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Glacial Thermohaline Circulation and Climate: Forcing from the North or South?

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ABSTRACT

Based on the evidence available from both observations and model simulations, the author proposes a view that may provide a unified interpretation of the North Atlantic thermohaline variability. Because of the slow response time of the Southern Ocean (millennia) and the relatively faster response time of the North Atlantic (centuries), the North Atlantic thermohaline circulation is controlled predominantly by the climate forcing over the Southern Ocean at the long glacial cycle timescales, but by the North Atlantic climate forcing at the short millennial timescales.

Key words: interhemispheric interaction, thermohaline circulation, glacial cycle, millennial variability

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1. Introduction

One of the most robust paleoceanographic changes is the shallowing of the North Atlantic thermohaline circulation (THC) at the Last Glacial Maximum (LGM, ∼21000 years ago) (Duplessy et al., 1988; Boyle, 1992; Lee and Boyle, 1990; McCave et al., 1995) (Fig. 1). Many studies suggest that the North Atlantic THC is sensitive to North Atlantic freshwater forcing (e.g., Broecker and Denton, 1989; Stocker, 2000; Schmittner et al., 2003). This may imply that the shallowing of the North Atlantic THC is caused by the North Atlantic climate forcing. Our recent study of LGM THC in the NCAR CCSM1 (Shin et al., 2003a,b; Liu et al., 2004) shows that the glacial THC can be driven by the lower glacial atmospheric CO_2, which cools the Southern Ocean, leading to an increased sea-ice cover and in turn salinity there. This view of Southern Ocean forcing of glacial THC appears to be consistent with stable isotope ($\delta^{18}O$) records from ice cores of Byrd (Antarctic) and GISP2 (Greenland); the Antarctic temperature leads the Greenland temperature by 1000–2000 years (Fig. 2), implying an earlier change of the high latitude climate forcing, and in turn the THC, in the Southern Hemisphere than in the Northern Hemisphere. However, the interpretation of the proxy data remains controversial; the same ice core records may also be interpreted as a lead of the Greenland to the Antarctic if the extreme millennial events are considered, implying a dominant role of the North Atlantic climate forcing (Alley et al., 2002; Steig and Alley, 2003; Schmittner et al., 2003). This controversy leads to a fundamental question: Is the North Atlantic THC driven by the North Atlantic or the Southern Ocean climate forcing? Here, based on available observational and modelling evidence, we propose a unified view that accounts for both the North Atlantic and Southern Ocean climate forcing: the THC is driven predominantly by the Southern Ocean climate forcing at orbital timescales, but by the North Atlantic climate forcing at millennial timescales. We will first briefly discuss a recent LGM simulation in section 2 and then the implication for interhemispheric climate interaction in section 3. A summary is given in section 4.

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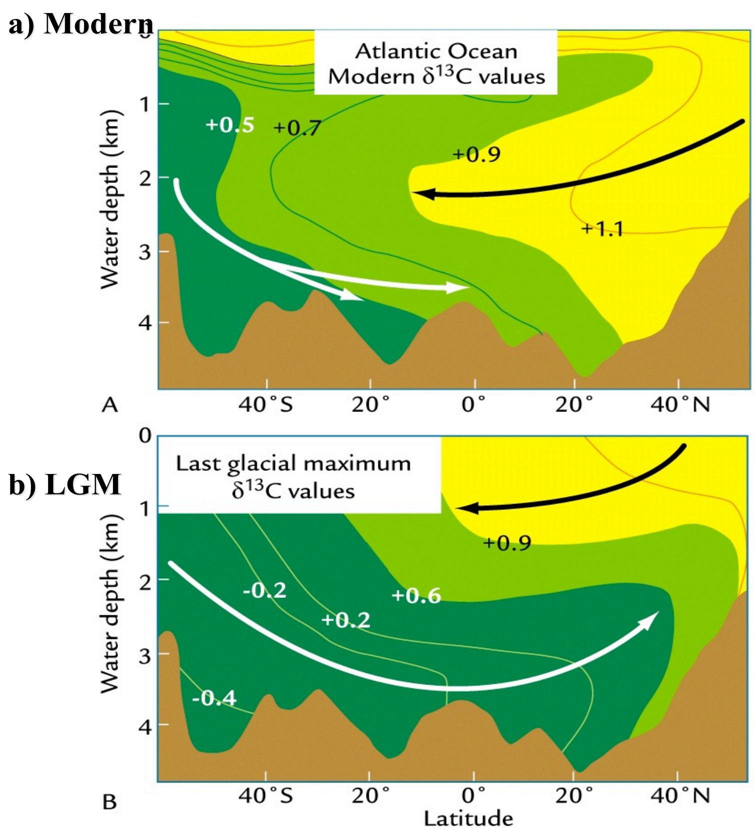

