On the control of glacial-interglacial atmospheric CO_2 variations by the Southern Hemisphere westerlies

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Introduction

According to Toggweiler et al. [2006, hereafter T06], the relationship between deep ocean circulation and ocean carbon budget can be schematically decomposed into two components: a northern, biologically productive upper ocean circulation on top of a southern unproductive deep ocean circulation (Fig. 1).

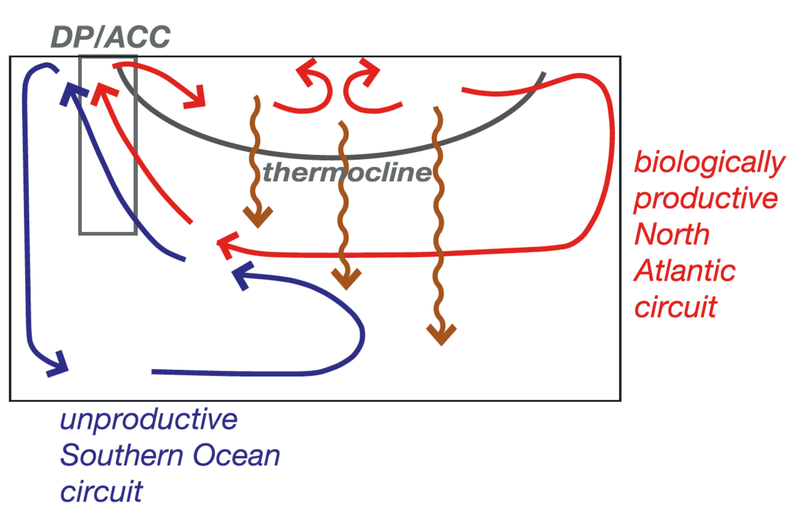


Figure 1. Conceptual view of T06. The southern deep branch upwells in the Southern Ocean and sinks along Antarctica with very limited biological production at the surface. The northern upper ocean branch encompasses all three basins with biologically productive surface waters and returns to the surface in the Southern Ocean. The northern branch is seen as the source of most of the organic sinking particles which are remineralized at depth in both the deep and upper ocean branches.

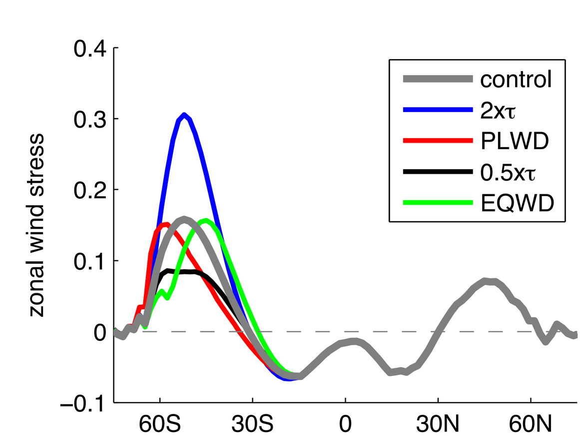
T06 assume that the biological carbon pump is independent of the southern deep circulation which brings back the deep carbon to the surface. Because the Southern Hemisphere Westerlies (SHW) directly control the majority of the upwelling of the deep ocean, T06 hypothesized that changes in the SHW can potentially explain a significant part of the atmospheric CO_2 variations seen between glacial and interglacial periods.

The hypothesis of T06 was first formally tested in 3D ocean models by *Tschumi et al.* [2008] and *Menviel et al.* [2008] who found only small changes of atmospheric CO_2 in response to SHW variations. Here, **we revisit T06's hypothesis** and the above studies. In particular, we give a different explanation as to why wind-driven changes to the deep ocean carbon reservoir do not necessarily entail changes in atmospheric CO_2.

Experimental design

We compare responses of a **poleward shift of the SHW** and of an **increase in their amplitude**. This comparison is particularly relevant here as T06 argued that both of these SHW changes would have essentially the same effect [see also Tschumi et al., 2008].

Figure 2.
Latitudinal profile of the annual average of zonal wind stress (in Pa), zonally averaged over the ocean, for the control simulation and the four perturbation experiments.



To this end, we use the University of Victoria Earth System Climate Model. The control simulation corresponds to a warm interglacial climate (CO_2 concentration of 300 ppm, present day ice-sheet and solar forcing) witha 8000 years spinup. Two 1000 year perturbation experiments are branched off, in which the monthly wind stress field is modified such that the SHW amplitude is doubled or such that the SHW latitudinal position is shifted poleward (**2 × τ** and **PLWD** respectively, Fig. 2).

