

RESEARCH LETTER

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Key Points:

- Warming of the upper and lower stratosphere could be observed this century despite increasing CO₂
- This is due to stratospheric heating from increases in ozone which offsets cooling from CO₂
- Some CMIP5 models have deficiencies in their future ozone which affects stratospheric temperatures

Supporting Information:

- Supporting Information S1

Correspondence to:

A. C. Maycock,
a.c.maycock@leeds.ac.uk

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The contribution of ozone to future stratospheric temperature trends

A. C. Maycock¹¹School of Earth and Environment, University of Leeds, Leeds, UK

Abstract The projected recovery of ozone from the effects of ozone depleting substances this century will modulate the stratospheric cooling due to CO₂, thereby affecting the detection and attribution of stratospheric temperature trends. Here the impact of future ozone changes on stratospheric temperatures is quantified for three representative concentration pathways (RCPs) using simulations from the Fifth Coupled Model Intercomparison Project (CMIP5). For models with interactive chemistry, ozone trends offset ~50% of the global annual mean upper stratospheric cooling due to CO₂ for RCP4.5 and 20% for RCP8.5 between 2006–2015 and 2090–2099. For RCP2.6, ozone trends cause a net *warming* of the upper and lower stratosphere. The misspecification of ozone trends for RCP2.6/RCP4.5 in models that used the International Global Atmospheric Chemistry (IGAC)/Stratosphere-troposphere Processes and their Role in Climate (SPARC) Ozone Database causes anomalous warming (cooling) of the upper (lower) stratosphere compared to chemistry-climate models. The dependence of ozone chemistry on greenhouse gas concentrations should therefore be better represented in CMIP6.

1. Introduction

Stratospheric temperature trends are a fingerprint for anthropogenic climate change. In recent decades, the global decline in ozone and rising CO₂ levels have driven stratospheric cooling, with a magnitude that peaks near the stratopause [Randel *et al.*, 2009]. Changes in ozone and CO₂ are estimated to have caused a global mean cooling at 1 hPa of 0.8 K dec⁻¹ and 1 K dec⁻¹, respectively, between 1980 and 2000 [Shine *et al.*, 2003]. Cooling due to CO₂ dominates in the mid-stratosphere, and cooling due to ozone dominates in the lower stratosphere. While observed trends in stratospheric water vapor are more uncertain [e.g., Rosenlof, 2001], the increases measured at some locations could have contributed a global mean cooling of up to ~0.2 K dec⁻¹ in the lower stratosphere, which is ~40% of the observed trend in this region over the last 30 years [Maycock *et al.*, 2014; Seidel *et al.*, 2011].

Stratospheric temperature trends over the 21st century will be largely determined by future changes in well-mixed greenhouse gases (in particular CO₂), ozone [Eyring *et al.*, 2013], stratospheric water vapor [Gettelman *et al.*, 2010], and changes in stratospheric dynamics [e.g., Manzini *et al.*, 2014]. Ozone abundances across much of the stratosphere are projected to return to pre-1980 levels this century because of the declining levels of ozone depleting substances (ODSs) in the atmosphere [e.g., World Meteorological Organization (WMO), 2014]. Thus, in contrast to the last few decades, where ozone and CO₂ trends have both caused stratospheric cooling, changes in ozone are expected to offset part of the cooling effect due to increasing CO₂ this century. The interplay between the effects of ozone and CO₂ has been proposed as an explanation for the apparent cessation of lower stratospheric cooling since the mid-1990s [Ferraro *et al.*, 2015]. It is therefore important to quantify the impact of these key drivers on future stratospheric temperature trends and to examine their dependence on future greenhouse gas emissions.

Stratospheric ozone abundances not only affect stratospheric temperatures, but are also affected by them due to the temperature dependence of catalytic loss cycles for ozone [e.g., Haigh and Pyle, 1982]. This is particularly important in the upper stratosphere where ozone is under photochemical control. Indeed, stratospheric cooling due to CO₂ is the primary driver of the projected “super recovery” of ozone in the upper stratosphere over the 21st century [e.g., Chipperfield and Feng, 2003; SPARC CCMVal, 2010; Eyring *et al.*, 2013]. This means that the contribution of ozone to future stratospheric temperature trends will depend on the evolution of greenhouse gas abundances. Changes in atmospheric transport may also be important for ozone in regions where the photochemical lifetime of ozone is long (e.g., in the tropical lower stratosphere) [e.g., SPARC CCMVal, 2010].

