Chapter 11

Climate Change, Ozone, and Ultraviolet Radiation

Contents

11.1. Introduction

Global climate change is a growing concern, especially in Arctic regions where increases in temperature from anthropogenic influences could be considerably higher than the global average. Climatic changes are not new to the Arctic or its peoples. Indigenous peoples of the far north have adapted to the austere climate; different groups have found their own unique ways to harvest food and provide clothing, tools, and shelter. At times the climate has warmed or cooled relatively suddenly and people have either adapted, moved, or died off. The paleo-archaeological record, indigenous peoples' oral history, and historical documents provide evidence of climatic changes for thousands of years. Today, people of the Arctic, whether they continue to live close to the land or live in urban centers, must again confront rapid changes in climate. Various records over the last 40 years confirm that the rate of global warming has been greatest over Eurasia and North America between 40°N and 70°N (IPCC 1996a). Arctic research substantiates these observations through direct and indirect indicators of climate change. Sea ice, snow cover, glaciers, tundra, permafrost, boreal forests, and peatlands are all responsive to subtle variations in sunlight, surface temperature, ocean heat transport, air and ocean chemistry, and aerosols in the atmosphere. Compared with the rest of the globe, the Arctic climate is very sensitive to change because of a complex series of interactions and positive feedback processes among the region's oceanic and atmospheric circulation patterns, temperature regime, hydrologic cycle, and sea ice formation (Barry *et al.* 1993a, Kellogg 1983, Mysak 1995).

Present models of the Arctic climate system suggest that positive feedbacks in high-latitude systems amplify anthropogenically-induced atmospheric changes and that disturbances in the circumpolar Arctic climate may substantially influence global climate (IPCC 1990a, 1992a, 1996a). The extreme sensitivity of the Arctic's climatic and ecological systems implies that the Arctic will be profoundly affected by anthropogenic climate change (Quadfasel *et al.* 1991, Walsh 1991). Both positive and negative feedbacks complicate Arctic climate change, making it difficult to model or predict. A number of different positive feedbacks have been identified for the Arctic. For example, sea ice and snow reflect a much larger fraction of incident sunlight than water and soil, so that a reduction of sea ice and snow causes a perturbation in the energy budget, amplifying warming in the Arctic. This warming is transferred globally and, at the same time, feeds back regionally to further reduce snow and ice extent. Concurrently, as temperature rises the air is able to hold more moisture which increases the greenhouse effect, adding to the temperature increase. Of course, actual processes of climate change are not so simple and data analyses reveal conflicting trends in climate and temperature. Other subtle systemic feedback mechanisms which could offset the primary feedbacks may be present. In summary, while strong positive feedback mechanisms have been identified and are expected to play major roles in climate change in the future, the complex interactions of these and other environmental feedbacks, both positive and negative, are not fully understood.

Another major factor affecting climate is stratospheric ozone. Stratospheric ozone is an important indicator, as well as an agent, of climate change. Stratospheric cooling, a direct result of what is generally referred to as global warming, allows for increased ozone destruction in the Arctic. As an absorber of solar radiation, ozone partly controls the temperature structure of the atmosphere, influencing dynamical as well as thermal properties of the atmosphere. Ozone depletion is an increasing concern in the Arctic as anomalously low levels of ozone have been recorded in recent years. The Scientific Assessment of Ozone Depletion: 1994 concluded that 'Chlorine- and bromine-catalyzed ozone loss has been confirmed in the Arctic Winter' (WMO 1995). The destruction of ozone increases UV radiation at the earth's surface, making ozone critically important to the well-being of the biosphere and human health.

The increase in UV has become more significant since evidence shows that changes in UV levels due to ozone depletion can be large, particularly in spring-time. Elevated UV levels adversely affect aquatic and terrestrial ecosystems, as well as humans. A combination of high early-summer biological activity and changes in UV levels makes the Arctic an area where ozone depletion may have notable effects in the near future. Many questions remain which need to be addressed. Little effort has been expended to study the effects of UV on the Arctic biosphere, including human health, although the International Arctic Science Committee concluded in 1995 that 'There is a pressing need to quantify ozone-dependent UV-B effects on diverse Arctic ecosystems under their current conditions.'

This chapter provides an overview of current research and knowledge on climate change, ozone depletion, and UV-B radiation in the Arctic. After this brief introduction, the second section gives an overview of climate change, including evidence and measurements of climate change as well as what is known about the various components of the climate change system in the Arctic. Section three addresses stratospheric ozone, and section four covers UV radiation in the Arctic. The general effects of climate change and UV radiation on aquatic and terrestrial ecosystems are examined in section five, and section six examines their effects on Arctic peoples and their communities. Section seven provides an overview of major international programs involving research and assessments. Recommendations for further international efforts are made in section eight.

11.2. Climate change

11.2.1. Dynamic interactions

The complexity of climate change in the Arctic can be observed through the strong and dynamic changes in the energy, trace gas and hydrological balances of the Arctic. Solar radiation inside the Arctic Circle varies seasonally from continuous sunlight to no sunlight. Depending on ambient conditions, the Arctic can be either a source or a sink of particular trace gases. The water balance of the Arctic is highly dynamic in its exchange between ice, water, and water vapor. Virtually all components and aspects of the Arctic climate system, from sea ice to surface temperature, are determined integrally by the energy, trace gas, and hydrological cycles. The balance of these three systems will be discussed first to allow an overview before the individual components of the Arctic climate system are covered in more detail.

11.2.1.1. Energy balance

Alteration of the Earth's radiation balance is the most direct way to affect climate. The Earth's surface temperature is a result of the balance between the energy fluxes in a small layer near the surface. The amount of solar radiation absorbed by the surface is determined by many factors. Apart from solar elevation angle and day length, the most important factors are atmospheric scattering and absorption by clouds, haze, atmospheric chemistry, and surface albedo. Net radiation (the balance between longwave and shortwave fluxes) in the Arctic is strongly positive in the summer and negative in the winter. The interactions and feedbacks between cloud cover, albedo (percentage of incoming radiation reflected), radiation, sea ice, and snow cover are important modulators of the radiation balance and represent key uncertainties in evaluating the role of the Arctic in the global climate system. The latent heat flux is near zero in the winter, when there is very little capacity for the air to hold moisture; in the summer it is usually negative, indicating evaporation. The sensible heat flux is a principal component of the surface energy balance in the winter over continents as is the conductive flux over the oceans. The conductive flux depends on the ice thickness and snow cover and is the cause of the relatively warm temperatures observed over the frozen oceans.

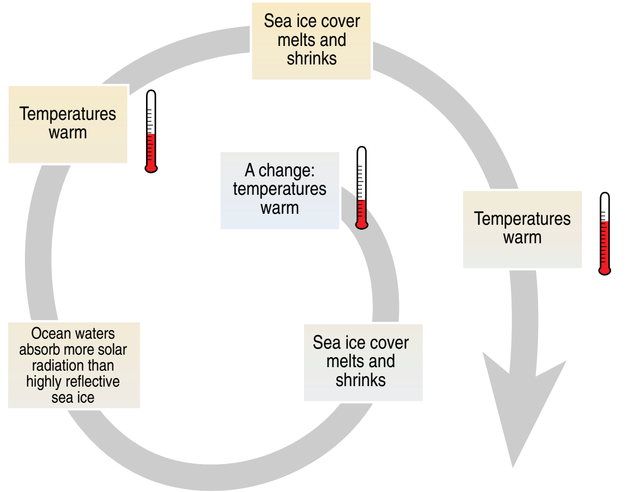


Figure 11·1. An example of a positive feedback loop. Warming leads to a decrease in sea ice cover which in turn leads to a decrease in albedo over the ocean, the result of which is further warming and further decreases in the sea ice cover. In the reverse situation, if sea ice cover, and hence albedo, were to increase, the feedback would lead to cooler air temperatures, thus promoting greater sea ice cover. In either case, the feedback loops are positive, i.e. the change is amplified by the system feeding-back onto itself. Negative feedbacks help regulate a system causing changes to be moderated or diminished.

Changes in sea ice extent have a major bearing on the energy related processes between ocean and atmosphere. This dependence has been described as sea ice-albedo feedback, i.e. the chain of events following an initial warming of the near-surface temperatures, a reduction in sea ice, and a subsequently enhanced energy transfer from atmosphere to ocean, which leads to further reductions in sea ice. This dependence is shown in Figure 11·1. A similar process operates over snow-covered land.

Cloud cover, extent of sea ice, and snow cover can influence the planetary albedo and large-scale albedo gradients, with consequent impacts on atmospheric circulation. As surface fluxes of solar and long-wave radiation are strongly influenced by cloud cover, variations in cloudiness affect polar sea ice and snow cover. Along with a direct effect on the radiation balance, clouds have an indirect bearing on the stability of the atmospheric boundary layer and thus on the sensible and latent heat fluxes.

11.2.1.2. **Trace gas balance**

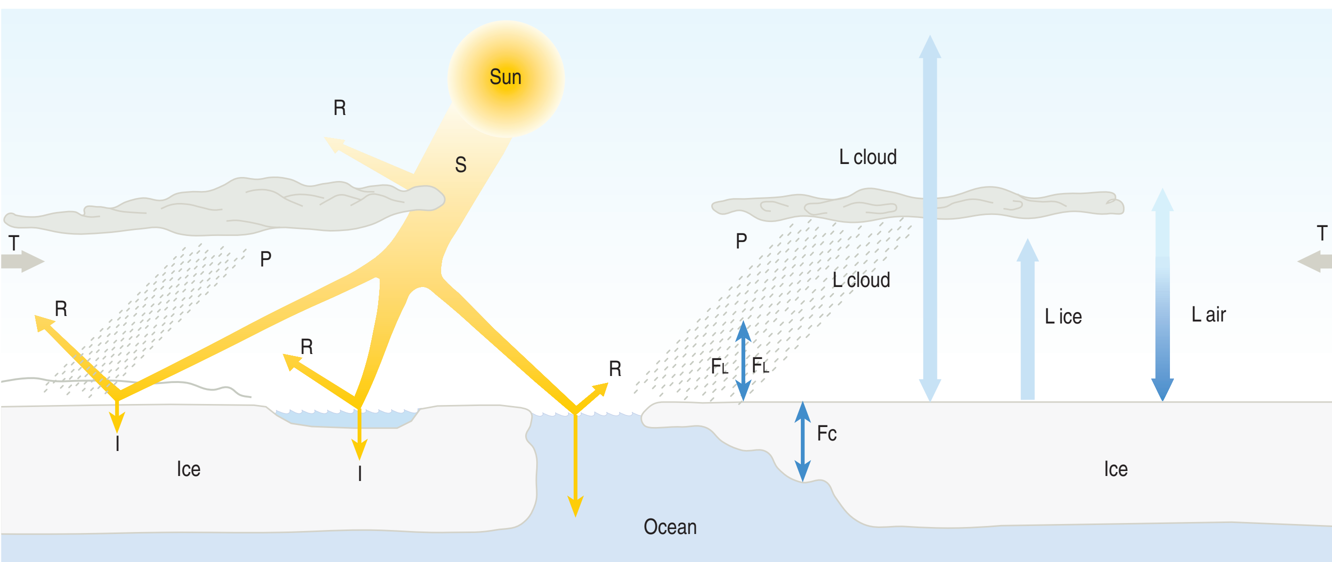
The greenhouse effect, i.e. the warming of the troposphere and the Earth's surface from the absorption of infrared radiation by certain gases, is one of the important forcing mechanisms of climate. Carbon dioxide (CO_2), methane (CH_4), nitrous oxide (N_2O), tropospheric ozone (O_3), and chlorofluorocarbons (CFCs) are the most important ones, and their atmospheric abundances have all increased since the last century. The calculated direct global mean change in the surface radiative heat balance is +2.5 W/m^2 due to the addition of the above gases (O_3 not included), without taking any potential feedbacks into consideration (IPCC 1996a). This amount is equal to one percent of the global mean solar radiative forcing at the surface.

Greenhouse gases are emitted, stored, and absorbed by ecosystems on the land and in the oceans as a result of natural processes. The concentrations of these gases can be altered as a result of industrial development. Long-lived trace gases (>1 year) have global influences because they are redistributed by transport in the atmosphere. A shorter-lived gas such as ozone has a more regional influence. Because of greater atmospheric stability, the concentrations of many trace gases in the Arctic tend to be somewhat higher in the cold season.

Perturbations to climate in the Arctic may increase the emissions of CO_2, CH_4, and N_2O, which will feed back into the global forcing of climate. The emissions are influenced by a host of factors, such as soil temperature and moisture, nutrient deposition/loss, snow cover, cloud cover, and sea ice extent. Acting slowly over decades, but no less important to the trace gas balance, are changes in the species composition of Arctic ecosystems that are likely to result from climate change. All of these processes are poorly or incompletely understood. Understanding the complex systems governing trace gases is crucial to predicting future climate changes.

11.2.1.3. **Hydrological cycle**

The hydrological cycle of the Arctic encompasses a variety of components (Figure 11·2) including precipitation, sea ice, river run-off, glaciers, icebergs, clouds, and humidity. Water



F: Turbulent heat flux in water (F_L - Latent heat flux due to evaporation and condensation, F_C - Convective flux) I: Absorbed radiation L: Thermal infrared radiation fluxes P: Precipitation R: Albedo (Reflection) S: Solar infrared radiation fluxes T: Net advection of moist static energy.

Figure 11·2. The hydrologic balance in the Arctic is highly dynamic on the seasonal time scale with large and rapid ice melts in the spring. On land, this ice and snow melting results in large rapid floods and surges in rivers. In the oceans, the ice melt results in large areas of the Arctic becoming available for biological growth and activity. On longer time scales, climate change in the Arctic could release glacial waters increasing the present sea level globally.

transport and storage in the Arctic are fundamentally different than in other parts of the world. Much of the water in the Arctic is frozen most of the time. Seasonal freeze/thaw cycles of sea ice, snow, and permafrost, and perennial freezing of the polar ice cap, glaciers, and below-ground ice control the exchange of water, trace gases, and water-borne materials.

Water vapor and temperature are closely related to the hydrological cycle in both the oceanic and terrestrial regions. Leads and polynyas in sea ice release a large flux of water vapor into the atmosphere. Similarly, warmer air temperatures and open water in lakes and drainage systems and melting of permafrost allow water vapor to be released into the atmosphere. This water vapor is then available for cloud formation and potential precipitation release elsewhere.

The Arctic region's oceans include 25% of the world's continental shelf areas. Of all of the world's coastal waters, the near-shore Arctic Ocean is the region most affected by the delivery of freshwater from the adjoining drainage system (Aagaard and Carmack 1989). These discharges may in turn exert important controls on the formation of Arctic sea ice and hence the albedo and radiation balance of the planet; changes in delivery of freshwater from a warming climate would thus affect the global energy balance (Semtner 1987, Rowntree 1989, Allard *et al.* 1995, cf. Hakkinen 1990).

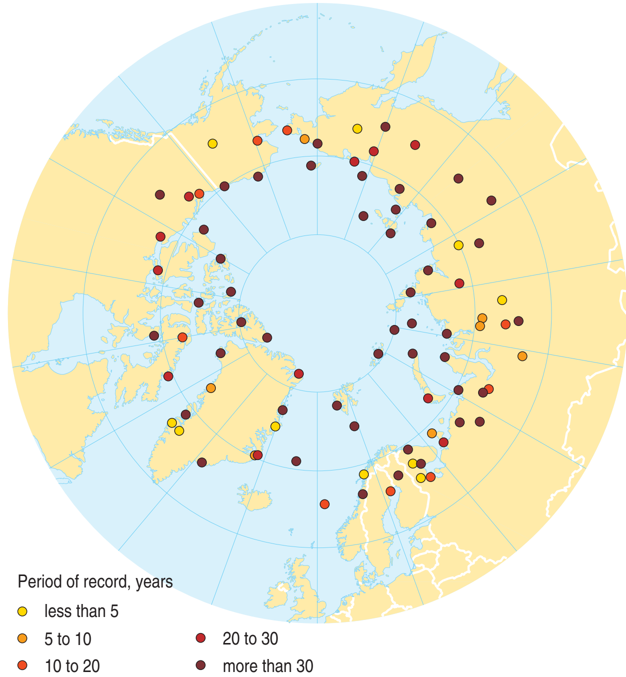


Figure 11·3. Locations of upper-air meteorological monitoring stations in the Arctic (from Kahl *et al.* 1992). Meteorological station density for surface observations is greater for inland areas of the Arctic, while observations of any kind are sparse over the Arctic Ocean.

11.2.2. Climate change: Methods of assessment and recent trends

A variety of direct and indirect methods are available to assess climatic and environmental changes over a range of spatial and temporal scales. It is clear that examination of both direct and indirect indicators of climate change on short and long time scales is necessary to understand the Arctic's climate history, to make policy decisions about anthropogenic effects on climate, and to make valid predictions about the future.

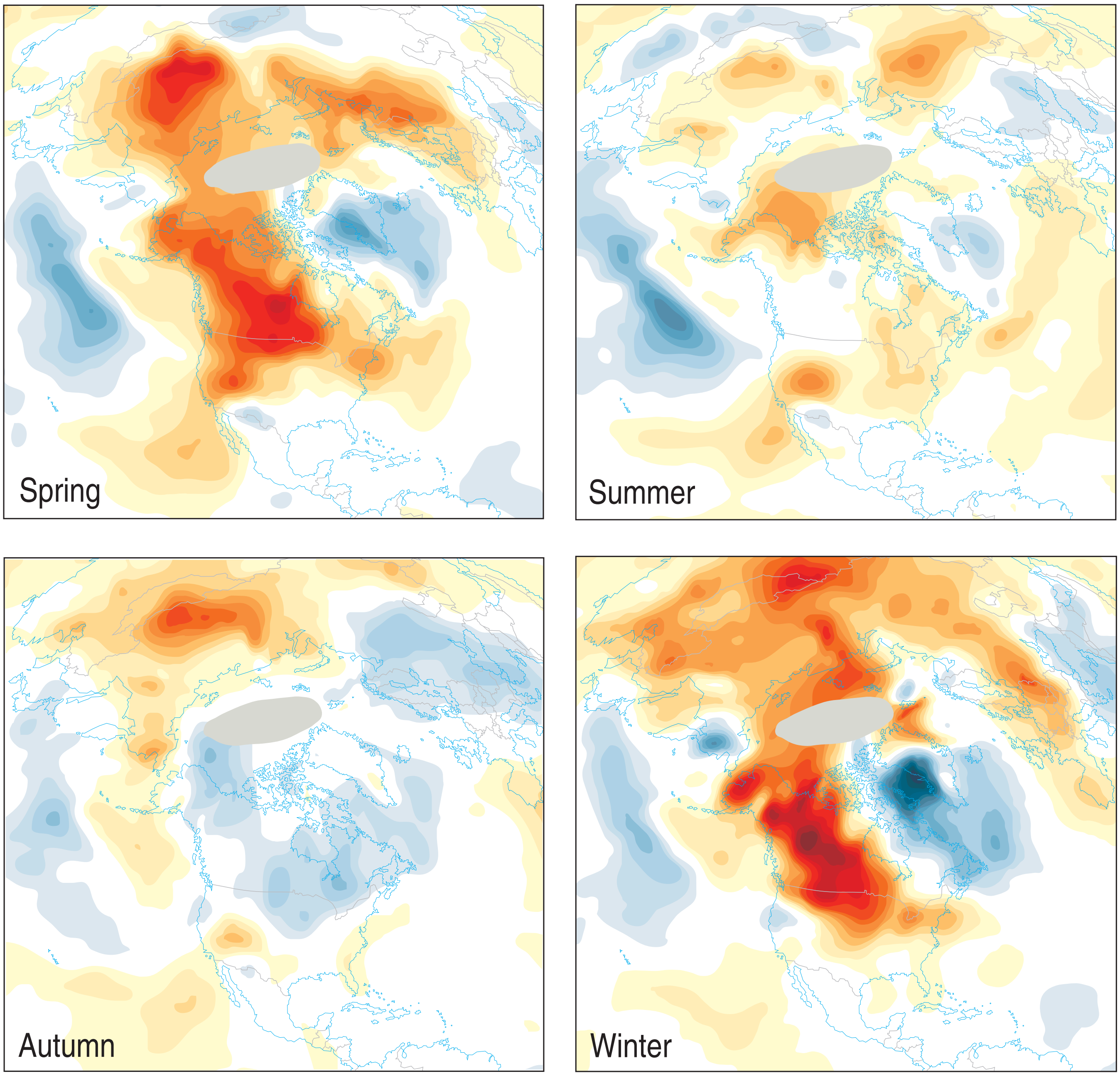
Sources of climate information are of varied quantity and quality and can be difficult to interpret. Recent trends of environmental change frequently conflict with one another and existing data are often insufficient to provide a clear picture. Surface air temperature is the most obvious direct indicator of climate change, yet circum-Arctic temperature records are not long and are particularly sparse over the Arctic Ocean. Indirect indicators such as borehole temperatures, snow cover extent, glacier recession, precipitation, and vegetation changes yield data that so far support direct indicators of warming (IPCC 1996a). While meteorological measurements provide temperature and precipitation records on short time scales, paleoecological studies and historical records provide evidence of past changes in climate. Early records of plant and animal life contained in ice, fossils, and sediment cores reveal long-term changes in the environment, while ice cores can provide evidence of mean or extreme temperatures and atmospheric gas composition over 200 000 years or more (Alley *et al.* 1996). Glacier mass has been used to derive estimates of warming since the end of the Last Ice Age. Indigenous peoples and historical records provide information about a variety of climate change indicators such as changes in animal populations, plant distribution, water levels, sea ice thickness and extent, and temperature and precipitation, while archaeological research can help reveal local and regional climatic changes.

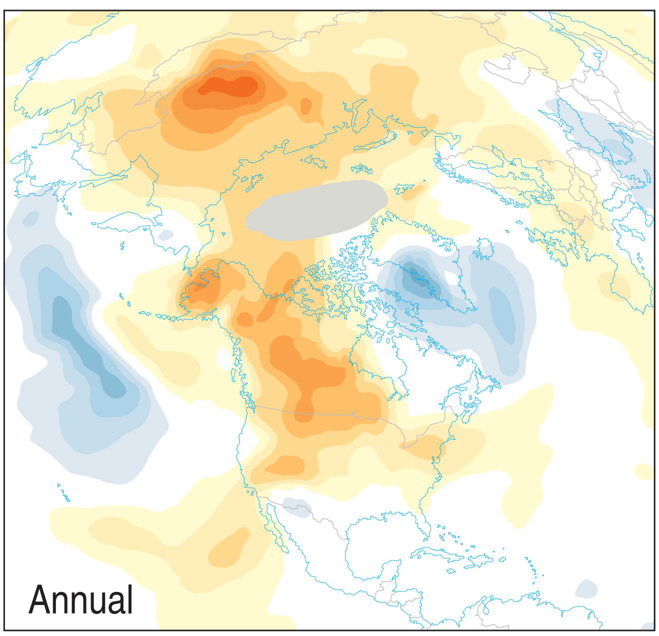
11.2.2.1. Temperature records

Surface and atmospheric temperatures in the Arctic have changed in complex and variable ways over the past several decades. Knowledge of long-term variations in Arctic temperatures is limited because of large areas that have never been monitored or have been monitored only sporadically in space and time (Figure 11·3). Few stations north of 70°N are available either on land or on sea ice, and data from them must be augmented with drifting-buoy data (Colony and Thorndike 1984) and remotely-sensed information (Maslanik *et al.* 1996). Multi-decadal trends have been determined for available surface and upper-air temperatures. Trend analyses of surface temperatures typically rely on gridded surface temperature datasets representing an assimilation of raw station observations (Chapman and Walsh 1993, IPCC 1990a, 1992a). Upper-air analyses are based on rawinsonde or dropsonde temperature profiles (Kahl *et al.* 1993a, 1993b) and satellite monitoring platforms (Christy 1995). The determination of long-term trends is hampered by non-climatic factors which can obscure real trends or produce artificial ones. These factors include changes in instrument response characteristics, balloon ascent rates, data reporting, and correction procedures (Gaffen 1994, Parker and Cox 1995, Skony *et al.* 1994).

Surface temperatures in the Arctic vary widely over land and sea. Analyses of surface observations indicate that warming has occurred over the northern land masses during the past century (Jones 1994, Parker *et al.* 1995, IPCC 1996a). Inland Arctic areas of central Siberia and North America have warmed by 1.5°C per decade (Jones and Briffa 1992). This warming trend is particularly evident in winter and spring (Chapman and Walsh 1993) (Figure 11·4). Cooling trends of 1.5°C per decade have been observed over eastern North America and through the North Atlantic. Surface temperature trends over Fennoscandia and the subarctic seas are smaller and even negative in the southern Greenland region (Jones *et al.* 1986).

For upper air temperatures, systematic balloon measurements (radiosondes) have been made since the 1950s for approximately 50 locations in the Arctic region (Parker and Cox 1995, Kahl *et al.* 1992). Since 1979, satellites have been able to monitor nearly all of the polar region, both for the lower troposphere and lower stratosphere





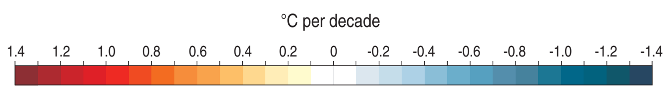


Figure 11·4. Seasonal and annual average surface temperature trends (°C per decade) over the northern hemisphere.

(Christy 1995). Analyses of radiosonde measurements for coastal and inland Arctic stations (Figure 11·3) from 1958-1986 indicate regions of warming and cooling, but there is no general circumpolar trend and trends at most stations are not statistically significant (Kahl *et al*. 1993b, Michaels *et al*. 1995). Tropospheric temperature profiles over the Arctic Ocean similarly fail to reveal widespread significant trends since 1950, although surface cooling has been detected in winter and autumn (Kahl *et al*. 1993a). Further analysis of lower tropospheric temperatures over the Arctic Ocean has revealed a strengthening of the thermal stability of the Arctic inversion layer (Kahl *et al*. 1996), which exerts a strong influence on Arctic pollution as well as on fluxes of mass and momentum between the atmosphere, ice, and ocean.

Satellite monitoring of lower tropospheric temperatures since 1979 (Christy 1995) suggests an Arctic warming trend of 0.05°C per decade, somewhat greater than that of the Earth

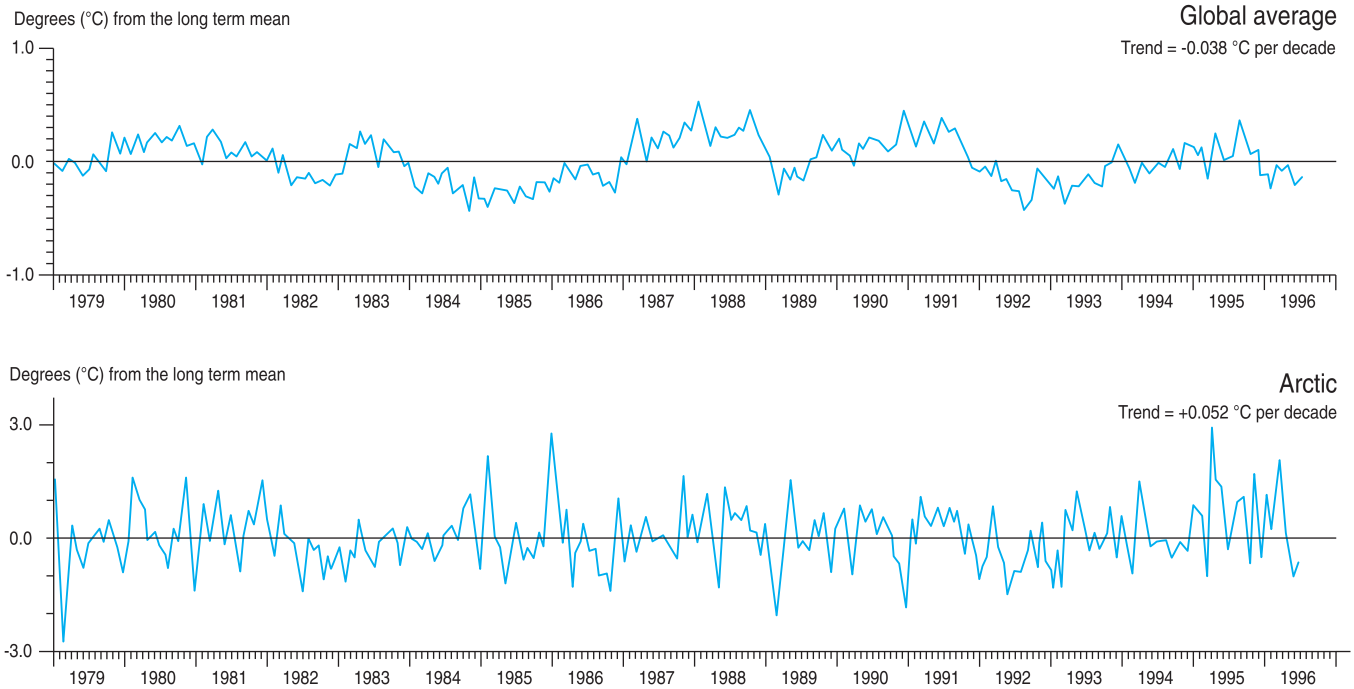


Figure 11·5. Lower tropospheric temperatures since 1979, from satellite monitoring (Christy 1995). High variability is characteristic of the Arctic region due to its isolation from the moderating influences of warm, lower latitude, ocean currents.

as a whole (Figure 11·5). However, trend values over short periods are very sensitive to natural fluctuations occurring at the beginning and end of the time series. In the Arctic, temperature comparisons of the lower troposphere with the surface indicate that the surface is warming more rapidly. The lower stratospheric temperature record reveals large fluctuations since 1979 (Christy 1995). The global record shows two episodes of remarkably sudden stratospheric warming caused by infrared-absorbing aerosols from volcanic eruptions: the combination of Nyamuragria and El Chichon in 1981 and 1982, and then Mount Pinatubo in 1991 (Figure 11·6). Globally, post-eruption surface temperatures fell to lower levels than pre-eruption values in both episodes.

The Arctic stratospheric cooling trend of -1.01°C per decade (Figures 11·6 and 11·7) is the largest decrease seen on the globe. Loss of stratospheric ozone is related to the decreasing temperature since ozone maintains the temperature in the stratosphere by absorbing heat. The geographic pattern of the stratospheric trends (Figure 11·7) is more uniform than in the troposphere (Figure 11·8), suggesting that the physical forcing mechanisms for tropospheric climate change are more strongly dependent upon regionally varying surface processes.

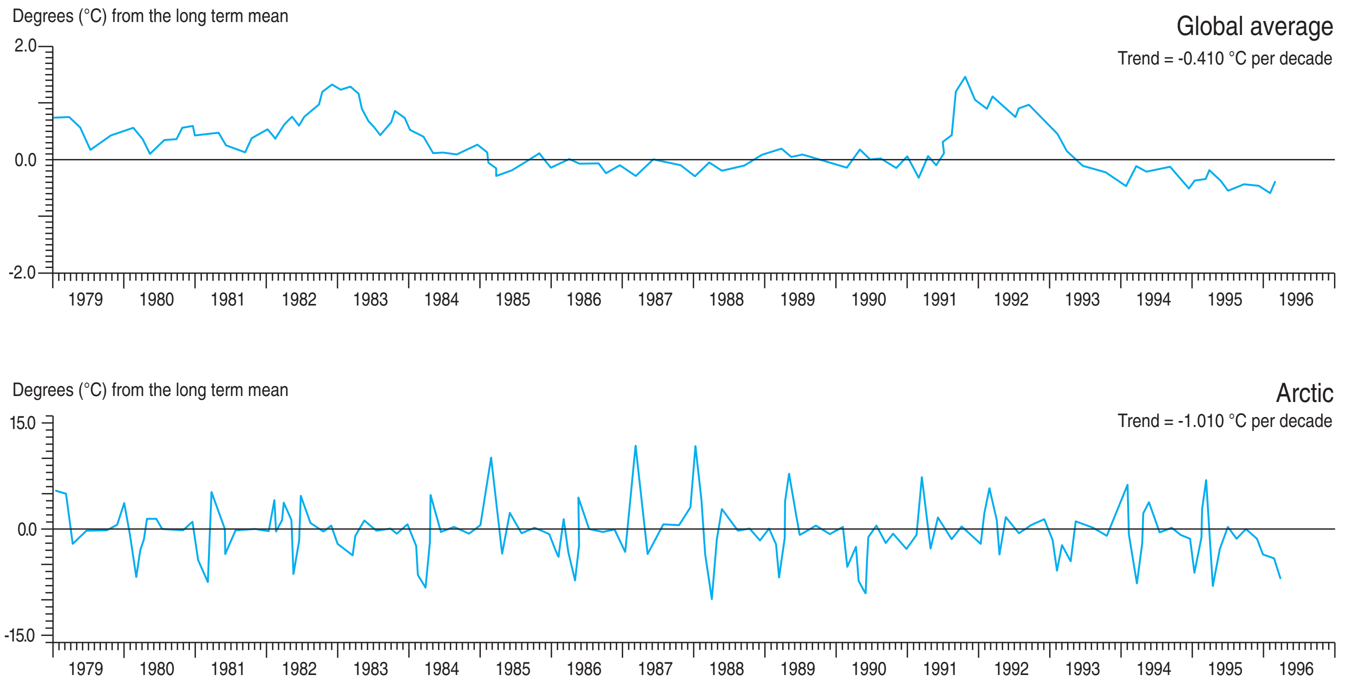


Figure 11·6. Lower stratospheric temperature record since 1979 (Christy 1995). Large variations in stratospheric temperatures are due to sudden stratospheric warming and cooling (SSWC) events which occur in Nother winter (and are related to tropospheric planetary waves), and the impact of volcanic aerosols. The north polar regions exhibit the most dramatic cooling trend of the planet, which is consistent with Arctic ozone depletion.

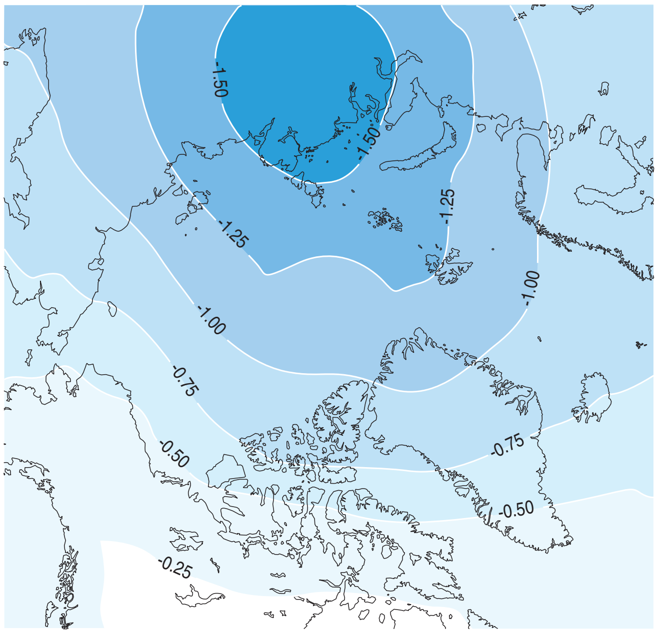


Figure 11·7. Lower stratospheric (ca. 120-40 hPa) Arctic temperature trends (°C per decade, January 1979 to February 1996), as monitored by MSUs on polar orbiting satellites. The entire Arctic stratosphere has experienced a cooling trend which is maximised over Siberia.

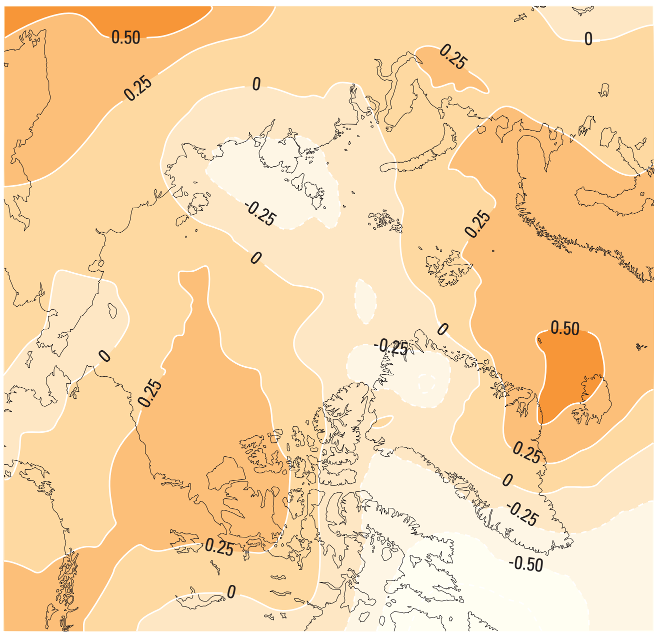


Figure 11·8. Lower tropospheric Arctic temperature trends (°C per decade, January 1979 to February 1996), as monitored by MSUs on polar orbiting satellites.

11.2.2.2. **Radiatively important trace substances**

Certain trace substances alter atmospheric radiative forcing through enhancement of the greenhouse effect or through scattering and absorption of solar radiation (e.g. aerosols and clouds). Records of these substances give an indirect method of assessing climate change. Because of the rapid increase in its levels globally over the last two centuries, carbon dioxide (CO_2) has been the major focus of research on effects of greenhouse gases on temperature. However, CH_4, N_2O, H_2O, O_3, and CFC's, whether their sources are in the Arctic or in the mid-latitudes, also contribute to the greenhouse effect (Wang 1986). The question of long-term changes in the composition of the Arctic atmosphere can only be addressed by continued monitoring of trace gases.

Currently the Arctic tundra as a whole is a sink for CO_2, yet changes in vegetation could reverse the net CO_2 flux and increase global warming. Already, results of studies of CO_2 on two tundra ecosystems (Oechel *et al.* 1993, 1994, Malmer and Wallen 1996) indicate they have switched from being sinks to being sources of atmospheric carbon due to recent warming. It is estimated that between 10 and 25% of global wetland emissions of CH_4 are from north of 60° (Matthews 1993). Whalen and Reeburgh (1992) showed that CH_4 emission in a very wet year could be four times that in a dry year, suggesting that increased precipitation and thawing permafrost could release a globally significant amount of CH_4 into the atmosphere.

Recent studies show that air from lower latitudes, containing high mixing ratios of N_2O, can be transported northward and entrained into the polar vortex while air masses containing low mixing ratios of N_2O can be stripped off of the polar vortex and transported to mid-latitudes (Kumer *et al.* 1993, Manney *et al.* 1994b, 1995b, Ruth *et al.* 1994, Sutton *et al.* 1994, Waugh *et al.* 1994). The extent to which this happens and the degree to which springtime mid-latitude ozone depletions are linked to these polar processes is still a matter of debate and needs further investigation. Vertical profile measurements (ozonesondes) over the Canadian Arctic show significant declines in ozone at all altitudes and at all locations since the mid-1980s (Tarasick *et al.* 1995, Logan 1994, Oltmans 1993) (see Figure 11·14, section 11.2.4.3.2). Of relevance to radiative forcing, ozone concentrations in the upper troposphere affect the energy budget, and a recent declining trend in tropospheric ozone may counteract warming caused by other substances.

11.2.2.3. **Water vapor**

Water vapor in the atmosphere is an integral part of the climate system. In the Arctic, water vapor is an important part of the energy and hydrological cycles, particularly near leads and polynyas. Water vapor is also a radiatively active gas; and changes in its concentration affect the Arctic climate. Globally, there are no available observations with which to judge long-term changes in water vapor (IPCC 1994). However, increases in methane are expected to be accompanied by increases in stratospheric water vapor.

11.2.2.4. **Precipitation**

Precipitation is a valuable and sensitive indicator of climate change because it is tied to temperature, the hydrological cycle, and the surface energy balance. Some regional studies have found evidence that a portion, but not all, of observed inter-annual snow cover fluctuations can be explained by large scale (i.e., continental to hemispheric) forcing. Precipitation has increased in high latitudes by up to 15% over the last 100 years, most of which has occurred in the winter in northern latitudes within the last 40 years (Bradley *et al.* 1987, Groisman 1991, Karl *et al.* 1993, Groisman and Easterling 1994, Dahlstrøm 1994, Hanssen-Bauer and Førland 1994).

Between 1988 and 1996, annual average snow cover extent was 24.4 million km^2, significantly lower than the 1972-1985 average of 25.9 million km^2 (Figures 11·9 and 11·10). The late 1980s and early 1990s saw a decrease in spring northern hemisphere snow cover compared with the previous 15 years, concurrent with warmer temperatures (Groisman *et al.* 1994b). However, no trends in snow extent are evident, and spring snow covers during 1995 and 1996 were as extensive as those observed in the earlier period

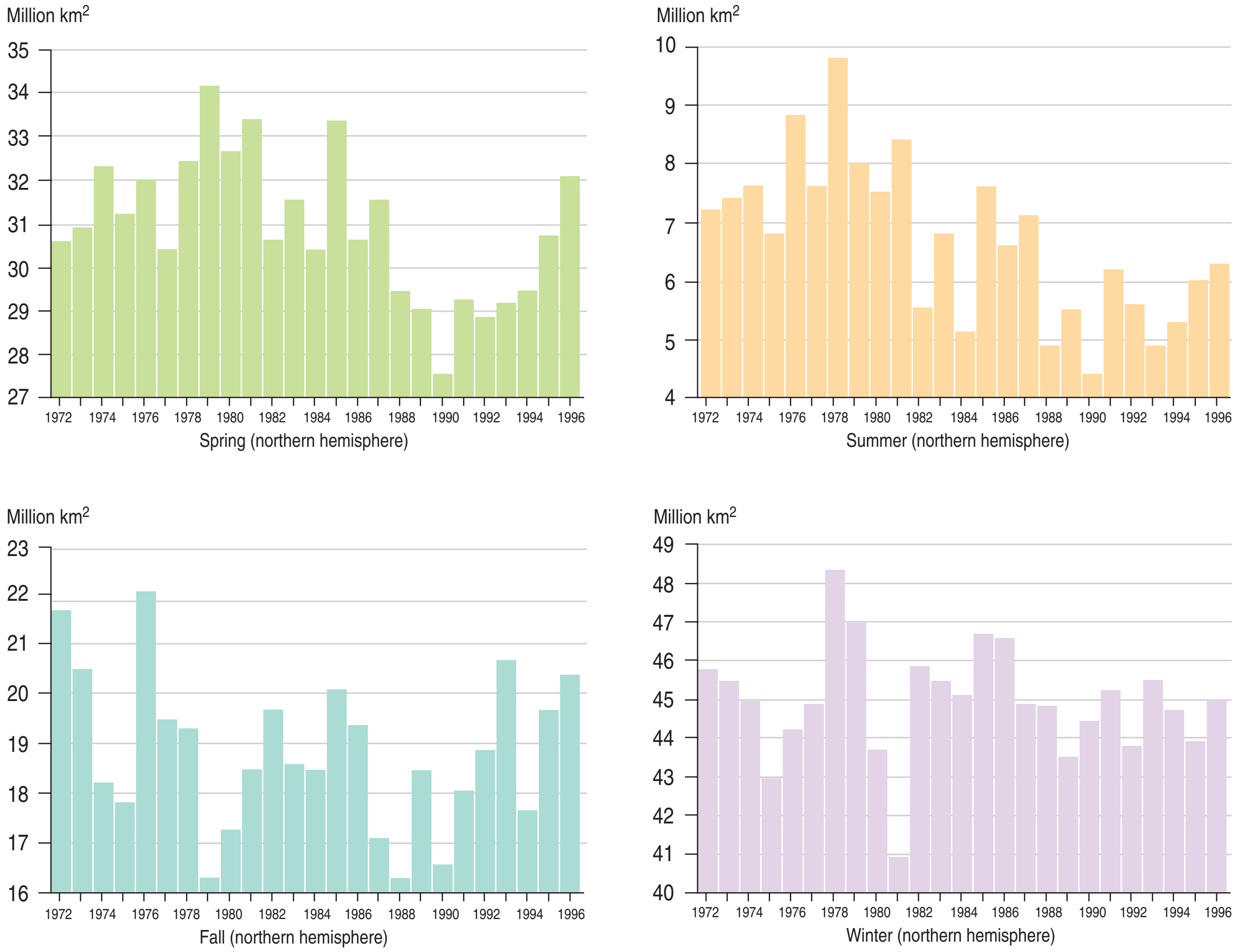


Figure 11·9. Snow cover over northern hemisphere lands between 1972 and 1996 for different seasons. Values are determined from analyses of NOAA snow charts created using visible satellite imagery. (D. Robinson, Rutgers University).

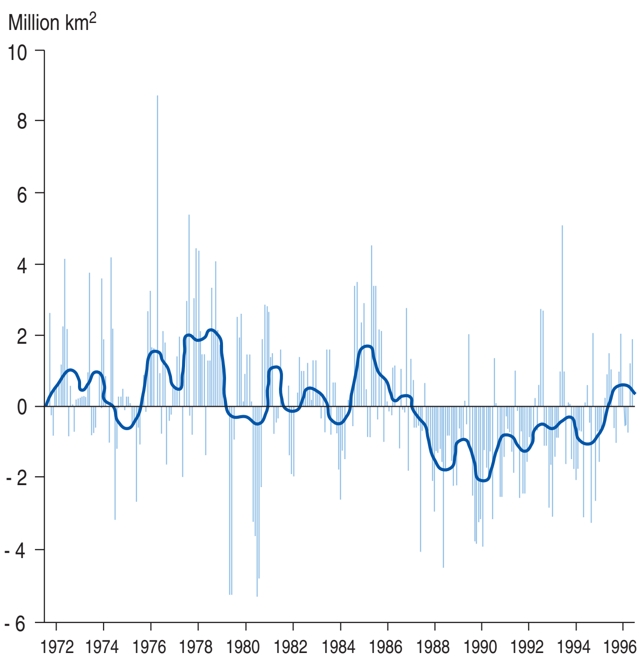


Figure 11·10. Monthly snow cover anomalies over northern hemispheric lands (including Greenland) between January 1972 and August 1995. Also shown are the 12-month running anomalies of hemispheric snow extent, plotted on the seventh month of a given interval. Anomalies are calculated from a mean hemispheric snow extent of 24.5 million km^2 for the full period of record.

(Figures 11·9 and 11·10). Recently, several datasets comprised of *in situ* station observations of snow cover have been examined (Foster 1989, Cao 1993, Brown *et al.* 1995, Hughes and Robinson 1996, Barry *et al.* 1995). Foster (1989) noticed a trend toward earlier spring melt since the late 1960s for much of the North American Arctic tundra.

Data on precipitation are difficult to collect due to the high degree of spatial variability in precipitation (Willmott *et al.* 1994, Willmott and Legates 1991). Gauges are sparse and records are generally biased, underestimating precipitation by over 50%. Satellite data, notably passive microwave, when calibrated by snow survey information, may offer an instrument for regional surveillance of the Arctic snow cover (Ferraro *et al.* 1994, Robinson *et al.* 1993, Woo *et al.* 1995, Grody and Basist 1996). However, the introduction of new instrumentation for measuring precipitation may inadvertently affect monitoring efforts and obscure subtle changes. Snow cover water equivalent measurements from satellite are as yet unreliable (Armstrong and Brodzik 1995), but airborne gamma-measurements are promising (Carroll and Carroll 1993, Carroll 1995).

Direct measurements of snow atop Arctic sea ice are relatively scarce (Barry *et al.* 1993a). No reliable means are yet available for remote sensing of snow depth or snow liquid-water-equivalent over sea ice. However, snow extent can be inferred from changes in surface albedo. Snow retreats northward over the pack ice during June and July, and the ice is essentially snow-free by mid summer (Robinson *et al.*

1992). Substantial inter-annual variability in rates of snow melt has been observed (Scharfen *et al.* 1987). Direct observations of snow cover are too limited in space and time to document natural variability and trends.

Current general circulation models (GCM's) predict a warming of 2-5°C and increased precipitation for Arctic environments in response to global climate change (Schlesinger and Mitchel 1987, IPCC 1990a, 1996a), but predictions of future changes in summer and winter precipitation have large uncertainties. Most GCMs predict enhanced precipitation at high latitudes with an average increase of 20-25% over all seasons in North America (Mitchell *et al.* 1990, IPCC 1992a, Maxwell 1992).

11.2.2.5. **Hydroclimatology**

Climatic warming and the increased melt of stored water will affect the present hydrologic balance with the likely direct effect of raising the sea level globally. Although there are estimates of the general water balance of the Arctic, the detailed hydroclimatology of the Arctic drainage system is still insufficiently known (WWB 1974, Baumgartner and Reichel 1975, Ivanov 1990, Ivanov and Yankina 1991). There is a sharp decline in the number of basic meteorological stations as one progresses northward. Although discharge measurements (Global Runoff Data Center, Koblenz; Roshydromet, St. Petersburg) are relatively much better than those for climatic driving variables within the region, improvements in the basic hydrometeorological data are necessary before assessing any further changes.

11.2.2.6. **Sea ice**

The responsiveness of sea ice cover to thermodynamical and transport processes suggests that changes in sea ice extent, concentration, thickness, and transport are sensitive indicators of climate change (Hall 1988, Barry *et al.* 1993b). However, regional and inter-annual variability complicate the detection of a general climate-change signal. The extent of sea ice has been lower than average during the 1990s (Johannessen *et al.* 1995, IPCC 1996a), including reductions in Arctic ice extent larger than those observed in the remote-sensing record through 1979 (Maslanik *et al.* 1996). Total mass of sea ice is a critical variable for determining possible impacts of climate change on Arctic atmospheric and oceanic conditions. However, measurements of sea ice thickness are relatively rare and not suitably distributed in space and time to detect trends (McLaren *et al.* 1992). Characteristics such as ice age, salinity, and surface roughness provide some information about climatic conditions and interactions since growth and decay of sea ice influence the temperature, circulation, and moisture content of the overlying atmosphere, all of which, in turn, affect sea ice mass (Polar Group 1980, IPCC 1990b, Meehl and Washington 1990, Raymo *et al.* 1990).

Relatively accurate ice extent and concentration information can be determined from satellites (Parkinson *et al.* 1987, Thomas 1990, Maslanik and Barry 1990, Barry *et al.* 1993b, Weaver *et al.* 1987, Serreze *et al.* 1995). These data are being used alone or in combination with longer records of ice extent obtained from surface observations to detect and interpret possible trends or lack of trends (Parkinson and Cavalieri 1989, Mysak and Manak 1989, Gloersen and Campbell 1991, Parkinson 1992, Zwally 1995) and to identify links to climate forcing (Cavalieri and Parkinson 1981, Serreze *et al.* 1990, Mysak *et al.* 1990, Power and Mysak 1994, Chapman and Walsh 1993, Serreze *et al.* 1995). The negative anomalies in Arctic ice extent can be linked to changes in atmospheric circulation, in particular to an increase in cyclonic activity in the Eurasian sector of the Arctic (Maslanik *et al.* in press). The relatively short record of satellite observations and the large inter-annual variability in the sea ice cover must be taken into account when assessing connections to climate change (Zwally 1995).

Estimates of ice transport as measured by drifting buoys (Colony and Thorndike 1984, McLaren *et al.* 1987), observed from satellites (Emery *et al.* 1995), or simulated by models (Walsh *et al.* 1985) are used to estimate mass budgets and possible connections between ice transportation and changes in ocean salinity and circulation (Hakkinen 1993). No techniques are yet available to observe ice thickness or the rate of change of ice thickness over large areas and extended time periods (Table 11·1). Existing measurements of sea ice thickness, while valuable for describing general conditions, are insufficient for documenting changes over inter-annual or climatological time scales (McLaren *et al.* 1990).

Although areas of thinner ice have been recorded (Bourke and Garrett 1987, Bourke and McLaren 1992, McLaren 1989, Wadhams 1990), Walsh *et al.* (1995) found no appreciable trend toward thicker or thinner ice over the North Pole in the period 1958 to 1992. One problem is that sea ice measurements taken by upward looking sonar (Vinje *et al.* 1989) provide average ice draft information within a sonar beam, but variations of the thinner and more fractured areas that may be the most sensitive indicators of climate change are not recorded. In addition to sonar data, airborne laser profilers and electromagnetic induction sounding techniques have provided ice thickness estimates (Nagurny 1995).

Too little is currently known about the natural variability of Arctic sea ice thickness to draw any conclusions with regard to climate change. Models suggest that warming temperatures will lead to substantial changes in sea ice extent. Simulations using coupled ocean-atmospheric models show that Arctic ice thickness decreases considerably when the atmospheric CO_2 content is increased (Manabe *et al.* 1992). The postulated warming from CO_2 could have important consequences for the thickness and extent of Arctic sea ice and consequently for the ice exported through Fram Strait.

11.2.2.7. **Vegetation**

The Arctic troposphere is profoundly affected by the large areas of vegetation which contribute to the strong seasonal cycle of trace gases seen in the Arctic. In addition to cycling trace gases, vegetation affects surface albedo as well as soil temperature and moisture, thus playing an important role in the energy balance and hydrological cycle.

Table 11·1. Various techniques used to date to measure ice thickness.