Fig. 1. δ^{13}C of present (top) and LGM (bottom) Atlantic Ocean. (adapted from Ruddiman, 2001).

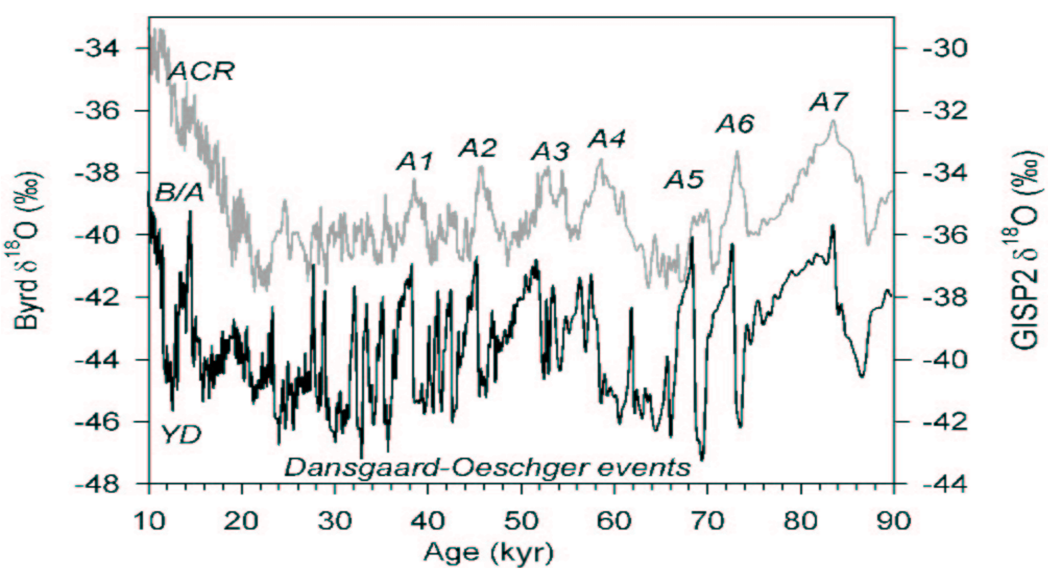

Fig. 2. Comparison of ice core paleotemperature (δ^{18}O) records from Greenland (GISP2, lower) and Antarctia (Byrd, upper), using the timescales of Blunier and Brook (2001). ACR, YD and B/A refer to Antarctic Cold Reversal, Younger Dryas and Bolling/Allerod respectively. (Adapted from Steig and Alley, 2002).

