ON THE ORIGIN OF THE INSTABILITY OF GLACIAL CLIMATES

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1. MOTIVATION

Reconstructions of oxygen isotopes from Greenland ice cores have revealed that the climate of the last glacial period was interrupted by rapid climate swings in the Northern Hemisphere, with an approximate spacing of 1,500 years. These Dansgaard-Oeschger (DO) stadial-interstadial (cold-warm) transitions were accompanied by rapid increases in surface air temperature of up to 16°C over Greenland followed by a long-lasting cooling. Although the volume of proxy data documenting these millennial-scale oscillations is continuously growing, the physical mechanism responsible for their existence and the causes of the major reorganization of the ocean-atmosphere coupled system remain poorly understood.

Several physical mechanisms have been proposed so far to explain the possible origin of these DO cycles and the relative stability of the Holocene. Jumps between cold stadials and warm interstadials were first reproduced in noise-free models with the help of a periodic surface freshwater forcing. The stochastic and coherence resonance mechanisms were then shown to produce stadial-interstadial transitions whose characteristics share many similarities with the observed DO cycles. However, noise-free millennial oscillations under steady freshwater (and solar) forcing were also found more than a decade ago in both zonally-averaged and 3D climate models. These oscillations are simply the so-called deep-decoupling oscillations described in detail by Winton and Sarachik (1993). Loving and Vallis (2005) suggested that glacial climates exhibit abrupt millennial-scale oscillations because of the reduced strength of the Atlantic Meridional Overturning Circulation. However, observations suggest that the strength of the circulation during the last glacial period was probably not very different from today, despite the large sea-ice changes (Lynch-Stieglitz et al., 2007). Here, we suggest instead that this is the reduced thermal stratification of the glacial ocean, mainly caused by upper ocean cooling, that is at the origin of the instability of glacial climates.

2. THE MODEL

We use an idealized coupled ocean-atmosphere-sea ice model set up in a single-hemisphere geometry. The choice of this geometry is first motivated by the strong evidence from the paleorecord that the signature of the DO events was prominent in the North Atlantic, and second by the many studies that identify precursors of these events in the North Atlantic. The ocean component is based on the planetary geostrophic equations, valid for time scales much larger than the inertial period and spatial scales much larger than the internal Rossby radius of deformation. This ocean model is coupled a simple one layer thermodynamic sea ice model and an atmospheric energy balance model. The domain is a flat-bottom sector of a sphere, extending from the equator to 84°N and is 63° wide in zonal extent. The horizontal resolution is 3° and there are 15 layers whose vertical thickness increases unevenly from 50 m at the surface to 550 m at the bottom (4500 m depth). Lateral boundaries are solid vertical walls where no-slip and no-flux boundary conditions are applied. Surface boundary conditions include a prescribed freshwater forcing and a surface wind-stress forcing (Fig. 1). We focus on the marginal boundaries between stable and unstable states, so that stochastic forcing is absent. We use two control parameters, namely the magnitude p of the freshwater forcing and the planetary emissivity ε_p.

3. MILLENNIAL OSCILLATIONS

The objective of this study is to compare the sensitivities of cold and warm climate solutions. The only differences between the two climates are the latitude of the land ice edge (70°N and 50°N for the warm and cold cases respectively) and the planetary emissivity ε_p (0.60 and 0.61 for the warm and cold cases respectively), which can be understood as a parameter expressing the level of atmospheric greenhouse gas concentration. The mean glacial climate generated with p=60 cm yr^{-1} appears to be close to marginal stability. Increasing the

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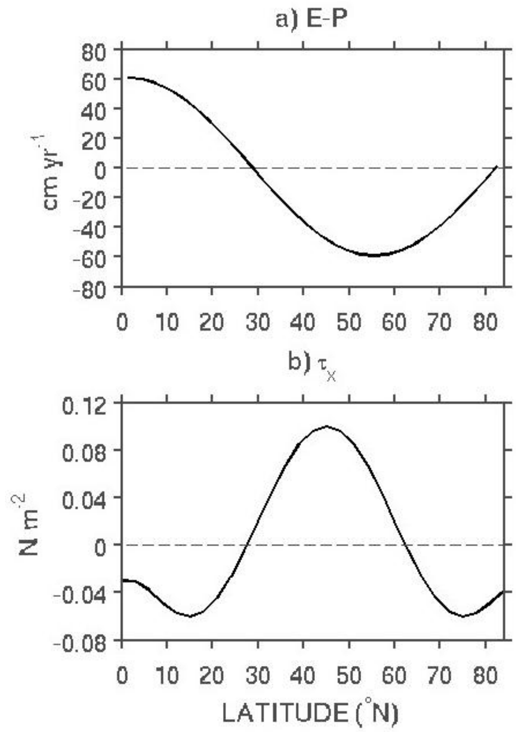

