Chapter 4. How Do Climate Change and Stratospheric Ozone Loss Interact?

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KEY ISSUES

Stratospheric ozone abundances are dependent on a balance of chemical processes that both produce and destroy ozone and dynamical processes that transport ozone throughout the stratosphere. The chemical processes depend on atmospheric temperatures and the abundances of ozone-depleting substances and other trace gases, such as water vapor and nitric oxides. Transport depends on heating in the atmosphere, which also depends on the distribution and abundance of trace gases, such as carbon dioxide, ozone-depleting substances, and ozone. Atmospheric temperature, transport, and trace gas amounts, for example, are all aspects of Earth's climate. As these and other climate parameters change as result of human activities and natural variability, ozone abundances will decrease or increase in response in a manner that depends on a variety of factors.

This complex coupling of ozone and climate parameters is not fully defined at present and has significant uncertainties associated with known key aspects. Chemistry climate models (CCMs) of the atmosphere are in development and use by researchers aiming to

reduce the uncertainty in the ozone-climate interaction and to explore other aspects of the interrelationship. Key questions related to the coupling of ozone and climate are:

- How do ozone-depleting substances and ozone depletion contribute to the radiative forcing (RF) of climate?
- How do long-term changes in greenhouse gases affect stratospheric ozone?
- How have stratospheric temperatures changed in recent decades and what is the cause of these changes?
- Is stratospheric water vapor changing in a way that influences ozone abundances?
- How do ozone changes influence the climate of the stratosphere and troposphere?
- Will volcanic eruptions influence future ozone amounts?

KEY FINDINGS

Linking of ozone and climate change

- Ozone and climate change are linked because ozone-depleting substances (ODSs) are also greenhouse gases, which contribute to radiative forcing. The ODS contribution to global radiative forcing between 1750 and 2005 is approximately 20% (0.34 W per m^{-2}) of that from carbon dioxide, the largest anthropogenic contribution. The ODS contribution is expected to decline in coming decades as ODS emissions and their atmospheric abundances continue to decline in the atmosphere.
- Each ODS contributes to ozone depletion and climate warming with different levels of effectiveness as represented, for example, by the ozone depletion potentials (ODP) and global warming potentials (GWP), respectively. For the

principal ODSs, these values vary over orders of magnitude for equal mass of emissions.

- The abundance of stratospheric ozone is dependent on a balance of production and loss processes. These processes are dependent on several features of the atmosphere: namely, its chemical composition, air motions, radiation, and temperatures. Climate change will lead to changes in these features, which in turn will affect ozone abundances. Feedback from climate change has the potential to increase or decrease ozone abundances depending on the region and extent of climate change.
- Chemical ozone depletion also contributes to climate change by modifying atmospheric radiative properties. Feedback from ozone losses can alter atmospheric temperatures and atmospheric transport. Ozone depletion can affect the climate of both the troposphere and stratosphere.

Impact of climate change on ozone

- The complexity of the interactions of ozone changes with climate parameters requires that coupled models of Earth's atmospheric chemistry and climate processes (called chemistry climate models, CCMs) be used to predict future ozone amounts. In addition, CCMs are needed to evaluate the sensitivity of ozone to climate parameters and the response of climate to ozone changes.
- Stratospheric temperatures have decreased in the observational records that begin in the 1960s. This is attributed to ozone depletion, increased carbon dioxide, and changes in water vapor.

- Stratospheric temperatures influence ozone amounts through chemical and transport processes. Future increases in CO_2 will continue to contribute to global stratospheric cooling. The photochemical loss of ozone is slowed in some regions when temperatures decrease with the result that ozone recovery may be accelerated.
- Human activities are expected to increase the future abundances of greenhouse gases that influence stratospheric ozone amounts, principally, carbon dioxide (CO_2), methane (CH_4) and nitrous oxide (N_2O).
- Stratospheric water vapor has increased in recent decades but since 2001 has been decreasing in the lower stratosphere. The oxidation of methane emissions is an important contribution to increasing water vapor trends. Tropical tropopause temperatures modulate dehydration of air entering the stratosphere, and recent decreases in water vapor are well correlated with negative tropical tropopause anomalies. Future trends are uncertain because stratospheric water vapor responds to both atmospheric methane emissions and the temperature of the tropical tropopause. Stratospheric water vapor influences stratospheric ozone through reactive hydrogen chemistry and polar stratospheric cloud formation.
- CCM simulations predict that the atmospheric circulation between the stratosphere and troposphere will increase in a changing climate in the coming decades. If this occurs, the increased circulation will increase the stratospheric flux of ozone to the troposphere.

Impact of changes in stratospheric ozone on climate change

- Depletion of stratospheric ozone since about 1980 has caused a *negative* radiative forcing of climate change (~0.05 W per m^2) because ozone is a greenhouse gas. The short-wave cooling of the lower stratosphere has in turn led to a long-wave cooling of the upper troposphere. Increases in ozone from pollution chemistry have increased average tropospheric ozone, causing a *positive* radiative forcing (~0.35 W per m^2).
- Ozone depletion causes changes to the temperature and circulation of the stratosphere and troposphere. Meteorological analyses indicate that stratospheric ozone depletion over Antarctica has caused strengthened circumpolar flow throughout the troposphere over Antarctica and caused surface temperature changes.

Importance of volcanic eruptions

- If explosive volcanic eruptions occur again in the coming decades, they will decrease stratospheric ozone levels for several years as a result of the heterogeneous reactions occurring on volcanic sulfate aerosols. These reactions enhance halogen loss processes by reducing the abundance of key reactive nitrogen compounds. For a given size eruption, the resulting effect on ozone will decrease as halogen amounts decrease in the coming decades.
- Explosive volcanic eruptions are expected to cause major temperature and circulation changes in the stratosphere as have occurred after past eruptions. These changes are a feedback response to the large increases in sulfate aerosol amounts in the stratosphere following such eruptions. The increases result in only

a short-term shift in stratospheric climate because natural processes remove the additional sulfate aerosols within 2-3 years after the eruption.

4.1 INTRODUCTION

Ozone occurs naturally in the atmosphere as a result of photochemical processes. In the stratosphere ozone is beneficial to life on Earth because it absorbs ultraviolet radiation from the sun. Ultraviolet light absorption heats the stratosphere. Ozone is also a greenhouse gas that helps trap terrestrial infrared radiation, which leads to heating of the troposphere and stratosphere. In the natural atmosphere, ozone's warming of the planet makes it the third most important greenhouse gas after water vapor and carbon dioxide (Kiehl and Trenberth, 1997). As a consequence, changes in ozone amounts have the potential to change climate parameters in the stratosphere and troposphere. Anthropogenic pollution has led to increased ozone production and abundances in the troposphere, particularly near Earth's surface. In contrast, emissions of ozone-depleting substances (ODSs) in recent decades have led to significant depletion of global stratospheric ozone with particularly high losses in polar regions. The Montreal Protocol has been established to protect the ozone layer by reducing the global production and consumption of ODSs.

The complex interrelationship between ozone and climate change is illustrated in Figure 4.1. Multiple radiative, chemical, and dynamical processes control ozone amounts and their distribution in the troposphere and stratosphere. Production and loss cycles of ozone involve many chemical species, as well as aerosols, and are influenced by atmospheric

parameters such as solar insolation and temperature. Natural and anthropogenic processes on Earth's surface influence ozone through changes caused in atmospheric composition and climate parameters. The chemical loss rate of ozone leads to an atmospheric lifetime that is relatively short compared to carbon dioxide, for example. As a result, the distribution of ozone is non-uniform in the atmosphere and dynamical processes such as planetary waves and the Brewer-Dobson circulation have important roles in establishing the distribution.

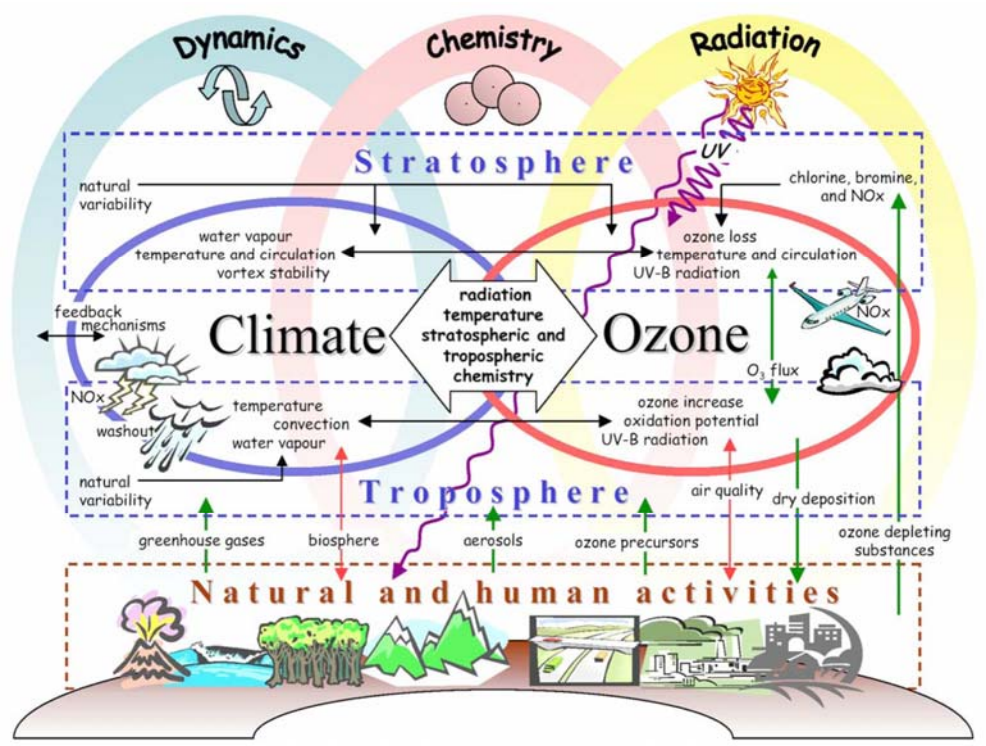


Figure 4.1 Schematic depiction of the processes that link climate change and ozone abundances (Isaksen, I.S.A., 2003).

