

Assessing the climate impact of trends in stratospheric water vapor

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[1] It is now apparent that observed increases in stratospheric water vapor may have contributed significantly to both stratospheric cooling and tropospheric warming over the last few decades. However, a recent study has suggested that our initial estimate of the climate impact may have *overestimated* both the radiative forcing and stratospheric cooling from these changes. We show that differences between the various estimates are *not* due to inherent problems with broadband and narrow-band radiation schemes but rather due to the different experimental setups, particularly the altitude of the water vapor change relative to the tropopause used in the radiative calculations. Furthermore, we show that if recent estimates for the observed water vapor trends are valid globally they could have contributed a radiative forcing of up to 0.29 Wm^{-2} and a lower-stratospheric cooling of more than 0.8 K over the past 20 years, with these values more than doubling if, as has been suggested, the trend has persisted for the last 40 years. This 40 year radiative forcing is roughly 75% of that due to carbon dioxide alone but, despite its high value, we find that the addition of this forcing into a simple model of climate change still gives global mean surface temperature trends which are consistent with observations. **INDEX TERMS:** 1610 Global Change (New Category): Atmosphere (0315, 0325); 1836 Hydrology: Hydrologic budget (1655); 3362 Meteorology and Atmospheric Dynamics: Stratosphere/troposphere interactions; 3359 Meteorology and Atmospheric Dynamics: Radiative processes

1. Introduction

[2] Initial interest in the climatic role of changes in stratospheric water vapor (SWV) [e.g. *Manabe and Wetherald*, 1967, *Rind et al.*, 1996] stemmed from the possibility that a fleet of supersonic aircraft could significantly perturb concentrations. This was followed by concern that an indirect impact of increased concentrations of methane would, via its oxidation in the stratosphere, lead to an increase in SWV; estimates of the size of this impact were divergent (see *IPCC* [1995], page 173). Balloon-based and satellite observations of increases in SWV has led to renewed interest in the climatic implications [*Forster and Shine*, 1999; *Smith et al.*, 2001; *Dvortsov and Solomon*, 2001; *Shindell*, 2001; *Oinas et al.*, 2001]. This letter re-examines these climatic implications in light of new trend estimates [e.g. *Oltmans et al.*, 2001, *Rosenlof et al.*, 2001]. It is structured as follows. First we provide a validation of the radiative transfer scheme used here. Second, we extend our earlier calculations using a better background water vapor climatology and an updated trend analysis. Third we consider the impact of these changes on estimates of stratospheric temperature changes. Finally, we note the consequences of our results on observations of both surface and stratospheric temperature change in recent decades.

[3] There are several areas that we can only touch on. We recognise that the evidence for sustained (i.e. decadal) global trends in SWV remains controversial, as do the possible causes. Second,

we treat SWV changes here as a radiative forcing, but recognise, as we did in *Forster and Shine* [1999] (FS99), that they could be, at least partly, a feedback.

2. Testing the Narrowband Radiation Code

[4] *Oinas et al.* [2001] (OLRSH) have suggested that broadband radiative schemes, including the scheme employed in our FS99 calculations, overestimate the radiative forcing and stratospheric cooling due to SWV increases. In the light of their comments it is important to test the codes we use here. We use our standard Malkmus narrow band model (NBM) at 10 cm^{-1} spectral resolution [*Forster and Shine*, 1997] with updated spectral line and continuum representations. We compare it against two other models. First, we use a line-by-line (lbl) code [*Dudhia*, 1997], which has been compared against both observations and other line-by-line codes [*Sihra et al.*, 2001]. Second, A. A. Lacis (pers. comm.) kindly provided results from the 500-k model used by OLRSH. Using the OLRSH global-mean profile and perturbing SWV from 6 to 6.7 ppmv at pressures below 150 hPa, yields an instantaneous longwave forcing at 190 hPa of 0.19 Wm^{-2} in the OLRSH model, 0.20 Wm^{-2} in our lbl calculations and 0.21 Wm^{-2} from our NBM. At 150 hPa, the values are, respectively, 0.26, 0.28 and 0.29 Wm^{-2} . These results show that the NBM works well compared against these other models. (We note also that *Forster et al.* [2001] showed general agreement in the SWV forcing using a range of different radiation codes; in general they found that the broader band models *underestimated* the radiative forcing (and stratospheric cooling) contrary to the general conclusion of OLRSH. The instantaneous change in radiative heating rate in the stratosphere using our NBM, which is important in determining the temperature change, was shown to agree with the lbl results to within 10% at all levels, and normally to within 5%.

