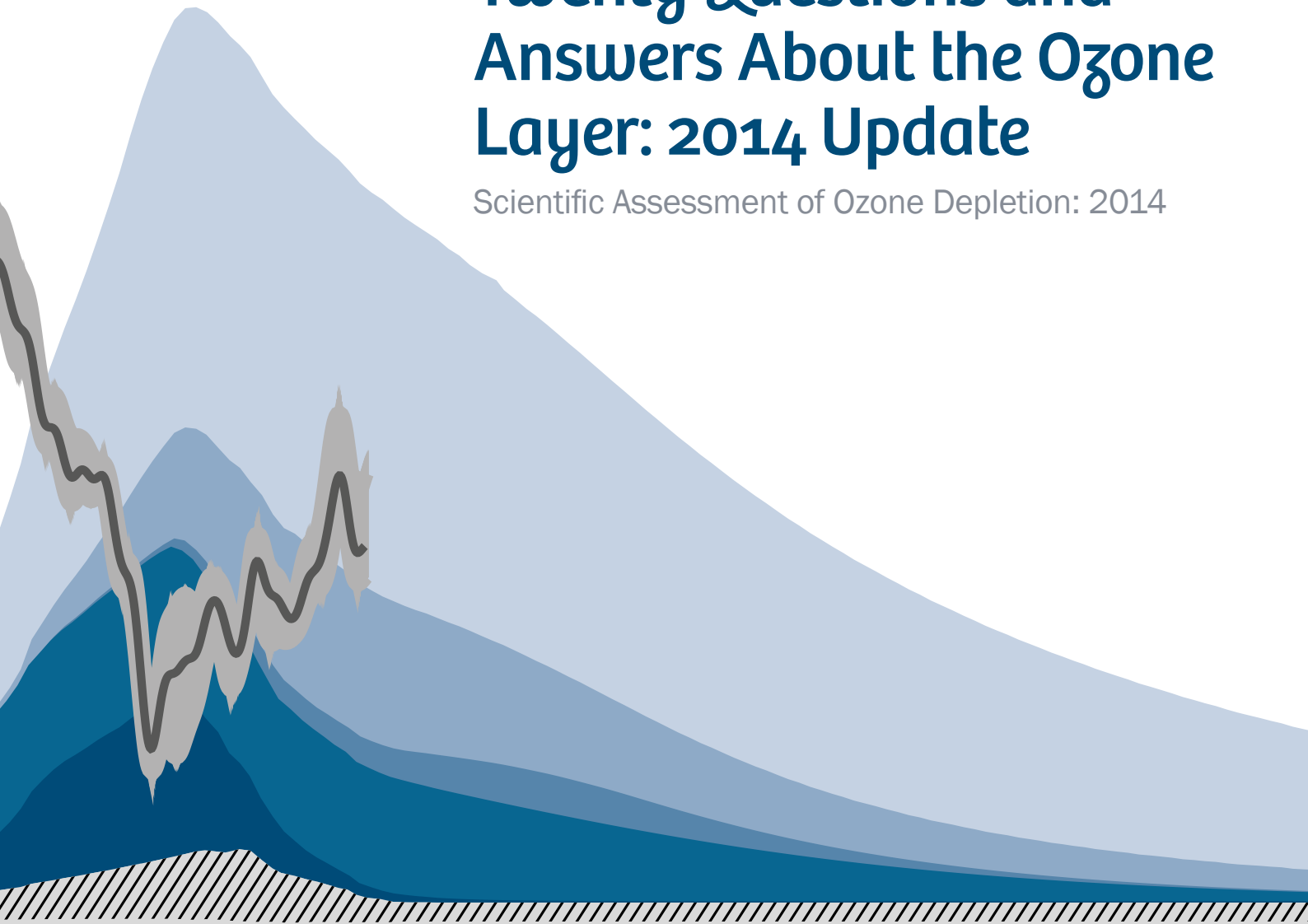# Twenty Questions and Answers About the Ozone Layer: 2014 Update

Scientific Assessment of Ozone Depletion: 2014

World Meteorological Organization
United Nations Environment Programme
National Oceanic and Atmospheric Administration
National Aeronautics and Space Administration
European Commission

NOAA
NATIONAL OCEANIC AND ATMOSPHERIC ADMINISTRATION
U.S. DEPARTMENT OF COMMERCE

UNEP

WMO OMM

NASA
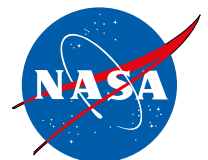
What is ozone and where is it in the atmosphere?

Ozone is a gas that is naturally present in our atmosphere. Each ozone molecule contains three atoms of oxygen and is denoted chemically as O_3. Ozone is found primarily in two regions of the atmosphere. About 10% of atmospheric ozone is in the troposphere, the region closest to Earth (from the surface to about 10–16 kilometers (6–10 miles)). The remaining ozone (about 90%) resides in the stratosphere between the top of the troposphere and about 50 kilometers (31 miles) altitude. The large amount of ozone in the stratosphere is often referred to as the "ozone layer."

Ozone is a gas that is naturally present in our atmosphere. Ozone has the chemical formula O_3 because an ozone molecule contains three oxygen atoms (see Figure Q1-1). Ozone was discovered in laboratory experiments in the mid-1800s. Ozone's presence in the atmosphere was later discovered using chemical and optical measurement methods. The word ozone is derived from the Greek word όζειν *(ozein)*, meaning "to smell." Ozone has a pungent odor that allows it to be detected even at very low amounts. Ozone reacts rapidly with many chemical compounds and is explosive in concentrated amounts. Electrical discharges are generally used to produce ozone for industrial processes such as air and water purification and bleaching of textiles and food products.

Ozone location. Most ozone (about 90%) is found in the stratosphere, which begins about 10–16 kilometers (6–10 miles) above Earth's surface and extends up to about 50 kilometers (31 miles) altitude. The stratospheric region with the highest ozone concentration is commonly known as the "ozone layer" (see Figure Q1-2). The ozone layer extends over the entire globe with some variation in altitude and thickness. Most of the remaining ozone, about 10%, is found in the troposphere, which is the lowest region of the atmosphere, between Earth's surface and the stratosphere.

