

PALEOCLIMATOLOGY

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Ice Cores

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Introduction

Ice cores drilled in glaciers and ice sheets are used to investigate atmospheric chemistry, climate and glacier dynamics in the past. Precipitation that falls as snow in the polar regions, and at high altitudes in the tropical and temperate regions, if it survives summer warmth, is eventually transformed to glacier ice. Aerosols and soluble gases that are scavenged from the atmosphere by precipitation (wet deposition), or accumulate on the snow surface as dry deposition, are preserved within the ice. Atmospheric gases are also preserved in the form of air bubbles trapped within the ice. Ice cores are retrieved by means of a mechanical or thermal drill that cuts an annulus around a central, vertical core that is typically 10–20 cm in diameter and a few tens of meters to several thousand meters in length. If the relationship between length along the core and time in the past is known, time-series of atmospheric composition can be derived. The best-known example is the record of carbon dioxide from Antarctic ice cores, which documents significant changes in concentration of this important greenhouse gas over the last several hundred thousand years of Earth history. Time-series have also been developed from ice cores for many other trace gases and aerosol species, and, indirectly, of atmospheric temperature, atmospheric dynamics, solar variability, and marine and terrestrial biogeochemical cycles. Because most of this information is unavailable from any other source, ice cores play a central role in our understanding of paleoclimate. Data from ice cores also serve as a baseline of natural variability against which contemporary changes in the atmosphere may be evaluated.

Glacier Ice and Ice Core Dating

Ice core studies are conducted at perennially cold sites where little or no melting occurs. Examples include Antarctica, the interior portions of Greenland, the small ice caps of Arctic Canada, and the high mountains of the Andes and the Himalaya. Under such conditions, there is net snow accumulation in most years, and a combination of mass loading and metamorphism leads to compaction of the snow and eventual transition to glacier ice. Buildup of ice masses reaches from hundreds of meters thick in alpine settings to several kilometers in the great polar ice sheets of Antarctica and Greenland. This buildup is balanced by the slow deformation of glacier ice under pressure. Flow by internal deformation and sliding along the bed brings ice into warmer, lower-elevation environments where it is lost to melting, sublimation, or iceberg calving. Because of compaction and flow, snow layers of original thickness λ at the surface become thinner with depth. The amount of time t for a layer to reach depth z cannot generally be determined directly, but can be estimated with a geophysical ice flow model, used to calculate a thinning function $\psi = \lambda(z, 0)/\lambda(0, -t)$, if the net snow accumulation rate $\dot{b}(-t)$ is known. Ice core drilling sites are usually chosen near an ice divide (analogous to a drainage divide in hydrology), where the dominant strain direction is vertical. This greatly simplifies the calculation of ψ , which in this case will be approximately exponential. Under favorable circumstances, age and layer thickness can be determined by direct visual observation, using the alternating layering of coarse-grained, dusty, and wind-packed summer snow and relatively uniform winter snow. Counting of visible annual layers and seasonal cycles in chemistry, coupled with the identification of volcanic eruptions of known age (using ash chemistry and sulfate concentration anomalies preserved in the ice), allowed researchers to determine the age–depth relationship in cores from Summit, Greenland, to a precision better than 2% for the last 10 000 years (10 ky) and 5% for the last 50 ky. Precise determination

of layer thickness and age also permitted the calculation of an accurate time series of accumulation (Figure 1).

Typical snow densities at the surface of an ice sheet are $\sim 300 \text{ kg m}^{-3}$, increasing to $\sim 900 \text{ kg m}^{-3}$ a few tens of meters below the surface. The zone of intermediate density, referred to as the firn, is a permeable layer through which atmospheric gases can move. As air channels in the firn become closed off, gases are isolated from the overlying atmosphere and are eventually trapped as air bubbles. Bubble close-off occurs at a density near 820 kg m^{-3} , which generally corresponds to a depth of 60–100 m. At a given depth, the age of the gas trapped within the ice is always younger than the ice itself because of advective and diffusive transport within the firn. The amount of time (eqn [1]) corresponding to the bubble-close-off depth (COD) is a nonlinear function of snow accumulation rate and temperature.

$$\Delta \text{age}(z) = t_{\text{ice}}(z) - t_{\text{gas}}(z) = t_{\text{ice}}(\text{COD}) \quad [1]$$

At high-accumulation sites like central Greenland ($\dot{b} > 200 \text{ kg m}^{-2} \text{ y}^{-1}$), Δage can be a few hundred years or less. At Vostok Station in central East Antarctica ($\dot{b} < 30 \text{ kg m}^{-2} \text{ y}^{-1}$), Δage is several thousand years. Several Antarctic ice core records have been dated indirectly by correlation with the layer-counted Greenland records, using variations in globally well-mixed trace gases as time markers (Figure 2). The major source of error in this cross-dating technique is uncertainty in the value of Δage .

