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Climate and ozone response to increased stratospheric water vapor

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Abstract

Stratospheric water vapor abundance affects ozone, surface climate, and stratospheric temperatures. From 30–50 km altitude, temperatures show global decreases of 3–6 K over recent decades. These may be a proxy for water vapor increases, as the GISS climate model reproduces these trends only when stratospheric water vapor is allowed to increase. Observations suggest that stratospheric water vapor is indeed increasing, however, measurements are extremely limited in either spatial coverage or duration. The model results suggest that the observed changes may be part of a global, long-term trend. Furthermore, the required water vapor change is too large to be accounted for by increased production within the stratosphere, suggesting that ongoing climate change may be altering tropospheric input. The calculated stratospheric water vapor increase contributes an additional $\approx$24% ($\approx$0.2 W/m^2) to the global warming from well-mixed greenhouse gases over the past two decades. Observed ozone depletion is also better reproduced when destruction due to increased water vapor is included. If the trend continues, it could increase future global warming and impede stratospheric ozone recovery.

Introduction

Observations in the middle and upper stratosphere from 30–50 km from satellites, rocketsondes and lidars all show 1–2 K/decade cooling over the two to three decades of data [*Golitsyn et al., 1996; Keckhutet al., 1999; Dunkerton et al., 1998; World Meteorological Organization, 1999* (hereafter WMO99)]. Several factors have likely influenced these trends. Observations show that upper stratospheric ozone has been decreasing. In percentage terms, the largest decreases have been at 40–50 km altitude (≈2–1 mb), and at 15–20 km (≈150–70 mb), where they are 6–10%/decade [*WMO99*]. Since ozone absorbs incoming solar radiation, heating the atmosphere, this depletion will have cooled these regions of the stratosphere [*Ramaswamy et al.*, 1996]. Concurrently, the well-mixed greenhouse gases CO_2, CH_4, N_2O, and the chlorofluorocarbons have increased These gases also cool the stratosphere.

Stratospheric water vapor changes would have a similar effect [*Rind and Lonergan, 1995; Forster and Shine, 1997,1999; WMO99*] The only significant long-term sources of stratospheric water vapor are *in situ* production resulting from methane oxidation, and transport from the troposphere, which takes place primarily through the tropical tropopause [*e. g. Dessler et al., 1995; Holton et al., 1995; Mote et al., 1996*] The latter is governed by the temperature minimum at the tropopause, and could be greatly altered by changes as small as a few tenths of a degree [*Evans et al.*, 1998]. Additionally, cross-tropopause transport of water vapor can be limited by water vapor availability in some seasons, so that an overall increase in tropospheric water vapor, predicted by most climate models [*International Panel on Climate Change*, 1995], could lead to an increase in stratospheric water vapor even without a tropopause temperature change. Though the observational record is limited, seven years of satellite (Halogen Occultation Experiment - HALOE) and ground-based observations of the middle and upper stratosphere [*Evans et al.*, 1998; *Nedoluha et al.*, 1998] and a 14-year set of lower stratospheric measurements over Boulder, Colorado [*Oltmans and Hofmann*, 1995] suggest that stratospheric water vapor has been increasing. Very recently, the trend at 1 mb slowed or stopped during 1996–1998, though the trend at 10 mb did not slow as much [*Randel et al.*, 1999]. A comparison of aircraft observations taken from 1993 to 1997 showed no significant trend in lower stratospheric water vapor [*Hurst et al.*, 1999]. Another aircraft study, however, did find large increases in water vapor abundances between 1986 and the mid-1990s [*Peter*, 1998]. Given these uncertainties, the question remains open as to current and future trends in stratospheric water vapor.

Model Description

We have used a version of the Goddard Institute for Space Studies (GISS) general circulation model (GCM) which includes a complete stratosphere to investigate the source of the observed long-term upper stratospheric cooling and ozone depletion [*e.g. Shindell et al., 1998a,b*]. The version used has 8° x 10° resolution, with vertical layers extending from the surface to 85 km. Use of a model with an upper boundary this high allows for a very good simulation of middle and upper stratospheric temperatures and circulation. Similar to previous simulations [*Shindell et al.*, 1998a], the model was run with increasing greenhouse gases (GHGs) based on observations through the 1980s, and using projected ghg emissions and chlorine trends. To en-

able simulation of pre-1979 and future ozone amounts, parameterized ozone chemistry derived previously from our 2D photochemical model [*Shindell et al.*, 1998b] was installed into the GCM. Ozone responds to anomalies in constituent abundances relative to the present, along with temperature changes and changes in local radiative flux resulting from differences in the overhead ozone column. Polar stratospheric clouds do not respond to changes in water vapor, however, and ozone transport changes do not interact with the radiation, though these are important primarily in the lower stratosphere. Though the chemistry is simplified, use of a full GCM allows us to simulate changes in tropical tropopause temperatures and in circulation, and to assess the climate response.

