Chapter 5

Ozone and Ultraviolet Radiation

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Summary

Depletion of stratospheric ozone over the Arctic can reduce normally high winter and spring ozone levels and allow more ultraviolet (UV) radiation to reach the surface of the earth. Arctic ozone levels exhibit high natural seasonal and interannual variability, driven primarily by atmospheric dynamics that govern the large-scale meridional transport of ozone from the tropics to high latitudes. The spatial distribution of total column ozone over the Arctic is less symmetric around the pole than is the case for ozone over the Antarctic. The large natural variability in arctic ozone complicates the interpretation of past changes and the projection of future ozone levels. Observations have shown substantial late winter and early spring reductions in arctic total column ozone over the last two decades. These reductions have been directly linked to chemical reactions occurring at low temperatures in the presence of anthropogenic chlorine and bromine compounds. Between 1979 and 2000, the trend in mean annual total column ozone over the Arctic was about -3% per decade (7% accumulated loss), while the trend in mean spring total column ozone was about -5% per decade (11% accumulated loss). Arctic ozone depletion is also strongly affected by stratospheric temperatures and polar stratospheric cloud formation. Climate change leading to lower temperatures in the stratosphere is likely to increase the frequency and severity of ozone-depletion episodes.

Ozone levels directly influence the amount of UV radiation reaching the surface of the earth. Surface UV radiation levels are also strongly affected by clouds, aerosols, altitude, solar zenith angle, and surface albedo. These different factors contribute to high variability in UV radiation levels and make it difficult to identify changes that result from ozone depletion. Because of the low solar elevation in the Arctic, the region is subject to an increased proportion of diffuse UV radiation, from scattering in the atmosphere as well as from reflectance off snow and ice. Reflectance off snow can increase the biologically effective irradiance by more than 50%. Changes in global climate are likely to result in changes in arctic snow cover and sea ice. Snow and ice cover strongly attenuate UV radiation, protecting organisms underneath. A reduction in snow and ice cover on the surface of rivers, lakes, or oceans is likely to increase the exposure of many organisms to damaging UV radiation. Loss of snow or ice cover earlier in the spring, when UV radiation is very likely to be at increased levels, would be stressful for both aquatic and terrestrial life.

Ground-based measurements of UV radiation levels are conducted in all arctic countries. However, the current monitoring network does not provide sufficient coverage over vast regions. Available individual measurements suggest localized increases in UV radiation levels reaching the surface, but the measurement time series are not yet long enough to allow trends to be detected. Reconstructed time series of surface UV radiation levels based on total column ozone, sunshine duration, and cloud cover suggest distinct increases, but reconstruction methods are less certain than direct measurements because they involve assumptions about the spectral characteristics of cloud and aerosol attenuation and surface reflectivity. The increases in UV radiation levels occur primarily in the spring, when ozone depletion reaches a maximum, and can result in spring UV radiation levels that are higher than those measured during the summer.

Atmospheric sampling indicates that the Montreal Protocol and its amendments have already resulted in a leveling off of some atmospheric halogen concentrations. However, climate change and other factors are likely to complicate the recovery of the ozone layer. Changes in both the overall meteorology of the region and in atmospheric composition may delay or accelerate the recovery of the arctic ozone layer. Ozone levels are projected to remain depleted for several decades and thus surface UV radiation levels in the Arctic are likely to remain elevated in the coming years. The elevated levels are likely to be most pronounced in the spring, when ecosystems are most sensitive to harmful UV radiation. Exposure to UV radiation has been linked to skin cancers, corneal damage, cataracts, immune suppression, and aging of the skin in humans, and can also have deleterious effects on ecosystems and on materials.

5.1. Introduction

Ultraviolet radiation levels reaching the surface of the earth are directly influenced by total column ozone amounts and other geophysical parameters. In the Arctic, UV radiation is of particular concern, particularly during the spring and summer when the region experiences more hours of sunshine compared to lower latitudes. Goggles found in archaeological remains suggest that indigenous peoples had developed protection from sunlight long before the onset of anthropogenic ozone depletion (e.g., Hedblom, 1961; Sliney, 2001). Although systematic measurements of UV radiation levels have been performed for little more than decade, analysis of fossil pigments in leaf sediments suggests that past UV radiation levels in the Arctic may have been similar to modern-day (pre-depletion) levels (Leavitt et al., 1997). In recent years, however, Arctic ozone depletion (which has sometimes been severe) has allowed more UV radiation to reach the surface. In the years since ozone depletion was first observed over the Arctic, UV radiation effects such as sunburn and increased snow blindness have been reported in regions where they were not previously observed (Fox, 2000).

Less than 10% of the solar energy reaching the top of the atmosphere is in the UV spectral region, with wavelengths between 100 and 400 nm. The shortest wavelengths (100–280 nm) are referred to as UV-C

radiation. Radiation at these wavelengths is almost entirely absorbed by atmospheric oxygen and ozone, preventing it from reaching the surface. Wavelengths between 280 and 315 nm comprise the UV-B portion of the spectrum (while some communities use 320 nm to mark the division between UV-B and UV-A radiation, it is the convention in this report to use 315 nm). Ultraviolet-B radiation is absorbed efficiently but not completely by atmospheric ozone. Wavelengths between 315 and 400 nm are referred to as UV-A radiation. Absorption of UV-A radiation by atmospheric ozone is comparatively small.

Although the intensity of solar UV-B radiation is low, the energy per photon is high. Due to this high energy, UV-B radiation can have several harmful impacts on human beings (i.e., DNA damage, skin cancers, corneal damage, cataracts, immune suppression, aging of the skin, and erythema), on ecosystems, and on materials (e.g., UNEP, 1998, 2003). These effects are discussed in detail in sections 7.3, 7.4, 8.6, 9.4, 14.12, 15.3.3, 16.3.1, and 17.2.2.3. Ultraviolet-B radiation also affects many photochemical processes, including the formation of tropospheric ozone from gases released into the environment by motor vehicles or other anthropogenic sources.

The amount of UV radiation reaching the surface of the earth is expressed in terms of irradiance, denoting the radiant power per unit area reaching a surface. Figure 5.1 shows typical spectral irradiance in the UV-A and UV-B wavelengths for the Arctic. The values were obtained using a radiative transfer model with a solar zenith angle of 56.5°, total column ozone of 300 Dobson units (DU), and surface albedo of 0.6.

The exposure necessary to produce some biological effect, such as erythema (skin reddening), at each

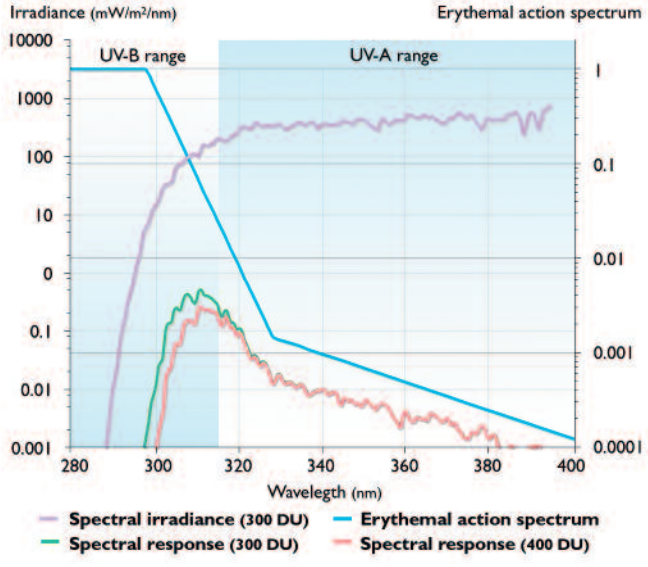

Fig. 5.1. Spectral UV irradiance in the UV-A and UV-B wavelengths (for total column ozone of 300 DU), the CIE erythemal action spectrum, and biological response curves for total column ozone of 300 and 400 DU.

wavelength in the UV spectral region is given by an action spectrum. In general, shorter UV-B wavelengths have greater biological effects than longer UV-A wavelengths, and action spectra account for this wavelength dependence. The action spectra are used to provide biological weighting factors to determine sensitivities to UV radiation exposure. The action spectrum often used to estimate human health effects is the McKinlay-Diffey erythemal response spectrum (McKinlay and Diffey, 1987). This curve is shown in Fig. 5.1 and represents the standard erythemal action spectrum adopted by the Commission Internationale de l'Eclairage (CIE) to represent the average skin response over the UV-B and UV-A regions of the spectrum (CIE, 1998).

The biological response is determined by multiplying the spectral irradiance at each wavelength by the biological weighting factor provided by the action spectrum. As ozone levels decrease, the biological response increases (see Fig. 5.1). Integrating the product of the spectral irradiance and the biological weighting factor over all wavelengths provides a measure of the biologically effective UV irradiance, or dose rate, with units W/m^2. This dose rate is scaled to produce a UV index value (WHO, 2002), which is made available to the public to provide an estimate of the level of UV radiation reaching the surface in a particular area at a particular time. Summing the dose rate over the exposure period (e.g., one day) results in a measure of the biologically effective radiation exposure, or dose, with units J/m^2. In the Arctic, the extended duration of sunlight during the summer can result in moderately large UV radiation doses. When considering biological impacts, it is important to distinguish that the definition of dose presented here differs slightly from that used by biologists, who refer to dose as the amount actually absorbed by the receptor. In addition, for some biological effects the cumulative dose model outlined above is too simple, because dose history also plays a role. In many cases, repair mechanisms cause the dose received over a longer time period to have less effect than a single, intense exposure.

Although some exposure to UV radiation can be beneficial, increases in surface UV radiation doses can have detrimental effects on humans and organisms in the Arctic. The levels of UV radiation reaching the surface are affected not only by total column ozone and solar zenith angle, but also by cloudiness, surface reflectance (albedo), and atmospheric aerosol concentrations. Climate change is likely to affect both future cloudiness and the extent of snow and ice cover in the Arctic, in turn leading to local changes in the intensity of solar UV radiation. It is very likely that climate change is already influencing stratospheric dynamics, which are very likely to in turn affect ozone depletion and surface UV radiation levels in the future. This chapter addresses some of the factors influencing total column ozone and surface UV irradiance, and describes both observed and projected changes in arctic ozone and UV radiation levels.

5.2. Factors affecting arctic ozone variability

Ozone in the atmosphere prevents most harmful UV radiation from reaching the biosphere. About 90 to 95% of atmospheric ozone is found in the stratosphere; the remaining 5 to 10% is in the troposphere. Most of the stratospheric ozone is produced by photochemical reactions in equatorial regions; at high latitudes, there is less photochemical ozone production and much of the stratospheric ozone is imported from low latitudes by the Brewer-Dobson circulation. This diabatic circulation also distributes ozone to lower altitudes in the high latitude regions, where, owing to a longer photochemical lifetime, it accumulates. For these reasons, total column ozone tends to exhibit global maxima near the poles. The atmospheric circulation varies seasonally, and oscillations in the circulation patterns explain some of the natural spatial, seasonal, and annual variations in the global total ozone distribution. In the Northern Hemisphere, the maximum total column ozone usually occurs in spring and the minimum in autumn. Solar activity also causes small fluctuations in total column ozone in phase with the solar cycle.

In addition to natural ozone production and destruction processes (WMO, 1995, 1999, 2003), stratospheric ozone is destroyed by heterogeneous chemical reactions involving halogens, particularly chlorine and bromine, which are derived from chlorofluorocarbons (CFCs) and other ozone-depleting substances. In the presence of solar radiation, extremely low stratospheric temperatures facilitate ozone depletion chemistry. Thus, ozone depletion can occur in relatively undisturbed polar vortices (see section 5.2.2, Box 5.1) with the return of sunlight in early spring. The fundamental processes governing ozone levels over the Arctic and Antarctic are the same, however, relative rates of production and destruction can differ. Low temperatures within the stable Antarctic vortex and the presence of ozone-depleting gases have led to an area of large-scale ozone depletion, the "ozone hole", which has been observed every spring since the 1980s. In contrast, the arctic polar vortex is less stable, resulting in arctic air masses that are on average warmer than air masses over the Antarctic. However, chemical ozone depletion has been observed over the Arctic during springs when temperatures in the arctic stratosphere were lower than normal. The decreases over the Arctic and Antarctic have both been sizeable (Fig. 5.2), although climatological spring ozone levels over the Arctic tend to be higher than those over the Antarctic, so that total column ozone after depletion events is higher in the Arctic than in the Antarctic. The depletion observed over the Antarctic in spring 2002 was not as severe as in previous years, but this was due to exceptional meteorological conditions and does not indicate an early recovery of the ozone layer.

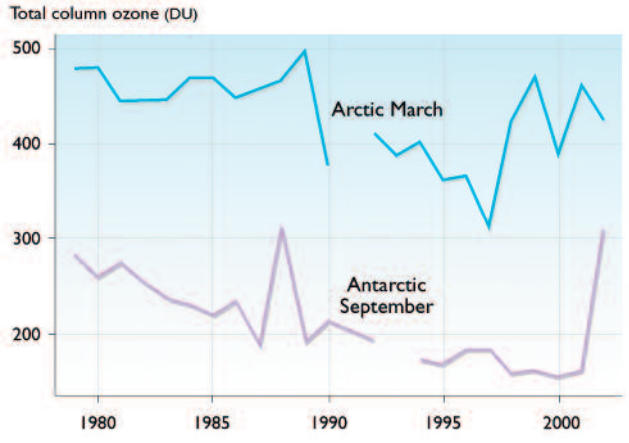

Fig. 5.2. Spring ozone depletion over the Antarctic and the Arctic between 1979 and 2002.

Since the late 1980s, much attention has been directed to studying ozone depletion processes over the Arctic. Arctic ozone levels have been significantly depleted in the past decade, particularly during the late winter and early spring (seasons when pre-depletion ozone levels were historically higher than at other times of the year). Several studies (Austin, 1992, 1994; Austin and Butchart, 1992; Austin et al., 1995; Guirlet et al., 2000) have focused on both the chemical and dynamic factors contributing to this depletion. These factors have combined to change the overall concentrations and distribution of ozone in the arctic stratosphere (e.g., Weatherhead, 1998; WMO, 1995, 1999, 2003), and the observed changes have not been symmetric around the North Pole. The greatest changes in ozone levels have been observed over eastern Siberia and west toward Scandinavia.

Ozone depletion can increase the level of UV radiation reaching the surface. These increased UV doses, particularly when combined with other environmental stressors, are very likely to cause significant changes to the region's ecosystems. Ozone depletion has been greatest in the spring, when most biological systems are particularly sensitive to UV radiation. The depletion has not been constant over time: very strong ozone depletion has been observed in some years while very little depletion has been observed in other years.

Transport of low-ozone air masses from lower latitudes can result in a few days of very low ozone and high UV radiation levels (Weatherhead, 1998). This transport of low-ozone air masses is often observed in late winter or early spring and is likely to have occurred naturally for decades. Climate change is likely to change transport patterns and is therefore likely to alter the frequency and severity of these events (Schnadt and Dameris, 2003).

5.2.1. Halogens and trace gases

Chlorine and bromine compounds cause chemically induced ozone depletion in the arctic stratosphere (E.C., 2003; Solomon, 1999; WMO, 1999, 2003). The source gases for these halogens are predominantly anthropogenic (E.C., 2003; WMO, 1999, 2003) and

are transported to the polar stratosphere over a period of 3 to 6.5 years (Harnisch et al., 1996; Schmidt and Khedim, 1991; Volk et al., 1997). In the stratosphere, the source gases are converted through photolysis and reaction with the hydroxyl radical to inorganic species of bromine, chlorine, and fluorine. The halogens are normally present as reservoir species (primarily hydrogen chloride – HCl, chlorine nitrate – $ClONO_2$, and bromine nitrate), which are efficiently converted into photochemically active species in the presence of sulfate aerosols or polar stratospheric clouds (WMO, 1999, 2003). Subsequently, in the presence of sunlight, reactive compounds (e.g., chlorine monoxide, bromine monoxide) are formed that react with and destroy stratospheric ozone in catalytic cycles.

