Mechanisms controlling the late twentieth century cooling and freshening of Antarctic Intermediate Water

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Abstract

The late twentieth century cooling and freshening of Antarctic Intermediate Water (AAIW) is examined in a coupled climate model of intermediate complexity. Processes that may have contributed to changes in AAIW properties include increased sea-ice melt, increased precipitation, higher heat fluxes into the ocean, and a poleward shift in the subpolar westerly winds. Each of these processes has been previously linked to anthropogenic climate change, although their impact on AAIW is yet to be examined in a coupled climate model. This thesis applies each forcing mechanism separately to determine their relative impact on AAIW temperature and salinity. To begin with, each perturbation forcing is applied over a 100 year time scale to examine the potential change on ocean circulation and specifically AAIW over a uniform time period. Then, each forcing is applied at a rate that more realistically captures the likely change in these processes in the latter half of the twentieth century. The model response is compared to observations of AAIW that show a cooling and freshening of 0.3°C and 0.03 psu over a period of 20-30 years. Analysis of the results is conducted on both geopotential (or “z”) coordinates and isopycnals to distinguish between changes due to “heaving” of isopycnals and genuine changes in water-mass properties. It is found that all four processes have the potential to cool and freshen AAIW, though most likely a combination of increased precipitation, higher heat fluxes and a poleward wind shift has enabled the water-mass to cool and freshen as observed.
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Chapter 1

Introduction

Antarctic Intermediate Water (AAIW) is a key part of the ocean’s thermohaline circulation, transporting heat, salt and carbon around the globe and regulating the earth’s climate. Mode and intermediate waters are a good indicator of climate change as they are influenced by air-sea fluxes and ventilated by surface waters over relatively rapid time scales. Observations of temperature and salinity profiles from the Pacific and Indian Oceans reveal an average maximum cooling and freshening of 0.3°C and 0.03 psu (practical salinity units) of AAIW in the late twentieth century (Bindoff and Church (1992), Johnson and Orsi (1997), Wong et al. (1999), Bindoff and McDougall (2000), Wong et al. (2001), Aoki et al. (2005)). Identifying the mechanisms that caused these changes in AAIW properties is critical to an understanding of climate change and its impact on the oceans. The aim of this study is to determine what physical mechanisms might have caused the late twentieth cooling and freshening of AAIW using a coupled climate model.

AAIW is one of the most prominent features of the Southern Ocean, found north of
the Subantarctic Front (SAF) and transported eastward with the Antarctic Circumpolar Current (ACC), extending equatorward into the Atlantic, Indian and Pacific basins. It is characterised by a fresh salinity tongue that extends north at approximately 800-1000 m depth. Its typical salinity and temperature range is 34.2-34.4 psu and 3-5°C (Piola and Georgi, 1982). The $27.2 \text{ kg m}^{-3}$ isopycnal ($\sigma_{27.2}$) more or less runs through the core of AAIW and outcrops near the maximum sea-ice margin in winter (Santoso and England, 2004).

There remains some debate over the formation mechanism of AAIW. Early studies (Wüst (1935), Deacon (1937), Sverdrup et al. (1942)) suggested that AAIW was formed by circumpolar subduction or sinking of surface waters along the Antarctic Polar Front Zone (APFZ). This idea has now been replaced by the view that formation occurs in specific regions in the Southern Ocean. McCartney (1977) proposed that formation of AAIW was linked to Subantarctic Mode Water (SAMW), which forms in a deep convective mixed layer in the southeast Pacific Ocean off southern Chile and the southwest Atlantic. In contrast Molinelli (1981) suggested that AAIW is formed by isopycnal exchange across the polar front with significant inputs from the southeast Pacific. There are observations to support either one or a combination of the two theories (see for example, Georgi (1979), Piola and Georgi (1982)).

Models of the ocean have also provided insight into the AAIW formation debate. England et al. (1993) produced a realistic representation of AAIW in a coarse resolution general circulation model. The renewal process involved convective overturning and along-isopycnal mixing of fresh surface waters in the southeast Pacific off southern
Chile, largely consistent with the view of McCartney (1977). An inverse box model, using observed temperature and salinity data to determine baroclinic or relative velocity, used by Sloyan and Rintoul (2001) revealed that diapycnal transport driven by air-sea fluxes and interior mixing were important for the circulation, renewal and modification of AAIW. While there is a general consensus that AAIW is formed in specific regions of the Southern Ocean, there is still some debate over the physical mechanisms involved.

Repeat hydrographic sections allow for comparison between historical data and more recent measurements. A number of studies comparing temperature and salinity properties of AAIW have been conducted, revealing significant cooling and freshening on density surfaces over a period of approximately 20-30 years. Bindoff and Church (1992) were the first to document a change in AAIW. They found a cooling of approximately 0.05-0.2°C and freshening of 0.01-0.04 psu on density surfaces at intermediate depths over 22 years in the Tasman Sea at 43°S and 28°S (estimated from their Fig. 3). It is interesting to note that a warming was found at intermediate depths on isobars, demonstrating the importance of using density surfaces to analyse temporal changes. Johnson and Orsi (1997) undertook a similar study on the only other South Pacific datasets available at the time. They found a freshening of the salinity minimum of approximately 0.03 psu and cooling of 0.1°C along density surfaces over 22 years along the meridian 170°W between 60°S and the equator. Wong et al. (1999) found a freshening of 0.021 psu and 0.064 psu at 17°S in the Pacific and 32°S in the Indian Oceans respectively. Their study focused on the freshening of AAIW, although a cooling on density surfaces can also be seen. Bindoff and McDougall (2000) found a cooling and freshening
of the salinity minimum at 32°S in the Indian basin of 0.33°C and 0.06 psu on density surfaces. Wong et al. (2001) used the zonal sections 17°S and 10°N and the 150°W meridian in the Pacific Ocean to show a freshening on density surfaces of 0.02 psu over a period of approximately 20 years. The only study specific to the Southern Ocean (between 30°E and 150°E; Aoki et al. (2005)) also showed a decrease of 0.05 psu and 0.6°C on density surfaces over 30 years. The reader is referred to the attached Literature review for a more detailed description of twentieth century observations of AAIW.

There have been several suggested mechanisms for the changes in AAIW detected in the late twentieth century; each one related to anthropogenic climate change via the effects of increased atmospheric greenhouse gases. For instance Wong et al. (1999) suggested their measured AAIW changes could be explained by an increase of freshwater into the Southern Ocean of about 31 mm yr\(^{-1}\). A warming on isobars accompanied by a cooling on isopycnals led Bindoff and McDougall (1992) to suggest a “pure warming” forcing mechanism; a somewhat counter-intuitive process described by Bindoff and McDougall (1994) whereby a warming of surface waters appears as a cooling on isopycnals when the waters subduct into the thermocline. The third mechanism proposed is due to changes in the subpolar westerly winds, causing Ekman transport changes, as examined by Oke and England (2004) in a climate model. The Ekman transport changes in their study produced a cooling and freshening of intermediate waters at depth. Banks and Bindoff (2003) imposed changes in greenhouse gases, tropospheric and stratospheric ozone, and sulphate aerosols in a coupled climate model to suggest surface warming to be the dominant factor in AAIW changes.
The physical mechanisms causing changes in AAIW temperature and salinity is to be investigated here. These include changes in freshwater fluxes such as sea-ice melt and increased precipitation, changes in surface heat fluxes (and in turn sea surface temperatures (SST)) and shifts in the subpolar westerly winds. The rest of this thesis is divided as follows: in chapter 2 the coupled climate model and experimental design is described. In chapter 3 results for the century scale experiments are presented. Late twentieth century results are shown in chapter 4. Finally, a discussion and conclusions are given in chapter 5.
Chapter 2

Methodology

2.1 The Climate Model

The model used to run the simulations is the University of Victoria Earth System Climate model (hereafter the UVic model) of intermediate complexity version 2.7 (see Weaver et al. (2001) for a more detailed description). The basic components of the model include an ocean general circulation model, a sea-ice model, and a simple 1-layer atmospheric model, all coupled via heat, freshwater and momentum fluxes. All components of the model have a resolution of 3.6° (zonal) by 1.8° (meridional) and a global coverage. A brief overview of the model will be given here.