Banerjee et al. [2016] separated the effects of future climate change (sea surface temperatures, sea ice, and CO₂ abundances) on ozone from those of future ODS changes for the Representative Concentration Pathway 4.5 (RCP4.5) and RCP8.5 scenarios [*Meinshausen et al.*, 2011]. As anticipated, changes in climate over the 21st century resulted in an increase in upper stratospheric ozone due to stratospheric cooling from CO₂. This increase in ozone was larger by around a factor of 2 in RCP8.5 compared to RCP4.5. The difference in upper stratospheric ozone between these two climate scenarios is of a similar magnitude to the effect of future decreases in ODSs [e.g., *WMO*, 2014; *Banerjee et al.*, 2016]. However, according to model simulations [e.g., *Banerjee et al.*, 2016], ODSs have a much larger impact on polar lower stratospheric ozone compared to the effects of climate change. Changes in both ODSs and greenhouse gases will therefore contribute to future stratospheric ozone trends.

In addition to the differences in CO₂, the RCP scenarios differ markedly in their representation of future CH₄ and N₂O abundances. CH₄ is an important chemical driver of stratospheric ozone through its effect on stratospheric water vapor (the main source of HO_x) and the formation of reservoir species that remove reactive radicals that destroy ozone (e.g., HCl) [*Fleming et al.*, 2011]. N₂O is also an important chemical driver of ozone as it is the main source of stratospheric NO_x [*Crutzen*, 1970]. RCP2.6 and RCP4.5 show decreases in CH₄ between 2006–2015 and 2090–2099 of 512 and 173 ppbv, respectively, while RCP8.5 shows a large increase of 1908 ppbv. All three RCP scenarios include increases in N₂O over this period of 21, 47, and 104 ppbv, respectively.

Revell et al. [2012] tested the impact of the differences in CH₄ and N₂O between RCP2.6 and RCP8.5 on stratospheric ozone. They found a complex vertical structure for the impacts of increased CH₄. Ozone abundances increase in the tropical lower stratosphere due to increased vertical advection from the troposphere and increased in situ chemical production [e.g., *Portmann and Solomon*, 2007; *Fleming et al.*, 2011]; this partly offsets the decrease in ozone in this region from a strengthened Brewer-Dobson circulation (BDC) under climate change [e.g., *Butchart*, 2014]. The higher CH₄ levels in RCP8.5 further cause a decrease in ozone between 30 and 50 hPa, an increase between 20 and 3 hPa of <5% in the tropics, and a decrease above 3 hPa where additional HO_x from CH₄ oxidation leads to increased ozone destruction. At polar latitudes, where changes in climate do not appear to have such a large impact on ozone [*Banerjee et al.*, 2016], there is an increase in ozone of 5–10% between 20 and 50 hPa because of the increase in CH₄ favoring the formation of reservoirs for NO_x and ClO_x. *Revell et al.* [2012] also showed that higher N₂O levels in RCP8.5 lead to increases in ozone in the tropical lower stratosphere and decreases elsewhere in the stratosphere through increased NO_x-driven loss. The maximum changes in ozone were ~5% at 5 hPa in the tropics and ~10% over the southern pole [*Revell et al.*, 2012]. In the upper stratosphere and mesosphere, the effects of differences in CH₄ on ozone were found to dominate over those of N₂O. Although their combined effects will not be additive because of coupling via the formation of reservoir species (e.g., ClONO₂), to first order, the higher levels of CH₄ and N₂O in RCP8.5 would both act to increase ozone in the tropical lower stratosphere, but have opposing effects in most of the mid-stratosphere. Differences in future ozone trends between the RCPs will therefore be influenced by differences in climate (i.e., stratospheric temperatures) and in CH₄ and N₂O concentrations.

Global climate models typically include the main factors that determine the forced component of stratospheric temperature changes either as externally prescribed forcings (e.g., CO₂) or in a manner that is consistent with the simulated climate state (e.g., stratospheric water vapor and dynamical processes). Ozone can fall into either of these groups, depending on whether it is externally prescribed or calculated online. Whichever approach is adopted, climate models must account for the factors that determine ozone trends (including ODSs and greenhouse gases) to realistically simulate future stratospheric climate.

The goal of this study is to quantify the contribution of ozone to future stratospheric temperature trends using model simulations from the Fifth Coupled Model Intercomparison Project (CMIP5) and offline radiative transfer calculations. This contribution is compared for CMIP5 models that either externally prescribe or interactively simulate stratospheric ozone. The importance of the results for model studies of future stratospheric climate is emphasized, in particular the relevance for the ozone database being developed for upcoming CMIP6 experiments (M. Hegglin, personal communication, 2016).