The concentrations of chlorine measured in the stratosphere correspond well with the concentrations of CFCs and related gases that have been measured in the troposphere (Chang et al., 1996; Russell et al., 1996; Zander et al., 1996). From the mid-1980s to the early 1990s, the atmospheric chlorine concentration increased approximately 3 to 4% per year (WMO, 1990, 1992), while between 1995 and 1997, the rate of stratospheric chlorine increase was estimated at 1.8±0.3% per year (WMO, 1999). An analysis of long-term trends in total column inorganic chlorine through 2001, based on 24 years of HCl and $ClONO_2$ data, showed a broad plateau in inorganic chlorine levels after 1996 (Rinsland et al., 2003). Some uncertainty remains concerning the time lag between reductions in emissions of chlorine-containing compounds at the surface and chlorine concentrations in the stratosphere (e.g., Waugh et al., 2001), although this lag is thought to be between 3 and 5 years on average. Other studies report an estimated total organic bromine growth rate of 2.2% per year (Butler et al., 1998; Wamsley et al., 1998), although errors in the experimental method make the stratospheric bromine mixing ratios more difficult to determine. More recently, Montzka et al. (2003) reported that total organic bromine amounts in the troposphere have decreased since 1998.

Changing concentrations of the trace gases nitrous oxide (N_2O), methane (CH_4), water vapor, and carbon dioxide (CO_2) directly affect ozone chemistry and also alter local atmospheric temperatures by radiative cooling or heating, influencing the reaction rates of ozone depletion chemistry and the formation of ice particles. All of these gases emit radiation efficiently to space from the stratosphere (although CO_2 and water vapor are the most important), so increases in the abundances of these gases are very likely to lead to stratospheric cooling. In the polar regions, this cooling is very likely to lead to ozone depletion through heterogeneous chemistry. Lower temperatures facilitate the formation of polar stratospheric cloud particles, which play a role in transforming halogens to reactive compounds that can destroy ozone very rapidly. Small changes in temperature have been shown to have a significant effect on ozone levels (e.g., Danilin et al., 1998; Tabazadeh et al., 2000).

The trace gases N_2O, CH_4, and water vapor are also important chemically. In the stratosphere, CH_4 acts as an important source of water vapor and is also a sink for reactive chlorine. In addition, stratospheric water vapor is an important source of hydrogen oxide radicals, which play an important role in ozone destruction. Evans et al. (1998), Dvortsov and Solomon (2001), Shindell (2001), and Forster and Shine (2002) have studied the effects of water vapor on homogeneous chemistry. Their model results suggest that increases in water vapor reduce ozone levels in the upper stratosphere, increase ozone levels in the middle stratosphere, and reduce ozone levels in the lower stratosphere. Ozone levels in the lower stratosphere dominate total column ozone, and the model results differ most in this region. In the simulations of Evans et al. (1998), reductions in lower-stratospheric ozone levels occur only in the tropics when water vapor increases, while in the other simulations, the reductions extend to the mid-latitudes or the poles. The models of Dvortsov and Solomon (2001) and Shindell (2001) projected a slower recovery of the ozone layer as a result of increased stratospheric water vapor, and a 1 to 2% reduction in ozone levels over the next 50 years compared to what would be expected if water vapor did not increase.

Water vapor affects heterogeneous chemistry by enhancing the formation of polar stratospheric clouds (PSCs). This effect may be much more important than the relatively small impacts of water vapor on homogeneous chemistry. Kirk-Davidoff et al. (1999) projected a significant enhancement of arctic ozone depletion in a more humid atmosphere. Much of this projected effect is based on the radiative cooling of the stratosphere assumed to be induced by water vapor, a value that is currently uncertain. A smaller value would imply a reduced role for water vapor in enhancing PSC formation. Even using a smaller cooling rate, however, the impact on ozone is likely to be large, as the ~3 ºC cooling of the stratosphere projected to occur if CO_2 concentrations double is of comparable magnitude to the cooling that would be caused by a water vapor increase of only ~2 ppmv. Although precise quantification of radiative forcing due to water vapor is difficult, an estimate by Tabazadeh et al. (2000) suggests that an increase of 1 ppmv in stratospheric water vapor (with constant temperature) would be equivalent to a ~1 ºC decrease in stratospheric temperature and would cause a corresponding increase in PSC formation. This comparison suggests that the radiative impact of water vapor is larger than its effects on chemistry or microphysics. Given the potential for atmospheric changes in the Arctic, and the large ozone losses that could result from a slight cooling (Tabazadeh et al., 2000), it is important both to understand trends in stratospheric water vapor, and to resolve differences in model projections of the radiative impact of those trends.

Changing concentrations of the trace gases N_2O and CH_4 may also affect ozone levels. Nitrous oxide breaks

down to release nitrogen oxide radicals, which are extremely reactive and play an important role in ozone chemistry. Increases in CH_4 concentrations lead to an increase in hydrogen oxide radicals but at the same time increase the sequestration of chlorine radicals into HCl. The effects of increases in these gases on ozone depletion are thought to be relatively small (e.g., Shindell et al., 1998b; WMO, 1999), although a recent study by Randeniya et al. (2002) suggests that increasing concentrations of N_2O may have a larger impact than previously thought.

5.2.2. Arctic ozone depletion and meteorological variability

Partitioning the transport and chlorine chemistry contributions to arctic ozone variability is a subject of much discussion (Shepherd, 2000). The degree of ozone depletion in the Arctic depends strongly on air temperatures and PSC formation. Several methods have been used to estimate the total column ozone depletion in the arctic polar vortex based on meteorological measurements (e.g., Goutail et al., 1999; Manney et al., 1996; Müller et al., 1997; Rex et al., 1998), and comparisons between the different studies show good agreement (Harris et al., 2002). Since 1988–1989, three winters (1994–1995, 1995–1996, and 1999–2000) have had particularly low stratospheric temperatures and were characterized by PSC formation in both the early and late parts of the season (Braathen et al., 2000; Pawson and Naujokat, 1999). Some of the most severe arctic ozone losses (up to 70% at 18 km altitude) were observed during those winters (Knudsen et al., 1998; Rex et al., 1999; 2002).

Chipperfield and Pyle (1998) used models to investigate the sensitivity of ozone depletion to meteorological variability, chlorine and bromine concentrations, denitrification, and increases or decreases in stratospheric water vapor. Although the models tended to underestimate observed rates of arctic ozone depletion, their results agreed at least qualitatively with empirical estimates of ozone depletion, which suggest that substantial arctic ozone depletion is possible when both early and late winter temperatures in the stratosphere are extremely low. Cold early winters or cold late winters alone are not enough to produce extensive ozone depletion, but can still cause depletion to occur. During the winters of 1993–1994 and 1996–1997, temperatures in the arctic stratosphere were very low in late winter compared to earlier in the season. Ozone losses at specific altitudes during these years were of the order of 40 to 50% (Braathen et al., 2000; Schulz et al., 2000).

Dynamic processes dominate the short-term (day-to-day) variability in winter and spring total column ozone at mid- and high latitudes. Local changes in total column ozone of the order of 100 DU have been frequently reported (e.g., Peters et al., 1995) and are linked to three main transport processes:

1. A shift in the location of the polar vortex leads to changes in total column ozone, because the polar vortex air masses are characterized by low ozone levels compared to air masses outside the vortex.
2. Tropical upper-tropospheric high-pressure systems moving to higher latitudes cause an increase in the height of the tropopause at those latitudes, and thus a reduction in the overall depth of the stratospheric air column, as a result of divergence, resulting in ozone redistribution and a decline in total column ozone (e.g., James, 1998).
3. Tropical lower-stratospheric or upper-tropospheric air masses may be mixed into the stratosphere at higher latitudes. Referred to as "streamers" (e.g., Kouker et al., 1999), these phenomena introduce lower ozone content to the high-latitude air masses.

These three transport processes are not independent and can occur simultaneously, potentially increasing total column ozone variability.

Box 5.1. The polar vortex and polar stratospheric clouds

Winter and early spring ozone levels in the Arctic are influenced by the *polar vortex,* a large-scale cyclonic circulation in the middle and upper troposphere. This circulation keeps ozone-rich mid-latitude air from reaching the vortex region and can also lead to very cold air temperatures within the vortex.

Cold temperatures allow the formation of *polar stratospheric clouds* (PSCs), which play two important roles in polar ozone chemistry. First, the particles support chemical reactions leading to active chlorine formation, which can catalytically destroy ozone. Second, nitric acid removal from the gas phase can increase ozone loss by perturbing the reactive chlorine and nitrogen chemical cycles in late winter and early spring (WMO, 2003).

As the stratosphere cools, two types of PSCs can form. *Type-1* PSCs are composed of frozen nitric acid and water and form at temperatures below 195 K. At temperatures below 190 K, *Type-2* PSCs may form. Type-2 PSCs are composed of pure frozen water and contain particles that are much larger than the Type-1 PSC particles. Both types of PSCs occur at altitudes of 15 to 25 km and can play a role in ozone depletion chemistry, although Type-2 PSCs are quite rare in the Northern Hemisphere.

The occurrence of ozone minima or ozone "mini-holes" at northern mid- and high latitudes caused by tropopause lifting (process 2) exhibits high interannual variability. James (1998) found no detectable trend in mini-hole occurrences using Total Ozone Mapping Spectrometer (TOMS) satellite data for the period from 1979 to 1993. However, an analysis of satellite data by Orsolini and Limpasuvan (2001) found an increase in the frequency of ozone mini-holes in the late 1980s and early 1990s. The increase may be linked to the positive phase of the North Atlantic Oscillation (NAO; see section 2.2.2.1), which displaces the westerly jet to higher latitudes, allowing pronounced northward intrusions of high-pressure systems (processes 2 and 3). A similar link between the NAO and the frequency of ozone mini-holes has been found in ground-based measurements (Appenzeller et al., 2000).

Coupled chemistry-climate models are currently able to simulate these meteorological phenomena (Eyring et al., 2003; Stenke and Grewe, 2003). Stenke and Grewe (2003) compared simulations from a coupled chemistry-climate model with TOMS data and showed that ozone minima were fairly well represented in the simulations. Such simulations suggest that the processes affecting PSC formation can significantly increase chemical ozone depletion, leading to mini-hole occurrences or other substantial ozone minima.

5.2.3. Large-scale dynamics and temperature

The Arctic is highly affected by atmospheric processes, and mid- and high-latitude dynamics can play an important role in arctic ozone depletion. The Northern Hemisphere is characterized by large landmasses and several high mountain ranges at middle and high latitudes. These geographic features generate planetary-scale atmospheric waves that disturb the northern polar vortex. As a result, the polar vortex tends to be less stable and less persistent over the Arctic than over the Antarctic. Ozone depletion over the Arctic has therefore been less severe than that over the Antarctic, but is still greater than the depletion observed at tropical or mid-latitudes. Ozone depletion in the Arctic is characterized by large interannual variability, depending largely on the strength of the polar vortex and on air temperatures within it. During years when the polar vortex was especially strong, substantial (up to 40%) total column ozone depletion was observed (Weatherhead, 1998; WMO, 2003).

Changes in the dynamics of the stratosphere play a role in long-term trends as well as in inter- and intra-annual variability in arctic ozone levels. The stratospheric circulation determines how much ozone is transported from the lower-latitude production regions, as well as the extent, strength, and temperature of the winter polar vortex. The variability of polar vortex conditions is strongly influenced by fluctuations in the strength of the planetary-wave forcing of the stratosphere. There is evidence from both observations and modeling studies that long-term trends in arctic ozone levels are not solely driven by trends in halogen concentrations, but are also a function of changes in wave-driven dynamics in the stratosphere (Fusco and Salby, 1999; Hartmann et al., 2000; Hood et al., 1999; Kodera and Koide, 1997; Kuroda and Kodera, 1999; Pitari et al., 2002; Randel et al., 2002; Shindell et al., 1998a; Waugh et al., 1999). During years in which planetary waves penetrate effectively to the stratosphere, the waves enhance the meridional Brewer-Dobson circulation, which brings more ozone from the low-latitude middle and upper stratosphere to the polar region and then down to the arctic lower stratosphere. At the same time, the planetary waves are likely to disrupt the polar vortex, reducing the occurrence of temperatures low enough for PSC formation. Increased planetary-wave activity is thus highly correlated with greater ozone levels, but projections of future wave forcing remain uncertain (WMO, 2003).

Extremely low stratospheric temperatures (below 190 K) in the polar regions can lead to the formation of PSCs (Box 5.1). Polar stratospheric clouds contribute significantly to ozone chemistry, leading to accelerated ozone destruction. Over the Antarctic, stratospheric temperatures are routinely lower than these thresholds every spring. Over the Arctic, stratospheric temperatures are often near these critical temperature thresholds, such that during periods when the temperatures are slightly lower than average, accelerated ozone depletion is observed, while during periods when the temperatures are slightly higher than average, ozone levels can appear climatologically normal. Current climate models suggest that stratospheric temperatures are likely to decrease in the coming decades as a result of increasing atmospheric concentrations of greenhouse gases, thus, it is likely that there will be more periods when accelerated ozone destruction could occur. The combination of dynamics, interannual variability, and the coupling between chemistry and radiative forcing makes projecting future arctic stratospheric temperatures and ozone depletion extremely challenging.

5.3. Long-term change and variability in ozone levels

In the early 1970s, scientists began projecting that anthropogenic emissions of CFCs and other halocarbons would lead to stratospheric ozone depletion (Molina and Rowland, 1974). These projections were confirmed when the Antarctic ozone hole was discovered in 1985 (Farman et al., 1985), and subsequent work (e.g., Anderson et al., 1989) identified and refined the chemical mechanisms that are responsible for ozone depletion. Since that time, decreases in stratospheric and total column ozone have been reported over both poles and in the mid-latitudes in both hemispheres.

5.3.1. Monitoring stratospheric ozone over the Arctic

Ground-based and satellite-borne instruments are used to monitor the concentrations and vertical distributions of stratospheric ozone. Ground-based Dobson spectroradiometers have been used since the 1920s, and Brewer spectroradiometers have been introduced more recently to provide both ozone and UV radiation monitoring. Currently, more than 30 Dobson and Brewer instruments are operated in or near the Arctic. In Russia, total ozone is monitored using filter radiometers. In addition to these measurements of total column ozone, the vertical ozone distribution or ozone profile can be measured using ozonesondes (balloon-borne measuring devices). Figure 5.3 shows the current network of regularly reporting total ozone and ozonesonde stations in or near the Arctic.
The ground-based monitoring network provides the longest and most accurate record of stratospheric ozone levels. In addition to ground-based monitoring, various satellite-borne instruments have been in orbit since the 1970s and are able to provide global spatial coverage not available from ground-based networks. Because the ground-based monitoring network does not cover all parts of the Arctic, monitoring arctic ozone levels relies on a combination of ground-based and satellite-borne instruments.

5.3.2. Total column ozone on a global scale

Total column ozone is a measure of the total number of ozone molecules in a column of atmosphere above a particular location. Total column ozone is important because of its direct, measurable effect on the amount of UV radiation reaching the surface. The variability in total column ozone at a single location is strongly influenced by the movement of air from one region to another. Thus, total column ozone averages over the entire globe, or over large regions, often show less variability than total column ozone at a specific location (Bodeker et al., 2001). Although ozone measurements have been made by satellite-borne instruments since the late 1970s, orbits and instrument capabilities have not always ensured year-round monitoring of conditions in the arctic stratosphere.

