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The increase in stratospheric water vapor from balloonborne, frostpoint hygrometer measurements at Washington, D.C., and Boulder, Colorado

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Abstract. Stratospheric water vapor concentrations measured at two midlatitude locations in the northern hemisphere show water vapor amounts have increased at a rate of 1-1.5% yr^{-1} (0.05-0.07 ppmv yr^{-1}) for the past 35 years. At Washington, D.C., measurements were made from 1964-1976, and at Boulder, Colorado, observations began in 1980 and continue to the present. While these two data sets do not comprise a single time series, they individually show increases over their respective measurement periods. At Boulder the trends do not show strong seasonal differences; significant increases are found throughout the year in the altitude range 16-28 km. In winter these trends are significant down to about 13 km.

Introduction

Water vapor mixing ratios in the stratosphere are roughly 1/1000 of those measured near Earth's surface, yet water vapor plays a key role in the chemical and radiative behavior of the upper troposphere and stratosphere. Long-term changes of water vapor amounts in the stratosphere have been found in both surface-based [*Oltmans and Hofmann*, 1995; *Nedoluha et al.*, 1998] and satellite [*Evans et al.*, 1998; *Nedoluha et al.*, 1998; *Abbas et al.*, 1996; *Michelsen et al.*, 2000] data sets. Inclusion of increases in stratospheric water vapor in climate models has shown varying impact on modeled atmospheric temperatures both in the stratosphere itself and at Earth's surface. In some cases [*Ramaswamy et al.*, 1998] the cooling due to stratospheric ozone losses dominates cooling from water vapor increases. Recent simulations [*Forster and Shine, 1999*] (V.L. Dvortsov and S. Solomon, Response of the stratospheric temperatures and ozone to past and future increases in stratospheric humidity, submitted to *J. Geophys. Res.*, 2000) find roughly equal cooling from ozone and water vapor in the lower stratosphere. At the surface the warming due to the increase in stratospheric water vapor found by *Forster and Shine* [1999] for the period 1979-1997 is about 40% of that found for carbon dioxide increases alone over that period.

This paper presents results for two time series of balloonborne water vapor profiles obtained at midlatitudes of the northern hemisphere. From 1964 to 1979 soundings were made near Washington, D.C., by the Naval Research Laboratory [*Mastenbrook*, 1968; *Mastenbrook and Oltmans*, 1983]. Beginning in 1980, and continuing to the present, profiles have been obtained at Boulder, Colorado. As discussed in the following, these two data sets cannot be considered as a single time series, but even as individual series, they represent relatively long measurement records for stratospheric water vapor. The long-term stratospheric trends are presented along with the seasonal pattern in these trends for Boulder.

Measurement Method

The water vapor profiles are obtained using balloonborne, chilled-mirror hygrometers. The measurement technique and its application for this sort of study were previously described [*Mastenbrook and Oltmans*, 1983; *Vömel et al.*, 1995; *Oltmans and Hofmann*, 1995]. The instrument measures water vapor by detecting the temperature of a chilled mirror on which a thin layer of frost is maintained. The frostpoint temperature is converted to mixing ratio using the Goff-Gratch formulation of the Clausius-Clapeyron relationship. Based on an analysis of potential instrument errors for measurements in the stratosphere, the precision is 10% in mixing ratio for an individual profile at Boulder and 15% at Washington, D.C. Over the 35-year span covered by these measurements the instrument has evolved, taking advantage of various improvements in electronic and optical components. The equipment used for calibrating the thermistor that measures the mirror face temperature has changed, but all of these methods are traceable to a U.S. National Institute of Standards and Technology reference, have been intercompared with each other, and have been found to give very similar results (within 0.1K).