3. Radiative Forcing

[5] FS99 estimated a global-mean adjusted radiative forcing due to an increase in SWV from 6 to 6.7 ppmv of 0.19 Wm^{-2} , whereas OLRSH estimated a lower (but still climatically significant) value of 0.12 Wm^{-2} . There are several aspects of the OLRSH calculation. First, their forcing calculation is based on a calculation using a global-mean profile. Second, they perturb their water vapor at pressures below 150 hPa, while their tropopause is at 190 hPa. Third, their global-mean profile is based on a radiative-convective model, rather than observations. We use our own global-mean profile, based largely on European Centre for Medium Range Weather Forecasts analyses, which has a 2 Kkm^{-1} tropopause at around 120 hPa, and perturb from 6 to 6.7 ppmv at all pressures lower than this. We obtain an adjusted forcing (including the effect of changes in shortwave radiation) of 0.15 Wm^{-2} . Further experiments indicated that moving the altitude of the water vapor perturbation by 20 hPa away from the tropopause decreased the forcing by more than 10%; hence much of the difference between our global-mean result and that of OLRSH could be due to different experimental designs whereby their perturbation of water vapor starts well above the tropopause. (Note that the instantaneous forcing has around double the sensitivity to changing the height

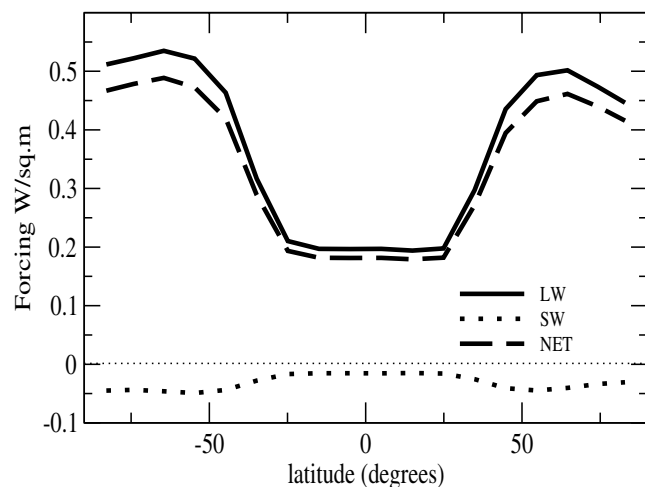


Figure 1. The annually averaged adjusted longwave (LW), shortwave (SW) and net (NET) radiative forcing calculated by the NBM, as a function of latitude, for SWV changes between 1980–2000 (1 ppmv SWV was added to 1980 background values). The thin dotted line indicates the zero forcing level.

at which the perturbation starts in the lower stratosphere; however, the impact of such changes on the adjusted forcing is reduced because the cooling of the lower stratosphere opposes the increased greenhouse effect of the extra water vapor.)

[6] Next, we include the latitudinal variation of the forcing, now perturbing the water vapor above the local tropopause. This significantly increases the forcing due to the decrease in tropopause height at mid to high latitudes. Using a 10° latitudinal resolution, and calculating the forcing for each of the mid-season months, we derive a forcing of 0.17 Wm^{-2} , closer to our original estimate than the OLRSH result. Hence, we believe that the OLRSH estimate is much lower than ours because it was based on a global-mean profile and the fact that their perturbation of SWV did not coincide with their tropopause.

