World Meteorological Organization
Global Ozone Research and Monitoring Project — Report No. 44

SCIENTIFIC ASSESSMENT OF OZONE DEPLETION: 1998

EXECUTIVE SUMMARY

TABLE OF CONTENTS

EXECUTIVE SUMMARY

SCIENTIFIC ASSESSMENT OF OZONE DEPLETION: 1998

PREFACE

The present document contains key summaries from the *Scientific Assessment of Ozone Depletion: 1998*. The full assessment report will be part of the information upon which the Parties to the United Nations Montreal Protocol will base their future decisions regarding protection of the stratospheric ozone layer.

Specifically, the Montreal Protocol on Substances that Deplete the Ozone Layer states (Article 6): ". . . the Parties shall assess the control measures . . . on the basis of available scientific, environmental, technical, and economic information." To provide the mechanisms whereby these assessments are conducted, the Protocol further states: ". . . the Parties shall convene appropriate panels of experts" and "the panels will report their conclusions . . . to the Parties."

Three assessment reports have been prepared during 1998 to be available to the Parties in advance of their meeting in 1999, at which they will consider the need to amend or adjust the Protocol. The two companion reports to the present scientific assessment focus on the environmental and health effects of ozone layer depletion and on the technological feasibilities and economic implications of various mitigation approaches.

The scientific assessment summarized in the present document is the latest in a series of eight scientific assessments prepared by the world's leading experts in the atmospheric sciences and under the international auspices of the World Meteorological Organization (WMO) and/or the United Nations Environment Programme (UNEP). The present assessment is the fifth in the set that has been prepared directly as input to the Montreal Protocol process. The chronology of the scientific assessments on the understanding of ozone depletion and their relation to the international policy process is summarized as follows:

Year	Policy Process	Scientific Assessment
1981		*The Stratosphere 1981. Theory and Measurements.* WMO No. 11.
1985	Vienna Convention	*Atmospheric Ozone 1985.* Three volumes. WMO No. 16.
1987	Montreal Protocol	
1988		*International Ozone Trends Panel Report 1988.* Two volumes. WMO No. 18.
1989		*Scientific Assessment of Stratospheric Ozone: 1989.* Two volumes. WMO No. 20.
1990	London Adjustments and Amendment	
1991		*Scientific Assessment of Ozone Depletion: 1991.* WMO No. 25.
1992		*Methyl Bromide: Its Atmospheric Science, Technology, and Economics* (Assessment Supplement). UNEP (1992).
1992	Copenhagen Adjustments and Amendment	

Year	Policy Process	Scientific Assessment
1994		*Scientific Assessment of Ozone Depletion: 1994.* WMO No. 37.
1995	Vienna Adjustment	
1997	Montreal Adjustments and Amendment	
1998		*Scientific Assessment of Ozone Depletion: 1998.* WMO No. 44.
1999	11th Meeting of the Parties (China)	

The genesis of *Scientific Assessment of Ozone Depletion: 1998* occurred at the 7th Meeting of the Conference of the Parties to the Montreal Protocol in Vienna, Austria, in December 1995, at which the scope of the scientific needs of the Parties was defined. The formal planning of the present report was started in January 1997 by an *ad hoc* international steering group who crafted the outline and suggested scientists from the world community to serve as authors. The first drafts of the chapters were examined at a meeting that occurred on 12 - 14 November 1997 in Washington, D.C., at which the Lead Authors and a small number of international experts focused on the content of the draft chapter and the coordination among the chapters.

The second drafts of the chapters were reviewed by 124 scientists worldwide in a mail peer review. These comments were considered by the authors. At a Panel Review Meeting in Les Diablerets, Switzerland, held on 1 - 5 June 1998, the responses to these mail review comments were proposed by the authors and discussed by the 73 participants. Final changes to the chapters were decided upon there, and the Executive Summary contained herein was prepared by the participants.

The group also focused on updating a set of questions that are frequently asked about the ozone layer. Based upon the scientific understanding represented by the assessments, answers to these frequently asked questions were updated. These questions and answers are included in this report.

As the accompanying list indicates, the *Scientific Assessment of Ozone Depletion: 1998* is the product of 304 scientists from the developed and developing world[1] who contributed to its preparation and review (218 scientists prepared the report and 148 scientists participated in the peer review process).

What follows is a summary of their current understanding of the stratospheric ozone layer and its relation to humankind.

---

[1] Participating were Argentina, Australia, Austria, Belgium, Brazil, Canada, Chile, Czech Republic, Denmark, Egypt, Finland, France, Germany, Greece, India, Iran, Ireland, Italy, Japan, Kenya, Malaysia, New Zealand, Norway, Poland, Russia, South Africa, Republic of Korea, Sweden, Switzerland, Taiwan, The Netherlands, The People's Republic of China, United Kingdom, United States of America, and Venezuela.

EXECUTIVE SUMMARY

The 1987 Montreal Protocol on Substances that Deplete the Ozone Layer commemorated its 10th anniversary in September 1997. Among the provisions of the Protocol was the requirement that the Parties to the Protocol base their future decisions on the available scientific, environmental, technical, and economic information as assessed by the worldwide expert communities. The advances of the understanding in ozone science over this decade were assessed in 1988, 1989, 1991, and 1994. This information was input to the subsequent Amendments and Adjustments of the 1987 Protocol. The Assessment summarized here is the fifth in that series.

Recent Major Scientific Findings and Observations

Since the *Scientific Assessment of Ozone Depletion: 1994*, significant advances have continued to be made in the understanding of the impact of human activities on the ozone layer, the influence of changes in chemical composition on the radiative balance of the Earth's climate, and, indeed, the coupling of the ozone layer and the climate system. Numerous laboratory investigations, atmospheric observations, and theoretical and modeling studies have produced several key ozone- and climate-related findings:

- **The total combined abundance of ozone-depleting compounds in the lower atmosphere peaked in about 1994 and is now slowly declining. Total chlorine is declining, but total bromine is still increasing.** As forecast in the 1994 Assessment, the long period of increasing total chlorine abundances – primarily from the chlorofluorocarbons (CFCs), carbon tetrachloride (CCl_4), and methyl chloroform (CH_3CCl_3) – has ended. The peak total tropospheric chlorine abundance was 3.7 ± 0.1 parts per billion (ppb) between mid-1992 and mid-1994. The declining abundance of total chlorine is due principally to reduced emissions of methyl chloroform. Chlorine from the major CFCs is still increasing slightly. The abundances of most of the halons continue to increase (for example, Halon-1211, almost 6% per year in 1996), but the rate has slowed in recent years. These halon increases are likely to be due to emissions in the 1990s from the halon "bank," largely in developed countries, and new production of halons in developing countries. The observed abundances of CFCs and chlorocarbons in the lower atmosphere are consistent with reported emissions.

- **The observed abundances of the substitutes for the CFCs are increasing.** The abundances of the hydrochlorofluorocarbons (HCFCs) and hydrofluorocarbons (HFCs) are increasing as a result of a continuation of earlier uses and of their use as substitutes for the CFCs. In 1996, the HCFCs contributed about 5% to the tropospheric chlorine from the long-lived gases. This addition from the substitutes offsets some of the decline in tropospheric chlorine associated with methyl chloroform, but is nevertheless about 10 times less than that from the total tropospheric chlorine growth rate throughout the 1980s. The atmospheric abundances of HCFC-141b and HCFC-142b calculated from reported emissions data are factors of 1.3 and 2, respectively, smaller than observations. Observed and calculated abundances agree for HCFC-22 and HFC-134a.

- **The combined abundance of stratospheric chlorine and bromine is expected to peak before the year 2000.** The delay in this peak in the stratosphere compared with the lower atmosphere reflects the average time required for surface emissions to reach the lower stratosphere. The observations of key chlorine compounds in the stratosphere up through the present show the expected slower rate of increase and show that the peak had not occurred at the time of the most recent observations that were analyzed for this Assessment.

- **The role of methyl bromide as an ozone-depleting compound is now considered to be less than was estimated in the 1994 Assessment, although significant uncertainties remain.** The current best estimate of the Ozone Depletion Potential (ODP) for methyl bromide (CH_3Br) is 0.4, compared with an ODP of 0.6 estimated in the previous Assessment. The change is due primarily to both an increase in the estimate of ocean removal processes and the identification of an uptake by soils, with a smaller contribution from the change in our estimate of the atmospheric removal rate. Recent research has shown that the science of atmospheric methyl bromide is complex and still not well understood. The current understanding of the sources and sinks of atmospheric methyl bromide is incomplete.

- **The rate of decline in stratospheric ozone at midlatitudes has slowed; hence, the projections of ozone loss made in the 1994 Assessment are larger than what has actually occurred.** Total column ozone decreased significantly at midlatitudes (25-60˚) between 1979 and 1991, with estimated linear downward trends of 4.0, 1.8, and 3.8% per decade, respectively, for northern midlatitudes in winter/spring, northern midlatitudes in summer/fall, and southern midlatitudes year round. However, since 1991 the linear trend observed during the 1980s has not continued, but rather total column ozone has been almost constant at midlatitudes in both hemispheres since the recovery from the 1991 Mt. Pinatubo eruption. The observed total column ozone losses from 1979 to the period 1994-1997 are about 5.4, 2.8, and 5.0%, respectively, for northern midlatitudes in winter/spring, northern midlatitudes in summer/fall, and southern midlatitudes year round, rather than the values projected in the 1994 Assessment assuming a linear trend: 7.6, 3.4, and 7.2%, respectively. The understanding of how changes in stratospheric chlorine/bromine and aerosol loading affect ozone suggests some of the reasons for the unsuitability of using a linear extrapolation of the pre-1991 ozone trend to the present.

- **The link between the long-term build-up of chlorine and the decline of ozone in the upper stratosphere has been firmly established.** Model predictions based on the observed build-up of stratospheric chlorine in the upper stratosphere indicate a depletion of ozone that is in good quantitative agreement with the altitude and latitude dependence of the measured ozone decline during the past several decades, which peaks at about 7% per decade near 40 km at midlatitudes in both hemispheres.

- **The springtime Antarctic ozone hole continues unabated.** The extent of ozone depletion has remained essentially unchanged since the early 1990s. This behavior is expected given the near-complete destruction of ozone within the Antarctic lower stratosphere during springtime. The factors contributing to the continuing depletion are well understood.

- **The late-winter/spring ozone values in the Arctic were unusually low in 6 out of the last 9 years, the 6 being years that are characterized by unusually cold and protracted stratospheric winters.** The possibility of such depletions was predicted in the 1989 Assessment. Minimum Arctic vortex temperatures are near the threshold for large chlorine activation. Therefore, the year-to-year variability in temperature, which is driven by meteorology, leads to particularly large variability in ozone for current chlorine loading. As a result, it is not possible to forecast the behavior of Arctic ozone for a particular year. Elevated stratospheric halogen abundances over the next decade or so imply that the Arctic will continue to be vulnerable to large ozone losses.

- **The understanding of the relation between increasing surface UV-B radiation and decreasing column ozone has been further strengthened by ground-based observations, and newly developed satellite methods show promise for establishing global trends in UV radiation.** The inverse dependence of surface UV radiation and the overhead amount of ozone, which was demonstrated in earlier Assessments, has been further demonstrated and quantified by ground-based measurements under a wide range of atmospheric conditions. In addition, the influences of other variables, such as clouds, particles, and surface reflectivity, are better understood. These data have assisted the development of a satellite-based method to estimate global UV changes, taking into account the role of cloud cover. The satellite estimates for 1979-1992 indicate that the largest UV increases occur during spring at high latitudes in both hemispheres.

• **Stratospheric ozone losses have caused a cooling of the global lower stratosphere and global-average negative radiative forcing of the climate system.** The decadal temperature trends in the stratosphere have now been better quantified. Model simulations indicate that much of the observed downward trend in lower stratospheric temperatures (about 0.6°C per decade over 1979-1994) is attributed to the ozone loss in the lower stratosphere. A lower stratosphere that is cooler results in less infrared radiation reaching the surface/troposphere system. Radiative calculations, using extrapolations based on the ozone trends reported in the 1994 Assessment for reference, indicate that stratospheric ozone losses since 1980 may have offset about 30% of the positive forcing due to increases in the well-mixed greenhouse gases (i.e., carbon dioxide, methane, nitrous oxide, and the halocarbons) over the same time period. The climatic impact of the slowing of midlatitude ozone trends and the enhanced ozone loss in the Arctic has not yet been assessed.

• **Based on past emissions of ozone-depleting substances and a projection of the maximum allowances under the Montreal Protocol into the future, the maximum ozone depletion is estimated to lie within the current decade or the next two decades, but its identification and the evidence for the recovery of the ozone layer lie still further ahead.** The falloff of total chlorine and bromine abundances in the stratosphere in the next century will be much slower than the rate of increase observed in past decades, because of the slow rate at which natural processes remove these compounds from the stratosphere. The most vulnerable period for ozone depletion will be extended into the coming decades. However, extreme perturbations, such as natural events like volcanic eruptions, could enhance the loss from ozone-depleting chemicals. Detection of the beginning of the recovery of the ozone layer could be achievable early in the next century if decreasing chlorine and bromine abundances were the only factor. However, potential future increases or decreases in other gases important in ozone chemistry (such as nitrous oxide, methane, and water vapor) and climate change will influence the recovery of the ozone layer. When combined with the natural variability of the ozone layer, these factors imply that unambiguous detection of the beginning of the recovery of the ozone layer is expected to be well after the maximum stratospheric loading of ozone-depleting gases.

Supporting Scientific Evidence and Related Issues

RECENT HALOGEN AND METHANE CHANGES

• Tropospheric abundances of total organic chlorine (Cl) contained in long- and short-lived halocarbons reached maximum values of 3.7 ± 0.1 parts per billion (ppb) between mid-1992 and mid-1994 and are beginning to decrease slowly in the global troposphere. The decline in the tropospheric abundance of methyl chloroform (CH_3CCl_3) (at a rate of about 40 to 42 parts per trillion (ppt) Cl yr^{-1} in 1996) is the principal cause of the decrease and reversal in the Cl growth rate. At the same time, chlorine from the sum of the major CFCs grew at 7 ppt Cl yr^{-1} (CFC-12, 9 ppt Cl yr^{-1}; CFC-11, -2 ppt Cl yr^{-1}; CFC-113, 0 ppt Cl yr^{-1}) and by 10 ppt Cl yr^{-1} from the three major hydrochlorofluorocarbons (HCFCs) (HCFC-22, 5 ppt Cl yr^{-1}; HCFC-141b, 4 ppt Cl yr^{-1}; HCFC-142b, 1 ppt Cl yr^{-1}). The rate of decay of CH_3CCl_3 is expected to slow down to less than 10 ppt Cl yr^{-1} by 2005. By that point its concentration should be so small that it will no longer be an important contributor to atmospheric organic chlorine.

