Resilience of the Present Climate to Antarctic Sea-Ice Loss

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ABSTRACT

The role played by Southern Hemisphere sea-ice in global climate is explored using an Earth system climate model of intermediate complexity. An ensemble of experiments is analysed where freshwater forcing equivalent to a complete 100 year meltback of Southern Hemisphere sea-ice is applied to a model run that simulates the present climate. This freshwater forcing acts to mildly subdue Southern Ocean deep overturning, reducing mean AABW export by 0.5 Sv in the ensemble average. The decreased vertical overturning cools the surface waters thereby increasing sea-ice volume, and thus forming a negative feedback that stabilises Antarctic sea-ice. In contrast, the reduced vertical overturn warms sub-surface waters in the Southern Ocean which, combined with the imposed freshening, results in a reduction in the meridional steric height gradient and hence a slow-down 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. Experiments are also examined where sea-ice is removed either by suppressing its formation or by artificially adding heat to the sea-ice. In both cases sea-ice exhibits a resilience to being removed, due to the fact that the atmospheric and oceanic conditions remain favourable for ice formation at high southern latitudes. The case in which ice formation is suppressed is more effective at removing sea-ice, but allows the surface waters to cool to below freezing, and hence to become unphysical. In the case when an artificial heat flux is applied, sea-ice loss only occurs once sufficient heat is added to the ice to heat the atmosphere or ocean as well, hence reducing the rate of sea-ice formation. Once the artificial heat flux is removed, the mean climate state recovers to its original condition
within a century.
1 Introduction

In the following we investigate the role played by Antarctic sea-ice in global climate through the use of 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 thermocline 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 melting of Antarctic ice has been suggested as a possible trigger for the Bollard-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 northwards, and within a time-scale of years can influence the North Atlantic Oscillation.

Recent observations suggest significant changes have occurred in deep Southern Ocean water masses over the past decade. Broecker et al. (1999) suggested that the production rate of AABW slowed considerably during the 20th Century. Bates et al. (2005) study changes in Southern Hemispheric 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 multi-century 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 isolated from the effects of global warming and CO$_2$ increases.

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 were 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. The results of these experiments are discussed in Section 3, while the model and experimental design are outlined in Section 2. A conclusion appears in Section 4.
2 The coupled model

The model used in this study is Version 2.7 of the University of Victoria 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 3-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 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, multi-level thermodynamics, and a rotated coordinate system which 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 parametrisation 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 $1.0 \times 10^7$ cm$^2$/sec.

Such a model has a number of advantages and disadvantages over those employing a full 3-dimensional dynamical atmosphere (e.g. Claussen et al., 2001). In our study the use of an intermediate complexity model allows multiple multi-century 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 Richards et al. (2005) who concentrate on the decadal response to high latitude freshwater forcing.

The experimental design consisted 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 pre-industrialised CO$_2$ level of 280 ppm and no additional forcing. 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 / year. The total amount of freshwater added during the simulation at each grid cell is limited by the mean maximum 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 years. 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.

We also study a more extreme scenario of MELT$_{FW}$, wherein this freshwater anomaly is applied essentially instantaneously over the same domain.

A second set of experiments (MELT$_{ICE}$) were designed to simulate the effect of actual sea-ice removal. A partial melt-back was first attempted by adding an additional heat flux to 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. Alternate approaches
investigated included artificially suppressing sea-ice formation and direct sea-ice thinning. As will be discussed below, each of these strategies were only partially successful in reducing the Southern Hemisphere sea-ice volume due to a self-stabilising feedback loop in the ice component of the coupled climate system.

Each simulation was run for 400 years. As mentioned above, in MELTFW the freshwater forcing ceased after the initial 100 years. In MELTICE, the amplitude of additional downward radiation was increased linearly over the initial 100 years, and then kept steady for the following 300 years. To remove any sensitivity to initial conditions, an ensemble of five simulations was performed for each experiment, initialised from distinct states of the CTRL run separated in time by 100 years. The results that follow represent averages over each ensemble set.

3 Results

3.1 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 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 timescales 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 in this model version.
Figure 1a shows the global meridional overturning circulation (MOC) for the CTRL run. Production of the major water masses is close to estimates based on observations, with the AABW and NADW cells transporting approximately 6 Sv and 22 Sv respectively. This is slightly stronger for NADW and weaker for AABW, compared to observations. The AABW production rate we quote is the commonly diagnosed meridional overturning circulation (MOC), used often by modellers to benchmark a simulation against the observed estimate of 8 - 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 et al., 2006) as water also upwells at these latitudes adjacent to regions of downwelling. The simulated Antarctic sea-ice field compares reasonably well with observations (Fig. 2a,b), although in general the model sea-ice extends further north than observed. The modelled Antarctic Circumpolar Current (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.

