Sensitivity of the Present-Day Climate to Freshwater Forcing Associated with Antarctic Sea Ice Loss

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ABSTRACT

The role played by Southern Hemisphere sea ice in the global climate system is explored using an earth system climate model of intermediate complexity. An ensemble of experiments is analyzed in which freshwater forcing equivalent to a complete 100-yr meltback of Southern Hemisphere sea ice is applied to a model run that simulates the present climate. This freshwater forcing acts to mildy subdue Southern Ocean deep overturning, reducing mean Antarctic Bottom Water (AABW) export by 0.5 Sv (1 Sv $\equiv 10^6$ m^3 s^{-1}) in the ensemble average. The decreased convective overturning cools the surface waters, thereby increasing sea ice volume and thus forming a negative feedback that stabilizes Antarctic sea ice. In contrast, the reduced convective overturn warms subsurface waters in the Southern Ocean, which, combined with the imposed freshening, results in a reduction in the meridional steric height gradient and hence a slowdown of the Antarctic Circumpolar Current (ACC). The reduction in ACC strength is, however, only modest at 1.5 Sv. These responses are thus of only weak magnitude, and the system recovers to its original state over time scales of decades. An extreme scenario experiment with essentially instantaneous addition of this meltwater load shows similar results, indicating the limited response of the climate system to the freshening implied by Antarctic sea ice melt. An additional experiment in which a much larger freshwater forcing of approximately 0.4 Sv is applied over 100 yr confirms the relatively weak response of the model's climate state to such forcing, relative to the well-documented climatic effects of freshwater forcing added to the North Atlantic.

1. Introduction

In this study we investigate the role played by Antarctic sea ice in global climate using a coupled climate model of intermediate complexity. Sea ice is an important component of the earth's climate system, affecting both the ocean and the atmosphere by its presence and through its formation. The albedo of sea ice is substantially higher than that of the ocean, so that sea ice growth in the open ocean increases the reflection of incoming solar radiation back to space. The presence of a layer of sea ice also profoundly modifies the fluxes of heat, salt, gases, and momentum between the ocean and atmosphere. Heat and salt fluxes to the upper ocean are further modulated by the growth (melting) of sea ice, through latent heat release (absorption) and brine rejection (freshwater release). In particular, the brine rejection that accompanies sea ice growth drives the formation of Antarctic Bottom Water (AABW), a key component of the global thermohaline circulation. Finally, sea ice shields the air–sea exchange of gases, including carbon dioxide, so it fundamentally controls the rate of CO_2 uptake by the world's oceans.

The effects upon climate of freshwater forcing due to ice melt have been discussed in a number of previous studies. A decay in the AABW and Antarctic Intermediate Water (AAIW) production rates due to melt-

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ing of Antarctic land ice has been suggested as a possible trigger for the Bolling–Allerod warming (Weaver et al. 2003). Saenko et al. (2003) describe such changes in the state of the thermohaline circulation as a function of the relative density differences between AAIW and North Atlantic Deep Water (NADW), showing that freshwater forcing off the southern tip of South America can control the extent of NADW ventilation. Ivchenko et al. (2004) find that positive salinity anomalies in the Southern Ocean can generate equatorial temperature anomalies within a number of months in an ocean model. Richardson et al. (2005) investigated the decadal-scale response of the climate system to Southern Ocean freshwater forcing. They found that surface waters cool due to inhibited ventilation caused by the increased stratification, in turn cooling the atmosphere in the Southern Hemisphere. This atmospheric signal propagates northward, and within a time scale of years can influence the North Atlantic Oscillation. Bitz et al. (2006) found that a retreat of Antarctic sea ice produced deep warming of the Southern Ocean in the Community Climate System Model, version 3 (CCSM3), due predominantly to weakened convection resulting from surface freshening. Stouffer et al. (2007) investigated the effects upon global climate of an idealized land ice melt applied to either hemisphere. The Atlantic thermohaline circulation was found to be relatively insensitive to freshening in the Southern Ocean, in contrast to the well-documented weakening that follows North Atlantic freshening. By contrast, the atmospheric response to freshening in each hemisphere was found to be similar, with cooling occurring in the hemisphere of surface freshening.

Recent observations suggest significant changes have occurred in deep Southern Ocean water masses over the past decade. Broecker et al. (1999) hypothesized that the production rate of AABW slowed considerably during the twentieth century. Bates et al. (2005) studied changes in Southern Hemisphere ocean circulation and climate for various global warming scenarios, finding an increased Antarctic Circumpolar Current (ACC) transport, and a century-scale decrease in AABW production, with eventual recovery over multicentury time scales. Goosse and Renssen (2001) suggest that under global warming the large heat capacity of the Southern Ocean will delay the loss of Antarctic sea ice, but result in a greater long-term loss relative to the Northern Hemisphere. Unlike previous studies, we concentrate exclusively on the role of Antarctic sea ice in the present-day climate system. In particular, we focus on the effect of the relatively weak freshwater forcing that would result from a complete Antarctic sea ice loss, but isolated from the effects of global warming and CO_2 increases.

To summarize, the effect of sea ice melt on the present climate state over century time scales is investigated using an intermediate-complexity earth system model. A set of experiments was performed involving addition of a freshwater flux equivalent to a complete meltback of Antarctic sea ice. We also study simulations that involve both this meltwater flux and an actual retreat in Antarctic sea ice extent. Various methods are tested to remove the sea ice directly. As the freshwater forcing due to Antarctic sea ice melt is relatively weak in comparison to that due to the melting of continental ice, we also perform an additional experiment with a strong freshwater forcing in order to determine to what degree the results are relevant to previous studies, especially the suite of papers exploring NADW response to freshening. The results of the experiments are discussed in section 3, while the model and experimental design are outlined in section 2. A summary and conclusions appear in section 4.

