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Southern Ocean Overturning Role in Modulating High Southern Latitude Climate and Atmospheric CO_2 on Millennial Timescales

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With 3 Figures

1. Introduction

During the last glacial period and particularly Marine Isotope Stage 3, North Atlantic Deep Water (NADW) formation weakened significantly on a millennial timescale leading to Greenland stadials (Kissel et al. 2008). Ice core records reveal that each Greenland stadial is associated with a warming over Antarctica, so-called Antarctic Isotope Maximum (AIM) (*EPICA and Community Members* 2006). Recent ice core records further suggest that atmospheric CO_2 increased with Antarctic temperature only during the long Greenland stadials, i.e. the Heinrich stadials (Ahn and Brook 2014).

Under present day conditions, the formation of NADW leads to an equatorward ocean heat transport in the South Atlantic. Previous studies (Stocker 1998) suggested that the bipolar seesaw pattern in surface air temperature during AIM was due to a heat redistribution in the Atlantic basin: as NADW weakens its associated northward heat transport also reduces. However, idealized experiments performed with coupled Atmosphere-Ocean General Circulation Models featuring a significant NADW weakening only display a small warming (0.5–1 °C) over Antarctica if any (Kageyama et al. 2012), in contrast with estimates for the large AIM (e.g. AIM12 and AIM8).

Antarctic Bottom Waters (AABW), presently formed under sea-ice on the Antarctic continental shelf, are an integral part of the global oceanic circulation and make significant contribution to deep ocean ventilation (Orsi et al. 2002). Due to difficulties in estimating past Southern Ocean ventilation, possible changes in AABW during the last glacial period and the deglaciation have received little attention.

Performing transient simulations of MIS3 with prognostic atmospheric CO_2 and comparing our results with paleoproxy records we suggest that enhanced AABW transport could have played a significant role in shaping the large AIM and the associated atmospheric CO_2 increase (Menviel et al. 2015). High resolution simulations of the Antarctic ice sheet further show that weakened AABW during the Antarctic Cold Reversal could have accelerated the retreat of the West Antarctic Ice Sheet (Golledge et al. 2014).

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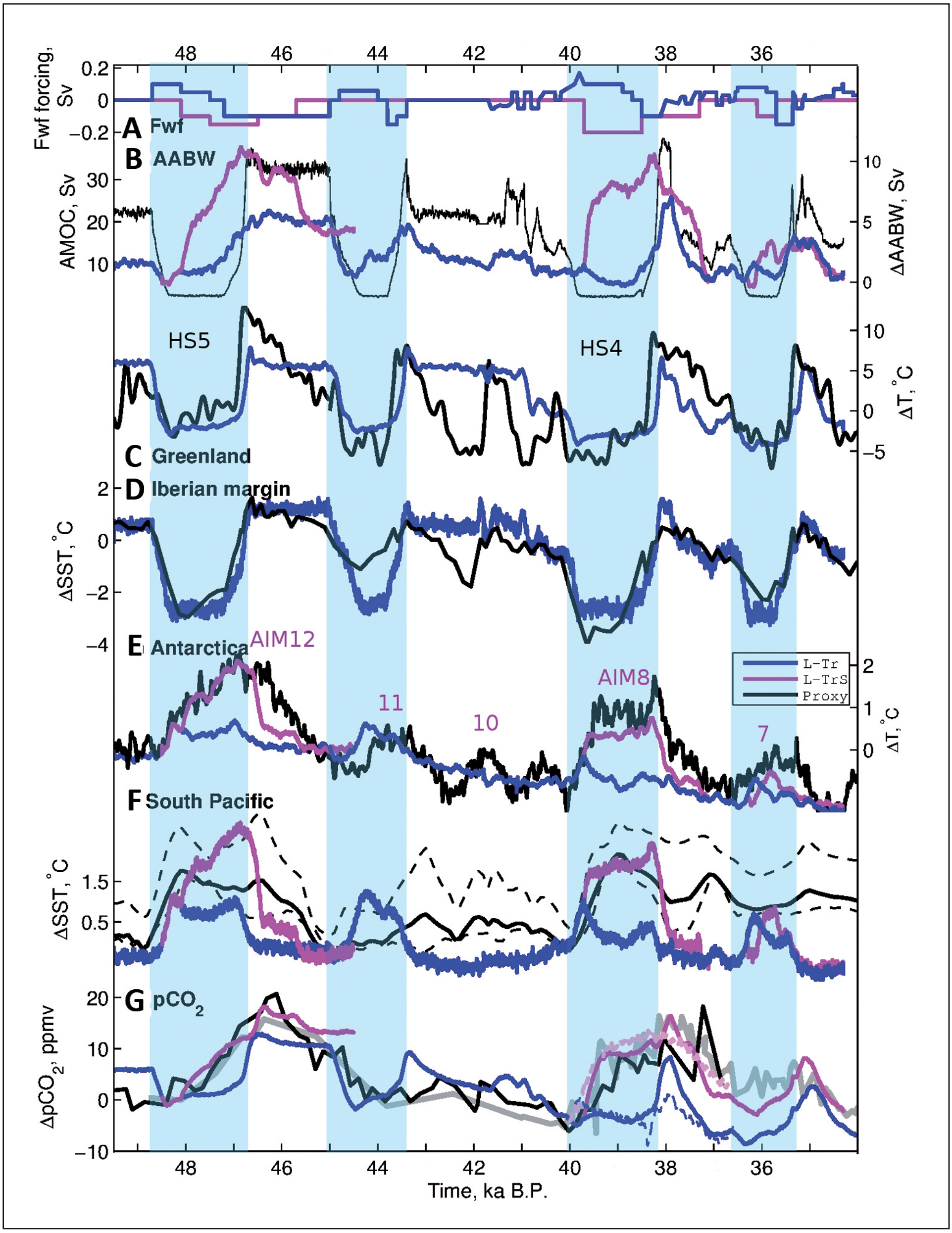