The model was run in four experimental setups: GHGs, with well-mixed greenhouse gas changes, but fixed water and ozone (G); Ozone, with ghg and chlorine changes, calculated ozone, and fixed water vapor (G + O); MethOx, with ghg and chlorine changes, calculated ozone, and water vapor increases due to methane oxidation, as derived from the 2D model (G + O + M); and Water, with ghg and chlorine changes, calculated ozone, and the GCM's internally calculated water vapor plus the contribution from increased methane oxidation (G + O + M + W). In the MethOx and Water runs, water is allowed to increase throughout the stratosphere and ozone is allowed to respond. Water vapor trends in the MethOx and Water runs are shown in Table 1, along with observed values. Model increases in the lower stratosphere lie between the balloon and satellite observations. Model values generally follow the vertical shape of the observed trends (greater increases at higher altitudes), but have smaller magnitudes than in the HALOE data. HALOE water increases even more rapidly than in the model as it includes increasing conversion of methane to water with time [*Randel et al.*, 1999], perhaps related to changes in circulation [*Nedoluha et al.*, 1998].

Simulated Ozone and Temperature Trends

The model parameterizations reproduce the overall spatial distribution and the magnitude of the ozone trends observed by satellite (Figure 1). The upper stratosphere is the region of greatest interest here, and where the observations are statistically significant. Agreement in this region is best when water vapor increases are included. Ozone trends are less certain in the lowermost stratosphere [*WMO99*], but the model's decreases in the Ozone run are near the lower end of depletion trend estimates, as for most photochemical models [*Dvortsov et al.*, 1999]. As in the upper stratosphere, inclusion of water vapor induced ozone destruction seems to improve the model's response. The net ozone change is made up of individual contributions from several different mechanisms (Figure 2). Increased chlorine loading plays the dominant role in upper stratospheric ozone depletion, with an additional contribution from water vapor. These are partially offset by temperature changes, whereby cooling due to increasing ghgs leads to slower rates for ozone destroying chemical reactions, and by the direct chemical effect of increasing methane (not shown), which removes chlorine from forms that destroy ozone [*e. g., Siskind et al., 1998*].

We compare modeled and observed temperature trends at 43 N, where the measurements are dense, in Table 2. The model's upper stratosphere cools in response to increasing ghgs and ozone depletion, but much

less than the observed trends at 40 km and above. Including a water vapor increase due to methane oxidation and the ozone response improves the result, but only with the inclusion of additional water vapor as calculated by the GCM internally can the model reproduce the observed upper stratospheric cooling trends within the 1σ measurement uncertainty. Similar results were found at other latitudes where observations were available.

While the temperature response to ghgs in the GISS model agrees well with other GCMs (e.g. WMO99, p. 12.30, showing the response of various models to increasing ghs and halogens, similar to the Ozone run), it differs from fixed dynamical heating (FDH) models [*Forster and Shine,* 1999; *Ramaswamy et al.,* 1992]. As shown in Figure 2 of [*Rind et al.,* 1998], the change in temperature by dynamics becomes quite important at and above about 50 km. At these levels, dynamics caused heating rates of roughly 0.25 K/day for a doubled CO2 scenario. The heating change by dynamics in these experiments was approximately one-tenth this value over a decade, leading to a decrease in the greenhouse gas induced cooling rate of about 0.2 K at 50 km. This dynamical effect causes the maximum cooling from greenhouse gas increases to occur at lower altitudes in the GCM than in the simpler FDH models. Additionally, the Forster and Shine model has a top at 1 mb, making the upper stratosphere subject to the unrealistic influence of the model top (they focus on the lower stratosphere in their paper, for which their model extends adequately high). This probably accounts for discrepancies between their results and those of the more extensive Ramaswamy et al FDH model.

Temperature trends in the Ozone, MethOx and Water runs also depend upon the imposed water vapor increases and the ozone changes. Our values maximize in the upper stratosphere, where the water vapor increases fastest (see Tables). In contrast, Forster and Shine's work included a uniform water increase throughout the stratosphere, at values appropriate for the lower stratosphere, their primary interest. It therefore makes sense that our temperature response in the upper stratosphere is larger than theirs. However, some differences must also be due to the particular models. In the GISS GCM, the vertical profiles of the response to a uniform increase of water vapor or ghgs are quite similar, both maximizing in the upper stratosphere [*Rind and Lonergan,* 1995], while in the FDH model with a top at 1 mb, they are not [*Forster and Shine,* 1999]. It makes physical sense that the response maximizes near the local temperature maximum at the stratopause, where each molecule is radiating more energy than the levels above or below, so this behavior seems logical for both greenhouse gases and water vapor.

