



Impacts of climate change on stratospheric ozone recovery

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[1] The impact of increasing greenhouse gases (GHGs) on the “recovery” of stratospheric ozone is examined using simulations of the Goddard Earth Observing System Chemistry-Climate Model. In this model, GHG-induced climate change has a large impact on the ozone evolution and when O₃ recovery milestones are reached. The two distinct milestones of “O₃ returning to historical values” and “O₃ being no longer significantly influenced by ozone depleting substances (ODSs)” can be reached at very different dates, and which occurs first varies between regions. GHG-induced cooling in the upper stratosphere causes O₃ to increase, and O₃ returns to 1980 or 1960 values several decades before O₃ is no longer significantly influenced by ODSs. In contrast, transport changes in the tropical and southern mid-latitude lower stratosphere cause O₃ to decrease. Here O₃ never returns to 1980 values, even when anthropogenic ODSs have been removed from the atmosphere. O₃ returning to 1960 (or 1980) values should not necessarily be interpreted as O₃ recovery from the effects of ODSs. **Citation:** Waugh, D. W., L. Oman, S. R. Kawa, R. S. Stolarski, S. Pawson, A. R. Douglass, P. A. Newman, and J. E. Nielsen (2009), Impacts of climate change on stratospheric ozone recovery, *Geophys. Res. Lett.*, *36*, L03805, doi:10.1029/2008GL036223.

1. Introduction

[2] There is considerable interest in how stratospheric ozone will evolve through the 21st century. Due to regulations imposed by the Montreal Protocol and its amendments, the concentration of the sum of all ozone-depleting substances (ODSs) peaked in the 1990s and is expected to decrease back to 1960 levels around the end of this century [*World Meteorological Organization (WMO)/United Nations Environment Programme (UNEP)*, 2007], with a corresponding “recovery” of stratospheric ozone. However, changes in temperature, transport, and nitrogen and hydrogen ozone-loss cycles caused by the anticipated continued increase in well-mixed greenhouse gases (GHGs) are likely to affect this ozone “recovery” [e.g., *Haigh and Pyle*, 1979; *Brasseur and Hitchman*, 1988; *Rosenfield et al.*, 2002; *Chipperfield and Feng*, 2003; *Austin and Wilson*, 2006; *Eyring et al.*, 2007]. The probable influence of

climate change on the evolution of stratospheric ozone complicates quantification of the recovery process, including the interpretation of reaching different milestones that have been used to define ozone recovery.

[3] Traditionally the date that ozone returns to a specified historical value (e.g., pre-1980 values) has been used to describe full ozone recovery [e.g., *Rosenfield et al.*, 2002; *Shindell and Grewe*, 2002; *Austin and Wilson*, 2006]. However, this milestone does not require attribution to ODSs, and may be reached because of natural variability or other climate changes, or may never be reached even when all anthropogenic ODSs are removed from the atmosphere. This lack of attribution can lead to misinterpretation of observations [e.g., *Kane*, 2008]. To demonstrate recovery from halogen-induced destruction it is necessary to attribute changes in ozone to changes in ODSs and to changes in climate [*Shindell and Grewe*, 2002]. An alternative, attributed, milestone for “full ozone recovery from ODSs” introduced by *WMO/UNEP* [2007] is the date when ozone is no longer significantly affected by ODSs. Recent studies of the simulated ozone evolution through the 21st century [e.g., *Rosenfield et al.*, 2002; *Chipperfield and Feng*, 2003; *Austin and Wilson*, 2006; *Eyring et al.*, 2007; *Shepherd and Jonsson*, 2008] have not examined the milestone proposed in *WMO/UNEP* [2007].

[4] This study uses simulations of NASA’s Goddard Earth Observing System Chemistry-Climate Model (GEOS CCM) [*Pawson et al.*, 2008] to examine the impacts of stratospheric climate change on ozone recovery, focusing on differences in the dates when the above “full recovery” milestones are reached.

