Article:
Forster, PM, Bodeker, G, Schofield, R et al. (2007) Effects of ozone cooling in the tropical lower stratosphere and upper troposphere. Geophysical Research Letters, 34 (23). ISSN: 0094-8276

https://doi.org/10.1029/2007GL031994

Effects of ozone cooling in the tropical lower stratosphere and upper troposphere

Piers M. Forster,[1] Greg Bodeker,[2] Robyn Schofield,[3,4,5] Susan Solomon,[3] and David Thompson[6]

Received 11 September 2007; accepted 22 October 2007; published 13 December 2007.

[1] In this paper, we examine the tropical lower stratosphere and upper troposphere and elucidate the key role of ozone changes in driving temperature trends in this region. We use a radiative fixed dynamical heating model to show that the effects of tropical ozone decreases at 70 hPa and lower pressures can lead to significant cooling not only at stratospheric levels, but also in the "sub-stratosphere/upper tropospheric" region around 150–70 hPa. The impact of stratospheric ozone depletion on upper tropospheric temperatures stems from reduced longwave emission from above. The results provide a possible explanation for the long-standing discrepancy between modeled and measured temperature trends in the uppermost tropical troposphere and can explain the latitudinal near-homogeneity of recent stratospheric temperature trends. **Citation:** Forster, P. M., G. Bodeker, R. Schofield, S. Solomon, and D. Thompson (2007), Effects of ozone cooling in the tropical lower stratosphere and upper troposphere, *Geophys. Res. Lett.*, *34*, L23813, doi:10.1029/2007GL031994.

1. Introduction

[2] Stratospheric temperature trends provide key evidence for attributing climate change to humans [e.g., *Ramaswamy et al.*, 2006]. Earlier discrepancies between satellite and model temperature trends in the free troposphere are now largely resolved [*Climate Change Science Program* (*CCSP*), 2006]. However, in the tropical upper troposphere and lower stratosphere, unresolved differences remain between observed and expected trends. A characteristic of climate models is that their temperature response in the troposphere increases with altitude, typically extending up to 100 hPa. In contrast, radiosonde temperature trend analyses show temperature increases only extending up to ~200 hPa with cooling above that level [e.g., *Santer et al.*, 2005; *CCSP*, 2006]. This issue remains a concern in detection/attribution studies, and is a primary focus of this paper.

[3] A second focus of the paper is explaining the latitudinal homogeneity of recent stratospheric temperature trends. Previous studies have shown that greenhouse gases alone do not reproduce the cooling that has been observed in the tropical lower stratosphere. In this paper we examine in more detail the role of ozone changes in such cooling.

[4] The region between typical convective outflow and the cold-point tropopause, can be referred to Tropical Tropopause Layer (TTL) or the tropical sub-stratosphere [*Folkins et al.*, 1999; *Gettelman and Forster*, 2002; *Thuburn and Craig*, 2000] and has many stratosphere-like attributes. Temperatures here appear to be predominantly controlled by radiative rather than convective processes. A recent study also found that tropical tropopause height changes are more closely related to stratospheric temperatures than to those in the troposphere [*Seidel and Randel*, 2006]. Climate models appear to overestimate the convective response in the tropical upper troposphere and thus their temperature changes here are closely tied to the moist-adiabatic temperature response [*Mitas and Clement*, 2006; *Cordero and Forster*, 2006; *Santer et al.*, 2003]. For these reasons we perform our attribution experiments with a detailed and tested radiative response model, which deliberately does not include dynamical or convective components. However, tropical upwelling could be increasing by a few percent per decade in the lower stratosphere [e.g., *Butchart et al.*, 2006; *Nikulin and Karpechko*, 2005] so we perform sensitivity tests to examine the possible role of these changes.

2. Datasets

[5] Key to our approach is using accurate and detailed ozone trends at high vertical resolution. These trends were calculated by applying a least squares variance-weighted regression model [*Bodeker et al.*, 2001] to version 6.2 Stratospheric Aerosol and Gas Experiment (SAGE) II ozone profiles [*McCormick et al.*, 1989]. The publicly available ozone profiles were subjected to further quality assurance to ensure that erroneous data were excluded from the trend analysis as prescribed by *Wang et al.* [1996, 2002]. Additionally, individual ozone profiles measured under cloudy or high aerosol loading conditions (i.e. conditions due to Mt Pinatubo) were screened according to recommendations [*Rind et al.*, 2005]. Variance weighted monthly mean, and latitude mean (5°), number density profiles of ozone, and their variances, were calculated and used as input to the trend analysis.