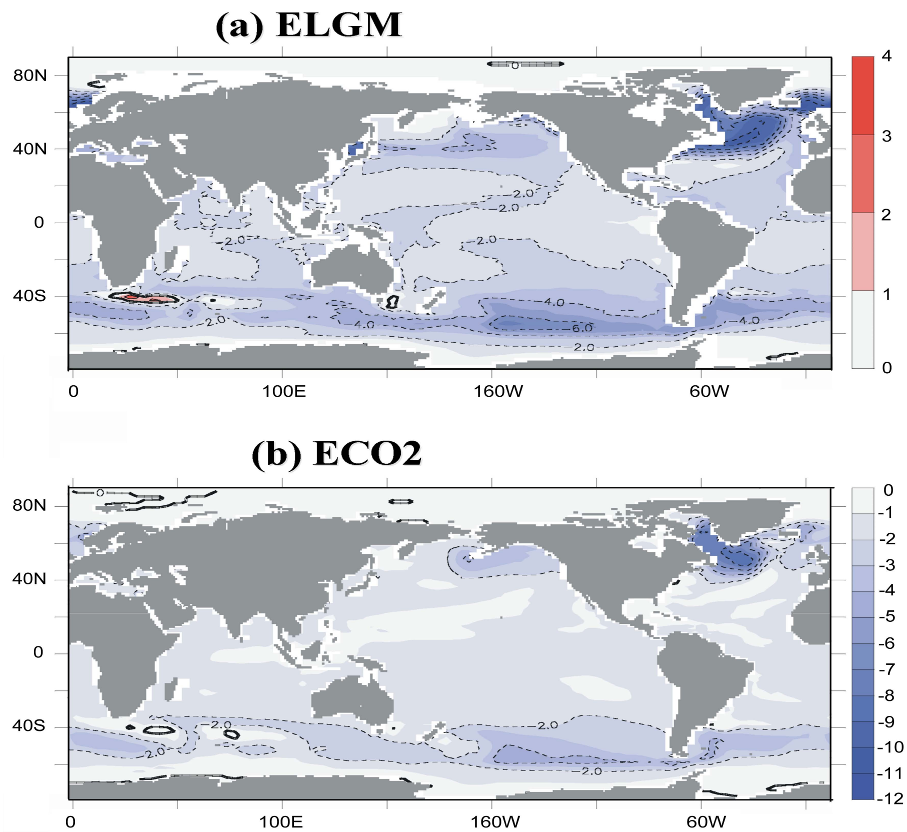

Fig. 3. Annual mean SST changes between (a) ELGM and CTRL and (b) ECO2 and CTRL. The CTRL is forced by the present climate forcing, including the present atmospheric CO_2 level. The results are for the average over the last 50 years of the 300-year simulation.

2. Glacial THC simulation and mechanism

We perform an LGM experiment (ELGM) in the NCAR-CCSM1 (Bouville and Gent, 1998; Otto-Bliesner and Brady, 2001), forced by the complete LGM climate forcing: the orbital forcing (Berger, 1978), continental ice sheet (Peltier, 1994), and the lowered atmospheric greenhouse gases, including a CO_2 of 200 ppm (Raynaud et al., 1999). * The glacial simulation reproduces the major features of the reconstructed LGM global climate (Shin et al., 2003a, b; Liu et al., 2004). The SST decreases by up to 10°C, with the maximum cooling occurring in the high latitude North Atlantic (Fig. 3a), while the sea surface salinity (SSS) increases by over 2 psu at high latitudes, predominantly around the Antarctic (Shin et al., 2003a). The simulated modern Atlantic is characterized by a deep overturning (Fig. 4a), a deep southward-directed tongue of relatively warm and saline North Atlantic Deep Water (NADW) and a weak northward tongue of fresh and cold Antarctic Bottom Water beneath (AABW). The simulated glacial Atlantic THC shallows by over 1 km with the water transport reduced by 30% (Fig. 4b). The most striking change occurs in the deep temperature/salinity field below 1000 m: the tongue of the NADW disappears completely, resulting in virtually uniform deep temperature and salinity outside polar regions (Liu et al., 2004), consistent with paleo reconstructions (Adkins et al., 2003). The simulation of the shallow LGM THC has remained a challenge for coupled GCMs. Although simulated in two coupled models of intermediate complexity (Ganoplski et al., 1998; Weaver et al., 1998), this shallower LGM THC has not been reproduced in other coupled GCMs (Hewitt et al., 2001; Kitoh et al., 2001) unless a substantial discharge of melt water into the North Atlantic

*ELGM is integrated with deep-water acceleration for 300 surface years. The final LGM state seems to be insensitive to initial conditions, because they converge towards another LGM simulation that is initiated from a glacial ocean state forced by the CLIMAP (CLIMAP, 1981) SST, but without deep ocean acceleration (for details see Liu et al., 2002; Shin et al., 2003a, b; Liu et al., 2004).