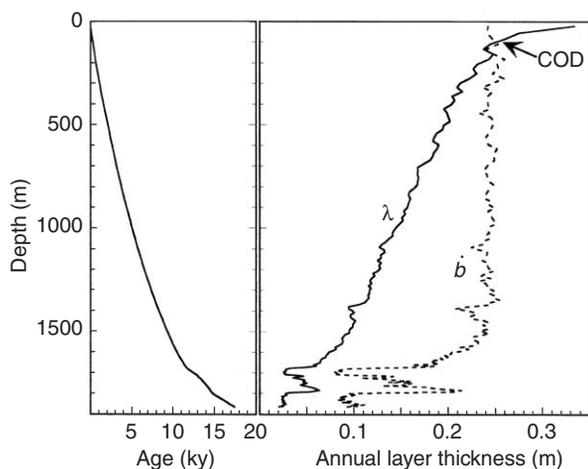


Figure 1 Depth versus age, measured annual layer thickness (λ) and calculated ice-equivalent annual accumulation (\dot{b}) in the GISP2 (Greenland Ice Sheet Project 2) ice core from Summit, central Greenland. The air bubble close-off depth (COD) is approximately at the intersection of \dot{b} and λ .

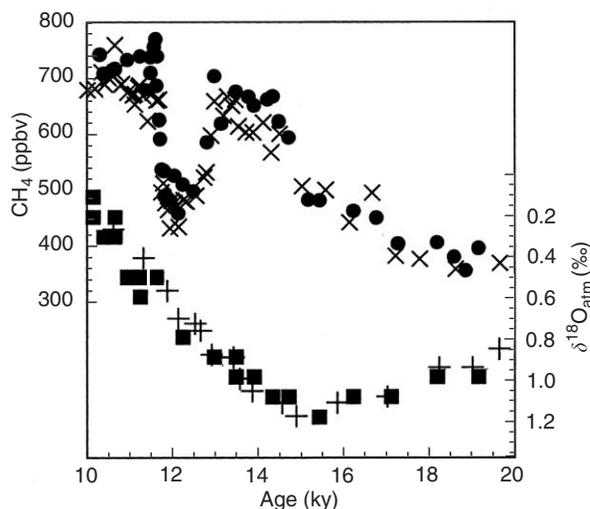


Figure 2 Concentrations of methane (CH_4) and isotopic ratio of molecular oxygen ($\delta^{18}\text{O}$ of O_2) from ice cores at Summit (GISP2), Greenland (\bullet , \blacksquare) and Taylor Dome, Antarctica (\times , $+$) between 10 ky and 20 ky ago. The oscillation in CH_4 between 13 ky and 11.5 ky is coincident with the Younger Dryas cold oscillation observed in paleoclimate records from around the North Atlantic region. The overall decrease in $\delta^{18}\text{O}$ of atmospheric O_2 (plotted on a reversed scale, by convention) reflects the increase in sea level from the influx of isotopically light (low $\delta^{18}\text{O}$) water from melting Northern Hemisphere ice sheets. (Adapted with permission from E J Steig *et al.* (1998) *Science* 282: 92–95.)

Stable Isotopes and Temperature

Time-series of past temperature change are obtained from ice cores using the ratios of stable isotopes of water. The major features of the global distribution of water isotopes can be represented by a simple Rayleigh distillation approximation of the hydrological cycle that works particularly well at high latitudes. Seasonal variations in temperature (T) are reflected in the $^{18}\text{O}/^{16}\text{O}$ and D/H ratios of atmospheric water vapor and precipitation according to nearly linear relationships (eqns [2] and [3]).

$$\frac{d(\delta^{18}\text{O})}{dT} = \alpha \quad [2]$$

$$\frac{d(\delta\text{D})}{d(\delta^{18}\text{O})} = 8 \quad [3]$$

The average value of α is about 0.7‰ K^{-1} . (‰ = parts per thousand; δ refers to the deviation in ‰ from the $^{18}\text{O}/^{16}\text{O}$ or D/H ratio of a standard, usually Standard Mean Ocean Water). Where temperatures are warm or snow accumulation rates are low, diffusion of water molecules within the firn layer will erase seasonal cycles after a few years to decades. Annual and interannual variations, however, are generally well