[6] Trends were calculated for 1984–2005 using a second order auto-correlation regression model. The quasi-biennial oscillation, solar cycle, and El Niño Southern Oscillation,

[1]School of Earth and Environment, University of Leeds, Leeds, UK.
[2]National Institute of Water and Atmospheric Research, Lauder, New Zealand.
[3]Chemical Sciences Division, NOAA Earth System Research Laboratory, Boulder, Colorado, USA.
[4]Cooperative Institute for Research in Environmental Sciences, University of Colorado, Boulder, Colorado, USA.
[5]Now at Alfred Wegener Institute for Polar and Marine Research, Potsdam, Germany.
[6]Department of Atmospheric Science, Colorado State University, Fort Collins, Colorado, USA.

0094-8276/07/2007GL031994

offset and trend basis functions were included, accounting for seasonal variability, as detailed by *Bodeker et al.* [1998]. A second order autocorrelation of the residuals was applied in the calculation of the coefficient uncertainties [*Bodeker et al.*, 2001]. The ozone profile trends are shown in Figure 1a and are similar to those given by *Randel et al.* [1999]. It is not the goal of this paper to identify the cause(s) of these tropical ozone decreases, which are a matter of current research, but rather to examine their implications for tropical temperature trends. While tropical ozone profile data suggest decreases in the lower stratosphere, total ozone column data appear to display little trend there, suggesting the possibility of cancellation at other altitudes [see, e.g., *Randel et al.*, 1999]. Note, that at altitudes lower than 18 km (roughly 70 hPa) SAGE trends become more uncertain and less statistically significant [see also *Randel and Wu*, 2007], but as will be shown below, our key findings are insensitive to changes in this uncertain region.

[7] As in previous studies [e.g., *Forster and Shine*, 2002] we use the Reading Narrow Band Model (RNBM) to compute temperature changes, assuming fixed dynamical heating. We make use of recent line databases and the latest ozone absorption cross sections. Timeseries of the long-lived greenhouse gases (LLGHGs), CO_2, N_2O, CH_4, CFC11 and CFC12, are also employed in the model, using data from *Forster et al.* [2007]. Of these CO_2 has the dominant cooling effect in the upper stratosphere, while the CFC changes contribute a small warming effect at the tropopause (see *Forster and Joshi* [2005] for further details). The model was run to equilibrium for each month in the two end years of the linear trend analysis and temperatures compared to determine the temperature trend per decade. Vertical resolution was around 500 m in the tropopause region. Temperatures were adjusted at pressures lower than 300 hPa. The choice of this level had minimal impact on the results.

[8] We compared our temperature trends to those from radiosonde analyses using subsets of the global radiosonde network as also used by *Thompson and Solomon* [2005] (the radiosonde data used by *Thompson and Solomon* [2005] and reproduced here are virtually identical to the RATPAC data described by *Free et al.* [2005]). We also used the AMSU/MSU T4 satellite data which covers the region from about 150 to 15 hPa (the T4 data used here were obtained from Remote Sensing Systems). Data from these are compiled into monthly average temperature anomalies and a similar linear trend analysis to the ozone case applied. While the radiosonde climatologies incorporate various adjustments to account for data inhomogeneities [*Free et al.*, 2004], a recent analysis by *Randel and Wu* [2006] suggests a systematic cold bias in the RATPAC tropical lower stratospheric data compared to the T4 satellite observations. This potential cold bias in the tropical lower

