# The Polar Regions and Climatic Change: Appendix (1984)

Pages
127

Size
5 x 8

ISBN
0309324548

Committee on the Role of the Polar Regions in Climatic Change; Polar Research Board; Commission on Physical Sciences, Mathematics, and Resources; National Research Council

# The Polar Regions and Climatic Change
## Appendix

Committee on the Role of the Polar Regions in Climatic Change
Polar Research Board
Commission on Physical Sciences, Mathematics, and Resources
National Research Council

NATIONAL ACADEMY PRESS
Washington, D.C. 1984

NOTICE: The project that is the subject of this report was approved by the Governing Board of the National Research Council, whose members are drawn from the councils of the National Academy of Sciences, the National Academy of Engineering, and the Institute of Medicine. The members of the committee responsible for the report were chosen for their special competences and with regard for appropriate balance.

This report has been reviewed by a group other than the authors according to procedures approved by a Report Review Committee consisting of members of the National Academy of Sciences, the National Academy of Engineering, and the Institute of Medicine.

The National Research Council was established by the National Academy of Sciences in 1916 to associate the broad community of science and technology with the Academy's purposes of furthering knowledge and of advising the federal government. The Council operates in accordance with general policies determined by the Academy under the authority of its congressional charter of 1863, which establishes the Academy as a private, nonprofit, self-governing membership corporation. The Council has become the principal operating agency of both the National Academy of Sciences and the National Academy of Engineering in the conduct of their services to the government, the public, and the scientific and engineering communities. It is administered jointly by both Academies and the Institute of Medicine. The National Academy of Engineering and the Institute of Medicine were established in 1964 and 1970, respectively, under the charter of the National Academy of Sciences.

Preparation of this report was made possible by a grant from the Andrew W. Mellon Foundation to the Polar Research Board and by the continuing support provided to the Board by the Office of Naval Research, the National Oceanic and Atmospheric Administration, and the Department of Energy.

Copies available in limited quantity from
Polar Research Board
2101 Constitution Avenue, N.W.
Washington, D.C. 20418

# Committee on the Role of the Polar Regions in Climatic Change

# Polar Research Board

Charles R. Bentley, Geophysical and Polar Research Center, University of Wisconsin, Chairman
W. Lawrence Gates, Department of Atmospheric Sciences, Oregon State University
Ben C. Gerwick, Jr., Department of Civil Engineering, University of California, Berkeley
Richard M. Goody, Division of Engineering and Applied Physics, Harvard University
Arnold L. Gordon, Lamont-Doherty Geological Observatory, Columbia University
Hans O. Jahns, EXXON Production Research Company, Houston
Philip L. Johnson, John E. Gray Institute, Lamar University
Arthur H. Lachenbruch, U.S. Geological Survey, Menlo Park
Louis J. Lanzerotti, AT&T Bell Laboratories
Chester M. Pierce, Harvard University
Juan G. Roederer, Geophysical Institute, University of Alaska
Robert H. Rutford, University of Texas at Dallas
John H. Steele, Woods Hole Oceanographic Institution
Ian Stirling, Canadian Wildlife Service, Edmonton, Alberta
Cornelius W. Sullivan, Department of Biological Sciences, University of Southern California

Ex Officio

Jerry Brown, Committee on Permafrost, Chairman
Mark F. Meier, Committee on Glaciology, Chairman
James H. Zumberge, U.S. Delegate to Scientific Committee on Antarctic Research, International Council of Scientific Unions

Agency Liaison Representatives

Thomas J. Gross, $CO_2$ Research Program, Department of Energy

G. Leonard Johnson, Arctic Programs, Office of Naval Research

Ned A. Ostenso, National Oceanic and Atmospheric Administration

Edward P. Todd, Division of Polar Programs, National Science Foundation

Staff

W. Timothy Hushen, Staff Director
Bertita E. Compton, Staff Officer
Muriel Dodd, Administrative Assistant

## Commission on Physical Sciences, Mathematics, and Resources

Howard E. Simmons, Jr., Central Research and Development Department, E. I. du Pont de Nemours & Company, Inc.
John D. Spengler, Department of Environmental Health Sciences, Harvard School of Public Health
Hatten S. Yoder, Jr., Geophysical Laboratory, Carnegie Institution of Washington

Raphael G. Kasper, Executive Director
Lawrence E. McCray, Associate Executive Director

# Preface

The climate of the Earth is perpetually variable and changing. Uncertainties about future climate, compounded by concerns that the accumulation of carbon dioxide in the atmosphere and other by-products of human activities may bring about inadvertent changes of climate in future decades, place a premium on improved understanding and prediction of climate to meet the needs of modern society. A major goal of climate research now in progress under the aegis of many national and international institutions is to develop a comprehensive theory of global climate as a first step to a rational climate prediction capability. Significant progress toward this goal has been achieved in the past decade.

In connection with global climate concerns, the polar regions of the Earth have become the focus of considerable attention in view of three distinct yet related circumstances:

- Growing evidence that the polar regions play a key role in the physical processes responsible for global climatic fluctuations and in some circumstances may be a prime mover of such fluctuations;
- Widening appreciation of the polar regions as a natural repository of information about past climatic fluctuations and about past Earth-environmental events causally related to climatic fluctuations; and
- Mounting concerns that future changes of climate, such as the general global-scale warming believed likely to result in future decades and centuries from the accumulation of carbon dioxide and other pollutants in the atmosphere, may disturb the equilibrium of polar ice masses and hence global sea levels.

These several matters clearly merit intensive parallel study based on separate research agendas. At the same time, however, the Polar Research Board of the National Research Council (NRC) has perceived a need to address them together in the context of a holistic view of planetary-scale climate, from which perspective the global interconnectedness of climate processes and their social implications is both the unifying theme and the key to setting overall goals and priorities in climate-related polar research. This perspective and its implications for future research are the focus of the report, to which this volume is an appendix.

The report was prepared by the Committee on the Role of the Polar Regions in Climatic Change, which was established by the Polar Research Board in 1980 to undertake a study in the Board's new "Polar Research - A Strategy" series. The committee's overall findings, conclusions, and recommendations are published in a separate volume. The volume in hand consists of three signed appendixes that together provide much of the background material and documentation considered by the committee in framing those conclusions and recommendations.

# Contents

# Appendix A.
# The Role of Polar Regions in Climate Dynamics

J. O. Fletcher and U. Radok

## 1. INTERNAL FACTORS AFFECTING CLIMATIC VARIATIONS

In the discussion of climatic variations it is often said that the ocean is the flywheel or the memory of the global climate system. Most climate planning documents include such a statement, but none ventures to explain how this memory works. For most of us it is a declaration of faith; we believe in it because it is logical.

We observe that climatic variations in a particular region are associated with changes in the prevailing atmospheric circulation and that substantial changes in atmospheric circulation seem to be global in extent. We know that the global atmospheric circulation is a response to the global pattern of thermal forcing, represented on the largest scale by the temperature contrast between tropical and polar regions. Air and ocean water, the two working fluids of this global thermodynamic engine, share the task of moving heat poleward, each accomplishing about half of the total transport. Movement of ocean water is forced directly by surface wind stress and by heat and moisture exchanges, which affect the water-density field. For the atmosphere the largest changes in the thermal forcing pattern probably arise from changes in sea-surface temperature and surface heat exchange, which in high latitudes are strongly affected by the presence of sea ice.

The ocean and the atmosphere thus affect each other, but the response times of the two fluids are very different. The atmosphere completes most of its response to changed forcing in a few weeks; the ocean responds on a range of longer time scales. If the wind stress changes suddenly to a different value that it maintains thereafter, the ocean-surface field of motion will take a few months to adjust. The surface temperature field, which must reflect horizontal transport times, will adjust in a decade or so, but the vertical temperature structure, or heat storage, requires several decades. The salinity structure, which influences vertical stability, may require centuries to adjust. Since adjustments on all these time scales can result from a simple change in wind stress, we have good reason to say that the ocean is the memory of the coupled system and that it resembles a flywheel in that it tends to persist in a particular state once that state has been established.

The reaction times of different ice forms range from the survival time of individual snowflakes in the atmosphere (minutes or hours) to the slow changes of the largest polar ice sheets (tens or hundreds of thousands of years). Between these extremes, probably the most significant interaction of ice and snow (collectively referred to as the cryosphere) with the ocean-atmosphere system occurs on the intermediate reaction time scale of sea ice (months or years). But also to be considered are the much slower interactions between the oceans and the large ice shelves of Antarctica, as well as the possibility of large masses of ice being injected into the ocean by "surging" of a major ice sheet.

Factors internal to the atmosphere that importantly affect its heat balance and pattern of thermal forcing are cloudiness, water vapor, carbon dioxide ($CO_2$), ozone, and aerosols. All of these, except perhaps ozone, can be influenced by the more slowly changing ocean and cryosphere.

## 2. EXTERNAL FACTORS AFFECTING CLIMATIC VARIATIONS

The solar energy reaching the earth per unit area is summarized as a function of latitude and time of year in Figure 1a, for the top of the atmosphere, and in Figure 1b, for the earth's surface, assuming an average attenuation by the atmosphere.

The radiation totals of Figure 1a all change in direct proportion to the intrinsic total energy output of the sun, or the "solar constant." Actual variations of the solar constant, of the order of 0.1 percent, that are associated with the formation and decay of groups of sunspots have so far been established beyond doubt for the relatively short period of accurate measurements;

(a)

(b)

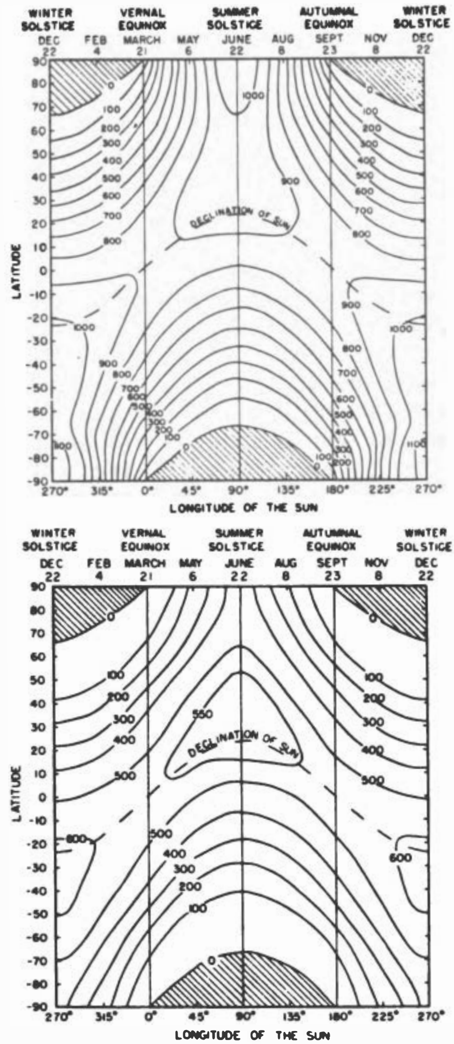


**Figure 1.** Daily radiation intake (cal/cm$^2$ d) (a) of the earth's surface in the absence of an atmosphere and (b) below an atmosphere with constant transmissivity of 0.7 (1 cal/cm$^2$ d = 0.486 W/m$^2$). (From Hess 1959.)

there is some indirect evidence that the solar emission may also have changed by a few tenths of a percent on time scales of the order of centuries. Such relatively small changes in the solar constant could be of significance to climate. Their effects, however, would be of the same order as those of quite modest changes in cloudiness or atmospheric particle loading. The effects of intensity variations in only a limited part of the solar spectrum such as the ultraviolet are even more difficult to assess and remain a subject of research and controversy under the heading of sun-weather relationships. For the purposes of this discussion, the "solar constant" is considered constant.

The distribution of solar radiation with latitude and with time of year is a different matter and changes slowly and systematically with time due to variations in the earth's orbit and in the inclination of the earth's axis of rotation to the orbital plane. The first quantitative treatment of these changes was given by Milankovitch (1941); an up-to-date reassessment has been undertaken by Berger (1978). For this discussion it suffices to note that over intervals of many centuries and millennia the orbital variations can change the solar energy reaching the earth by as much as 10 percent in key latitudes and seasons and thereby exert a significant influence on regional climatic characteristics, such as the length of the growing season and the number of days with temperatures above freezing. These so-called "Milankovitch climate changes," however, are slow and probably produce no climatic effects on time scales shorter than a few centuries.

The radiation reaching the surface (Figure 1b) undergoes additional changes inside the earth's atmosphere due to variations in water vapor, clouds, and aerosol loading. Changes in optically active trace gases ($H_2O$, $CO_2$, $O_3$, oxides of nitrogen or "$NO_x$," chlorofluorocarbons, etc.) affect climate both directly, by altering the clear-sky radiation, and indirectly, by modifying clouds, precipitation processes, and biotic systems (through their influence on the surface energy balance and on atmospheric composition).

Volcanic eruptions and heavy dust concentrations over deserts may give rise to significant short-term changes in climate, but the evidence for this is not yet conclusive (see, e.g., Miles & Gildersleeves 1978; Turco et al. 1982).

A fundamental difficulty in assessing the climatic effects of changing atmospheric constituents is that in most cases the trace gases and aerosols involved are modulated by both human activities and natural events. For example, episodes of atmospheric haze over the Arctic are thought to be produced jointly by dust from the Gobi Desert and particles from industrial activities in Asia, North America, and Europe. The nature of the aerosol can make a great deal of difference to the climate effects produced by a persistent change in aerosol concentration; but for a given ratio of absorption to backscattering, the effect will be to warm areas with a surface reflectance (albedo) above a certain threshold value. This value is exceeded throughout much of the Arctic (Kellogg et al. 1975). This finding implies a warming effect of most types of aerosol occurring in the polar regions. Past changes in atmospheric particle loading are recorded in the ice of glaciers and ice sheets and offer possibilities for separating recent industrial effects from the changes in the natural background (Thompson & Mosley-Thompson 1980).

Other problems arise in assessing the climatic effect of changes in atmospheric trace gases, notably atmospheric $CO_2$. Efforts to model the climatic effect of increasing concentration of $CO_2$ have not yet progressed beyond a limited range of physical realism; in particular, clouds and the ocean have not yet been incorporated in a manner permitting them to play their full modulating roles. Here again, the polar regions deserve special attention. For one thing, $CO_2$ model calculations suggest that surface temperatures should rise more in polar latitudes than elsewhere. For another, polar ice cores have recently provided the first evidence that atmospheric $CO_2$ concentrations were appreciably less than present values during the last glaciation, by as much as 30 to 40 percent, and possibly also somewhat greater than current values during the warm climate of from 6000 to 8000 years ago (Delmas et al. 1980). Finally, the surface exchanges related to formation and decay of sea ice are mechanisms draining $CO_2$ from the atmosphere into the bottom layer of the polar oceans (Budd 1980).

## 3. POLAR PROCESSES AND CLIMATIC VARIATIONS

Crucial questions on the relations between polar processes and climatic dynamics are the following:

- o How do the polar oceans and ice influence the behavior of the atmosphere?
- o What are the feedback mechanisms?
- o How do these mechanisms operate, and under what circumstances are they most influential?

On a regional basis, climatic unrest probably has many causes. On the global scale, climatic unrest may also have various ultimate causes, but in each case it strongly reflects a change in the thermal gradient between the tropics and the polar regions. To such a change, the tropics contribute primarily variations in precipitation, cloudiness, and ocean surface temperature, whereas polar effects include also changes in boundary-layer structure, in the extent of the highly reflective snow cover, and in the fluxes of heat, moisture, and momentum through sea ice, a major dynamic barrier between the polar oceans and the atmosphere.

Climatic anomalies are associated with changes in the global pattern of thermal forcing. That forcing is dominated by the surface heat balance, which is determined by the solar radiation absorbed and reflected, by the infrared emission to space from the atmosphere and surface, and by the transport of heat by atmosphere and ocean circulations. The manner in which the surface heat balance can be disturbed is somewhat different in the two polar regions.

### 3.1 Arctic

The energy balance of the Arctic Basin is dominated by the radiation components, summarized in Figure 2. The existence of a sea-ice cover (about 3 m thick) effectively insulates the heat reservoir of the Arctic Ocean from the atmosphere. The net annual radiation balance at the surface is small in comparison to the individual heat budget components. The large amount of incoming shortwave radiation during the long polar summer days (100 $W/m^2$) suffices, even after two thirds of it has been reflected, to melt more than 1 cm of sea ice per day. The melting is accelerated in midsummer by the sudden lowering of

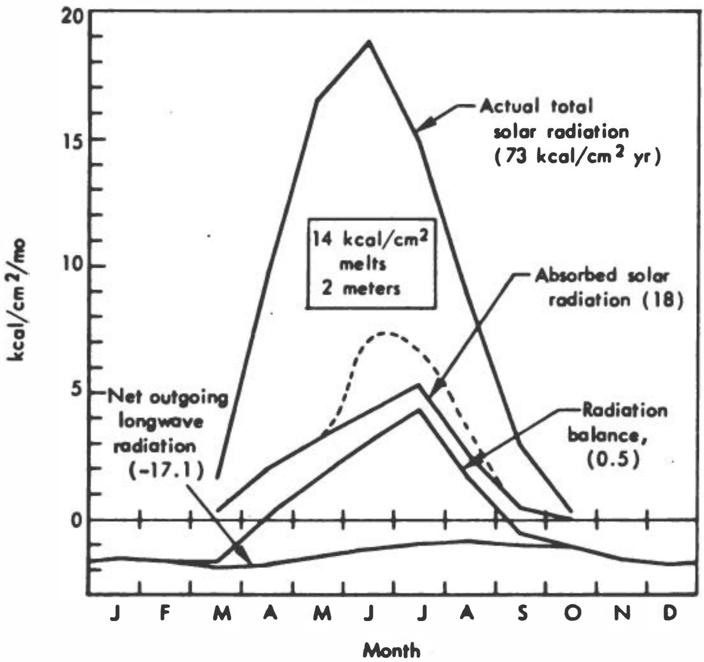


Figure 2. Radiation components at the surface in the central Arctic (from Fletcher 1965).

albedo, which accompanies collapse of the snow cover and puddling on the ice. Thus, relatively small anomalies in the radiation budget can have large effects on the ice, and interannual variations in the date of puddling exert a great leverage on the radiation budget. The same applies to turbulent heat fluxes, although in the atmospheric boundary layer of the central Arctic these are normally between 1 and 2 orders of magnitude smaller than individual radiative fluxes (Doronin 1963; Fletcher 1965).

A major complication arises from stresses applied to the ice by the wind and the water, which cause the ice to crack and part in places. New ice forms in the open "leads," resulting in an ice matrix with a wide range of different thicknesses, each having a different surface temperature and hence different heat balance. During winter, initial heat loss from open water exceeds that from perennial ice by 2 orders of magnitude (Badgley 1966). During summer, the heat balance of the surface of the pack ice is dominated by the persistent and extensive stratus clouds that cover the Arctic Basin. These cloud layers interact with the streams of solar and terrestrial radiation (Herman & Goody 1976) and ultimately affect the rate of melting at the surface.

## 3.2 Antarctic

In contrast, the Antarctic contains a high ice plateau surrounded by an enormous area of seasonal sea ice. Over the plateau, dense clouds are infrequent, and surface and free-air temperatures are considerably lower than in the Arctic. The surface energy balance of the Antarctic region is shown in Figure 3. The principal heat loss to space comes from the cloudy regions around the continent and from the free atmosphere over the high plateau where the cooling produces a large-scale direct thermal circulation (Figure 4). The thermal gradient between Antarctica and the tropics forces the Southern Hemisphere westerlies, which maintain great strength during both summer and winter. The surface wind regime over the ice sheet is forced by the intense radiational cooling, which greatly reinforces the surface temperature inversion that is characteristic of the polar regions in general. In the inversion layer above the sloping ice-sheet surface, the slope-parallel component of the buoyancy force gives rise to a "katabatic" pressure gradient (Ball 1960). Even where slopes are as gradual as $10^{-3}$, this gradient is comparable in magnitude to the pressure gradients of synoptic weather systems and creates a very steady regime of surface winds crossing the elevation contours at a small angle (Figure 5). On the steep slopes near the edges of the ice sheet the direction of the katabatic flow approaches that of the fall line, and the flow becomes intermittent, due in part to instability mechanisms (Ball 1956) and in part to the fact that the surface cooling is inadequate to renew the supply of cold air as rapidly as it is drained away (Schwerdtfeger 1970). However, in some regions of Antarctica the ice-sheet topography is such as to produce regular jet systems, which have been observed to maintain, without interruption, wind speeds of 30 m/s or more for an entire month (Loewe 1956).

Even in regions with less extreme conditions, the katabatic winds give the Antarctic climate its unique character. The precise role in the surface energy and mass balance of the ice sheet played by these winds and

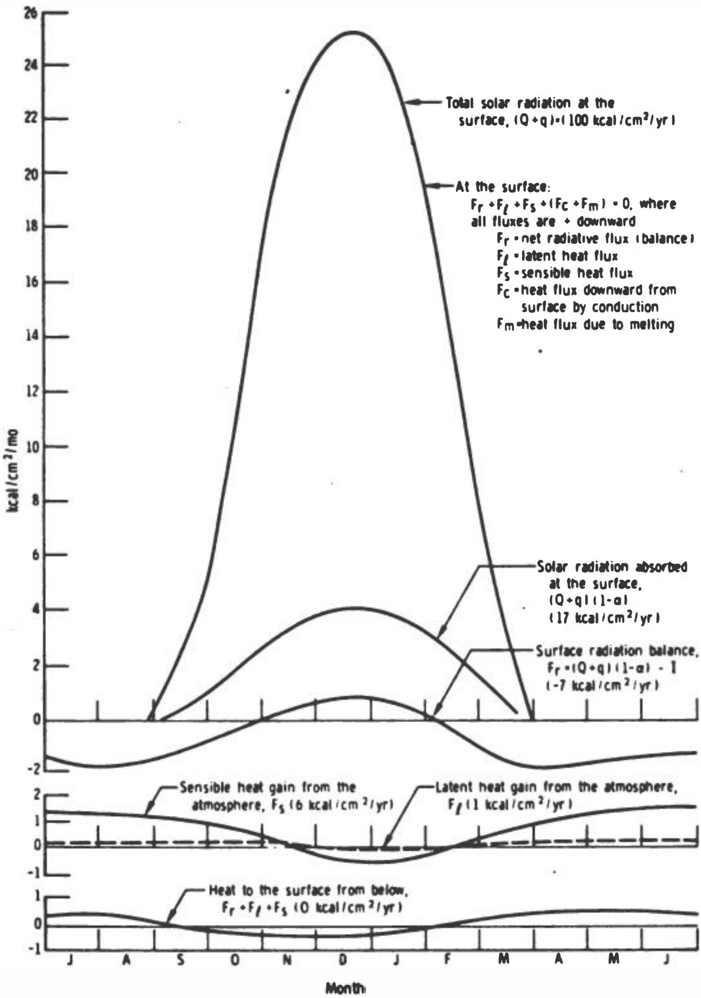


**Figure 3. Estimated surface heat balance of the Antarctic continent (from Fletcher 1969).**

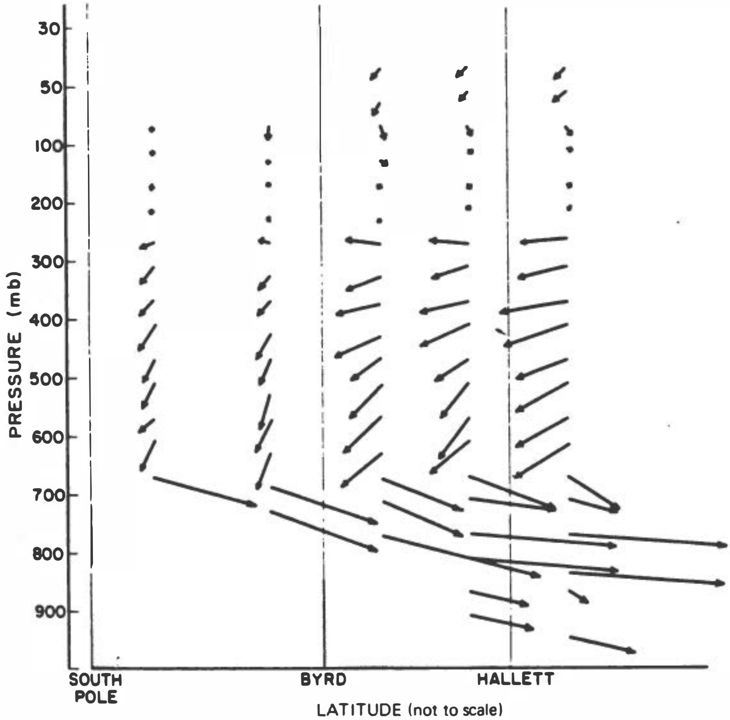


Figure 4. Meridional flow in the Antarctic atmosphere, which compensates for its radiational cooling by adiabatic warming during June and July (from White & Bryson 1967). The largest vectors correspond to velocities of the order of 135 cm $s^{-1}$.

their associated "aeolian" drift snow transport has not yet been fully established. This forms an unsolved parameterization problem for atmospheric modelers. Another effect of the katabatic wind is that it cools and mixes the water surrounding the Antarctic continent, transports drift snow and its associated latent heat into the ocean, and affects the manner in which the surface water is able to freeze. The persistent SE surface winds force the motion of the westward flowing "east wind drift" near the continent. Further, the winds play an important role in the production of open-water "polynyas" favoring the formation of sea ice and of Antarctic bottom water near the continent. These processes allow the circumpolar ocean to exchange heat with the atmosphere

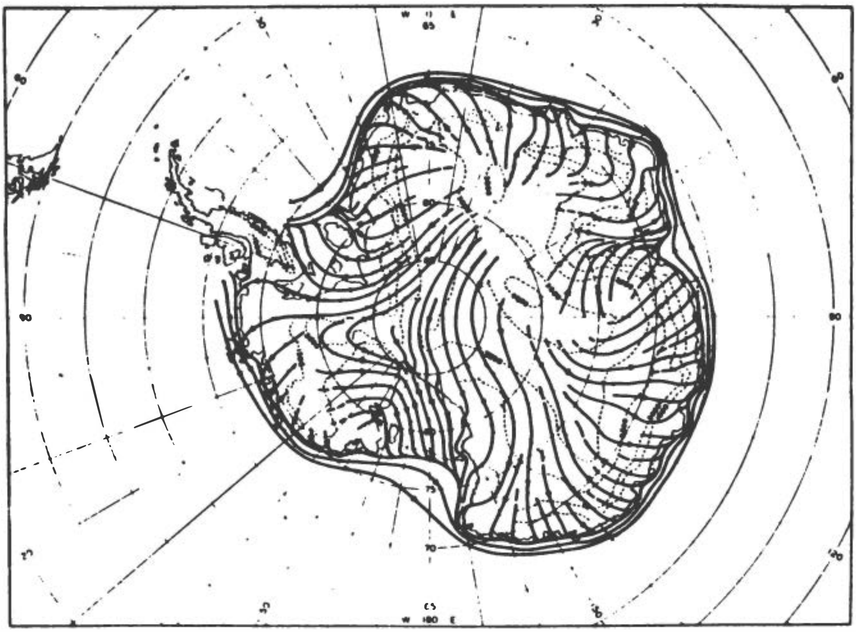

Figure 5. Average surface wind directions on the Antarctic ice sheet, inferred from station and traverse observations and from sastrugi (from Mather & Miller 1967).

much more actively than in the Arctic, where the ocean is ice covered all year, is very stable in its upper layers, and is severely restricted by geography in exchanging water with other areas. The climatic consequence of this contrast between the Arctic and the Antarctic will be elaborated in sections that follow.