Technique	Applicability	Platform	Reference
Surface drilling through ice	Local surveys	Surface	Ackley *et al.* 1976, Kovacs 1983, Koerner 1973
Upward-looking sonar from mobile platforms	Local, basin surveys	Submarine, ROVs	McLaren 1989, Walsh *et al.* 1995, Wadhams *et al.* 1991
Upward-looking sonar from moored buoy	Local	Moored array	Moritz 1992
Laser profiler	Basin surveys	Aircraft	Krabill 1992, Comiso *et al.* 1991, Wadhams *et al.* 1991
Electromagnetic impulse sounders	Local surveys	Helicopter	Kovacs 1983
Satellite estimation of thin ice types	Basin surveys	LANDSAT, ERS-1 SAR, Radarsat	Steffen and Heinrichs 1994, Steffen and Schweiger 1990

Vegetation can also provide an indirect record of climate changes. With warmer temperatures, the extent of boreal forest will probably expand and the treeline will move farther north into regions that are currently tundra (Sveinbjörnsson *et al.* 1992). Tree ring studies can give a record of response to climatic conditions. At the northern edge of boreal forests, trees exhibited a positive growth response during the warming of the 1930s and 1940s, but not during the recent period of warming (Jacoby and d'Arrigo 1995). However, because of the influence of microclimates, growth patterns are idiosyncratic, making the interpretation of tree-ring data difficult (Mason and Gerlach 1995).

11.2.2.8. Soils and permafrost

Concern over the impacts of climate change and altered UV on soils is related to emission of greenhouse gases. Nadelhoffer *et al.* (1992) highlighted the important differences in response of dry, moist, and wet Arctic soils to changes in air temperature and precipitation. However, accurate predictions of the effects these changes will have in soils are complicated by changes in vegetation and the additional effects of other changes such as enhanced CO_2, UV-B, and increased atmospheric pollution. Soil temperatures are critical to active layer waterlogging, aeration, and nutrient cycling. Consequently, soil temperature affects plant communities and the balance of CO_2 and CH_4 emission.

Permafrost is both an indirect indicator and an archive of climate change because increases in air temperature can thaw the soil. Permafrost terrains display evidence of previous warm periods, such as the formation of thermokarst and deepening of the active layer during the Holocene climatic optimum in the western Arctic (Burn and Smith 1990). One of the best examples of the impact of climate change comes from northern Alaska where Lachenbruch *et al.* (1988) have documented evidence of 2-4°C warming over the last 100 years based on an inversion in the upper part of several deep ground temperature profiles. Indigenous residents of northern Alaskan villages also report thawing of previously frozen ground. While thermokarst is a natural part of the evolution of permafrost landscapes and has been linked to forest fires rather than climate change (Burn 1990, Burn and Smith 1990, Harry and MacInnes 1988, Mackay 1995), more recent observations by Osterkamp (1994) have confirmed warming and thawing of discontinuous permafrost in Alaska. Lewkowicz (1992) has also linked the high number of recent active layer detachments on the Fosheim Peninsula of Ellesmere Island to unusually warm summer conditions.

11.2.2.9. Glaciers and ice sheets

Glacial history of the large Greenlandic ice sheet and of smaller glaciers throughout the Arctic provide both long- and short-term evidence of climate change. Historical records and photographs often include the size of glaciers; these can be used to infer changes. Because the length of a glacier is tied to snowfall, which in turn is tied to temperature and other factors, moraine positions provide evidence of warming and cooling episodes. Indications of global warming from glacier length are consistent with, but independent of, other records of global warming during this century (Oerlemans 1994).

Increased temperatures can cause glaciers to melt, but can also cause increased precipitation. The balance of these two processes determines the net storage of water in glaciers and, thus, whether sea levels will rise. Over the last 100 years, glaciers have receded globally, contributing to part of the concurrent sea level rise of about 15-30 cm (Revelle 1983). Measuring changes of ice sheets in Antarctica and Greenland by conventional surveying methods is beyond present means. Satellite mounted radar altimeters can measure changes in surface elevation of the ice sheets. The measurements are not very precise, however, and await the launch of a laser altimeter on future satellite missions such as EOS (Earth Observing System). Present indications show that the mass balances of Greenland and possibly Antarctica are positive, i.e. they are growing, not shrinking (Zwally 1989).

11.2.2.10. Ice and sediment cores

The reconstruction of past climatic conditions from the sedimentary records of ice caps, lakes, ponds, and peats is an important element of climate change research. These records offer time series that are typically one to three orders of magnitude longer in duration than the instrumented climate record, thus affording unique perspectives on long-term natural variability. This context is invaluable when attempting to discern natural from anthropogenic climate forcing (IPCC 1996a). Paleoclimatic data are useful in validating how accurately climate models simulate climatic forcing; they provide a testing ground for how well the same models are likely to perform in predicting future conditions (Overpeck *et al.* 1991). It is generally recognized, however, that the forcing mechanisms of anthropogenic climate change, and perhaps the rates of envisaged change, have no direct geological analogs (Overpeck 1995, IPCC 1996b). Because an understanding of the past is critical to predicting the future, and because the historical perspective is a robust means to decipher processes occurring on time scales from decades to millennia, paleoclimatic studies are extremely useful, particularly in the northern high latitudes.

11.2.2.10.1. Ice cores

Ice cores have been collected from several locations in the Arctic, providing reasonable geographic coverage for Greenland and the Canadian Archipelago (e.g., Körner and Fisher 1990, Grootes *et al.* 1993). Ice cores are unequalled archives of atmospheric gases and particulates. However, they are expensive to collect and analyze, and are restricted to stable ice masses. Paleotemperature indicators from ice cores include stable isotopes of oxygen, borehole temperature profiles, and melt-layer frequencies. This information is used to derive paleo-temperatures, as well as the thicknesses of annual accumulation corrected for ice flow, which are used to derive past snow accumulation (Figure 11·11) (Alley *et al.* 1993, Alley and Anandakrishnan 1995, Cuffey *et al.* 1995). Complementary indicators include aeolian sea salt, dust, biogenic sulfonates from the ocean, particulates from boreal forest fires, volcanic ash, and geochemical proxies of solar activity (e.g., Mayewski *et al.* 1993, Stuiver *et al.* 1995, Taylor *et al.* 1996, Saltzman *et al.* in press).

Although ice core records offer outstanding resolution for the last glaciation and the Holocene, ice compaction and flow compromises the record of the previous interglacial period (Grootes *et al.* 1993). Air bubbles in ice cores are a source of information on paleo-chemistry of atmospheres. Analysis of the trapped air reveals gradual atmospheric changes on the glacial-interglacial time scale that are linked to variations in Earth's orbital parameters. More than 20°C of warming, for example, are inferred since the coldest portion of the last glacial period (Figure 11·11) (Cuffey *et al.* 1995). A greater contribution of the ice core record, perhaps, has been toward understanding the character and rates of abrupt climate

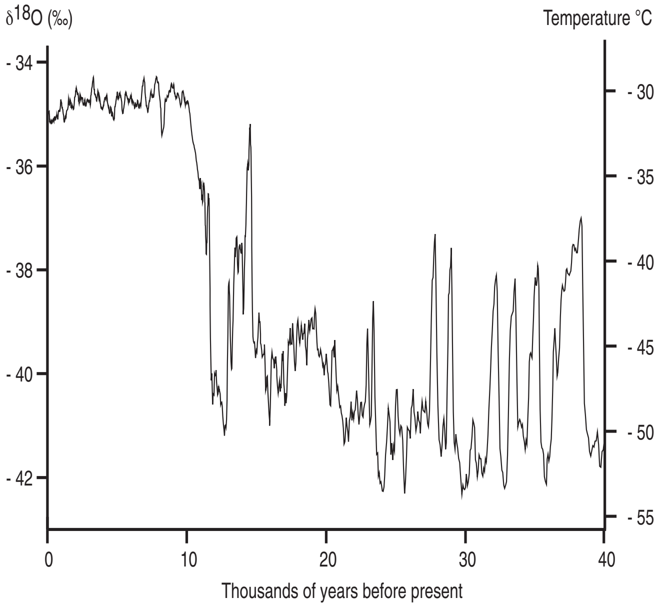


Figure 11·11. History of temperature changes in central Greenland over the last 40 000 years, from the GISP2 core, showing very large and probably abrupt changes (after Cuffey *et al.* 1995). The stable-isotopic ratios of ice were calibrated, as shown on the vertical scales, using borehole temperatures in the ice sheet. The record shows a warming of more than 20°C since the coldest time of the last ice age, and jumps of many °C over decades.

changes, unrelated to external (orbital) forcing, that occurred during the last glaciation and toward its termination. Examples include the documentation of Heinrich/Bond events, with roughly 7000 year periodicity, that are related to surges of the Laurentide Ice Sheet from Hudson Bay (Bond *et al.* 1993), as well as the millennial Dansgaard/ Oeschger oscillations linked to changes in oceanic thermohaline circulation (Broecker *et al.* 1990). Further, ice cores have been instrumental in demonstrating that 1) abrupt climate changes may occur on time scales relevant to humans (Figure 11·12) (Dansgaard *et al.* 1989, Alley *et al.* 1993); 2) large changes are not restricted to ice ages, but may equally affect the relatively warm interglacial climate state (Dansgaard *et al.* 1993, O'Brien *et al.* 1995); and 3) there are strong global teleconnections of several discrete paleoclimatic events (Bender *et al.* 1994, Brook *et al.* 1996).

11.2.2.10.2. **Paleoecological records**

Unlike in ice cores, the climate signal is not recorded directly in the paleoecological record. Instead, plant and animal fossils preserved in sedimentary deposits provide a record of past climatic conditions. The prevalence of lakes and ponds throughout much of the Arctic enables the development of relatively dense networks of coring sites. Over broad geographic areas, the time-transgressive nature of climatic and ecological changes may be determined. This is particularly

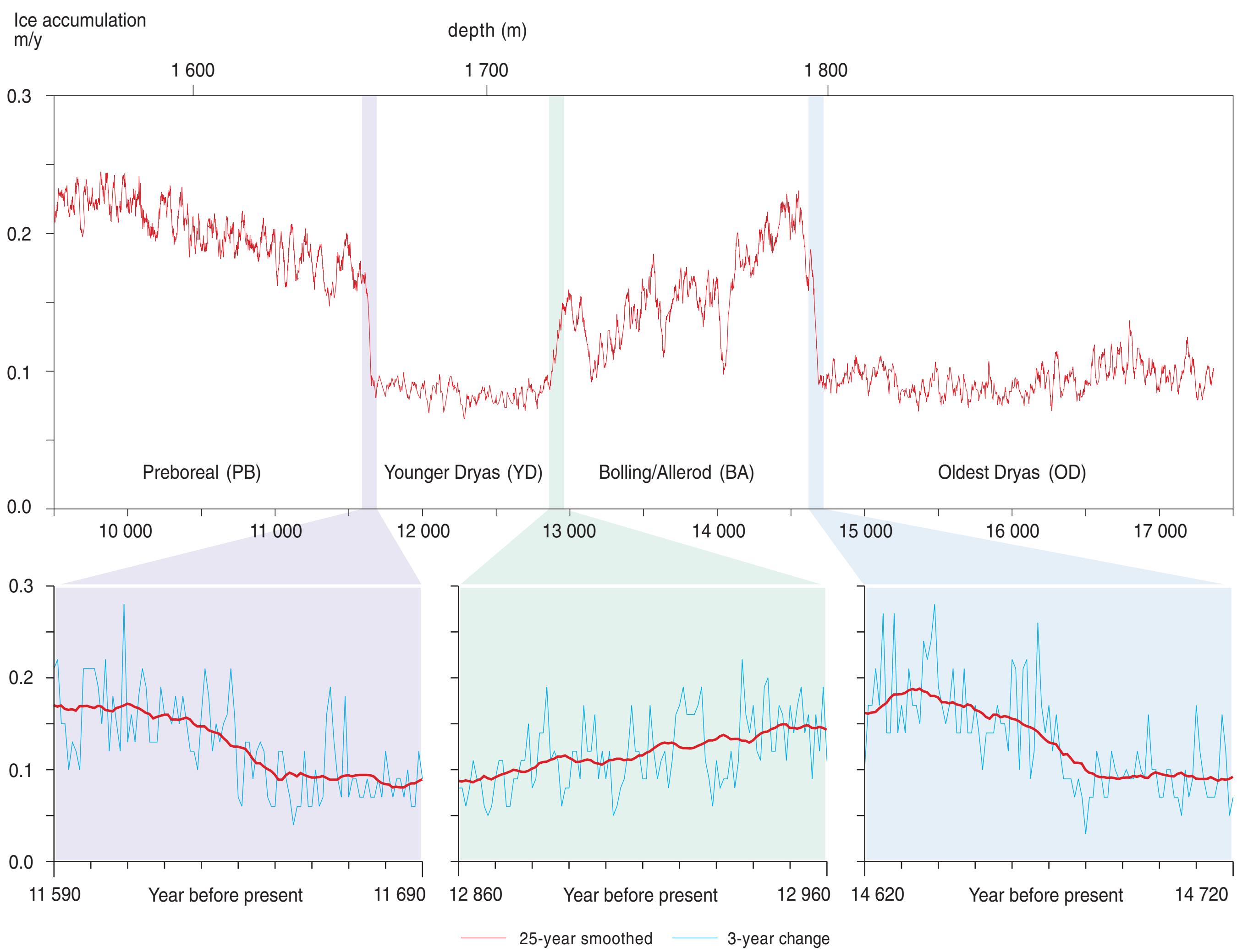


Figure 11·12. History of snow accumulation in central Greenland during the most recent glaciation, as measured in the GISP2 core (after Alley *et al.* 1993). The 25-year running mean, represented in all panels, shows very large and rapid climate changes. Shorter-term variability is shown in the lower panels. Age is relative to present = 1950. Accumulation doubled at the end of the Younger Dryas cold event, as shown in the middle-lower panel, and a somewhat larger change occurred almost as rapidly at the onset of the Bolling warm event.

relevant in the Arctic where regional variations in the northward transport of heat by the atmosphere and oceans cause spatial heterogeneity and strong latitudinal gradients in climate. Palynology is the paleoecological technique that has been most frequently applied to climate reconstructions (Bartlein *et al.* 1986, Prentice *et al.* 1991). However, palynology has some limitations in many Arctic regions (e.g., the High Arctic), owing to the low pollen production rates of many local tundra plants, and the related over-representation of exotic pollen types of uncertain, distant provenances (Gajewski *et al.* 1995). Plant macrofossils (seeds, leaves, buds) extracted from lake and peat cores have the advantage of testifying with certainty the past presence of plant taxa. Unfortunately, rich macrofossil sites are not common in the Arctic, so that species representation may be selective and the record may be more strongly influenced by local edaphic factors than by climate. For sites with well preserved and diverse plant macrofossils, paleoclimatic inferences may be made with reasonable confidence (Birks 1991), especially when coupled with co-occurrent fossil groups, such as insects (Elias *et al.* 1996).

The siliceous fossils of freshwater diatoms are ubiquitous in the sediments of both lakes and ponds, and appear to respond rapidly and sensitively to climatically controlled limnological changes (Smol *et al.* 1991, 1995, Pienitz and Smol 1993, Pienitz *et al.* 1995). Although recent advances have been made in Arctic freshwater diatom ecology (Douglas and Smol 1995, Wolfe 1996a), many regions remain understudied. The rigorous application of diatom-based paleolimnology to the study of climate change in the Arctic is therefore still in its infancy, although progress is being made. For example, Douglas and Smol (1994) have documented dramatic diatom stratigraphic changes, believed to be climatically induced, in the sediments deposited over the last 200 years in ponds on Ellesmere Island. These results, and others in progress, clearly demonstrate the strong potential of paleolimnological approaches toward addressing questions of climate change in Arctic regions.

One of the most significant advances in paleoecology has been the improvement of sediment dating control through direct ^{14}C measurements by accelerator mass spectrometry (Andrée *et al.* 1986). Very small samples of plant remains, as well as selectively extracted organic compounds (e.g., humic acid) may now be dated with confidence to about 40 000 years. Most Arctic lakes are typified by low sediment accumulation rates relative to their temperate counterparts, so that the ability to date specific levels, instead of increments of core several centimeters thick, becomes especially relevant (e.g., Snyder *et al.* 1994).

Due to the complex glacial history of the Arctic, most continental sediment records are limited to the Holocene (the last 10 000 years), although longer records do exist in certain regions (Wolfe 1994, Elias *et al.* 1996). This in no way detracts from the utility of these records for paleoclimate research, since both warm (Overpeck 1996) and cold (Maslin and Tzedakis 1996) climate 'surprises' are thought to punctuate the interglacial climate state, at rates relevant to humans.

11.2.2.11. **Historical and archaeological evidence**

Documentary historical records can provide information about past variations in climate, the impact of climate on past societies, human responses to climatic stress, and human perceptions of climate and climate change (Ingram *et al.* 1981). In the Arctic, historical records have been used to reconstruct past climate in, for example, Iceland (Ogilvie 1991, 1992, 1996) and the Hudson Bay area (Catchpole 1992). Additional investigations could yield climate reconstructions for other areas such as Arctic Russia and Siberia.

Traditional ecological knowledge and especially oral history provide a further source of climate information from the recent past. Elders, as oral historians, often have good memories of environmental changes that have occurred since their childhood and may know stories related to them by an earlier generation. Potentially useful sources of climate information might come from traditional knowledge and oral history studies, yet little work has been done in this area.

Archaeologists are frequently able to integrate a variety of research methods to provide a long-term perspective on human-resource-climate interactions. Northern archaeology has greatly expanded its field of interest to include zooarchaeology, paleobotany, human paleobiology and new methods of relative and chronometric dating (McGovern *et al.* 1988, Mason and Gerlach 1995). An interesting recent development is a desire among northern archaeologists, historians, and others working in related fields to combine their knowledge and efforts to increase understanding of the relationship between humans and climate change. An example of this is the interdisciplinary work currently being undertaken on the Norse settlements in Greenland (Buckland *et al.* 1996, Barlow *et al.* 1997).

11.2.3. **Ability to predict**

Computer models based on physical principles help understand the Earth's climate and assess the prospect of climate change. A variety of classes of models are in use today, enabling the study of various components of the climate. Coupled General Circulation Models (CGCMs) are the most sophisticated of these, consisting of sub-models of the atmosphere, ocean, cryosphere, and land surface. Assessment of climate change requires consideration of each component of the climate system, and CGCMs are therefore the most powerful tools available for this purpose. CGCMs have developed rapidly in the 1990s (IPCC 1996a), but there remain problems that must be resolved before they can be used to confidently project regional climate change. Simulation of the Arctic climate is particularly challenging, in part because of the extreme sensitivity of the models to sea ice.

In recent years climate modelers have begun to work more closely by comparing the results of their respective models, a process that helps to identify systematic model errors (cf. Walsh *et al.* 1995, Tao *et al.* 1996). Over the Arctic Ocean, the mean bias of 19 models simulating surface air temperature compared with observations was found to be small (<1°C), except in spring when it was +3°C. This springtime bias may be sufficient to prematurely break-up the Arctic ice in a coupled model if the bias persists over decades. The precipitation predicted by every CGCM studied was larger than observed (Vowinckel and Orvig 1970, WWB 1974, Elliot *et al.* 1991, Ross and Elliott in press), particularly during the winter months, when it was more than twice what is observed. Therefore, empirically derived climatologies contain large uncertainties. For example, although evaporation in winter was found to be excessive, in summer the simulated evaporative flux did not differ substantially from observations. The freshwater flux (precipitation minus evaporation) was approximately twice that observed, suggesting that the models' input of freshwater to the Arctic Ocean was too large. Walsh *et al.* (1995) point out that this may have important implications for the stratification, stability, and dynamics of the Arctic Ocean in coupled model simulations. The simulated total cloud cover in the Arctic was found to vary tremendously from one model to the next. During the summer, for example, the cloud cover was found to vary from 30% to more than 90%. Finally, in most models the snow depth of Greenland was not in equilibrium, because in each successive year of a 10-year simulation there was more snowfall than the sum of evaporation and melting.

From the studies of Walsh *et al.* (1995) and Tao *et al.* (1996) it is clear that state-of-the-art atmospheric models do not adequately simulate the present Arctic climate. Improvements in models will be necessary before they can produce credible predictions regarding future temperatures, circulation patterns, or consequent ozone concentrations. Efforts are underway to more closely study CGCM simulations and improve the understanding of coupled model simulations. Further development of atmosphere, ocean, cryosphere, and land surface models are all high priority efforts in the climate community, as it is clear that advancements in each are vital to the development of the coupled climate models used to understand the prospects of climate change in the polar regions.

11.2.4. Components of the Arctic

The Arctic climate system is one of the most complex and dynamic climate systems on the earth. The strong interactions between various components make understanding the Arctic, and therefore predicting future changes, extremely complex. Each component is dynamically tied to the other components through energy, water, and trace gases exchange.

11.2.4.1. Oceanic regime

The Arctic Ocean influences global climate change through its effects on surface heat balance and thermohaline circulation in the Arctic Basin (Aagaard and Carmack 1994). Both processes are closely linked to the salinity structure and sea ice cover of the Arctic Ocean. The atmosphere-ocean exchange of energy and matter, which is strongly modified by the presence or absence of sea ice, controls the flow of heat and water vapor from low to high latitude regions. In addition, the presence of sea ice and its associated snow cover significantly affects the surface albedo and radiation budget. Any discussion of climate change must include consideration of salinity (cf. Rooth 1982), sea ice (cf. Lange *et al.* 1990), and the Arctic Ocean circulation (cf. Stigebrandt 1981, Aagaard and Carmack 1994).

11.2.4.1.1. Ocean stratification and water circulation

Water tends to evaporate in regions where temperatures are high and condense in regions where temperatures are low. For this simple reason, the cold polar oceans maintain freshened surface waters floating over more saline deep waters. Through brine rejection, freezing serves to further separate salt and fresh water, acting as the high-latitude counterpart to evaporation (Aagaard and Carmack 1989, Macdonald and Carmack 1991). This basic stratification acts to inhibit thermally forced convection and thus is a dominant factor controlling ice cover, surface albedo, and material transport in high-latitude oceans. Changes in freshwater flux to the oceans associated with melting ice and other environmental changes could trigger large and abrupt climatic changes (Broecker 1994, Alley 1995). Sea level rise of 1-2 mm per year in restricted and sensitive regions of the North Atlantic from melting of land ice is nearly as big as the freshwater fluxes modeled to cause major reductions of deepwater formation (Rahmstorf 1995). Reductions in deep water formation in the North Atlantic would have a direct impact on climates throughout the rest of the world.

The principal waters entering the Arctic Ocean are the relatively warm and saline waters from the Atlantic via Fram Strait and the Barents Sea and the relatively fresh waters from the North Pacific via Bering Strait. These waters flow in a counterclockwise direction around the four major basins where they are subsequently modified by air/sea/ice interactions, river inflow, and exchange with surrounding shelf regions. The large volume of relatively fresh water stored in the upper 200-300 m of the Arctic Ocean reflects inputs from river inflow, sea-ice melt, net precipitation, and the inflow of Pacific water (Aagaard and Carmack 1989, Treshnikov 1985, Macdonald and Bewers 1996).

Underneath the surface mixed layer is a cold isothermal layer with markedly increasing salinity called the Arctic halocline. This halocline layer overlies warm Atlantic water, and inhibits convection, which would otherwise release heat from the Atlantic layer and hinder ice formation in the Arctic (Aagaard *et al.* 1981, Melling and Lewis 1982, Wallace *et al.* 1987). The halocline is weaker in the eastern Arctic than in the western Arctic which is more strongly affected by low-salinity Pacific waters. Thickness and horizontal extent of the halocline layer also affects the areal extent of sea-ice cover, and thus has direct implications for climate. Recent observations suggest that the boundary between these two regimes may move across ridge systems in the vicinity of the Lomonosov Ridge (McLaughlin *et al.* 1996).

Salinity distributions dominate the density structure within the Arctic Ocean, and thus determine the large-scale thermohaline flow (Aagaard *et al.* 1985, Coachman and Aagaard 1974, Semtner 1987, Hakkinen 1993). Pathways of flow are created by salinity-dominated buoyancy fluxes (both positive and negative) around the basin perimeter (Gawarkiewicz and Chapman 1995). Fresh water components, such as brackish seawater and ice, exit the Arctic Ocean through Fram and Davis straits into the North Atlantic, where they may supply buoyancy to the upper ocean, slow the rate of deepwater formation, and thus impact the global thermohaline circulation by affecting rates of water mass transformation in the Greenland, Iceland, and Labrador Seas (Dickson *et al.* 1988, Aagaard and Carmack 1989). The thermohaline circulation is, in turn, responsible for as much as half of the Earth's poleward heat transport, affecting all aspects of global climate (Broecker *et al.* 1985a).

The formation of deep water in the North Atlantic, and more specifically in the basins of the Greenland, Icelandic, and Norwegian Seas (Nordic Seas), is a major component of global oceanic circulation. This is one of only three places in the world's oceans where an exchange between warm, less-saline water and deeper, more-saline water takes place. Thus, the existence and strength of this exchange has major consequences for the world ocean and the marine biosphere, as deep water formation enables entrainment and transport of oxygen and nutrients to deeper parts of the world ocean. The traces of this entrainment can be seen well beyond the equatorial region of the Atlantic.

The importance of deep water formation and related processes in the Nordic Seas goes beyond the oceanic regime. It also controls the strength and geometry of thermohaline circulation in the North Atlantic and basically controls the amount of warm water exported by the Gulf Stream to northern Europe and into the Barents Sea. Thus, climate conditions in northern Europe depend on and may be changed by alterations in deep water formation in the North Atlantic.

The convective processes in the Nordic Seas occur in different stable configurations, each of which is accompanied by significantly different climatic conditions in northern Europe and the European Arctic. If, however, the condition of the warm water is changed, e.g. through a lesser salt content or even higher water temperature (an expected consequence of climatic warming), a major part of the convective system in the Nordic Seas breaks down. This leads to a decrease in the magnitude of thermohaline convection, i.e. a decrease in warm water flow to the North Atlantic, and a possible enlargement of the southward extent of sea ice. The balance leading to one or the other of the stable conditions is extremely fragile and can be considered as a 'switch' for northern European climate. It is hypothesized that alterations in this balance in the past have led to short-term changes in regional climate such as that during the Younger Dryas.

11.2.4.1.2. Sea ice

Sea ice is one of the defining properties of the Arctic oceans; it affects the biosphere, human populations, the energy balance, and water vapor exchange between the ocean and the atmosphere. The Arctic Ocean sea ice cover is a mass of different amounts and concentrations of ice of various ages, surface characteristics, and properties. It extends over about 15×10^6 km^2 of the ocean in winter but shrinks to 8×10^6 km^2 in summer (Gloersen *et al.* 1992). Regional variations in open water within the sea ice cover contribute significantly to modifying ocean salinity (Martin and Cavalieri 1989) and heat and moisture fluxes into the atmosphere (LeDrew *et al.* 1991).

One of the major characteristics of sea ice in the polar seas is its seasonal variation in thickness and extent. The annual cycle of sea ice growth and decay is a driver of temperature and salinity variations in the upper layers of the Arctic Ocean. The structure of sea ice largely reflects its evolution. The presence of brine creates an ice microstructure which is very sensitive to temperature changes, resulting in large seasonal variations within the ice that directly affect such properties as strength, thermal conductivity, albedo, microwave emissivity and optical extinction (Weeks and Ackley 1982). For example, in midwinter leads, the brine released downward from rapidly growing thin ice induces and intensifies thermohaline circulation in the mixed layer (Carmack 1986); brine rejected upward toward the surface of the ice affects its brightness, temperature, and other microwave properties.

Sea ice modulates the energy exchange between the ocean below and atmosphere above, effectively insulating the ocean from the atmosphere. Sea ice mass balance is determined by calculating ice extent, concentration, thickness, and motion over time (Thorndike *et al.* 1992). The mass of sea ice contributes most of the fresh water supplied to the Greenland Sea and thereby impacts deep convection, bottom water formation, and global thermohaline circulation. Climatic warming could lead to a decrease in both the thickness and the areal extent of sea ice, significantly affecting ocean heating and the ocean-atmosphere heat flux at high latitudes.

Changes in climate variables other than temperature can also directly act upon sea ice. Drag forces exerted by surface wind on the sea ice act upon atmosphere-sea ice interactions. Surface winds are mostly responsible for the smaller scale motion of the sea ice cover, while large scale geostrophic circulation patterns account for about half of the large scale motion of sea ice; the other half is influenced by ocean circulation (Thorndike and Colony 1982). The major drift patterns of Arctic sea ice and its mean velocities are characterized by a linear component (the Transpolar Drift) and a circular movement pattern (the Beaufort Gyre). On time scales of days, the winds and their changes associated with synoptic systems can significantly advect pack ice (e.g., Dey 1980) at roughly 2% of the surface wind speed, produce coastal polynyas (Pease 1987), and affect the organization and fractional coverage of leads (e.g., Walter and Overland 1993).

Leads and polynyas

Leads and polynyas exert a prevalent and immediate effect on the surface energy balance as well as the hydrological cycle. They are often locations of plankton blooms and feeding areas for marine mammals. Inuit continue to use both leads and polynyas for hunting marine mammals as they have for the past three millennia (Schledermann 1980).

Leads are long, roughly linear openings, or breaks, in the pack ice caused by variations in the wind stress. They range in width from a few meters to a few kilometers and in length from a few kilometers to hundreds of kilometers. In the central Arctic the areas of open water and thin ice, or the effective lead fraction, averages about 3% in the winter and spring, rising to 6% in the summer (Lindsay and Rothrock 1994). The effective lead fraction in the peripheral seas was found to be between 5% and 10% in the winter, while in the summer the peripheral seas are often ice free. In the summer, open water is much more prevalent, occupying as much as 10% of the area in the interior of the pack, although there is little or no temperature differential between the ice and the open water.

The effect of leads on the surface energy balance occurs due to the large difference in albedo between thick ice and open water or thin ice in leads, and to the large contrasts possible between lead surface temperatures and the overlying air. In winter, open water has a temperature of –1.8°C, as much as 40°C or more warmer than the surrounding ice and atmosphere. The sensible heat flux from the open water is high, and the open water in a lead often freezes within hours of the lead's formation. In the winter, the sensible heat flux changes sign, from roughly 10 or 20 W/m^2 downward over thick ice to as much as 500 W/m^2 upward over open water with high winds (Andreas and Murphy 1986). The sensible heat flux from leads is sometimes enough to change the sign of the regional-average sensible heat flux from a net downward flux to a net upward flux. In a warming climate this can have important ramifications in the structure of the atmospheric boundary layer, which becomes unstably stratified instead of being stably stratified (Serreze *et al.* 1992).

Cloud plumes originating from leads are often observed in satellite images. Leads are a vast source of moisture for the atmosphere. While they may not provide enough to increase snowfall, they can significantly increase cloud cover. An increase in effective lead area from climatic warming is likely to increase cloud cover in the Arctic.

Polynyas are larger regions of thin ice or open water that occur repeatedly in similar locations throughout the polar winter. Changes in air temperature will impact their size and location and the magnitude of their influence on ocean circulation, ice formation, and the energy balance. Polynyas are maintained by divergence in the ice drift or melt from oceanic heat flux, and they range in size from a few hundred square meters to hundreds of square kilometers (Smith *et al.* 1990). Common locations are on the lee sides of islands or peninsulas or at the edge of fast ice. In both cases, wind blows the ice away from the open-water area, maintaining the polynya. The near-surface air temperature is much warmer over the polynya than over the surrounding ice, yet much colder than the surface water temperature. Thus polynyas will drive a strong sensible heat flux from the water that can amount to a few hundred W/m^2. Polynyas are large sources of brine rejection caused by the high rates of ice production. This brine contributes to water-mass formation and ocean-current dynamics.

Snow cover

Snow is an effective insulator that controls the rate of heat conduction from the base to the surface of the ice and thus affects ice growth rates (Maykut 1978). An increase in precipitation, and thus snow cover, impacts not only ice growth but also the surface energy balance. Fresh snow on sea ice increases surface albedo and further reduces any energy transfer between ocean and atmosphere due to its low thermal conductivity. In addition, snow may become part of the sea ice cover through the formation of snow-ice, thus influencing the mass balance of the sea ice cover (Lange *et al.* 1990). The distribution of snow depth on sea ice is affected chiefly by sea ice topography, which is in turn a function of ice age and ice motion (Buzuev and Dubovtsev 1978). Snow depth measurements from the drifting ice stations of the Former Soviet Union indicate that the maximum snow depth on thick ice is in mid-May and averages 0.35 m. The snow usually, but not always, disappears in July and August. Snow cover is a reservoir of fresh water and plays a key role in melt pond formation.

Melt ponds

The seasonal variation of the sea ice surface directly affects the radiation balance and is one of the key elements of the ice-albedo feedback mechanism. A substantial portion of the sea ice is covered by dark melt ponds in the summer. This lowers the albedo but does not substantially change the surface temperature. The area of melt ponds reaches a maximum of 50-60% in early summer, and then decreases to a minimum of about 10% at the end of summer, causing a strong seasonal contrast in the average surface albedo (Maykut 1986, Grenfell and Maykut 1977, Grenfell and Perovich 1984, Perovich 1996). As the area of melt ponds increases, the area of lower albedo increases which leads to an increase in solar radiation absorption at the surface. However, later in the summer as the ponds warm and increase in size and depth, they drain through the base and/or margins of the ice floes; consequently, total pond coverage decreases, increasing albedo. Numerical, thermodynamic ice-growth models show that the simulated ice thickness is sensitive to the fraction of meltwater that runs off floes and to the melt pond coverage (Ebert and Curry 1993).

11.2.4.2. **Terrestrial regime**

The vast land masses of the Northern Hemisphere provide the Arctic climate system with a complex set of variables that make climate change in the Arctic difficult to predict or model. The land masses contribute warmer air to the polar region, disrupting the polar vortex. Thus, the Arctic's atmospheric dynamics and processes of climate change are distinctly different than those in the Antarctic. Soil and permafrost dynamics affect trace gas fluxes, surface temperatures, and the hydrological balance. Precipitation and runoff influence ocean temperature, salinity, and circulation. Vegetation types and snow cover affect albedo and thus surface radiation balance. Under a warming climate the terrestrial surface plays an important role in trace gas feedback effects. Vegetation along with soil moisture and temperature affect trace gas fluxes and thus cause large seasonal differences in chemical atmospheric dynamics. Historically, terrestrial ecosystems have stored CO_2, helping to cool the Earth. Glaciers and ice sheets have a large influence on sea level and on ocean circulation patterns. All of these processes will be affected by changes in climate. Changes to the Arctic have the potential to affect global climate through the response of the Arctic trace gas and hydrologic systems.

11.2.4.2.1. **Soil**

The soils of Arctic regions are particularly vulnerable to changes in climate because of the dominant role of temperature on their physical properties and biological processes. Arctic soils are highly varied in pH balance, particle size, and moisture content. Because of the differing soil properties, the effect of climate change varies considerably; while some responses will be large and rapid, others will be barely detectable. Depth of the active layer and the associated temperature and moisture regimes are the primary controls of processes of decomposition which determine the rate and amount of organic matter accumulation and trace gas flux from soils. An increase in the depth of thaw and the length of the active season from warming temperatures will be negatively related to the moisture content (positively correlated to the soil moisture deficit) (Nadelhoffer *et al.* 1992). Changes in moisture content are hard to predict because changes in precipitation are uncertain and soil moisture is also affected by snow melt, drainage, and evapotranspiration. Moist and wet soils may become drier as a result of increased evapotranspiration and lowering of the water table.

Soil moisture, to a large extent, determines the temperature profile, depth of permafrost, and soil organic matter content. Thermal conductivity, which is strongly regulated by the moisture content of the soil, is an important element of soil/atmosphere temperature dynamics. Water has a lower thermal conductivity than air and organic soils have a thermal conductivity almost an order of magnitude lower than that of frozen, nearly saturated mineral soils (Kane *et al.* 1992). Dry soils with cushion, heath, and lichen vegetation and thin organic mats tend to occur on slopes or well drained stony areas in the High Arctic. They have a relatively deep thaw or active layer because of limited insulation by the organic layer and their high thermal conductivity. Moist soils constitute much of the patterned ground and tussock tundra, with low relief and moderate depths of organic matter, which are typical of much of the low Arctic. The permafrost is thinner in these soils because of insulation by the surface organic layer and high water content. Wet and waterlogged soils dominated by sedges and mosses occur in poorly drained areas of the low Arctic. The thick organic layers which have accumulated because of the waterlogged and anaerobic conditions insulate the soil and help to retain moisture, thus reducing thermal diffusion and accounting for the shallow active layer.

11.2.4.2.2. **Permafrost**

Permafrost is 'ground that remains at or below 0°C for at least 2 years' (Permafrost Subcommittee 1988). It occurs extensively in Arctic and subarctic regions, affecting up to 80% of Alaska and 50% of Canada and Russia. Permafrost is divided into continuous and discontinuous zones defined by location and areal extent of cryotic ground. Depths range from > 1000 m (1450 m in Siberia) to only a few meters near the southern limit (Heginbottom *et al.* 1993, Smith 1993, Williams and Smith 1989). During the summer, the surface of the ground thaws. The depth of this active layer ranges from a few decimeters in the High Arctic to more than 2 meters in parts of the discontinuous zone. The seasonal freezing and thawing of the active layer and the seasonal pattern of temperature change in the upper part of permafrost produce distinctive features unique to Arctic tundra, such as patterned ground, gelifluction lobes, active layer detachments, seasonal frost mounds, and frost cracks.

Permafrost reflects a thermodynamic balance between ground surface temperature and the geothermal gradient. The spatial distribution and depth of permafrost are closely related to climate and are expected to change with climate changes. Much extant permafrost is close to 0°C and inherently unstable. Thawing could cause the degradation of existing discontinuous permafrost, a decrease in the distribution of continuous permafrost, and potential terrain instability (Smith 1993). There are serious concerns that there will be increased erosion, mass wasting, disruption of surface vegetation, and changes in surface drainage systems including drainage of critical wetlands (Fitzharris 1996). Permafrost degradation due to climate warming is complicated by its effect on vegetation and potential increases in trace gas fluxes (Nisbet 1989); these changes could then create conditions that would further affect climate.

Locally, the nature of the ground surface, including slope angle and aspect, surface water, vegetation, and snow cover, creates boundary layer conditions that determine the degree to which air temperature controls ground thermal regimes. Geologic, tectonic, and subsurface hydrologic conditions further influence permafrost.

The freezing of water in the ground produces an assortment of ground ice forms ranging from disseminated ice crystals in a soil matrix (pore ice) to thick (10-20 m), horizontally layered bodies of nearly pure ice which extend for several square kilometers. The type and rate of permafrost formation involve a complex set of processes including soil particle size, temperature, water content and chemistry, and water transfer processes and rates. Ground ice content sometimes exceeds the saturated moisture content of its host sediments. When permafrost containing excess ice thaws, the ground subsides in proportion to the volume of excess ice, forming thermokarst.

Of central importance in the investigation of ground ice and climate change are the distribution, nature, and origin of ice-rich permafrost and massive ground ice (e.g. Harry *et al.* 1988, Lawson 1983, Mackay 1992, Pollard and French 1980, Pollard and Dallimore 1988). Few investigations have been made of massive ground ice in the Arctic Archipelago (French *et al.* 1986, Lorrain and Demeur 1985, Pollard 1991, Barry and Pollard 1992), but considerable literature does exist for ground ice occurring in northern Russia. Of particular interest are studies on buried ice (Astakhov and Iasayev 1988), the gas content of ground ice (Arkhangelov and Novgorodova 1991), and ice petrography (Solomatin 1986).

11.2.4.2.3. **Runoff**

The potential impacts on Arctic terrestrial water balance of elevated temperatures associated with the greenhouse effect occur through the complex interactions of land surface hydrology and the freeze/thaw cycle. Winter freezing of the active layer of permafrost as well as shallow groundwater, lakes, and rivers inhibits the movement of water. During the early thaw period, melting of snow above the frozen surface entrains and accelerates surface run-off of both water and constituents, with overland flow (not interflow or baseflow) serving as the dominant input into river systems (Kane and Hinzman 1988).

Woo and Steer (1983) show that overland flow predominates in sites with a shallow active layer, while infiltration prevails in those with greater thaw depths. Water balance dynamics, specifically run-off generation, evapotranspiration, and changes in soil moisture, are intimately connected to permafrost dynamics. The most important factor controlling evapotranspiration in the Arctic has been reported to be soil moisture in the active layer (Kane and Hinzman 1988). Increase in active layer depth would simultaneously increase infiltration, soil moisture, evapotranspiration, and the lateral redistribution of water through enhanced groundwater flow. Woo and Marsh (1990) found that groundwater inflow into fen environments was also critical in maintaining soil moisture during the summer. Rainfall run-off studies in the Arctic suggest that drainage basins underlain by permafrost maintain ‘flashier’ stream hydrographs compared with non-permafrost areas (Haugen *et al.* 1982).

Because meltwater is stored in the snowpacks, on the basin slopes, and within channels, the initiation of streamflow often lags behind the commencement of snowmelt by several days or more. Once the flow begins, it usually exhibits a marked diurnal rhythm, reflecting the daily snowmelt cycle. In some instances, slush flow along a valley may convey water and snow rapidly downstream (Barsch *et al.* 1993). Annual peak flows are common during this period (Kane *et al.* 1992) and many rivers are prone to flooding. After the snow is depleted in the basin, streamflow subsides except during some heavy rain events. Warming temperatures and ensuing thinner ice, increased precipitation, and earlier spring melt could alter run-off dynamics and mean residence times. The situation is far from straightforward, however, as increases in precipitation, also predicted by numerous GCM’s for the region between 50-70°N (IPCC 1990a, Boer *et al.* 1992a, 1992b), could enhance the formation of saturation overland flow.

11.2.4.2.4. **Snow**

Because both the deposition and duration of snow on the ground during all seasons are associated with atmospheric dynamics and thermodynamics, and high latitudes are believed to be especially sensitive to global warming due to cryospheric feedbacks, snow cover fluctuations can potentially be used as an indicator of global climate change. Snow cover changes the surface albedo, regulates surface temperature, affects air mass formation, and insulates the ground and vegetation from severe cold temperatures. The snow cover-albedo feedback amplifies any climatic changes and impacts global climate variability (Cess *et al.* 1991, Randall *et al.* 1994, Groisman *et al.* 1994a, 1994b). The huge difference in albedo between snow cover and vegetation significantly impacts the energy balance. The albedo of fresh snow may exceed 0.80 and be as much as 4-5 times the value for open water or tundra (Kondratyev 1969, Grenfell and Warren 1994). Inter-annual fluctuations of snow cover over the Northern Hemisphere, though much smaller than seasonal ones, are sufficiently large to affect global and regional radiative and thermal energy budgets (Barry 1985, Shine *et al.* 1990). Snow contributes to the hydrologic cycle regionally and globally and is a key element in ocean thermohaline circulation (Broecker 1991).

The insulation value of the snow is important to soil processes and to the winter survival of many species of plants and animals. The accessibility of vegetation to herbivores is partly determined by the depth, density, and location of snow, and thus snow affects their populations and distribution. A persistent increase or decrease in the extent, or a change in the type, of snow cover could severely affect survival of vegetation and animals.

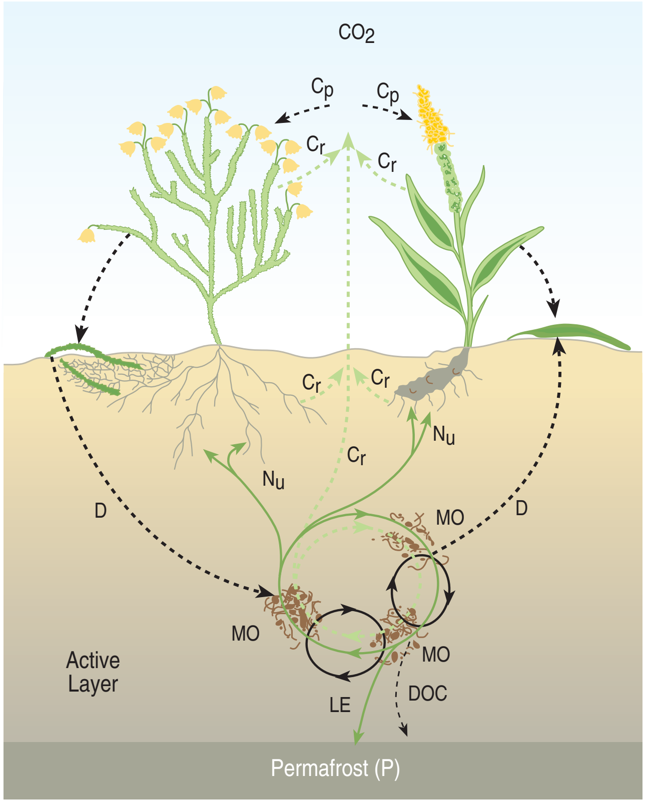


Figure 11·13. Schematic illustration of ecosystem carbon fluxes and nutrient circulation in the tundra (adapted from Callaghan and Jonasson 1995). Carbon and nutrient pathways are shown by dashed and continuous lines, respectively. Atmospheric CO_2 is fixed in the plants through photosynthesis (C_p). Dead parts of plants with organic carbon and nutrients fall to the ground as litter and enter the decomposition cycle (D) where they are transformed into soil organic matter. The organic matter with its nutrients undergoes complicated transformations in the soil microbial biomass (MO). Carbon is continuously lost to the atmosphere as CO_2 through microbial and plant respiration (C_r) and made available directly for plant photosynthesis or added to the atmospheric pool (CO_2). Nutrients are circulated within the microbial system of the soil or traded-off and taken-up by the plants (Nu). Microbial activity is limited by, among other things, low temperatures due to permafrost (P). Some carbon and nutrients can be lost from the system as dissolved organic matter (DOC) or nutrient leachate (LE) and transported to adjacent systems. Climate change acts on the system mainly by controlling the rate of CO_2 uptake by plants, and the activity of micro-organisms. Tundra ecosystems differ from others in that the cold soils have large stores of nutrients and carbon, and the microbial biomass has low rates of CO_2 uptake, i.e. low productivity.

11.2.4.2.5. Terrestrial ecosystems: Physical properties

The terrestrial ecosystems of the Arctic currently cool the Earth in three basic ways: by reflecting incoming radiation from high-albedo surfaces such as snow and ice; by exchanging energy and water with the atmosphere; and by taking CO_2 out of the atmosphere and storing the carbon in organic soils and peats. Terrestrial ecosystems will respond to climatic changes, resulting in positive feedbacks to the atmosphere through mechanisms such as alterations in the balance of trace gas emissions, particularly CO_2 and methane; altered albedo as vegetation cover increases and forest extends into tundra areas; and plant responses to UV and enhanced CO_2. The net result of the array of feedback effects may be positive or negative; overall effects remain inconclusive because of the complexity of the many interactions.

For approximately the last 10 000 years, many tundra terrestrial ecosystems have taken carbon out of the atmosphere, reducing the natural greenhouse effect, because the respiration of CO_2 (dry habitats) and CH_4 (wet habitats) from soil microbial respiration during the decomposition cycle is slower than the uptake of atmospheric CO_2 in green plant photosynthesis (Figure 11·13). This process has resulted in the tundra and boreal regions storing 14% of the world's organic soil carbon. However, if the soils warm and microbial activity increases, there is concern that much of this carbon will be returned to the atmosphere in another positive feedback by increasing the concentrations of the greenhouse gas CO_2.

The tundra has a high albedo compared with the neighboring dark boreal coniferous forest or taiga. Changes in species composition might greatly affect albedo and thus the radiative balance. A northward migration of the tree line would reduce albedo and provide a positive feedback to the climate system, i.e., warming would encourage trees to replace the short tundra vegetation which would further decrease albedo and increase climatic warming. However, pollen records from the end of the last ice age indicate that trees of the taiga move more slowly than the rates of predicted climate change. Possible feedback effects from a change in vegetation would be dampened by this lag. Permafrost disturbance also affects plant species composition and thus albedo.

The greatest change in Arctic ecosystems and thus feedbacks to climate, should be expected where species are at their higher rather than lower temperature-related distribution limits, i.e. at their southern and lower altitudinal boundaries, and where they are in proximity with potential competitors. The tundra/taiga ecotone is, therefore, a critical region of potential change.

11.2.4.2.6. Arctic glaciers and ice sheets

Glaciers and ice sheets are among the defining elements of the Arctic. They are tightly coupled to climate change and have been shrinking, contributing to sea-level rise; positive feedbacks associated with global warming will accelerate this melting of fresh water. Glaciers also play a role in global radiation balance and are an important 'climate archive'. Land ice in the Arctic includes the massive Greenland ice sheet and smaller ice caps and glaciers. The Greenland ice sheet covers 1.7×10^6 km^2, or 80% of the island of Greenland; an additional 0.1×10^6 km^2 of the island is under smaller ice caps and glaciers. Ice cover in other areas of the Arctic (including coastal Alaska) amounts to almost 0.4×10^6 km^2 (Meier 1984).

Glacier mass is tied to precipitation and temperature. Warming will cause most glaciers to recede. Many models assume that, for the Greenland ice sheet, warming will produce both increased melting at low altitudes and increased snowfall at high altitudes, the latter linked to increased saturation vapor pressure (e.g., IPCC 1996a). However, paleoclimate records do not show much dependence of high-altitude accumulation on temperature (e.g., Cuffey and Clow in press). A change in mass balance will produce a time-lagged change in ice flow and glacier extent. Glaciers which terminate at tideline can advance over long distances as a function of submarine moraines (Powell 1990). It is possible that climate changes can influence maintenance of a glacier on its protective moraine. Run-off from glaciers that grow should increase due to summer melting. Eventually, however, temperature increases will lead to glacial recession.

A compilation of studies (Meier 1993) suggests that a global warming of 1°C will lead to ≈1 mm per year of sea-level rise from small ice caps and glaciers, with the Arctic supplying over half of the total, and an additional 0.3-0.4 mm per year contributed from the Greenland ice sheet (IPCC 1990a). Over longer time scales (a century or more) the small glaciers will shrink as the ice melts. The total volume of ice in smaller glaciers is not known with high accuracy due to a scarcity of thickness measurements, but it is probably enough to raise sea level ≈0.3 m if melted (IPCC 1996a). Long-term persistence of warming of more than 3°C is predicted to cause the Greenland ice sheet to shrink and split into northern and southern domes. Warming of more than 6°C, a plausible value, is predicted to remove the ice sheet entirely over ≈10 000 years (Letreguilly *et al.* 1991). The 2.6×10^6 km^3 of water in the ice sheet, plus up to an additional 0.1×10^6 km^3 in smaller glaciers on Greenland, would raise global sea level over 7 m if completely melted.

The residence time (how long typical snowfall spends in the ice, obtained by dividing the volume of the glacier by the input rate) averages a few centuries for the ice caps and glaciers of the Arctic, but several thousand years for the Greenland ice sheet. Because glaciers are controlled by the physical processes of ice motion as well as by snowfall and melting, changes in glacier size typically lag climatic changes (by years to decades for typical mountain glaciers), and record smoothed versions of any change (see Paterson 1994). The Greenland ice sheet typically responds much more slowly to climate change than other ice masses in the Arctic. If climate change is amplified in polar regions as expected (IPCC 1990a, 1992a, 1996a, Cuffey *et al.* 1995), Arctic glaciers will have a greater influence on the global water balance than estimated here.

11.2.4.3. Atmospheric regime

The atmosphere, along with the oceans, directly ties the Arctic environment to the rest of the world. Atmospheric transport of trace gases, including greenhouse gases, aerosols, and water, directly affects the Arctic environment. Many of the observed changes to the atmosphere are anthropogenically driven and it may, to some extent, be possible to control them through international agreements. The effects of atmospheric changes on the Arctic are particularly strong because of positive feedback processes.

11.2.4.3.1. Atmospheric structure and components

The troposphere is the lowest portion of the atmosphere and is where the interaction with the surface and the ocean/land interactions are most dynamic. The troposphere can be divided into the boundary layer, which is close to the earth's surface, and the free troposphere, where the exchange of gases with the stratosphere can be important. The atmospheric boundary layer contains moisture, trace gases, and aerosols from localized sources and is characterized by rapid circulation. The height of the boundary layer in the Arctic can range from a few meters to a few kilometers. The biosphere is immediately affected by processes and constituents in the boundary layer.

The stratosphere, roughly 10-30 km above the earth's surface, is the portion of the atmosphere where most of the total column ozone is located. The dynamic nature and chemical composition of the stratosphere exert a major influence on ozone concentrations and are extremely important to climate change. The stratosphere is less affected by local sources and sinks of gas and is therefore more sensitive to global emissions than the troposphere. At higher levels in the Arctic stratosphere, the cooling effects of both ozone depletion and greenhouse gas increases should be clearly evident against the background of natural variability. In the Arctic winter, stratospheric flow is dominated by extensive, and more or less circumpolar, cyclonic systems. These systems are referred to as the 'polar vortices'.

Temperature

The cold polar night is a dominant characteristic of the Arctic climate. Indeed, the Arctic is often defined solely in terms of low temperature, i.e. as that area where the average temperature for the warmest month is below 10°C. Temperature disparities in the Arctic, between oceans and continents, across latitudes, and between seasons, are complex. The contrast between summer and winter climates in the interior land masses of the subarctic is particularly extreme. For example, Verkhoyansk, Siberia recorded an absolute January low of –67.8°C and a summer high of 36.7°C. Steep temperature inversions characterize the Arctic boundary layer in winter; these inversions are particularly strong over land.

