300 m from 0 to 4200 m (each level is assumed to be at the grid box altitude for the surface energy balance). These calculations are performed following the same code as the original CLIMBER radiative budget code. We obtain, for each CLIMBER grid box, a vertical profile of surface temperatures. Similarly, total precipitation profiles are computed for the same altitudes using the basic CLIMBER formulation. Finally, the snowfall profile is computed as a fraction of the total precipitation which depends on the surface temperature at a given level. Tann, Tjja and Snw are computed at sea-level over the open ocean and sea-ice CLIMBER surface types (the weighted average for these two types gives an oceanic value of the variable), and their vertical profile is computed for each CLIMBER land surface type: land-ice, forest, grass, desert (the weighted average for the last 3 types gives an ice-free continental value). The interpolation from the CLIMBER climatic fields on the GREMLINS fine grid is a trilinear interpolation, with a dependence on the GREMLINS surface type (ocean, land-ice and ice-free continents).

[7] The surface type fractions and altitudes computed by GREMLINS are averaged over each CLIMBER gridbox to provide new surface boundary conditions. The amount of water used to build the ice-sheets is subtracted from the runoff, and fresh water from ice-sheet melting is directed to the oceans which closes the water budget.

3. Simulation of the Last Inception: 126–106 ky BP

[8] Our baseline simulation (STD) is obtained by forcing our AOV-NHIS model with 126 to 106 ky BP orbital parameters [*Berger*, 1978] and CO_2 from the Vostok ice core [*Petit et al.*, 1999]. The initial state of the ice-sheet at 126 ky BP is the present situation, which is justified by sea-level data [*Waelbroeck et al.*, 2002]. The initial climate state is obtained by running the climate model in equilibrium with this ice-sheet for the orbital parameters of 126 ky BP for 5 ky.

[9] During the first 1800 y of the run (Figure 1b), the Greenland ice-sheet (GIS) volume decreases due to the insolation forcing (i.e., increased summer insolation. The GIS actually vanishes in 11800 y in a perpetual 126 ky BP insolation fully coupled run). At 124.2 ky BP, the GIS southern tip has melted, consistent with ISM studies [e.g., *Tarasov and Peltier*, 2003] that predict a sea level rise of a few m compared to the modern sea-level due to the GIS melting at the Eemian. This can be considered as the adjustement of the ice-sheet to Eemian conditions.

[10] From 126 ky BP, the 60°N North Atlantic SST decreases. By 120 ky BP it has cooled by 1.3°C compared to the initial conditions, 0.2°C more than with ice-sheets fixed at the initial state and 0.5°C more than with ice-sheets and vegetation fixed at their initial states and a constant 280 ppm CO_2 [*Khodri et al.*, 2003]. Thus, accounting for all the feedbacks in the AOV-NHIS system yields a better agreement with *Cortijo et al.* [1999]. Meanwhile, the 30°N SST decreases by 0.5°C, both with interactive ice-sheets and with ice-sheets fixed at their initial values, but 0.3°C more than with fixed ice-sheets and vegetation, and a CO_2

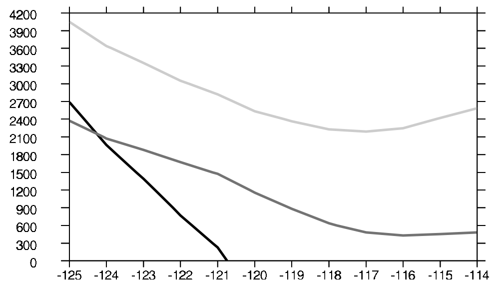

Figure 2. Highest (summer) altitude of the 0°C isotherms for the ice-free surface type between 126 and 114 ky BP. Canadian Archipelago in black, NW Rockies in dark grey, Fennoscandia in light grey.

of 280 ppm. This results in an increase in the meridional temperature gradient and in energy and moisture fluxes from the tropics to the extratropics as suggested by *Cortijo et al.* [1999].