3.2 MELT$_{FW}$

3.2.1 Years 0 - 100

The addition of freshwater to Southern Ocean sea-ice locations during the initial 100 years 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 suppressing the overturning of surface water into the ocean interior. With relatively warm water at depth this reduced vertical 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). This negative feedback mechanism is also seen to occur in coupled climate models (Richardson et al. 2005). In the MELTFW experiment Southern Ocean sea-ice volume is actually seen to increase by 2% almost immediately after application of the freshwater perturbation, declining gradually toward the unperturbed value over the initial century, in-keeping with the reduction in freshwater forcing at the same time (Fig 5c). 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 $0.5^\circ$ 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 (Figure 2a).

As above, the surface freshening suppresses overturning, reducing the annual average production of AABW by up to 0.5 Sv in year 50 (Fig 1b, Fig 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 $0.1^\circ$ 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), i.e. 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 Antarctic Intermediate Water (AAIW), similar to that discussed by Saenko et al. (2003). The additional meltwater spreads northwards at the surface and into the zone of AAIW production, so that both Antarctic Shelf Water and AAIW become anomalously fresh (Figure 4b). AAIW in particular also
becomes anomalously cool, but still experiences a drop in density (Figure 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 (Fig. not shown). The cool atmospheric temperature anomalies propagate northwards, 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 Antarctic Circumpolar Current (ACC). Transport through Drake Passage drops on average by approximately 1.5 Sv from an unperturbed mean value of 109 Sv (Figure 5b). 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.

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

3.2.3 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 where 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 (Figs not shown). However, the system quickly recovers to its unperturbed state within around 50 years. 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.

3.3 MELT\textsubscript{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\textsuperscript{2} directly to 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 melt-back only occurs once atmospheric and oceanic temperatures have warmed sufficiently to alter the heat balance. After 400 years 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. Further attempts were made to inhibit the formation of sea-ice by simply suppressing its formation in the model. Such experiments, however, have the side-effect of producing surface ocean temperatures well below freezing, due to unabated air-sea heat exchange. Unphysical SSTs then develop due to the fact that continued heat loss to the atmosphere occurs without the insulating ice layer forming.

4 Summary and Conclusions

The set of simulations discussed above suggest that Antarctic sea-ice is relatively stable in the present-day climate system. 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). The effect of meltwater suppressing vertical ocean overturn, that would otherwise draw heat up from depth to the surface, acts as a negative feedback to limit further sea-ice melt (refer to Fig. 6 for details). 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 melt-back 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. 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), and that 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 melt-back of Antarctic sea-ice causes 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.

Our finding that Antarctic sea-ice is difficult to melt-back, and that the global climate response to its melt-back is 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), yet agrees with other Antarctic based studies. A number of reasons exist that may explain the discrepancy between the Antarctic and the 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). In addition, 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. 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 in an "on" state, so that a reduction in AABW production and freshening of AAIW serve to reinforce rather than destabilise 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 that has a marked impact on heat transfer into the Northern Hemisphere.

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Figure 1: Zonal mean meridional overturning (Sv, 1 Sv = 10^6 m^3 sec^{-1}) 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 Sv and 10 Sv, and 10 Sv thereafter. In b) the contour interval is 0.5 Sv.
Figure 2: Mean fractional sea-ice coverage (a) from observations (HadISST, Rayner et al. 2003) and (b) as simulated in the CTRL run. The mean difference in sea-ice volume between CTRL and MELT_{FW}, expressed as MELT_{FW} -CTRL, is shown over the initial 25 years of the perturbation phase in (c).
Figure 3: Difference in a) temperature, b) salinity, and c) potential density at the surface between the MELT<sub>FW</sub> and CTRL runs averaged over the initial 25 years of the perturbation phase. Fields are shown as MELT<sub>FW</sub> minus CTRL.
Figure 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.
Figure 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.
Figure 6: Schematic showing the negative feedback mechanism for maintaining stable Antarctic sea-ice. Freshwater from sea-ice meltback inhibits heat overturn, and thereby cools surface waters and encourages further ice growth.