Laboratory Antarctica: Research Contributions to Global Problems

GUNTER WELLER, CHARLES R. BENTLEY, DAVID H. ELLIOT, LOUIS J. LANZEROTTI, PATRICK J. WEBBER

Research in Antarctica is becoming increasingly important in the large interdisciplinary studies of connections within the earth's geosphere-biosphere system. Four examples of broad research areas are discussed. Upper atmosphere research explores the sun-earth interactions, which are most intense in the polar regions. The mass balance and dynamics of the large Antarctic ice sheet, and its paleoclimatic records recovered from deep ice cores, are important indicators of past and present global changes. Antarctica and sediment cores from the Southern Ocean contain the history of inception and growth of the ice masses and their subsequent fluctuations, and the long-term history of paleoclimate. The remarkable adaptations of Antarctic biota to extreme cold and drought may allow, through biotic monitoring, the detection of changes in the ocean and climate of Antarctica.

AWARENESS OF ANTARCTICA (FIG. 1) HAS REACHED THE typical American newspaper reader rather infrequently in the last decade or so. Newsworthy items tended to be related to disasters, such as the Air New Zealand crash of a DC-10 on Mount Erebus in 1981, or to political issues, such as a possible review of the Antarctic Treaty (due in 1991), the clamor of Third World nations to have a piece of the antarctic "action," the establishment of a Greenpeace base on the continent, and rumors about the allegedly rich petroleum and mineral resources of the region. More recently, however, a number of scientific issues have also proved newsworthy. These include the Antarctic ozone "hole," the possible collapse of the West Antarctica ice sheet in response to global "greenhouse" warming and the consequent significant (7-meter) rise in sea level, the large swarms of high protein krill in the nutrient-rich Southern Ocean, and the geological and paleontological evidence for biogeographic links with the other southern continents. Antarctica's time for world attention has perhaps again arrived; this time the focus may be more on science than on heroic explorations, as in the first half of this century.

G. Weller is professor of geophysics at the University of Alaska, Fairbanks, AK 99775, and chairman of the National Research Council's Polar Research Board. C. D. Bentley is A. P. Crary Professor of Geophysics at the University of Wisconsin, Madison, WI 53706, and director of the Geophysical and Polar Research Center. D. H. Elliot is the director of the Byrd Polar Research Center at Ohio State University, Columbus, OH 43210. L. J. Lanzerotti is a distinguished member of the technical staff with AT&T Bell Laboratories, Murray Hill, NJ 07974. P. J. Webber is professor of biology at the University of Colorado, Boulder, CO 80309.

The Upper Atmosphere

Nowhere are the complex interactions among the elements of the earth's environment more apparent than in the upper atmosphere of the polar regions, where particles and fields from the solar wind interact most strongly with the earth. In the 30 years that have elapsed since the International Geophysical Year (IGY), there has been a revolution in understanding of the physics and chemistry of upper atmospheric phenomena. In the context of Antarctic research, the upper atmosphere is defined broadly to cover regions from an altitude of about 30 km and up, as far as the sun and, more recently, astronomical studies of the galaxy and beyond. Much of the revolution of understanding of near-earth space was produced by the advent of scientific spacecraft, both earth orbiting and earth escape, data from which determined the morphology of the magnetosphere of the earth (Fig. 2) and the interplanetary medium. The magnetosphere is that region of space in which the electric and magnetic fields of the earth normally dominate the motions of the charged particles—the plasmas—that populate these regions (*1*).

In parallel, research on upper atmosphere science in the Antarctic has changed fundamentally from the basic exploration mode that was appropriate for the IGY era. The continent is now used as an

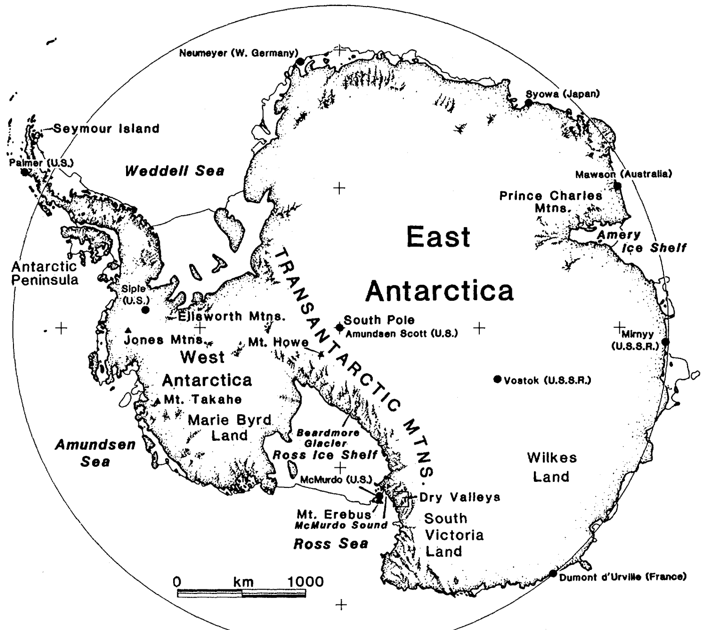


Fig. 1. Map of Antarctica, showing major geographical features and some of the scientific stations.

ideal staging location for active and passive studies of micro- and macroscale time and spatial variations of magnetosphere processes, particularly those associated with the plasmasphere and the auroral and polar cusp regions shown in Fig. 2. Geomagnetic fields threading through the outer regions of the magnetosphere intercept the ionosphere (altitude, ~100 km) and the surface of the earth in high latitude geographical regions. Because the offset between the earth's rotation axis and the magnetic dipole axis is only about 12°, magnetic fields from the inner magnetosphere also intersect the earth at rather high geographical latitudes in some locations. Indeed, as indicated in Fig. 2, the Antarctic continent is well placed for U.S. research because it can be conducted in geomagnetic latitudes ranging from about 53°S at Palmer Station to 79°S at McMurdo. Siple Station, at a geomagnetic latitude of about 65°S is located near the average position of the boundary of the plasmasphere (traced along field lines to the ionosphere). The daily rotation of the earth alternately brings the auroral zone on the night side and the magnetosphere cusp during local magnetic day over the South Pole Station (magnetic latitude, ~78°S). Figure 2 illustrates the nominal auroral zone in the Southern Hemisphere for geomagnetic local noon at the South Pole.

Magnetic fields traced from some areas of the Antarctic continent are found to intercept Northern Hemisphere land masses in eastern Canada, in Greenland, and in Iceland. Japanese, British, and U.S. investigators have made good use of this feature of the Antarctic continent in recent years, deploying sensors in both hemispheres to study "conjugate" phenomena and thereby better to delineate magnetosphere processes from the ground. The phenomena studied in detail include auroral displays, ultralow frequency (ULF) and very low frequency (VLF) variations in magnetic and electric fields, changes in the characteristics of the ionosphere under conditions of solar-induced disturbances, and galactic cosmic rays (*2*).