[7] The next aspect we explore is the impact of using our highly idealized perturbation of SWV (6 to 6.7 ppmv), noting that *Smith et al.* [2001], using more refined trends and background climatology, derived a forcing consistent with our initial estimate of 0.19 Wm^{-2} .

[8] The degree of approximation employed for our applied perturbation in FS99 was felt justified by the paucity of long-term SWV measurements. This crude early estimate has since been refined, for 1992–1999, using trends derived from the HALogen Occultation Experiment (HALOE) [*Smith et al.*, 2001]. Recently *SPARC* [2000] and *Rosenlof et al.* [2001] have revisited past observations of SWV. Although still extremely uncertain, these works show 10 different datasets which give a positive trend in SWV (between 200–10 hPa) of up to 0.1 ppmv/year with a mean trend of around 0.05 ppmv/year. There is even some evidence, from a few aircraft flights in the 1950s and 1970s, that perhaps SWV has been increasing for the past 40–50 years (see *SPARC* [2000] and *Rosenlof et al.* [2001] and references therein for details). Trends calculated near the tropopause are very uncertain, but are possibly larger. This is the region where the previous section found the climate to be most sensitive. Further, the non-satellite observations tend to confine themselves to the Northern extratropics. However, if these trends are representative of the entire stratosphere, the original perturbation, of 0.7 ppmv over a 17 year period in FS99, would underestimate the SWV change. In addition to this there is a need to choose an appropriate pre-SWV-change background level. For example, we find that perturbing SWV by 0.7 ppmv from a more realistic present day SWV amount increases the

effects compared to a perturbation from 6 ppmv: for a global-mean profile the radiative forcing increases by 14% and the lower stratospheric cooling by about 20% (also see discussion in OLRSH). In FS99 the perturbation was added to a background value of 6 ppmv. In retrospect this background is too high as present day values range between 4–6 ppmv, with the smallest values between 100–50 hPa [*SPARC*, 2000], and the few observations of water vapor from the 1950s show values as low as 1–3 ppmv at 100 hPa.

[9] Using a climatology of SWV, based on HALOE data for 1992–2000 (Rosenlof, pers. comm.) to represent SWV values of 1996, we reduced stratospheric values by 0.05 ppmv/year to obtain background states representative of 1960 and 1980. We then calculated the adjusted radiative forcing (Figure 1, shown for 1980–2000) by adding 2 ppmv and 1 ppmv onto these backgrounds respectively to simulate changes until the year 2000.

[10] We choose a constant 0.05 ppmv/year increase in SWV (as did *Dvortsov and Solomon* [2001]) as representative of the water vapor trend at pressures of 100–10 hPa [*SPARC*, 2000]. Outside of this region the same trend is applied although the observed trend may be more than twice this value [*SPARC*, 2000]. We do this because water vapor changes at pressures below 10 hPa have only a small effect on the radiative forcing. At pressures above 100 hPa the higher trend may be the result of “seeing” tropospheric water vapor; therefore we did not feel confident in including it as part of the stratospheric change.

[11] Our results show a globally and seasonally averaged net adjusted radiative forcing of 0.29 Wm^{-2} for the 1980–2000 scenario and 0.63 Wm^{-2} for the 1960–2000 scenario. Figure 1 illustrates that the high latitude forcing is double the tropical forcing. This is simply due to a lower tropopause and therefore a greater column water vapor change at high latitudes. It also shows that the extra absorption of solar radiation moderates the longwave forcing by roughly 10%. These radiative forcings are much larger than all the already-quoted previously published estimates. These are values for a global trend in SWV and as such perhaps represent an upper bound to the radiative forcing. If the positive trend has been limited to Northern Hemisphere latitudes above 30° N the radiative forcings are reduced to 0.10 Wm^{-2} and 0.21 Wm^{-2} for the 1980–2000 and 1960–2000 water vapor changes respectively. However, even these magnitudes are comparable to and of the opposite sign to the radiative forcing due to stratospheric ozone depletion [e.g. *Forster and Shine*, 1997].