The use of observed ozone changes would not fully separate out the influence of water on temperature trends, as those ozone changes include chemical effects of water on ozone. This is especially important in the lower stratosphere and above about 5 mb (Fig. 2). Attribution of temperature trends in those regions is therefore especially complex.

Long-term monitoring of stratospheric water vapor will be necessary to definitively establish trends. However, our model-observation comparisons, based on well-established, long-term temperature trends, suggest that there must have been an increase in stratospheric water vapor over the past several decades. Furthermore, the increase was larger than that due to *in situ* production from methane oxidation alone, in agreement with the di-

rect water vapor measurements [*Evans et al.*, 1998; *Nedoluha et al.*, 1998; *Oltmans and Hofmann*, 1995]. This implies that a portion of the increase resulted from enhanced transport of water from the troposphere to the stratosphere. This occurs in the GCM due to a warming of the tropical tropopause caused by the increasing greenhouse gases. Note that observations do not indicate an overall warming trend in tropical tropopause temperatures [*Simmons et al.*, 1999]. Given the complexity of cross-tropopause transport of water vapor, however, this does not rule out an increase in the annual flux of water vapor into the stratosphere due to climate change.

Model simulations continued to 2070. Temperature trends vary markedly between the different simulations, especially in the future. The greenhouse gas run shows a steady decrease with time at 0.7 mb (≈50 km) (Figure 3). Including ozone depletion accelerates the trend from 1959 to about 2000, but then the trend levels off as greenhouse gas induced cooling is offset by warming due stratospheric ozone recovery at this level. Thus if there were no increases in stratospheric water vapor, the rapid decrease in upper stratospheric temperatures should soon level out. However, when water vapor increases, we see a long-term continuation of upper stratospheric cooling. Similar behavior is seen at other altitudes, with larger effects with increasing altitude.

Ozone trends also show large differences in the future. At high altitudes, hydrogen oxides dominant the chemical reactions that destroy ozone, so that increased water leads to an ozone decrease (Fig. 2). In the middle stratosphere though, ozone is less sensitive to hydrogen oxides, so there is a net ozone increase due to the water-induced cooling and increased radiation resulting from the reduced overhead ozone column. These effects are similar to those computed by *Evans et al* [1998]. In the lower stratosphere, which dominates the ozone column, ozone chemistry is again quite sensitive to the abundance of water, which participates in homogeneous and heterogeneous reactions. Nitrogen removal through heterogeneous hydrolysis of dinitrogen pentoxide increases the ability of chlorine to destroy ozone, so that in the current atmosphere, with abundant chlorine, water increases lead to a net chemical loss of ozone. The importance of this heterogeneous chemistry was unequivocally confirmed when it was enhanced in response to the injection of stratospheric aerosols by Mt. Pinatubo [*Solomon et al.*, 1998]. Decades from now, if chlorine is greatly reduced, the removal of nitrogen should increase ozone by reducing the ozone depleting reactions with nitrogen oxides. The predicted behavior of the ozone column over time in the simulations shows that the influence of increasing water vapor is to delay the recovery of stratospheric ozone past the peak chlorine loading of 2000–2005 (Figure 4). While increasing greenhouse gases in the ozone run speed up the recovery of ozone, when water vapor increases are also included, the time of recovery to 1979 levels at mid-latitudes is extended past the end of the simulations.

While the modeled ozone changes shown in Figure 1 agreed fairly well with observations, the observations themselves show significant disagreements in the lowermost stratosphere [*WMO99*]. Given the simplified ozone anomaly model used here, along with the difficulties of both defining and reproducing mid-latitude ozone trends near the tropopause [*Dvortsov et al.*, 1999], the future ozone column amounts should be treated primarily as qualitative predictions of the influence of water vapor.

In the polar regions, "ozone holes" are created by massive ozone destruction resulting from chemical processing on the surfaces of polar stratospheric clouds. Polar stratospheric cloud formation could be enhanced by increased stratospheric water vapor [*WMO99, Kirk-Davidoff et al., 1999*], leading to even greater ozone losses and a further delay relative to that predicted due to ghg-induced cooling alone [*Rind and Lonergan, 1995; Forster and Shine, 1997,1999; Shindell et al., 1988a; Dameris et al., 1998*].