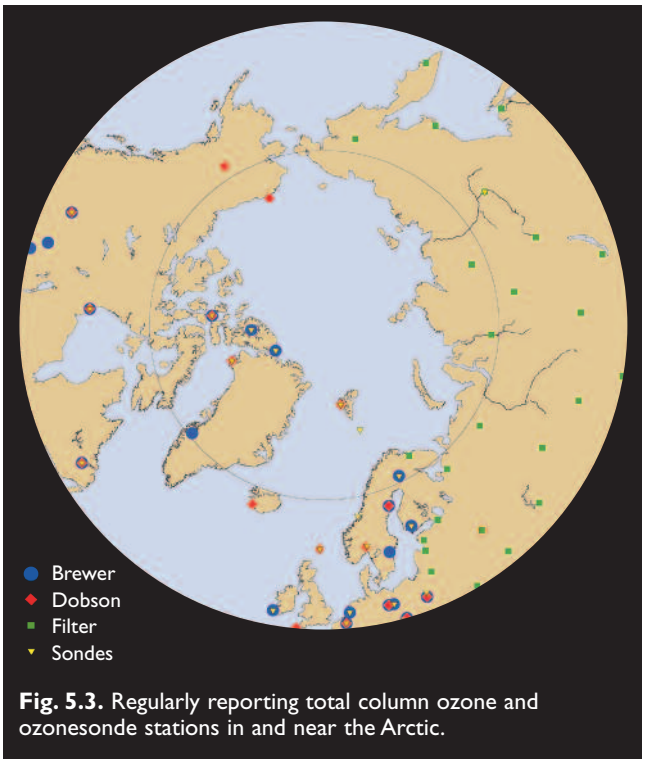

Fig. 5.3. Regularly reporting total column ozone and ozonesonde stations in and near the Arctic.

Instrument drift, problems with calibration, and other issues influencing data continuity can all affect estimates of ozone levels derived from satellite data. Careful comparison with well-calibrated ground-based instruments has helped resolve many of these difficulties, and the satellite data have been used in many analyses of ozone depletion (e.g., Herman and Larko, 1994; McPeters et al., 1996; Newman et al., 1997; Reinsel et al., 1994; Staehelin et al., 2002; Weatherhead et al., 2000). The results indicate strong downward trends in stratospheric ozone amounts, particularly during the late winter and spring. The data show strong latitudinal variability as well as observable longitudinal variations.

Several datasets of zonal total column ozone values were compared and used to estimate long-term changes in total column ozone. The datasets were prepared by different groups and are based on TOMS, Solar Backscatter Ultraviolet (SBUV, SBUV/2), Global Ozone Monitoring Experiment, and ground-based measurements (Fioletov et al., 2002). To avoid problems of missing data at high latitudes, and to estimate global total ozone, it was assumed that deviations from the long-term mean over regions with no data (such as over the poles) were the same as the deviations in the surrounding latitude belts. The results suggest that global average total column ozone in the late 1990s was 3% lower than in the late 1970s.

5.3.3. Total column ozone trends

The decline in total column ozone is a function of the solar cycle, atmospheric dynamics, chemistry, and temperatures. In general, the agreement between the long-term trends in total column ozone obtained from satellite and ground-based data is very good: both indicate a latitudinal variation in the trends, with values close to zero over the equator and substantial declines outside the 35° S to 35° N zone.

Satellite data indicate that variations in the total column ozone trends are predominantly latitudinal, with some smaller longitudinal differences. The greatest decrease in total column ozone over the Northern Hemisphere high latitudes (7% per decade) occurred in the spring (March–May) over the subpolar regions of Siberia, northern Europe, and the Canadian Arctic. These longitudinal differences correspond at least partially to large relative decreases during the winter and

spring, which occur when air masses with relatively low ozone concentrations typical of the polar vortex are transported over regions with high climatological ozone values. In these situations, the decrease in total column ozone is not limited to the polar vortex area alone (WMO, 1999, 2003). Unlike the winter and spring depletion, the summer and autumn decrease in total column ozone over the Northern Hemisphere has been smaller and more uniform with longitude.

5.3.4. Variations in arctic total column ozone

Variations and trends in total column ozone over the Arctic are similar to those over mid-latitudes. However, a strong polar vortex in late winter and early spring leads to an additional decrease in total column ozone. Extremely large decreases in total column ozone over the Arctic were observed in certain years, for example, in 1993 and 1997 (Fig. 5.4), which have been partly attributed to a strong polar vortex during those years. Because of the large interannual variability in the strength of the vortex, ozone decreases in the late 1990s and early 2000s were not as large. Decreases in total column ozone associated with the polar vortex can be as large as 45% over vast areas and can last longer than two weeks (Weatherhead, 1998). These traits make vortex-related decreases different from local anomalies or mini-holes, which are caused by advections of tropical and polar air into the mid-latitudes. Mini-holes can be as deep as 35 to 40%, but last only a few days (Weatherhead, 1998).

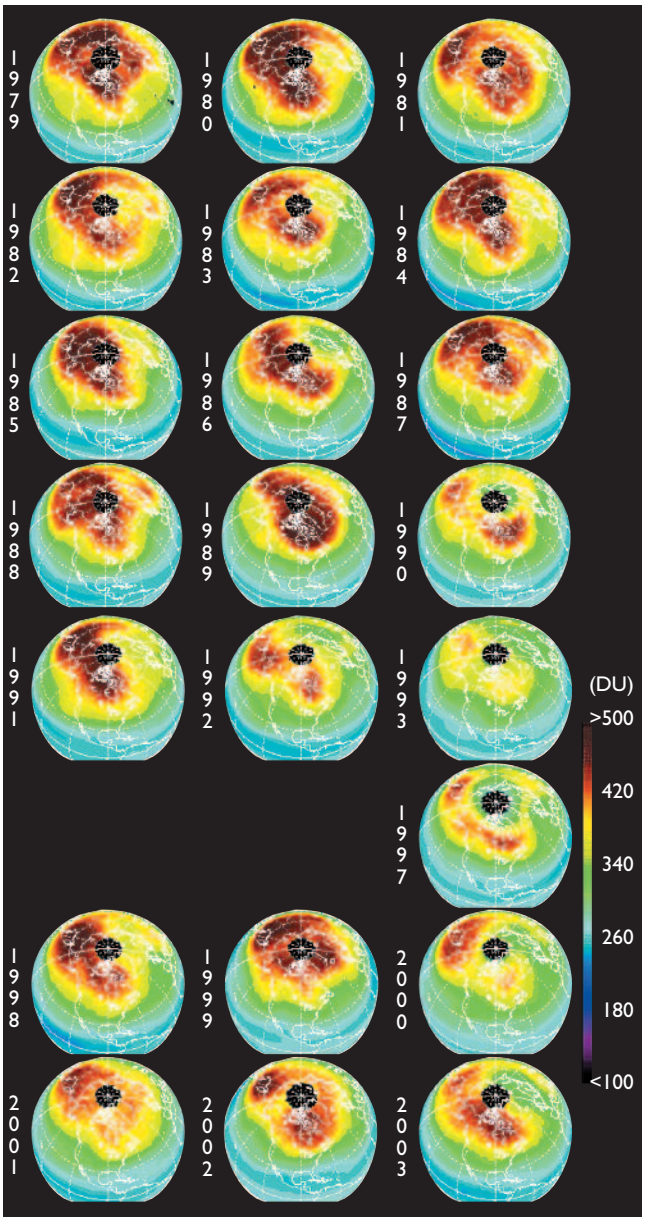

Fig. 5.4. March total column ozone (monthly mean) from 1979 to 2003, from the merged TOMS+SBUV dataset (NASA Goddard Space Flight Center, 2004).

The trend in mean annual total column ozone over the Arctic was approximately -3% per decade for the period from 1979 to 2000 (a total decrease of about 7%). Trends depend on season; the trend in mean spring total column ozone was approximately -5% per decade for the period from 1979 to 2000 (a total decrease of 11%). Large mean monthly decreases in total column ozone (30–35% below pre-depletion levels) were reported in March 1996 and 1997. Some of the daily total column ozone values during these months were below 270 DU, or 40 to 45% below pre-depletion levels.

5.3.5. Ozone profiles

The vertical distribution of ozone within the column plays a lesser role than the total column ozone in determining surface UV radiation levels. At the present time, approximately 20 stations measure vertical ozone profiles during the winter and spring. Measurements of the vertical profile of ozone concentration using ozonesondes have been made weekly since 1980 at several sites in Canada (Edmonton, Goose Bay, Churchill, and Resolute), since 1987 at Alert, and since 1992 at Eureka. Ozone soundings are also performed regularly at Sodankylä, Finland; Ny Ålesund, Norway; Scoresbysund and Thule, Greenland; and Yakutsk, Russia; and occasionally at Bear Island, Norway. In 1988, Europe, Canada, and Russia coordinated an ozonesonde network to measure ozone amounts within the polar vortex. The network consists of 19 stations and has provided assessments of chemical ozone loss for almost every winter since 1988–1989 (Rex et al., 2002). Preliminary analysis of the profiles suggests that trends in ozone concentrations as a function of altitude are most significant in the lower and middle stratosphere, at pressure altitudes of approximately 100 to 25 hPa.

5.4. Factors affecting surface ultraviolet radiation levels in the Arctic

The factors that affect UV radiation levels in the Arctic are generally well established (WMO, 2003), and are illustrated in Fig. 5.5. Atmospheric ozone levels, solar zenith angle, clouds, aerosols, and altitude are all major factors affecting UV radiation levels reaching the surface of the earth. In the Arctic, snow and ice cover add

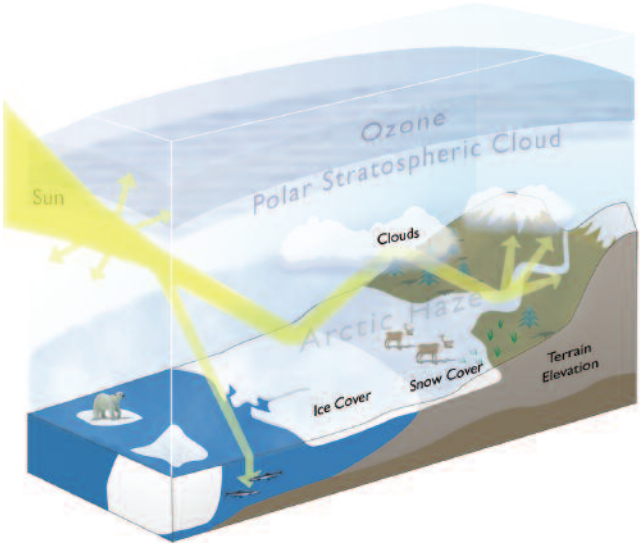

Fig. 5.5. Factors affecting UV radiation in the Arctic.

further complexity to the estimation of UV radiation exposure. When UV radiation passes through the atmosphere it is partially absorbed by ozone, and scattered by air molecules, aerosol particles, and clouds. Attenuation of UV-B radiation as it passes through the ozone layer is primarily a consequence of the sharp increase in the ozone absorption cross section at shorter wavelengths. The ratio of diffuse to global (direct and diffuse) radiation is greater in the UV than in the visible spectrum, primarily due to the wavelength dependence of Rayleigh scattering. Moreover, the ratio is usually higher in the Arctic than at lower latitudes due to large solar zenith angles and frequent snow cover.

Many of the factors affecting UV radiation have large natural variations, which makes it difficult to discern changes in UV radiation levels that result from ozone depletion. Furthermore, the factors are not independent but interact in complex ways. For example, enhancement of surface UV irradiance by multiple scattering depends on both surface albedo and cloud conditions. These features make polar regions, including the Arctic, unique and complex in terms of their UV radiation environments.

Table 5.1 summarizes the factors that affect surface UV radiation levels in the Arctic.

5.4.1. Extraterrestrial solar spectrum

The radiation output of the sun varies over a range of timescales. Over the last century, the largest variation has been the 11-year solar cycle, which can be estimated by the average number of sunspots. The variation in solar irradiance is dependent on wavelength, with greater variability at shorter wavelengths (Solanki and Unruh, 1998). It has been estimated, using models and data from the Upper Atmosphere Research Satellite Solar Stellar Irradiance Comparison Experiment instrument, that although the total solar irradiance varies by only about 0.1% over the 11-year solar cycle, the amplitude of variation is as high as 8.3% for wave-

Table 5.1. Factors affecting surface UV irradiance in the Arctic.

Factor	Correlation with UV doses	Summary remarks
Solar activity	Negative	In the past century, changes in solar activity have caused fluctuations in surface UV irradiance on the order of a few percent.
Solar zenith angle	Negative	Diurnal and seasonal changes in solar zenith angle depend on latitude. In the Arctic, seasonal variations are extreme while diurnal variations are smaller than those at lower latitudes.
Atmospheric ozone	Negative	The amount of ozone in the stratosphere directly affects the amount of UV radiation reaching the troposphere and the surface of the earth.
Cloudiness	Negative/Positive	Thick clouds can attenuate UV radiation reaching the surface of the earth by tens of percent. Multiple reflections between clouds and snow-covered surfaces can lead to increases in surface UV irradiance, also of the order of tens of percent.
Atmospheric aerosols	Negative	Aerosols can attenuate UV radiation reaching the surface of the earth.
Altitude	Positive	Estimated changes in erythemal UV irradiance with altitude vary from 7 to 25% per 1000 m altitude gain.
Surface albedo	Positive	Reflection off snow can increase surface UV doses by more than 50%.
Snow and ice cover	Negative/Positive	Changes in the extent and duration of snow or ice cover can expose organisms currently shielded from UV radiation. Organisms living above the snow or ice cover will receive lower UV doses as melting snow or ice reduces the surface albedo.
Water quality	Not applicable	The amount of UV radiation penetrating through water is affected by UV-absorbing dissolved organic carbon. Organisms in the near-surface layer experience the greatest exposure to UV radiation.
Receptor orientation	Not applicable	The UV radiation doses received by a vertical surface (such as eyes or face) in the Arctic can be substantially higher than those that are received by a horizontal surface.

lengths in the 200 nm range and 0.85% for wavelengths in the 300 nm range (Lean, 2000).

Fligge and Solanki (2000) reconstructed solar spectral irradiance from 1700 to the present using a model of the magnetic features of the surface of the sun. Their results suggest that since the Maunder solar activity minimum in 1700, solar irradiance has increased by approximately 3% at wavelengths shorter than 300 nm. According to Rozema et al. (2002), the increased solar activity since 1700 has led to enhanced atmospheric ozone production and reduced surface UV-B irradiance. Thus, while the 11-year solar cycle has only a small effect on surface UV-B irradiance, longer-term variations in solar activity have the potential to affect future UV radiation levels.

The amount of UV radiation reaching the earth also depends on the distance between the earth and the sun. Due to the eccentricity of the orbit of the earth, this distance varies throughout the year. The earth is closest to the sun on 3 January (perihelion) and farthest away on 4 July (aphelion). The difference between the perihelion and aphelion distances is about 3%, and therefore extraterrestrial irradiance is about 7% higher during the austral (Southern Hemisphere) summer than it is during the boreal (Northern Hemisphere) summer.

