Climate Model Simulations of Effects of Increased Atmospheric CO$_2$ and Loss of Sea Ice on Ocean Salinity and Tracer Uptake

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ABSTRACT

Recent observations show a decrease in the extent of Northern Hemisphere sea ice; this decrease has been attributed to human activities. Climate model simulations are presented that examine how loss of sea ice affects the ocean salinity and density structure, and rates of uptake of an idealized transient tracer. The latter results are indicative of how loss of sea ice might affect the ocean’s rate of uptake of anthropogenic carbon from the atmosphere. In simulations in which there is no freshwater forcing due to sea ice forming or melting, the salinity minimum associated with Antarctic Intermediate Water is much weaker than in simulations of the present-day ocean. This suggests that this salinity minimum is maintained in part by a steady supply of freshwater from melting of Antarctic sea ice. In addition, in simulations with no freshwater forcing due to sea ice, vertical salinity and density gradients in the Southern and Arctic Oceans are weaker than in simulations of the present-day ocean. This supports the notion that these gradients are maintained in part by freshwater forcing due to the seasonal cycle of formation and melting of sea ice. As a result, loss of sea ice due to global warming would tend to decrease the stability in parts of the ocean; this opposes the well-known tendency of global warming to increase ocean stability by warming and freshening the upper ocean. Simulations of ocean uptake of an idealized transient tracer in both constant-CO$_2$ and increasing-CO$_2$ environments are performed to investigate the effects of physical changes in ocean and sea ice on transient tracer uptake. In the Southern Ocean, physical changes to the ocean and sea ice are found that result in slower transient tracer accumulation in most locations. When averaged over the entire Southern Ocean, however, these reductions are small, because changes in convective activity due to increased atmospheric CO$_2$ are relatively small, and because transient tracer uptake is relatively insensitive to changes in convective activity. These results suggest that Southern Ocean uptake of anthropogenic CO$_2$ may decrease less than previously supposed as global warming progresses.

1. Introduction

Recent observations (Johannessen et al. 1999) indicate a significant decrease in the extent of Northern Hemisphere sea ice since 1978. Climate model simulations (Vinnikov et al. 1999) suggest that this decrease is due to anthropogenic climate change. If so, the decrease in sea ice extent is likely to continue and perhaps even accelerate. It is therefore important to understand how loss of sea ice might affect the physical structure of the ocean, and the rate at which the ocean removes fossil fuel carbon from the atmosphere (which affects future rates of climate change). For the latter problem, recent attention has focused in particular on the Southern Ocean. Here, Caldeira and Duffy (2000) showed that fluxes of anthropogenic carbon into the ocean are highest (although storage is very low), and Sarmiento et al. (1998) predicted a significant decrease in carbon uptake due to effects of increasing atmospheric CO$_2$. This predicted decrease occurred in part as a result of reduced vertical overturning caused by warming the surface. (Other factors affecting ocean CO$_2$ uptake are changes in biological activity and in the solubility of CO$_2$ in seawater.)

In this paper we identify another effect of global warming on the stratification of the high-latitude oceans, which acts in the opposite direction. Loss of sea ice would tend to reduce ocean stratification because, as mentioned by Pierce et al. (1995) and by Duffy and Caldeira (1997), the seasonal cycle of freezing and melting of sea ice tends to strengthen vertical salinity and density gradients in the present-day ocean. This occurs because ice formation tends to cause sinking of salty water, whereas melting ice freshens the surface ocean. The net effect is to make the surface fresher, and the
Fig. 1. Latitude–depth sections of annual mean salinity, from (a) the Levitus and Boyer (1994) climatology, (b) the Plume2, (c) NIFWF2, (d) NIFWF2 minus Plume2, (e) Plume1, (f) NIFWF1, and (g) NIFWF1 minus Plume1 model simulations. In the bottom panels, a constant (0.1631) has been added to the results to force the mean salinity difference to equal zero; this allows for easier comparison to the NIFWF2 minus Plume2 results, where the mean difference is also nearly zero. All results are averaged in longitude over the world ocean.
In the Southern Hemisphere, salinities within the AAIW salinity minimum increase rapidly. This suggests that this minimum is maintained at least in part by introduction of freshwater from melting of sea ice. This therefore tends to increase density stratification; coupled with more air–sea gas exchange due to reduced sea ice cover, it tends to produce faster uptake of anthropogenic CO$_2$ and other transient tracers.

