# 1

# The Earth within the Solar System

## 1.1 The Sun and its constancy

Any account of the Earth's atmosphere and ocean cannot be regarded as complete without a discussion of the Sun, the solar system and the place of the Earth within this system. The Sun supplies the energy absorbed by the Earth's atmosphere-ocean system. Some of Sun's energy is converted directly into thermal energy, which drives the atmospheric circulation. A small portion of this energy appears as the kinetic energy of the winds which, in turn, drives the ocean circulation. Some of the intercepted solar energy is transformed by photosynthesis into biomass, a large proportion of which is ultimately converted into heat energy by chemical oxidation within the bodies of animals and by the decomposition and burning of vegetable matter. A very small proportion of the photosynthetic process produces organic sediments which may eventually be transformed into fossil fuels. It is estimated that the solar radiation intercepted by the Earth in seven days is equivalent to the heat that would be released by the combustion of all known reserves of fossil fuels on the Earth. The Sun, therefore, is of fundamental importance in the understanding of the uniqueness of the Earth.

The Sun is a main sequence star in the middle stages of its life and was formed $4.6 \times 10^9$ years ago. It is composed mainly of hydrogen (75% by mass) and helium (24% by mass); the remaining 1% of the Sun's mass comprises the elements oxygen; nitrogen; carbon; silicon; iron; magnesium and calcium. The emitted energy of the Sun is $3.8 \times 10^{26}$ W and this energy emission arises from the thermonuclear fusion of hydrogen into helium at temperatures around $1.5 \times 10^7$ K in the core of the Sun.

In the core, the dominant constituent is helium (65% by mass) and the hydrogen content is reduced to 35% by mass as a direct result of its consumption in the fusion reactions. It is estimated that the remaining hydrogen in the Sun's core is sufficient to maintain the Sun at its present luminosity and size for a further $4 \times 10^9$ years. At this stage it is expected that the Sun will

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expand into a red giant and engulf all of the inner planets of the solar system (i.e. Mercury, Venus, Earth and Mars).

There exists a high-pressure gradient between the core of the Sun and its perimeter, and this is balanced by the gravitational attraction of the mass of the Sun. In the core, the energy released by the thermonuclear reaction is transported by energetic photons but, because of the strong absorption by peripheral gases, most of these photons do not penetrate to the surface. This absorption causes heating in the region outside the core. In contrast, the outer layers of the Sun are continually losing energy by radiative emission into space in all regions of the electromagnetic spectrum. This causes a large temperature gradient to develop between the surface and the inner region of the Sun. This large temperature gradient produces an unstable region and large scale convection currents are set up that transfer heat to the surface of the Sun. The convection currents are visible as the fine grain structure, or granules, in high resolution photos of the Sun's surface. It is thought that the convection currents have a three-tier structure within the Sun. The largest cells, $200 \times 10^3$ km in diameter are close to the core. In the middle tier the convection cells are about $30 \times 10^3$ km in diameter and at the surface they are $1 \times 10^3$ km. The latter cells have a depth of 2000 km. In each cell, hot gas is transported towards the cooler surface, whilst the return flow transports cooler gases towards the interior.

Almost all of the solar radiation emitted into space, approximately 99.9%, originates from the visible disc of the Sun, known as the photosphere. The photosphere is the region of the Sun where the density of the solar gas is sufficient to produce and emit a large number of photons, but where the density of the overlying layers of gas is insufficient to absorb the emitted photons. This region of the Sun has a thickness of approximately 500 km but no sharp boundaries can be defined. The radiative spectrum of the Sun, when fitted to a theoretical black body curve, gives a black body temperature of 6000 K, although the effective temperature, deduced from the total energy emitted by the Sun, gives a lower temperature of 5800 K.

The photosphere is not uniform in temperature. The lower regions of the photosphere have temperatures of 8000 K, whilst the outer regions have temperatures of 4000 K. Furthermore, the convection cells produce horizontal temperature variations of 100 K between the ascending and descending currents of solar gas. Larger convection cells also appear in the photosphere and they have diameters of 30 000 km. They appear to originate from the second tier of convection within the Sun. The appearance of sunspots gives rise to horizontal variations of 2000 K within the photosphere. The inner regions of the sunspots have black body temperatures of 4000 K. Sunspots have diameters of 10 000–150 000 km and they may last for many weeks. It is thought that the 'sunspot' causes a localised suppression of the convection and therefore leads to a reduction in the transfer of heat into the photosphere from the interior. However, although the sunspot features are dramatic, they occupy less than 1% of the Sun's disc and therefore the effect on the luminosity of the Sun is small.

Beyond the photosphere lies the chromosphere where the temperature decreases to a minimum of 4000 K at 2000 km above the photosphere and then increases sharply to a temperature of $10^6$ K at a height of 5000 km in the region of the corona. However, because of the low density of the gas in this region, the radiation emitted from the chromosphere and the corona only amount to 0.1% of the total radiation from the Sun.

These different temperature zones of the Sun can be observed in the solar spectrum (Figure 1.1). The visible and infra-red radiation emitted from the photosphere follow reasonably closely the black body curve for a temperature of 5800 K. However, substantial deviations from the theoretical curve occur in the X-ray and radio wavebands, and lesser deviations occur in the ultraviolet spectrum. The high temperature of the corona is responsible for the intense X-ray band, whilst the high radio frequency energy is associated with the solar wind and solar activity. However, the energy in these wavebands is a negligible fraction of the total emitted energy and therefore these very variable regions of the spectrum have little direct influence on the total solar energy received on the Earth. The depletion of energy in the ultraviolet spectrum is the result of emission at a lower temperature than the photosphere and therefore probably originates in the temperature minimum of the lower

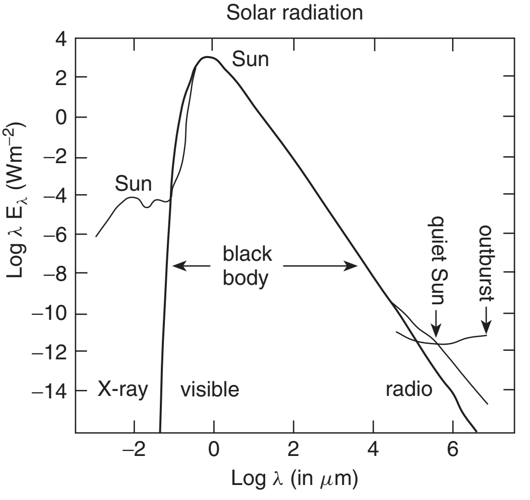


**Figure 1.1** Solar spectrum and blackbody curve: energy distribution of the Sun and a black body at 5800 K. Reproduced, with permission, from Allen, C.W., 1958, Quarterly Journal of the Royal Meteorological Society, 84: page 311, figure 3

chromosphere. Recent observations have shown that the ultraviolet energy is not constant and shows considerable variability in the short-wave part of the spectrum amounting to 25% of its average value at 0.15 μm, and 1% at 0.23 μm. Again, the amount of energy in this band is relatively small and contributes less than 0.1% of the total energy. Furthermore, satellite observations of the solar spectrum since 1978 have demonstrated a 0.1% variation of the solar output over the 11 year sunspot cycle. These variations are rather small when compared to those produced by orbital variations of the Earth around the Sun.

## 1.2 Orbital variations in solar radiation

Let $S$ be the total solar output of radiation in all frequencies. At a distance, $r$, from the centre of the Sun, imagine a sphere of radius $r$ on which the flux of radiation will be the same (assuming the radiation from the Sun is equal in all directions). If the flux of radiation per unit area at a distance $r$ is given by $Q(r)$, then the total radiation on the imagined sphere is $4\pi r^2 Q(r)$.