2. Model and Simulations

[5] The GEOS CCM includes representations of atmospheric dynamics, radiation, and stratospheric chemistry and their coupling through transport and radiative processes. *Pawson et al.* [2008] show that the climate structure and ozone in GEOS CCM agree quite well with observations. Two deficiencies are a high bias in total O₃ when chlorine loading is low (in the 1960s) and the anomalous nature of the Antarctic vortex [*Pawson et al.*, 2008], but these should have only a minor effect on the present analysis, which focuses on relative ozone changes. Additional evaluations of GEOS CCM [*Eyring et al.*, 2006; *Pertlitz et al.*, 2008; *Oman et al.*, 2008] reveal good comparisons with observations. Further, *Waugh and Eyring* [2008] demonstrate that GEOS CCM is one of the better-performing models from the CCMs included by *Eyring et al.* [2006].

[6] This study uses a pair of GEOS CCM simulations with identical surface concentrations of CO₂, N₂O, and

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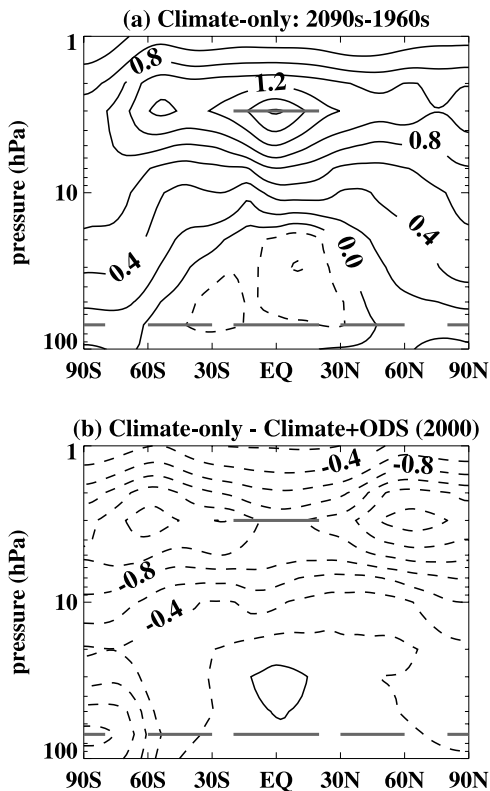


Figure 1. Difference in annual mean ozone (a) between 1960s (1960–1969 average) and 2090s (2090–2099 average) for *climate-only* simulation, and (b) for 1995–2005 between the *climate-only* and *climate + ODS* simulations, showing ozone loss due to ODSs. The contour interval is 0.2 ppm, and the grey horizontal bars show the regions shown in Figure 2.

CH₄, and sea surface temperatures (SST) but different ODS concentrations. The first simulation has ODSs fixed at their 1960 values, so that changes in O₃ are due to climate change (this is referred to as the *climate-only* simulation). In the second simulation the surface ODS concentrations vary with time according to the WMO/UNEP [2003] Ab scenario, and O₃ changes are due to changes in both climate and ODSs (this is referred to as the *climate + ODS* simulation). The difference in O₃ between the two simulations is then the ODS-induced change. These ODS-induced changes include both the direct ODS chemical impact and ‘indirect’ feedbacks, such as that in the upper stratosphere with respect to temperature decreases caused by ODS-induced ozone loss.

[7] The pair of simulations include separate runs for the past (1950–2004) and the future (2000–2099) [Pawson *et al.*, 2008; Perlwitz *et al.*, 2008]. The past runs used the observed surface concentrations of GHGs, and observed SSTs and sea ice concentrations from Rayner *et al.* [2003], whereas the future runs used GHGs from the IPCC scenario A1b, and SST and sea ice data from an AR4 integration of CCSM3.0 [Collins *et al.*, 2006] using the same GHG scenario. Output from 1960 to 2000 from the past run and 2001 to 2099 output from the future run were used to form

single time series for each of the *climate-only* and *climate + ODS* simulations.