5.4.2. Solar zenith angle

The solar zenith angle (SZA) is the angle between zenith and the position of the sun. Its cosine is approximately inversely proportional to the path length that the direct solar beam has to travel through the atmosphere to reach the surface of the earth. At large SZAs, when the sun appears low in the sky, atmospheric gases and aerosols absorb more UV radiation owing to the longer path length that photons must travel. Variations in the SZA cause clear diurnal and annual variations in surface UV radiation levels. The SZA is also responsible for most of the latitudinal variation in surface UV radiation levels. The percentage change between summer and winter UV radiation levels is higher in the Arctic than at lower latitudes, while diurnal variations in the SZA are smaller at higher latitudes. In general, SZAs are large in the Arctic and therefore, arctic UV irradiances are typically lower than those at lower latitudes. However, when daily integrated doses are compared, the length of arctic summer days somewhat compensates for the effect of large SZAs. The annual variation of the clear-sky daily erythemal dose at latitudes of 50°, 70°, and 80° N is shown in Fig. 5.6. The values are based on radiative transfer calculations assuming moderate polar ozone levels (300 DU), snow-free conditions with a surface albedo of 0.03, and clear skies. The seasonal variation in erythemal dose is caused solely by the seasonal variation in the SZA.

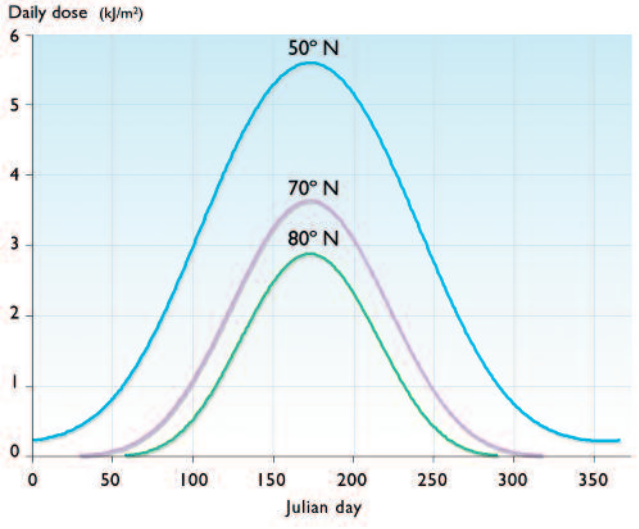

Fig. 5.6. Modeled clear-sky daily erythemal UV radiation dose at latitudes of 50°, 70°, and 80° N.

5.4.3. Ozone levels

Absorption by ozone causes attenuation of UV-B irradiance. It has been repeatedly demonstrated that a decrease in total column ozone leads to an increase in UV radiation levels (WMO, 2003). The relationship depends somewhat on the vertical distribution of ozone in the atmosphere. At small SZAs, a redistribution of ozone from the stratosphere to the troposphere leads to a decrease in UV-B radiation levels at the surface (Brühl and Crutzen, 1989). At very large SZAs, this redistribution leads to an increase in UV-B radiation levels (Krotkov et al., 1998). Lapeta et al. (2000) and Krzyscin (2000) further quantified this effect, and concluded that the erythemally weighted UV dose rate varies by a maximum of 5% owing to changes in the ozone profile.

The change in surface UV irradiance as a result of a change in total column ozone depends highly on the wavelength of the radiation. Traditionally, radiation amplification factors (RAFs) have been used to quantify the change in biologically effective irradiances as a result of a change in total column ozone (e.g., Booth and Madronich, 1994; van der Leun et al., 1989; WMO, 1989). These factors can also be used to indicate the sensitivity of a particular UV radiation effect to a change in total column ozone. Values of RAFs depend largely on the biological effect and vary between 0.1 and approximately 2.5 (Madronich et al., 1998). The RAF for the standard erythemal action spectrum (CIE, 1998) is 1.1 at small SZAs (Madronich et al., 1998), indicating that a 1% decrease in total column ozone leads to a 1.1% increase in erythemal UV radiation. For large changes in total ozone, the relationship is nonlinear, and a more complex relationship is required to estimate the corresponding changes in biologically effective UV radiation (Booth and Madronich, 1994). In the Arctic, where SZAs are often large, RAFs should be used with caution due to their pronounced dependence on the SZA and on total column ozone at large SZAs (Micheletti et al., 2003). For example, at an 80° SZA and total column ozone of

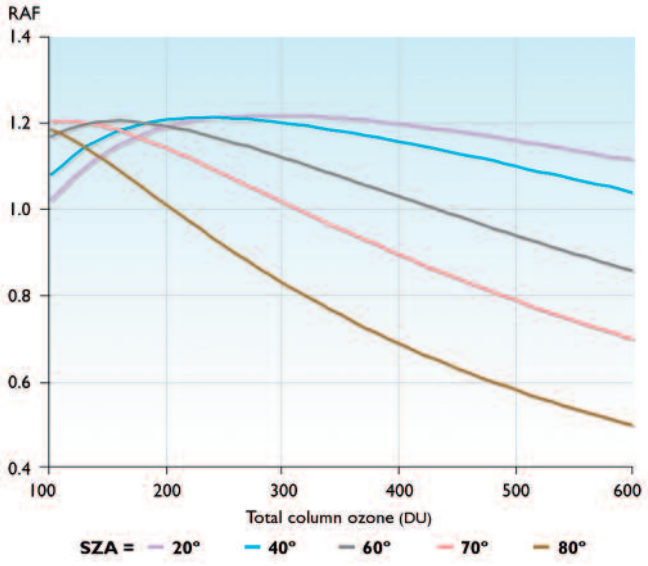

Fig. 5.7. Erythemal radiation amplification factor (RAF) as a function of total column ozone and solar zenith angle (SZA).

300 DU, the erythemal RAF is reduced to approximately 0.8, which is about 27% lower than that for smaller SZAs typical of the mid-latitudes (Fig. 5.7).

5.4.4. Clouds

The effect of clouds on UV radiation is difficult to quantify because of their complex three-dimensional character and rapid temporal variation. A uniform cloud layer generally leads to a decrease in irradiance at the surface of the earth, because part of the radiation that is reflected upward by the cloud layer escapes into space. However, local surface UV irradiance can be increased if clouds are not obstructing the disk of the sun and additional radiation is reflected from the side of a broken cloud field toward the ground (Mims and Frederick, 1994; Nack and Green, 1974). In meteorology, cloud cover is traditionally measured in "octas". The sky is divided into eight sectors and the octa number, between zero and eight, is based on the number of observed sectors containing clouds. Bais et al. (1993) and Blumthaler et al. (1994a) showed that when the solar disk is clear of clouds, cloud amounts up to six octas have little effect on irradiance compared to clear-sky situations. Thiel et al. (1997) and Josefsson and Landelius (2000) have further parameterized the attenuation of UV irradiance as a function of cloud cover and type.

Cloud transmittance of UV radiation depends on wavelength (Frederick and Erlick, 1997; Kylling et al., 1997; Seckmeyer et al., 1996). The maximum transmittance occurs at approximately 315 nm, although the actual location of this maximum depends on the cloud optical depth, the amount of tropospheric ozone, and the SZA (Mayer et al., 1997). In general, clouds in the Arctic tend to be optically thinner than clouds at lower latitudes owing to reduced atmospheric water vapor content. When the ground is covered by snow, attenuation of UV radiation by clouds is further diminished owing to multiple scattering between the ground surface and the cloud base (Nichol et al., 2003).

5.4.5. Aerosols

Aerosols are solid or liquid particles suspended in the atmosphere, found primarily in the lower part of the troposphere. The attenuation of surface UV irradiance by aerosols depends on the aerosol optical depth (AOD), single scattering albedo, asymmetry factor, and aerosol profile. Measurements of AOD are routinely carried out at visible and UV-A wavelengths (e.g., Holben et al., 1998). The AOD is generally assumed to follow Ångström's law, which states that AOD is proportional to $\lambda^{-\alpha}$, where λ is wavelength and α is the Ångström coefficient. Converting the AOD measured at longer wavelengths to an AOD value for the UV-B spectrum is not straightforward, however, because α is not easy to measure and is likely to have some wavelength dependence. The single scattering albedo is the ratio of the scattering cross section of the aerosol to its extinction cross section, and is typically greater than 0.95 in relatively unpolluted areas of the Arctic (d'Almeida et al., 1991).

Episodes of long-range transport of pollutants have been observed in the Arctic. These episodes, combined with the lower rates of particle and gas removal in the cold and stable arctic atmosphere, can lead to a phenomenon called "arctic haze" (Shaw, 1985, 1995). Arctic haze events result in increased aerosol concentrations and mostly occur in winter and spring.

Relatively few studies have addressed the role of aerosols in attenuating solar UV radiation in the Arctic. Wetzel et al. (2003) conducted field investigations at Poker Flat, Alaska, and sampled different air mass types originating from sources outside the region. The measured AOD at 368 nm ranged from 0.05 to 0.25, and

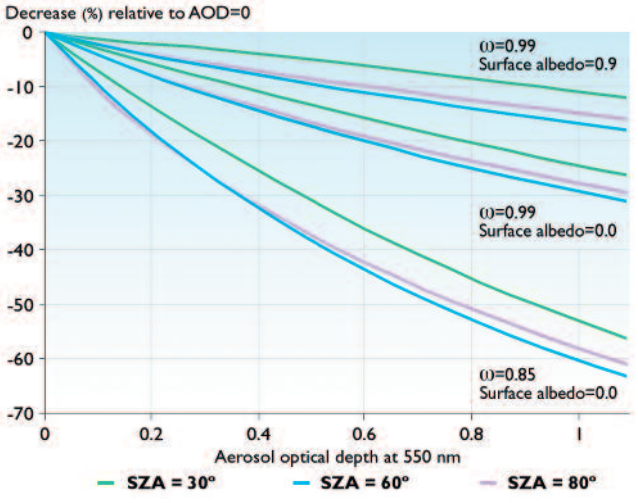

Fig. 5.8. Percentage decrease in erythemal UV irradiance as a function of aerosol optical depth (AOD), based on theoretical calculations with a radiative transfer model (ω=aerosol single scattering albedo).

estimates for the single scattering albedo varied from 0.63 to 0.95, the former being for spring air masses originating from Asia and the latter for cleaner air masses of marine origin. Herber et al. (2002) summarized eight years of measurements of AOD at the Koldeway station in Ny Ålesund, Norway, and reported strong arctic haze events, mainly in late winter and spring. The mean AOD at 371 nm during arctic haze conditions was about 0.18, while in the autumn the average AOD was only 0.05. Quinn et al. (2002) presented results from three years of simultaneous measurements of aerosol chemical composition and light scattering and absorption at Barrow, Alaska. They found that sulfate concentrations were highest at Barrow and decreased with latitude from Poker Flat to Denali to Homer, suggesting a north–south gradient. Ricard et al. (2002) studied the chemical properties of aerosols in northern Finland, and found that, compared to other arctic sites, the aerosols reflect smaller contributions from arctic haze and marine events in winter and larger contributions from biogenic sources in summer. For the range of aerosols sampled at Poker Flat, Alaska, Wetzel et al. (2003) found that the attenuation of UV radiation at 305 nm and 368 nm ranged from a few percent up to about 11%.

Figure 5.8 illustrates the decrease in erythemal UV irradiance as a function of AOD, based on theoretical calculations with a radiative transfer model (Mayer et al., 1997). The figure indicates that in the Arctic, where SZAs tend to be high, the reduction of erythemal UV irradiance by aerosols depends strongly on aerosol properties, including the single scattering albedo, and on surface properties, including surface albedo. In practice, AOD and single scattering albedo cannot be directly translated into UV attenuation, as the asymmetry factor, vertical distribution, and other factors must also be taken into account.

5.4.6. Altitude

Ultraviolet radiation levels increase with altitude for several reasons. At higher elevations, the atmosphere is optically thinner, and therefore fewer particles exist to absorb or scatter radiation. Higher elevations also experience a reduced influence from tropospheric ozone or aerosols in the boundary layer. In the Arctic and in mountainous regions, the ground is more likely to be covered by snow at higher altitudes, which leads to higher albedo and increased UV reflectance. Clouds below a mountain summit have a reflective effect similar to snow-covered ground, and will therefore increase UV radiation levels at the summit. In contrast, the same cloud may reduce UV radiation levels in a valley below the mountain. The variation of UV radiation levels with altitude depends on several factors, all of which have different wavelength dependencies; therefore, this variation cannot be expressed by a simple relationship. Changes in erythemal UV irradiance with altitude reported in the literature vary between 7 and 25% per 1000 m of altitude gain (Blumthaler et al., 1994b, 1997; Gröbner et al., 2000; McKenzie et al., 2001a).

5.4.7. Surface albedo

The extent and duration of snow cover in the Arctic has a significant effect on surface UV radiation doses. An increase in surface albedo leads to an increase in downwelling UV radiation, as part of the radiation that is reflected upward is backscattered by air molecules or clouds. Snow is particularly efficient at reflecting UV radiation; multiple reflections between snow-covered ground and clouds, therefore, can lead to a significant increase in surface UV radiation levels compared to a snow-free situation (Kylling et al., 2000a).

Surface albedo at UV wavelengths is generally low, except in the presence of snow cover. Blumthaler and Ambach (1988) measured erythemally weighted surface albedos for various snow-free surfaces and reported values ranging between 0.01 and 0.11. Spectral measurements by Feister and Grewe (1995) and McKenzie and Kotkamp (1996) confirm these low values. For snow-covered surfaces, the measurements suggest values ranging from 0.50 to 0.98. In general, dry new snow has the highest albedo, which ranges from 0.90 to 0.98 (Grenfell et al., 1994). The albedo of a snow-covered surface depends not only on snow depth and condition, but also on topography, vegetation, and man-made structures (Fioletov et al., 2003). Albedo is an important factor affecting UV radiation levels in the Arctic, where the ground is covered by snow for extensive periods of the year. Figure 5.9 shows the spectral amplification of surface UV irradiance by surface albedo for clear-sky conditions. The figure indicates that snow cover, with an albedo that can be greater than 0.8, can increase erythemal irradiance by up to 60% compared to a snow-free case (albedo 0.2 or less). The amplification is greatest at short UV wavelengths, and thus increases the ratio of UV-B to UV-A radiation.

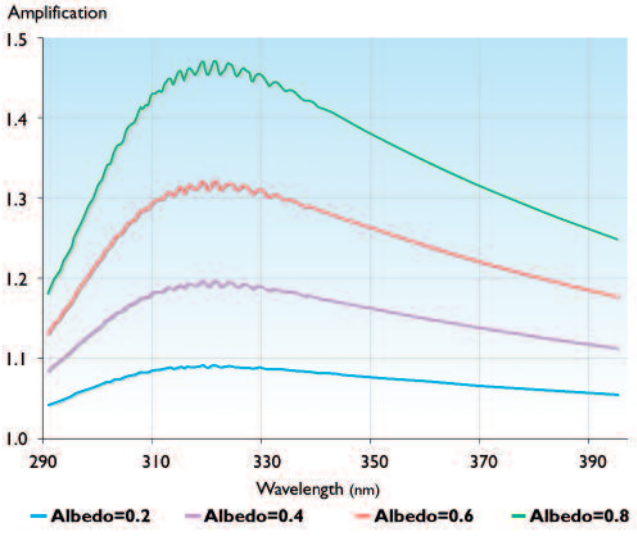

Fig. 5.9. Spectral amplification of surface UV radiation by surface albedo under clear-sky conditions (adapted from Lenoble, 1998).

Scattering in the atmosphere may occur far away from the location of interest; therefore, the ground properties of a large area around the measurement site must be considered. The regionally averaged albedo is often referred to as the "effective albedo" (Gröbner et al., 2000; Kylling et al., 2000b), and can be considered the albedo estimate that gives the best agreement between measured and modeled irradiances when used in a radiative transfer model. Three-dimensional radiative transfer models have shown that the area of significance, defined by an increase in UV irradiance of more than 5% when effective albedo is taken into account, can extend more than 40 km around the point of interest (Degünther et al., 1998; Lenoble, 2000; Ricchiazzi and Gautier, 1998).