preserved and provide a means of extending instrumental temperature records back in time. Regional or global climate events such as the Little Ice Age between *c.* AD 1400–1900, and the transition from the cold temperatures of the last glacial period, about 20 ky ago, to the current, Holocene interglacial period are clearly seen in the isotopic time-series (Figure 3). Also evident in Arctic ice cores are very rapid isotopic changes, known as Dansgaard–Oeschger (D-O) events after their discoverers. The calculation of temperature from isotope ratios is based on the assumption that the slope, α , of the δ/T relationship is constant. Calibration of the isotope paleothermometer has been accomplished using spatial gradients on instrumental time-series, and with paleotemperatures calculated from measured temperatures in liquid-filled boreholes, using geophysical inverse methods. The

calibration results show that α varies by about a factor of 2 over glacial–interglacial time scales because of changes in the seasonality of precipitation, moist air mass transport trajectories, and ocean surface conditions. Where such changes can be quantified, $\delta^{18}\text{O}$ and δD may be considered reliable quantitative proxies for past temperature change. Ocean surface conditions, specifically, can be investigated using variations in ‘deuterium excess’, defined in

$$d = \delta\text{D} - 8 \times \delta^{18}\text{O} \quad [4]$$

Deviations of d from the average meteoric water value of 10‰ reflect the difference in sensitivity to phase changes of δD versus $\delta^{18}\text{O}$. Ice core time series of d are used to identify the ocean source areas for polar and high-altitude water vapor, and to quantify changes in evaporation temperature and humidity at those source areas.

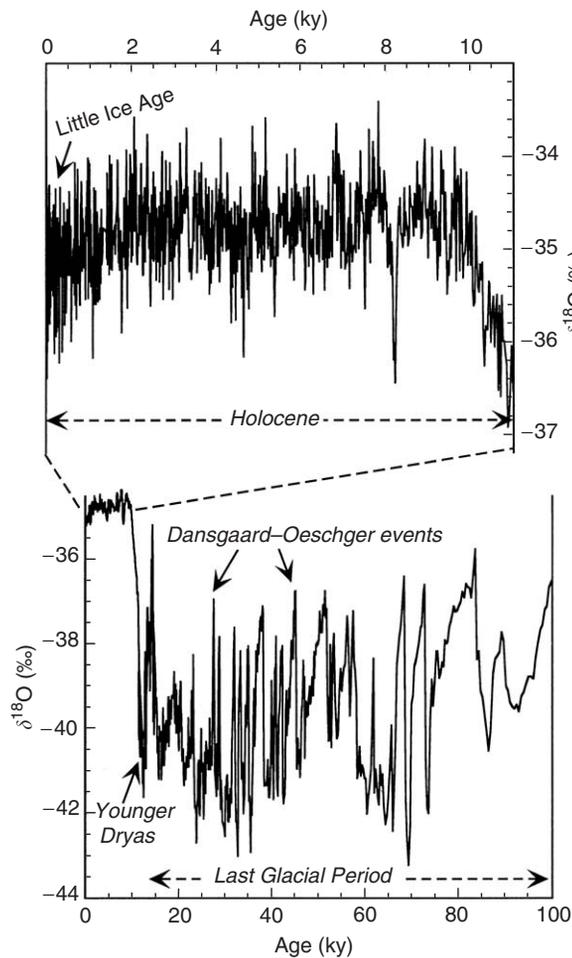


Figure 3 Oxygen isotope ratios of water ($\delta^{18}\text{O}$) from the GISP2 ice core, Greenland, from 100 ky ago to the present. The transition from the highly variable glacial period to the relative stable Holocene interglacial occurs at ~ 11.5 ky. The most recent 11 ky are shown on an expanded scale.

Trace Gases

Ice cores provide an essentially direct record of past concentrations of atmospheric gases. Depending on the magnitude of Δage , each gas sample analyzed typically represents an average of a few hundred years. Measurement of CO_2 from Greenland ice cores has been problematic because of the relatively warm temperature and high carbonate content of the ice. That CO_2 records from Antarctic ice cores are reliable is demonstrated by the excellent agreement with direct atmospheric measurements during the period of overlap (Figure 4). Over long time scales CO_2 is linearly related (correlation coefficient, $r^2 > 0.7$) to stable-isotope records of air temperature (Figure 5). Natural

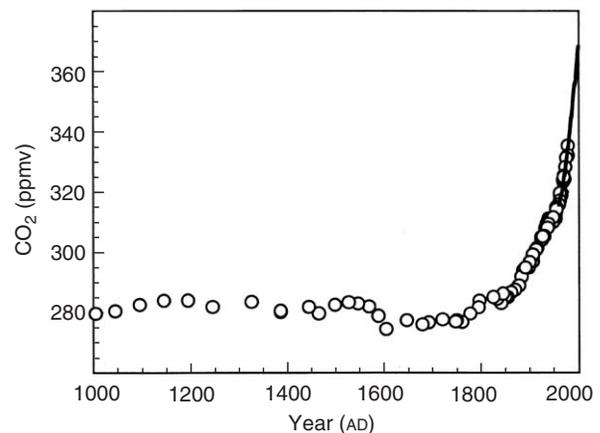


Figure 4 Carbon dioxide (CO_2) concentrations (\circ) from air bubbles in an ice core from Law Dome, Antarctica, for the last 1000 years compared with the direct measurements of atmospheric CO_2 (—) at Mauna Loa, Hawaii since AD 1959.

