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## Comment on “Tropospheric temperature response to stratospheric ozone recovery in the 21st century” by Hu et al. (2011)

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**Abstract.** In a recent paper Hu et al. (2011) suggest that the recovery of stratospheric ozone during the first half of this century will significantly enhance free tropospheric and surface warming caused by the anthropogenic increase of greenhouse gases, with the effects being most pronounced in Northern Hemisphere middle and high latitudes. These surprising results are based on a multi-model analysis of CMIP3 model simulations with and without prescribed stratospheric ozone recovery. Hu et al. suggest that in order to properly quantify the tropospheric and surface temperature response to stratospheric ozone recovery, it is necessary to run coupled atmosphere-ocean climate models with stratospheric ozone chemistry. The results of such an experiment are presented here, using a state-of-the-art chemistry-climate model coupled to a three-dimensional ocean model. In contrast to Hu et al., we find a much smaller Northern Hemisphere tropospheric temperature response to ozone recovery, which is of opposite sign. We suggest that their result is an artifact of the incomplete removal of the large effect of greenhouse gas warming between the two different sets of models.

### 1 Introduction

Stratospheric ozone depletion has had a radiative effect on global mean surface climate, although the sign of the effect is uncertain due to the large compensation between the short-wave warming due to increased penetration of solar radiation and the long-wave cooling due to reduced downwelling infrared radiation from the colder stratosphere (Intergovernmental Panel on Climate Change (IPCC), 2007; Chapter 10

of SPARC CCMVal, 2010). But all recent estimates (IPCC, 2007; SPARC CCMVal, 2010) are considerably smaller in magnitude than  $0.1 \text{ W m}^{-2}$ , and thus represent a small number compared to the total radiative forcing. On the other hand, the Antarctic ozone hole, which is a huge perturbation to the Southern Hemisphere (SH) stratosphere, has been the dominant driver of past changes in high-latitude SH tropospheric climate in summer (e.g. Arblaster and Meehl, 2006; Fogt et al., 2009), with ozone recovery expected to offset the effects of climate change over the next half-century (e.g. Son et al., 2010). While similar physics might be expected to be at work at high latitudes in the Northern Hemisphere (NH), no such effect has so far been detected there, partly because of the smaller magnitude of ozone depletion in the Arctic, and partly because of the larger impact of greenhouse warming due to melting sea ice (see discussion in Chapter 4 of WMO, 2011).

In a recent study, Hu et al. (2011, henceforth H11) investigated the possible impact of stratospheric ozone recovery on tropospheric temperatures using Coupled Model Intercomparison Project (CMIP3) general circulation model (GCM) simulations of the 21st century. They did this by comparing one set of model projections in which ozone recovery is prescribed with another set of projections (employing different models) in which ozone concentrations are held fixed, with both sets of projections having identical increases in well-mixed greenhouse gas (GHG) concentrations. Focusing on the period from 2001 to 2050, H11 find a significant enhancement of tropospheric warming in the GCM ensemble with prescribed ozone recovery. This enhanced warming is largest in the upper troposphere, with a global and annual

mean change of  $\sim 0.41$  K over 50 yr ( $\sim 0.08$  K decade $^{-1}$ ). They also find relatively large enhanced warming in the extratropical and polar regions in summer and autumn in both hemispheres, as well as a significant warming at the surface with a global and annual mean change of  $\sim 0.16$  K over 50 yr ( $\sim 0.03$  K decade $^{-1}$ ). In fact, the largest warming is found in the NH, which is very surprising given that the changes in stratospheric ozone are much larger in the SH. Furthermore the NH high-latitude surface warming maximizes in late fall/early winter, which is also very surprising since the ozone increase maximizes in spring. H11 compare their GCM results to results from a radiative-convective model, and find that although the latter predicts increased warming as ozone levels recover, the tropospheric warming is weaker by a factor of four than that determined from the ensembles of GCMs. They attribute this warming difference to the simplicity of their radiative-convective model.

