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Stratospheric water vapor trends in a coupled chemistry-climate model

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[1]   Stratospheric H_2O trends are examined in a detailed, coupled chemistry-climate model (CCM). The modeled H_2O trend in the upper stratosphere (US) is mainly caused by CH_4 oxidation while the trend in the lower stratosphere (LS) is largely related to changes in temperature and transport near the tropopause. Incomplete CH_4 oxidation leads to a maximum upward H_2O trend in the US of about $1.4\times$ the imposed tropospheric CH_4 trend. Cross-tropopause exchange of water vapor gives rise to much larger trends ($\sim$+50 ppbv/yr) in the LS. A trend of +0.44 K/decade in the 100 hPa temperature (T) contributes up to 70% (+35 ppbv/yr) of the model LS H_2O trend while the remainder (around 30%) can be caused by changes in transport processes near the tropopause. In the LS the maximum modeled trend is close to observations although the globally averaged value is smaller than observed. Given an observed decrease in tropopause temperatures, our results suggest that the observed LS water vapor increases would require a significant change in transport. **Citation:** Tian, W., and M. P. Chipperfield (2006), Stratospheric water vapor trends in a coupled chemistry-climate model, *Geophys. Res. Lett.*, *33*, L06819, doi:10.1029/2005GL024675.

1. Introduction

[2]   Although unusually low water vapor anomalies have been observed in the LS for 2001–2003 [*Randel et al.*, 2004], several studies have shown a general increase of water vapour in the LS in the past few decades [e.g., *Nedoluha et al.*, 1998; *Oltmans et al.*, 2000; *Rosenlof et al.*, 2001]. Due to data limitations the processes responsible for the increase are not easily diagnosed. Consequently, the reason(s) for the observed increase over the past few decades are still widely debated [e.g., *Dvortsov and Solomon*, 2001] and the future trend of stratospheric H_2O is uncertain. Various processes have been proposed to explain the long-term increases including warming of tropical tropopause, changing strength of the Brewer-Dobson (BD) circulation, and enhanced CH_4 oxidation [*Kirk-Davidoff et al.*, 1999; *Smith et al.*, 2000; *Röckmann et al.*, 2004; *Randel et al.*, 2004; *Fueglistaler et al.*, 2005]. However, the stratospheric H_2O increases are not compatible with the current understanding of changes in tropical tropopause T which have decreased from 1980–2000 [*Randel et al.*, 2000; *Seidel et al.*, 2001; *Zhou et al.*, 2001]. Although CH_4 oxidation is the principal source of stratospheric water vapor, the reported trend in H_2O over the past few decades of about 50 ppbv/yr [e.g., *Rosenlof et al.*, 2001] is about twice that expected based on the CH_4 trend.

[3]   In recent years, some CCMs have simulated the long-term changes of stratospheric trace gases [*Stenke and Grewe*, 2005; *Dameris et al.*, 2005; *Shindell*, 2001]. While those CCM simulations have shed additional light on the long-term variation of stratospheric water vapor, the underlying mechanisms have still not been clarified. *Shindell* [2001] used a low resolution ($8° \times 10°$) CCM with para-meterised chemistry to investigate relative contributions of different processes to the H_2O trend. He found an increase in water vapour due to both methane oxidation and increased input from the troposphere, due to warming of the model tropopause, and argued that climate change must be altering the stratospheric input.

[4]   In this paper, we investigate further the long-term stratospheric H_2O trend using a higher resolution, coupled, full chemistry CCM. The relative significance of CH_4 oxidation and cross-tropopause exchange of water vapor for the trend is estimated from output of two 40-year CCM transient runs. Although our CCM, in common with others, may not capture accurately the trends in all relevant parameters, we can diagnose them to quantify the relative contributions of processes which can help us understand the impact of these parameters on observed H_2O trends. We focus on the stratospheric water trend in the next few decades under IPCC scenario A2 of greenhouse gases (GHGs).