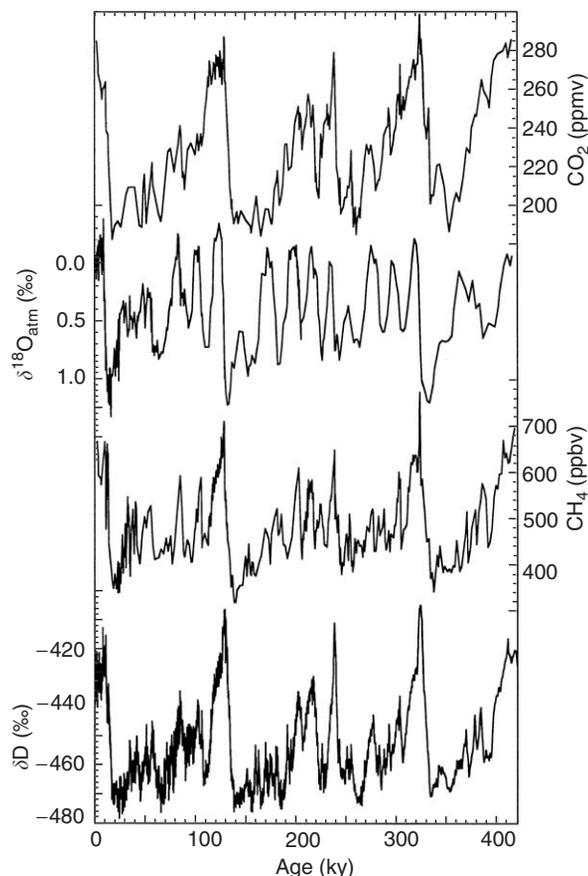


Figure 5 Concentrations of CO_2 , $\delta^{18}\text{O}$ of O_2 , and methane (CH_4) in air bubbles and hydrogen isotopes ratios (δD) in ice from the Vostok ice core, central East Antarctica, for the last 420 ky.

variations over the last 420 ky remained within 180–300 ppmv, well below the early twenty-first century value of >370 ppmv. Analysis of data from multiple ice and ocean sediment cores indicates a lead of temperature over CO_2 of perhaps a few hundred years for glacial–interglacial time scales. However, the phase relationship appears to be nonstationary and its average value remains poorly known because of large uncertainties in Δage .

Concentrations of other atmospheric trace gases such as methane (CH_4) and nitrous oxide (N_2O) show long-term variations similar to those of CO_2 . Comparisons of records from ice cores in Antarctica and Greenland allows determination of changes in the inter-polar gradients of these gases, which reflects different biogenic production rates in the Northern versus Southern Hemispheres. Methane additionally shows rapid variations that correspond to the fast warming and cooling of the D-O events observed in $\delta^{18}\text{O}$ records. Measurements of $^{15}\text{N}/^{14}\text{N}$, $^{40}\text{Ar}/^{36}\text{Ar}$, and $^{84}\text{Kr}/^{36}\text{Ar}$ ratios reveal the preservation of ther-

mally driven isotope ratio anomalies that permit independent calculation of the magnitude of temperature change and the value of Δage during these events. The results indicate that some of the D-O warmings were extremely rapid ($>1^\circ\text{C}/\text{y}$) and of even larger magnitude ($>15\text{ K}$) than would be predicted from $\alpha \sim 0.7\text{‰ K}^{-1}$, and show that temperature leads CH_4 by no more than a few decades.

Another important set of data obtained from ice core air bubbles is the $^{18}\text{O}/^{16}\text{O}$ ratio of atmospheric molecular oxygen (O_2), reported as $\delta^{18}\text{O}_{\text{atm}}$ to distinguish it from $\delta^{18}\text{O}$ of ice. The primary control on $\delta^{18}\text{O}_{\text{atm}}$ is the isotopic composition of average surface ocean water ($\delta^{18}\text{O}_{\text{sw}}$), which in turn is determined to first order by global ice volume – the amount of water stored on land in the form of ice sheets and glaciers. Time-series of $\delta^{18}\text{O}_{\text{atm}}$ from ice cores are used to tie ice core chronologies to each other and to marine sediment records of $\delta^{18}\text{O}_{\text{sw}}$. The connection between $\delta^{18}\text{O}_{\text{atm}}$ and ocean water $\delta^{18}\text{O}_{\text{sw}}$ is indirect, via photosynthesis and respiration by phytoplankton; there is a consequent variability in the lag between $\delta^{18}\text{O}_{\text{sw}}$ and $\delta^{18}\text{O}_{\text{atm}}$, known as the ‘Dole effect’, of up to ~ 2000 years.

