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Modelling the climate evolution from the last interglacial to the start of the last glaciation: The role of Arctic Ocean freshwater budget

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[1] The Earth's climate has alternated between glacial and interglacial stages, but we still do not understand how the changes in insolation have been amplified by the climate system to lead to such a large climatic response. Here we use a climate model of intermediate complexity to show that the orbitally induced shift from an interglacial to a pre-glacial climate is highly sensitive to the high northern latitudes moisture budget. Using the insolation forcing alone, the model produces a first order response of sea surface temperatures and thermohaline circulation evolution which agrees with the available geological data. Under the same gradual insolation evolution, adding a small freshwater input into the Arctic Ocean induces a rapid climate switch from an interglacial to a pre-glacial mode. These results suggest a simple connection between orbital forcing and the thermohaline circulation through a freshwater threshold within the ocean-atmosphere-sea-ice system. **INDEX TERMS:** 3344 Meteorology and Atmospheric Dynamics: Paleoclimatology; 3309 Meteorology and Atmospheric Dynamics: Climatology (1620); 3339 Meteorology and Atmospheric Dynamics: Ocean/atmosphere interactions (0312, 4504); 4267 Oceanography: General: Paleoceanography; 1620 Global Change: Climate dynamics (3309). **Citation:** Khodri, M., G. Ramstein, J. C. Duplessy, M. Kageyama, D. Paillard, and A. Ganopolski, Modelling the climate evolution from the last interglacial to the start of the last glaciation: The role of Arctic Ocean freshwater budget, *Geophys. Res. Lett.*, 30(12), 1606, doi:10.1029/2003GL017108, 2003.

1. Introduction

[2] How seasonal and latitudinal variations of the incident solar radiation initiate internal feedbacks that produce a shift from an interglacial to a glacial mode is still a matter of debate. Studies based on Atmospheric General Circulation Models (AGCMs) forced by modern sea surface temperatures (SSTs) or coupled to mixed layer ocean [Phillipps and Held, 1994; Rind et al., 1989], have not been successful in simulating perennial snow cover, while accounting for ocean and vegetation feedbacks has produced results in better agreement with proxy records [de Noblet et al., 1996; Khodri et al., 2001]. Nevertheless, beyond the Milankovitch theory that argues for an orbitally induced shift from an

interglacial to a glacial climate, the available proxy records give important clues that have still not been fully explored through modelling studies. Recent work from North Atlantic sediment cores clearly identifies a cooling step correlated to a significant change in deep water circulation around 115 kyr BP [Shackleton et al., 2002; Adkins et al., 1997; Hall et al., 1998]. There is more evidence, from both land and marine records, for an "intra-Eemian" cooling event in the high latitudes of the Northern Hemisphere around 120 kyr BP [Cortijo et al., 1999; Sánchez-Goñi et al., 1999; Tzedakis et al., 2002]. It appears that this early "cold reversal" within the Eemian was associated to low salinity values in the Northern Seas [Cortijo et al., 1999] and to a significant shallowing of North Atlantic Deep Water [Maslin et al., 2001]. Superimposed to those abrupt events, on longer time scales there is an increasing trend in the NADW production during Marine Isotope Stage 5e [Duplessy and Shackleton, 1985]. After the mid-Eemian cold reversal, it seems that the NADW recovers, but the climate warming does not return to full interglacial conditions before the second cooling step, 115 kyr BP, which marks the beginning of the latest glacial stage. However, during the whole last interglacial period mid- to low latitudes surface ocean temperatures remain warm [McManus et al., 2002].

[3] The lag between insolation forcing and northern high latitudes cooling extrema at 120 kyr BP emphasises the necessity of exploring the mechanisms responsible for the changes occurring through the last interglacial and their implication for the last glacial inception. We propose to explore this issue by combining an Earth system Model of Intermediate Complexity (EMIC), a coupled ocean-atmosphere GCM (OAGCM), and the comparison to paleoclimatic data. First, we simulated the climate evolution from the stable interglacial (126 kyr BP) to the period of major ice sheet build-up (115 kyr BP) and to compare the results to data. For this, we use CLIMBER-2 [Petoukhov et al., 2000] forced only with the evolution of the orbital parameters between those two dates (Table 1). Second, since OAGCM results have shown the importance of the ocean in amplifying the initial insolation forcing [Khodri et al., 2001], we performed sensitivity experiments to determine the thresholds in freshwater that may convert the insolation decrease of 115 kyr BP into a maximum northern high latitudes cooling.

