CHAPTER 4

How Do Climate Change and Stratospheric Ozone Loss Interact?

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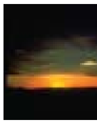
KEY ISSUES

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 nitrogen 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 a result of human activities and natural variability, ozone abundances will decrease or increase 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 are the causes 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?



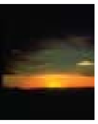
KEY FINDINGS

Linking of ozone and climate change

- Ozone and climate change are linked because both ozone and ozone-depleting substances (ODSs) are greenhouse gases, which contribute to radiative forcing. The ODS contribution to global radiative forcing between 1750 and 2005 is approximately 20 percent (+0.34 W per m^2) of that from carbon dioxide, the largest human 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 a different level of effectiveness as represented, for example, by the Ozone Depletion Potentials (ODPs) and Global Warming Potentials (GWPs), 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. 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. Ozone losses can also alter atmospheric temperatures and atmospheric transport. Ozone depletion can affect the climate of both the troposphere and stratosphere. The ozone depletion due to ODSs is expected to offset the direct climate forcing by ODSs. The extent of the offset is uncertain, with an estimated value of –0.05 W per m^2 with an uncertainty range of –0.15 to + 0.05 W per m^2.

Impact of climate change on ozone

- The complexity of the interactions of ozone changes with climate parameters requires development and evaluation of coupled models of Earth's atmospheric chemistry and climate processes (called chemistry climate models, CCMs). These CCMs are 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.
- Global average stratospheric temperatures have decreased in the observational records that begin in the 1960s. The decrease is attributed mainly to ozone depletion, increased carbon dioxide (CO_2), 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, 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 contributor 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 water vapor trends are uncertain because of uncertainties in projecting methane emissions and the temperatures of the tropical tropopause. Stratospheric water vapor influences stratospheric ozone through reactive hydrogen chemistry and polar stratospheric cloud formation.



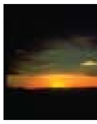
- Chemistry climate model simulations predict that the atmospheric circulation between the stratosphere and troposphere will increase in a changing climate in the coming decades. Increased circulation will change stratospheric ozone amounts and increase the stratospheric flux of ozone to the troposphere.

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 (approximately –0.05 W per m^2, with an uncertainty that encompasses a range from −0.15 to +0.05 W per m^2). The forcing is a balance between a shortwave cooling of the lower stratosphere and a longwave cooling below the region of ozone depletion. In comparison, increases in ozone from pollution chemistry in the troposphere have caused a positive radiative forcing (approximately +0.35 W per m^2).
- Ozone depletion causes changes to the temperature and circulation of the stratosphere and troposphere. Observational analyses indicate that stratospheric ozone depletion over Antarctica has strengthened circumpolar flow throughout the troposphere over Antarctica and caused surface temperature changes. This effect has been well simulated using many different general circulation models (GCMs).

Importance of volcanic eruptions

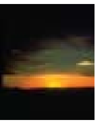
- If volcanic eruptions that inject material into the stratosphere, referred to here as explosive eruptions, occur again in the coming decades, they will decrease stratospheric ozone levels for several years as a result of the heterogeneous reactions occurring on volcanic sulfate aerosols. These reactions increase halogen loss processes by reducing the abundance of key reactive nitrogen compounds. For a given eruption size, the resulting enhanced ozone destruction from halogens will decrease as halogen amounts in the atmosphere 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 in 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 most of the additional sulfate aerosols within two to three 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 (UV) 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 longwave 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.

In the natural atmosphere, ozone's warming of the planet makes it the third most important longwave greenhouse gas after water vapor and carbon dioxide.



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 processes and human activities influence ozone through changes 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, dynamical processes such as planetary waves and the Brewer-Dobson circulation have important roles in establishing the observed non-uniform distribution of ozone in the atmosphere.

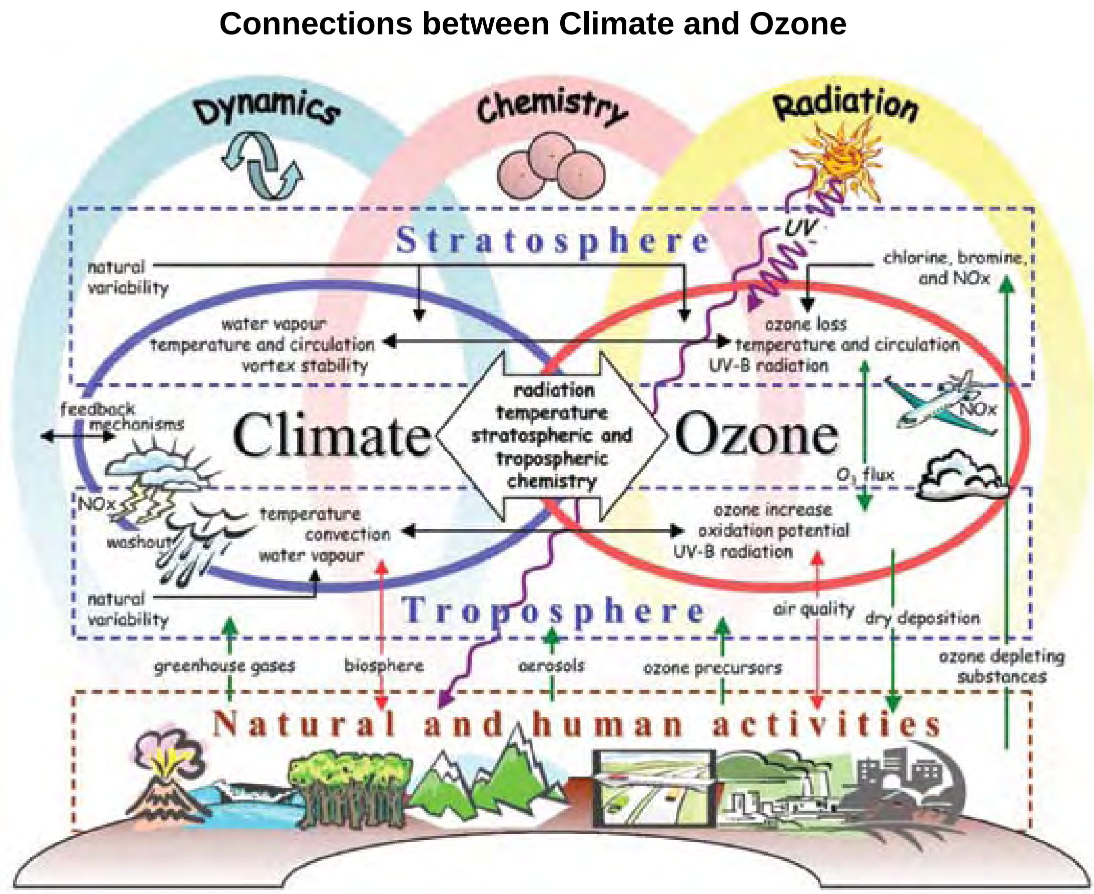


Figure 4.1 Schematic depiction of how climate change and ozone abundances are linked to each other in the stratosphere and troposphere and to natural and human activities at Earth's surface (Isaksen, 2003.)