Climate Response and Conclusions

Changes in both stratospheric water vapor and ozone affect surface climate. From 1959 to 1999, simulated stratospheric ozone depletion caused a global-mean annual average change of surface air temperature of -0.09 K (-0.07 K for 1979–1999). This is somewhat less than the -0.12 to -0.20 K calculated with the standard (limited stratosphere) GISS climate model using observed 1979–1994 ozone changes [*Hansen et al.*, 1997]. This is due to the photochemical model's bias toward underestimation of ozone losses in the lowermost stratosphere, where the climate sensitivity to ozone change is largest. During 1959–1999, the simulated water vapor increase in the Water run (water increases due to climate change and methane oxidation) led to an increase of +0.26 K (+0.07 K/decade). *Forster and Shine* [1999], who used the observed lower stratospheric water vapor trends, found a similar increase in global-mean annual average surface air temperature: 0.11 K for the eighteen year period 1979–1997 (+0.06 K/decade). The net change from stratospheric water and ozone was thus +0.17 K, approximately 33% of the observed increase. Given the large uncertainties in climate forcings due to clouds and aerosols [*Cess et al.*, 1997; *Hansen et al.*, 1998], such a large value is not inconsistent with current understanding.

The argument made here, for an indirect confirmation of a water vapor trend based upon the long-term, well-established temperature observations in the upper stratosphere, suggests that impacts of stratospheric water vapor must be accounted for in climate and ozone studies. Changes in water and ozone are partially "forcings", resulting directly from changes in anthropogenic emissions, and partially "feedbacks", resulting from the meteorological response to emission changes. For water vapor, a portion can be attributed directly to methane, while the remainder must be attributed to all the greenhouse gases (including ozone) that affect circulation and tropopause temperature. For ozone changes, after another few decades, the influence of ghgs and water vapor on ozone becomes larger than that of halogens, so that attribution should progressively switch from halogens to climate change. Given the potential importance of water vapor trends, a thorough observational program is clearly needed.

Future ozone recovery will affect its climate forcing. In the model simulations, ghg induced cooling is strongest in the upper stratosphere, leading to a much more rapid recovery there than in the lower stratosphere. Furthermore, increased upper stratospheric ozone absorbs more incoming radiation, reducing ozone production in the lower stratosphere. Since upper stratospheric ozone increases lead to surface cooling [*Rind and Lacis,* 1993], they largely offset the radiative impact of the slowly increasing abundances of lower stratospheric ozone. By 2050, the change in surface air temperature due to

ozone is reduced to -0.03 K, while the change due to water vapor has increased to +0.43 K. The net stratospheric water-plus-ozone forcing may thus more than double over the coming half-century, contributing an additional 10–15% to the global warming due to the well-mixed greenhouse gases.

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Figure 1. Modeled and observed ozone changes during the 1980s and early 1990s. Values are annual average trends in percent per year. Model results are shown for the Ozone (G + O) and Water (G + O + M + W) runs, in the left and center panels, respectively. Shading in the data from the Stratospheric Aerosol and Gas Experiment (SAGE) satellite (right) indicates that the results are not statistically significant. Note that that data only extends to about 60 degrees.

Figure 2. Influence of individual factors on ozone. Values are 1979 to 1996 annual average trends in percent per decade from the MethOx run. Trends are from temperature, chlorine, radiation (local change due to change in overhead column) and water changes, based on the prescribed trends given in the text. Methane and nitrous oxide effects were smaller than those shown here. Reactions on PSCs caused large ozone changes, but only at high latitudes in the lower stratosphere (see Figure 1). The effect of water vapor shown here is the chemical effect, which is partially canceled by its contribution to greenhouse cooling included under the temperature effect.

Figure 3. Temperature trends annually averaged over 60N-60S, at 0.7 mb (≈50 km altitude). Values are given relative to 1980. Model trends have been smoothed with a ten-year weighting (simulations described in text). Observations taken by the Stratospheric Sounding Unit (SSU) over roughly 44–56 km altitude are shown as open circles. There is little latitudinal structure to the modeled temperature changes, as in the observations.