[11] At 121 ky BP, two ice-sheets start growing over North America: one over the northwestern Rockies, and the other over the Canadian Archipelago. By 118 ky BP, all land in the Canadian Archipelago north of 70°N is ice-covered. At 117 ky BP, the two ice-sheets merge and by 113 ky BP ice extension in the 60–70°N American sector reaches its maximum, and final, extent. The ice thickness then grows, reaching 4000 m at 106 ky BP (Figure 1a), its volume being equivalent to a 17 m sea-level drop. The North American ice volume simulated at 110 ky is similar to that simulated by *Wang and Mysak* [2002], but we find no ice-sheet growth over Eurasia, in agreement with data [*Mangerud et al.*, 1979; *Turon*, 1984]. Processes helping or preventing ice growth are analysed in the following sections.

4. What Makes the NH Ice Sheets Grow?

[12] Figure 2 shows the evolution of the altitude of the summer 0°C isotherm in the temperature profiles, computed as described in section 2. It reaches typical surface altitudes over the Canadian Archipelago (<500 m) and NW Rockies (from 1000 to 2000 m) by 121 ky BP. However, the lowest summer 0°C isotherm altitude over northern Eurasia always remains above 2100 m, too high for an ice-sheet to start growing. Vegetation changes over America (forest gradually disappears before the inception) amplify the insolation-induced summer cooling. This does not occur over Eurasia because the initial climate is warmer, and vegetation is further away from the taiga-tundra threshold. Crossing this threshold is crucial for the inception: if vegetation is fixed at its initial interglacial value, inception does not occur (Figure 1b). Moreover, the 0°C isotherm then remains high enough (between 300 and 900 m higher than in the full AOV-NHIS run) for inception never to occur.

[13] To quantify the role of each variable passed from CLIMBER to GREMLINS, we ran sensitivity experiments in which one variable is kept fixed at its 122 ky BP value (Figure 3a) from 122 ky BP onwards. If Snw is kept fixed, the final North American ice volume is 25% lower than in STD. If Tann is kept fixed (with variations of ablation only linked to the seasonal cycle amplitude decrease), the ice volume is maximal at 112 ky BP, 39% lower than in STD. Lastly, if Tjja is kept fixed (with variations of ablation due to a smooth decrease of Tann, resulting in an amplification of the seasonal cycle but with maximum temperatures imposed at same level), no ice-sheet grows over North America, or anywhere else in the ISM domain. This shows that Tjja is the limiting factor for the initiation of glaciation, as expected from Milankovitch theory. Our fixed vegetation run shows that vegetation is one of the main factors explaining this sensitivity to Tjja. The other variables are important for the final ice volume, not only for its value (Snw and Tann), but also for its timing (Tann).

5. Feedbacks of the Ice-Sheets on Climate

[14] Once the ice-sheets start growing, at 121 ky BP, land-ice feedbacks come into play. The 1st feedback involves the albedo: as the ice-sheet extends, the albedo averaged over this land area increases, favouring a colder temperature. The 2nd feedback involves the altitude: an average altitude increase, given the usual temperature lapse rate, implies further surface cooling. The 3rd feedback is linked to the ice sheet build-up restricting the fresh water input to the adjacent oceans. Its impact could be positive [*Wang and Mysak*, 2002]: less fresh water to the ocean acts to strengthen the North Atlantic deep water formation, which favours moisture transport to the ice-sheets, or negative: as the thermohaline circulation strengthens, North Atlantic SSTs warm up and prevent the ice-sheets from growing. All these feedbacks directly operate in our model, via the variables that are passed from the ISM to CLIMBER: the ice-sheet fraction (for the albedo feedback), the average

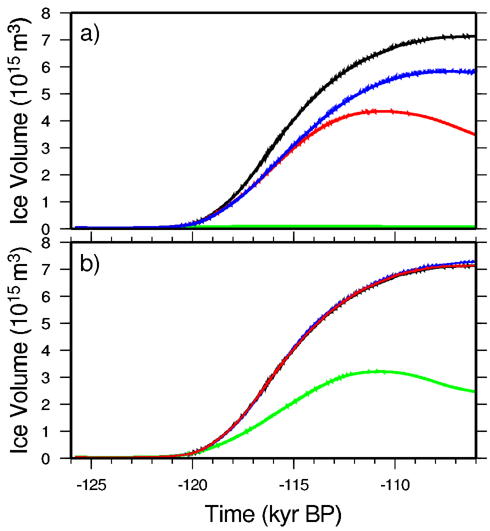

Figure 3. North American ice volume in 126–106 ky BP runs all forced by the same insolation and CO_2 as the STD run (in black), but with one coupling variable fixed after 122 ky BP. a) sensitivity to variables passed from CLIMBER to the ISM: green: fixed Tjja, red: fixed Tann, blue: fixed Snw; b) sensitivity to the variables passed from the ISM to CLIMBER: green: fixed glaciers' extent. The curves for STD (black), fixed glacier altitude (blue) and no correction in the fresh water going to the ocean (red) are superimposed.