• Space-based remote measurements of hydrogen chloride (HCl), hydrogen fluoride (HF), and total chlorine in the stratosphere, as well as column abundances of HCl, chlorine nitrate ($ClONO_2$), HF, and carbonyl difluoride (COF_2) from the ground, are consistent with the content and rate of change of the total organic chlorine and fluorine abundance of the troposphere. These observations provide evidence that the rate of increase of stratospheric chlorine loading has slowed in recent years.

- Growth in the tropospheric concentrations of HCFCs and hydrofluorocarbons (HFCs) has been observed as expected from continuation of previous uses and from their use as replacements for chlorofluorocarbons (CFCs). Emissions calculated by industry from sales and use data are in accordance with the current global abundances of HCFC-22 and HFC-134a. For HCFC-141b and -142b, the industry data underestimate the current global abundances by factors of approximately 1.3 and 2 respectively. No production and sales data are currently available for other HCFCs and HFCs being used as CFC alternatives.

- New studies suggest a major reduction in the magnitude of the estimated oceanic source of methyl chloride (CH_3Cl). As a result, the sum of known sources is inadequate to explain the observed atmospheric burden of CH_3Cl, thus requiring a larger contribution from other sources, either natural or anthropogenic.

- Tropospheric bromine loading continues to rise largely because of the ongoing growth of Halon-1211 (almost 6% yr^{-1}), Halon-2402 (2% yr^{-1}), and Halon-1301 (1% yr^{-1}). Possible causes are the large "banking" in developed countries of that compound during the 1980s and its subsequent use and release during the 1990s, and new production in developing countries. Continued increases of halons over the next few years could cause the abundance of equivalent chlorine to decline more slowly than predicted in the 1994 Assessment.

- Recent measurements and intercomparisons of calibration standards have confirmed that the average global mixing ratio of methyl bromide (CH_3Br) is between 9 and 10 ppt and that the interhemispheric ratio is 1.3 ± 0.1 (north/south). New estimates of methyl bromide losses yield magnitudes of 77 Gg yr^{-1} (ranging from 37 to 133 Gg yr^{-1}) for ocean uptake; 42 Gg yr^{-1} (ranging from 10 to 214 Gg yr^{-1}) for soil uptake; and 86 Gg yr^{-1} (ranging from 65 to 107 Gg yr^{-1}) for removal by hydroxyl radical (OH), for a total removal rate of 205 Gg yr^{-1} with a range of about 110 to 450 Gg yr^{-1}. The current best estimate of the lifetime of atmospheric CH_3Br, as calculated from losses within the atmosphere, to the ocean, and to soils, is 0.7 years, with a range of 0.4 to 0.9 years. The Ozone Depletion Potential (ODP) of methyl bromide is 0.4, with a range of 0.2 to 0.5.

- No new important sources of methyl bromide have been identified. The ocean now appears to be a net sink, with an estimated net flux from the atmosphere of -21 Gg yr^{-1} (ranging from -3 to -32 Gg yr^{-1}). Estimates of ocean emissions of order 60 Gg yr^{-1} can be directly deduced from the above estimates for uptake and net ocean flux. The total emission of CH_3Br from identified sources is 122 Gg yr^{-1}, with a range of 43 to 244 Gg yr^{-1}. The best-quantified source is fumigation, with a magnitude of 41 Gg yr^{-1} and a range of 28 to 64 Gg yr^{-1}. Other anthropogenic sources include biomass burning (20 Gg yr^{-1}, ranging from 10 to 40 Gg yr^{-1}) and leaded gasoline use (5 Gg yr^{-1}, ranging from negligible to 10 Gg yr^{-1}). Identified sources of CH_3Br thus constitute only about 60% of identified sinks on a globally averaged basis. This disagreement is difficult to reconcile with estimated uncertainties in the source and sink terms. The short lifetime of methyl bromide, coupled with the inhomogeneity of its sources and sinks, complicates the interpretation of its global budget.

- Based on the most recent analysis of the methyl chloroform (CH_3CCl_3) observational record (including a refinement in calibration), the estimated atmospheric lifetimes (with respect to reactive removal by OH) of CH_3CCl_3, HCFCs, HFCs, and CH_4 have been reduced by about 15% since the 1994 Assessment. The 1995 assessment of the Intergovernmental Panel on Climate Change (IPCC) mostly reflected these revisions, with a slightly smaller correction factor of about 10%. For species whose chemical lifetime is shorter than 1 to 2 years, the use of a global-mean lifetime may not be appropriate.

- The atmospheric abundance of CH_4 continues to increase, but with a declining growth rate. The average growth rate between 1980 and 1992 of about 10 ppb yr^{-1} can be compared with the 1996-1997 rate of approximately 3 to 4 ppb yr^{-1}. The current best estimate for the total atmospheric lifetime of methane has been lowered to 8.9 ± 0.6 years.

STRATOSPHERIC PARTICLES

• Observations and models have further confirmed that stratospheric sulfate aerosol (SSA) and polar stratospheric clouds (PSCs) play a key role in ozone loss chemistry through heterogeneous reactions that activate halogen species and deactivate nitrogen species.

• Observations have increased our knowledge of particle formation processes, the dispersal and decay of volcanic SSA, and particle climatology. They show that supercooled ternary solution (STS) droplets that form from SSA without a nucleation barrier are an important class of PSC particles. The formation processes of solid PSC particles that play a significant role in denitrification of the polar vortices remain uncertain. Recent studies suggest that mesoscale temperature fluctuations, especially over mountain ranges, may be important in PSC formation processes, particularly in the Arctic.

• The two most recent major volcanic eruptions, El Chichón (1982) and Mt. Pinatubo (1991), both temporarily increased SSA amounts by more than an order of magnitude.

• There is no clear trend in SSA abundances from 1979 to 1997, demonstrating that any anthropogenic contribution must be smaller than thought in the 1994 Assessment. SSA models including known tropospheric sulfur sources underpredict 1979 values, which were thought to represent the non-volcanic background, but it is not clear that this period was truly free of volcanic influence.

OZONE IN THE MIDLATITUDES AND TROPICS

• As noted in the 1994 Assessment, Northern Hemisphere midlatitude column ozone decreased markedly in 1992-1993, following the large enhancement of stratospheric aerosol caused by the eruption of Mt. Pinatubo in 1991. Column ozone has now reached amounts higher than a linear extrapolation of the pre-Pinatubo trend would predict. Between 25 and 60˚N, ozone abundances for 1994-1997 averaged about 4% below 1979 values, although with large variability, while extrapolation of the pre-1991 trend would predict current (1997) abundances about 5.5% below 1979 values. The corresponding winter/spring and summer/fall losses average about 5.4 and 2.8%, respectively, while a linear extrapolation would predict 7.6 and 3.4%, respectively. The average ozone abundances between 25 and 60˚S are currently about 4% (satellite) or 5% (ground) below 1979 values, while the linear extrapolation would predict 7.2% (both satellite and ground).

• Our understanding of how changes in halogen and aerosol loading affect ozone suggests some of the reasons for the unsuitability of using a linear extrapolation of the pre-1991 ozone trend to the present. For example, observations of stratospheric HCl and $ClONO_2$ show a build-up of stratospheric chlorine in recent years consistent with halocarbon emissions, but slower than would have been predicted by the chlorine trends observed before 1992. In addition, enhanced stratospheric aerosol was also present throughout much of the decade of the 1980s due to earlier volcanic eruptions (e.g., El Chichón and Ruiz), likely enhancing the downward trend of ozone observed even before Pinatubo.

• There are no statistically significant trends in total ozone in the equatorial regions (20˚S to 20˚N).

• The amplitude of the annual cycle of ozone at middle to high latitudes has decreased by approximately 15% in the last decades because larger declines have occurred during the season of maximum ozone values.

• For northern midlatitudes, combined vertical profile ozone trends through 1996 are negative at all altitudes between 12 and 45 km and are statistically significant at the 2σ level. The downward trend is largest near 40 and 15 km (approximately 7% per decade) and is smallest at 30 km (2% per decade). The bulk of column ozone decline is between the tropopause and 25 km.

- The re-evaluation of the Stratospheric Aerosol and Gas Experiment (SAGE) I/II satellite data indicates that there are no significant interhemispheric differences in upper stratospheric trends through 1996. Agreement is good, within estimated uncertainties, between SAGE I/II and ozonesonde trends in the lower to middle stratosphere in northern midlatitudes.

- The total ozone and the vertical profile trends derived for the northern midlatitudes are consistent with each other over the periods studied.

- Most of the midlatitude column ozone decline during the last two decades arose because of depletion in the lower stratosphere. That region is influenced by local chemical ozone loss that is enhanced by volcanic aerosol, and by transport from other regions. The vertical, latitudinal, and seasonal characteristics of the depletion of midlatitude ozone are broadly consistent with the understanding that halogens are the primary cause. The expected low ozone amounts in the midlatitude lower stratosphere following the Mt. Pinatubo eruption further strengthened the connection between ozone destruction and anthropogenic chlorine.

- Models that represent processes affecting ozone are able to calculate variations in ozone abundances that are broadly consistent with the observed midlatitude column ozone trend as well as the response to volcanic enhancement of stratospheric sulfate aerosol. In particular, models reproduce the lower ozone abundances observed immediately following Mt. Pinatubo and the subsequent increases as the aerosol disappeared.

- Current two-dimensional (2-D) assessment models that allow for the observed build-up of stratospheric chlorine calculate reductions in ozone that are in good quantitative agreement with the altitude and latitude dependence of the measured decline in upper stratospheric ozone during the past several decades. This clearly confirms the hypothesis put forth in 1974 that release of CFCs to the atmosphere would lead to a significant reduction of upper stratospheric ozone, with the peak percentage decline occurring around 40 km.

- Comparison of recent observations and model results shows that the overall partitioning of reactive nitrogen and chlorine species is well understood for the upper stratosphere. The previously noted discrepancy for the chlorine monoxide/hydrogen chloride (ClO/HCl) ratio has been resolved based on new kinetic information. Balloonborne observations of OH and hydroperoxyl radicals (HO$_2$) agree well with theory, but satellite and ground-based observations of these species exhibit systematic differences compared with model calculations.

- An improved understanding of the relevant kinetic processes has resulted in a close balance between the calculated production and loss of ozone at 40 km (i.e., the long-standing difference between calculated and observed ozone abundance has been mostly resolved).

- Constituent measurements show that the tropics are relatively isolated from midlatitudes in the lower stratosphere. The extent of isolation affects the budgets (and lifetimes) of chemical species that affect ozone abundance.

Ozone in High-Latitude Polar Regions

- The large ozone losses in the Southern Hemisphere polar region during spring continued unabated with approximately the same magnitude and areal extent as in the early 1990s. In Antarctica, the monthly total ozone in September and October has continued to be 40 to 55% below the pre-ozone-hole values of approximately 320 m-atm cm ("Dobson units"), with up to a 70% decrease for periods of a week or so. This depletion occurs primarily over the 12- to 20-km altitude range, with most of the ozone in this layer disappearing during early October. These ozone changes are consistent overall with our understanding of chemistry and dynamics.

- In the Arctic vortex, low column ozone values were observed in the late-winter/spring for 6 out of the last 9 years. Monthly mean values were about 100 m-atm cm below 1960-1970 averages, with shorter-period differences exceeding 200 m-atm cm (equivalent to about 20 to 45% of values found in the 1960s and early 1970s). Within the column, the largest ozone differences were observed in the lower stratosphere.

- Years with large seasonal ozone depletion in the late-winter/spring Arctic are characterized by specific meteorological conditions. These conditions are lower-than-normal late-winter Arctic temperatures, which lead to enhanced activated chlorine, and a more isolated vortex and weaker planetary-wave driving, which lead to less transport of ozone-rich air into the Arctic. Low temperatures, an isolated vortex, and reduced wave driving are coupled processes that occur in concert in the stratosphere. Chemical ozone losses have been identified within the Arctic vortex and are associated with activated chlorine augmented by bromine. The total seasonal chemical ozone losses within the vortex have been estimated to be approximately 100 m-atm cm.

- With the present high abundances of chlorine loading, late-winter/spring Arctic chemical ozone loss is particularly sensitive to meteorological conditions (temperature and vortex isolation) because minimum vortex temperatures are at a critical value in terms of activating chlorine. Winter vortex temperatures in the 1990s have been particularly low. In the absence of low temperatures and an isolated vortex, reduced chemical ozone loss would be expected. However, such a reduced ozone loss would not indicate chemical recovery. The Arctic will remain vulnerable to extreme seasonal loss as long as chlorine loading remains high.

- Chlorine activation in liquid particles in the lower stratosphere (both SSA and liquid PSCs) increases strongly with decreases in temperature and is at least as effective as that on solid particles. Thus, chlorine activation is to a first approximation controlled by temperature and water vapor pressure and only secondarily by particle composition.

- Rapid polar ozone loss requires enhanced chlorine monoxide in the presence of sunlight. Maintenance of elevated ClO in late-winter/spring is dependent upon temperature and requires either repeated heterogeneous processing or denitrification. Since the 1994 Assessment, new understanding has shown that cold liquid aerosol can maintain elevated ClO in non-denitrified air.

STRATOSPHERIC TEMPERATURES

- Radiosonde and satellite observations indicate a decadal cooling trend of the global, annual-mean lower stratosphere (approximately 16 to 21 km) since about 1980. Over the period 1979 to 1994, its amplitude is approximately 0.6°C per decade. At midlatitudes the trend is larger (approximately 0.75°C per decade) and broadly coherent among the various datasets with regard to the magnitude and statistical significance.

- Substantial cooling (approximately 3°C per decade) is observed in the polar lower stratosphere during late-winter/spring in both hemispheres. A decadal-scale cooling is evident in the Antarctic since the early 1980s and in the Arctic since the early 1990s. However, the dynamical variability is large in these regions, particularly in the Arctic, and this introduces difficulties in establishing the statistical significance of trends.

- The vertical profile of the annual-mean stratospheric temperature change observed in the Northern Hemisphere midlatitudes is robust for the 1979-1994 period within the different datasets. The trend consists of an approximately 0.75°C per decade cooling of the 15- to 35-km region, a slight reduction in the cooling at about 35 km, and increased cooling with height above 35 km (approximately 2°C per decade at 50 km).

- Model simulations based on known changes in the stratospheric concentrations of various radiatively active species indicate that the depletion of lower stratospheric ozone is the dominant radiative factor in the explanation of the observed global-mean lower stratospheric cooling trends for the period 1979-1990 (approximately 0.5°C per decade). The contribution to these trends from increases in well-mixed greenhouse gases is estimated to be less than one-fourth that due to ozone loss.

- Model simulations indicate that ozone depletion is an important causal factor in the latitude-month pattern of the decadal (1979-1990) lower stratospheric cooling. The simulated lower stratosphere in Northern and Southern Hemisphere midlatitudes and in the Antarctic springtime generally exhibit a statistically significant cooling trend over this period consistent with observations.