The ODSs that destroy ozone are also greenhouse gases. Thus, the radiative effect of accumulated ODS emissions is partially offset by the subsequent reduction in global

ozone amounts. Systematic and long-term ozone depletion can change atmospheric circulation patterns and contribute to climate change. Furthermore, changes in climate can potentially alter ozone amounts. Changes in temperature, amounts of trace gases such as methane, nitrous oxide, and water vapor, and atmospheric circulation can all potentially lead to ozone changes in the stratosphere and troposphere. Finally, large volcanic emissions can alter both ozone and climate for temporary periods of several years. This chapter assesses these interconnections, schematically shown in Figure 4.2, by outlining what is known about ODS and volcanic emissions and the processes through which ozone influences climate and through which climate change will influence ozone amounts. Further detail on the coupling of climate and ozone changes can be found in recent scientific assessments (IPCC/TEAP, 2005; WMO, 2007).

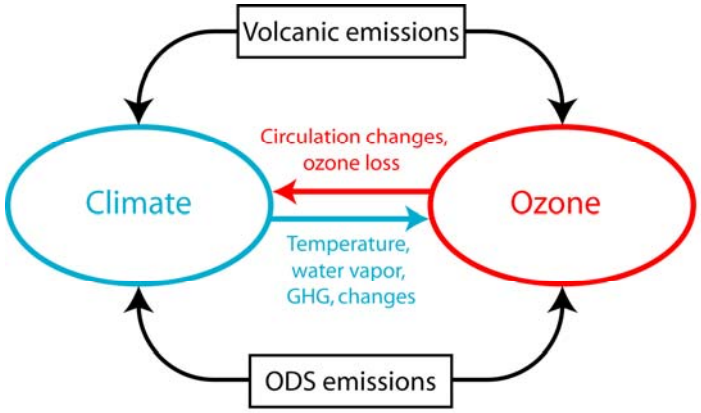


Figure 4.2 Schematic of specific processes that interconnect and influence atmospheric ozone amounts and climate parameters and that are addressed in this assessment.

4.2 RADIATIVE FORCING OF CLIMATE BY OZONE-DEPLETING SUBSTANCES AND OZONE CHANGES

4.2.1 Radiative Forcing by Ozone-Depleting Substances

The accumulation of ozone-depleting substances in the atmosphere has contributed to the

radiative forcing of climate because all ODSs are greenhouse gases. The ODS efficiencies as ozone-depleting substances (*i.e.*, ODPs) and as greenhouse gases (*i.e.*, GWPs) are contrasted in Figure 4.3. A large range is found in both metrics for the principal gases. The continuous measurements of ODS abundances in the atmosphere over the last 2-3 decades allow an accurate evaluation of their contributions to ozone depletion and climate change. Projections of emissions allow the future contribution of ODSs to be estimated. Time series of ODP-weighted and GWP-weighted ODS emissions in Figure 4.4 show that both weighted emissions grew substantially in recent decades but peaked in the late 1990s (Velders *et al.*, 2007). The decline is in response to the provisions of the Montreal Protocol, which requires a staged phase-out of all principal ODS use in developed and developing nations. The RF contribution from ODSs likely would have been approximately twice as large in 2010 in the absence of the Montreal Protocol or other regulation (Velders *et al.*, 2007).

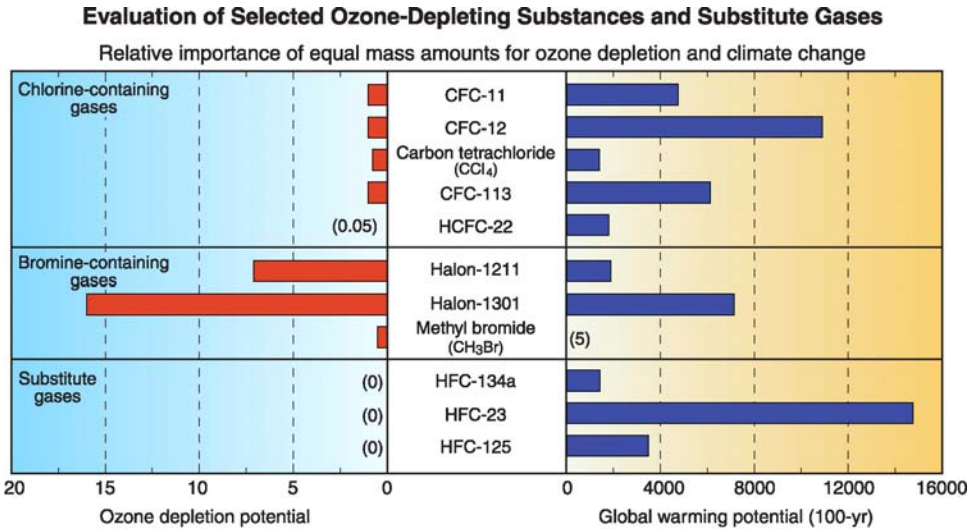


Figure 4.3 Comparison of the ozone depletion potentials (ODPs) and global warming potentials (GWPs) for principal ozone-depleting substances (ODSs) and HFCs. The contributions of emissions to ozone depletion and climate change increase with the ODP and GWP values, respectively. HFCs are ODS substitute gases which do not destroy ozone (*i.e.*, ODP = 0). The comparison is for emissions of equal mass. The GWPs are evaluated for a 100-yr period after emission. The ODPs of CFC-11 and CFC-12, and the GWP of CO_2 are defined to have values of 1.0 (WMO, 2007)

****** BEGIN BOX ******

BOX 4.1: The Ozone Depletion Potential and Global Warming Potential[1]

Ozone Depletion Potential. Ozone-depleting substances are compared in their effectiveness to destroy stratospheric ozone using the "Ozone Depletion Potential" (ODP), as shown in Figure 4.3. A gas with a larger ODP has a greater potential to destroy ozone over its lifetime in the atmosphere. The ODP is calculated on a "per mass" basis for each gas relative to CFC-11, which has an ODP defined to be 1. Halon-1211 and halon-1301 have ODPs significantly larger than CFC-11 and most other emitted gases, because bromine is much more effective overall (about 60 times) on a per-atom basis than chlorine in chemical reactions that destroy ozone in the stratosphere. The gases with small ODP values generally have short atmospheric lifetimes or fewer chlorine and bromine atoms. The production and consumption of all principal halogen source gases by humans are regulated under the provisions of the Montreal Protocol.

Global Warming Potential. The climate impact of a given mass of a halocarbon emitted to the atmosphere depends on its radiative properties and atmospheric lifetime. The two can be combined to compute the global warming potential (GWP), which is a proxy for the climate effect of a gas relative to the emission of a pulse of an equal mass of CO_2. Multiplying emissions of a gas by its GWP gives the CO_2-equivalent emission of that gas over a given time horizon. A value of 100 yrs is often chosen as a reference time horizon for intercomparisons of GWPs.

GWPs are most useful as relative measures of the climate response due to direct radiative forcing of well-mixed greenhouse gases whose atmospheric lifetimes are controlled by similar processes, which includes most of the halocarbons.

[1]Excerpted from WMO (2007) and IPCC/TEAP (2005).

****** END BOX 4.1 ******

The radiative forcing of individual ODSs varies because of differences in emissions, lifetimes, and GWPs. The RF values attributable to individual ODSs for the period 1970-2000 are shown in Table 4.1 along with values for CO_2 and CH_4. CFCs as a group form the largest contribution to RF amongst all ODSs. A comparison of the RF from halocarbon gases as a group with values associated with other aspects of natural and anthropogenic climate forcing is shown in Figure 4.5. Halocarbon gases represent 13% of the RF from all long-lived greenhouse gases and 21% of the total anthropogenic RF.

ODSs account for 94% (0.32 W per m^2) of the halocarbon term in Figure 4.5. The balance (6%) is due to the accumulation of hydrofluorocarbon (HFC) emissions, which are included in the Kyoto Protocol (UNFCC, 1997). Emissions of HFCs are increasing because they are ODS substitute gases. HFCs do not deplete ozone (ODP = 0) but can have substantial GWPs (Figure 4.3).

