The future evolution of the Southern Ocean CO$_2$ sink

by Nicole S. Lovenduski$^{1,2}$ and Takamitsu Ito$^1$

ABSTRACT

We investigate the impact of century-scale climate changes on the Southern Ocean CO$_2$ sink using an idealized ocean general circulation and biogeochemical model. The simulations are executed under both constant and changing wind stress, freshwater fluxes, and atmospheric $p$CO$_2$, so as to separately analyze changes in natural and anthropogenic CO$_2$ fluxes under increasing wind stress and stratification. We find that the Southern Ocean sink for total contemporary CO$_2$ is weaker under increased wind stress and stratification by 2100, relative to a control run with no change in physical forcing, although the results are sensitive to the magnitude of the imposed physical changes and the rate of increase of atmospheric $p$CO$_2$. The air-sea fluxes of both natural and anthropogenic CO$_2$ are sensitive to the surface concentration of dissolved inorganic carbon (DIC) which responds to perturbations in wind stress and stratification differently. Spatially averaged surface DIC scales linearly with wind stress, primarily driven by changes in the Ekman transport. In contrast, changes in the stratification cause non-linear and more complex responses in spatially averaged surface DIC, involving shifts in the location of isopycnal outcrop for deep and thermocline waters. Thus, it is likely that both wind stress and stratification changes will influence the strength of the Southern Ocean CO$_2$ sink in the coming century.

1. Introduction

Oceanic models (Lovenduski et al., 2008), atmospheric data (Le Quéré et al., 2007), and oceanic observations (Metzl, 2009) indicate that the Southern Ocean sink for atmospheric CO$_2$ has substantially weakened in the last few decades, relative to the expected sink from rising atmospheric CO$_2$ and fixed physical climate. It has been suggested that the primary cause of the sink reduction is a trend in the position and intensity of the Southern Hemisphere westerly winds and the subsequent increase in the upwelling and equatorward transport of CO$_2$-rich waters (Le Quéré et al., 2007; Lovenduski et al., 2008). As simulations of future climate from coupled models consistently find a trend toward stronger, poleward shifted winds over the Southern Ocean during the next century (Miller et al., 2006; Meehl et al., 2007), it is possible that a further weakening of the Southern Ocean CO$_2$ sink will occur.

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The impact of future secular trends in Southern Hemisphere wind stress on Southern Ocean air-sea CO\textsubscript{2} flux are predicted to be accompanied by changes in climate-driven effects, such as the stratification of the global oceans (Fung et al., 2005). The Southern Ocean in particular will likely experience a large increase in stratification due to enhanced melting of land and sea ice, warmer surface ocean temperatures, and enhanced precipitation at high latitudes (Manabe and Stouffer, 1993; Sarmiento et al., 1998; Hirst, 1999; Sarmiento et al., 2004b; Bitz et al., 2006). A stratification change of this type can simultaneously reduce the upward flux of natural CO\textsubscript{2} (Toggweiler, 1999) and the downward flux of anthropogenic CO\textsubscript{2} (Sarmiento et al., 1998) in the Southern Ocean, making it difficult to predict the impact of stratification on the total CO\textsubscript{2} sink.

Accurate predictions for the future evolution of the Southern Ocean CO\textsubscript{2} sink therefore require understanding how both wind stress and stratification impact carbon cycling and CO\textsubscript{2} exchange in this region. It has been suggested by Matear and Lenton (2008) that, over the past 5 decades, increased wind stress and increased stratification of the Southern Ocean have led to an equal and opposite CO\textsubscript{2} flux response. Furthermore, Matear and Lenton (2008) find that changes in Southern Ocean natural CO\textsubscript{2} exchange have always been countered by much smaller changes in anthropogenic CO\textsubscript{2} uptake over this time period. It is of critical importance to know if these compensatory processes will continue to operate in a future characterized by increased atmospheric CO\textsubscript{2} concentrations, when wind stress and stratification changes are likely to continue.

Unfortunately, predictions for the future evolution of the Southern Ocean CO\textsubscript{2} sink are plagued by uncertainties in the estimated future state of the atmosphere and the strength of climate feedback processes. While it is well-accepted that wind stress and stratification will somewhat increase in this region, the magnitude of the changes in these quantities differs substantially among the general circulation models in the IPCC AR4 (W. Lefebvre, pers. comm., 2008; Meehl et al., 2007), making it difficult to accurately represent the future state of CO\textsubscript{2} fluxes in such models. One possible route to characterizing future behavior is by performing a suite of sensitivity studies in a simplified framework, in order to better understand the mechanisms leading to change.

Here, we use an idealized model to explore the possible future evolution of the Southern Ocean natural and anthropogenic carbon cycles under both enhanced wind stress and increased stratification. We simulate the ocean carbon cycle using an ocean general circulation model with a simplified geometry and biogeochemistry. We then probe the model with various configurations of atmospheric forcing and CO\textsubscript{2} boundary conditions to separately examine these processes in the most simplified and theoretical manner.