2. The coupled model

The model used in this study is version 2.7 of the University of Victoria (UVic) Earth System Climate Model (Weaver et al. 2001), with a spherical grid resolution of 1.8° (meridional) by 3.6° (zonal). The model is of intermediate complexity, with a dynamic atmospheric energy-moisture-balance model (EMBM), a thermodynamic/dynamic sea ice model, and a three-dimensional ocean general circulation model, all coupled through heat fluxes and the exchange of water and momentum at the air–sea and sea ice interfaces. The single-layer atmospheric EMBM transports heat through diffusion and moisture through advection and diffusion. The atmospheric wind fields are prescribed from the National Centers for Environmental Prediction–National Center for Atmospheric Research (NCEP–NCAR) reanalysis (Kalnay et al. 1996). Wind stress is applied over both the ocean and sea ice, with sensible heat and evaporative fluxes being controlled by wind speed. The sea ice model combines dynamics, multilevel thermodynamics, and a rotated coordinate system that allows a comprehensive evaluation of processes occurring at high latitudes. The 19-level ocean model is the Geophysical Fluid Dynamics Laboratory (GFDL) Modular Ocean Model version 2 (Pacanowski et al. 1995) based on the Navier–Stokes equations and the Boussinesq and hydrostatic approximations, with depth levels ranging from 25 m at the surface to 500 m at the bottom. In the ocean the Gent and McWilliams

(1990) mixing scheme has been included to allow for a parameterization of the effects of eddy-induced mixing and to reduce Southern Ocean circulation errors caused by the model's coarse resolution. The isopycnal and isopycnal thickness diffusion coefficients are both set to 1.0×10^7 cm^2 s^{-1}.

Such a model has a number of advantages and disadvantages over those employing a full three-dimensional dynamical atmosphere (e.g., Claussen et al. 2002). In our study the use of an intermediate-complexity model allows multiple multicentury simulations to be performed, permitting an ensemble analysis of climate sensitivity over century time scales. The same experiments would have to be 1–2 orders of magnitude shorter if a comprehensive coupled GCM were used. Thus, in this study it is possible to extend the results of Richardson et al. (2005) who concentrate on the decadal response to high-latitude freshwater forcing.

The experimental design consists of a control case (CTRL) and several perturbation experiments that incorporate differing Southern Hemisphere ice melt scenarios. The CTRL experiment represents an equilibrium solution under the preindustrialized CO_2 level of 280 ppm and no additional forcing. The CTRL case is spun up from an ocean state with initial T-S set to observations and equilibrated over 4500 yr, by which time a stable climate consistent with present-day conditions has been established. The first ice melt scenario ($MELT_{FW}$) consisted of adding a freshwater flux to all Southern Hemisphere sea-ice-covered grid boxes at a set rate of approximately 3 cm yr^{-1}, corresponding to a freshwater flux of approximately 0.004 Sv (1 Sv $\equiv$ 10^6 m^3 s^{-1}). The total amount of freshwater added during the simulation at each grid cell is limited by the mean minimum ice volume at that cell in the control state, so that the net freshwater anomaly added corresponds to that which would occur under a complete Antarctic sea ice meltback. As the melt rate is constant and the sea ice thickness is spatially variable, some regions at the ice margin are subjected to relatively short-lived freshwater anomalies. In addition, the area-integrated freshwater forcing declines over the 100-yr period as the implied ice margin contracts poleward. All freshwater forcing ceases after 100 yr. This run isolates the effect of the freshwater flux accompanying a complete melt-off of Southern Hemisphere sea ice, with none of the additional effects of a sea ice retreat, such as albedo changes and increased air–sea heat and freshwater fluxes.

In addition, we study three alternate scenarios of $MELT_{FW}$: 1) the freshwater anomaly is applied over the course of a single year over the same domain; 2) the freshwater anomaly is applied rapidly as in 1), but completely at the continental margin; and 3) the freshwater forcing is increased to 3 m of ice melt per year, or approximately 0.4 Sv. In 1), by simulating the extreme scenario of a sudden complete sea ice meltback, we test the sensitivity of the results to the time scale of melting. In 2) we further test the sensitivity to the location of the applied freshwater flux, as it is likely that an Antarctic sea ice meltback would produce positive freshwater fluxes predominantly at the continental margin. Finally, as the freshwater flux signified by Antarctic sea ice melt is relatively small compared to other sources of anomalous freshwater, such as precipitation and land ice melt, the significantly larger freshwater forcing applied in 3) allows us to gauge the relevance of our results to previous studies, especially those focused over the North Atlantic.

A further set of experiments ($MELT_{ICE}$) was designed to simulate the effect of actual sea ice removal. A partial meltback was first attempted by adding an additional surface heat flux to the surface of all Antarctic sea ice of 10 W m^{-2}. This has the primary effect of raising the temperature of all Antarctic sea ice. In the absence of feedbacks, this forcing would be approximately twice that needed to melt the annual mean Antarctic sea ice volume during the year. Alternate approaches investigated included artificially suppressing sea ice formation and direct sea ice thinning. As will be discussed below, each of these strategies was only partially successful in reducing the Southern Hemisphere sea ice volume due to a self-stabilizing feedback loop in the ice component of the coupled climate system.