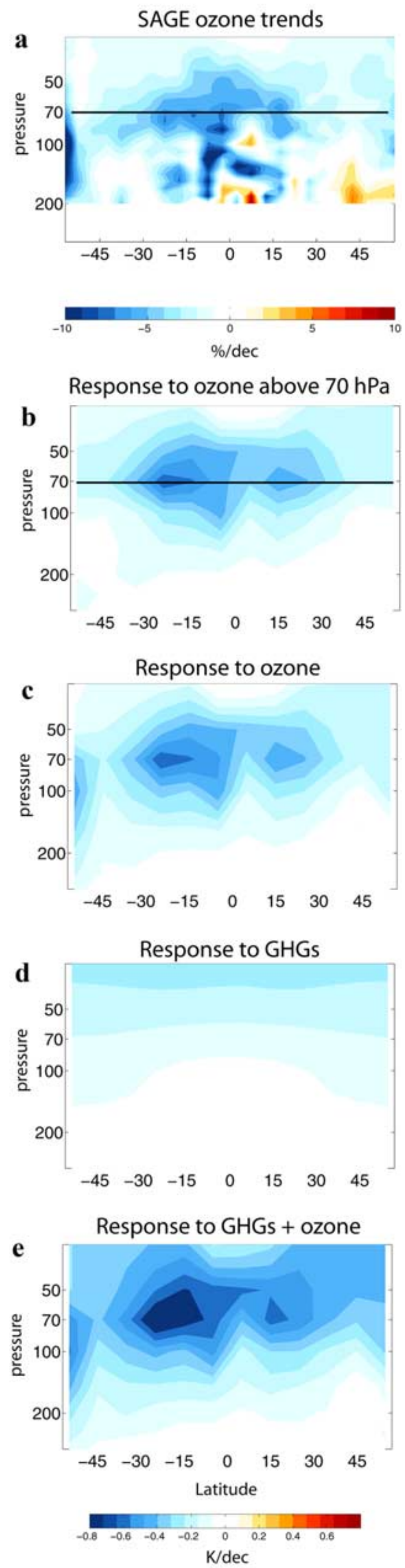

Figure 1. (a) 1984–2005 SAGE II linear ozone trends in % per decade. Other panels show 1984–2005 linear temperature trends (K/decade) from fixed dynamical heating calculations with the RNBM model. Calculations employ various ozone and LLGHG trends. (b) Ozone trends employed at pressures less than 70 hPa. (c) Ozone trends employed at pressures less than 200 hPa. (d) LLGHG changes (CO_2, CH_4, N_2O, CFC11, and CFC12). (e) LLGHG changes and ozone changes from Figure 1c.

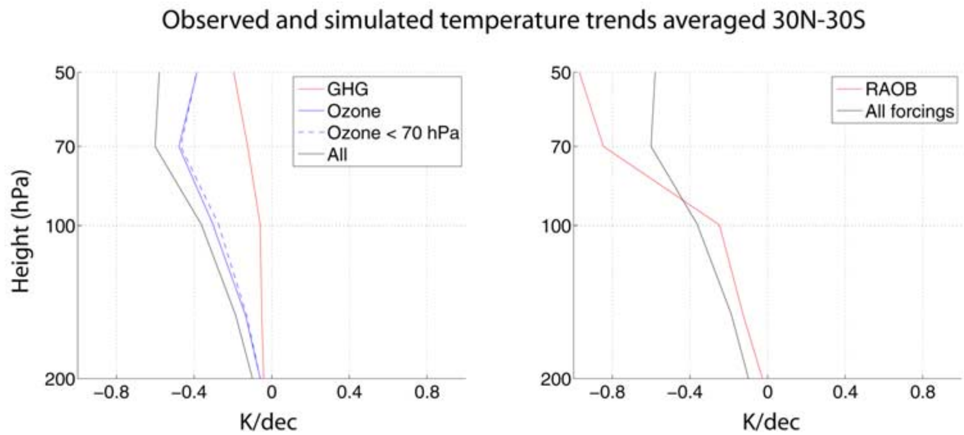

Figure 2. Vertical profiles of 1984–2005 linear temperature trends (K/decade) for the tropics (30S-30N). (left) For the same LLGHG and ozone combinations as Figure 1. (right) For model combined LLGHG and ozone changes (black line) and from the radiosonde analyses of *Thompson and Solomon* [2005] (red line).

stratosphere radiosonde observations is considered in the subsequent model comparisons.

