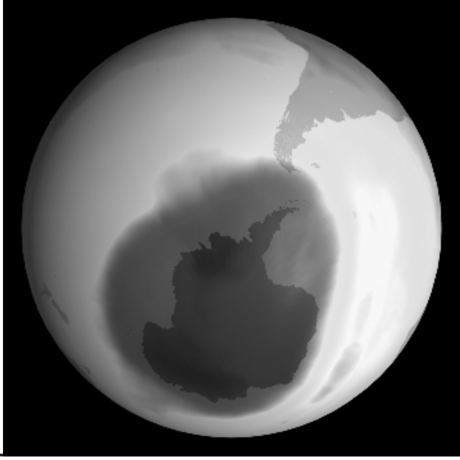
1 The climate of the Earth



JOHN LOCKWOOD

4 Woodthorne Croft, Leeds, Yorkshire, LS17 8XQ,
UK (jglockwood1@mac.com)

The causes, history and distributions of the Earth's climates are introduced in this chapter. The combination of the distribution of incoming solar radiation across the Earth's surface and the Earth's rotation, drive and shape the observed atmosphere–ocean circulation. Important factors, which determine changes in climate, include palaeogeography, greenhouse gas concentrations, changing orbital parameters, and varying ocean heat transport. One of the major controls of climatic changes is the greenhouse gas concentration of the atmosphere, in particular that of carbon dioxide. Before the Eocene–Oligocene boundary (≈ 34 Myr ago) the atmosphere–ocean circulation supported a warm atmosphere and ocean, with both poles free of permanent ice. At the Eocene–Oligocene boundary the atmosphere–ocean circulation changed to a form similar to the present, and the first evidence of an Antarctic ice sheet is found. Falling atmospheric carbon dioxide levels probably caused this change. The waxing and waning of massive temperate latitude continental ice sheets characterize the climate of the past million years. Recent climate changes are described and evidence produced that they are largely driven by anthropogenic generated atmospheric carbon dioxide. In particular recent climate changes are causing the expansion of the tropical zone and a retreat of the polar zones.

The major climate zones of the world are described, with particular attention to interannual variability, and the causes of droughts and heavy rainfalls. This includes discussions of the climatic effects of the North Atlantic Oscillation and El Niño–Southern Oscillation.

1.1 Basic climatology

The climate of a particular place is the average state of the atmosphere observed as the weather over a finite period (e.g. a season) for a number of different years. The so-called **climate system**, which determines the weather, is a composite system consisting of five major interactive adjoint components: the atmosphere, the hydrosphere, including the oceans, the cryosphere, the lithosphere, and the biosphere (Fig. 1.1). All the subsystems are open and non-isolated, as the atmosphere, hydrosphere, cryosphere and biosphere act as cascading systems linked by complex feedback processes. The climate system is subject to two main external forcings that condition its behaviour, solar radiation and the Earth's rotation. Solar radiation must be regarded as the primary forcing mechanism, as it provides almost all the energy that drives the climate system.

The distribution of climates across the Earth's surface is determined by its spherical shape, its rotation, the tilt of the Earth's axis of rotation in relation to a perpendicular line through the plane of the Earth's orbit round the Sun, the eccentricity of the Earth's orbit, the greenhouse gas content of the atmosphere and the nature of the underlying surface. The spherical shape creates sharp north–south temperature differences, while the tilt is responsible for month-by-month changes in the amount of solar radiation reaching each part of the planet, and hence the variations in the length of daylight throughout the year at different latitudes and the resulting seasonal weather cycle.

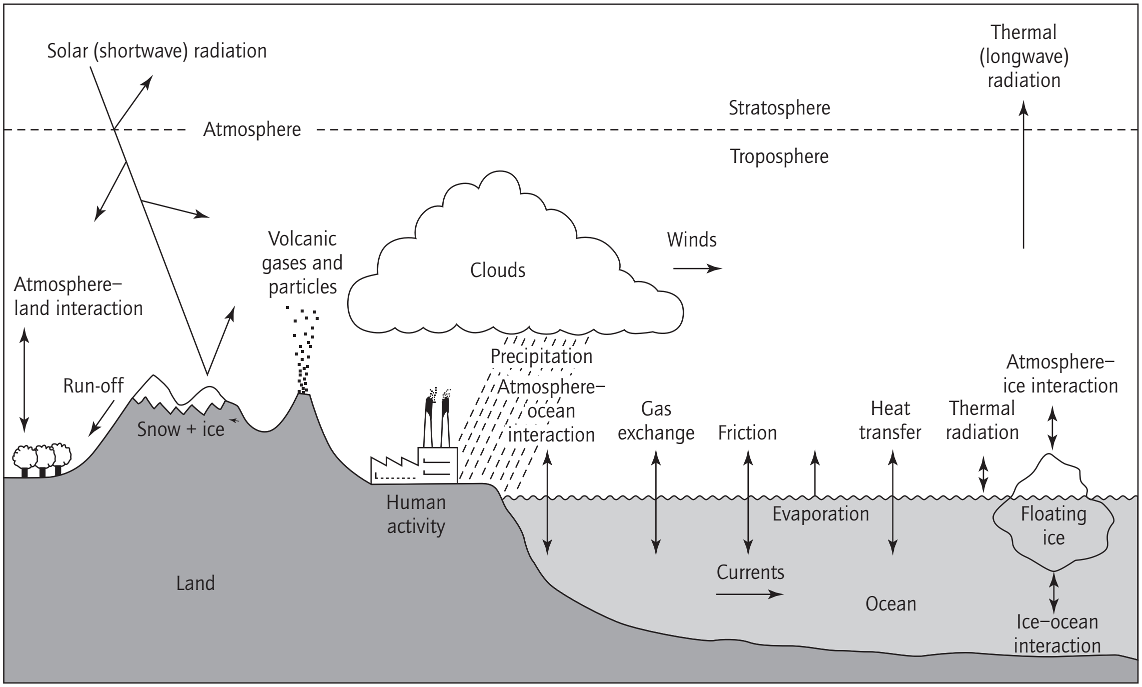


Fig. 1.1 The climate system (Houghton, 2005).

The present orbit of the Earth is slightly elliptical with the Sun at one focus of the ellipse, and as a consequence the strength of the solar beam reaching the Earth varies about its mean value. At present the Earth is nearest to the Sun in January and farthest from the Sun in July. This makes the solar beam near the Earth about 3.5% stronger than the average mean value in January, and 3.5% weaker than average in July. The gravity of the Sun, the Moon and the other planets causes the Earth to vary its orbit around the Sun (over many thousands of years). Three different cycles are present, and when combined, produce the rather complex changes observed. These cycles affect only the seasonal and geographical distribution of solar radiation on the Earth's surface, yearly global totals remaining constant. Surplus in one season is compensated by a deficit during the opposite one; surplus in one geographical area is compensated by simultaneous deficit in some other zone. Nevertheless, these Earth orbital variations can have a significant effect on climate and are responsible for some major long-term variations.

Firstly, there are variations in the **orbital eccentricity**. The Earth's orbit varies from almost a complete circle to a marked ellipse, when it will be nearer to the Sun at one particular season. A complete cycle from near circular through a marked ellipse back to near circular takes between 90 000 and 100 000 years. When the orbit is at its most elliptical, the intensity of the solar beam reaching the Earth must undergo a seasonal range of about 30%. Secondly, there is a wobble in the Earth's axis of rotation causing a phenomenon known as the **precession of the equinoxes**. That is to say, within the elliptical orbits just described, the distance between Earth and Sun varies so that the season of the closest approach to the Sun also varies. The complete cycle takes about 21 000 years. Lastly, the **tilt of the Earth's axis of rotation** relative to the plane of its orbit varies at least between 21.8° and 24.4° over a regular period of about 40 000 years. At present it is almost 23.44° and is decreasing. The greater the tilt of the Earth's axis, the more pronounced the difference between winter and summer. Technically these three mechanisms are known as the **Milankovitch** mechanism.

If the Earth did not rotate relative to the Sun, that is it always kept the same side toward the Sun, the most likely atmospheric circulation would consist of rising air over an extremely hot, daylight face and sinking air over an extremely cold, night face. The diurnal cycle of heating and cooling obviously would not exist since it depends on the Earth's rotation. Surface winds everywhere would blow from the cold night face towards the hot daylight face, while upper flow patterns would be the reverse of those at lower levels. Whatever the exact nature of the atmospheric flow patterns, the climatic zones on a non-rotating Earth would be totally different from anything observed today. Theoretical studies suggest that if this stationary Earth started to rotate, then as the rate of rotation increased, the atmospheric circulation patterns would be progressively modified until they resembled those observed today. In very general terms, these circulation patterns take the form of a number of meridional overturning cells in the atmosphere, with separate zones of rising air motion at low and middle latitudes, and corresponding sinking motions in subtropical and polar latitudes.

1.2 General atmospheric circulation

A schematic representation of the mean meridional circulation between Equator and pole is shown in Fig. 1.2. A simple direct circulation cell, known as the **Hadley cell**, is clearly seen equatorward from 30° latitude (Lockwood, 2003). Eastward angular momentum is transported from the equatorial latitudes to the middle latitudes by nearly horizontal eddies, 1000 km or more across, moving in the upper troposphere and lower stratosphere. This transport, together with the dynamics of the middle latitude atmosphere, leads to an accumulation of eastward momentum between 30° and 40° latitude, where a strong meandering current of air, generally known as the subtropical westerly jet stream, develops (Fig. 1.3). The cores of the subtropical westerly jet streams in both hemispheres and throughout the year occur at an altitude of about 12 km. The air subsiding from the jet streams forms the belts of subtropical anticyclones at about 30 to 40° N and S (Fig. 1.4). The widespread subsidence in the descending limb of the Hadley cell should be contrasted with the rising limb, where ascent is restricted to a few local areas of intense convection. More momentum than is necessary to maintain the subtropical jet streams against dissipation through internal friction is transported to these zones of upper strong winds. The excess is transported downwards and polewards to maintain the eastward-flowing surface winds (temperate latitude westerlies) against ground friction. The middle latitude westerly winds are part of an indirect circulation cell known as the **Ferrel cell**. The supply of eastward momentum to the Earth's

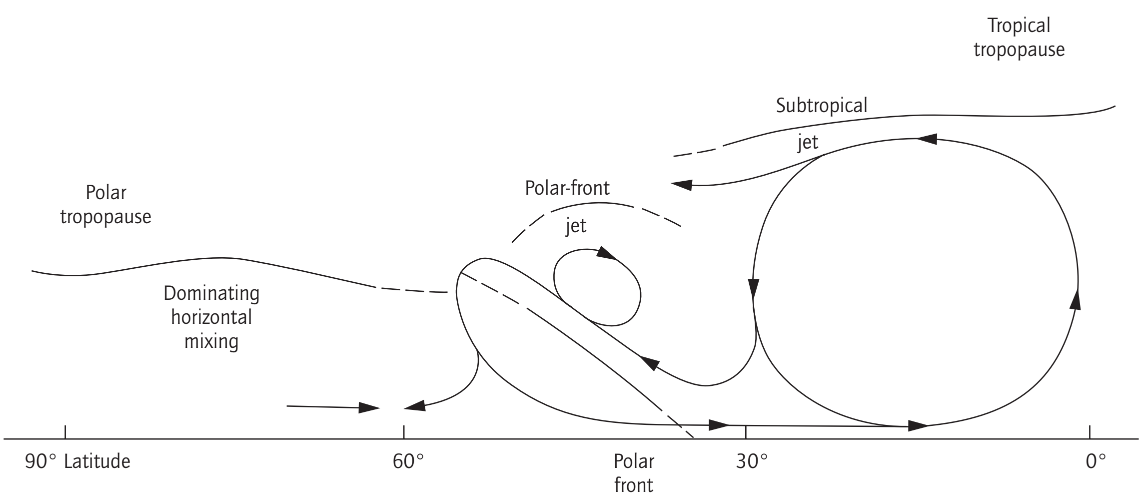


Fig. 1.2 Schematic representation of the meridional circulation and associated jet-stream cores in winter. The tropical Hadley Cell and middle latitude Ferrel Cell are clearly visible. (From Palmen, 1951.)