Each perturbation simulation was run for 400 yr using as initial conditions an equilibrated state from the CTRL run. As mentioned above, in most $MELT_{FW}$ cases the freshwater forcing ceased after the initial 100 yr, while in two experiments the freshwater anomaly is added abruptly within the first model year. In $MELT_{ICE}$, the amplitude of additional downward radiation was increased linearly over the initial 100 yr, and then kept steady for the following 300 yr. To remove any sensitivity to initial conditions, an ensemble of five simulations was performed for each perturbation experiment, initialized from distinct states of the CTRL run separated in time by 100 yr. The results that follow represent averages over each ensemble set.

3. Results

a. CTRL

The UVic model has previously been shown to simulate the key components of the earth's climate system reasonably well (Weaver et al. 2001). Version 2.7 of the UVic model used here improves on earlier versions in a

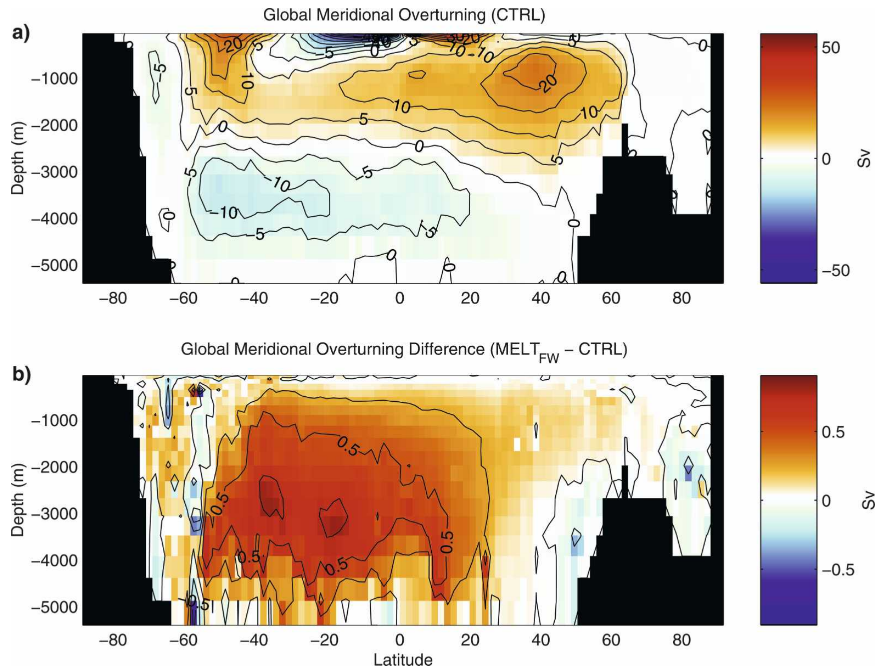

FIG. 1. Zonal mean meridional overturning (Sv) in (a) the CTRL experiment and (b) the difference between the $MELT_{FW}$ experiment and CTRL, averaged over years 50 to 60, corresponding to the time of maximum decline in the AABW cell. Contour interval in (a) is 5 Sv between −10 and 10 Sv, and 10 Sv thereafter. In (b) the contour interval is 0.5 Sv.

number of aspects. An improved simulation of meridional moisture transport has meant that lower values of vertical viscosity could be used without compromising the accurate simulation of the ocean salinity field. The reduced vertical viscosity allows a substantial increase in climate variability over time scales of decades and beyond. In addition, the tendency of earlier versions of the model to produce unrealistically high surface salinities in the southern Ross Sea due to an overly active sea ice formation–brine rejection process has been corrected without the need for flux corrections through small improvements to the model geometry.

Figure 1a shows the global meridional overturning circulation (MOC) for the CTRL run. We estimate the intensity of the AABW (NADW) cell by calculating the minimum (maximum) value of the MOC south of 60°S (between 20° and 55°N). Production of the major water masses is close to estimates based on observations, with the AABW and NADW cells transporting approximately 6 and 22 Sv, respectively. Compared to observations, this is slightly stronger for NADW and weaker for AABW. The AABW production rate we quote is the commonly diagnosed MOC, used often by modelers to benchmark simulations against the observed estimate of 8.0–9.5 Sv (Orsi et al. 1999). However, as this MOC quantity represents a zonal integral, the transport value may underestimate the actual net downwelling rate of AABW adjacent to Antarctica (England 1992) as water also upwells at these latitudes adjacent to regions of downwelling. The simulated Antarctic sea ice field compares reasonably well with observations (Figs. 2a,b), although in general the model sea ice extends farther north than observed. The modeled ACC transports 109 Sv through the Drake Passage. Overall the CTRL case captures the present-day climate state of the Antarctic and Southern Ocean to a reasonable degree of accuracy.

b. $MELT_{FW}$

We first discuss the $MELT_{FW}$ case wherein the equivalent Antarctic minimum sea ice volume is completely melted over 100 yr.