4. Stratospheric Temperature Response

[12] Since FS99 we have found an error in the implementation of the *Christidis* [1999] broadband scheme in the General Circulation Model (GCM) used in that study, which is particularly evident in the calculation of stratospheric temperatures due to water vapor changes. This leads to an overestimate of the stratospheric temperature response to SWV changes. However, fixed dynamical heating (FDH) calculations indicate that there is not an inherent problem with broadband schemes in calculating the stratospheric temperature response. Replacing the GCM radiation scheme with that of *Zhong et al.* [1996] and then repeating the SWV experiment in FS99 we obtain a cooling in the lower-stratosphere of 0.3–0.4 K (not shown), bringing our results closer to those of OLRSH.

[13] In this letter we use the FDH approximation [e.g. *Forster and Shine*, 1997] to model the stratospheric temperature response. This allows us to employ a more sophisticated radiation scheme and perform many more experiments than would be possible with a GCM. The FDH approximation has been shown to quantitatively agree reasonably well with results from models which include the full dynamical response of the

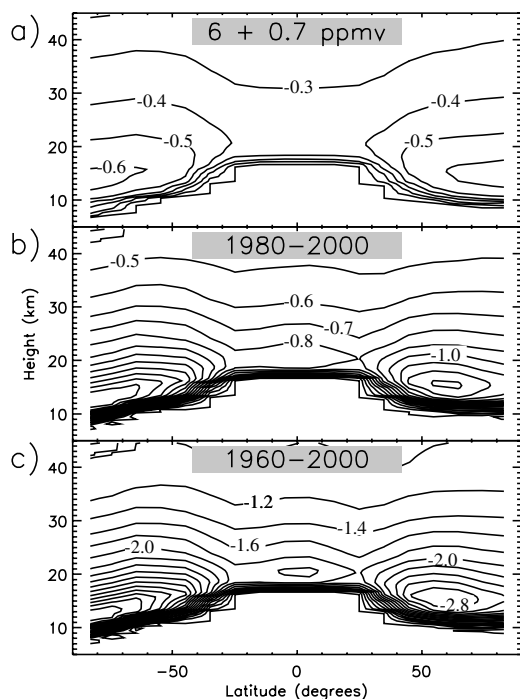


Figure 2. The annually averaged FDH stratospheric temperature change (in K) calculated by the NBM as a function of latitude for: (a) a 0.7 ppmv increase in SWV, from a 6 ppmv background value; (b) a 1 ppmv increase from a 1980 background, to simulate the 1980–2000 change in SWV and (c) a 2 ppmv increase from a 1960 background, to simulate the 1960–2000 change in SWV. (a) and (b) have a contour interval of 0.1 K and (c) a contour interval of 0.2 K.

stratosphere [e.g. *Ramaswamy et al.*, 2001]. The stratospheric temperature changes for FDH calculations with the narrowband scheme are shown in Figure 2, using the climatology described in the previous section. Figure 2a shows the response of a 6 to 6.7 ppmv change in SWV. The temperature response in the lower stratosphere is in reasonable agreement with the GCM results of OLRSH. Adopting the more refined water vapor perturbations discussed in the previous section, the 1980–2000 scenario FDH temperature response (Figure 2b) shows a cooling of 0.8 K in the tropical lower stratosphere with up to a 1.4 K cooling at high latitudes. This cooling can be seen to more than double for the 1960–2000 scenario (Figure 2c) to give coolings of 2.0 K in the tropical lower stratosphere and 4.0 K at high latitudes. These cooling are consistent with those found by *Smith et al.* [2001], if we allow for the use of different background water vapor profiles and a 20–30% underestimate of lower stratospheric cooling with their results, due to limitations of their particular broadband scheme [*Forster et al.*, 2001]. An equivalent 0.05 ppmv/year increase in SWV modeled by *Dvortsov and Solomon* [2001] led to a very similar temperature response at 43°N.