altitude in a CLIMBER grid-box, and the runoff which is modified in relation to ice-sheet build up (or decay). We have studied the effect of the first 2 feedbacks by running the fully coupled model to 122 ky BP, fixing the value of each variable (in separate runs) from 122 ky BP onwards. The 3rd feedback's impact is evaluated in a run in which the runoff is not modified to take the ice volume change into account.

[15] If the albedo feedback is inhibited (Figure 3b), the maximum North American ice volume is reached at 111 ky BP and is 55% lower than in STD. Inhibiting the other two feedbacks makes no difference. Thus, the most important feedback in our model is the albedo feedback, contrary to the results from *Gallée et al.* [1992] and *Wang and Mysak* [2002], who found the altitude feedback to be crucial. The fact that, in our run, the thermohaline circulation is not sensitive to the fresh water deficit related to ice-sheet growth could be linked to it being smaller than in the work of *Wang and Mysak* [2002].

[16] In addition to these feedbacks, we have seen that ice-sheets, when they become large enough, can modify the atmospheric circulation and moisture transport. In our simulations, the moisture transport is modified via the response of the ocean and sea-ice to the insolation forcing [see *Khodri et al.*, 2003]. By 106 ky BP, the 30 to 50°N moisture and heat transports are slightly amplified in the STD run compared to a run uncoupled to any ice-sheet (by 0.01 Sv and 0.1 PW, respectively). Thus, the ice-sheet feedbacks operate via the local surface energy budget, but do not have a large impact on the atmospheric circulation or the ocean circulation. This could be a reason why the ice-sheets in our model do not build up very fast.

6. Discussion and Future Work

[17] We have developed an AOV-NHIS model capable of simulating the last glacial inception using only insolation and CO_2 forcings between 126 and 106 ky BP. Inception occurs over the Canadian Archipelago, in agreement with data, and over the NW Rockies, for which data is controversial (see discussion by *Wang and Mysak* [2002]). No ice appears over Eurasia. Therefore, compared to the study of *Khodri et al.* [2003], we show that it is very important to couple the climate model to an ISM to evaluate if the climatic variations simulated by CLIMBER indeed result in ice sheet inception. A large enough summer temperature cooling, as in the Milankovitch theory, is a necessary condition for inception in our model, but annual temperature and snowfall are important to set the timing of the maximum and its amplitude. Vegetation is necessary to get this strong enough cooling to initiate glaciation over northern North America. Then, once inception has started, the ice extent feedback accelerates ice-sheet growth. We find that altitude and fresh water feedbacks are not crucial.

[18] The main deficiency of our model is that the ice-sheet growth rate is not large enough compared to sea-level data. In this model version, there is no coupling to an Antarctic ISM. However, we believe most of the sea-level drop to be due to the NH ice-sheets build-up and that our model must be improved to obtain faster ice-sheet growth. Our next step is to investigate why the meridional atmospheric moisture flux does not increase as much as suggested by GCMs (as explained by *Khodri et al.* [2003]). Discrepancies in the precise location of ice-sheet inception could be linked to snowfall not being redistributed over mountain ranges as a function of the atmospheric circulation. We plan this development for the near future. In addition, future work will include studying the role of the ocean not only for ice-sheet inception, but also for the vegetation changes that are a pre-requisite to glacial inception.

[19] **Acknowledgments.** This work has been funded by the BIOCLIM European Project and the French national programme PNEDC IMPAIRS. The authors are endebted to Andrey Ganopolski and the CLIMBER team at PIK for providing their model. The two reviewers are thanked for their critical reading which helped improving the manuscript and Andrew Friend is thanked for his efficient corrections. LSCE contribution #1124.