1) Years 0–100

The addition of freshwater to Southern Ocean sea ice locations during the initial 100 yr decreases sea surface salinity (SSS) in the sea ice covered regions by up to 0.1 psu (Fig. 3b). The decreased SSS, accompanied by weak changes in sea surface temperature (SST) (Fig. 3a), increases the buoyancy of the surface waters (Fig. 3c), hence discouraging sinking and thereby reducing the convective overturning of surface water into the ocean interior. With relatively warm water at depth

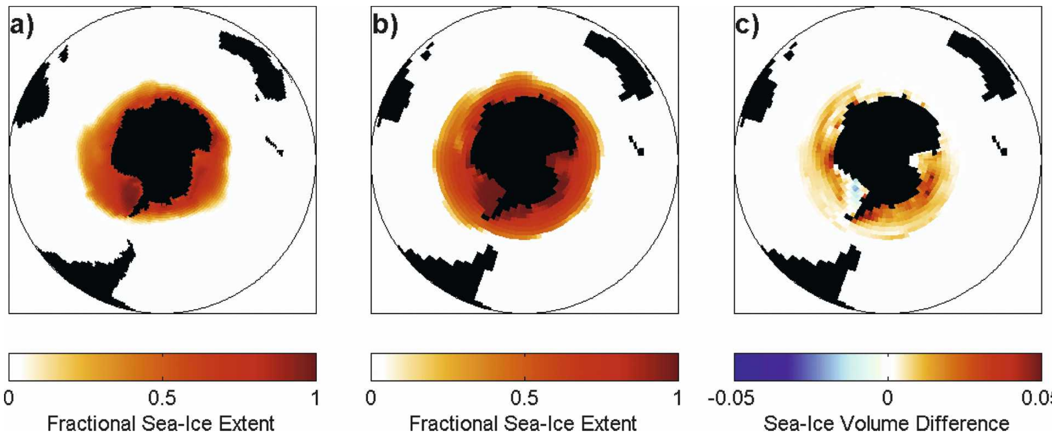

FIG. 2. Mean fractional sea ice coverage (a) from observations [Hadley Centre Sea Ice and Sea Surface Temperature dataset (HadISST); Rayner et al. 2003] and (b) as simulated in the CTRL run. (c) The mean difference in fractional sea ice volume between CTRL and $MELT_{FW}$, expressed as $MELT_{FW}$ − CTRL, shown over the initial 25 yr of the perturbation phase. Fractional sea ice volume was determined as the product of fractional ice coverage and ice thickness and is shown in units of meters.

this reduced convective overturning within the Southern Ocean results in warm anomalies at depths greater than 200 m (Fig. 4a) and cooler water at the surface (Fig. 3b), especially at the sea ice margin, thereby allowing for greater sea ice formation (Fig. 2c). Such interactions between the ocean and Antarctic sea ice are discussed in Martinson (1990). This negative feedback mechanism is also seen to occur in coupled climate models (Richardson et al. 2005). In the $MELT_{FW}$ experiment Southern Ocean net sea ice volume is seen to increase by 2% almost immediately after application of the freshwater perturbation, declining gradually toward the unperturbed value over the initial century (Fig. 5c), in keeping with the reduction in freshwater forcing at the same time. While SST in the sea ice zone does not decrease greatly, being already close to freezing and prevented from significant heat loss to the atmosphere by the insulating properties of the sea ice, SST at the sea ice margin experiences a cooling of up to 0.5°C due to the suppressed oceanic convection and mixing (Fig. 3a). Over time the surface cooling trend propagates northward through advection of the anomalously cool water from the sea ice margin. The integrated heat content of the water column at high latitudes increases, however, as the ocean's ability to lose heat to the atmosphere is diminished by the increased stratification and sea ice coverage (Fig. 2a). A similar deep warming of the Southern Ocean due to reduced surface convection has also been reported by Bitz et al. (2006).

In terms of water mass production, the surface freshening described above reduces the annual average production of AABW by up to 0.5 Sv in year 50 (Figs. 1b and 5a). During the same period a small increase in NADW production of the order of 0.1 Sv develops, leading to a slight warming and salinity increase in the northern North Atlantic of up to 0.1°C and 0.05 psu, respectively. This response in the North Atlantic is unlikely to be due to the mechanism of Richardson et al. (2005), that is, propagation via the atmosphere, due to the inability of the atmospheric component of the model to support wave propagation. It is instead likely to be due to a decade-scale response to density changes in AAIW, similar to that discussed by Saenko et al. (2003). The additional meltwater spreads northward at the surface and into the zone of AAIW production, so that both Antarctic Shelf Water and AAIW become anomalously fresh (Fig. 4b). AAIW in particular also becomes anomalously cool, but still experiences a drop in density (Fig. 4c). As noted above, surface cooling in the region of AAIW formation (Fig. 3a) is not consistent with suppressed vertical convective mixing, but rather is a result of northward advection of anomalously cool waters from the sea ice margin.

Increased sea ice extent and lower SST at the sea ice margin result in reduced atmospheric temperatures, in particular above and south of the sea ice edge, with anomalies of up to 0.3°C (figure not shown). The cool atmospheric temperature anomalies propagate northward, predominantly in the Pacific, while the slight northern North Atlantic SST warming results in positive atmospheric temperature anomalies of up to 0.1°C.

