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Possible role of atmosphere-biosphere interactions in triggering the last glaciation

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Abstract. We coupled a global biome model iteratively with an atmospheric general circulation model to study the possible role of vegetation in the climate system, at the time of glacial inception 115,000 years ago. Orbital forcing alone was not sufficient to initiate glaciation when other components of the climate system were kept as present (atmospheric composition, oceans, biosphere and cryosphere). Summers were however cold enough to induce major vegetation shifts in high northern latitudes. Southward migration of the boreal forest/tundra limit helped to create favourable conditions for continental ice-sheet growth, with increasing snow depth and duration in Labrador, Arctic Canada and northern/western Fennoscandia. These results support a role for biogeophysical feedback in initiating glaciations.

Introduction

According to the Milankovitch theory [*Hays et al.*, 1976; *Berger*, 1988; *Imbrie et al.*, 1992-1993], glacial periods begin under orbital configurations when northern-hemisphere summer insolation is at a minimum. Atmospheric general circulation models (AGCMs) however do not predict the required increase in permanent snow cover at the start of the last glaciation as a consequence of orbital forcing alone [*Royer et al.*, 1984; *Rind et al.*, 1989; *Phillipps and Held*, 1994]. Positive feedbacks from other components of the climate system have been proposed, including changes in sea-surface temperatures and sea-ice distribution [*Phillipps and Held*, 1994; *Syktus et al.*, 1994; *Dong and Valdes*, 1995; *Gallimore and Kutzbach*, 1995], in the thermohaline circulation [*Imbrie et al.*, 1992-1993; *Cortijo et al.*, 1994], and climate-induced vegetation shifts [*Gallée et al.*, 1992]. Diagnostic calculations [*Harrison et al.*, 1995] show vegetation changes large enough to induce further perturbations in the atmosphere [*Street-Perrott et al.*, 1990; *Foley et al.*, 1994; *Gallimore and Kutzbach*, 1996; *TEMPO Members*, 1996], which could in turn modify the distribution of biomes. Coupled climate/biome model experiments have already been carried out for present-day climate [*Henderson-Sellers*, 1993; *Claussen*, 1994]. *Claussen and Gayler* [subm.] have further applied a coupled model to the climate of the mid-Holocene and were able to simulate a large northward migration of savannas in the western Sahara, as seen in paleodata [*Jolly et al.*, in press].

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Experimental set-up

The atmospheric general circulation model (AGCM) developed at the Laboratoire de Météorologie Dynamique (version 5.3) [*Sadourny and Laval*, 1984; *Harzallah and Sadourny*, 1995] and the global biome model BIOME (version 1.0) [*Prentice et. al.*, 1992] were coupled as in Figure 1. The AGCM includes an advanced land-surface scheme, based on SECHIBA [*Ducoudré et al.*, 1993], but with more realistic canopy conductance and seasonal leaf area variations [*Haxeltine and Prentice*, 1996]. Vegetation distributions for the climate of 115 kyr BP were based on modern climatology modified using mean monthly anomalies (departures from the control simulation) smoothly interpolated on to a 0.5° grid. Each AGCM grid cell is assigned fractional covers of biomes. Fluxes are calculated separately for each biome and aggregated to provide lower boundary conditions for the AGCM.

Each AGCM simulation was 16 years long, with sea-surface temperatures and sea-ice distribution prescribed as present. Climatic variables were averaged over the last 15 years. The paleoclimate modelling intercomparison project [PMIP, *Joussaume and Taylor*, in press] recommends averaging over at least 10 years, and initial studies [*de Noblet et al.*, unpublished] showed that at least 10 years are needed to achieve stable high-latitude biomes.

Five iterations were necessary for the atmosphere/biosphere system to reach equilibrium under 115 kyr BP orbital conditions prescribed from *Berger and Loutre* [1991]. The first palaeoclimate simulation (G1) was obtained using a biome distribution based on modern climatology (long-term monthly means of temperature, precipitation and sunshine derived from *Leemans and Cramer* [1991]). In each subsequent iteration (G2-G5) biome distribution was changed based on the mean climate from the previous iteration. Atmospheric CO_2 concentration was 345 ppm in the control simulation and 280 ppm in the

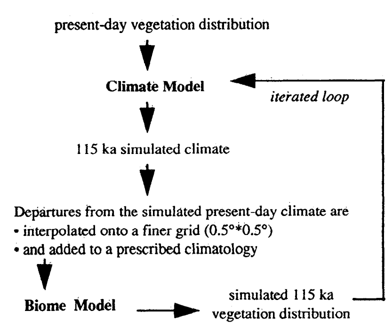

Figure 1. Schematic of the coupling between the atmospheric general circulation model and the biome model.

