Projected Changes to the Southern Hemisphere Ocean and Sea Ice in the IPCC AR4 Climate Models

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ABSTRACT

Fidelity and projected changes in the climate models, used for the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4), are assessed with regard to the Southern Hemisphere extratropical ocean and sea ice systems. While individual models span different physical parameterizations and resolutions, a major component of intermodel variability results from surface wind differences. Projected changes to the surface wind field are also central in modifying future extratropical circulation and internal properties. A robust southward shift of the circumpolar current and subtropical gyres is projected, with a strong spinup of the Atlantic gyre. An associated increase in the core strength of the circumpolar circulation is evident; however, this does not translate into robust increases in Drake Passage transport. While an overarching oceanic warming is projected, the circulation-driven poleward shift of the temperature field explains much of the midlatitude warming pattern. The effect of this shift is less clear for salinity, where, instead, surface freshwater forcing dominates. Surface warming and high-latitude freshwater increases drive intensified stratification, and a shoaling and southward shift of the deep mixed layers. Despite large intermodel differences, there is also a robust weakening in bottom water formation and its northward outflow. At the same time the wind intensification invigorates the upwelling of deep water, transporting warm, salty water southward and upward, with major implications for sequestration and outgassing of CO2. A robust decrease is projected for both the sea ice concentration and the seasonal cycling of ice volume, potentially altering the salt and heat budget at high latitudes.

1. Introduction

The Southern Ocean acts as the engine room to the world’s overturning circulation, drawing its energy from the powerful midlatitude westerlies acting over a zonally continuous ocean. A high-latitude wind-driven divergence draws deep waters to the surface. Subsequently, these upwelled waters are driven both north and south via Ekman fluxes and buoyancy-driven sinking around the Antarctic margin. The northward flow subducts to form the basis of the thermocline, mode, and intermediate water masses that comprise the upper limb of the global overturning circulation. The southward flow, modified by a salt injection from sea ice development and large winter heat losses, forms the densest bottom water masses that ventilate the abyssal ocean. These vertical movements of water help to set the density structure of the Southern Ocean.

Intimately linked to this overturning circulation is the zonally continuous Antarctic Circumpolar Current (ACC), which connects the three ocean basins with a transport of ~135 Sv (1 Sv = 10^6 m^3 s^{-1}) through the Drake Passage (e.g., Cunningham et al. 2003). Surface wind and buoyancy forcing, internal circulation, the underlying bathymetry and the action of mesoscale eddies all play a role in determining its characteristics (Gent et al. 2001; Rintoul et al. 2001; Olbers et al. 2004). At lower latitudes, gradients in the surface wind stress largely determine the dynamics of the subtropical gyres that control the flow pathways that extend from the surface down to intermediate depths. The Southern Ocean also hosts the most expansive volume of homogeneous mode waters, formed through vigorous winter convection. Annual communication between the surface and ocean interior mean that this water mass sequesters large
quantities of heat and gases from the atmosphere. This is complemented by the formation and northward export of Antarctic Intermediate Water (AAIW) into the ocean middepths and Antarctic Bottom Water (AABW) into the abyssal ocean. Observational studies have shown that these water masses play a major role in the removal of additional anthropogenic CO₂ (Sabine et al. 2004) and heat (Gille 2002; Alley et al. 2007), and that, furthermore, the properties of these water masses have been subject to significant trends over recent decades (e.g., Bindoff and McDougall 2000; Bryden et al. 2003; Aoki et al. 2005a,b; Johnson et al. 2007; Rintoul 2007).

The Southern Hemisphere (SH) boasts one of the most profound and robust climate trends observed over the past few decades, characterized by a poleward shift and strengthening of the midlatitude jet (Kushner et al. 2001), and more generally of the midlatitude westerly winds. This is often described in terms of a shift to an increasingly positive phase of the dominant mode of extratropical variability, the Southern Annular Mode (SAM; e.g., Thompson et al. 2000; Thompson and Solomon 2002; Marshall et al. 2004). This trend has been linked to changes in both stratospheric ozone and greenhouse forcing (Gillett and Thompson 2003; Thompson and Solomon 2002). Moreover, the surface signature of this atmospheric rearrangement has the potential to significantly modify characteristics of the ocean and sea ice systems (Hall and Visbeck 2002; Sen Gupta and England 2006). This modification of the general circulation is on the backdrop of, and intimately tied to, unprecedented, large-scale increases in global temperatures and modifications to the hydrological cycle (Alley et al. 2007).

In light of both the observed and projected changes to the surface forcing, there has been a growing focus on the Southern Ocean, driven in part by a recognition of the potential changes to its enormous sequestration capacity. This has profound consequences when considering requirements for the mitigation of future climate change (e.g., Russell et al. 2006a; Toggweiler and Russell 2008; Friedlingstein 2008). Scientific progress has been greatly aided by the availability of model output from the Coupled Model Intercomparison Project (CMIP3), used as part of the Intergovernmental Panel on Climate Change (IPCC) Fourth Assessment Report (AR4), which covers a wide range of control and forced scenarios. Most of the AR4 models show robust changes in the SH midlatitude westerlies over the twentieth century, however, the poleward intensification tends to be underestimated compared to recent observations (Yin 2005). Furthermore, Fyfe and Saenko (2006) show that for a subset of the AR4 models, this projected trend continues under increasing greenhouse forcing.

In model experiments using both an ocean-only model (Oke and England 2004) and an intermediate-complexity climate model (Fyfe et al. 2007), changes in the position and strength of the midlatitude westerlies are shown to induce major changes in Southern Ocean temperature, extending from the surface to intermediate depths, independent of any surface warming. As such, one might expect an associated modification of the circulation into the future. Such shifts were found by Fyfe and Saenko (2006) in the Southern Ocean across a subset of CMIP3 models, with a poleward contraction and strengthening of the zonal (ACC) and meridional (Ekman) flow. While part of the circulation change can be directly attributed to the altered wind field, other factors, including surface warming, an intensified hydrological cycle, and modifications to the Antarctic sea ice, can also cause a reorganization of the internal density field, and thus the internal flow. For instance, Russell et al. (2006b) investigate the factors affecting the fidelity of the Southern Ocean simulations in the CMIP3 models. They find that the most important gauges of realistic circulation are the strength of the westerlies over the Drake Passage latitudes, the heat flux gradient over this region, and the salinity gradient across the ACC through the water column. The authors suggest that it is this latter property—modulated by the upwelling of North Atlantic Deep Water (NADW) to the south of the ACC—that is the most important in affecting intermodel variability in the ACC. There is modeling evidence that the future potential for the Southern Ocean to sequester atmospheric heat and gases in climate simulations is also strongly modulated by the present-day position and strength of the midlatitude westerlies (Russell et al. 2006a). Despite rising temperatures causing increased surface stratification, stronger westerlies and greater high-latitude divergence help to maintain a strong connection between the surface ocean and the interior, mitigating against sequestration decreases in anthropogenic CO₂ resulting from the increased stratification. This result does not, however, make allowance for changes in the natural carbon cycle. Recent modeling (Lovenduski et al. 2007) and observational (Le Quéré et al. 2007) studies suggest that a positive SAM is actually associated with increased CO₂ outgassing as a result of enhanced wind-driven upwelling of old carbon-rich subsurface waters, resulting in a potential positive feedback. This is an issue of ongoing debate (Zickfeld et al. 2008; Le Quéré et al. 2008; Law et al. 2008).

The subtropical gyre circulation is more directly controlled by the curl of the wind stress via Sverdrup dynamics. This circulation extends to considerable depth and controls the flow pathways of surface, waters,
but also the thermocline, mode, and intermediate water masses. Roemmich et al. (2007) find an increase in the observed wind-driven transport of the South Pacific subtropical gyre since 1993, which they link to an intensification of the wind stress curl to the east of New Zealand, possibly associated with the positively trending SAM. Cai et al. (2005) investigate projected changes to the wind stress curl in one of the AR4 models CommonwealtScientific and Industrial Research Organisation Mark version 3.0 (CSIRO Mk3.0) and find an intensification of wind stress curl over the subtropical gyres. This manifests itself as a spinup of the gyre, with a significant increase in the East Australian Current (EAC) transport and an associated enhanced warming in the Tasman Sea. A similar shift in the wind stress curl and subtropical gyres was found by Saenko et al. (2005) for the Canadian Centre for Climate Modelling and Analysis (CCMa) Coupled General Circulation Model, version 3.1 (T47 resolution) [CGCM3.1(T47)].