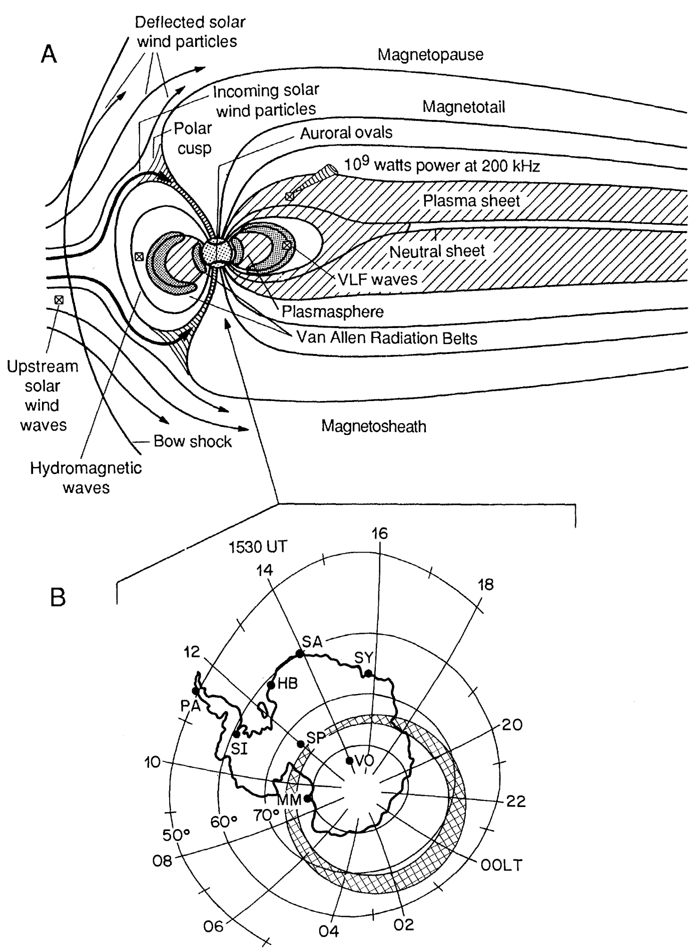


Fig. 2. The magnetosphere of the earth. (**A**) Important features of the plasmas and waves in the magnetic fields that surround the earth. (**B**) Nominal auroral zone for geomagnetic local noon at the South Pole (SP). Other stations shown are Vostok (VO), McMurdo (MM), Siple (SI), Palmer (PA), Halley (HB), Sanea (SA), and Syowa (SY).

One of the more innovative uses of the Antarctic ice sheet for upper atmosphere studies has involved the establishment, at Siple Station, of a long (42 km) dipole antenna that transmits coherent VLF signals (about 2 to 5 kHz) at a power level of 175 watts into the ionosphere and magnetosphere (*3*). The ice sheet thickness beneath the antenna (average, ~2000 m) makes it possible to achieve an adequate transmitted efficiency (several percent). Controlled transmission experiments have revealed a number of new plasma physics processes in the ionosphere and magnetosphere. For example, a highly nonlinear, narrowband wave growth process, called the "coherent wave instability," has been found where, for input signals exceeding a certain threshold, coherent waves are found to grow at exponential rates of 25 to 250 decibels per second (*4*). The mechanism for producing the wave growth appears to be based on Doppler-shifted cyclotron wave resonances that occur along magnetic fields near the equatorial plane between the input waves and the counter-streaming, ambient magnetosphere electrons. Following saturation of the wave growth, narrowband wave emissions into the magnetosphere are found to be spontaneously triggered; these emissions persist for a few tenths of a second and rise and fall in frequency by several hundred hertz or more with respect to the input frequency (*4*). The triggered waves can be suppressed or entrained by other man-made signals found in the magnetosphere, such as those produced by VLF communication and navigation transmitters or by the northern hemisphere power grids.

In addition to their intrinsic interest for fundamental plasma physics studies, the active wave injection experiments at Siple have provided new insights into naturally occurring plasma processes in the magnetosphere and ionosphere. In particular, new understandings have been achieved in a variety of research areas that can ultimately be related to the studies of natural phenomena, such as the relation of natural VLF and ULF activity to the losses of trapped particles from the radiation belts (*5, 6*) and the production of x-rays (*7*) and light emissions (*8*) in the ionosphere by the bombarding particles.

Also in recent years the South Pole site, with its high altitude, very low humidity, and continuous daylight during austral summer, has been used to excellent advantage for specialized studies of the sun, especially for studies of the regular, periodic (minutes to hours) oscillations of the sun as it transfers heat from its internal nuclear source to the surface (*9*). Long-period (tens of minutes to hours) oscillations are especially difficult to measure at a single mid-latitude observatory because the rotation of the earth introduces noise at frequencies of one per day and its harmonics into the spectrum. Thus, measurements at South Pole, where continuous daylight exists during austral summer, provides highly advantageous, long intervals of uninterrupted observing time. Weather conditions tend to be the limiting factor, though generally allowing as much as 5 to 6 days of continuous observing.

Analyses of data from South Pole Station for the wide range of frequencies over which the sun oscillates—in both period of oscillation and spatial scale on the observed disk—show distinct, discrete patterns in a frequency-spatial scale diagram (*10, 11*). Such results can be used, with sophisticated models of the possible modes of solar oscillation, to deduce important information about the solar core and the solar convective zone—the region between the solar surface and a depth of about 0.7 of a solar radius where the heat generated by the nuclear reactions at the center of the sun is convected to the surface. The larger the spatial scale of the oscillation, the deeper the sun is effectively being "probed." The situation, by analogy with studies of the interior of the earth, is often called "helioseismology."

Interpretation of recent measurements has led to the conclusion that the outer approximately half of the sun rotates with essentially the same angular velocity as that measured for the photosphere (*12*). Continued solar studies at South Pole will provide further insight into the fundamental physics of the nearest star. In addition, other recent austral summer measurements (*13*) show that during times of clear weather conditions, the effective sky temperature at 3 mm wavelength above the South Pole is the lowest measured to date anywhere on the earth ($T_{sky} < 7$ K). It is likely that in the future the South Pole will be increasingly utilized for astronomical studies in the infrared and millimeter wavelength regions.

The Antarctic Ice Sheet

The largest ice mass on the earth is the Antarctic ice sheet, more than 5 km in maximum thickness with a volume of 25×10^6 km^3. There have been several significant scientific advances in the study of this ice mass since the IGY. Perhaps the most striking and rewarding has been the development of radar sounding of ice thickness, particularly from the air. The importance to glaciology of this advance is comparable to the importance to oceanography of the development of the sonic echo sounder to replace the lead line. During the IGY an individual sounding of the ice sheet, by seismic shooting, took 2 or 3 hours, and was followed a day later by another sounding some tens of kilometers away. Radar sounding, in contrast, takes less than 50 μsec for an individual measurement and yields a continuous profile of ice thickness at aircraft flying speed along the flight path. To date, more than half the ice sheet has been mapped at a flight-line spacing of 100 km or better (*14*).

Radar sounding yields much more information than simply the ice thickness (*15*). Of particular interest are the many internal layers in the Antarctic (and Greenland) ice sheets that are associated with small density changes, presumably depositional or diagenetic, and with changes in electrical conductivity associated with the fallout of volcanic ash. The internal reflections provide time horizons within the ice sheet that can be invaluable in studying the patterns of ice flow. Changes in the character of the basal echo have also proven useful, particularly in delineating what appear to be lakes beneath the ice sheet where the ice is thick enough that its bed is at the melting point. Crevassing near the surface is a particular feature of the giant fast-moving "rivers" of ice within the ice sheet called "ice streams," which are now known to transport most of the ice from the Antarctic interior. The crevasses produce a striking difference in the radargram recorded over them compared to their surroundings, a difference which makes it possible to identify such streams easily even where the crevasses are buried and invisible.

