Ocean circulation response to anthropogenic-aerosol and greenhouse gas forcing in the CSIRO-Mk3.6 coupled climate model

Mark A. Collier¹, Leon D. Rotstayn¹, Kwang-Yul Kim², Anthony C. Hirst¹ and Stephen J. Jeffrey³

¹Centre for Australian Weather and Climate Research, CSIRO Marine and Atmospheric Research, Aspendale, Australia
²School of Earth and Environmental Sciences, Seoul National University, Seoul, South Korea
³Department of Science, Information Technology, Innovation and the Arts, Dutton Park, Australia

(Manuscript received July 2012; revised March 2013)

We use the CSIRO-Mk3.6 coupled climate model to examine the impact of anthropogenic aerosols (AAs) and long-lived greenhouse gases (GHGs) on aspects of the global ocean circulation. Focusing on the second half of the twentieth century, we compare multiple ten member ensembles of historical climate change, which are forced by different combinations of forcing agents; these different simulations enable us to separately diagnose the effects of changes in AAs and GHGs. We also compare ten member 21st century ensembles driven by Representative Concentration Pathways 4.5 (RCP4.5) and 8.5 (RCP8.5).

To a large degree the pattern of change in the oceans due to the impact of AAs is similar to the effect of increasing GHGs, but of opposite sign. The Atlantic Meridional Overturning Circulation (and associated North Atlantic Deep Water formation) strengthens in response to historical changes in AAs but weakens in response to increasing GHGs. Similarly, the Indonesian Throughflow strengthens in response to AAs, and weakens in response to increasing GHGs. The Drake Passage Transport, however, shows a small weakening (strengthening) due to historical changes in AAs (GHGs) in the ensemble mean. The change in the Drake Passage Transport is much clearer in the 21st century, in which it increases strongly in response to increasing GHGs and decreasing AAs in both RCP4.5 and RCP8.5.

The results suggest that without the influence of AAs, changes in ocean circulation would have already followed a path much more like one dominated by increasing GHGs. Considering that the AA levels are expected to decrease during the next few decades, the effects of increasing GHGs on ocean circulation will be amplified accordingly.

Introduction

Anthropogenic aerosols (AAs) exert a net negative radiative forcing at the top of the atmosphere. This forcing shows strong spatial variations, e.g. it is stronger in the northern hemisphere than the southern hemisphere (Rotstayn et al. 2012, their Fig. 3). The implications of this forcing have been studied with atmospheric General Circulation Models (GCMs) for over a decade. In an atmospheric GCM with a slab ocean model, which includes no ocean dynamics, the hemispheric imbalance in radiative forcing induces large changes in tropical circulation (Rotstayn et al. 2000, Williams et al. 2001, Ming and Ramaswamy 2011). These can be summarised as a southward shift of the intertropical convergence zone, a weakening of the Hadley circulation in the northern hemisphere and a strengthening of the Hadley circulation in the southern hemisphere.

Less understood are the implications of AA forcing on circulation in coupled ocean–atmosphere GCMs. The response in these models tends to be much more complex, since both atmospheric and ocean dynamics can respond to the forcing. Rotstayn et al. (2012) reviewed a number of studies of the response of coupled GCMs to transient forcing from AAs, though most of these studies focused primarily on the atmospheric component.
Regarding the ocean, some attention has been paid to the effects of AA forcing on the Atlantic Meridional Overturning Circulation (MOC). It is known that most GCMs simulate a weakening of the Atlantic MOC in response to increasing greenhouse gases due to warming and (to a lesser extent) freshening of the upper ocean in the North Atlantic (Gregory et al. 2005). Delworth and Dixon (2006) found that the direct radiative effects of AAs strengthened the Atlantic MOC in their coupled GCM. They argued that aerosols have exerted a ‘protective’ effect against the expected greenhouse gas-induced weakening of the Atlantic MOC. It is physically plausible that AA forcing might have this effect, since AA forcing is strongly negative in the vicinity of the North Atlantic. Since Delworth and Dixon (2006) considered direct aerosol effects only, it is also plausible that the impact would be stronger in a model that includes indirect effects.

Several other recent studies have explored the role of AAs in driving climate change or multi-decadal variability in the Atlantic basin region. Chang et al. (2011) found that sulfate aerosol changes contributed substantially to simulated 20th century changes in the tropical Atlantic interhemispheric gradient in sea surface temperature (SST) in a multi-model ensemble. They also concluded that changes in sulfate may have significantly altered the tropical Atlantic rainfall climate, supporting the findings of earlier studies that used simpler models (Rotstayn and Lohmann 2002). Booth et al. (2012) found that, in their GCM, aerosol emissions and periods of volcanic activity explained 76 per cent of the simulated multi-decadal variance in de-trended 1960–2005 North Atlantic SSTs. Villarini and Vecchi (2012) argued that projected changes in North Atlantic tropical storm activity during the 21st century are substantially modulated by radiative forcing from AAs. They also called for coordinated experiments to isolate changes in different forcing agents in climate projections, in order to improve our understanding of these effects.