(a)

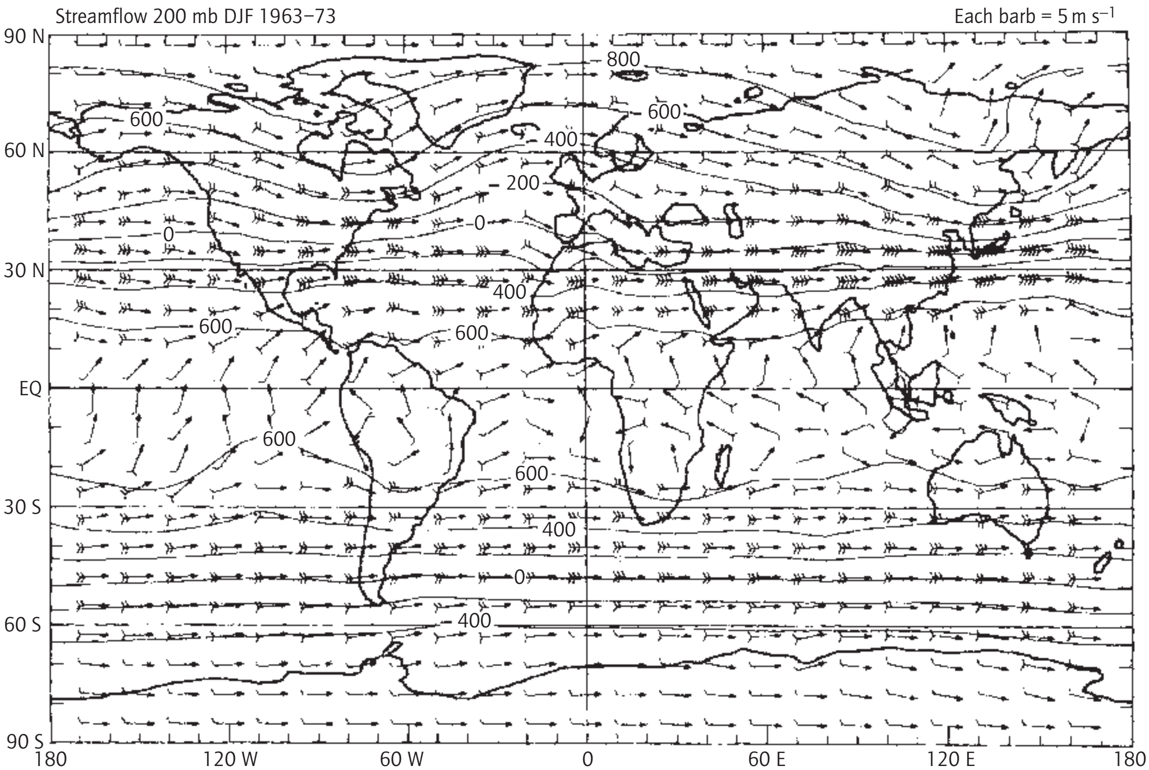


Fig. 1.3 Global distribution of the mean height (1963–73) of the 200 hPa pressure field represented as mean height minus 11 784 gpm for (a) December, January, February, (b) June, July, August. Wind speed and direction shown by arrows. Each barb on the tail of an arrow represents a wind speed of 5 m s^{-1}. (From Peixoto & Oort, 1992.)

surface in middle latitudes tends to speed up the Earth's rotation. Counteracting such potential speeding up of the Earth's rotation, air flows from the subtropical anticyclones toward the equatorial regions, forming the so-called **trade winds**. The trade winds, with a strong flow component directed toward the west (easterly winds), tend to retard the Earth's rotation, and in turn gain eastward momentum.

The greatest atmospheric variability occurs in middle latitudes, from approximately 40 to 70º N and S, where large areas of the Earth's surface are affected by a succession of eastward moving cyclones (frontal depressions) and anticyclones or ridges. This is a region of strong north–south thermal gradients with vigorous **westerlies** in the upper air at about 10 km, culminating in the polar-front jet streams along the polar edges of the Ferrel cells (Fig. 1.2). The zone of westerlies is permanently unstable and gives rise to a continuous stream of large-scale eddies near the surface, the cyclonic eddies moving eastward and poleward and the anticyclonic ones moving eastward and equatorward. In contrast, at about 10 km, in the upper westerlies, smooth wave-shaped patterns are the general rule. Normally there are four or five major waves around the hemisphere, and superimposed on these are smaller waves that travel through the slowly moving train of larger waves. The major waves are often called **Rossby waves**, after Rossby who first investigated their principal properties. Compared with the Hadley cells the middle latitude atmosphere is highly disturbed and the suggested meridional circulation shown in Fig. 1.2 is largely schematic.

The extension into very high latitudes and the northward narrowing of the northern North Atlantic have consequences on the Atlantic Ocean

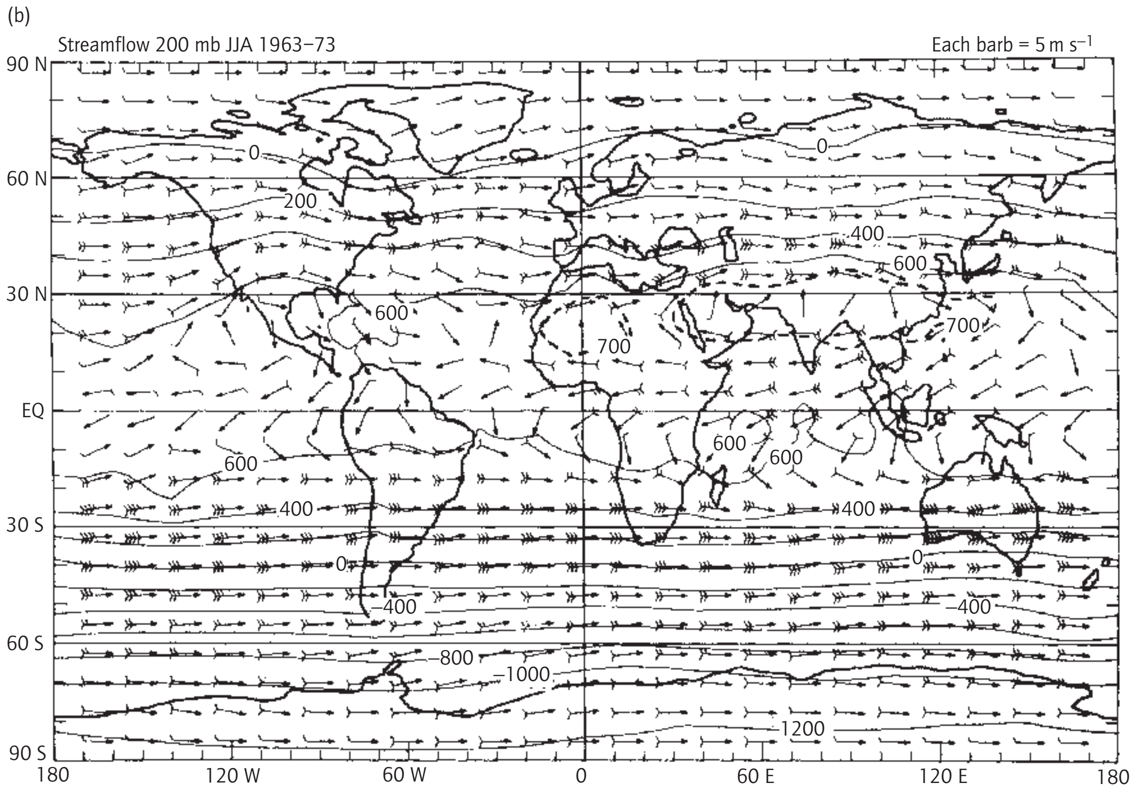


Fig. 1.3 (*Continued*)

circulation, which in turn has a series of unique effects on the climate system. This is in complete contrast to the much more benign North Pacific Ocean. Warm, saline surface water flows into the northern North Atlantic, after travelling from the Caribbean Sea, via the Gulf Stream and the North Atlantic Drift. This inflowing water, which is more saline than anywhere else in the high-latitude oceans, is finally advected to sites in the Greenland and Norwegian Seas, where extreme cooling to the atmosphere occurs and surface water sinks to the ocean depths. When cooled, water with the salinity normal in the world's oceans becomes denser, but unlike fresh water does not reach its maximum density until near its freezing point, at about −2º C. Thus the salt water of the deep oceans, when cooled at the surface, goes into convective patterns, the coldest and densest portions gradually sinking from the surface to the ocean depths. Low-density surface layers in the oceans can arise either because of surface heating, or the addition of relatively fresh continental runoff or precipitation onto the ocean surface. In the cold oceans, sea-ice will form only when a layer of the ocean close to the surface has a relatively low salinity. The existence of this layer allows the temperature of the surface water to fall to freezing point, and ice to form, despite the lower levels of the ocean having a higher temperature.

1.3 Palaeoclimates

The major controls of very long-term climatic change include palaeogeography, greenhouse gas concentrations, changing orbital parameters, and varying ocean heat transport. One of the major controls on long-term climatic changes is the greenhouse gas concentration of the atmosphere and in particular that of carbon dioxide. Atmospheric carbon dioxide concentration has varied markedly during the Earth's history. Atmospheric

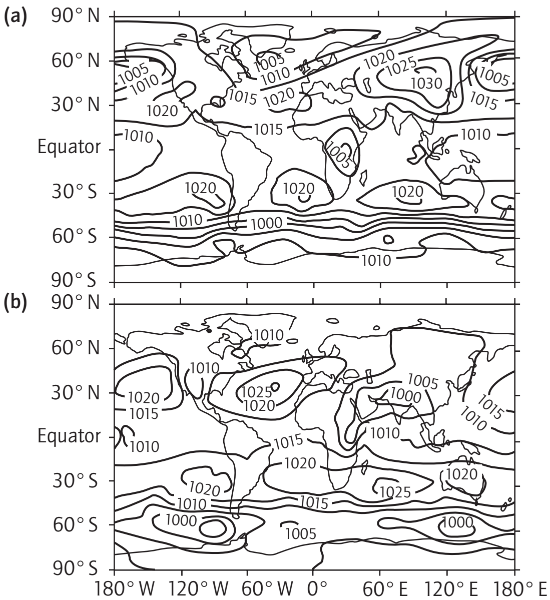


Fig. 1.4 Mean sea-level pressure (hPa) (1963–73) averaged for (a) December, January, February and (b) June, July, August. (From Henderson Sellers & Robinson, 1986; Oort, 1983.)

CO_2 concentrations are controlled by the carbon cycle, and the net effect of slight imbalances in the carbon cycle over tens to hundreds of millions of years has been to reduce atmospheric CO_2. Atmospheric CO_2 concentrations remained relatively high up to about 60 Myr ago when there was a very marked fall. Atmospheric concentrations continued to fall after about 60 Myr ago and there is geochemical evidence that concentrations were less than 300 ppm by about 20 Myr ago.

Available evidence is that during the Mesozoic era, temperatures ranged from 10 to 20° C at the poles to 25 to 30° C at the Equator, that is the poles were free of permanent ice fields, and the atmosphere–ocean circulation was different in some important aspects from that observed today. Slight cooling took place at the start of the Jurassic Period and marked high-latitude warming during the first half of the Cretaceous Period. Global cooling again took place towards the end of Cretaceous time, and a long-term cooling trend commenced at the start of the Eocene Epoch, some 55 Myr ago.

