S279 Our dynamic planet: Earth and life
Science: Level 2

The Open University

Book 2 Part 1
Earth and Life

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Cover image: Folded layers of a banded iron formation (BIF or banded ironstone). Banded ironstone is a rock formed from alternating layers of the iron oxide mineral hematite (red) and tiger eye and jasper. At greater than 3 billion years of age these are among the oldest rocks on Earth. They were laid down in shallow seas where primitive bacteria caused iron to be oxidised and precipitated. Subsequent deep burial of the rocks, and tectonic movements, have caused the rock to alter and deform. This small sample is around 2 cm across. The rock has been cut open to expose the banding (strata), and also polished for display. Many BIFs come from Western Australia. (Dirk Wiersma/Science Photo Library)

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Chapter 1

A habitable planet

No one knows whether life flourishes on planets orbiting stars more distant than our own Sun. Despite the excitement in 1996 about possible fossils in Martian meteorites, it is still not known whether life exists, or has ever existed, on Mars. It is known, however, that within the Solar System, the profusion of life in the forms that are recognised and understood on Earth could not be supported on any of the other planets. Why is this? In this introductory chapter you will explore some of the conditions that make the Earth habitable to life.

1.1 How is the Earth different from other planets?

If you could observe the Solar System from far off, you would see the Sun and its orbiting planets, all virtually in the same plane. This is a legacy of their common origin from the same spinning nebular disc. If you were to look more closely at the four planets nearest to the Sun, you would see a small planet, Mercury, then a very bright one, Venus, then our own blue Earth with its swirls of white cloud, and finally the reddish globe of Mars. These four planets have very different surface environments, yet formed in more or less the same part of the solar nebula, and so are likely to be made up of the same elements in very similar proportions. Looking closer still at the blue planet, it has a thin envelope of hazy atmosphere. Through this, below the clouds, not only the blue of oceans and seas and the bright white of the ice caps can be seen, but also the greens, greys and browns of land. Even the night hemisphere is illuminated by billions of tiny light sources grouped into cities, by flashes of bluish lightning, and by the occasional red glow of fire or erupting magma.

The habitability of this mysterious and beautiful environment is the result of the complex interplay of many processes – physical, chemical, biological and geological – acting over a vast range of temporal and spatial scales. Looked at simply, however, the Earth is hospitable principally as a result of its position in relation to the Sun. Earth is half as far again from the Sun as Venus, where the average surface temperature is 460 °C. The Earth is about two-thirds of the distance of Mars, where the average surface temperature is –50 °C. By contrast, the average temperature at the surface of the Earth is a moderate 15 °C. Averages can, of course, conceal enormous ranges. On the Earth, at the present time, surface temperatures rarely rise above 50 °C, and rarely fall below –50 °C, although there are geographical variations, such as with latitude. However, some microorganisms on Earth can grow at temperature extremes much greater than even these (see Box 1.1). Most life on Earth, particularly multicellular life, is confined to a much narrower range. The relatively small temperature range of most of the Earth's surface is a result of the form and content of the Earth's atmosphere and ocean. As you will see, the full story is very complicated (and by no means completely understood) but, in terms of regulating the Earth's surface temperature, the two most important atmospheric constituents are carbon dioxide and water.

Box 1.1 The diversity of microbial life in Earth's environments

The habitability of the Earth can be defined by the boundaries of environmental conditions in which life can grow. Many environments, once thought to be too extreme for life, are now known to support microbial life.

1. Currently, the highest temperature limit for life is set by the microorganism known as Strain 121, which was isolated from hot fluid emanating from a hydrothermal vent on the floor of the northeast Pacific Ocean. The previous record holder was *Pyrolobus fumarii*, which can grow at 114 °C. Strain 121 can grow at a temperature of 121 °C (the temperature used for sterilising objects). It could survive at 130 °C, but it needs to be at the lower temperature of 121 °C to grow. Although organisms with higher growth temperatures might eventually be found, it is likely that, at much higher temperatures, the energy imparted to molecules makes the energetic cost of repairing or synthesising them prohibitive to growth.

   Many of the microorganisms that can grow at high temperatures (known as hyperthermophiles) belong to the domain **Archaea**. Some of these organisms have been recovered from the deep subsurface, for example at 3 km depth in African mines, showing that life is not limited to the surface of the Earth and oceans. It has been postulated that the characteristics of many of these groups of microorganisms reflect the conditions on the early Earth, when both volcanism and asteroid and comet impacts, and thus hot environments, were more common.

2. At the other end of the scale, frozen permafrost can provide an environment for metabolising bacteria. At temperatures below 0 °C, saline liquid water can exist because salt depresses the freezing point of water and at temperatures of –10 °C, bacteria that can metabolise and grow, albeit slowly, have been recovered from Siberian permafrost.

3. Microorganisms have also been found growing at extremes of pH. *Ferroplasma acidarmanus* can grow at a pH of 0, living in acidic waters at Iron Mountain in California. Similarly, organisms have been found growing at high pH in alkaline soda lakes.

4. Some microorganisms, such as *Deinococcus radiodurans*, also have high radiation resistance, using highly efficient DNA repair processes to reverse radiation damage caused to their DNA. This organism can tolerate a radiation dose at least three orders of magnitude higher than that which is lethal to a human.

These **extremophiles** help define the envelope of life on Earth, and thus the extremes beyond which the planet becomes uninhabitable.

On the Earth, surface temperatures are such that most of the planet's surface water is in liquid form, with the remainder in the ice caps and in the atmosphere. The atmosphere contains a small amount of CO_2, the oceans a good deal more, and large amounts are effectively 'locked up' in crustal rocks. By contrast, on Venus the atmosphere is largely CO_2, there is a minute amount of atmospheric water vapour (i.e. H_2O gas) and, at the prevailing temperatures, none of it can condense on the planet's surface. The Martian atmosphere, like the Venusian one, largely consists of CO_2, but some CO_2 and most of the planet's water is in the form of ice. Were it not for the presence of liquid water on the surface of the Earth, life – at least in familiar forms – could not exist here.

Look at Figure 1.1. The cloud cover, which obscures much of the Earth's surface, has been removed. Immediately obvious are bright areas of ice cover, the enormous area of ocean in the Southern Hemisphere, and the green of forests and crops.

- To what extent does this map tell you about life on Earth?
- It indicates where life is concentrated because much of the biologically productive life on Earth, and on land in particular, is based on plants and other **phototrophs** – organisms that can build their own organic material by harnessing the energy of sunlight by **photosynthesis**.

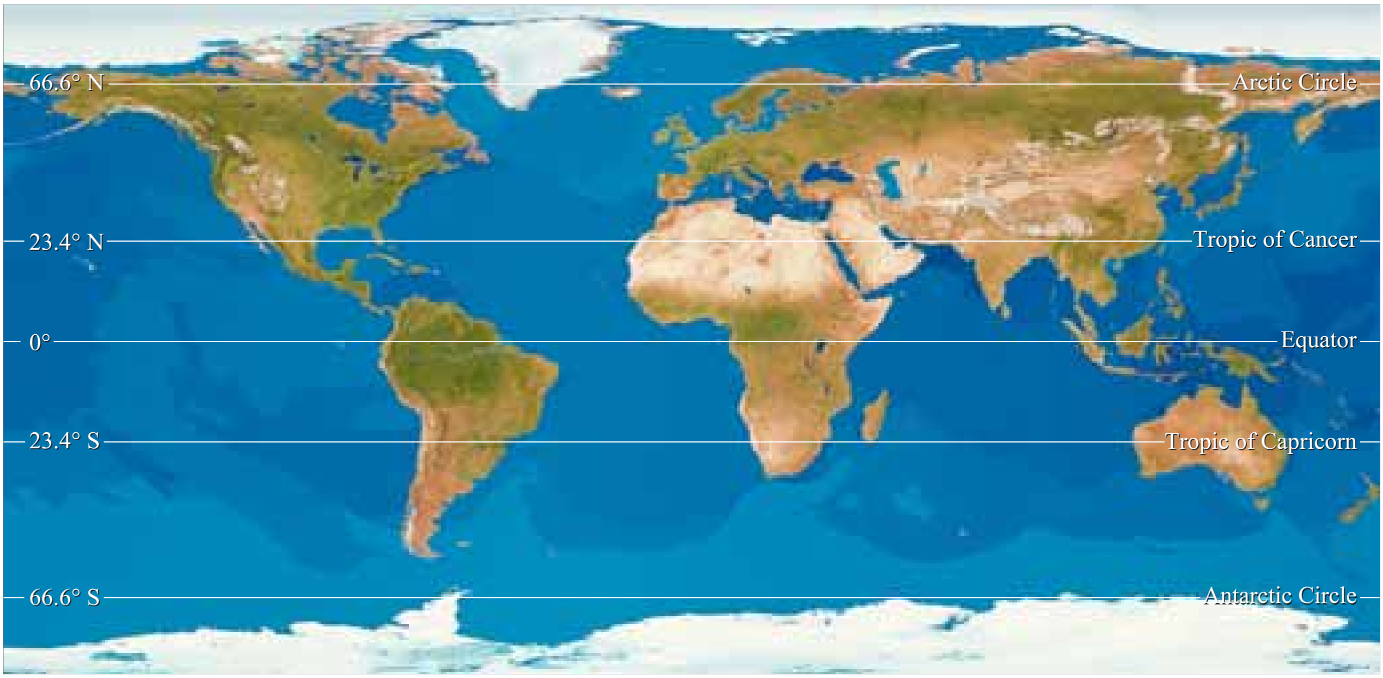


Figure 1.1 Composite satellite image of the Earth's surface. The green areas (e.g. Central Brazil and Western Europe) are forests and crops; the light-coloured areas in the tropics are mostly savannah, semi-arid scrub and desert; and the brown areas at high northern latitudes are tundra. At high latitudes, and at lower latitudes where there are mountains (e.g. the Rockies, Andes and Himalaya), white corresponds to ice and snow. *Note*: the area of ice at high latitudes is grossly exaggerated by this type of projection.

Expressed simply, the chemical equation for photosynthesis can be written as follows:

$$\underset{\substack{\text{carbon dioxide}\\ \text{from atmosphere}}}{nCO_2(g)} + nH_2O \xrightarrow{\text{light energy}} \underset{\substack{\text{organic}\\ \text{matter}}}{(CH_2O)_n} + \underset{\text{oxygen}}{nO_2(g)} \qquad (1.1)$$

where n can have various values, $(CH_2O)_n$ represents a range of carbohydrate materials of which glucose ($C_6H_{12}O_6$) is the simplest, and (g) is gas. We could rewrite Equation 1.1, for instance, with glucose, $C_6H_{12}O_6$, representing organic matter in the form:

$$\underset{\substack{\text{carbon dioxide}\\ \text{from atmosphere}}}{6CO_2(g)} + 6H_2O \longrightarrow \underset{\substack{\text{organic}\\ \text{matter}}}{C_6H_{12}O_6} + \underset{\text{oxygen}}{6O_2(g)} \qquad (1.2)$$

This process of taking free carbon from the atmosphere and combining it into living organic material is referred to as '**fixing**' carbon, and the process of building living material by fixing carbon is known as **primary production**. You will look at this in much more detail in Chapter 3. Animals cannot fix carbon and so can live only by consuming **primary producers**, either directly or indirectly.

But Figure 1.1 tells only a small part of the story; for one thing, the oceans like the land also support abundant phototrophs. Figure 1.2 shows not only the geographical variation of the potential for primary production on land (i.e. the potential for carbon to be incorporated into terrestrial living material), but also the average concentration of chlorophyll in algae living in surface waters. Chlorophylls are the light-collecting pigments found in photosynthesising organisms that give them their green colouration. Although both Figures 1.1 and 1.2 are composites of many satellite images, they are akin to snapshots in time. They provide information about the standing stock of plant material, but by themselves they do not reveal anything about the *rates* at which plant material is being made – the primary *productivity* – in different environments. Nor do they indicate anything about the rates at which the plant material is being eaten by animals, decomposing or being recycled. As you will see, because organic material is essentially carbon, almost all of which can eventually find its way back into the atmosphere as gaseous CO_2, all these processes are important influences on the Earth's climate and thus its habitability.

■ Returning to Figure 1.1 for a moment, suggest why the vegetation patterns shown are not wholly reliable as indicators of local climatic conditions.

■ (i) Humans have direct effects on vegetation through for example the removal of forests (particularly in temperate latitudes) and irrigation of land in arid regions; their domesticated animals graze the vegetation. (ii) Seasonal changes in local climatic conditions are not reflected in the figure.

Nevertheless, the patterns of primary production seen in Figures 1.1 and 1.2 are to a large extent determined by the movement of air and water over the surface of the Earth – including the swirling clouds. You will see *how* shortly, but first one of the primary influences on the continual motion of the Earth's atmosphere and oceans, i.e. energy from the Sun, will be discussed.

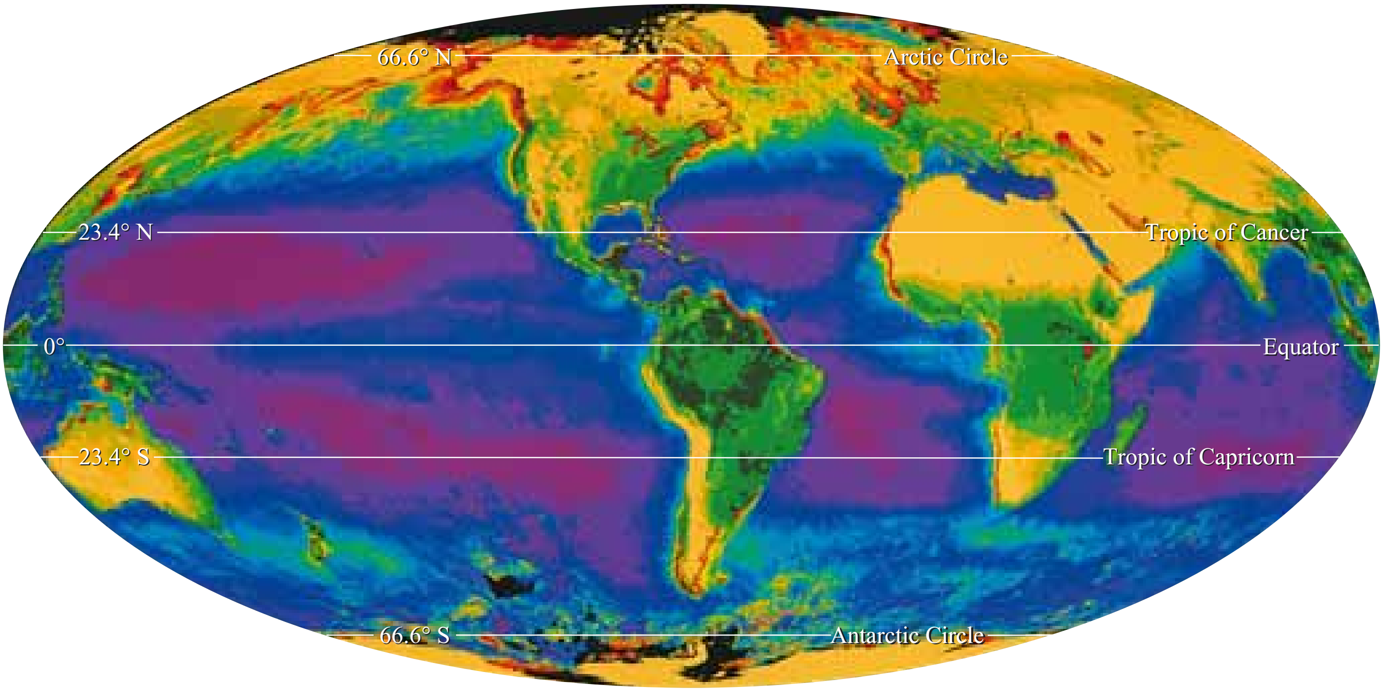


Figure 1.2 Global distribution of the potential for primary production on land and in surface waters, as indicated by chlorophyll concentration (determined using satellite-borne sensors). On land, the darkest green areas (e.g. Brazil) correspond to the greatest potential for production of new plant material; decreasing production potential is indicated by increasingly paler greens. Least productive of all are the paler deserts, high mountains and arctic regions shown in yellow. In surface waters, regions of highest productivity (bright red, e.g. Canadian coast) are mostly around the coasts, followed by yellow, green and blue. Least productive oceanic regions (e.g. large areas in the tropics) are shown in purplish-red.

1.2 Energy from the Sun

The Earth is, on average, about 150 × 10^6 km from the Sun. From this, the average amount of solar energy reaching the top of the atmosphere can be calculated. The amount of solar energy that would fall on a surface at right angles to the Sun's rays, known as the **solar flux** (or solar irradiance, or solar constant), is ~1370 W m^{-2}. This means that the amount of solar radiation that the Earth intercepts is ~1370 W $m^{-2} \times \pi r^2$ W, where πr^2 is the area of a disc with the same radius as the Earth. However, the Earth is roughly *spherical*, so the area presented to the incoming solar radiation by the rotating Earth (over any period longer than a day) is $4\pi r^2$, i.e. four times as great. The average flux of solar energy is therefore effectively only a quarter of the solar flux:

$1 \text{ W} = 1 \text{ J s}^{-1}$

Surface area of a sphere
$A = 4\pi r^2$, where r is the radius

$$\frac{1370 \text{ W m}^{-2}}{4} \approx 343 \text{ W m}^{-2}.$$

So the average amount of solar energy reaching the top of the atmosphere, i.e. the *effective solar flux* is ~343 W m^{-2} (Figure 1.3).

Not all of this incoming solar energy is available to heat the Earth–atmosphere system: about 30% of it is reflected back into space, mainly from the tops of clouds. In other words, the **albedo** of the Earth as a whole (i.e. the fraction of

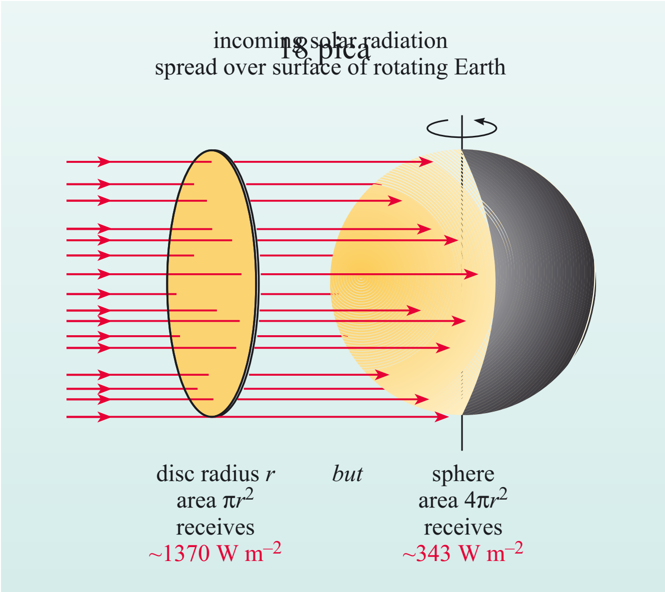


Figure 1.3 Diagram to illustrate the average flux of solar energy reaching a disc and the Earth (with the same radii).

incoming solar radiation that is reflected from it) is about 30%. This means that the Earth–atmosphere system receives ~70% of the solar flux, i.e. 240 watts of solar energy per square metre (240 W m^{-2}). But this figure of 240 W m^{-2} is an average, and it is the uneven distribution of this heating that drives the Earth's climate engine.

■ Look at Figure 1.4. Bearing in mind that the atmosphere absorbs a proportion of incoming solar energy, suggest two reasons why the intensity of solar radiation at the Earth's surface, and hence the surface temperature, is generally lower at high latitudes than at low latitudes.

■ The intensity of solar radiation at the Earth's surface (on the diagram, the number of rays per unit area) depends on the angle of the rays with respect to the surface: the more oblique the angle, the larger the area over which the solar energy will be spread. Furthermore, the more oblique the rays, the greater the thickness of atmosphere through which the rays will have to travel.

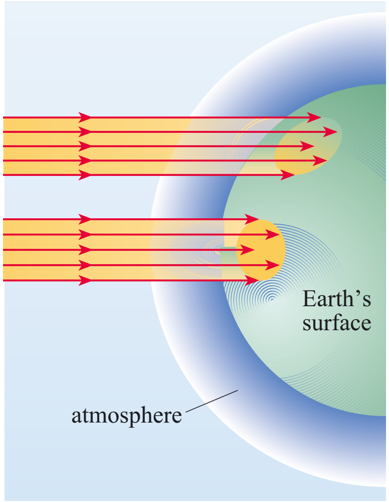


Figure 1.4 Schematic diagram to show why there is a difference in the intensity of solar radiation reaching the Earth's surface at different latitudes (the atmosphere is not to scale).

The relationship between the angles the Sun's rays make with the ground and their ability to warm it is, in fact, the origin of the word 'climate' (from the Greek, meaning slope). Slopes that face equatorwards (i.e. southwards in the Northern Hemisphere) – where the Sun's rays meet the ground at a steeper angle – are warmer.

If the Earth's axis of rotation were at right angles to the plane of its orbit, for any given latitude, the angle at which the rays of the noonday Sun fell upon the surface would remain constant, with higher latitudes in both hemispheres always receiving less solar radiation than lower latitudes. In other words, the Earth's surface at, for example, 10° N and 10° S would always receive the same amount of solar radiation, with this always being more than that received at, for example, 40° N and 40° S. The Sun would be overhead at noon only at the Equator, the poles would have perpetual twilight, and night and day would always be the same length (i.e. each 12 hours long) everywhere on the globe.

But the Earth's axis of rotation is *tilted* with respect to the plane of its orbit, currently at an angle of 23.4°. As a result, the latitude at which the noonday Sun is overhead migrates between 23.4° N (the Tropic of Cancer) and 23.4° S (the Tropic of Capricorn), passing the Equator only twice a year, at the equinoxes (when the lengths of night and day are equal). This change in the position of the noonday Sun throughout the year is the cause of the seasons (Box 1.2).

Box 1.2 The cause of seasons

Figure 1.5a shows the passage of the seasons for the Northern Hemisphere.

- Along the Tropic of Cancer (23.4° N), the noonday Sun is overhead, and maximum solar radiation is received during the summer solstice (the longest day), which is 21 June.

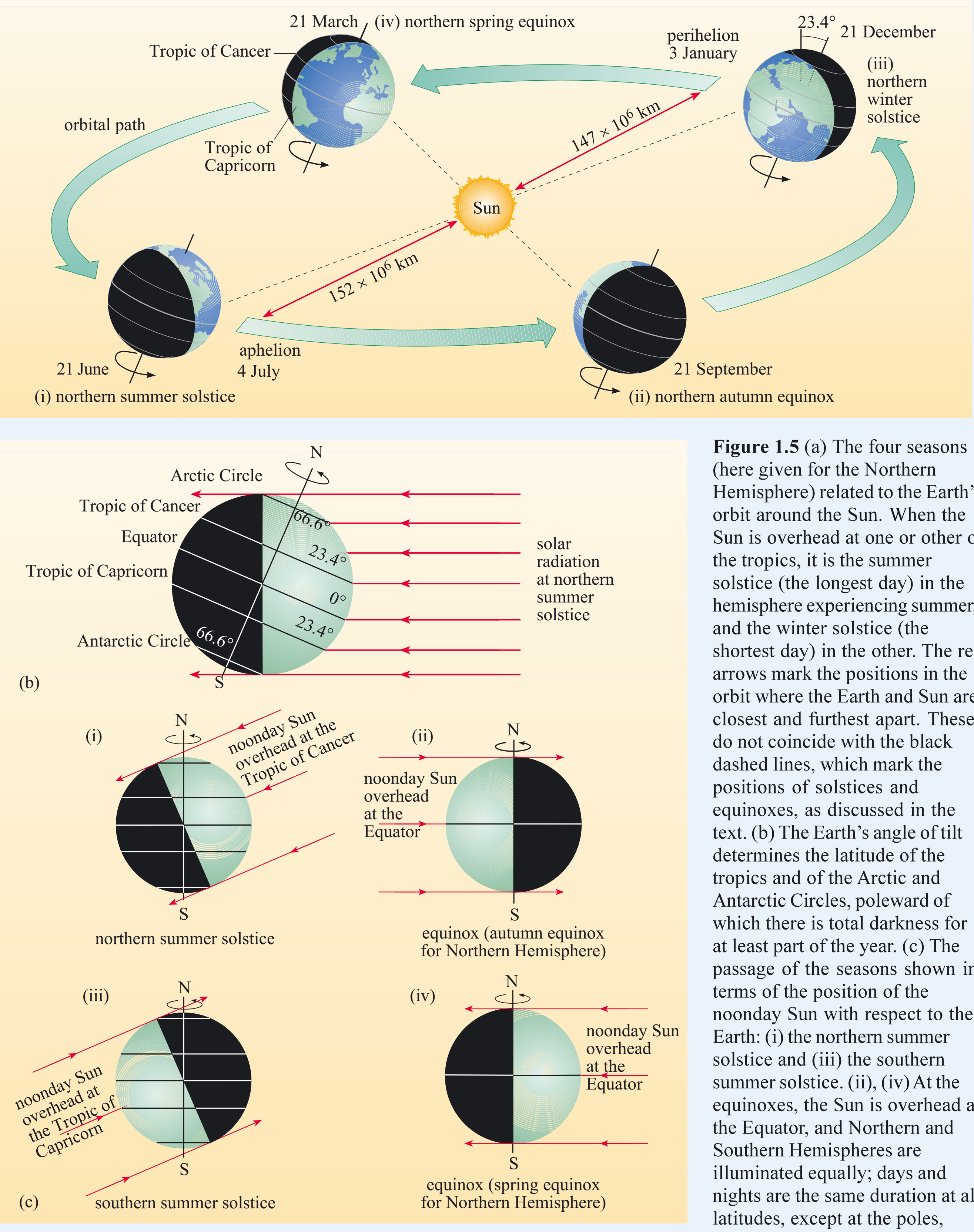


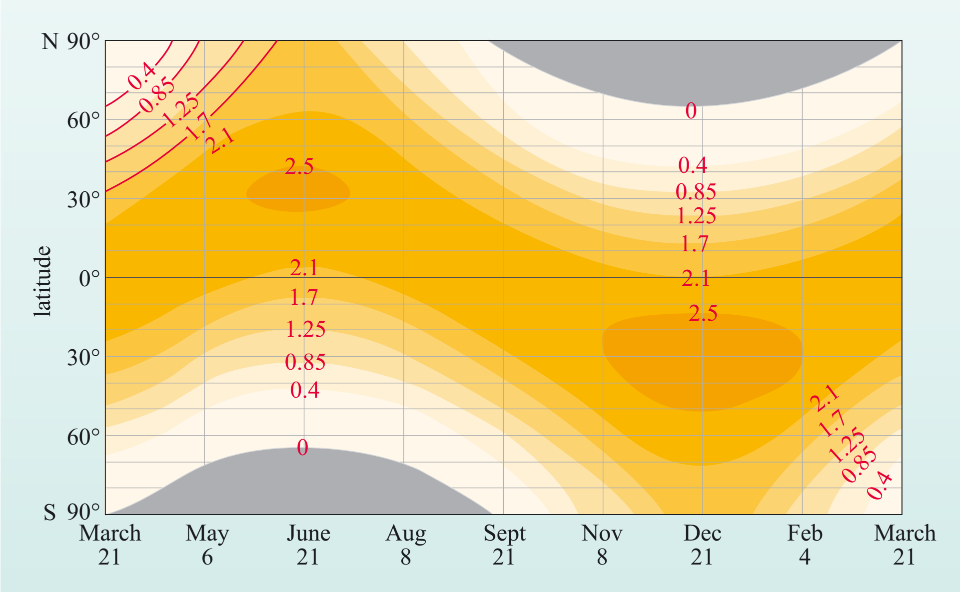
Figure 1.5 (a) The four seasons (here given for the Northern Hemisphere) related to the Earth's orbit around the Sun. When the Sun is overhead at one or other of the tropics, it is the summer solstice (the longest day) in the hemisphere experiencing summer, and the winter solstice (the shortest day) in the other. The red arrows mark the positions in the orbit where the Earth and Sun are closest and furthest apart. These do not coincide with the black dashed lines, which mark the positions of solstices and equinoxes, as discussed in the text. (b) The Earth's angle of tilt determines the latitude of the tropics and of the Arctic and Antarctic Circles, poleward of which there is total darkness for at least part of the year. (c) The passage of the seasons shown in terms of the position of the noonday Sun with respect to the Earth: (i) the northern summer solstice and (iii) the southern summer solstice. (ii), (iv) At the equinoxes, the Sun is overhead at the Equator, and Northern and Southern Hemispheres are illuminated equally; days and nights are the same duration at all latitudes, except at the poles, which are grazed by the Sun's rays for 24 hours.

- After 21 June, the days begin to shorten, until at the autumn equinox, on 21 September, day and night are of equal length.
- After 21 September, day-lengths continue to shorten, until the shortest day (the winter solstice, on 21 December), after which the days begin to lengthen again.
- At the Equator, maximum solar radiation is received at the March and September equinoxes, when the noonday Sun is overhead, and day and night are of equal length. Poleward of the tropics, the Sun is *never* overhead, although it is at its highest at the summer solstice. The poles themselves are wholly illuminated in summer and wholly dark in winter.

The Earth's angle of tilt (presently 23.4°) determines the latitude of the tropics (where the Sun is overhead at one of the solstices) and of the Arctic and Antarctic Circles at 66.6° N and 66.6° S respectively (90° – 23.4°), poleward of which there is total darkness for at least part of the year (Figure 1.5b). The seasons in terms of the position of the noonday Sun with respect to the Earth are shown in Figure 1.5c.

Figure 1.6 shows the seasonal variation in the amount of solar radiation received daily at the Earth's surface (i.e. taking into account the amount absorbed in the atmosphere). The zero contour corresponds to 24-hour darkness. At the North Pole (90° N), it encompasses the period between 21 September and 21 March, and at the South Pole it encompasses the period between 21 March and 21 September. In each case, the first of these dates is the autumn equinox and the second is the spring equinox (compare with Figure 1.5). Don't worry if this type of diagram seems strange – you should be able to see how it works when you have tried Question 1.1.

Figure 1.6 Seasonal variation of daily incoming solar radiation (in 10^7 J m^{-2}) at the Earth's surface, taking account of absorption by the atmosphere but ignoring the effect of topography. Note that this is not an ordinary spatial map, but a map, or plot, of incoming solar energy against latitude on the one axis and time of year on the other.



Question 1.1

(a) (i) With reference to Figure 1.6, describe briefly how the incoming solar radiation changes over the course of the year at 50° N.

(ii) The units used in Figure 1.6 are J m^{-2} because it shows values of incoming solar radiation *per day*. Convert the contour values so that they are for *average* incoming solar radiation in W m^{-2}.

(b) Over the year as a whole Figure 1.6 shows that, on average, the Equator receives the most solar radiation. Which latitudes receive the most solar radiation at any one time? Why is this?

One aspect of Figure 1.6 that may be initially puzzling is that the maximum amount of solar energy received by southern mid-latitudes in the southern summer is *greater* than the maximum amount received by northern mid-latitudes in the northern summer (e.g. the areas enclosed by the 2.5×10^7 J m^{-2} contour). Furthermore, careful study of Figure 1.6 indicates that in the southern summer *all* latitudes receive more energy than the corresponding latitudes in the other hemisphere in the northern summer. This is because the Earth's orbit is elliptical, and at the present time the Earth comes closest to the Sun (i.e. is at **perihelion**) during the southern summer (on 3 January), and is furthest from the Sun (i.e. is at **aphelion**) during the northern summer (on 4 July). It is because the Sun is at one of the two foci of the ellipse, rather than at its geometric centre, that perihelion and aphelion only occur once a year rather than twice a year.

Because of the varying gravitational attraction of the Sun and of the other planets (notably Jupiter and Saturn), the degree of ellipticity (*eccentricity*, or off-centredness) of the Earth's orbit varies with time, and over a period of about 110 000 years changes from its most elliptical (maximum eccentricity) to nearly circular and back again (Figure 1.7a). This 110 000-year cycle is the longest of three astronomical cycles that affect the amount and distribution of solar radiation reaching the Earth's surface, and it is the only one that affects the *total* amount of solar radiation reaching the Earth. The two shorter cycles (tilt and precession: Figure 1.7b) involve the orientation of the Earth's spin axis, and so affect the *distribution* of solar radiation over the Earth's surface.

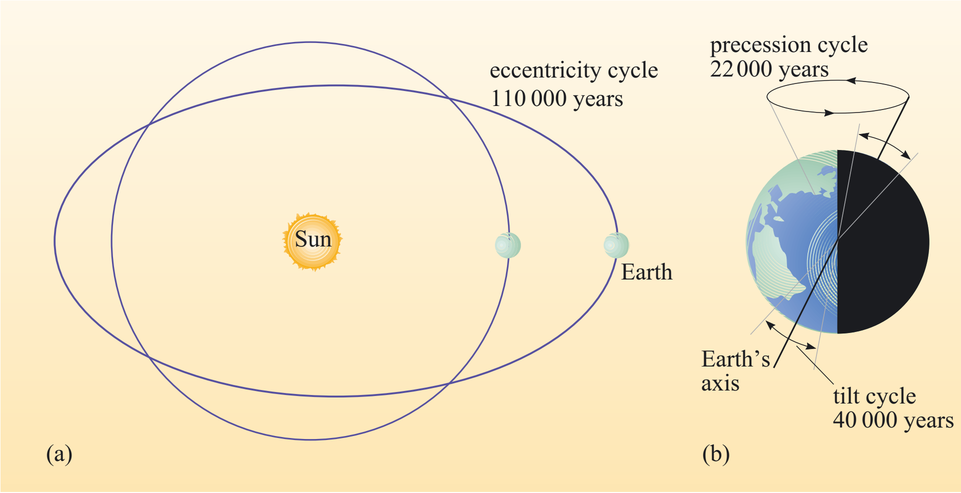


Figure 1.7 The component Milankovich cycles. (a) Plan view of the Earth's orbit to show how it changes shape from circular to most elliptical and back again, over the course of the 110 000-year eccentricity cycle. (b) The Earth showing the 40 000-year tilt cycle and the 22 000-year precession cycle.

These three astronomical cycles (eccentricity, tilt and precession) are usually known as **Milankovich cycles**, after Milutin Milankovich, a Serbian astronomer. Milankovich's work in the 1930s and 1940s was an improvement and refinement of the work of Scotsman James Croll, for this reason the cycles are sometimes referred to as Milankovich–Croll cycles.

Over the course of about 22 000 years, the direction in which the Earth's spin axis points, traces out a circle. Therefore, from about 11 000 years from now, the positions in the orbit of the northern and southern summer will be reversed as the seasons (i.e. the solstices and the equinoxes) will have moved clockwise round the orbit. In another 11 000 years, they will be back in their current positions. This phenomenon is often referred to as the *precession of the equinoxes*. (Of course, our calendars will continually have to be adjusted to take account of this so that, for example, the northern summer solstice will remain in June and not gradually drift towards December.) Eccentricity will be roughly as it is now, thus the Earth will be at perihelion in northern, rather than southern, summer.

At the same time that the Earth's spin axis traces out a circle, the angle it makes with the normal to the orbital plane varies between about 21.8° and 24.4°, and back again, with a periodicity of about 40 000 years; at the moment, the angle of *tilt* is about 23.4°.

■ Bearing in mind Figure 1.5, what do you think the latitude of the tropics would be if the angle of tilt increased so that it was 24.4°, rather than 23.4°? What effect would this have on Figure 1.6?

■ If the angle of tilt increased to 24.4°, the tropics (over which the Sun would be directly overhead during the summer solstice) would be at 24.4° N and 24.4° S. This would mean that on a diagram like Figure 1.6, the areas of maximum incoming solar radiation corresponding to summer months would be shifted polewards slightly, and for the winter hemisphere, the zero contour (for example) would extend a little further towards the Equator.

So the greater the angle of tilt, the greater is the difference between winter and summer. At present the angle of tilt is in fact *decreasing*, so summers should be very gradually becoming cooler and winters should be very gradually becoming warmer.

The form of the three cycles over the past 800 000 years can be seen in Figure 1.8.

- The 110 000-year eccentricity cycle (Figure 1.8a) is actually a combination of a 100 000-year cycle with a much weaker 413 000-year cycle, and is sometimes referred to collectively as the 100 000-year cycle.
- The tilt cycle (Figure 1.8b) is sometimes referred to as the 40 000-year or 41 000-year cycle
- The 22 000-year precession cycle (Figure 1.8c) can be broken down into a 19 000-year cycle and a stronger 23 000-year cycle.

At present the eccentricity of the Earth's orbit is such that the Earth–Sun distance is about 147 × 10^6 km at perihelion and about 152 × 10^6 km at aphelion (Figure 1.5a).

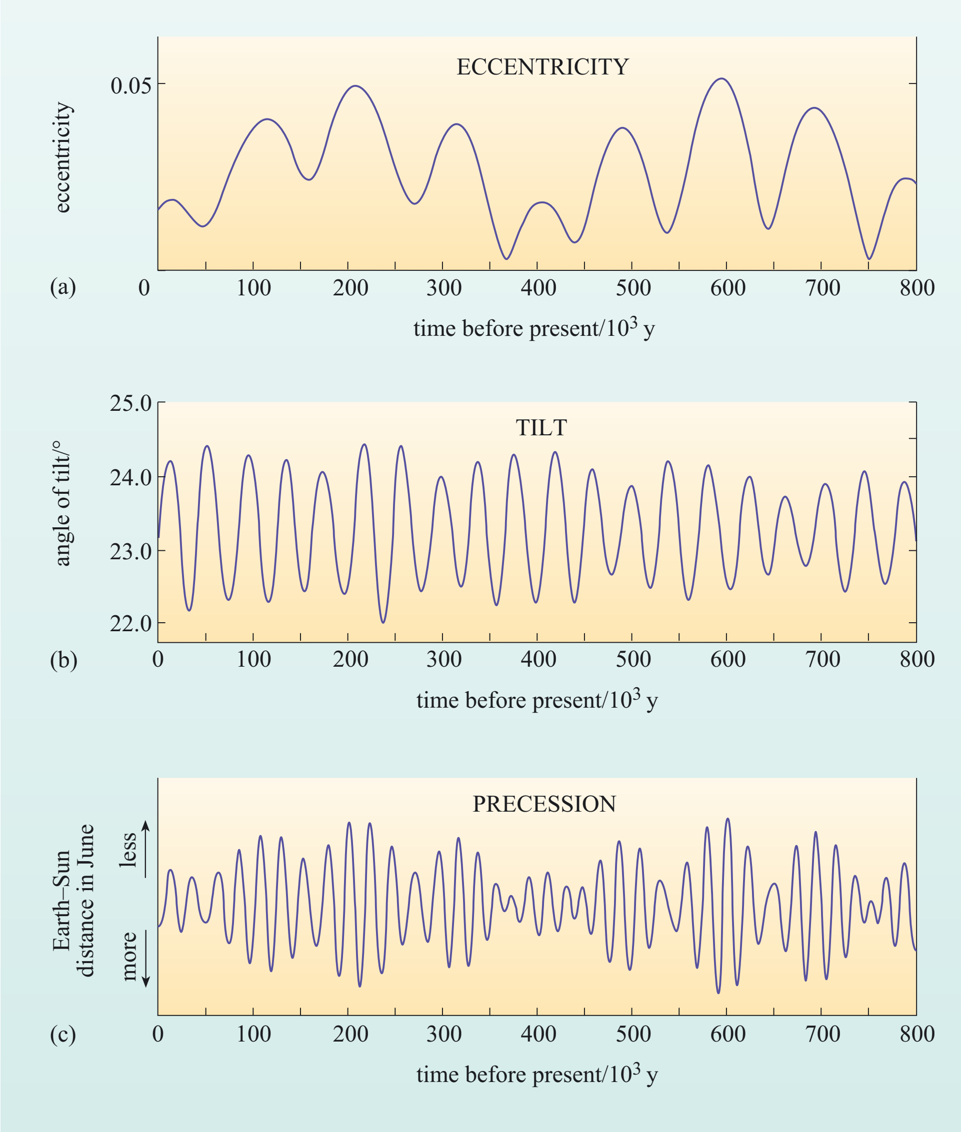


Figure 1.8 Milankovich orbital changes over the past 800 000 years, based on astronomical data. (a) Eccentricity (the higher the value, the more elliptical the orbit: an eccentricity of zero corresponds to circular). (b) Tilt. (c) Precession expressed in terms of the Earth–Sun distance in June.

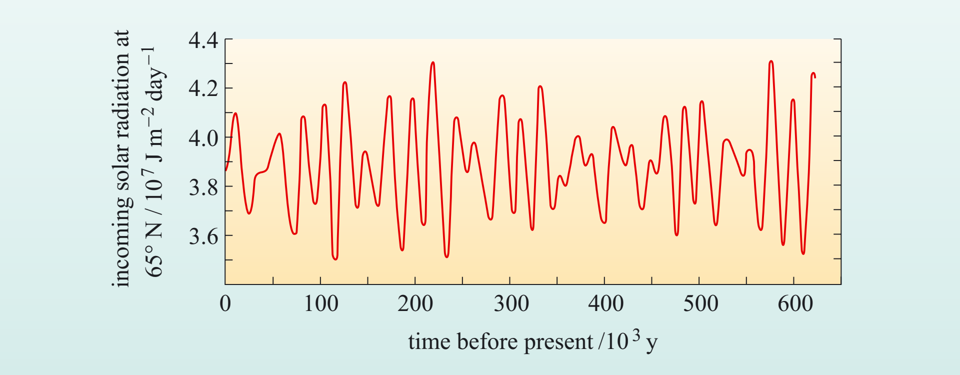
■ According to Figure 1.8, is the Earth's orbit currently becoming more or less elliptical? Approximately when was it last nearly circular?

■ It is currently becoming less elliptical. At the present, the eccentricity curve (Figure 1.8a) is tending towards smaller values, and observation of the wave-like shape of the curve shows that it last passed through a minimum (i.e. a period when the orbit is almost circular) about 50 000 years ago.

Note that an elliptical orbit tends to exaggerate the seasons in one hemisphere (the one for which winter occurs during aphelion and summer occurs during perihelion) and to moderate them in the other; the more elliptical the orbit, the more extreme this effect is.

Figure 1.9 was computed using the information shown in Figure 1.8 and shows how the intensity of summer sunshine at high northern latitudes (65° N) has varied in response to Milankovich orbital changes over the past 650 000 years.

Figure 1.9 The variation in incoming solar radiation in summer at northern latitudes (65° N) over the past 625 000 years.



There is good geological evidence that Milankovich cycles have been important agents of climatic variation over much of the last ~500 Ma. However, the complexity of the climate system and interactions between the different components, including both **positive** and **negative feedback** effects, mean that it is difficult to predict the effect of a given change in incoming solar radiation. Despite this, cycles play an important role in linking incoming solar radiation and its variations to the climate, and thus habitability, of the surface of the Earth.

One way in which astronomical variations may affect the Earth's climate is through the influence they have on the growth and decay of the polar ice caps and, hence, the amount of solar radiation reflected by the Earth. As mentioned earlier, the planetary albedo (i.e. the fraction of incoming solar radiation reflected from the Earth) is about 30%. Albedos for various surfaces are given in Table 1.1. Note that it is not possible to provide precise values because the reflectivity of a surface depends not only on what it is made of, its colour and its roughness, but also on the angle of the incoming radiation and the wavelength.

Table 1.1 Some typical albedos.

Type of surface	Albedo
fresh snow and sea-ice	80–90%
clouds	average ~55% (~15–80%, depending on type and thickness)
thawing snow	~45%
desert	35%
grassland	25–33%
forest, bare soil, rock, cities	~10–20%
water:	
moderate–high Sun (elevation >40°)	<5%
low Sun (elevation ~10°)	>50%

■ Apart from changes in snow and ice cover, how might changes in climate cause changes in planetary albedo?

■ Changes in cloud cover and vegetation type would also affect planetary albedo.

1.2.1 The atmosphere – a protective filter

So far, illumination and heating by solar radiation have been taken as being almost interchangeable, but visible and thermal radiation occupy different parts of the Sun's spectrum (i.e. they have different frequencies and wavelengths). As frequency and wavelength of radiation have very important climatic implications you should now read Box 1.3, which describes important facts about the **electromagnetic spectrum** and thermal energy.

Box 1.3 The electromagnetic spectrum and thermal energy

The Sun emits electromagnetic radiation over a wide range of the electromagnetic spectrum, from gamma-rays to radio waves (Figure 1.10), but ~99.9% of the energy is in the wavelength range 0.15–5 μm (Figure 1.11), which includes ultraviolet radiation, visible radiation (violet to red), and infrared (heat) radiation. As with any propagating waves, the shorter the wavelength (λ), the higher the frequency (f), and multiplying the two together will give the speed (c):

$$c = f\lambda \qquad (1.3)$$

where the term '*ultra*violet' refers to radiation with a higher frequency than that of violet, and the term '*infra*red' refers to radiation with a lower frequency than that of red.

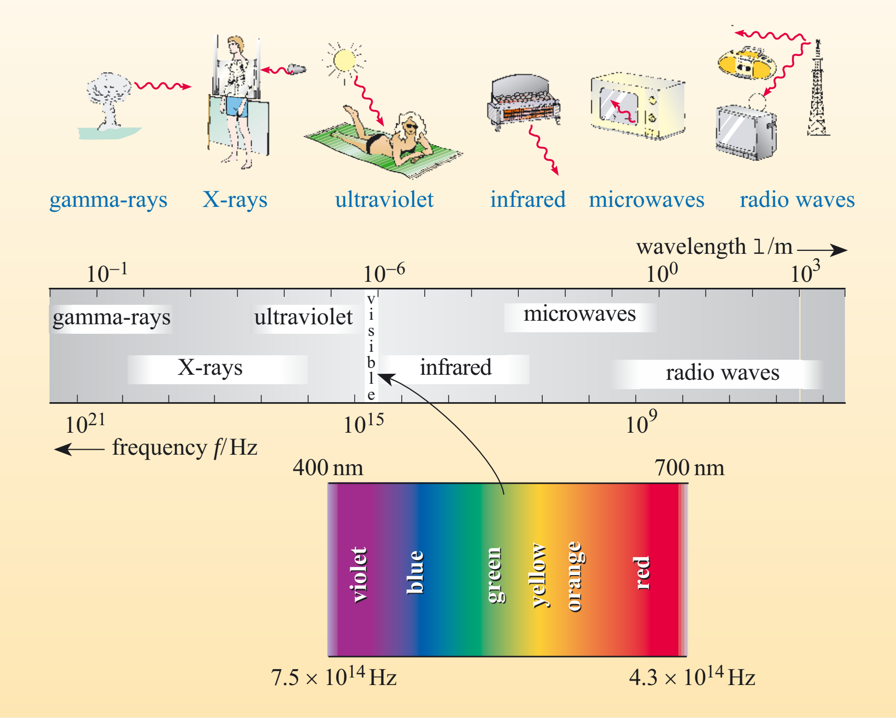


Figure 1.10 The electromagnetic spectrum. Wavelength is given in metres. For the expanded visible spectrum, wavelength is given in nanometres (1 nm = 10^{-9} m); frequency is given in hertz (Hz = cycles s^{-1}).

Note that the process of radiation is completely different from reflection, in which energy is not absorbed and reradiated, but ‘bounced off’, with its frequency and wavelength unaffected (think of light reflected by a mirror).

All molecules vibrate, and when a molecule absorbs electromagnetic energy, the amplitude of its vibration increases. Once ‘excited’ in this way, molecules can lose the energy again, either by re-emitting infrared radiation or by converting it into kinetic energy, by colliding with other molecules. The effect of such collisions is to raise the internal energy of the material, i.e. to raise its temperature. In the case of a solid or liquid, the vibrations of the excited atoms and/or molecules not only cause the surface to radiate longwave radiation, but the collisions of excited atoms (particularly electrons) with adjacent less energetic atoms result in the transfer of energy down the temperature gradient, by **conduction**.

It is often scientifically convenient to consider the Sun (and indeed the Earth) as a black body, i.e. a body that is radiating energy at the maximum possible rate for its temperature. If this assumption is made, then energy radiated per unit area per unit time is proportional to T^4, where T is the absolute temperature in kelvin (K). This expression is known as the *Stefan–Boltzmann law*, and it states:

1. Any body above absolute zero emits radiant energy.
2. The energy emitted per unit area per unit time is proportional to the fourth power of the temperature in kelvin. Remember that the kelvin scale starts at absolute zero, or –273.15 °C, so 1 °C = 274.15 K.

Figure 1.11a shows the spectral curves for the Sun and the Earth, assuming that they radiate like black bodies of 6000 K and 255 K, respectively. These curves demonstrate a general principle: the higher the temperature of a surface, the more the maximum in the spectrum of energy it radiates is shifted towards shorter wavelengths and higher frequencies. As a practical example, think of a lump of coke in a furnace. As the coke starts to get hot it glows red; as it gets hotter and hotter, shorter and shorter wavelengths are emitted, until all of the visible part of the spectrum is being emitted, resulting in white heat.

■ To what extent do the ranges of wavelengths emitted by the Sun and by the Earth overlap?

□ The curves overlap only very slightly. Incoming solar radiation is in the ultraviolet, visible and infrared (thermal) bands; outgoing radiation is all in the infrared, and there is a small overlap between the two curves at wavelengths of 4.0–4.5 μm.

■ Is it reasonable to use shortwave and longwave as shorthand for solar radiation and thermal energy radiated by the Earth, respectively?

□ In the context of radiation incident on and emitted by the Earth, it is reasonable to think of solar radiation as shortwave and terrestrial radiation as longwave.

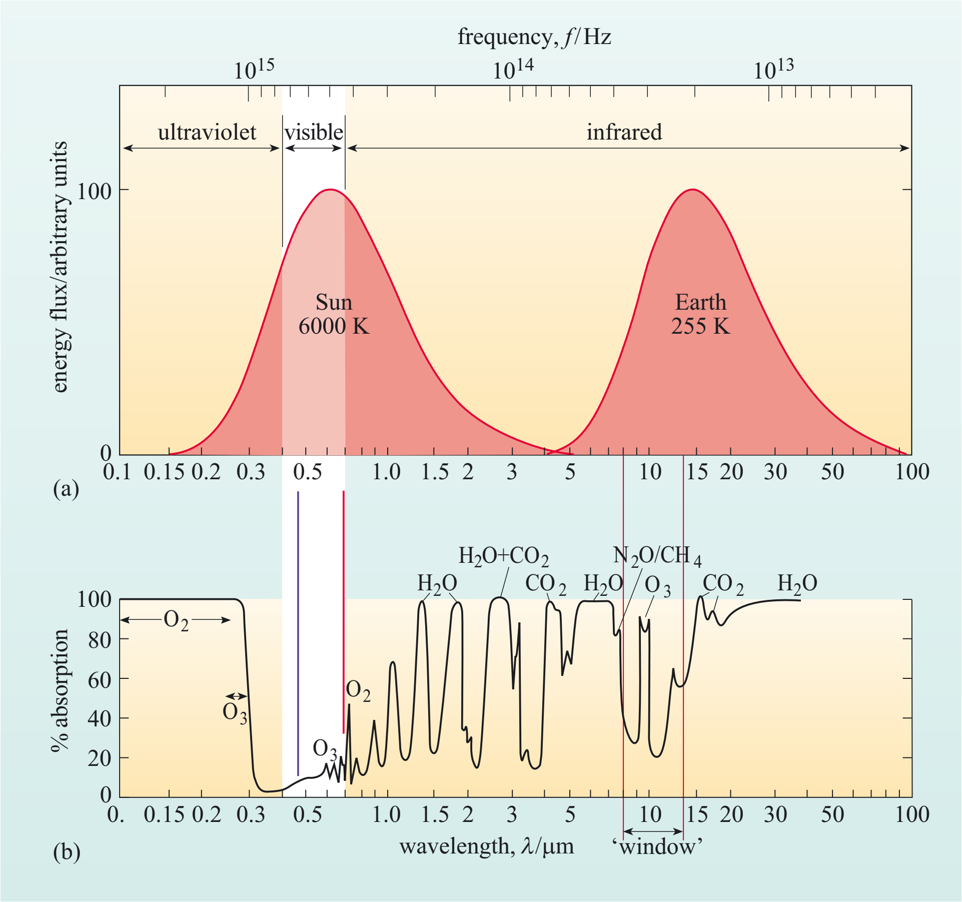


Figure 1.11 (a) The spectrum of electromagnetic radiation emitted by the Sun (assuming that it is a black body at 6000 K) and the Earth (assuming that it is a black body at 255 K). This latter value (255 K = –18 °C) is the Earth's *effective* planetary temperature, consistent with the radiation it emits to space. Because the Earth has an atmosphere, the average temperature *at its surface* is much higher, at about 288 K (= 15 °C). *Note*: Both horizontal scales are logarithmic. Wavelength and frequency are inversely related, so while the wavelength increases from left to right, frequency increases from right to left (Equation 1.3). (b) The absorption spectrum of the Earth's atmosphere, showing the total percentage of radiation absorbed on passing through the atmosphere as a function of wavelength. This is a composite curve with the effects of the different atmospheric gases added together (CH_4 is methane and N_2O is nitrous oxide). Also shown within the visible band are the wavelengths of the radiation most used by photosynthesising organisms (~0.45 µm and ~0.65–0.7 µm). *Note*: these are the wavelengths *absorbed*, not the ones reflected (it is the latter that determine the colour of chlorophyll pigment).

The curves in Figure 1.11a are emission curves. They show the spectra of radiation emitted by the Sun and Earth. The spectrum of solar radiation reaching the Earth's surface, however, is strongly modified by absorption and reradiation within the atmosphere. The atmosphere acts as a filter, removing dangerous ultraviolet radiation and maintaining habitable conditions at the surface. To understand how it does so, you need to consider its gaseous composition.

Table 1.2 lists the gases in the atmosphere in order of abundance. By far the most abundant gases are nitrogen and oxygen, which together make up ~99% of the total. Apart from nitrogen and argon, all of these gases affect the Earth's climate through their interaction with incoming (shortwave and longwave) radiation, outgoing longwave radiation, or both. As shown in Figure 1.11b, each gas absorbs particular wavelengths, often in more than one part of the spectrum.

Figure 1.11a shows, in effect, the spectrum of wavelengths emitted by the Sun, and the spectrum of wavelengths reradiated by the Earth. The curve in Figure 1.11b shows the percentage of absorption of radiation passing through the atmosphere, as a function of wavelength. This is a composite curve with the effects of the different atmospheric gases added together – some peaks are entirely attributable to a particular gas, others result from the combined effect of two or more gases.

Table 1.2 Gases naturally present in the atmosphere, and current concentration of each.

Gas	Approximate concentration by volume*
nitrogen, N_2	78%
oxygen, O_2	21%
argon, Ar (and other noble gases)	1%
water vapour, H_2O	~ 3000 ppm (0.3%), but very variable
carbon dioxide, CO_2	~ 380 ppm (0.038%)†
ozone, O_3	~ 0.01–0.1 ppm
methane, CH_4	1.8 ppm§
nitrous oxide, N_2O	0.3 ppm

*Measured as the proportion of the total number of molecules in the atmosphere contributed by each component, equivalent to its proportion by volume in the atmosphere (ppm = parts per million). *Note*: these values are averages.

†Value in 2006, increasing by ~1–3 ppm annually.

§Value in 2006, increasing by 0–0.01 ppm annually.

As far as oxygen and ozone are concerned, most absorption of incoming solar radiation occurs in the stratosphere and affects the short-wavelength ‘tail’ of the Sun’s spectrum (i.e. the ultraviolet). Its absorption has important implications for life on Earth, because proteins and nucleic acids are damaged by radiation with wavelengths less than about 0.29 μm. As Figure 1.11b shows, for most ultraviolet radiation, absorption by oxygen provides an effective filter, but for frequencies approaching 0.29 μm only ozone has any impact.

Question 1.2

(a) To what extent are (i) visible radiation and (ii) incoming infrared radiation absorbed by atmospheric gases? Which gases are mainly responsible in each case?

(b) Which atmospheric gases are mainly responsible for the absorption of outgoing longwave radiation?

The gases that absorb most energy from incoming solar radiation are oxygen and ozone, water vapour and carbon dioxide. For outgoing longwave radiation, ozone, water vapour and carbon dioxide are again important, along with nitrous oxide (N_2O) and methane (CH_4). Much of the outgoing longwave radiation is in the range 8–13 μm, where there is relatively little absorption. This range is therefore known as the **atmospheric window**. Gases that absorb radiation reradiate energy *in all directions* and the absorption of outgoing longwave radiation and the subsequent reradiation, much of it back towards the Earth’s surface, is what keeps the Earth’s average surface temperature (~15 °C) so much higher than the effective planetary temperature (~ –18 °C) at which the system as a whole emits radiation to space. This is referred to as the **greenhouse effect**.

Figure 1.12 summarises the overall energy budget for the Earth and its atmosphere, including all the factors considered so far. The 100 ‘units’ of

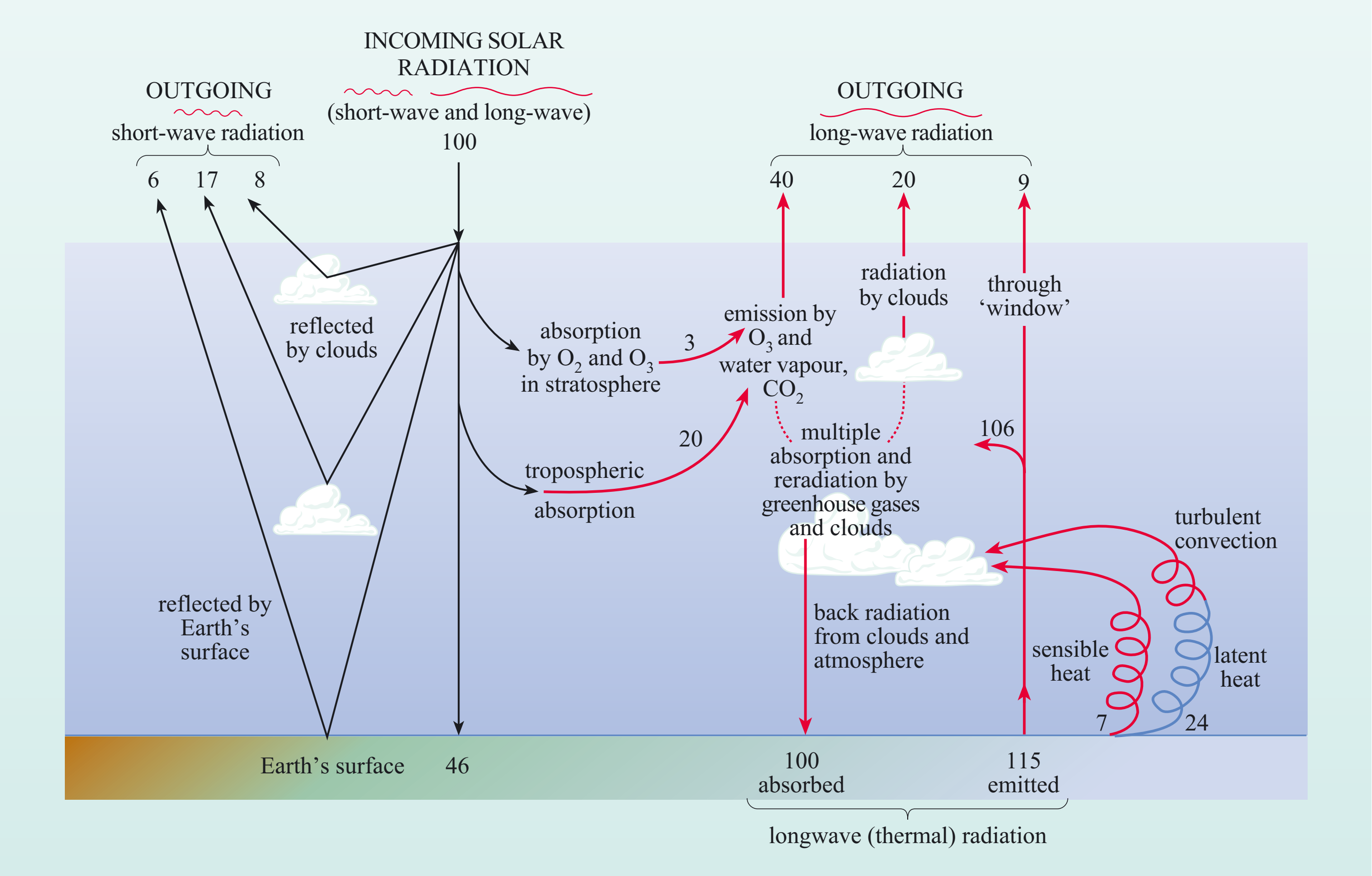


Figure 1.12 Schematic diagram to show the overall energy budget of the atmosphere. Values for the outgoing radiation have been measured by satellite-borne radiometers; whereas the *reradiated* radiation (back radiation) has a longer wavelength than the incoming radiation, *reflected* radiation has the same wavelength after reflection as before. Other values are derived from model calculations or measurements, and you may find slightly different values given elsewhere. The effects of greenhouse gases and clouds result in the energy that the Earth's surface radiates (115 units) being greater than that originally absorbed as solar radiation (46 units).

incoming solar radiation represent the effective solar flux at the top of the atmosphere of ~ 343 W m^{-2} (Section 1.2). Note that:

- *at the top of the atmosphere*, incoming and outgoing radiation are in balance (100 units = 6 + 17 + 8 units reflected + 40 + 20 + 9 units reradiated)
- *within the atmosphere*, total energy absorbed, from both the Sun and back radiation (20 + 3 + 106 + 24 + 7 = 160 units), balances the total energy emitted, both as back radiation and from the top of the atmosphere (100 + 40 + 20 = 160 units).

The percentage of incoming solar energy reflected by the Earth (i.e. 6 + 17 + 8 = 31%) is close to our earlier estimate for the Earth's albedo (Section 1.2).

Note that liquid water in the form of clouds absorbs and reradiates longwave radiation in the same way as water vapour does. Clouds thus play a very important role in the greenhouse effect but, unlike gases, they also strongly affect the Earth's albedo. The net contribution of clouds to climate change is extremely difficult to gauge.

In summary, although around 30% of the shortwave energy input from the Sun is reflected back to space, while some of the remainder is absorbed directly by the atmosphere, most of it is absorbed at the Earth's surface, only to be re-emitted at longer wavelengths to the atmosphere. Little of this longwave radiation is radiated directly into space however, as most is absorbed in the atmosphere, particularly by carbon dioxide, water vapour and cloud droplets. The atmosphere is therefore heated primarily from below and the overall energy budget of the Earth system is balanced by the longwave radiation re-emitted into space, mostly from the top of the cloud cover.

Incoming and outgoing radiation are thus closely balanced for the whole Earth, but that does not mean that they are in balance for a given region, or range of latitudes. The solid curve in Figure 1.13 shows the average daily amount of solar energy absorbed by the Earth and atmosphere, as a function of latitude. As temperatures at the top of the cloud cover do not vary much with latitude, neither does the intensity of longwave radiation emitted to space: this can be seen from the dashed curve in Figure 1.13.

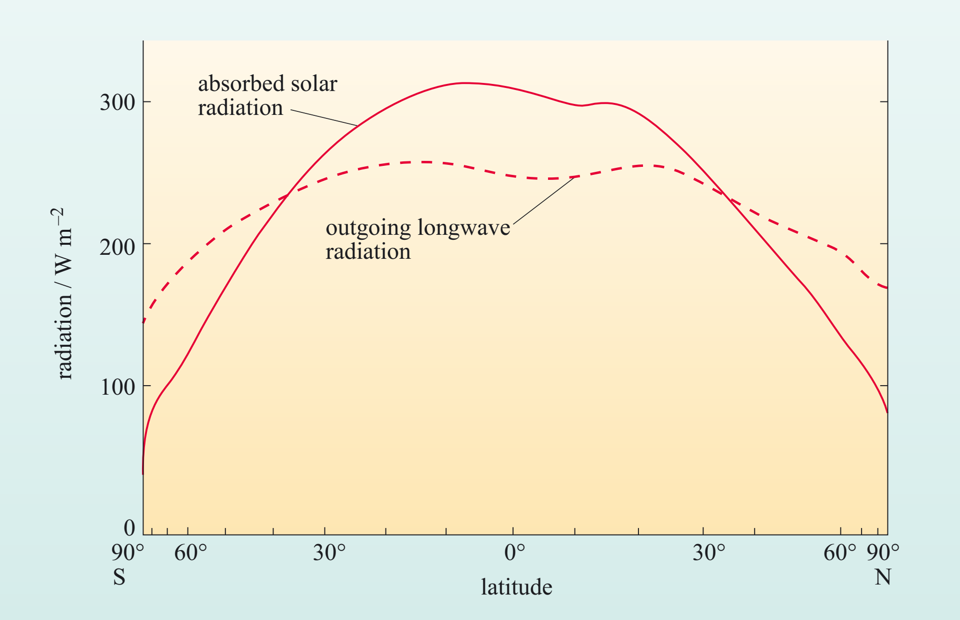


Figure 1.13 The variation with latitude of the solar radiation absorbed by the Earth–atmosphere system (solid curve) and the outgoing longwave radiation lost to space (dashed curve). Values are averaged over the year, and the latitude axis is scaled according to the area of the Earth's surface in different latitude bands so that the area under the graph is proportional to the net radiation of the planet in each band.

Question 1.3

(a) Looking first at the full curve for the solar radiation absorbed by the Earth, give two reasons why it has the general shape that it does. *Hint*: Refer to Figures 1.1 and 1.4, and Table 1.1.

(b) Over what latitudes does the Earth–atmosphere system have a net gain of heat, and over which latitudes does it have a net loss?

Note that together the two regions in Figure 1.45 (in the Answers to questions section towards the end of this book) marked as net loss are approximately equal in area to the region marked as net gain, demonstrating that – at least over short timescales – the overall Earth–atmosphere radiation budget is in balance, so the Earth is neither cooling down nor heating up. Over timescales of decades or more however, small imbalances may lead to net heating or cooling.

Despite the positive radiation balance at low latitudes and the negative one at high latitudes, there is no evidence that low-latitude regions are steadily heating up while high-latitude regions are steadily cooling as a result of natural processes. The reason for this is the continual redistribution of heat over the globe by winds in the atmosphere and currents in the ocean, which form the subject of the next section.

It may have struck you that in this text an implicit assumption has been made: namely that the intensity of radiation emitted by the Sun remains constant. Is this a valid assumption? The answer is no, it isn't. For one thing, according to theories of stellar evolution, the amount of radiation emitted by the Sun at the formation of the Solar System 4600 Ma ago would have been only 70–75% of what it is now. Changes in solar intensity have also been shown to affect climate on timescales of decades to centuries; indeed, while the dominant cause of present global warming is human activity, climatic changes in the 20th–21st centuries can only be fully explained when both natural and **anthropogenic** effects are taken into account.

Radiation from the Sun is an external input to the Earth's climate system that drives or forces much of the activity in the system. Such an external input is described as a **forcing function**, where the word 'function' refers to the distribution or map of the forcing in time or space, represented in this case by Figure 1.6. The human emission of pollutant gases can similarly be considered as being an external input (in this case external to the natural part of the Earth system), and would similarly give rise to a forcing function with a certain distribution in time and space.

The size of the response of a system to a forcing of given magnitude is referred to as its **sensitivity**, the response of the Earth's surface temperature to a change in solar forcing being one example. Unfortunately, the Earth's sensitivity to climate forcing is hard to measure, and (as implied in discussion of the effect of Milankovich cycles) hard to predict. Observation and modelling suggest that an increase in the average amount of incoming solar radiation of 4.0–4.5 W m^{-2} could cause a change in temperature of anything from between 1.5 °C and 5.5 °C, which means that the climate's sensitivity is of the order of 0.5–1.0 °C for 1 W m^{-2} of radiative forcing (say 0.75 °C per W m^{-2} of forcing).

Notice the uncertainty in this estimate. One reason for this is the time it takes for the Earth to respond to a change: as you will shortly see, its systems have huge inertia. If the Sun were to be completely extinguished for a few minutes (as it is locally during eclipses), the effects would be negligible, but if the solar flux were to decrease abruptly and permanently by 0.1%, this change would ultimately show up in all aspects of the climate system, for example the temperature distribution within the atmosphere and ocean, the extents of polar ice caps and of tropical rainforests, and so on. Each of these changes would feed back on other parts of the system. Thus, in trying to assess the sensitivity of the Earth's climate to changes in forcing, all of the various feedback processes that may come into play have to be taken into account, along with their characteristic timescales. An important implication of this is that the sensitivity of a system relates the change in a particular system property or output, to the change in a particular input, over a particular timescale. The sensitivity for a different input, output or timescale will generally be different. Nevertheless, the term 'climate sensitivity' is often used as shorthand for the change in average surface temperature resulting from a given change in atmospheric CO_2 concentration or solar forcing over a timescale of up to a few millennia.

Question 1.4

During the Little Ice Age, which altogether lasted from the 15th century until well into the 19th century, different parts of the Earth experienced unusually cold periods, e.g. icebergs became common off Norway, glaciers advanced down valleys and the Thames froze in winter. It has been suggested that these cold periods, which were most marked from 1640 to 1720, were caused by a prolonged period of low solar activity known as the 'Maunder minimum'. Given that average global temperatures were about 1 °C cooler than today, by how much (as a percentage) would incoming solar radiation (i.e. the effective solar flux) have been below its present value (~343 W m^{-2})?

Note: Assume a climate sensitivity of 0.75 °C per W m^{-2} of solar forcing; both theoretical modelling and studies of the Earth's glacial history indicate that this is a reasonable average value.

In fact, various other factors have been proposed as causing (or contributing to) the Little Ice Age, including natural changes in concentrations of greenhouse gases in the atmosphere and changes in patterns of ocean circulation. Although internal to the natural Earth system, such factors can be considered as forcing agents for the atmospheric part of the climate system.

■ Can Milankovich variation in orbital parameters be described as a forcing function?

■ Yes, because their origin is external to the Earth system. In this case the variation is principally a function of time, as described by Figure 1.8. This is often described as 'Milankovich', 'astronomical', or 'orbital' forcing.

Examples of other forcing functions will be discussed later, but first we will take a look at how the winds and currents combined form the Earth's air-conditioning and/or central heating/cooling system, which in some cases moderates and in others mediates the effects of the various forcing factors and their influence on the habitability of the Earth.

1.3 The Earth's surface temperature pattern

The redistribution of solar energy from low to high latitudes is the principal driving force behind the motions of the atmosphere and oceans and, as such, strongly controls the diversity and habitability of environments on the Earth's surface. Not surprisingly, given the complexity of the Earth's geography and the turbulent, chaotic motions of the atmosphere and oceans, the smooth annual variations of solar forcing seen in Figure 1.6 transform into a much more complicated pattern of average climatic conditions. This section begins by investigating how and why the Earth's 'climate engine' performs this transformation by considering the geographical variation of average surface temperature.

Figure 1.14 shows daytime surface temperatures as measured by a satellite-borne radiometer. In Figure 1.14a and b, temperatures below 0 °C are green and blue (and occur mostly at high latitudes); the highest temperatures are shown by red and dark brown (and are at lower latitudes). In January (Figure 1.14a), temperatures at high northern latitudes are very low, having fallen below 0 °C in Eastern Europe and the northern USA, and are approaching –30 °C over Siberia and most of Canada. In the Southern Hemisphere it is summer, with mid-latitude temperatures of 20–30 °C. By July (Figure 1.14b), areas of the Northern Hemisphere have warmed by 10–20 °C. The Greenland ice cap remains frozen, but temperatures are considerably lower in the Antarctic, where there is now a large area of sea-ice. Temperature differences between January and July are shown in Figure 1.14c. Areas of greatest increase in temperature are red and dark brown (between the Tropic of Cancer and the North Pole); areas of greatest decrease in temperature are bright blue and dark blue (ranging from the continental areas between the Tropic of Capricorn and the Antarctic Circle, and the Antarctic Circle to the South Pole). The greatest changes are over land in mid-latitudes, while low latitudes are more stable.

