On the control of glacial-interglacial atmospheric CO$_2$ variations by the Southern Hemisphere westerlies

M. d’Orgeville, W. P. Sijp, M. H. England, and K. J. Meissner

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[1] The University of Victoria Earth System Climate Model is used to investigate the effects of changes in Southern Hemisphere Westerlies (SHW) on atmospheric CO$_2$. It is shown that a northward shift of the SHW and a decrease of their amplitude have the same qualitative effect on deep ocean carbon storage, which increases because of a deceleration of the bottom meridional overturning circulation. A southward shift or a strengthening of the SHW has the opposite effect. However, latitudinal shifts of the SHW and changes in their amplitude are not equivalent in terms of atmospheric CO$_2$. In particular, while doubling the SHW amplitude increases atmospheric CO$_2$ by 36 ppm and halving it by 20 ppm, the latitudinal shifts (north- or southward) have no significant impact on atmospheric CO$_2$. These different CO$_2$ responses are due to different dynamical responses of the upper ocean circulation which, in the case of latitudinal shifts, produce a carbon storage change opposite to the one observed for the deep ocean. In all experiments, the changes in the biological carbon pump in response to a redistribution of the nutrients by the modified oceanic circulation remain small. Ultimately, the atmospheric CO$_2$ response depends on the control the SHW exert on both the ventilation of the deep ocean and the depth of the upper ocean pycnocline.


1. Introduction

[2] 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 (T06’s Figure 3). The southern deep branch upwells in the Southern Ocean and sinks along Antarctica with therefore very limited biological production at the surface. By contrast, the northern upper ocean branch encompasses all three basins with biologically productive surface waters, sinking in the North and returning to the surface in the Southern Ocean. The northern branch is therefore seen as the source of most of the organic sinking particles which are remineralized at depth in both the deep and upper ocean branches. In other words, T06 assume that the biological carbon pump is independent of the southern deep circulation which brings back the deep carbon to the surface.

[3] In this context, a mechanism able to spin up or spin down, the southern deep circulation without modifying the upper ocean branch and the biological pump can directly lead to (i) a change in the deep ocean carbon reservoir and, consequently, (ii) a change of atmospheric CO$_2$ concentration. Because the Southern Hemisphere Westerlies (SHW) directly control the majority of the upwelling of the deep ocean [Toggweiler and Samuels, 1995] and indirectly control the production of Antarctic Bottom Water (AABW) and Circumpolar Deep Water (CDW), 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.

[4] The hypothesis of T06 was first formally tested in 3-dimensional 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. Especially relevant to T06’s conceptual view, Menviel et al. [2008] demonstrated that stronger SHW bring extra nutrients to the surface (originally trapped as preformed nutrients in the deep ocean), increasing the biological production (in the northern upper-ocean branch) which is then capable of taking up some or all of the CO$_2$ released to the atmosphere by the stronger deep circulation.

[5] In the present study, we revisit the results of these previous studies and, in particular, give a different explanation as to why wind-driven changes to the deep ocean carbon reservoir do not necessarily entail changes in atmospheric CO$_2$. To this end, we compare responses of a poleward shift of the SHW and of an increase in their amplitude (as well as a comparison between an equatorward shift and an amplitude weakening). 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]. The results obtained highlight the primary role of the upper ocean circulation in response to SHW variations for the global ocean carbon budget.

2. Model and Experimental Design

[6] The UVic ESCM (Version 2.8) consists of an ocean general circulation model coupled to a vertically integrated two dimensional energy-moisture balance model of the atmosphere and a dynamic-thermodynamic sea ice model [Weaver et al., 2001]. It includes a fully coupled carbon cycle with terrestrial and oceanic components [Meissner et al., 2003; Ewen et al., 2004; Schmittner et al., 2008]. The climate boundary conditions of the control simulation correspond to a warm interglacial climate (CO$_2$ concentration of 300 ppm, present day ice-sheet distribution, solar
forcing of year 2000). The model is integrated for 8000 years, the last 1000 years is taken as the control simulation.

