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Cross-references

Bolide Impacts and Climate
Cretaceous Warm Climates
Sea Level Change, Last 250 Million Years

CRYOSPHERE

Introduction

The cryosphere (derived from the Greek *kryos* for ''cold'') comprises all of the frozen water and soil on the surface of the Earth. While this term is often used by climatologists, a more accurate name for the study of all aspects of ice and snow is *glaciology*. Ice in clouds is also an important element of climate but is not considered part of the cryosphere. Because the freezing point of water lies in the center of the range of terrestrially achieved temperatures, the cryosphere is a very changeable feature of the Earth's surface. Its six main elements, in diminishing order of area for today's climate, are: (a) seasonal snow cover, (b) sea ice, (c) permafrost, (d) ice sheets, (e) river and lake ice, and (f) mountain glaciers and small ice caps.

Seasonal snow cover is the most dynamic element of the cryosphere. It responds rapidly to atmospheric conditions on timescales of days to weeks and, in fact, exhibits the greatest seasonal variation of any geophysical element on the Earth's surface. Although it has a small seasonal heat storage capacity, its main effect on climate comes from its high albedo (fraction of incident solar radiation reflected back to space). Fresh snow has an albedo of 0.9. *Sea ice* forms when sea water is cooled below its freezing point. This freezing is a complex process because the freezing point of water depends on the salt concentration of the ocean. A sea ice field usually consists of numerous individual floes, typically 0.5–4 m thick, with diameters between 1 meter and many kilometers. Sea ice has a high albedo and therefore the effect its seasonal cycle exerts on the surface heat balance of the Earth is similar to, though smaller than, that exerted by seasonal snow cover. Sea ice also acts as a barrier to the exchange of heat, moisture, and momentum between the atmosphere and the ocean, and furthermore plays a role in the formation of deep water masses in some oceanic areas. Sea ice is covered by snow for much of its lifetime. *River and lake ice* form in a similar manner to sea ice but without the complications of high salt concentrations. Their climatic behavior resembles that of sea ice. *Ice sheets* are quasi-permanent features, responding to environmental changes on time scales of millennia. Today, only the two polar ice sheets on Antarctica and Greenland remain. Together, they contain by far the greatest proportion of mass of the cryosphere but are least sensitive to climate change on the shorter timescales. Ice sheets have a high elevation and high albedo, and so act as elevated cooling surfaces for atmospheric heat balance. They also play an important role in the freshwater balance of the oceans. Changes in their volume, though slow, are the main cause for sea-level changes on geological timescales. *Mountain glaciers* and small ice caps are a small component of the cryosphere. They are distinguished from the large ice sheets by their size and by the fact that they are topographically controlled. They are too small to have a direct influence on climate; however, their high rates of mass turnover allow rapid changes in response to climatic change (timescales of the order of decades to centuries). *Permafrost* is ground material that remains below freezing for at least two consecutive years, whether or not it contains ice. It may be continuous, underlying the whole of a region, or discontinuous, occurring in patches. In many regions the uppermost, so-called active layer, thaws annually to a depth of 1 to a few meters. Permafrost is a product of heat exchange between the land surface and the atmosphere, changing on timescales of centuries. It affects surface ecosystems and river discharge, and changes in its extent can control rates of trace gas emissions, especially of methane. Permafrost records temperature changes, hence its role as geo-indicator for monitoring and assessing environmental change.

Current distribution of ice in the cryosphere

Today, nearly 80% of the Earth's freshwater is stored on the continents in the form of ice, almost all of it in the polar regions and in high mountain areas, cf. Table C9 and Figure C78. Perennial ice covers 10% of the Earth's land surface and 4% of the oceans. The seasonal distribution of snow, lake and river ice, and to a lesser extent, sea ice, is closely linked with atmospheric conditions, in particular with temperature.