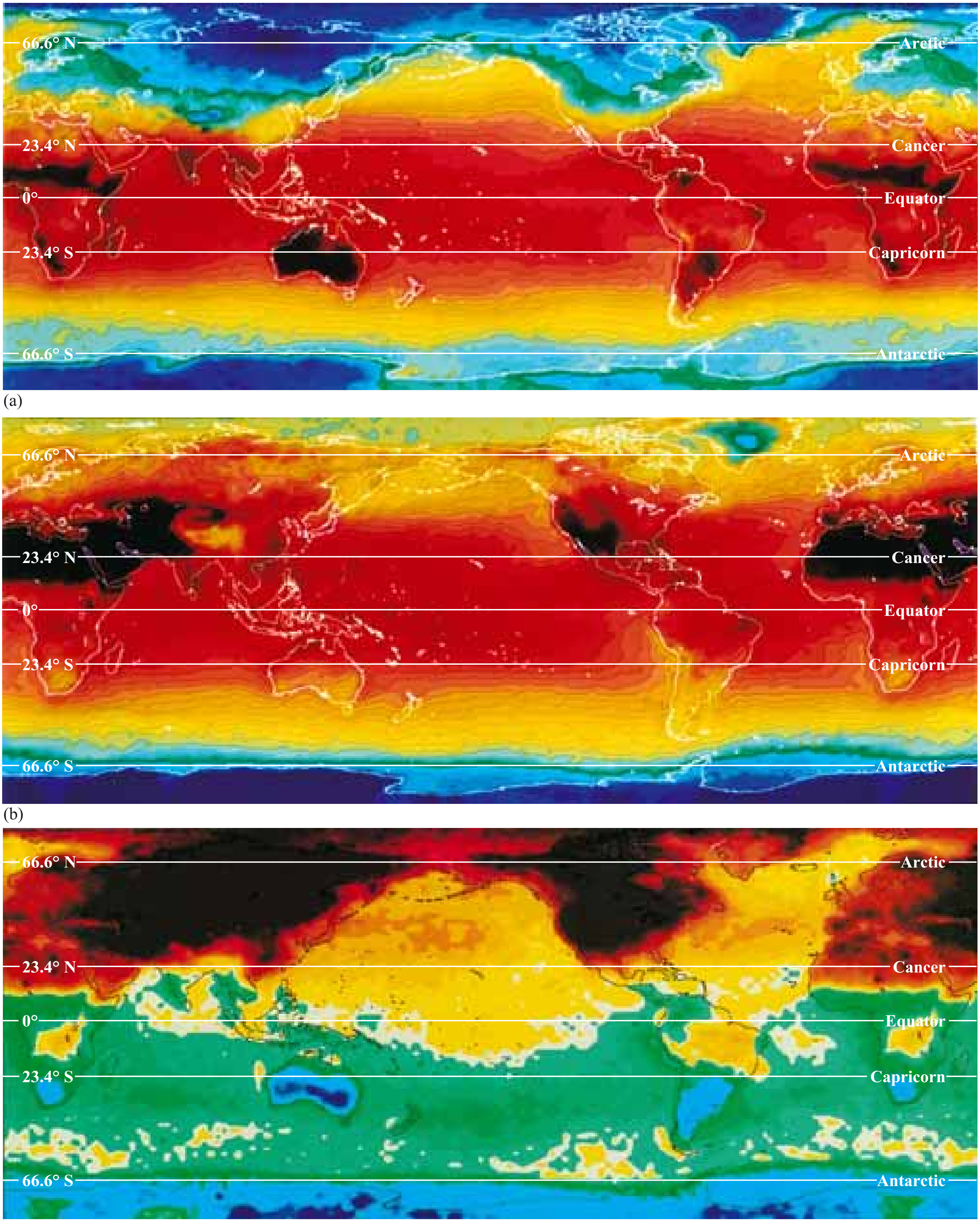
■ With what are the most extreme departures from a simple east–west trend associated?

□ The distribution of continents and oceans. In the hemisphere experiencing summer, at a given latitude continental temperatures are generally *higher* than sea-surface temperatures; in the hemisphere experiencing winter, continental temperatures are generally *lower* than sea-surface temperatures.

This point is even more forcefully borne out in Figure 1.14c. The large areas of red and brown in the Northern Hemisphere and the great areas of blue and green in the Southern Hemisphere show that the greatest warming and cooling occurs over the continents. Seasonal changes of up to 30 °C occur over land in both hemispheres; by contrast, seasonal changes in sea-surface temperature, which are greatest in mid-latitudes, rarely exceed 8–10 °C.

■ Look at Box 1.4. Which one of the properties of water listed explains the contrast in seasonal temperature changes on land and sea?

□ The high specific heat of water, which results in the thermal capacity of the oceans being much greater than that of the continents. In other words, a much greater heat input is needed to raise the temperature of a mass of ocean by 1 °C than is needed to raise the temperature of the same mass of continental rock by 1 °C. As a result, continental areas heat up and cool down more quickly than oceanic areas.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(b)
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Arctic
23.4° N
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Equator
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Capricorn
66.6° S
Antarctic

(c)

◀ **Figure 1.14** (a) and (b) Daytime surface temperatures for January and July respectively, as measured by a satellite-borne radiometer (which measures thermal radiation emitted by the Earth). Surface temperatures generally decrease from low to high latitudes, but the **isotherms** (contours of equal temperature) do not run simply east–west (blue denotes low temperature). (c) Differences in surface temperature between January and July (large temperature differences are denoted by reds and browns). See text for explanation.

Box 1.4 Properties of water that are of importance for the climate

The relevance of some of these properties of water to the Earth's climate may already be obvious; the importance of others will become clear as you read further.

- At the range of temperatures found at the surface of the Earth, water can exist as a gas (water vapour), as a liquid, and as a solid (ice).
- Its specific heat, i.e. the amount of heat needed to raise the temperature of 1 kg of water (in any state) by 1 °C (or 1 K) is 4.18×10^3 J kg^{-1} °C^{-1}. This is the highest specific heat of all solids and liquids except ammonia.
- Its latent heat of fusion, i.e. the amount of heat needed to convert 1 kg of ice to water at the same temperature (and the amount of heat *given up* to the surrounding environment when 1 kg of ice forms) is 3.3×10^5 J kg^{-1}. This is the highest latent heat of fusion (or freezing) of all solids and liquids except ammonia.
- Its latent heat of evaporation, i.e. the amount of heat needed to convert 1 kg of liquid water to water vapour at the same temperature (and the amount of heat *released* to the surrounding environment when 1 kg of water vapour condenses) is 2.25×10^6 J kg^{-1}. This is the highest latent heat of evaporation (condensation) of all substances.
- It dissolves more substances, and in greater quantities, than any other liquid.
- It conducts heat more efficiently than most other liquids naturally occurring on the Earth.
- Its temperature of maximum density decreases with increasing salt content; pure water has its maximum density at 4 °C, but the density of seawater increases down to its freezing point at about −1.9 °C.
- The density of ice is less than that of water, the result of which is that ice occupies more space than the water from which it formed, and ice floats on water. (For most substances, the solid phase is denser than the liquid phase.)
- Compared with other liquids, it is relatively transparent.

The property that the oceans have of heating up and cooling down slowly is sometimes referred to as their 'thermal inertia'. It is part of the reason why the range of temperatures found in the oceans is less than half that which occurs on land and one of the principal reasons for zonal (east–west) variations in surface temperatures. In the next section, you will look at another reason for this temperature difference: atmospheric and oceanic circulation patterns.

1.4 The Earth's air-conditioning and heating systems

In Section 1.2.1, you saw that the Earth's radiation budget has an excess at low latitudes and a deficit at high latitudes (Figure 1.13). Figure 1.15 shows, very schematically, how heat is redistributed over the surface of the Earth. Put simply, the three principal processes involved are:

1 **Wind-driven surface currents**

Under the influence of winds, surface ocean water warmed at low latitudes flows polewards in surface currents, while that cooled at high latitudes flows equatorwards.

2 **Winds in the atmosphere**

Moving air takes up heat from the surface of the oceans and continents. Warm air rising at low-pressure regions such as the Equator and moving polewards in the upper troposphere transports heat from low to high latitudes, as does any warm air moving polewards. Net poleward heat transport results

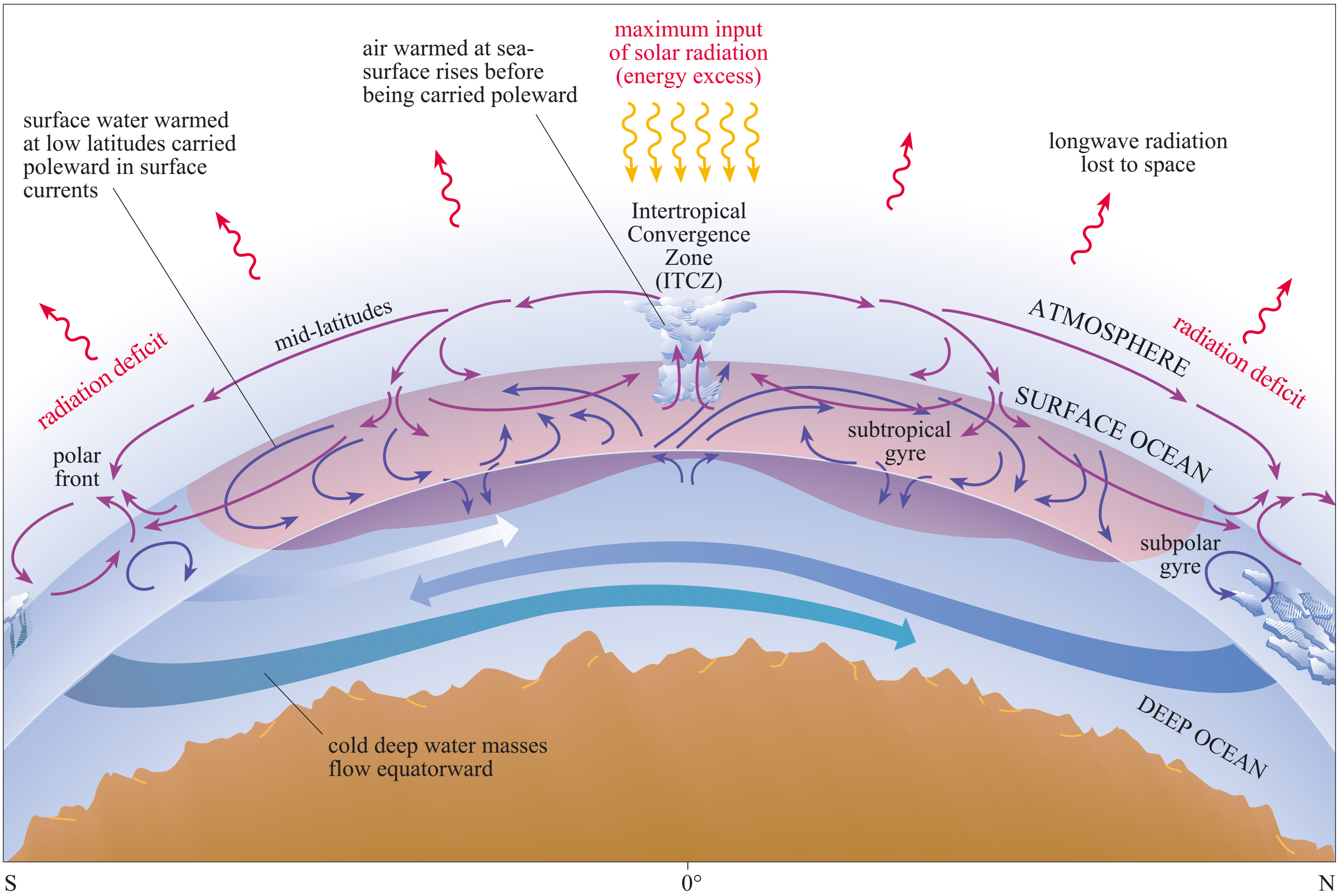


Figure 1.15 Schematic diagram of the Earth's heat-redistribution system (not to scale). The three interlinked circulatory systems are: wind-driven surface currents, winds in the atmosphere, and density-driven currents in the deep ocean. The **Intertropical Convergence Zone** (**ITCZ**) is the region where the wind systems of the two hemispheres meet. (See text for more details.)

because the air that moves towards the Equator at lower level to compensate is relatively cool from contact with ice and cold land and sea surfaces. Furthermore, warm, humid air from the Equator releases latent heat due to the condensation of water during its journey polewards.

3 **Density-driven currents in the deep ocean**

Surface ocean water cooled at high latitudes increases in density, sinks and flows equatorwards in the deep ocean.

Another way of looking at these three processes is as interrelated circulatory systems, largely driven by the transfer of heat and of momentum (i.e. mechanical or frictional forcing) at the ocean surface. An important feature of both ocean and atmospheric circulation systems is **convection**. When a pan of water is heated from below, heat is transmitted through the pan to the water at the bottom of the pan by conduction. Being heated, this water expands, becomes less dense and rises. On reaching the surface, the warmed water begins to lose heat to the air; it cools, becomes denser and sinks, then is heated again and rises, and so on (Figure 1.16), so forming convection ‘cells’. (You can sometimes see such cells when cooking spaghetti, because the spaghetti strands tend to become aligned with the flow.)

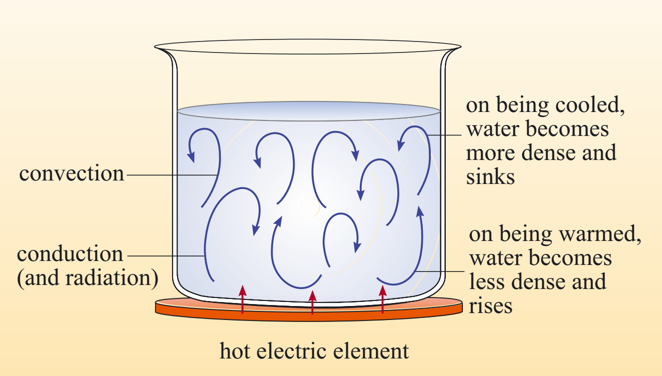


Figure 1.16 The circulatory pattern in a pan of water heated on an electric ring.

■ Does the convection occurring (i) in the atmosphere, and (ii) in the ocean (see Figure 1.15) resemble that occurring in a pan of water being heated (Figure 1.16)?

■ Convection occurring in the atmosphere does resemble that occurring in a pan of water, because it is driven by heating from below. Convection in the ocean is apparently driven by cooling from above, but technically the process is identical, as it is only the temperature difference between the top and bottom that matters.

1.4.1 Transport of heat and water by the atmosphere

Warm air rises or, to be more precise, air that is warmer than its surroundings (and is therefore less dense) rises. As the analogy with the pan of water demonstrates, it is the convective bulk mixing of water that distributes the heat supplied at the bottom of the pan so that eventually all of the water becomes warm. So it is for the atmosphere: when air is warmed by contact with a warm sea or land surface and rises, it is replaced by cooler air, which is warmed in turn. By contrast, the transfer of energy by conduction occurs at the molecular level (see Box 1.3) – if you had to rely on conduction to heat a pan of water, you would have to wait a very long time indeed.

What happens in practice, however, is complicated by two factors:

- air, like all fluids, is compressible
- air contains variable amounts of water vapour.

First consider compression: when a fluid is compressed, the internal energy that it possesses per unit volume by virtue of the motions of its constituent atoms, and which determines its temperature, is increased. Thus, a fluid heats up when compressed (a well-known example is the compression of air in a bicycle pump), and cools (i.e. undergoes a decrease in energy per unit volume) when it expands (this is what occurs in the cooling system of a refrigerator). Changes in temperature that occur in this way, and not as a result of a gain or loss of heat from the surroundings, are described as **adiabatic**. When warmed air rises, the atmospheric pressure it is subjected to decreases (see black curve in Figure 1.17), and so it expands and becomes less dense, it therefore cools by adiabatic expansion. It will only continue to rise as long as it continues to be warmer than its surroundings.

Imagine a parcel of air heated by contact with the ground and beginning to move upwards. Temperature decreases with height in the lower atmosphere (see dashed curve in Figure 1.17), but as long as the adiabatic decrease in temperature of a rising parcel of air is *less* than the decrease of temperature with height in the lower atmosphere, the rising parcel of air will be warmer and less dense than its surroundings and will continue to rise: the situation will be *unstable*, in the sense that convection, once initiated, is reinforced. Furthermore, any small random motions (that will always occur) will initiate convection. Such conditions are said to be ‘conducive to convection’. On the other hand, if the adiabatic cooling of the rising parcel of air is sufficient to reduce its temperature to below that of the surrounding air, conditions are said to be *stable*, in that the air will sink back to its original level and convection will be inhibited.

So far, we have been assuming that the rising parcel of air is dry, that is it contains no gaseous water vapour. Rising air, particularly over the ocean, may be saturated with water vapour or become saturated as a result of adiabatic cooling (warm air can hold more water vapour than cool air). Continued rise and associated adiabatic cooling result in cloud formation (Box 1.5). This condensation releases latent heat to the rising air (compare with Box 1.4), offsetting the effect of adiabatic cooling. In other words, humid air convects much more easily than dry air because condensation of water vapour releases additional heat energy, keeping the rising air less dense than the surrounding air for longer than would be the case for dry air.

Rising air warmed locally by conduction and/or convection becomes part of the global-scale atmospheric circulatory system shown schematically in Figure 1.15 (and Figure 1.20), whose horizontal flows over the surface of the Earth are the surface winds. Wind systems redistribute heat partly by the **advection** (bulk transport) of warm air masses into cooler regions (and vice versa), and partly by

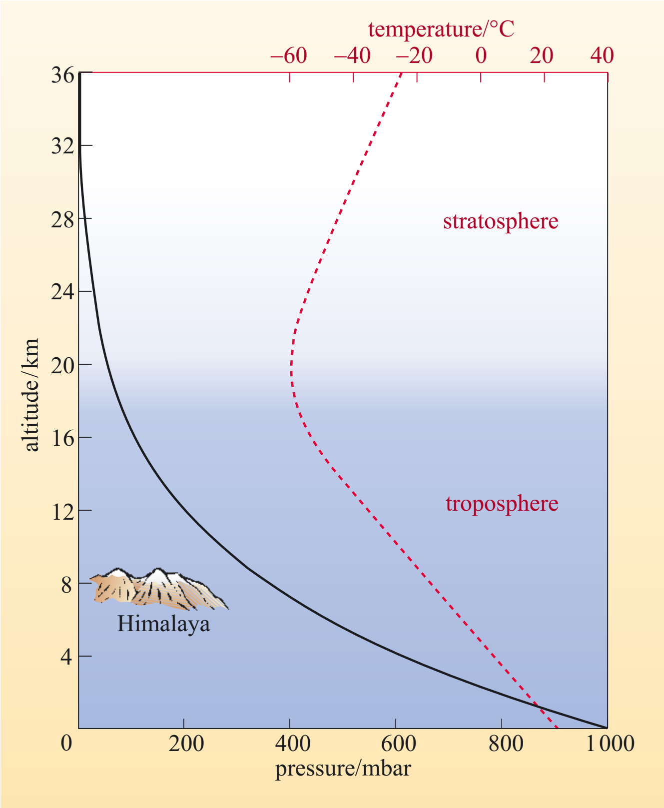


Figure 1.17 Schematic diagram to show how both pressure (solid curve) and temperature (dashed curve) decrease with increasing height in the lowermost atmosphere, or troposphere. Above that, in the stratosphere, temperature increases again. The curves are generalised, and intended to illustrate the general principle only. The thickness of the troposphere and actual values of temperature and pressure near the Earth’s surface vary with latitude and location; the temperature curve shown is appropriate for a latitude of about 30°. One millibar (mbar) is one-thousandth of a bar, where 1 bar is 1 atmosphere (atm).

the transfer of latent heat bound up in water vapour, which is released when the water vapour condenses to form cloud in a cooler environment, perhaps thousands of kilometres from the site of evaporation. Most of this moisture comes from the surface of the ocean; indeed, at any one time a large proportion of the water in the atmosphere (water vapour and clouds) has only recently evaporated from the tropical ocean, and poleward transport of warm humid air is the most important way in which heat from the ocean at low latitudes is transferred to higher latitudes.

The amount of evaporation from land depends on the moisture content of the exposed soil or rock. Vegetation is also a source of atmospheric moisture, both through simple evaporation from surfaces and through **transpiration**, whereby water drawn up from the soil by roots is lost to the atmosphere through pores in leaves; together, these two processes are known as **evapotranspiration**.

■ From which type of land areas would you expect transfer of latent heat to the atmosphere to be greatest?

■ From tropical rainforests, where evapotranspiration releases large amounts of water vapour to the atmosphere.

Indeed, being both warm and a good source of atmospheric moisture, rainforests behave climatically rather like the tropical ocean.

As mentioned earlier, heat is also transferred from the Earth's surface to the atmosphere by conduction and convection. This is referred to as transfer of **sensible heat** (heat which can be felt, or sensed) and it increases as the temperature difference between the Earth's surface and the overlying atmosphere increases.

■ By reference to Figure 1.14a–c, would you expect loss of sensible heat to the atmosphere to be greatest from land surfaces or from the ocean?

■ From land surfaces, especially at mid- and low latitudes in summer when the landmasses have heated up.

At low latitudes, sensible heat loss from land surfaces to the atmosphere is an order of magnitude greater than heat loss by evaporation. It is true that there is also a loss of sensible heat from the sea to the air, if the sea-surface is warmer than the air above it (which is the case more often than not), but as far as the ocean is concerned, *an order of magnitude more* heat is transferred to the atmosphere via evaporation and subsequent condensation than is transferred by conduction, even when enhanced by convection.

Cumulus and cumulonimbus clouds (over both the land and sea), are visible evidence of convection in the atmosphere (see Box 1.5). Cloud formation, the turbulence of atmospheric convection and the winds that redistribute heat over the Earth's surface are mostly confined to the lower atmosphere. This is known as the **troposphere** (from the Greek word *tropos* meaning 'turn', i.e. causing mixing), and makes up ~80% of the total mass of the atmosphere. Temperature decreases with height within the troposphere (Figure 1.17) up to its upper boundary (the *tropopause*) then begins to increase again in the overlying stratosphere because of absorption of radiation by ozone there. It is this temperature *inversion* that generally prevents convection in the lower atmosphere from reaching any higher, since rising air from below is almost always denser than warmer air above. There is some interchange of air between the troposphere and stratosphere, particularly in mid-latitudes, and in certain circumstances rapidly rising air masses can overshoot the tropopause. In general, however, stratospheric winds do not interact strongly with the winds of the lower atmosphere and so do not directly affect conditions at the Earth's surface. It is within the troposphere that the Earth's weather occurs.

Box 1.5 Clouds

Clouds form when water vapour in the atmosphere condenses around solid particles or nuclei (e.g. pollen grains, dust or salt from sea spray) to form tiny water droplets, or at greater heights and lower temperatures, ice crystals. Condensation can also occur on **aerosols**, minute droplets, often of sulfate compounds, notably sulfuric acid formed from water vapour and sulfur dioxide (SO_2) emitted from volcanoes or produced by industrial processes.

As pressure decreases with height in the atmosphere (solid curve in Figure 1.17), water droplets can exist in clouds at temperatures down to –12 °C, as a mixture of water droplets and ice crystals at temperatures between –12 °C and –30 °C, and predominately as ice crystals at temperatures less than –30 °C. Clouds consist entirely of ice crystals at temperatures below –40 °C.

As shown in Figure 1.18, clouds can have a variety of forms: for example, cumulus and cumulonimbus are associated with convection and have flat bottoms (marking the height at which condensation began to occur) and bubbly tops; stratus clouds form by more gentle uplift of air at a front or as a result of the presence of mountains and are layered. Wispy cirrus-type clouds are made of ice and are usually associated with flow in jet streams; cumulus-type clouds consist of water droplets, though the tops of towering cumulonimbus clouds consist of ice crystals. Stratus-type clouds, which at ground level are what we refer to as fog, can be made of either.

All clouds are highly reflective, being white in colour when illuminated (they only appear grey if their undersides are in shadow). However, the veil-like structure of cirrus clouds means that they have a fairly low albedo (<20%), while stratocumulus and cumulonimbus, which are thicker and consist of more densely packed droplets and/or ice crystals, have fairly high albedos of about 50% and 70%, respectively.

The role of clouds in the climate system is still unclear. Because of their high albedo, they may greatly reduce the amount of solar radiation reaching the Earth's surface, particularly at high and low latitudes (see discussion in connection with Table 1.1). As you will shortly see, however, they absorb radiation as well as reflect it. Last but not least, their complex three-dimensional shapes make their effect on the radiation budget much more difficult to quantify than, say, land ice. The role of clouds in the climate system will be discussed in more detail later in this chapter.

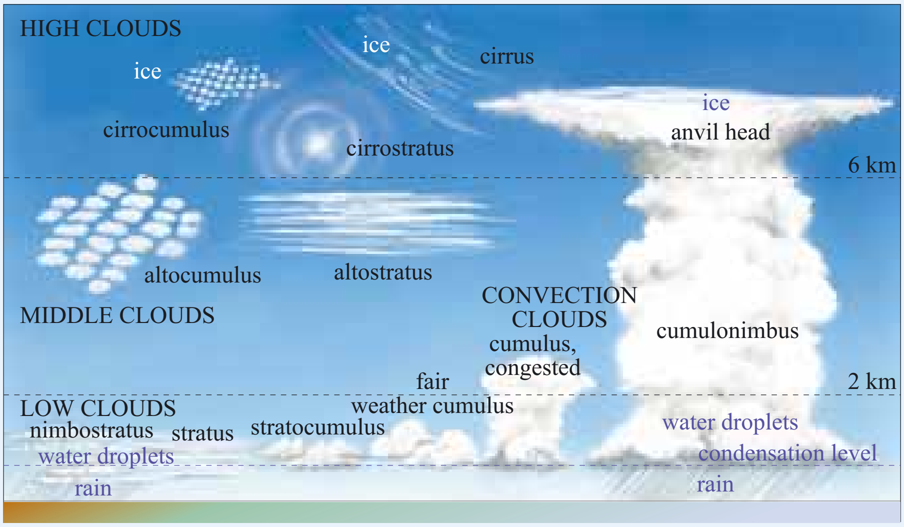


Figure 1.18 Various types of clouds and their characteristic levels and/or extents in the atmosphere.

Figure 1.19 shows how temperature varies with height in the atmosphere as a whole (compare with Figure 1.17 for the troposphere), and that above the troposphere and the stratosphere there are two more changes in temperature gradient within two more 'spheres', although it is generally thought that these outer 'spheres' are not important influences on the Earth's climate system. Part of the reason for this is that the atmosphere is so rarefied at these altitudes that the heat content of the air is very low, even when the temperature (determined by the vibrations of individual molecules, Box 1.3) is high. The increase of temperature in the stratosphere is caused by absorption of ultraviolet radiation there by oxygen and ozone. This absorption of energy results in that layer heating up directly, rather than by absorption of longwave radiation from below, as occurs in the troposphere.

Figure 1.20 shows what the wind system and surface atmospheric pressure pattern would be if the Earth were completely covered by ocean.

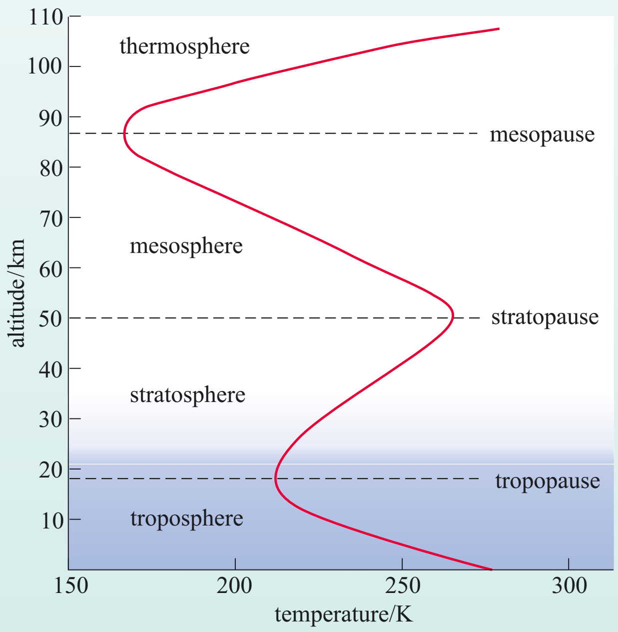


Figure 1.19 The vertical temperature 'structure' of the atmosphere, defined by the way that temperature varies with altitude. In each successive 'sphere', the temperature gradient is reversed: temperature decreases with height in the troposphere, so that conditions are conducive to convection; by contrast, temperature increases with height in the stratosphere, and vertical motions are suppressed. Note that, as indicated by the vertical air movements in Figure 1.20, the tropopause is higher at low latitudes (where it is at ~17 km) than at high latitudes (where it is at ~8–10 km).

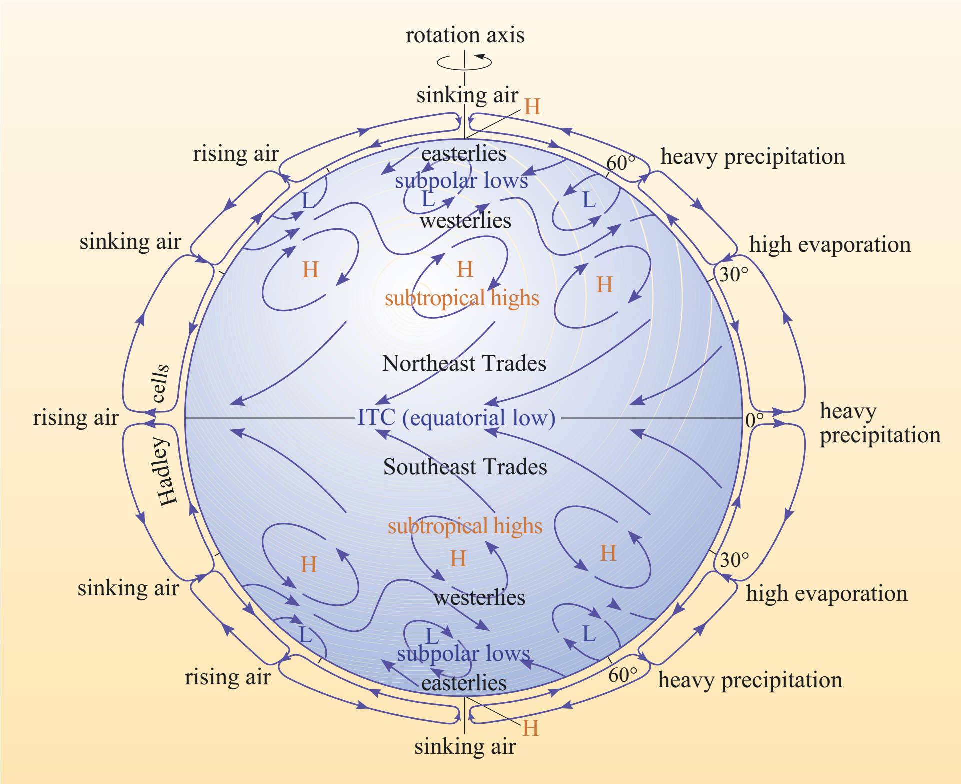


Figure 1.20 Wind system for a hypothetical water-covered Earth, showing the major surface winds and the zones of low and high pressure. Vertical air movements are indicated on the left-hand side of the diagram; characteristic surface conditions are given on the right-hand side.

Question 1.5

(a) By reference to Figure 1.20, describe how *surface* wind directions relate to regions of high and low surface pressure.

(b) By reference to the vertical air motions shown on Figure 1.20, describe how vertical air movements relate to regions of high and low surface pressure.

Air sinks and flows anticyclonically *outwards* from regions of high surface pressure, and flows cyclonically *inwards* and rises at regions of low surface pressure (Figure 1.21).

But why are the paths of winds curved? And why isn't the global wind system a simple convection system in which air sinks only at the poles, and surface winds blow directly from the poles to the Equator, as illustrated in Figure 1.22? The answer lies in the rotation of the Earth. Air masses moving above the surface of the Earth are bound extremely weakly to it by friction, and as they move, the Earth turns beneath them. The resulting deflection of winds relative to the surface of the Earth is known as the **Coriolis effect** and it applies equally to ocean currents as to the atmosphere. The effect increases with latitude, from no deflection at the Equator to a maximum at the poles. The Coriolis effect can be understood in terms of an imaginary Coriolis *force*, acting at right angles to the direction of flow:

- in the Northern Hemisphere, the Coriolis force acts to the right of a flow and, in the absence of balancing forces, turns it clockwise
- in the Southern Hemisphere, the Coriolis force acts to the left of the flow, and in the absence of balancing forces turns it anticlockwise.

The origin of the Coriolis effect is explained in more detail in Box 1.6.

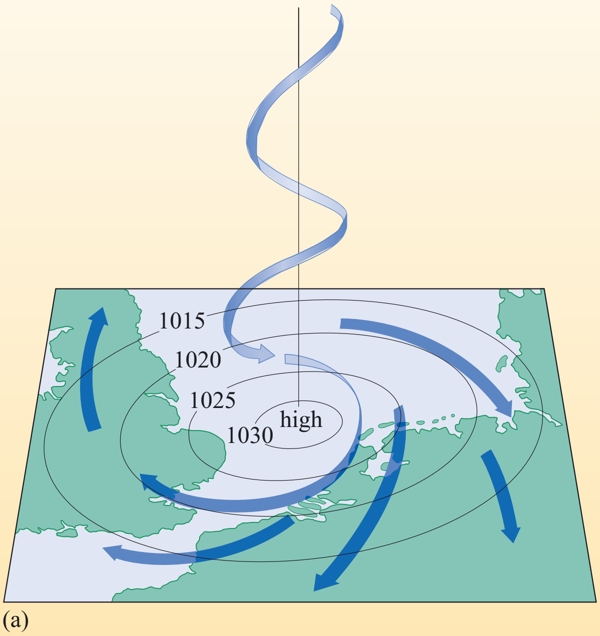


Figure 1.21 (a) Air spiralling downwards and outwards from an atmospheric high (an anticyclone). (b) Air spiralling inwards and upwards towards an atmospheric low (a cyclone or depression). Note that the flow directions shown are for the Northern Hemisphere; in the Southern Hemisphere, flow around anticyclones is anticlockwise, and flow around cyclones is clockwise (Figure 1.20). Contour values are typical atmospheric pressures at sea level, expressed in millibars (10^{-3} bar). Surface wind speeds typically reach a few tens of m s^{-1}.

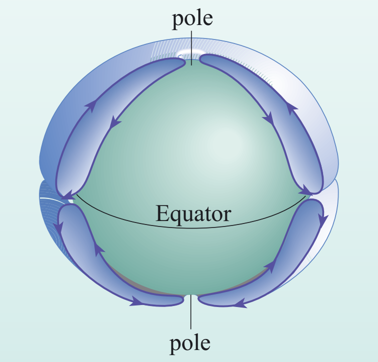


Figure 1.22 Simple hypothetical wind system for a non-rotating Earth.

Box 1.6 The Coriolis effect

An object which is stationary relative to the Earth is, in fact, moving in a circle as the Earth rotates. If it starts to move relative to the Earth, its total velocity is the sum of two parts:

- its velocity relative to the Earth
- the circular velocity due to the Earth's rotation.

According to Newton's laws, a change of velocity is by definition an acceleration and therefore requires a force.

Now consider the situation illustrated in Figure 1.23. If the velocity of the object is constant relative to the Earth, then the velocity itself is also moving in a circle because of the Earth's rotation, but this constitutes a change of absolute velocity, and hence an acceleration. This is the Coriolis effect. Technically, the effect is doubled because the velocity of the object itself relative to the Earth changes the speed and radius of the circular motion and again, this amounts to a change of velocity and hence an acceleration.

In the ocean and atmosphere, there is no force to supply the acceleration (which is to the left of the motion in the Northern Hemisphere), and hence water and air do not generally move with constant velocity relative to Earth, but show a deflection to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. Thus, in the absence of

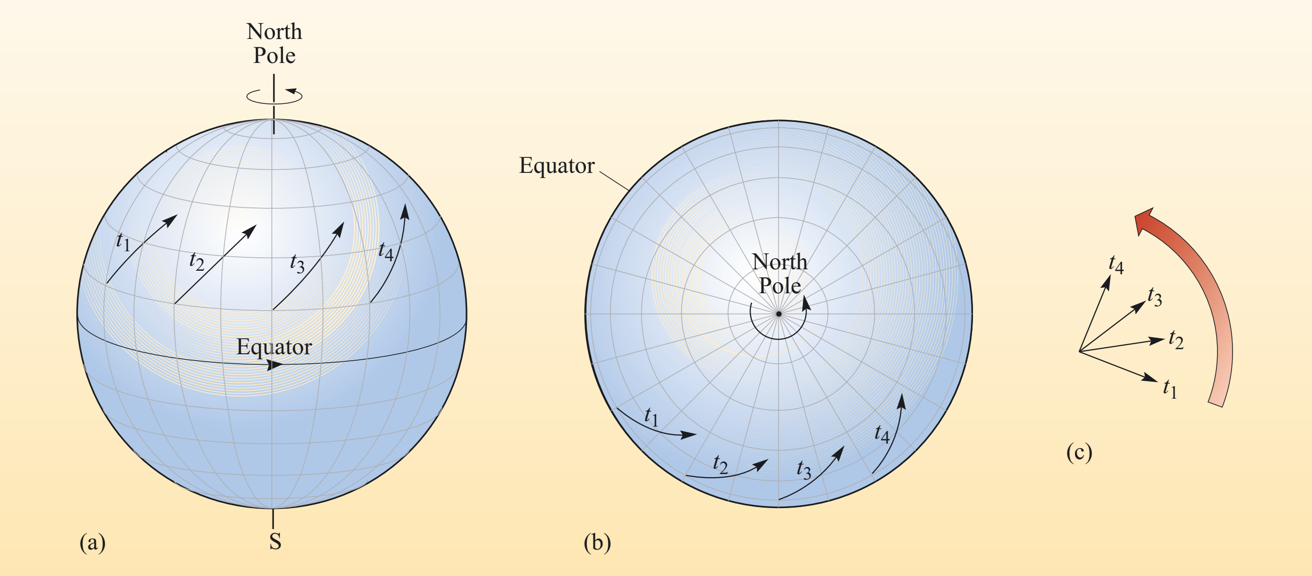


Figure 1.23 As the Earth rotates, the velocity vector of an object moves and changes, even for an object with constant velocity relative to the Earth: (a) shows the velocity of such an object at successive times, assuming, for convenience, it moves a small distance relative to the Earth; (b) shows the same scene looking down on the North Pole, illustrating how the velocity rotates towards the axis; (c) again shows the change in velocity vector, but relative to a single point. This change amounts to the Coriolis effect.

a *real* force to supply the acceleration, there is a deflection (as if by an *imaginary* force) in the opposite direction. Note that it is only the horizontal component of the acceleration that is relevant, because any vertical part is swamped by other forces.

The change in velocity is always towards the axis of rotation; thus, on the Equator it has only a vertical component and is thus irrelevant.

Note, too, that the Coriolis effect is only felt for large horizontal scales or rapid rotations. The familiar centripetal effect, which you feel when driving round a bend, for instance, is not important for understanding global-scale flows because you simply experience it as a modification to gravity.

■ So, what can be said about the Coriolis effect for motion over a hypothetical *cylindrical* Earth, rotating on its axis as shown in Figure 1.24?

■ For a rotating cylindrical Earth, there would be no visible Coriolis effect, except at the flat ends of the cylinder.

Since the acceleration is always towards the axis, the horizontal deflection on a cylinder would be zero, as it is on the Equator, and both Coriolis and centripetal accelerations would simply amount to changes in the effective force of gravity. The ends, however, would still experience deflections, as on a sphere.

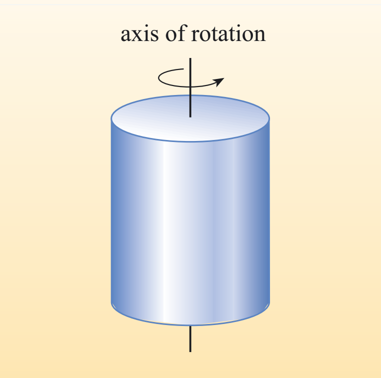


Figure 1.24 Hypothetical cylindrical Earth, spinning about its axis.

The characteristic conditions indicated at the right-hand side of Figure 1.20 are a result of the fact that while sinking air is generally dry, rising air often has a high moisture content. The most dramatic common manifestations of rising (i.e. convecting) moist air are cumulonimbus clouds, the tallest of which form at low latitudes in the Intertropical Convergence Zone (ITCZ), where the wind systems of the two hemispheres meet (Figures 1.15 and 1.20). Here, moist air carried in the Trade Winds converges and rises, resulting in the formation of towering thunderheads.

As a result of the Coriolis effect, the Trade Winds blow towards the Equator not from the north and the south, but from the northeast and the southeast. The convention for naming winds is to use the direction *from* which they blow, and so these winds are known as the Northeast Trades and the Southeast Trades. As mentioned above, in the lower atmosphere, pressure is low along the Equator (or, more accurately, along the ITCZ, which corresponds more or less to the region of highest surface temperature), and here air converges and rises. It then moves polewards in the upper troposphere, with much of it sinking in mid-latitudes where, as a result, the pressure at the Earth's surface is high. Thus, the Trade Winds are the surface expression of a helical circulatory system known as the **Hadley circulation** or Hadley cells (Figure 1.25). Strictly speaking, only the latitudinal-vertical component of the circulation, as shown on Figure 1.20, is named the Hadley circulation.

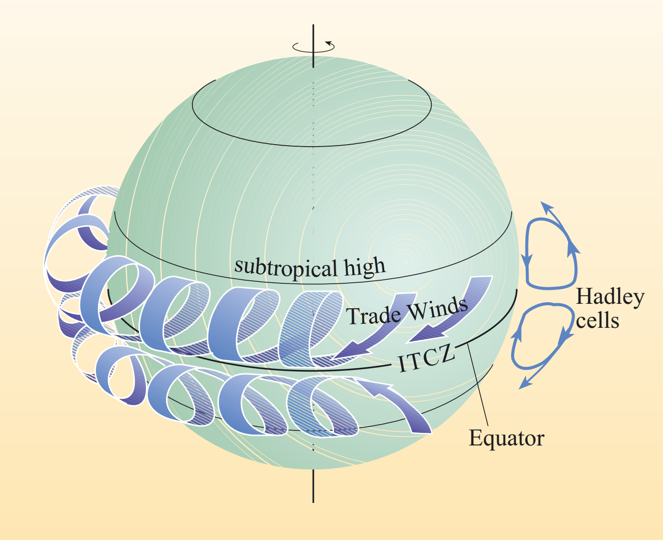


Figure 1.25 The helical circulation patterns of which the Trade Winds form the surface expression; the north–south component of this helical circulation is known as the Hadley circulation; the two 'Hadley cells' can be seen on either side of the Equator (see also left-hand side of Figure 1.20).

As a result of the Coriolis effect increasing with increasing latitude, there is a change in the *style* of atmospheric circulation with latitude. As the Trade Winds blow in tropical latitudes where the Coriolis effect is relatively small, they are deflected laterally only slightly. At higher latitudes, where the degree of deflection is much greater, vortices tend to form in the lower atmosphere with a predominantly horizontal, or slantwise, circulation. As shown in Figure 1.21, these are the anticyclonic and cyclonic winds familiar from weather charts.

- ■ How can mid-latitude cyclones and anticyclones contribute to the poleward transport of heat?
- ■ Moving air masses mix with adjacent air masses along *atmospheric fronts*, and heat is exchanged between them. Air moving northwards in a Northern Hemisphere cyclone or anticyclone will transport relatively warm air polewards, while air returning equatorwards will have been cooled. This is, in effect, a kind of large-scale stirring, and is shown schematically in Figure 1.26.

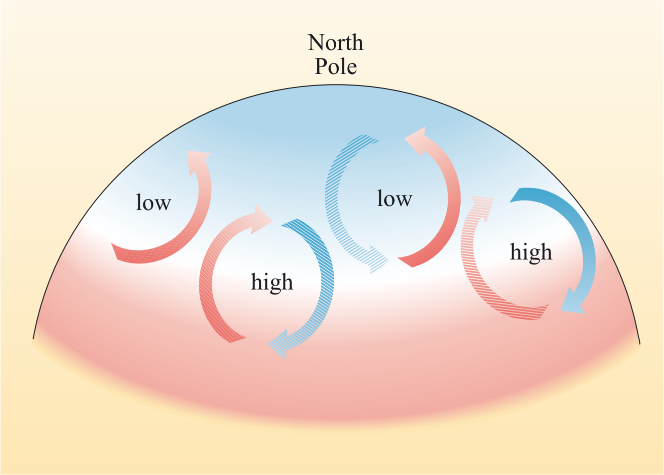


Figure 1.26 Highly schematic diagram to show the poleward transport of heat through the action of mid-latitude cyclones and anticyclones.

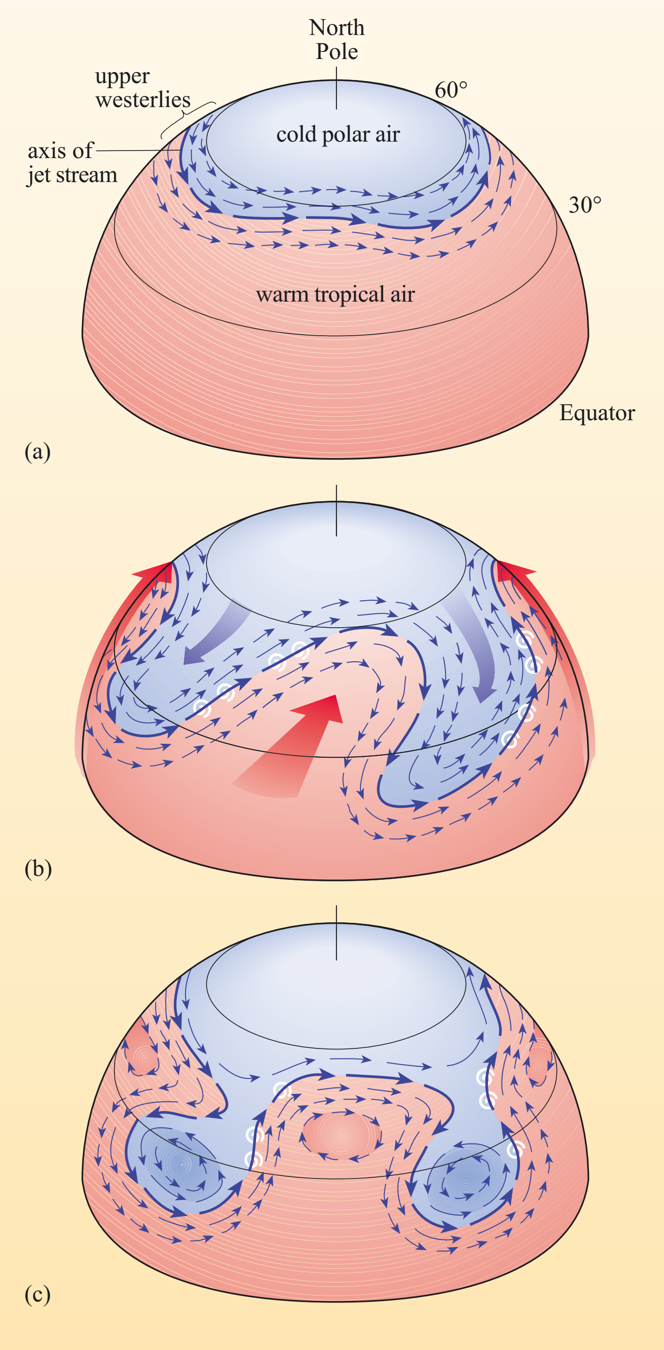


Figure 1.27 Schematic diagram showing stages in the development of waves in the northern polar jet stream, which flows eastwards along the polar limit of the upper westerlies near the tropopause at heights of ~10 km: (a) the jet stream begins to undulate; (b) waves become more extreme; (c) large cells of polar and tropical air become isolated. Meanwhile below, cyclones (i.e. depressions, shown here as white 'swirls' of cloud) form along the poleward-trending parts of the front, and anticyclones (not shown) form along the equatorward-trending parts.

The paths taken by mid-latitude depressions and anticyclones are determined by the behaviour of the **polar jet stream** (Figure 1.27). This is a high-level, fast air current that in each hemisphere flows around the Earth above the boundary between warm tropical air and the underlying cold polar air, known as the **polar front**. The jet stream tends to develop large undulations, typically three to six in number, which eventually become so extreme that cells of tropical air become isolated at relatively high latitudes, and cells of polar air become isolated at relatively low latitudes (Figure 1.27c). This mechanism results in the horizontal transport of enormous amounts of warm air polewards and cold air equatorwards. Meanwhile, the vertical air movements that of necessity accompany these large-scale horizontal wave motions (known, incidentally, as *Rossby waves*) lead to the development of cyclonic flow and low-pressure centres, and anticyclonic flow and high-pressure centres near the Earth's surface.

■ What would be the effect of the undulations in the jet stream becoming more pronounced?

□ Cold polar air would flow to lower latitudes than normal, while warm tropical air would flow to higher latitudes than normal.

The areas of the globe that are affected by pronounced undulations in the jet stream depend on the configuration of the jet stream at the time; the undulations themselves travel westwards around the Earth, relative to the air flow. In recent years, unusual behaviour of the jet stream (perhaps related to global warming) has resulted in droughts and crop failures in the United States. The effect of droughts on agriculture, whether in the United States or the African Sahel, is a reminder that whether a particular type of vegetation flourishes or dies is determined by the prevailing climatic conditions. Natural ecosystems are more robust than modern agricultural monocultures, but they are attuned to the prevailing climate, and if that changes, then so will they.

Now look at Figure 1.28, which shows the prevailing winds at the Earth's surface. Note the zones of westerly winds between the subpolar low-pressure regions and the subtropical highs. These are the parts of the globe that, depending on the configuration of the polar jet stream (Figure 1.27), may be affected by cold polar air or warm tropical air. These regions are often referred to as temperate.

■ Compare the wind patterns in Figure 1.28 with those shown in Figure 1.20. Over which parts of the globe are the subtropical highs and subpolar lows most evident? What features of the real world cause the actual wind pattern to differ from the hypothetical one shown in Figure 1.20?

■ Subpolar lows and subtropical highs are most evident over the oceans; this is not surprising given that Figure 1.20 shows the wind pattern appropriate to an Earth completely covered in water. It is the presence of the continental landmasses that makes the real wind pattern so much more complicated.

The presence of the continents 'distorts' the global wind system from the hypothetical east–west zones of polar high/subpolar low/subtropical high/ equatorial low. Question 1.6 investigates the effects of the distribution of land and sea on the global wind system in more detail.

Question 1.6

The ITCZ, where the wind systems of the two hemispheres meet (Figure 1.28), generally follows the zone of highest temperature at the Earth's surface.

(a) In general terms, what are the main differences between the position and/or shape of the ITCZ in July (Figure 1.28a) and January (Figure 1.28b)? How do Figure 1.14 and the related discussion help to explain these differences?

(b) The ITCZ moves seasonally northwards and southwards between the extreme positions shown in Figure 1.28. What implication does the changing position of the ITCZ have for the prevailing wind direction at the Earth's surface? (To answer this, describe how the wind direction changes over tropical West Africa (Mauritania, Senegal and Guinea) between July and January.)

(c) Over what region is the north–south seasonal shift in the position of the ITCZ greatest?

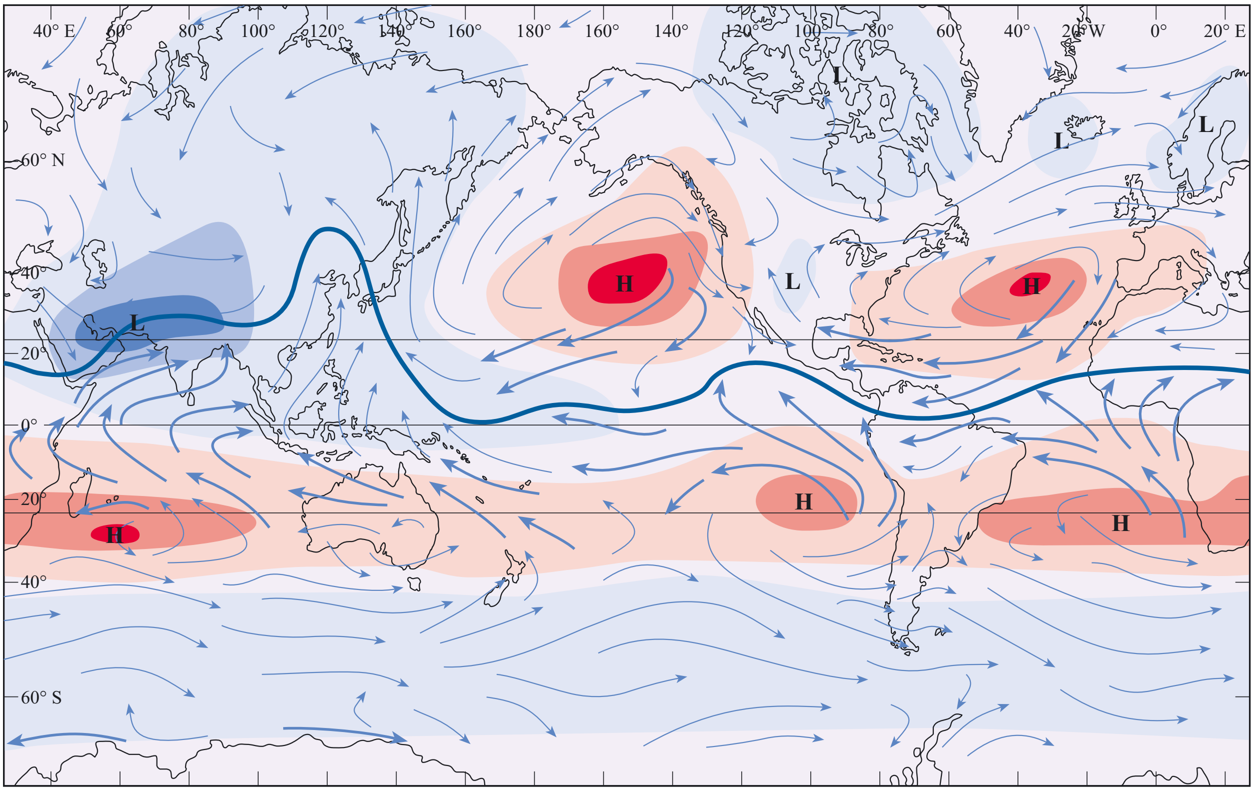
Many low-latitude regions experience such seasonal changes in wind direction. Air masses that travel over the oceans pick up moisture, while those that move over arid regions are dry, so seasonally reversing winds often bring with them contrasting climatic conditions (Figure 1.29).

■ Bearing this in mind, how might the conditions in the south of the Eurasian continent differ between July and January?

■ Conditions will be wet in July when the prevailing winds are south-westerlies blowing off the Indian Ocean and dry in January when the prevailing winds are north-easterlies blowing off the Eurasian continent.

The rains that fall over southern Asia in the summer are often referred to as the monsoon rains. However, as the word 'monsoon' derives from the Arabic for 'winds that change seasonally' there are really two Asian monsoons:

- the warm, moist South-West Monsoon
- the relatively cool, dry North-East Monsoon.



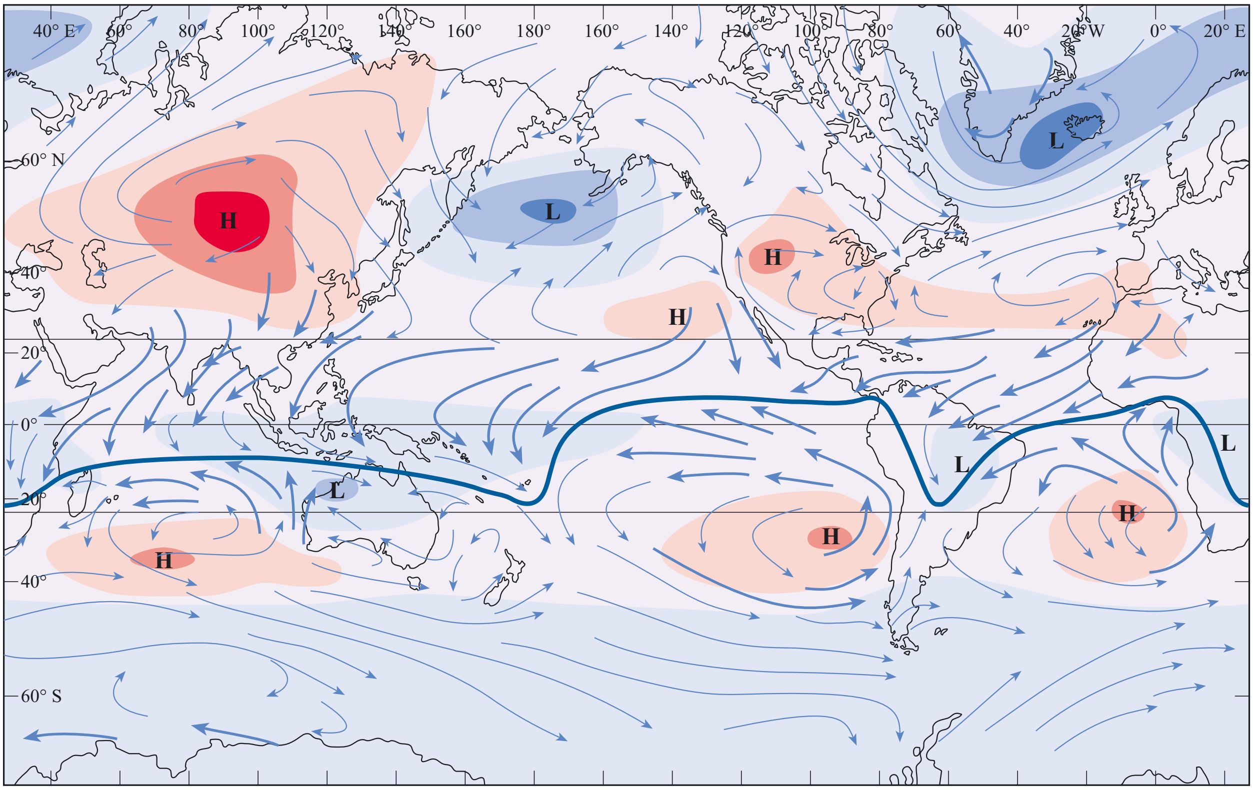
(a) JULY

KEY

mean position of ITCZ

most frequent wind direction

prevailing wind direction (≥50% of observations)



(b) JANUARY

◀ **Figure 1.28** The prevailing winds at the Earth's surface, and the position of the Intertropical Convergence Zone (ITCZ) where the wind systems of the two hemispheres meet, in (a) July (northern summer/southern winter) and (b) January (southern summer/northern winter). Also shown are the positions of the main regions of high (H) and low (L) surface pressure for these seasons of the year.

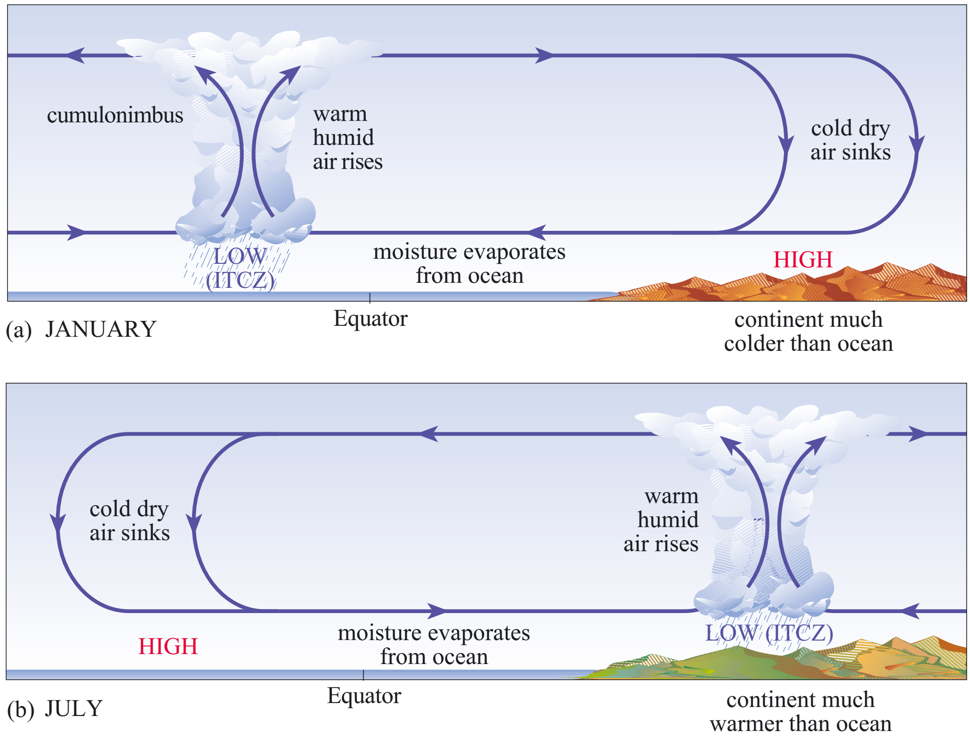
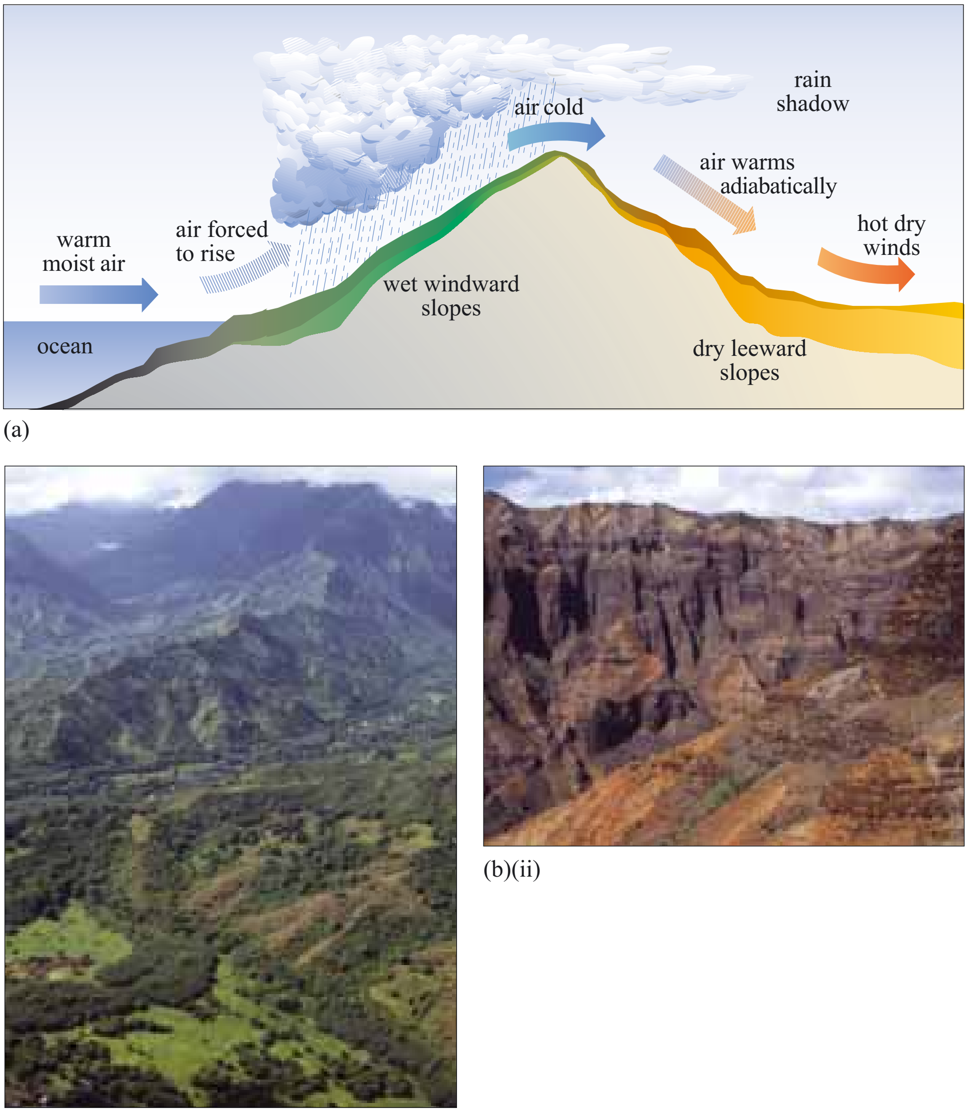


Figure 1.29 Schematic diagrams to illustrate the changing conditions over continent and ocean in a monsoon climate, drawn assuming that the continent is in the Northern Hemisphere. (a) Northern winter: dry, cooled air subsides over the continent, which is a region of high surface pressure; winds blow off the continent picking up moisture from the much warmer ocean and eventually moist air rises at the Intertropical Convergence Zone, causing abundant rain. (b) Northern summer: land is now much warmer than the ocean, and so the region of low surface pressure corresponding to the ITCZ and its zone of rain has moved northwards towards the interior of the continent. (*Note*: the latitudes shown are notional, as the northernmost and southernmost positions of the ITCZ vary from place to place, as can be verified from Figure 1.28, which should be studied in conjunction with this figure.)

It has been worth considering monsoonal reversals in some detail because, as will be seen later, the interaction between winds and continents may also have a more subtle and longer term influence on global climate.

The large-scale reversals in atmospheric circulation over southern Asia and the Indian Ocean are particularly marked because the continental landmass involved is not only large, but in places very high. In general, it is important to remember that climatic conditions are affected not only by the distribution of land and sea, but by the shape of the landmasses in the vertical dimension (i.e. their topography). Air that is forced to rise over high ground may be triggered to convect, leading to rainstorms. Furthermore, moisture-laden air that is forced to rise by the presence of a topographic high will cool adiabatically and, being cooler, will no longer be able to hold as much moisture. If cooling is sufficient, condensation will occur, clouds will form and rain or snow will result (Figure 1.30a). By contrast, air subsiding over the leeward slope will warm adiabatically and having no moisture source may become very dry, resulting in the formation of a *rain shadow*. Rain and snowfall triggered by mountains (known as *orographic* precipitation) occurs at all scales – from small volcanic islands in mid-ocean to mountain chains on land. In southwestern USA, the western slopes of the Coast Range and the Sierra Nevada receive ample precipitation from the moisture-laden north-westerlies blowing off the Pacific (Figure 1.28); to the

Figure 1.30 (a) Orographic precipitation: mountains cause moisture-laden air to rise, with the result that the windward slopes are wet and well vegetated, while to leeward is an area of rain shadow. (b) Two locations on the Hawaiian Island of Kauai: (i) the verdant eastern side of the island, thought to be one of the wettest places on Earth; (ii) Waimea Canyon on the western side, which has a much drier climate and desert-like vegetation.



(a)

(b)(ii)

(b)(i)

southeast of these mountains, in their lee, are the desert regions of Nevada and eastern California (also visible on Figure 1.1). Figure 1.30b shows the dramatic effect of orographic precipitation and rain shadow on the vegetation of the mountainous island of Kauai.

Before moving on to look more closely at the role of the ocean in the climate system, it is important to remember the influence that living terrestrial organisms have on their environment. This is most dramatically illustrated by the effects of deforestation, so often shown on our television screens. Removal of large areas of rainforest means that rainwater is no longer trapped but runs away, carrying with it the topsoil (itself largely a product of the forest). As a result, productive ecosystems that trapped and recycled water, and provided moisture and heat to the overlying atmosphere, are all too often converted to arid wastes.

1.4.2 Heat transport by the upper ocean

As shown by the summary diagram in Figure 1.15, the ocean has a surface current system that is primarily wind driven and a deep current system driven primarily by heating and cooling at the surface. The surface current system is confined to the uppermost layers of the ocean, which are separated from deeper colder water by a zone of marked decrease in temperature known as the **thermocline**.

Comparisons between Figure 1.31, which shows the average surface current pattern, and the winds on Figure 1.28 (particularly (b)), reveal a fairly close correspondence between the two, although current patterns are of necessity modified by the presence of landmasses, which cause flow within ocean basins to form more-or-less closed circulatory systems, or **gyres**. In the Atlantic, anticyclonic wind systems blowing around the high-pressure regions in mid-latitudes give rise to anticyclonic gyres (often referred to as subtropical gyres), in both hemispheres; similar subtropical gyres can be seen in the Pacific.

■ To what extent can cyclonic gyres be identified that correspond to the cyclonic winds associated with subpolar low-pressure regions (Figure 1.20)?

■ Cyclonic gyres can be seen in the subpolar regions of both the North Atlantic and the North Pacific, but *not* in similar latitudes in the Southern Hemisphere. In the Southern Ocean, where there are no land barriers, the westerly winds drive the Antarctic Circumpolar Current eastwards around the globe.

To see how surface currents influence the distribution of heat, compare Figure 1.31 with Figure 1.14a and b, showing surface temperature. Although it is quite hard to see from Figure 1.14, sea-surface temperatures in low to mid-latitudes are generally higher on the western sides of the Atlantic and the Pacific than on the eastern sides.

■ Why is this?

■ Water warmed at low latitudes is carried polewards in the western ‘limbs’ of the subtropical gyres. Although some water is carried to high latitudes, some must also circulate in the subtropical gyres for a considerable time. However, because flow is anticyclonic, the western sides of the gyres will have been warmed more recently than the eastern sides.

One of the clearest manifestations of surface current flow in the temperature distributions in Figure 1.14 is the shape of the isotherms in the North Atlantic, which follow the northeastwards flow of the Gulf Stream (Figure 1.31). By contrast, in the South Pacific, the finger of low temperatures extending northwards up the west coast of South America (particularly clear in Figure 1.14b) may be correlated with flow in the cold Peru or Humboldt Current, carrying water cooled at high latitudes equatorwards; a similar though less-marked influence on the temperature distribution may be seen in the South Atlantic associated with the Benguela Current (Figure 1.31).

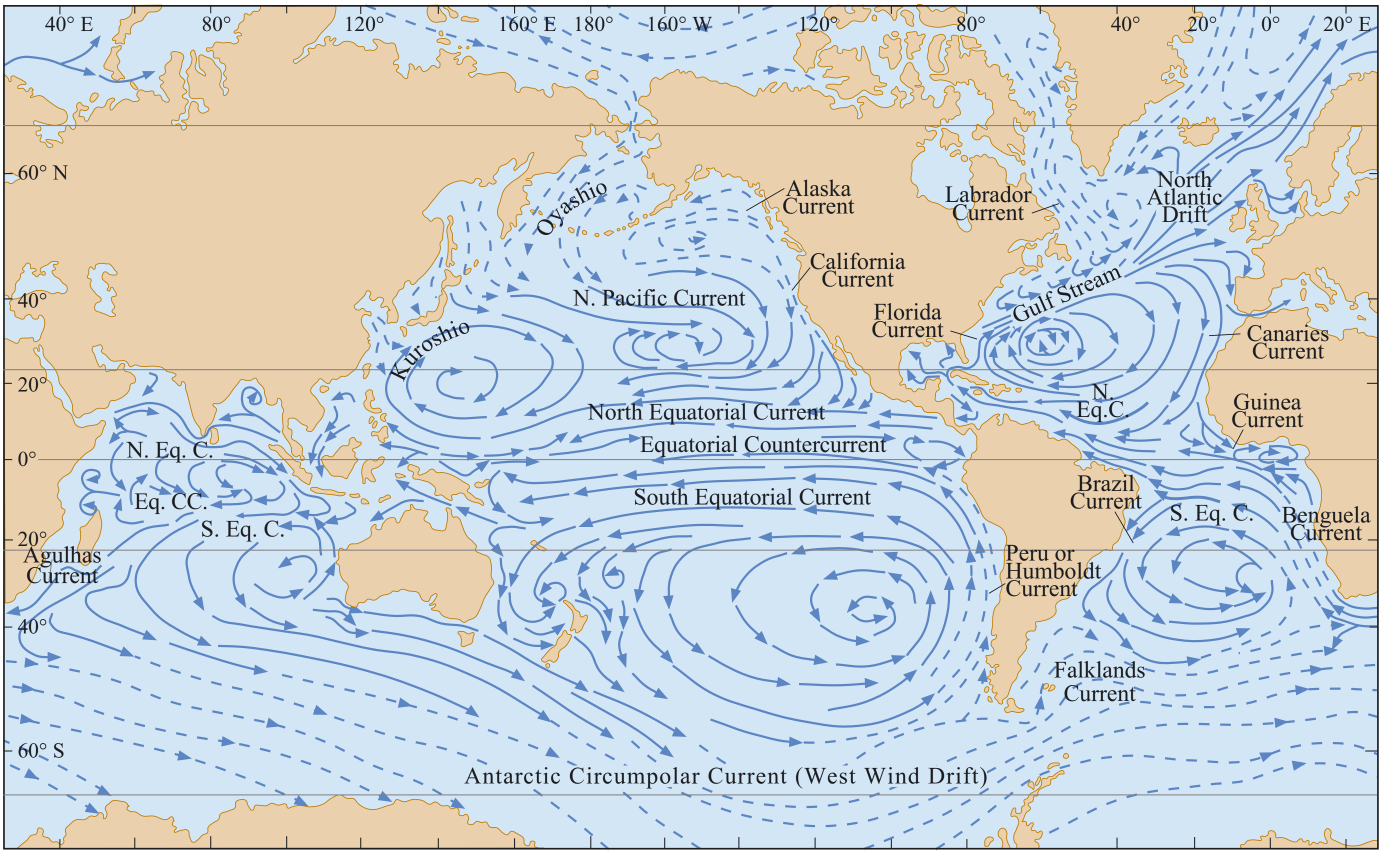


Figure 1.31 The global surface current system in the northern winter; this is the long-term average pattern – at any one time, the pattern will differ in detail. There are local differences in the northern summer, particularly in regions affected by monsoonal reversals; see also Figure 1.14. Cold currents are dashed. Note that even in strong currents, such as the Gulf Stream, current speeds rarely exceed a few m s^{-1}.

The flow of warm water northeastwards across the North Atlantic in the Gulf Stream moderates the climate of Britain and northwest Europe, making the region much warmer than it would otherwise be. The Gulf Stream is the narrow, fast-flowing western side of the North Atlantic subtropical gyre; similar western boundary currents are found in all the ocean basins. They owe their intensity to the fact that the Coriolis effect increases with latitude. Were it not for the speed at which the Gulf Stream flows polewards, much more of its heat would be lost to the atmosphere and adjacent ocean en route, and so be unavailable to warm higher latitudes.

The gyres and the Gulf Stream, and other current patterns on maps like Figure 1.28, are examples of large-scale, long-term features of the oceanic circulation. They represent the *average* situation in the ocean (at any one time the current in a particular place could be flowing in the opposite direction to that shown on the current map), and are often thought of as the ocean's climate. The ocean also has 'weather' (i.e. short-term, smaller-scale features). Most significant in this context are eddies about 50–250 km across (i.e. about a quarter of the size of mid-latitude atmospheric cyclones and anticyclones), which because of their intermediate size are known as **mesoscale eddies**.