Four different perturbation experiments are branched off after the 7000 year spinup period, in which the standard monthly NCEP wind stress field is modified such that the SHW amplitude is doubled or halved (experiments $2 \times t$ and $0.5 \times t$), or such that the SHW latitudinal position is shifted poleward or equatorward by 7.2 degrees (experiments PLWD and EQWD). The zonal wind stress anomaly is computed monthly and is zonally symmetric (Figure 1a). The perturbation experiments are integrated for 1000 years. The differences between the average of the last 100 years of each perturbation experiment with the average of the last 100 years of the control simulation are described below. Note that the wind modification is applied to both the ocean and sea-ice momentum transfer but not to atmospheric variables such as moisture advection and evaporative heat fluxes, unlike other studies interested in how SHW changes can affect future warming scenarios [Zickfeld et al., 2007]. In the remainder of this letter, references to wind changes will implicitly refer to changes in wind stress only.

To examine the resultant changes in the ocean circulation, we conducted “dye experiments” for which passive tracers (dye) are released in different regions. From the start of each 1000 year perturbation experiment, dye concentration at the surface is restored continuously to 1 in the region of interest and 0 elsewhere. Changes in dye concentrations at depth relative to the control simulation can then be interpreted as a measure of the change in ventilation rate from this region.

3. Results

Figure 1b displays the change in surface partial pressure of carbon dioxide ($pCO_2$) spatially averaged over the ocean as a function of the change in total ocean carbon content. A strengthening of the SHW ($2 \times \tau$ - blue) leads to an outgassing of $CO_2$ by the ocean (as predicted by T06) and a 36 ppm increase in atmospheric $CO_2$. Note that 90 GtC of the 168 GtC released by the ocean is captured by the land (i.e., soil and vegetation), with only 78 GtC remaining in the atmosphere (right axis). Conversely, a weakening of the SHW ($0.5 \times \tau$ - black) leads to an enhanced drawdown of carbon by the ocean and a net decrease of 20 ppm of atmospheric $CO_2$. By contrast, there is no significant change in the total ocean carbon content or in the average $pCO_2$ for the shifted SHW cases.

Figures 1c and 1d present the vertical profiles of the change in oceanic carbon content. As predicted by T06, the deep ocean carbon reservoir below 2000m depth is decreased for a strengthening or a poleward shift of the SHW (Figure 1c) and increased for a weakening or an equatorward shift (Figure 1d). However, in all experiments, the largest changes in oceanic carbon content take place in

Figure 2. Latitude-depth diagrams of DIC concentration in mol/m$^3$ zonally averaged over the entire ocean: initial values (gray lines, values in Figure 2d) and anomaly (in color). Change of the global MOC (dark lines): only the 0.5 (solid) and $-0.5$ Sv (dotted) isolines are plotted to highlight the sign of the change. Fields are shown for experiments with a change in SHW amplitude (a) $2 \times \tau$ and (b) $0.5 \times \tau$ and in their latitudinal position (c) PLWD and (d) EQWD. Note the irregular depth axis. Magenta and purple boxes in (d) define the upper and deep ocean regions of Table 1 respectively (limits at 40°S and at 550 and 2200 m depth).
the upper ocean between 500 and 2000m depth. In $2 \times \tau$ and $0.5 \times \tau$, this upper ocean change is of the same sign as in the deep ocean, therefore leading to a net exchange of carbon with the atmosphere. But for experiments PLWD and EQWD, carbon changes in the upper ocean counterbalance those of the deep ocean resulting in almost no net exchange of CO$_2$ with the atmosphere.