## 4. HEAT BUDGET FEEDBACK MECHANISMS

The clearest polar feedback mechanism influencing climate variability is the changing extent of the highly reflective area covered by snow and ice (Kukla & Kukla 1974; Williams 1975, Matson & Wiesnet 1981). A shrinking of the snow and ice cover, such as might follow a warming trend, would result in more solar radiation being absorbed at the surface. This extra heat should further warm the region. This is an example of a "positive feedback" mechanism that has appropriately been included in virtually all of the current latitude-dependent climate models, starting with those of Budyko (1969) and Sellers (1973). The feedback greatly enhances the model sensitivity to cooling and more than doubles the high-latitude response of the mean surface temperature to an increase in heat available to the system from any source (including a hypothetical increase in solar constant, or warming by the greenhouse effect of increasing atmospheric $CO_2$).

Another positive feedback mechanism that has been identified in the polar regions (Wetherald & Manabe 1975) is the joint effect of atmospheric stability and changing moisture content. In the case of a general climatic warming trend, warmer air will evaporate more water at mid-latitudes. As more moist air is advected to the polar regions, it will increase atmospheric opacity to terrestrial infrared radiation, thereby decreasing the net radiation loss at higher levels of the atmosphere. This effect is further enhanced by the suppression of vertical mixing by a stable lower troposphere. Thus, a general global-scale warming would tend to be amplified in the polar regions, and this polar amplification would then be likely to have a feedback effect on the global climate system. However, at all stages of this line of reasoning, associated changes in cloudiness and precipitation that have not been allowed for could lead to a different outcome.

A similarly complex but more transparent system of feedback loops results from significant changes in the atmospheric circulation pattern of high latitudes. Outstanding examples are polar "blocking highs," such as that which developed over the Ross Sea in December 1956, where record high pressures were observed at the start of the International Geophysical Year (IGY), and the persistent Alaskan blocking high of January 1977. The events preceding the formation of a blocking high are well documented; however, their physical causes remain obscure. Their origins have been sought in hydraulic jump processes (changes from supercritical to subcritical flow (Rossby 1950), in marked zonal temperature gradients (Smith 1973), in vertical circulations produced by the splitting of an upper jet stream (Green 1977), and most recently in nonlinear interaction of forced waves with slowly moving free waves (Egger 1978; Charney & DeVore 1979). Whether such factors or more esoteric phenomena such as solar flares (Ramanathan 1977) are responsible for the formation of high-latitude blocking highs, their preferred locations of formation undoubtedly have marked effects on the global climate. In particular, the formation of blocking highs at Australian longitudes seems to be a feature of both interannual fluctuations associated with the Southern Oscillation and longer-term climate changes.

Knox (1982) has identified seasonal variations in the frequency and location of Northern Hemisphere blocks and decadal changes in these climatic features. Similar information for the Southern Hemisphere was collected by Southern Hemisphere meteorologists before and during the IGY but remained tentative for lack of reliable synoptic charts prior to the modern satellite era. In the last 15 years synoptic weather analyses over the Southern Ocean have become increasingly realistic. Pending a comprehensive synthesis of new First GARP Global Experiment (FGGE) material, the prevailing view (Van Loon et al. 1972) remains that the majority of the Southern Hemisphere blocking highs occur southeast of New Zealand, downstream of the most northerly position of the polar front in the central Indian Ocean. The Ross Sea blocking high of December 1956 was an extreme example of that blocking tendency.

## 5. SEA ICE AND CLIMATE

The largest variations in the intensity of the polar heat sinks are associated with variations of ice extent on the ocean. Ice cover on the ocean is a thermal "valve" that very effectively regulates heat exchange between the atmosphere and the ocean, both winter and summer.

In the central Arctic in January, the heat reaching the surface from below is small even though the air is very cold, while the ocean waters only a few feet below are above freezing. When and where ice is absent over a small area, as in an open "lead," the upward heat flow is typically as much as 100 times greater than over the ice. Should the ice cover of the whole Arctic somehow be removed, the total upward heat flow would be 5 or 6 times greater than at present. This would supply most of the usual planetary heat loss to space, with potentially profound implications for the climate in lower latitudes.

Conversely, in summer the high reflectivity of ice reduces the solar heat input to the ocean by a factor of about 4, so that, on the whole, both upward and downward heat exchange are drastically reduced by an ice cover.

In brief, the extent of ice on the ocean regulates heat exchange between ocean and atmosphere and influences the pattern of net atmospheric cooling, thereby influencing the thermal forcing of the dynamic system. To put the matter in global perspective, we need to ask: How much of the earth's surface is subject to thermal regulation by sea ice, and how much does the ice-covered area vary over the year and from one year to another?

The maximum and minimum areas and volumes of sea ice and the total snow/ice covers that occur on the average in the two hemispheres are given in Table 1, based on Untersteiner (1983). In the Northern Hemisphere, the annual maximum extent of sea ice is about 6 percent of the hemispheric area, and the annual variation is about one half of this. In the Southern Hemisphere the maximum sea-ice area is about 8 percent of the hemispheric area, and the annual variation is much larger, seven eighths of this. Thus, the annual variation in area covered by sea ice is roughly two and one half times greater in the

Table 1. Comparison of Area and Volume in Arctic and Antarctic Sea Ice

| | Area ($10^6$ km$^2$) | | Volume ($10^4$ km$^3$) | |
|---|---|---|---|---|
| | Max. | Min. | Max. | Min. |
| Antarctic sea ice | 20 | 2.5 | 3.0 | 0.5 |
| Arctic sea ice | 15 | 8.0 | 5.0 | 2.5 |
| Total snow and ice | | | | |
| N. Hemisphere | 64 (25%) | 10 (4%) | | |
| S. Hemisphere | 34 (13.5%) | 16.5% (6.5%) | | |

From Untersteiner (1983).

Antarctic than in the Arctic. We might expect that year-to-year variations are also larger in the Southern Hemisphere, and the first few years of satellite microwave observations from the Antarctic give a direct verification of this (Zwally et al. 1983).

Records for the Arctic are much more extensive. During the warming of the early part of this century, variations in the mean area of Arctic sea ice were roughly from 10 to 15 percent. Changes in mean thickness may have been as much as one third. These are large variations from a regional viewpoint, but, since the mean Arctic ice extent is only about 11 percent of the hemisphere, the long-term variation in area was only about 1 percent of the area of the hemisphere.

A shrinking ice cover permits greater thermal participation by the ocean and thus reduces poleward temperature gradients and forcing of atmospheric motion. The period of Arctic warming (1917-1938) was a period of relatively weak global atmospheric circulation; the cooling of more recent decades has been accompanied by vigorous global circulation (Barry 1983).

Examination of the surface radiation budget (Figure 2) shows why the Arctic pack ice is sensitive to small changes in atmospheric heat advection during early summer. Maximum absorption is in July rather than in June (when solar energy is maximum), because a sudden increase in absorptivity occurs in midsummer when melting produces puddles of water on the ice. A warmer-than-usual summer advances the melting date, and heat budget variations during this brief summer season are as great as 10 percent of the total annual heat advection by the atmosphere from lower latitudes.

During the northern summer the kinetic energy of the Southern Hemisphere circulation is about 4 times greater than that of the Northern Hemisphere. Also at that time, the meteorological equator is displaced far to the north of the geographic equator, and momentum is being transported across the equator from south to north. It is thus to be expected that Southern Hemisphere trends would be dominant in influencing the global system, partly because of the greater energy of the Southern Hemisphere circulation, but also because the winds over the equatorial zone, which cause important year-to-year ocean variations, are mainly a feature of the Southern Hemisphere circulation.

Variations in the extent of ice on the Antarctic Ocean also seem to be an important factor influencing global climate. The great vigor of the Southern Hemisphere circulation is partly due to the high, cold, white continent of Antarctica, which provides a major heat sink for the global system even in midsummer. However, because it is always cold and white, its properties do not vary much from year to year. On the other hand, the area of ocean surrounding the continent that is covered by sea ice expands in winter to about 8 percent of the hemisphere (more than one-and-one-half times the area of the continent) and shrinks in summer to only about one eighth as much. Variations in sea-ice extent around Antarctica thus offer a high potential for influencing the global system. Their influence should be in the sense that greater ice extent corresponds to stronger thermal forcing and more vigorous global circulation.

Atmospheric blocking highs and their associated distortion of the normal storm-track patterns have been linked to sea ice as both a direct and indirect source of climatic unrest. According to several studies, statistical relationships exist between ice extent in the North Atlantic and certain regional features of the atmospheric pressure field (Brennecke 1904; Meinardus 1906; Koch 1945; Strübing 1967). In their turn, large variations of the ice boundary in the North Atlantic have been shown to be associated with fluctuations in average high-latitude temperatures (e.g., Scherhag 1936). The difficulty in interpreting these observational studies is that ice anomalies are themselves forced by atmospheric anomalies.

An empirical orthogonal function analysis by Walsh (1980) suggests that there may be well-defined situations where sea-ice anomalies lead (and hence could cause) atmospheric circulation anomalies, as the early European climatologists first proposed. The possibility that sea-ice anomalies could force global-scale atmospheric variations was investigated by Herman and Johnson (1978) using an atmospheric general circulation model. One significant result of their experiments was that sea-ice fluctuations in both the North Atlantic and North Pacific caused statistically significant modifications of the Aleutian and Icelandic lows in the model, which in turn affected the intensity of the major Northern Hemisphere subtropical highs.

This type of inferred climatic effect remains perilously close to the noise range of the model used. Somewhat more definite results have emerged from calculations assuming an ice-free Arctic Ocean. Such calculations have been performed by Fletcher et al. (1973), Warshaw and Rapp (1973), and Newson (1973). Although differing in details, these studies found similar and significant differences between the "ice in" and "ice out" features of Arctic and mid-latitude climates. However, these studies do not yet provide conclusive evidence for or against the predictions of Donn and Shaw (1966) and Budyko (1972), namely, that the Arctic Ocean would remain free of ice once that condition had been established by a temporary climatic anomaly. No evidence exists either in recorded history or from the interpretation of sedimentary records that the Arctic Ocean was ever free of ice during the last 0.7 million years; in fact, the beginnings of at least a seasonal ice cover seem to have appeared as long as 2.5 million years ago (Herman & Hopkins 1980).

Major climatic effects can be expected to arise from variations of the Antarctic sea ice. Budd (1975) has described the variability of its pronounced annual expansion and recession and established the empirical effects of sea-ice extent and timing on the climatic conditions along the Antarctic coast. Earlier, Fletcher (1969) pointed out that the rate of decrease of sea-ice area is a maximum in austral midsummer when the solar radiation input is at a maximum (Figure 6). This means that minor changes in ice extent could have magnified effects on the partitioning of surface radiation between atmosphere and ocean, as well as on the total hemispheric heat intake.

The advection of sea ice by the wind is a further variable polar process and is concentrated in certain key regions. One of these is the western Weddell Sea, where intense "barrier winds" are created by the same mechanism as the katabatic winds, that is, in a cold air mass with sloping upper boundary periodically piling up against the eastern side of the Antarctic Peninsula (Figure 7). These winds produce large variations in the flow of sea ice from the Weddell Sea into the Atlantic Ocean. The ice export from the Arctic Ocean through the Greenland-Spitzbergen Passage undergoes similar variations. Such fluctuating sea-ice transports represent another contribution to climatic unrest by the polar oceans.

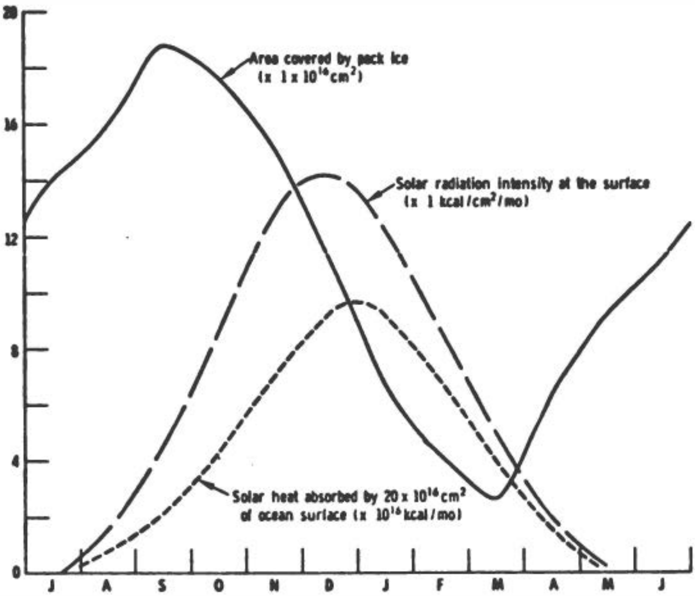


Figure 6. Annual variation of pack-ice area and solar heat absorbed at the surface of the Antarctic sea ice (from Fletcher 1969).

## 6. CLIMATIC ROLES OF THE POLAR OCEAN

The Arctic and Antarctic are both regions of intense air-sea exchange that initiates the main process by which the ocean contributes to the balancing of the global energy budget and to longer-term climatic fluctuations. The role of oceanic heat and mass transports in the polar energy balance is, nevertheless, different for the two polar regions. The differences are of three kinds:

- In the Arctic, heat and mass exchange with adjacent seas occurs mainly between Greenland and Spitzbergen. Other Arctic passages may be locally important, and all appear to show a large temporal variability. In the Antarctic, such exchange is across open boundaries.
- In the Arctic, the inputs of fresh or low-salinity water from a large number of sources, with a large seasonal variability, are particularly significant because they affect the ocean density stratification, which in turn exerts strong control over vertical ocean

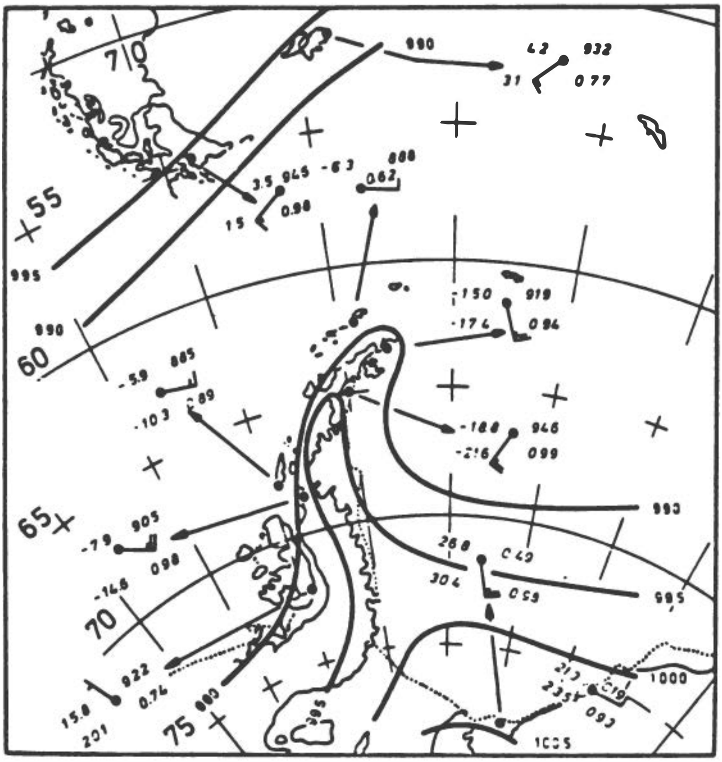

Figure 7. Mean surface winds, temperatures, and surface pressures during an extended period of barrier wind on the east side of the Antarctic Peninsula (1-8 May 1968) (from Schwerdtfeger 1979).

heat and mass fluxes. These inputs at present are important in the Arctic and not in the Antarctic.

o In the Arctic, the shallow and extensive shelf seas favor intensive water modification by horizontal mixing. In the Antarctic, the deeper ocean waters are regions of vertical thermohaline convection. This convection creates the Antarctic bottom water, which may surface centuries later at the rim of the Arctic Ocean (cf. Figure 8).

In general, the most intense interaction between atmosphere and ocean occurs in regions where cold polar air masses first come into contact with the unfrozen ocean surface. These regions are a small fraction of the

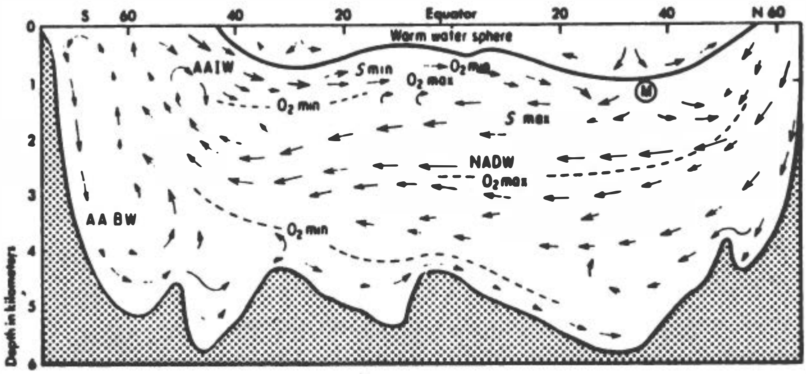


Figure 8. Major water masses and flow directions of the Atlantic Ocean. AAIW marks the Antarctic intermedial water; AABW, Antarctic bottom water; and M, Mediterranean water flowing from east to west (from Turekian 1968).

ocean area and occur in the northwest Pacific, the northwest Atlantic, the Norwegian Sea, and, especially, the Southern Ocean around Antarctica. In such regions the ocean releases enormous amounts of heat and moisture to the atmosphere, quickly transforming the atmosphere to a much warmer and moister state. The heat released to the atmosphere in these regions constitutes a substantial fraction of the total poleward heat transport by the ocean, which, on a global scale, is roughly comparable to the heat transport by the atmosphere (Trenberth 1979).

In each region of such intense air-sea heat exchange, identifiable cold water masses are created that sink to intermediate or deep levels and spread slowly toward the equator. These water masses conserve their heat deficits until they reach low latitudes and interact with the atmosphere there. On a global scale this water-mass forming process leads to a net interoceanic transport of heat deficits from as far away as the North Atlantic to the tropical Pacific via the Southern Ocean. This is the equivalent of a net transport of heat from the tropical Pacific to the North Atlantic, by a route that involves long time lags for achieving equilibrium.

The picture that emerges invites us to think of ocean heat transport in terms of several basin-wide systems, which, individually, do not balance the heat gained or lost from the atmosphere but which do achieve an overall heat balance--at least in the longer term--when viewed as a coupled global system. Because the oceans are coupled to the atmosphere through heat fluxes on both a basin-wide and a global basis, heat anomalies in any part of the total ocean system can influence climate at a distance in both space and time. The time-lag relationships involved suggest that polar ocean anomalies may provide the first signs of a major global-scale climatic fluctuation. This assumption underlines the importance of improved understanding of the formation and subsequent history of the water masses formed in subpolar regions.

## 7. SNOW AND ICE ON LAND

Snow cover is obviously a result of climatic conditions but may have causal roles as well. Snow has already been mentioned as the heart of the so-called "albedo feedback" mechanism. At its maximum extent in the Northern Hemisphere, snow cover extends over an area even larger than that of sea ice (cf. Table 1). A change in that area and in the timing of the snow-cover growth and decay, if sustained over a number of years, could be a significant influence on climate. It should be noted that the contribution of polar snow to raising the planetary global albedo is not its only climatic effect, perhaps not even its most important one. The insulating properties of snow, which allow its surface temperatures to drop and hence reduce the overall radiative heat loss in higher latitudes, are also significant. Other important climatic influences of snow cover may be the change from a net cooling to a net heating effect from atmospheric aerosols when the surface albedo is increased, the latent energy absorbed in the process of ripening and melting of the snow in spring, and the role of snow cover in altering the form and extent of clouds.

The capability of modeling the climatic influences of snow cover is only just coming within reach. It is also to be recognized that the snow cover could be changed through human activity. For example, the albedo change from extensive deforestation would be accentuated by snow cover on the exposed land, and the snow albedo is reduced by deposition of aerosol particles (Warren & Wiscombe 1980).

Mountain glaciers cover too small a total area to have any significant effect on global-scale climate. The polar ice sheets are a very different matter, however. Particularly on the time scales of millennia, the potential effects of ice-sheet variations on the global climate are large. But opinions differ on details and even on the reality of some of the effects.

The most direct effect of ice sheets has already been mentioned: the control exerted by the ice-sheet topography on the surface wind system, which in turn is linked through the structure of the boundary layer to the surface temperature and accumulation rate, key factors in the dynamics of the ice sheet. The complete feedback involved is beyond the reach of current models but finds expression in the Antarctic surface temperatures and especially in their height gradients. Figure 9 shows that these gradients define three broad zones: a coastal zone where the katabatic wind is fully developed and associated with the "isentropic" warming of 1°C/100 m surface descent; a flat central zone with temperature gradients ranging from warming of around 0.5°C/100 m descent to cooling in closed depressions where cold air can accumulate (Kane 1970); and a transitional zone with gradients of 2°C/100 m and more, which alternates between the central and katabatic regimes. The extent and location of this transitional zone should be significantly different for different ice-sheet shapes and must be allowed for in the climatic interpretation of ice-core data.

The factor currently regarded as most directly responsible for changes in the polar ice sheets is the "Milankovitch" variation in the planetary distribution of solar radiation (insolation), associated with changes in the earth's orbit. Especially relevant here is the summer-season insolation, which controls the extent of surface melting and the associated changes in the surface energy balance. Several studies (Weertman 1976; Pollard 1978; Budd & Smith 1981) have demonstrated that the waxing and waning of the Pleistocene ice sheets in lower polar latitudes of the Northern Hemisphere could be explained in this way, allowing also for interactions

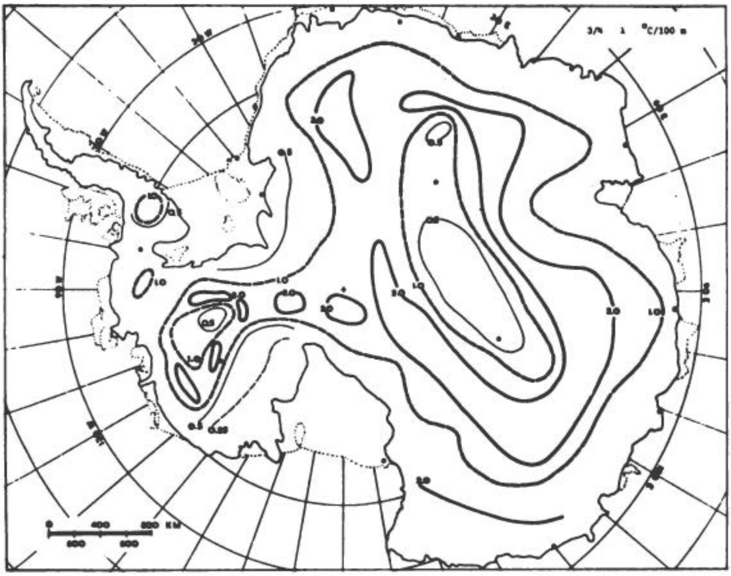

Figure 9. Vertical temperature gradient along the surface of the Antarctic ice sheet (from Budd et al. 1970).

with climate and with the isostatic resilience of the earth's crust. Linked with the Milankovitch effects through the ice sheets are secular changes in sea level. For the Antarctic ice sheet in particular, sea-level changes could modify the dominant ice-flow mechanism on the continental shelves and near the edges of the large bedrock depressions beneath the ice sheet. As a result, the ice sheet could undergo a massive buildup to the edge of the continental shelf during periods of low sea level and develop large floating ice shelves during periods of high sea level.

The most spectacular but also the most controversial way in which Antarctica is visualized as able to disturb the global climate arises from the (as yet unproven) possibility of quasi-periodic ice-sheet "surges." Wilson (1964) has suggested that surges of most or all of the Antarctic ice sheet may occur, filling the Southern Ocean with floating ice that could initiate a global glaciation by raising the earth's albedo. There has been little support for this theory in its initial form, but modeling studies suggest that surges of individual drainage systems in the Antarctic ice sheet may be possible (Budd & McInnes 1978).

Flohn (1974) has speculated on the meteorological effects of a large injection of Antarctic ice into the ocean. The volume of ice he assumes is somewhat larger than the amount transferred in even the largest model surges described by Budd and McInnes but could be produced by the simultaneous surging of two or more separate drainage systems. Converted to icebergs, the injected ice would significantly raise the albedo of an area of ocean aggregating to from 6 to 7 percent of the earth's surface; however, that increase would be nullified by clouds if they normally covered that area. The ocean cooling by the ice would occur mainly north of the present Antarctic convergence and especially in the relatively restricted space of the South Atlantic. Flohn's chain of reasoning then leads through ocean heat-balance changes, a lowering of water temperatures in the Gulf of Mexico and the Caribbean, and the development of a semipermanent trough in atmospheric circulation along the American east coast, with consequent cold summers, to the initiation of a new glaciation of North America and northern Europe. This remains a speculation of how the effect of the south polar region might conceivably spread through ocean processes into the Northern Hemisphere; none of the steps in this chain of events has yet been established.

One particular corollary of the surge theory is that a relatively sudden rise in global sea level should immediately precede each new ice age. The geomorphological record has been intepreted by some observers to show such rises (Hollin 1980; Aharon et al. 1980). Surges have also figured in the model simulations of the Laurentide Ice Sheet (Budd & Smith 1981) in order to account for the recorded southernmost extent of the ice sheet. Surges, moreover, could produce significant changes in the high latitudes occupied by the Antarctic ice sheet, where the radiational control becomes weak.

This discussion emphasizes the fact that the ice sheets take part in a complex system of glacial-climate feedbacks, involving bedrock isostasy and sea-level changes, as well as temperature and precipitation changes. A realistic assessment of the role of the polar ice sheets in long-term climatic unrest must be based on the further development of combined ice-sheet/climate models.

These models are needed also to interpret ice-core records of the climatic past. Ice cores provide a compound record of changes in atmospheric temperature and precipitation as well as in ice-cap shape; therefore, their reliable interpretation should be based on reconstruction of regional histories for ice catchments, or for an entire ice sheet, using as time-space boundary conditions all available core records (plus new cores from key areas suggested by the modeling).