Another possible explanation for the apparently large impact of stratospheric ozone recovery on NH temperatures is that their multi-model approach is flawed. Attributing differences between the two sets of simulations to the effects of ozone recovery is questionable if the signal one is looking for is small. Since greenhouse warming is expected to dominate the effects of ozone recovery in the NH, small differences in the tropospheric temperature trends between the two sets of models may simply be a reflection of differences in the GHG-induced warming, and have nothing to do with ozone recovery. Although H11 claim that the mean transient climate response (TCR) of the two sets of models is the same (1.7 K), we compute a difference of 0.22 K for the models used for the future changes, based on the incomplete information provided in Table 8.2 of IPCC (2007), with the models with ozone recovery having the larger mean TCR. It is therefore plausible that a relatively small difference in the mean TCRs could account for the different rates of tropospheric warming in their two sets of model simulations. Furthermore, there were differences in the radiative forcings applied to the two sets of models, notably with respect to tropospheric ozone where there is a close correspondence between whether or not models included stratospheric ozone recovery and tropospheric ozone changes. Since the projected increase of tropospheric ozone provides a significant GHG warming, especially in the NH, this could also contribute to the NH warming found by H11. Finally, the rate of Arctic warming, which is not encapsulated in a global metric like the TCR, differs from model to model because of different rates of Arctic sea ice loss. In fact, Crook and Forster (2011) show that GCMs with large Arctic amplification factors do not necessarily have large TCRs. Thus, even if the mean TCRs of the two sets of models were identical, the mean Arctic amplification factors will almost certainly differ. The enhanced surface warming in Arctic winter found by H11 for the models with imposed ozone recovery may therefore be a reflection of that.

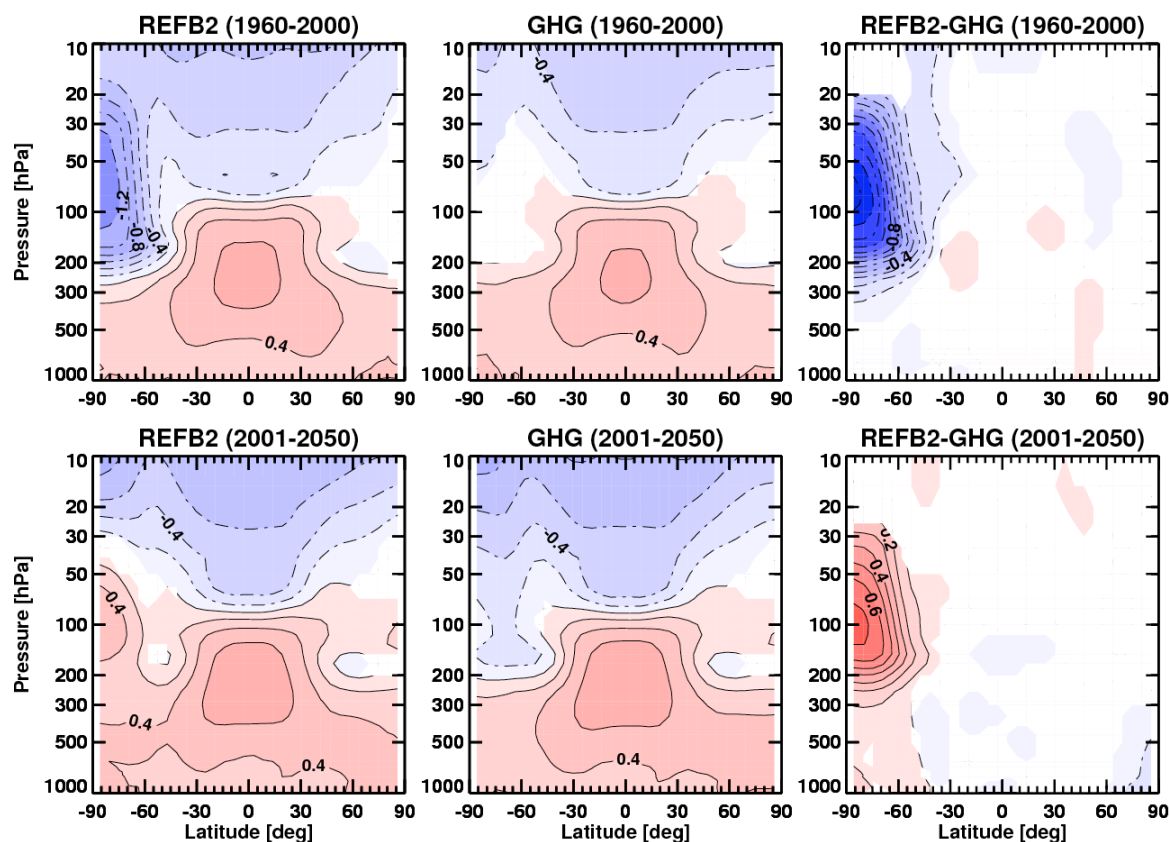
In their Conclusion, H11 acknowledge the limitations in their approach and suggest that coupled atmosphere-ocean models including stratospheric ozone chemistry are needed to properly investigate the tropospheric and surface temperature responses to stratospheric ozone recovery, in order to avoid this “small difference of large terms” problem. Here, we describe results from such an exercise, using simulations from the Canadian Middle Atmosphere Model (CMAM). By comparing an ensemble of simulations with increasing GHG concentrations and time-varying ozone-depleting substances (ODSs) to an ensemble of simulations with only increasing GHG concentrations (i.e., ODS concentrations held fixed), using the same coupled model, we are able to assess the impact of ozone recovery on tropospheric temperatures in a self-consistent manner. Contrary to the results of H11, we find only a small NH tropospheric temperature response to ozone recovery, which is in fact opposite in sign to theirs.