Air temperatures over the sea ice cover tend to be warmer relative to continental temperatures in the winter and a near constant 0°C in the summer. The uniform and relatively warm winter temperature is maintained by a balance between the net radiation, which amounts to about –30 W/m^2, and the conductive flux and sensible heat flux, which average about +15 W/m^2 each (Maykut 1982). In the summer, the strong positive net radiation, 80 W/m^2 in July, is mostly balanced by melt.

Air temperatures over the winter pack ice are colder near the Canadian archipelago than on the Siberian side of the Arctic Ocean. The ice-free regions in the peripheral seas warm above freezing in summer, depending on location and local wind and ocean current regimes. In the winter these areas are covered for the most part by thin, first-year ice, resulting in both warmer surface temperatures and more leads. Strong horizontal temperature gradients are observed near the ice edge where there are large variations in the surface fluxes and surface temperatures, depending on whether the wind flows on or off the ice (Overland and Guest 1991).

The diurnal cycle influences air temperature in the spring, summer, and fall seasons except near the Pole where the sun maintains a constant height above the horizon. Synoptic-scale variations, with periods between 2 and 8 days, dominate the air temperature variability, except in summer when fluctuations are small. The winter air temperature is strongly correlated with cloud cover and wind speed. In the winter, the air temperature is an average of about 9°C warmer under cloudy skies than under clear skies, and 1.2°C warmer for every meter per second increase in the wind speed.

The temperature increase caused by stronger wind speeds is due to breakdown of the low-level temperature inversion through mixing. This increase in temperature with altitude is a characteristic feature of the polar boundary-layer climate. Primarily driven by radiation deficits at the surface and warm air advection aloft, the inversion typically extends to altitudes of ≈1 km over the Arctic Ocean, with temperature differences of ≈10°C. It is strongest during winter, although a secondary peak exists during the late summer due to melting at the ice-air interface. The seasonal and inter-annual variability in the Arctic inversion layer has been reported in several studies (Serreze *et al.* 1992, Skony *et al.* 1994, Kahl *et al.* 1996).

Surface winds

Surface winds are an important climatic parameter because of their influence on ambient temperatures, as well as energy, trace gas, and humidity fluxes. Longer period changes in atmospheric circulation, including both air temperature and winds, drive variations in sea ice extent (Rogers and van Loon 1979, Walsh and Johnson 1979, Walsh and Sater 1981, Fang and Wallace 1994). The surface stresses at the air–ice interface are communicated to the water and ultimately are responsible for the gross aspects of the basin-wide circulation and local transports, such as through Bering Strait (Aagaard *et al.* 1985). Surface winds tend to be suppressed with clear skies, and stronger (closer to geostrophic) during cloudy conditions due to the effects of downwelling longwave radiation on boundary layer stability.

The northern boundary of the westerlies lies at about 70°N, but even in a general sense an 'Arctic high' and 'polar easterlies' are only weakly indicated and, compared with mid-latitudes, circulation is generally sluggish. Vowinckel and Orvig (1970) report typical wind speeds of 4-5 m/s over the central Arctic, with greater speeds more common at the periphery, especially at the exits of storm tracks from mid-latitudes. Seasonal and geographical variations in the Arctic sea-level pressure and wind fields show that cyclonic systems are relatively strong and prevalent in the Norwegian and Barents Seas in winter, and are more uniformly distributed and weaker in the summer (Serreze and Barry 1988, Walsh and Chapman 1990, Serreze *et al.* 1993). Anticyclones are prevalent over Siberia, Alaska, the Yukon, and central Arctic during the winter, and along the margins of the Arctic Ocean during the summer. The spring and fall transition seasons generally feature high pressure near the pole and more frequent, anticyclonic circulation centers in the western Arctic.

Significant temporal variations occur in the seasonal mean winds. Variability on time scales of a few days is dominated by transient cyclones and anticyclones, and their associated fronts (e.g., Shapiro *et al.* 1989). Higher wind speeds are found with cyclones or fronts; these storms either migrate into the Arctic from lower latitudes or develop locally from generally baroclinic processes (LeDrew 1985, 1988). Anomalous atmospheric conditions within the Arctic have been attributed to both variations at lower latitudes, e.g., modulations in the Aleutian and Icelandic lows (Agnew 1993), and to anomalous boundary forcing due to sea-ice within the Arctic (Mysak *et al.* 1990). Interdecadal trends for the Arctic include the tendencies for a decreasing intensity of the Siberian anticyclone during winter (Sahsamanoglou *et al.* 1991), and an increasing frequency for cyclones and anticyclones north of 65°N throughout the year (Serreze *et al.* 1993) which will affect the mean winds and their variations.

GCM simulations indicate that warming may cause reduced sea-level pressure in the Arctic (e.g., Gregory 1993) which will combine with a poleward shift in the mid-latitude baroclinicity and storm tracks (e.g., Hall *et al.* 1994) to cause more cyclonic storm systems, especially in the winter. This implies more variability in the day-to-day winds, but it is not certain whether there would be any systematic changes in the intensity of storms or in the typical strength of winds. The present Arctic climate is punctuated by inter-annual to decadal scale variations of substantial amplitude. These variations will probably continue and at any particular time their effects on atmospheric circulation and surface winds are liable to outweigh the slow changes associated with climatic warming.

Understanding of surface winds in the Arctic is hindered by sampling problems and a scarcity of observations. In principle, wind speed and direction can be derived from the sea-level pressure field. Given the geostrophic flow, the surface wind depends on the surface roughness and on the atmospheric boundary layer structure, especially the vertical profile of static stability (e.g., Overland 1985).

Clouds

Clouds have a strong impact on the energy and hydrologic balance of the Arctic. Even small changes in cloud cover can alter the radiative balance. Total cloud cover during winter ranges from 40 to 70% (Vowinckel and Orvig 1962, Huschke 1969, Gorshkov 1983, Warren *et al.* 1986, 1988); the greatest cloud cover is found over the Atlantic side of the Arctic Ocean where cyclonic activity is most frequent and water vapor is relatively abundant (Serreze *et al.* 1995). Cloud cover is typically more limited over land, sea ice cover, and the central Arctic Ocean, areas where water vapor is less abundant, the lower boundary layer is more stable, and anticyclonic atmospheric conditions are common (Shine *et al.* 1984, Serreze *et al.* 1993). Convective clouds (cumulus and even cumulonimbus) are fairly common in winter over the open waters of the Norwegian and Barents seas.

Total cloud amount increases to 70-90% in summer, with the most rapid increase between May and June. Extensive low-level, optically-thin Arctic stratus prevail over the ocean when relatively warm, moist air moves over the melting sea ice cover and cold water and condenses (Herman and Goody 1976). An increase in cyclonic activity over the central Arctic Ocean in summer is associated with increases in poleward water vapor transport and vapor flux convergence (Serreze *et al.* 1995).

There are still significant gaps in the understanding of cloud formation and maintenance in the Arctic (Moritz *et al.* 1993). Related to this are uncertainties in assessing the importance of surface and advective fluxes of moisture through open water. These uncertainties make it difficult to predict climate change in the Arctic.

11.2.4.3.2. **Radiatively important trace substances**

Changes in the atmospheric concentrations of a few trace gases (CO_2, CH_4, H_2O, N_2O, CFCs, and ozone) and aerosols can strongly influence changes in the global energy balance. These gases contribute to greenhouse warming in various ways. While some of them are emitted and absorbed in the Arctic, most of the changes in mixing ratios of these gases are due to changes in anthropogenic activity at lower latitudes. The atmospheric cycles of many of the radiatively important trace substances are affected by the hydroxyl radical (OH), which is the result of complex atmospheric chemistry involving production by ozone photolysis and destruction by reaction with CH_4, CO, and a host of atmospheric trace substances (e.g., non-methane hydrocarbons). Changes in the Arctic climate are likely to alter the trace gas balance within the Arctic and thus directly affect global climate.

The Arctic peatlands, boreal forests, and permafrost contain potentially important sources of trace gases which could amplify the effects of warming. The 7.5 million km^2 Arctic region holds a vast store of carbon in the soil, mainly in the wetlands which cover about 1.5 million km^2 north of 60°. This store represents the imbalance, accumulated over thousands of years, between carbon fixed annually by plants in net production and that released back to the atmosphere as CO_2 and CH_4 through decomposition. The balance of the processes of production and decomposition and of CO_2 and CH_4 emission are strongly influenced by climatic factors and may have an important effect on greenhouse gases. Evidence indicates that the Arctic Ocean currently contributes an insignificant amount of CH_4 to the atmosphere yet stores larger concentrations under the polar ice cover.

The dominant environmental controls over fluxes of water, energy, and trace gases from Arctic ecosystems vary seasonally, often changing completely from summer to winter. Soil and vegetation may be a source or a sink of CO_2 in summer but are a net source in winter (Oechel and Vourlitis 1994), while CH_4 flux is a summer phenomenon. Winter variations in energy budgets are determined by slope and aspect as well as by albedo, whereas summer energy budgets are governed by evapotranspiration, which also governs runoff in rivers.

Trace gas and energy fluxes also change with spatial scale. At the plot scale, vegetation type, nutrient availability, and moisture are important predictors. At the landscape scale, moisture and hydrology, as governed by slope and aspect, are the major controls. At the regional scale, temperature and vegetation become the dominant controls. Moisture has opposing effects on the two major trace gases: CH_4 flux declines with soil drying while CO_2 flux initially increases (Reeburgh 1985).

Carbon dioxide

Globally, carbon dioxide concentrations have increased by almost 30% since the late 18th century. Most of this increase has been attributed to combustion of fossil fuel, cement production, and land use changes. Continued increase of CO_2 is expected due to continued anthropogenic activities.

Ecosystems in the Arctic store large amounts of carbon in soils that turn over very slowly. Global warming could be exacerbated by the loss of stored carbon emanating as CO_2 from terrestrial ecosystems. Interactions between carbon and nutrients (nitrogen, phosphorus) are a major constraint on the carbon balance of all ecosystems, including the Arctic (Shaver *et al.* 1992). Increased soil temperature and increased decomposition are likely to increase annual emissions of CO_2 and reduce the proportion of carbon emitted as CH_4. Other important factors affecting CO_2 release are more subtle climatic changes such as cloudiness and the diurnal temperature range.

Schlesinger (1977) estimates the amount of carbon stored in tundra (and alpine) soils and in boreal forest soils combined is equal to 166 ppm of CO_2 in the global atmosphere. For comparison, the pre-industrial atmosphere contained 280 ppm of CO_2, and the total cumulative amount of fossil fuel carbon released to the atmosphere through 1995 is equivalent to 118 ppm of CO_2 (Keeling 1994, Marland *et al.* 1994).

Raich and Schlesinger (1992) summarized data on soil respiration rates and found positive correlations between respiration, precipitation, and annual mean and seasonal temperature. An increase in temperature was correlated with a median increase of CO_2 flux of 9% per degree (Raich and Schlesinger 1992). If warming is accompanied by increased precipitation, respiration would be boosted further. On the other hand, primary productivity is also positively correlated with both temperature and precipitation. Net carbon uptake by ecosystems depends on the difference between production and respiration.

Methane (CH_4)

Methane is a radiatively active trace gas that has been increasing in concentration at a rate of about 0.9% per year for at least the past 100 years (Khalil and Rasmussen 1983b, Steele *et al.* 1987, Blake and Rowland 1988), although recently the rate has decreased to 0.5% per year (Steele *et al.* 1992). On a per molecule basis, CH_4 exerts a greenhouse forcing that is an order of magnitude stronger than that of CO_2 (Shine *et al.* 1990). Sources of atmospheric methane are primarily located outside of the Arctic and include wetlands, rice paddies, ruminants, biomass burning, gas production and transmission, termites, landfills, coal mining, and gas hydrates (Khalil and Rasmussen 1983b, Cicerone and Oremland 1988). However, the Arctic tundra is a potential source of methane and there is concern that if the soil active layer increases, CH_4 may increase through the microbial production of methane. The relationship between methane production and its release to the atmosphere is controlled by soil moisture and vegetation type so that processes which determine plant community composition influence methane release (Whalen and Reeburgh 1990a, 1990b). Soils may also consume methane through methane oxidation: there is increasing evidence that the large areas of dry and mesic High Arctic tundra may form a significant sink for methane (Christensen *et al.* 1995).

Present information suggests that the Arctic Ocean has only a minimal impact on the global methane budget (Kvenvolden *et al.* 1993, Lammers *et al.* 1995). Methane concentrations in Alaskan waters which freeze over in winter are 3 to 28 times greater than in Canadian near-surface waters where the methane is approximately in equilibrium with the atmosphere (Kvenvolden *et al.* 1993). The surface water under the polar ice cover is supersaturated with methane that usually exceeds the atmospheric equilibrium concentration by factors of about 1.3 to 4.0. The polar ice cap, however, forms a barrier to the atmosphere (Kvenvolden and Lorenson 1995). These differences in oceanic methane concentrations suggest that methane builds up in water when ice is present and is released to the atmosphere when ice is absent. Because methane is released in the Arctic from coastal waters over a very short time period each year, it adds to the seasonal methane cycle and may have some regional impact.

Carbon monoxide

Carbon monoxide, while not a greenhouse gas, plays an important role in the chemistry of the lower atmosphere largely because of its reaction with the hydroxyl radical (OH), a strong oxidant which controls the distributions of many chemically-reduced species. There is an inverse relationship between the concentrations of these two molecules, such that high levels of CO tend to deplete OH and high levels of OH deplete CO (Logan *et al.* 1981).

The greatest CO mixing ratios in the background atmosphere are found in the high latitudes of the northern hemisphere. CO mixing ratios in the Arctic boundary layer have a well-defined seasonal cycle, with the highest levels (200-250 ppb) observed in late winter/early spring and lowest levels (80-90 ppb) found in summer (Novelli *et al.* 1992).

This distribution of CO in the Arctic is determined by both natural and anthropogenic processes. CO has a long lifetime in winter and high concentrations of man-made pollutants found in the Arctic during winter are the result of their transport from the northern mid-latitudes (Khalil and Rasmussen 1983a, Rahn and Lowenthal 1986). In summer the lifetime of CO in the Arctic is shortened because OH levels increase; CO decreases until August, after which it starts to increase again. The hydroxyl radical is responsible for the removal of many climatically-significant trace gases in addition to CO, including dimethyl sulfide, methane, and the HCFCs (replacements of the CFCs). Thus, changes of CO levels in the atmosphere are expected to have an impact on Arctic climate and global trace gas budgets.

Nitrous oxide (N_2O)

N_2O is an important greenhouse gas because of its long atmospheric lifetime and a radiative forcing larger than CO_2 on a per-molecule basis. Global levels of N_2O continue to rise in the background atmosphere. The main natural source of N_2O is emission from the oceans and from wet soils where N_2O is produced by bacterial activities; the main anthropogenic sources are from fertilizers in the cultivation of soils. The mean tropospheric mixing ratio of N_2O was about 310 ppbv in 1992 (WMO 1994), with an annual increase of about 0.8 ppbv per year over the past four decades (IPCC

1994). Over the past decade, annual increases ranging from 0.5 to 1.2 ppbv per year have been observed (Khalil and Rasmussen, 1992). The main (and the only quantified) sinks of atmospheric N_2O are photolysis and reaction with $O(^1D)$ in the stratosphere. A possible sink of tropospheric N_2O is uptake by soils (Donoso *et al.* 1993).

Understanding the correlation between N_2O and NO_y is particularly important for the investigation of the budget of the nitrogen family and also for the identification of denitrification in the Arctic winter (Kondo *et al.* 1994, 1996). Denitrification has a very strong influence on the chemistry of stratospheric ozone destruction in the polar region (Fahey *et al.* 1990a, 1990b, Loewenstein *et al.* 1993, Kawa *et al.* 1992b, Oelhaf *et al.* 1995, Nakajima *et al.* 1996). Correlations between N_2O and ozone have been used by several investigators to infer chemical ozone loss in the Arctic (Proffitt *et al.* 1990, Manney *et al.* 1994a, Bregman *et al.* 1995).

Due to its long photochemical lifetime, N_2O can be used as a tracer to study vertical and horizontal transport in the stratosphere. A series of *in situ* measurements of N_2O in the polar Arctic vortex has been used to investigate the morphology of N_2O distribution (Schmidt and Khedim 1991, Loewenstein *et al.* 1990), both with respect to the position relative to the polar vortex (Podolske *et al.* 1993) and with respect to time (Bauer *et al.* 1994). These studies show the descent of N_2O-poor air inside the polar vortex and a strong gradient of N_2O across the vortex boundary.

Chloroflourocarbons (CFCs)

Chloroflourocarbons (CFCs) are direct greenhouse gases; their contemporary radiative heating amounts to about 10% of the global total of radiative forcing from all anthropogenic gases (IPCC 1996a). CFCs also influence the atmospheric energy budget indirectly because they affect the abundance of stratospheric ozone, particularly in polar regions. The flux of CFCs to the atmosphere from Arctic terrestrial sources, however, is negligible in comparison to sources from mid-latitudes (McCulloch *et al.* 1994). The direct effect of CFCs on ozone concentrations are dealt with in section 11.3.2.

With the exception of very small quantities from volcanic vents, these substances are present in the atmosphere as a result of human activity (Isodorov *et al.* 1990). CFCs are removed from the atmosphere by photolysis in the stratosphere so that tropospheric concentrations are governed by mass transport to the stratosphere and the rates of removal there (WMO 1994). The decrease in the anthropogenic CFCs produced and emitted since 1990 is reflected in stable or decreasing levels (Simmonds *et al.* 1996, UNEP 1994). This has the effect of reducing the range of concentrations close to the global mean value, particularly in remote locations such as the Arctic.

Recent calculations show that the direct radiative heating from CFCs is, to an extent, offset by indirect effects arising from their interaction with stratospheric ozone (Ramaswamy *et al.* 1992, Daniel *et al.* 1995, Solomon and Daniel 1996). The magnitude of the indirect effect is liable to be larger toward the poles, since the greatest ozone losses are observed in the polar stratosphere. The major concern for the Arctic is the extent to which polar stratospheric ozone depletion offsets local greenhouse enhancement.

Water vapor

Water vapor affects greenhouse warming, particularly because it absorbs strongly in the infra-red. This is especially important in the Arctic where humidity is low due to cold temperatures. As the air temperature of the Arctic increases, the atmosphere will be able to hold more water vapor, and increases in water vapor will cause additional regional warming. Water vapor is difficult to measure at cold temperatures and atmospheric monitoring often does not address it.

The Intergovernmental Panel on Climate Change (IPCC) (1990a) stated that GCMs show that an increase in humidity will parallel an increase in warming under a doubling of CO_2. However, data show a decrease in precipitable water from the surface to 500 mb over northern Canada (Elliot *et al.* 1991). More recent analyses show that 20-year trends of relative humidity are positive for North America but negative at 500 mb in interior Alaska and northern Canada; at the same time, the dewpoint in Canada and Alaska is increasing (Ross and Elliot *et al.* in press). These seemingly contradictory results emphasize the need for improved climate models and continued measurements in the Arctic.

Aerosols

Aerosols in the troposphere have the potential to modify climate by perturbing the earth's radiative balance (Wigley 1989, Charlson *et al.* 1992). Climate forcing can occur directly as a result of particulate scattering and absorption of solar radiation, or indirectly as a result of aerosol-induced changes in the prevalence, location, and albedo of clouds. The removal of these atmospheric particles by deposition to snow and ice has the potential to then influence climate by alteration of the surface albedo. Averaged over the entire northern hemisphere, the annual direct forcing due to sulfate aerosols has an estimated value of about 0.5-1.0 W/m^2, which is similar in magnitude but of the opposite sign to the forcing by current levels of all anthropogenic greenhouse gases combined (Charlson *et al.* 1991).

The absorptive black carbon component of aerosols is much more effective in altering the atmospheric energy budget in the Arctic than elsewhere due to the high albedo of the Earth's surface in the Arctic. In the period March to May, these aerosols have a net positive radiative forcing (Blanchet 1989, 1991). This is opposite to the net effect of these aerosols on a global basis. Thus, there are large spatial and temporal gradients in aerosol forcing associated with aerosols over northern, high albedo areas (i.e., the Arctic and much of the continents), and these need to be addressed in climate change studies.

Natural aerosols in the Arctic include terrestrial dusts, marine salt particles, ice crystals, smoke particles, and biogenic particles. Aerosols from major volcanic eruptions such as El Chichon or Mount Pinatubo are important to the Arctic and are transported predominantly in the stratosphere while small, local volcanic eruptions are transported in the troposphere and are less important.

Aerosols in the Arctic are of climatic importance because of their relatively long tropospheric lifetimes (Ogren and Charlson 1983). The comparatively low levels of precipitation and low mixing ratios of water vapor in the High Arctic mean that particles are unlikely to be scavenged by precipitation or to absorb atmospheric water and grow to sizes conducive to rapid settling. Typically, aerosol concentrations increase in the winter and early spring when northern mid-latitude pollution sources are strongest, meteorological conditions are favorable for aerosol transport into the Arctic, and the lower troposphere is extremely stable. Components of the springtime aerosol maximum, known as Arctic haze (Shaw 1995), reach concentrations similar to those found in populated mid-latitude regions (Rosen *et al.* 1981), a remarkable finding considering the sparse population of the Arctic.

In the Arctic, winter and springtime aerosols have a major anthropogenic component, probably originating in the mid-latitudes (Barrie *et al.* 1981, Ottar 1981, Rahn 1981, Lowenthal and Rahn 1985). Summer and fall aerosols show generally lower overall concentrations, have a smaller anthropogenic component, and contain relatively more natural aerosols (Lannefors *et al.* 1983). March-April anthropogenic aerosol levels may have followed a statistically-significant decrease from 1982 to 1992, hypothesized to result from a reduction in pollution emissions (Bodhaine *et al.* 1993, 1995). Others, however, challenge these conclusions (Jaffe *et al.* 1995). Increasing aerosol concentrations are suggested (Barrie *et al.* 1988) as a possible explanation for winter and spring surface cooling (Finlayson-Pitts *et al.* 1990), via an intensification of the atmospheric dehydration cycle (McConnell *et al.* 1992).

The presence of one aerosol type can influence the prevalence of another. For example, ice crystals are not typically present in Arctic haze. One explanation is that the haze aerosols, composed largely of sulfate and sulfate coated particles, de-activate scavenging ice-forming nuclei (Clarke and Noone 1985, Borys 1989). Consequently, clear sky precipitation increases, and the airmass dehydration rate is accelerated. Fewer activated particles also increase the lifetime of the aerosol, cooling surface temperatures (Blanchet and Girard 1994, 1995, Bradley *et* al. 1993), causing cold anomalies exceeding –2°C over the ice covered Arctic Ocean, and strengthening the surface temperature inversion by about +4°C during fall and winter over the last 40 years (Kahl *et al.* 1993a, 1993b). This is contradictory to the effects of greenhouse gases alone.

A modest amount of carbonaceous aerosol (0.2-0.6 mg/m^3) increases diabatic heating (0.1-0.2 K per day) when compared with the effects of water vapor (Valero *et al.* 1988). Atmospheric warming is further enhanced by reflected visible light from the snow covered surface. Temperature profile analyses (Kahl *et al.* 1993a, 1993b) indicate a springtime warming tendency in the lower troposphere of the High Arctic. In parallel, early model simulations estimate a springtime warming of 1-2°C (Blanchet 1989). Model simulations indicate that the excess heat produced by soot aerosols will increase temperature but will also preferentially alter the surface energy balance, snow and ice cover, and regional scale circulation. According to Kahl *et al.*'s (1993a, 1993b) results, this amount of warming is marginally significant given the natural variability of temperature in the Arctic.

Another potential influence of tropospheric sulfate aerosols is a role in hosting chemical reactions that lead to tropospheric ozone destruction. There is evidence that reservoirs of acidic chlorine and bromine compounds that are inactive in ozone destruction are converted to active chlorine and bromine gases on sulfuric acid aerosols (Barrie *et al.* 1994, McConnell *et al.* 1992).

Large uncertainties in the magnitude of, and strong spatial gradients in, aerosol forcing make it crucial that aerosol influences be actively represented in climate models, rather than simply represented as a passive anti-greenhouse influence (Charlson *et al.* 1992). The results of Wigley and Raper (1992) indicate that the neglect of anthropogenic particles in GCMs will result in an uncertainty in temperature projections for the Northern Hemisphere in the year 2050 of about 2°C.

Tropospheric ozone

Ozone in the Arctic troposphere comprises roughly 5-10% of the total overhead burden of atmospheric ozone. Despite this relatively small percentage, tropospheric ozone (usually below about 8-10 km in the Arctic) plays a key role in both the chemical and radiative behavior of the atmosphere. In the upper troposphere, ozone acts as a greenhouse gas, trapping infrared radiation that would otherwise escape to space, leading to heating of the lower atmosphere. In the lower troposphere, ozone is a pollutant and acts as a poison to both plants and animals.

Figure 11·14. Trends in (annual average) ozone mixing ratio over Canada from 1980 to 1993 (from Tarasick *et al.* 1995).

Evidence from vertical profile measurements of ozone in the Arctic suggests that the stratosphere is an important source of ozone to the upper and middle troposphere (Browell *et al.* 1992, 1994, Oltmans *et al.* 1989b, Raatz *et al.* 1985, Bachmeister *et al.* 1994). Tropospheric ozone has a spring maximum (except in the boundary layer over and near the frozen Arctic Ocean) and an autumn minimum. The spring buildup may have more than one source, although the late winter and spring maximum of ozone in the lower stratosphere probably contributes substantially to the peak in the troposphere (Oltmans 1991). A strong low-level temperature inversion and proximity to the Arctic Ocean determine the distribution of ozone in the troposphere.

The significant decline in ozone in the lower stratospheric reservoir over the past decade may reduce tropospheric ozone levels, since in recent years a decline in tropospheric ozone parallels that seen in the lower stratosphere. Though the tropospheric trends (Figure 11·14) have not been mechanistically linked to the stratospheric decline, the parallel nature of the changes is clearly suggestive. In polar regions, combinations of cold winter temperatures, aerosols, and long periods of darkness and sunlight facilitate the chemical alteration of ozone. These chemical processes combine with transport regimes that may either incorporate ozone precursors into polar latitudes or mix ozone from the stratosphere into the troposphere.

Episodes of very low tropospheric ozone observed during the spring in the Arctic Ocean basin are unique (Barrie *et al.* 1988, Oltmans *et al.* 1989b). Shortly after the appearance of sunlight over the frozen Arctic Ocean, ozone concentrations have been measured at near zero for up to a few days (Oltmans *et al.* 1989a). At the onset of these episodes a strong inverse correlation was noted between ozone and atmospheric bromine (Barrie *et al.* 1988, Oltmans *et al.* 1989a). This ozone depletion has not been observed inland from the Arctic Ocean or at sites where the ocean remains open during the winter (Oltmans and Levy 1994, Barrie *et al.* 1994). The bromine that plays a key role in the ozone loss is of biogenic origin and is emitted through leads in the ice (Sturges *et al.* 1993). The exact mechanism of the ozone-bromine reaction has not been established, but may involve heterogeneous chemical reactions as in the stratosphere (McConnell *et al.* 1992). Ozone replenishment from above is prevented by a strong temperature inversion which keeps air trapped near the surface (Mickle *et al.* 1989) and limits the impact of biogenic bromine on the higher levels of the Arctic troposphere.

Local sources can also add to the ozone concentrations in the troposphere. Forest fires over Canada produce ozone enhancements within the plume (Browell *et al.* 1994, Anderson *et al.* 1994). In areas with a large population or industrial activity, local ozone production might be expected when sufficient sunlight and ozone precursors are available (Jaffe 1991). The effect of fire- or industrially-produced ozone on ozone distribution within the Arctic is expected to be small. Mauzerall *et al.* (1996) suggest that during summer the largest source of tropospheric ozone below 5 km is dispersed, *in situ* photochemical production.

11.3. Arctic stratospheric ozone

Changes in stratospheric composition have been observed in both the Arctic and Antarctic polar regions. Changes in stratospheric temperature, aerosol concentration, and ozone concentration are important to the energy balance of the Arctic as well as to the UV which reaches the biosphere. Upper-air measurements may offer clearly discernible relationships to such phenomena as global warming due to the enhanced greenhouse effect and to the effects of ozone depletion. At higher levels in the Arctic stratosphere, the cooling effects of both ozone depletion and greenhouse gas increases should be clearly evident against the background of natural variability. Measurements of change in these upper atmospheric layers, in combination with quantities observed at the surface, provide an ensemble of information to give the most robust opportunity for climate change detection. In the Arctic winter, stratospheric flow is dominated by extensive, and more or less circumpolar, cyclonic systems. This flow is referred to as the 'polar vortex'. The coherence of these polar vortices determines the exchange of stratospheric air between lower latitudes and the Arctic.

11.3.1. Arctic stratospheric ozone

Ozone is an important trace gas in the earth's atmosphere because it blocks harmful UV radiation from reaching the biosphere and it acts as a greenhouse gas. Significant changes have occurred in the Arctic ozone layer during the past decade: long-term trends have been measured, and short-term, geographically isolated events of extremely low ozone levels have been observed. Some studies have confirmed the importance of halogen chemistry in the destruction of the ozone layer; however, the causes and the magnitude of recent changes in Arctic halogen processes are not completely understood. Recent, unusually cold stratospheric temperatures and changes in circulation patterns have also been cited as important factors in observed low ozone concentrations. Despite the potential impact on the environment and humans, little scientific effort has been made to understand the changes in the Arctic stratospheric ozone layer.

Most of the ozone in the atmosphere is in the 'ozone layer', a region of the stratosphere between 10 and 30 km above the earth's surface. Because ozone is the major absorber of near-UV radiation in the stratosphere, it helps determine the temperature structure of the stratosphere and thus affects atmospheric circulation processes. Stratospheric ozone is destroyed by heterogeneous chemistry involving chlorofluorocarbons (CFCs) and other ozone-depleting substances in reactions that are facilitated by the occurrence of extremely low stratospheric temperatures at the end of winter when the molecules are energized by the return of sunlight. The unique temperature and dynamics of the Arctic atmosphere, in combination with chemical processes, contribute to greater ozone loss in the Arctic than at mid-latitudes.

Predictions for future quantities of ozone in the Arctic are unreliable due to a lack of understanding of photochemical and dynamical processes forming the Arctic atmosphere (WMO 1994). Since the 1994 WMO report, further reductions in the Arctic ozone layer have been observed, and questions remain concerning the observed changes.

11.3.2. Chemistry of ozone depletion – Polar vortex dynamics

A great deal of scientific effort has gone into understanding the physical and chemical processes contributing to the Antarctic ozone hole. Less is known about processes of Arctic ozone depletion because, while similar in its general climate, the Arctic does not form a distinct seasonal ozone hole. This is primarily due to the instability of the Arctic polar vortex, a consequence of larger land masses in the northern middle hemisphere than in the southern middle hemisphere. The vast land area leads to greater land-sea temperature contrasts and, hence, more planetary-scale waves that disturb the northern polar vortex and result in frequent major stratospheric warming. In Antarctica, the lowest ozone depletions are observed inside the polar vortex, with clear indications that heterogeneous chlorine and bromine chemistry is the dominant factor behind ozone deviations (WMO 1994). The strong polar winds in the Antarctic isolate the air masses, allowing them to reach much cooler temperatures than are observed in the Arctic. In the Arctic, the picture is more complex because the dynamic exchange of air masses across the Arctic vortex does not allow for cooling as severe within the Arctic vortex as is observed within the Antarctic vortex. Nevertheless, anomalously low ozone levels have been observed in the Arctic during the last decade.

Both concentration and distribution of ozone in the Arctic stratosphere have undergone changes in the 1990s (Taalas 1993). Some of these changes may be linked to aerosols from the eruption of Mount Pinatubo, but some of the loss observed in the Arctic has been attributed to halogen chemistry. Halogen containing compounds, primarily chlorine oxide (ClO), chlorine nitrate ($ClONO_2$), hydrogen chloride (HCl), bromine oxide (BrO), and bromine nitrate ($BrONO_2$),

Table 11·2. Total column ozone measurement stations [a].

Toronto database St. no.	Year measurements started	Location name	Location country	Latitude, °N	Longitude, ° (negative, if west)
5	1957	Dikson Island	Russia	73.50	80.23
18	1957	Alert	Canada	82.50	−62.30
24	1957	Resolute	Canada	74.72	−94.98
42	1957	St. Petersburg	Russia	59.97	30.30
43	1952	Lerwick	UK	60.13	−1.18
51	1952	Reykjavik	Iceland	64.13	−21.90
52	1943	Tromsø	Norway	69.65	18.95
89	1990	Ny-Ålesund, Svalbard	Norway	78.93	11.88
105	1963	Fairbanks	USA	64.82	−147.87
114	1974	Heiss Island	Russia	80.62	58.10
117	1961	Murmansk	Russia	68.97	33.05
118	1962	Nagaevo	Russia	59.58	150.78
123	1973	Yakutsk	Russia	62.08	129.75
129	1973	Pechora	Russia	65.12	57.10
140	1990	Thule	Greenland	76.52	−68.76
142	1973	Igarka	Russia	67.47	86.57
144	1973	Markovo	Russia	64.68	170.42
145	1974	Olenek	Russia	68.50	112.43
148	1973	Vitim	Russia	59.45	112.58
150	1974	Mansijsk	Russia	60.97	69.07
165	1946	Oslo	Norway	59.91	10.72
186	1975	Tiksi	Russia	71.58	128.92
189	1970	Hornsund, Svalbard	Norway	77.00	15.55

a. Total column ozone levels have been measured for the past several decades in or near the Arctic. Despite their importance for scientific research and long-term monitoring, many of these stations are under threat of being shut down due to budget considerations. The data from these stations are available from the World Ozone and UV Data Center in Toronto, Canada.

appear to be most important. Cold temperatures and polar stratospheric clouds or aerosols found in the Arctic activate the halogen chemical processes that facilitate ozone destruction.

Stratospheric halogens arise primarily from anthropogenic emission of chlorine- and bromine-containing substances in the troposphere. These CFCs do not react in the troposphere and are transported into the stratosphere. This transport occurs primarily across the tropical tropopause and can take 2-3 years (Schmidt and Khedim 1991). The CFCs then circulate toward the polar regions and to higher altitudes in the stratosphere, where they are photolysed by ultraviolet radiation (SORG 1987). CFCs photolyse as they pass poleward: the stratospheric concentration of CFC-11, for example, at 20 km is 10-20% of that in the troposphere (Fraser *et al.* 1994). CFCs can also act as tracers: the shape of the CFC-12 concentration profile has been used by Notholt (1994) to show the subsidence of stratospheric air during winter above Ny-Ålesund, Spitsbergen.

Almost all of the CFC-11 and a significant portion of the other abundant CFCs (CFC-12 and CFC-113) are decomposed by photolysis during transport from the troposphere to the stratosphere (Kawa *et al.* 1992a, Woodbridge *et al.* 1995). The photolysis products that most affect stratospheric chemistry are chlorine (Cl) and chlorine oxide (ClO), which react with ozone in a catalytic cycle that regenerates the chlorine atom. Chlorine eventually forms stable reservoir species that are inactive toward ozone, such as hydrogen chloride and chlorine nitrate. In the presence of polar stratospheric clouds, however, these reservoir species can decompose and augment the abundance of active chlorine in the polar stratosphere. Current global concentrations of active stratospheric chlorine are about 3.5 ppbv, substantially higher than the natural background of 0.6 ppbv which arises mainly from the stratospheric degradation of methyl chloride (WMO 1995).

The importance of ClO lies in its ability to catalyze the destruction of ozone. By a sequence of reactions involving ClO, BrO, and sunlight, the ozone concentration in the polar stratosphere is reduced. These reactions are particularly significant in the stratosphere over the South Pole during the austral spring, where ozone depletion is constrained within the Antarctic polar vortex, leading to the ozone hole. They are potentially more important in the Arctic region because of the high level of biological activity in the Arctic which would be directly affected by increased UV radiation. Reductions of over 40% have been observed in the winter stratosphere at high northern latitudes (SESAME 1995). The impact of these reductions in ozone on UV radiation and its effect on biological systems are not yet fully understood.

11.3.3. Measurements of stratospheric ozone

Ground-based, airborne (such as balloon and aircraft), and satellite instruments are used to monitor ozone and supply data to determine ozone trends in the stratosphere. Some instruments monitor only total atmospheric ozone, while others measure ozone profiles with varying degrees of vertical resolution and range. The ground-based Dobson network, used for measuring total ozone above the location of each instrument, was first established in 1958 (Dobson 1968a). Today, data from about 50 stations are used for analyzing global trends in ozone (Bojkov *et al.* 1995a) (Table 11·2).

Since the late-1970s, satellites have been used to measure ozone globally. Many of these satellites are not appropriate for Arctic research, however, because of limited latitudinal range or dependence on sunlight (Miller 1989, WMO 1988). Satellite measurements have helped advance understanding of both dynamical and chemical processes that determine the distribution of stratospheric ozone. The satellite instruments listed in Table 11·3 use several different techniques to measure ozone in the stratosphere. Not all of the satellite instruments are appropriate for year-round monitoring of the polar ozone layer. Two satellites presently in orbit contain ozone monitoring equipment appropriate for the Arctic. The Global Ozone Monitoring Experiment (GOME) on ERS-2 satellite, launched in 1995, takes measurements of Arctic ozone daily. The EarthProbe satellite, launched in 1997, also has a Total Ozone Mapping System (TOMS) instrument aboard.

11.3.4. Results of measurements

A variety of analyses of ozone data have shown a decrease in Arctic ozone over the past three decades, most of which has occurred since 1980. Results from some of these analyses are summarized in Table 11·4. A major part of the observed ozone decrease is attributed to increased amounts of stratospheric chlorine (ClO) and resulting heterogeneous chemical losses on polar stratospheric cloud particles and sulfate aerosols (WMO 1992, 1995). Many aspects of the observed changes in Arctic ozone are not well understood, and the relative importance of additional processes, including stratospheric cooling and dynamical changes, is an area of active research. Complicating the search for long-term trends are natural, nonlinear disturbances, such as the significant decrease in ozone which occurred during the early 1990s following the injection of sulfate aerosols by the Mount Pinatubo volcanic eruption (Randel *et al.* 1995). Longer-term variations in other variables such as temperature and aerosol surface area distribution may also have contributed to the magnitude and timing of ozone losses in the northern hemisphere (Solomon and Daniel 1996).

Because of the temperature dependence of heterogeneous reaction rates, decreased temperature in the lower stratosphere may further accelerate chemical losses of ozone (e.g., Herman and Larko 1994). Randel and Cobb (1994) sug-

Table 11·3. Satellite instruments for ozone measurements [a].

Instrument	Platform	O_3 measurement	Latitude range [b]	Time period
HALOE	UARS	O_3 profile, 15-80 km (Δz = 2 km)	80°N to 80°S	1991 to present
LIMS	Nimbus 7	O_3 profile, 10-65 km (Δz = 2.8 km)	84°N to 64°S	1978-1979
MLS	UARS	O_3 profile, 15-50 km (Δz = 4 km)	80°N to 80°S	1991 to present
POAM II	SPOT-3	O_3 profile, 10-50 km (Δz = 1 km)	55°N to 71°N 63°S to 88°S	1993 to present
SAGE I	AEM-2	O_3 profile, 10-55 km (Δz = 1 km)	80°N to 80°S	1979-1981
SAGE II	ERBS	O_3 profile, 10-65 km (Δz = 1 km)	80°N to 80°S	1984 to present
SBUV	Nimbus 7	O_3 profile, 25-55 km (Δz = 8 km)	80°N to 80°S	1978-1987
SBUV/2	NOAA-11 NOAA-9 [c]	O_3 profile, 25-55 km (Δz = 8 km)	80°N to 80°S	1988 to present 1985-1990
SME	SME	O_3 profile, 45-65 km (Δz = 3.5 km)	85°N to 85°S	1982-1986
TOMS	Nimbus 7 Meteor 3	Total ozone	90°N to 90°S	1978-1993 1991 to present
GOES	ERS-2	Total ozone	90°N to 90°S	1995 to present
TOMS	ADEOS	Total ozone	90°N to 90°S	1996 to present
TOMS	Earth Probe	Total ozone	90°N to 90°S	1996 to present

a. Satellites for ozone monitoring have not always covered the polar regions. Since the discovery of the Antarctic ozone hole, satellites have been launched to cover the polar regions with the added benefit of providing coverage for the Arctic. Nevertheless, the instruments usually require sunlight, thus preventing full coverage of the Arctic ozone layer for the entire year.

b. The latitude ranges are, in general, not realized during all seasons. Many of the measurements are daylight measurements, thus cannot be obtained for the highest northern latitudes during the local winter (Miller 1989).

c. The calibration system for the SBUV/2 instrument on the NOAA-9 satellite was unsuccessful. A third SBUV/2 instrument was on board the NOAA-13 satellite, which was launched in August 1993 but failed soon after launch.

gested that lower-than-normal temperatures measured by the Microwave Sounding Unit at polar latitudes (65°to 90°N) during 1993 were due to radiative effects of decreased ozone, as measured by Total Ozone Mapping System (TOMS) and Solar Backscattering Ultraviolet (SBUV).

Dynamical processes can result in concurrent, transport-related changes in ozone and temperature in the lower stratosphere (e.g., Roldugin and Henriksen 1996, Finger *et al.* 1995). There is evidence that the observed strong longitudinal dependence of total ozone and lower stratosphere temperature trends in winter (e.g., Randel and Cobb 1994) are a consequence of long-term changes in stratospheric dynamics, specifically in the structures of quasi-stationary waves propagating up from the troposphere (Hood and Zaff 1995). Long-term changes in lower stratospheric circulation and ozone transport may also contribute significantly to zonal mean ozone trends.

Measurements of the vertical profiles of ozone are valuable for interpreting the cause of observed ozone trends. For example, ozone in the Arctic vortex derived from Microwave Limb Sounder (MLS) observations showed no decrease in total column ozone during the winter of 1992/93 (Froidevaux *et al.* 1994), yet MLS profile measurements showed a decrease of about 20% in the lower stratosphere. Since the ozone in the Arctic polar mid-stratosphere was increasing at this time, it was concluded that the increase at the higher altitudes masked the decrease in the lower stratosphere in the column measurements (Manney *et al.* 1994a).

Because of the combined influence of stratospheric temperature, dynamics, aerosols, and chemistry on stratospheric ozone concentrations, linking ozone changes within the Arctic directly and exclusively to anthropogenic activity is difficult. Nonetheless, evidence exists that anthropogenic activity has had an impact on the observed ozone depletion. Ozone depletion over the Former Soviet Union was measured using re-evaluated filter-ozonometer data from 1973-1995 (Bojkov *et al.* 1995b). Ozone depletions averaged 12% for winter and spring and 5% for summer. These depletions in ozone could be explained neither by natural ozone fluctuations nor by instrumental errors. Although the LIMS (Limb Infrared Monitor of the Stratosphere) measurements were of short duration, they are particularly valuable since they were made before lower stratosphere ozone losses due to chlorine chemistry were significant. Manney *et al.* (1994a) compared the evolution of ozone on the 465 K potential temperature surface in the Arctic vortex derived from LIMS data during the 1978/79 winter with MLS observations in the winters of 1991/92 and 1992/93. One of their conclusions was that the LIMS data should be used cautiously when interpreting that recent decreases in vortex-averaged ozone are due to chemical rather than dynamical processes. The LIMS data showed a delayed increase in vortex ozone early in the 1978/79 winter, and a short decrease in January 1979; both of these observations could be interpreted as evidence for chemical depletion, which would not have occurred at a significant level

Table 11·4. Ozone trends in the Arctic. A variety of studies examining ozone in the Arctic show consistent evidence for a decrease in mean ozone levels for the Arctic. The variety of different instruments, platforms, and analyses all showing similar downward trends in ozone demonstrate a strong consensus as to current changes in ozone.

Location	Instrument	Trend in % per decade	Time period	Reference
Western Siberia	Dobson	−3.5±0.8	1973-1994	Bojkov *et al.* 1994
Eastern Siberia and the Far East	Dobson	−3.2±0.8	1973-1994	Bojkov *et al.* 1994
Arctic	Dobson	−7.5±3.8	1964-1994 (winter/spring)	Bojkov *et al.* 1995a
Arctic	Dobson	−5.6±2	1964-1994	Bojkov *et al.* 1995a
50-65°N	Dobson	−1.37±0.34	1964-1991	Reinsel *et al.* 1994
50-65°N	Dobson	−1.54±0.42	1971-1991	Reinsel *et al.* 1994
50-65°N	Dobson	−2.71±0.43	1978-1991	Reinsel *et al.* 1994
55°N	TOMS and SBUV	−5.0	1979-1991	Rusch *et al.* 1994
60°N	SBUV	−5 (lower stratosphere)	1978-1990	Hood and McCormack 1992
50-60°N	TOMS	−3.96±1.15	1978-1991	Reinsel *et al.* 1994
50-60°N	TOMS	−5.03±1.46	1978-1991 (winter)	Reinsel *et al.* 1994
65°N	TOMS	−5.1±1.1 to 8.7±2.0	1978-1993	Herman and Larko 1994
60°N	Sage I/II	−3.5	1979-1991	McCormick *et al.* 1992
63-90°N	TOMS/SBUV	−10.7±1.8	1978-1997	Newman *et al.* 1997
50-60°N	TOMS	−1.8 to −6.7	1978-1994	McPeters *et al.* 1996
60-70°N	TOMS	−2.4 to −6.9	1978-1994	McPeters *et al.* 1996
70-80°N	TOMS	−1.5 to −8.8	1978-1994	McPeters *et al.* 1996

during the 1978/79 winter. After considering the behavior of long-lived tracers in order to separate chemical and dynamical effects, Manney *et al.* (1994b) concluded that MLS ozone measurements in February and March of 1993 indicated that Arctic ozone depletion by chlorine chemistry was indeed occurring.

The vertical distribution of ozone in and out of the Arctic vortex during the winters of 1993/94 and 1994/95 shows that ozone concentrations at the beginning of the winter in the lower stratosphere were higher inside than outside the vortex. Subsequently, concentrations inside the vortex decreased so that by the end of winter the mixing ratios inside were lower than outside. This behavior is consistent with chemical depletion, but further knowledge of the relevant dynamics is required before making a stronger statement (Randall *et al.* 1995). Evidence from MLS ozone measurements suggests chemical depletion of ozone in the lower stratosphere inside the vortex during the winters of 1991/92, 1992/93, and 1993/94 although dynamical mechanisms were not ruled out (Manney *et al.* 1995a). By comparing data from ozonesondes and lidars over Thule, Greenland during the winters of 1991/92 through 1993/94, di Sarra *et al.* (1995) concluded that dynamical perturbations caused a negative correlation between ozone and aerosol concentrations, and masked possible chemical effects such as enhanced ozone depletion on aerosol surfaces. Studies such as these highlight the complementary nature of observational studies of the chemical behavior of long-lived trace gases and theoretical studies exploring the dynamical properties of the stratosphere.

11.3.5. Arctic ozone anomalies

While the decadal trend of ozone in the Arctic is negative, there is inter-annual variability and, more importantly, short-term episodes of severe depletion in areas of a few hundred kilometers in diameter. Data suggest that ozone depletions as large as 20% may have occurred inside the northern polar vortex in the 1980s (Brune 1990, Profitt *et al.* 1990) and as much as 40% in the 1990s (SESAME 1995). Recent studies indicate that both dynamics and chemistry play important roles in these anomalies. Dynamically induced variations include those associated with the Quasi-Biennial Oscillation (QBO) and the 11-year solar cycle (WMO 1992, 1995). The QBO modulation at middle and high latitudes is most pronounced during the winter–spring season (e.g., Tung and Yang 1994, Hood and McCormack 1992). Strong negative ozone anomalies at high northern latitudes occur during the westerly phase of the equatorial QBO as observed in winter-spring of 1993 and 1995.

The 1995 winter–spring negative ozone anomaly over northern middle and high latitudes was particularly prominent and included a new record low minimum of 25% over Siberia and parts of the Arctic (Bojkov *et al.* 1995a). At the same time, the polar vortex was displaced over northern Siberia and temperatures in the lower stratosphere were well below normal. The relative importance of dynamical transport and chemical losses accelerated by the reduced temperatures in producing this anomaly is being investigated. A comparison of northern winters since the 1991 launch of Upper Atmosphere Research Satellite (UARS) with those of earlier years indicates that the 1994/95 winter was marked by an unusually early cold spell and a more isolated vortex (Zurek *et al.* 1996). These characteristics would seem to be more conducive to ozone loss by heterogeneous chemistry (Chipperfield *et al.* 1996). On the other hand, a more isolated vortex might also result in reduced eddy ozone and heat transport from lower latitudes.

Two types of ozone anomalies have been observed in the Arctic over the past decade: short term type 1, lasting a few days, and more long term type 2 lasting for several weeks (Taalas *et al.* 1995, 1996, 1997). For both types, the losses in total ozone have exceeded 35%. The location of ozone anomalies in relation to the location of the polar vortex and the stratospheric temperatures provide information as to the possible cause of the anomalies. While these two categories for describing ozone anomalies are useful, all ozone depletion events are not easily divided into dynamic or chemically driven, nor inside or outside of the vortex.

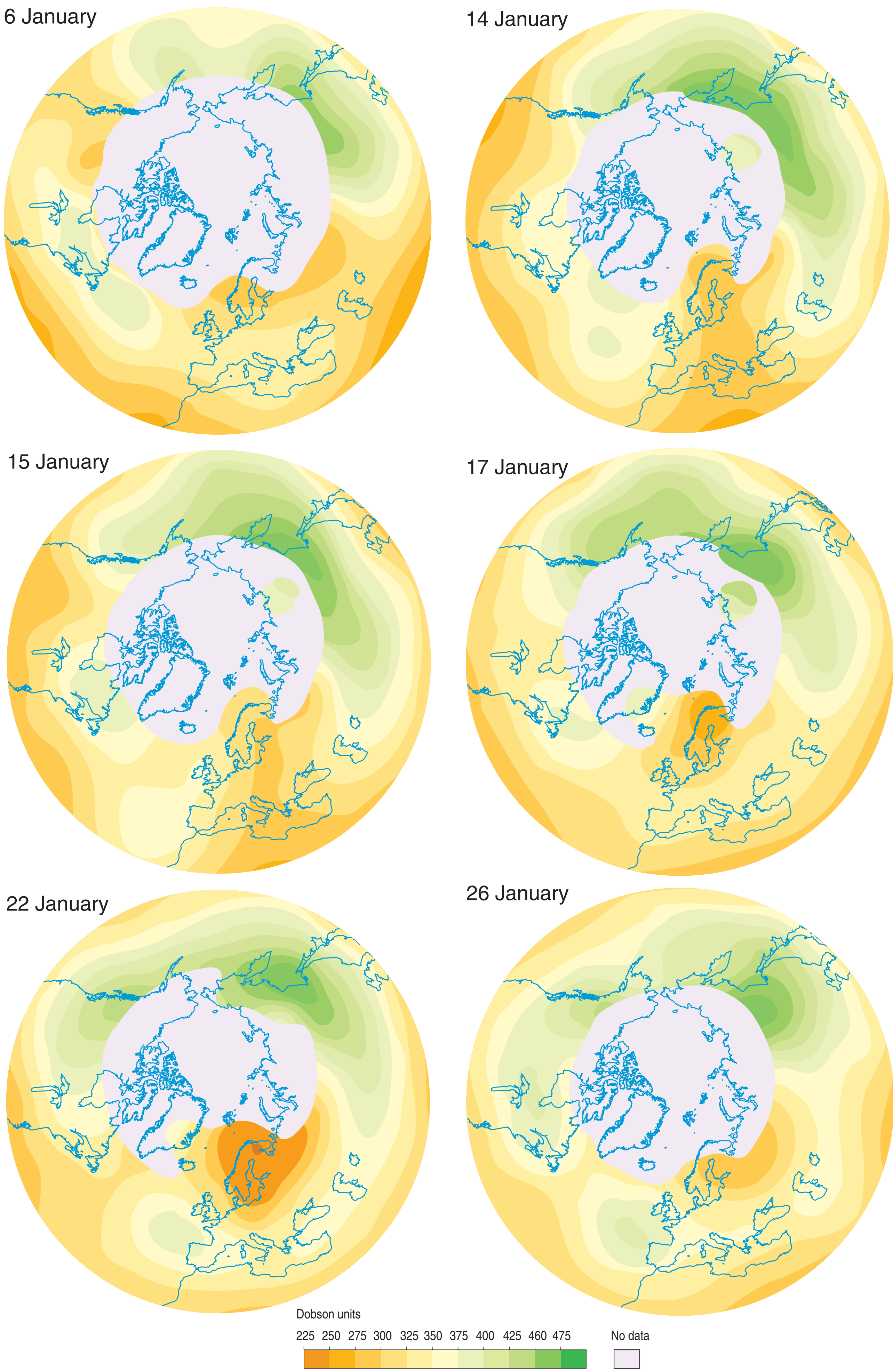
Finnish data from Sodankylä have been compared with the long-term (1935-1969) total ozone means from Tromsø, as recalculated by Bojkov *et al.* (1995a). Anomalies of 10% and greater from those at Sodankylä were chosen for further analyses. The anomalies were sorted according to stratospheric temperature observed at Sodankylä. Most of the anomalies observed in the European Arctic during recent years are connected to 'warm' stratospheric temperatures. This may mean low probability for the occurrence of Polar Stratospheric Clouds (PSC's) in the airmass, which suggests a considerably different mechanism for ozone loss in the Arctic compared with the Antarctic. For example, substantial chemical ozone loss in the 1993/94 Arctic winter occurred mainly during an unusually cold ten-day period in late February and early March (Manney *et al.* 1995c). According to a detailed study of inter-hemispheric differences using UARS data (Santee *et al.* 1995), although ClO is enhanced over the Arctic as well as over the Antarctic, denitrification and subsequent rapid ozone loss is inhibited in the Arctic due to the generally higher temperatures. Future cooling of the Arctic lower stratosphere could lead to more severe anomalous ozone losses in the northern spring.