Seasonal snow

Seasonal snow reaches its maximum extent in late winter, when it covers almost 50% of the land surface of the Northern Hemisphere. It is almost absent in the Southern Hemisphere, where it is limited to the mountainous areas of New Zealand, southeastern Australia, and South Africa; it is more widespread only in Patagonia and ice-free regions of Antarctica and on sub-Antarctic islands.

Sea ice

The seasonal variation of sea ice is much smaller in the Arctic than in the Antarctic owing to the different geography of the two polar regions. The Arctic Ocean is a truly polar ocean surrounded by land, with only a limited range of longitudes within which sea ice can expand seasonally into lower latitudes. By contrast, in the Southern Hemisphere the central polar region is covered by the continent of Antarctica, preventing sea ice from extending to the highest latitudes in summer but at the same time allowing winter sea ice to extend in a roughly concentric zone around Antarctica at almost all longitudes. A second contrast between the two hemispheres is the larger upward oceanic heat flux in the Antarctic, as less fresh water enters the ocean to cause a low-salinity layer of polar surface water

Table C9 Size of the present-day cryosphere and average mass turnover times of its main components as given by the ice flux and average residence time (adapted from Fitzharris et al., 1996; Vaughan et al., 1999; Zhang et al., 1999; Huybrechts et al., 2000; Church et al., 2001; Lythe et al., 2001)

Cryospheric element	Area (10^6 km^2)	Ice volume (10^6 km^3)	Ice flux (km^3 yr^{-1})	Residence time[a] (yr)
Seasonal snow cover	47.2	~0.01	~25,000	~0.4
Northern Hemisphere (max)	46.3			
Southern Hemisphere (max)	0.9			
Sea ice				
Northern Hemisphere (max/min)	16.0/9.0	0.05/0.03	~40,000	~1
Southern Hemisphere (max/min)	22.0/4.0	0.03/ <0.01	~20,000	~1
Permafrost	25.4	~0.02	~40	~500
Ice sheets				
Antarctic Ice Sheet (grounded ice)	12.37	25.71	2,100	~12,500
Antarctic ice shelves	1.49	0.66	540	~1,200
East Antarctic Ice Sheet	10.09	22.59	1,350	~17,000
West Antarctic Ice Sheet	2.28	3.12	750	~4,000
Greenland Ice Sheet	1.71	2.85	570	~5,000
River and lake ice	<1.0			~0.3
Mountain glaciers and small ice caps	0.68	0.18	~750	~250

[a]The residence time is defined as the ice volume divided by the ice flux.

hampering vertical heat exchange from below. For these reasons, the Arctic features a thicker permanent sea ice cover of multiyear ice, augmented in winter by a seasonal ice cover of first-year ice extending into the surrounding subpolar seas. Around Antarctica, however, almost all of the winter ice melts in summer, leaving multiyear ice confined only to the western Weddell Sea, the Bellingshausen and Amundsen Seas, and coastal embayments. Arctic sea ice generally reaches its maximum extent in March and minimum extent in September. By contrast, Antarctic autumnal ice growth is slower and spring decay about one month faster.

Polar ice sheets

Present-day continental ice cover is entirely dominated by the Antarctic Ice Sheet, which is responsible for 90% of the volume and 85% of the area of all land ice. Most of this ice is in the East Antarctic Ice Sheet, which constitutes a vast, relatively flat dome with a maximum elevation of 4,030 m and a maximum known thickness of 4,776 m. The Transantarctic Mountains divide the East Antarctic Ice Sheet from the West Antarctic Ice Sheet and the Antarctic Peninsula. The West Antarctic Ice Sheet rests on bedrock below sea level and largely drains into two big ice shelves in the Ross and Weddell Seas. These ice shelves are up to 1,000 m thick and float in equilibrium with the ocean water. Smaller ice shelves fringe the Antarctic coastline elsewhere. Ice flow towards the margin results mainly from internal ice deformation under the action of gravity. At the margin, the flow is often channeled in outlet glaciers and in ice streams, in which the dominant flow mechanism is basal sliding, and that are responsible for the bulk of the ice discharge across the grounding line into the ice shelves. Mass is lost from the ice shelves by bottom melting and by calving of icebergs from their ice fronts. Current Antarctic surface temperatures are so low that there is virtually no surface melting. These low temperatures also limit the amount of water vapor that can be advected inland. Accumulation rates over the vast interior of the continent are only a few cm per year and this makes the central Antarctic Plateau one of the driest places on Earth.