2. Results and Discussion

[4] The increasing axial tilt and gradual shift of the date of perihelion into the winter months from 126 kyr BP to 115

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Table 1. Orbital Parameters of the Earth Used in the Simulations for Incoming Solar Radiation Calculation at the Top of the Atmosphere

Month	Control	115 kyr BP	126 kyr BP
Eccentricity	0.016724	0.041421	0.04
Obliquity (axial tilt °)	23.446	22.404	23.928
Perihelion-180° (°)	102.04	111.01	291.24
CO ₂ (ppm)	280	280	280

The transient experiment starts with the initial conditions of a 126 kyr BP equilibrium experiment and is run for 11000 years with the corresponding orbital parameters varying according to *Berger* [1978]. The present day simulation is referred to as the control run. For all the simulations, the concentration of the atmospheric CO₂ is kept at its pre-industrial level, as shown in the Vostok records [*Raynaud et al.*, 1993].

kyr BP induces a progressive decrease of the summer insolation at the high latitudes of the Northern Hemisphere (Figures 1a and 1c) and an increase in insolation during winters, centred in the subtropical latitudes. The summer SSTs over the North Atlantic (at 75°N) cool by 1.2°C in summer and by 0.8°C in annual mean, while the low latitude annual mean SSTs remain roughly unchanged, since the winter warming counterbalances the summer cooling (Figures 1e and 1f). In the mean time, in response to the decreasing summer insolation forcing, the sea-ice melt is progressively reduced, leading to an extension and thickening of Arctic Ocean sea-ice, reaching a plateau at 120 kyr BP (not shown). The associated albedo increases by about 10% during this time period, enhancing the orbitally-forced cooling in northern high latitudes (Figures 1a–1f and caption). Therefore, the high latitude SSTs are at their

coolest early within the last interglacial thanks to the sea-ice albedo increase, in agreement with the SSTs inferred from North Atlantic marine sediment cores [*Cortijo et al.*, 1999]. This first order response to the insolation forcing simulated by CLIMBER-2 provides a possible explanation for the early Eemian cooling. However, the model does not succeed in simulating any abrupt cold reversal and shallowing of NADW neither at 120 kyr BP or at 115 kyr BP.

[5] Instead, during the simulated last interglacial, the coupled ocean-atmosphere system responds to the orbital forcing by enhancing the northward heat transport by both the ocean and atmosphere. The strength of the thermohaline circulation (THC) increases from 126 kyr BP to 120 kyr BP and then remains constant and vigorous until 115 kyr BP (Figure 1g). The stronger THC increases the amount of heat transported northward by the Atlantic Ocean during this time period (by 7.5%) and the atmospheric maximum northward heat transport, at 115 kyr BP, experiences a 6.5% increase mainly due to the dry static heat component. Both the increased ocean and atmospheric northward heat transports are driven by the equator-to-pole insolation gradient. Between 126 kyr and 120 kyr BP, the total fresh water flux into the Arctic Ocean decreases by 0.01 Sverdrup (Sv) mainly because of decreased precipitation (Table 2 and caption). This leads to an increase in salinity of Arctic waters (Figures 2a–2c), which gradually penetrate in the Greenland-Iceland-Norwegian (GIN) seas, inducing higher winter deep mixing rate at 120 kyr BP (Figure 1g). In the meantime, the subtropical SSTs warm during winter because of the solar forcing. At 120 kyr BP, in response to the