The outline of our paper is as follows. In Sect. 2 we describe CMAM and the simulations we use. In Sect. 3 we discuss our results. For easy comparison we present many of our results in a similar format to that used by H11. In Sect. 4 we discuss in greater depth the potential causes for the disagreement between our results and those of H11.

## 2 Description of model and simulations

CMAM is the upward extension of the Canadian Centre for Climate Modelling and Analysis (CCCma) third generation coupled GCM (CGCM3). The ocean component of CMAM is described in McLandress et al. (2010). The atmospheric component has 71 vertical levels, with a resolution that varies from several tens of meters in the lower troposphere to  $\sim 2.5$  km in the mesosphere. A T31 spectral resolution is used in the horizontal, which corresponds to a grid spacing of  $\sim 6^\circ$ . Detailed descriptions of the stratospheric chemistry scheme and the atmospheric component of CMAM are provided in de Grandpré et al. (2000) and Scinocca et al. (2008), respectively.

The two sets of simulations we use are described in detail in McLandress et al. (2010), and the evolution of ozone in the simulations is described in Plummer et al. (2010). The first set is the “REF-B2” simulation, which employs time varying concentrations of GHGs and ODSs, with the GHGs prescribed according to the moderate SRES A1B scenario (IPCC, 2001) and the ODSs according to the A1 scenario (WMO, 2007). The second set is the “GHG” simulation, which uses identical forcings as REF-B2 (i.e., GHGs, aerosols, etc.) with the exception of the ODSs, whose concentrations are held fixed at their 1960 values in the chemistry scheme. Note that in the GHG simulation time-varying concentrations of CFC-11 and CFC-12 are used in the radiation scheme, as in the REF-B2 simulation. The impact of the ODS changes (and therefore the impact of the stratospheric ozone changes) is inferred by differencing the REF-B2 and



**Fig. 1.** Annual and zonal mean temperature trends for 1960–2000 (top) and 2001–2050 (bottom): REF-B2 (left), GHG (middle), and REF-B2 minus GHG (right). Contour intervals are 0.2 and 0.1 K decade<sup>-1</sup> in the two left columns and right columns, respectively, with red denoting positive values and blue negative. Only trends that are significant at the 95 % level are shown.