Such eddies are frequently generated by meanders in strong currents like the Gulf Stream or the Antarctic Circumpolar Current. As these currents flow along boundaries between bodies of water with contrasting temperatures, they are known as **frontal currents**; the formation of eddies from meanders in frontal currents is similar to the formation of weather systems along atmospheric fronts (Figure 1.32). Mesoscale eddies clearly play an important role in transporting heat around the oceans, particularly across oceanic frontal regions, but until recently, computer simulations of the ocean have not been of sufficiently fine resolution to include realistic representation of the eddies. Indeed, computer modelling of global oceanic flows poses particular problems because the high density of water compared to air means that the length-scales of current patterns are generally shorter and the timescales longer than is the case for winds.

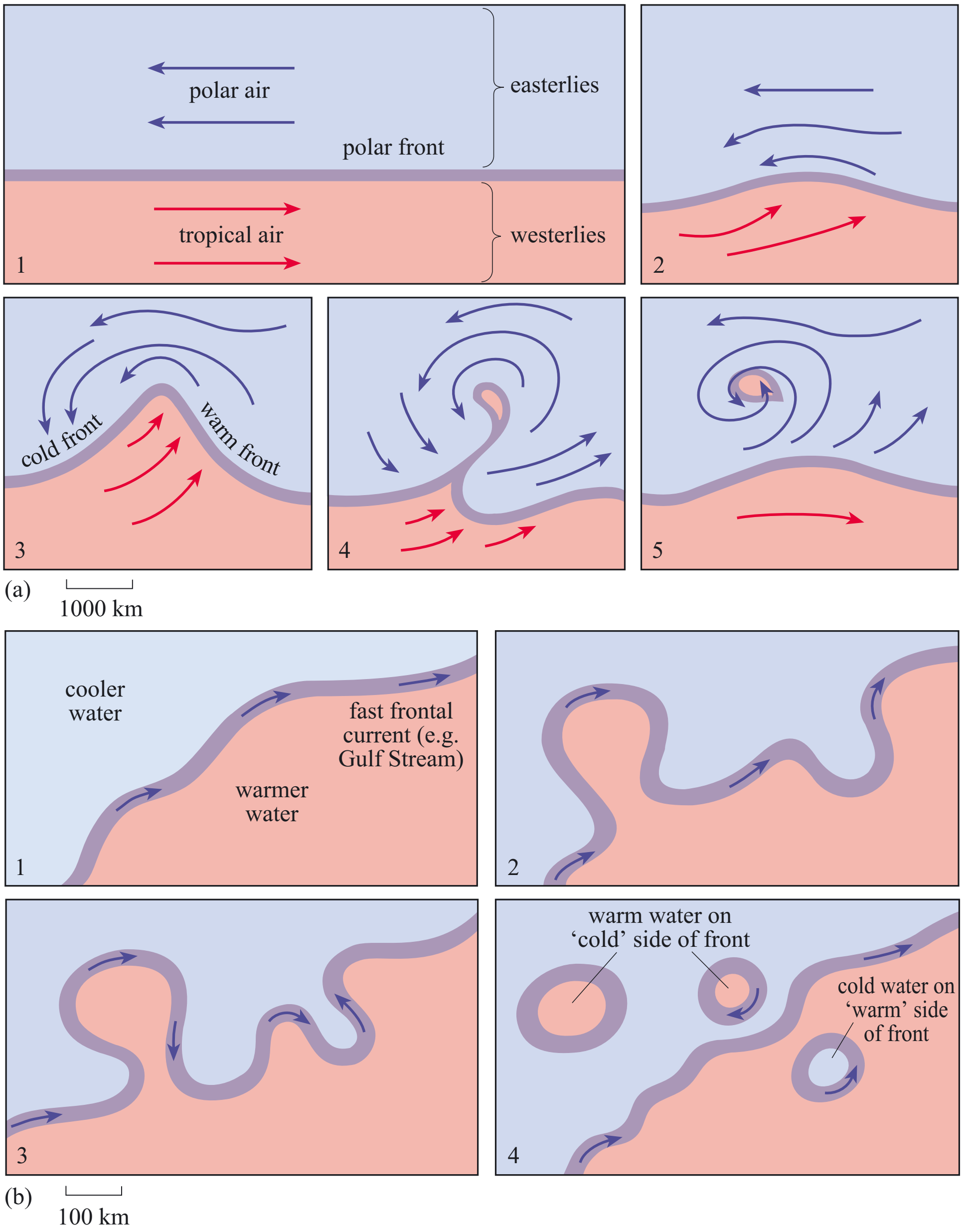


Figure 1.32 Plan view diagrams showing weather in the atmosphere and in the ocean. (a) Stages in the development of a mid-latitude cyclone (depression) in the Northern Hemisphere, showing how it contributes to the poleward transport of heat. (b) The formation of mesoscale eddies at a strong frontal current like the Gulf Stream, showing how they act to transport heat from one side of an oceanic front to the other.

1.4.3 The thermohaline circulation

The previous section considered how the ocean can transport heat by horizontal recirculations in the upper ocean, but the ocean also has vertically overturning circulations similar to the Hadley cells of the atmosphere. Where these overturning cells consist of relatively warm upper-layer water moving polewards above relatively cool deep water moving equatorwards, as is most often the case, the net effect is a poleward heat transport. Although the flows are far slower than in the atmosphere, the much greater heat capacity of water means that enormous quantities of heat can be transported. The properties of the slowly moving deep ocean currents are therefore surprisingly important for climate. These properties are mostly set in high latitudes in winter where, in certain locations, cold winds cool surface water to such an extent that it becomes denser than the water beneath it and sinks. This displaces the water beneath it, which in turn rises to be cooled. Convection cells are set up and the deep mixing leads to the formation of homogeneous bodies of water known as **water masses**. In particular regions, the water may become well mixed all the way to the seabed. The dense water masses that result from deep mixing circulate at great depths in the ocean and are known as *deep* (or *bottom*) *water masses*.

Figure 1.33 shows the main features of the deep circulation in the Atlantic. The two northward-flowing water masses shown are known as Antarctic Bottom Water and Antarctic Intermediate Water; both of these water masses extend northwards in all three ocean basins. The Antarctic Bottom Water flows along the seabed, and Antarctic Intermediate Water flows at a depth of 1000 m or so.

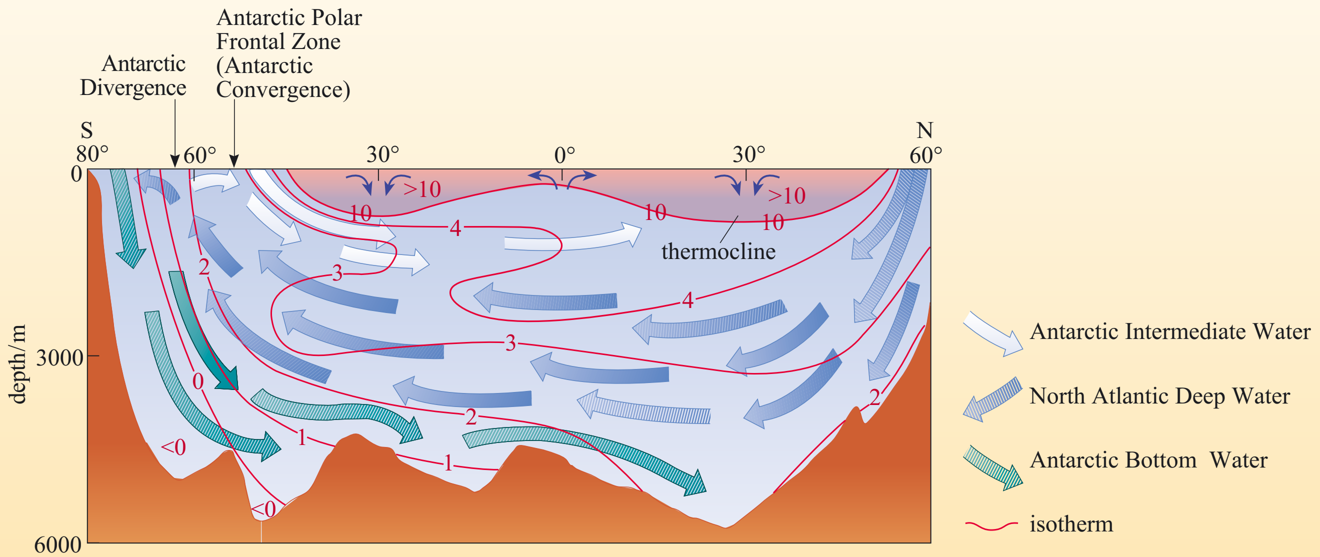


Figure 1.33 Generalised north–south cross-section of the Atlantic between 80° S and 60° N, showing flow of main water masses and resulting temperature distribution; values on the isotherms are in °C. At the Antarctic Polar Frontal Zone (or Antarctic Convergence), Antarctic surface waters converge with warmer subtropical waters and sink beneath them: at the Antarctic Divergence, water that sank in high northern latitudes eventually reaches the surface again. Small downward arrows near the surface at about 30° of latitude represent convergence in the subtropical gyres, and small upward arrows near the surface at the Equator represent upwelling at the Equatorial Divergence (see Figures 1.36 and 1.34). *Note*: vertical scale is greatly exaggerated.

Observation of the temperature contours between about 40° S and the Equator reveals that the Antarctic Intermediate Water may be distinguished as a 'tongue' of water that is cooler than both the overlying surface water and the underlying deep water mass – North Atlantic Deep Water.

■ Why, at first sight, would this seem to be an unlikely situation?

■ Cooler water is normally denser and would therefore be expected to mix very rapidly with the warmer water by convection.

The solution to this conundrum is that the Antarctic Intermediate Water is relatively fresh (for seawater), both because it forms in the subpolar regions where precipitation is high (Figure 1.20), and because it contains ice meltwater. By contrast, North Atlantic Deep Water is relatively saline; that is to say, its concentration of dissolved salts (its **salinity**) is relatively high. So, despite being cooler than North Atlantic Deep Water, Antarctic Intermediate Water is less dense and flows above it. Deep circulations in the ocean, in particular the overturning cells, are driven essentially by small variations in density. Since the density of seawater is determined by its temperature and its salinity, the ocean's deep circulation system is known as the **thermohaline** circulation, where 'thermo-' refers to its dependence on temperature, and '-haline' to its dependence on salinity.

Salinity of seawater may be altered in two ways:

- by addition of fresh water by precipitation (rain or snow) or its removal by evaporation
- by addition of meltwater from ice, or removal of freshwater by freezing.

When sea-ice forms (Figure 1.34), the ice itself is freshwater, so droplets of brine-rich water collect beneath it. This *brine-rejection* in cold polar waters may produce water sufficiently dense to sink to the seabed, and occurs extensively around Antarctica, contributing to formation of Antarctic Bottom Water. It also

Figure 1.34 Newly forming sea-ice (also known as 'grease ice' or 'frazil ice'). Note that this ice, forming from seawater, should not be confused with the polar and Greenland ice caps, which form from snowfall.

occurs around the Arctic Ocean, where ice forms over the continental shelf, and here it leads to formation of a cold, saline water mass that sinks to fill all the deep basins of the Arctic. Eventually, it escapes into the North Atlantic either by flowing through the Canadian Arctic islands to the west of Greenland and out into the Labrador Sea, or by flowing out into the deep basins to the east of Greenland, and then into the northeastern Atlantic.

Either way, this cold, saline Arctic water flows below regions in the northernmost Atlantic where surface water may be cooled by winter winds to such an extent that deep convection is triggered. Surface water may sink down to mix with the Arctic water at several thousand metres depth, and the resulting mixture is North Atlantic Deep Water. As the water sinking from the surface has been carried north in the Gulf Stream, it is initially fairly warm, at about 12–15 °C, and it has been estimated that in winter about 1300 km^3 of water sinks from the surface of the northern North Atlantic each day, having been cooled by winds from 12–15 °C down to 1–4 °C.

Question 1.7

Assuming that the warm surface water cools by 11 °C, how much heat is given up each day to the atmosphere over the northern North Atlantic? (*Hint*: use information from Box 1.4 along with the fact that 1 m^3 of water has a mass of 10^3 kg and there are 10^9 m^3 in 1 km^3.)

This is an enormous amount of heat – four orders of magnitude more than that supplied by solar radiation at ~55° N in winter (Figure 1.6).

Having sunk, the waters flow equatorwards as North Atlantic Deep Water still carries heat, but much less than it carried polewards. Meanwhile, much of the heat given up to the atmosphere is carried eastward in the prevailing winds, so formation of North Atlantic Deep Water has important implications for the climate of Europe.

Furthermore, it is thought that as water sinking from the surface draws in yet more water in to fill its place, North Atlantic Deep Water formation effectively increases flow in the Gulf Stream. This idea, combined with what is known about the paths of water through the ocean, has led to the concept of the thermohaline conveyor belt, illustrated in Figure 1.35. Overall, heat is carried northward in both the North and the South Atlantic, and this is compensated for by a net transfer of heat into the South Atlantic from the Pacific and Indian Oceans. Of course, the flow pattern shown is a generalisation (e.g. a particular parcel of water would be very unlikely to follow this particular path). To give some idea of the timescales involved, however, a parcel of water sinking in the northern North Atlantic would take on the order of a thousand years to come to the surface again in the northern Indian Ocean or the Northern Pacific.

Interestingly, the Pacific has no water mass analogous to North Atlantic Deep Water. The reason for this seems to be that because of high precipitation, the salinity of surface seawater in the northern Pacific is so low that cooling and evaporation can never increase its density sufficiently for sinking to occur. So does this mean that formation of North Atlantic Deep Water can be ‘turned off’ by a decrease in the salinity of Atlantic surface waters? Perhaps so. Indeed,

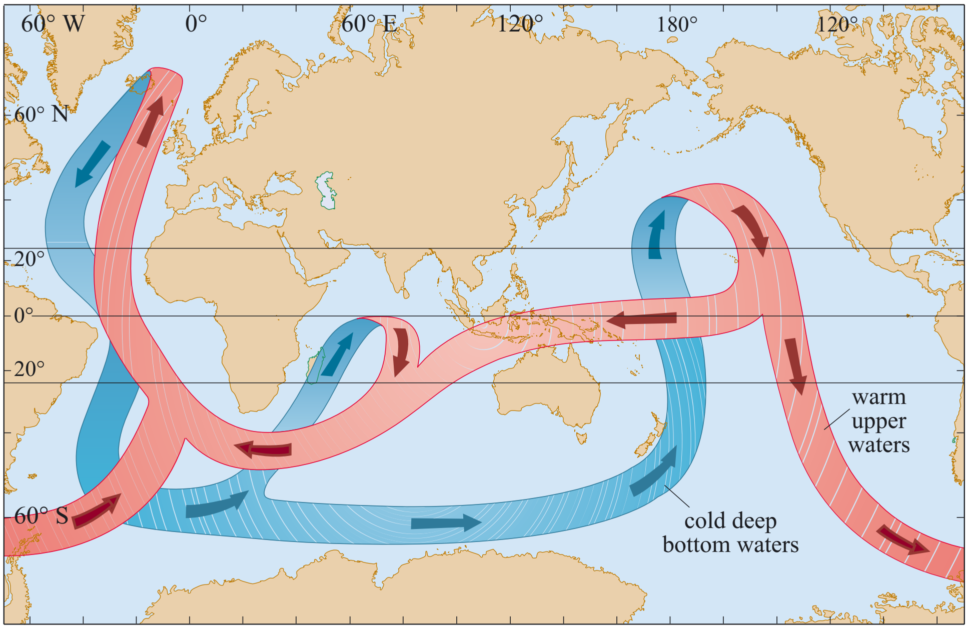


Figure 1.35 Schematic diagram of the global thermohaline conveyor, driven by the sinking of cold water at high latitudes. Warm water in the upper 1000 m or so of the ocean generally follows a pathway towards the northern North Atlantic. After sinking, cold water generally follows a path towards the northern Pacific and the northern Indian Ocean. (As the current flows in the Pacific and Indian Oceans (and the seas between) are not as well known as those in the Atlantic, different versions of this diagram are used.) (Broecker, 1991)

many scientists believe that North Atlantic Deep Water formation was turned off by a lid of low-salinity meltwater between 12 900 and 11 600 years ago, causing the period of cold climate known as the Younger Dryas.

As they flow through the oceans, water masses gradually mix with adjacent waters, but they may nevertheless still be identified many thousands of kilometres from their sites of formation. Since they acquire their characteristic temperatures and salinities when they are in contact with the atmosphere at the surface and maintain these characteristics over long periods of time, water masses – or at least imprints of them – are invaluable for studying past climate. (This topic will be explored in more detail in Chapter 5.)

Upwelling, downwelling and convection

The global-scale patterns of overturning seen in Figures 1.33 and 1.35 are driven by heating, cooling and freshwater forcing at the surface and only involve sinking or downwelling in very restricted, high-latitude regions with widespread upwelling elsewhere. This asymmetry, like the narrowness of the western boundary currents, is a consequence of the change in Coriolis effect with latitude. Localised downwelling is closely related to localised high-latitude convection, but there is a subtle difference between the two. Convection involves an exchange or mixing of denser surface water with lighter water from below and there is no net transfer of water downwards. Convection on its own, therefore, cannot supply the net downwelling needed to complete the overturning loop of the thermohaline circulation. It is, however, still closely linked to thermohaline downwelling because it is responsible for creating dense water masses. One reason for the strong connection is that convection results in a powerful *positive* feedback: when surface cooling causes convection, warmer water is lifted to the surface from below, thus releasing yet more heat to the atmosphere.

Downwelling occurs when surface waters move together, or converge. Similarly, diverging surface waters result in upwelling. Such convergence and divergence

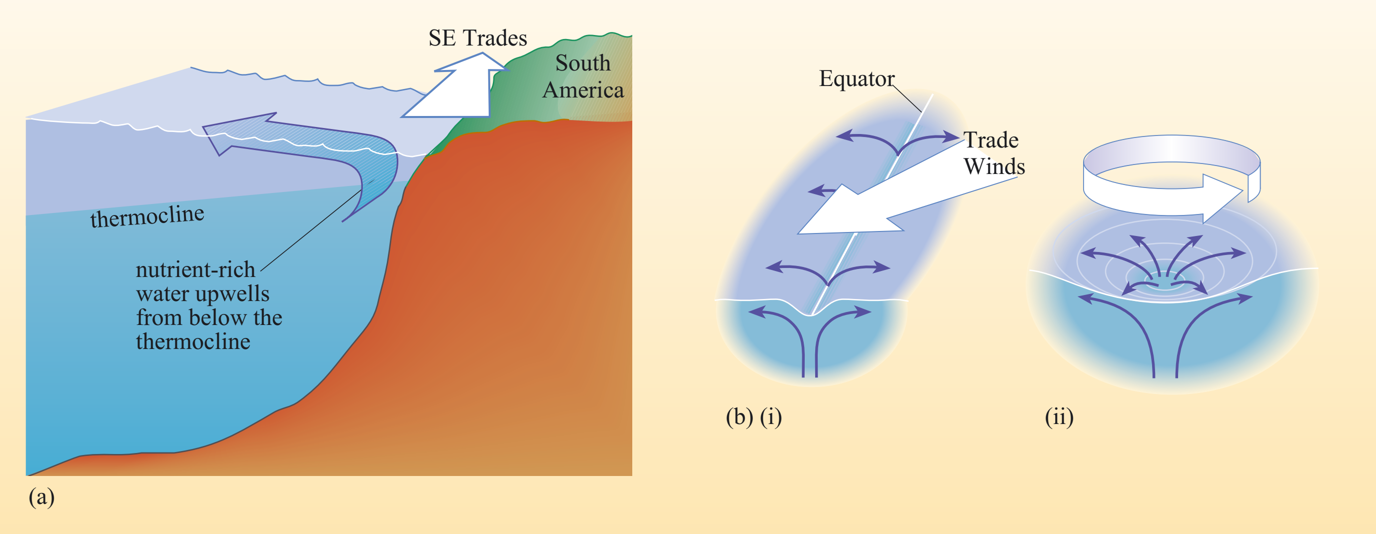


Figure 1.36 (a) Equatorward winds along the coast of Peru and Chile lead to offshore current flow (i.e. to the left of the wind, as this is in the Southern Hemisphere), causing divergence of water from the coast and upwelling of nutrient-rich water from below the thermocline. (b) Schematic diagrams to show other types of wind fields that lead to upwelling: (i) Trade Winds crossing the Equator (where the Coriolis force is zero) lead to a zone of upwelling along the Equatorial Divergence. (ii) Cyclonic winds lead to divergence of surface water and mid-ocean upwelling (here shown for the Northern Hemisphere); by contrast, anticyclonic winds lead to convergence of surface water and downwelling.

can be driven by density variations, as in the thermohaline circulation, or by wind forcing. Wind-driven upwelling often occurs at coasts, where land–sea temperature contrasts lead to winds blowing roughly parallel to the shore. As a result of the Coriolis force, water acted on by the wind moves at an angle to it, so for example the Southeast Trades blowing *along* the western side of South America cause surface water to diverge from the coast, leading to upwelling of water from below. You do not need to worry about the details of exactly *how* coastal upwelling occurs, but instead look at the various wind patterns that lead to upwelling along coasts in Figure 1.36a and in the open ocean (Figure 1.36b). As will be seen in later chapters, patterns of wind-driven upwelling and downwelling have important consequences for life in the oceans, and indirectly affect the climate.

Figure 1.37, which shows the upwelling and downwelling flow in July, was computed using the known wind field. It can clearly be seen that the Equatorial Divergence (the band of upwelling that occurs where the Southeast Trades cross the Equator, Figure 1.36b(i)), and the areas of coastal upwelling along the western coasts of North and South America and of Africa, is caused by the Trade Winds (Figure 1.36a).

■ By reference to Figures 1.20 and 1.28, can you suggest why there is also some open ocean upwelling at latitudes above about 45°?

■ Divergence and upwelling will occur in the subpolar gyres, which are driven by the cyclonic winds of the subpolar low-pressure systems (Figure 1.35bii).

By contrast, the centres of the subtropical gyres beneath the anticyclonic winds of the subtropical highs are sites of convergence of surface water and sinking.

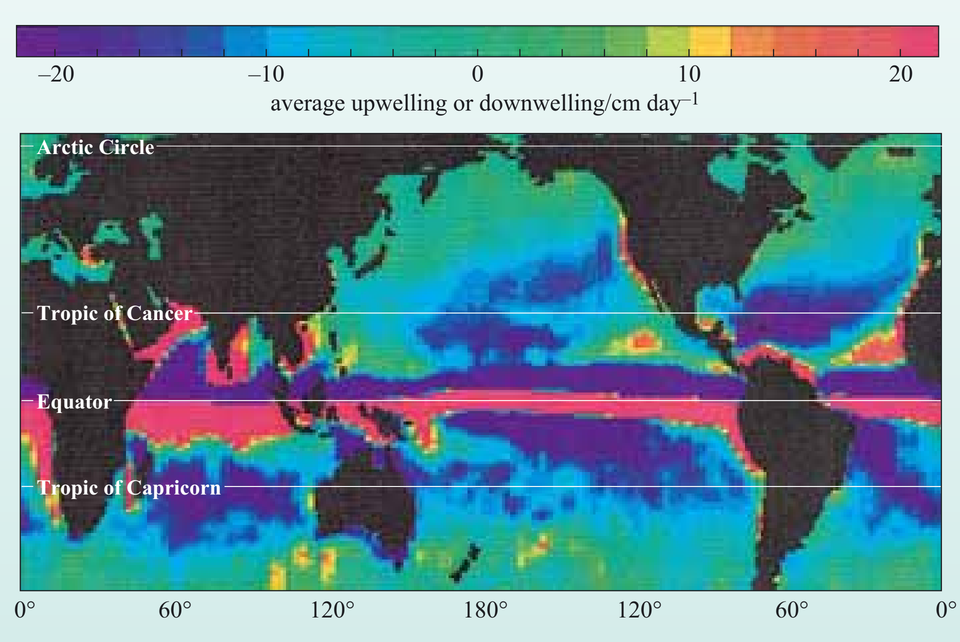


Figure 1.37 Upwelling and downwelling (vertical current flow in cm day^{-1}) for July, computed from the wind field. Areas of strongest upwelling (large positive values in the key, most notably along the West African, S.Indian and Sri Lankan coasts and close to the Equator) are pink; areas of strongest downwelling (large negative values – notably the tropical oceans away from the Equator) are dark blue.

1.5 Earth–ocean–atmosphere: the support system for life

In Section 1.4, you considered the Earth's heating and air-conditioning system in terms of two connected fluid envelopes – the atmosphere and ocean – meeting and interacting at the sea surface by transfer of heat and momentum. You have already seen that the transfer of fresh water is important too, but atmosphere, ocean, and indeed the solid Earth and living organisms are interlinked by the movement between them of many other materials. Though the timescale of movement can be months, millennia or even millions of years, depending on the circumstances and the elements involved, this continual redistribution may be described in terms of **biogeochemical cycles** – a concept that will be fundamental to the discussions regarding carbon in Chapter 3.

The transport of water (and hence heat) by means of evaporation (particularly from the sea-surface at low latitudes), followed by condensation in cooler environments, forms just one part of the **hydrological cycle** (Figure 1.38). It is the movement of water (which is a powerful solvent) that underpins the cycling of many other constituents through the Earth–atmosphere–ocean system. In the atmosphere, gases, including CO_2 and SO_2, dissolve in rainwater, resulting in a weak acid which, when it falls onto land, is neutralised by reaction with minerals in soil and rocks, releasing their constituent ions into solution in a process known as **chemical weathering**. The hydrological cycle can thus be viewed as a simple but special example of a biogeochemical cycle, another indication of the central role of water in Earth's unique environment.

Question 1.8

(a) According to Figure 1.38, what percentage of the Earth's water inventory is in the oceans at the present time? What percentage is in ice caps and glaciers?

(b) There is currently much concern about the rise in sea level as a result of global warming. Assuming one-quarter of the ice caps eventually melted, what percentage increase in the volume of the oceans would result?

So while the ice caps and glaciers contain most of the water that is on the land surface, they contain only a few per cent of the global water inventory: all but 3% of the Earth's water is in the oceans.

A concept that can usefully be applied to many different biogeochemical cycles is that of average **residence time** which, in the case of the hydrological cycle depicted in Figure 1.38, is a measure of the average length of time an individual water molecule spends in any particular stage or **reservoir** of the cycle. It is calculated by dividing the amount of water in that particular reservoir by the amount entering (or leaving) in unit time. It should be pointed out that although

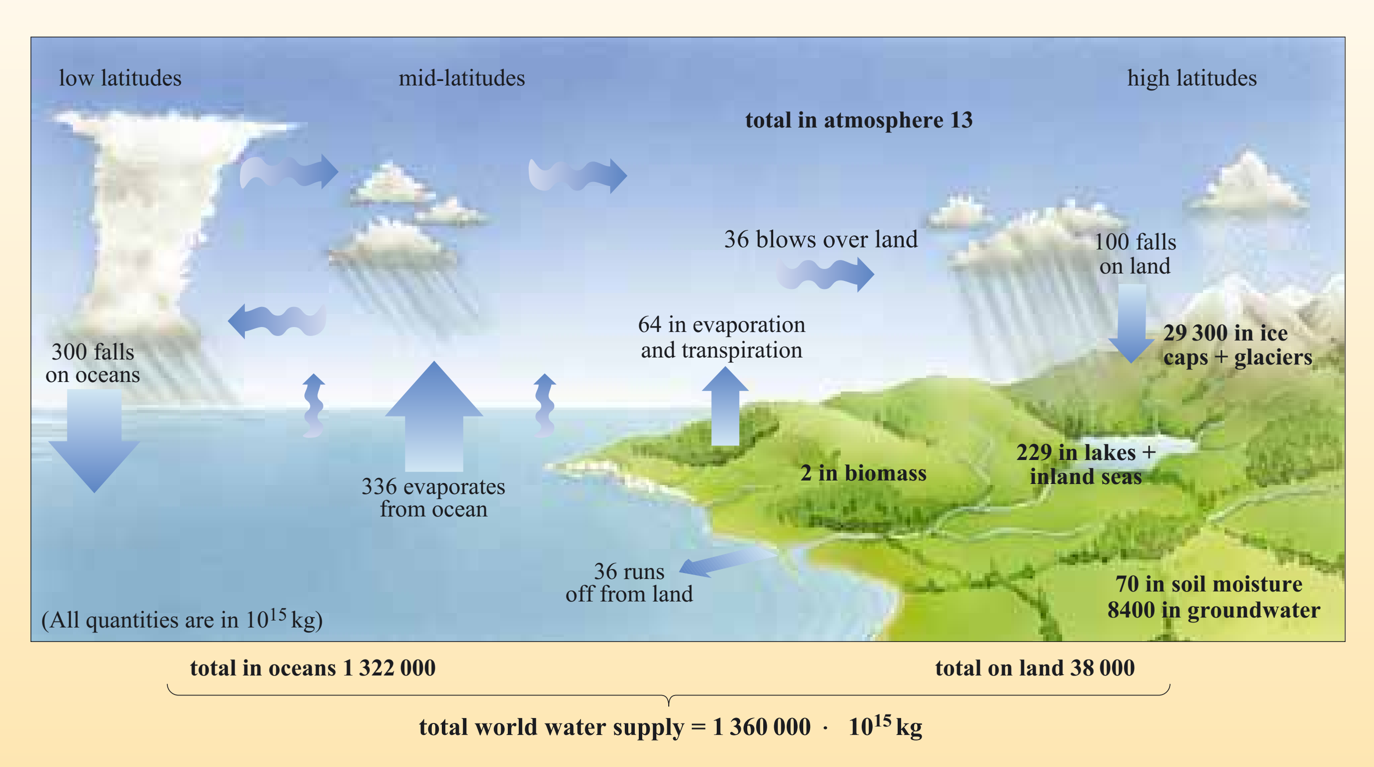


Figure 1.38 The hydrological cycle, showing annual movements of water through the cycle and amounts of water stored in different parts of the cycle (often referred to as reservoirs and here shown in bold type). Note the role played by vegetation in storing and cycling water. All quantities are × 10^{15} kg; note that 10^{15} kg liquid water occupies 10^3 km^3. The values given are estimates, and you may see different values elsewhere. For example, some authorities consider that groundwater storage accounts for a much larger proportion of the total water on land than shown here. The labels 'low latitudes', 'mid-latitudes' and 'high latitudes' are intended to remind you of the high precipitation at the ITCZ and in subpolar regions, and the high net evaporation in mid-latitudes; they are not supposed to imply that the processes shown are unique to those areas.

this is a useful concept, it is only an indication of average behaviour; individual water molecules may spend longer or shorter times in any given reservoir.

■ How much water (in 10^{15} kg) moves through the atmosphere annually? What is the average residence time of water in the atmosphere?

□ According to Figure 1.38, the amount of water entering the atmosphere yearly is 336×10^{15} kg + 64×10^{15} kg = 400×10^{15} kg (or the amount leaving it per year is 100×10^{15} kg + 300×10^{15} kg = 400×10^{15} kg); in other words, 400×10^{15} kg of water moves through the atmosphere annually. On average, 13×10^{15} kg of water resides in the atmosphere at any one time, so the residence time of water in the atmosphere is:

$$\frac{13 \times 10^{15}\ \text{kg}}{400 \times 10^{15}\ \text{kg y}^{-1}} = 0.033\ \text{y or about 12 days.}$$

Figure 1.38 and the above calculations are appropriate for the present time on Earth. In the past, and presumably in the future, under radically different climatic conditions, the relative magnitudes of the different processes and fluxes could be very different.

Looking at Figure 1.38, it would be tempting to conclude that because rivers flow into the ocean, seawater is simply a concentrated form of river water. But is it? Figure 1.39 shows the average chemical composition of rainwater, river water and seawater. Study these histograms carefully, taking note of the different vertical axes (and for the moment ignoring the fact that some columns of the river water histogram consist of two parts).

Comparison of the top and bottom histograms in Figure 1.39 shows that the relative concentrations of seawater compared to rainwater is:

$$\frac{34.4\ \text{g l}^{-1}}{7.1\ \times 10^{-3}\ \text{g l}^{-1}} = 4800\ \text{(to two sig. figs).}$$

■ Just comparing the histograms by eye (and ignoring the differences in concentration), would you say that seawater is closer in composition to rainwater or to river water?

□ Rainwater; the rainwater and seawater histograms bear a striking similarity to one another.

One of the main differences between rainwater and seawater on the one hand, and river water on the other, is the relatively high concentrations of dissolved Ca^{2+}, HCO_3^- and SiO_2 in the latter. These are supplied to rivers as a result of chemical weathering of carbonate and silicate minerals, both common constituents of the rocks making up the Earth's crust. What happens to remove these constituents from solution in seawater will become clear in Chapter 3.

Another marked difference is the relatively large amount of chloride in rainwater and seawater and the fairly small amount in river water. Only a tiny proportion of the chloride in the oceans comes from weathering. It originated as HCl gas released in volcanic eruptions (especially early in Earth's history when volcanism was more widespread than it is now) and is continually recycled via oceanic aerosols. These originate as droplets of seawater ejected into the atmosphere

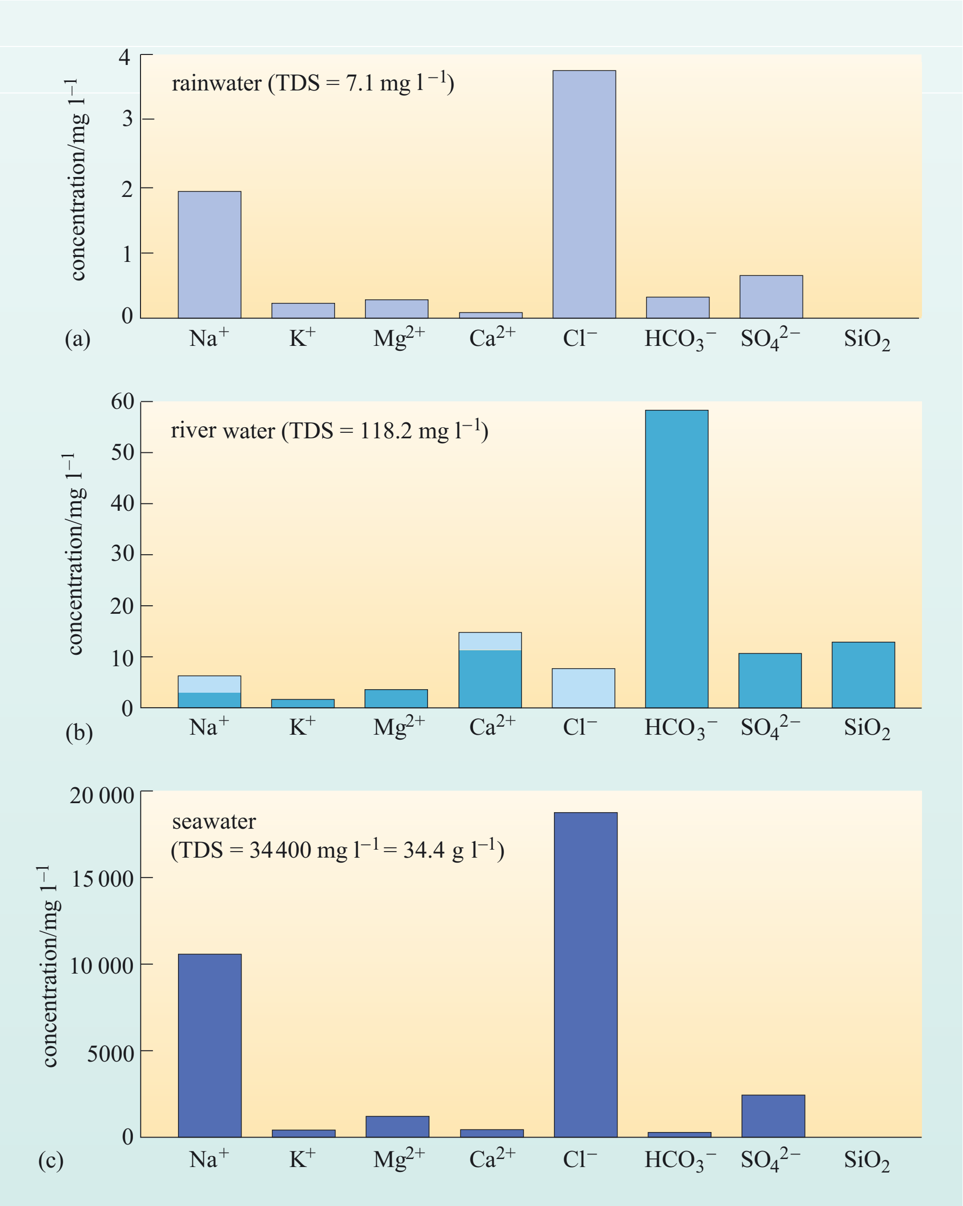


Figure 1.39 The average chemical composition of (a) rainwater, (b) river water and (c) seawater, shown in terms of the eight most abundant dissolved constituents, some at concentrations too low to appear. TDS = total dissolved salts. In (b), the darker lower parts of the columns are the contributions from weathering; the lighter upper parts correspond to salts supplied via aerosols (see text). Note the changes in scale of ×15 from (a) to (b), and × 400 from (b) to (c). Concentrations increase partly because of the addition of new dissolved constituents during weathering, etc., and partly because of loss of water through evaporation and transpiration.

when conditions are rough, along with any dissolved constituents, gases and particulate matter that they may contain. Particles of sea salt that end up in the atmosphere in this way are amongst those acting as nuclei for cloud condensation (Box 1.5), and their constituents are eventually rained out. As shown in Figure 1.39, while most dissolved ions in river water result from weathering (darker parts of histogram columns), a significant proportion of both sodium (Na^+) and calcium (Ca^{2+}), in addition to virtually all the chloride (Cl^-), comes from the ocean via rainwater (lighter parts of columns).

In total, all of the 92 naturally occurring elements found on Earth have been detected in seawater, in widely varying concentrations. Figure 1.40 summarises the various ways in which dissolved constituents enter and leave the oceans. Note that for some elements, hydrothermal vents at spreading axes are an important source. (Sedimentation, biological cycling and volcanism are discussed in Chapters 3 and 4.)

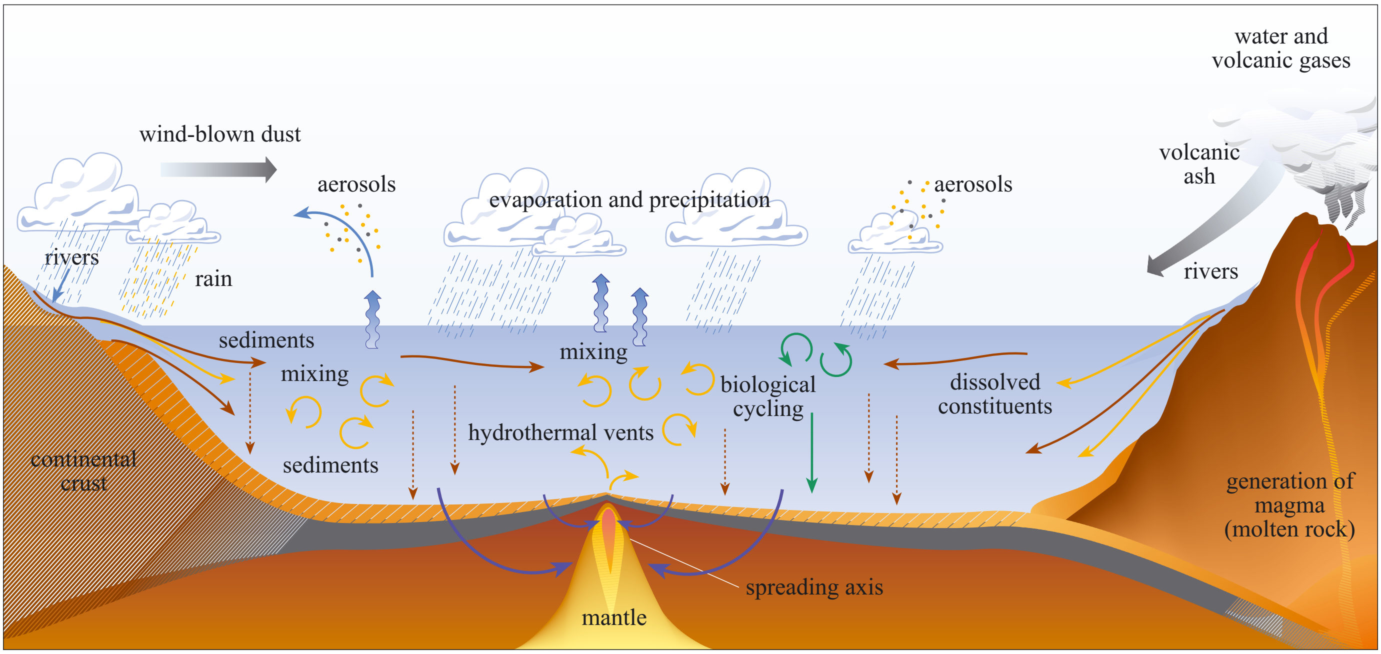


Figure 1.40 Diagrammatic cross-section illustrating the important role of the ocean in the global cycling of elements.

So, if water and all the elements are continuously recycled through the Earth system, what are the implications for life? The term biogeochemical cycles suggests that life plays a leading role in the story. Earlier, you saw how the tilt of the Earth's axis and the shape of its orbital path affect the distribution of incoming solar energy over the surface of the globe, and how the resulting pattern of surface temperature is modified by winds and currents. Intimately linked to the transport of heat over the globe is the transport of water.

■ To what extent is the following statement true?

Together, energy from the Sun and water-supply on Earth determine the distribution of primary production (mainly of plants) over the surface of the globe (Figure 1.1).

■ It is only partly true. As any gardener or farmer knows, some areas of land are more fertile (i.e. will support more plant growth) than others.

A fertile soil is one that contains an abundance of elements necessary for plant growth (i.e. **nutrients**) in a form that can be used by plants. Some landscapes will never be very green, irrespective of how much rain falls on them, because the minerals in the bedrock cannot be made *available* for plant growth (i.e. as dissolved ions) in sufficient amounts to support large stands of vegetation. Examples of such barren regions are steep rocky surfaces where no soil can accumulate, sand dunes and limestone pavement. To a large extent, however, the distribution of land plants is limited by the availability of light for photosynthesis, temperature and the availability of water. In the ocean, water is never in short supply; furthermore, temperatures are never too low to allow plant growth (see Figure 1.2). Nutrients, on the other hand, are not so plentiful. A crucial role of biogeochemical cycling for life, particularly in the ocean, is thus to provide the nutrients essential for growth.

Biogeochemical cycles also apply to elements that are normally gaseous, and indeed seawater contains dissolved gases as well as dissolved solids, particles and ions. In some cases, these too have important implications for the distribution of life. Figure 1.41 shows the concentration in seawater of the four most abundant gases in the atmosphere. The data in (b) are given for a particular temperature and pressure because the solubility of gases generally *decreases* with increasing temperature and salinity, and *increases* with increasing pressure.

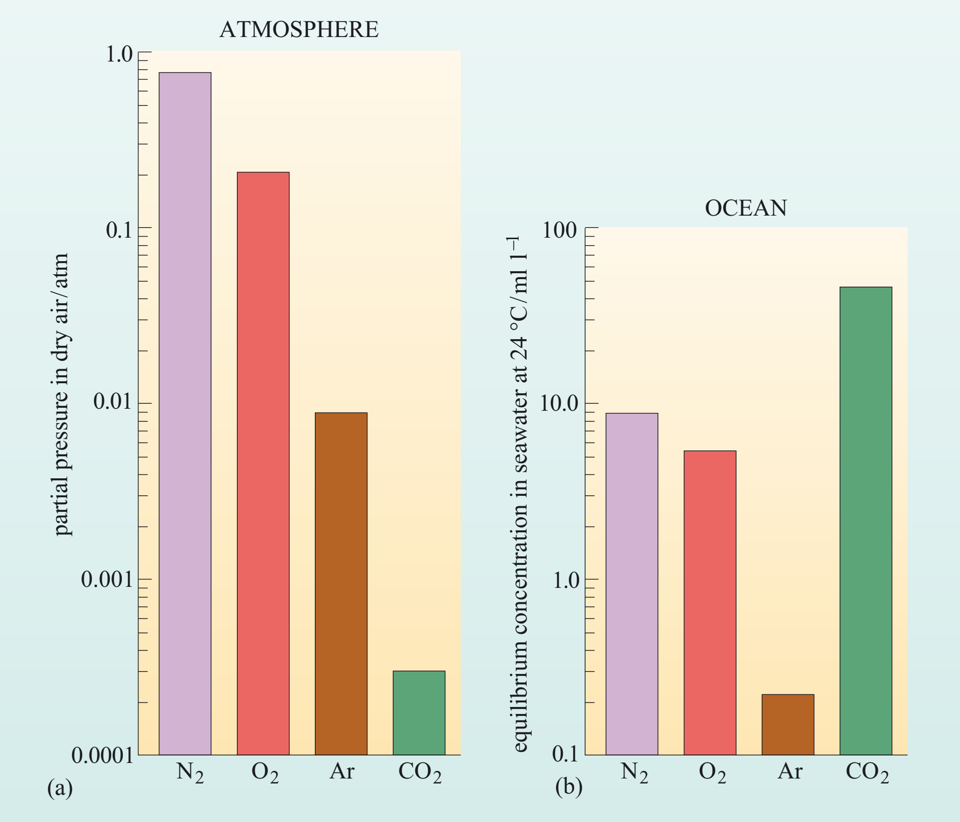


Figure 1.41 (a) Proportions by volume of the four most abundant gases in the atmosphere (together totalling 99.9%). (b) Concentrations by volume of these gases dissolved in seawater at 24 °C as controlled by the solubilities of the gases at their atmospheric concentrations. Note the different logarithmic scales on the vertical axes. (b) has been plotted assuming that there is equilibrium between atmosphere and ocean across the air–sea interface; for the four gases shown, this is probably valid, but only to a first approximation, for reasons that will become clear later on.

■ Which one of the gases shown in Figure 1.41 is the most soluble in seawater? Which one is essential for respiration of marine organisms?

■ CO_2 is the most soluble, by a long way (the reason for this will become clear in the next chapter). O_2 is essential for respiration.

Atmospheric gases dissolve into or escape from the surface ocean, depending on the relative concentrations either side of the air–sea interface (this is discussed further in Chapter 3). If it is very windy, and the sea-surface is rough, air and water mix together and the rate of gas transfer between them is increased. Once a gas is dissolved in the upper ocean, its transport throughout the ocean is the result of current flow and turbulent mixing.

■ Given what you know about ocean currents (Figures 1.15 and 1.33), would you expect to find living organisms at great depths in the ocean?

■ Yes, because dissolved oxygen is carried down into the deep ocean in the cold water masses sinking in polar regions.

So the ocean does not just contain dissolved constituents, particulate material and dissolved gases – it is also the largest 'living space' on Earth, a soup of living organisms, large and small. If asked to name a marine organism, many people might say 'fish' or 'whale', but the most important organisms are too small to be seen by the naked eye. These are:

- the *primary producers*, which are the minute floating algae or **phytoplankton** (Figure 1.42a), sometimes called 'the grass of the sea', and
- the *decomposers*, which are the **bacteria**.

Zooplankton are minute animals that feed on these smaller organisms and on each other (Figure 1.42b).

Not surprisingly, biological activity plays an important role in changing the concentrations of oxygen and carbon dioxide below the sea surface. Respiration by marine organisms, including bacteria, takes place throughout the ocean at all depths.

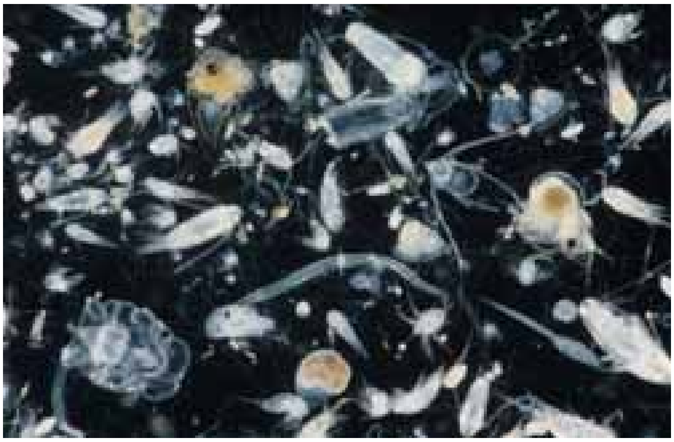
■ Is this also true of photosynthesis?

■ No. Photosynthesis uses light, and can therefore only occur in near-surface waters.

The oceans are pitch-black below about 1000 m; light is too weak for photosynthesis below ~250 m even in the clearest water. Since phytoplankton are plants and need to photosynthesise to grow (Equations 1.1 and 1.2), they (and the zooplankton that feed on them) live in the sunlit surface waters of the **photic zone**. As far as phytoplankton are concerned, the most important of the non-gaseous constituents dissolved in seawater are nitrate (NO_3^-), phosphate (PO_4^{3-}) and silica (SiO_2), i.e. the nutrients. Mysteriously, N and P occur in seawater in the same molar ratio (15 : 1) that they occur in the soft tissues of marine organisms. This suggests a certain level of feedback operates in the marine biosphere (probably on very long timescales) to maintain the dynamic balance between ocean and biospheric composition.

Other constituents are also used by living organisms, for example calcium, which, like silica, is used to build shells and skeletons, and sodium, which is one of a number of elements used in soft tissues. Calcium and sodium are available in great abundance in seawater and so their precise concentrations are not important for living organisms. Nitrate, phosphate and silica, on the other hand, may be completely used up by phytoplankton in surface waters and so their concentration may be the limiting factor determining whether or not phytoplankton can grow. For this reason, these three constituents are described as **biolimiting**.

(a)



(b)

Figure 1.42 (a) Living phytoplankton: the needle-shaped, cylindrical and chain-like organisms are diatoms, which are encased in silica; the irregular ones with 'horns' or spines are dinoflagellates. The field of view is about 1.75 mm across. (b) Living zooplankton, including copepods (planktonic crustaceans) and the planktonic larvae of various animals. The field of view is about 1.75 cm across, so is ×10 that in (a). (These photographs show samples collected by special plankton nets that concentrate the organisms; they are not naturally found so closely packed in the open sea.)

Most of the organic matter forming the soft tissue of phytoplankton is recycled in near-surface waters through consumption by zooplankton (and other pelagic animals) and bacterial breakdown of detritus and excretion products. A small amount of debris does however escape from surface waters and sinks towards the deep ocean. As a result of decomposition of organic remains and dissolution of shells and skeletons, their constituents, including the nutrient elements, are gradually returned to seawater at greater depths.

■ How might these constituents be returned to surface waters, where they can again be used by planktonic organisms?

□ By upwelling of subsurface water in response to the divergence of water nearer the surface (Figure 1.36).

Question 1.9

Figure 1.43 shows the concentration of chlorophyll pigment in the surface waters of the North Atlantic in May and December. Bearing in mind Figures 1.6 and 1.28:

(a) Explain why there is such a difference in the primary productivity in the northern North Atlantic at the two seasons of the year.

(b) Explain why, even in May, there is very little chlorophyll pigment in the central North Atlantic.

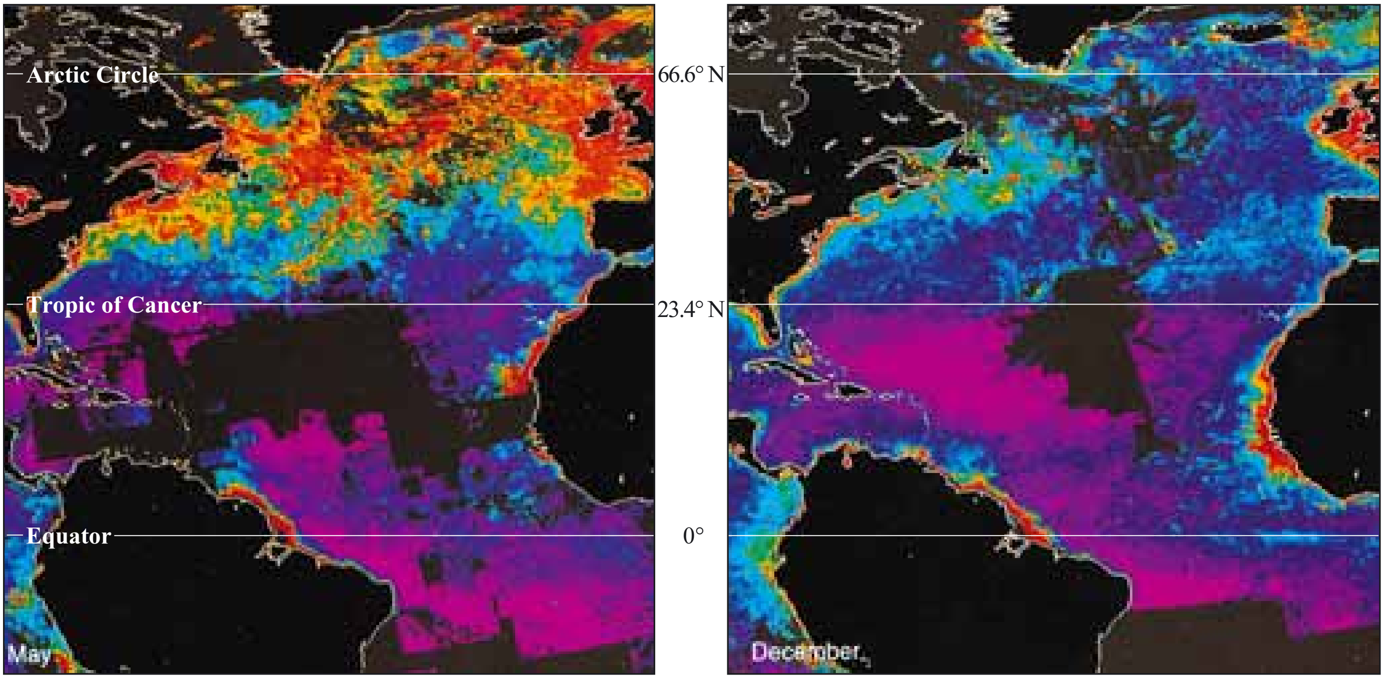


Figure 1.43 Seasonal changes in primary productivity in the North Atlantic, as indicated by concentration of chlorophyll pigment in (a) May and (b) December, measured by the satellite-borne Coastal Zone Color Scanner. Pink corresponds to lowest chlorophyll concentrations as in most open-ocean areas in December, and low-latitude regions in May (e.g. off the coast of eastern Brazil); bright red corresponds to highest chlorophyll concentrations widespread in mid- to high latitudes in May and in many coastal areas (e.g. off the coast of West Africa in May). Note that the areas of bright colour around the British Isles may be due partly to large amounts of suspended sediment and partly to real high primary productivity in response to nutrients (including agricultural fertiliser) contained in river runoff.

Therefore, as mentioned earlier, vertical current flows in the ocean have important implications for phytoplankton populations. Incidentally, it is important to bear in mind that the only reason we can so easily observe changes in marine primary productivity using satellite images is that phytoplankton increase their populations very fast but only live a short time, i.e. they have a very fast turnover rate. It would not be so easy to observe the changes in productivity of a forest of long-lived trees.

Recently, there has been much speculation as to whether, in certain circumstances and certain locations (e.g. the Southern Ocean around Antarctica), phytoplankton growth might be limited by lack of dissolved iron. Iron is found in seawater and in organisms in concentrations several orders of magnitude less than those of N, P and Si; it is therefore sometimes referred to as a *micronutrient.*

The clear blue waters of the tropical oceans are often described as barren because of their small phytoplankton populations. To test the hypothesis that tropical waters are iron-limited, one thousand square kilometres of the tropical Pacific was 'fertilised' by the addition of dissolved iron in 1995. The area concerned was to the southeast of the Galápagos Islands, at about 4° S, 104° W and the experiment was named IRONEX II. The large phytoplankton bloom that resulted from the addition of iron showed conclusively that primary production in tropical waters *is* iron-limited. Since then, a multitude of iron-fertilisation experiments have extended this result to other areas, in particular the Southern Ocean. Such controls on the amount of marine primary production may have important implications for climate.

As more is discovered about planet Earth, more links are discovered between the biosphere and the other components of the climate system – the atmosphere, hydrosphere and geosphere. In the late 1980s, a link was proposed between phytoplankton in the ocean and cloud formation. Many phytoplankton have population explosions or blooms during spring and summer (Figure 1.43), and blooms of certain species produce the volatile sulfur compound dimethyl sulfide (DMS). Once in the atmosphere, DMS undergoes a series of photochemical oxidation reactions, to produce sulfur dioxide (SO_2) and then sulfuric acid aerosols, which act as cloud condensation nuclei. This discovery excited many researchers, including environmental scientist James Lovelock (see Box 2.4), because increased concentrations of 'greenhouse' CO_2 in the atmosphere might also lead to increased marine primary productivity, i.e. more phytoplankton and bigger plankton blooms.

Question 1.10

How and why might such increased primary productivity affect the Earth's albedo?

By producing bigger blooms and hence more clouds, phytoplankton might counteract, or at least moderate, the effect of increased amounts of CO_2 in the atmosphere. This is an example of *negative* feedback, stabilising the climate system. This idea was favourably received when it was first proposed in the 1980s, but despite two decades of energetic research, conclusive evidence for or against the proposed feedback has remained elusive.

Finally, it should be noted that primary production can and does occur in the dark of the deep sea, by bacteria using chemical energy rather than the energy of sunlight. Like their counterparts photosynthesizing in surface waters, these primary producers form the bases of whole ecosystems, of which the best known are the communities of animals found at hydrothermal vents. Unlike phytoplankton, however, **chemosynthetic bacteria** do not interact with the climate system in any significant way (at least as far as is known).

Figure 1.44 summarises much of what has been discussed about conditions at the surface of the Earth at the present time. As you will see later in this book, in the past, the Earth has been very different. This map is not intended to be comprehensive, but to remind you of some of the main points described in this chapter.

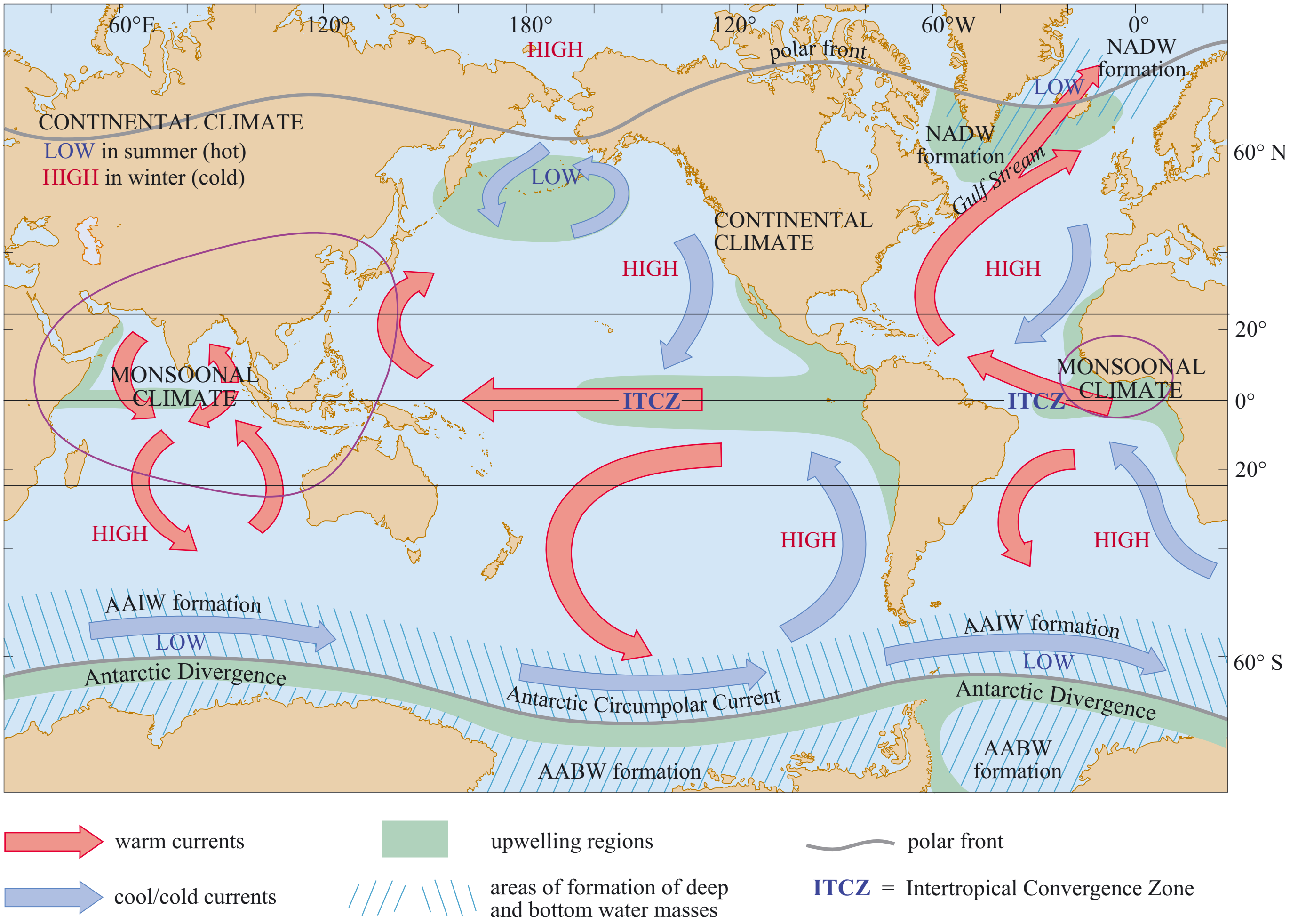


Figure 1.44 Highly schematic map showing general climatic features for land and sea, and the general position of the polar fronts in the atmosphere. HIGH and LOW refer to atmospheric pressure at the Earth's surface. Regions of deep and bottom water-mass formation are shown by blue hatching. Antarctic Bottom Water (AABW) formation, in particular, is localised, occurring only at very high southern latitudes, while North Atlantic Deep Water (NADW) formation occurs only in the northern North Atlantic. Antarctic Intermediate Water (AAIW) is formed throughout the Southern Ocean north of the Antarctic Divergence. Regions of upwelling, principally in equatorial and sub-polar gyre regions and along the Antarctic Divergence, are shown in green.

Summary of Chapter 1

1 The Earth is hospitable to life because of its particular orbit around the Sun, which determines the amount of solar radiation that reaches it. The amount of solar radiation actually available to warm the Earth's surface is determined by how much is reflected rather than absorbed, i.e. by its albedo. At the temperatures obtained at the Earth's surface, water can exist as solid, liquid or gas; and were it not for the presence of liquid water, life could not exist on Earth.

2 The fixation of carbon by primary producers (mainly plants) is the basis of all life on Earth.

3 Seasonal changes in incoming solar radiation are a result of the tilt of the Earth's axis in relation to the orbital plane (i.e. if the Earth's axis were at right angles to the orbital plane, there would be no seasons). Over timescales of tens of thousands to hundreds of thousands of years, there are cyclical variations in incoming solar radiation caused by changes in the degree of *eccentricity* of the Earth's orbit, in the *angle* of tilt of the Earth's axis, and in the *direction* in which the axis points. These cycles have periodicities of ~110 000 years, ~40 000 years and ~22 000 years respectively, and are known as the Milankovich cycles.

4 Most of the radiative energy emitted by the Sun is in the wavelength range 0.15–5 μm, which includes ultraviolet radiation, visible radiation and infrared radiation. Solar radiation is often referred to as shortwave radiation, to distinguish it from longer-wavelength (thermal) radiation emitted by the Earth (and any other body) as a consequence of having been warmed. Certain gases in the atmosphere, notably carbon dioxide and water vapour, absorb outgoing longwave radiation and reradiate energy, much of it back to the Earth's surface. This trapping of radiant energy (which also involves clouds) is known as the greenhouse effect.

5 Although the Earth's surface is heated unevenly by the Sun, the redistribution of heat by winds and currents ensures that low latitudes do not continually heat up and high latitudes do not continually cool down. The continents have a much lower thermal capacity than the oceans, and so heat up and cool down much faster. This strongly affects the Earth's surface temperature distribution, particularly in the land-dominated Northern Hemisphere; in the oceanic areas of the Southern Hemisphere, temperature decreases more smoothly from Equator to poles, the temperature zones simply shifting northwards and southwards with the seasons.

6 Heat is redistributed over the surface of the Earth by winds and ocean currents, and by the evaporation, transport and condensation of fresh water. Rising air generally has a high moisture content (as at the Intertropical Convergence Zone); sinking air (as at the subtropical highs) is usually dry.

7 Clouds both reflect incoming solar radiation and absorb longwave radiation. Although they are a very important component of the climate system, and

potentially the main driving mechanism for changes in the Earth's radiation budget, their role is by no means fully understood. Cloud condensation nuclei include sulfate aerosols, produced by volcanoes and industry, as well as forming from dimethyl sulfide, a waste product of phytoplankton populations.

8 The Coriolis force acts to deflect flows to the right in the Northern Hemisphere and to the left in the Southern Hemisphere. The Coriolis effect increases with increasing latitude, leading to more circulatory atmospheric flow in mid-latitudes than nearer the Equator. The increase in Coriolis effect is also responsible for the narrowness of the western boundary currents found in all ocean basins. The formation of mid-latitude weather systems is determined by the behaviour of the polar jet stream, whose undulations at the top of the troposphere lead to the transport of enormous amounts of warm air polewards and cold air equatorwards. Extreme undulations of the polar jet stream may lead to severe droughts and unusual weather patterns. The ocean also has weather in the form of mesoscale eddies, which are an order of magnitude smaller than atmospheric cyclones and anticyclones.

9 Tropical regions where the ITCZ moves seasonally over land are subject to seasonally reversing winds (monsoons) and hence dry seasons alternating with seasons of heavy rainfall. The atmosphere and ocean are tightly coupled by positive feedback loops, especially in tropical areas. Positive feedback loops cause instability in systems: negative feedback loops are stabilising.

10 While light, suitable temperatures and availability of water are the principal factors limiting primary production (and hence most other life) on land, in the oceans the limiting factors are light and the availability of nutrients. Nutrients depleted from the sunlit surface layer may be returned to it via upwelling. Upwelling occurs along the Equator, along western coastlines under the Trade Winds, and as a result of cyclonic winds.

11 While phytoplankton live in near-surface waters, in the photic zone, animals and bacteria live throughout the ocean, at all depths. This is possible because the cold water masses sinking at high latitudes carry dissolved oxygen down into the deep ocean. Some organisms, using chemical sources of energy, thrive in deep-sea hydrothermal vents and seeps.

Learning outcomes for Chapter 1

You should now be able to demonstrate a knowledge and understanding of:

1.1 How interactions between physical, chemical, biological and geological processes on Earth, along with its position relative to the Sun, make it a habitable plant for life to develop, exist and evolve.

1.2 The range of factors influencing the amount of solar radiation that reaches the Earth over both short and geological timescales, and difficulties associated with calculating such values.

1.3 The role of wind and oceanic currents in maintaining the Earth's surface temperature conditions and how perturbations in these circulatory systems can result in extreme short (and longer-term) changes in weather patterns.

1.4 The different circulatory systems that operate within the distinct oceanic and atmospheric layers (or spheres), and the physical and chemical processes that result in interaction and mixing between these layers.

1.5 How the concept of biogeochemical cycling and residence time can be used to investigate the extent and rate of redistribution of materials between different biological, chemical, geological and physical spheres on Earth.

Chapter 2

The emergence and persistence of life

> I look at the natural geological record, as a history of the world imperfectly kept, and written in a changing dialect; of this history we possess the last volume alone.
>
> (Darwin, 1859, p. 311)

One aspect of Darwin's disappointment with the then known fossil record of evolutionary change was the apparent absence of Precambrian fossils. The only hint of such ancient life that he knew of, when he made the statement quoted above, was 'the presence of phosphatic nodules and bituminous matter' in them. Nevertheless, he firmly believed that there had to have been a long history of evolution prior to the seemingly abrupt first appearance of marine shelly fossils just over 500 Ma, as the latter already represented many different major groups of animals. His conviction has been vindicated, for there is a fairly extensive, if partial, Precambrian fossil record stretching back to Archaean times – though the story it tells might have come as something of a surprise to Darwin himself. It turns out that animals (multicellular organisms that ingest their food) do seem to have made a relatively late appearance in the **Neoproterozoic**, followed by the evolutionary explosion of shelly forms that announced the dawn of the Phanerozoic. Yet it was during this long earlier history of the Archaean that the fundamental pattern of feedbacks between life and the Earth, particularly the atmosphere, became established, and the precursors to the later forms of Phanerozoic life, the **eukaryotes** (whose cells contain nuclei), evolved. In Chapter 1 you learned about some of the factors of the Earth system that regulate the habitability of the surface; in this chapter you will look at some of the history of life on Earth and its evolution.

The Precambrian or 'Cryptozoic' is the period of the geological timescale from the formation of Earth (around 4.6 Ga) to the evolution of abundant macroscopic fossils, which marked the beginning of the Cambrian, ~542 Ma.

2.1 Former worlds

As you try to understand the past conditions on the Earth you must remember that the environment has apparently stayed within modest limits for most of the Earth's history, although some have questioned this, postulating mean temperatures of over 60 °C and even higher oceanic temperatures for the Archaean. Other aspects – most notably atmospheric composition – have undergone radical changes (see Box 2.1). Life has apparently been present throughout virtually all the time considered from near the beginning of the Archaean, and it has been implicated both in the regulation of global conditions and in their changes. Individual complex multicellular organisms are nevertheless of relatively recent vintage, arriving approximately 600 Ma ago during the beginning of the Phanerozoic. Today, they present a kaleidoscopic web of interactions, both with each other and with their environments.

Box 2.1 Oxygen through time

The concentration of oxygen in the Earth's atmosphere has not remained constant through time. Prior to about 2.4 Ga, evidence suggests that the atmosphere was essentially oxygen-free (although some researchers doubt this). The presence of reduced minerals such as pyrite and uraninite in riverbeds older than 2.4 Ga attest to **anoxic** conditions. The early geological record also contains banded iron formations (BIFs), which are thought to have been formed by reduced iron that welled up from the anoxic deep ocean and oxidised in surface waters. BIFs begin to disappear from the geological record at about 1.8 Ga. Their continuation for about 600 Ma after the first rise in oxygen may reflect the time it took for the deep oceans to become oxygenated.

One conundrum with this view of the early Earth is the fossil evidence for oxygen-producing life before 2.4 Ga. Fossil stromatolites, carbon isotopic ratios consistent with oxygenic photosynthetic organisms, and **biomarkers** (specific biological molecules) suggest that oxygenic photosynthetic life had evolved by 3.5 Ga. Why was the rise in oxygen delayed long after oxygen-producing life appeared? One solution to this apparent problem is that in the early stages much of the oxygen being produced reacted with reductants in the atmosphere such as hydrogen and hydrogen sulfide (H_2S). Only when the concentrations of these reductants declined could the net atmospheric concentration of oxygen increase.

The lack of oxygen in the early atmosphere would have meant that there was no stratospheric ozone to screen UV radiation between 200 nm and 290 nm, resulting in potentially higher radiation damage to surface-dwelling organisms. In the worst-case, UV radiation would have been about one thousand times more damaging to DNA than it is today, although some researchers have suggested that the early atmosphere might have contained a hydrocarbon smog, which could have provided UV shielding.

A second rise in oxygen can be recognised from the geological record at about 800–600 Ma. Like the first rise in oxygen about 2.4–2.3 Ga ago, this transition was associated with a global freezing or 'snowball' event. One explanation for these dramatic environmental changes associated with the rise in oxygen is the consumption and conversion of the greenhouse gas methane (CH_4) by oxygen into CO_2 and H_2O, causing a global cooling. The period following the second rise in oxygen is also coincidental with the Cambrian explosion and the proliferation of animal diversity.

Investigating what controls the concentration of oxygen in the atmosphere, past and present, and how this might be linked to biological evolution is an important challenge in understanding the habitability of the Earth.

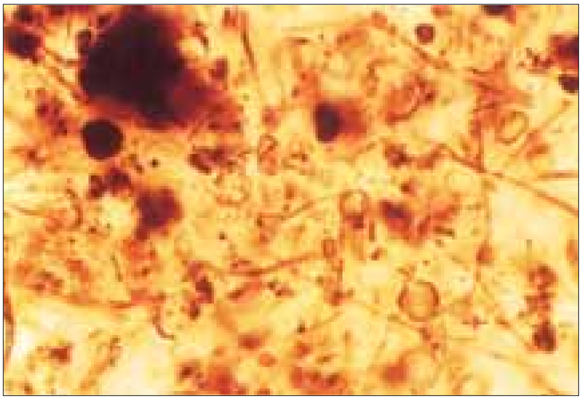
Question 2.1

What conditions would you expect to find if you were to visit a mid-latitude, lowland area on Earth at (a) 4 Ga, (b) 2.5 Ga, (c) 500 Ma, (d) 100 Ma, and (e) 100 000 years ago? To help you think about your answer, decide which of the following items you would need to take with you: (i) breathing apparatus (with a supply of oxygen), (ii) suncream (factor 100), (iii) a thermally insulated self-contained capsule, (iv) a crabbing net, (v) weapons for fighting large predators, (vi) sandwiches, and (vii) a good book.

For the first three-quarters of life's 3.8 billion year history, there is frustratingly little fossil evidence (Figure 2.1a), despite its profound influence on the composition of the atmosphere, but what there is shows surprisingly little change in outward form and diversity. Yet this was a time of fundamental innovations in cell structure, giving rise to the first eukaryotes around ~2.7 Ga. Much later, at 1 Ga, a prolific burst of eukaryote diversification ensued. Such meagre fossil evidence as there is, together with other kinds of chemical biomarker, judiciously spiced with speculation, is the diet on which understanding of life on the early Earth is based.