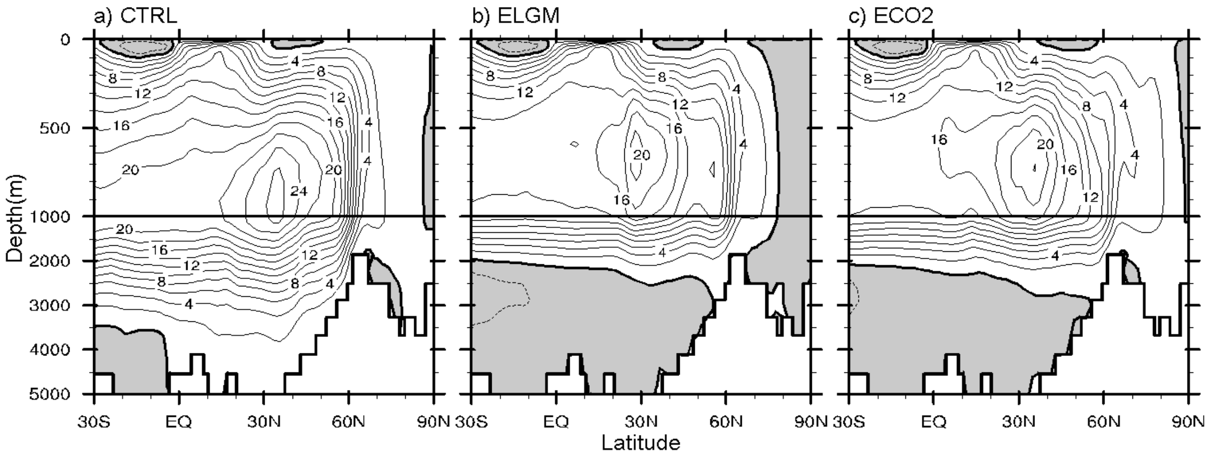

Fig. 4. North Atlantic overturning streamfunction in (a) CTRL, (b) ELGM and (c) ECO2. The contour interval is 2 Sv. (1 Sv=10^6 m^3 s^{-1}). (Adapted from Liu et al., 2004).

is imposed (Kim et al., 2003). The success of the CCSM1 simulation of LGM THC therefore seems to agree better with the available evidence[†].

Shin et al. (2003b) proposed that the shallow THC at LGM is caused predominantly by a stronger Southern Ocean sea-ice sensitivity. In response to the LGM cooling forcing, in the initial 100 surface years, SST decreases rapidly in the North Atlantic, but slowly in the Southern Ocean due to the buffer effect on SST cooling associated with deep mixing there. Later, the cooling slows down significantly in the North Atlantic, but continues to intensify in the Southern Ocean because of the increased surface albedo associated with the expanded sea-ice there, which is caused by a stronger sensitivity of the sea ice cover, and in turn ice albedo feedback in the Southern Ocean. The deep mixing over the Southern Ocean limits the thickness of the sea ice to about 1 m. This thin ice coverage over the Southern Ocean is much more sensitive to climate forcing than the thick sea ice over the North Atlantic (Stouffer, 2004). The enhanced sea-ice formation leads to a greater brine injection and, in turn, a denser buoyancy flux into the Southern Ocean, forming a stronger AABW. This denser AABW penetrates into the North Atlantic along the ocean bottom, shallowing the North Atlantic THC. This coupled experiment suggests that, at glacial timescales, the Southern Ocean climate forcing is able to overwhelm the North Atlantic forcing to dominate the THC, consistent with a recent OGCM study (Knorr and Lohmann, 2003). This Southern Ocean control effect is related to the fact that the present THC is controlled predominantly by the pole-to-pole density gradient, rather than by the pole-to-equator density gradient (Rooth, 1982; Hughes and Weaver, 1994; Marotzke and Klinger, 2000).