Cai et al. (2006, 2007) took a more global perspective regarding aerosols and ocean circulation. They examined AA-induced effects on ocean circulation in historical transient simulations with a low resolution version of the CSIRO GCM (Mk3A; Rotstayn et al. 2007). Cai et al. (2006) found that aerosols induce a strengthening of the Atlantic MOC and increase of northward cross-equatorial heat transport in the Atlantic and Pacific Oceans, with the majority of change taking place in the Atlantic Ocean. Cai et al. (2007) argued further that AAs strengthen the global ocean conveyor (Gordon 1986), and thereby improve the simulated historical trends in the temperature structure of the southern Indian Ocean. These studies suggest that AAs have the potential to substantially affect ocean circulation in the southern hemisphere, as well as the northern hemisphere.

It is of interest to further quantify the extent to which AAs could be modulating the global ocean circulation. This is especially so, in view of the projected decrease in AAs over the 21st century, which would lead to a rapid reduction in any such effects (Lamarque et al. 2011).

In this paper we explore the impact of AAs and long-lived greenhouse gases (subsequently referred to as GHGs for brevity) on ocean circulation in the CSIRO-Mk3.6 climate system model (Rotstayn et al. 2012). We use simulations carried out for the Coupled Model Intercomparison Project Phase 5 (CMIP5); see Jeffrey et al. (2013) for more details of the experimental design. We use single-forcing historical simulations, described in the next section, to distinguish the effects of AAs and GHGs. The AA-induced effects are compared with those obtained using a low-resolution (spectral R21) model by Cai et al. (2006); to our knowledge, this is the only previous study that used single-forcing simulations to assess the effects of direct and indirect aerosol forcing on the Atlantic MOC. We also compare the simulated historical changes with 21st century changes when the model is forced by two Representative Concentration Pathways (RCPs).

**Model and experiments**

The CSIRO-Mk3.6 (hereafter Mk36) atmospheric model has 18 vertical levels and a horizontal resolution of approximately 1.9 degrees latitude and longitude (spectral T63). The dynamic/thermodynamic sea-ice model is imbedded in the atmospheric model component (O’Farrell 1996). The ocean component is the NOAA/GFDL MOM2.2 ocean model (Pacanowski 1996). It has 31 vertical levels and horizontal resolution that specifies twice as many latitudes when compared to the atmospheric component. Mk36 was developed from the earlier CSIRO-Mk3.5 version (Mk35; Gordon et al. 2002, 2010). The main differences between Mk35 and Mk36 are the inclusion of an interactive aerosol treatment and an updated radiation and atmospheric boundary layer scheme in Mk36. The AA forcing in Mk36 (the difference in top-of-atmosphere radiation between simulations with year 2000 and year 1850 aerosol emissions) is –1.41 +/- 0.09 W m$^{-2}$ (see Rotstayn et al. 2012) for further details about the model and aerosol forcing.

A large number of CMIP5 experiments has been performed with Mk36, including six ‘individually’ forced attribution experiments covering the 1850–2012 historical period; each of these has ten ensemble members (Jeffrey et al. 2013). For the work conducted in this paper a subset of experiments has been used, as summarised in Table 1. The AA-induced effects in a warming climate are calculated from the difference of ALL minus NoAA. The GHG experiment is the same as the all forcing historical experiment except that it is forced by long-lived GHGs only (GHG-only for brevity). We extended all the historical runs from 2006 to 2010 following the RCP4.5 forcing scenario.

Note that the AA-induced effect is likely to be non-linear, in the sense that ALL minus NoAA is not identical to an experiment forced only by anthropogenic aerosols. This has been shown in studies that used atmospheric GCMs coupled to slab ocean models (Feichter et al. 2004, Ming and Ramaswamy 2009). Diagnosing the AA-induced effect
by removing anthropogenic aerosols from an all forcings simulation may be more realistic than using an experiment forced only by changes in anthropogenic aerosols, since aerosols in the real world act against a background of a warming climate. We plan to investigate these complex feedbacks in a separate study.

