How important are Southern Hemisphere wind changes for low glacial carbon dioxide? A model study – supplementary material

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1. Model Setups and Spinup

Sea surface temperatures and salinities are constrained by using either restoring (RBC) or mixed boundary conditions (MBC). Restoring compels sea surface temperatures and salinities to remain close to observational values. MBC however permit surface salinities to depart from those such that circulation-salinity feedbacks are allowed to develop upon

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perturbation of the model state. Note that surface temperatures are always restored towards observations thereby largely preventing temperature-mediated feedbacks.

The driving forces of the meridional overturning circulation (MOC) are still a matter of intense investigation. Presently, two major mechanisms are discussed (*Kuhlbrodt et al.* [2007]). The first one is turbulent mixing of heat from the surface into the ocean's interior, the second is wind-driven upwelling in the Southern Ocean. In this perspective, the choice of diapycnal diffusivity κ_{dia} controls the relative importance of thermohaline versus wind-driven MOC forcing in a given model setup.

The sinking of dense salty water in the North Atlantic is an important organizing feature of the modern deep ocean circulation. Modeled deep water mass properties, ventilation time-scales and tracer distributions in different ocean basins depend quite sensitively upon the modeled rate of deep water formation in the North Atlantic. The anomalous surface freshwater flux $F_{Atl-Pac}$ permits a systematical increase (decrease) in Atlantic seasurface salinities and a strengthening (weakening) of the Atlantic MOC. This allows for the creation of a series of model states differing in the rate of the AMOC to probe the robustness of the model response to different initial states.

For every model setup a steady state has been computed in the following manner: First, the physical model is spun-up for 10'000 model years using RBC. Subsequently, air-sea freshwater-fluxes are diagnosed and averaged over the last 1000 years to provide mixed boundary conditions. The model is now switched to mixed boundary conditions and anomalous freshwater fluxes $F_{Atl-Pac}$ are included. In the case of the setups that employ restoring by definition RBC are maintained at this point and no additional freshwater flux $F_{Atl-Pac}$ is applied. Then, a biogeochemical spinup phase with a fixed atmospheric CO_2 mixing ratio of 278 ppm is run for another 10'000 years. The resulting state is finally perturbed by instantaneously switching from the standard to a modified windstress scenario and run for further 5000 model years to achieve a new steady state. During this last step, atmospheric CO_2 is allowed to evolve freely. Annual means are used for the analysis.

2. Model Representation of Additional Mechanisms

2.1. CO_2 solubility

To estimate the effect of increased CO_2 solubility we impose a reconstruction of LGM SST and SSS on the carbonate chemistry routine that calculates the speciation of DIC and pCO_2 in the seasurface. Note that the temperature and salinity fields relevant to ocean dynamics are kept unchanged. The fields for glacial SST and SSS are derived through interpolation from the data set provided by *Paul and Schäfer-Neth* [2003] which is a merge of CLIMAP and GLAMAP data with modeling results.

2.2. Sea Ice Cover

In glacial times, increased sea ice cover around Antarctica might have drastically decreased outgassing of CO_2 in the Southern Ocean. Using a box model, *Stephens and Keeling* [2000] have demonstrated a substantial lowering of atmospheric CO_2 resulting from reduced air-sea gas exchange in the Antarctic region. However, *Archer et al.* [2003] showed that in contrast to box models, such an effect cannot be found in GCMs. In this study we examine the mechanism by reducing gas exchange in the domain around Antarctica where sea ice was present at the Last Glacial Maximum (LGM). Based on the data set provided by *Paul and Schäfer-Neth* [2003] a monthly map of fractional sea ice cover was obtained through area-weighted interpolation onto the coarse Bern3D model grid. Fractional sea ice cover was then used to linearly scale down gas transfer velocities according to:

$$k_w^{ice} = (1 - f_{ice}) k_w^{open}, \tag{1}$$

where k_w^{open} denotes the gas transfer velocity for an open ocean surface and f_{ice} is the fractional sea ice cover. Circulation and biological uptake were held unaffected by the addition of sea ice.

2.3. $CaCO_3$ Compensation

The mechanism of $CaCO_3$ compensation calls for a deep ocean drop in CO_3^{2-} due to an addition of excess remineralized CO_2 from the upper ocean (*Marchitto et al.* [2004]). The resulting decline in the saturation of deep waters with respect to $CaCO_3$ initiates the dissolution of seafloor $CaCO_3$ sediments. This process adds alkalinity and DIC to the seawater in a 2:1 ratio and thus shifts the speciation of DIC away from dissolved CO_2. When this signal propagates to the ocean surface, pCO_2 is reduced, causing a further uptake of atmospheric CO_2.

$CaCO_3$ compensation acts as homeostat for the deep ocean carbonate concentration. We simulate the effect of this process by restoring CO_3^{2-} at the sea floor deeper than 2000 meters towards CO_3^{2-*}, the carbonate concentrations diagnosed in the initial state:

$$J^{sed}_{CO_3^{2-}}(i,j,k_{i,j}) = \begin{cases} \frac{1}{\tau_{rest}}(CO_3^{2-*}(i,j,k_{i,j}) - CO_3^{2-}(i,j,k_{i,j})), \\ \text{if depth}(k_{i,j}) \ > \ 2000\text{m}, \\ \\ 0, \ \text{otherwise.} \end{cases} \tag{2}$$

$$SMS^{sed}_{DIC}(i,j,k_{i,j}) = J^{sed}_{CO_3^{2-}}(i,j,k_{i,j}) \tag{3}$$

$$SMS^{sed}_{ALK}(i,j,k_{i,j}) = 2J^{sed}_{CO_3^{2-}}(i,j,k_{i,j}) \tag{4}$$

SMS^{sed}_{DIC} and SMS^{sed}_{ALK} are the source-minus-sink terms for DIC and alkalinity that result from sediment dissolution or accumulation. The restoring timescale τ_{rest} is set to 10 years as we are not interested in the transient response, but want to bring the model into the new equilibrium as quickly as possible. The latitudinal and meridional indices of the model grid are denoted by i and j, while $k_{i,j}$ represents the depth-index of the deepest ocean grid cell above the sea floor.

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