Finally, ice-sheet models clarify the discussion of how polar events might shape our climatic future. A key question follows from climate model predictions, namely, that substantially higher polar temperatures will result from increasing atmospheric concentrations of $CO_2$. The possibility that such temperatures (as well as associated changes in cloudiness and precipitation, which have received much less attention so far) could substantially accelerate the flow of ice sheets, especially in the West Antarctic, requires quantitative checking with adequate ice-sheet models. For the longer-term discussion these models need to be run in conjunction with realistic models of the polar climate to anticipate the course of events as the planet slowly chills toward the next in the long succession of glacial periods.

## 8. RESEARCH STRATEGIES

### 8.1 Central Objectives

As succinctly stated by the SCAR Group of Specialists on Antarctic Climate Research (1981, page 5):

> Two central issues in understanding the role of Antarctica in the dynamics of climate variability emerge:
>
> (1) What factors influence the intensity of the continental heat sink?
>
> (2) How do the resulting anomalies of oceanic heat storage feed back into the ocean/atmosphere system?
>
> The first question calls for a thorough understanding of the factors influencing the radiation budget over the continent, a continuing monitoring programme to observe year-to-year variability, and analysis of ice cores to determine past variations.

The second question calls for a combined modeling and observational program to understand how the ocean transports heat storage anomalies formed in Antarctica, and climate diagnostic studies to understanding how such anomalies feed back on the climate system (possibly by influencing the tropical sea surface temperature distribution).

In a broad sense, the above statements define both the central issues and strategies for dealing with them. An elaboration of the strategies follows.

## 8.2 Factors Influencing the Intensity of the Continental Heat Sink

To determine the time variability of the Antarctic heat sink, a full understanding is required of the following aspects.

1. What is the relative importance of each component of the heat budget? The answer can be based on measurements at only a few locations.
2. How does each component vary with time, and what factors are responsible for the variations? The answer requires continuous measurements of all components and theoretical modeling of the effects of such factors as $CO_2$, $O_3$, $H_2O$, ice crystals, and volcanic aerosols.
3. To what degree do measurements at the few locations of continuous observations represent conditions over the whole heat-sink area? The answer requires comparative observations at other locations, but these need not be continuous nor simultaneous. In general, these requirements apply to measurements of present conditions and also to interpretation of past climatic conditions from ice cores. The two go hand in hand.

The only U.S. station that supports continuous heat-budget measurements, Amundsen-Scott South Pole Station, is not ideally located for sampling the main center of cooling (East Antarctica). Thus, it is especially important to determine that these measurements can represent the heat budget over wider areas.

It is equally important to collect and analyze shallow cores over wider areas in order to decipher the climatic record. For obtaining a climatic record for East

Antarctica, a series of shallow cores at about 100-km intervals from Amundsen-Scott Station toward Mizuho Station would be a good sampling pattern. This could be accomplished by surface transport or by aircraft from Amundsen-Scott.

## 8.3 Transport of Oceanic Heat-Storage Anomalies

This second "main issue" is likely to be most important to climate variability on time scales of decades and centuries. The transport time of Antarctic intermediate water from its region of formation to its mixing to the surface in the tropics is a few decades. The feedback time for deeper water masses is longer. Polar sea-ice processes are the obvious starting point for investigating the polar ocean-atmosphere feedback.

## 8.4 Polar Clues to the Climate of the Coming Centuries

The prime objects for research and monitoring on the time scale of centuries are the large ice shelves of Antarctica and their key processes: surface and internal warming or cooling, melting and freezing on the surface and at the ocean-ice interface, and movement of the line along which ice, rock, and ocean meet ("grounding line"). Since the differential effects on the time scale of centuries are linked with the integrated effects of the decadal time scale, it should be feasible to use model simulation studies for anticipating some of the changes in the ice shelves and for deducing the likely consequences for the grounded ice sheet upstream. In this way, key areas can be identified in which precise monitoring might reveal the onset of long-term trends.

## 8.5 Polar Clues to the Climate of Coming Millennia

On the time scale of millennia we are completely dependent on models for predictions and on the climatic and geological records for verification. The predictions require a marriage of well-matched climate and ice-sheet models, which must involve a minimum of ad hoc assumptions and be capable of reconstructing the Pleistocene and Holocene climatic histories before being turned around to look into the future. To achieve this historical matching, it will probably be necessary to tune the models experimentally with runs using a range of assumptions for the lesser-known model parameters and historical aspects. Thus, the problems of assessing the effects of the polar regions on climate multiply and expand the further we look ahead, and the overall strategy must include priorities for the time scales to be addressed.

## REFERENCES

Aharon, P., J. Chappell, and W. Compston, 1980. Stable isotope and sea-level data from New Guinea supports Antarctic ice-surge theory. Nature, 280, 649-651.

Badgley, F. G., 1966. Heat budget at the surface of the Arctic Ocean. Proceedings of the Symposium on Arctic Heat Budget and Atmospheric Circulation, Lake Arrowhead (pp. 267-277). Rand Corporation-University of California at Los Angeles.

Ball, F. K., 1956. The theory of strong katabatic winds. Australian Journal of Physics, 9, 373-386.

Ball, F. K., 1960. Winds on the ice slopes of Antarctica. In Director of Meteorology (ed.), Antarctic Meteorology (pp. 9-16). Pergamon, London.

Barry, R. G., 1983. Arctic ocean ice and climate: perspective on a century of polar research. Annals, Association of American Geographers, 73 (4), 485-501.

Berger, A., 1978. Long-term variations of caloric insolation resulting from the earth's orbital elements. Quaternary Research, 9, 139-167.

Brennecke, W., 1904. Beziehungen zwischen der Luftdruckverteilung und den Eisverhaltnissen des Ostgronlandischen Meeres. Ann. Hydrogr. Mar. Meteorol., 32, 49-62, 1904.

Budd, W. F., 1975. Antarctic sea ice from satellite sensing in relation to climate. Journal of Glaciology, 15, 417-427.

Budd, W. F., 1980. The importance of the polar regions for the atmospheric carbon dioxide concentrations. In G. I. Pearman (ed.), Carbon Dioxide and Climate: Australian Research, Australian Academy of Science, Canberra.

Budd, W. F. and B. J. McInnes, 1978. Periodic surging of the Antarctic ice sheet--an assessment by modeling. Hydrological Sciences Bulletin, 24, 95-104.

Budd, W. F. and I. N. Smith, 1981. The growth and retreat of ice sheets in response to orbital radiation changes. In I. Allison (ed.), Symposium on Sea Level, Ice Sheets, and Climatic Change. IAHS Publication 131. International Association of Hydrological Science, Gentbrugge, Belgium.

Budd, W. F., D. Jenssen, U. Radok, et al. 1970. Derived physical characteristics of the Antarctic ice sheet, Mark I. Publication 18. Meteorology Department, University of Melbourne 1970. (Publication 120, Australian National Antarctic Research Expeditions, 1971).

Budyko, M. I., 1969. The effect of solar radiation variations on the climate of the earth. Tellus, 21, 611-619.

Budyko, M. I., 1972. The future climate. EOS, 53, 868-874.

Charney, J. G. and J. G. DeVore, 1979. Multiple flow equilibria in the atmosphere and blocking. Journal of Atmospheric Sciences, 36, 1205.

Delmas, R. J., J. M. Ascencio, and M. Legrand, 1980. Polar ice evidence that atmospheric $CO_2$ 20,000 yr BP was 50% of present. Nature, 284, 155-157.

Donn, W. L. and D. M. Shaw, 1966. The stability of an ice-free Arctic Ocean, Journal of Geophysical Research, 71, 1087-1095.

Doronin, Yu. P., 1963. On the heat balance of the central Arctic. Reports of the Arctic and Antarctic Research Institute, 253, 1963. (Translation by Israel Program of Scientific Translations, Jerusalem.)

Egger, J., 1978. Dynamics of blocking highs. Journal of Atmospheric Sciences, 35, 1788-1801.

Fletcher, J. O., 1965. The heat budget of the Arctic Basin and its relation to climate. Report R-444-PR. Rand Corporation, Santa Monica, California.

Fletcher, J. O., 1969. Ice extent on the Southern Ocean and its relation to world climate. Rep. RM-5793-NSF. Rand Corporation, Santa Monica, California.

Fletcher, J. O., Y. Mintz, A. Arakawa, and T. Fox, 1973. Numerical simulation of the influence of Arctic sea ice on climate. Energy Fluxes over Polar Surfaces. WMO Technical Note 129. World Meteorological Organization, Geneva.

Flohn, H., 1947. Background of a model of the initiation of the next glaciation. Quaternary Research, 4, 385-404.

Green, J. S. A., 1977. The weather during July 1976: some dynamical considerations of the drought. Weather, 32, 120-126.

Herman, G. F. and R. M. Goody, 1976. Formation and persistence of summertime arctic stratus clouds. Journal of Atmospheric Sciences, 33, 1537-1553.

Herman, G. F. and W. T. Johnson, 1978. The sensitivity of the general circulation to arctic sea ice boundaries: A numerical experiment. Monthly Weather Review, 106, 1649-1664.

Herman, Y. and D. M. Hopkins, 1980. Arctic ocean climate in later Cenozoic time. Science, 209, 557-562.

Hess, S. L., 1959. Introduction to Theoretical Meteorology. Holt-Dryden, New York.

Hollin, J. T., 1980. Climate and sea level in isotope stage 5; on East Antarctic ice surge at 95,000 BP? Nature, 283, 629-633.

Kane, H. S., 1970. A study of 10 m firn temperatures in central East Antarctica. IAHS Publication, 86, 165-174.

Kellogg, W. W., J. A. Coakley, Jr., and G. W. Grams, 1975. Effect of anthropogenic aerosols on the global climate. Proceedings of the WMO/IAMAP Symposium on Long-Term Climatic Fluctuations. WMO 421 (pp. 323-330). World Meteorological Organization, Geneva.

Knox, J. L., 1982. Atmospheric blocking in the Northern Hemisphere. Canadian Climate Center Report, Downsview, Ontario.

Koch, L., 1945. The East Greenland ice. Meddelelser om Gronland, 130, 373.

Kukla, G. J. and H. J. Kukla, 1974. Increased surface albedo in the Northern Hemisphere. Science, 183, 709-714.

Loewe, F., 1956. Etudes de Glaciologie en Terre Adelie, 1951-52. Actualities Scientifiques et Industrielles #1247. Hermann, Paris.

Mather, K. B. and G. S. Miller, 1967. Notes on topographic factors affecting the surface wind in Antarctica with special reference to katabatic winds; and bibliography. Technical Report, Grant GA-400. Geophysical Institute, University of Alaska, Fairbanks.

Matson, K. B. and D. R. Wiesnet, 1981. New data base for climatic studies. Nature, 289, 451-456.

Meinardus, W., 1906. Periodische Schwankungen der Eistrift bei Island, I, II, III. Ann. Hydrogr. Mar. Meteorol., 34 (4,5,6), 148-162, 227-285.

Milankovitch, M., 1941. Canon of Insolation and the Ice Age Problem. Royal Serbian Academy, Belgrade. (Translation by Israel Program for Scientific Translations, Jerusalem, 1969.)

Miles, M. K. and P. B. Gildersleeves, 1978. Volcanic dust and changes in Northern Hemisphere temperature. Nature, 271, 735-736.

Newson, R. L., 1973. Response of a general circulation model of the atmosphere to removal of the Arctic ice-cap. Nature, 241, 39-40.

Pollard, D., 1978. An investigation of the astronomical theory of the ice ages. Nature, 272, 233-235.

Ramanathan, V., 1977. Troposphere-stratosphere feedback mechanism: stratospheric warming and its effect on the polar energy budget and the tropospheric circulation. Journal of Atmospheric Sciences, 34, 439-447.

Rossby, C. G., 1950. On the dynamics of certain types of blocking waves. Journal of Chinese Geophysical Society, 2, 1-13.

SCAR Group of Specialists on Antarctic Climate Research, 1981. Basis for an Action Plan on Antarctic Climate Research. Scientific Committee on Antarctic Research, Cambridge, United Kingdom.

Scherhag, R., 1936. Eine Bemerkenswerte Klimaanderung uber Nordeuropa. Ann. Hydrogr. Mar. Meteorol., 64, 96-100.

Schwerdtfeger, W., 1970. The climate of the Antarctic ice. In S. Orvig (ed.), Climates of the Polar Regions. Volume 14 of World Survey of Climatology, Elsevier, New York.

Schwerdtfeger, W., 1979. Meteorological aspects of the drift of ice from the Weddell Sea toward the mid-latitude westerlies. Journal of Geophysical Research, 84, 6321-6328.

Sellers, W. D., 1973. A new global climatic model. Journal of Applied Meteorology, 12, 241-254.

Smith, R. F. T., 1973. A note on the relationship between large scale energy functions and characteristics of climate. Quarterly Journal of the Royal Meteorological Society, 99, 693-703.

Strübing, K., 1967. Über Zusammenhänge zwischen der Eisführung des Ostgrönlandstroms under der Atmosphärischen Zirkulation über dem Nordpolarmeer, Deutsche Hydrographische A., 20. 257-265.

Thompson, L. G. and E. Mosley-Thompson, 1980. Microparticle concentration variations linked with climatic change: evidence from polar ice cores. Science, 212, 812-815.

Trenberth, K. E., 1979. Mean annual poleward energy transport by the oceans in the Southern Hemisphere. Dynamics of Atmopsheres and Oceans, 4, 56-64.

Turco, R. P., R. C. Whitten, and B. Toon, 1982. Stratospheric aerosols: observations and theory. Reviews of Geophysics and Space Physics, 20, 233-279.

Turekian, K. K., 1968. Oceans. Prentice-Hall, Englewood Cliffs, New Jersey.

Untersteiner, N., 1983. Climate Research Reviews. Joint Scientific Committee, World Climate Research Program, World Meteorological Organization, Geneva.

Van Loon, H., J. J. Taljaard, T. Sasamori, J. London, D. V. Hoyt, K. Labitzke, and C. W. Newton, 1972. Meteorology of the Southern Hemisphere. American Meteorological Society, Boston, Massachusetts.

Walsh, J. E., 1980. Empirical orthogonal functions and the statistical predictability of sea ice extent. In R. S. Pritchard (ed.), Proceedings of the ICSI/AIDJEX Symposium on Sea Ice Processes and Models, September 1977. University of Washington Press, Seattle.

Warren, S. G. and W. J. Wiscombe, 1980. A model for the spectral albedo of snow. Part II: Snow containing atmospheric aerosols. Journal of Atmospheric Sciences, 37 (12), 2734-2745.

Warshaw, M. and J. R. Rapp, 1973. An experiment on the sensitivity of a global circulation model. Journal of Applied Meteorology, 12, 43-49.

Weertman, J., 1976. Milankovitch solar radiation variations and ice-age sheet sizes. Nature, 261, 17-20.

Wetherald, R. T. and S. Manabe, 1975. The effects of changing the solar constant on the climate of a general circulation model. J. Atmos. Sci., 32, 2044-2059.

White, F. D., Jr. and R. A. Bryson, 1967. The radiation factor in the mean meridional circulation of the Antarctic atmosphere during the polar night. In S. Orvig (ed.), Polar Meteorology, WMO Technical Note 87 (pp. 199-224). World Meteorological Organization, Geneva.

Williams, J., 1975. The influence of snow cover on the atmospheric circulation and its role in climatic change; an analysis based on results from the NCAR global circulation model. Journal of Applied Meteorology, 14, 137-152.

Wilson, A. T., 1964. Origin of ice ages: an ice shelf theory for Pleistocene glaciation. Nature, 101, 147.

Zwally, H. J., J. C. Cosimo, C. L. Parkinson, W. J. Campbell, F. D. Carsey, and P. Gloersen, 1983. Arctic Sea Ice Cover 1973-76: Satellite Passive Microwave Observations. NASA-SP-459 (206 pp.). NASA Goddard Space Flight Center, Greenbelt, Maryland.

# Appendix B. Polar Regions as Windows on the Past

J. D. Hays, J. T. Andrews, C. C. Langway, and T. L. Péwé

## 1. INTRODUCTION

Physical models of the climate system have become very sophisticated, yet underlying all models is a set of assumptions and simplifications. The success of climate models is generally judged on their ability to simulate the present climate of the earth, but this kind of validation test does not necessarily assure us that the models display the right sensitivity to a change of conditions.

A complementary approach to understanding the causes of climatic change--and for predicting future changes--is the careful documentation of how climate has changed in the past. Since the climate system is global, it is best studied on a global scale. Nevertheless, present information clearly suggests that the polar regions play a major, perhaps pivotal, role in long-term climatic change.

The record of past climatic changes is preserved in various ways, sometimes in remarkable detail, in the sedimentary deposits and glacial ice of high latitudes. Only recently have skills been developed that allow the extraction of some of this information from these deposits. The success achieved during the last two decades indicates the richness of the information these deposits contain and suggests that future work will yield even more significant findings.

Each of the polar sedimentary and ice-core sequences provides paleoclimatic information about specific aspects of the climate system on various temporal and spatial scales. The limitations of the record depend on the dating accuracy of the chronology and the degree to which a measured sedimentary or ice-core parameter represents an aspect of the climate system.

## 2. SOURCES OF PALEOCLIMATIC INFORMATION

Four sedimentary environments (including ice sheets) contain important paleoclimatic information; in the subsequent sections we discuss each of these. In addition, trees rings provide valuable information.

First, we consider ocean-floor sediments, which contain the longest record of climatic change, with a potential length of about 50 million years. Since these deposits cover vast areas, ocean-floor sediments provide climatic information on a broad spatial scale as well.

The skeletal remains of microscopic organisms are the primary source of climatic information in cores obtained from the ocean floor by drilling and coring. From these records, information about past sea-ice cover, sea-surface temperature, surface circulation patterns, bottom-water temperature, and sea level can be gleaned.

Second, the frozen sedimentary sequences of ice sheets contain one of the most detailed records of past climates, with an annual record that goes back some 8000 years. Climatic information less well controlled chronologically has been recovered for more than 100,000 years, and the potential length of the record may be 10 times longer. The large ice sheets of Greenland and Antarctica, together with the smaller ice caps of Canada and Spitsbergen, are the primary sampling areas. Information about air temperature at the top of the ice sheet, accumulation rates of snow, and the concentration of atmospheric aerosols and trace gases can be drawn from these frozen deposits.

Third, lakes and bogs in polar regions contain detailed records of past physical and biotic changes that may be interpreted in terms of climatic events. Although the Pleistocene ice sheets of both America and Europe obliterated the record beyond the past 10,000 to 15,000 years in some areas, longer records are available from unglaciated areas in Alaska, Arctic Canada, and Siberia. The pollen and other microfossils contained in these sediments provide information about changing summer temperature and precipitation. Lake and bog deposits are widely spaced in polar regions of the Northern Hemisphere, providing excellent spatial control.

Fourth, polar soils contain another data set, the structure built by permafrost-induced movements and the products of glacial ice movements. Since frost-induced features can sometimes be related to specific temperature regimes, these structures (when dated) can provide information on past temperature regimes. Moraines (when dated) give information on the extent of past ice sheets at specific times.

Finally, trees growing in subpolar areas contribute information about variations of climate during the past few centuries, revealed by variations in their annual growth rings. In high latitudes, the annual growth of indigenous conifers and deciduous species is confined to the relatively short summer growing season; ring variations reflect primarily temperature conditions in that season.

## 2.1 Ocean-Floor Sediments

Ocean-floor sediments provide information on two important aspects of the past climate system: (a) the ocean--its surface temperature, ice cover, and circulation patterns--and (b) sea level which, to a good first approximation, is inversely related to the total volume of water in ice sheets on land.

2.1.1 Establishing a chronology. It is obviously essential to know the time when past events occurred, and oceanographic and ice-volume changes can be placed in an increasingly accurate chronology based on the radiometrically dated globally synchronous record of magnetic field reversals (Figure 1). Additional synchronous levels are provided by the variation of the ratio between two isotopes of oxygen ($^{18}O/^{16}O$) found in the carbonate portion of the sediments.

This ratio in the shells of calcareous bottom-living fossils (Foraminifera) is controlled primarily by the isotopic composition of the water from which they build their shells. Because fractionation occurs between $^{18}O$ and $^{16}O$ as water is removed from the ocean and deposited as snow on ice sheets, the ratio of $^{18}O/^{16}O$ of ocean water changes as ice sheets grow and contract. As the ocean is thoroughly mixed in about a thousand years, changes in its isotopic composition are spread evenly throughout the ocean within this time. As a consequence, the ratio of $^{18}O/^{16}O$ locked in the shells of benthic foraminifera record in some complex fashion the growth and decay of land-based ice; therefore, these changes in the $^{18}O/^{16}O$ ratio represent synchronous levels in sea-floor sediments within the mixing time of the ocean and the disturbing effects of sediment-burrowing

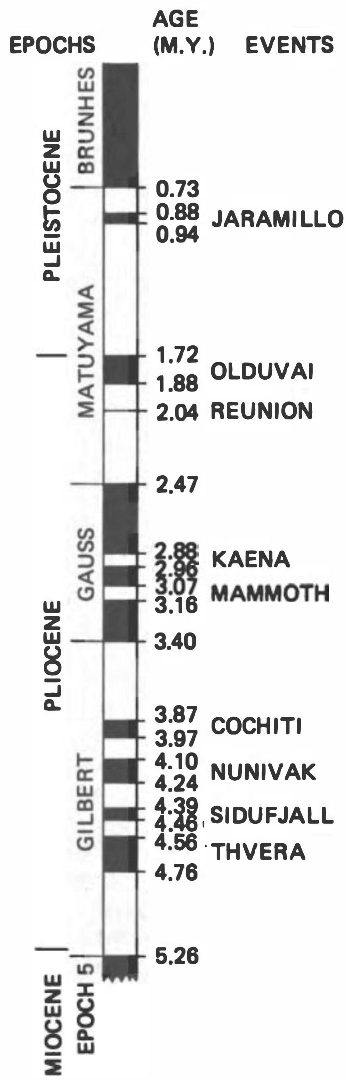


**Figure 1. The sequence of magnetic field reversals and their age, determined largely from ratios of various radioactive decay products. These magnetic events occurred at the same time everywhere. (From Berggren et al. 1980.)**

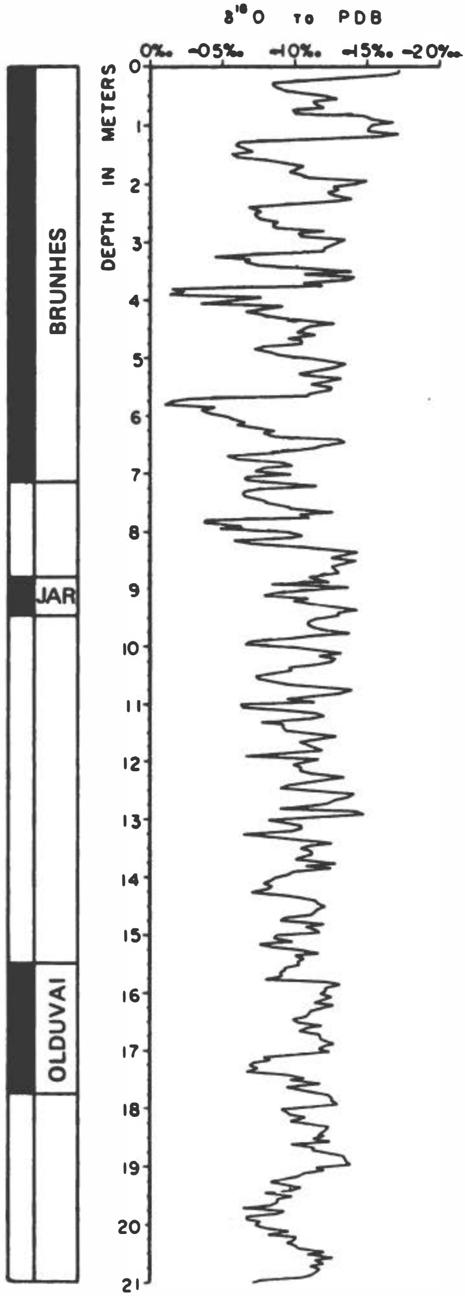


**Figure 2. Sample of an ocean-bottom core analyzed in terms of the $^{18}O/^{16}O$ isotope ratio. (From Berggren et al. 1980.)**

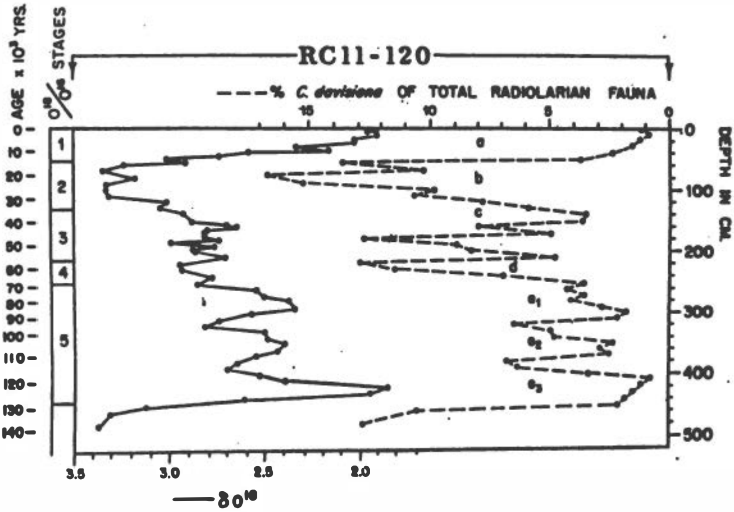


Figure 3. An example of a core analyzed for both $^{18}O/^{16}O$ and the fraction of a particular radiolarian fauna in the total population, both as a function of depth or age. Note how the two factors tend to be closely related.

organisms (Figure 2). (The latter effect is thought to amount to usually less than 10 cm.)

Since high-latitude ocean sediments frequently contain little or no calcium carbonate, the stratigraphic control in these areas is often dependent on fossil changes that can be shown to be regionally synchronous and can be directly correlated with changes in the oxygen isotope ratio (Figure 3).

Ages for various climatic events during the Quaternary are controlled by $^{14}C$ dating during the last 20,000 years, by uranium series dates on raised terraces that can be correlated with the $^{18}O/^{16}O$ record in the deep sea at about 80,000, 105,000, and 127,000 years B.P., and by potassium/argon dating of the last reversal of the earth's magnetic field--730,000 years B.P. Between these discrete levels in ocean-floor cores, ages of climatic events are estimated by interpolation, which involves the assumption of uniform sedimentation rates.

This chronology is currently being improved by making use of the fact that certain frequency components in the climatic record are known to match those of Earth's orbital parameters, with a constant phase relationship between them (the so-called Milankovitch effect). By this means, the absolute dating of ocean-floor cores has the potential of reaching an accuracy of plus or minus a few thousand years over the last million years.

There are approximately 10,000 ocean-floor sediment cores housed at various oceanographic institutions that can be dated by these methods.

2.1.2 Estimates of past water-mass positions and sea-surface temperatures. The distribution of microfossil assemblages in the deposited sea-floor sediments show coherent patterns with today's surface ocean-water masses and major current systems. These assemblages have been successfully used by CLIMAP, for example, to estimate the past position of surface water masses at the maximum of the last glaciation (~18,000 years B.P.).