The T-S and density changes across the Southern Ocean (Fig. 4) conspire to decrease the meridional steric height gradient and thus the strength of the ACC. Transport through the Drake Passage drops on average

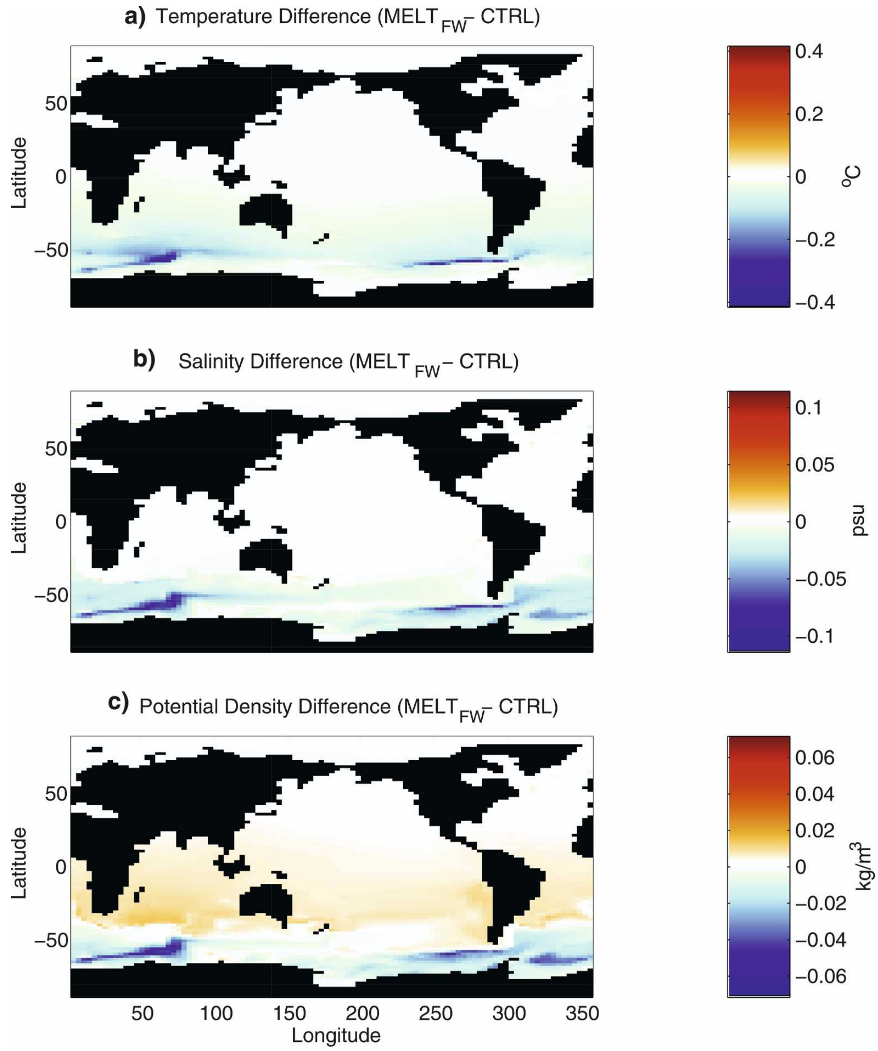

FIG. 3. Difference in (a) temperature, (b) salinity, and (c) potential density at the surface between the $MELT_{FW}$ and CTRL runs averaged over the initial 25 yr of the perturbation phase. Fields are shown as $MELT_{FW}$ minus CTRL.

by approximately 1.5 Sv from an unperturbed mean value of 109 Sv (Fig. 5b). This is consistent with Bates et al. (2005) who found that, while the net effect of increased greenhouse forcing was to increase the strength of the ACC, the freshwater forcing component acted to weakly reduce ACC transport. Changes in the path of the ACC are too small to account for the general surface cooling seen in the Southern Ocean (Fig. 3a), suggesting that anomalous heating/cooling patterns are the result of air–sea and ice–sea heat flux anomalies and the advection of anomalous SST (as opposed to anomalous advection of mean temperatures). The Brazil and East Australian Currents also experience a small decline in transport, while the other major current systems remain little affected.

2) Years 100–400

The remainder of the simulation represents a gradual return to initial conditions (Fig. 5). After the freshwater forcing ceases at year 100, the negative salinity anomalies propagate northward into the west wind drift and are carried to the equator in the Peru–Chile Current, while positive salinity anomalies of up to 0.1 psu develop at low latitudes south of Africa (figure not shown). By year 200 the surface waters of the South Pacific have freshened by approximately 0.05 psu and the positive salinity anomalies in the North Pacific and south of Africa are within the range of natural variability in the model. At the same time a positive sea surface salinity anomaly of up to 0.1 psu develops in the North