The ocean component of the model is based on the Geophysical Fluid Dynamics Laboratory (GFDL) Modular Ocean Model (MOM) 2.2 (Pacanowski, 1995). MOM is a z-coordinate, primitive equation model based on the Naiver-Stokes equations subject to the Boussinesq and hydrostatic approximations. It has 19 vertical levels with increased resolution near the surface for improved representation of upper ocean pro-
cesses. Brine parameterisation is also incorporated during sea-ice formation, which improves the representation of intermediate waters (Duffy et al., 1999). The surface is driven by both buoyancy forcing and a prescribed present-day wind stress field. All fluxes are seasonally-varying. Vertical mixing is achieved using diffusivity that increases with depth from 0.6 cm²s⁻¹ at the surface to 1.6 cm²s⁻¹ at the bottom. The model also includes isopycnal mixing of Redi (1982) and Gent and McWilliams (1990) to approximate the effect of eddies on the mean flow. Horizontal diffusivity is realistically set to zero.

The atmospheric component is a one-layer reduced complexity model based on the energy-moisture balance model of Fanning and Weaver (1996). The model uses vertically-integrated thermodynamic equations to allow for the distribution of thermal energy and moisture content. Precipitation is assumed to occur whenever the relative humidity is greater than 85%. Precipitation falls as snow when the surface air temperature falls below a critical value (usually -5°C). Precipitation that falls over land, or snow that melts on land, is returned instantaneously to the ocean via one of the 33 drainage basins. A specified lapse rate is used to represent the cooling of surface temperatures over topography.

The sea-ice model assumes that the ice has no heat capacity and that the surface temperature is in instantaneous balance with external forcing. The model predicts ice thickness, areal fraction, and ice surface temperature. Two categories of ice are used: open water and ice-covered, allowing for grid cells to be partially covered by sea-ice.
2.2 Experimental Design

Improvements to the previous version of the model, version 2.6, include increased moisture diffusion in the atmosphere to better simulate the fronts across the Southern Ocean. Also, the isopycnal diffusion has been decreased, resulting in a somewhat slower and more realistic response of the ocean to anomalies. The along isopycnal diffusivity used in this model is \(4 \times 10^6 \text{ cm}^{-2} \text{s}^{-1}\). The representation of the current climate in the control (CNTRL) case will be examined in Section 3.1.

2.2 Experimental Design

The perturbations for the following experiments are motivated by both observational and model evidence of climate change over the past century. This includes changes in freshwater and heat fluxes and shifts in the wind stress. Possible sources of freshwater include a decline in the sea-ice extent and increased precipitation. Curran et al. (2003) and de la Mare (1997) report a decrease in the sea-ice extent around Antarctica of approximately 1.5-2.8° latitude (equivalent to 20-25% meltback). Climate change experiments (Allen and Ingram (2002)) also suggest an increase in precipitation of approximately 0.1-0.2 mm day\(^{-1}\) over the Southern Ocean over a multi-decadal year period (it should be noted that direct measurements of rainfall over the Southern Ocean are extremely sparse). Changes in the heat-fluxes are motivated by global warming studies and a warming of the upper water column in the latitude range of the ACC of approximately 0.17°C since the 1950’s (Gille (2002)). The Intergovernmental Panel on climate Change (IPCC, 2001), a broad review of all available literature in climate science, report an increase of 2.5 W m\(^{-2}\) in radiative forcing since 1750
2.2 Experimental Design

Figure 2.1: A summary of the areas (shaded black) where forcing is to be applied in experiments (a)ICE, (b)PRECIP and (c)HF. The 5.4° latitude shift in zonally averaged $\tau_x$ for experiment WINDS is shown in (d) with the CNTRL case in black, the shifted WINDS case in blue dashed, and the difference, WINDS - CNTRL, in red dashed.

to 2000 due to greenhouse gases. Finally, a poleward shift in the subpolar westerly winds, linked to the Southern Annular Mode (SAM) trending towards a more positive phase (Thompson et al. (2000), Thompson and Solomon (2002)) will also be considered.

This report will focus on two main sections based on four experiments to be presented in Chapters 3 and 4. Both sections will involve four main simplified forcing experiments. Specifically, we will perturb the control experiment sequentially by a separate mode of forcing corresponding to each of the changes referred to above. This will include adding freshwater to the winter sea-ice extent (not including the permanent ice region; see Fig. 2.1(a)) to simulate increased sea-ice melt equal to 25% of the
2.2 Experimental Design

mean winter sea-ice volume (denoted as ICE). In the second experiment freshwater will be added linearly to the ocean at 55-65°S (Fig. 2.1(b)) up to an increase of 0.2 mm day\(^{-1}\) in precipitation at the end of the simulation (denoted as PRECIP). In the third experiment an extra heat flux of 2.5 W m\(^{-2}\) will be added to the area of the Southern Ocean where Gille (2002) found the most pronounced warming (Fig. 2.1(c); denoted as HF). Finally, an anomaly will be added, in the fourth experiment to the subpolar westerlies to shift them south by 5.4° (Fig. 2.1 (d); denoted as WINDS).

The above changes will be at first added linearly over a 100 year time period to allow the general response of the ocean and climate system to be compared between the four processes (century scale experiments will be denoted with the subscript \(_{100}\)). These century scale experiments will then be re-run with some of the perturbations applied over a shorter time period to simulate the enhanced effect of these processes during the late part of the twentieth century. This enables a more quantitative comparison with observations of AAIW (denoted by the subscript \(_{40}\)). Specifically, each of the sea-ice, precipitation and winds anomalies are thought to have appeared most prominently in the last three to four decades of the twentieth century (de la Mare (1997), Allen and Ingram (2002), Marshall (2003)). The century scale heat-flux experiment, in contrast, is already appropriately configured to take account of the enhanced greenhouse warming of the past 100 years in comparison to the rest of the pre-industrial period. Experiments ICE\(_{40}\), PRECIP\(_{40}\) and WINDS\(_{40}\) will have their perturbations applied over a 40 year time period whilst HF\(_{100}\) will be used to assess the causes of the late twentieth century changes in AAIW properties. Differences for the 40 year experiments will be based on the last 30 years of integration to be compared with the observations of Wong.
et al. (1999). All results presented are based on 3 year means; i.e years 38-40 minus 1-3. The experiments are summarised in Tables 2.1 and 2.2.

It has become common to report temporal changes of water masses on density surfaces or isopycnals rather than on isobars to interpret the cause of climate change. This is because the differences on isobars (or z-coordinates, where 1dbar = 1m) are influenced by both changes in the water-mass properties and the shoaling or deepening of isopycnals, also referred to as “heave” (Bindoff and McDougall (1994)). Thus, to eliminate the effects of heave, differences on isopycnals or density surfaces are often used. The results presented here for temperature and salinity difference fields will be calculated for AAIW depths on both z-levels (i.e geopotential, or depth coordinates) and isopycnal levels. Differences on isopycnals are calculated by first interpolating the fields onto isopycnal surfaces, taking the differences on those surfaces and then interpolating back to depth coordinates (as done in Oke and England (2004)). This is to remove the effects of heaving of isopycnals, leaving only pure water-mass changes in the difference fields.
### 2.2 Experimental Design

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Physical Process</th>
<th>Description</th>
<th>Amount</th>
<th>Duration (years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>CNTRL</td>
<td>Pre-industrial climate</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>ICE(_{100})</td>
<td>Increased sea-ice melt</td>
<td>Add freshwater to the sea-ice margin</td>
<td>25% of current sea-ice</td>
<td>100</td>
</tr>
<tr>
<td>PRECIP(_{100})</td>
<td>Increased precipitation</td>
<td>Add freshwater to 55-65(^\circ)S</td>
<td>0.2 mm d(^{-1})</td>
<td>100</td>
</tr>
<tr>
<td>HF(_{100})</td>
<td>Increased SST</td>
<td>Add extra heat flux to the ocean at 55-65(^\circ)S</td>
<td>2.5 W m(^{-2})</td>
<td>100</td>
</tr>
<tr>
<td>WINDS(_{100})</td>
<td>Shifting westerly winds</td>
<td>Shift winds westerly winds</td>
<td>Shift south by 5.4(^\circ) latitude</td>
<td>100</td>
</tr>
</tbody>
</table>

**Table 2.1:** Summary of experiments for the century-scale perturbations described in Chapter 3. See also Fig. 2.1 for areas of perturbation.