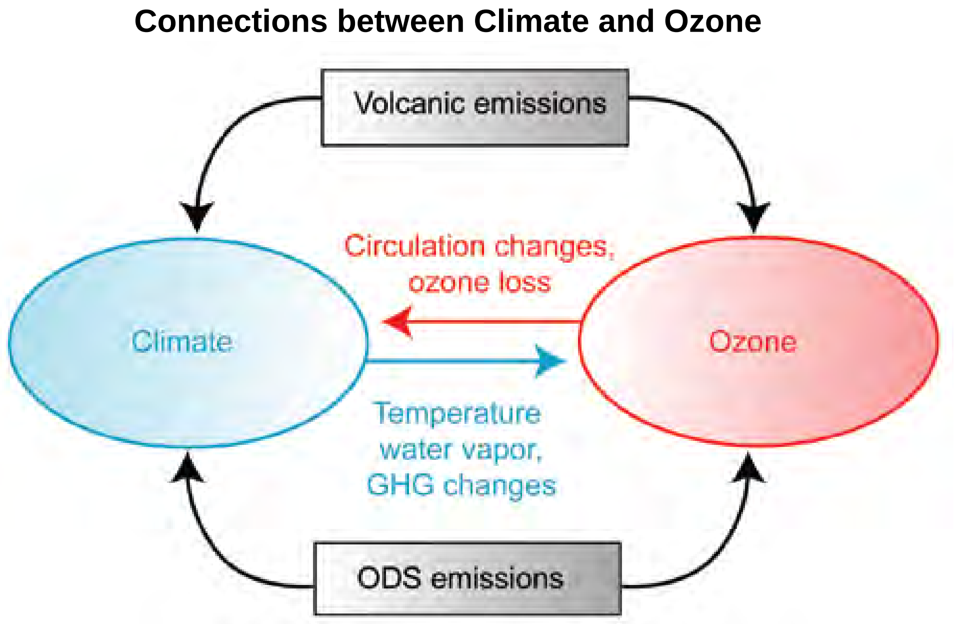


Figure 4.2 Schematic of specific processes addressed in this assessment that interconnect and influence atmospheric ozone amounts and climate parameters.

The accumulation of ozone-depleting substances in the atmosphere causes a direct radiative effect, similar to that from carbon dioxide.

The ODSs are also greenhouse gases. The radiative effect of accumulated ODS emissions is partially offset by the reduction in global ozone amounts caused by these emissions (see Section 4.2). Systematic and long-term ozone depletion can change atmospheric circulation patterns and contribute to climate change. Furthermore, changes in climate can 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 ODSs 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).

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

Ozone and ODSs are greenhouse gases. The accumulation of ODS emissions in the atmosphere causes a direct radiative effect in a manner similar to that from emissions of CO_2, the leading anthropogenic greenhouse gas. ODS emissions also cause an indirect radiative effect, because ODSs cause the destruction of stratospheric ozone. These radiative effects are discussed in the following sections.

4.2.1 Direct Radiative Forcing by Ozone-Depleting Substances

The accumulation of ozone-depleting substances in the atmosphere contributes to the direct radiative forcing of climate. The ODS efficiencies as ozone-depleting substances (*i.e.*, ODPs) and as greenhouse gases (*i.e.*, GWPs) are contrasted in Figure 4.3 (also see Box 4.1). A large range is found in both metrics for the principal gases. The continuous measurements of ODS abundances in the atmosphere over the last two to three decades allow an accurate evaluation of their contributions to ozone depletion and climate change. Projections of emissions allow the future contribution of ODSs to ozone depletion and climate change to be estimated. ODP-weighted and GWP-weighted

ODS emissions grew substantially in recent decades but peaked in the late 1990s, as shown in Figure 4.4 (IPCC/TEAP, 2005; 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 direct radiative forcing (RF) contribution from ODSs likely would have been approximately twice as large in 2010 in the absence of the Montreal Protocol or other regulation (Figure 4.4, right panel) (Velders *et al.*, 2007).

The radiative forcing of individual ODSs varies because of differences in emissions, lifetimes, and radiative efficiencies. The RF values attributable to individual ODSs for the period 1970 to 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. The RF from halocarbon gases in 2005 is +0.34 [±0.03] W per m^2 (warming), which represents 13 percent of the RF from all long-lived greenhouse gases and 21 percent of the total anthropogenic RF in 2005.

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

A comprehensive evaluation of the protection of climate afforded by increases or reductions in ODS emissions must take into account two indirect effects or offsets. The case of reduced ODS emissions from actions under the Montreal Protocol were analyzed recently by Velders *et al.* (2007). As ODS emissions are reduced, global stratospheric ozone levels 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

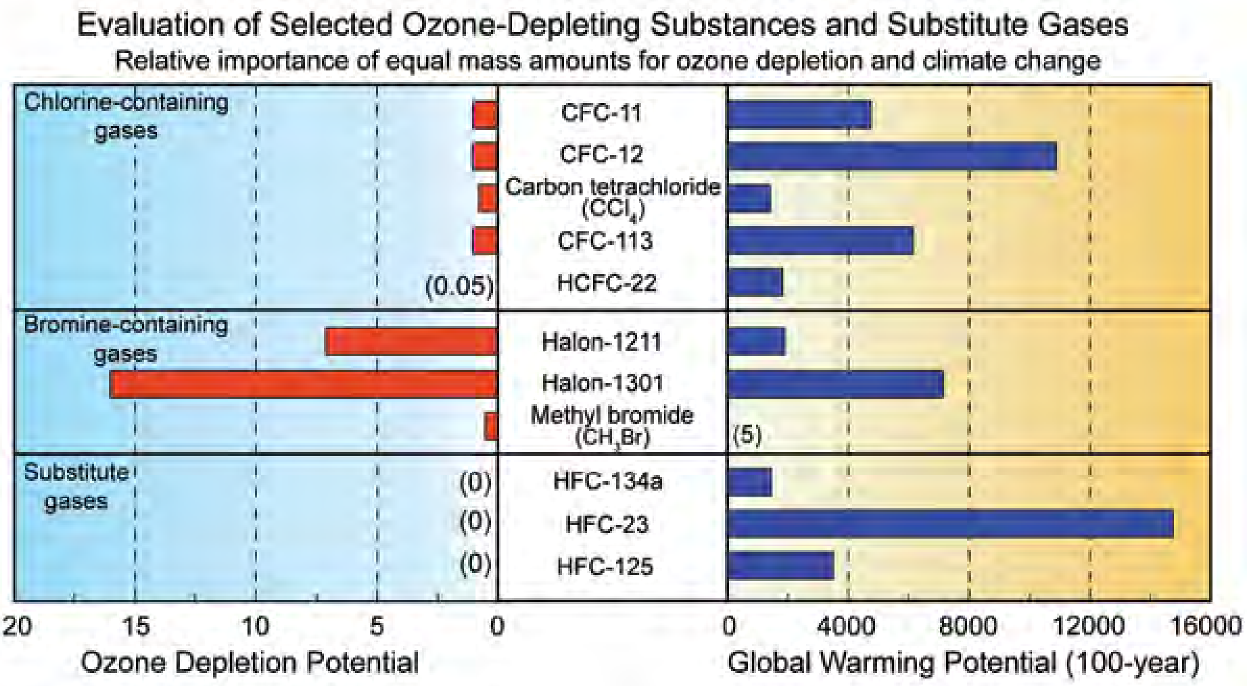


Figure 4.3 Comparison of the Ozone Depletion Potentials (ODPs) and Global Warming Potentials (GWPs) for principal ozone-depleting substances (ODSs) and hydrofluorocarbons (HFCs) (see Box 4.1). The contributions of emissions to ozone depletion and climate change increase with the ODP and GWP values, respectively. HFCs are ODS substitute gases that do not destroy ozone (*i.e.*, ODP = 0). The comparison is for emissions of equal mass. The GWPs are evaluated for a 100-year 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 (Daniel *et al.*, 1995; IPCC/TEAP, 2005; WMO, 2007).