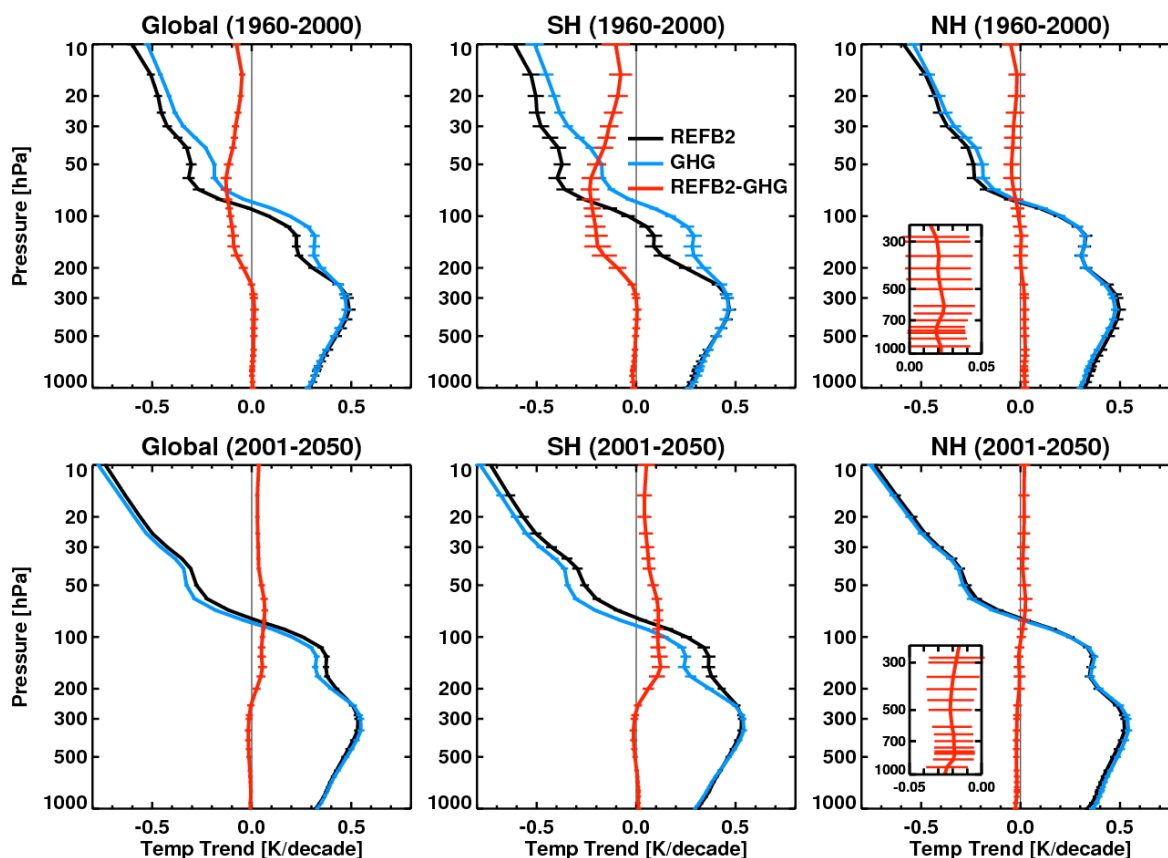
GHG simulations, as in Plummer et al. (2010). The simulations extend from 1960 to 2099, with each set of simulations comprising an ensemble of three. Details of the spin-up procedure are given in McLandress et al. (2010).

We present results both for the 1960–2000 (“ozone depletion” or “past”) period and the 2001–2050 (“ozone recovery” or “future”) period. Since the sign of the trends driven by changes in stratospheric ozone is expected to change from past to present (e.g., McLandress et al., 2010, 2011), comparing these two periods helps in assessing the robustness of the results. Linear trends are computed from ensemble mean time series, and their statistical significance is computed using the standard t-test (i.e., assuming independent and Gaussian-distributed residuals). All figures show ensemble averages.