### Results

#### (a) Changes in global-mean temperature

Figure 1 shows that, when all climate forcings (GHGs, AAs, ozone, solar, volcanic) are included in the historical experiment (ALL; black line), we get the best match for the surface air temperature trend when compared to the observations (OBS; red line) of Brohan et al. (2006). There is some indication of an underestimate of global-mean warming by Mk36 towards the end of the historical simulation. This was not evident in the earlier low-resolution simulations with Mk3A (Rotstayn et al. 2007); a likely explanation is the stronger AA forcing in Mk36 (−1.4 W m⁻² compared to −1.1 W m⁻² in Mk3A), discussed further below. The NoAA (green line) and GHG (blue line) experiments overestimate the warming, especially from about 1950 for NoAA and a decade or so later for the GHG experiment. The NoAA experiment result has much greater inter-decadal variability than that of the GHG experiment, likely due to the effect of the extra forcings in NoAA (i.e. ozone, solar and volcanic forcing). This variability carries over to the ALL experiment, although (as expected) the ten member ensemble means show less variability than observed.

#### (b) Changes in the Atlantic Overturning Circulation

North Atlantic Deep Water (NADW) is formed by the sinking of cold and salty water in the North Atlantic Ocean, forming a crucial arm of the global ocean circulation. One of the primary functions of the Atlantic Meridional Overturning Circulation (AMOC) is to transport warm upper water from the tropical South Atlantic into the North Atlantic. Here we show the relative impact of AAs and GHGs on the basin wide (depth versus latitude) mass transports in the Atlantic Ocean. In Fig. 2 the MOC linear trend (1951–2000) is shown for the AA-induced (upper left), GHG-only (upper right) and ALL forcing (lower left) cases. The 500-year annual MOC climatology from experiment PICNTRL (Table 1) is included so that the trends can be put in the perspective of changes compared to the pre-industrial period. The climatological MOC pattern (Fig. 2(d)) is similar to the results from other contemporary models (e.g. Danabasoglu et al. 2012, Delworth et al. 2012). The main point to be seen from the trends in Fig. 2 is that the Atlantic MOC intensifies in response to AAs, and that the response to GHG forcing is of opposite sign but smaller magnitude. The AA-induced impact on the Atlantic is a strengthening of the overall meridional overturning cell (maximum trend of about 8 × 10⁹ kg s⁻¹ century⁻¹ located at 1500 m). The GHG-only result is generally of opposite sign, but the changes are more concentrated in the northern hemisphere and the absolute value of the central trend, approximately 4 × 10⁹ kg s⁻¹, is much smaller.

The GHG weakening of the Atlantic MOC is broadly consistent with the results of other models, though there is substantial model-dependence of the response (Gregory et...
The magnitude of the AA-induced increase in the Mk36 Atlantic MOC is three times greater than that modelled in the Mk3A study (Cai et al. 2006). This is a striking result, given that the overall AA forcing at the top of the atmosphere is only moderately stronger in Mk36 (−1.4 W m\(^{-2}\); Rotstayn et al. 2012) than in Mk3A (−1.1 W m\(^{-2}\); Rotstayn et al. 2007). However, further examination shows that the increase in heat loss from the ocean surface of the northern hemisphere in Mk36 is about three times greater than in Mk3A during the period 1951–2000. The net AA forcing in Mk36 is substantially stronger over oceans than over land (Rotstayn et al. 2012; their Fig. 2), whereas it is the other way around in Mk3A. These differences are summarised in Table 2. The difference between the total AA forcing at the top of the atmosphere over the oceans of the northern hemisphere (−2.1 W m\(^{-2}\); for

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Mk36 versus −1.3 W m\(^{-2}\) for Mk3A is substantial. The bulk (−0.6 W m\(^{-2}\)) of this difference can be attributed to indirect aerosol effects, since the direct AA forcing over the oceans of the northern hemisphere is −0.36 (−0.54) W m\(^{-2}\) in Mk36 (Mk3A). The relatively weaker indirect aerosol effect over oceans in Mk3A is at least partly related to a ‘tuning’ that was applied to reduce the magnitude of the effect (over oceans only) in that model; this was discussed by Rotstayn et al. (2007; their section 2.1).

A further contribution to the difference in the surface...
heat flux changes between the two models may be due to black carbon, which strongly absorbs solar radiation in the atmosphere. The global-mean loading of black carbon in Mk3A (0.28 mg m$^{-2}$) was thought to be an underestimate, whereas Mk36 has a larger, and likely more realistic value of 0.57 mg m$^{-2}$ (Rotstayn et al. 2007, 2012). In Mk36 the anthropogenic black carbon direct forcing at the ocean surface is $-0.90$ W m$^{-2}$ in the northern hemisphere. Although we don’t have this information for Mk3A, assuming a linear relation between black carbon loading and surface forcing implies an additional surface heat flux difference of order $-0.45$ W m$^{-2}$ between the two models in the northern hemisphere.