The glacial THC change is further found to be forced predominantly by the reduction of the atmospheric CO_2. This is seen in a CO_2 sensitivity experiment (ECO2) that is identical to ELGM except that only the LGM CO_2 forcing is applied. ECO2 is found to capture most of the changes of the ELGM simulation including SST (Fig. 3b), SSS, THC (Fig. 4c) and deep water properties (Liu et al., 2004). The dominant control by the atmospheric CO_2 on the Southern Ocean at glacial timescales, combined with the potential of a strong Southern Ocean control on glacial atmospheric CO_2 (Sarmiento and Toggweiler, 1984; Archer et al., 2000), suggests an important positive feedback between global climate and the carbon cycle during the glacial/interglacial cycle. The cooling induced by a reduction of atmospheric CO_2 increases the sea-ice cover and surface ocean stratification over the Southern Ocean (Stouffer and Manabe, 2003), which in turn work to further lower the atmospheric CO_2 by preventing gas exchange (Stephens and Keeling, 2002) and reducing the carbon flux from the deep ocean to the surface (Francois et al., 1998; Toggweiler, 1998). This positive feedback can be further enhanced by other processes, such as the enhanced iron-rich dust input from the drier climate (Martin, 1990), and contribute to the amplification of glacial cycles.

3. Implication to interhemispheric climate interactions

The view of the Southern Ocean forcing of the

[†]This version of CCSM used a simple precipitation-scaling scheme instead of river runoff, to close the hydrological cycle. However, in the most recent version CCSM3 that does use explicit river runoff, the LGM thermohaline exhibits similar behaviour (B. Otto-Bliesner, personal communication, 2005).

glacial THC (Weaver et al., 2002; Shin et al., 2003b; Knorr and Lohmann, 2003) differs from earlier studies which tend to emphasize the role of the fresh water forcing over the North Atlantic Ocean (Stocker, 2000; Kim et al., 2003; Alley et al., 2002; Schmittner et al., 2003). This Southern Ocean view appears to be consistent with recent paleo evidence that shows a lead of 1000–2000 years of the Antarctic temperature over Greenland temperature (Blunier and Brook, 2001) at orbital timescales (Steig and Alley, 2003; Wunsch, 2003). More generally, our LGM simulations is consistent with paleoclimatic evidence of glacial cycles, which shows an evolution of atmospheric CO_2 in phase with the air temperatures in the Antarctic and Tropics, but leading the Greenland air temperature and continental ice volume by a few thousand years (Shackleton, 2000; Visser et al., 2003).

The dominant Southern Ocean forcing of the North Atlantic THC is interesting, because the climate forcing, dominated by the CO_2 forcing, is largely uniform over the globe. As pointed out above (Shin et al., 2003b), the thin sea ice, and the resulting strong albedo feedback, is the key factor that empowers the Southern Ocean as the dominant (salinity forcing) climate forcing that drives the North Atlantic THC through interhemispheric climate interaction. However, this Southern Ocean forcing mechanism, we believe, is inefficient at shorter timescales comparable to millennia, because the full response of the Southern Ocean takes thousands of years. This response time when combined with that of the carbon cycle may become even longer. At millennial timescales, the Southern Ocean mechanism is likely to be overwhelmed by the climate forcing over the North Atlantic, which has a rapid response of multi-decadal to centennial timescales, due to the presence of the meridional boundary of the ocean basin, which establishes a shallow thermocline and therefore prevents a deep mixing. Furthermore, this boundary allows a fast communication from the North Atlantic high latitudes towards lower latitudes, and in turn a rapid response of the North Atlantic THC. With a North Atlantic climate forcing, the oceanic signal penetrates rapidly as a Kelvin wave first equatorward along the western boundary, then eastward along the equator and finally northward along the eastern boundary. Along the eastern boundary, most of the energy leaks westward as planetary Rossby waves (Johnson and Marshall, 2002), whose cross-basin propagation establishes the THC. In contrast, the absence of a meridional boundary in the upper 2 km of the Southern Ocean enables the Antarctic Circumpolar Current (ACC) to extend deep into the Southern Ocean and mix strongly with the deep waters there. Furthermore, the equatorward propagation of climate signals across the ACC is also very slow, because it has to rely on eddy mixing rather than Kelvin waves. All these lead to a strong difference between the response times to the Southern Ocean and North Atlantic forcing. This timescale disparity can be seen explicitly in the recent fully-coupled GCM study of Stouffer (2004) (Fig. 5). This study shows that, under a CO_2 climate forcing, the hemispheric surface air temperature reaches equilibrium rapidly in hundreds of years in the Northern Hemisphere, but slowly in thousands of years in the Southern Hemisphere.