### 3 Results

#### 3.1 Annual mean

Figure 1 shows latitude-height sections of annual and zonal mean temperature trends for the REF-B2 (left) and GHG (middle) simulations and their difference (right) for the past (top) and future (bottom). REF-B2 and GHG both show tropospheric warming and stratospheric cooling over both periods as a result of increasing GHG concentrations in those two simulations. Only trends that are statistically significant at the 95 % level are plotted. The difference between the two shows large trends in the SH polar stratosphere, which change sign from past to future as the Antarctic ozone hole recovers. In the troposphere there are several regions of weak trends in the future, with warming at high southern latitudes and cooling in the Arctic, and a tendency for oppositely signed trends in the past. The lower right panel in Fig. 1 is directly comparable to Fig. 6a of H11. In contrast to their results, we see no evidence of enhanced tropospheric warming during the ozone recovery period, and, as stated above, we in fact find weak cooling in the NH.



**Fig. 2.** Annual mean temperature trends for 1960–2000 (top) and 2001–2050 (bottom): global average (left), Southern Hemisphere (middle) and Northern Hemisphere (right) for REF-B2 (black), GHG (blue) and REF-B2 minus GHG (red). Error bars denote the 95 % confidence levels of the trends. The insets in the two right panels show blow-ups of the REF-B2-minus-GHG trends in the troposphere.

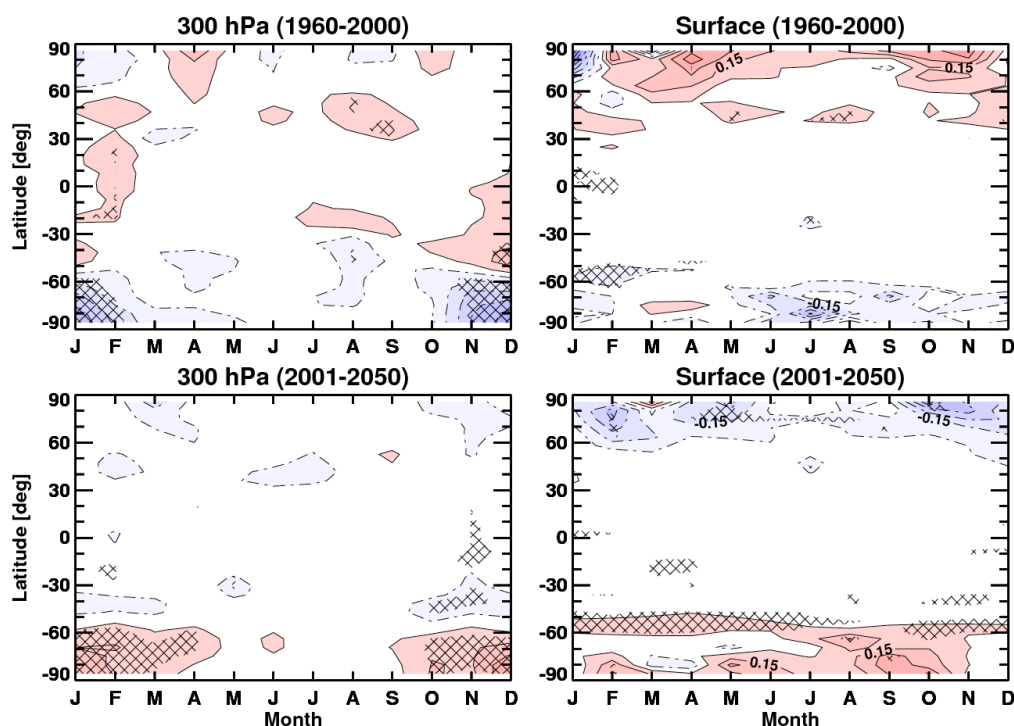
A more compact way of presenting the annual mean temperature trends is by plotting latitudinal averages, as is done in Fig. 2. Shown here are global averages (left), SH average (middle) and NH average (right) for REF-B2 (black), GHG (blue) and REF-B2 minus GHG (red) for the past and future. The two left and bottom right panels are directly comparable to Figs. 2 and 4 of H11. The maximum impact of the ozone changes occurs at  $\sim 70$  hPa, with the effect being much larger in the SH than in the NH, as expected. We also note that the magnitude of the trends in REF-B2 minus GHG is larger for the past than for the future because the ozone recovery process is not completed by 2050 (Plummer et al., 2010).

