Declining Aerosols in CMIP5 Projections: Effects on Atmospheric Temperature Structure and Midlatitude Jets

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ABSTRACT

The effects of declining anthropogenic aerosols in representative concentration pathway 4.5 (RCP4.5) are assessed in four models from phase 5 the Coupled Model Intercomparison Project (CMIP5), with a focus on annual, zonal-mean atmospheric temperature structure and zonal winds. For each model, the effect of declining aerosols is diagnosed from the difference between a projection forced by RCP4.5 for 2006–2100 and another that has identical forcing, except that anthropogenic aerosols are fixed at early twenty-first-century levels. The response to declining aerosols is interpreted in terms of the meridional structure of aerosol radiative forcing, which peaks near 40°N and vanishes at the South Pole.

Increasing greenhouse gases cause amplified warming in the tropical upper troposphere and strengthening midlatitude jets in both hemispheres. However, for declining aerosols the vertically averaged tropospheric temperature response peaks near 40°N, rather than in the tropics. This implies that for declining aerosols the tropospheric meridional temperature gradient generally increases in the Southern Hemisphere (SH), but in the Northern Hemisphere (NH) it decreases in the tropics and subtropics. Consistent with thermal wind balance, the NH jet then strengthens on its poleward side and weakens on its equatorward side, whereas the SH jet strengthens more than the NH jet. The asymmetric response of the jets is thus consistent with the meridional structure of aerosol radiative forcing and the associated tropospheric warming: in the NH the latitude of maximum warming is roughly collocated with the jet, whereas in the SH warming is strongest in the tropics and weakest at high latitudes.

1. Introduction

Climate projections in phase 5 of the Coupled Model Intercomparison Project (CMIP5) are forced by four representative concentration pathways (RCPs). Declining aerosol emissions are a feature of all the RCPs during the twenty-first century, because emission controls are assumed to increase as income rises (Smith et al. 2005). The decline is much larger than assumed in earlier scenarios (van Vuuren et al. 2011), although it is unclear how realistic it is. Declining aerosols may have important implications for projections carried out during CMIP5, since their effects are unlikely to be identical to those of increasing greenhouse gases. However, there have been few studies of these implications.

Since all the RCPs include sharp declines in aerosols, the range of plausible aerosol futures is not adequately sampled by the standard RCPs (Kirtman et al. 2014). The effects of projected aerosol changes can be identified by running simulations that are similar to the standard RCPs, except that emissions or concentrations of anthropogenic aerosols are held fixed at present-day levels. A few groups have recently published analyses of such simulations, which have generally been based on single models (Levy et al. 2013; Gillett and von Salzen

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To varying degrees, these studies show that declining aerosols increase projected twenty-first-century warming.

A similar conclusion was reached by Meehl et al. (2013), who analyzed CMIP5 projections carried out with the Community Earth System Model, version 1 (CESM1). Compared to the previous version of the model (which only treated direct aerosol effects), they found that inclusion of indirect aerosol effects in CESM1 added close to 1 W m$^{-2}$ of positive radiative forcing during the twenty-first century.

In addition to analysis of a single model, Rotstayn et al. (2013) considered intermodel correlations between historical aerosol radiative forcing and projected changes in temperature and precipitation in RCP4.5 from 13 CMIP5 models; they found that models with stronger (more negative) historical aerosol forcing tend to simulate less warming during 1850–2005 and more warming during 2006–2100. In other words, models with more aerosol cooling in the historical period tend to have more projected warming due to declining aerosols. However, until now there have been no conventional multimodel analyses of the effects of declining aerosols.

Some insights into the possible effects of declining aerosols on large-scale circulation can be gained from analyses of the effects of historically increasing aerosols. Several studies have focused on the effects of stronger aerosol forcing in the Northern Hemisphere (NH) than the Southern Hemisphere (SH). In response to the interhemispheric gradient in forcing, models generally simulate more cooling in the NH than the SH, and an associated southward shift of tropical precipitation. This has been seen in atmospheric global climate models (GCMs) coupled to slab ocean models (Rotstayn et al. 2000; Williams et al. 2001; Rotstayn and Lohmann 2002; Kristjánsson et al. 2005; Jones et al. 2007; Yoshimori and Broccoli 2008; Ming and Ramaswamy 2011), and more recently in fully coupled ocean–atmosphere GCMs (Chang et al. 2011; Wilcoxon et al. 2013; Hwang et al. 2013). Various studies also suggest large regional impacts on circulation and rainfall from historical increases in aerosols, especially in the Sahel (Rotstayn and Lohmann 2002; Kawase et al. 2010; Ackerley et al. 2011), South Asia (Ramanathan et al. 2005; Meehl et al. 2008; Bollasina et al. 2011), and East Asia (Cheng et al. 2005; Liu et al. 2011).

These studies have generally emphasized the effects of spatially inhomogeneous aerosol forcing. On the other hand, Xie et al. (2013) analyzed CMIP5 climate simulations, and concluded that the spatial patterns of climate response to increasing greenhouse gases and aerosols are qualitatively similar (but of opposite sign) over the ocean. They argued that common feedbacks occur in response to both forcing agents, leading to similar spatial patterns of sea surface temperature (SST) and rainfall response, despite different patterns of forcing.

This raises questions about whether declining aerosols have similar effects on climate to those of increasing greenhouse gases, or whether it is necessary to account for the unique spatial distribution of aerosol radiative forcing in climate projections. One reason why this is important is that aerosol effects are highly uncertain; for example, the global-mean radiative forcing due to anthropogenic aerosols is estimated to have an uncertainty range from $-0.1$ to $-1.9$ W m$^{-2}$ (Boucher et al. 2014). Also, because all the RCPs have sharply declining aerosols, CMIP5 projections may show systematic differences from CMIP3 projections because of spatially inhomogeneous aerosol changes (Kirtman et al. 2014).

A topic that has recently received some attention is the effect of aerosol forcing on the jet streams, which are important features of extratropical climate. The wind shear associated with the jet streams contributes to baroclinic instability, so they are associated with midlatitude eddy formation; thus, projected changes in the jets have implications for changes in storm development and extreme weather (Frederiksen et al. 2011; Francis and Vavrus 2012). In the SH, future changes in the associated surface westerlies may weaken the uptake of CO$_2$ by the Southern Ocean, so reliable projections of changes in the jets are also important for understanding the carbon cycle (Lenton et al. 2009).