In contrast to the clear long-term decrease in observed Arctic sea ice cover in the Northern Hemisphere (NH; e.g., Serreze et al. 2007; Alley et al. 2007), SH sea ice exhibits large interannual variations but no consistent large-scale trends (Alley et al. 2007). Only at regional scales is there evidence that annual sea ice duration and monthly sea ice concentration have decreased (increased) in the western Antarctic Peninsula and southern Bellingshausen Sea (western Ross Sea) region since 1979 (Stammerjohn et al. 2008). These changes are postulated to be related to decadal variations in the SAM and the extratropical response to El Niño–Southern Oscillation (ENSO). For the AR4 models, Parkinson et al. (2006) find that their ability to reproduce the observed annual cycle in sea ice cover is better in the NH than the SH. They attribute this to a more realistic representation of the ocean circulation in the NH. Arzel et al. (2006) and Lefebvre and Goosse (2008) investigate sea ice fidelity in the AR4 model simulations and find that there is a reasonable reproduction of the observed Antarctic sea ice trends over the twentieth century by the multimodel mean, especially for the decrease along the Antarctic Peninsula. Again, an associated link with a positively trending SAM is suggested (Lefebvre and Goosse 2008).

In this study we compare large-scale simulated ocean and sea ice characteristics from the last two decades of the CMIP3 twentieth-century control runs with observations. Future projections of these characteristics are also assessed for the Special Report on Emissions Scenarios (SRES) A1B scenario. The SRES A1B scenario contains the largest selection of data available for the present study. By assessing the largest possible set of models, various relationships may be drawn between characteristics of the oceanic forcing (particularly pertaining to differences in wind stress) and oceanic intermodel differences. A primary aim is to provide a self-consistent explanation for the projected changes throughout the extratropical ocean, given the projected changes to surface forcing. In section 2, we briefly describe the CMIP3 models used in our analysis. In section 2a we discuss the need to correct for model drift in order to investigate future projections, particularly in the deeper ocean. In section 3, we present comparisons of simulated SH properties with observations, and we show the projected changes to the atmospheric forcing (section 3a), various oceanic fields (sections 3b–3g), and sea ice (section 3h). Finally, in section 4, we present our conclusions and summary.

2. CMIP3 models

The CMIP3 is an initiative by the World Climate Research Programme to bring together output from an unprecedented array of climate models used as part of the IPCC AR4 [for details on the initiative see Meehl et al. (2005, 2007)]. Output is stored and disseminated by the Program for Climate Model Diagnoses Intercomparison (PCMDI). Outputs from three standard experiments are used in this study for the available models. First, the twentieth-century control run (20C3M) is an integration from 1850 to ~2000 using realistic natural and anthropogenic forcing. This integration is generally initialized from a long preindustrial control run. For the analysis of a given variable, we use all possible realizations available for the last 20 yr of the twentieth century (20C). Sensitivity studies related to the time period employed suggests that 20 yr is sufficient to account for interannual variability, providing a robust mean state. Second, to investigate future changes, we use output from the SRES A1B scenario (Nakicenovic and Swart 2000), which is initialized from the end of the 20C3M integrations. This scenario has increasing greenhouse forcing out to 2100, reaching doubled present-day values; thereafter, concentrations are kept constant. We analyze the last 20 yr of the twenty-first century (21C). Finally, to assess model drift we use 100 yr of the preindustrial control run, corresponding to the twenty-first-century time period, but with constant preindustrial forcing. The variables, time spans, and number of realizations are not necessarily the same across the various models or experiments used. For any given variable we generally present analyses from all available models and realizations.

The CMIP3 repository contains a large set of disparate models (although a few of the component models are implemented in more than one model) encompassing a
broad range of resolutions and incorporating a variety of different physical parameterizations. Table 1 shows model information with a focus on ocean/sea ice components that supplement information presented in Alley et al. (2007). For the ocean components, resolutions vary from the eddy-permitting Model for Interdisciplinary Research on Climate 3.2, high-resolution version [MIROC3.2(hires); 0.28° × 0.19° × 47 levels] to the coarse-resolution Goddard Institute for Space Studies Model E-R (GISS-ER; 5° × 4° × 13 levels). Most models employ a z-level vertical coordinate, although isopycnal, sigma, and various hybrid schemes are also represented. The subgrid-scale mixing parameterization is implemented at different levels of sophistication. Most models implement some form of the Gent and McWilliams (1990; GM) scheme, to account for the mixing of tracers by subgrid-scale eddies. A major improvement over the models used for the third IPCC assessment is that all of the models [except for Meteorological Research Institute Coupled General Circulation Model, version 2.3.2 (MRI CGCM2.3.2), Institute of Numerical Mathematics Coupled Model, version 3.0 (INM-CM3.0), and both CGCM3.1 models] can maintain a realistic present-day climate without the need for flux correction.

Model drift

A model integrated forward in time from some set of initial conditions will tend to “drift” until a quasi-equilibrium state is reached. The equilibration time scale is highly dependent on the model component, with the slowly ventilated deep ocean taking thousands of years to reach steady conditions. Unfortunately, it is not feasible to integrate these computationally expensive climate models for such periods, and the various forced simulations are generally started from control runs that have been integrated out for a few hundred years at best. As a result, some model drift is to be expected, particularly within the ocean. For many of the models, the control simulations, from which the forced scenarios are initialized, are integrated out to provide an overlap with the forced runs. By subtracting out linear trends from the control integration, for the concurrent twenty-first-century period, a first-order correction for the drift is made. This assumes that any low-frequency variability in the ocean is small compared to the drift. Where possible, we apply such corrections to ocean temperature, salinity, barotropic and overturning streamfunctions, and to any subsequent derived variables. Preindustrial datasets are not, however, available for all model–variable combinations. In general, only a single realization of the control integration is available to make drift corrections. However, for the Flexible Global Ocean–Atmosphere–Land System Model gridpoint version 1.0 (FGOALS-g1.0) model, three control realizations are available. The spatial pattern of trends in these realizations are very similar, indicating that, for this model at least, the linear trend is insensitive to initial conditions and the long-term variability is adequately sampled, providing a representative drift.

Figure 1 shows the trend in the zonally averaged temperature across the 100 yr of the preindustrial control run for individual models. A sizeable drift is evident in many of the models, with regional drift sometimes exceeding 0.5°C century⁻¹. To put this into perspective, these regional drifts may exceed both the range of natural variability (not shown) and the magnitude of projected change over the 100 yr of the forced run (Fig. 1, superimposed). This is most evident in the deeper ocean, where ventilation by new water is weak and subsequently any forced change is small. In regions of strong ventilation, on the other hand, forced trends generally exceed the drift. It is clearly of great importance to take account of any model drift when investigating changes in the ocean interior, particularly when analyzing changes at smaller spatial scales or in the deep ocean. Averaging across the models (Fig. 1, bottom panel) shows that the multimodel mean drift is small, indicating that there is no systematic bias in the sign of the drift. Drift in the interior circulation is discussed in section 3f.

3. Results

a. Atmospheric forcing

We begin by assessing the present-day state and twenty-first-century projections for the surface atmospheric fields that force the ocean and sea ice. Figure 2 shows the multimodel and time-averaged 20C means for these variables, the associated standard deviations across the models, the biases between the observations and multimodel mean (20C), and the differences between the 21C and 20C. Multimodel averages are calculated as unweighted means of all available models (unless otherwise stated), where means over multiple realizations for a given model are computed first. Associated zonal means are presented in Fig. 3.

1) Wind stress

Large-scale features of the surface wind fields are well represented in all the models (Figs. 2, 3), with easterlies in the subtropics and strong westerlies at midlatitudes, with maximum strengths in the Indian Ocean basin (although the maximum generally tends to extend too far east) and the most poleward position in the Pacific basin. In the multimodel mean, it is clear that the band