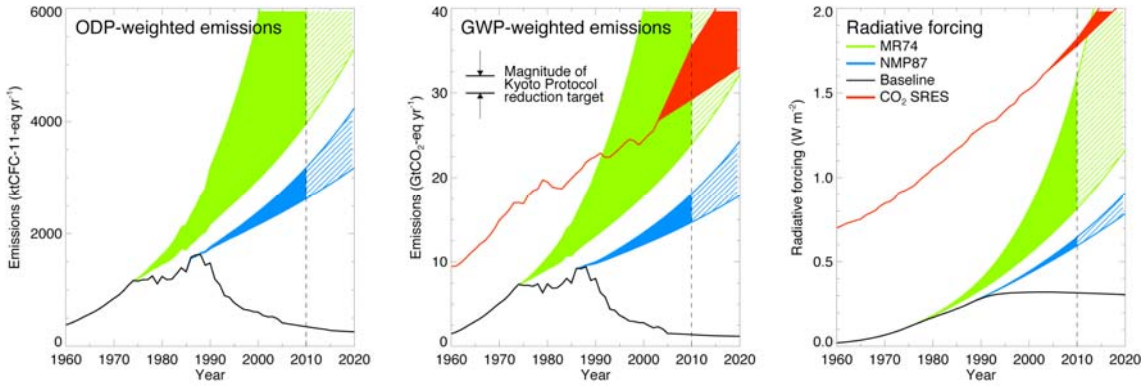


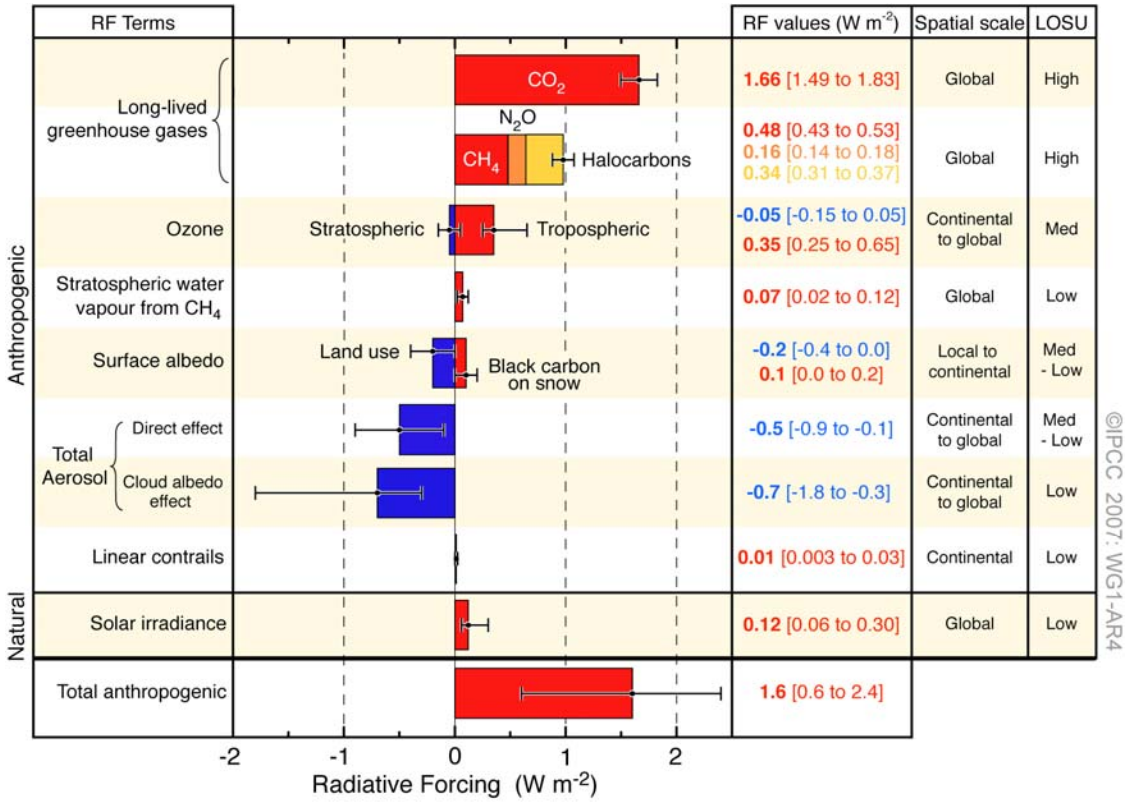
Figure 4.4 ODP-weighted emissions (left panel), GWP-weighted emissions (100-yr) (middle panel), and radiative forcing (right panel) for ODS and CO_2 scenarios for 1960–2020. Four scenarios are used: the baseline which represents ODS observations to date and projections for the future; the emissions that plausibly would have occurred in absence of the Molina and Roland warning that ODSs deplete ozone (MR74), the emissions that plausibly would have occurred in absence of the implementation of the Montreal Protocol (NMP87), and the IPCC SRES scenario for CO_2 emissions beyond 2003 (see legend). ODS emissions are normalized by their direct GWPs to form units of "equivalent $GtCO_2$ yr^{-1}". The shaded regions reflect uncertainties in the MR74 and NMP87 scenarios in projecting ODS growth rates. The striped shaded regions indicate larger scenario uncertainties past 2010. The CO_2 emissions for 1960–2003 are from global fossil fuel and cement production. All RF values represent net changes from the start of the industrial era (1750) to present. The reduction target of the first commitment period of the Kyoto Protocol is shown in the middle panel for reference. (Velders *et al.*, 2007)

Table 4.1 Radiative Forcing of CO_2, CH_4 and principal ODSs

Gas	Radiative forcing (W per m^{-2})
CO_2	0.67
CH_4	0.13
N_2O	0.068
CFC-11	0.053
CFC-12	0.136
CFC-113	0.023
CFC-114	0.003
CFC-115	0.002
HCFC-22	0.0263
HCFC-141b	0.0018
HCFC-142b	0.0024
Halon-1211	0.0012
Halon-1301	0.0009
CCl_4	0.0029

For accumulated emissions in the period 1970-2000. (Adapted from IPCC/TEAP, 2005).

A comprehensive evaluation of the protection of climate afforded by reductions in ODS emissions must take into account two compensating factors or offsets (Velders *et al.*, 2007). As ODS emissions are reduced, global stratospheric ozone are restored from their depleted state. Since ozone is a greenhouse gas, ozone RF increases as ozone levels are restored, thereby offsetting the reductions in RF from ODS reductions. The second offset is the increase in emissions of HFCs, all potent greenhouse gases, which is intrinsically tied to ODS reductions because HFCs are key substitute gases for ODSs. Thus, the net gain from reducing the RF contribution of ODSs must include negative offsets due to the reversal of some ozone depletion and increased abundances of other greenhouse gases. In 2010, these factors offset about 25% of the RF decrease attributable to reductions of



5208 **Figure 4.5** Radiative forcing values for the principal contributions to climate change from anthropogenic
5209 activities and natural processes. Each numerical value listed and indicated with a bar is a global mean value
5210 representing the change between preindustrial times (ca, 1750) and the present (2005). The error bars
5211 indicate the uncertainty ranges. The spatial scale and level of scientific understanding (LOSU) is also
5212 indicated for each value. (IPCC, 2007)
5213

5214 ODSs under the Montreal Protocol since 1990 (Velders *et al.*, 2007). As ODS
5215 abundances continue to decline in the atmosphere after 2010, the relative size of the
5216 ozone offset is likely to remain unchanged while the HFC offset might increase
5217 depending on growth in production and use of HFCs, which are not regulated by the
5218 Montreal Protocol.

5219

5220 **4.2.2 Radiative Forcing From Ozone Changes**

5221 Stratospheric ozone depletion and increases in tropospheric ozone both contribute to the
5222 RF of climate. The response of surface climate to ozone changes is complex, in general,

because of the balance between short-wave and long-wave radiative effects. For example, when ozone is increased in the troposphere or lower stratosphere, surface temperatures tend to increase due to increased long-wave forcing (Forster and Shine, 1997; IPCC/TEAP, 2005). Overall, surface temperatures are most sensitive to changes in ozone concentrations near the tropopause.

Stratospheric ozone depletion has occurred primarily at extratropical latitudes with substantially larger changes in the Southern Hemisphere. Southern Hemisphere ozone values over the period 2000-2003 are on average 6% below pre-1980 values, while Northern Hemisphere values are 3% lower. The net RF change from these observed depletions has been assessed by the IPCC to be near -0.05 W per m^2 (Figure 4.5) (IPCC, 2007; Hansen *et al.*, 2005). The instantaneous response to ozone depletion in the lower stratosphere is a positive forcing because solar flux significantly increases below the tropopause and downwelling longwave radiation decreases slightly. In addition, less solar and longwave radiation is absorbed in the lower stratosphere when ozone amounts are reduced, thereby cooling the region and further reducing the downwelling longwave flux from ozone and other gases. When all effects and feedbacks are taken into account, ozone depletion causes a net reduction in RF at the tropopause and a cooling effect on the atmosphere.

With an uncertainty of ±0.1 W per m^2, the IPCC best estimate of the RF from stratospheric ozone lies between -0.15 and +0.05 W per m^2 (IPCC, 2007). The possibility of a positive RF reflects that stratospheric ozone may have increased in some regions

since preindustrial times, despite losses related to ODSs. The RF value is particularly sensitive to ozone changes in the tropical lower stratosphere, which are small compared to changes at high latitudes. Observational and modeling studies indicate how Northern Hemisphere ozone amounts are also influenced by changes in atmospheric dynamics, such as changes in tropopause heights (Pyle *et al.*, 2005), in addition to increased amounts of ODSs. The model results show large differences in how ozone column amounts have responded at mid latitudes to changes in ODSs and other parameters, such as circulation (Gauss *et al.*, 2006).