It is unclear whether ozone anomalies have an impact on trends. Longitudinally averaged trends may obscure the impact of these events and it is even possible that no change will take place in the monthly or yearly mean ozone levels, despite the occurrence of isolated events of severe ozone depletion. Nevertheless, the timing and severity of the ozone anomalies may have significant impacts on biological activity. These issues are addressed in sections 11.5.1.2 and 11.5.2.3.

11.3.5.1. Type 1 Arctic ozone anomaly

Type 1 Arctic ozone anomalies are dynamically induced, generally as a result of low ozone air masses transported northward from mid-latitudes. These anomalies usually occur outside of the Arctic polar vortex and are generally short-lived, lasting less than two weeks, and often less than a few days. The complicated nature of these events is illustrated in Figure 11·15, which shows the evolution of a type 1 ozone anomaly. The anomaly is formed by the intrusion of low-ozone air from lower latitudes, which then becomes isolated inside the polar vortex, where chemical reactions cause a further reduction in the ozone concentrations.

→
Figure 11·15. Development of a type 1 ozone anomaly. The series of satellite images show total column ozone for a period of days spanning 6-26 January 1996. The anomaly was primarily caused by the dynamic atmospheric circulation, and then augumented by chemical reactions. January 6: a 'normal' Arctic image with typically higher ozone (green colours) nearer the pole and an assymetric distribution around the pole due to the unstable polar vortex. January 14-15: insurgence of lower-ozone air (orange colours) from mid-latitudes coming northward over central Europe. January 17: the migrated low-ozone air is isolated by the polar vortex. January 22: the ozone concentration within this isolated air mass (most visible over Scandinavia) is lowered to approximately 200 Dobson units, presumably due to chemical destruction. January 26: the dissolution of this ozone event after a few days. (Source of data: SBUV-2 on Nimbus 14).


6 January
14 January
15 January
17 January
22 January
26 January
Dobson units
225 250 275 300 325 350 375 400 425 460 475
No data

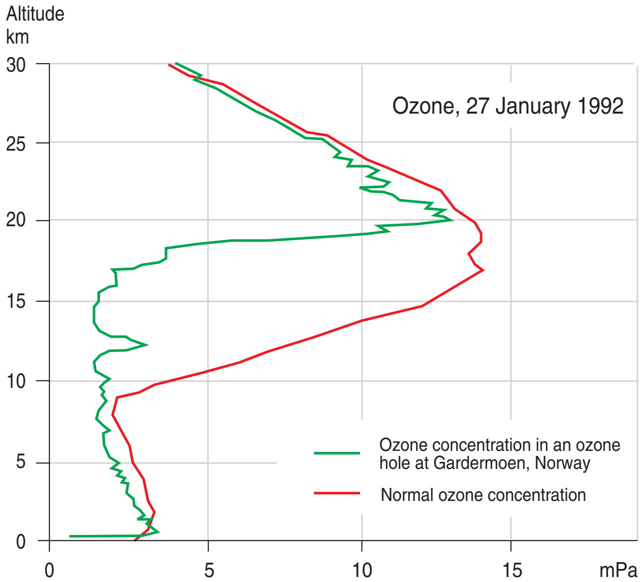


Figure 11·16. Ozone soundings at Gardermoen, southern Norway, 27 January 1992. The green line denotes the partial pressure of ozone observed during the (type 1) anomaly, when low-ozone air passed over the area. The red line indicates the 1989-91 January mean ozone profile at Sodankylä, Finland.

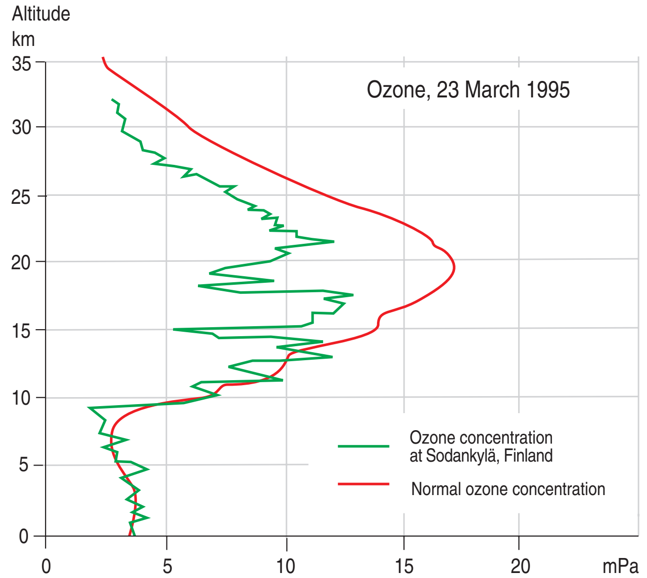


Figure 11·17. Ozone soundings at Sodankylä, Finland, 23 March 1995. The green line denotes the profile of partial pressure of ozone observed during the (type 2) anomaly. The red line shows the monthly mean ozone profile for March.

Another example of a strong total ozone anomaly with a minimum of 196 Dobson Units (DU) was observed moving eastward over the European Arctic during the period 25-29 January 1992 (Taalas *et al.* 1995). The diameter of the anomaly was approximately 2000 km. The potential vorticity analyses for the 475 K and 380 K surfaces indicate an air mass with low ozone values, and a 3-dimensional trajectory analysis for the lower stratosphere indicated a subtropical origin for the airmass before reaching Gardermoen in southern Norway. An ozone sounding performed while the anomaly area was passing over Gardermoen (Figure 11·16), showed that the primary ozone loss was observed in the lower stratosphere below the ozone maximum and well below the levels where low temperatures are typically observed.

Similar features may be found to have been behind the other strong deviations observed in the European Arctic during early 1992 (Taalas and Kyrö 1994). These findings demonstrate the importance of large-scale circulation as a modifier of the global ozone distribution in contrast to the dominant chemical ozone loss observed in Antarctica.

11.3.5.2. **Type 2 Arctic ozone anomaly**

Type 2 Arctic ozone anomalies are primarily chemically induced and are generally observed in connection with a strong polar vortex and the presence of Polar Stratospheric Clouds (PSCs). In 1995 a cold polar vortex was observed over the European Arctic until late March. Both PSC's and ClO were observed inside the Arctic polar vortex and severe ozone depletion was measured. The resultant ozone loss was recorded with a variety of techniques including an ozone sounding made inside the polar vortex (Figure 11·17). Analyses of temperature and potential vorticity indicate that the ozone anomaly was located inside the polar vortex at temperatures below 195 K (Taalas *et al.* 1997). The ozone loss was strongest in the cold airmass between 17 and 21 km altitude, with maximum negative ozone deviation exceeding 60% from the 1989-1991 March mean partial pressure of ozone in the same altitude range. Figure 11·18 shows an example of a type 2 ozone anomaly when record low ozone levels of 30-35% below the long-term mean were recorded over Siberia from February-March of 1995 (Bojkov *et al.* 1995b).

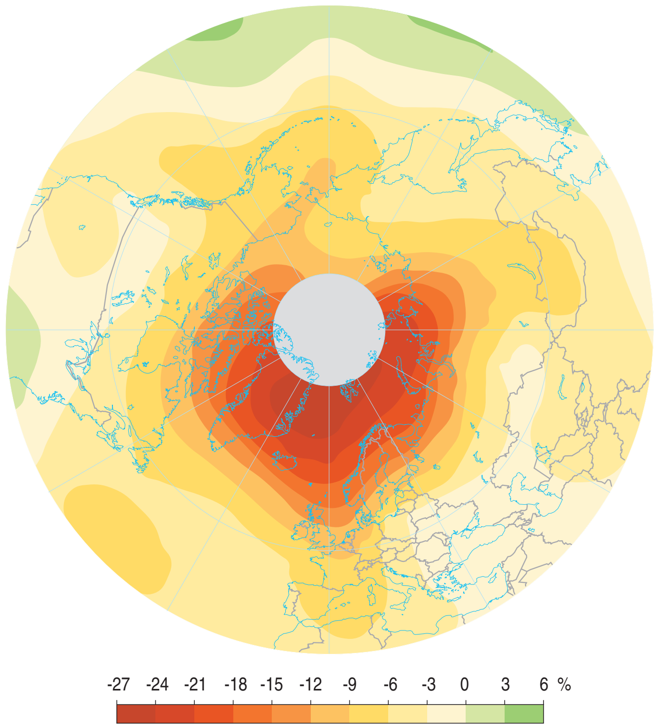


Figure 11·18. An example of a type 2 ozone anomaly, showing levels of ozone depletion relative to the long-term mean.

11.4. **UV radiation**

The most direct effect of changes in ozone is a change in UV radiation reaching the biosphere. Despite the low sun angle in the Arctic, UV radiation has always been an environmental threat, even before ozone depletion. Snow blindness (photokeratitis), which is directly caused by UV radiation, is an environmental risk which indigenous Arctic populations have always needed to protect themselves against. Recent changes in the Arctic ozone layer have heightened concern about the effects of UV radiation on indigenous peoples and on the environment in which they live (Wulf 1994). Direct and diffuse radiation both affect biological systems, and the amount of

UV and visible radiation at the earth's surface is highly variable due to many factors. The amount of ozone, and hence UV, can vary greatly from day to day, and cloud cover, including Arctic haze, can cause variations on even shorter time scales. In the Arctic, decreases in ozone in the winter result in increased UV during a time when many plants are protected by snow and the sun is low on the horizon; springtime decreases in ozone, however, are likely to result in UV increases during critical phases of biological productivity. Recently observed ozone anomalies have occurred in the spring when some biological organisms are most sensitive.

UV radiation is composed of all radiation between 100 and 400 nm. It is further broken down into sub-categories:

UV-A 315-400 nm
UV-B 280-315 nm
UV-C 100-280 nm

Visible radiation, or Photosynthetically Active Radiation (PAR), is above 400 nm.

Variations in stratospheric ozone modify the amount of solar UV radiation available for absorption in the atmosphere and at the surface, and lead to changes in a) atmospheric composition through altered photochemistry, b) atmospheric circulation through changes in warming/cooling rates, and c) terrestrial and aquatic ecosystems through modulations in biologically effective UV radiation reaching the biosphere (IASC 1995). The response of the atmosphere-hydrosphere-biosphere system to such photochemical and radiative changes is not well understood.

Surface albedo, clouds, total column ozone, vertical distributions of ozone, pollutants, and Arctic haze can affect the transmission of UV through the Arctic atmosphere (Tsay and Stamnes 1992). The effects are not always apparent because the diffuse and direct components of the radiation field can be differentially altered. Henriksen *et al.*(1989) demonstrated clearly that diffuse radiation comprises the majority of the total UV radiation at the surface in the summertime Arctic. For low solar elevations, stratospheric aerosols may lead to an increase in UV dose rate, while Arctic haze results in a decrease. A redistribution of ozone from the stratosphere to the troposphere also tends to decrease UV exposure, in agreement with previous findings of Brühl and Crutzen (1989) and Frederick *et al.* (1989), except for low solar elevations when an increase may instead occur (for a physical explanation, see Tsitas and Yung 1996).

UV and visible irradiance at the surface depend strongly on surface albedo for clear, aerosol-loaded atmospheres as well as cloudy atmospheres. Snow is one of the few natural materials whose reflectance (local albedo) is high enough to substantially increase the UV radiation reaching the ground. For fresh snow the UV reflectance may be as high as 94% (Blumthaler and Ambach 1988, Grenfell and Warren 1994). In other studies, albedos for snow range from 0.76 (Feister and Grewe 1995) to 0.97 at 300 nm for the South Pole (Grenfell and Warren 1994). This is considerably higher than the albedo for open water and vegetation (McKenzie and Kotkamp 1996).

Under cloudy conditions, multiple reflections may occur between the snow covered surface and the clouds, increasing UV considerably from that which would have occurred if the surface were not snow covered. Perovich (1993) made laboratory and field observations of the reflection of ultraviolet radiation by young and first-year sea ice (Figure 11·19). Ultraviolet albedos for cold, bare, first-year sea ice are between 40 and 60%. The presence of even a thin snow cover was found to significantly increase the ultraviolet albedo and greatly reduce the transmission.

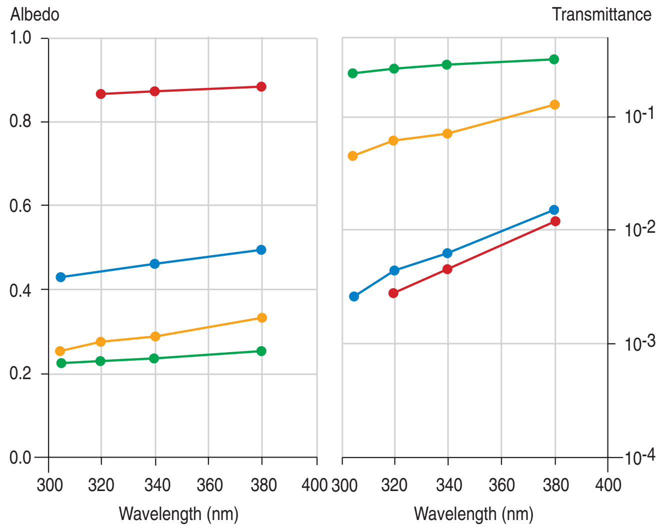


Figure 11·19. Ultraviolet albedo and transmittance values for A) thin (0.3 m) bare sea; B) cold 1.6 m thick ice with a 0.1 m deep snow cover in April; C) bare 1.7 m thick ice in May; and, D) ponded ice with 0.1 m of water over 1.5 m thick ice. A was measured in the laboratory and B, C and D were measured in shorefast first-year ice near Barrow, Alaska (Perovich 1993).

The contribution from the surface to the downward irradiance (due to multiple reflections between the atmosphere and the surface) depends on the spherical albedo, which is the same as the flux reflectance for uniform illumination of the atmosphere from below (Stamnes 1982). For a cloudy atmosphere, spherical albedo is large and gives rise to a nonlinear behavior of the downward irradiance as a function of surface albedo (Kylling *et al.* 1995). This yields a much more pronounced dependence on surface albedo for cloudy than for clear atmospheric conditions.

11.4.1. Measurements

Ground-based measurements of UV radiation are the best method available for assessing UV levels. Unfortunately, these measurements are difficult, expensive, and cannot be made throughout the Arctic. A variety of UV measurements have been taken in the Arctic for many years (e.g., Wester 1996, Hisdal 1986, Stamnes *et al.* 1988). Solar UV measurements have been the subject of several reviews (cf. Blumthaler 1993), but in the early 1990s, few national programs for monitoring UV irradiance included Arctic installations.

Several types of instruments are presently used to measure UV in the Arctic. Spectrometers and photometers have been used in the Arctic since 1927 (Dobson 1968b) to make measurements of the direct solar beam, or the zenith sky. These contrast with radiometric UV measurements where the intensity of UV radiation is measured in Watt/m^2 or Watt/m^2/nm or in some other related measure such as a biological dose. UV measurements can be further divided into narrow-band that are spectrally resolved (e.g., by spectroradiometers) and broad-band meters that present a single measure, such as 'UV-B' or 'erythemal dosage'. Spectroradiometers have the advantage of being able to compute a wide variety of single measures, allowing separation of the effects of changes in cloud cover or atmospheric clarity and total column ozone (Stamnes *et al.* 1991), while single measure devices do not.

There is little agreement in the international community on the kind of instrument to use, nor is there a single standardized spectroradiometer or even a standardized operating protocol or calibration method. Several national and inter-

Table 11·5. UV monitoring sites inside the Arctic Circle. Most UV monitoring efforts in the Arctic were started within the last decade, making long-term assessments of UV difficult. The lack of sites in the Russian Arctic limits our ability to monitor changes in Arctic UV levels today.

Location	Coordinates	Operator	Instrument	Established
Tromsø, Norway	70°N, 19°E	University of Tromsø	JY32D, GUV-511	1987
Ny-Ålesund, Norway	79°N, 12°E	NPI	GUV-511, SL500, SL501	1990
Longyearbyen, Norway	78°N, 16°E	Univ. Tromsø	JY H32D	1991
Barrow, Alaska	71°N, 156°W	NSF, (Biospherical)	SUV-100	1990
Resolute, Canada	75°N, 95°W	AES Canada	Brewer	1992
Alert, Canada	82°N, 62°W	AES Canada	Brewer	1992
Eureka, Canada	80°N, 86°W	AES Canada	Brewer	1993
Thule, Greenland	76°N, 69°W	DMI	DMI (.22 m double)	1994
Søndrestrøm, Greenland	67°N, 51°W	DMI	Brewer MkII	1990
Abisko, Sweden	68°N, 19°E	Univ. Lund	SL501	
Kiruna, Sweden	67°N, 21°E	SMHI	SL500	1989
Sodankylä, Finland	67°N, 27°E	FMI	Brewer MkII	1989
Varrio, Finland	67°N, 30°E	University of Helsinki	SL501A	1995

national UV instrument intercomparisons have been held to address problems with measuring UV (Koskela 1994, McKenzie *et al.* 1993, Seckmeyer *et al.* 1994, Gardiner and Kirsch 1995). There are several national UV monitoring networks that include Arctic sites (see Table 11·5) (Kerr and Wardle 1993, Johnsen 1996). The Nordic countries, Sweden, Finland, Norway, Denmark, and Iceland, are organized into the Nordic Ozone Group; each country operates its own UV monitoring network with strong cooperation between the scientists.

11.4.2. Modeling

Models of UV radiation, using satellite data to estimate the relevant components of the atmosphere, provide estimates of UV in regions where UV monitoring does not exist. However, the complexity of the atmospheric and environmental conditions makes estimates difficult in the Arctic, particularly in partly cloudy situations. Considerable progress has been made in understanding the relationship between atmospheric conditions and the UV irradiance incident on the ground for mid-latitudes. But, the limitations in predicting or modeling ground level UV are greatest in polar regions where a combination of uncertainty in albedo due to changing snow or ice conditions, frequent cloud cover, and the increased uncertainty in radiative transfer models at low solar angles greatly complicate the estimation of surface UV using satellites. The use of ground-based UV measurements is crucial to the development of reliable algorithms for using satellite information to estimate UV levels in the Arctic.

One of the least understood components of this complex system is the radiative effects of particulate clouds, as well as the environmental conditions leading to their formation and dissipation. The effects of molecular multiple scattering, solar elevation, surface albedo, and cloud properties (such as optical depth, altitude, and cloud fraction) on ozone absorption have been studied using radiative transfer models of varying degrees of sophistication (Shettle and Green 1974, Luther and Gelinas 1976, Spinhirne and Green 1978, Nicolet *et al.* 1982, Thompson 1984, Madronich 1987, Frederick and Lubin 1988, Frederick and Snell 1990, Smith *et al.* 1992, Stamnes *et al.* 1990, 1992).

11.4.3. Biologically relevant UV

The present understanding of the relationship between the UV climatology and biological exposure in snow covered regions is inadequate to fully assess the impacts of ozone depletion on the Arctic. Three factors make assessing biologically relevant UV difficult in the Arctic: a) the high surface albedo; b) extremely low sun angle; and c) many of the biological receptors in the Arctic are under snow, water, or ice. Because different biological processes respond differently to the various components of UV-A and UV-B radiation, it is difficult to summarize UV radiation for biological use. Furthermore, standard methods of monitoring UV are inappropriate for research on UV effects in Arctic conditions.

11.4.3.1. Spectral considerations

The total amount of UV radiation does not adequately describe the amount of harmful UV radiation because of the differing efficacies of photons within the UV spectrum: the shorter wavelength photons are several orders of magnitude more damaging than the longer wavelengths (UNEP 1994, SCOPE 1993). The relative harmfulness of the various regions within UV may be summarized by a Biological Weighting Function (BWF) which is unique to the process being studied. Biologically relevant UV is determined by weighting the spectral radiation by its biological effectiveness.

Biologically effective UV can be summarized in terms of its response to changes in ozone. At mid-latitudes, a change of one percent in ozone may result in a change of between one and three percent in biologically weighted UV, depending on the action spectra being considered. The percent increase in UV dose rate per percent decrease in ozone amount (radiation amplification factor) is largest (and tends to increase with solar zenith angle) for biological responses which are heavily weighted toward UV-B radiation (e.g. plant response), while biological effects with significant response to UV-A radiation (e.g. erythema) will have a smaller amplification factor that decreases with increasing solar zenith angle. Large changes in ozone result in non-linear changes in UV (Booth and Madronich 1994). These relationships were derived by examining the relationship between ozone and UV at mid-latitudes; further work needs to be done to determine the relevance of these relationships in polar regions.

11.4.3.2. Geometrical considerations

To reach biological receptors in the Arctic, UV radiation must often travel through snow, water, or ice. Even biological receptors which are on open land require special considerations in parameterizing UV radiation. Traditionally, the exposure of human populations and ecosystems to solar UV radiation is commonly described using measurements of the biologically weighted irradiance hitting horizontal surfaces which is at best an approximation because UV hits biological organisms at many different angles. New methods of parameterizing UV become particularly important when trying to assess UV effects in the Arctic. Previous methods have proven inadequate and vastly under-estimate the impact of UV on Arctic ecosystems and human inhabitants because of the increased fraction of UV radiation that is scattered relative to direct UV radiation.

11.4.4. UV on land

Measurement of UV striking a horizontal surface is not always appropriate to describe UV irradiance in the Arctic. For many biological systems, particularly in the Arctic, the UV reaching a vertical surface which is rotated over 360° to give the average irradiance reaching a vertical surface may give a more relevant measure of biologically important UV. The difference between UV reaching a horizontal and a vertical surface can be particularly critical in the Arctic when snow cover is present. In environments of high UV reflectivity and especially when the terrain is open and not covered by trees, buildings, etc., UV exposure to the face and eyes is strongly affected by reflected UV. Under these conditions, measurements of exposure on vertical surfaces give good estimates of the exposure received by different parts of the face and eyes.

Theoretical models are available for measuring UV on a vertical surface (Dahlback and Moan 1990, Hay and Davies 1978, Bird and Riordan 1986, Schauburger 1992), and have been used to estimate the effect of ozone depletions on skin cancer (Moan and Dahlback 1993) and the increase of UV due to the reflection from snow (Jokela *et al.* 1993, Jokela *et al.* 1995). Sliney (1986, 1994) has studied the different environmental and anatomical features contributing to UV exposure of the eye. The results from these studies may be applicable to UV effects on other biological systems. Simple algorithms have been proposed (Hay and Davies 1978, Bird and Riordan 1986) and tested (Jokela *et al.* 1993) for converting irradiance on a horizontal surface, as is typically measured, to irradiance on a vertical surface for a given albedo. These algorithms are particularly useful in the high-albedo, low-sun-angle conditions often found in the Arctic.

While UV reaching a vertical surface may give a more accurate dose than UV on a horizontal surface for some biological systems in the Arctic, measurements of biological exposure are complicated further by natural protective mechanisms. Snow has an even greater effect on radiation reaching the eyes than that reaching the face due to the anatomy of the head and eyes. The viewing angle of the eyes, which varies according to the amount of squinting, is considerably less than 30 degrees vertical above the horizon and less than 180 degrees in azimuth, and the eyes normally look toward the ground or horizon (Sliney 1986, 1994). In bright sunlight over snow, the squinting of the lids mimics Inuit slit goggles and allows only a narrow vertical field of view of a few degrees. This form of protection may be more effective than many modern types of sunglasses which allow considerable scattered light into the eye from the side.

Because ocular damage by UV radiation has always been a health hazard in the Arctic, ozone depletion is likely to raise the importance of UV as an environmental health concern for the Arctic populations. Figure 11·20 illustrates the effect of snow on erythemally effective UV received by horizontal and vertical surfaces in northern Finland (Saariselkä, 68.4°N, 27.5°E) (Taalas *et al.* 1996). The irregularly plotted line shows the results computed by using the actual total ozone measured by the nearby Meteorological Observatory in Sodankylä (67.4°N, 26.6°E). For the winter, a reflectivity of 0.8 was adopted, which decreased linearly to the summer value of 0.04 during the period of 25 days before the end of the snow season (16 May). In the autumn the change was assumed to be abrupt (24 October).

In Saariselkä, snow albedo increases total UV exposure to the face in winter and early spring by a factor greater than two. Exposure to these surfaces in April is equal to midsummer exposure. Snow has a much smaller effect on UV dose reaching a horizontal surface than on the UV dose reaching a vertical surface. In southern Finland, the effect of ground reflectivity on both horizontal and vertical exposure is relatively small because the snow melts before UV increases to significant levels. Also, the terrain is covered by trees and buildings, which effectively reduce reflectivity. The snow albedo effectively smooths out geographical differences in the dose of UV: the computed dose on a horizontal surface is about 29% higher in Helsinki than in Saariselkä, while the difference is only 12% for UV on a vertical surface.

UV has been thought to be considerably less in the Arctic than at mid-latitudes. However, when one examines the difference in UV to a vertical surface as a proxy for UV reaching a human face or eyes, as a function of latitude, the difference between mid-latitudes and the Arctic is not as great as one might expect. Figure 11·21 shows a comparison of the computed annual UV doses to vertical and horizontal

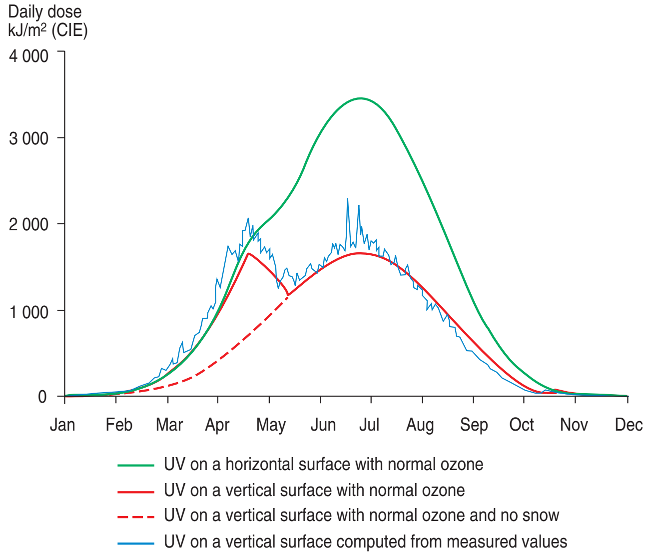


Figure 11·20. Theoretical clear-day UV dose to horizontal and vertical surfaces, for Arctic Finland. The amount of energy from UV radiation reaching a horizontal surface peaks around midsummer, while that reaching a vertical surface peaks in early spring, in part due to efficient reflection by snow. The irregular line shows the results obtained by using measured values for total ozone. The smooth lines were calculated with the mean (climatological) total ozone. The dashed line depicts the mean vertical exposure with summer albedo (0.04). The theoretical curves are realistic only for clear skies with fresh snow.

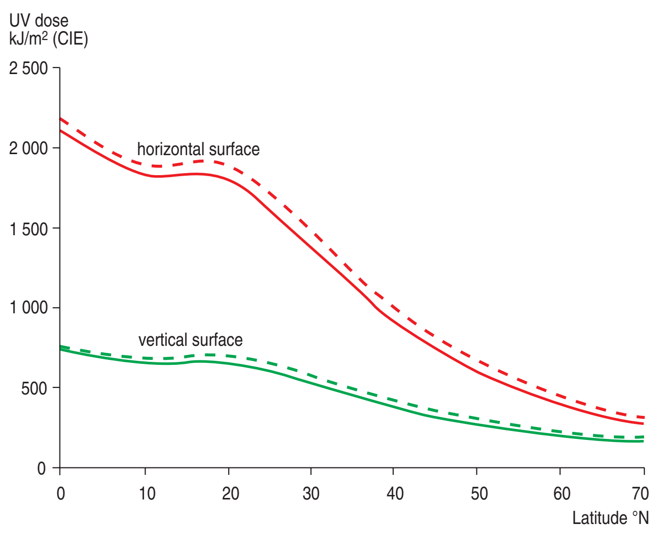


Figure 11·21. Average annual UV dose falling on vertical and horizontal surfaces as a function of latitude. The solid lines represent the situation for the normal UV level with undisturbed ozone (prior to 1978). The dashed lines are for the UV level following the maximum estimated global ozone depletion (in the year 2000).

surfaces as a function of latitude. Additionally, the estimated effect of global ozone depletion is shown (Taalas *et al.* 1996). Annual doses were corrected for the effects of clouds (Frederick and Lubin 1988) and reflection from the snow at 60°N and at 70°N. The latitudinal difference of the ratio of vertical to horizontal dose is 0.35 at the equator and increases to 0.5 at high latitudes. This is mainly due to the relative increase of diffuse radiation from the sky and ground compared with the direct radiation from the sun.

The relative increase of the UV dose on a vertical surface at high latitudes may have some influence on the relationship between the ozone depletion and the incidence of skin cancers at different latitudes. Atmospheric, epidemiological and animal studies show that the relative incidence of a skin carcinoma increases exponentially:

$$I_1/I_2 = (D_1/D_2) \times BAF = (O3_1/O3_2) \times RAF \times BAF$$

where I_1/I_2 is the relative cancer incidence, D_1/D_{22} is the relative annual UV dose, $O3_1/O3_2$ is the relative total ozone. Index 1 refers to the depleted and index 2 to undepleted conditions. BAF is the Biological Amplification Factor and RAF is the Radiation Amplification Factor. When the ozone depletion is below 10%, the amplification factors are approximately the factors by which the percentage change of the UV dose or ozone depletion must be multiplied in order to obtain the percentage change in the cancer incidence or UV dose, respectively. For Squamous Cell Cancer (SCC) the current estimate for the BAF is BAF = 2.5 ± 0.7 (de Gruijl and van der Leun 1993). The erythemally effective UV radiation RAF decreases from approximately 1.2 to 1 from the equator to high latitudes. The BAFs have been derived from the measured erythemal horizontal UV at different latitudes.

Estimates of vertical BAF can be made by making assumptions about the angular distribution of radiation and applying a conversion factor derived from the calculations of annual doses on both horizontal and vertical surfaces. This conversion results in the increase of the BAF for the SCC from 2.5 to 3.2. It is interesting that the increased BAF is closer to the murine BAF, which varies from 3.2 to 4.4 (de Gruijl and van der Leun 1993). The increase of the BAF based on the UV reaching a vertical surface is in qualitative agreement with the results of Moan and Dahlback (1993), which examined the dependence of the amplification factors from the receiving surface for the Norwegian population. They found that the RAF does not significantly depend on the exposure geometry, but the total amplification factor increases slightly from the exposure to horizontal and vertical surfaces. Since the total amplification factor is the product of RAF and BAF, the latter must increase. Although the irradiation on a vertical surface describes better the actual UV exposure than the horizontal, it is still a very crude approximation of the complex exposure patterns.

11.4.5. UV penetration in aquatic systems

The penetration of radiation in the PAR and UV regions in ocean and lake water depends on dissolved substances and particles that attenuate light. Natural levels of UV-B radiation are known to penetrate to ecologically significant depths in many bodies of water (cf. Smith *et al.* 1992, Scully and Lean 1994, Morris *et al.* 1995, Hessen 1996). Increases in UV-B radiation due to ozone depletion alter the spectral balance of UV-A, UV-B, and PAR and increase the exposure of aquatic ecosystems to UV radiation (Smith *et al.* 1992).

The maximum penetration of UV-B is found in oceanic waters, while in coastal seawater the penetration is less due to a larger amount of Dissolved Organic Matter (DOM), especially humic substances (Smith and Baker 1979). For the estuarine areas of the Arctic Ocean this is of major importance (Höjerslev and Aas 1991). In the waters off Svalbard and western Greenland, diffuse attenuation coefficients representing 10% radiation levels at 310 nm ($K_{d(310)}$) were shown to be around 0.2, or a little more than 10 m, while in the North Sea the corresponding irradiation levels were at a depth of around 5 m and in Hardangerfjord, Norway levels were 1-2 m (calculated from Højerslev and Aas 1991).

Theoretical estimates indicate that the presence of a sea ice cover causes a significant reduction in the amount of UV transmitted and that UV-B radiation is attenuated more than UV-A radiation or visible light. Even a modest snow cover (≈0.15 m) reduces UV levels by nearly two orders of magnitude (Perovich 1993). For young, thin sea ice UV transmittance is large, roughly 20%, while for cold first year ice transmittances are less than 2%. Extinction coefficients for cold bare ice are between 2.5 and 3 per meter, in the range of visible values observed for the interior of white ice and the surface scattering layer of white ice (Grenfell and Maykut 1977). In all cases, a larger portion of PAR is transmitted than of the UV radiation. Wavelength-integrated extinction coefficients for PAR are smaller than UV values.

Models for solar radiative transfer through the coupled atmosphere-snow/ice-ocean system have also been developed (Jin *et al.* 1994a, 1994b, Zeng *et al.* 1993). However, data on optical properties for snow, ice, and water in the UV are scarce (see Trodahl and Buckley 1990) and few spectral irradiance measurements exist for testing of such models. Simultaneous spectral measurements of Arctic surface (snow and ice) and underwater UV optical properties, as well as UV irradiance and radiance measurements, are needed. Progress awaits comprehensive experiments executed in the Arctic environment.

Recently there has been growing concern about the impact of enhanced UV levels on ice biota in the Arctic as well. Diatom mats cover the undersurface of the ice over vast expanses and form the basis of a food chain that extends from copepods and amphipods to cod, seals, and polar bears. Determining how much enhanced UV radiation at the surface reaches biological organisms living in and under the ice cover is difficult. This problem is complicated by the temporal and spatial variability in the physical and therefore optical properties of sea ice.

Penetration of UV in lakes and ponds varies greatly as a function of amount of dissolved organic matter; ice and snow cover also absorb UV-B for much of the year. Hessen (1996) found 10% attenuation at 10 m for 310 nm in alpine, subarctic lakes, while Arctic ponds that are slightly more influenced by Dissolved Organic Carbon (DOC) have a 10% attenuation at a depth of 2-4 m for 310 nm. A number of these lakes have maximum depths of less than 2 m. For Arctic lakes of Alaska, more influenced by humic compounds, Morris *et al.* (1995) also recorded low attenuation depths relative to high UV-B transparency in clear alpine lakes.

There is a limited number of underwater measurements from Arctic and subarctic sites (Scully and Lean 1994, Hessen 1993, 1996). In general, high latitude and high altitude localities are low in DOC, implying low absorbance of UV-B. Some alpine lakes are extremely UV transparent (Hessen 1993, Schindler *et al.* 1997), while High Arctic localities at 80°N span from extremely clear lakes to shallow ponds slightly influenced by colored organic matter (Hessen 1996).

11.5. Effects of climate change and UV radiation on the biosphere

Climate change will affect biological processes and physical properties of the oceans and land, directly impacting the lives of humans living in Arctic regions. The most obvious effects of warming in the Arctic will most likely involve changes to the cryosphere (glaciers, permafrost, sea ice cover, and snow), with consequent changes in biotic and biogeochemical processes. Accurate predictions of change are complicated by complex interactions: on land between permafrost, the feedback effects of soil changes on vegetation, and alterations in snow cover and precipitation; in the atmosphere between trace gas amounts, changes in ozone and atmospheric pollutants, and seasonal temperature and light regimes; and in the oceans between the sea ice albedo feedback, and changes in sea ice extent and thermohaline circulation.

Temperature and precipitation are not the only environmental variables that are changing and the balance between opposing responses to different environmental factors is not clear yet. Increases in CO_2 and UV radiation have the potential to change the chemical quality of plant tissues which might reduce microbial decomposition rates by as much as 10% (Couteaux *et al.* 1991, Gehrke *et al.* 1995) and slow the release of carbon to the atmosphere. Increased UV coupled with climatic warming and acidification may increase the rate at which DOM is cycled in fresh water and increase the CO_2 flux to the atmosphere (Schindler *et al.* 1997). Even subtle changes in plant community distribution may have effects on canopy roughness and the turbulent transfer of energy and water between vegetation and atmosphere. Effects of climate change and UV are often counterintuitive and long term *in situ* studies are needed.

The Mackenzie Basin Impact Study (MBIS), which consisted of 30 research activities and consultations with stakeholders, represents one of the first attempts at an integrated assessment of the impact of climate change on a region and its inhabitants (Cohen 1996, 1997). Climate warming scenarios of 4-5ºC were obtained from three GCMs and a composite analogue based on paleoecological data. Results from two studies from the ecosystem component on boreal peatlands are provided in Nicholson and Gignac (1995) and Nicholson *et al.* (1996).

11.5.1. Terrestrial ecosystems

11.5.1.1. Climate change effects on terrestrial ecosystems

The impacts of climate change on Arctic terrestrial ecosystems are complex and difficult to predict because of the many interactions which exist within ecosystems and between many concurrently changing environmental variables (Chapin *et al.* 1992, Callaghan and Jonasson 1995). Primary production can be affected by higher temperatures, atmospheric CO_2 concentrations, and UV-B radiation; impacts from such changes are uncertain and depend on the responses of soil microbial decomposer activity to changes in soil temperature, moisture, and plant litter quality. Increases in atmospheric CO_2 are expected to increase plant production, increase translocation of carbon to below ground, reduce nitrogen and increase lignin and tannin concentrations in plant litters. Some of these changes are closely tied to UV-B increases. Thawing permafrost can disturb the soil, which would warm and further disturb vegetation, resulting in a change in species composition. A change in species composition due to competition effects from less sensitive local species and from intruding southerly species is also expected (Chapin *et al.* 1992, Körner 1994).

Many Arctic plant species are currently stressed by the harsh environment. Vegetation is subject to periodically frozen soil, strong winds and ice crystal abrasion, short growing seasons, and extreme temperature variations. Microbial decomposer activity is limited by cold, shallow soils overlying permafrost so that few nutrients are available for plants. Soils are often either dry as in the polar deserts, or waterlogged and anaerobic as in mires and polygonal sedge meadows; much of the year they are frozen. Changes in air temperature will warm the soil and will induce changes in the depth and moisture of the active layer where hydrologic, geomorphic, chemical, and biological processes occur that profoundly affect ecosystems at the landscape level (Reynolds and Tenhunen 1996). In many systems, warming may increase soil biological processes of trace gas emission, decomposition of soil organic matter, and nutrient availability to plants but the resulting increase in leaf area index may reduce the penetration of solar radiation to the soil thereby negating longer term soil warming or even resulting in soil cooling (Callaghan and Jonasson 1995). Increased soil moisture content in already wet soils may reduce the rates of these processes through a shift in the balance of aerobic and anaerobic conditions.

Harsh Arctic conditions have acted as a filter to biodiversity, which is further constrained by the relative youth of most tundra ecosystems. While biodiversity is low at the species level (Chapin and Körner 1995), populations are often large and geographical distributions extensive which should mitigate against losses in biodiversity due to climate changes. Arctic species have successfully survived changes in the past, such as climatic warming in medieval times, the recovery from the Little Ice Age over the past 150 years, and considerable variability in temperature between years in the present period (Havström *et al.* 1995).

11.5.1.1.1. Vegetation

Plant communities

Within the Arctic, the variation in vegetation ranges from closed, relatively productive communities of the subarctic which support a diversity of herbivores, to sparse, unproductive vegetation of the High Arctic. This geographical variation in structure of plant communities and environments within the Arctic will result in different responses to future climatic changes by different plant community types. Although Arctic organisms can survive current harsh environments, many can also tolerate higher temperatures expected to occur following climatic warming.

Plants exhibiting traits of slow growth, compact forms, and parsimonious nutrient use are predisposed to tolerate Arctic conditions but are also sensitive to competition from opportunistic, aggressive species from farther south. In a warmer Arctic, the competition from more responsive species would probably threaten Arctic flora more than the environment *per se* because many Arctic plants are sensitive to shade and other competitive mechanisms (Callaghan and Jonasson 1995). However, Arctic latitudes are broad and the migrations of plants from southern latitudes will occur more slowly than temperature changes. For example, a warming of 2°C could result in a 4-5° latitude northward shift of the climatic zone associated with the taiga of Eurasia (Velichko *et al.* 1990), i.e., a shift of 400-500 km by the year 2020. If the taiga forests could migrate at this rate, tundra would be totally displaced from the Eurasian mainland. However, expected tree migration rates of 10-1000 meters per year are insufficient to accomplish this. The disparity between rates of predicted climate change and tree migration will result in some tundra areas experiencing supra-optimal thermal regimes which could adversely affect plant carbon balance and survival.

At a much smaller scale, those genotypes, or ecotypes, within individual species which are more responsive to change may out-compete those best adapted to the harshest environments. Different ecotypes of the same species can grow close to each other but have different distributions; some, for example, prefer sheltered, moist depressions, while others prefer more exposed sites. (Crawford *et al.* 1995) Again however, slow rates of spread, i.e. small scale migration, will limit change. Plants of the harshest Arctic environments, sparsely located in a landscape, growing at 2 mm per year (Molau 1996), may take centuries to replace a neighboring genotype.

Nutrient availability

In many tundra ecosystems the response of plants to climate changes will be influenced by the availability of nutrients, including those released by decomposition of organic matter and mineralization from parent materials. Thus it is possible that the growth potential of plants may not be realized because of limitations in nutrient availability if the supply of nutrients from soil is less than or asynchronous with demand. The available evidence indicates that nitrogen and phosphorus mineralization will increase in mesic more than in dry tundra. Also, the increased depth of the active layer in organic soils will tend to allow deeper rooting to access the mineral horizons which were previously inaccessible (Nadelhoffer *et al.* 1992, Berendse and Jonasson 1992). Projected increases in nitrogen fixation of 65-85% (Chapin and Bledsoe 1992) are also likely to minimize nitrogen limitations to plant growth under projected climate changes. However, extrapolation of nutrient information over time and space must be treated with caution. Enhanced release of nutrients does not necessarily lead to increased plant uptake and there is considerable variation between responses in different sites (Jonasson *et al.* 1993, Parsons *et al.* 1994). Transport from lower latitudes of atmospheric pollutants, especially nitrogen compounds, may also influence nutrient availability.

Permafrost and vegetation

A major cause of disturbance to vegetation is, and will increasingly be, the dynamics of permafrost. Thawing of the active layer each summer allows plant roots to function. Vegetation, and mosses in particular, restrict the degree of soil warming in continuous summer daylight, but when the upheaval of the permafrost and soil results in drainage and lack of moss cover, the permafrost can thaw at the surface and retreat. This leads to soil slumping and active layer detachment by which the soil can slide down the surface of sloping permafrost at a rate of up to 5 m per week (Edlund 1989). At the coast, thermal abrasion of soil and permafrost can occur at 14 m per year.

A particular threat to Arctic biota from soil warming is an increase in thermokarst disturbance. The disturbance to vegetation from thawing permafrost opens up new habitats for colonization and the low biodiversity of the tundra could increase, with the risk of competition from immigrating species displacing the existing species. Slow plant growth and sporadic recruitment of individuals to populations could result in recovery taking decades or centuries.

Water stress

Vegetation composition and structure in the Arctic are tied to soil moisture and nutrient availability. Actual effects of climatic warming and increased precipitation on soil moisture are hard to predict and may be site or regionally specific. Increased temperatures may, however, desiccate moist soils and hence cause water stress to plants. Although little research has been conducted on Arctic species, an increase in CO_2 may also affect tundra ecosystems via plant water relations because, in general, water use efficiency increases with enhanced atmospheric CO_2 (Oberbauer and Dawson 1992). Increased efficiency might offset soil drying and allow species to colonize marginal or barren areas of the High Arctic which would change the albedo and thus affect the energy balance (Lashof 1989). Plant communities along rivers and streams may be negatively affected by climate change. Decreased river and streamflow affects riparian ecosystems and the animals that inhabit them. A decrease in flooding and subsequent drying of deltaic ecosystems will alter plant composition and reduce migratory bird habitat.

11.5.1.1.2. **Invertebrates**

One of the likely effects of climate change accompanying enhanced temperature is a change in the pattern of outbreaks of pests and pathogens. The species involved may be indigenous but currently benign, or they may be immigrants from the lower latitudes.

Invertebrates with relatively short generation times may show great population dynamic responses to increased temperature (Strathdee *et al.* 1995). Populations of some invertebrate pests which defoliate large areas of Fennoscandian subarctic birch forests are controlled by the sensitivity of over-wintering eggs to low temperatures (Tenow and Holmgren 1987); warmer winters will likely lead to increased pest populations and greater deforestation. Which species will become harmful is unpredictable, but the probability of occurrence is high. Such outbreak events could be an important factor determining the future pattern and timing of vegetation change.

Increasing insect outbreak may also result in greater fuel loads, linking insect outbreak and wildfire. The coupling of more frequent insect outbreaks and fire is likely to decrease surface albedo because there will be less green foliage in winter; this change in albedo may cause positive feedbacks to climate (Bryant *et al.* 1991).

11.5.1.1.3. **Vertebrates**

Animals of the Arctic are ultimately dependent on plants for energy and nutrients. Climate change can affect animals directly through an increase in extreme events such as drought or ice layer formation over forage in winter and through slower processes such as changes in the amount and quality of food available. Altered winter snow conditions will affect not only the availability of moisture, food, and shelter for winter residents, but also the spring and summer moisture regime. Quality and quantity of forage for herbivores can decline through an increase in plant defense mechanisms or by a change in species composition (Bryant *et al.* 1991). Migratory mammals and birds can probably adjust more readily than non-migratory species to changes in the quantity and distribution of their food plants or prey in the Arctic, but vertebrate and invertebrate herbivores may face problems with changes in the quality of their food plants.

Long-term experiments by Chapin and Shaver (1996) show warming of Arctic tundra is likely to result in replacement of cotton grass (*Eriophorum vaginatum*) by deciduous shrub and, in particular, dwarf birch (*Betula exilis* – formerly *B. nana* ssp. *nana*). This vegetation change could be very detrimental to populations of caribou and reindeer (*Rangifer tarandus*). At the time of calving and in early lactation, these animals depend heavily first on flowers of cotton grass and, later, leaves of willow for a high protein food source. High protein food is necessary for calf production and survival and hence herd maintenance (Kuropat 1984). This shift in species composition would therefore likely result in significant herd declines, especially in Siberia and North America.

Warming in high latitudes may also result in migration of taiga woody species into regions now occupied by tundra. Taiga woody species appear to be more heavily defended chemically against mammalian herbivores than are tundra shrubs. An increase in the defense level of tundra vegetation would also stress herds of tundra herbivores such as caribou (Bryant *et al.* 1991).

11.5.1.2. **UV effects on terrestrial ecosystems**

The literature on effects of UV radiation on Arctic terrestrial ecosystems is limited and many assumptions are extrapolated from data on organisms in temperate regions (IASC 1995, Björn *et al.* 1997). In polar regions, the low air and ground temperatures and low solar zenith angle, even during the vegetation growth period, affect ecosystems in various ways. Under unperturbed ozone and cloud conditions, UV-B irradiance is low. Consequently, Arctic plants have fewer protective pigments and are more UV-B susceptible than comparable plants from other regions of the world (Robberecht *et al.* 1980, Barnes *et al.* 1987). Repair processes are slower at cooler temperatures, while photochemical damage to DNA and other molecules can proceed as rapidly as at higher temperatures, thus sensitivity to UV-B is more pronounced at lower temperatures (Takeuchi *et al.* 1993). Therefore, an increase in UV radiation is an ecological concern in the Arctic, and further ozone depletion is a potential threat.

11.5.1.2.1. **Dwarf shrubs, mosses, and lichens**

The responses of subarctic plants to UV-B are subtle and sometimes surprising, vary from species to species, and cannot be predicted without detailed experimentation. There is some evidence that stem growth inhibition from increased UV-B in perennial plants increases over time. Although it is too early to make any detailed statements, the effects of altered UV-B radiation on certain plant species is expected to have an impact on complex ecological interactions.

Growth and morphology of dwarf shrubs exhibit changes in leaf thickness and relative shoot growth under increases in UV-B radiation. Species-specific changes have implications for plant community composition and plant nutritive value for herbivores. In UV enhancement experiments, the leaves of *Vaccinium vitis-idaea* became thicker (by 4 to 9% depending on year) and those of the deciduous dwarf shrubs thinner (by 4 to 10% depending on year and species) (Johanson *et al.* 1995a). Enhanced UV radiation caused increases of 24% to leaf dry weight and 29% to leaf area for *V. uliginosum* (Johanson *et al.* 1995b).

The relative longitudinal shoot growth (i.e. shoot growth under enhanced UV-B divided by growth of the same shoot during the year before application of enhanced UV-B) was reduced in *Empetrum hermaphroditum* by 14% after one year of enhanced UV-B and by 33% after two years. For the *Vaccinium* species, no significant effects on longitudinal growth of shoots were found after one year of irradiation, but after two years there was a reduction of 14% for *V. myrtillus*, of 12% for *V. uliginosum*, and of 27% for *V. vitis-idaea* (Johanson *et al.* 1995b).

Mosses as well as dwarf shrubs react to increased exposure to UV radiation. Mosses and lichens are particularly important components of Arctic vegetation, and are often critical to ecosystem function. Mosses mediate the exchange of water and energy between soil and atmosphere while lichens are the winter forage of caribou and reindeer *(Rangifer tarandus)*. Quite unexpectedly, growth of the moss *Hylocomium splendens* was strongly stimulated by UV-B, provided the moss received additional water (Gehrke *et al.* 1996). The stimulation of annual growth in length by enhanced UV-B (15% enhancement calculated for cloudless skies) was, in three successive growing periods in the field, 15%, 31%, and 27%. The stimulation of growth in length by UV-B was observed also in a greenhouse experiment. For *in situ H. splendens* receiving only water naturally available, however, there was no effect (first year of irradiation), or an inhibition of growth in length (by 25% during the second year and 18% during the third year).

Thalli of lichens (*Cladonia arbuscula, Cetraria islandica* and *Steroecaulon paschale*) collected at Abisko (68°N) were compared to thalli of the same species collected in southern Sweden (56°N). Experiments were carried out under 350, 600, and 1000 ppm CO_2. UV-B exposure increased the photochemical quantum yield of photosystem II as measured with pulse modulated fluorimetry, (F'm-Ft)/F'm, under low CO_2 (except late in the season, i.e. August), but not under the highest concentration. The ratio was greater for lichens from Abisko than for those from southern Sweden (Sonesson *et al.* 1995).

11.5.1.2.2. **Decomposition**

The relative weight of indirect effects of UV-B on decomposition (via changes in litter chemistry) versus direct effects (via photodegradation of lignin, etc.) are liable to differ between tundra habitats. In High Arctic habitats and fellfields with open plant canopies, direct effects may outweigh indirect effects when litter intercepts UV-B. In closed vegetation regions of the mid- and subarctic, direct effects of photochemical degradation may predominate when living leaves intercept UV-B (Moorhead and Callaghan 1994). Whichever pathway prevails, alterations in nutrient availability to plants are significant in the nutrient-limited ecosystems of the Arctic.

Exposure of *V. uliginosum* leaf litter to enhanced UV-B changes the litter composition toward a decrease in cellulose and the cellulose/lignin ratio and an increase in tannins (Gehrke *et al.* 1995). UV-B exposure during decomposition decreases the proportion of lignin and decreases fungal colonization and total microbial respiration. Of three fungal species investigated, *Mucor hiemalis* and *Truncatella truncata* were more UV-B sensitive than was *Penicillium brevicompactum* (Gehrke *et al.* 1995). Leaf litter from treated (UV-B supplemented) *V. myrtillus* plants is decomposed more slowly by microorganisms than is litter from control plants (Gehrke *et al.* 1995). The increase in UV-B radiation, although possibly lethal for some soil microorganisms, will have little direct effect because of its very limited transmission into the soil. Changes in plant chemistry in response to UV-B will probably have much greater influence on soil processes (Caldwell *et al.* 1989, Gehrke *et al.* 1995).

Positive responses of mosses to enhanced UV-B could counteract future impacts of climatic warming, but, as in the example of *H. splendens*, predictions of future water balance under warming in the Arctic suggest greater stress which would result in repressed growth due to UV-B.

11.5.1.2.3. Animals

Animals are likely to be more affected by the impact of changing UV-B levels on their food supplies and general environment than by physiological harm. Nevertheless, ocular damage and systemic damage may harm larger mammals, particularly under large changes in UV-B. Increased UV-B could lead to a deterioration in the quality of forage which in turn could lead to nutritional stress on herbivores. For example, UV-B is causally linked to synthesis of certain phenolics that may act as feeding deterrents or toxins to herbivores including mammals (Caldwell *et al.* 1989). Over the past 10 years, an increase in tannin and other phenols in some Arctic species has been linked to a decline in productivity of moose (Bo and Hjeljord 1991) and might be associated with a decline in productivity of caribou. It remains unknown whether the increase in phenolics is from an increase in sunny days and a resultant increase in PAR or from an increase in UV-B or interannual differences in temperature.