2. Data and Methods

Data are used from CMIP5 model simulations with emissions following the RCP2.6, RCP4.5, and RCP8.5 greenhouse gas scenarios [*Meinshausen et al.*, 2011]. Results for the RCP6.0 emissions scenario are not analyzed

because most of the CMIP5 models analyzed here did not perform this simulation. The analysis focuses on changes or trends in ozone and temperature calculated from differences between the periods 2006–2015 and 2090–2099 (divided by a factor of 8.5 for trends). These represent the first and last decades of the standard future CMIP5 experiments. The changes in atmospheric CO₂ concentration between these periods are 42, 159, and 557 ppmv in the RCP2.6, RCP4.5, and RCP8.5 scenarios, respectively.

Data are analyzed from eight CMIP5 models that fully resolve the stratosphere (defined here as having a model lid above 1 hPa): CESM1-WACCM, CMCC-CMS, GISS-E2-H (p2), GFDL-CM3, MIROC-ESM-CHEM, MPI-ESM-MR, MRI-CGCM3, and MRI-ESM1. Where more than one ensemble member is available for each simulation these are averaged to create an ensemble mean (see supporting information Table S1). The MRI-ESM1 model only performed the RCP8.5 experiment. The CMCC-CMS model did not perform the RCP2.6 experiment. Data for GISS-E2-H (p2) were only available for levels up to 10 hPa. Because of this the GISS-E2-H (p2) ozone data are not included in the analysis.

Of these eight models analyzed, five include interactive chemistry (CESM1-WACCM, GISS-E2-H (p2), GFDL-CM3, MIROC-ESM-CHEM, and MRI-ESM1) and the remaining models (CMCC-CMS, MPI-ESM-MR, and MRI-CGCM3) used the International Global Atmospheric Chemistry (IGAC)/Stratosphere-troposphere Processes and their Role in Climate (SPARC) CMIP5 Ozone Database [Cionni *et al.*, 2011; Eyring *et al.*, 2013]. The future part of this database for the stratosphere was generated using chemistry-climate model (CCM) simulations from the Chemistry-Climate Model Validation Exercise (CCMVal-2), which include the effects of projected changes in well-mixed greenhouse gases and ODSs [SPARC CCMVal, 2010]. However, only one future greenhouse gas scenario was used to generate the stratospheric ozone data for all RCP scenarios. This was the SRES (Special Report on Emission Scenarios) A1b scenario, which is most similar to RCP6.0 in the new scenarios adopted in CMIP5. Thus, for some of the RCPs there will be an inconsistency between the future evolution of ozone and greenhouse gases in models that used the IGAC/SPARC Ozone Database.

Additional offline fixed dynamical heating (FDH) calculations are performed using the Edwards and Slingo radiative transfer code [Edwards and Slingo, 1996; Cusack *et al.*, 1999] and the method for adjustment of stratospheric temperatures described in detail by Maycock *et al.* [2011]. The offline FDH calculations use a monthly and zonal mean atmospheric climatology calculated from ERA-Interim reanalysis data for the period 1980–2009 [Dee *et al.*, 2011]. Perturbations in CO₂ and ozone from the three RCP scenarios are added to this climatology, and stratospheric temperatures are adjusted until a new radiative equilibrium state is established under the assumption that the dynamical component of heating remains fixed [Maycock *et al.*, 2011; see also Fels *et al.*, 1980]. The FDH calculations implicitly neglect the contribution to temperature changes from dynamical processes, which may be important on regional scales (e.g., in polar regions), but provide a good estimate of global mean stratospheric temperature changes where radiative equilibrium is a valid assumption.

3. Results

Figure 1 shows latitude-pressure cross sections of zonal and annual mean changes in ozone mixing ratios (ppmv) in the three RCP scenarios. Figures 1a–1c show the multimodel mean of the CMIP5 CCMs analyzed here (excluding GISS-E2-H (p2)—see section 2), and Figures 1d–1f show the IGAC/SPARC CMIP5 Ozone Database. Note that Figures 1d–1f show identical fields because the impact of differences in greenhouse gas concentrations between the RCP scenarios on ozone was not accounted for in the database [Cionni *et al.*, 2011]. Ozone increases across most of the stratosphere in all three RCP scenarios. The increases in upper stratospheric ozone peak at ~3 hPa, which is above the ozone layer maximum (~10 hPa). There is a double peaked structure about the equator, which is related to the effects of declining ODSs because the concentration of ClO_x peaks off the equator [e.g., Banerjee *et al.*, 2016]. Ozone decreases in the tropical lower stratosphere predominantly because of an increase in the strength of the BDC under climate change [e.g., SPARC CCMVal, 2010; Nowack *et al.*, 2015]. This pattern is qualitatively consistent with other recent studies [e.g., Banerjee *et al.*, 2016].