Solid-state electronics and optical components were introduced into the instrument beginning in 1977. Between 1977 and the conclusion of the measurement program at Washington, D.C., in 1980, only this solid-state version of the instrument was flown. From the measurement time series at Washington, D.C., it appears that, beginning in 1977 with the advent of the newly designed instruments, water vapor mixing ratios were systematically lower than those measured immediately prior to the switch. Unfortunately there were no overlapping flights of the two instrument types, and there is a gap of 16 months between the June 1976 flight and the first flight with the redesigned instrument in November 1977. Because of the discontinuity with the instrument change and the small number of profiles (11 in 4 years) after June 1976, the time series analysis was done for the period 1964-1976.

At Boulder only the solid-state version of the instrument has been used. In 1991 the hygrometer was adapted to digital signal transmission. No shifts are evident as a result of the conversion from the analog to digital telemetry system or the change in air temperature or pressure measurement techniques. Because of the

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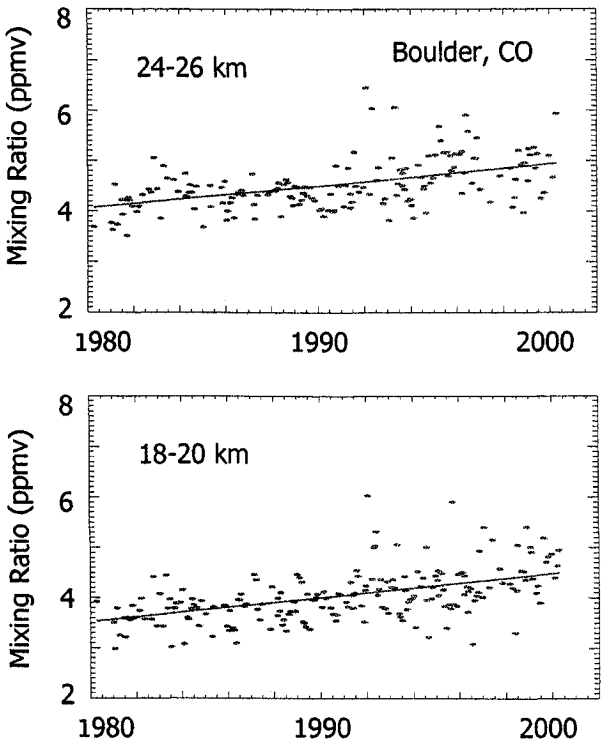

Figure 1. Time series of individual soundings in two layers (18-20 km and 24-26 km) at Boulder, Colorado. The linear trends and 95% confidence intervals are 0.048 ± 0.011 ppmv yr^{-1} (18-20 km) and 0.044 ± 0.012 ppmv yr^{-1} (24-26 km).

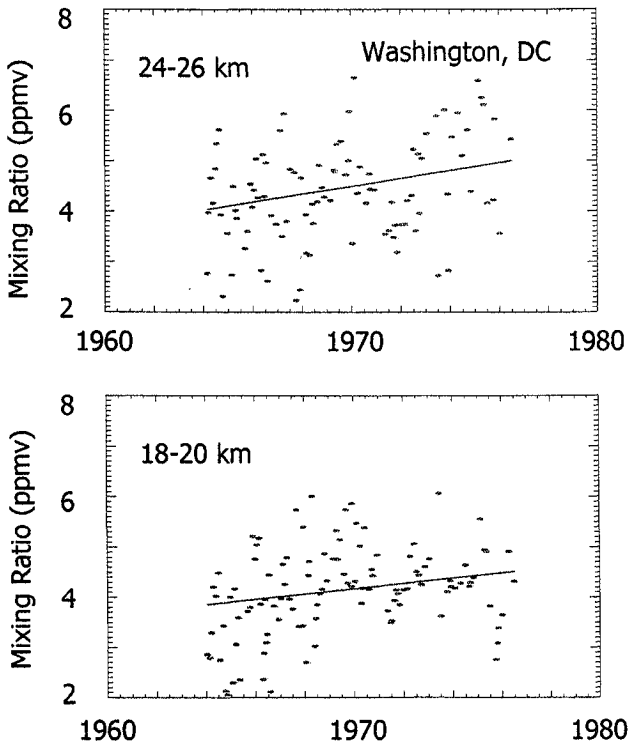

Figure 3. Time series of individual soundings in two layers (18-20 km and 24-26 km) at Washington, D.C. The linear trends and 95% confidence intervals are 0.053 ± 0.043 ppmv yr^{-1} (18-20 km) and 0.079 ± 0.048 ppmv yr^{-1} (24-26 km).

