

Separating the Influence of Halogen and Climate Changes on Ozone Recovery in the Upper Stratosphere

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Abstract. While halogens from chlorofluorocarbons (CFCs) dominated past ozone changes, our simulations show climate change playing an increasingly important role over coming decades. Including potential climate-induced stratospheric water vapor increases, the ozone change relative to 1980 attributable to climate change surpasses that due to halogens in the 2030s for both the upper stratosphere and the total column. Its overall impact may delay recovery of total ozone to 1980 values. These results emphasize that ozone recovery is not only a detection issue, but a detection and attribution issue, as trends must be ascribed to climate and halogens. Exploiting their different spatial patterns, we demonstrate a technique to isolate the halogen signature in the upper stratosphere by contrasting equatorial and mid-latitude ozone trends at 40 km altitude. This may allow early detection of the effectiveness of halogen emission reductions, which would show that a delayed recovery does not represent a CFC treaty failure, but rather highlights the growing influence of greenhouse gases on the Earth's atmosphere.

1. Introduction

Stratospheric ozone shields the Earth's surface from biologically harmful ultraviolet (UV) light, and acts as a greenhouse gas. Concern over ozone depletion has led to an international agreement limiting emissions of halogen-containing CFCs. These halogens catalytically destroy ozone, and have been largely responsible for the ozone destruction observed since the 1970s. Decreased halogen abundance will therefore lead to a recovery from this depletion. At the same time, however, stratospheric ozone is also influenced by climate change. Yet in contrast to the global warming case, in which studies have focused on detection and attribution of change [e.g., *Santer et al.*, 1993; *Hegerl et al.*, 1996], ozone trend studies have largely focused on trends induced by halogens, placing relatively little emphasis on the separate attribution of non-halogen influences. While much useful work has been done to determine the optimal locations to detect ozone recovery [*Hofmann et al.*, 1997; *World Meteorological Organization*, 1999 (hereinafter referred to as WMO99); *Weatherhead et al.*, 2000], ozone increases alone will not demonstrate recovery from halogen-induced destruction. We show here that climate change is likely to play a major role in future ozone changes, and so must be taken into account in predictions of future ozone amounts, in defining future measurement priorities, and in attribution of any detected recovery. The treaty to protect the

ozone layer from halogen-induced depletion, often hailed as a singular triumph of international environmental cooperation, required great scientific and policy making efforts, followed by economic and industrial adjustments in order for implementation. Adequate verification of its effectiveness is an important final step in this process.

The ozone response to climate change is complex, encompassing the many direct and indirect ways in which greenhouse gases affect the atmosphere. Greenhouse gas increases cool the stratosphere [*Fels et al.*, 1980; *Boville*, 1986; *Rind et al.*, 1990], slowing down the reaction rates of ozone depletion chemistry, and therefore indirectly leading to more ozone. Increases in methane directly affect ozone chemistry by removing chlorine from reactive, ozone depleting species into stable molecules, especially via the reaction $\text{CH}_4 + \text{Cl} \rightarrow \text{HCl} + \text{CH}_3$, again leading to ozone increases. Increased methane also causes ozone destruction by enhancing production of water vapor via methane oxidation. Additionally, changes in circulation, tropospheric water vapor amounts, or in the temperature of the tropopause may affect the amount of water vapor entering the stratosphere, potentially leading to large increases, as have recently been observed [*Oltmans and Hofmann*, 1995; *Nedoluha et al.*, 1998; *Evans et al.*, 1998; *Randel et al.*, 1999; *Oltmans et al.*, 2000]. This would affect both ozone chemistry directly, and also alter local temperatures (as well as surface climate [*Forster and Shine*, 1999; *Shindell*, 2001]). While the exact strength of climate change's impact on the stratosphere is uncertain, it is clear from temperature observations that an impact is already present [WMO99].

The obvious places to look for ozone recovery are in the upper or lower stratosphere, where halogen chemistry is most effective and therefore ozone depletion over recent decades has been greatest [WMO99; *Newchurch et al.*, 2000]. Recovery from halogen-induced depletion will likely take place at both levels over coming years. Definitive attribution of lower stratospheric ozone increases to CFC limitations will be very difficult, however, for several reasons. Firstly, there are large disagreements between ozone trends observed by different techniques in that region [WMO99], so that determining what is actually occurring there is difficult. Secondly, the processes behind ozone changes in this region are not fully understood, as models that include all the known chemical, radiative and dynamical processes still have trouble reproducing observations [*Dvortsov et al.*, 1999]. Thirdly, there are large uncertainties in the future behavior of key processes such as atmospheric circulation and heterogeneous chemistry, which is extremely sensitive to temperature changes. Since heterogeneous chemistry has a large impact on lower stratospheric ozone, future volcanic eruptions that inject aerosol into the stratosphere could affect ozone trends, further complicating attribution in that region.

The upper stratosphere should therefore be the best place to unravel the relative importance of each of the underlying factors controlling ozone's behavior during the anticipated recovery. The depletion there has been just as large as in the lower

stratosphere (in percentage), but the processes controlling ozone abundances are much simpler and the natural variability is much less. Neither heterogeneous chemistry nor ozone transport are important at higher altitudes. Current models are able to reproduce the trends observed there [WMO99; Shindell, 2001], and agree reasonably well with one another, showing that the chemistry is fairly well understood in that region. Excluding the high latitudes, where uncertain dynamical trends can have a large effect, models also agree quite well with one another in their simulations of climate change impacts in this region (as these are driven primarily by the well-understood radiative response to greenhouse gas increases). The upper stratosphere outside the polar regions is thus an ideal region for attribution studies, as the signal is large, while variability and model uncertainties are minimized. Used along with models, data acquired by long-term monitoring of this region by both satellite and ground-based instruments should therefore be useful for determining the causes of ozone changes over the coming decades. Demonstrating that climate change has a significant effect on ozone in the upper stratosphere would imply that the lower stratosphere is likely subject to sizeable impacts as well, since the physical processes that are important in the upper stratosphere are active in at lower altitudes as well.

