Anthropogenic effects on the subtropical jet in the Southern Hemisphere: aerosols versus long-lived greenhouse gases

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Abstract
We use single-forcing historical simulations with a coupled atmosphere–ocean global climate model to compare the effects of anthropogenic aerosols (AAs) and increasing long-lived greenhouse gases (LLGHGs) on simulated winter circulation in the Southern Hemisphere (SH). Our primary focus is on the subtropical jet, which is an important source of baroclinic instability, especially in the Australasian region, where the speed of the jet is largest. For the period 1950 to 2005, our simulations suggest that AAs weaken the jet, whereas increasing LLGHGs strengthen the jet. The different responses are explained in terms of thermal wind balance: increasing LLGHGs preferentially warm the tropical mid-troposphere and upper troposphere, whereas AAs have a similar effect of opposite sign. In the mid-troposphere, the warming (cooling) effect of LLGHGs (AAs) is maximal between 20S and 30S; this coincides with the descending branch of the Hadley circulation, which may advect temperature changes from the tropical upper troposphere to the subtropics of the SH. It follows that LLGHGs (AAs) increase (decrease) the mid-tropospheric temperature gradient between low latitudes and the SH mid-latitudes. The strongest effects are seen at longitudes where the southward branches of the Hadley cell in the upper troposphere are strongest, notably at those that correspond to Asia and the western Pacific warm pool.

Keywords: global climate modeling, climatology, climate change and variability, particles and aerosols, Asia, Australia, Southern Hemisphere, subtropical jet

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1. Introduction
A prominent feature of the winter circulation in the Southern Hemisphere (SH) is the upper-tropospheric subtropical jet (Van Loon 1972). It is strongest near 30S over Australia and the western Pacific Ocean, where it provides an important source of baroclinic instability (Hoskins and Hodges 2005). A weakening of the jet in recent decades has been identified in reanalyses, and the observed winter rainfall decline in south-western Western Australia has been linked to the weaker jet (Frederiksen and Frederiksen 2007, Frederiksen et al 2011).
An important question is the role of anthropogenic forcing in driving such circulation changes. Increasing concentrations of long-lived greenhouse gases (LLGHGs) are thought to induce a strengthening and poleward shift of the SH jet streams (Kushner et al. 2001). The importance of stratospheric ozone depletion for SH mid-latitude circulation changes has also been recognized for some time, especially in summer and autumn (Thompson and Solomon 2002).

Changes in LLGHGs and stratospheric ozone are both thought to drive the Southern Annular Mode (SAM) towards a more positive state, and the effects of ozone depletion in austral summer may be substantially larger than those of LLGHGs in recent decades (Shindell and Schmidt 2004, Arblaster and Meehl 2006, Polvani et al. 2011). However, ozone effects are thought to be much smaller in winter than summer (Miller et al. 2006).

Simulations suggest that the jet streams and storm tracks in the Northern Hemisphere (NH) tend to shift equatorward in response to the cooling effects of AAs (Kristjánsson et al. 2005, Ming et al. 2011). However, possible effects of anthropogenic aerosols (AAs) on SH mid-latitude circulation have had little attention, perhaps because AAs are mostly concentrated in the NH. Arblaster and Meehl (2006) analysed the response of the SAM to different forcing agents in an atmosphere–ocean global climate model (AOGCM). They found a relatively weak response to AAs, though the only aerosol effects in their model were the direct radiative effects of sulfate. Using a low-resolution AOGCM with direct and indirect aerosol effects, Cai and Cowan (2007) focused on annual-mean AA-induced changes in the SH. They found a southward shift of oceanic and atmospheric circulations and a positive trend in the SAM, and argued that these represented a positive air–sea feedback initiated principally by the oceanic changes.

From an atmospheric perspective, in austral winter strong AA radiative forcing in the NH is collocated with the upward branch of the Hadley circulation. The Hadley circulation weakens in response to AA forcing in this season, both in atmospheric GCMs (Williams et al. 2001) and AOGCMs (Rotstayn et al. 2012). Since the primary descending branch of the Hadley cell is in the SH in JJA (e.g., Baines 2006), this suggests the possibility of AA-induced effects on atmospheric circulation in the SH.