5. Climate Consequences

[14] Our upper bound for the radiative forcing due to SWV is 0.63 Wm^{-2} between 1960 and 2000. To crudely investigate the impact of such a forcing, a simple energy-balance model coupled to a deep ocean via diffusion is used. We employ the time series of all the natural and anthropogenic forcings given by *Myhre et al.* [2001], together with a climate sensitivity parameter of $0.67 \text{ K (Wm}^{-2}\text{)}^{-1}$ (approximately equivalent to a 2.5 K equilibrium surface warming for doubling carbon dioxide). The time-dependent forcing estimate includes forcings due

to well mixed greenhouse gases, ozone, direct and indirect aerosol, solar and volcanism, but not SWV. It is similar to that in *IPCC* [2001] (Fig. 6.8) but, in addition, includes a mid-range estimate for the indirect aerosol forcing. The results are compared to the combined global-mean land and sea-surface temperature anomaly dataset [e.g. *Parker and Alexander*, 2001]. Figure 3 shows the temperature anomaly both with and without the SWV forcing, with the water vapor forcing assumed to increase linearly from zero to 0.63 Wm^{-2} between 1960 and 2000. It is recognized that both the forcing series and the climate sensitivity are significantly uncertain [*IPCC*, 2001] and so strong positive statements about the agreement with observations cannot be made, at least using global-mean surface temperature alone. Nevertheless, the important message from Figure 3 is that a significantly large SWV forcing certainly cannot be ruled out on the basis of present understanding. For this calculation the SWV change is included as a forcing. If the SWV change was, at least in part, due to a feedback, then it would remain significant. If as an upper limit all of the 0.6 Wm^{-2} change in 40 years is assumed to result from approximately 0.4 K warming of the surface (see Figure 3), this would imply a contribution to the inverse climate sensitivity parameter of $1.5 \text{ Wm}^{-2}\text{K}^{-1}$, which is similar in size to the tropospheric water vapor feedback [see, e.g. *IPCC*, 1990].

[15] The size of the stratospheric cooling, indicated by Figure 2, is less easy to reconcile with observations; however, current understanding of the causes of stratospheric temperature trends is not sufficiently complete to rule out a significant SWV contribution. For example, *Ramaswamy et al.* [2001, especially Plate 2] present a comparison of 50 hPa temperature trends from a number of analyses of radiosonde data. Although there is a large spread in these results, coolings are reported of up to, and occasionally exceeding, 0.5 Kdecade^{-1} in low-to-mid latitudes over the period 1966 to 1994. If this is representative of the 1960–2000 trends, then the 50 hPa level could have cooled by up to 2 K in this period. This is larger than could be accounted for by other mechanisms. FDH calculations with the 1960–2000 increases in the well-mixed greenhouse gases indicate a cooling of 0.55 K at 50 hPa. Two recent GCM studies using imposed

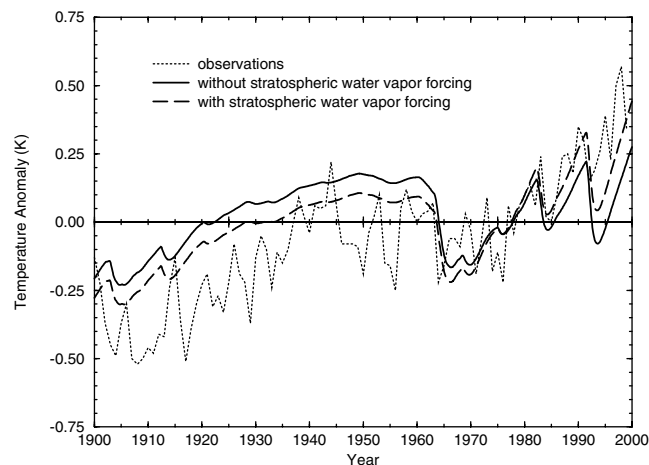


Figure 3. Evolution of global-mean surface temperature, using a simple global-mean energy balance model (with a climate sensitivity of $0.67 \text{ K (Wm}^{-2}\text{)}^{-1}$), expressed as an anomaly from the 1961–1990 mean. The solid line shows the temperature anomaly using all natural and anthropogenic radiative forcings from *Myhre et al.* [2001]; the dashed-dot line shows the impact of adding a stratospheric water vapor forcing increasing linearly from zero in 1960 to 0.63 Wm^{-2} in 2000. The dotted line shows the observations.