2. Model Simulations

The model used here is a version of the Goddard Institute for Space Studies (GISS) climate model with $8^\circ \times 10^\circ$ horizontal resolution, and 23 vertical layers extending into the mesosphere (0.002 mb, ~ 85 km) [Rind *et al.*, 1988a; 1988b]. The model includes an interactively calculated stratospheric ozone response to changes in temperature, radiation, and the abundance of methane, water vapor, nitrous oxide, and halogens. Ozone responses are based on two-dimensional (2D) model-derived chemistry parameterizations [Shindell *et al.*, 1998b], calculated off-line due to computational resource limitations. The 2D model includes detailed stratospheric photochemistry, with 39 molecular species interacting via 131 reactions [Shindell and de Zafra, 1996], with updated reaction rates [DeMore *et al.*, 1997]. Sensitivity functions for ozone chemical responses were stored in two-dimensional (latitude \times altitude) look-up tables. Since the perturbations are not necessarily independent, we also investigated various combinations of changes. One particularly interesting feature was that as the concentration of chlorine diminished, the direct chemical influence of methane began to reverse itself. At a chlorine loading of one-fifth the current level (near background, non-anthropogenic conditions), a 10% increase in methane led to ozone losses down to ~ 45 km (2 mb), in contrast to the present day situation, in which greater methane abundances cause ozone increases by sequestering reactive chlorine. Ozone's response to methane perturbations therefore included a chlorine dependence as well (a three-dimensional look-up table). Changes over the next several decades were not so large as to require additional second-order dependencies. This is somewhat fortuitous, as chlorine is changing more rapidly than any other species important to ozone depletion, and its abundances are

projected to return to their mid-1980s level in 2040. The system therefore never changes greatly from the 1985 conditions used in deriving the response functions, and calculation of ozone perturbations works quite well. This will not necessarily be true in the future, however, as chlorine levels decline toward their natural amounts while greenhouse gases continue to increase. The independent ozone responses are then added together in the climate model to form an “ozone equation” describing the abundance of ozone as a function of temperature (T, which is affected by both greenhouse gases and ozone itself), radiative flux (Θ), halogen, methane, water vapor, and nitrous oxide abundances:

$$\frac{dO_3}{dt} = A \times \frac{dT(\text{GHG})}{dt} + A \times \frac{dT(O_3)}{dt} + B \times \frac{d\Theta}{dt} + C \times \frac{dCl}{dt} + D(Cl_y) \times \frac{dCH_4}{dt} + E \times \frac{dH_2O}{dt} + F \times \frac{dN_2O}{dt}$$

Responses to changes in constituent amounts were calculated by running the 2D model without allowing any ozone changes to feedback on temperature or radiation, so that the calculated changes strictly result from the abundance variations. The sensitivity of ozone to temperature or constituent changes in the GISS 2D photochemical model is quite similar to other 2D models, based on the intercomparison shown in [WMO99].

Ozone trends have been simulated with the chemistry-climate model from 1970 through 2069, including the effects of both halogen and greenhouse gas changes. Constituent trends are based on observations in the past, and on projections for greenhouse gas emissions following the IS92a scenario [*International Panel on Climate Change*, 1995] and for chlorine [WMO99]. Projected bromine trends are included as well, though this is primarily important in the lower stratosphere, so we refer to the total halogen trends as chlorine. These trends are similar to those used previously [*Shindell et al.*, 1998a]. Based on these trends, the ozone chemistry parameterizations enable the climate model to reproduce zonal mean stratospheric ozone depletion during the 1980s and 1990s in good agreement with observations [*Shindell*, 2001]. Ozone transport changes are not included interactively in these simulations, but are calculated afterwards. This should have a minimal impact on our results, however, as our focus is on the upper stratosphere, where transport is not important, and our calculated ozone does interact with the radiation, and hence the dynamics in the climate model. Indeed, postprocessing calculation of ozone changes resulting from altered transport shows that they are unimportant in the extrapolar upper stratosphere.

Other studies have revealed many potential interactions between climate change and polar ozone depletion which are relatively uncertain [*Austin et al.*, 1992; *Shindell et al.*, 1998a; *Dameris et al.*, 1998; *Kirk-Davidoff et al.*, 1999; *Tabazedeh et al.*, 2000], so we concentrate here on regions outside the polar areas (in contrast to studies such as *Hofmann et al.* [1997]). As noted above, variability is much less in the extrapolar atmosphere. Furthermore, severe ozone losses in the polar regions are caused by unique chemistry which is extremely sensitive to temperature and water vapor abundances. Definitive attribution of the impacts of either climate change or CFC limitations will therefore likely occur somewhat later there

[Weatherhead *et al.*, 2000]. In fact, predicted trends are not consistent among models even in the relatively stable Antarctic region [Shindell *et al.*, 1998a; Austin *et al.*, 2001]. We have therefore continued to use our representation of polar heterogeneous chemistry based on a simple on-off threshold at 195 K [Shindell *et al.*, 1998a], though this is too simplistic to account for many important processes that may be impacted by climate.