<table>
<thead>
<tr>
<th>Experiment</th>
<th>Physical Process</th>
<th>Description</th>
<th>Amount (total at end)</th>
<th>Duration (years)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ICE(_{40})</td>
<td>Increased sea-ice melt</td>
<td>Add freshwater to the sea-ice margin</td>
<td>25% of the current sea-ice</td>
<td>40</td>
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<td>PRECIP(_{40})</td>
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<td>0.2 mm d(^{-1})</td>
<td>40</td>
</tr>
<tr>
<td>WINDS(_{40})</td>
<td>Shifting westerly winds</td>
<td>Shift winds</td>
<td>Shift south by 5(^\circ) latitude</td>
<td>40</td>
</tr>
</tbody>
</table>

**Table 2.2:** Summary of experiments for the late twentieth century perturbations described in Chapter 4. See also Fig. 2.1 for areas of perturbation.
Chapter 3

Century Scale Experiments

3.1 Control Experiment (CNTRL)

The CNTRL experiment has been integrated for 3400 model years until a statistical equilibrium was reached. The CNTRL run is in good agreement with the preindustrial climate. The main components of the ocean will be reviewed here. The horizontal transport streamfunction is shown in Fig. 3.1 and the maximum transport rates of some major currents shown in Table 3.1. The major western boundary currents compare well with reported estimates from observations. The maximum strength of the ACC when taken through Drake Passage is 119.9 Sv ($1 \text{ Sv} \equiv 10^6 \text{ m}^3 \text{ s}^{-1}$), the Agulhas Current transport is 74.6 Sv and the Brazil Current is 29.3 Sv. The maximum strength of the East Australian Current (EAC) when taken to be the main southward flowing branch is 40.4 Sv. This is within the range of estimates from observations although much of this is recirculated and the total transport into the Tasman Sea is reported to be 8-25.5 Sv (Chiswell et al., 1997). For a coarse resolution model there is a good representation of western boundary current transport rates.
Transport values for the global zonally averaged meridional overturning (MOT) (Fig. 3.2(a)) are also in reasonable agreement with observations. The Deacon Cell in the Antarctic Circumpolar region reaches down to approximately 1500 m and shows the main downwelling branch at 35-50°S that is associated with subduction. The maximum transport in the Deacon Cell is 33.5 Sv which is typical of ocean climate models. The AABW production rate has been estimated from observations to be around 8 Sv (Orsi et al., 1999). The AABW production for experiment CNTRL is only 2.6 Sv which is somewhat smaller than the observed, although this mismatch could be due in part to the zonally-averaging of the overturning cell to estimate AABW production in models (England et al. (2006) in preparation). Overturning in the North Atlantic in
### Table 3.1: Transport values for selected currents in the Southern Hemisphere: Antarctic Circumpolar Current (ACC); East Australian Current (EAC); Agulhas Current and the Brazil Current. Values given in Sv.

<table>
<thead>
<tr>
<th>Current</th>
<th>CNTRL Maximum Transport (Sv)</th>
<th>Observed Transport Range (Sv)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ACC</td>
<td>119.9</td>
<td>98-154 Whitworth et al. (1982)</td>
</tr>
<tr>
<td>EAC</td>
<td>40.4</td>
<td>22.2-42.2 Chiswell et al. (1997)</td>
</tr>
<tr>
<td>Agulhas</td>
<td>74.6</td>
<td>40-100 Beal and Bryden (1999)</td>
</tr>
<tr>
<td>Brazil</td>
<td>29.3</td>
<td>30-56 Maamaatuaiahutapu et al. (1998)</td>
</tr>
</tbody>
</table>

Fig. 3.2(b) shows North Atlantic Deep Water (NADW) production and outflow into the Southern Ocean. Maximum formation rates of NADW are 20.6 Sv which is just above the range of observed values for NADW (Gordon, 1986). The outflow rate of approximately 15 Sv is in good agreement with observed estimates.

The temperature and salinity profile of the model is in reasonable agreement with the Levitus Climatology (Levitus et al. (1994), Levitus and Boyer (1994)). The fresh salinity tongue is well captured, with the $\sigma_{27.2}$ surface running through the centre of the salinity minimum (Fig. 3.3). Zonally averaged fields of temperature and salinity in the AAIW domain are also shown (Fig. 3.4). This reveals the fresh salinity tongue and the steep temperature gradient across the Southern Ocean.
Figure 3.2: Zonally averaged meridional overturning (MOT) (a) globally and (b) across the Atlantic basin for the CNTRL experiment. Contour interval is 2 Sv with every second contour labeled. Negative values (dashed) indicate anti-clockwise motion and positive values (continuous) indicate clockwise motion. The zero contour is in bold. Note that there is a difference horizontal scale for (a) and (b).
Figure 3.3: Salinity fields through the Atlantic (24 °W), Pacific (150 °W) and Indian (60 °E) sectors in the (left) model and (right) Levitus climatology. The thick line represents the mean position of $\sigma_{27.2}$ density surface. Note that the same colour scale has been used for each ocean sector and the contour interval is 0.1 psu.

Figure 3.4: Zonally-averaged (a) potential temperature and (b) salinity for CNTRL, focusing on the AAIW region. $\sigma_{27.2}$ is shown in bold in (b). Contour interval is (a) 1 °C and (b) 0.05 psu.
3.2 Sea-ice (ICE$_{100}$)

The sea-ice melt experiment shows a maximum cool and fresh anomaly of 0.2°C and 0.06 psu at the surface at approximately 55°S (Fig. 3.5). This then weakens as it extends into the interior of the ocean along isopycnals, as expected, to a maximum depth of approximately 1300-1500 m at 40-50°S. The anomaly at intermediate depths is 0.063°C and 0.0218 psu. The anomaly then shoals as it spreads further north to the equator. The prominent band of warming to the south of 60°S indicates a reduction in the convective overturning in this region caused by the addition of fresh and buoyant water to the system.
For salinity the differences on isopycnals are similar to those on z-coordinates (compare Figs 3.5(c),(d)). In contrast for temperature the differences on isopycnals reveal an increased cooling. This is best explained with the aid of a temperature-salinity (T-S) diagram (Fig. 3.6). When a water-mass is freshened, the T-S curve is displaced to the left to lower salinity values. It can be seen that the blue curve is fresher than the black curve. Water at the same z-level is marked with a cross for the fresher curve and a circle for CNTRL. The difference in temperature between the points marked is $T_2 - T_1$. When comparing water-masses on the same isopycnal as the blue cross this water-mass is now at the intersection of the red dotted (representing an isopycnal) and black solid lines. The difference is now $T_3 - T_1 (> T_2 - T_1)$ i.e larger than the difference.
calculated on z-coordinates. Hence, a more significant cooling is found on isopycnals. This is similar to the “pure freshening” process described by Bindoff and McDougall (1994) where a freshening of surface waters results in a cooling and freshening on isopycnals when subducted into the thermocline.

