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Changes in dissolved oxygen in the Southern Ocean with climate change

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[1] **Abstract:** Climate models predict that increasing greenhouse gases in the atmosphere will change our climate. However, the range of observations available to assess these model predictions is limited. We have utilized an ocean biogeochemical model to project the impact of climate change on the oxygen cycle in the ocean and the air-sea exchange of oxygen. With climate change, our model predicts an outgassing of oxygen from the ocean. In the model, the Southern Ocean is the region that displays the largest change in the air-sea exchange of oxygen with climate change. The reduced oxygen uptake in the Southern Ocean is dominated by a reduction in convective mixing and in the subduction rate of bottom water which decreases the amount of dissolved oxygen in the Southern Ocean. Our model results suggest that dissolved oxygen is sensitive to changes in Southern Ocean circulation revealing the potential to use dissolved oxygen to test model predictions of climate change. From Southern Ocean observations collected 28 years a part we observed a decrease in dissolved oxygen that was consistent with model climate change simulation. This provides credibility to the climate model predictions. Additional analysis and observations of dissolved oxygen are needed to substantiate our climate model simulations.

Keywords: Dissolved oxygen; climate change; outgassing; climate change detection; climate model validation.

Index terms: Biogeochemical processes; climate dynamics; oceans; biogeochemical cycles.

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1. Introduction

[2] The concentration of CO_2 and other greenhouse gases are increasing in the atmosphere and changing the radiative properties of the atmosphere. Climate model simulations with increasing atmospheric greenhouse gases predict a slow down in the ocean thermohaline

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circulation [*Manabe and Stouffer*, 1993; *Hirst*, 1999; *Wood et al.*, 1999]. The climate model predictions differ in magnitude and timing of the change, but the change is a robust feature of global warming simulations [*Rahmstorf*, 1999]. The reduction in thermohaline circulation is caused by an increase in density stratification in the high latitudes of both hemispheres. The increased density stratification reduces convective mixing and decreases deepwater formation in the North Atlantic and Southern Ocean. Both of these processes are important for the ventilation of the high-latitude Ocean.

[3] We use the CSIRO climate model output [*Gordon and O'Farrell*, 1997; *Hirst et al.*, 2000] to investigate the impact of climate change on ocean biogeochemical cycles. The CSIRO climate model uses the *Gent and McWilliams* [1990] (GM) scheme for parameterization of eddy-induced transport in the ocean. The GM parameterization reduces deep convective mixing and produces realistic density stratification in the Southern Ocean [*Danabasoglu et al.*, 1994; *Hirst et al.*, 1996; *Hirst and McDougall*, 1996]. These improvements in the representation of the Southern Ocean circulation are crucial in achieving realistic oceanic ventilation rates [*Robitaille and Weaver*, 1995; *England and Hirst*, 1997; *Caldeira and Duffy*, 2000].