Tropospheric ozone since preindustrial times has increased as a result of increased emissions of anthropogenic pollutants, primarily nitrogen oxides, carbon monoxide, and organic compounds, including methane. Photochemical and radiative transfer models are used to calculate ozone changes and the associated RF, respectively. The changes include the net transport of ozone from the stratosphere to the troposphere, which can be altered by climate change and stratospheric ozone depletion. The tropospheric ozone RF (0.35 W per m^2) from human activities is larger than the stratospheric ozone term and associated with large uncertainties (Figure 4.5).

4.3 THE RESPONSE OF OZONE TO CLIMATE CHANGE PARAMETERS

Ozone responds to climate change parameters in a variety of ways because ozone is photochemically produced and destroyed in the atmosphere and thus dependent on the abundance of other gases emitted by natural and anthropogenic processes. The complexity of the interaction of ozone with climate change parameters (Figure 4.1)

requires the use of chemistry climate models (CCMs) to diagnose the sensitivity of ozone to climate change parameters and to predict future ozone amounts in a changing climate.

BOX 4.2: Model Used To Study Climate Processes[1]

Atmospheric General Circulation Model (AGCM): A three-dimensional model of large-scale (spatial resolution of a few hundred km) physical, radiative, and dynamical processes in the atmosphere over years and decades. An AGCM is used to study changes in natural variability of the atmosphere and for investigations of climate effects of radiatively active trace gases (greenhouse gases) and aerosols (natural and anthropogenic), along with their interactions and feedbacks. Usually, AGCM calculations employ prescribed concentrations of radiatively active gases, *e.g.*, carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), chlorofluorocarbons (CFCs), and ozone (O_3). Changes of water vapor (H_2O) concentrations due to the hydrological cycle are directly simulated by an AGCM. Sea surface temperatures (SSTs) are prescribed. An AGCM coupled to an ocean model, commonly referred to as an AOGCM or a climate model, is used for investigation of climate change. More recently, climate models may also include other feedback processes (*e.g.*, carbon cycle, interaction with the biosphere).

Chemistry-Climate Model (CCM): An AGCM that is interactively coupled to a detailed chemistry module. In a CCM, the simulated concentrations of the radiatively active gases are used in the calculations of net heating rates. Changes in the abundance of these gases due to chemistry and advection influence heating rates and, consequently, variables describing atmospheric dynamics such as temperature and wind. This gives rise to a dynamical-chemical coupling in which the chemistry influences the dynamics (via radiative heating) and vice versa (via temperature and advection). Not all CCMs have full coupling for all chemical constituents; some radiatively active gases are specified in either the climate or chemistry modules. Ozone is always fully coupled, as it represents the overwhelmingly dominant radiative-chemical feedback in the stratosphere.

[1]Excerpted from WMO (2007).

END BOX 4.2

4.3.1 Calculating the Response of Ozone to Climate Change Parameters with CCMs

The approach to CCM use is schematically shown in Figure 4.6. Transport, radiation, dynamics, and chemistry and microphysics are the four principal aspects of a CCM. A CCM requires as input specific knowledge of natural process and their trends, such as in emissions, solar irradiance, and volcanic eruptions; and of human activities and their trends, primarily for emissions. These inputs define and constrain the current and future state of climate parameters. The CCM output includes a wide array of parameters and diagnostics in addition to ozone that can be compared to observations and other models.

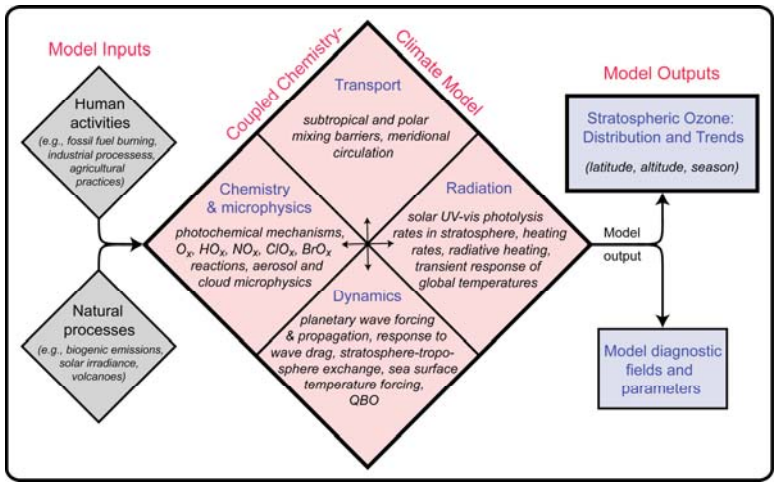


Figure 4.6 The processes in a Chemistry Climate Model (CCM) are represented by four basic groups: transport, dynamics, radiation, and stratospheric chemistry & microphysics. Significant interactions occur between aspects within each group. The CCM requires inputs describing human activities and natural processes. The CCM provides as output projections of future ozone abundances and their distribution along with a large variety of other parameters and diagnostics. See Box 4.2. (Eyring *et al.*, 2005).

CCMs are complex because simulating the atmosphere requires interdependence and interaction between the core aspects of the model. Important examples include the coupling between transport and radiation. Transport depends in part on atmospheric temperature gradients that are established by the distribution of radiative heating. Radiative heating is determined, in part, by long-lived greenhouse gases and ozone. Photochemical reaction rates also depend on ambient temperatures. Thus, the photochemical balance controlling the abundances of ozone and other species depends substantially on the atmospheric abundances of greenhouse gases. Sea surface temperatures, land-sea temperature differences and other factors influence wave propagation into the stratosphere, thereby affecting meridional transport rates. These couplings are discussed in more detail in WMO (2007) and Eyring *et al.* (2005).

The validation of CCM output for ozone and other parameters has become a focus topic because of the heightened need to project future ozone abundances with reliable uncertainty estimates (Eyring *et al.*, 2005; 2006). Reasonable agreement is found between many CCMs and global ozone trends, but for polar ozone trends the CCMs show a large spread in results. Uncertainties in CCM results reflect limitations in our understanding of how to represent atmospheric processes and their feedbacks in model simulations and, therefore, limit the precision and accuracy of our projections of future ozone amounts and the influence of climate change.

4.3.2 Stratospheric Temperature Changes

Stratospheric temperatures have decreased over the last 3 to 4 decades. Observations from satellites beginning in 1979 and radiosonde observations from about 1960 both reveal the cooling. The trend is about -0.5K/decade in the lower stratosphere and about 1 – 2K/decade in the middle stratosphere (~25-30 km). The latitude dependence of the temperature trends is not fully consistent across the various datasets, especially in the tropics, and remains a topic of research (WMO, 2007). The time series of temperatures reveals a non-monotonic decrease in the lower stratospheric temperatures (Figure 4.7). Volcanic aerosols formed in the aftermath of explosive volcanic eruptions led to a warming of the stratosphere for a few years following an eruption. Both the El Chichón and Mt. Pinatubo (1991) volcanic eruptions increased stratospheric aerosol amounts (McCormick *et al.*, 1995; Pawson *et al.*, 1998). The temperature trend decreases after 1995 after the gradual decay of volcanic aerosol amounts removed a large fraction from

the stratosphere (IPCC, 2007). In the evolution of the global lower stratospheric temperature, a sharp increase lasting for ~2 years is found immediately following the El Chichón (1982) eruption and is followed by a period of quasi-steady temperatures that are lower than the pre-eruption value. After the eruption of Mt. Pinatubo (1991), temperatures again increased sharply and were followed by a steady period in which the temperatures became lower than before this eruption. There is a slightly reduced cooling towards the end of 1990s and beginning of 2000s (Mears *et al.*, 2003).