3. Results

[9] Stratospheric temperature trends were computed from the RNBM model employing ozone trends from our SAGE analysis in Section 2 and LLGHG changes for the 1984–2005 period, and are illustrated in Figures 1b, 1c, 1d and 1e. Figures 1b and 1c show the annually averaged temperature changes from Figure 1b applying ozone changes only at lower pressures (higher altitudes) than 70 hPa, and Figure 1c, applying all ozone trends at pressures lower than 200 hPa. Differences are evident in the mid latitude Southern Hemisphere lower stratosphere, where the significant ozone depletion at pressures higher than 70 hPa drives cooling at these lower altitudes (see Figure 1c). At other latitudes, cooling occurs for pressures higher than 70 hPa irrespective of whether or not local ozone losses are imposed at those levels, as discussed below. Large coolings of over 0.5 K/decade are seen in the tropical and subtropical lower stratosphere (100–50 hPa) especially in the Southern Hemisphere. Inspecting the seasonal trends, this tropical cooling occurs year round. Another sensitivity test employed ozone trends from the latest *Randel and Wu* [2007] SAGE analysis, which has slightly less tropical ozone reduction. Employing these data, our model gave temperature trends in the lower stratosphere that had a very similar vertical and horizontal structure but were around 0.05 K/decade more positive compared to those employing our own analysis.

[10] LLGHG trends have a small effect near the tropopause and give a cooling that increases with altitude. Between 30–100 hPa (33–15 km) the combined effect of ozone and LLGHGs (Figure 1e) is of more-or-less uniform cooling with latitude. There is a slight tropical minimum in the cooling around 30 hPa, and a maximum cooling of slightly over 0.5 K/decade in the mid latitude Southern Hemisphere lower stratosphere. The generally accepted view [e.g., *Shine et al.*, 2003] has long been that ozone cooling is at a minimum in the tropics, having its largest effect at higher latitudes. However, the current results clearly point to a large ozone induced tropical lower stratospheric cooling over 1984–2005, agreeing with *Thompson and Solomon* [2005] who noted that the tropics were cooling at a similar or greater rate than higher latitudes. *Forster and Shine* [1997] noted a tropical lower stratospheric cooling over 1979–1995, employing an earlier version of combined SAGE I and SAGE II trends. However, due to uncertainties in merged satellite timeseries, these were not deemed trustworthy. Here confidence is strengthened both by a longer record than these previous studies, and by examination of the correspondence of the resulting modeled temperature trend structure to improved observations – see Discussion and Conclusions.

[11] Figure 2 shows vertical profiles of trends in the tropics. The blue dashed line on Figure 2 (left) shows the temperature changes calculated when ozone trends at pressures higher than 70 hPa are excluded. The resulting temperature change is similar to that applying ozone changes everywhere at pressures lower than 200 hPa, especially in the tropics. This demonstrates that the changes in temperature in this region are linked to ozone changes, but not at the local altitudes. The ozone trends at 70 hPa are subject to much smaller uncertainties than those at higher pressures, so that this finding makes the link of temperature changes in the key region from about 150 to 70 hPa to ozone changes more robust. Analyses of radiation fields shows that the cooling in the 150 hPa–100 hPa region is caused by decreases in down-welling longwave radiation from the ozone decrease and subsequent cooling of the 70–30 hPa layer. The cause of the cooling in this higher layer is split roughly equally between the effects of a decrease in shortwave absorption from reduced ozone at these levels (40%), and a decrease of the absorption of upwelling longwave radiation by the same reduced ozone (60%).

[12] At 70 hPa ozone and LLGHG changes cannot account for all the cooling seen in the observations, and with decreasing pressure the discrepancy gets worse (Figure 2, right). Since recent work does not show evidence of a long-term change in stratospheric water vapour over this time period [*Randel et al.*, 2006] it is unlikely that this cooling could have been caused by any water vapour increase. It is possible that ozone changes in the extratropics