Closer inspection of the right panels of Fig. 2 reveals that below about 300 hPa the 95 % confidence error bars on the red curve do not cross the zero line (see insets), indicating that there is a statistically significant impact of both ozone depletion and ozone recovery on NH average tropospheric temperature. Interestingly, our model results suggest that NH ozone depletion has led to a small tropospheric warming, which would be consistent with ozone depletion exerting a net positive radiative forcing (Chapter 10 of SPARC CCMVal, 2010). Our simulations also suggest that ozone re-

covery will lead to a small tropospheric cooling. However, the magnitude of both the past and future NH tropospheric temperature trends are small ( $\sim 0.02$  K decade $^{-1}$  in the upper troposphere, i.e., about a factor of four smaller than the future warming found by H11).

### 3.2 Seasonal variation

Turning now to the seasonal variation of the ODS-induced temperature changes in the troposphere, the left panels in Fig. 3 show latitude-month cross sections of the REF-B2 minus GHG temperature trends at 300 hPa. Note that unlike Fig. 1 statistical significance is denoted here by cross hatching. Opposite-signed trends between past and future are seen at high southern latitudes in late spring and early summer. These are due to the delayed breakdown of the SH vortex during the ozone depletion period and the return to earlier breakdown dates during the ozone recovery period (e.g. McLandress et al., 2010). Comparing the bottom left panel to Fig. 8a in H11, one can clearly see the above-mentioned SH features in H11's CMIP3 model results. However, the warming at low and middle latitudes in the NH in summer seen in



**Fig. 3.** Zonal mean temperature trend versus month and latitude for REF-B2 minus GHG for 1960–2000 (top) and 2001–2050 (bottom) at 300 hPa (left) and at the surface (right). Contour interval is  $0.1 \text{ K decade}^{-1}$ , with red denoting positive values and blue negative. Trend magnitudes less than  $0.05 \text{ K decade}^{-1}$  are not plotted. Cross hatching denotes regions where the 95 % significance level is exceeded.

H11 is absent in our results. Although there are patches of past warming and future cooling in the NH, which are consistent with the NH average results shown in Fig. 2, they are not statistically significant when considered regionally and seasonally.

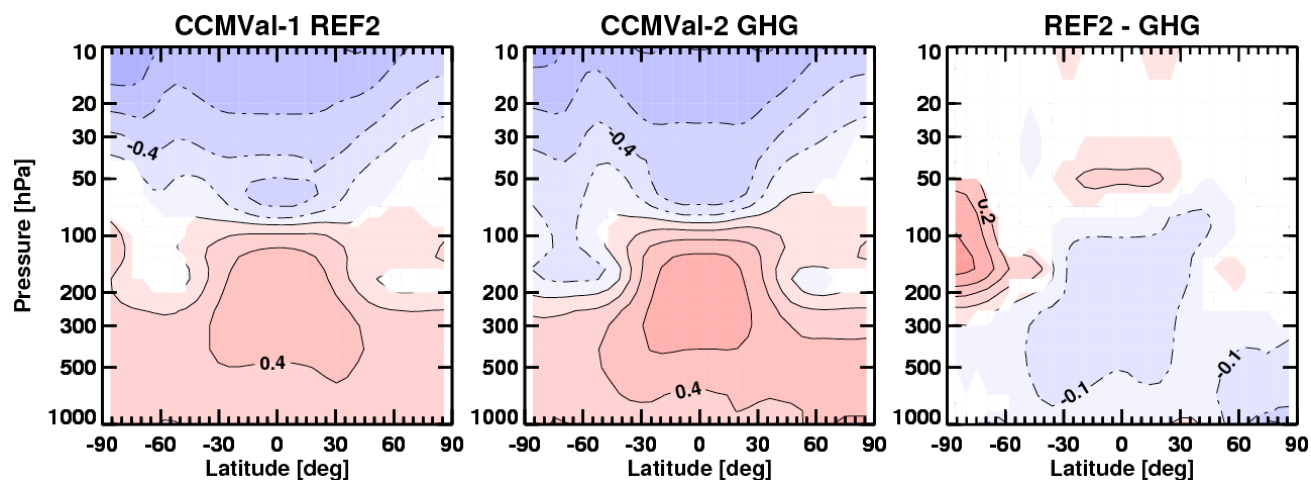
H11 also found large enhanced surface warming in the Arctic during the period of ozone recovery, and suggested that the increasing ozone concentrations are somehow amplifying the high-latitude response to global warming. The right panels of Fig. 3 show the zonal mean temperature trends at the surface. A comparison of the bottom right panel to Fig. 11 of H11 reveals major differences. H11 reported strong warming in the Arctic, especially in fall and winter, while CMAM shows cooling at these latitudes. The Arctic (average over  $60\text{--}90^\circ \text{N}$ ) surface temperature difference trends averaged from September to January – the time period H11 found to exhibit the maximum warming – exhibit a weak but statistically significant cooling in the future ( $-0.136 \pm 0.130 \text{ K decade}^{-1}$ ).