2.1.3 Estimates of sea ice. Although estimates of the extent of sea ice are on a less firm quantitative footing than estimates of temperature at the sea surface, a number of lines of evidence are consistent with substantially increased sea-ice cover during the last glacial maximum in both hemispheres.

Since sea ice is known to inhibit primary productivity if it is present during the summer months, strongly reduced biogenic remains, combined with slower accumulation rates, suggest areas where summer sea ice occurred during the last glacial maximum. Additional evidence from sea-ice-rafted terrestrial and volcanic detritus in both hemispheres further supports this line of reasoning.

2.1.4 Sea-level changes. Since sea-level changes are caused by the transfer of water from the ocean basins to ice sheets on the continents, the volume of this transfer can be calculated if the change in ice volume can be measured. This transfer can be estimated in two ways: first, by mapping terminal moraines and calculating equilibrium ice-sheet profiles, from which thickness can be estimated and a volume calculation made. (See Appendix C.) This approach has been attempted by Hughes et al. (1980). Large uncertainties in this method of inferring changes in sea level arise because of the lack of information about the extent of (marine-based) ice sheets in both the Northern and Southern Hemispheres. To bracket these uncertainties, Hughes et al. (1980) developed maximum and minimum estimate models. Second, because changes of the ratio of $^{18}O/^{16}O$ in the shells of benthic foraminifera are to a large extent controlled by changes in volume of land ice, changes in sea level can be estimated if changes in $^{18}O/^{16}O$ can be calibrated to changes of sea level. Raised terraces on the island of Barbados and elsewhere provide the information needed to do this. Rates of tectonic uplift on Barbados are uncertain, but if they are assumed to be uniform and near the average ($\sim$25 cm per 1000 years) between successive low and high stands, changes in isotopic values on the reef-crest coral (*Acropora palmata*) from recent and late Pleistocene terraces can be related to elevation differences (Shackleton & Matthews 1977).

Comparison of several successive low-sea-level and high-sea-level stands give a calibration factor of $-0.011$°/°° per meter in the ratio of $^{18}O/^{16}O$ ($\delta^{18}O$) (R. K. Matthews personal communication).

Application of this calibration factor to the 0.65°/°° range of $\delta^{18}O$ in high-deposition-rate, deep-sea cores yields an estimate of 150 m for the amplitudes of the last glacial/interglacial sea-level change (Shackleton 1977).

This calculation rests on the assumption that there were not large floating ice shelves, which would alter somewhat the isotopic composition of the ocean while having no effect on sea level. Errors produced by this assumption, if corrected, would reduce by perhaps 30 m the estimate of sea-level lowering during an ice age.

Both the maximum ice reconstruction of Hughes et al. (1980) and the isotopic estimate (Shackleton 1977) yield similar values for lowering of sea level. Probably lowering of sea level did not exceed 150 m at the height of the last ice age, some 18,000 years B.P. and may well have been less. A significantly lower estimate is suggested by a third approach, based on the reasoning that the amount of crustal isostatic rebound (the rising of the land surface after the weight of an ice sheet has been lifted) is proportional to former ice thickness; thus, the rebound numbers can be used to calculate the former ice thickness (Peltier & Andrews 1976; Peltier 1982). The results of these calculations based on isostatic rebound determinations indicate thinner ice sheets than those calculated by Hughes et al. (1980).

2.1.5 Future needs. To make better use of the deep-sea record for climatic interpretations, there are two primary needs: first, to study more intensively, and with new techniques, existing deep-sea cores, and second, to collect, through use of the hydraulic piston corer, much longer undisturbed sedimentary sequences in high latitudes.

In the first category, methods to improve time control at high latitudes must precede detailed studies of the frequency distribution of climatic records from deep-sea sediments. With better time control, detailed comparisons can be made between the two hemispheres. The development and utilization of techniques to measure the past distribution of sea ice and to determine the frequencies of its fluctuations are of prime importance. Comparison of these fluctuations between the two hemispheres would provide insight into the role of these variations as amplifiers of global climatic change.

The sedimentary sections available for these kinds of studies are limited at present to 10-20 m. The hydraulic piston corer makes possible the recovery of sequences hundreds of meters long. The use of this device in high latitudes would revolutionize deep-sea studies of climate. It should receive the highest priority.

## 2.2 Glacier Ice Records

2.2.1 Introduction to ice cores. The massive ice caps of Antarctica and Greenland and the smaller ice caps of some Arctic Islands contain a high-resolution record of climatic change at specific locations. The key to the stratigraphy of these deposits is the recognition of annual layers. Table 1 summarizes the various kinds of climatic and environmental history that can be extracted from different ice-core studies and Figures 4 and 5 show where these drilling sites were located. The investigations range in scope from the direct physical, mechanical, and chemical analyses of continuous samples, or samples from selected depth intervals, to the indirect borehole and airborne geophysical sensing techniques covering the particular vertical profile being considered (Dansgaard 1977; Oeschger 1977; Gudmandsen 1975; Gudmandsen & Overgaard 1978; Langway 1977).

For convenience, ice cores are categorized according to depth of recovery, as shown in Table 1: shallow ice cores range downward to about 100 m and, depending on the

Table 1. Glaciological and Climatological Investigations of Polar Ice Cores (From Langway et al. 1984)

| Parameter Measured | Research Objectives | Time Resolutions | Galciological, Climatic, and Environmental Information |
|---|---|---|---|
| **Shallow Depth (0-100 m), 100-500 Years or More of Record** | | | |
| Physical features<br>Stratigraphy<br>Structure<br>Density<br>Crystal size<br>Light transmission<br>Air bubbles | Direct observation and counting of annual layers. Record of textural variations. Quantitative measurements of features. | Better than ± 10% | Seasonal and annual accumulation. Identification of unusually warm/cold periods. |
| Chemical species | Identification and counting of annual concentration minima and maxima. Identification and correlation of index horizons. | Better than ± 10% | Seasonal and annual accumulation rates, annual layer thickness, atmospheric turbidity and residence times, volcanic record, pollutant production and transport, aerosol sources. Model testing. |
| Stable isotopes | Continuous measurements and counting of annual layers. | Better than ± 10% | Seasonal and annual accumulation rates. Relative temperature variations. Model testing. |
| Radioactive isotopes | Absolute dating from natural radioactive decay. Artificial radioactive fallout horizons. Seasonal variations. | Varies with isotope | Accumulation rates, wind transport and residence times. Solar activity. Dating. Model testing. |
| Gas composition | Radiogenic dating of gases. Total and trace gas analyses. | Varies with isotope | History of atmospheric composition. $CO_2$ changes. $^{14}C$ dating. |
| Microparticles (terrestrial and extra-terrestrial) | Counting annual concentration variations. Dust influx. | Variable | Seasonal and annual layer Atmospheric turbidity, wind transport, residence times. |
| Pollen | Concentration peaks. Species changes | Variable | Vegetation cover. Transport processes. |
| Temperature (borehole) | Identification of recent climatic events (i.e., Climatic Optimum). | Variable | Mean annual temperature. Recent variations in temperature regime. |

Table 1 (continued)

| Parameter Measured | Research Objectives | Time Resolutions | Galciological, Climatic, and Environmental Information |
|---|---|---|---|
| **Intermediate Depth (0-500 m), 1,000-11,000 or More Years of Record** | | | |
| All the above | All above | Above | All above. |
| Mechanical properties | Confined compression, tension, and shear testing to determine variation of strain with stress, temparature, and time. | | Flow model and deformation history, time scale. Natural behavior of large ice masses. |
| Physical properties | | | |
| Crystal size | Thin section analysis. Identification of abrupt changes in crystalline texture creating distinct time horizon. | Decades to centuries | Changes in surface environmental conditions, variations in abundance of nucleating agents. |
| Fabrics | Universal stage measurements of crystal orientations. Ultrasonic velocity measurements. | Centuries | Ice rheology, variations in strain rate, flow history. |
| Air volume | Determining variation of pore close-off density with temperature and elevation. .Time horizons marked by air volume changes. | Decades to centuries | Paleoelevations. Ice age/interglacial boundaries. |
| Stable isotopes | Tima horizons; long-term climatic variations. | Decades to centuries | Delineation of ice age boundaries. |
| Radioactive isotopes | Measurement of long-lived isotopes. | Centuries | Dating "old" glacier ice. |
| Electrical properties | Dielectrics, conductivity, time horizons. | Decades to centuries and distinct horizons | Ice age/interglacial boundaries, volcanic horizons. |

Table 1 (continued)

| Parameter Measured | Research Objectives | Time Resolutions | Galciological, Climatic, and Environmental Information |
|---|---|---|---|
| **Deep core (to 3000 m) and 1500 to 2 x $10^6$ or More Years of Record** | | | |
| All of the above | All above | Above | All above |
| Remote sensing | Electromagnetic properties (UHF and SHF frequencies) magnetometry, IR radiometry. | Centuries | Ice age/interglacial boundaries, Pre-Holocene stratification and glacier structure. Time horizons for correlations with other locations. Extremely useful for drill site location. |
| Geophysics | | | |
| Surface strain network | Velocity, strain, and accumulation rates. | Decades to centuries | Areal variations in dynamic behavior, model calculations. |
| Borehole inclination | Strain measurements over entire profile. | | Flow history, check on model calculations. |
| Elastic waves | Ultrasonic velocity measurement, seismic studies. | | Anisotropic characteristics. Flow behavior. |
| **Interface and Sub-Ice Material** | | | |
| Sedimentary petrology | Fabric, lithology, included carbonaceous material. | Centuries to millenia | History of glaciation, maximum extent of ice sheet, subaerial weathering, age of till. |
| Interstitial ice chemistry and gas composition | Determine possibility of water phase in past. | | Thermal history. Model testing. |
| Rock fragments and bedrock | Radiomatric dating. | | Age of bedrock. Degree or rate of erosion, weathering, scouring. |
| Geophysics | Seismic, gravity, electrical resistivity profiles. | | Ancient landforms, drainage patterns, bottom topography and roughness. |
| Temperature | Basal and borehole measuraments. | | Thermal and climatic history and regime, geothermal gradient. |


**Figure 4. Map of Greenland showing logistic bases and sites of recent Greenland Ice Sheet Project (GISP) activities.**


**Figure 5. Map of Antarctica showing logistic bases near which cores have been obtained.**

net snow accumulation at the site, extend back in time from 100 to 1000 years or more. This time frame provides several important "index horizons" or stratigraphic levels. For example, by some of the techniques listed in Table 1, ice-core studies can reveal major volcanic episodes (e.g., Krakatoa in 1883 and Katmai in 1912). Ice cores of 100 m also span the dust bowl catastrophe of the 1930s as well as the period when massive fluxes of anthropogenic pollutants were expelled into the atmosphere.

The ice cores of intermediate depth range downward from the surface to 600-900 m and bridge time spans from over 1000 years to the end of the Wisconsin Stage (11,000 years B.P.) or more. In other words, they extend throughout historically recorded time, through the "climatic optimum" (4500 to 8000 years B.P.) and the Holocene period on into a global ice age. Intermediate-depth ice cores provide data that are not only distinctive for the region from which they were obtained but also reflect general trends of global significance. Intermediate and deeper ice cores are also invaluable in the investigation of secular variations that must be understood for proper interpretation of a much longer temporal study of deeper ice cores (to bedrock).

Calculations based on glaciological and geophysical parameters show that the ice at the base of the ice sheet in south-central Greenland could have ages up to $10^6$ years and that Antarctic ice may be as old as 2 to 3 x $10^6$ years. Ice cores as old as this would extend through at least two, and perhaps more, of the glacial/interglacial periods of the Pleistocene Epoch and permit verification of the low-frequency cyclic nature of the data obtained in a variety of studies on the deep ice cores from Camp Century and Dye-3, Greenland, and from Byrd Station, Antarctica.

The parameters measured in ice cores vary from the physical, chemical, and isotopic properties to the gaseous and wave-propagation characteristics of the ice. As shown in Table 1, depending on the particular parameter or features measured, different information is revealed. Most data from 100-m core studies reveal seasonal or annual accumulation rates or other environmental conditions. These are useful, however, because a great majority of the natural (volcanic) and anthropogenic (bomb testing and other pollutant ejections into the atmosphere) events occurring during the past century are well documented. As we go deeper into the polar ice sheets, we not only measure all of the parameters listed for the shallow ice cores but also, through necessity, turn to other indirect and remote investigative techniques. For example, glacial/interglacial boundaries are suggested by airborne radar-sounding techniques. Multiple internal reflections that extend laterally for many miles are also revealed by continuous flight profiling. These signals probably contain climatic information, but their interpretation is not yet sufficiently clear.

In short, it is possible to count back in time from the surface to about 100 to 150 years using classical stratigraphic techniques, mainly physical property determinations (explained in the next section), and back to about 8000 years B.P. using the stable isotope, chemical, and microparticle methods. The limits are primarily based on the diffusion properties of the stable isotope, the minimum detection level of a chemical species, and the dynamic flow characteristics of the location. At this time, accurate dating of ice older than about 12,000-15,000 years awaits further development of radiometric methods and refinement of mathematical and geophysical models.

2.2.2 Stratigraphy. The precipitation that falls in the dry snow zone of a high polar glacier buries the previous net accumulation. The density of the underlying snow increases by compaction, and with increased depth the snow or "firn" gradually transforms into glacier ice. (By definition, the transition from "firn" to ice occurs when the intercommunicating pore spaces close off into isolated bubbles, which usually occurs when the density increases to about 0.83 g/cm$^3$.) The resulting stratigraphic sequences not only consist of polycrystalline aggregates of snow, firn, or ice crystals but also contain atmospheric air, entrapped when the pore spaces close off, together with all of the organic, inorganic, soluble, and insoluble foreign material that fell with the precipitation or as dry fallout on the snow surface. Thus, ice sheets preserve a record of a considerable variety of paleoenvironmental variables and of their changes over time. Cores into the ice provide samples of this record, amenable to systematic measurement as a continuous function of depth.

Stratigraphic sequences in firn cores consist of summer and winter deposits determined by variations in density, hardness, grain size, melt phenomena, and depth hoar features. The study of snow, firn, or ice stratigraphy is comparable to the investigation of annual lake varves, laminations in a sedimentary formation, or tree rings. In all cases, individual layers are deposited or formed, and the specific characteristics or properties of the material allow one to differentiate between summer and winter or between other time intervals.

Seasonal and annual layers in ice cores are readily differentiated from the surface to 100-m depths (to ages of 100-150 years) by means of stratigraphic examination and careful physical property measurements on cores from high polar glaciers (Langway 1970). Physical property studies useful in identifying annual layers and the history of glaciers include grain and crystal sizes and shapes; structural seasonal features (melt layers, wind crusts, depth hoar); air-bubble size, shape, grain boundary relationships, and volume; bubble pressure; and ice-crystal orientations.

2.2.3 Stable isotopes. The stable isotope technique has become one of the standard tools in glaciology (Dansgaard 1953, 1954, 1964; Dansgaard et al. 1969, 1973, 1975; Johnsen et al. 1972; Epstein & Sharp 1959;

Picciotto et al. 1960; Epstein et al. 1970; Patterson et al. 1977; Koerner 1977; Koerner & Fisher 1981). It is especially useful in polar regions for measurements of accumulation rates, dating of ice cores, and recording of past temperature changes (Dansgaard 1977).

As mentioned previously, the major isotopes studied are those of oxygen: $^{18}O$ and $^{16}O$. The ratio of these isotopes, when expressed as a deviation from the ratio for present seawater in parts per thousand, is defined as $^{18}O$ (or simply $\delta$). Seasonal snow accumulation cycles can also be identified by measuring the seasonal $\delta$ variations in firn and ice. In high-accumulation regions (more than 25 cm of ice per year), this method of discriminating annual layers is applicable many thousands of years back in time.

The principal limiting factor in applying the method to older ice is the tendency for the seasonal $\delta$ variations to be gradually erased over time by isotopic diffusion in the solid ice. Beyond the depth where the seasonal variations are no longer obvious, statistical deconvolution techniques may allow identification of the seasonal cycles at somewhat greater depths (Johnsen 1977). When extended to depths corresponding to ice formed in the last glaciation, detailed $\delta$ records should reveal at least the longer-term changes of the accumulation rates on the ice sheet and the relationship of accumulation rates to temperature.

Ice dating by the stable-isotope method is possible by counting summer maxima in $\delta$ downward from the surface as long as the seasonal oscillations are preserved. This technique gives a much higher dating accuracy (a few years) than any radioisotope method, and depending on the accumulation rate in late Wisconsin times, it may allow dating of ice layers as old as 15,000 or even 20,000 years. This ability, with its potential for inferring snow accumulation rates back into the last glacial period, may be matched as a dating technique, only by microparticle and chemical analyses (Herron & Langway 1979; Herron 1982; Hammer 1977; Hammer et al. 1978).

Atmospheric temperature variations can also be determined from the stable-isotope technique because the isotopic fractionation governing the ratio $^{18}O/^{16}O$ in the snow deposited on the ice sheet is influenced by the temperature where and when the snowflakes were formed. This temperature, in turn, is the same as the air temperature at the altitude of snow formation. When looking at long-term $\delta$ records, such as those from the

Camp Century core, the inferred temperature changes compare quite favorably with those established from other studies of glacial chronology (Dansgaard et al. 1969). In addition, some of the more pronounced temperature fluctuations within historical time and during this century are recognizable, particularly in the Crete station record from central Greenland.

Thus, $\delta$ profiles obtained from long ice cores provide extremely valuable information in the form of:

1. Climatic records of temperature, more detailed than those obtained by any other method and absolutely dated as far back as 10,000-20,000 years;
2. Data sufficient to estimate time lags and coupling coefficients between climatic changes within the Northern and Southern Hemispheres and between hemispheres;
3. Comparisons with particle profile data to determine whether correlations exist between atmospheric turbidity and global climate change;
4. Estimates on the corrections to the $^{14}C$ scale beyond the range of dendrochronology (which only goes back about 6000 years).

2.2.4 Radioactive isotopes. Although counting annual layers through stable-isotope and other stratigraphic dating techniques (physical properties, total $\beta$ activity, dust concentrations, and elemental chemistry) is considered accurate to within a few years, perhaps back to the late Wisconsin, independent checks on these techniques are desirable. In addition, other absolute methods must be developed for ice older than about 30,000 years (Oeschger et al. 1966, 1967, 1976).

For radioactive isotopes produced by cosmic radiation in the atmosphere, a near equilibrium exists between production and decay. Dating by these radioisotopes is based on the assumption that their specific activities at the time of deposition or occlusion (withdrawal from contact and exchange with the atmosphere) have always been the same and that during subsequent downward movement of the ice layers the specific activities decay exponentially with time according to their half-lives (T 1/2).

The following radioactive isotopes are found in ice as components of:

- water molecules $^{3}H$,
- occluded gases $^{14}C$ in $CO_2$, $^{39}Ar$, $^{85}Kr$, $^{81}Kr$, and in particulate and dissolved matter, $^{32}Si$, $^{26}Al$, $^{36}Cl$, $^{53}Mn$, and $^{10}Be$.

The different radioactive dating techniques available for glaciological purposes are listed below according to their age ranges (Oeschger et al. 1975, 1977; Lehman et al. 1977; Siegenthaler & Oeschger 1978; Siegenthaler et al. 1979; Stauffer & Berner 1978; Berner et al. 1978).

1. Short term (0-100 years) on ice cores. Short-lived artificial isotopes (e.g., fission products $^{90}Sr$ and $^{137}Cs$, and $^{3}H$ from fusion bomb tests) and natural isotopes (e.g., "pre-bomb" $^{3}H$ and $^{210}Pb$) can be used for dating near-surface snow and firn layers of up to 100 years of age. Tritium and other fission products have been deposited on the Greenland Ice Sheet according to a well-known time sequence. Pronouced deposition horizons are found in the layers from 1954 to 1963, corresponding to certain atomic tests.

2. Medium term (100-1000 years) on ice in situ. At somewhat greater depths, longer-lived isotopes must be used. For the time period between 50 and 1500 years, $^{32}Si$ (T 1/2 = 295 years) and $^{39}Ar$ (T 1/2 = 269 years) are useful for dating purposes.

3. Long term (more than 1000 years). At present, $^{14}C$ offers the only possibility for reliable radioactive dating of ice up to the age of 25,000 years, although several potential isotopes are available: $^{53}Mn$, $^{36}Cl$, $^{81}Kr$, $^{26}Al$, and $^{10}Be$ (Loosli & Oeschger 1968; Oeschger et al. 1977; Oeschger 1977). Samples for $^{36}Cl$ measurements were taken at Crete in 1974, and methods for the radioactivity measurements are being studied.

2.2.5 Chemistry. Polar snow layers retain much of their original chemical composition for up to 120,000 years, even though these layers are subject to physical alteration by stress, flow, and diagenetic processes. Within these layers are dissolved and insoluble particulate matter (deposited aerosols), which reached the snow surface as ice-crystal nuclei, condensation nuclei (scavenged by falling snowflakes), and dry fallout. Ice sheets are generally remote from sources of natural and artificial aerosols; therefore, the chemical composition of the impurities within the ice tends to reflect regional or global atmospheric burdens of constituents and their changes over time (Windom 1969; Cragin et al. 1977; Herron et al. 1977). These constituents may be of marine, terrestrial, or extraterrestrial origin, including volcanic, biological, and anthropogenic matter carried to the ice sheet by atmospheric transport and deposited on it by selective removal mechanisms.

Ice sheets are similar to the oceans in that they act as temporary reservoirs for these materials. However, oceanic sedimentary records are often complicated by postdepositional chemical and biological modification. The ice sheets contain only atmospherically transported material, and because of the low temperatures, chemical substances remain more or less as they were at the time of deposition. Chemical analysis of deep ice-core profiles is an ideal way to investigate the nature and composition of the baseline global aerosol and its changes, to model the interaction of aerosols and climate, and to quantitatively assess the impact on air quality of human activities throughout historic time (Hammer 1977, 1979; Hammer et al. 1978; Herron 1980).

A recent development of interest to the climate community has been the ability to determine the carbon dioxide ($CO_2$) content of the air at the time of deposition of the snow. This determination is made by the careful recovery of entrapped gas released by melting an ice core, followed by an analysis of the gas (Stauffer & Berner 1978; Berner et al. 1978, 1980; Raynard & Delmas 1977; Delmas et al. 1980). Results indicate that the $CO_2$ content of the atmosphere during the height of the Wisconsin ice age was only about half the present content and that during the warm period 5000-8000 years B.P., it may have been somewhat greater than it is now.

Seasonal variations in chemical concentrations are known to occur in ice cores and surface snow (Murozumi et al. 1969; Weiss et al. 1971a,b, 1975; Langway et al. 1977). These variations provide information about elemental sources and their atmospheric pathways to the ice sheet (Koide & Goldberg 1971). Seasonal cycles based on chemical analyses persist longer than stable-isotope variations (specifically $^{18}O/^{16}O$ ratios), since the diffusion rates for the elements of interest are significantly less than for self-diffusion of ice. Elements of diverse origin, such as dust-derived Al and marine Na, showing concentration maxima in different seasons, may be used for cross-dating purposes in the same way as dust and stable-isotope variations.

2.2.6 Future needs and directions for glacial ice cores. The greatest emphasis should be directed toward improving capabilities in two general areas: (a) field collecting equipment and devices, and (b) more comprehensive laboratory investigation on ice cores by experienced and competent scientists.

In regard to equipment, there is immediate need for, and much international interest in, "off-the-shelf" shallow (100 m) and intermediate depth (500 to 1000 m), large diameter (minimum 10 cm), fast operating, and lightweight ice-core drilling devices capable of air transport. Several high-performance operational devices exist, but shop drawings necessary to make replicas of them are not readily available. We suggest that a comprehensive engineering study of all available shallow and intermediate rigs be made by appropriate specialists and that a universal model be designed incorporating the best features of available rigs.

Until 1968, only one operating deep-core drilling rig existed in the scientific community that was capable of drilling to depths of more than 1000 m. In connection with the GISP program, a new, unique, lightweight deep-ice core drilling rig for use in Greenland was conceived, developed, and successfully operated between 1979 and 1981 by the Danish, Swiss, and U.S. participants of the program (Langway et al. 1984). We suggest a careful U.S. engineering study be made of this GISP rig for possible standardization as "the deep core drill" and drawings be offered to all nations engaged in polar or subpolar core drilling. This procedure will conserve drill development and testing costs for the total community and allow savings to be directed to a stronger and more vigorous and comprehensive core-study program.

Remote penetrating probes that could be used to measure in situ physical parameters and to collect large liquid, gaseous, and solid samples at selected depths in a glacier have been under development in several nations since the early 1960s. Limited success has been achieved, probably due to low-priority attention and little financial support of the projects. We suggest full review of the ratio of development costs to the return of significant data and comparison to borehole and core studies before giving further attention to remote systems.

In regard to laboratory studies, we note that since the advent and successful recovery of largely continuous, undisturbed, intermediate and deep ice cores during the pre-IGY (International Geophysical Year) period (1956 in Greenland), many countries (e.g., the United States, Denmark, Switzerland, Soviet Union, Australia, France, Japan, New Zealand, Belgium, England, Canada, Chile, and Austria) have drilled for ice cores in the polar, subpolar, and alpine regions in both the Northern and Southern Hemispheres. We note, however, that about 25 or so ice cores deeper than 100 m have been recovered from the different localities since 1956, and only three deeper than 1000 m. This finding illustrates the need to obtain more ice cores from all areas of the globe for comprehensive and integrated investigation with current, advanced analytical techniques.

A few of the first-order scientific problems related to climatic change and paleoenvironmental conditions that could be answered by ice-core studies include those listed in Table 2.

### 2.3 Lakes and Bogs in the Tundra: The Pollen Record in Polar Regions

2.3.1 Introduction of palynology. The use of pollen analysis, or palynology, as a tool for climatic reconstruction is becoming increasingly well established in Arctic areas. It has been used infrequently in polar regions of the Southern Hemisphere, primarily because of the general rarity on the Antarctic continent of organic deposits with useful microfossil diversity.

In the Northern Hemisphere, interpretation of "pollen diagrams" (plots of relative species abundances as functions of time) from tundra areas has advanced slowly. By and large, the majority of palynologists have not considered the specific problems of interpreting diagrams from tundra regions; that is, relatively little effort has been expended on identification of tundra taxa to the species level, and there are few studies on the trends and diversity of the modern "pollen rain" (deposition of airborne pollen) onto the Arctic tundra.