<table>
<thead>
<tr>
<th>Model</th>
<th>Oceanic model</th>
<th>Oceanic resolution</th>
<th>Vertical coordinate</th>
<th>Mixing parameterization</th>
<th>Ice dynamic, rheology</th>
<th>Atmospheric resolution</th>
<th>Flux correction</th>
<th>Forcing</th>
<th>Reference</th>
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<td>KT, KPP</td>
<td>H</td>
<td>T63 L31</td>
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<td>Z</td>
<td>GM, EVD</td>
<td>FH, H, S</td>
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<td>H, F</td>
<td>(-)</td>
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<td>Z</td>
<td>GM, TKE</td>
<td>HD</td>
<td>T63 L45</td>
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<td>Z</td>
<td>GM*, BL, PP</td>
<td>S, FH, OF</td>
<td>T63 L18</td>
<td>N</td>
<td>(-)</td>
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<td>Z</td>
<td>GM*, BL, PP, V, KT</td>
<td>S, FH, OF</td>
<td>T63 L18</td>
<td>N</td>
<td>(-)</td>
<td></td>
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<tr>
<td>GFDL CM2.0</td>
<td>OM3P4</td>
<td>( 1^\circ (1/3)^\circ \times 1^\circ, ) L50</td>
<td>Z</td>
<td>GM*, BL, BBL-BD, KPP</td>
<td>S*, HD</td>
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<td>N</td>
<td>ALL</td>
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<td>GM*, KPP</td>
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<td>E</td>
<td>GM, PP</td>
<td>S</td>
<td>T42 L26</td>
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<td>PP</td>
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<td>HOPE-G</td>
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of westerlies is situated too far equatorward, particularly south of Australasia (cf. \(0.18 \pm 0.03 \, \text{N m}^{-2}\) and \(49.2 \pm 3.1^\circ\text{S}\) for the simulations, and \(0.18 \, \text{N m}^{-2}\) and \(53.2^\circ\text{S}\) for the observations). Only the CSIRO Mark version 3.5 (CSIRO Mk3.5) and ECHAM4 models have poleward biases. Projected wind stress over the twenty-first century shows a robust southward displacement (\(1.0^\circ \pm 0.3^\circ\) in the multimodel mean; see Table 2) and intensification (\(8.0\% \pm 0.1\%)\) of the maximum zonal wind stress band across the models. Using a subset of 12 CMIP3 models, Fyfe and Saenko (2006) found a considerably stronger response in the position and strength of the maximum westerlies (~1.9° southward shift, ~15% strength increase) for the more extreme A2 emissions scenario. Of the 20 models analyzed here, only 2—FGOALS-g1.0 and the Parallel Climate Model (PCM)—exhibit a northward displacement; and only 1—the PCM—simulates a weakening of the westerlies (Fig. 4b). For most models, this poleward intensification is most pronounced over the Pacific (Table 2). The wind shift projects strongly onto the positive phase of the SAM (Thompson and Wallace 2000; Alley et al. 2007). Despite large differences in model resolutions and physics, there are strong correlations between the position and strength of the maximum (20°C) wind stress (Fig. 4a) and between the associated changes in these variables over the twenty-first century (Fig. 4b). We also note that models with more poleward 20°C wind stress maxima tend to have smaller southward shifts into the future (Fig. 4c). This suggests an increasing resistance to successively greater poleward wind stress maximum migration.

2) WIND STRESS CURL

In the open ocean, the depth-integrated meridional transport can be estimated from the curl of the wind stress. The large-scale wind stress curl distribution is captured by individual models, in particular, the circumpolar maximum situated near 40°S (Figs. 2, 3). The largest positive values are seen around 35°S, 11°E, in the southeastern Indian Ocean, consistent with Harrison (1989), and in the southeastern South Atlantic, suggesting strong transport associated with the Agulhas Current. Figure 2f shows small standard deviation in the tropics and subtropics, revealing good agreement among the models where the curl magnitudes are large. The largest intermodel disagreement occurs near 50°S, where the meridional gradient of the wind stress curl is large. In the multimodel mean the maximum curl is stronger than observed and situated too far to the north (Fig. 3b). Like the wind stress, the largest biases occur south of Australia and New Zealand (Fig. 2g). The projected southward intensification of the wind stress is mirrored by the wind stress curl, although the maximum...
change occurs \(\sim 10^\circ\) further north, over the southern parts of the subtropical gyres. In the zonal and multimodel mean there is an increase of \(\sim 8.5\%\) and a poleward shift of \(1^\circ\) in the curl maximum.

3) Precipitation

Large biases in precipitation exist across all the models. While the 20C multimodel mean captures the low precipitation over the eastern side of the subtropical oceans (Fig. 2i), where subsidence occurs, the AR4 models are generally unable to correctly simulate the observed rainfall band associated with the subtropical convergence zones. As a consequence, the zonally averaged precipitation is generally underestimated over the midlatitudes, compared to the Climate Prediction Center (CPC) Merged Analysis of Precipitation.
Fig. 2. (Left to right) Multimodel (20C) mean, standard deviation, difference from the observation, and projected change (21C–20C) of (a)–(d) wind stress (N m$^{-2}$), (e)–(h) wind stress curl (N m$^{-1}$), (i)–(l) precipitation (mm day$^{-1}$), (m)–(p) latent heat flux (W m$^{-2}$), (q)–(t) sensible heat flux (W m$^{-2}$), and (u)–(x) net heat flux (W m$^{-2}$). Positive heat fluxes indicate gain to the ocean. Observations of wind stress and wind stress curl are based on the 40-yr European Centre for Medium-Range Weather Forecasts (ECMWF) Re-Analysis (ERA-40; 1979–2001, see Uppala et al. 2005). Precipitation observations are based on the CMAP gridded dataset (1979–2006, see Xie and Arkin 1997). Heat fluxes are based on the SOC gridded observations (Josey et al. 1999).
(CMAP) climatology (Fig. 3c). Biases in the representation of the midlatitude convergence zones are known to be present in both general circulation models and reanalysis products (e.g., Quartly et al. 2007; Taschetto and Wainer 2008). Along the midlatitude storm track, the simulated precipitation is generally overestimated (Figs. 2i,k). It should be noted, however, that large disagreement also exists between observed rainfall products (Quartly et al. 2007; Gruber et al. 2000).

Precipitation is projected to increase at tropical latitudes associated with increased evaporation and enhanced convection in a warmer world (see changes in latent heat flux). At midlatitudes, between 15° and 40°S, rainfall tends to decrease, particularly over the central to eastern Pacific (Fig. 2i). Farther south, the band of maximum precipitation across the Southern Ocean intensifies and shifts southward, mirroring the poleward intensification of the maximum wind stress. Previous studies have reported a poleward shift in storm-track location and increased storm intensity over the last few decades (e.g., Simmonds and Keay 2000; Simmonds et al. 2003; Yin 2005). These changes in extratropical cyclones have again been associated with the SAM trend (Simmonds and Keay 2000).

4) HEAT FLUX

Turbulent fluxes generally result in heat loss from the ocean. In agreement with the observations, the largest contribution is from latent heat losses in the subtropics (see Fig. 3d), in particular, over the Indian Ocean and at higher latitudes over the warm boundary currents. The latent heat loss steadily decreases south of 15°S with the reduction in SST. Sensible heat losses are generally of considerably smaller magnitude, with largest magnitudes between 20° and 40°S. Minimum heat loss is centered along 50°S (Fig. 3e) because of the enhanced poleward heat transport by the transient eddies in the atmosphere. Intermodel variability for sensible heat is particularly large. For both turbulent latent and sensible fluxes, the largest intermodel differences are associated with discrepancies in the positions of the boundary currents and the ACC (discussed below) and in regions of coastal upwelling. Both terms show consistent low biases across most models compared to the observations (although this is much improved when comparing against the reanalysis, which also uses bulk formulas to estimate fluxes). Observed fluxes are known to be poorly constrained, particularly at higher latitudes, with sizeable differences between products (Taylor 2000).

The projections show an enhancement of the latent heat loss from the ocean, particularly at lower latitudes, indicating higher evaporation in a warmer world (Fig. 2p). There is a weaker intensification in the Southern Ocean consistent with the smaller increases in SST at higher latitudes (see section 3b). The largest regional intensification occurs in the Agulhas Current retroflection, the
Brazil–Malvinas confluence regions, and in the Tasman Sea, again consistent with projected SST changes. While SST is generally projected to rise, air temperatures increase more rapidly. As such, there is a general reduction in the air–sea temperature gradient and an associated reduction in ocean sensible heat losses (Fig. 3e). This effect is accentuated south of 50°S, where increased poleward eddy heat transport resulting from the poleward-intensified wind field contributes to increased air temperatures, and relatively weak SST increases, further reducing the air–sea temperature gradient. As a consequence, only at higher latitudes is there a substantial net heat flux into the ocean (Fig. 3g).

b. Sea surface temperature and salinity

Figure 5a shows the difference between observed and modeled (20C) long-term mean SST for the multimodel ensemble. All observations of temperature and salinity (throughout the water column) are taken from the CSIRO Atlas of Regional Seas 2006 (CARS2006) gridded dataset (Ridgway et al. 2002), derived from all available historical subsurface ocean property measurements for the SH extratropics through to 2006. While the use of profiling floats in particular is rapidly increasing the SH data coverage, high-latitude observations are still sparse, particularly for subsurface properties. For individual models, the zonally averaged SST distribution (Fig. 5g) is well represented with biases, compared to observations, generally less than 1°C difference. However, regional differences for individual models (not shown) can exceed 5°C. While regional biases are, in general, considerably reduced in the multimodel mean, compared to any individual model, certain regions still retain sizeable biases, indicating locations of systematic bias over a number of the models. Strong midlatitude meridional gradients mean that small biases in the position of the ACC translate into large SST biases. In particular, the Malvinas Confluence region is generally simulated too far to the north and removed from the continental margin, resulting in too much cold freshwater in this region, with warm, salty subtropical water adjacent to the continent. A warm bias along the eastern boundaries is associated with the common problem of overly sluggish eastern boundary currents (e.g., Large and Danabasoglu 2006). Close to steep topography (e.g., along the Chilian margin) there are also problems with the representation of upwelling-favorable winds, which lead to biases in low cloud formation and radiative heat transfer Randall et al. (2007). These biases are common to a number of the models, suggesting similar model deficiencies. The pattern of the biases are reflected in the standard deviation across the

Table 2. Position of the maximum zonal wind stress for 20C, zonally averaged over the Southern Ocean (Total) and over each ocean basin and the associated projected differences between 21C and 20C. Bold numbers, for individual models, indicate an equatorward displacement of the maximum wind stress position.