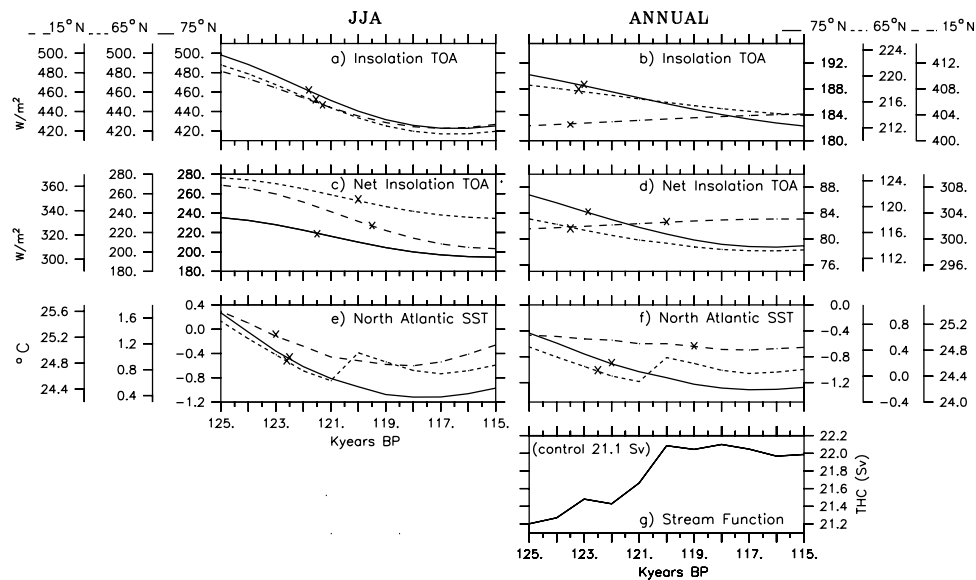


Figure 1. Time series of the North Atlantic Sea Surface Temperature (SST) and thermohaline circulation strength (THC) in response to changing insolation forcing at the top of the atmosphere (TOA) between 125 kyr BP and 115 kyr BP. **a–b**, June–July–August (JJA, a) and Annual mean (b) of incoming solar radiation at the top of the atmosphere at 15°N (dashed curve), 65°N (dotted curve) and 75°N (solid curve) in W/m² [*Berger*, 1978]. **c–d**, The same as in a,b but for the net values at the top of the atmosphere simulated by CLIMBER-2; **e–f**, June–July–August (JJA, e) and Annual mean (f) of North Atlantic SSTs at 15°N (dashed curve), 65°N (dotted curve) and 75°N (solid curve) in °C. **g**, Evolution of the annual mean values of intensity of the THC in Sverdrup (Sv, i.e. 10⁶.m³.s⁻¹). In **a–f**, the corresponding value for the control is indicated by a cross. The value of intensity of the THC for the control run is also indicated in **g**. By comparing the rate of summer insolation decrease at 75°N, 65°N and at 15°N (**a–d**), the rate (the first derivative) of cooling is twice more rapid in high latitudes than to lower latitudes.

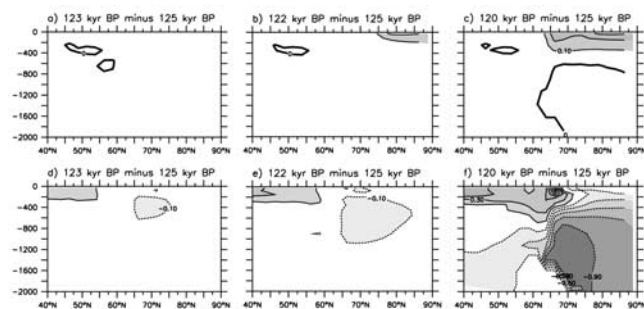
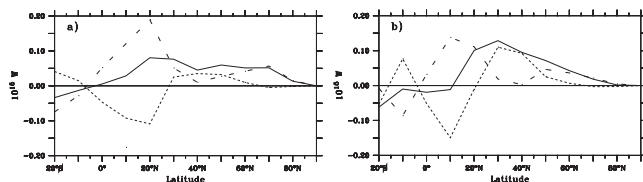
Table 2. Freshwater Budget Evolution in the Arctic Ocean in Response to the Changing Insolation of the Last Interglacial