The only Arctic ice mass of significance is the Greenland Ice Sheet. It is a relict from the ice ages that overlies a bowl-shaped continent and has a maximum surface elevation of 3,230 m. In contrast to Antarctica, summer temperatures on Greenland are high enough to cause widespread summer melting, resulting in a negative mass balance below elevations of about 500 m in the north and 1,800 m in the south. Greenland has no major ice shelves apart from a few small ones along the north and northeast coast. Ice not lost by ablation is discharged into the ocean by iceberg calving from glaciers, in roughly the same amount as runoff. Complete melting of both polar ice sheets would raise global sea level by about 70 m.

Glaciers and ice caps

The water currently contained in glaciers and ice caps (excluding Antarctica and Greenland) is equivalent to about 0.5 m of global sea level. Half of it is in small ice caps, mostly in the Arctic, and the other half is stored in mountain glaciers elsewhere. Most mountain glacier ice is in Alaska, followed by Central Asia and Patagonia. Ice volume contained in the Alps is less than 1% of total glacier volume, but its glaciers have been studied the most. There are more than 160,000 glaciers worldwide. Glaciers enlarge when the accumulation of snow and ice exceed the loss by melting, and sometimes calving. Net accumulation occurs at higher altitude, net ablation at lower altitude. To compensate for net accumulation and ablation, the ice flows downhill by internal deformation and basal sliding. The primary controls on glacier mass balance are temperature, especially in summer, and accumulation, especially in winter. The nature of the response varies from glacier to glacier depending on mass-balance gradient, hypsometry, and glacier length.

Permafrost

About 24% of the exposed land area of the Northern Hemisphere is currently underlain by permafrost, most of it in the tundra of Canada, Alaska, and Russia (Zhang et al., 1999). It is also present under shallow polar sea beds, in ice-free areas in Antarctica, on some sub-Antarctic islands, and in many mountain areas and high plateaus of the world. Its distribution roughly follows the pattern of air temperature, though some of it is relict, having formed during colder glacial periods and having melted only very slowly since then. Sizeable areas below the polar ice sheets and small ice caps are also below freezing. This thermal condition of the ground extends to maximum depths of about 1,500 m in Siberia, 1,000 m in the Canadian Arctic, and 680 m in Alaska, but often it is also less than 20 m thick. Generally, permafrost with high ice content (>20% by volume) is found at high latitudes, and permafrost with low ice content (<10% by volume) mainly in mountainous regions and high plateaus. Almost half of the total permafrost area is continuous permafrost. The remaining part is made up of discontinuous and sporadic permafrost or occurs in isolated patches.