Aerosols and Soluble Gas-Phase Species

The major ions present in most ice cores are Na^+ , H^+ , Ca^{2+} , Mg^{2+} , Cl^- , NO_3^- and SO_4^{2-} . Typical concentrations are at the ppb (10^{-9}) level. In coastal areas of the polar ice sheets, Na^+ , Cl^- , Mg^{2+} , Ca^{2+} , and SO_4^{2-} represent $>80\%$ of the ionic budget. Nitrate and non-sea-salt sulfate dominate the budget at more inland sites. All of these species show major variations that parallel those in stable isotopes and trace gas concentrations (Figure 6). Determining whether observed changes in concentration reflect changes in wet and dry deposition processes or real changes in atmospheric mixing ratios can be problematic. For example, a large fraction of the glacial–interglacial change in SO_4^{2-} concentrations probably reflects dilution by the twofold increase in snow accumulation rate at the beginning of the Holocene. The set of transfer functions describing air-to-snow relationships for all chemical species can be expressed as a formal mathematical inverse problem. Because of the high level of covariance in most ice core time-series, this problem must currently be considered under-terminated for most species. Nevertheless, significant changes in atmospheric mixing ratios have clearly occurred and are particularly large for terrestrially derived ions such as Ca^{2+} .

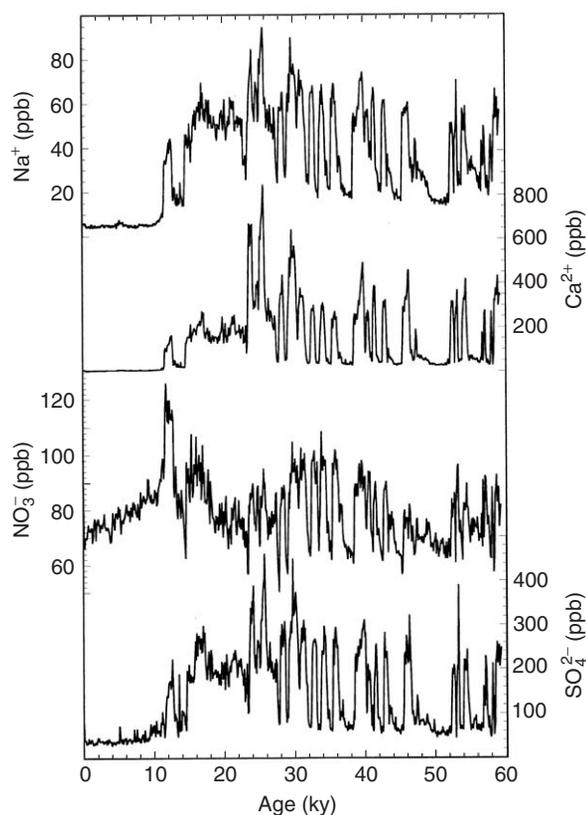


Figure 6 Concentrations of sodium, calcium, nitrate, and sulfate ions in the GISP2 ice core, Greenland, from 60ky ago to the present.

Increases in terrestrial species reflect increases in the overall dustiness of the atmosphere due to enhanced continental aridity and aeolian transport during cold climate periods. This effect is amplified by increased wind speeds and the consequent greater efficiency of transport through the troposphere to the surfaces of glaciers and ice sheets. Specific source regions for continental dust have been identified through the use of isotopic and rare-earth element analysis of mineral grains preserved in the ice. Increased concentrations of marine-derived species in ice cores reflect increases both in transport efficiency and in biogenic production rates. More than 50% of SO_4^{2-} in inland Antarctic snow is derived from the oxidation of dimethyl sulfide, an algal waste product. The ratio of sulfate to methanesulfonic acid, an intermediate oxidation product that is preserved in the ice, has been used to identify biogenic versus nonbiogenic sources of SO_4^{2-} . Changes in the biogenic contribution of SO_4^{2-} are of particular interest because of the role that sulfate aerosols play in climate forcing via modification of cloud radiative properties. Emissions from volcanoes are an important source of SO_4^{2-} concentration

anomalies in ice cores; comparison with the stable-isotope record of temperature provides strong independent evidence for the cooling effect of sulfate aerosols. In central Greenland cores, time-series analysis shows that geochemical variations can be resolved into two nearly orthogonal components that are dominated by marine and terrestrial species, respectively. The ratio of marine to terrestrial components is believed to reflect variations in the relative strength of meridional versus zonal tropospheric flow, possibly related to changes in the strength of the polar vortex.