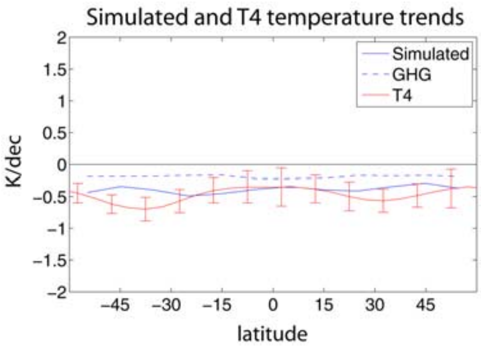

Figure 3. Shows the synthetic T4 weighted model trend for ozone and greenhouse gas changes (blue line), greenhouse gas changes only (blue dashed line), and the satellite derived trends as a function of latitude with 95% error bars (red line; for details of the T4 trends and error calculations shown here, see *Thompson and Solomon* [2005]).

or an increase in wave driving may induce changes in the meridional circulation and act to cool the tropical stratosphere, which could add to the radiative effects shown here [e.g., *Ramaswamy et al.*, 2001]. However, radiosonde errors have been shown to increase with altitude and the potential for errors in the sonde data at these higher altitudes of the tropics has recently been highlighted [e.g., *Sherwood et al.*, 2005; *Randel and Wu*, 2006; *Free et al.*, 2005]. Figure 3 suggests that the trends calculated here are in good agreement with the T4 satellite data (150–15 hPa) for the tropics.

[13] The results presented in Figures 2 and 3 suggest that it is likely the cooling trend in the tropical lower stratosphere is overestimated in the radiosonde record. *Randel and Wu* [2006] and *Sherwood et al.* [2005] found that a subset of the best tropical stations that had little evidence of bias gave coolings of around 0.5 K/decade at 70 hPa and lower pressures. Using such a subset would bring observations and model into closer agreement.

4. Discussion and Conclusions

[14] The two principle conclusions of the study are the following:

[15] (1) Radiative cooling due to ozone depletion in the tropical lower stratosphere above 70 hPa induces cooling which extends down to at least 150 hPa, well into the tropical upper troposphere. The simulated cooling of the upper tropical troposphere is due to a reduction in downwelling longwave radiation from the ozone depleted region above. The results suggest decreasing ozone in the tropical stratosphere may be responsible for observations of cooling in the tropical sub-stratosphere region.

[16] (2) The simulated response to the observed ozone trends drives temperature trends that are roughly comparable at all latitudes, consistent with the latitudinal near-homogeneity of recent stratospheric temperature trends revealed by *Thompson and Solomon* [2005].

[17] Analysis of the *Randel and Wu* [2007] trends that are time dependent show that this ozone trend and subsequent cooling is quite gradual over the time period – there are no sudden jumps and the magnitude is roughly the same size as the biennial variation in ozone and temperature caused by the Quasi-Biennial Oscillation. Ozone trends do not explain the sudden decrease in tropical tropopause temperatures during 2001 noted by *Randel et al.* [2006].

[18] A caveat of this work is that we do not account for the influence of convection or dynamics in the sub-stratosphere/upper troposphere. As noted in the introduction, climate model convection schemes are uncertain. The 150–100 hPa region is above the maximum level of convective outflow around 200 hPa and convective vertical mixing drops off rapidly in the sub-stratosphere [*Gettelman and Forster*, 2002; *Folkins et al.*, 1999]; we therefore argue that excluding changes in convective mixing is acceptable, especially as there is currently no adequate model of vertical mixing this region. However, changes in convective heating at lower altitudes would effect the temperature change at the bottom boundary which would have a small impact higher up. More significant would be any changes in the overturning circulation of the stratosphere. Butchart et al. found an average increase of 4% per decade in tropical upwelling mass flux from CO2 warming scenarios over the 20th century, and we used this estimate as a basis for sensitivity tests. This upwelling was incorporated into the model by modifying the background heating rates. This increase in upwelling gave a maximum cooling trend of 0.8 K/decade in the 100–70 hPa region. Adding this additional cooling would therefore significantly impact the comparison of our radiative response model with observations.

[19] The impact of ozone depletion in the stratosphere on temperature trends in the upper troposphere revealed in this study does not appear to be captured in most climate models. Climate models would be expected to capture the change in longwave emissions form ozone and the resulting temperature change, but they do not have enough ozone reduction in the tropical lower stratosphere for the mechanism to operate. In addition they may also overestimate the influence of convection in the upper troposphere. Our results therefore suggest that climate change simulations could over-estimate the warming of the upper troposphere. This, in turn, could lead to an overestimation of the water vapour feedback [see, e.g., *Cordero and Forster*, 2006]. Additionally, the ozone-induced cooling will likely affect the static stability of the region and the amount of water vapour entering the stratosphere, introducing a small negative radiative forcing. In closing, we note that this study has shown that (1) incorporating accurate ozone trends and (2) improving our understanding of the convective, dynamical, and radiative influences on the tropopause region is key to understanding the spatial patterns of temperature changes, which in turn are critical for climate change.