The Arctic tundra does not consist of a uniform assemblage of plants. Botanists have long recognized that the tundra can be subdivided into three or four

Table 2. Needed Studies of Climatic and Paleoenvironmental Conditions

| Study | Scientific Objectives |
|---|---|
| Stable isotopes | Paleoclimates, annual accumulation |
| Radioactive isotopes | Dating, global circulation (natural and artificial pollution) |
| Chemical elements, ions, heavy metals | Baselines, global circulation, pollution |
| Terrestrial dust and aerosols | Volcanic index, dating, global circulation |
| Extraterrestrial dust | Fluctuations in meteoric deposition |
| Halogenated hydrocarbons | Pesticides, ozone layer |
| $CO_2$ | Greenhouse effect, its change with time |
| Total gas composition | Bulk and trace analyses |
| Organics | Pollen & plant fragments as indices of biotic events and changes, elemental carbon |
| Physical properties | Environmental conditions of past |
| Stratigraphy | Fluctuations in particle deposition, surface melt features of past |
| Fabrics | History of deposition and flow |
| Mechanical properties | Flow characteristics related to time scale |

major vegetational zones. These can be characterized by the occurrence, or absence, of specific species. Young (1971) discusses his subdivision of the Arctic tundra in terms of the climate critical to plant growth, which consists simply of the sum of mean monthly temperatures above 0°C. Figure 6 illustrates Young's zonation, where the following definitions of the boundaries of summer warmth zones apply:

| Zone | Summer Warmth (Sum of monthly mean temperatures above 0°C) |
|---|---|
| 1 | 0-6 |
| 2 | 6-12 |
| 3 | 12-20 |
| 4 | 20-35 |
| Within timberline | 35 |

Although palynologists have qualitatively recognized some variations in Arctic tundra pollen, these have rarely been associated with detailed vegetational studies (see Birks 1973 for a review).

Extensive palynological research has been conducted in Greenland by European palynologists (e.g., Fredskild 1973, and in press; Funder 1978). In Arctic Canada the area north of the tree line has been studied in a few localities (Rampton 1971; Ritchie & Hare 1971; Nichols 1975). In northern Alaska, palynological research is undergoing renewed emphasis (Brubaker et al. 1983; Walker et al. 1981) following the pioneer work of Livingston (1955).

2.3.2 Pollen identification. Pollen diagrams from Northern Hemisphere polar regions are available from northern Alaska, Arctic Canada, Greenland, Spitsbergen, and the USSR Arctic. Pollen identification from tundra areas is more difficult than that from sites within the forest limit because of less generic diversity and the lack of accepted criteria for distinguishing pollen of different species in a genus (e.g., Carex) or different genera in a family (e.g., grasses). Much information is

**Figure 6. Boundaries of the four proposed Arctic floristic zones. Zones 2 and 4 are shaded, zones 1 and 3 are unshaded. (After Young 1971.)**

being lost by the absence of pollen keys that enable identification to the species level. Work of this detail is best exemplified by the papers from Greenland. However, other workers are now realizing the need for this type of analysis. It should be noted that this type of detailed pollen identification takes considerable time and patience.

2.3.3 Accuracy and precision of pollen counts. Little work has been undertaken on the problem of the accuracy and precision of pollen counts on tundra materials. The production of pollen in Arctic areas is low. Estimates of pollen influx, measured as grain/cm$^2$/yr, indicate that values range between $10^1$ and $10^2$; thus, many slides, when they are prepared, have relatively few pollen grains. In many cases whole slides have to be counted to get some statistically satisfying numbers. In areas within Young's Zones 3 and 4 (Figure 6), pollen productivity is significantly higher than in Zones 2 and 1. In Zone 4 considerable tree pollen is blown into the area and deposited, and in both Zones 4 and 3 the shrubs of *Betula* and *Alnus* are prolific producers. However, in Zones 2 and 1 the major components are the grasses, sedges, and heaths.

2.3.4 Types of sediment sampled. Pollen analyses have been performed on sediments of six major kinds: lake bottoms, marine muds, ice cores, peat of bogs, frozen ground, and surface soils. The first two are of special interest here.

Probably the majority of pollen diagrams from polar areas come from pollen extracted from lake muds. Sampling this type of environment usually involves the use of a piston corer, using ice as a platform, or coring from a small inflatable boat or raft. Length of core recovery from this sort of operation varies between 1 and 4 m. Longer cores would appear feasible in some situations. Lake cores from polar areas frequently cover whole or part of the Holocene.

Pollen also has been reported from ice cores obtained from Camp Century (Fredskild & Wagner 1974), Devon Island (Lichti-Federovitch 1974), and the Penny Ice Cap (Baffin Island) (Short, 1979). Pollen influx is low and is dominated by exotic pollen taxa. Work is in progress on a pollen stratigraphy from the Devon Island ice cores, and a pollen stratigraphy is planned for a deep hole of bedrock through the Penny Ice Cap (1986?). Pollen preservation is often excellent from such an environment. This type of pollen investigation is just starting.

2.3.5 Radiometric control. Dating control of Arctic pollen diagrams ranges from poor (one date per diagram) to excellent (greater than eight dates). The dating of Arctic lake and peat sediments encounters problems associated with the low biological productivity, the presence of permafrost, and rapid rootlet penetration at sites that are characterized by excessively well-drained sediments.

Peaty sediments have not always given reliable dates in polar regions. Bomb carbon appears to be preferentially deposited in Arctic regions, and in some peats, $^{14}C$ activity greater than recent levels has been reported at depths of 50 cm and for sites that, on archeological grounds, should date from 2800 years B.P.

Ice-core pollen analysis has at least the possibility of providing annual estimates of pollen flux at those levels where annual layers can be detected. However, the pollen concentrations are sufficiently low that this analysis would demand a dedicated core solely for pollen use. The potential information in this type of record suggests that such a core should be included in future science plans for ice-coring programs from sub-Arctic or temperate latitudes where there are sufficient pollen and spore cones.

Resolution in Holocene core materials is limited by the standard error on the radiocarbon dates. This is commonly ±200 years at the 95 percent confidence level. If time series are to be developed that can take into account the possible resolution implicit in sedimentation rates (for example, one sample over 10-50 years), we suggest that multiple dates are required at critical horizons. This approach could lower the standard error to ±60 years at the 95 percent confidence level if ten replicate dates were run, or ±100 if only four dates per level were obtained. Annual pollen variations are possible, of course, if annual lake laminea can be sampled; such sites are probably available.

In general, the resolution of many Arctic pollen diagrams is such that variance spectrum analysis should be feasible to gain some information on periodicities (if any) from about 100 to 2500 years (e.g., Nichols et al. 1978).

2.3.6 Transfer functions and climatic information. Climatic information in polar pollen diagrams has been inferred from:

- presence or absence of various thermophilous shrubs or herbs (Funder 1978);
- presence of far-travelled exotic pollen taxa that are interpreted to reflect an increase in southerly advection into a region (Hyvarinen 1972; Nichols et al. 1978);
- transfer functions based on a statistical relationship between pollen species percentages in modern surface samples and climatic parameters (Kay 1979; Andrews et al. 1979; Andrews & Nichols 1981).

Funder's (1978) study of lake cores in East Greenland indicates the substantial level of paleoclimatic information that can be obtained through a detailed analysis. Kay's (1979) results are derived from the application of transfer functions based on surface moss polsters to lake sediment pollen variations. The results of Andrews et al. (1979, 1980) were obtained by applying transfer functions based on surface mosses to peats. The latter approach is preferred because it is not clear how closely the tops of lake-sediment cores match the pollen rain in surface mosses (which should be a more comparable material to peats in terms of pollen catch). An example of the kind of climatic information that can be recovered is the decline of July temperatures between about 4500 and 3500 years B.P. in the eastern Canadian Arctic (Andrews et al. 1979). Furthermore, modern July temperatures there are warmer by 2°C or so than at any time since about 2500 B.P. A time-series analysis suggests a periodicity of about 250 years in the exotic pollen influx that may be statistically significant (see also Nichols et al. 1978).

2.3.7 Future needs and research areas for polar palynology. Future research efforts into the climatic information held in pollen spectra from the tundra regions should focus on a number of specific problems.

These have been alluded to in earlier sections, but they are summarized here.

A better data base is required for the geographic variation in the modern pollen rain. Because the geological record is interpreted from a variety of materials (peats, lake muds, marine muds, and ice), the modern variability and trends in surface pollen occurrences need to be examined through an analysis of core top data, annual snow layers, and surface mosses.

More and longer lake cores are required at strategic sites throughout the polar regions. In the Northern Hemisphere, workers should seek sites outside of the area covered by the late Wisconsin ice sheets so that the history of vegetation and climate prior to the Holocene can be evaluated. Colinveaux's success in obtaining long cores from North Alaska indicates that, with the proper equipment and some luck, these attempts are feasible.

The geographic coverage of pollen sites should be improved. Large blank areas exist even if just the Holocene Series is considered. Spitsbergen, North and East Greenland, the Queen Elizabeth Islands, the USSR Arctic islands, and the northern region of Alaska have produced few Holocene pollen spectra. Longer continuous core records are even more sparse. Rampton (1971) has recovered a core from the Yukon territory of Canada that extends back to about 24,000 years B.P., and Colinveaux has several long core records. This coverage needs to be drastically improved so that the climatic conditions during the period 10,000 to 20,000 B.P. can be properly evaluated.

Every effort should be made to increase the number, accuracy, and precision of radiocarbon dates on Arctic peat sections and cores. Several dates should be obtained at the same level in order to reduce the statistical uncertainty in the laboratory age determinations. With better chronological control, sampling of cores and peats should be undertaken in such a way that the data are immediately available for time-series analysis, which would mean that the dating should proceed prior to sampling for pollen. In many cases, however, the need for dating is determined by the pollen spectra.

If the above needs can be met, we foresee within the next decade a series of synoptic views of the Holocene climate based on polar palynological data processed through appropriate transfer function equations. More information will be needed from USSR sources.

## 2.4 Periglacial Features as Climatic Indicators

2.4.1 Geomorphic indicators. Pingos are geomorphic indicators of perennially frozen ground. They are conical ice-cored hills or mounds, round to oval in plan, 20-400 m in diameter, and 10-70 m high, that formed when large, massive layers of ice grew near the surface in permafrost. Pingos are of two distinct generic types: the closed system and the open system. Most closed-system pingos occur under Arctic climate conditions and a mean annual air temperature (MAAT) lower than -5°C. Open-system pingos are almost entirely confined to sub-Arctic climate conditions and a MAAT between 0°C and -5°C (Péwé 1983b; Mackay 1978).

In areas now subpolar, or even temperate, polar conditions can be shown to have existed in the past by the presence of pingo scars, which are microtopographic features formed when the pingo ice melts as permafrost disappears. They indicate past permafrost conditions and the mean annual temperature, depending on the type of pingo, if this can be determined.

Ice wedges are masses of foliated ground ice that grow in thermal contraction cracks in permafrost. Active cracking is to be expected, and wedges continue to grow where the ground temperature at the top of the permafrost layer is about -15°C or lower in winter (Péwé 1966). Most foliated ice masses occur as wedge shape, vertical, or inclined dikes, 1 cm to 3 m wide and 1-10 m high. Ice wedges grow only in permafrost and form best with a MAAT in the vicinity of -6°C and -7°C. However, for purposes of generalization, we will use here the figure of -6°C for an area where ice wedges regularly form. It should be emphasized that the growth of ice wedges depends not on MAAT but rather on the rate at which the temperature drops at the time of cracking (Péwé 1966, 1983a; Lachenbruch 1966). Ice wedges may be inactive or dormant and remain in permafrost even though the MAAT may be higher than -6°C. They will exist until the permafrost thaws. They are part of the three-dimensional polygonal network of ice that causes the formation of a microrelief pattern on the surface called polygonal ground. These polygonal surface patterns are widespread in polar areas today.

With the thawing of the permafrost and melting of the ice, space formally occupied by ice wedges is replaced with sediment that falls into the opening, forming that which has been termed an "ice wedge cast." Ice wedge casts are widespread in subpolar and temperate areas where polygonal networks of such casts are observable on the surface, a relic from the Wisconsin ice age. The ice wedge cast and ice wedge polygons indicate the condition of past permafrost and that the MAAT was -6°C or lower when they formed.

A widespread land form in unglaciated polar areas that is indicative of past permafrost consists of cryoplanation terraces (Reger 1975; Reger & Péwé 1976). These are large bedrock steps or terraces on ridge, crests, and hilltops; the terraces possess at least one scarp and a "flat" surface. The trend or "flat" area is 10 to several hundred meters wide and long and slopes from 1° to 15°, parallel to the ridge crest. Terrace scarps are from 1 to 30 m high. They form below the snowline and appear to require a mean annual air temperature colder than -10°C to -12°C. The mean summer temperature was probably 2°C to 6°C (Reger & Péwé 1976).

Rock glaciers, tongue-shaped or lobate masses of unsorted, angular, frost-rived material, with interstitial ice (if active) and with steep lichen-free fronts 10-100 m high, are one of the most spectacular periglacial deposits but limited in aerial extent. Active rock glaciers occur below the snowline and exist in areas of permafrost. Therefore, they indicate that the MAAT at the time they existed was lower than 0°C (Péwé 1983a,b).

2.4.2 Stratigraphic indicators. In addition to the geomorphic phenomena we have described, exposure of the underlying stratigraphy in some areas tells us something about the past or present climate. For example, in areas south of the permafrost border there are deep-lying relics of permafrost indicative of temperatures in the past with a MAAT colder than 0°C. Also there are the well-known frozen fossil carcasses of extinct animals, which indicate that the MAAT was less than 0°C at the time of death, and that the climate has been that cold or colder until the present time. If the climate had become warmer since the death of the animal, the carcass would have disintegrated as the permafrost thawed (Péwé 1975).

2.4.3 Geothermal indicators. Measurements of ground temperature as a function of depth indicate that permafrost is not necessarily in equilibrium with present climate at the sites of measurements. In some areas, about 25 percent of the present permafrost is the product of an extinct climate, when it was colder than at present (Lachenbruch et al. 1966, p. 158). A study of temperature profiles through deep permafrost near Barrow indicates that the mean annual ground surface temperature has increased about 4°C since the mid-nineteenth century, about half of this increase having occurred since 1930. The mean annual surface temperature at Barrow 100 years ago was about -12°C, and with the recent warming, it is now about -9°C.

2.4.4 Geochemical indicators. A relatively new approach to the study of perennially frozen ground is a chemical investigation of ice and sediments associated with permafrost. Most of the work in this field has evolved around the distribution of soluble salts and sodium, magnesium, calcium, and potassium in perennially frozen sediments as a reflection of past thermal and leaching regimes. If the soluble salt content of existing perennially frozen ground has been leached in the past, it indicates a former thaw of the now-perennial frozen ground; therefore, the MAAT must have risen to greater than 0°C to permit this to occur. These events can be dated by standard geologic stratigraphic techniques (Péwé 1975).

2.4.5 Cold climate features: Semiqualitative temperature data. There are many microgeomorphic features common to polar areas that may or may not be associated with permafrost. Repeated freezing and thawing of the ground throughout the year produce small-scale pattern ground, a feature of rigorous climate that does not necessarily need underlying permafrost, thus a climate with a MAAT less than 0°C. Also solifluction, the slow downslope movement of detrital debris under the influence of gravity and wet conditions, is common in cold regions. Boulder or rock fields and other mass-movement features indicate rigorous climate but not necessarily one where the MAAT is colder than 0°C. These features occur today in subpolar and temperate areas and are no longer active. They are indicators of the past and denote a colder period when perennially frozen ground was present (Péwé 1983a).

2.4.6 Glacial constructional features as indicators of past climate. Glacial deposits yield three types of primary data that give indirect information about paleoclimates: the areal distribution of glacier ice at a specific time in the past, the fluctuations of glacier termini through time, and the former equilibrium line altitudes (ELAs) of mountain glaciers. All three types of data are difficult--if not impossible--to interpret in terms of specific climatic parameters because they represent the complex dynamic reaction of glaciers to changes in mass balance, which in turn has a complex relation to climate. Rather, the data give general trends and relative magnitudes of climatic changes. Also, the data on glacier distribution, when analyzed on a global scale, afford the basis for reconstructing the areal and vertical extent of past ice sheets, ice caps, and mountain glaciers. Such reconstructions are used as input to numerical models that attempt to simulate past conditions of the atmosphere. Because this aspect of glacial geology is so important, its basis is reviewed in Tables 3 and 4.

## 2.5 Tree Rings

The great boreal forests of the north circumpolar regions, populated mainly by evergreen conifers but often intermixed with a variety of hardy deciduous species, extend northward to fairly well-defined Arctic tree lines in both North America and Eurasia. Relatively isolated stands of such trees are found also in more polar latitudes, for example, in the Brooks Range in Alaska. These trees, like those in alpine areas at lower latitudes, add new growth each year only during the short summer growing season and with a vigor that varies from year to year depending primarily on variations of local air temperatures, snow conditions, and insolation. The changes in growth are reflected in the varying widths and other characteristics of the new wood added each year, evident in cores extracted from their trunks by increment boring tools to reveal the sequence of annual growth rings over the lifetime of each tree. By a judicious combination of measurements of such rings from many trees of the same species in an area, in some cases including measurements on fallen dead wood that tends to be well preserved in the dry polar cold, precisely dated tree-ring indices extending back through several

**Table 3. Type and Validity of Data Used for Dafining the Former Extent of Glaciers**

| Type of Feature | Chronologic Control | Reliability |
|---|---|---|
| Moraine ridge | Radiocarbon dating. Only as valid as the basic stratigraphic control. Also, the reliability depends critically on the type of material being dated. Wood is best. Lacustrine deposits and bulk shell samples commonly give misleading ages.<br><br>Tepdrochronology. As good as the basic data. Commonly afford only limiting ages. | Questionable, unless the internal characteristics of the moraine can be studied carefully to ascertain that overriding has not occurred and that till sheets are not stacked in the moraine. Such features might suggest that the moraine is complex and does not repreeent the terminal position of the advance assumed to have produced the moraine. This situation is commonplace in the classic Great Lakee region and has led to numerous examples of misinterpretation of glacial history. |
| Till sheets | Radiocarbon dating. Only as valid as the basic stratigraphic control. Also, the reliablity depends critically on the type of material being datad. Wood is best. Lacustrine deposits and bulk shell samples commonly give misleading ages.<br><br>Tepdrochronology. As good as the basic data. Commonly afford only limiting ages. | This is commonly the best type of data from which to recon- struct former glacier margins, perticularly if the physical characteristics and stratigraphy of the till sheets are carefully studied. |
| Marginal channels | Usually very difficult to date. | Least reliable under most circumstances because the origin of the channels is commonly debatable. |
| "Weathering zones" | Various weathering studies of landscapes, including weathering of bedrock and drift. In some cases, radiocarbon dates are used to date portions of young weathering zones. In general, the chronologic control of weathering zones is very poor. | Weathering zones have been widely employed to delimit late Wisconsin ice extent along the eastern seaboard of Canada. Unless moraines are preeent precisely at the contact between weathering zones, the method is suspect. This is because the weathering zones could as well result from differing basal ice conditions as from differing exposure intervals. |

Table 3. (continued)

| Type of Feature | Chronologic Control | Reliability |
|---|---|---|
| Change in surface | Chronologic control variable | Depends on field area, detail of mapping, and chronologic control of the morphologic break. |
| Distribution of radiocarbon dates, particularly those of samples that do not appear to have been over-ridden by glacier ice. | Radiocarbon dating | This method has been used widely in the Arctic, commonly in the absence of stratigraphic control to delimit late Wisconsin ice. Commonly, the inferences drawn from accepting these dates at face value are later shown to be incorrect. The samples turn out to be composed of mixed shells of different ages, redistributed driftwood, or lacustrine sediments contaminated with old carbon from various sources. This method is of questionable reliability, particularly when it is realized that available dates of this nature could be used to eliminate most Northern Hemisphere late Wisconsin ice sheets. |
| Emerged marine features | Radiocarbon dating | The distribution of emerged marine features is widely used in the Arctic to postulate the extent of late Wisconsin ice. The use of this feature depends critically on two factors: dating control and interpretation of response of the Earth's crust to glacial loading and unloading. In general, there are serious questions involved with both of these factors. Until these questions can be resolved, emerged marine features should be used with caution to infer former glacier extent. |

Table 4. Periglacial Features as Indicators of Past Climates

| Parameters Measured | Climatic Inferences | Length of Record | Reliability: Inferences of Climate | Reliability: Dating | Dating Methods |
|---|---|---|---|---|---|
| **Geomorphic Indicators of Past Climatic Regimes** | | | | | |
| Pingos (ice present) | MAAT* | mid-Wisconsin to present | | | Pollen $^{14}C$ |
| Closed system | < -5°C Permafrost present | | Good | Good | Geologic stratigraphic methods |
| Open system | < 0°C > -4°C Permafrost present | | | | |
| Pingo scars | | Wisconsin to present | Good | Good | Pollen $^{14}C$ |
| Closed system | < -5°C Permafrost present | | | | Geologic stratigraphic methods |
| Open system | < 0°C > -4°C Permafrost present | | | | |
| Ice wedges | | Wisconsin to present | | | |
| Active | < -6°C Permafrost present | | Good | Good | Presently growing |
| Dormant | < 0°C Permafrost present | | Good | Good | Pollen $^{14}C$ Geologic stratigraphic methods |
| Ice and sand wedge casts and polygons | < -6°C Permafrost present | Early Pleistocene to present (some pre-Cambrian reported) | Good | Good | Pollen $^{14}C$ Geologic stratigraphic methods |

Table 4. (continued)

| Parameters Measured | Climatic Inferences | Length of Record | Reliability: Inferences of Climate | Reliability: Dating | Dating Methods |
|---|---|---|---|---|---|
| Cryoplanation terraces | Required MAAT* at least < -10°C Mean summer temperature +2-6°C Permafrost present | Illinoian to present | Good | Good | Geologic stratigraphic methods |
| Rock glaciers | < 0°C Permafrost present | Wisconsin | Good | Good | Geologic stratigraphic methods |
| **Stratigraphic Indicators of Past Climatic Conditions** | | | | | |
| Permafrost relics (exist at depth) | < 0°C | Wisconsin | Good | Good | $^{14}C$, Geologic stratigraphic methods |
| Frozen carcasses | < 0°C from time of death to present | mid-Wisconsin to present | Good | Good | $^{14}C$ |
| **Geothermal Indicators of Paleoclimatic Variations** | | | | | |
| Geothermal profile of the ground | MAAT changes | Last 400 years | Good | Good | Geologic stratigraphic methods, $^{14}C$ |
| **Geochemical Indicators of Past Changes in Climate** | | | | | |
| Leaching of salts when frozen ground previously thawed | MAAT < 0°C | Wisconsin | Fair to good | Good | $^{14}C$ Geologic stratigraphic methods |
| **Cold Climate Features--Semiqualitative Temperature Data** | | | | | |
| Patterned ground, boulder or rock fields and streams, solifluction deposits | Indicative of a rigorous climate, but no definite temperature figures can be given | Illinoian to present | Fair | Good | Pollen $^{14}C$ Geologic stratigraphic methods |

All sampling intervals are discrete events.

*MAAT, mean annual air temperature.

centuries can be developed for some polar regions. Multivariate statistical techniques can then be applied to these indices, using recent weather records for calibration, to extract information about the interannual variations and longer-term changes of temperature and precipitation in the Arctic.

## REFERENCES

Andrews, J.T., and H. Nichols, 1981. Modern pollen deposition and Holocene paleotemperature reconstructions, central northern Canada. Arctic and Alpine Research, 13 (4), 387-408.

Andrews, J.T., P.J. Webber, and H. Nichols, 1979. A late Holocene pollen diagram from Pangnirtung Pass, Baffin Island, N.W.T., Canada. Review of Palaeobotany and Palynology, 27, 1-28.

Andrews, J.T., W.N. Mode, and P.T. Davis, 1980. Holocene climate based on pollen transfer function equations, eastern Canadian Arctic. Arctic and Alpine Research, 12, 41-64.

Andrews, J.T., P.T. Davis, W.N. Mode, H. Nichols, and S.K. Short, 1981. Relative departures in July temperatures in northern Canada for the past 6000 yr. Nature, 289, 164-167.

Berggren, W.A., et al., 1980. Towards a Quaternary time scale, Quarternary Research, 13(3), 277-302.

Berner, W., B. Stauffer, and H. Oeschger, 1978. Past atmospheric composition and climate, gas parameters measured on ice cores. Nature, 276, 3-55.

Berner, W., H. Oeschger, and B. Stauffer, 1980. Information on the $CO_2$ cycle from ice core studies. Proceedings of the 10th International Radio-carbon Conference, Bern and Heidelberg, August 1979. Radiocarbon, 22, 227-235.

Birks, H.J.B., 1973. Modern pollen rain studies in some Arctic and alpine environments. In J.J.B. Birks and R.G. West (eds.), Quarternary Plant Ecology (pp. 143-169). Blackwell, Oxford.

Brubaker, L.B., H.L. Garfinkel, and M.E. Edwards, 1983. A Late Wisconsin and Holocene vegetation history from the Central Brooks Range: implications for Alaskan palaeoecology. Quarternary Research, 20, 194-214.

Cragin, J.H., M.M. Herron, C.C. Langway, Jr., and G.A. Klouda, 1977. Interhemispheric comparison of changes in the composition of atmospheric precipitation during the Late Cenozoic era. Proceedings of the SCOR/SCAR Polar Oceans Conference, Montreal, May 1974 (pp. 617-631).

Dansgaard, W., 1953. The abundance of $^{18}O$ in atmospheric water and water vapor. Tellus, 5, 461-469.

Dansgaard, W., 1954. The $^{18}O$ abundance in fresh water. Geochimica et Cosmochimica Acta, 6, 241-260.

Dansgaard, W., 1964. Stable isotopes in precipitation. Tellus, 16, 436-468.

Dansgaard, W., 1977. Summary of activities at the Geophysical Isotope Laboratory, Copenhagen. IUGG International Symposium on Isotopes and Impurities in Snow and Ice, Grenoble, France, August 28-30, 1975, IAHS Publication 118 (pp. 401-407). International Association of Hydrological Sciences, Gentbrugge, Belgium.

Dansgaard, W., S.J. Johnsen, J. Moller, and C.C. Langway, Jr., 1969. One thousand centuries of climatic record from Camp Century on the Greenland Ice Sheet. Science, 166, 377-381.

Dansgaard, W., S.J. Johnsen, H.B. Clausen, and N. Gundestrup, 1973. Stable isotope glaciology. Meddelelser om Gronland, 197 (2), 53.

Dansgaard, W., S.J. Johnsen, N. Reeh, N. Gundestrup, H.B. Clausen, and C.U. Hammer, 1975. Climatic changes: Norsemen and modern man. Nature, 255 (5503), 24-28.

Delmas, R.J., J.M. Ascencio, and M. Legrand, 1980. Polar ice evidence that atmospheric $CO_2$ 20,000 years BP was 50% of present. Nature, 284, 155-157.

Epstein, S. and R.P. Sharp, 1959. Oxygen-isotope studies. Transactions of the American Geophysical Union, 40, 81-84.

Epstein, S., R.P. Sharp, and A.J. Gow, 1970. Antarctic Ice Sheet: stable isotope analyses of Byrd Station cores and interhemispheric climatic implications. Science, 168, 1570-1572.

Fredskild, B., 1969. A postglacial standard pollen diagram from Peary Land, North Greenland. Pollen et Spores, XI (3), 573-583.

Fredskild, B., 1973. Studies in the vegetational history of Greenland. Palaeobotanical investigations of some Holocene lake and bog deposits. Meddelelser om Gronland, 198 (4), 1-245.