We present simulations with an ocean–sea ice model and with an independent ocean–atmosphere–sea ice model that examine the importance of sea ice in maintaining vertical salinity and density gradients in the present-day Southern Ocean. We also present an increasing-CO$_2$ simulation performed with the ocean–atmosphere–sea-ice model that investigates how this and other physical changes to the ocean–sea ice system caused by increasing atmospheric CO$_2$ affect rates of uptake and storage by the ocean of an idealized transient tracer.

Several other papers (e.g., Stossel et al. 1998; Goosse et al. 1997; Goosse and Fifechet 1999) have looked at effects of eliminating freshwater forcing due to sea ice in global ocean models. We are reassessing this issue because our models give relatively realistic simulations of the present-day Southern Ocean; as discussed below, this is in part due to these models’ use of instantaneous mixing of salt rejected during formation of sea ice, which mimics subgrid-scale brine-induced convection. These models should therefore allow us to obtain a more accurate assessment of the effects of eliminating salt rejection. As shown below, our models predict very different (and much more dramatic) effects of eliminating freshwater forcing due to sea ice than previous studies have.

Section 2 describes the models used in this study. Section 3 presents our results. In section 4 we give conclusions.

2. Description of models and simulations

In this paper we present simulations performed with two independent models, an ocean–sea ice model and an ocean–atmosphere–sea ice model. The ocean–sea ice model we use is the Lawrence Livermore National Laboratory (LLNL) three-dimensional ocean general circulation model, coupled to a massively parallel version of the Oberhuber (1993) sea ice model. The ocean component of this model is based on the Geophysical Fluid Dynamics Laboratory’s Modular Ocean Model (MOM) version 1.1 (Pacanowski et al. 1991), but has significant scientific and computational enhancements relative to that model. For the simulations presented here each grid cell has a size of 4° of longitude by 2° of latitude. We use a maximum of 23 levels in the vertical. The model bathymetry is essentially realistic, consistent with the model resolution; some smoothing was performed to...
minimize symptoms of numerical problems. Coefficients of vertical mixing of tracers are prescribed. Between the top two model levels (each 25 m thick) the mixing coefficient is 1.0 cm$^2$ s$^{-1}$, to represent the surface mixed layer. Below this, vertical diffusion coefficients vary from 0.2 cm$^2$ s$^{-1}$ near the surface to 1.5 cm$^2$ s$^{-1}$ at the ocean bottom. We use the “Gent–McWilliams” or GM90 eddy parameterization (Gent and McWilliams 1990). Isopycnal and thickness diffusivities are 2.0 $\times$ 10$^7$ and 1.0 $\times$ 10$^7$ cm$^2$ s$^{-1}$, respectively. Following Robitaille and Weaver (1995), we use a latitude-dependent horizontal viscosity that increases with the cosine of latitude from 1 $\times$ 10$^9$ cm$^2$ s$^{-1}$ at the poles to 6 $\times$ 10$^9$ cm$^2$ s$^{-1}$ at the equator. The higher viscosity at the equator is intended to minimize Peclét-type instabilities associated with high vertical velocities (Weaver and Sarachik 1990; Duffy et al. 1997). (Since vertical velocities are calculated via a continuity equation, increasing horizontal viscosities reduces vertical velocities by reducing the divergence of horizontal velocities.) To further minimize these problems, within 30 degrees of the equator, we increase vertical diffusivities and vertical viscosities as needed to ensure that neither of the grid-Peclét stability conditions given by Weaver and Sarachik (1990) are violated.

Surface fluxes of momentum are prescribed, and are obtained by interpolating between monthly mean wind stresses of Hellerman and Rosenstein (1983). Surface fluxes of freshwater are handled as equivalent salt fluxes. Except under sea ice, these fluxes are calculated by restoring to Levitus and Boyer (1994) monthly climatological salinity values, with a restoring time constant of 60 days. Under sea ice, freshwater fluxes are calculated by the sea-ice model based on rates of formation and melting of ice. We use no salinity restoring under sea ice. Surface fluxes of heat are calculated using the approach of Oberhuber (1993). Here, sensible, latent, longwave, and shortwave components of the heat flux are calculated based upon observed atmospheric climatological data and simulated ocean sea surface temperatures. The ice component of our ocean–sea ice model is a version of the Oberhuber (1993) dynamic–thermodynamic sea ice model that has been modified to run on massively parallel computers. We run this model on the same grid as our ocean model. The behavior of this ice model has been documented by Stoessel et al. (1996) and Holland et al. (1993).