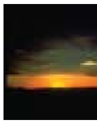
BOX 4.1: Ozone Depletion Potential and Global Warming Potential[1]

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 about 60 times more effective overall 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 ozone-depleting substances (ODSs) in human activities 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 both its radiative properties and its 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 emission of that gas over a given time horizon. A value of 100 years 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] Adapted from WMO (2007) and IPCC/TEAP (2005).



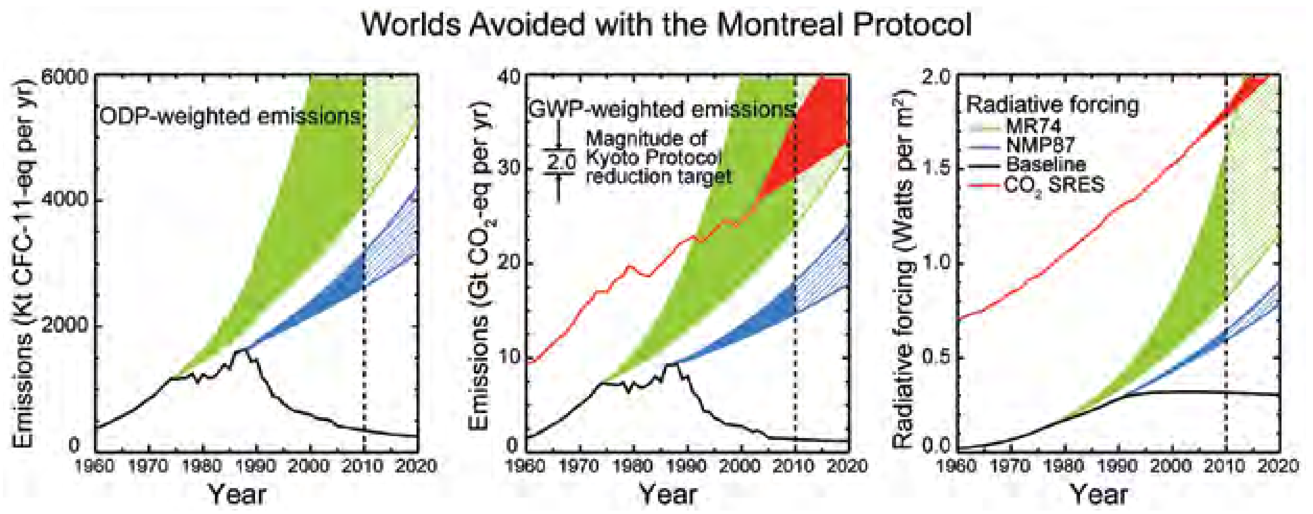


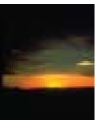
Figure 4.4 ODP-weighted emissions (left panel), GWP-weighted emissions (100-year) (middle panel), and radiative forcing (right panel) for ODS and CO_2 scenarios for 1960–2020. Four scenarios are used: (1) the baseline which represents ODS observations to date and projections for the future (black line); (2) the emissions that plausibly would have occurred in absence of the Molina and Roland warning that ODSs deplete ozone (MR74) (green shaded and striped region), (3) the emissions that plausibly would have occurred in absence of the implementation of the Montreal Protocol (NMP87) (blue shaded and striped region), and (4) the IPCC SRES scenario for CO_2 emissions beyond 2003 (red line). ODS emissions are normalized by their direct GWPs to form units of equivalent gigatons (Gt) CO_2 per year. The shaded regions reflect uncertainties in projecting ODS growth rates in the MR74 and NMP87 scenarios. The striped regions indicate larger uncertainties are associated with the scenarios 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 (2005). The magnitude of the reduction target of the first commitment period of the Kyoto Protocol (2008 to 2012) is shown in the middle panel for reference. The target represents average emission reductions in six key greenhouse gases by Annex-1 countries (5.8 percent below 1990 values) plus the expected growth in emissions between 1990 and 2008 to 2012 (adapted from Velders *et al.*, 2007).

Table 4.1 Direct radiative forcing of CO_2, CH_4, and principal ODSs[1]

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

[1] For accumulated emissions in the period 1970-2000. (Adapted from IPCC/TEAP, 2005, Table TS-1.)

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 were estimated to offset about 25 percent of the RF decrease attributable to reductions of ODSs under the Montreal Protocol since 1990 (Velders *et al.*, 2007). There is a large uncertainty in this offset in radiative forcing (see below). As ODS abundances continue to decline in the atmosphere after 2010, the relative size of the ozone offset is likely to remain unchanged while the HFC offset might increase depending on growth in production and use of HFCs, which are not regulated by the Montreal Protocol.