3. Ozone Changes 1960 to 2100

[8] Figure 1 illustrates the latitude- and height-dependence of ozone change caused by GHG and ODS changes. The ozone change resulting from GHG increases between 1960 and 2100 in the *climate-only* simulation (Figure 1a) varies with location and is broadly consistent with that in prior studies, as discussed below. There is more O₃ in the upper stratosphere at the end of the 21st century than in the 1960s, but less O₃ in the tropical lower stratosphere and through the middle latitudes. Because the peak ODS loading of the stratosphere is expected to occur around 2000, the ozone change caused by ODSs is shown for 1995–2005 as the difference in ozone concentrations between the *climate-only* and *climate + ODS* runs (Figure 1b). In much of the stratosphere, the GHG-induced increase in O₃ between 1960 and 2100 has comparable magnitude to the ODS-induced loss for 1995–2005. The notable exception is in the Antarctic lower stratosphere where the ODS-induced ozone loss is much larger than that due to any other process in the lower stratosphere.

[9] Given the spatial variations in the impact of increased GHGs on O₃, it is necessary to consider the relative impacts of ODS and GHG changes in different regions separately. The spatial variation in the evolution of O₃ is illustrated in Figure 2 which shows O₃ from the *climate-only* (red solid curves) and *climate + ODS* (black) simulations for the regions marked by horizontal bars in Figure 1.

[10] In the tropical upper stratosphere (Figure 2a) the O₃ increase in the *climate-only* simulation occurs steadily from 1960 to 2100. This increase is due to cooling caused by increasing CO₂ and the resulting slowing of gas-phase chemical reactions that destroy O₃ [e.g., Haigh and Pyle, 1979; Brasseur and Hitchman, 1988; Shindell *et al.*, 1998; Rosenfield *et al.*, 2002]. This steady increase throughout the period does not occur in the *climate + ODS* simulation: O₃ decreases rapidly from 1960 to 2000, then increases through the 21st century. By 2100, O₃ is significantly higher (~20%) than in the 1960s, but slightly lower than in the *climate-only* simulation.

[11] The changes in tropical upper stratospheric O₃ due to ODSs (black dashed curve in Figure 2a) closely match the concentration of the equivalent effective stratospheric chlorine (EESC; blue curve in Figure 2a, plotted as a negative with respect to the dashed line to illustrate the correspondence with O₃). The EESC is defined as Cl_y + 60Br_y, and varies slightly from the more traditional calculation of EESC in that location specific Cl_y and Br_y is used rather than inferring these from surface halogen concentrations and assumptions about transport times. Comparing the changes due to ODSs with the O₃ change in the *climate-only* simulation shows that the relative contributions of ODSs and climate changes vary with time. The decrease in O₃ between 1960 and 2000 is due primarily to increases in ODSs, which is only partially offset by changes related to increases in GHGs. Changes in ODSs and GHGs both contribute to the O₃ increase over the 21st century.

[12] The long-term change in tropical lower stratospheric O₃ (Figure 2b) is very different from that in the upper