We present two simulations with the ocean–sea ice model; these differ only in their treatment of the effects of ice formation and melting upon the ocean salinity (Table 1). In our Plume1 simulation, salt rejected when sea ice forms is spread uniformly between the ocean surface and a maximum depth that is calculated from a prescribed density contrast relative to the surface; the maximum depth of mixing of rejected salt is whatever depth is 0.4 kg m$^{-3}$ denser than the surface, or the ocean bottom, whichever is shallower. This instantaneous mixing of rejected salt into the subsurface ocean simulates the fact that brine-induced convection occurs on horizontal spatial scales that are much smaller than our model grid cells; Duffy and Caldeira (1997) and Caldeira and Duffy (1998) showed that instantaneous mixing of rejected salt into the subsurface ocean dramatically improves simulated salinities and simulated uptake of CFC-11 in the LLNL ocean–sea ice model. In addition, Duffy et al. (1999) showed that in the ocean–atmosphere–sea ice model used here, the brine-mixing scheme improved simulated salinities in the Southern Ocean and elsewhere, improved simulated temperature throughout the deep ocean, and improved simulated sea ice extents and surface air temperatures in the Southern Ocean region. Finally, Caldeira and Duffy (2000) showed that the brine-rejection scheme allows good agreement between our ocean–sea ice model and observation-based estimates of concentrations of anthropogenic CO$_2$.

Despite these successes, it should be kept in mind that the brine-rejection scheme used here is simplified in several important respects. First, in the real ocean, brine-induced convection mixes heat and other tracers as well as salt. In our scheme, only salt is mixed. Second, the “density contrast” of 0.4 kg m$^{-3}$ that determines the maximum depth of mixing of rejected salt is not derived from first principles; rather, this number was arrived at empirically, as a value that produces relatively realistic results in the LLNL ocean–sea ice model and in the University of Victoria (UVic) ocean–atmosphere–sea ice model. Thus, the good results obtained with our brine-rejection scheme are to some extent the result of “tuning.”

In the Plume1 simulation, freshwater released when ice melts is placed entirely in the top model level (because this water has low density and little tendency to sink). In our No Ice Freshwater Forcing 1 (NIFWF1) simulation, we eliminate any effect of freezing or melting of sea ice on ocean salinity; there is no salt rejection when ice forms and no addition of meltwater to the ocean when ice melts. The Plume1 and NIFWF1 simulations were initialized from observed climatological temperatures and salinities (Levitus and Boyer 1994). Initial velocities were zero. As discussed by Bryan (1984), our simulations used longer time steps in the deep ocean, to speed the solutions’ approaches to steady states. Both simulations were run for 1000 simulated surface years, equivalent to 7500 simulated years in the deepest model layer. After this “spinup,” the simulations were run for 25 simulated years with no deep-ocean acceleration, to minimize distortions to the seasonal cycle (Danabasoglu et al. 1996) Results presented below are annual means from the end of this 25-yr period.

A disadvantage of using our ocean–sea ice model to assess the effects of sea ice on the ocean density structure is that this model, like others of its type, uses a “restoring” type surface boundary condition on freshwater (except, as discussed above, under sea ice). This
boundary condition has the unrealistic property that the freshwater flux depends on the surface ocean salinity. In part to avoid this problem, and in part to see if our results can be reproduced with an independent model, we present a second pair of simulations, also examining the effects of freshwater forcing due to sea ice, using the UVic coupled ocean–atmosphere–sea ice model. This model consists of an ocean general circulation model, a dynamic–thermodynamic sea ice model, and a simplified representation of the atmosphere. The ocean component of the UVic model is the MOM2 ocean general circulation model (Pacanowski 1995). This version of the model uses a rotated coordinate system (Eby and Holloway 1994) in which both coordinate singularities occur over land; this eliminates the need to filter the solution at high latitudes. In all simulations with the UVic model, the Gent–McWilliams eddy parameterization (Gent and McWilliams 1990) is used, with thickness and isopycnal diffusivities of $2 \times 10^{-7}$ cm$^2$ s$^{-1}$. No deep-ocean acceleration is used in these simulations. The ice model uses the thermodynamics of Hibler (1979) and dynamics of Hunke and Dukowicz (1997), as detailed by Bitz et al. (1999, manuscript submitted to J. Geophys. Res.).