In the absence of additional energy sources to the Sun

$$S = 4\pi r^2 Q(r) \tag{1.1}$$

Rearranging:

$$Q(r) = \frac{S}{4\pi r^2} \tag{1.2}$$

In practice, the solar radiation cannot be measured at the Sun but it can be measured by satellites above the Earth's atmosphere. Recent determinations of the flux of radiation per unit area, $Q$, give a value of 1360 W m$^{-2}$. Given that the Earth is approximately $1.5 \times 10^{11}$ m away from the Sun, S can be calculated to be $3.8 \times 10^{26}$ W.

Though $S$ is the true solar constant, in meteorology $Q$ is defined as the solar constant of the Earth. Table 1.1 shows the value of the solar constant obtained for other planets in the solar system. It is noted that the dramatic changes in the solar constant between the Earth and our nearest planetary neighbours, Mars and Venus, merely serve to highlight the uniqueness of the position of the Earth in the solar system. At the radius of Pluto (some 39 Earth-Sun distances), the flux of the radiation from the Sun is less than 1 W m$^{-2}$.

The radiation incident on a spherical planet is not equal to the solar constant of the planet. The planet intercepts a disc of radiation of area $\pi a^2$, where a is the planetary radius, whereas the surface area of the planet is $4\pi a^2$. Hence the solar radiation per unit area on a spherical planet is

$$\frac{Q\pi a^2}{4\pi a^2} = \frac{Q}{4} \tag{1.3}$$

**Table 1.1** Radiative properties of terrestrial planets. $Q$ is the solar irradiance at distance $r$ from the Sun, $\alpha$ is the planetary albedo, $T_e$ is the radiative equilibrium temperature and $T_s$ is the surface temperature

| | $r(10^9 m)$ | $Q(W m^{-2})$ | $\alpha$ | $T_e(K)$ | $T_s(K)$ |
|---|---|---|---|---|---|
| Venus | 108 | 2623 | 0.75 | 232 | 760 |
| Earth | 150 | 1360 | 0.30 | 255 | 288 |
| Mars | 228 | 589 | 0.15 | 217 | 227 |

The average radiation on the Earth's surface is $340\,W\,m^{-2}$. The above discussion assumes the total absence of an atmosphere and that the Earth is a perfect sphere in a spherical orbit.

The three geometrical factors which determine the seasonal variation of solar radiation incident on the Earth are shown in Figure 1.2. The Earth revolves around the Sun in an elliptical orbit, being closest to the Sun, i.e. at perihelion, about 4 January and farthest from the Sun, i.e. at aphelion, on 4 July. The eccentricity, represented by the symbol $e$, of the present day orbit is 0.017. At aphelion, it can be shown that

$$r_{aphelion} = (1 + e)\overline{r}$$

where $\overline{r}$ is the mean distance between the Earth and the Sun.

As the incident radiation is inversely proportional to the square of the Earth's distance from the Sun, it can also be shown that, at aphelion, the radiation received by the Earth is 3.5% less than the annual average, whilst

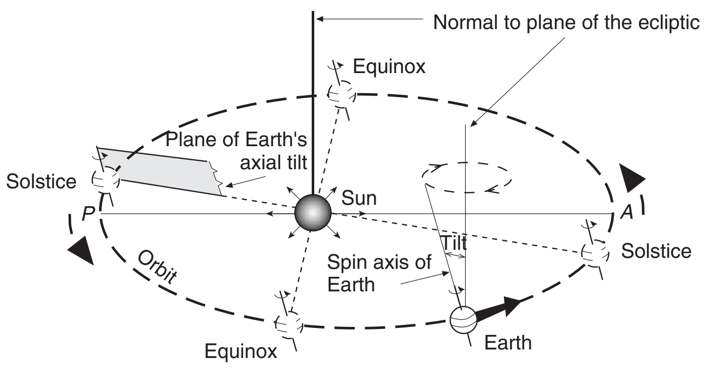


**Figure 1.2** Geometry of the Earth–Sun system. The Earth's orbit, the large ellipse with major axis AP and the Sun at one focus, defines the plane of the ecliptic. The plane of the Earth's axial tilt (obliquity) is shown passing through the Sun corresponding to the time of the southern summer solstice. The Earth moves around its orbit in the direction of the solid arrow (period one year) whilst spinning about its axis in the direction shown by the thin curved arrows (period one day). The broken arrows shown opposite the points of aphelion (A) and perihelion (P) indicate the direction of the very slow rotation of the orbit. Reproduced, with permission, from Pittock, A.B. *et al.* eds, 1978, Climate Change and Variability: A Southern Perspective, Cambridge University Press: page 10, figure 2.1

at perihelion it is 3.5% greater than the annual average. Therefore the total radiation incident on the Earth over the course of one year is independent of the eccentricity of the Earth's orbit. The long-term variation in eccentricity indicates changes of 0.005–0.060 occurring with a period of 100 000 years. This would produce changes of up to 10% in the radiation incident on the surface of the Earth at perihelion and aphelion.

However, the angle of tilt ($\varepsilon$) of the Earth with respect to the plane of rotation of the Earth's orbit, i.e. the obliquity of the ecliptic, is the dominant influence on the seasonal cycle of solar radiation (Figure 1.3). It not only determines the march of the seasons, but it is also important in determining the latitudinal distribution of the climatic zones. At the present time, the angle is 23.5º, which is similar to the obliquity of Mars. It can be shown that the amplitude of the seasonal variation in the solar radiation is directly proportional to the obliquity. If the obliquity were reduced, there would be a reduction in the amplitude of the seasonal solar radiation cycle, and if

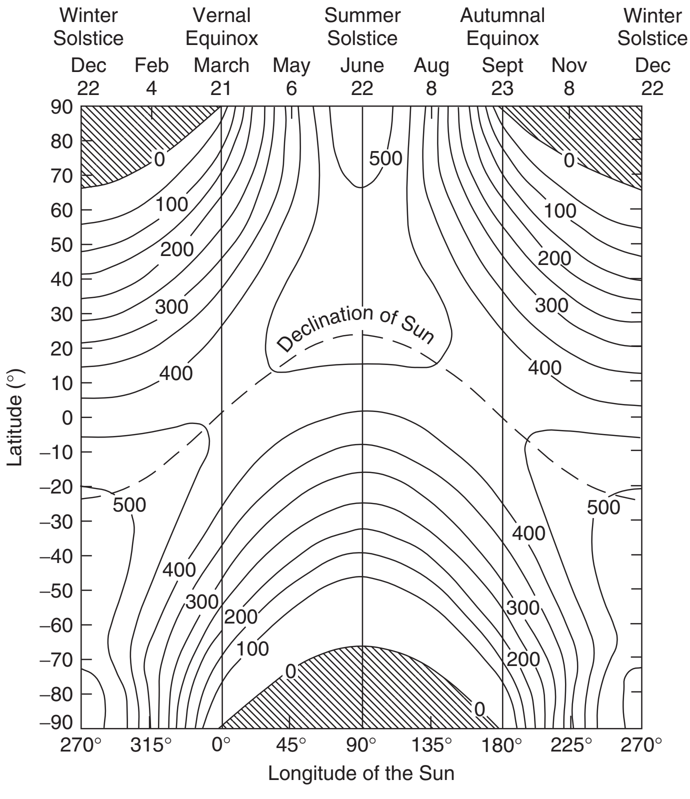


**Figure 1.3** Solar radiation in W m$^{-2}$ arriving at the Earth's surface, in the absence of an atmosphere, as a function of latitude and time of year at 2000 A.D. Reproduced, with permission, from Hess, S.L., 1959, Introduction to Theoretical Meteorology, Holt, Rinehart & Winston: page 132, figure 9.1

the obliquity were zero, the seasonal variation would depend solely on the eccentricity of the orbit.