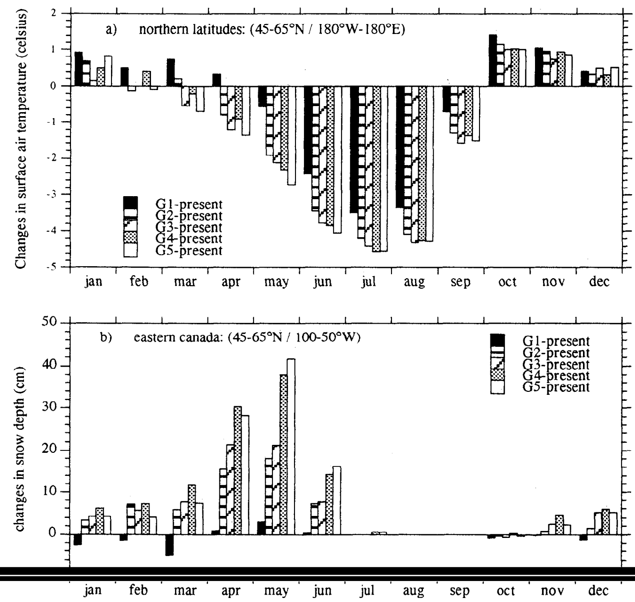

Figure 2. Simulated changes in monthly mean *a,* surface air temperatures (°C) in the 45-65°N latitude band, *b,* snow depth (cm) in eastern Canada (45-65°N, 100-50°W).

palaeoclimate simulations, consistent with ice-core evidence [*Barnola et al.*, 1987] that CO_2 did not fall to glacial levels until well after 115 kyr BP. One may argue that the CO_2 level should have been set to its preindustrial value in the control simulation as well, but since sea-surface temperatures are not changed, the sensitivity of the model to lowering CO_2 concentration is severely restricted [*Hewitt and Mitchell*, 1996]. Moreover solar forcing changes seasonally and latitudinally and vegetation changes are sensitive to changes in the seasonal cycle.

Sensitivity experiments carried out with AGCMs coupled to interactive mixed-layer ocean models [*Foley et al.*, 1994; *Dong and Valdes*, 1995] have shown enhancement of orbitally-induced climate changes when the ocean surface is set free. Fixed ocean surface conditions may limit the response of the simulated climate. Our goal was to isolate the sensitivity of the atmosphere to changes in the global distribution of vegetation, while recognizing that atmosphere/ocean interaction may produce further feedbacks.

Results

Perihelion at 115 kyr BP was in northern-hemisphere winter, as at present, while eccentricity was greater and obliquity less than present [*Berger and Loutre*, 1991]. Both factors imply a reduced seasonal insolation contrast in the northern hemisphere. Orbital forcing alone (simulation G1) produced conditions in the 45-65°N band that were cooler than present in the summer months, with maximum (3-4 K) cooling in July, but warmer than present from late autumn to early spring (Figure 2*a*). Snowfall and snow cover were generally less than present (Figures 3*a*, 2*b*), due to the warm winters and springs. These results confirm that 115 kyr BP orbital forcing of this AGCM is insufficient to start glaciation.

The cool summers simulated in G1 however produced a major change in vegetation patterns (Figures 4*a*,*b*). Present-day taiga and cold deciduous forests were replaced by tundra over large areas. These vegetation changes would raised surface albedo throughout the year and most of all during winter and spring, due to snow masking by forest [*Bonan et al.*, 1992; *Chalita and Le Treut*, 1994; *Foley et al.*, 1994]. The northern continents would therefore have absorbed considerably less solar energy than was calculated in G1, above all during spring snowmelt.

Simulation G2 shows this feedback. Simulated winter temperatures for 45-65°N (Figure 2*a*) were lower in G2 than in G1. April temperatures became lower than present, although they were warmer than present in G1. Winter snowfall remained less than present, but spring snowfall was greater and snow lasted longer (Figure 3*b*). Summer temperatures were further lowered and tundra extended even further south than in G1 (Figure 4*c*). In subsequent simulations tundra continued to expand but at a declining rate (Figure 4*d*-*f*). Mean temperatures for the summer months (Figure 2*a*) showed a downward trend, levelling towards G5. Winter temperatures (Figure 2*a*) and the