Fredskild, B. and P. Wagner, 1974. Pollen and fragments of plant tissue in core samples from the Greenland Ice Cap. Boreas, 3, 105-108.

Fredskild, B., in press. Holocene pollen records from West Greenland. In J.T. Andrews (ed.), Quarternary Environments: Eastern Canadian Arctic, Baffin Bay and West Greenland. George Allen and Unwin, London.

Funder, S., 1979. Holocene stratigraphy and vegetation history in the Scoresby Sund area, East Greenland. Gronlands Geologiske Undersogelse Bulletin, 129, 1-66.

Gudmandsen, P., 1975. Layer echoes in polar ice sheets. Journal of Glaciology, 15 (73), 95-101.

Gudmandsen, P. and S. Overgaard, 1978. Establishment of time horizons in polar ice sheets by means of radio-echo sounding. Lyngby, Electromagnetic Institute, Technical University of Denmark.

Hammer, C.U., 1977. Past volcanism revealed by Greenland Ice Sheet impurities. Nature, 270 (5637), 482-486.

Hammer, C.U., 1979. Acidity of polar ice cores in relation to absolute dating, post volcanism and radio-echos. Journal of Glaciology, 25, 369-372.

Hammer, C.U., H.B. Clausen, W. Dansgaard, N. Gundestrup, S.J. Johnsen, and N. Reeh, 1978. Dating of Greenland ice cores by flow models, isotopes, volcanic debris and continental dust. Journal of Glaciology, 20 (82), 3-26.

Herron, M.M., 1980. The impact of volcanism on the chemical composition of Greenland Ice Sheet precipitation. Ph.D. thesis (158 pp.). State University of New York at Buffalo.

Herron, M.M., 1982. Glaciochemical dating techniques. Nuclear and Chemical Dating Techniques: Interpreting the Environmental Record. Symposium Series 176 (pp. 303-318). American Chemical Society, New York.

Herron, M.M. and C.C. Langway, Jr., 1979. Dating of Ross Ice Shelf cores by chemical analysis. Journal of Glaciology, 24 (90), 345-357.

Herron, M.M., C.C. Langway, Jr., H.V. Weiss, and J.H. Cragin, 1977. Atmospheric trace metals and sulphate in the Greenland Ice Sheet. Geochimica et Cosmochimica Acta, 41, 915-920.

Hyvarinen, H., 1972. Pollen-analytical evidence for Flandrian climatic change in Svalbard. Acta University, Series A, Geology, 1, 225-238.

Johnsen, S.J., 1977. Stable isotope profiles compared with temperature profiles in firn and with historical temperature records. IUGG International Symposium on Isotopes and Impurities in Snow and Ice, Grenoble, France, August 28-30, 1975. IAHS Publication 118 (pp. 388-392). International Association of Hydrological Sciences, Gentbrugge, Belgium.

Johnsen, S.J., W. Dansgaard, H.B. Clausen, and C.C. Langway, Jr., 1972. Oxygen isotope profiles through the Antarctic and Greenland Ice Sheets. Nature, 235 (5339), 429-434.

Kay, P.A., 1979. Multivariate statistical estimates of Holocene vegetation and climate change, forest-tundra transition zone, N.W.T. Canada. Quarternary Research, 11, 125-140.

Koerner, R.M., 1977. Devon Island ice cap: core stratigraphy and paleoclimate. Science, 196, 15-18.

Koerner, R.M., and D.A. Fisher, 1981. Studying climatic change from Canadian High Arctic ice cores. In C.R. Harington (ed.), Syllogeus, Volume 33 of Climate Change in Canada (pp. 195-218). National Museum of Man, Ottawa.

Koide, M. and E.D. Goldberg, 1971. Atmospheric sulfur and fossil fuel combustion. Journal of Geophysical Research, 76, 7689-7696.

Lachenbruch, A.H., 1966. Contraction theory of ice wedge polygons: a qualitative discussion. Permafrost International Conference, Proceedings. Publication 1287 (pp. 63-71). National Academy of Sciences/National Research Council, Washington, D.C.

Lachenbruch, A.H., G.W. Greene, and B.V. Marshall, 1966. Permafrost and geothermal regimes. In N.J. Wilimovsky (ed.), Environment of the Cape Thompson Region, Alaska (pp. 149-163). U.S. Atomic Energy Commission, Oak Ridge, Tennessee.

Langway, C.C., Jr., 1970. Stratigraphic analysis of a deep ice core from Greenland. Geological Society of America Special Paper, 125, 186 pp.

Langway, C.C., Jr., 1977. The development, scope and future of core drilling and ice core studies. IUGG International Symposium on Isotopes and Impurities in Snow and Ice, Grenoble, France, August 28-30, 1975. IAHS Publication 118 (p. 411). International Association of Hydrological Sciences, Gentbrugge, Belgium.

Langway, C.C., Jr., J.H. Cragin, G.A. Klouda, and M.M. Herron, 1977. Seasonal variations of chemical constituents in annual layers of Greenland deep ice deposits. U.S. Army CRREL Research Report 347, and IUGG International Symposium on Isotopes and Impurities in Snow and Ice, Grenoble, France, August 28-30, 1975. IAHS Publication 118 (pp. 302-306). International Association of Hydrological Sciences, Gentbrugge, Belgium.

Langway, C.C., Jr., H. Oeschger and W. Dansgaard, 1984. GISP 1971-1981: A summary. In The Greenland Ice Sheet Program 1984. Geophysical Mongraph Series. American Geophysical Union, Washington, D.C.

Lehmann, B., M. Wahlen, R. Zumbrunn, and H. Oeschger, 1977. Isotope analysis by infrared laser absorption spectroscopy. Applied Physics, 13, 153-158.

Lichti-Federovitch, S., 1974. Pollen analysis of surface snow from the Devon Island Ice Cap. Geological Survey of Canada Paper, 74-1 (Part A), 197-199.

Livingstone, D.A., 1955. Some pollen profiles from Arctic Alaska. Ecology, 36, 587-600.

Loosli, H.H. and H. Oeschger, 1968. Detection of $^{39}Ar$ in atmospheric argon. Earth and Planetary Science Letters, 5 (3), 191-198.

Mackay, J.R., 1978. Contemporary pingos: a discussion. Biuletyn Peryglacjalny, 27, 133-154.

Murozumi, M., T.J. Chow, and C. Patterson, 1969. Chemical concentrations of pollutant lead aerosols, terrestrial dusts and sea salts in Greenland and Antarctic snow strata. Geochimica et Cosmochimica Acta, 33, 1247-1294.

Nichols, H., 1975. Palynological and paleoclimatic study of the Late Quaternary displacement of the boreal forest-tundra ecotone in Keewatin and Mackenzie, N.W.T., Canada. Institute of Arctic and Alpine Research, Occasional Paper, 15, 1-87.

Nichols, H., P.M. Kelly, and J.T. Andrews, 1978. Holocene palaeo-wind evidence from palynology in Baffin Island. Nature, 273 (5658), 140-141.

Oeschger, H., 1977. Summary of work by the Physics Institute, Bern, on alpine and polar ice. IUGG International Symposium on Isotopes and Impurities in Snow and Ice, Grenoble, France, August 28-30, 1975. IAHS Publication 118 (pp. 408-410). International Association of Hydrological Sciences, Gentbrugge, Belgium.

Oeschger, H., B. Alder, H. Loosli, and C.C. Langway, Jr., 1966. Radiocarbon dating of ice. Earth and Planetary Science Letters, 1 (2), 49-54.

Oeschger, H., B. Alder, and C.C. Langway, Jr., 1967. An in situ gas-extraction system to radio-carbon date glacier ice. Journal of Glaciology, 6, 939-942.

Oeschger, H., U. Siegenthaler, U. Schotterer, and A. Gugelmann, 1975. A box diffusion model to study the carbon dioxide exchange in nature. Tellus, 27, 168-192.

Oeschger, H., B. Stauffer, P. Bucher, and M. Moell, 1976. Extraction of trace components from large quantities of ice in boreholes. Journal of Glaciology, 17 (75), 117-128.

Oeschger, H., B. Stauffer, P. Bucher, and H.H. Loosli, 1977. Extraction of gases and dissolved and particulate matter from ice in deep boreholes. IUGG International Symposium on Isotopes and Impurities in Snow and Ice, Grenoble, France, August 28-30, 1975. IAHS Publication 118 (pp. 307-311). International Association of Hydrological Sciences, Gentbrugge, Belgium.

Patterson, W.S.B., et al., 1977. An oxygen-isotope climatic record from the Devon Island ice cap. Nature 266, 508-511.

Peltier, W.R., and J.T. Andrews, 1976. Glacial-isostatic adjustment, I, The forward problem. Geophysical Journal of the Royal Astronomical Society, 46, 605-646.

Péwé, T.L., 1966. Ice-wedges in Alaska: classification, distribution, and climatic significance. Permafrost International Conference, Proceedings. Publication 1287 (pp. 76-81). National Academy of Sciences/National Research COuncil, Washington, D.C.

Péwé, T.L., 1975. Quaternary geology of Alaska. U.S. Geological Survey Professional Paper, 835, 145 pp.

Péwé, T.L., 1983a. The periglacial environment of the United States in Late Wisconsin time. In H.E. Wright, Jr., ed., Late Quaternary Environments of the United States. Volume 1 (pp. 157-189) in S.C. Porter (ed.), The Late Pleistocene, University of Minnesota Press, Minneapolis.

Péwé, T.L., 1983b. Alpine permafrost in the contiguous United States. Arctic and Alpine Research, 15 (2), 145-156.

Picciotto, E., X. DeMaere, and I. Friedman, 1960. Isotopic composition and temperature of formation of Antarctic snows. Nature, 187, 857-859.

Rampton, V., 1971. Late Quaternary vegetational and climatic history of the Snog-Klvane Area, Southwestern Yukon Territory, Canada. Bulletin, Geological Society of America, 83, 959-978.

Raynaud, D. and R. Delmas, 1977. Composition des gaz contenus dans la glace polairé. IUGG International Symposium on Isotopes and Impurities in Snow and Ice, Grenoble, France, August 28-30, 1975. IAHS Publication 118 (pp. 377-381). International Association of Hydrological Sciences, Gentbrugge, Belgium.

Reger, R.D., 1975. Cryoplanation terraces of interior and western Alaska. Ph.D. dissertation (326 pp.). Arizona State University, Tempe.

Reger, R.D. and T.L. Péwé, 1976. Cryoplanation terraces: indicators of a permafrost environment. Quaternary Research, 6, 99-109.

Ritchie, J.C., and F.K. Hare, 1971. Late-Quaternary vegetation and climate near the Arctic tree line of northwestern North America. Quaternary Research, 1 (3), 331-342.

Siegenthaler, U. and H. Oeschger, 1978. Predicting future atmospheric carbon dioxide levels. Science, 199, 388-395.

Siegenthaler, U., M. Heimann, and H. Oeschger, 1979. $^{14}C$ variations caused by changes in the global carbon cycle. Proceedings of the 10th International Radiocarbon Conference, Bern and Heidelberg, August 19-26, 1979.

Stauffer, B. and W. Berner, 1978. $CO_2$ in natural ice: Journal of Glaciology, 21 (85), 291-300.

Walker, D.A., S.K. Short, J.T. Andrews, and P.J. Webber, 1981. Late Holocene pollen and present-day vegetation, Prudhoe Bay and Atigun River, Alaskan North Slope. Arctic and Alpine Research, 13 (2), 153-172.

Weiss, H.V., M. Koide, and E.D. Goldberg, 1971a. Mercury in a Greenland Ice Sheet: evidence of a recent input by man. Science, 174, 692-694.

Weiss, H.V., M. Koide, and E.D. Goldberg, 1971b. Selenium and sulfur in a Greenland Ice Sheet. Science, 172, 261-263.

Weiss, H.V., K.K. Bertine, M. Koide, and E.D. Goldberg, 1975. The chemical composition of a Greenland glacier. Geochimica et Cosmochimica Acta, 39 (1), 1-10.

Windom, H.L., 1969. Atmospheric dust records in permafrost snowfields: implications to marine sedimentation. Geological Society of America Bulletin, 80, 761-782.

Young, S.B., 1971. The vascular flora of St. Lawrence Island with special reference to floristic zonation in the arctic regions. Contribution, Gray Herbarium, 201, 11-115.

# Appendix C.
# The Polar Regions: A Concern for the Future

C. R. Bentley, R. G. Barry,
J. H. Mercer, and T. J. Hughes

In any consideration of the evolution of present-day polar ice bodies and of their possible response to future climatic changes, it is important first to recognize that glacial systems are dynamic and behave in at least four fundamentally different ways. The first is stable steady state, wherein the system remains in the same state unless there is a change in some boundary condition. For a small change (e.g., a change in surface temperature) the response (e.g., change in marginal extent) is approximately linear and reversible. Therefore, doubling the change in the boundary condition will double the response, and returning the boundary condition to its original value will result in a return of the glacier to its original configuration. The response does not occur instantaneously; the time lag depends on the size of the glacier and can be many thousands of years for a large ice sheet.

The second, cyclic steady state, involves inherent cyclic processes whereby the extent of the ice changes in a periodic or pseudoperiodic way due entirely to internal causes, without a change in the externally imposed boundary conditions. In this case the state of the system is constantly changing, although the changes may be quite small most of the time. Both instantaneously and averaged over an interval that is long compared with a typical cyclic period, however, the system may be stable in the sense previously described. The primary example of such behavior is the phenomenon of glacial surging, known to occur in valley glaciers and recognized as at least a possibility in major ice sheets. (The effects of surges of the polar ice sheets on climate are reviewed in Appendix A.)

In the third category of behavior, unstable steady state, the system can continue without change if boundary conditions do not change, but the response to an external change is irreversible; the system will then move toward a new steady state position and will not return to its original value.

Transiency, the fourth category, refers to a dynamic state that is currently changing, without any current change in boundary conditions, and generally represents a transition toward a new steady state--although in reality, there is no guarantee that steady state behavior will ever be reached.

Of course, real glacial systems are complex and may involve combinations of types of behavior. They may be stable with respect to some boundary conditions and unstable with respect to others. Furthermore, the boundary conditions themselves may be dependent on the system response; for example, change in surface elevation of an ice sheet can change the surface temperature and mass balance.

A further complication and possible source of confusion arises from the consideration of numerical modeling calculations in the course of which the imposed boundary conditions are changed. It may be possible to change the behavior of the numerical model from, say, stable steady state to cyclic steady state (e.g., surging) by a particular change in boundary conditions, but it does not follow that the surging behavior is caused by the change in boundary conditions, since cycling can continue without further externally imposed change. The numerical situation is analogous to transforming an individual, real, nonsurging glacier to surging behavior, a transformation that has not yet been observed in nature. However, at least one example of the reverse transition from surging to nonsurging behavior is on record--the Vernagtferner in the Austrian Alps (Kruss & Smith 1982).

It is important to keep these basic distinctions in mind, particularly with respect to cyclic steady state and unstable steady state. The word "surge," commonly used in conjunction with or to describe a cyclic steady state of glaciers, has recently also been used to include the response of an ice sheet to a push from a position of unstable steady state. This use is unfortunate, particularly in a discussion of interaction with climate, because cyclic steady state changes proceed, to the extent that the whole cyclic process is stable, without or even despite changes in the environment, whereas displacement from unstable steady state occurs only because of changes in the environment. Because the mechanisms of cycling and unstable response may overlap in some cases, it is especially important to keep the terminology straight. Here, we will use "surge" for cyclic changes only.

The problem of long-term evolution of Arctic and Antarctic ice bodies and their possible responses to future climatic trends can (and should) be attacked in two quite different ways: by examining the empirical evidence of conditions during past climatic epochs, and by theoretical analyses of the interactions among atmosphere, ice, and ocean.

Several aspects of climatic change will be considered here. The most obvious are those relating directly to the atmosphere, such as air temperature, precipitation rates, and cloudiness. Equally or even more important for some mechanisms are the oceanic aspects: water temperature, currents, and sea level. Third, there are those aspects related to the cryosphere, especially the extent and thickness of the sea-ice cover, which may interact directly with the margins of the more permanent ice.

We will lcok first at specific response mechanisms and then consider the evidence for past and present changes and the expectations for the future. The discussions of sea-ice cover and grounded ice sheets, which must be separated because of quite different mechanisms and time scales, are presented in parallel.

## 1. RESPONSE MECHANISMS

### 1.1 Sea-Ice Cover

Most model studies of ice-ocean-atmosphere interaction have been concerned with the response of the atmosphere to changes in the extent of the ice cover. The problem of initial interest here, however, is the converse question: how does the sea ice respond to different atmospheric and oceanic conditions? Briefly, the answer is that the response could be dramatic but that we really do not yet know. For example, two simplified models of the Arctic sea ice suggest that 5 years after a 5°C warming a steady state would be reached in which the Arctic Ocean is ice-free in the summer, with the ice reforming each winter, whereas other models of the same warming point to a continuation of permanent, year-round ice cover. However, the geological record of sediments suggests that a more or less permanent sea-ice cover existed throughout the globally warm interglacial period of the Pleistocene (Clark 1983).

The actual mechanisms involved in the interaction include the energy transfer among atmosphere, ice, and ocean; the effects of winds and currents; and the impact of melting and freezing of the ice and of freshwater runoff from land on the salinity and temperature structure of the ocean. An ice cover insulates the atmosphere from the ocean, although there are still large net transfers to the air through open leads that form due to ice motion and through young ice. The ice and its overlying snow cover reflect about 80 percent of incoming radiation in the autumn and spring until snowmelt, puddle formation, and ice breakup begin to modify the surface energy budget in June-July (in the Arctic) by a threefold to fourfold increase in the absorbed solar radiation. Data on such fundamental features as snow depth on the ice or the spatial extent of melt ponds and open-water areas in the Arctic Ocean are limited (Barry 1983). In Antarctica there is very little melt of the snow on the sea ice, but the summer disappearance of the ice is rapid. The possible contribution of ocean heat flux and of wind and wave effects on ice melt requires additional study. The melt process greatly modifies the summer climate in the Arctic, leading to high average cloudiness. A major outlet for Arctic ice is the East Greenland Current system, which may remove annually the equivalent of a 60-cm layer of ice from the Arctic Ocean (Maykut 1982).

It is not possible to isolate the dynamic and thermodynamic effects of the atmosphere on sea ice, or vice versa (Allison 1982). Ice responses to the atmosphere are comparable to atmospheric responses to ice extent during the ice-growth season in the Arctic, with lags of several months in each case. Thus, short-term (weekly or monthly) ice anomalies can be attributed to wind and atmospheric temperature effects, but there are also indications that more extensive ice increases the frequency of cyclone formation, at least regionally. Longer-term changes in ice extent may involve changes in ocean heat transport, but data are not yet sufficient for reliable tests of this possibility.

Given a sustained reduction in ice extent, through natural variation or externally induced (for example, by $CO_2$ warming effects or the proposed diversion of part of the flow of Siberian rivers that now enter the Arctic Ocean), substantial modifications of the circulation and climate may be expected in high and middle latitudes. Preliminary modeling experiments indicate that cyclone tracks would shift northward, perhaps reducing precipitation in Europe. The intensity of the subpolar, low-pressure centers and large-scale poleward energy transports by the atmosphere might also be reduced, although ocean responses to such changes have not yet been taken into account.

## 1.2 Snow Cover

Only since the mid-1960s has information been available on extent of snow cover and its variability on a global scale. Relationships between amounts of snowfall and winter temperature anomalies have been little explored. Milder winter months tend to be more snowy in some areas of the Sub-Arctic, yet an extensive snow cover is usually produced by cyclonic precipitation in a large-scale outbreak of polar air. Duration of snow cover is primarily determined by spring and summer temperatures. The melt period lasts only about 10 days on the Arctic tundra, but from 3 to 4 weeks on the Arctic pack ice, where air temperatures cannot rise significantly above 0°C.

There are two principal climatic effects of a snow cover (Kukla 1981). The high reflectivity of snow cover for solar radiation (80-90 percent for new snow, 60 percent for old, melting snow) and its low thermal conductivity, isolating the underlying surface, cause temperatures to be low over snow cover. Snow exerts a local cooling effect on air crossing it, and an extensive snow cover modifies the atmospheric circulation by establishing or reinforcing an upper trough pattern in the westerly wind circulation. This pattern tends to steer cyclone paths and displace the zones of cyclogenesis equatorward, particularly in the colder seasons. Hence, there are complex feedbacks between the atmosphere and a snow cover; little analytical or modeling work on such interactions has yet been carried out. The only modeling study of these effects is described by Williams (1975) for an equilibrium general circulation model experiment. Any overall warming trends, due to $CO_2$, atmospheric aerosols, or other factors, will be likely to lengthen the snow-free season and thereby substantially modify the surface energy and moisture budgets, especially over land.

## 1.3 Ice Sheets and Glaciers

In considering the response of glaciers and ice sheets to climatic change, one must first define the terrestrial ice bodies (including Arctic glaciers and large parts of the Greenland and East Antarctic Ice Sheets), which rest on a substrate close to or above sea level, and marine ice sheets (principally the major part of the West Antarctic Ice Sheet), which are ice grounded below sea level, and the floating ice shelves. The distinction between terrestrial and marine ice sheets is somewhat diffuse. Both Greenland and East Antarctica also have large areas of deep bedrock, whereas West Antarctica features two extensive regions of elevated bedrock. However, the distinction is made because fully terrestrial ice bodies only interact directly with the atmosphere, so probably change their form and extent rather slowly in response to climatic changes, whereas marine ice sheets also interact directly with the ocean and therefore may be sensitive to, and capable of rapid response to, changes in sea level, ocean temperature, and sea-ice cover. The significance of this distinction is a matter of vigorous debate among glaciologists; nevertheless, it must be understood, for it is fundamental to arguments that propose the possibility of catastrophic instability.

In the following sections we will, therefore, first consider terrestrial ice sheets, ignoring smaller ice caps and glaciers that can have only a minor effect on world climate, and then marine ice sheets, comprising the grounded inland ice and the adjoining ice shelves.

1.3.1 Terrestrial ice sheets. The term "ice sheet" implies ice flow from a high central region. In a series of papers, Weertman (1976, and references therein) first showed that an idealized, two-dimensional ice sheet resting on uniform rock above sea level is in stable steady state, if the position of the snowline, which divides the interior zone of net mass gain from the peripheral zone (if any) in which melting predominates, depends primarily on the distance from the center of the ice sheet, but that the ice sheet is unstable if the position of the snowline depends primarily on elevation above the ice-sheet margin. He then developed a model that incorporated both possibilities and examined the effect of changing the surface mass balance, the extent of a basal meltwater layer, and the strength of the solar radiation on the size and rates of growth or decay of terrestrial ice sheets. A common feature of his and similar models is that very large snow accumulation rates would have been needed to produce growth rates of the Pleistocene ice sheets that correspond with the rates of sea-level lowering indicated by studies of seafloor sediments, and that unreasonably large melt rates would have been required to explain the rapid retreat of the Laurentide Ice Sheet. This last point has suggested to one group of glaciologists that much more rapidly reacting marine ice sheets, particularly in and around Hudson Bay, played a crucial role in ice-sheet growth and retreat (Denton & Hughes 1981).

Quite a different point of view is held by another group. Using a "frictional lubrication model" for basal sliding that includes the important factor of warming and melting of the ice by internal friction, particularly at and near the glacial bed, they find that rapid flow can occur naturally in any ice sheet (Budd & Jenssen 1975), that large ice sheets can exhibit cyclic surge behavior, and that the growth and decay history of the Laurentide Ice Sheet in response to variations in the earth's orbital parameters can be simulated without recourse to marine ice sheets or anomalously large melt rates (Budd & Smith 1979).

1.3.2 Marine ice sheets. The particular importance of marine ice sheets is suggested by a theoretical analysis showing that a simple two-dimensional model ice sheet resting on an initially flat bed below sea level would be inherently unstable (Weertman 1974). Depending on whether the depth of the bed below sea level was more or less than a certain critical value (on the order of 200 m), the ice sheet would either shrink until it disappeared, or grow until its edges reached deep water. The reason is that for a deep enough bed, the marginal portions of the ice sheet would tend to float, with a consequent flattening out and spreading over the water. This occurrence would increase the surface slope upstream, accelerating the flow of the ice, with consequent continued thinning and retreat of the "grounding line" that separates the floating and grounded portions of the ice sheet.

This idealized marine ice sheet can be stable if the bed slopes downward away from the center of the ice sheet, but if the bed slopes the other way, the instability is accentuated. It has also been argued that restraints on an ice-shelf portion might produce stability, particularly if the movement of the ice shelf is constrained by lateral boundaries, or pinned by islands and shoals.

The dependence of growth or shrinkage on the depth of the bed below sea level is of critical importance, since, if this model is even qualitatively correct, it provides a mode of direct response of a marine ice sheet to externally imposed sea-level changes. Such changes could be of particular importance if there were a relatively shallow sill or shoal near the margin of an extended marine ice sheet. In that case, if sea level should rise sufficiently to produce a bed depth exceeding the critical one on the sill, unstable retreat of the ice sheet would then follow as the grounding line moved back into regions of deeper water behind the sill. Such shoals, currently found along the edge of the continental shelf in the Ross Sea, may have marked one margin of the late Pleistocene West Antarctic Ice Sheet.

If, instead of a simple two-dimensional ice sheet, one considers a three-dimensional model, then ice streams (fast-flowing "glaciers" within ice sheets) should be included, because their very high activity compared to the relatively sluggish sheet flow elsewhere makes it likely that they will play an important role in any response of the ice. Ice streams occur in terrestrial as well as marine ice sheets, and some progress has been made in modeling an ice stream as an individual unit. However, attempts to model the dynamics of ice sheets that include ice streams are just beginning. Ultimately, an adequate model must relate movement of the ice-sheet grounding line, particularly in the ice streams, to such factors as downdraw of the ice-stream drainage basins, behavior of any fringing ice shelves, rates of growth or shrinkage of other ice sheets, processes of subglacial deposition and erosion (particularly at the grounding line), and response of the earth's mantle to crustal loads.

One of the three-dimensional model studies that have been made of the retreat of marine ice sheets in response to rise in sea level at the beginning of the Holocene supports the suggestion that a marine ice sheet depends for its existence on the presence of the ice shelves that surround it. The vulnerability of marine ice sheets to climatic warming would then depend on the vulnerability of their ice shelves. If so, and if rising temperatures, especially of the surrounding ocean water, were to lead to the disappearance of the ice shelves, then rapid shrinkage of the marine ice sheet would follow. However, in another three-dimensional model simulation (Budd & Smith 1982), the rise in sea level made the ice sheet retreat gradually to the high ground of West Antarctica without any dramatic intervention by the (parameterized) ice shelves. Moreover, there are widely divergent opinions on the speed with which such ice shelves as those of the West Antarctic would respond to temperature changes; the presence or absence of sea-ice cover could also be a factor in speed of response.

