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Sensitivity of simulated salinity in a three-dimensional ocean model to upper ocean transport of salt from sea-ice formation

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Abstract. We show that explicit representation of sinking of salt rejected during sea-ice formation dramatically improves simulated salinity in an ocean general circulation model (OGCM). In our "control" simulation, rejected salt goes into the top model layer, and simulated salinities are typical of OGCMs: the deep ocean is too fresh, and the intermediate-depth salinity minimum associated with Antarctic Intermediate Water is absent. These problems are eliminated in our "test" simulation, in which we distribute rejected salt uniformly over the upper 160 m. Also, the strength of the Antarctic Circumpolar Current is more realistic in this simulation. These results show the need for, but do not provide, a better representation of sinking of rejected salt. The sensitivity of our model to sinking of rejected salt suggests that a similar sensitivity may exist in the real ocean, and that loss of Antarctic sea ice might have major effects.

Introduction

Inability to realistically simulate salinity is a chronic failing of three dimensional ocean general circulation models (OGCMs). Typically, simulated salinities are too low in the deep ocean, and the intermediate-depth salinity minimum associated with Antarctic Intermediate Water (AAIW) is absent or too weak. These problems occur in many simulations (e.g., those of Duffy et al. 1997, Danabasoglu and McWilliams 1995, Toggweiler et al. 1989) using the GFDL model and its relatives, as well as in the LSG model (Maier-Reimer and Hasselmann, 1987), the primitive equation model of Drijfhout et al. (1995), and the OPYC isopycnal model of Oberhuber (1993; Oberhuber, personal communication). They do not occur in the isopycnal model of Hu (1997), or in one simulation described by Large et al. (1997). The reasons why some simulations do not produce realistic salinities while others do are not well understood, but in general involve some combination of unrealistic surface forcing and unrealistic internal transport.

In many OGCM simulations, surface salinities are "restored" to prescribed values, using a term of the form $dS_1/dt = (S^* - S_1)/\tau$ in the evolution equation for salinity. Here, S_1 is the salinity in the top model layer, τ is a prescribed time constant, and S^* are prescribed (usually observed) salinities. Simulated salinities can be improved by increasing S^* values near Antarctica (e.g., *England, 1993*; Robitaille and Weaver, 1995; Hirst and Cai, 1994). However, Toggweiler and Samuels (1993) showed that increasing S^* values can cause unrealistic ocean-atmosphere fluxes of fresh water. Also, this approach cannot be used in coupled ocean-atmosphere models, which do not use "restoring"-type boundary conditions. Although Southern Ocean S^* values should not be *arbitrarily* increased, wintertime Southern Ocean salinities of Levitus and Boyer 1994 may be too fresh (see e.g. Gordon and Huber, 1990), due to a paucity of observations under ice. If so, this could contribute to unrealistically low salinities in our control simulation and in other simulations.

We identify a different way to improve simulated salinities in OGCMs, involving sinking of salt rejected during sea-ice formation. Local salinities can increase significantly during times of sea ice formation, since sea ice is largely free of salt. "Salt rejection" can cause deep convection and bottom-water formation (see e.g. references cited in Killworth, 1979, and Toggweiler and Samuels 1993). This paper deals not with these processes, however, but with sinking of rejected salt to depths of up to 160 m. This mimics transport of salt within the mixed layer. We compare results of a "test" OGCM simulation in which salt released by sea ice formation is forced to sink to depths of up to 160 m to results of a "control" simulation in which this salt is allowed to remain in the top model layer. Simulated salinities throughout the ocean are much more realistic in the test simulation. Also, the strength of the simulated Antarctic Circumpolar Current is increased, and agrees better with observations. Our findings support the conclusion of England (1991) that modeling sea ice is critical in obtaining accurate simulated salinities.

Sinking of salty water created by sea-ice formation should occur automatically in OGCMs via convection. However, in OGCMs convection occurs only when the mean density in a grid cell (of order one hundred km on a side) exceeds the mean density in the cell beneath it. By contrast, mixing triggered by ice formation probably occurs on much smaller horizontal scales (Denbo and Skyllingstad, 1996, and references therein). Small-scale vertical mixing can occur in regions where the ocean is vertically stable on scales the size of a GCM grid cell. Thus, sinking of salt rejected during sea ice formation is not adequately represented by traditional OGCM convective adjustment algorithms.