A frequently observed zonal section, 32°S across the Indian Ocean (e.g Wong et al. (1999), Bindoff and McDougall (2000)) is shown in Fig. 3.7 to demonstrate the nature of the impact each perturbation has on the T-S curve. It can be seen that the curve is displaced to the left to lower salinity values for experiment ICE\textsubscript{100}. The difference is more pronounced at lower densities as this is closer to the surface, and the source of the forcing.
Figure 3.7: T-S curves zonally averaged at 32°S across the Indian sector. CNTRL is shown in solid black and the dashed blue curve represents the corresponding experiment at the end of the century-scale integration for (a) ICE$_{100}$, (b) PRECIP$_{100}$, (c) HF$_{100}$ and (d) WINDS$_{100}$. 

3.2 Sea-ice (ICE$_{100}$)
3.3 Precipitation (PRECIP\textsubscript{100})

Results for the PRECIP\textsubscript{100} experiment are similar to those of ICE\textsubscript{100} except that the maximum surface anomaly is further north and overall there is a more significant cooling and freshening (Fig. 3.8). This is consistent with the location and amount of freshwater that was added to each experiment. The rate of freshwater added to the ICE\textsubscript{100} and PRECIP\textsubscript{100} experiments is equivalent to 62 cm y\textsuperscript{-1} and 73 cm y\textsuperscript{-1}, respectively.

Fig. 3.8 reveals the maximum zonal-mean cooling and freshening at the surface at approximately 50°S to be more than 0.1°C and 0.08 psu. The maximum anomaly at intermediate depths is 0.1236°C and 0.0325 psu. The band of warming to the south, as in ICE\textsubscript{100}, is due to a reduction in convective overturning due to surface freshwater anomalies. Again, a stronger cooling signal is found on isopycnals for reasons explained in Section 3.2. The impact of the precipitation anomalies on the T-S curve (Fig. 3.7(b)) is also similar to that of ICE\textsubscript{100}, as expected.
Figure 3.8: Temperature and salinity differences as in Fig. 3.5, only for PRECIP\textsubscript{100} - CNTRL.
3.4 Heat-flux (HF\textsubscript{100})

The addition of 2.5 W m\textsuperscript{-2} over 100 years in experiment HF\textsubscript{100} causes the global surface air temperature (SAT) to rise by a maximum of 0.68 °C in the Southern Hemisphere (Fig. 3.9). This is most concentrated over the southwest Indian Ocean. The SST in the Southern Ocean increases by a maximum of 0.89°C (Fig. 3.10(a)).

Fig. 3.11(a) reveals how the zonal-average water column is warmed due to surface heat fluxes. The maximum temperature anomaly is found at the surface at approximately 55°S. This then extends into the interior of the ocean via along isopycnal mixing, as expected, to warm intermediate waters and extends beyond. The maximum warming at intermediate water depths on z-coordinates is 0.0920°C. A weak freshening of 0.0028 psu can also be observed on z-coordinates. This is due to the addition of freshwater from increased sea-ice melt (see Fig. 3.12) due to the increase in SST and SAT. This meltwater is advected north via Ekman dynamics into the area of AAIW formation.

On isopycnals the warming signal is not apparent in the latitudes north of 50°S.
Figure 3.10: (a) Sea surface temperature (SST) for HF$_{100}$ - CNTRL and (b) sea surface salinity (SSS) for HF$_{100}$ - CNTRL shown for the Southern Hemisphere. Units are in (a) °C and (b) psu.
Figure 3.11: Temperature and salinity differences as in Fig. 3.5 only for HF$_{100}$ - CNTRL.

Figure 3.12: Difference in ice area fraction for HF$_{100}$ - CNTRL.
above intermediate depths. In this area there is a weak zonal-mean cooling up to 0.0632°C. This result can be explained by the counter intuitive process of “pure warming” described by Bindoff and McDougall (1994). This involves a warming of the water column due to surface fluxes. When the warmer water is subducted into the thermocline, it is actually both cooler and fresher when compared on isopycnals. This also explains why the salinity anomaly is amplified to 0.013 psu on isopycnals. To verify this process in these results, a T-S diagram comparing HF\textsubscript{100} and CNTRL is examined in Fig. 3.13. The HF\textsubscript{100} curve is displaced upwards to higher temperature values. By comparing water at the same z-level it can be seen that a water parcel on the HF\textsubscript{100} curve \((S_2, T_2)\) is warmer than a parcel at the same z-level on the CNTRL curve \((S_1, T_1)\). However, if parcel \((S_2, T_2)\) on the HF\textsubscript{100} curve is compared to water at the same density on CNTRL, marked by \((S_3, T_3)\), then it can be seen that the parcel on HF\textsubscript{100} is now cooler than the parcel on CNTRL. This explains why a warming on z-coordinates can manifest as a cooling on isopycnals.

The complete T-S curve along 32°S across the Indian sector for HF\textsubscript{100} is shown in Fig. 3.7(c). This shows the T-S curve displaced to both warmer and fresher values (although cooler on isopycnals). This is more pronounced at lower densities’, close to the surface where the warming anomalies are applied.
Figure 3.13: A section of a T-S diagram for HF$^{100}$ zonally-averaged across 32°S showing how a warming of the water column can be described as a cooling on isopycnals. HF$^{100}$ is shown in blue dashed and CNTRL in solid black. Measurements taken at the same z-level is shown by a cross labelled ($S_2, T_2$) for HF$^{100}$ and a circle labelled ($S_1, T_1$) for CNTRL. The measurement for CNTRL at the same isopycnal level as ($S_2, T_2$) is marked with a star labelled ($S_3, T_3$) and is cooler compared to ($S_2, T_2$).
3.5 Winds (WINDS\textsubscript{100})

The zonally-averaged wind stress, \(\tau_x\), is shifted 5.4\degree poleward in the WINDS\textsubscript{100} experiment. This perturbation produces many changes to the climate system and is not as easily explained as the previous experiments. A range of fields will be reviewed here to account for these results.

The ocean’s interior temperature and salinity response is shown in Fig. 3.14. A band of warming can be seen at approximately 45-55\degree S that extends down from the surface on z-coordinates. This warming also dominates surface layers. A cool anomaly is found from approximately 500 m depth down below intermediate depths. This cooling is not linked directly to the surface. The maximum cooling of 0.5\degree C is found at around 600 m and at 45\degree S. Cooling can also be found south of 60\degree S below around 200 m. These patterns of temperature change can be linked to changes in the MOT circulation. Fig. 3.15 shows the temperature change of Fig. 3.14(a) with contours of MOT overlaid. Generally speaking, regions of warmer water correspond to anomalous downwelling, and cooler water to regions of anomalous upwelling. A high saline anomaly can also be found on z-coordinates that extends down from the surface at 45-50\degree S. A freshening up to 0.0388 psu is found in the same area as that of maximum cooling, (see Figs 3.14(a), (c)).

On isopycnals the warming band at 45-65\degree S is weakened but is still up to 0.5\degree C. The areas of cooling and freshening at 500 m have become less significant, now only 0.0991\degree C and 0.0091 psu, indicating the effects of “heaving” of isopycnals via the wind stress
Figure 3.14: Temperature and salinity differences as in Fig. 3.5 only for WINDS\textsubscript{100} - CNTRL.