It is important to make clear that the theoretical models of marine ice sheets discussed in this section include inherent instability as a feature of the model, so that they cannot be used to prove, or even test, the basic concept. The same is true of the frictional lubrication model (see Section 1.2.1); it includes an inherent instability stemming from the regenerative friction-lubrication feedback.

That the deglaciation of a marine ice sheet can proceed rapidly is suggested by one reconstruction of the "gutting" of the remnant of the North American Laurentide Ice Sheet that depicts its central portion, grounded below sea level in Hudson Bay, as disappearing in no more than about 200 years. The analogy cannot be pushed too far, however, because of fundamental differences between the late-glacial Hudson Bay ice and the present West Antarctic Ice Sheet.

## 2. EVIDENCE FOR PAST CHANGES

The record of the interplay of past changes in the extent of sea ice, the volume of land ice, and climate, offers possible analogs of future events. The most likely outcome of human activities is climatic warming (as discussed in Appendix A) so that the state of the cryosphere during past intervals that were warmer than the present is of particular interest. (The techniques for reconstructing the climatic history of the polar regions are explained in Appendix B.)

## 2.1 Sea Ice

Empirical evidence of sea-ice extent during the Pleistocene and Holocene epochs, reviewed by Barry (in press), is limited because sedimentological indicators of ice cover and the nature of their seasonal response are not reliably established. The difference between terrestrial debris rafted by ice floes or by icebergs is also uncertain, and planktonic marine organisms from ocean cores provide only a broad indication of ice extent.

If we allow for those and other uncertainties, however, we can infer a limit of icebergs at about 50°N in the North Atlantic at the last Glacial Maximum 18,000 years ago. The sedimentary record in the Arctic Ocean indicates the general presence of a sea-ice cover, similar to the present one, for at least the last 700,000 years, although glacial/interglacial regimes during this interval may have included alterations between permanent pack-ice cover and a seasonally partially ice-free ocean.

For Antarctica, the sedimentary and marine isotopic records indicate that the Ross Ice Shelf, and presumably a seasonal sea-ice cover, formed in the Late Miocene (about 15 million years ago), but with subsequent warm and cold intervals (Frakes 1978). The sedimentary record further suggests that during the last glaciation (between about 100,000 and 15,000 years ago) the sea ice in spring extended substantially farther north than at present (Burckle et al. 1982).

## 2.2 Snow Cover

There is little evidence specifically on snow cover during the Pleistocene and Holocene, except as it relates to changes in the altitude of the permanent snowline and the extents of ice sheets and other glaciers. Snowlines in many areas were lowered generally between 600 and 900 m during the last Glacial Maximum some 18,000 years ago, when summer temperatures over the northern continents fell by at least 5°C. During the so-called Little Ice Age, from the seventeenth to nineteenth centuries, snowlines in Europe and other alpine areas were some

100-200 m lower than today, representing about 15 percent of the full glacial depression (Porter 1981).

Historical snow-cover records for Great Britain over the last 300 years show that the duration of snow on the ground was at a minimum in the 1920s and 1930s, corresponding to the period of highest temperatures. However, records at several stations in Europe since the 1890s show that there are regional variations in the timing of the decades when maximum and minimum durations of snow cover occur. Further compilation of available historical records would help to shed light on the space and time scales of snow-cover variability (Barry, in press).

## 2.3 Ice Sheets

Analyses of stable-isotope ratios of microfossils in deep-sea cores have been interpreted in terms of global ice volume on land. As discussed in Appendix B, they show that a large ice sheet, inferred to have been the land-based East Antarctic Ice Sheet, accumulated between about 15.3 and 13.3 million years ago and has remained in place ever since. Probably the stage was set for the formation of this ice sheet by the thermal isolation of Antarctica that followed the opening of Drake Passage and the resultant inception of the circum-Antarctic Current at some time during the Early Miocene, which must have greatly reduced the flow of warm ocean water to the margins of the Antarctic continent.

The history of the West Antarctic Ice Sheet, however, may have been very different. It must have formed by some combination of thickening sea ice and coalescence and thickening of initial ice shelves growing out from glaciers in the mountainous regions of Marie Byrd and Ellsworth Lands, and from the Ellsworth and Transantarctic Mountains, across the sea that lay between. Because the ice shelves had to grow at sea level, the ice sheet presumably could not have formed until summer temperatures were close to freezing at sea level, so it may have formed substantially later than the East Antarctic Ice Sheet. Conversely, during a time of rising temperatures it might have started to shrink when the climate at its margins became too warm for cold ice at sea level (summer temperatures perhaps about 5°C above today's), and, since it might have been able to respond much more quickly and more profoundly to a short interval of unfavorable climate than could the terrestrial East Antarctic Ice Sheet, it may have been temporarily absent during exceptionally warm intervals after its initial formation.

The Greenland Ice Sheet, like the East Antarctic Ice Sheet, is mainly land based. Although smaller than the West Antarctic Ice Sheet, its melting would produce about the same rise in sea level because virtually none of it is displacing ocean water. Little is known about the past history of the Greenland Ice Sheet, neither when it first formed, nor the extent of its depletion during subsequent warm intervals. During the present interglacial, deglaciation has been most extensive in the southwest part, where the ice margin receded about 100 km during the Hypsithermal Interval after 9500 B.P. Since 4800 B.P. the ice margin has oscillated within a few kilometers of its present position (Ten Brink & Weidick 1974). This modest loss of volume of the climatically most vulnerable part of the Greenland Ice Sheet during the warmest part of the present interglacial suggests that the ice sheet is unlikely to have been seriously depleted during any previous warmer interglacial, unless the warmth was very prolonged.

From the foregoing some glaciologists conclude that under a climate moderately warmer than that of today, the East Antarctic Ice Sheet would remain virtually unaffected. The West Antarctic Ice Sheet also would be unaffected until warming exceeded a critical level, above which it would be vulnerable to a relatively rapid reduction in extent. The Greenland Ice Sheet would undergo slow depletion, its extent depending on the duration of warmth.

On the other hand, calculations using the frictional lubrication model suggest that the response of the bulk of East Antarctic and Greenland ice may be more dynamic and that, especially in Greenland, melting could have contributed equally or even predominantly to past rises in sea level. One potential basis for a judgment on these views is better knowledge of the climatic record of the last interglacial that ended about 100,000 B.P., deduced from biological indicators and from inferred changes in ice volume. That interglacial appears to have been the only occasion during the last three fourths of a million years or so when sea level was higher than it is today, suggesting a lesser global ice volume on land.

## 2.4 Climate of the Last Interglacial

2.4.1 Evidence from biological changes. The last interglacial began about 125,000 years ago and lasted no more than about 15,000 years, perhaps considerably less. Complete stratigraphic records are preserved in some ocean cores, possibly in the ice of East Antarctica (although this has not yet been demonstrated), and in some terrestrial accumulations of peat and lake sediments. Ocean cores provide information on ocean paleotemperatures and on the volume of terrestrially stored ice. Fossil coral reefs, especially on uplifted coasts, provide information about paleo-sea levels, which give an independent check on the inferred ice-volume signal from ocean cores; furthermore, suitable coral specimens are amenable to radiometric dating techniques, which form the basis for the estimated age of the last interglacial. A few terrestrial deposits containing proxy paleoclimatic data are believed to cover the entire interglacial, but suitable dating techniques are hard to apply to deposits of their age and composition.

Because of predictions that future climatic warming in response to $CO_2$ increases will be greater at the poles than at the equator, information about past latitudinal differences in temperature trends is of interest. What little is known about the climate of the last interglacial in tropical areas suggests that the climate then was similar to today's. For example, in the Caribbean, surface water temperatures at the maximum of the interglacial probably were little higher than at present. Most studies of terrestrial deposits during the last interglacial age have been done in middle latitudes of the Northern Hemisphere, where they generally show that the interglacial summers were somewhat warmer than they are today: by 2-3°C in Western Europe and at least 3°C in southern Ontario. Information from the Arctic is consistent with a poleward magnification of temperature changes, but the data are too scant and ambiguous to confirm it. The temperatures reached in the Northern Hemisphere during the warmest part of the last interglacial seem to have been somewhat less than models suggest would follow a doubling of $CO_2$ levels (see Section 4).

Evidence about climatic conditions in Antarctica during the last interglacial is both direct and indirect. The sole direct evidence comes from Taylor Valley, Victoria Land (latitude 78°S), where algal limestones were deposited in enlarged lakes between 130,000 and 90,000 years ago (according to $^{234}U/^{230}Th$ dating) (Hendy et al. 1979). The enlarged lakes are believed to imply greater warmth than today, but no estimate has been made of the actual temperature reached. Indirect evidence comes from changes in ocean volume and world sea level during the last interglacial.

2.4.2 Evidence from changes in sea level and ice volume. Two independent lines of evidence point to great ocean volume--and thus, by implication, to smaller ice-sheet volume--during the last interglacial: raised strandlines and low oxygen isotopic ratios of benthic foraminifera. Raised strandlines at elevations of 5 $\pm$ 3 m on coasts thought to be more or less stable (e.g., Bahamas and Australia) suggest greater ocean volume at the time they were formed, but they are not completely conclusive evidence because other mechanisms, such as changes in the ocean bed of the geoid, can be invoked. Oxygen isotopic composition of formaminifera in ocean cores and of molluscs and corals from the strandline deposits, however, also suggest that less glacier ice and, therefore, more ocean water by a few meters of sea level was present during the height of the last interglacial than today. The additional water was about what would have been added by the melting of either the entire West Antarctic or Greenland Ice Sheets, or by contributions from all the ice sheets.

If the West Antarctic Ice Sheet was absent for a few thousand years during the warmest part of the last interglacial, the most likely reason is that Antarctic temperatures, especially of the near-surface ocean, had risen too high for its survival on the low-level bedrock between the Ellsworth and Marie Byrd Land mountains. The possibility of a major surge of the Antarctic Ice Sheet has also been suggested, (Hollin 1980), although that would require either a fortuitous coincidence or a causal link between a cyclic phenomenon with the thermal maximum of the interglacial.

## 3. NATURAL CHANGES

### 3.1 Sea Ice

The seasonal and extreme extents of sea ice in both hemispheres are shown in Figures 1 and 2. In the Southern

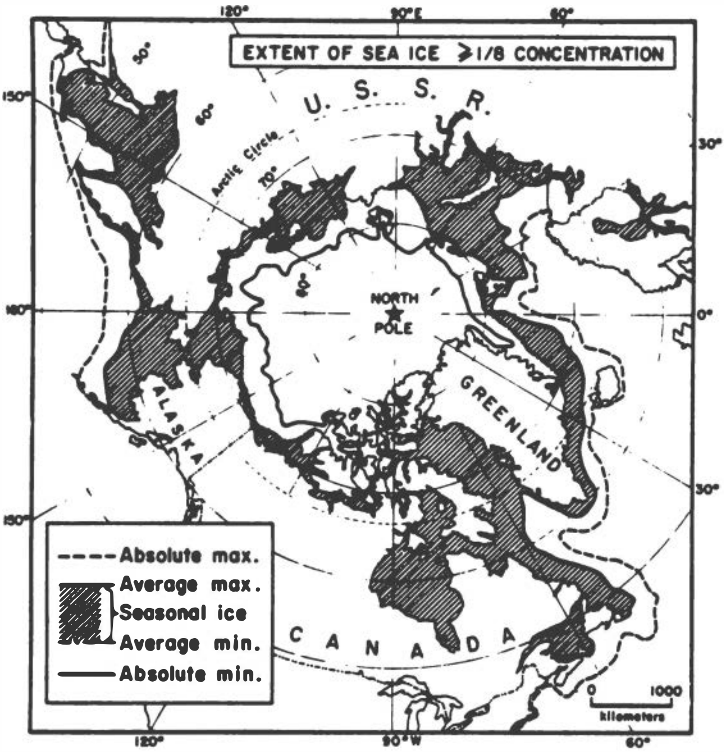


**Figure 1. Northern Hemisphere seasonal sea-ice zone (from Barry 1980).**

Hemisphere, at the seasonal maximum in September, the sea ice extends over 20 x $10^6$ $km^2$, or 8 percent of the hemisphere. However, adjusted for the actual area fully covered by ice, the figure is 15 x $10^6$ $km^2$ (Zwally et al. 1983). At the minimum extent in March, the ice affects 2.5 x $10^6$ $km^2$, giving a seasonal range of 8 (or 6):1. In contrast, the seasonal range in the Northern Hemisphere is only 2:1 (14 x $10^6$ $km^2$ and 7 x $10^6$ $km^2$) (Walsh & Johnson 1979). The pattern of seasonal expansion and contraction is fairly regular around Antarctica, where there is an open ocean, but in the Northern Hemisphere the distribution of land causes marked longitudinal asymmetries in the seasonal sea-ice zones. For example, in late winter there is ice to the south of 50°N in the Sea of Okhotsk and the Labrador Sea, whereas open water occurs north of 75°N in the Norwegian-Barents sea.

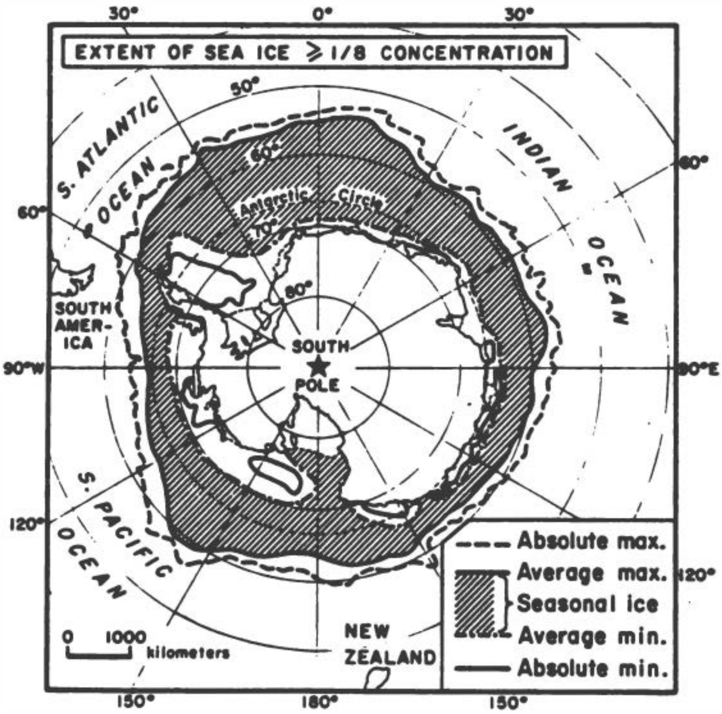


**Figure 2. Southern Hemisphere seasonal sea-ice zone. Ice limits for 1971-1976. (From Barry 1980.)**

Annual anomalies in ice extent are known to occur on a regional scale, with substantial antiphase behavior between different sectors; that is, more extensive ice in the Alaskan sector is frequently associated with less ice in the Baffin Bay-Greenland Sea sector, and vice versa. There is similar antiphase behavior between the Ross Sea and Weddell Sea ice in Antarctica. Longer (decadal) trends occur in total Arctic ice extent. There was an increase in mean ice area in the late 1960s and early 1970s compared, for example, with the late 1950s or late 1970s. These trends show a broad similarity with hemispheric temperature fluctuations, although regional ice anomalies are primarily determined by anomalous atmospheric circulation patterns. Information for Antarctica is less certain due to the limited number of sectors where ice conditions were regularly observed.

Representative ice thicknesses in the Arctic are poorly known because of the mixture of first-year and multiyear floes, areas of open water, and pressure ridges. The mean thickness ranges generally from about 2 m on the Siberian side of the basin to from 4 to 6 m off the Canadian Arctic Archipelago, where pressure-ridging is most common and intensive (Hibler 1981). Even less is known abut the thickness of Antarctic sea ice, but, since most of the ice is less than 1 year old, the mean thickness is surely less than 2 m. Submarine sonar measurements and satellite microwave radiometry mapping of the age structure of the pack ice should eventually provide such data, but at the present time we are unable to determine reliably whether there are trends in net ice production and mean equilibrium thickness.

## 3.2 Snow Cover

For the winter months in the Northern Hemisphere, snow cover on land is very much a mid-latitude phenomenon (Figure 3). Satellite data reveal that at its maximum, snow covers about 60 x $10^6$ $km^2$ (24 percent) of the Northern Hemisphere, but there is interannual variability, particularly over Eurasia (Matson & Wiesnet 1981). Interannual variability in the location of the snow boundary over Eurasia is largest in the transition seasons.

The short record length has restricted analytical studies of the interaction between snow cover and climate. However, recent analyses of North American records shows that snow cover contributes some 10-20 percent of the variance of the concurrent monthly temperature in a broad belt across the United States centered around the snow margin (Walsh et al. 1982). Anomalies of North American snow extent modify the atmospheric long-wave structure over about one third of the hemisphere, a fact that illustrates the complexity of the interaction process.

Figure 3. World snow cover showing duration in months. The small montane areas depicted in the southern continents have annual snow covers of varying duration. The stippled areas are ice sheets with snow cover. Snow-covered sea ice is omitted (modified after Rikter, from a map published by Environment Canada, Inland Waters Directorate).

### 3.3 Ice Sheets

3.3.1 Greenland. It is still not known whether the total mass of the Greenland Ice Sheet is increasing, decreasing, or unchanging. Although the mass input in the form of snow accumulation on the surface is fairly well measured, the rate of ice loss is uncertain. Iceberg discharge is reliably known only for west coast outlet glaciers and even there only for the summer. LANDSAT imagery of the Jacobshavn Glacier, which accounts for 40 percent of the discharge from these outlet glaciers, suggests that discharge is considerably reduced during the winter months. Summer meltwater discharged beneath the glaciers is substantial but unmeasured. It is quite possible that the Greenland Ice Sheet is presently growing in the west and shrinking in the east (Malzer & Seckel 1976). However, the average rate of growth or shrinkage, if any, is very unlikely to exceed 0.01 percent of the total ice mass per year; dramatic changes from natural causes do not appear to be occurring.

3.3.2 East Antarctica. Studies of the mass balance of the East Antarctic Ice Sheet suggest a mass increase over most of the interior regions of an amount that corresponds to about a 1 mm $yr^{-1}$ drop in world sea-level change, yet actual sea-level measurements show a rise of about that amount. The coastal regions have much higher snow input and output rates (per unit area) than the interior, so that, despite a lesser areal extent, they could easily be undergoing a net loss twice as large as the interior gain. Evidence for dramatic rates of surface lowering, as large as 1 m $yr^{-1}$, has been reported in certain coastal areas, yet in one of these same areas, as in several others, the measured mass balance is strongly positive, which should mean surface buildup. These results suggest interior buildup after a surge, although, of course, their correct interpretation is strongly debated. In any case, as in Greenland so a fortiori in East Antarctica, the percentage change in the total mass is very slow--at most one part in $10^5$ $yr^{-1}$. It should be noted, however, that a substantial fraction of the East Antarctic Ice Sheet lies on bedrock far below sea level and could behave like marine ice sheets where a sufficiently deep connection to the ocean exists.

3.3.3 West Antarctica. The present mass balance of the West Antarctica Ice Sheet is unknown, but it is the subject of widespread interest because of the unstable behavior of which deep-bedded marine ice sheets are probably capable. Most of the West Antarctic Ice Sheet is marine and deep bedded--if the ice were suddenly removed, the Weddell, Amundsen, and Ross Seas would all be openly connected by water at least 500 m deep, much of it deeper than 1000 m. Conclusions drawn from analyses of both the current state and that over the last 20,000 years, which encompass the end of the last glacial epoch, have differed in the extreme from no major changes at all to drastic shrinkage even at present; however, a growing preponderance of field evidence indicates that major shrinkage has occurred.

A substantial body of evidence suggests that the grounded ice-sheet margin has retreated in the last 12,000 years or so from a position near the edge of the continental shelf in the Ross Sea (Stuiver et al. 1981). On the other hand, some investigators interpret the sedimentary record in the Ross Sea as showing that no grounded ice sheet has overlain the area at all in the last 20,000 years (Drewry 1979), and analysis of the shape of internal (radar-reflecting) layers within the central West Antarctic Ice Sheet suggests that no major change in form has taken place at that site in the last 10,000 years (Whillans 1976).

Evidence relating to present-day changes is also ambiguous. Field evidence from the Ross Ice Shelf has been interpreted as showing that the ice shelf is at present thickening in some areas near the grounding line; however, recent evidence suggests instead that the ice shelf is currently in a nearly steady state but has thinned in the last few hundred years. Local mass-balance studies near the center of the inland ice sheet show a slow thinning (~ 30 mm of ice per year), part or all of which may be attributable to the slow penetration of a surface warming at the end of the last glacial epoch. On the other hand, calculations using the latest data continue to indicate a strongly positive mass balance for the whole part of the West Antarctic Ice Sheet that flows into the Ross Ice Shelf. These findings again suggest postsurge conditions (Bentley & Jezek 1982). Preliminary numerical studies employing the frictional lubrication model suggest that either an oscillatory regime, with the ice sheet currently in a postsurge state, or a steady state, fast-flow regime, may be appropriate.

In its extreme form the drastic-shrinkage model for the West Antarctic leads to a sea-level rise of about 5 m in as little as 200 years (Hughes 1973; Mercer 1978). However, a careful numerical assessment by Bentley (quoted in Revelle 1983) has increased the minimum time to about 500 years, although other model simulations (Budd & Smith 1982) suggest much more gradual changes. Nevertheless, because such a rise in sea level would have disastrous consequences for the coastal areas of the world, the more extreme model predictions must be considered seriously. They are not literally applicable at present, for the simple reason that sea level is rising an order of magnitude more slowly than the implied rate. The problem continues to be studied with a variety of models that can be expected to produce firmer projections soon. An essential contribution will come from additional observations in West Antarctica. Whether the ice sheet is currently poised in unstable steady state is an open question. Field work aimed at determining the present situation continues and could be aided immeasurably by the launching of a polar-orbiting satellite carrying a special ice-sensing and topographic surveying instrument package (Science and Applications Working Group 1979)--though one such satellite (NOSS), originally planned for the mid-1980s, was deferred indefinitely due to the 1982 budget reduction.

3.3.4 A hypothetical scenario for the West Antarctic ice sheet. With the drastic-shrinkage model, a hypothetical scenario for the past and future of the West Antarctic Ice Sheet can be constructed. As the earth's climate warmed from the ice-age conditions that peaked between 22,000 and 18,000 years ago, the marine portion of the West Antarctic Ice Sheet shrank from its maximum extent, one that covered the entire Antarctic continental shelf north of the Transantarctic Mountains, to the present ice sheet that is grounded only over the Byrd Subglacial Basin. The present fringe of floating ice shelves, notably the Ross Ice Shelf in the Ross Sea and the Filchner-Ronne Ice Shelf in the Weddell Sea, formed in the process. Shrinkage was initially rapid, because the scarcity of islands and seamounts beyond the Antarctic continental shelf prevented widespread formation of pinned ice shelves. Shrinkage slowed as the ice-stream grounding lines retreated across the Antarctic continental shelf, creating large embayments in the ice sheet where ice-stream tongues merge to create the present ice shelves.

At present, West Antarctic surface flowlines radiate from three major ice domes. Most converge on 15 major ice streams, which drain about 90 percent of the grounded ice sheet and, except for the two that find their coastal expressions in Thwaites Glacier and Pine Island Glacier, merge to form the Ross and Filchner-Ronne Ice Shelves. Ice tongues forming from Thwaites and Pine Island Glaciers calve into the open water of Pine Island Bay.

In this scenario of what might happen during a polar warming trend, since there is now no ice shelf in the Amundsen Sea to buttress these ice streams, the grounding lines would irreversibly retreat into the vast Byrd Subglacial Basin that underlies most of the present grounded West Antarctic Ice Sheet. Grounding-line retreat would be accompanied by lowering surface elevations in the drainage basins for Thwaites and Pine Island Glaciers. This, in turn, would cause the ice divide of the Amundsen Sea sector of the West Antarctic Ice Sheet to migrate inland, shrinking the Ross Sea and

Weddell Sea sectors of the ice sheet. Reducing the volume of ice in these sectors would reduce the velocities of the ice streams supplying the present Ross and Ronne Ice Shelves. The ice shelves would then acquire a strongly negative mass balance, and their grounding lines would begin to retreat up the ice streams. In this sequence of events, shrinkage of the Amundsen Sea sector leads to shrinkage of the Ross Sea and Weddell Sea sectors as well--the end result is the elimination of the entire marine portion of the inland West Antarctic Ice Sheet, in perhaps a few hundred years. Either an ice shelf (a remnant of the inland ice sheet) or seasonally open seas would then occupy the Byrd Subglacial Basin, and there would be a flow of ocean water between the Ross and Weddell Seas.

3.3.5 An alternative West Antarctic history based on model simulations. To put the hypothetical scenario of the preceding section into perspective, we have the less dramatic sequence of events resulting from the most advanced three-dimensional model calculations so far carried out for Antarctica (Budd & Smith 1982). In these, both a drop in sea level of the size probably associated with the growth of the Northern Hemisphere ice sheets and the corresponding rise in sea level at the end of the glaciation were used as inputs to the model, which also made allowance for climatic changes in mass balance and temperature (ice viscosity), for the delayed bedrock response to the changing ice load, and to different backpressures exerted by the ice shelves. The results lend support to the advance of the West Antarctic Ice Sheet to the edge of the continental shelf around 18,000 B.P., inferred by Hughes et al. (1981), but suggest that the present extent of the ice sheet represents a minimum and a precursor to a readvance as the bedrock continues to recover from the ice load of the recent glaciation.

# 4. HUMAN-INDUCED CHANGES AND CONSEQUENCES

## 4.1 Modification of the Atmosphere and Its Climatic Consequences

Many workers have pointed out that the known duration of earlier interglacials (10,000 to 15,000 years) suggests that the present one has nearly run its course; in the natural course of events, temperatures would probably "soon"--sometime during the next few millennia--start to decline toward eventual full-glacial levels. However, it seems increasingly likely that the climatic effects of human activities, particularly those involving $CO_2$ production, could by the end of the century start to overwhelm any natural cooling trends that may be in progress, causing warming that will last at least several centuries, because of the long residence time of $CO_2$ in the atmosphere.

The effects of human activities on climate are discussed in Appendix A; here, we will briefly review the main factors. Human activities that are likely to affect climate are, in order of decreasing importance, those that (a) add to the $CO_2$ content of the atmosphere (fossil fuel combustion, lime kilning, deforestation, and agricultural oxidation of soil humus); (b) add other infrared-absorbing gases to the atmosphere (fossil fuel combustion, certain industrial processes, use of nitrate fertilizers, use of chlorofluorocarbons as refrigerants and aerosol propellants); (c) emit particles into the atmosphere (fossil fuel combustion, industrial processes, deforestation, agricultural slash-and-burn practices); and (d) change the albedo of the earth's surface (deforestation and overgrazing). Man-made heat production is at present of only local importance and will remain so for at least several decades.