- In the middle and upper stratosphere, both the well-mixed greenhouse gases and ozone change contribute in an important manner to the cooling. However, the computed cooling due to these gases underestimates the observed decadal trend.

Tropospheric Ozone

- Trends in tropospheric ozone since 1970 in the Northern Hemisphere show large regional differences, with increases in Europe and Japan, decreases in Canada, and only small changes in the United States. The trend in Europe since the mid-1980s has reduced to virtually zero (at two recording stations). In the Southern Hemisphere, small increases have now been observed in surface ozone.

- Recent field studies have shown that anthropogenic emissions of ozone precursors (nitrogen oxides, carbon monoxide, and hydrocarbons) lead to large-scale production of ozone, which, through long-range transport, influences the ozone concentration in large regions of the troposphere in both hemispheres. However, significant uncertainties remain in the budget of tropospheric ozone, its precursors, and the chemical and physical processes involved. Large spatial and temporal variability is observed in tropospheric ozone, resulting from important regional differences in the factors controlling its concentration.

- Important improvements in global chemical transport models (CTMs) have allowed better simulations of tropospheric ozone distributions and of ozone perturbations resulting from anthropogenic emissions.

- Considerable progress has been made in testing tropospheric photochemistry through field measurements. Our theoretical understanding of tropospheric OH is nevertheless incomplete, specifically in regard to sources of upper tropospheric OH and polluted conditions.

- Increases in air traffic and the resulting emissions could have impacts on atmospheric chemistry and cloud formation, with implications for the ozone layer and the climate system. The understanding of the effects of aircraft emissions are currently being assessed as part of the Intergovernmental Panel on Climate Change (IPCC) special report *Aviation and the Global Atmosphere: 1999*. Consequently, this topic is not included in the scope of the present Assessment.

Changes in UV Radiation

- The inverse correlation between ozone column amounts and ultraviolet-B (UV-B) irradiance has been reconfirmed and firmly established by numerous ground-based measurements. The ground-based measurements have increased our understanding of additional effects such as albedo, altitude, clouds and aerosols, and geographic differences on UV irradiance at the Earth's surface.

- A controversy concerning anomalous UV-trend estimates from the Robertson-Berger (RB) meter network located in the continental United States. (1974-1985) has been explained in terms of poor calibration stability. The reanalysis of this U.S. RB-meter dataset shows that the errors are too large for determining UV-irradiance trends over that period.

- Increases in UV-B irradiance (e.g., 1989-1997; 1.5% yr^{-1} at 300 nm, 0.8% yr^{-1} at 305 nm) have been detected with a few ground-based spectroradiometers at midlatitudes (near 40˚) and are consistent with expected changes from the decreasing amounts of ozone. Although these UV changes are consistent with those estimated from satellite data, the ground-based data records from suitably stable and calibrated instruments are not yet long enough to determine decadal trends. Local irradiance changes, not seen in the coarse-spatial-resolution satellite data, caused by pollution and aerosols have been detected in both UV-B (280 to 315 nm) and UV-A (315 to 400 nm).

- New satellite estimates of global (±65˚) UV irradiance that now include cloud, surface reflectivity, and aerosol effects have been estimated from measured backscattered radiances from the Total Ozone Mapping Spectrometer (TOMS) using radiative transfer models. Climatological maps of UV irradiance can be produced from the daily data. In addition, the satellite data have been used to estimate zonally averaged global and seasonal trends in UV irradiance from 1979 to 1992. For this period, annual erythemal UV-irradiance decadal increases were estimated to be $3.7 \pm 3\%$ at 60˚N and $3 \pm 2.8\%$ at 40˚N. Larger decadal increases were observed in the Southern Hemisphere: $3.6 \pm 2\%$ at 40˚S and $9 \pm 6\%$ at 60˚S. No statistically significant trends were observed between ±30˚ latitude. Zonally averaged UV-A irradiances have not changed.

- Current zonal-average UV-irradiance trend estimations from satellite data that include cloud effects are nearly identical to clear-sky estimates. The currently estimated trends are slightly lower than the clear-sky trend estimates in the 1994 Assessment because of the new TOMS retrieval algorithm.

- Instrument intercomparison and newly developed calibration and database centers have improved the quality and availability of ground-based data.

CHANGES IN CLIMATE PARAMETERS

- Increased penetration of UV radiation to the troposphere as a result of stratospheric ozone depletion influences key photochemical processes in the troposphere. Model results suggest that a 1% decrease in global total ozone leads to a global increase of 0.7 to 1% in globally averaged tropospheric OH, which would affect the lifetimes of several climate-related gases.

- The global average radiative forcing due to changes in stratospheric ozone since the late 1970s, using extrapolations based on the ozone trends reported in the 1994 Assessment for reference, is estimated to be -0.2 ± 0.15 Wm^{-2}, which offsets about 30% of the forcing due to increases in other greenhouse gases over the same period. The climatic impact of the slowing of midlatitude trends and the enhanced ozone loss in the Arctic has not yet been assessed. Recovery of stratospheric ozone would reduce the offset to the radiative forcing of the other greenhouse gases. The ozone recovery will therefore lead to a more rapid increase in radiative forcing than would have occurred due to increases in other greenhouse gases alone.

- The global average radiative forcing due to increases in tropospheric ozone since preindustrial times is estimated to be $+0.35 \pm 0.15$ Wm^{-2}, which is about 10 to 20% of the forcing due to long-lived greenhouse gases over the same period.

- Coupled ocean-atmosphere general circulation models (GCMs) have been used to calculate the impact of stratospheric ozone loss on the thermal structure of the atmosphere. The calculated altitude of the transition from tropospheric warming to stratospheric cooling due to increases in well-mixed greenhouse gases is in better agreement with observations when ozone depletion is taken into account.

- Radiative forcings and Global Warming Potentials (GWPs) are now available for an expanded set of gases. New categories include fluorinated organic molecules. The CFC-11 radiative forcing has been revised by +12% from the value used since IPCC (1990), primarily because of the use of an improved vertical profile of CFC-11 mixing ratio. This and other updates lead to GWPs relative to CO_2 that are typically 20% higher than those in IPCC (1995).

FUTURE HALOGEN CHANGES

- Large reductions in the production and atmospheric release of ozone-depleting substances (ODSs) have been achieved by international regulations (Montreal Protocol and its Amendments and Adjustments). Without such controls, and assuming a (conservative) 3% annual growth rate in production, ODSs would have led to an equivalent effective chlorine loading of around 17 ppb in 2050. The control measures of the original Montreal Protocol (1987) reduce this to approximately 9 ppb; the Amendments of London (1990) to about 4.6 ppb; and the Amendments of Copenhagen (1992) to approximately 2.2 ppb (but with stratospheric halogen loading increasing again in the second half of the 21st century). The Adjustments of Vienna (1995) and the Amendments of Montreal (1997) further reduce this to about 2.0 ppb (approximately the 1980 abundance) around the year 2050.

- Stratospheric halogen loading lags tropospheric loading by up to 6 years. Given that tropospheric halogen loading peaked around 1994 and assuming a scenario with a 3-yr lag time, the equivalent effective stratospheric chlorine loading is estimated to have peaked in 1997, at an abundance 1.7 times higher than in 1980. If annual ozone trends observed in the 1980s are attributed solely to these halogen increases, the peak ozone reductions in 1997, relative to 1980, are estimated to be about 5% at 45°N and 6% at 45°S. The corresponding increases in erythemally weighted UV radiation in 1997 are estimated to be 5% at 45°N and 8% at 45°S relative to the 1980 values.

RECOVERY OF THE OZONE LAYER

- In the absence of other changes, stratospheric ozone abundances should rise in the future as the halogen loading falls in response to regulation. However, the future behavior of ozone will also be affected by the changing atmospheric abundances of methane (CH_4), nitrous oxide (N_2O), water vapor (H_2O), sulfate aerosol, and changing climate. Thus, for a given halogen loading in the future, the atmospheric ozone abundance may not be the same as found in the past for that same halogen loading.

- Several two-dimensional models were used to look at the response of ozone to past and future changes in atmospheric composition. Future global ozone abundances are predicted to recover only slowly toward their 1980 values. The return toward 1980 ozone values in the models depends sensitively on the emission scenarios used. The CH_4 scenario used here has a lower growth rate than in previous assessments, which slows the modeled ozone recovery significantly. Understanding the methane trend is an important priority for understanding the future ozone recovery.

- Temperatures in the Arctic winter lower stratosphere are generally close to the threshold for substantial chlorine activation, making Arctic ozone particularly sensitive to small changes in temperature (e.g., cooling of the lower stratosphere by changes in greenhouse gases). Preliminary calculations with coupled chemistry/climate models suggest that recovery in the Arctic could be delayed by this cooling and, because of the large natural variability, recovery will be difficult to detect unambiguously until well into the next century.

- The detection of the onset of ozone recovery from halogen-induced depletion should be possible earlier in the Antarctic than in the Arctic or globally because there is less variability in the ozone loss in the Antarctic. Estimates of the timing of the detection of the onset of ozone recovery are uncertain. However, it is clear that unambiguous detection of the beginning of recovery will be delayed beyond the maximum loading of stratospheric halogens.

Implications for Policy Formulation

The results from more than two decades of research have provided a progressively better understanding of the interaction of human activities and the chemistry and physics of the global atmosphere. New policy-relevant insights to the roles of trace atmospheric constituents have been conveyed to decision-makers through the international state-of-the-understanding assessment process. This information has served as a key input to policy decisions by governments, industry, and other organizations worldwide to limit the anthropogenic emissions of gases that cause environmental degradation: (1) the 1987 Montreal Protocol on ozone-depleting substances, and its subsequent Amendments and Adjustments, and (2) the 1997 Kyoto Protocol on substances that alter the radiative forcing of the climate system.

The research findings that are summarized above are of direct interest and significance as scientific input to governmental, industrial, and other policy decisions associated with the Montreal Protocol (ozone layer) and the Kyoto Protocol (climate change):

- **The Montreal Protocol is working.** Global observations have shown that the combined abundance of anthropogenic chlorine-containing and bromine-containing ozone-depleting substances in the lower atmosphere peaked in 1994 and has now started to decline. One measure of success of the Montreal Protocol and its subsequent Amendments and Adjustments is the forecast of "the world that was avoided" by the Protocol:

  – The abundance of ozone-depleting gases in 2050, the approximate time at which the ozone layer is now projected to recover to pre-1980 levels, would be at least 17 ppb of equivalent effective chlorine (this is based on the conservative assumption of a 3% per annum growth in ozone-depleting gases), which is about 5 times larger than today's value.

  – Ozone depletion would be at least 50% at midlatitudes in the Northern Hemisphere and 70% at midlatitudes in the Southern Hemisphere, about 10 times larger than today.

  – Surface UV-B radiation would at least double at midlatitudes in the Northern Hemisphere and quadruple at midlatitudes in the Southern Hemisphere compared with an unperturbed atmosphere. This compares to the current increases of 5% and 8% in the Northern and Southern Hemispheres, respectively, since 1980.

Furthermore, all of the above impacts would have continued to grow in the years beyond 2050. It is important to note that, while the provisions of the original Montreal Protocol in 1987 would have lowered the above growth rates, recovery (i.e., an improving situation) would have been impossible without the Amendments and Adjustments (London, 1990; Copenhagen, 1992; and Vienna, 1995).

17

- **The ozone layer is currently in its most vulnerable state.** Total stratospheric loading of ozone-depleting substances is expected to maximize before the year 2000. All other things being equal, the current ozone losses (relative to the values observed in the 1970s) would be close to the maximum. These are:

  - about 6% at Northern Hemisphere midlatitudes in winter/spring;
  - about 3% at Northern Hemisphere midlatitudes in summer/fall;
  - about 5% at Southern Hemisphere midlatitudes on a year-round basis;
  - about 50% in the Antarctic spring; and
  - about 15% in the Arctic spring.

  Such changes in ozone are predicted to be accompanied by increases in surface erythemal radiation of 7, 4, 6, 130, and 22%, respectively, if other influences such as clouds remain constant. It should be noted that these values for ozone depletion at midlatitudes are nearly a factor of 2 lower than projected in 1994, primarily because the linear trend in ozone observed in the 1980s did not continue in the 1990s. However, springtime depletion of ozone in Antarctica continues unabated at the same levels as observed in the early 1990s, and large depletions of ozone have been observed in the Arctic in most years since 1990, which are characterized by unusually cold and protracted winters.

  Some natural and anthropogenic processes that do not in themselves cause ozone depletion can modulate the ozone loss from chlorine and bromine compounds, in some cases very strongly. For example, in coming decades midlatitude ozone depletion could be enhanced by major volcanic eruptions, and Arctic ozone depletion could be increased by cold polar temperatures, which in turn could be linked to greenhouse gases or to natural temperature fluctuations. On the other hand, increases in methane would tend to decrease chlorine-catalyzed ozone loss.

  The current vulnerability to ozone depletion over the next few decades is primarily due to past use and emissions of the long-lived ozone-depleting substances. The options to reduce this vulnerability over the next two decades are thus rather limited. The main drivers of ozone change could be natural and anthropogenic processes not related to chlorine and bromine compounds, but to which the ozone layer is sensitive because of the elevated abundances of ozone-depleting substances.

- **The ozone layer will slowly recover over the next 50 years.** The stratospheric abundance of halogenated ozone-depleting substances is expected to return to its pre-1980 (i.e., "unperturbed") level of 2 ppb chlorine equivalent by about 2050, assuming full compliance with the Montreal Protocol and its Amendments and Adjustments. The atmospheric abundances of global and Antarctic ozone will start to slowly recover within coming decades toward their pre-1980 levels once the stratospheric abundances of ozone-depleting (halogen) gases start to decrease. However, the future abundance of ozone will be controlled not only by the abundance of halogens, but also by the atmospheric abundances of methane, nitrous oxide, water vapor, and sulfate aerosols and by the Earth's climate. Therefore, for a given halogen loading in the future, atmospheric ozone abundance is unlikely to be the same as found in the past for the same halogen loading.

- **Few policy options are available to enhance the recovery of the ozone layer.** Relative to the current, but not yet ratified, control measures (Montreal, 1997), the equivalent effective chlorine loading above the 1980 level, integrated from now until the 1980 level is re-attained, could be decreased by:

  - 9% by eliminating global Halon-1211 emissions in the year 2000, thus requiring the complete elimination of all new production and destruction of all Halon-1211 in existing equipment;

  - 7% by eliminating global Halon-1301 emissions in the year 2000, thus requiring the complete elimination of all new production and destruction of all Halon-1301 in existing equipment;

– 5% by eliminating the global production of all HCFCs in the year 2004;

– 2.5% by eliminating the global production of all CFCs and carbon tetrachloride in the year 2004;

– 1.6% by reducing the cap on HCFC production in developed countries from 2.8% to 2.0% in the year 2000, by advancing the phase-out from the year 2030 to 2015, and by instituting more rapid intermediate reductions; and

– about 1% by eliminating the global production of methyl bromide beginning in 2004.