Figure 3.15: Zonally-averaged temperature differences for WINDS\textsubscript{100} - CNTRL shown in colour in °C. Overlaid are contours for MOT differences. Contour interval is 1 Sv.
Figure 3.16: Upper parts of T-S curves for CNTRL (black solid) and WINDS\textsubscript{100} (blue dashed) zonally-averaged along 32\(^\circ\)S. Measurements at the same z-level are marked with circles for CNTRL and crosses for WINDS\textsubscript{100}. Dotted lines are isopycnals surfaces. 1-4 have the same temperature tendency on z-coordinates and isopycnals. At the fifth level a cooling can be seen on z-coordinates. When compared along isopycnals the difference is minimal and shows a slight warming.
Figure 3.17: Meridional overturning (MOT) for (a) CNTRL and (b) the difference WINDS$_{100}$ - CNTRL. Solid lines are positive values (clockwise) and negative values are dashed (anti-clockwise). The zero contour is bold. Contour interval is (a) 2 Sv and (b) 1 Sv with every second contour labeled.
Figure 3.18: Horizontal transport streamfunction for WINDS\textsubscript{100} (top) and the difference (CNTRL - WINDS\textsubscript{100}) overlaid on SST differences (bottom). Contour interval is 2 Sv with every second contour labeled. Colour bar is in °C.
(and Ekman pumping) changes. This result contrasts those found for HF$_{100}$ whereby the “pure warming” process was apparent and a cooling was seen on isopycnals. This is examined in Fig. 3.16 showing a zonally-averaged T-S diagram along 32$^\circ$S. On z-coordinates this section shows warming at the surface and cooling below 250 m depth. Results taken at the same z-level in Fig. 3.16, marked by circles (CNTRL) and crosses (WINDS$_{100}$) show this. When comparing water along the same isopycnal the first 3 levels lie on approximately the same isopycnal, thus differences on z-coordinates and isopycnals are roughly the same. The fourth model level taken begins to show a slightly increased warming when differences are taken along isopycnal surfaces. For the fifth z-level, the WINDS$_{100}$ temperature field is cooler on z-coordinates. However, when compared along the same isopycnal, the difference between WINDS$_{100}$ and CNTRL is close to zero. This is an example of “pure heave” as described by Bindoff and McDougall (1994). Mathematically this is

$$\alpha T\prime|_n = \beta S\prime|_n = 0$$  \hspace{1cm} (3.1)

where changes in temperature and salinity on neutral surfaces (or isopycnals), $S\prime|_n$ and $T\prime|_n$, multiplied by the thermal expansion and haline contraction coefficients, $\alpha$ and $\beta$ respectively, results in a neutral change.

Both horizontal and vertical transport in the ocean is affected by the wind shift in WINDS$_{100}$. The wind shift has a significant impact on the MOT, particularly the Deacon cell (Fig. 3.17). The Deacon cell shifts south corresponding to the change in zonal winds and Ekman transport. The topography of Drake Passage also allows the
cell to deepen. The poleward shift of the winds also shifts the subtropical gyres in the Southern Hemisphere to the south (Fig. 3.18). This has the effect of transporting more warmer waters from the subtropical region further south. The ACC is also shifted southward and the transport through the Drake Passage increased by 8 Sv. Shifts in the gyres also results in a change the areas of upwelling and downwelling (as will be described below).

Fig. 3.19 shows SSS is fresher in WINDS\textsubscript{100} near Antarctica in the Ross Sea and extending along the Antarctic Peninsula. This can be attributed to a decline in sea-ice (Fig. 3.20) from increased SST and warmer SAT. A zonal band of higher salinity water is found at approximately 65-70°S in WINDS\textsubscript{100} (Fig. 3.19(b)). The zonally averaged horizontal advection of salt due to wind-driven Ekman transport can be approximated by

$$\sum_{i=1}^{\text{latitude}} (v_E S_E)$$

(3.2)

where $v_E$ is the y-component of the surface velocity (dominated by Ekman flow at 40-56°S, Santoso and England (2004)) and $S_E$ is salinity at the surface. Fig. 3.21 shows the zonally-averaged salinity transport in CNTRL and WINDS\textsubscript{100}. The northward salinity transport has been shifted south and has also decreased suggesting that fresher waters have been advected from further south.

A significant warming of SST is apparent for the majority of the Southern Hemisphere in WINDS\textsubscript{100}(Fig. 3.19(a)). This warming is highest in the Southern Ocean to the southwest of Australia and to the west of South America. There is a neutral
3.5 Winds (WINDS\textsubscript{100})

Figure 3.19: Difference fields for WINDS\textsubscript{100} - CNTRL for (a) sea surface temperature (SST) in °C and (b) sea surface salinity (SSS) in psu. Overlaid are differences for surface velocity vectors for WINDS\textsubscript{100} - CNTRL.
response in the Ross Sea and the Antarctic Peninsula (i.e. under the permanent sea-ice zone). To investigate the cause of this warming the terms in the heat budget for the mixed layer (Eqn. 3.3), similar to that presented by Rintoul and England (2002), will be considered:

\[
\frac{\partial T_s}{\partial t} = \frac{Q_{\text{net}}}{\rho_0 C_p h_s} - u \frac{\partial T_s}{\partial x} - v \frac{\partial T_s}{\partial y} - w \frac{\partial T}{\partial z} + \text{mixing terms}
\]  

(3.3)

where \( T_s \) is the surface temperature, \( Q_{\text{net}} \) is the net air-sea heat exchange, \( \rho_0 \) is density, \( C_p \) is the specific heat, \( h_s \) is the surface layer depth, \( u \) is the east-west velocity, \( v \) the north-south velocity, \( w \) is the vertical velocity, \((x, y, z)\) is three-dimensional space and \( t \) is time. In the above equation we do not estimate the magnitude of mixing terms, such as \( T_s \) changes due to vertical convective overturning. The terms \( u \frac{\partial T}{\partial x} \) and \( v \frac{\partial T}{\partial y} \) are the horizontal advection of temperature, and \( w \frac{\partial T}{\partial z} \) is the vertical advection of temperature.

Fig. 3.22 shows changes in \( Q_{\text{net}} \) in the WINDS\(_{100}\) experiment relative to CNTRL. The negative areas reveal regions where more heat is being lost from the ocean to the atmosphere in WINDS\(_{100}\). This is indicative of increased convective overturning. Vectors in Fig. 3.19(a) show that contributions from \( u \frac{\partial T}{\partial x} \) are only weak as there is little zonal temperature gradient (i.e. \( \frac{\partial T}{\partial x} \) is small; see Fig. 3.19). Velocity vectors in Fig. 3.19 suggest that cooler water from the south would be transported north via \( v \frac{\partial T}{\partial y} \), as is verified by Fig. 3.23 showing changes in temperature due to northward transport of cooler waters from the south. This cooling is absent from total changes in SST (Fig. 3.19(a)), so it must be dominated by another term that warms the surface of the ocean. Fig. 3.24 shows a band of increased downwelling that would have the effect of
warming surface waters by either suppressing the upwelling of cooler waters or by increasing the vertical dispersion of heat from the surface. The heat flux due to vertical advection $w \frac{\partial T}{\partial z}$ is shown in Fig. 3.25, and reveals a band of warming across 40-60°S, symptomatic of increased downward Ekman pumping in this region.

Fig. 3.26 shows how the convection depth has changed in the WINDS$_{100}$ experiment. In the areas corresponding to the two main regions of SST warming the convection depth has increased quite markedly (up to 1000 m in the far southeast Indian Ocean). Because there is a temperature inversion in the region, increased convection has the effect of bringing warmer and saltier waters up to the surface from below.

In summary, much of the surface ocean temperature warms in WINDS$_{100}$, due to a combination of north-south advection (north of 50°S, Fig. 3.23(b)), Ekman pumping change (40-60°S, Fig. 3.25) and convective overturning changes (Fig. 3.26).
Figure 3.20: Sea-ice area fraction difference for WINDS_{100} - CNTRL.

Figure 3.21: Zonally-averaged SSS multiplied by surface northward ocean velocity. WINDS_{100} is shown by the blue dashed line and CNTRL is the solid black.
3.5 Winds ($\text{WINDS}_{100}$)

**Figure 3.22:** Heat flux, $Q_{\text{net}}$, difference for $\text{WINDS}_{100}$ - CNTRL. Units are in W m$^{-2}$.

**Figure 3.23:** Northward advection of heat $-v \frac{dT}{dy}$ ($^\circ$C s$^{-1}$) for (a) CNTRL and (b) the difference in $\text{WINDS}_{100}$ - CNTRL.
Figure 3.24: Difference in $w$ (m s$^{-1}$) from WINDS$_{100}$ - CNTRL where $w$ is taken to be positive upward.

Figure 3.25: Difference in the vertical heat fluxes $w \frac{dT}{dz}$ (${^\circ}$C s$^{-1}$) for WINDS$_{100}$ - CNTRL.

