Is Antarctic climate most sensitive to ozone depletion in the middle or lower stratosphere?

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[1] Antarctic stratospheric ozone depletion has been associated with an observed downward trend in tropospheric geopotential height and temperature. Stratospheric ozone depletion peaks in October–November, whereas tropospheric trends are largest in December–January, concurrent with maximum ozone changes close to the tropopause. Surface temperatures are most sensitive to ozone loss near the tropopause, therefore it has been suggested that the observed tropospheric response is forced mainly by ozone depletion in the lower stratosphere. In this study the climate response to ozone depletion exclusively below 164 hPa is simulated using HadSM3-L64, and compared with simulations in which ozone depletion is prescribed exclusively above 164 hPa. Results indicate that the tropospheric response is dominated by ozone changes above 164 hPa, with ozone changes in the lowermost stratosphere playing an insignificant role. A tropospheric response is also seen in fall/winter which agrees well with observations and has not been found in modeling studies previously. **Citation:** Keeley, S. P. E., N. P. Gillett, D. W. J. Thompson, S. Solomon, and P. M. Forster (2007), Is Antarctic climate most sensitive to ozone depletion in the middle or lower stratosphere?, *Geophys. Res. Lett.*, *34*, L22812, doi:10.1029/2007GL031238.

1. Introduction

[2] In recent decades, Antarctic stratospheric ozone depletion has caused a strong cooling and strengthening of the Antarctic vortex during spring [*Ramaswamy et al.*, 2001; *Zhou et al.*, 2000; *Shindell and Schmidt*, 2004; *Sexton*, 2001; *Baldwin et al.*, 2007]. Observations [*Thompson and Solomon*, 2002] and models [*Kindem and Christiansen*, 2001; *Sexton*, 2001; *Gillett and Thompson*, 2003] have shown a significant tropospheric response to these changes, with decreases in Antarctic geopotential height and temperature occurring around one month after the maximum stratospheric cooling at 70 hPa. While initial studies suggested that the lag may arise as a result of dynamical processes analogous to those associated with the downward-propagating response to stratospheric anomalies [*Baldwin and Dunkerton*, 1999; *Thompson et al.*, 2005], more recent work has suggested an alternate hypothesis. Observations from radiosonde data from Syowa and South Pole show that ozone losses at lower altitudes occur about a month later than the maximum depletion at 18 km (70 hPa) which occurs in October [*Solomon et al.*, 2005]. It has been shown that the surface temperature is most sensitive to ozone losses (with the same local percentage change) at 11 km, near the tropopause, and also in the region of 20 km; there is a local minimum response to percentage ozone changes around 18 km [see *Forster and Shine*, 1997, Figure 11]. *Baldwin et al.* [2007] hence conclude that "this suggests that surface cooling is radiatively induced, and that the apparent lag between stratospheric and tropospheric responses is due to the downward transport of ozone-depleted air toward the tropopause, rather than any dynamical effect". In this study we aim to test this hypothesis by simulating the response to lower stratospheric (tropopause region) ozone losses compared to those at higher altitudes where the maximum ozone depletion occurs. We also examine changes in the components of the thermodynamic energy equation to explore the causes of the tropospheric cooling.

2. Experimental Setup

[3] We use a high vertical resolution version (64 vertical levels) of version 4.5 of the Hadley Centre Atmospheric model coupled to a 50-m slab ocean. The model extends to 0.01 hPa and has 30 levels between 240 hPa and 1 hPa. This model is almost identical in setup to HadSM3-L64 as used by *Gillett and Thompson* [2003] (hereinafter referred to as GT03), which used version 4.4 of the Hadley Centre model, and we use the same monthly varying, zonal mean ozone forcing fields, based on *Randel and Wu* [1999] ozone trends over the period 1979–1997. As a mixed-layer ocean is used we consider the difference between a control and perturbed equilibrium run which are both 40 years long. The control experiment is run with an ozone field which is comparable to 1970s ozone concentrations. There are three perturbation experiments; the first with ozone depletion throughout the depth of the stratosphere, the second with depletion restricted to the region below 164 hPa and the third with depletion only at and above 164 hPa. In all cases, wherever the ozone is depleted its concentration is representative of the late 1990s. Percentage changes in ozone are shown in Figure 1. It should be noted that absolute changes in column ozone associated with the midstratospheric depletion are much larger than those associated with the lower stratospheric depletion, since the ozone concentration is much higher in the midstratosphere. The maximum column ozone change associated with the lower stratospheric depletion is only 13% of the maximum column ozone change associated with the mid-stratospheric depletion at 70°S. The model level of