This paper is a sensitivity study, because our representation of sinking of ice-related salt is oversimplified in several respects. First, the depth to which we sink that salt is chosen arbitrarily. Second, sinking of ice-related salt in the real ocean (but not in our model) results in mixing of heat and other tracers. Thus, we show the need for, but do not provide, a realistic parameterization of sinking of ice-related salt.

Models including sinking of ice-rejected salt have been used before, by Oberhuber (1993) and by Stocker and Wright (1996).

Model Description

Our runs use the LLNL ocean general circulation model; this is based on the GFDL Modular Ocean Model (MOM) version

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1.1, but has several important additions: (1) the dynamic/thermodynamic sea ice model of Oberhuber (1993); (2) a free surface approach to solving for barotropic velocities; (3) the numerical scheme of Webb (1995), which minimizes errors in vertical advection of momentum; (4) the Gent-McWilliams parameterization (Gent and McWilliams, 1990) of transport of tracers by subgrid scale eddies; (5) a new term in the vertical diffusion equation which partially cancels numerical diffusion due to vertical advection (Yin and Fung, 1991). To minimize symptoms of numerical problems, we slightly modified the model topography to eliminate isolated regions where unphysical tracer values tend to occur. Also, we increased (relative to the values described below) both horizontal and vertical diffusivities in any grid cell where the temperature on the time step in question was less than -2.1 deg. C (indicating numerical problems). Vertical and horizontal diffusivities were increased by 0.5 cm^2/s and 2.0 $\times 10^7$ cm^2/s, respectively.

Our model is run here on a mesh of 2° (latitude) by 4° (longitude) with 23 vertical levels. The grid is the same as that of Duffy et al. (1997), and has vertical layer thicknessed ranging from 25 m near the surface to 450 m in the deep ocean. As in many GFDL-based models, there is an artificial island at the North Pole. Coefficients of vertical diffusivity are prescribed. Between the top two model layers, the vertical diffusivity is 1.0 cm^2/s. Below that, vertical diffusivities vary according to $A_v = 0.8 + (1.2786/\pi) * \tan^{-1} [4.5 \times 10^{-3}(z - 2.5 \times 10^{-3})]$ cm^2/s (Bryan and Lewis, 1979). This gives vertical diffusivities ranging from 0.2 cm^2/s in the upper ocean to 1.5 cm^2/s in the deep ocean. Isopycnal and thickness diffusivities are 2.0×10^7 cm^2/s and 1.0 $\times 10^7$ cm^2/s; respectively. The vertical viscosity is 20 cm^2/s. Following Robitaille and Weaver (1995), we use a latitude-dependent horizontal viscosity which is higher near the equator to minimize Peclet-type instabilities in that region (Weaver and Sarachick, 1991; Duffy et al., 1997). Our horizontal viscosity is $1.0 \times 10^9 \times [1 + 6 \cos(\varphi)]$ cm^2/s, where φ is latitude. To further suppress Peclet-type instabilities, we increase vertical mixing as needed to prevent the grid-Peclet conditions given by Weaver and Sarachick (1991) from being violated within 30 degrees of the equator.

Our model is forced at the surface with wind stresses of Hellerman and Rosenstein (1983), linearly interpolated in time between monthly means. Averaging and interpolating the wind stresses reduces the maximum values applied to the model, and may unrealistically reduce motions, especially of sea ice. Surface fluxes of heat are calculated based on climatological atmospheric data and calculated SSTs, as in Oberhuber (1993). Surface salinities are restored to monthly mean observed values (Levitus and Boyer, 1994), with a time constant of 60 days. In addition to restoring salinities, salinities are modified in response to melting and freezing of sea ice, as discussed below. No subsurface restoring of any tracer is performed.

We present two simulations which are identical except for their treatment of salt rejected during sea-ice formation. We assume sea ice contains no salt. (Since sea ice does contain some salt, we may overestimate slightly the effects of salt rejection.) In the control run, salt freed by formation of sea ice is put into the top model layer. Thus, if a depth of water Δh_{ice} is frozen into sea ice, the salinity of the surface layer is adjusted according to $S_1^{new} = S_1^{old}\, \Delta h_1/(\Delta h_1 - \Delta h_{ice})$ In the "test" simulation, salt released by ice formation is put uniformly into the 5 top model layers, down to a depth of 160 meters. That is, the salinity S_k in layers 1 through $k_{max} = 5$ is modified according to

$$S_k^{new} = S_k^{old} + S_1 \times \frac{\Delta h_{ice}}{\sum_{k'=1}^{k_{max}} \Delta h_{k'}} \quad (1)$$

Here, Δh_k is the thickness of layer k, and S_1 is the original salinity in layer one.