The sudden, widespread glaciations of Antarctica and the associated shift towards colder temperatures at the Eocene–Oligocene boundary (approximately 34 Myr ago) is one of the most fundamental reorganizations of global climate and ocean circulation known in the geological record. Prior to the Eocene–Oligocene boundary there is little evidence of the deep cold water in the world ocean that is so common today. Indeed before the boundary, atmospheric and particularly oceanic circulation conditions were probably different from those observed today. After the boundary they are probably rather similar to present-day conditions. Oceanic bottom water is formed in small regions by convective buoyancy plumes that transfer relatively dense ocean water from near the surface to the ocean depths. Deep-ocean temperatures are therefore closely related to ocean surface temperatures in key regions. The surface density and salinity is usually increased by evaporation and heat transfer to the atmosphere, therefore virtually all deep-water formation seems to be over continental shelves in low latitudes or at high latitudes. During the Cretaceous Period and up to the end of the Eocene Epoch, the ocean bottom-waters were warm, saline and formed in shallow subtropical marginal seas. At the Eocene–Oligocene boundary, ocean bottom-water temperatures decreased rapidly to approximately present-day levels. Deep-sea cores suggest that this change occurred within 100 000 years, which is remarkably abrupt for pre-glacial Tertiary times, and is considered to represent the time when large-scale freezing conditions developed at sea-level around Antarctica, forming the first significant sea-ice. At this time cold-water plumes forming off Antarctica started to dominate ocean bottom-water formation, and together with Arctic Ocean plumes they have dominated until the present day. Thus from early Oligocene times onwards it may be considered that world climates were in the present cold or semi-glacial state.

The initial growth of the East Antarctic Ice Sheet near the Eocene–Oligocene boundary is often attributed to the opening by continental drift of ocean gateways between Antarctica and Australia (Tasmanian Passage) and Antarctica and South America (Drake Passage), leading to the organization of the Antarctic Circumpolar current and the 'thermal isolation' of Antarctica. This notion has been challenged because although most tectonic reconstructions place the opening of the

Tasmanian Passage close to the Eocene–Oligocene boundary, the Drake Passage may not have provided a significant deep-water passage until several million years later. Recent model simulations (DeConto & Pollard, 2003) of the glacial inception and early growth of the East Antarctic Ice Sheet suggest that declining Cenozoic carbon dioxide first leads to the formation of small, highly dynamic ice caps on high Antarctic plateaux. At a later time, a carbon dioxide threshold is crossed, initiating various feedbacks that cause the ice caps to expand rapidly with large orbital variations, eventually coalescing into a continental-scale East Antarctic Ice sheet. According to this simulation the opening of the two Southern Ocean gateways plays a secondary role in this transition, relative to changing carbon dioxide concentration.

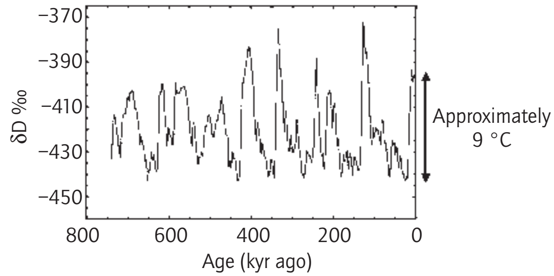


Fig. 1.5 The figure shows measurements deduced from ice-cores drilled from the Antarctic ice-sheet, and analysed by the British Antarctic Survey and others as part of the European programme EPICA. The actual measurement is of the concentration of deuterium in air bubbles, and this can be related to local temperatures. The figure shows that temperature rise between the depth of the last ice age 20 000 years ago and the current interglacial is about 9°C. (Hadley Centre for Climate Prediction and Research. From EPICA Community Members, *Nature*, 2004.)

Quaternary glaciations

Continental ice-sheets probably appeared in the Northern Hemisphere about 3 Myr ago, occupying lands adjacent to the North Atlantic Ocean. The time of the formation of the Greenland ice-sheet is not well known from terrestrial evidence, but the presence of glacial marine sediments in North Atlantic marine cores first appeared around 3 Myr ago. The oldest glacial moraines in Iceland are dated to approximately 2.6 Myr ago. For at least the past million years the Earth's climate has been characterized by an alternation of glacial and interglacial episodes, marked in the Northern Hemisphere by the waxing and waning of continental ice-sheets and in both hemispheres by rising and falling temperatures (Fig. 1.5). The present dominant cycle is one of about 100 kyr and is seen in the growth and decay of the continental ice-sheets. The last major glacial episode started about 110 kyr ago and finished only about 10 kyr ago. These fluctuations or cycles are found in a large number of **proxy** data records, analysis of which suggests that Antarctic air temperature, atmospheric CO_2 content and deep-ocean temperatures are dominated by variance with a 100 kyr period and vary in phase with orbital eccentricity. In contrast global ice volume lags changes in orbital eccentricity (Shackleton, 2000). Hence, the 100-kyr ice-sheet cycle does not arise from ice-sheet dynamics; instead, it is probably the response of the global carbon cycle to changes in orbital eccentricity that generates the eccentricity signal in the climate record by causing changes in atmospheric carbon dioxide concentrations.

Proxy data records can be grouped into two climatic regimes with the transitional zone about 430 kyr ago (Fig. 1.5). The earlier shows higher-frequency cycles (dominance of 40-kyr cycles), with less coherence among the various proxy climatic records, than the later one (dominance of 100-kyr cycles). This may be due to a decrease in average atmospheric CO_2 levels over the past 2 million years (Brook, 2005). There are differences in the amplitudes of deuterium and CO_2 oscillations before and after 430 kyr ago. The atmospheric concentration of CO_2 did not exceed 300 ppmv for the 650 000 years before the beginning (around 1750) of the industrial era. Since the Industrial Revolution, atmospheric carbon dioxide concentrations have increased by 33%. Before 430 kyr ago concentrations of CO_2 did not exceed 260 ppmv.

The transition from glacial to interglacial conditions about 430 kyr ago resembles the transition into the present interglacial period at about 10 kyr ago in terms of the magnitude of changes in temperature and greenhouse gases. As commented above the transition 430 kyr ago also delimits the frontier between two different patterns of climate, and has been identified by recent investigations as a unique and exceptionally long interglacial. Some workers (see Brook, 2005) suggest that because the orbital parameters (low eccentricity and consequently weak precessional forcing) are similar to those of the present and next tens of thousands of years, the interglacial 430 kyr ago may be the best analogue available for present and future climate without human intervention. Long interglacials with stable conditions are not therefore exceptional, short interglacials such as the past three are not the rule and hence cannot serve as analogues of the present Holocene interglacial.

Sudden and short-lived climate oscillations giving rise to warm events occurred many times during the generally colder conditions that prevailed during the last glacial period between 110 000 and 10 000 years ago (Lockwood, 2001). They are often known as 'interstadials' to distinguish them from the cold phases or 'stadials'. Between 115 000 and 14 000 years ago there were 24 of these oscillations, as recognized in the Greenland ice-core records where they are called **Dansgaard–Oeschger oscillations**. These can be viewed as oscillations of the climate system about an extremely ill defined mean state. Each oscillation contains a warm interstadial, which is linked to and followed by a cold stadial. Ice-core and ocean data suggest that the oscillations began and ended suddenly, although in general the 'jump' in climate at the start of an oscillation was followed by a more gradual decline that returned conditions to the colder 'glacial' state. Warming into each oscillation occurred over a few decades or less, and the overall duration of some of these warm phases may have been just a few decades, whereas others vary in length from a few centuries to nearly 2000 years.

Of totally different nature to Dansgaard–Oeschger oscillations are extreme and short-lived cold events, known as **Heinrich events**. These events occurred against the general background of the glacial climate and represent the climatic effects of massive surges of fresh water and icebergs from melting ice sheets into the North Atlantic, causing substantial changes in the thermohaline circulation (Lockwood, 2001). Several massive ice-rafting events show up in the Greenland ice-cores as a further 3–6° C drop in temperature from already cold glacial conditions. Many of these events have also been picked up as particularly cold and arid intervals in European and North American pollen records. The most recent Heinrich event is known as the Younger Dryas and appears as a time of glacial readvance in Europe after the end of the main ice-age.

The recent climate record

Extensive instrumental temperature records exist only for the period after about 1860, but recently multiproxy data networks (Mann *et al.,* 1999) have been used to reconstruct Northern Hemisphere temperatures back to AD 1000 (Fig. 1.6). This reconstruction shows a long-term cooling trend in

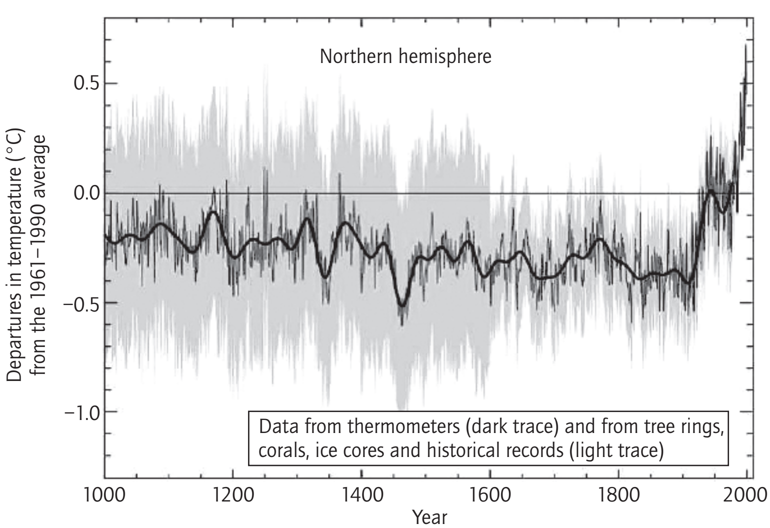


Fig. 1.6 Reconstructed Northern Hemisphere average surface temperatures back to AD 1000. Up to the mid-nineteenth century the record is based on proxy data (e.g. tree rings, etc.; see Mann *et al.*, 1999). Since the mid-nineteenth the record is based on instrumental data.

the Northern Hemisphere prior to industrialization of –0.02° C per century, possibly related to orbital forcing, which is thought to have driven long-term temperatures downward since the mid-Holocene at a rate within the range from –0.01 to –0.04° C per century. The temperature reconstruction also shows that the late eleventh, twelfth and fourteenth centuries rival mean twentieth century temperature levels, while cooling following the fourteenth century can be viewed as the initial onset of the cold period known as the Little Ice Age. There is general agreement that the Little Ice Age came to an abrupt end around 1850, while studies in Switzerland indicate that overall the coldest conditions of the past 500 years were in the late seventeenth and early nineteenth centuries. The early nineteenth century was especially cold and can be considered as the 'climatic pessimism' of the past 1000 years.

Global mean surface temperature has increased since the climatic pessimism of the past 1000 years in the early nineteenth century, but not in a uniform manner (Fig. 1.7). The global increase in temperature since 1880 occurred during two sustained periods, one beginning around 1910 and another during the 1970s. Best available estimates at present (Jones & Moberg, 2003) give global temperature trends from 1910 to 1945 of 0.11° C per decade, –0.01° C per decade from 1946 to 1975 and 0.22° C per decade from 1976 to 2000. The warmest year so far in this time series was 1998, followed by 2002, 2003 and 2004. All the 'top ten' years have been since 1990, and all the 'top twenty' warmest years have been since 1981, with the exception of one (1944). The largest recent warming is in the winter extratropical Northern Hemisphere, with a faster rate of warming over land compared with the ocean. Using satellite-borne microwave sounding units, Qiang Fu *et al.* (2006) have examined atmospheric temperature trends for 1979–2005. They found that relative to the global-mean trends of the respective layers, both hemispheres have experienced enhanced tropospheric warming and stratospheric cooling in the 15 to 45° latitude belt. This suggests a widening of the tropical circulation zone and a poleward shift of the subtropical jet streams and their associated subtropical dry zones.

Minimum temperatures for both hemispheres increased abruptly in the late 1970s, coincident with an apparent change in the character of the El Niño-Southern Oscillation (ENSO) phenomenon, giving persistently warmer sea temperatures in the tropical central and east Pacific. The more reliable data sets show that it is likely that total atmospheric water vapour has increased several per cent per decade over many regions of the Northern Hemisphere since the early 1970s. Similarly it is likely that there has been an increase in total cloud cover of about 2% over many mid- to high-latitude land areas since the beginning of the twentieth century. The increases in total cloud amount are positively correlated with decreases in the diurnal temperature range.