2. Model configuration and experiments

We investigate the impact of changes in wind stress and stratification on the Southern Ocean CO\textsubscript{2} sink using a sector version of the MIT ocean general circulation and biogeochemical model (Ito and Follows, 2005). The model is configured with an idealized
Figure 1. Configuration of the sector model used in the study.

bathymetry, such that it represents an interhemispheric, rectangular ocean basin with an open channel in the Southern Hemisphere (Fig. 1). A periodic boundary condition is applied to this open channel, simulating the flow of the Antarctic Circumpolar Current, which extends to 2000 m depth. The full model domain extends from 65°N to 65°S, and to a flat bottom at 3000 m. We run the model with a horizontal resolution of 2° and 30 vertical layers of varying thickness. Eddies are parameterized according to the Gent and McWilliams (1990) scheme with a constant isopycnal thickness diffusivity. The small domain size and coarse resolution allow the model to be run at a very low computational cost, making it ideal for use in sensitivity studies.

The physical model is forced with a sinusoidal zonal wind stress profile, and surface temperature and salinity are restored on a 1 month timescale to prescribed sinusoidal profiles, which are broadly consistent with the modern ocean climatology. There is no seasonal variation in these forcing profiles, and no sea ice component in the model.

The biogeochemical component of this model is also highly idealized. It consists of 5 passive tracers: dissolved inorganic carbon (DIC), alkalinity, dissolved oxygen, phosphate, and dissolved organic phosphorous. These tracers evolve with physical transport, biological uptake and remineralization, and air-sea gas exchange. Biological uptake is parameterized using Michaelis-Menton kinetics, with a uniform, constant maximum uptake rate and half-saturation coefficient for all simulations. Air-sea CO₂ exchange is parameterized according to Wanninkhof (1992), with an evolving ocean pCO₂ and prescribed atmospheric pCO₂.
The model is initialized with uniform distributions of all tracers and spun up for a period of 3000 years under constant physical forcing and restoring profiles and an atmospheric $p$CO$_2$ boundary condition of 278 ppmv. This long integration period allows ocean CO$_2$ to come into equilibrium with the atmosphere, and there is no measurable drift in the modeled, globally integrated air-sea flux of CO$_2$ (not shown).

Following the spinup, we conduct three separate sets of simulations with different atmospheric $p$CO$_2$ boundary conditions (Table 1, Fig. 2a). In the first, atmospheric $p$CO$_2$ is set to the constant, pre-industrial spin-up value of 278 ppmv. In the second and third, this boundary condition is increased as piecewise linear fits to the measured and reconstructed atmospheric $p$CO$_2$ from 1765 until 2000, and from 2000 until 2100, with stabilized values after 2100 of 450 ppmv and 900 ppmv, respectively. These endpoints were chosen because they represent an extreme range of atmospheric CO$_2$ concentrations in 2100, as predicted by various economic development scenarios. From the first set of simulations, we obtain information about the fluxes of natural CO$_2$, whereas from the second and third set, we obtain information about the fluxes of contemporary CO$_2$. The difference between the fluxes of contemporary and natural CO$_2$ is considered to be the fluxes of anthropogenic CO$_2$ (Lovenduski et al., 2007). Throughout this manuscript, when making reference to the anthropogenic and contemporary CO$_2$ fluxes, we are specifically referring to fluxes from this second set of simulations (450 ppmv), unless otherwise stated.

In each set of simulations, we conduct one control run with constant physical forcing, one run with increasing wind stress, and another run with increasing freshwater flux from 1980 to 2100 (Table 1). Figures 2b and 2c show the beginning and ending profiles used for and derived from these perturbation runs. In our wind perturbation experiment, zonal wind is linearly interpolated from the solid to the dashed line over a 120-year period. The

Table 1. Sensitivity experiments conducted with the sector model.

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Figure 2. (a) Atmospheric $p$CO$_2$ (ppmv) boundary conditions used for the sensitivity experiments. (b) Wind stress (N m$^{-2}$) profiles used in the control and perturbation experiments, where the magnitude of the wind stress at 50°S was increased linearly from the control to the perturbed value over a 120-year period. (c) Zonal-mean surface salinity resulting from the control and freshwater perturbation experiments, where the Southern Ocean surface freshwater flux was increased linearly for a 120-year period. The magnitude of the wind stress perturbation was determined by extrapolating the approximate historical (1958–2004) trend in wind stress magnitude from the NCEP/NCAR reanalysis into the future. Note that the mean position of maximum wind stress does not change with time in this set of experiments. We conducted one additional experiment where the position of maximum wind stress moves poleward with time (Table 1). We mimic the effect of
increased stratification by increasing the high latitude freshwater flux input with time. This imposed change in freshwater flux lowers the surface salinity (Fig. 2c) and causes the high latitude regions (<50°S) of the Southern Ocean to stratify nearly as much as predicted by the IPCC AR4 multi-model mean estimates in the year 2100 (Table 2). We purposely chose to force the system with freshwater flux boundary conditions rather than a uniformly imposed salinity to allow the model ocean to find weak points in stratification through which its ventilation can occur (Toggweiler and Bjornsson, 2000).