From around 540 Ma, a proliferation of shelly animals began to litter the sea floor with their remains (Figure 2.1b), commencing the rich fossil record of the Phanerozoic. The Phanerozoic has yielded a sharply focused picture of patterns in the history of life relative to the Precambrian record. Very uneven rates of turnover (e.g. appearance and disappearance of fossil groups) have been revealed, with mass extinctions episodically cutting across long phases of slower evolutionary change. One eminent palaeontologist aptly commented, 'the history of any one part of the Earth, like the life of a soldier, consists of long periods of boredom and short periods of terror'. Indeed, this episodic pattern was widely acknowledged by geologists from the beginning of the 19th century and was utilised as a basis for the relative geological timescale that is still in use today:

- the most devastating mass extinction was used to mark the close of the Palaeozoic Era (i.e. the Permo-Triassic boundary)
- a less pervasive but more abrupt mass extinction (including the dinosaurs) was taken to mark the end of the Mesozoic Era (i.e. the **K/T boundary**).


(a)


(b)

Figure 2.1 (a) Early Proterozoic fossil assemblage. Photomicrograph of prokaryotic cells in chert from the Gunflint Iron Formation, Ontario, Canada. Spheres are about 10 µm across. (b) An assemblage of shells of the trilobite *Redlichia* from the Lower Cambrian of Kangaroo Island, Australia. Fossil at bottom centre is 5 cm long.

Some of the geological periods within the eras were likewise terminated with mass extinctions of varying severity, though not all were. Needless to say, some of these extinctions radically reset the biological agenda for subsequent evolution, with the richness of the Phanerozoic fossil record and the accompanying wealth of geological data placing constraints upon hypotheses

K/T boundary: Cretaceous–Tertiary boundary (C was already used for Carboniferous, hence the use of K).

seeking to explain such events and their aftermaths. These will be explored in greater detail in subsequent chapters, but for the present you need to consider Precambrian times. An understanding of this history is essential to investigate the origins of eukaryotes and to make sense of the part they have played in the Earth–life system.

2.2 Mat world

Life on Earth at 2 Ga (which was conjured up in Question 2.1) comprised greenish-grey mats along the coastline. The laminated calcareous structures they formed (called **stromatolites**) dominated the macroscopic evidence of life in all but the last couple of hundred million years or so of the Proterozoic fossil record. They had already become quite widespread and diverse by early Proterozoic times. In rare cases, fossil stromatolites contain tiny filamentous and spheroidal fossil remains (Figure 2.1a). But how representative are these microscopic fossils (**microfossils**) of the life forms of that time? One of the best-known examples of such preservation is the 2 Ga old Gunflint Iron Formation that straddles the boundary between Canada and the USA around Lake Superior (Figure 2.1a). The exquisite preservation of the fossil structures in this instance is due to mineralisation in chert (a form of silica (SiO_2)), precipitated from mineral-rich waters seeping through the algal mats. The mineralisation must have taken place soon after deposition and before significant decomposition of the organic materials could occur, and was probably due to molecules of silicic acid (H_4SiO_4) formed by the earlier dissolution of silica, bonding with hydroxyl ions (OH^-) on the surface of the organic matter.

The preservation of similar associations in Proterozoic rocks found elsewhere in North America, Siberia and Western Australia suggests that the organisms might have been widely distributed. The minute size of the fossil structures is consistent with a microbial origin, the filamentous sheaths resembling those of living cyanobacteria (although the cyanobacterial interpretation of some Archaean fossils has recently become a matter of some controversy), and the spheroidal structures (2–15 μm across) also resembling cyanobacteria, although many other bacteria also have a spheroidal (coccoid) morphology. The filaments do not display the preferred orientations (alternations of horizontally and vertically orientated clusters) characteristic of mat-forming microbes in younger stromatolites, so interpretation of their relationships and ecology must be tentative.

There are also some non-stromatolitic assemblages of microfossils in the Gunflint Iron Formation and elsewhere, preserved in silicified mudrocks. These include star-shaped forms interpreted as iron- and manganese-oxidising bacteria, as well as simple, rather larger (6–30 μm across) spheroidal forms of uncertain evolutionary relationships, which were probably planktonic.

A yet wider variety of stromatolites (ranging from flat-layered to domed, conical, columnar and even branched forms) is known from younger Proterozoic rocks, reaching a peak in those deposited around 1 Ga. Microfossils also testify to a rich diversity of microbes existing at that time. The association of various fossils with differing sediments suggests some diversification among habitats. Many of the microfossils are quite similar to living cyanobacteria or, more rarely, to bacterial **heterotrophs**. Larger eukaryotic forms also contributed to the diversity of that time (Figure 2.2) and Figure 2.3 shows a diagrammatic transect across a range of shelf-sea environments that can be interpreted from sedimentary rocks of mid- to late Proterozoic age, illustrating the variety of life forms associated with them. The restricted inshore to supratidal zone (marked (1) on Figure 2.3) was

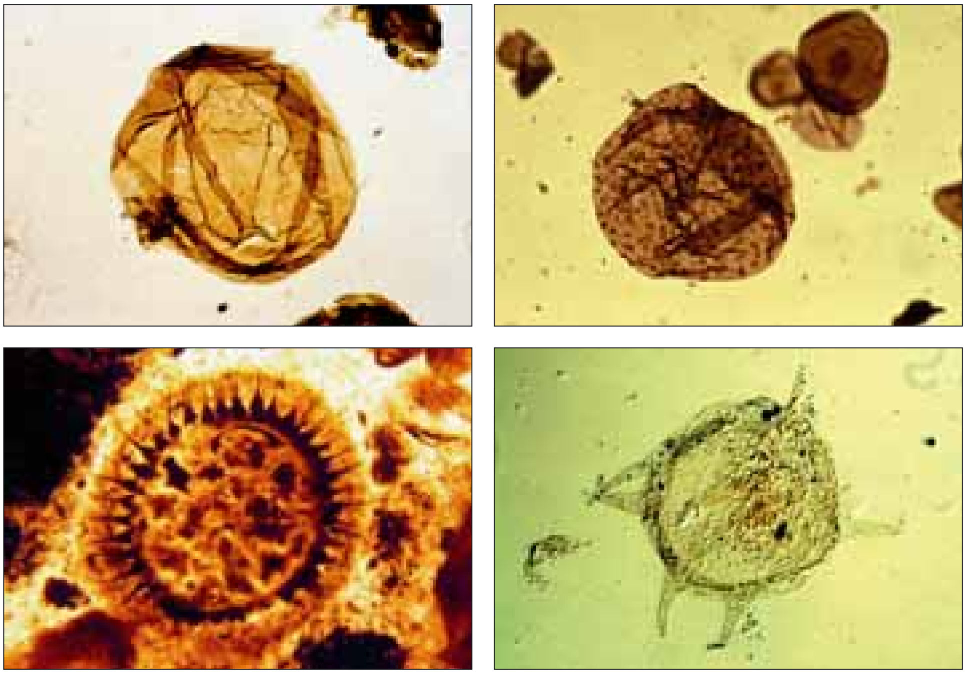


Figure 2.2 Fossils of early eukaryotes: a variety of cysts (resting stages), known as **acritarchs**, from rocks of late Proterozoic age. (Each box in the figure is about 300 μm high.)

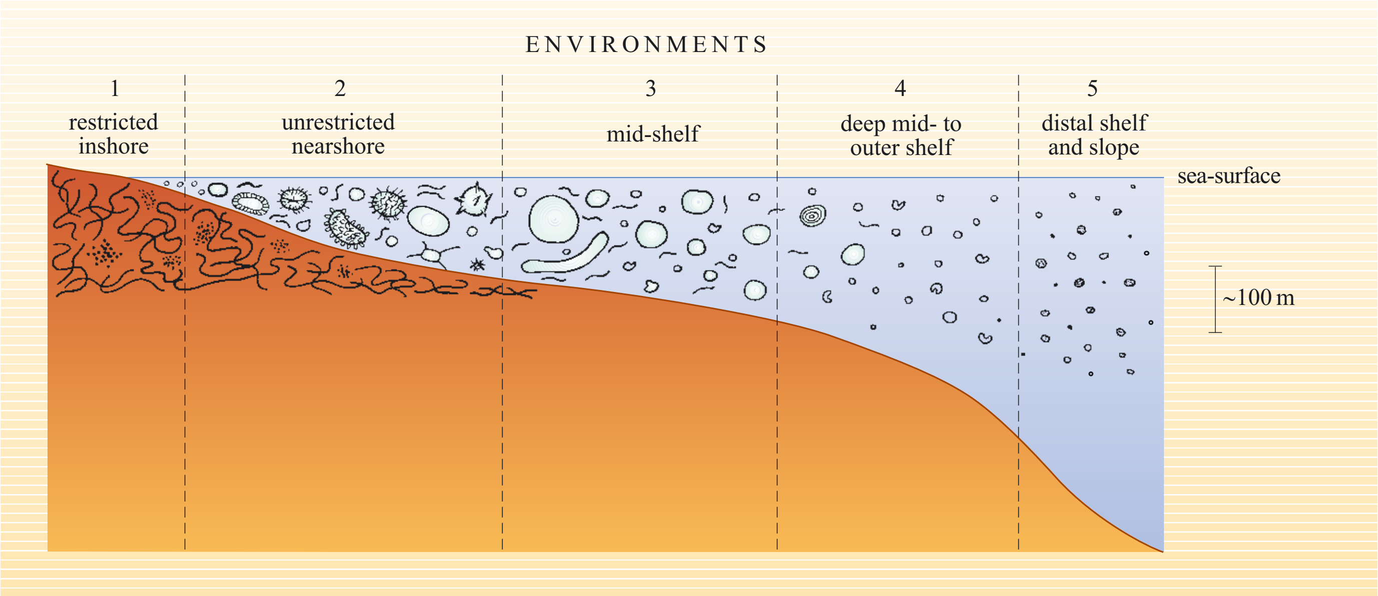


Figure 2.3 The reconstructed distribution of life forms in a range of shelf to basin environments in mid- to late Proterozoic times. The five numbered environments and their inhabitants are described in the text. *Note*: figure not drawn to scale.

dominated by tiny filamentous and spheroidal **prokaryotes** that produced stromatolites (like the earlier Gunflint examples). In unrestricted nearshore zones (2), prokaryotic filaments were joined by larger (>50 µm), elaborately ornamented eukaryotic cysts called acritarchs (discussed below), which were probably planktonic. In somewhat deeper waters on the mid-shelf (3), less ornamented, spheroidal acritarchs predominated. Over the deep mid- to outer shelf (4) and distal shelf and slope (5), there was a low diversity of small (<50 µm), simple, spheroidal plankton, with larger, probably eukaryotic forms yielding entirely to clusters of smaller prokaryotic cells out in the basins. As a result of the lack of preservation of ancient ocean floor, little is known about the life forms that lived there.

By mid-Proterozoic times, microbial ecosystems had already established most of the fundamental biogeochemical feedbacks that shape the modern world. Foremost among these was the evolution of plankton, which supplied a constant rain of dead organic material to the sea floor. There, it could be incorporated in the accumulating sediment, relatively undisturbed by current activity, and not see the light of day again until geological processes returned the sedimentary rocks to the surface.

Question 2.2

Why was this burial of organic material of crucial importance in the evolution of the Earth’s atmosphere?

The stromatolites that dominate the Proterozoic fossil record are not without their own biogeochemical significance. Their widespread development within the photic zone (from open shelf to supratidal settings) trapped large quantities of calcium carbonate ($CaCO_3$), to form beds of limestone. This represented an important geological sink for carbon (in the carbonate), although it was not accompanied by the release of molecular oxygen, unlike photosynthesised organic matter. Nevertheless, the carbonate story was, in time, to be taken much further by the eukaryotes, as will be seen later in this chapter.

2.3 Empire of the eukaryotes

2.3.1 What is a eukaryote?

The cells of eukaryotes possess orders of magnitude more DNA than do those of prokaryotes, which is packaged together with proteins to form a number of thread-like bodies in the nucleus called **chromosomes** (Figure 2.4a). Besides the nucleus, eukaryote cells also contain a host of discrete **organelles** that carry out specific functions. Some of these organelles, such as the energy-evolving mitochondria, are postulated to have been prokaryotes that were incorporated by **symbioses** into early eukaryotic cells. Another feature of the eukaryote cell is an internal framework of protein rods (microfilaments and microtubules) that plays an important role in its internal organisation. In contrast, most prokaryote cells are surrounded by a rigid outer wall, and it is to the inner surface of this that their simple loop of DNA is anchored (Figure 2.4b). In addition they also lack organelles.

Question 2.3

What is the main difference in the organisation of DNA in these two major groups of organisms, i.e. eukaryotes and prokaryotes?

For reasons that will be explained shortly, these differences in cell architecture, particularly in the arrangement of the DNA, have vastly expanded the scope of eukaryote evolution relative to that of the prokaryotes, and the contrast is readily apparent from living examples (Box 2.2). Most organisms that can be distinguished by the naked eye are made up of eukaryotic cells. Most are multicellular, including plants, fungi and animals, though there are also many single-celled forms that reach tens to hundreds of micrometres in maximum dimension. Single-celled eukaryotes include amoebae, foraminifers, *Plasmodium* (the parasite that gives you malaria) and many kinds of unicellular algae. The vast majority of prokaryotes (e.g. bacteria), by contrast, are unicellular, though a few may group together to form multicellular filaments. Their cells are usually at least an order of magnitude smaller than eukaryote cells (they are rarely more than 10 μm in maximum dimension and, in most forms, are less than 1 μm). The innovations ushered in by the eukaryotes were therefore key events in ecological history.

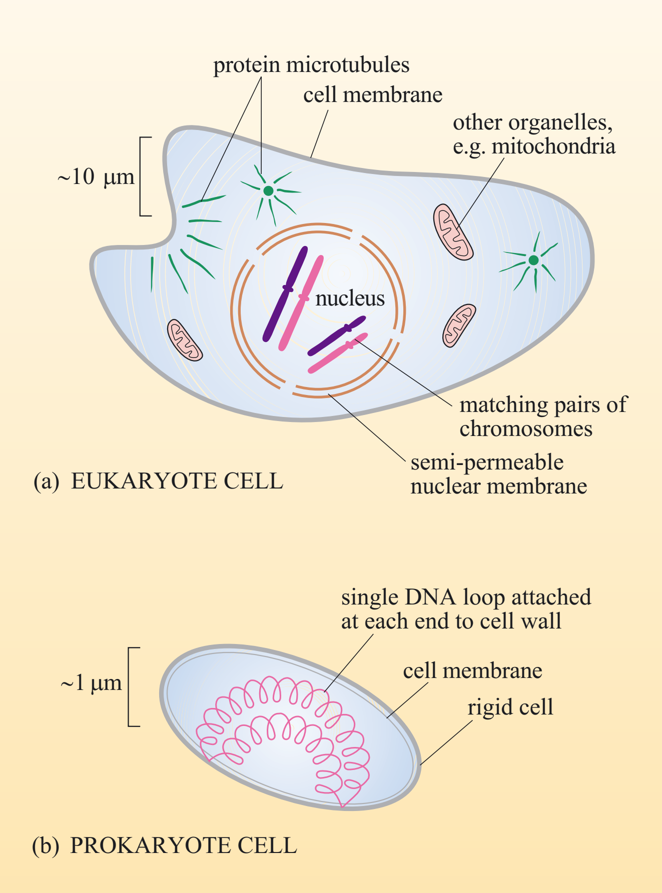


Figure 2.4 The arrangement of DNA in (a) a eukaryote cell and (b) a prokaryote cell.

Box 2.2 The kingdoms of organisms

Until well into the 20th century, it seemed suitable to divide organisms into only two broad kingdoms: plants and animals. Plants are mostly **autotrophic** (from *auto* meaning self and *trophic* meaning feeding) and **sessile** (fixed), while animals are mostly heterotrophic and mobile.

As techniques of study improved, particularly with the advent of electron microscopes, it became clear that bacteria, which had traditionally been lumped with the plants, differed profoundly in structure from all other organisms. Hence, the fundamental division between prokaryotes and eukaryotes was established. However, many eukaryotes still posed problems for the old plant/animal division. Fungi, for example, contradict their traditional ‘plant’ status by being heterotrophic, as well as showing various unique chemical and structural traits. Moreover, unicellular eukaryotes present a highly confusing picture. For example, some eukaryotic forms swim by means of a whip-like structure called a flagellum, yet possess chloroplasts for photosynthesis, while other closely related flagellate forms feed heterotrophically by ingesting prey. For a time, a convenient solution seemed to be to put all the unicellular eukaryotes in another kingdom, Protista, so leaving the multicellular forms to be divided between plants, animals and fungi. Further detailed studies of cell structure and differences in DNA sequences have caused even this neat compromise to break down as further significant divisions were recognised, some cutting across the unicellular/multicellular divide. At present, there is no universally agreed scheme and evolutionary relationships are still debated. To avoid getting lost in detailed discussion, the pragmatic (if not strictly evolutionary) divisions of the eukaryotes shown in Table 2.1 will be adopted in this book.

Table 2.1 Divisions of the eukaryotes.

Kingdom	Description
Animalia	Multicellular, heterotrophic forms that ingest their food.
Fungi	Largely multicellular (though with incomplete dividing walls), heterotrophic forms that feed on organic molecules, which they either absorb directly or break down externally with enzymes and then absorb.
Plantae	Largely multicellular, autotrophic forms, comprising land plants (all multicellular) and algae (both seaweeds, which are multicellular, and related unicellular forms).
Protista	Other unicellular forms, which may be either autotrophic or heterotrophic. As employed in this book, this grouping is still a motley assembly of several only remotely related groups, some of which share relatively more recent common ancestors with certain of the other kingdoms.

2.3.2 Eukaryote beginnings

Fossil evidence of the earliest eukaryotes is extremely sparse. The first forms are likely to have been single-celled and devoid of any skeletal hard parts and as such, their potential for preservation would have been slim. The discovery of exceptional fossils of such organisms, therefore, has a strong component of luck, but each year new records are discovered/appear that push the inferred schedule for various evolutionary accomplishments further back in time. Even when they are found, the fossils may be difficult to interpret as little critical structural information tends to be preserved. Criteria for recognising eukaryotes are themselves prone to reassessment as more is learnt about the characteristics of living forms and the processes involved in the preservation of the fossils themselves. Figure 2.5 summarises the evidence available at the time of writing – you may choose to add any subsequent information that you encounter.

What features should be used to identify eukaryote fossils? Complex histories of chemical alteration of the original organic materials of the organisms, if preserved at all, effectively limit the analysis of body fossils to their structural features. The scale of resolution of these depends upon the circumstances of preservation. The finest preservation is found where organic materials have been directly coated by microscopic mineral crystals very soon (maybe only days or even hours) after death. The potential resolution of this style of preservation depends on the size of the mineral crystals, which is analogous to the 'grain' in a photograph affecting the detail of the image. Rare examples of petrified soft tissues are known from Phanerozoic deposits in which even the shapes of cell nuclei can be discerned. However, the preservation of structure is usually somewhat cruder. Where only impressions remain in the enclosing sedimentary rock, the grain size of the host rock itself limits the available resolution.

Question 2.4

From what you have learnt so far about the differences between prokaryotes and eukaryotes, which diagnostic features of eukaryotes might you have some chance of detecting in the fossil record?

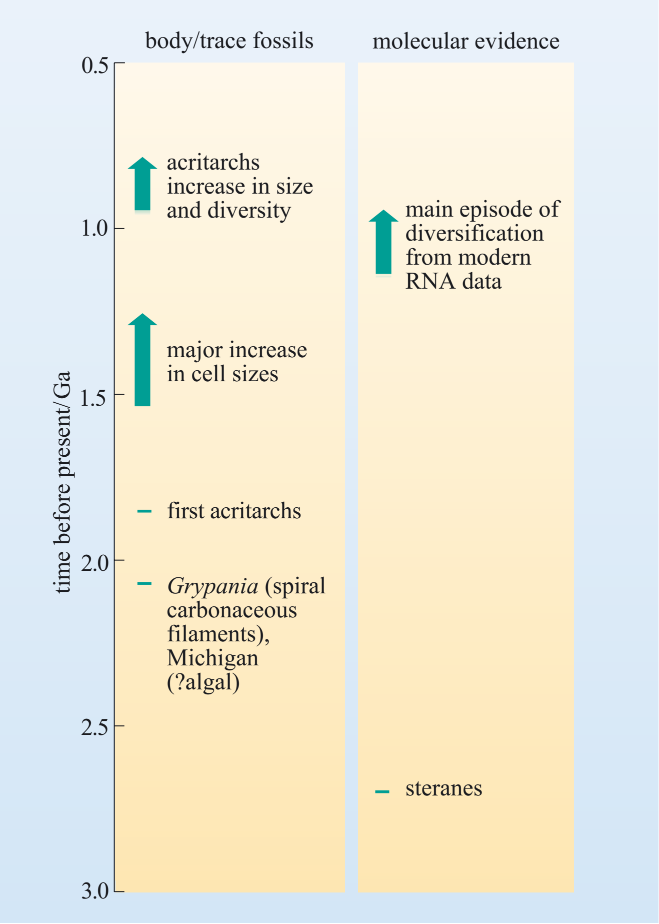


Figure 2.5 Evidence for the early evolution of eukaryotes.

You might also have thought of the characteristic product of **meiosis** (e.g. a quartet of **haploid** cells, Box 2.3), and some Proterozoic fossil tetrahedral cell clusters have indeed been interpreted as such. However, because such clusters might equally well have been produced through successive fissions by prokaryotes that failed to separate, this criterion is regarded as unreliable.

In practice, all three possible criteria given in the answer to Question 2.4 have proved problematical in the search for the oldest eukaryote fossils. Dark spots within some Proterozoic fossil cells preserved in cherts have been interpreted by some researchers as representing organelles, but the development of similar features produced through cell collapse in rotting prokaryotes gives cause for doubt. Nor is evidence for cellular organisation particularly pertinent in this context, as tissue differentiation only occurred much later in multicellular eukaryotes. Even cell size is not strictly diagnostic, as there is some overlap in the size ranges of prokaryote and eukaryote cells. Significantly increased size in fossil cells is, however, regarded as reflecting the spread of eukaryotes. In Proterozoic rocks deposited over the time interval from about 1.6 Ga until about 1.4–1.2 Ga, the average size of fossil cells more than doubles. Marine shales dating from about 1.4 Ga, moreover, yield resistant organic cell coats, called acritarchs (Figure 2.2), most of which probably represent the cysts (thick-walled resting stages) of eukaryotic algae that floated in the seawater.

Chemical fossils (chemofossils) push the origins of the eukaryotes even further back in time than the microfossil evidence cited above. Substances called **steranes** (Figure 2.5), formed from sterols (of which a familiar example is cholesterol), are known only from eukaryotic cell membranes that have been recorded from Australian petroleum deposits dated to at least 1.7 Ga. The interpretation of yet older fossils remains inconclusive. Carbonaceous filaments abundant in some formations dating from about 1.3 Ga onwards have also been recorded in rocks as old as 2.1 Ga (e.g. *Grypania*, Figure 2.5). These are commonly interpreted as multicellular eukaryotes that were probably algae of some sort, although a prokaryotic identity for at least some of them cannot be ruled out. Recently reported examples from China dating to about 1.7 Ga are considered to be eukaryotes.

All that can be reliably gleaned from the fossil record then, is that the eukaryotes, as recognised on the limited criteria discussed above, had arisen at least by the middle of the mid-Proterozoic (by about 2.1 Ga), if not some time earlier. How they evolved remains open to conjecture, although a fascinating hypothesis first proposed some time ago, and which is consistent with a number of features seen in living eukaryotes, has been championed by the American biologist Lynn Margulis. Some eukaryote organelles bear striking resemblances to bacterial cells; for example **mitochondria** (singular mitochondrion), which are responsible for the energy supply derived from aerobic (oxygen-consuming) respiration, contain small amounts of DNA in a simple loop, similar to that found in bacteria, although some of the DNA responsible for a functional mitochondrion is housed in the cell nucleus. **Chloroplasts**, the organelles that effect photosynthesis in green plant cells, show a similar pattern, and can even undergo **binary fission**. From this and the evidence of other resemblances, Margulis suggested that these organelles were indeed once independent prokaryotes that took up symbiotic residence inside ancestral eukaryotic host cells. This is known as the **endosymbiotic hypothesis**, proposed for the origin of eukaryotic organelles.

Not all the organelles can be so readily explained; the nucleus in particular has its long chains of DNA in protein-bound chromosomes (Section 2.3.1) and shows other marked differences from bacteria. It probably originated from within the original cell structure of some distinct kind of prokaryotic ancestor (which then became the primordial eukaryote host cell type for the later acquisitions mentioned above). One effect of replacing a single DNA loop with multiple chromosomes is that the process of DNA replication during cell division can be carried out at the same time on different chromosomes. It is thought that this parallel processing approach may have been an adaptation for speeding up the replication of increasing amounts of DNA.

If at least some eukaryote organelles were thus secondarily acquired through endosymbiosis, the original ancestral host may have differed from other prokaryotes only in the possession of a nucleus. This primordial eukaryote line may well have extended back into the Archaean as a product of the early evolutionary diversification of the prokaryotes.

Question 2.5

Casting your mind back to the earlier discussion of the structural differences between eukaryote and prokaryote cells, what difference from the usual prokaryote condition would have been a prerequisite for endosymbiosis?

The living bacterium *Thermoplasma* (one of the Archaebacteria), which is adapted to extreme conditions of high temperature and acidity, lacks a rigid cell wall, and in this respect provides a plausible model for the eukaryote ancestral host cell. The ancestral form is thought likely to have been an anaerobically respiring heterotroph (Figure 2.6) that in due course adapted to rising oxygen levels through symbiosis with aerobic bacteria capable of mopping up the oxygen. These eventually became the mitochondria.

- What immediate and later advantages would have been gained by the host from such an acquisition?
- The aerobic bacteria would have helped to remove molecular oxygen, which would have been toxic to the anaerobic host. Later, the host could exploit the energy yield of its aerobic guests.

Photosynthesis may also have arisen in eukaryotes through the acquisition of autotrophic symbionts that eventually became chloroplasts. These various acquisitions of organelle precursors were not necessarily unique events and the diversity of chloroplast types in different groups of algae implies independent acquisition in several different lines. In some cases it appears that the chloroplast precursor was itself a chloroplast-bearing eukaryote because a nuclear relic is present within the chloroplast structure; the brown algae (e.g. kelp), common along many shorelines, are one such group. There is as yet no evidence for mitochondria having been acquired more than once, but then again nothing is known of the mitochondria in the myriads of organisms that have become extinct.

The timing of this hypothesised evolutionary assembly of the variously 'kitted-out' eukaryotes can only currently be guessed at, as direct fossil evidence for the appearance of the organelles in question has not yet been found. Indeed, the succession of endosymbiotic unions may well have extended over a long interval, although in the case of the spread of mitochondria, any time after the appearance of molecular oxygen in the atmosphere and oceans would seem plausible.

Question 2.6

On this basis, very approximately, when might mitochondria-bearing eukaryotes first have begun to proliferate?

Such early evolutionary advances appear to have had only a limited initial impact on the Earth's ecosystems – or at least this appears to be the case for those ecosystems represented by the fossil record. As noted earlier, the largely prokaryote-dominated mats that yielded stromatolites continued to dominate shallow sea-floor communities for another 1000 million years, until their rapid decline from around 850 Ma. The main evolutionary divergence of eukaryote stocks, including most of the various chloroplast-bearing groups, seems to have occurred only after a prolonged period of time.

2.3.3 Eukaryote architecture and evolution

Two attributes of eukaryote cells have been of paramount importance in the evolution of their diversity and complexity. First, their flexible cell membrane is supported internally by the molecular framework of protein rods. These microfilaments and microtubules can grow or shorten and so act rather like a system of extendible tent poles inside a tent. The eukaryote cell, unlike the prokaryote cell, can therefore change shape and engulf external objects. This

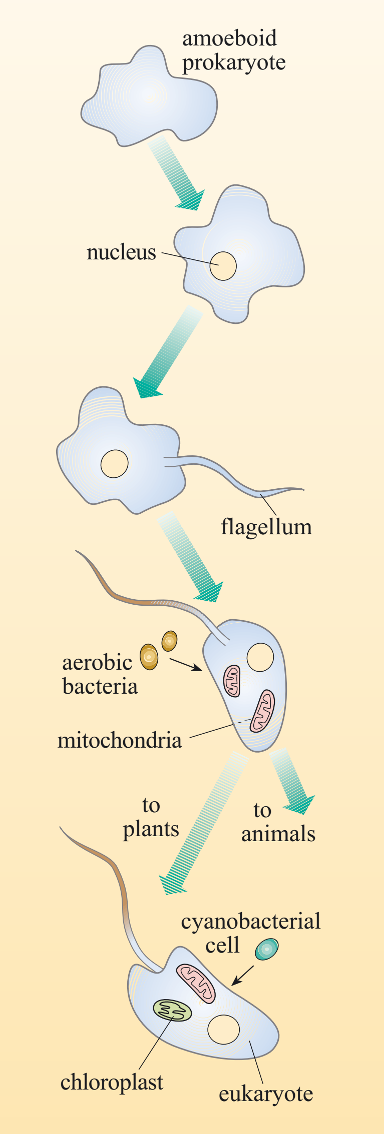


Figure 2.6 The sequence of possible acquisitions of organelles by eukaryotes according to a modern version of the endosymbiotic hypothesis. The first step (initiating the primordial eukaryote line) was the formation of a nucleus. Thereafter, some of the organelles at least may have been acquired through endosymbiosis of various prokaryotes, as shown.

difference has permitted the addition of consumption by physical ingestion to the repertoire of feeding modes, and ultimately, to the build-up of multi-tiered **trophic pyramids**. The microtubules also play other crucial roles, for example in cell division by controlling internal structure (described below).

The second attribute is the vastly greater number of genes in eukaryotic cells that permits more specialised development. Multicellular forms, in particular, may contain a huge variety of cell types, making up tissues with widely different functions. Just consider those that make up your own body, for example. Specialised cells produce bones, skin and connective tissue for support and protection, while muscle cells allow movement, itself coordinated by nerve cells. A variety of secretory cells produce everything from hormones, controlling growth and behaviour, to digestive juices and mucus, while blood cells are involved in gas exchange and the immune system. Reproduction is effected by sex cells. And so the list goes on.

All these functions are underwritten by genes and nearly all cells possess the genetic information to make potentially any type of cell in the body. Development from a single initial cell means that the genes have not only to be reliably replicated as cells repeatedly divide, but their activity must also be regulated through interactions between genes to produce different types of cell. Such a complex set of instructions involves a lot more DNA than that found in the simple loop within a prokaryote. Whereas a typical bacterium may have some 4 million **base pairs** in its DNA, that of a human, for example, has about 3.5 billion base pairs (though probably only up to a quarter of this DNA actually codes for proteins). The management of so much DNA poses its own problems. For example, the total number of mutations is unavoidably increased, despite the existence of complex enzyme-controlled DNA repair mechanisms. For although the probability of mutation at any one DNA base site is extremely low (around one in a billion per replication), it is now compounded over billions of sites, so mutations are likely to arise at almost every cell division. Where these affect the functions of the proteins coded for, most will be of detrimental effect, and only a few, by chance, will be beneficial. Without some means of compensating for such errors, viable development of multicellular forms would be virtually prohibited.

One way the problem of harmful mutations is mitigated in eukaryotes involves a doubling up of chromosomes in the cell nucleus to yield what is termed a **diploid** complement of chromosomes: each chromosome thus has a homologous partner, containing corresponding genes. If a gene in one set becomes corrupted by a mutation then, to some extent, its function can be covered by the matching gene from the other set. Given the low probability of mutation of any one gene, the chances of both homologous genes being coincidentally affected is almost negligible. A twin-engined aeroplane provides an analogy: it may be able to continue flying even if one engine fails, whereas a single-engined craft will crash if its engine fails. Nevertheless, chromosome doubling by itself is an insufficient insurance against the risk of disadvantageous mutation when multiplied by the colossal span of geological time.

Question 2.7

Can you think why this should this be so?

A further eukaryote innovation, **sexual reproduction**, gets around the problem of disadvantageous mutation by regularly mixing together genes from different individuals. Two steps, **mitosis** and **meiosis**, are entailed in sexual reproduction (Figure 2.7). Fusion of special parental cells called **gametes** (e.g. sperm and egg cells), each containing only a single or **haploid** set of chromosomes, produces a new diploid cell (the fertilised egg). For the process to be repeated in subsequent generations without continually doubling chromosome numbers, there clearly has to be a corresponding process to halve the number of chromosomes again between generations, i.e. to form haploid gametes. Normal cell division, known as mitosis (see Box 2.3), which is involved in body growth, conserves chromosome numbers in the nucleus, whereas meiosis, a special kind of cell division, halves the number of chromosomes, whilst also exchanging segments of DNA between the matching chromosome pairs of the diploid parent (see Box 2.3). Together, meiosis to yield haploid gametes and fusion of the gametes to generate new diploid combinations, shuffle and distribute the genes of parents so as to deal a unique genetic hand to each of their offspring (except in the case of identical twins).

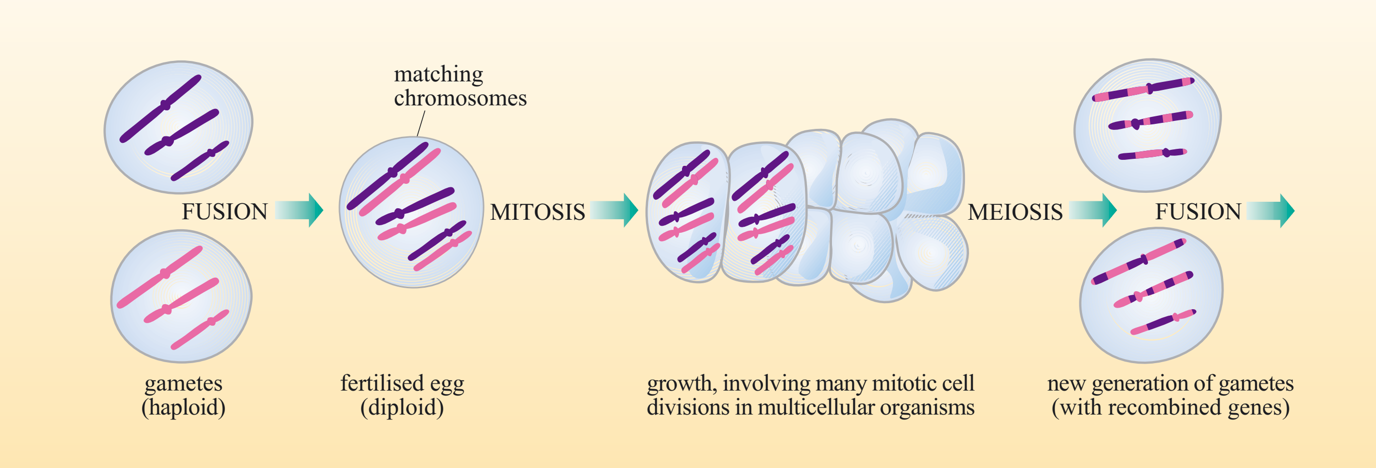


Figure 2.7 The cycle of sexual reproduction. See text for explanation.

Box 2.3 Cell division in eukaryotes – mitosis and meiosis

Mitosis is an elaborate mechanism for cell division that has evolved in eukaryotes and ensures the rapid duplication of DNA in the daughter cells with a high degree of accuracy. Prior to cell division, the DNA sequence of each chromosome (Figure 2.8a) replicates and the pairs of replicate strands for the time-being stay together (Figure 2.8b). A spindle-shaped structure of protein microtubules then forms within the cell, with the chromosomes aligned along its equator (Figure 2.8c). At about the same time, the nuclear membrane breaks down. The replicates of each chromosome separate and are pulled by the microtubules to the opposite poles of the spindle (Figure 2.8d). There, each set of chromosomes becomes re-enclosed by a new nuclear membrane, while the spindle disintegrates (Figure 2.8e). Inward pinching of the outer cell membrane completes the process of mitotic cell division. This mode of cell division ensures that the huge amounts of replicated DNA are correctly divided between daughter cells, such that each receives an identical full set of chromosomes.

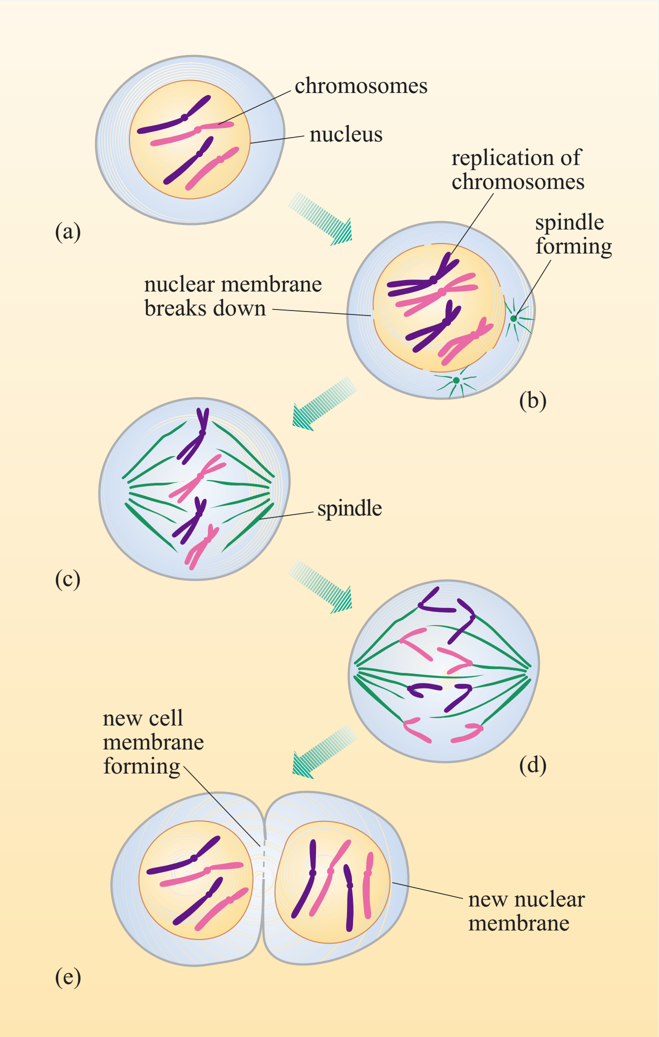


Figure 2.8 The process of mitosis. The chromosomes in (a) are shaded differently to indicate that they were originally from different individuals. See text for explanation.

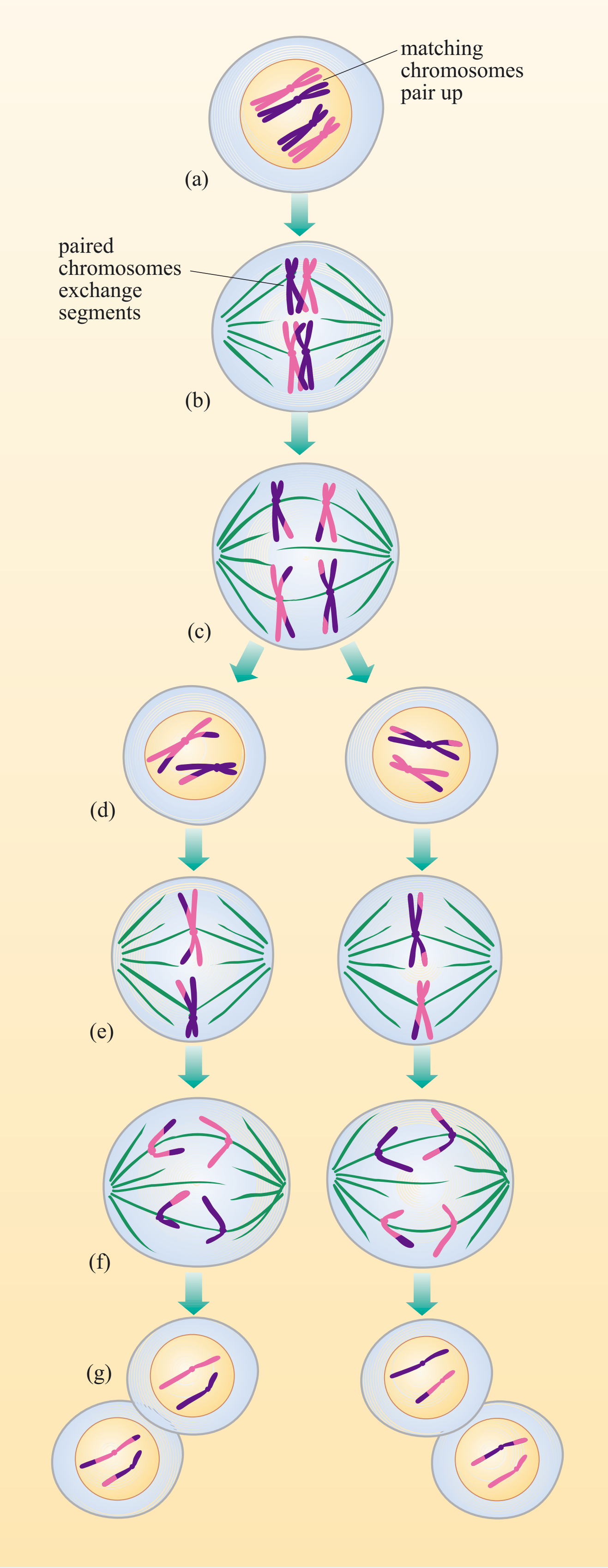


Figure 2.9 The process of meiosis. See text for explanation.

A second type of division called meiosis produces haploid daughter cells (i.e. each daughter cell has half the original number of chromosomes) and forms the counterpart to fusion in the cycle of sexual reproduction (Figure 2.7). Meiosis again employs spindles, though in two successive divisions. As before, the already replicated chromosomes (Figure 2.9a) become aligned along the equator of a spindle for the first division (Figure 2.9b). However, in contrast to mitosis, homologous chromosomes, derived from the fusion of parental gametes, are now also paired up alongside each other. A very curious process then ensues. Each homologous pair (there are now four strands in all in each bundle, as each chromosome is already replicated) proceeds to twist along its length, in the manner of a towel being wrung out (Figure 2.9b). The replicated strands repeatedly snap, but their loose ends re-attach, though not necessarily to those from which they have just separated. Consequently, the four entwined strands randomly exchange segments of their gene sequences with each other. Eventually, the chromosome pairs separate again (Figure 2.9c) and migrate to opposite poles of the spindle to form two daughter cells (Figure 2.9d). The second division of the daughter cells is essentially mitotic in character (Figure 2.9d–g), separating the two strands of each chromosome (with their shuffled sequences). A quartet of haploid daughter cells results, with each cell containing only a single unpaired set of chromosomes (Figure 2.9g). Together, the crossing over of bits of the homologous chromosomes and the independent assortment of the paired chromosomes into the daughter cells bring about a thorough shuffling, or recombination, of the parental genes. No two daughter cells (e.g. sperm or egg cells) receive the same half share of genes. Harmful mutations can therefore be effectively screened out by natural selection.

The evolution of sexual reproduction in eukaryotes has had an enormous impact on the history of life, for it has altered the very rules by which evolution through natural selection proceeds (Figure 2.10). To a large extent, prokaryote evolution is a straight contest between clones (races of genetically identical individuals). Their reproduction by simple fission (involving straightforward replication of their DNA loop) produces virtually identical offspring, varying through (rare) mutation only, and many individuals must be produced to bring about a favourable combination of genes by coincidental mutation (Figure 2.8a). In contrast, the small size and rapid reproduction of prokaryotes means that the race can certainly be fast and furious as clones with new advantageous mutations supplant others. Sexual reproduction in eukaryotes provides a means for regularly varying the genetic complement that an individual passes on to its offspring through recombination (Figure 2.8b). However, things are not quite as simple as that, for some exchange of DNA between individuals can occur, providing an additional, if irregular, source of genetic variation. Potentially, all possible permutations of the genes available in a population can be assembled in different individuals and played off against one another in the Darwinian struggle for existence. Those genes that generally combine to yield the best adapted individuals will be selectively preferred through the generations. A beneficial mutation can be rapidly promoted (i.e. its frequency in the population increased) by natural selection, unhampered by the other genes present in the original individual in whom it arose. The great leap in the efficacy of natural selection that this represents means that even large and complex multicellular organisms, with relatively slow rates of reproduction (compared with those of prokaryotes), can nevertheless evolve rapidly.

There are, however, several reasons why, at the individual level, asexual reproduction can offer advantages over sexual reproduction. These are:

- energy (hence resources) is not wasted on finding a suitable mate and achieving fertilisation
- a female in a sexual population who switched to asexual reproduction would endow each of her offspring with 100% rather than only 50% of her genes, and thus stand to pass on twice as many of her genes to subsequent generations
- proven favourable combinations of genes risk being broken up by sexual reproduction.

In view of these and other advantages of asexual reproduction, it is difficult to see how sexual reproduction could have gained a foothold and spread through populations at all; natural selection should surely suppress sexual reproduction because of the disadvantages at the individual and population levels. Several hypotheses have, however, been proposed by evolutionary biologists to counteract such arguments, referring to the possible advantages of sexual reproduction at the individual level. One model, for example, considers the effect of parasites; rapidly reproducing parasites can soon adapt to a given combination

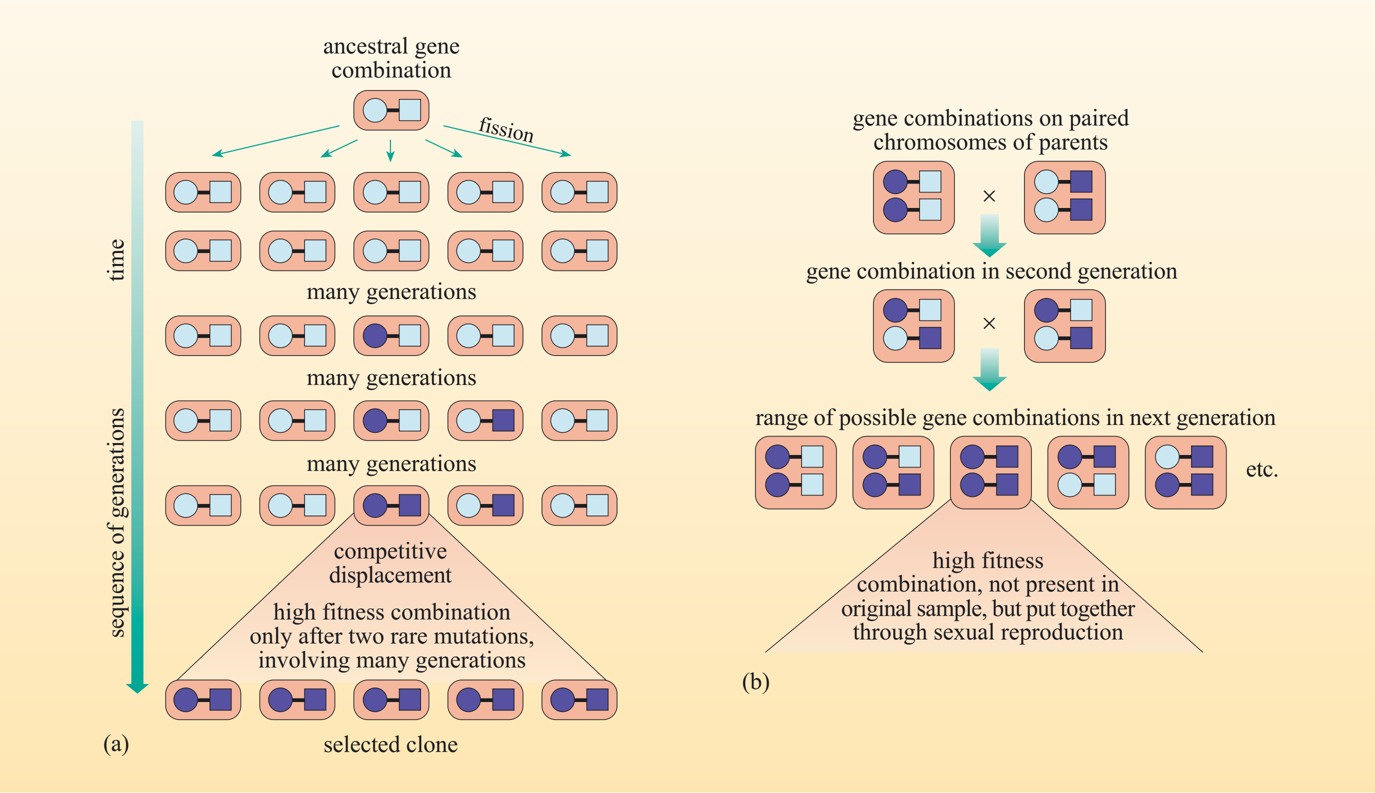


Figure 2.10 The effect of sexual reproduction on natural selection: (a) competition in asexually reproducing organisms is between clones, with variant offspring produced by rare mutations (represented by the dark symbols); (b) sexual reproduction furnishes each individual with its own unique permutation of genes, offering greater genetic variety for natural selection to act upon. The light and dark symbols represent alternative gene types. Only two genes on each chromosome are shown for the sake of simplicity. A high fitness combination of genes is one that, through preferential survival and/or reproduction of individuals bearing it, becomes relatively more frequent in subsequent generations.

of genes in their host organism and can thus threaten the survival and reproductive success of all potential host individuals with that combination of genes. Asexual reproduction provides virtually no escape from this threat; the offspring will automatically carry the same combination of genes. Sexual reproduction, on the other hand, does offer an escape by varying the genetic complements of offspring, such that some at least may prove robust if not immune to infection by the parasites. So there may well be advantages to sexual reproduction at the individual level offsetting the disadvantages, which could explain how it became established and maintained in populations. Once there, it can provide the fortuitous gift of evolutionary flexibility to those species that have adopted it, and it is to this effect that the eukaryotes owe much of their evolutionary diversity.

2.3.4 Diversification of the eukaryotes

The fossil record of eukaryotes points to a marked proliferation of types in the late Proterozoic era. Though known from older rocks, acritarchs for example become abundant and diverse in rocks dating from around 1 Ga onwards. Andrew Knoll, from Harvard University in the USA, has led an intensive study of a sedimentary succession of marine origin, ranging in age from 850 Ma to 600 Ma, in the Arctic island of Spitsbergen. Fossil records reveal the presence of both prokaryotes and eukaryotes from a variety of habitats. However, while the prokaryotes look very much like those living in corresponding environments today and show little evidence for change, the eukaryote fossils present a different story. They reveal a marked diversification, with a relatively rapid turnover of species up through **sedimentary succession**. Alongside single-celled forms similar to living unicellular green algae and dinoflagellates are various multicellular algae (seaweeds). Conspicuous by its absence, however, is any fossil evidence for animal life. Subsequent studies elsewhere appear to corroborate this story.

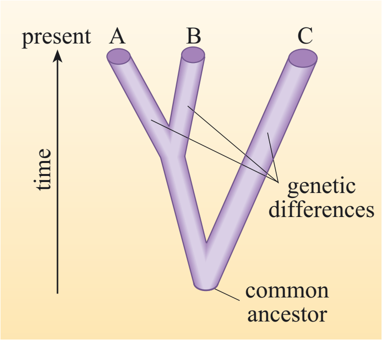


Figure 2.11 The relationship of genetic differences between organisms and their times of divergence from common ancestors, according to the molecular clock theory. As the lineage leading to C split first from a common ancestor to A and B, its genetic sequence would have had more time to change independently than the lineages leading to A and B. Providing that the average rate of change in all lineages stayed the same, the genetic sequences of A and B would remain more similar to each other than to that of C. Such reconstructions of evolutionary trees are only partial because fragments of fossilised DNA sequences are exceedingly rare and analysis of them is fraught with problems.

Another kind of evidence comes from comparing the DNA and/or RNA sequences of living eukaryotes (i.e. the amino acid sequences of proteins derived from DNA and RNA). The hypothesis is that as evolutionary lineages diverge from a common ancestor, so too do the amino acid sequences, as a consequence of cumulative (advantageous) mutational change in each lineage. There are theoretical reasons backed up by empirical evidence for presupposing that the average rate of such divergence remains approximately constant, at least for particular parts of the sequences. This proposal is known as the **molecular clock theory**, and refers to the supposedly clock-like gradual change of the sequences. In so far as this theory is correct (and there is plenty of debate about that), it implies that, for any three organisms, the pair that shows the smallest mutual divergence of sequences shares the most recent common ancestor (Figure 2.11). On this basis, a hierarchy of evolutionary relationships can be established for large numbers of living species, from which the shape of their evolutionary family tree can be partially reconstructed.

Absolute dating of some of the evolutionary branching points reconstructed from molecular data may be established from fossil evidence where the record is sufficiently complete. Such calibration allows the average rate of change of the molecular clock in question to be estimated and the whole evolutionary tree accordingly set against an absolute timescale. At the very least, the method provides a means of producing hypotheses concerning the pattern and chronology of evolutionary history that can be further tested against the direct fossil record.

Comparisons of certain RNA sequences from living organisms suggest that the majority of the major eukaryote groups separated from one another in a flurry of evolutionary diversification from common ancestors between about 1.2 Ga and 1.0 Ga. What sparked this burst of evolutionary activity is again open to conjecture. As with the origin of the eukaryotes, the fossil record bears witness to the resultant evolutionary explosion, but is mute on how it came about. There is no geological evidence for any significant environmental change (e.g. of atmospheric composition or climate) around the time that might be considered to have precipitated this evolutionary explosion. A pertinent biological observation though, is that those groups of living single-celled organisms that appear to represent the earliest branches of the eukaryote evolutionary tree are all effectively asexual.

Question 2.8

What hypothesis might be deduced from this observation to explain the main burst of eukaryote diversification (i.e. that the earliest branches were effectively asexual)?

Though only a hypothesis, the answer to Question 2.8 highlights the important point that not all major evolutionary events need be explained by reference to external environmental triggers. Internal modifications may fortuitously open up major new opportunities for further evolutionary change, and the evolution of sexual reproduction in eukaryotes is a likely example, offering an explanation for what might be thought of as one of the biggest gear changes in evolutionary history.

Whatever the circumstances, from 1.2–1.0 Ga onwards, the takeover of the Earth's ecosystems by eukaryotes seems to have become almost unstoppable. Meanwhile, the diversity of stromatolites, which had reached a peak at ~1.0 Ga, began to decline, especially from about 850 Ma, although they remained widespread until Phanerozoic times, when they became largely restricted to stressful environments (e.g. highly salty bays or tidal flats) or cryptic habitats (such as cavities within reefs). Whether these changes reflect direct interactions between the prokaryotes and the eukaryotes, or whether stromatolite-building microbes declined in response to other environmental stresses, with various eukaryotes moving in on their vacated habitats, remains unclear. This is not to say that prokaryotes became unimportant – far from it! Rather, they adapted to the eukaryote world in a variety of ways. Some entered into symbiotic relationships (including disease), and others adopted specialised roles, often in extreme conditions. Indeed, all life continues to rely upon the fundamental biochemical activity of prokaryotes for its continued existence.

2.4 The carnival of the animals

2.4.1 Ediacaran faunas

Section 2.3 described how fossil evidence for animals in the first major episode of eukaryote diversification is absent. Fossils widely interpreted as ‘animals’ make a late but spectacular entry in the stratigraphic record of the **Ediacaran Period** (610–542 Ma), representing the last 70 Ma or so of the Proterozoic Era. From many parts of the world, assemblages of characteristic types of body fossils (Figure 2.12) have been recovered, which are collectively known as **Ediacaran faunas** after the Ediacara Hills in South Australia, from where many such fossils were first described in detail.

With the exception of a few small, simple forms in northwestern Canada, dated to ~600 Ma, diverse Ediacaran assemblages range in age from at least 565 Ma until the end of the **Vendian** period (650 Ma to 542 Ma). A few distinctive types are known to have survived until mid-Cambrian times, especially in deeper-water habitats.

2.4.2 Interpreting the Ediacaran fossils

What were these Ediacaran organisms like? Perhaps their most striking feature is their size, with some of the later forms reaching up to a metre or so in length. Yet there is no evidence for them possessing any supporting skeletal hard parts. The fossils are found in sedimentary rocks originating from a variety of environments, ranging from deep marine areas to inshore, even tidal flat, settings. The fossils are preserved only as flattened impressions at the base of sandy to silty sedimentary layers. Palaeontologists, therefore, have to interpret the Ediacaran fossils on the basis of their gross anatomical features. Needless to say, opinions have differed quite sharply. Earlier descriptions usually saw them being allocated to those phyla (major groups) still surviving today with which they seemed to have most in common (even if only superficially). Thus, various kinds of segmented annelid worms (similar to ragworms), primitive elongate arthropods (shrimp-like animals) and bun-like echinoderms (relatives of sea-urchins) were all recognised in this the Ediacaran fauna, along with assorted cnidarians (jellyfish and colonial polyps). However, many of these identifications remained problematical. For example in only one form, *Dickinsonia* (Figure 2.12a), have structures interpreted as guts been detected. Moreover, the intact condition of the fossils, even in sandstones that were deposited in shallow, agitated and presumably well-aerated water, raises questions as to how they could have been preserved. Today, a combination of predation, scavenging and decay, not to mention disturbance by burrowing organisms, virtually rules out such a style of preservation of soft-bodied organisms in equivalent environments.

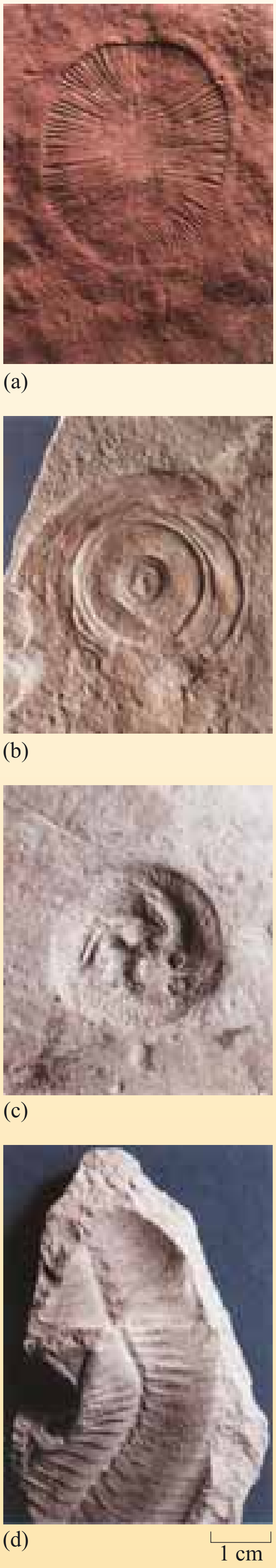


Figure 2.12 Examples of some Ediacaran fauna fossils: (a) *Dickinsonia*, an elongate pancake-shaped worm; (b) *Cyclomedusa*, a jellyfish; (c) *Tribrachidium*, a bun-shaped organism with three spiral tracts on its upper surface; (d) *Inkrylovia*, an elongate bag-like form with transverse partitions. All photographs are to the same scale.

One explanation given for these exceptional assemblages is that large-scale predators, scavengers and deep burrowers simply did not exist at that time, making the Vendian a sort of privileged window of opportunity for preservation. Another proposal views the majority of Ediacaran organisms as a distinct eukaryote group separate from the existing kingdoms, and of very simple construction. They are portrayed as having had a quilted mattress-like body filled with fluid enclosed in a tough cuticle (explaining their good preservation), and to have not yet evolved a digestive system consisting of guts. It is suggested that they either absorbed dissolved food molecules over their broad body surfaces or that they may have contained autotrophic symbionts within their tissues, avoiding the need for a digestive system. At least in the case of *Dickinsonia* (Figure 2.12a), however, this interpretation can be rejected as, apart from the possible gut structure, one specimen has been described as having been preserved with its segments torn across prior to burial. The impression it created in the underlying sediment is just as deep as that made by intact specimens, showing that the body was firm and not, after all, fluid-filled. Another researcher has suggested that the Ediacaran organisms may perhaps have been lichens (consisting of symbiotic associations of algae and fungi). Most authorities, however, would regard the pendulum as having swung rather too far from the conventional view in such interpretations.

The most widely accepted hypothesis at present is that most of the Ediacaran fossils represent primitive animals, with the majority having a grade of organisation equivalent to that of the cnidarians. Living cnidarians have a sac-like body wall directly surrounding a digestive cavity that is reached via a single orifice as, for example, found in a sea anemone. The tough body wall consists of outer and inner cellular layers that are separated by a gelatinous layer, and in a few forms segmentation and bilateral symmetry of the body seem to be well developed (Figure 2.12a). This may perhaps be regarded as the primitive relatives of more advanced animals. Certain problematical forms, however, may still turn out to be algae rather than animals.

In addition to these originally soft-bodied forms, a few kinds of skeletal components are also known from the Late Vendian period, including some networks of sponge spicules (mineralised needles that provide a skeletal framework for many sponges).

The recognition of Vendian animals is consistent with molecular evidence that suggests an even earlier time of origin, using the molecular clock theory. Figure 2.13 shows the estimated divergence of the amino acid sequences of the protein haemoglobin between various pairs of living animal species plotted against the minimum possible ages for the original divergences from common ancestors based on fossil evidence. The minimum age for divergence in each case is taken as that of the oldest fossil that can be attributed to one, but not to the other, of the evolutionary branches in question: the existence of such a fossil shows that the split from a common ancestor must already have taken place. As there may have been successive amino acid substitutions at any given site on a haemoglobin molecule, values for the *cumulative* amounts of divergence may exceed 100% and the maximum is 190%.

Haemoglobin is the red pigment that transports respiratory gas molecules in the human bloodstream.

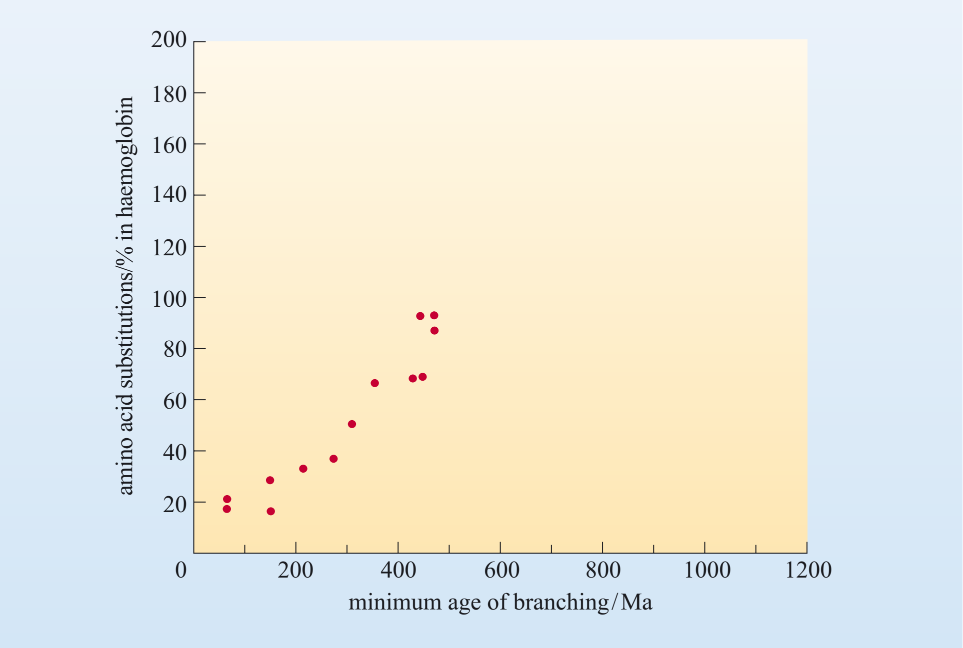


Figure 2.13 Measures of the relative amounts of divergence in the amino acid sequences of haemoglobin between pairs of living animal species plotted against the time by which divergence from a common ancestor must have occurred, based on fossil evidence.

Question 2.9

Using the molecular clock theory as a guide, extrapolate from the data shown in Figure 2.13 to deduce the inferred latest time of origin of the animals, based on the maximum value of haemoglobin divergence given above. (*Hint*: note that the origin of the animals can be assumed to have preceded the first divergence to have taken place within the group.)

The value derived in Question 2.9 is, of course, only a minimum estimate due to the possibility of yet more divergent forms having become extinct. On the other hand, the molecular clock may not itself have been regular, so the conclusion is by no means infallible.

The Ediacaran animals thus appear to represent a major evolutionary proliferation of large-bodied animals some time after the appearance of the first animals, which may date back to around 1.0 Ga.

2.5 The rules of the new evolutionary game

But what was it about the evolution of the eukaryotes and, ultimately, the Ediacarans that was so innovative? One of the most important developments was sexual reproduction (Section 2.3.3). In the final part of this chapter it is worth following through the theoretical implications of the innovation of sexual reproduction for the evolution of adaptations in organisms. Analysing who or what benefits from adaptations will provide a useful perspective for ongoing considerations of the relationship between the Earth and life.

The characteristic adaptedness of organisms is a product of evolution by natural selection, which may be summarised as follows. The reproductive potential of a population usually well exceeds the numbers that can be sustained, giving rise to Darwin's struggle for existence. Individuals possessing heritable variations that promote their survival and reproduction in the prevailing circumstances relative to other individuals, preferentially endow subsequent generations with their offspring and hence their genes. Given the genetic mixing of sexual reproduction, genes associated with favourable traits tend to increase in frequency in the population. Through time, selection of these favourable traits leads to adaptations of form and behaviour with recognisable functions in relation to survival and/or reproduction. The process is hierarchical, with individuals being selected, and genes associated with adaptive features consequentially sorted, generation by generation.

Who, or what, then, are the beneficiaries of the process – individuals, genes or whole populations whose gene pools are affected? To try to answer this question, it is crucial to consider cause and effect. The efficient cause of the evolutionary process described above is natural selection acting on individuals. There is an immediate pay-off in the relative size of an individual's genetic legacy in the next generation. This is expressed as an individual's production of offspring, themselves surviving to be capable of reproduction, relative to other individuals – referred to as its **fitness**. The sorting of the genes, and any consequences for populations, however, are the incidental effects of such fitness differences. Hence, the only logically necessary beneficiaries of the adaptations that evolve are the individuals possessing them, who are rewarded with higher average values of fitness relative to other individuals. (The 'selfish gene' concept of Richard Dawkins, set out in his book with that title in 1976, argues that it is the selected genes that get the ultimate benefit. Yet that is, nevertheless, still contingent on the successes of the well-adapted individuals carrying them.)

At first sight, two evolutionary phenomena may seem to contradict the conclusion above concerning the benefits of adaptations. The first is what biologists call 'altruism' within species (though with no implication of intentionality), and it involves one individual behaving in such a way as to benefit the reproductive success of another, though at a cost to itself. The second is the coevolution of different species to yield mutually beneficial features, such as that between flowers and their insect pollinators. Closer analysis of both phenomena shows them to conform, albeit in subtle ways, to the conclusion above.

A classic example of altruism, for example, is the worker honeybee, which feeds the queen and her offspring, and may die defending them, while herself failing to reproduce. How could such apparently selfless behaviour have evolved? The answer to this conundrum was provided in the 1960s by the British evolutionary biologist W. D. Hamilton. In most sexual organisms, the mean amount of genetic overlap between siblings (i.e. genes received in common from the parents) is 50%, because each individual gets a random 50% of each parent's genes (Box 2.3). Social insects such as honeybees, however, have a rather curious pattern of genetic inheritance because males are haploid and females are diploid. Hence, females (including the workers) receive all of their father's genetic complement, together with 50% of their mother's genetic complement. Consequently, the majority (75% on average) of the genes received by each worker are shared with its siblings – all the queen's progeny, including new queen grubs. By helping the queen and her daughter queens who are especially adapted for copious reproduction, a worker ensures a far greater representation by proxy of her (shared) genes in subsequent generations than she might achieve by her own reproduction. So her effective genetic legacy or 'inclusive fitness', which takes those shared genes into account, is increased by her apparently altruistic behaviour. Such traits are said to have arisen by a special form of natural selection termed **kin selection**.

■ Why do you suppose the cells making up your own body do not compete with your sex cells to generate new individuals?

■ Since all the cells in your body are derived from a single fertilised egg, they all share the same genes. The sex cells, therefore, transmit the genes of all the other cells by proxy, leaving the other cells dedicated to their various specialised functions.

Evolved adaptations are effectively of selfish, transient benefit to the individual genetic entities concerned, and their rewards are strictly dependent on the prevailing circumstances. The adaptations provide no guarantee for the future welfare of the population (or even species) as a whole. A change in circumstances can precipitate extinction, as the testimony of the fossil record abundantly illustrates. Some adaptations may tend to prove disadvantageous to populations in the longer term, notwithstanding individual benefits in the shorter term. For example, in some environmental circumstances the acquisition of an asexual mode of reproduction by normally sexually reproducing organisms can lead to greater reproductive success (Section 2.3.3). However, those species that become fully asexual often appear to become extinct more rapidly than related sexually reproducing species. This is probably because of the associated decline in evolutionary flexibility of the asexual species. The short-term gain of individuals may eventually be at the longer-term expense of the population.

What is the implication of these arguments for the way in which living organisms interact with the Earth? Of great importance among the adaptations of organisms are systems of self-regulation, or **homeostasis**, which maintain stable conditions within the body in the face of a range of environmental perturbations. A familiar example of such a system is that which maintains your normal body temperature. In response to feeling cold, your body produces heat by various means (such as shivering), while it prevents overheating by, for example, sweating. Thus you are equipped with an integrated system of sensors (temperature-sensitive nerves in this case) and compensatory devices that maintains a constant core temperature in the body. Such internal constancy is advantageous because enzymes, which are themselves adapted to function optimally in the body's normal state, may be highly sensitive to fluctuating conditions. It is tempting to try to draw a parallel between such self-regulation within organisms and the system of feedbacks that appear to regulate conditions on the Earth. Such a parallel has been explicitly proposed in the Gaia hypothesis (Box 2.4).

Box 2.4 The relationship between life and Earth, according to the Gaia hypothesis

The Gaia hypothesis, first formulated by James Lovelock in 1972, asserts that ‘the climate and chemical composition of the Earth’s surface environment is, and has been, regulated at a state tolerable for the biota’ (Lovelock, 2000). The Earth and its biota are regarded as having evolved together as a tightly coupled system, with self-regulation of important properties, such as climate and chemical composition, arising as emergent properties. The whole system is thus explicitly likened in this respect to a ‘super-organism’. The kind of process Lovelock had in mind when proposing this hypothesis was the long-term maintenance of levels of molecular oxygen in the atmosphere sufficient to keep us alive, despite the short residence time of this highly reactive gas.

No theoretical explanation has been advanced as to why the feedbacks involved should necessarily serve to regulate conditions in the interests of the biota. Instead, a computer model to suggest how such a system might operate has been developed. Called ‘Daisyworld’, it envisages an idealised world similar to our own, receiving energy from a gradually warming Sun. The planet is imagined to have been seeded by two sorts of daisies, one black and one white. The essence of the model is that, to start with, when solar heating is modest and the surface temperature is below the optimum for the daisies, the black daisies warm up more quickly than the white ones and their growth is advantaged. As the black daisies spread, they reduce the planetary albedo, allowing more heat to be retained, and so help warm the atmosphere to an optimum level for the growth of daisies. Later, with increasing solar flux, optimum atmospheric temperatures may be surpassed in places, and it is the turn of the white daisies, whose greater reflectance helps keep them cool, to be at an advantage. As they now spread, displacing the black daisies, they increase the planetary albedo, and so serve to counteract the overheating. Thus, the planet and its biota together furnish a self-regulating system that, for a time at least, maintains optimum conditions for the daisies. Further refinements were later added to this model (with the addition of, for example, grey daisies and a fauna of rabbits and foxes), none of which was found to disrupt the self-regulatory nature of the system. The Daisyworld concept came to prominence in the influential 1980s nuclear thriller *Edge of Darkness* written by Troy Kennedy Martin.

To test whether Daisyworld is a valid model of the real world, it is necessary to see if the feedbacks between life and Earth have consistently operated over geological time in such a way as to maintain optimal conditions for life. (This question will be revisited in the final chapter of this book.)

Question 2.10

Would you expect homeostatic systems like those found in living organisms to evolve by means of natural selection at a higher level than the individual, operating, say, for the benefit of whole populations, ecosystems or even life as a whole?

It is important to distinguish between evolved homeostasis in organisms of the kind explained above and simple equilibration brought about by feedbacks. A mixture of ice and water will equilibrate at the freezing temperature of water, for example, because the bonds that hold water molecules together in the ice crystals provide a means of negative feedback. Any heat added to the system is taken up by some of the ice melting (a process that absorbs heat as the bonds break). Loss of heat is compensated for by the heat released through ice formation. So, for as long as both pure water and ice are present, the temperature stays constant. This system has no checks and balances built into it, however, to guarantee constancy in the face of other perturbations. The equilibrium temperature in this case can, for example, be altered simply by adding salt, which modifies the energy balance and lowers the freezing temperature. Without intervention, the system will not rid itself of the salt to retrieve the previous equilibrium temperature. By contrast, core body temperature is not expected to fluctuate (except temporarily, when ill), for example according to changes of diet. Evolution by natural selection has furnished each human being with homeostatic devices that compensate for all the kinds of environmental changes that were faced by our ancestors (who did survive them and reproduce, as our own existence shows). Equilibria in systems that do not reproduce in the manner of organisms, and have not therefore been honed by natural selection, lack such built-in safeguards to their stability.

Certain aspects of ecosystems, and indeed the Earth and life as a whole, may certainly equilibrate because of the balancing effects of feedbacks (e.g. interactions between levels in the trophic pyramid) within them. But there is no reason to expect the essentially selfish adaptations of organisms to underwrite a given environmental equilibrium. Should conditions change, the winners in the ‘struggle for existence’ will simply be those that adapt to the new circumstances. Hence, no support is forthcoming from the natural selection theory, at least, for the claim of the Gaia hypothesis that the Earth and its life together regulate conditions *in the manner of a living organism*.