Fig. 1 Results of transient MIS3 simulations performed with LOVECLIM. Standard transient simulation (L-Tr, blue) and transient simulation with enhanced AABW (L-TrS, magenta) performed with LOVECLIM (Menviel et al. 2015) compared to paleoproxy records (black). Timeseries of (*A*) applied North Atlantic (blue) and Southern Ocean (magenta) freshwater forcing (Sv); (*B*) simulated maximum meridional overturning circulation in the North Atlantic (Sv, black) and simulated changes in Antarctic Bottom Water (AABW, Sv); (*C*) simulated NE Greenland air temperature anomalies (40ºW–10ºE, 66ºN–85ºN) compared to the NGRIP temperature reconstruction (Huber et al. 2006) on the GICC05 timescale (Svensson et al. 2008); (*D*) simulated SST anomalies (ºC) off the Iberian

2. Methodology

We perform transient simulations spanning the period 50–34 ka BP with two Earth System Models (LOVECLIM and the UVic ESCM) to understand the possible link between changes in NADW, changes in high latitude Southern Hemispheric climate and evolution of atmospheric CO_2.

The transient experiments are run with continuously varying orbital and ice sheet forcing (topography and albedo) and with prognostic atmospheric CO_2. As both models do not include an interactive ice sheet, the impact on oceanic circulation of the time-evolution of ice sheet growth and freshwater release is simulated by applying a freshwater forcing to the North Atlantic region.

Moderate changes in the mid/high Southern latitudes hydrological cycle can significantly impact surface salinity. As the atmospheric models used here are simple, it is thus possible that our standard experiments might not capture the past variability of AABW accurately. Additional transient experiments are performed, in which a salt flux is added over the Southern Ocean during Heinrich stadials 5 and 4, corresponding to AIM12 and AIM8 respectively.

Given the importance of mesoscale eddies for meridional heat transport in the Southern Ocean, we further investigate the relationship between enhanced AABW formation and SST changes using a global eddy-permitting ocean sea-ice model (GFDL-MOM025). From a present-day control run, a 40 years long simulation is performed with 0.5 psu increase in the sea surface salinity restoring climatology within 4º latitude of the Antarctic coastline.