Figure 3.26: Convection depth differences (m) for WINDS$_{100}$ - CNTRL.
Chapter 4

Twentieth Century Simulations

A key unanswered question in physical oceanography is the cause of the late twentieth century cooling and freshening of AAIW, and whether this can be associated with anthropogenic climate change. To investigate this perturbations were applied for a 40-year period in experiments ICE$_{40}$, PRECIP$_{40}$, and WINDS$_{40}$ to capture the accelerated response of the climate system to global warming in the latter half of the twentieth century. In contrast, HF$_{100}$ was not re-run over 40 years, because a century scale increase of heat fluxes to 2.5 W m$^{-2}$ is already a reasonable approximation of the twentieth century warming trend due to greenhouse gases.

The interior ocean temperature and salinity response over the last 30 years of each simulation, calculated on isopycnals, is presented in Fig. 4.1 and Fig. 4.2 respectively. It can be seen that a modest cooling of AAIW north of 50°S is apparent in ICE$_{40}$. The cooling at intermediate levels is a maximum of 0.0694°C. This cool and fresh anomaly is concentrated at the surface near the ice-melt regions. Temperature changes for PRECIP$_{40}$ are similar to those of ICE$_{40}$ but stronger. A maximum zonal-mean
Figure 4.1: Zonally-averaged temperature differences (°C) for the last 30 years on isopycnals for (a) ICE$_{40}$, (b) PRECIP$_{40}$, (c) HF$_{100}$ and (d) WINDS$_{40}$. Contour interval is 0.1 °C and the same colour bar is used for all plots. The bold dashed line represents $\sigma_{27.2}$ from CNTRL.
Figure 4.2: Same as Fig. 4.2 except for salinity differences. Contour interval is 0.01 psu.
cooling of 0.1575°C is found as far down as 800 m depth at 50°S in this experiment with forced precipitation anomalies. The cooling in both ICE_{40} and PRECIP_{40} reaches as far north as the equator, although generally appearing above the depths of AAIW once the anomaly gets north of approximately 20°S.

A fresh anomaly is found to mirror the cooling on isopycnals for both ICE_{40} and PRECIP_{40} (Fig. 4.2(a), (b)). At the surface the maximum freshening is 0.04 psu for ICE_{40} and 0.06 psu for PRECIP_{40}. The anomaly for PRECIP_{40} is stronger at intermediate depths where a zonal-mean freshening of 0.01-0.02 psu is found. For ICE_{40} the freshening is slightly weaker at around 0.01 psu. Like the cooling anomaly in both experiments the freshening shoals above \( \sigma_{27.2} \) at around 25°S.

For HF_{100} a cooling on isopycnals is apparent over the last 30 years as a result of the “pure warming” process as discussed in Section 3.4. This cooling is found at all intermediate water depths north of 50°S, though it is only weak at less than 0.1°C. A freshening of approximately 0.01 psu in the zonal-mean is also found in the same area as this cooling for HF_{100} (Fig. 4.2(c)). Freshening of intermediate waters can also be attributed to the “pure warming” process as described previously.

A warming dominates AAIW temperature changes on isopycnals for WINDS_{40} at 40-60°S and also north of 25°S. Unlike WINDS_{100}, where the cooling anomaly below 200 m that was apparent on z-coordinates vanished on isopycnals, a cooling of up to 0.1337°C in the zonal-mean persists in this area (40-30°S). This is mirrored by a freshening in the same area of up to 0.0285 psu. These cool and fresh anomalies are found in
an area that overlaps the locations of AAIW and Subantarctic Mode Water (SAMW). A warm and saline anomaly dominates the upper water column above these interior regions. A summary of the maximum cooling and freshening of AAIW on isopycnals caused by all twentieth century experiments is given in Table 4.1.

T-S curves across the sections observed by Wong et al. (1999) are also presented to allow for comparison to observations of AAIW in the late twentieth century (Figs 4.3 - 4.6). At 10°N and 17°S across the Pacific both ICE$_{40}$ and PRECIP$_{40}$ show very little changes. At 32°S across the Pacific, both of these curves begin to show signs of freshening. At 43°S across the Tasman a freshening can be more readily identified. The most significant changes to the T-S curves for the last 30 years of HF$_{100}$ are at 43°S. T-S curves for WINDS$_{40}$ fail to show any significant changes at 10°N in the Pacific. The T-S curve at 17°S is displaced to higher salinity values. At 32°S and 43°S the curve is displaced to higher salinity values for the lighter isopycnals at the surface. At AAIW levels, $\sigma_{27.2}$, the curve is displaced to fresher values. This is consistent with the results presented in Fig. 4.2(d). These T-S curves will be discussed further in Section 5.1.2.

Both the experiments ICE$_{40}$ and PRECIP$_{40}$ had similar impacts on the temperature and salinity properties of AAIW. PRECIP$_{40}$ had a larger cooling and freshening and is closer to the observations of AAIW late twentieth century change. The “pure warming” process that was dominant in HF$_{100}$ managed to cool and freshen AAIW although only to a small extent compared to other experiments. The WINDS$_{40}$ experiment has a significant impact on the temperature and salinity of AAIW. Although a
significant cooling and freshening of AAIW was found there was also some warming and higher salinity waters indicating that this would not be the sole mechanism causing the observed changes in AAIW.

Figure 4.3: Zonally-averaged T-S curves across the Pacific Basin at (a) 10° N and (b) 17° S; across (c) across the Indian at 32° S; and (d) across the Tasman Sea at 43° S. Solid lines show the results after 10 years and dashed blue lines show results for 40 years.
<table>
<thead>
<tr>
<th>Experiment</th>
<th>Maximum cooling (°C)</th>
<th>Maximum Freshening (psu)</th>
</tr>
</thead>
<tbody>
<tr>
<td>ICE$_{40}$</td>
<td>0.0694</td>
<td>0.0156</td>
</tr>
<tr>
<td>PRECIP$_{40}$</td>
<td>0.1575</td>
<td>0.0254</td>
</tr>
<tr>
<td>HF$_{100}$</td>
<td>0.0809</td>
<td>0.0165</td>
</tr>
<tr>
<td>WINDS$_{40}$</td>
<td>0.1337</td>
<td>0.0285</td>
</tr>
</tbody>
</table>

**Table 4.1:** Summary of the maximum cooling and freshening of AAIW on isopycnals for the final 30 years for experiments ICE$_{40}$, PRECIP$_{40}$, HF$_{100}$ and WINDS$_{100}$.

**Figure 4.4:** Same as in Fig. 4.3 only for PRECIP$_{40}$. 


Figure 4.5: Same as in Fig. 4.3 only for the last 30 years of HF$_{100}$. 
Figure 4.6: Same as in Fig. 4.3 only for WINDS$_{40}$. 
Chapter 5

Discussion and Conclusions

5.1 Discussion of Results

5.1.1 Century-Scale Experiments

The role of the century-scale experiments was to examine the potential change to the climate system the four processes would have over a uniform time scale. Measurements from observations and climate model predictions were used as a guide when applying the forcing. It should be noted that these experiments are not predictions of the future climate over the coming century, nor have they been run to equilibrium with the additional fluxes.

The experiments involving the addition of freshwater, ICE$_{100}$ and PRECIP$_{100}$, produced qualitatively similar results. In both cases the surface anomalies propagated along isopycnals into the interior of the ocean to cool and freshen AAIW on both $z$-coordinates and along isopycnals. The cool and fresh anomalies began to upwell above
intermediate depths and approached the surface closer to the equator. The cooling and freshening in ICE\textsubscript{100} was less than PRECIP\textsubscript{100} due to the total amount and location of the source of freshwater.