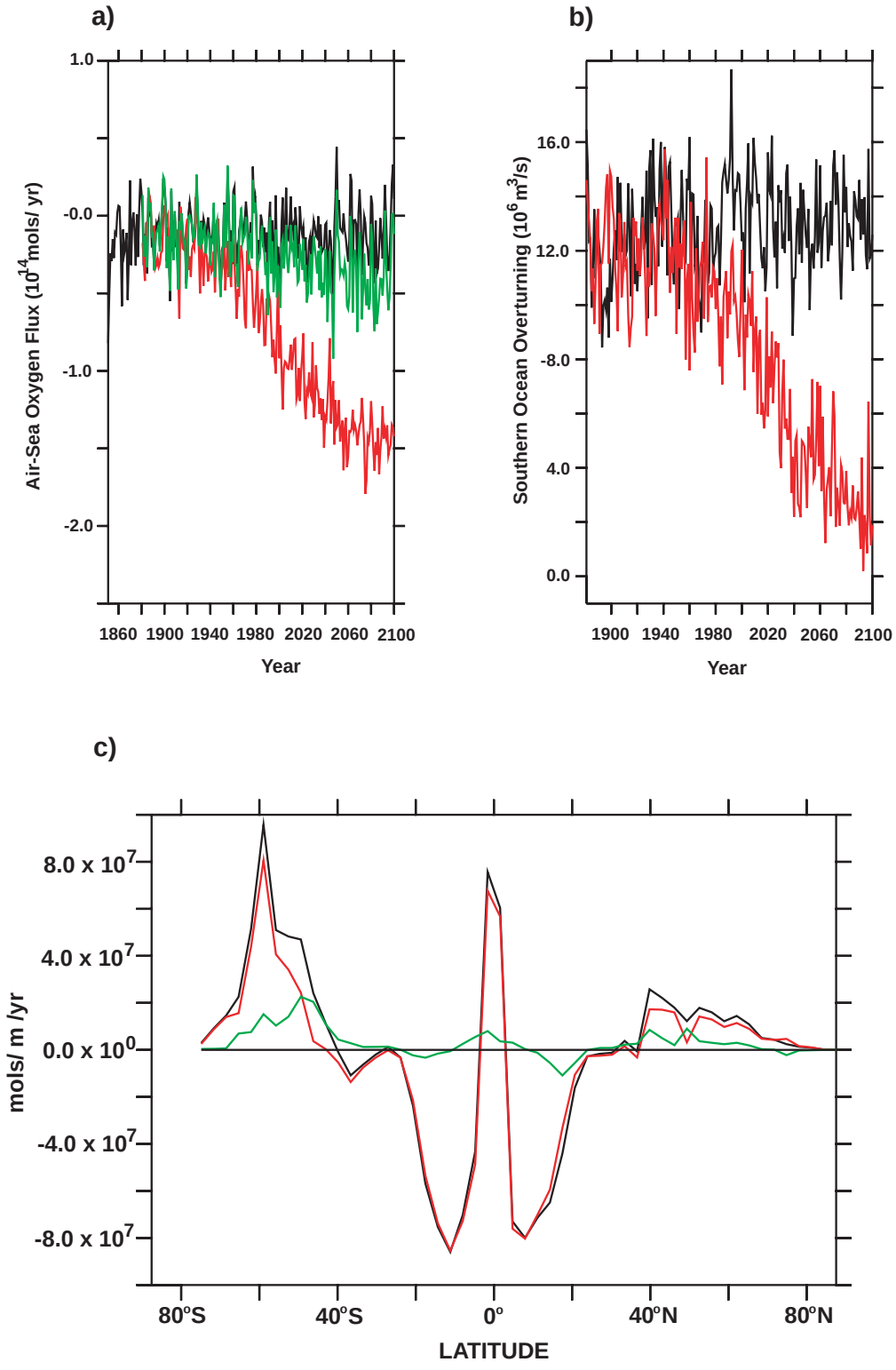
[4] We performed ocean carbon cycle runs using archived climate model output following a technique described by *Aumont et al.* [1998]. The archived data used are the three-dimensional fields of ocean temperature, salinity and currents, and the surface fields of wind speed and sea ice cover. Archived monthly mean data provided the forcing fields for our ocean carbon cycle model, which determines distributions of dissolved inorganic carbon (DIC), alkalinity, phosphate, dissolved oxygen, and dissolved nitrogen in the ocean [*Matear and Hirst*, 1999]. An initial study, which describes our modeling approach, focused on the impact of climate change on the future oceanic uptake of CO_2 [*Matear and Hirst*, 1999]. Here we explore the impact of climate change on the oxygen cycle and demonstrate the potential of using biogeochemical tracers to test the climate model predictions.

2. Results

[5] We utilized model output from two experiments of the CSIRO climate model: (1) a control experiment with a constant atmospheric level of CO_2 and (2) a greenhouse gas forcing experiment where the equivalent CO_2 followed the IS92a radiative forcing scenario [*Houghton et al.*, 1995] from 1850 until 2100. We refer to climate change as the difference between experiment 2 and 1.

[6] With climate change, our model predicted an outgassing of oxygen from the oceans (Figure 1). The simulated oxygen outgassing from the ocean with climate change may reflect changes in ocean circulation, biological production and the solubility of oxygen in the upper ocean. Climate change leads to a global warming of the surface ocean [*Matear and Hirst*, 1999]. The warming of the surface ocean

Figure 1. (a) Global averaged oceanic uptake of oxygen (10^{14} mols/yr), (b) intermediate and deep water overturning in the Southern Ocean (10^6 m^3/s), and (c) annual mean zonal averaged oceanic uptake of oxygen (mol m^{-1} yr^{-1}) for period of 1990–2000. In these plots the black line denotes the Control experiment, and the red line denotes the Greenhouse Gas forcing experiment. In Figure 1a the green line denotes the control experiment with the solubility of oxygen calculated using the sea surface temperature from the Greenhouse Gas forcing experiment. In Figure 1c the green line denotes the effect of climate change (black minus red line).


a)
1.0
-0.0
-1.0
-2.0
Air-Sea Oxygen Flux (10^{14}mols/ yr)
1860 1900 1940 1980 2020 2060 2100
Year
b)
16.0
12.0
-8.0
4.0
0.0
Southern Ocean Overturning (10^6 m^3/s)
1900 1940 1980 2020 2060 2100
Year
c)
8.0 x 10^7
4.0 x 10^7
0.0 x 10^0
-4.0 x 10^7
-8.0 x 10^7
mols/ m /yr
80°S 40°S 0° 40°N 80°N
LATITUDE