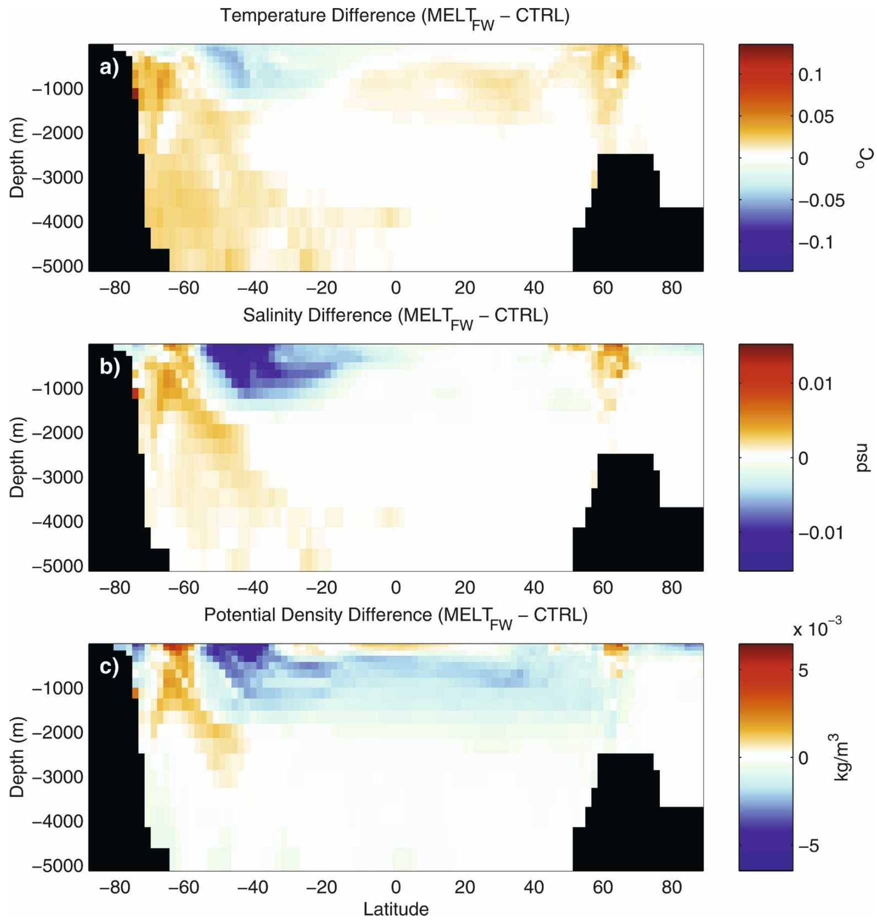

FIG. 4. Zonal mean difference between MELT$_{FW}$ and CTRL in (a) ocean temperature, (b) salinity, and (c) potential density, averaged over years 50 to 75.

Pacific south of Alaska. By year 400, the temperature and salinity of all waters have returned close to the initial unperturbed values.

c. *Additional experiments*

1) Abrupt sea ice melt

To test the sensitivity of our results to the time scale of the sea ice melt, an extreme case was run wherein the total freshwater forcing was imposed within the initial year of the simulation. In this case significant changes occur during the initial decade, including a reduction in the AABW cell by 3 Sv and a reduction in the ACC of 2.5 Sv. Interior T-S changes are up to 0.02°C and 0.03 psu in both AAIW and AABW (figures not shown). However, the system quickly recovers to its unpeturbed state within around 50 yr. In summary, imposition of an abrupt freshwater pulse equivalent to a complete meltback of Antarctic sea ice has no long-term consequences for the global thermohaline circulation in the model.

2) Continental margin sea ice melt

As Antarctic sea ice tends to be advected equatorward, its melt may be expected to concentrate the anomalous freshwater flux near the continental margin. In an experiment in which MELT$_{FW}$ is concentrated at the southernmost grid boxes, however, we find largely similar results to those described above. In this case, the anomalous freshwater spreads quickly over the surface of the sea ice zone, creating essentially the same conditions seen in the standard MELT$_{FW}$ case. We note that in the real system the exact location of freshwater forcing may have a greater influence than what is seen in the model, as simulated AABW formation is much less localized than observed. However, ancillary experiments with even larger FW perturbations still fail to permanently collapse the Southern Ocean thermohaline circulation.

3) Continental ice melt

The freshwater flux added in the previous experiments, corresponding to a complete meltback of all

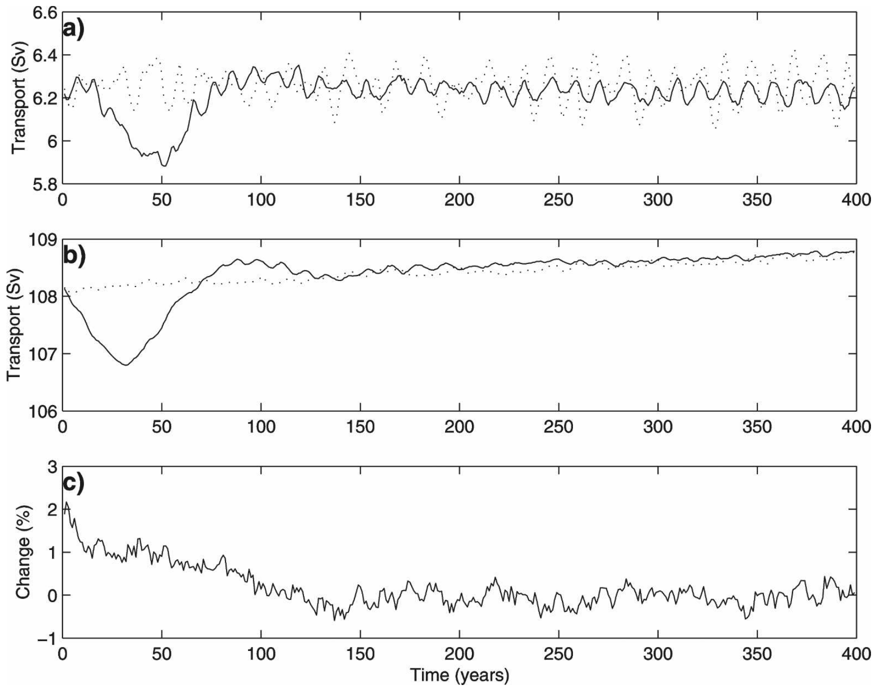

FIG. 5. (a) Meridional overturning transport in the AABW cell in CTRL (dashed) and $MELT_{FW}$ (solid); (b) transport in the ACC in CTRL (dashed) and $MELT_{FW}$ (solid); and (c) percent increase in net Antarctic sea ice volume in $MELT_{FW}$ relative to CTRL.