#### 4 Conclusions

A self-consistent analysis of the possible impact of stratospheric ozone recovery on tropospheric temperatures has been undertaken using a version of the Canadian Middle At-

mosphere Model (CMAM) that is coupled to an ocean model. Two sets of simulations are performed: one with time-varying concentrations of GHGs and ODSs, the other with time-varying GHGs and constant ODSs. Although our simulations show the expected large differences in stratospheric temperature changes, we find only a small impact on tropospheric temperatures, consistent with the small estimated radiative forcing of stratospheric ozone changes (IPCC, 2007; SPARC CCMVal, 2010). Interestingly, the effect in the NH is such that ozone depletion leads to a tropospheric warming, and ozone recovery to a tropospheric cooling, which is consistent with ozone depletion representing a positive radiative forcing as has been suggested in recent studies (SPARC CCMVal, 2010). It could also represent a climate feedback.

Our results are in striking contrast to those of Hu et al. (2011), who suggest that ozone recovery will have a substantial warming effect in the troposphere (a global and annual mean change of  $\sim 0.41 \text{ K}$  over 50 yr ( $\sim 0.08 \text{ K decade}^{-1}$ ) in the upper troposphere, compared with the cooling of  $\sim 0.02 \text{ K decade}^{-1}$  found here), which is largest in the NH. H11 base their findings on an analysis of CMIP3 models with and without ozone recovery. This approach has been used successfully to determine the impact of stratospheric ozone changes on SH summertime circulation changes (e.g., Perlwitz et al., 2008; Fogt et al., 2009; Son et al., 2009), as confirmed by a multi-model comparison of CCMVal models



**Fig. 4.** Annual and zonal mean temperature trends for 2001–2050: CCMVal-1 REF2 (left), CCMVal-2 GHG (middle), and CCMVal-1 REF2 minus CCMVal-2 GHG (right). Contour intervals are 0.2 and 0.1 K decade<sup>-1</sup> in the two left panels and the right panel, respectively, with red denoting positive values and blue negative. Only trends that are significant at the 95 % level are shown.

(e.g., Son et al., 2010) and sensitivity studies using single models (McLandress et al., 2011; Polvani et al., 2011). The reason why this approach works in the summertime SH is because the Antarctic ozone hole is such a large perturbation to the SH circulation. However, using such an approach in the NH, in particular the Arctic, as H11 do, is problematic since the stratospheric ozone changes in northern high latitudes are considerably weaker, and the GHG-induced warming (which needs to be removed in order to isolate the effects of ozone recovery) is larger.