3. Impact of AABW Changes during Heinrich Stadials 5 and 4

The simulated NADW weakening during MIS3 yield a 9 ºC surface air temperature decrease over Greenland and a ~3 ºC sea surface cooling off the Iberian margin (Fig. 1) during Heinrich stadials, in good agreement with paleoproxy records (Huber et al. 2006, Martrat et al. 2007, Menviel et al. 2014a).

However, we find that changes in NADW alone are not sufficient to explain the temperature anomaly estimated from Antarctic ice cores during the largest AIM (namely AIM12 and AIM8). In both models, NADW cessation leads to ~0.6 ºC air temperature increase over Antarctica, which is in reasonable agreement with Antarctic temperature anomaly estimates for AIM10 and AIM7 but is less than half of the estimated anomaly for the large AIM (i.e. AIM12 and AIM8) (Fig. 1).

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margin (15ºW–8ºW, 37ºN–43ºN) compared to alkenone-based SST anomalies from marine sediment core MD01-2444 (Martrat et al. 2007) on the GICC05 timescale; (*E*) air temperature anomalies (ºC) averaged over Antarctica (90ºS–75ºS) compared to temperature anomaly estimates from EPICA Dome C ice core (Jouzel et al. 2007) on the EDC3 time-scale (Parrenin et al. 2007); (*F*) SST anomalies averaged over the South Pacific Ocean (120ºE–285ºE, 55ºS–35ºS) compared to a SST anomaly composite from South Pacific marine sediment cores (Pahnke et al. 2003, Kaiser et al. 2005, Caniupan et al. 2011, Lopes Dos Santos et al. 2013). Dashed black lines represent the +1 and –1 standard deviation of the composite; (*G*) simulated atmospheric CO_2 (ppmv) anomalies compared to pCO_2 anomalies measured in EDML and Talos Dome ice cores (Bereiter et al. 2012, black) and Siple Dome ice core (Ahn and Brook 2014, grey). Dashed blue and magenta lines represent experiments respectively similar to L-Tr and L-TrS but with fully coupled changes in terrestrial carbon.

Stronger AABW during AIM12 and AIM8 doubles the simulated warming at high Southern latitude thus leading to ~1.5 °C temperature increase over Antarctica in better agreement with paleoproxy records (Fig. 1, Menviel et al. 2015).

The robustness of this southern warming response is tested using an eddy-permitting coupled ocean sea-ice model (GFDL-MOM025). We find that stronger Antarctic Bottom Water formation contributes to Southern Ocean surface warming by increasing the Southern Ocean meridional heat transport (Menviel et al. 2015).

In the standard MIS3 transient experiments, NADW weakening decreases the ventilation in the Atlantic, thus increasing the Dissolved Inorganic Carbon (DIC) content in that basin (Fig. 2). A strengthening of North Pacific Deep Water (NPDW) increases the ventilation of the Pacific above ~2000m depth, which decreases the DIC content in that region. The net

