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Contribution of enhanced Antarctic Bottom Water formation to Antarctic warm events and millennial-scale atmospheric CO_2 increase

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ABSTRACT

During Marine Isotope Stage 3, the Atlantic Meridional Overturning Circulation (AMOC) weakened significantly on a millennial time-scale leading to Greenland stadials. Ice core records reveal that each Greenland stadial is associated with a warming over Antarctica, so-called Antarctic Isotope Maximum (AIM), and that atmospheric CO_2 increases with Antarctic temperature during the long Greenland stadials. Here we perform transient simulations spanning the period 50–34 ka B.P. with two Earth System Models (LOVECLIM and the UVic ESCM) to understand the possible link between changes in the AMOC, changes in high latitude Southern Hemispheric climate and evolution of atmospheric CO_2. We find that oceanic carbon releases due to the AMOC resumption during stadial/interstadial transitions lead to an atmospheric CO_2 increase. However, the atmospheric CO_2 increases observed during the first parts of AIM12 (∼47.6 ka B.P.) and AIM8 (∼39.8 ka B.P.) occur during periods of weak AMOC (HS5 and HS4 respectively) and could instead be explained by enhanced Antarctic Bottom Water transport. Enhanced Antarctic Bottom Water formation is shown to effectively ventilate the deep Pacific carbon and lead to CO_2 outgassing into the atmosphere. In addition, changes in the AMOC alone are not sufficient to explain the largest Antarctic Isotope Maxima (namely AIM12 and AIM8). Stronger formation of Antarctic Bottom Water during AIM12 and AIM8 would enhance the southern high latitude warming and lead to a better agreement with high southern latitude paleoproxy records. The robustness of this southern warming response is tested using an eddy-permitting coupled ocean sea–ice model. We show that stronger Antarctic Bottom Water formation contributes to Southern Ocean surface warming by increasing the Southern Ocean meridional heat transport.

1. Introduction

During Marine Isotope Stage 3 (MIS3, 59.4–27.8 ka B.P.), Greenland and North Atlantic climate oscillated between stadial and interstadial conditions on a millennial time-scale (Masson-Delmotte et al., 2013), in what are called Dansgaard–Oeschger (DO) cycles. DO cycles are characterized by an abrupt warming of 8 to 15 °C within a few decades in Greenland (Dansgaard et al., 1993; Huber et al., 2006), followed by a gradual cooling and then a Greenland stadial. During some Greenland stadials, named Heinrich stadials (Sánchez-Goñi and Harrison, 2010), thick layers of ice-rafted debris are found in marine sediment cores from the North Atlantic (e.g. Hemming, 2004). Paleoproxy records and modeling studies have suggested that DO cycles and Heinrich stadials were part of a continuum of variability that is generated through ice sheet-driven changes in the Atlantic Meridional Overturning Circulation (AMOC) (e.g. Elliot et al., 1998, 2002; Grousset et al., 2001; Sarnthein et al., 2001; Timmermann et al., 2003; Dokken et al., 2013; Menviel et al., 2014b).

During Greenland stadials cold and dry conditions prevailed over Greenland (Alley, 2000; Huber et al., 2006), the North Atlantic (e.g. Bard, 2002; Martrat et al., 2007) and Europe (e.g. Sánchez-Goñi et al., 2002; Fleitmann et al., 2009; Harrison and Sánchez Goñi, 2010). Concurrently, Antarctic ice cores have revealed that during each Greenland stadial of MIS3, air temperature over Antarctica rose by 1 to 3 °C (EPICA and community members, 2006), in what are called Antarctic Isotope Maxima (AIM)[1] (EPICA and community members, 2006). Marine sediment cores from the Southern Ocean further suggest that the surface of the Southern

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[1] For simplicity, in this manuscript, AIM denotes both the singular maximum and plural maxima.

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Ocean warmed during the large AIM events (Pahnke et al., 2003; Crosta et al., 2004; Kaiser et al., 2005; Caniupán et al., 2011; Lopes dos Santos et al., 2013).

In contrast to other subtropical basins, under present day conditions the ocean heat transport in the South Atlantic is equatorward due to the presence of a strong AMOC (Perez et al., 2011). Previous studies (Crowley, 1992; Stocker, 1998) suggested that the bipolar seesaw pattern in surface air temperature during AIM was due to heat redistribution in the Atlantic basin: as the AMOC weakens its associated northward heat transport also reduces. In addition, while atmospheric CO_2 increased during the large AIM (e.g. AIM12 and AIM8), little atmospheric CO_2 changes were observed during the small amplitude AIM (e.g. AIM10) (Ahn and Brook, 2008, 2014; Bereiter et al., 2012). It has thus been widely accepted that the Southern Hemispheric warming observed during Greenland stadials was the result of meridional heat transport reorganization due to AMOC changes and due to atmospheric CO_2 increase (e.g. Timmermann et al., 2010).

Idealized experiments featuring a shutdown of the AMOC under constant Last Glacial Maximum boundary conditions have been performed with several coupled Atmosphere–Ocean General Circulation Models. However, not all models display a warming signal over Antarctica and the Southern Ocean (Kageyama et al., 2013) concomitant with an AMOC shutdown. Furthermore, when a surface warming is simulated over Antarctica it is of the order of $\sim 1\,^{\circ}C$ (Kageyama et al., 2009; Otto-Bliesner and Brady, 2010; Buiron et al., 2012), which is less than temperature anomaly estimates from Antarctic ice cores for the large AIM (e.g. AIM12 and AIM8) (Brook et al., 2005; Jouzel et al., 2007; Uemura et al., 2012). Finally, it is questionable whether the bipolar seesaw effect (i.e. meridional heat transport reorganization) can explain the duration of the warming observed over Antarctica during AIM12, which is over 1000 yr longer than Heinrich stadial 5 (HS5).