The advent of radar sounding has not, however, relegated seismic sounding of ice sheets to the trash bin. On the floating ice shelves, whose importance to the dynamic balance of the ice sheet becomes ever more apparent, seismic sounding can be combined with radar sounding to give both ice thickness and water depth below the ice (radio waves do not penetrate seawater). A 5-year project in the mid-1970s on the Ross Ice Shelf yielded a subglacial morphology map of the entire "Ross embayment" (*16*), which shows no indication in the subglacial morphology of either the outer edge of the ice shelf or the boundary between the grounded and floating ice, a fact that points up the impermanence of the glacial boundaries.

Another technological revolution in glaciology resulted from the development of Doppler-satellite positioning. During the IGY it was not possible to measure the velocity of ice movement anywhere except close to rock outcrops, which in Antarctica are few and far between. Satellite positioning now has made it possible to obtain a position accurate to a few meters in just a few hours. Measurements repeated at the same spot on the ice after a suitable time period (a few months to a few years, depending on the speed of ice movement) yield precise flow velocities. For example, a velocity-vector map of the entire Ross Ice Shelf was produced in this way (*17*).

The ability to map the ice thickness and ice movement in detail has made it possible to examine the dynamic behavior of the ice sheet, particularly that of West Antarctica. The importance of the West Antarctic ice sheet stems from its expected response to climatic warming produced by increasing carbon dioxide, methane, and other greenhouse gases in the atmosphere. If shrinkage of the Antarctic ice sheet is going to contribute to a significant rise of global sea level, it will not be primarily because of an increased rate of melting in expanding ablation areas, but rather because the "marine" parts of the ice sheet, those parts that rest on a bed far below sea level and therefore, at least theoretically, tend to be unstable, might begin to thin and retreat, causing accelerated discharge of ice into the ocean (*18*). The immense East Antarctic ice sheet is mostly terrestrial, but the smaller ice sheet of West Antactica is mostly marine and is therefore the focus of attention in studying the response of the Antarctic ice masses to climatic change.

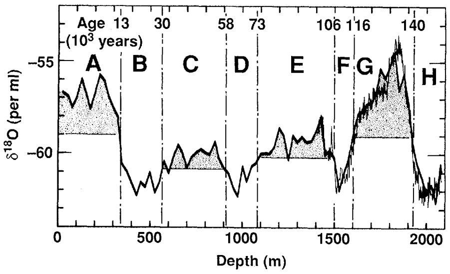
A key question in evaluating the stability of the West Antarctic ice sheet is why the ice streams, which transport most of the interior ice, move so fast. Exciting new results relating to that question have recently been reported. Seismic sounding has shown that beneath one ice stream there is a subglacial till layer several meters thick that is charged with water; the deformation of this layer by the overriding ice is a likely mechanism for fast "sliding" (*19*). Another intriguing aspect of ice stream behavior is that what clearly was once a major Antarctic ice stream has stopped its rapid motion, probably within the last few centuries (*20*).

The precipitation that falls in the dry snow zone of a high polar glacier (where there is no summer melting) buries the previously fallen snow; with increased depth the snow gradually transforms into glacier ice. The resulting annual stratigraphic sequences comprise not only aggregates of ice crystals but also atmospheric air, trapped when the pore spaces closed off, together with all of the foreign material that fell either with the precipitation or as dry fallout on the snow surface. Thus, ice sheets of Greenland and Antarctica preserve a widely varied record of climatic change and environmental history that can be examined by deep core drilling (*21*).

Five deep core holes now exist that penetrate at least into ice that fell on the surface of the ice sheet during the last ice age. No aspect of Antarctic glaciology has produced more important results than the analyses of these deep ice cores. We will briefly consider only four aspects—oxygen isotope ratios, carbon dioxide and methane levels, and microparticle deposits.

Variations in oxygen isotopes with depth yield extremely valuable information about paleotemperatures. Studies have shown that a close correlation exists between the $^{18}O:^{16}O$ isotope ratio and the surface temperature on the ice sheets at the time of deposition. Seasonal variations in the temperature, and therefore in the isotope ratio, produce seasonal variations that are preserved in the ice core and can be used, simply by counting annual layers, to date ice as old as 15,000 years. But the most striking feature of the isotopic record is the clear and detailed evidence of the alternation between glacial and interglacial epochs. The best record of paleotemperature that has been produced to date comes from an ice core at the Soviet station at Vostok, where the record extends back 150,000 years (*22*). The record in that core shows that the present Holocene epoch

Fig. 3. Oxygen-18 versus depth in an ice core from Vostok, in central East Antarctica. The age scale is approximate. More negative values of ^{18}O mean colder temperatures (*25*).



was preceded by a long glacial time marked by two relatively warm interstadials (Fig. 3). The well-marked last interglacial (section G in Fig. 3) at Vostok was significantly warmer than the Holocene, and the late stages of the previous glacial event, about 150,000 years ago, were similar in temperature to the most recent glacial maximum around 20,000 years ago.

Analyses of the air extracted from the bubbles in the ice have revealed the remarkable fact that the atmospheric CO_2 concentration during the last glacial maximum was only 60% of today's value and only 75% of its value even before anthropogenic contamination (*23*). Furthermore, the change from low to high CO_2 concentration occurred rapidly. One detailed study of ice age cores from Greenland indicates that the CO_2 content of the atmosphere and atmospheric temperature changed concordantly with a time lead or lag of no more than about a century (*24*). Whether the CO_2 variations are merely an effect of climate change or are a contributor to, or even an initiator of, climate change is not known yet.

Another surprising discovery of the studies of entrapped gas has been the demonstration that the concentration of methane about 200 years ago was less than half the present level (*25*). This result, which has been attributed by some to englacial chemical reactions, has recently been confirmed as an atmospheric effect by measurements of solar spectra at a high Alpine station (*26*). Whether the increase in methane with time is due entirely to increased emissions or at least in part to chemical processes in the atmosphere (that is, reduction in the hydroxyl radical concentration) is not yet known. In either case, however, the implications for climate are important, since methane is one of the greenhouse gases.

Microparticles are transported aerodynamically, so their distribution and subsequent deposition is related closely to the local and global atmospheric circulation regimes. In addition, temporal fluctuations in the number of particles deposited depend upon the frequency and intensity of volcanic eruptions and variations in continental aridity. The deposition rate of small particles in Greenland, most of them terrestrially (rather than volcanically) derived, was up to 100 times as great in the last ice age as at present (*27*); in the Antarctic it was an order of magnitude greater (*23*). The larger flux of terrestrially derived particles suggests a more intense atmospheric circulation and more exposed, unvegetated land areas, presumably on the present continental shelves, during the last glacial age than at present.

Cenozoic Paleoclimatic History

Geological research in Antarctica has made some major contributions to our understanding of earth history. Among the more significant is the discovery of fossil vertebrates, all of Triassic age, which provided the compelling paleontological evidence for continental drift (*28*). Another major discovery was the occurrence of silicified peats of Permian and Triassic age. These deposits are similar to the coal balls of the Carboniferous of Europe and North America and can be expected to yield similar detailed information on such aspects as the anatomy and pathology of the Permian and Triassic plants (*29*). But perhaps the single most significant contribution lies in the history of Cenozoic glaciation because of its importance to global paleoclimatology.