The model output is analyzed using spatial integration, spatial averages, and zonal-mean properties. Spatial integration and averaging is done over the entire Southern Ocean, in the region between 30°S and 65°S, while taking into account the relative area of each grid cell. Zonal averages of a property, however, are calculated simply as an arithmetic mean at each latitude, and are displayed without consideration for the area of that region.

### 3. Evaluation of the modeled mean state

Conducting idealized experiments with a simple ocean biogeochemical model helps to clarify the important mechanisms affecting air-sea CO₂ flux. However, it is necessary to first evaluate the sector model in its mean state to determine its strengths and weaknesses. Here, we compare output from the sector model with observations and a global ocean model with complex biogeochemical parameterizations. We limit our evaluation to zonal mean air-sea CO₂ fluxes, as the physical circulation has been previously described in a former version of the sector model (Ito and Follows, 2005).

Our simple model can capture broad patterns and magnitudes of the annual-mean (year 2000), zonal-mean contemporary CO₂ flux in the Southern Hemisphere, as compared to observations (Takahashi et al., 2009) and simulations of the ECO-CCSM model (Lohmann et al. (2007), Fig. 3a). Oceanic uptake of contemporary CO₂ occurs in the region between 20 and 50°S, where natural CO₂ uptake is strong (Fig. 3b), whereas we find the region south of 50°S to be neither a source nor sink of contemporary CO₂, as the strong natural CO₂ outgassing is compensated by anthropogenic CO₂ uptake (Fig. 3b). The sector model underestimates the outgassing south of 60°S relative to the observations and forward ocean model; however, the ocean inversion model described in Gruber et al. (2009) predicts uptake of CO₂ in the region south of 58°S. Also, our simple sector model does not

### Table 2. Southern Ocean stratification changes from the freshwater perturbation experiment with the sector model, and predicted stratification changes from the IPCC AR4 models (W. Lefebvre, pers. comm., 2008). Stratification calculated as the difference in potential density between 300 m and the surface, and change calculated as the difference between years 2091–2100 and 1991–2000 (kg m⁻³).

<table>
<thead>
<tr>
<th>Latitude</th>
<th>Sector model</th>
<th>IPCC AR4</th>
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<tbody>
<tr>
<td>30–40°S</td>
<td>-0.02</td>
<td>0.14</td>
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<tr>
<td>40–50°S</td>
<td>0.04</td>
<td>0.14</td>
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<tr>
<td>50–60°S</td>
<td>0.11</td>
<td>0.13</td>
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<td>60–65°S</td>
<td>0.10</td>
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account for the freezing and melting of sea-ice, a process that may play an important role in controlling the $pCO_2$ budget in this high-latitude region (Metzl et al., 2006). Both the sector and ECO-CCSM models under estimate the CO$_2$ uptake in the mid latitudes of the Southern Hemisphere.

The zonal-mean natural and anthropogenic CO$_2$ fluxes found in our sector model also agree remarkably well with fluxes estimated from the ECO-CCSM (Fig. 3b). The natural outgassing peak in the ECO-CCSM at 60°S is not found at the correct latitude in the sector model, perhaps due to the absence of sea ice and our idealized configuration of the model domain (Fig. 1).

The temporal evolution of the zonal-mean fluxes of natural, anthropogenic, and contemporary CO$_2$ from the control (constant physical forcing) runs is shown in the first row of Figure 4. Our model Southern Ocean is a source of natural CO$_2$ to the atmosphere in the region south of 50°, and a sink of natural CO$_2$ in the region north of this latitude. These patterns of natural CO$_2$ exchange do not change with time, owing to the fixed boundary condition for atmospheric $pCO_2$ in this run. Our model Southern Ocean is everywhere absorbing anthropogenic CO$_2$ from the atmosphere, with a higher uptake occurring in high latitudes and the region north of 30°S. Absorption of anthropogenic CO$_2$ is increasing with time in our model due to the prescribed increase in the atmospheric $pCO_2$ boundary condition. The contemporary fluxes of CO$_2$ are the sum of the natural and anthropogenic fluxes,
Figure 4. Temporal evolution of the zonal-mean fluxes of (1st column) natural, (2nd column) anthropogenic, and (3rd column) contemporary CO$_2$. (1st row) Control experiment plotted with contours every 0.1 mol m$^{-2}$ yr$^{-1}$. Anomalies in CO$_2$ flux from the (2nd row) wind perturbation and (3rd row) freshwater perturbation experiments plotted with contours every 0.05 mol m$^{-2}$ yr$^{-1}$. Positive fluxes indicate outgassing.