The atmospheric component of the UVic model is an updated version of that described by Fanning and Weaver (1996). The model atmosphere is a vertically integrated, energy and moisture balance model that diffusively transports heat and moisture on the same computational grid as the ocean model. Precipitation occurs whenever relative humidity exceeds 85%. Terrestrial runoff is routed into the ocean using geographically quasi-realistic “river routing;” runoff occurring in major drainage basins is instantaneously transported to the appropriate coastal ocean grid cell. Latent and sensible ocean–atmosphere heat fluxes are parameterized via bulk formulas. Radiative fluxes are parameterized based on cloud fractions derived from an empirical relationship between fractional cloudiness and relative humidity. No “flux adjustments” or restoring of salinity or temperature are employed.

The latest version of the UVic coupled model includes an interactive thermomechanical ice sheet model on land (based on the University of British Columbia Marshall–Clarke model). The UVic model simulations described here use a simpler treatment, however. Here, the UVic model grows snow to a maximum of 10 m (more than can melt in a year—which means permanent ice as far as the albedo is concerned). Any additional precipitation over 10 m of snow is treated as rain to avoid any long-term climate drift. This rain (which should have been snow) approximates the effects of calving icebergs from the edge of the ice sheet. In the LLNL model, glacial melt runoff is accounted for indirectly, via surface re-storing of salinity to observed values.

The Plume2 and NIFWF2 simulations, both performed with the ocean–atmosphere–sea ice model, are analogous to the Plume1 and NIFWF1 simulations performed with the ocean–sea ice model and described above. That is, in the Plume2 simulation, meltwater is placed in the top model layer, but rejected salt is mixed instantaneously into the subsurface ocean. As in the Plume1 simulation, the maximum depth of mixing is whatever depth has a potential density 0.4 kg m$^{-3}$ greater than the surface waters, or the ocean bottom, whichever is shallower; rejected salt is spread uniformly between the surface and this maximum depth. Our Plume2 simulation is the same as the “Plume” simulation of Duffy et al. (1999). In the NIFWF2 simulation, as in the NIFWF1 simulation, neither formation nor melting of ice has any effect on the salinity of surface water. Except for these differences in the treatment of freshwater forcing due to ice freezing and melting, the Plume2 simulation is identical to the NIFWF2 simulation. As discussed by Duffy et al. (1999), the Plume2 simulation was initialized from the final state of a 4880-yr run with the UVic coupled model, and run for 1000 simulated years. The NIFWF2 simulation was initialized from the final state of the Plume2 simulation, and run for another 1150 simulated years.

The final simulation discussed in this study is a global warming (GW) simulation, in which atmospheric CO$_2$ is prescribed to increase by 5 ppm per yr for 200 yr and then remains constant. This simulation was performed with the UVic ocean–atmosphere–sea ice model. Both in this simulation and in the Plume2 simulation, we simulate the ocean uptake and storage of an idealized transient tracer. This allows us to quantify effects of CO$_2$-induced changes in ocean circulation and sea ice on transient tracer uptake.

3. Results

a. Effects of freshwater forcing due to sea ice on salinity

In the Southern Hemisphere, the salinity minimum associated with Antarctic Intermediate Water (AAIW) is represented well (by the standards of present-day climate models) in both the Plume1 (ocean–sea ice model) and Plume2 (ocean–atmosphere–sea ice model) simulations. As a result of this salinity minimum, a strong salinity contrast exists between the deep Southern Ocean and intermediate depths, both in these simulations and in observed salinities (Fig. 1). As discussed by Duffy and Caldeira (1997) and Duffy et al. (1999), the relatively realistic salinity structure in these models is made possible in large part by the use of the Gent–McWilliams eddy parameterization and by instantaneous mixing of rejected salt into the subsurface ocean.