Crucially, the magnitudes of the ozone changes in Figures 1a–1c become larger with increased greenhouse gas forcing. For example, the increase in ozone in the upper stratosphere approximately doubles from ~0.75 ppmv to ~1.5 ppmv between RCP2.6 and RCP8.5. These relative differences do not scale with

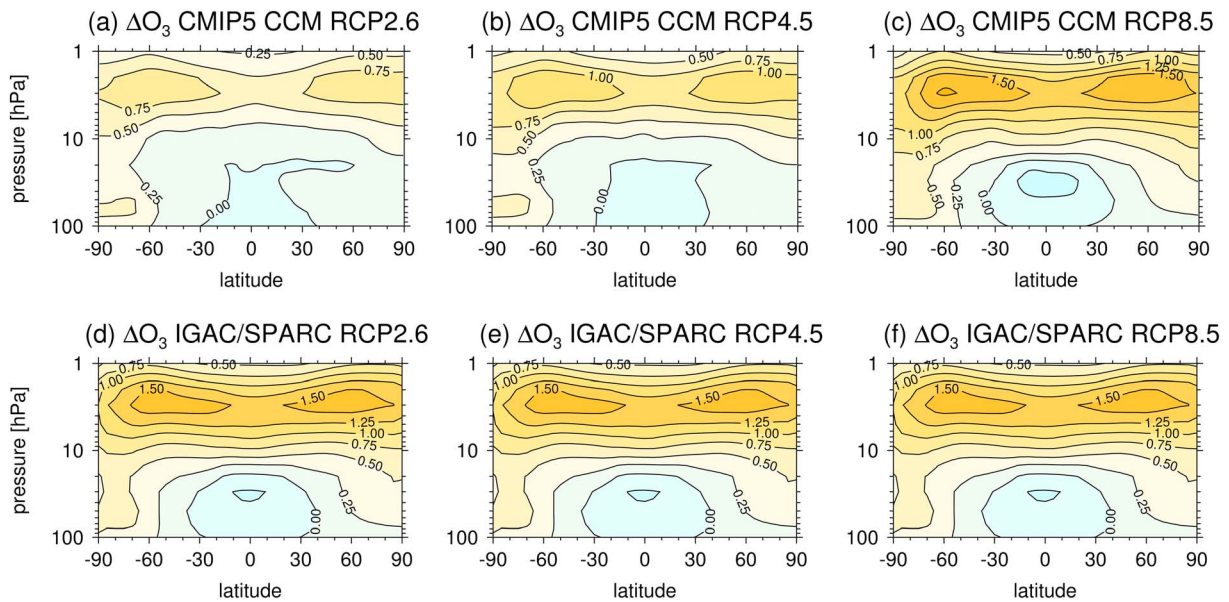


Figure 1. Differences in annual and zonal mean ozone mixing ratios (ppmv) between 2006–2015 and 2090–2099 for the (a–c) multimodel mean of chemistry-climate models from CMIP5 and (d–f) the IGAC/SPARC ozone database [Cionni *et al.*, 2011] recommended for CMIP5 models without chemistry. Data show the RCP2.6 (Figures 1a and 1d), RCP4.5 (Figures 1b and 1e), RCP8.5 (Figures 1c and 1f) experiments. The contour interval is 0.25 ppmv.

CO₂ forcing, which is larger by a factor of 13 in RCP8.5 compared to RCP2.6. This is because ODSs make a substantial contribution to the changes in ozone and there are also differences in CH₄ and N₂O concentrations between the scenarios (see section 1). The decrease in ozone in the tropical lower stratosphere is also enhanced with increased greenhouse gas forcing due to the accompanying increase in the strength of the BDC [e.g., Hardiman *et al.*, 2014]. Note that the results in Figure 1c show slightly larger increases in upper stratospheric ozone over the 21st century if the MRI-ESM1 model, which only performed the RCP8.5 simulation, is excluded from the multimodel mean. However, the effect of including this model is considerably smaller than the differences in the ozone changes between the individual RCPs.