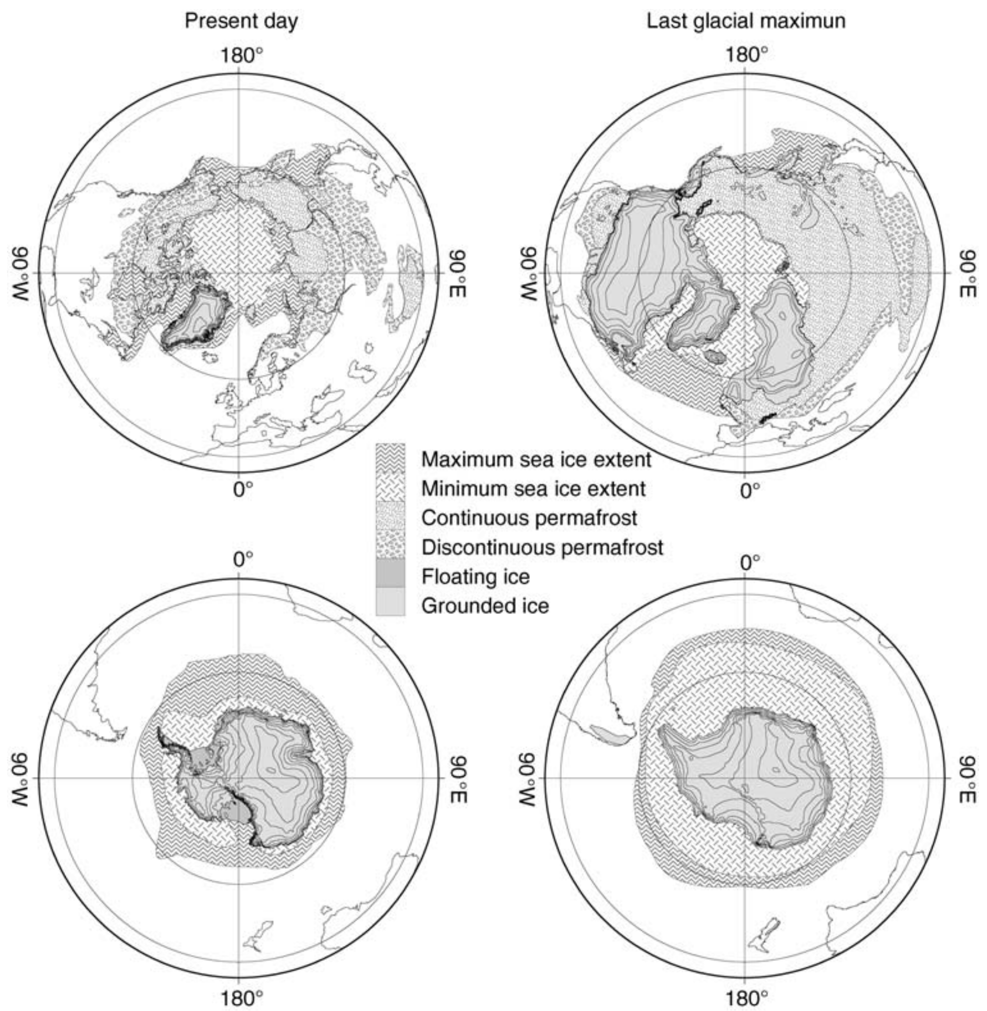

Figure C78 Geographic extent of main components of the cryosphere in both hemispheres. The *left panels* are for the present distribution, the *right panels* depict the situation during the Last Glacial Maximum between 23,000 and 19,000 years ago (adapted from maps presented in Cooke and Hays, 1982; Untersteiner, 1984; Frenzel et al., 1992; Fitzharris et al., 1996; Van Vliet-Lanoë and Lisitsyna, 2001, and the author's own ice sheet model results).

Current evolution of the cryosphere

The current evolution of many elements of the cryosphere, especially of those with the shorter response times, appears well correlated with the recent global warming trend of about 0.6 °C since the late nineteenth century.

Snow covers in temperate regions are usually thin and often close to their melting point, hence they are sensitive indicators of climate change. Monitoring of seasonal snow is practical only with satellites, so there are no reliable records prior to 1971. Visible-band and passive microwave data since then show considerable variability from year to year, but also reveal a decline in the extent of snow cover in the Northern Hemisphere by about 10% since the mid-1980s, largely due to a decrease in spring extent (Robinson et al., 1993). This reduction in snow cover is highly correlated with increasing air temperature. Snow-cover observations at ground stations have been recorded for many decades and suggest that Northern Hemisphere spring extents are currently at their lowest values in the past 100 years. Numerous small perennial snow patches have disappeared over the twentieth century, commensurate with the general warming trend.