The zonally averaged dissolved inorganic carbon (DIC) concentration changes are presented in Figure 2. Comparing simulations $2 \times \tau$ with PLWD and $0.5 \times \tau$ with EQWD, there is a common change in DIC below 3000m depth and an opposite change in DIC between 2000 and 5000m depth north of 40°S. The remainder of this letter will concentrate on explaining these two changes for simulations $2 \times \tau$ and PLWD, with a similar but opposite mechanism explaining the results obtained for simulations $0.5 \times \tau$ and EQWD.

### 3.1. Deep Ocean Ventilation

The downward carbon fluxes due to the biological pump into the deep ocean increase slightly in $2 \times \tau$ and do not change in PLWD (Table 1). Changes in the biological pump can therefore be ruled out as the driver for the deep DIC decrease, common to both experiments (see also section 3.3). On the other hand, the negative bottom meridional overturning circulation (MOC) is accelerated in $2 \times \tau$ and PLWD (Figures 2a and 2c), thus decreasing the residence time of the deep ocean. This can therefore explain the decrease in deep DIC concentration seen in both $2 \times \tau$ and PLWD.

Figure 3 presents the difference in concentration of dye released in the Southern Ocean (south of 35°S). The maximum increase in dye concentration fraction is as large as 0.4 in the deep ocean in both $2 \times \tau$ and PLWD, meaning that up to 40% more water originates south of 35°S in those simulations. The deep ventilation from the Southern Ocean has therefore increased in $2 \times \tau$ and PLWD as expected by T06. This increased deep ventilation comes primarily from an increase of AABW production adjacent to Antarctica (dye south of 65°S - not shown) as well as an increase in the portion of CDW mixed with the surface water of the Antarctic Circumpolar Current (dye south of 55°S - not shown). The acceleration of the bottom circulation in $2 \times \tau$ and PLWD can also be related to the increased upwelling visible in the MOC changes from 2000m depth to the surface around 65°S in $2 \times \tau$ and around 70°S in PLWD (Figures 3a and 3c), due to a change in Ekman pumping at the surface [Toggweiler and Samuels, 1995]. Ultimately, an increase in AABW production and CDW mixing, together with an increased upwelling of deep water in the Antarctic Circumpolar Current region drive the increase in the MOC bottom circulation and the decrease in DIC concentration seen below 3000m depth in both $2 \times \tau$ and PLWD (Figures 2a and 2c).

### 3.2. Pycnocline Depth

The second important feature seen in Figures 2a and 2c is the DIC change of opposite sign occurring in the upper ocean, defined as north of 40°S between 550 and 2200m depth. The changes in the strength of the biological carbon pump into the upper ocean remain small in both $2 \times \tau$ and PLWD (Table 1) and are therefore unlikely to explain these DIC changes (see also section 3.3). The upper ocean depth range is also the transition region between the light, low-latitude surface waters and the denser, deep water, with 1000m being roughly the depth of the permanent oceanic pycnocline. This depth range and the pycnocline can also be seen as separating the carbon depleted water close to the surface from the carbon rich water of the abyss. The DIC changes of opposite sign with extrema located in the upper ocean are indicative of opposing changes in the pycnocline depth. In $2 \times \tau$, the upper ocean waters warm (Figure 4a), the pycnocline deepens and the upper ocean waters become more depleted in carbon (Figure 2a). Conversely in PLWD, the upper ocean waters cool (Figure 4c), the pycnocline shoals and the upper ocean waters are more enriched in carbon (Figure 2c).

As suggested by Gnanadesikan [1999], the pycnocline depth is dependent on the rate of North Atlantic Deep Water (NADW) production, the diffusivity across the pycnocline and the rate of wind-induced upwelling in the

![Figure 3](image)

**Figure 3.** Same as Figure 2, but for concentration of dye released south of 35°S. Dimensionless values range from 0 to 1. See text for details.
Southern Ocean. In our experiments, variations in the pycnocline are unlikely to be due to changes of diffusion (the same background diffusivity is used in all experiments) or to the change in NADW production (which increases in both experiments). On the other hand, the surface northward Ekman transport is decreased in PLWD from 55°S to 35°S due to the poleward displacement of the SHW, and increased in $2 \times T$ because of the overall strengthening of the SHW. Changes of the wind north of 55°S seem therefore to control the variation of the pycnocline depth and subsequently the upper ocean carbon content in our model experiments.