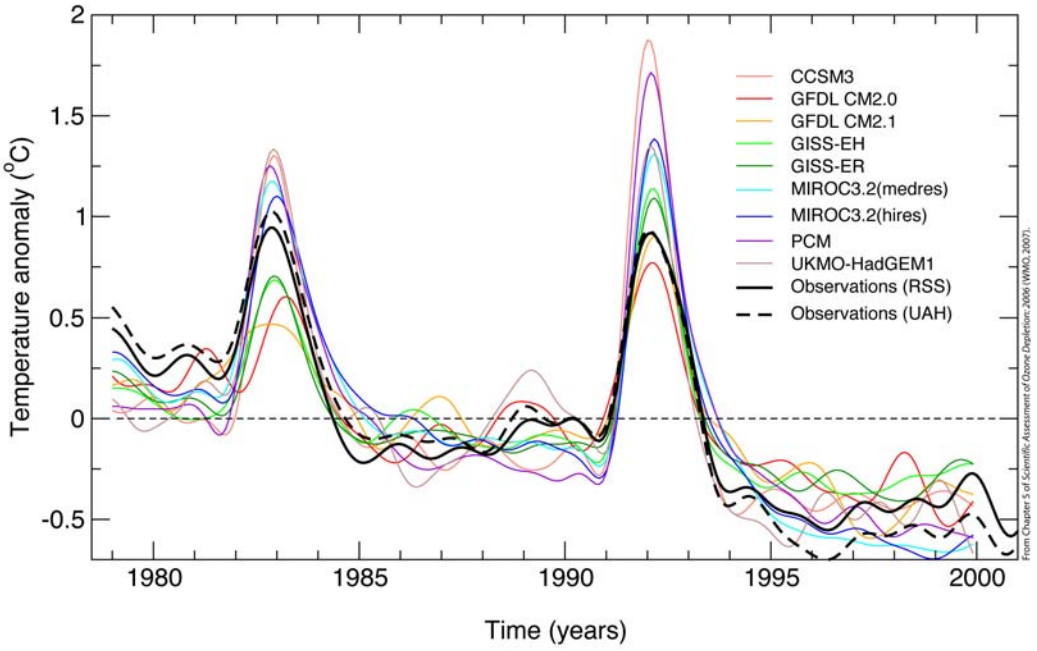


Figure 4.7 Temperature anomalies in the lower stratosphere as calculated by an ensemble of climate models. Results are compared with observations derived from satellite datasets. Further details are in Section 4.3.2. (From Santer *et al.*, 2006 and (WMO, 2007: Figure 5-3)

Climate model simulations also show that the combined influences of the agents that are known to "drive" the climate system offer a reasonable quantitative explanation of the observed non-monotonic decrease of the temperatures in the global lower stratosphere (Seidel and Lanzante, 2004; Dameris *et al.*, 2005; Ramaswamy *et al.*, 2006). The global

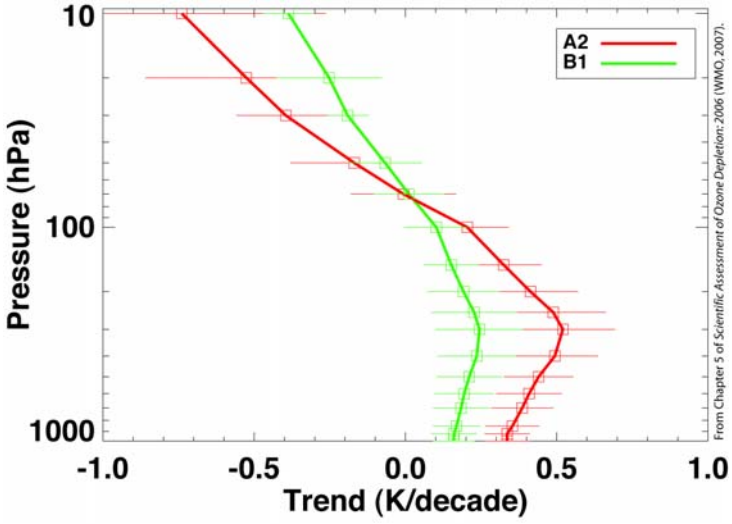
stratospheric temperature trends over the past 2-3 decades are attributed in modeling studies to a combination of increases in greenhouse gases and water vapor, and decreases in ozone (Ramaswamy and Schwarzkopf, 2002; Schwarzkopf and Ramaswamy, 2002; Langematz *et al.*, 2003; Shine *et al.*, 2003; Santer *et al.*, 2006). The above studies indicate that attribution of the cooling trend is possible on the global-annual and zonal-annual scales, and for the springtime Antarctic, but smaller spatial scales and seasonal behavior pose problems in attribution owing to the dynamical variability present. The results from an ensemble of climate models are shown in Figure 4.7 for stratospheric temperature anomalies calculated as global and monthly means (Santer *et al.*, 2006). In addition to stratospheric ozone depletion, the models all include climate forcings from changes in WMGHGs sulfate aerosol, volcanic aerosol, and solar irradiance. The temperature anomalies are differences from a 1979-1999 reference period. The models in general account for the long-term decrease in stratospheric temperatures and the short-term increases caused by two large volcanic eruptions.

The effects on climate of a wide range of forcings have been examined by Hansen *et al.* (2005), which shows that different forcings produce different response patterns in the vertical temperature profile. Results from climate model simulations as outlined in CCSP (2006) show:

- increases in greenhouse gases warm the troposphere and cool the stratosphere
- volcanic aerosols warm the stratosphere and cool the troposphere
- increase in solar forcing warms most of the atmosphere
- increases in tropospheric ozone warm the troposphere

- decreases in stratospheric ozone cool the stratosphere
- sulfate aerosols cool the troposphere and slightly warm the stratosphere.

The projections for the 21st century by coupled atmosphere-ocean general-circulation models (AOGCMs) using IPCC emissions scenarios show that average global temperatures continue to decrease in the stratosphere and increase in the troposphere (Figure 4.8). This result is primarily a consequence of increases in well-mixed greenhouse gases (WMGHGs) (mainly CO_2). Changes in the thermal gradients in the stratosphere and troposphere, initiated by greenhouse gas and aerosol changes, could additionally alter stratosphere-troposphere interactions and the state of the stratosphere. Changes of water vapor in the stratosphere arising from tropospheric warming, possible changes in convective activity and transport of water to the stratosphere, also can affect the stratospheric thermal state. Increased frequency of explosive volcanic events and ones much more intense than the Mt. Pinatubo eruption, should they occur, can be expected to substantially alter the climate and chemistry of the stratosphere for a few years through the particulates produced and impacts on atmospheric circulation.



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5414 **Figure 4.8** Temperature trends in the troposphere and stratosphere calculated as global and annual means
5415 for the 21st century using atmosphere-ocean GCMs (with no ozone chemistry) (AOGCMs). The
5416 calculations were made for two IPCC emission scenarios: A2 (high) and B1 (low). The symbols indicate
5417 the average trend computed for all models, while the thin horizontal lines indicate the range. (WMO,
5418 2007).
5419

5420 **4.3.2.1 Response of ozone to stratospheric temperature changes**

5421 With the coupling of ozone with climate parameters as outlined above, the effect of
5422 temperature changes on ozone is difficult to isolate. However, model simulations reveal
5423 some strong tendencies arising due to temperature changes. In the upper stratosphere,
5424 ozone amounts are controlled primarily by photochemical processes rather than transport
5425 and these processes are considered well understood. When temperatures decrease, ozone
5426 loss slows in the dominant photochemical cycles (NO_X, ClO_X, and HO_X) (Figure 4.9).
5427 For example, 15-20% ozone increases were calculated in the upper stratosphere for a
5428 climate with doubled CO_2 concentrations (Jonsson *et al.*, 2004). In the lower
5429 stratosphere, lower temperatures also decrease the destruction rate; but, production and
5430 destruction rates are lower than in the upper stratosphere and transport plays a more

important role. As a result, temperature changes have less influence on steady state ozone values in the lower stratosphere than in the upper stratosphere.

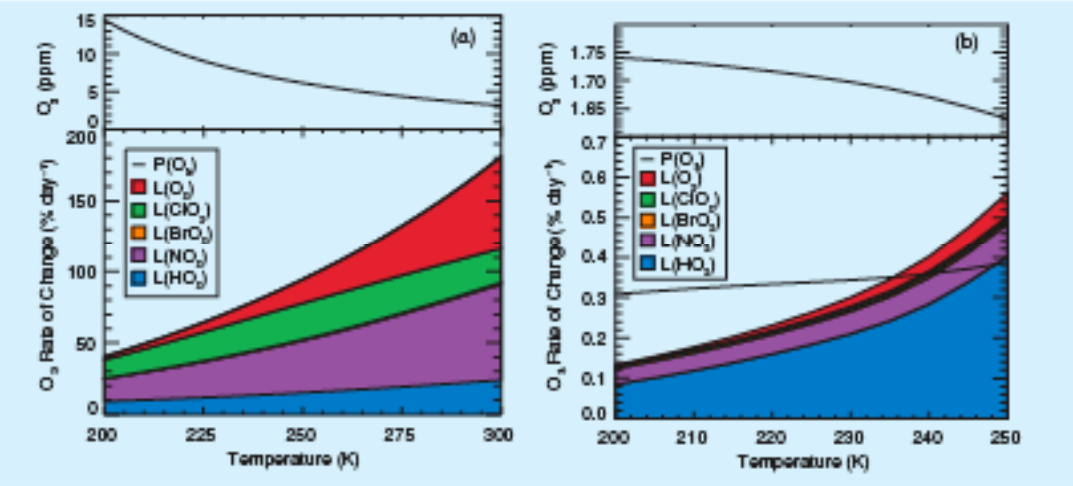

Figure 4.9 Comparison of ozone production and loss rates as a function of stratospheric temperature for 40-km altitude (left panel) and 20-km altitude (right panel) conditions at 45°N at equinox (end of March). The colored regions indicate the contribution from the principal loss cycles of ozone: odd-oxygen (O_X), reactive chlorine (ClO_X), reactive bromine (BrO_X), reactive nitrogen (NO_X), and reactive hydrogen (HO_X). The fractional contribution of each cycle varies with temperature differently in the two regions. The top trace in each panel is the ozone value at the end of 20-day runs of a chemical box model starting from climatological values for ozone, other constituents, and temperature (250K at 40 km; 215K at 20 km). At 40 km, the production rate coincides with the sum of loss rates because ozone is in photochemical balance at all temperatures shown. At 20 km, ozone production can be higher or lower than total loss depending on temperature because transport plays a more important role. The changes in ozone after 20-day runs are much smaller at 20 km than at 40 km, confirming that small temperature changes in the upper stratosphere will significantly alter ozone abundances. (IPCC/TEAP, 2005)

In the polar lower stratosphere, the reduction in photochemical loss with lower temperatures can be completely offset by increased activation of reactive chlorine and bromine, which increases ozone loss. Lower temperatures promote the formation of polar stratospheric clouds (PSCs), which facilitate heterogeneous reactions that form reactive halogens from reservoir gases. In the Arctic region, increased reactive halogens have the largest effect in controlling the ozone response to lower temperatures. For Northern

Hemisphere winters from 1993 to present, a strong linear relationship is found between winter/early spring ozone depletion and the volume of air containing PSCs during the winter. The ODS abundances are nearly constant during this time period (Rex *et al.*, 2004). Arctic ozone depletion might increase if further reductions occur in Arctic stratospheric temperatures because temperature decreases can lead to increases in the duration and frequency of PSCs (Douglass *et al.,* 2006).