Baines and Folland (2007) documented an abrupt climate shift across the late 1960s, which was most evident in austral winter. The extensive list of changes included an interhemispheric difference in SST (with a relatively cooler NH), patterns of tropical rainfall such as a drying in the Sahel, changes in various branches of the Hadley circulation and a weakening of the subtropical jet stream in the SH. They suggested that the most likely causes of the shift were a quasi-interhemispheric oscillation of temperature variability on multidecadal timescales (centred on the Atlantic region), or increases of AAs; for further discussion, see the recent review by Chiang and Friedman (2012).

There is an interesting analogy with the palaeoclimate modelling study by Lee et al. (2011), who used an atmospheric GCM to simulate the atmospheric response to an imposed cooling within the glacial North Atlantic, representative of the last glacial termination. They found that the southern branch of the Hadley circulation weakened, and this caused the SH subtropical jet to weaken, which in turn drove stronger sub-polar westerlies over the Southern Ocean. An interesting implication of this study is that extra-tropical cooling in the NH can affect the Hadley circulation, and potentially SH circulation, through the development of an interhemispheric temperature gradient (Chiang and Friedman 2012).

Here we use single-forcing simulations with a coupled AOGCM to compare the simulated effects of AAs and increasing LLGHGs on the SH subtropical jet. Complex aerosol processes are very simplified in AOGCMs (e.g., Rotstayn et al. 2007), and the radiative forcing due to AAs is highly uncertain, even in the global mean (Forster et al. 2007). Also, as indicated above, the effects of AAs in driving recent observed climatic changes can be difficult to disentangle from natural variability, and the magnitudes of reanalysed circulation changes are themselves uncertain, especially in the SH (Marshall 2003). For these reasons, we do not attempt a formal attribution of changes of the subtropical jet in this letter. Rather, we contrast the effects of changes in AAs and LLGHGs in a single AOGCM, and discuss the implications. We find that the jet weakens in response to AAs and strengthens in response to increasing LLGHGs. Our results suggest that changes in AAs are a significant driver of changes in the jet, and potentially of changes in baroclinicity and rainfall in the subtropics and mid-latitudes of the SH.

2. Model and experiments

CSIRO Mark 3.6 (CSIRO-Mk3.6) is a coupled atmosphere–ocean model with dynamic sea ice. It was developed from the earlier Mk3.5 version (Gordon et al. 2010). The main differences between Mk3.5 and Mk3.6 are the inclusion of an interactive aerosol treatment and an updated radiation scheme in Mk3.6. The model treats the direct radiative effects of sulfate, organic carbon, black carbon, dust and sea salt and the indirect effects of sulfate, sea salt and carbonaceous aerosol on liquid–water clouds. CSIRO-Mk3.6 is described in more detail by Rotstayn et al. (2012), who also evaluate the aerosol simulation with regard to observations and other models.

The anthropogenic (i.e., 2000 minus 1850) aerosol radiative forcing in CSIRO-Mk3.6 is $-1.4 \text{ W m}^{-2}$ in the annual mean (Rotstayn et al. 2012). This includes direct radiative effects, the first and second indirect effects on liquid–water clouds, and the effect of black carbon on snow albedo. As in earlier studies, the forcing is substantially stronger in the NH than in the SH (Rotstayn et al. 2012, their figure 2).

A substantial number of historical experiments were performed with CSIRO-Mk3.6 as part of CMIP5 (Jeffrey et al. 2012). We use a subset of these experiments, each of which is a 10-member ensemble of transient runs for the period 1850–2005:

- HIST: standard historical run with ‘all forcings’, namely LLGHGs, ozone, AAs, volcanic and solar forcing.
Figure 1. Zonal-mean JJA temperature trends (shaded) for 1950–2005 (K per century) from (a) GHGAS, (b) HIST minus NoAA. The mean 1950–2005 field from HIST is contoured. Stippling denotes trends significant at 5%.