The observed temperature record since the early 1900s can be partly explained in terms of natural

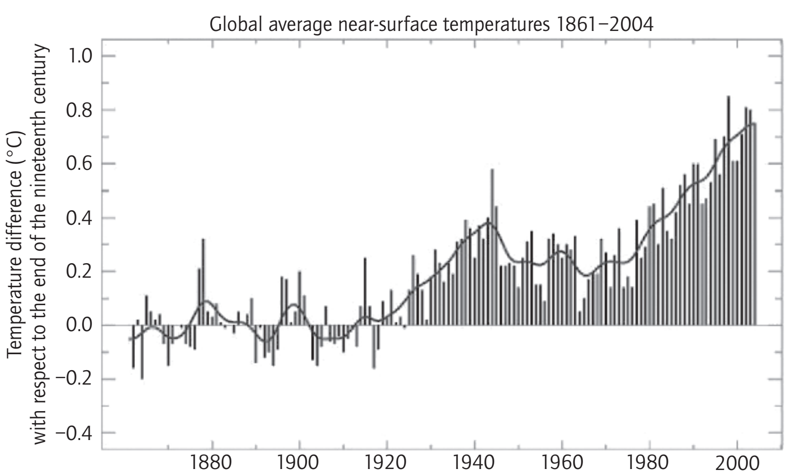


Fig. 1.7 Change in global average surface temperature from 1861 to 2004. Bars show the individual annual averages, and the continuous line shows a smoothed trend, with changes shown relative to temperatures over the last decades of the nineteenth century. (Hadley Centre for Climate Prediction and Research.)

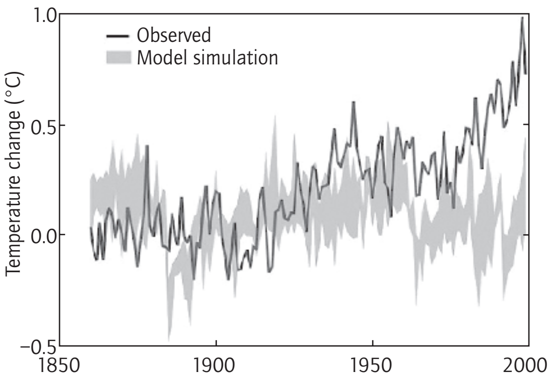


Fig. 1.8 Natural factors cannot explain recent warming. Shaded area shows result of the Hadley Centre climate model driven by natural factors, but excluding anthropogenic carbon dioxide and sulphate particles. Continuous line shows actual observations. (Hadley Centre for Climate Prediction and Research.)

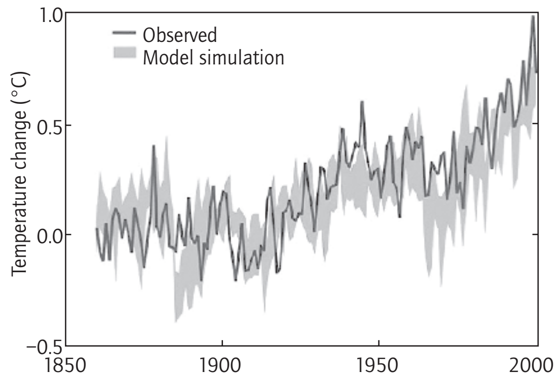


Fig. 1.9 The Hadley Centre climate model is now driven by changes in anthropogenic factors – changes in greenhouse gas concentrations and sulphate particles – in addition to natural factors; observations (line) and model simulations (shaded area) are in much better agreement. (Hadley Centre for Climate Prediction and Research.)

factors such as the chaotic variability of climate due largely to interactions between atmosphere and ocean; changes in the output of the Sun and changes in the optical depth of the atmosphere from volcanic emissions. Some of the computations made on this basis are shown in Fig. 1.8 which clearly do not agree with the observations, particularly in the period since about 1970 when observed temperatures have risen by about 0.5° C, but those simulated from natural factors have hardly changed at all.

Sulphate aerosol particles scatter some sunlight, which would otherwise reach the surface of the Earth and heat it, back to space. They therefore have a cooling influence on climate. They are formed in the lower atmosphere when sulphur dioxide, emitted by human activities such as power generation and transport, is oxidized. Other sources of sulphur dioxide are 'background' non-explosive volcanoes, and natural di-methyl sulphate (DMS) from ocean plankton. If the climate model used in Fig. 1.9 is now driven by changes in human-made factors, such as sulphate aerosols and changes in greenhouse gas concentrations, in addition to natural factors, observations and model simulation are in much better agreement (Fig. 1.9). In particular, the warming since about 1970 is clearly simulated (Stott *et al.*, 2000). Further work has recently demonstrated that warming over individual continental areas in the past few decades can be explained only if human activities are included.

1.4 Polar climates

The most distinctive climatic features in both North and South Polar Regions are the presence of ice, snow, deeply frozen ground and a long winter period of continuous darkness (Orvig, 1970). The radiation budget of the polar surfaces is nearly always negative because of the absence in winter of any solar radiation, the low angle of solar incidence in the summer and the high albedo of the ice fields. Indeed, the polar regions serve as global sinks for the energy that is transported poleward from the tropics by warm ocean currents and by atmospheric circulation systems, particularly travelling cyclones and blocking anticyclones. However, there are major differences between the Arctic and Antarctic regions in terms of the distribution of land, sea and ice. The Arctic is a largely ice-covered, land-locked ocean, whereas Antarctica consists of a continental ice-sheet over 2000 m thick surrounded by ocean.

Surrounding the Antarctic continent is a zone of floating sea-ice, about 1.5 m, thick which undergoes a marked annual cycle from a minimum in February–March to a maximum in September–

October. In the summer the Antarctic sea-ice melts almost back to the coastline of the continent, so only slightly more than 10% survives the summer. The winter advance of the sea-ice is not restricted by land at its equatorward boundary and it crosses the Antarctic circle.

In contrast, in winter the southern limits of the Arctic sea ice are constrained by the northern coastlines of Asia and North America, but pack-ice extends into middle latitudes off eastern Canada and eastern Asia where northerly winds and cold currents transport ice far southward. West of Norway there is open water to about 78° N because of the warm waters of the North Atlantic Drift and thermohaline circulation. The North Atlantic thermohaline circulation is particularly vigorous and its climatic effect is often illustrated by comparing the surface temperature of the northern Atlantic with comparable latitudes of the Pacific, since the former is 4 to 5° C warmer. During summer the sea-ice melts back towards the pole, but much of the Arctic ice survives at least one summer melt season, and as a consequence its mean thickness is 3–4 m. The thickness is highly variable locally, with narrow openings (leads) throughout the pack-ice even in winter, and larger openings (polynyas) adjacent to coastal areas where winds blow offshore. While there appears to have been no significant long-term trend in Antarctic sea-ice extent, Arctic sea-ice extent has decreased by about 2.5% per decade since 1970. More than 80% of Greenland is covered by an extensive ice sheet that rises above 3000 m in elevation.

Arctic

In a climatological sense, the Arctic corresponds approximately to the land areas beyond the northern limits of forests and ocean areas with at least a seasonal sea-ice cover. The boundary between Boreal forest and tundra vegetation corresponds closely to the 10° C mean isotherm of the warmest month and to the average July location of the Arctic Front separating Arctic and temperate air masses. At high latitudes, the total daily solar radiation depends largely on day-length, which in turn varies widely with the season of the year. The radiation climate produced with continuous darkness in winter and continuous daylight in summer is distinctly different from that found at lower latitudes (Serreze & Barry, 2005).

There are two main areas of semi-permanent low level inversions (atmospheric temperature increases with height) in the world: the subtropical belt and the polar regions. The main cause of the polar inversions is the negative energy balance at the surface, but the inversion is also maintained by warm air advection aloft with associated subsidence. The strongest vertical temperature gradients are therefore often found near the surface and over the Central Polar Ocean the temperature inversion persists for at least 60% of the days in all months, reaching a maximum in late winter of 100%. Because of this the surface air temperature is primarily dependent on the nature of the surface conditions. The prevailing summer melting of snow and ice holds the surface air temperature close to 0° C, but slightly positive temperatures are usually observed near the pole in the second half of June (Fig. 1.10). Relatively warm water below the sea-ice keeps winter temperatures around −30 to −35° C, since sensible heat is conducted upwards through the sea-ice to warm the lowest air layers. Indeed, the lowest winter surface temperatures are not found over the central Polar Ocean, but rather over the land areas of northeast Siberia and the Yukon (Fig. 1.10a). Here, under the influence of the continental anticyclones, calm, clear, subsiding air becomes trapped in hollows and the temperature fall below −50° C on occasion and winter monthly means below −40° C are common. Rapid loss of heat during the long winter nights in the form of longwave radiation cools the ground surface but in contrast to the sea-ice, this heat cannot be replaced. Similar low temperatures are found on the high plateau of the Greenland ice-sheet. Solar forcing of the diurnal temperature cycle is at a minimum in polar regions, making the polar diurnal temperature cycle extremely weak and difficult to compare directly with cycles at locations nearer to the Equator with a stronger diurnal solar radiation signal. During the winter months of December to February there is almost continuous darkness in the Arctic, therefore the conventional concept of solar radiative forced maximum and minimum temperatures does not apply and diurnal temperature range is governed by the atmospheric circulation.

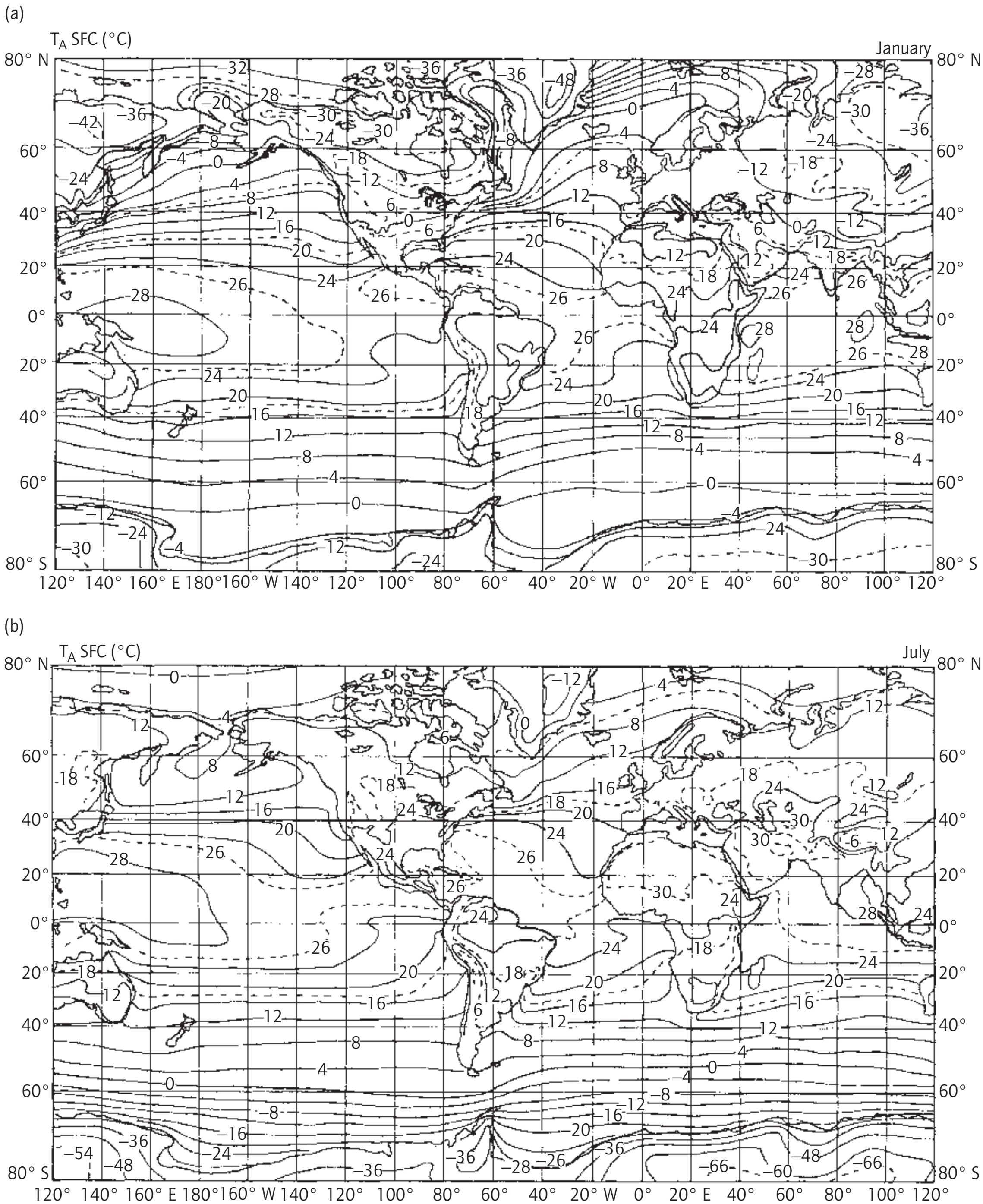


Fig. 1.10 Mean surface air temperature for (a) January and (b) July. (From Natural Climate Data Center, 1987; Peixoto & Oort, 1992.)