The pattern of ozone changes in the IGAC/SPARC Ozone Database most closely resembles the CMIP5 CCM response for RCP8.5 (Figure 1c) amongst the scenarios considered here, although the total radiative forcing in the SRES A1b scenario used to generate this distribution is closest to RCP6.0 [Meinshausen *et al.*, 2011]. The changes in ozone imposed in the RCP2.6 and RCP4.5 experiments are, therefore, larger in magnitude than those simulated in the CMIP5 CCMs. This difference is likely to affect the simulated differences in stratospheric climate in the models that used the IGAC/SPARC Ozone Database.

Figures 2a–2i show latitude and pressure cross sections of annual and zonal mean stratospheric temperature trends (K dec⁻¹) between 2006–2015 and 2090–2099 for the three RCPs calculated using the offline FDH calculations. For each RCP scenario, temperature trends are calculated for the effect of changes in CO₂ (Figures 2a–2c), and for changes in CO₂ + ozone, where ozone is taken from either the CMIP5 CCM multimodel mean (Figures 2d–2f) or from the IGAC/SPARC Database (Figures 2g–2i). Figures 2j–2l show vertical profiles of the global mean temperature changes for each combination of forcings and for each RCP scenario.

The increases in CO₂ drive stratospheric cooling that increases in magnitude with height [Fels *et al.*, 1980] (Figures 2a–2c). The maximum global mean cooling trend due to CO₂ at 1 hPa is 0.17, 0.65, and 1.65 K dec⁻¹ over the time period considered in the three RCPs (dash-dotted line in Figures 2j–2l). The maximum cooling due to CO₂ in RCP2.6 is therefore around 4 times smaller than in RCP4.5 and around 10 times smaller than in RCP8.5. The corresponding differences in the change in CO₂ concentration between the scenarios are a factor of 4 and a factor of 13. The maximum temperature trend therefore approximately scales with CO₂ between RCP2.6 and RCP4.5, but appears to be sublinear for the larger amplitude RCP8.5 forcing. This can also be seen