These policy actions would advance the date at which the abundance of effective chlorine returns to the 1980 value by 1-3 years. A complete and immediate global elimination of all emissions of ozone-depleting substances would result in the stratospheric halogen loading returning to the pre-1980 values by the year 2033. It should also be noted that if the currently allowed essential uses for metered dose inhalers are extended from the year 2000 to 2004, then the equivalent effective chlorine loading above the 1980 level would increase by 0.3%.

• **Failure to comply with the international agreements of the Montreal Protocol will affect the recovery of the ozone layer.** For example, illegal production of 20-40 ktonnes per year of CFC-12 and CFC-113 for the next 10-20 years would increase the equivalent effective chlorine loading above the 1980 abundance, integrated from now until the 1980 abundance is re-attained, by about 1-4% and delay the return to pre-1980 abundances by about a year.

• **The issues of ozone depletion and climate change are interconnected; hence, so are the Montreal and Kyoto Protocols.** Changes in ozone affect the Earth's climate, and changes in climate and meteorological conditions affect the ozone layer, because the ozone depletion and climate change phenomena share a number of common physical and chemical processes. Hence, decisions taken (or not taken) under one Protocol have an impact on the aims of the other Protocol. For example, decisions made under the Kyoto Protocol with respect to methane, nitrous oxide, and carbon dioxide will affect the rate of recovery of ozone, while decisions regarding controlling HFCs may affect decisions regarding the ability to phase out ozone-depleting substances.

FREQUENTLY ASKED QUESTIONS ABOUT OZONE

Ozone is very rare in our atmosphere, averaging about three molecules of ozone for every 10 million air molecules. In spite of this small amount, ozone plays vital roles in the atmosphere. This appendix to the Executive Summary of the *Scientific Assessment of Ozone Depletion: 1998* provides answers to some of the questions that are most frequently asked about ozone and the changes that have been occurring in recent years. These questions and answers are based on the information presented in this 1998 report, which was prepared by 304 scientists from 35 countries worldwide. Therefore, the information presented here represents the knowledge of a large group of experts from the international scientific community.

Ozone is mainly found in two regions of the Earth's atmosphere. Most ozone (about 90%) resides in a layer that begins between 8 and 18 kilometers (5 and 11 miles) above the Earth's surface and extends up to about 50 kilometers (30 miles). This region of the atmosphere is called the stratosphere. The ozone in this region is commonly known as the ozone layer. The remaining ozone is in the lower region of the atmosphere, which is commonly called the troposphere. The figure below shows an example of how ozone is distributed in the atmosphere.

The ozone molecules in these two regions are chemically identical, because they all consist of three oxygen atoms and have the chemical formula O_3. However, they have very different effects on humans and other living beings. Stratospheric ozone plays a beneficial role by absorbing most of the biologically damaging ultraviolet sunlight (called UV-B), allowing only a small amount to reach the Earth's surface. The absorption of ultraviolet radiation by ozone creates a source of heat, which actually forms the stratosphere itself (a region in which the temperature rises as one goes to higher altitudes). Ozone thus plays a key role in the temperature structure of the Earth's atmosphere. Without the filtering action of the ozone layer, more of the Sun's UV-B radiation would penetrate the atmosphere and would reach the Earth's surface. Many experimental studies of plants and animals and clinical studies of humans have shown the harmful effects of excessive exposure to UV-B radiation.

At the Earth's surface, ozone comes into direct contact with life-forms and displays its destructive side. Because ozone reacts strongly with other molecules, high levels of ozone are toxic to living systems. Several studies have documented the harmful effects of ozone on crop production, forest growth, and human health. The substantial negative effects of surface-level tropospheric ozone from this direct toxicity contrast with the benefits of the additional filtering of UV-B radiation that it provides.

The dual role of ozone leads to two separate environmental issues. There is concern about *increases* in ozone in the troposphere. Low-lying ozone is a key component of photochemical smog, a familiar problem in the atmosphere of many cities around the world. Higher

Atmospheric Ozone

Stratospheric Ozone
(The Ozone Layer)

- Contains 90% of Atmospheric Ozone
- Beneficial Role: Acts as Primary UV Radiation Shield
- Current Issues:
  — Long-Term Global Downward Trends
  — Springtime Antarctic Ozone Hole Each Year
  — Springtime Arctic Ozone Losses in Several Recent Years

Tropospheric Ozone

- Contains 10% of Atmospheric Ozone
- Harmful Impact: Toxic Effects on Humans and Vegetation
- Current Issues:
  — Episodes of High Surface Ozone in Urban and Rural Areas

"Smog" Ozone

Altitude (kilometers)

Ozone Amount
(pressure, milli-Pascals)

amounts of surface-level ozone are increasingly being observed in rural areas as well.

There is also widespread scientific and public interest and concern about *losses* of ozone in the stratosphere. Ground-based and satellite instruments have measured decreases in the amount of stratospheric ozone in our atmosphere. Over some parts of Antarctica, up to 60% of the total overhead amount of ozone (known as the column ozone) is depleted during Antarctic spring (September-November). This phenomenon is known as the Antarctic ozone hole. In the Arctic polar regions, similar processes occur that have also led to significant chemical depletion of the column ozone during late winter and spring in 6 out of the last 9 years. The ozone loss from January through late March has been typically 20-25%, and shorter-period losses have been higher, depending on the meteorological conditions encountered in the Arctic stratosphere. Smaller, but still significant, stratospheric ozone decreases have been seen at other, more-populated regions of the Earth. Increases in surface UV-B radiation have been observed in association with local decreases in stratospheric ozone, from both ground-based and satellite-borne instruments.

The scientific evidence, accumulated over more than two decades of study by the international research community, has shown that human-produced chemicals are responsible for the observed depletions of the ozone layer. The ozone-depleting compounds contain various combinations of the chemical elements chlorine, fluorine, bromine, carbon, and hydrogen and are often described by the general term halocarbons. The compounds that contain only chlorine, fluorine, and carbon are called chlorofluorocarbons, usually abbreviated as CFCs. CFCs, carbon tetrachloride, and methyl chloroform are important human-produced ozone-depleting gases that have been used in many applications including refrigeration, air conditioning, foam blowing, cleaning of electronics components, and as solvents. Another important group of human-produced halocarbons is the halons, which contain carbon, bromine, fluorine, and (in some cases) chlorine and have been mainly used as fire extinguishants. Governments have decided to eventually discontinue production of CFCs, halons, carbon tetrachloride, and methyl chloroform (except for a few special uses), and industry has developed more "ozone-friendly" substitutes.

Two responses are natural when a new problem has been identified: cure and prevention. When the problem is the destruction of the stratospheric ozone layer, the corresponding questions have been the following ones: Can we repair the damage already done? How can we prevent further destruction? Remedies have been investigated that could (1) remove CFCs selectively from the atmosphere, (2) intercept ozone-depleting chlorine before much depletion has taken place, or (3) replace the ozone lost in the stratosphere (perhaps by shipping the ozone from cities that have too much smog or by making new ozone). However, because ozone reacts strongly with other molecules, it is too unstable to be made elsewhere (e.g., in the smog of cities) and transported to the stratosphere. Considering the huge volume of the Earth's atmosphere and the magnitude of global stratospheric ozone depletion, the suggested remedies quickly become much too expensive, too energy consuming, impractical, and potentially damaging to the global environment.

Repair involves the internationally agreed-upon Montreal Protocol and its Amendments and Adjustments. This agreement regulates the production of CFCs and other ozone-depleting substances. Production of the most damaging ozone-depleting substances was eliminated, except for a few critical uses, by 1996 in developed countries and will be eliminated by 2010 in developing countries. As a result, the total concentration of chlorine in the lower atmosphere that can be carried to the stratosphere has peaked already. The concentrations in the stratosphere will likely peak by the end of this decade and then will start to decrease slowly as natural processes remove the ozone-depleting substances. All other things being equal, and with adherence to the international agreements, the ozone layer is expected to recover over the next 50 years or so.

How Can Chlorofluorocarbons (CFCs) Get to the Stratosphere If They're Heavier than Air?

CFCs reach the stratosphere because the Earth's atmosphere is always in motion and mixes the chemicals added into it.

CFC molecules are indeed several times heavier than air. Nevertheless, thousands of measurements from balloons, aircraft, and satellites demonstrate that the CFCs are actually present in the stratosphere. This is because winds and other air motions mix the atmosphere to altitudes far above the top of the stratosphere much faster than molecules can settle according to their weight. Gases such as CFCs that do not dissolve in water and that are relatively unreactive in the lower atmosphere are mixed relatively quickly and therefore reach the stratosphere regardless of their weight.

Measured changes in the concentration of constituents versus altitude teach us more about the fate of compounds in the atmosphere. For example, the two gases carbon tetrafluoride (CF_4, produced mainly as a by-product of the manufacture of aluminum) and CFC-11 (CCl_3F, used in a variety of human activities) are both heavier than air.

Carbon tetrafluoride is completely unreactive at altitudes up to at least 50 kilometers in the atmosphere. Measurements show it to be nearly uniformly distributed throughout the atmosphere (as illustrated in the figure below, the abundance of CF_4 is nearly the same at all altitudes where measurements have been made). There have been measurements over the past two decades of several other completely unreactive gases, both lighter than air (neon) and heavier than air (argon and krypton), that show that they also mix upward through the stratosphere regardless of their weight.

CFC-11 is unreactive in the lower atmosphere (below about 15 kilometers) and is similarly uniformly mixed there, as shown in the figure. However, the abundance of CFC-11 decreases as the gas reaches higher altitudes, because it is broken down by high-energy solar ultraviolet radiation. Chlorine released from this breakdown of CFC-11 and other CFCs remains in the stratosphere for several years, where every chlorine atom destroys many thousands of molecules of ozone.

Atmospheric Measurements of CFC-11 and CF_4
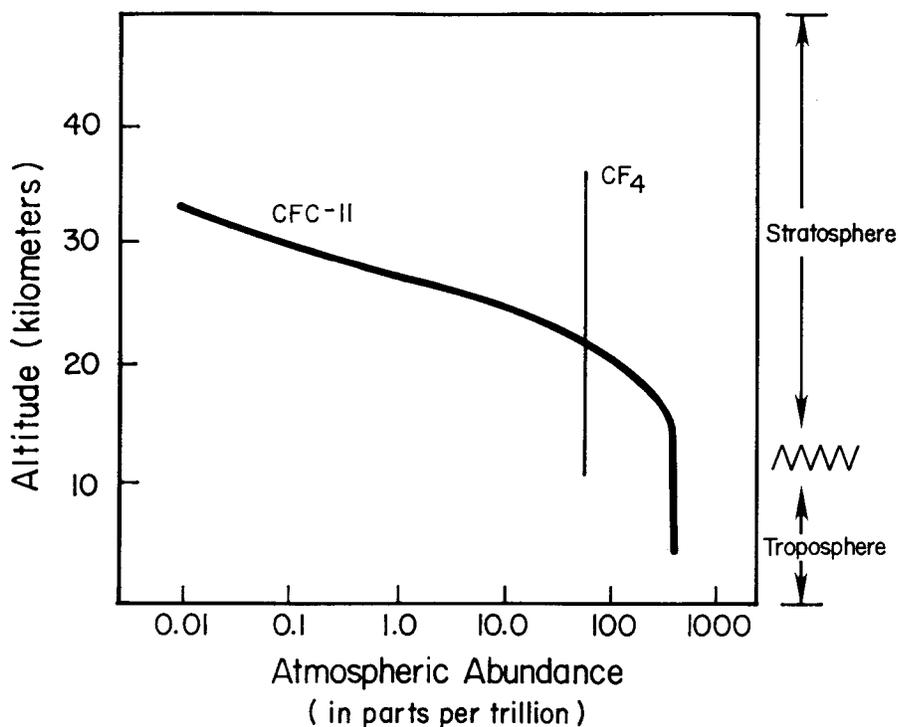
What is the Evidence that Stratospheric Ozone is Destroyed by Chlorine and Bromine?

Numerous laboratory investigations and analyses of worldwide measurements made in the stratosphere have demonstrated that chlorine- and bromine-containing chemicals destroy ozone molecules.

Research studies in the laboratory show that chlorine (Cl) reacts very rapidly with ozone. They also show that the reactive chemical chlorine monoxide (ClO) formed in that reaction can undergo further processes that regenerate the original chlorine, allowing the sequence to be repeated very many times (a chain reaction). Similar reactions also take place between bromine and ozone.

But do these ozone-destroying reactions occur in the "real world"? All the accumulated scientific experience demonstrates that the same chemical reactions do take place in nature. Many other reactions (including those of other chemical species) are often also taking place simultaneously in the stratosphere. This makes the connections among the changes difficult to untangle. Nevertheless, whenever chlorine (or bromine) and ozone are found together in the stratosphere, the ozone-destroying reactions are taking place.

Sometimes a small number of chemical reactions are so dominant in the natural circumstance that the connections are almost as clear as in laboratory experiments. Such a situation occurs in the Antarctic stratosphere during the springtime formation of the ozone hole. Independent measurements made by instruments from the ground and from balloons, aircraft, and satellites have provided a detailed understanding of the chemical reactions in the Antarctic stratosphere. Large areas reach temperatures so low (less than -80°C, or -112°F) that

stratospheric clouds form, which is a rare occurrence, except during the polar winters. These polar stratospheric clouds allow chemical reactions that transform chlorine species from forms that do *not* cause ozone depletion into forms that do cause ozone depletion. Among the latter is chlorine monoxide, which initiates ozone destruction in the presence of sunlight. The amount of reactive chlorine in such regions is therefore much higher than that observed in the middle latitudes, which leads to much faster chemical ozone destruction. The chemical reactions occurring in the presence of these clouds are now well understood from studies under laboratory conditions that mimic those found naturally in the atmosphere.

Scientists have repeatedly observed a large number of chemical species over Antarctica since 1986. Among the chemicals measured were ozone and chlorine monoxide, which is the reactive chemical identified in the laboratory as one of the participants in the ozone-destroying chain reactions. The satellite maps shown in the figure below relate the accumulation of chlorine monoxide observed over Antarctica and the subsequent ozone depletion that occurs rapidly in a few days over very similar areas.

Similar reactions involving chlorine and bromine have also been shown to occur during winter and spring in the Arctic polar regions, which leads to some chemical depletion of ozone in that region. Because the Arctic is not usually as persistently cold as the Antarctic, fewer stratospheric clouds form, and therefore there is less ozone depletion in the Arctic, which is the subject of a later question.