References

Bodeker, G. E., I. S. Boyd, and W. A. Matthews (1998), Trends and variability in vertical ozone and temperature profiles measured by ozonesondes at Lauder, New Zealand: 1986–1996, *J. Geophys. Res.*, *103*, 28,661–28,681.

Bodeker, G. E., J. C. Scott, K. Kreher, and R. L. McKenzie (2001), Global ozone trends in potential vorticity coordinates using TOMS and GOME intercompared against the Dobson network: 1978–1998, *J. Geophys. Res.*, *106*, 23,029–23,042.

Butchart, N., et al. (2006), Simulations of anthropogenic change in the strength of the Brewer–Dobson circulation, *Clim. Dyn.*, *27*, 727–741.

Climate Change Science Program (CCSP) (2006), Temperature trends in the lower atmosphere: Steps for understanding and reconciling differences, edited by T. R. Karl et al., report, Clim. Change Sci. Program and Subcomm. on Global Change Res., Washington, D. C.

Cordero, E. C., and P. M. de F. Forster (2006), Stratospheric variability and trends in models used for the IPCC AR4, *Atmos. Chem. Phys.*, *6*, 5369–5380.

Folkins, I., M. Lowewenstein, J. R. Podolske, S. J. Oltmans, and M. H. Proffitt (1999), A barrier to vertical mixing at 14 km in the tropics: Evidence from ozonesondes and aircraft measurements, *J. Geophys. Res.*, *104*, 22,095–22,102.

Forster, P. M., and M. Joshi (2005), The role of halocarbons in the climate change of the troposphere and stratosphere, *Clim. Change*, *71*, 249–266.

Forster, P. M., and K. P. Shine (1997), Radiative forcing and temperature trends from stratospheric ozone changes, *J. Geophys. Res.*, *102*, 10,841–10,855.

Forster, P. M. de F., and K. P. Shine (2002), Assessing the climate impact of trends in stratospheric water vapor, *Geophys. Res. Lett.*, *29*(6), 1086, doi:10.1029/2001GL013909.

Forster, P., et al. (2007), Changes in atmospheric constituents and in radiative forcing, in *Climate Change 2007: The Physical Science Basis. Contribution of Working Group I to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change*, edited by S. Solomon et al., pp. 129–234, Cambridge Univ. Press, New York.

Free, M., J. K. Angell, I. Durre, J. R. Lanzante, T. C. Peterson, and D. J. Seidel (2004), Using first differences to reduce inhomogeneity in radiosonde temperature datasets, *J. Clim.*, *17*, 4171–4179.

Free, M., D. J. Seidel, J. K. Angel, J. Lanzante, I. Durre, and T. C. Peterson (2005), Radiosonde Atmospheric Temperature Products for Assessing Climate (RATPAC): A new dataset of large-area anomaly time series, *J. Geophys. Res.*, *110*, D22101, doi:10.1029/2005JD006169.

Gettelman, A., and P. M. d. F. Forster (2002), Definition and climatology of the tropical tropopause layer, *J. Meteorol. Soc. Jpn.*, *80*, 911–924.

McCormick, M. P., J. M. Zawodny, R. E. Veiga, J. C. Larsen, and P. H. Wang (1989), An overview of SAGE I and SAGE II ozone measurements, *Planet. Space Sci.*, *37*(12), 1567–1586.

Mitas, C. M., and A. Clement (2006), Recent behavior of the Hadley cell and tropical thermodynamics in climate models and reanalyses, *Geophys. Res. Lett.*, *33*, L01810, doi:10.1029/2005GL024406.

Nikulin, G., and A. Karpechko (2005), The mean meridional circulation and midlatitude ozone buildup, *Atmos. Chem. Phys.*, *5*, 3159–3172.