and as such, are somewhat difficult to interpret. Generally, our model Southern Ocean is a sink for total contemporary CO$_2$, and this sink strength is growing with time. The spatially-integrated Southern Ocean ($<$30°S) CO$_2$ fluxes exhibit this behavior as well (Fig. 5). We investigate in the following sections how these zonal and spatial mean fluxes evolve under increasing wind stress and increasing stratification.
4. Wind perturbation

a. Natural CO₂ changes

The spatially-integrated (<30°S) Southern Ocean sink of natural CO₂ weakens by about 50% of its control run value by 2100 in the wind perturbation experiment (Fig. 5a). The weaker sink under enhanced wind stress can also be expressed as a positive anomaly in natural CO₂ flux from this region (Fig. 5b). This anomalous outgassing is linearly increasing with time, as wind stress over the Southern Ocean is also linearly increasing. The spatial pattern of this anomalous natural CO₂ flux is shown in Figure 4d. Wind stress causes anomalous outgassing of natural CO₂ nearly everywhere in the Southern Ocean, but there is a concentrated region of outgassing between 40°S and 50°S, and at about 60°S.

We investigate the processes driving the anomalous outgassing of natural CO₂ under enhanced wind stress by decomposing the modeled air-sea CO₂ exchange, \( F_{\text{CO}_2} \), into its contributing parameters (Lovenduski et al., 2007),

\[
F_{\text{CO}_2} = G(p\text{CO}_2^{oc} - p\text{CO}_2^{atm}),
\]
Table 3. Estimated contributions to the change in natural ocean \( pCO_2 \), \( \Delta pCO_2 \), during the wind perturbation experiments, [\( \mu \) atm], averaged over the Southern Ocean (<30°S). \( \Sigma \) is the sum of all five terms, and \( \Delta pCO_{2,mod} \) is the modeled change in \( pCO_2 \).

<table>
<thead>
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<th>year</th>
<th>( \frac{\partial pCO_2}{\partial sDIC} )</th>
<th>( \Delta sDIC )</th>
<th>( \frac{\partial pCO_2}{\partial sAlk} )</th>
<th>( \Delta sAlk )</th>
<th>( \frac{\partial pCO_2}{\partial ffw} )</th>
<th>( \Delta ffw )</th>
<th>( \frac{\partial pCO_2}{\partial \frac{S}{T}} )</th>
<th>( \Delta \frac{S}{T} )</th>
<th>( \frac{\partial pCO_2}{\partial \frac{S}{S}} )</th>
<th>( \Delta \frac{S}{S} )</th>
<th>( \Sigma )</th>
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<td>2000</td>
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<td>0</td>
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<td>2050</td>
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<td>-0.01</td>
<td>-0.11</td>
<td>-0.01</td>
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<td>4.36</td>
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<td>2100</td>
<td>7.27</td>
<td>-2.33</td>
<td>-0.02</td>
<td>-0.20</td>
<td>-0.02</td>
<td>7.30</td>
<td>6.92</td>
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where G is a measure of gas exchange (a function of gas transfer velocity, sea ice fraction, and solubility), \( pCO_2^{oc} \) is the surface ocean \( pCO_2 \), and \( pCO_2^{atm} \) is the \( pCO_2 \) of the overlying atmosphere. As G changes very little between our control and wind-perturbed experiments (constant gas transfer velocity and no sea ice fraction), and \( pCO_2^{atm} \) is constant, the primary mechanism creating anomalies in the natural \( CO_2 \) fluxes under enhanced wind stress are anomalies in the ocean surface \( pCO_2 \), \( \Delta pCO_{2,oc} \). We consider the total derivative for this quantity and its decomposition into contributions from DIC, alkalinity (Alk), temperature, and salinity,

\[
\Delta pCO_{2,oc} = \frac{\partial pCO_{2,oc}}{\partial DIC} \Delta DIC + \frac{\partial pCO_{2,oc}}{\partial Alk} \Delta Alk + \frac{\partial pCO_{2,oc}}{\partial T} \Delta T + \frac{\partial pCO_{2,oc}}{\partial S} \Delta S,
\]

(1)

where \( \frac{\partial pCO_{2,oc}}{\partial DIC} \), \( \frac{\partial pCO_{2,oc}}{\partial Alk} \), \( \frac{\partial pCO_{2,oc}}{\partial T} \), and \( \frac{\partial pCO_{2,oc}}{\partial S} \) are determined from the equations in Sarmiento and Gruber (2006) and the spatially averaged Southern Ocean values in the control simulation (see Appendix for more details). \( \Delta DIC \), \( \Delta Alk \), \( \Delta T \), and \( \Delta S \) are the differences of the spatially-averaged Southern Ocean DIC, Alk, T, and S, respectively in the wind perturbed experiment and the control experiment.