When ice melts, all the fresh water goes into the surface layer, in both runs. (This fresh water does not sink because it is "lighter" than sea water.) Thus, when ice melts, the salinity S_1 in layer 1 is reduced according to $S_1^{new} = S_1^{old}\left(\Delta h_1 - \Delta h_{ice}\right)/\Delta h_1$ where Δh_{ice} is a thickness of water created by melting ice and is assumed to be positive. These equations show that, in the control run, formation and subsequent melting of sea ice would leave salinities unchanged absent other perturbing influences.

The runs presented here use longer time steps in the deeper model layers to accelerate the convergence of the model solution towards equilibrium (Bryan 1984). Each run was integrated for 1500 simulated surface years, equal to 11,250 simulated years in the deepest model layer. To minimize distortions to the seasonal cycle, both runs were next integrated for 25 years without deep-ocean acceleration (Danabasoglu et al., 1996). Results presented here are annual means from the last 5 years of this period.

Results

Simulated salinities in our control run are unrealistic. The deep ocean is much too fresh, and the intermediate-depth salinity minimum associated with Antarctic Intermediate Water (AAIW) is almost entirely missing (Figures 1, 2 and 3). This is typical of results obtained with OGCMs. The simulated deep ocean is too fresh in part because fresh, cold water is transported downward from the surface ocean off Antarctica and fills the Southern Ocean with unrealistically fresh water. In addition, North Atlantic Deep Water (NADW) does not penetrate deeply enough; thus, the deep ocean fills too much with relatively fresh AABW and not enough with relatively salty NADW.

Simulated salinities are much more realistic in our "test" simulation. The problem of the deep ocean being too fresh is completely gone; in fact, the deep ocean is slightly too salty (Figure 3). This is by design, in that we chose the maximum depth (160 m) of sinking of ice-related salt to achieve this result. The structure of the intermediate-depth salinity minimum, which is almost completely missing from our control run, is basically correct in our test simulation (Figures 1 and 2).

How does sinking salt released by sea ice formation improve simulated salinities? In our control run and typical OGCM simulations, which are initialized from observed temperatures and salinities, the deep ocean becomes too fresh because the strong vertical gradient in observed salinities in near-surface waters off Antarctica is gradually eliminated. In our "test" run, the seasonal cycle of melting and freezing of sea ice maintains this salinity gradient and keeps the deep Southern Ocean salty. As ice forms, salt is added uniformly to the upper 160 m of the ocean; when ice melts, the surface layer is freshened. The net effect is to reduce the salinity of the top model layer and increase salinities in the four layers below that; this increases the near-surface vertical salinity gradient. Another important factor is that our test run has less convective activity in the Southern Ocean than our control run does. Measured in terms of the rate of loss of potential energy, convective activity in the test run is about 4 times weaker on average south of latitude -50 deg. than in the control run.; this helps maintain the near-surface vertical salinity gradient in the