In addition to changes in surface heat flux, changes in the Atlantic MOC in the AA-induced case could be driven by changes in the hydrological cycle (Delworth and Dixon 2006). A rigorous assessment of the relative importance of heat and water fluxes would require partial coupling experiments, in the manner of Gregory et al. (2005), who concluded that changes in heat fluxes are more important than changes in water fluxes. Here we present a qualitative assessment, with a focus on heat-related quantities in Fig. 3, and water-related quantities in Fig. 4.

SST increases north of about 45°N (Fig. 3(a)), and the net heat flux trend into the ocean (Fig. 3(b)) is negative; this suggests that the warming is due to the stronger MOC, and the heat flux trend is in part a response to the warmer SSTs. The spatial pattern of the net heat flux trend strongly resembles that of the latent heat flux (Fig. 3(c)), and in some areas it is strongly augmented by the sensible heat flux trend (Fig. 3(d)). In the absence of feedbacks, the effect of AAs on the surface heat flux is expected to occur via shortwave radiation, so it is interesting that the net short-wave radiation trend is relatively weak, and is of the opposite sign (i.e. a warming tendency at the surface, rather than a cooling). In the eastern part of the North Atlantic, this occurs due to a decrease of cloud cover, whereas the short-wave radiation changes off the coast of Greenland and in the Arctic are due to decreasing sea-ice fraction (not shown). Trends in net surface long-wave radiation (Fig. 3(f)) are also relatively small and negative; these tend to oppose the shortwave effects, due to decreasing cloud over the eastern part of the basin, and increasing net long-wave emission from a warming surface the coast of Greenland.

Sea surface salinity (SSS) also increases north of about 45°N (Fig. 4(a)), suggesting that salinity changes also contribute to the stronger AA-induced Atlantic MOC. Fig. 4(b) indicates that evaporation changes (consistent with increased SST in Fig. 3(a)) contribute to the trend of increasing SSS, whereas precipitation changes (Fig. 4(c)) are relatively small. Although aerosol forcing is broadly expected to decrease precipitation, this may be offset in the North Atlantic region by warmer SSTs induced by the strengthening MOC. Thus atmosphere–ocean feedbacks cause the AA-induced effect on the hydrological cycle to be expressed regionally as increased evaporation, rather than decreased precipitation. Evaporation may also exert a positive feedback that further strengthens the MOC, by increasing the salinity of surface waters. When the effects of ice- and snow-melt and continental runoff are included in the surface water budget (Fig. 4(d)), the net water flux trend shows marked differences from the evaporation trend near the coast of Greenland. However, the salinity trend is positive throughout most of the mid- and high-latitude North Atlantic. This may be because the MOC brings salty water northwards into the higher latitude North Atlantic, where a large fraction sinks and forms deep water (e.g. Visbeck 2007) and the remainder gradually gets freshened by net (downward) water flux and input of fresher Arctic waters (Dickson et al. 2002, Talley et al. 2003). Thus a spin-up of the MOC would tend to increase SSS in that region, and a strong enough spin-up could plausibly cause an increase in SSS even in locations where the net water flux to the ocean increases.

It is interesting that an initially negative radiative perturbation leads to an SST increase in parts of the North Atlantic. To first order, this can be understood as the inverse of the ‘warming hole’ that appears over the North Atlantic due to a slowdown of the AMOC in CMIP5 models in response to global warming (Drijfhout et al. 2012). When comparing future projections under RCP4.5 with other CMIP5 models (not shown), we find that Mk36 simulates a marked SST cooling in the sub-polar North Atlantic, whereas most (but not all) other models merely simulate weaker warming; this suggests that the AMOC response to anthropogenic forcing in Mk36 may be stronger than that in the majority of CMIP5 models. This is also indicated by Weaver et al. (2012, their Fig. 2), in which the AMOC in Mk36 weakens quite sharply in the 21st century. Thus, the more vigorous response of the AMOC in Mk36 compared to Mk3A may be substantially due to a large sensitivity of the AMOC, as well as stronger AA forcing at the ocean surface.