There remains much debate over the stability of the sea-ice in Antarctica with some studies suggesting that southern hemisphere sea-ice is quite stable. The IPCC (2001) report that the Antarctic sea-ice has increased since 1978 after an initial decline in the mid-1970s. Thus, the experiments involving sea-ice melt are likely to be an over estimate of their impact on AAIW. There is also sparse data available for precipitation over the Southern Ocean, so actual increases over the past century are poorly known. We applied a change in precipitation at the high end of available estimates, so the PRECIP\textsubscript{100} results may be seen as an over estimate. The main result from both experiments, ICE\textsubscript{100} and PRECIP\textsubscript{100}, was that the addition of freshwater to the climate system, whether it be from increased precipitation, increased sea-ice melt or a combination of the two, has the potential to both cool and freshen AAIW. Other sources of meltwater, such as ice-shelf melt and glacial melt, could also act to cool and freshen AAIW.

The heat flux added to the ocean was applied linearly over the same period for both the century-scale and the late twentieth century experiments. Applying this forcing linearly over 100 years approximates the exponential increase in heat flux due to greenhouse gasses over a 250 year period. Although the forcing was only applied to the Southern Ocean, when applied on a larger global scale the impact on AAIW was markedly similar (global heat flux experiment not presented here). The results for this
experiment also demonstrated the importance of using isopycnals or density surfaces to diagnose water-mass variability, with a warming on z-levels transforming to a cooling on isopycnal surfaces.

The temperature anomaly for HF$_{100}$ shows a warming of 0.1°C of the upper 800 m on z-coordinates. This equates to a warming of 0.001-0.002°C yr$^{-1}$. This is just below the warming reported by Gille (2002) between 45°S and 60°S of 0.004°±0.001°C yr$^{-1}$.

An unavoidable consequence of applying the heat flux to the Southern Ocean was that some Antarctic sea-ice melted back. Thus, the freshening seen in the HF$_{100}$ experiment is not entirely due to the simple warming of the ocean; it is also affected by the resulting addition of meltwater at the sea-ice margin.

Experiment WINDS$_{100}$ involved looking solely at latitudinal shifts in the winds and their impact on AAIW temperature and salinity. A positive SAM phase corresponds to a stronger and more poleward contracted air jet (Sen Gupta and England, 2006). This increase in the strength of the winds is not examined here, instead we focus solely on the latitudinal shift in the winds. The results for WINDS$_{100}$ were much more complicated and unpredictable than the other experimental results. There were significant changes in the interior circulation of the ocean as well as the horizontal transport streamfunction and convective overturning. The changes in temperature and salinity for AAIW were somewhat similar to those found by Oke and England (2004) with the exception that the warming of the upper layers due to increases in SST was much more significant in the results presented here. The salinity anomaly was also much less sig-
5.1 Discussion of Results

Significant and is not connected to the surface. The temperature and salinity differences with Oke and England (2004) can be attributed to the unrealistic “restoring” forcing applied in their ocean-only modelling study. Restoring boundary conditions calculate fluxes of heat and freshwater based on the difference between the simulated state of the ocean and the observed climatology. In doing this, the climate response can be erroneous due to incorrect time lags, errors in amplitude and artificial damping of higher frequency components of the forcing (Pierce, 1996). The UVic model used in this study in contrast is coupled to an energy-moisture atmospheric model and provides a more realistic representation of the coupled ocean-atmosphere response to wind forcing.

Perhaps the most unforeseen result in WINDS$_{100}$ was the significant surface warming. Warming of the surface was expected where the Ekman flux of cool waters from the south was shifted poleward. An anticipated consequence of this was that a more pronounced cooling should be seen where the northward Ekman flux of water was shifted too. This was not the case.

Possible processes affecting heat content at the surface are described in equation (3.3). Although the change in surface temperature due to northward Ekman flux is evident in $v \frac{\partial T}{\partial y}$, this was overshadowed by other sources of heat. The loss of sea-ice due to the increase in SST and SAT was another factor to be considered. As the sea-ice extent decreases the ocean becomes more exposed to the cold atmosphere, thus enabling a greater release of heat. It can be seen that areas of significant sea-ice loss correspond to areas of negative heat flux into the ocean; i.e ocean heat loss. Thus, changes in $Q$ could not explain the surface heating in WINDS$_{100}$. The term $w \frac{\partial T}{\partial z}$ describes the heat
brought to the surface from changes in upwelling beneath the upper layer. Looking at \( w \) alone also reveals a band of increased downwelling (Fig. 3.24). Analysis of \( w \frac{\partial T}{\partial z} \) further confirmed that vertical advection of heat was an important factor in the warming the surface of the ocean in WINDS\(_{100}\). The areas for convection depth have also deepened, particularly in the southeast Pacific. This becomes the main area of formation of AAIW in the WINDS\(_{100}\) experiment. Convective overturning is also one of the main mechanisms for AAIW formation in the ocean. By increasing convection depth warmer subsurface water is brought to the surface while cooler SST is overturned to depth. The two major regions for SST warming in the Southern Ocean (below Western Australia and in the southeast Pacific) correspond to the areas of increased convection depth. This accounts for increases in SST south of 50°S. Further north, warming can be attributed to the shifts in the gyres changing regions of upwelling and thereby affecting the vertical advection of heat.

### 5.1.2 Twentieth Century Simulations

The aim of the late twentieth century simulations was to produce results that could be compared to observations of AAIW change (notably Wong et al. (1999) and Bindoff and Church (1992)). The forcing was applied over a 40-year period (with the exception of HF\(_{100}\) which remained over 100 years) to capture the increased effect of climate change in the latter half of the twentieth century. Differences were taken over the last 30 years as most observations of AAIW change report differences over a 20-30 year time period.
The results for changes in temperature and salinity over the last 30 years for ICE$_{40}$ produced a globally averaged change of 0-0.1°C and 0.01 psu on isopycnals. This is less than what was found by Bindoff and Church (1992) and Wong et al. (1999) for the Pacific. Zonally-averaged changes in the final 30 years for PRECIP$_{40}$ are more significant, with a zonal-mean freshening of 0.02 psu on isopycnal surfaces. Cooling of up to 0.1°C on isopycnals was similar to ICE$_{40}$ but was stronger in the core of AAIW. This is closer to what has been reported in observations.

The freshwater added to the system for ICE$_{40}$ and PRECIP$_{40}$ is equivalent to an increase of 62 mm yr$^{-1}$ and 73 mm yr$^{-1}$ respectively at the end of each simulation. Wong et al. (1999) estimate an increase of 31 mm yr$^{-1}$ to account for the changes in temperature and salinity in the source region of AAIW, in a particular freshening of 0.02 psu in the Pacific. It should also be noted that changes in T-S diagrams for ICE$_{40}$ and PRECIP$_{40}$ are qualitatively similar to those presented by Wong et al. (1999). Changes at 10°N in the Pacific are insignificant. Results at 17°S for ICE$_{40}$ and PRECIP$_{40}$ were also minimal in accordance with Wong et al. (1999), reporting a small freshening of the salinity minimum. At 32°S across the Indian Ocean ICE$_{40}$ still shows minimal change while a freshening of AAIW is beginning to appear in PRECIP$_{40}$, albeit weaker than the observations. Finally, at 43°S across the Tasman Sea, a freshening of the water-mass is apparent in both ICE$_{40}$ and PRECIP$_{40}$. There are significant differences at 43°S across the Tasman Sea. This is perhaps due to the small area that is zonally-averaged across and might be expected from a coarse resolution model.

The final 30 years of HF$_{100}$ shows a warming of 0.1°C of the upper 500 m on z-
coordinates. This equates to a warming of 0.003°C yr⁻¹. This rate of warming is more rapid than what was calculated over the full century, even though the addition of heat increases linearly in the experiment. This is due to the high specific heat of the ocean, which causes a delay in the ocean’s response to SAT changes.

The T-S curves for HF₁₀₀ of the last 30 years show minimal changes at 10°N and 17°S across the Pacific. At 32°S across the Indian the curve is displaced to slightly higher temperature values. The observations for AAIW at this transect clearly show a freshening of the salinity minimum. This is not apparent in the results for HF₁₀₀. Results for 43°S across the Tasman show a freshening above the salinity minimum.