Sverdrups (Sv)	126 kyr BP	120 kyr BP	115 kyr BP	115 kyr BP m 126 kyr BP
P-E	0.107	0.102	0.098	-0.009
Runoff	0.085	0.088	0.089	+0.004
Net ice melting	-0.026	-0.03	-0.031	-0.005
Total	0.166	0.16	0.156	-0.01

During the transient experiment the precipitation minus evaporation (P-E) becomes the major operating distillation process. Indeed, the water distillation rate by sea-ice freezing, due to increased sea-ice export from production area (-0.005 Sv from 126 kyr BP to 115 kyr BP) is compensated by the increased runoff (+0.004 Sv).

increased ocean meridional overturning, the warmer surface waters reach the deep convection sites (Figures 2d–2f). Then, the competition between the increased salinity and the warmer surface temperatures at the deep convection sites leads to a self-stabilised THC from 120 kyr BP to 115 kyr BP (Figure 1g). The warming effect of the increased northward ocean heat transport affects surface temperatures as far as 65°N at 120 kyr BP (Figures 1e–1f, and Figure 2f). As shown on Figures 1e–1f, at 120 kyr BP, during summer, there is indeed a step up of SST at 65°N. This excursion to warmer temperatures during summer illustrates a short attempt of the ocean heat transport to go further north in response to the strengthened THC which peaks at 120 kyr BP. However it is immediately followed by a decrease in temperature because of the summer insolation decrease and the sea ice extent feedbacks. In annual mean, the decreasing high latitude SSTs therefore meanly reflect the summer response.

[6] According to the data reviewed in the introduction section there is indeed a small global trend of increased production of NADW during the stage 5e [Duplessy and Shackleton, 1985] in agreement with our results and with those of Wang and Mysak [2002] who used another EMIC. The THC and SSTs changes simulated here therefore seem realistic. However, there is no snow accumulation over Northern Hemisphere continent and no abrupt cooling associated to a reduced THC at 120 kyr BP nor at 115 kyr BP. The crude land surface resolution in CLIMBER, and in EMICS in general, might explain why we do not obtain snow nucleation over Northern Hemisphere continents. Also, the present model does not include a coupling

**Figure 2.** Response of the winter North Atlantic salinity and temperature to changing insolation forcing. Latitude-depth distribution of January–February means of the salinity in PSU (a, b, c) and temperature in °C in the North Atlantic (d, e, f).**Figure 3.** (a) Annual mean zonal average difference between 115 kyr BP and control simulations of the atmospheric heat transport in PW, for total (solid line), dry static (dashed line) and latent heat (dotted line) simulated by CLIMBER-2. (b) Same as (a) but obtained by the IPSL model [Khodri *et al.*, 2001].

to an ice-sheet model, which could help in starting a glaciation.

[7] Using a finer grid OAGCM forced by the 115 kyr BP insolation, Khodri *et al.* [2001] have shown that the surface ocean cooling north of 60°N, the Arctic sea ice extension, the southward shift of the polar front and the shallowing of the NADW, help maintain perennial snow over the Canadian Archipelago. This reorganisation, due to a positive feedback operating via the increased atmospheric moisture transport from the warmer tropical ocean to the northern seas therefore corroborates the data which demonstrate such an event at 115 kyr BP [Shackleton *et al.*, 2002; Adkins *et al.*, 1997; Hall *et al.*, 1998]. In both the transient interglacial EMIC simulation and the snapshot OAGCM simulation, the increased meridional gradient of SSTs leads to an enhanced northward atmospheric heat transport; but, at 115 kyr BP in the transient simulation the annual mean atmospheric heat transport at about 30°N is almost half that in the OAGCM simulation (0.07 PW and 0.13 PW respectively), and is mainly represented by the dry static heat (71.4% vs. 30% for the OAGCM simulation, Figure 3). The smaller increase of northward atmospheric moisture transport in the EMIC (+0.03 Sv) compared to the OAGCM (+0.06 Sv), is not large enough to induce a decrease of THC at 115 kyr BP (Figure 3). Therefore small variations of the hydrologic cycle over the North Atlantic may have dramatic consequences. Since Cortijo *et al.* [1999] also reported a freshening of the Northern oceans from 120 kyr BP until 115 kyr BP, it seems that the hydrological cycle response in CLIMBER is too weak to capture such an event. Moreover, Cuffey and Marshall [2000] also argue that the sea level rise during the last interglacial was partly due to meltwater from Greenland. As the version of CLIMBER-2 used here is not coupled to an ice sheet model, this might also explain why the model, when only forced by the orbital parameters does not reproduce the mid-Eemian cold reversal.