5.4.8. Snow and ice cover

Many arctic ecosystems are shielded from UV radiation for much of the year by snow or ice cover. The transmission of UV radiation through snow or ice depends on wavelength, the thickness of the cover, and the optical properties of the snow or ice. In general, radiation is attenuated by a factor that changes exponentially with the thickness of the snow or ice cover. Shorter wavelengths are more strongly attenuated than longer wavelengths. Field measurements conducted at Alert, Canada, suggest that a 10 cm snow-cover depth reduces the amount of transmitted 321 nm UV radiation by two orders of magnitude (King and Simpson, 2001). According to Perovich (1993), approximately 1.3 m of white ice would be required to achieve a similar attenuation of transmitted 300 nm UV radiation. It is difficult to project how changes in snow and ice cover will affect the amount of UV radiation to which terrestrial and aquatic life forms in the Arctic are exposed. Observed and projected changes in sea-ice and snow cover are discussed in sections 6.3 and 6.4, respectively.

5.4.9. Water quality

The water quality parameters that are known to affect underwater UV radiation levels are dissolved organic carbon and chlorophyll a (Kuhn et al., 1999; Laurion et al., 1997; Morris et al., 1995; Scully and Lean, 1994). A general optical characterization of water columns is obtained from diffuse attenuation coefficients ($K_d(\lambda)$), which are calculated from measurements of spectral irradiance at various depths. For comparative purposes, wavelength-specific 10% depths (the depth to which 10% of the below-surface irradiance penetrates) are often derived from the $K_d(\lambda)$ values. It is important to note that the choice of 10% depth is arbitrary and is not based upon any correlation with biological effects. Figure 5.10 shows 10% penetration depths in Lake Cromwell, the St. Lawrence River estuary, and the Gulf of St. Lawrence, Canada. In clear ocean water, 10% of the radiation at longer UV wavelengths can penetrate to a depth of nearly 100 m. In shallower water, this depth may be of the order of only a meter, but organisms living within this 1 m layer would still be at risk. At 310 nm, 10% penetration depths are 20 m for clear ocean water (Smith and Baker, 1979); 1 to 4 m for the Gulf of St. Lawrence and for coastal zones (Booth and Morrow, 1997; Kuhn et al., 1999); 0.5 m for estuarine waters; and 0.1 m for Lake Cromwell, Québec (Scully and Lean, 1994). Measurements made in arctic waters suggest that 10% penetration depths are typically less than 5 m (Aas et al., 2001). Ultraviolet-A radiation generally reaches greater depths. Organisms residing in the near-surface layer experience the greatest exposure to UV radiation.

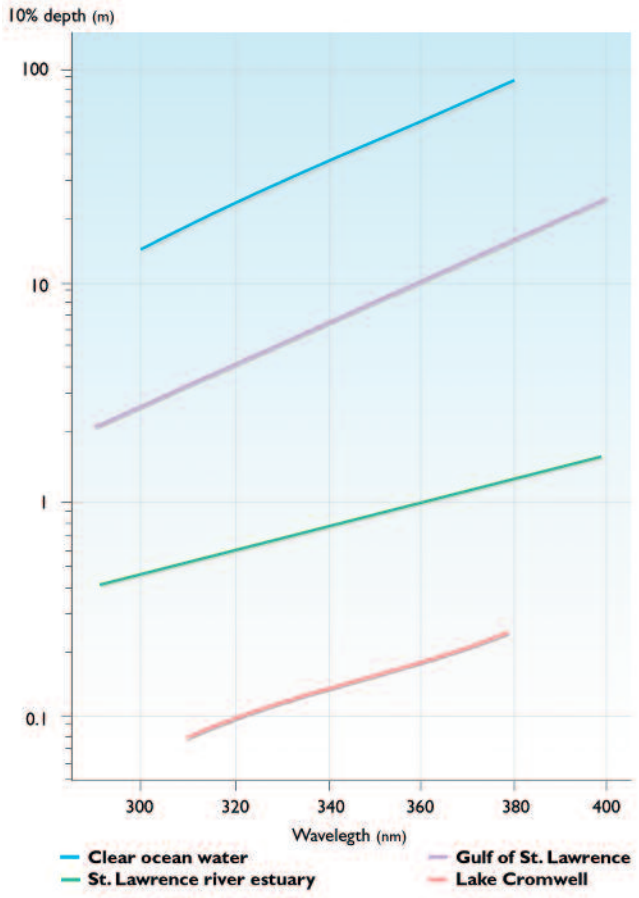

Fig. 5.10. Ten percent depth penetrations at selected locations in eastern Canada (data from Booth and Morrow, 1997; Kuhn et al., 1999; Scully and Lean, 1994; and Smith and Baker, 1979).

5.4.10. Receptor orientation

Ultraviolet irradiance has traditionally been measured on a flat, horizontal surface. While this approach has sound physical merit, it does not accurately represent the UV irradiance that reaches many biological receptors. The amount of UV radiation incident on a vertical (as opposed to horizontal) surface has important biological implications, particularly in terms of effects on the eye (Meyer-Rochow, 2000; Sliney, 1986, 1987). Recent studies have explored both the effect of high snow reflectivity (e.g., McKenzie et al., 1998; Schmucki et al., 2001) and the orientation of the receptor on UV radiation doses. Some investigators

have measured the amount of UV radiation incident on a surface oriented perpendicular to the rays of the sun (e.g., Philipona et al., 2001) while others have measured the amount incident on a vertical surface (Jokela et al., 1993; Weatherhead, 1998; Webb et al., 1999). As reported by Webb et al. (1999), the relationship between irradiance measured on a horizontal surface and that measured on a vertical surface depends on the orientation of the vertical surface, the SZA, and the wavelength. For wavelengths shorter than 400 nm, Webb et al. (1999) found distinct maxima in the vertical to horizontal irradiance ratios during the morning and afternoon. Under cloudless, snow-free conditions, the maximum ratios ranged from 1.4 at 300 nm to 7 at 500 nm. Snow cover, which increases surface albedo, may substantially increase these ratios. For example, in the presence of fresh snow cover and at SZAs greater than 60°, Philipona et al. (2001) reported a 65% increase in erythemal UV irradiance on a surface oriented perpendicular to the sun compared to the irradiance observed on a horizontal surface under the same conditions.

Similar results were obtained by Jokela et al. (1993), who pointed UV radiometers azimuthally South, North, West, and East to assess UV dose rates on vertical surfaces in Saariselkä, Finland. The results indicated a snow albedo of 0.83, in good agreement with data presented by Blumthaler and Ambach (1988). The ratios of vertical to horizontal dose rates varied from about 0.25 to 1.4, depending on direction and on whether the ground was barren or covered with fresh snow. In general, the observations indicated that spring ozone depletion could greatly increase ocular UV radiation doses because of the significant effect of snow reflection. The measurements by Jokela et al. (1993) show that ocular doses of UV radiation in Saariselkä can be higher at the end of April than at any other time of the year. These high doses suggest that the amount of UV radiation incident on the eye when looking toward the horizon can be equivalent or greater than the amount of UV radiation incident on the eye when looking directly upward.

For many biological systems, the actinic flux (the radiation incident at a point) is a more relevant quantity than the horizontal irradiance. Only recently have measurements of the spectral actinic flux become more common (Hofzumahaus et al., 1999; Webb et al., 2002). Webb et al. (2002) found that the ratio of actinic flux to horizontal irradiance varied between 1.4 and 2.6 for UV wavelengths, and depended on wavelength, SZA, and the optical properties of the atmosphere.

5.5. Long-term change and variability in surface UV irradiance

Several instruments and methods have been used to determine surface UV irradiance in the Arctic. Spectral and broadband radiometers, as well as narrowband multi-filter instruments, are used to measure surface UV irradiance. Quality-controlled measurements of UV irradiance have been available for little more than a decade in the Arctic, with limited spatial coverage. In addition to the direct ground-based UV irradiance measurements, surface UV irradiance can be reconstructed using satellite data, or by using observed total column ozone combined with commonly available meteorological data. In addition, historic UV-B radiation levels can be reconstructed using biological proxies.

5.5.1. Ground-based measurements

Surface UV irradiance measurements have been made in the Arctic for many years (Hisdal, 1986; Stamnes et al., 1988; Wester, 1997), and a discernible improvement in the quality of these measurements occurred during the 1990s. Lantz et al. (1999) and Bais et al. (2001) reported that similarly calibrated spectroradiometers typically differ by less than 5% in the UV-A spectrum, and by 5 to 10% in the UV-B spectrum. However, reliable data have only been available since the 1990s, which is inadequate for long-term trend analyses (Weatherhead et al., 1998). Nevertheless, the measured surface UV irradiance time series illustrate the variability of UV radiation and the role of different processes that affect UV irradiance at each of the monitoring sites. The ground-based UV irradiance records are also crucial for validating the indirect methods of estimating UV irradiance.

The current network of ground-based UV irradiance measurement stations in the Arctic is shown in Fig. 5.11. Only those installations operated on a regular basis are shown. The measuring instruments fall

Fig. 5.11. Stations monitoring surface UV irradiance in and near the Arctic.

into three categories: spectroradiometers, multi-filter instruments, and broadband instruments. Spectroradiometers scan radiation in narrow wavelength bands with a typical resolution of 1 nm or less (Seckmeyer et al., 2001). They provide the greatest measurement accuracy and the spectral information enables versatile data use. However, spectroradiometers are costly and require trained personnel for maintenance and operation. Multi-channel or multi-filter instruments typically consist of several filtered photodetectors that measure radiation in selected wavelength bands (Bigelow et al., 1998; Dahlback, 1996; Harrison et al., 1994). They provide much faster sampling than conventional scanning spectroradiometers, allowing the evaluation of both short- and long-term changes in UV irradiance. Methods have also been developed to reconstruct the high-resolution spectrum from these multi-filter instrument measurements (Dahlback, 1996; Fuenzalida, 1998; Min and Harrison, 1998). Broadband instruments collect radiation over a portion of the spectrum and apply a weighting function to the measurements. Many of these instruments are designed to measure the erythemal irradiance. Broadband instruments are comparatively inexpensive and easy to maintain. However, their spectral response often deviates from the one that they are attempting to estimate. Thus, the calibration of broadband instruments depends on solar zenith angle and total column ozone: determining the calibration (Bodhaine et al., 1998; Mayer and Seckmeyer, 1996) and maintaining the long-term stability of the broadband instruments can be very demanding (Borkowski, 2000; Weatherhead et al., 1997). In addition to the instruments described previously, the UV radiation dose (the biological dose rate integrated over time), can be measured using biological UV dosimeters, which directly quantify the biologically effective solar irradiance affecting certain processes by allowing biological systems to act as UV radiation sensors. A number of different sensors have been developed, based on triggers that include direct DNA damage, the inactivation of bacterial spores and bacteriophages, photochemical reactions involving vitamin D photosynthesis, and the accumulation of polycrystalline uracil. Agreement with weighted spectroradiometer measurements has been shown for some of these systems (Bérces et al., 1999; Furusawa et al., 1998; Munakata et al., 2000).

The Canadian UV-monitoring network includes four arctic measurement sites: Churchill, Resolute, Eureka, and Alert. Monitoring at these sites began in 1992, 1991, 1997, and 1995, respectively. There are reports of increased UV-B radiation levels in the Arctic during six winters in the 1990s (Fioletov and Evans, 1997; Kerr and McElroy, 1993), but according to Tarasick et al. (2003), the maximum UV indices (section 5.1) measured at the Canadian Arctic monitoring sites were 8 in Churchill, 4 in Resolute and Eureka, and 3 in Alert. These values are relatively small compared to the UV indices measured at Canadian monitoring sites outside of the Arctic. However, the maximum observed UV index may not be the best indicator of UV radiation doses received in the Arctic because it does not take into account day length or receptor orientation. Moreover, the Canadian UV measurements imply that snow is a major factor affecting UV irradiance at high latitudes, and may enhance the dose received by a horizontal surface by 40%.

Measurements of spectral UV irradiance at Sodankylä, Finland began in 1990. Lakkala et al. (2003) analyzed the Sodankylä data for the period from 1990 to 2001, incorporating corrections for temperature, cosine error, noise spikes, and wavelength shifts. No statistically significant changes in any month were found over the 12-year period, which may be a result of the relatively short period of analysis coupled with the high natural interannual variability of UV irradiance. The lack of a distinct trend over this particular period may also be linked to ozone levels, which decreased in the early 1990s, reached a minimum in the middle of the analysis period, and then increased in the very late 1990s and early 2000s. Arola et al. (2003) applied methods to separate the effects of the different factors that affect short- and long-term changes in UV irradiance to the Sodankylä data and found that in some cases, ozone levels accounted for nearly 100% of the short-term variability in monthly mean irradiance, although on average they accounted for about 35% of the variability. The effects of clouds were smaller, accounting for a maximum of 40% and an average of 12% of the short-term irradiance variability. Albedo-related effects were strongest at Sodankylä during the month of May, accounting for a maximum of 21% and an average of 7% of the short-term irradiance variability.

Measurements are also available from Tromsø, Norway and were recently used by Kylling et al. (2000a) to analyze the effects of albedo and clouds on UV irradiance. The results, shown in Fig. 5.12, indicate that snow

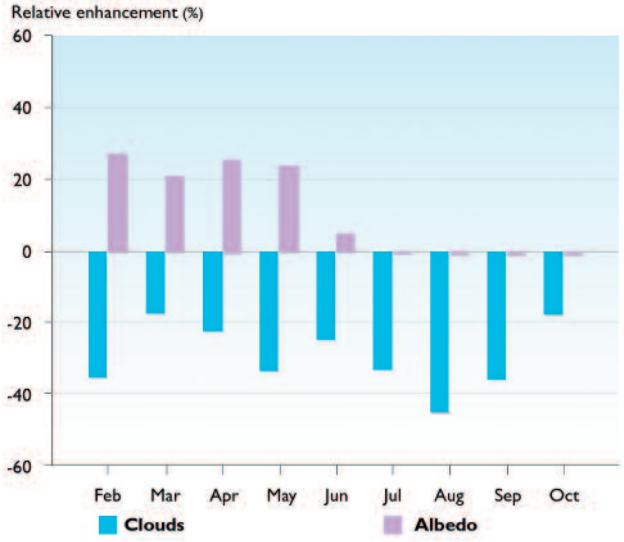

Fig. 5.12. Relative change in monthly erythemal UV radiation doses owing to the presence of clouds and albedo from snow cover (adapted from Kylling et al., 2000a).

increases monthly erythemal UV radiation doses by more than 20% in spring and early summer. The albedo effect at Tromsø is relatively small because the area is influenced by the Gulf Stream, which prevents the adjacent ocean from freezing over. In addition, the terrain around Tromsø is inhomogeneous, consisting of open fjords surrounded by high mountains. Results from a three-dimensional radiative transfer model (Kylling and Mayer, 2001) suggest that the inhomogeneous surface albedo reduces the effective albedo to only 0.55 to 0.60. On specific days, clouds can reduce daily erythemal UV radiation doses at Tromsø by up to 85%. Clouds reduce average monthly erythemal UV radiation doses by 20 to 40% (Fig. 5.12) and this attenuation is fairly constant throughout the year. Multiple scattering between the ground and cloud base links the effects of surface albedo and cloud cover. Attempts to separate the two factors (e.g., Arola et al., 2003 and Kylling et al., 2000a) are only approximations.