Species such as hydrogen peroxide (H_2O_2), hydrochloric acid (HCl), and nitric acid (HNO_3), which have an appreciable vapor pressure at ambient surface temperatures, additionally complicate the interpretation of ice core geochemical records. While deposition rates are controlled by kinetic processes in clouds and at the snow surface, loss rates following deposition are dominated by the tendency toward equilibrium with the overlying atmosphere. Final concentrations archived in the ice core are determined by the extent to which equilibrium is reached. The approach to equilibrium depends on a combination of snow accumulation rate, temperature and, for photoreactive species, irradiance levels. Ambient oxidative conditions determined by hydroxyl radical ($\dot{\text{O}}\text{H}$) and ozone (O_3) mixing ratios also play a role. The measurement of multiple reactive and nonreactive species and their isotopic composition in ice cores offers a means for obtaining quantitative information on the oxidizing capacity of the atmosphere in the past.

Cosmogenic Radionuclides and Geomagnetic and Solar Variability

Rare isotopes produced in the atmosphere by cosmic rays, including ^{10}Be , ^{26}Al , ^{36}Cl , and ^{14}C , are all present in ice cores in measurable concentrations. These cosmogenic radionuclides (CRNs) are of interest as a means to extend the record of geomagnetic and solar variability beyond the instrumental and historical time periods and as a source of information on snow accumulation and atmospheric dynamics. Solar modulation of CRN production is reflected in variations in concentration that have dominant periodicities at ~ 11 years and ~ 90 years, and that closely track the record of atmospheric ^{14}C variations, independently derived from tree-ring studies (Figure 7). Over longer time scales, geomagnetic modulation may also be important in affecting CRN deposition rates. Differences among records from Arctic, tropical, and Antarctic ice cores can in principle be used to separate solar and geomagnetic modulation of CRN production, since the latter is unimportant at high latitudes.

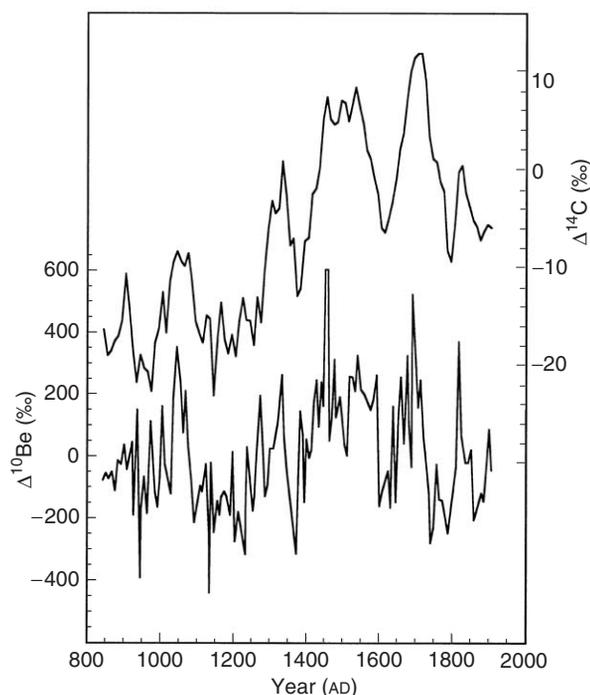


Figure 7 Anomalies of ^{10}Be in an ice core from the South Pole compared with atmospheric ^{14}C anomalies from analyses of tree rings between AD 800 and 1950. The Δ notation refers to relative deviation in parts per thousand. (Adapted with permission from E. Bard *et al.*, *Earth and Planetary Science Letters* 150: 453–462.)

In practice, the difficulty of separating production variations from meteorologically influenced changes in deposition rate has prevented an unequivocal geomagnetic record from being obtained from CRN measurements. An exception is the large excursion observed in ^{10}Be and ^{36}Cl in several ice cores, which coincides with the Laschamp geomagnetic excursion 40 ky ago, when the field strength may have approached zero. Most other large changes in CRN concentrations reflect changes in snow accumulation rate. Assuming that production rates are approximately constant over time scales greater than those expected from solar variability (i.e. >1000 years), ^{10}Be can be used as an independent variable in separating changes in atmospheric loading of other aerosol species from changes in precipitation scavenging and dry deposition efficiency. The assumption of constant production is reasonable in Antarctica, where limited meridional air-mass exchange prevents a significant amount of geomagnetically modulated, low-latitude-produced ^{10}Be from reaching the ice sheet surface. Because significant CRN production occurs in the stratosphere (about two-thirds of the total production), CRN can also provide information on past stratosphere/troposphere exchange processes.