The establishment of the Antarctic ice sheets and the associated climatic cooling is the most significant environmental event in the Cenozoic. It has had a profound effect on world climate and thus on the history and development of marine and terrestrial biota. The factors that led to the inception of the ice cover are far from clear, though the isolation and polar position of Antarctica, attained as a result of breakup of Gondwanaland and subsequent plate movements (Fig. 4), must have played a role. One consequence has been the development of circumpolar circulation of water masses and unrestricted circulation of the atmosphere, the two combining to give a physical isolation that is reflected in the present climatic extremes of the continent and one of the most pronounced oceanic boundaries on the earth, the Antarctic Convergence.

How the present climate of the earth, with strong latitudinal temperature gradients, evolved from the much more equable climate of the late Mesozoic and early Cenozoic is a question for which Antarctica may hold some of the answers. The inception of the ice cover and the subsequent fluctuations in ice volume are events of great importance.

The long-term climatic trends during the Cenozoic have been inferred from a variety of sources. Probably the most widely used and recognized is the oxygen isotope record obtained by analysis of foraminiferal tests recovered from deep-sea cores. The record shows a gradual increase in $\delta^{18}O$ values of both planktonic and benthic foraminifera from a low in the Middle Eocene to a culmination in an abrupt enrichment near the Eocene-Oligocene boundary. Fluctuating isotopic ratios followed until another abrupt increase in benthic $\delta^{18}O$ values in the early Middle Miocene. Two major factors contribute to oxygen isotopic variations: isotopic fractionation between water and calcite, controlled mainly by temperature, and changes in the isotopic ratio of the ocean, influenced by the amount of ^{18}O-impoverished ice on land. The traditional interpretation is that the first ^{18}O enrichment at the Eocene-Oligocene boundary represents the establishment of cold bottom waters throughout the oceans, resulting from the initiation of sea-ice formation around Antarctica, and that the mid-Miocene enrichment reflects the formation of the Antarctic Ice Sheet (*30*). This view has been challenged (*31*) and arguments advanced for major ice buildup (ice sheet formation or expansion) at the Eocene-Oligocene boundary, with the mid-Miocene ^{18}O enrichment reflecting a change in average bottom water temperature.

Intensive study of the late Quaternary to Recent record contained in deep-sea piston cores from around Antarctica has yielded estimates of global conditions at the last glacial maximum and geological support for the Milankovitch theory of Ice Ages (*32*). Detailed studies of cores have also demonstrated fluctuations in the northward extent of ice-rafted debris, the biosiliceous zone, and the Polar Front (*33*). Widespread disconformities and sediment regime fluctuations reflect the influence of Antarctic Bottom Water and other circum-Antarctic water masses, all of which show significant variations through time (*34*). The floras and faunas from Deep Sea Drilling Project (DSDP) cores exhibit changes at or near the Eocene-Oligocene and Oligocene-Miocene boundaries that mark, respectively, the establishment of assemblages with distinct polar characteristics, and the transition to present-day steep faunal and floral diversity gradients (*35*). The inferred paleoenvironmental changes in the Southern Ocean through time have been argued to reflect climatic changes that are directly related to ice conditions on and around the Antarctic continent.

The deep-sea record provides relatively indirect evidence for glacial conditions; more direct evidence of glacial advance and retreat in the Southern Hemisphere must be sought in New Zealand, Patagonia, and Antarctica. The antiquity of Cenozoic glaciation is now well supported, though questions remain about the extent and location of the ice cover through time. Arguments for an early Cenozoic ice cover should take into account the implications of the derived flora contained in shelf sediments from around Antarctica (*36*). Recent paleontological and paleomagnetic work by U.S. scientists on sediment cores recovered by the New Zealand Antarctic Research Program (MSSTS-1 drill core) from McMurdo Sound suggests that glacial sedimentation had begun by at least 30 million years ago (*37*). This sediment and others as old as 25 million years recorded by DSDP in the Ross Sea (*38*) suggest the presence of floating ice that calved from tidewater glaciers, but give no direct evidence of an ice sheet in either West or East Antarctica. On land, subglacially erupted volcanic rocks (hyaloclastites) are common in Marie Byrd Land and are as old as 27 million years (*39*). Some hyaloclastite occurrences suggest eruption beneath an ice sheet rather than beneath glaciers flanking volcanoes. Physical evidence for terrestrial ice, such as glacial grooves and striated pavements, is widely scattered, but the oldest age that can be established is only 7.0 million years (*40*). As always, it is difficult to determine whether glaciation was local or regional in extent.

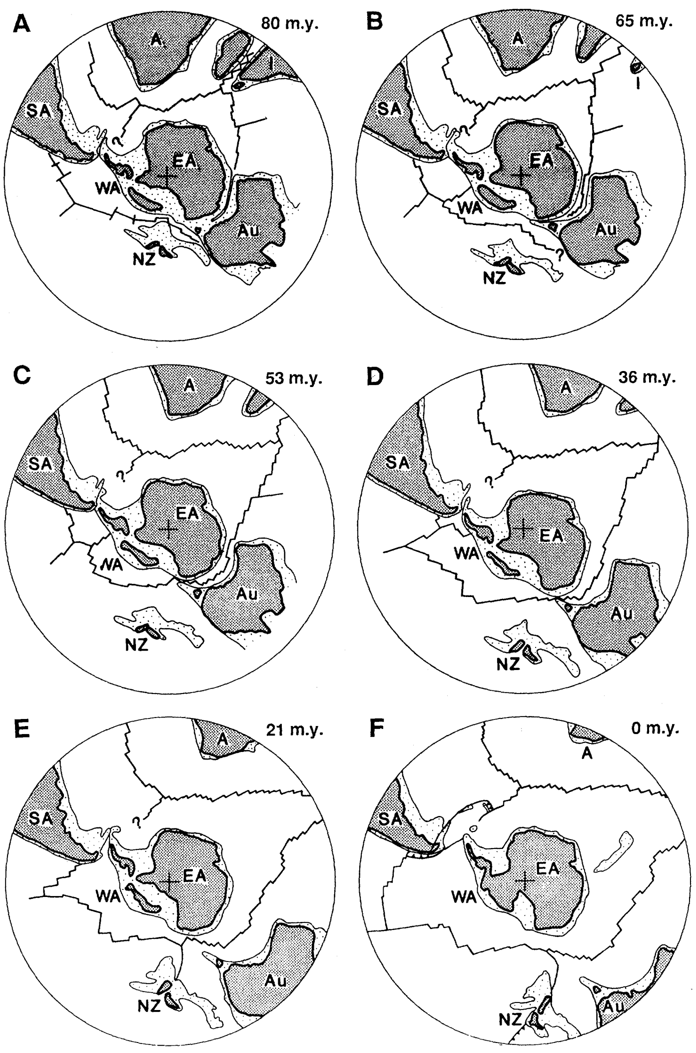


Fig. 4. Evolution of the Southern Ocean from Late Cretaceous time to the present to show the growing physical isolation of the Antarctic continent: (**A**) 80 million years (m.y.), Late Cretaceous (Campanian); (**B**) 65 m.y., early Paleocene (Danian); (**C**) 53 m.y., early Eocene; (**D**) 36 m.y., early Oligocene; (**E**) 21 m.y., early Miocene; (**F**) present day. Light stipple indicates continental shelves; heavy stipple, continents (both present day). Lines in oceanic regions represent mid-ocean ridge spreading centers and connecting transform faults; lines with teeth represent subduction zones. SA, South America; A, Africa; I, India; Au, Australia; NZ, New Zealand; EA, East Antarctica; WA, West Antarctica; West Antarctica is illustrated as it would be today if the ice cover were removed. The paleogeography of West Antarctica during the Late Cretaceous and at least the early Cenozoic is poorly known but it is likely to have included shallow seaways cutting across the margin and connecting with the Pacific as well as seaways across the interior; East Antarctica is likely also to have had seaways allowing shallow water movement across the continental interior. [Adapted from Kennett in (*35*)]