Antarctic climates

The Antarctic continent consists of two domed ice sheets covering about 97.6% of its area at an average elevation of about 2200 m. The first evidence of the formation of Antarctic ice-sheets is at the Eocene – Oligocene boundary (≈ 34 Myr ago). The high plateau of East Antarctica has extensive areas above 3000 m. This desolate world of ice is the coldest and windiest of the world's great land masses. On 24 August 1960, Vostock (78.5° S, 106.9° E; 3488 m) recorded a temperature of −88.3° C, which is probably the lowest temperature recorded on the Earth's surface. Antarctica and its environs play an important role in the workings of climate on a global scale because the region acts as a large sink in the global energy cycle (King & Turner, 1997; Simmons, 1998).

Mean surface temperatures at coastal stations typically range, depending on location and other factors, between 0 and −5° C in January and −15 and −25° C in July. The seasonal cycle of surface air temperature at South Pole Station (Table 1.1) shows a number of interesting features that are peculiar to Antarctica. The temperature rises rapidly in the southern spring (October and November), then averages −28° C for the summer months of December and January. In February and March the temperature falls rapidly by nearly half a degree Celsius per day, reaching winter levels by the March equinox.

Table 1.1 Mean temperature data (°C) for South Pole (1957–1966): latitude 90° S, elevation 2800 m

Month	Daily mean
January	–28.8
February	–40.1
March	–54.4
April	–58.5
May	–57.4
June	–56.5
July	–59.2
August	–58.9
September	–59.0
October	–51.3
November	–38.9
December	–28.1

Source: After Orvig (1970).

The shortness of the Antarctic summer is caused by the variations of global radiation and the surface albedo. The maximum global radiation is on average around the time of the summer solstice. In the weeks proceeding the solstice the surface albedo decreases, due to the increasing solar elevation above the horizon and a slight metamorphosis of the snow cover. It is normal for the albedo of many surfaces to decrease with increasing solar elevation, thus more solar radiation can be absorbed at the snow surface. After the solstice, the global radiation decreases, and the first light snowfall or influx of drifting snow restores surface conditions to those favouring higher albedo values. Hence there are good reasons to expect absorbed solar radiation to decrease rapidly in January and the surface temperature to fall quickly. During the Antarctic winter the intense radiational cooling and the highly transmissive atmosphere generates a persistent low-level temperature inversion that reaches its greatest depth over the higher elevations of the ice-sheet, where it is present almost the entire year. The inversion is no stronger at the end than at the beginning of the polar night. This is because equilibrium is reached between the surface longwave radiation loss, which decreases as temperature falls, and the downward radiation from the inversion layer, which changes relatively little with time. The absence of a clearly defined time for the occurrence of the surface minimum temperature is a feature of the Antarctic, comprising the so-called 'coreless winter'.

Katabatic winds occur when cooling causes a shallow blanket of air adjacent to the surface to become colder and therefore heavier than the atmosphere above; the air then drains downslope under the influence of gravity. These winds are most persistent where the ground is covered by ice and snow and they dominate the surface wind regime of Antarctica. Nearly everywhere on the ice-sheet the surface winds are directed downslope from the interior towards the surrounding ocean. The surface katabatic winds are related to the energy balance of the low-level inversion over Antarctica. Radiosonde measurements at the South

Pole (Schwerdtfeger, 1984) indicate an average radiative cooling rate of 4° C per day at the inversion level. This is the cooling rate that appears to be occurring from radiative balance considerations alone, the energy is lost by longwave radiation to space. The real cooling rate in the atmosphere obtained from temperature observations is 1° C or less per month. Thus other atmospheric processes must be compensating for the longwave radiative loss to space. These additional atmospheric processes are identified as horizontal advection and adiabatic sinking. Air converges over Antarctica at middle levels, sinks and warms due to adiabatic compression, causing the low level inversion over the ice surface. The sinking air cools rapidly by infrared emission in the inversion layer, as explained earlier. The radiatively cooled air sinks through the inversion layer, is replaced by sinking air from above and flows away as the surface katabatic wind.

1.5 Temperate latitude climates

Daily weather charts for any extensive region in middle latitudes reveal well-defined **synoptic systems** that normally move from west to east with a speed that is considerably smaller than their constituent air currents. At the surface the predominant features are closed cyclonic and anticyclonic systems of irregular shape, while at higher altitudes in the atmosphere smooth wave-shaped patterns are the general rule. These upper Rossby waves are important because surface synoptic systems tend to move in the direction of the broad upper flow with a velocity that is proportional to the intensity of the upper flow. So an intense upper flow pattern is associated with the rapid passage eastward of frequent surface frontal depressions and anticyclones. Surface frontal depressions also tend to form slightly downwind of upper troughs and similarly surface anticyclones tend to form slightly downwind of upper ridges. Major temperate, climatological cyclone development regions are therefore located just downwind of upper climatological troughs.

Northern temperate latitude Rossby waves tend to be locked in preferred locations which are shown in Fig. 1.3. These preferred locations may arise because the atmospheric circulation is influenced not only by the thermal properties of land and sea, but also by high mountain ranges and highlands in general. When a westerly air stream crosses a north–south aligned mountain range such as the Rocky Mountains, anticyclonic deformation of flow is found over the mountain range and cyclonic deformation to leeward. These orographically generated ridges and troughs propagate downstream and can reinforce similar features generated by the direct thermal effects of land and sea.

Europe

European climate is particularly interesting because it is likely that much of the southern and central parts of the continent could experience over the twenty-first century marked decreases in annual precipitation, and corresponding increases in temperature. In contrast Scandinavia could experience increasing annual precipitation. Of the elements governing the distribution of climate, the most basic is radiation from the Sun and its balance with outgoing longwave radiation from the Earth. The southernmost rim of Europe lies at about 35° N, almost within the subtropics, and its northernmost island extensions reach 78° N, almost in the Arctic. One important consequence of this large latitudinal spread is that the length of daylight at the solstices varies significantly over the region. Except along the southern rim, the net radiation balance of the continental surface in winter is negative, but the energy lost by the continual cooling of the land is replaced by vigorous warm air flow from the Atlantic Ocean to the west. Variations in the vigour of this westerly flow because of the **North Atlantic Oscillation** are of great significance in relation to the level of European winter temperatures and are discussed in detail later. Because the warm air masses from the Atlantic Ocean cool as they pass eastward over the winter continental surface, the mean isotherms in January run north–south, with temperatures falling both eastward away from the Atlantic and northward toward the Arctic (Fig. 1.10a). In contrast, in summer the net radiation balance over the continental surface is positive: values fall towards the north, so isotherms are aligned east-west (Fig. 1.10b).

During the summer months, the prevailing westerlies retreat northward from southern Europe, leaving it under the influence of the subsiding air of the subtropical anticyclones and creating an intense dry season with only occasional rainfall from convective showers. In winter the westerlies return to southern Europe, bringing frequent depressions that cause a precipitation maximum in the winter half of the year.

A major source of interannual variability in the atmospheric circulation over the North Atlantic and western Europe is the North Atlantic Oscillation (NAO), which is associated with changes in the strength of the oceanic surface westerlies. Its influence extends across much of the North Atlantic and well into Europe and it is usually defined through the regional sea-level pressure field, although it is readily apparent in mid- height fields in the troposphere. The NAO's amplitude and phase vary over a range of time scales from intraseasonal to interdecadal; the NAO is present throughout the year but the largest amplitudes typically occur in winter. The NAO is often indexed by the standardized difference of December to February sea-level atmospheric pressure between Ponta Delgado, Azores (37.8° N, 25.5° W) or Lisbon, Portugal (38.8° N, 9.1° W) and Stkkisholmur, Iceland (65.18° N, 22.7° W). Statistical analysis reveals that the NAO is the dominant mode of variability of the surface atmospheric circulation in the Atlantic and accounts for more than 36% of the variance of the mean December to March sea-level pressure field over the region from 20° to 80° N and 90° W to 40° E, during 1899 through to 1994. Marked differences are observed between winters with high and low values of the NAO. Typically, when the index is high the Icelandic low is strong, which increases the influence of cold Arctic air masses on the northeastern seaboard of North America and enhances westerlies carrying warmer, moister air masses into western Europe (Hurrell, 1995). During high NAO winters, the westerlies directed onto northern Britain and southern Scandinavia are over 8 m s^{-1} stronger than during low NAO winters, with higher than normal pressures south of 55° N and a broad region of anomalously low pressure across the Arctic. In winter, western Europe has a negative radiation balance and mild temperature levels are maintained by the advection of warm air from the Atlantic. Thus strong westerlies are associated with anomalously warm winters, weak westerlies with anomalously cold winters, and NAO anomalies are related to downstream wintertime temperature and precipitation anomalies across Europe, Russia and Siberia.

Overlying the interannual variability there have been four main phases of the NAO index during the historical record: prior to the 1900s the index was close to zero; between 1900 and 1930, strong positive anomalies were evident; between the 1930s and 1960s, the index was low; and since the 1980s, the index has been strongly positive. The recent persistent high positive phase of the NAO index, extending from about 1973 to 1995, is the most persistent and high of the historical record. Preliminary reconstructions from tree ring data of a 1000-year-long record suggest that the recent extended period of high values in the NAO index series may not be unique. During each positive phase, higher than normal winter temperatures prevail over much of Europe, culminating in the unprecedented strongly positive NAO index values and mild winters of 1989 and 1990. During high NAO index winters, drier conditions also prevail over much of central and southern Europe and the Mediterranean, while enhanced rainfall occurs over the northwestern European seaboard (Hurrell, 1995). The recent high positive phase of the NAO index has resulted in an extended dry period in Morocco with as much as 35% reduction in runoff from major river catchments and serious reduction in cereal crop production.

Interior North America and Asia

The interiors of both continents are far removed from oceanic influences and because they are located in middle latitudes experience extreme continentality of climate. The main effect of extreme continentality is to produce large seasonal temperature variations. Thus Bergen (60° 24′ N, 5° 19′ E, 44 m) on the Atlantic coast of Norway has a mean annual temperature of 7.8° C and an annual range of 13.7° C, while at approximately the same latitude the central Asian city of Omsk (54° 56′ N, 73° 24′ E, 105 m) has a mean annual temperature of 0.4° C and a range of 38.4° C. Since the oceans are the main atmospheric moisture source, the continental interiors also tend to be dry, allowing the subtropical deserts to extend into temperate latitudes (see Fig. 1.11). In the USA the region between the Rockies and 100° W has annual precipitation means between 300 and 500 mm, whereas vast areas of central Asia receive less than 500 mm, but in both cases amounts increase towards the east coast. In North America, the mountain ranges along the west coast curtail the penetration of moist Pacific air-masses inland and have the same drying effect as the whole of lowland western Europe has on air-masses entering Asia. The interiors of both continents are open to invasions by air currents of Arctic origin. North America is also open to invasion by air currents of subtropical origin that contrast strongly with the cold northern air-masses, generating at times extremely active cold fronts, convective activity and tornadoes. The mid-western and western plains' states of the USA are visited by more severe tornadoes than any other area in the world; they are most frequent during spring and summer months. In contrast, moist tropical air is blocked from penetrating into central Asia by the Tibetan Plateau and its associated mountain ranges.