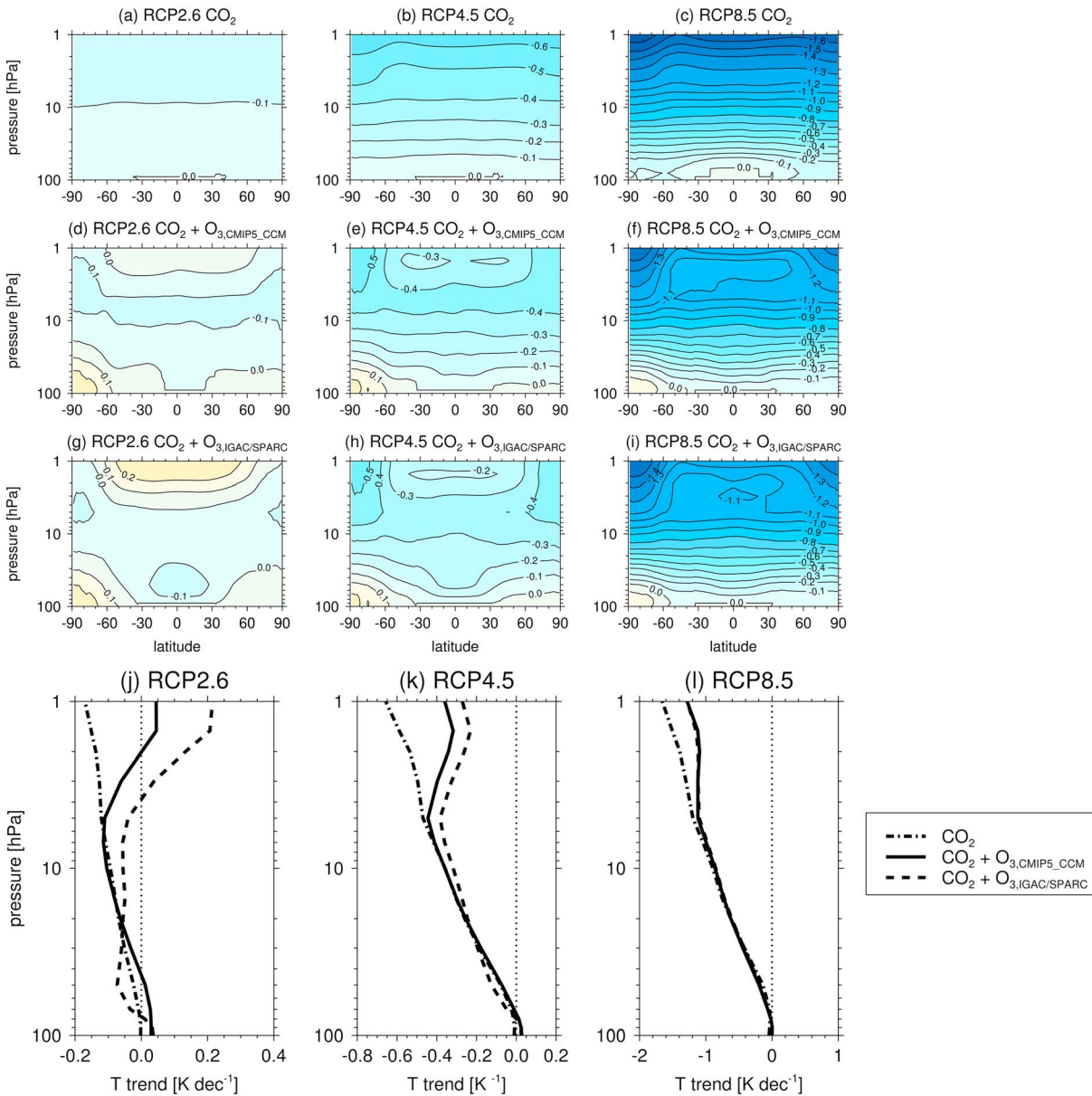


Figure 2. (a–i) Latitude/pressure cross sections of annual and zonal mean stratospheric temperature trends (K dec^{-1}) between 2006–2015 and 2090–2099 from the FDH calculations. Data show (a, d, g, j) RCP2.6, (b, e, h, k) RCP4.5, and (c, f, i, l) RCP8.5 for changes in CO_2 only (Figures 2a–2c), CO_2 plus the CMIP5 CCM multimodel mean ozone changes (Figures 2d–2f) shown in Figures 1a–1c, and CO_2 plus the IGAC/SPARC ozone changes (Figures 2g–2i) shown in Figures 1d–1f. The contour interval is 0.1 K dec^{-1} . (j–l) Vertical profiles of the annual and global mean stratospheric temperature trends due to changes in CO_2 ; the solid lines denote CO_2 plus the CMIP5 CCM ozone changes from Figures 1a–1c; and the dashed lines denote CO_2 plus the IGAC/SPARC ozone changes from Figures 1d–1f.

by comparing the maximum temperature trends due to CO_2 from the RCP4.5 and RCP8.5 scenarios, which show a difference in cooling by a factor of 2.5 but a difference in CO_2 increase by a factor of 3.5. Such nonlinearity in the temperature response could potentially arise from saturation effects in the center of the strong $15 \mu\text{m}$ absorption band for CO_2 [Zhong and Haigh, 2013].

We now consider the impact of changes in ozone on future stratospheric temperature trends. For the RCP2.6 scenario, the ozone changes simulated by the CMIP5 CCMs (Figures 1a–1c) contribute a relative warming trend in the upper stratosphere by up to 0.2 K dec^{-1} in the global and annual mean relative to the CO_2 -induced changes only (compare dash-dotted and solid line in Figure 2j). The largest impact of the ozone changes on temperature is in the tropical upper stratosphere (Figure 2d). For the IGAC/SPARC ozone database,

in which the increase in upper stratospheric ozone is considerably larger for the RCP2.6 scenario, the magnitude of the warming effect is up to 0.35 K dec^{-1} (compare dash-dotted and dashed lines in Figure 2j). In both cases, the warming effect due to ozone outweighs the cooling from the relatively small increase in CO_2 in RCP2.6, resulting in a net *warming* in the uppermost stratosphere. This net warming trend extends down to lower altitudes in the case of ozone changes from the IGAC/SPARC database (4 hPa) than for the CMIP5 CCMs (2 hPa). The CMIP5 CCM ozone changes are small between 30 and 5 hPa and thus have little effect on temperature trends at these levels, whereas the IGAC/SPARC ozone changes contribute a relative warming compared to the effects of CO_2 down to 20 hPa. At pressures greater than 30 hPa, where CO_2 has a smaller effect on temperature, the CCM ozone changes also lead to a net *warming* of the lower stratosphere. This means that, depending on the future greenhouse gas emissions trajectory, warming of the upper and lower stratosphere could be observed this century, despite continued increases in CO_2 ; this would be in contrast to observed stratospheric temperature trends over the recent past [Randel *et al.*, 2009]. In contrast, the IGAC/SPARC ozone changes have a *cooling* effect below 30 hPa because of the anomalously large decreases in ozone imposed in the tropical lower stratosphere (compare Figures 1a and 1d), which are primarily related to increases in the strength of the BDC. Thus, the approach for representing ozone adopted by a model can substantially alter the magnitude of simulated global temperature trends in the upper stratosphere and alter the sign of temperature trends in the lower stratosphere.