In the Antarctic lower stratosphere, winter temperatures are well below the thresholds for heterogeneous conversion of halogen reservoirs for much longer periods and for much larger fractions of the polar vortex than found in the Arctic. Antarctic ozone depletion is currently much more extensive and complete, and decreasing temperatures would have less of an effect (Tilmes *et al.,* 2006). Under current conditions, seasonal Antarctic ozone depletion is more sensitive to reductions in ODS amounts than to small decreases in temperature (See Chapter 3) (Newman *et al.*, 2004). As ODS abundances decrease in the coming decades, polar ozone destruction due to reactions with halogen species ultimately will decrease in both hemispheres regardless of changes in the frequency and duration of PSCs.

4.3.3 Stratospheric Water Vapor Changes

The amounts of atmospheric water vapor, the most important and abundant greenhouse gas, serve as a feedback in the climate system. Change in the global distribution of water vapor is one of the important responses to the anthropogenic climate forcings summarized in Figure 4.5. Water vapor enters the stratosphere primarily through the

tropical tropopause. The water vapor abundance is reduced in dehydration processes involving low tropical tropopause temperatures and the formation and sedimentation of ice particles. Methane, released in the troposphere and oxidized in the stratosphere, is the underlying cause of the water-vapor component of anthropogenic radiative forcing as summarized in Figure 4.5.

Stratospheric water vapor has been measured by a wide variety of instruments and platforms, including balloons, aircraft, and satellites. The longest time series of continuous measurements is from small-balloon observations beginning in 1980. These measurements show that water vapor has increased at all levels between 15 and 26 km. At 17-22 km altitude, the increase can be expressed as a trend of ~5-10% per decade (Figure 4.10) (Oltmans *et al.*, 2000; Rosenlof *et al.*, 2001). Other stratospheric observations up to 30-35 km also show increasing trends, but over shorter time periods and with a high degree of variability (SPARC, 2000; Rosenlof *et al*., 2001). Part of the long-term increase in water vapor is attributable to increases in methane abundances due to anthropogenic emissions. Methane, which has increased by about 0.55 ppmv since the 1950s, is oxidized in the stratosphere producing two water molecules for each molecule of methane. The methane water vapor source in the stratosphere increases radiative forcing from water vapor by an estimated 0.1 W per m^{-2} (Myhre *et al.*, 2007). The implications of this indirect effect are not clear for the interpretation of water vapor and temperature trends in the stratosphere.

Since about 2000 the water vapor in key balloon and satellite observations in the mid to

lower stratosphere has shown significant decreases (Randel *et al.*, 2004). As a possible explanation, an analysis of the tropical tropopause temperatures for 1992-2005 shows that satellite water vapor amounts are consistent with interannual changes in the cold point temperatures and with the occasions of anomalously low tropopause temperatures (Randel *et al.*, 2004; 2006). Tropopause temperatures modulate the dehydration of air entering the lower stratosphere from the troposphere. In contrast, the earlier, longer water record from balloon measurements is not fully consistent with the record of tropopause temperatures (Seidel *et al.*, 2001). In general, the attribution of the causes of observed water vapor changes and trends in the stratosphere is incomplete, suggesting that projections of future amounts are uncertain.

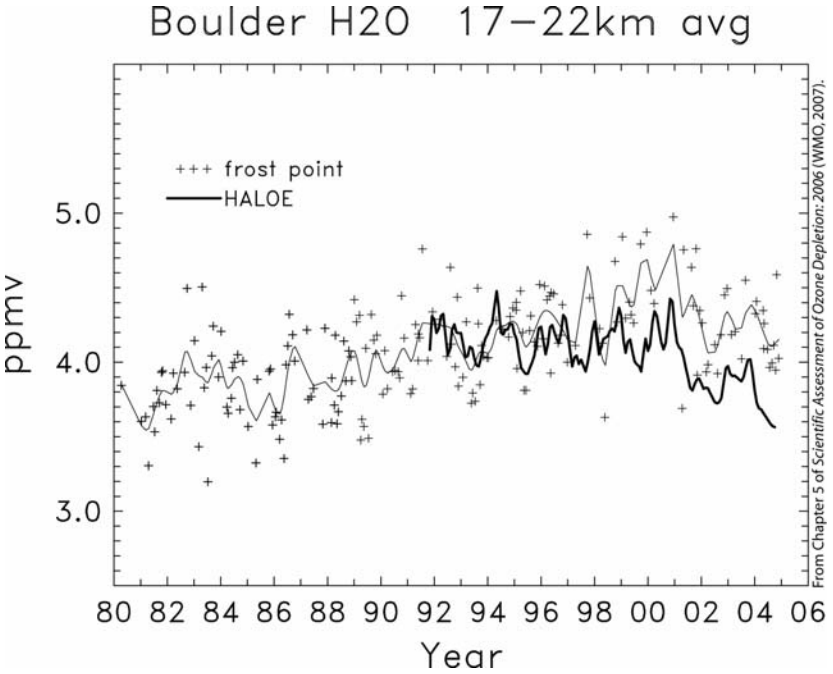

Figure 4.10 Time series of stratospheric water vapor mixing ratios (ppm, parts per million by volume) for the period 1980 to 2005. The measurements were made with a balloon-borne frost point hygrometer over Boulder, Colorado (40°N, 105°W). The data points are averages over 17-22 km altitudes. The thin line is a smoothed fit to the measurements. HALOE satellite observations for 1992-2005 are shown with the heavy line for the same altitude near Boulder (latitude 35°N-45°N, longitude 80°W-130°W). Preliminary revisions to the frost-point data reveal a slightly smaller trend (Scherer *et al.*, 2007). Updated from Randel

et al. (2004). (WMO, 2007)

4.3.3.1 Response of Ozone to Stratospheric Water Vapor Changes

Increases in stratospheric water lead to increases in reactive hydrogen species (HO_X), which catalyze the chemical destruction of ozone (Wennberg *et al.*, 1994; Brasseur and Solomon, 1986). Ozone destruction is chemically buffered with a combination of loss cycles so that the response to increased HO_X is generally not linear and varies with location in the stratosphere (Figure 4.9). Model simulations show that a 1%/year, long-term trend in water vapor would increase ozone loss due to increases in HO_X and delay the recovery of the ozone layer (Dvortsov and Solomon, 2001). Increased water vapor also increases the temperature threshold for PSC formation in both polar regions because PSC are formed, in part, from water-vapor condensation. A higher threshold increases heterogeneous conversion of chlorine and extends the time period over which PSCs can form in the winter season (Stenke and Grewe, 2005). Both effects lead to increased ozone destruction in polar regions at constant ODS amounts. However, the sensitivity of ozone to PSCs will decrease as ODS amounts decrease, because less chlorine and bromine will be available to participate in ozone destruction reactions.

4.3.4 Changes in Ozone from Increases in Long-Lived Gases in the Stratosphere

The atmospheric concentrations of the three long-lived greenhouse gases, CO_2, CH_4, and N_2O, have increased significantly due to human activities since 1750 and are expected to continue increasing in the 21st century (IPCC, 2007). These continuing increases have consequences for ozone amounts and, hence, also indirectly influence climate through the changes they produce in ozone (Portmann and Solomon, 2007). Calculations with a two-

dimensional, chemical-radiative-dynamical model illustrate the sensitivity of ozone to each of these gases (Figure 4.11). CO_2 increases, as discussed above, reduce stratospheric temperatures and ozone loss rates, and consequently, increase ozone amounts in the mid to upper stratosphere. The increased ozone in the upper stratosphere can lead to reduced ozone in the lower stratosphere because of the reduced penetration of solar UV into the lower stratosphere. Increases in N_2O lead to increases in the NO_X catalytic loss cycle for ozone in the mid to upper stratosphere, because N_2O decomposes to form NO_X in the stratosphere. The effect of increased NO_X is less in the lower stratosphere, because the NO_X loss cycle plays a less prominent role, competing with the HO_X and ClO_X catalytic loss cycles (Wennberg *et al.*, 1994). Finally, the oxidation of CH_4 increases H_2O and ozone losses in the HO_X catalytic cycle in the upper stratosphere and lower mesosphere. In the troposphere, ozone is increased because oxidation of CH_4 catalyzed by NO_X produces ozone.