This has been accomplished qualitatively, using the magnitude of solar-modulated ^{10}Be variations, and the ratio of ^{36}Cl to ^{10}Be , which have different tropospheric lifetimes, as tracers of air mass origin. Radiocarbon (^{14}C) is a special case of CRN in ice cores because the predominant fraction is produced *in situ* by neutron activation of ^{14}N . Concentrations of ^{14}CO and $^{14}\text{CO}_2$ in air bubbles provide a determination of snow accumulation rate that agrees well with independent evidence.

The Causes of Ice Ages and Rapid Climate Change Events

Geochemical time-series from ice cores have had a profound influence on our understanding of climate change. Particularly important is the opportunity provided by ice cores to evaluate the characteristics of paleoclimate time-series in the frequency domain. The high-precision, independent dating achieved with ice cores has allowed a more rigorous test of Milankovich theory than was previously possible with ocean sediment cores, and confirms that climate change exhibits significant power at orbital frequencies. The observation from the Vostok ice core that CO_2 varies approximately linearly with temperature addresses a critical problem for Milankovich theory, which is that the amplitude of solar insolation variations resulting from Earth's orbital changes is too small to alone account for the large magnitude of climate changes. At least 50% of the required amplification of small insolation changes over glacial–interglacial cycles can be attributed to observed CO_2 variations alone. The $\delta^{18}\text{O}_{\text{atm}}$ record shows a strong coherence with precession (~ 23 ky period) supporting the proposed linear relationship between orbital forcing and global ice-sheet volume.

Ice core data also demonstrate that large-magnitude climate changes have occurred in the past that are not related to Milankovich forcing. Although there is no evidence for rapid CO_2 changes in the past (variations originally identified in the Greenland Dye 3 ice core are now known to be artifacts), the D-O events clearly reflect real, rapid changes in atmospheric composition and dynamics. The D-O events are evident not only in $\delta^{18}\text{O}$ and CH_4 in Greenland ice cores, but also in dust and aerosol concentrations, and have been identified in ocean sediment cores from the North Atlantic region, from chemical tracers of deep water circulation, and from sea surface temperature. These events recur throughout the last glacial period with a nominal frequency of $1/1500\text{ y}^{-1}$. There is considerable variation in the length of this cycle: in most data sets the spectral power of D-O events rises significantly

above red noise only for bandwidths of several hundred years. On average, every fourth D-O event is associated with a coarse-grained 'Heinrich' layer in North Atlantic sediment cores. Heinrich layers reflect the transport of debris by icebergs from the margins of the ice sheets that covered much of Europe and North America during the last glaciation. Many paleoclimatologists believe that comparison of paleoclimate records from both hemispheres is a key to understanding the D-O events and their association with ocean thermohaline circulation and the dynamics of large ice sheets. Quasi-periodic climate changes with comparable timing have been identified in South American and Antarctic ice cores and Southern Ocean sediment cores, but these are generally smaller in amplitude and, with a few exceptions, do not show the rapid warmings characteristic of D-O events.

Natural Climate Variability and the Anthropogenic Impact on the Atmosphere

Ice core records provide an important baseline against which anthropogenic changes to the atmosphere can be measured. The high temporal resolution obtained with ice core geochemical and isotopic measurements has been used to extend records of the North Atlantic Oscillation and other important climate indices beyond the short period available from instrumental records. Arrays of ice cores covering the last 200–2000 years at annual resolution have been used to document the spatial patterns of interannual variability of both Arctic and Antarctic climate, for which instrumental data are particularly limited. On longer time scales, stable-isotope records show that temperatures during the twentieth century were slightly cooler, on average, than during the peak warm intervals of both the last interglacial period (~125 ky ago) and the Holocene. A mid-Holocene warm period is especially pronounced in ice cores from northern Greenland and the Canadian Arctic. The ice cores provide little evidence, however, for temperatures as high as the last two decades of the twentieth century having occurred at any time in the past 100 000 years. Similarly, ice core records show that the background of natural variability in atmospheric chemistry is large. However, increases in most chemical species in the last 100–200 years appear to be unprecedented in magnitude and rate of change. Particularly clear signals of the anthropogenic impact on the chemistry of the troposphere include the following examples. Ice core measurements show that, relative to the AD 1900 value, concentrations of CH₄ have doubled, CO₂ has