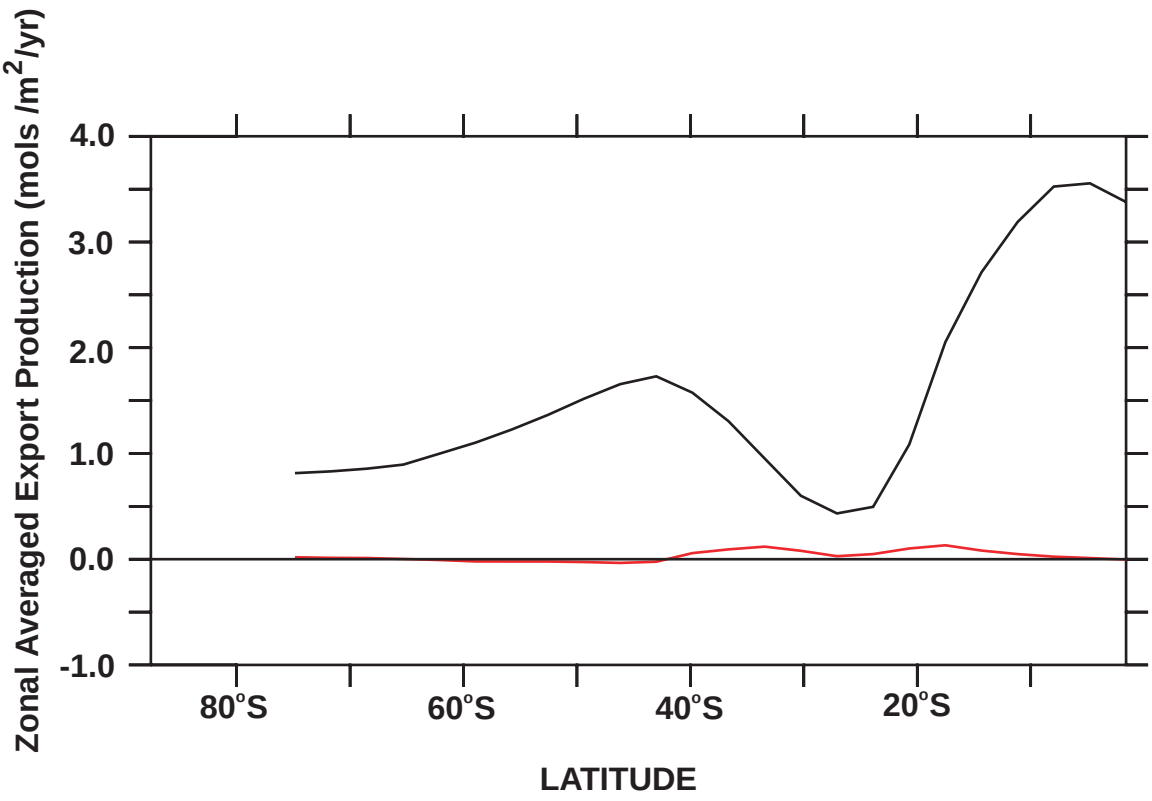


Figure 2. For the 1990–2000 period the zonal averaged export production of the control experiment (black line) and the difference between the greenhouse gas experiment and the control experiment (red line).

reduces the solubility of O_2 in the surface ocean, which decreases the sea surface dissolved oxygen. To assess the impact of sea surface warming, we performed a simulation that used the control experiment forcing data with the sea surface temperatures predicted by the greenhouse gas experiment to determine the solubility of oxygen in the surface water. The simulation demonstrates that oxygen outgassing by the ocean with climate change is weakly influenced by solubility changes associated with sea surface warming (Figure 1a).

[7] The outgassing of oxygen from the ocean with climate change is dominated by reduction in the oxygen uptake by the ocean south of 40°S (Figure 1c) and we focus on this region. South of 40°S, our model estimated little change in biological export production with climate change (Figure 2). Therefore the simulated outgassing of oxygen primarily reflects changes in the ocean circulation in the Southern Ocean with climate change.

[8] Our climate model predicts an increase in density stratification in the Southern Ocean and a reduction in both the rate of subduction of Antarctic Bottom Water (AABW) and convective mixing with climate change [*Hirst*, 1999; *Matear and Hirst*, 1999]. The control experiment displays strong downwelling near Antarctica (south of 60°S) with $10–16 \times 10^6$ m^3 s^{-1} (Figure 1b) of water flowing through 1250 m in the Antarctic region. With greenhouse gas forcing, this downwelling greatly diminishes to 2×10^6 m^3 s^{-1}. In the Southern Ocean the