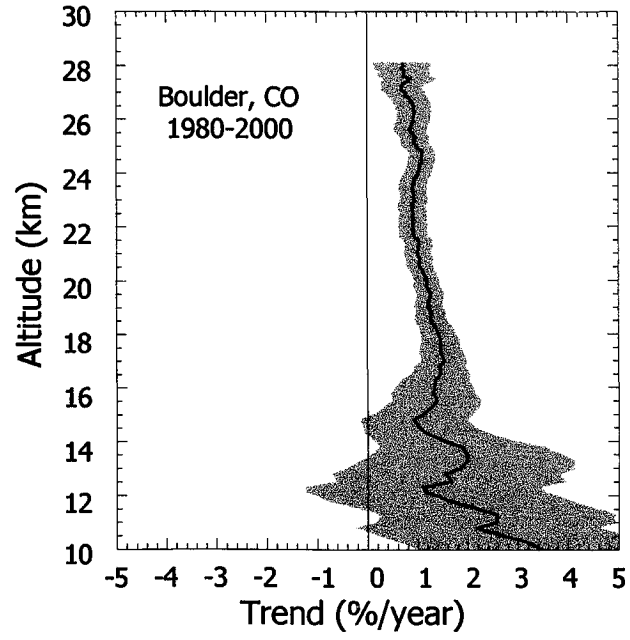

Figure 2. Summary of vertical profile percentage trends in the stratosphere over Boulder for all data. The solid line is the trend computed at 250 m altitude increments. The shaded area is the 95% confidence region.

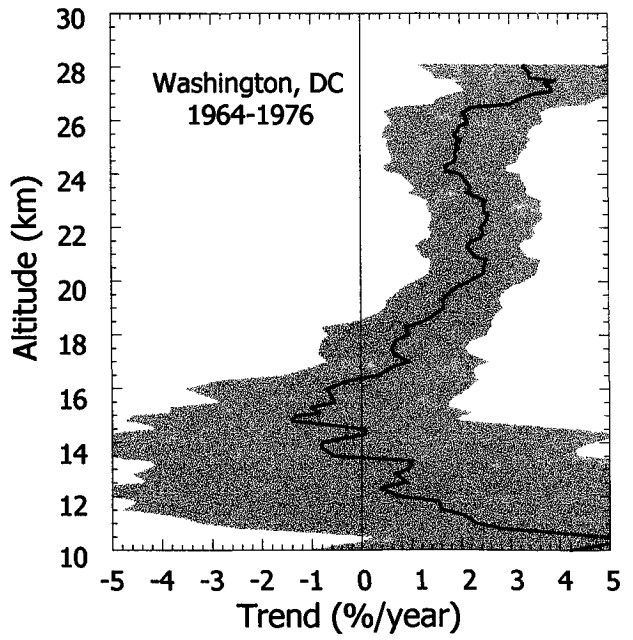

Figure 4. Summary of vertical profile percentage trends in the stratosphere over Washington, D.C., for all data. Solid line is the trend computed at 250 m altitude increments. The shaded area is the 95% confidence region.