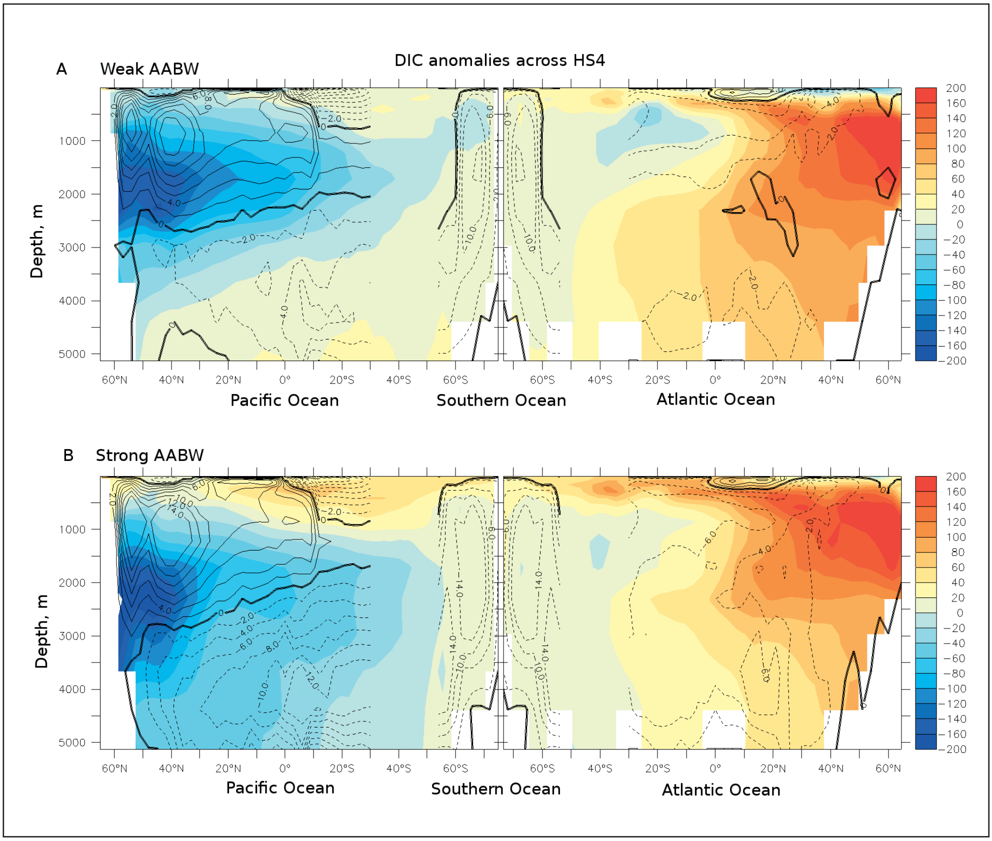

Fig. 2 Dissolved Inorganic Carbon (DIC) anomalies (μmol/L) across HS4 (39.1 ka BP compared to 39.9 ka BP) as simulated in transient experiments performed with LOVECLIM (Menviel et al. 2015) and averaged over (*left*) the Pacific and (*right*) Atlantic basins. (*A*) Results of standard transient experiment (L-Tr) and (*B*) results of transient experiment with enhanced AABW (L-TrS). Overlaid is the overturning streamfunction (Sv).

result is a slight atmospheric CO_2 decrease during a shutdown of NADW formation. This is in contrast with the atmospheric CO_2 increases observed during the first parts of AIM12 (~47.6 ka BP) and AIM8 (~39.8 ka BP), which occur during periods of weak NADW formation (HS5 and HS4, respectively).

Transient experiments in which AABW is enhanced display a ~13 ppmv atmospheric CO_2 increase during HS5 and HS4 in better agreement with ice core records. Enhanced AABW formation is shown to effectively ventilate the deep Pacific carbon, bringing DIC rich waters to the surface of the Southern Ocean and leading to CO_2 outgassing into the atmosphere (Fig. 2, Menviel et al. 2014b).