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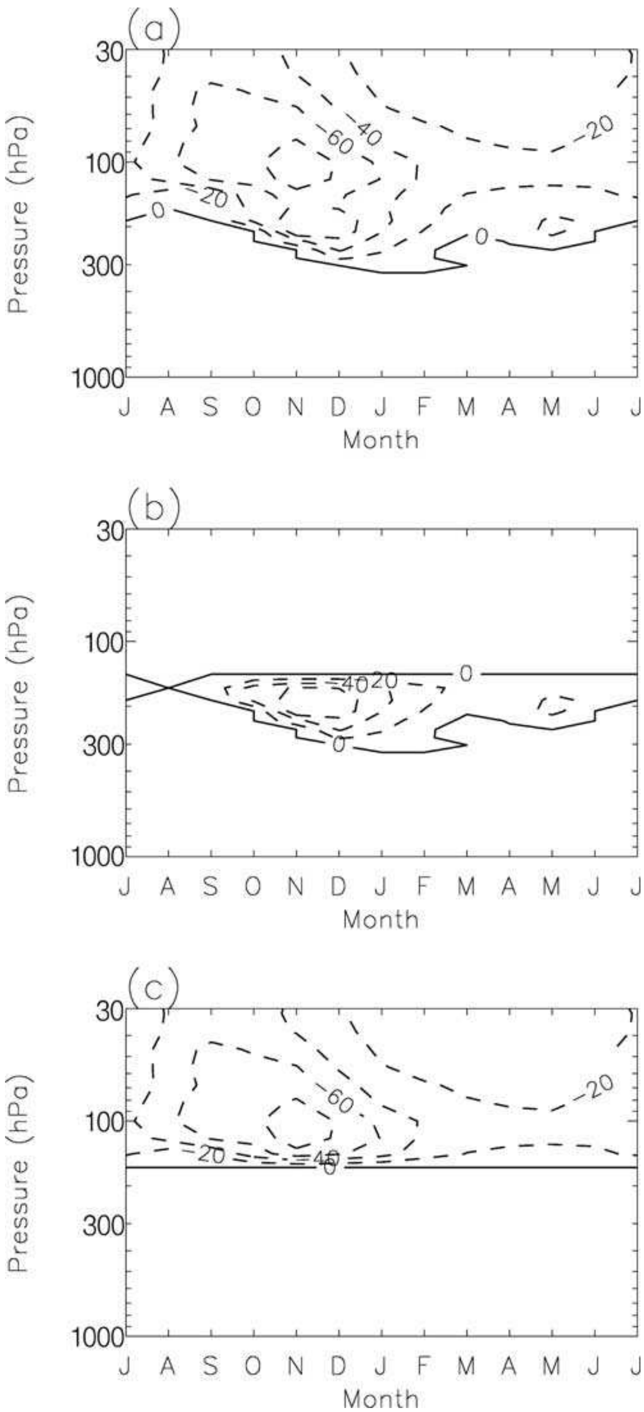

Figure 1. Percentage change in ozone for each month and pressure level at 70°S for (a) the whole stratosphere, (b) the lowermost stratosphere below 164 hPa (the ozone concentration in the troposphere remains constant so the zero contour marks the climatological tropopause in Figure 1a), and (c) the midstratosphere, above 164 hPa. The ozone changes shown in Figure 1a are identical to those used by GT03.

164 hPa was chosen as a representative level separating the lower and middle stratosphere, and is above the 200 hPa level examined by *Solomon et al.* [2005]. Except for the ozone depletion, the control and perturbed model integrations are identical. In each case the model is run for 40 years and data are taken from the last 30 years of the model integration, after the model has reached equilibrium.