A large uncertainty in projecting climate change's impact on the stratosphere is predicting what will happen to stratospheric water vapor. Though observational records are extremely limited in spatial coverage or duration, they suggest that stratospheric water vapor has been increasing. Seven years of global satellite observations of the middle and upper stratosphere, supported by coincident ground-based data [Nedoluha *et al.*, 1998; Randel *et al.*, 1999], and lower stratospheric measurements over Boulder, Colorado and Washington, DC covering a period of 37 years [Oltmans and Hofmann, 1995; Oltmans *et al.*, 2000] all show very large increases, in the range of 8-20% per decade. A summary analysis of ten data sets taken by five types of instruments which have reported observations of lower stratospheric water vapor during 1954-2000 shows an overall increase of 1% yr⁻¹, [Rosenlof *et al.*, 2001]. Further supporting this apparent upward trend, we have shown that the very large, well-observed, long-term upper stratospheric cooling can only be reproduced in our model by including an increase in water vapor [Shindell, 2001]. Though the evidence indicates the likelihood of an increase over the coming decades, such a trend must still be regarded as speculative compared with the better constrained surface emissions of greenhouse gases and CFCs. Two simulations are therefore discussed here. Both include projected trends in greenhouse gases and halogens, but one has fixed water vapor amounts, while the other includes a gradual increase in water vapor. The water vapor trend is created by letting the model calculate changes from methane oxidation and those arising from cross-tropopause transport, which is enhanced in the model by the increasing greenhouse gases, which induce a more rapid circulation and warm the tropopause [Shindell, 2001]. This yields an increase at 15 mb of 40 ppbv/yr at 43° N, in good agreement with the 44 ppbv/yr at 40° N seen in several decades of observations [Oltmans *et al.*, 2000]. Averaged over 60° S to 60° N, the model's increase is 50-80 ppbv/yr over the altitude range 30-50 km, on the conservative side of the 60-115 ppbv/yr seen in five years of satellite observations taken at these altitudes [Randel *et al.*, 1999]. Both the model and the satellite data (as well as ground-based observations [Nedoluha *et al.*, 1998]) show a faster rate of water vapor increases at higher altitudes due to methane oxidation. We concentrate here on the increasing water run, as the most likely scenario, but include results from both of these simulations.

In summary, we ran two long-term transient simulations including trends in both greenhouse gases and halogens, one with increasing water vapor and one without. In addition to these, two sensitivity simulations were performed leaving out greenhouse gas trends in one, and halogen trends in the other, both without water vapor increases. These provide control runs for each forcing. The parameterized

ozone chemistry was included in all four simulations.

3. Results

The progression from halogen to climate control is clear in the simulations. Figure 1 shows modeled ozone changes for 1996 and 2036 relative to 1979, when satellite observations of ozone began, in the simulation with increasing water vapor. We define ozone recovery as a return to the satellite values observed in 1979, although limited ozone depletion likely occurred prior to the onset of global observations. In the upper stratosphere, strong depletion takes place during the 1980s and 1990s as chlorine abundances grow. By 2036, the depletion there has been replaced by an increase in ozone outside the polar regions. Examination of the terms in the “ozone equation” describing ozone’s response to perturbations reveals the relative influence of each individual forcing. Figure 2 shows the two dominant terms for the simulation with increasing water vapor: ozone changes due to chlorine and to temperature differences arising from greenhouse gas increases. The enhanced ozone in 2036 is clearly due to a combination of decreased chlorine and cooling from increasing greenhouse gases. Note that an increase in meridional circulation in the model leads to heating in the upper stratosphere at high northern latitudes, resulting in continued net ozone decreases there.

Details for the mid-latitude upper stratosphere are shown in Figures 3a and 3b. We give Northern Hemisphere values, but the two hemispheres are in fact quite similar. Through 2020, chlorine has by far the largest impact on ozone (Figure 3a). Note that the total change includes the ozone response to temperature changes induced by ozone changes themselves, a negative feedback by which ozone depletion leads to colder temperatures and therefore reduced chemical ozone destruction. Roughly 20% of the ozone change induced by each individual factor is compensated for by this mechanism. While chlorine catalyzed depletion, damped by the ozone-temperature feedback, accounts for nearly all the twentieth century trend, this is not the case in the twenty-first century. Initial recovery during the current decade is almost exclusively due to climate change, as chlorine amounts hold relatively steady (Figure 3b). In subsequent decades, chlorine decreases and climate change combine to increase ozone amounts. By the 2030s, the impact of stratospheric cooling has surpassed that of chlorine (Figure 3a), so that there is more ozone than in 1979 despite significant chlorine-induced depletion. Overall, the chlorine signal diminishes during the twenty-first century, approaching zero in the 2040s (Figure 3a), while the climate change signal grows.

At around 30 km and at 50km and above, regions where halogens have less influence on ozone (Figure 2), the effect of climate change is even larger relative to that of chlorine. In the uppermost stratosphere, where water vapor induces large chemical ozone losses, the simulation with increasing water vapor shows continued ozone decreases through at least 2050. In the lower stratosphere, which dominates the total column, both temperature changes and water vapor increases lead to significant ozone depletion, in addition to that caused by chlorine. Though

uncertainties are large in this region, as noted above, several similar conclusions can be drawn regarding the relative influence of halogens and climate on ozone. As in the upper stratosphere, initial changes in ozone during the present decade arise primarily from climate change. With the strong influence of water vapor on ozone, the total impact from climate may exceed that from halogens by the 2020s, even earlier than in the upper stratosphere. Though the exact values of lower stratospheric ozone depletion remain difficult to model, these chemistry-climate GCM simulations indicate that the total column recovery may perhaps take much longer than it would in the absence of climate change (Figure 3c).