In the NIFWF1 and NIFWF2 simulations, salinity patterns are quite different from in the Plume1 and Plume2 simulations. The differences are consistent with the net annual mean transport of freshwater equatorward by the growth, transport, and melt of sea ice. Thus, for example, in the Southern Ocean, the net effect of sea
ice is to add salt to the Ross and Weddell Seas, and to add freshwater farther north. Eliminating freshwater forcing due to sea ice increases salinities in much of the upper kilometer or so of the ocean (Fig. 1). On the other hand, the deep Southern Ocean, the near-surface Arctic Ocean, and much of the abyssal ocean, become fresher. This occurs because the lack of salt rejection from Antarctic Sea causes Antarctic Bottom Water...
Fig. 4. Latitude-depth sections of an idealized transient tracer, in the (top) Plume2 simulation, (middle) GW simulation, and (bottom) GW minus Plume2. Results are averaged in longitude over the World Ocean, and are shown 195 yr after the start of transient tracer uptake (and, in the GW simulation, 195 yr after the start of increasing atmospheric CO$_2$ concentrations).
(AABW) to become fresher; the large extent of the area that becomes fresher may be symptomatic of too much of the deep ocean filling up with AABW, as opposed to North Atlantic Deep Water (NADW), in the models. The deep Arctic Ocean becomes much saltier when freshwater forcing due to sea ice is turned off. This is attributable to an increase in salinity in the North Atlantic, which results from loss of freshwater from melting Arctic sea ice.

In the Southern Hemisphere, the intermediate-depth salinity minimum associated with AAIW is much weaker in the NIFWF1 and NIFWF2 simulations than in Plume1 and Plume2 simulations. (Indeed, it is essentially absent in the NIFWF1 simulation.) Thus, both models used here suggest that freshwater forcing associated with sea ice formation and melting are important in maintaining this salinity minimum, in both the models and the real ocean. [Although it should be kept in mind that, as shown by Duffy and Caldeira (1997) and Duffy et al. (1999), the representation of the AAIW salinity minimum in models is sensitive to treatment of salt rejected when sea ice forms.] Two related mechanisms appear to be responsible for this. First, as mentioned by Pierce et al. (1995) and by Duffy and Caldeira (1997), the seasonal cycle of freezing and melting of sea increases vertical salinity gradients in the Southern Ocean. When ice forms, rejected salt is mixed into the subsurface ocean and increases salinities there. When ice melts, freshwater is added to the surface ocean and decreases salinities there. Thus, the net effect is to increase the vertical gradient in salinity. This may explain in part why the vertical salinity gradient in the upper few hundred meters of the Southern Ocean does not diffuse away—it is constantly being renewed by the seasonal cycle of freezing and melting of sea ice.

The second mechanism is related to the fact that ice tends to form near the Antarctic continent, move north, and melt away from the continent. Sea ice therefore adds freshwater to the ocean in regions where isopycnal surfaces associated with AAIW outcrop. This freshwater diffuses along these isopycnal surfaces and creates a salinity minimum that is strongest near the Antarctic continent and weaker farther north. In other words, our results suggest that the salinity minimum associated with AAIW (although not the AAIW itself) exists at least in part because of a constant input of freshwater due to melting of Antarctic sea ice. Figure 2 supports this hypothesis. This shows that in the decades immediately following shutoff of freshwater due to ice freezing/melting, salinities within the AAIW salinity minimum increase rapidly.

As a result primarily of these salinity changes, vertical gradients in density in the Southern Ocean (Fig. 3) are less in the NIFWF2 simulation than in the Plume2 simulation. One might therefore expect more convective activity in the Southern Ocean in the NIFWF2 simulation. In fact, convective activity originating from the surface (as measured by the mean ventilation depth) is slightly greater in the NIFWF2 simulation than the Plume2 simulation (Table 2). However, overall convective activity (including that originating below the surface), as measured by the mean number of model levels involved in convection, is much less in the NIFWF2 simulation. This reduction occurs because in the Plume2 simulation the introduction of rejected salt into the subsurface ocean causes significant grid-scale convection to originate below the surface. Thus, the Southern Ocean may be slightly more susceptible to convection in the NIFWF2 simulation, but less convection actually occurs, because the triggering mechanism of sea ice formation is absent.

### b. Global warming simulation and transient tracer uptake

Our results shown in the previous section support the suggestion of Pierce et al. (1995) and Duffy and Caldeira (1997) that the seasonal cycle of freezing and melting of sea ice plays a significant role in maintaining vertical gradients in salinity and density in the Southern Ocean. This suggests that loss of Antarctic sea ice, such as due to global warming, could tend to decrease the stability of the high-latitude ocean, at least on multicitycles of years. This would oppose the oft-cited ef-

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**Table 2. Rates of convective activity and fractional ice coverage, in simulations with the UVic model, in the Southern Ocean.** All values are averaged vertically over the whole ocean, averaged horizontally over all ocean area south of −50° lat, and are annual means. For the GW simulation, the value shown is 195 yr after the start of increase of atmospheric CO₂ concentrations.