[8] To explore such a sensitivity to the freshening of the Northern Seas, we increased the northward atmospheric moisture transport by about 0.03 Sv at 115 kyr BP in the EMIC simulation, to mimic the OAGCM results. In this sensitivity experiment, the THC weakens, as obtained in the OAGCM simulation and perennial snow over Northern Canada is simulated (not shown). This last result shows that both the OAGCM and the EMIC lead to the same simulated climate when applying the same hydrological cycle changes. To identify the threshold at which such a reversal is achieved, in an other sensitivity experiment, the freshwater transport by the atmosphere from low to high

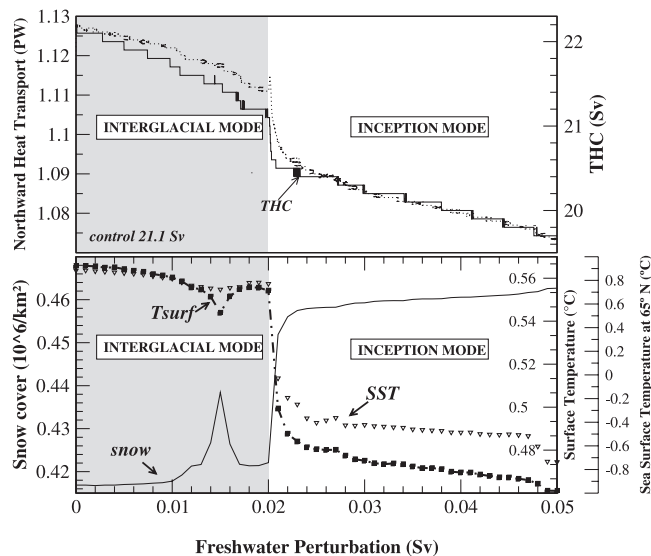


Figure 4. Response of the THC, Maximum Atlantic Northward heat transport (top panel) to freshwater supply (Sv, x axis) in the northern seas and the impact (bottom panel) over the Atlantic SSTs at 65°N, the northern Canada surface temperature and snow cover (70°–80°N) at 115 kyr BP. In this sensitivity experiment, the freshwater transport by the atmosphere from low to high latitudes is progressively increased from 0 to 0.05 Sv.

latitudes is progressively increased from 0 to 0.05 Sv in the climate simulated by CLIMBER at 115 kyr BP (Figure 4). At 0.015 Sv, there is a short reversal in land surface temperature and snow cover which corresponds to a first attempt for the climate to shift to a colder mode. From this result we can suppose that the mid-Eemian cooling recorded by the data may have been such an attempt, since according to environmental reconstructions, the climate that followed this event did not return to full interglacial conditions. But the most important results lies into the non-linear response of the simulated climate at 115 kyr BP to the freshwater forcing. Indeed, when the small threshold in freshwater input of 0.02 Sv into high latitude ocean is reached, the climate switches abruptly from an interglacial mode with warm high latitudes temperatures and snow melted in summer over northern Canada to a pre-glacial mode characterised by cooler land, perennial snow cover and weakened THC (Figure 4).

[9] We have shown that the maximum cooling in North Atlantic ocean is achieved before the minimum of summer insolation as recorded by data thanks to the sea-ice albedo feedbacks. However, to obtain a North Atlantic freshening and cooling at 115 kyr BP, consistent with the data and to the OAGCM simulation it is necessary to amplify the hydrological cycle. Here, we did not consider the ice sheet feedback, but our results show that such an interaction could have important impacts through land ice melting and albedo

effects. We are currently developing the model with which we will be able to study such interactions.

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