Results

Comparing DIC changes in experiments 2 × τ and PLWD, there is a common decrease below 3000m depth and an opposite change between 2000 and 500m depth north of 40°S (Fig. 3).

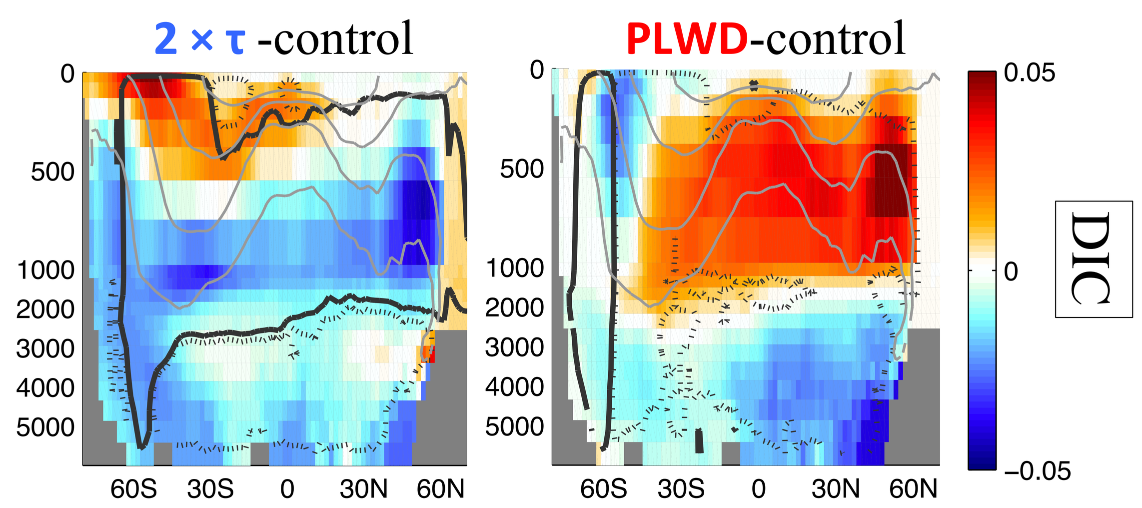


Figure 3. Difference between perturbation and control of DIC in mol/m^3 zonally averaged over the entire ocean. Initial values of DIC (gray lines). Change of the global MOC: 0.5 Sv (solid black) and −0.5 Sv (dotted black). Experiments with a change in SHW amplitude 2 × τ (left) and with a poleward shift PLWD (right).

The **common deep decrease in DIC** is due to a decrease in the residence time of the deep ocean water while the downward carbon fluxes due to the biological pump into the deep ocean increase slightly in 2 × τ and do not change in PLWD (not shown). The acceleration of the bottom cell of the MOC (Fig. 3) can also be visualized by the difference in concentration of dye released in the Southern Ocean (south of 35°S) at the beginning of each experiment and of the control (Fig. 4). The **deep ventilation from the Southern Ocean has therefore increased** in 2 × τ and PLWD as expected by T06 (increase in AABW production, in mixing of CDW and in upwelling in the Southern Ocean).