Ozone abundance. Ozone molecules have a low relative abundance in the atmosphere. In the stratosphere near the peak concentration of the ozone layer, there are typically a few thousand ozone molecules for every *billion* air molecules (1 billion = 1,000 million). Most air molecules are either oxygen (O_2) or nitrogen (N_2). In the troposphere near Earth's surface, ozone is even less abundant, with a typical range of 20 to 100 ozone molecules for each billion air molecules. The highest surface values result when ozone is formed in air polluted by human activities.
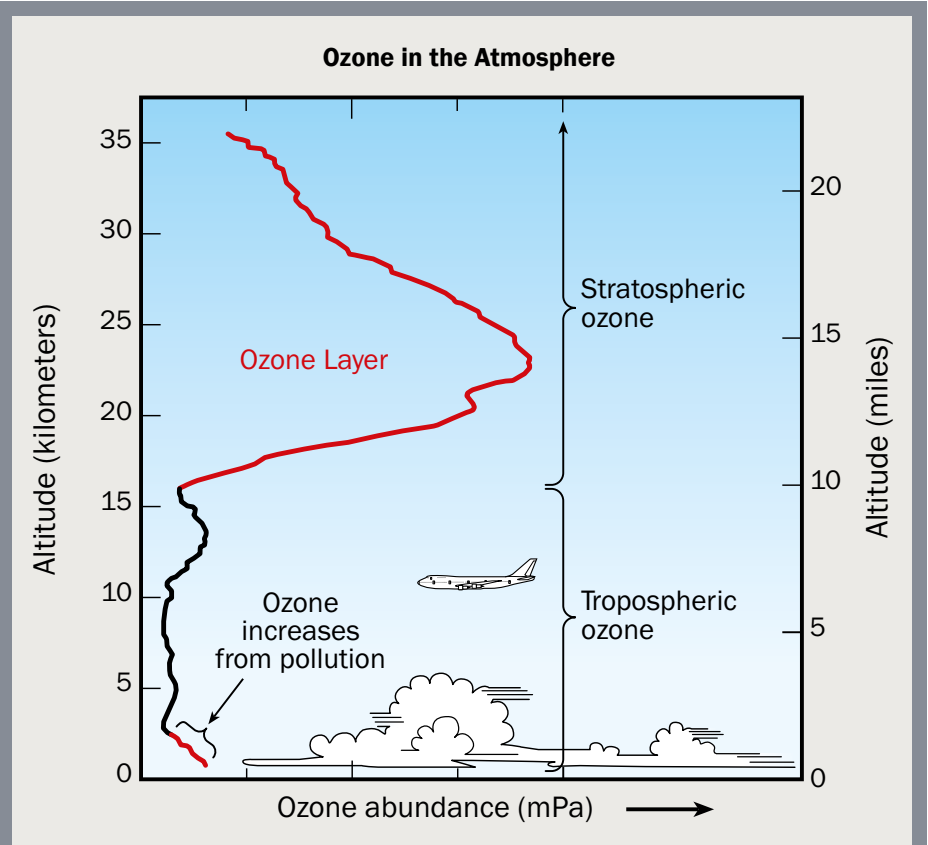
Figure Q1-1 Ozone and oxygen.
A molecule of ozone (O_3) contains three oxygen atoms (O) bound together. Oxygen molecules (O_2), which constitute 21% of the gases in Earth's atmosphere, contain two oxygen atoms bound together.

Ozone and Oxygen

Oxygen atom (O)

Oxygen molecule (O_2)
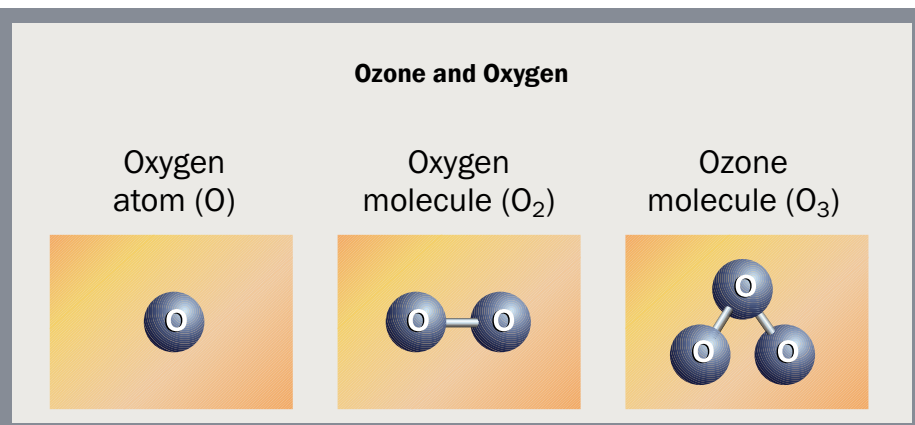
Ozone molecule (O_3)

As an illustration of the low relative abundance of ozone in our atmosphere, one can imagine bringing all the ozone molecules in the troposphere and stratosphere down to Earth's surface and forming a layer of pure ozone that extends over the entire globe. The resulting layer would have an average thickness of about three millimeters (about one-tenth inch) (see Q4). Nonetheless, this extremely small fraction of the atmosphere plays a vital role in protecting life on Earth (see Q3).

Figure Q1-2 Ozone in the atmosphere.
Ozone is present throughout the troposphere and stratosphere. This profile shows schematically how ozone changes with altitude in the tropics. Most ozone resides in the stratospheric "ozone layer." The vertical extent or thickness of this layer varies from region to region and with season over the globe (see Q4). Increases in ozone occur near the surface as a result of pollution from human activities.

Ozone in the Atmosphere

Ozone Layer

Stratospheric ozone

Tropospheric ozone

Ozone increases from pollution

Altitude (kilometers)

Altitude (miles)

Ozone abundance (mPa)

Q2

How is ozone formed in the atmosphere?

Ozone is formed throughout the atmosphere in multistep chemical processes that require sunlight. In the stratosphere, the process begins with an oxygen molecule (O_2) being broken apart by ultraviolet radiation from the Sun. In the lower atmosphere (troposphere), ozone is formed by a different set of chemical reactions that involve naturally occurring gases and those from pollution sources.