The obliquity has varied between 22° and 24.5° over the past million years and this variation has a total period of 41 000 years. Such variations cannot influence the total radiation intercepted by the Earth, but they will exert an influence on the seasonal cycle.

The third parameter which affects the seasonal cycle is the longitude of the perihelion, measured relative to the vernal equinox. It precesses by 360° in a period of 21 000 years. Therefore, in about 10 500 years' time, the Earth will be closest to the Sun in July and farthest away in January.

The sensitivity of the global temperature to the seasonal cycle, which arises from ice, cloud and ocean feedback processes, indicates that these changes may be sufficient to initiate long term changes in climate and, in the most extreme cases, ice ages (Figure 1.4).

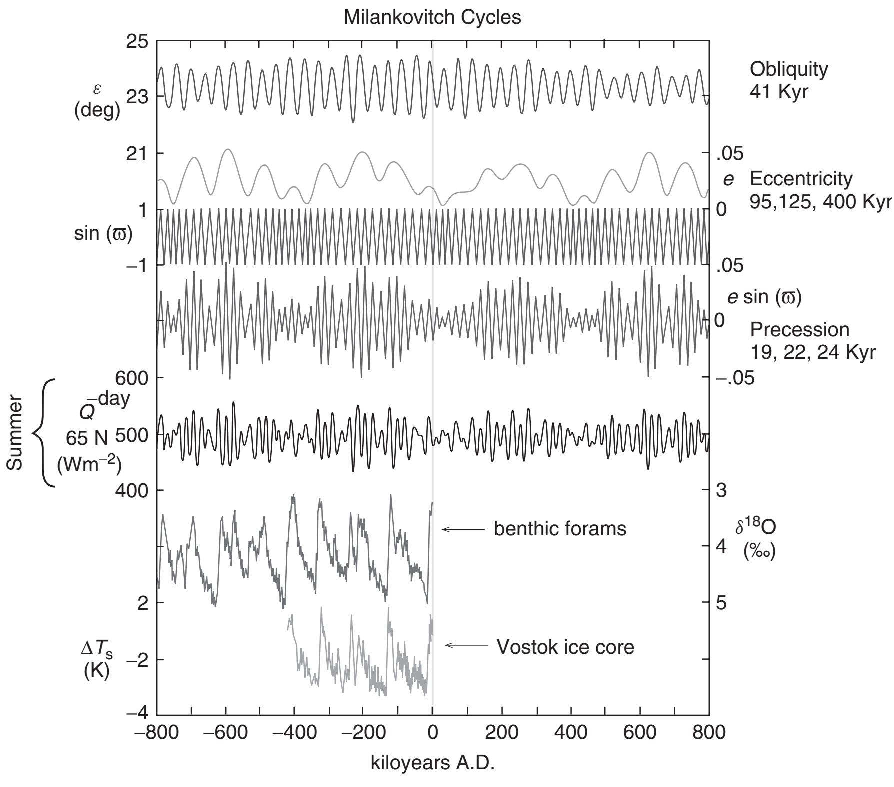


**Figure 1.4** Past and future Milankovitch cycles $\varepsilon$ is obliquity (axial tilt), e is eccentricity, $\varpi$ is longitude of perihelion. e $\sin(\varpi)$ is the precession index, which together with obliquity, controls the seasonal cycle of insolation. $\overline{Q}^{day}$ is the calculated daily-averaged insolation at the top of the atmosphere, on the day of the summer solstice at 65 N latitude. Benthic forams and Vostok ice core show two distinct proxies for past global sea level and temperature, from ocean sediment and Antarctic ice respectively. Vertical gray line is current condition, at 2 ky A.D. See plate section for a colour version of this image

Milankovitch, a Serbian scientist, recognized the relationship between the variations in the Earth's orbit around the Sun and climate change over the last two million years. In particular, he determined the orbital conditions which would lead to the initiation of an ice age over the Northern Hemisphere continents. He reasoned that a reduction in solar radiation during the polar summer would lead to a reduction in summer melting of the snow cover. This, in turn, over many years, would lead to the build up of ice over the continents. The optimum orbital conditions for this reduction in the summer solar radiation are:

(i) That aphelion occurs during the Northern Hemisphere summer.
(ii) That the eccentricity be large, to maximise the Earth–Sun separation at aphelion.
(iii) That the obliquity be small.

These three factors would act to decrease the amplitude of the solar radiation cycle and therefore reduce the solar radiation in summer at high latitudes in the Northern Hemisphere. However, all of these factors would tend to lead to an increase in solar radiation in the Northern Hemisphere winter. Comparison of the orbital parameters with long term temperature records, deduced from the dating of deep ocean sediment cores, indicate that there is a good statistical agreement with the theory.

## 1.3 Radiative equilibrium temperature

The inner planets of the solar system have one common attribute – the lack of a major source of internal energy. It is, of course, true that there is a steady geothermal flow of energy from the interior of the Earth as a result of radioactive decay in the Earth's core, but this energy flow is less than $0.1\,\mathrm{W\,m^{-2}}$ and can be compared with an average solar heat flux of $340\,\mathrm{W\,m^{-2}}$. What, then, is the fate of this steady input of solar energy? On average, about 30% is scattered back into space by clouds; snow; ice; atmospheric gases and aerosols, and the land surface, leaving 70% available for the heating of the atmosphere, the ocean and the Earth's surface. However, the annual average of the Earth's temperature has not changed by more than 1 K over the past 100 years. Why, then, have the temperatures of the atmosphere and ocean not increased? The observation that both the temperatures of the atmosphere and the ocean have remained relatively constant implies that energy is being lost from the ocean, the atmosphere and the Earth's surface at approximately the same rate as it is being supplied by the Sun. The only mechanism by which this heat can be lost is by the emission of electromagnetic radiation from the Earth's atmosphere into space. If it is assumed that the Earth can be represented by a black body, then it is possible to apply Stefan's law to the electromagnetic emission of the Earth. Stefan's law states that the total emission of radiation from a black body over all wavelengths is proportional to the fourth power of the temperature, i.e. $E = \sigma T^4$, where $\sigma$ is the Stefan – Boltzmann constant. For the emission by the Earth of an amount of radiation equal to that received from the Sun, there can be defined a radiative equilibrium temperature, $T_e$.

If $Q/4$ is the energy flux from the Sun and $\alpha$ is the fraction of solar radiation reflected back into space, known as the planetary albedo, then for the Earth's system in thermal equilibrium:

$$\left(\frac{Q}{4}\right)(1-\alpha) = \sigma T_e^4 \tag{1.4}$$

where $\sigma = 5.67 \times 10^{-8}\,\mathrm{W\,m^{-2}\,K^{-4}}$. By inserting $\alpha = 0.3$ and $Q/4 = 340\,\mathrm{W\,m^{-2}}$, a radiative temperature of 255 K or $-18^\circ$C can be calculated.