We will consider these categories in reverse order: the change in the earth's surface albedo caused by human activities is uncertain but may have amounted to $\pm 10\%$ since the Neolithic Age. The effect on climate of atmospheric particles resulting from human activities is difficult to determine, mainly because of uncertainties about their distribution and optical properties. Recent research suggests that industrial particles in general have a net warming effect when over land, especially at high latitudes. Recently, attention has been paid to the heavy haze that affects the Arctic, particularly in winter and spring. This haze is now believed to result from the transport of man-made pollutants from North America, Europe, and Asia. The haze may affect the Arctic climate by its influence on cloud formation and also by absorbing additional sunlight. It may also lower the surface albedo of snow on the Arctic pack ice if soot particles are deposited on the surface.

As possible man-made climatic modifiers the most important gases after $CO_2$ are the chlorofluorocarbons, which, if present rates of injection are maintained, may cause an average global warming of about 1°C by A.D. 2030. An additional warming of about 0.5°C may result from increasing levels of nitrous oxide ($N_2O$) derived from nitrogen fertilizers and fossil fuels. There are other industrial gases with smaller but not negligible climatic impacts. None of these gases has the long residence time in the atmosphere of $CO_2$, and our knowledge of the rates of injection and removal are too meager to place much confidence in estimates of future concentrations, but it seems clear that they could substantially reinforce the warming due to $CO_2$.

The present rising atmospheric $CO_2$ level is caused by burning fossil fuels, deforestation, soil humus oxidation, and lime kilning. Of these, the burning of fossil fuel is generally believed to be the most important, but there is some disagreement about the relative contributions of the other sources. The rate at which the $CO_2$ level is rising, however, is being monitored at Point Barrow and the South Pole, on Mauna Loa, and elsewhere. These measurements show that everywhere the level has risen about 6 percent since 1958; estimates of the production from the burning of fossil fuel are about twice this amount, suggesting that about half the excess $CO_2$ produced has remained in the atmosphere, the other half presumably having been taken up by the ocean. If the present rate of increase in the burning of fossil fuel continues, the $CO_2$ level would double by the second half of the next century (Climate Board 1982), but it is most probable that this rate of increase will slacken in the decades ahead and the time for doubling move farther into the future, perhaps into the second half of the next century. If all fossil fuel that is accessible by present techniques were burnt, the atmospheric $CO_2$ level would rise between fivefold and eightfold. This concentration of $CO_2$ would probably not be completely removed from the atmosphere for from 5,000 to 10,000 years, due to the slow rate of mixing of the deep-ocean water and slow removal mechanisms, and the concentration would be likely to remain at more than twice the preindustrial level for several centuries at least.

Several attempts have been made to estimate the likely warming effect of increased atmospheric $CO_2$. At first there was a wide diversity in the figures obtained by different investigators, but a consenus seems to have been reached that the average global warming from a doubling of $CO_2$ would probably be between 2°C and 3°C. Modeling also suggests that warming at high latitudes may be 2 or 3 times greater than the average, with a greater increase in the Arctic than in the Antarctic. This enhancement of the warming at the poles seems to be due to the lesser amount of convective mixing compared to lower latitudes and the feedback effect of decreased albedo as sea-ice cover and snow (primarily in the Arctic) recede poleward. This theoretical conclusion is reinforced by the fact that the pattern of Northern Hemisphere warming during the first half of this century has been greatest in high latitudes, especially in winter.

## 4.2 Climatic Warming--Interaction with the Cryosphere

General assessments of potential effects of $CO_2$-induced warming on the cryosphere have been provided by Barry (1978, 1982) and Hollin and Barry (1979), based on historical-geological data, empirical results, and modeling studies.

4.2.1 Sea ice. If the increasing amount of $CO_2$ in the atmosphere were to increase temperatures by several degrees in the polar regions, the next concern would be the resulting effect on the thickness and extent of the sea ice. Estimates of change in ice thickness due to warming can be based on simple empirical relationships between ice thickness and the number of degree-days below freezing in the Arctic, which lead to a decreasing thickness of about 10 percent for each 3° rise in winter temperature. This estimate suggests that the extent of the ice is not very sensitive to a winter temperature change of the magnitude that could reasonably be expected. On the other hand, other empirical relationships suggest that a rising mean summer temperature will move the summer southern margin of the Arctic sea ice approximately 1° north for each 1°C in change (Rogers 1978). If linear extrapolation is valid, it follows that a coverage only half as large as that existing at present would remain at the end of the summer if the temperature increased 3°, and that an approximately 10° temperature change would be needed to remove the summer ice cover completely. So large a temperature increase would probably require nearly a quadrupling of atmospheric $CO_2$ content; however, at least two other theoretical analyses have suggested a much greater sensitivity--namely, that about a 5° increase in temparature would be sufficient to melt the Arctic sea ice completely in midsummer (Parkinson & Kellogg 1979; Manabe & Stouffer 1979)--but concluded that it would probably form again in winter.

In the Antarctic, most of the sea ice is less than 1 year old, is thin by Arctic standards, and undergoes large seasonal changes in area. Therefore, it seems likely that climatic warming would have a relatively greater effect on Antarctic than on Arctic sea ice. One analysis leads to an estimate of a 2° latitude variation in maximum extent of the ice corresponding to a 1°C change in annual mean temperature (Budd 1975b), and it is likely that interaction with ocean heating would act to augment the simple temperature effect. Assuming that the summertime extent of the ice would decrease proportionally, it appears that in the Antarctic the doubling of the $CO_2$ content of the atmosphere could quite conceivably diminish the summertime sea-ice cover, and perhaps the wintertime cover as well, by one half.

4.2.2 Snow cover. The effects of a warming on snowfall and snow cover will differ according to latitude. In low and middle latitudes, where the occurrence of snow rather than rain is frequently marginal, warming will decrease the frequency of snowfall and the duration of snow cover on the ground. In high latitudes, where snowfall is limited by the frequency of cyclonic incursions and the moisture flux, there is a tendency for warmer winters to be more snowy. At Barrow, Alaska, for example, for the 30 winters 1946-1976, there is a -0.57 correlation between freezing degree-days and snowfall during December-February. In Labrador-Ungava and Keewatin, snowy winter months tend to be associated with an average temperature departure of +1° to +3° C (Brinkmann & Barry 1972), although this relationship is apparently nonlinear and it does not appear to hold on the time scale of daily data. While increased snowfall will provide a deeper snowpack, the duration of snow cover in the Arctic is likely to be only marginally affected by this increase. Estimated ablation rates during the melt period at Barrow, Alaska, are 9 mm $day^{-1}$. Here the 30-40 cm snow cover disappears in about 10 days from the start of melt, and evaporation accounts for only 2 percent of total ablation, although slightly higher contributions may apply as the snow cover becomes patchy (Weller & Holmgren 1974). On a hemispheric scale, the general circulation model simulations of $CO_2$ effects (Manabe & Wetherald 1980; Manabe & Stouffer 1980) indicate a broad retreat of the snow and ice line and a concomitant 30-40 percent decrease in surface albedo at about 75°-85°N for a $CO_2$ doubling.

4.2.3 Glaciers and ice sheets. Mountain glaciers in some arid high polar regions might initially respond to general warming by advancing, as a consequence of increased snowfall; indeed, this may have occurred recently in parts of the Canadian Arctic islands and in Antarctica. In most parts of the world, however, and in the long run, mountain glaciers would shrink.

The Greenland Ice Sheet would shrink or expand, depending on the changes in net mass balance resulting from increased snowfall in the central part and increased melting and ablation at its edges. The East Antarctic Ice Sheet would probably be little affected until warming had lasted for a long time. Even if snowfall increased while melting was still insignificant, there would be little immediate change because of the very long response time of the ice sheet as a whole. However, there are portions of this ice sheet that are grounded below sea level, and they could behave differently from the main body.

The West Antarctic Ice Sheet probably would be little affected until or unless its ice shelves, exposed to higher ocean temperatures and less protected by sea ice, were destroyed. As discussed in Section 3.2.3, what would happen next is a matter of controversy. The marine ice sheet might then "rapidly" disappear (in a few centuries), leading to about a 5 m rise in sea level, or there might be little change for the next millennium.

A rough estimate of the temperature rise that might correspond to ice-shelf destruction can be made by an analog (Mercer 1978). Along the west coast of the Antarctic Peninsula, ice shelves abruptly cease north of the midsummer surface water isotherm of -1.5°C, which approximately coincides with the northern limit of sea ice at its minimum extent in late summer and with the 0°C isotherm for midsummer air temperature of 0°C. Midsummer air temperatures along the Ross and Filchner-Ronne ice barriers and the coast of the Bellingshausen Sea are now about -5°C, so that, by analogy, the critical temperature increase for these bodies is about 5°C. If temperatures rose further and were sustained, the West Antarctic ice shelves, by this argument, would eventually be destroyed, but this could take many centuries.

It should be recalled that model studies suggest that a 5° warming in the polar regions may be exceeded with a doubling of $CO_2$ (see Appendix A and Section 4.1). Of course, the likelihood of ice-shelf destruction would increase still further as the $CO_2$ level continued to rise. The most vulnerable part of the periphery of the West Antarctic Ice Sheet may be in the Amundsen Sea, where the ice streams terminate in floating ice tongues that are currently less restrained than the large ice shelves. Destruction of these ice tongues, therefore, should no longer be of major consequence, and new observations (Crabtree & Doake 1982) suggest that the inland mass balance is positive at present, lending no support, for the moment at least, to the hypothetical scenario in Section 3.2.4.

These arguments and suggestions are neither backed nor contradicted by any definitive quantitative study. Even the most sophisticated models yet produced involve critical simplifications and assumptions, modification of which could be important. For example, the preliminary numerical modeling results already referred to simply show that a model can be devised that yields a credible result by reasonable adjustment of certain parameters, parameters whose actual values are not well known. Thus, the whole matter is very much undecided at present. Still, the crucial point is that drastic responses of the West Antarctic marine ice sheet (and portions of the East Antarctic Ice Sheet), although certainly not proven, cannot be ruled out. Furthermore, the unstable nature of the dynamic response of a marine ice sheet is such that the process, if it were set in motion, would be impossible to reverse.

## 4.3 Global Implications of a Polar Warming

The consequences of a $CO_2$-induced warming on the global cryosphere are numerous, involving decreases in winter snow-cover extent and duration, thinning and recession of glaciers, ground ice, and sea ice, and the possible rapid disappearance of marine-based ice sheets causing a 5 m or so rise of global sea level (Hollin & Barry 1979). Although such results seem quite possible, the development of coupled ocean-atmosphere-cryosphere climate models is still in its infancy; feedbacks among various parts of the climate system may modify the currently recognized interactions in unsuspected ways.

For example, the temperature-ice-albedo coupling included in most models is widely accepted as a major positive feedback effect (see Appendix A). Reduced snow and ice cover should allow storm tracks to shift poleward and perhaps further accelerate warming trends in higher latitudes as well as affecting the precipitation patterns there. Model experiments and empirical evidence suggest that an ice-free Arctic Ocean would cause higher winter temperatures and more precipitation over the Arctic Basin and possibly lower temperatures and less precipitation over the continents, but there are no good analogs in postglacial time for these conditions.

Warmer polar regions would weaken the equator-pole temperature gradients and tend to reduce the intensity of the atmospheric circulation. In turn, weaker wind systems would slacken oceanic current systems that transport heat poleward, and this slowing could conceivably lessen the general warming in higher latitudes. Global atmospheric teleconnections can be expected, as shown by recent work linking the character of winter conditions in the Greenland region with the intensity of the Gulf Stream, the strength of the trade winds, and precipitation amounts in south-central Africa (Namias 1980), but cause and effect cannot yet be separated.

The impact of these climatic changes on mankind would be many and varied--beneficial in some regions and damaging in others. Nevertheless, the international dislocations would be serious and extensive (Kellogg 1978; Kellogg & Schware 1981; American Association for the Advancement of Science 1980).

Probably the most serious effect to be considered is the possible rise in sea level. A 5 m rise would have a disastrous effect on the low-lying coastal regions of the world, where a large fraction of the world's population lives. Of course, the rise would occur relatively slowly on a human time scale even in the most drastic scenario; nevertheless, the social and economic implications of flooding out many of the largest cities of the world are truly staggering. If such a disaster is going to occur, it may not be preventable, but it certainly behooves mankind to be as long forewarned as possible.

# 5. PROBLEMS AND RESEARCH OBJECTIVES

## 5.1 Sea Ice and Snow Cover

A number of outstanding problems that urgently require attention can be identified. While, partly as a result of Project AIDJEX (Arctic Ice Dynamics Joint Experiment), the dynamics and thermodynamics of large-scale sea-ice behavior in the Arctic are now reasonably well understood (Polar Group 1980), a fully coupled ocean-ice-atmosphere model has yet to be developed. Moreover, the processes at work in the marginal ice zone around the Arctic, and especially in the Antarctic, are little known. In the Antarctic even the mechanisms of the overall seasonal fluctuations remain in doubt. The role of the numerous recently identified polynyi should be determined. The question of interannual variability and long-term trends in ice extent has not yet been addressed in detail, although a start has been made for the Arctic. Ice-thickness distribution in both polar regions is still poorly known; earlier suggestions of long-term trends in ice thickness are not generally accepted (Barry in press). Studies of snow cover and climate interactions are hampered by the limited observational data on large-scale snow-cover distribution and depth. Little work has yet been done on modeling such interactions in general circulation models despite recognition of the importance to global climate of snow and ice albedo feedbacks.

Three broad objectives can be proposed as a basis for improved understanding of sea-ice and snow-cover responses to climate:

1. Better empirical data sets are needed on the short-term (1-10 year) and long-term ($10^3$ years) variation of snow and ice extent and ice concentration in relation to climatic conditions.

2. Studies of atmospheric effects on snow cover and sea ice, and *vice versa*, are needed on synpotic and seasonal time scales. The availability of microwave and other satellite imagery offers considerable promise here, although more synchronous ground-truth data are required to assist in the interpretation of these data. Commensurate atmospheric measurements from data buoys are also needed.

3. Modeling experiments are needed to explore the range of possible climatic effects of increases in $CO_2$ and other atmospheric pollutants on snow cover and sea ice.

## 5.2 Present Ice Sheets

The polar ice sheets pose a special problem in the $CO_2$/climate context. Our knowledge about their present state is limited, and about their past it is at best circumstantial. Modeling their future is like predicting developments of the general circulation from a single weather chart.

This situation implies a need for analog arguments (e.g., using surging glaciers, which can be observed, as models for an ice sheet potentially capable of surging), and also a broad, numerical approach: varying key parameters in ice-sheet models and, for the lesser understood parameters, constructing entire solution fields instead of single projections. This procedure will also establish some order in the wide range of currently available projections.

Atmospheric modelers could contribute a great deal by examining the summer scenario of a persistent 0°C temperature over a broad coastal belt of the Antarctic ice sheet. This case should be the easiest to model atmospherically, since it reduces to a minimum the importance of the surface inversion that may decouple the boundary-layer flow from the rest of the atmosphere in the other seasons, and especially in winter.

No foreseeable $CO_2$-induced warming will turn the polar ice sheets into "banana belts" in winter, but since their elevation changes may have some effect on the global circulation, the katabatic boundary layer needs improved representation in global circulation models. A theoretical derivation of the ice-sheet surface-pressure distribution will be needed for this, and is being developed.

Oceanic modelers might address the difficult question of heat transfer at the base of an ice shelf by modeling the circulation in a basin, with a lid, that opens on a larger ocean next to the ice shelf with or without gyral circulation.

The ice-sheet modelers should coordinate their efforts to the extent of using similar inputs for comparable reconstructions of the past and future behavior of the ice sheet. Some of the existing models will require a good deal of development work for this purpose, but this work is justified by the prospect of achieving the most powerful projection tool that can be constructed from present knowledge of ice and ice sheets.

Evidence should continue to be sought on whether the West Antarctic Ice Sheet was present or absent during the last interglacial. A definite answer to that question would not only shed light on the past behavior of that marine ice sheet but would help test predictions concerning the relative magnitude of climatic warming in the Arctic, the Antarctic, and lower latitudes. This information could in turn provide a partial analog of a $CO_2$-induced warming.

Finally, the critical need for continuing collection of field data on the ice sheets themselves is obvious. Such data as ice thickness, surface elevation and slope, bed slope, surface mass balance, basal mass balance, internal structure, ice velocity, iceberg calving rates, changing sea level, and changing temperatures are needed. Many of these data can be obtained by remote sensing, particularly if a dedicated ice-sensing satellite materializes (Science and Applications Working Group 1979) eventually becomes reality. Even with the best of remote-sensing capabilities, however, a vigorous field program will still be required.

## REFERENCES

Alford, D., 1974. Snow. In J. D. Ives and R. G. Barry (eds.), Arctic and Alpine Environments (pp. 85-220). Methuen, London.

Allison, I., 1982. The role of sea ice in climatic variations. Report of the WMO/CAS-JSC-CCCO Meeting of Experts on the Role of Sea Ice in Climatic Variations. WCP-26. World Meteorological Organization, Geneva.

American Association for the Advancement of Science, 1980. Environmental and Societal Consequences of a Possible $CO_2$-Induced Climate Change: A Research Agenda, Vol. 1. Prepared by American Association for the Advancement of Science for the Carbon Dioxide Effects Research and Assessment Program, U.S. Department of Energy, doe/EV/10019-01, Washington, D.C.

Barry, R. G., 1978. Cryospheric responses to a global temperature increase. In J. Williams (ed.), Carbon Dioxide, Climate and Society (pp. 169-180). Pergamon Press, London.

Barry, R. G., 1980. Meteorology and climatology of the seasonal sea ice zone. Cold Regions Science and Technology, 2, 133-150.

Barry, R. G., 1982. Snow and ice indicators of possible climatic effects of increasing atmospheric carbon dioxide. Proceedings of the Workshop on First Detection of Carbon Dioxide Effects. DOE/CONF-8106214 (pp. 207-236). U.S. Depertment of Energy, Washington, D.C.

Barry, R. G., 1983. Arctic Ocean ice and climate: perspectives on a century of polar research. Annals, Association of American Geographers, 73 (4), 485-501.

Barry, R. G., in press. The sea ice data base. In N. Untersteiner et al. (eds.), The Geophysics of Sea Ice. Proceedings of the NATA Advanced Study Institute, Maratea, Italy, 1981. Plenum, New York.

Bentley, C. R. and K. C. Jezek, 1982. RISS, RISP, and RIGGS: post-IGY glaciological investigations of the Ross Ice Shelf in the U.S. program. Journal of the Royal Society of New Zealand, 11 (4), 355-372.

Brinkmann, W. A. R. and R. G. Barry, 1972. Paleoclimatological aspects of the synoptic climatology of Keewatin, Northwest Territories, Canada. Paleogeography, Paleoclimatology, Paleoecology, 11, 77-91.

Budd, W. F., 1975a. A first simple model for periodically self-surging glaciers. Journal of Glaciology, 14 (70), 3-21.

Budd, W. F., 1975b. Antarctic sea-ice variations from satellite remote sensing in relation to climate. Journal of Glaciology, 15 (73), 417-427.

Budd, W. F. and I. N. Smith, 1979. The growth and retreat of ice sheets in response to orbital radiation changes (abstract). Presented at IUGG XVII General Assembly, Canberra, Australia, December 1979.

Budd, W. F. and I. N. Smith, 1982. Large-scale numerical modelling of the Antarctic ice sheet. Annals of Glaciology, 3, 42-49.

Burckle, L. H., P. Robinson, and D. Cooke, 1982. Reappraisal of sea-ice distribution in the Atlantic and Pacific sectors in the Southern Ocean at 18,000 yr B.P. Nature, 299, 435-437.

Clark, D., 1983. Origin, nature and world climate effect of Arctic Ocean ice cover. Nature, 300, 321-325.

Climate Board, 1982. Carbon Dioxide and Climate: A Second Assessment. National Academy Press, Washington, D.C.

Crabtree, R. D. and C. S. M. Doake, 1982. Pine Island glacier and its changing basin: results from radio echo-sounding. Annals of Glaciology, 3, 65-70.

Denton, G. H. and T. Hughes, 1981. The Arctic Ice Sheet: an outrageous hypothesis. In G. H. Denton and T. Hughes (eds.), The Last Great Ice Sheets (pp. 437-467). Wiley-Interscience, New York.

Dewey, K. F. and R. Heim, Jr., 1981. Satellite observations of variations in northern hemisphere seasonal snow cover. NOAA Technical Report NESS 87. U.S. Department of Commerce, Washington, D.C.

Dickson, R. R. and J. Posey, 1967. Maps of snow-cover probability for the northern hemisphere. Monthly Weather Review, 95, 347-52.

Drewry, D. J., 1979. Late Wisconsin reconstruction for the Ross Sea region, Antarctica. Journal of Glaciology, 24 (90), 231-244.

Frakes, L. A., 1978. Cenozoic climates: Antarctica and the Southern Ocean. In A. B. Pittock et al. (eds.), Climatic Change and Variability: A Southern Perspective (pp. 53-69). Cambridge University Press, New York.

Hendy, C. H., E. M. Healy, E. M. Rayner, J. Shaw, and A. T. Wilson, 1979. Late Pleistocene glacial chronology of the Taylor Valley, Antarctica, and the global climate. Quaternary Research, 11, 172-84.

Hibler, W. D., III, 1981. Sea ice growth, drift and decay. In S. Colbeck (ed.), Dynamics of Snow and Ice Masses. Academic Press, New York.

Hollin, J. T., 1980. Climate and sea level in isotope stage 5: an East Antarctic ice surge at 95,000 BP? Nature, 283, 629-633.

Hollin, J. T. and R. G. Barry, 1979. Empirical and theoretical evidence concerning the response of the earth's ice and snow cover to a global temperature increase. Environment International, 2, 437-444.

Hughes, T. J., 1973. Is the West Antarctic Ice Sheet disintegrating? Journal of Geophysical Research, 78, 7889-7910.

Hughes, T. J., G. H. Denton, B. G. Anderson, D. H. Schilling, J. L. Fastook, and C. S. Lingle, 1981. The last great ice sheets: a global view. In G. H. Denton and T. J. Hughes (eds.), The Last Great Glacial Ice Sheets (pp. 262-317). Wiley-Interscience, New York.

Kellogg, W. W., 1978. Global influences of mankind on the climate. In J. Gribbin (ed.), Climate Change (pp. 205-227). Cambridge University Press, New York.
Kellogg, W. W. and R. Schware, 1981. Climate Change and Society. Westview Press, Boulder, Colorado.
Kruss, P. D. and I. N. Smith, 1982. Numerical modeling of the Vernagtferner and its fluctuations. Zeitschrift fur Gletscherkunde und Glazialgeologie, 18 (1), 97-110.
Kukla, G., 1981. Climatic role of snow covers. In I. Allison (ed.), Sea Level, Ice, and Climatic Change. IAHS Publication 131 (pp. 79-107). International Association of Hydrological Sciences, Gentbrugge, Belgium.
Malzer, H. and H. Seckel, 1976. Das geometrische Nivellement uber das Inlandeis-Hohennanderungen zwischen 1959 und 1968 im Ost-West-Profil der EGIG. Zeitschrift fur Gletscherkunde und Glazialgeologie, 11 (2), 245-252.
Manabe, S. and R. Stouffer, 1979. Study of climatic impact of $CO_2$ increase with a mathematical model of global climate, Nature, 282, 491-493.
Manabe, S. and R. J. Stouffer, 1980. Sensitivity of a global climate model to an increase of $CO_2$ concentration in the atmosphere. Journal of Geophysical Research, 85, 5529-5554.
Manabe, S. and R. T. Wetherald, 1980. On the distribution of climate change resulting from an increase in the $CO_2$ content of the atmosphere. Journal of Atmospheric Science, 37, 99-118.
Matson, M. and D. R. Wiesnet, 1981. New data base for climate studies. Nature, 289, 451-456.
Maykut, G. A., 1982. Large-scale heat exchange and ice production in the central Arctic. Journal of Geophysical Research, 87, 7971-7984.
Mercer, J.H., 1978. West Antarctic ice sheet and $CO_2$ greenhouse effect: a threat of disaster. Nature, 271, 321-325.
Namias, J., 1980. Some concomitant regional anomalies associated with hemispherically averaged temperature variations. Journal of Geophysical Research, 85, 1585-1590.
Parkinson, C. L. and W. W. Kellogg, 1979. Arctic sea-ice decay simulated for $CO_2$-induced temperature use. Climatic Change, 2, 149-162.
Polar Group, 1980. Polar atmosphere-ice-ocean processes: review of polar problems in climate

research. Review of Geophysics and Space Physics, 18, 525-543.

Porter, S., 1981. Glaciological evidence of Holocene climatic change. In T. M. L. Wigley, M. J. Ingram, and G. Farmer (eds.), Climate and History (pp. 82-110). Cambridge University Press, New York.

Revelle, R. R., 1983. Probable future changes in sea level resulting from increasing carbon dioxide. In Carbon Dioxide Assessment Committee, Changing Climate (pp. 433-448). National Academy Press, Washington, D.C.

Rogers, J. C., 1978. Meteorological factors affecting interannual variability of summertime ice extent in the Beaufort Sea. Monthly Weather Review, 106, 890-897.

Science and Applications Working Group, 1979. Ice and Climate Experiment (ICEX). NASA Goddard Space Flight Center, Greenbelt, Maryland.

Stuiver, M., G. H. Denton, T. J. Hughes, and J. L. Fastook, 1981. History of the marine ice sheet in West Antarctica during the last glaciation: a working hypothesis. In G. H. Denton and T. Hughes (eds.), The Last Great Ice Sheets (pp. 319-346). Wiley-Interscience, New York.

Ten Brink, N. W. and A. Weidick, 1974. Greenland ice sheet history since the last glaciation. Quaternary Research, 4, 429-430.

Walsh, J. E. and C. M. Johnson, 1979. An analysis of Arctic sea ice fluctuations, 1953-77. Journal of Physical Oceanography, 9, 580-591.

Walsh, J. E., D. R. Tucek, and M. R. Peterson, 1982. Seasonal snow cover and short-term climate fluctuations over the United States. Monthly Weather Review, 110, 1474-1485.

Weertman, J., 1974. Stability of the junction of an ice sheet and an ice shelf. Journal of Glaciology, 13, 3-11.

Weertman, J., 1976. Milankovitch solar radiation variations and ice age ice sheets. Nature, 261 (5555), 17-20.

Weller, G. and B. Holmgren, 1974. The microclimate of the Arctic tundra. Journal of Applied Meteorology, 13, 854-862.

Whillans, I. M., 1976. Radio-echo layers and the recent stability of the West Antarctic ice sheet. Nature, 264 (11), 152-155.

Williams, J., 1975. The influence of snowcover on the atmospheric circulation and its role in climatic change: an analysis on results from the NCAR Global Circulation Mode. Journal of Applied Meteorology, 14, 137-152.

Zwally, J. H., J. C. Comiso, C. L. Parkinson, W. J. Campbell, F. Carsey, and P. Gloersen, 1983. Antarctic Sea Ice Cover, 1973-78. Satellite Passive Microwave Observations. NASA-SP-459 (206 pp.). NASA Goddard Space Flight Center, Greenbelt, Maryland.