Stratospheric ozone. Stratospheric ozone is formed naturally by chemical reactions involving solar ultraviolet radiation (sunlight) and oxygen molecules, which make up 21% of the atmosphere. In the first step, solar ultraviolet radiation breaks apart one oxygen molecule (O_2) to produce two oxygen atoms (2 O) (see Figure Q2-1). In the second step, each of these highly reactive atoms combines with an oxygen molecule to produce an ozone molecule (O_3). These reactions occur continually whenever solar ultraviolet radiation is present in the stratosphere. As a result, the largest ozone production occurs in the tropical stratosphere.
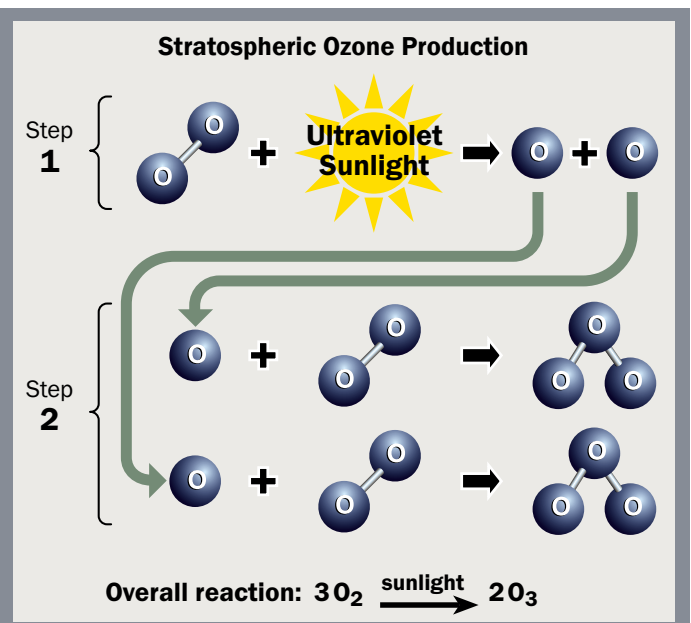
The production of stratospheric ozone is balanced by its destruction in chemical reactions. Ozone reacts continually with sunlight and a wide variety of natural and human-produced chemicals in the stratosphere. In each reaction, an ozone molecule is lost and other chemical compounds are produced. Important reactive gases that destroy ozone are hydrogen and nitrogen oxides and those containing chlorine and bromine (see Q8).

Some stratospheric ozone is regularly transported down into the troposphere and can occasionally influence ozone amounts at Earth's surface, particularly in remote, unpolluted regions of the globe.

Tropospheric ozone. Near Earth's surface, ozone is produced by chemical reactions involving gases from natural and pollution sources. Ozone production reactions primarily involve hydrocarbon and nitrogen oxide gases, as well as ozone itself, and all

Figure Q2-1. Stratospheric ozone production.
Ozone is produced naturally in the stratosphere by a two-step reaction process. In the first step, solar ultraviolet radiation (sunlight) breaks apart an oxygen molecule to form two separate oxygen atoms. In the second step, each atom collides with another oxygen molecule and forms an ozone molecule in a binding reaction. In the overall process, three oxygen molecules plus sunlight react to form two ozone molecules.

Stratospheric Ozone Production

Step 1

Ultraviolet Sunlight

Step 2

Overall reaction: $3 O_2 \xrightarrow{\text{sunlight}} 2 O_3$

require sunlight for completion. Fossil fuel combustion is a primary source of pollutant gases that lead to tropospheric ozone production. The production of ozone near the surface does not significantly contribute to the abundance of stratospheric ozone. The amount of surface ozone is too small in comparison and the transport of surface air to the stratosphere is not effective enough. As in the stratosphere, ozone in the troposphere is destroyed by naturally occurring chemical reactions and by reactions involving human-produced chemicals. Tropospheric ozone can also be destroyed when ozone reacts with a variety of surfaces, such as those of soils and plants.

Balance of chemical processes. Ozone abundances in the stratosphere and troposphere are determined by the balance between chemical processes that produce and destroy ozone. The balance is determined by the amounts of reactive gases and how the rate or effectiveness of the various reactions varies with sunlight intensity, location in the atmosphere, temperature, and other factors. As atmospheric conditions change to favor ozone-producing reactions in a certain location, ozone abundances increase. Similarly, if conditions change to favor other reactions that destroy ozone, abundances decrease. The balance of production and loss reactions, combined with atmospheric air motions that transport and mix air with different ozone abundances, determines the global distribution of ozone on timescales of days to many months (see also Q4). Global ozone has decreased during the past several decades (see Q13) because the amounts of reactive gases containing chlorine and bromine have increased in the stratosphere due to human activities.

Why do we care about atmospheric ozone?

Ozone in the stratosphere absorbs a large part of the Sun's biologically harmful ultraviolet radiation. Stratospheric ozone is considered "good" ozone because of this beneficial role. In contrast, ozone formed at Earth's surface in excess of natural amounts is considered "bad" ozone because it is harmful to humans, plants, and animals. Natural ozone near the surface and in the lower atmosphere plays an important beneficial role in chemically removing pollutants from the atmosphere.

Good ozone. Stratospheric ozone is considered good for humans and other life forms because it absorbs ultraviolet-B (UV-B) radiation from the Sun (see Figure Q3-1). If not absorbed, UV-B radiation would reach Earth's surface in amounts that are harmful to a variety of life forms. In humans, increased exposure to UV-B radiation increases the risks of skin cancer, cataracts, and a suppressed immune system. UV-B radiation exposure before adulthood and cumulative exposure are both important health risk factors. Excessive UV-B exposure also can damage terrestrial plant life, single-cell organisms, and aquatic ecosystems. Other UV radiation, UV-A, which is not absorbed significantly by ozone, causes premature aging of the skin.

Protecting good ozone. In the mid-1970s, it was discovered that gases containing chlorine and bromine atoms released by human activities could cause stratospheric ozone depletion (see Q6 and Q7). These gases, referred to as halogen source gases, and as ozone-depleting substances (ODSs), chemically release their chlorine and bromine atoms after they reach the stratosphere. Ozone depletion increases surface UV-B radiation above naturally occurring amounts. International efforts have been successful in protecting the ozone layer through controls on ODS production and consumption (see Q15 and Q16).

Bad ozone. Ozone near Earth's surface in excess of natural amounts is considered bad ozone. It is formed by reactions involving human-made pollutant gases. Increasing surface ozone above natural levels is harmful to humans, plants, and other living systems because ozone reacts strongly to destroy or alter many biological molecules. High ozone exposure caused by air pollution reduces crop yields and forest growth. In humans, exposure to high levels of ozone can reduce lung capacity; cause chest pains, throat irritation, and coughing; and worsen pre-existing health conditions related to the heart and lungs. In addition, increases in tropospheric ozone lead to a warming of Earth's surface because ozone is a greenhouse gas (see Q18). The negative effects of excess tropospheric ozone contrast sharply with the protection from harmful UV-B radiation afforded by an abundance of stratospheric ozone.