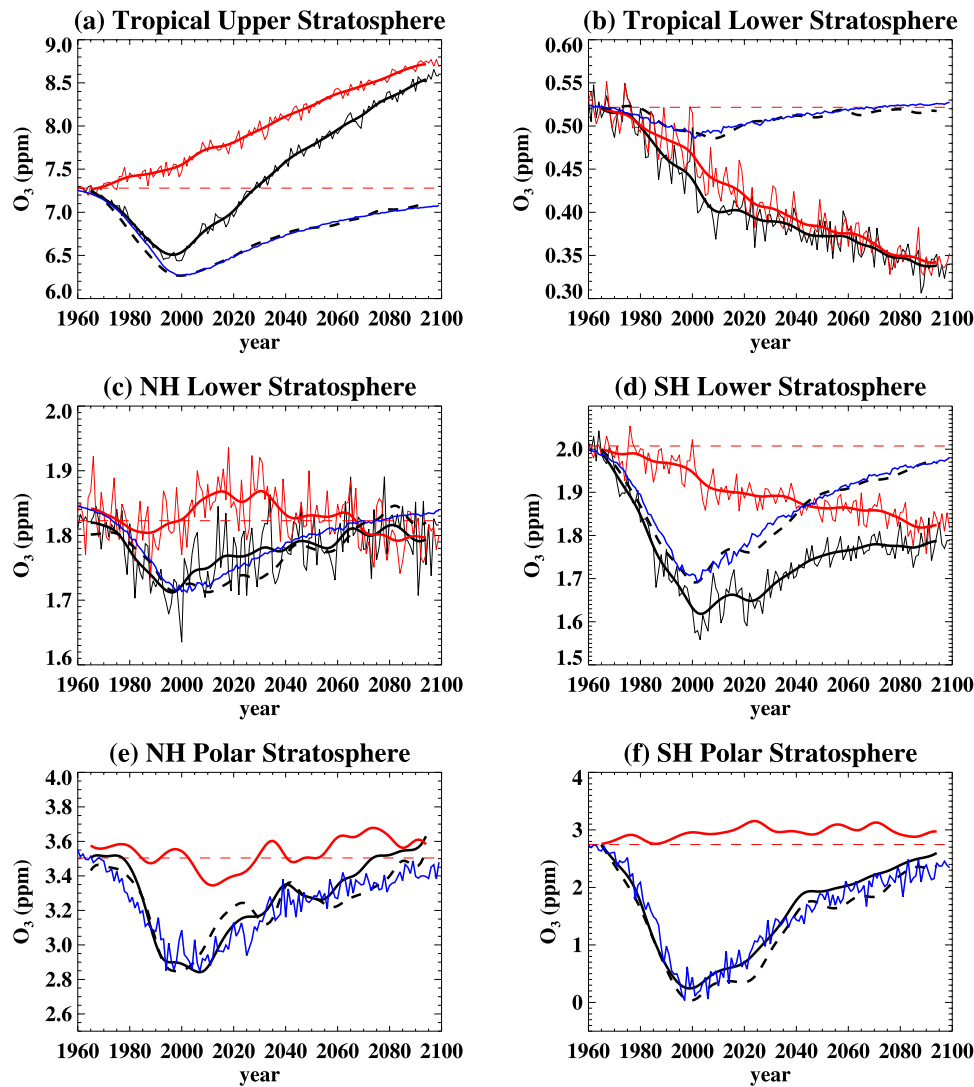


Figure 2. Evolution of O_3 from *climate + ODS* (black) and *climate-only* (red curves) simulations, for (a) 3hPa, 20°S–20°N, (b) 70hPa, 20°S–20°N, (c) 50hPa, 30°–60°N, (d) 50hPa, 30°–60°S, (e) 50 hPa, 80°–90°N, and (f) 50 hPa, 80°–90°S. Figures 2a–2d show annual-mean, Figure 2e shows March-mean, and Figure 2f shows October-mean values. Thick curves are smoothed versions of the thin curves, calculated by applying a 1:2:1 filter iteratively 30 times [see Eyring *et al.*, 2007]. For clarity, only smoothed curves are shown for the O_3 fields in Figures 2e and 2f. The black dashed curves show the O_3 loss due to ODSs, calculated as differences between the red and black thick curves, added to 1960 O_3 (horizontal dashed line). The blue curve shows the evolution of the negative of EESC, $Cl_y + 60Br_y$, which has been scaled so the minimum and maximum values match those of the black dashed curve.

stratosphere. In both the *climate-only* and *climate + ODS* simulations there is a steady decrease in the decadal-averaged O_3 . As found in other CCM simulations [e.g., Butchart *et al.*, 2006; Austin *et al.*, 2007; Garcia and Randel, 2008], there is an increase in tropical upwelling in both these GEOS CCM simulations [Oman *et al.*, 2009], which causes the tropical O_3 decrease. Although the variations in the decadal-averaged tropical lower stratospheric O_3 are similar in the two simulations, there are some differences (see black dashed curve in Figure 2b) that indicate that changes in ODSs do play a minor role in O_3 changes in this region. These O_3 changes are, however, related to polar O_3 influence on tropical upwelling rather than in-situ halogen chemistry [Oman *et al.*, 2009].