Fig. 1. (a) Meridional distribution of the freshwater forcing for p=60 cm yr^{-1} and (b) prescribed wind-stress forcing.

freshwater forcing by a small amount, until p=70 cm yr^{-1}, leads to finite amplitude millennial oscillations with a period of about 1700 years. Note that the warm climate solution is fully in the stable regime for this particular value of freshwater forcing and switches into an oscillatory regime for p=90 cm yr^{-1}.

One of the most prominent features of the DO events is their saw-tooth shaped temperature profile, as recorded in the NGRIP Greenland ice core. This aspect of the behavior of the high-latitude climate during the last glacial period is well captured by our model (Fig. 2): that is, there is a clear abrupt and significant warming followed by a plateau phase with slow cooling lasting several centuries, and finally a rapid decline back to stadial conditions.

4. RESULTS

We mentioned above that the cold climate oscillates for a lower freshwater flux than the warm climate (70 and 90 cm yr^{-1} respectively). Figure 3 reveals that this result is robust across a wide range of planetary emissivity values ε_p. Why do cold climates oscillate for significantly lower freshwater fluxes than warm climates ?

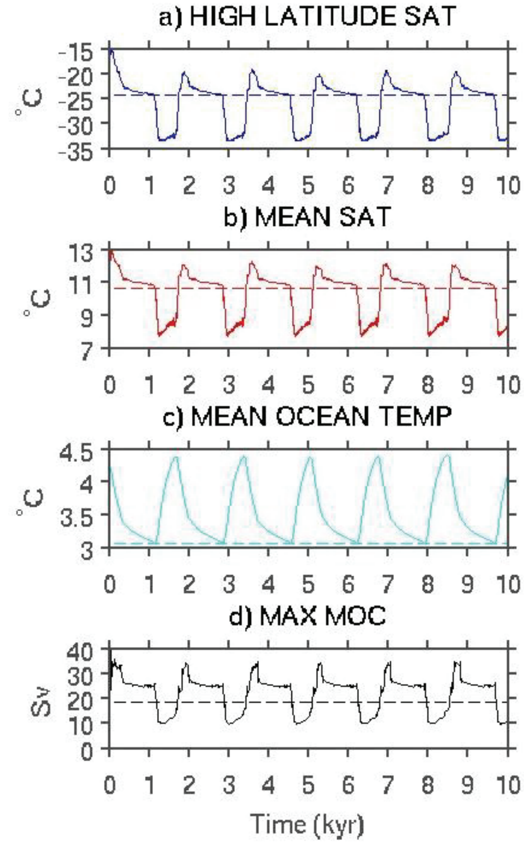

Fig. 2. Millennial oscillations of the glacial climate obtained with p=70 cm yr^{-1}. (a) High-latitude SAT (75°N-84°N), (b) mean SAT, (c) mean oceanic temperature, (d) maximum of the MOC estimated below 850 m. The dashed lines indicate the values of the nearby stable state obtained with p=60 cm yr^{-1}.

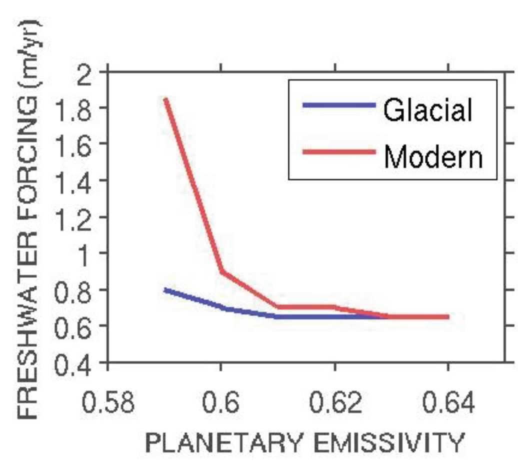

Fig. 3 The position of the transition from the stable thermal state to the oscillatory regime under cold and warm climate conditions in the freshwater forcing – planetary emissivity

parameter space.