Anomalies in DIC and Alk can be driven by changes in ocean transport/mixing and biological uptake/remineralization, but also by changes in freshwater fluxes (\( \Delta ffw \)). We separate these two components of change by normalizing DIC and Alk with salinity (sDIC and sAlk), and expanding our total \( pCO_{2,oc} \) derivative (see Appendix for more details),

\[
\Delta pCO_{2,oc} = \frac{S}{S_0} \frac{\partial pCO_{2,oc}}{\partial DIC} \Delta sDIC + \frac{S}{S_0} \frac{\partial pCO_{2,oc}}{\partial Alk} \Delta sAlk + \frac{\partial pCO_{2,oc}}{\partial ffw} \Delta ffw + \frac{\partial pCO_{2,oc}}{\partial T} \Delta T + \frac{\partial pCO_{2,oc}}{\partial S} \Delta S,
\]

(2)

whose individual terms are shown in the top section of Table 3.

As the anomalous surface ocean \( pCO_2 \) under increased wind stress exhibits a linear behavior, we show the contributions to \( \Delta pCO_2 \) for three time slices (Table 3). In each
slice, it is clear that the primary driver for the positive anomalies in \( pCO_2 \) is the large positive contribution from the sDIC term, which is being somewhat mitigated by the negative contribution from the sAlk term. Contributions from freshwater, T, and S play comparatively smaller roles.

sDIC is an important constituent of the total \( pCO_2 \) anomaly budget because its surface distribution is linearly increasing with time in the wind perturbation experiment (Fig. 6b). The surface increase is largest in the region between 40°S and 50°S, where wind stress drives anomalous advection of natural DIC via increased upwelling and northward Ekman transport (Fig. 6a, Fig. 7). The linear increase in sDIC, and thus \( pCO_2 \) is driven by the linear change in the overturning circulation, which is responding to the increased wind stress. Ultimately, this process causes a linear decrease in the strength of the Southern Ocean sink for natural \( CO_2 \) (Fig. 5). The decreasing sink strength is most prominent in the regions of anomalous upwelling (\( \sim 60°S \)) and anomalous Ekman transport (\( \sim 45°S \), Fig. 4d), where the surface distribution of sDIC is changing the most.

b. Anthropogenic \( CO_2 \) changes

Our model Southern Ocean sink for anthropogenic \( CO_2 \) strengthens by 0.01 Pg C yr\(^{-1}\) in 2100 under increasing wind stress and with atmospheric \( CO_2 \) stabilization at 450 ppmv (Fig. 5a). This strengthening sink can also be expressed as a negative anomaly in anthropogenic \( CO_2 \) exchange, which exhibits a linear behavior with time (Fig. 5b). The anomalous uptake is found nearly everywhere in the Southern Ocean, but is highest in the region just north of 60°S during the final 50 years of the experiment (Fig. 4e).

The changes in meridional overturning brought on by the increasing wind stress (Fig. 7) cause large changes in the distribution of anthropogenic sDIC in our model. Anthropogenic
Figure 7. Anomaly in residual-mean (eulerian + bolus) meridional overturning streamfunction during the wind perturbation experiment, year 2100. Contours are every 0.5 Sv.

sDIC has a mean distribution that is higher at the surface than at depth (Fig. 6d). When increased wind drives enhanced upwelling in the high latitude regions, there is an anomalous increase in the supply of waters low in anthropogenic sDIC to the surface. This supply lowers the surface concentration of anthropogenic sDIC (Fig. 6e) and allows the ocean to absorb more anthropogenic CO$_2$ from the atmosphere. The anthropogenic CO$_2$ fluxes do not respond as dramatically to changes in wind stress as the natural CO$_2$ fluxes, as the mean horizontal and vertical gradients in anthropogenic sDIC are much smaller than those of natural sDIC (compare Figs. 6a and d).