This temperature is lower than might be expected. The mean surface temperature of the Earth is 15°C, or 288 K, and so it is clear that the radiative temperature bears little direct relationship to the observed surface temperature. This is because most of the planetary radiation is emitted by the atmosphere, whilst only a small fraction originates from the surface of the Earth. The temperature of the atmosphere decreases by $6.5\,\mathrm{K\,km^{-1}}$ from the surface to the tropopause, 10 km above the Earth's surface, and therefore the radiative equilibrium temperature corresponds to the temperature at a height of 5 km.

The Wien displacement equation enables the calculation of the wavelength of maximum radiation, $\lambda_{max}$, and it states that

$$\lambda_{max} T = 2897\,\mu\mathrm{m\,K} \tag{1.5}$$

If the radiative temperature, $T_e = 255$ K, is substituted in this equation, then $\lambda_{max}$ is calculated to be approximately 11 μm, which lies in the middle part of the infra red spectrum. In this region of the electromagnetic spectrum, water and carbon dioxide have large absorption bands so that all substances containing water are particularly good absorbers of radiation in the middle infra red. As far as emission is concerned, water has an emissivity of 0.97, which means that it emits radiation at the rate of 97% of the theoretical black body value. Hence clouds, which are composed mainly of water droplets, are good infra red emitters. The prevalence of liquid water and water vapour in the Earth's atmosphere therefore leads to the strong absorption and re-emission of radiation. This ability of water to absorb and re-emit radiation back to the Earth's surface results in the higher observed mean surface temperature. If the atmosphere was transparent to the emitted planetary radiation, then the surface temperature would be close to the radiative equilibrium temperature.

The temperature of the surface of a planet without an atmosphere, such as Mercury, would be observed to have a surface equilibrium temperature equal to that of the radiative equilibrium temperature. Table 1.1 shows the radiative equilibrium temperatures for Mars and Venus, as well as their surface temperatures. The surface temperatures are higher than the radiative temperatures. On Mars the atmospheric mass is smaller than that of Earth by two orders of magnitude and therefore, though carbon dioxide absorbs and re-emits planetary radiation back to the surface, a surface warming of only 19 K is produced. However, on Venus the surface temperature of 760 K is sufficient to melt lead, in spite of the fact that the radiative equilibrium temperature is less than that of the Earth. This low radiative equilibrium temperature is caused by the reflection of 77% of the incident solar radiation back into space by the omnipresent clouds, which are not composed of water droplets as they are on Earth. Therefore, although the planet receives less net solar radiation than the Earth, it has a surface temperature 472 K higher. This is the result of the massive carbon dioxide atmosphere, which absorbs virtually all of the radiation emitted by the surface and re-emits it back to the surface. Furthermore, it is known that clouds of sulphuric acid also enhance this warming effect.

The ability of an atmosphere to maintain a surface temperature above the radiative equilibrium temperature is commonly known as the ‘greenhouse effect’.

It is clear that, although the distance from the Sun determines the energy incident on an atmosphere, the mass and the constituents of that atmosphere are also important factors in the determination of the surface temperature of the planet.

## 1.4 Thermal inertia of the atmosphere

The thermal inertia of the atmosphere gives an indication of how quickly the atmosphere would respond to variations in solar radiation and it is therefore of importance in the understanding of climatic change. Consider an atmosphere having a mass $M$, per unit area, and a specific heat at constant pressure, $C_p$. The thermal inertia of the atmosphere is $MC_p$.

Initially, the atmosphere is in thermal equilibrium and therefore the solar radiation absorbed by the atmosphere is in balance with the emission of long wave planetary radiation into space (equation 1.4).

Now consider a situation where heat is suddenly added to the atmosphere, for example, as the result of burning fossil fuels or by a large thermonuclear explosion. How long would the atmosphere take to regain its former equilibrium? If all the heat were liberated at the same time, then the atmosphere’s temperature would increase suddenly by $\Delta T$. The planetary radiation emitted into space would now be $\sigma(T_e + \Delta T)^4$ and therefore more radiation would be emitted than received from the Sun. This deficit implies that the atmosphere would cool. The rate of temperature change, $dT/dt$, is proportional to the net difference between the emitted planetary radiation and the solar radiation and thus:

$$MC_p \frac{dT}{dt} = -\sigma(T_e + \Delta T)^4 + \frac{Q}{4}(1-\alpha)$$

Rearranging and substituting from equation 1.4:

$$MC_p \frac{dT}{dt} = -\sigma T_e^4 \left(1 + \frac{\Delta T}{T_e}\right)^4 + \sigma T_e^4$$

Expanding by the binomial expansion

$$MC_p \frac{dT}{dt} = -\sigma T_e^4 \left(1 + \frac{4\Delta T}{T_e} + 6\left[\frac{\Delta T}{T_e}\right]^2 + 4\left[\frac{\Delta T}{T_e}\right]^3 + \left[\frac{\Delta T}{T_e}\right]^4\right) + \sigma T_e^4$$

Providing that $\Delta T/T_e \ll 1$, then the higher order terms in the above expression can be neglected. Therefore, retaining only the first order terms:

$$MC_p \frac{dT}{dt} = -\sigma T_e^4 \frac{4\Delta T}{T_e}$$

Since $T = T_e + \Delta T$, where $T_e$ is independent of time, then:

$$\frac{dT}{dt} = \frac{d\Delta T}{dt}$$

Hence:

$$\frac{d\Delta T}{dt} = -\left(\frac{4\sigma T_e^3}{MC_p}\right)\Delta T \tag{1.6}$$

The solution of the above equation is:

$$\Delta T(t) = \Delta T_0 \exp(-t/\tau) \tag{1.7}$$

where $\Delta T_0$ is the initial temperature perturbation at $t = 0$ and $\tau_R$ is known as the radiation relaxation time constant.

$$\tau_R = \frac{MC_p}{4\sigma T_e^3} \tag{1.8}$$

For Earth's atmosphere, where $M = 10.316 \times 10^3 \mathrm{kg\,m^{-2}}$, $\sigma = 5.67 \times 10^{-8}\,\mathrm{W\,m^{-2}\,K^{-4}}$, $C_p = 1004\,\mathrm{J\,kg^{-1}\,K^{-1}}$ and $T_e = 255\,\mathrm{K}$, $\tau_R$ can be calculated to be $2.7 \times 10^6$s or 32 days.

From equation 1.7:

$$\Delta T = \Delta T_0 e^{-1} \text{ at } t = \tau_R$$

$$\Delta T = \Delta T_0 e^{-2} \text{ at } t = 2\tau_R$$

$$\Delta T = \Delta T_0 e^{-3} \text{ at } t = 3\tau_R$$

For an initial temperature perturbation of 1 K, equivalent to the instantaneous burning of $160 \times 10^9$ t of coal, then after approximately 32 days, the temperature perturbation will be reduced to 0.37 K; after 65 days it will be 0.13 K and after 96 days it will be 0.05 K. For the Earth's atmosphere, therefore, the colossal excess of heat will be lost by planetary radiation to space within 100 days and the Earth's radiative equilibrium will be restored.

Although this particular problem is hypothetical, the concept is a useful one for understanding the response of the atmosphere to changes in the radiative balance caused by daily and seasonal variations in solar radiation. For instance, because the radiative time constant is much greater than 1 day, the lower part of the Earth's atmosphere (wherein is located the major part of the mass of the atmosphere) does not respond significantly to the daily variation in solar radiation. However, for seasonal variations, where the solar radiation changes are of a longer period than the radiative time constant, there will be a significant response in the radiation temperature of the atmosphere.

Table 1.2 shows a comparison of the radiative time constants of Mars, Venus and the Earth. On Mars, because of the small thermal inertia, the diurnal variation of temperature at the surface is very large and, as a result, large thermal tides are set up in the atmosphere.