Like seasonal snow, *sea ice* is also thought to be sensitive to climate change but the mechanisms are less well understood as additional processes such as ice transport, ocean circulation, and convection depths come into play. Warming is nevertheless thought to cause a reduction in both area and thickness of sea ice, which sets in motion a positive feedback, as more solar radiation is absorbed and more oceanic heat is transferred to the atmosphere. Satellite passive microwave data show a marked decrease of perennial Arctic sea ice extent for the period 1978–2000 of 9% per decade consistent with a summer warming of 1.2 °C per decade over the same period (Comiso, 2002). Antarctic records of sea-ice extent, on the other hand, show large variability but a negligible, albeit slightly positive, overall trend. Knowledge of sea-ice thickness largely comes from upward sonar profiling from submarines in the Arctic and more recently, from buoys moored to the sea floor. Observations of sea ice thickness have also been derived from satellite altimetry of the freeboard height, but geographical coverage and/or record length do not yet allow for strong conclusions. Summer Arctic sea ice cover and thickness have decreased in recent years (e.g., Lindsay and Zhang, 2005; Serreze et al., 2007).

Observations of *lake* and *river ice* exist primarily in the form of freeze-up and break-up dates. Records for the Northern Hemisphere since 1850 show significant evidence of later autumn freeze-up and earlier spring break-up of on average 5.8 days and 6.5 days per 100 years, respectively, consistent with the warming trend from the beginning of the twentieth century (Magnuson et al., 2000).

The present evolution of the *polar ice sheets* is still not known with confidence but a picture is starting to emerge from a combination of mass budget studies, satellite altimetry, and modeling (Rignot and Thomas, 2002). Current evidence indicates that the Antarctic Ice Sheet has probably lagged behind in the last glacial cycle of retreat. Most of the ongoing shrinking probably occurs in West Antarctica as a delayed response to post-glacial sea-level rise, whereas the East Antarctic Ice Sheet may be close to a stationary state, or growing slightly in response to increased accumulation rates. Superimposed on this long-term trend is a likely thickening effect of modern accumulation increases. Larger imbalances have recently been detected in West Antarctica, positive for the Siple Coast ice stream area and negative for the Pine Island/Thwaites catchment area. These probably represent local ice-dynamic effects that are unrelated to climate. The Greenland Ice Sheet is believed to be close to an overall neutral equilibrium. Current data show a consistent picture of a small thickening of the accumulation zone offset by larger thinning rates at lower altitudes. According to the Intergovernmental Panel on Climate Change (IPCC), both ice sheets together may have contributed between −2 cm and +6 cm to the observed twentieth century sea-level rise (Church et al., 2001).

A wealth of data exists on the geometry of valley *glaciers*. A large amount of information is available from sketches, etchings, paintings, and old photographs. In many cases, information from terminal moraines and trimlines is available to construct the history of a glacier over the last few centuries. An overwhelming characteristic of global glacier length records is the almost uniform recession of glaciers, which started about 1850 at the end of the Little Ice Age and is still continuing today. Rising global temperature is the most likely explanation and so the worldwide glacier retreat is probably one of the strongest measures of global warming. The only exceptions today are Norwegian and New Zealand glaciers, which are currently advancing, due to local increases in precipitation. However, this is an exceptional state that can probably not be sustained in the long term. Global glacier melt assessments indicate that since the late nineteenth century, glaciers and ice caps have lost between 5 and 10% of their volume and a little less than that in area. The IPCC reports a glacier contribution to global sea-level rise of between 2 and 4 cm over the last 100 years, less than the contribution from thermal expansion but similar to the combined contribution from the Antarctic and Greenland Ice Sheets. Findings of ancient remains in the European Alps (e.g., the 5,000-year old Oetztal "ice man") indicate that current glacier recession is reaching levels not seen for probably several millennia. Recent studies suggest that the twentieth century rate of glacier wastage in Alaska and the Patagonian Andes doubled after the mid-1990s (Arendt et al., 2002; Rignot et al., 2003).