11.5.2. Aquatic ecosystems

11.5.2.1. Climate change and marine ecosystems

Climate change will have a profound impact on the marine ecosystem through changes in temperature, sea ice, cloud cover, and ocean circulation patterns. The effects of warming ocean temperatures and decreased extent or thickness of sea ice on primary productivity in marine ecosystems are unclear. Sea ice thickness and extent has a bearing for the communities of phyto- and zoo-plankton that utilize it as a specific habitat (Horner 1985). A possible consequence of more open water and thinner ice is increased primary productivity which would affect higher trophic levels. Conversely, reduction in the stratospheric ozone layer and increased UV penetration as discussed in section 11.4.3 are likely to have adverse effects on productivity. Organisms higher up the food chain will also be affected by warmer waters; community composition of marine fish may be altered, which would affect predatory mammals, birds, and commercial fisheries.

11.5.2.1.1. Marine fish

An increase in ocean water temperatures is expected to affect species distribution and population numbers, thereby altering species composition in marine waters. The Barents Sea can be used as a model to show the effects of warming on fish populations since there is historical documentation of the effects of relatively warm and cool temperatures. In the Barents Sea, where temperature is the limiting factor for population distribution of many fish species, warming will allow new areas to be colonized by cod (*Gadus morhua*), herring (*Clupea harengus*), haddock (*Melanogrammus aeglefinus*), and Greenland halibut (*Reinhardtius hippoglossoides*), among others. In contrast, herring and gadoids (cod and haddock) in the North Sea might slowly decline as they are pushed north out of important feeding and spawning grounds by competing species from the south. Mackerel (*Scomber scombrus*) and sardine (*Sardina* spp.) may increase and the area become more Mediterranean-like in its fish populations (Øiestad 1990). Salmonids may benefit from changes in both ocean temperatures and stream habitat (see section 11.5.2.4.2).

Predicted climate changes that cause increased ice extent and cooling in some marine areas might decrease habitat for marine fish species. Cooling deep water could limit gadoid production, much as occurred in the Norwegian and Barents Seas from around 1900 into the 1920s (cf. Øiestad 1990).

11.5.2.1.2. Larger animals

Climate change is likely to affect the habitats and therefore the areal extent of larger animal populations. Sea ice provides critical habitat for ringed seals (*Phoca hispida*), walrus (*Odobenus rosmarusa*), and polar bears (*Ursus maritimus*). Changes in sea ice will affect each of these marine mammals in various ways. Annual variations in sea ice distribution and abundance affect ringed seal and polar bear reproduction and survival (Stirling and Derocher 1993). Changes in distribution and abundance of polar bears as a result of climatic fluctuation have been demonstrated for a population of polar bears in Greenland (Vibe 1967). Polar bears at the southern extent of their range are likely to be most severely affected. Climate warming may impact the ability of polar bears to successfully hunt seals by reducing access to seals because of a decrease in sea ice surface and seal population.

11.5.2.2. Photochemical effects of UV: Dissolved organic matter

Dissolved organic matter makes up the bulk of organic material in both lakes and oceanic water. It also supplies available carbon (Walsh 1995, Wheeler *et al.* 1996). Because of its absorptive properties, dissolved organic matter (measured as dissolved organic carbon, hereafter DOC) in the water column influences how well biota are shielded from UV exposure (Scully and Lean 1994, Schindler *et al.* 1997, 1996b). Autochthonous DOC is generated via excretion by plants and animals within the aquatic ecosystem and blocks very little UV (Curtis and Adams 1995). In contrast, allochthonous DOC supplied from terrestrial sources is highly colored and is the principal UV-attenuating substance in marine and fresh waters.

The biogeochemical cycling of DOC is partially regulated by UV radiation. In contrast to earlier beliefs, it has been shown during the last decade that dissolved humic matter (DHM), largely of terrestrial or littoral origin, is available to pelagic bacteria, making it an important carbon source in addition to autochthonous organic matter (Tranvik 1988, Hessen *et al.* 1990, Moran and Hodson 1994). Since DOC absorbs light, particularly UV radiation, radiation can be ex-

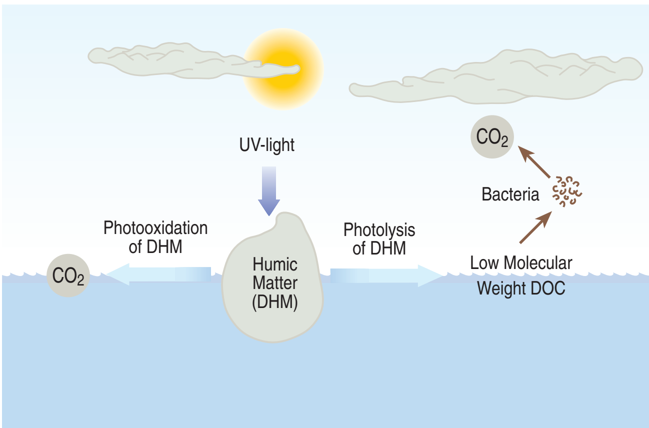


Figure 11·22. Illustration of the role of UV radiation in the biogeochemical cycling of DOC. Under the influence of solar UV light, dissolved humic matter (DHM, organic matter largely derived from terrestrial ecosystems) in streams, lakes and coastal marine waters can be transformed to CO_2. This can occur through purely abiotic photooxidation of DHM or through photolysis (cleavage) of DHM into smaller molecules more easily utilizable as a substrate for aquatic bacteria than the original high molecular weight DHM. Increased availability of bacterial substrate then leads to enhanced CO_2 production through bacterial respiration. However, direct negative effects of UV light on bacteria, causing reduced growth and respiration, also influences the process. The balance between these separate processes in nature is unknown.

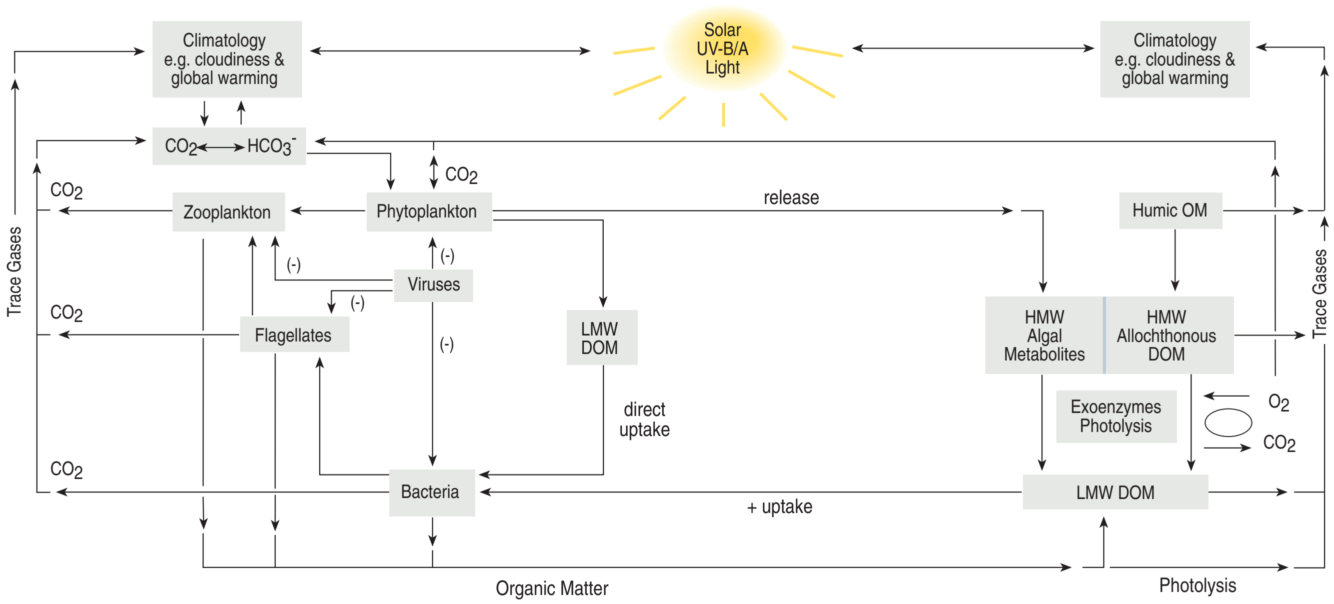


Figure 11·23. Illustration of the role of UV radiation in various light-dependent processes. Natural UV radiation impacts the numbers, distribution and activity of several aquatic ecosystem targets and thus, theoretically, their interactions (adapted from Karentz *et al.* 1994). There is a large uncertainty as to the magnitude of UV effect and the sign (feedback/feedforward) of the UV-dependent loops between the ecosystem/atmospheric components.

pected to transform autochthonous and allochthonous DOC abiotically. There are two main pathways (see Figure 11·22): 1) production of Dissolved Inorganic Carbon (DIC)* through photooxidation of DOC (Miles and Brezonik 1981, Kulovaara and Backlund 1993, Frimmel 1994, Salonen and Vahatalo 1994) and, 2) production of low molecular weight organic compounds and CO through photolysis of DOC (Mopper and Stahovec 1986, Kieber *et al.* 1989, Backlund 1992, Kulovaara and Backlund 1993, Allard *et al.* 1994, Conrad and Seiler 1980, Miller and Zepp 1995).

The net result of exposure of DOC to radiation would, in both cases, be increased DIC production and a higher turnover rate of DOC, as Pathway 1 would directly produce DIC and Pathway 2 would supply pelagic bacteria with readily utilized substrates, increasing bacterial production rate, biomass, and rate of respiration, i.e., CO_2 production.

Mopper *et al.* (1991) suggest that photochemical degradation is the rate-limiting step for the removal of a large fraction of oceanic DOC. Since oceanic DOC is one of the largest global carbon reservoirs, increases in its rate of cycling could potentially significantly increase atmospheric CO_2 (Wetzel *et al.* 1995). In addition, model calculations indicate that increased atmospheric CO_2 can cause the formation of an Arctic ozone hole (Austin *et al.* 1992), further increasing the positive feedback between increased UV-B and CO_2 in the atmosphere. If UV light, and especially UV-B, is primarily responsible for photooxidation and photolysis of DOC, then increased UV-B radiation would cause an increase in photo-oxidative and respiratory production of DIC in surface waters (lakes, wetlands, rivers, marine waters), escalating CO_2 transport to the atmosphere and thus reinforcing the greenhouse effect.

In ocean waters, DOC may increase under global warming due to the thaw of terrestrial permafrost which might increase run-off to the Arctic Ocean in some areas (Jones and Briffa 1992, Goryachkin *et al.* 1994). However, climatic warming may decrease DOC in freshwater systems and oceans due to a reduction of export of DOC from terrestrial and wetland catchments as reduced precipitation and increased evaporation cause streamflow to decrease and water tables to fall. An increase in UV-B concurrent with a decrease in DOC increases the removal of DOC from freshwater. Acidification compounds these effects (Schindler *et al.* 1997).

Apart from its optical properties, DOC also influences water column chemistry by chelating trace metals such as iron, manganese, copper, and aluminum that are either essential nutrients or can even be toxic in their free state (e.g., Al). UV radiation can alter the redox state of trace elements via formation of reactive oxygen species. Such changes will affect the solubility and availability of biogenic elements for organisms.

11.5.2.3. **UV and the marine ecosystem**

An increase in UV radiation due to stratospheric ozone depletion has measurable, direct impacts on the marine environment. Light-dependent processes in phytoplankton (i.e., plant photosynthesis, photoinhibition, phototaxis, photoprotection, trace gas emission, etc.), animals (i.e. vision, behavior and reproductive cycles, sensitivity to epidermal disease, etc.), as well as microbial and photochemical transformations of various organic molecules and redox states of metals are all regulated by the spectral composition of the light field (Figure 11·23) (cf. IASC 1995, SCOPE 1992, 1993, UNEP 1991, 1994, IASC 1995). Owing to seasonality, depth refugia, vertical mixing, and pronounced natural annual variability, detection of early signs of increased UV stress require careful, focused experiments.

Special attention has been paid to the sensitivity of primary production in marine ecosystems to UV radiation because planktonic algae are the base of the marine food web and they constitute an important CO_2 sink in the global carbon cycle. The considerable dilution of Arctic Ocean surface waters by seasonal melting of ice edges and year round freshwater inflow causes pronounced stratification of the upper mixed layer (30-50 m thick) in the Arctic Ocean at all times of year (cf. Barry 1989). This is in striking contrast to other major world oceans except near Antarctica (Smith *et al.* 1992). Coastal waters of the Arctic Ocean carry high loads of particulate and dissolved organic matter (comprised of minerals, dissolved organic carbon, chelating metals, and humic substances recalcitrant in nature). As a result of constraints on nutrient availability, planktonic organisms and

* DIC is defined as the sum of dissolved CO_2, H_2CO_3, HCO_3^- and CO_3^{2-}, but not including CO, which has also been shown to be produced photo chemically from DOC in water (Conrad and Seiler 1980, Miller and Zepp 1995)

larvae are confined to the upper reaches of the water column and thus may be particularly vulnerable to increased UV radiation, especially in the spring months when ozone depletion appears most severe and when migratory species of animals come to breed in the Arctic Ocean and its freshwater tributaries.

Each marine plant ecosystem, whether free-floating, embedded in the undersides of ice, or attached to the surface of the sediments, serves the needs of different sets of animals, plays an important and often distinct role in the biogeochemistry of the Arctic Ocean, and is likely differently affected by changing UV climatology over Arctic latitudes (IASC 1995). UV also directly affects the vitality and chemical activity of viruses, bacterioplankton, protozoans, and microzooplankton (Herndl *et al.* 1993, Hunter *et al.* 1979, Williamsen and Zagarese 1994, Helbling *et al.* 1995, Müller-Niklas *et al.* 1995, Vetter 1996). Because the amount of UV-B reaching marine plants varies, the effects of increased UV-B will also vary. In the long run, some organisms may adapt to enhanced UV-B radiation and thereby a succession toward more tolerant species will occur. Adaptive mechanisms such as synthesis of screening pigments, capacity to avoid UV-B, and increased enzymatic repair might reduce the inhibitory effects of UV-B if phytoplankton are exposed during longer time periods (Karentz 1994). However, the sensitivity to UV-B radiation and the adaptive mechanisms differ widely between species.

11.5.2.3.1. **Primary producers**

Physiological responses of primary producers to UV have been observed in cell DNA, protein, nitrogen metabolism, lipid metabolism, motility, screening pigments (mycosporine-like amino acids), photosynthetic machinery (photosystem II, the carboxylating enzyme, the Calvin-Benson cycle), and repair mechanisms (e.g., Holm-Hansen *et al.* 1993a, Vincent and Roy 1993, Häder *et al.* 1995). UV can inhibit motility in flagellated phytoplankton (Ekelund 1990, Häder 1993) and alter cell composition (Vosjan *et al.* 1990, Goes *et al.* 1994, 1995) and cell structure (Karentz *et al.* 1991, Behrenfeld *et al.* 1992). UV-B radiation is known to reduce short-term growth rates and primary production of a number of polar aquatic phytoplankton communities (cf. Holm-Hansen *et al.* 1993a, Vincent and Roy 1993, Karentz *et al.* 1994, Prézelin *et al.* 1994a, 1994b, 1995, 1996, 1997, Boucher and Prézelin 1996).

There is a nonlinear relationship between changing UV climatology and UV sensitivity of natural phytoplankton communities, making the accuracy of future spectral modeling of UV effects on primary production dependent upon the availability of *in situ* BWFs (see section 11.4.3.1) for the aquatic ecosystem under study. To extrapolate results of enhanced UV-B on phytoplankton growth into consequences of increased UV-B radiation, the correct BWF needs to be applied. Results from Antarctica show that BWFs vary on time intervals as short as two hours, reflecting diel periodicity in both cell biology and the magnitude and spectral balance of the natural light field. Also, BWFs differ significantly over the austral springtime and may reflect different light histories, taxon-specific biology, or changing physiological states of the phytoplankton communities.

The Arctic and subarctic surface waters exhibit low nutrient (nitrogen, silicate, and phosphate) concentrations which usually become exhausted during summer. Both nitrate and silicate have been considered to limit primary production (Kristiansen and Farbot 1991, Harrison and Cota 1991). The nitrogen transformations liable to be affected by UV-B radiation are uptake, assimilation, mineralization, and nitrogen fixation, since all of these occur in the upper part of the water column. Phytoplankton productivity in the Arctic Ocean varies widely from the least productive Central Polar Basin to ice-free coastal upwelling areas and episodic blooms, especially red tides, during periods of spring/summer stratification (cf. Andersen 1989, Smith and Sakshaug 1990, Smith and Grebmeier 1994).

Little effort has been made to understand the effects of UV-B on *in situ* primary production in the Arctic Ocean. Effects of UV-B radiation on *in situ* primary production in the oceans near Antarctica have been examined by incubating either unialgal cultures or the ambient natural populations. By modeling the results from these experiments, often based on irradiation from inside and outside an ozone hole, scenarios for reduced primary production have been calculated (e.g., Smith *et al.* 1980, 1992, Smith and Baker 1982, Holm-Hansen *et al.* 1993b, Cullen *et al.* 1992, Cullen and Neale 1994). In general, it is assumed that UV-B irradiation reduces total water column photosynthesis by 5-6% under normal Antarctic ozone levels, while during an ozone hole the reduction can increase to 6-18% (cf. Schofield *et al.* 1995). It is evident that ambient radiation (PAR, UV-A and UV-B) during normal levels of stratospheric ozone inhibits primary production close to the surface, and thereby enhanced UV-B radiation will increase the UV-B dependent inhibition. There are no indications that similar scenarios could not be expected in the Arctic. Recent research indicates that post-bloom phytoplankton communities from northern Norway are more sensitive to ambient UV-B radiation than those from the southern ocean (Helbling *et al.* 1996).

The above-mentioned conclusions are all based on short-term experiments but results suggest that UV-B radiation will induce changes in species composition that will cause long-term effects at a community level. Unfortunately, experimental data describing the long-term effects of UV-B radiation on phytoplankton communities are very limited, but changes in species composition of phytoplankton communities have been shown in long-term experiments with natural phytoplankton communities (Worrest *et al.* 1981, Helbling *et al.* 1992, Wängberg *et al.* 1996, El-Sayed *et al.* 1990). Evaluation of the effects of UV-B irradiation on the marine ecosystem lack long-term experiments on the biological and chemical processes in planktonic food webs. Not enough data are available to make a valid risk assessment (Wängberg *et al.* 1996).

Experiments with natural assemblages of phytoplankton using additions of ^{15}N-labeled nutrients have shown nitrogen uptake and assimilation to be negatively affected by UV-B (e.g., Döhler *et al.* 1991, Döhler 1992). Enhanced UV-B might then increase the nitrogen demand in phytoplankton because of protein degradation. Proteins are known to be absorbers of UV-B radiation (Karentz 1994); an increase in protein degradation followed by resynthesis in order to replace UV-B-sensitive proteins could thus be expected during UV-B exposure (Cullen and Neale 1994). Nitrogen-depleted phytoplankton might therefore be more susceptible to UV-B radiation. In addition, UV has a regulatory effect on the uptake of inorganic nitrogen and phosphorous required for growth (Dohler 1992, Tyagi *et al.* 1992, Behrenfeld *et al.* 1995). On longer time scales, if phytoplankton fix less carbon, grow slower, or increase rates of organic excretion in the presence of UV radiation, there may be indirect consequences to the biogeochemical cycling of nutrients and therefore to the vitality of higher trophic levels within aquatic ecosystems.

Phytobenthos and macroalgae play important roles in the extensive coastal areas of the Arctic Ocean, contributing ≥50% of total primary production of shelf systems. Macroalgae produce organohalogens that are released to the water. It has been suggested that these volatile gases play a role in trace gas chemistry of the atmosphere (Gschwend *et al.* 1985). Hence, the influence of UV radiation on macroalgal growth might have various secondary consequences of concern for climate change issues (IASC 1995). Both laboratory studies under artificial light sources and field studies under natural light suggest that UV radiation inhibits production and growth in many different benthic, lithotrophic, and macroalgal taxa, as well as seagrasses, but that the magnitude of the impact may well vary between taxa and as a function of a photoadaptive state (Trocine *et al.* 1981, Wood 1987, Larkum and Wood 1993, Evens *et al.* 1994, Grobe and Murphy 1994).

11.5.2.3.2. **Bacteria**

The capacity of bacteria to resist or adapt to UV-B irradiation varies widely. Ambient UV-B irradiation can inhibit bacterial DNA replication, protein synthesis, and degradative enzyme activities in some species (Herndl *et al.* 1993, Helbling *et al.* 1996, Müller-Niklas *et al.* 1995). Other genera seem to have an extreme resistance to UV-B irradiation and some are even stimulated in population growth by low-level exposure (Herndl *et al.* 1993, Karentz 1994).

Photochemical degradation of DOC, causing increased availability of low molecular weight compounds, could also stimulate bacterial activity. Lindell *et al.* (1995) and Wetzel *et al.* (1995) found enhanced bacterial numbers and cell volumes concomitantly with a reduction in DOC when exposing natural limnic water to moderate UV-B irradiation; they interpreted this to be a result of enhanced availability of bacterial substrates. The net effect of these processes may depend on the humic content and absorption of the water (Lindell *et al.* 1995), although the balance between these effects needs to be further evaluated before the long-term response of an ecosystem can be determined (Karentz *et al.*, 1994). Heterotrophic bacteria compete efficiently with phytoplankton for nutrients (Goldman and Dennett 1991). An enhanced nitrogen uptake relative to carbon fixation rates in natural plankton communities exposed to UV-B radiation during a one week experiment indicated a proliferating bacterial population. A 35-40% increase in thymidine incorporation activities has also been found in *in situ* enclosure (5 m^3) experiments with natural pelagic communities exposed to enhanced UV-B radiation for a fortnight. Although individual phytoplankton species may not be harmed by or may be adapted to increased UV-B radiation, phytoplankton productivity in nitrogen limited waters might decrease due to competition by bacteria for nutrients.

11.5.2.3.3. **Zooplankton**

Natural irradiances of UV-B have been shown to be lethal to zooplankton; the numbers, survival and fecundity of their eggs and nauplii are reduced by short-term UV-B exposure (Karanas *et al.* 1979, Damkaer *et al.* 1980, 1981, Damkaer and Dey 1983, Ringelberg *et al.* 1984, Thomson 1986). That certain zooplankton will be able to adapt, or already have protective systems, is suggested by the lower mortality rates often encountered in zooplankton containing photoprotective compounds such as astaxanthin, melanin, and mycosporine-like amino acids (Hairston 1976, 1980, Ringelberg 1980, Ringelberg *et al.* 1984, Dunlap *et al.* 1986, Herbert and Emery 1990, Karentz *et al.* 1991). Increased UV could cause a change in species composition and thereby a succession that may alter the planktonic food webs.

11.5.2.3.4. **Invertebrates**

Little research has been done on the responses of invertebrates to enhanced UV in the Arctic and few publications are available on invertebrate responses to UV radiation in any region of the globe. Exceptions are the work of Hatcher and Paul (1994) and Newsham *et al.* (1996) which show that invertebrate herbivores can be affected both by changes in food plant quality and by apparent artifacts associated with methods of increasing UV exposure. Recent studies suggest that UV inhibits blackfly (*Simulidae* and *Diptera*) colonization in boreal and alpine streams (Bothwell *et al.* 1994, Williamson *et al.* 1996).

11.5.2.3.5. **Fish populations**

Fish species of highest risk in Arctic areas could be those species that settle in shallow waters in the early spring and species with pelagic drifting eggs and larvae. Many important fish stocks (i.e., herring, pollock, cod, and salmonids) spawn in fully exposed shallow waters where they are exposed to epidermal and pineal effects that may be aggravated by UV-B radiation. Fish larvae are generally very susceptible to UV-B damage, and even ambient solar radiation has the potential to affect their viability (Hunter *et al.* 1979). The larvae of several fish species, including salmonids and herring, are particularly vulnerable to UV radiation since spawning occurs in very shallow waters and the larvae remain in the vicinity during their early stages of growth. Skin and gill lesions caused by UV-B radiation have been observed in adult fish in lower latitudes (Robberts and Bullock 1981, Bullock 1985, Bullock and Coutts 1985). Research needs to be carried out on Arctic organisms that have evolved in low UV environments. Important for evaluating the potential effects of UV radiation on fish are timing of the spawn, the distribution of species within the water column (Holm-Hansen *et al.* 1993a), and the degree of vertical mixing in the water column.

Indirect effects of enhanced UV-B on the fisheries may arise from changes in the planktonic food webs. Perturbations in food webs, however, are difficult to predict because they depend on the long-term adaptation of species and changes in community structure. Long-term experiments on natural communities and ecosystems are warranted in order to predict such changes.

11.5.2.3.6. **Larger animals**

If the secondary productivity of the Arctic aquatic ecosystem is challenged, then so too is the viability of sea birds and land predators (i.e., seals, foxes, bears) which feed on aquatic organisms. The ocular effects of increased UV on anadromous fish and also birds, seals, and polar bears need to be considered as well (IASC 1995).

11.5.2.4. **Climate change and Arctic freshwater**

11.5.2.4.1. **Climate change and Arctic lakes and ponds**

The amount of ice and snow accumulated on a lake or pond, and the duration of ice cover, have profound effects on the limnological characteristics of all northern aquatic systems. Especially in the High Arctic, where the entire ice free season may only be a few weeks in duration (Schindler *et al.*

1974, Douglas and Smol 1994), even modest climatic warming can significantly increase the growing season.

Arctic ponds (*sensu* Sheath 1986) are defined as shallow (generally < 2 m deep, and in many cases <1 m) systems that freeze to the bottom each winter. Fish and other vertebrate predators are generally absent (with the exception of some migratory birds), and thus food webs are relatively simple. These shallow ponds are a dominant feature of many Arctic landscapes, yet few data are available on the limnology of these systems (e.g., Hobbie 1980, Douglas and Smol 1994, Havas and Hutchinson 1983).

The most obvious effect of a warmer Arctic would be to increase pondwater temperatures. Because the vast majority of these sites are so shallow, the ponds closely track ambient air temperatures (Douglas and Smol 1994). With warmer temperatures and reduced ice cover, many aspects of the pond system would change (Rouse *et al.* 1997, Douglas and Smol 1993, 1994). Reduced ice and snow cover may result in a lengthened growing season, which will affect many chemical, physical, and biological processes. In some lakes total annual primary production may increase, and more complex periphytic assemblages may develop (Douglas and Smol 1993, 1995). In deeper lakes, however, primary production may decrease after break-up so that earlier break-up would have a negative effect on production (Kalff and Welch 1974).

On the other hand, if precipitation does not increase significantly, evaporation from ponds would increase with climatic warming and the net water balance could be affected. As a result, pondwater conductivity values might increase, as might pondwater pH (Douglas and Smol 1994). A more significant manifestation of increased evaporation rates would be that some of the shallower ponds would desiccate completely.

The relationship between lakewater acidity and climate change may prove to be a significant determinant of Arctic ecosystems. Climate might control lakes via pH through the effects on water renewal and relative yields of base cations and strong acid anions (Psenner and Schmidt 1992, Schindler *et al.* 1996a). Climatic warming and increased evaporation cause increased in-lake removal of sulfate and return of base cations from sediments, both of which increase as water renewal time increases; thus alkalinity increases. In areas that are strongly affected by acid rain, however, the effects of climatic warming and increased UV-B may increase cycling of DOC (Schindler *et al.* 1996a).

High Arctic lakes would similarly be affected by climatic warming, although maximum water temperatures would be less affected than in ponds, as deeper lakes have much higher thermal capacities. High Arctic lakes are characterized by extended snow and ice covers, with some lakes remaining covered by a central float of ice throughout the short summer. Production in High Arctic lakes occurs whenever snow is absent and the sun is above the horizon. Much of the annual production occurs under total ice cover, since many lakes are not clear of ice until late July or early August (Kalff and Welch 1974, Welch *et al.* 1989). Strong winds and low snowfalls generally keep lakes snow free for much of the year.

Temperature has been found to immediately affect productivity. Some preliminary results from a littoral zone of a boreal lake indicate higher primary production with 2°C rise in temperature, even in the absence of additional nutrient loading (Kankaala *et al.* 1994). However, Schindler *et al.*'s (1997) 20-year study found no change in production with 2°C temperature increase in boreal lakes. Phytoplankton biomass declined in proportion to reduced phosphorus concentrations and inputs caused by declining streamflow.

Although several environmental gradients exist in the Arctic, the location of Arctic treeline is climate-related, closely matching the mean July position of the Arctic Front (Bryson 1966). The presence or absence of coniferous trees that are associated with the Arctic treeline (e.g., spruce) have profound effects on the limnology of lakes draining these catchments (MacDonald *et al.* 1993, Pienitz and Smol 1993). If the tree line moves north, patterns of wind mixing and subsequent lake stratification may be altered by the presence of trees (Rhodes and Davis 1995, Schindler *et al.* 1990), and DOC released from litter may increase (Schindler *et al.* 1992). Changes in DOC can affect lake and pond water clarity and color, heat absorption, PAR and UV transmission, and nutrient cycling, which in turn affect the biota of the lakes (Salonen *et al.* 1992, Schindler *et al.* 1990, 1992, 1996a, 1997).

Although generally considered rare in the Arctic, athalassic lakes (inland, saline, non-marine lakes) occur in some regions (Pienitz *et al.* 1992, Veres *et al.* 1995). These closed-basin systems are especially sensitive to climate change because slight alterations in evaporation and precipitation can greatly affect lakewater salinity levels, which then can strongly influence the biota in these systems (Hammer 1986). Because many of these lakes are shallow, the possibility exists of total desiccation for some systems with increased evaporation rates.

Climatic warming could offset a positive feedback on light regimes and UV penetration in lakes, since warming and drought strongly reduces DOC and increases UV penetration. This has been demonstrated for North American lakes (Schindler *et al.* 1996a, 1997). These effects could influence Arctic lakes and coastal areas as well.

11.5.2.4.2. Climate change and Arctic rivers and streams

Climatic warming and resultant changes in precipitation and glacial melt are likely to have a major impact on Arctic streams and rivers and organisms inhabiting them. Thermal regimes in glacier-fed rivers could become colder while smaller streams could become warmer. Sediment and flow regimes are likely to change as a result of glacier and permafrost melting, soil drying, and changes in spring breakup (Oswood 1989).

Some of the most obvious consequences of a warming climate are increases in ice-free days and water temperatures in non-glacial freshwater systems. The effect of a milder thermal regime may change the biogeography of freshwater fish. Increased water temperatures combined with year-round groundwater flow (from decreased permafrost) might change the suitability of streams for some fish species (Eaton and Scheller 1996).

Studies at the freshwater ecosystem level are few. Allard *et al.* (1994) noted that in the subarctic river Kalix (northern Sweden) there was a higher humic fraction DOC during winter (dark period) than during summer. The authors interpret this as an effect from photodegradation of dissolved humic matter in river water during the long daylight periods of the Arctic summer.

11.5.2.5. UV and Arctic freshwater

The effects of UV on Arctic freshwater will be highly dependent on the transparency and DOC level of the water. Arctic freshwater may be expected to be altered from increased UV, especially water of high transparency and low DOC. It may be anticipated that most species and populations are locally adapted to present-day irradiance; however, plankton in both Arctic and alpine waters are commonly light stressed and exhibit low repair capacities (Luecke and O'Brien 1983, Hebert and Emery 1990, Hessen *et al.* 1990). A further increase in UV-B could thus cause detrimental effects on various taxa. Organisms in shallow ponds without depth refugia could, in particular, be susceptible to UV. Recent studies of UV-B effects on bacteria and zooplankton have been undertaken in experimental enclosures in the Canadian Arctic (Lean unpubl.). Combinations of *in situ* and lab studies have been performed with phytoplankton in subarctic alpine areas (Hessen *et al.* 1995, Van Donk and Hessen 1996), and particularly with zooplankton in the Canadian Arctic (Hebert and Emery 1990) and Norwegian subarctic and High Arctic at Svalbard (79°N) (Hessen 1993, 1994, 1996). The work with phytoplankton demonstrates high susceptibility of flagellum status, phosphorus uptake, growth rate, and cell wall morphology. A particularly intriguing discovery is that UV exposure induces cell wall changes that reduce digestibility of phytoplankton for zooplankton (Van Donk and Hessen 1995). The zooplankton studies focus on the widespread key genus *Daphnia*, and studies on melanic (UV tolerant) and hyaline (UV susceptible) natural clones have been performed. These studies are intimately linked to population genetics and evolutionary history of these clones.

11.6. Effects of climate change and UV radiation on Arctic peoples

An understanding of contemporary patterns of settlement and resource use combined with evidence from historical and archeological records can help estimate the magnitude of potential impacts of climate change to Arctic communities. To the extent that Arctic peoples continue to harvest plant and animal foods, live permanently or seasonally on coastal spits and along river banks and lake shorelines, and use snow mobiles and dog sleds for transportation, climate changes will directly impact their lives.

Climate change affects people on different scales; globally, regionally, and subregionally. While a general warming of the Arctic is predicted, regional weather patterns are probably more important to human use and habitation as some areas will become cooler while others warm. Should glaciers and ice sheets melt significantly over the coming decades, sea level rise will impact village location, transportation, tourism, water resources, and other facets of Arctic life. If climatic warming results in less sea ice and better access to northern oceans, it is reasonable to project that activities such as shipping, mining, and on-and offshore oil development will increase. To understand the impacts of climate change on communities and regions, it is important to identify associated regional policy concerns (Yin and Cohen 1994).

Potential effects of climate change on settlement and resource abundance can be inferred from the past record. To the peoples living in the North, normal seasonal variation has always been considerable, commonly demanding or enabling routine changes of settlement area, social grouping, house form, subsistence activity, mode of transportation, and diet. While access to modern resources may mitigate the effects of climate change, impacts upon the ability of northern populations to continue traditional ways of life might be overwhelming. Establishment of permanent villages and towns may also compound the effects as people today are tied to villages and less able to move with shifting resources. In some areas then, climate change might further remove people from culturally important hunting, gathering, and fishing activities and associated spiritual and social relationships, knowledge transference, and other traditional economic endeavors.

11.6.1. Prehistorical and historical effects of climate change

Oceanographic and terrestrial factors related to indigenous occupation of the Arctic center around cyclic resource abundance or biological productivity. In Northwest Alaska, past periods of warming are associated with intensive occupation of coastal sites while cold periods supported inland occupation by Iñupiat (Mason and Gerlach 1995). Ocean dynamics and especially areas of nutrient upwelling influence location and hence availability of marine mammals and fish. Areas in close proximity to polynyas have historically and prehistorically been productive areas for human settlement, as have inland areas supporting large numbers of caribou and fish. Changes in ocean dynamics resulting in changes in polynya location have caused shifts in settlement patterns (Schledermann 1980, Mason and Gerlach 1995). Sudden decreases in caribou populations or sudden changes in caribou migration routes have forced indigenous peoples to abandon some areas for more productive ones; some past changes in caribou concentrations have been attributed to climate change.

Variations in climate may crucially alter societies living in regions which are environmentally and climatically marginal as certain economic activities, particularly those based on agriculture, become inviable (Parry 1978). In pre-twentieth century Iceland, for example, periods of cooling were frequently associated with poor harvests, and subsequent malnutrition and death among livestock and humans (Ogilvie 1982, 1984). In Greenland, the Norse settlement was abandoned around 1350 AD. Computer modeling of the climate and agricultural data indicate that a combination of reduced summer growing season for fodder and an extended winter feeding period for imported European domestic animals may have placed the Norse economy under extreme stress (Buckland *et al.* 1996, Barlow *et al.* 1997). Inuit living in the same area of Greenland, with seal hunting forming their main economic base, continued to thrive. From these and other examples, it is possible to infer that dependence upon local resources and the vulnerability of local environments and geography to changes in parameters such as precipitation, temperature, sea level, and ice cover can lead to large-scale shifts in resource use and settlement patterns in the Arctic.

11.6.2. Settlement and resource use

Indigenous patterns of resource use and settlement location are based on stable patterns of resource availability. Settlements have been established where food, water, shelter, and transportation are reliable. Today hunting and fishing camps are set where resource abundance can be predicted for certain times of the year; as resource availability shifts so do hunting and fishing camps. While today many would describe the economy of northern peoples as mixed, the robustness of the indigenous economies depends on their abilities to accommodate the annual variability in local plant and animal resources as well as the physical environment (Langdon 1986).

Many coastal settlements in the Arctic, particularly in Alaska and Russia, are on low-lying coastal plains and river deltas. Storms during summer and fall have historically caused damage to these communities. Coastal erosion threatens several villages in Alaska, and some have already moved to more protected areas. Changes in ice formation and breakup could increase flooding along rivers. A rise in sea level will threaten many more communities as coastal plains and spits are flooded and the geography of major river deltas transformed. Detailed information on community response to floods is provided by Newton (1995).

Loss of permafrost from climate warming may also threaten existing structures such as buildings, underground storage, roads, and pipelines. Areas that presently have large amounts of massive ground ice could become particularly unstable, causing increased problems for existing roads and buildings and new construction. At the same time, construction methods may become easier in some areas as soils thaw, dry out, and become more stable.

Precipitation and river flow are also important considerations for village location. Temperature and humidity changes may affect both the energy balance of ecosystems and the availability of water and snow cover. Much of the Arctic receives little precipitation, and a decrease in snowfall or rain could reduce water supplies for villages and towns. In the large areas of coastal plain and wet tundra, an increase in precipitation could make lands unusable for settlement. Changes in snow cover and sea ice would also affect traveling conditions over the land and sea respectively, altering the ability of hunters to reach other villages and hunting locations in winter. Increased precipitation and run-off could also interfere with springtime travel.

Thus, climatic changes would impact geophysical and biological processes. Changes in air and water temperature, precipitation, and storm patterns could affect availability of fish, land and marine animals, and plant foods to humans. Indigenous hunting strategies are the result of the interplay of complex variables such as the distribution of numerous game species, dietary needs, transportation possibilities, and energy expenditures of different hunting forms (Freeman 1984). Animal and fish species which often seem abundant are largely composed of seasonal migrants to specific feeding grounds where food production can be intense but short.

Changes in vegetation and precipitation could shift the migration of terrestrial mammals and alter the breeding and molting areas of birds. Cooling or warming of rivers and lakes would probably affect the distribution and abundance of all types of freshwater and anadromous fish.

Marine mammals are important cultural and dietary resources for indigenous people of the Arctic (Lantis 1938, Worl 1980). Human exploitation of marine mammals is dependent on sea ice conditions, including lead and polynya formation, ice extent, and location of shorefast ice. At the same time, the population density of many marine mammals including ringed seals and walrus is positively correlated with the distribution and condition of coastal sea ice (MacLaren 1961). Climate change might cause ringed seals, whales, or walrus to migrate in new paths, thereby significantly affecting subsistence harvesting. All of the above changes may directly impact the way indigenous peoples are able to obtain food resources and maintain their economies.

Increases in off-shore oil development and shipping due to a decrease in sea ice might affect populations of marine mammals in general. Of particular concern are culturally and nutritionally important animals such as ringed seals, whales, walrus, and polar bears (Stirling and Calvert 1983, Stirling 1988, Stirling and Derocher 1993). An increased ability for humans to live in the north may also increase human-bear interactions, and bears could be pushed out of parts of their present range.

11.6.3. Economic activities

A warmer climate will impact communities economically and socially. Accounting for these factors requires consideration of the nature of Arctic communities, which often include both wage and non-wage (subsistence) economies and lifestyles (Peterson and Johnson 1995). Adaptation responses will be shaded by other factors besides climate change, including global economic forces, technological changes, and political developments. This is evident in the Mackenzie Basin Impact Study (MBIS) case from northwest Canada (Cohen 1997).

11.6.3.1. Commercial fisheries

The effects of both warmer and cooler ocean temperatures on fish populations are discussed in section 11.5.2.1.1. A warmer ocean could bring large economic benefit to Arctic residents if fish populations increase. However, some populations might decline as species composition changes. Aquaculture would probably profit from a temperature increase. However, toxic algae and new fish diseases might cause increasing problems in a warmer ocean as might occasional episodes of high surface water temperatures (Øiestad 1990). Beneficial influences on wild salmon populations might come from river and stream temperatures, flow rate, and sediment amount. As discussed in section 11.5.2.4.2, some fish species, including salmonids, are likely to benefit from effects of a warming climate. In some areas where ocean temperatures decrease and the extent of sea ice expands fisheries would be expected to collapse (Øiestad 1990).

11.6.3.2. Reindeer herding

In Fennoscandia and Eurasia domestic reindeer herding is an important livelihood among indigenous pastoralists. It is an important food source, a foundation of cultural heritage, and a vital economic resource. Changes to reindeer herding caused by increased precipitation and temperature and resultant changes in vegetation composition and nutritive value are difficult to predict. Displacement of high quality forage by low quality or non-food species as described in section 11.5.1.1 may severely reduce grazing lands impacting the economies of reindeer herders (Bryant *et al.* 1991). Changes in timing of break-up or increased river levels could make it difficult to move animals between pasturage.

11.6.3.3. Transportation

Warmer temperature and a possible reduction in seasonal and annual sea ice cover may increase the opportunities for shipping in the Arctic through the Northern Sea Route and the Northwest Passage as well as by increasing ice-free days for some harbors. At the same time, regional decreases in ocean temperature may have the opposite effect and close off areas to year-round shipping. Changes to regional shipping practices including cruise ships will also depend on regional economic activity, which may be affected by a change in ocean access or other climate-affected parameters.

11.6.3.4. Forestry

The northern treeline is expected to move north as the climate warms and conditions become more favorable (MacDonald *et al.* 1993, Sveinbjornsson *et al.* 1992). In areas of nutrient availability trees are also expected to grow more quickly; this could supply more harvestable trees in the future. However, as a result of decreased summer precipitation and drier foliage and leaf litter, outbreaks of defoliating insects might increase resulting in increased forest fires as discussed in section 11.5.1.1.2.

Fire may have beneficial effects if the intensity of fire is not too great. Fire can fertilize the forest floor and stimulate growth of early successional deciduous trees that provide habitat for moose and other large herbivores. On the other hand, drier, windy conditions that may coincide with climatic warming may cause intense fires which burn off the organic layer and release nutrients from the system (Kasischke *et al.* 1995, Starfield and Chapin 1996, Flannigan and Van Wagner 1991).

11.6.3.5. Agriculture

Arable land in the Arctic is mainly limited by surface and soil temperatures. Temperature records from four sites in Alaska show a general increase in the agricultural growing season between 1924 and 1989 although regional patterns were inconclusive (Sharratt 1992). Lengthened growing season was due to earlier last spring freezes, later first fall freezes, or both. These trends are similar to mid-latitude trends and do not indicate an Arctic amplification of growing season length. In a study based on soil surveys and climate in Alaska and northwestern Canada, the impact of a doubling of CO_2 on arable land was modeled (Mills 1994). Potentially arable lands showed increased moisture deficits and an improved climatic capability for agriculture with an increase of nearly 160 000 km^2 (16 M Ha) of arable land in the study area due to warming temperatures. Moisture limitations, it is predicted, would limit the use of some of this land, and serious questions remain about the adaptation of existing plant species to higher CO_2 levels (Mills 1994).

11.6.4. Effects of UV radiation on human health

UV radiation is linked to a variety of skin disorders and an increase in UV dose to northern populations is a health concern. The effects of UV radiation on human health are covered more fully in chapter 12, section 12.2.4.2. The health effects of UV radiation include skin disorders (cancer, changes in elasticity and hypersensitivity), eye damage, and immune suppression. Skin disorders have been shown to affect different northern populations in different ways. Both carcinomas and malignant melanomas of the skin are less common in the Greenland Inuit population than in the general Danish population. Several reasons have been posed for this, including the low UV radiation historically received in the Arctic and the pigmentation of the native people. Changes in elastosis of the skin, which can be linked to UV exposure, have been observed in Inuit (Wulf *et al.* 1989). Erythema (sunburn) while a temporary irritant, has also been linked to malignant melanoma. Young children who get severe sunburns are particularly prone to skin cancer later in life (WHO 1993). A preliminary investigation has suggested that polymorphic light eruption, a photodermatosis resulting from sun exposure, is more common in the Arctic than at lower latitudes. The ratio of UV-B to UV-A has been proposed as an explanation for the disorder at high latitudes. Actinic prurigo, another idiopathic photodermatosis, has been found to be frequent in the Canadian Inuit population and has been linked to UV-B sensitivity (Orr and Birt 1984). Ozone depletion may, therefore, reduce the incidence of polymorphic light eruption and increase the incidence of actinic prurigo in the Arctic. Some of these effects are mitigated in the Arctic because most skin is covered by clothing for much of the year. Changes in personal exposure patterns can have a greater effect than small changes in stratospheric ozone.

Ocular damage linked to UV exposure includes photokeratitis, cataracts, and pterygium. The highly reflective snow covered surfaces in the Arctic can help reflect ambient UV to the eye, increasing the UV exposure to levels higher than those observed at lower latitudes. Photokeratitis, commonly referred to as snow blindness, has always been a problem in the Arctic, as evidenced by Inuit ocular shields. Cataract formation has been linked to a variety of causes, including UV radiation. There has been some evidence to link cortical cataracts with chronic ocular UV-B exposure (Dolin 1995). Increasing UV-B should be considered a problem which could be costly for remote health care (IASC 1995). Other eye diseases such as spheroidal degeneration of the cornea and conjunctivea (climatic droplet keratopathy) and pterygium may or may not be associated with UV-B exposure. However, pterygium mainly occurs under conditions of climatic extremes and might, therefore, be of special relevance for UV-B in the Arctic (IASC 1995, WHO 1993).

UV-B radiation is able to suppress the immune system which may exacerbate a variety of existing health threats. This immune suppression can cause the reactivation of viruses including herpes, chicken pox, Epstein-Barr, and the wart virus. Immunosuppression appears to be independent of skin pigmentation, and very small doses of UV-B are able to start the reaction (Olivarius *et al.* 1996). It is unclear how much exposure is necessary to induce a systemic immune suppression (IASC 1995).

11.7. International efforts

11.7.1. Agreements

The Montreal Protocol on Substances That Deplete the Ozone Layer and its amendments have set standards to reduce the production of CFCs and other ozone depleting substances. The impetus for these international agreements has been the discovery of ozone holes in Antarctica. Projections regarding the Arctic ozone layer are not directly addressed by these agreements but, because of the likely link between stratospheric cooling and ozone depletion in the Arctic, further investigation is warranted to determine if measures need to be taken to assure the protection of the Arctic stratospheric ozone layer. Compliance with the Montreal Protocol is still of some concern. Bulgaria, Poland, Russia, Ukraine, and Belarus have each said that they may not be able to reach their agreed phase-out goals. Evidence of illegal trade in the controlled substances may undermine the goals of the Montreal Protocol and amendments. Continued re-examination of the available science and compliance information is necessary to assure the success of the Montreal Protocol's efforts to reduce ozone depleting substances.

The United Nations convened the Framework Convention on Climate Change at the Rio Earth Summit in 1992. 162 countries signed the agreement to implement national policies to monitor and attempt to reduce greenhouse gas emissions. Significant to the Arctic climate, it was also agreed that sustainable management of greenhouse gas sinks (natural ecosystems which can remove greenhouse gases from the atmosphere) should be promoted. At the more recent Berlin conference, timeframes to reduce greenhouse gases were set, but no new targets were agreed upon. Appropriate greenhouse targets are critical to the Arctic environment because of its high sensitivity to climate change. While research to determine the appropriate targets may take many years, efforts should be made immediately to assure the most cautionary levels are agreed upon as soon as possible. The presently proposed system of management depends on individual countries reaching concentration goals within their country limits. This form of regulation may not be sufficient to protect the Arctic, which is under no single national control. The newly-formed Arctic Council, however, may take up this issue. Further efforts will be needed to assure that Arctic levels due to transport do not exceed set standards. The treaty signed in 1979 at the Geneva Convention concerning long-range transports may be invoked to aid in the control of greenhouse gases in the Arctic. Furthermore, while the United States and the European Union have agreed that a binding treaty should be drawn up with regard to greenhouse gas emissions, many other countries prefer that this agreement remain voluntary.

11.7.2. Programs

The Arctic has many unique attributes which make it particularly susceptible to climate change, ozone depletion, and UV enhancement, more so than mid-latitude and even Antarctic environments. While many programs address climate change at mid-latitudes and ozone in the Antarctic, the international community has only recently begun to address these problems in the Arctic. Individual efforts to understand climate change, ozone, and UV radiation in the Arctic are on-going at a number of institutions around the world. Several international programs have been started in the last decade to address issues of climate change, ozone, and UV radiation in the Arctic.

The EU has been active in climate change issues and has sponsored a number of projects related to ozone and UV in the European Arctic. The European Commission's Environment Research Programme has supported European Arctic Stratospheric Ozone Experiment (EASOE) and Second European Stratospheric Arctic and Mid-latitude Experiment (SESAME) as well as a recent project, Ultraviolet Radiation in the Arctic; Past, Present and Future (UVRAPPF) to examine ozone and UV in northern Europe. The European Commission has also funded a program of experimental research into the impacts of enhanced UV-B radiation on European heathlands (UVECOS), which includes several different ecosystems within the European Arctic from Lapland to High Arctic Svalbard. These efforts have revealed and brought into focus many questions which still need to be answered. Because issues of climate change, ozone depletion, and UV radiation affect the entire Arctic, these programs would benefit from expansion or coordination with North American and Russian research communities in order to obtain more complete information on the Arctic environment.

The World Meteorological Organization (WMO) has sponsored and coordinated a number of scientific activities related to climate change, ozone, and UV radiation. The Global Atmosphere Watch (GAW) program coordinates monitoring sites around the globe, including in the Arctic. The World Climate Research Programme within the WMO has established Stratospheric Processes and their Role in Climate (SPARC). These programs need to be expanded to give more emphasis to Arctic issues in line with the importance of the Arctic in the global climate system. The World Climate Research Programme has also established the Arctic Climate System Study (ACSYS) to examine freshwater balance, runoff, and sea ice in the Arctic. A useful product which may result from this program is the ACSYS precipitation data center to be established in Germany.

Toronto hosts the WMO's Ozone and UV Database center which offers data over the Internet. There is increased need to include more ozone and UV data from the Arctic in this center. The responsibility for inputting ozone and UV data lies with the individual countries and agencies collecting the data. The data, once available, need to be examined in a regional, rather than national, perspective in order to understand observations with respect to the Arctic. This is beyond the present scope of the WMO data center.

The Nordic Ozone and UV Group has coordinated research on ozone and UV for the Nordic countries. This group has been very effective in focusing research needs and intercomparing results. Further benefits would be gained from participation from Canada, the United States, and Russia.

The International Arctic Science Committee (IASC), a non-governmental organization established in 1990 to encourage and facilitate Arctic research, has proposed a comprehensive program to assess the effects of UV radiation on the Arctic environment and the regional impacts of climate change in the Barents and Bering Seas. The program is presently in search of funding to implement the proposed research program.

11.7.3. Assessments

This AMAP document is the first document sponsored by a large, international committee to assess what is known about climate change, ozone, and UV, and their effects in the Arctic. Some of the information presented in this chapter has been presented in other assessments. For the most part, however, the information has not previously been brought together in the context of Arctic research. Several important assessments have been written concerning global climate change, ozone, and UV radiation, yet most effort has been placed on northern mid-latitude and Antarctic changes. Considerably less effort has been placed on the Arctic. In some cases, the Arctic environment is not addressed at all; in other cases, the Arctic environment is addressed as part of the polar environments. Such grouping ignores the vast differences between Arctic and Antarctic environments and concerns.

11.7.3.1. Climate change

The Intergovernmental Panel on Climate Change (IPCC), jointly established by the World Meteorological Organization and the United Nations Environmental Programme in 1988, has produced a number of documents, most notably four scientific assessments (IPCC 1990a, 1992a, 1994, 1996a) summarizing the present understanding of the science of climate change, and four assessments of the impacts of climate change (IPCC 1990b, 1992b, 1996b). These assessments have focused on modeling and model predictions for future levels. The IPCC documents have illuminated serious problems in the existing models for Arctic climate. At present no international programs exist to promote the advancements of the climate models for future predictions in the Arctic, although scientists continue to work on improving their individual models. The IPCC scientific assessments take a more general view and do not focus on any region of the Earth. Because of the unique qualities of the Arctic environment and its interactive processes, many of the general results are not applicable to the Arctic. Focused effort, on the part of the IPCC, to examine the Arctic's responsivity to climate change through available data from the Arctic would be appropriate and useful to understanding global climate change.