McMurdo Sound, the Dry Valleys of South Victoria Land, and the central Transantarctic Mountains have yielded much information on glacial history. Microfossils from marine sediments show the Dry Valleys were fjords sometime during the late Miocene and Pliocene (*41*). Extensive mapping of deposits and erosional features in the Dry Valleys have allowed a largely relative chronology of glacial events to be established (*42*). In addition to the detailed chronology of young events, evidence is now accumulating for one or more periods of submergence of the Transantarctic Mountains beneath the ice (*43*). Past surface elevations, and hence volumes, of the ice sheet are difficult to establish because the ages of overriding are poorly constrained and only long-term average rates of uplift of the mountains can be inferred from fission track dating and other data (*44*). Glacial deposits and erosional features are found at present elevations of 2987 m in South Victoria Land, and over 4000 m in the Beardmore Glacier region (*43, 45*). Higher ice levels are also recorded in the Ellsworth Mountains and adjacent nunataks (*46*), and elsewhere.

Tills that lack a direct relation to present glacial drainage occur in the Transantarctic Mountains at elevations up to 2700 m. These tills have yielded marine microfossil floras and faunas and wood fragments and other plant debris, including pollen and spores (*47*). The marine microfossils, contained in mud clasts as well as dispersed in the tills, include a diatom flora dominated by Pliocene taxa. The diatoms require open marine waters for at least part of the year, and Webb and others (*47*) argue that the open water lay over basins located in East Antarctica. Such open waters point to much greater ice volume changes than previously thought. Furthermore, the presence of wood and a sparse, low diversity pollen and spore assemblage suggest not only deglaciation, but also temperatures warm enough at 85°S to support a sparse vegetation, possibly of tundra type (*47*). Webb and others (*48*) argue that the flora grew at sea level and that substantial uplift has occurred since deposition of the beds. Although conditions considerably warmer than at present have been inferred for the early Pliocene (*49*), neither the magnitude of deglaciation suggested by the marine microfossils, nor the warmth of the climate inferred from the plant material are indicated in the commonly accepted interpretation of the oxygen isotope record from deep-sea cores.

Many problems remain in establishing a well-dated chronology of glacial events in Antarctica from which paleoclimatic inferences can be drawn. Why the inferred magnitude of the Pliocene deglaciation (that is, the inferred absolute minimum ice volume) is not evident in the $\delta^{18}O$ curves from deep-sea cores is an unanswered question. The extent of climatically forced faunal and floral migrations across latitude and the diachroneity of microfossil range are unresolved issues. The dating of the glacial deposits themselves is a major problem as are also the timing and rates of uplift of the Transantarctic Mountains (*48*). The impetus provided by these recent discoveries will surely lead in the coming years to some major advances in understanding climatic history and should help clarify and strengthen the links being postulated between the Antarctic ice sheet, the late Cenozoic climate and the evolution of biota—particularly the evolution of man (*50*).

Biological Adaptations

An adaptation is a genetically determined characteristic that improves an organism's ability to survive and reproduce successfully under prevailing environmental conditions. Studies of adaptation at all levels of the biological hierarchy from cellular enzyme systems through ecosystems have been a major focus in the Antarctic life sciences (*51*). This is seen in the biological research of all nations active in Antarctica and is likely to remain so for two principal reasons. First, adaptation through natural selection is perhaps the most important notion in biological evolution and second, most organisms in Antarctica are living at the extreme end of the primary resource gradients of energy and nutrients.

By highlighting studies of adaptation to extreme cold and drought, we are not denying the value of other areas of Antarctic biological research, such as the role played by the southern oceans in biospheric functions, but we are rather seeking to discuss this most intriguing and often publicized aspect of Antarctic biology. Close scrutiny of many of the reported adaptations that sustain life, even in endemic and possibly autochthonous taxa, shows that these adaptations are found not only in other polar regions and in cold temperate systems, but they may even be found in tropical systems (*52*). Often an adaptation under discussion is one of degree of operation rather than uniqueness to Antarctica. Of course at the gene and enzyme homolog level unique molecular configurations and functions do occur. Regardless of their novelty all Antarctic adaptations are justifiably studied in the natural Laboratory Antarctica. The 1906 statement of Brown (*53*) remains true, "it is . . . extremely desirable that . . . physiological and morphological questions should be studied on the spot, indeed, the impracticability of satisfactory investigation in any other circumstances is most obvious."

Biogeographers recognize two broad regions in Antarctica: maritime and continental (*54*). Temperatures in the Maritime Antarctic Peninsula and the associated islands to the north average about −10°C in winter and in summer exceed freezing by only a little for 1 to 4 months. Precipitation falls mostly as rain in summer and ranges from 350 to 500 mm water equivalent per year. Continental Antarctica is colder and drier, and most life is restricted to mountains and nunataks that rise above the ice caps and glaciers or to the coastal fringes where temperatures are above freezing for only a few days or weeks each year and where total annual precipitation is between 100 and 150 mm of water equivalent.

In the Maritime Antarctic, there is a decreasing gradient of terrestrial organism diversity and productivity from north to south. In the north of the Antarctic Peninsula, a complete cover of vegetation may occur in favorable places. This vegetation may include moss carpets and the only two species of native vascular plants. In some locations, the balance of moss growth and decay may be such that peat accumulates. In the south, the two vascular plants are less frequent but still occur, closed stands of vegetation are rare, and there is no accumulation of peat. In the Continental Antarctic, vascular plants do not occur and peat does not accumulate. Here, terrestrial ecosystems are restricted to nunataks where surfaces become free of snow and ice, and to the coastal areas where there are sporadic beaches and glacier forelands. In southern Victoria Land there are extensive snow and ice-free dry valleys. Most of the continental Antarctic is covered with permanent snow and ice and is devoid of life, apart from patches of algae in melting snow and occasional microorganisms and stray birds. Nevertheless, lichen are found throughout the Transantarctic Mountains, including areas only about 300 km from the South Pole at Mount Howe.

The ice-free terrestrial environment, which represents only 2.4% of the continent, has associated biological communities that are patchy and heterogeneous. Nevertheless, whenever and wherever the appropriate resources for life exist the site is occupied. For example, in the Dry Valleys endolithic lichens live beneath rock surfaces (*54*). Adélie penguins occupy pebble beaches of sufficient size and nearness to the ocean (*55*), and mosses grow where there are sufficient water, dissolved nutrients, and sunshine.

The marine environment fluctuates less than the terrestrial system. Water temperatures of the ocean surrounding the continent seldom exceed 2°C. McMurdo Sound, the site of much of the research on marine adaptations, remains largely ice-covered. Water temperatures average −1.9°C with a small yearly range of less than 1°C so that even during the austral summer the sea water temperature remains less than 0°C. Through all but the summer, when the sea ice usually breaks up, ice crystals abound beneath the surface—that is, in shallow water and on the sea bottom, down to depths of about 30 m.