This phenomenon is also found in the upper reaches of the Earth's atmosphere, where the atmospheric mass is small and the thermal capacity is reduced. However, on Venus the diurnal variation in temperature is only 1–2 K because of the very large radiative time constant, even though the Venusian day is 243 Earth days long.

It must, of course, be remembered that the seasonal change in temperature on Earth is modified by the thermal capacity of the ocean. The thermal capacity of the atmosphere is equivalent to 2.6 m of sea water, and therefore it can be seen that the thermal capacity of the ocean is equivalent to approximately

**Table 1.2** Radiative time constant, $\tau_R$ for the terrestrial planets. $C_p$ is the specific heat at constant pressure, p is the surface pressure and g is the acceleration due to gravity

| | $C_p$ (J kg$^{-1}$ K$^{-1}$) | p(hPa) | g(m s$^{-2}$) | $\tau_R$(days) |
|---|---|---|---|---|
| Venus | 830 | 93 000 | 8.9 | 3550 |
| Earth | 1004 | 1013 | 9.8 | 32 |
| Mars | 830 | 6 | 3.7 | 1 |

1600 atmospheres. However, during the seasonal cycle, only the upper 100 m of the ocean is affected significantly and therefore the thermal capacity is equivalent to only 38 atmospheric masses, giving a thermal relaxation time of approximately 3.5 years. This simple sum suggests that the ocean is responsible for a major modification of the seasonal cycle of temperature in the atmosphere. An example of the oceanic influence is shown in Figure 1.5. The flux of the long wave radiation from the top of the atmosphere gives an indication of the seasonal change in temperature. In the Northern Hemisphere, where the largest ratio of continents to land exists (Figure 1.10 shows about 70% is land at 60°N), there is a large change of approximately 40 K in the radiation temperature during the seasonal cycle, whilst in the Southern Hemisphere between 30° and 60°S, where the continental area is less than 10%, there is only a small change in radiation temperature during the year. The most continental region benefits from amelioration in its seasonal climate

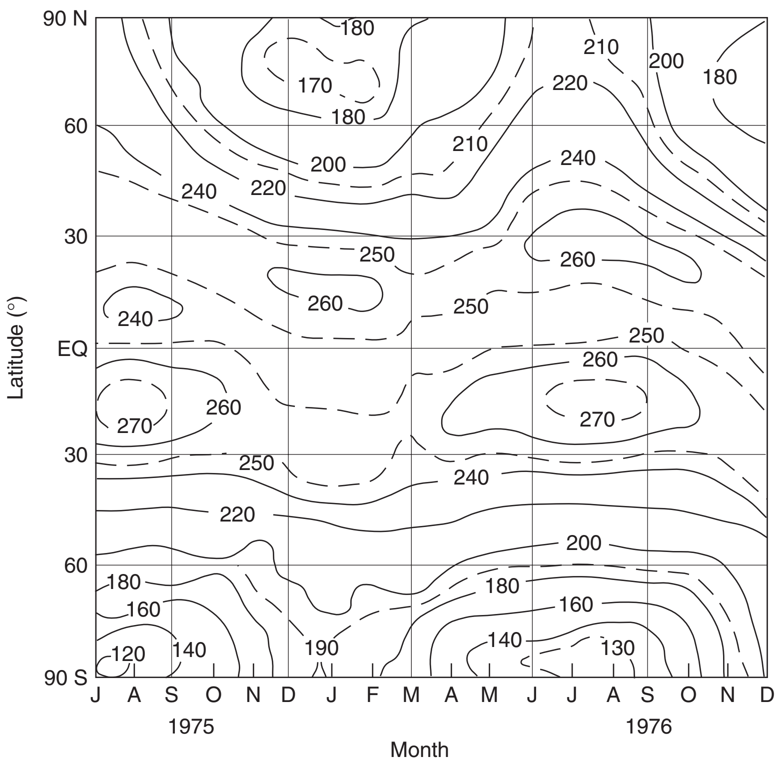


**Figure 1.5** Seasonal variation of the zonal mean long-wave radiation flux, $W\,m^{-2}$, measured by satellite from July 1975 to December 1976. A flux of $260\,W\,m^{-2}$ corresponds to a radiation temperature of 260 K, whilst a flux of $160\,W\,m^{-2}$ is equivalent to a temperature of 230 K. Reproduced, with permission, from Jacobowitz, H. *et al.*, 1979, Journal of Atmospheric Science, 36: page 506, figure 6

brought about by the ocean. Even in Siberia, the winters are not as severe as they would be if the ocean did not exist.

## 1.5 Albedo

The planetary albedo is the ratio of the reflected (scattered) solar radiation to the incident solar radiation, measured above the atmosphere. As this reflected radiation is lost immediately from the Earth-atmosphere system, the accurate determination of planetary albedo is important in the calculation of the amount of solar radiation absorbed by the system, and therefore the climate of the Earth. On average, the global albedo is 30% (±2%). This value has been determined from a number of satellite measurements over a period of many years. The global albedo is not constant but varies on both seasonal and interannual time scales by approximately 2%. Observations have shown that the albedo is a maximum in January and a minimum in July. If the cloud and surface distributions were similar in both hemispheres, then no annual variation would be expected. However, because of the large seasonal cycle in snow extent over the Northern Hemisphere continents, particularly over the Eurasian land mass, and the seasonal cycle of cloud in the Northern Hemisphere, the global albedo has a seasonal component.

The latitudinal variation in albedo is determined by:

(i) The elevation of the Sun.
(ii) The distribution and type of cloud.
(iii) The surface albedo.

At latitudes less than 30°, the planetary albedo is relatively constant at 25% but it then increases with latitude to a maximum value of 70% over the Poles. However, because the polar regions occupy less than 8% of the Earth's surface, the value of the global albedo is weighted to the albedo of the middle and low latitudes. It is noted that the latitude distributions in the albedo in the Northern and Southern Hemispheres are similar, despite the difference in land and sea distribution between the two hemispheres. This indicates that the cloud distributions in both hemispheres have a dominant influence over the surface variations in albedo in determining the planetary albedo (see Figure 1.6).

The long term latitudinal variation in absorbed solar energy and the emitted planetary variation are depicted in Figure 1.7a. The absorbed solar energy has a maximum value of 300 W $m^{-2}$ in low latitudes. The planetary radiation, which is dominated by emission from the lower troposphere, shows a decrease with latitude that occurs at a slower rate than the decrease in absorbed solar radiation. If the atmosphere were in radiative balance at each latitude, the

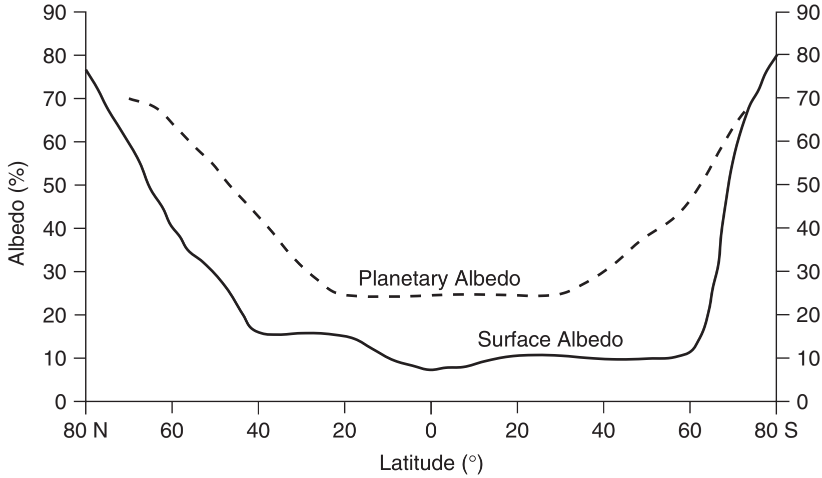


**Figure 1.6** Planetary and surface albedo (%) expressed as a function of latitude. The planetary albedo is the fraction of incident solar energy reflected back into space, by the atmosphere and the Earth's surface

two curves should be identical. However, the solar absorption exceeds the planetary emission between 40°N and 40°S, and therefore there is a net excess in low latitudes and a net deficit at high latitudes (i.e. poleward of 40°). This latitudinal imbalance in radiation implies that thermal energy (heat) must be transferred from low to high latitudes by the circulation of the atmosphere and ocean. In effect, the atmosphere and oceanic circulations are maintaining higher atmospheric temperatures polewards of 40° than can be maintained by solar absorption.