increased by 25%, and N₂O by 10%. In Northern Hemisphere ice cores, SO₄²⁻ concentrations increased by more than a factor of 4 between 1900 and 1970. This is in good agreement with the documented increase in SO₂ emissions from industrialized nations. Northern Hemisphere ice cores also document an overall decrease in SO₄²⁻ since 1980, probably reflecting abatement measures in the United States and Europe. Nitric acid deposition has similarly increased through most of the twentieth century and continues to increase. About 50% of the current NO₃⁻ deposited to Arctic snow can be attributed directly to anthropogenic NO_x emissions. Concentrations of heavy metals in Northern Hemisphere ice cores have increased since the late eighteenth century in parallel with growth in use since the industrial revolution. Isotopic studies have been used to identify source regions for these contaminants and show that United States emissions contributed about two-thirds of the deposition of lead from the atmosphere over Greenland until the late 1970s when the use of leaded gasoline declined. Finally, analysis of ice cores from several Arctic and temperate alpine locations shows that deposition rates of organochlorine pesticides remain at levels comparable to those of the 1970s when production and use was much higher than at present. The continued high concentration of these compounds in Arctic snowfall likely reflects their long residence times in soils.

See also

Anticyclones. Arctic Climate. Carbon Dioxide. Climate Variability: North Atlantic and Arctic Oscillation. **Ice Ages (Milankovitch Theory). Land-Atmosphere Interactions:** Trace Gas Exchange. **Paleoclimatology:** Varves. **Sea Ice. Snow (Surface).**

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Varves

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Introduction

Varves are laminated sediments formed annually in aquatic environments by seasonal climatic changes. Each varve is made of two thin, alternating light and dark sedimentary laminae representing one year of deposition (**Figure 1**). The coupling of varve sedimentation with seasonality makes varved sequences useful as high-resolution archives of climate. The geologist de Geer first defined varves in the late nineteenth century from glacial sediments in Sweden. He and his students surveyed Swedish glacial lakes and found that sediment thickness patterns correlated between adjacent lakes. From the regularity, continuity, and broad distribution of the laminated sediments, he concluded that their deposition was driven by a widespread, regular, and strong forcing mechanism, i.e. annual climate.

Varve Formation

Varves form in depositional basins characterized by seasonal variation in sediment composition and by low-oxygen bottom waters. Changes in sediment flux result in laminae of alternating composition, while low-oxygen waters prevent burrowing animals from mixing the sediment and obliterating the laminated structure. The two necessary conditions required for the formation of varves are found in both fresh water and salt water environments. In glacial lakes, for example, meltwater carries fine sand, clay, and silt into the lake along the glacier margin during spring and early summer (**Figure 2**). Finer particles are kept suspended in the water column during summer by wind-driven currents, whereas coarser, heavier sedi-

ment falls to the lake floor. As ice melts, nutrients are released into the lake, resulting in diatom blooms during the productive spring season. The coarser particles and the diatom frustules form a distinct light layer in spring and early summer. During summer,

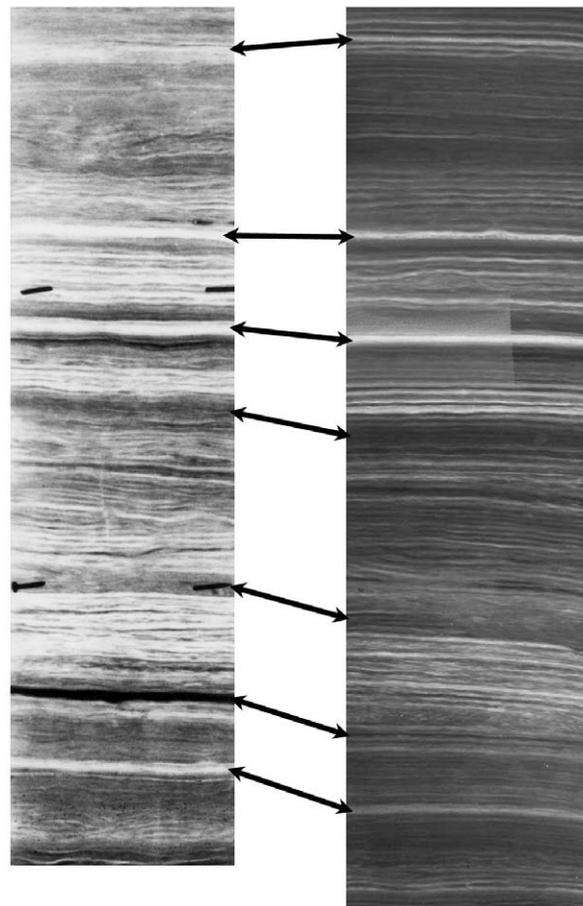


Figure 1 Contact prints of X-radiographs from two sediment cores taken from the bottom of Santa Barbara Basin, off southern California. Even though the cores were taken from different locations within the basin, it is possible to cross-match varve features (arrows).