Nevertheless, the feedbacks between the Earth and its life have certainly yielded conditions quite unlike those that would have prevailed had the Earth been lifeless. The innovations introduced by the eukaryotes discussed above have allowed them in their turn to add their own significant twists to the tale.

Summary of Chapter 2

1 While some conditions at the Earth's surface (e.g. mean temperature) may have remained within modest limits for most of its history, others (e.g. oxygen concentrations) have undergone radical changes. Yet life has apparently been present almost throughout, although the main diversification of eukaryotes only occurred over the last 1000 Ma.

2 Stromatolites dominate all but the last part of the Proterozoic fossil record, reaching a peak around 1.0 Ga and declining thereafter. Exceptionally well preserved early examples contain microfossils of tiny filamentous and spheroidal prokaryotes. By mid- to late Proterozoic times, there was a clear differentiation of communities in different habitats, including both prokaryotes and simple eukaryotes. Planktonic forms supplied organic material to the deep sea floor.

3 Fossils of probable eukaryotes show that they had evolved at least by the mid-Proterozoic (around 2.1 Ga). Organelles such as mitochondria and chloroplasts probably evolved from endosymbiotic prokaryotes that took up residence in the ancestral eukaryotic cells. The nucleus, by contrast, probably evolved within the original cell structure. The acquisition of mitochondria may have been in response to the first appearance of molecular oxygen.

4 Both fossils and molecular data point to an evolutionary explosion of eukaryotes commencing between about 1.2 Ga and 1.0 Ga. This may reflect the evolution of sexual reproduction: no particular environmental trigger has been implicated.

5 The last 70 Ma of the Proterozoic (the Ediacaran Period) was marked by the appearance of assemblages of large, enigmatic fossils collectively referred to as Ediacaran faunas. They have been subject to a variety of alternative interpretations, but most are likely to have been primitive animals. Molecular data confirm the existence of animals from about 1.0 Ga.

6 Eukaryote cell nuclei contain orders of magnitude more DNA in chromosomes than the simple loop attached to the inside of the rigid outer wall of prokaryote cells. The eukaryote cell also possesses an internal framework of protein rods, which can alter cell shape and control its internal structure. These differences have vastly expanded the relative scope of eukaryote evolution to include multicellularity with cell differentiation, allowing them to build up multi-tiered trophic pyramids. Prokaryotes, meanwhile, have been confined to production and decomposition, though under a wide range of conditions.

7 There is an increased probability of mutation associated with the increased amounts of DNA in eukaryotes. Compensation for this is provided by doubling of the chromosomes, together with the mixing of genes from different individuals through sexual reproduction.

8 Sexual reproduction has in turn increased the efficacy of natural selection, potentially allowing all possible permutations of the genes available in a population to be tested in the struggle for existence. This allows even complex multicellular organisms with slow rates of reproduction to evolve rapidly.

9 Adaptations that evolve through natural selection remain of selfish benefit, in terms of fitness, to the individuals (or genetic entities) possessing them. Adaptation may influence higher levels, e.g. populations and ecosystems, but only by way of incidental effect, which may be good or bad. Hence, while natural selection can explain adaptations for homeostasis in individuals, the theory does not predict the emergence of analogous systems at higher levels.

Learning outcomes for Chapter 2

You should now be able to demonstrate a knowledge and understanding of:

2.1 How changing physical conditions on Earth (and in particular increasing oxygen concentrations) can be linked with the development of early life on Earth through biogeochemical feedback mechanisms.

2.2 The main differences between prokaryotes (single-celled organisms) and eukaryotes (multicellular organisms) in terms of their cell size, structure, function and methods of reproduction.

2.3 The important role played by eukaryotic cells in the evolution of increasingly diverse and complex of life on Earth, and how all life can be classified into four main kingdoms.

2.4 The scarcity of information on the earliest life forms and how the explosion in diversification of the Ediacaran fauna has been interpreted and compared to modern organisms.

2.5 The evolutionary process of natural selection and how aspects such as altruism and adaptiveness can favour an individual within a population, as well as alter the characteristics of the population over time.

Chapter 3

The carbon cycle

All life, from bread mould to beetles, and from streptococci to professors, is composed of similar combinations of carbon-containing molecules and water. Life on this planet is based on the chemistry of carbon, which is why carbon-based compounds are referred to as ‘organic’.

3.1 Carbon and life

There are good reasons for carbon’s unique role in the living world. Here are the main ones:

1 *Carbon can form both soluble and other insoluble compounds*. Life is thought to have originated in water, and water is essential to life. Elements fundamental to the development of living organisms must be able to interact freely with one another, and that occurs most readily in the presence of water. Carbon dioxide in the atmosphere readily dissolves in water, as do many organic molecules that make up living organisms.

2 *Carbon can combine with itself and other elements to make more compounds than any other element*, and these compounds may be solid, liquid or gaseous. It is continually cycled through living organisms, and between living, dead and inorganic components of the Earth.

3 *Carbon permits the storing and passing on of information*. The chemical properties of carbon allow the construction of large, complex three-dimensional molecules (RNA and DNA) that are unique in the extent to which they can store, replicate and convey large amounts of information.

4 *Carbon compounds can store and release energy*. Carbon exists in both oxidised and reduced compounds. The energy released when carbon is transformed from a reduced to an oxidised state can be used for biochemical reactions in cells.

The underlying reason for all of these characteristics is the unique chemistry of the carbon atom (Box 3.1).

Chemical reactions can be of two types:

- **endothermic** reactions, which *require* energy to proceed
- **exothermic** reactions, which *release* energy as they proceed.

Reactions involving carbon are no exception. For instance, burning coal is an *exothermic* reaction in which the carbon in coal is combined with oxygen in air, releasing energy in the form of heat and light. For life, the most important *endothermic* reaction of carbon is photosynthesis, whereby plants make organic compounds and oxygen.

■ The outside energy source is light from the Sun, but what are the raw materials used in photosynthesis?

□ Carbon dioxide from the air and hydrogen in water.

The carbon dioxide diffuses into the plant through **stomata**, which are special cells (mainly on the underside of leaves) that can open and close to the air and, along with the roots, regulate the gas and water balance of the plant.

Box 3.1 Carbon compounds

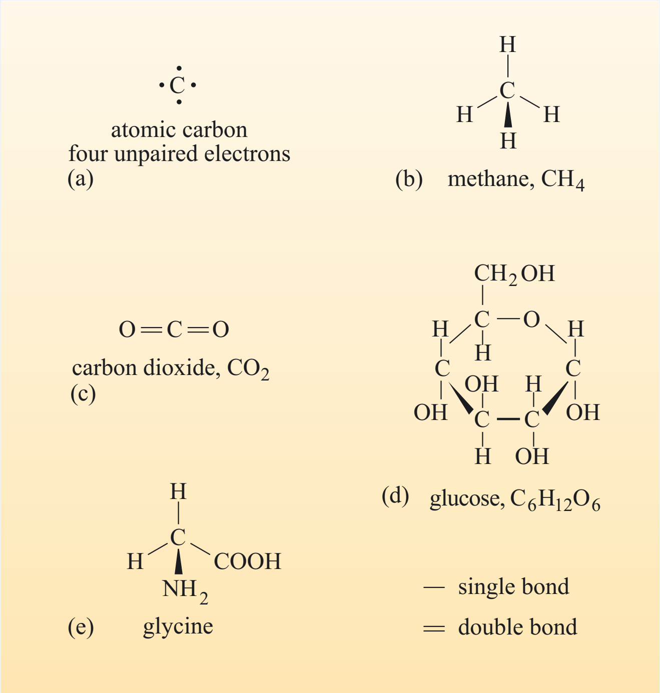


Figure 3.1 The carbon atom and some common carbon compounds: (a) atomic carbon, with its four unpaired electrons that enable each carbon atom to form up to four bonds; (b) methane, which is a tetrahedral molecule; (c) carbon dioxide; (d) glucose, which is a sugar; (e) glycine, which is an amino acid. The structures in b, d and e use a solid wedge symbol for groups and atoms projecting forwards (e.g. $–NH_2$ and –H).

The carbon atom has four unpaired electrons in its outer shell and forms covalent bonds by sharing these four electrons with other atoms (Figure 3.1a). As a result, carbon atoms may form up to four separate bonds with other atoms and so build complex three-dimensional structures. For example, carbon can share four electrons with four hydrogen atoms to form the simplest organic carbon compound, methane (CH_4), or share two pairs of electrons with two oxygen atoms to form carbon dioxide (CO_2) (Figures 3.1b and c). Moreover, carbon readily forms similar shared-electron bonds with oxygen, nitrogen, phosphorus and sulfur, which are all key elements in living organisms. Most important of all, carbon can share electrons with other carbon atoms. This enables it to form large molecules such as sugars (e.g. glucose, Figure 3.1d), chain-like molecules such as **carbohydrates** (which are sugars strung together), and cyclic molecules. Figure 3.1e shows the structure of glycine, which is an amino acid and one of the building blocks of the large and complex protein molecules.

Photosynthesis is a complex process involving many chemical reactions but, as mentioned in Section 1.1, it may be expressed as follows:

$$\underset{\substack{\text{carbon dioxide}\\\text{from atmosphere}}}{6CO_2(g)} + 6H_2O \longrightarrow \underset{\substack{\text{organic}\\\text{matter}}}{C_6H_{12}O_6} + \underset{\text{oxygen}}{6O_2(g)} \qquad (1.2)$$

where organic matter is represented by the carbohydrate molecule $C_6H_{12}O_6$ (see Figure 3.1d). The *total* amount of carbon fixed in this way by plants is called **gross primary production**. Primary production (and by extension, organic matter) is usually expressed in terms of grams of carbon (gC) or kilograms of carbon (kgC).

Some of the CO_2 taken up by plants is returned to the atmosphere through the plant **respiration**, which is essentially the reverse reaction of photosynthesis:

$$C_6H_{12}O_6 + 6O_2 \longrightarrow 6CO_2 + 6H_2O + \text{energy} \qquad (3.1)$$

Respiration releases the energy stored in organic matter and provides energy for the plant to sustain itself by, for example, synthesising tissues and reproduction. The carbon that is *not* released back into the atmosphere via plant respiration is called the **net primary production** (or productivity). In other words, net primary production is the carbon that becomes incorporated into plant tissue. (Note that the term respiration encompasses what is commonly referred to as breathing *and* eating. Eating supplies organic matter to be oxidised with oxygen, releasing energy and CO_2 as a by-product. Breathing brings in the oxygen and removes the carbon dioxide. Both processes are combined in Equation 3.1.)

Plants are autotrophic (i.e. they manufacture their own food). Plant tissue may be consumed by organisms that cannot photosynthesise and depend ultimately upon plants, the primary producers, for energy. These include carnivores that consume animals that consume plants. Such consuming organisms (e.g. fungi, cows and humans) are heterotrophic. Like plant respiration, aerobic (oxygen-using) heterotrophic respiration uses oxygen to convert organic carbon back into CO_2 and water, releasing stored energy in the process (Equation 3.1). So plants cannot live without sunlight and humans cannot live without plants.

Question 3.1

How would you classify the chemical reactions involved in (a) photosynthesis (Equation 1.2) and (b) respiration (Equation 3.1) in terms of oxidation or reduction?

The oxygen liberated in photosynthesis comes from breakdown of the *water* molecules. The hydrogen so produced reduces the CO_2 to form carbohydrate molecules. In addition to being a reducing reaction, the conversion of CO_2 into organic matter during photosynthesis is an endothermic reaction (as seen above), taking in energy in the form of sunlight. Taking in energy is a primary characteristic of living things; on death (and during respiration), the energy is released to their surroundings.

Incidentally, you may wonder why, if it releases energy, Equation 3.1 does not simply occur spontaneously. The answer is that it does, but very slowly. Both plants and animals secrete enzymes (biological catalysts) that allow them to regulate the rate of this reaction.

3.2 Carbon and climate

As mentioned in Section 1.2, CO_2 and CH_4 are greenhouse gases that absorb longwave (infrared) radiation emitted by the Earth's surface and reradiate energy at longer wavelengths, much of it back towards the Earth (Figure 1.12). The warming caused by these and other natural greenhouse gases (e.g. water vapour, ozone and nitrous oxide) means that rather than being about –18 °C, the average surface temperature of our planet is about +15 °C, which is favourable for the maintenance of liquid water and life. Indeed, without the 'blanket' of greenhouse gases, the Earth's surface would be completely covered in ice.

The current interest in atmospheric CO_2 in the context of global warming is heightened because there is good evidence that its concentration has varied over the history of the Earth and that these variations are at least partially linked to changes in the Earth's climate. But what are the factors that control atmospheric CO_2? For reasons that will become clear later, the best way to begin to attack this problem is to study the fluxes of carbon through the natural carbon cycle.

3.3 The natural carbon cycle: a question of timescale

Largely because of its involvement in living processes, carbon continually cycles through the different spheres of the Earth, i.e. the biosphere, the atmosphere, the hydrosphere and even the outer part of the solid Earth – the lithosphere. This global cycling of an element used by organisms (e.g. carbon, oxygen, nitrogen, phosphorus and sulfur) in and out of living, dead and inorganic reservoirs (Figure 1.4) is a biogeochemical cycle (Section 1.5). Such elements periodically move between reservoirs and rearrange themselves into different compounds and may be thought of as the currency exchanged between Earth and life.

There are six main carbon reservoirs, as shown in Figure 3.2.

■ What are these reservoirs in order of decreasing size?

■ They are: the deep ocean; deep-sea sediments; soil, including organic debris; the upper ocean (including its **biota**); the atmosphere; and terrestrial organisms (the land biota).

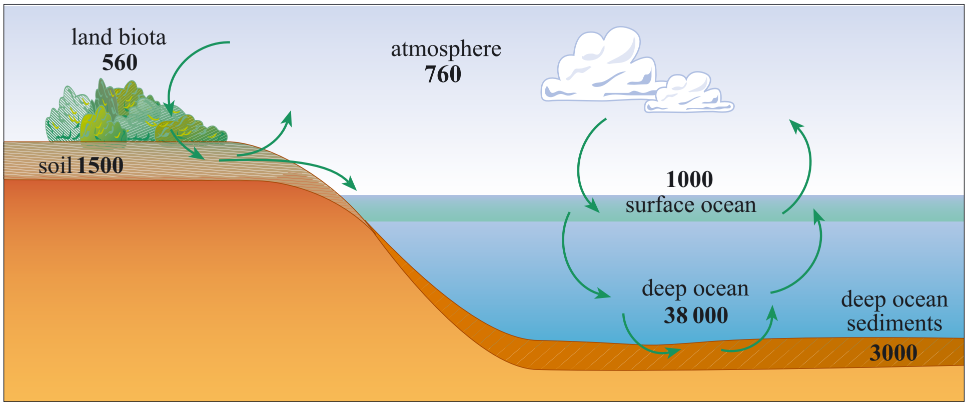


Figure 3.2 A schematic representation of the natural carbon cycle, with six main reservoirs as described in the text. 'Biota' means living organisms – in this context mainly plants. Bold numbers give the sizes of the reservoirs in 10^{12} kgC. In the remainder of this chapter, estimated values will be added to the fluxes indicated by the arrows.

The different amounts of time that carbon atoms spend on average in each of these reservoirs have enormous implications for their potential to affect climate.

The overall natural carbon cycle is actually a number of cycles that link together and occur on a number of different timescales (Figure 3.2):

- a **short biological timescale** of months/years to decades
- an **intermediate geological timescale** of up to hundreds of thousands of years involving chemical, biological and physical components
- a **long geological timescale** of up to hundreds of millions of years, involving rocks and sediments (not shown on Figure 3.2).

In the next section, you will look at the carbon cycle driven by land-based organisms, which is often referred to as the *terrestrial carbon cycle*.

3.3.1 Short timescales: the terrestrial carbon cycle

The terrestrial carbon cycle is dominated by the processes of photosynthesis and respiration and involves three major reservoirs:

- the atmosphere
- plant biomass
- soil organic matter (including detritus).

The reservoir sizes of plant biomass and the atmosphere are similar at approximately 560 × 10^{12} kgC and 760 × 10^{12} kgC respectively (Figure 3.3). Soils store about three times as much carbon as plants (about 1500 × 10^{12} kgC).

■ There is no box shown on Figure 3.3 for terrestrial animal biomass. Why do you think this has been omitted?

■ Animals are an insignificant carbon reservoir when compared with plant biomass (in fact they comprise only about 0.01% of the carbon resevoir).

The land surface of the Earth is about 150 × 10^6 km^2, of which a large proportion is covered by plants (Figure 1.4).

■ If plant biomass were spread evenly over the land surface, how much (in kilograms of carbon, kgC) would there be per square metre?

■ $$\frac{560 \times 10^{12}\ \text{kgC}}{150 \times 10^{6}\ \text{km}^2} = 3.7 \times 10^{6}\ \text{kgC km}^{-2}$$

or 3.7 kgC m^{-2} (to 2 sig. figs).

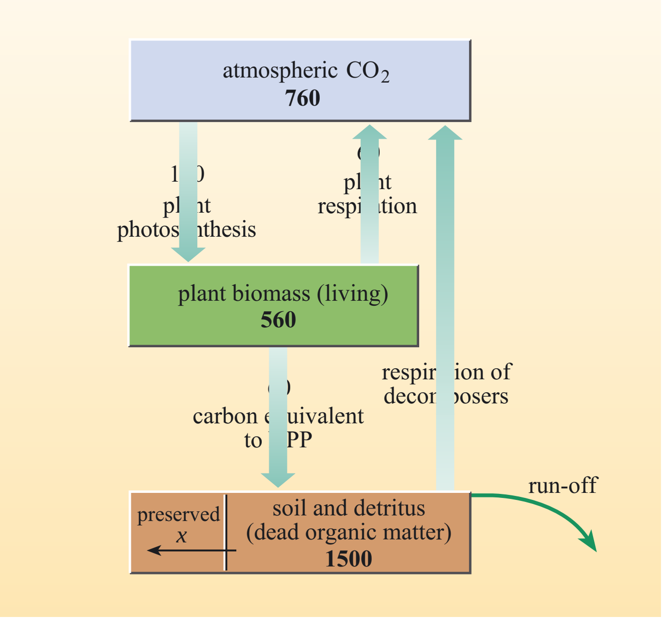


Figure 3.3 The biospheric carbon cycle on land. The bold numbers are the sizes of the reservoirs and the numbers on the arrows are the sizes of the annual fluxes (both in 10^{12} kgC). x denotes the carbon that becomes sequestered in forms that are not readily returned to the system, e.g. plant residues. NPP is net primary production.

Clearly, this average conceals huge variations – think of the amount of plant biomass in 1 m^2 of lawn compared with the average in 1 m^2 of forest.

Plants convert about 120 × 10^{12} kg of carbon from atmospheric CO$_2$ into organic carbon each year. This is the annual terrestrial gross primary production. About half of this is returned to the atmosphere and soil as CO$_2$ through plant respiration, so that *net* primary production (i.e. the annual *net* fixation of carbon from the atmosphere into terrestrial biomass) is about 60 × 10^{12} kg y^{-1}. Most of this carbon goes into soil organic matter when vegetation dies or trees shed leaves or needles, so the amount of carbon deposited in the soil is also about 60 × 10^{12} kg y^{-1}. This is balanced globally by an approximately equal amount of carbon released back into the atmosphere when plant debris decomposes or vegetation burns in natural fires (ignoring the effect of human activities, such as deforestation, on this flux). Remember the amount of carbon that goes into animal biomass is trivial compared with plant biomass; respiration fluxes from animals – particularly the less numerous 'larger' animals – are also small. The return of CO$_2$ to the atmosphere from recycled plant material is therefore almost entirely accounted for by the respiration of decomposers (bacteria, etc.) in the soil.

As with the hydrological cycle (Figure 1.38), if the size of a reservoir and the flux of a substance into and out of it are known, it is a simple matter to calculate the **residence time** of the substance in the reservoir:

$$\text{residence time} = \frac{\text{mass of substance in reservoir}}{\text{flux into (or out of) it}} \qquad (3.2)$$

In doing such a calculation, it is assumed that an equilibrium has been reached, i.e. that the fluxes in and out are equal. Imagine a plant that has been uprooted. Within a short time it wilts: the flux of water molecules into the roots has been cut off, but the flux out of the leaves has not. The fluxes of water are no longer in balance and so the size of the reservoir (the amount of water in the plant) decreases.

Question 3.2

(a) What is the average residence time of carbon in plant biomass? What is it in soil organic matter?

(b) In global terms, 560×10^{12} kg is a relatively small reservoir of carbon (Figure 3.2) and 60×10^{12} kg y^{-1} is a relatively large flux. Would you expect large fluxes through a small reservoir to result in long or short residence times?

The terrestrial biospheric carbon cycle is characterised by relatively large global fluxes through small reservoirs of biota (living or dead), with relatively short residence times. Although the average residence time of organic carbon in soils is about 25 years, it varies greatly depending on the composition of the organic matter and the location. Most carbon in fresh litter is decomposed in a year or two, but highly resistant organic matter in the same soil (such as the carbon in plant structural material, **lignin**) may take much longer to decompose. In fact, carbon in anoxic soils may never fully decompose to release CO_2 back to the atmosphere and may be stored as residues in peat bogs, swamps or similar environments, perhaps eventually becoming coal. The amount of carbon preserved in this way is a very small proportion of the total plant debris and very difficult to quantify; it is indicated by x in the ‘soil and detritus’ reservoir in Figure 3.3.

■ What is the size of the respiration of decomposers flux in Figure 3.3?

■ It must be $(60 - x) \times 10^{12}$ kg y^{-1}.

In other words, Figure 3.3 shows that the carbon in the CO_2 produced by respiration of organisms decomposing plant debris in the soil over the course of a year must equal the carbon added in plant debris *minus* the amount that has been preserved and effectively removed from the cycle *plus* that carried away in run-off. Both of these values are very small, so the carbon added to the soil in plant debris and the carbon removed from the soil in the CO_2 produced by respiration of decomposers are therefore more or less equal (both are about 60×10^{12} kg y^{-1}) and together equal the amount of carbon fixed as (gross) primary production (about 120×10^{12} kg y^{-1}).

Although the amount of organic carbon removed from soils in run-off and exported by rivers to the sea each year is small, removal rates are locally higher if there is soil erosion because of exceptionally high rainfall and floods, or as a consequence of human activities such as farming, mining or forestry. This organic carbon may be either **particulate organic carbon** (**POC**) (fragments of soil or organic debris) or **dissolved organic carbon** (**DOC**). This, together with inorganic carbon removed in the weathering of rocks (**dissolved inorganic carbon** (**DIC**), see Box 3.2), gives a run-off flux of about 0.4×10^{12} kgC y^{-1}.

Difficulties in estimating the *sizes* of the different reservoirs (i.e. the amount of carbon stored globally in each of the various categories identified) impose significant limitations on the accuracy of any calculations of residence time. At its simplest, making a global estimate of a carbon reservoir involves multiplying two numbers: the global area (or volume) of the reservoir and the average amount of carbon in a representative area (or volume) of the reservoir. In practice, arriving at these two numbers is difficult and requires making assumptions and approximations. For example, to estimate the amount of carbon stored globally as plant biomass, one could begin by categorising the vegetation of the Earth into a number of distinct ecosystem types (i.e. **biomes)**, which characteristically store different amounts of carbon. Figure 3.4 shows a global distribution of terrestrial biomes based on climatological and geographical considerations. In practice, satellite images are often used to determine such distributions (Figure 1.4). For each biome, one could:

1 Lay out sampling grids (e.g. 10 m^2, 1 km^2, etc. depending on the type of biome) in numerous different parts of the biome, covering all the known ecological variability.

2 Measure the amount of carbon contained in each of the grid squares.

3 Convert that amount to an average for the entire biome.

4 Multiply that average amount by the global area of the biome.

At best, the final number can only represent an approximate biomass for an extremely varied reservoir. Total global biomass would be obtained by repeating the process for every biome defined and then adding the individual totals together.

In Table 3.1, which shows estimates of terrestrial primary production and net primary productivity, the fourth column shows the results of one such calculation for the total mass of carbon in terrestrial biomass. Here, the land surface is divided into 14 different ecosystem types (column 1), the area of each ecosystem type is calculated (column 2) and then, for each, the mean plant biomass (in terms of carbon) per square metre is estimated (column 3). Columns 2 and 3 are multiplied together to give an estimate of the total mass of carbon in this ecosystem type at any one time. All such estimates may then be summed to provide a total global estimate. The last two columns show the net primary productivity (NPP) for each of the ecosystem types, i.e. the global net fixation of carbon into plant material per year.

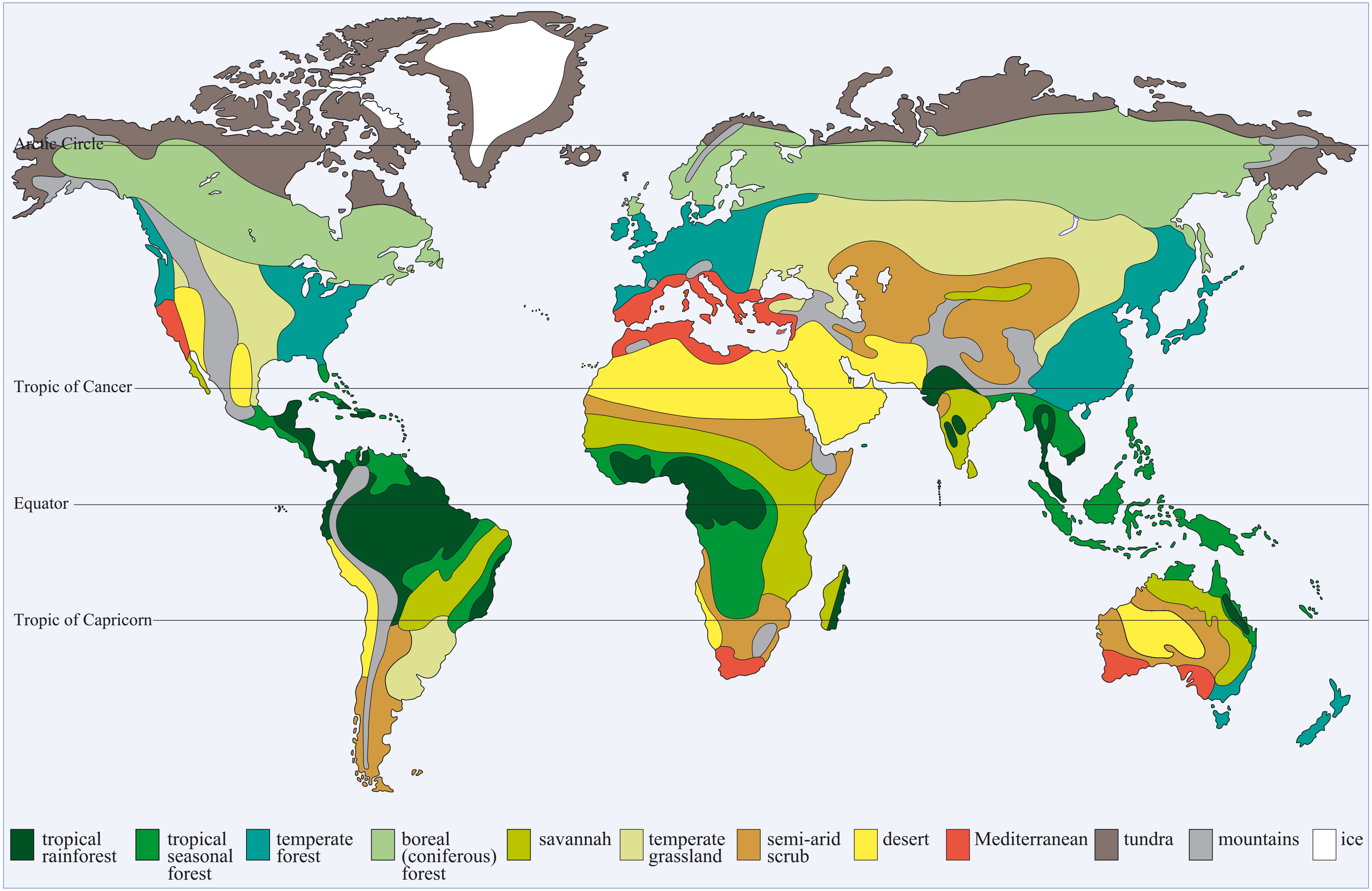


Figure 3.4 Geographical distribution of the major natural regional ecological communities or biomes (compare with Figures 1.1 and 1.2). An anotated version of this figure is in Appendix D.

Table 3.1 Estimates of terrestrial primary production (standing stock) and net primary productivity (i.e. rates of primary production). Mean NPP is to 2 significant figures. NPP = net primary productivity.

Ecosystem type	Area /10^6 km^2	Mean plant biomass per unit area /kgC m^{-2}	Total plant biomass /10^{12} kgC	Mean NPP per unit area /kgC m^{-2} y^{-1}	Total NPP /10^{12} kgC y^{-1}
swamp, marsh	2	6.8	14	1.125	2.2
lake, stream	3	0.01	0	0.225	0.6
cultivated land	14	0.5	7	0.290	4.1
rock, ice, sand	24	0.01	0	0.002	0.04
desert scrub	18	0.3	5	0.032	0.6
tundra, alpine meadow	8	0.3	2	0.065	0.5
temperate grassland	9	0.7	6	0.225	2.0
woodland, shrubland	8	2.7	22	0.270	2.2
savannah	15	1.8	27	0.315	4.7
boreal forest	12	9.0	110	0.360	4.3
temperate deciduous forest	7	14	98	0.540	3.8
temperate broadleaf evergreen forest	5	16	80	0.585	2.9
tropical seasonal forest (deciduous)	8	16	120	0.675	5.1
tropical rainforest (evergreen)	17	20	340	0.900	
total	**150**		**830**		**48.3**
global weighted mean		**5.55**		**0.324**	

According to Table 3.1, the global total for carbon in terrestrial vegetation reservoirs is estimated as 830 × 10^{12} kgC, which is rather different from the 560 × 10^{12} kgC used in Figure 3.3. Variations between estimated totals should not be surprising as there are serious difficulties in making the individual estimates in columns 3 and 4 of Table 3.1. The value of 830 × 10^{12} kgC is in fact at the upper end of the range of estimates for the total global carbon in terrestrial plants; some estimates are as low as ~420 × 10^{12} kgC. All the estimates are, however, of the same order of magnitude and the biggest is only about twice the smallest – quite impressive for a quantity that is so hard to assess. The differences among these estimates are almost entirely due to variation in values for relatively under-studied tropical forests.

Even identifying the biomes is not straightforward, as an apparently homogeneous ecosystem may in fact be made up of several subcategories, and errors may be introduced by ‘smoothing’ the borders between biomes. Other problems include mapping or locating the biomes, measuring the amount of carbon in each grid, choosing representative grid samples, and measuring sufficient samples. In addition, there may be errors in laboratory analysis and inconsistencies arising because different researchers use different techniques.

Other reservoirs are more uniform and easier to estimate. For example, the global reservoir of atmospheric carbon is known to high accuracy because the air in the troposphere is well mixed and therefore fairly homogeneous (in contrast to the very 'patchy' and heterogeneous distribution of vegetation), its volume is known and many direct and precise measurements of atmospheric carbon concentrations (in CO_2 and CH_4) have been taken around the Earth. By contrast, the total amount of carbon in soils and detritus, or in marine sediments, may be significantly larger or smaller than the current best estimate.

Estimating fluxes in and out of a carbon reservoir is even trickier than estimating the reservoir size because it is often difficult to quantify the rates of the processes (such as photosynthesis, respiration and decomposition) that contribute to these fluxes. In practice, many of the processes are measured indirectly. For example, the rate at which various types of organic material decompose may be estimated in the field by the loss in mass over a given time period of a bag containing a known quantity of litter such as leaves; the bag is made of a mesh through which organisms can move freely, decomposing the litter to CO_2. Similarly, gross primary production may be measured by monitoring the CO_2 decrease or O_2 increase in a sealed box containing a plant or leaf (Figure 3.5), and net primary productivity may be estimated by the change in the weight of plant biomass over a given area at the beginning of and at the peak of the growing period.

Before moving on, look more closely at the information in Table 3.1 and see how it relates to the geographical distribution of ecosystems/biomes shown in Figure 3.4.

Figure 3.5 Estimating gross primary productivity by monitoring the decrease in CO_2 concentration in a sealed chamber (supported on a tower) enclosing a branch of Norway spruce (*Picea abies*).

Question 3.3

(a) In terms of net primary productivity per square metre, tropical rainforests (darkest green (TR in figure in Appendix D), shown mainly between the tropics in Figure 3.4) are by far the most productive ecosystems.

(i) What is the annual flux of carbon into and out of tropical rainforests?

(ii) According to the data in the last row and last column of Table 3.1, what percentage of the total global net primary productivity is contributed by these forests?

(b) (i) Area for area, how much more productive are tropical forests than boreal forests (i.e. forests that grow at high northern latitudes, see Figure 3.4)?

(ii) Bearing in mind what you have read earlier, particularly in relation to Figures 1.9 and 2.7, briefly discuss the various possible reasons for this difference in productivity.

(iii) How does the residence time of carbon in living plants in swamps and marshes compare with that in boreal forests?

Of course, in reality it cannot be assumed that areas of the globe covered by particular ecosystems remain constant. For example, tropical rainforests are known to be decreasing in area – adding to the widespread concern about global warming. This apart, the assumption that reservoirs and fluxes are constant does not in practice exclude small-scale fluctuations that average out over relatively short periods of time. The amount of carbon stored in living Northern Hemisphere forests for instance, may greatly fluctuate from summer to winter, but over a number of years the average amount stays relatively constant. Likewise, as will be seen later in this chapter, the concentration of atmospheric CO_2 fluctuates seasonally.

■ If, on average, the input fluxes currently equal the output fluxes, how do you think these various carbon reservoirs built up in the first place?

■ At various times in the Earth's history the fluxes have not been in balance and inputs have been larger than outputs.

For example during the Carboniferous (359–299 Ma) the terrestrial reservoir of carbon grew dramatically as plants spread over the land.

3.3.2 Intermediate timescales: the marine carbon cycle

Although the only one-way flux of carbon to the ocean is in rivers and other run-off (Figure 3.2), by far the greatest carbon fluxes into and out of the ocean are those to and from the atmosphere, across the air–sea interface. For this reason, it is convenient to start a discussion of the marine carbon cycle in the atmosphere. It is therefore important to understand how gaseous CO_2 interacts with liquid water.

Carbon dioxide in water

Carbon dioxide is the most soluble of atmospheric gases (Figure 1.41b). This is because some of the CO_2 molecules that diffuse into water then react with it to produce a variety of dissolved ions (sometimes referred to as ionic species). Box 3.2 sets out the details of the reactions that occur between gaseous carbon dioxide and its various aqueous forms, known collectively (for reasons that should become clear) as the **carbonate system**. You do not need to remember all the details of Box 3.2, but you should understand the following two main points:

1. Very little of the CO_2 dissolved in, for example, rainwater or the ocean is in the form of dissolved gas; most is in the form of bicarbonate and carbonate ions (HCO_3^- and CO_3^{2-} respectively) or as molecules of carbonic acid (H_2CO_3).
2. Given time, the carbonate system will always tend to adjust to an equilibrium situation.

Box 3.2 Carbon dioxide and water: the carbonate system

Very little of the CO_2 dissolved in natural waters is in the form of dissolved gas as such. When molecules of CO_2 gas diffuse into water, some react with water to produce the weak acid, carbonic acid (H_2CO_3), but most occur as hydrated CO_2 (written as $CO_2(aq)$), where each CO_2 molecule is surrounded by water molecules. Since it is difficult to distinguish analytically between $CO_2(aq)$ and $H_2CO_3(aq)$ in practice, dissolved carbon dioxide is normally referred to simply as carbonic acid. Following this convenient shorthand, the chemical equation for the solution of CO_2 gas in water can be written as:

$$CO_2(g) + H_2O \longrightarrow H_2CO_3(aq) \qquad (3.3a)$$

At any particular temperature, the amount of CO_2 that diffuses into the water depends upon the concentration of CO_2 in the atmosphere and the concentration of carbonic acid in the water.

When enough H_2CO_3 builds up in the water, some of the dissolved carbon is released as CO_2 back to the atmosphere in the reverse reaction:

$$H_2CO_3(aq) \longrightarrow CO_2(g) + H_2O \qquad (3.3b)$$

Eventually, the forward reaction (Equation 3.3a) and the reverse reaction (Equation 3.3b) occur at equal rates and a state of dynamic balance or **chemical equilibrium** is established. At equilibrium, *reactants* (on the left-hand side of the equation) are forming *products* (on the right-hand side of the equation) at the same rate as products are decomposing back into their constituent reactants. Although the concentrations of product and reactant do not change, at the molecular level there is a constant and equal exchange between them. To represent this, equilibrium systems are written with two half-headed arrows pointing in opposite directions:

$$CO_2(g) + H_2O \rightleftharpoons H_2CO_3(aq) \qquad (3.3c)$$

The system is, however, more complicated than this. Carbonic acid, like all acids, tends to dissociate, i.e. lose a hydrogen ion. A second equilibrium is established between the carbonic acid and its dissociation products – a hydrogen ion (H^+) and a **bicarbonate ion** (HCO_3^-):

$$H_2CO_3(aq) \rightleftharpoons H^+(aq) + HCO_3^-(aq) \tag{3.4}$$

Furthermore, bicarbonate itself dissociates to form a **carbonate ion** (CO_3^{2-}) and a hydrogen ion. This third equilibrium is written:

$$HCO_3^-(aq) \rightleftharpoons H^+(aq) + CO_3^{2-}(aq) \tag{3.5}$$

So carbon dissolved in water achieves a state of dynamic chemical equilibrium between the CO_2 in the air and that partitioned among three dissolved carbon compounds (shown in bold):

$$\mathbf{CO_2}(g) + H_2O \rightleftharpoons \mathbf{H_2CO_3}(aq) \rightleftharpoons H^+(aq) + \mathbf{HCO_3^-}(aq) \rightleftharpoons H^+(aq) + \mathbf{CO_3^{2-}}(aq) \tag{3.6}$$

This dynamic equilibrium among dissolved carbon compounds is called the carbonate equilibrium system, and the dissolved carbon compounds H_2CO_3, HCO_3^- and CO_3^{2-} are collectively termed dissolved inorganic carbon. At equilibrium, the *concentrations* of the different carbon compounds are constant (but not necessarily equal) and, for each pair of compounds, the amounts of carbon *exchanged* are equal. Therefore, the relative proportions of the reactant and product remain constant. You may visualise this by imagining four boxes, each containing a different number of marbles but with a constant exchange of two marbles back and forth between them (Figure 3.6). In the same way (Figure 3.7), at equilibrium the amount of

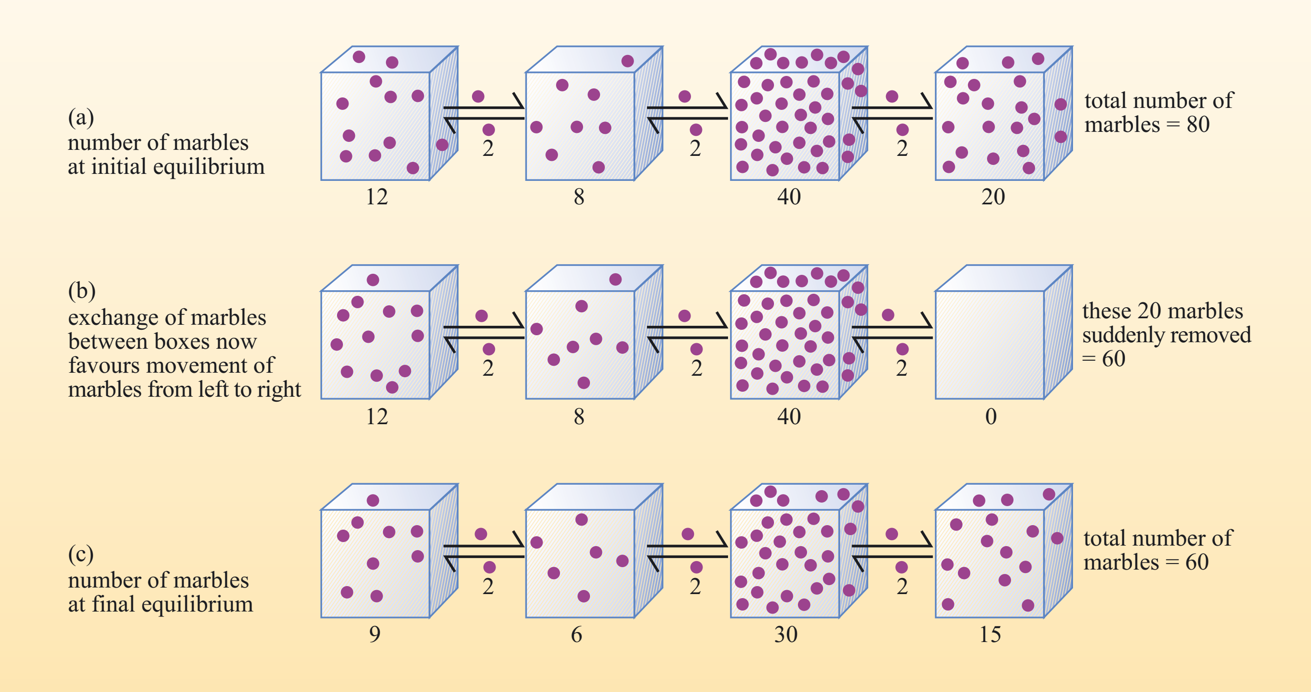


Figure 3.6 (a) Equilibrium among four reservoirs in a closed system; as far as the carbonate system is concerned, the four reservoirs are loosely analogous to atmospheric CO_2, H_2CO_3, HCO_3^- and CO_3^{2-}. (b) If one of the reservoirs is depleted, more reactants will form products in a direction tending to re-establish equilibrium. (c) A new equilibrium is established. The total number of marbles corresponds to the number in the boxes. The marbles on the arrows denote the marbles that will move from one box to another.

CO_2 leaving the surface of the water is the same as the amount going into the water (Equation 3.3c), and the amount of carbonate being formed is the same as that being recombined to form bicarbonate, HCO_3^- (Equation 3.5). Note that the concentration of the components is determined by two equilibrium reactions, Equations 3.4 and 3.5.

When a chemical system in equilibrium is subjected to an external constraint, the system responds in a way that tends to lessen the effect of the constraint; this expression of a fundamental observation is known as **Le Chatelier's principle**, where in a reaction, any factor that removes some or all of the product(s) will cause the reactant(s) to form more product(s) and re-establish the equilibrium.

Returning to our marble analogy, according to Le Chatelier's principle, if the box at the right-hand end is depleted (Figure 3.6b), the exchange of marbles between boxes will begin to favour the movement of marbles towards the right-hand end so that, eventually, equilibrium – a different equilibrium from the first – is established (Figure 3.6c).

■ As far as the carbonate system is concerned, what would be the effect on the rate of diffusion of CO_2 into the water of any process that removes H_2CO_3 from solution?

□ Anything that removes carbonic acid from solution will increase the rate of diffusion of CO_2 *into* the water; hence the rate of dissolution of atmospheric CO_2 will increase.

Conversely, if the concentration of atmospheric CO_2 increased, more carbon dioxide would dissolve in water, forming more H_2CO_3 and ultimately partitioning itself among the various dissolved inorganic carbon compounds (Figure 3.7).

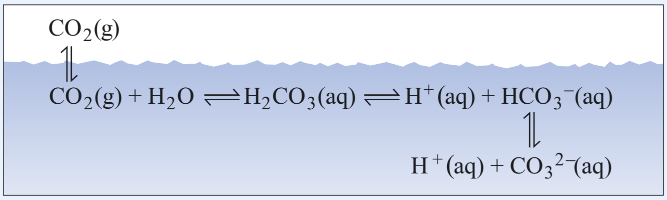


Figure 3.7 Pictorial representation of the carbonate system expressed by Equation 3.6. (In reality, H_2CO_3 is mostly $CO_2(aq)$.)

As mentioned at the end of Box 3.2, if the concentration of atmospheric CO_2 increased, more would dissolve in natural waters. Figure 3.8 is a schematic representation of this situation applied to a body of water such as the ocean. The flux into the ocean would remain greater than the flux out (Figure 3.8b) only until a new equilibrium had been established. Conversely, if the concentration of atmospheric CO_2 decreased, more would come out of solution in the ocean than would dissolve until a new equilibrium had been established (Figure 3.8c).

Question 3.4

The important chemical processes in the global carbon cycle are in approximate balance. What are the equivalent 'reverse reactions' for (a) photosynthesis and (b) weathering of limestones on land?

- Bearing in mind what you read in Section 2.3.2 about the solubility of gases, suggest another circumstance that would lead to an increased flux of CO_2 from the ocean to the atmosphere.
- As all gases are more soluble in cold water than in warm water, warming of surface waters would cause more CO_2 to come out of solution, increasing the flux of CO_2 from ocean to atmosphere.

In other words, the equilibrium situation for warmer water has a lower concentration of dissolved CO_2. This new equilibrium would take a while to establish itself and, until this happened, there would be a net flux from sea to air.

A component of the carbonate equilibrium system that has not yet been discussed is the hydrogen ion (H^+). This has an important role as its concentration in solution determines the relative proportions of the different DIC compounds (H_2CO_3, HCO_3^- and CO_3^{2-}). Hydrogen ion concentration in water (often written using square brackets: $[H^+]$) is generally expressed as **pH**, where a low pH corresponds to a high hydrogen ion concentration and vice versa. For example, a pH of 5 represents a hydrogen ion concentration, $[H^+]$, of 10^{-5} mol l^{-1}, while a pH of 10 represents a $[H^+]$ of 10^{-10} mol l^{-1}. Low pH (below about 5) favours the formation of H_2CO_3, whilst higher pH (7 to 8) favours the formation of HCO_3^- and still higher values (above 9) favour the formation of CO_3^{2-}. These relationships are shown schematically in Figure 3.9, along with the chemical equilibria that are affected (Equation 3.6).

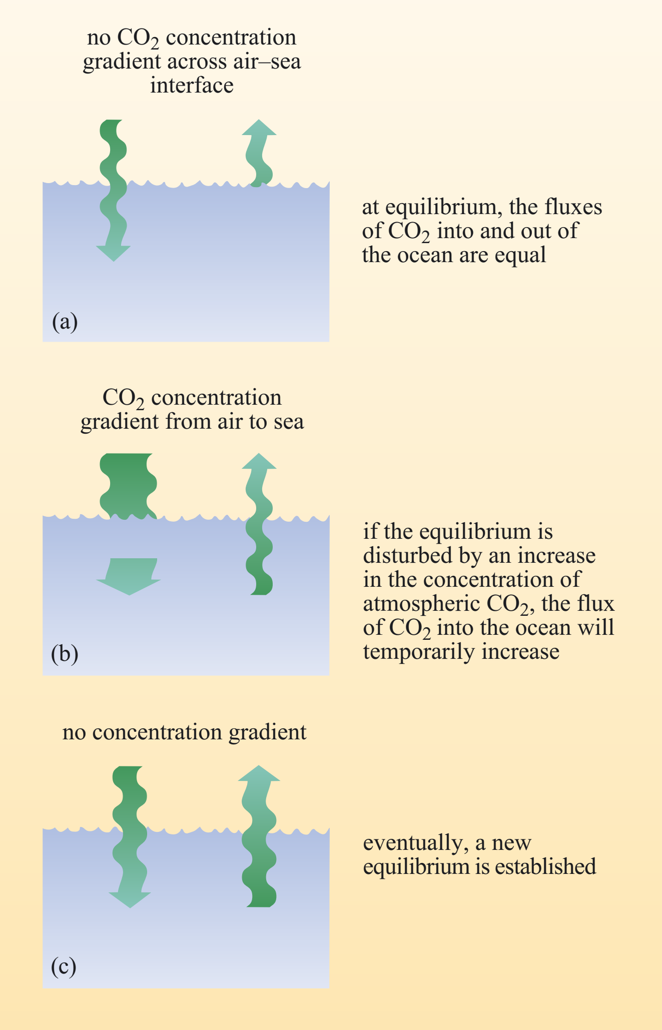


Figure 3.8 Schematic illustration of the change from one equilibrium situation to another. (a) At equilibrium, the fluxes of CO_2 into and out of the ocean are equal. (b) If the equilibrium is disturbed by an increase in the concentration of CO_2 in the atmosphere, the flux of CO_2 into the ocean (i.e. the rate at which CO_2 goes into the ocean) will temporarily increase. (c) Eventually, a new equilibrium is established in which more CO_2 is dissolved in the ocean.

- According to Figure 3.9, what are the main contributions to dissolved inorganic carbon in seawater (pH around 7.7) and rainwater (pH 5.6 or less)?
- In seawater, most (~85%) of the DIC is as HCO_3^- (about 10–15% is as CO_3^{2-}, and less than 1% is as H_2CO_3), whilst in rainwater about half the DIC is in the form of H_2CO_3 and about half as HCO_3^-. (Remember, however, that the positions of the curved lines can vary – see caption.)

In the next section, you will see how weakly acidic rainwater is returned to the oceans.

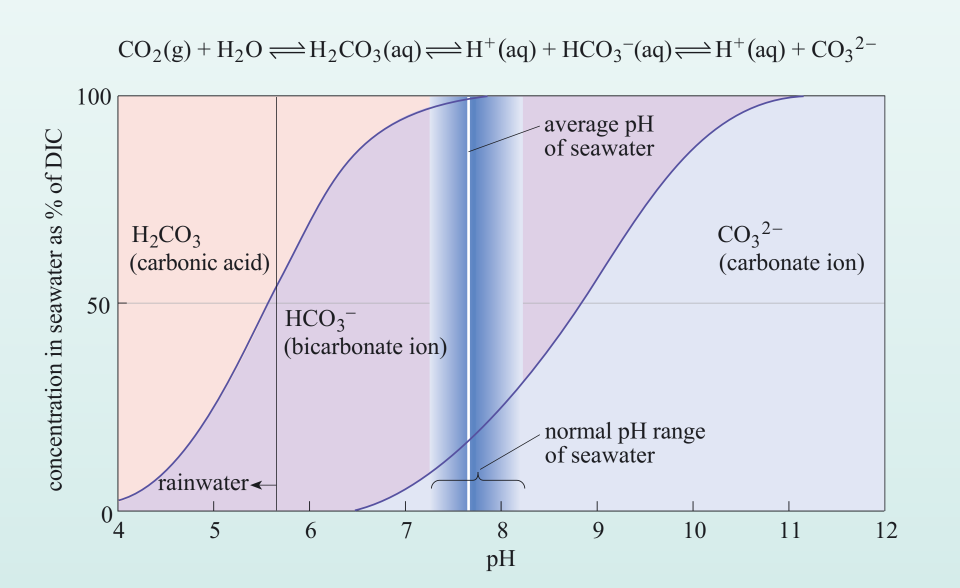


Figure 3.9 Generalised diagram showing approximately how the relative proportions of the three principal components of DIC in the aqueous carbonate system vary with pH in natural waters. The pH of rainwater is 5.6 or less; the pH of seawater averages about 7.7 and can range from about 7.2 to 8.2. The positions of the curved lines can vary with temperature and, in the ocean, also with salinity and pressure.

From rain to soil to river to ocean

Imagine weakly acidic rain falling onto soil where microbes are decomposing organic matter in respiration reactions that use oxygen and release CO_2 (Equation 3.1). Plants are also releasing CO_2 from respiration (most of it through their roots). As a result of both plant respiration and microbial activity pumping CO_2 into pore spaces in the soil, CO_2 concentrations are typically 10 to 100 times higher than in the atmosphere. As a result of this, concentrations of CO_2 in soil water are also raised, moving the reactions in Equation 3.6 to the right, producing additional hydrogen ions. This, plus the addition of various organic acids (e.g. humic acid and fulvic acid) from the decomposition of plant remains, means that soil water is often significantly more acidic than rainwater.

Figure 3.10 Soil forming in a limestone crevice around a newly established thrift (*Armeria maritima*) plant.

Acidic soil water is the main agent of chemical weathering of minerals in soil and rock. In general, carbonic acid plays a larger role in tropical forests where lower concentrations of organic acids remain after surface organic litter has decomposed, while organic acids dominate the weathering processes in cool temperate forests where weathering processes are slow and incomplete. Thus, especially in cool conditions, weathering rates are increased by the activities of living organisms, including small flowering plants (Figure 3.10), fungi and lichens.

Cations such as Na^+ and Ca^{2+} dissolved from rocks are essential for plant growth. Humic material that collects in a developing soil also plays an important part because the essential elements (i.e. nitrogen (as nitrate), phosphorus (as phosphate) and other nutrients) are retained in the soil and are made available to plants largely through their organic content. As every good gardener knows, a high content of humus also helps retain water in the soil; in addition, the relatively dark colour of soil helps it absorb heat and a warm soil generally increases plant growth rates.

As discussed in Section 2.3.2, compared with rainwater, river water contains high concentrations of HCO_3^-, Ca^{2+} and SiO_2 that have been released from rocks by chemical weathering. Most of the Earth's crustal rocks are made up of silicate or carbonate minerals, so the two basic chemical weathering reactions can be represented by just two equations. Equation 3.7a represents the reaction of rainwater with calcite, a form of calcium carbonate that is common in sedimentary rocks (especially limestone and chalk), to release calcium and bicarbonate ions in solution. In Equation 3.7b, rainwater reacts with albite, a common mineral in igneous and metamorphic rocks, to produce a clay mineral, along with sodium and bicarbonate ions in solution, plus silica which is only partly in solution (note that the carbon atoms have been emboldened in red).

For the weathering of carbonates:

$$\underbrace{Ca\mathbf{C}O_3}_{\text{calcite (or any carbonate mineral in rock/soil)}} + \underbrace{\mathbf{C}O_2 + H_2O}_{\text{from rainwater}} \rightarrow \underbrace{Ca^{2+}(aq) + 2H\mathbf{C}O_3^-(aq)}_{\text{in solution in soil/stream water}} \tag{3.7a}$$

For the weathering of silicates:

$$\underbrace{2NaA1Si_3O_8}_{\text{albite (or any silicate mineral in rock/soil)}} + \underbrace{2\mathbf{C}O_2 + 3H_2O}_{\text{from rainwater}} \rightarrow \underbrace{A1_2Si_2O_5(OH)_4}_{\text{kaolinite (a clay mineral)}} + \underbrace{2Na^+(aq) + 2H\mathbf{C}O_3^-(aq)}_{\text{in solution in soil/stream water}} + \underbrace{4SiO_2(aq)}_{\text{silicate partly in solution}} \tag{3.7b}$$

Albite is a sodium-rich silicate mineral, but similar equations could be written for other silicates, such as calcium-rich anorthite ($CaAl_2Si_2O_8$), potassium-rich orthoclase ($KAlSi_3O_8$), or magnesium–iron-rich olivine ($(Mg,Fe)_2SiO_4$).

Each of these two weathering reactions results in two bicarbonate ions ($2H\mathbf{C}O_3^-$) on the right-hand side.

- How do the two reactions differ in terms of where the two carbon atoms in the $2H\mathbf{C}O_3^-$ come from?
- In the weathering of *carbonates* (Equation 3.7a), one carbon atom comes from atmospheric CO_2 and one comes from the carbonate mineral itself. In *silicate* weathering (Equation 3.7b), *both* carbon atoms come from the atmosphere.

It is important to make a note of this distinction, as it will become crucial later in the chapter.

Soil-water, carrying ions released by weathering, collects in streams and eventually reaches the sea via rivers.

- Apart from being a much more concentrated solution, how, generally speaking, does seawater differ from river water?
- The relative proportions of dissolved ions in seawater are different from those in river water (and are, in fact, closer to those in rainwater).

As discussed in Section 1.5, river water has relatively high concentrations of HCO_3^-, Ca^{2+} and SiO_2 (Equations 3.7a and 3.7b), whereas the most abundant ions in seawater are Cl^-, Na^+ and SO_4^{2-}. As will be seen shortly, the apparent shortfall of HCO_3^-, Ca^{2+} and SiO_2 in seawater is intimately related to the cycling of carbon in the ocean.

Carbon in the oceans

Figure 3.11 shows a diagram of the marine carbon cycle. Unlike the terrestrial carbon cycle, which is dominated by the formation, decomposition and recycling of *organic* (mainly plant) material (made of molecules built up of carbon, hydrogen and *oxygen*), the marine carbon cycle is dominated by the chemistry of *inorganic* carbon, i.e. minerals (or dissolved ions) containing carbon. The ocean can be envisaged as a soup of living organisms (particularly in the sunlit surface layers) and the chemical transformations involved in the marine carbon cycle take place largely through their activity. Nevertheless, the flux of carbon to and from the atmosphere – in other words the direction of the net exchange of CO_2 across the air–sea interface (Figures 3.7 and 3.8) – is not primarily regulated by photosynthesis (as it is on land), but by carbonate equilibrium reactions (Equation 3.6) that are affected by both biological and physical factors, which vary from place to place and season to season.

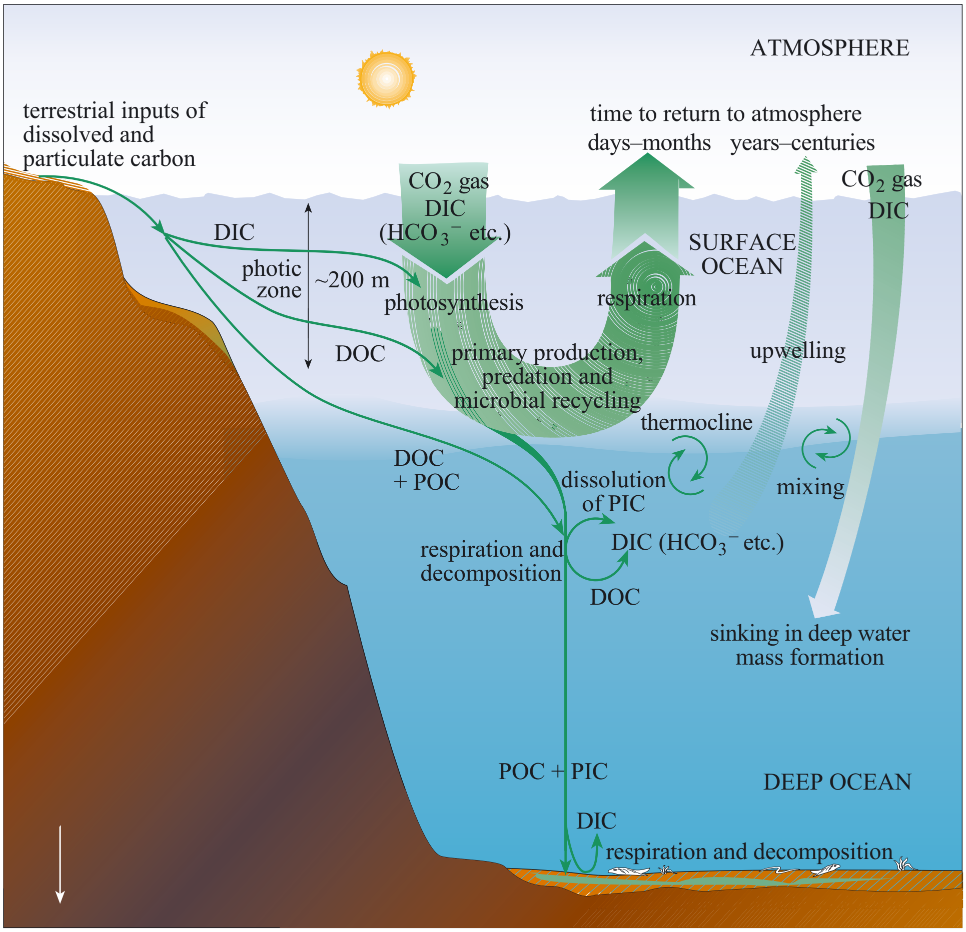


Figure 3.11 Pictorial representation of processes contributing to the marine carbon cycle (not to scale). Remember that while both physical and biological processes affect the fluxes of CO_2 across the air–sea interface, they occur through the action of *chemical equilibria* (Figures 3.7 and 3.8). DIC and PIC = dissolved and particulate inorganic carbon; DOC and POC = dissolved and particulate organic carbon, respectively. Aspects of this diagram not discussed here will be addressed later.

POC = particulate organic carbon
DOC = dissolved organic carbon
PIC = particulate inorganic carbon
DIC = dissolved inorganic carbon

The ocean can be divided into two 'layers' (Figure 1.33): warm, well-mixed surface waters that are underlain by the cold deep ocean. CO_2 exchanges across the air–sea interface (Figure 3.7) at a rate dependent on its concentration in the atmosphere and in the mixed surface layer (Figure 3.11). Windy conditions enhance surface mixing (Section 1.4.2).

Figure 3.12 shows data collected in the North Atlantic in the spring of 1989, when light levels had risen sufficiently for phytoplankton to have begun to multiply (compare with Figure 1.43). The lower (red) plot shows how the concentration of chlorophyll pigment in surface water (an indication of phytoplankton biomass) varied along the ship's track; the highs and lows correspond to fronts, eddies, and so on. The upper (black) plot shows the difference in the partial pressure of carbon dioxide between the atmosphere and the underlying ocean along the same track. Given the complications of estimating the carbon dioxide concentration in water, all you need to know is that the negative values correspond to concentrations in the atmosphere being higher than those in the ocean.

Question 3.5

(a) By reference to the right-hand axis of Figure 3.12, explain whether or not the CO_2 dissolved in the surface ocean was in equilibrium with the CO_2 in the overlying atmosphere at the time in question (Figures 3.7 and 3.8). Was there a net flux of CO_2 across the air–sea interface and, if so, was it into or out of the ocean?

(b) Why do you think the shapes of the two plots are similar?

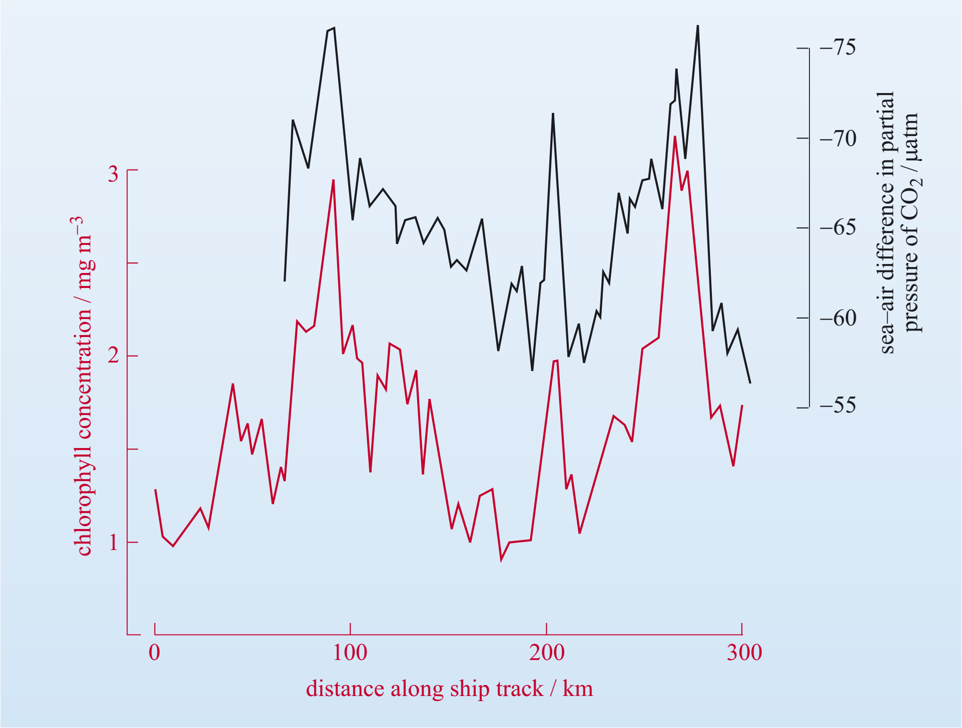


Figure 3.12 Variation in the concentration of chlorophyll in surface water along a ship's track in the North Atlantic (lower red plot) and the difference between the partial pressure of carbon dioxide in surface waters and that in the overlying air (upper black plot), expressed in micro-atmospheres (µatm or 10^{-6} atm), along the same track. Negative values correspond to concentrations in the atmosphere being higher than those in the surface ocean. (Source of data: the North Atlantic Bloom Experiment, which was part of BOFS, the Biogeochemical Ocean Flux Study)

Phytoplankton populations, therefore, act as *sinks* for atmospheric carbon dioxide, as do microscopic benthic (i.e. bottom-living) algae and seaweeds (sometimes referred to as macro-algae). Table 3.2 includes data on carbon fixation and productivity for both phytoplankton and benthic plants.

Table 3.2 Estimates of marine primary production (standing stock) and net primary productivity.

Region	Area /10^6 km^2	Mean plant biomass per unit area /kgC m^{-2}	Total plant biomass /10^{12} kgC	Mean NPP per unit area /kgC m^{-2} y^{-1}	Total NPP /10^{12} kgC y^{-1}
open ocean	332	0.0014	0.46	0.057	19
upwelling areas	0.4	0.01	0.004	0.225	0.1
continental shelf	27	0.005	0.14	0.162	4.3
algal beds and reefs	0.6	0.90	0.54	0.900	0.5
estuaries	1.4		0.63	0.810	1.1
total	**361**		**1.76**		**25**
global weighted mean		**0.0049**		**0.069**	

- ■ Which of the regions listed in Table 3.2 will *not* support any benthic plants?
- ■ Benthic plants can only grow where the bottom is within the photic zone, so the row for ‘open ocean’ (depth 2000–5000 m) cannot include data for benthic plants.

The photic zone may be as much as 250 m deep in the open ocean, but in turbid coastal waters it is a lot less. Algal beds and reefs only grow within the photic zone and support largely benthic production, while the data for estuaries, upwelling areas and continental shelf will include a variable amount of benthic plants depending on the clarity of the water. Clear nutrient-rich coastal waters can support highly productive stands of vegetation (notably kelp forests, Figure 3.13).

Figure 3.13 A kelp forest, characteristic of nutrient-rich, near-shore waters under the influence of coastal upwelling.

It is difficult to define satisfactory categories for compilations such as Table 3.2 (e.g. many upwelling areas are over continental shelves) and, like land vegetation, the estimates are subject to considerable error. Nevertheless, Table 3.2 shows that the *net* amount of carbon converted each year into *new* plant (mainly phytoplankton) tissue (i.e. the net primary productivity) is less than that for terrestrial plants, it is only ~20×10^{12} kgC y^{-1}. Furthermore, the actual amount of carbon in the marine plant reservoir at any one time is very small, only ~$1–2 \times 10^{12}$ kgC (see fourth column), indicating a rapid turnover (i.e. a short residence time) of carbon in marine plants.

Phytoplankton are consumed by zooplankton, and both phytoplankton and zooplankton are consumed by larger animals, and all these organisms respire, releasing CO_2 back into the surrounding water, mainly as HCO_3^-. On death, soft tissues are decomposed by bacteria, releasing organic molecules into solution. Bacterial respiration oxidises both tissue and dissolved organic molecules, returning carbon dioxide into solution; thus, decomposition returns both inorganic and organic carbon to solution (both DIC and DOC are produced), which is available for reuse by organisms. In these various ways, most of the carbon in organisms of the upper ocean is recycled many times into dissolved inorganic and organic carbon and back again.

So how is it that the carbon fixed by photosynthesis does not all find its way back into the atmosphere, in a very short time via respiration?

The reason is that the recycling system is not 100% efficient and some $4–5 \times 10^{12}$ kgC y^{-1} escapes from surface waters, mostly as particles of organic debris, zooplankton faecal pellets (particulate organic carbon, POC) and skeletal remains. Most of the detritus consists of very small particles (i.e. dead and dying algal cells and bacteria) and would take months to sink to the seabed were it not for the fact it forms clumps or aggregates of fluffy debris, often referred to as **marine snow** (Figures 3.14a and b). These aggregates sink to the seabed in days or weeks and arrive below where production occurred in surface waters rather than being dispersed by currents. As it sinks, and after arriving on the deep seabed, the particulate organic carbon is decomposed by bacteria, and both bacteria and organic remains provide food for animals in the water column (i.e. *pelagic* animals) and on the seabed (i.e. benthic, or bottom-dwelling, animals) (Figure 3.14c); all of these organisms respire, returning CO_2 to solution. Generally, soft tissue is almost completely decomposed before it reaches the sea floor (Figure 3.15), but if productivity is very high, as during the spring bloom in the North Atlantic (Figure 1.43a), much phytoplankton debris aggregates into marine snow and sinks out of surface waters without being consumed, providing a food bonanza for bottom-living animals (Figure 3.14d). This transfer of carbon from surface waters to the deep ocean (which means that more has to be supplied from the atmosphere to support plankton growth) is known as the **biological pump**.

Before moving on to look at the *physical* mechanisms that control the flux of carbon into and out of the ocean, another biological process will be considered that uses carbon in seawater. Many, but by no means all, planktonic organisms in the ocean have protective shells or skeletons (as indeed do other larger marine organisms). Many of these hard parts are composed of calcium carbonate

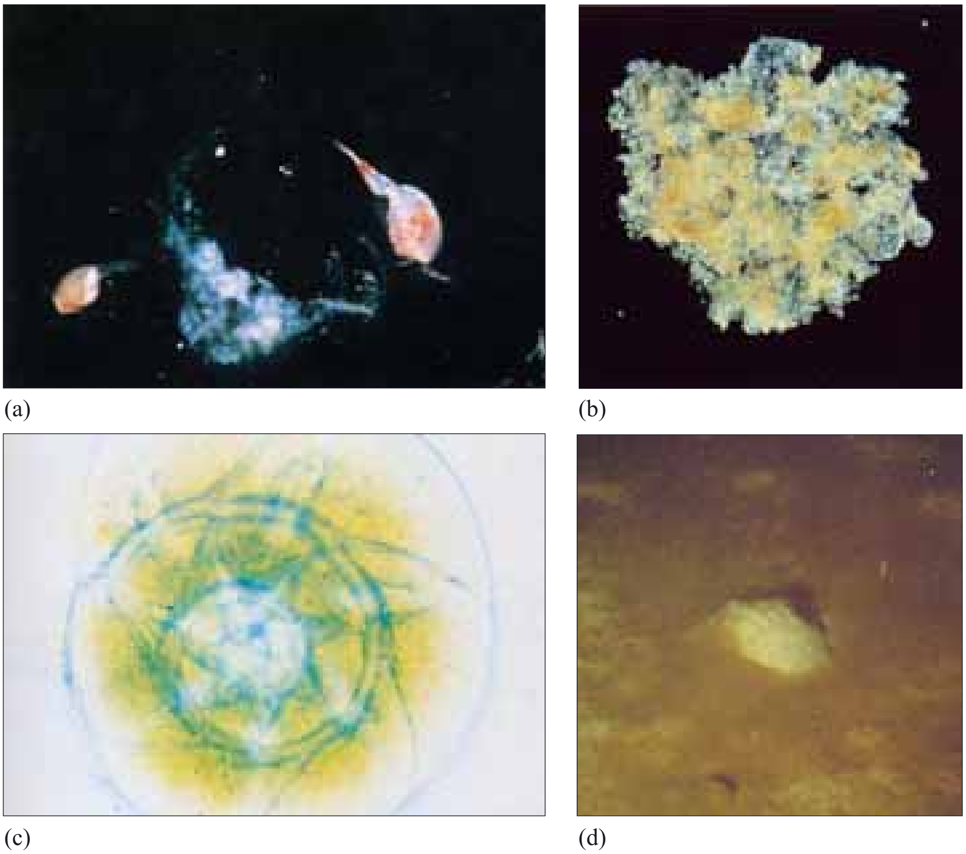


Figure 3.14 (a) Particle of marine snow being consumed by copepods (small planktonic crustaceans). Aggregation of organic material into marine snow speeds up its descent to the seabed, so that there is much less time for it to be decomposed (or eaten) en route. The field of view is ~1–2 cm across. (b) An individual 'snowflake', an isolated aggregate of marine snow ~1 cm across. (c) The ultimate fate of most marine snow is to be consumed by benthic animals. Here you can see phytoplankton remains actually within a benthic foraminiferan (which is about 200 μm across). (d) Marine snow carpeting the seabed and partially burying a mound made by an animal. Some benthic animals time their reproductive cycle so that their young can take advantage of organic debris from the spring bloom in surface waters. The field of view is ~0.5 m across.

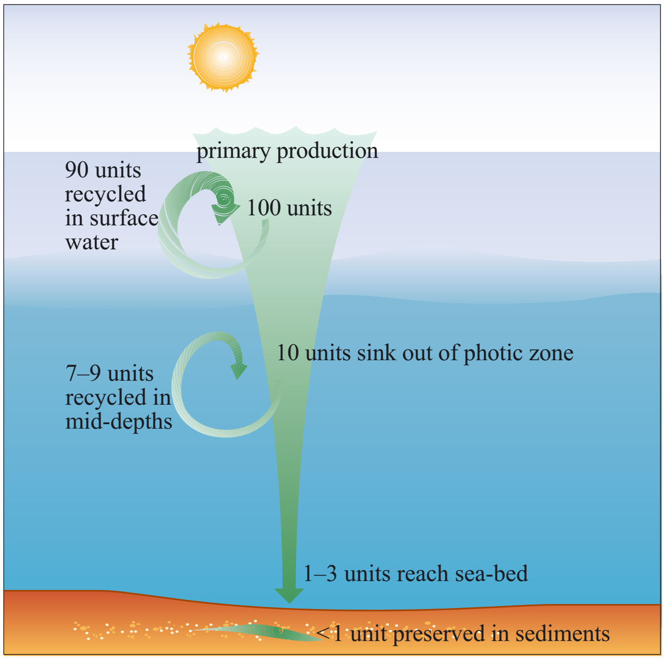


Figure 3.15 Sketch (not to scale) to illustrate the progressive decrease with depth of the carbon initially fixed in the photic zone, and its recycling in the upper and mid-ocean. Generally, only 1–3% reaches the seabed and less than one-third of that might be preserved in sediments.