To obtain the transport of heat by the atmosphere and ocean, the net radiation at each latitude, i.e. the difference between the absorbed solar radiation and the emitted planetary radiation, is integrated from pole to pole (Figure 1.7b). Thus, it is possible to determine the poleward transport of heat from satellite radiation measurements, but it is not possible to distinguish what proportion of heat is transferred by the atmosphere and what proportion by the ocean. It is noted that the maximum transfer of heat occurs between 30° and 40° of latitude and is approximately $5 \times 10^{15}$ W.

Seasonal variations in the surface albedo are significant, particularly in the middle and high latitudes of the Northern Hemisphere, and arise from changes in snow and cloud cover. At 60°N the albedo varies from 65% in February to less than 40% in July, as a result of the seasonal variation in snow cover. In the Southern Hemisphere the seasonal cycle is not so pronounced but a minimum in albedo can be observed in March at 60°S, which is probably associated with the minimum sea ice extent around Antarctica. The emitted planetary radiation shows a strong asymmetry between the Southern and

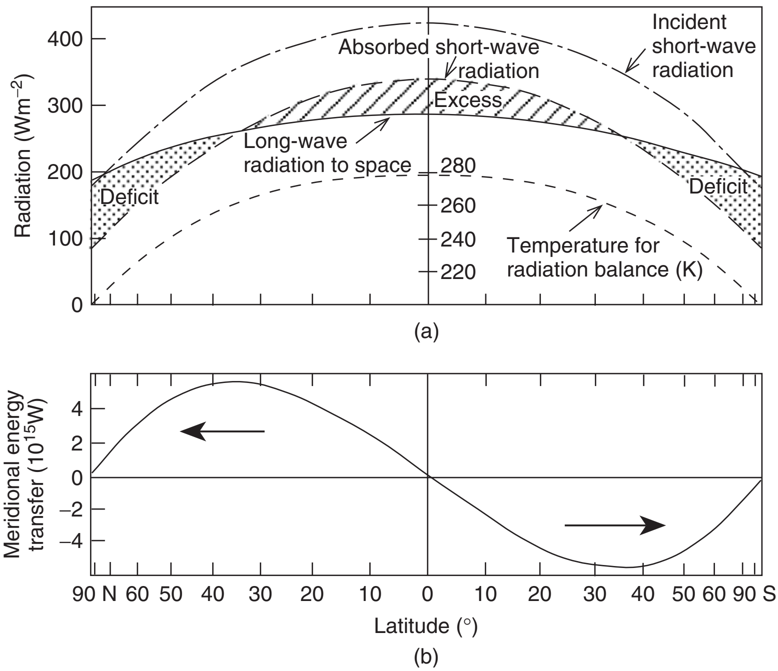


**Figure 1.7** (a) Radiation balance of the Earth–atmosphere system as a function of latitude. The thin broken curve indicates an equivalent radiative temperature (centre scale) for each latitude band to be individually in balance with the absorbed solar radiation (heavy broken curve). The thin solid curve shows the actual observed long-wave radiation emitted to space as a function of latitude; (b) Poleward energy transfer within the Earth – atmosphere system needed to balance the resulting meridional distribution of net radiative gains (hatched) and losses (stippled) as indicated in (a). Note that the horizontal scale is compressed towards the poles so as to be proportional to the surface area of the Earth between latitude circles. Reproduced, with permission, from Pittock, A.B. *et al.* eds, 1978, Climate Change and Variability: A Southern Ocean Perspective, Cambridge University Press: page 12, figure 2.1.2

the Northern Hemispheres, which is the result of the different land-sea distributions in each hemisphere.

The net radiation budget, illustrated in Figure 1.8, indicates substantial seasonal variation as the result of the seasonal solar distribution. The zero contour varies from 15° to 70° latitude, thus showing that the zone for maximum heat transport will also vary during the seasonal cycle. The maximum net radiation gain varies between 30°S and 30°N, and is larger in the Southern Hemisphere than in the Northern Hemisphere.

The surface albedo, as has been shown, makes an important contribution to the planetary albedo although the cloud distribution is, without doubt, the dominant influence. However, the amount of solar radiation absorbed by the surface, and therefore the surface temperature, are strongly influenced by

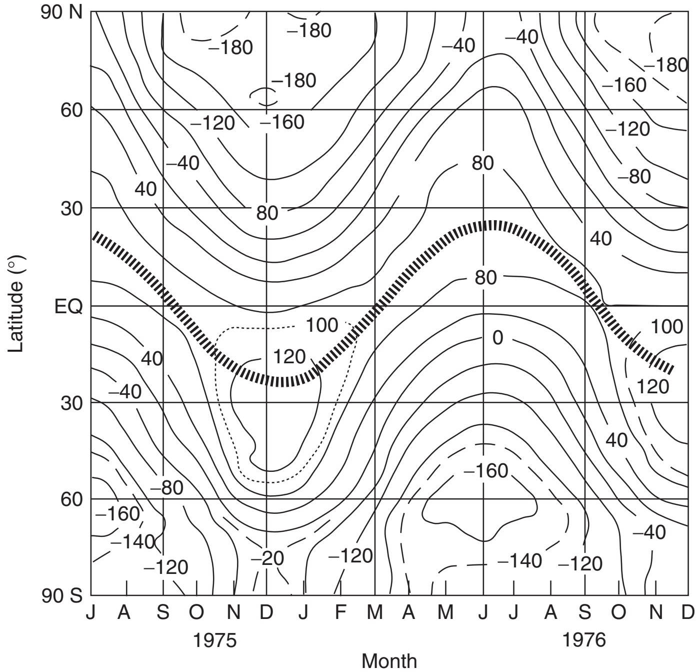


**Figure 1.8** Seasonal variation of zonal mean net (solar-longwave) radiation budget, W m$^{-2}$, measured by satellite from July 1975 to December 1976. Reproduced, with permission, from Jacobowitz, H. *et al.*, 1979, Journal of Atmospheric Science, 36: page 506, figure 7

the albedo of that surface. The albedo of any particular surface depends on the following factors:

(i) The type of surface.
(ii) The solar elevation and the geometry of the surface relative to the Sun.
(iii) The spectral distribution of the solar radiation and the spectral reflection.