[13] The evolution of O_3 in the mid-latitude lower stratosphere differs from that in the tropical lower stratosphere, and also differs between hemispheres (Figures 2c and 2d). In southern mid-latitudes the O_3 in 2100 in the *climate + ODS* simulation is much lower than it was in the 1960s, whereas in the northern mid-latitude the values in 2100 are similar to 1960 values. The evolution of halogen-induced changes (black dashed curves) is similar between hemispheres and closely matches the evolution of EESC. However, the impact of GHGs (red curves) differs between hemispheres, and is the primary cause of the inter-hemispheric differences in the *climate + ODS* simulation. Lower stratospheric O_3 averaged over southern mid-latitudes decreases in the *climate-only* simulation, whereas the northern mid-

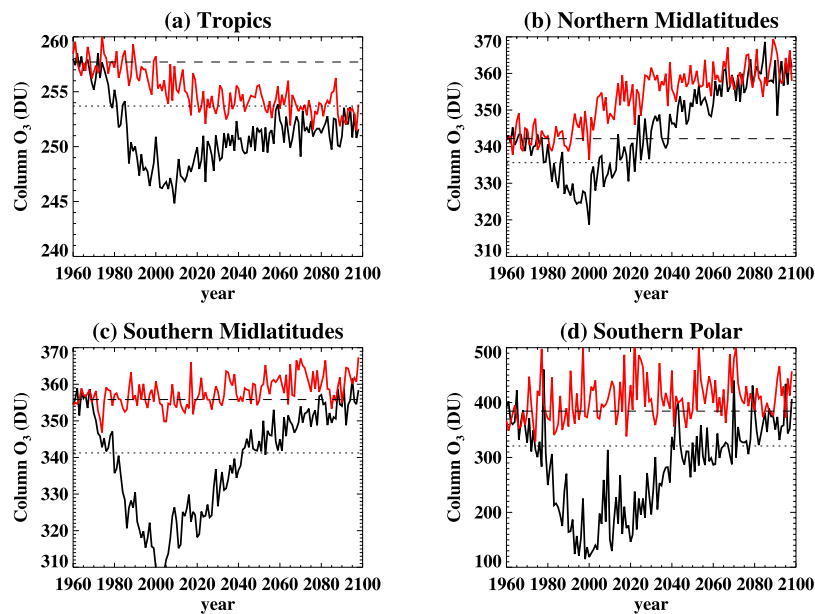


Figure 3. As in Figure 2 except for total column O_3 for (a) 20°S – 20°N , (b) 30° – 60°N , (c) 30° – 60°S , and (d) 80° – 90°S .

latitude average shows no significant long-term trend in the same simulation (see also Figure 1a). This inter-hemispheric difference is likely related to differences in transport [Li *et al.*, 2008].

[14] In polar regions there are larger seasonal variations in long-term O_3 trends than in the tropics and mid-latitudes, with much larger changes in spring than other seasons. We therefore focus on spring polar O_3 . In both hemispheres spring O_3 decreases from 1960s to minimum values around 2000, and then increases to around or slightly less than 1960s values by 2100, see Figures 2e and 2f). This long-term evolution is dominated by the rise and fall in ODS, with only small decadal-scale variations in the *climate-only* O_3 . Changes in GHGs have the potential, through changes in polar temperatures, to modify polar O_3 depletion, but increases in GHGs do not have a significant impact on polar temperatures or O_3 in the GEOS CCM simulations.

[15] There are also differences in the evolution of total column O_3 between latitude regions. In all regions column O_3 in the *climate + ODS* simulation decreases rapidly from 1960 to around 2000, and then increases through the 21st century (Figure 3). However, there are quantitative differences in the rate of increase over the 21st century between regions. As might be expected from differences in the evolution of lower stratospheric O_3 , the increase over the 21st century is most rapid for northern mid-latitudes and slowest for the tropics.

4. Recovery Dates

[16] The analysis above highlights how the climate and ODS impacts lead to spatial variations in the O_3 evolution. We now examine differences in the date that recovery milestones occur. For both the date O_3 returns to historical values (the traditional definition of full recovery) and the date when O_3 is no longer significantly affected by ODSs (the *WMO/UNEP* [2007] definition of full recovery due

to ODSs) reference levels need to be defined. Traditionally 1980 values are used as the reference levels for full recovery. However, there had already been a significant build up of anthropogenic ODSs by 1980, and referencing to an earlier date (e.g., 1960) might be a better measure of anthropogenic activities. We therefore consider recovery relative to both 1960 and 1980 levels.