2. Model and Experiments

[5]   Our CCM is based on the UK Met Office Unified Model (UM) with a well-tested detailed chemistry scheme from the SLIMCAT model [*Chipperfield*, 1999]. Details of the CCM are given by *Tian and Chipperfield* [2005]. The model was run with a horizontal resolution of $2.5° \times 3.75°$ and 64 levels from the surface to 0.01 hPa. Two 40-year transient runs were performed; one with coupled chemistry and one without. In the coupled run the chemical fields of N_2O, CH_4, H_2O are fed back to the UM's radiation scheme, and GHG values of CO_2, N_2O, CH_4, CFC-11, and CFC-12 from IPCC scenario A2 [see *World Meteorological Organisation* (WMO), 2003] are used in the radiation scheme wherever the corresponding values from the chemistry module are not available. The model was run from 1979 to 2020. Sea-surface temperatures are taken from AMIPII [*Gates et al.*, 1999] until 1996 and then from previous UM climate runs. To avoid the effect of the model spin-up on the trend estimation the results presented in section 3 are based on the model output from 1990 to 2020.

[6]   There are two H_2O fields in the CCM. The strato-spheric chemistry water vapor in the stratosphere (hereafter $CCM(H_2O)$) has no interaction with the specific humidity field of the UM (hereafter UM(q)). The $CCM(H_2O)$

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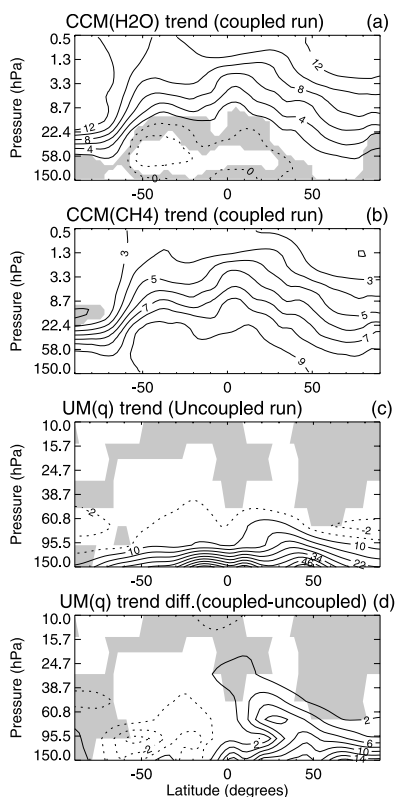
Figure 1. Height-latitude cross sections of linear trends (ppbv/yr) in (a) $CCM(H_2O)$ and (b) CH_4 estimated from the coupled model run for the period from 1990–2020. (c) The trends in $UM(q)$ from the uncoupled run and (d) the differences in $UM(q)$ trends between the coupled and uncoupled runs are also shown. Regions where the derived trend is not significant at the 2σ level are shaded. The contour interval is ± 2 ppbv/yr for water vapor and ± 1 ppbv/yr for CH_4. Note different y-axis ranges.

assumes a fixed H_2O mixing ratio entering from the troposphere and, therefore, any long-term trends in the $CCM(H_2O)$ result solely from changes in tropospheric methane which is based on observations until 2000 and then, in the A2 scenario, increases at a rate of 10 ppbv/yr from 2000 to 2020. It should be noted that this rate exceeds recent observations. In the chemistry scheme CH_4 oxidation is assumed to produce 2 H_2O molecules, i.e., H_2 production is ignored along with H_2O loss in the mesosphere. $UM(q)$ does not include any chemical source (e.g., from CH_4 oxidation) and is used as a proxy to study the trend caused by the cross-tropopause transport of water vapor from the troposphere.

[7] Previous studies have shown that the UM humidity field ($UM(q)$) is able to reproduce many aspects of the water vapor structure and evolution in the upper troposphere and lower stratosphere (UTLS) [e.g., *Pope et al.*, 2001]. The modeled climatologies of $CCM(H_2O)$ and CH_4, which were calculated from the output for 1980–2000 for comparison with observations, show a good agreement with the observations both in distribution and magnitude though some differences exist (see auxiliary material[1]). In the US the modeled water vapor is slightly lower than observed while

the modeled methane is a bit larger, suggesting that the vertical transport in the model is slightly too strong in the tropics.