3. Results

[4] The monthly zonal mean response of geopotential height and temperature averaged over the region south of 65°S is found by taking the difference between the control and perturbed experiments, each averaged over the 30 years of model data. Figures 2a and 3a show the response when ozone is depleted throughout the depth of the stratosphere - this is a repeat of the experiment carried out by GT03. The results show a similar response to the previous results of GT03, although the tropospheric response seems to occur later than in simulations of GT03: Any differences must result either from our use of a different model version, our use of a different computer, or natural variability. The effect of ozone depletion is to strengthen the polar vortex and maximum changes in geopotential height occur in the stratosphere in November coinciding with the maximum ozone depletion (Figure 1). The maximum cooling response is also seen in the stratosphere in November and the timing and spatial location coincide with the maximum ozone depletion. A delayed response is seen in the troposphere in January–February in temperature and geopotential height. There is also a significant tropospheric temperature response in May, coinciding with a small local maximum in ozone depletion (Figure 1) which was not seen in the previous modeling study performed by GT03. This is, however, a robust result seen even when the data is sub-sampled, and corresponds well to the observational results of *Thompson and Solomon* [2002]. Figures 2b and 3b show the response when the ozone is only perturbed in the region below 164 hPa. The temperature is only significantly affected locally and the response is dominated by the shortwave cooling in that region (not shown). There is no significant geopotential response to ozone changes within the stratosphere and the apparently significant response in the troposphere in September is not found to be robust when sub-samples of the data are analyzed.

[5] Figures 2c and 3c show the geopotential height and temperature response when the ozone is depleted from 164 hPa upwards. The pattern of response in space and time is very similar to that of the full perturbed ozone simulations shown in Figures 2a and 3a. The geopotential height change in the troposphere is significant at the surface in this experiment throughout summer and a large change in geopotential height is simulated in the midtroposphere. The temperature response to depletion above 164 hPa shown in Figure 3c is similar to that of 3a, with only slightly reduced cooling in the lowermost stratosphere. Overall there is very little evidence of any significant differences between the geopotential height (Figure 2d) and temperature (Figure 3d) responses to the midstratospheric depletion only and the full stratospheric ozone depletion. A comparison of the temperature response to lower stratospheric depletion only (Figure 3b) with the difference between response to full stratosphere depletion and the response to midstratosphere depletion (Figure 3d) indicates that the responses to ozone depletion in the two regions add linearly.