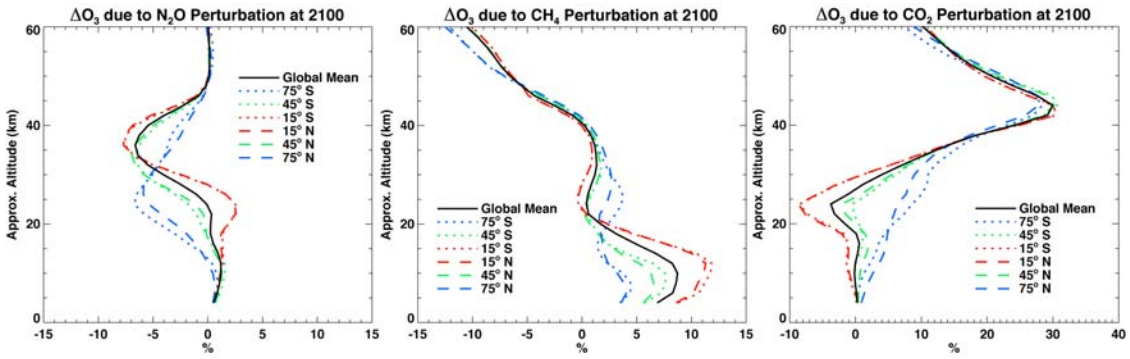

Figure 4.11 Comparison of perturbations to ozone amounts for changes in three principal greenhouse gases: CO_2, N_2O, and CH_4, as a function of altitude in the troposphere, stratosphere, and mesosphere. The changes are computed for six 30°-wide latitude bands with the NOCAR 2-D model and expressed as the percent change from 2000 to 2100. The global mean change is shown with the black line in each panel. The halogen changes follow the WMO (2003) scenario and greenhouse gas increases follow the IPCC A2

scenario. The latter is high compared to other scenarios but was chosen to maximize the ozone response in the model. (Portmann and Solomon, 2007)

4.3.4.1 Changes in Ozone from Stratospheric Circulation Changes

The net mass exchange between the troposphere and stratosphere is associated with the large-scale Brewer-Dobson circulation (Holton *et al.*, 1995) with a net upward flux in the tropics balanced by a net downward flux in the extratropics. Model studies indicate that climate change will impact the mass exchange rates across the tropopause. For a doubled CO_2 concentration, all 14 climate-change model simulations analyzed by Butchart *et al.* (2006) showed an increase in the annual mean troposphere-to-stratosphere exchange rate, with a mean trend of about 2% per decade. Consequences of such an increase include shorter lifetimes and more rapid removal from the atmosphere for long-lived gases, including CFCs, CH_4 and N_2O (Butchart and Scaife, 2001) and increased mass flux of ozone from the stratosphere to the troposphere at mid and high latitudes. A model simulation by Zeng and Pyle (2003) shows that a strengthened Brewer Dobson circulation would increase the flux of ozone to the troposphere. A larger flux results from increased transport across the tropopause and enhanced ozone amounts in the extra-tropical lower stratosphere. The enhanced ozone results from the strengthened circulation and decreases in ODSs and temperatures.

4.4 THE EFFECT OF OZONE CHANGES ON CLIMATE PARAMETERS

Ozone and climate change are highly coupled as illustrated in Figures 4.1 and 4.2. The response of ozone to changes in stratospheric temperature and water vapor is discussed in Section 4.3 above. In this section, changes in atmospheric temperatures and circulation are described as examples of the feedback responses in the climate system to ozone

depletion.

4.4.1 Response of Stratospheric and Tropospheric Temperatures to Ozone Depletion

Temperatures have decreased throughout the stratosphere in recent decades as described above in Section 4.3.2. Furthermore, model simulations show that a combination of increases in greenhouse gases and water vapor and decreases in ozone can account for observed temperature changes. A more detailed examination of the ozone feedback on temperature was carried out with the SKYHI GCM for ozone decreases observed in the period 1979-1997 (Ramaswamy and Schwarzkopf, 2002). The results in Figure 4.12 indicate that in the lower to middle stratosphere (5 – 100 hPa) ozone changes create a larger decrease in temperature than increases in WMGHGs. In this case, these include CO_2, CH_4, N_2O, CFC11, CFC-12, CFC-113 and HCFC-22. However, above about 5 hPa (~38 km) changes in both ozone and WMGHGs contribute significantly to temperature decreases. Thus, depletion in stratospheric ozone plays a significant role throughout the stratosphere in creating a reduction in stratospheric temperatures in GCM simulations.

The feedback on temperature for stratospheric ozone depletion extends into the upper troposphere. The Reading Narrow Band Model was used to calculate temperature changes for observed ozone depletion with the assumption of fixed dynamical heating (Forster *et al.*, 2007). Model cooling occurs in the 30 - 70 hPa (25-13 km) region due to ozone depletion. Short-wave absorption and upwelling long-wave radiation are both reduced and contribute comparably to the cooling in this region. The missing ozone also causes a decrease in the down-welling long-wave radiation that causes reduced

temperatures at altitudes below the ozone depletion region (100-150 hPa, 21-14 km). This feedback response or coupling of temperatures in different altitude regions is found at all latitudes in the model and may be a cause of upper tropospheric temperature trends.

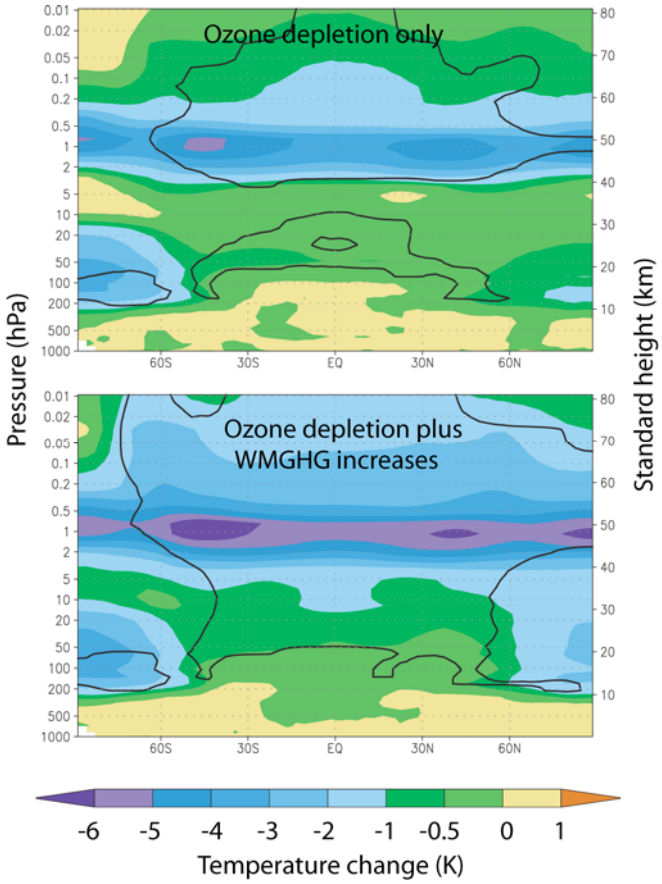

Figure 4.12 GCM feedback response of stratospheric temperatures to ozone depletion as observed between 1979 and 1997. Upper panel: ozone changes alone. Lower panel: ozone changes plus increases in WMGHGs over the same period. Solid lines enclose regions of statistical significance. (Adapted from Ramaswamy *et al.*, 2002)

4.4.2 Response of Surface Temperatures to Antarctic Ozone Depletion

The largest depletion in stratospheric ozone is found over Antarctic in late winter/early spring. Studies of Antarctic ozone depletion have revealed strong evidence for feedbacks on the temperatures and circulation of the Antarctic troposphere (Gillet and Thompson, 2003; Thompson and Solomon, 2002). Severe ozone depletion strengthens the

circumpolar winds of the Antarctic winter vortex in many model simulations. Recent observations show that strengthened circumpolar winds extend to the surface, especially in the summer months, with changes in geopotential heights serving as a proxy. A model with high vertical resolution was used to show anomalies in geopotential height in the troposphere could be well simulated in intensity and seasonality (Gillet and Thompson, 2003). Changes in surface circulation also lead to cooling over most of the Antarctic continent and modest warming of the Antarctic Peninsula. Figure 4.13 shows model results compared to observed changes in 500-hPa geopotential height over a 22-year period and in surface temperature over a 32-year period (1969 to 2000), both averaged over December to May. The observed and simulated patterns in geopotential height and surface temperature show strong similarities, reinforcing the conclusion that stratospheric circulation influences tropospheric circulation patterns and, hence, that intense stratospheric ozone depletion can effect changes in surface climate parameters.

4.5 IMPORTANCE OF VOLCANOES

4.5.1 The Effect of Volcanic Aerosol on Ozone

Large volcanic eruptions are those that inject significant quantities of SO_2 into the stratosphere. SO_2 is subsequently oxidized to sulfuric acid, which condenses onto preexisting aerosols, causing significant increases in aerosol surface area and volume in the lower stratosphere. As a consequence, heterogeneous reactions occurring on these surfaces gain prominence in the chemical production and loss balance of ozone, leading to decreased ozone amounts (WMO, 2007: Figures 3-26). These reactions convert nitrogen oxides (NO_x) to a more stable form, nitric acid (HNO_3). In the lower

stratosphere, reduced NO_x increases the role that reactive halogen compounds (ClO_x) play in destroying ozone. Analysis following the most recent large volcanic eruption, that of Mt. Pinatubo in 1991, shows that ozone amounts reached record lows and that halogen reactions, aided by temperature variability could explain the observed losses (Solomon *et al.*, 1998; Tie and Brasseur, 1995).