Figure 4. Total column ozone trends at 43 N. Percentage depletion for the three simulations with calculated ozone are shown on the left axis, smoothed with a ten-year weighting. On the right axis is the prescribed atmospheric chlorine loading in parts per billion by volume. Results from the Ozone run are similar to those predicted at northern mid-latitudes by 2D models. Satellite and ground-based observations from 1979 through 1997 show a decrease of about 3.3 ± 1.5 percent per decade at this latitude, with indications of a decreasing loss rate since 1997 [*WMO99*]

Table 1. Water vapor trends (ppbv/yr)

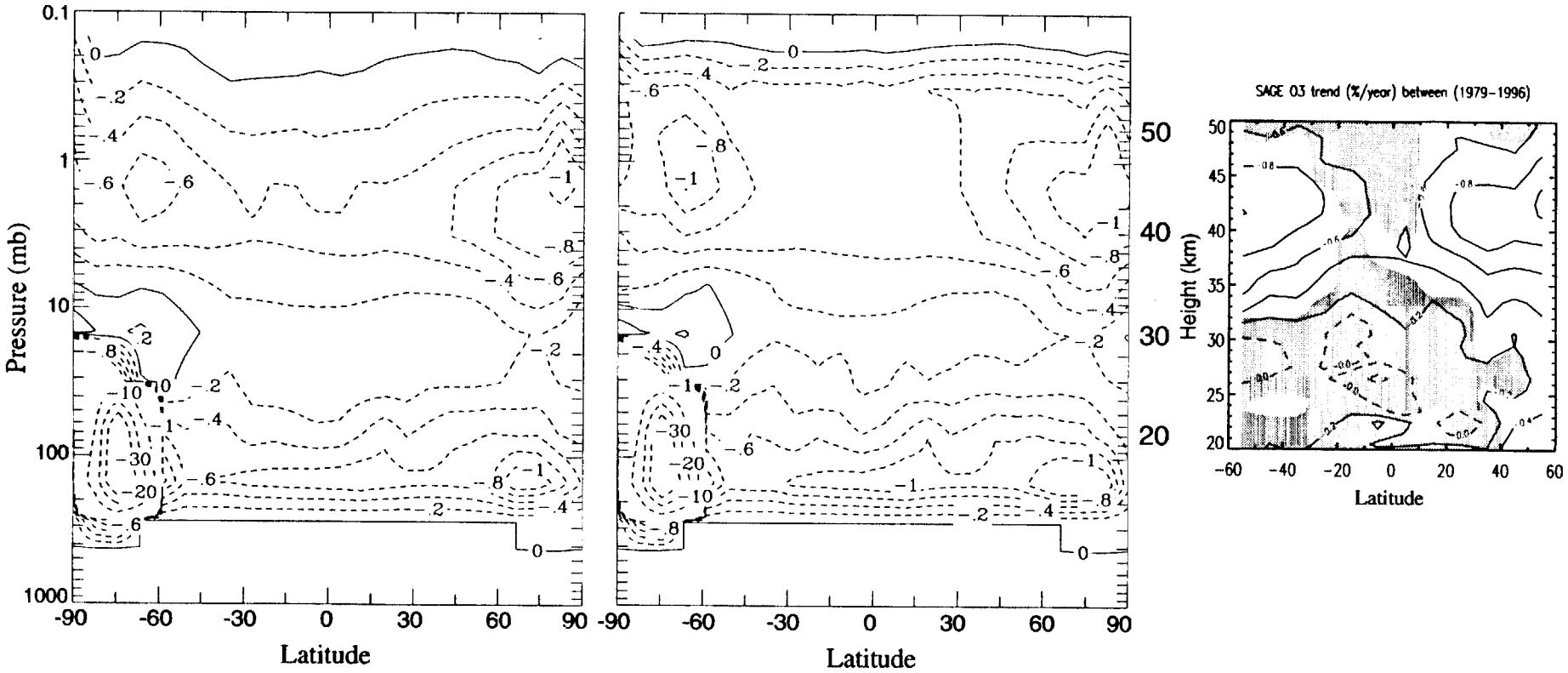
Pressure (mb)	Altitude (km)	MethOx 43N	Water 43N	MethOx 60S-60N	Water 60S-60N	Balloons 40N	HALOE 60S-60N
15	29	19	40	20	51	15	60 - 65
6.8	34	20	44	21	57	na	85 - 90
3.2	39	22	61	21	66	na	90 - 95
1.5	44	24	75	23	78	na	110 - 115
0.7	50	27	67	26	69	na	100 - 105

Model altitudes are approximate equivalent altitude for the GCM layers centered at the given pressure levels. Balloon data are those reported by *Oltmans and Hofmann* (1995) for Boulder, Colorado for 24–26 km altitude (the highest level reported). Uncertainty on that data is ± 15 ppbv/yr at the 95 percent confidence level. HALOE data are those reported by *Randel et al.* (1999), and are fits to 1993–1997 measurements. Uncertainty on the HALOE data is given as approximately ± 15 ppbv/yr at 32 mb, and ± 20 ppbv/yr at 1 mb.