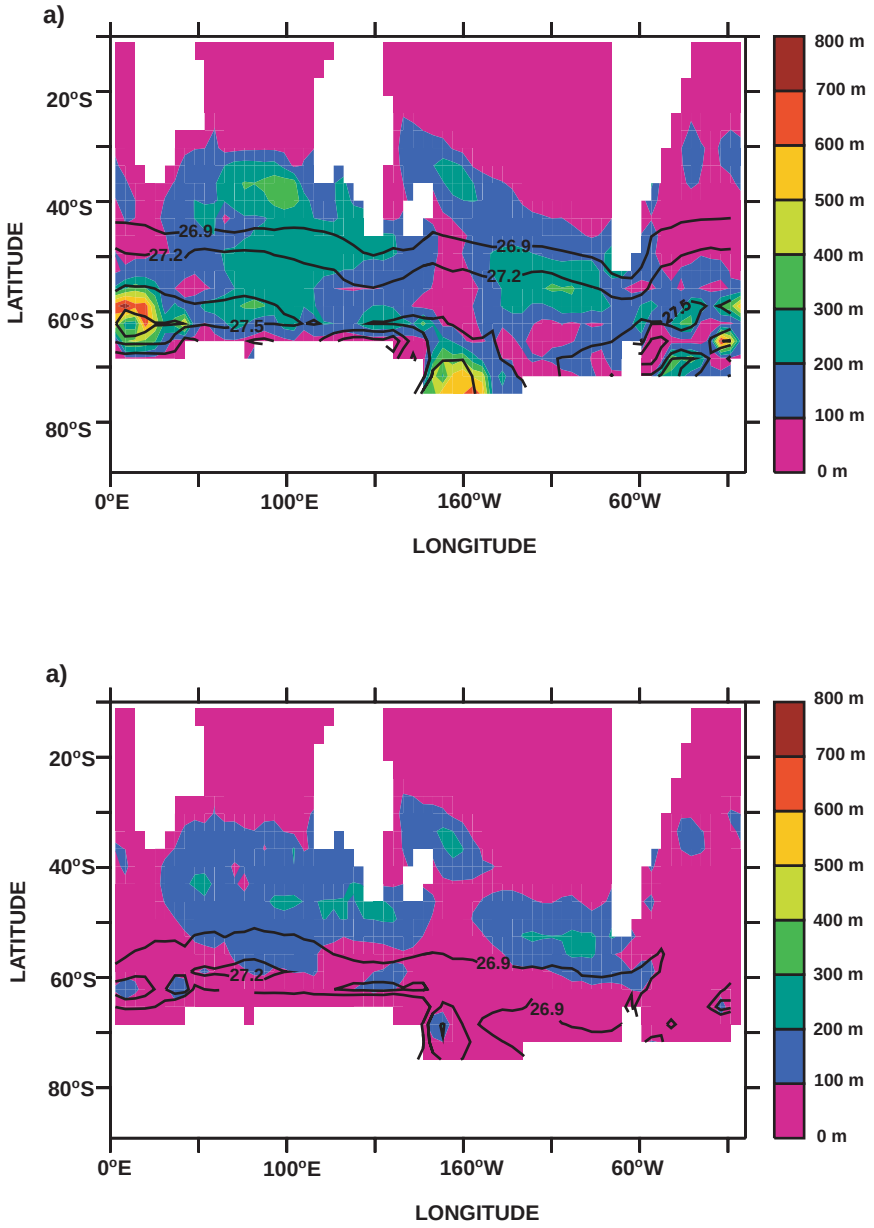


Figure 3. Averaged maximum depth of winter convection for the 2080–2100 period for the (a) Control experiment and (b) Greenhouse Gas forcing experiment. The contour lines (27.2, 27.5, and 27.6) show the maximum potential density of the surface water in the winter. The AAIW is formed in the region where the surface water density is between 27.2 and 27.5, and AABW is formed where density is greater than 27.5.

control experiment produces winter convective depths locally approaching 800 m (Figure 3a). In the greenhouse gas forcing experiment the depth of winter convection is reduced to less than 300 m with the largest reductions occurring south of 60°S (Figure 3b).

[9] Both the reductions in the rate of subduction of AABW and depth of winter convection would reduce the oceanic uptake of oxygen in the Southern Ocean and the supply of oxygen to the ocean interior. These changes would reduce the dissolved oxygen levels in the ocean interior. The zonal averaged change in dissolved oxygen with climate change for the 2080–2100 period shows a maximum decrease in oxygen of 70 umol kg^{-1} (Figure 4a). The maximum decrease occurs south of 60°S at a depth of ~350 m. In contrast, changes in the surface concentrations are much smaller, with an actual increase in dissolved oxygen in the surface waters south of 60°S due to a regional cooling (but freshening) of the surface water. Clearly, this pattern is not consistent with a passive response to a solubility-related surface oxygen loss but instead implies that changes in ocean circulation play a major role. At the surface the oxygen anomaly is controlled only by changes in solubility because the equilibration of surface water with atmospheric oxygen (≈7 days) is much faster than the influence of ocean circulation. This prevents the ocean circulation from influencing surface dissolved oxygen. Below the surface, dissolved oxygen levels are controlled by the supply of oxygen through water mass subduction and winter ventilation and the consumption of oxygen by the remineralization of organic matter. The change in export production in the Southern Ocean with climate change was small (Figure 2) and integrating the change in export production from 1880 to 2100 would produce an average change in dissolved oxygen in the upper 1000 m of less than 5 $\mu mol/kg^{-1}$. Therefore it is the reduction in the supply of oxygen to the ocean interior and the maintenance of a constant consumption of oxygen that generates the middepth decreases in oxygen with climate change.

[10] The presence of the maximum decrease in dissolved oxygen south 60°S at 350 m is consistent with the reduction of convective ventilation of Antarctic waters with climate change (Figure 3). The oxygen anomaly propagates northward into the subtropics with the subduction of the Antarctic Intermediate Water and downward into the deep ocean with the subduction of AABW. The oxygen anomaly in the Antarctic Intermediate Water is associated with negative salinity anomaly that also moves into the subtropics with the subduction of this water mass (Figure 4d).