3. Long-Term Change in Stratospheric H_2O

[8] Figure 1 shows the linear trends in $CCM(H_2O)$ and CH_4 from the coupled CCM run. Also shown are the linear trends in $UM(q)$ from the uncoupled run and the differences in the $UM(q)$ trends between the coupled and uncoupled runs (trends which are not significant at the 2σ level are shaded). The modeled stratospheric water vapor trends are not uniform. The observed trends in water vapor also have a large variation with height [e.g., *Evans et al.*, 1998; *Randel et al.*, 2004]. The largest trend in $CCM(H_2O)$ is seen in the high latitude US with a maximum increase of around 12 ppbv/yr. An interesting feature is the downward trend of about 2 ppbv/yr in $CCM(H_2O)$ in the tropical LS.

[9] The latitude-altitude variations of modeled CH_4 trends are consistent with observations up to the late 1990s [e.g., *Evans et al.*, 1998] with maximum and horizontal uniform values in the LS and minimum values in the US. The overall trend in the troposphere of 8–10 ppbv/yr includes the period of observations before 2000 when the growth rate was below 10 ppbv/yr. Figure 1 indicates that the $CCM(H_2O)$ trends in the LS do not reflect the rate of the methane change in the UTLS, while the $CCM(H_2O)$ trends in the US are around 70% of $2\times$ the CH_4 trend specified at the model's lower boundary as CH_4 is not fully oxidised (i.e., non-zero trend). Figure 1 also implies that methane oxidation contributes only a small fraction to the water vapor trends in the LS, and thus that observed water vapor trends in the LS may be caused mainly by changes in the injection of tropospheric water vapor and other processes.

[10] The vertical and horizontal structure of the trends in $UM(q)$ also display a marked variation with height and latitude. Consistent with the observations of *Evans et al.* [1998], the maximum trend in the coupled run occurs in LS northern midlatitudes with an increase reaching 50 ppbv/yr at around 150 hPa, 48°N. The magnitude of this maximum trend is actually close to the observed values although the globally averaged trend in the LS is smaller than observed. The trends in $UM(q)$ have an evident seasonal variation with a maximum trend in the northern hemisphere summer which reaches about 150 ppbv/yr and 120 ppbv/yr in the coupled and uncoupled run, respectively (see auxiliary material). *Smith et al.* [2000] also found a significant seasonal variation from HALOE data. However, their results indicate a maximum trend in autumn of more than 250 ppbv/yr. Also noticeable is that coupling of the chemistry leads to an increase in the $UM(q)$ trends by 5–10 ppbv/yr at 100 hPa in northern middle and high latitudes. It is apparent from Figure 1 that the modeled water vapor trend in the LS is largely related to temperature and transport near the tropopause, although the processes of water vapor exchange between stratosphere and troposphere are complicated.

[11] The modeled $UM(q)$ trends in the mid-upper stratosphere are small and not significant. The combined trend in