Antarctic sea ice, is substantially smaller than that associated with land ice melt and commonly applied in studies of climate sensitivity. To investigate to what degree the results of the earlier experiments may be relevant to large-scale continental ice melt, the $MELT_{FW}$ experiment was repeated but with a substantially increased freshwater forcing of 0.4 Sv per year over a century. This freshwater flux is similar in magnitude to those applied to the North Atlantic and/or Antarctic in previous studies (e.g., Rahmstorf and Willebrand 1995; Weaver et al. 2003; Stouffer et al. 2007). We do not provide a thorough analysis of this experiment, but rather concentrate on the major differences between this case and the standard $MELT_{FW}$ experiment. Figure 6 shows the response of AABW, the ACC, Antarctic sea ice volume, and NADW to this larger land ice anomaly.

Although quantitative differences exist between this experiment and $MELT_{FW}$, the two experiments are broadly consistent qualitatively. As in the standard $MELT_{FW}$ case, during application of the stronger anomalous freshwater forcing, AABW production and ACC transport decrease, while NADW production and Antarctic sea ice volume increase. The mechanisms responsible for these changes were found to be the same as those discussed above for $MELT_{FW}$. Qualitative differences to the standard $MELT_{FW}$ did occur following cessation of the stronger freshwater forcing, as the system does not return monotonically to the original state. Between years 100 and 200 NADW production is reduced by close to 10 Sv, while AABW production increases rapidly to 15 Sv before declining equally rapidly to 4 Sv. However, after 400 yr both the AABW and NADW cells continue to approach their original intensities and are within 2 Sv of their values in CTRL.

4) $MELT_{ICE}$

In a series of additional experiments an attempt was made to directly remove Southern Ocean sea ice while perturbing the rest of the climate system as little as possible. As may be expected from the negative sea ice feedback mechanism described above, this proved a difficult task. One approach was to add an artificial heat flux of 10 W m^{-2} directly at the surface of all Antarctic sea ice, resulting primarily in increased sea ice temperatures. However, as the temperature of the atmosphere and ocean remain initially unchanged, the conditions for sea ice growth remained, so that the artificially removed sea ice was rapidly replaced, and potentially at greater thickness due to the negative feedback associated with meltwater forcing. Thus, the heat added to the sea ice is rapidly lost to the atmosphere and ocean, so that little sustained sea ice loss occurs. Significant meltback only occurs once atmospheric and oceanic

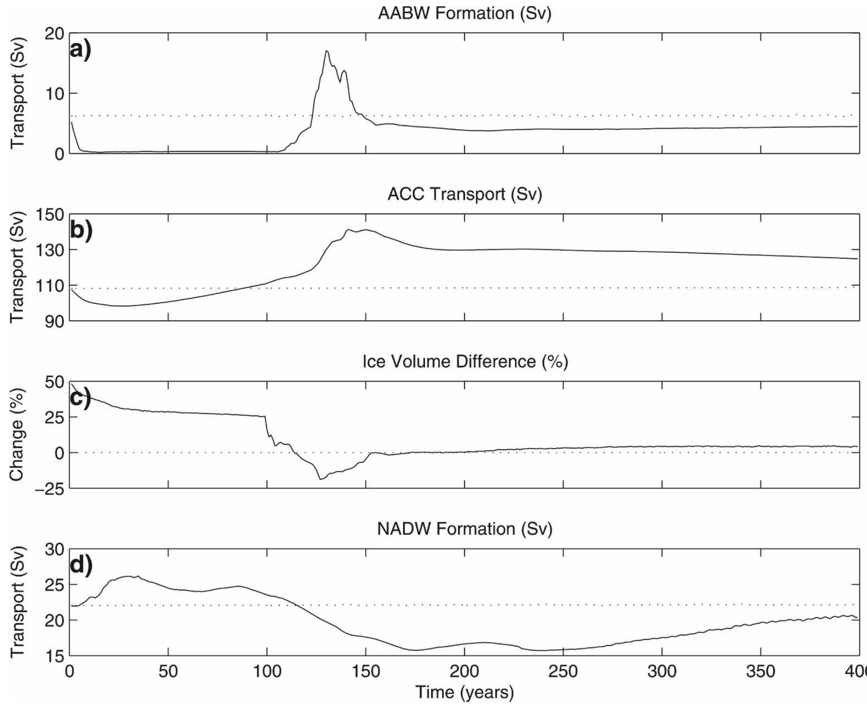

FIG. 6. Time series from the land ice melt simulation (solid) compared to CTRL (dashed). (a) Meridional overturning transport in the AABW cell; (b) transport in the ACC; (c) percent increase in net Antarctic sea ice volume; and (d) meridional overturning transport in the NADW cell.

temperatures have warmed sufficiently to alter the heat balance. After 400 yr of this heat flux forcing, sea ice volume had only decreased by about 5% relative to the control experiment, while mean ocean surface and atmospheric temperatures had increased by up to 1°C. Attempts to inhibit the formation of sea ice by simply suppressing its formation in the model had the side effect of producing surface ocean temperatures well below freezing, due to unabated air–sea heat exchange. Previous studies that induced sea ice retreat without creating an imbalance in surface fluxes by instead decreasing the sea ice albedo (Hansen and Nazarenko 2004; Bitz et al. 2006) also suggest a resilience of Antarctic sea ice to melting. In the case of Bitz et al. (2006) a reduction in sea ice albedo of approximately 8% generated a loss of close to 50% in Arctic sea ice volume, but only less than 10% decrease in Antarctic sea ice volume.