Additional experiments (not shown here) under mixed boundary conditions revealed that this fundamental difference can be solely explained in terms of internal ocean dynamics. We then compare the characteristics of the oceanic states for the cold and warm climate solutions at marginal stability of the cold case, that is for a value of freshwater forcing that is just below that corresponding to the transition from the stable thermal state to the oscillatory regime. The stable thermal states generated with p=60 cm yr^{-1} are characterized by overturning circulations with similar strength (18 Sv, not shown).

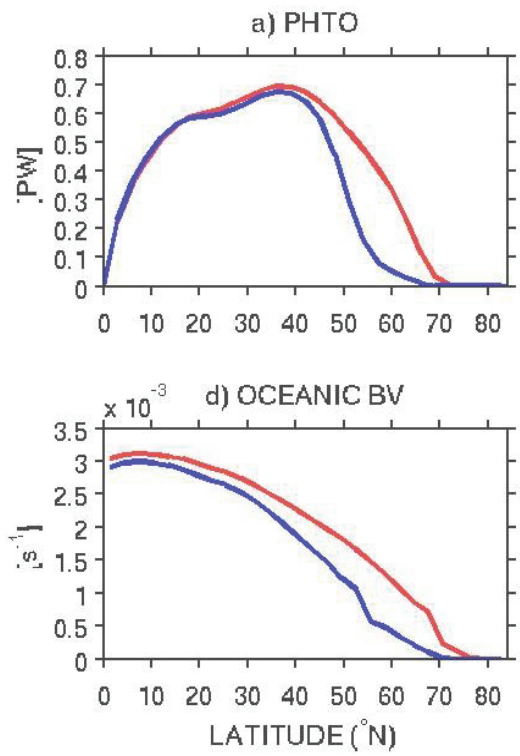

Fig. 4 Climatology of the cold and warm climate solutions obtained with p=60 cm yr^{-1}. (a) Poleward oceanic heat transport, (b) vertically-averaged Brunt-Vaisala frequency.

This feature is in qualitative agreement with observations that suggest that the strength of the glacial ocean circulation was probably not very different from today (Lynch-Stieglitz et al., 2007). However the poleward oceanic heat transport is less in the cold than in the warm case (Fig. 4a). A weaker baroclinic heat transport with the same mass transport is only possible if the vertical temperature contrast is less, which is indeed the case (Fig. 4b). This reduced density stratification is mainly caused by upper ocean cooling, salinity effects being negligible.

5. SUMMARY AND DISCUSSION

With similar overturning strengths between the cold and warm climates, this weaker temperature stratification implies a weaker baroclinic heat transport that ultimately leads to a weaker stabilization of the circulation by the negative advective-temperature feedback. As such, the polar water column stability is reduced in a cold climate and becomes therefore unstable for a smaller input of buoyancy (i.e freshwater in the present case) than warmer, interglacial climates. We are therefore led to hypothesize that the magnitude of the overall freshwater flux during the last glacial period was stronger than the value at the transition from the steady thermal state toward the oscillatory regime (see the blue and red curves respectively in Fig. 3). The converse can be hypothesized for the Holocene. This provides a simple explanation for the absence of DO events during the Holocene.

Ganopolski and Rahmstorf (2001) showed that the transition from the stable thermal state to the state of totally suppressed overturning occurs for lower freshwater fluxes in the glacial than in the modern case. The argument is similar here, but applies for the transition from the stable thermal state to the oscillatory regime. The absence of millennial oscillations in the hysteresis diagrams of Ganopolski and Rahmstorf (2001) may be due to the relatively high rate of variation of freshwater forcing used to build the stability diagrams (0.1 Sv / 1000 years). Timmermann et al. (2003) suggested that a rate of variation of freshwater forcing of about 0.001 Sv / 1000 years must be used to yield reasonable approximation of the true bifurcation structure. This is because of the critical slowing down effect, becoming very important in the vicinity of bifurcation points.

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