Ramaswamy, V., et al. (2001), Stratospheric temperature trends: Observations and model simulations, *Rev. Geophys.*, *39*, 71–122.

Ramaswamy, V., M. D. Schwarzkopf, W. Randel, B. D. Santer, B. Soden, and G. Stenchikov (2006), Anthropogenic and natural influences in the evolution of lower stratospheric cooling, *Science*, *311*, 1138–1141.

Randel, W. J., and F. Wu (2006), Biases in stratospheric and tropospheric temperature trends derived from historical radiosonde data, *J. Clim.*, *19*, 2094–2104.

Randel, W. J., and F. Wu (2007), A stratospheric ozone profile data set for 1979–2005: Variability, trends, and comparisons with column ozone data, *J. Geophys. Res.*, *112*, D06313, doi:10.1029/2006JD007339.

Randel, W. J., R. S. Stolarski, D. M. Cunnold, J. A. Logan, M. J. Newchurch, and J. M. Zawodny (1999), Trends in the vertical distribution of ozone, *Science*, *285*, 1689–1692.

Randel, W. J., F. Wu, H. Vömel, G. E. Nedoluha, and P. Forster (2006), Decreases in stratospheric water vapor after 2001: Links to changes in the tropical tropopause and the Brewer-Dobson circulation, *J. Geophys. Res.*, *111*, D12312, doi:10.1029/2005JD006744.

Rind, D., J. Lerner, and J. Zawodny (2005), A complementary analysis for SAGE II data profiles, *Geophys. Res. Lett.*, *32*, L07812, doi:10.1029/2005GL022550.

Santer, B. D., et al. (2003), Behavior of tropopause height and atmospheric temperature in models, reanalyses, and observations: Decadal changes, *J. Geophys. Res.*, *108*(D1), 4002, doi:10.1029/2002JD002258.

Santer, B. D., et al. (2005), Amplification of surface temperature trends and variability in the tropical atmosphere, *Science*, *309*, 1551–1556, doi:10.1126/science.1114867.

Seidel, D. J., and W. J. Randel (2006), Variability and trends in the global tropopause estimated from radiosonde data, *J. Geophys. Res.*, *111*, D21101, doi:10.1029/2006JD007363.

Sherwood, S. C., J. R. Lanzante, and C. L. Meyer (2005), Radiosonde daytime biases and late 20th century warming, *Science*, *309*, 1556–1559.

Shine, K. P., et al. (2003), A comparison of model-simulated trends in stratospheric temperature, *Q. J. R. Meteorol. Soc.*, *129*, 1565–1588.

Thompson, D. W., and S. Solomon (2005), Recent stratospheric climate trends as evidenced in radiosonde data: Global structure and tropospheric linkages, *J. Clim.*, *18*, 4785–4795.

Thuburn, J., and G. C. Craig (2000), Stratospheric influence on tropopause height: The radiative constraint, *J. Atmos. Sci.*, *57*, 17–28.

Wang, H. J., D. M. Cunnold, and X. Bao (1996), A critical analysis of stratospheric aerosol and gas experiment ozone trends, *J. Geophys. Res.*, *101*(D7), 12,495–12,514.

Wang, H. J., D. M. Cunnold, L. W. Thomason, J. M. Zawodny, and G. E. Bodeker (2002), Assessment of SAGE version 6.1 ozone data quality, *J. Geophys. Res.*, *107*(D23), 4691, doi:10.1029/2002JD002418.

G. Bodeker, National Institute of Water and Atmospheric Research, Private Bag 50061, Omakau, Lauder, State Highway 85, Central Otago 9182, New Zealand.

P. M. Forster, School of Earth and Environment, University of Leeds, Leeds LS2 9JT, UK. (piers@env.leeds.ac.uk)

R. Schofield, Alfred Wegener Institute for Polar and Marine Research, Telegrafenberg A43, D-14473 Potsdam, Germany.

S. Solomon, Chemical Sciences Division, NOAA Earth System Research Laboratory, 325 Broadway, Boulder, CO 80305, USA.

D. Thompson, Department of Atmospheric Science, Colorado State University, 200 West Lake Street, Fort Collins, CO 80523, USA.