Ming et al. (2011) considered the effects of increasing anthropogenic aerosols on winter extratropical circulation in the NH in an atmospheric model coupled to a slab ocean model. The zonal-mean response of their model included an equatorward shift of the subtropical jet. Since the aerosol-induced cooling was roughly collocated with the NH jet, it strengthened the meridional temperature gradient on the equatorward side of the jet, and weakened it on the poleward side; consistent with thermal wind balance, the jet strengthened (weakened) on its equatorward (poleward) side.

Allen and coworkers have studied the effects on atmospheric circulation of increases in scattering or absorbing aerosols, with a primary focus on the width of the tropics and associated jet displacement in the NH. In simulations by Allen and Sherwood (2011), the direct radiative effects of increases in scattering aerosols shifted the NH subtropical jets equatorward, whereas absorbing aerosols had the opposite effect. Allen et al. (2012a) showed that recent NH tropical expansion has been primarily driven by short-lived tropospheric warming.
agents (black carbon and ozone); further, poleward expansion of the tropics and the NH tropospheric jet was consistent with an “expansion index” (Allen et al. 2012b), which compared midlatitude tropospheric warming with that at other latitudes. Allen et al. (2012a,b) noted that warming of midlatitudes relative to others displaces the maximum meridional temperature gradient poleward, so the jet also shifts poleward [similar to the explanation of Ming et al. (2011)]. Recently, Allen and Ajoku (2014, manuscript submitted to Geophys. Res. Lett.) showed that projected declines in anthropogenic aerosol amplify simulated twenty-first-century warming in the NH midlatitudes, and cause a poleward shift of NH circulation.

Considering the SH, a weakening of annual-mean surface zonal winds at sub-Antarctic latitudes in response to historically increasing aerosols has been noted in CMIP5 models (Xie et al. 2013; Collier et al. 2013); this suggests that increasing aerosols induce a weakening and/or equatorward shift of the eddy-driven jet in the SH, a response that resembles the inverse of the effects of increasing greenhouse gases (Fyfe et al. 1999; Kushner et al. 2001). Rotstayn (2013) considered the simulated effects of projected aerosol declines on SH extratropical circulation in austral summer and found a poleward shift of the eddy-driven jet, which is qualitatively consistent with the results of Xie et al. (2013) and Collier et al. (2013).

Here we consider projections based on RCP4.5 from four GCMs that participated in CMIP5. For each of these models, we compare the standard RCP4.5 projection with a modified RCP4.5, in which anthropogenic aerosols are held fixed at present-day values. By comparing the standard and modified RCP4.5 projections, we diagnose the effects of declining aerosols during the twenty-first century. The focus of this study is on annually and zonally averaged atmospheric temperature structure and the midlatitude jets in both hemispheres. Since aerosol effects are complex and uncertain, we aim to identify broad, large-scale responses that are common to the models, but we also identify differences among the models.

2. Models, experiments, and methods

We use simulations from four CMIP5 models: CanESM2, CSIRO Mk3.6, GFDL CM3, and IPSL-CM5A-LR (references and expansions of these model names are given in Table 1). Anthropogenic aerosols treated in all models are sulfate, organic aerosol, and black carbon. In IPSL-CM5A-LR, aerosol concentrations are prescribed, based on offline calculations with the atmospheric model (Szopa et al. 2013); in the other models, anthropogenic aerosol concentrations are calculated interactively, using prescribed emissions based on Lamarque et al. (2011). In CSIRO Mk3.6, emissions of organic aerosol and black carbon are increased by 50% and 25%, respectively (Rotstayn et al. 2012). All models treat direct aerosol radiative effects and the first indirect effect, but there are differences in the inclusion of other indirect effects; see Table 2 for a summary. Further details of the forcing agents used in each model are given in Table 12.1 of Collins et al. (2014).

Our analysis of twenty-first-century projections is based on the following experiments:

- RCP45 is a projection for 2006–2100 forced by the medium-low RCP4.5 pathway (Thomson et al. 2011). Anthropogenic forcing in this experiment is based on time-varying, prescribed concentrations of ozone and well-mixed greenhouse gases (WMGHGs) and emissions of anthropogenic aerosols and their precursors (for CanESM2, CSIRO Mk3.6, and GFDL CM3) or prescribed concentrations of anthropogenic aerosols (for IPSL-CM5A-LR). Note that anthropogenic sources include biomass burning. Land use change is also included in all models except CSIRO Mk3.6. Ensemble sizes range from 3 (GFDL CM3) to 10 (CSIRO Mk3.6); see Table 1.

- RCP45fixAA is identical to RCP45, except that anthropogenic aerosol emissions (for CanESM2, CSIRO...
In addition, standard historical simulations are used to derive climatological fields for 1986–2005, which are used for reference in some of the figures. For three of the models, further details of the RCP45fixAA simulations have been given by Gillett and von Salzen (2013) (CanESM2), Rotstayn et al. (2013) (CSIRO Mk3.6), and Levy et al. (2013) (GFDL CM3). We diagnose the effects of declining aerosols from RCP45 minus RCP45fixAA, consistent with the approach used in these studies; we shall use the term “declining aerosols” to refer to aerosols from anthropogenic and biomass-burning sources, noting that natural aerosols (such as dust and sea salt) may also change, but do not necessarily decline. The large-scale climate response in RCP45fixAA is expected to be mostly dominated by the effects of increasing WMGHGs. In some respects this assumption is not valid; a notable example is circulation in the SH in austral summer, when the effects of stratospheric ozone recovery are important (Arblaster et al. 2011; Eyring et al. 2013).