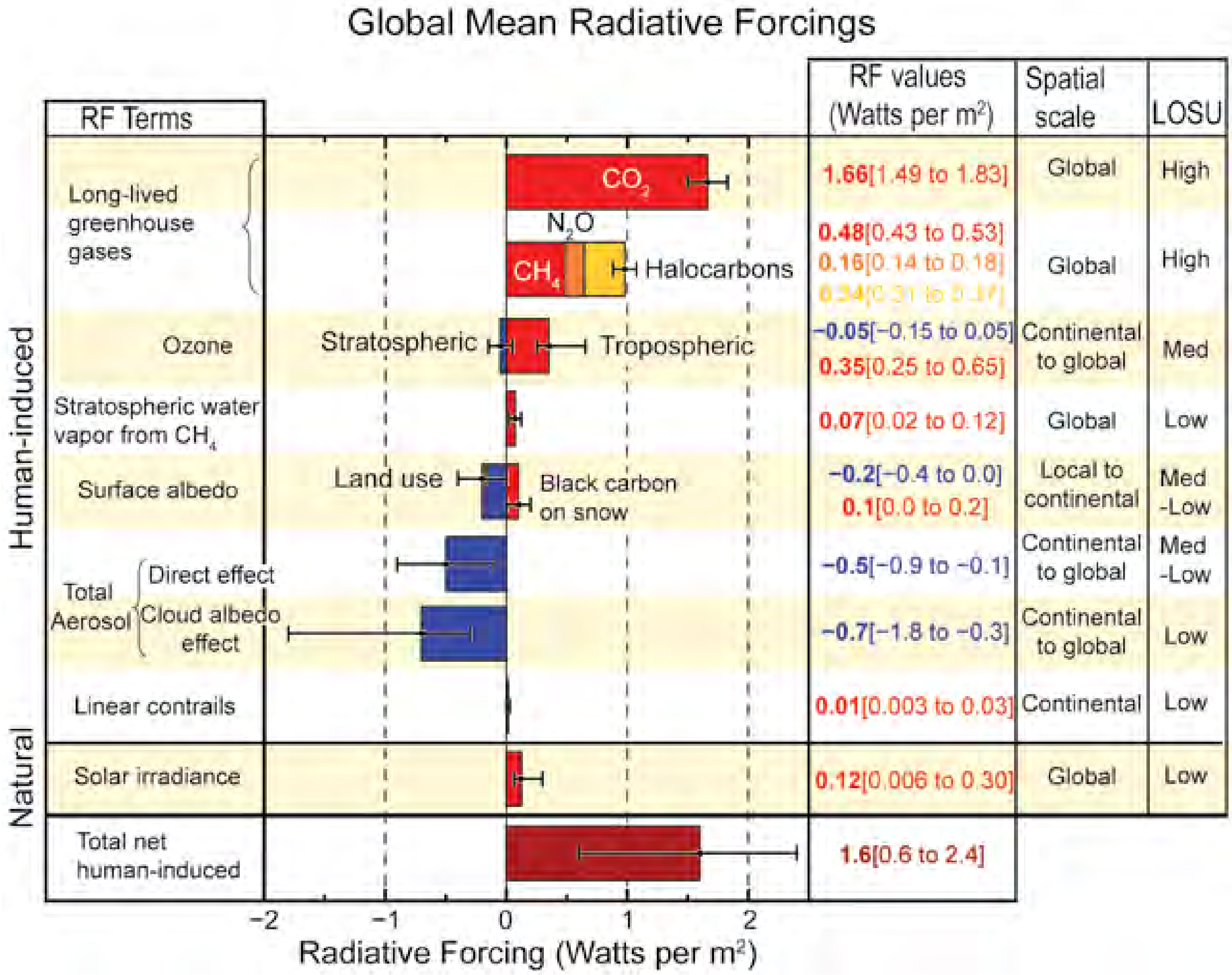


Figure 4.5 Radiative forcing values for the principal contributions to climate change from anthropogenic activities and natural processes. Each numerical value listed and indicated with a bar is a global mean value representing the best estimate of the change between preindustrial times (ca, 1750) and the present (2005). The error bars indicate the uncertainty ranges. The spatial scale and level of scientific understanding (LOSU) is also indicated for each value (adapted from IPCC, 2007, Figure SPM-2).

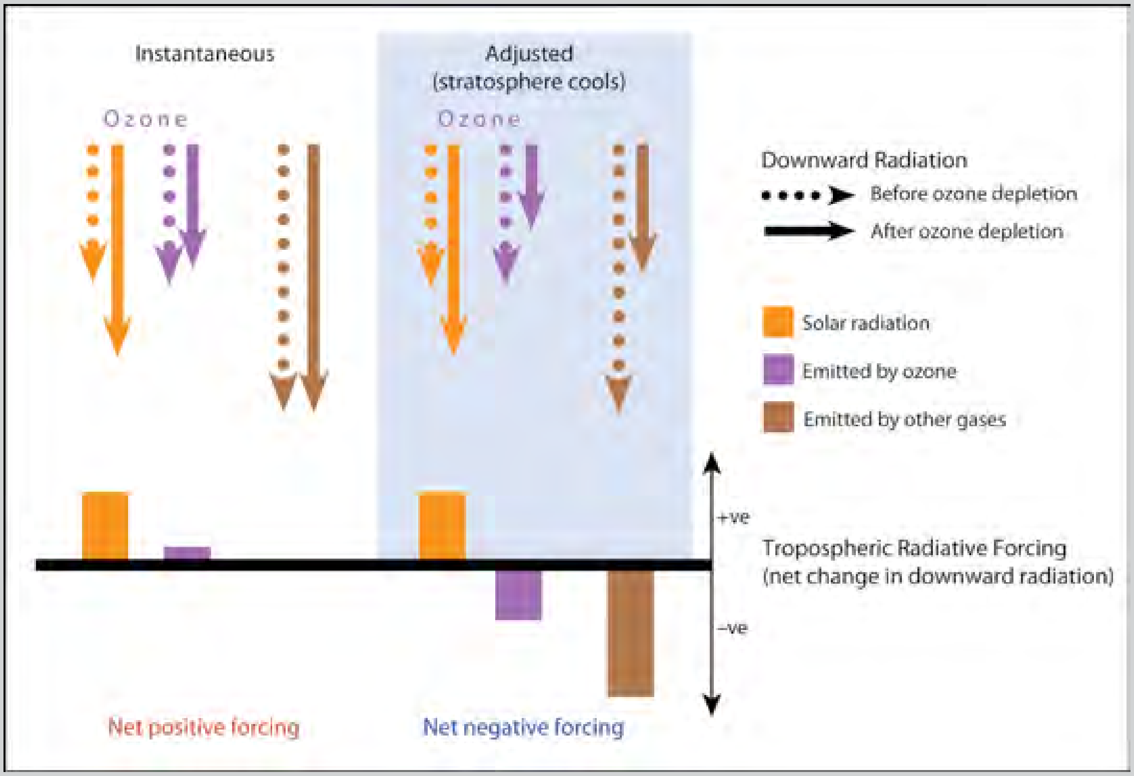
4.2.2 Radiative Forcing From Ozone Changes

Stratospheric ozone depletion and increases in tropospheric ozone both contribute to the radiative forcing of climate. Stratospheric ozone depletion is an indirect radiative effect of ODS emissions. The response of surface climate to ozone changes is complex, in general, because it depends on the balance between shortwave and longwave radiative effects. For example, when ozone is increased in the troposphere or lower stratosphere, surface temperatures tend to increase due to increased longwave forcing (Forster and Shine, 1997; IPCC/TEAP, 2005) (see Box 4.2). 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 from 2000 to 2003 are on average 6 percent below pre-1980 values, while Northern Hemisphere values are 3 percent lower. The net RF change from these

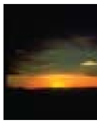
BOX 4.2: Why is the radiative forcing of stratospheric ozone negative?[1]

The radiative forcing is defined as the change in net radiative flux at the tropopause after allowing for stratospheric temperatures to readjust to radiative equilibrium (IPCC, 2001, Chapter 6). The details of this definition are crucial for stratospheric ozone, and are explained in the figure.



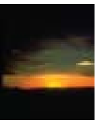
Box Figure 4.2 The instantaneous effect of stratospheric ozone depletion (left-hand side of schematic) is to increase the shortwave radiation from the Sun reaching the tropopause (because there is less ozone to absorb it), and to slightly reduce the downward longwave radiation from the stratosphere, as there is less ozone in the stratosphere to emit radiation. This gives an instantaneous net positive radiative forcing. However, in response to less absorption of both shortwave and longwave radiation in the stratosphere, the region cools, which leads to an overall reduction of thermal radiation emitted downward from the stratosphere (right-hand side of schematic). The size of this adjustment term depends on the vertical profile of ozone change and is largest for changes near the tropopause. For the observed stratospheric ozone changes the adjustment term is larger than the positive instantaneous term, thus the stratospheric ozone radiative forcing is negative.

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



observed depletions has been assessed by the Intergovernmental Panel on Climate Change (IPCC) to be −0.05 [±0.1] W per m^2 (cooling) (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. The calculation of the net change in forcing is associated with large uncertainties because net cooling is sensitive to the location where ozone depletion occurs (see Box 4.2). When all effects and feedbacks are taken into account, stratospheric ozone depletion is estimated to cause a net reduction in RF at the tropopause and lead to a cooling effect on the atmosphere.

Stratospheric ozone depletion is estimated to cause a net reduction in radiative forcing at the tropopause and lead to a cooling effect on the atmosphere.



Stratospheric ozone depletion is included in the IPCC best estimate of the RF from global stratospheric ozone changes that have occurred between 1750 and 2005 (IPCC, 2007). The net global ozone RF estimates used in the IPCC include the possibility of a positive RF contribution due to a modeled increase in stratospheric ozone 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 midlatitudes to changes in ODSs and other parameters, such as circulation (Gauss *et al*., 2006).

The total RF associated with ODS emissions is the sum of three effects: the direct warming summed over the accumulation of the contributing ODS gases (+0.34 [±0.03] W per m^2); the direct warming from HFC emissions that occurred as a result of ODSs reductions; and the indirect cooling associated with stratospheric ozone changes (−0.05 [±0.1] W per m^2). The warming due to HFCs is currently small (approximately +0.02 W per m^2) but is expected to grow in the future as ODS warming declines (Velders *et al*., 2007). The uncertainty in the indirect effect is large compared to the direct effects. As previously discussed, in some cases the indirect effect as evaluated in atmospheric models includes effects not related to ODS emissions. In addition, the regional distribution of the radiative forcing from ODSs generally is different from that due to ozone changes. Thus, there are currently large uncertainties in the magnitude of global ozone cooling effects and the degree to which they offset the warming due to ODSs, as well as where and how these offsets would vary regionally.