Spectroradiometer measurements at Barrow, Alaska commenced in January 1991 (Booth et al., 2001). Increased levels of UV radiation were observed in 1993, which were probably due to low ozone levels related to the injections of aerosols into the stratosphere from the eruption of Mt. Pinatubo (Gurney, 1998). Photochemically induced ozone depletion events at Barrow have primarily occurred during March and April. The most notable case occurred on 18 April 1997, when the daily erythemal UV radiation dose was 36% higher than the climatological average dose. This increase is still small compared to spikes seen in UV radiation measurements at sites affected by the Antarctic ozone hole. By comparing measured and modeled spectra, Bernhard et al. (2003) estimated that Barrow's effective UV albedo (surface albedo for UV wavelengths) during the winter and early spring is 0.85 ± 0.10. The albedo is greater than that of Tromsø because the ocean adjacent to Barrow is frozen until late in the spring, and the treeless tundra surrounding the measurement site is covered by snow until June. The higher albedo leads to a 40 to 55% enhancement of erythemal UV radiation when skies are clear. No statistically significant changes in UV irradiance were found at Barrow for the period from 1991 to 2001 (Booth et al., 2001), for reasons similar to those for Sodankylä. It should be noted that the observed reductions in ozone levels have been much greater in the European and Russian sectors of the Arctic than over Alaska. UV radiation increases at Barrow are therefore less pronounced compared to those measured at the other arctic sites.

5.5.2. Reconstructed time series

In order to assess variations in UV irradiance over longer time periods, various empirical methods for reconstructing UV irradiance data have been developed. These methods are usually based on total column ozone or other commonly available weather data, including global (direct and diffuse) radiation levels. Bodeker and McKenzie (1996), for example, presented a model based on total column ozone, broadband radiation measurements, and radiative transfer calculations. Similar methods were developed and used by McArthur et al. (1999), Kaurola et al. (2000), and Feister et al. (2002) for estimating UV irradiance at various locations in Canada and Europe. Gantner et al. (2000) estimated clear-sky UV irradiance using an empirically determined statistical relationship between clear-sky UV irradiance and total column ozone. The results obtained with reconstruction methods generally compare well with observations, but there are some sources of uncertainty. For example, accounting for the influence of aerosols on the amount of UV radiation reaching the surface of the earth is difficult, as is accounting for sulfur dioxide from volcanic and anthropogenic sources, although Fioletov et al. (1998) reported that sulfur dioxide had a negligible influence on erythemal UV irradiance at 12 out of 13 Canadian and Japanese sites. The importance of examining the homogeneity of the data cannot be overestimated when using long time series of measurements as input data for estimating past UV radiation levels. Likewise, it is of great importance to properly validate the empirical methods against independent measurement data.

Fioletov et al. (2001) developed a statistical model using global radiation levels, total column ozone, dew point temperature, and snow cover data to reconstruct past UV irradiance at several Canadian sites, including Churchill. The model is based on previous efforts to estimate UV-A irradiance from pyranometer measurements (McArthur et al., 1999). Fioletov et al. (2001) produced UV irradiance estimates for the period from 1965 to 1997, and found a statistically significant increase in the erythemally weighted UV irradiance (~11% per decade) over the period from 1979 to 1997. This increase follows a reported decline in UV radiation levels from 1965 to 1980. The increase in UV irradiance between 1979 and 1997 is more than twice the increase that would be expected to result from the observed decline in total column ozone, because concurrent changes in surface albedo and cloud conditions have enhanced the increase in surface UV irradiance over this period. Fioletov et al. (2001) also examined the frequency of extreme values of UV irradiance at Churchill, and reported that the number of hours when the UV index exceeds five increased from an average of 26 per year during the period 1970 to 1979 to an average of 58 per year during the period 1990 to 1997.

Díaz et al. (2003) estimated narrowband (303.030–307.692 nm) irradiances for Barrow from pyranometer data and satellite-measured total column ozone, and found that spring UV irradiance increased between 1979 and 2000. Lindfors et al. (2003) presented a method for estimating daily erythemal UV radiation doses using total column ozone, sunshine duration, and snow depth as input data. They estimated UV radiation doses at Sodankylä for the period from 1950 to 1999, and found a statistically significant March increase

(3.9% per decade). April also showed an increasing trend. For both March and April, the trend was more pronounced during the latter part of the period (1979–1999), suggesting a connection to stratospheric ozone depletion. For July, a statistically significant decreasing trend of 3.3% per decade was found, attributed to changes in total column ozone and sunshine duration (defined by the World Meteorological Organization as the time during which direct solar radiation exceeds 120 W/m^2). June and August exhibited decreasing trends, although not statistically significant, and trends for May and September were negligible.

In addition to atmospheric observations, biological proxies can be used to estimate past UV irradiance. For example, Leavitt et al. (1997) proposed that the past UV radiation environment of freshwater ecosystems could be reconstructed by studying fossil pigments in lake sediments, because some algae and other aquatic organisms produce photoprotective pigments when exposed to UV radiation. However, the long-term variation of underwater UV irradiance is primarily controlled by the amount of dissolved organic matter, which limits the use of fossil pigment sediments for reconstructing surface UV irradiance without ancillary information about dissolved organic matter (Leavitt et al., 2003).

Archaeological findings offer further evidence that UV radiation has long been a concern in the Arctic. Goggles found in the ancestral remains of indigenous peoples have thin slits, which allow the wearer to see but limit the amount of sunlight reaching the eyes (Hedblom, 1961; Sliney, 2001). The indigenous peoples of the Arctic have historically relied heavily on the environment for their livelihood and well-being, and have therefore been acutely aware of changes in climate-related variables. Fox (2000) documented some of these observations, which include reports of sunburn and increased incidence of snow blindness in populations not normally experiencing these effects. The existence of goggles confirms that arctic peoples have long sought protection from the sun and its glare, but the reports of sunburn and other effects of UV radiation suggest that the recent changes in UV irradiance are unusual when considered in this longer context. This information serves as a useful proxy of UV irradiance over time, and provides firsthand evidence of UV radiation impacts resulting from ozone depletion in the Arctic. The information also helps to corroborate available UV radiation and ozone measurements.

5.5.3. Surface estimates from satellite data

Estimates of surface UV irradiance derived from satellite data use a radiative transfer model, with input parameters determined from the satellite measurements. In addition to total column ozone, satellite-derived information about clouds, aerosols, and surface albedo is needed to accurately model these parameters, and various sources of data have been used for that purpose (Herman et al., 1999; Krotkov et al., 2001; Li et al., 2000; Lubin et al., 1998; Meerkoetter et al., 1997; Verdebout, 2000). The advantages of satellite measurements include global or near-global spatial coverage and long-term data continuity. However, there are limitations: the best spatial resolution of the currently available data from ozone-monitoring satellite instruments is 40×40 km and satellites usually provide only a single overpass per day. Satellite instruments also have difficulty probing the lower atmosphere, where UV-absorbing aerosols or tropospheric ozone can significantly affect surface UV irradiance. Fortunately, neither absorbing aerosols nor tropospheric ozone are major factors affecting UV radiation in the Arctic, with the exception of air masses occasionally transported from lower latitudes (section 5.4.5).

Several studies have assessed the accuracy and limitations of satellite-based UV irradiance estimates by comparing them to ground-based data (Arola et al., 2002; Chubarova et al., 2002; Fioletov et al., 2002; Kalliskota et al., 2000; McKenzie et al., 2001b; Slusser et al., 2002; Wang et al., 2000). For noon irradiance, the root mean square difference between the TOMS-derived estimates and ground-based measurements is of the order of ±25%. Estimating surface UV irradiance in the Arctic is more uncertain than that at lower latitudes: when the ground is snow-covered, the satellite-derived UV irradiance estimates have been systematically lower than the ground-based measurements. The bias originates from underestimates of surface albedo, which is a critical parameter for accurately estimating surface UV irradiance in the Arctic. Some studies have addressed surface albedo variations related to snow and ice cover (Arola et al., 2003; Lubin and Morrow, 2001; Tanskanen et al., 2003), but none of the proposed methods have completely resolved the issue. Snow or ice cover also complicates the determination of cloud optical depth properties (Krotkov et al., 2001), and the radiative transfer models are less accurate at high solar zenith angles. Because of these limitations, the accuracy of current satellite retrieval algorithms is not adequate for monitoring arctic surface UV irradiance, although some information on variations in UV irradiance over time can be inferred. The satellite-derived daily erythemal UV radiation dose in and near the Arctic on 21 March 2000 is shown in Fig. 5.13.

Long-term trends in surface UV irradiance were determined using time series derived from TOMS data. The analyses were performed for clear-sky estimates (i.e., the modulation of surface UV radiation by clouds or aerosols was not considered). The trend analysis methods applied were similar to those used to study long-term changes in total column ozone. The trend models take into account changes in solar activity and oscillations in the atmospheric circulation that affect stratospheric ozone levels. Using TOMS Version 7 ozone and reflectivity data from 1979 to 1992 and the corresponding TOMS surface UV retrieval algorithm, Herman et al. (1996) estimated that zonally averaged

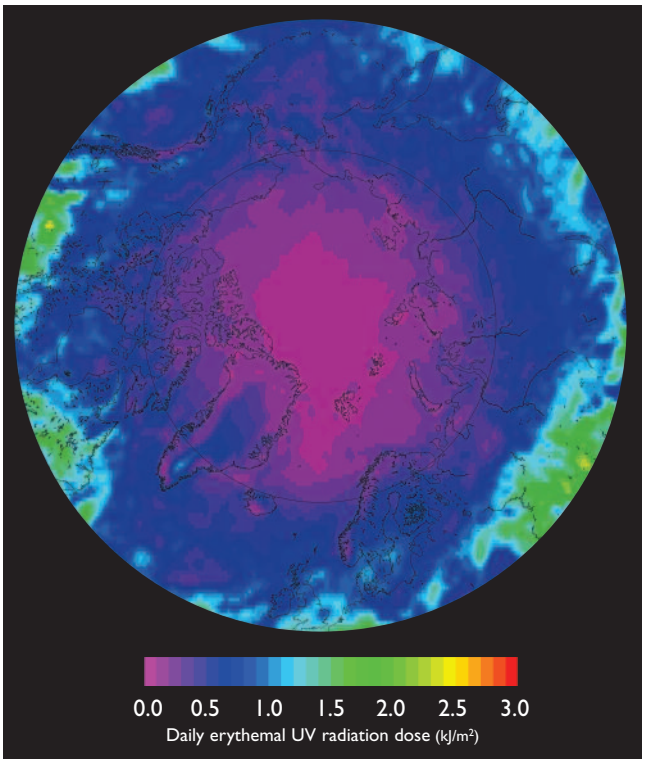

Fig. 5.13. Distribution of the daily erythemal UV radiation dose in and near the Arctic on 21 March 2000 calculated from TOMS total ozone data (Finnish Meteorological Institute, 2004, using data from the National Aeronautics and Space Administration).

mean annual surface UV irradiances at 300, 310, and 320 nm increased by 15%, 6%, and 2% per decade, respectively, for the latitude band 60° to 70° N. The greatest increases were found in winter and spring. Ziemke et al. (2000) studied the spatial distribution of the trends in erythemally weighted surface UV irradiance at northern latitudes using TOMS/Nimbus-7 data and found positive trends exceeding 10% per decade at high latitudes in Eastern Siberia. Unfortunately, the studies based on long-term satellite data did not include the region north of 65° N, and the retrieval algorithm used for the calculations is known to underestimate surface UV irradiance over snow-covered terrain.

5.6. Future changes in ozone

Stratospheric ozone levels over the polar regions are very different from levels over the mid-latitudes. Total column ozone over the Arctic in winter and spring is usually higher than that over the equator and northern mid-latitudes. Ozone levels over the Arctic are marked by a strong annual cycle, with a peak in the spring, and a decrease in late summer and throughout the autumn. Low stratospheric temperatures provide the potential for substantial ozone depletion in the winter and early spring, reducing ozone levels at a time when they would normally be high and when reproduction and new growth leave ecosystems particularly vulnerable. The same physical and chemical processes govern ozone levels and ozone depletion over both the Arctic and Antarctic. However, a stronger, less disturbed polar vortex over Antarctica and thus uniformly lower stratospheric temperatures, have resulted in greater percentage ozone losses over the past two decades in the Antarctic compared to the Arctic. In the years where dynamic conditions allow for similarly cold stratospheric temperatures in the Arctic, significant ozone losses have been observed at northern high latitudes as well.

The Montreal Protocol and its amendments have already resulted in a decrease in the atmospheric concentrations of some ozone-depleting substances (Anderson et al., 2000; Montzka et al., 1999). Although scientific understanding of the dynamics and other factors influencing ozone depletion remains incomplete, most projections suggest that mid-latitude ozone levels will gradually recover over the next 50 years (WMO, 2003). Confirming either a change in ozone trends or an actual increase in ozone levels is likely to require some time because of natural variability and intrinsic measurement errors (Reinsel et al., 2002; Weatherhead et al., 1998, 2000). In polar regions, the projections of recovery are complicated by the effects of dynamic processes and climate change.

Recovery of the ozone layer is likely to occur in stages. The first signs of recovery should be a reduction in the downward trend followed by an increase in ozone levels. Final recovery may be defined as an overall return to pre-depletion ozone levels or as the determination that ozone levels are no longer being affected by anthropogenic ozone-depleting substances. Newchurch et al. (2003) reported evidence of a reduction in the downward trend in ozone levels based on satellite estimates averaged over 60° S to 60° N. Their analysis indicates that since 1997 there has been a slowdown in mid- and low-latitude stratospheric ozone losses at altitudes of 35 to 45 km. These changes in loss rates are consistent with the slowdown in total stratospheric chlorine increases, and, if they continue, will represent the first stage of a mid-latitude ozone recovery. No evidence of this change in loss rates has been reported for polar latitudes, and it is also important to note that ozone at the altitudes where these reduced loss rates are being observed plays a lesser role than total column ozone in absorbing UV radiation.

Several models have been used to project future ozone levels (WMO, 1999, 2003). Intercomparison of these models (WMO, 2003) shows qualitative agreement between the projections, although specific projections of recovery rates can disagree significantly. Two-dimensional models have been used for global ozone level projections, while three-dimensional chemistry–climate models are useful for simulating polar processes (Austin et al., 2003; WMO, 2003). Three-dimensional models can provide multi-year time slice simulations, which have the advantage that several realizations are available for a single year, allowing a better assessment of the projections. Three-dimensional models are also able to address dynamic changes in well-mixed greenhouse

gases and provide a more detailed evolution of ozone levels based on the mechanisms that are likely to occur in the atmosphere. These models can also provide information on the expected range of interannual variability.

Austin et al. (2003) compared several chemistry–climate models used in recent ozone assessments (WMO, 2003). These include the Unified Model with Eulerian Transport and Chemistry (UMETRAC; Austin, 2002), the Canadian Middle Atmosphere Model (de Grandpre et al., 2000), the Middle Atmosphere European Centre Hamburg Model (ECHAM) with chemistry (Manzini et al., 2002; Steil et al., 1998, 2003), the ECHAM model with chemistry run at Deutschen Zentrum für Luft- und Raumfahrt (E39/C; Hein et al., 2001; Schnadt et al., 2002), the University of Illinois at Urbana-Champaign (UIUC) model (Rozanov et al., 2001; Yang et al., 2000), the Center for Climate System Research/National Institute for Environmental Studies (CCSR/NIES) model (Nagashima et al., 2002; Takigawa et al., 1999), the Goddard Institute for Space Studies (GISS) model (Shindell et al., 1998b), and the Università degli Studi dell'Aquila (ULAQ) model (Pitari et al., 2002). The models were compared based on their ability to simulate ozone climatologies for the current atmosphere. For the Northern Hemisphere, all the models tended to overestimate the area-weighted hemispheric total column ozone, by an average of 7.2%. The models were unable to simulate the observed loss rates within the arctic polar vortex (Becker et al., 1998, 2000; Bregman et al., 1997; Hansen et al., 1997; Woyke et al., 1999), so the modeled ozone depletion is less than that observed. Uncertainties in the model projections include temperature biases, leading to the "cold pole problem" (Pawson et al., 2000); these biases are worse in some models than in others. In the Northern Hemisphere, the biases are sometimes positive at certain altitudes, resulting in projections of insufficient ozone depletion in early winter, but excess depletion in spring. The cold pole problem is due largely to the absence of gravity-wave forcing, which many models now include. Other uncertainties include the inability of the models to accurately simulate PSCs and to account for all aspects of constituent (chemical) transport, including processes occurring at the upper boundary of the model. Changes in planetary waves and heat flux also pose uncertainties, and are discussed in greater detail in section 5.6.2.