The CMIP5 protocol also includes runs that enable calculation of aerosol effective radiative forcing (ERF) for the year 2000 relative to 1850. Aerosol ERF is an estimate of total aerosol forcing, which includes direct and indirect radiative effects as well as rapid adjustments of the atmosphere (such as the semidirect effect). Aerosol ERF can be calculated from pairs of atmospheric model runs with the same prescribed climatological SSTs. In CMIP5, these are referred to as sstClim and sstClimAerosol (Taylor et al. 2012). The first (sstClim) has all forcing agents set to 1850 values, while the second (sstClimAerosol) differs from sstClim only with respect to anthropogenic aerosols, which have emissions or concentrations appropriate for the year 2000. The difference in net radiation at the top of the atmosphere between the two runs gives aerosol ERF for 2000 relative to 1850. A further run (sstClimSulfate) enables calculation of aerosol ERF for 2000 relative to 1850 for sulfate only (Taylor et al. 2012). As noted by Boucher et al. (2014), the method is not perfect, because of the nonzero change in land surface temperature between the two runs; one effect of this is to artificially perturb the land–sea temperature gradient, which can induce changes in circulation in monsoonal regions (Rotstayn et al. 2013). Some results from these runs are discussed in section 3a. In addition, we use another pair of analogous runs from CSIRO Mk3.6 for the years 2005 and 2100 (using emissions from RCP4.5 in 2100); the difference of these two runs enables calculation of twenty-first-century aerosol ERF in that model.

Changes in twenty-first-century climate are expressed as least squares linear trends over the period 2006–2100; note that the use of linear trends does not imply that changes are necessarily linear over the time period. Statistical significance of the ensemble mean from each model is assessed using a two-sided t test, taking the individual trend values from each ensemble member as independent data points. The multimodel ensemble (MME) mean is calculated as the average of the ensemble means from the four models. When plotting zonally averaged trends, model agreement is assessed following the method of Tebaldi et al. (2011). Specifically, stippling is used to denote agreement in plots of the MME-mean trend if both of the following criteria are satisfied: 1) at least three of the four models have zonal-mean trends that are significantly different from zero at the 5% level, and 2) at least three of the models
with significant trends (i.e., three out of three or three out of four) agree on the sign of the trend.

3. Results and discussion

a. Aerosol radiative forcing

A useful guide to the magnitude of the effects of declining aerosols is the aerosol ERF, which can be calculated from pairs of runs with prescribed SSTs (section 2). This quantity is not generally available from models for the twenty-first century, but in RCP4.5 (and the other RCPs), assumed emissions of sulfur and carbonaceous aerosols return to roughly nineteenth-century levels by 2100 (Lamarque et al. 2011; Smith and Bond 2014). It follows that global-mean aerosol ERF almost returns to its preindustrial value by 2100 (Bond 2014). Thus, to first order, published CMIP5 aerosol ERF for 2000 relative to 1850 can be used as a proxy (with sign reversed) for aerosol ERF during the twenty-first century (Rotstayn et al. 2013). In other words, the change in aerosol ERF during the twenty-first century is similar to the historical change in aerosol ERF, except that it is positive rather than negative.

Global-mean values of aerosol ERF from published CMIP5 runs are given in Table 2; note that the two models that include treatments of the cloud-lifetime effect (CSIRO Mk3.6 and GFDL CM3) have markedly stronger aerosol ERF than CanESM2 and IPSL-CM5A-LR. The modeled values can be compared with the best estimate (−0.9 W m⁻²) and uncertainty range (from −0.1 to −1.9 W m⁻²) from the recent Intergovernmental Panel on Climate Change (IPCC) report (Boucher et al. 2014); note that the IPCC values apply to the period 1750–2010, and exclude the effects of black carbon on snow and ice albedo. The mean aerosol ERF from the four models is −1.15 W m⁻², which is essentially the same as the mean value from the 13 models considered by Rotstayn et al. (2013).

Table 2 shows an interesting difference between CSIRO Mk3.6 and the other three models. Assuming linearity, the difference between aerosol ERF and sulfate ERF gives an estimate of ERF from carbonaceous aerosol (organic aerosol and black carbon). In CanESM2, GFDL CM3, and IPSL-CM5A-LR, carbonaceous aerosol ERF is close to zero, but in CSIRO Mk3.6 it is −0.3 W m⁻². This implies that CSIRO Mk3.6 has positive ERF from declining carbonaceous aerosol in the twenty-first century (because of organic aerosol, since ERF from black carbon is positive). As discussed by Rotstayn et al. (2012), negative ERF from carbonaceous aerosol may be caused by the 50% increase of organic aerosol emissions in CSIRO Mk3.6, as well as a substantial indirect effect (which is parameterization dependent). Note that CMIP5 models show considerable uncertainty in carbonaceous aerosol ERF, ranging from roughly zero to −0.6 W m⁻² (Boucher et al. 2014, their Table 7.5).

The models with stronger aerosol ERF in the historical period (CSIRO Mk3.6 and GFDL CM3) have larger projected trends in global-mean surface air temperature (SAT) due to declining aerosols than CanESM2 and IPSL-CM5A-LR (Table 2). The fraction of projected warming due to declining aerosols is 19% in IPSL-CM5A-LR, 25% in CanESM2, 38% in GFDL CM3, and 48% in CSIRO Mk3.6, reflecting the uncertainty associated with aerosol effects. This range also suggests that the magnitude of changes in circulation caused by declining aerosols will vary substantially among the models. Comparing CSIRO Mk3.6 and GFDL CM3, it is initially surprising that the fraction of projected warming is larger in CSIRO Mk3.6, even though historical aerosol ERF is stronger in GFDL CM3; this occurs because future aerosol ERF is not exactly the inverse of historical aerosol ERF, as discussed below.

Simulated temperature changes depend on climate sensitivity as well as forcing. Transient climate response (TCR) is the change in SAT at the time of CO₂ doubling in an experiment forced by CO₂ concentrations that increase at 1% yr⁻¹; it depends on equilibrium climate sensitivity and ocean heat uptake (e.g., Forster et al. 2013). Values of TCR in the four models (last column of Table 2) range from 1.8° to 2.4°C. For comparison, Forster et al. (2013) found a MME-mean value of 1.8° ± 0.63°C from 23 CMIP5 models (where the range is a 90% confidence interval); thus the four models in this study tend to have larger TCR than the average from CMIP5.

One might expect the models’ different TCR values to be reflected in their SAT trends in RCP45fixAA, in which the main forcing agent is increasing WMGHGs. Consistent with having the lowest TCR, CSIRO Mk3.6 also has the smallest SAT trend in RCP45fixAA (1.27°C century⁻¹). The other models all have similar SAT trends in RCP45fixAA (between 1.67° and 1.77°C century⁻¹), and there is no obvious correlation with their TCR values, which range from 2.0°C (GFDL CM3 and IPSL-CM5A-LR) to 2.4°C (CanESM2). It is unclear why the larger TCR in CanESM2 is not reflected in its SAT response in RCP45fixAA; it may be that forcing agents other than CO₂ (non-CO₂ greenhouse gases and land use change) cause more warming in that model.