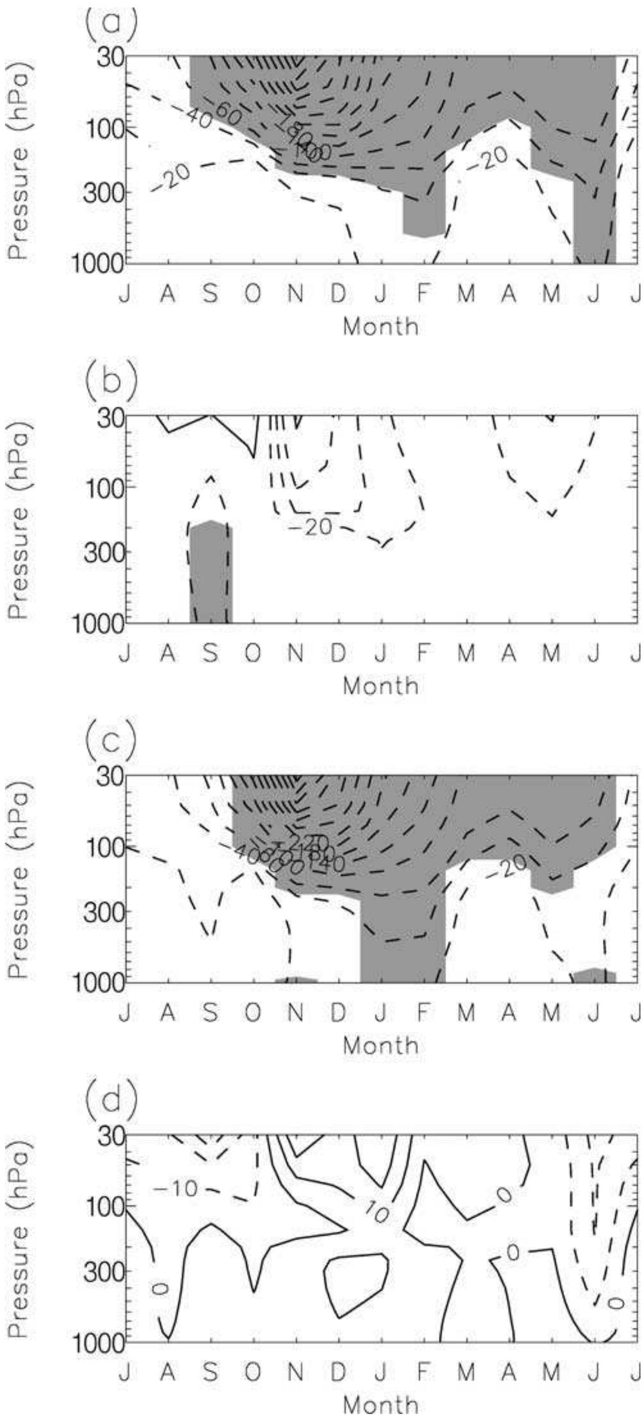

Figure 2. Average monthly mean geopotential height response (m) for the region south of 65°S. Anomalies relative to the control simulation are shown for the (a) full, (b) lower, and (c) midstratospheric ozone depletion experiments. (d) The difference between simulated geopotential height in the full and midstratospheric depletion experiments. Shading indicates regions of significance at the 95% level based on a two-sample t-test.

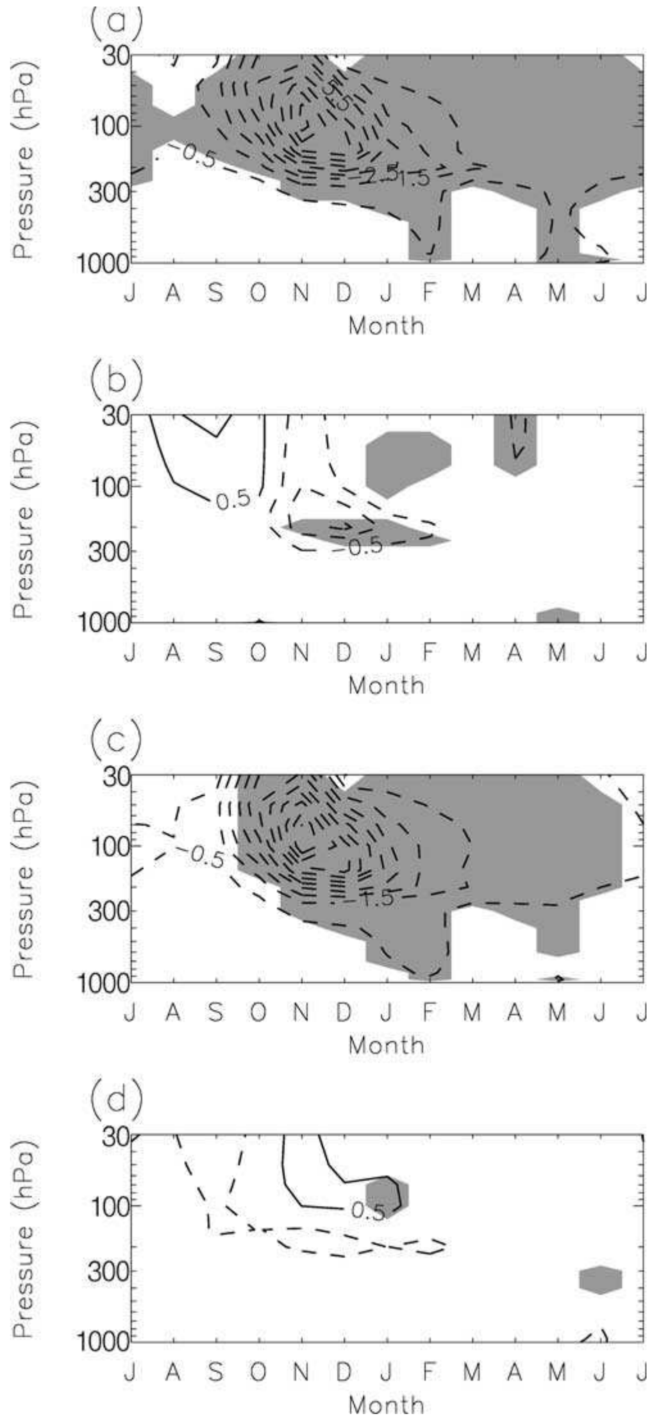

Figure 3. As Figure 2, except for the temperature response (K). The contour level separation is 1 K.