[11] Changes in export production could give rise to subsurface oxygen anomalies. By defining a new tracer, OP, as O_2-phosphate × $R_{O/P}$ (where $R_{O/P}$ is −136, which is the ratio used in the model and is based on *Redfield et al.* [1963]), one obtains a tracer that is conserved in the ocean interior. The OP tracer is conserved in the ocean interior because any change in the amount of remineralized organic material would consume oxygen and produce phosphate at the O/P ratio prescribed in the model (136). The OP tracer is equivalent to the PO_4* tracer [*Broecker et al.*, 1998] except we express it in dissolved oxygen units to compare it to the simulated dissolved oxygen changes. The change in OP with climate change (Figure 4b) shows the same features as the oxygen change (Figure 3a) and demonstrates that changes in export production are not contributing significantly to the modeled oxygen anomalies in the Southern Ocean.

[12] Observational evidence shows a warming [*Bindoff and Church*, 1992] in the sub-Antarctic Mode Water (300–700 m) and freshening [*Wong et al.*, 1999] in the Antarctic Intermediate Water (AAIW) of the South Pacific. The temperature and salinity changes are inferred to reflect the warming and freshening of the subducted sub-Antarctic Mode Water [*Bindoff and Church*, 1992] and Antarctic Intermediate Water [*Wong et al.*, 1999], respectively. The temperature and salinity anomalies are related

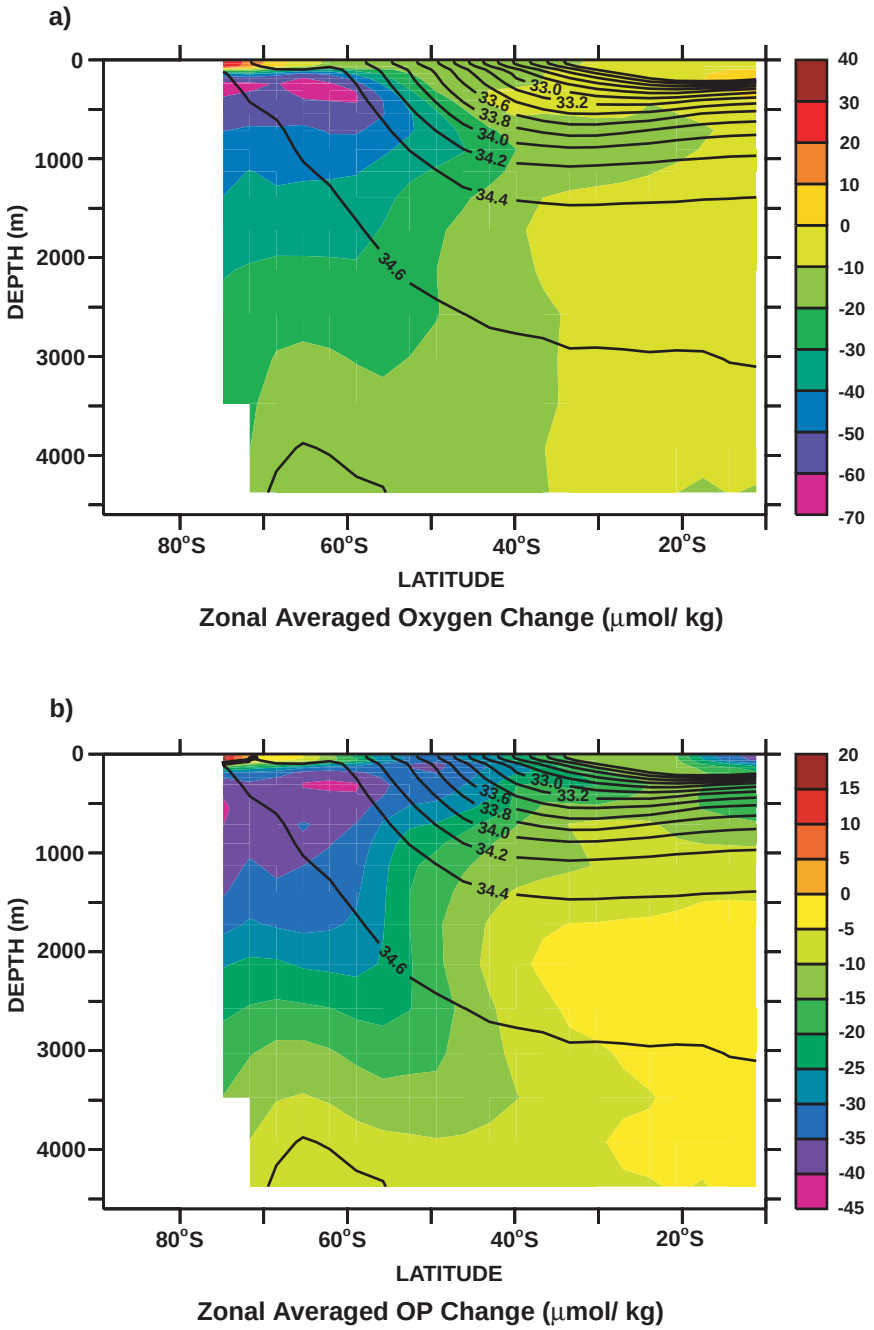


Figure 4. Zonal averaged change along depth surfaces between the Greenhouse Gas forcing experiment and the Control experiment during the period 2080–2100 for (a) dissolved oxygen (μmol kg^{-1}), (b) OP (μmol kg^{-1}), (c) temperature, and (d) salinity. Shown in these sections are the averaged potential density surfaces referenced to 1500 dbars from the control experiment for the 2080–2100 period. The 1500 dbar reference was chosen to adequately display density stratification in both the deep and surface water. This choice of reference level is also more representative of isopycnal flow in the ocean, as defined by neutral surfaces [*McDougall*, 1987], than a surface reference level.