For the RCP4.5 scenario, the changes in ozone simulated by the CCMs contribute a relative warming of the upper stratosphere by up to 0.3 K dec^{-1} , which offsets around 50% of the cooling due to CO_2 in this scenario. In contrast, ozone has no detectable impact on global temperatures below 5 hPa (Figure 2k), where the ozone changes are smaller and there is a compensation between warming and cooling effects at high and low latitudes (Figure 2e). As a result of the larger increases in ozone in the upper stratosphere compared to the CCMs, the IGAC/SPARC ozone changes in RCP4.5 lead to a slightly larger relative warming in the upper stratosphere of up to 0.4 K dec^{-1} , which offsets around 60% of the cooling due to CO_2 . In this case ozone has a relative warming effect down to 20 hPa. As was found in the RCP2.6 experiment, the IGAC/SPARC ozone causes a cooling at pressures greater than 30 hPa because of the larger ozone decreases in the tropics (Figure 2h). For RCP8.5, the ozone changes simulated by the CCMs and the IGAC/SPARC Database are very similar (compare Figures 1c and 1f) and therefore their effects on stratospheric temperature trends are virtually indistinguishable. In both cases, there is a peak warming effect due to ozone of $\sim 0.4 \text{ K dec}^{-1}$, which offsets the much larger upper stratospheric cooling due to increases in CO_2 by around 20%.

In all three RCPs, the changes in ozone substantially modify the stratospheric temperature trends due to CO_2 alone. Since the impact of ozone on temperatures is more spatially inhomogeneous than for CO_2 , this has the potential to affect projected trends in stratospheric winds [e.g., Manzini *et al.*, 2014]. The projected changes in stratospheric ozone also have a modest impact on estimates of future radiative forcing trends, with ozone acting to slightly reduce the positive forcing from CO_2 (see supporting information Table S2). It is therefore important to represent ozone in climate models in a manner that is consistent with the adopted scenario for future greenhouse gas emissions.

The comparison between the impacts of the CMIP5 CCM simulated ozone and the IGAC/SPARC Ozone Database on stratospheric temperature trends reveals a source of error in CMIP5 models. In particular, models that employed the IGAC/SPARC Database would be expected to show stronger warming of the uppermost stratosphere and cooling in the lower stratosphere for RCP2.6, and show weaker upper stratospheric cooling in RCP4.5 than the mean behavior of models containing interactive chemistry schemes. To test this, Figure 3 shows vertical profiles of global and annual mean stratospheric temperature trends over 2006–2015 to 2090–2099 in the eight CMIP5 models described in section 2 (see also supporting information Table S1). Each line denotes one model, with solid lines showing models that calculate ozone online and dashed lines showing models that used the IGAC/SPARC Ozone Database. The equivalent thick black lines show multimodel averages for the respective subsets of models. For RCP2.6 and RCP4.5, the differences between the two subsets of models strongly resemble the differences between the solid and dashed lines in Figures 2j–2l. It therefore appears extremely likely that the prescription of ozone is responsible for the major differences in stratospheric temperature trends between these models in the RCP2.6 and RCP4.5 scenarios. Differences in stratospheric water vapor trends could also play a role [e.g., Gettelman *et al.*, 2010], but this is likely to be most important in the lowermost stratosphere [e.g., Maycock *et al.*, 2011, 2014].