<table>
<thead>
<tr>
<th>Zonal wind stress position</th>
<th>Total</th>
<th>Atlantic sector, 50°W–20°E</th>
<th>Indian sector, 30°–100°E</th>
<th>Pacific sector, 180°E–90°W</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>20C</td>
<td>21C–20C</td>
<td>20C</td>
<td>21C–20C</td>
</tr>
<tr>
<td>Observation (ERA-40)</td>
<td>-53.2</td>
<td>-50.0</td>
<td>-49.9</td>
<td>-57.5</td>
</tr>
<tr>
<td>Multimodel mean</td>
<td>-49.3</td>
<td>-1.2</td>
<td>-48.6</td>
<td>-1.2</td>
</tr>
<tr>
<td>CGCM3.1(T47)</td>
<td>-48.0</td>
<td>-2.4</td>
<td>-47.7</td>
<td>-1.7</td>
</tr>
<tr>
<td>CGCM3.1(T63)</td>
<td>-50.5</td>
<td>-2.1</td>
<td>-49.2</td>
<td>-1.7</td>
</tr>
<tr>
<td>CNRM-CM3</td>
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<td>-0.7</td>
<td>-46.1</td>
<td>-0.9</td>
</tr>
<tr>
<td>CSIRO Mk3.0</td>
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<td>-0.7</td>
<td>-49.7</td>
<td>-1.2</td>
</tr>
<tr>
<td>CSIRO Mk3.5</td>
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<td>-0.5</td>
<td>-52.0</td>
<td>-1.5</td>
</tr>
<tr>
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<td>-1.2</td>
</tr>
<tr>
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<td>-0.9</td>
<td>-50.1</td>
<td>-1.1</td>
</tr>
<tr>
<td>GISS-AOM</td>
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<td>-44.2</td>
<td>-0.6</td>
</tr>
<tr>
<td>FGOALS-g1.0</td>
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<td>-48.3</td>
<td>0</td>
</tr>
<tr>
<td>ECHAM4</td>
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<td>-0.8</td>
<td>-52.2</td>
<td>-1.0</td>
</tr>
<tr>
<td>INM-CM3.0</td>
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<td>-0.5</td>
<td>-44.3</td>
<td>-2.3</td>
</tr>
<tr>
<td>IPSL-CM4</td>
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<td>-3.1</td>
<td>-44.8</td>
<td>-2.3</td>
</tr>
<tr>
<td>MIROC3.2(hires)</td>
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<td>-1.1</td>
<td>-44.3</td>
<td>-2.3</td>
</tr>
<tr>
<td>MIROC3.2(medres)</td>
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<td>-1.8</td>
<td>-46.9</td>
<td>-1.5</td>
</tr>
<tr>
<td>ECHO-G</td>
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<td>-4.2</td>
<td>-47.3</td>
<td>-3.1</td>
</tr>
<tr>
<td>ECHAM5</td>
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<td>-0.8</td>
<td>-49.4</td>
<td>-1.1</td>
</tr>
<tr>
<td>MRI CGCM2.2.3</td>
<td>-50.5</td>
<td>-0.7</td>
<td>-49.5</td>
<td>-1.0</td>
</tr>
<tr>
<td>CSM3</td>
<td>-52.1</td>
<td>-1.0</td>
<td>-50.7</td>
<td>-1.2</td>
</tr>
<tr>
<td>PCMI</td>
<td>-51.4</td>
<td>0.2</td>
<td>-50.8</td>
<td>0.1</td>
</tr>
</tbody>
</table>
models (Fig. 5c), indicating that model SST biases are controlled, to a large extent, by biases in the circulation of the ACC and boundary currents.

Projected changes in SST are robust at a large scale, with all of the models showing significant warming everywhere, except in the vicinity of Antarctica, where some models show areas of weak regional cooling, particularly in the Weddell gyre region (figure not shown). In the zonal mean, the degree of warming generally increases equatorward to about 45°S, north of which the magnitude of warming remains stable. These meridional differences are consistent with changes to the northward midlatitude Ekman transport. Most models show an increase in the northward transport of cold high-latitude water to the south of 45°S, with an anomalous decrease north of this latitude (Fig. 5e). Between 20° and 45°S, individual models generally show a warming of between 1.3° and 2.5°C, except for MIROC3.2(hires), in which temperature increases exceed 3°C (in the zonal mean). Many of the models [e.g., the Bjerken Centre for Climate Research (BCCR) Bergen Climate Model version 2.0 (BCM2.0), CGCM3.1

FIG. 4. Scatterplots for (a) zonally averaged wind stress maximum position (τR) vs zonally averaged wind stress maximum strength (|τR|, N m⁻²), (b) Δ|τR| vs Δ|τd|, (c) τR vs position of ACC core (ACCP; calculated from zonally averaged depth-integrated zonal velocity), (d) |τd| vs ACC core strength (ACCS, Sv), (e) Δ|τd| vs ΔACCS, (f) Δ|τd| vs ΔACCP, (g) |τd| vs position of minimum zonally averaged barotropic streamfunction (STF), (h) Δ(|τd| × τR) vs ΔSTF. Where either ordinate or abscissa values is missing for a given model, the value for the available variable is plotted along the side of the panel. The numbers next to the plotted circles refer to: 1: Obs, 2: CGCM3.1(T47), 3: CGCM3.1(T51), 4: CNRM-CM3, 5: CSIRO Mk3.0, 6: CSIRO Mk3.5, 7: GFDL CM2.0, 8: GFDL CM2.1, 9: Goddard Institute for Space Studies Atmosphere–Ocean Model (GISS-AOM), 10: GISS-ER, 11: FGOALS-g1.0, 12: ECHAM4, 13:INM-CM3.0, 14: IPSL-CM4, 15: MIROC3.2(hires), 16: MIROC3.2(medres), 17: ECHO-G, 18: ECHAM5, 19: MRI CGCM2.3.2, 20: CCSM3.0, 21: PCM, and 22: HadCM3. Models where ozone hole recovery over the twenty-first century is simulated are highlighted with an “X” (based on information from Son et al. 2008); those where information regarding ozone recovery is unavailable are highlighted with a “-”; all other models do not simulate a recovery. Correlation coefficients (r), p values (p), and best-fit linear regression lines are also shown.
T63 resolution; T63), and Community Climate System Model, version 3 (CCSM3) show an enhanced surface warming along the path of the ACC, particularly in the western Atlantic. This becomes a common feature of most of the models when subsurface warming is examined (see section 3e). Some of the regional structure of surface warming can be associated with the distribution of the deep mixed layers. Regions associated with deep mixing generally show weaker surface warming. This is particularly evident along the southeastward band of deep mixed layers stretching across the Indian Ocean with an associated band of relatively weak warming in the ensemble mean (Fig. 5e, superimposed contours). The strongest warming is evident in the southwestern Pacific, and in particular in the Tasman Sea. Cai et al. (2005) note that this is likely due to increased transport in the EAC as part of a spin up of the subtropical gyre.

In general, sea surface salinity (SSS) is more poorly represented than SST (Fig. 5b). Even in the zonal mean, salinity biases (compared to CARS06) can exceed ±0.5 psu (Fig. 5h). In the tropical regions, where precipitation rates far exceed those in the extratropics, salinity biases are readily explained by discrepancies in surface freshwater fluxes (e.g., Large and Danabasoglu 2006). However, in the mid- and high latitudes the biases are likely to also reflect discrepancies resulting in the upper-ocean circulation. The multimodel mean shows that in general SSS is too fresh at midlatitudes, although
some models do have a strong salty bias [particularly ECHAM4, L’Institut Pierre-Simon Laplace Coupled Model, version 4 (IPSL-CM4), and GISS-ER, not shown]. Also evident in many of the models are extended regions of high salinity along parts of the South American margin, again associated with a poor representation of local atmospheric processes and upwelling. At higher latitudes, surface waters tend to be too saline, particularly at some regions around the Antarctic margin. Unlike SST, the largest intermodel variability occurs at lower latitudes (Fig. 5d) and likely results from intermodel differences in surface freshwater fluxes. The very large intermodel variability is clearly evident in the zonally averaged profiles (Fig. 5h).