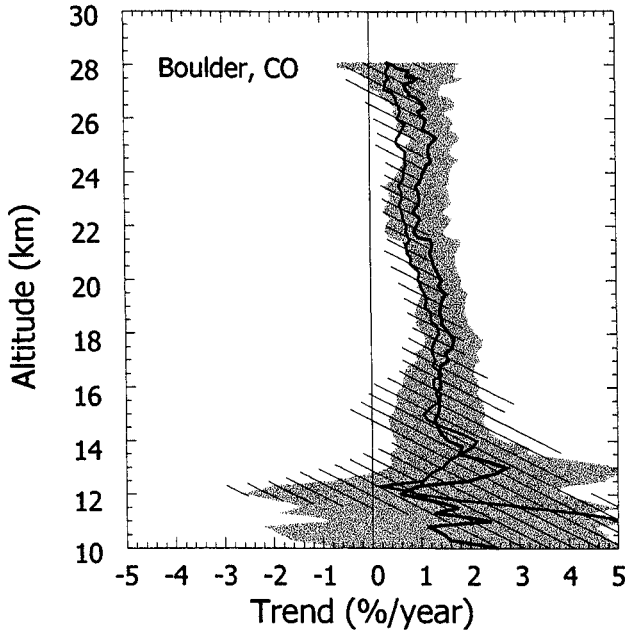

Figure 5. Vertical profile percentage trends for two seasons (Dec.-March and Aug.-Nov.) over Boulder. The thicker solid line and the shading are for Dec.-March. The thinner solid line and crosshatching are for Aug.-Nov.

potential for contamination from instrument and balloon outgassing of water vapor, the instrument is flown in a configuration that allows partial valving of helium from the balloon at the balloon ceiling altitude to give a controlled balloon descent. This allows the air inlet to be ahead of the instrument and balloon package, minimizing potential contamination. Improvement to the inlet system in recent years gives uncontaminated ascent data up to about 25 km rather than the 18-20 km of earlier flights. For this analysis where ascent and descent data are both available and of good quality, an average for a layer includes ascent and descent data up to the altitude where the mixing ratios diverge.

Trend Results

The time series of water vapor mixing ratio (ppmv) at Boulder for two layers (18-20 km and 24-26 km) that are above the altitude where rapid transport from the tropics produces a pronounced seasonal variation are shown in Figure 1. The linear fit plotted for the individual sounding values is 0.048 ± 0.011 ppmv yr^{-1} at 18-20 km and 0.044 ± 0.012 ppmv yr^{-1} at 24-26 km where the uncertainties are 95% confidence intervals using the t-statistic. Percentage trends computed at 250 m intervals over the entire altitude range from 10-28 km show that increases over Boulder (Figure 2) above 16 km are all statistically significant and are rather uniform at about 1% per year. The uncertain trends below 16 km reflect the strong variability seen in the lowermost stratosphere and the variation in tropopause height.

For Washington, D.C., the mixing ratios for the same two layers as Boulder are shown in Figure 3 with the fit to the data that yields a linear trend of 0.053 ± 0.043 ppmv yr^{-1} in the 18-20 km layer and 0.079 ± 0.048 ppmv yr^{-1} at 24-26 km. The variability for this site is significantly larger than at Boulder. This likely reflects the lower precision of the Washington, D.C., soundings, since it is very unlikely that the variability in the stratosphere at these altitudes would be different at the two sites. The trends computed at 250 m intervals between 10-28 km (Figure 4) show significant increases between 18-28 km. This is a similar altitude range over which the trends in Boulder were significant. The increase is somewhat larger at Washington, D.C., averaging about 1.5% yr^{-1} over the altitude range of significant increases, although this difference may not by statistically significant.

Discussion and Conclusion

Trends for two contrasting seasons (Dec.-March and Aug.-Nov.) as a function of altitude are shown in Figure 5. The intermediate season (April-July) shows trends generally between the two seasons shown in Figure 5. There are essentially no times of the year or altitude regions where water vapor is decreasing. Trends above 16 km are larger during winter than in other seasons, but given the generally overlapping significance estimates, these differences may not be significant. During the winter the trends are significant down to altitudes of about 13 km. This likely reflects the inclusion of this region in the stratosphere in winter. A recent analysis of the seasonal trends for the shorter record (1992-1999) of HALOE [*Smith et al.*, 2000] suggests that the autumn (Sept.-Nov.) tropical upper troposphere may be the source of the increases seen over this time period. This change is then reflected at higher altitudes and latitudes in other seasons through the action of the global circulation. Autumn is also a time of the year when the tropical tropopause is not saturated, as noted below, allowing for the flux of additional water vapor into the stratosphere. A change in the extremity of the Asian summer monsoon [*Evans et al.*, 1998; *Smith et al.*, 2000] has also been proposed as a possible cause of the autumn increase seen in the HALOE data.