height-resolved stratospheric ozone trends [Langematz, 2000; Rosier and Shine, 2000] find a cooling of around 0.5 K at 50 hPa between about 1980–2000, and tropospheric ozone increases may have led to a lower stratospheric cooling of around 0.2 K [Ramaswamy *et al.*, 2001]. Assuming insignificant stratospheric ozone losses prior to 1980, the sum of these coolings is then about 1.25 K, which could leave room for a significant contribution from SWV, albeit smaller than that indicated in Figure 2c. It is recognized that there are considerable uncertainties in both the analysis of radiosonde data, and in particular, the estimates of the height-resolved stratospheric ozone and water vapor changes in the lower stratosphere.

[16] In addition, both Langematz [2000] and Rosier and Shine [2000] find that the observed temperature trends in the lower stratosphere in the Arctic springtime are several Kdecade⁻¹ more negative than those simulated with the observed (1977 to mid-1990s) ozone trends. There is a question as to the significance of the observed trends, and as to whether the trends over this period are representative of longer time periods [Langematz, 2000], but this extra cooling may also be a “fingerprint” of a contribution from water vapor.

6. Conclusions

[17] Prompted by recent SWV trend estimates and OLRSH, we re-examined our earlier estimates of the radiative forcing and stratospheric temperature change due to SWV trends. We partly agree with the OLRSH conclusion that our values of stratospheric cooling were, in the context of the simple experiment we performed, an overestimate. However, experiments with a more refined background climatology of water vapor and its trends, yielded temperature changes not dissimilar to our original estimates. If, as has been speculated, the trends have been sustained since 1960 and are global in nature, water vapor changes would have been the dominant cause of lower stratospheric temperature trends. In some regions there are indications that the observed trends could be significantly in excess of those that can be explained from greenhouse gases and ozone changes alone.

[18] We do not agree with OLRSH that our original radiative forcing estimate was a significant underestimate, and believe that much of the difference may be due to different experimental design. Using the updated water vapor climatology and trends, and assuming that the SWV change is a forcing rather than a feedback, we derive an upper limit of the forcing as 0.63 Wm⁻² since 1960, or about 40% of the forcing due to the well-mixed greenhouse gases, or about 75% of the forcing due to CO₂ alone. Although large, such a forcing certainly cannot be ruled out, given the uncertainties in the size of other forcings and in the climate sensitivity.

[19] The prime conclusion of this work is that SWV remains *potentially* a very significant contributor to recent climate change. It is therefore important that efforts continue to quantify the size and reliability of SWV trends and their possible causes.

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References

Christidis, N., Halocarbon radiative forcing in radiation and general circulation models *Ph.D Thesis*, University of Reading, U.K., 1999.