In this context, the upper ocean DIC changes can be linked to pycnocline depth variations described above either by the changes in residence time of the upper ocean waters (for instance, because of a change in the production rate of intermediate waters as found in dye experiments not presented here) or by the solubility pump response to changes in upper ocean temperature (warmer in $2 \times T$, colder in PLWD). We refer the reader to Sipp and England [2008, 2009] for a thorough discussion of the changes in temperature and production of intermediate water for the SHW shift cases. Although it would be interesting to disentangle the relative importance of residence time versus solubility effect in a future study, both factors are due to a direct change of the ventilation of the upper ocean and will not be discussed further here.

### 3.3. Biological Pump

For the ocean carbon budget, when the SHW are modified, the first-order impact is via the ventilation of the upper ocean and deep water masses. The secondary effect is to modify the distribution of nutrients, which in turn impacts the strength of the biological carbon pump. The balance between these two effects will ultimately determine the net change in ocean carbon content in response to the applied wind perturbations.

In $2 \times T$, the secondary effect is a global increase of surface nutrients which enhances the biological export production into both the upper and deep oceans (Table 1). In terms of total ocean carbon content, this stronger biological pump is opposite in sign to the primary effect of increased ventilation of the ocean, and therefore this partially offsets the ocean carbon loss due to ventilation changes. Similar results were obtained by Menviel et al. [2008] for a smaller increase of the SHW strength.

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 (Table 1). The decrease in deep ocean carbon content is therefore solely due to the increased deep water ventilation. The increase of upper ocean carbon content is due primarily to the reduced ventilation, with the slightly enhanced biological pump playing a secondary role.

### 4. Conclusions

We have demonstrated that, although changes in Southern Hemisphere Westerlies (SHW) lead to a large vertical redistribution of carbon inside the ocean, only in certain cases this redistribution results in a significant change of atmospheric CO$_2$. A latitudinal shift of the SHW results in compensating changes in the deep and upper ocean carbon reservoirs leading to no significant changes in atmospheric CO$_2$. On the other hand, changes in the SHW strength result in a global change of the oceanic carbon reservoir that will therefore impact on atmospheric CO$_2$.

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. The present analysis suggests, moreover, that the wind amplitude south of 55°S controls the ventilation of the deep ocean, and therefore the deep ocean carbon reservoir (T06), but the wind amplitude north of 55°S controls the depth of the upper ocean pycnocline [Gnanadesikan, 1999], and therefore the upper ocean carbon reservoir.

Because of small atmospheric CO$_2$ responses to wind changes in their respective models, both Menviel et al. [2008] and Tschumi et al. [2008] concluded that SHW changes were an unlikely candidate to explain a significant part of the glacial-interglacial CO$_2$ variations, and therefore rejected T06’s hypothesis. In the present study, instead of focusing only on the magnitude of the atmospheric response, SHW changes have been shown to efficiently redistribute the carbon content within the ocean. SHW changes could therefore potentially explain the observation of abrupt $\Delta^{14}$C changes during glacial-interglacial cycles that are not necessarily related to net atmospheric CO$_2$ changes. Furthermore, SHW changes have been considered in isolation in this study, but their impact in combination with changes in other key parameters (such as, for instance, a shutdown of NADW formation) could either mitigate or exacerbate the corresponding atmospheric CO$_2$ response.
during deglaciation [Toggweiler and Lea, 2010]. This should be explored in future studies.

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References

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M. d’Orgeville, M. H. England, K. J. Meissner, and W. P. Sijp, Climate Change Research Centre, University of New South Wales, Sydney, NSW 2052, Australia. (marcdo@unsw.edu.au)