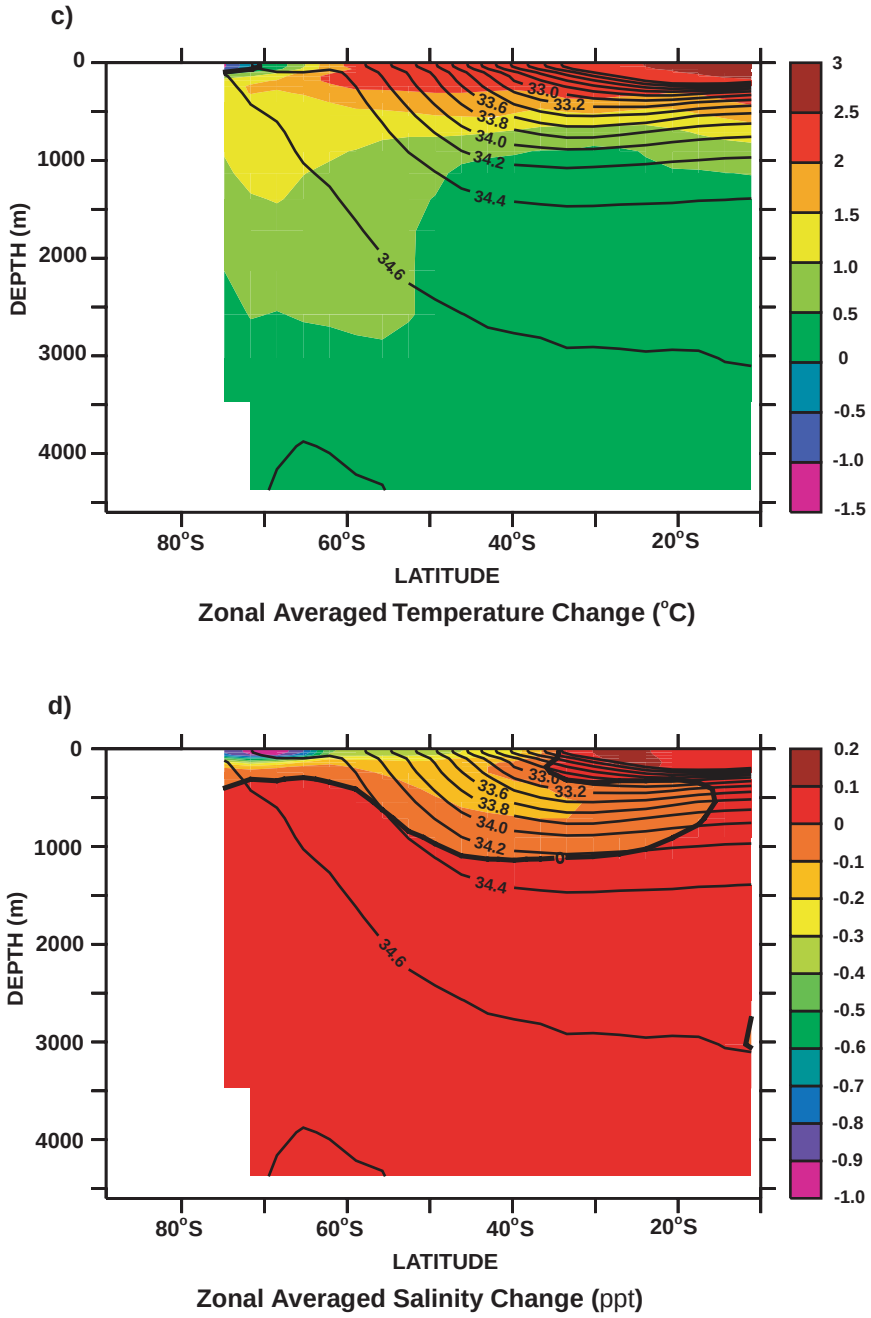


Figure 4. (continued)

to changes in surface properties, whereas the oxygen anomalies are driven more by changes in the rate of ventilation. Therefore it is expected that the climate change anomaly pattern for dissolved oxygen changes would differ from temperature and salinity. The zonal averaged change in temperature and salinity (Figure 4c and 4d) confirms this conclusion. The tem-

Table 1. Comparison of the Observed Average Change in Ocean Temperatures With the Simulated Climate Change Values for the South Hemisphere for the 1950–1995 Period[a]

Depth Interval, m	Observed Temperature Change, °C	Simulated Temperature Change, °C
0–300	0.3 ± 0.1	0.33
0–1000	0.1 ± 0.03	0.15
0–3000	0.05 ± 0.02	0.07

[a] The observed changes were extracted from *Levitus et al.* [2000].