5.6.1. Considerations for projecting future polar ozone levels

The chemical contributions to ozone depletion are generally understood well enough to describe the annual ozone losses observed over Antarctica as a result of efforts to understand the ozone hole observed there. Over the Arctic, however, ozone depletion processes are often much more complicated and depend greatly on climate conditions and climate change. For example, when potential increases in stratospheric water vapor and corresponding stratospheric cooling resulting from climate change are included in models, the resulting projected mid-latitude ozone decrease in the 2030s surpasses that resulting from the projected amounts of CFC-derived halogens (Shindell and Grewe, 2002). At high latitudes, the effects of stratospheric water vapor and stratospheric cooling on the ozone column are anticipated to be even larger due to the effects of PSCs. Separating the chemical and dynamic/climate-related contributions to ozone depletion is not a simple task, and many questions concerning the future of ozone over the Arctic remain unanswered.

Most model projections suggest small but continuing ozone losses over the Arctic for at least the next two decades (Austin et al., 2000; WMO, 2003). Ozone depletion in the Arctic is strongly influenced by the dynamics of the polar atmosphere: changes in circulation, and particularly changes that affect air temperatures in the polar region, can have a substantial effect. For example, a strong polar vortex results in decreasing stratospheric temperatures, which further strengthen the polar vortex. This positive feedback effect contributes to increased ozone depletion, and is likely to be exacerbated by the stratospheric cooling projected to occur as a result of future climate change.

While dynamics determine the onset of ozone depletion and also influence the rate and severity of the depletion processes, the main driver for upper stratospheric (~40 km) ozone loss and for the spring losses in the polar stratosphere is the chemistry associated with chlorine and bromine (Solomon, 1999; WMO, 1999, 2003). The Montreal Protocol and its amendments have led to a reduction in atmospheric chlorine concentrations, and concentrations of ozone-depleting halogens are expected to continue to decrease between 2000 and 2050. The decreases were first reported in the troposphere (Montzka et al., 1996, 1999), but have also been observed in the upper stratosphere (Anderson et al., 2000). Bromine is another halogen particularly effective at destroying ozone, and its overall levels may increase or remain high because of shorter-lived substances, such as bromoform (Dvortsov et al., 1999). However, the magnitude of ozone loss will depend greatly on dynamic and climate conditions (section 5.2), with low temperatures contributing to the formation and persistence of PSCs. Over the polar regions, heterogeneous chemistry in or on these clouds converts stable chlorine and bromine reservoirs to more active forms that can deplete ozone. Future volcanic eruptions could also change stratospheric ozone levels worldwide for at least several years, and could have a large effect on arctic ozone levels as long as halogen loading remains large (Tabazadeh et al., 2002).

5.6.2. The role of climate change in arctic ozone recovery

Projections of the recovery of ozone levels in the Arctic depend on projections of global climate change.

Understanding long-term changes (natural and anthropogenic), will be essential to improving assessments and projections of the dynamic structure of the stratosphere (E.C., 2003). Current chemical and dynamics models project that climate change resulting from increased atmospheric concentrations of carbon dioxide and other greenhouse gases will warm the troposphere, but will cool the stratosphere. In the Arctic, this cooling is likely to lead to increased ozone destruction, as lower temperatures are likely to result in the formation and persistence of PSCs, which aid in the activation of ozone-depleting compounds and can therefore accelerate ozone depletion. Stratospheric cooling resulting from climate change is therefore likely to lead to an increased probability of larger and longer-lasting ozone holes in the Antarctic and extensive, more severe ozone losses over the Arctic (Dameris et al., 1998). On the other hand, climate change could possibly trigger an increase in planetary waves, enhancing the transport of warm, ozone-rich air to the Arctic (Schnadt et al., 2002). This increased transport would counter the effects of heterogeneous chemistry and possibly hasten recovery of the ozone layer. Understanding this "dynamic effect on chemistry" requires improved information about the effects of increasing greenhouse gas concentrations so that the balance between dynamics and radiation can be deduced. If radiative effects dominate, planetary wave activity would be more likely to decrease, resulting in more ozone depletion at arctic latitudes.

Another climate feedback affecting ozone is a potential increase in stratospheric water vapor due to changes in tropopause temperatures (Evans et al., 1998). Few long-term datasets of water vapor concentrations are available, but previous studies of existing observations have suggested that stratospheric water vapor has been increasing (Oltmans and Hofmann, 1995; Oltmans et al., 2000; Randel et al., 1999). Analyses of 45 years of data (1954–2000) by Rosenlof et al. (2001) found a 1% per year increase in stratospheric water vapor concentrations. Analyses of satellite data, however, have shown less evidence of a water vapor increase (Randel et al., 2004). Increased water vapor is likely to contribute to increased ozone destruction by affecting the radiation balance of the stratosphere (Forster and Shine, 2002; Shindell, 2001). Greater water vapor concentrations in the stratosphere can raise the threshold temperatures for activating heterogeneous chemical reactions on PSCs, and can cause a decrease in the temperature of the polar vortex (Kirk-Davidoff et al., 1999).

Ozone itself is central to climate change science: it is an important greenhouse gas in the infrared part of spectrum and is the primary absorber of solar UV radiation. Ozone is critical to the radiation balance of the atmosphere, and to the dynamics of the stratosphere. Indeed, recent observational findings confirm that "the stratosphere is a major player in determining the memory of the climate system" (Baldwin et al., 2003). Stratospheric ozone levels play a role in determining many properties of the polar atmosphere, including the strength of the polar vortex. Observations show that the strengths of the polar vortices affect surface temperatures in the polar regions and at mid-latitudes in both hemispheres. Connections between ozone levels and other properties of the stratosphere can alter weather processes in the troposphere, with an effect whose magnitude is comparable to that of the El Niño–Southern Oscillation (Gillett and Thompson, 2003; Hartmann et al., 2000).

Projected future changes in ozone levels over the polar regions differ from projected changes over the rest of the globe, where stratospheric temperatures do not reach the low thresholds necessary for the formation of PSCs. In recent years, the arctic polar vortex has increased in strength and has become more persistent (Waugh et al., 1999; Zhou et al., 2000). A strong polar vortex can enhance the amount of depletion experienced over the Arctic. If these strong polar vortex conditions continue in future years, arctic ozone recovery is likely to be substantially delayed (Shindell et al., 1998b). For example, in an analysis of approximately 2000 ozonesonde measurements, Rex et al. (2004) found that each 1 °C cooling of the arctic stratosphere resulted in an additional 15 DU of chemical ozone loss. Their findings indicate that over the past four decades, the potential for the formation of PSCs increased by a factor of three, resulting in stratospheric conditions that have become significantly more favorable for large arctic ozone losses. This relationship between potential amounts of PSCs and ozone loss is not well-represented in current chemistry–climate models. If the arctic stratosphere continues to cool as a result of climate change, the region is likely to continue to experience severe ozone depletion until chlorine and bromine loadings have returned to background levels. Any delay in the recovery of the ozone layer over polar regions means a longer-lasting, and perhaps more severe, threat of ecosystem damage due to increased UV irradiance.

5.6.3. Projected changes in ozone amounts

A number of two-dimensional models using specified scenarios of atmospheric halocarbon concentrations were used to estimate future ozone levels for the most recent Scientific Assessment of Ozone Depletion (WMO, 2003). These included the Atmospheric and Environmental Research (AER) model (Weisenstein et al., 1998), the Max-Planck Institute (MPI) model (Grooß et al., 1998), the Goddard Space Flight Center (GSFC) model (Fleming et al., 1999), GSFC-INT (interactive version of the GSFC model; Rosenfield et al., 1997), the National Oceanic and Atmospheric Administration/National Center for Atmospheric Research (NOCAR) model (Portmann et al., 1999), the University of Oslo (OSLO) model (Stordal et al., 1985), the National Institute for Public Health and the Environment (RIVM) model (Velders, 1995), the State University of New York – St. Petersburg (SUNY-SPB) model (Smyshlyaev et al., 1998), UIUC (Wuebbles et

al., 2001), and ULAQ (Pitari and Rizi, 1993). Figure 5.14 shows the spring (March–May) changes in ozone for the latitude band 60° to 90° N (relative to 1980 levels) projected by these two-dimensional models. The spring is interesting because ozone depletion reaches its most severe levels and UV irradiance can also be relatively high during what is the beginning of the growth period for many biological systems. The model results shown are for the greenhouse gas scenario MA2 and baseline halocarbon scenario AB (WMO, 2003). All the models except RIVM include arctic chemistry, while only the MPI and UIUC models include the 11-year solar cycle. The projected spring changes in arctic ozone levels are about twice as large as those projected for the Northern Hemisphere mid-latitudes and about three times as large as projected changes in the 60° N to 60° S annual average. Generally, the models simulate local minimums in arctic ozone levels in the late 1990s, followed by a gradual increase. The majority of the models project significantly lower ozone levels in 2020 compared to 1980.

Because the two-dimensional models are unable to incorporate dynamic effects, their results are considered very rough projections for the polar regions, where ozone levels are strongly influenced by atmospheric dynamics. The model simulations used in the 2002 assessment (WMO, 2003) differed from those used in prior assessments in that they incorporate a lower level of stratospheric aerosols and thus project a more rapid recovery of the ozone layer. About half of the models project recovery to 1980 levels by 2050. A two-dimensional model simulation with the GSFC-INT model by Rosenfield et al. (2002) projects that arctic ozone recovery will be slowest in the spring, with total column ozone returning to 1980 levels after 2050, and earliest in the autumn, with total column ozone returning to 1980 levels before 2035. The results from the two-dimensional models project a range of arctic ozone recovery rates, from about 0.5% per decade to about 2% per decade.

Three-dimensional model simulations for the Arctic are also presented in the assessment (WMO, 2003). These models offer greater insight into the dynamic factors affecting current and future arctic ozone levels. Figure 5.15 shows the spring (March–May) average change in ozone relative to 1980 for the latitude band 60° to 90° N projected by the UMETRAC and GISS models, which are transient simulations, and E39/C, which is a time-slice simulation. In general, the three-dimensional models simulate larger ozone depletion over the Arctic between 1980 and 2000 than do the two-dimensional models. The different three-dimensional models project quite different future ozone levels. The UMETRAC model provides projections through 2020; these projections indicate slow recovery (a few percent) between 2000 and 2020. The E39/C model provides simulations for 1960, 1980, and 1990, and a projection for 2015. The ozone levels simulated by the E39/C model for 1980 are about 6% lower than the 1960 levels, but this large decrease between 1960 and 1980 is not corroborated by observations. The E39/C model simulates the same rate of decrease between 1980 and 1990, while the projections for 2015 show ozone levels above those of 1980 but still lower than 1960. The GISS model is the only three-dimensional model that provides projections beyond 2020; these projections indicate further ozone depletion between 1995 and 2015, with only modest recovery in 2045.

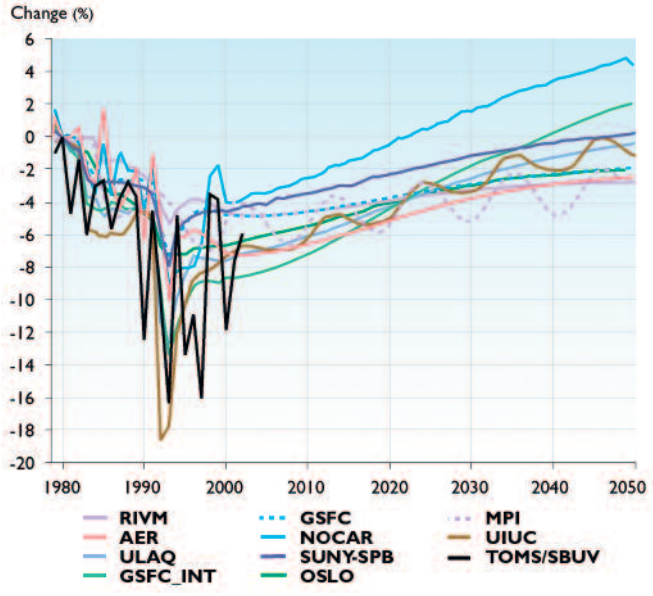

Fig. 5.14. Two-dimensional model projections of the change in spring (March–May) total column ozone relative to 1980; averaged over the band from 60° to 90° N. TOMS/SBUV observations are shown for comparison.

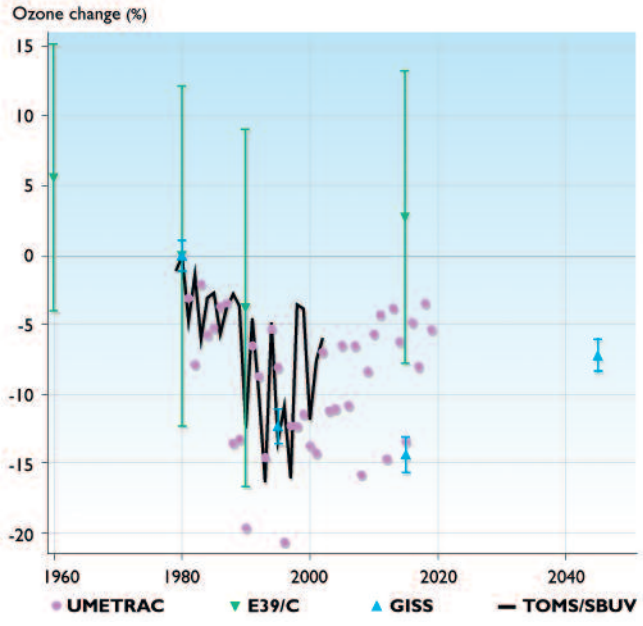

Fig. 5.15. Three-dimensional model projections of the change in spring (March–May) total column ozone relative to 1980; averaged over the band from 60° to 90° N. Error bars represent variability in model projections averaged over 5 years for the GISS model, and twice the standard deviation of the 20 individual years within each model sample for the E39/C model. TOMS/SBUV observations are shown for comparison.

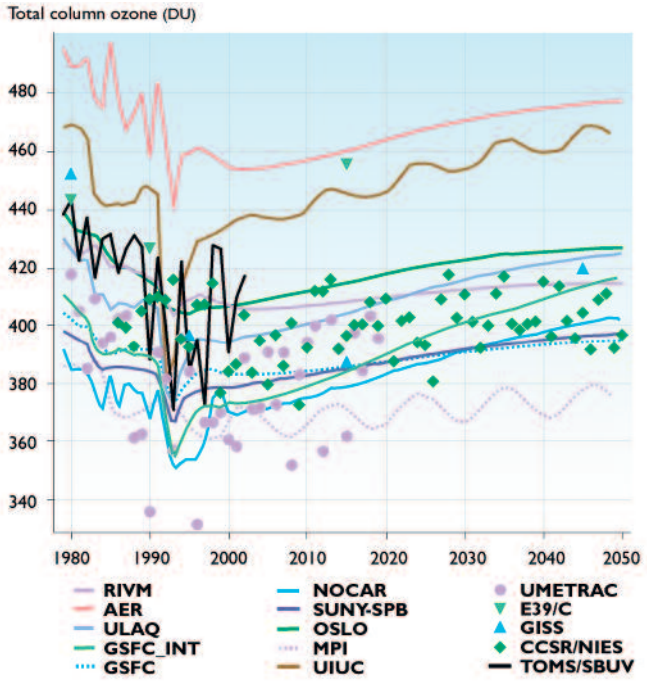

Fig. 5.16. Two- and three-dimensional model projections of spring (March–May) total column ozone averaged over the band from 60° to 90° N. TOMS/SBUV observations through 2002 are shown for comparison.