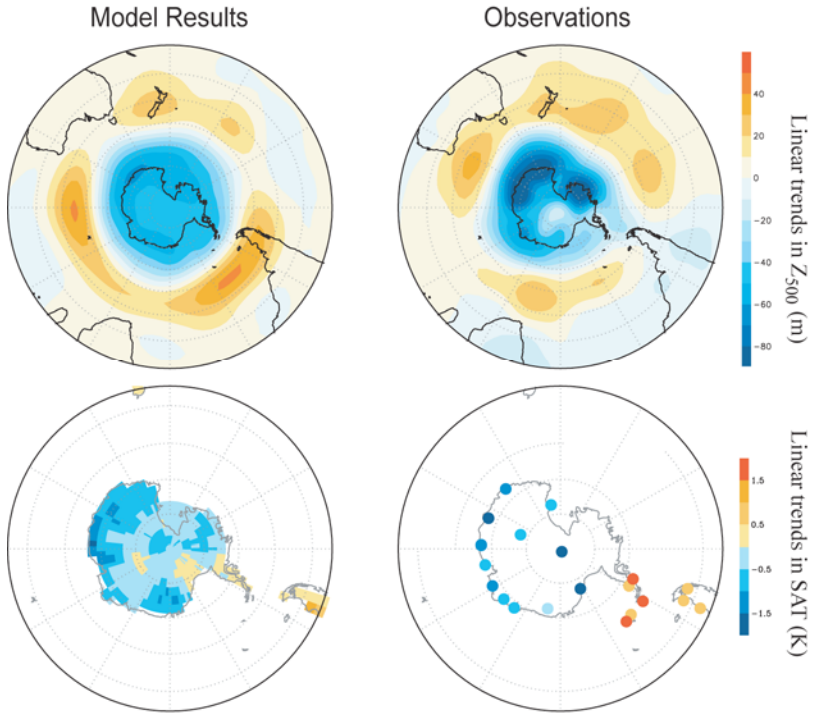

Figure 4.13 Comparison of model results (left column) with observations (right column) for changes in 500-hPa geopotential height (m) (upper row) and near-surface temperature (K) (lower row) in the Southern Hemisphere in response to stratospheric ozone depletion between 1979 and 1997. (Adapted from Gillett and Thompson, 2003)

As ODS amounts decrease to pre-1980 levels in the coming decades, the sensitivity of ozone to depletion caused by volcanic aerosol reactions will also decrease. Global ozone levels decreased by about 2% following the 1991 Mt. Pinatubo eruption. Sedimentation and transport removal of volcanic aerosol occurs over a 2- to 3-year period following an

eruption, so the effects are short lived compared to ODS atmospheric lifetimes which are 45-100 yrs for principal species (*e.g.*, CFC-11 and CFC-12). Thus, expectations for the long-term recovery of ozone are not significantly affected by episodic volcanic eruptions.

The plumes of large volcanic eruptions contain significant amounts of HCl that are removed in the troposphere by uptake and sedimentation of liquid aerosols formed (Tabazadeh and Turco, 1993). However, some eruptions inject non-negligible amounts of HCl into the stratosphere, adding to inorganic chlorine. For example, temporary increases in HCl column amounts of up to 40% were observed after the Mt. Pinatubo eruption (Coffey, 1996). Overall, the frequency of explosive volcanic eruptions has been low in the past two decades, thereby precluding significant volcanic enhancements in global stratospheric chlorine.

Volcanic aerosols have a direct radiative impact that increases stratospheric temperatures while decreasing the surface temperatures. Volcanic aerosols scatter incoming solar radiation and absorb solar infrared radiation, heating the lower stratosphere. The tropospheric cooling that results can be expected to change the tropospheric circulation, as well as the interaction between the stratosphere and the troposphere. Lower stratospheric temperatures, for example, following the eruptions of El Chichón and Mt. Pinatubo, were observed to increase by about 1K near 20 km altitude (Figure 4.7). The loss of ozone following an eruption also adds to the temperature perturbation. Lower stratospheric temperatures influence water vapor amounts through dehydration of air parcels entering the stratosphere from the troposphere and influence ozone amounts

through the sensitivity of ozone chemical reaction rates. Climate-chemistry model simulations of the temperature perturbations after the eruptions of El Chichón and Mt. Pinatubo often show larger increases than observed (Figure 4.7). The elevated temperatures are evident for several years and are followed by an overall slow cooling. The strength of the volcanic signal varies substantially between the different CCM and climate models (see also Eyring *et al.*, 2006).

Volcanic eruptions are not predictable but expected to be a feature of the future atmosphere. A large volcanic eruption is likely to occur in the next 30 years based on the historical record (Roscoe, 2001). Infrequent large volcanic eruptions would affect ozone with timescales as observed for previous large eruptions. A period of frequent large eruptions in the next century could enhance ozone depletion from ODSs for many years but the enhancement would lessen as global ODS abundances decline in the coming decades. Whenever the stratosphere is cleansed of volcanic aerosol, ozone abundances are expected to recover fully from volcanic effects.

4.6 SUMMARY

Stratospheric ozone and climate change are linked through a variety of processes. Radiative forcing of climate occurs from the depletion of stratospheric ozone, as well as the increases in ozone-depleting substances. Global ozone depletion is a principal cause of decreasing temperature trends in the stratosphere and upper troposphere. Severe ozone depletion over Antarctica has changed the circulation over the continent in both the stratosphere and troposphere and altered surface temperatures. Other important

components of anthropogenic climate change arise from emissions of long-lived greenhouse gases, such as carbon dioxide. Observed and anticipated changes in climate parameters include decreases in stratospheric temperatures and increases in stratospheric water vapor, carbon dioxide, methane, and nitrous oxide. Lower stratospheric temperatures reduce ozone loss rates in the mid to upper stratosphere, thereby aiding the recovery from ozone depletion. Enhanced water vapor alters ozone destruction rates in reactive hydrogen photochemistry and can increase the frequency and extent of polar stratospheric clouds, which aid ozone destruction. These varied composition changes contribute to circulation changes in the stratosphere and between the stratosphere and troposphere that can cause significant changes in the ozone distribution. The increases in stratospheric aerosols that follow explosive volcanic eruptions create several-year changes in climate parameters in the stratosphere and troposphere and increase ozone depletion.

The complexity of the interactions between ozone and climate involving changes in atmospheric composition pose a challenge to our understanding of basic stratospheric and tropospheric processes. Tools of the complexity of Chemistry Climate Models (CCMs) are required to combine stratospheric transport, dynamics, radiation, and chemistry and microphysics to analyze past ozone amounts and project future amounts. CCMs guided by atmospheric observations will help define the sensitivity of ozone to future climate changes and reduce the uncertainties in our understanding of ozone and climate interactions. As ozone depletion slows and ozone amounts recover from ODSs in the coming decades, changes in climate parameters will increase in importance in influencing

stratospheric ozone amounts.

4.6.1 Relevance for the United States

Human activities have led to changes in ozone abundances and climate parameters. Ozone depletion is attributed primarily to the accumulation of ozone-depleting substances and climate change is attributed to increases in long-lived greenhouse gases and surface albedo changes (Figure 4.5). Ozone is further influenced by changes in climate parameters such as stratospheric temperatures and composition, and atmospheric circulation. Since activities in the United States have caused significant emissions of greenhouse gases and ozone-depleting substances, the changes in ozone and climate attributable to human activities are, in part, attributable to the United States.

Decisions initiated or supported by United States policymakers have great potential to influence ozone and climate in the future. Important decisions could be taken on the following topics or issues:

- *Increased stringency of Montreal Protocol regulations.* The Montreal Protocol regulates production and consumption of ODSs in developed and developing nations. Stratospheric ODS amounts will decline to pre-1980 values around the middle of this century based on current regulations. More stringent regulations could accelerate this decline. For example, recent unratified regulation, supported by the United States, accelerates HCFC production in developed and developing nations.

- *Increased destruction or capture of ODS banks.* Banks of ODS compounds represent large sources of future ODS emissions (see Chapters 2 and 5).
- *Increased climate protection under the Montreal Protocol.* ODSs compounds are also greenhouse gases. Reducing ODS production and consumption under the Montreal Protocol has led to significant reductions in ODS atmospheric abundances and their associated radiative forcing of climate. Further reductions in ODS production, as well as emissions, will further protect climate. The accelerated HCFC phase-out under the Montreal Protocol represents a large potential benefit to climate. In addition, promotion of low-GWP compounds as replacement for ODSs in widespread applications can help minimize the climate consequences of new and existing Montreal Protocol regulations.
- *Reductions in the future growth rates of methane and nitrous oxide emissions*. Ozone and climate are strongly influence by methane and nitrous oxide emissions. These emissions affect the photochemical production and loss of ozone in both the troposphere and stratosphere. Both gases are greenhouse gases that have increased significantly due to human activities.

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