Reducing bad ozone. Limiting the emission of certain common pollutants reduces the production of excess ozone in the air surrounding humans, plants, and animals. Natural emissions from the biosphere, mainly from trees, also participate in reactions that produce ozone. Major sources of pollutants include large cities where fossil fuel consumption and industrial activities are greatest. Many programs around the globe have already been successful in reducing or limiting the emission of pollutants that cause production of excess ozone near Earth's surface.

Natural ozone. In the absence of human activities, ozone would still be present near Earth's surface and throughout the troposphere and stratosphere because ozone is a natural component of the clean atmosphere. Ozone plays important roles in the atmosphere beyond absorbing UV radiation. For example, ozone initiates the chemical removal of many pollutants, such as carbon monoxide (CO) and nitrogen oxides

(NO_x), as well as some greenhouse gases, such as methane (CH_4). In addition, the absorption of UV-B radiation by ozone is a natural source of heat in the stratosphere, causing temperatures to increase with altitude. Stratospheric temperatures affect the balance of ozone production and destruction processes (see Q2) and air motions that redistribute ozone throughout the stratosphere (see Q4).

UV Protection by the Ozone Layer

Figure Q3-1. UV protection by the ozone layer.
The ozone layer resides in the stratosphere and surrounds the entire Earth. UV-B radiation (280- to 315-nanometer (nm) wavelength) from the Sun is strongly absorbed in this layer. As a result, the amount of UV-B reaching Earth's surface is greatly reduced. UV-A (315- to 400-nm wavelength), visible light, and other solar radiation are not strongly absorbed by the ozone layer. Human exposure to UV-B radiation increases the risks of skin cancer, cataracts, and a suppressed immune system. UV-B radiation exposure can also damage terrestrial plant life, single-cell organisms, and aquatic ecosystems.
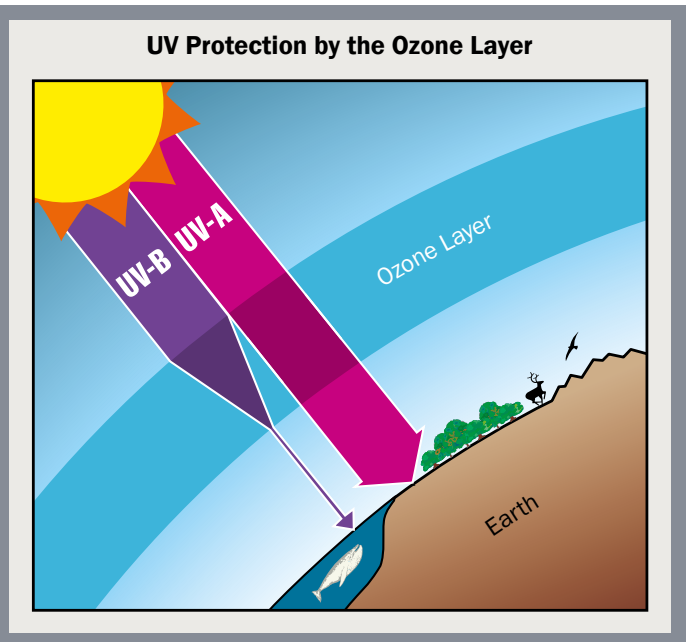
Q8

What are the reactive halogen gases that destroy stratospheric ozone?

Emissions from human activities and natural processes represent a large source of chlorine- and bromine-containing gases that enter the stratosphere. When exposed to ultraviolet radiation from the Sun, these halogen source gases are converted to more reactive gases containing chlorine and bromine. Some reactive gases act as chemical reservoirs that convert to form the most reactive gases, namely chlorine monoxide (ClO) and bromine monoxide (BrO). The most reactive gases participate in catalytic reactions that efficiently destroy ozone. Most volcanoes emit some reactive halogen gases that readily dissolve in water and are usually washed out of the atmosphere before they can reach the stratosphere.

Halogen-containing gases present in the stratosphere can be divided into two groups: *halogen source gases and reactive halogen gases* (Figure Q8-1). The source gases, which include ozone-depleting substances (ODSs), are emitted at Earth's surface by natural processes and by human activities (see Q7). Once in the stratosphere, the halogen source gases chemically convert at different rates to form the reactive halogen gases. The conversion occurs in the stratosphere for most gases instead of the troposphere because solar ultraviolet radiation (a component of sunlight) is more intense in the stratosphere. Reactive gases containing the halogens chlorine and bromine lead to the chemical destruction of stratospheric ozone.