Increasing wind stress, then, allows the Southern Ocean to absorb slightly more anthropogenic CO$_2$ than it would with constant wind, due to enhanced upwelling of waters low in anthropogenic CO$_2$ and faster exposure of these waters to the atmosphere. These results are in agreement with the study by Russell et al. (2006). When the same wind stress increase acts on a larger gradient in air-sea $p$CO$_2$, however, the magnitude of the changes in anthropogenic CO$_2$ uptake can be even larger. This is shown in Figure 8, where the experiments were run with a more rapid increase in the atmospheric $p$CO$_2$ boundary condition, so that it reached 900 ppmv by 2100. The relative increase in anthropogenic CO$_2$ uptake under enhanced wind stress in this case is 0.04 Pg C yr$^{-1}$ by 2100, approximately four times the increase in the 450 ppmv simulation.

c. Contemporary CO$_2$ changes

Overall, we find the Southern Ocean to be a weaker sink for total contemporary CO$_2$ by the year 2100 in our wind perturbation experiment (Fig. 5a). This is because the natural CO$_2$ outgassing anomaly is growing with time at a faster rate than the anthropogenic CO$_2$
uptake anomaly, so the net effect is an anomaly toward outgassing of total contemporary CO2 by 2100 (Fig. 4f, Fig. 5b).

Using simple box models and coupled general circulation models, several authors (Zickfeld et al., 2007; Le Quéré et al., 2008; Matear and Lenton, 2008) have hypothesized that the coming century will be characterized by a reversal of this contemporary CO2 sink anomaly. They suggest that increasing wind stress, coupled with the predicted rise in atmospheric CO2 will act to first weaken, then strengthen the Southern Ocean sink for contemporary CO2. While we do not observe this behavior when our atmospheric pCO2 boundary condition reaches 450 ppmv by 2100 (Fig. 5), our model does exhibit this behavior when this boundary condition rises to 900 ppmv by 2100 (Fig. 8). In this case, natural CO2 uptake continues to weaken under enhanced wind stress, but anthropogenic CO2 uptake strengthens at a faster rate, causing a reversal in the net contemporary CO2 flux anomaly at \(\sim2050\). Thus, our results are sensitive to the atmospheric pCO2 boundary condition used in our simulations.

d. Wind position perturbation

In order to separately examine the relative importance of wind magnitude and position changes, we conduct an additional experiment where the magnitude of the maximum wind
stress remains constant, but its position moves poleward by 0.75° over 120 years (Table 1). Moving the maximum wind stress reduces the surface ocean $pCO_2$ by a small amount, primarily as a consequence of lower DIC in the surface waters (Table 3). This wind movement is accompanied by a 0.005 kg m$^{-3}$ reduction in high latitude surface density, which decreases the size of the outcrop area for DIC-rich waters (Tschumi et al., 2008), and lowers the natural CO$_2$ outgassing from this region. While the sign of this change in natural CO$_2$ flux is opposite to that of the wind strength perturbation experiment (Section 4a), its magnitude is much smaller.

5. **Freshwater perturbation**

a. **Natural CO$_2$ changes**

The spatially-integrated Southern Ocean sink for natural CO$_2$ strengthens in the freshwater perturbation simulation, as compared to the control run (Fig. 5a). This negative anomaly in natural CO$_2$ flux is evident everywhere in the Southern Ocean, but is concentrated in the high latitude regions (Fig. 4g). A linear change in freshwater supply causes a non-linear change in the integrated flux anomalies with time (Fig. 5b). At first, increased freshwater causes the sink strength to increase dramatically, but this behavior does not continue beyond 2060, where the anomalous time series stops decreasing and begins to slightly increase.

We investigate the processes responsible for the behavior of the freshwater perturbation anomalies in natural CO$_2$ using the technique described in Section 4a. We decompose the anomaly in surface ocean $pCO_2$, $\Delta pCO_2$ into the components contributing to change at each time step and display the results in Figure 9. This analysis reveals that anomalies in surface ocean $pCO_2$ are primarily a result of contributions from the sDIC term, with the sAlk term slightly mitigating the response. As was the case for increasing wind
stress, changes in freshwater, T, and S play comparatively smaller roles in the overall budget.

Adding freshwater to the Southern Ocean decreases surface density, resulting in dramatic changes in stratification (Fig. 10). Increased stratification causes the isopycnal surfaces to flatten, reducing the outcrop area for upwelling water and changing the entire surface and interior ocean distribution of natural sDIC (Fig. 6c). The result is a surface ocean characterized by much lower sDIC, and therefore $pCO_2$, and an accumulation of sDIC in the deep parts of the high latitudes. Stratification also lowers surface ocean sAlk, although its effect on $pCO_2$ is smaller.

The non-linearity in the surface ocean $pCO_2$ anomaly is closely tied with the non-linear behavior in the surface ocean sDIC. This behavior appears a result of the slow flattening of the isopycnal surfaces under increased freshwater forcing (Fig. 10), which changes the characteristics of the upwelling/outcropping water in the high-latitudes. After a certain time (~2060 in our simulation), the deep water enriched in sDIC is prevented from reaching the surface and continued flattening of isopycnals has no impact on surface sDIC concentrations.