- NoAA: same as HIST, but with AA emissions fixed at 1850 levels.
- GHGAS: same as HIST, but forced only by historical changes in LLGHGs.

The effect of anthropogenic aerosol changes (hereafter ‘AA-induced’ effect) is diagnosed from HIST minus NoAA. The AA-induced effect is likely to be non-linear, in the sense that HIST minus NoAA is not identical to an experiment forced only by anthropogenic aerosols. However, using HIST minus NoAA may be more realistic, because in the real world AAs act in a warming climate. Experiment GHGAS is included so that the effects of changes in LLGHGs can be contrasted with those of AAs.

We calculate ensemble-mean trends for June–August (JJA) 1950–2005, using ordinary least squares. As in Rotstayn et al (2012), statistical significance of ensemble means is assessed using a two-sided $t$-test, taking the 10 individual trend values as independent data points. For the difference of two ensemble means at each grid point (i.e. HIST minus NoAA), a two-sample $t$-test for the difference of the mean of two populations is used.

3. Results

Since the upper-tropospheric circulation is central to some of our arguments, it is useful to briefly evaluate this aspect of the simulation. Figure S1 (available at stacks.iop.org/ERL/8/014030/mmedia) compares climatological fields of JJA zonal wind at 300 hPa from the HIST ensemble and the NCEP-DOE Reanalysis 2 (NCEP2; Kanamitsu et al 2002). The SH subtropical jet is represented rather well by the model, including its enhancement in the Australasian region. Compared to NCEP2, the jet in the model is a little too strong, e.g. it exceeds 45 m s$^{-1}$ over Australia and the western Pacific, but in the reanalysis it just touches 45 m s$^{-1}$ in a small area to the north of New Zealand.

Figure S2(a) (available at stacks.iop.org/ERL/8/014030/mmedia) shows climatological fields of JJA meridional wind at 200 hPa from the HIST ensemble. The strongest southward branch of the Hadley circulation occurs at longitudes that correspond to the Asian continent and the warm pool in the tropical western Pacific; this suggests that the strong heating over Asia and the warm pool is likely to be the cause of the strong upper-level southward flow at similar longitudes (Baines 2006). A similar feature is seen in NCEP2, shown in figure S2(b) (available at stacks.iop.org/ERL/8/014030/mmedia); compare also Baines (2006) and Baines and Folland (2007). Weaker southward branches exist near South America and Africa. Some distortion of the Hadley circulation is seen in the model, e.g. to the northeast of Australia. Also, the southward flow near South America is too weak, and displaced to the north compared to NCEP2.

The model’s simulation of mid-tropospheric vertical velocity (omega) also qualitatively captures the main features seen in NCEP2 (figure S3 available at stacks.iop.org/ERL/8/014030/mmedia). Some errors are seen, especially over the equatorial western Pacific, where the model’s sea-surface temperature field (not shown) suffers from the well-known cold-tongue bias (e.g., Zheng et al 2012). Regions of strong ascent over Southeast Asia and the tropical north-western Pacific are qualitatively captured by the model, as are the subsidence regions in the subtropics of the SH.

We now present temperature trends from GHGAS and HIST minus NoAA, followed by trends in zonal winds in the SH. We will interpret the wind trends in terms of thermal wind balance.