In addition to global-mean values of aerosol ERF, it is instructive to also consider its variation with latitude. Figure 1 shows annual, zonal-mean aerosol ERF at the
top of the atmosphere for 2000 relative to 1850, multiplied by \(-1\) (solid curves); the sign is changed so that the values are generally positive, as would be the case for twenty-first-century aerosol ERF.

Consistent with the global-mean values given in Table 2, CSIRO Mk3.6 and GFDL CM3 tend to show stronger aerosol ERF than IPSL-CM5A-LR and CanESM2. An exception is in the Arctic region, where inclusion of BC effects on snow and ice albedo causes aerosol ERF in CSIRO Mk3.6 to change sign. The curves tend to be somewhat noisier in CSIRO Mk3.6 and GFDL CM3 than the other models; this may be due to the shorter run length in these models, as well as the larger magnitude of aerosol ERF. The individual models and the MME mean have largest aerosol ERF near 40°N, while in the SH aerosol ERF tends to decrease toward the pole; we shall return to this important point in section 3b, when we consider the response of atmospheric temperature to declining aerosols.

Also shown (as the dashed blue curve) is aerosol ERF from CSIRO Mk3.6 for 2100 relative to 2005. The meridional variation of future aerosol ERF resembles the inverse of the historical pattern, with peak magnitude near 40°N. However, there are also some differences; for example, future aerosol ERF in CSIRO Mk3.6 is stronger in NH midlatitudes but weaker near the equator than historical aerosol ERF.

Why is future aerosol ERF not exactly the inverse of historical aerosol ERF? In part this occurs because of different effects from sulfate and carbonaceous aerosol. Global-mean organic aerosol emissions in RCP4.5 return to slightly less than 1850 levels by 2100, but sulfur and black carbon emissions in 2100 are comparable to those in 1900 (Fig. 2); note that sulfur emissions are roughly 5 times larger in 1900 than in 1850. Comparing CSIRO Mk3.6 and GFDL CM3, this suggests why CSIRO Mk3.6 has a larger fraction of projected warming due to declining aerosols, even though its historical aerosol ERF is weaker: in CSIRO Mk3.6 there is additional positive ERF from the decline in organic aerosol (which exerts a substantial negative ERF in that model). As a result, global-mean twenty-first-century (2100 minus 2005) aerosol ERF in CSIRO Mk3.6 is 1.5 W m\(^{-2}\), which is larger than the corresponding value from GFDL CM3 (1.3 W m\(^{-2}\)).

This model-dependent balance between sulfate and carbonaceous aerosol also affects the geographical pattern of future aerosol ERF, since there are larger sources of organic aerosol than sulfur dioxide in the tropics. Some specific geographical differences between historical and future changes in emissions have been noted in previous studies—for example, biomass-burning aerosol emissions in central Africa (and the associated ERF) are not assumed to decrease as much as in other regions (Takemura 2012; Rotstayn et al. 2013); this contributes
to the weaker future aerosol ERF in the deep tropics in CSIRO Mk3.6.

In comparison to aerosol ERF, the radiative forcing due to increasing WMGHGs is more uniform in space (not shown). When evaluated as the traditional “stratosphere adjusted” forcing (i.e., as the change in radiative flux at the tropopause, after allowing the stratosphere to adjust), WMGHG forcing is quasi-uniform in space, with only a modest enhancement in the subtropics (Shindell et al. 2013, their Fig. 20). When evaluated as an ERF (i.e., using an analogous method as used to generate the aerosol ERF in Fig. 1), rapid adjustments of the atmosphere and land surface allow more variation of the forcing with latitude, although this variation is still much less than for aerosol ERF (Xie et al. 2013, their Fig. 1).

b. Atmospheric temperature trends

Although our main focus is on the MME mean, we first compare atmospheric temperature trends due to declining aerosols in each of the four models. Figure 3 shows ensemble-mean trends in zonal-mean atmospheric temperature from RCP45 minus RCP45fixAA. Tropospheric warming is larger in the models with stronger aerosol forcing (CSIRO Mk3.6 and GFDL CM3), while the model with the weakest aerosol forcing (IPSL-CM5A-LR) shows less tropospheric warming.

To varying degrees, all the models also show stratospheric cooling, which bears some resemblance to the effect of increasing WMGHGs (e.g., Lorenz and DeWeaver 2007). A similar but opposite response (i.e., stratospheric warming) is seen in the MME mean of historical simulations forced only by increasing aerosols (Xie et al. 2013). However, the process that causes stratospheric cooling due to increasing WMGHGs (increasing longwave emission) cannot explain this response: anthropogenic aerosols only interact with shortwave radiation in these models, except for CanESM2, in which the longwave effect is small (Li et al. 2001). This is discussed further in section 3d.

We now compare the zonal-mean structure of MME-mean atmospheric temperature trends due to increasing WMGHGs (RCP45fixAA) and declining aerosols (RCP45 minus RCP45fixAA), while also noting the effects of stratospheric ozone recovery in RCP45fixAA; these are shown in Fig. 4. To first order, the patterns are similar in RCP45fixAA and RCP45 minus RCP45fixAA.

Warming in the lower Antarctic stratosphere in RCP45fixAA (Fig. 4a) is due to recovery of ozone. The first-order temperature balance in the stratosphere is between shortwave absorption by ozone and longwave emission by CO2 (Held 1993); thus stratospheric ozone recovery warms the lower stratosphere (especially at high latitudes in the SH), offsetting the cooling due to...
increasing WMGHGs (Butchart et al. 2010; Eyring et al. 2013).