Table 2. Temperature change, K/decade, 1979–1994

Pressure (mb)	Altitude (km)	Observations $\pm 1\sigma$	GHG run (G)	Ozone run (G+O)	MethOx run (G+O+M)	Water run (G+O+M+W)
15	29	-0.80 ± 0.29	-0.24	-0.29	-0.40	**-0.53**
6.8	34	-0.88 ± 0.30	-0.47	**-0.63**	**-0.73**	**-0.84**
3.2	39	-1.23 ± 0.33	-0.46	-0.70	-0.84	**-1.43**
1.5	44	-1.81 ± 0.37	-0.45	-0.74	-0.94	**-1.68**
0.7	50	-2.55 ± 0.40	-0.33	-0.63	-1.37	**-2.32**

Altitudes for the GCM runs are approximate, as in Table 1. The observations are from several, independent techniques, with an average latitude of 45 N, and altitudes at 5 km intervals beginning at 30 km [*WMO99*]. Modeled values lying within 1σ of the observations are shown in bold type. Latitudinal structure in the GCM does not alter the conclusion that water vapor increases are required to reproduce the observed trends. For comparison, the 60S-60N temperature trend averages for the GHG run are -0.33, -0.46, -0.55, -0.60, and -0.64 with increasing altitude, while for the Ozone run, they are -0.39, -0.60, -0.75, -0.85 and -0.92. See also Figure 3 for 60S-60N averages.


Pressure (mb)
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Latitude
-90 -60 -30 0 30 60 90
Latitude
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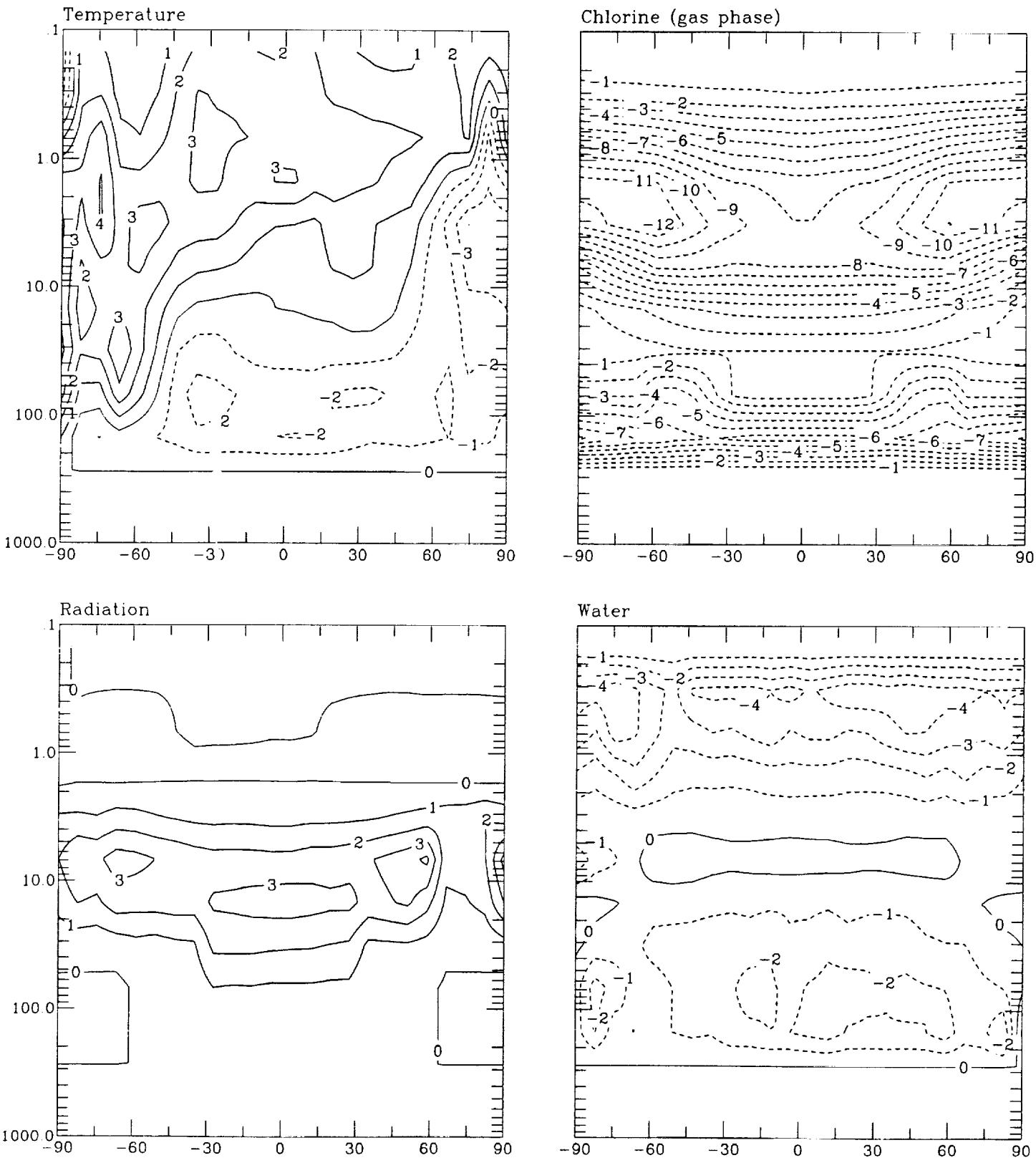
Height (km)
SAGE 03 trend (%/year) between (1979-1996)
50
45
40
35
30
25
20
-60 -40 -20 0 20 40 60
Latitude