Chlorine Monoxide and the Antarctic Ozone Hole: Late August 1996
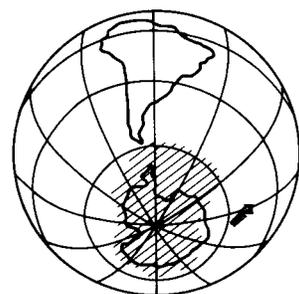
Region of
High Chlorine Monoxide (ClO)
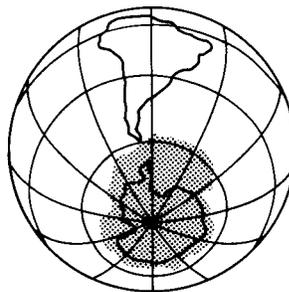
Region of
Low Ozone (O_3)

# Does Most of the Chlorine in the Stratosphere Come from Human or Natural Sources?

Most of the chlorine in the stratosphere is there as a result of human activities, as the figure below illustrates.

Many compounds containing chlorine are released at the ground. Those that dissolve in water cannot reach stratospheric altitudes in significant amounts because they are "washed out" of the atmosphere in rain or snow. For example, large quantities of chlorine are released from evaporated ocean spray as sea salt (sodium chloride) particles. However, because sea salt dissolves in water, this chlorine is taken up quickly in clouds or in ice, snow, or rain droplets and does not reach the stratosphere. Another ground-level source of chlorine is from its use in swimming pools and as household bleach. When released, this chlorine is rapidly converted to forms that dissolve in water and therefore are removed from the lower atmosphere. Such chlorine never reaches the stratosphere in significant amounts. Volcanoes can emit large quantities of hydrogen chloride, but this gas is rapidly converted to hydrochloric acid, which dissolves in rain water, ice, and snow and does not reach the stratosphere. Even in explosive volcanic plumes that rise high in the atmosphere, nearly all of the hydrogen chloride is removed by precipitation before reaching stratospheric altitudes. Finally, although the exhaust from the Space Shuttle and from some rockets does inject some chlorine directly into the stratosphere, the quantities are very small (less than 1% of the annual input from halocarbons in the present stratosphere).

In contrast, the major ozone-depleting human-produced halocarbons — such as chlorofluorocarbons (CFCs) and carbon tetrachloride (CCl_4) — do not dissolve in water, do not react with snow or other natural surfaces, and are not broken down chemically in the lower atmosphere. Therefore, these and other human-produced substances containing chlorine do reach the stratosphere.

Several pieces of evidence combine to establish human-produced halocarbons as the primary source of stratospheric chlorine. First, measurements have shown that the chlorinated species that rise to the stratosphere are primarily manufactured compounds [mainly CFCs, carbon tetrachloride, methyl chloroform, and the hydrochlorofluorocarbon (HCFC) substitutes for CFCs], together with small amounts of hydrochloric acid (HCl) and methyl chloride (CH_3Cl), which are partly natural in origin. Second, researchers have measured nearly all known gases containing chlorine in the stratosphere. They have found that the emissions of the human-produced halocarbons, plus the much smaller contribution from natural sources, could account for all of the stratospheric chlorine. Third, the *increase* in total stratospheric chlorine measured between 1980 and 1998 corresponds to the known increases in concentrations of human-produced halocarbons during that time.

Primary Sources of Chlorine Entering the Stratosphere in the Early 1990s
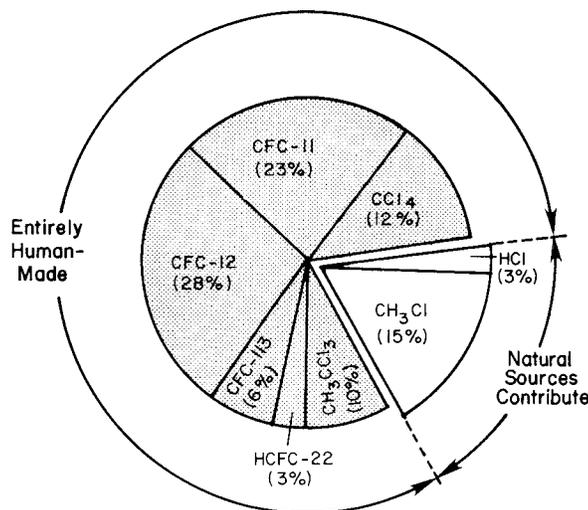
Can Natural Changes Such As the Sun's Output and Volcanic Eruptions Be Responsible for the Observed Changes in Ozone?

Although there are natural forces that cause fluctuations in ozone amounts, there is no evidence that natural changes are contributing significantly to the observed long-term trend of decreasing ozone.

The formation of stratospheric ozone is initiated by ultraviolet (UV) light coming from the Sun. As a result, the Sun's output affects the rate at which ozone is produced. The Sun's energy release (both as UV light and as charged particles such as electrons and protons) does vary, especially over the well-known 11-year sunspot cycle. Observations over several solar cycles (since the 1960s) show that total global ozone levels vary by 1-2% from the maximum to the minimum of a typical cycle. However, changes in the Sun's output cannot be responsible for the observed long-term changes in ozone, because the ozone downward trends are much larger than 1-2%. As the figure below shows, since 1978 the Sun's energy output has gone through maximum values in about 1980 and 1991 and minimum values in about 1985 and 1996. It is now increasing again toward its next maximum around the year 2002. However, the trend in ozone was downward throughout that time. The ozone trends presented in this and previous international scientific assessments have been obtained by evaluating the long-term changes in ozone after accounting for the solar influence (as has been done in the figure below).

Major, explosive volcanic eruptions can inject material directly into the ozone layer. Observations and model calculations show that volcanic particles cannot on their own deplete ozone. It is only the interaction of human-produced chlorine with particle surfaces that enhances ozone depletion in today's atmosphere.

Specifically, laboratory measurements and observations in the atmosphere have shown that chemical reactions on and within the surface of volcanic particles injected into the lower stratosphere lead to enhanced ozone destruction by increasing the concentration of chemically active forms of chlorine that arise from the human-produced compounds like the chlorofluorocarbons (CFCs). The eruptions of Mt. Agung (1963), Mt. Fuego (1974), El Chichón (1982), and particularly Mt. Pinatubo (1991) are examples. The eruption of Mt. Pinatubo resulted in a 30- to 40-fold increase in the total surface area of particles available for enhancing chemical reactions. The effect of such natural events on the ozone layer is then dependent on the concentration of chlorine-containing molecules and particles available in the stratosphere, in a manner similar to polar stratospheric clouds. Because the particles are removed from the stratosphere in 2 to 5 years, the effect on ozone is only temporary, and such episodes cannot account for observed long-term changes. Observations and calculations indicate that the record-low ozone levels observed in 1992-1993 reflect the importance of the relatively large number of particles produced by the Mt. Pinatubo eruption, coupled with the relatively higher amount of human-produced stratospheric chlorine in the 1990s compared to that at times of earlier volcanic eruptions.

Global Ozone Trend, Major Volcanic Eruptions, and Solar Cycles

26
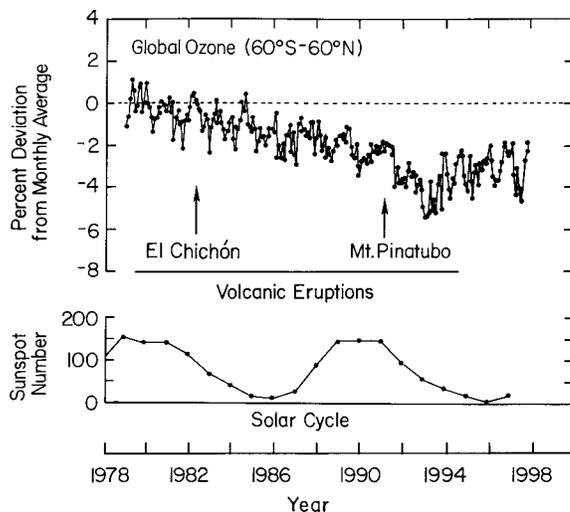
When Did the Antarctic Ozone Hole First Appear?

The springtime Antarctic ozone hole is a new phenomenon that appeared in the early 1980s.

The observed average amount of ozone during September, October, and November over the British Antarctic Survey station at Halley, Antarctica, first revealed notable decreases in the early 1980s, compared with the preceding data obtained starting in 1957. The ozone hole is formed each year when there is a sharp decline (currently up to 60%) in the total ozone over most of Antarctica for a period of about three months (September-November) during spring in the Southern Hemisphere. Late-summer (January-March) ozone amounts show no such sharp decline in the 1980s and 1990s. Observations from three other stations in Antarctica and from satellite-based instruments reveal similar decreases in springtime amounts of ozone overhead. Balloonborne ozone instruments show dramatic changes in the way ozone is distributed with altitude. As the figure below from the Syowa site shows, almost all of the ozone is now depleted at some altitudes as the ozone hole forms each springtime, compared to the normal ozone profile that existed before 1980. As explained in an earlier question (page 24), the ozone hole has been shown to result from destruction of stratospheric ozone by gases containing chlorine and bromine, whose sources are mainly human-produced halocarbon gases.

Before the stratosphere was affected by human-produced chlorine and bromine, the naturally occurring springtime ozone levels over Antarctica were about 30-40% lower than springtime ozone levels over the Arctic. This natural difference between Antarctic and Arctic conditions was first observed in the late 1950s by Dobson. It stems from the exceptionally cold temperatures and different winter wind patterns within the Antarctic stratosphere as compared with the Arctic. This is not at all the same phenomenon as the marked downward trend in ozone over Antarctica in recent years.

Changes in stratospheric meteorology cannot explain the ozone hole. Measurements show that wintertime Antarctic stratospheric temperatures of past decades had not changed prior to the development of the ozone hole each September. Ground, aircraft, and satellite measurements have provided, in contrast, clear evidence of the importance of the chemistry of chlorine and bromine originating from human-made compounds in depleting Antarctic ozone in recent years.

Springtime Depletion of the Ozone Layer over Syowa, Antarctica
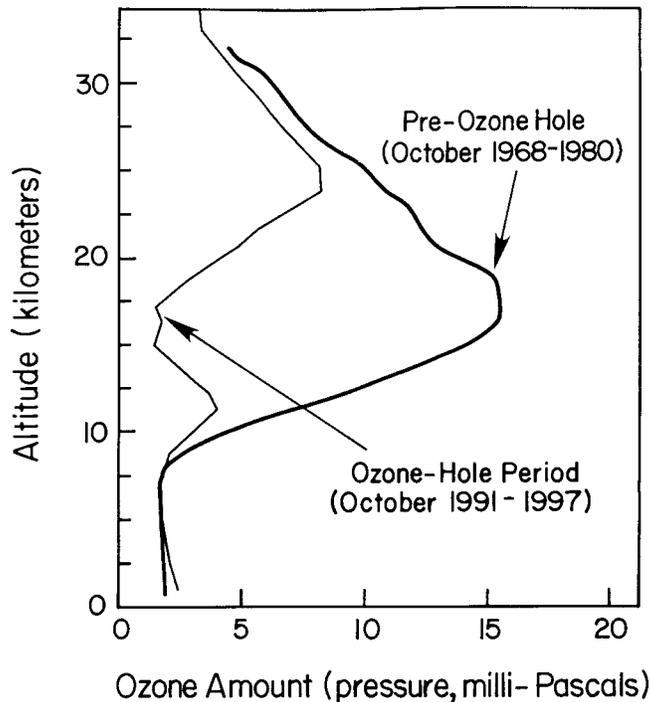
Why Has an Ozone Hole Appeared over Antarctica When CFCs and Halons Are Released Mainly in the Northern Hemisphere?

The Earth's atmosphere is continuously stirred over the globe by winds. As a result, ozone-depleting gases get mixed throughout the atmosphere, including Antarctica, regardless of where they are emitted. The special meteorological conditions in Antarctica cause these gases to be more effective there in depleting ozone compared to anywhere else.

Human emissions of chlorofluorocarbons (CFCs) and halons (bromine-containing gases) have occurred mainly in the Northern Hemisphere. About 90% have been released in the latitudes corresponding to Europe, Russia, Japan, and North America. Gases such as CFCs and halons, which are insoluble in water and relatively unreactive, are mixed within a year or two throughout the lower atmosphere. The CFCs and halons in this well-mixed air rise from the lower atmosphere into the stratosphere mainly in tropical latitudes. Winds then move this air poleward—both north and south—from the tropics, so that air throughout the global strato-sphere contains nearly equal amounts of chlorine and bromine.

In the Southern Hemisphere, the South Pole is part of a very large land mass (Antarctica) that is completely surrounded by ocean. This symmetry is reflected in the meteorological conditions that allow the formation in winter of a very cold region in the stratosphere over the Antarctic continent, isolated by a band of strong winds circulating around the edge of that region. The very low stratospheric temperatures lead to the formation of clouds (polar stratospheric clouds) that are responsible for chemical changes that promote production of chemically active chlorine and bromine. This chlorine and bromine activation then leads to rapid ozone loss when sunlight returns to Antarctica in September and October of each year, which then results in the Antarctic ozone hole. As the figure below depicts, the magnitude of the ozone loss has generally grown through the 1980s as the amount of human-produced ozone-depleting compounds has grown in the atmosphere.

Similar conditions do not exist over the Arctic. The wintertime temperatures in the Arctic stratosphere are not persistently low for as many weeks as over Antarctica, which results in correspondingly less ozone depletion in the Arctic (see the next question).

Schematic of the Growth of the Antarctic Ozone Hole

1979   1986

1991   1996
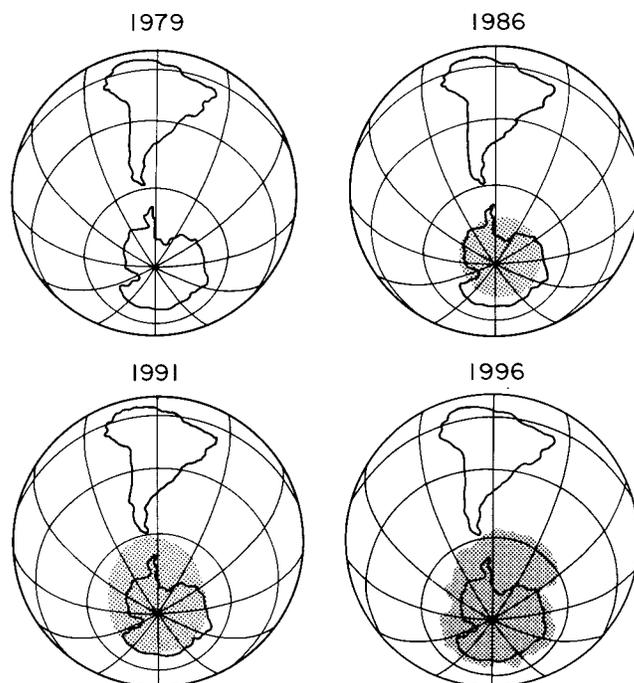
Is There an Ozone Hole over the Arctic?