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[1]Auxiliary material is available at ftp://ftp.agu.org/apend/gl/2005gl024675.
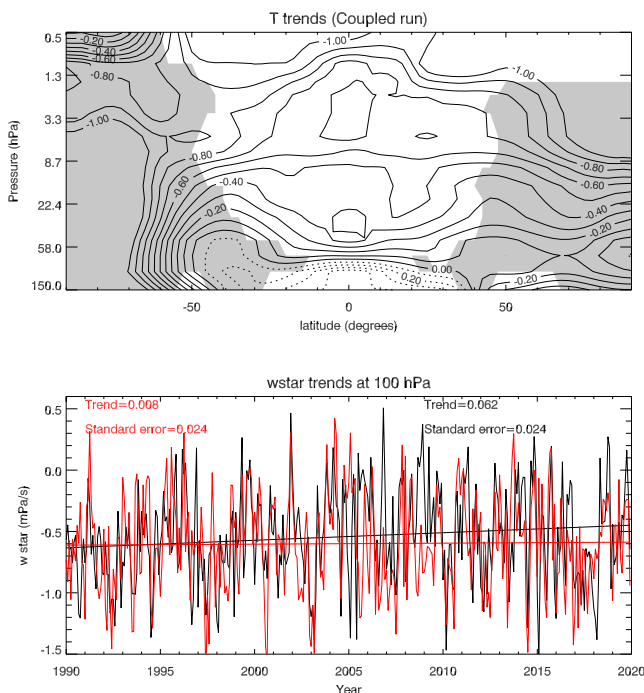
Figure 2. (top) Trends in T in the coupled run. The contour interval is ±0.1 K/decade and regions where the derived trend is not significant at the 2σ level are shaded. (bottom) Time series of the vertical velocity of the BD circulation averaged between 5°N and 5°S at 100 hPa for the coupled (black line) and uncoupled (red line) model runs. Negative values denote upward motion. The linear trend fit is also shown.

the US caused by CH_4 oxidation and cross-tropopause transport is no more than 20 ppbv/yr which is small compared with the observed trend which reaches about $40-50$ ppbv/yr [e.g., *Rosenlof et al.*, 2001; *Oltmans et al.*, 2000]. *Stenke and Grewe* [2005] also found that the simulated water vapor trend in their model is 35% weaker than values observed by *Oltmans et al.* [2000]. On the other hand, the observed water vapor trends are affected not only by the time period of the data record but also by data quality as was pointed out by *Randel et al.* [2004]. It should be kept in mind that our main objective is to evaluate the likely significance of the two different processes which can contribute to the trend in the stratospheric water vapor. Despite the uncertainties in the magnitude of the simulated water vapor trends, we can note that stratospheric water vapor may increase over the next two decades.

[12] The different CCM(H_2O) trends in the lower and upper stratosphere are related to the differences in chemical processes and the large-scale circulation in the stratosphere. The stratospheric chemical CCM(H_2O) trend will balance the chemical CH_4 trend although dynamical effects can change the balance between these two overall rates. The negative CCM(H_2O) trend in the tropical LS and large positive CCM(H_2O) trend in the US high latitudes are probably caused by the differences in large-scale transport processes. Figure 2 shows the trends in T in the coupled run and time series of the vertical velocity of the BD circulation, $\overline{\omega}^\star$, averaged between 5°N and 5°S at 100 hPa. T exhibits negative trends in the tropical mid-upper stratosphere, but

positive trends in the tropical UT/LS. The trend differences in the tropical T between the coupled and uncoupled runs are relatively large in the US but small in the LS (not shown). Note that most of the $\overline{\omega}^\star$ trends are not significant at the 2σ level. However, at 100 hPa, $\overline{\omega}^\star$ averaged between 5°N and 5°S has a positive trend (i.e., decreasing upward velocity) of 0.062 mPa/s/decade in the coupled run which is significant at the 2σ level, but the trend in the uncoupled run is not significant. At 10 hPa level, a weak negative trend in $\overline{\omega}^\star$ can be noted (see auxiliary material).

[13] The large-scale transport and methane oxidation process in the upper and lower stratosphere lead to a larger upward trend in CCM(H_2O) in the US but a weak negative trend in the tropical LS. It should be pointed out that $\overline{\omega}^\star$ in the tropical UT has a weak positive trend in both the coupled and uncoupled run although the trends do not reach the 2σ significance level (see auxiliary material). A UM study by *Butchart and Scaife* [2001] showed that GHG-induced cooling tends to increase the tropical upwelling at 68 hPa as a result of increased extratropical planetary-wave driving. However, our results show no significant coherent increasing trend in large-scale upwelling in the LS although in the US the vertical velocity of the BD circulation shows a negative trend. Also note that the dynamical feedback of our results differ from *Shindell* [2001]. In his runs, which did not have O_3 feedback, the H_2O trend due to CH_4 oxidation was near 20 ppbv/year (i.e., $2\times$ the imposed CH_4 trend) throughout the stratosphere.