While the upper winter ridge over western North America directly influences the weather over the interior of that continent, a similar relationship does not exist between the ridge over western Europe and central Asia (Fig. 1.3). This is because a further winter trough–ridge pair of small amplitude are located between the main ridge over western Europe and the east Asian trough, with the secondary ridge located about 85° E. In January, almost the whole of Canada, the north interior and northeast have snow covers, similarly all Asia north of about 40° N is snow covered. In winter, the interiors of both continents are dominated by large anticyclones (Fig. 1.4a), the result of cold air ponding over the cold continental surfaces. The most intense is over Asia and is located on average to the east of the 85° E meridian, that is just downwind of the upper climatological ridge. The anticyclone over North America is a

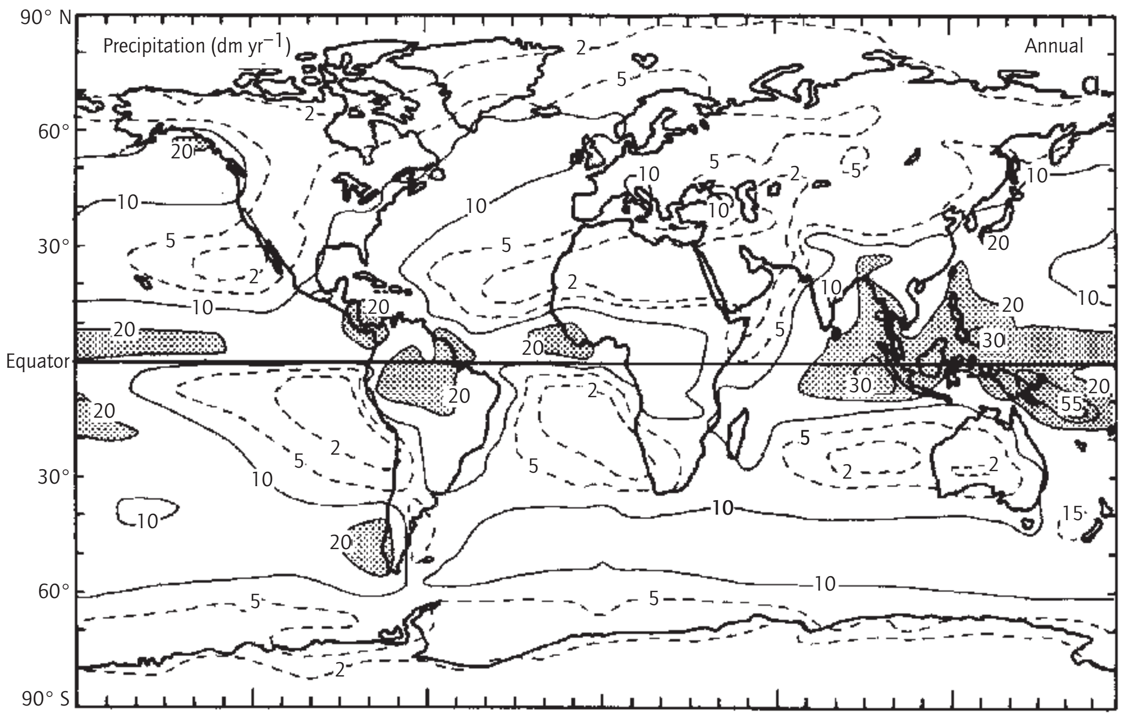


Fig. 1.11 Global distribution of precipitation in decimetres per year. (From Peixoto & Oort, 1992.)

weak and unstable feature which can be interpreted as a statistical average rather than as a **semi-permanent** and **quasi-stationary system**. Both systems act as the sources of continental polar air, which is intensely cold and dry. In summer, the continental anticyclones vanish and are replaced by shallow heat lows.

1.6 Tropical climates

The tropical world may be considered as bounded at the surface by the two belts of subtropical high pressure, at about 30° N and 30° S, and in the upper air by the corresponding subtropical westerly jet streams (Figs 1.3 and 1.4). Under this definition tropical climates are found in those areas dominated by the tropical Hadley cell circulations or tropical monsoon circulations.

The tropical atmosphere differs meteorologically from the middle latitude atmosphere in a number of important aspects (McGregor & Nieuwolt, 1998). In the tropics temperatures tend to be uniform in the horizontal over vast areas, so contrasting air currents are rare. In middle latitudes there are marked north–south temperature gradients, and air currents with differing origins therefore have contrasting temperatures and humidities. That is the middle latitude atmosphere is strongly **baroclinic** with large horizontal temperature gradients and therefore strong thermal winds (measure of wind shear with height). The strongly baroclinic nature of the middle latitude atmosphere explains the formation of frontal depressions and jet streams, with the latent heat released by the condensation of water vapour only exerting a modifying influence. In the tropics, where atmospheric moisture values are often considerable, rainfall and therefore condensation rates can be high resulting in the release of large amounts of latent heat, which can be of importance in the development of tropical circulation systems. Because the **Coriolis** parameter is small in the tropics, small pressure gradients can generate air currents as intense as those found in middle latitudes.

In the tropical world the Sun is nearly always almost overhead at midday, with little seasonal variation in day length from about 12 hours. Seasonal temperature variations in the humid tropics are therefore generally small, particularly near the Equator (Fig. 1.10). The annual temperature range (mean January minus mean July) is very small at 3° C or less over the oceans in the equatorial zone, and is only 2 or 3° C greater at 30° N and 30° S. While values of annual temperature range are equally small over the equatorial continents, values increase rapidly towards the subtropics to reach, for example, 20° C in the Sahara Desert and 15° C in central Australia. Since horizontal temperature gradients in the tropical atmosphere are small, heating of the lower atmosphere often causes convective overturning because it cannot be compensated by horizontal cold air advection from elsewhere. Therefore tropical circulation patterns are particularly influenced by heat inputs from such sources as warm ocean surfaces acting through latent heat released in deep cumulus convection. Other heat sources, such as high tropical plateaus and equatorial rain forest, are also important. These heat sources show marked latitudinal and longitudinal variations in their distributions, and also a marked tendency to vary on both annual and, in the case of the tropical oceans, on greater than annual scales. One consequence of this is that tropical rainfall patterns show both annual and greater-than-annual variations and also marked teleconnections with distant locations.

Mean winds in the tropical world reflect the mean pressure patterns (Fig. 1.4). In the centres of the subtropical anticyclones and in the equatorial trough, winds are normally light and variable. Between the equatorial trough and the subtropical anticyclones there is a region of easterly winds, with a small deflection towards the Equator, usually known as the **trade winds**. A seasonal reversal of wind direction takes place over southern Asia and the northern Indian Ocean; these are the Asian summer monsoon circulations.

Tropical storms

Atmospheric disturbances such as easterly waves and tropical storms (cyclones, hurricanes and typhoons) are frequent over the western parts of the subtropical oceans (see Table 1.2). Tropical storms are driven largely by latent heat released by condensing water vapour associated with intense

Table 1.2 Tropical storm* development regions

Area/location	**Average number of tropical storms per year**
Northeast Pacific	10
Northwest Pacific	22
South Pacific	7
Northwest Atlantic (including western Caribbean and Gulf of Mexico)	7
Bay of Bengal	6
South Indian Ocean	6
Arabian Sea	2
Off northwest Australian coast	2

*Tropical storms are defined by World Meteorological Organization as a warm-core vortex circulation with sustained maximum winds of at least 20 m s^{-1}.
Source: After Lockwood (1974).

rainfall. Rainfall rates near the centre of a tropical cyclone can exceed 500 mm per day, and this continuous inflow of water vapour is supplied by warm ocean surfaces with temperatures above about 26.5° C, which are normally found in the tropical oceans in the late summer and autumn. The storms decay rapidly when the supply of water vapour is cut off by the passage across cooler water surfaces or land areas. In summer and autumn they frequently bring high winds, storm surges and intense rainfall to the eastern coasts of the USA and Central America, subtropical southeast Asia, the Indian subcontinent and northern Australia.

Walker circulations, the Southern Oscillation and El Niño

Satellite imagery and rainfall data clearly show (Fig. 1.11) three equatorial regions of maximum cloudiness and rainfall: the so-called 'Maritime Continent' of the Indonesian Archipelago, the Amazon river basin in South America and the Zaire river basin in Africa. The rest of the equatorial region is comparatively dry, and some, such as the coasts of Peru, even desert. These longitudinal variations in rainfall are associated with east–west regional circulations along the Equator, the most important being the **Walker circulation**, which involves rising air motion over the Indonesian Archipelago and sinking over the eastern Pacific (Fig. 1.12). The rising air motion takes place mostly in deep convective clouds and is associated with intense convective rainfall and therefore the wet humid climates of the Indonesian Archipelago. The subsiding air suppresses cloud formation and rainfall, therefore it is associated with the coastal deserts of Peru. The Hadley cell circulation refers to the north–south component of these circulations: equatorward motion at low levels, rising in the convective regions near the Equator and poleward flow aloft. The Walker circulation refers to the east–west component, which is particularly prominent in the equatorial plane. Both circulations are driven by the release of latent heat in deep convective shower clouds.

The Walker circulation is closely coupled with the sea-surface temperature distribution over the Pacific, with relatively cool water in the east and warm in the west. When the Pacific Ocean off the coast of South America is particularly cold, the air above is too stable to take part in the ascending motion of the Hadley cell circulation. Instead, the equatorial air flows westward between the Hadley cell circulations of the two hemispheres to the warm West Pacific where, having been heated and supplied with moisture from the warmer waters, the equatorial air can take part in large-scale ascent (Fig. 1.12). The easterly winds that blow along the Equator and the northeasterly winds that blow along the Peru and Ecuador coasts both tend to drag the surface water along with them. The Earth's rotation then deflects the resulting surface currents toward the right (northward) in the Northern Hemisphere and to the left (southward) in the Southern Hemisphere. The surface waters are therefore deflected away from the Equator in both hemispheres and also away from the coastline. Where the surface water moves away under the influence of the trade winds, colder, nutrient-rich water upwells from below to replace it. Since the newly upwelled water is colder than its surroundings, its signature in infrared satellite images takes the form of a distinctive 'cold tongue' extending westward along the Equator from the South American coast. The winds that blow along the Equator also affect the ocean **thermocline**, the

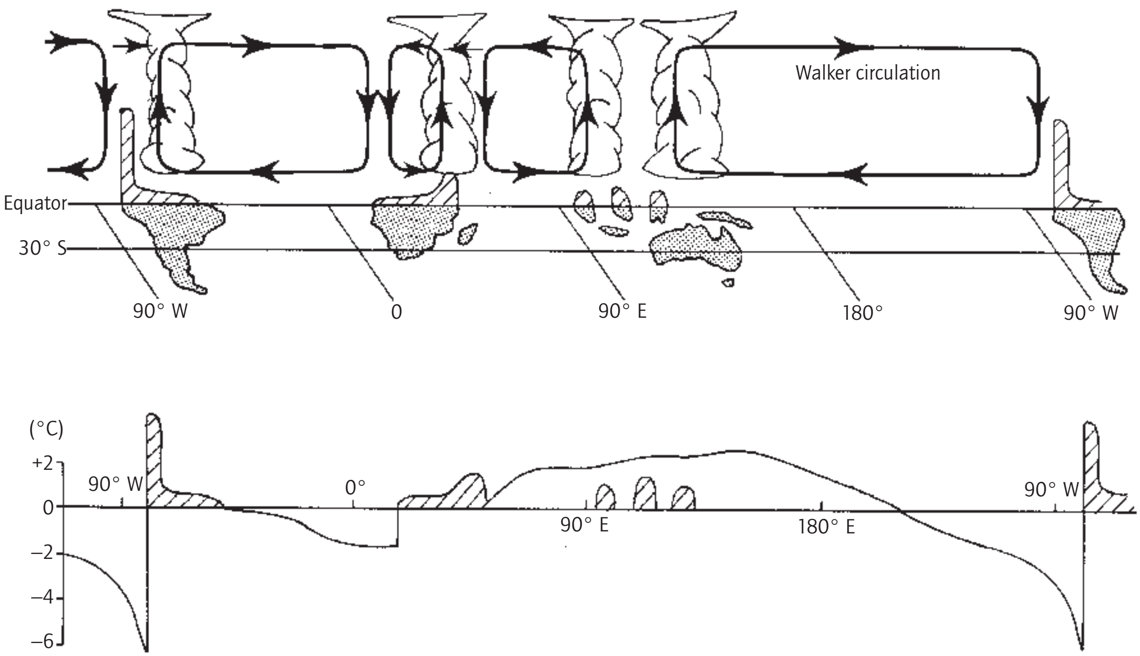


Fig. 1.12 Schematic representation of the Walker circulation along the Equator during non-ENSO (El Niño–Southern Oscillation) conditions. The sea-surface temperature departures from the zonal-mean along the Equator are shown in the lower part of the figure. (From Peixoto & Oort, 1992; Wyrtki, 1982.)

boundary between the warm surface water and the deep cold water. In the absence of the wind the thermocline would be nearly horizontal; but the trade winds drag the surface water westward, raising the thermocline nearly to the ocean surface in the east and depressing it in the west. The situation in the equatorial Atlantic is analogous to the equatorial Pacific in that the warmest part is in the west, at the coast of Brazil, but west–east contrasts of water temperature are much smaller than in the Pacific. However. in January a thermally driven Walker circulation may operate from the Gulf of Guinea to the Andes, with the axis of the circulation near the mouth of the Amazon.