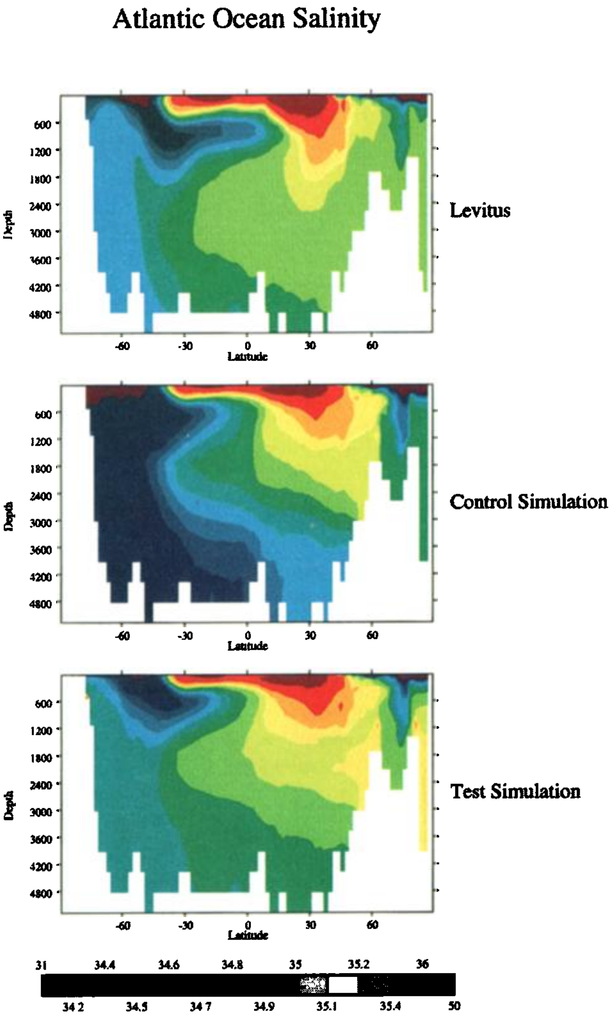

Figure 1. Latitude depth sections of annual mean salinity in the Atlantic ocean, (a) as observed by Levitus and Boyer (1994), (b) in our control simulation, (c) in our test simulation. Results are averaged in longitude over the Atlantic basin.

test run. Surface fluxes of freshwater into the ocean from the restoring boundary condition are slightly *higher* in the Southern Ocean in the test run. Thus, the improved simulated salinities in this run are due to improved transport of salt within the simulated ocean, not to changed surface fluxes.

Our conclusions may depend on the seasonal cycle of sea ice being about right in our model. Southern Hemisphere sea ice areas are reported to oscillate during the year between about 2 and 15×10^6 km^2 (Gloersen et al., 1992) and between about 4 and 20 $\times 10^6$ km^2 (Zwally et al., 1983); our simulated range is about 2 to 19×10^6 km^2. In the Northern Hemisphere, observed ice areas fluctuate between about 6 and 14×10^6 km^2 (Gloersen et al., 1992), while those in our model range from about 4 to 14×10^6 km^2. Relatively few measurements of ice thicknesses have been published. Our mean ice thicknesses range seasonally from about 1.6 m to 3.7 m in the Northern Hemisphere, compared to "generally accepted average values of 2.5 to 3 m" (Mellor and Hakkinen, 1994) and measured mean thicknesses (over limited areas) of about 2.5 to 4.5 m (McLaren et al., 1994). In the Antarctic, drilling surveys (which may not yield typical thicknesses) show thicknesses of 0.6 to 1.0 m for first year ice, and 1.0 to 2.5 m for multi-year ice (Wadhams, 1994); our model has a seasonal range of about 1.25 m to 2.0 m for mean Southern Hemisphere thicknesses. Ice thicknesses and distributions are nearly identical in the test and control runs. The maximum thickness is less than 6 m. We conclude that the improved salinities in our test run are not an artifact of poor simulated sea ice.

The sinking of salt in our test simulation makes the simulated Antarctic Circumpolar Current (ACC) stronger, in better agreement with observations. In our test run, the ACC through the Drake Passage is 108 Sv., versus 84 Sv. in our control run and 118-146 Sv. inferred from observations (Whitworth, 1983).

Our results are not sensitive to the details of the surface boundary condition on salinity under sea ice. Runs identical to those presented here only without surface restoring of salt under ice produce results very similar to those presented here. Unlike when surface fresh-water fluxes are specified, when a restoring boundary condition is used, the global mean simulated salinity may be incorrect. It is much too low in our control run, but about right in our test run. The vertical gradients in salinity in the upper Southern Ocean, which are much more realistic in our test run than in our control run, should not dependent strongly on the type of surface boundary condition used.