Stratospheric Halogen Gases
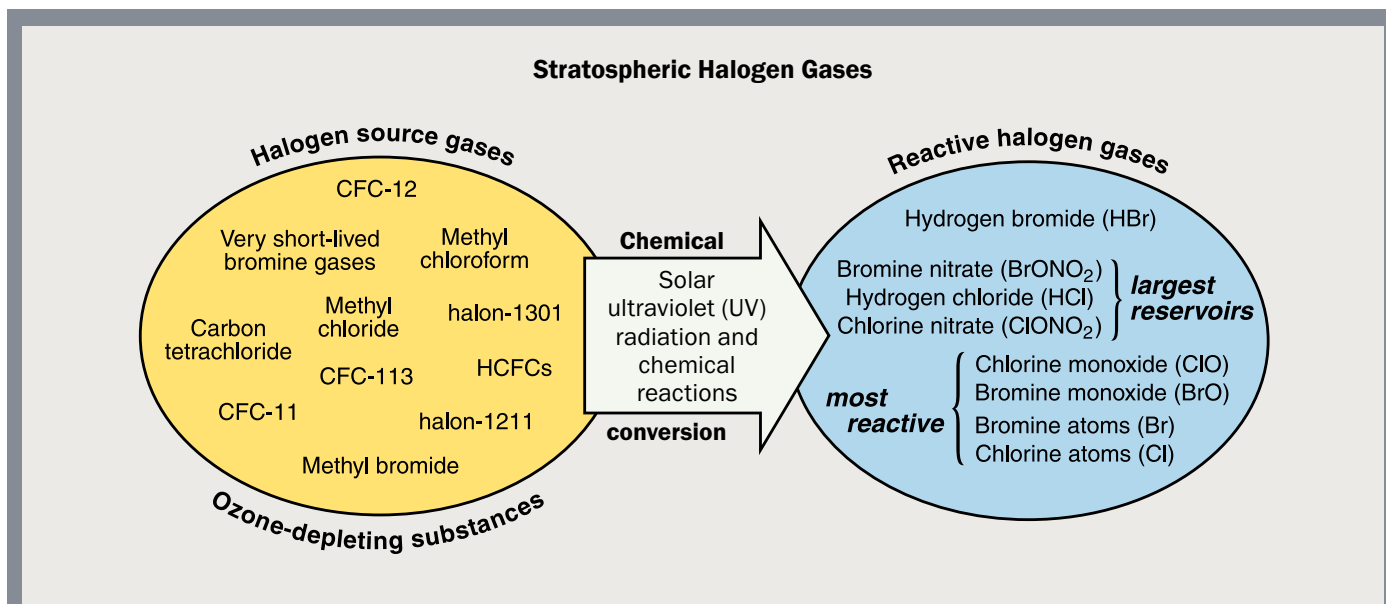
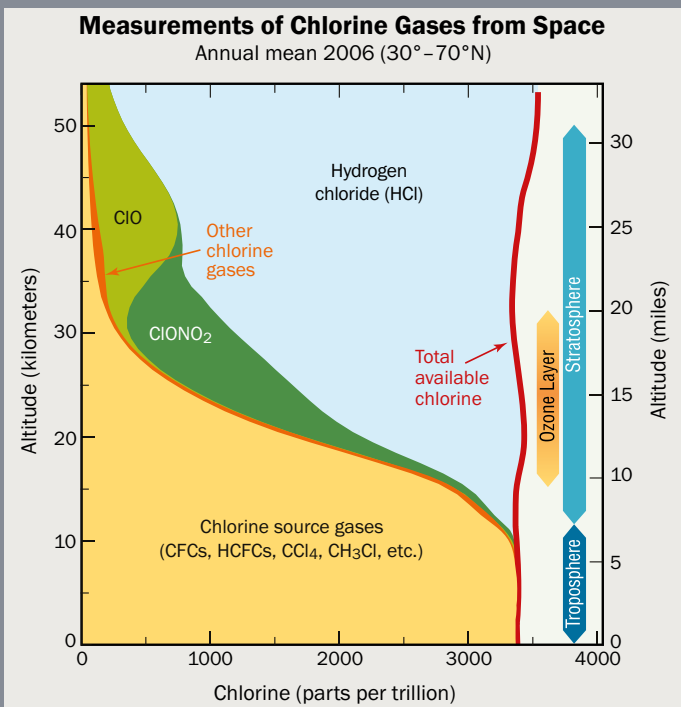**Figure Q8-1. Conversion of halogen source gases.**
Halogen source gases containing chlorine and bromine are chemically converted to reactive halogen gases primarily in the stratosphere. The conversion requires solar ultraviolet radiation and a few chemical reactions. The shorter-lived gases undergo some conversion in the troposphere. The reactive halogen gases contain all the chlorine and bromine originally present in the source gases before conversion. The reactive gases can be grouped into the reservoir gases, which do not directly destroy ozone, and the most reactive gases, which participate in ozone destruction cycles (see Q9). A principal reactive gas, ClO, is formed by reactions of the reservoir gases HCl and ClONO$_2$ that occur on the surfaces of liquid and solid polar stratospheric clouds (PSCs) (see Q10).

Figure Q8-2. Reactive chlorine gas observations. The abundances of chlorine source gases and reactive chlorine gases as measured from space in 2006 are displayed with altitude for a midlatitude location. In the troposphere (below about 10 kilometers), all chlorine is contained in the source gases. In the stratosphere, the abundances of reactive chlorine gases increase with altitude as the amounts of chlorine source gases decline. This is a consequence of chemical reactions initiated by solar ultraviolet radiation that convert source gases to reactive gases (see Figure Q8-1). The principal reactive chlorine gases formed are HCl, $ClONO_2$, and ClO. Summing the source gases with the reactive gases gives "Total available chlorine," which is nearly constant with altitude throughout the stratosphere. In the ozone layer (15-25 km), chlorine source gases are still present and HCl and $ClONO_2$ are the most abundant reactive chlorine gases. (The unit "parts per trillion" is defined in the caption of Figure Q7-1.)

Reactive halogen gases. The chemical conversion of halogen source gases, which involves solar ultraviolet radiation and other chemical reactions, produces a number of reactive halogen gases. These reactive gases contain all of the chlorine and bromine atoms originally present in the source gases. The most important reactive chlorine- and bromine-containing gases that form in the stratosphere are shown in Figure Q8-1. Throughout the stratosphere, the most abundant are typically hydrogen chloride (HCl) and chlorine nitrate ($ClONO_2$). These two gases are considered important *reservoir* gases because, while they don't react directly with ozone, they can be converted to the *most reactive* forms that do chemically destroy ozone. The most reactive forms are chlorine monoxide (ClO) and bromine monoxide (BrO), and chlorine and bromine atoms (Cl and Br). A large fraction of available reactive bromine is generally in the form of BrO, whereas usually only a small fraction of reactive chlorine is in the form of ClO. The special conditions that occur in the polar regions in winter cause the reservoir gases $ClONO_2$ and HCl to undergo nearly complete conversion to ClO in reactions on polar stratospheric clouds (PSCs) (see Q10).

Reactive chlorine at midlatitudes. Reactive chlorine gases have been observed extensively in the stratosphere using both local and remote measurement techniques. The measurements from space displayed in Figure Q8-2 are representative of how the amounts of chlorine-containing gases change between the surface and the upper stratosphere at mid- to high latitudes. Available chlorine (see red line in Figure Q8-2) is the sum of chlorine contained in halogen source gases and in the reactive gases (e.g., CFC-11, $ClONO_2$, ClO). Available chlorine is constant to within about 10% from the surface to above 50 kilometers (31 miles) altitude. In the troposphere, available chlorine is contained almost entirely in the source gases described in Figure Q7-1. At higher altitudes, the source gases become a smaller fraction of available chlorine as they are converted to the reactive chlorine gases. At the highest altitudes, available chlorine is all in the form of the reactive chlorine gases.

In the altitude range of the ozone layer at midlatitudes, as shown in Figure Q8-2, the reservoir gases HCl and $ClONO_2$ account for most of the available chlorine. ClO, the most

reactive gas in ozone depletion, is a small fraction of available chlorine. The low abundance of ClO limits the amount of ozone destruction that occurs outside of polar regions.

Reactive chlorine in polar regions. Reactive chlorine gases in polar regions undergo large changes between the fall and late winter. Meteorological and chemical conditions in both polar regions are now routinely observed from space in all seasons. Fall and winter conditions over the Antarctic are contrasted in Figure Q8-3 using seasonal observations made near the center of the ozone layer (about 18 km (11.3 miles) altitude) (see Figure Q12-3).