The most notable difference between the century-scale WINDS₁₀₀ experiment and the final 30 years of WINDS₄₀ was that the cooling on isopycnals persists on both z-coordinates and on isopycnals, although it is less significant on isopycnals and is in reasonable agreement with the cooling anomaly found by Oke and England (2004). The T-S curves for WINDS₄₀ at 17°S and 32°S is displaced to higher salinity values unlike those in the observations. The curve at 43°S shows a shift to significantly more saline waters at lighter densities and a small freshening around the salinity minimum. The small area over which this T-S curve is zonally averaged over also contributes to the differences seen at 43°S.

The temperature changes on z-coordinates and isopycnals should also be investigated when comparing results to observations. Bindoff and Church (1992) clearly report a warming on isobars (z-coordinates) at 43° and 28°S across the Pacific of approximately
0.1°C at intermediate depths (their Fig. 2). When this is converted to differences on neutral surfaces (similar to isopycnal surfaces) a cooling of 0.1°C is found. Bindoff and McDougall (2000) also found a mean warming of the upper 900 dbar of 0.5°C at 32°S (Indian Ocean). On isopycnals, AAIW had cooled. Both ICE$_{40}$ and PRECIP$_{40}$ had an increased cooling on isopycnals that could counter-act any warming of the water column. The results for WINDS$_{40}$ in fact show the opposite to the observations, with the cooling anomaly on z-coordinates becoming less significant when interpolated onto isopycnals, leaving a warming effect dominating the temperature change. HF$_{100}$ was consistent with the “pure warming” process.

5.2 Conclusions

The century-scale experiments suggest that the addition of freshwater has the potential to both cool and freshen AAIW on both z-coordinates and isopycnals, with an increased cooling apparent on isopycnals. This flux of freshwater could be either at the AAIW formation region via precipitation or at the sea-ice margin (from where Ekman transport sends this anomaly northward). The addition of an extra heat flux to the Southern Ocean also showed that the “pure warming” process can cool and freshen AAIW on isopycnals, where there is a warming on z-coordinates. Shifting the winds south has significant implications for the climate system, causing a warming of the Southern Hemisphere and many changes in ocean circulation patterns. AAIW was cooler and fresher on z-coordinates, although this was significantly reduced along isopycnals indicating the effect of heaving of isopycnals.
Observations reveal a cooling and freshening of AAIW of 0.03 psu and 0.3°C on average across the Indian and Pacific basins over a 20-30 year period (Bindoff and Church (1992), Johnson and Orsi (1997), Wong et al. (1999), Bindoff and McDougall (2000), Wong et al. (2001), Aoki et al. (2005)). The results presented using the late twentieth century simulations show each process has the capacity to cool and freshen AAIW on isopycnals. This cooling and freshening is in the range of 0-0.1°C and 0-0.02 psu. There was a broad cooling and freshening that weakened as it approached the equator in the ICE$_{40}$ and PRECIP$_{40}$ experiments. HF$_{100}$ exhibited cooling and freshening for all AAIW north of 55°S over the last 30 years, and WINDS$_{40}$ had a localised area of cooling and freshening centred at 30°S.

To answer the question of what caused the late twentieth century cooling and freshening of AAIW, the experiments in chapter 4 suggest that all four processes have the potential to cool and freshen intermediate waters on isopycnals. The response of the global thermohaline circulation to change ranges from decadal to millennial timescales. Changes in AAIW are an early indication of a response to a changing climate. Interpreting AAIW change is confounded by the fact that all four possible mechanisms for cooling and freshening intermediate waters are consistent with observed changes in forcing over the late twentieth century.
References


Banks, H. T. and N. L. Bindoff, 2003: Comparison of observed temperature and salinity changes in the indo-pacific with results from the coupled climate model HadCM3: Processes and mechanisms. *Journal of Climate*, 16, 156–166.


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# Appendix A

## Abbreviations and notation

### A.1 Abbreviations

<table>
<thead>
<tr>
<th>Abbreviation</th>
<th>Description</th>
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<tbody>
<tr>
<td>AABW</td>
<td>Antarctic Bottom Water</td>
</tr>
<tr>
<td>AAIW</td>
<td>Antarctic Intermediate Water</td>
</tr>
<tr>
<td>ACC</td>
<td>Antarctic Circumpolar Current</td>
</tr>
<tr>
<td>APFZ</td>
<td>Antarctic Polar Front Zone</td>
</tr>
<tr>
<td>CNTRL</td>
<td>Control Experiment</td>
</tr>
<tr>
<td>EAC</td>
<td>East Australian Current</td>
</tr>
<tr>
<td>GFDL</td>
<td>Geophysical Fluid Dynamics Laboratory</td>
</tr>
<tr>
<td>HF\textsubscript{100}</td>
<td>Heat-flux experiment run over 100 years</td>
</tr>
<tr>
<td>ICE\textsubscript{100}</td>
<td>Sea-ice experiment run over 100 years</td>
</tr>
<tr>
<td>ICE\textsubscript{40}</td>
<td>Sea-ice experiment run over 40 years</td>
</tr>
<tr>
<td>IPCC</td>
<td>Intergovernmental Panel on Climate Change</td>
</tr>
<tr>
<td>MOM</td>
<td>Modular Ocean Model</td>
</tr>
<tr>
<td>Abbreviation</td>
<td>Description</td>
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<tr>
<td>--------------</td>
<td>-------------</td>
</tr>
<tr>
<td>MOT</td>
<td>Meridional overturning</td>
</tr>
<tr>
<td>NADW</td>
<td>North Atlantic Deep Water</td>
</tr>
<tr>
<td>PRECIP$_{100}$</td>
<td>Precipitation experiment run over 100 years</td>
</tr>
<tr>
<td>PRECIP$_{40}$</td>
<td>Precipitation experiment run over 40 years</td>
</tr>
<tr>
<td>psu</td>
<td>practical salinity units</td>
</tr>
<tr>
<td>SAM</td>
<td>Southern Annular Mode</td>
</tr>
<tr>
<td>SAMW</td>
<td>Subantarctic Mode Water</td>
</tr>
<tr>
<td>SAT</td>
<td>Surface air temperature</td>
</tr>
<tr>
<td>SST</td>
<td>Sea surface temperature</td>
</tr>
<tr>
<td>SSS</td>
<td>Sea surface salinity</td>
</tr>
<tr>
<td>Sv</td>
<td>Sverdrup ($10^6 \text{m}^3\text{s}^{-1}$)</td>
</tr>
<tr>
<td>UVic model</td>
<td>The University of Victoria Earth System climate model</td>
</tr>
<tr>
<td>WINDS$_{100}$</td>
<td>Winds shift experiment run over 100 years</td>
</tr>
<tr>
<td>WINDS$_{40}$</td>
<td>Winds shift experiment run over 40 years</td>
</tr>
</tbody>
</table>
A.2 Notation

\( \alpha \)  
Thermal expansion coefficient

\( \beta \)  
Haline contraction coefficient

\( C_p \)  
Specific heat

\( h_m \)  
Mixed layer depth

\( Q_{\text{net}} \)  
Net air-sea heat exchange

\( \rho_0 \)  
Density

\( \sigma_{27.2} \)  
1027.2 kg m\(^{-3}\) density surface

\( S \)  
Salinity

\( S'|_n \)  
Salinity anomaly on neutral surfaces

\( S_E \)  
Surface salinity

\( \tau_x \)  
Wind stress in \( x \) direction

\( T \)  
Potential temperature

\( T'|_n \)  
Potential temperature anomaly on neutral surfaces

\( T_{s(\text{anomaly})} \)  
Sea surface temperature anomaly

\( u \)  
Zonal velocity

\( u_e \)  
Ekman velocity \((u_e, v_e)\)

\( u_g \)  
Geostrophic velocity \((u_g, v_g, w_g)\)

\( w_e \)  
Entrainment velocity

\( v \)  
Meridional velocity

\( v_E \)  
Surface meridional velocity, dominated by Ekman flow

\( w \)  
Vertical velocity
Appendix B

Literature Review