Significant reductions in ozone content in the stratosphere above the Arctic have been observed during the late winter and early spring (January-March) in 6 of the last 9 years. However, these reductions, typically 20-25%, are much smaller than those observed currently each spring over the Antarctic (the ozone hole).

The difference between ozone content in the two polar regions (see figure below) is caused by dissimilar weather patterns. The Antarctic continent is a very large land mass surrounded by oceans. This symmetrical condition produces very low stratospheric temperatures within a meteorologically isolated region, the so-called polar vortex, which extends from about 65˚S to the pole. The cold temperatures lead in turn to the formation of clouds, known as polar stratospheric clouds. These clouds provide surfaces that promote production of forms of chlorine and bromine that are chemically active and can rapidly destroy ozone. The conditions that maintain elevated levels of chemically active chlorine and bromine persist into September and October in Antarctica, when sunlight returns over the region to initiate ozone depletion.

The winter meteorological conditions in the Northern Hemisphere, just like in the Southern Hemisphere, lead to the formation of an isolated region bounded by strong winds, in which the temperature is also cold enough for polar stratospheric clouds to form. However, the geographic symmetry about the North Pole is less than about the South Pole. As a result, large-scale weather systems disturb the wind flow, making it less stable over the Arctic region than over the Antarctic continent. These disturbances prevent the temperature in the Arctic stratosphere from being as cold as in the Antarctic stratosphere, and fewer polar stratospheric clouds are therefore formed. Nevertheless, chemically active chlorine and bromine compounds are also formed over the Arctic, as they are over Antarctica, from reactions at the surface of the clouds. But the cold conditions rarely persist into March, when sufficient sunlight is available to initiate large ozone depletion.

In recent years, there has been a string of unusually cold winters in the Arctic, compared with those in the preceding 30 years. The cold and persistent conditions have led to enhanced ozone depletion, because the atmospheric concentrations of ozone-depleting gases have also been relatively large during these years. However, the cause of the observed change in meteorological conditions is not yet understood. Such conditions might persist over the coming years, further enhancing ozone depletion. But it is also possible that, in the next few years, they could revert to conditions characteristic of a decade ago. In the latter case, chemical ozone depletion in the Arctic would be expected to diminish.

Therefore, although there has been significant ozone depletion in the Arctic in recent years, it is difficult to predict what may lie ahead, because the future climate of the Arctic stratosphere cannot be predicted with confidence.

A Schematic of the Ozone over the Arctic and Antarctica in 1996

300-320 Dobson Units
(Pre-1980 Levels: About 450)

Less than 210 Dobson Units
(Pre-1980 Levels: About 320)

ARCTIC (March 1996)   ANTARCTIC (September 1996)
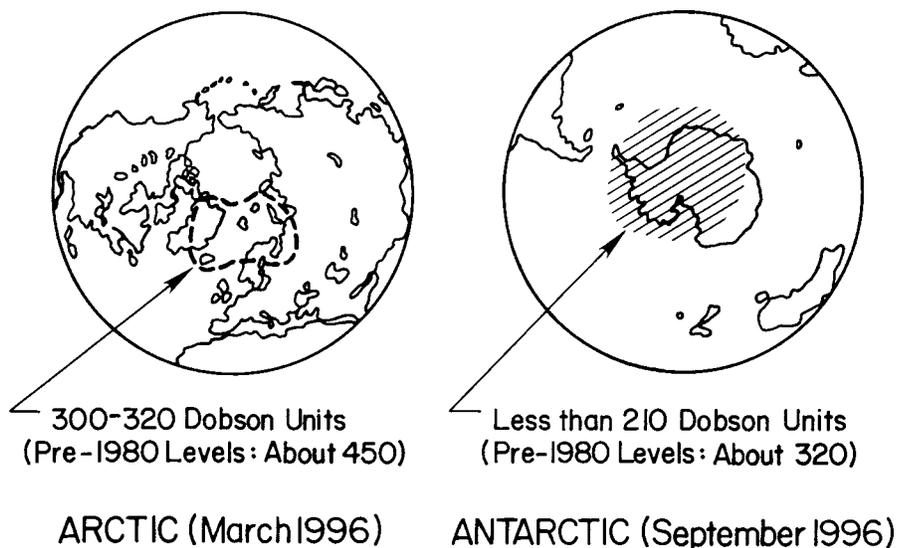
Is the Depletion of the Ozone Layer Leading to an Increase in Ground-Level Ultraviolet Radiation?

The depletion of the ozone layer leads, on the average, to an increase in ground-level ultraviolet radiation, because ozone is an effective absorber of ultraviolet radiation.

The Sun emits radiation over a wide range of energies, with about 2% in the form of high-energy, ultraviolet (UV) radiation. Some of this UV radiation (UV-B) is especially effective in causing damage to living beings, for example, sunburn, skin cancer, and eye damage to humans. The amount of solar UV radiation received at any particular location on the Earth's surface depends upon the position of the Sun above the horizon, the amount of ozone in the atmosphere, and local cloudiness and pollution. Scientists agree that, in the absence of changes in clouds or pollution, decreases in atmospheric ozone lead to increases in ground-level UV radiation.

The largest decreases in ozone during the past 15 years have been observed over Antarctica, especially during each September and October when the ozone hole forms. During the last several years, simultaneous measurements of UV radiation and total ozone have been made at several Antarctic stations. In the late spring, the biologically damaging ultraviolet radiation in parts of the Antarctic continent can exceed that in San Diego, California, where the Sun is much higher above the horizon.

In areas or the world where smaller ozone depletion has been observed, UV-B increases are more difficult to detect. In particular, detection of trends in UV-B radiation associated with ozone decreases can be further complicated by changes in cloudiness, by local pollution, and by difficulties in keeping the detection instrument in precisely the same operating condition over many years. Prior to the late 1980s, instruments with the necessary accuracy and stability for measurement of small long-term trends in ground-level UV-B were not available. Therefore, the data from urban locations with older, less-specialized instruments provide much less reliable information, especially since simultaneous measurements of changes in cloudiness or local pollution are not available. When high-quality measurements have been made in other areas far from major cities and their associated air pollution, decreases in ozone have regularly been accompanied by increases in UV-B. This is shown in the figure below, where clear-sky measurements performed at six different stations demonstrate that ozone decreases lead to increased UV-B radiation at the surface in amounts that are in good agreement with that expected from calculations (the "model" curve).

Increases in Erythemal (Sunburning) Ultraviolet Radiation Due to Ozone Decreases
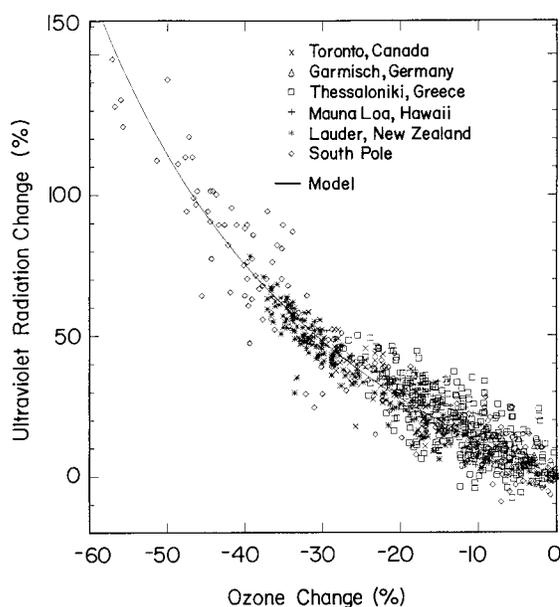
Does Ozone Depletion Cause Climate Change?

Ozone depletion and climate change are linked in a number of ways, but ozone depletion is not a major cause of climate change.

Atmospheric ozone has two effects on the temperature balance of the Earth. It absorbs solar ultraviolet radiation, which heats the stratosphere. It also absorbs infrared radiation emitted by the Earth's surface, effectively trapping heat in the troposphere. Therefore, the climate impact of changes in ozone concentrations varies with the altitude at which these ozone changes occur. The major ozone losses that have been observed in the lower stratosphere due to the human-produced chlorine- and bromine-containing gases have a cooling effect on the Earth's surface. On the other hand, the ozone increases that are estimated to have occurred in the troposphere because of surface-pollution gases have a warming effect on the Earth's surface, thereby contributing to the "greenhouse" effect.

In comparison to the effects of changes in other atmospheric gases, the effects of both of these ozone changes are difficult to calculate accurately. In the figure below, the upper ranges of possible effects from the ozone changes are indicated by the open bars, and the lower ranges are indicated by the solid bars.

As shown in the figure, the increase in carbon dioxide is the major contributor to climate change. Carbon dioxide concentrations are increasing in the atmosphere primarily as the result of the burning of coal, oil, and natural gas for energy and transportation. The atmospheric abundance of carbon dioxide is currently about 30% above what it was 150 years ago. The relative impacts on climate of various other "greenhouse" gases are also shown on the figure.

There is an additional factor that indirectly links ozone depletion to climate change; namely, many of the same gases that are causing ozone depletion are also contributing to climate change. These gases, such as the chlorofluorocarbons (CFCs), are greenhouse gases, absorbing some of the infrared radiation emitted by the Earth's surface, thereby effectively heating the Earth's surface.

Conversely, changes in the climate of the Earth could affect the behavior of the ozone layer, because ozone is influenced by changes in the meteorological conditions and by changes in the atmospheric composition that could result from climate change. The major issue is that the stratosphere will most probably cool in response to climate change, therefore preserving over a longer time period the conditions that promote chlorine-caused ozone depletion in the lower stratosphere, particularly in polar regions. At present, the amplitude and extent of such a cooling, and therefore the delay in the recovery of the ozone layer, still have to be assessed.

Relative Importance of the Changes in the Abundance of Various Gases in the Atmosphere
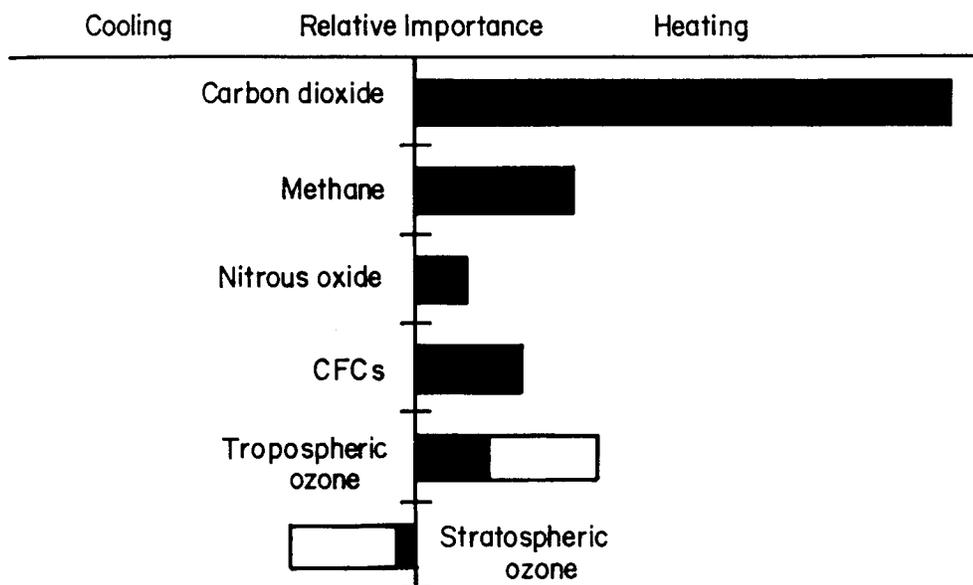
How Severe Is the Ozone Depletion Now?

Stratospheric ozone depletion, caused by increasing concentrations of human-produced chemicals, has increased since the 1980s. The springtime loss in Antarctica is the largest depletion. Currently, in nonpolar regions, the ozone layer has been depleted up to several percent compared with that of two decades ago.

As the figure below indicates, the magnitude of ozone depletion varies between the regions of the Earth. For example, there has been little or no ozone depletion in the tropics (about 20 degrees north and south of the equator). The magnitude of the depletion also depends on the season. From 1979 to 1997, the observed losses in the amount of ozone overhead have totaled about 5-6% for northern midlatitudes in winter and spring, about 3% for northern midlatitudes in summer and fall, and about 5% year round for southern midlatitudes. Since the early 1980s, the ozone hole has formed over Antarctica during every Southern Hemisphere spring (September to November), in which up to 60% of the total ozone is depleted. Since the early 1990s, ozone depletion has also been observed over the Arctic, with the ozone loss from January through late March typically being 20-25% in most of the recent years. All of these decreases are larger than known long-term natural variations.

The large increase in atmospheric concentrations of human-made chlorine and bromine compounds is responsible for the formation of the Antarctic ozone hole. Furthermore, the overwhelming weight of evidence indicates that those same compounds also play a major role in the ozone depletion in the Arctic and at midlatitudes.

In addition to these long-term changes, transient effects have also been observed in the stratospheric ozone layer following major volcanic eruptions such as Mt. Pinatubo in 1991. During 1992 and 1993, ozone in many locations dropped to record low values. For example, springtime depletions exceeded 20% in some populated northern midlatitude regions, and the levels in the Antarctic ozone hole fell to the lowest values ever recorded. These unusually large, but short-term, ozone decreases of 1992 and 1993 are believed to be related in part to the large amounts of volcanic particles injected into stratosphere, which temporarily increased the ozone depletion caused by human-produced chlorine and bromine compounds, much as polar stratospheric clouds increase these chemicals' effect on ozone depletion in polar regions. Because these particles settle out of the stratosphere within a few years, the ozone concentrations have largely returned to the depleted levels consistent with the downward trend observed before the Mt. Pinatubo eruption. Should a similar eruption occur in the coming decade, ozone losses of the same magnitude might be expected, because the chlorine levels in the stratosphere will still be high.

Schematic of the North-to-South Ozone Depletion: 1979-1997
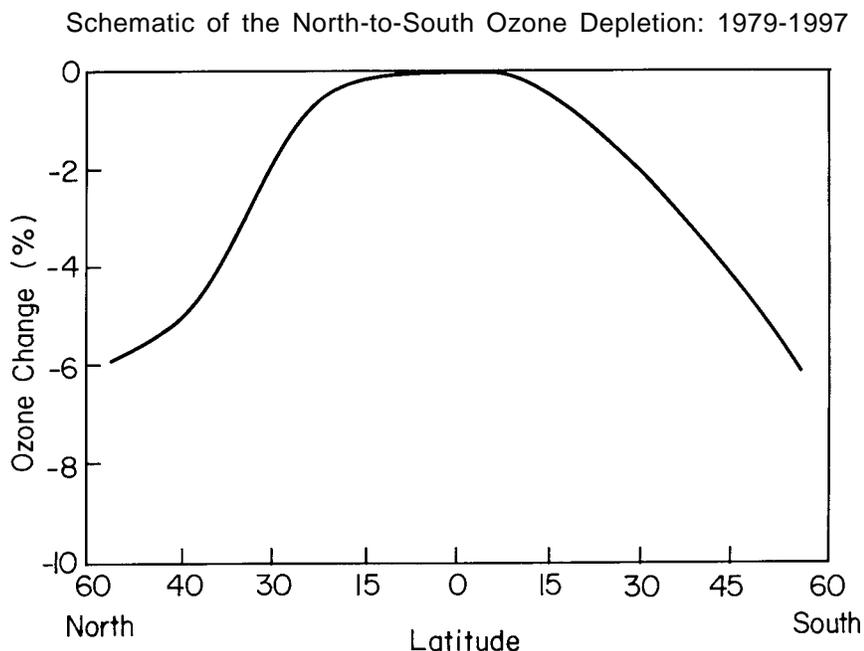
Is the Ozone Layer Expected to Recover? If So, When?