- Dudhia, A., Reference Forward Model v3 Software User’s Manual, Department of Atmospheric, Oceanic and Planetary Physics, Univ. of Oxford, United Kingdom, Oxford, 1997.
- Dvortsov, V. L., and S. Solomon, Response of the stratospheric temperatures and ozone to past and future increases in stratospheric humidity, *J. Geophys. Res.*, *106*, 7505–7514, 2001.
- Forster, P. M. de F., and K. P. Shine, Radiative forcing and temperature trends from stratospheric ozone changes, *J. Geophys. Res.*, *102*, 10,841–10,855, 1997.
- Forster, P. M. de F., and K. P. Shine, Stratospheric water vapor changes as a possible contributor to observed stratospheric cooling, *Geophys. Res. Lett.*, *26*, 3309–3312, 1999.
- Forster, P. M. de F., M. Ponater, and W. Y. Zhong, Testing broadband radiation schemes for their ability to calculate the radiative forcing and temperature response to stratospheric water vapour and ozone changes., *Meteorol. Z.*, *10*, 387–393, 2001.
- IPCC, Climate Change 1990. Intergovernmental Panel on Climate Change, Cambridge University Press, Cambridge, UK, 1990.
- IPCC, Climate Change 1994. Intergovernmental Panel on Climate Change, Cambridge University Press, Cambridge, UK, 1995.
- IPCC, Climate Change 2001: The Scientific Basis. Intergovernmental Panel on Climate Change, Cambridge University Press, Cambridge UK, 2001.
- Langematz, U., An estimate of the impact of observed ozone losses on stratospheric temperature, *Geophys. Res. Lett.*, *27*, 2077–2080, 2000.
- Manabe, S., and R. T. Wetherald, Thermal equilibrium of the atmosphere with a given distribution of relative humidity, *J. Atmos. Sci.*, *24*, 241–259, 1967.
- Myhre, G., A. Myhre, and F. Stordal, Historical evolution of radiative forcing of climate, *Atmos. Environ.*, *35*, 2361–2373, 2001.
- Oinas, V., A. A. Lacis, D. Rind, D. T. Shindell, and J. E. Hansen, Radiative cooling by stratospheric water vapor: big differences in GCM results, *Geophys. Res. Lett.*, *28*, 2791–2794, 2001.
- Oltmans, S. J., H. Vomel, D. J. Hofmann, K. H. Rosenlof, and D. Kley, The increase in stratospheric water vapor from balloonborne, frostpoint hygrometer measurements at Washington, DC, and Boulder, Colorado, *Geophys. Res. Lett.*, *27*, 3453–3456, 2000.
- Parker, D. E., and L. V. Alexander, Global and regional climate in 2000, *Weather*, *56*, 255–267, 2001.
- Ramaswamy, V., M. L. Chanin, J. Angell, J. Barnett, D. Gaffen, M. Gelman, P. Keckhut, Y. Koshelkov, K. Labitzke, J. J. R. Lin, A. O’Neill, J. Nash, W. Randel, R. Rood, K. Shine, M. Shiotani, and R. Swinbank, Stratospheric temperature trends: Observations and model simulations, *Rev. Geophys.*, *39*, 79–122, 2001.
- Rind, D., P. Lonergan, and K. Shah, Climatic effect of water vapor release in the upper troposphere, *J. Geophys. Res.*, *101*, 29,395–29,405, 1996.
- Rosenlof, K. H., S. J. Oltmans, D. Kley, J. M. Russell, E. W. Chiou, W. P. Chu, D. G. Johnson, K. K. Kelly, H. A. Michelsen, G. E. Nedoluha, E. E. Remsberg, G. C. Toon, and M. P. McCormick, Stratospheric water vapor increases over the past half-century, *Geophys. Res. Lett.*, *28*, 1195–1198, 2001.
- Rosier, S. M., and K. P. Shine, The effect of two decades of ozone change on stratospheric temperature, as indicated by a general circulation model, *Geophys. Res. Lett.*, *27*, 2617–2620, 2000.
- Shindell, D. T., Climate and ozone response to increased stratospheric water vapor, *Geophys. Res. Lett.*, *28*, 1551–1554, 2001.
- Sihra, K., M. D. Hurley, K. P. Shine, and T. J. Wallington, Updated radiative forcing estimates of sixty-five halocarbons and non-methane hydrocarbons, *J. Geophys. Res.*, *106*, 20,493–20,506, 2001.
- Smith, C. A., J. D. Haigh, and R. Toumi, Radiative forcing due to trends in stratospheric water vapour, *Geophys. Res. Lett.*, *28*, 179–182, 2001.
- SPARC, SPARC Assessment of Upper Tropospheric and Stratospheric Water Vapour, pp. 312, Stratospheric Processes and their Role in Climate, WCRP-113, WMO-TD No 1043, SPARC Report 2, Verrieres Le Buisson Cedex, 2000.
- Zhong, W. Y., R. Toumi, and J. D. Haigh, Climate forcing by stratospheric ozone depletion calculated from observed temperature trends, *Geophys. Res. Lett.*, *23*, 3183–3186, 1996.

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