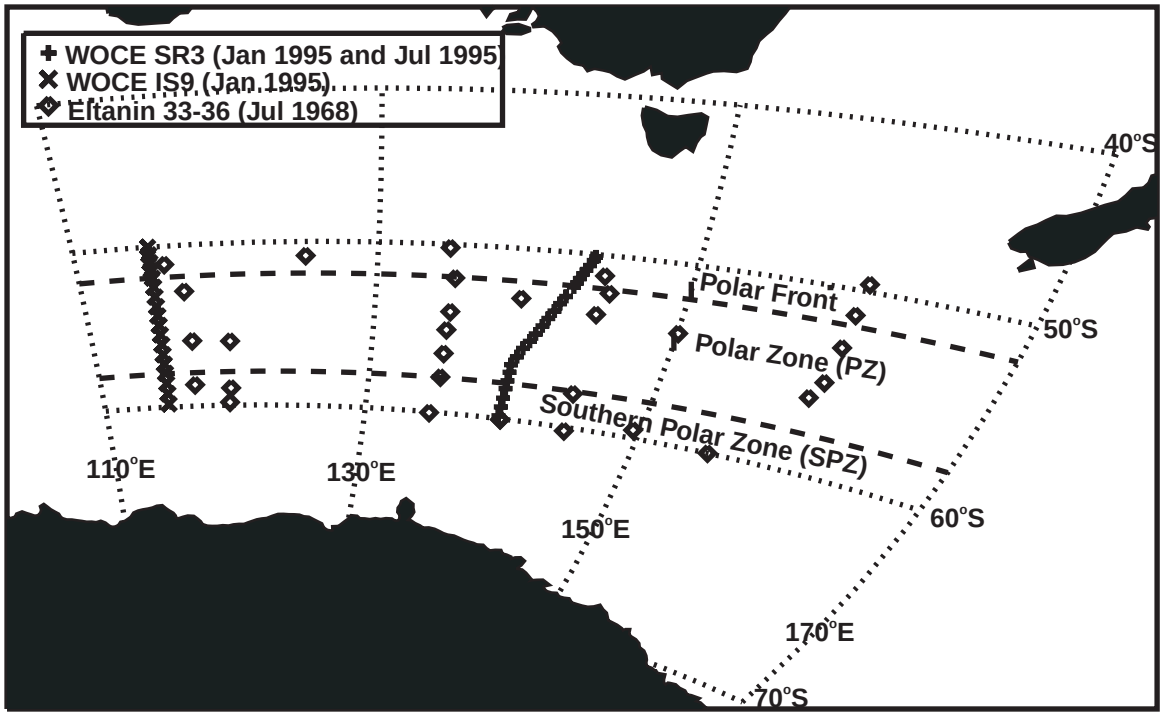
perature and salinity changes are greatest at the surface and decrease with depth. The dissolved oxygen decrease displays a prominent subsurface maximum. The unique information in the oxygen anomaly implies that oxygen data contains independent information, and it is a valuable complement to the temperature and salinity data.

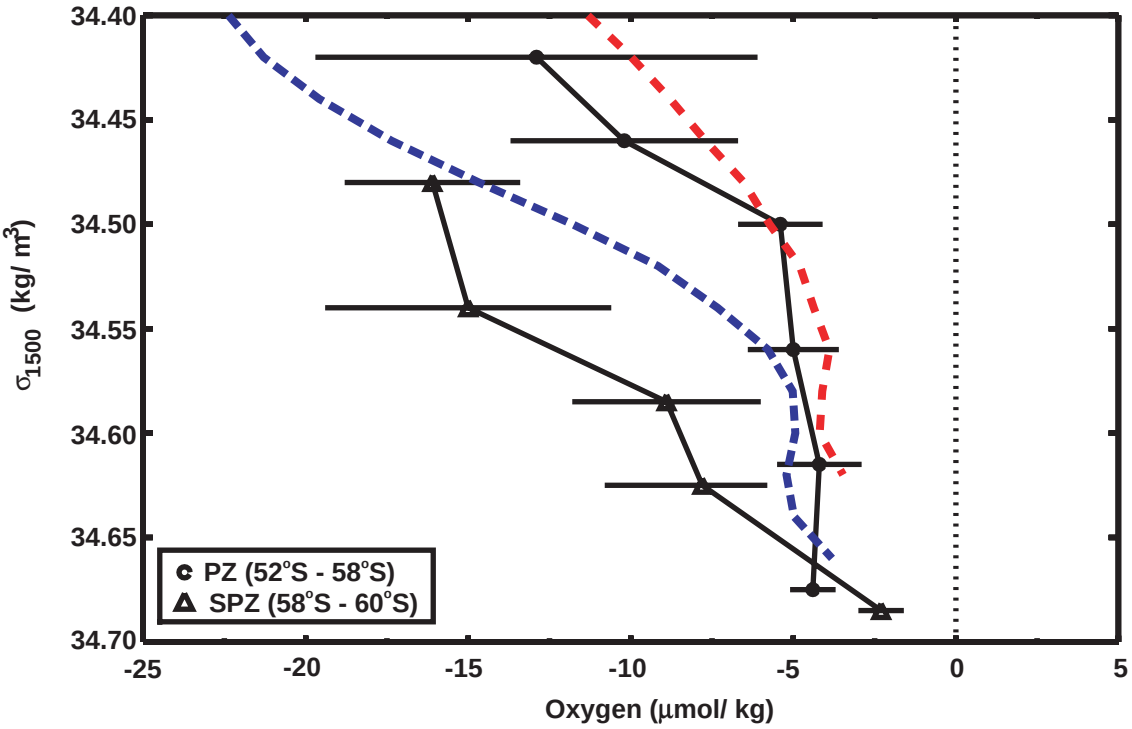
[13] To assess the model performance, we compared the simulated increase in ocean temperatures with climate change in the South Hemisphere for the 1950–1995 period to the observed changes [*Levitus et al.*, 2000]. Although our model simulated greater increase in temperatures than observed (Table 1), the values were comparable to the observations and imply that our climate model is not projecting an unrealistic climate change response.