Table 1.3 shows typical values of the surface albedo. For a calm tropical sea surface, where the Sun's elevation is high, the albedo can be as low as 2%, and therefore the oceans are generally very good absorbers of solar radiation in comparison with most types of land surface. Forests and wet surfaces generally have low albedos whilst deserts and snow-covered surfaces have high albedos. The surface albedo is also a function of the spectral reflectivity of the surface, and variations between visible and near infra red can be substantial. For instance, though freshly fallen snow has an albedo greater than 90% in the visible, at 1.5 μm, in the near infra red, its albedo is only 25%. Vegetation usually has a low albedo in the visible region but a high albedo in the near infra red. Although it is useful to be able to derive a single albedo for

**Table 1.3** Albedos for different terrestrial surfaces

| Type of surface | Albedo% |
|---|---|
| Ocean | 2–10 |
| Forest | 6–18 |
| Grass | 7–25 |
| Soil | 10–20 |
| Desert (sand) | 35–45 |
| Ice | 20–70 |
| Snow (fresh) | 70–80 |

the solar spectrum, it must be remembered that the solar spectrum incident on the surface changes with solar elevation and cloud cover, and that the albedo will also change.

## 1.6 The topography of the Earth's surface

The Earth is a spheroid which is slightly distorted by rotation. This distortion amounts to approximately 0.3% of the Earth's mean radius, with the equatorial radius being greater than the polar radius, Because of the irregularities in the Earth's surface, geophysicists have defined a surface called the geoid. The geoid is a surface of equipotential gravity. The height of this surface varies with position relative to the centre of mass of the Earth but *along* the surface there is no gravitational force, i.e. the direction of the gravity force is everywhere perpendicular to the equipotential surface.

The ocean surface would be an equipotential surface if winds; tides; currents and density variations did not produce deformations in its surface. Oceanographers, therefore, have defined their reference level in terms of mean sea level, after the removal of tides and meteorological influences. The mean sea level has been obtained by long-term observations of the sea level at coastal stations around the world. With reference to the mean sea level, the continental surface has a mean altitude of 840 m, whilst the mean depth of the ocean is 3800 m. This implies that the average height of the Earth's crust is 2440 m below sea level.

The relative topography of the Earth's surface is depicted in Figure 1.9, where it is noted that the frequency distribution has two maxima; one corresponding to the mean height of the continents and the second corresponding to an ocean depth of approximately 4500 m. This diagram shows the most frequent height of the Earth's crust corresponds to the ocean bottom. The ocean occupies 70.8% of the Earth's surface, and 76% of this ocean has depths between 3000 m and 6000 m.

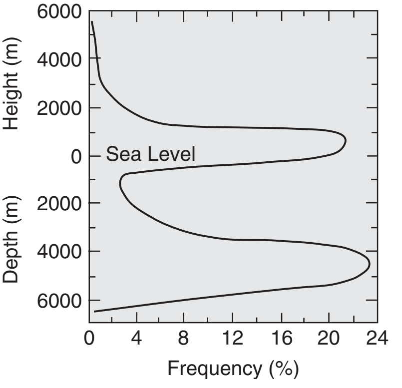


**Figure 1.9** Frequency distribution of elevation intervals of the Earth's surface. Reproduced, with permission, from McLellan, H.J., 1965, Elements of Physical Oceanography, Pergamon Press: page 5, figure 1.2

These depths are known as the abyssal depths of the ocean. Of the remaining 24% of the ocean area, 7.5% is occupied by the continental shelf, which is roughly confined to depths less than 200 m, and 15.3% is occupied by the continental slope, which is the transition region between the continental shelf and the abyssal ocean. Only 1.1% of the ocean has depths greater than 6000 m and only 0.1% has depths greater than 7000 m. This small percentage is associated with ocean trenches, whose depths plunge down to 11 500 m in the Mindanao Trench.

The relative distributions of land and ocean as a function of latitude are shown in Figure 1.10. It is again noted that the land/sea distribution is asymmetrical, with over 80% of the total global land area being located in the Northern Hemisphere and 63% of the global ocean area in the Southern Hemisphere. The continental land masses of North America and Eurasia contribute to the dominance of land over ocean between 45$^{o}$ and 70$^{o}$N. At corresponding latitudes in the Southern Hemisphere are regions where land occupies less than 3% of the total area, with essentially no land between 55$^{o}$ and 60$^{o}$S. Poleward of 70$^{o}$ latitude, the situation is reversed, with the Northern Hemisphere being occupied by the Arctic Ocean and the Southern Hemisphere being occupied by the continent of Antarctica. It is believed that the presence of a continent in a polar region is a significant factor in causing glacial cycles over the past 60 million years. An interesting property of the land/ocean distribution is that 95% of all land points have antipodes in the ocean.

The distribution of the major topographic features of the globe is shown in Figure 1.11. The most extensive feature is the continental slope or escarpment which extends for more than 300 000 km, with an average slope of 4$^{o}$. The continental slope marks the geological boundary between the older, relatively thick granitic continental crust and the young, basaltic ocean crust.

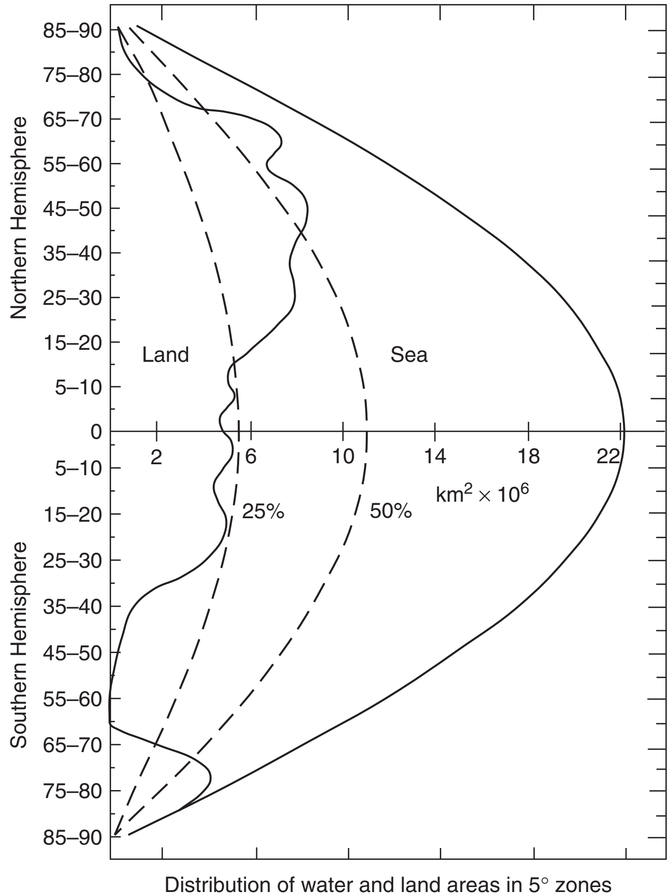


**Figure 1.10** Distribution of water and land areas in 5° zones for the Northern and Southern Hemispheres. Reproduced, with permission, from McLellan, H.J., 1965, Elements of Physical Oceanography, Pergamon Press: page 4, figure 1.1

The second most important feature is the mid-ocean ridge system which has a length of approximately 75 000 km. The typical width of the ridge is about 500 km and the height ranges from 1 to 3 km above the abyssal depths. The ridges are regions of newly formed ocean crust, with the crust moving outwards from the central part of the ridge system. A rift valley approximately 20–50 km wide, along the central part of the system, is the region of major crust formation and it is frequently associated with volcanic and seismic activity. The ocean ridges are of importance in the movement of deep water masses, because they split the oceans into smaller ocean basins.