($CaCO_3$) made from bicarbonate ions (HCO_3^-) and calcium ions (Ca^{2+}) extracted from seawater. The chemical reaction for the precipitation of calcium carbonate shells is:

$$\underbrace{Ca^{2+}(aq) + 2HCO_3^-(aq)}_{\text{in solution in seawater}} \rightarrow \underbrace{CaCO_3(s)}_{\substack{\text{precipitated}\\ \text{by organisms}}} + H_2O + CO_2 \qquad (3.8)$$

Note that Equation 3.8 is Equation 3.7a in reverse. The reaction proceeds exothermically in only one direction; the other direction requires energy (i.e. it is endothermic).

■ Which direction proceeds exothermically and which proceeds endothermically?

■ The weathering reaction (Equation 3.7a) proceeds spontaneously at the surface of the Earth (think of crumbling limestone walls) and so must be exothermic. The biological reaction (Equation 3.8) requires energy – the organism is investing energy for its defence, and so is endothermic.

Two important examples of organisms that secrete $CaCO_3$ are the planktonic algae known as coccolithophores, which secrete plates ('coccoliths') of calcium carbonate (Figures 3.16a and b), and the zooplanktonic foraminiferans (Figure 3.16c). Other planktonic organisms precipitate *silica* from solution to form their shells and skeletons. These include *diatoms*, which are phytoplankton (Figure 1.42a), and radiolarians, which are zooplankton. While organisms that build hard

Figure 3.16 (a) A coccolithophore, which is a single-celled alga ~10 µm across, encased in calcite plates; this species is *Emiliania huxleyi*. Coccolithophores are often the dominant alga in nutrient-poor waters. (b) Debris from within seabed sediments, with coccoliths from several species of coccolithophore (magnification ×2300). (c) Remains of foraminiferans, extracted from seabed sediments. These are ~50 µm across, but some species are much larger. Such shells are often a major carbonate component of sediments below areas of high productivity.

parts of calcium carbonate use Ca^{2+} and HCO_3^-, those organisms that have silica shells remove SiO_2 from solution in seawater. Precipitation of shells and skeletons, along with precipitation of similar minerals involving chemical reactions with sediments and rocks at the seabed, is sometimes referred to as *reverse weathering*. (It is processes such as these that help maintain the difference in composition between seawater and river water.)

Despite their long journey to the deep sea floor, the calcium carbonate remains of planktonic organisms are found in sediments over much of the seabed (Figure 3.16b). Whether calcium carbonate is eventually dissolved or incorporated into sea-floor sediments depends on:

- the chemistry of the seawater (its acidity and the extent to which it is saturated with calcium carbonate)
- the speed of sinking of remains and the depth of the sea floor.

Calcium carbonate debris generally begins to dissolve some distance above the seabed, but by that stage of its descent it has already been incorporated into marine snow and is sinking fairly fast, so there is insufficient time for it to dissolve en route. Most dissolution, therefore, occurs at the seabed. The extent to which remains are dissolved before being protected by further layers of sediment (including more calcareous debris) increases with increasing depth.

The depth at which the proportion of calcium carbonate remains falls to less than 20% of the total sediment (biogenic sediments plus terrigenous clays) is known as the **carbonate compensation depth** (**CCD**). Under areas of high surface productivity, the 'rain' of calcareous debris is such that carbonate in seabed sediments is preserved at greater depths than it would otherwise be, and so the CCD is also at greater depth and is said to be 'depressed'.

Planktonic organisms are not the only ones that extract constituents from seawater to make hard parts, and this process is not only confined to the deep ocean, but also occurs:

- in shallow nearshore waters, where bivalves such as mussels and oysters often grow together, forming large accumulations of shells of calcium carbonate;
- in clear tropical waters, where certain algae form carbonate-rich accumulations, while coral skeletons build up to form substantial reefs.

The physical factors that control fluxes of carbon dioxide across the air–sea interface and its solubility within the ocean are considered next.

■ Referring back to Section 1.5, does the solubility of CO_2 increase or decrease with (i) increasing temperature and (ii) increasing pressure?

□ As discussed in connection with Figure 1.41, the solubility of gases decreases with increasing temperature and increases with increasing pressure.

If you have trouble remembering which way these controls on solubility act, think of the following everyday examples.

1. If you want to check that an electric kettle is working, look at the submerged element: if bubbles of gas are collecting there, air is being driven out of solution as the water warms and the kettle is working.

2 When a can or bottle of gasified drink is opened, it fizzes as gas comes out of solution: the solubility is reduced as pressure is reduced.

Question 3.6

Figure 3.17 shows fluxes of CO_2 across the air–sea interface in January–March and April–June, in the Northern Hemisphere. Areas acting as a net sink for atmospheric CO_2 (net flux of CO_2 from atmosphere to ocean) are shown in green and blue, and areas acting as a source (net flux of CO_2 from ocean to atmosphere) are shown in brown, red and yellow. Study the map, concentrating particularly on the Atlantic.

(a) Explain the direction of the CO_2 flux in the northeastern North Atlantic during January–March. (*Hint*: Look at Figure 1.33.)

(b) Bearing in mind when the data were collected, discuss possible reasons for changes in the CO_2 flux in the North Atlantic between January–March and April–June. (*Hint*: Look at Figures 1.37, 1.43 and 3.11.)

(c) Explain the direction of the flux in tropical and equatorial areas.

The answer to Question 3.6 is illustrated in Figure 3.18, which extends the area of consideration to include the Southern Hemisphere. The main point to appreciate is that net fluxes of carbon dioxide are both into and out of the ocean, and that while regions of deep water mass formation generally act as sinks, other regions may act as sinks at some times and sources at others, depending on the time of year and the physical and biological processes occurring in surface waters.

While surface waters may be in equilibrium with the overlying atmosphere as far as its gas content is concerned (Figure 3.7), the cold waters filling the deep ocean were last in contact with the atmosphere when they were at the surface hundreds of years earlier.

■ In general, therefore, would you expect deep waters to contain higher or lower concentrations of CO_2 (as dissolved inorganic carbon) than surface waters?

■ Deep water generally contains higher concentrations of dissolved inorganic carbon. When deep water masses were last at the surface, they were at high latitudes where, because of the low temperatures, large amounts of CO_2 could dissolve from the atmosphere.

The decomposition of organic remains as they sink into the deep ocean and the dissolution of skeletons and shells also supply dissolved inorganic carbon to deep ocean waters, making them slightly more acidic than surface waters. This increased acidity, along with increased pressures, partly explains why calcium carbonate debris dissolves more readily at depth.

Deep ocean waters are not an endless sink for carbon dioxide because there is exchange of water between the surface and deep ocean. As well as localised regions of downwelling and upwelling (mostly from relatively shallow depths, but also from greater depths at the Antarctic Divergence), there is mixing between

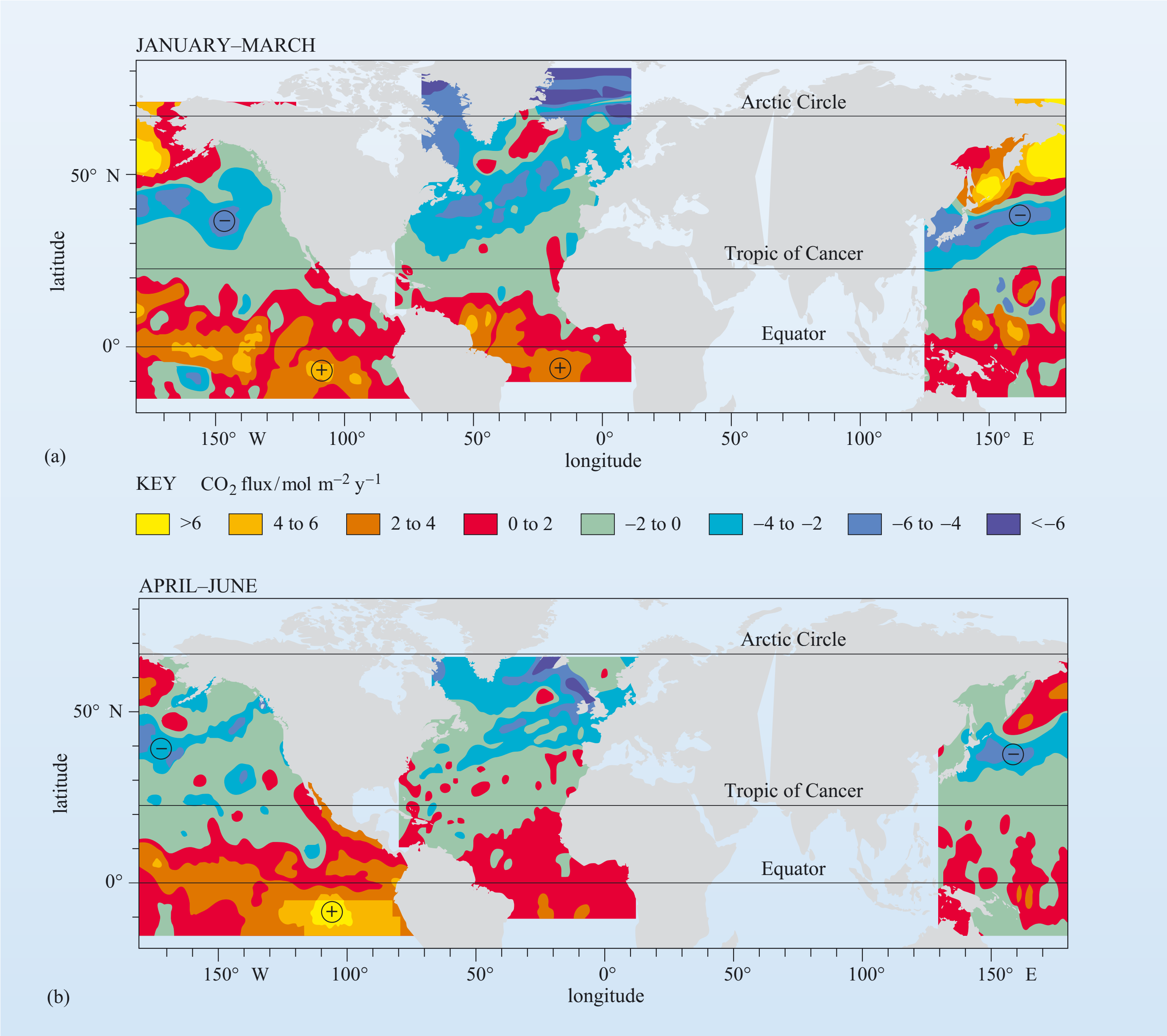


Figure 3.17 Fluxes of CO_2 across the air–sea interface at two seasons of the year: (a) averaged over the months January–March; (b) averaged over the months April–June. Fluxes are given in moles of CO_2 m^{-2} y^{-1}. Positive values (brown, red and yellow marked ⊕) correspond to a net flux of CO_2 out of the ocean. Negative values (greens and blues marked ⊕) correspond to a net flux into the ocean.

the water masses flowing at various depths in the ocean. To compensate for the large volumes of water sinking at high latitudes, the net direction of mixing in the oceans as a whole is upwards. Due to the general movement of deep water towards the northern Pacific and Indian Oceans in the 'global thermohaline conveyor' (Figure 1.35), a relatively large proportion of carbon-rich water eventually comes to the surface in these regions (as can be seen by the yellow and red colours in Figure 3.17). As a result of all the various physical, biological and chemical processes acting on it, a carbon atom will, on average, reside in the oceans for about a thousand years and most of this time will be spent in the deep ocean below the thermocline.

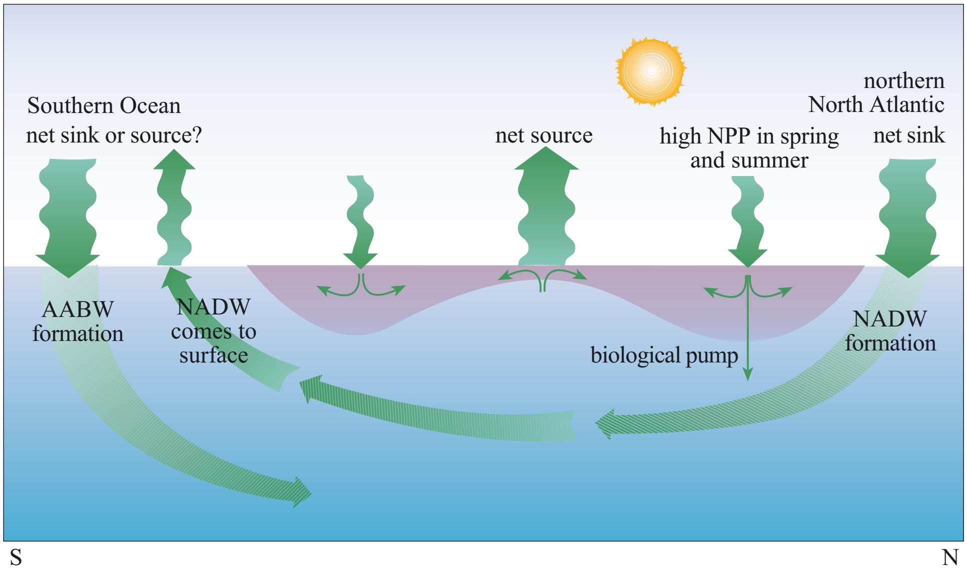


Figure 3.18 Sketch cross-section of the Atlantic Ocean to illustrate the carbon dioxide fluxes across the air–sea interface. Northern and southern high latitudes act as sinks for atmospheric CO_2 because they are regions of deep and bottom water mass formation (for North Atlantic Deep Water (NADW) and Antarctic Bottom Water (AABW) respectively). North Atlantic Deep Water eventually comes to the surface again at the Antarctic Divergence (Figure 1.33), and it is not clear whether the Southern Ocean is a net source or a net sink of atmospheric CO_2. (The diagram is drawn for the northern summer and so high primary productivity, contributing to drawdown of CO_2, is shown for the North Atlantic.)

Figure 3.19 shows how the marine carbon cycle links with the terrestrial carbon cycle (Figures 3.2 and 3.3). The deep ocean is by far the largest reservoir of carbon, storing some 38 000 × 10^{12} kgC at any one time. The other carbon reservoirs in the ocean are the surface waters, which at any one time contain approximately 1000 × 10^{12} kgC (including 2 × 10^{12} kgC in phytoplankton), and the uppermost seabed sediments, which store some 3000 × 10^{12} kgC. These sediments are sometimes called ‘reactive sediments’ because the carbon in them may undergo physical, biological or chemical reactions, so releasing it into the overlying water. Also shown in Figure 3.19 are estimates of annual fluxes between the various reservoirs.

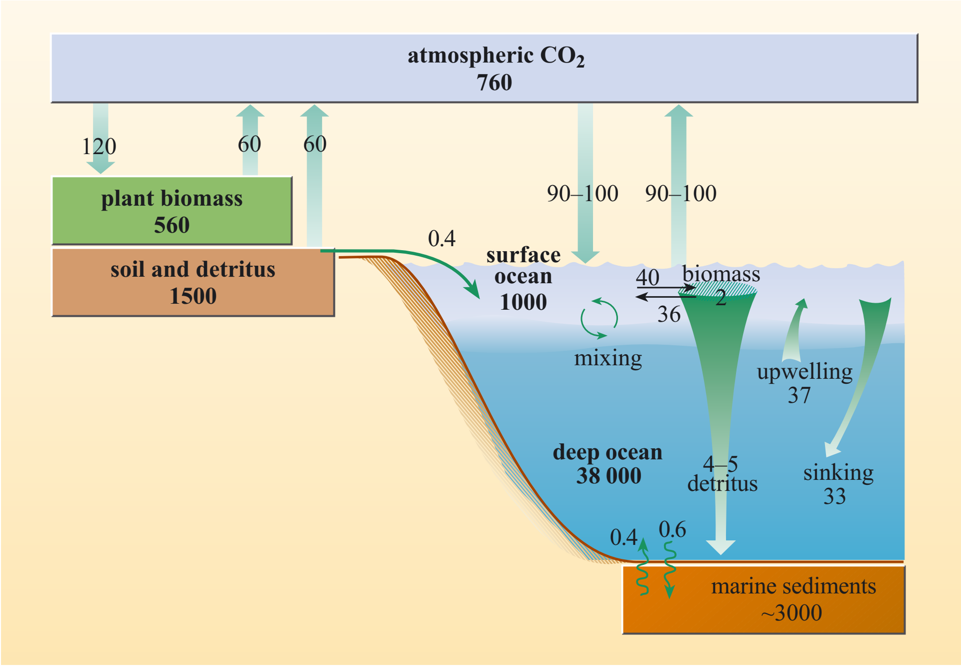


Figure 3.19 The carbon cycle (terrestrial plus marine), including those processes acting over timescales of up to hundreds of thousands of years. Carbon reservoirs are given in 10^{12} kgC and fluxes in 10^{12} kgC y^{-1}.

To calculate the residence time of carbon in ocean waters (reservoir size ~38 000 × 10^{12} kgC, see above), it might seem that the best course of action would be to use the input from rivers as the flux because, although small (only ~0.4 × 10^{12} kgC) the river flux of carbon is one way, while fluxes across the air–sea interface are in both directions.

■ Calculate the residence time using the input from rivers as the flux.

■ Using Equation 3.2:

$$\text{residence time} = \frac{\sim 38\ 000 \times 10^{12}\ \text{kgC}}{0.4 \times 10^{12}\ \text{kgC y}^{-1}} \approx 95\ 000\ \text{y} \approx 10^5\ \text{y}.$$

As with the deep oceans, observations and measurement show that a carbon atom actually resides in all ocean waters for about 10^3 years. The fate of carbon in the ocean (shown schematically in Figure 3.11) will help you understand this conundrum.

About half the carbon entering the ocean in run-off originates as organic carbon (DOC and POC) and about half is inorganic carbon released in the weathering of carbonate and silicate minerals. Once in the ocean both DOC and DIC enter into biological cycling and eventually become part of the bicarbonate–carbonate system in equilibrium with atmospheric CO_2 or are precipitated to form shells, skeletons, etc. River-borne particulate material (such as the remains of leaves, dead freshwater plankton and sewage) accumulates in estuaries or nearshore sediments, where the carbon may be released to the overlying water through biological oxidation. Some of the carbon accumulating in these sediments may become deeply buried and eventually isolated from the biospheric carbon cycle. These deposits, together with organic remains buried in deep-sea sediments and buried organic deposits on land (in swamps, etc.), account for the removal of ~0.05 × 10^{12} kgC^{-1} (not shown on Figure 3.19).

Some approximate values can be added on the fluxes of carbon to the deep ocean, discussed a little earlier in connection with Figures 3.14 and 3.15. The biological pump transfers 4–5 × 10^{12} kgC from surface to deep water every year, and downward mixing and sinking adds another 33 × 10^{12} kgC. Every year, however, upward mixing and upwelling bring approximately 37 × 10^{12} kgC in solution up to the surface ocean. In other words, there is a net transfer of about 4 × 10^{12} kgC y^{-1} from the deep to the surface ocean, balancing the carbon transported to the deep ocean via the biological pump. Therefore, over intermediate timescales (decades to centuries) the cycle is roughly in balance.

It is clear that ocean waters cannot really be simplified into one reservoir as far as the marine carbon cycle is concerned as it is a number of interlinked reservoirs with fluxes between them. Most important of all, the surface ocean not only has fluxes in and out across the air–sea interface, but also has fluxes across its lower boundary, the thermocline; and within the surface ocean is the reservoir of living biomass where carbon resides, on average, for a matter of days. Moreover, although fluxes across the air–sea interface are 90–100 × 10^{12} kgC y^{-1} in both directions, at any one location at any one time fluxes into and out of the surface ocean are by no means equal. Furthermore, carbon entering the ocean as CO_2 dissolving in cold surface waters at high latitudes is carried straight down into the deep ocean – in fact, at high latitudes the surface and deep oceans are effectively joined (Figure 1.33).

As a consequence of marine-snow formation, the time between the fixation of carbon in phytoplankton and the arrival of organic remains at the seabed can be as little as a few weeks, so this aspect of the marine carbon cycle (i.e. the biological pump) can reflect what is happening in surface waters on very short timescales. Of the 4–5 × 10^{12} kgC y^{-1} transported to the deep ocean in biological debris, about 4 × 10^{12} kgC y^{-1} is redissolved in deep ocean water, so that only a small percentage of the carbon originally fixed in surface waters is preserved in sediments (Figures 3.11 and 3.15). Of the 0.6 × 10^{12} kgC y^{-1} deposited or precipitated from deep ocean water within marine sediments, ~0.4 × 10^{12} kgC y^{-1} eventually finds its way back into deep ocean water. This means that about 0.2 × 10^{12} kgC y^{-1} becomes decoupled from the deep ocean water, and as in the case of carbon buried in sediments on land or in shallow water, this small amount of carbon 'leaks' from the cycle. This 0.2 × 10^{12} kgC y^{-1} is the amount preserved at the present time, but because of the short timescales involved, short-term changes in biological productivity at the surface are likely to be reflected in the amount of carbon preserved in sediments (though the chemistry of deep ocean waters would also be affected, and this would in turn affect the preservation of material). Over the long term, however, the rate of accumulation of carbon in sediments will reflect its rate of supply in rivers. For this reason, although small, the river flux of carbon is not a trivial player in the global carbon cycle. In the next section, you will consider the fate of the carbon that is preserved.

Question 3.7

(a) According to Table 3.2, what is the approximate total mass of carbon in marine plant material (i.e. the standing stock), and how does this compare with the average standing stock of plant material on land?

(b) According to Figure 3.19, what is the average residence time of carbon in living phytoplankton? Does carbon cycle through the marine biomass reservoir faster or slower than through the terrestrial biomass reservoir?

3.3.3 Long timescales: the geological carbon cycle

The carbon cycle described so far involves cycling over timescales of years or decades (the terrestrial carbon cycle) to up to hundreds of thousands of years (the marine carbon cycle). Since both the terrestrial and marine carbon cycle 'leak' (i.e. they are not closed systems) over long time periods, carbon can accumulate deep within sediments and is removed from these cycles.

As discussed in the previous section, there are two different types of carbon-rich accumulations.

■ What are the two types?

■ Sediments containing carbonate (carbon in *inorganic* form) and those containing soft tissue (*organic*) remains.

The first of these types of sediment is often described as *calcareous*. Such sediments are generally marine deposits containing accumulations of calcium carbonate shells and skeletons, which have been made by organisms using dissolved inorganic carbon, mainly HCO_3^-. They usually also contain variable amounts of land-derived sediments (mainly clay) and, particularly below upwelling regions, the remains of shells of silica. Those with a significant siliceous component are referred to as *calcareous–siliceous* sediments. (There are also sediments that are predominantly clay or predominantly siliceous, and these contain negligible amounts of carbonate.)

Question 3.8

Is the following statement true or false? ‘Limestone deposits are examples of inorganic carbon, but skeletal remains and shells are examples of organic carbon.’ Give a reason for your answer.

Over millions of years, because of high pressures due to increasing thicknesses of overlying sediment and high temperatures due to heat loss from within the Earth, chemical and structural changes occur in the sedimentary accumulations and they become lithified (i.e. converted into rock). The kind of rock that is formed depends on the initial composition of the sediments, for example carbonate-rich sediments become rocks such as chalk and limestone.

The second kind of carbon-rich accumulation containing organic material (i.e. that made up of molecules of carbon, hydrogen and oxygen) is often described as *carbonaceous*. As organic-rich material is covered with a mixture of more organic matter and/or other sediment, the weight of the overlying deposits causes compaction, which squeezes out the water and residual air from the pore spaces. In the resulting oxygen-poor environment, a dense residue enriched in carbon, known as **kerogen**, is formed. Under continued deposition of organic matter and sediments (often related to the subsidence of continental crust), the original material may be buried to depths of several kilometres. As in the case of calcareous deposits, high temperatures and pressures eventually cause the carbonaceous sediments to become lithified. Large accumulations of land plants (e.g. the remains of trees that have accumulated in anoxic swamps) may become coal; marine sediments containing very high concentrations of phytoplankton debris can produce petroleum. If the organic matter is dispersed through sediments, oil shales may result.

At the present time:

- about 0.05×10^{12} kgC y^{-1} of buried organic matter accumulates on land and sea
- about 0.2×10^{12} kgC y^{-1} of inorganic carbonates accumulates in shallow waters and the deep ocean.

Over the whole Earth, there is approximately $10\,000\,000 \times 10^{12}$ kgC y^{-1} of carbon in carbonaceous rock (including fossil fuels) and four times that in limestones, chalks, etc. (Figure 3.20). Collectively, rocks are the largest reservoir of carbon on Earth. The residence time for carbon within this reservoir is of the order of 100–200 Ma, but even carbon incorporated into rocks can eventually enter the atmosphere again, as geological processes, particularly mountain

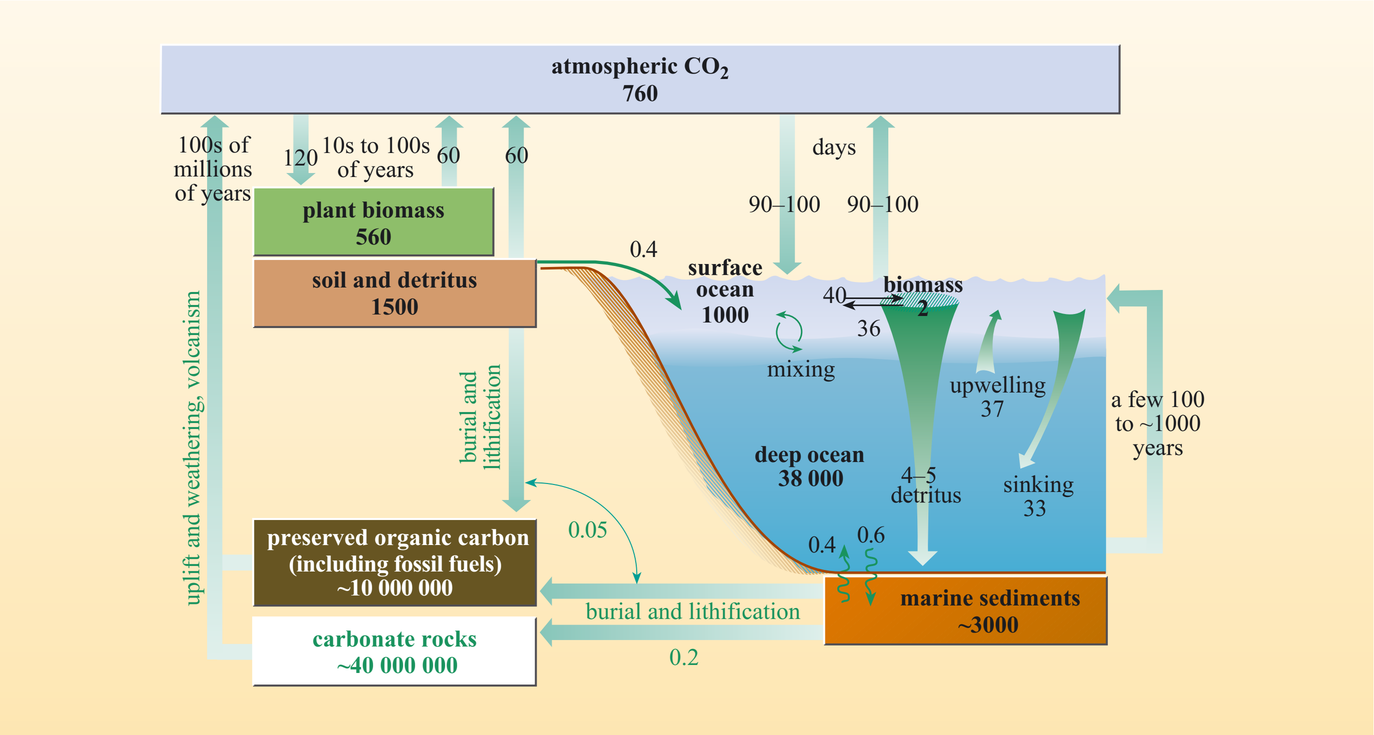


Figure 3.20 Summary diagram of the carbon cycle showing burial and preservation of carbon in sediments (as both calcareous and carbonaceous remains), its eventual return to the atmosphere, and the different timescales involved. Note that the carbonate rocks include lithified remains of calcareous organisms that lived in shallow waters as well as those that lived in the oceans. The approximate mass of carbon in each reservoir is given in units of 10^{12} kgC, and the fluxes are given in units of 10^{12} kg C y^{-1}. (Anthropogenic contributions and current imbalances in the overall system are not included.)

building, bring deeply buried rocks to the surface. The cycle then begins again as the rocks are weathered and eroded by wind, water or biological activity. Section 3.3.2 discussed how weathering of rocks by rain and soil water produces dissolved inorganic carbon, particularly HCO_3^- (Equation 3.7). Carbonaceous sediments exposed to the air may oxidise directly to CO_2, may be oxidised by bacterial respiration, or may be chemically weathered by soil-water and rainfall to release carbon (DOC and POC) into streams. Sometimes, the carbon in rocks finds its way back into the atmosphere in volcanic gases (which will be discussed later). Either way, the carbon from the rock reservoir is eventually returned to the atmosphere and/or the ocean.

Long-term controls on atmospheric CO_2

What affects the concentration of atmospheric CO_2? As discussed in Section 3.3.1, the main flux of CO_2 into the atmosphere through the terrestrial carbon cycle is via respiration by plants and decomposers, which is almost exactly balanced by that removed by photosynthesis (Figure 3.3), for which the chemical reaction is:

$$\underset{\text{carbon dioxide from atmosphere}}{6CO_2(g)} + 6H_2O \longrightarrow \underset{\text{organic matter}}{C_6H_{12}O_6} + \underset{\text{oxygen}}{6O_2(g)} \qquad (1.2)$$

If the current rate of burial of organic carbon into deep sediments roughly equals the rate of carbon release by oxidation and weathering of organic-rich sedimentary rocks (e.g. exposed coal seams), there will be no net gain or loss of atmospheric CO_2. The rate of burial must more or less match the rate of release by oxidation and weathering because a large imbalance would change the concentration of atmospheric oxygen, and such a change has not occurred in the recent geological past. The early Earth had very little, if any, oxygen in its atmosphere, and it was only the cumulative effect of burial of small amounts of organic carbon that allowed oxygen to build up in the atmosphere.

Whereas the flux of carbon into carbonaceous rocks can be balanced by the return flux from weathering, this is not the case for the flux of carbon into carbonate rocks. There is a discrepancy of about 0.03×10^{12} kgC y^{-1} between the carbon that is stored in carbonate rocks (0.2×10^{12} kgC y^{-1}, Figure 3.20) and that eventually returned to the atmosphere and ocean by weathering of carbonate rocks on land (currently estimated at 0.17×10^{12} kgC y^{-1}).

■ Bearing in mind that *silicate* rocks as well as carbonate rocks are weathered (Equations 3.7a and 3.7b, and associated text), suggest what the underlying reason for this discrepancy might be.

■ The answer lies in the fact that, in the weathering of carbonate minerals, only *one* of the carbon atoms that forms the bicarbonate $2HCO_3^-$ comes from the atmosphere, the other comes from the carbonate mineral itself; in the weathering of silicate minerals, *both* carbon atoms come from the atmosphere.

So, weathering a carbonate mineral on land removes one atom of carbon from the atmosphere (in a molecule of CO_2) (Equation 3.7a). Precipitating a carbonate mineral in the ocean (e.g. as part of a shell or skeleton) returns that molecule to the upper ocean, which is in equilibrium with the overlying atmosphere:

$$\underbrace{Ca^{2+}(aq) + 2HCO_3^-(aq)}_{\text{in solution in seawater}} \rightarrow \underbrace{CaCO_3(s)}_{\substack{\text{precipitated}\\ \text{by organisms}}} + H_2O + CO_2 \tag{3.8}$$

Weathering of a carbonate mineral followed by precipitation of a carbonate mineral results in *no net* gain or loss of carbon (in effect CO_2) to the atmosphere.

Now, consider the case of silicate weathering; the reaction for weathering of a silicate mineral (e.g. $NaAlSi_3O_8$, as in Equation 3.7b) requires *two* molecules of atmospheric CO_2. Precipitating calcium carbonate in the sea returns *only one* of those molecules to the upper ocean–atmosphere equilibrium system (Equation 3.8). The burial of this $CaCO_3$ in the sediments removes it from the terrestrial and marine carbon cycles and thus represents a *net depletion* of atmospheric CO_2. These arguments are summarised in Figure 3.21.

Silicate weathering on land followed by carbonate precipitation and burial in the sea removes about 0.03×10^{12} kgC from the atmosphere each year.

The geological record shows that weathering of silicate minerals has occurred on Earth for at least the last 3.8 Ga and that the atmosphere has always contained some CO_2. There must be some return flux of atmospheric CO_2, and this is

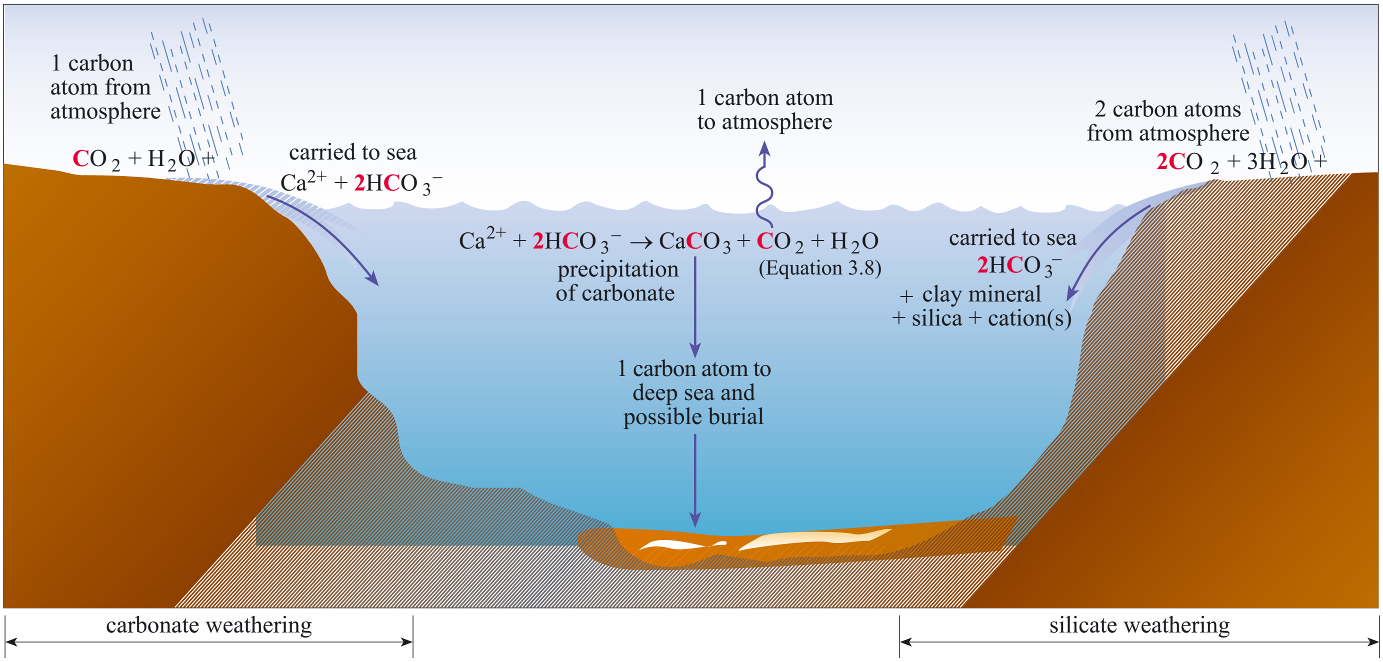


Figure 3.21 Diagram illustrating why burial of sediments containing carbon from the weathering of silicates can result in a net removal of CO_2 from the atmosphere, but burial of carbonates does not. (Note that this diagram does not include processes involving *organic* carbon.)

thought to be via volcanoes, which are part of the plate tectonic cycle that operates over timescales of hundreds of millions of years (as will be discussed in Chapter 4). Indeed, about 0.02×10^{12} kgC y^{-1} is estimated to be released to the atmosphere by volcanoes at the present time.

3.4 A system in balance?

You have seen that the global carbon cycle is highly complex: carbon may be in gaseous, dissolved or solid form, and may reside in various reservoirs on timescales ranging from decades (or less) to hundreds of millions of years. With such complexity, is it possible to determine whether the cycle is in balance?

To explore this question, some kind of reference is needed, such as a reservoir that can be used as an 'indicator' of *any changes* in the cycle.

■ Would changes in the cycle show up first in a large reservoir or a small one? (Imagine a box containing a thousand marbles and a box containing ten marbles. If five marbles are added to or removed from each of these boxes, in which one would you notice it?)

■ The addition or removal of five marbles would be much more noticeable in the box with only ten marbles initially. By extension, changes in the carbon cycle would show up more clearly in a small reservoir.

Atmospheric carbon dioxide is just such a small reservoir. Exchange reactions between the atmosphere and the other carbon reservoirs tend toward equilibrium, with the result that atmospheric CO_2 responds rapidly to changes in the larger reservoirs. In other words, changes in the larger reservoirs are detectable via the atmospheric CO_2 record long before they are noticeable in the sources and/or sinks themselves. Furthermore, the atmosphere is the only reservoir with direct links to the short, intermediate *and* long-term cycles of carbon (Figures 3.19 and 3.20).

The CO_2 in the atmosphere would in theory be depleted in 25 000 years if there was no return flux of CO_2 via volcanism. Over that time period, the amount of carbon in the carbonate rock reservoir would *increase* (in theory) by 760×10^{12} kgC, the mass of carbon originally in the atmosphere.

By this point you may be feeling a little suspicious.

■ Suggest one reason as to why the above calculations would not be accurate, even if volcanism *were* to stop entirely.

■ The calculations effectively assume that all the other carbon equilibrium reactions would be unaffected, but this is unrealistic. For example, decreasing atmospheric CO_2 would cause the ocean carbonate equilibria to shift so that more oceanic CO_2 would be released to the atmosphere (Figures 3.7 and 3.8, the latter of which illustrates the opposite situation), and this would slow the rate of depletion of atmospheric CO_2.

It seems that the concentration of CO_2 in the atmosphere has remained at a few hundreds to several thousands of ppm over the past few hundred million years. This relative stability of atmospheric CO_2 and of the carbon cycle in general is because changes in one reservoir cause repercussions in other reservoirs, and the *negative* feedback loops between them tend to stabilise them.

A property of atmospheric CO_2 that could provide negative feedbacks within the carbon cycle is its role as a greenhouse gas.

Question 3.9

Imagine that the supply of CO_2 to the atmosphere was suddenly increased (e.g. by enhanced volcanic activity); while CO_2 was being added to the atmosphere much faster than it was being removed, its concentration there would be increasing.

(a) Based on evidence in this chapter and by reference to Figure 1.38, suggest at least two ways in which greenhouse warming resulting from such an increase could indirectly affect other processes forming part of the global carbon cycle.

(b) Which, if any, of these processes would result in extra carbon being preserved, and so removed from cycling for hundreds of millions of years?

If CO_2 levels on the Earth were, therefore, say ten times higher than they are today, it could be said with confidence that the rate of CO_2 *removal* from the atmosphere would be increased (Equations 1.2 and 3.8). Precipitation and accumulation of carbonate resulting from silicate weathering, and preservation of organic material (mainly in the sea but also on land) both act as long-term sinks of carbon. Although it is not known how rapidly silicate weathering and global net primary productivity may respond to changes in atmospheric CO_2, it can be reasonably ascertained that a ‘greenhouse’ feedback loop similar to that investigated in Question 3.6 keeps fluctuations of atmospheric CO_2 within a relatively narrow range.

3.4.1 Short-circuiting the geological carbon cycle

If the scale of observation is changed from millions of years to decades, however, it is clear that the carbon cycle is not in balance today; there is more carbon entering the atmosphere than there is being removed from it. This is one of the few certainties in the global carbon cycle (Box 3.3).

Box 3.3 The record on the mountain

In 1957, the scientist Charles David Keeling, then still a student, set up two stations for the continuous monitoring of CO_2 in the atmosphere: one in Hawaii on the volcanic peak of Mauna Loa and the other at the South Pole. The results were dramatic. The CO_2 concentration of the atmosphere at Hawaii averaged around 317 ppm but oscillated around that value on an annual basis, increasing in the winter (e.g. due to the decay of vegetation) and decreasing in the summer (e.g. due to photosynthesis); this provided the first clear evidence that life on Earth profoundly influences the atmosphere.

But this was not the end of the story; within about five years, it became obvious that the annual average CO_2 concentration was steadily increasing (Figure 3.22). No natural phenomenon could be found to account for this spectacularly rapid rate of increase (geologically speaking) and the most likely source appeared to be the release of CO_2 from fossil-fuel combustion. Therefore not only was the growth and decay of plants recorded in the atmosphere, but so too was human industrial activity.

The Mauna Loa and South Pole records are continuing to be added to and, over the years, have been augmented by a worldwide network of measurement stations. All the records of CO_2 concentration show the same rising trend. By 2006, the global average atmospheric CO_2 concentration stood at just over 380 ppm, and it is still rising.

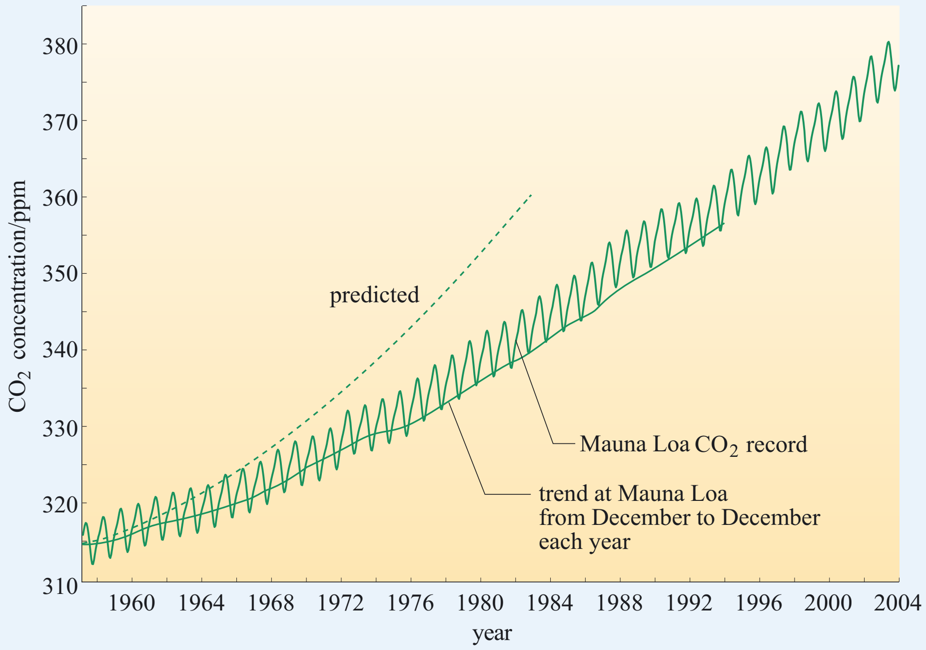


Figure 3.22 The increase in the concentration of CO_2 in the atmosphere, as measured at the Mauna Loa observatory in Hawaii. Also shown is the increase predicted on the basis of fossil-fuel combustion.

Whether they are from Mauna Loa, the South Pole or the newer monitoring stations, all records of atmospheric CO_2 concentration show the same striking features (Figure 3.22):

- a strong trend of increasing atmospheric levels of CO_2 over time, with the current rate of increase at about 1.5 ppm per year or about 0.4% per year
- seasonal oscillations, with a CO_2 peak in the winter and a CO_2 trough in the summer.

All monitoring stations do not, however, produce exactly the same results. Figure 3.23 is a composite diagram showing the variation of atmospheric CO_2 concentrations over the course of four years (1981–1984) over all latitudes.

■ Apart from the peaks and troughs, what is the most striking aspect of this diagram?

🟦 The seasonal oscillations are marked in the Northern Hemisphere, but extremely damped in the Southern Hemisphere with only slight peaks occurring during minima in the Northern Hemisphere.

The peaks and troughs and their global variations both correspond with issues discussed in Question 3.10.

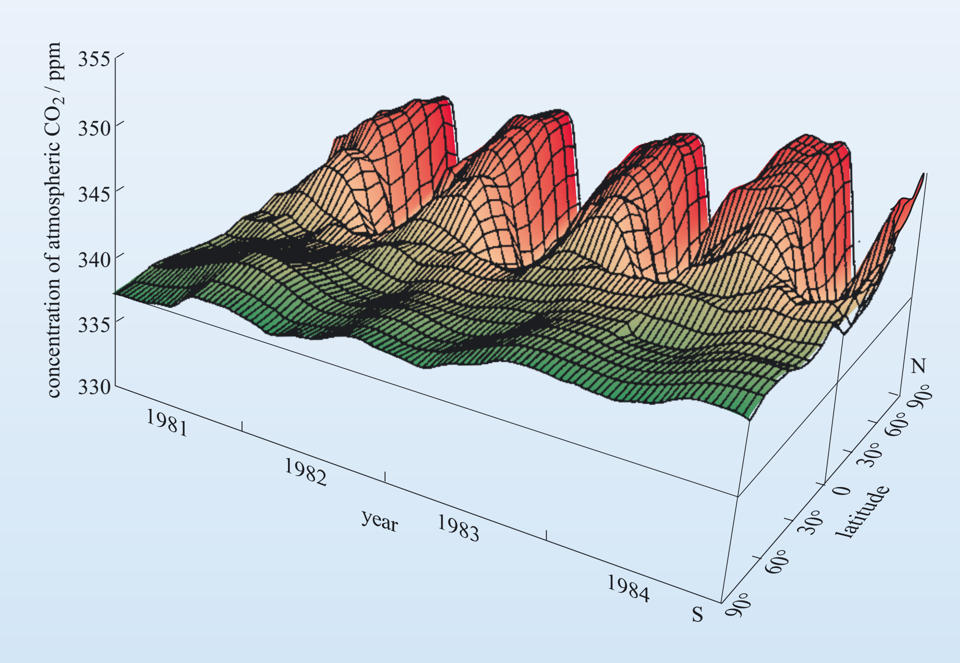


Figure 3.23 Sea-level seasonal fluctuations in atmospheric concentrations of CO_2 from 1981 to 1984, as a function of 10° latitude bands. Note that lows in the Northern Hemisphere correspond to (small) highs in the Southern Hemisphere.

Question 3.10

(a) What is the reason for the seasonal fluctuations in atmospheric CO_2 concentration?

(b) The patterns shown in Figures 3.22 and 3.23 have been dubbed 'the Earth breathing'. State one reason why you think this is, or is not, an appropriate description.

(c) By reference to any global map (e.g. Figure 3.4), suggest why this pattern is dampened in the Southern Hemisphere.

This annual cycle of CO_2 uptake and release emphasises the influence of living organisms on Earth's geochemical cycles. The seasonal variations in CO_2 fluxes into and out of the atmosphere can be quite significant: at Hawaii (latitude 19° N) for example, seasonal variations are sufficient to cause the atmospheric concentration of CO_2 to fall by about 7 ppm over the course of spring and summer, whilst in Barrow, Alaska (latitude 71° N), the winter to summer decrease in CO_2 is double that.

If seasonal decreases and subsequent increases remained equal in magnitude from year to year, the atmospheric reservoir should remain constant. However, the rising trends apparent in Figures 3.22 and 3.23 indicate that there is a persistent disequilibrium in fluxes into/out of the atmospheric CO_2 reservoir. Analyses of air trapped in ice cores has shown that this rapid increase in atmospheric CO_2 is a modern phenomenon; it began approximately 100 years ago, and has greatly accelerated in recent decades. The current rate of increase is about 1.5 ppm per year, with $\sim 3 \times 10^{12}$ kgC added to the atmosphere each year. This is rapid enough to make textbook values for atmospheric CO_2 obsolete; for example in 1958, atmospheric CO_2 stood at 315 ppm, by 1994 it was about 357 ppm. It is now (late 2006) just over 380 ppm. The overwhelmingly likely cause for the increase in atmospheric CO_2 concentrations is the extraction and burning of fossil fuels.

■ Referring to Figure 3.19, briefly explain *why* extraction and burning of fossil fuels would cause an increase in atmospheric CO_2 concentrations.

■ Extracting fossil fuels (organic carbon) and burning them rapidly adds CO_2 to the atmosphere, and so the geological carbon cycle is being 'short-circuited' (Figure 3.24).

The important point here is not that carbon is being returned to the atmosphere (this would eventually happen anyway), but the *rate* at which carbon is being added. Concentrations in the atmosphere of CO_2 (and other greenhouse gases – notably methane, which also contains carbon, Box 3.1) are increasing faster than fluxes into the other reservoirs (i.e. the major short-term buffering processes) can accommodate them. Simply speaking, human activity has caused an increase in the rate of return of carbon to the atmosphere that was previously stored in the Earth's crust.

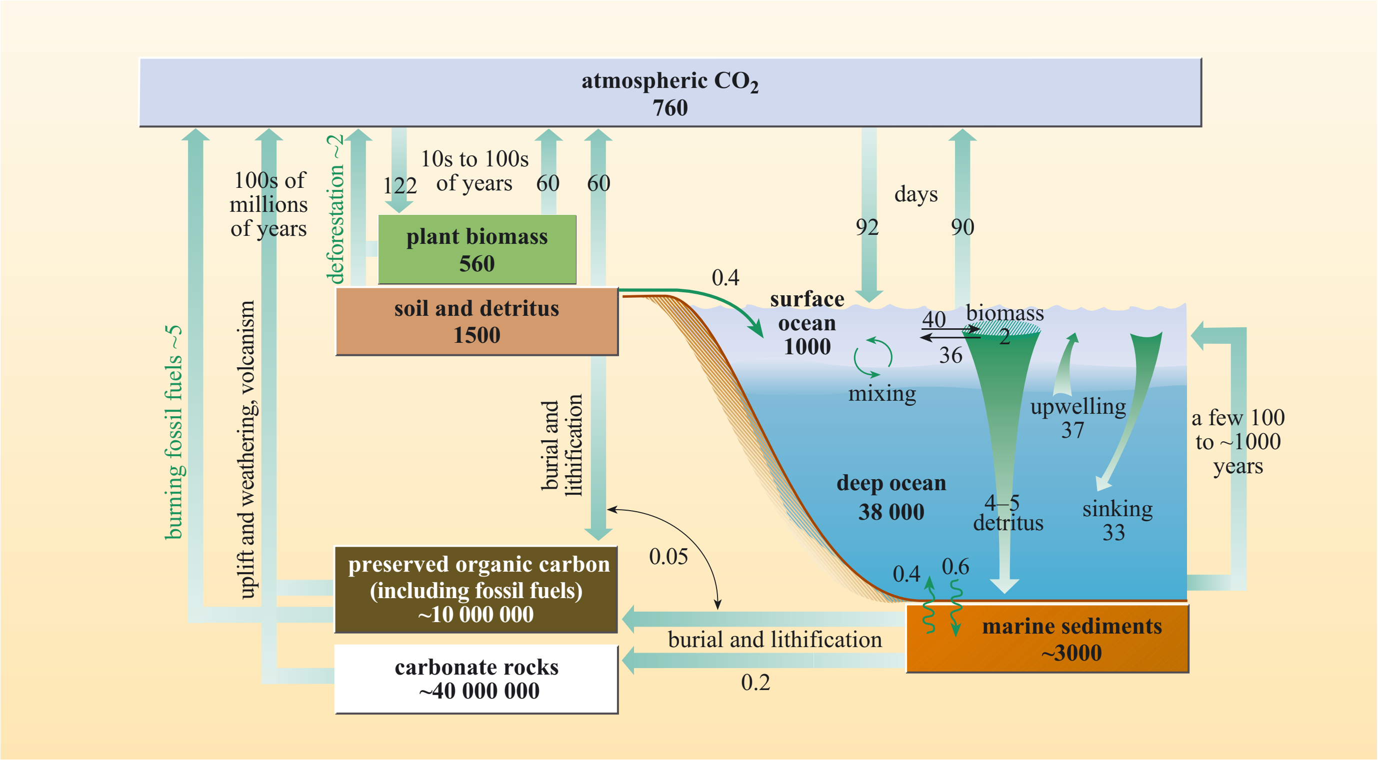


Figure 3.24 Summary diagram of the carbon cycle (Figure 3.20) now also showing the fluxes back to the atmosphere caused by human (anthropogenic) activities. Carbon reservoirs are given in 10^{12} kgC and fluxes in 10^{12} kgC y^{-1}.

As human activity has been directly responsible for this flux through their activities, its size is known to a high degree of accuracy and, according to global production figures for coal and oil, it is ~5×10^{12} kgC y^{-1}.

- How does the current yearly rate of increase of CO_2 in the atmosphere (see above) compare with the flux from fossil fuels?
- The current annual increase is only about 3×10^{12} kgC , which is about a third of that being added from fossil fuels.

So there is actually more carbon being *released* into the atmosphere by human activities than is *accumulating* there – hence the difference between the observed and predicted increases shown in Figure 3.22. This discrepancy is even greater when the amount of carbon added to the atmosphere each year by logging and burning of forest vegetation (currently estimated at ~2.0×10^{12} kgC y^{-1}) is also included.

Figure 3.24 shows the global carbon cycle as perturbed by human activity; it is similar to Figure 3.20 but now includes not only estimates of the fluxes directly resulting from human activities (fossil-fuel burning and deforestation), but also the changes that might feed through into other fluxes.

- Referring to Figure 3.24, suggest where the ‘missing’ carbon might have gone.
- The only possibilities seem to be into the ocean or into plant biomass (terrestrial and/or marine). (Figure 3.24 shows possible increased fluxes into these reservoirs from the atmosphere.)

As illustrated in Figure 3.8 (and discussed above), an increase in the concentration of atmospheric CO_2 will push the equilibria shown in Figure 3.7 to the right and cause an increase of the flux of CO_2 into the ocean (though it would take a long time for the oceans to equilibrate with the atmosphere). Computer models of ocean circulation and calculations of carbonate equilibria suggest that an extra 1.6×10^{12} kgC y^{-1} could eventually be removed from the atmosphere to the oceans in this way. When the total fluxes of carbon are calculated, that still leaves an estimated 2.2×10^{12} kgC y^{-1} unaccounted for.

Conclusive evidence as to the identity of this 'missing sink' has so far eluded scientists, partly because, with the exception of the atmosphere, none of the reservoirs in the carbon cycle is small enough to detect a change of the order of 10^{12} kgC. As mentioned above, however, increased rates of fixation of carbon in plant material on land and in the oceans is a possibility. The answer to Question 3.6a shows that an increase in the concentration of atmospheric carbon dioxide can (at least under experimental conditions) increase rates of carbon fixation (i.e. net primary productivity), thus storing some of the missing carbon in vegetation or soil organic matter. Current indications somewhat surprisingly suggest that the forests of Amazonia may be responsible for taking up a large proportion of the 'extra' carbon, both into continuing growth of existing trees and regrowth in cleared areas. Of course, the missing sink may actually turn out to be a combination of sites, including the huge volume of the ocean.

The rapid increase in atmospheric CO_2 observed today simply reflects the fact that the carbon cycle is currently not in balance. Eventually, coal and oil deposits will be exhausted and the rock-to-atmosphere CO_2 'shunt' will be closed. Some time after this, the various reservoirs of the carbon cycle may re-establish equilibrium with one another and atmospheric CO_2 will stop increasing, re-establishing itself at a higher concentration than previously.

Finally, it is worth considering methane (CH_4), whose atmospheric concentration is some two orders of magnitude less than that of CO_2 (Table 1.2). It is in fact a more powerful greenhouse gas than CO_2, but its very low concentration means that it has a negligible effect on global systems. Methane is emitted from swamps, wetlands, rice paddies and cattle. It is also the principal constituent of the natural gas used for domestic heating, and substantial amounts are lost during extraction of fossil fuels. Since methane is relatively quickly oxidised to CO_2 in the atmosphere, its residence time in the atmosphere is only about 10 years, so at present it is not considered a significant agent of present-day global warming.

However, atmospheric methane concentrations were greater at various times in the geological past, and could be so in the future. Large quantities are presently locked up as gas hydrates in frozen wetlands (tundra) and in some deep-sea sediments. There is good evidence that at various times in the geological past (e.g. the Palaeocene/Eocene boundary, 55 Ma), this frozen methane released catastrophic amounts of gas into the atmosphere at rates sufficient to cause a positive feedback effect, which exacerbated the warming because of the greenhouse gas characteristics of methane.

Artificial carbon compounds (e.g. chlorofluorocarbons (CFCs)), used as spray propellants and refrigerants are also greenhouse gases, but have a much more powerful effect in the stratosphere, where they are the principal cause of the ozone layer depletion that has received so much publicity in recent years. As far as the global carbon cycle is concerned, however, they can be neglected because their atmospheric concentrations are even less than that of methane.

Summary of Chapter 3

1 The chemistry of the carbon atom means that it has a unique role in the living world. Its ability to share electrons with other carbon atoms allows the construction of large, complex molecules of carbon, hydrogen and oxygen of which organic material is built.

2 Carbon fulfils two essential roles in the biosphere: it is the primary component of living tissue (see point 1 above) and, in its gaseous forms (CO_2 and CH_4), it warms the Earth's surface enough to support life.

3 The main biogeochemical connection between Earth and life is the global carbon cycle, i.e. the movement of carbon through the atmosphere, biosphere, lithosphere and ocean. The global carbon cycle involves interlinking cycles over three major timescales:
   (i) the terrestrial carbon cycle, driven by biological processes and acting over timescales of months/years to decades
   (ii) the marine carbon cycle, involving chemical, biological and physical components acting over an intermediate timescale of up to hundreds of thousands of years
   (iii) the geological carbon cycle, involving rocks and sediments, and acting on timescales of up to hundreds of millions of years.

4 The terrestrial carbon cycle is driven by the fixation of atmospheric carbon into organic matter via photosynthesis, and involves large fluxes of carbon between the atmosphere and fairly small organic reservoirs (e.g. living organic matter and organic debris). The average residence time of carbon in plant biomass is about nine years. Estimates of annual net primary production (flux of carbon into the terrestrial biomass reservoir) and of standing stock are problematic, and involve determining the areas occupied by different biomes.

5 The marine carbon cycle is linked to the terrestrial carbon cycle through the carbon carried to the sea in rivers in organic and inorganic form (both dissolved and particulate). The surface ocean is, on average, in equilibrium with the overlying atmosphere and so the fluxes between the two are broadly in balance. CO_2 flux across the air–sea interface may be driven by physical processes (particularly sinking of cold water masses in high latitudes, and upwelling) and by biological activity (i.e. algal photosynthesis which, when combined with rapid sinking of organic matter into the deep sea, constitutes the *biological pump*). In both cases, the mechanism whereby CO_2 is 'pushed out' or 'drawn into' the ocean is chemical. The linked equilibria determine the relative proportions of CO_2 gas and the various forms of *dissolved inorganic carbon*, i.e. H_2CO_3 (carbonic acid), HCO_3^- (bicarbonate ion, the main constituent in seawater) and CO_3^{2-} (carbonate ion); these equilibria are together known as the carbonate system. Biological

activity produces inorganic particulate carbon in the form of shells and skeletons that eventually sink to the seabed; whether they are dissolved or accumulate depends on the chemistry of the deep ocean water and the rate of supply of remains. The depth at which the proportion of calcium carbonate that remains falls to less than 20% of the total sediment is known as the *carbonate compensation depth*.

6 In the geological carbon cycle, carbon in the organic and inorganic products of weathering on land is carried to the sea in rivers, takes part in the marine carbon cycle, and, in the case of a very small proportion, is preserved and buried, eventually to be returned to the atmosphere or ocean via volcanism or via weathering and/or oxidation. The geological carbon cycle acts on a timescale of up to hundreds of millions of years, and involves small fluxes between large reservoirs; carbonate and organic sedimentary rocks together store more than 99.9% of the carbon on the Earth.

7 Extracting and burning fossil fuels shortcuts the geological-scale return flux of carbon and has currently brought the global carbon cycle into disequilibrium. As a result, the concentration of CO_2 in the atmosphere is rising (as are those of other greenhouse gases), but it is not rising as much as expected. Some of the ‘extra’ CO_2 has almost certainly been taken up by the ocean; the rest (or at least some of it) is being taken up by increased net primary productivity of land plants, notably in forests.

8 Long-term stability of the carbon cycle is controlled by negative feedbacks acting on the atmospheric carbon reservoir via rates of silicate weathering on land followed by deposition and preservation of carbonates in marine sediments, and (possibly) the deposition and preservation of organic carbon, mainly in marine sediments.

Learning outcomes for Chapter 3

You should now be able to demonstrate a knowledge and understanding of:

3.1 The important role carbon plays on Earth, in relation to the systems within the biosphere, atmosphere, hydrosphere and geosphere, and in turn how these spheres are interconnected by various biogeochemical feedback mechanisms.

3.2 The differences between the terrestrial, marine and geological carbon cycles in terms of the timescales involved in cycling carbon between different reservoirs, and the resultant impacts these cycles have on the Earth's climate.

3.3 How changes in the global carbon budget can most easily be identified by investigating variations in a relatively small reservoir (e.g. atmospheric carbon dioxide concentrations), and how this has revealed seasonal fluctuations in concentrations due to natural processes, superimposed on annual increases attributed to anthropogenic activities.

3.4 The negative feedbacks that control the long-term stability of the carbon cycle and how the system increasingly appears to be out of balance due to anthropogenic activity.

Chapter 4

Plate tectonics, climate and life

Before the 1980s, most geoscientists saw life as a mere passenger on a dynamic and evolving Earth. Since then, the view that life is partly responsible for maintaining the conditions necessary for its survival has become widely accepted. In its most extreme form this view, which perceives the planet as a holistic, self-regulating system, has been termed the Gaian hypothesis, after the Greek goddess of the Earth. There is no doubt that life on Earth today plays a key role in the feedbacks that maintain climatic conditions within the limits that life can survive. However, you should not forget that life was created from, and was nurtured by, an inorganic world. This implies there must be inorganic feedbacks that keep climate sufficiently stable for this to have been possible. Such feedbacks remain very much in play, operating on longer timescales than most biological feedbacks, and underpin the stability of the Earth's climate over the past 4 Ga.

Since the very early history of the Earth, the underlying geological cycle of the Earth has been driven by plate tectonics. The outermost solid shell of the Earth, termed lithosphere, is segmented into several fragments that cover the face of the Earth; each of these is in constant motion, driven by the internal heat of the planet, which, over millions of years, redistributes the positions of the continents. Plate tectonics profoundly affects climate and life in many ways, three of which you will look at in some detail in the following chapters. Firstly, the energy driving the plates is released largely along plate boundaries as seismic and magmatic activity, releasing gases and particles into the atmosphere that impact dramatically on surface temperatures. Secondly, the migrating continents affect the geometry of the world's oceans and therefore their efficiency in moderating temperature differences through ocean circulation patterns; the distribution of ocean basins also affects sea level and the global extent of coastlines, both of which have implications for the diversity of life. And thirdly, the mountain ranges thrown up during the collision between continents affect the rate at which rock weathering takes place, which in turn affects the carbon dioxide reservoir in the atmosphere (Chapter 5). In this chapter, you will examine the first two of these issues.

4.1 Volcanism and the Earth system

Like the workings of the carbon cycle, most of the processes occurring on and in the Earth are quiet and unspectacular. Volcanism can be very different: Earth's history has been frequently punctuated by massive explosive eruptions the consequences of which are discernible worldwide. Predictably, these occasionally catastrophic episodes have been linked with everything from the death of the dinosaurs to the triggering of Ice Ages. While some of the more extravagant speculations are hard to sustain, volcanism has played, and continues to play, a crucial role in the evolution of the Earth system.

During volcanic activity, heat and material from the Earth's interior rise to the surface. This movement is triggered by the same internal processes that drive the motions of the Earth's lithospheric plates. Figure 4.1 shows the locations of current or recent volcanic activity in relation to plate boundaries.

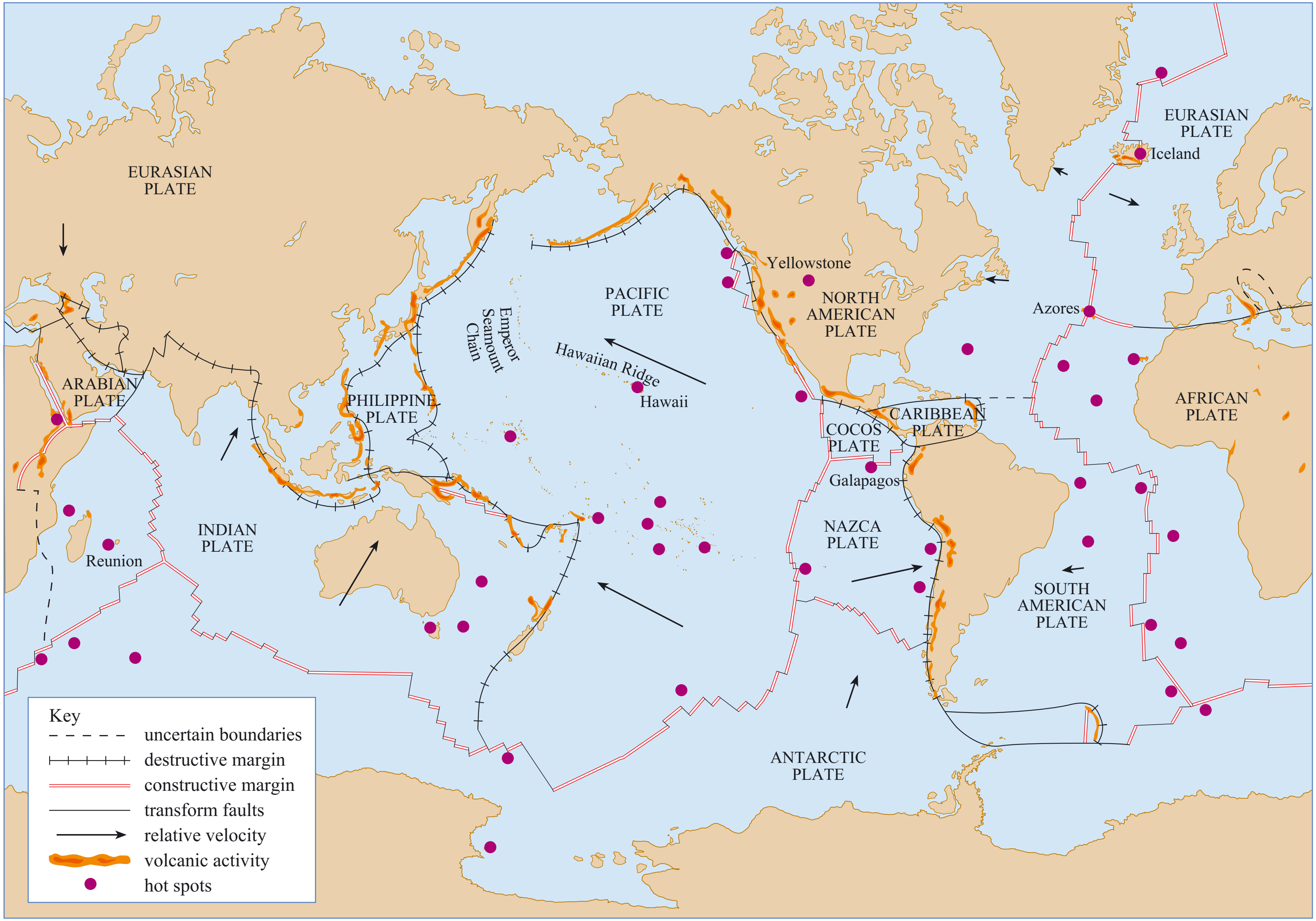


Figure 4.1 The distribution of active (or recently active) volcanic activity in relation to the boundaries of the lithospheric plates, of which there are three types: constructive boundaries (mid-ocean spreading ridges), destructive boundaries (subduction zones), and conservative boundaries (transform faults). The arrows indicate the relative motions of the plates, assuming the African Plate to be stationary; the arrow shown in the key corresponds to a relative velocity of 50 mm y^{-1}.

- ■ Volcanic eruptions related to plate boundaries can be divided into two main groups. What are they?
- ■ Volcanic eruptions at constructive plate boundaries (i.e. mid-ocean spreading ridges) and those in the vicinity of destructive plate boundaries (i.e. subduction zones).

Subduction zone eruptions often take place above sea level, and so they attract more attention than submarine eruptions, particularly as they are intrinsically more explosive. Both types of eruption, however, are important in the Earth system.

At mid-ocean ridges, volcanism is sporadic on short timescales (decades), but over timescales of hundreds of years, ridges are continually active.

A third type of volcanic activity unrelated to plate boundaries occurs at hot spots resulting from mantle plumes. Also shown on Figure 4.1 are the positions of hot spots, which are places where volcanism is triggered by **mantle plumes** (i.e. upcurrents of magma rising from the boundary between the Earth's core and mantle). This source remains fixed as the lithosphere moves over it, so that over time a chain of volcanoes is generated (Figure 4.2). Such hot-spot volcanism is responsible for linear island chains, including those in the northern Pacific, of which the best known is the Hawaiian group (visible in Figure 4.1), with the island of Hawaii itself at the 'young' end of the chain.

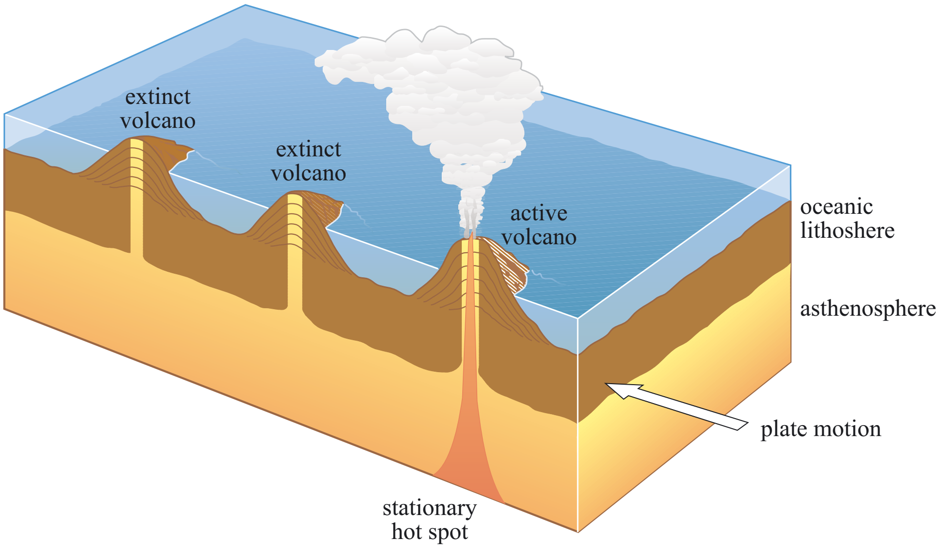


Figure 4.2 Schematic diagram (not to scale) illustrating how a volcanic island chain such as the Hawaiian Chain, forms over a stationary hot spot. The volcano on the right (the equivalent of Hawaii) is still active, and the age of the volcanoes increases towards the left.

Hawaiian eruptions have been well documented, so they offer a useful starting point for this chapter. Hawaii contains two major active volcanoes, Mauna Loa and Kilauea. Kilauea is the world's most persistently active volcano, steadily erupting molten lava into the ocean at about 5 $m^3 s^{-1}$. Such a production rate makes it easy to appreciate the role that volcanism plays in simply forming new land – the island of Hawaii has been built up from the sea floor over the past 1–2 Ma (Figure 4.3). But rocks are only part of the story; Table 4.1 lists the composition of a typical Hawaiian lava at the time of its eruption.

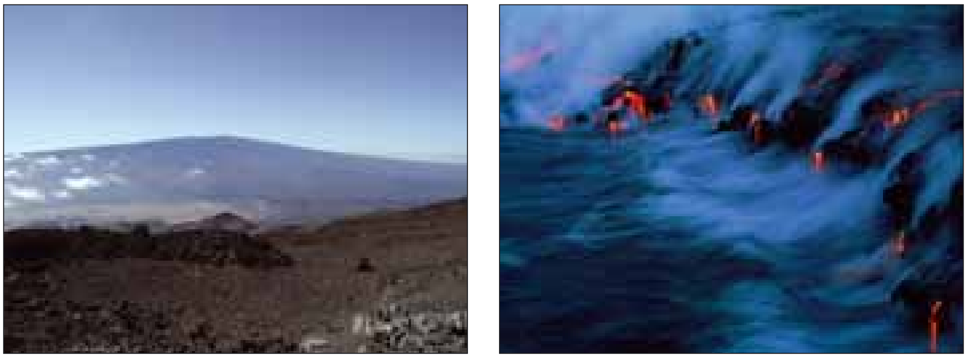

(a) (b)

Figure 4.3 (a) Mauna Kea volcano (altitude 4205 m), seen from the flanks of Mauna Loa, Hawaii. (b) Erupting magma flows into the Pacific Ocean on the southeast coast of Hawaii.

Table 4.1 Chemical composition of the solid and gaseous components of a typical Hawaiian basalt lava erupted from Mauna Loa in 1859.

Elements present as silicate or oxide minerals in magma. The data have been presented to show the relative percentages in mass of different elements.	
Element	**Mass %**
silicon, Si	24.15
aluminium, Ai	7.31
iron, Fe	8.23
magnesium, Mg	4.35
calcium, Ca	7.55
sodium, Na	1.71
potassium, K	0.35
manganese, Mn	0.13
titanium, Ti	1.24
phosphorus, P	0.10
oxygen, O	44.26
Dissolved gases	
water vapour, H_2O	0.27
carbon dioxide, CO_2	0.32
sulfur dioxide, SO_2	0.18

■ Look at the analysis of the magma in Table 4.1. Which of the constituents or elements are most likely to affect the environment some distance from the volcano?