Increased stratification also has an impact on nutrient supply and biological production in our model, although its impact on the air-sea CO$_2$ flux is only a small component of change. Fig. 9 (dashed lines) shows the $\Delta pCO_2$ budget results from an identical freshwater simulation, but one where export production was fixed, rather than evolving with surface phosphate concentration. A visual comparison of these time series with the ones from the original simulation indicates that the increase in stratification slowly decreased the supply of phosphate to the surface ocean, and slowed the biological drawdown of DIC and Alk. However, the overall influence of this biological slow-down is of little consequence for the total natural $pCO_2^\text{oc}$ budget.

Figure 10. Annual-mean, zonal-mean potential density in the (a) control and (b) freshwater perturbation experiments in the year 2100. Contour intervals are 0.2 kg m$^{-3}$. 
b. Anthropogenic CO₂ changes

The Southern Ocean sink for anthropogenic CO₂ weakens considerably in our freshwater perturbation experiment, relative to the control experiment (Fig. 5a). This results in positive anthropogenic CO₂ flux anomalies which exhibit a non-linear behavior with time (Fig. 5b). The anomalies are concentrated in the high latitude regions of the Southern Ocean (Fig. 4h).

The increased stratification brought on by increases in surface freshwater slows the upwelling of waters depleted in anthropogenic sDIC (Fig. 6d, 6f), reducing the ability of the surface ocean to absorb anthropogenic CO₂ from the atmosphere. This result is consistent with that of Sarmiento et al. (1998), who find that century-scale changes in stratification reduce the downward flux of anthropogenic carbon in the Southern Ocean.

The stratification-imposed weakening of the Southern Ocean sink for anthropogenic CO₂ is amplified as the ocean-atmosphere gradient in pCO₂ increases. Figure 8 shows the anthropogenic CO₂ flux response to the same stratification increase in a simulation where the atmospheric boundary condition for pCO₂ grows to 900 ppmv by 2100. The anomalous response to stratification is even more dramatic in this case, with the anthropogenic CO₂ sink weakening to about 50% of its value in the control run by 2100.

c. Contemporary CO₂ changes

The non-linear response of natural and anthropogenic CO₂ flux to increased freshwater supply causes the total sink for contemporary CO₂ to first strengthen, then weaken, as compared to the control simulation (Fig. 4i, 5). At the beginning of the freshwater perturbation simulation, anomalies in natural CO₂ uptake dominate the anomalies in anthropogenic CO₂ outgassing, but after ∼2040, the anthropogenic CO₂ outgassing anomalies begin to dominate, and the net response is reversed. While enhanced stratification may increase the drawdown of atmospheric CO₂ at first, by 2100, our simulation suggests that the Southern Ocean will become a weaker sink for total contemporary CO₂.

Increasing atmospheric CO₂ amplifies the weakening of the anthropogenic CO₂ flux in response to higher stratification, overwhelming the anomalous natural CO₂ uptake, and causing the Southern Ocean sink for contemporary CO₂ to weaken throughout the course of the simulation, as compared to the control run (Fig. 8). This implies that a future characterized by increased Southern Ocean stratification and rapidly increasing atmospheric CO₂ concentrations will also be characterized by a weakening Southern Ocean CO₂ sink.

6. Simultaneous wind and freshwater perturbations

We have demonstrated that increasing wind and increasing stratification separately lead to large changes in the Southern Ocean sink for contemporary CO₂ by 2100, through a complex interaction of the circulation with natural and anthropogenic CO₂. It is worth contemplating, then, how a future characterized by both increased wind stress and increased stratification will impact the Southern Ocean CO₂ sink. Figure 11 shows results from two sets of experiments where both wind stress and freshwater forcing linearly increase with
time. In the first (second) set, the atmospheric $p$CO$_2$ boundary condition increases to 450 (900) ppmv by 2100. Both sets of simulations indicate that the Southern Ocean sink for total contemporary CO$_2$ will be weaker under simultaneous increases in wind stress and stratification, as compared to control simulations with no changes in physical forcing. The net response of the CO$_2$ fluxes to these simultaneous changes is essentially a linear superposition of the responses to the individual changes. Simultaneous changes in wind and stratification are likely to occur in the future, as shifts in the position and intensity of wind stress and precipitation are naturally correlated in the storm track.