References

Berger, A. L. (1978), Long-term variations of daily insolation and quaternary climatic changes, *J. Atmos. Sci.*, *35*, 2362–2367.
Charbit, S., C. Ritz, and G. Ramstein (2002), Simulations of northern hemisphere ice-sheet retreat: Sensitivity to physical mechanisms involved during the Last Deglaciation, *Quatern. Sci. Rev.*, *21*, 243–265.
Cortijo, E., et al. (1999), Changes in meridional temperature and salinity gradients in the North Atlantic Ocean (30°–72°N) during the last interglacial period, *Paleoceanography*, *14*, 23–33.
Crucifix, M., and M.-F. Loutre (2002), Transient simulations over the last interglacial period 126–115 kyr BP: Feedback and forcing analysis, *Clim. Dyn.*, *19*, 417–433.
DeNoblet, N. I., et al. (1996), Possible role of atmosphere-biosphere interactions in triggering the last glaciation, *Geophys. Res. Lett.*, *23*, 3191–3194.
Gallée, H., et al. (1992), Simulation of the Last Glacial Cycle by a coupled, sectorially averaged climate-ice sheet model: 2. Response to insolation and CO_2 variations, *J. Geophys. Res.*, *97*, 15,713–15,740.
Khodri, M., et al. (2001), Simulating the amplification of orbital forcing by ocean feedbacks in the last glaciation, *Nature*, *410*, 570–574.
Khodri, M., et al. (2003), Modelling the climate evolution from the last interglacial to the start of the last glaciation: The role of the Arctic ocean freshwater budget, *Geophys. Res. Lett.*, *30*(12), 1606, doi:10.1029/2003GL017108.
Mangerud, J., et al. (1979), Correlation of the Eemian (interglacial) stage and the deep-sea oxygen-isotope stratigraphy, *Nature*, *277*, 189–192.
Petit, J.-R., et al. (1999), Climate and atmospheric history of the past 420,000 years from the Vostok ice core, Antarctica, *Nature*, *399*, 429–436.
Petoukhov, V., et al. (2000), CLIMBER-2: A climate system model of intermediate complexity. Part I: Model description and performance for present climate, *Clim. Dyn.*, *16*, 1–17.
Ritz, C., A. Fabre, and A. Letréguilly (1997), Sensitivity of a Greenland ice sheet model to ice flow and ablation parameters: Consequences for evolution through the last climatic cycle, *Clim. Dyn.*, *13*, 11–24.
Tarasov, L., and W. R. Peltier (2003), Greenland glacial history, borehole constraints and Eemian extent, *J. Geophys. Res.*, *108*(B3), 2143, doi:10.1029/2001JB001731.
Turon, J.-L. (1984), Direct land/sea correlations in the last interglacial complex, *Nature*, *309*, 673–676.
Vettoretti, G., and W. R. Peltier (2003), Post-Eemian glacial inception. Part II: Elements of a cryospheric pump, *J. Clim.*, *16*, 912–927.
Vettoretti, G., and W. R. Peltier (2004), Sensitivity of glacial inception to orbital and greenhouse gas climate forcing, *Quatern. Sci. Rev.*, *23*, 499–519.
Waelbroeck, C., et al. (2002), Sea-level and deep water temperature changes derived from benthic foraminifera isotropic records, *Quatern. Sci. Rev.*, *21*, 295–305.
Wang, Z., and L. A. Mysak (2002), Simulation of the Last Glacial Inception and rapid ice sheet growth in the McGill paleoclimate model, *Geophys. Res. Lett.*, *29*(23), 2102, doi:10.1029/2002GL015120.

S. Charbit, M. Kageyama, and G. Ramstein, LSCE-IPSL CE Saclay L'Orme des Merisiers Bât. 701 91191 Gif-sur-Yvette Cedex, France. (masa.kageyama@cea.fr)

M. Khodri, Geophysical Fluid Dynamics Laboratory, U.S. Department of Commerce, National Oceanic and Atmospheric Administration, Forrestal Campus, U.S. Route 1, 201 Forrestal Road, P. O. Box 308, Princeton, NJ 08542-0308, USA.

C. Ritz, Laboratoire de Glaciologie et Géophysique de l'Environnement, 54, rue Moliére, 38402 Saint-Martin-d'Hères Cedex, France.