The ozone response to greenhouse gas-induced stratospheric cooling and increasing water vapor is broadly consistent with those of the 2D model of *Evans et al* (1998). That model found ozone increases in response to doubled CO₂ which peaked at around 10-12 percent at altitudes of 2-3 mb, and showed little variation with latitude outside high latitude regions. Our model also finds an ozone increase of 10-12 percent due to greenhouse gases at 2-4 mb for the year 2070, when the model's CO₂ has doubled. Similarly, the impact of water vapor increases is relatively small at these altitudes, but increases substantially in the uppermost stratosphere [*Shindell*, 2001]. In the lower stratosphere, however, there are substantial differences. While the increased water vapor induces a similar enhancement of ozone destruction in the 2D model and the GCM, the 2D model finds that increases in greenhouse gases lead to reduced ozone only from about 25° N to 40° S. In contrast, these decreases extend to mid-latitudes in our simulations. This has a large effect on the total column response, which would be positive in the 2D model at mid-latitudes (as in another 2D model study that shows a faster mid-latitude ozone recovery caused by increasing carbon dioxide [WMO99]). The differences in the lower stratosphere stem from the uncertainty of the temperature response to increasing greenhouse gases in the tropopause region. The radiative influence of greenhouse gases is fairly straightforward: they lead to heating in the troposphere by trapping of outgoing radiation, and cooling in the stratosphere by a net emission of longwave radiation out to space. Dynamics plays a significant role as well, however, as convection keeps even the upper troposphere closely tied to the surface in the tropics. In the GCM, increasing greenhouse gases lead to warming not only in the troposphere, but up into the lower stratosphere as well, leading to ozone decreases there (in contrast to the limited observational record in the tropical tropopause region). This result depends upon changes in the strength of convective activity, the location of the tropopause, and the overall residual circulation. Though a GCM is better able to simulate these changes than a 2D model, the response of these factors to greenhouse gases increases is poorly understood, making this result very uncertain. This once again highlights the value of focusing on the upper stratosphere for detection and attribution of ozone changes.

4. Attribution of Ozone Changes

Given the large projected impact of climate change on ozone, how will we be

able to interpret ozone observations over the coming decades? The effects of reduced stratospheric halogen loading and of increasing greenhouse gases can be separated by exploiting the differing spatial patterns of their impacts on ozone (as a function of latitude and altitude). Model simulations are able to isolate these two patterns (Figure 2). While there are many alternatives, we propose the following comparison of ozone's behavior as a function of latitude, altitude, and time as ideal.

To isolate halogen impacts, we focus on 40 km altitude (~ 3 mb), in the area where they have been largest (Figure 2), where ozone depletion has also been very large over recent decades (in percentage), and where both observations and model data are available. There are marked differences between ozone's response to halogens at the equator and at mid-latitudes at this level. Note that since ozone's chemical lifetime is very short in the upper stratosphere, abundances there are determined almost exclusively by local conditions. The strong latitudinal dependence of chlorine catalyzed ozone destruction is due primarily to the greater fraction of chlorine present in reactive forms at higher latitudes [*Waters et al.*, 1999]. This dependence is quite different from that of other factors that influence ozone. The cooling at this altitude due to increasing greenhouse gases varies only weakly with latitude, a robust result across various GCMs [WMO99], leading to relatively uniform ozone increases (except in the polar regions, e.g., Figure 2). Methane's impact on ozone also has a meridional dependence, since it operates partially by sequestering chlorine. However, this merely damps out a small portion of the halogen-induced latitudinal variation. Combined with the fact that methane's overall influence is quite small (Figure 3), it has a negligible effect on isolation of the halogen signal in the upper stratosphere. Modeled water vapor changes also have a fairly uniform effect with latitude [*Shindell*, 2001] (though the spatial distribution of water vapor trends is not yet well characterized in observations). Solar variability can affect ozone in the upper stratosphere, but it too has only a fairly weak ($\sim 1\%$) latitudinal dependence between the equator and middle latitudes at around 40 km [*McCormack and Hood*, 1996]. Thus the meridional difference in ozone depletion rates at this level reflects almost exclusively the influence of halogens.

Modeled ozone depletion at 40 km due to chlorine has proceeded at a 30-40 percent faster rate at mid-latitudes than at the equator, where the halogen-induced loss is a minimum (Figure 2). Satellite data show a similar pattern [WMO99]. Greenhouse cooling of the stratosphere has increased ozone at this altitude, canceling out a fraction of the halogen-catalyzed ozone depletion. The net result is a present day overall mid-latitude ozone depletion 50-60 percent greater than at the equator (Figure 1). Since the largest effect of halogens is at around $50-75^\circ$ in each hemisphere, we compare ozone trends at 0 and 50 degrees, which maximizes the signal strength while avoiding the influence of uncertain dynamical changes in the high latitude regions.

The rates of simulated ozone depletion agree quite well with satellite observations at both 50°N and the equator (Figure 4, top two panels). We use data

from the Stratospheric Aerosol and Gas Experiment (SAGE) II satellite. This data set provides the longest available record, 1984 to 1999, from a single instrument, eliminating problems associated with inter-instrument calibration. While these measurements may nevertheless be subject to drift, an extensive comparison with multiple independent data sets, including those from ground-based lidars and umkehrs, as well as three other satellites, shows no statistically significant biases between trends in the upper stratosphere [*World Meteorological Organization, 1998; WMO99*].

5. The Halogen Signature in Models and Observations

The halogen signature, given by the difference between 40 km ozone depletion rates at 50N and the equator, agrees well with SAGE measurements (Figure 4, bottom panel). The positive slope during the 20th century shows the more rapid depletion of ozone by halogens at mid-latitudes. As halogen abundances decrease during the 21st century, the difference between depletion at mid-latitudes and the equator will also decrease, reflected in a turnover of the rate of change of this quantity (negative slope in bottom panel of Figure 4). The removal of the climate signal is demonstrated by two test runs we performed. A run with ozone changes due only to halogens is indistinguishable from the runs with ozone response to both increasing greenhouse gases and halogens, while a run with no halogen changes shows no trend in the ozone difference (Figure 4, bottom panel). This meridional difference at 40 km has therefore effectively isolated the halogen signal.