In fall (May), ozone values are high over the entire Antarctic continent and beyond. Temperatures are mid-range, HCl and nitric acid (HNO_3) are high, and ClO is very low. High HCl indicates that substantial conversion of halogen source gases has occurred in the stratosphere. In the past decades, the reservoir gases HCl and $ClONO_2$ have increased substantially in the stratosphere following increased emissions of halogen source gases. HNO_3 is an abundant, naturally occurring stratospheric compound that plays major roles in stratospheric ozone chemistry by both moderating ozone destruction and condensing to form polar stratospheric clouds (PSCs). Low ClO indicates that little conversion of the reservoir gases occurs in the fall, thereby limiting catalytic ozone destruction.

By late winter (September), ozone amounts reflect substantial depletion at 18-km altitude over an area larger than the Antarctic continent. Antarctic ozone holes arise from similar chemical destruction throughout much of the altitude range of the ozone layer (see altitude profile in Figure Q12-3). The meteorological and chemical conditions in late winter are very different from those found in fall: very low temperatures, very low HCl and HNO_3, and very high ClO. Low stratospheric temperatures are characteristic of winter, when solar heating is reduced. Low HCl and high ClO reflect the conversion of the reactive halogen reservoir compounds, HCl and $ClONO_2$, to the most reactive chlorine form, ClO. This conversion occurs selectively in winter on PSCs, which form at the very low temperatures (see Q10). Low HNO_3 is indicative of its condensation to form PSCs, some of which subsequently descend to lower altitudes through gravitational settling. High ClO abundances generally cause ozone depletion to continue in the Antarctic region until mid-October (spring), when the lowest ozone values usually are observed (see Q11). As temperatures rise at the end of the winter, PSC formation is halted, ClO is converted back into the reservoir species HCl and $ClONO_2$ (see Q10), and ozone destruction is curtailed.

Similar though less dramatic changes in meteorological and chemical conditions are also observed between fall and late winter in the Arctic, where ozone depletion is less severe than in the Antarctic.

Reactive bromine observations. Fewer measurements are available for reactive bromine gases in the lower stratosphere than for reactive chlorine, in part because of the lower abundance of bromine. The most widely observed bromine gas is bromine monoxide (BrO), which can be observed from space. Estimates of reactive bromine abundances in the stratosphere are larger than expected from the conversion of the halons and methyl bromide source gases, suggesting that the contribution of the very short-lived bromine-containing gases to reactive bromine must also be significant (see Q7).

Other sources. Other emission sources of reactive halogen gases exist that are associated with natural processes and human activities. Most emissions become trapped in the troposphere when dissolved in water, and ultimately are returned to Earth's surface before they can reach the stratosphere. Volcanoes are an important episodic source of reactive halogen gases (e.g., HCl) that do not reach the stratosphere in appreciable quantities (see Q14). Other examples include reactive chlorine produced by evaporation of ocean spray. Sea salt products dissolve in water and are removed in the troposphere. Solid rocket motors, such as those used on the Space Shuttle, release
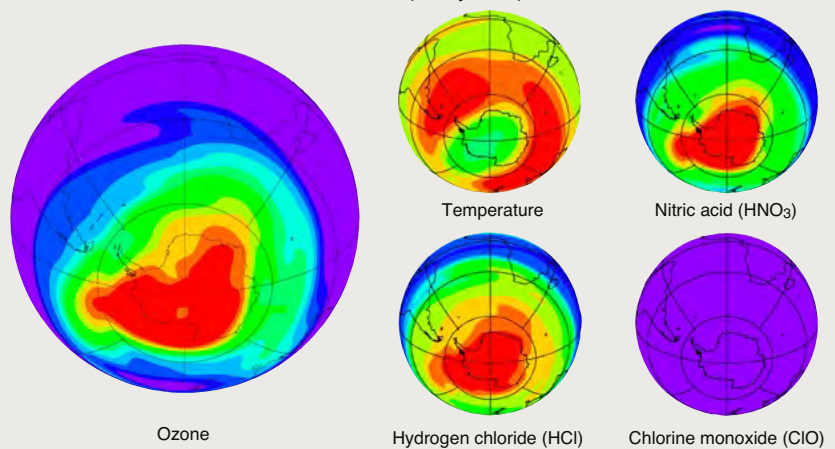
reactive chlorine gases directly into the troposphere and stratosphere. The quantities emitted by the Space Shuttle were small in comparison with halogen emissions from other human activities.

Figure Q8-3. Chemical conditions in the ozone layer over Antarctica.
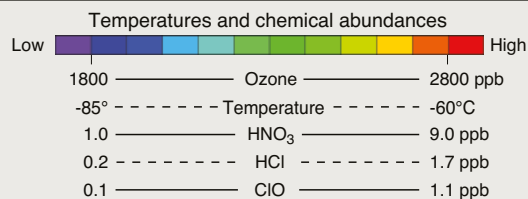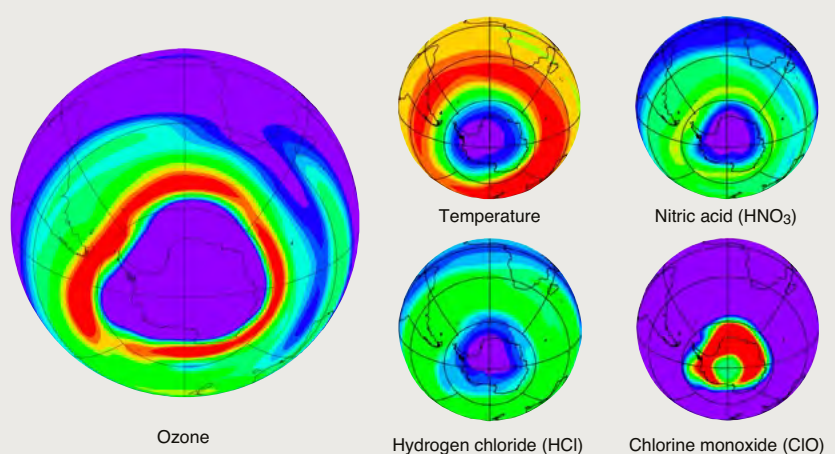
Observations of the chemical conditions in the Antarctic region highlight the changes associated with severe ozone depletion that forms the ozone hole. Satellite instruments now routinely monitor ozone, reactive chlorine gases, and temperatures in the global stratosphere. Results are shown here for fall (May) and late winter (September) seasons in Antarctic winter for a narrow altitude region near 18 kilometers (11.2 miles) within the ozone layer (see Figure Q12-3). Ozone has naturally high values in fall, before the onset of ozone destruction reactions that cause widespread depletion. The high ozone is accompanied by moderate temperatures, high values of HCl and HNO_3, and very low ClO amounts. When ClO is not enhanced, significant ozone destruction from ozone-depleting substances does not occur. Chemical conditions are quite different in late winter when ozone undergoes severe depletion. Temperatures are much lower, HCl has been converted to ClO (the most reactive chlorine gas), and HNO_3 has been removed by the gravitational settling of PSC particles. ClO values closely surrounding the South Pole are low in September because ClO formation requires sunlight, which is still gradually returning to the highest latitudes. The high ClO values in late winter covering an extensive area that at times exceeds that of the Antarctic continent can last for 1 to 2 months, and efficiently destroy ozone in sunlit regions in late winter/early spring. Ozone typically reaches its minimum values in early to mid-October (see Q12). Note that the first and last colors in the color bar represent values outside the indicated range of values. (The unit "parts per billion," abbreviated "ppb," is used here as a measure of the relative abundance of a gas in air: 1 part per billion equals the presence of one molecule of a gas per billion (=10^9) total air molecules (compare to ppt in Figure Q7-1).)