Fig 1


Temperature
Chlorine (gas phase)
Radiation
Water
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30
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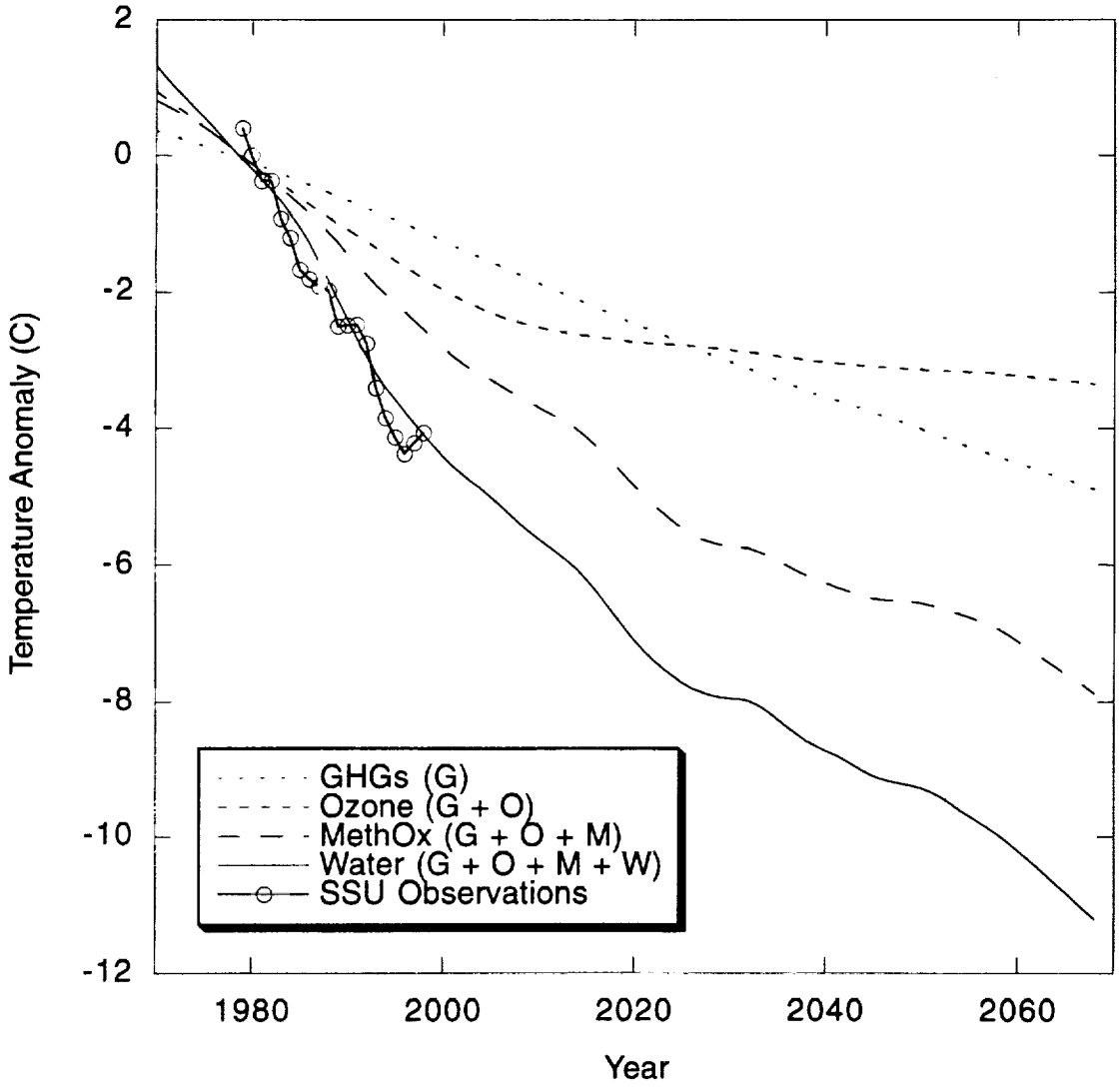
90