It is also intriguing that the largest contributions to the net heat flux trend shown in Fig. 3(b) are from latent and sensible heat fluxes, rather than the short-wave radiative flux (which is more closely related to the initial AA-induced perturbation). In fact, the trend in the AA-induced net heat flux is much larger than the radiative forcing due to AAs, which is calculated without coupled ocean-atmosphere feedbacks (e.g. compare Fig. 2 of Rotstayn et al. 2012). This suggests that the strong AA-induced spin-up of the AMOC in our model may be partly driven by a positive feedback, involving changes in heat and/or water fluxes.

(c) Key indices of the global ocean circulation

In Fig. 5 the impacts of changes in atmospheric forcing on three indices of the ocean circulation are presented. These indices provide the strength of water mass transports at three locations, and have been measured by various direct and indirect means over recent decades due to their crucial and hypothesised roles in maintaining global climate. Note that to compare the results here with the earlier discussion of AA-induced change, we must consider the difference
Fig. 3. 1951–2000 ensemble-mean trends in aerosol-induced (a) sea surface temperature, and surface fluxes of (b) net heat, (c) latent heat, (d) sensible heat, (e) net short-wave radiation, and (f) net long-wave radiation. The mean 1980–2005 field from the historical ALL experiment is contoured, contour range and interval is provided. Units for the trends are provided in each panel; the mean field is in the same units, without the time dependence (century)$^{-1}$. The fluxes (panels b–f) are positive downward.
modeled NADW is approximately \(21.5 \times 10^9\) kg s\(^{-1}\), compared to the (not maximum) observational (at 24°N) value of \(15.75 \times 10^9\) kg s\(^{-1}\) (\(\pm 1.6 \times 10^9\) kg s\(^{-1}\)) from Ganachaud and Wunsch (2000) or (at 26.5°N) \(18.7 \times 10^9\) kg s\(^{-1}\) (\(\pm 5.6 \times 10^9\) kg s\(^{-1}\)) from Cunningham et al. (2007). The NADW, commensurate with numerous other studies (Schmittner et al. 2005, Delworth and Dixon 2006, Cai et al. 2006), decreases in the NoAA and GHG historical experiments during the 20th century, and by 2005 the NoAA experiment index strength has decreased fractionally more than the GHG-only one. The difference between ALL and NoAA shows that AAs act to strengthen the NADW, and, as shown in Table 3, this enhancing effect is stronger than the weakening caused by GHGs.

In Fig. 5(a) it is also noteworthy that in ALL, NADW increases in the period 1960 to 1980 when AAs increase strongly in the Atlantic sector (Smith et al. 2011), and

Fig. 4. As Fig. 3, but for trends in (a) sea surface salinity, (b) net water flux into the surface, (c) evaporation, and (d) precipitation. The net water flux includes contributions from sea-ice melt and continental runoff in addition to precipitation minus evaporation.
Variability of the ITF is modulated by a complex interplay between local and remote oceanic and atmospheric forcing from both the Pacific and the Indian Oceans, including the Madden-Julian Oscillation, Kelvin and Rossby waves, the Asian-Australian monsoon, and ENSO (Susanto et al. 2012). The response of the ITF to multi-decadal increases in GHGs (or other anthropogenic forcing) has had relatively little attention, although a majority of models from CMIP3 do project a weakening of the ITF in the 21st century (Sen Gupta et al. 2012). On an interannual and inter-decadal basis, the ITF tends to weaken in response to weaker easterly winds in the equatorial western Pacific (e.g. Meyers 1996). There is evidence for such a weakening in the GHG experiment (Fig. 6(a)). The effect is stronger in RCP4.5, and even more so in RCP8.5 (not shown); this is consistent with an eastward shift of the main equatorial convection centre under increasing GHGs in Mk36, from the Indonesian region towards the central-western Pacific (Rotstayn et al. 2012, their Fig. 16). However, the inverse of this pattern is not seen in the AA-induced case (Fig. 6(b)), even though the model simulates an AA-induced strengthening of the ITF. This suggests that the mechanism may be more complex, e.g. the AA-induced ITF changes may be part of a pan-oceanic adjustment as hypothesised by Cai et al. (2006).

An interesting feature of Fig. 6(b) is that the AA-induced then reaches a plateau. After 1980, there were substantial decreases in sulfate aerosols, due to emission controls on sulfur dioxide motivated by concerns about acid rain (Stern 2006).

Beyond the end of the historical ALL experiment, when GHGs become the dominant forcing, the NADW strength weakens dramatically. This likely reflects the impact of both increasing GHGs and decreasing aerosols in RCP4.5 and RCP8.5.