Terrestrial organisms mostly adapt through avoidance or passive tolerance. Many, if not all, of their adaptations are preadaptations, often in the form of phenotypic plasticity that evolved in a former environment and few, if any, of the adaptations are known only from Antarctica (*54, 56*). Physiological adaptations certainly occur; in plants, photosynthesis and respiration can operate at low temperatures (*54*) presumably using the same mechanisms that are better known in alpine and arctic plants (*57*). The best known terrestrial adaptation is that of the cryptoendolithic organisms, which have been carefully studied in the frigid Dry Valleys of Antarctica (*58*). These organisms (bacteria, blue-green algae, green algae, yeasts, filamentous fungi, and lichens) live in the airspace system of porous, light-colored rocks. They live in a protected nannoclimate within a few millimeters of the rock surface where light and gases can still penetrate. The endolithic habit has been considered as a possible strategy for some hypothetical Martian organisms, since the Dry Valleys may be the closest terrestrial analogs of the Martian environment (*59*).

Adaptations reported for Antarctic marine systems are active and more striking than those for terrestrial systems. This is especially true for fish. Enzymatic adaptations, that ensure adequate metabolism at low temperatures, including lower activation energy and greater substrate affinity, have been well documented for both Arctic and Antarctic fish and invertebrates (*60*). These adaptations come under the heading of capacity extension. Some endemic Antarctic fish have no blood hemoglobin (*61*). Other polar (Arctic and Antarctic) and some temperate fish may also have reduced hemoglobin levels, especially in winter (*62*). A partial explanation of this reduction in hemoglobin is that oxygen dissolves more readily in cold water than in warm water, turbulent Antarctic waters are often oxygen saturated, and certain other hemoglobin functions are not required in the cold and thermally stable waters of locales such as McMurdo Sound (*63*). The adaptive significance of the absence of hemoglobin in blood is not clear, since these fish have modifications to their circulation systems, such as larger hearts, wider blood vessels, greater surficial capillary systems, and modified gill structures, apparently as compensation for loss of the oxygen-carrying capacity of hemoglobin (*64*). The most famous fish adaptation is the lowering of body freezing point with increased solutes and antifreeze agents such as salt, proteins, glycoproteins, and glucose. It has been found that proteins and glycoproteins do not lower the freezing point by the colligative properties of the antifreeze agents alone, but do so by preventing ice crystal growth down to temperatures of −2.2°C (*65*). Arctic fish show seasonal fluctuations in antifreeze proteins whereas Antarctic fish do not. The fish of McMurdo Sound are obligate stenotherms, that is, they cannot live in warmer water or water with a fluctuating temperature and have little power of acclimation compared to Arctic fish with similar antifreeze mechanisms (*65, 66*). It would appear that these fish have evolved in, or in conditions similar to, McMurdo Sound where the temperature is nearly constant at −1.9°C and where acclimation has been unnecessary.

It has been pointed out that most of the successful polar organisms are widely distributed groups that use preadaptations to survive (*67*). This may well be a result of the relative youth of present polar biota compared to the more temperate biota of lower latitudes (*52, 68*) and of insufficient time in polar regions for natural selection of new adaptations. Certainly more unusual adaptations are seen in the marine systems than in the younger terrestrial systems, which have been subject to glacial scouring and are only relatively recently colonized. Also, the Antarctic marine system, which is much older than the Arctic marine system, has more endemics (*69*) and more highly specialized adaptations (*55*).

The effect of global climatic change and the predictions of increased precipitation and temperature at the poles could have a significant effect on polar biology and are worth monitoring and long-term study. Certainly Antarctica is not isolated from the rest of the planet, as demonstrated by long-distance transport of man-made chlorinated hydrocarbons to Antarctica (*70*). One would expect greater production in terrestrial systems where heat, length of growing season, and water are limiting to growth and where considerable powers of organism plasticity and acclimation are known. In time one would also expect greater biological diversity. The majority of Antarctic marine systems would probably show similar trends. The fate of the obligate stenotherms is a matter of greater conjecture and they may be threatened. There may not always be sufficient appropriately frigid and thermally stable ocean water available around Antarctica to support these restricted species.

Conclusion

Of the numerous fields of research involving U.S. scientists in Antarctica, only a few have been selected for review in this article. Much of the research in the Antarctic is of the "big science" variety for scientific and logistic reasons. Antarctic research will likely be of even greater importance, and will play a more significant international role in the coming era of the large, coordinated global geoscience projects.

As global and international connections become more important in understanding planet earth, so does the importance of research on these connections in "laboratory Antarctica."