Changes in atmospheric CO_2 observed during MIS3 have been tentatively explained by previous modeling experiments. While some studies suggest that the carbon source to the atmosphere is of terrestrial origin (Scholze et al., 2003; Köhler et al., 2005; Obata, 2007; Menviel et al., 2008; Bozbiyik et al., 2011) others attribute it to an oceanic carbon release (Marchal et al., 1999; Schmittner et al., 2007; Schmittner and Galbraith, 2008; Bouttes et al., 2012; Matsumoto and Yokoyama, 2013). However these studies were very idealized as they were performed from either constant Last Glacial Maximum or pre-industrial conditions and used an idealized freshwater forcing.

Significant uncertainties remain regarding the mechanism leading to the large AIM and their relationship with changes in atmospheric CO_2. Menviel et al. (2014a) recently suggested that a strengthening of Antarctic Bottom Water (AABW) transport during Heinrich stadial 4 (HS4) could release deep Pacific Ocean carbon and lead to an atmospheric CO_2 increase in general agreement with the one recorded in Siple Dome ice core at ∼39.8 ka B.P. (Ahn and Brook, 2014). Here we extend this hypothesis and show that enhanced AABW transport during Heinrich stadials could explain part of the observed changes in high Southern latitude temperature and atmospheric CO_2.

The goal of this study is to better understand the origin of the large AIM, the associated changes in atmospheric CO_2 and their relationship with Greenland stadials. We simulate the sequence of high southern latitude millennial-scale events as well as the atmospheric CO_2 variations during the period 50–34 ka B.P. with two Earth System Models of Intermediate Complexity (LOVECLIM and the UVic ESCM). We use two different Earth System Models to test the robustness of the mechanism leading to high southern latitude warming and changes in atmospheric CO_2.

Given that most of the observed ocean kinetic energy occurs on the mesoscale (10–100 km length scales) (Wunsch, 2007), there is considerable debate surrounding the adequacy of ocean meridional heat transport simulated at coarse resolution, particularly in the Southern Ocean where mesoscale eddies are ubiquitous (Spence et al., 2012; Morrison et al., 2013; Bryan et al., 2014). Therefore, we further investigate the relationship between enhanced AABW formation and SST changes using a global eddy-permitting ocean sea–ice model.

2. Models and experimental setup

2.1. Transient simulations of MIS3 performed with LOVECLIM and the UVic ESCM

LOVECLIM consists of a free surface ocean general circulation model (CLIO) with a horizontal resolution of 3° longitude, 3° latitude and 20 depth layers (Goosse et al., 2010). The atmospheric component (ECBilt) is a spectral T21, three-level model based on quasi-geostrophic equations of motion and ageostrophic corrections. LOVECLIM also includes a dynamic–thermodynamic sea–ice model, a land surface scheme, a dynamic global vegetation model (VECODE) (Brovkin et al., 1997) and a marine carbon cycle model (LOCH) (Mouchet, 2011; Menviel et al., 2008).

The UVic Earth System Climate Model (UVic ESCM v2.9) (Weaver et al., 2001) consists of an ocean general circulation model (Modular Ocean Model, Version 2) with a resolution of 3.6° longitude and 1.8° latitude, coupled to a vertically integrated two dimensional energy-moisture balance model of the atmosphere including a parameterization of geostrophic wind stress anomalies, a dynamic–thermodynamic sea–ice model, a land surface scheme, a dynamic global vegetation model, a marine carbon cycle model (Schmittner et al., 2008). Sediment processes are represented using an oxic model of sediment respiration (Archer, 1996; Eby et al., 2009). The oceans barotropic momentum equations are solved with a rigid lid formulation and surface freshwater fluxes are converted to fluxes of salt with a constant salt to freshwater mass ratio of 3.49×10^{-2}.

Initial conditions for both LOVECLIM and the UVic ESCM were obtained by conducting a 15,000 yr equilibrium spin-up simulation using an atmospheric CO_2 content of 207.5 ppmv, orbital forcing for the time 50 ka B.P. and an estimate of the 50 ka B.P. ice sheet orography and albedo, which were obtained from a 130 ka off-line ice sheet model simulation (Abe-Ouchi et al., 2007).

In LOVECLIM the organic matter that is not remineralized as well as the carbonate and opal that are not dissolved are permanently preserved in the sediments. This leads to a loss of alkalinity, carbon, phosphates and silicates, which is compensated by the river influx of these components (Menviel et al., 2008). For the LOVECLIM equilibrium run, the alkalinity over Dissolved Inorganic Carbon (DIC) ratio in the riverine input was set to 1.6, thus allowing for a greater oceanic carbon reservoir at 50 ka B.P. than at pre-industrial time. In a subsequent 3000 yr equilibrium run atmospheric CO_2 becomes a prognostic variable and the riverine input of alkalinity and DIC then compensates for the loss due to organic matter and carbonate sedimentation. As the UVic ESCM includes a sediment model, no such riverine input adjustment is made and the oceanic carbon reservoir simulated at 50 ka B.P. is smaller than the pre-industrial one. A 3000 yr equilibrium run with prognostic atmospheric CO_2 was also run with the UVic ESCM.

Transient simulations of the period 50–34 ka B.P. (L–Tr and U–Tr) were run with continuously varying orbital and ice sheet forcing following the methodology of Timm et al. (2008), but without any acceleration, and with prognostic atmospheric CO_2. As both models do not include an interactive ice sheet, freshwater withholding from the ocean during phases of ice sheet growth and freshwater release into the ocean as a result of ice sheet calving and ablation are not explicitly captured. To mimic the time-