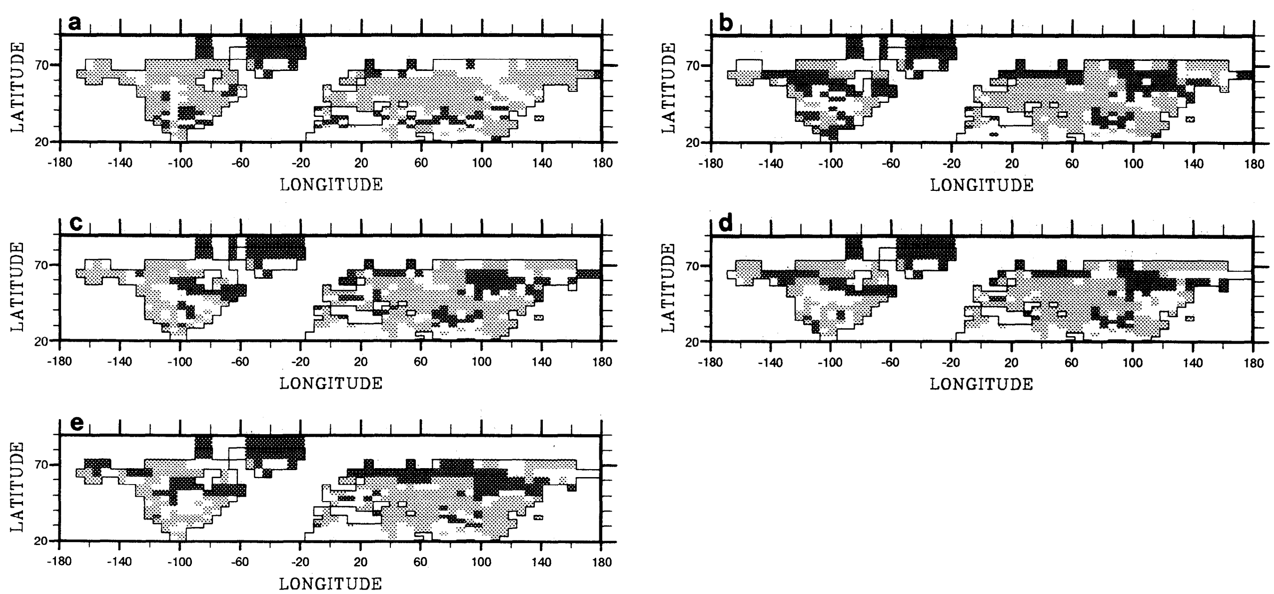

Figure 3. Simulated simultaneous changes in snow depth and duration: *a,* G1, departures from present; *b,* G2, departures from present; *c,* G3, departures from present; *d,* G4, departures from present; *e,* G5, departures from present. Dark shading indicates regions where both snow depth and duration increase; light shading indicates regions where they both decrease.