REFERENCES AND NOTES

1. L. J. Lanzerotti and S. M. Krimigis, *Phys. Today* **38** (no. 11), 24 (1985).
2. Much of the U.S. upper atmosphere research in Antarctica, following the IGY to the late 1970s, was summarized in L. J. Lanzerotti and C. G. Park, Eds. [*Upper Atmosphere Research in Antarctica* (American Geophysical Union, Washington, DC, 1978)]. The role of Antarctic investigations during the International Magnetosphere Study (1976–1980) are summarized by T. J. Rosenberg [in *The IMS Source Book*, C. T. Russell and D. J. Southwood, Eds. (American Geophysical Union, Washington, DC, 1982), p. 182], M. J. Rycroft [in (*3*), p. 196], and T. Nagata and T. Hirasawa [in (*3*), p. 188)] for research sponsored by the United States, the United Kingdom, and Japan, respectively.
3. R. A. Helliwell and J. P. Katsufrakis, in *Upper Atmosphere Research in Antarctica*, L. J. Lanzerotti and C. G. Park, Eds. (American Geophysical Union, Washington, DC, 1978).
4. R. A. Helliwell, D. L. Carpenter, T. R. Miller, *J. Geophys. Res.* **85**, 3360 (1980).
5. R. A. Helliwell, *Radio Sci.* **6**, 801 (1983).
6. R. L. Arnoldy, K. Dragoon, L. J. Cahill, Jr., S. B. Mende, T. J. Rosenberg, *J. Geophys. Res.* **87**, 10449 (1982); L. J. Lanzerotti and T. J. Rosenberg, *ibid.* **88**, 9115 (1983); U. S. Inan, D. L. Carpenter, R. A. Helliwell, J. P. Katsufrakis, *ibid.* **90**, 7457 (1985).
7. T. J. Rosenberg, R. A. Helliwell, J. P. Katsufrakis, *ibid.* **76**, 8445 (1971); J. L. Roeder, E. A. Bering, J. R. Benbrook, W. R. Sheldon, *ibid.* **90**, 10975 (1985).
8. R. H. Eather, *ibid.* **90**, 1569 (1985).
9. J. Christensen-Dalsgaard, D. O. Gough, J. Toomre, *Science* **229**, 923 (1985); J. W. Leibacher, R. W. Noyes, J. Toomre, R. K. Ulrich, *Sci. Am.* **253**, 48 (March 1985); F.-L. Deubner and D. O. Gough, *Ann. Rev. Astron. Astrophys.* **22**, 593 (1984).
10. G. Grec, E. Fossat, M. Pomerantz, *Nature (London)* **288**, 541 (1980); G. Grec, E. Fossat, M. Pomerantz, *Solar Phys.* **74**, 59 (1981); J. Harvey, in *Proceedings of the ESA Workshop on Future Missions in Solar, Heliosphere and Space Plasma Physics* (ESA SP-235, European Space Agency, Paris, 1985), p. 199.
11. T. L. Duvall, Jr., and J. W. Harvey, *Nature (London)* **302**, 24 (1983).
12. ——, M. A. Pomerantz, *ibid.* **321**, 500 (1986).
13. M. Draganovic and A. A. Stark, personal communication.
14. D. J. Drewry, Ed., *Antarctica: Glaciological and Geophysical Folio* (Scott Polar Research Institute, University of Cambridge, Cambridge, 1983).
15. V. V. Bogorodsky, C. R. Bentley, P. E. Gudmandsen, *Radioglaciology* (Reidel, Dordrecht, Netherlands, 1985).
16. C. R. Bentley and K. C. Jezek, *J. Roy. Soc. N. Z.* **11**, 35 (1981).
17. R. H. Thomas, D. R. MacAyeal, D. H. Eilers, D. R. Gaylord, *Antarctic Res. Ser.* **42**, 21 (1984).
18. Polar Research Board, *Glaciers, Ice Sheets, and Sea Level: Effects of a CO_2-Induced Climatic Change* (National Academy Press, Washington, DC, 1985).
19. D. D. Blankenship, C. R. Bentley, S. T. Rooney, R. B. Alley, *Nature (London)* **322**, 54 (1986); R. B. Alley, D. D. Blankenship, C. R. Bentley, S. T. Rooney, *ibid.*, p. 57.
20. R. A. Bindschadler, D. R. MacAyeal, S. N. Stephenson, *Dynamics of the West Antarctic Ice Sheet*, C. J. van der Veen and J. Oerlemans, Eds. (Reidel, Dordrecht, Netherlands, 1987), p. 161; I. M. Whillans, personal communication.
21. G. de Q. Robin, Ed., *The Climatic Record in Polar Ice Sheets* (Cambridge Univ. Press, Cambridge, 1983).
22. C. Lorius *et al.*, *Nature (London)* **316**, 591 (1985).
23. C. Lorius, D. Raynaud, J.-R. Petit, J. Jouzel, L. Merlivat, *Ann. Glaciol.* **5**, 88 (1984).
24. B. Stauffer, H. Hofer, H. Oeschger, J. Schwander, U. Siegenthaler, *ibid.*, p. 160.
25. R. A. Rasmussen and M. A. K. Khalil, *J. Geophys. Res.* **32**, 176 (1984).
26. C. P. Rinsland, J. S. Levine, T. Miles, *Nature (London)* **318**, 245 (1985).
27. C. U. Hammer, H. B. Clausen, W. Dansgaard, A. Neftel, P. Kristinsdottir, N. Reeh, *Am. Geophys. Union Monogr.* **33**, 90 (1985).
28. D. H. Elliot, E. H. Colbert, W. J. Breed, J. A. Jensen, J. S. Powell, *Science* **169**, 1197 (1970); J. W. Kitching, J. W. Collinson, D. H. Elliot, E. H. Colbert, *ibid.* **175**, 524 (1972); W. R. Hammer and J. W. Cosgriff, *J. Paleontol.* **55**, 410 (1981).
29. J. M. Schopf, *Science* **169**, 274 (1970); T. N. Taylor, E. Taylor, J. W. Collinson, *Antarct. J. U.S.* **21** (no. 5) 26 (1986); T. N. Taylor, E. Taylor, J. W. Collinson, D. H. Elliot, *ibid.*, p. 27.
30. N. J. Shackleton *et al.*, *Init. Rep. Deep Sea Drill. Project* **29**, 743 (1975); J. P. Kennett, *J. Geophys. Res.* **82**, 3843 (1978).
31. R. K. Matthews and R. Z. Poore, *Geology* **8**, 501 (1980).
32. J. D. Hays, J. A. Lozano, N. Shackleton, G. Irving, *Geol. Soc. Am. Mem.* **145**, 337 (1976); J. D. Hays, J. Imbrie, N. J. Shackleton, *Science* **194**, 1121 (1976).
33. P. F. Ciesielski, M. T. Ledbetter, B. B. Ellwood, *Mar. Geol.* **46**, 1 (1982).
34. P. F. Ciesielski *et al.*, *Init. Rep. Deep Sea Drill. Project* **71**, 461–477 (1983); N. I. Osborn, P. F. Ciesielski, M. T. Ledbetter, *Geol. Soc. Am. Bull.* **94**, 1345 (1983); M. T. Ledbetter, P. F. Ciesielski, *Palaeogeogr. Palaeoclimat. Palaeoecol.* **52**, 1985 (1986).
35. J. P. Kennett, *Mar. Micropalaeontol.* **3**, 301 (1978); R. M. Leckie and P. N. Webb, *Init. Rep. Deep Sea Drill. Project* **90**, 1093 (1985).
36. E. M. Truswell, D. J. Drewry, *Mar. Geol.* **59**, 187 (1984).
37. D. M. Harwood, *N. Z. Dept. Sci. Indust. Res. Misc. Bull.* **237**, 69 (1986); P. N. Webb, *ibid.*, p. 115.
38. D. E. Hayes *et al.*, *Init. Rep. Deep Sea Drill. Project* **28** (1975).
39. W. E. LeMasurier and D. C. Rex, in *Antarctic Earth Science*, R. L. Oliver, P. R. James, J. B. Jago, Eds. (Australian Academy of Sciences, Canberra, 1983), pp. 663–670.
40. R. H. Rutford, C. Craddock, C. M. White, R. L. Armstrong, in *Antarctic Geology and Geophysics*, R. J. Adie, Ed. (Universitetsforlaget, Oslo, 1972), pp. 239–243.
41. P. N. Webb, *J. Foram. Res.* **4**, 184 (1974); H. T. Brady, *Mem. Natl. Inst. Polar Res. Spec. Iss.* **13**, 150 (1979); L. H. Burckle, M. L. Prentice, G. H. Denton, *Eos* **67**, 295 (1986); S. E. Ishman and P. N. Webb, *Third International Symposium on Benthic Foraminifera* (Museum of Natural History, Geneva, 1986), abstract.
42. M. Stuiver, G. H. Denton, T. J. Hughes, J. L. Fastook, in *The Last Great Ice Sheets*, G. H. Denton and T. J. Hughes, Eds. (Wiley, New York, 1981), pp. 319–336.
43. G. H. Denton, M. L. Prentice, D. E. Kellogg, T. B. Kellogg, *Geology* **12**, 203 (1984).
44. A. J. W. Gleadow, B. C. McKelvey, K. U. Ferguson, *N. Z. J. Geol. Geophys.* **27**, 457 (1984).
45. G. H. Denton, personal communication.
46. ——, J. G. Bockheim, R. H. Rutford, B. G. Anderson, *Geol. Soc. Am. Mem.*, in press.
47. P. N. Webb, D. M. Harwood, B. C. McKelvey, J. H. Mercer, L. D. Stott, *Geology*, **12**, 287 (1984); R. A. Askin and V. Markgraf, *Antarct. J. U.S.* **21** (no. 5), 34 (1986); P. N. Webb, D. M. Harwood, B. C. McKelvey, M. C. G. Mabin, J. H. Mercer, *ibid.*, in press.
48. P. N. Webb, D. M. Harwood, B. C. McKelvey, M. C. G. Mabin, J. H. Mercer, *Antarct. J. U.S.* **21** (no. 5), 99 (1986).
49. J. Keany, *Mar. Micropaleontol.* **3**, 35 (1978).
50. G. H. Denton, *S. Afr. J. Sci.* **81**, 224 (1985).
51. G. A. Llano, *Adaptations Within Antarctic Ecosystems* (Smithsonian Institution, Washington, DC, 1977).
52. M. J. Dunbar, in *Antarctic Ecology*, M. W. Holdgate, Ed. (Academic Press, New York, 1970), vol 1, p. 105; P. F. Scholander, W. Flagg, V. Walters, L. Irving, *Physiol. Zool.* **26**, 67 (1953).
53. R. N. R. Brown, *Scott. Geogr. Mag.* (1906), p. 473.
54. R. I. Lewis Smith, in *Antarctic Ecology*, R. M. Laws, Ed. (Academic Press, London, 1984), p. 61.