Tropospheric ozone since preindustrial times has increased as a result of increased emissions of human 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 continually photochemically produced and destroyed in the atmosphere and thus dependent on the abundance of other gases emitted by natural processes and human activities. The complexity of the interaction of ozone with climate change parameters (Figure 4.1) requires the use of chemistry climate models (CCMs) (see Box 4.3) to diagnose the sensitivity of ozone to climate change parameters and to predict future ozone amounts in a changing climate.

BOX 4.3: Models Used to Study Climate Processes[2]

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 human-made), 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.

[2] Excerpted from WMO (2007).

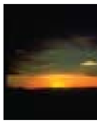
4.3.1 Calculating the Response of Ozone to Climate Change Parameters with CCMs

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.

Chemistry climate models 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.

A chemistry climate model requires as input specific knowledge of natural processes and their trends, such as in emissions, solar irradiance, and volcanic eruptions. It also requires knowledge of human activities and their trends, primarily for emissions.



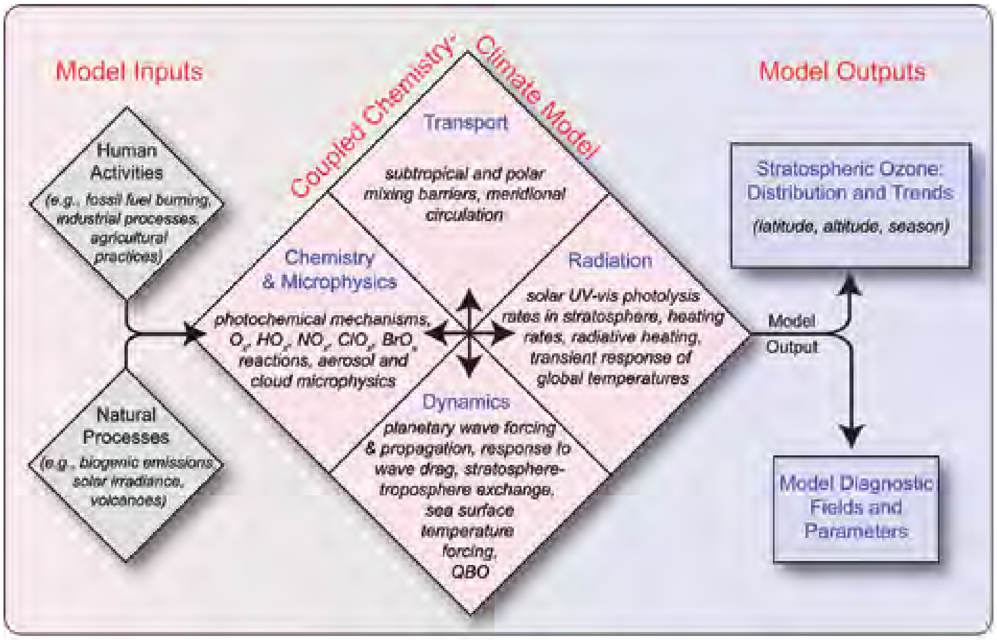
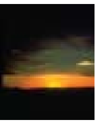


Figure 4.6 The processes in a Chemistry Climate Model (CCM) are represented by four basic groups: transport, dynamics, radiation, and stratospheric chemistry and microphysics. Significant interactions occur between aspects within each group. The CCM requires inputs describing human activities and natural 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 4.3 (Eyring *et al.*, 2005).

Stratospheric temperature has decreased by about 0.5 K per decade in the lower stratosphere and by about 1 to 2 K per decade in the middle stratosphere.



4.3.2 Stratospheric Temperature Changes

Stratospheric temperatures have decreased over the last three to four decades. Observations from satellites beginning in 1979 and radiosonde observations from about 1960 both reveal the cooling. The trend is about −0.5K per decade in the lower stratosphere and about −1 to −2K per decade in the middle stratosphere (about 25 to 30 km in altitude). 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 a volcanic eruption that injects sulfur dioxide in to the stratosphere lead to a warming of the stratosphere for a few years following an eruption. Both the El Chichón (1982) and Mt. Pinatubo (1991) volcanic eruptions increased stratospheric aerosol amounts (Figure 4.7) (McCormick *et al.*, 1995; Pawson *et al.*, 1998). In the evolution of the global lower stratospheric temperature, a sharp increase, lasting for approximately two 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).

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 two to three decades are attributed in modeling studies to a combination of increases in well-mixed 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

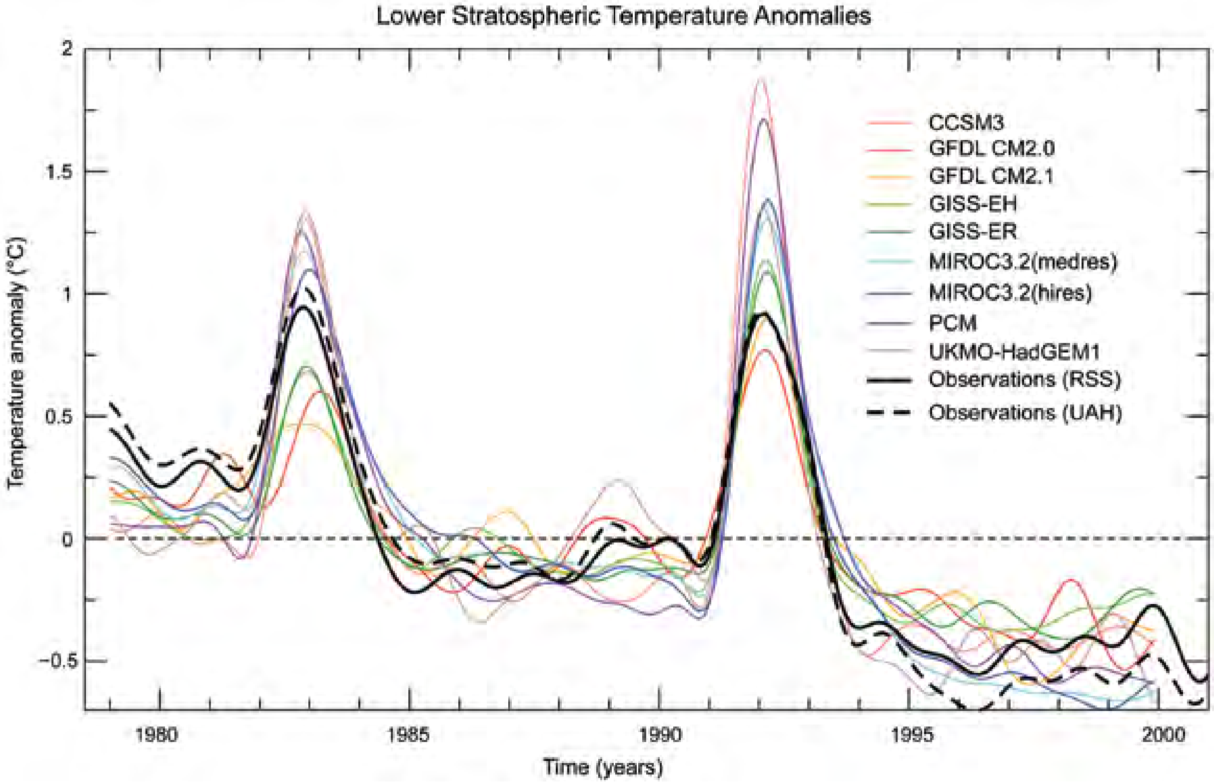


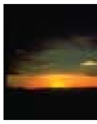
Figure 4.7 Temperature anomalies in the lower stratosphere as calculated by an ensemble of climate models. All anomalies are expressed as departures from a 1979 to 1999 reference period average. These models were chosen because they satisfy certain minimum requirements in terms of the forcings applied in the model run. All were driven by changes in well-mixed greenhouse gases (WMGHGs), sulfate aerosol direct effects, tropospheric and stratospheric ozone, volcanic aerosols, and solar irradiance. Model results are compared with observations derived from satellite datasets (RSS and UAH). Further details are discussed in Section 4.3.2 (From Santer *et al*., 2006: Figure 5.2 and WMO, 2007: Figure 5-3).

seasonal behavior pose problems in attribution owing to the dynamical variability. 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 well-mixed greenhouse gases (WMGHGs), sulfate aerosol, volcanic aerosol, and solar irradiance. The temperature anomalies are differences from a 1979 to 1999 reference period. The models, in general, account for the long-term decrease in stratospheric temperatures and the short-term increases caused by the two large volcanic eruptions.