3.1. Temperature trends

Figure 1 shows trends in zonally averaged temperature from GHGAS and HIST minus NoAA. The enhanced warming in the tropical upper troposphere in GHGAS (panel a) is a robust feature of GCM simulations forced by increasing concentrations of LLGHGs, because tropical convection tends to adjust the troposphere towards the moist adiabatic lapse rate (Wilson and Mitchell 1987). To first order, the tropical pattern in HIST minus NoAA (panel b) resembles the inverse of that in GHGAS, with enhanced cooling in the tropical upper troposphere; see also Ming et al (2011) (their figure 2). This is expected, because aerosols cool the surface, and the moist adiabatic lapse rate amplifies the effect in the upper troposphere. (Note that the cooling of the tropical troposphere in figure 1(b) is generally smaller than the warming in panel a, so the sum of changes due to LLGHGs and AAs would be a net warming.)
We now compare the spatial patterns of mid-tropospheric temperature trends in more detail (figure 2). The climatological field from HIST is also plotted; note that the warmest region at 500 hPa is collocated with the Asian monsoon. It is noteworthy that in GHGAS (HIST minus NoAA) the maximum warming (cooling) trend occurs just south of 20S in the zonal mean (right panels). This is not seen in the annual mean, in which the mid-tropospheric temperature trends are much more symmetric about the Equator (figure S4 available at stacks.iop.org/ERL/8/014030/mmedia), they are relatively small in both experiments: the largest changes over 1950–2005 are only of order 2% of the mean value in this latitude range. This suggests that the large temperature trend at 500 hPa over the Indian Ocean and Australia (figure 2) may be more related to advection of temperature anomalies by the mean Hadley circulation and/or (2) anomalies in the Hadley circulation induced by changes in LLGHGs or AAs. Between 20S and 30S, the trends in mid-tropospheric vertical velocity (figure S6(a) available at stacks.iop.org/ERL/8/014030/mmedia) are relatively small in both experiments: the largest changes over 1950–2005 are of order 2% of the mean value in this latitude range. This suggests that the large temperature trend at 500 hPa over the Indian Ocean and Australia (figure 2) may be more related to advection of temperature anomalies by the mean Hadley circulation, rather than local changes in the Hadley circulation. This is illustrated in figure S7 (available at stacks.iop.org/ERL/8/014030/mmedia), which shows temperature trends from both experiments, zonally averaged between 60E and 150E, overlaid on streamlines representing the mean meridional wind in the latitude range. In both panels there is an area between roughly 20S and 30S, where the strong tropical upper-tropospheric temperature trends extend downward into the mid-troposphere; these ‘descending lobes’ are collocated with the subsiding branch of the Hadley cell.

The temperature trends in figure 2 show substantial departures from the zonal means. In GHGAS, the largest temperature increases between 20S and 30S occur in a wide band that extends across the Indian Ocean and Australia, and a smaller band (slightly further north) in the vicinity of South America. In HIST minus NoAA, the largest aerosol-induced temperature decreases also occur in these regions. The variation with longitude of the temperature response in figure 2 is generally similar to the zonal structure of the Hadley circulation, which is especially strong south of Asia, and to a lesser extent near South America (figure S2 available at stacks.iop.org/ERL/8/014030/mmedia). One exception is the strong cooling in the central South Pacific in HIST minus NoAA, which is not associated with a southward branch of the upper-level Hadley circulation; rather, it appears to be caused by a positive trend in meridional wind speed in that region, which advects cooler air from higher latitudes (not shown).

3.2. Possible mechanism for temperature changes

The strong mid-tropospheric temperature response between 20S and 30S over the Indian Ocean and Australia in both experiments is intriguing. It is collocated with the descending part of the strongest southward branch of the upper-level Hadley circulation (figures S2 and S3 available at stacks.iop.org/ERL/8/014030/mmedia). This suggests that temperature trends in the corresponding ascending branch in the NH may be a key part of the associated mechanism. In fact, the strongest simulated temperature trends at 200 hPa in JJA occur over Southeast Asia and the tropical north-western Pacific in both experiments (figure S5 available at stacks.iop.org/ERL/8/014030/mmedia); this is roughly across the equator from the strong mid-tropospheric temperature response between 20S and 30S. If air parcels at 20S to 30S are advected via the Hadley circulation, then in GHGAS, increased potential temperature due to diabatic heating in the ascending branch would be conserved in the absence of changes in other diabatic processes. The converse would apply in HIST minus NoAA.