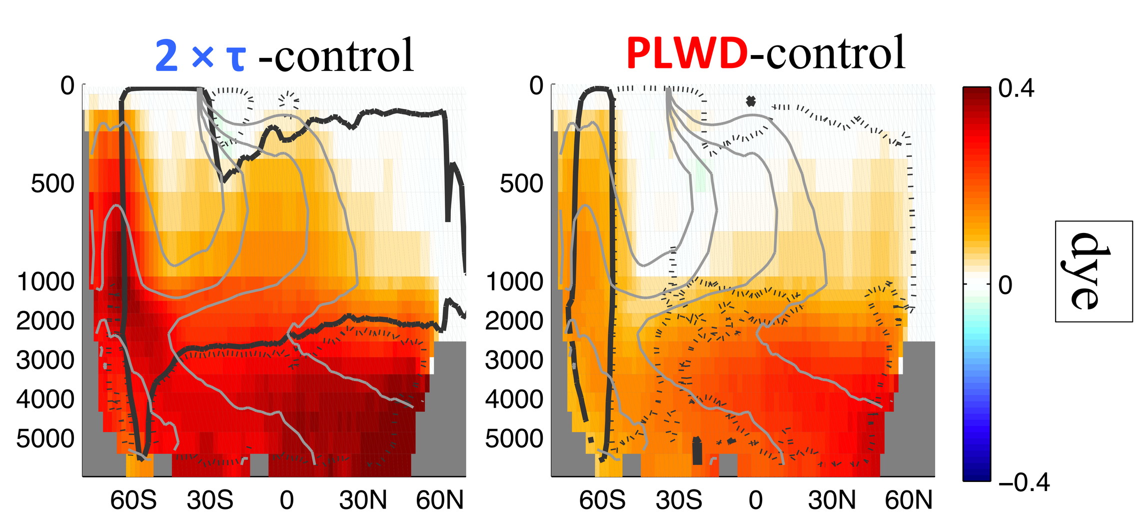


Figure 4. Same as Figure 3, but for concentration of dye released south of 35°S. Dimensionless values range from 0 to 1.

The **DIC change of opposite sign in the upper ocean** (Fig. 3) is due to an **opposite change of the pycnocline depth**, with a deepening in 2 × τ and a shoaling in PLWD, as seen by the warmer and colder upper ocean respectively (Fig. 5). The deepening (shoaling) of the pycnocline depth corresponds to an increased (decreased) ventilation and a reduced (enhanced) carbon solubility of the upper ocean water, both bringing down (up) DIC between 500 and 2000m.

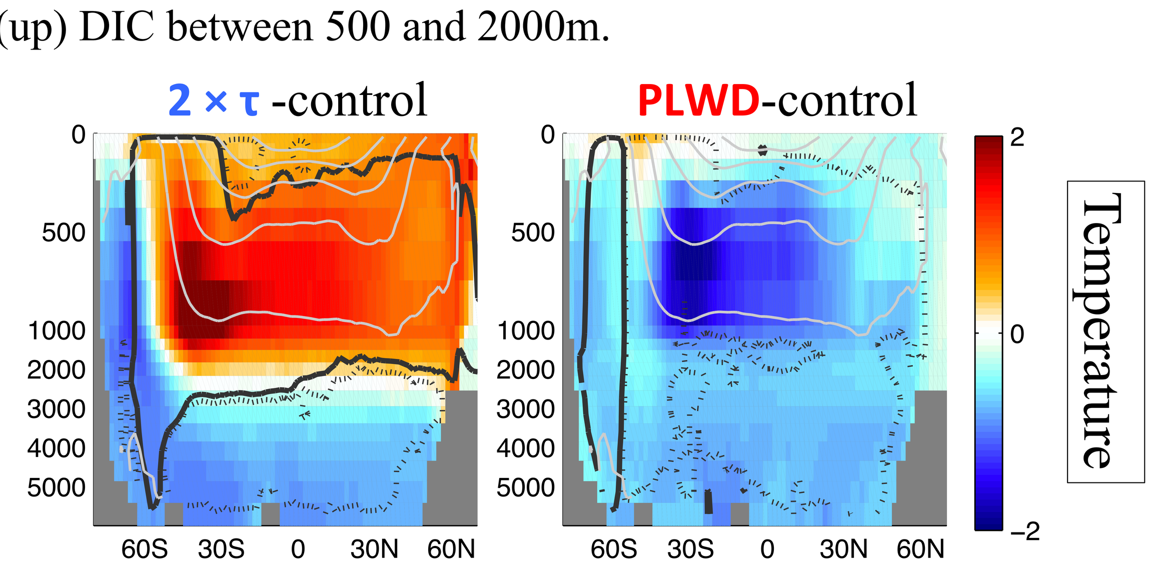


Figure 5. Same as Figure 3, but for temperature in °C.

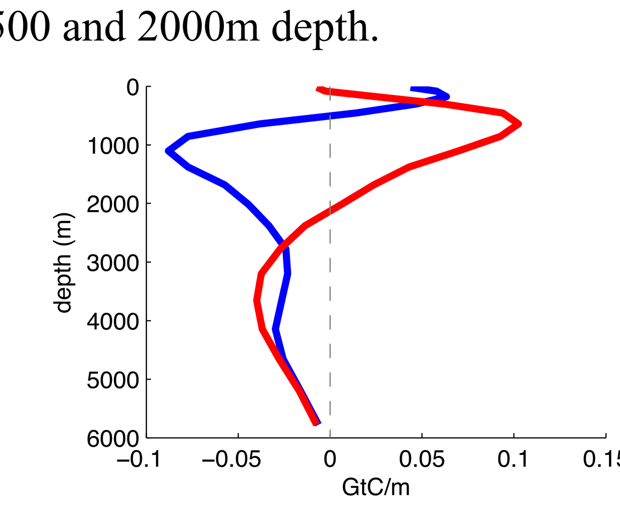
The secondary effect of the modified SHW is to modify the distribution of nutrients, which in turn impacts the strength of the biological carbon pump.