Compared to the intermodel differences and model biases, the projected changes in SSS are relatively modest. The most robust feature is a freshening of surface waters around Antarctica. This is consistent with the large-scale increase in precipitation over the Southern Ocean and enhanced runoff from Antarctic. These models do not simulate changes to land ice cover, and as such a large potential source of additional freshwater runoff is likely absent. Salinity changes resulting from altered sea ice conditions are discussed in section 3h.

c. Mixed layer depth

The SH midlatitude oceans are home to some of the deepest, most expansive mixed layers in the world, formed through vigorous winter convection. These deep mixed layers are fundamental to the formation of Subantarctic Mode Water (SAMW) in the Pacific and Indian Oceans and of AAIW in the southeastern Pacific, and in setting thermocline properties. These water masses are also associated with the sequestration and storage of large quantities of anthropogenic CO$_2$ (e.g., Sabine et al. 2004). As noted above, the mixed layer depth (MLD) distribution also affects the simulated pattern of projected surface warming. A realistic representation of MLD is therefore a vital component of any coupled climate model.

To compare model and observed depths we estimate the maximum annual mixed layer depth, calculated as the depth where the buoyancy difference between the near surface and the mixed layer base exceeds 0.003 m s$^{-2}$. While many of the models actually diagnose MLD via a variety of schemes (see, e.g., Griffies et al. 2009), this diagnostic is unavailable as part of the PCMDI repository and the use of a common definition throughout means that we are comparing like with like. Deep observed MLD stretches in a band from the central Indian Ocean southeastward toward the Drake Passage, with weaker enhanced convection in the southwest Atlantic (Fig. 6, last panel). The deepest midlatitude mixed layers are evident south of Australia and New Zealand and in the eastern Pacific. Very deep mixing also occurs in the Weddell Sea and at other locations around the Atlantic margin, although little in situ data are available to constrain the observational products used in the estimation of “observed” MLD.

Most models simulate the enhanced midlatitude mixed layers forming a southeastward band across the Indian and Pacific regions. The GISS-ER shows the greatest midlatitude MLD bias across all basins with a basin-wide zonally averaged depth that is too great by a factor of 3. On the other hand, FGOALS-g1.0 has only a weakly enhanced midlatitude maximum MLD, reaching to only $\sim$250 m, where in the observations depths exceed 700 m in places. Figure 7 shows the surface area within the midlatitude domain that has MLD within (20 m) binned ranges for the 20C and 21C, with the observed distribution superimposed for reference. The right-hand tails of the distributions represent the important deepest mixed layers associated with water mass formation. Many of the models [particularly BCCR BCM2.0, CSIRO Mk3.5, Geophysical Fluid Dynamics Laboratory Climate Model version 2.0 (GFDL CM2.0), FGOALS-g1.0, MIROC3.2(hires), MRI CGCM2.3.2, and the Met Office’s third climate configuration of the Met Office Unified Model (HadCM3)] lack the deepest mixed layers (>400 m) that are evident in the observations; although this is often compensated by more substantial areas of medium-depth (200–400 m) MLD. On the shallow end of the distribution, nearly all models lack a substantial area of very shallow (<40 m) mixed layers. It should be stressed that the lack of observational constraints could affect both ends of the observed distribution. Despite large regional intermodel differences, integrated over the full midlatitude domain, the volume of water contained within the winter MLD envelope varies by less than 30% from that observed ($\sim$1.4 x 10$^{16}$ m$^3$) across all of the models, except for CSIRO Mk3.0 and GISS-ER, whose volumes are too large by $\sim$50% and 150%, respectively.

A comparison of the surface area of deep mixed layers (>200 m) in the Pacific versus the Indian basin (where MLD reach considerable depths) across the models reveal a significant correlation ($r = 0.8$ using all models), that is, a model with a deep bias in the Pacific will also tend to have a deep bias in the Indian basin. This suggests that a large proportion of the model MLD bias is unrelated to local forcing, and instead results from different mixing parameterizations or large-scale forcing biases. In an intercomparison of ocean–ice GCMs forced using a common Coordinated Ocean-Ice Reference Experiments (CORE) climatology, Griffies et al. (2009) suggest that a large component of the intermodel
variability might be explained by differences in the treatment of the neutral physics scheme (Gent and McWilliams 1990) rather than different vertical coordinate systems. The standard deviation across the models indicate that the largest intermodel variability is associated with the depth of the deepest mixed layers in the eastern Pacific, but is more related to the position of these layers in the east Indian basin. Large variability also exists around the Antarctic margin, as some models exhibit very deep MLD, while others have essentially no enhanced deep mixing.

Figures 6 and 7 also show the projected changes for the twenty-first century. In general, there is a consistent shoaling of the deepest midlatitude mixed layers. Similarly, around the Antarctic there is an extensive shoaling of deep mixed layers. Considering the midlatitude domain only (see Fig. 7), the total volume contained within the MLD is projected to decrease for all the models, ranging from a drop of −5% for GFDL CM2.0 and MRI CGCM2.3.2 to over 20% for the Centre National de Recherches Météorologiques Coupled Global Climate Model, version 3 (CNRM-CM3), ECHAM and the global Hamburg Ocean Primitive Equation (ECHO-G). The anomalous projected midlatitude shoaling is situated slightly south of the maximum MLD, in the multimodel mean, indicating a small poleward shift in position in addition to the shoaling.

d. Horizontal circulation

A major component of the oceanic circulation is driven by the wind. As a result, the robust shifts in Southern Hemisphere winds, both observed and simulated, are likely to drive major changes to the ocean circulation. Modified heat and moisture fluxes associated with enhanced greenhouse forcing will also modify the circulation via buoyancy changes.

The barotropic streamfunction represents the depth-integrated ocean circulation. Figure 8 shows the multimodel mean barotropic streamfunction and observed streamfunction derived from the University of Maryland’s Simple Ocean Data Assimilation Reanalysis (SODA) version 2.0.2-3. The main features are well captured in the multimodel mean and across individual models (not shown). The simulated ACC has the
characteristic southeastward pathway from the western Atlantic through to the eastern Pacific, with regional meandering as a result of topographic steering (particularly over the Macquarie Ridge system and the Kerguelen Plateau), and a contraction in the flow through the Drake Passage followed by a sharp northerly excursion. In general, the Weddell gyre is realistically situated, albeit slightly weaker than observed; however, the Ross gyre is very weak and confined too far to the south in the ensemble mean. The simulated subtropical gyre in the Indian Ocean basin is generally situated too far to the north, consistent with the equatorward bias in the wind stress curl, and the western boundary currents are overly sluggish, a common feature of coarse-resolution models. For the most part, the individual models capture the general lateral circulation features described for the multimodel mean, that is, western intensification in the subtropical gyre and a strong circumpolar circulation, although, there are significant regional intermodel differences.

In a selection of the AR4 models, Fyfe and Saenko (2006) found that associated with the poleward shift in the

Fig. 7. Midlatitude mixed layer depth frequency distributions for domain 30°–60°S (except between 160° and 300°E, where 30°–65°S is used), showing the area within the domain with MLD of a given depth (in bins of 20 m, units: m²). Observations (gray lines) and models (black bars). Twenty-first-century decrease (blue) and increase (red) in the area associated with a given MLD range.
midlatitude westerlies, over the twentieth and twenty-first centuries, there is a consistent poleward intensification of both the ACC and high-latitude meridional Ekman flow. They fit the depth-integrated zonal velocity and the zonal wind stress to a Gaussian distribution at each longitude to obtain an estimate of the latitude and magnitude of the maximum flow. While this method works well for the wind stress (see Fyfe and Saenko 2006, Fig. 1), we find that the structure of the ACC is less amenable to this approach because the distribution is often heavily skewed or has multiple cores (commonly associated with sub-Antarctic and polar fronts in observations). We implement an alternative approach whereby at each longitude, an average of latitude weighted by the depth-integrated zonal flow strength is calculated to obtain an estimate of the average core position (using only those latitudes where the zonal flow exceeds 50% of the maximum eastward flow magnitude). Core strength is then simply the mean depth-integrated velocity over these values. While this strength estimate does not correspond to the maximum core strength (as it includes everything to 50% of the maximum), it provides a representative magnitude for the ACC core.