Deep *permafrost* temperatures change by conduction and therefore are a manifestation of past climatic conditions over long spans of time. Permafrost is also very sensitive to changes in its surface energy balance. Very small changes in surface climate can produce robust changes in permafrost temperatures. Long-term measurements in deep boreholes in Alaska, Canada, and elsewhere demonstrate a distinct but spatially heterogeneous warming trend in low-land permafrost (Romanovsky and Osterkamp, 2001). Temperature measurements in northern Alaska have demonstrated a warming of the permafrost of 2–4 °C over the last century. In North America, the southern extent of the discontinuous permafrost zone has migrated northward in response to warming after the Little Ice Age, and continues to do so today. In China, both an increase in the lower altitudinal limit of mountain permafrost and a decrease in areal extent have been observed. Increases of the thickness of the active layer in response to climate warming have also been reported.

The cryosphere during the last glacial cycle

During the last ice age, which was most severe from about 23,000 to 19,000 years ago, an extensive ice complex stretched across North America, Greenland, the polar seas, and parts of northern Eurasia. It consisted of huge land-based glaciers, marine-based ice sheets, and either permanent pack-ice or shelf ice. Ice increased synchronously over Antarctica and the Southern Ocean. Sea levels were 125 m lower than today because water was bound up in continental ice sheets that increased their total mass by about a factor of three (Clark and Mix, 2002; Figure C78 and Table C10).

The surface of the Earth during the Last Glacial Maximum (LGM) was comprehensively mapped by the CLIMAP (Climate: Long-range Investigation, Mapping, and Prediction) project in the 1970s from a combination of biological and lithological evidence from deep-sea cores and from the land (CLIMAP, 1976). Best documented is the chronology of Northern Hemisphere ice-sheet extent. The largest glacier complex was situated over North America. It is inferred to have consisted of several domes centered over the high plateaus of eastern Canada and the Rocky Mountains. An ice-free corridor separated the Cordilleran Ice Sheet from the Laurentide Ice Sheet until late in the glacial

Table C10 Size of main cryospheric elements during the Last Glacial Maximum (adapted from Denton and Hughes, 1981; Cooke and Hays, 1982; Van Vliet-Lanoë and Lisitsyna, 2001; Huybrechts, 2002; Clark and Mix, 2002)

Cryospheric element	Area (10^6 km^2)	Ice volume (10^6 km^3)	Excess ice-equivalent sea level (m)
All land ice	41.8	75–88	110–135
Laurentide Ice Sheet	14.6	31–38	75–88
Fennoscandian Ice Sheet	6.2	5.8–7.5	14–18
Greenland Ice Sheet	2.6	4.0–4.5	2–3
Antarctic Ice Sheet (grounded)	15.1	33–36	15–20
Other ice caps and glaciers	3.3	1.4–2.2	4–6
Northern Hemisphere sea ice (max/min)	18/11	–	–
Southern Hemisphere sea ice (max/min)	35/25	–	–
Exposed permafrost (continuous and discontinuous)	30	–	–

period. The glaciation of Eurasia had an ice sheet centered over the British Isles in the west and that extended beyond the present day shoreline onto the then emerged shelf of the North Sea. The Fennoscandian Ice Sheet spread into the same area from the east but did not coalesce into the British Isles at the LGM. To the east, the Fennoscandian Ice Sheet extended into the northern part of the Russian plain and the Barents Sea, which in the northwest was alsocovered by ice from Novaya Zemlya. The Kara Sea, except at its western margin, was not glaciated at the LGM, but likely at an earlier stage of the last glacial cycle. The thickness of the ice sheets cannot be reconstructed from glacial-geological observations but it has been inferred to have been up to 4 km from indirect evidence such as uplift rates of the land, and from modeling.