- In **2 × τ**, this secondary effect is a global increase of surface nutrients which **enhances the biological export production** into both the upper and deep oceans (not shown), therefore **partially offsetting the ocean carbon loss due to ventilation** changes [Menviel et al., 2008].
- In PLWD, the wind shift decreases the concentration of nutrients at the surface of the Southern Ocean and increases surface nutrients elsewhere. These surface nutrient changes lead to negligible changes of export production into the deep ocean and to a small increase into the upper ocean (not shown).

Conclusions

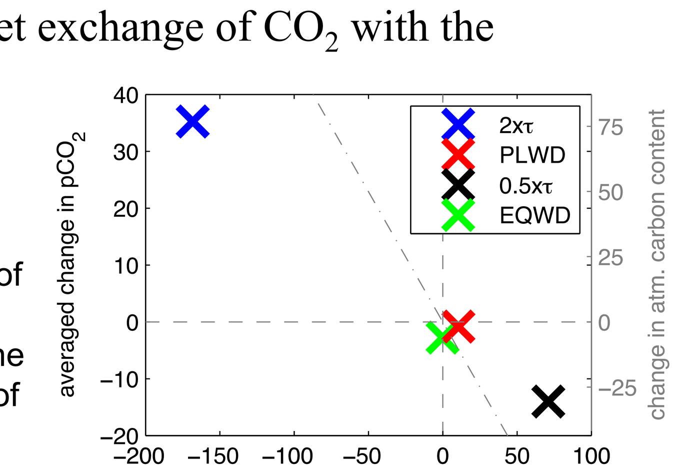
- As predicted by T06, the deep ocean carbon reservoir below 2000m depth is decreased for a strengthening or a poleward shift of the SHW (Fig. 6). However, in all experiments, the **largest changes in oceanic carbon content take place in the upper ocean** between 500 and 2000m depth.

Figure 6.
Vertical profiles of the change in oceanic carbon content by depth (in GtC/m) for simulations 2 × τ and **PLWD**.



- For the case of a **poleward shift of the SHW, carbon changes in the upper ocean counterbalance those of the deep ocean** resulting in almost no net exchange of CO_2 with the atmosphere.

Figure 7.
Globally averaged change in pCO_2 (in ppm) as a function of the change in total oceanic carbon content (in GtC) for the four pertubation experiment of Figure 2 (**2 × τ** and **PLWD**).

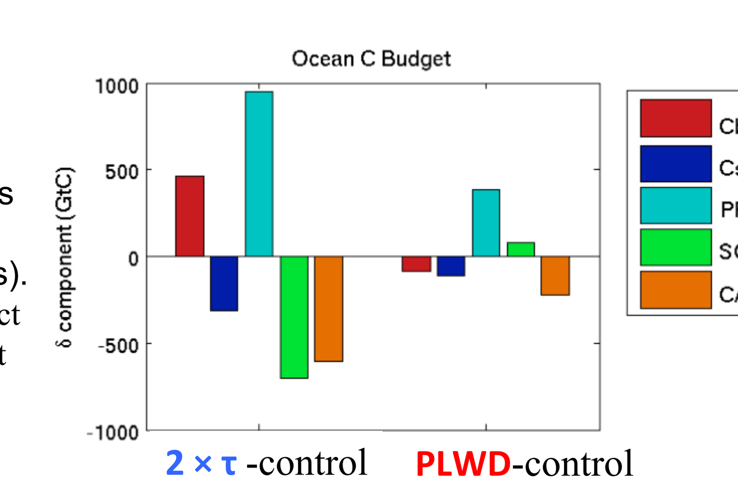


- In all wind perturbation experiments the secondary effect on the biological pump due to changes in the oceanic circulation remains small. **The primary driver of ocean carbon content changes is the change in ocean ventilation forced by the wind.**

Future work

Figure 8. Separation of different oceanic components (in GtC) obtained with 12 passive tracers (t=3000years).

- Cbuff – ocean buffering effect
- Csurf –alk,T,S surf.ace effect
- PREF –solubility pump
- SOFT – soft-tissue pump
- CARB – carbonate pump



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