Figure 9 shows the position of the ACC core for the different models, with the SODA reanalysis ACC core superimposed. Two of the models, CSIRO Mk3.0 and GISS-ER, have the largest overall biases, with the ACC core over 5° too far south (in the zonal mean), but with regional differences exceeding 10°. Conversely, the IPSL-CM4 and MIROC3.2(hires) have ACC cores in excess of 3° too far to the north. The multimodel mean of the ACC position (Fig. 10), however, closely tracks the observed pathway across most of its domain. The largest bias is evident in the western Atlantic, where the ACC for the majority of models tracks too far to the north. This bias was also manifest in the SST (Fig. 5, section 3b), where a large, cold bias occurs as high-latitude water is advected too far northward.

Figure 9 also shows the projected change in the position of the ACC core over the twenty-first century. Our findings, which comprise an expanded set of 15 models, are essentially in agreement with the 8 models analyzed by Fyfe and Saenko (2006). The majority of models show an overall poleward contraction of the ACC core, although a number of models have regions where the core moves equatorward. Four models show a net equatorward movement (GISS-ER, CSIRO Mk3.0, FGOALS-g1.0, and CNRM-CM3, although the change is small in the first two). The FGOALS-g1.0 change is associated with the only net northward wind shift across the models (Fig. 4f) and the CNRM-CM3 change is related to a large northward movement in the mid-Pacific, where the core is weak and poorly defined.

The model with the largest southward shift in the maximum westerlies (ECHO-G) also shows the greatest overall (zonally averaged) ACC shift (almost 3° in the zonal mean). Across the models there is a strong link between the magnitude of the wind stress shift and the ACC shift, with a correlation of $r = 0.7$ ($p = 0.01$; see Fig. 4). A regression analysis indicates that the shift in the position of the ACC is approximately half that of the wind stress shift. This may relate to the fact that the ACC flow is somewhat constrained by contours of $f/H$ ($H =$ bottom depth), although the strength of flow along lines of constant $f/H$ may change, leading to a meridional shift in the ACC core. The multimodel average shows a poleward shift at all longitudes, with a zonally averaged shift of $-0.6° \pm 0.4°$. 
In addition to the poleward shift in the ACC, Fyfe and Saenko (2006) found a consistent strengthening of the maximum depth-integrated flow at all longitudes. While we also find an increase in the strength of the ACC over the majority of the models, there is a high degree of regional and intermodel variability. CSIRO Mk3.0 is again an outlier and shows a zonally averaged weakening of the ACC core strength with a reduction of over 5%. All of the other models show an increased core magnitude with many models having projected increases exceeding 10% (Fig. 4g). The change in core strength does not necessarily correspond to a change in ACC transport, however, as the total flow width may also contract or expand. An important and well-measured metric for ACC flow is the volume transport passing through the Drake Passage. Table 3 shows the 20C and 21C mean volume transports. Common estimates of Drake Passage transport are ~135 Sv (e.g., Cunningham

![Figure 9](image-url)
et al. 2003). A model’s ability to capture this transport varies considerably, with present-day estimates ranging from less than 50 Sv for IPSL-CM4 to in excess of 300 Sv for CSIRO Mk3.0. Projected changes in the transport are also highly variable with a range of almost a 30% increase for INM-CM3.0 to almost a 30% decrease for CNRM-CM3. There is essentially no projected change in the ACC core position passing through the Drake Passage for any of the models.

Cai et al. (2005) find that for the CSIRO Mk3.0 there is a projected spinup of the midlatitude gyre circulation into the future, particularly along the path of the EAC, as the SAM trends toward an increasingly positive state. Figure 9 shows selected streamfunction contours representative of the position of the midlatitude gyres for the 20C and 21C periods. Zonally averaged streamfunction and wind stress curl are also shown for the individual models. As expected, a significant link exists between the position of the wind stress curl minimum and the cores of the subtropical gyres across the models (Fig. 4). A southward shift occurs across all the models over the twenty-first century. A strong correlation exists across the models between the magnitude of the shift in the zonally averaged position of the wind stress curl core and the associated change in the position of the zonally averaged streamfunction core (Fig. 4; $r = 0.8$, $p = 0.01$). As with the shift in the position of the ACC, a regression analysis indicates that the shift in the wind stress curl position is approximately twice the magnitude of shift in the position of the gyre cores.

To provide a measure of the gyre strength, Fig. 11 tracks the strength (calculated as the component of the depth-integrated core strength traveling along the coast) of the western boundary current core for the three subtropical gyres. The projected changes have quite different characteristics within the different basins. Along the EAC there is often a strong increase in flow at higher latitudes with a compensating decrease at lower latitudes, indicative of a southward shift in the circulation. There is no robust increase in the maximum flow however, although, as noted by Cai et al. (2005), CSIRO Mk3.0 does show an enhanced flow into the Tasman Sea region. Along the Brazil Current, on the other hand, as
well as a southward shift, there is a robust increase in the strongest flow, indicative of a spinup of the gyre. In the Agulhas region changes are less robust, with some models showing weak projected reductions in the higher-latitude flow.

e. Heat content

Figures 9 and 10 also show the change in the depth-integrated heat content for individual models and the ensemble mean, respectively. The pattern of heating appears to be strongly modulated by changes in the lateral circulation. For most models, the largest heating occurs at midlatitudes, broadly constrained by the ACC to the south and the eastward-flowing limbs of the subtropical gyres to the north. This is clearly seen in the multimodel mean, where both the ACC core and the subtropical gyres show obvious shifts to the south.

Were the change in heat content mainly attributable to altered surface heat fluxes, the midlatitude region might be expected to be an area of enhanced warming because the deep overturning would mix additional surface heat into the interior. However, the largest heat content changes occur in the western Atlantic and Indian Ocean regions, where the MLDs are generally shallower than in the eastern Indian and Pacific regions. Furthermore, the primary input of additional surface heating occurs to the south of 50° (Fig. 3f). However, the western Atlantic and Indian Ocean regions are generally associated with the strongest meridional temperature gradients (figure not shown). Thus, any poleward
shift in the circulation would preferentially enhance heat content change at these locations.

f. Vertical circulation

The extratropical large-scale Eulerian meridional overturning circulation (MOC) is characterized by the following three main cells: the Antarctic overturning cell, which is related to the sinking of AABW; the abyssal cell, which extends north of approximately 40°S, representing the northward flow of AABW and the compensating southward flow of lower Circumpolar Deep Water (CDW) and lower NADW; and the Deacon cell, which represents the upwelling of upper CDW and northward Ekman transport. The Deacon cell overturning should be interpreted with care because by construction a part of the cell strength is an artifact of zonal averaging in a region where isopycnal depths change with longitude (Döös and Webb 1994). The MOC diagnostic is not readily calculable in the observations because it requires a full three-dimensional knowledge of the ocean meridional velocity fields.

Figure 12 shows the 20C mean state MOC overlaid on the projected twenty-first-century changes. The three-cell configuration is clear in the multimodel mean, but is also present in all the models except for the INM-CM3.0 where, for example, a negative subsurface cell exists at the characteristic position of the Deacon cell. In IPSL-CM4 both the Deacon and Antarctic cells are situated in a much more northward position than the other models. This bias is associated with the largest northward zonal wind stress (and ACC) position (Fig. 4) and manifests itself as the most extreme midlatitude cold bias (figure not shown). Dynamical reasons would suggest a relationship between wind stress position–strength and overturning position–strength, via changes in the strength and position of Ekman pumping, and indeed significant correlations exist across the models. The standard deviation in MOC across the models (Fig. 12) indicates that large intermodel variability is associated not only with the strength of the Deacon circulation, but also with the magnitude and position of the deep water exchange. By comparing the mean and standard deviation of individual cell strengths (Fig. 13), we also find significant intermodel differences in the weaker cells. In particular, the magnitude of the standard deviation and the mean of the Antarctic cells are of comparable magnitude. These differences are partly associated with the strength of high-latitude density stratification. A significant correlation \( r \sim -0.7 \) exists between the mean vertical density gradient over the top 100 m, south of 60°S, and the AABW overturning across the models (excluding INM-CM3.0). This implies that models with weak stratification, which also tend to have low vertical resolution, tend to exhibit strong AABW overturning.

The sign of the projected changes for the three overturning cells are, however, robust across almost all of the models. In all but CSIRO Mk3.5, there is a spinup of the Deacon cell (Fig. 13). The behavior of this outlier is related to the equatorward shift of the zonal wind stress in the Pacific sector (Table 2). The projected multimodel mean increase of the Deacon cell is \( \sim -4 \) Sv (Fig. 12), indicating an intensified upwelling of relatively warm and saline upper CDW. Significant correlations exist between the change in both the position and the strength of the maximum zonally averaged wind stress and the corresponding changes in the position and strength of the Deacon cell across the models \( (r > 0.75, p < 0.007 \) for position, and \( r > 0.8, p < 0.001 \) for strength, excluding INM-CM3.0). Despite substantial differences in resolution, bathymetry, and physical parameterizations, intermodel differences in the midlatitude MOC is primarily determined by biases in wind field. Previous studies have shown a strong short-term response in the Deacon cell driven by changes in the SAM (Hall and Visbeck 2002; Sen Gupta and England 2006). For a positive SAM this cell spins up and shifts poleward. The long-term trend toward a more positive SAM phase seen in the models would therefore be consistent with these projected changes.