By comparison, the Antarctic and Greenland Ice Sheets underwent only small changes in volume and extent during the last glacial cycle. Volume changes of the Antarctic Ice Sheet were largely concentrated in West Antarctica and the Antarctic Peninsula. At the LGM, grounding lines advanced close to the continental shelf break almost everywhere as global sea level dropped. It is unclear whether extensive ice streams continued to exist in the Ross basin and elsewhere. Ice over interior East Antarctica was generally thinner than today because of lower accumulation rates. The Greenland Ice Sheet also extended beyond the present coastline to cover at least the inner continental shelf. Summer melting ceased and mass was lost by calving only. The Greenland Ice Sheet extended northwest to join the Laurentide Ice Sheet flowing out of Canada. Ice retreat of the Antarctic and Greenland Ice Sheets was in full progress between 13,000 and 8,000 years ago, several thousand years later than for the North American and European Ice Sheets (Huybrechts, 2002).

To date, no criteria have been developed to recognize paleo-ice shelves. It has been speculated that the Antarctic continent was entirely encircled by a vast ice shelf that merged imperceptibly with the permanent pack ice and possibly extended as far as 800 km beyond the continental shelf (Denton and Hughes, 1981). Russian scientists have hypothesized a 1 km thick Arctic ice shelf from glacial scouring of ridges at the sea floor, but its dating is uncertain (Polyak et al., 2001).

Small ice caps and glaciers merged to larger glacier complexes in Patagonia, the Alps, and many presently glaciated areas elsewhere (Pyrenees, Caucasus, eastern Siberia, New Zealand Alps, and other mountain areas). There is no convincing evidence for the occurrence of a Tibetan Ice Sheet, but glacier fronts in the Himalayas descended by about 1,000 m elevation.

In the Southern Hemisphere, the most striking difference was the winter extent of sea ice, which was significantly greater than it is today (Cooke and Hays, 1982). Summer extent was probably similar to today's winter extent. Permanent sea-ice cover also existed in the Norwegian, Greenland, and Labrador Seas coincident with a large shift in the path of the Gulf Stream, which flowed due east at 40 °N during the glacial maximum. A striking and unexplained difference was the comparably small amount of cooling in the North Pacific and the associated southward extension of sea ice at LGM compared with equivalent latitudes in the North Atlantic.

There are no proxy data to infer the seasonal variation of snow cover and river and lake ice. Nevertheless, winter cover must have been greatly expanded, commensurate with Northern Hemisphere continental coolings of up to –20 °C (CLIMAP Project members, 1981; Frenzel et al., 1992). There are also many fossil periglacial features at the surface that show that permafrost at the LGM was more extensive than today. In North America, discontinuous permafrost extended up to 1,000 km south of the southern margin of the Laurentide Ice Sheet. Permafrost was also present on most of Europe, the Atlas Mountains, Turkey, Iran, the whole of Siberia, the Tibetan Plateau, northern China, and Hokkaido. In the Southern Hemisphere, widespread permafrost extended at mid altitude in southern America, as far north as 40 °S, and was present in Patagonia, the southern island of New Zealand, and on all sub-Antarctic islands. It occurred isolated in mountainous areas of Tasmania and southeast Australia, and even in the highest mountains of East Africa and the high Andes. It never existed in South Africa, with the exception of a small spot of mountain permafrost in the Drakensberg (Van Vliet-Lanoë and Lisitsyna, 2001).