A long-term change in the water vapor content of the lower stratosphere over Boulder could be related to several possible causes. Because methane oxidation is a source of water vapor in the stratosphere [*Le Texier et al.*, 1988] increasing methane levels in the atmosphere [*Dlugokencky*, 1998] should lead to rising stratospheric water vapor amounts [*Singer*, 1971]. However, as water vapor has continued to increase in the stratosphere, there has been a slowing in the growth rate of methane in the troposphere from about 0.014 ppmv yr^{-1} in 1984 to about 0.003 ppmv yr^{-1} in 1996 [*Dlugokencky*, 1998]. It seems likely that another source in addition to methane oxidation is contributing significantly to the continuing water vapor increase over Boulder.

One possible cause of a change in water vapor is a modification in transport characteristics that can alter the amount of tropical air reaching Boulder. At present evidence of such a change has not been suggested or demonstrated. Alteration of the water vapor content of air from the lower stratosphere that is transported to Boulder might occur as a result of a change in tropical tropopause region characteristics. The required tropopause (or cold point) temperature increases needed to produce the observed water vapor increase over Boulder are quite small. For an average tropopause temperature and altitude for air crossing the tropical tropopause at water saturation, less than a 1 K change in the tropical tropopause temperature would be required to account for about one half of the water vapor change seen at Boulder. Recent studies have actually found decreases in tropical cold-point [*Zhou et al.*, 2000] or tropopause [*Simmons et al.*, 1999, *Randel et al.*, 2000] temperatures. Other characteristics of the tropical tropopause or cold-point temperature may also change. An increase in the altitude of the cold-point of about 15 hPa can also cause a moistening of the lower stratosphere equivalent to about a 1 K temperature increase of the tropical tropopause, but this has not been observed [*Randel et al.*, 2000]. Since the tropical tropopause is not always at saturation, a moistening of the upper troposphere can also lead to a wetter strato-

sphere without a warming of the cold-point temperature [*Vömel and Oltmans*, 1999].

With both of the sites considered here located at midlatitudes of the northern hemisphere, a possible moistening of the lowermost stratosphere by subsonic aircraft that spend a portion of their flight time in the stratosphere is another potential source of increasing water. Although increases in the aircraft emissions of water vapor into the stratosphere have been substantial over this period, it does not seem likely that this has been the source of the trend that is seen to the highest levels measured by the balloon [*Danilin et al.*, 1998; *Fahey and Schumann*, 1999].

A moistening of the tropical lower stratosphere eventually leads to a wetter stratosphere in general as the extratropical wave driven circulation transports air upward and poleward [*Holton et al.*, 1995]. The trend above 24 km is consistent with the results seen from the shorter HALOE record on the UARS and ground-based millimeter wavelength measurements [*Evans et al.*, 1998; *Nedoluha et al.*, 1998]. In the lowermost stratosphere (below about 20 km), the HALOE measurements do not show a significant increase [*Randel et al.*, 1999], but the satellite data do suggest that the increase [*Evans et al.*, 1998, *Randel et al.*, 1999] seen over Boulder is not confined to a particular location.

The trend of increasing stratospheric water vapor reported earlier for Boulder [*Oltmans and Hofmann*, 1995] is continuing at a rate of about 1% per year through the altitude range covered by the balloon measurements (14–28 km). This increase is well above that expected from rising methane levels. The corroborating increases seen in the earlier measurements from Washington, D.C., suggest that increases of water vapor extend back at least 35 years and thus are a result of longer-term changes than might be expected with decadal scale circulation fluctuations. While the results from Boulder and Washington, D.C., are intriguing, it is clear in light of the possible significance of stratospheric water vapor changes in climate forcing [*Forster and Shine*, 1999] that additional measurements are essential.

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