In the multimodel mean, the AABW cell weakens by \( \sim -2 \) Sv (Fig. 13). This spindown is consistent with the projected surface-intensified buoyancy gain around the Antarctic margin. A comparable reduction in the abyssal cell strength reflects the weaker AABW formation and its subsequent outflow. A weak correlation exists between changes to the abyssal and AABW cell strengths. As an aside, Fig. 13 also shows the overturning calculated without any correction for drift. It is evident that for some models drift explains an important proportion of the raw MOC change, particularly in the abyssal cell. Figure 12 also shows the standard deviation in the projected overturning changes. Large differences in the projected response are associated with the Deacon cell and the inflow of deep water. There are also large surface-intensified intermodel differences along 40°S, presumably associated with wind-driven differences in the convergence zone.

g. Internal properties

Figure 14 shows the projected changes to the zonally averaged potential temperature. A general warming pervades the extratropics with enhanced surface warming particularly between 40° and 60°S that extends to middepth; and for almost all models there is an enhanced abyssal warming near the Antarctic margin. The
FIG. 12. Modeled projected change in MOC (21C – 21C, color maps). Only differences that are significant above the 95% level are displayed. The significance level is determined by a two-tailed $t$ test. The 20C mean state of MOC is superimposed.

(bottom left) Multimodel mean of 20C MOC [solid contours for positive MOC (4-Sv interval), dashed contours for negative (1-Sv interval)] and projected changes (color map). (bottom right) Multimodel standard deviation for both the 20C mean states (contours with 1-Sv interval) and projected changes (color map).
weakest warming, and in some models a weak cooling, occurs at middepth north of 40°S.

It was suggested in section 3d that a large component of the change in heat content can be attributed to the poleward shift of the midlatitude circulation. Superimposed on Fig. 14 are contours indicating the change in temperature that would arise from a 1°C southward translation of the 20°C temperature distribution. At the midlatitudes, as far north as 20°S, and primarily for the upper 1000 m, there is a very close correspondence between the projected warming and the warming that would result from a pure southward shift. This technique was used by Alory et al. (2007) to show that projected warming in the Indian Ocean could be explained, in part, by a simple southward translation of the temperature field. A recent analysis of upper Southern Ocean warming by Gille (2008) over the latter part of the twenty-first century also supports the idea that the warming trend is intimately tied to a southward shift in the ACC.

The corresponding changes in salinity across the models (Fig. 15) are more varied. However, certain general features are evident. There is a freshening signal that tracks the upper part of the AAIW salinity minimum and at lower latitudes in the thermocline. Furthermore, most of the models show a saline anomaly in the surface layer at lower latitudes. At locations associated with upwelled CDW, there is generally a poleward-intensified salination (except in the CNRM-CM3 and CSIRO Mk3.0), which in some models [e.g., CGCM models and MIROC3.2(hires)] feeds into the lower layers of AAIW. In most models abutting this region of salination, both at middepths and north of 40°S there is a freshening signal that coincides with the region of cooling (or relatively weak warming) in Fig. 14. This
signal is most prominent in the Atlantic basin (Fig. 16), indicating that these changes are associated with a modification in the southward export of NADW. In other basins, changes in this region are weak, consistent with areas of old, poorly ventilated waters (e.g., Pacific Deep Water) where recent surface changes would not have penetrated.

Unlike temperature, which possesses a strong monotonic meridional gradient at all longitudes, zonally averaged salinity changes are less clearly attributable to a southward shift in the wind-driven circulation. Other factors are evidently more important in part because salinity is much more spatially heterogeneous than temperature. Only in the upper ocean near 40°S, where there are strong meridional salinity gradients, is there an obvious contribution by this shift. This pattern originates in the Atlantic region (Fig. 16) where, as noted, there is a large poleward shift and a spinup of the subtropical gyre. The projected salination and warming signal seen in most models south of 40°S reflect changes in CDW. While changes in CDW could occur through property changes in NADW or via changes in the rate of diapycnal mixing with neighboring AAIW/AABW (Santoso et al. 2006), such mechanisms would not result in changes consistent with those projected. Instead, the dominant mechanism appears to be an increased southward and upward transport of freshwater and heat, within the CDW, associated with the spinup of the Deacon circulation.

North of 40°S the freshening of the deep water is consistent with the slowdown of the North Atlantic overturning cell that transports warm and salty water southward. This signal appears to be decoupled from changes in the Southern Ocean, south of the region of NADW separation. At shallower depths enhanced advection of warm and salty water into the Southern Ocean occurs primarily in the Atlantic basin via the poleward-intensified Brazil Current (Fig. 16) across the Brazil–Malvinas confluence where strong T–S gradients mean that the poleward
intensification of the circulation generates substantial $T$–$S$ changes. This warm, salty signal connects into the upper CDW. This is consistent with a modeling study by Santoso et al. (2006), which showed that a significant portion of upper CDW $T$–$S$ variability originates within the Brazil–Malvinas confluence via meridional advection. The anomalous signal then propagates into the Southern Ocean CDW via the ACC.

As noted in section 3f, in polar regions AABW overturning is generally projected to weaken. This is consistent with an increase in buoyancy in the surface regions (see below) as a result of surface warming, increased surface freshwater flux, and changes to the seasonal cycle of sea ice (see section 3h). Combined with a generally negative vertical gradient in the mean state $T$ and $S$ at polar latitudes (i.e., colder and fresher conditions near the surface), this slow down of the overturning translates into a warm, saline signal close to the Antarctic margin. These modifications reinforce the changes in $T$–$S$ noted for the CDW. Changes in upper-ocean temperature and salinity result in robust large-scale reductions in potential density (Fig. 17). These changes are most pronounced toward the surface, resulting in a large-scale increase in the stratification of the Southern Ocean. This is consistent with the findings of Lefebvre and Goosse (2009, manuscript submitted to Climate Dyn.), which show a robust increase in the density gradient in the upper ocean for the CMIP3 models. At midlatitudes, as well as the poleward shift in the steep isopycnals, there is also an increase in the meridional density gradient, consistent with the simulated acceleration of the zonal flow, through the thermal wind balance and a strengthening of the frontal gradients (Wainer et al. 2004). Figure 17 also indicates the regions where the density changes are primarily controlled by salinity. This occurs primarily around Antarctica, where there is an increased surface freshwater input. Salinity also plays an important role at middepth in the subtropics, driven by changes in the subduction of freshwater into upper intermediate water.

FIG. 15. Same as Fig. 14, but for salinity. Salinity contour representative of the salinity minimum (yellow contour), associated with the pathway of intermediate water. Line contour interval is 0.01 psu with an absolute maximum contour value of 0.1 psu.
h. Sea ice

For completeness we briefly discuss the fidelity and projections associated with sea ice. All models except for FGOALS-g1.0 have a reasonable representation of the seasonal cycle in the sea ice concentration (SIC) surrounding the Antarctica sea ice. This outlier has a highly unrealistic seasonal cycle (with double the integrated SIC in wintertime) resulting from high levels of numerical filtering applied at high latitudes (see http://www-pcmdi.llnl.gov/ipcc/model_documentation/more_info_iap_fgoals.pdf) and is excluded from the multimodel analysis. Simulated maximum SIC generally occurs between August and October and the minimum SIC extent is between January and March, in line with observations. During wintertime, all models except CGCM3.1(T63) (which strongly overestimates SIC) are within 30% of the observed total concentration. During summer, a number of models have very low ice concentrations, in particular, CNRM-CM3, MIROC3.2(hires), MIROC3.2(medres), and IPSL-CM4 have between 6% and 30% of the observed amount. Holland and Raphael (2006) look at the representation of sea ice in six of the AR4 models and find little systematic bias in regional SIC across the models. This is borne out to some extent by the larger ensemble set evaluated here. The spatial patterns of the multimodel SIC are in reasonable agreement with observations in both seasons (Fig. 18).

The sea ice trends into the future are robust across all simulations. All of the models show a loss of sea ice cover over the twenty-first century in both summer and winter (Table 4). The reduction in the integrated SIC ranges from 10% [ECHO-G and CGCM3.1(T47)] to almost 50% (CNRM-CM3 and BCCR BCM2.0) during winter. For the majority of the models, larger fractional losses in summer ranges from 33% in CSIRO Mk3.0 to almost complete loss of sea ice cover in the MIROC3.2(hires). Using their smaller subset of AR4 models, Holland and Raphael (2006) found a possible link between the
zonally averaged meridional sea level pressure (SLP) gradient at 60°S for April–June and sea ice extent. In our larger sample (15 models), this relationship does not appear to hold. In addition, no significant correlation was found between ice extent and the position of the SLP minimum, or between ice area and zonal wind stress at 60°S.