[6] The results described so far indicate that the tropospheric temperature response is driven mainly by ozone depletion in the midstratosphere, but they do not indicate whether the temperature change is predominantly driven by dynamical or radiative processes. To test this we calculated means of all the terms in the thermodynamic energy equation for each month of each simulation, and evaluated their changes in response to the prescribed ozone depletion. We used the Eulerian zonal mean thermodynamic energy equation [*Holton*, 2004, equation 10.12], but retained terms associated with spherical geometry, horizontal advection by the mean meridional circulation and vertical eddy flux

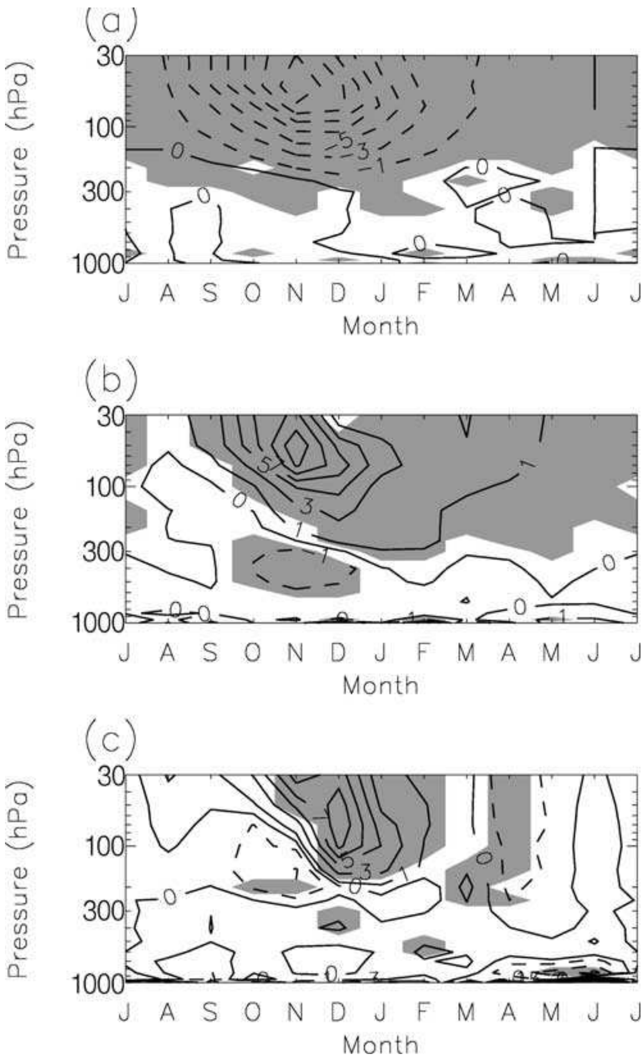

Figure 4. Heating components of the full ozone change (shown in Figure 1a) for (a) shortwave, (b) longwave, and (c) dynamical heating. Contour interval 2 K month^{-1} (zero contour marked as well). Shading indicates regions of significance at the 95% level based on a two-sample t-test.

divergence. Expressing this equation in terms of p, the pressure, ω, the vertical wind speed in Pa s^{-1} and potential temperature θ we obtain:

$$\frac{\partial \overline{T}}{\partial t} = \frac{\overline{J}}{c_p} - \frac{\overline{v}}{a} \cdot \frac{\partial \overline{T}}{\partial \phi} - \left(\frac{p}{p_s}\right)^{\kappa} \cdot \overline{\omega} \frac{\partial \theta}{\partial p} - \frac{1}{a} \cdot \frac{\partial \overline{v'T'}}{\partial \phi} + \frac{\tan \phi}{a} \cdot \overline{v'T'} - \frac{\partial \overline{\omega' T'}}{\partial p} + \frac{\kappa \overline{\omega' T'}}{p} \quad (1)$$

where an overbar indicates a zonal mean quantity and a prime is the deviation from the zonal mean. T is temperature, t is time, $\frac{J}{c_p}$ is the total radiative heating term, v is the meridional wind speed, a is the radius of the Earth, p_s the surface pressure, κ is equal to $\frac{c_p - c_v}{c_p}$ and c_p and c_v are the specific heat capacity of dry air at constant pressure and volume respectively. We also found that it was necessary to output the vertical velocity and temperature on the model grid, rather than on pressure levels, otherwise interpolation errors prevented the thermodynamic energy equation from balancing. While tropospheric radiative heating changes are diagnosed directly by the model and therefore robust, the heat budget in the troposphere did not close exactly, perhaps due to undiagnosed latent heating and boundary layer heating contributions, therefore tropospheric dynamical heating changes should be interpreted with caution. The main response to ozone depletion in the lowermost stratosphere was found to be a small local shortwave cooling, and the responses to ozone depletion above and below 164 hPa were found to add linearly, therefore we show only the components of the heating response to full stratospheric ozone depletion here.