<table>
<thead>
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<th>Simulation name</th>
<th>Model used</th>
<th>What is simulated</th>
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<tr>
<td>Plume2</td>
<td>UVic ocean–atmosphere–sea ice</td>
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<td>NIFWF1</td>
<td>LLNL ocean–sea ice</td>
<td>no freshwater forcing from ice freezing/melting</td>
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<td>NIFWF2</td>
<td>UVic ocean–atmosphere–sea ice</td>
<td>5 ppm/yr atmospheric CO₂ increase (otherwise same as Plume2)</td>
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<td>GW</td>
<td>UVic ocean–atmosphere–sea ice</td>
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Plume2 NIFWF2 GW

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<td>Fractional ice coverage</td>
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</table>
fect that global warming, by warming and freshening the surface ocean, would tend to increase vertical stability. To quantify these and other effects of increasing CO$_2$ on ocean stability and sea ice, we performed an idealized increasing CO$_2$ simulation with our ocean–atmosphere–sea ice model. This GW simulation starts from the steady-state present-day climatology obtained in the Plume2 simulation. The initial atmospheric CO$_2$ concentration of 350 ppm is increased by 5 ppm per yr for 200 yr, and then is held constant at 1350 ppm.

Changes in stability are related to changes in rates of transient tracer uptake, because reduced stability creates greater susceptibility to vertical mixing. Therefore, to quantify how physical changes to the ocean and sea ice brought about by increasing atmospheric CO$_2$ affect the rate of uptake and storage of transient tracers (e.g., anthropogenic CO$_2$) we performed simulations of ocean uptake of an idealized transient tracer, in both the GW and Plume2 simulations. The atmospheric concentration of this tracer is initially zero, and, like the atmospheric CO$_2$ concentration in our GW simulation, increases by 5 units per year for 200 yr. Thus, the atmospheric concentration of this tracer is proportional to the “excess” atmospheric CO$_2$ in the GW simulation. The transient tracer has the same atmospheric concentration history in the Plume2 simulation, even though atmospheric CO$_2$ is held constant in this simulation. Thus, by differing transient tracer concentrations in the GW and Plume2 simulations, we isolate effects of physical changes to ocean and sea ice on transient tracer uptake. While this is informative, it tells only part of the story of how ocean uptake of anthropogenic CO$_2$ will be affected by global warming. In addition to physical changes to the ocean and sea ice, changes in biological transport of carbon and changes in the solubility of carbon in seawater will influence future ocean uptake and storage of anthropogenic CO$_2$; these effects are not examined here.

Entry of the transient tracer into the ocean is parameterized by restoring the transient tracer concentrations in the surface ocean level to prescribed atmospheric values (described above). The restoring time constant was 5 days over open ocean, $10^{20}$ days (effectively infinity) under sea ice. In grid cells with partial sea-ice coverage, the restoring time constant $\tau$ is given by

$$1/\tau = C(1/10^{20}) + (1 - C)(1/5),$$

where $C$ is the fraction of the grid cell covered by ice. This mimics rapid gas exchange over open ocean, no gas exchange in ice-covered regions, and intermediate values in partially covered cells. Thus, reduced ice cover increases ocean uptake of the transient tracer, as would occur in the real ocean.

Increasing atmospheric CO$_2$ concentrations in the GW simulation result in significant changes to the ocean–sea-ice system: the surface ocean warms, ice thickness and extent decrease, and the depth of penetration of NADW is reduced. As discussed below, these changes result in significant changes to the ocean uptake and storage of an idealized transient tracer (Figs. 4, 5, and 6).

Perhaps the most noticeable of these is reduced tracer storage in the North Atlantic (Figs. 4 and 5), which is a result of shallower penetration of NADW. (This occurs due to warming and freshening of surface waters, which both tend to reduce its density.) In the Arctic, tracer column inventories north of Asia are reduced due to increased river runoff and resulting reduced convection. On the other hand, column inventories east of Greenland increase, as a result of reduced ice export out of the Arctic and increased convection in this region.