In the troposphere, both panels of Fig. 4 show features that have previously been identified in the atmospheric response to increasing WMGHGs. Suppressed warming near the surface in the SH extratropics is due to large ocean heat uptake (Kuhlbrodt and Gregory 2012; Lu and Zhao 2012). Amplified warming over the Arctic is caused by temperature-related feedbacks and (to a lesser extent) surface–albedo feedbacks (Pithan and Mauritsen 2014). Amplified warming is seen in the tropical upper troposphere, because tropical convection tends to adjust the troposphere toward a moist adiabat (Wilson and Mitchell 1987; Bony et al. 2006). These features are also seen in the RCP4.5 MME-mean temperature change from 41 CMIP5 models (Collins et al. 2014, their Fig. 12.12). The idea that similar feedbacks cause the response pattern to be similar for WMGHGs and aerosols (at least to first order) is consistent with the arguments of Xie et al. (2013).

In the next few paragraphs we show that, although the temperature response patterns for increasing WMGHGs and declining aerosols have these similar features, the aerosol-induced response also shows the signature of the distinct meridional pattern of aerosol radiative forcing. In effect, the aerosol-induced temperature response has a component that is similar to the effect of increasing WMGHGs (governed by the feedbacks described in the previous paragraph), and another component that reflects the pattern of aerosol radiative forcing.

First, we wish to take the difference of the two patterns in Fig. 4. Since the MME-mean temperature response is larger for increasing WMGHGs (Fig. 4a) than declining aerosols (Fig. 4b), we first normalize the MME-mean temperature trends by dividing them by the corresponding MME-mean, global-mean surface air temperature trends. Using the results in Table 2, these surface air temperature trends amount to 1.60°C century\(^{-1}\) for RCP45fixAA and 0.80°C century\(^{-1}\) for RCP45 minus RCP45fixAA.

Figure 5 shows the difference (declining aerosols minus increasing WMGHGs) of MME-mean, normalized, zonal-mean atmospheric temperature trends. Relative to the pattern induced by increasing WMGHGs, declining aerosols cause more warming near 40°N (where aerosol ERF is largest) and less warming in the SH, especially at high latitudes (where aerosol ERF is smallest). Warming because of ozone recovery is also expected to contribute to the differences seen at high SH latitudes (e.g., Shindell and Schmidt 2004).

Is the tropospheric warming trend due to declining aerosols larger in the NH midlatitudes than in the tropics? It is not completely clear from Fig. 5, because it...
shows differences relative to the pattern from increasing WMGHGs, or from Fig. 4b, because the contour levels do not clearly resolve the temperature trends in the NH subtropics. One way to see this is to plot normalized temperature trends, vertically averaged between the surface and 250 hPa (Fig. 6); normalizing the temperature trends (as in Fig. 5) makes it easier to compare the patterns due to declining aerosols and increasing WMGHGs. The warming trend due to declining aerosols (blue curve) shows maxima at 35°N (slightly equatorward of the peak aerosol ERF) and over the Arctic. This confirms that (in a vertically integrated sense) the tropospheric warming due to declining aerosols is indeed stronger in the NH subtropics than in the tropics. In contrast, the warming trend due to increasing WMGHGs (red curve) shows a broad tropical maximum, with a moderate enhancement in the NH compared to the SH; this enhancement is expected given the larger land fraction in the NH (e.g., Joshi et al. 2008). Note that if the integration was terminated at the tropopause, instead of 250 hPa, both curves would show more amplification in the deep tropics; we chose to terminate the integration at 250 hPa because that is roughly the level of the jet stream at around 200 hPa (shown in Fig. 8). Consistent with thermal wind balance, there is a reversal of the MTG above the level of the climatological jets.

Trends in the MTG show similarities between RCP45fixAA (Fig. 7a) and RCP45 minus RCP45fixAA (Fig. 7b), but also some interesting differences. In the SH tropics, RCP45 minus RCP45fixAA shows a reversed MTG trend between about 10° and 30°N, extending up to roughly 300 hPa; this is a consequence of the subtropical warming maximum referred to above (Fig. 6). There is only a very weak indication of a similar feature in RCP45fixAA. At higher latitudes in the NH, MTG trends in the lower troposphere in both panels show the effect of enhanced Arctic warming; the effect is much stronger in RCP45fixAA than in RCP45 minus RCP45fixAA, perhaps because aerosol ERF is small over the Arctic (Fig. 1).

c. Zonal wind trends

In this section, we consider trends in zonal winds, and discuss the link to atmospheric temperature structure via the thermal wind equation, namely

\[ \frac{\partial u}{\partial p} = \frac{R}{fp} \left( \frac{\partial T}{\partial y} \right)_p \]  

(e.g., Holton 1992). Here, \( u \) is zonal-mean zonal wind, \( p \) is pressure, \( R \) is the gas constant for dry air, \( f \) is the Coriolis parameter, and \( (\partial T/\partial y)_p \) is the zonal-mean MTG on constant pressure surfaces.

MME-mean zonal wind trends are shown in the upper panels of Fig. 8, with corresponding thermal wind trends in the lower panels. Thermal winds at a given pressure level are obtained by integrating Eq. (1) upward from 1000 hPa, using centered finite differences. There is generally good agreement between the zonal winds and thermal winds, both in the climatology and trends; this indicates that the zonal wind trends are consistent with
a geostrophic adjustment to the modified atmospheric temperature structure.

Zonal wind trends in the upper panels of Fig. 8 show strengthening midlatitude jets, which are a well-known feature of simulations forced by increasing WMGHGs (Lorenz and DeWeaver 2007; Collins et al. 2014). The patterns are qualitatively similar, but there are some marked differences. The strengthening of the jets is more substantial in RCP45fixAA than in RCP45 minus RCP45fixAA, consistent with larger temperature changes for increasing WMGHGs than declining aerosols in the MME mean. In the Antarctic stratosphere, the jet weakens in RCP45fixAA (Fig. 8a). This is a response to the warming effect of recovering ozone, and is also consistent with thermal wind balance (Figs. 8c and 7a). At high northern latitudes there are easterly wind trends in RCP45fixAA, consistent with strong warming near the Arctic surface (Woolings 2008); in RCP45 minus RCP45fixAA there is only a hint of this effect. In RCP45fixAA the strengthening of the jets is more...
symmetric between the hemispheres, although there is a somewhat larger response in the SH than the NH. In the NH, the jet shows a distinct response in RCP45 minus RCP45fixAA (Fig. 8b); it weakens on its equatorward, lower flank and strengthens on its poleward, upper flank, so it shifts poleward and upward. This is also seen in the corresponding thermal wind trends (Fig. 8d). The thermal wind relation does not—in itself—distinguish between cause and effect, but the above discussion suggests that there is a direct causal link from the peak in aerosol ERF near 40°N, to the corresponding temperature response and then to the change in zonal wind structure. A similar (but opposite) response in NH winter was found by Ming et al. (2011) for increasing aerosols in an atmospheric model coupled to a slab ocean model (their Fig. 3b). Our explanation in terms of thermal wind balance is consistent with theirs.