■ Water, carbon dioxide and sulfur dioxide, because they can escape as volatiles (gases) and mingle freely with the atmosphere. (*Note*: the first 10 components (Si to P), combined with oxygen in silicates and oxides, all contribute to the rocky lava.)

At Hawaii, the magma originates deep within the mantle so that the carbon dioxide emitted represents a net addition to the atmosphere. Early in the history of the Earth, virtually the whole of the first stable atmosphere is believed to have formed by volcanic outgassing in this manner. First though, the volcanic input of carbon dioxide (which is a potent greenhouse gas) into the atmosphere will be considered.

The total amount of CO_2 liberated to the atmosphere during Mauna Loa's formation so far is about 2×10^{15} kg. If Mauna Loa has been building up over 1.5 Ma, the flux into the atmosphere will have been 1.4×10^{9} kg per year (about the same as a 200 MW coal-fired power station) and equivalent to about 0.01% of the annual anthropogenic contribution. Although some volcanoes (e.g. Mount Etna) exhale more CO_2 than Kilauea, the total global volcanic contribution to the atmosphere is thought to be about 6.5×10^{10} kg y^{-1}, which is still less than 0.3% of the total anthropogenic (human-generated) flux. This figure is a striking illustration of how human activity is altering the natural environment. It is important to note, however, that although volcanoes presently account for the addition of a relatively small amount of CO_2 to the atmosphere, at certain times in the past the impact of volcanic CO_2 was much greater (and this will be examined in more detail later in this chapter).

■ Of the three types of volcanism shown in Figure 4.1 (volcanism at mid-ocean ridges, subduction zones or mantle plumes), which will emit the most CO_2 along with erupted magma?

■ It is likely to be volcanism associated with mid-ocean ridges, which is much more widespread, and overall produces much more lava than either volcanism associated with subduction zones or mantle plumes.

Current estimates of magma production by these three modes of volcanism have considerable uncertainties attached to them, but mid-ocean ridges are generally thought to account for around 85% of global magma production. The other two modes, being more episodic in nature, are difficult to estimate, but probably generate roughly equal volumes of magma on average. While both subduction zone and mantle plume volcanism are predominately subaerial, releasing CO_2 directly into the atmosphere, mid-ocean ridges are located on average at depths of about 2.5 km underwater, where pressures of several hundred atmospheres prevent CO_2 escaping from the lavas in gaseous form. Even so, CO_2 is still an important component of the gases dissolved in the fluids that are expelled at hydrothermal vents found wherever there is submarine volcanism. The huge volumes of magma erupted at mid-ocean ridges carry with them correspondingly vast quantities of CO_2, which will eventually be released into the atmosphere after circulating in the ocean in solution for on average about 1000 years. Therefore, on geological timescales, mid-ocean ridges contribute around 75% of the CO_2 from volcanic sources to the atmosphere.

So far, the text has concentrated on volcanic emissions of CO_2 *on short timescales*. Sulfur dioxide (SO_2, which is a poisonous gas) has far more dramatic effects on the atmosphere, and this will be the subject and discussion of the next section.

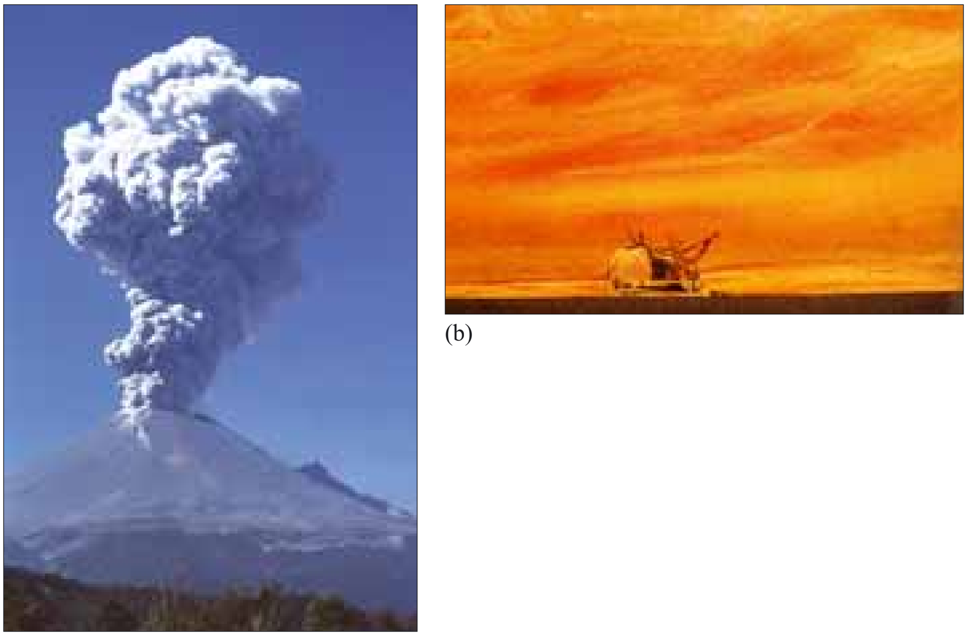
4.2 Volcanic aerosols and climatic change

> The ashes now began to fall upon us, though in no great quantity … darkness overcame us, not like that of a cloudy night, or when there is no moon, but of a room when it is all shut up and all the lights extinguished.
>
> (Pliny the Younger, AD 79)

This brief extract from Pliny the Younger's classic account of the eruption of Vesuvius in AD 79 reminds us that volcanic ash clouds can be highly effective at blocking out sunshine. Major explosive eruptions (still called Plinian eruptions to commemorate Pliny's observations) blast huge amounts of volcanic ash high into the atmosphere, often reaching the stratosphere (Figure 4.4a). Awe-inspiring though these eruption columns are, it is not the ash particles that cause the most significant climatic effects. Ash fragments are relatively large (ranging from shards 1–2 μm across to fist-sized lumps of pumice) and they fall back to the ground over timescales ranging from seconds to days. The longest-lasting environmental effects are produced by the volcanic gases.

As mentioned above (Table 4.1), volcanoes produce large quantities of sulfur dioxide as well as carbon dioxide. After eruption, sulfur dioxide reacts with water vapour in the atmosphere to form aerosols (i.e. tiny airborne droplets) of sulfuric acid, H_2SO_4. These acid aerosols form by a complex series of photochemical reactions that may continue for months, replenishing the aerosol cloud so that new particles form while older, larger ones settle out of the atmosphere. Aerosol particles are very tiny (initially only 0.1–1 μm in diameter) so any reaching the stratosphere may remain suspended there for months or years, far longer than solid ash fragments.

Volcanic aerosols often produce gloriously colourful sunsets (and sunrises), which can be enjoyed around the world (Figure 4.4b). During the day, the sky looks blue


(a)

(b)

Figure 4.4 (a) A Plinian-type explosive eruption of Popocatepetl (Mexico) in December 1998; the eruption column is about 5 km high. (b) During the autumn of 1883, after the eruption of Krakatau, spectacular sunsets around the world attracted the attention of artists. This is one of a series of six paintings by William Anscom of the sun setting on a November evening seen from Chelsea, London. It was published in the contemporary Royal Society report on the great eruption.

because gas molecules scatter sunlight in all directions. These molecules are smaller (~0.03 μm) than the wavelength of light and scatter blue wavelengths (~0.4 μm) more effectively than red (~ 0.7μm). Thus, blue light from the sun actually reaches your eye even when you are not looking directly at it, whereas the other wavelengths do not. When the size of particles is about the same as the wavelength of light, as in aerosols and smoke, the scattering process is more complex but the dominant effect is a shift to longer visible wavelengths, around orange to red in the visible spectrum. It is less intense than the blue-sky effect, and is most noticeable at sunset and dawn when the sun is low in the sky and its rays have to traverse a much longer atmospheric path (Figure 1.4).

In detail, the physics of light scattering by aerosols is complex, but the overall effect is that when a layer of volcanic aerosols gets between the Earth and the sun, a fraction of the sun's radiation is scattered back to space. In other words, the presence of the aerosols effectively increases the Earth's albedo (Figure 1.12). To estimate the climatic consequences of increased aerosols in the atmosphere, the balance between incoming solar radiation and outgoing longwave (thermal) radiation from the Earth and atmosphere has to be taken into consideration. As Figure 1.12 showed, a fraction of the incoming radiation is reflected back to space; of the remainder, some passes directly through to the surface of the Earth, and some is **forward-scattered**, also reaching the ground. The ground heated by the sunshine warms up, and reradiates longwave thermal energy, some of which is reradiated back to Earth by the atmosphere and clouds, while the remainder escapes to space. Incoming radiation (shortwave, S_{in}) and outgoing radiation (reflected shortwave, S_{ref}, plus thermal longwave, L) must balance:

$$S_{\text{in}} = S_{\text{ref}} + L \tag{4.1a}$$

Figure 4.5a applies these symbols to a simplified version of Figure 1.12: part (b) of Figure 4.5 is the same as part (a), but also includes the effect of aerosols, which not only scatter incoming solar radiation back to space but also scatter longwave thermal radiation back towards the Earth's surface. If solar radiation scattered back into space by the aerosol layer (S_{a}) is now included in the equation along with longwave thermal radiation **back-scattered** to Earth by the aerosol layer (L_{a}), this equation can be modified to read:

$$S_{\text{in}} = S_{\text{ref}} + S_{\text{a}} + (L - L_{\text{a}}) \tag{4.1b}$$

S_{a} will add a component to the sunlight already normally reflected by clouds, while L_{a} will reduce the amount of longwave thermal radiation escaping to space (Figure 4.5b).

■ What would be the overall effect of increasing the term $S_{\text{ref}} + S_{\text{a}}$, bearing in mind that the radiation budget would eventually reach a state of balance?

■ In order to maintain balance, the term $L - L_{\text{a}}$ would have to decrease.

$L - L_{\text{a}}$ is the thermal radiation escaping from the Earth; if this term is decreasing, the Earth's global temperature must be decreasing. It follows that increasing the term S_{a} (and hence $S_{\text{ref}} + S_{\text{a}}$) will lead to global cooling. In other words, until proved otherwise, volcanic aerosols will cause global cooling.

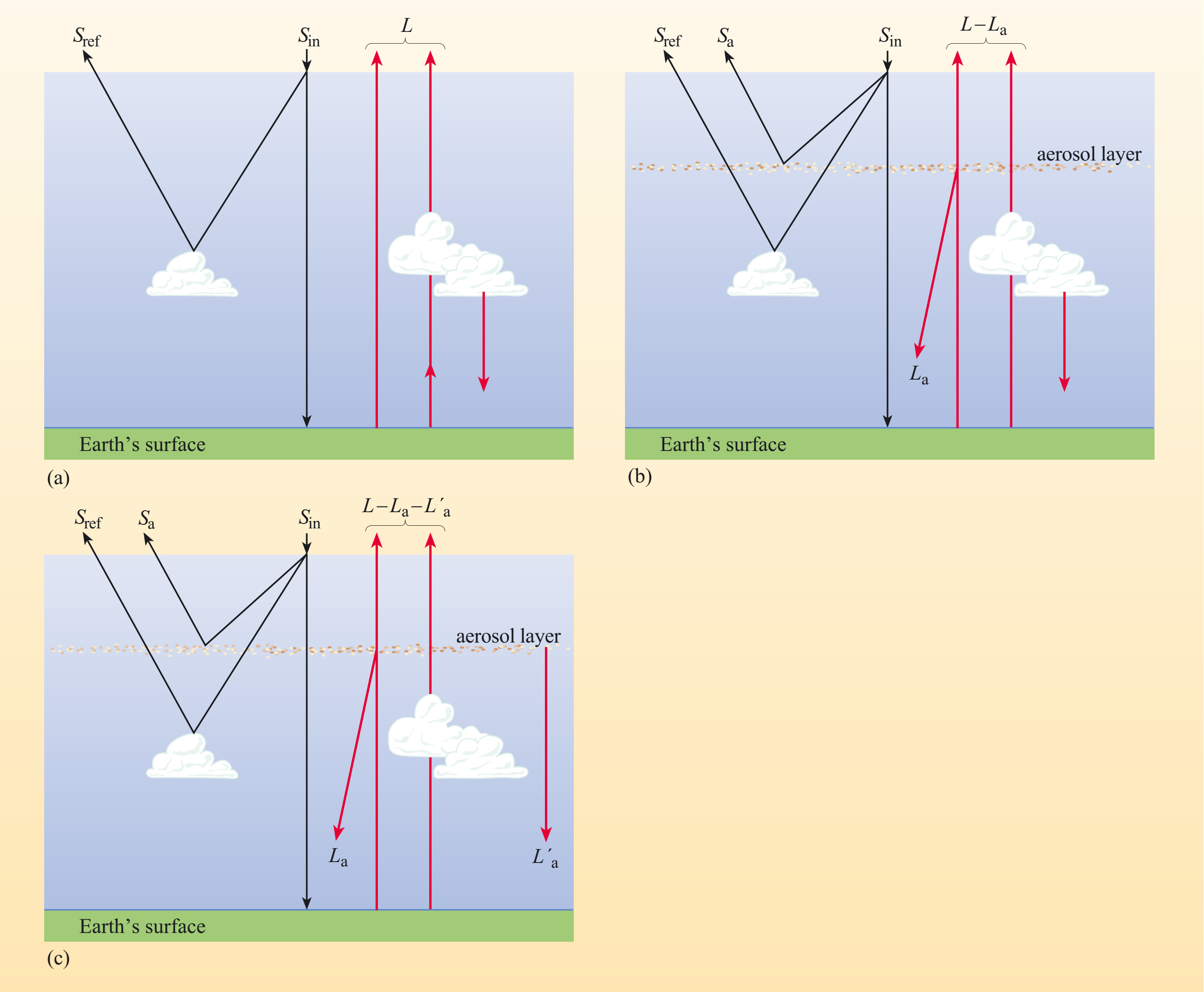


Figure 4.5 (a) Simplified version of Figure 1.12 where S_{in} represents incoming solar radiation, S_{ref} represents the solar radiation reflected by the Earth (i.e. by clouds, the Earth's surface and the atmosphere), and L represents the longwave (i.e. thermal) radiation that escapes to space. (b) If a layer of aerosols is added to the situation shown in (a), the amount of solar radiation reflected back to space becomes $S_{ref} + S_a$, and the amount of longwave radiation that escapes to space becomes $L - L_a$. (c) If the aerosol layer also absorbs outgoing thermal radiation and reradiates an amount L'_a towards the Earth, the amount of longwave radiation escaping to space becomes $L - L_a - L'_a$. Whether the net result is global cooling or global warming depends on the relative sizes of S_a, L_a and L'_a; in general though, S_a is greater than $L_a + L'_a$ and the net result is global cooling.

Things are more complicated in detail and, for a given aerosol layer, the fraction of solar radiation that is actually scattered back into space depends on both the amount of aerosol present and the size of the droplets. In general, an increased albedo caused by aerosol-scattering cuts down the sunlight reaching the surface of the Earth, causing cooling. While droplets of all sizes may scatter radiation, they can, however, also absorb it if their diameters are of the same order as the wavelength of the radiation. As reradiated thermal radiation is of relatively long wavelength, it can be absorbed by large aerosols, which will themselves emit longwave radiation, emitting some of this back towards the Earth (Figure 4.5c).

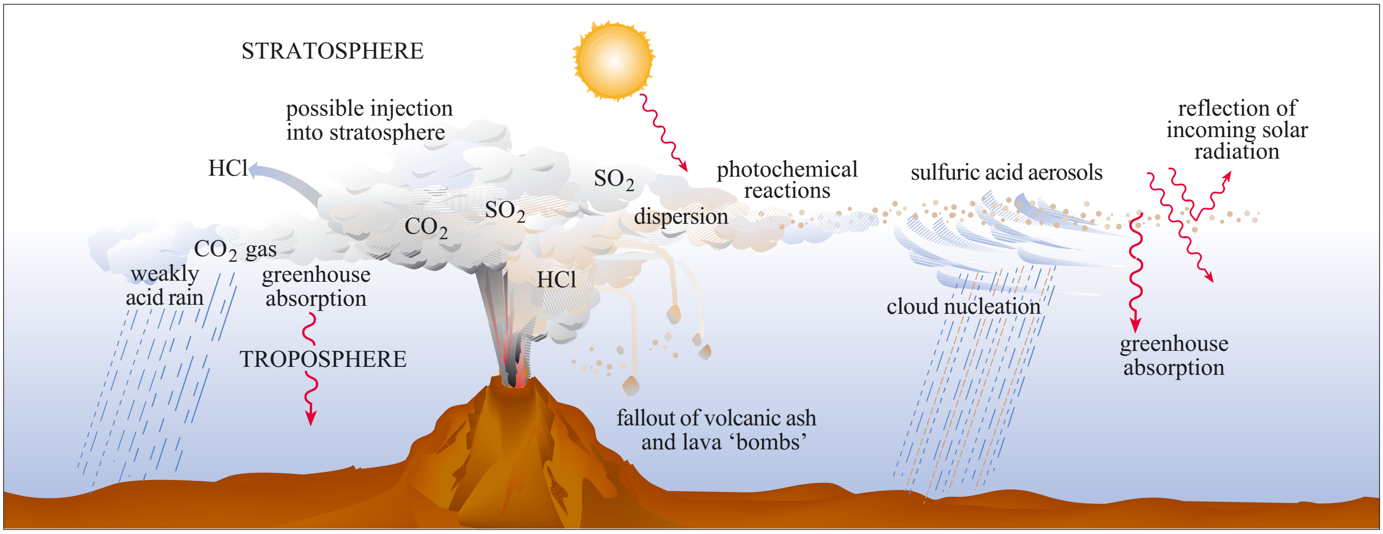


Figure 4.6 Physical and chemical interactions of a volcanic eruption with the atmosphere. The eruption releases sulfur dioxide (SO_2), which is eventually converted to sulfuric acid (through reaction with atmospheric oxygen and water), producing a mist of droplets, or aerosols. These reflect a fraction of the incoming solar radiation, cooling the troposphere; they also absorb some solar radiation, warming the stratosphere. If the aerosols are sufficiently large, the aerosol layer may absorb and reradiate thermal energy, warming the lower atmosphere (Figure 4.5). Note that they are sometimes referred to as *sulfate* aerosols rather than sulfuric acid aerosols. The eruption also produces large amounts of CO_2, which mixes in the atmosphere, forms a much weaker acid with atmospheric water, and does not form aerosols. Smaller amounts of hydrogen chloride (HCl) are also released into the atmosphere, forming hydrochloric acid.

An aerosol layer can therefore act as a 'greenhouse', preventing some thermal radiation escaping and so causing surface warming.

It is difficult to make precise predictions about the effects of an aerosol layer on the Earth's surface temperature. Indeed, the physical and chemical interactions of an eruption are complex (Figure 4.6). For most long-lived volcanoes, however, the volcanic aerosols are relatively small such that they cause a net cooling as can be seen from Table 4.2.

Table 4.2 Computed climatic forcings and the average cooling that resulted from some major recent eruptions.

Location of eruption and year	Mass of material erupted /10^{11} kg	Mass of aerosols in stratosphere /10^{9} kg	Climatic forcing /W m^{-2}	Observed Northern Hemisphere temperature change /°C
Katamai, Alaska (58° N), 1912	240	25	–3.5	–0.2
El Chichón, Mexico (17° N), 1982	7.5	12	–1.7	–0.2
Pinatubo, Philippines (15° N), 1991	120	30	–4.3	–0.5
Krakatau, Indonesia (6° S), 1883	247	55	–7.8	–0.5
Agung, Indonesia (8° S), 1963	24	20	–2.86	–0.3
Tambora, Indonesia (8° S), 1815	2400	150	–21.4	–1.0

The precise climatic impact of a volcano is related not only to the flux and composition of the volatiles, but also the altitude to which they are ejected and the latitude of the volcano (Figure 4.6). If the character of an eruption is such that the erupted gases are not injected into the stratosphere, then there is little chance of the resulting aerosols having any global effect, as they will be rapidly rained out of the troposphere.

Question 4.1

(a) Referring to Figure 1.20, explain why the effect of aerosols injected into the troposphere from a volcano in (for the sake of argument) the Southern Hemisphere will be largely confined to that hemisphere.

(b) Why is an eruption from a volcano at high latitudes more likely to eject gases (and hence aerosols) into the stratosphere than a volcano at low latitudes?

For example, about 40 days after the eruption of Mount Pinatubo (at ~15° N), stratospheric aerosols had reached high latitudes in both hemispheres. An eruption sending ejecta into the stratosphere at *high* latitudes, however, will have less effect in the other hemisphere than one taking place in the tropics because stratospheric circulation patterns are mainly zonal (east to west, or west to east) and so inhibit the spread of aerosols from one hemisphere into the other (north to south or vice versa).

4.3 Flood basalts and their effects on climate and life

So far you have considered only recent, short-lived volcanic events. Due to the long response time of the Earth system, it is unlikely that such events have had a sustained impact on global climate. A different kind of eruption that may have had more serious environmental effects than the explosive kind is the effusion of huge volumes of lava in the geologic past. These are termed **flood basalts**.

Accumulations of flood basalt lava flows cover large parts of the Earth. Some of the statistics are impressive: India's 65 Ma Deccan Traps cover 0.5 million km^2 (Figure 4.7), and may have covered 1.5 million km^2 when first erupted. They have an average thickness of at least one kilometre. Most of their huge volume may have been erupted in less than 0.5 million years (although this is controversial) and they may consist of many hundreds of individual lava flows. On erosion, they acquire a distinctive topography resembling flights of steps. In northwestern USA, the Columbia River Province consists of 240 000 km^3 of flood basalts erupted between 17 Ma and 12 Ma, in individual flows 20–50 m thick, covering an area of more than 200 000 km^2 (Figure 4.8). Approximately 90% of the total volume of the province may have been erupted during a period of less than 1.5 million years between 16.5 Ma and 15 Ma. Work on individual eruptive units of basalt lava in the Columbia River Province has shown that >1000 km^3 could have been erupted in continuous, individual eruptions lasting a decade or longer.

The next step is to consider what impact such massive eruptions would have on global climate. At any given time in Earth history, the amount of carbon dioxide in the atmosphere is the result of the balance of fluxes between sources and sinks in the carbon cycle. If the average global flux of carbon dioxide to the

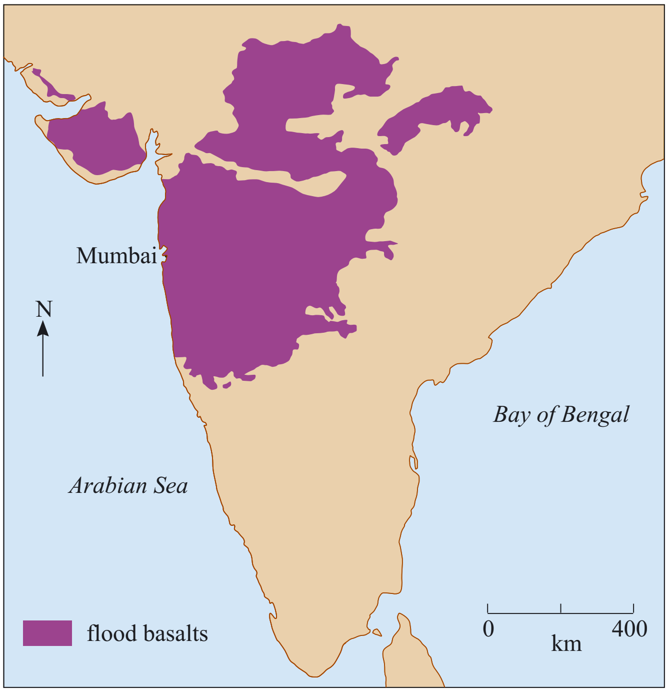


Figure 4.7 The areal extent of the Deccan Traps. A significant fraction of India's land surface is covered by the outcrop of these 65 Ma basalt lavas.

Figure 4.8 Layers of stacked lava flows in the Columbia River Province of the Columbia River. The flows shown represent just a small thickness of the voluminous and rapidly erupted Grande Ronde Formation part of the succession. (Steve Self/Open University)

atmosphere from volcanoes should increase for some reason, and if that increased flux is sustained, then the carbon cycle will settle into a new equilibrium, with important implications for climate (Section 3.4).

But is there any evidence that the flux of carbon dioxide into the atmosphere has varied over geological time? There are some pointers in this direction. Most attention has been focused on the Cretaceous Period (145–65 Ma), which seems to have been an exceptionally productive period in terms of creation of ocean crust and eruption of massive basaltic plateaux. So much new material was erupted in the Pacific Basin during the Cretaceous that it is difficult to find any older oceanic crust or sediments there – they have been simply swamped by

Cretaceous lavas. Two huge submarine volcanic plateaux were formed at this time: the Ontong–Java Plateau at about 122 Ma, and the Kerguelen Plateau at about 112 Ma (Figure 4.9). Both of these probably grew partly above sea level when they formed. The Ontong–Java Plateau is estimated to be 25 times bigger than the Deccan Traps, and one result of so much new lava being erupted was that it elevated the global sea level by about 10 m, displacing seawater like a body in a bath tub.

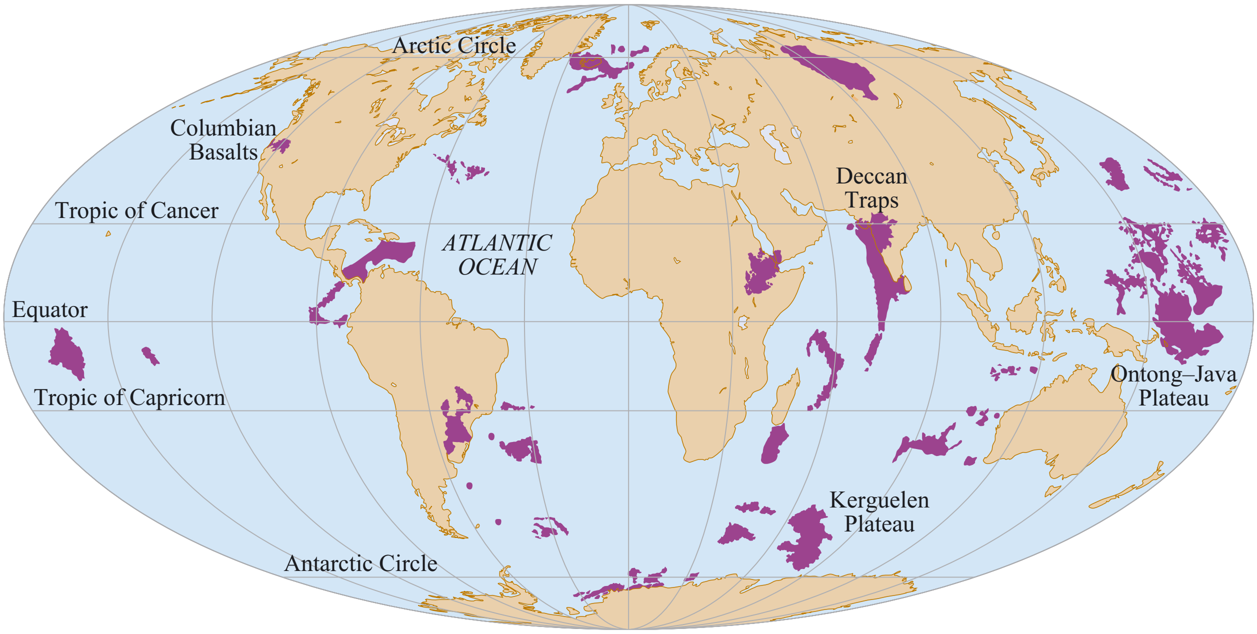


Figure 4.9 Global distribution of large igneous provinces, including the large oceanic plateaux.

Estimates of the average global rate of ocean crust production over the past 150 Ma range between 16 and 27 $\text{km}^3\ \text{y}^{-1}$. Estimates of the production rate of the Ontong–Java Plateau on its own are around 12–15 $\text{km}^3\ \text{y}^{-1}$, i.e. one flood basalt province produced the same order of magnitude of basalt as the entire oceanic ridge network.

From the graph of estimated oceanic crust production rate through time (Figure 4.10), it can be seen that global rates during the early Cretaceous (145–100 Ma) were relatively high. Such estimates are hotly debated, with some researchers claiming that mid-ocean ridge crust production has been constant for 150 Ma, but the abrupt formation of the enormous Ontong–Java and Kerguelen Plateaux cannot be denied. These two plateaux, and some lesser ones in the southwest Pacific, certainly contributed to the higher rates of crust production at this time.

What were the effects of all this volcanic activity on CO_2 in the atmosphere? There may have been both direct and indirect consequences. As discussed at the start of this chapter, as long as it remained below sea level, the volcanism taking place at mid-ocean ridges, and leading to creation of new oceanic crust, would have had little immediate effect on the atmosphere. Increased sea-floor spreading, however, has to be balanced by increased subduction – if it were not, the total area of the Earth's lithosphere would increase, so that the Earth would have to expand. When oceanic crust is subducted, carbonate-rich sediments are

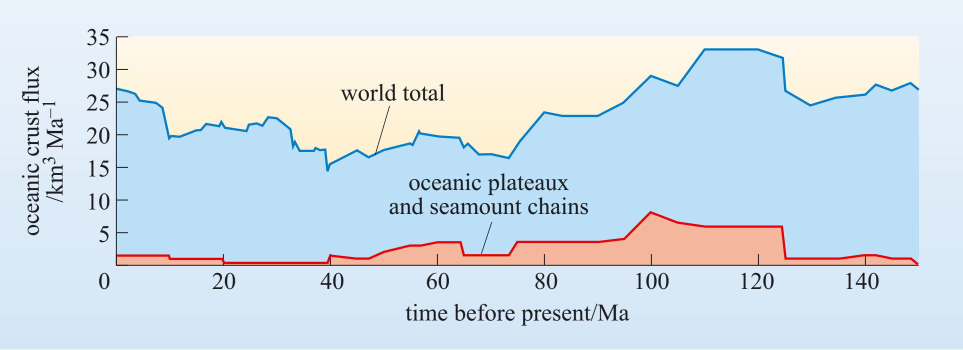


Figure 4.10 Variation with time of the estimated rate of production of oceanic crust. Note the peak between about 120 Ma and 80 Ma.

also subducted, and their carbon may ultimately be exhaled as CO_2 at subduction zone volcanoes. More importantly, if mantle plumes lead to the formation of several large volcanoes (e.g. Hawaii) extending above sea level, then a direct flux of CO_2 from the mantle into the atmosphere could be sustained.

Estimates of the abundance of CO_2 during the Cretaceous suggest that atmospheric CO_2 could have reached a level at least four times the modern pre-industrial level of 280 ppm, leading to global greenhouse warming of between 3 °C and 8 °C. This would have been accompanied by other changes, such as an increase in sea level. A runaway greenhouse effect was counteracted by increased deposition and preservation of carbon in the warm oceans and shallow seas.

It is important to consider whether this increase in atmospheric CO_2 concentration could be the direct effect of the eruption of flood basalt or whether other sources need to be considered. Initial estimates based on the (admittedly sparse) knowledge of how much mantle CO_2 is actually released into the atmosphere suggest that these volcanic fluxes are fairly low. For example, the estimated annual flux of CO_2 to the atmosphere from the eruption of 1000 km^3 of basalt over a decade is about 3% of the present-day natural land–atmosphere flux – too small to have a significant effect on the CO_2 concentration in the atmosphere. Such a modest flux would probably allow carbon cycle processes to offset the volcanic addition. This will still result in some global warming, however, and could trigger other mechanisms such as the destabilisation of **methane hydrate** deposits and the release of methane into the atmosphere. There is no doubt that the short-term impact of aerosol emission (i.e. over a period of a few years) is often one of global cooling (Table 4.1), but there remains considerable uncertainty over the extent of long-term warming (i.e. over a period of millions of years) associated with the eruption of large-scale flood basalts.

Question 4.2

Some volcanologists and climatologists argue that volcanic eruptions usually affect weather, not climate. What aspect of volcanic eruptions underlies this point of view? In what circumstances might it be appropriate to speak of the effects of volcanism on just climate?

It is reasonable to ask how life on Earth was affected during this, and similar, periods of sustained volcanic activity. At the end of the Cretaceous Period (65 Ma) there was worldwide extinction of many animal species, i.e. a **mass**

extinction, both in the sea and on land. The demise of the dinosaurs has captured the public imagination, but many less-notorious groups of species also became extinct, notably many planktonic foraminiferans. An increasing body of evidence indicates that at that time a major impact event covered the Earth in a film of iridium-rich ejecta, now seen in the sedimentary record at the Cretaceous–Tertiary (K/T) boundary. Researchers have even zeroed-in on the site of the impact, at Chicxulub on the Yucatan peninsula of Mexico.

At the same time as the Chicxulub impact, at least 10^6 km^3 of basalts were being erupted to form the Deccan Traps of India (Figure 4.7). Were these two remarkable events connected? Did the impact extinguish the dinosaurs or was it the basaltic flood eruption, or the combined effect of both? Did the impact event trigger the basaltic flood or did the two extraordinary events reinforce each other in some lethal way to cause mass extinctions? These highly contentious issues have sparked vigorous debates for more than 20 years.

As long ago as the early 1970s, some scientists argued that widespread basaltic volcanism might have been implicated in the extinction of the dinosaurs. A round-number estimate of the volume of the Deccan basalts is 10^6 km^3, suggesting that a total of about 10^{16} kg of SO_2 gas was released, which has the potential to generate $\sim 2 \times 10^{16}$ kg of sulfuric acid aerosols. It seems inevitable that the huge mass of acid aerosols produced by the Deccan eruptions (with each one of the ~1000 eruptions producing $\sim 2 \times 10^{13}$ kg of aerosols) had far-reaching consequences.

An even greater mass extinction defines the Permian–Triassic boundary, at 251–245 Ma, separating the Palaeozoic ('ancient life') Era from the Mesozoic ('middle life') Era. Huge numbers of species, including trilobites (marine arthropods, probably related to the ancestors of the crustaceans), became extinct by the end of the Permian, while the Mesozoic saw the blossoming of the age of reptiles, including the dinosaurs. Is it purely coincidental that the Palaeozoic–Mesozoic boundary approximately coincides with the vast outpouring of the Siberian flood basalts, currently the largest subaerial flood basalt province known (Table 4.3)? More than 3×10^6 km^3 of basalt may have been erupted within a period of less than one million years, an average rate of more than 3 km^3 y^{-1} (about 100 m^3 s^{-1}).

■ Does the extinction event at the Tortonian–Serravallian boundary (Table 4.3) correspond with a flood basalt eruption, within the limits of error of the ages cited?

■ No. None of the flood basalt eruptions fall in the period 9.5–11.5 Ma (within error of this extinction event). The closest is the Columbia River eruption at 17.0 ±0.2 Ma.

Question 4.3

Which extinction events within the past 300 Ma (listed in Table 4.3) correspond with a flood basalt eruption, within the limits of error of the ages cited?

These correlations indicate that it is plausible that the flood basalts may be implicated in at least some extinction events.

Table 4.3 Table of peak eruption ages for flood basalt provinces and ages of extinction and mass extinction events for the most recent 260 Ma of Earth history. Mean ages are derived mainly from Courtillot and Renne (2003), and stratigraphic boundaries quoted from Gradstein, et al. (2003). M = Mass extinction event, E = extinction event.

Flood basalt or large igneous province	Mean age (Ma)	± (Ma)	Stage or epoch boundary at which extinction occurred	Age (Ma)	± (Ma)
Emeishan Traps	259	3	Guadalupian–Lopingian (M)	260.4	0.7
Siberian Traps	250	1	Permian–Triassic (M)	251	0.4
Central Atlantic Magmatic Province	201	1	Triassic–Jurassic (M)	199.6	0.6
Karoo Ferrar	183	2	Pliensbachian–Toarcian (E)	183	1.5
			Bajocian–Bathonian	168	3.5
			Tithonian–Berriasian (E)	145	4
Parana Etendeka	133	1	Valanginian–Hauterivian	136.4	2
Ontong Java 1	122	1	Early Aptian (E)	125	1.0
Rajmahal/Kerguelan	118	1			
Ontong Java 2	90	1	Cenomanian–Turonian (E)	93.5	0.8
Caribbean Plateau	89	1			
Madagascar Traps	88	1			
Deccan Traps	65.5	0.5	Cretaceous–Palaeogene (M)	65.5	0.3
North Atlantic 1	61	2			
North Atlantic 2	56	1	Paleocene–Eocene (M)	55.8	0.2
Ethiopia and Yemen	30	1	Oi1 event	30	2.5
Columbia River	16	1	Early Miocene–Mid Miocene (E)	16.0	0.1
			Serravallian–Tortonian	11.6	0.3
			Pliocene–Pleistocene (E)	1.81	0.02

4.4 Continental drift and climate

Volcanism is only one means by which plate tectonics affects global climate. The process by which continents are separated is known as sea-floor spreading which occurs at rates of a few centimetres a year, such that major ocean basins open and close on timescales of 100–200 Ma. In other words, continents break up, are carried over the surface of the globe, and eventually are brought together again, at intervals of hundreds of millions of years. Not surprisingly, such movements cause the climate of individual continents to change through time. For example, at 300 Ma the British Isles was located at the Equator, and since that time has migrated northwards to its present position at latitude 50–55° N. During this time, the climate affecting the British Isles has become became markedly cooler. Is it possible though, that *global* climate could itself be affected by the migrating continents?

The periods when the Earth is partially covered by extensive ice sheets are referred to as ‘Ice Ages’. We are presently within an Ice Age, known as the Quaternary Ice Age: there is a large ice cap over the landmass of Antarctica, another over Greenland, and permanent ice cover over much of the Arctic Ocean. If the present Ice Age is typical, you can assume that, within Ice Ages, ice caps grow and retreat many times, as glacial periods alternate with interglacial periods. At the present time, the Earth's ice caps are by no means at their maximum extents and the climate, even at fairly high latitudes, is moderate – in other words, we are currently experiencing an interglacial period.

Over the history of the Earth there have been five, or possibly six, Ice Ages, three of which were in the last 500 Ma. They occur at intervals of hundreds of millions of years.

Question 4.4

Time- and length-scales are vital in considering the climatic implications of different phenomena (Figure 4.11).

(a) Where would major periods of extended glaciation (Ice Ages) plot on Figure 4.11?

(b) Why might this suggest that large-scale changes in the positions of the continents could be at least a contributory cause of the initiation and ending of Ice Ages?

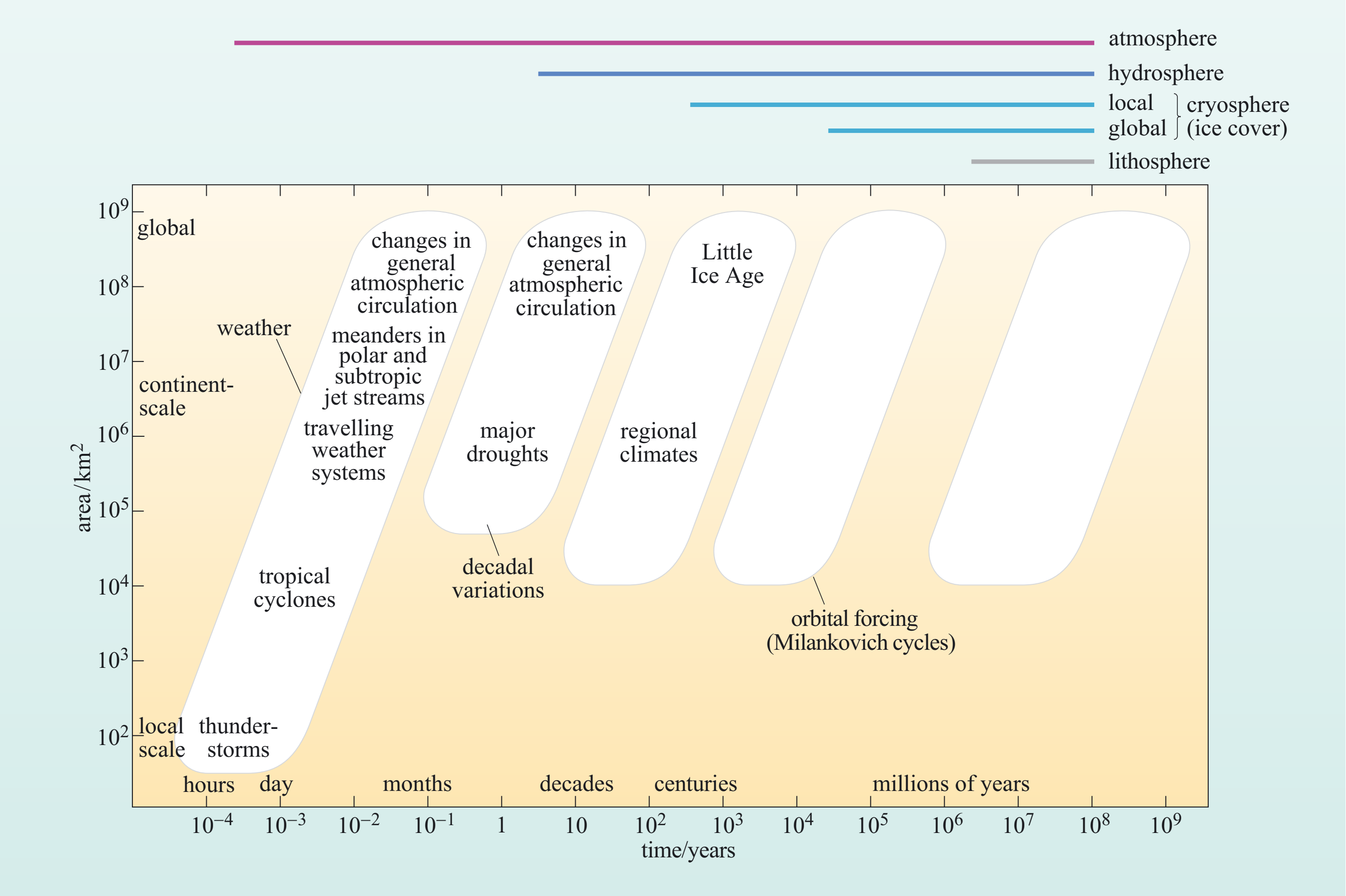


Figure 4.11 Schematic diagram showing the time- and length-scales of various phenomena affecting conditions on Earth. (The continual increase in solar luminosity since the formation of the Solar System is not included.) The bars along the top show the range of timescales on which each of the major parts of the Earth system vary. These bars illustrate the general principle that while short-term variations are mostly limited to the atmosphere, longer-term variations involve progressively more components of the Earth system. To be completed in Question 4.4.

In other words, plate motions would seem to be an obvious candidate for a major forcing factor implicated in the coming and going of Ice Ages, because they act over a similar timescale. Plate motions are most unlikely to be the only factor, however, as of themselves they cannot bring about climate change. What they can do though, is change the distribution and relative positions of continents and oceans in ways that enable other influences to come into play. The ways in which the distribution of land and sea might affect climate can be explored by looking at models that use greatly simplified situations to simulate conditions that might have been obtained in the past, when distributions of continents and oceans were different (Box 4.1).

Box 4.1 Model worlds and their climates

Although contrasting global climates in the geological past appear to be associated with different continental distributions, the only way the relationship between continental configuration and climate can be quantitatively investigated is through the use of computer simulations. Before considering how the two are related, it is worth glancing at the imaginative speculations of Charles Lyell (the 'Father of Geology'). In his book *Principles of Geology*, first published in 1837, Lyell redrew the map of the world to show the present-day continents distributed either in an equatorial/tropical 'ring' (Figure 4.12a) or in two polar 'caps' (Figure 4.12b). This was long before geologists had begun to think about continental drift, but Lyell's vision was sufficiently far-sighted for him to recognise the consequences of shifting continents relative to climatic belts. It was a prodigious intellectual leap, and it is further evidence of Lyell's genius that he also proposed the inverse effect: that shifting continents might themselves be an agent of global climate change.

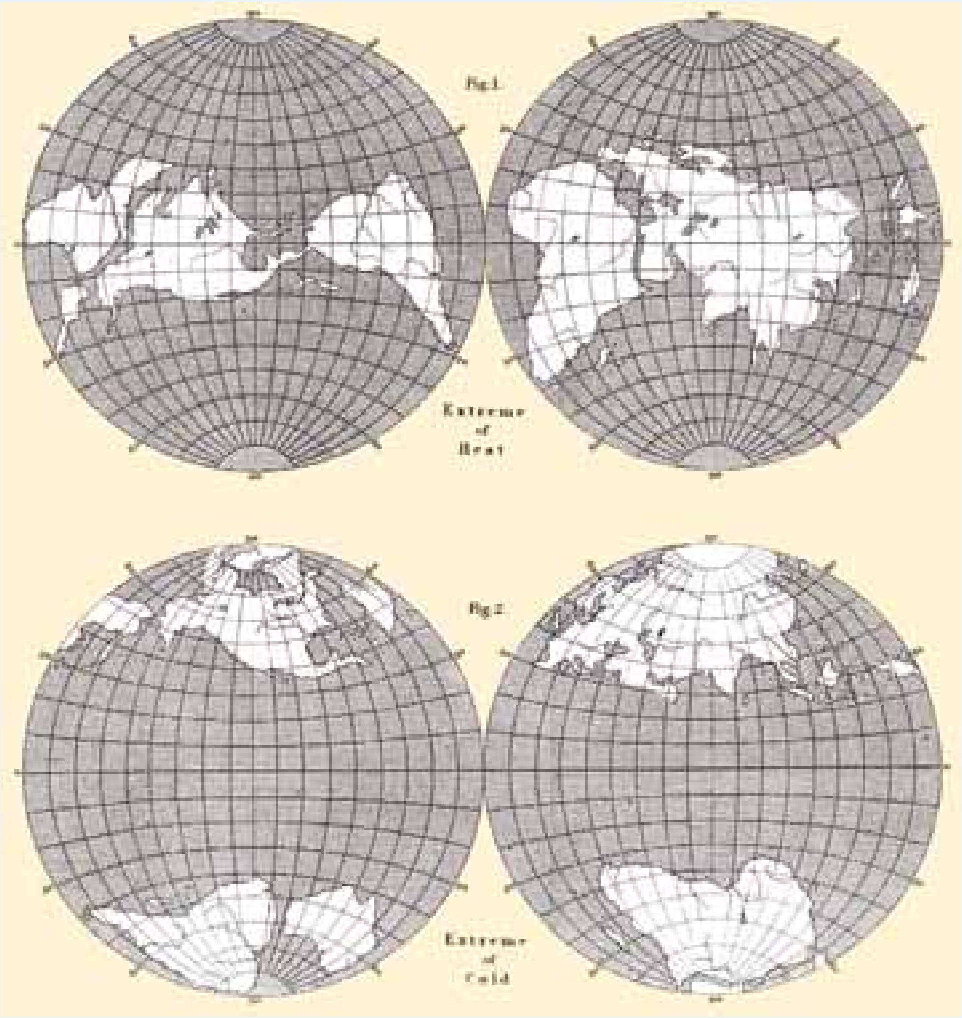


Figure 4.12 Facsimiles of Charles Lyell's maps of present-day continents redistributed to form (a) a 'ring world' with the continents concentrated in the tropics; (b) a 'cap world' with the continents gathered together round the two poles.

Today, if a problem can be written down mathematically, for instance as a number of simple heat budget equations, it can usually be solved using computers. Below, idealised extremes of continent–ocean distribution have been used to investigate the effects of continental configurations on climate by means of a model consisting of a set of simple, connected balances of heat and moisture in different latitude bands. The model simulates some of the familiar features of the Earth's climate system including seasonal and latitudinal variations of incoming solar radiation, surface temperatures, cloud cover, precipitation and evaporation, and snow and ice cover.

The model was run using two idealised continental geometries based on the present-day total land area:

- two 'caps' of land extending from the poles to 45° of latitude (both with and without ice caps, Figure 4.13a and b);
- a tropical 'ring' of land extending 17° north and south of the Equator (Figure 4.13c).

These idealised continental distributions are nothing like those of today, but the geography of the Earth may have approximated a tropical 'ring world' between 700 Ma and 600 Ma, and could have approximated very roughly a 'cap world' with one polar ice cap in the Late Carboniferous at 300 Ma.

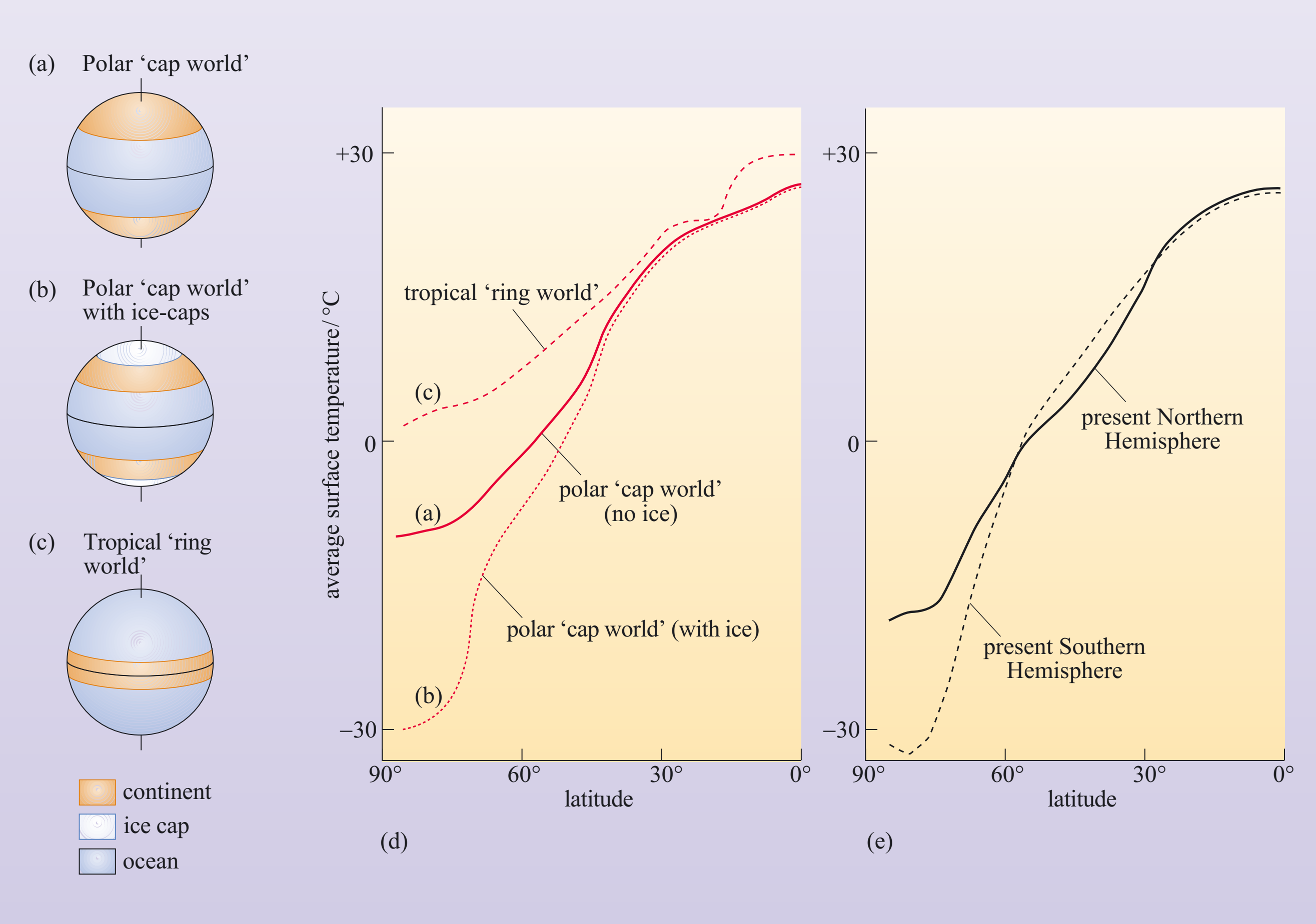


Figure 4.13 Simplified models of the Earth with present-day total land area redistributed to form (a) a polar 'cap world' without ice caps, (b) a polar 'cap world' with ice caps extending equatorwards to latitude 70°, and (c) a tropical 'ring world'. (d) Mean annual surface temperatures (zonally averaged), simulated for the cap and ring worlds shown in (a). (e) Mean annual surface temperatures (zonally averaged) for each hemisphere, simulated for present-day geography.

As you might expect, for a polar cap world with ice caps (Figure 4.13b) the meridional variation (i.e. variation along a line of longitude) in temperature (Figure 4.13d) looks rather similar to a simulation of the temperature distribution for the present-day Southern Hemisphere (Figure 4.13e) and both of the cap worlds give average equatorial temperatures close to those of the present day. By contrast, the tropical ring world (Figure 4.13c) is significantly warmer than either of the cap worlds.

■ In terms of radiation balance, will the addition of ice caps lead to a cooler world?

■ Ice has a much higher albedo than exposed or vegetated continental crust (Table 1.1), so an ice cap world reflects more solar radiation back into space than an ice-free cap world.

In summary then, a simple model of the kind shown in Figure 4.13a–c can confirm the original supposition that the relative distribution of continents and oceans can influence global climate. This particular model will not be considered further because it ignores important aspects of the Earth's climate system.

As demonstrated in Box 4.1, computer models can be useful in helping to isolate the effects of different factors (in this case, changes in albedo and moisture balance), while keeping other variables (e.g. the total continental area, and total incoming solar radiation) constant. They can also be used to combine the effects of different factors in a quantitative way, and to test hypotheses against physical laws. Climate models are based on a myriad of approximations, and while they can often demonstrate that certain events may possibly have occurred, it is much more difficult to attempt to prove that they must certainly have occurred.

In this case, two important components of the climate system not incorporated into the model are:

- heat transport by surface ocean currents and by the deep thermohaline circulation
- changes in the concentrations of greenhouse gases in the atmosphere.

The second of these factors will be considered later in this chapter, but first it is important to understand how ocean currents might be affected by changes in the global distribution of continents and oceans.

4.4.1 Ocean currents and climate change

Cap, ring and slice worlds (Figures 4.13a–c and 4.14) are very simplistic representations of the Earth, but do help to show how continental configurations might influence ocean current patterns and hence climate.

■ Which configuration – cap world, ring world or slice world – would affect the surface current pattern so as to intensify cold conditions in polar regions?

■ Cap world, where currents carrying heat from lower latitudes would be unable to penetrate to very high latitudes, with the result that there would be a stronger temperature contrast between equatorial and polar regions.

In a cap world, the polar regions would be thermally isolated, making the development of polar ice caps more likely: strong eastward currents, comparable with today's Antarctic Circumpolar Current (Figure 1.31), could flow around the

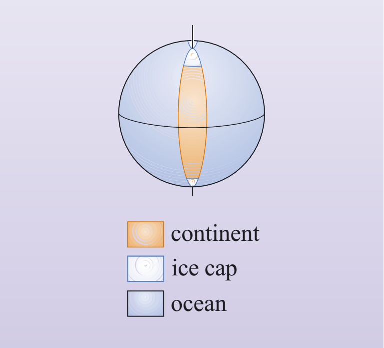


Figure 4.14 A 'slice world', showing the longitudinal configuration of one of the continental areas and parts of two polar ice caps.

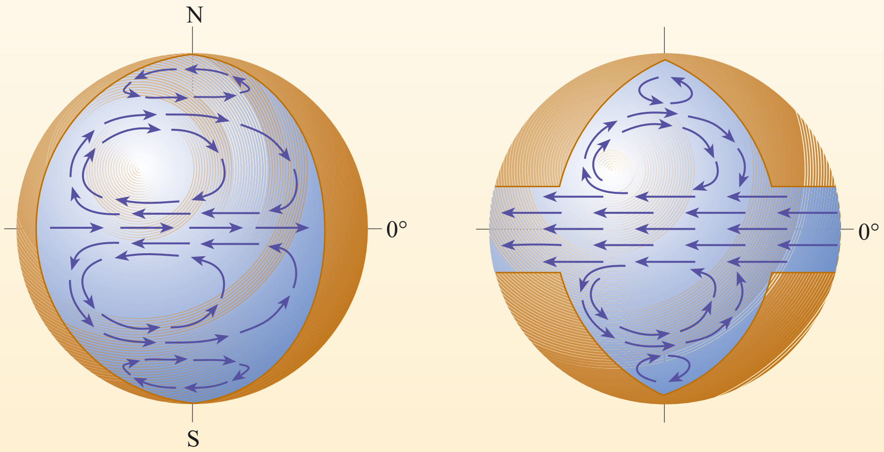
polar continents under the influence of westerly winds, further isolating the polar continents from warm currents flowing from low latitudes.

■ Look carefully at Figure 1.31. In today's world, are northern polar regions similarly isolated?

□ No, but they are nevertheless largely cut off from warm currents flowing from lower latitudes. The Arctic Ocean is almost completely surrounded by land, with the result that the only warm water penetrating the region is the North Atlantic Drift (the downstream extension of the Gulf Stream), which flows northwards through the Norwegian and Greenland Seas.

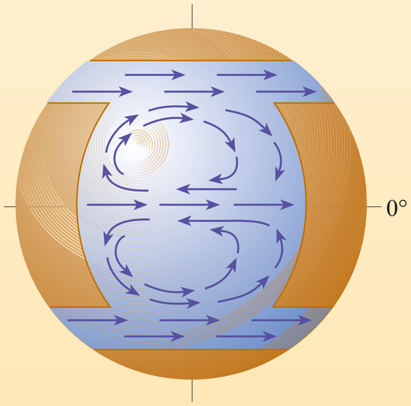
By contrast, in a tropical ring world (Figure 4.13c), ocean circulation between tropical and polar regions would be possible, so heat could be transferred from mid-latitudes to high latitudes, resulting in lower meridional temperature contrasts.

Figure 4.15 shows three more configurations of oceans and continents, not quite as simple as those shown in Figures 4.13 and 4.14. In the past, the configuration of oceans and continents has been very different, so it is worth using these simple models to aid your thinking about how current patterns may affect the distribution of temperature over the Earth's surface.



(a) Double-slice world

(b) Double-slice world with low-latitude seaway



(c) Double-slice world with high-latitude seaways

Figure 4.15 Three simple configurations of oceans and continents, with appropriate generalised surface current patterns: (a) a 'double-slice' world, with two continents and two oceans (only one visible here); (b) as (a), but with open oceans around the globe at low latitudes; (c) as (a) but with open ocean around the globe at relatively high latitudes.

■ On the basis of what you know about winds and currents on the real Earth, you could yourself have sketched surface current patterns similar to those shown in Figure 4.15. Identify features of the circulatory pattern, visible in all three configurations, that are likely consequences of fundamental characteristics of the rotating Earth and its fluid envelopes.

□ Features you may have recognised as being fundamental to a rotating Earth are: anticyclonic subtropical gyres; cyclonic subpolar gyres; poleward-flowing western boundary currents like the Gulf Stream; westward-flowing North and South Equatorial Currents. (Note that only configurations (a) and (c) have eastward-flowing Equatorial Counter-currents, because these are a result of westward flow in the vicinity of the Equator being deflected back *along the Equator* (where the Coriolis force is zero) by the western boundary (which is not present in configuration b).

■ Would you expect the temperature distributions over the oceans in Figure 4.15 to be symmetrical, with temperatures along the western boundary the same as those along the eastern boundary at similar latitudes?

□ No, the temperature distribution would not be symmetrical in any of the configurations shown. As discussed in connection with Figure 1.31, over much of an ocean, water warmed at low latitudes flows polewards along the western side, while water cooled at high latitudes flows equatorwards in the eastern part of the ocean. This results in the western sides of oceans being generally warmer than eastern sides (though, at the present day, the northeasterly flow of the Gulf Stream causes the northeastern North Atlantic to be warmer than the northwestern part).

Of course, discussing the transport of heat around the Earth by means of currents alone is unrealistic – as the redistribution of heat by winds, including the effects of evaporation, transport and condensation of water vapour is being ignored. Nevertheless, the high specific heat of water (Box 1.4) means that heat transport in the ocean is an extremely important influence on climate.

Implicit in the discussion of Figure 4.15 is the effect of gateways on the pattern of surface and deep ocean currents. Gateways are gaps between the continents that permit significant longitudinal or latitudinal connections to be made between oceans. Their opening and closing, and the resulting changes in heat transport, can cause climate for a particular landmass to change much more rapidly than the slow drift of the continent across climatic belts.

Mention of deep currents should remind you that an extremely important aspect of the oceanic heat transport system has not yet been considered. The effect of wind-driven surface currents in transferring heat from low to high latitudes is reinforced by the transport of cold deep water away from polar regions in the density-driven thermohaline circulation (Section 1.4.3). It may be facilitated by the opening of gateways in the deep ocean, or impeded by the development of topographic barriers such as mid-ocean ridges and seamounts. Barriers and gateways may be affected by crustal uplift or volcanic activity on the one hand, and subsidence related to lithospheric plate movements on the other.

4.4.2 The breakup of Pangea

Having used our simple models to consider some possible effects of different continental configurations on ocean circulation and global climate, you can now consider how events in the breakup of the supercontinent known as Pangea (Greek for '*all land*') might have influenced climate change. Pangea formed at ~250 Ma, when what is now North America, most of Europe, and Asia came together with the pre-existing southern supercontinent (comprising South America, Africa, India, Antarctica and Australia), which is referred to as Gondwana. Gondwana had been extensively glaciated about 50 million years before (i.e. at 300 Ma), but by 250 Ma, the Earth was probably already beginning to warm up, and by 100 Ma it was a 'greenhouse' planet some 10 °C warmer (on average) than today. As Pangea broke up and the continents separated, the Earth subsequently began to cool again, albeit gradually, to arrive at its present 'icehouse' state.

Before Pangea began to break up, the supercontinent must have been accompanied by a superocean. This superocean (which has been named Panthalassa, '*all ocean*'), extended from the North Pole to high southern latitudes (about 50–60° S), and is thought to have extended for some four-fifths of the Earth's circumference around the Equator. There is no reason to suppose that the factors determining the global wind and surface current systems were any different from what they are today, and some likely wind and current patterns have been proposed. If present-day oceanic current patterns are any guide, there could have been a number of linked subtropical gyres in each hemisphere (the present-day Pacific Ocean has a more complex gyral system than the narrower Atlantic, Figure 1.31), and presumably also a number of subpolar gyres.

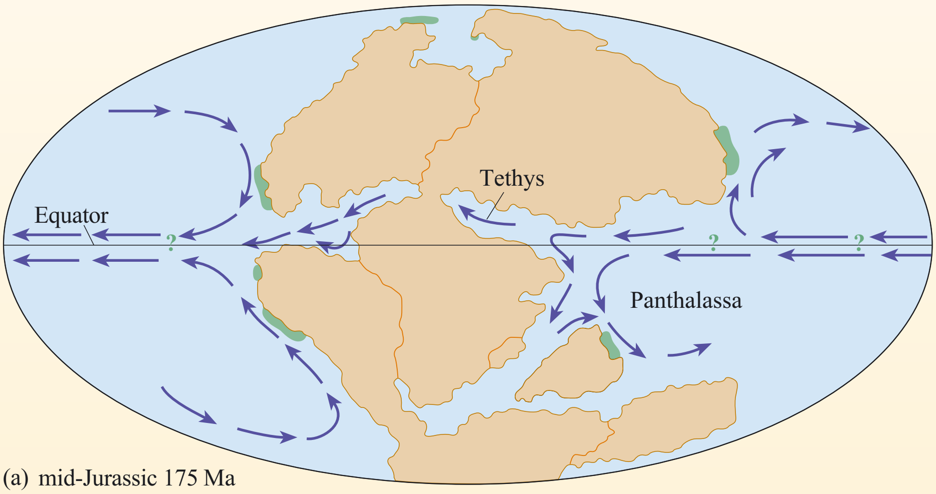
Figure 4.16 shows a series of maps that illustrate successive stages in the breakup of Pangea over the past 175 Ma. On the basis of what is known about global winds and the effects of landmasses on winds and currents, we can not only infer surface current patterns but also propose regions where deep and bottom water masses might have formed.

The breakup of Pangea began at about 200 Ma, and an equatorial gateway started to open within the supercontinent. By 175 Ma, in Jurassic times (Figure 4.16a), circum-equatorial flow was blocked by a relatively narrow isthmus at what is now Gibraltar, and the ancestral Mediterranean (known as the Tethys Ocean) had formed. Surface flow is shown as westwards on both sides of this isthmus in the reconstruction. It is possible that east of Gibraltar, the return flow eastwards was at depth, in the form of a dense water mass.

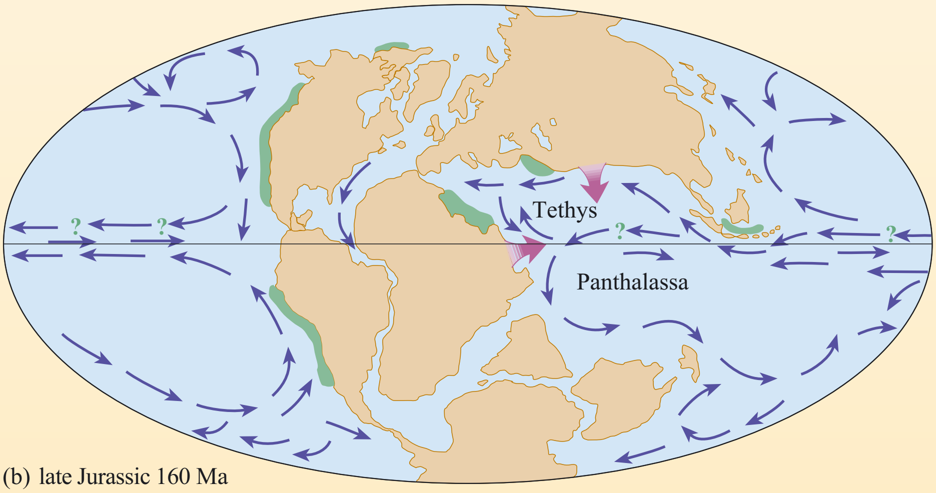
■ In a relatively warm Earth, how might surface water at low latitudes be made sufficiently dense to sink?

■ High rates of evaporation would remove freshwater and so increase the salinity.

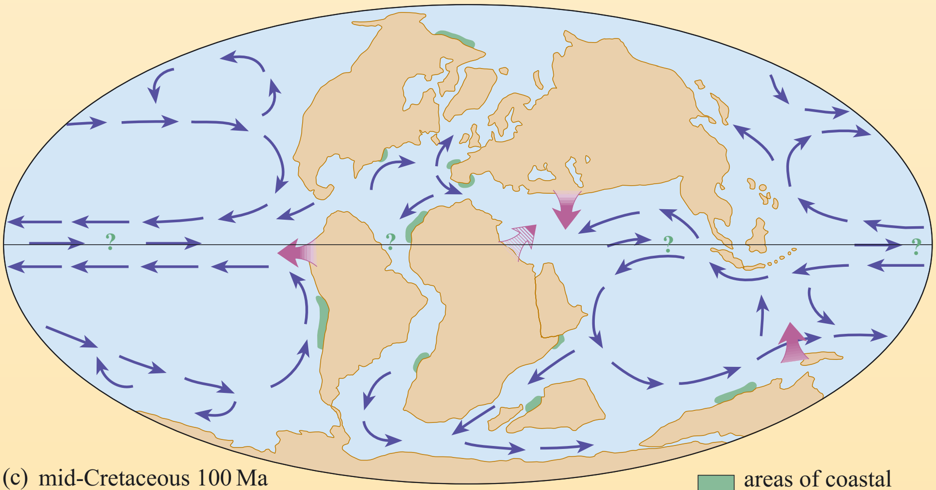
In fact, in today's Mediterranean, a warm saline deep water mass is formed in this way, aided by some cooling in winter. It is thought that by 160 Ma, similar saline but relatively warm water masses could have been forming in shallow



(a) mid-Jurassic 175 Ma



(b) late Jurassic 160 Ma

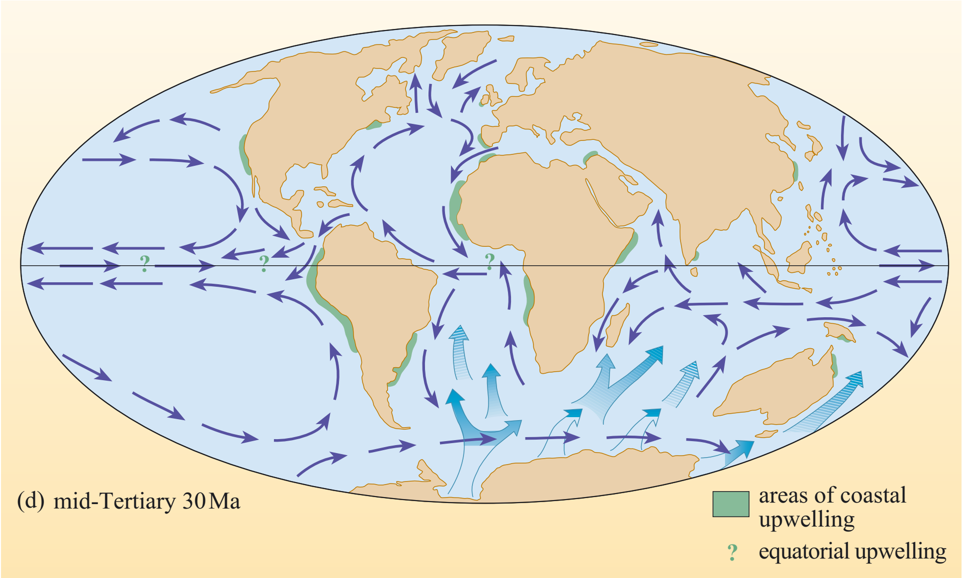


(c) mid-Cretaceous 100 Ma

🟩 areas of coastal upwelling

? equatorial upwelling

Figure 4.16 Maps showing the changing distribution of the continents and changing surface current patterns during the breakup of Pangea (no attempt has been made to show actual coastlines or shelf seas). In (b) and (c), broad arrows indicate locations where warm, saline water masses may have flowed down into the ocean from shallow evaporating basins. (a) 175 Ma, there was no Atlantic Ocean, only the great ocean Panthalassa. (b) By 160 Ma, the North Atlantic was a long narrow ocean but the South Atlantic had barely begun to open. (c) By 100 Ma, the opening of the Straits of Gibraltar and the submergence of 'Central America' had provided a low-latitude seaway. (d) By 30 Ma (overleaf), the oceans were approaching their present configuration. Gyres were now established in the North and South Atlantic, and currents could flow unimpeded around Antarctica under the influence of westerly winds. The climate was by now cooler, and cold bottom waters formed near Antarctica and flowed north into all the ocean basins – the equivalent of today's Antarctic Bottom Water.



coastal basins, subject to high rates of evaporation in the areas shown in Figure 4.16b. At 160 Ma, there was a land bridge between North and South America.

By the early Cretaceous, the Straits of Gibraltar had opened and rising sea levels had caused ‘Central America’ to become submerged, providing a shallow-water gateway. Circum-equatorial currents could now flow unimpeded round the Earth, and by the mid-Cretaceous (100 Ma, Figure 4.16c) this seaway was quite wide.

Ocean circulation patterns became more complicated as the Atlantic and Indian Oceans opened, but it would seem that by the mid-Tertiary (30 Ma, Figure 4.16d) much of the surface circulation resembled the pattern seen today. Closure of the Tethys Ocean was well advanced, shutting off the equatorial seaway. Submergence of the Tasman Ridge opened a gateway between Australia and Antarctica at about 40 Ma; this reduced the flow between the Pacific and Indian Oceans, but allowed surface currents to flow around Antarctica. Opening of the deep-water channel between the southern end of South America and Antarctica (today’s Drake Passage) at around 25 Ma resulted in the circumpolar circulation known today. As mentioned above, the Antarctic Circumpolar Current isolates Antarctica from poleward transport of warm water; its initiation is generally considered to have led to significant cooling of Antarctica, and the growth of the southern polar ice cap.

The equatorial gateway between the Atlantic and Pacific Oceans closed relatively recently, at about 3 Ma, when Central America emerged above sea level (so providing a land bridge linking the long-isolated mammalian fauna of North and South America). Equatorial flow would no longer have been continuous from the eastern Atlantic to the western Indian Ocean and closure may well have contributed to cooling in the Northern Hemisphere and the initiation of Arctic glaciation. In fact, cooling of Antarctic waters began as long ago as 60 Ma earlier (Figure 4.17), possibly in response to closure of the ‘Tethyan gateway’ at Gibraltar. Global temperatures roughly stabilised from then until the early Miocene, when the Drake Passage opened and Antarctica became isolated.