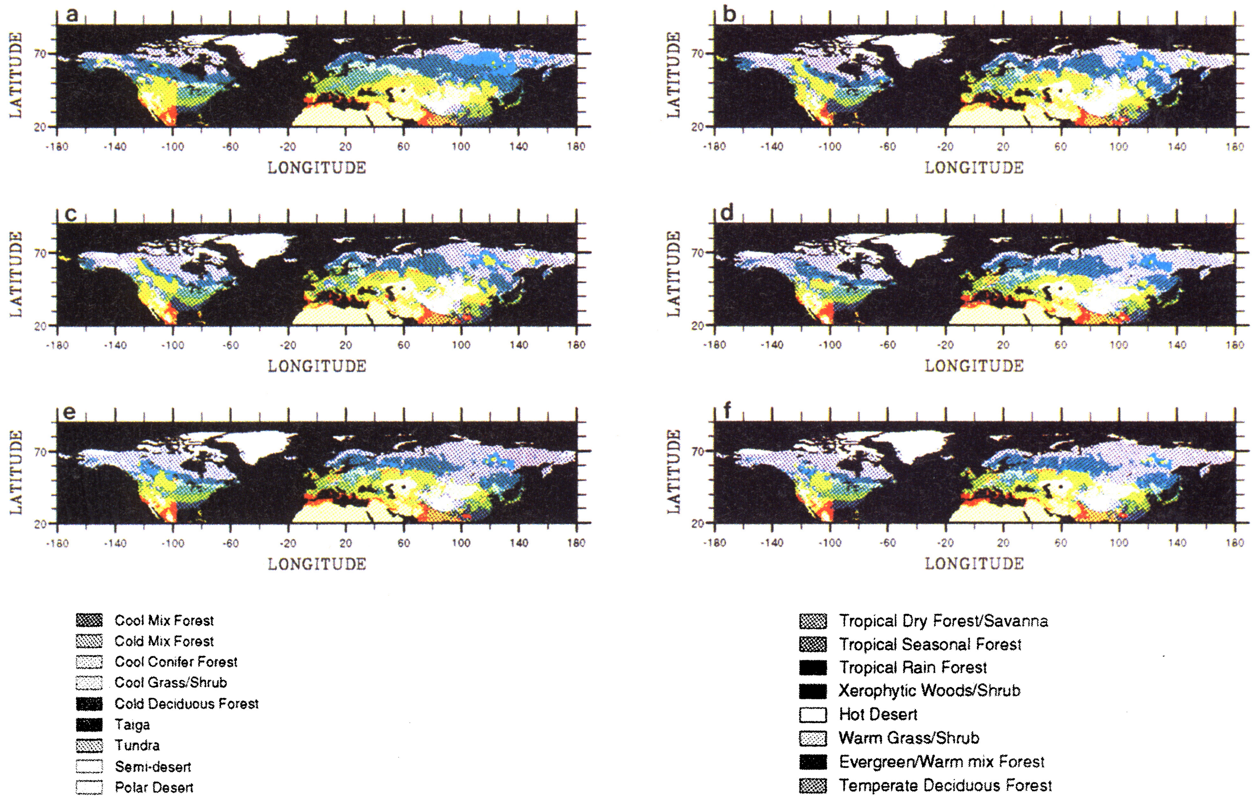

Figure 4. Simulated vegetation distribution (20-90°N): *a,* present, based on climatological data [*Leemans and Cramer*, 1991]; *b,* from experiment G1, based on 115 kyr BP orbital conditions and present vegetation; *c,* from experiment G2, based on 115 kyr BP orbital conditions and vegetation from G1; *d,* from experiment G3, based on 115 kyr BP orbital conditions and vegetation from G2; *e,* from experiment G4, based on 115 kyr BP orbital conditions and vegetation from G3; *f,* from experiment G5, based on 115 kyr BP orbital conditions and vegetation from G4.

spatial distribution of snow cover (Figure 3) varied more between iterations, due to the strength of advective effects [*Harrison et al.*, 1992]. In western Canada, for example, although the tundra area increased steadily (Figure 4), non-directional changes in the meridional position of the Aleutian Low induced variability in the snow statistics (Figure 3). Certain areas nevertheless showed a consistent trend towards greater snow accumulation (as shown for eastern Canada in Figures 2*b* and 3): Québec/Labrador, central Canada, Greenland and the Canadian Arctic, northern/western Fennoscandia, northern Siberia and Tibet (Figure 3). The high-relief areas of Labrador and the Canadian Arctic (e.g. Baffin Island) and the mountains of Fennoscandia have been considered as likely initial growth centres for the Laurentide and Fennoscandian ice sheets, respectively [*Ives et al.*, 1975; *Andrews and Mahaffy*, 1976]. Thus, our results point to a mechanism by which biogeophysical feedback could have helped to create conditions favourable to the growth of mid-latitude ice sheets.

Discussion

There can be many reasons why most AGCMs do not initiate glaciation when sea-surface temperature and sea-ice distribution are kept as present, see *Dong and Valdes* [1995] and *Phillipps and Held* [1994]. One of them is the present-day land-surface temperatures simulated in spring and summer. Our control values are about 5°C too warm in the mid and high northern latitudes. The cooling we simulate may be correct, but the ground is still too warm in the model to prevent snow from melting. Another possible problem is in the snowmelt parameterization. *Oglesby* [1990] and *Dong and Valdes* [1995] showed that if all the available excess energy is applied to snowmelt, as in our AGCM and many others, then the snow melts too quickly.

Our simulations cannot give a complete picture of the climatic conditions around the end of stage 5e. Changes in the thermohaline circulation may also have contributed to the start of glaciation [*Imbrie et al.*, 1992, 1993]. However substantial cooling in the North Atlantic did not occur until the ice sheets were well developed [*Ruddiman and McIntyre*, 1979; *Cortijo et al.*, 1994]. A persistently warm sea-surface favoured the delivery of moisture to the growing ice-sheets.

Models that include interactive sea-surface conditions have shown that the low mean annual insolation in high latitudes at 115 kyr BP would have led to increased Arctic sea ice thickness and duration and thus to lower temperatures and greater snow persistence on the northern continents [*Phillipps and Held*, 1994; *Syktus et al.*, 1994; *Harrison et al.*, 1995; *Gallimore and Kutzbach*, 1995; *Dong and Valdes*, 1995]. Moreover, vegetation-induced spring and summer cooling might have been amplified by the biosphere-atmosphere-ocean interaction mechanism [*Bonan et al.*, 1992; *Foley et al.*, 1994; *Gallimore and Kutzbach*, 1996].

Our results support the hypothesis [*Gallée et al.*, 1992] that consideration of biosphere-atmosphere interactions, which may be further modified by atmosphere-ocean interactions, is essential for understanding and modelling the dynamics of glacial-interglacial cycles. These interactions can only be fully explored by explicit interactive coupling of models of the atmosphere and biosphere.

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