7. Discussion and conclusions

We conclude that the Southern Ocean CO$_2$ sink is very sensitive to changes in wind stress and stratification. A linear increase in wind stress drives a linear decrease in contemporary CO$_2$ uptake, as anomalies in natural CO$_2$ outgassing overwhelm enhanced anthropogenic CO$_2$ uptake. These anomalies are caused by changes in the upwelling and northward surface transport of DIC. Wind-induced changes in anthropogenic CO$_2$ are sensitive to the atmospheric $p$CO$_2$ boundary condition. A poleward shift in the wind stress reduces the high-latitude isopycnal outcrop area and slightly decreases natural CO$_2$ outgassing. We find that a linear increase in freshwater forcing leads to first an increase, then a decrease in contemporary CO$_2$ uptake, caused by a complex interaction of the non-linear natural and anthropogenic CO$_2$ anomalies. Increased stratification causes a shift in the location of isopycnal outcrop, creating changes in the DIC distribution. Simultaneous increases in wind stress and stratification lead to a weaker sink for atmospheric CO$_2$ in the Southern Ocean by 2100.
Our simple approach to providing a mechanistic understanding of the CO$_2$ flux changes with changes in wind stress and stratification does not come without caveats. The coarse resolution of our model domain does not permit eddy-scale interactions to occur, and their effects must be parameterized. Recent literature highlights the importance of these eddies in controlling the long-term circulation response to wind stress (Böning et al., 2008), and the mean anthropogenic CO$_2$ transport in the Southern Ocean (Ito et al., 2010). Also, our results de-emphasize the role of biological uptake in controlling the CO$_2$ flux response in the Southern Ocean, in contrast to idealized simulations on millennial timescales (Tschumi et al., 2008). Restricting our analysis to the Southern Ocean and to short time scales precludes us from investigation of the global biological response to the slow change in thermocline nutrients that may originate in the Southern Ocean (Sarmiento et al., 2004a; Dutkiewicz et al., 2005).

Despite these caveats, and the somewhat unrealistic formulation of our model and perturbation experiments, results from our idealized experiments highlight the potential for the Southern Ocean to experience large changes in the coming century, and argue strongly for improved observational effort in this region.

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APPENDIX

$pCO_2^c$ partial derivatives

Anomalies in surface ocean $pCO_2$ during the perturbation experiments, ($\Delta pCO_2^c$) are decomposed into the contributions from DIC, Alk, T, and S according to (1). We use the following set of equations to approximate the Southern Ocean ($<30^\circ$S) mean values of the $pCO_2^c$ partial derivatives (Sarmiento and Gruber, 2006),

$$\frac{\partial pCO_2^c}{\partial DIC} = \frac{pCO_2^c}{DIC} \times \gamma_{DIC}$$

$$\frac{\partial pCO_2^c}{\partial Alk} = \frac{pCO_2^c}{Alk} \times \gamma_{Alk}$$

$$\frac{\partial pCO_2^c}{\partial T} \approx pCO_2^c \times 0.0423^\circ C^{-1}$$

$$\frac{\partial pCO_2^c}{\partial S} \approx \frac{pCO_2^c}{S},$$
where the buffer factors can be approximated with

\[
\gamma_{\text{DIC}} \approx \frac{3\text{Alk} \cdot \text{DIC} - 2\text{DIC}^2}{(2\text{DIC} - \text{Alk}) (\text{Alk} - \text{DIC})}
\]

\[
\gamma_{\text{Alk}} \approx -\frac{\text{Alk}^2}{(2\text{DIC} - \text{Alk}) (\text{Alk} - \text{DIC})}.
\]

**Salinity normalization**

To separate the contribution from freshwater fluxes on DIC and Alk, we use the following two equations for the first and second terms of (1),

\[
\frac{\partial pCO_2^{\text{ec}}}{\partial \text{DIC}} \Delta \text{DIC} = \frac{\partial pCO_2^{\text{ec}}}{\partial (S/S_0 \text{ sDIC})} \Delta (S/S_0 \text{ sDIC}) = \frac{s\text{DIC}}{S_0} \frac{\partial pCO_2^{\text{ec}}}{\partial \text{DIC}} \Delta S + \frac{S}{S_0} \frac{\partial pCO_2^{\text{ec}}}{\partial \text{DIC}} \Delta \text{sDIC}
\]

\[
\frac{\partial pCO_2^{\text{ec}}}{\partial \text{Alk}} \Delta \text{Alk} = \frac{\partial pCO_2^{\text{ec}}}{\partial (S/S_0 \text{ sAlk})} \Delta (S/S_0 \text{ sAlk}) = \frac{s\text{Alk}}{S_0} \frac{\partial pCO_2^{\text{ec}}}{\partial \text{Alk}} \Delta S + \frac{S}{S_0} \frac{\partial pCO_2^{\text{ec}}}{\partial \text{Alk}} \Delta \text{sAlk}.
\]

We extract the first terms from the above two equations, as they represent the contribution from freshwater forcing (fw) on \(pCO_2^{\text{ec}}\),

\[
\frac{\partial pCO_2^{\text{ec}}}{\partial \text{fw}} \Delta \text{fw} = \frac{s\text{DIC}}{S_0} \frac{\partial pCO_2^{\text{ec}}}{\partial \text{DIC}} \Delta S + \frac{s\text{Alk}}{S_0} \frac{\partial pCO_2^{\text{ec}}}{\partial \text{Alk}} \Delta S.
\]

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