4. Summary and conclusions

The set of simulations discussed above suggest that Antarctic sea ice may be relatively stable in the present-day climate system. The freshwater forcing due to sea ice melt suppresses the convective overturn that would otherwise draw heat up from depth to the surface, and thus acts as a negative feedback to limit further sea ice melt. Figure 7 illustrates this process. This is consistent with the fact that no significant trends in Antarctic sea ice extent have been observed despite a clear signal of global warming (e.g., Cavalieri et al. 1997; Parkinson et al. 1999; Cavalieri et al. 2003). We note that in the real system a number of other factors not included in the model would complicate this basic mechanism. For example, under enhanced global radiative forcing (not included in the simulations presented here) this process would reinforce the mechanism discussed by Bitz and Roe (2004), whereby ice displays greater resilience to meltback by radiative forcing as it thins. The fact that in all experiments considered here the simulations recovered to the initial unperturbed climate state suggests that Antarctic sea ice is relatively difficult to perturb to a new state under present-day thermal forcing. This is true even for individual ensemble runs for which AABW collapse was substantially greater, and for the case of an artificially large freshwater forcing. This is consistent with earlier studies that suggest that the mid- to high-latitude Southern Hemisphere climate is largely driven by external factors (e.g., Goosse and Renssen 2001; Bates et al. 2005; Arzel et al. 2006), and that

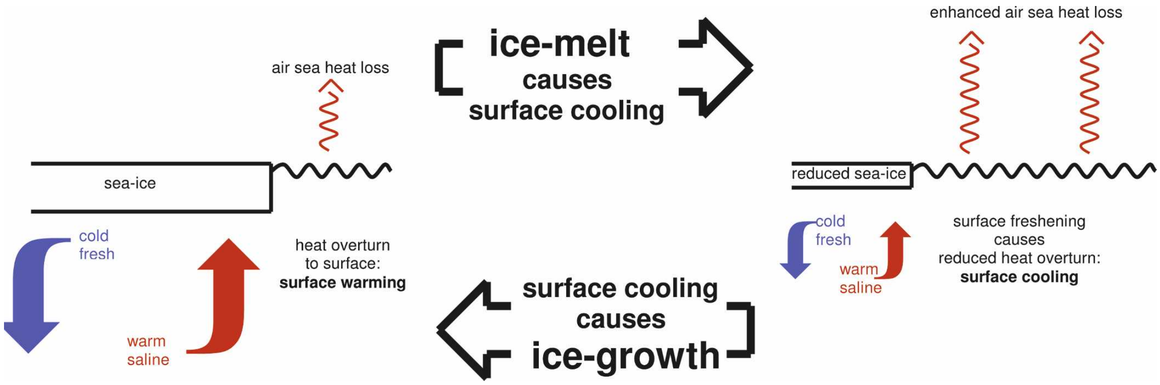

Fig. 7. Schematic showing the negative feedback mechanism for maintaining stable Antarctic sea ice. Freshwater from sea ice meltback inhibits convective overturn, and thereby cools surface waters and encourages renewed sea ice growth.

Southern Ocean sea ice loss is likely to occur only as a second stage of hemispheric-wide warming. The freshwater pulse resulting from a complete meltback of Antarctic sea ice caused only relatively small changes in the simulated climate, including a temporary 0.5-Sv reduction in AABW production and a 1.5-Sv slowing of the ACC. A simulation of the larger-scale freshwater forcing that would accompany a continental ice melt produced more dramatic temporary changes to the overturning circulation, during and immediately following application of the forcing, but without major long-term changes.

Our finding that Antarctic sea ice is difficult to melt back, and that the long-term global climate response to even a large anomalous freshwater flux is relatively small, is in contrast to numerous studies that suggest a strong sensitivity to meltwater addition in the North Atlantic (e.g., Manabe and Stouffer 1993; Rahmstorf and Willebrand 1995). A number of reasons exist that may explain the discrepancy between the Antarctic and North Atlantic studies. For example, there are significant differences in the sea ice formation conditions, with sea ice formation being associated with katabatic winds off Antarctica, but with more moderate conditions in the North Atlantic. Sea ice also forms at lower latitudes in the North Atlantic compared to the Antarctic, with a resulting change in the importance of downward shortwave radiation (e.g., Bitz and Roe 2004). Also, a fundamental difference exists between the thermohaline circulation in the Antarctic and North Atlantic regions: while AABW is formed due to its very low temperature, NADW forms due to its very high salinity. In addition, due to their differing geometries, freshwater anomalies reside substantially longer in the North Atlantic than in the Southern Ocean, allowing AABW production to recover much more rapidly from applied freshwater forcing than NADW production (see also Stouffer et al. 2007). The stability of the present climate to Southern Ocean freshwater forcing may also be a result of the fact that the North Atlantic overturning cell is currently in an "on" state, so that a reduction in AABW production and freshening of AAIW serve to reinforce rather than destabilize the system. It may also be that the modest present-day AABW formation rates, along with the Drake Passage effect (Toggweiler and Samuels 1995; Sijp and England 2004), result in the AABW cell not affecting the global distribution of poleward heat transport in any profound way, unlike NADW, which has a marked impact on heat transfer into the Northern Hemisphere.

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