Another interesting aspect of the zonal wind response to declining aerosols is that the jet strengthens more in the SH than the NH. Since aerosol concentrations are generally lower in the SH than the NH, this may seem counterintuitive. However, Fig. 7b shows that this result is consistent with the more widespread increase of the MTG in the subtropics and midlatitudes of the SH than in the NH. Thus it is the very fact that aerosol forcing is stronger in the NH that causes this effect: aerosol ERF peaks near 40°N, and in the MME mean it is relatively flat between the equator and 40°N (Fig. 1). On the other hand, south of the equator its magnitude falls away rapidly with increasing latitude.

In the SH, MME mean strengthening westerlies in RCP45fixAA and RCP45 minus RCP45fixAA extend down to the surface; this equivalent barotropic structure is consistent with a strengthening and poleward shift of the eddy-driven jet (Kushner et al. 2001). However, there is little stippling below the upper troposphere in Fig. 8, denoting a lack of model agreement. In RCP45fixAA (figure not shown), different responses of the SH westerlies reflect different balances between the effects of increasing WMGHGs, which shift them poleward, and recovering stratospheric ozone, which shifts them equatorward (Arblaster et al. 2011; Eyring et al. 2013). Since declining aerosols are the main focus of this paper, the lack of model agreement in RCP45 minus RCP45fixAA merits a closer look.

Figure 9 compares zonal wind trends from RCP45 minus RCP45fixAA in the four models. In CanESM2, CSIRO Mk3.6, and IPSL-CM5A-LR, wind trends near the surface indicate a poleward shift of SH surface westerlies (although the trends are not significant in IPSL-CM5A-LR). On the other hand, in GFDL CM3 strengthening surface westerlies occur farther north, at roughly the same latitude as the climatological westerlies. Comparing Fig. 3, GFDL CM3 is also the only model that does not show SH polar stratospheric cooling...
in RCP45 minus RCP45fixAA; this is discussed further in section 3d.

There are several qualitative features that all models have in common in Fig. 9. At middle-to-high SH latitudes, the zonal wind response has an equivalent barotropic vertical structure in all models, consistent with a strengthening eddy-driven jet. In the subtropics of the SH, there is an indication of a baroclinic response, with a stronger subtropical jet and easterly wind trends near the surface. In the NH, the subtropical jet in all models moves upward and poleward, but only CSIRO Mk3.6 shows significant strengthening of the eddy-driven jet (i.e., stronger midlatitude westerlies at the surface).

Figure 10 shows the MME-mean zonal wind response in RCP45, which (by construction) equals the sum of the trends in Figs. 8a and 8b. In several respects, the zonal wind trends caused by increasing WMGHGs and declining aerosols are essentially additive. This applies to the strengthening and upward shift of the jets, the enhanced midlatitude westerlies that extend down to the surface in the SH (and to a lesser extent in the NH), and the low-level easterly wind trends in the SH tropics and subtropics. All of these features are broadly consistent with the CMIP5 MME mean shown by Collins et al. (2014, their Fig. 12.19); thus, declining aerosols tend to enhance these aspects of the dynamical response to increasing WMGHGs.

On the other hand, the weakening of the NH jet on its lower and equatorward flank is not stippled in Fig. 10; the lack of model agreement in RCP45 occurs because CanESM2 has positive zonal wind trends in RCP45fixAA in that region (not shown), and this offsets the negative wind trends due to declining aerosols. The RCP4.5 MME-mean in Collins et al. (2014) does show negative wind trends between the surface and 300 hPa in the vicinity of 30°N, although the feature does not satisfy their criterion for significance (with changes of consistent sign in 90% of models). This may be caused (at least in part) by the wide range of aerosol treatments in CMIP5 models; for example, some models only treat direct aerosol effects.

A southward shift of the eddy-driven jet in the SH is suggestive of a positive trend in the southern annular mode (SAM), which is associated with positive trends of sea level pressure at midlatitudes and negative trends of sea level pressure at high latitudes (Gong and Wang 1999). These changes in sea level pressure reflect an anomalous meridional circulation (with subsidence at midlatitudes and ascent at high latitudes) associated with the southward shift of the eddy-driven jet (Thompson and Wallace 2000).

Annual, zonal-mean trends of sea level pressure are shown in Fig. 11. In RCP45fixAA (Fig. 11a), sea level pressure trends are generally consistent with a positive SAM trend, (although only weakly so in CSIRO Mk3.6), indicating that in these models the effects of increasing WMGHGs tend to outweigh the effects of recovering ozone on the SAM.

Declining aerosols (Fig. 11b) also induce positive SAM-like sea level pressure trends, which are relatively similar among the models. This result is consistent with Rotstayn (2013), who found a positive SAM trend in austral summer due to declining aerosols in CSIRO Mk3.6, and noted that associated circulation changes in the SH were similar to those induced by increasing WMGHGs. It is also consistent with Sigmond and Fyfe (2014), who found a decreased SAM in response to historical aerosol increases in austral summer in a majority of CMIP5 models. In GFDL CM3, positive sea level pressure trends in the SH are located farther north than in the other models, similar to the response of zonal wind trends in that model. MME-mean sea level pressure trends in RCP45 minus RCP45fixAA in the SH are rather similar to those in RCP45fixAA; this suggests that without recovering ozone in RCP45fixAA, which opposes the trends induced by increasing WMGHGs, there would be a stronger positive SAM trend due to the effects of increasing WMGHGs than the effects of declining aerosols.