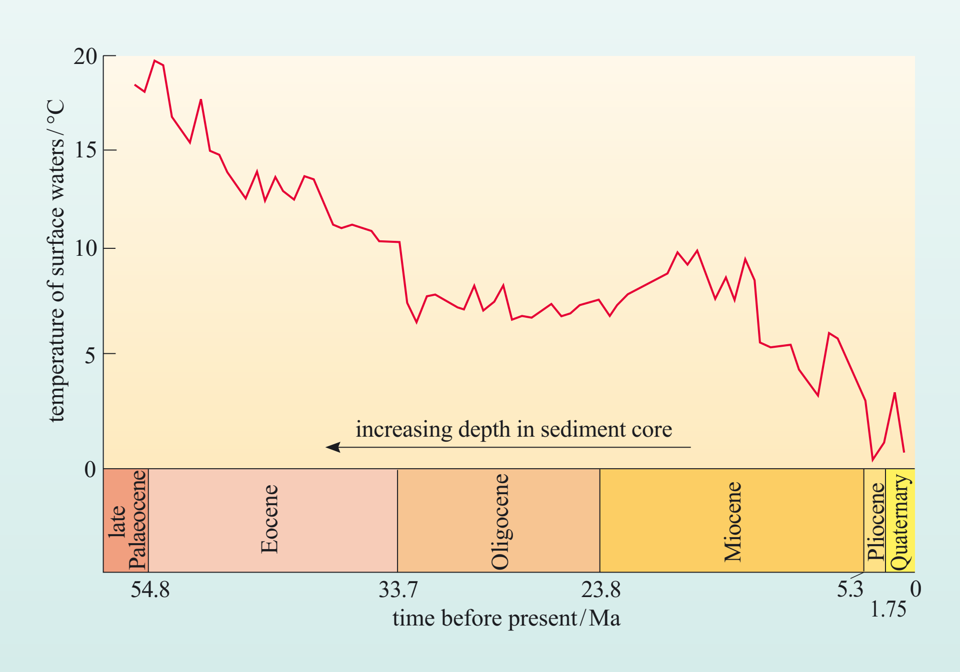


Figure 4.17 Surface water temperatures in the vicinity of Antarctica during the past ~60 Ma, as determined from oxygen isotope ratios of the remains of planktonic foraminiferans. See Box 4.2 for an explanation of the use of oxygen isotopes as a proxy for past sea temperatures.

Box 4.2 Oxygen isotopes and the climate record

Oxygen has three stable isotopes with relative atomic masses 16, 17 and 18. Over 99% of natural oxygen is made up of ^{16}O, with most of the balance being ^{18}O. Water that evaporates from the ocean eventually condenses as cloud and falls as rain or snow. When seawater evaporates from the ocean, water molecules with the lighter oxygen isotope ($H_2{}^{16}O$) evaporate more readily, so atmospheric water vapour is relatively enriched in the lighter isotope. When water vapour condenses and is precipitated back into the ocean, the water containing the heavier isotope ($H_2{}^{18}O$) condenses preferentially. Both processes deplete water vapour in the atmosphere in $H_2{}^{18}O$ relative to $H_2{}^{16}O$. When ^{18}O-depleted water vapour is precipitated as snow in polar regions, the snow will also be depleted in ^{18}O relative to the oceans. The larger the ice caps, the higher the relative proportion of ^{18}O in seawater and the lower the relative proportion of ^{18}O in ice caps.

Marine organisms that form hard parts (shells or skeletons) of calcium carbonate incorporate different proportions of ^{16}O and ^{18}O from dissolved ions in the seawater around them according to the temperature: the lower the temperature, the greater the $^{18}O : {}^{16}O$ ratio in the calcium carbonate secreted. Importantly, all organisms secreting calcium carbonate have higher $^{18}O : {}^{16}O$ ratios in cold than in warm water, though some species shift the isotope ratio by greater amounts than others. This shift in the calcium carbonate $^{18}O : {}^{16}O$ ratio modifies the ratio inherited from seawater, which as described above itself varies according to global ice volume.

The fossils in sediment cores used for oxygen-isotope analysis are usually microorganisms with calcium carbonate shells, often foraminiferans. The amount of ^{18}O in their shells is very small, but it can be measured accurately by mass spectrometry. The result is not given as a simple ratio, but as a delta (δ) value, which is determined by comparison of the sample with a standard, and results in a value expressed in parts per thousand (‰ or 'per mil'):

$$\delta^{18}\text{O} = \frac{(^{18}\text{O}/^{16}\text{O})_{\text{sample}} - (^{18}\text{O}/^{16}\text{O})_{\text{standard}}}{(^{18}\text{O}/^{16}\text{O})_{\text{standard}}} \times 1000 \qquad (4.2)$$

The standard generally used nowadays is Standard Mean Ocean Water (SMOW), or Vienna Standard Mean Ocean Water (VSMOW), which superseded SMOW in 1995. Snowfall in polar regions has δ^{18}O values of –30‰ to –50‰, the negative (or 'light') values indicating depletion of ^{18}O. The higher (or less negative, 'heavier') the measured δ^{18}O value in marine fossils, the greater the enrichment of ^{18}O in seawater and the larger the ice caps on land at the time the organisms were alive.

δ^{18}O values may be determined for the hard parts of both planktonic and benthic species of foraminiferans.

■ Bearing in mind that oxygen-isotope ratios are affected by the temperature of the water in which an organism lived, which would give the most reliable estimates of global ice volumes over the course of glacials and interglacials: planktonic foraminiferans or benthic foraminiferans?

□ Benthic foraminiferans, because the temperature variation of the cold bottom waters is less than that of surface waters.

Useful information about past climate can also be obtained from planktonic foraminiferans, particularly those living at very high latitudes where seasonal temperature variations are quite small. Their δ^{18}O values reflect changes in ice volume and global surface temperature and, of course, when global temperatures are lower, ice caps are larger.

As well as changing ocean circulation, the evolution of global climate since the breakup of Pangea would also have been strongly affected by

- the concentrations of CO_2 in the atmosphere, influenced by the changing rates of subduction through time, which affects volcanic eruption rates
- the intensity and distribution of rainfall and the hydrological cycle.

These factors in turn influence the distribution of living organisms, which affects chemical weathering rates, a major sink for CO_2 over geological timescales (Section 3.3.3). It has been suggested that during the existence of supercontinents such as Pangea, there are relatively low levels of volcanic outgassing delivering less CO_2 to the atmosphere. A 'runaway' cooling effect is prevented largely by the arid conditions within the supercontinent, yielding low chemical weathering rates, and hence a lower rate of removal of CO_2 from the atmosphere. During periods of continental dispersal (as today), there are increased levels of volcanic activity but higher levels of continental weathering due to wetter conditions. You will see in the next chapter that the topography of the continents also plays a significant role.

Climate models can include all these effects but, because of the vast range of timescales involved, models that accurately represent short-term atmospheric circulations, for example, ignore longer-timescale carbon-cycle processes. Similarly, models that represent changes on very long timescales include only an approximation of faster processes such as atmospheric circulation.

4.4.3 Plate tectonics and life

Plate tectonic processes can have more direct effects than simply those of moving continents relative to climatic belts, transporting species across the globe or causing them to become isolated. Over relatively short periods of time they can also alter the relative proportions of *types* of environment available as living space.

In considering how ocean circulation has been affected by changing continental configurations, one aspect of great importance for life, namely upwelling, was omitted.

■ Why is upwelling of particular importance for life? Suggest how it might be possible to deduce where upwellings occurred in the distant past.

■ Upwelling is important for life as it brings nutrient-rich water up into the photic zone where it can support high levels of primary production, i.e. large phytoplankton populations, and hence directly and indirectly, support other organisms. Areas where it might have occurred in the past could be deduced from the ways in which the positions of the continents affect the wind field (Figure 1.28), and from the distribution of organic remains in marine sediments. High productivity results in large fluxes of organic debris to the seabed, some of which may be preserved in the sedimentary sequence.

Areas where coastal upwelling may well have occurred in the Jurassic and mid-Cretaceous are shown in Figure 4.16a–c. Those along the western sides of the 'American' continents would have resulted from the same equatorward winds (the equivalent of the present-day Trade Winds) that drove the currents along the eastern side of the gyres (Figure 1.36).

Although the open oceans support a large plant biomass, the continental shelves support a much higher biomass per unit area (Table 3.1). Figure 4.18 illustrates how the numbers of families in nine phyla of invertebrate shelf-dwelling animals have varied since the beginning of the Cambrian, ~542 Ma. Compare the shape of the curve with the diagrammatic representation of changes in the numbers of continents over the same period. Note that stages D to F correspond to the coming together and breakup of Pangea, shown in Figure 4.16.

■ Is the information in Figure 4.18 consistent with the hypothesis that the global diversity of shelf-dwelling animals is related at least in part to the availability of shores and shallow seas?

Yes. In general, the greater the number of continental fragments, the greater the total length of shoreline available for marine animals to colonise, and the greater the number of separate, isolated environments in which endemic coastal faunas could evolve. The number of families increased from Cambrian to Ordovician times as an ancient supercontinent broke up (A, B), remained more or less constant until Pangea began to be assembled (C, D, E), and decreased dramatically during the Permian, when the total length of shoreline was at a minimum. Thereafter, as Pangea fragmented (F, G), diversity rose sharply once more.

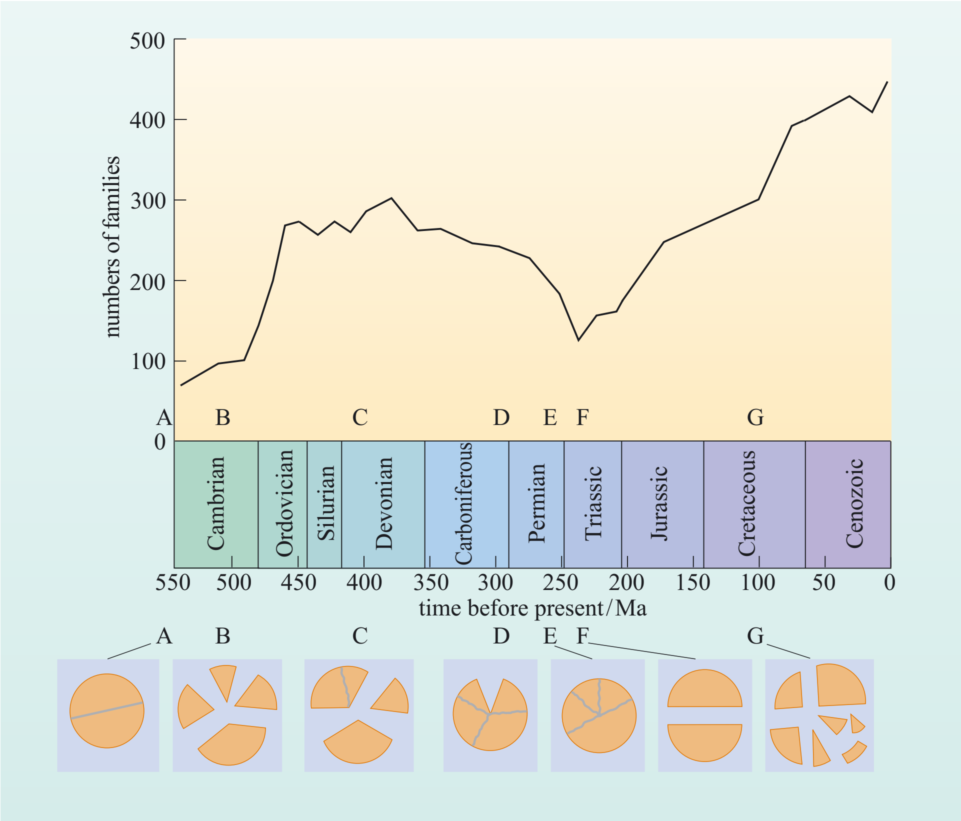


Figure 4.18 Changes in numbers of families in nine phyla of benthic shelf-dwelling invertebrate animals with hard parts, and changes in number of continental fragments (shown diagrammatically), from Cambrian (~542 Ma) to the present. Positions of letters on the graph indicate approximate times when the continents had the numbers of fragments shown (below).

Clearly, the correlation is a fairly general one and shoreline length is only one possible factor in determining diversity of shelf-dwelling organisms. It would be unsound, for example, to attribute the great Permian extinction of species (responsible for the dip at F) mainly to this cause, though some scientists think it could have been a contributory factor. Perhaps the most important factor affecting the distribution and nature of shelf-dwelling organisms is sea level, which is examined in the next section.

4.5 Sea-level changes: causes and consequences

Plate tectonics has a profound effect on sea level. It does this in a myriad of ways, such as changing the sizes and distribution of ocean basins, by promoting uplift of land masses and by impacting on climate and so changing the size of the polar ice caps.

■ Why is sea level related to ice volume?

■ Water evaporated from the Earth's surface and transported polewards may fall as snow at high latitudes, where it can accumulate to form ice caps. Since sea ice displaces its own mass of water, its growth or decline does not alter sea level. By contrast, continental ice sheets such as Antarctica remove water entirely from the oceans as they grow, so sea level falls.

Sea-level changes are made up of two components:

1 **Eustatic changes** of sea level are worldwide changes that affect all oceans and have the potential to cause global climatic changes. Over the past 2 Ma or so, such changes have been caused largely by ocean water becoming frozen into, or melted from, continental ice caps. Over longer timescales, as hinted above, eustatic changes of sea level are caused by other mechanisms (This will be discussed in Section 4.5.1).

2 **Isostatic (or epeirogenic) changes** of sea level are caused by vertical movements of the crust. Such movements may be caused by changes in the thickness and/or density of the lithosphere, and by loading or unloading with ice or sediments. Such changes cause lithosphere to ride higher or lower on the underlying asthenosphere, rather as blocks of wood may float higher or lower in water (Box 4.3).

Box 4.3 Isostasy

Consider a block of wood floating in water (Figure 4.19). Notice the thicker the wooden block, the greater the thickness of wood that emerges above the water. Similarly with an iceberg: the larger it is, the more of it can be seen above the sea-surface. The tendency for the Earth's lithosphere to behave in a similar manner with respect to the underlying asthenosphere is known as **isostasy**.

Continental crust (i.e. the upper part of continental lithosphere) is mostly granitic in composition. Its average thickness is about 40 km, but beneath mountain ranges it can be as much as 90 km thick. Oceanic crust is mostly basaltic in composition with an average thickness of about 7–8 km, and it is denser than continental crust. By analogy with the blocks of wood and icebergs, at isostatic equilibrium, continental lithosphere 'rides' or 'floats' higher on the underlying asthenosphere than oceanic lithosphere, which is why the ocean floors are below sea level.

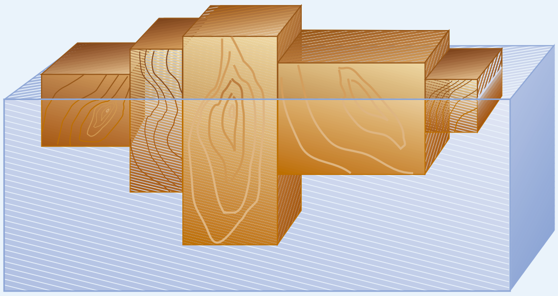


Figure 4.19 Wooden blocks floating in water, an approximate analogy for continental lithosphere of different thicknesses, illustrating why mountains have deep 'roots'.

The analogy with wooden blocks is a simplification because the real lithosphere increases in density with depth, but for the purposes of this discussion you can imagine that the density of the blocks is equivalent to the average density of the lithosphere.

Crustal thickening occurs during mountain-building (through magmatic intrusions and/or convergence of plates at subduction zones). This is why mountains are high. Crustal thinning occurs by stretching during continental breakup, and results in subsidence of the crust that may continue over long periods of time, allowing thick sequences of sediment to accumulate. Loading the lithosphere (by infilling a basin with sediment or by piling ice on top of the crust) will cause its surface to be lowered, and unloading it (by eroding a mountain belt or by melting an ice sheet) will cause it to rise.

Continental shelves and shelf seas

As outlined above, continental crust is thinned by stretching during the breakup of continents, and it therefore subsides. This explains why many continental margins are low lying, with coastal plains bordered by shallow waters (about 200 m deep on average) overlying continental shelves, which in many cases can be extensive enough to be called shelf seas – the North Sea and the Baltic are good examples. Beyond the edge of the continental shelf, where the thinned continental crust ends and oceanic crust begins, water depths increase into the ocean basins up to a few kilometres deep. Substantial thicknesses of sediment can accumulate on continental shelves, deposited from rivers running off the adjacent land. This additional load upon the crust causes it to continue subsiding isostatically, but average water depths remain of the order of a couple of hundred metres.

Continental shelves are important in the context of sea-level changes, because as they typically have a rather flat topography, relatively small rises or falls can flood or expose substantial areas of shelf.

Figure 4.20 compares changes in sea level and (estimated) global temperature over the past 540 Ma. During past glacial periods, sea level was as much as 150 m below its present level whereas during the Cretaceous it was considerably higher than today. The plots of sea level and global temperature change (warming and cooling) correlate reasonably well during the Jurassic, Cretaceous and Tertiary. Earlier in Earth history, the correlation between the two curves is less good, though cooling during the Carboniferous is initially matched by falling sea level.

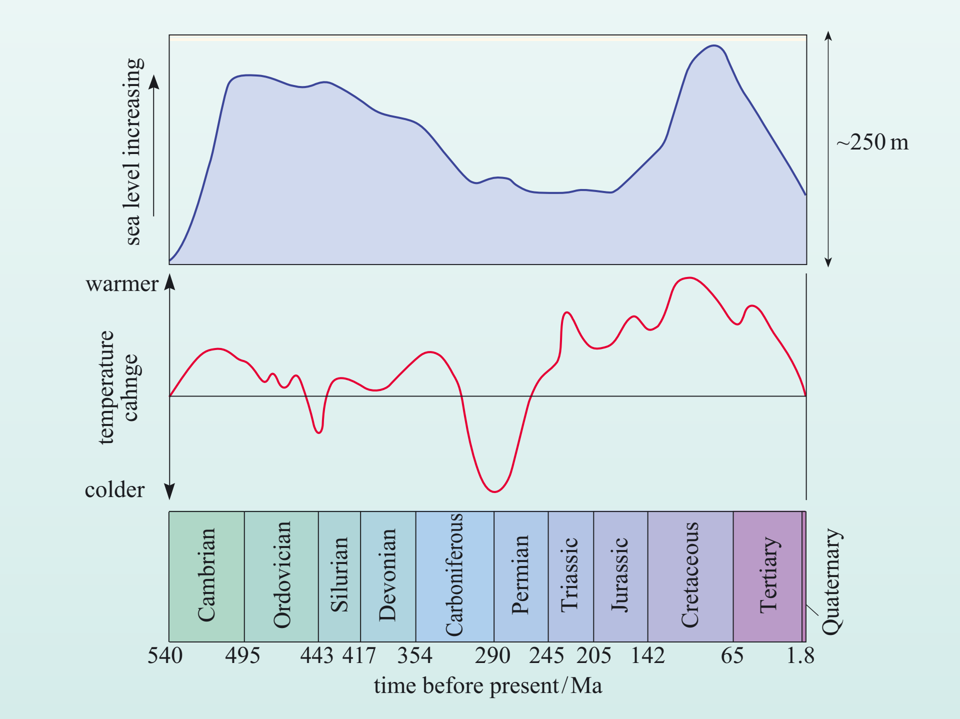


Figure 4.20 Variations in sea level and average global temperature during the past 540 Ma (the Phanerozoic). Sea level has been determined from variations in the extent of shallow-water sediments (limestones, sandstones, mudstones, etc.).

When examining Figure 4.20, you may have wondered whether climate controlled sea level during the Phanerozoic, or whether sea level controlled climate. It is fairly clear that climate has controlled sea level during the last couple of million years of the Phanerozoic, i.e. during the current Ice Age. Global cooling triggered the growth of polar ice sheets which removed large amounts of water from the oceans to the land, resulting in a drop in sea level. Global warming leads to melting of the ice, and sea level rises again. In addition, changes in the global mean temperature of the oceans will increase or decrease the water volume by thermal expansion or contraction. Expansion and contraction cause much smaller sea-level changes than the formation or melting of ice sheets; for example an increase of 10 °C throughout the water column in all oceans would cause a eustatic sea-level rise of about 10 m.

An important aspect of sea-level change, however, is that it has a significant effect on the total area of 'emergent' continental crust, i.e. of continental crust forming land above sea level. Today, 30% of the Earth's surface is emergent continental crust, but during glacial periods this increased to as much as about 35%, whereas in the Cretaceous (Figure 4.20) it was probably as little as about 25%. The reason for such large changes in area above or below sea level is the extensive areas of continental shelf and coastal plains, of which large areas can be flooded or exposed by even relatively small rises or falls of sea level.

Question 4.5

How could changes in relative areas of land and sea either reinforce or counteract climate changes caused by other forcing factors? Think back to discussions in previous sections concerning (a) the opening and closing of oceanic gateways and (b) the Earth's albedo.

Implicit in the answers to these questions is the existence of feedback mechanisms linking sea-level change and climate. The corollary of this is that if sea level were raised or lowered by some process *independent* of climate, then the changing sea level might itself initiate climate change. To explore this aspect, it is important to consider other processes that can affect sea level.

4.5.1 Causes of eustatic sea-level change

There are two basic processes that change global sea level. The first involves changing the volume of water filling the ocean basins, and the second results from changes in shape and size of the ocean basins themselves.

Changes in ocean water volume

The principal cause of changes in ocean water volume is the formation and melting of ice caps. If the entire present-day Antarctic ice sheet melted, global sea level would rise by 60–75 m. Disappearance of the Greenland ice sheet would add about another 7 m. However, over tens of thousands of years, the load of the extra water in the oceans would depress the ocean crust and increase the depth of the ocean basins, so the overall rise would be in the range of 40–50 m. As already mentioned, an increase of 10 °C in the average temperature of water in the oceans would raise sea level by about a further 10 m through expansion of seawater.

Changes in shape and size of ocean basins

It has been proposed that a major cause of rising sea level could be an increase in the rate of formation of oceanic crust at spreading axes, which would increase the volume of ocean ridges, thus displacing water onto the continents. This suggestion needs to be treated with some caution for the following reason. Sea-floor spreading is believed to be the major means of heat loss from the Earth's convecting interior: the rate of heat loss has declined since the Earth formed about 4.6 Ga, so any relatively short-term increases in heat loss would have been superimposed on the extremely long-term decrease. Nonetheless, many scientists do consider the expansion of ridges associated with increased rates of spreading to be a major cause of sea-level rise.

Another explanation for rising sea levels – and a much simpler one – involves simply displacing water from a contracting deep basin into a number of shallow ones. Consider the breakup of Pangea (Figure 4.16a), with new oceans opening between the dispersing continental fragments and the Panthalassa Ocean contracting as the continental fragments disperse. A shrinking deep basin would not become significantly deeper, so water in it would have to be displaced elsewhere. If it could only be displaced into a small number of other basins (newly forming and still shallow) then the net result would be a global rise of sea level. Eventually, deepening of the new ocean basins would become more rapid than shrinkage of the original Panthalassa Ocean, which would approach the dimensions of the present-day Pacific. This probably occurred in the mid- to late Cretaceous (at about 100 Ma) when sea level was at a maximum (Figure 4.16c).

Distributions of shallow-water marine sediments suggest that at that time global sea level could have been ~200 m higher than at present (Figure 4.20), and some estimates are much higher than that. It seems likely that such high sea levels might have been at least partly caused by production of new ocean floor, despite the reservations discussed above concerning spreading ridges.

The oceanic crust can also grow in response to the eruption of enormous amounts of basaltic lavas to form oceanic plateaux (Figure 4.10) and significant volumes of rock were known to be added to the oceanic crust in this way during the Cretaceous.

■ What side-effect of the increase in volcanism during the Cretaceous might have had an additional effect on sea level?

□ The release into the atmosphere of large amounts of carbon dioxide, leading to 'greenhouse' warming.

The rise in sea level in response to greenhouse warming would have been modest, because it would have been due only to thermal expansion of ocean waters. There would have been no melting of ice caps, as there were none present at that time.

Note that although CO_2 would have been *directly* supplied to the atmosphere by eruptions above sea level (e.g. those forming volcanic islands and subduction zone volcanoes), on timescales of 1000 years or so, much of the CO_2 released into the ocean at mid-ocean ridges (via hydrothermal vents and volcanoes) could also escape to the atmosphere. In other words, on geological timescales it may not matter whether the volcanoes are above or below sea level.

Eustatic sea-level rise can also be caused by the deposition of oceanic sediments transported from continental areas because this will effectively decrease the size of the ocean basins by displacing water (though this effect may be offset by isostatic depression of the crust by the weight of the sediment). Sea-level *falls*, on the other hand, could occur during the aggregation of supercontinents (the reverse of the breakup of Pangea, discussed above), when continental collisions result in thickened continental crust and the formation of isostatically elevated mountains (Box 4.3).

4.5.2 Sea level, climate and atmospheric CO_2

Earlier in Section 4.5, it was proposed that changes in sea level might lead to secondary effects which would, in turn, affect climate. For instance, changing albedo is a positive feedback, leading to instability in the climate system, while changing currents due to gateways opening or closing could be a positive or negative feedback. Both these effects are essentially physical in nature, so how do chemical and biological consequences of sea-level change compare? In the following discussion and question, three aspects of the Earth system that are biologically and chemically sensitive to changes in sea level will be examined, all of which exert a strong influence on climate.

The first is continental weathering and its effect on atmospheric CO_2 (Figure 3.21). During periods of low sea level, larger areas of continental crust (notably silicate minerals) are exposed to weathering processes, which will tend to increase the rate of removal of atmospheric CO_2, further reinforcing any tendency towards global cooling. The upper parts of mountains are mainly sites of vigorous *physical* weathering. However, the weathering products are carried away (exposing yet more rock surface to be weathered) and are deposited on the warmer lower slopes and coastal plains, where they are subjected to chemical weathering – particularly if there is abundant vegetation.

When sea levels rise, the area of coastal plains available for (chemical) weathering is reduced, so the rate of removal of atmospheric CO_2 will tend to decrease. Consequently, the concentration of CO_2 in the atmosphere would eventually begin to rise again, favouring global warming. (Note, however, that unless carbon added to the ocean through weathering is preserved in sediments such as carbonaceous deposits and limestones, it will be returned to the atmosphere in 1000 years or so.)

■ When sea-level changes affect climate via continental weathering, is this a positive or a negative feedback mechanism, and what general effect might it be expected to have on the climate system?

■ Considered in isolation, the effect of sea-level change on climate via weathering is one of positive feedback. As discussed in Chapter 1, positive feedback mechanisms lead to instability in the climate system.

Question 4.6 asks you to examine two further mechanisms, both of which involve parts of the biosphere, in a similar way to continental weathering.

Question 4.6

(a) Suggest possible ways in which changing the relative areas of land and sea might indirectly influence climate through resulting changes in:
  (i) the areal extent of environments found in low-lying coastal regions
  (ii) the areal extent of shallow seas.

(b) Explain whether each of the feedback mechanisms you identified in (a) is positive or negative, and hence what general effect it might be expected to have on the climate system.

Of course, in reality no feedback effect resulting indirectly from sea-level change would be acting on its own, and the real situation would be much more complicated than suggested here. Nevertheless, it is interesting that the two negative feedbacks that are best understood are those that involve the activities (i.e. lives and deaths) of organisms, through their effects on the global carbon cycle. This finding might be consistent with a 'Gaian' view of the world (Box 2.4), where the activities of organisms contribute to keeping conditions (in this case temperatures) within a range suitable for life.

Summary of Chapter 4

1 Volcanoes are conduits linking deep mantle processes to atmospheric composition and hence climate; they exhale large amounts of water, CO_2 and SO_2. Over geological timescales, volcanoes have an important role in the carbon cycle, but in the recent past their contribution to atmospheric CO_2 has been overwhelmed by the anthropogenic flux.

2 The radiative effects of major explosive eruptions can be large, reducing the amount of solar radiation reaching the Earth's surface by more than 10%. The effects of most explosive eruptions, however, are short-lived (2–3 years), as aerosols fall out of the stratosphere. Due to the long response time of the Earth's climate system as a whole, brief volcanic events do not have a significant effect on climate.

3 Eruptions of flood basalts such as those of the Columbia River Province may involve effusion of more than 1000 km^3 of sulfur-rich basalt lava over periods of 10–100 years. As these eruptions are sustained over longer periods than great explosive events, their radiative and environmental effects may be more profound. The close coincidence in timing between great episodes of flood basalts in Earth history with major mass extinctions, such as those at the Cretaceous–Tertiary and Palaeozoic–Mesozoic boundaries, suggests that mass extinctions may be linked with the environmental and climatic effects of the flood basalts.

4 Climate models using highly simplified continental configurations suggest that a tropical 'ring world' would be significantly warmer than polar 'cap worlds'. It is thought that the geography of the Earth may have approximated to the former between 700 Ma and 600 Ma, and to the latter (with one polar ice cap) during the late Carboniferous.

5 It is thought that the changing distribution of continents in response to plate-tectonic processes affects global climate on a million-year timescale, through its effect on the radiation budget, and hence indirectly on the hydrological cycle and weathering (which affects the CO_2 concentration of the atmosphere). Furthermore, the closing and opening of oceanic gateways as continents change their relative positions has a significant effect on shallow and deep oceanic circulation, and can contribute to global warming and cooling on significantly shorter timescales.

6 The breakup of Pangea began at about 200 Ma. The opening of a low-latitude seaway may have contributed to global warming in the Cretaceous, at about 100 Ma. Around this time, deep water masses were probably warm and very saline, having formed at low latitudes. It is thought that thermal isolation resulting from the initiation of the Antarctic Circumpolar Current at 25 Ma accelerated cooling of Antarctica and the growth of the south polar ice cap.

7 Relative sea-level change results from a combination of eustatic and isostatic sea-level change. Eustatic changes are global in extent, whereas isostatic changes result from local or regional uplift or subsidence of the lithosphere. Eustatic sea-level changes are due either to changes in the volume of ocean waters (resulting from the formation and melting of ice sheets), or to changes in the size and shape of the ocean basins (resulting either from the formation of new oceanic crust, notably as submarine plateaux, or from the replacement of a few large ocean basins by a number of smaller ones). During periods of global warming, some sea-level rise is attributable to expansion of the ocean water; however, an increase of 10 °C throughout the water column in all oceans would lead to a sea-level rise of only ~10 m.

8 Global warming during the Cretaceous may have been related to the addition to the atmosphere of huge amounts of CO_2 as a result of volcanism (as flood basalts and as a consequence of increased rates of sea-floor spreading). However, the warming was counteracted by the removal of atmospheric CO_2, through deposition and preservation of carbon in the ocean.

Learning outcomes for Chapter 4

You should now be able to demonstrate a knowledge and understanding of:

4.1 The influence and importance of different volcanic gases on short- and long-term climate change, and the physical and chemical factors that determine the extent of any such change.

4.2 The potential link between the eruption of flood basalts and mass extinctions at specific points throughout geological time.

4.3 The mechanisms by which the changing positions and configurations of the continents and oceans have affected global climate over time, as a result of changing circulatory systems, albedo and weathering patterns.

4.4 The positive and negative feedback mechanisms that connect changes in the global temperature and/or eustatic and isostatic sea levels, with preservation of carbon in atmosphere, oceans, biosphere and geosphere.

Answers to questions

Question 1.1

(a) (i) At 50° N, the incoming solar radiation on 21 March (the spring equinox) is about 10^7 J m^{-2}. It increases to a maximum of 2.3–2.4 × 10^7 J m^{-2} in late June (around the time of the summer solstice/longest day), then declines to about 0.3 × 10^7 J m^{-2} towards the end of December (around the time of the winter solstice/shortest day), after which it begins to rise again.

(ii) Watts are joules per second; therefore, to convert J m^{-2} (incoming solar energy per unit area per day) to W m^{-2} (average solar power) you must divide the contour values by the number of seconds in a day, i.e. 8.64 × 10^4 s. The contour values would therefore range from:

$$\frac{0.4\times10^{7}\ \text{J m}^{-2}}{8.64\times10^{4}\ \text{s}}\ \text{to}\ \frac{2.5\times10^{7}\ \text{J m}^{-2}}{8.64\times10^{4}\ \text{s}},\ \text{or} \sim 46\ \text{to}\ 289\ \text{J s}^{-1}\ \text{m}^{-2}$$

which is 46–290 W m^{-2} (to 2 sig. figs).

(b) Mid-latitudes (between ~30° N and S) receive the most solar radiation at any one time: $> 2.5 \times 10^7$ J m^{-2} in summer. This is because of the long days at this time of year, when the noonday Sun is high.

Question 1.2

(a) (i) Approximately 10–20% of visible radiation is absorbed in the atmosphere, with absorption at the red end of the spectrum being greater than that at the blue/violet end. The gas responsible is ozone.

(ii) About half of incoming infrared radiation is absorbed by the atmosphere. The gases mainly responsible in this case are water vapour and carbon dioxide.

(b) Outgoing longwave radiation is absorbed mainly by water vapour and carbon dioxide, with lesser absorption by methane, nitrous oxide and ozone.

Question 1.3

(a) The main and most obvious reason is that more solar radiation reaches the Earth's surface at low latitudes than at high latitudes, for reasons demonstrated by Figure 1.4. The second reason is that, of the solar radiation that reaches the Earth's surface, much more is reflected at high latitudes than at low latitudes, particularly because of the high albedos of ice and snow (Table 1.1).

(b) The Earth–atmosphere system has a net gain of heat between about 40° S and 35° N (Figure 1.45 overleaf); poleward of those latitudes, there is a net loss of heat.

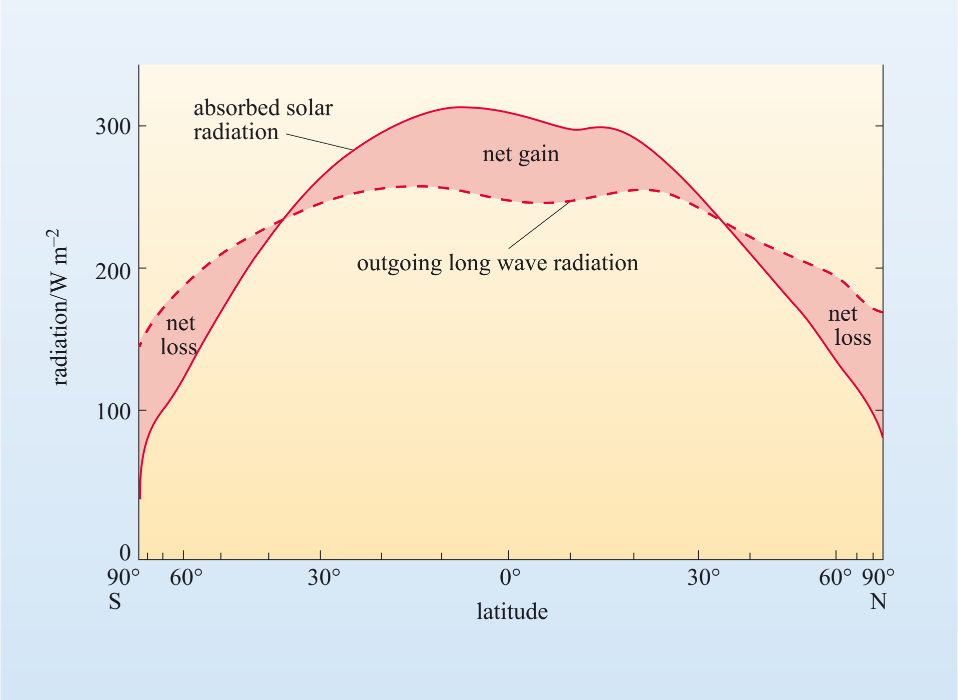


Figure 1.45 The variation with latitude of the solar radiation absorbed by the Earth–atmosphere system (solid curve) and the outgoing radiation lost to space (dashed curve). Values are averaged over the year, and are scaled according to the area of the Earth's surface in different latitude bands.

Question 1.4

If during the Little Ice Age average global temperatures were about 1 °C lower, and this were attributable entirely to a decrease in solar luminosity, there would have to have been a reduction in incoming solar radiation of about:

$$\frac{1\ ^{\circ}\mathrm{C}}{0.75\ ^{\circ}\mathrm{C\ per\ W\ m^{-2}}} \approx 1.3\ \mathrm{W\ m^{-2}}$$

because the climate sensitivity is given as the change in temperature per unit change in forcing – thus the change in forcing is the change in temperature divided by the sensitivity.

Expressed as a percentage of the present-day average value:

$$\frac{1.3\ \mathrm{W\ m^{-2}}}{343\ \mathrm{W\ m^{-2}}} \times 100 = 0.389\% \approx 0.4\% \text{ to 1 sig. fig.}$$

Question 1.5

(a) Generally, winds blow from regions of high pressure to regions of low pressure. This is most clearly seen in the case of the Trade Winds blowing towards the zone of low surface pressure along the Equator. In addition, winds blow clockwise around high-pressure regions in the Northern Hemisphere and anticlockwise around them in the Southern Hemisphere; such flow is referred to as anticyclonic. Winds blow in the opposite direction (i.e. are cyclonic) around low-pressure regions.

(b) In regions of high surface pressure, vertical air motion will result in the air sinking whereas in regions of low pressure the air will be rising.

Question 1.6

(a) The ITCZ is (mostly) further north in July and further south in January; in other words, it moves into the hemisphere experiencing summer, with this being particularly prevalent over large landmasses. This would be expected because the ITCZ, which is an area of vigorous upward convection, tends to be located over the warmest parts of the Earth's surface and, as discussed in connection with Figure 1.14, continental masses heat up much faster than the oceans in summer and cool down much faster than the oceans in winter.

(b) In July, when the ITCZ is at its most northerly position, the winds over tropical West Africa are mainly southwesterly (i.e. from the southwest). In January, when the ITCZ is at its most southerly position, the winds over the region are mainly easterly or northeasterly (i.e. from the east or northeast).

(c) The north–south shift in the position of the ITCZ is most marked over the northern Indian Ocean and over the western tropical Pacific and South-East Asia.

Question 1.7

The value needed from Box 1.4 is the specific heat of water, $4.18 \times 10^3\ \mathrm{J\ kg^{-1}\ {}^\circ C^{-1}}$. $1300\ \mathrm{km}^3$ is $1300 \times 10^9\ \mathrm{m}^3$, which has a mass of 1300×10^{12} kg. The total amount of heat given up each day, on average, is:

$$\begin{aligned}\text{heat given up each day} &= \text{specific heat} \times \text{mass of water} \times \text{temperature drop}\\ &= (4.18 \times 10^3\ \mathrm{J\ kg^{-1}\ {}^\circ C^{-1}}) \times (1300 \times 10^{12}\ \mathrm{kg}) \times 11\ {}^\circ\mathrm{C}\\ &= 5.9774 \times 10^{19}\ \mathrm{J} = 6.0 \times 10^{19}\ \mathrm{J}\ \text{to two sig. figs.}\end{aligned}$$

Therefore, $\sim 6 \times 10^{19}$ Joules of heat are released to the atmosphere over the northern North Atlantic every day.

Using the same conversion as in Question 1.1 ($1\ \mathrm{J\ s^{-1}} = 1\ \mathrm{W}$), $6 \times 10^{19}\ \mathrm{J\ day^{-1}}$ is about 0.7×10^{15} W, which is equivalent to the heat output of about half a million large power stations. In the winter months, the heat lost to the atmosphere would be even greater than this average value.

Question 1.8

(a) The percentage of the Earth's water presently in the ocean:

$$= \frac{1\,322\,000 \times 10^{15}\ \mathrm{kg}}{1\,360\,000 \times 10^{15}\ \mathrm{kg}} \times 100 \approx 97\%.$$

The percentage of the Earth's water presently in ice caps and glaciers:

$$= \frac{29\,300 \times 10^{15}\ \mathrm{kg}}{1\,360\,000 \times 10^{15}\ \mathrm{kg}} \times 100 \approx 2\%.$$

(b) Even if all the ice caps and glaciers melted, the volume of water in the oceans would only increase by about 2%, so if a quarter of them melted it would increase by 0.5%.

This may not sound much, but it would mean a rise in sea level of about 15 m. Given that a large proportion of human habitations are concentrated around the coasts, such a sea-level rise would be disastrous.

Question 1.9

(a) Figure 1.6 shows that in December only low levels of solar radiation reach high latitudes in the Northern Hemisphere, which is insufficient to allow phytoplankton populations to grow. By May, light levels have risen considerably in the Northern Hemisphere – and the image in Figure 1.43a illustrates the burst in primary productivity, known as the 'spring bloom'.

(b) This region is the centre of the Atlantic subtropical gyre, driven by anticyclonic winds. As discussed in the text in connection with Figure 1.36, such gyres are regions of downwelling, and hence can only support limited primary productivity.

By contrast, the cyclonic subpolar gyres are regions of divergence and upwelling (Figure 1.36b(ii)), which is another reason why primary productivity can be high there once light levels rise in spring.

Question 1.10

Increased primary productivity would mean bigger plankton blooms and a greater flux of dimethyl sulfide to the atmosphere. This in turn could lead to more cloud formation, resulting in an increase in the Earth's albedo.

Question 2.1

(a) The lunar cratering record suggests that 4.0 Ga ago, the Earth was experiencing a storm of impacts that may have repeatedly boiled off any incipient oceans. It would be ill-advised to leave behind the thermally insulated capsule on this trip, and indeed your time-travel agent might well ask you if you had some alternative destinations in mind. If you risked a visit and succeeded in spotting any standing water through the window of your capsule you might find it fringed with greenish slime.

(b) By contrast, you could probably alight from your capsule on the trip back to 2.5 Ga, but not without the breathing apparatus because of the scarce molecular oxygen in the atmosphere, and you would need lashings of suncream because of the fierce ultraviolet radiation let in by the lack of an effective ozone screen. Your sandwiches would be preferable to the slimy greenish-grey mats you might encounter along the coast, and the book could be handy unless you really like looking at barren landscapes.

(c) Breathing might be manageable, if somewhat strenuous, on the trip back to 500 Ma, so you could possibly leave the suncream behind now as an effective ozone layer has been developed at this stage. Unless you took some sandwiches, you would need the crabbing net to catch your lunch at the seaside (but with no guarantees as to what it would taste like). Meanwhile the book would still be useful, as you would have to wait another 100 Ma or so for the appearance of complex and visually interesting life-forms on land.

(d) Top on the list for the trip back to 100 Ma would be your weapons to fend off dinosaurs, if not to help you catch your lunch.

(e) The weapons would again prove useful for your trip back to 100 000 years, the threat now being from large mammals or aggressive earlier members of your own species, resenting your presence.

Question 2.2

The long-term burial of organic material allowed molecular oxygen to build up in the atmosphere. If all the material produced through photosynthesis remained available for respiration (or combustion), it would have consumed the same amount of oxygen as had originally been released during its production. With the removal of such organic material, however, photosynthesis could yield a net excess of oxygen.

Question 2.3

The genetic material (DNA) in a eukaryote cell is contained in a nucleus, whereas the DNA in a prokaryote cell is confined to a single loop within the cell.

Question 2.4

The diagnostic features most likely to be detected are:

- cell size (eukaryotes tend to be larger than prokaryotes)
- presence of organelles
- cellular organisation (differentiated cells in multicellular forms).

Other molecular (e.g. genetic) attributes are unlikely to be preserved.

Question 2.5

The ability to engulf other cells, which is the first step towards endosymbiosis, would require loss of the constraining rigid prokaryotic cell wall.

Question 2.6

The most likely time for mitochondria-bearing eukaryotes to proliferate is in the early Proterozoic, when the deposition of banded iron formations began to decline and red beds started to appear in the rock record , e.g. around 2.0 Ga. These changes in the sedimentary record are thought to reflect the appearance of molecular oxygen in the surface waters of the oceans.

Question 2.7

Single mutations whose effects were hidden by an unaffected partner would still accumulate through the generations, increasing the eventual probability of double mutations at matching sites on the paired chromosomes. (To return to the aeroplane analogy, this is rather like the risk of continuing to make further flights in a twin-engined aeroplane with only one functioning engine.)

Question 2.8

The observation implies that sexual reproduction evolved only after there was diversification between the asexual forms and the new sexual ones from a common ancestral group. It must have arisen in the common ancestor of the remaining eukaryote groups. The main burst of eukaryote evolutionary history may, therefore, have been triggered by the appearance of sexual reproduction.

Question 2.9

The molecular clock theory supposes a constant average rate of divergence between corresponding sequences. Therefore extrapolating a straight line in Figure 2.13 to the maximum divergence value known among animals (~190%) allows the latest time of the origin of the animals to be estimated at around 1.0 Ga.

Question 2.10

No, because as discussed above, the focus of natural selection is on the individual genetic entity, which is the only *necessary* beneficiary of any adaptations. What happens above that level (e.g. at the population level) as a consequence of adaptations of individuals is dependent on whether changes are either beneficial or disadvantageous to the maintenance of the system under consideration.

Question 3.1

(a) In Equation 1.2 (photosynthesis), oxygen is being liberated by breaking C—O bonds in CO_2. In addition the carbon in CO_2 is linked to two oxygen atoms, whereas in organic matter (glucose in Equation 1.2) each carbon is linked to other carbon atoms and hydrogen atoms as well as oxygen atoms. This reaction is therefore a reducing reaction.

(b) In respiration (Equation 3.1) the reverse process is going on, in which the C—C and C—H bonds in glucose are being broken and the hydrogen and carbon are combining with oxygen to produce CO_2 and water. This is therefore an oxidising reaction. (However, you should note that the oxygen molecules are being broken up and recombined with C and H so as far as the oxygen is concerned, the reaction is reducing.)

Bonds with oxygen are very stable and require a large amount of energy to break them so photosynthesis requires an input of energy from the Sun to take place – it is an endothermic reaction. By contrast the oxidation of glucose releases energy because C—C and C—H bonds can be broken relatively easily. The reaction is therefore exothermic.

Question 3.2

(a) According to Figure 3.3, the amount of carbon in terrestrial plant material is about 560×10^{12} kgC and the net fixation of carbon from the atmosphere (i.e. the net annual input of carbon to the reservoir of living plants) is $60 \times 10^{12}\ \text{kgC y}^{-1}$. The mean residence time for carbon in plant biomass is therefore:

$$\frac{560 \times 10^{12}\ \text{kgC y}^{-1}}{60 \times 10^{12}\ \text{kgC y}^{-1}} \approx 9 \text{ years to 1 sig. fig.}$$

The amount of carbon deposited in soil is also $60 \times 10^{12}\ \text{kgC y}^{-1}$, while the amount of carbon in the soil reservoir at any one time is about

1500×10^{12} kgC (Figure 3.3). The mean residence time for carbon in soil organic matter is therefore:

$$\frac{1500 \times 10^{12}\ \text{kgC}}{60 \times 10^{12}\ \text{kgC y}^{-1}} = 25\ \text{years.}$$

(b) You could expect short residence times because small numbers divided by large numbers give even smaller numbers.

Question 3.3

(a) (i) The annual flux of carbon into tropical rainforests is the total amount of carbon fixed in them annually, i.e. their total net primary productivity, which is $(17.0 \times 10^6\ \text{km}^2) \times (0.9\ \text{kgC m}^{-2}\ \text{y}^{-1}) = 15.3 \times 10^{12}\ \text{kgC y}^{-1}$. (Having now calculated this value, you can complete the last column of Table 3.1.)

(ii) The percentage of total global net primary productivity contributed by the rainforest is:

$$\frac{15.3 \times 10^{12}\ \text{kgC y}^{-1}}{48.3 \times 10^{12}\ \text{kgC y}^{-1}} \times 100 = 31.68\% = 31.7\%\ \text{to 3 sig. figs.}$$

As you will probably have noticed in working out your answer, this high percentage is partly due to the large area of the globe occupied, even now, by tropical rainforest. (Note, however, that in doing this calculation, it is assumed that the area of the globe covered by rainforests remains constant, whereas in reality it is decreasing.)

(b) (i) To make a comparison area for area, the column headed 'Mean NPP per unit area' must be used. The value for tropical evergreen forests is $0.900\ \text{kgC m}^{-2}\ \text{y}^{-1}$, and that for boreal forests is $0.360\ \text{kgC m}^{-2}\ \text{y}^{-1}$, so tropical evergreen forests are:

$$\frac{0.900\ \text{kgC m}^{-2}\ \text{y}^{-1}}{0.360\ \text{kgC m}^{-2}\ \text{y}^{-1}} = 2.5\ \text{times more productive than boreal forests.}$$

(ii) One factor to be considered is rainfall. Tropical forests grow at low latitudes where there is heavy rainfall associated with the ITCZ. However, boreal forests grow in the subpolar regions where precipitation is also fairly high, so this cannot be the main reason. Another important difference between the two areas is the amount of light they receive during the year: tropical regions have high daytime light levels all year, but subpolar regions have very low light levels (or are dark) for much of the year (Figure 1.9) and hence have a very short growing season. (*Note*: because of length of the growing season, even temperate savannah and tall grassland are more productive than boreal woodland and boreal forest.) The most important reason, however, is that growth rates are greatly affected by temperature. At the present time, this is the factor that limits terrestrial primary productivity at high latitudes (see Section 2.3.2).

(iii) Residence time in living plant material is given by the total plant biomass divided by the total net primary productivity. Therefore, according to Table 3.1, the residence time of carbon in swamps and marshes is:

$$\frac{13.6\times10^{12}\ \text{kgC}}{2.2\times10^{12}\ \text{kgC y}^{-1}} = 6.18\ \text{y} \approx 6\ \text{years.}$$

In boreal forests the residence time is:

$$\frac{108\times10^{12}\ \text{kgC}}{4.3\times10^{12}\ \text{kgC y}^{-1}} \approx 25\ \text{years}$$

which is effectively the average lifetime of a tree. The residence time in boreal forests is approximately 4 times that in swamps and marshes. (Again, for this question, it is assumed that the areal extents of swamps and marshes, and of boreal forests, remain constant.)

Question 3.4

(a) The reverse reaction for photosynthesis is that for respiration (Equation 3.1).

(b) The reverse reaction for carbonate weathering is carbonate precipitation (Equation 3.8).

Question 3.5

(a) The values on the right-hand axis are all negative, and as explained in the caption, negative values correspond to CO_2 concentrations in the atmosphere being higher than those in surface waters. This means that the situation is out of equilibrium, with a concentration gradient across the air–sea interface and a net flux of CO_2 *into* the ocean (Figure 3.8).

(b) As it was the time of the spring bloom, the phytoplankton were multiplying and fixing carbon (i.e. there was high net primary productivity). Where there were more phytoplankton in the surface water (i.e. chlorophyll concentrations were high), more carbon was being fixed, causing more CO_2 gas to enter surface waters from the atmosphere; where there were fewer phytoplankton, the reverse was true.

(As you may have realised, the phytoplankton – and indeed zooplankton feeding on them – would also have been respiring, releasing CO_2 into the water; but at times of high primary productivity, sufficient organic debris would be falling out of the surface layers to drive a net flux of CO_2 into the ocean.)

Question 3.6

(a) In the northeastern North Atlantic (as well as at high southern latitudes), the net flux of CO_2 is into the ocean. This is a region where deep-water formation is occurring as a result of intense cooling of surface water (and brine-rejection) (Figure 1.33). Cold water can take up a relatively large amount of CO_2 before becoming saturated and on sinking down from the surface would allow yet more CO_2 to dissolve.

(b) At both seasons of the year there is a flux of CO_2 into the ocean in the central North Atlantic, partly because surface waters converge and sink there (Figure 1.37). By April–June, however, the water is being warmed and so the region where there is a net flux of CO_2 out of surface water is beginning to spread northwards. In addition, at this time of year there is high net primary productivity in the northern part of the ocean (Figure 1.43a), which also contributes to a net flux of CO_2 from air to sea.

(c) In low latitudes, water is upwelling to the surface and warming. Both the decrease in pressure and the increase in temperature cause CO_2 to come out of solution, so that there is a net flux from sea to air (despite any high primary productivity along the Equator that would tend to draw CO_2 down into the ocean).

Question 3.7

(a) According to Table 3.2, the total mass of carbon in marine plant material is 1.76×10^{12} kgC, i.e. about 2×10^{12} kgC. This is a small percentage of the standing stock of biomass on land, whether the value of 560×10^{12} kgC (from Figure 3.3) or 827×10^{12} kgC (from Table 3.1) is used. In the first case, the answer is ~0.36%, in the second it is ~0.24%.

(b) If an average of the input and output fluxes shown in Figure 3.19 is used, the residence time for carbon in living phytoplankton is:

$$\frac{2 \times 10^{12}\ \text{kgC}}{38 \times 10^{12}\ \text{kgC y}^{-1}} \approx 0.05 \text{ years or about 19 days.}$$

By contrast, the residence time of carbon in the terrestrial biomass reservoir is about nine years (Question 3.2a), so carbon cycles much faster through the marine biomass reservoir than through the terrestrial biomass reservoir.

Question 3.8

The statement is only partly true. Both limestone and shells/skeletal remains (that eventually become limestones) are forms of inorganic carbon. They are precipitated from dissolved inorganic carbon in seawater (mainly HCO_3^-), and are not composed of large organic molecules made up of carbon, hydrogen and oxygen.

Question 3.9

(a) Your answer may have included any two of the following ways in which greenhouse warming could indirectly affect the global carbon cycle.

If the Earth's surface became warmer, evaporation of water (from both the sea-surface and land) would increase, there would be more water in the atmosphere, and so precipitation would increase; in other words, the hydrological cycle would become more active. (To some extent, this is being seen at the present time, in the form of more droughts at lower latitudes and increased rainfall in subpolar latitudes.)

As a result of warmer, wetter conditions, terrestrial biological productivity would increase (consider how you encourage plants to grow). This would fix more carbon from the atmosphere, supply more to soil in organic debris and perhaps increase the rate at which carbon is preserved on land (e.g. in swamps). Marine productivity might also increase as a result of higher sea-surface temperatures (though other factors might counteract this). (There is also some (inconclusive) evidence that an increase in the concentration of atmospheric CO_2 increases net primary productivity.)

Warmer, wetter conditions would also increase rates of chemical weathering, particularly if plant growth were increased. Increased weathering of silicates would result in an increase of the net flux of carbon to the ocean (Figure 3.21).

(b) Increased weathering of silicates and increased primary production on land and in the sea both result in increased rates of sedimentation of calcareous and organic sediments in the ocean. However, although an increased flux of organic-rich material is likely to result in more rapid burial and preservation of carbon, the same cannot necessarily be said for inorganic carbon. This is because an increase in the concentration of CO_2 in the atmosphere and increased primary productivity would both lead to an increase of dissolved inorganic carbon in the deep ocean, and hence an increase in the acidity of deep waters and an increase in the rate of dissolution of calcareous remains (i.e. a deeper carbonate compensation depth).

You were not expected to think of the last aspect; it has been included to illustrate the complexity of the global carbon cycle and how difficult it is to make predictions about the results of changing fluxes in any part of the cycle.

Question 3.10

(a) The fluctuations in atmospheric CO_2 concentration are a result of the uptake of CO_2 by plants during photosynthesis in spring and summer, i.e. removal of carbon from the atmosphere and its fixation in living plant material. (Note that it is the lows that correspond to spring and summer, and the highs that correspond to winter.)

(b) It is a very evocative description, but rather misleading. Breathing – effectively *respiration* – involves the uptake of oxygen and the release of carbon dioxide (Equation 3.2). Respiration of biomass continues all year round (though more so in the spring and summer), but high rates of primary production and hence *net* primary productivity (production minus respiration) *only* occur in spring and summer. Thus (as discussed in (a)), the pattern is primarily a manifestation of photosynthesis rather than of respiration.

(c) The pattern is dampened in the Southern Hemisphere because primary productivity per unit area in the ocean is much less than on land (e.g. compare the columns for 'Mean NPP per unit area' in Tables 3.1 and 3.2), and the Southern Hemisphere is largely ocean.

Question 4.1

(a) As shown in Figure 1.20, the tropospheric wind systems of the two hemispheres are to a large extent separate, meeting at the Intertropical Convergence Zone (ITCZ). For example, air being carried northwards across the Equator by the Trade Winds is likely to rise at the ITCZ and then be carried southwards again, rather than be carried northwards to the Northern Hemisphere.

(b) As shown schematically in Figure 1.20, the tropopause (i.e. the boundary between the troposphere and the stratosphere) is higher at low latitudes than at high latitudes. As discussed in the caption for Figure 1.19, it is generally at a height of about 17 km in low latitudes but only 8–10 km high at high latitudes. As a result, a volcano is more likely to be able to send gases (and aerosols) into the stratosphere at high latitudes than at low latitudes.

Question 4.2

If the gases produced by an eruption do not get into the stratosphere, any aerosols are formed in the troposphere and are soon washed out. Even the biggest explosive volcanic eruptions – such as that of Mount Pinatubo in 1991 – are short-lived, and their aerosols remain in the stratosphere for only a year or two. Thus, although their short-term effects may be profound, the climate forcing they cause is not sustained for sufficiently long for the climate system to respond. For this to happen, feedback loops (e.g. involving growth of vegetation or spread of ice sheets) have to have time to become established, over periods of tens or hundreds of years.

Flood basalt eruptions, however, may be exceptional in that they involve the steady effusion of sulfur-rich magmas over periods of decades or more.

Question 4.3

Table 4.3 Answer to Question 4.3.

Flood basalt or large igneous province	Mean age (Ma)	± (Ma)	Stage or epoch boundary at which extinction occurred	Age (Ma)	± (Ma)
Emeishan Traps	259	3	Guadalupian–Lopingian (M)	260.4	0.7
Siberian Traps	250	1	Permian–Triassic (M)	251	0.4
Central Atlantic Magmatic Province	201	1	Triassic–Jurassic (M)	199.6	0.6
Karoo Ferrar	183	2	Pliensbachian–Toarcian (E)	183	1.5
			Bajocian–Bathonian	168	3.5
			Tithonian–Berriasian (E)	145	4
Parana Etendeka	133	1	Valanginian–Hauterivian	136.4	2
Ontong Java 1	122	1	Early Aptian (E)	125	1.0
Rajmahal/Kerguelan	118	1			
Ontong Java 2	90	1	Cenomanian–Turonian (E)	93.5	0.8
Caribbean Plateau	89	1			
Madagascar Traps	88	1			
Deccan Traps	65.5	0.5	Cretaceous–Palaeogene (M)	65.5	0.3
North Atlantic 1	61	2			
North Atlantic 2	56	1	Paleocene–Eocene (M)	55.8	0.2
Ethiopia and Yemen	30	1	Oi1 event	30	2.5
Columbia River	16	1	Early Miocene–Mid Miocene (E)	16.0	0.1
			Serravallian–Tortonian	11.6	0.3
			Pliocene–Pleistocene (E)	1.81	0.02

The eight extinction events and continental flood basalt–large igneous province (CFB–LIP) ages highlighted in red and emboldened show a high degree of correlation in that their ages are within the uncertainties of each other. Other CFB–LIPs are not associated with any extinction events and some extinctions are not closely associated with known CFB–LIPs. These coincidences suggest a link between basalt eruptions and some extinctions but, clearly, extinctions can have other causes.

Question 4.4

(a) Given that they occur at intervals of hundreds of millions of years (and the whole globe is affected), Ice Ages would plot in the upper part of the right-hand bubble (Figure 4.22).

(b) If major ocean basins open and close on timescales of 100–200 millions of years, then the timescale on which the disposition of continents over the surface of the globe changes significantly is of the same order as the timescale at which Ice Ages come and go.

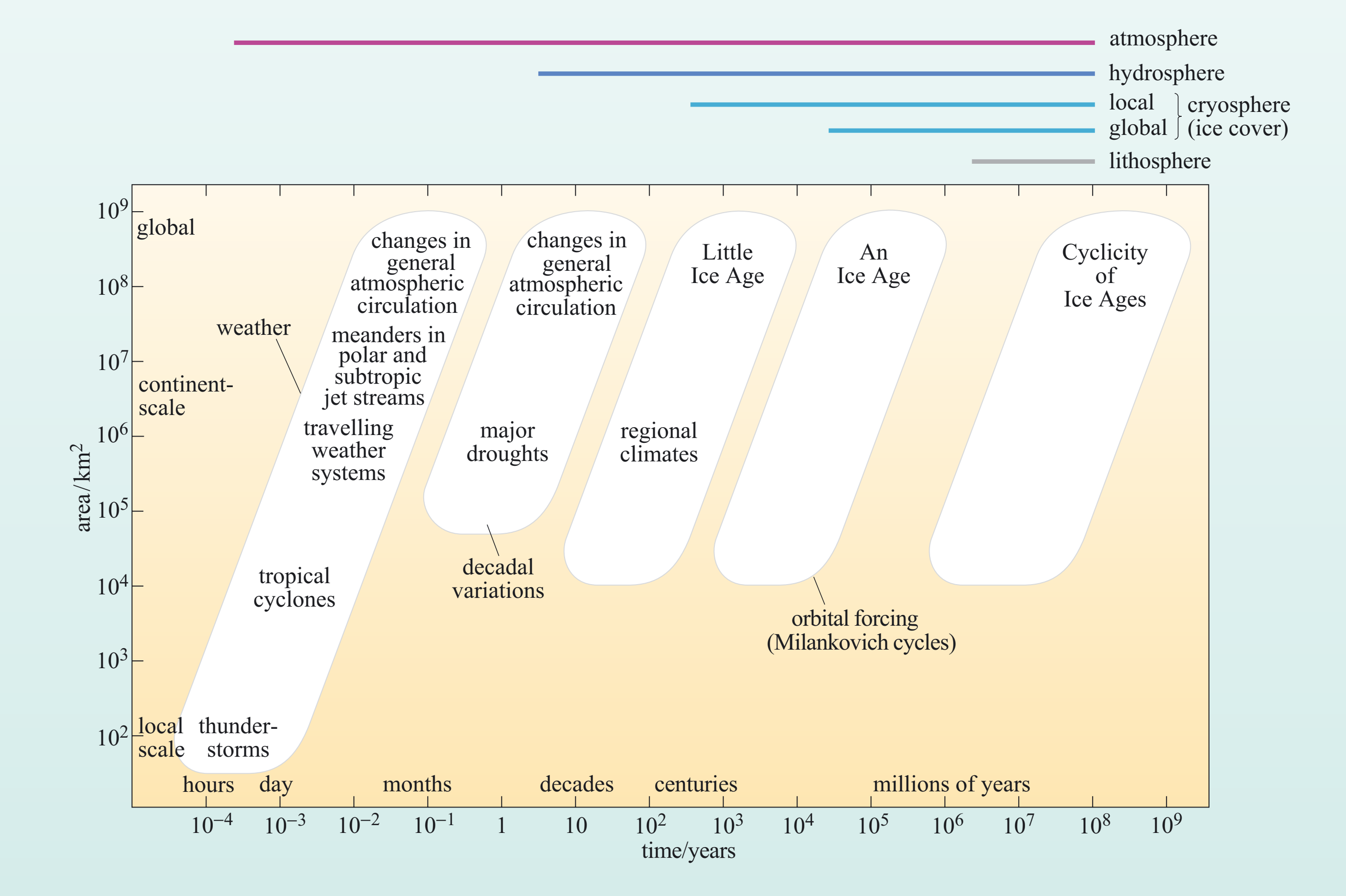


Figure 4.22 Answer to Question 4.4a.

Question 4.5

(a) Changing sea levels may open or close oceanic gateways to surface (or even deep) currents. This could influence how much heat is carried poleward and thus affect the climate as a whole.

(b) Generally speaking, during periods of high sea level, the Earth's average albedo is reduced, as sea-surfaces generally have a lower albedo than land. More solar radiation is absorbed, which would reinforce any global warming trends. Conversely, falling sea level exposes more land that has a higher-albedo and would encourage global cooling.

Note, however, that this argument is somewhat oversimplified, as both land and (especially) sea have a greater albedo for low sun elevations (i.e. at high latitudes).

Question 4.6

(a) (i) As continental shelves have a flat topography, relatively small changes in sea level can result in relatively large areas being flooded or exposed. Assuming sea level is starting from a low level, a rise in sea level would initially result in a significant increase in the area of low-lying, swampy coastal land, where organic remains (e.g. peat and coal) could be preserved. As discussed in Chapter 3, preserved organic carbon is a long-term sink for carbon, which would remove carbon dioxide from the atmosphere on geological timescales. Thus, an increase in sea level could lead to a decrease in the concentration of atmospheric carbon dioxide and global cooling.

Conversely, falling sea levels could expose organic remains to oxidation and erosion, returning carbon dioxide to the atmosphere and encouraging global warming.

(ii) A rise in sea level would increase the area of shallow seas, where shallow-water carbonate-secreting organisms (such as certain algae, bivalves and corals) could flourish. Accumulation of these carbonate remains represents a long-term sink for carbon, removing CO_2 from the atmosphere on geological timescales and leading to global cooling.

Carbonate rocks exposed by falling sea levels would be subject to oxidation and erosion in the same way as organic carbon and with the same result: release of carbon dioxide to the atmosphere and subsequent global warming. On long (geological) timescales, the organic carbon and carbonates exposed by a fall in sea level would not necessarily be the same ones that accumulated in the previous period of high sea level, as much tectonic activity might have occurred in between.

(b) In contrast to the chemical mechanism of continental weathering, the effects described for (i) and (ii) (both of which involve parts of the biosphere) are mechanisms for negative feedback, and therefore act so as to bring about stability in the climate system.

Appendix A

The elements

Atomic number, *Z*	Name	Chemical symbol	Atomic number, *Z*	Name	Chemical symbol	Atomic number, *Z*	Name	Chemical symbol
1	hydrogen	H	32	germanium	Ge	63	europium	Eu
2	helium	He	33	arsenic	As	65	terbium	Tb
3	lithium	Li	34	selenium	Se	66	dysprosium	Dy
4	beryllium	Be	35	bromine	Br	67	holmium	Ho
5	boron	B	36	krypton	Kr	68	erbium	Er
6	carbon	C	37	rubidium	Rb	69	thulium	Tm
7	nitrogen	N	38	strontium	Sr	70	ytterbium	Yb
8	oxygen	O	39	yttrium	Y	71	lutetium	Lu
9	fluorine	F	40	zirconium	Zr	72	hafnium	Hf
10	neon	Ne	41	niobium	Nb	73	tantalum	Ta
11	sodium	Na	42	molybdenum	Mo	74	tungsten	W
12	magnesium	Mg	43	technetium	Tc[a]	75	rhenium	Re
13	aluminium	Al	44	ruthenium	Ru	76	osmium	Os
14	silicon	Si	45	rhodium	Rh	77	iridium	Ir
15	phosphorus	P	46	palladium	Pd	78	platinum	Pt
16	sulfur	S	47	silver	Ag	79	gold	Au
17	chlorine	Cl	48	cadmium	Cd	80	mercury	Hg
18	argon	Ar	49	indium	In	81	thallium	Tl
19	potassium	K	50	tin	Sn	82	lead	Pb
20	calcium	Ca	51	antimony	Sb	83	bismuth	Bi
21	scandium	Sc	52	tellurium	Te	84	polonium	Po[a]
22	titanium	Ti	53	iodine	I	85	astatine	At[a]
23	vanadium	V	54	xenon	Xe	86	radon	Rn[a]
24	chromium	Cr	55	caesium	Cs	87	francium	Fr[a]
25	manganese	Mn	56	barium	Ba	88	radium	Ra[a]
26	iron	Fe	57	lanthanum	La	89	actinium	Ac[a]
27	cobalt	Co	58	cerium	Ce	90	thorium	Th[a]
28	nickel	Ni	59	praseodymium	Pr	91	protoactinium	Pa[a]
29	copper	Cu	60	neodymium	Nd	92	uranium	U[a]
30	zinc	Zn	61	promethium	Pm[a]			
31	gallium	Ga	62	samarium	Sm			

[a] No stable isotopes.

Appendix B

SI fundamental and derived units

Table B1 SI fundamental and derived units.

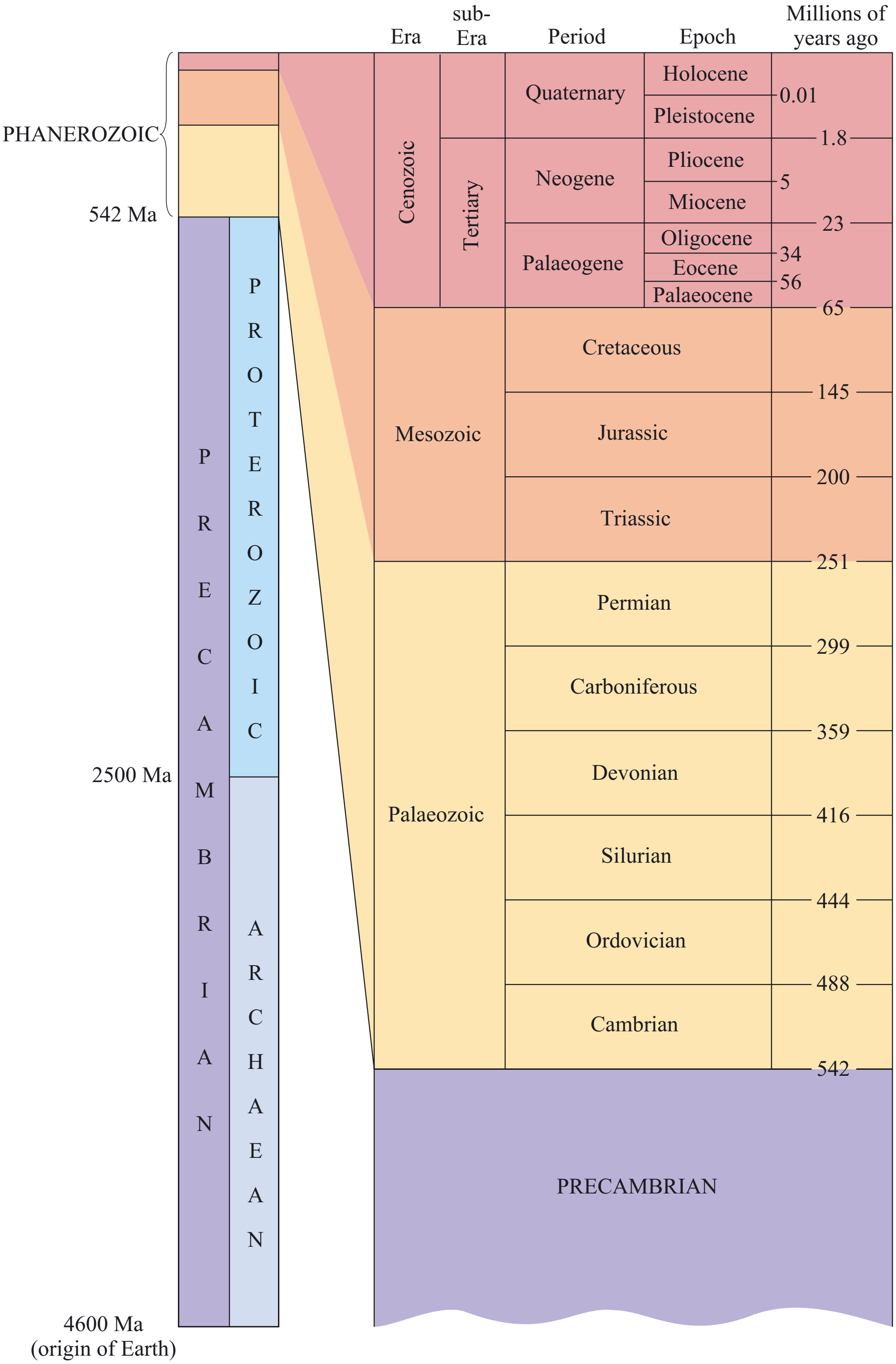
Quantity	Unit	Abbreviation	Equivalent units
mass	kilogram	kg	
length	metre	m	
time	second	s	
temperature	kelvin	K	
angle	radian	rad	
area	square metre	$\mathrm{m^2}$	
volume	cubic metre	$\mathrm{m^3}$	
speed, velocity	metre per second	$\mathrm{m\ s^{-1}}$	
acceleration	metre per second squared	$\mathrm{m\ s^{-2}}$	
density	kilogram per cubic metre	$\mathrm{kg\ m^{-3}}$	
frequency	hertz	Hz	(cycles) $\mathrm{s^{-1}}$
force	newton	N	$\mathrm{kg\ m\ s^{-2}}$
pressure	pascal	Pa	$\mathrm{kg\ m^{-1}\ s^{-2}}$, $\mathrm{N\ m^{-2}}$
energy	joule	J	$\mathrm{kg\ m^2\ s^{-2}}$
power	watt	W	$\mathrm{kg\ m^2\ s^{-3}}$, $\mathrm{J\ s^{-1}}$
specific heat capacity	joule per kilogram kelvin	$\mathrm{J\ kg^{-1}\ K^{-1}}$	$\mathrm{m^2\ s^{-2}\ K^{-1}}$
thermal conductivity	watt per metre kelvin	$\mathrm{W\ m^{-1}\ K^{-1}}$	$\mathrm{m\ kg\ s^{-3}\ K^{-1}}$

The Greek alphabet

Table C1 The Greek alphabet.

Name	**Lower case**	**Upper case**	**Name**	**Lower case**	**Upper case**
Alpha	α	A	Nu (new)	ν	N
Beta (bee-ta)	β	B	Xi (cs-eye)	ξ	Ξ
Gamma	γ	Γ	Omicron	ο	O
Delta	δ	Δ	Pi (pie)	π	Π
Epsilon	ε	E	Rho (roe)	ρ	P
Zeta (zee-ta)	ζ	Z	Sigma	σ	Σ
Eta (ee-ta)	η	H	Tau (torr)	τ	T
Theta (thee-ta; 'th' as in theatre)	θ	Θ	Upsilon	υ	Y
Iota (eye-owe-ta)	ι	I	Phi (fie)	ϕ	Φ
Kappa	κ	K	Chi (kie)	χ	X
Lambda (lam-da)	λ	Λ	Psi (ps-eye)	ψ	Ψ
Mu (mew)	μ	M	Omega (owe-me-ga)	ω	Ω

Appendix D



Era	sub-Era	Period	Epoch	Millions of years ago
Cenozoic		Quaternary	Holocene	0.01
			Pleistocene	1.8
	Tertiary	Neogene	Pliocene	5
			Miocene	23
		Palaeogene	Oligocene	34
			Eocene	56
			Palaeocene	65
Mesozoic		Cretaceous		145
		Jurassic		200
		Triassic		251
Palaeozoic		Permian		299
		Carboniferous		359
		Devonian		416
		Silurian		444
		Ordovician		488
		Cambrian		542
PRECAMBRIAN				

Note that time intervals are not drawn to scale. You may see other versions of this timescale with minor differences, as the subdivisions and radiometric age dates are subject to revision.

Further reading

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Acknowledgements

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