If the large mid-tropospheric temperature response between 20S and 30S is linked to the Hadley circulation, then it could be related to (1) advection of temperature anomalies by the mean Hadley circulation and/or (2) anomalies in the Hadley circulation induced by changes in LLGHGs or AAs. Between 20S and 30S, the trends in mid-tropospheric vertical velocity (figure S6(a) available at stacks.iop.org/ERL/8/014030/mmedia) are relatively small in both experiments: the largest changes over 1950–2005 are of order 2% of the mean value in this latitude range. This suggests that the large temperature trend at 500 hPa over the Indian Ocean and Australia (figure 2) may be more related to advection of temperature anomalies by the mean Hadley circulation, rather than local changes in the Hadley circulation. This is illustrated in figure S7 (available at stacks.iop.org/ERL/8/014030/mmedia), which shows temperature trends from both experiments, zonally averaged between 60E and 150E, overlaid on streamlines representing the mean meridional circulation in this longitude range. In both panels there is an area between roughly 20S and 30S, where the strong tropical upper-tropospheric temperature trends extend downward into the mid-troposphere; these ‘descending lobes’ are collocated with the subsiding branch of the Hadley cell.

Although trends in mid-tropospheric vertical velocity are relatively small in the subtropics of the SH (figure S6(a) available at stacks.iop.org/ERL/8/014030/mmedia), they are substantial in the equatorial region. For example, north of the equator, trends (per century) in both experiments reach magnitudes of more than 10 hPa d^{-1}, which is more than 25% of the mean value. These tropical circulation changes are coupled to trends in rainfall and latent heating, which are...
discussed below. Importantly, there are also associated trends in meridional wind in the tropical upper troposphere, with a strengthening (weakening) of the climatological southward flow at 200 hPa in GHGAS (HIST minus NoAA) (figure S6(b) available at stacks.iop.org/ERL/8/014030/mmedia). These changes act in the right direction to contribute to the mid-tropospheric warming (cooling) in the SH subtropics in GHGAS (HIST minus NoAA).

Why are the strongest upper-tropospheric temperature trends seen over Southeast Asia and the tropical north-western Pacific? At the surface, there is relatively strong warming (cooling) over Asia in GHGAS (HIST minus NoAA); see Rotstayn et al (2012) (their figure 15). In GHGAS, this is due to rapid transient warming of the Asian continent, whereas in HIST minus NoAA it is due to strong (negative) regional aerosol radiative forcing. In the upper troposphere, temperature trends in this region also reflect changes in convective latent heating associated with rainfall trends. Simulated rainfall trends for boreal summer are shown in figure S8 (available at stacks.iop.org/ERL/8/014030/mmedia). GHGAS (HIST minus NoAA) simulates a substantial increase (decrease) of rainfall over Southeast Asia and the tropical western Pacific during the last few decades; the rainfall increase in GHGAS is of the same sign as the 21st century multi-model mean change from Meehl et al (2007), and the AA-induced rainfall decrease is broadly consistent with several other modelling studies (e.g., Bollasina et al 2011, Ganguly et al 2012).

The above argument may explain why such a strong mid-tropospheric temperature response is seen at 20S–30S over the Indian Ocean and Australia: In essence, changes of convective latent heating in the ascending branch of the Hadley cell are approximately conserved in the dry descending branch. Furthermore, changes in the strength of the southward upper branch of the Hadley cell tend to enhance the effect in both experiments.

A rigorous evaluation of this argument would require calculation of changes in all the terms in the temperature tendency equation, some of which were not saved in our simulations. According to Rodwell and Hoskins (2001), the dominant temperature balance in the subtropical descent regions is between subsidence-induced (adiabatic) warming and cooling by horizontal advection and diabatic effects, with relatively small contributions from eddy heat flux convergence. In our simulations, trends in meridional eddy heat flux convergence equatorward of 30S are an order of magnitude smaller than trends in $\nabla \cdot VT$, where $V$ and $T$ are respectively the monthly mean three-dimensional wind vector and temperature (not shown). This suggests that it is valid to explain temperature changes between 20S and 30S in terms of the mean circulation and its changes.