Associated with the Walker circulation is the **El Niño phenomenon**, when every few years the tropical Pacific Ocean off the coasts of Peru and Ecuador occasionally becomes much warmer than average for periods of several months. Under El Niño conditions the Walker circulations become reversed, resulting in heavy rain in the normally arid areas of Peru, and drought in the western Pacific. The Southern Oscillation is dominated by an exchange of air between the Southeast Pacific subtropical high and the Indonesian equatorial low, with a period that varies between roughly 1 and 5 years. During one phase of this oscillation, the trade winds are intense and converge into the warm western tropical Pacific, where rainfall is plentiful and sea-level pressures low. At such times the atmosphere over the eastern tropical Pacific is cold and dry. During the complementary phase, the trade winds relax, the zone of warm surface-waters and heavy precipitation shifts eastwards, and sea-level pressure rises in the west while it falls in the east. The latter phase is the more unusual and in the eastern Pacific has become known as El Niño, whereas vigorous episodes of the former are often termed La Niña. The combined atmospheric–oceanic conditions that give rise to these changes in rainfall across the Pacific and neighbouring areas are referred to as El Niño–Southern Oscillation (ENSO) events. The ENSO is important climatologically for two main reasons. First, it is one of the most striking examples of interannual climatic variability on an almost global scale. Second, in the Pacific it is associated with considerable fluctuations in rainfall and sea-surface temperature, and also with extreme weather events around the world (Glantz, 1996).

The Southern Oscillation may be defined in terms of the difference in sea-level pressure between Darwin, Australia and Tahiti. Records are available, with the exception of a few years and occasional months, from the late 1800s. The El Niño phenomenon is associated with extreme negative Southern Oscillation values, but for much of the time the series exhibits continuous transitions from high to low values, with most values being positive. Of great importance to the development of the Southern Oscillation is the difference in the ways the ocean and atmosphere respond to changes in the winds and sea-surface temperature patterns, respectively. During La Niña, intense trade winds drive warm surface waters of the equatorial Pacific westward while exposing cold water at the surface in the east. When the winds relax during El Niño, the warm waters move back eastward, overflowing the cold water. The response of the ocean to changing winds involves not only changes in currents but also the excitation of waves that travel back and forth across the Pacific. These waves have a large signature, not at the surface, but at the thermocline, which separates the warm waters of the upper ocean from the colder water at depth. These waves eventually bring the ocean to a new equilibrium after a change in the winds, and the time it takes them to propagate across the Pacific is important in determining the time scale of the Southern Oscillation.

The monsoon circulations of Southern Asia and Eastern Africa

The characteristics of the monsoon climate are to be found mainly in the Indian subcontinent, where over much of the region the annual changes may conveniently be divided into the northeast (dry) and southwest (wet) seasons (Pant & Rupa Kumar, 1997). Over many tropical oceans, the atmospheric circulation undergoes very little seasonal variation. In contrast, over tropical continents the atmospheric circulation displays distinct seasonal rhythms and variations. Over the oceans, evaporation consumes a high proportion of the incident radiation, which is also rapidly absorbed by the water surface and spread over great depths by waves and turbulence, and dissipated to other latitudes by ocean currents. Over the tropical continents, particularly if they are dry because of low rainfall, solar radiation is used mostly to warm the Earth's surface, so that surface temperatures in these regions reach very high values owing to the much lower heat capacity of the soil compared with water. This warming over the tropical continents in early summer produces thermal lows which gradually take on some functions of the equatorial trough, thus forming a new region of convergence. The trade winds from the winter hemisphere cross the Equator, slow down and create a minor area of convergence owing to their change in direction, caused by the reversal of the Coriolis force. In winter, the tropical continents experience relatively low temperatures and high surface pressure; winds are re-established flowing towards the Equator, which are similar to the trade winds. Thus the tropical continents and adjacent oceans can experience a semi-annual reversal in wind direction characteristic of monsoons.

The southwest monsoon current in the lower 5000 m near India consists of two main branches: the Bay of Bengal branch, influencing the weather over the northeast part of India and Burma; and the Arabian Sea branch, dominating the weather over the west, central and northwest parts of India. The low-level flow across the Equator during the southwest monsoon is not evenly distributed between the longitudes 40° E and 80° E, but flows intermittently during the southwest monsoon from the vicinity of Mauritius through Madagascar, Kenya, eastern Ethiopia, Somalia, and then across the Indian Ocean toward India. A particularly important feature of this flow is the strong southerly current, with a mean wind speed of about 14 m s^{-1} observed at the Equator over eastern Africa from April to October. The strongest flow occurs near the 1000–5000 m level, but it often increases to more than 25 m s^{-1} and occasionally to more than 45 m s^{-1} at heights between 1000–2000 and 2000–5000 m.

During the northern winter season the subtropical westerly jet stream crosses southern Asia, with its core located at about 12 000 m altitude. It divides in the region of the Tibetan Plateau, with one branch flowing to the north of the plateau and the other to the south. The two branches merge to the east of the plateau and form an immense upper convergence zone over China. In May and June the subtropical jet stream over northern India slowly weakens and disintegrates, causing the main westerly flow to move north into central Asia. While this is occurring, an easterly jet stream, mainly at about 14 000 m, builds up over the equatorial Indian Ocean and expands westward into Africa. The formation of the equatorial easterly jet stream is connected with the formation of an upper-level high-pressure system over Tibet. In October the reverse process occurs; the equatorial easterly jet stream and the Tibetan high disintegrate, while the subtropical westerly jet stream reforms over northern India. The Himalayan–Tibetan plateau is of importance because it appears to accelerate the onset of the Asian monsoon and to increase its ultimate intensity. Central and southeastern parts of Tibet remain free of snow throughout most of the year, hence, the plateau must heat rapidly during the northern spring. This direct warming of the middle troposphere creates an upper-level anticyclone, which is readily observed on synoptic charts, with upper-level divergence and low-level convergence. Thus, suitable conditions are produced for the Asian monsoon in the northern spring. Latent heat released in intense tropical storms over India keeps the system functioning during the northern summer. Since a complex feedback system produces the southwest monsoon, failures in the system are common and produce extensive breaks in the monsoon rains when the whole system shows signs of collapse. Variations in winter snow cover over Tibet will influence the start and intensity of the southwest monsoon. General cooling over Southern Asia at the end of the northern summer causes its collapse.

Nearly half the Indian subcontinent is arid or semi-arid (Fig. 1.11), and by far the larger part of this dry area is located in northwest India and southwestern Pakistan. The Arabian Sea summer monsoon circulation is linked with a heat low over Arabia, Pakistan and northwest India. This heat low, which develops during May, establishes the low-level westerly monsoon wind regime a full month before heavy monsoon rains start over western India. In mid-July the heat low is deepest, the southwest monsoon is strongest, and the west Indian rains are heaviest. The question then arises as to why the heat low remains cloud-free while rain falls to the south. The unique summer aridity of the desert belt from the western Sahara to Pakistan is strongly correlated with the forced descending motion on the northern side of the exit region of the equatorial easterly jet stream. During the northern summer, the equatorial easterly jet stream extends at about 14 000 m in the latitude belt 5–20° N from the Philippines across southern Asia and northern Africa to almost the western Atlantic. Over the whole exit region from India westwards the very gradual deceleration of the jet stream core results in widespread sinking motions on the northern side and rising motions on the southern. Since the Hadley cell circulations already generate desert conditions over North Africa and the Middle East, the deceleration of the equatorial easterly jet stream intensifies the aridity of these deserts. Similarly, numerical simulations suggest that large areas of northeast Africa would have considerably more rainfall if the Tibetan Plateau and the equatorial easterly jet stream were absent. The reverse flow is observed in the entrance region over Southeast Asia, where air sinks to the south and rises to the north, with a rainy area over southern Asia. Radiosonde ascents at Karachi and Jodhpur indicate the frequent presence of a low-level inversion, with moist air originating over the Arabian Sea below the inversion. Subsidence limits the height to which the surface air from the Arabian Sea can ascend, restricting cloud development, and thus favouring strong solar heating.

Australia

The chief determinant of the climate and wind field over Australia is the subtropical anticyclonic belt (Sturman & Tapper, 1996). In general the continent is affected by mid-latitude westerlies on the southern fringe, tropical convergence on the northern fringe in summer, and stable subsiding air under the subtropical anticyclones over the interior. The result is that most of the continent is covered by arid or semi-arid climates. The average altitude of Australia is only 300 m, with 87% of the continent less than 500 m and 99.5% less than 1000 m. In general the low relief of Australia does not significantly obstruct the movement of the atmospheric systems that control the climate.

Most of Australia is warm to hot (Fig. 1.10), with the exception of the alpine area in the southeast where there is seasonal snow. The month with the highest temperature varies from November in the far north to February in the south. In the north the build-up of monsoon clouds cools the latter part of the summer, while in the south the time taken to warm the ocean delays the peak temperature until late summer. July is the coldest month throughout the country. Australia is the driest continent, excluding Antarctica, and no continent has less runoff from its rivers. More than a third of the country receives on average less than 250 mm of rainfall annually, and only 9% receives more than 1000 mm. Most of the area south of 35° S receives mainly winter rains, while north of 25° S; most rain falls in summer, associated with the summer monsoon and tropical cyclones. Australian rainfall averages disguise an extremely variable rainfall, with droughts and flooding being very common.

Australian rainfall is more variable than could be expected from similar climates elsewhere in the world, mainly due to the impact of ENSO. Australian rainfall fluctuations, as well as being more severe because of ENSO's influence, also operate on very large spatial scale. High rainfall totals in Australia occur when the Southern Oscillation Index (SOI) is large and positive (La Niña events). In contrast when the SOI is strongly negative (El Niño years) drought occurs over much of the continent. Thus the continental scale of the 1982/83 drought is typical of many years, although it was more severe than most. Extended periods of drought or extensive rains in Australia do not occur randomly in time, in relation to the annual cycle. The ENSO phenomenon, and Australian rainfall fluctuations associated with it, are phase-locked with the annual cycle. Thus the heavy rainfall of an anti-ENSO event tends to start early in the calendar year and finish early in the following year. The dry periods associated with ENSO events tend to occupy a similar time period. For example, the 1982/83 drought started about April 1982 and broke over much of the country in March and April 1983.

Rainfall is not the only aspect of Australian climate affected by ENSO. Frosts tend to be more common in inland Australia during ENSO events, because low rainfall is associated with decreased cloud cover, allowing increased radiative cooling at night. The decreased cloud cover also causes higher maximum temperatures during ENSO events. It is also reported that north of about 25° S low-level winds in winter during ENSO events tend to be about three to four times stronger than during anti-ENSO events. In the southeast the winds in summer tend to be two to three times stronger in ENSO events. The higher maximum temperatures and stronger winds associated with ENSO related droughts increase the likelihood of wildfires. The latitudinal variations in the season in which the ENSO-amplified winds occur (winter in the north, summer in the south) means that the stronger winds occur at the time of year when fires are most likely.