FIG 2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Temperature Anomaly (C)
1980
2000
2020
2040
2060
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GHGs (G)
Ozone (G + O)
MethOx (G + O + M)
Water (G + O + M + W)
SSU Observations

FIG 3

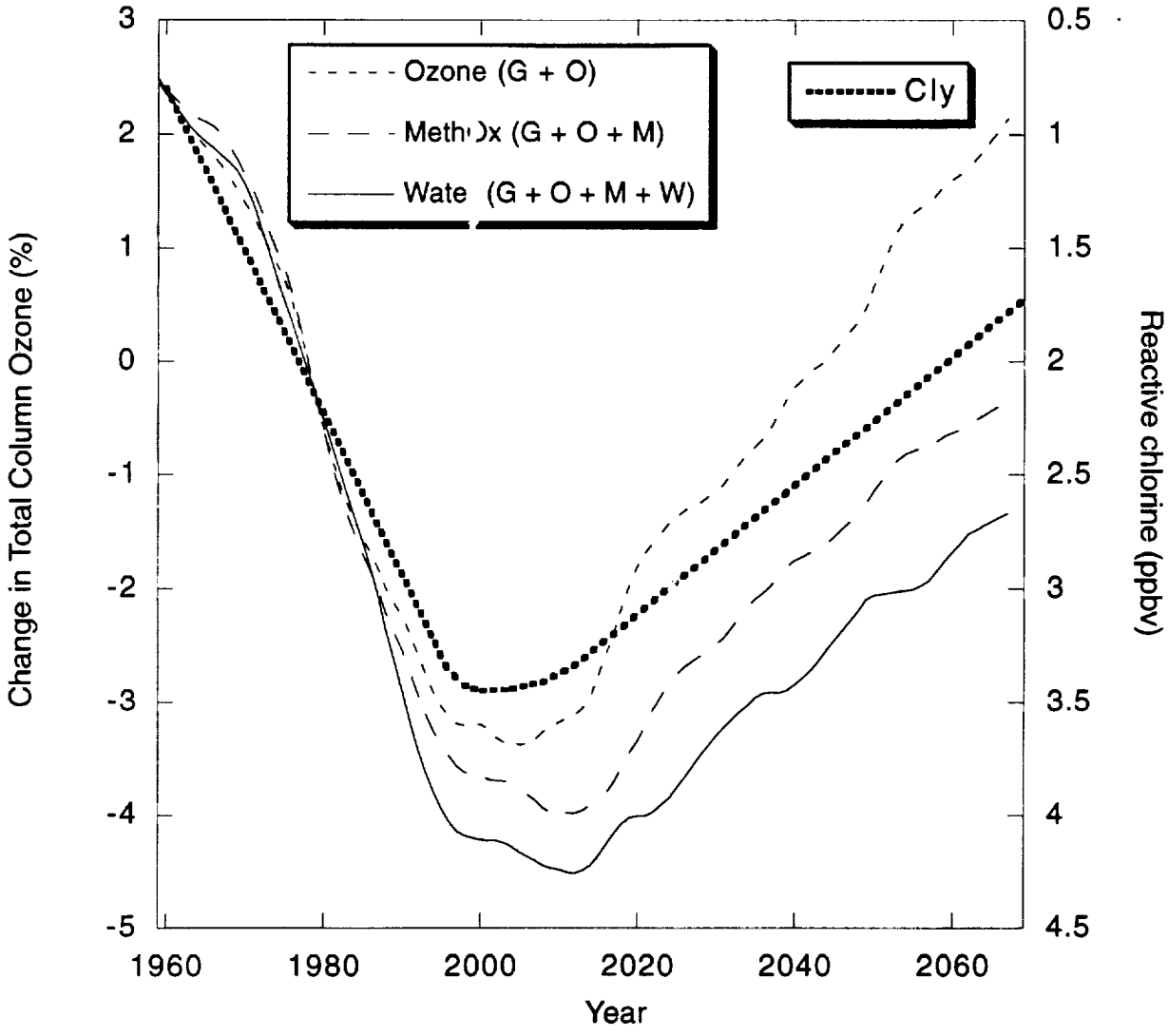
Ozone (G + O)
Meth Ox (G + O + M)
Wate (G + O + M + W)
Cly
Change in Total Column Ozone (%)
Reactive chlorine (ppbv)
Year
3
2
1
0
-1
-2
-3
-4
-5
0.5
1
1.5
2
2.5
3
3.5
4
4.5
1960
1980
2000
2020
2040
2060

FIG 4