3. Discussion and Conclusions

[14] We investigated changes in dissolved oxygen in the Southern Ocean south of Australia by comparing recent World Ocean Circulation Experiment (WOCE) measurements (SR3 and IS9 sections [*Rosenberg et al.*, 1997]) to the 1968 Eltanin Cruise [*Jacobs et al.*, 1970]. On the basis of the WOCE sections we confined the comparison to two regions, the Polar Zone (53°S–59°S) and the Southern Polar Zone (59°S–61°S). The combined WOCE data, which included two SR3 sections (January 1995 and July 1995) and the IS9 section (January 1995) displayed ±1 μmol/kg^{-1} variability in dissolved oxygen along σ_{1500} (potential density surfaces referenced to 1500 m) surfaces that were deeper than ~400 m. Above 400 m, seasonal cycle of dissolved oxygen was too large to enable the detection of decadal changes in dissolved oxygen. However, below 400 m the spatial and temporal consistency in the WOCE observations suggested that the data would be suitable for detecting decadal changes. By comparing the dissolved oxygen along σ_{1500} surfaces we observed a decrease in subsurface dissolved oxygen from the Eltanin Cruise to the WOCE data (Figure 5). The dissolved oxygen was greatest in the intermediate water, but changes were also evident in the deep water. In the intermediate water the magnitude of the oxygen changes increased as one moved south. The observed dissolved oxygen decreases were comparable to the pattern and

Figure 5. Dissolved oxygen changes from 1968 to 1996 along isopycnal surfaces referenced to 1500 dbars in the Polar Zone (PZ), 53°S–59°S, and Southern Polar Zone (SPZ), 59°S–61°S. The error bars denote the 95% confidence interval. The dashed lines show the modeled change (control run — greenhouse gas forcing run) in the zonal averaged oxygen (110°E to 170°E) for the 1990–1999 period at 60°S (blue) and 55°S (red). For the model simulation the interannual variability in dissolved oxygen was a maximum at a density of 33.4, where it was ±4 and ±3 μmol kg^{-1} for 55°S and 59°S, respectively. At densities greater than 33.5 the modeled interannual variability in dissolved oxygen was less than ±1 μmol kg^{-1}.


WOCE SR3 (Jan 1995 and Jul 1995)
WOCE IS9 (Jan 1995)
Eltanin 33-36 (Jul 1968)
40°S
50°S
60°S
70°S
Polar Front
Polar Zone (PZ)
Southern Polar Zone (SPZ)
110°E
130°E
150°E
170°E


34.40
34.45
34.50
34.55
34.60
34.65
34.70
σ1500 (kg/ m3)
-25
-20
-15
-10
-5
0
5
Oxygen (µmol/ kg)
PZ (52°S - 58°S)
SPZ (58°S - 60°S)

magnitude estimated by the climate change simulation for 1990s (Figure 5). The observations suggest a climate change reduction in the ventilation of intermediate and deep water in the Southern Ocean. Recent analysis of CFC and nutrient data also suggests that the rate of AABW formation may have decreased over the last century [*Broecker et al.*, 1998] with the inferred rate of AABW from CFC data approximately half the 16×10^6 m^3 s^{-1} estimated from the nutrient data.

[15] Although we hypothesized that the observed decrease in dissolved oxygen in the Southern Ocean south of Australia reflected climate change, other processes may be responsible. The observed changes may reflect multiyear variability in Southern Ocean ventilation or export production. One potential source of multiyear variability in the Southern Ocean is the circumpolar wave [*White and Peterson*, 1996]. Further analysis of the repeat observations of dissolved oxygen may help to verify the hypothesis that climate change is reducing dissolved oxygen in the Southern Ocean. Such work would also provide valuable observations to test global warming predictions of climate models.

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