The trend values calculated with a least-squares regression and their 95% confidence intervals are shown in Table 1. The increase during the last three decades of the twentieth century is significant in all the simulations with increasing halogens, regardless of the ghg and water vapor forcing applied, but not in the run without halogen changes. Clearly, neither the model's internal variability nor differences between model runs prevent retrieval of significant trends. During the period of SAGE II data, these three simulations agree with the observed changes to within their 95% confidence levels. Note, however, that this time period is so short that trends are subject to large changes with different sampling. For example, the trend in the SAGE II data from 1984 through 1994 was $0.92\% \text{ yr}^{-1}$ (0.78 to $1.05\% \text{ yr}^{-1}$ 95% confidence interval). Thus the quantitative value of the slopes determined over this brief time period should be judged with caution, as they will depend on the precise interval chosen (perhaps the slowdown in the trend during the latter half of the 1990s already signals the slower rate of increase in halogen loading). Nevertheless, the data and the model both show statistically significant increases during the past, followed by statistically significant modeled decreases in the future. While the model statistics are based upon a single simulation for each scenario, as opposed to the ideal (but computationally expensive) ensemble of simulations for each set of conditions, the similarity between the three simulations with projected halogen loading shown in Table 1 lends confidence to our results. Furthermore, ozone's behavior in the extrapolar upper stratosphere depends

primarily on radiatively induced cooling (and of course the prescribed composition changes), which is a very robust feature across model runs, especially compared with dynamical changes.

Model trends in the twenty-first century become statistically different from the twentieth century trends in 2013, 2015, and 2018 for the three simulations with varying halogens, so it may require one to two decades for the measurements to reveal a signal over the variability. Further short-term excursions such as that seen in the 1992 SAGE data could extend this time frame, though the model's decadal variability is actually quite similar to that seen in observations (Table 1). Ozone observations over the coming decades should therefore be able to discriminate between the increasing slope of the twentieth century and the decreasing slope of the twenty-first. Such an observation of a change in the slope of this difference in ozone depletion rates will be a clear indication that recovery from halogen-induced depletion is indeed taking place, and demonstrate the success of the international protocols limiting emissions of substances that deplete the ozone layer.

6. Conclusion

The behavior of stratospheric ozone over coming decades will thus be governed by both halogen abundances and climate change. Taking into account limitations in current observational capacity and understanding of relevant processes, long-term monitoring of upper stratospheric ozone should yield the best information on the effects of each factor individually, and should be given a high priority along with measurements of total column ozone. Observations of upper stratospheric temperatures and halogens would also be useful, and could be used with models to ascribe ozone changes to various factors as a cross-validation of this pattern recognition technique. The advantage of this method is its dependence upon only a single data set, ozone observations (which typically can be validated more easily against independent data than other quantities), instead of relying upon contemporaneous and co-located measurements of temperature and halogen loading from other instruments.

Using spatial patterns in the upper stratosphere is quite robust to uncertainties in future atmospheric composition as well. Even a large volcanic eruption that greatly perturbs lower stratospheric ozone should have minimal effect on the upper stratosphere. Unexpected variations in future halogen loading or rates of greenhouse gas emissions would also not affect the suitability of this technique, as it is designed to separate these factors regardless of their magnitude. Significant uncertainty is also associated with future rates of stratospheric water vapor change, leading to uncertainty in the projected rate of ozone change. Even in the case of a large water vapor increase that leads to a delay in ozone increases, however, the identification techniques discussed here will be able to reveal an ozone recovery from halogen catalyzed losses. This will allow us to verify the effectiveness of the Montreal Protocol's limitations on substances that deplete the ozone layer, and also to improve our understanding of the magnitude of climate change's influence on the

ozone layer. The latter is important not only because of ozone's role as a UV shield for the Earth, but also because ozone is itself a greenhouse gas. As stratospheric ozone recovers, it will change from a negative climate forcing to a positive one [Solomon and Daniel, 1996]. The rate of this change will depend on the magnitude of climate change's impact on the stratosphere, and on the relative importance of halogen-induced and greenhouse gas-induced changes. Accurate prediction of future climate therefore requires an understanding of all the factors influencing ozone amounts.

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Figure 1. Modeled zonal mean ozone changes. Percentage changes are given for the 5 year averages centered around 1996 and 2036 relative to 1979 for the simulation with increasing water vapor. Tropospheric ozone was not changed in these runs. Calculations for the lower stratosphere are highly uncertain, given the difficulties models have in this region [Dvortsov *et al.*, 1999] and the lack of interactive ozone transport in these climate model simulations.

Figure 2. Individual components of modeled ozone changes. The ozone response to greenhouse gas induced temperature changes and to chlorine changes (bromine plus chlorine, gas-phase only) are shown for 1996 (top row) and 2036 (bottom row) relative to 1979, as in Figure 1, again for the simulation with increasing water vapor.

Figure 3. Ozone changes from individual forcings in the simulation with increasing water vapor. (a) Changes at 40 km altitude due to chemical interactions with chlorine (chlorine plus bromine), methane and water vapor, cooling induced by increasing greenhouse gases, and changes in flux at this altitude resulting from ozone changes at higher altitudes. The last factor is affected by both halogens and greenhouse gases. The total change includes the negative feedback resulting from ozone changes themselves. Curves are based on decadal averages relative to 1979 climatological ozone amounts. Results for the run with fixed water vapor are quite similar, so are omitted for clarity. (b) As in (a), but the net change for each decade relative to the previous decade. (c) As in (a), but for the total ozone column. Ozone changes due to methane are extremely small, so are not shown. Note that the column values are relatively uncertain, and that the model underestimates ozone depletion up to the present relative to observations, a common deficiency in chemical models.