### 3.3. Trends in zonal winds

Figure 3 shows zonal-mean trends in zonal wind speed in the SH from both experiments. In GHGAS (figure 3(a)), there is a significant strengthening of the subtropical jet, as well as a small upward and poleward shift. This resembles a weaker version of the CMIP3 multi-model ensemble-mean change shown in response to stronger 21st century forcing by Lorenz and DeWeaver (2007) (their figure 3(c)). Conversely, the AA-induced effect on the subtropical jet (figure 3(b)) is a significant weakening, as well as a small downward and equatorward shift. The opposing effects of LLGHGs and AAs on the subtropical jet are qualitatively consistent with the temperature changes shown in the right panels of figure 2, since the increase of westerly wind with height is in thermal wind balance with the meridional temperature gradient.

Figure 4 shows the spatial pattern of trends in zonal wind speed at 300 hPa, together with the mean climatological field from HIST. There is substantial zonal asymmetry, reflecting the zonal asymmetry in mid-tropospheric temperature trends (figure 2). The regions of strongest decrease in zonal wind in HIST minus NoAA coincide with strong temperature changes, in the manner expected from thermal wind balance, i.e., large wind changes occur on the flanks of regions of large temperature change. Over the Australasian region, the changes in the subtropical jet in both experiments are most evident on its poleward flank.

We have explained the response of the subtropical jet in terms of thermal wind balance, but it must also be possible to explain it in terms of the momentum equations. Lee et al (2011) used the quasi-geostrophic zonal momentum equation to qualitatively explain the zonal-mean response of the subtropical jet in their GCM in terms of changes in
upper-tropospheric meridional wind. Following their notation,

\[
\frac{\partial \bar{u}}{\partial t} = f \bar{v} - \frac{\partial (\bar{u}'v')}{\partial y} + \bar{X} ,
\]

where the overbar denotes zonal means, and primes denote the eddy terms; \( u \) and \( v \) are respectively the zonal and meridional winds, \( f \) is the Coriolis parameter, and \( X \) represents unresolved turbulent effects (‘friction’). Equation (1) shows that (instantaneously) the mean meridional circulation (\( f\bar{v} \)) accelerates the subtropical jet, and the divergence of momentum flux by transient eddies (\( \partial (\bar{u}'v')/\partial y \)) decelerates it; long-term trends can arise from relatively small changes in the balance between these effects. Figure S9 (available at stacks.iop.org/ERL/8/014030/mmedia) shows trends in these two terms at 300 hPa from HIST minus NoAA; equatorward of about 40\(^\circ\)S, changes in \( f\bar{v} \) tend to drive a weakening of the jet, while eddies have a damping effect on these changes. Although Lee et al (2011) were not able to calculate the effects of eddies, our conclusion regarding the effect on the jet of changes in the mean meridional circulation is similar to theirs.

Equation (1) is valid only in the zonal mean, and cannot be used to examine regional trends in terms of momentum balance. In principle, analysis of the momentum budget in the primitive equations would explain regional trends in the jet, although we do not have access to all the necessary terms from our runs. With this in mind, it is noteworthy that Berbery and Nogués-Paegle (1993) linked accelerations of the Australasian subtropical jet to enhanced poleward flow in the upper troposphere, associated with local Hadley cells driven by enhanced monsoonal convection in the NH.

4. Discussion and conclusion

Our results suggest that the largest mid-tropospheric temperature response to forcing from LLGHGs and AAs in JJA occurs in the subtropics of the SH, especially in the vicinity of Australia. In the case of AAs, this is possibly counter-intuitive, because AAs are concentrated mostly in the NH. However, we showed that these largest responses were collocated with strong southward branches of the Hadley circulation, and argued that this circulation advects changes in upper-tropospheric temperature in the NH tropics to the subtropics of the SH. A similar argument was used recently by Haarsma et al (2012), to explain a region of simulated maximum warming in the NH subtropics over the eastern Atlantic in boreal winter.