The two- and three-dimensional model projections of spring total column ozone for the 60° to 90° N band are shown in Fig. 5.16. The results indicate large differences in the projected total ozone column amounts, with most models projecting lower ozone levels than have been observed. All of the models project that ozone levels will remain substantially below pre-depletion levels for at least the next two decades.

The three-dimensional models can offer insight into the spatial distribution of ozone depletion and recovery. As Fig. 5.17 shows, both the E39/C and UMETRAC models project greater changes near Greenland than elsewhere in the Arctic, although the UMETRAC model projects continued depletion and the E39/C model projects earlier recovery for this region. These differences indicate the uncertainty in the spatial distribution of future ozone levels. Zonally symmetric dynamics in the GISS model result in near-zonally symmetric ozone loss and recovery.

Austin et al. (2003) summarized the uncertainties in many of the chemistry–climate models that are used to project future ozone levels. Some of the most important uncertainties related to arctic projections are due to the cold temperature biases in the arctic winter that most models have, and that the models are forced with less than half the observed trend in stratospheric water vapor. Differences in gravity-wave and planetary-wave simulations as well as model resolution can lead to very different projections of polar temperatures and transport of ozone to the poles. Arctic ozone depletion is also subject to large natural variability, complicating definitive projections of how ozone levels will evolve (Austin et al., 2003; WMO, 2003).

As this section suggests, modeling past and future ozone levels, particularly in the Arctic, is challenging. One of the primary challenges is the difficulty of simulating observed polar temperatures, which are essential for determining the severity of chemical ozone depletion by anthropogenic chlorine and bromine. Many of the current chemistry–climate models do not reproduce the observed occurrence of PSCs or the large observed increase in PSC occurrence since the 1960s (Austin et al., 2003, Pawson and Naujokat, 1999; WMO, 2003). Some models are also unable to accurately reproduce the observed ozone loss rates within the arctic polar vortex. As reported by Rex et al. (2004), the limitations of accurately simulating the relationship between potential amounts of PSCs and ozone depletion may be leading to more optimistic projections of arctic ozone levels than are likely to occur given the influences of climate change. The difficulties in simulating arctic stratospheric temperatures stem partly from the strong influence of polar dynamics, and until these processes are better understood, future changes in arctic dynamics and ultimately in arctic ozone levels will be difficult to project.

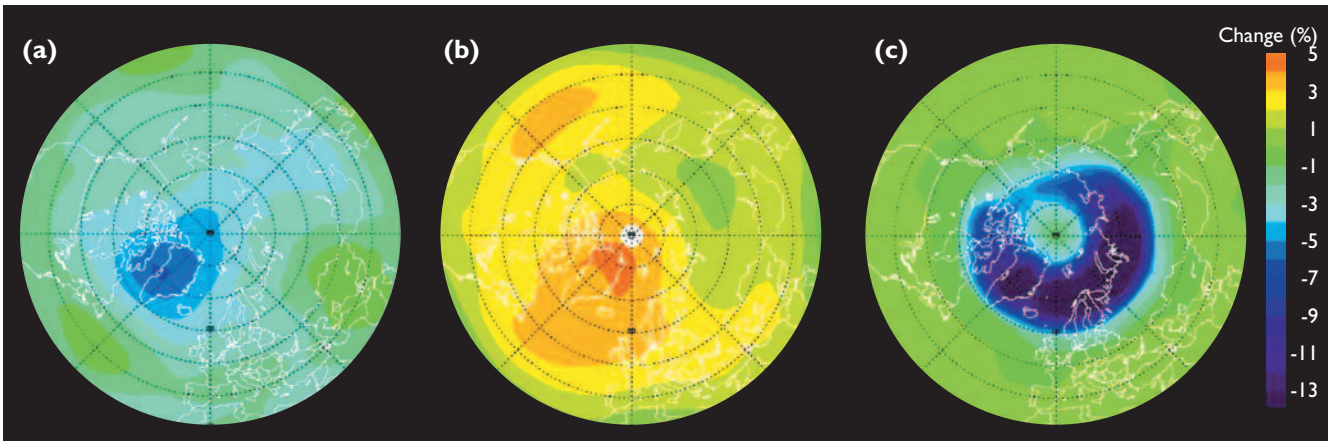

Fig. 5.17. Percentage change in average annual total column ozone projected by (a) the UMETRAC model for the period 2010–2019 (average over the time slice) relative to 1980–1984; (b) the E39/C model for 2015 relative to 1980; and (c) the GISS model for the period 2010–2019 (average over the time slice) relative to 1980.

Because of the anticipated decline in stratospheric chlorine and bromine concentrations resulting from the Montreal Protocol and its amendments, an increase in arctic ozone levels is expected to occur eventually. However, any quantitative statements concerning the timing and magnitude of arctic ozone layer recovery are highly uncertain.

5.7. Future changes in ultraviolet radiation

While there are early signs that the Montreal Protocol and its amendments are working, a return to normal ozone levels is not likely to occur for several decades. Scientists primarily concerned with chemical contributions may be interested in the earlier signs of ozone recovery (for example, a reduction in the downward trend in ozone levels), while those studying UV radiation and its effects are likely to focus on an overall recovery from depleted values. Delayed recovery of total column ozone in the polar regions implies the possibility of increased UV irradiance at northern latitudes for another one to five decades. Although there are many uncertainties in the models, current projections indicate that ozone depletion over the Arctic is likely to continue for 50 years or more (WMO, 1999, 2003).

Because ozone depletion at polar latitudes is expected to persist, and possibly worsen, over the next few decades (WMO, 2003), UV radiation reaching the surface is likely to remain at levels greater than those observed in the past. Using ozone projections from the GISS chemistry–climate model, which projects the most severe, longest-lasting ozone depletion of any of the models, Taalas et al. (2000) estimated that the worst-case spring erythemal UV radiation doses averaged over the period from 2010 to 2020 will increase by up to 90% relative to average 1970–1992 conditions. By comparison, the annual UV radiation dose increases for the entire Northern Hemisphere are estimated to be 14% for 2010 to 2020, and 2% for 2040 to 2050. Ultraviolet radiation projections depend on future ozone levels, which are highly uncertain. Future UV radiation levels will also be complicated by changes in snow and ice cover and albedo and are likely to vary locally as well as regionally. Reuder et al. (2001) simulated ozone levels using various CFC emissions scenarios, as well as changes in future temperatures and dynamics and used these results to project UV radiation conditions over central Europe. Their results indicate a slight increase in spring UV radiation levels between the present and 2015, and the potential for continued above-normal late winter and spring UV radiation levels through approximately 2050, although future climate-induced forcings of arctic ozone recovery still need to be better explored.

Continued increased UV levels are likely to have profound effects on human health. Slaper et al. (1996) estimated that reaching minimum ozone levels around the year 2000 (based on the Copenhagen Amendments to the Montreal Protocol) would be likely to result in a 10% increase in skin cancer 60 years later. While these results were based on analyses at mid-latitudes, they illustrate the long-range effects of increased UV radiation levels on human health. These human health effects are discussed in more detail in section 15.3.3.

While humans can choose clothing, sunglasses or goggles, and other protection to reduce their exposure to UV radiation, plants and animals in the Arctic must adapt to their environment through slower, biological means or by migrating or seeking shelter to reduce their exposure. Section 7.3.2 describes plant and animal adaptations to UV radiation exposure in greater detail. These adaptations include thick leaves and protective pigments in plants and reflective white feathers and fur in arctic animals. Sections 8.6 and 9.4 also address potential adaptations and protections for organisms experiencing increased UV radiation exposure. If UV radiation levels in the Arctic remain high into the 21st century, the relative incidence and time frame of effects on human health and on plants and animals would also be extended.

Ozone levels and future changes in these levels are not the only factors affecting anticipated UV irradiance in the Arctic; aerosol concentrations and cloud cover also play a role (see section 5.4). These factors are likely to change, at least on a regional basis, in the future. A more active hydrological cycle in the Arctic, projected to occur as a result of climate change (IPCC, 2001), is likely to result in changes in cloud cover. In general, increases in cloud cover will reduce UV irradiance at the surface, except in certain conditions when multiple reflections between clouds and a snow-covered surface may enhance the UV radiation dose. Current model projections suggest that cloud cover over the Arctic is likely to increase in some areas and decrease in others. Uncertainties in projections of future aerosol or cloud changes and in the understanding of aerosol–cloud–UV radiation interactions complicate projections of future surface UV irradiance.

Sea ice and snow cover are also likely to be affected by climate change, and can have major effects on incident UV radiation, both by reflecting radiation and by protecting organisms buried beneath it. Almost all climate models project increases in precipitation in the Arctic (IPCC, 2001). Model projections indicate that overall temperatures in the Arctic are likely to be warmer, suggesting that for late spring through early autumn, much of the precipitation increase is likely to be in the form of rain, or of rain falling on existing snow cover, possibly enhancing the rate of snowmelt. The extent and duration of snow cover in the Arctic is important in part because of its relation to UV radiation doses. In the polar regions, UV radiation doses affecting biological organisms depend greatly on the local surface albedo. Reflection off snow, for example, can increase the amount of UV radiation reaching an organism's face, eyes, or other surface. These amplified doses can

be particularly pronounced at low solar elevations, or in the presence of increased multiple scattering by non-absorbing aerosols. Any shift in the extent or duration of snow cover, particularly during the critical spring months, is likely to amplify the biologically effective UV radiation doses received by ecosystems potentially already stressed by climate change. In areas normally covered by snow, early spring snowmelt, such as has been observed by Stone et al. (2002), is very likely to leave organisms at ground level vulnerable to increased UV irradiance during periods of spring ozone depletion.

UV radiation exposure can have a range of effects on humans and on the overall arctic environment. Human health concerns include skin cancers, corneal damage, cataracts, immune suppression, and aging of the skin. Ultraviolet radiation can also have deleterious effects on both terrestrial and aquatic ecosystems, and is known to affect infrastructure through damage to plastics, wood, and other materials. Many of these effects require increased study. The combination of future climate change and the likelihood of prolonged increases in arctic UV radiation levels present a potentially challenging situation for the people and environment of the Arctic. These effects and some of their expected consequences are discussed in greater detail in sections 7.3, 7.4, 8.6, 9.4, 14.12, 15.3.3, 16.3.1, and 17.2.2.3.

5.8. Gaps in knowledge, future research, and observational needs

Four key areas of research activity will improve the ability of the scientific community to assess the changes in and effects of ozone depletion and UV radiation in the Arctic: addressing unanswered scientific questions concerning variability and long-term changes in both ozone levels and UV irradiance; ensuring accurate and comprehensive monitoring of ozone and UV radiation levels; improving analysis of emerging data and incorporating this new understanding into modeling efforts; and undertaking cross-disciplinary studies to determine the effects of changes in UV irradiance. All four areas are important both to improve scientific understanding and to provide relevant information for policy decisions.

There are a number of unanswered scientific questions concerning the sources of variability in ozone and UV radiation levels in the Arctic. Improved knowledge is needed to quantify the effects of trace gases, dynamics, and temperature on arctic ozone levels. The influence of climate change on both ozone and UV radiation levels needs to be better understood. Understanding the controls on and interactions between various processes will greatly improve projections of future ozone levels. In addition to ozone levels, several other factors, including cloud conditions, aerosol concentrations, and surface albedo, affect surface UV irradiance. The interactions between and overall influence of these factors are still the subject of much uncertainty, but future changes in any one parameter – for example, in cloudiness or snow melt timing – could substantially affect UV radiation levels in the Arctic. Quantifying these factors across the Arctic will provide opportunities to more realistically assess changes in UV irradiance and their effects.

Although many of the questions regarding the cause of ozone depletion have been addressed and confirmed in a number of studies both within and outside the Arctic, questions still remain concerning the future of ozone and UV radiation levels in the Arctic. Current model projections vary widely in terms of future ozone and UV radiation levels, with the large differences due mainly to uncertainties regarding the roles of dynamics, temperature, and trace gases. Because of these uncertainties, continued monitoring of both ozone and UV radiation levels in the Arctic is important. Ozone levels and the stratosphere are also known to play a key role in influencing and modulating climate. Monitoring efforts are necessary both to document the evolution of ozone and UV radiation levels over time and to validate model projections. Past monitoring of ozone and UV radiation levels has shown that changes occur on a regional basis even within the Arctic, indicating that regional observations are critical for assessing the overall status of the Arctic. Satellite monitoring of the Arctic for most times of the year has been ongoing for the past few decades, but continued observations will be necessary to understand the evolution of arctic ozone. The continuation of ground-based monitoring of ozone levels depends on available funding and is highly uncertain at this time. Adding UV radiation monitoring in the Russian Arctic and coordinating the existing surface UV radiation and ozone level monitoring throughout the Arctic would allow for a more accurate assessment of the changes that are occurring.

Analyses of emerging ozone and UV radiation data continue to reveal new information concerning the relative importance of trace gases, dynamics, and temperatures in the Arctic. Continued studies are likely to add to the understanding of UV radiation levels in the Arctic. Campaigns with intensive measurements of a variety of parameters as well as detailed monitoring of trace gases and vertically resolved ozone concentrations are important for advancing this understanding. Analyzing these measurements and using the available information in conjunction with model results will help achieve the best possible insight into future ozone and UV radiation levels, as well as the impacts of specific changes in these levels. Recent studies support the idea that advanced three-dimensional models will be fundamental for obtaining improved projections of future ozone levels over the Arctic.

One of the most important issues to address in the Arctic is determining the impacts of increased UV irradiance, particularly as levels are likely to remain higher than normal in the coming decades. This work requires coordinated cooperation between the UV radiation

monitoring community and the biological and impacts communities. Few cross-disciplinary efforts have been implemented so far, but the collaboration that has taken place has resulted in many of the impact studies cited in other chapters of this assessment. Many questions remain concerning the impacts of UV radiation on individual species, ecosystems, and human health. Because the Arctic is likely to have elevated spring UV radiation levels for some time, understanding the magnitude of potential impacts will be critical for future policy decisions. Because of projected future changes in emissions and atmospheric concentrations of various trace gases, future arctic ozone levels are highly uncertain, not only for the next few decades but throughout the rest of the 21st century. Policy decisions regarding trace gas emissions are likely to directly influence ozone levels in the Arctic, and should be based on not only improved understanding of how ozone and UV radiation levels are likely to evolve, but also on increased knowledge of how the ecosystems, infrastructure, industries, and people of the Arctic will be affected.

In addition to the scientific requirements outlined above, there is ongoing concern about maintaining the existing international legislation to protect the ozone layer. Most developed countries have ratified the Montreal Protocol and its amendments, but economic and political pressures have led some countries to suggest that they cannot meet their obligations under the current agreements. The atmospheric concentrations of most ozone-depleting substances are decreasing due to compliance with the Protocol and its amendments, but future compliance is uncertain and requires vigilance and cooperation. In addition, because arctic ozone and UV radiation levels are strongly influenced by climate change, including the effects of changes in temperature, trace gas concentrations, and dynamics, international legislation regarding climate change is likely to directly affect arctic ozone levels in the coming decades. Any climate change policies need to be considered in light of the impacts both on arctic climate and on arctic ozone and UV radiation levels.

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