In the Atlantic Ocean, the eastern abyssal basins are separated from the western abyssal basin by the mid-Atlantic ocean ridge. Connections between the basins, however, occur in the fracture zones which are orientated

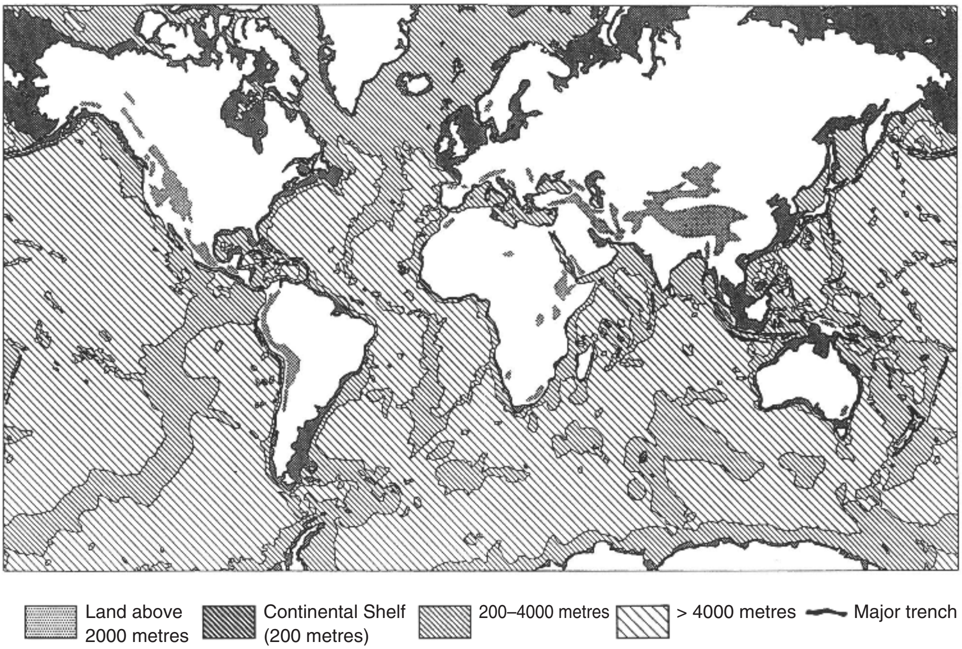


**Figure 1.11** The major topographic features of the ocean and the continents

perpendicular to the axis of the ridge. Two important fracture zones in the Atlantic Ocean are the Romanche gap, close to the equator, which allows Atlantic Bottom Water to flow into the eastern basins, and the Gibbs fracture zone at about 53°N, which allows the North Atlantic Deep Water to flow westward into the Labrador basin. Furthermore, the ridge system has an influence on oceanic flows above the level of the ridge topography and in some regions, particularly in the Southern Ocean, its effect extends to the ocean surface currents.

The relatively low relief of the land masses has already been noted and only 11% of the land area is above 2000 m. However, there are two significant features associated with Tertiary mountain building that are worthy of further comment. These are the Western Cordillera of the American continent, which has an altitude of over 6 km above sea level, and the Himalaya mountain chain, with altitudes in excess of 8 km. These mountain ranges are very important in their influence on the general circulation of the atmosphere. Although the atmosphere extends well beyond the highest peaks, 50% of the atmospheric mass occurs below 5 km, and therefore these large mountain chains have a major disruptive effect on the flow of the lower atmosphere. The Western Cordillera has a predominantly north-south orientation and this disrupts the westerly flow for over 5000 km downstream. The Himalayas have an east-west orientation which results in an intensification of the zonal flow and an inhibition of the northward penetration of the Asian monsoon. Not only is the flow directly affected by the mountain barriers but, because most of the water vapour in the atmosphere is confined to the lowest 5 km, the geographical distribution of the hydrological cycle is also significantly affected. Furthermore, the release of latent heat by condensation into the atmosphere provides a major source of energy for the general circulation of the atmosphere and the mountain chains will influence the location of heat sources in the atmosphere and the associated low-level atmospheric flow.

A feature worthy of note is the comparison of the horizontal scales of the ocean and the atmosphere with their vertical scales. The ocean basins have horizontal dimensions of 5000 km in the Atlantic Ocean to over 15 000 km in the Pacific Ocean, whilst the average depths are approximately 3.8 km. Thus the vertical scale is small compared with the horizontal scale and the ratio between them, known as the aspect ratio, is at most 1 part in 1000 and, more typically, 1 part in 10 000. The atmosphere has 99% of its mass between the sea surface and a height of 30 km, and compared with the circumference of the globe, the aspect ratio is again less than 1 in 1000. Thus both atmosphere and ocean are very thin when compared with their horizontal dimensions.

It has been shown that the topography of the land masses has a significant effect on the circulation of the lower atmosphere. In a different way, the shape and size of the ocean basins have profound effects on the circulation of water masses. Except for the region lying between 50° and 60°S, the ocean is bounded by continental land masses and therefore zonal (i.e. west-east) flows are restricted to the Southern Ocean. The presence of continental margins produces intense meridional (i.e. north-south) flows in the western regions of the basins, as evidenced by the Gulf Stream in the North Atlantic and the Kuroshio current in the North Pacific. All the ocean basins are connected via the circumpolar flow of the Southern Ocean. In contrast, the flow into the Arctic Ocean from the Pacific Ocean is inhibited by the shallowness of the Bering Sea, whilst from the Atlantic there are ‘sills’ which control the flow between Greenland and Iceland, and between Iceland and Scotland. A sill is a relatively shallow region which connects either two deep ocean basins or a deep marginal sea with an ocean basin.

There is a deep-water connection between the Indian Ocean and the Pacific Ocean, through the Indonesian Archipelago, which has important influences on the flow around the Australian continent, and on the exchange of surface and intermediate water masses between these two ocean basins.

The width of the ocean basins has a marked effect on the ocean-atmosphere interaction. The relatively narrow North Atlantic Ocean is exposed to drier air from the surrounding continents than the Pacific Ocean, and therefore evaporation is approximately twice as large in the North Atlantic Ocean as in the Pacific Ocean. This high evaporation rate is a factor in the larger heat loss from the North Atlantic Ocean than from the Pacific Ocean, and hence dense, cold water masses are formed in the North Atlantic Ocean rather than in the Pacific Ocean. Thus, the deep-water vertical circulation in the North Atlantic Ocean is much more intense than the sluggish vertical circulation observed in the Pacific Ocean. It will be shown in Chapter 5 that this has consequences for the heat transport in the ocean. In the context of evaporation, the marginal seas, such as the Mediterranean Sea and the Red Sea, are important regions for the formation of dense water masses because of their high evaporation rates.

The continental shelf, though occupying a relatively small area of the ocean, is of considerable importance both to human kind and to life in the ocean. The intense flows, associated with winds and tides in the continental shelf region, result in the mixing of the nutrients from continental shelf sediment into the water column. These nutrients, in conjunction with readily available oxygen and sunlight in the shallow sea areas, produce an abundance of phytoplankton. These regions of high primary production are very important in the maintenance of marine food chains and contain the most intensive fisheries in the world. The sedimentation rates on the continental shelves are also high, being of the order of 1 m per 1000 years in comparison with 0.1–10 mm per 1000 years in the deep ocean. Much of the sediment deposited is derived from river-borne material and this contains large quantities of phosphorus which is an element necessary for the maintenance of productivity in the ocean.

In conclusion, it has been shown that the surface topography and the shape and interlinkage of the ocean basins have a significant effect on the circulation of both the atmosphere and ocean. This, in turn, affects the biological productivity of the land and ocean. In the geological past, different ocean geometry and surface topography would have been partly responsible for the many varied climatic patterns that have occurred since the formation of the Earth.