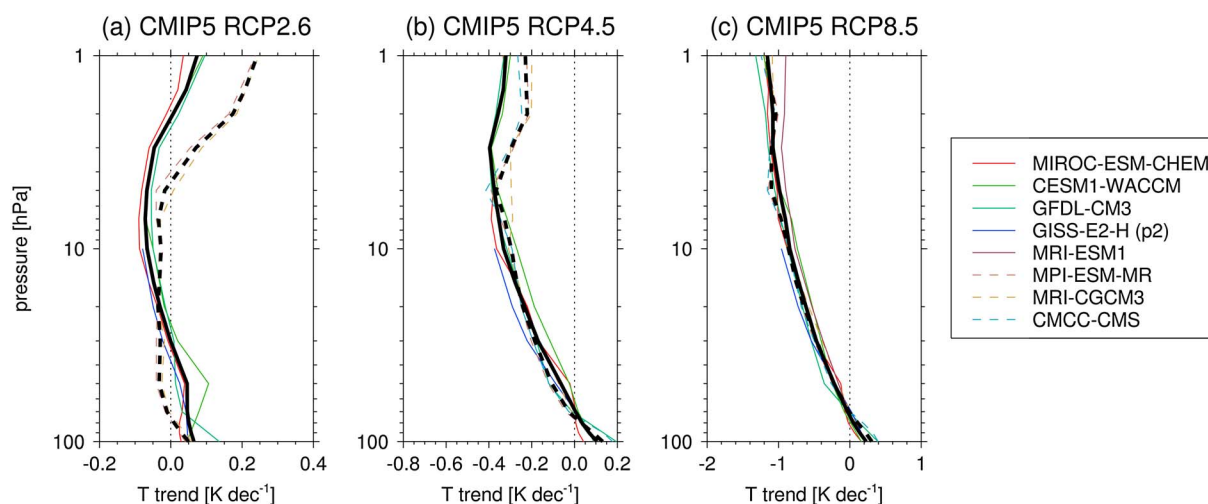


Figure 3. Vertical profiles of annual, global mean stratospheric temperature trends (K dec^{-1}) between 2006–2015 and 2090–2099 for the (a) RCP2.6, (b) RCP4.5, and (c) RCP8.5 experiments. The colored lines show individual models; solid lines denote models that calculate ozone online (MIROC-ESM-CHEM, CESM1-WACCM, GFDL-CM3, MRI-ESM1 (RCP8.5 only), and GISS-E2-H (p2) (only up to 10 hPa)); and dashed lines denote models that prescribe the IGAC/SPARC Ozone Database or an extended version of this database (MPI-ESM-MR, MRI-CGCM3, and CMCC-CMS (not RCP2.6)). The equivalent solid and dashed black lines show the multimodel averages of the model subsets.

4. Conclusions

Stratospheric temperature trends are a key fingerprint for anthropogenic climate change. Over the past few decades, observed stratospheric cooling has been mostly driven by the global decline in stratospheric ozone and increasing CO_2 abundances [Shine *et al.*, 2003], with a more uncertain, but potentially significant, role for stratospheric water vapor changes in the lower stratosphere [Maycock *et al.*, 2014]. In the future, the projected recovery of stratospheric ozone is expected to modulate the cooling due to continued increases in CO_2 . It is therefore important to quantify the interplay between these drivers, and their dependence on future greenhouse gas emissions, to understand potential scenarios for future stratospheric temperature trends.

This study has quantified the contributions of CO_2 and ozone to future stratospheric temperature trends for three representative concentration pathway (RCP) scenarios that were used within CMIP5: RCP2.6, RCP4.5, and RCP8.5. In all scenarios, the projected future increases in upper stratospheric ozone offset a substantial part of the cooling due to CO_2 . The fractional offset is largest for the most conservative greenhouse gas emissions scenario (RCP2.6). In this case, the effect of ozone on globally averaged temperatures overwhelms the cooling due to CO_2 and results in a net *warming* in both the upper and lower stratosphere. The interplay between the effects of ozone and CO_2 has been pointed to as an explanation for the recent slowdown in the rate of lower stratospheric cooling [Ferraro *et al.*, 2015]; the results presented here show that small, or even positive, stratospheric temperature trends could be observed in the future depending on which greenhouse gas emissions pathway is followed. For the RCP4.5 and RCP8.5 scenarios, ozone offsets around 50 and 20% of the cooling in the uppermost stratosphere due to CO_2 , respectively.

The dependence of future ozone trends on greenhouse gas abundances was not properly accounted for in the IGAC/SPARC Ozone Database that was recommended for CMIP5 models that do not calculate ozone online. In particular, ozone trends that most closely resemble the response to the RCP8.5 scenario were imposed for all RCPs. This misrepresentation of ozone causes an anomalous warming trend in the upper stratosphere and cooling in the lower stratosphere in the RCP2.6 and RCP4.5 experiments, which can be identified in a subset of CMIP5 models that resolve the stratosphere. The impact of this misrepresentation of ozone on other aspects of future stratospheric climate change [e.g., Manzini *et al.*, 2014] remains to be tested. Nevertheless, the impacts highlighted here motivate that the ozone database being created for CMIP6 should more realistically account for the greenhouse gas dependence of future ozone trends.

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