We suggest that the enhanced tropospheric warming found by H11 results from the comparison of groups of models having different rates of GHG-induced warming; specifically, that differencing the two groups of models does not remove the effect of GHG-induced warming as is needed in order to isolate the effects of ozone recovery. Important regions where such sensitivity to GHG changes becomes obvious are the upper tropical troposphere and the Arctic surface. The rate of upper tropical tropospheric warming is closely related to the rate of surface warming (Arblaster et al., 2011), which is closely linked to the climate sensitivity of the model. For the Arctic, surface warming is strongly determined by the rate of Arctic sea ice loss. Stroeve et al. (2007) showed that CMIP3 models exhibit a large range of declining sea ice extent trends for the period 1953–2006. Thus, compositing two model sets with different sea ice loss rates will result in large apparent effects in Arctic surface temperatures. The seasonality of the Arctic warming determined by H11, with maximum surface warming during late fall/early winter, is consistent with the seasonality expected from the impact of Arctic sea ice loss (Deser et al., 2010). This seasonality is not consistent with the effect of stratospheric ozone changes, which maximize in spring.

We provide here a simple yet illustrative example of why the method of H11 is inappropriate in the tropical and NH troposphere where the impact of ozone forcing is expected to be small relative to that of other processes. We do this by computing differences in two ensembles of simulations produced using two different versions of CMAM. The first is the “REF2” simulation generated using the CCMVal-1 version of CMAM (Eyring et al., 2007). Like REF-B2, the REF2 ensemble of three simulations uses time-varying concentrations of GHGs and ODSs, but unlike REF-B2 it employs prescribed sea-surface temperatures and sea-ice distributions generated using an earlier version of the CCCma coupled atmosphere-ocean model on which that version of CMAM was based. The second set is the GHG simulation using the CCMVal-2 version of CMAM, which has been discussed above. Differencing the two ensemble means is thus analogous to H11 differencing the means of the two different sets of AR4 models with and without ozone recovery.

The results of this exercise are given in Fig. 4, which shows annual and zonal mean temperature trends for the 2001–2050 period for the two sets of simulations and for the corresponding difference. As with the differences between REF-B2 and GHG shown previously (bottom right panel in Fig. 1), the impact of ozone recovery is clearly seen in the Antarctic lower stratosphere. However, large statistically significant trends (cooling) are also found in the troposphere, with larger values in the NH than in the SH and with a strong surface signal in the Arctic, much as in H11 but of opposite sign. The reason why there are such large differences in the troposphere is because the GHG-induced warming is stronger in the CCMVal-2 version of CMAM than in the CCMVal-1 version, with tropical (20° S to 20° N) sea-surface temperature trends from 2001–2050 of  $\sim 0.27$  K decade<sup>-1</sup>



and 0.20 K decade<sup>-1</sup>, respectively. Thus, differencing the two sets of simulations yields the cooling trends seen in the right panel of Fig. 4. The fact that H11 find enhanced warming, while Fig. 4 shows cooling, is immaterial since the mean rate of GHG-induced global warming in the CMIP3 models with ozone recovery may simply be larger than in those without.

A recent study by Prevedi and Polvani (2012) confirms our hypothesis. They examined the near surface temperature trends in CMIP3 model experiments in which CO<sub>2</sub> increased by 1 % per year until doubling. When differencing the same two groups of models as H11, they found remarkably similar results to H11. Since the stratospheric ozone forcing in their two sets of models was identical, the temperature differences could only have arisen from the different responses to the GHG forcing.

In addition, Table 10.1 in IPCC (2007) shows that in most cases, the CMIP3 simulations that include stratospheric ozone recovery also include tropospheric ozone changes, and vice versa. Since the projected increase of tropospheric ozone provides a significant GHG warming, especially in the NH, this could also contribute to the differences found by H11. There were also differences in the groups of models considered by H11 in their treatment of other radiative forcing agents such as black carbon, indirect aerosol effect, etc., which further complicates the attribution of enhanced tropospheric warming in a specific CMIP3 model group to a single forcing factor like stratospheric ozone recovery.

Although our results are for only a single model (and so are subject to the potential weaknesses of that model), they clearly illustrate the pitfalls in analysing CMIP3 models with and without ozone recovery when trying to quantify the impacts of ozone recovery on tropospheric temperatures in the NH. A more definitive analysis would require a multi-model approach using coupled chemistry-climate models or IPCC-like models in which each model performs simulations with and without ozone recovery, and where the ocean and sea ice models coupled to the atmospheric model can respond.

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