Hansen *et al.* (2005) examined the effects on climate of a wide range of forcings, showing 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 twenty-first century by coupled atmosphere-ocean general-circulation models (AOGCMs) using IPCC emissions scenarios show that average global temperatures will continue to decrease in the stratosphere and increase in the troposphere (Figure 4.8). This result is primarily a consequence of increases in WMGHGs (mainly CO_2). Changes in the thermal gradients in the stratosphere and troposphere, initiated by greenhouse gas and aerosol changes, could additionally



For the twenty-first century, average global temperatures are projected to continue to decrease in the stratosphere and increase in the troposphere, primarily as a consequence of increases in well-mixed greenhouse gases.

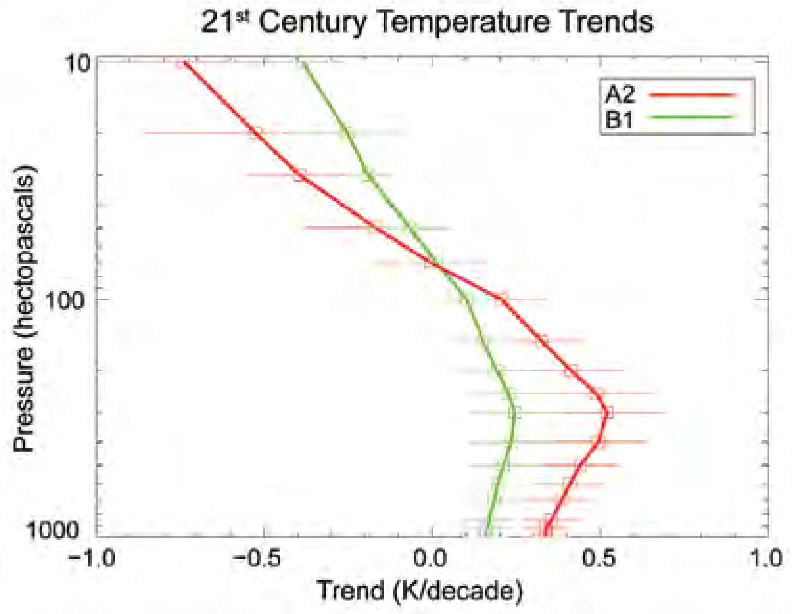
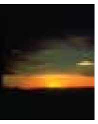


Figure 4.8 Temperature trends in the troposphere and stratosphere calculated as global and annual means for the twenty-first century using atmosphere-ocean GCMs (AOGCMs, with no ozone chemistry). Temperatures are expected to continue to increase in the troposphere and decrease in the stratosphere. The calculations were made for two IPCC emission scenarios: A2 (high) and B1 (low). The symbols indicate the average trend computed for all models, while the thin horizontal lines indicate the range. The vertical scale represents altitude from the surface (1000 hectopascal, hPa) to approximately 30 km (10 hPa). The pressure of the tropopause, which divides the stratosphere above from the troposphere below, varies between 100 and 300 hPa over most of the globe (WMO, 2007).

alter stratosphere-troposphere interactions and the state of the stratosphere. Changes of water vapor in the stratosphere arising from tropospheric warming, and possible changes in convective activity and transport of water to the stratosphere, also can affect the stratospheric thermal state. Volcanoes 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 if they erupt more frequently and/or if they inject much more material into the stratosphere than the Mt. Pinatubo eruption.



When temperatures decrease in the upper stratosphere, ozone loss slows in the dominant photochemical cycles.

4.3.2.1 RESPONSE OF OZONE TO STRATOSPHERIC TEMPERATURE CHANGES

With the coupling of ozone with climate parameters as outlined above, the effect of temperature changes on ozone is difficult to isolate. However, model simulations reveal some strong tendencies arising due to temperature changes. In the upper stratosphere, ozone amounts are controlled primarily by photochemical processes rather than transport, and these processes are considered well understood. When temperatures decrease, ozone loss slows in the dominant photochemical cycles (reactive nitrogen [NO_x], reactive chlorine [ClO_x], and reactive hydrogen [HO_x]) (Figure 4.9). For example, 15-20 percent ozone increases were calculated in the upper stratosphere for a climate with doubled (CO_2) concentrations (Jonsson *et al.*, 2004). In the lower stratosphere, the destruction rate decreases for lower temperatures, but production and 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.

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

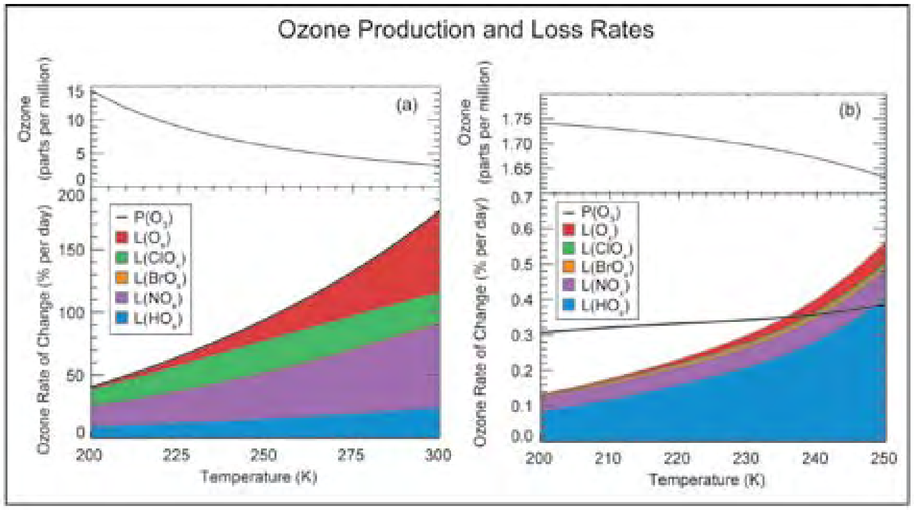
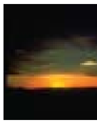


Figure 4.9 Comparison of ozone production and loss rates as a function of stratospheric temperature for 40-km altitude (left panel) and 20-km altitude (right panel) conditions at 45°N at equinox (end of March). The colored regions indicate the contribution from the principal loss cycles of ozone: odd-oxygen (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 (adapted from IPCC/TEAP, 2005).

Water vapor has increased about 5 to 10% per decade at a height of 17 to 22 km since 1980.