The disparity of the response times implies that the North Atlantic THC may react fully to a North Atlantic climate forcing at millennial or shorter timescales, but may need thousands of years or longer to react fully to a Southern Ocean climate forcing. This leads us to propose that the North Atlantic climate forcing is important for the THC at millennial or shorter timescales. Only at the much longer orbital timescales, can the THC fully react to a Southern Ocean climate forcing. The full response to the Southern Ocean, combined with the dominant control of the Southern Ocean on the carbon cycle, excites a strong positive feedback at glacial/interglacial timescales, overwhelming the North Atlantic climate forcing in the glacial/interglacial climate change.

This proposal seems to be consistent with the available observations. Figure 6 (Steig and Alley, 2003) shows the lagged correlation of the $\delta^{18}O$ records between the Antarctic and Greenland. To illustrate the dependency of the correlation with timescale, the lagged correlation is done after the high-pass filtering. It is seen that at the glacial/interglacial timescales

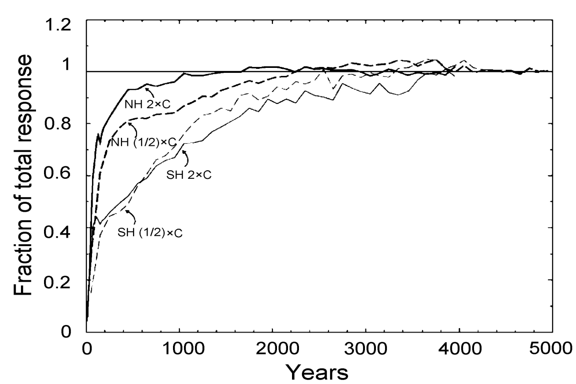

Fig. 5. Time series of the fraction of the total equilibrium, hemispheric mean surface air temperature response (thick lines: Northern Hemisphere values, thin lines: Southern Hemisphere). The solid lines are values from the $2\times CO_2$ integration and the dashed lines are from the $(1/2)\times CO_2$ integration. The experiments are performed using the GFDL low resolution fully coupled ocean-atmosphere model (adapted from Stouffer, 2004).

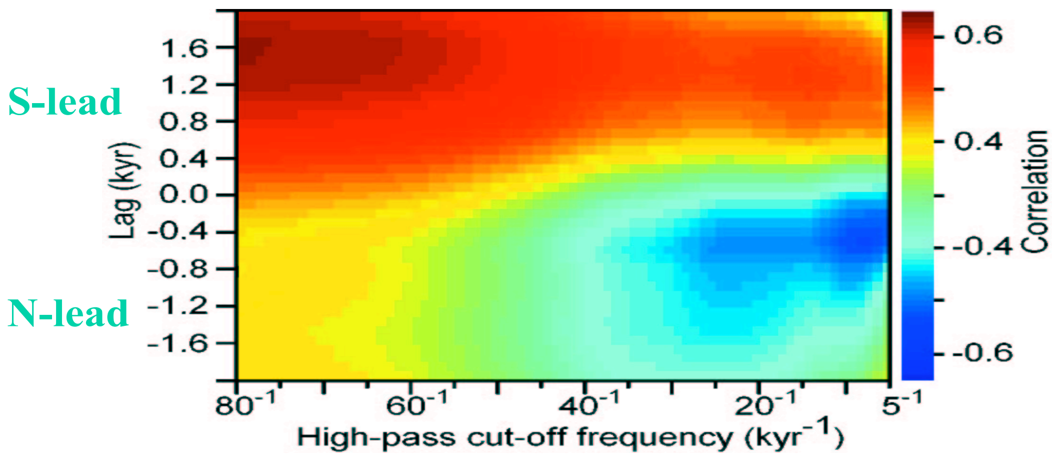

Fig. 6. Lag correlation space for the Byrd and GISP2 δ^{18}O records, showing correlation coefficients of high-pass filtered data as a function of high-pass cutoff frequency, 1/80 kyr^{-1} to 1/5 kyr^{-1} (adapted from Steig and Alley, 2003).