The last glacial cycle was enclosed by two periods when the cryosphere was smaller than today. During the Last Interglacial, which culminated 125,000 years ago, Northern Hemisphere temperatures at mid-latitudes were 2–3 °C warmer than today, increasing towards the pole to more than 4 °C. The Greenland Ice Sheet may have retracted to a small central ice dome, thus providing most of the melt to explain the inferred sea-level stand of approximately 6 m higher than today (Letréguilly et al., 1991; Cuffey and Marshall, 2000). The Holocene Climatic Optimum occurred around 6,000 years ago. The warming was of a similar magnitude as that of the Eemian warming, but it was shorter-lived. At that time, the Greenland Ice Sheet margin receded inland from its present position. The other components of the cryosphere, except the Antarctic Ice Sheet, were also smaller.

The cryosphere before the last glacial cycle

In geological terms, the last ice age ended a short time ago. It was the last episode of a period of climate oscillations between glacial and relatively ice-free environments that began about 2.3 million years ago. Over the last 800,000 years, such ice ages had a period of about 100,000 years; before that, 41,000-year cycles dominated. The penultimate glacial cycle, which ended 130,000 years ago, was probably the most severe. Periods of extensive glaciation also existed prior to those of the Pleistocene epoch. Though the record of early climates is dominated by evidence of global warmth and likely almost total absence of land and sea ice, there were intervals of widespread ice cover during Paleoproterozoic times, at around 2.3 billion years ago and perhaps also between 3.1 and 2.9 billion years ago.

The most extreme glacial climate in the Earth's history probably occurred during the Late Precambrian between 900 and 600 million years ago, and is known as the *Proterozoic ice house*. In its extreme version, ice sheets extended to the equator and the world's ocean was virtually entirely frozen over (Hoffman et al., 1998). At that time, the Earth experienced anomalously low atmospheric carbon dioxide concentrations, possibly as a result of mountain building and continental weathering, in addition to the Sun being 6% fainter than it is today. Recovery from such a *snowball Earth* could have resulted from volcanic outgassing of greenhouse gases. Glaciation of the Earth also occurred in the Late Ordovician at 450 million years ago, and during Carboniferous-Permian times between 320 and 250 million years ago. At this time, a single supercontinent, Gondwanaland, lay over the South Pole, which suggests that the presence of a large land mass at high latitude is a critical element for glaciation.

Glaciation on Earth started again during the Cenozoic Era, from 65 million years ago to the present. It started first in the

Southern Hemisphere and was punctuated by two periods of significant ice growth. The first ice increase occurred near the Eocene-Oligocene boundary 36 million years ago and saw frequent ice sheets growing and decaying on East Antarctica, reaching the margin of the continent in a few places (Barker et al., 1999). Global ice sheet volume increased again in the middle Miocene around 15 million years ago with the formation of a quasi-permanent East Antarctic Ice Sheet that may have reached its maximum extent. The first period of ice increase was probably driven by declining levels of atmospheric carbon dioxide, while the second period has been linked to the thermal isolation of Antarctica, as newly formed continents moved to the north and the Antarctic Circumpolar Current developed. The West Antarctic Ice Sheet likely formed later, during the Late Miocene. By 2 million years ago, the Antarctic glacial regime was much as at present and ice ages started to occur on the Northern Hemisphere, as did mountain glaciers in midlatitude highlands.

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Cross-references

Antarctic Glaciation History
Borehole Climatology: Climate Change from Geothermal Data
CLIMAP
Cordilleran Ice Sheet
Glaciations, Pre-Quaternary
Glaciations, Quaternary
Icehouse (cold) Climates
Last Glacial Maximum
Laurentide Ice Sheet
Mountain Glaciers
Periglacial Geomorphology
Pleistocene Climates
Proterozoic Climates
Scandinavian Ice Sheet

CYCLIC SEDIMENTATION (CYCLOTHEM)

The term "cyclic sedimentation" is generic and can be applied to any type or scale of repetitive sedimentation (Einsele et al., 1991). Conversely, the term "cyclothem" has a much more specific application. The concept has mostly been applied to stratigraphic successions of middle to late Pennsylvanian (upper Carboniferous) rocks that were deposited in cratonic basins of the eastern and mid-continental USA. Well-developed