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

Atmospheric water vapor is the most important and abundant greenhouse gas. Change in the global distribution of water vapor is one of the important responses to the human-caused 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 human-caused 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 altitude. At 17 to 22 km, the increase can be expressed as a trend of about 5-10 percent per decade (Figure 4.10) (Oltmans *et al.*, 2000; Rosenlof *et al.*, 2001). Other stratospheric observations up to 30 to 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 human emissions. Methane, which has increased by about 0.55 parts per million by volume (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).

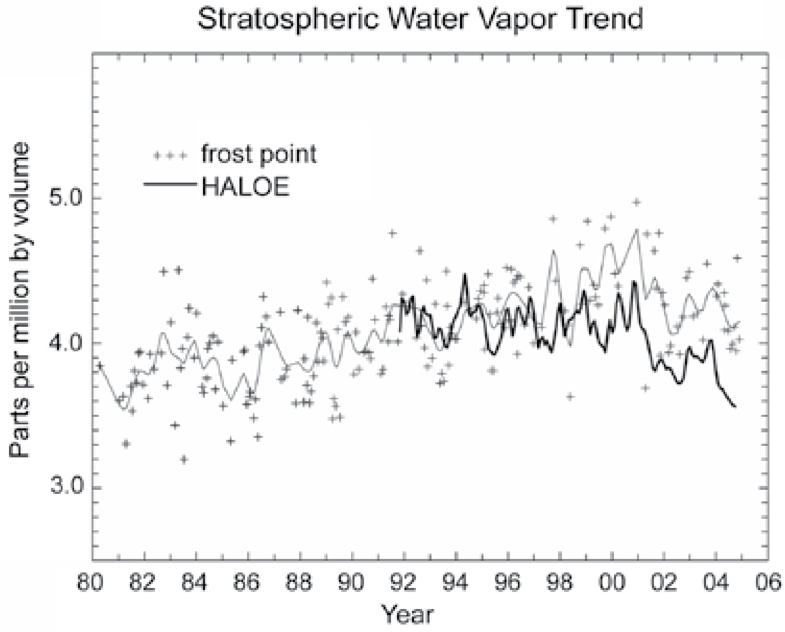
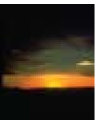


Figure 4.10 Time series of stratospheric water vapor mixing ratios (ppm) for the period 1980 to 2005 showing increases up to about 2001 followed by a decreasing trend. The measurements were made with a balloon-borne frost point hygrometer over Boulder, Colorado (40°N, 105°W). The data points are averages over 17 to 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 to 45°N, longitude 80°W to 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).



Increases in reactive hydrogen (HO_X) and the temperature threshold for polar stratospheric cloud (PSC) formation lead to increased ozone destruction in polar regions at constant ozone-depleting substance (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.

Since about 2000, the water vapor in key balloon and satellite observations in the mid- to lower stratosphere have shown significant decreases (Randel *et al.*, 2004). As a possible explanation, an analysis of the tropical tropopause temperatures for 1992 to 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.

4.3.3.1 RESPONSE OF OZONE TO STRATOSPHERIC WATER VAPOR CHANGES

Increases in stratospheric water lead to increases in reactive hydrogen (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 temperature and location in the stratosphere (Figure 4.9). Model simulations show that a 1 percent increase per 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, 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). Carbon dioxide increases, as previously discussed, 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

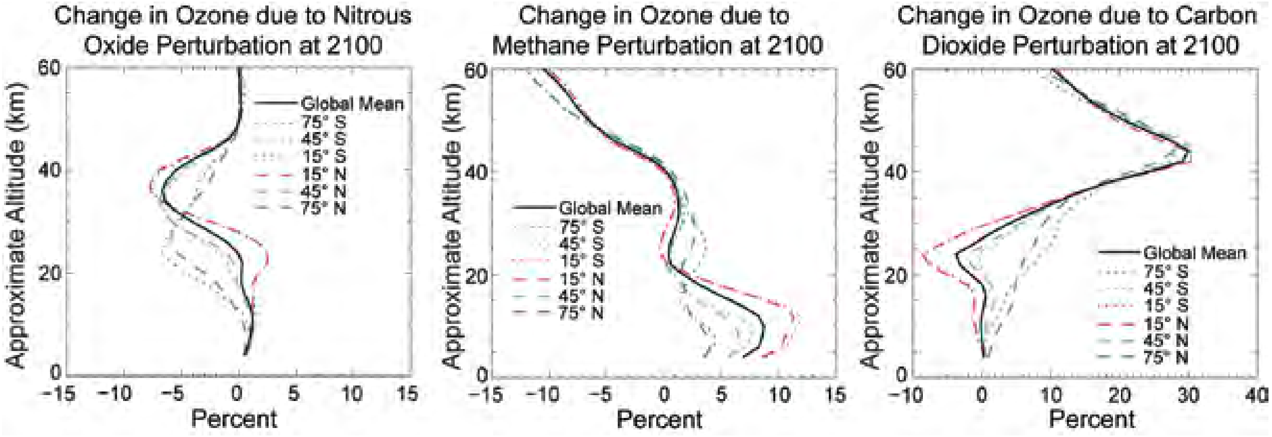


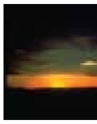
Figure 4.11 Comparison of perturbations to ozone amounts for increases in three principal greenhouse gases, N_2O, CH_4, and CO_2, as a function of altitude in the troposphere, stratosphere, and mesosphere. The changes are computed for six 30°-wide latitude bands with the NOCAR 2-D model and expressed as the percent change from 2000 to 2100. The ozone changes vary widely in magnitude and show increases as well as decreases. The changes result from a variety of chemical and radiative processes. 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 (adapted from Portmann and Solomon, 2007).

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 (Figure 4.7) (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.

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 percent 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 for 2100 with projected climate change (Zeng and Pyle, 2003) shows that a strengthened Brewer-Dobson circulation would increase ozone amounts in the lower stratosphere and the flux of ozone to the troposphere. A larger flux results from increased transport downward across the tropopause and enhanced ozone amounts in the extratropical lower stratosphere. The enhanced ozone results from the strengthened circulation and decreases in ODSs and temperatures.

Model studies indicate that climate change will impact the mass exchange rates across the tropopause.



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. In this section, changes in atmospheric temperatures and circulation are described as examples of the responses in the climate system to ozone depletion.

A combination of increases in greenhouse gases and water vapor and decreases in ozone can account for observed temperature changes in the stratosphere in recent decades.

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 influence of ozone on temperature was carried out with the SKYHI GCM for ozone decreases observed in the period from 1979 to 1997 (Ramaswamy and Schwarzkopf, 2002). The results in Figure 4.12 indicate that in the lower to middle stratosphere (100 to 5 hPa; 21 to 38 km), ozone changes create a larger decrease in temperature than increases in WMGHGs. In this case, these include CO_2, CH_4, N_2O, CFC-11, CFC-12, CFC-113, and HCFC-22. However, above about 5 hPa (about 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.