Table 4 also shows the changes in integrated sea ice volume (SIV). Again, there is a robust decrease in volume across the models during both seasons. For the majority of models the fractional decrease in volume is larger than the fractional change in integrated SIC, indicating that the average ice thickness is also reducing. The multimodel mean (Fig. 18) indicates that for the winter season, substantial sea ice loss occurs at all longitudes, with the greatest loss occurring in the western Antarctic, particularly on the outer edge of the Weddell region (0°–30°W) and between 90° and 150°W. In summer, the greatest ice loss is to the east of the Antarctic Peninsula. Arzel et al. (2006) suggests an increase in the seasonal cycle of sea ice over the twenty-first century because the percentage decrease in both integrated SIC and SIV is greater in summer than winter. However, the much larger quantity of ice present in the winter means that in real terms there is actually a robust reduction in the seasonal change in integrated SIC (−23% in the ensemble mean) and SIV (−30%) across the models. A reduction in winter ice formation would reduce both the wintertime brine rejection adjacent to Antarctica, and also the subsequent formation of freshwater during the melt season, which is subsequently advected northward. The net result would be a relative freshening close to the margin (less brine rejection) and more saline conditions north of the ice edge (from reduced freshwater export).

4. Summary and conclusions

Output from an ensemble of state-of-the-art coupled climate models used as part of the IPCC AR4 process are compared with recent observations and are investigated for projected changes over the twenty-first century. Our goal is to provide an overarching description of large-scale changes to the SH extratropical ocean in light of changes projected to occur in the atmosphere. Unlike in the tropics, changes to these higher-latitude oceans can be considered largely atmospherically forced phenomena (at least to first order), rather than having to consider highly coupled air–sea interactions.

There is large variability across the model set, as a result of different model configurations (e.g., resolution and bathymetry) and different treatment of the underlying physics (Table 1). Nevertheless, except in certain regions biases are often not systematic, and a simple, unweighted multimodel mean provides a realistic representation of the observed system. Moreover, future projections tend to be qualitatively robust across most models. It is apparent that a large component of individual model bias, associated with ocean dynamics and
thermodynamics, results from biases in the overlying wind field. Further, it is changes to the wind field that can explain many of the modifications projected for the ocean. Over the twenty-first century, both the wind stress maximum (at \( \sim 50^\circ S \)) and the wind stress curl (at \( \sim 40^\circ S \)) are projected to intensify and shift southward. At the same time, there is an intensification of the hydrological cycle with increased precipitation in the tropical and mid- to high-latitude regions and reduced precipitation in the subtropics. At the midlatitudes this is related to a poleward intensification of the midlatitude precipitation linked to a shift in the Southern Hemisphere storm track. These wind-induced modifications are consistent with an increasingly positive SAM. The pattern of heat flux change is more complex. Overall, there is a shift in the balance of turbulent heat fluxes from the ocean to the atmosphere: a reduction in sensible heat loss is counteracted by an increase in latent heat loss, although regional variations abound. In general, there is an increased radiative flux into the ocean, which is most pronounced at high latitudes where there is also a reduction in summer albedo. Most of the additional heat entering the SH oceans does so to the south of 50°.

Figure 19 shows a schematic diagram of the more robust projected changes across the models. Associated
with the southward intensification of the westerlies there is a southward shift in the position of the ACC core. As the ACC position is strongly affected by bathymetry, its poleward shift is, in general, only about half as large as the shift in the position of the wind stress. Sen Gupta and England (2006) showed modest changes in the barotropic flow of the ACC resulting from changes in surface slope associated with high-frequency SAM variability. At these short time scales they found little change in flow associated with modifications to the ocean density structure. For the long-term changes discussed here, an increase in the meridional density gradient is evident, which is consistent with an intensified ACC, driven in part by changes in surface Ekman pumping and an associated spinup of the Deacon circulation. At lower latitudes the maximum wind stress curl also moves southward and intensifies, driving a southward shift in the position of the subtropical gyres. Only in the Atlantic basin, where the gyre core has a more southerly mean position, is there evidence of a robust increase in the depth-integrated flow of the western boundary current. The overall southward shift of the lateral circulation manifests itself as an increase in the heat content of the water column, primarily between the southern limb of the gyres and the core of the ACC. This can be understood as a simple poleward displacement in the temperature distribution in a region where the meridional temperature gradients are large. The largest increase in heat content occurs in the western Atlantic, where, in addition to the circulation shift, there is increased transport of warm water in the boundary flow.

Around the Antarctic margin enhanced heating from above and below results in a melt back of the summer and winter sea ice. As noted, this feeds back to the radiative balance through a decrease in the surface albedo. The seasonal cycling of sea ice acts as a factory for the production of freshwater during the summer melt, which is advected northward in the Ekman flow, and saline water during wintertime ice formation and brine rejection, which is exported to depth. A reduction in the output of the factory by the reduced projected seasonal cycle in ice volume, will therefore result in relatively fresher conditions at high latitudes and saltier conditions north of the ice edge. The latter process appears to be completely offset by increased high-latitude precipitation over the ocean and increased runoff from water precipitated onto Antarctica. The surface freshening and surface-intensified warming is consistent with a weakening of AABW formation, which in turn is manifest as a warming signal close to the Antarctic margin that reaches abyssal depths. Possible changes to the Antarctic ice sheet are not accounted for in the IPCC AR4 models.

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Note: The table shows the total Antarctic sea ice area (× 10^14 m^2) and volume (× 10^14 m^3) for summer [January–March (JFM)] and winter [August–October (ASO)] seasons for 20C and 21C periods. Last column shows the percentage change over the twenty-first century in the ASO – JFM difference in ice volume (i.e., a measure of the change in the seasonal cycle).
While the poleward displacement of the circulation can account for much of the temperature distribution change, it explains less of the salinity changes. Instead, surface changes are closely tied to the projected intensification of the hydrological cycle and of Ekman transport. This is consistent with the idea that inter-model variability in SST is primarily driven by circulation differences, while SSS variability is more related to differences in freshwater fluxes (section 3b). Only in the Atlantic is there a clear “circulation shift” signal in the salinity as the Brazil–Malvinas confluence, which acts as a barrier between salty subtropical and fresh Southern Ocean water, moves southward. In the interior, south of \(-40^\circ S\), there is a subsurface increase in salinity that is consistent with an intensified southward and upward movement of relatively warm and saline CDW, associated with the more rapid Deacon cell overturning. This is further enhanced by an injection of warm and saline upper CDW entering from the Atlantic sector as a result of the southward-intensified Brazil Current coupled with the large meridional $T$-$S$ gradients in the Brazil–Malvinas confluence zone. At lower latitudes, primarily in the Atlantic basin, there is a freshening signal at depth, consistent with the projected slowdown of the relatively salty NADW circulation. Throughout the extratropics there is an increase in stratification driven by a surface-intensified density reduction. At high latitudes, in the upper ocean, this is primarily a result of salinity changes, while at lower latitudes it is primarily driven by temperature changes. This results in a large-scale shoaling of the deep mixed layers at midlatitudes.

While this paper provides a consistent picture of future changes, our analysis is essentially qualitative. A more quantitative accounting of the various changes must follow. For instance, by looking at changes along density surfaces (e.g., Bindoff and McDougall 1994), the contributions of heat and freshwater fluxes and pure displacements of isopycnals via “heave” effects can be...
disentangled. This will be the subject of future work. We also note two caveats in interpreting projections from the AR4 models. First, the dynamics of the ocean circulation still lacks certain physical processes. For example, subgrid-scale eddies could be critical to realistically capturing future oceanic changes. Recent modeling and observational studies (e.g., Meredith and Hogg 2006) suggest that a proportion of the additional energy entering the ocean via a momentum flux would energize the eddy field rather than enhance the circumpolar circulation. Second, recent evidence based on models with more realistic depictions of the stratosphere (Son et al. 2008) suggests that the poleward intensification of the wind field may halt, or even reverse, at least during austral summer as a result of the recovery of stratospheric ozone over Antarctica. Only a few of the AR4 models incorporate any form of ozone recovery (see Fig. 4), and these tend to underestimate the resulting modification in the atmospheric circulation compared to models with more realistic stratospheres (Son et al. 2008). Thus, a definitive assessment of the changes in certain oceanic properties may be beyond the scope of the AR4 models. Nevertheless, it is encouraging to find that despite sometimes large intermodel differences, systematic biases are generally small and projected changes are for the most part qualitatively robust.

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