[17] To apply the *WMO/UNEP* [2007] definition, “significant impact on O_3 ” needs to be defined. If pre-1960 levels of ODSs are deemed not to have a significant impact on O_3 then the convergence of (decadal-averaged) O_3 from the *climate + ODS* simulation with that from the *climate-only* simulation (with ODSs fixed at 1960 levels) can be used to define “full recovery due to ODSs”. If the sensitivity of O_3 to EESC does not change with time this milestone is equivalent to the return of EESC in the *climate + ODS* simulation to 1960 levels. This is generally the case in the GEOS CCM simulations, where the evolution of EESC matches the changes in O_3 due to ODSs (see Figures 2c, 2e, and 2f), and the return of EESC to historical values can be used as a proxy for “full recovery due to ODSs”. This latter method has the advantage that it can be used to define recovery relative to pre-1980 levels of ODSs without a simulation with ODSs fixed at 1980 value, and will be used here.

[18] The relative timing of the two recovery milestones varies between regions, for both 1960 and 1980 reference levels. In the upper stratosphere O_3 returns to 1980 (1960) values around 2015 (2030), but EESC does not return to 1980 (1960) values until around 2060 (2100) (Figure 2a). This means that full recovery from ODSs occurs many decades later than the return of O_3 to historic levels (around 45 (70) years later if referenced to 1980 (1960) values).

[19] The opposite difference is found in the tropical and southern mid-latitude lower stratosphere. Here, EESC is below 1960 levels by 2100, but O_3 has yet to return to even 1980 values by this date (Figures 2b and 2d). In fact, these

simulations suggest O₃ in these regions may never return to 1980 values, even after removal of anthropogenic ODSs.

[20] In contrast to the above regions, there is limited net impact of climate change in the lower stratosphere northern mid-latitudes (Figure 2c) and the polar lower stratosphere in spring (Figures 2e and 2f), and the two milestones occur around the same time.

[21] There are also differences in the occurrence of the recovery milestones when considering total column O₃. Column O₃ from the two GEOS CCM simulations converge around the end of the 21st century for all regions, but there is a large variation in the return of column values to historical values. For example, in northern mid-latitudes column O₃ returns to 1960 values before 2040, but tropical column O₃ is still below 1960 (and even 1980) values at the end of the 21st century (e.g., Figure 3). Thus, even for column O₃ which recovery milestone occurs first varies between regions.

5. Conclusions

[22] GEOS CCM simulations indicate that stratospheric climate change caused by projected increases in GHGs will have a large impact on stratospheric O₃ recovery, and that the impact will differ substantially between regions. This means that climate change must be considered when examining O₃ recovery and that different regions need to be examined separately to create a meaningful picture of the recovery process.

[23] This is especially true for interpreting established milestones of full ozone recovery. These milestones may be reached at different times, and which occurs first will likely vary between regions. In some regions the milestone of O₃ returning to 1960 (or 1980) values may be reached many decades before the milestone of full recovery from ODSs, e.g., upper stratospheric O₃ and northern mid-latitudes column O₃ in the GEOS CCM simulation exceed 1960 values over 60 years before the ODSs return to 1960 values. However, in other regions O₃ may never return to 1980 values even when anthropogenic ODSs have been removed from the atmosphere, e.g., tropical and southern mid-latitude lower stratosphere O₃ in the GEOS CCM simulation. These examples show that the milestone of O₃ returning to 1960 (or 1980) values should not necessarily be interpreted as O₃ recovery from the effects of anthropogenic ODSs, and to demonstrate O₃ recovery from anthropogenic ODSs it is necessary to make an attribution to ODSs in the recovery milestone [WMO/UNEP, 2007].

[24] The results presented above are based on a single CCM and also single ODS and GHG scenarios: Other models and other GHG scenarios may yield alternative projected dates that the different milestones are reached. However, the O₃ changes in the GEOS CCM *climate + ODS* simulation are qualitatively similar to that in other CCMs [see Eyring *et al.*, 2007], and it is likely that the qualitative features presented here are robust. It would nevertheless be of great value to examine the relative role of climate change and ODSs in other CCMs and in simulations using other GHG scenarios. It is also of interest to quantify the relative role of different GHG-related mechanisms in causing changes in O₃, i.e. role of changes in temperatures, circulation, and nitrogen and hydrogen ozone-loss cycles.

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