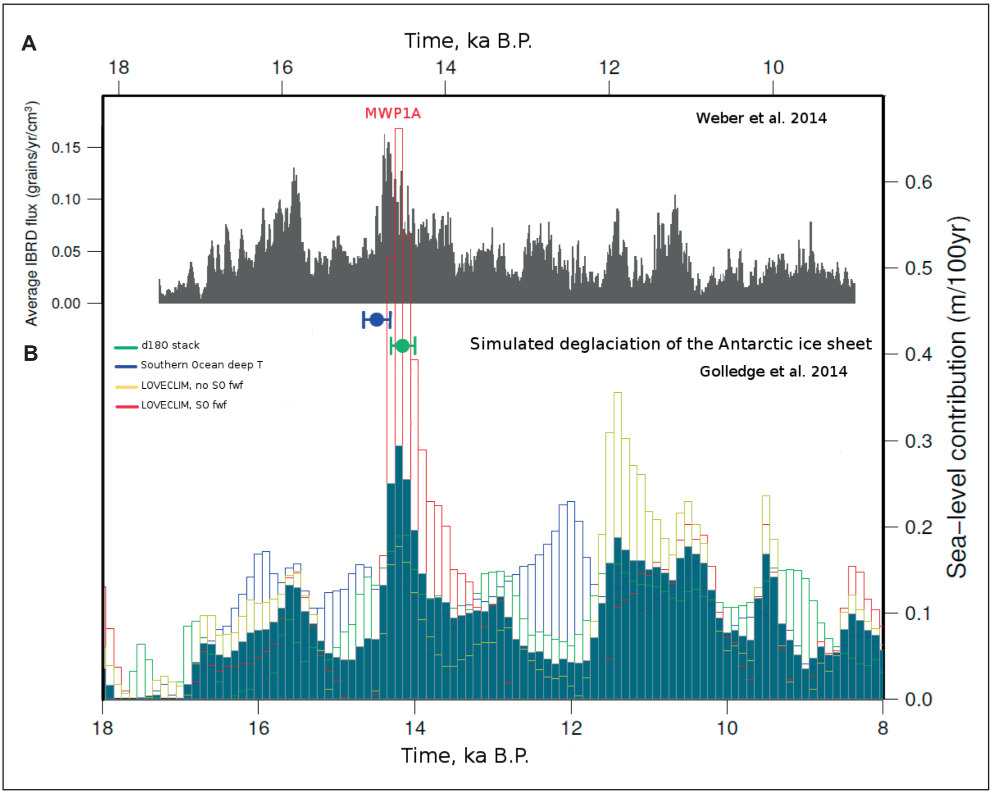

Fig. 3 (*A*) Stack of iceberg-rafted debris flux measured in marine sediment cores from the Scotia Sea (Weber et al. 2014) compared to (*B*) Sea-level contribution through the period 18–8 ka BP, binned at 100-year intervals simulated by a high resolution Antarctic Ice Sheet model for group means and ensemble mean (solid grey). Experiments are forced with ocean heat fluxes based on a benthic $\delta^{18}O$ global stack (green), a Southern Ocean deep temperature record (blue), a transient simulation of the deglaciation (yellow) and a transient simulation of the deglaciation which includes weak AABW formation at the time of the Antarctic Cold Reversal (red) (Golledge et al. 2014). The blue and green circles represent the timing of meltwater pulses 1*A* as estimated by Deschamp et al. 2012 and Lui et al. 2004.

4. AABW Changes during the Last Deglacation

Transient simulations of the last deglaciation performed with LOVECLIM have shown that a weakening of AABW formation at the end of Heinrich stadial 1 would lead to cooling over Antarctica and at the surface of the Southern Ocean, thus initiating the Antarctic Cold Reversal (Menviel et al. 2011). Increased Southern Ocean stratification during times of weak AABW induces a subsurface warming in the Southern Ocean (Menviel et al. 2010).

Using a high resolution Antarctic ice sheet model (PISM), we show that this subsurface warming can thermally erode grounded marine-based ice and instigate a positive feedback that further accelerate the ice-sheet retreat (Golledge et al. 2014). Coincident with iceberg-rafted debris records from the Scotia Sea (Weber et al. 2014, Fig. 3), we simulate up to 4 m rise in sea level equivalent during the broad period of meltwater pulse 1A (15–13 ka BP), mainly due to a retreat of the West Antarctic Ice Sheet in the Weddell Sea and Antarctic Peninsula sectors.

5. Summary

We suggest that strong Antarctic Bottom Water formation during HS5 and HS4 could enhance the bipolar seesaw effect and lead to a warming of Antarctica and the Southern Ocean in better agreement with paleoproxy records. In addition strong AABW can effectively ventilate the deep Pacific Ocean and release oceanic carbon into the atmosphere, thus leading to atmospheric CO_2 increase during HS5 and HS4 (AIM12 and AIM8, respectively, Menviel et al. 2015).

Conversely, weak AABW formation between 14.8 and 13 ka BP would lead to a cooling at high Southern latitudes, which could explain the Antarctic Cold Reversal. The associated Southern Ocean stratification induces a subsurface warming, which leads to an accelerated deglacial retreat of the West Antarctic ice sheet (Golledge et al. 2014).

Enhanced AABW formation during Greenland stadials could be due to changes in surface buoyancy forcing over the Southern Ocean and/or to stronger/poleward shifted southern hemispheric westerlies. The amplitude of the Greenland stadial, the origin and timing of the meltwater pulse/iceberg discharge as well as the background climatic conditions might further modulate the AABW response.

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