The ozone depletion caused by human-produced chlorine and bromine compounds is expected to gradually disappear by about the middle of the 21st century as these compounds are slowly removed from the stratosphere by natural processes. This environmental achievement is due to the landmark international agreement to control the production and use of ozone-depleting substances. Full compliance would be required to achieve this expected recovery.

In 1987, the recognition of the potential for chlorine and bromine to destroy stratospheric ozone led to the Montreal Protocol on Substances that Deplete the Ozone Layer, as part of the 1985 Vienna Convention for the Protection of the Ozone Layer, to reduce the global production of ozone-depleting substances. Subsequently, global observations of significant ozone depletion have prompted amendments to strengthen the treaty. The 1990 London Amendment calls for a ban on the production of the most damaging ozone-depleting substances by 2000 in developed countries and 2010 in developing countries. The 1992 Copenhagen Amendment changed the date of the ban to 1996 in developed countries. Further restrictions on ozone-depleting substances have been agreed upon in Vienna (1995) and Montreal (1997).

The figure on the right shows past and projected stratospheric abundances of chlorine and bromine without the Protocol, under the Protocol's original provisions, and under its subsequent agreements. Without the Montreal Protocol and its Amendments, continuing use of chlorofluorocarbons (CFCs) and other ozone-depleting substances would have increased the stratospheric abundances of chlorine and bromine tenfold by the mid-2050s compared with the 1980 amounts. Such high chlorine and bromine abundances would have caused very large ozone losses, which would have been far larger than the depletion observed at present.

In contrast, under the current international agreements that are now reducing the human-caused emissions of ozone-depleting gases, the net tropospheric concentrations of chlorine- and bromine-containing compounds started to decrease in 1995. Because 3 to 6 years are required for the mixing from the troposphere to the stratosphere, the stratospheric abundances of chlorine are starting to reach a constant level and will slowly decline thereafter. With full compliance, the international agreements will eventually eliminate most of the emissions of the major ozone-depleting gases. All other things being constant, the ozone layer would be expected to return to a normal state during the middle of the next century. This slow recovery, as compared with the relatively rapid onset of the ozone depletion due to CFC and bromine-containing halon emissions, is related primarily to the time required for natural processes to eliminate the CFCs and halons from the atmosphere. Most of the CFCs and halons have atmospheric residence times of about 50 to several hundred years.

However, the future state of the ozone layer depends on more factors than just the stratospheric concentrations of human-produced chlorine and bromine. It will also be affected to some extent by the changing atmospheric abundances of several other human-influenced constituents, such as methane, nitrous oxide, and sulfate particles, as well as by the changing climate of the Earth. As a result, the ozone layer is unlikely to be identical to the ozone layer that existed prior to the 1980s. Nevertheless, the discovery and characterization of the issue of ozone depletion from chlorine and bromine compounds and a full global compliance with the international regulations on their emissions will have eliminated what would have been, as the figure illustrates, a major deterioration of the Earth's protective ultraviolet shield.

Effect of the International Agreements on Ozone-Depleting Stratospheric Chlorine/Bromine
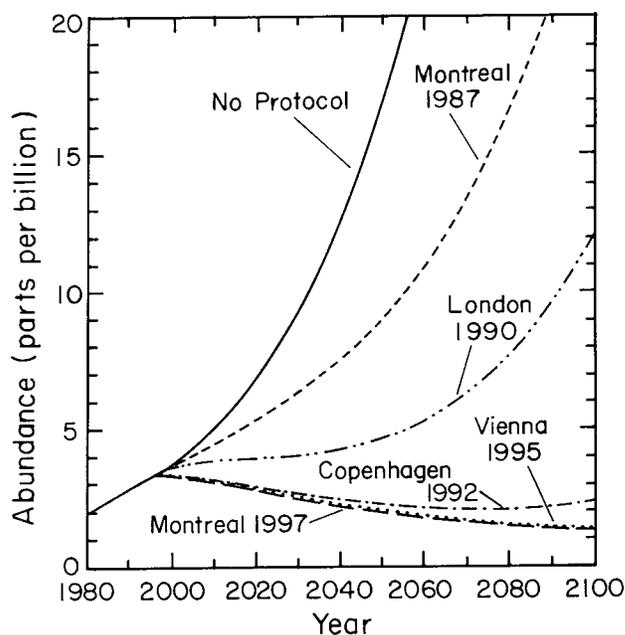
List of International Authors, Contributors, and Reviewers

Assessment Cochairs

Daniel L. Albritton, Pieter J. Aucamp, Gérard Mégie, and Robert T. Watson

Chapters and Lead Authors

Part 1. Halocarbon, Ozone, and Temperature Changes
Chapter 1. Long-Lived Ozone-Related Compounds *(Ronald G. Prinn and R. Zander)*
Chapter 2. Short-Lived Ozone-Related Compounds *(Michael J. Kurylo and José M. Rodríguez)*
Chapter 3. Global Distributions and Changes in Stratospheric Particles *(Sophie Godin and Lamont R. Poole)*
Chapter 4. Ozone Variability and Trends *(Rumen D. Bojkov and Robert Hudson)*
Chapter 5. Trends in Stratospheric Temperatures *(Marie-Lise Chanin and V. Ramaswamy)*

Part 2. Advances in Understanding the Processes Involved
Chapter 6. Upper Stratospheric Processes *(Rolf Müller and Ross J. Salawitch)*
Chapter 7. Lower Stratospheric Processes *(A.R. Ravishankara and Theodore G. Shepherd)*
Chapter 8. Tropospheric Ozone and Related Processes *(Jos Lelieveld and Anne M. Thompson)*

Part 3. Impacts of Ozone Changes
Chapter 9. Ultraviolet Radiation at the Earth's Surface *(Jay R. Herman and Richard L. McKenzie)*
Chapter 10. Climate Effects of Ozone and Halocarbon Changes *(Claire Granier and Keith P. Shine)*

Part 4. Predictions of Future Changes
Chapter 11. Halocarbon Scenarios for the Future Ozone Layer and Related Consequences
(Sasha Madronich and Guus J.M. Velders)
Chapter 12. Predicting Future Ozone Changes and Detection of Recovery
(David J. Hofmann and John A. Pyle)

Coordinator: Frequently Asked Questions about Ozone

Gérard Mégie   Service d'Aéronomie du CNRS   France

Authors, Contributors, and Reviewers

Alberto Adriani	Consiglio Nazionale della Richerche Istituto di Fisica dell'Atmosfera	Italy
Daniel L. Albritton	NOAA Aeronomy Laboratory	US
Georgios T. Amanatidis	European Commission	Belgium
James G. Anderson	Harvard University	US
Meinrat O. Andreae	Max-Planck-Institut für Chemie	Germany
James K. Angell	NOAA Air Resources Laboratory	US
Roger Atkinson	Bureau of Meteorology	Australia
Elliot L. Atlas	National Center for Atmospheric Research	US
Pieter J. Aucamp	Ptersa Environmental Consultants	South Africa
John Austin	U.K. Meteorological Office	UK
Linnea Avallone	University of Colorado	US
Dimitrios S. Balis	Aristotle University of Thessaloniki	Greece
John Barnes	NOAA Climate Monitoring and Diagnostics Laboratory	US

J. Barnett	Oxford University	UK
Darrel Baumgardner	National Center for Atmospheric Research	US
Slimane Bekki	Service d'Aéronomie du CNRS	France
Germar Bernhard	Fraunhofer-Institut für Atmosphärische Umweltforschung	Germany
Richard Bevilacqua	Naval Research Laboratory	US
P.K. Bhartia	NASA Goddard Space Flight Center	US
Lane Bishop	AlliedSignal Inc.	US
Donald R. Blake	University of California at Irvine	US
Mario Blumthaler	Universität Innsbruck	Austria
Greg Bodeker	National Institute of Water & Atmospheric Research Ltd. (NIWA)	New Zealand
Barry A. Bodhaine	NOAA Climate Monitoring and Diagnostics Laboratory	US
Rumen D. Bojkov	World Meteorological Organization	Switzerland
C. Rocky Booth	Biospherical Instruments, Inc.	US
Stephan Borrmann	Universität Mainz	Germany
Byron Boville	National Center for Atmospheric Research	US
Guy Brasseur	National Center for Atmospheric Research	US
Christoph Brühl	Max-Planck-Institut für Chemie	Germany
William H. Brune	Pennsylvania State University	US
Neal Butchart	U.K. Meteorological Office	UK
James H. Butler	NOAA Climate Monitoring and Diagnostics Laboratory	US
Pablo O. Canziani	Universidad de Buenos Aires	Argentina
Daniel Cariolle	Météo-France, Centre National de Recherches Météorologiques	France
Ken Carslaw	Max-Planck-Institut für Chemie	Germany
Edward Celarier	Software Corporation of America	US
Marie-Lise Chanin	Service d'Aéronomie du CNRS	France
Martyn Chipperfield	Centre for Atmospheric Science, University of Cambridge	UK
H. Claude	Meteorological Observatory Hohenpeißenberg	Germany
Cathy Clerbaux	Service d'Aéronomie du CNRS	France
Michael Coffey	National Center for Atmospheric Research	US
Peter S. Connell	Lawrence Livermore National Laboratory	US
David B. Considine	University of Maryland; NASA Goddard Space Flight Center	US
R.A. Cox	Centre for Atmospheric Science, University of Cambridge	UK
Paul J. Crutzen	Max-Planck-Institut für Chemie	Germany
Derek M. Cunnold	Georgia Institute of Technology	US
Martin Dameris	Deutsche Forschungsanstalt für Luft- und Raumfahrt	Germany
John S. Daniel	NOAA Aeronomy Laboratory	US
Christine David	Service d'Aéronomie du CNRS	France
Richard Derwent	U.K. Meterological Office	UK
Terry Deshler	University of Wyoming	US
Roseanne D. Diab	University of Natal	South Africa
Susana Diaz	Centro Austral de Investigaciones Cientificas (CADIC)	Argentina
Ed Dlugokencky	NOAA Climate Monitoring and Diagnostics Laboratory	US
Guido Di Donfrancesco	Consiglio Nazionale della Richerche Istituto di Fisica dell'Atmosfera	Italy
Thomas Duafala	Tri-Cal Research Division	US
Dieter H. Ehhalt	Institut für Chemie und Dynamik der Geosphäre	Germany
James W. Elkins	NOAA Climate Monitoring and Diagnostics Laboratory	US
Andreas Engel	Universität Frankfurt	Germany
Christine A. Ennis	NOAA Aeronomy Laboratory	US
David E. Etheridge	CSIRO Division of Atmospheric Research	Australia

Simon J. Evans	Imperial College	UK
David W. Fahey	NOAA Aeronomy Laboratory	US
Vitali Fioletov	Atmospheric Environment Service	Canada
Jack Fishman	NASA Langley Research Center	US
Eric L. Fleming	Space Applications Corporation	US
Lawrence E. Flynn	Software Corporation of America	US
Piers M. de F. Forster	University of Reading	UK
J. Paul F. Fortuin	Koninklijk Nederlands Meteorologisch Instituut (KNMI)	The Netherlands
Paul J. Fraser	CSIRO Division of Atmospheric Research	Australia
Lucien Froidevaux	California Institute of Technology/Jet Propulsion Laboratory	US
Michael Fromm	Computational Physics, Inc.	US
Jan Fuglestvedt	Universitetet I Oslo/CICERO	Norway
Dian J. Gaffen	NOAA Air Resources Laboratory	US
Marvin A. Geller	State University of New York at Stony Brook	US
Melvyn E. Gelman	NOAA NWS Climate Prediction Center	US
Sophie Godin	Service d'Aéronomie du CNRS	France
Allen Goldstein	College of Natural Resources	UK
George S. Golitsyn	Russian Academy of Sciences	Russia
Claire Granier	Service d'Aéronomie du CNRS; NOAA Aeronomy Laboratory	France/US
William B. Grant	NASA Langley Research Center	US
Jens-Uwe Grooß	Forschungszentrum Jülich	Germany
Robert Guicherit	Netherlands Organization for Applied Scientific Research (TNO)	The Netherlands
Michael R. Gunson	California Institute of Technology/Jet Propulsion Laboratory	US
James Hansen	NASA Goddard Institute for Space Studies	US
Jochen Harnisch	Max-Planck-Institut für Aeronomie	Germany
Neil R.P. Harris	European Ozone Research Coordinating Unit	UK
Dana E. Hartley	Georgia Institute of Technology	US
Fumio Hasebe	Ibaraki University	Japan
Alain Hauchecorne	Service d'Aéronomie du CNRS	France
Didier Hauglustaine	National Center for Atmospheric Research	US
Peter Haynes	Centre for Atmospheric Science, University of Cambridge	UK
Jay R. Herman	NASA Goddard Space Flight Center	US
Mark Hervig	University of Wyoming	US
David J. Hofmann	NOAA Climate Monitoring and Diagnostics Laboratory	US
Stacey M. Hollandsworth	Space Applications Corporation	US
James R. Holton	University of Washington	US
Øystein Hov	Norsk Institutt for Luftforskning (NILU)	Norway
Robert Hudson	University of Maryland	US
Drusilla Hufford	U.S. Environmental Protection Agency	US
Dale F. Hurst	NOAA Climate Monitoring and Diagnostics Laboratory/CIRES	US
Abdel Moneim Ibrahim	Egyptian Meteorological Authority	Egypt
Mohammad I. Ilyas	Universiti Sains Malaysia	Malaysia
Ivar S.A. Isaksen	Universitetet I Oslo	Norway
Tomoyuki Ito	Japan Meteorological Agency	Japan
Charles H. Jackman	NASA Goddard Space Flight Center	US
Daniel J. Jacob	Harvard University	US
Evgeny A. Jadin	Central Aerological Observatory	Russia
Horst Jäger	Fraunhofer-Institut für Atmosphärische Umweltforschung	Germany
Nicholas B. Jones	National Institute of Water & Atmospheric Research Ltd. (NIWA)	New Zealand