South America

Of particular climatological significance in South America are the Andes Mountains along the west coast and the tapered shape of the continent, with much of its area lying in tropical latitudes. The latter means that the oceans are of importance in the climates over large areas of the continent.

Three circulation regimes dominate the climate of the continent (Cerveny, 1998):

1. the prevailing westerly winds of the extreme southern continental latitudes;
2. the semi-permanent subtropical high-pressure cells positioned over the South Atlantic and South Pacific oceans;
3. the location of the intertropical convergence zone (ITCZ), a migrating band of maximum convergence, convection cloudiness and rainfall.

The movements of the ITCZ strongly influence the climates of tropical South America. The ITCZ reaches its northernmost location in June and September/October, causing the season of greatest rainfall for northern South America and the Caribbean. At the height of the northern winter, the ITCZ extents southward into the central Amazon Basin. Consequently, the months of January and February mark the dry season for much of northern South America and the tropical Atlantic, while the Amazon Basin receives much of its annual rainfall in this season. In general, the onset of the rainy season occurs first in southeast Amazonia, with onset dates occurring progressively later toward the northwest. The demise of the rainy season occurs first in the southeast and progressively later to the northwest. Associations with the extremes of the Southern Oscillation (SO) index are most marked in central Amazonia and near the mouth of the Amazon. Years with the low/high SO phase are consistent with anomalously dry/wet rainy seasons in these two regions, due mainly to a late/early onset of the rainy season. During the El Niño, the near-surface Atlantic trade winds are weaker, consistent with an anomalously northward displaced ITCZ, thus the moisture input from the Atlantic into the Amazon basin is weak, the moisture convergence and convection is also weak over Amazonia, and lower rainfall is observed especially in the central and mouth of Amazon River regions. The sea-surface temperature dipole in the tropical Atlantic indicates anomalously warmer surface water to the north of the Equator. During La Niña, these patterns are reversed, with a southward displacement of the ITCZ, strong Atlantic trades winds and moisture transport into Amazonia, and an inverted sea-surface temperature dipole (anomalously warmer surface water south of the Equator). Over the western part of the basin there is a descending branch of the Walker cell during El Niño, with subsidence affecting a zonal band from the Andes to the Atlantic. There is also strong upward flow, convection and rainfall over northern Peru to the west of the Andes, associated with anomalously warm Pacific sea-surface temperatures, implying compensatory subsidence and reduced rainfall over western Amazonia. This flow pattern is inverted during La Niña. The upper level Bolivian tropospheric high is weaker during El Niño, with the subtropical westerly jet stronger and located anomalously northward of its average position, whereas during La Niña the jet is weaker and anomalously southward of its average position.

Northeast Brazil is an exceptional area in which rainfall diminishes from in excess of 1000 mm along the coast to less than 400 mm inland (Fig. 1.11). The relative dryness of northeast Brazil is caused primarily by the flow patterns of the general atmospheric circulation, particularly by thermally driven circulations of the Hadley–Walker types. To the immediate west of northeast-Brazil is the Amazon basin, where the high rainfall is associated with vigorous upward convection in cumulonimbus clouds. The water vapour condensing in these clouds releases latent heat, which warms the air and thereby maintains the ascending motion. The air rising over Amazonia descends in a Walker type circulation over the eastern subtropical Atlantic Ocean, including northeast Brazil and possibly the western coast of Africa. The circulation patterns are reflected in the rainfall and cloud distributions. Similarly, the north–south circulation patterns of the Hadley cells, with ascending motion in the convective clouds of the ITCZ and descending motion over the subtropical Atlantic of both hemispheres, also contributes to the aridity of northeast Brazil. The interannual variability, both in strength and geographical position, of the Hadley–Walker circulations results in the high interannual variability of precipitation in northeast Brazil.

In oceanic regions the ITCZ generally lies over or near the highest sea-surface temperatures. Therefore a relationship should be expected to exist between the general sea-surface temperature distribution in the tropical Atlantic and the rainfall over northeast Brazil. Warmer (colder) sea-surface temperatures in the tropical South Atlantic and colder (warmer) ones in the tropical North Atlantic are associated with wet (dry) years in northeast Brazil. Thus in wet years (e.g. 1964, 1967, 1984 and 1985) the sea-surface temperature anomalies were positive in the tropical South Atlantic and negative or near zero in the tropical North Atlantic. In contrast, in dry years (e.g. 1951, 1953, 1958, 1970 and 1979) the sea-surface anomalies were negative in the tropical South Atlantic and positive in the tropical North Atlantic.

Africa

Of all the continents, Africa is the most symmetrically located with regard to the Equator, and this is reflected in the climatic zonation. The continent may be regarded as a giant plateau for there is a relative absence of very pronounced topography, although some high mountains exist, especially in the East African region. The latitudinal position of Africa means that the continent is influenced by tropical, subtropical and mid-latitude pressure and wind systems. As the near-equatorial trough migrates with the seasons, the ITCZ lies south of the Equator in the northern winter and north of the Equator in the summer. Connections between the equatorial easterly jet stream and North African rainfall were discussed in the section on monsoons.

Of particular interest in Africa is the Subsaharan region (Sahel) because it shows a dramatic decrease in rainfall since the late 1960s and continued severe drought conditions from the early 1970s up to at least the late 1980s. As in the case of northeast Brazil, this is probably connected with regional circulation changes and North Atlantic sea-surface temperatures. Indeed, colder than average tropical North Atlantic sea-surface temperatures (SST) appear to be strongly associated with drought in the African Sahel (Folland *et al.* 1986; Hastenrath, 1995).

References

Brook, E.J. (2005) Tiny bubbles tell all. *Science*, **310**: 1285–1287.

Cerveny, R.S. (1998) Present climates of South America. In *Climates of the Southern Continents* (Eds J.E. Hobbs, J.A. Lindesay and H.A. Bridgman), pp. 107–135. Wiley, Chichester.

DeConto, R.M. and Pollard, D. (2003) Rapid Cenozoic glaciation of Antarctica induced by declining atmospheric CO_2. *Nature*, **421**: 245–249.

EPICA Community Members. (2004) Eight glacial cycles from an Antarctic ice core. *Nature*, **429**: 623–628.

Folland, C.K., Palmer, T.N. and Parker, D.E. (1986) Sahel rainfall and worldwide sea temperatures 1901–85. *Nature*, **320**: 602–607.

Glantz, M.H. (1996) *Currents of Change: El Niño's Impact on Climate and Society.* Cambridge University Press, Cambridge.

Hastenrath, S. (1995) Recent advances in tropical climate prediction. *Journal of Climate*, **8**: 1519–1532.

Henderson-Sellers, A. and Robinson, P.J. (1986) *Contemporary Climatology*. Longman, Harlow, Essex.

Houghton, J. (2005) Global warming. *Reports on Progress in Physics*, **68**: 1343–1403.

Hurrell, J.W. (1995) Decadal trends in the North Atlantic Oscillation: regional temperature and precipitation. *Science*, **269**: 676–679.

Jones, P.D. and Moberg, A. (2003) Hemispheric and large scale surface air temperature variations: an extensive revision and an update to 2001. *Journal of Climate*, **16**: 206–23.

King, J.C. and Turner, J. (1997) *Antarctic Meteorology and Climatology.* Cambridge University Press, Cambridge.

Lockwood, J.G. (1974) *World Climatology: an Environmental Approach.* Edward Arnold, London.

Lockwood, J.G. (2001) Abrupt and sudden climatic transitions and fluctuations: a review. *International Journal of Climatology*, **21**: 1153–1179.

Lockwood, J.G. (2003) The Earth's climates. In *Handbook of Atmospheric Science: Principles and Applications* (Eds C.N. Hewitt and A.V. Jackson), pp. 59–89. Blackwell, Oxford.

Mann, M.E., Bradley, R.S. and Hughes, M.K. (1999) Northern Hemisphere temperatures during the past millennium: inferences, uncertainties and limitations. *Geophysical Research Letters*, **27**: 1519–1522.

McGregor, G.R. and Nieuwolt, S. (1998) *Tropical Climatology: an Introduction to the Climates of the Low Latitudes.* Wiley, Chichester.

National Climatic Data Center (1987) *Monthly Climatic Data for the World,* Vol. 40, Nos 1 and 7. National Oceanic and Atmospheric Administration, Asheville, NC.

Oort, A.H. (1983) *Global Atmospheric Circulation Statistics, 1958–1973*. National Oceanic and Atmospheric Administration, Washington, DC.

Orvig, S. (Ed.) (1970) *Climate of the Polar Regions*, Vol. 14, *World Survey of Climatology.* Elsevier, Amsterdam.

Pant, G.B. and Rupa Kumar, K. (1997) *Climates of South Asia.* Wiley, Chichester.

Palmen, E. (1951) The role of atmospheric disturbances in the general circulation. *Quarterly Journal Royal Meteorological Society*, **77**: 337–54.

Peixoto, J.P. and Oort, A.H. (1992) *Physics of Climate*. American Institute of Physics, New York.

Qiang Fu, Johanson, C.M., Wallace, J.M. and Reicher, T. (2006) Enhanced mid-latitude tropospheric warming in satellite measurements. *Science*, **312**: 1179

Shackleton, N.J. (2000) The 100 000-year ice-age cycle identified and found to lag temperature, carbon

dioxide, and orbital eccentricity. *Science*, **289**: 1897–1902.
Schwerdtfeger, W. (1984) *Weather and Climate of the Antarctic.* Elsevier, Amsterdam.
Serreze, M.C. and Barry, R.G. (2005) *The Arctic Climate System.* Cambridge, Cambridge University Press.
Simmons, I. (1998) The climate of the Antarctic Region. In *Climates of the Southern Continents: Present, Past and Future* (Eds J.E. Hobbs, J.A. Lindesay and H.A. Bridgman), pp. 137–160. Wiley, Chichester,
Stott, P.A., Tett, S.F.B., Jones, G.S., Allen, M.R., Mitchell, J.F.B. and Jenkins, G.J. (2000) External control of twentieth century temperature variations by natural and anthropogenic forcings. *Science*, **15**: 2133–2137.
Sturman, A.P. and Tapper, N.J. (1996) *The Weather and Climate of Australia and New Zealand.* Oxford University Press, Melbourne.
Wyrtki, K. (1982) The Southern Oscillation, ocean–atmosphere interaction, and El Niño. *Marine Technology Society Journal*, **16**: 3–10.

Further reading

Houghton, J.T. (2004) *Global Warming: the Complete Briefing* (3rd edition). Cambridge University Press, Cambridge.
IPCC. (2001) *Climate Change 2001: The Scientific Basis. Contribution of Working Group 1 to the Third Assessment Report of the Intergovernmental Panel on Climate Change* (Eds J.T. Houghton, Y. Ding, D.J. Griggs, *et al.*). Cambridge University Press, Cambridge.
IPCC. (2007) *Climate Change 2007: The Physical Science Basis. Contribution of Working Group 1 to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* (Eds S. Solomon, D. Qin, M. Manning, *et al.*). Cambridge University Press, Cambridge.
IPCC. (2007) *Climate Change 2007: Climate Change Impacts, Adaptation and Vulnerability. Contribution of Working Group 2 to the Fourth Assessment Report of the Intergovernmental Panel on Climate Change* (Eds N. Adger, P. Aggarwal, S. Agrawala, *et al.*). Cambridge University Press, Cambridge.
Munn, T. (Ed.-in-Chief) (2002) *Encyclopedia of Global Environmental Change* (5 vols). John Wiley, Chichester.
Oliver, J.E. (Ed.) (2005) *Encyclopedia of World Climatology.* Springer-Verlag, Dordrecht.

Questions

1 Sulphate aerosols originating from industrial areas cool the atmosphere but cause both ill health and acid rain. Attempts to remove such aerosols from the atmosphere will accelerate global warming. What therefore should be the policy on industrial pollution?
2 Summarize the main changes in the climate of your home area over the past 100 000 years.
3 How does ENSO affect local climate and human activity across the world?
4 How do major mountain ranges and plateaus affect large-scale climate?
5 What is meant by a ‘proxy’ indicator of climate? Give examples of how these can be used to reconstruct records of past climate.