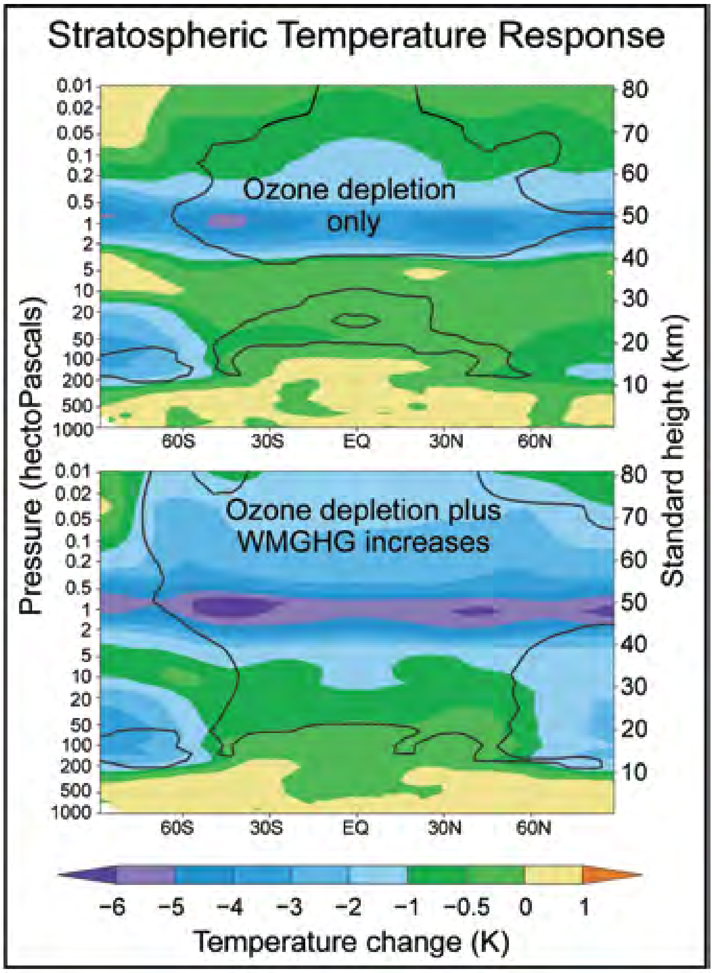
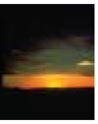


Figure 4.12 The response of stratospheric temperatures calculated with a global climate model (GCM) using observed ozone depletion and changes in well-mixed greenhouse gases (WMGHGs) between 1979 and 1997. The upper panel shows modeled temperature response to ozone changes alone and the lower panel shows the response to ozone changes plus increases in greenhouse gases over the same period. Ozone changes alone cause a significant temperature response in the period. Solid lines enclose regions of statistical significance. The vertical scale shows pressure (left) and altitude (right) (adapted from Ramaswamy and Schwarzkopf, 2002).



The response of temperature to 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 70 to 30 hPa (13 to 25 km) region due to ozone depletion. Shortwave absorption and upwelling longwave radiation are both reduced and contribute comparably to the cooling in this region. The missing ozone also causes a decrease in the downwelling longwave radiation that causes reduced temperatures at altitudes below the ozone depletion region (150 to 100 hPa; 14 to 21 km). This 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.

4.4.2 Response of Surface Temperatures to Antarctic Ozone Depletion

The largest depletion in stratospheric ozone is found over the Antarctic in late winter/early spring. Studies of Antarctic ozone depletion have revealed strong evidence for responses in 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 (a vertical coordinate system referenced to Earth's mean sea level) serving as a proxy. A model with high vertical resolution was used to show that anomalies in geopotential height

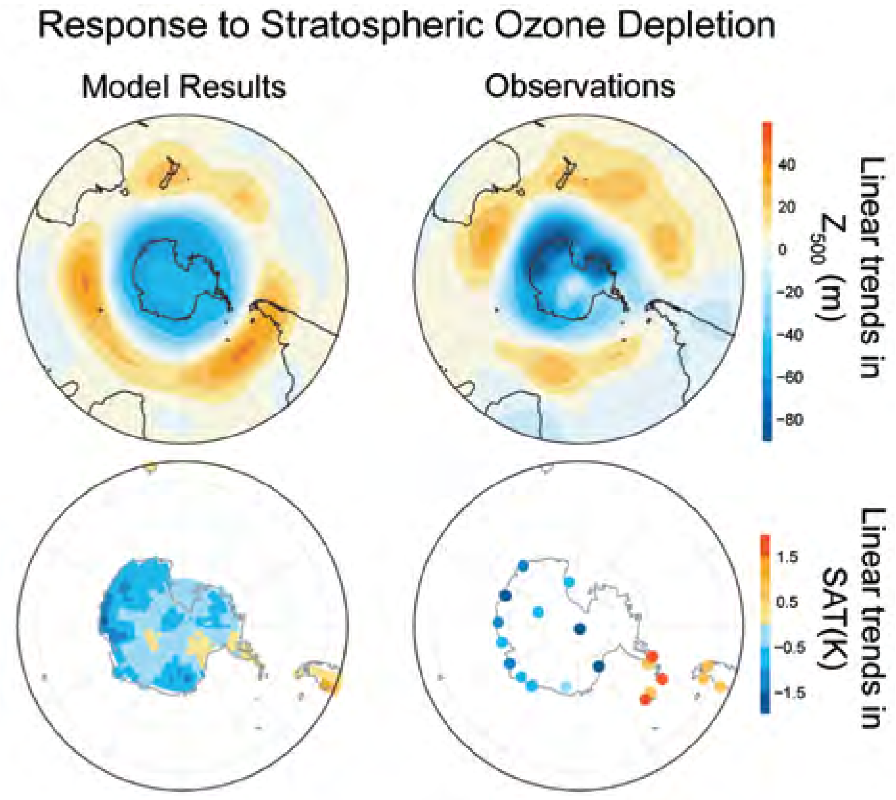
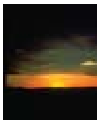


Figure 4.13 Comparison of model results (left column) with observations (right column) for changes in 500-hPa geopotential height (m) (upper row) and near-surface temperature (K) (lower row) in the Southern Hemisphere in response to stratospheric ozone depletion between 1979 and 1997. The observed and 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 (adapted from Thompson and Solomon, 2002).

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 bring about changes in surface climate parameters.

The intense ozone depletion above Antarctica brought about changes in surface climate parameters of that continent.



4.5 IMPORTANCE OF VOLCANOES

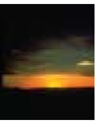
4.5.1 The Effect of Volcanic Aerosol on Ozone

Large volcanic eruptions are those that inject significant quantities of sulfur dioxide (SO_2) into the stratosphere. The 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 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, 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).

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 percent following the 1991 Mt. Pinatubo eruption. Sedimentation and transport removal of volcanic aerosol occurs over a two to three year period following an eruption, so the effects are short lived compared to ODS atmospheric lifetimes, which are 45 to 100 years 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.

A period of frequent volcanic eruptions in the next century could enhance ozone depletion from ozone-depleting substances (ODSs) but the potential for enhancement will lessen as global ODS abundances decline.



The plumes of large volcanic eruptions contain significant amounts of hydrochloric acid (HCl) that are removed in the troposphere by uptake and sedimentation of the 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 percent 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, long term, volcanic enhancements in global stratospheric chlorine.

Volcanic aerosols have a direct radiative impact by scattering incoming solar radiation and absorbing solar infrared radiation. Absorption increases stratospheric temperatures while decreasing surface temperatures. 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 temperature response varies substantially between the different CCM and climate models (see Eyring *et al.*, 2006).

Volcanic eruptions, while not predictable, are expected to occur in 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 (Figure 4.7). A period of frequent large eruptions in the next century could enhance ozone depletion from ODSs for many years but the potential for enhancement will lessen as global ODS abundances decline in the coming decades (Figure 2.12). 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 depletion of stratospheric ozone, as well as 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 human-caused 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 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, expected 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, changes in aerosols and clouds, 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 U.S. 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 hydrochlorofluorocarbon (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.* ODS 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 replacements 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 carbon dioxide, methane, and nitrous oxide emissions.* Ozone and climate are strongly influenced by carbon dioxide, methane and nitrous oxide emissions. Emissions of these gases ultimately affect the photochemical production and loss of ozone in the troposphere and/or stratosphere. All are greenhouse gases that have increased significantly due to human activities.