When the atmospheric CO$_2$ concentration reaches its maximum value of 1350 ppm, convective activity in the Southern Ocean has reduced to about half its original value. (This is true no matter which of several measures of convective activity are used; Table 2.) This reduction in convective activity is much more modest than that found by Hirst (1999), although differences in the CO$_2$ concentration histories between his work and ours make exact comparisons impossible. The relatively modest reduction in our simulations occurs apparently because levels of convective activity in our present-day climate are lower than in other models due to our use of the Gent–McWilliams eddy parameterization and instantaneous vertical mixing of rejected salt. The relatively small reduction in Southern Ocean convection in our simulations seems to be consistent with a relatively small change in Southern Ocean stratification (Fig. 3).

Changes in Southern Ocean transient tracer storage between the Plume2 and GW simulations are even more modest (only a few percent when averaged over ocean areas south of latitude $-$ 50 degrees; Fig. 6). The largest difference occurs in part of the Atlantic sector, where tracer column inventories are around 30% smaller in the GW simulation than in the Plume2 simulation; this is in part a result of slower rates of sea-ice formation, which results in less convection. The smaller column inventories in the Atlantic sector are partly cancelled by larger column inventories in the Ross Sea in the GW simulation than in the Plume2 simulation. This occurs due to slightly increased rates of ice formation in this region. Although these differences result in large percentage differences in tracer concentrations in the deep Southern Ocean (Fig. 4), differences in horizontally averaged tracer column inventories are small compared to the total (Fig. 6), at least for the first roughly 200 yr after atmospheric CO$_2$ concentrations start to increase. This occurs because Southern Ocean column inventories are dominated by near-surface regions, where concentrations are much higher than in the deep ocean.

In summary, increasing atmospheric CO$_2$ causes small changes in Southern Ocean storage of an idealized transient tracer in part because reductions in Southern Ocean convection due to increasing atmospheric CO$_2$ are relatively small (Table 2), and in part because (as Fig. 4 suggests) the main mechanism for transient tracer uptake in the Southern Ocean is isopycnal transport, not
convection. Thus, at least during the first 200 or so yr of increasing atmospheric CO$_2$, storage of this idealized transient tracer is relatively insensitive to reductions in Southern Ocean convection caused by increasing atmospheric CO$_2$.

c. Comparison to previous results

Our results appear to contrast those of Sarmiento et al. (1998), who predicted that changes in ocean circulation and sea ice would reduce global storage of an-
thoropogenic CO$_2$ by around 22% compared to a simulation in which increasing atmospheric CO$_2$ was not allowed to affect ocean circulation or sea ice (their Table 1). This global decrease is independent of changes in solubility and in the “biological pump” of carbon, and was driven primarily by a large reduction in Southern Ocean convection and vertical mixing (due to surface warming and freshening). In our simulations, reductions in Southern Ocean convection are more modest, and have an even more modest effect on transient tracer column inventories. Similarly, Hirst (1999) shows dramatic reductions in Southern Ocean convection as atmospheric CO$_2$ increases. Why the apparent difference? In our simulations, the main tracer transport mechanism in the Southern Ocean is flow along isopycnal surfaces (Fig. 4); convection is less dominant than in the simulations of Sarmiento et al. (1998). This is likely due to two parameterizations that are used in our model but not in that of Sarmiento et al.: the Gent–McWilliams eddy parameterization (Gent and McWilliams 1990) and instantaneous mixing of rejected salt into the subsurface ocean (Duffy and Caldeira 1997). Both these parameterizations significantly reduce convective activity in the Southern Ocean (Duffy et al. 1999). Thus, in our simulations the potential to reduce transient tracer storage by reducing convection is relatively limited.

One further point should be made about the effect of the instantaneous mixing of rejected salt used in our simulations. Since Sarmiento et al. used a (simplified) sea ice model, one might think that their results would include the effect described above wherein the seasonal cycle of formation and melting of sea ice tends to stabilize the upper Southern Ocean. However, a key distinction between other models (including that of Sarmiento et al.) and ours is that in other models all convection, including that induced by brine rejection, operates on the scale of entire grid cells. Thus if ice formation (i.e., brine rejection) triggers significant convection, vertical gradients in salinity, temperature, and density under the sea ice are erased. By contrast, in our model rejected salt is mixed into the subsurface ocean without disturbing the other salt and heat beneath the ice. Thus in our model mixing of rejected salt does not erase vertical gradients in temperature or salinity or density beneath sea ice. Thus our model can (at least crudely) represent the buildup of vertical density gradients due to sea ice better than other models (like that of Sarmiento et al.) that use only grid-scale convection.