Attribution of changes in the SH subtropical jet is complicated by the so-called 1960s climate shift, which was initiated by a cooling of the NH relative to the SH. Baines and Folland (2007) concluded that the most likely causes of the shift were a (natural) reduction in the northward heat transport by the Atlantic thermohaline circulation, and a coincident rapid increase of AAs in the vicinity of Europe and North America. Based on models from CMIP3, Chang et al (2011) estimated that roughly half of the observed 20th Century trend in the Atlantic interhemispheric temperature gradient is due to a forced response, most probably to sulfate aerosols. They also noted that models that include aerosol indirect effects...
simulated stronger trends than models that treat direct effects only. Recently, Booth et al (2012) found that in a CMIP5 AOGCM with a more detailed aerosol treatment than found in most CMIP3 models, aerosols were the most important driver of 20th century North Atlantic SST variability. However, the relative importance of natural variability and anthropogenic forcing is still uncertain (Chiang and Friedman 2012).

Considering the possible role of AAs on the subtropical jet, it is interesting that the model assessed by Frederiksen et al (2011) as best able to capture the change in the jet—MIROC3.2(medres)—includes a diagnostic aerosol scheme, which treats the first and second indirect effects as well as direct effects of several aerosol species (Takemura et al 2005). This is in marked contrast to many CMIP3 models, which only included very rudimentary aerosol treatments.

On the other hand, the influence of natural (unforced) decadal variability may also explain why Frederiksen et al (2011) found that none of the CMIP3 models were able to capture the full extent of the reanalysed changes in 300 hPa zonal wind strength since the late 1960s. This is also true for CSIRO-Mk3.6 in the HIST ensemble mean; figure S10 (available at stacks.iop.org/ERL/8/014030/mmedia) shows zonal-mean trends in zonal wind speed from HIST (i.e., in response to ‘all forcings’). The subtropical jet does weaken in the upper troposphere, but the effect is smaller than in figure 3(b), and not statistically significant.

Our 10-member ensemble-mean trends do not show the extent of natural decadal variability, such as that associated with changes in the Atlantic Multidecadal Oscillation. Although it is beyond the scope of this letter to address the effects of decadal variability in detail, it is interesting to briefly consider trends in zonal wind speed at 300 hPa from individual runs. These are shown in figure S11 (available at stacks.iop.org/ERL/8/014030/mmedia) for GHGAS and figure S12 (available at stacks.iop.org/ERL/8/014030/mmedia) for HIST minus NoAA (with contour levels doubled relative to those in figure 4). There are large variations in the trends among the individual runs that comprise each experiment, even in the zonal means. This is a good reminder that observed trends over the last few decades may strongly reflect natural variability, and do not necessarily represent the forced response of the climate system.

We have considered changes in the subtropical jet, but it is important to note that baroclinicity depends on static stability as well as wind shear. For example, the Phillips criterion used by Frederiksen et al (2011) comprises the difference of two such terms. A warmer climate is expected to enhance static stability in the mid-latitudes (Frierson 2008, Lim and Simmonds 2009), so one expects increasing LLGHGs to broadly increase (decrease) static stability. Thus changes in baroclinicity in the future climate may represent a subtle interaction between competing effects; this analysis is deferred to a later study.

One caveat is that our results are based on only a single model; an important extension of this work will be to compare such ‘single forcing’ simulations across a range of CMIP5 models. For example, our suggested mechanism involved simulated rainfall changes in the tropics of the NH, which are likely to be model-dependent. Another caveat is the large uncertainty about aerosol forcing, which could easily be under- or over-estimated in our simulations.

In summary, our results suggest that AAs have contributed to a weakening of the SH subtropical jet, especially in the vicinity of Australia (and, to a lesser extent, South America). This is in marked contrast to increasing LLGHGs, which tend to strengthen the jet. Understanding such effects is important, as LLGHGs are expected to increase, while AAs are projected to decrease during the next few decades. AAs may have been a substantial driver of recent circulation changes in the SH; further research is needed to systematically explore these effects, and the implications for future climate change.

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