Figure 5(b) shows the evolution of the strength of the Indonesian Throughflow (ITF), a current of water climatologically moving from east to west but observed to have significant interannual variability (Sprintall et al. 2009). The model’s simplified land–sea mask in the Indonesian seas region allows transports to be calculated across a single passage between northwestern Australia and the eastern tip of East Java. The ITF is then a summation of individual horizontal transports for all depths. The climatological strength of the modelled ITF is about 17.5 × 10^9 kg s^{-1}, comparable to the observed estimate of 15 × 10^9 kg s^{-1} (Sprintall et al. 2009). Both the NoAA and GHG experiments suggest a weakening of the ITF (i.e. a decreasing westward transport of water mass) by a similar amount by the end of the historical ALL experiment. The AA-induced increase is slightly smaller than the decrease induced by GHGs. The impacts of GHGs beyond 2005 are evident in the model, with an even stronger weakening in RCP8.5 compared to RCP4.5.

Variability of the ITF is modulated by a complex interplay between local and remote oceanic and atmospheric forcing from both the Pacific and the Indian Oceans, including the Madden-Julian Oscillation, Kelvin and Rossby waves, the Asian-Australian monsoon, and ENSO (Susanto et al. 2012). The response of the ITF to multi-decadal increases in GHGs (or other anthropogenic forcing) has had relatively little attention, although a majority of models from CMIP3 do project a weakening of the ITF in the 21st century (Sen Gupta et al. 2012). On an interannual and inter-decadal basis, the ITF tends to weaken in response to weaker easterly winds in the equatorial western Pacific (e.g. Meyers 1996). There is evidence for such a weakening in the GHG experiment (Fig. 6(a)). The effect is stronger in RCP4.5, and even more so in RCP8.5 (not shown); this is consistent with an eastward shift of the main equatorial convection centre under increasing GHGs in Mk36, from the Indonesian region towards the central-western Pacific (Rotstayn et al. 2012, their Fig. 16). However, the inverse of this pattern is not seen in the AA-induced case (Fig. 6(b)), even though the model simulates an AA-induced strengthening of the ITF. This suggests that the mechanism may be more complex, e.g. the AA-induced ITF changes may be part of a pan-oceanic adjustment as hypothesised by Cai et al. (2006).

An interesting feature of Fig. 6(b) is that the AA-induced...
trends in zonal wind stress, averaged across the equatorial Pacific, resemble a weakening of the Walker circulation; this is contrary to the hypothesis from Ming and Ramaswamy (2011), who argued that AAs should strengthen the Pacific Walker circulation, by offsetting the expected GHG-induced weakening. However, the prevailing view that the Walker circulation weakens in response to increasing GHGs has recently been challenged by Meng et al. (2012); these authors argue that increasing GHGs tend to drive an enhanced equatorial SST gradient between the Maritime Continent and the eastern Pacific, and this in turn strengthens the Walker circulation. The possible role of AAs in this puzzle is an intriguing topic for further research.

Figure 5(c) presents a measure of the strength of the ACC at its narrowest width, commonly referred to as the Drake Passage Transport (DPT). The calculation for DPT is essentially the same as for the ITF; the summation in this case is performed on a north–south section between the southern tip of South America and the Antarctic continent. The DPT and other similarly formed indices along the path of the ACC provide a quantitative measure of the strength of the transfer of water, heat and salt and other properties between the three major basins. The climatological strength of the modelled DPT is about $110 \times 10^9$ kg s$^{-1}$, weaker than the observed estimate of approximately $134 \times 10^9$ kg s$^{-1}$ from Cunningham et al. (2003), and a more recent estimate of $154 \times 10^9$ kg s$^{-1}$ ± $38 \times 10^9$ kg s$^{-1}$ by Firing et al. (2011) suggests that an even larger value is possible. There is a strong increase of the DPT in the RCP experiments, even for the moderate RCP4.5 case in which atmospheric emissions are stabilised by 2100.

An examination of zonal wind stress trends (Fig. 6) shows an increase (decrease) in the surface wind stress at the latitude of the ACC in the GHG (AA-induced) experiment. Note that Cai and Cowan (2007) suggested that AAs induce a positive trend in the Southern Annular Mode, and (by implication) stronger sub-polar westerly winds. This suggests that the result may be model dependent, and further work is needed to understand the underlying mechanisms. These sub-polar westerlies are an important driver of the ACC and DPT (Meredith et al. 2004). A strengthening and poleward shift of the mid-latitude storm track is a robust response of GCMs forced by increasing GHGs (Yin et al. 2005). Although the underlying dynamical processes are still an active area of research, a key part of the mechanism under increasing GHGs is an increase in the meridional temperature gradient (Arblaster et al. 2011). This is a plausible explanation for the opposing responses of the DPT in the GHG- and AA-induced cases, because AAs tend to mitigate the enhanced tropical tropospheric warming caused by increasing GHGs (Rotstayn et al. 2013). Note that Sen Gupta et al. (2009) found that, even though the 21st century projections show an
associated increase in the core strength of the ACC, this does not translate into robust increases in the DPT across models. Wang et al. (2011) studied CMIP3 models which exhibited a small ensemble mean weakening much smaller than the inter-model standard deviation. Thus, the DPT change in our simulations is likely to be model-dependent; see also the discussion below regarding the effects of drift.