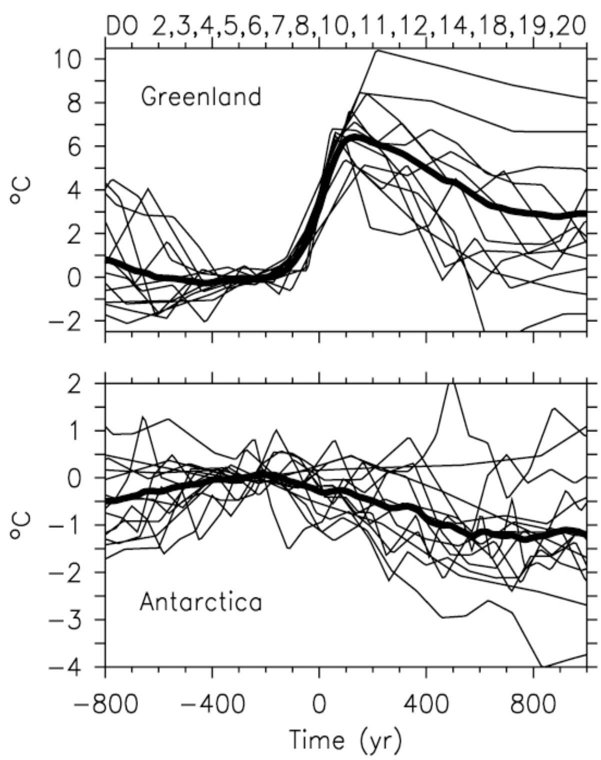

Fig. 7. Reconstructed surface air temperature changes in the northern (upper) and southern (lower) hemispheres for individual DO events (thin lines) and their composite (thick lines) for warming events using δ^{18}O from the ice cores of GISP2 and Byrd. The rapid warming of the Greenland temperature and the slow cooling of the Antarctic temperature generates a lagged correlation similar to that at the high frequency limit in Fig. 6. (adapted from Schmitner et al., 2003).

(longer than 10,000 years), a single maximum correlation is achieved with the Antarctic leading by about 1.6 kyr. This lead time has been confirmed to be statistically significant by Wunsch (2003), but only for variability of timescales much longer than millennia. This suggests that for glacial/interglacial changes, the Antarctic clearly leads Greenland and therefore is likely to be a more important forcing. In contrast, towards the high frequency end of millennial timescales for Heinrich events and Dansgaard/Oeschger events, the lagged correlation shows a much more complex structure. It appears that the maximum positive correlation is achieved with the Antarctic leading by 1200 years, while an equally strong (negative) maximum correlation is achieved with Greenland leading by about 500 years. [These correlations, however, appear to be statistically insignificant in the coherence analysis of Wunsch (2003)]. This correlation structure, according to the modelling study and composite data analysis of Schmittner et al. (2003), can be interpreted as a Northern lead of an anti-phase Southern ("seesaw") response, with the former varying much more rapidly than the latter, as shown in the composite of Dansgaard/Oeschger events in Fig. 7 (Schmittner et al., 2003). Therefore, for high frequency variability, the North Atlantic dominates the THC, with a close coupling with the Southern Ocean through the interhemispheric climate interaction of the bipolar seesaw (Stocker, 2000; Schmittner et al., 2003).

4. Summary

Our coupled GCM simulations suggest that the glacial THC is likely to be caused by the lower atmospheric CO_2, which, at the orbital timescale, can drive a full response of the Southern Ocean, which in turn drives the North Atlantic THC and results in a lead time of 1000–2000 years between the Antarctic and Greenland temperatures. This full Southern Ocean response could also be coupled with the marine carbon cycle feedback, because the carbon cycle is determined predominantly by the Southern Ocean and therefore may also have a timescale comparable with the Southern Ocean response. However, only at timescales longer than the Southern Ocean response can the CO_2 feedback and Southern Ocean driving become fully functional. Therefore, for fast millennial variability, the Southern Ocean forcing mechanism is overwhelmed by the North Atlantic climate forcing. Hence, our proposal, which is based on the different response times of the Southern Ocean and North Atlantic, may help to clarify previous debates on the interhemispheric interaction, providing a physically more comprehensive view of the dynamic response of the THC variability. This comprehensive view, however, remains speculative at this stage and will need to be substantiated with many further studies.

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