Our results for the effects of eliminating freshwater forcing associated with ice disagree with the results of some others who have performed similar simulations. Goosse and Fichefet (1999) found that eliminating freshwater forcing due to sea ice had less effect on ocean salinity than we found. We speculate that this is due to excessive open-ocean convection in the Southern Ocean in their control simulation (a problem they themselves recognize). This convection tends to erase vertical salinity gradients and therefore produces a salinity structure with weak vertical salinity gradients at high latitudes, which tends (perhaps paradoxically) to resemble that with no freshwater forcing due to sea ice. By contrast, our Plume1 and Plume2 simulations have relatively strong vertical salinity gradients at high latitudes; since these gradients are maintained in part by freshwater forcing due to sea ice, a larger change occurs when this forcing is eliminated.

Stoessel et al. (1998) found that eliminating salt–freshwater rejection slightly increased salinities at 4000-m depth (Table 3). This occurs because in their reference simulation, like many others, excessive grid-scale convection tends to tie salinities at 4000 m toward surface salinities, which in turn are tied to (relatively fresh) observed values through a restoring boundary condition. That the convection in their reference simulation is excessive is shown by the salinities in this simulation, which are much too fresh in the deep Southern Ocean (their Fig. 2), and by the fact that their deep ocean is significantly too cold (Table 3). In their sim-

![Image](image-url)
ulation without salt or freshwater rejection, a reduction in grid-scale convection allows salinities at 4000 m to become slightly saltier. By contrast, in our simulations with both the LLNL and UVic models, the “no ice freshwater forcing” simulations are significantly fresher at 4000 m than the respective reference simulations (Plume1 or Plume2). This occurs because in the Plume1 and Plume2 simulations, rejected salt is transported into the subsurface ocean without inducing massive grid-scale convection. This means that in these simulations—-as in the real ocean—the seasonal cycle of freezing and melting of sea ice increases vertical salinity gradients and makes the deep Southern Ocean saltier. This effect is eliminated in the “no ice freshwater forcing” simulations, and the deep ocean thus is fresher in these simulations. Thus, the disagreement between our results and those of Stoeessel et al. is more in simulations of the present-day ocean than in simulations without salt–meltwater rejection. In particular, our simulations of the present-day Southern Ocean have more vertical stratification (in better agreement with observations).

4. Conclusions

Simulations performed with an ocean–sea ice model and with an independent ocean–atmosphere–sea ice model both show that freshwater forcing due to sea ice formation and melting has important effects on the ocean salinity structure. Perhaps most notably, eliminating freshwater forcing due to sea ice eventually largely eliminates the strong salinity minimum associated with AAIW, suggesting that this minimum is maintained in part by an influx of freshwater from melting Antarctic sea ice. This melting occurs predominately away from the Antarctic continent, where isopycnals that connect to AAIW outcrop; freshwater from melting sea ice diffuses along these isopycnals and contributes to the AAIW salinity minimum. More generally, eliminating freshwater forcing due to sea ice results (after hundreds of simulated years) in major reductions in vertical salinity gradients at high latitudes. This supports the notion that the seasonal cycle of freezing and melting of seawater acts to increase vertical salinity gradients. It also suggests that loss of sea ice due to global warming would tend to decrease density stratification high-latitude oceans. This would oppose increases in stratification due to surface warming and freshening, and suggests that decreases in the rate of uptake of transient tracers (such as anthropogenic CO₂) might be relatively modest.

Simulations of an idealized increasing CO₂ scenario show a roughly a factor of 2 reduction in Southern Ocean convective activity at the time atmospheric CO₂ reaches 1350 ppm. Compared to a simulation in which atmospheric CO₂ is held fixed, Southern Ocean storage of an idealized transient tracer over 200 years decreases by only a few percent in both the Southern and global oceans. This modest reduction occurs because the main tracer transport mechanism in the Southern Ocean is isopycnal transport, not convection; thus transient tracer storage is relatively insensitive to a reduction in convective activity.

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