55. W. S. Benninghoff and W. N. Bonner, *Man's Impact on the Antarctic Environment* (Scientific Committee on Antarctic Research, International Council of Scientific Unions, Scott Polar Research Institute, Cambridge, United Kingdom, 1985), p. 56.
56. V. Ahmadjian, in *Antarctic Ecology*, M. W. Holdgate, Ed. (Academic Press, New York, 1970), vol. 2, p. 801; E. D. Rudolph, in *Antarctic Soils and Soil Forming Processes*, J. C. F. Tedrow, Ed. (American Geophysical Union, Washington, DC, 1966), vol 8, p. 105.
57. F. S. Chapin III and G. R. Shaver, in *Physiological Ecology of North American Plant Communities*, B. F. Chabot and H. A. Mooney, Eds. (Chapman & Hall, New York, 1985), p. 16; L. C. Bliss, in *ibid.*, p. 41.
58. E. I. Friedman, *Science* **215**, 1045 (1982).
59. N. J. Horowitz, R. E. Cameron, J. S. Hubbard, *ibid.* **176**, 242 (1972).
60. B. L. Umminger, in *Adaptations Within Antarctic Ecosystems*, G. A. Llano, Ed. (Proceedings of the Third SCAR Symposium on Antarctic Biology, Smithsonian Institution, Washington, DC, 1977), p. 347.
61. J. C. Hureau, D. Petit, J. M. Fine, M. Marneux, *ibid.*, p. 459.
62. R. M. G. Wells, M. D. Ashby, S. J. Duncan, J. A. MacDonald, *J. Fish Biol.* **17**, 517 (1980).
63. A. M. Slicher and G. E. Pickford, *Physiol. Zool.* **41**, 293 (1968).
64. E. A. Hemmingsen and E. L. Douglas, in *Adaptations Within Antarctic Ecosystems*, G. A. Llano, Ed. (Smithsonian Institution, Washington, DC, 1977), p. 479.
65. A. L. DeVries, in *Polar Research to the Present and the Future*, M. A. McWhinnie, Ed. (Westview, Boulder, CO, 1978), p. 175; A. L. DeVries, *Comp. Biochem. Physiol.* **73A**, 627 (1982).
66. P. W. Hochachka and G. N. Somero, *Biochemical Adaptations* (Princeton Univ. Press, Princeton, NJ, 1984).
67. P. J. Darlington, *Zoogeography: The Geographical Distribution of Animals* (Wiley, New York, 1966), pp. 675.
68. L. C. Bliss, in *Tundra Ecosystems: A Comparative Analysis*, L. C. Bliss, O. W. Heal, J. J. Moore, Eds. (Cambridge Univ. Press, Cambridge, 1981), p. 813.
69. M. G. White, in *Adaptations Within Antarctic Ecosystems*, G. A. Llano, Ed. (Smithsonian Institution, Washington, DC, 1977), p. 197.
70. R. W. Risebrough and M. Carmigniani, in *Conservation Problems in Antarctica*, C. Parker, Ed. (Allen, Lawrence, KS, 1972), p. 63.
71. We wish to thank the following for their comments, suggestions, and critical reviews of various parts of the manuscript: J. W. Harvey, R. A. Helliwell and M. A. Pomerantz commented on the upper atmosphere; D. H. Harwood, J. H. Mercer, M. L. Prentice and P. N. Webb commented on Cenozoic paleoclimatic history; and W. S. Benninghoff, J. C. Halfpenny, V. Komárková, K. A. Salzberg, and C. W. Sullivan commented on biological adaptations.

Gas-Phase Polymerization: Ultraslow Chemistry

HOWARD REISS

The mechanism of formation of polymer molecules in the gas phase is difficult to study because the involatile polymers tend to condense out of that phase. However, new techniques, involving the use of cloud chambers, have enabled workers to use the nucleation of liquid drops in supersaturated monomer vapors to detect single polymer molecules and therefore to work with so few simultaneously growing polymers that aggregation and condensation are avoided. Chain polymerization in which the chain carriers are either radicals or ions can therefore be studied in the vapor. Furthermore, the ability to work with such small concentrations of growing polymeric radicals, for example, makes it possible to avoid encounters between them that lead to recombination and the formation of "dead" polymers that are incapable of further growth. Many aspects of gas-phase polymerization can be studied including, besides radical and ion chains, ring-opening polymerization, initiation, radiation-induced polymerization, and especially "ultraslow" chemistry.

CHAIN POLYMERIZATION IN THE CONDENSED PHASE (usually liquid) has been studied extensively (*1*, chapters 3 through 6) and is of major importance to the polymer industry. The steps involved in a typical process are shown for the production of polystyrene from styrene ($\phi CH{=}CH_2$, where ϕ represents a phenyl group) via a free radical mechanism:

$$I \xrightarrow{k_d} 2R\cdot \qquad \text{(thermal or photoinitiation)}$$

$$R\cdot + \phi CH{=}CH_2 \xrightarrow{k_p} RCH_2\text{–}\underset{\phi}{\dot{C}H}$$

$$RCH_2\underset{\phi}{\dot{C}H} + \phi CH{=}CH_2 \xrightarrow{k_p}$$

$$RCH_2\text{–}\underset{\phi}{CH}\text{–}CH_2\text{–}\underset{\phi}{\dot{C}H} \qquad \text{(propagation)}$$

and so forth

$$R(CH_2\underset{\phi}{CH})_n\text{–}CH_2\underset{\phi}{\dot{C}H} + R(CH_2\underset{\phi}{CH})_m\text{–}CH_2\underset{\phi}{\dot{C}H}$$

$$\xrightarrow{k_t} R(CH_2\underset{\phi}{CH})_{n+1}\text{–}(\underset{\phi}{CH}CH_2)_{m+1}R \qquad \text{(termination)}$$

Scheme 1

where k_d, k_p, and k_t are rate constants for dissociation, propagation, and termination, respectively; I is an initiator that can decompose thermally or photolytically; and R· is a free radical. Other elementary steps are possible [such as chain transfer (*1*)], in which both a "dead" polymer and a monomeric radical that can still propagate are formed. A separate initiator may not be necessary with photochemical initiation. Termination not only leads to smaller polymers but also complicates both theory and measurement. In liquids it can be forestalled by the use of emulsion polymerization (*1*), and recently by the use of plasma-induced polymerization (*2*). Propagation that involves ion chains is also possible.

The author is in the Department of Chemistry and Biochemistry, University of California, Los Angeles, CA 90024.