In the NH, both RCP45fixAA and RCP45 minus RCP45fixAA show a suggestion of a positive trend in the northern annular mode (NAM) in the MME mean, with negative trends in sea level pressure at high latitudes relative to the subtropics. (The NAM can be defined as the difference of sea level pressure between 65° and 35°N; Li and Wang 2003; Gillett and Fyfe 2013.) However, it is much weaker than the corresponding trend in the SH, and in some models (especially CanESM2) it is absent.
Previous studies have shown that the sea level pressure response to increasing WMGHGs projects strongly onto the SAM (Fyfe et al. 1999; Kushner et al. 2001) and a positive change in the SAM is consistent across models (e.g., Arblaster et al. 2011). On the other hand, the response to increasing WMGHGs projects only weakly onto the NAM (Fyfe et al. 1999; Woollings 2008) and the change in the NAM is model dependent (Shindell et al. 1999; Miller et al. 2006). Gillett and Fyfe (2013) found that CMIP5 models generally simulate positive projected changes in the SAM and NAM under RCP4.5, with larger changes in the SAM than the NAM. A stronger response of the SAM than the NAM is consistent with what we observe here (for declining aerosols as well as increasing WMGHGs).

d. Further discussion

There is a “chicken and egg” relationship between temperature and circulation changes, so use of the thermal wind relationship does not (in itself) demonstrate cause and effect. However, since tropospheric temperature trends were broadly consistent with the meridional structure of aerosol radiative forcing, thermal wind balance provides a satisfying explanation for the overall strengthening of the SH jet and the poleward shift of the NH jet in response to declining aerosols. Regarding the NH jet, our explanation is consistent with Ming et al. (2011) and Allen et al. (2012a,b). It is harder to explain the tendency for the SH zonal winds throughout the troposphere to strengthen predominantly on the poleward flank of the climatological westerlies (giving a positive trend in the SAM), since it is not clear why the aerosol forcing should induce a stronger zonal wind response at, for instance, 55°S than farther north.

It may be that the mechanism underlying the poleward shift of the SH westerlies is similar for increasing WMGHGs and declining aerosols; indeed, this was the conclusion of Rotstayn (2013), who noted the similarity of circulation changes associated with the positive SAM trend induced by increasing WMGHGs and declining aerosols in CSIRO Mk3.6. The dynamical mechanisms that cause a poleward shift of the SH westerlies in response to increasing WMGHGs are complex and are still a topic of active research. Fundamentally, it is related to the change in the latitude of eddy momentum flux convergence (Chen and Held 2007; Chen et al. 2008). Some authors have argued that the chain of events is triggered by upper-tropospheric warming and lower-stratospheric cooling, which (because of the sloping tropopause) accelerates the zonal winds near the tropopause (Chen and Held 2007; Chen et al. 2008). Other studies have focused on meridional heating gradients in the lower troposphere (Butler et al. 2011; Allen et al. 2012b) or the meridional SST gradient (Lu et al. 2010; Chen et al. 2010) as the primary trigger for the change in eddy flux convergence. Another line of research has considered the effect of increased static stability in the midlatitude troposphere (Frierson 2008; Lu et al. 2008). From an aerosol perspective, it is easier to rationalize mechanisms related to meridional heating or temperature gradients than to an increase of midlatitude static stability, since aerosol radiative forcing is quite small in the SH extratropics.

As discussed by Rotstayn (2013), the mechanisms by which aerosol changes affect stratospheric temperatures are currently unclear. All four models show this to some extent, although the details differ. For example, GFDL CM3 does not give a cooling of the SH polar stratosphere in response to declining aerosols, and this may be linked to the more northward location of increasing westerly winds in that model. It should be noted that cause and effect are not obvious; while it is tempting to assume that the different temperature response of
GFDL CM3 causes the dynamical response to differ, it is also possible that the temperature response is caused by the different circulation change. For example, Thompson and Wallace (2000) showed that positive fluctuations in the SAM on interannual time scales are associated with lower temperatures in the polar stratosphere (their Fig. 7c); they attributed this to adiabatic cooling associated with anomalous ascent. Considering direct radiative arguments as an explanation for stratospheric cooling due to declining aerosols, it is possibly caused by decreasing shortwave absorption by black carbon, although it is also possible that this effect is overestimated. There are other plausible explanations, such as changes in stratospheric water vapor (Rotstayn 2013). This is an interesting topic for further research.

A caveat concerning this study is that none of the models treat nitrate aerosol. Nitrate is expected to offset declines in other aerosol species, since emissions of ammonia from agriculture are assumed to be insensitive to emission controls in the RCPs (van Vuuren et al. 2011; Bellouin et al. 2011; Lamarque et al. 2011). However, few CMIP5 models treat nitrate; Collins et al. (2014) (their Table 12.1) list only 6 models out of 47 that included nitrate. So in this respect our results are broadly representative of most CMIP5 models.

The models in this study all have substantial negative aerosol ERF at the top of the atmosphere (from $-0.7$ to $-1.6 \text{W m}^{-2}$ for 2000 relative to 1850). We have not explicitly considered the role of solar absorption by black carbon, which is generally underestimated by CMIP5 models (Shindell et al. 2013; Allen et al. 2013). Allen et al. (2012a) showed that solar absorption by black carbon and tropospheric ozone causes a northward expansion of the tropics and the NH jet during 1979–99 in CMIP3 models. Also, Allen and Ajoku (2014, manuscript submitted to Geophys. Res. Lett.) found that in twenty-first-century projections with the Community Atmospheric Model, declining sulfate and declining black carbon exert opposing effects on large-scale NH circulation. Thus it is possible that stronger black carbon absorption in the models used here would tend to drive an equatorward shift of the NH jet in response to declining aerosols (i.e., an opposite response to what we found).

All four models used in this study include a treatment of indirect aerosol effects: two include only the cloud-albedo effect, and two include both the cloud-albedo and cloud-lifetime effects. Table 12.1 in Collins et al. (2014) shows that a number of CMIP5 models (15 out of 47 listed) do not treat indirect aerosol effects. These models are likely to have weaker aerosol radiative forcing (and climatic response) than the models considered here.

There are other aerosol–cloud interactions that may be climatically important and are not resolved by the models in this study. For example, cloud-resolving simulations by Zhang et al. (2007) suggested that pollution aerosols from Asia have caused an intensification of the winter storm track over the North Pacific Ocean. The mechanism involves interactions between aerosols and deep convection, which are generally not treated in GCMs. Note that the extent to which aerosols invigorate (i.e., deepen) convective clouds is very uncertain; see Altaratz et al. (2014) for a recent review.