[14] The stratospheric trends in UM(q), which are caused solely by processes related to the cross tropopause injection of water vapor from the troposphere, are closely related to the tropopause T and other transport processes near the tropopause such as quasi-horizontal stratosphere-troposphere exchange (STE) and large-scale upwelling. The question that arises here is what is the contribution of the tropopause T change to the differences in the UM(q) trend between the coupled and uncoupled run. The corresponding model T trends averaged from 5°N–5°S are +0.44 (coupled) and +0.42 K/decade (uncoupled) at 100 hPa and −0.48 (coupled) and −0.65 K/decade (uncoupled) at 10 hPa. The model trends at 10 hPa are well within the range of the stratospheric T trends summarized by *Shine et al.* [2003] and the cooling effect of stratospheric water vapor is also evident from the difference in average T between the coupled and uncoupled run. The trends at 100 hPa are not consistent with the cooling trend in observations [e.g., *Simmons et al.*, 1999; *Randel et al.*, 2000]. However, it has been pointed out that there are uncertainties in the observed trends due to changes in instrumentation and methods.

[15] Although the model predicts a heating trend at 100 hPa, which clearly partially explains the trend in UM(q), the contribution of the other processes to the water vapor trend appears also significant. Figure 1 indicates that the maximum difference in the UM(q) trend between the coupled and uncoupled runs reaches a maximum of 15 ppbv/yr at 150 hPa and about $5-10$ ppbv/yr at 100 hPa. However, the differences in the temperature and temperature trend near the tropical tropopause between the coupled and uncoupled runs are about 0.1 K and 0.02 K/decade, respectively. According to the estimation of *Kirk-Davidoff et al.* [1999], a water trend of 40 ppbv/yr is equivalent to a tropopause T trend of 0.05 K/yr. Therefore, the 0.02 K/

decade difference is too small to cause a significant change in the H_2O trend. The absence of significant differences in the T trend between the coupled and uncoupled run at 100 hPa implies that transport changes near the tropopause, such as changes in the BD circulation and quasi-horizontal STE processes may account for the water vapor trend differences between the coupled and uncoupled run, i.e., about 30% of the maximum UM(q) trend in the coupled run. We can see from Figure 2 that the difference in the trend of $\overline{\omega}^\star$ at 100 hPa is significant, but Figure 2 (bottom) implies no increasing trend in large-scale upwelling at 100 hPa. On the other hand, the largest differences in the UM(q) trend occur at midlatitudes (Figure 1d) suggesting that quasi-horizontal STE processes may be important. However, due to uncertainties in the circulation trend and limitations in the model setup, the relative contributions of different transport processes to the water vapor trend can not be further diagnosed and clarified in this study.

[16] From another point of view, the water vapor entering from troposphere into stratosphere is proportional to the saturation mixing ratio, Q_s, at the tropical tropopause. The trend in Q_s reflects most of the trend in stratospheric water vapor resulting from the tropopause T change. The Q_s trend is 35 and 34 ppbv/yr in our coupled and uncoupled runs (not shown), respectively. As the maximum trend in UM(q) in the coupled run is about 50 ppbv/yr, the T trend at 100 hPa accounts for roughly 70% of the UM(q) trend. This value is in accordance with that inferred from Figure 1 and the modeled T trends.

4. Summary

[17] Overall, the model does reproduce many aspects of observed past H_2O increases. Although, part of the modeled LS increase is due to the increase in tropopause T, we can still draw conclusions about the cause of the observed trend. Water vapor trends in the United States can be mostly explained by CH_4 oxidation while in the LS the trends may be mainly related to exchange of water vapor between troposphere and stratosphere. While the tropopause T change may mainly be responsible for the water vapor trend in the LS, the contribution from the change in the BD circulation and other processes are also important. From our results we can say that reconciling an observed tropopause cooling with increases in LS H_2O would require a significant change in transport in the very low stratosphere, unless other mechanisms are invoked [e.g., *Sherwood*, 2002; *Notholt et al.*, 2005].

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