11.7.3.2. Ozone and UV

The World Meteorological Organization, in conjunction with the United Nations Environmental Program (UNEP), the US National Oceanic and Atmospheric Administration (NOAA), and the US National Aeronautic and Space Agency (NASA), has produced several assessments addressing the present understanding of ozone depletion. The most recent report, Ozone Depletion: 1994 (WMO 1994), documents observed ozone loss in the Arctic which is consistent with photochemical model calculations. However, the recent ozone anomalies, which had not been well documented when the last report was written, need to be addressed in future documents. The Arctic has been grouped together with the Antarctic in past reports. In light of recent understanding of the different character of ozone in the Arctic versus the Antarctic, future assessments should address the Arctic separately, with full attention to what is known about the Arctic and the observed anomalies. Future assessments should also address the issue of UV radiation in the Arctic in more depth than previous reports.

UNEP has co-sponsored several important reports on UV radiation and its effects. However, very little attention has been given to the Arctic environment in any of these reports. The Scientific Committee on Problems of the Environment (SCOPE) has also sponsored several important reports on UV radiation, again with no significant attention to the Arctic. The effects of UV radiation in the Arctic were reviewed by the International Arctic Science Committee (IASC). The resulting reports offer an extensive evaluation of what is known about UV effects in the Arctic and underlines how little is known about the effects of UV on Arctic species, ecosystems, and human inhabitants (IASC 1995, 1996).

11.8. Conclusions and recommendations

11.8.1. Climate change: Conclusions

The climate of the Arctic affects both the inhabitants of the Arctic and the global climate system. Anthropogenically-driven climate change is likely to be most severe in the Arctic because of the strong feedback mechanisms. The Arctic has undergone rapid changes in the past. Recent changes have been linked to anthropogenic forcing; but despite the global consequence of changes in the Arctic climate, monitoring of basic climatic parameters, such as sea ice, precipitation, and the atmosphere, are not adequate to assess these recent changes. Present monitoring in the Arctic is not adequate to assess the impacts of global climate change. The uneven geographical location of present monitoring stations limits the present ability to understand the climate system.

In preparing this document it became evident that knowledge from Arctic people is underutilized in climate change research. Arctic peoples have passed down information through generations on climate change which has been useful for their survival. Traditional knowledge is a valuable source of information on present climate dynamics and needs to be documented and included in future assessments.

Changes in sea ice coverage and thickness are the primary driving forces in the polar amplification of global warming forecast by many climate models. Yet there are few continuous and ongoing measurements of the surface energy balance, of the ice thickness distribution, or of the structure of the upper ocean. In the past, the drifting ice stations of the Former Soviet Union provided many key measurements, but that series of manned camps has ended and none from any nation are likely to replace them.

Prediction of Arctic climate is not reliable at this time. The complex interactions of the various components of the Arctic make climate modeling and therefore climate predictions extremely difficult and uncertain at this time. The climate models in existence do not agree well with current measurements, nor do they agree with each other with respect to predictions when applied to the Arctic. Sea ice dynamics and trace gas balance, particularly over land, are very important aspects of the climate system in the Arctic, yet very little is known about changes taking place in either of these components.

11.8.2. Climate change: Recommendations

There is a strong need to improve understanding of the climate system and to develop reliable methods of climate prediction for the Arctic. Baseline information is needed throughout the Arctic to be able to assess any future changes. Ice and lake sediment cores, peat cores, biological records, archaeological records, and historical documents should be examined to determine the climate of the past few hundred years for the Arctic. Traditional knowledge needs to be documented and incorporated into the present scientific knowledge of climate change. Crucial aspects of climate in the Arctic, particularly the hydrological and trace gas cycles, are insufficiently understood. Increased knowledge of these systems is necessary before reliable predictions can estimate future changes. The role of gaseous exchange in Arctic terrestrial areas requires further investigation to understand both the Arctic climate and the global climate. Existing atmospheric models need to better incorporate the dynamical nature of the Arctic feedback mechanisms to provide valid estimates.

Every effort should be made to increase the number of climatologically useful observations on, above, and below the ice pack, with a concentration of effort in critical regions that are particularly sensitive to climate change, such as shelf regions and the Arctic Ocean's primary outflow region, Fram Strait. Buoys, moorings, aerial surveys, and satellite remote sensing are likely sources of the information needed to monitor the state of the Arctic ice pack. Increased support of such measurement programs is important to understanding changes in Arctic climate.

11.8.3. Ozone: Conclusions

In the last 20 years, ozone concentrations in the Arctic have been changing more than those at mid-latitudes, with observed trends of in excess of 10% per decade. Of more concern than the average trend is the occurrence of short-term episodes of extreme ozone depletion. There are indications that the occurrence of these episodes is increasing, yet their full cause is not well understood. Because ozone depletion in the Arctic is linked to stratospheric cooling, future climate change is expected to have a strong influence on future ozone levels. The reductions of CFC's alone may not be enough to restore the Arctic ozone layer to pre-1980s levels.

11.8.4. Ozone: Recommendations

The recent changes in the ozone concentrations in the Arctic require further examination. Both the mean trends and distributions in the Arctic ozone layer need to be understood before confidence may be placed on predictions about the Arctic ozone layer. There is a pressing need to accurately explain the mechanisms responsible for the recently documented ozone anomalies in the Arctic. The cause and impact of these anomalies need to be further understood. The potential link between climate change, in particular the cooling of the stratosphere, and Arctic ozone concentrations requires immediate further research.

11.8.5. UV: Conclusions

UV radiation in the Arctic is not negligible, as is often assumed, because of the large amount of diffuse radiation.
In the Arctic, UV is difficult to measure and difficult to estimate from satellite information, yet existing UV ground-monitoring stations in the Arctic are very unevenly distributed, so that the present spatial coverage of UV monitoring

is insufficient. The increase of UV during the observed ozone depletion events needs further study, particularly because these events often occur in spring-time when snow cover is decreasing and ecosystems are extremely sensitive.

11.8.6. UV: Recommendations

The existing UV monitoring efforts in the Arctic need to be expanded to give more broad geographical coverage; monitoring is particularly needed in the Russian Arctic. Merging the available UV data with satellite information to develop models for estimating UV throughout the entire Arctic is necessary for determining a broad UV climatology of the Arctic. Both the UV reaching vertical surfaces and that reaching horizontal surfaces are relevant to the ecosystems in the Arctic, and more research is needed to determine the character and climatology of ambient UV levels. UV during episodes of severe ozone depletion events needs to be understood further.

11.8.7. Climate change and UV effects on ecosystems: Conclusions

Despite the general understanding of possible climate changes to the Arctic, little is known about the specific effects of climate change on marine or terrestrial ecosystems or on individual species. Little is known about widespread effects of climate change which have occurred already because of the lack of ecosystem monitoring and the complexity of the feedback mechanisms in the Arctic. Changes in sea ice, snow, and permafrost extent will determine the available habitat for plants and animals in the Arctic. Climate changes are likely to take place faster than Arctic ecosystems are likely to be able to respond.

The effects of UV radiation on Arctic ecosystems are not well understood. Many of the assumptions made for the Arctic are based on UV studies carried out in the Antarctic or at mid-latitudes. Those UV effects that have been observed appear to be tied to such factors as water stress and other environmental factors that are also influenced by climate change. The springtime changes in ozone concentrations could expose primary production in the Arctic Ocean as well as the terrestrial ecosystems to harmful UV radiation because of the timing and intensity of the changes; changes in primary production would impact higher trophic levels.

11.8.8. Climate change and UV effects on ecosystems: Recommendations

Climate change effects on ecosystems need to be examined in terms of an integrated assessment, taking into account relevant environmental factors such as acidification, UV and pollutants, as well as the effects of climate change on competing species within an ecosystem. The potential for climate change to alter the Arctic environment demands immediate assessment of available baseline information as well as fundamental research into the effects of change on particular species as well as ecosystems. UV effects need to be examined in terms of an integrated assessment, taking into account relevant environmental factors, such as acidification and water stress, as well as the effects of UV on competing species within an ecosystem. Both aquatic and terrestrial ecosystems need to be examined through efforts such as IASC's proposal to establish and coordinate focused centers of research in the Arctic for the purpose of studying UV effects.

11.8.9. Climate change and UV effects on humans: Conclusions

The effects of climatic warming on the Arctic ecosystems are likely to be large and to have a noticeable impact on present human ways of life. Climate change, through its effect on physical properties of the land and ocean, could greatly impact infrastructure, especially transportation and large structures; this could make the Arctic more accessible to development. Agriculture, forestry, and fisheries will likely be altered. While these changes may include some economic benefits, they are likely to permanently impact traditional ways of life. Sea level rise could displace many permanent communities.

Effects of UV radiation on human health in the Arctic are larger than previously suspected. The strong seasonal cycle often exposes populations to high UV levels in the early spring when they have not yet developed protective pigmentation. UV has been linked to dermatological, ocular, and immunosuppression effects, all of which can have severe impacts on the health of Arctic people. The low sun angle and high reflectivity of snow make UV even more hazardous for Arctic populations than previously thought. Because of the high cost of administering health care in the Arctic, UV effects due to decreasing ozone are likely to have high economic impact.

11.8.10. Climate change and UV effects on humans: Recommendations

Of particular importance is an examination of the effects of UV and climate change on Arctic peoples and on the plants, mammals, and fish they harvest for food, on grazing animals, and on the physical environment. It is important to include indigenous people in future policy planning and research efforts in order to understand the impacts of both UV and climate change on traditional ways of life.

The examination of the direct effect of UV radiation on human health requires immediate attention, particularly with respect to ocular damage and additionally to immunosupression effects and dermatological disorders.

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Authors
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References

Aagaard, K. and E.C. Carmack, 1989. The role of sea ice and other fresh water in the Arctic circulation. *J. Geophys. Res.* **94**: 11485-11498.

Aagaard, K. and E.C. Carmack, 1994. The Arctic ocean and climate: A perspective in the polar oceans and their role in shaping the global environment. pp 5-20, *In*: O.M. Johannessen, R.D. Muench and J.E. Overland (eds.). The Nansen Centennial Volume, Geophys. Monogr. Ser. 85, AGU, Washington, D.C.

Aagaard, K., A.T. Roach and J.D. Schumacher, 1985. On the wind-driven variability of the flow through Bering Strait. *J. Geophys. Res.* **90**: 7213-7221.

Aagaard, K., L.K. Coachman and E.C. Carmack, 1981. On the halocline of the Arctic Ocean. *Deep-Sea Res.* **28**: 529-545.

Ackley, S.F., W.D. Hibler III, F.K. Kugzruk, A. Kovacs and W.F. Weeks, 1976. Thickness and roughness variations of Arctic multiyear sea ice. CRREL Report 76-18, U.S. Army Cold Regions Research and Engineering Laboratory, Hanover, New Hampshire. 25p.

Agnew, T., 1993. Simultaneous winter sea-ice and atmospheric circulation anomaly patterns. *Atmosphere-Ocean* **31**: 259-280.

Akagawa, S. and M. Fakuda, 1991. Frost heave mechanism in welded tuff. *Permafrost and Periglacial Processes* **2**: 301-309.

Allard, B., H. Born, C. Pettersson and G. Zhang, 1994. Degradation of humic substances by UV radiation. *Environ. Internat.* **20**: 97-101.

Allard, R., R. Preller and S Piacsek, 1995. Variations of river influences in the Arctic basin. Fourth Conference on Polar Meteorology and Oceanography, Dallas, Texas, American Meteorological Society, 193-199.

Alley, R.B., 1995. Resolved: the Arctic controls global climate change. pp. 263-283, *In*: W.O Smith Jr. and J.M. Grebmeier (eds.). Arctic Oceanography: Marginal ice zone and continental shelves. American Geophysical Union Coastal and Estuarine Studies 49.

Alley, R.B. and S. Anandakrishnan, 1995. Variations in melt-layer frequency in the GISP2 ice core: implications for Holocene summer temperatures in central Greenland. *Annals of Glaciology* **21**: 64-70.

Alley, R.B., D.A. Meese, C.A. Shuman, A.J. Gow, K.C. Taylor, P.M. Grootes, J.W.C. White, M. Ram, E.D. Waddington, P.A. Mayewski and G.A. Zielinski, 1993. Abrupt increase in Greenland snow accumulation at the end of the Younger Dryas event. *Nature* **362**: 527-528.

Alley, R.B., P. Mayewski, D. Peel and B. Stauffer, 1996. Twin ice cores from Greenland reveal history of climate change. *EOS (American Geophysical Union)* **77**(**22**): 209-210.

Andersen, O.G.N., 1989. Primary production, chlorophyll, light and nutrients beneath the Arctic sea ice. pp. 147-191, *In*: Y. Herman (ed.). The Arctic Seas: Climatology, Oceanography, Geology, and Biology. Ban Nostrand Reinhold Co., N.Y.

Anderson, B.E., G.L. Gregory, J.D.W. Barricks, J.E. Collins Jr., G.W. Sachse, M.C. Shipham and C.H. Hudgins, 1994. Summertime tropospheric ozone distributions over central and eastern Canada. *J. Geophys. Res.* **99**: 1781-1792.

Anderson, S.P., 1988. The upfreezing process: experiments with a single clast. *Geol. Soc. of Amer. Bull.* **100**: 609-21.

Andreas, E.L. and B. Murphy, 1986. Bulk transfer coefficients for heat and momentum over leads and polynyas. *J. Phys. Oceanogr.* **16**(**11**): 1875-1883.

Andrée, M., H. Oeschger, U. Siegenthaler, T. Riesen, M. Moell, B. Ammann and K. Toboloski, 1986. ^{14}C dating of plant macrofossils in lake sediments. *Radiocarbon* **28**: 411-416.

Arkhangelov, A.A. and E.V. Novgorodova, 1991. Genesis of massive ice at "Ice Mountain", Yenesei river, western Siberia, according to results of gas analyses. *Permafrost and Periglacial Processes* **2**(**2**): 167-170.

Armstrong, R.L. and M.J. Brodzik, 1995. An earth-gridded SSM/I data set for cryospheric studies and global change monitoring. *Advances in Space Research* **16**(**10**): 155-163.

Astakhov, V. and L. Isayeva, 1988. The "Ice Hill"; an example of retarded glaciation in Siberia. *Quaternary Science Reviews* **7**: 29-40.

Austin, J. and N. Butchart, 1994. The influence of climate change and the timing of stratospheric warmings on Arctic ozone depletion. *J. Geophys. Res.* **99**: 1127-1145.

Austin, J.H., N. Butchart and K.P. Shine, 1992. Possibility of an Arctic ozone hole in a doubled-CO_2 climate. *Nature* **360**: 221-225.

Bachmeister, A.S., M.C. Shipham, E.V. Browell, W.B. Grant and J.M. Klassen, 1994. Stratospheric/tropospheric exchange affecting the northern wetlands regions of Canada during summer 1990. *J. Geophys. Res.* **99**: 1793-1804.

Backlund, P., 1992. Degradation of aquatic humic material by UV light. *Chemosphere* **25**: 1869-1878.

Barlow, L.K., J.P. Sadler, A.E.J. Ogilvie, P.C. Buckland, T. Amorosi, J.H. Ingimundarson, P. Skidmore, A.J. Dugmore, T.H. McGovern, 1997. Bad weather, bad luck or bad planning? Interdisciplinary investigations of the end of the Norse Western Settlement in Greenland. *The Holocene*. In review.

Barnes, P.W., S.D. Flint and M.M. Caldwell, 1987. Photosynthesis damage and protective pigments in plants from a latitudinal arctic/alpine gradient exposed to supplemental ultraviolet-B radiation in the field. *Arctic Alpine Res.* **19**: 21-27.

Barrie, L.A., J.W. Bottenheim and W. Hart, 1994. An overview of Polar Sunrise Experiment 1992. *J. Geophys. Res.* **99D**: 25313-25314.

Barrie, L.A., J.W. Bottenheim, R.C. Schnell, P.J. Crutzen and R.A. Rasmussen, 1988. Ozone destruction and photochemical reactions at polar sunrise in the low arctic atmosphere. *Nature* **334**: 138-141.

Barrie, L.A., R.M. Hoff and S.M. Daggupaty, 1981. The influence of mid-latitudinal pollution sources on haze in the Canadian Arctic. *Atmos. Environ.* **15**(**8**): 1407-1419.

Barry, P. and W. Pollard, 1992. Ground probing radar investigation of ground ice on the Fosheim Peninsula, Ellesmere Island, N.W.T. Musk-Ox, 39: 59-67.

Barry, R G., M.C. Serreze, J.A. Maslanik, and R.H. Preller, 1993a. The Arctic sea ice climate system: observations and modelling. *Rev. Geophys.* **31**: 397-422.

Barry, R.G, J.A. Maslanik, K. Steffen, R.L. Weaver, V. Troisi, D.J. Cavalieri, and S. Martin, 1993b. Advances in sea ice research based on remotely sensed passive microwave data. *Oceanography* **6**(**1**): 5-13.

Barry, R.G., R.L. Armstrong and J.-M. Fallot, 1995. Assessing decadal changes in the cryosphere: Eurasian snow cover, Proc. Fifth Symposium of Global Change Studies, Nashville TN, American Meteorological Society, 148-155.

Barry, R.G., 1989. The present climate of the Arctic Ocean and possible past and future states. pp. 1-46, *In*: Y. Herman (ed.). The Arctic Seas: Climatology, Oceanography, Geology, and Biology. Ban Nostrand Reinhold Co., New York.

Barry, R., 1985. The cryosphere and climatic change. pp. 109-148, *In*: M.C. MacCracken and F.M. Luther (eds.). Detecting the Climatic Effects of Increasing Carbon Dioxide. U.S. Dept. Energy, Washington, D.C. DOE/ER-0235.

Barsch, D., M. Gude, R. Mausbacher, G. Schukraft, A. Schulte and D. Strauch, 1993. Slush stream phenomena – Process and impact. *Zeitschrift für Geomorphologie* **92**: 39-53.

Bartlein, P.J., I.C. Prentice and T. Webb III, 1986. Climate response surfaces from pollen data for some eastern North Amercan taxa. *J. of Biogeography* **13**: 35-57.

Bauer, R., A. Engel, H. Franken, E. Klein, G. Kulessa, C. Schiller, U. Schmidt, R. Borchers and J. Lee, 1994. Monitoring the vertical structure of the Arctic polar vortex over northern Scandinavia during EASOE: Regular N_2O profile observations. *Geophys. Res. Lett.* **21**: 1211-1214.

Baumgartner, A. and E. Reichel, 1975. The World Water Balance. Mean annual global, continental, and maritime precipitation, evaporation and runnoff. Elsevier, Amsterdam. 179p.

Behrenfeld, M.J., D.R.S. Lean and H. Lee, 1995. Ultraviolet-B radiation effects on inorganic nitrogen uptake by natural assemblages of oceanic plankton. *J. Phycol.* **31**: 25-36.

Behrenfeld, M.J., J.T. Hardy and H. Lee, 1992. Chronic effects of ultraviolet-B radiation on growth and cell volume of *Phaeodactylum tricornutum* (Bacillariophyceae). *J. Phycol.* **58**: 757-760.

Beltrami, H. and D. Chapman, 1994. Drilling for a past. *New Scientist* **142**, April 23: 36-40.

Bender, M.L., T. Sowers, M.-L. Dickson, J. Orchardo, P. Grootes, P.A. Mayewski and D.A. Meese, 1994. Climate correlations between Greenland and Antarctica during the past 100,000 years. *Nature* **372**: 663-666.

Berendse, F. and S. Jonasson, 1992. Nutrient use and nutrient cycling in northern ecosystems. pp 337-356, *In*: F.S. Chapin III, R.L. Jefferies, J.F. Reynolds, G.R. Shaver and J. Svoboda (eds.). Arctic ecosystems in a changing climate; an ecophysiological perspective. Academic Press, San Diego, California.

Bird, R.E. and C. Riordan, 1986. Simple solar spectral model for direct and diffuse irradiance on horizontal and tilted planes at the earth's surface for cloudless atmospheres. *J. of Climate and Applied Meteorology.* **25**: 97-97.

Birks, H.H., 1991. Holocene vegetational history and climate change in West Spitsbergen: plant macrofossils from Skadtj_rna, an Arctic lake. *The Holocene* **1**: 209-218.

Bjorn, L.O., T.V. Callaghan, C. Gehrke, D. Gwynn-Jones, B. Holmgren, U. Johanson and M. Sonesson, 1997. Effects of enhanced ultraviolet-B radiation on subarctic vegetation. Chapter 11, pp 241-248, *In*: S. Woodin and M. Marquiss (eds.). Ecology of Arctic Environments, Special Publication No. 13 of the British Ecological Society. Blackwell Science, Oxford.

Blake, D.R. and Rowland, F.S., 1988. Continuing worldwide increase in tropospheric methane, 1978 to 1987. *Science* **240**: 634-637.

Blanchet, J.P., 1989. Toward estimation of climatic effects due to Arctic aerosols. *Atmos. Environ.* **23**(**11**): 2609-2625.

Blanchet, J.P., 1991. Potential climate change from Arctic air pollution. *In*: W.T. Sturges (ed.). Pollution of the Arctic Atmosphere. Elsevier Press, London.

Blanchet, J.P. and E. Girard, 1994. Arctic Greenhouse Effect. *Nature* **371**: 383-.

Blanchet, J.P. and E. Girard, 1995. Water vapour-temperature feedback in the formation of continental Arctic air: Its implication for climate. *Sci. Total Environ.* **160/161**: 793-802

Blumthaler, M., 1993. Solar UV measurements. *In*: M. Tevini (ed.). Ultraviolet-B Radiation and ozone depletion: Effects on humans, animals, plants, microorganisms and materials. Lewis Pub., Boca Raton, Florida.

Blumthaler, M. and W. Ambach, 1988. Solar ultraviolet-B-albedo of various surfaces. *Photochem. and Photobiol.* **48**: 85-88.

Bo, S. and O. Hjeljord, 1991. Do continental moose ranges improve during cloudy summers? *Can. J. Zool.* **69**: 1875-1879.

Bodhaine, B. A. and E. G. Dutton, 1993. A long-term decrease in arctic haze at Barrow, Alaska. *Geophys. Res. Lett.* **20**(**10**): 947-950.

Bodhaine, B.A. and E.G. Dutton, 1995. Reply. *Geophys. Res. Lett.* **22(6)**: 741-742.

Boer, G.J., K. Arpe, M. Blackburn, M. Deque, W.L. Gates, T.L. Hart, H. le Treut, E. Roeckner, D.A. Sheinin, I. Simmonds, R.N.B. Smith, T. Tokioka, R.T. Wetherald and D. Williamson, 1992a. Some results from an intercomparison of the climate simulated by 14 atmospheric general circulation models. *J. Geophys. Res.* **97**: 12771-12786.

Boer, G.J., N.A. McFarlane and M. Lazare, 1992b. Greenhouse gas-induced climate change simulated with the CCC second-generation general circulation model. *J. Climate* **5**: 1045-1077.

Bojkov, R.D., L. Bishop, W.J. Hill, G.C. Reinsel and G.C. Tiao, 1994. A statistical trend analysis of revised Dobson total ozone data over the northern hemisphere. *J. Geophys. Res.* **95**:9785-9807.

Bojkov, R.D., V.E. Fioletov, D.S. Balis, C.S. Zerefos, T.V. Kadygrova and A.M. Shalamjansky, 1995a. Further ozone decline during the northern hemisphere winter-spring of 1994-1995 and the new record low ozone over Siberia. *Geophys. Res. Lett.* **22**: 2729-2732.

Bojkov, R.D., V.E. Fioletov, T.V. Kadygrova, K.I. Romashkina and A.M. Shalamyanskii, 1995b. Estimating ozone depletion of Eurasia in 1973-1995 using reevaluated filter ozonometer data. *Meteorologia I. Gidrologia* **9**: 30-40.

Bond, G., W. Broecker, S. Johnsen, J. McManus, L. Labeyrie, J. Jouzel and G. Bonani, 1993. Correlations between climate records from North Atlantic sediments and Greenland ice. *Nature* **365**: 143-147.

Booth, C.R. and S. Madronich, 1994. Radiation amplification factors: Improved formulation accounts for large increases in ultraviolet radiation associated with Antarctic ozone depletion. pp. 39-42, *In*: C.S. Weiler and P.A. Penhale (eds..), Ultraviolet Radiation in Antarctica: Measurements and Biological Research, AGU Antarctic Research Series, volume 62, American Geophysical Union, Washington, D.C.

Borys, R., 1989. Studies of ice nucleation by Arctic aerosol on AGASP-II. *J. Atmos. Chem.* **9, 1-3**: 169-185.

Bothwell, M.L., D.M.J. Sherbot, and C.M. Pollock, 1994. Ecosystem response to solar ultraviolet-B radiation; influence of trophic level interactions. *Science* **265**: 97-100.

Boucher, N.P. and B.B. Prézelin, 1996. Spectral modeling of UV inhibition of in situ Antarctic primary production using a field-derived biological weighting function. *Photochem. and Photobiol.* **64** (**3**): 407-418.

Bourke, R.H. and R.P. Garrett, 1987. Sea ice thickness distribution in the Arctic Ocean. *Cold Reg. Sci. Technol.* **13**: 259-280.

Bourke, R.H. and A.S. McLaren, 1992. Contour mapping of Arctic Basin ice draft and roughness parameters. *J. Geophys. Res.* **97**: 17715-17728.

Bradley, R.S, F.T Keimig and H.F. Diaz, 1993. Recent changes in the North American Arctic boundary layer in winter. *J of Geophys. Res.* **98D**: 8851-8858.

Bradley, R.S, H.F. Diaz, J.K. Eischeid, P.D. Jones, P.M. Kelly and C.M. Goodess, 1987. Precipitation fluctuations over Northern Hemisphere land areas since the mid-19th century. *Science* **237**: 171-175.

Bregman, A., P.F.J. van Velthoven, F.G. Wienhold, H. Fischer, T. Zenker, A. Waibel, A. Frenzel, F. Arnold, G.W. Harris, M.J.A. Bolder and J. Lelieveld, 1995. Aircraft measurements of O_3, HNO_3 and N_2O in the winter Arctic lower stratosphere during the stratosphere-troposphere experiment by aircraft measurements (STREAM) 1. *J. Geophys. Res.* **100**: 11245-11260.

Broecker, W.S., 1991. The great ocean conveyor. *Oceanography* **4(2)**: 79-89.

Broecker, W.S., 1994. Massive iceberg discharges as triggers for global climate change. *Nature* **372**: 421-424.

Broecker, W.S., G. Bond and M. Klas, 1990. A salt oscillator in the glacial Atlantic? 1. The concept. *Paleoceanography* **5**: 469-477.

Broecker, W.S., T.- H. Peng, G. Ostlund and M. Stuiver, 1985a. Distribution of bomb radiocarbon in the ocean. *J. of Geophys. Res.* **90(C4)**: 6953-6970.

Broecker, W. S., D.M. Peteet and D. Rind, 1985b. Does the ocean-atmosphere system have more than one stable mode of operation. *Nature* **315(6014)**: 21-26.

Brook, E.J., T. Sowers and J. Orchardo, 1996. Rapid variations in atmospheric methane concentrations during the past 110,000 years. *Science* **273**: 1087-1091.

Browell, E.V., C.F. Butler, S.A. Kooi, M.A. Fenn, R.C. Harriss and G.L. Gregory, 1992. Large-scale variability of ozone and aerosols in the summertime arctic and subarctic troposphere. *J. Geophys. Res.* **97**: 16433-16450.

Browell, E.V., M.A. Fenn, C.F. Butler, W.B. Grant, R.C. Harriss and M.C. Shipham, 1994. Ozone and aerosol distributions in the summertime troposphere over Canada. *J. Geophys. Res.* **99**: 1739-1755.

Brown, R., M. Hughes and D. Robinson, 1995. Characterizing the long-term variability of snow cover extent over the interior of North America. *Annals of Glaciology* **21**: 45-50.

Brühl, C. and P.J. Crutzen, 1989. On the disproportionate role of tropospheric ozone as a filter against solar ultraviolet-B radiation. *Geophys. Res. Lett.* **16**: 703-706.

Brune, W.H., 1990. Ozone crisis: The case against chlorofluorocarbons. Weatherwise 43: 136-143.

Bryant, J.P., P.B. Reichardt and T.P. Clausen, 1991. Effects of global warming on woody plant secondary metabolism in high latitude ecosystems. *In*: F.S. Chapin III, R. Jefferies, G. Schaver, J. Reynolds and J. Sovboda (eds.). Physiological ecology of Arctic plants: Implications for climate change. Academic Press Inc.

Bryson, R.A., 1966. Air masses, streamlines, and the boreal forest. *Geographical Bull.* **8**: 228-269.

Buckland, P.C., T. Amorosi, L.K. Barlow, A.J. Dugmore, P.A. Mayewski, T.H. McGovern, A.E.J. Ogilvie, J.P. Sadler, and P. Skidmore, 1996. Bioarchaeological and climatological evidence for the fate of Norse farmers in medieval Greenland. *Antiquity* **70**: 88-96

Bullock, A.M., 1985. The effect of ultraviolet-B radiation upon the skin of plaice *Pleuronectes platessa* infested with the bodonid ectoparasite *Ichtyobodo necator*. *J. Fish. Dis.* **8**: 547-550.

Bullock, A.M. and R.R. Coutts, 1985. The impact of ultraviolet-B radiation upon the skin of rainbow trout, *Salmo gairdneri* Richardson, farmed at high altitude in Bolivia. *J. Fish. Dis.* **8**: 263-272.

Burn, C., 1990. Implications for palaeoenvironmental reconstruction of recent ice-wedge development at Mayo, Yukon Territory. *Permafrost and Periglacial Processes* **1**: 3-14.

Burn, C. and M. Smith, 1990. Development of thermokarst lakes in the Holocene at sites near Mayo, Yukon Territory, Canada. pp. 700-705, *In*: Permafrost, Fifth International Conference, Proceedings, Volume 1, Trondheim, Norway. Tapir Publishers, Trondheim.

Buzuev, A.I.A. and V.F. Dubovtsev, 1978. Certain patterns in the distribution of snow and ice cover thickness on the Arctic seas. Soviet meteorology and hydrology, Mar. 1978; No. 3: 44-48.

Caldwell, M.M., A.H. Teramura and M. Tevini, 1989. The changing solar ultraviolet climate and the ecological consequences for higher plants. Trends in Ecology and Evolution 4, 363-367.

Callaghan T.V. and S. Jonasson, 1995. Arctic terrestrial ecosystems and environmental change. *Philosophical Transactions of the Royal Society, London* **352**: 259-276.

Cao, M., 1993. The response of seasonal snow cover of Xinjiang Uygur Autonomy in western China to climate change. Exchange Processes at the Land Surface for a Range of Space and Time Scales, Proc. Yokohama Symp., Yokohama, Japan. IAHS Publ. 212: 521-526.

Carmack, E.C., 1986. Circulation and mixing in ice-covered waters. pp. 641-712, *In*: N. Untersteiner (ed.). The geophysics of sea ice. NATO ASI B146. Martinus Nijhoff Publ., Dordrecht, Netherlands.

Carroll, S.S. and T.R. Carroll, 1993. Increasing the precision of snow water equivalent estimates obtained from spatial modeling of airborne and ground-based snow data. Proceedings of the 50th Eastern Snow Conference, Quebec City, 83-87.

Carroll, T.R., 1995. GIS used to derive operational hydrologic products from in situ and remotely sensed snow data. pp. 335-342, *In*: A. Carrara and F. Guzzetti (eds.). Geographical Information Systems in Assessing Natural Hazards, Kluwer Academic Publ., Netherlands.

Catchpole, A.J.W., 1992. Hudson's Bay Company ships' log-books as sources of sea-ice data, 1751-1870. pp. 17-39, *In*: R.S. Bradley and P.D. Jones (eds.). Climate Since A.D. 1500, Routledge, London.

Cavalieri, D.J. and C.L. Parkinson, 1981. Large-scale variations in observed Antarctic sea ice extent and associated atmospheric circulation. *Mon. Wea. Rev.* **109**: 2323-2336.

Cess, R.D., G.L. Potter, M.-H. Zhang, J.-P. Blanchet, S. Chalita, R. Colman, D.A. Dazlich, A.D. Del Genio, V. Dymnikov, V. Galin, D. Jerrett, E. Keup, A.A. Lacis, H. le Treut, X.-Z. Liang, J.-F. Mahfouf, B.J. McAvaney, U.P. Meleshko, J.F.B. Mitchell, J.-J. Morcrette, P.M. Norris, D.A. Randall, L. Rikus, E. Roeckner, J.-F. Royer, U. Schlese, D.A. Sheinin, J.M. Slingo, A.P. Sokolov, K.E. Taylor, W.M. Washington, R.T. Wetherald and I. Yagai, 1991. Interpretation of snow-climate feedback as produced by 17 general circulation models. *Science* **253**: 888-892.

Chandrasekhar, S., 1960. Radiative Transfer, Dover Publications, Inc., New York.

Chapin III, F.S. and C. Körner (eds.), 1995. Arctic and alpine biodiversity: patterns, causes and ecosystem consequences. Springer Verlag, Heidelberg.

Chapin III, F.S. and G.R. Shaver, 1996. Physiological and growth responses of arctic plants to a field experiment simulating climatic change. *Ecology* **77**: 822-840.

Chapin III, F.S., R.L. Jefferies, J.F. Reynolds, G.R. Shaver and J. Svoboda, 1992. Arctic ecosystems in a changing climate: An ecological perspective. Academic Press, San Diego.

Chapin, D.M., and C.S. Bledsoe, 1992. Nitrogen fixation in arctic plant communities. pp. 301-319, *In*: F.S. Chapin III, R.L. Jefferies, J.F. Reynolds, G.R. Shaver, and J. Svoboda (eds.). Arctic ecosystems in a changing climate: An ecophysiological perspective. Academic Press, San Diego.

Chapman, W.L. and J.E. Walsh, 1993. Recent variations of sea ice and air temperature in high latitudes. *Bull. Amer. Meteor. Soc.* **73**: 34-47.

Charlson, R.J., J. Langner, H. Rodhe, C.B. Leovy and S.G. Warren, 1991. Perturbation of the northern hemisphere radiative balance by backscattering from anthropogenic sulfate aerosols. *Tellus* **43AB**: 152-163.

Charlson, R.J., S.E. Schwartz, J.M. Hales, R.D. Cess, J.A. Coakley Jr., J.E. Hansen and D.J. Hofmann, 1992. Climate forcing by anthropogenic aerosols. *Science* **255**: 423-430.

Chipperfield, M.P., A.M. Lee and J. Pyle, 1996. Model calculations of ozone depletion in the Arctic polar vortex for 1991/92 to 1994/95. *Geophys. Res. Lett.* **23**: 559-562.

Christensen, T.R., S. Jonasson, T.V. Callaghan and M. Havstrom, 1995. Spatial variation in high-latitude methane flux along a transect across Siberian and European tundra environments. *J. of Geophys. Res.* **100**: 21035-21045.

Christy, J.R., 1995. Temperature above the surface layer. *Clim. Change* **31**: 455-474.

Christy, J.R. and S.J. Drouilhet, 1994. Variability in daily, zonal mean lower-stratospheric temperatures. *J. Climate.* **7**: 106-120.

Christy, J.R., R.W. Spencer and R.T. McNider, 1995. Reducing noise in the daily MSU lower tropospheric temperature data set. *J. Climate.* **8**: 888-896.

Cicerone, R.J. and R.S. Oremland, 1988. Biogeochemical aspects of atmospheric methane. Global Biogeochemical Cycles, 2: 299-327.

Clarke, A.D. and K.J. Noone, 1985. Soot in the Arctic snowpack: a cause for perturbation in radiative transfer. *Atmos. Environ.* **19**: 2045-2053.

Coachman, L.K. and K. Aagaard, 1974. Physical oceanography of Arctic and subarctic seas. pp 1-72, *In*: Y. Herman (ed.). Marine Geology and Oceanography of the Arctic Seas. Springer Verlag, New York.

Cohen, S.J., 1996. Integrated assessment of global climatic change: Lessons from the Mackenzie Basin Impact Study (MBIS). Global and Planetary Change, 11: 179-185.

Cohen, S.J. (ed.), 1997. Mackenzie Basin Impact Study (MBIS): Final Report. Environment Canada, Downsview, Canada.

Colony, R. and A.S. Thorndike, 1984. An estimate of the mean field of Arctic sea ice motion. *J. of Geophys. Res.* **89(C6)**: 10623-10629.

Comiso, J.C., P. Wadhams, W.B. Krabill, R.N. Swift, J.P. Crawford and W.B. Tucker III., 1991. Top/bottom multisensor remote sensing of Arctic sea ice. *J. Geophys. Res.* **96**: 2693-2709.

Conrad, R. and W. Seiler, 1980. Photooxidative production and microbial consumption of carbon monoxide in seawater. *FEMS Microbiology Letters* **9**: 61-64.

Couteaux M.M, M. Mousseau, M.I. Celerier and P. Bottner, 1991. Increased atmospheric CO_2 and litter quality: decomposition of sweet chestnut leaf litter with animal food webs of different complexities. *Oikos* **61**: 54-64.

Crawford, R.M.M., H.M. Chapman and L.C. Smith, 1995. Adaptations to variation in growing season length in Arctic populations of Saxifraga oppositifolia L. *Bot. J. Scotl.* **47**: 177-192.

Cuffey, K.M. and G.D. Clow, in press. Temperature, accumulation and ice sheet elevation in central Greenland through the last deglaciation. *J. Geophys. Res.*, in press.

Cuffey, K.M., G.D. Clow, R.B. Alley, M. Stuiver, E.D. Waddington and R.W. Saltus, 1995. Large Arctic temperature change at the Wisconsin-Holocene glacial transition. *Science* **270**: 455-458.

Cullen J.J., P.J. Neale and M.P. Lesser, 1992. Biological weighting function for the inhibition of phytoplankton photosynthesis by ultraviolet radiation. *Science* **258**: 646-650.

Cullen, J.J., P.J. Neale, 1994. Ultraviolet radiation, ozone depletion and marine photosynthesis. *Photosynth. Res.* **39**: 303-320.

Curtis, P.J. and H.E. Adams, 1995. Dissolved organic matter quantity and quality from freshwater and saltwater lakes in east-central Alberta. *Biogeochemistry* **30**: 59-76

Dahlback, A. and J. Moan, 1990. Annual exposures to carcinogenic radiation from the Sun at different latitudes and amplification factors related to ozone depletion. The use of different geometrical representations of the skin surface receiving the ultraviolet radiation. *Photochem. and Photobiol.* **52**: 1025-1028.

Dahlström, B., 1994. Short term fluctuations of temperature and precipitation in Western Europe. Climate Variations in Europe. Proceedings of the European Workshop on Climate Variations held in Kirkkonummi, Finland, 15-18 May 1994, Publ. of the Academy of Finland, 30-38.

Damkaer, D.M. and D.B. Dey, G. A. Heron, 1981. Dose/dose rate response of shrimp larvae to ultraviolet-B radiation. *Oecologia* **60**: 169-175.

Damkaer, D.M. and D.B. Dey, 1983. UV damage and photoreactivation potentials of larval shrimp, *Pandalus platyceros*, and adult euphasiids, *Thysanoessa rashii. Oecologia* **60**: 169-175.

Damkaer, D.M., D.B. Dey, G.A. Heron and E.F. Prentice, 1980. Effects of ultraviolet-B radiation on near surface zooplankton of Puget Sound. *Oecologia* **44**: 149-158.

Daniel J.S., S. Solomon and D.L. Albritton, 1995. On the evaluation of halocarbon radiative forcing and global warming potentials. *J. Geophys. Res.* **100(D1)**: 1271-1285.

Dansgaard, W., J.W.C. White and S.J. Johnsen, 1989. The abrupt termination of the Younger Dryas climate event. *Nature* **339**: 532-534.

Dansgaard, W., S.J. Johnsen and H.B. Clausen, 1993. Evidence for general instability of past climate from a 250-kyr ice-core record. *Nature* **364(6434)**: 218-220.

de Gruijl, F.R. and J.C. van der Leun, 1993. Influence of ozone depletion on the incidence of skin cancer - Quantitative prediction. *In*: A.R. Young, L.O. Björn, J.M. Moan and W. Nultsch (eds.). Environmental UV photobiology, Plenum Press, New York and London.

Dey, B., 1980. Variation of August ice cover in the Beaufort Sea and related weather conditions. *Bull. Amer. Meteor. Soc.* **61**: 213-217.

di Sarra, A., M. Cacciani, G. Fiocco, D. Fu, T.S. Jorgensen, B. Knudsen, N. Larsen and I.S. Mikkelsen, 1995. Ozone and aerosol correlated observations at Thule, Greenland, in the period 1991-1994. *J. Geophys. Res.* **100**: 25965-25977.

Dickson, R.R., J. Meinke, S.A. Malmberg and A.J. Lee, 1988. The "great salinity anomaly" in the northern North Atlantic. *Prog. Oceanogr.* **20**: 103-151.

Dobson, G.M.B., 1968a. Forty years research on atmospheric ozone at Oxford: a History. *Applied Optics* **7**: 387-405.

Dobson, G.M.B., 1968b. Exploring the Atmosphere. Clarendon Press, Oxford.

Döhler, G., 1992. Impact of ultraviolet-B radiation on uptake of ^{15}N-ammonia and ^{15}N-nitrate by phytoplankton of the Wadden Sea. *Mar. Biol.* **112**: 485-489.

Döhler, G., G.E. Hagmeier, K.-D. Krause, 1991. Impact of solar UV radiation on uptake of ^{15}N-ammonia and ^{15}N-nitrate by marine diatoms and natural phytoplankton. *Biochem. Physiol. Pflanzen* **187**: 293-303.

Dolin, P.J., 1995. Assessment of the epidemiological evidence that exposure to solar ultraviolet radiation causes cataract. *Documenta Ophthalmologica* **88**: 327-337.

Donoso, L., R. Santana and E. Sanhueza, Seasonal variation of N_2O fluxes at a tropical savannah site: soil consumption of N_2O during the dry season, 1993. *Geophys. Res. Lett.* **20**: 1379-1382.

Douglas, M.S.V. and J.P. Smol, 1994. Limnology of High Arctic ponds (Cape Herschel, Ellesmere Island, N.W.T.). *Arch. Hydrobiologie* **131**: 401-434.

Douglas, M.S.V. and J.P. Smol, 1995. Periphytic diatom assemblages from High Arctic ponds. *J. of Phycology* **31**: 60-69.

Douglas, M.S.V. and J.P. Smol, 1993. Freshwater diatoms from High Arctic ponds (Cape Herschel, Ellesmere Island, N.W.T.). *Nova Hedwigia* **57**: 511-552.

Dunlap, W.C., B.E. Chalker and J.K. Olivier, 1986. Barthymetric adaptations of reef-building corals at Davies Reef, Great Barrier Reef, Australia. III. ultraviolet-B absorbing compounds. *J. Exp. Mar. Biol. Ecol.* **104**: 239-248.

Eaton, J.G. and R.M. Scheller, 1996. Effects of climate warming on fish thermal habitat in streams of the United States. *Limnol Oceanog.* **41**: 1109-1115.

Ebert, E.E. and J.A. Curry, 1993. An intermediate one dimensional thermodynamic sea ice model for investigating ice atmosphere interactions. *J. Geophys. Res.* **98**: 10085-10109.

Edlund S., 1989. Vegetation indicates potentially unstable Arctic terrain. *Geos.* **18**: 9-13.

Ekelund, N.G.A., 1990. Effects of ultraviolet-B radiation on growth and motility of four phytoplankton species. *Physiol. Plant.* **78**: 590-594.

El-Sayed, S.Z., F.C. Stephens, R.R. Bidigare and M.E. Ondrusek, 1990. Effect of ultraviolet radiation on Antarctic marine phytoplankton. pp. 379-385, *In*: K.R. Kerry, G. Hempel (eds.). Antarctic Ecosystems. Springer-Verlag, Berlin Heidelberg.

Elias, S.C., S.K. Short, C.H. Nelson and H.H. Biris, 1996. Life and times of the Bering landbridge. *Nature* **382**: 60-61.

Elliott, W.P., M.E. Smith and J.K. Angell, 1991. On monitoring tropospheric water vapor changes using radiosonde data. pp. 311-328, *In*: M. Schlesinger (ed.). Greenhouse-gas induced climate change: A critical appraisal of simulations and observations. Elsevier, Amsterdam, 615p.

Emery, W., T. Kelley, J. Dozier and P. Rotar, 1995. On-line access to weather satellite imagery and image manipulation software. *Bull. Amer. Meteor. Soc.* **76(6)**: 923-932.

Evens, T.J., S. Roll, B. Golden and B.B. Prézelin, 1994. Icecolors '93: Epilithic productivity by microalgae exhibits a potentially high sensitivity to natural levels of ultraviolet-B radiation. *Antarctic Journ. U.S.* **29**: 275-277.

Fahey, D.W., K.K. Kelly, S.R. Kawa, A.F. Tuck, M. Loewenstein, K.R. Chan and L.E. Heidt, 1990a. Observations of denitrification and dehydration in the winter polar stratospheres. *Nature* **344**: 321-324.

Fahey, D.W., S. Solomon, S.R. Kawa, M. Loewenstein, J.R. Podolske, S.E. Strahan and K.R. Chan, 1990b. A diagnostic for denitrification in the winter polar stratosphere. *Nature* **345**: 698-702.

Fang, Z. and J.M. Wallace, 1994. Arctic sea ice variability on a timescale of weeks and its relation to atmospheric forcing. *J. Climate* **7**: 1897-1914.

Feister, U. and R. Grewe, 1995. Spectral albedo measurements in the UV and visible region over different types of surfaces. *Photochem. and Photobiol.* **62(4)**: 736-744.

Ferraro, R., N. Grody, D. Forsyth, R. Carey, A. Basist, J. Janowiak, F. Weng, G.F. Marks and R. Yanamandra, 1994. Microwave measurements produce global climatic hydrologic data. *EOS (American Geophysical Union)* **75**: 337-338.

Finger, F.G., R.M. Nagatani, M.E. Gelman, C.S. Long and A.J. Miller, 1995. Consistency between variations of ozone and temperature in the stratosphere. *Geophys. Res. Lett.* **22**: 3477-3480.

Finlayson-Pitts, B.J., F.E. Livingston and H.N. Berko, 1990. Ozone destruction and bromine photochemistry at ground level in the Arctic spring. *Nature* **343**: 622-625.

Fitzharris, B.B., 1996. The cryosphere: Changes and their impacts. pp. 243-265, *In*: T.R. Watson *et al.* (eds.). Climate Change 1995: Impacts, adaptations and mitigation of climate change: scientific technical analysis. Cambridge University Press.

Flanagan, P.W. and F.L. Bunnell, 1980. Microflora activities and decomposition. pp 291-334, *In*: J. Brown, P.C. Miller, L.L. Tieszen and F.L. Bunnell (eds.). An Arctic Ecosystem. The Coastal Tundra at Barrow, Alaska. Dowden, Hutchinson and Ross, Stroudsburg.

Flannigan, M.D.F. and C.E. Van Wagner, 1991. Climate Change and wildfire in Canada. *Can. J. of Forest Res.* **21**: 66-72.

Foster, J., 1989. The significance of the date of snow disappearance on the Arctic tundra as a possible indicator of climate change. *Arctic and Alpine Res.* **21**: 60-70.

Fraser P.G., S.A. Penkett, M. Gunson, R. Weiss and F.S. Rowland, 1994. Measurements. Chapter 1, *In*: J. Kaye, S.A. Penkett and F. Ormond (eds.). NASA Report on Concentrations, Lifetimes and Trends of CFCs, Halons and Related Species. NASA Reference Publication 1339, Washington D.C.

Frederick, J.E. and D. Lubin, 1988. The budget of biologically active ultraviolet radiation in the Earth-Atmosphere system. *J. Geophys. Res.* **93**: 3825-3832.