4. Summary and conclusions

We compared the projected effects of declining aerosols in RCP4.5 using four models, with a focus on annual, zonal-mean atmospheric temperature structure and zonal winds. We interpreted the response in terms of the meridional structure of aerosol radiative forcing, which peaks near $40^\circ$N and vanishes at the South Pole. Our main emphasis was on features that were consistent across a majority of models, although we also considered differences among the models. Consistent aspects of the response to declining aerosols are summarized schematically in Fig. 12, and in more detail in the next few paragraphs.

To first order, declining aerosols warm the troposphere in a manner akin to the effects of increasing
WMGHGs: similar features include amplified warming in the tropical upper troposphere and near the Arctic surface, and less warming in the extratropical lower troposphere in the SH. These common features are due to feedbacks that act similarly in response to both forcing agents (Xie et al. 2013). It should be noted that feedbacks are not expected to be identical for changes in aerosols and WMGHGs: Shindell (2014) recently showed that the larger weighting of aerosol forcing toward the NH extratropics provokes stronger feedbacks, and a larger change in surface temperature per unit forcing for aerosols than WMGHGs.

The effect of the meridional structure of aerosol forcing is also seen in the tropospheric temperature response to declining aerosols, which has a different structure to that caused by increasing WMGHGs. Relative to the pattern of warming caused by increasing WMGHGs, declining aerosols cause more warming in NH midlatitudes, and less warming in the SH, especially at high latitudes. Averaged between the surface and 250 hPa, the maximum warming due to declining aerosols is at 35°N (whereas the warming due to increasing WMGHGs shows a broad tropical maximum). As a consequence, declining aerosols cause the MTG in the troposphere to generally increase in the SH, whereas in the NH it decreases in the tropics and subtropics.

Declining aerosols cause the midlatitude jets to strengthen in both hemispheres, but more substantially in the SH. In the NH, the jet strengthens on its upper and poleward flank, but weakens on its lower and equatorward flank, so it shifts poleward and upward. We showed that these effects are broadly consistent with thermal wind balance. Thus the jet strengthens more in the SH than the NH because aerosol forcing has its largest magnitude in the NH midlatitudes, and is relatively flat between the equator and 40°N, whereas in the SH aerosol forcing generally decreases between the equator and the pole. The meridional structure of aerosol forcing is then reflected in the response of the MTG and the midlatitude jets.

The response to declining aerosols in the SH shows increasing sea level pressure at midlatitudes and decreasing sea level pressure at high latitudes (i.e., a positive trend in the southern annular mode). In the NH there is a suggestion of a positive trend in the northern annular mode in the MME mean, but the response is not consistent among the models. In the SH, the effects of increasing WMGHGs and declining aerosols on annual-mean zonal winds and the annular mode are broadly additive. The picture is more complex in the NH, because the jet is roughly collocated with the latitude of maximum aerosol forcing.

Changes in circulation (zonal winds and sea level pressure) caused by declining aerosols are of comparable magnitude to those caused by increasing WMGHGs in the MME mean, even though corresponding temperature changes are substantially smaller: global-mean surface air temperature trends are 1.6°C century⁻¹ in RCP45fixAA and 0.8°C century⁻¹ in RCP45 minus RCP45fixAA, and tropospheric temperature trends are also smaller in Fig. 4b than Fig. 4a. Based on the above discussion, the explanation is that trends in the MTG are of comparable magnitude for declining aerosols and increasing WMGHGs.

The jets are a source of baroclinic instability, so the effects of declining aerosols on the jets may be important for understanding projected changes in storm tracks and precipitation. For example, Chang et al. (2012) compared projected changes in storm tracks (based on filtered meridional wind variance) from CMIP5 and CMIP3. In the SH they found similar results in CMIP3 and CMIP5, namely a poleward and upward shift of the storm track. However, in the NH the CMIP5 projections differed substantially from CMIP3. CMIP5 models projected a modest poleward and upward shift of the storm track, but mainly a weakening of the storm track on its lower and equatorward flank (which was much more pronounced in CMIP5 than CMIP3). Although Chang et al. (2012) did not consider the role of aerosols, the qualitative similarity of their CMIP5 NH storm track changes to our results for the NH jet suggests that declining aerosols contribute to the weakening NH storm track in CMIP5 projections.

The sensitivity of the SAM to changing aerosols is intriguing. Long-term trends in the SAM are thought to affect precipitation in the middle-to-high latitudes of the SH (Fyfe et al. 2012), the Southern Ocean carbon sink (Lenton et al. 2009), and Antarctic sea ice extent (Sigmond and Fyfe 2014). Thus, aerosol forcing, which is principally located in the NH, is likely to be important for understanding climate change in the most remote and pristine parts of the SH.

We analyzed the annual- and zonal-mean response to declining aerosols, and found both similarities to and differences from the effects of increasing WMGHGs, which are relatively well known. We did not consider seasonal effects or zonal variations in circulation; the latter are especially important in the NH (e.g., Barnes and Polvani 2013). Studies of the effects of historically increasing aerosols, such as those reviewed in section 1, suggest that tropical and regional circulation will also show a distinct response to declining aerosols in climate projections. These are interesting topics for further research.

The magnitude of aerosol forcing varies substantially among the four models we considered, and so does the response. Global-mean historical aerosol ERF ranges
from $-0.7 \text{ W m}^{-2}$ in IPSL-CM5A-LR to $-1.6 \text{ W m}^{-2}$ in GFDL CM3. Although aerosol ERF for the period 2006 to 2100 in RCP4.5 is not available, the global-mean surface air temperature change due to declining aerosols is larger in the models with stronger aerosol ERF in the historical period (CSIRO Mk3.6 and GFDL CM3) than in IPSL-CM5A-LR or CanESM2. This is consistent with the argument that historical aerosol ERF (with sign reversed) is a reasonable proxy for twenty-first-century aerosol ERF (Rotstayn et al. 2013). Trends in atmospheric temperature and zonal winds were also generally stronger in the models with stronger aerosol forcing.

However, the range of aerosol forcing in these models does not fully capture the uncertainty, even in the global mean. The uncertainty is even larger when considering regional effects. In view of this, and the fact that the RCPs do not adequately sample the phase space of possible future aerosol emissions, systematic efforts are needed to investigate the role of declining aerosols in climate projections.

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