Conclusions

We find that simulated salinities in our OGCM are highly sensitive to transport in the upper 160 m of salt rejected during sea-ice formation. If rejected salt is sunk to depths of 0 to 160 m, simulated salinities are much more realistic than those in a control simulation in which rejected salt is left in the top model layer (which is 25 m thick). Sinking of rejected salt increases the strength of our simulated Antarctic Circumpolar Current (through the Drake Passage) from about 84 Sv in the control run

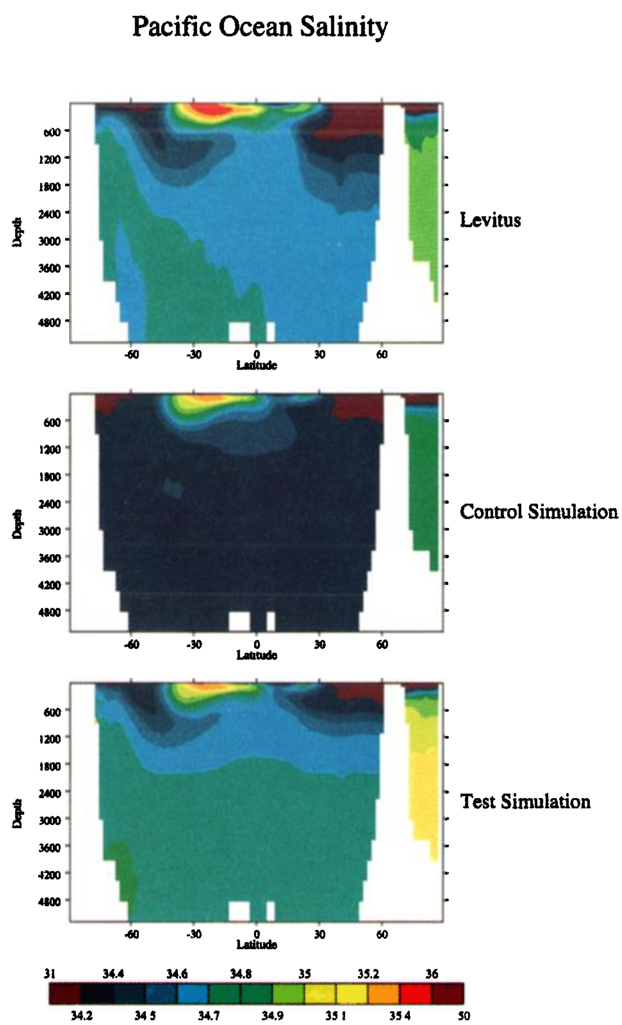

Figure 2. Same as Figure 1, except for the Indo-Pacific basin.

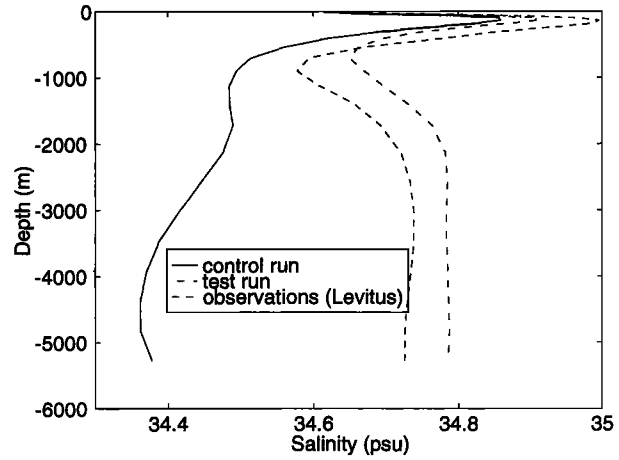

Figure 3. Vertical profiles of global mean salinity in our control run, 'test' run, and observed (Levitus and Boyer 1994).

to about 108 Sv; thus, agreement with observation-based estimates observations (118 - 146 Sv) is much improved.

Although sinking of ice-rejected salt should in principle be represented by the convective adjustment parameterization in OGCMs, the horizontal spatial scale of salt-related convection in the real ocean is likely much smaller than the horizontal dimension of a typical OGCM grid cell; thus standard OGCM convection parameterizations do not adequately represent this process.

Although the concept of salt-sinking is physically motivated, the formulation used here (distributing the salt evenly over the upper 160 meters) is not; therefore, the runs presented here are sensitivity studies which demonstrate the importance of transport of ice-rejected salt in the upper ocean. A better representation of this process is needed. This might involve a parameterization of sinking of ice-related salt and/or improvements to the ice model, in particular the use of daily wind forcing (to more realistically represent ice motions), or more realistic ice dynamics which better represent the opening and closing of leads.

Finally, our results suggest that mixing of ice-related salt in the upper 100-200 m of the ocean is also important in determining properties (especially salinity) of the real ocean. This suggests that loss of Antarctic sea ice (such as through global warming) could significantly affect global ocean properties.

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