Frederick, J.E. and H.E. Snell, 1990. Tropospheric influence on solar ultraviolet radiation: The role of clouds. *J. Clim.* **3**: 373-381.
Frederick, J.E., H.E. Snell and E.K. Haywood, 1989. Solar ultraviolet radiation at the earth's surface. *Photochem. and Photobiol.* **50**: 443-450.
Freeman, M., 1984. Arctic ecosystems. *In*: D. Dumas (ed.). Vol. 5: Arctic. *In*: W. Sturtevant (ed.). Handbook of North American Indians. Smithsonian Institution, Washington, D.C.
French, H.M., L. Bennet and D.W. Hayley, 1986. Ground ice conditions at Rae Point and Sabine Peninsula, Eastern Melville Is. *Can. J. of Earth Sciences* **22**: 1389-1400.
Frimmel, F.H., 1994. Photochemical aspects related to humic substances. *Environ. Internat.* **20**: 373-385.
Froidevaux, L., J.W. Waters, W.G. Read, L.S. Elson, D.A. Flower and R.F. Jarnot, 1994. Global ozone observations from the UARS MLS: An overview of zonal-mean results. *J. Atmos. Sci.* **51**: 2846-2866.
Gaffen, D.J., 1994. Temporal inhomogeneities in radiosonde temperature records. *J of Geophys. Res.* **99**: 3667-3676.
Gajewski, K., M. Garneau and J.C. Bourgeois, 1995. Paleoenvironments of the Canadian High Arctic derived from pollen and plant macrofossils: problems and potentials. *Quaternary Science Reviews* **14**: 609-629.
Gardiner, B.G. and P.J. Kirsch, 1995. Setting standards for European ultraviolet spectroradiometers. EUR 16153. Air pollution research report 53. European Commission, Luxembourg.
Gawarkiewicz, G. and D.C. Chapman, 1995. A numerical study of dense water formation and transport on a shallow, sloping continental shelf. *J. Geophys. Res.* **100**: 4489-4507.
Gehrke, C., U. Johanson, D. Gwynn-Jones, L.O. Björn, T.V. Callaghan and J.A. Lee, 1996. Effects of enhanced ultraviolet-B radiation on terrestrial subarctic ecosystems and implications for interactions with increased atmospheric CO_2. *Ecol. Bull.* **45**: 192-203.
Gehrke, C., U. Johanson, T.V. Callaghan, D. Chadwick and C.H. Robinson, 1995. The impact of enhanced ultraviolet-B radiation on litter quality and decomposition processes in Vaccinium leaves from the subarctic. *Oikos* **72**: 213-222.
Gloersen, P. and W.J. Campbell, 1991. Recent variations in Arctic and Antarctic sea ice covers. *Nature* **352**: 33-36.
Gloersen, P., W.J. Campbell, D.J. Cavalieri, J.C. Comiso, C.L. Parkinson and H.J. Zwally, 1992. Arctic and Antarctic Sea Ice, 1978-1987: Satellite Passive Microwave Observations and Analysis. NASA SP511. National Aeronautics and Space Administration, Washington, D.C.
Goes, J.I., N. Handa, S. Taguchi and T. Hama, 1994. Effect of ultraviolet-B radiation on the fatty acid composition of the marine phytoplankter Tetraselmis sp.: Relationship to cellular pigments. *Mar. Ecol. Prog. Ser.* **114**: 259-274.
Goes, J.I., N. Handa, S. Taguchi and T. Hama, 1995. Changes in the patterns of biosynthesis and composition of amino acids in a marine phytoplankter exposed to ultraviolet-B radiation: nitrogen limitation implicated. *Photochem. and Photobiol.* **62**: 703-710.
Goldman, J.C., M.R. Dennett, 1991. Ammonium regeneration and carbon utilization by marine bacteria grown on mixed substrates. *Mar. Biol.* **109**: 369-378.
Gorshkov, S.G. (ed.), 1983. World Ocean Atlas, Vol. 3, Arctic Ocean. Pergamon Press, Oxford. 184p.
Goryarchin, S.V., R.I. Zlotin and G.M. Tertitsky (eds.), 1994. Russian-Swedish expedition Tundra Ecology-94: Diversity of natural ecosystems in the Russian Arctic, a guidebook. University of Lund, Sweden, 106p.
Gregory, J.M., 1993. Sea level changes under increasing atmospheric CO_2 in a transient coupled ocean-atmosphere GCM experiment. *J. Climate* **6**: 2247-2262.
Grenfell, T.C. and G.A. Maykut, 1977. The optical properties of ice and snow in the Arctic Basin. *J. of Glaciology* **18**: 445-463.
Grenfell, T.C. and D.K. Perovich, 1984. Spectral albedos of sea ice and incident solar irradiance in the southern Beaufort Sea. *J. of Geophys. Res.* **89**: 7447-7450.
Grenfell, T.C. and S.G. Warren, 1994. Reflection of solar radiation by the Antarctic snow surface at ultraviolet, visible and near-infrared wavelengths. *J. Geophys. Res.* **99**: 18669-18684.
Grobe, C.W. and T.M. Murphy, 1994. Inhibition of growth of *Ulva expansa* (Chlorophyta) by ultraviolet-B radiation. *J. Phycol.* **30**: 783-790.
Grody, N.C. and A.N. Basist, 1996. Global identification of snowcover using SSM/I measurements. IEEE Transactions on Geoscience and Remote Sensing, 34: 237-249.
Groisman P.Ya., 1991. Data on present-day precipitation changes in the extratropical part of the Northern hemisphere. pp. 297-310, *In*: M.E. Schlesinger (ed.). Greenhouse-Gas- Induced Climatic Change: A Critical Appraisal of Simulations and Observations. Elsevier, Amsterdam. 615p.
Groisman, P.Y. and D.R. Easterling, 1994. Variability and trends of precipitation and snowfall over the United States and Canada. *J. Climate* **7**: 184-205.
Groisman, P.Y. and T. R. Karl and R.W. Knight, 1994a. Observed impact of snow cover on the rise of continental spring temperatures. *Science* **263**: 198-200.
Groisman, P.Y. and T.R. Karl, R.W. Knight and G. Stenchikov, 1994b. Changes of snow cover, temperature and the radiative heat balance over the Northern Hemisphere. *J. Climate* **7**: 1633-1656.
Grootes, P.M., M. Stuiver, J.W.C. White, S. Johnsen and J. Jouzel, 1993. Comparison of oxygen isotope records from the GISP2 and GRIP Greenland ice cores. *Nature* **366**: 552-554.
Gschwend, P.M., J.K. MacFarlane and K.A. Newman, 1985. Volatile halogenated organic compounds released to seawater from temperate marine macroalgae. *Science* **227**: 1033-1035.
Häder, D.-P., 1993. Effects of enhanced solar ultraviolet radiation on aquatic ecosystems. pp. 152-192, *In*: M. Tevini (ed.). Ultraviolet-B Radiation and Ozone Depletion. Effects on Humans, Animals, Plants, Microorganisms and Materials. Lewis, Boca Raton, Florida.
Häder, D.-P., R.C. Worrest, H.D. Kumar, R.C. Smith, 1995. Effects of increased solar ultraviolet radiation on aquatic ecosystems. *Ambio* **24**: 174-180.
Hairston Jr., N.G., 1976. Photoprotection by carotenoid pigments in the copepod *Diatomus nevadensis*. Proc. Natl. Acad. Sci. USA 73: 971-974.
Hairston Jr., N.G., 1980. The vertical distribution of diaptomid copepods in relation to body pigmentation. pp. 98-110, *In*: W.C. Keerfoot (ed.). Ecology and Evolution of Zooplankton Communities. University of New England Press, Hanover, New Hampshire.
Hakkinen, S., 1990. Models and their applications to polar oceanography. Polar Oceanography, Part A: Physical Science. Academic Press. pp. 335-385.
Hakkinen, S., 1993. An Arctic source for the Great Salinity Anomaly: A simulation of the Arctic ice/ocean system for 1955-1975. *J. Geophys. Res.* **98**: 16397-16410.
Hall, D.K., 1988. Assessment of polar climate change using satellite technology. Reviews of Geophysics, 26(1): 26-39.
Hall, N.M.J., B.J. Hoskins, P.J. Valdes and C.A. Senior, 1994. Storm tracks in a high-resolution GCM with doubled carbon dioxide. *Quart. J. Roy. Met. Soc.* **120**: 1209-1230.
Hallet, B., 1990. Self-organization in freezing soils: from microscopic ice lenses to patterned ground. *Can. J. of Phys.* **26**: 842-52.
Hallet, B. and E. Waddington, 1992. Buoyancy forces induced by freeze-thaw in the active layer: implications for diapirism and soil circulation. pp. 251-279, *In*: J.C. Dixon and A.D. Abrahams (eds.). Periglacial geomorphology. J. Wiley and Sons Ltd.
Hammer, U.T., 1986. Saline lake ecosystems of the World. Dr. W. Junk Publishers, Boston. 616p.
Hanssen-Bauer, I. and E.J. Førland, 1994. Homogenizing long Norwegian precipitation series. *J. Climate* **6**(**7**): 1001-1011.
Harrison, W.G. and G. F. Cota, 1991. Primary production in polar waters: relation to nutrient availability. *Polar Res.* **10**: 87-104.
Harry, D.G. and K. MacInnes, 1988. The effect of forest fires on permafrost terrain stability, Little Chicago-Travaillant Lake area, Mackenzie Valley, NWT. pp. 91-94, *In*: Current Research Part D, Geological Survey of Canada Paper 88(1D).
Harry, D.G., H.M. French and W.H. Pollard, 1988. Massive ground ice and ice-cored terrain near Sabine Point, Yukon Coastal Plain. *Can. J. of Earth Sciences* **25**: 1846-1856.
Hatcher, P.E. and N.D. Paul, 1994. The effects of elevated UV-B radiation on herbivory of pea by Autographagamma. *Entomol. Exp. Appl.* **71**: 227-233.
Haugen, R.K., S.I. Outcalt and J.C. Harle, 1982. Relationships between estimated mean annual air and permafrost temperatures in northern central Alaska. pp. 462-467, *In*: Permafrost-Fourth International Conference Proceedings, vol. 1, Washington, D.C., National Academy of Science.
Havas, M. and T.C. Hutchinson, 1983. Effect of low pH on the chemical composition of aquatic invertebrates from tundra ponds at the Smoking Hills, N.W.T., Canada, *Can. J. Zool.* **61**: 241-249.
Havström, M., T.V. Callaghan, J. Svoboda and S. Jonasson, 1995. Little ice age temperature estimated by growth and flowering differences between subfossil and extant shoots of *Cassiope tetragona*, an Arctic heather. *Functional Ecology* **9**: 650-654.
Hay, J. and J.C. Davies, 1978. Calculation of the solar radiation incident on an inclined surface. *In*: J.E. Hat and T.K. Won (eds.). Proc. First Canadian Solar Radiation Data Workshop, Toronto, Ontario, Canada.
Hebert, P.D.N. and C.J. Emery, 1990. The adaptive significance of cuticular pigmentation in Daphnia. *Functional Ecology* **4**: 703-710.
Heginbottom, J.A., E.S. Brown, and O. J. Ferrians Jr., 1993. Circumarctic map of permafrost and ground ice conditions. pp. 1132-1136, *In*: Sixth International Conference on Permafrost (Beijing, China, 5-9 July 1993), Proceedings, V. 2. South China University of Technology Press, Wushan, China.
Helbling, E.W., E.R. Marguet, V.E. Villafale and O. Holm-Hansen, 1995. Bacterioplankton viability in Antarctic waters as affected by solar ultraviolet radiation. *Mar. Ecol. Prog. Ser.* **126**: 293-298.
Helbling, E.W., H.C. Eilersen, V.E. Villafale and O. Holm-Hansen, 1996. Effects of UV radiation on post-bloom phytoplankton populations in Kvalsund, North Norway. *J. Photochem. Photobiol. B: Biology* **99**(**33**): 255-259.
Helbling, E.W., V. Villafale, M. Ferrario and O. Holm-Hansen, 1992. Impact of natural ultraviolet radiation on rates of photosynthesis and on specific marine phytoplankton species. *Mar. Ecol. Prog. Ser.* **80**: 89-100.
Henriksen, K., K. Stamnes and P. Ostensen, 1989. Measurements of solar UV, visible and near IR irradiance at 78 N. *Atmos. Environ.* **23**: 1573-1579.
Herbert, P.D.N., C.J. Emery, 1990. The adaptive significance of cuticular pigmentation in Daphnia. *Functional Ecology.* **4**: 703-710.
Herman, G. and R. Goody, 1976. Formation and persistence of Arctic stratus clouds. *J. Atmos. Sci.* **33**: 1537-1553.
Herman, J.R. and D. Larko, 1994. Low ozone amounts during 1992-1993

from Nimbus 7 and Meteor 3 total ozone mapping spectrometers. *J. Geophys. Res.* **99**: 3483-3496.
Herndl, G.J., G. Müller-Niklas, J. Frick, 1993. Major role of ultraviolet-B in controlling bacterioplankton growth in the surface layer of the ocean. *Nature* **361**: 717-718.
Hessen, D.O., 1993. DNA-damage and pigmentation in alpine and Arctic zooplankton as bioindicators of UV-radiation. *Verh. Internat. Verein. Limnol.* **25**: 482-486.
Hessen, D.O., 1994. Daphnia responses to UV-light. *Arch. Hydrobiol. Beih. Ergebn. Limnol.* **43**: 185-195.
Hessen, D.O., 1996. Competitive trade-off strategies in Arctic Daphnia linked to melanism and ultraviolet-B stress. *Polar Biology* **16**: 573-579.
Hessen, D.O., T. Andersen and A. Lyche, 1990. Carbon metabolism in a humic lake: pool sizes and cycling through zooplankton. *Limnol. Oceanogr.* **35**: 84-99.
Hessen, D.O., E. Van Donk and T. Andersen, 1995. Growth responses, P-uptake and loss of flagellae in *Chlamydomonas reinhardtii* exposed to ultraviolet-B. *J. Plankton Res.* **17**: 17-27.
Hisdal, V., 1986. Spectral distribution of global and diffuse solar radiation in Ny-Alesund, Spitsbergen. *Polar Res.* **5**: 1-27.
Hobbie, J.E. (ed.), 1980. Limnology of tundra ponds. Dowden, Hutchinson and Ross, Inc., Stroudsburg, Pa.. 514p.
Højerslev, N.K. and E. Aas, 1991. A relationship for the penetration of ultraviolet-B radiation in the Norwegian Sea. *J. Geophys. Res.* **96**: 17003-17005.
Holm-Hansen, O., D. Lubin and E.W. Helbling, 1993a. UV radiation and its effects on organisms in aquatic environments. pp. 379-425, *In*: A.R. Young, L.O. Björn, J. Moan and W. Nultsch (eds.). Environmental UV Photobiology. Plenum Press, New York.
Holm-Hansen, O., E.W. Helbing and D. Lubin, 1993b. Ultraviolet radiation in Antarctica: inhibition of primary production. *Photochem. and Photobiol.* **58**: 567-570.
Hood, L.L. and J.P. McCormack, 1992. Components of interannual ozone change based on Nimbus 7 TOMS data. *Geophys. Res. Lett.* **19**: 2309-2312.
Hood, L.L. and Da.A. Zaff, 1995. Lower stratospheric stationary waves and the longitude dependence of ozone trends in winter. *J. Geophys. Res.* **100**: 25791-25800.
Horner R.A., 1985. Ecology of sea ice microalgae. pp. 83-103, *In*: R.A. Horne (ed.). Sea Ice Biota. CRC Press, Boca Raton, Florida.
Hughen, K.A., J.T. Overpeck, R.F. Anderson and K.M. Williams, 1996. The potential for paleoclimate records from varved Arctic lake sediments: Baffin Island, Eastern Canadian Arctic. pp. 57-71, *In*: A.E.S. Kemp (ed.). Palaeoclimatology and Palaeoceanography from Laminated Sediments. Geological Society Special Publication, London. 116p.
Hughes, M. and D. Robinson, 1996. Historical snow cover variability in the Great Plains region of the United States: 1910 through 1993. *Int. J. Climatol.* **16**(**9**): 1005-1032.
Hunter, J.R., J.H. Taylor, H.G. Moser, 1979. Effect of ultraviolet irradiation on eggs and larvae of the northern anchovy, Engraulis mordax and the Pacific mackerel, *Scomber japonicus*, during the embryonic stage. *Photochem. and Photobiol.* **29**: 325-338.
Huschke, R.E., 1969. Arctic cloud statistics from air calibrated surface weather observations. Mem. RM-6173-PR. Rand Corp., Santa Monica, California. 79p.
IASC (International Arctic Science Committee), 1995. Effects of increased ultraviolet radiation in the Arctic. IASC Report no. 2. 56p.
IASC (International Arctic Science Committee), 1996. Ultraviolet international research centers (UVIRCs): A proposal for interdisciplinary UVB research in the Arctic. IASC Report no. 7. 36p.
Ingram, M.J., G. Farmer and T.M.L. Wigley, 1981. Past climates and their impact on man: A review. pp. 4-19, *In*: T.M.L. Wigley, M.J. Ingram and G. Farmer (eds.). Climate and History: Studies of past climates and their impact on Man. Cambridge University Press, Cambridge, U.K.
IPCC, 1990a. Climate Change: The IPCC Scientific Assessment. J.T. Houghton, G.J. Jenkins and J.J. Ephraums (eds.). Cambridge University Press, Cambridge, U.K.
IPCC, 1990b. Climate Change: The IPCC Impacts Assessment, Cambridge University Press, Cambridge, U.K.
IPCC, 1992a: Climate Change 1992: The Supplementary Report to the IPCC Scientific Assessment. J.T. Houghton, B.A. Callander and S.K. Varney (eds.). Cambridge University Press, Cambridge, U.K. 200p.
IPCC, 1992b. Climate Change 1992: The Supplementary Report to the IPCC Impacts Assessment. Cambridge University Press, Cambridge, U.K.
IPCC, 1994. Climate Change 1994: Radiative forcing of Climate Change and an evaluation of the IPCC IS92 emission scenarios. Cambridge University Press, Cambridge, U.K.
IPCC, 1996a. Climate Change 1995: Second Assessment Report on Climate Change. Cambridge University Press, Cambridge, U.K. 572p.
IPCC, 1996b. Climate Change 1995: Impacts, Adaptations and Mitigation of Climate Change: Scientific-Technical Analyses. Cambridge University Press, Cambridge, U.K. 878p.
Isodorov V.A., I.G. Zenkevich and B.V. Ioffe, 1990. Volatile organic compounds in solfataric gases. *J. Atmos. Chem.* **10**: 329-340.
Ivanov, V.V., 1990. The Arctic water resources and their studies: The state-of-the-art and future plans. Arctic Research, Advances and Prospects, Proceedings of the Conference of Arctic and Nordic Countries on Coordination of Research in the Arctic. Nauka, Leningrad, Moscow.
Ivanov, V.V. and V.A. Yankina, 1991. Water resources of the Arctic. *Problemy Arkt. Antarkt.* **66**: 118-128. (In Russian).
Jacoby, G.C. and R.D. D'Arrigo, 1995. Tree ring width and density evidence of climatic and potential forest change in Alaska. *Global Biogeochemical Cycles* **9**: 227-234.
Jaffe, D.A., 1991. Local sources of ozone in the Arctic: From Prudhoe Bay to the Taz Peninsula. pp. 255-287, *In*: W.T. Sturges (ed.). Pollution of the Arctic Atmosphere. Elsevier Applied Science, New York.
Jaffe, D., T. Iverson and G. Shaw, 1995. Comment on a long-term decrease in Arctic haze at Barrow, Alaska by B.A. Bodhaine and E.G. Dutton. *Geophys. Res. Lett.* **22**(**6**): 739-740.
Jin, Z., K. Stamnes and W.F. Weeks, 1994a. Transport of photosynthetically active radiation in sea ice and ocean. pp. 954-964, *In*: J. Jaffe (ed.). Ocean Optics XII, Proc. SPIE 2258.
Jin, Z., K. Stamnes, W.F. Weeks and S.C. Tsay, 1994b. The effect of sea ice on the solar energy budget in the atmosphere-sea ice-ocean system: A model study. *J. Geophys. Res.* **99**: 25281-25294.
Johannessen, O.M., M. Miles and E. Bjorgo, 1995. Arctic's shrinking sea ice. *Nature* **376**: 126-127.
Johanson, U., C. Gehrke, L.O. Björn and T.V. Callaghan, 1995b. The effects of enhanced ultraviolet-B radiation on the growth of dwarf shrubs in a subarctic heathland. *Functional Ecology* **9**: 713-719.
Johanson, U., C. Gehrke, L.O. Björn, T.V. Callaghan and M. Sonesson, 1995a. The effects of enhanced ultraviolet-B radiation on a subarctic heath ecosystem. *Ambio* **24**(**2**): 106-111.
Johnsen, B., 1996. UV monitoring in Norway: past, present and future. *In*: B.L. Diffey (ed.). Measurement and trends of terrestrial ultraviolet-B radiation in Europe. OEMF, Milan.
Jokela, K., K. Leszczynski and R. Visuri, 1993. Effects of Arctic ozone depletion and of snow on UV exposure in Finland. *Photochem. and Photobiol.* **58**: 559-566.
Jokela, K., K. Leszczynski, R. Visuri and L. Ylianttila, 1995. Increased UV exposure in Finland in 1993. *Photochem. and Photobiol.* **62**: 101-107.
Jonasson, S., M. Havstrom, M. Jensen and T.V. Callaghan, 1993. In situ mineralization of nitrogen and phosphorus of Arctic soils after perturbations simulating climate change. *Oecologia* **95**: 179-186.
Jones, P.D., 1994. Hemispheric surface air temperature variations: A reanalysis and update to 1993. *J. Climate* **7**: 1794-1802.
Jones, P.D. and K.R. Briffa, 1992. Global surface air temperature variations during the twentieth century: Part 1. Spatial, temporal and seasonal detail. *The Holocene* **2**: 165-179.
Jones, P.D., S.C.B. Raper, R.S. Bradley, H.F. Diaz, P.M. Kelly, T.M.L. Wigley, 1986. Northern Hemisphere surface air temperature variations, 1851-1984. *J. Clim. Appl. Meteor.* **25**: 161-179.
Kahl, J.D.W., D.A. Martinez and N.A. Zaitseva, 1996. Long-term variability in the low-level inversion layer over the Arctic Ocean. *Int. J. Climatol.* **16**: 1297-1313.
Kahl, J.D., D.J. Charlevoix, N.A. Zaitseva, R.C. Schnell and M.C. Serreze, 1993a. Absence of evidence for greenhouse warming over the Arctic Ocean in the past 40 Years. *Nature* **361**: 335-337.
Kahl, J.D., M.C. Serreze, S. Shiotani, S.M. Skony and R.C. Schnell, 1992. In-situ meteorological sounding archives for Arctic studies. *Bull. Amer. Meteor. Soc.* **73**: 1824-1830.
Kahl, J.D., M.C. Serreze, R.S. Stone, S. Shiotani, M. Kisley and R.C. Schell, 1993b. Tropospheric temperature trends in the Arctic: 1958-86. *J. Geophys. Res.* **98**: 12825-12838.
Kalff, J. and H.E. Welch, 1974. Phytoplankton production in Char Lake, a natural polar lake in Meretta Lake, a polluted polar lake, Cornwallis Island, Northwest Territories. *J. Fish. Res. Board Canada* **31**: 621-636.
Kane, D.L. and L.D. Hinzman, 1988. Permafrost hydrology of a small Arctic watershed. pp. 590-595, *In*: K. Sennest (ed.). Permafrost, Fifth International Conference, Proceedings, Trondheim, Norway. Tapir Publishers, Trondheim.
Kane, D.L., L.D. Hinzman, M.K. Woo, K.R. Everett, F.S. Chapin III, R.L. Jefferies, J.F. Reynolds, G.R. Shaver, J. Svoboda, E. . Chu (eds.), 1992. Arctic ecosystems in a changing climate: an ecophysiological perspective. Academic Press, San Diego.
Kankaala, P., L. Ojala, L. Arvola and T. Tuolonen, 1994. Impact of climate change on the carbon cycle in freshwater ecosystems. pp. 140-145, *In*: M. Kanninen and P. Heikinheimo (eds.). The Finnish Research Programme on Climate Change, Second Progress Report. Publications of the Academy of Finland 1/94. Helsinki.
Karanas, J.J., H. van Dyke, R.C. Worrest, 1979. Mid-ultraviolet (ultraviolet-B) sensitivity of *Acartia claussii* Giesbrecht (Copepoda). *Limnol. Oceanogr.* **24**: 1104-1116.
Karentz, D., 1994. Ultraviolet tolerance mechanisms in Antarctic marine organisms. pp. 93-110, *In*: C.S. Weiler and P.A. Penhale (eds.). Ultraviolet radiation in Antarctica: measurements and biological effects. Antarctic Research Series 62.
Karentz, D., J.E. Cleaver and D.L. Mitchell, 1991. Cell survival characteristics and molecular responses of antarctic phytoplankton to ultraviolet-B radiation. *J. Phycol.* **27**: 326-341.
Karentz, D., M.L. Bothwell, R.B. Coffin, A. Hanson, G.J. Herndl, S.S. Kilham, M.P. Lesser, M. Lindell, R.E. Moeller, D.P. Morris, P.J. Neale, R.W. Sanders, C.S. Weiler and R.G. Wetzel, 1994. Impact of ultraviolet-B radiation on pelagic freshwater ecosystems: Report of working group on bacteria and phytoplankton. pp. 31-69, *In*: C.E. Williamson and Zagarese (eds.). Advances in Limnology: Impact of ultraviolet-B radiation on pelagic ecosystems. *Arch. Hydrobiologie* **43**.
Karl, T.R., P.Y. Groisman, R.R. Heim Jr. and R.W. Knight, 1993. Recent variations of snow cover and snowfall in North America and their rela-

tion to precipitation and temperature variation. *J. Climate* **6**: 1327-1344.
Kasischke, E.S., N.L. Christiensen and B.J. Stocks, 1995. Fire, global warming and the carbon balance of boreal forests. *Ecological Applications* **5**: 437-451.
Kawa, S.R., D.W. Fahey and L.E. Heidt., 1992a. Photochemical partitioning of the reactive nitrogen and chlorine reservoirs in the high-latitude stratosphere. *J. Geophys. Res.* **97**: 7905-7923.
Kawa, S.R., D.W. Fahey, K.K. Kelly, J.E. Dye, D. Baumgardner, B.W. Gandrud, M. Loewenstein, G.V. Ferry and K.R. Chan, 1992b. The Arctic polar stratospheric cloud aerosol: aircraft measurements of reactive nitrogen, total water and particles. *J. Geophys. Res.* **97**: 7925-7938.
Keeling, C.D., 1994. Global historical CO_2 emissions. pp. 501-504, *In*: T.A. Boden, D.P. Kaiser, R.J. Sepanski and F.W. Stoss (eds.). Trends '93: A Compendium of Data on Global Change, ORNL/CDIAC-65. Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, Oak Ridge, Tennessee.
Kellogg, W.W., 1983. Feedback mechanisms in the climate system affecting future levels of carbon dioxide. *J. Geophys. Res.* **88**: 1263-1269.
Kerr, J.B. and D.I. Wardle, 1993. Ozone and ultraviolet-B monitoring program in Canada. pp 160-170, *In*: Atmospheric Ozone, SPIE Volume 2047. SPIE, Bellingham, WA, USA.
Khalil, M.A.K. and R.A. Rasmussen, 1992. The global sources of nitrous oxide. *J. Geophys. Res.* **97**: 14651-14660.
Khalil, M.A.K. and R.A. Rasmussen, 1983a. Gaseous tracers of Arctic haze. *Environ. Sci. Technol.* **17**: 157-164.
Khalil, M.A.K. and R.A. Rasmussen, 1983b, Sources, sinks and seasonal cycles of atmospheric methane. *J. of Geophys. Res.* **88(C9)**: 5131-5144.
Kieber, D.J., J. McDaniel and K. Mopper, 1989. Photochemical source of biological substrates in sea water: implications for carbon cycling. *Nature* **341**: 637-639.
Koener, R.M., 1973. The mass balance of the sea ice in Arctic Ocean. *J. Glaciol.* **12 (65)**: 173-185.
Kondo, Y., U. Schmidt, T. Sugita, P. Aimedieu, M. Koike, H. Zierreis and Y. Iwasaka, 1994. Total reactive nitrogen, N_2O and ozone in the winter Arctic stratosphere. *Geophys. Res. Lett.* **21**: 1247-1250.
Kondo, Y., U. Schmidt, T. Sugita, A. Engel, M. Koike, P. Aimedieu, M.R. Gunson and J. Rodriguez, 1996. Noy correlation with N_2O and CH_4 in mid-latitude stratosphere. *Geophys. Res. Lett.* **23**: 2369-2372.
Kondratyev, K.A., 1969. Radiation in the Atmosphere. Academic, San Diego, California.
Körner, C., 1994. Impact of atmospheric changes on high mountain vegetation. pp. 155-166, *In*: M. Beniston (ed.). Mountain environments in changing climates. Routledge, London,.
Körner, R.M. and D.A. Fisher, 1990. A 10,000 year record of summer climate: Agassiz Ice Cap, northern Ellesmere Island, Northwest Territories, Canada. *Nature* **343**: 630-633.
Koskela, T. (ed.), 1994. The Nordic intercomparison of ultraviolet and total ozone instruments at Izana from 24 October to 5 November 1993. Final Report. Meteorological Publication No. 27. Finnish Meteorological Institute, Helsinki.
Kovacs, A., 1983. Characteristics of multiyear pressure ridges. pp. 172-182, *In*: Proceedings of the 7th International Conference on Port and Ocean Engineering Under Arctic Conditions (POAC-83), Helsinki.
Krabill, W.B., 1992. Airbourne laser altimetry measurements for sea ice thickness. *In*: A.S. Thorndike, C. Parkinson and D.A. Rothrock (eds.). Report of the Sea Ice Thickness Workshop, 19-21 November, 1991.
Kristiansen, S. and T. Farbot, 1991. Nitrogen uptake rates in the phytoplankton and in the ice algae in the Barents Sea. *Polar Res.* **10**: 187-192.
Kulovaara, M. and P. Backlund, 1993. Effects of simulated sunlight on aquatic humic matter. *Vatten* **49**: 100-103.
Kumer, J.B., J.L. Mergenthaler and A.E. Roche, Claes 1993. CH_4, N_2O and CCl_2F_2 (F12) global data. *Geophys. Res. Lett.* **20**: 1239-1242.
Kuropat, P., 1984. Foraging behavior of caribou on a calving ground in northwestern Alaska. M. S. Thesis, University of Alaska, Fairbanks.
Kvenvolden, K.A., 1993. Gas hydrates - geological perspective and global change. *Reviews in Geophysics* **31**: 173-187.
Kvenvolden, K.A. and T.D. Lorenson, 1995. Arctic oceanic methane and global climate change: Transactions, American Geophysical Union, 76(46F): 120.
Kvenvolden, K.A., M.D. Lilley, T.D. Lorenson, P.W. Barnes and E. McLaughlin, 1993. The Beaufort Sea continental shelf as a seasonal source of atmospheric methane. *Geophys. Res. Lett.* **20**: 2459-2462.
Kylling, A., K. Stamnes and S.C. Tsay, 1995. A reliable and efficient two-stream algorithm for radiative transfer: Documentation of accuracy in realistic layered media. *J. Atmos. Chem.* **21**: 115-150.
Lachenbruch A.H., T.T. Cladouhos and R.W. Saltus, 1988. Permafrost temperatures and changing climate. pp. 9-17, *In*: Permafrost, Fifth International Conference, Proceedings, Volume 3, Trondheim, Norway. Tapir Publishers, Trondheim.
Lammers, S., E. Suess, M.N. Mansurov and V.V. Anikiev, 1995. Variations of atmospheric methane supply from the Sea of Okhotsk induced by the seasonal ice cover. *Global Biogeochemical Cycles* **9**: 351-358.
Langdon, S.J. (ed.), 1986. Contemporary Alaskan Native economies. University Press of America. Lanham. Mayland ix+183p.
Lange, M.A., P. Schlosser, S.F. Ackley, P. Wadhams and G.S. Dieckmann, 1990. 18O concentrations in sea ice of the Weddell Sea, Antarctica. *J. Glaciology* **36**: 315-323.
Lannefors, H., J. Heintzenberg and H.-C. Hansson, 1983. A comprehensive study of physical and chemical parameters of the Arctic summer aerosol; results from the Swedish expedition Ymer-80. *Tellus* **35B**: 40-54.
Lantis, M., 1938. The Alaskan whale cult and its affinities. *American Anthropologist* **40(3)**: 438-464.
Larkum, A.W.D. and W.F. Wood, 1993. The effect of ultraviolet-B radiation on photosynthesis and respiration of phytoplankton, benthic macoralgae and seagrasses. *Photosynthesis Res.* **36**: 17-23.
Lashof, D.A., 1989. The dynamic greenhouse: Feedback processes that may influence future concentrations of atmospheric trace gases and climatic change. *Climate Change* **14**: 213-242.
Lawson, D.E., 1983. Ground ice in perennially frozen sediments, northern Alaska. pp. 695-700, *In*: Permafrost, Fourth International Conference, Proceedings, Washington, D.C. National Academy Press.
LeDrew, E.F., 1985. The dynamic climatology of the Beaufort to Laptev sector of the polar basin for the winters of 1975 and 1976. *J. Climatol.* **5**: 253-272.
LeDrew, E.F., 1988. Development processes for five depression systems within the polar basin. *J. Climatol.* **8**: 125-153.
LeDrew, E.F., D. Johnson and J.A. Maslanik, 1991. An examination of atmospheric mechanisms that may be responsible for the annual reversal of the Beaufort Sea ice field. *Internat. J. of Climatol.* **11(8)**: 841-859.
Letreguilly, A., P. Huybrechts and N. Reeh, 1991. Steady-state characteristics of the Greenland ice sheet under different climates. *J. of Glaciology* **37**: 149-157.
Lewkowicz, A.G., 1992. Factors influencing the distribution and initiation of active-layer detachment slides on Ellesmere Island, Arctic Canada. pp. 223- 251, *In*: J.C. Dixon and A.D. Abrahams (eds.). Periglacial geomorphology. J. Wiley and Sons Ltd.
Lindell, M., W. Granéli, L. Tranvik, 1995. Enhanced bacterial growth in response to photochemical transformation of dissolved organic matter. *Limnol. Oceanogr.* **40**: 195-199.
Lindsay, R.W. and D.A. Rothrock, 1994. Arctic sea ice albedo from AVHRR. *J. Climate* **7**: 1737-1749.
Loewenstein, M., J.R. Podolske and K.R. Chan, 1990. N_2O as a dynamical tracer in the arctic vortex. *Geophys. Res. Lett.* **17**: 477-480.
Loewenstein, M., J.R. Podolske, D.W. Fahey, E.L. Woodbridge, P. Tin, A. Weaver, P.A. Newman, S.E. Strahan, S.R. Kawa, M.R. Schoeberl and L.R. Lait, 1993. New observations of the NO_y/N_2O correlation in the lower stratosphere. *Geophys. Res. Lett.* **20**: 2531-2534.
Logan, J.A., 1994. Trends in the vertical distribution of ozone: An analysis of ozonesonde data. *J. Geophys. Res.* **99**: 25553-25585.
Logan, J.A., M.J. Prather, S C. Wofsy, M.B. McElroy, 1981. Tropospheric chemistry: A global perspective. *J. Geophys. Res.* **86**: 7210-7254.
Lorrain, R.D. and P. Demeur, 1985. Isotopic evidence for relic Pleistocene glacier ice on Victoria Island, Canadian Arctic archipelago. *Arctic and Alpine Res.* **17**: 89-98.
Lowenthal, D.H. and K.A. Rahn, 1985. Regional sources of pollution aerosol at Barrow, Alaska during winter 1979-80 as deduced from elemental tracers. *Atmos. Environ.* **19(12)**: 2011-2024.
Luecke, C. and J.W.J. O'Brien, 1983. Photoprotective pigmentation in a pond morph of *Daphnia middendorffiana*. *Arctic* **36**: 365-368.
Luther, F.M. and R.J. Gelinas, 1976. Effect of molecular multiple scattering and surface albedo on atmospheric photodissociation rates. *J. Geophys. Res.* **81**: 1125-1132.
MacDonald, G., T. Edwards, K. Moser, R. Pienitz, and J.P. Smol, 1993. Rapid response of treeline vegetation and lakes to past climate warming. *Nature* **361**: 243-246.
Macdonald, R.W. and J.M Bewers, 1996. Contaminants in the Arctic environment; priorities for protection. *ICES J. mar. Sci.* **53**: 537-563.
Macdonald R.W. and E.C. Carmack, 1991. The role of large-scale under-ice topography in separating estuary and ocean on an arctic shelf. *Atmos. Ocean* **29**: 37-53.
Mackay, J.R., 1992. The frequency of ice-wedge cracking (1967-1987) at Garry Island, western Arctic coast, Canada. *Can. J. of Earth Sciences* **29**: 236-248.
Mackay, J.R., 1995. Active layer changes (1968 to 1993) following the forest-tundra fire near Inuvik, NWT., Canada. *Arctic and Alpine Res.* **27**: 323-336.
MacLaren, I., 1961. Methods of determining the numbers and availability of ringed seals in the Canadian Eastern Arctic. *Arctic* **14(3)**: 162-175.
Madronich, S., 1987. Photodissociation in the atmosphere: Actinic flux and the effects of ground reflections and clouds. *J. Geophys. Res.* **92**: 9740-9752.
Malmer, N. and B. Wallen. 1996. Peat formation and mass balance in subarctic ombrotrophic peatlands around Abisko, northern Scandinavia. *Ecol. Bull.* **45**: 79-89.
Manabe, S., M.J. Spelman and R.J. Stouffer, 1992. Transient responses of a coupled ocean-atmosphere model to gradual changes of atmospheric CO_2. Part II: Seasonal response. *J. Climate* **5**: 105-126.
Manney, G.L., L. Froidevaux, J.W. Waters and R.W. Zurek, 1995a. Evolution of microwave limb sounder ozone and the polar vortex during winter. *J. Geophys. Res.* **100**: 2953-2972.
Manney, G.L., L. Froidevaux, J.W. Waters, R.W. Zurek, W.G. Read, L.S. Elson, J.B. Kumer, J.L. Mergenthaler, A.E. Roche, A. O'Neill, R.S. Harwood, I. MacKenzie and R. Swinbank, 1994a. Chemical depletion of ozone in the Arctic lower stratosphere during winter 1992-1993. *Nature* **370**: 429-434.
Manney, G.L., R.W. Zurek, L. Froidevaux and J.W. Waters, 1995c. Evi-

dence for Arctic ozone depletion late February and early March 1994. *Geophys. Res. Lett.* **22**: 2941-2944.

Manney, G.L., R.W. Zurek, W.A. Lahoz, R.S. Harwood, J.C. Gille, J.B. Kumer, J.L. Mergenthaler, A.E. Roche, A. O'Neill, R. Swinbank and J.W. Waters, 1995b. Lagrangian transport calculations using UARS data. Part I: passive tracers. *J. Atm. Sci.* **52**: 3049-3068.

Manney, G.L., R.W. Zurek, A. O'Neill, R. Swinbank, J.B. Kumer, J.L. Mergenthaler and A.E. Roche, 1994b. Stratospheric warmings during February and March 1993. *Geophys. Res. Lett.* **21**: 813-816.

Marland, G., R.J. Andres and T.A. Boden, 1994. Global, regional and national CO_2 emissions. pp. 505-584, *In*: T.A. Boden, D.P. Kaiser, R.J. Sepanski and F.W. Stoss (eds.). Trends '93: A Compendium of Data on Global Change, ORNL/CDIAC-65. Carbon Dioxide Information Analysis Center, Oak Ridge National Laboratory, Oak Ridge, Tennessee.

Martin, S. and D.J. Cavalieri, 1989. Contributions of the Siberian shelf polynyas to the Arctic Ocean intermediate and deep water. *J. Geophys. Res.* **94(C9)**: 12725-12738.

Maslanik, J.A. and R.G. Barry, 1990. Remote sensing in Antarctica and the Southern Ocean: Applications and developments. *Antarctic Sci.* **2(2)**: 105-121.

Maslanik, J.A., M.C. Serreze and R.G. Barry, 1996. Recent decreases in Arctic summer ice cover and linkages to atmospheric circulation anomalies. *Geophys. Res. Lett.* **23**: 13, 1677- 1680.

Maslanik, J.A., C. Fowler, J. Key, T. Scambos, T. Hutchinson and W. Emery, 1997. AVHRR-based Polar Pathfinder products for modeling applications. *Annals of Glaciol.* **25**: 388-392.

Maslin, M.M. and C. Tzedakis, 1996. Sultry last interglacial gets sudden chill. *EOS (American Geophysical Union)* **77**: 353-354.

Mason, O.K. and S.C. Gerlach, 1995. Chukchi hot spots, paleo-polynyas and caribou crashes: climatic and ecological dimensions of north Alaska prehistory. *Arctic Anthropology* **32(1)**: 101-130.

Matthews, E., 1993. Wetlands. pp. 314-361, *In*: M.A.K. Khalil (ed.). Atmospheric methane: sources, sinks and role in global change. NATO ASI Series 1: Global Environmental Change, Vol. 13. Springer Verlag, Berlin.

Mauzerall, D.L., D.J. Jacob, S.-M. Fan, J.D. Bradshaw, G.L. Gregory, G.W. Sachse and D.R. Blake, 1996. Origin of tropospheric ozone at remote high northern latitudes. *J. Geophys. Res.* **101**: 4175-4188.

Maxwell, B., 1992. Arctic climate: potential for change under global warming. pp. 11-34, *In*: F.S. Chapin III, R.L. Jefferies, J.F. Reynolds, G.R. Shaver and J. Svoboda (eds.). Arctic ecosystems in a changing climate. An ecophysiological perspective. Academic Press, San Diego, California.

Mayewski, P.A., L.D. Meeker, S. Whitlow, M.S. Twickler, M.C. Morrison, R.B. Alley, P. Bloomfield and K. Taylor, 1993. The atmosphere during the Younger Dryas. *Science* **261**: 195-197.

Maykut, G. A., 1978. Energy exchange over young ice in the Arctic. J. of Geophys. Res. 83: 3646-3658.

Maykut, G.A., 1986. The surface heat and mass balance. pp. 395-463, *In*: N. Untersteiner (ed.). The Geophysics of Sea Ice, NATO ASI Series B: Physics, Vol. 146. Plenum Press, New York.

Maykut, G.A., 1982. Large-scale heat exchange and ice production in the Central Arctic. *J. Geophys. Res.* **87**: 7971-7984.

McConnell, J., G.S. Hendeson, L.A. Barrie, J. Bottenheim, H. Niki, C.H. Langford and E.M.J. Templeton, 1992. Photochemical bromine production implicated in Arctic boundary layer ozone depletion. *Nature* **355**: 150-152.

McCormick, M.P., R.E. Veiga and W.P. Chu, 1992. Stratospheric ozone profile and total ozone trends derived from the SAGE I and SAGE II data. *Geophys. Res. Lett.* **19**:269-272.

McCulloch A., P M. Midgley and D.A. Fisher, 1994. Distribution of emissions of chlorofluorocarbons (CFCs) 11, 12, 113, 114 and 115 among reporting and non-reporting countries in 1986. *Atmos. Environ.* **28(16)**: 2567-2582.

McGovern, T.H., G.F. Bigelow, T. Amorosi, D. Russell, 1988. Northern islands, human error, and environmental degradation: a view of social and ecological change in the medieval North Atlantic. *Human Ecology* **(16)3**: 225-270.

McKenzie, R.L. and M. Kotkamp, 1996. Upwelling UV irradiance and surface albedo measurements at Lauder, New Zealand. *Geophys. Res. Lett.* **23(14)**: 1757-1760.

McKenzie, R.L., M. Kotkamp, G. Seckmeyer, R. Erb, C.R. Roy, H.P. Gies and S.J. Toomey, 1993. *Geophys. Res. Lett.* **20**: 2223-2226.

McLaren, A.S., 1989. The under-ice thickness distribution of the Arctic Basin as recorded by USS Nautilus in 1958 and USS Queenfish in 1970. *J. Geophys. Res.* **94**: 4971-4983.

McLaren, A.S., R.G. Barry and R.H. Bourke, 1990. Could Arctic ice be thinning? *Nature* **345**: 762.

McLaren, A.S., M.C. Serreze and R.G. Barry, 1987. Seasonal variations of sea ice motion in the Canada Basin and their implications. *Geophys. Res. Lett.* **14**: 1123-1126.

McLaren A.S., J.E. Walsh, R.H. Bourke, R.L. Weaver and W. Wittman, 1992. Variability in sea-ice thickness over the North Pole from 1979-1990. *Nature* **358**: 224-226.

McLaughlin, F.A., E.C. Carmack, R.W. Macdonald and J.K.B. Bishop, 1996. Physical and geochemical properties across the Atlantic/Pacific water mass front in the southern Canadian Basin. *J. Geophys. Res.* **101**: 1183-1197.

McPeters, R.D., S.M. Hollandsworth, and J.R. Herman, 1996. Long-term trends derived from the 16-year combined Nimbus-7/Meteor-3 TOMS version 7 record. *Geophys. Res. Lett.* **23**:3699-3702.

Meehl, G.A. and W.M. Washington, 1990. CO_2 climate sensitivity and snow-sea-ice albedo parameterization in an atmospheric GCM coupled to a mixed-layer ocean model. *Climatic Change* **16(3)**: 283-306.

Meier, M.F., 1984. Contribution of small glaciers to global sea level. *Science* **226**: 1418-1421.

Meier, M.F., 1993. Ice, climate and sea level: do we know what is happening? pp. 141-160, *In*: W.R. Peltier (ed.). Ice in the Climate System, NATO ASI Series, v. I12. Springer-Verlag, Berlin,.

Melling, H. and E.L. Lewis, 1982. Shelf drainage flows in the Beaufort Sea and their effect on the Arctic Ocean Pycnocline. *Deep-Sea Res.* **29**: 967-985.

Michaels, P.J., P.C. Knappenberger and D.A. Gay, 1995. Predicted and observed long night and day temperature trends. *Atmos. Res.* **37**: 257-266.

Mickle, R.E., J.W. Bottenheim, W.R. Leaitch and W. Evans, 1989. Boundary layer ozone depletion during AGASP-II. *Atmos. Environ.* **23**: 2443-2449.

Miles, C.J. and L. Brezonik, 1981. Oxygen consumption in humic-colored waters by a photochemical ferrous-ferric catalytic cycle. *Environ. Sci. Technol.* **15**: 1089-1095.

Miller, A.J., 1989. A review of satellite observations of atmospheric ozone. *Planet. Space Sci.* **37**: 1539-1554.

Miller, W.L. and R.G. Zepp, 1995. Photochemical production of dissolved inorganic carbon from terrestrial organic matter: Significance to the oceanic organic carbon cycle. *Geophys. Res. Lett.* **22**: 417-420.

Mills, P.F., 1994. The Agricultural potential of Northwestern Canada and Alaska and the impact of climatic change. *Arctic* **47(2)**:115-123.

Mitchell, J., S. Manabe, V. Meleshko and T. Tokioka, 1990. Equilibrium climate change - and its implications for the future. pp. 131-172, *In*: J. Houghton, G. Jenkins and J. Ephraums (eds.). Climate change: the IPCC Scientific Assessment. Cambridge University Press, Cambridge, U.K.

Moan, J. and A. Dahlback, 1993. Ultraviolet radiation and skin cancer: Epidemiological data from Scandinavia. *In*: A.R. Young, L.O. Björn, J.M. Moan and W. Nultsch (eds.). Environmental UV photobiology. Plenum Press, New York and London.

Molau, U., 1996. Climatic impacts on flowering, growth, and vigour in an arctic-alpine cusion plant, *Diapensia lapponica*, under different snow cover regimes. *Ecol. Bull.* **45**: 210-219.

Moorhead D.L. and Callaghan T.V., 1994. Effects of increasing ultraviolet-B radiation on decomposition and soil organic matter dynamics: a synthesis and modelling study. *Biology and Fertility of Soils* **18**: 19-26.

Mopper, K. and W.L. Stahovec, 1986. Sources and sinks of low molecular weight organic carbonyl compounds in seawater. *Mar. Chem.* **19**: 305-321.

Mopper, K., X. Zhou, R.J. Kieber, D.J. Kieber, R.J. Sikorski, R.D. Jones, 1991. Photochemical degradation of dissolved organic carbon and its impact on the oceanic carbon cycle. *Nature* **353**: 60-62.

Moran, M.A. and R.E. Hodson, 1994. Support of bacterioplankton production by dissolved humic substances from three marine environments. *Mar. Ecol. Prog. Ser.* **110**: 241-247.

Moritz, R.E., J.A. Curry, A.S. Thorndike and N. Untersteiner, 1993. SHEBA: A research program on the surface heat budget of the Arctic Ocean. NSF ARCSS Workshop on Clouds, Radiation and Surface Energy Balance of the Arctic Ocean, Orlando, Florida, 8-10 February 1993. Polar Science Center, University of Washington, Seattle.

Morris, D.P., H. Zagarese, C.E. Williamson, E. Balseiro, B.R. Hargreaves, B. Modenutti, R. Moeller and C. Queimalinos, 1995. The attenuation of solar UV radiation in lakes and the role of dissolved organic carbon. *Limnol. Oceanogr.* **40**: 1381-1391.

Müller-Niklas, G., A. Heissenberger, S. Puskaric and G.J. Herndl, 1995. Ultraviolet-B radiation and bacterial metabolism in coastal waters. *Aquatic Microbial Ecol.* **9**: 111-116.

Mysak, L.A., 1995. Interdecadal variability in the Arctic and northern North Atlantic: Observations and models. Fourth Conference on Polar Meteorology and Oceanography, Dallas, Texas, American Meteorological Society J9: 26-27.

Mysak, L.A. and D.K. Manak, 1989. Arctic sea-ice extent and anomalies, 1953-1984. *Atmosphere-Ocean* **27**: 376-405.

Mysak, L.A., D.K. Manak and R.F. Marsden, 1990. Sea-ice anomalies observed in the Greenland and Labrador seas during 1902-1984 and their relation to an interdecadal climate cycle. *Climate Dynamics* **5**: 111-133.

Nadelhoffer, K.J., A.E. Giblin, G.R. Shaver and A.E. Linkins, 1992. Microbial processes and plant nutrient availability in Arctic soils. pp. 281-300, *In*: F.S. Chapin, R.L. Jeffries, U.R. Shaver and J. Svoboda (eds.). Arctic Ecosystems in a Changing Climate. An Ecophysiological Perspective. Academic Press, San Diego.

Nagurny, A.P., 1995. Space-time distribution of temperature inversions in the Arctic atmospheric boundary layer. *Ann. Geophysicae* **13**: 1087-1092.

Nakajima, H., Y. Kondo, T. Sugita, M. Koike, U. Schmidt, A. Engel and P. Aimedieu, 1996. Correlation of stratospheric total reactive nitrogen and nitrous oxide in the Arctic winter during SESAME campaign. Polar stratospheric ozone, Proc. 3rd European. Workshop, 201-204.

Newman, P., J. Gleason, R. McPeters and R. Stolarski, 1997. Anomalously low ozone over the Arctic. *Geophys. Res. Lett.* **24**:2689-2692.

Newsham, K.K., A.R. McCleod, P.D. Greenslade and B.A. Emmett, 1996. Appropriate controls in outdoor UV-B supplementation experiments. *Global Change Biology* **2**: 319-324.

Newton, J., 1995. An assessment of coping with environmental hazards in Northern aboriginal communities. *Canadian Geographer* **39(2)**: 112-120.
Nicholson, B.J., and L.D. Gignac, 1995. Ecotope dimensions of peatland bryophyte indicator species along gradients in the Mackenzie River Basin, Canada. *The Bryologist* **98**: 437-451.
Nicholson, B.J., L.D. Gignac, and S.E. Bayley, 1996. Effects of ecological and climatic gradients on the distribution of peatland bryophyte communities in the Mackenzie River Basin. *Vegetatio* **1-6(2)**: 119-133.
Nicolet, M., R.R. Meier and D.E. Anderson Jr., 1982. Radiation field in the troposphere and stratosphere from 240 to 1000 nm, II - Numerical analysis. *Planet. Space Sci.* **30**: 935-983.
Nisbet, E.G., 1989. Some northern sources of atmospheric methane: production, history and future implications. *Can. J. of Earth Sciences* **26**: 1603-1611.
Notholt J., 1994. FTIR measurements of HF, N_2O and CFCs during the Arctic polar night with the moon as light source, subsidence during winter 1992/3. *Geophys. Res. Lett.* **21(22)**: 2385-2399.
Novelli, P.C., L.P. Steele, and P.P. Tans, 1992. Mixing ratios of carbon monoxide in the troposphere. *J. Geophys. Res.* **97**: 20731-20750.
O'Brien, S.R., P.A. Mayewski, L.D. Meeker, D.A. Meese, M.S. Twickler and S.I. Whitlow, 1995. Complexity of Holocene climate as reconstructed from a Greenland ice core. *Science* **270**: 1962-1964.
Oberbauer, S.F. and T.E. Dawson, 1992. Water relations of arctic vascular plants. pp. 259-279, *In*: F.S. Chapin III, R.L. Jefferies, J.F. Reynolds, G.R. Shaver and J. Svoboda (eds.). Arctic Ecosystems in Changing Climate: An ecophysiological perspective. Academic Press: San Diego, California.
Oechel, W.C. and G.L. Vourlitis, 1994. The effects of climate change on land-atmosphere feedbacks in arctic tundra regions. *Trends in Ecology and Evolution* **9**: 304-329.
Oechel, W.C., S. Cowles, N. Grulke, S.J. Hastings, B. Lawrence, T. Prudhomme, G. Riechers, B. Strain, D. Tissue and G. Vourlitis, 1994. Transient nature of CO_2 fertilization in Arctic tundra. *Nature* **371**: 500-503.
Oechel, W.C., S.J. Hastings, M. Jenkins, G. Reichers, N. Grulke and G. Vorlitis, 1993. Recent change of Arctic tundra ecosystems from a carbon sink to a source. *Nature* **361**: 520-526.
Oelhaf, H., T.V. Clarmann and H. Fischer, 1994. Stratospheric $ClONO_2$ and HNO_3 profiles inside the Arctic vortex from MIPA's-B Limb emission spectra obtained during EASOE. *Geophys. Res. Lett.* **21(13)**: 1263-1273.
Oerlemans, J., 1994. Quantifying global warming from the retreat of glaciers. *Science* **264**: 243- 245.
Ogilvie, A.E.J., 1982. Climate and society in Iceland from the medieval period to the late eighteenth century. Ph. D. Thesis, School of Environmental Sciences, University of East Anglia, Norwich, U.K.
Ogilvie, A E.J., 1984. The impact of climate on grass growth and hay yield in Iceland: A.D. 1601 to 1780. pp. 343-352, *In*: N.-A. Morner and W. Karlen (eds.). Climate Changes on a Yearly to Millennial Basis. Reidel, Dordrecht.
Ogilve, A.E.J., 1991. Climatic changes in Iceland A.D. c. 865 to 1598. pp. 233-251, *In*: G.F. Bigelow (presentor). The Norse of the North Atlantic. *Acta Archaeologica* **61**-1990. Munksgaard, Copenhagen.
Ogilve, A.E.J., 1992. Documentary evidence for changes in the climate of Iceland, A.D. 1500 to 1800. pp. 92-117, *In*: R.S. Bradley and P.D. Jones (eds.). Climate Since A.D. 1500. Routledge, London,.
Ogilvie, A.E.J., 1996. Sea-ice conditions off the coasts Iceland A.D. 1601-1850 with special reference to part of the Maunder Minimum period (1675-1715). AmS-Varia 25, Archaeological Museum of Stavanger, Norway, 9-12.
Ogren, J.A. and R.J. Charlson, 1983. Elemental carbon in the atmosphere: Cycle and lifetime. *Tellus* **35B**: 241-254.
Oiestad, F., 1990. Effects of climate change on fisheries and fish farming, Univ. of Tromsø. FiskenHav, vol 2. (In Norwegian).
Olivarius, F. de Fine, H.C. Wulf, J. Crosby and M. Norval, 1996. The usnscreening effect of urocanic acid. *Photodermatol. Photoimmunol. Photomed.* **22(3)**: 95-105.
Oltmans, S.J., 1991. Arctic ozone chemistry. pp.195-215, *In*: W.T. Sturges (ed..). Pollution of the Arctic Atmosphere. Elsevier, London.
Oltmans, S.J., 1993. Climatology of Arctic and Antarctic tropospheric ozone. pp. 26-40, *In*: H. Niki and K.H. Becker (eds.). The tropospheric chemistry of ozone in the polar regions. NATO ASI series, vol. 17. Springer-Verlag, Berlin.
Oltmans, S.J. and H. Levy II, 1994. Surface ozone measurements from a global network. *Atmos. Environ.* **28**: 9-24.
Oltmans, S.J., W.E. Raatz and W.D. Komhyr, 1989a. On the transfer of stratospheric ozone into the troposphere near the North Pole. *J. Atmos. Chem.* **9**: 245-253.
Oltmans, S.J., R.C. Schnell, P.J. Sheridan, R.E. Peterson, S.M. Li, J.W. Winchester, P.P. Tans, W.T. Sturges, J.D. Kahl and L.A. Barrie, 1989b. Seasonal surface ozone and filterable bromine relationship in the High Arctic. *Atmos. Environ.* **23**: 2431-2441.
Orr, P.H. and A.R. Birt, 1984. Hereditary polymorphic light erruption in Canadian Inuit. *Int. G. Dermatol.* **7**: 472-475.
Osterkamp, T., 1994. Evidence for warming and thawing of discontinuous permafrost in Alaska. *EOS (American Geophysical Union)* **75(44)**: 85.
Oswood, M., 1989. Community structure of benthic invertebrates in interior Alaskan (USA) streams and rivers. *Hydrobiologia*, v. **172**.
Ottar, B., 1981. The transfer of airborne pollutants to the Arctic region. *Atmos. Environ.* **15(8)**: 1439-1445.
Overland, J.E., 1985. Atmospheric boundary layer structure and drag coefficients over sea ice. *J. Geophys. Res.* **90**: 9029-9049.
Overland, J.E. and P.S. Guest, 1991. The Arctic snow and temperature budget over sea ice during winter. *J. Geophys. Res.* **96**: 4651-4662.
Overpeck, J.T., 1995. Paleoclimatology and climate system dynamics. Review of Geophysics 33, Supplement B: 863-871.
Overpeck, J.T., 1996. Warm climate surprises. *Nature* **271**: 1820-1821.
Overpeck, J.T., P.J. Bartlein and T. Webb III, 1991. Potential magnitude of future vegetation change in eastern North America: comparisons with the past. *Science* **254**: 692-695.
Parker, D.E. and D.I. Cox, 1995. Towards a consistent global climatological radiosonde data base. *Internat. J. of Climatology* **15**: 473-496.
Parker, D.E., C.K. Folland and M. Jackson, 1995. Marine surface temperature: Observed variations and data requirements. *Clim. Change* **31**: 559-600
Parkinson, C.L., 1992. Spatial patterns of increases and decreases in the length of the sea ice season in the north polar region, 1979-1986. *J. Geophys. Res.* **97**: 14377-14388.
Parkinson, C.L. and D.J. Cavalieri, 1989. Arctic sea ice 1973-1987. seasonal, regional and interannual variability. *J. Geophys. Res.* **94**: 14499-14523.
Parkinson, C.L., J. Comiso, H.J. Zwally, D.J. Cavalieri, P. Gloersen and W.J. Campbell, 1987. Arctic sea ice, 1973-1976: Satellite passive microwave observations, NASA Special Publication, SP-489. 295p.
Parry, M.L., 1978. Climatic Change, Agriculture and Settlement, Studies in Historical Geography. Dawson, Folkestone.
Parsons, A.N., J.M. Welker, P.A. Wookey, M.C. Press, T.V. Callaghan and J.A. Lee, 1994. Growth responses of four sub-arctic dwarf shrubs to simulated environmental change. *J. of Ecology* **82**: 307-318.
Paterson, W.S.B., 1994. The Physics of Glaciers, Third Edition. Pergamon Press, Oxford.
Pease, C.H., 1987. The size of wind-driven coastal polynyas. *J. Geophys. Res.* **92**, 7049-7059.
Permafrost Subcommittee, 1988. Glossary of permafrost and related ground-ice terms. Associate Committee on Geotechnical Research, National Research Council of Canada, Ottawa. 156p.
Perovich, D.K., 1993. A theoretical model of ultraviolet light transmission through Antarctic sea ice. *J. Geophys. Res.* **98**: 22579-22587.
Perovich, D.K., 1996. The optical properties of sea ice. CRREL Monograph 96-1, U.S. Cold Regions Research and Engineering Laboratory, Hanover, NH. 25p.
Peterson, D.L. and D.R. Johnson (eds.), 1995. Human Ecology and Climate Change: People and Resources in the Far North. Taylor and Francis, Washington DC.
Petzoldt, K., B. Najjokat and K. Neugebohren., 1994. Correlation between stratospheric temperature, total ozone and tropospheric weather systems. *Geophys. Res. Lett.* **21(13)**: 1203-1206.
Pienitz, R. and J.P. Smol, 1993. Diatom assemblages and their relationship to environmental variables in lakes from the boreal forest-tundra ecotone near Yellowknife, Northwest Territories, Canada. *Hydrobiologia* **269/270**: 391-404.
Pienitz, R., I.R. Walker, B.A. Zeeb, J.P. Smol and P.R. Leavitt, 1992. Biomonitoring past salinity changes in an athalassic sub-arctic lake. *Int. J. Salt Lake Res.* **1(2)**: 91-123.
Pienitz, R., J.P. Smol and H.J.B. Birks, 1995. Assessment of freshwater diatoms as quantitative indicators of past climatic change in the Yukon and Northwest Territories, Canada. *J. Paleolimnology* **13**: 21-49.
Podolske, J.R., M. Loewenstein, A. Weaver, S.E. Strahan and K.R. Chan, 1993. Northern hemisphere nitrous oxide morphology during the 1989 AASE and the 1991-1992 AASE II campaigns. *Geophys. Res. Lett.* **20**: 2535-2538.
Polar Group, The, 1980. Polar atmosphere-ice-ocean processes: A review of polar problems in climate research. D.J. Baker, U. Radok and G. Weller (eds.). Reviews of Geophysics and Space Physics 18: 525-543.
Pollard, W.H., 1991. Observations on massive ground ice on the Fosheim Peninsula, N.W.T. pp. 223-231, *In*: Current Research, Part E, Geological Survey of Canada Paper 91(1E).
Pollard, W.H. and S.R. Dallimore, 1988. Petrographic characteristics of massive ground ice, Yukon Coastal Plain, Canada. pp. 224-229, *In*: Permafrost, Fifth International Conference, Proceedings, Volume 1, Trondheim, Norway. Tapir Publishers, Trondheim.
Pollard, W.H. and H.M. French, 1980. A first approximation of the volume of ground ice, Richard Is., Pleistocene Mackenzie Delta, NWT. *Canadian Geotechnical Journal* **17**: 509-16.
Powell, R.D., 1990. Advance of glacial tidewater fronts in Glacier Bay, Alaska. pp. 67-73, *In*: A.M. Milner, J.D. Wood Jr. (eds.). Proceeding of the Second Glacier Bay Science Symposium.. U.S. Dept of the Interior, National Park Service. Anchorage, Alaska 99503.
Power, S.B. and L.A. Mysak, 1994. On the interannual variability of Arctic sea-level pressure and sea ice. *Atmosphere-Ocean* **33**: 1-23.
Prentice, I.C., P.J. Bartlein and T. Webb III, 1991. Vegetation and climate change in eastern North America since the last glacial maximum. *Ecology* **72**: 2038-2056.
Prézelin, B.B., N.P. Boucher and O. Schofield, 1994a. Evaluation of field studies of ultraviolet-B radiation effects on Antarctic marine primary productivity. pp. 181-194, *In*: H. Biggs and M. E. B. Joyner (eds.). Stratospheric Ozone Depletion/ ultraviolet-B Radiation in the Biosphere. NATO ASI (Advance Study Institute) Series 118. Springer-Verlag, Heidelberg.