There is a clear separation of the DPT values (when compared to the ITF and to some degree NADW) for the NoAA and GHG experiments by the end of the 20th century and for the few years that follow, with the former experiment having a slightly stronger DPT by 2005. This likely reflects the impact of ozone depletion in the NoAA experiment, which also contributes to stronger sub-polar westerlies in GCMs (Oke and England 2004).

Note that the transports in Fig. 5 do not take into account drift in the PICNTRL experiment. The linear trends in the three indices, taken over the 500-year PICNTRL experiment, are 0.03, 0.05 and $0.85 \times 10^6$ kg s$^{-1}$ per century for NADW, ITF and DPT, respectively. Using the GHG experiment as an example, corresponding forced trends during 1950–2000 are $-1.92$, $0.84$ and $1.90 \times 10^6$ kg s$^{-1}$ per century. Thus the drift in NADW and ITF is small compared to the forced trend in GHG, whereas the drift in DPT is almost half the forced trend. The forced trends are much stronger in the 21st century; e.g. the DPT trend in RCP4.5 is $11.4 \times 10^6$ kg s$^{-1}$ per century during 2006–2100. Also, it should be noted that when AA-induced effects are calculated from ALL minus NoAA, the effects of drift are removed.

Figure 5 demonstrates that the transport indices throughout the 20th century respond in an opposite way to GHGs and AAs in Mk3.6. The magnitudes of these opposing effects are shown in Table 3. For the NADW, the simulated aerosol-induced strengthening is larger than the GHG-induced weakening. For the ITF, AAs and GHGs have effects of similar magnitude. The nominal effects of GHGs and AAs on the DPT are of opposite sign to their effects on the other two indices, in the sense that GHGs appear to induce a strengthening, whereas AAs induce a weakening. However, the percentage changes in the historical period are small. The signal-to-noise ratio is quite low for the DPT in the historical period: the ensemble range of the DPT shown in Table 3 due to AA or GHG forcing changes sign, unlike the other transport indices presented here. Further, DPT trends in GHG do not account for the effects of drift. The change in the DPT is much stronger in the 21st century; however, as noted above, this is model-dependent (Sen Gupta et al. 2009).

In Fig. 5 evolutions of all transport indices in the ALL experiment are shown to have a smaller slope throughout the second half of the 20th century than their counterparts in the NoAA and GHG experiments. This suggests that, without the overall cooling effect of AAs, the impact on the global ocean circulation by GHGs would be more evident. Further, for the RCP runs which feature strong increase of GHGs and decrease of AAs in the 21st century, the simulated trends of the three transport indices are even stronger than the NoAA or GHG experiments. This suggests that the response of ocean circulation to increasing GHGs will be enhanced by the ‘unmasking’ effect of decreasing aerosols.

Conclusions

In this paper we have examined aspects of the large-scale oceanic response in the CSIRO-Mk3.6 coupled climate model to both anthropogenic aerosol and greenhouse gas forcing. Historical anthropogenic aerosol-induced effects (against a background of global warming) were diagnosed from the difference of an experiment with ‘all forcings’ and another in which aerosol forcing was fixed at the 1850 level. Historical greenhouse gas-induced effects were diagnosed from an experiment in which greenhouse gases were allowed to vary as observed, with all other forcings fixed at the 1850 level. We also compared two projections for the 21st century (RCP 4.5 and RCP 8.5).

We found that historical changes in anthropogenic aerosols induce a stronger Atlantic Meridional Overturning Circulation and the associated North Atlantic Deep Water formation, opposing the effects of increasing greenhouse gases. We found anthropogenic aerosols lead to a warming of the SSTs in the North Atlantic even when the net heat flux there is negative, suggesting additional northward heat transport by a more active NADW overturning cell. The magnitude of the aerosol-induced effect is stronger than that of greenhouse gases, and also stronger than that found in a previous study based on a low-resolution GCM (Cai et al. 2006, 2007). We tentatively attributed the stronger response in this model to stronger aerosol forcing at the ocean surface, and a vigorous response of the Atlantic Meridional Overturning Circulation to perturbed forcing in Mk3.6. A more rigorous comparison is deferred to a later study.