[7] Figure 4 shows the change in heating rates for shortwave, longwave and dynamical heating terms for ozone depletion in the full stratosphere (as shown in Figure 1a). As expected the ozone depletion produces a local shortwave cooling effect (Figure 4a), due to reduced absorption of ultraviolet solar radiation [*Fels et al.*, 1980; *Forster and Shine*, 1997; *Baldwin et al.*, 2007]. The resulting decrease in temperature leads to reduced emission in the longwave, giving a warming effect in the stratosphere due to longwave radiation (Figure 4b). Ozone is also an effective absorber and emitter of longwave radiation, and its changed concentration must therefore also have an effect on the longwave heating, but it is impossible to separate the effects of temperature and ozone change on longwave heating in this experiment. The net change in dynamical heating is shown in Figure 4c. The changes in stratospheric dynamical heating are dominated by changes in the meridional eddy heat flux, $v'T'$ (not shown). As *Rosier and Shine* [2000] conclude, the dynamical response to ozone depletion has a warming influence on the stratosphere in December. However, they do not investigate the seasonality of this response, and our results indicate that while a large warming is indeed simulated in December and January, the dynamics act to cool the Antarctic lower stratosphere in October and November. Comparison with the mean seasonal cycle indicates that this dipole in the dynamical heating response to ozone depletion corresponds to a delay in the peak dynamical heating associated with the final warming. A delay in the Antarctic final warming has indeed been observed [*Waugh et al.*, 1999], and a preliminary analysis of trends in upward EP-flux at 100 hPa averaged south of 79°S in the NCEP reanalysis indicates a decrease in wave driving in November, followed by an increase in December and January (M. Rex, personal communication, 2007), consistent with the response simulated here.

[8] In the troposphere, the dominant response is one of longwave cooling in November and December; this cooling is significant throughout the troposphere and extends to the surface, which is at ~700 hPa at the South Pole. However, the longwave cooling precedes the largest surface temperature changes by several months (compare Figures 3a and 4b), so it appears the longwave forcing cannot be driving all of the simulated surface temperature changes. Dynamical heating changes do not appear to play a large role in the tropospheric cooling. Due to interactions of changes in wind speed with the surface inversion layer and radiative effects at the surface, results may be different for surface temperature. We restrict our attention to the free troposphere here, and will address surface effects in future work. The longwave cooling response in the troposphere seen in Figure 4b

is not found in response to the lower stratosphere only ozone depletion, but is seen when ozone in the region above 164 hPa is depleted (not shown), indicating that the changes in tropospheric longwave radiation are coming mainly from ozone changes above 164 hPa.

4. Conclusions

[9] Increasing evidence suggests that Antarctic stratospheric ozone depletion can induce a tropospheric response, which appears to lag the maximum stratospheric ozone depletion by 1–2 months [*Baldwin et al.*, 2007]. We also find that ozone depletion contributes to the change in geopotential and in circulation not only in summer, but also to the observed changes in May/June [*Thompson and Solomon*, 2002]. Such a response was not seen in GT03, but the simulations analyzed here are 10 years longer, which may improve the signal-to-noise ratio, but further investigation is required to determine if this result is robust. However the mechanism underlying these tropospheric responses, and the reason for its lag compared to the stratospheric forcing, remain open to debate. Based on an analysis of observations by *Solomon et al.* [2005], showing that ozone depletion close to the tropopause peaks in December and January, *Baldwin et al.* [2007] suggest in the WMO Ozone Assessment that the tropospheric response to ozone depletion may be an instantaneous radiative response to ozone changes close to the tropopause. Our results indicate that this is not the case, and that the tropospheric response is dominated by ozone changes above 164 hPa, with depletion below this level having no significant effect on tropospheric climate. However our results are consistent with the suggestion that decreased downwelling longwave contributes to driving the tropospheric cooling.

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