Prézelin, B.B., N.P. Boucher and R.C. Smith, 1994b. Marine primary production under the influence of the Antarctic ozone hole: Icecolors '90. pp. 159-186, *In*: S. Weiler and P. Penhale (eds.). Ultraviolet Radiation and Biological Research in Antarctica. Antarctic Research Series 62.

Prézelin, B.B., S. Madronich and H.A. Matlick, 1995. Springtime variability and predictive accuracy of in situ biological weighting functions for UV inhibition of primary production in Antarctic phytoplankton. ASLO, abst. June Mtg, Reno. Limnology and Oceanography, A:32. Seattle, Washington.

Prézelin, B.B., M.A. Moline and H.A. Matlick, 1997. Icecolors '93: spectral UV effects on Antarctic frazil ice algae. *In*: M. Lizotte and K. Arrigo (eds.). Sea Ice Biology. Antarctic Research Series. American Geophysical Union, Washington, D.C.

Proffitt, M.H., J.J. Margitan, K.K. Kelly, M. Loewenstein, J.R. Podolske and K.R. Chan, 1990. Ozone loss in the Arctic polar vortex inferred from high-altitude air-craft measurements. *Nature* **347**: 31-36.

Psenner, R. and R. Schmidt, 1992. Climate-driven pH control of remote alpine lakes and effects of acid deposition. *Nature* **356**: 781-783.

Quadfasel, D., A. Sy, D. Wells and A. Tunik, 1991. Warming in the Arctic. *Nature* **350**: 385.

Raatz, W.E. ; R.C. Schnell, B.A. Bodhaine, 1985. Atmospheric cross sections for the Arctic gas and aerosol sampling program, March-April 1983. NOAA Technical Memorandum, ERL ARL-134. U.S. Department of Commerce, Washington, DC. iv, 50p.

Rahmstorf, S., 1995. Bifurcation of the Atlantic thermohaline circulation in response to changes in the hydrological cycle. *Nature* **378**: 145-149.

Rahn, K.A., 1981. Relative importances of North America and Eurasia as sources of Arctic aerosol. *Atmos. Environ.* **15**(**8**): 1447-1455.

Rahn, K.A. and D.H. Lowenthal, 1986. Who's polluting the Arctic? Why is it so important to know? An American perspective. pp. 85-95, *In*: B. Stonehouse (ed.). Arctic Air Pollution. Cambridge University Press, Cambridge, U.K.

Raich, J.W. and W.H. Schlesinger, 1992. The global carbon dioxide flux in soil respiration and its relationship to vegetation and climate. *Tellus* **44B**: 81-99.

Ramaswamy V., M.D. Schwarzkopf and K.P. Shine, 1992. Radiative forcing of climate from halocarbon-induced global stratospheric ozone loss. *Nature* **355**: 810-812.

Randall, C.E., D.W. Rusch, R.M. Bevilacqua, J. Lumpe, T.L. Ainsworth, D. Debrestian, M. Fromm, S.S. Krigman, J.S. Hornstein, E.P. Shettle, J.J. Olivero and R.T. Clancy, 1995. Preliminary results from POAM II: Stratospheric ozone at high northern latitudes. *Geophys. Res. Lett.* **22**: 2733-2736.

Randall, D.A., R.D. Cess, J.P. Blanchet, S. Chalita, R. Colman, D.A. Dazlich, A.D. Del Genio, E. Keup, A. Lacis, H. Le Treut, X.-Z. Liang, A.J. McAvaney, J.F. Mahfout, V.P. Meleshkoe, J.-J. Morcrette, P.M. Norris, G.L. Potter, L. Rikus, E. Roeckner, J.F. Royer, U. Schlese, D.A. Sheininin, A.P. Sokolov, K.E. Taylor, R.T. Wetherald, I.Yagai, M.-H. Zhang, 1994. Analysis of snow cover feedbacks in 14 general circulation models. *J. Geophys. Res.* **99**(**D10**): 20757-20767.

Randel, W.J. and J.B. Cobb, 1994. Coherent variations of monthly mean total ozone and lower stratospheric temperature. *J. Geophys. Res.* **99**: 5433-5447.

Randel, W.J., F. Wu, J.M. Russell III, J.W. Waters and L. Froidevaux, 1995. Ozone and temperature changes in the stratosphere following the eruption of Mount Pinatubo. *J. Geophys. Res.* **100**(**16**): 753-764.

Raymo, M.E., D. Rind and W.F. Ruddiman, 1990. Climatic effects of reduced Arctic sea ice limits in the GISS II general circulation model. *Paleoceanography* **5**(**3**): 367-382.

Reeburgh, W. S., 1985. Fluxes associated with brine motion in growing sea ice. Polar Biology 3(1): 29-33.

Reinsel, G.C., G.C. Tiao, D.J. Wuebbles, J.B. Kerr, A.J. Miller, R.M. Nagatani, L. Bishop and L.H. Ying, 1994. Seasonal trend analysis of published ground-based and TOMS total ozone data through 1991. *J. Geophys. Res.* **99**:5449-5464.

Revelle, R.R., 1983. Probable future changes in sea level resulting from increased atmospheric carbon dioxide. pp. 433-448, *In*: Changing Climate. National Academy Press, Washington D.C.

Reynolds J.F. and J.D. Tenhunen (eds.), 1996. Landscape Function and Disturbance in Arctic Tundra. Ecological Studies 120. Springer-Verlag, Berlin, Heidelberg.

Rhodes, T.E. and R.B. Davis, 1995. Effects of late halocene forrest disturbance and vegetation change on acidic mud pond, Maine, USA. *Ecology* **76**: 734-746.

Ringelberg, J., 1980. Aspects of red pigmentation in zooplankton, especially copepods. pp 91-97, *In*: W.C. Kerfoot (ed.). Ecology and evolution of zooplankton communities. University of New England, Hanover, New Hampshire.

Ringelberg, J., A.L. Keyser and B.J.G. Flik, 1984. The mortality effect of ultraviolet radiation in a translucent and in a red morph of *Acanthodiaptomus denticornis* (Crustacea, Copepoda) and its possible ecological relevance. *Hydrobiologia* **112**: 217-222.

Robberecht, R., M.M. Caldwell and W.D. Billings, 1980. Leaf ultraviolet optical properties along a latitudinal gradient in the arctic-alpine life zone. *Ecology* **61**: 612-619.

Robberts, R.J. and A.M. Bullock, 1981. Recent observations on the pathological effects of ultraviolet light on fish skin. *Fish. Pathol. Tokyo.* **15**: 237-239.

Robinson, D.A., K.F. Dewey and R.R. Heim Jr., 1993. Global snow cover monitoring: an update. *Bull. Amer. Meteorol. Soc.* **74**: 1689-1696.

Robinson, D., M. Serreze, R. Barry, G. Scharfen and G. Kukla, 1992. Large-scale patterns and variability of snow melt and parameterized surface albedo in the Arctic basin. *J. Climate* **5**: 1109-1119.

Robinson, R.S., H.F. Diaz, J.K. Eischeid, P.D. Jones, P.M. Kelly and C.M. Goodess, 1987. Precipitation fluctuations over Northern Hemisphere land areas since the mid-19th century. *Science* **237**: 171-275.

Rogers, J.C. and H. van Loon, 1979. The seesaw in winter temperatures between Greenland and northern Europe. Part 2: Some oceanic and atmospheric effects in middle and high latitudes. *Mon. Wea. Rev.* **107**: 509-519.

Roldugin, V.C. and K. Henriksen, 1996. Horizontal movements in the ozone layer. *J. of Atmospheric and Terrestrial Physics* **58**(**12**): 1323-1333.

Rooth, C., 1982. Hydrology and ocean circulation. *Prog. Oceanogr.* **11**: 131-149.

Rosen, H., T. Novakov and B. A. Bodhaine, 1981. Soot in the Arctic. *Atmos. Environ.* **15**: 1371-1375.

Ross, R. and W. Elliott, 1997. Tropospheric water vapor climatology and trends over North America: 1973-1993. *J. Climate and App. Met.* **9**(**12**): 3561-3571.

Rouse, W.R., M. Douglas, R. Hecky, G. Kling, L. Lesack, P. Marsh, M. McDonald, B. Nicholson, N. Roulet and J. Smol, 1997. Effects of climate change on fresh waters of Region 2: Arctic and Subarctic North America. *Hydrologic Processes* **11**: 873-902.

Rowntree, P., 1989. The needs of climate modelers for water runoff. The Global Runoff Data Project, Workshop on the Global Runoff Data Set and Grid Estimation (Appendix C), June 1989, Koblenz, Germany.

Rusch, D.W., R.T. Clancy and P.K. Bhartia, 1994. Comparison of satellite measurements of ozone and ozone trends. *J. Geophys. Res.* **99**: 20501-20511.

Ruth, S.L., J.J. Remedios, B.N. Lawrence and F.W. Taylor, 1994. Measurements of N_2O by the UARS improved stratospheric and mesospheric sounder during the early northern winter 1991/92. *J. Atm. Sci.* **51**: 2818-2828.

Sahsamanoglou, H.S., T.J. Makrogiannis and P.P. Kallimopoulos, 1991. Some aspects of the basic characteristics of the Siberian anticyclone. *Int. J. Climatol.* **11**: 827-839.

Salonen, K. and A. Vahatalo, 1994. Photochemical mineralisation of dissolved organic matter in Lake Skjervatjern. *Environ. Internat.* **20**: 307-312.

Salonen, K., T. Kairesalo and R.I. Jones, 1992. Dissolved organic matter in lacustrine ecosystems: Energy source and system regulator. Kluwer Academic Publishers, Dordrecht. 291p.

Saltzman, E.S., P.-Y. Whung and P.A. Mayewski, 1998. Methanesulfonate in the GISP2 core. *J. Geophys. Res.* (In press).

Santee, M.L., W.G. Read, J.W. Waters, L. Froidevaux, G.L. Manney, Da.A. Flower, R.F. Jarnot, R.S. Harwood and G.E. Peckham, 1995. Interhemispheric differences in polar stratospheric HNO_3, H_2O, ClO and O_3. *Science* **267**: 849-852.

Scharfen, G., R.G. Barry, D.A. Robinson, G. Kukla and M.C. Serreze, 1987. Large-scale patterns of snow melt on Arctic sea ice mapped from meteorological satellite imagery. *Annals of Glaciology* **9**: 200-205.

Schauburger, G., 1992. Anisotropic model for the diffuse biologically-effective irradiance of solar UV radiation on inclined surfaces. *Theor. Appl. Climatol.* **46**: 45-51.

Schindler, D.W., S.E. Baylery, B.R. Parker, K.G. Beaty, D.R. Cruikshank, E.J. Fee, E.U. Schindler and M.P. Stainton, 1996b. The effects of climatic warming on the properties of boreal lakes and streams at the Experimental Lakes Area, Northwestern Ontario. *Limnol. Oceanogr.* **41**(**5**): 1004-1014.

Schindler, D.W., S.E. Baylery, P.J. Curtis, B.R. Parker, M.P. Stainton and C.A. Kelly, 1992. Natural and man-caused factors affecting the abundance and cycling of dissolved organic substances in precambrian shield lakes. *Hydrobiologia* **229**: 1-21.

Schindler, D.W., K. Beaty, E.J. Fee, D.R. Cruishank, E.R. DeBruyn, D.L. Findlay, G.A. Lindsey, J.A. Shearer, M.P. Stainton and M.A. Turner, 1990. Effects of climatic warming on the lakes of the central boreal forest. *Science* **250**: 967-970.

Schindler, D.W., P.J. Curtis, S.E. Baylery, B.R. Parker, K.G. Beaty and M.P. Stainton, 1997. Climate induced changes in the dissolved organic carbon budgets of boreal lakes. *Biogeochemistry* **36**: 9-28.

Schindler, D.W., P.J. Curtis, B. Parker and M.P. Stainton, 1996a. Consequences of climate warming and lake acidification for ultraviolet-B penetration in North American boreal lakes. *Nature* **379**: 705-708.

Schindler, D.W., H.E. Welch, J. Kalff, G.J. Brunskill and N. Kritsch, 1974. Physical and chemical limnology of Char Lake, Cornwallis Island (75N Lat.). *J. Fish. Res. Board, Canada* **31**: 585-607.

Schledermann, P., 1980. Polynyas and prehistoric settlement patterns. *Arctic* **33**: 292-302.

Schlesinger, M.E. and J.F.B. Mitchel, 1987. Climate model simulations of equilibrium climate response to increased carbon dioxide. *Review of Geophysics* **25**: 760-798.

Schlesinger, W.H., 1977. Carbon balance in terrestrial detritus. *Ann. Rev. Ecol. Syst.* **8**: 51-81.

Schmidt U. and A. Khedim, 1991. In situ measurements of carbon dioxide in the winter arctic vortex and at midlatitudes: an indicator of the 'age' of stratospheric air. *Geophys. Res. Lett.* **18**(**4**): 763-766.

Schofield, O., B.M.A. Kroon, B.B. Prézelin, 1995. Impact of ultraviolet-B radiation on photosystem II activity and its relationship to the inhibition of carbon fixation rates for Antarctic ice algae communities. *J.*

Phycol. **31**: 703-715.
SCOPE, 1992. Effects of increased ultraviolet radiation on biological systems. Scientific Committee on Problems of the Environment (SCOPE), 51 bd de Montmorency, 75016 Paris, France.
SCOPE, 1993. Effects of increased ultraviolet radiation on global ecosystems. Scientific Committee on Problems of the Environment (SCOPE), 51 bd de Montmorency, 75016 Paris, France.
Scully, N.M. and D.R.S. Lean, 1994. The attenuation of ultraviolet radiation in temperate lakes. pp. 135-144, *In*: C.E. Williamson and Zagarese (eds.). Advances in Limnology: Impact of ultraviolet-B radiation on pelagic ecosystems. *Arch. Hydrobiologie* **43**.
Seckmeyer, G., S. Thiel, M. Blumthaler, P. Fabian, S. Geber, A.G. Helminger, D.-P. Häder, C. Kettner, U. Kohler, P. Kopke, H. Maier, J. Schafer, P. Suppan, E. Tamm and E Thomalla, 1994. Intercomparison of spectral UV radiation measurement systems. *Applied Optics* **33**: 7805-7812.
Semtner, A.J., 1987. A numerical study of sea ice and ocean circulation in the Arctic. *J. Phys. Ocean.* **17**: 1077-1099.
Serreze, M.C. and R.G. Barry, 1988. Synoptic activity in the Arctic basin, 1979-1985. *J. Climate* **1**: 1276-1295.
Serreze, M.C., J.E. Box, R.G. Barry and J.E. Walsh, 1993. Characteristic of Arctic synoptic activity, 1952-1989. *Meteorol. Atmos. Phys.* **51**: 147-164.
Serreze, M.C., J.D. Kahl and R.C. Schnell, 1992. Low-level temperature inversions of the Eurasian Arctic and comparisons with Soviet drifting station data. *J. Climate* **5**: 615-630.
Serreze, M.C., J.A. Maslanik, J.R. Key and R.F. Kokaly, 1996. Diagnosis of the record minimum in Arctic sea ice area during 1990 and associated snow cover extremes. *Geophys. Res. Lett.* **22(16)**: 2183-2186
Serreze, M.C., J.A. Maslanik, R.H. Preller and R.G. Barry, 1990. Sea ice concentrations in the Canada Basin during 1988: comparisons with other years and evidence of multiple forcing mechanisms. *J. Geophys. Res.* **95(C12)**: 22253-22267.
Serreze, M.C., M.C. Rehder, R.G. Barry, J.D. Kahl and N.A. Zaitseva, 1995. The distribution and transport of atmospheric water vapor over the Arctic Basin. *Int. J. Climatol.* **15**: 709-727.
SESAME (Second European Stratospheric Arctic and Middle Latitude Experiment 1994-1995), 1995. Confirmed ozone loss over the Arctic. Europe Environment, 452 (ptIII), 4.
Shapiro, M.A., T. Hampel and L.S. Fedor, 1989. Research aircraft observations of an Arctic front over the Barents Seas. pp. 279-290, *In*: P.F. Twitchell, E.A. Rasmussen and K.L. Davidson (eds.). Polar and Arctic Lows. A. Deepak.
Sharratt, B.S., 1992. Growing season trends in the Alaskan Climate Record. *Arctic* **45(2)**: 124-127.
Shaver, G.S., W.D. Billings, F.S. Chapin, A.E. Giblin, K.J. Nadelhoffer, W.C. Oechel and E.B. Rastetter, 1992. Global change and carbon balance of Arctic ecosystems, *BioScience* **42**: 433-441.
Shaw, G.E., 1995. The Arctic haze phenomenon. *Bull. Amer. Met. Soc.* **76**: 2403-2413.
Sheath, R.G., 1986. Seasonality of phytoplankton in northern tundra ponds. *Hydrobiologia* **138**: 75-83.
Shettle, E.P. and A.E.S. Green, 1974. Multiple scattering calculation of the middle ultraviolet reaching the ground. *Applied Optics* **13**: 1567-1581.
Shine, K.P., A. Henderson-Sellers and R.G. Barry, 1984. Albedo-climate feedback: The importance of cloud and cryosphere variability. Chapter 4, pp. 135-155, *In*: A. Berger and C. Nicolis (eds.). New Perspectives in Climate Modeling, Developments in Atmospheric Sciences, 16, Elsevier.
Shine, K., R. Derwent, D. Wuebbles and J. Morcrette, 1990. Radiative forcing of climate. pp. 41-68, *In*: J. Houghton, G. Jenkins and J. Ephraums (eds.). Climate Change: the IPCC Scientific Assessment. Cambridge University Press, Cambridge, U.K.
Simmonds P.G., R.G. Derwent, A. McCulloch, S. O'Doherty and A. Gaudry, 1996. Long-term trends in concentrations of halocarbons and radiatively active trace gases in Atlantic and European air masses monitored at Mace Head, Ireland from 1987-1994. *Atmos. Environ.* **30(23)**: 4041-4050.
Skony, S.M., J.D. Kahl and N.A. Zaitseva, 1994. Differences between radiosonde and dropsonde temperature profiles over the Arctic Ocean. *J. Atmos. Ocean. Tech.* **11**: 1400-1408.
Sliney, H.D., 1986. Physical factors in cataractogenesis: Ambient ultraviolet radiation and temperature. *Investigative Ophthalmology and Visual Science* **27**: 781-790.
Sliney, H.D., 1994. Dosimetry for ultraviolet radiation exposure of the eye. *In*: D.H. Sliney and M. Belkin (eds.). Proc. Ultraviolet Radiation Hazards SPIE. The International Society for Optical Engineering, SPIE Vol. 2134B, 26-27 January 1994. SPIE, Los Angeles.
Smith, M.W., 1993. Climate change and permafrost: pp. 291-312, *In*: H.M. French and O. Slaymaker (eds.). Canada's Cold Environments. McGill-Queens University Press.
Smith, R.C. and K.S. Baker, 1979. Penetration of ultraviolet-B and biologically effective dose-rates in natural waters. *Photochem. and Photobiol.* **29**: 311-323.
Smith, R.C. and K. S. Baker, 1982. Assessment of the influence of enhanced ultraviolet-B on marine primary productivity. *In*: J. Calkins (ed.). The role of ultraviolet radiation in marine ecosystems. NATO conference series. IV, Marine sciences, 7. Plenum Press, New York. xvi + 724p.
Smith, R.C., K.S. Baker, O. Holm-Hansen, R. Olson, 1980. Photoinhibition of photosynthesis in natural waters. *Photochem. and Photobiol.* **31**: 585-592.
Smith, R.C., B.B. Prézelin, K.S. Baker, R.R. Bidigare, N.P. Boucher, T. Coley, D. Karentz, S. MacIntyre, H.A. Matlick, D. Menzies, M. Ondrusek Z. Wan and K.J. Waters, 1992. Ozone deletion: Ultraviolet radiation and phytoplankton biology in Antarctic waters. *Science* **255**: 952-959.
Smith, S.D., R.D. Muench and C. H. Pease, 1990. Polynyas and leads: An overview of physical processes and the environment. *J. Geophys Res.* **95**: 9461-9479.
Smith Jr., W.O. and J. M. Grebmeier (eds.), 1994. Arctic oceanography: marginal ice zones and continental shelves. Vol. 49 of Coastal and Estuarine Studies, AGU, Washington, D.C. 287p.
Smith Jr., W.O. and E. Sakshaug, 1990. Polar phytoplankton. pp. 477-526, *In*: W.O. Smith Jr. (ed.). Polar Oceanography, Part B, Chemistry, Biology and Geology. Academic Press, Inc., San Diego, California.
Smol, J.P., B.F. Cumming, M.S.V. Douglas and R. Pienitz, 1995. Inferring past climatic changes in Canada using paleolimnological techniques. *Geoscience Canada* **21**: 113-118.
Smol, J.P., I.R. Walker and P.R. Leavitt, 1991. Paleolimnology and hindcasting climatic trends. *Verh. Internat. Verein. Limnol.* **24**: 1240-1246.
Snyder, J.A., G.H. Miller, A. Werner, A.J.T. Jull and T.W. Stafford, 1994. AMS-radiocarbon dating of organic-poor lake sediments, an example from Linn Vatnet, Spitsbergen. *The Holocene* **4**: 413-421.
Solomatin, N., 1986. The petrogenisis of underground ice, Novosibirsk, USSR Academy of Sciences.
Solomon, S. and J.S. Daniel, 1996. Impact of the Montreal Protocol and its amendments on the rate of change of global radiative forcing. *Climatic Change* **32**: 7-17.
Sonesson, M., T.V. Callaghan and L.O. Björn, 1995. Short-term effects of enhanced ultraviolet-B and CO_2 on lichens at different latitudes. *Lichenology* **27**: 547-557.
SORG (U.K. Stratospheric Ozone Review Group), 1987. Stratospheric Ozone: the first report of SORG. HMSO, London.
Spencer, R.W. and J.R. Christy, 1992a. Precision and radiosonde validation of satellite gridpoint temperature anomalies, Part I. MSU channel 2. *J. Climate* **5**: 847-857.
Spencer, R.W. and J.R. Christy, 1992b. Precision and radiosonde validation of satellite gridpoint temperature anomalies, Part II. A tropospheric retrieval and tends during 1979-1990. *J. Climate* **5**: 847-857.
Spencer, R.W. and J.R. Christy, 1993. Precision lower stratospheric temperature monitoring with the MSU: technique, validation, and results 1979-1991. *J. Climate* **6**: 1194-1204.
Spinhirne, J.D. and A.E.S. Green, 1978. Calculation of the relative influence of cloud layers on received ultraviolet and integrated solar radiation. *Atmos. Environ.* **12**, 2449-2454.
Spunkuch, D. and E. Schulz, 1995. Diagnosing and forecasting total column ozone by statistical relations. *J. Geophys. Res.* **100**: 18873-18885.
Stamnes, K., 1982. Reflection and transmission by a vertically inhomogeneous planetary atmosphere. *Planet. Space Sci.* **30**: 727-732.
Stamnes, K., K. Henriksen and P. Ostensen, 1988. Simultaneous measurement of UV radiation received by the biosphere and total ozone amount. *Geophys. Res. Lett.* **15**: 784-787.
Stamnes, K., Z. Jin, J. Slusser, C.R. Booth and T. Lucas, 1992. Several-fold enhancement of biologically effective ultraviolet radiation at McMurdo Station Antarctica during the 1990 ozone hole. *Geophys. Res. Lett.* **19**: 1013-1016.
Stamnes, K., J. Slusser and M. Brown, 1991. Derivation of total ozone abundance and cloud effects from spectral irradiance measurements. *Applied Optics* **30**: 4418-4425.
Stamnes, K., J. Slusser, M. Bowen, C. Booth and T. Lucas, 1990. Biologically effective ultraviolet radiation, total ozone abundance and cloud optical depth at McMurdo Station, Antarctica, September 15, 1988 through April 15, 1989. *Geophys. Res. Lett.* **17**: 2181-2184.
Starfield, A.M. and F.S. Chapin, 1996. Model of transient changes in arctic and boreal vegetation in response to climate and land use change. *Ecological Applications* **6**: 842-864.
Steele, L.P., E.J. Dlugokencky, P.M. Lang, P.P. Tans, R.C. Martin and K.A. Masarie, 1992. Slowing down of the global accumulation of atmospheric methane during the 1980s. *Nature* **358**: 313-316.
Steele, L.P., P.J. Fraser, R.A. Rasmussen, M.A. K. Khalil, T.H. Conway, A.J. Crawford, R.H. Gammon, K.A. Masarie and K.W. Thoning, 1987. The global distribution of methane in the troposphere. *J. of Atmospheric Chemistry* **5**: 125-171.
Steffen, K.S. and J. Heinrichs, 1994. Feasibility of sea ice typing with synthetic aperture radar (SAR): Merging of Landsat thematic mapper and ERS-1 SAR satellite imagery. *J. Geophys. Res.* **99(C11)**: 413-424.
Steffen, K.S. and A. Schweiger, 1990. Multisensor approach to sea ice classification for Validation of SSM/I Sea Ice estimates. *J. Photogrammetric Eng.* **56(1)**: 75-82.
Stigebrandt, A., 1981. A model for the thickness and salinity of the upper layer of the Arctic Ocean and the relationship between the ice thickness and some external parameters. *J. Phys. Oceanogr.* **11**: 1407-1422.
Stirling, I.M., 1988. Attraction of polar bears (*Ursus maritimus*) to offshore drilling sites in the eastern Beaufort Sea. *Polar Record* **24**: 1-8.
Stirling, I.M. and W. Calvert, 1983. Environmental threats to marine mammals in the Canadian Arctic. *Polar Record* **21**: 433-449.
Stirling, I.M. and A.E. Derocher, 1993. Possible impacts of climate warming on polar bears. *Arctic* **46**: 240-245.
Stirling, I.M., M. Kingsley and W. Calvert, 1982. The distribution and abundance of seals in the eastern Beaufort Sea, 1974-69. Canadian

Wildlife Service, Occasional Paper 47: 23.
Strathdee A.T., J.S. Bale, F.C. Strathdee, W.C. Block, S.J. Coulson, N.R. Webb and I.D. Hodkinson, 1995. Climatic severity and the response to temperature elevation of Arctic aphids. *Global Change Biology* **1**: 23-28.
Stuiver, M., P.M. Grootes and T.F. Braziunas, 1995. The GISP2 delta 18O climate record of the past 16,500 years and the role of the sun, ocean and volcanoes. *Quaternary Research* **44**: 341-354.
Sturges, W.T., R.C. Schnell, S. Landsberger, S.J. Oltmans, J.M. Harris and S.-M. Li, 1993. Chemical and meteorological influences on surface ozone destruction at Barrow, Alaska, during spring 1989. *Atmos. Environ.* **27A**: 2851-2863
Sutton, R.T., H. Maclean, R Swinbank, A. O'Neill and F.W. Taylor, 1994. High-resolution stratospheric tracer fields estimated from satellite observations using Lagrangian trajectory calculations. *J. Atm. Sci.* **51**: 2995-3005.
Sveinbjörnsson, B., O. Nordell and H. Kauhanen, 1992, Nutrient relations of mountain birch growth at and below the elevation tree line in Swedish lapland. *Functional Ecology* **6(2)**: 213.
Taalas, P., 1993. Factors affecting the behavior of tropospheric and stratospheric ozone in the European Arctic and in Antarctica. Report No. 10, Finnish Meteorological Institute Contributions. Finnish Meteorological Institute, Helsinki, Finland.
Taalas, P., and E. Kyrö, 1994. Stratospheric winter of 1991/92 at Sodankyla in the European Arctic as compared with 1965-92 meteorological and 1988-91 ozone sounding statistics. *Geophys. Res. Lett.* **21(13)**: 1207-1210.
Taalas, P., J. Damski, A. Korpela, T. Koskela, E. Kyro and G. Braathen, 1995. Connections between atmospheric ozone, the climate system and ultraviolet-B radiation in the Arctic. pp. 411-426, *In*: W.C. Wang and I.S.A. Isaksen (eds.). Ozone as a climate gas. NATO ASI Series 32. Springer-Verlag, Heidelberg.
Taalas, P., J. Damski, E. Kyro, M. Ginzburg and G. Talamoni, 1997. The effect of stratospheric ozone variations on UV radiation and on tropospheric ozone at high latitudes. *J. Geophys. Res.* **102(D1)**: 1533-1543.
Taalas, P., E. Kyrö, K. Jokela, T. Koskela, J. Damski, M. Rummukainen, K. Leszczynski and A. Supperi, 1996. Stratospheric ozone depletion and its impact on UV radiation and on human health. *Geophysica* **32**: 127-165.
Takeuchi, Y., S. Ikeda and H. Kasahara, 1993. Dependence on wavelength and temperature of growth inhibition induced by ultraviolet-B irradiation. *Plant Cell Physiology* **34**: 913-917.
Tao, X., J. E. Walsh and W. L. Chapman, 1996. An assessment of global climate model simulations of Arctic air temperatures. *J. Climate* **9**: 1060-1076.
Tarasick, D.W., D.I. Wardle, J.B. Kerr, J.J. Bellefleur and J. Davies, 1995. Tropospheric ozone trends over Canada: 1980-1993. *Geophys. Res. Lett.* **22**: 409-412.
Taylor, A.E., 1991. Holocene paleoenvironmental reconstruction from deep ground temperatures: a comparison with paleoclimate derived from 18O record from an ice core from the Agassiz Ice Cap, Canadian Arctic Archipelago. *J. of Glaciology* **37**: 209-219.
Taylor, K.C., P.A. Mayewski, M.S. Twickler and S.I. Whitlow, 1996. Biomass burning recorded in the GISP2 ice core: A record from eastern Canada? *The Holocene* **6**: 1-6.
Tenow, O. and B. Holmgren, 1987. Low winter temperatures and an outbreak of Epirrita autumnata along a valley of Finnmarksvidda, the "cold-pole" of northern Fennoscandia. pp. 203-216, *In*: H Axelsson and B Holmgren (eds.). Climatological extremes in the mountain, physical background, geomorphological and ecological consequences. Department of Physical Geography of the University of Uppsala, UNGI Rapport 15.
Thomas, R.H., 1990. Polar research from satellites. Joint Oceanographic Institutions, Inc., Washington, D.C. 91p.
Thompson, A.M., 1984. The effect of clouds on photolysis rates and ozone formation in the unpolluted troposphere. *J. Geophys. Res.* **89**: 1341-1349.
Thomson, B., 1986. Is the impact of ultraviolet-B radiation on marine zooplankton of any significance? *In*: J.G. Titus (ed.). Effects of changes in stratospheric ozone and global climate: Stratospheric ozone 2: 203.
Thorndike, A.S. and R. Colony, 1982. Sea ice motion in response to geostrophic winds. *J. Geophys. Res.* **87**: 5845-5852.
Thorndike, A.S., C. Parkinson and D.A. Rothrock (eds.), 1992. Report of the sea ice thickness workshop, 19-21 November, 1991.
Tranvik, L., 1988. Availability of dissolved organic carbon for planktonic bacteria in oligotrophic lakes of differing humic content. *Microbial Ecology* **16**: 311-322.
Treshnikov, A.F., 1985. Arctic Atlas. Arkt.-Antarkt. Naucho-Issled. Inst. Moscow. 204p. (In Russian).
Trocine, R.P., J.D. Rice and G.N. Wells, 1981. Inhibition of seagrass photosynthesis by ultraviolet-B radiation. *Plant Physiol.* **68**: 74-81.
Trodahl, H.J. and R.G. Buckley, 1990. Enhanced ultraviolet transmission of Antarctic sea ice during the austral spring. *Geophys. Res. Lett.* **17**: 2177-2179.
Tsay, S.-C. and K. Stamnes, 1992. Ultraviolet radiation in the Arctic: The impact of potential ozone depletions and cloud effects. *J. Geophys. Res.* **97**: 7829-7840.
Tsitas, S.R. and Y.L. Yung, 1996. The effect of volcanic aerosols on ultraviolet radiation in Antarctica. *Geophys. Res. Lett.* **23(2)**: p. 157.
Tung, K.K. and H. Yang, 1994. Global QBO in circulation and ozone, Part 1: Reexamination of observational evidence. *J. Atmos. Sci.* **51**: 2699-2707.
Tyagi, R., G. Srinivas, D. Vyas, A. Kumar and H.D. Kumar, 1992. Differential effect of ultraviolet-B radiation on certain metabolic processes in a chromatically adapting Nostoc. *Photochem. and Photobiol.* **55**: 401-407.
UNEP, 1991. Environmental effects of stratospheric ozone depletion – 1991 update. J.C. van der Leuen, M. Tevini and R.C. Worrest (eds.). United Nations Environmental Programme (UNEP), Nairobi, Kenya.
UNEP, 1994. Environmental effects of ozone depletion - 1994 update. *Ambio* **24**: 137-196.
UNEP, 1994a. Handbook for the Montreal Protocol on substances that deplete the ozone layer. Ozone Secretariat of the United Nations Environment Programme, Nairobi, Kenya.
UNEP, 1995. Environmental effects of ozone depletion: 1994 assessment. *Ambio* **24(3)**, May 1995. 198p.
Valero, F.P.J., T.P. Ackerman and W.J.Y. Gore, 1988. Radiation studies in the Arctic. Aerosol and Climate. P.V. Hobbs and M.P. McCormick (eds.). Deepak Publishing, Hampton, Virginia, 271-275.
Van Donk, E. and D.O. Hessen, 1995. Reduced digestability of ultraviolet-B stressed and nutrient-limited algae by *Daphnia magna*. *Hydrobiologia* **307**: 147-151.
Van Donk, E. and D.O. Hessen, 1996. Loss of flagella in the green alga *Chlamydomonas reinhardtii* due to in situ UV-exposure. *Scientia Marina* **60**: 107-112.
Velichko A.A., O.K. Borisova, E.M. Zelikson and V.P. Nechaev, 1990. An assessment of dynamics of natural geosystems in the forest and tundra zones under anthropogenic climatic change. Quoted in: W.J. McG Tegard, T.W. Sheldon and D.C. Griffiths (eds.). Climate change: The IPCC Impacts Assessment. Ch. 3. Australian Government Publishing Service, Canberra.
Veres, A., R. Pienitz and J.P. Smol, 1995. Lake water salinity and periphytic diatom succession in three subarctic lakes, Yukon Territory, Canada. *Arctic* **48**: 63-70.
Vetter, R.D., 1996. Ultraviolet-B irradiance and DNA damage in pelagic fish eggs and larvae: temperate and polar habitats. *EOS (American Geophysical Union)* **76(3)**: 164.
Vibe, C., 1967. Arctic animals in relation to climatic fluctuations. Meddelelser om Grønland, Bd. 170: 5.
Vincent, W.F. and S. Roy, 1993. Solar ultraviolet-B radiation and aquatic primary production: damage, protection and recovery. *Environ. Rev.* **1**: 1-12.
Vinje, T., K.B.E. Axelsson, L.A. Fransson (eds.), 1989. An upward looking sonar ice draft series. POAC 89. The 10th International Conference on Port and Ocean Engineering under Arctic Conditions. June 12-16, Lulea, Sweden. Proceedings; Tekniska Högskolan i Lulea: 178-187.
Vosjan, J.H., G. Dohler and G. Nieuwland, 1990. Effects of ultraviolet-B irradiance on the ATP content of microorganisms of the Weddel Sea (Antarctica). *Netherlands J. of Sea Res.* **25**: 391-393.
Vowinckel. E. and S. Orvig, 1962. Insolation and absorbed solar radiation at the ground in the Arctic. Publications in Meterology, No. 53. McGill University, Montreal. 32p.
Vowinckel, E. and S. Orvig, 1970. The climate in the north polar basin. pp. 129-252, *In*: Climate of the Polar Regions, World Survey of Climatology, 14. Elsevier, Amsterdam.
Wadhams, P., 1990. Evidence for thinning of the Arctic ice cover north of Greenland. *Nature* **345**: 795-797.
Wadhams, P., N.R. Davis, J.C. Comiso, R. Kutz, J. Crawford, G. Jackson, W. Krabill, C.B. Sear, R. Swift and W.B. Tucker III., 1991. Current remote sensing of Arctic sea ice from submarine and aircraft. *Int. J. Rem. Sens.* **12**: 1829-1840.
Wallace, D.W., R.M. Moore and E.P. Jones, 1987. Ventilation of the Arctic Ocean cold halocline: rates of diapycnal and isopycnal transport, oxygen utilization and primary production inferred using chlorofluoromethane distribution. *Deep-Sea Res.* **34**: 1957-1980.
Walsh, J.E., 1991. The Arctic as a bellwether. *Nature* **352**: 19-20.
Walsh, J.E. and W.L. Chapman, 1990. Short-term climatic variability of the Arctic. *J. Climate* **3**: 237-250.
Walsh, J.E. and C.M. Johnson, 1979. Interannual atmospheric variability and associated fluctuations in Arctic sea ice. *J. Geophys. Res.* **84**: 6915-6928.
Walsh, J.E. and J.E. Sater, 1981. Monthly and seasonal variability in the ocean-ice-atmosphere systems of the North Pacific and the North Atlantic. *J. of Geophys. Res.* **86(C8)**: 7425-7445.
Walsh, J.E., W.D. Hibler III and B. Ross, 1985. Numerical simulation of Northern Hemisphere sea ice variability, 1951-1980. *J. Geophys. Res.* (Oceans) **90**: 4847-4865.
Walsh, J.E., R.S. Weaver, R. Colony, A.S. McLaren and R.H. Bourke, 1995. Correlation of ice thickness variability to atmospheric circulation over the north pole 1958-1992. Fourth Conference on Polar Meteorology and Oceanography 15-20 Jan. 1995. Dallas, TX. AMS. 156-161.
Walsh, J.J., 1995. DOC storage in Arctic Seas: The role of continental shelves. pp. 203-230, *In*: W.O. Smith Jr. and J.M. Grebmeier (eds.). Arctic Oceanography: Marginal Ice Zones and Continental Shelves. Coastal and Estuarine Studies 49.
Walter, B.A. and J.E. Overland, 1993. The response of lead patterns in the Beaufort Sea to storm-scale wind forcing. *Ann. Glaciol.* **17**: 219-226.
Wang, P., 1986. Preliminary chemical study on snow and ice in mountain glaciers of China. *J. of Glaciology and Geocryology* **8(1)**: 40-51.
Wängberg, S.-A., J.S. Selmer and K. Gustavson, 1996. Effects of ultraviolet-B radiation on biomass and composition in marine phytoplankton

communities. pp. 81-88, *In*: F.L. Figueroa, C. Jiménez, J.L. Pérez-Llorens and F.X. Niell (eds.). Underwater light and agal photobiology. *Sci. Mar.* **60** (**Supl. 1**).
Warren, S.G., C.J. Hahn, J. London, M. Chervin and R. Jenne, 1988. Global distribution of total cloud cover and cloud type amounts over the ocean. NCAR Tech. Note TN-317+STR. Boulder, CO. 29p.
Warren, S.G., C.J. Hahn, J. London, R.M. Chervin and R. Jenne, 1986. Global distribution of total cloud cover and cloud type amounts over land, NCAR Tech. Note TN-273+STR. Boulder, CO. 29p.
Washburn, A.L., 1989. Hear-surface soil displacement in sorted circles, Resolute area, Cornwallis Island. *Can. J. of Earth Sciences.* **26**: 941-55.
Waugh, D.W., R.A. Plumb, P.A. Newman, M.R. Schoeberl, L.R. Lait, M. Loewenstein, J.R. Podolske, J.W. Elkins and K.R. Chan, 1994. Fine-scale, poleward transport of tropical air during AASE2. *Geophys. Res. Lett.* **21**: 2603-2606.
Weaver, R., C. Morris and R.G. Barry, 1987. Passive microwave data for snow and ice research: Planned products from the DMSP SSM/I system. *EOS (American Geophysical Union)* **68**: 39.
Weeks, W.F. and S.F. Ackley, 1982. The growth, structure and properties of sea ice. CRREL Monograph 82-1.
Welch, H.E., J.A. Legault and H.J. Kling, 1989. Phytoplankton nutrients and primary production in fertilized and natural lakes at Saqvaqjuac, N.W.T. *Can. J. Fish. Aquat. Sci.* **46**: 90-107.
Wester, U., 1996. UV monitoring in Sweden, past present and future. *In*: B.L. Diffey (ed.). Measurement and trends of terrestrial ultraviolet-B radiation in Europe. OEMF, Milan.
Wetzel, R.G., Hatcher, P.G. and T.S. Bianchi, 1995. Natural photolysis by ultraviolet irradiance of recalcitrant dissolved organic matter to simple substrates for rapid bacterial metabolism. *Limnol. Oceanogr.* **40**: 1369-1380.
Whalen, S.C. and W.S. Reeburgh, 1990a. A methane flux transect along the Trans-Alaska Pipeline Haul Road. *Tellus* **42B**: 237-249.
Whalen, S.C. and W.S. Reeburgh, 1990b. Consumption of atmospheric methane by tundra soils. *Nature* **346**: 160-162.
Whalen, S.C. and W.S. Reeburgh, 1992. Interannual variations in tundra methane emission: A 4-year time-series at fixed sites. *Global Biogeochemical Cycles* **6**: 139- 159.
Wheeler, P.A., E. Sherr, M. Gosselin, D. Thibault, D.L. Kirchman, R. Benner and T.E. Whitledge, 1996. Biological production in the Central Arctic Ocean: An overview of results from the 1994 Arctic Ocean Section. *EOS (American Geophysical Union)* **76**(**3**): 59.
WHO, 1994. Environmental Health Criteria # 160: Ultraviolet Radiation. World Health Organization, Geneva.
Wigley, T.M.L., 1989. Possible climate change due to SO_2-derived cloud condensation nuclei. *Nature* **339**: 365-367.
Wigley, T.M.L. and S.C.B. Raper, 1992. Implications for climate and sea level of revised IPCC emission scenarios. *Nature* **357**: 293-300.
Williams, P.J. and M.W. Smith, 1989. The frozen Earth: Fundamentals of geocryology. Cambridge University Press, Cambridge, U.K.
Williamson, C.E. and H.E. Zagarese (eds.), 1994. Proceedings of a special workshop on the impact of ultraviolet-B radiation on pelagic freshwater ecosystems. *Archiv Fur Hydrobiologie* **43**: 1-226.
Williamson, C.E., R.S. Stemberger, D.P. Morris, T.M. Frost, and S.G. Paulsen, 1996. Ultraviolet radiation in North American lakes: Attenuation estimates from DOC measurements and implications for plankton communities. *Limnol. Oceanogr.* **41**: 1024-1034.
Willmott, C.J. and D.R. Legates, 1991. Rising estimates of terrestrial and global precipitation. *Climate Res.* **1**: 179-86.
Willmott, C.J., S.M. Robeson and J.J. Feddema, 1994. Estimating continental and terrestrial precipitation averages from rain-gauge networks. *Int. J. of Climatology* **14**.
WMO, 1988. Report of the international ozone trends panel, 1988. World Meteorological Organization Global Ozone Research and Monitoring Project, Report No. 18, Volume I. World Meteorological Organization, Geneva.
WMO, 1991. World Meteorological Organization, Scientific assessment of ozone depletion: 1991, Report 25. World Meteorological Organization, Geneva.
WMO, 1992. E. Raschke, H. Cattle, P. Lemke and W. Rossow (eds.). Report on the workshop on polar radiation fluxes and sea-ice modelling. World Climate Research Programme, WMO/TD No. 442, Bremerhaven, Germany, 5-8 November. 20p.
WMO, 1992. Scientific assessment of ozone depletion: 1991, World Meteorological Organization Global Ozone Research and Monitoring Project, Report No. 23. World Meteorological Organization, Geneva.
WMO, 1994. World Meteorological Organization, Scientific assessment of ozone depletion: 1991, Report 33. World Meteorological Organization, Geneva.
WMO, 1995. Scientific assessment of ozone depletion: 1994. World Meteorological Organization, Global Ozone Research and Monitoring Project, Report No. 37. World Meteorological Organization, Geneva.
Wolfe, A.P., 1994. Late Wisconsin and Holocene diatom stratigraphy from Amarok Lake, Baffin Island, N.W.T., Canada. *Journal of Paleolimnology* **10**: 129-139.
Wolfe, A.P., 1996a. Spatial patterns of modern diatom distribution and multiple paleolimnological records from a small non-glacial Arctic lake, Baffin Island, Northwest Territories. *Can. J. of Botany* **74**: 435-449.
Wolfe, A.P., 1996b. A high resolution late-glacial and early Holocene diatom record from Baffin Island, Northwest Territories. *Can. J. of Earth Sciences* **33**: 928-937.
Woo, M.K. and P. Marsh, 1990. Response of soil moisture change to hydrological processes in a continuous permafrost environment. *Nordic Hydrology* **21**: 235-252.
Woo, M.K. and P. Steer, 1983. Slope hydrology as influenced by thawing of the active layer, Resolute, N.W.T. *Can. J. of Earth Sciences* **29**: 58-68.
Woo, M.K., A. Walker, D. Yang and B.E. Goodison, 1995. Pixel-scale ground snow survey for passive microwave study of the Arctic snow cover. pp. 51-57, *In*: Proc. Eastern Snow Conference, 52nd Meeting, Toronto, June 8-10, 1995.
Wood, W.F., 1987. Effect of solar ultra-violet radiation on the kelp Ecklonia radiata. *Mar. Biol.* **96**: 143-150.
Woodbridge, E.L., J.W. Elkins, D.W. Fahey, L.E. Heidt, S. Solomon, T.J. Baring, T.M. Gilpin, W.H. Pollock, S.M. Schauffler, E.L. Atlas, M. Loewenstein, J.R. Podolske, C.R. Webster, R.D. May, J.M. Gilligan, S.A. Motzka, K.A. Boering and R.J. Salawitch, 1995. Estimates of total organic and inorganic chlorine in the lower stratosphere from in situ and flask measurements during AASE. *J. Geophys. Res.* **100**: 3057-3064.
Worl, R., 1980. The North Slope Iñupiat whaling complex. *In*: Y. Kotani and W.B. Workman (eds.). Alaskan native culture and history. National Museum of Ethnology, Senri Ethnological Studies 4. Osaka, Japan.
Worrest, R.C., B.E. Thomson, and H. van Dyke, 1981. Impact of ultraviolet-B radiation upon estuarine microcosms. *Photochem. and Photobiol.* **33**: 861-867.
Wulf, H.C., 1994. Effects of ultraviolet radiation from the Sun on the Inuit population. *Arctic Medical Research* **53**(**2**): 416-422.
Wulf, H.C., Poulsen, T. Davies, R.E., Urbach, F., 1989. Narrow band UV radiation and induction of dermal elastosis and skin cancer. *Photodermatol.* **6**: 44-51.
WWB, 1974. World Water Balance and Water Resources of the Earth (WWB). 1974-1978: Leningrad, Gidrometeoizdat.
Yin, Y. and S.J. Cohen, 1994. Identifying regional policy concerns associated with global climate change. *Global Environmental Change* **4**: 246-260.
Zeng, J., Z. Jin, and K. Stamnes, 1993. Impact of stratospheric ozone depletion on UV penetration into the ocean at high latitudes. pp. 56-53, *In*: H.C. Eilertsen (ed.). Underwater Light Measurements, Proc. SPIE 2048.
Zurek, R.W., G.L. Manney, A.J. Miller, M.E. Gelman and R.M. Nagatani, 1996. Interannual variability of the north polar vortex in the lower stratosphere during the UARS mission. *Geophys. Res. Lett.* **23**: 289-292.
Zwally, H.J., 1989. Growth of Greenland ice sheet: interpretations. *Science* **246**: 1589-1591.
Zwally, H.J., 1995. Observing changes in polar ice. Wadati Conf. on Global Change and the Polar Climate, Tsukuba Science City, Japan, 7-10 Nov. 1995. Geophysical Institute, University of Alaska, Fairbanks. 22-24.