Chemical Conditions Observed in the Ozone Layer Over Antarctica

Normal ozone amounts in fall (1 May 2008) at 18-km altitude

Ozone — Temperature — Nitric acid (HNO_3) — Hydrogen chloride (HCl) — Chlorine monoxide (ClO)

Large ozone depletion in late winter (15 September 2008) at 18-km altitude

Ozone — Temperature — Nitric acid (HNO_3) — Hydrogen chloride (HCl) — Chlorine monoxide (ClO)

Temperatures and chemical abundances

| Low		High
1800	Ozone	2800 ppb
-85°	Temperature	-60°C
1.0	HNO_3	9.0 ppb
0.2	HCl	1.7 ppb
0.1	ClO	1.1 ppb

Are there controls on the production of ozone-depleting substances?

Yes, the production and consumption of ozone-depleting substances are controlled under a 1987 international agreement known as the "Montreal Protocol on Substances that Deplete the Ozone Layer" and by its subsequent Amendments and adjustments. The Protocol, now ratified by all 196 United Nations members, establishes legally binding controls on national production and consumption of ozone-depleting substances (ODSs). Production and consumption of all principal ODSs by developed and developing nations will be almost completely phased out before the middle of the 21st century.

Montreal Protocol. In 1985, a treaty called the *Vienna Convention for the Protection of the Ozone Layer* was signed by 20 nations in Vienna. The signing nations agreed to take appropriate measures to protect the ozone layer from human activities. The Vienna Convention was a framework agreement that supported research, exchange of information, and future protocols. In response to growing concern, the *Montreal Protocol on Substances that Deplete the Ozone Layer* was signed in 1987 and, following sufficient country ratification, entered into force in 1989. The Protocol has been successful in establishing legally binding controls for developed and developing nations on the production and consumption of halogen source gases known to cause ozone depletion. Halogen source gases containing chlorine and bromine controlled under the Montreal Protocol are referred to as ozone-depleting substances (ODSs). National consumption of an ODS is defined as the amount that production and imports of the substance exceed its export to other nations. The Protocol provisions are structured for developed countries to act first and for developing countries to follow with some financial assistance. In 2009, the Montreal Protocol became the first international treaty to achieve universal ratification by all 196 United Nations members.

Effect of the Montreal Protocol
Long-term changes in equivalent effective stratospheric chlorine (EESC)
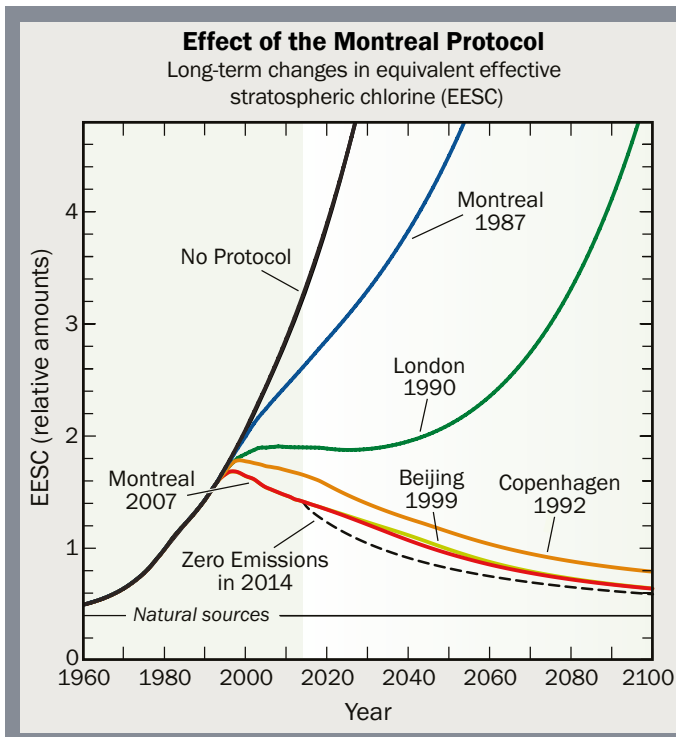
Figure Q15-1. Effect of the Montreal Protocol. The objective of the Montreal Protocol is the protection of the ozone layer through control of the global production and consumption of ODSs. Projections of the future abundances of ODSs expressed as equivalent effective stratospheric chlorine (EESC) values (see Q16) are shown separately for the midlatitude stratosphere for (1) no Protocol provisions, (2) the provisions of the original 1987 Montreal Protocol and some of its subsequent Amendments and adjustments, and (3) zero emissions of ODSs starting in 2014. The city names and years indicate where and when changes to the original 1987 Protocol provisions were agreed upon (see Figure Q0-1). EESC is a relative measure of the potential for stratospheric ozone depletion that combines the contributions of chlorine and bromine from ODS surface observations (see Q16). Without the Protocol, EESC values are projected to have increased significantly in the 21st century. Only with the Copenhagen (1992) and subsequent Amendments and adjustments did projected EESC values show a long-term decrease.