Figure 4. Change of stratospheric ozone at 0.7 mb (40 km altitude) relative to 1979. Trends at 0°N, 50°N, and the difference between the two are shown. Observations from the SAGE satellite and values from model runs with constant and increasing water vapor are presented in each panel. Ozone changes simulated in the model run with increasing greenhouse gases and constant water vapor were used to determine the offset due to ozone changes during the 1979-1984 period before SAGE data became available. The values in 1985 are therefore arbitrary, and only the slopes have meaning. The bottom panel, presenting differences, also includes model runs in which greenhouse gas concentrations were held constant while halogens varied, and the opposite case of constant halogens and increasing greenhouse gases.

Table 1. Ozone trends (% per yr) and statistics at 40 km altitude.

Description of simulation or data	1970-2000 trend	1984-1999 trend	2000-2060 trend
SAGE II observations		0.34 (0.26 - 0.43)	
GISS GCM with halogen and GHG trends	0.20 (0.13 - 0.27)	0.13 (-0.06 - 0.32)	-0.09 (-0.11 - -0.06)
GISS GCM with halogen, GHG, and water vapor trends	0.18 (0.15 - 0.21)	0.17 (0.04 - 0.29)	-0.07 (-0.08 - -0.06)
GISS GCM with halogen trends only	0.20 (0.15 - 0.26)	0.27 (0.10 - 0.44)	-0.09 (-0.12 - -0.06)
GISS GCM with GHG trends only	-0.01 (-0.05 - 0.03)	0.02 (-0.11 - 0.14)	0.01 (0.00 - 0.02)

Trends over the indicated time periods based on the data shown in Figure 4. The values shown in parentheses are the 95% confidence limits on these trends, assuming each point (i.e. each year) is independent, which seems reasonable given the short lifetime of ozone in the upper stratosphere.