Rod Jones	Centre for Atmospheric Science, University of Cambridge	UK
Jan-Eiof Jonson	Universitetet I Oslo	Norway
Torben S. Jørgensen	Danmarks Meteorologiske Institut	Denmark
Kenneth W. Jucks	Harvard-Smithsonian Center for Astrophysics	US
Igor Karol	A.I. Voeikov Main Geophysical Observatory	Russia
David Karoly	Monash University	Australia
Randy Kawa	NASA Goddard Space Flight Center	US
Jack A. Kaye	NASA Headquarters	US
Philippe Keckhut	Service d'Aéronomie du CNRS	France
Hennie Kelder	Koninklijk Nederlands Meteorologisch Instituut (KNMI)	The Netherlands
James B. Kerr	Atmospheric Environment Service	Canada
Jhoon Kim	Korea Aerospace Research Institute	Republic of Korea
Douglas E. Kinnison	Lawrence Livermore National Laboratory	US
Volker Kirchhoff	Instituto Nacional de Pesquisas Espaciais (INPE)	Brazil
Dieter Kley	Forschungszentrum Jülich	Germany
Malcolm K.W. Ko	Atmospheric and Environmental Research, Inc.	US
Kunihiko Kodera	Meteorological Research Institute	Japan
Yutaka Kondo	Nagoya University	Japan
Yuri Koshelkov	Central Aerological Observatory	Russia
Nickolay Krotkov	Raytheon STX Inc.	US
Janusz Krzyścin	Polish Academy of Sciences	Poland
Michael Kuhn	Fraunhofer-Institut für Atmosphärische Umweltforschung	Germany
Michael J. Kurylo	National Institute of Standards and Technology; NASA Headquarters	US
Karin Labitzke	Freie Universität Berlin	Germany
G.J. Labow	Raytheon STX Inc.	US
William A. Lahoz	University of Reading	UK
Murari Lal	Indian Institute of Technology	India
Shyam Lal	Physical Research Laboratory	India
Neils Larsen	Danmarks Meteorologiske Institut	Denmark
David J. Lary	Centre for Atmospheric Science, University of Cambridge	UK
Yuan-Pern Lee	National Tsing Hua University	Taiwan
Franck Lefèvre	Météo-France, Centre National de Recherches Météorologiques	France
Jos Lelieveld	Institute for Marine and Atmospheric Research Utrecht (IMAU)	The Netherlands
Jacqueline Lenoble	Université des Sciences et Technologies de Lille 1	France
Kirsti Leszczynski	Radiation and Nuclear Safety Authority (STUK)	Finland
Joel M. Levy	NOAA Office of Global Programs	US
J.-J. Roger Lin	Research and Data Systems Corporation	US
Jennifer A. Logan	Harvard University	US
Craig S. Long	NOAA NWS Climate Prediction Center	US
Beiping Luo	Zhenjiang Rietschle Machinery Company, Ltd.	China
E.V. Lysenko	Central Aerological Observatory	Russia
Sasha Madronich	National Center for Atmospheric Research	US
Emmanuel Mahieu	Université de Liège	Belgium
Jerry D. Mahlman	NOAA Geophysical Fluid Dynamics Laboratory	US
Manfred Maiss	Max-Planck-Institut für Chemie	Germany
Gloria L. Manney	California Institute of Technology/Jet Propulsion Laboratory	US
Andrew Matthews	National Institute of Water & Atmospheric Research Ltd. (NIWA)	New Zealand
Bernhard Mayer	Fraunhofer-Institut für Atmosphärische Umweltforschung	Germany
M. Patrick McCormick	Hampton University	US

Archie McCulloch	ICI Chemicals & Polymers Ltd.	UK
Mack McFarland	DuPont Fluoroproducts	US
Danny McKenna	Forschungszentrum Jülich	Germany
Richard L. McKenzie	National Institute of Water & Atmospheric Research Ltd. (NIWA)	New Zealand
Richard D. McPeters	NASA Goddard Space Flight Center	US
Gérard Mégie	Service d'Aéronomie du CNRS	France
Pauline M. Midgley	M & D Consulting	Germany
Jana B. Milford	University of Colorado	US
Alvin J. Miller	NOAA NWS Climate Prediction Center	US
Forrest Mims	Sun Photometer Atmospheric Network	US
Philippe Mirabel	Université Louis Pasteur	France
Masa Miyauchi	Japan Meteorological Agency	Japan
Mario J. Molina	Massachusetts Institute of Technology	US
Stephen A. Montzka	NOAA Climate Monitoring and Diagnostics Laboratory	US
Gary A. Morris	Valparaiso University	US
Mike Mozurkewich	York University	Canada
Rolf Müller	Forschungszentrum Jülich	Germany
Gunnar Myhre	Universitetet I Oslo	Norway
Hideaki Nakane	National Institute for Environmental Studies	Japan
John Nash	U.K. Meteorological Office	UK
Patrick Neale	Smithsonian Environmental Research Center	US
Gerald E. Nedoluha	Naval Research Laboratory	US
Roland Neuber	Alfred Wegener Institute	Germany
Cynthia D. Nevison	National Center for Atmospheric Research	US
Paul A. Newman	NASA Goddard Space Flight Center	US
Ole J. Nielsen	Risø National Laboratory	Denmark
Paul C. Novelli	NOAA Climate Monitoring and Diagnostics Laboratory	US
Alan O'Neill	University of Reading	UK
Donal O'Sullivan	NorthWest Research Associates, Inc.	US
Samuel J. Oltmans	NOAA Climate Monitoring and Diagnostics Laboratory	US
Michael Oppenheimer	Environmental Defense Fund	US
David E. Oram	University of East Anglia	UK
John J. Orlando	National Center for Atmospheric Research	US
Mary Osborn	Science Applications International Corporation	US
Stuart A. Penkett	University of East Anglia	UK
Joyce Penner	University of Michigan	US
Thomas Peter	Max-Planck-Institut für Chemie	Germany
Ruben Piacentini	Planetario y Observatorio de Rosario	Argentina
Michel Pirre	Service d'Aéronomie du CNRS	France
Giovanni Pitari	Università degli Studi dell'Aquila	Italy
Alan Plumb	Massachusetts Institute of Technology	US
Jean-Pierre Pommereau	Service d'Aéronomie du CNRS	France
Lamont R. Poole	NASA Langley Research Center	US
Bob Portmann	NOAA Aeronomy Laboratory	US
Gilles Poulet	Laboratoire de Combustion et Systèmes Réactifs du CNRS	France
Michael Prather	University of California at Irvine	US
M. Margarita Préndez	Universidad de Chile	Chile
Ronald G. Prinn	Massachusetts Institute of Technology	US
John A. Pyle	Centre for Atmospheric Science, University of Cambridge	UK

V. Ramaswamy	NOAA Geophysical Fluid Dynamics Laboratory	US
William J. Randel	National Center for Atmospheric Research	US
Lakshman Randeniya	CSIRO Telecommunications and Industrial Physics	Australia
A.R. Ravishankara	NOAA Aeronomy Laboratory	US
Claire E. Reeves	University of East Anglia	UK
George C. Reid	NOAA Aeronomy Laboratory	US
Wolfgang Renger	Deutsche Forschungsanstalt für Luft- und Raumfahrt	Germany
Markus Rex	Alfred Wegener Institute	Germany
Curtis P. Rinsland	NASA Langley Research Center	US
Henning Rodhe	Stockholms Universitet	Sweden
José M. Rodríguez	Atmospheric and Environmental Research, Inc.	US
Bjoerg Rognerud	Universitetet I Oslo	Norway
Richard B. Rood	NASA Goddard Space Flight Center	US
James M. Rosen	University of Wyoming	US
Michel Rossi	Ecole Polytechnique Fédérale de Lausanne	Switzerland
James M. Russell III	Hampton University	US
Keith R. Ryan	CSIRO Telecommunications and Industrial Physics	Australia
Ken G. Ryan	Industrial Research Limited	New Zealand
Nelson A. Sabogal	United Nations Environment Programme	Kenya
Karen H. Sage	NASA Langley Research Center	US
Ross J. Salawitch	California Institute of Technology/Jet Propulsion Laboratory	US
Stan Sander	California Institute of Technology/Jet Propulsion Laboratory	US
Eugenio Sanhueza	Instituto Venezolano de Investigaciones Cientificas (IVIC)	Venezuela
Michelle L. Santee	California Institute of Technology/Jet Propulsion Laboratory	US
K.M. Sarma	United Nations Environment Programme	Kenya
Sue M. Schauffler	National Center for Atmospheric Research	US
Ulrich Schmidt	Universität Frankfurt	Germany
Ulrich Schumann	Deutsche Forschungsanstalt für Luft- und Raumfahrt	Germany
M. Daniel Schwarzkopf	NOAA Geophysical Fluid Dynamics Laboratory	US
Gunther Seckmeyer	Fraunhofer-Institut für Atmosphärische Umweltforschung	Germany
Theodore G. Shepherd	University of Toronto	Canada
Drew Shindell	NASA Goddard Institute for Space Studies	US
Keith P. Shine	University of Reading	UK
Masato Shiotani	Hokkaido University	Japan
Howard W. Sidebottom	University College Dublin	Ireland
Peter G. Simmonds	University of Bristol	UK
David E. Siskind	Naval Research Laboratory	US
Harry Slaper	National Institute of Public Health and the Environment (RIVM)	The Netherlands
James R. Slusser	Colorado State University	US
Herman Smit	Forschungszentrum Jülich	Germany
Sergei P. Smyshlyaev	Russian State Hydrometeorological Institute	Russia
Susan Solomon	NOAA Aeronomy Laboratory	US
Boris Soukharev	St. Petersburg State University	Russia
Johannes Stähelin	Eidgenössische Technische Hochschule Hönggerberg	Switzerland
L. Paul Steele	CSIRO Division of Atmospheric Research	Australia
Leopoldo Stefanutti	Istituto di Riccrea sulle Onde Elettromagnetiche del Consiglio Nazionale della Richerche	Italy
William R. Stockwell	Fraunhofer-Institut für Atmosphärische Umweltforschung	Germany

Richard Stolarski	NASA Goddard Space Flight Center	US
Frode Stordal	Norsk Institutt for Luftforskning (NILU)	Norway
William T. Sturges	University of East Anglia	UK
B.H. Subbaraya	Physical Research Laboratory	India
Michael E. Summers	Naval Research Laboratory	US
Richard Swinbank	NASA Goddard Space Flight Center	US
Petteri Taalas	Finnish Meteorological Institute	Finland
Xiao-Yan Tang	Peking University	China
Hubert Teyssèdre	Météo-France, Centre National de Recherches Météorologiques	France
Larry Thomason	NASA Langley Research Center	US
Anne M. Thompson	NASA Goddard Space Flight Center	US
Thayne M. Thompson	NOAA Climate Monitoring and Diagnostics Laboratory	US
Brian Toon	University of Colorado	US
Geoffrey C. Toon	California Institute of Technology/Jet Propulsion Laboratory	US
Ralf Toumi	Imperial College	UK
Kleareti Tourpali	Aristotle University of Thessaloniki	Greece
Michael Trainer	NOAA Aeronomy Laboratory	US
Jörg Trentmann	Max-Planck-Insitut für Chemie	Germany
Adrian Tuck	NOAA Aeronomy Laboratory	US
Osamu Uchino	Japan Meteorological Agency	Japan
Jan C. van der Leun	University Hospital	The Netherlands
Karel Vanicek	Czech Hydrometeorological Institute	Czech Republic
Guus J.M. Velders	National Institute of Public Health and the Environment (RIVM)	The Netherlands
Guido Visconti	Università degli Studi dell'Aquila	Italy
C. Michael Volk	Universität Frankfurt	Germany
Andreas Volz-Thomas	Forschungszentrum Jülich	Germany
Peter von der Gathen	Alfred Wegener Institute	Germany
Ray H.J. Wang	Georgia Institute of Technology	US
Wei-Chyung Wang	State University of New York at Albany	US
Joe Waters	California Institute of Technology/Jet Propulsion Laboratory	US
Robert T. Watson	The World Bank	US
Darryn Waugh	Johns Hopkins University	US
Debra K. Weisenstein	Atmospheric and Environmental Research, Inc.	US
Ray F. Weiss	University of California at San Diego	US
Paul O. Wennberg	California Institute of Technology	US
James C. Wilson	University of Denver	US
Martin Wirth	Deutsche Forschungsanstalt für Luft- und Raumfahrt	Germany
Russell Wooldridge	IDEA Corporation	US
Doug Worsnop	Aerodyne Research, Inc.	US
Donald J. Wuebbles	University of Illinois	US
Youko Yokouchi	National Institute for Environmental Studies	Japan
Vladimir Yushkov	Central Aerological Observatory	Russia
Shari A. Yvon-Lewis	NOAA Atlantic Oceanographic and Meteorological Laboratory	US
Ahmed Zand	Tehran University	Iran
R. Zander	Université de Liège	Belgium
Christos Zerefos	Aristotle University of Thessaloniki	Greece
Xiuji Zhou	Chinese Academy of Meteorological Sciences	China
Jerry Ziemke	Software Corporation of America	US

Chapter Editorial Contributors

Chapter 1: Long-Lived Ozone-Related Compounds

Nada Derek	CSIRO Division of Atmospheric Research	Australia
L. Kubrick	Massachusetts Institute of Technology	US
Kathy A. Wolfe	Computer Sciences Corporation	US
Diane Zander	Université de Liège	Belgium

Chapter 2: Short-Lived Ozone-Related Compounds

Kathy A. Wolfe	Computer Sciences Corporation	US

Chapter 5: Trends in Stratospheric Temperatures

Gail Haller	NOAA Geophysical Fluid Dynamics Laboratory	US
Cathy Raphael	NOAA Geophysical Fluid Dynamics Laboratory	US
Jeff Varanyak	NOAA Geophysical Fluid Dynamics Laboratory	US

Chapter 6: Upper Stratospheric Processes

Stephan Böttcher	Tel Aviv University	Israel

Chapter 7: Lower Stratospheric Processes

LeAnn Droppleman	NOAA Aeronomy Laboratory	US

Chapter 8: Tropospheric Ozone

Matthew G. Seybold	Steven Myers and Associates Corporation/ NASA Goddard Space Flight Center	US

Chapter 11: Scenarios for the Future Ozone Layer and Related Consequences

Chris Fischer	National Center for Atmospheric Research	US

Chapter 12: Predicting Future Ozone Changes and Detection of Recovery

Ann Thorne	NOAA Climate Monitoring and Diagnostics Laboratory	US

Sponsoring Organizations Liaisons

Rumen D. Bojkov   World Meteorological Organization   Switzerland

K.M. Sarma   United Nations Environment Programme   Kenya

Daniel L. Albritton   National Oceanic and Atmospheric Administration   US

Michael J. Kurylo   National Institute of Standards and Technology; NASA Headquarters   US

Georgios T. Amanatidis   European Commission   Belgium