Amendments and adjustments. As the scientific basis of ozone depletion became more certain after 1987 and substitutes and alternatives became available to replace ODSs, the Montreal Protocol was strengthened with Amendments and adjustments. Each Amendment is named after the city in which the meeting of the Parties to the Montreal Protocol took place and by the year of the meeting. The timeline in Figure Q0-1 shows some of the major decisions that have been adopted in the last three decades. These decisions have put additional ODSs under control, accelerated the timing of existing control measures, and prescribed phase-out dates for the production and consumption of certain gases. The initial Protocol called for a 50% reduction in chlorofluorocarbon (CFC) production and a freeze on halon production. The 1990 London Amendment called for a phase-out of the production and consumption of the most damaging ODSs in developed nations by 2000 and in developing nations by 2010. The 1992 Copenhagen Amendment accelerated the phase-out date to 1996 in developed nations. Further controls on ODSs were agreed upon in later meetings in Vienna (1995), Montreal (1997, 2007), and Beijing (1999).

Influence of the Montreal Protocol. Montreal Protocol controls are based on several factors that are considered separately for each ODS. The factors include (1) the effectiveness in depleting ozone in comparison with other substances (see Ozone Depletion Potential, ODP, in Q18), (2) the availability of suitable substitutes for domestic and industrial use, and (3) the potential impact of controls on developing nations. The influence of Montreal Protocol provisions on stratospheric ODS abundances can be demonstrated with long-term changes in *equivalent effective stratospheric chlorine* (EESC). Calculations of EESC combine the amounts of chlorine and bromine present in surface air to form a measure of the potential for ozone destruction in a particular stratospheric region on an annual basis (see definition in Q16). The long-term changes in EESC at midlatitudes are shown in Figure Q15-1 for several cases:

- **No Protocol.** Without the Montreal Protocol the production, use, and emissions of CFCs and other ozone-depleting substances is expected to have increased after 1987 with an annual growth rate of about 3% (business-as-usual scenario). As a result, EESC is projected to have increased nearly 10-fold by the mid-2050s compared with the 1980 value. Computer models of the atmosphere indicate that such high EESC values would have at least doubled global total ozone depletion between 1990 and 2010 and increased it far beyond that by midcentury. As a result, harmful UV-B radiation would have increased substantially at Earth's surface, causing damage to ecosystem health and a global rise in skin cancer and cataract cases (see Q17).

- **Montreal Protocol provisions.** International compliance with only the 1987 provisions of the Montreal Protocol and the later 1990 London Amendment would have substantially slowed the projected growth of EESC. Not until the 1992 Copenhagen Amendments and adjustments did the Protocol projections show a decrease in future EESC values. The provisions became more stringent with the Amendments and adjustments adopted in Beijing in 1999 and Montreal in 1997 and 2007. Now, with full compliance to the Protocol, most ODSs will be almost completely phased out, with some exemptions for critical uses (see Q16). Global EESC is slowly decaying from its peak value in the late 1990s and is expected to reach 1980 values in the mid-21st century. The success of the Montreal Protocol to date is demonstrated by the decline in ODP-weighted emissions of ODSs shown in Figure Q0-1. Total emissions peaked in 1987 at values about 10-fold higher than natural emissions. Between 1987 and 2014, ODS emissions from human activities have decreased by almost 80%.

- **Zero emissions.** EESC values in the coming decades will be influenced by (1) the slow natural removal of ODSs still present in the atmosphere, (2) emissions from continued production and use of ODSs, and (3) emissions from currently existing *banks* containing a variety of ODSs. ODS banks are associated with applications that involve long-term containment of halogen gases. Examples are CFCs in refrigeration equipment and insulating foams, and halons in fire-fighting equipment. New emissions are projected based on continued production and consumption of ODSs, particularly in developing nations, under existing Protocol provisions.

  The zero-emissions case demonstrates the EESC values that would occur if it were possible to set all ODS emissions to zero beginning in 2014. This would eliminate the contributions from new production and bank emissions. Significant differences from the Montreal 2007 projections are evident in the first decades following 2014 because the phase-out of all ODS production under the Protocol is not yet complete and bank emissions are substantial. Zero emissions would bring forward the return of EESC to 1980 levels by about 13 years.

HCFC substitute gases. The Montreal Protocol provides for the use of hydrochlorofluorocarbons (HCFCs) as transitional, short-term substitute compounds for ODSs with higher ODPs, such as CFC-12. HCFCs are used for refrigeration, in making insulating foams, and as solvents, all of which were primary uses of CFCs. HCFCs are generally more reactive in the troposphere than other ODSs because they contain hydrogen (H) in addition to chlorine, fluorine, and carbon. HCFCs are 88 to 98% less effective than CFC-12 in depleting stratospheric ozone because their chemical removal occurs primarily in the troposphere (see ODPs in Table Q7-1). This removal protects stratospheric ozone from most of the halogen content of HCFC emissions. In contrast, CFCs and some other ODSs release all of their halogen content in the stratosphere because they are chemically inert in the troposphere (see Q6).

Under the provisions of the Montreal Protocol, developed and developing countries may continue to use HCFCs as ODS substitutes in the coming decades before they are ultimately phased out. In the most recent adjustment to the Protocol (Montreal 2007), the phase-out of HCFCs for all Parties was accelerated so that it will be complete in 2030, a decade earlier than in previous provisions. In adopting this decision, the Parties reduced the contribution of HCFC emissions to both long-term ozone depletion and future climate forcing (see Q18 and Q19).

HFC substitute gases. Hydrofluorocarbons (HFCs) are used as substitute compounds for CFCs, HCFCs, and other ODSs. HFCs contain hydrogen, fluorine, and carbon. HFCs do not contribute to ozone depletion because they contain no chlorine, bromine, or iodine. As a consequence, HFCs are not ODSs and are not subject to the provisions of the Montreal Protocol. HFCs and all ODSs are radiatively active gases that contribute to human-induced climate change based on their accumulation in the atmosphere (see Q18). HFCs are included in the basket of gases controlled by the Kyoto Protocol of the United Nations Framework Convention on Climate Change (UNFCCC). The Kyoto Protocol is an international treaty designed to protect climate by controlling emissions of carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), HFCs, perfluorocarbons (PFCs), and sulfur hexafluoride (SF_6). HFC emissions are expected to grow substantially in coming decades as the demand for their use as substitute gases and in new applications increases in developed and developing countries. The climate change contribution from future HFC emissions would be minimized if HFCs with very low Global Warming Potentials (GWPs) (less than 100) were chosen to meet the growing demand (see Q18 and Q19).