FIGURE 1

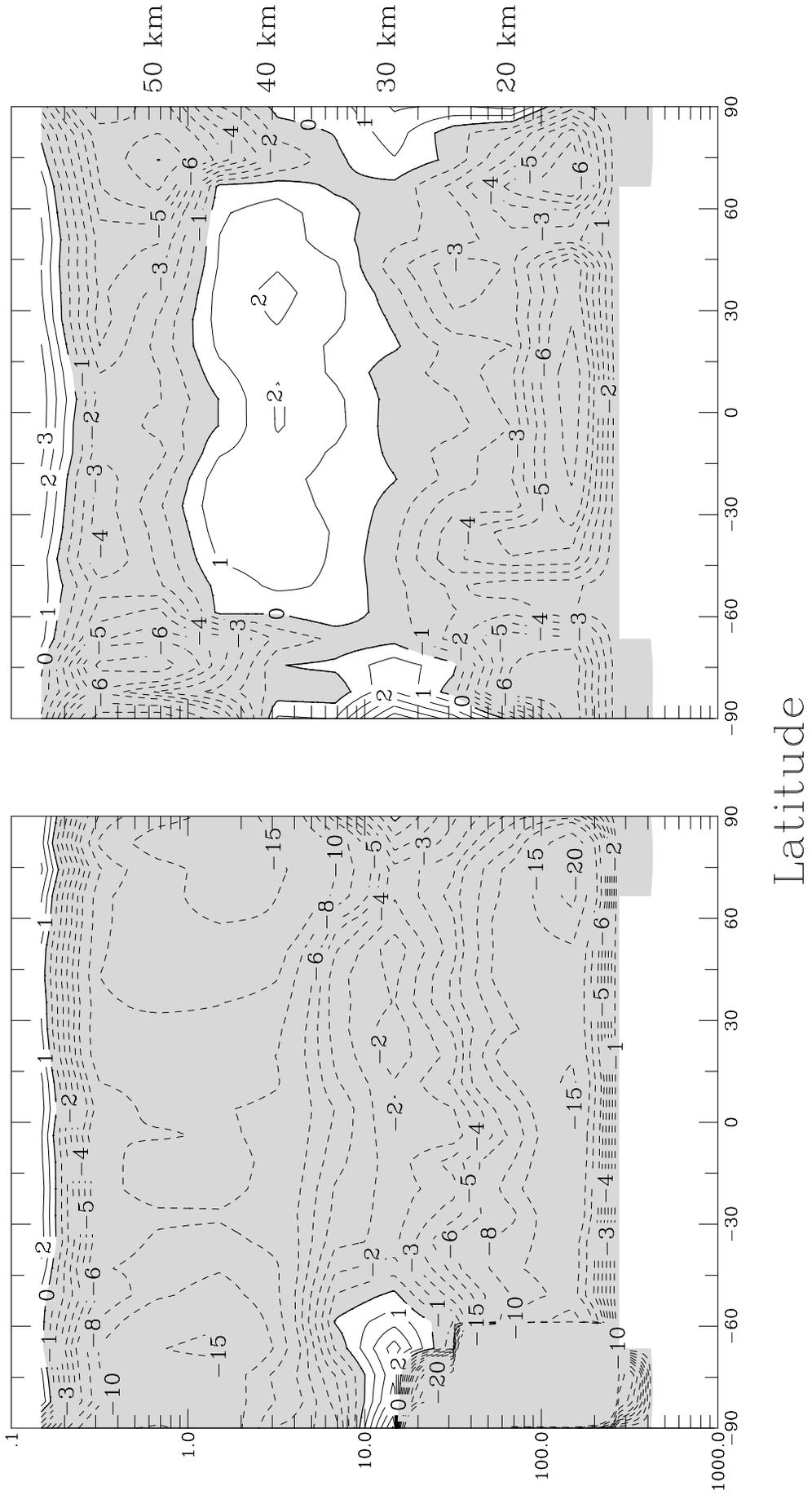
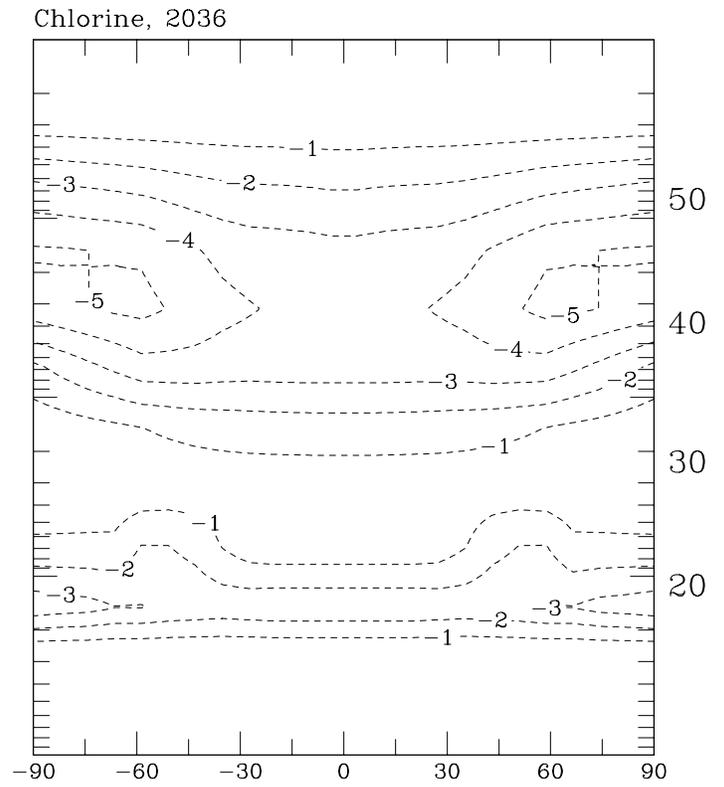
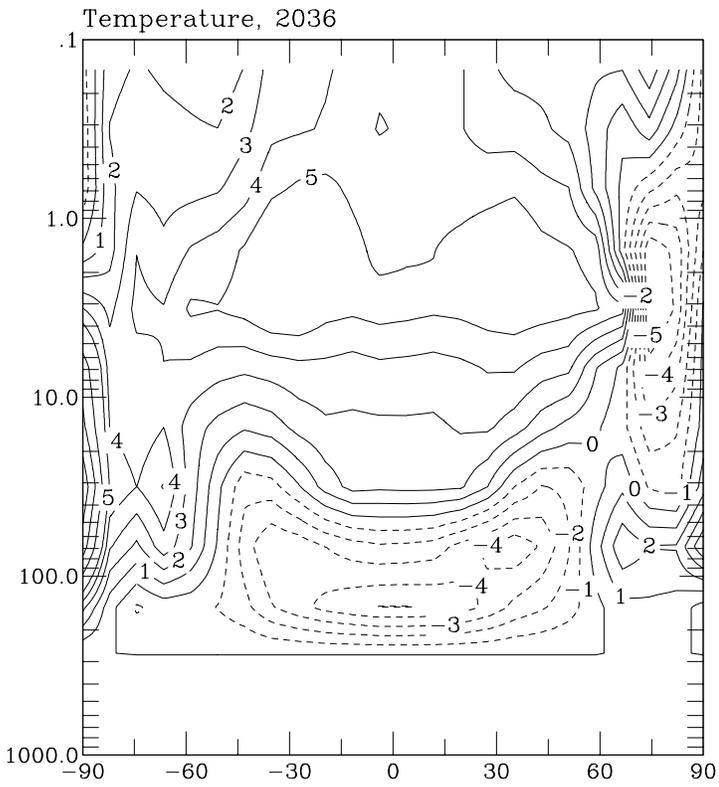
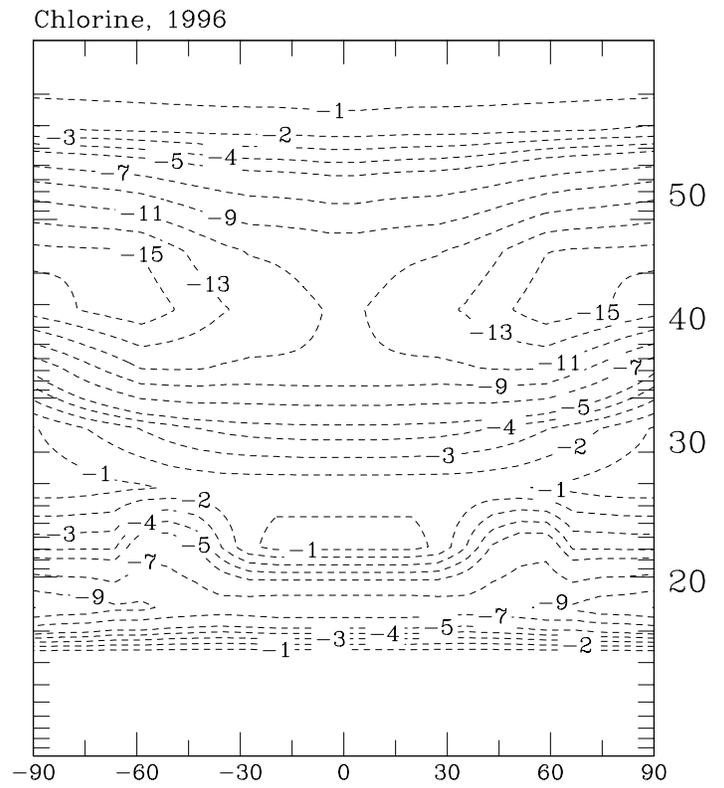
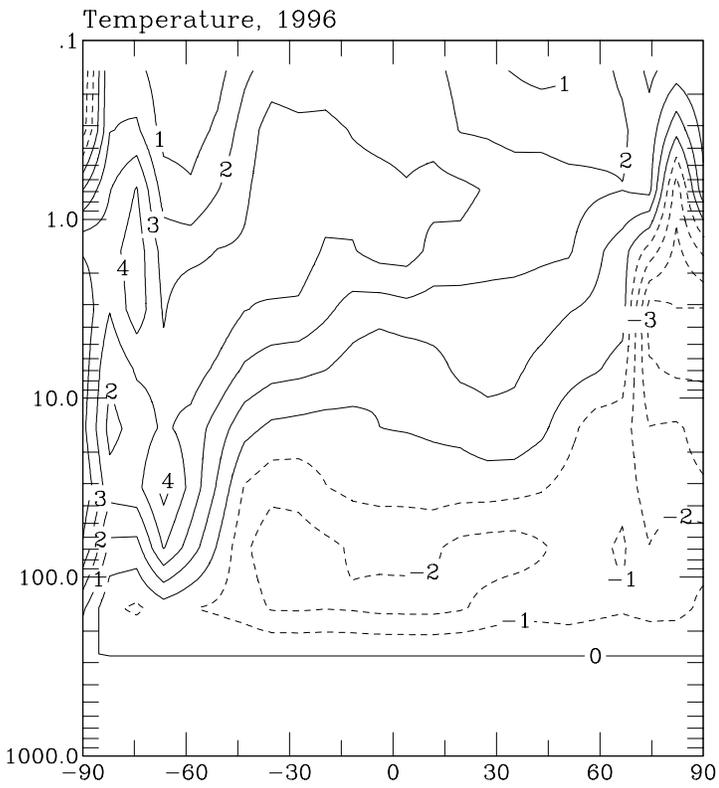