Two other important indices describing regional features of the global ocean circulation display a consistent response to anthropogenic aerosols, in the sense that the anthropogenic aerosol-induced change is of opposite sign to the greenhouse gas-induced change. For each of the Indonesian Throughflow and Drake Passage Transport, we identified plausible, regional mechanisms, whereby the aerosol-induced change is expected to be of opposite sign to the greenhouse gas-induced change. The DPT trend is relatively small in the historical experiment, the ten member ensemble average over the period 1950–2000 was found to be about a quarter for that of the GHG experiment, only slightly less than the unforced control experiment and over 20 times stronger in RCP4.5 over the period 2006–2100. We expect, however, that the DPT response to be significantly model-dependent, whether it be during forced or unforced experiments.

It is noteworthy that the simulated effects of anthropogenic aerosols on the three indices we used to characterize the global ocean circulation are all comparable to or larger than those of increasing greenhouse gases in the historical period. Whereas it is possible that the global-mean
anthropogenic aerosol forcing (~1.4 W m⁻²) in CSIRO-Mk3.6 is too strong, it is only slightly stronger than the estimate (~1.2 W m⁻²) from Forster et al. (2007). It is also smaller in magnitude than a recent ‘top down’ estimate of ~1.6 W m⁻² from Hansen et al. (2011).

Into the late 21st century, when the atmospheric warming impact of greenhouse gases is expected to dominate over the overall cooling effect of anthropogenic aerosols, we identified a decrease in the strength of the North Atlantic Deep Water and the Indonesian Throughflow, whereas the Drake Passage Transport increases. Examination of the time series of these indices for the NoAA or GHG experiments in the period 1950–2005 shows a slope stronger than the historical experiment with ‘all forcings’. This suggests that without the influence of anthropogenic aerosols the earth’s recent climate would have already followed a path much more like one dominated by greenhouse gases. This is especially important given the expectation that there will be a smaller concentration of anthropogenic aerosols in the future atmosphere (Bellouin et al. 2011, Lamarque et al. 2011).

Acknowledgments

The first author would like to acknowledge the financial help from a CSIRO Mid-Career Capability Development Fund Proposal obtained in 2012 to undertake the work in this paper during a visit to the School of Earth and Environmental Sciences, Seoul National University, South Korea. KYK acknowledges support for inter-institute collaborative research from the Research Institute of Basic Sciences, College of Natural Sciences, Seoul National University. This work was supported by CSIRO Marine and Atmospheric Research and the Queensland Climate Change Centre of Excellence (Queensland Government Department of Science, Information Technology, Innovation and the Arts). QCCE provided the high performance computing facilities for the model experimentation. The National Computational Infrastructure (National Facility) provided computing resources for post-processing the raw model datasets, hardware for hosting the CSIRO-Mk3.6 CMIP5 data on the Earth System Grid, and technical support. The NCAR Command Language (Version 6.0.0) [Software]. (2012). Boulder, Colorado: UCAR/NCAR/CSIL/VETS. http://dx.doi.org/10.5065/D6WD3XHS was used to perform most of the analysis and graphics needed to complete this paper. We thank and acknowledge the constructive comments from the reviewers.

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## Appendix: Acronyms and abbreviations

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<tr>
<th>Acronym</th>
<th>Full Form</th>
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<tbody>
<tr>
<td>AAs</td>
<td>Anthropogenic aerosols</td>
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<tr>
<td>ACC</td>
<td>Antarctic Circumpolar Current</td>
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<tr>
<td>CMIP5</td>
<td>Coupled Model Intercomparison Phase 5</td>
</tr>
<tr>
<td>DPT</td>
<td>Drake Passage Transport</td>
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<tr>
<td>GCMs</td>
<td>General circulation models</td>
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<td>GHGs</td>
<td>Greenhouse gases</td>
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<td>ITCZ</td>
<td>Intertropical Convergence Zone</td>
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<td>ITF</td>
<td>Indonesian Throughflow</td>
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<td>GHGs</td>
<td>Greenhouse Gases</td>
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<td>Mk3A</td>
<td>CSIRO-Mk3A</td>
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<td>MOC</td>
<td>Meridional Overturning Circulation</td>
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<tr>
<td>NADW</td>
<td>North Atlantic Deep Water</td>
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<tr>
<td>RCP</td>
<td>Representative Concentration Pathway</td>
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<tr>
<td>SSS</td>
<td>Sea surface salinity</td>
</tr>
<tr>
<td>SST</td>
<td>Sea surface temperature</td>
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