FIGURE 2



Latitude

FIGURE 3

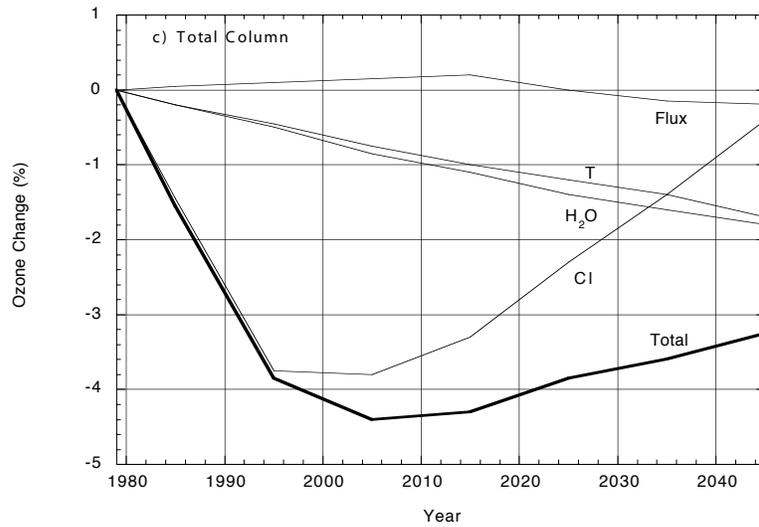
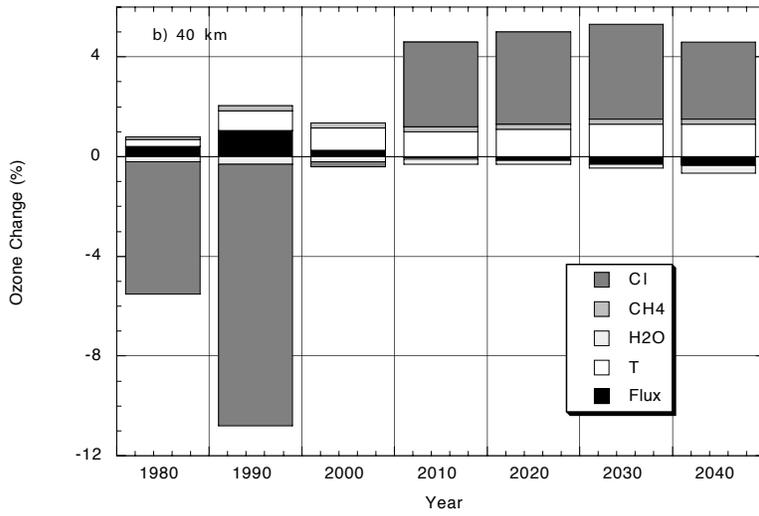
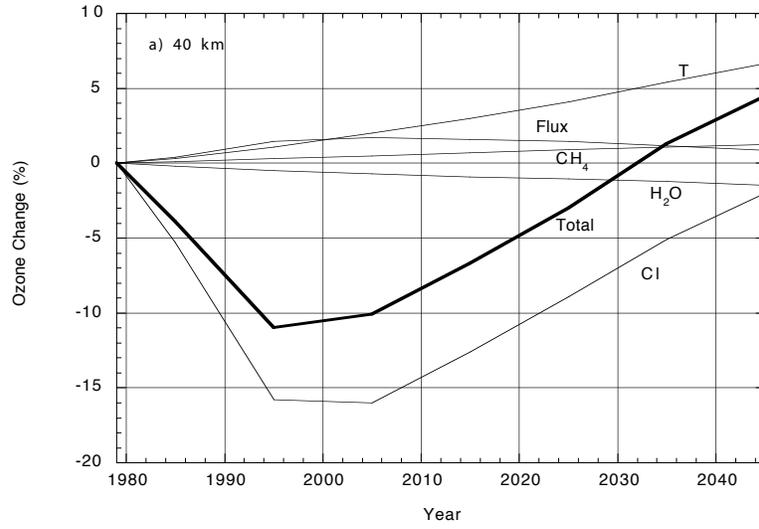


FIGURE 4

