

1 *The climate system*

Traditionally climate was defined as the average atmospheric state over at least a score of years, modulated by the seasonal cycle.

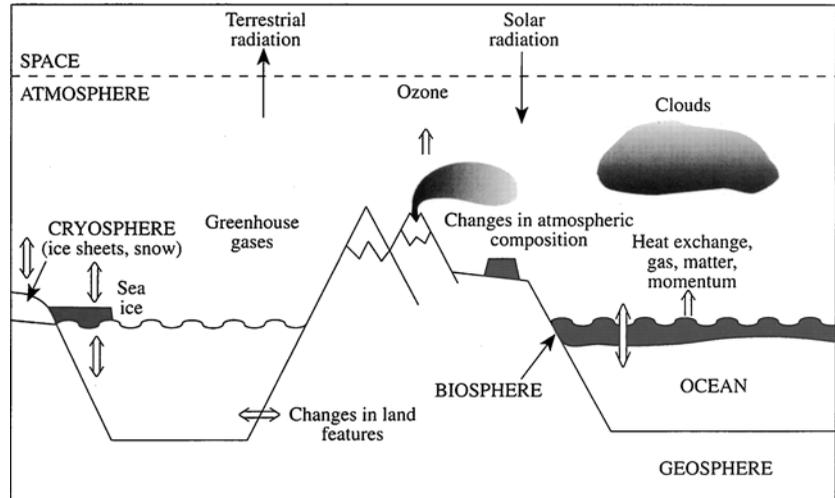
Such a definition concealed the temporal variability which produced the mean state, and the complexity of the underlying physical, chemical, biological, geographical and astronomical processes contributing to the climate system. It also implicitly assumed that the climate of a locality does not change over decades, while accepting long-term changes such as glacial periods. In recent decades this perspective of climate has changed. The strong coupling of different constituents of the climate system is now widely recognized, as is the fragility of 'stable' climate.

The traditional definition of climate contains two elements which lead us towards the concept of climatic change. First, the seasonal cycle demonstrates one scale of change through the direct impact of the annual change in solar radiation on the atmosphere and the *biosphere* (the collective term for plant and animal life). Contrastingly, the implication of stability over a number of years implies that basic balances exist within the system. It also implies that any decadal changes in the surface forcing of the atmosphere, from the land or ocean, are small; Chapters 5 and 6 will show that this is not always the case but in general this is a reasonable claim.

In recent decades concern over potential climatic changes due to humanity's activities has arisen, although acknowledgement of the possibility was first made early in the nineteenth century. This book will explore the natural climate system, and potential changes, man-made (*anthropogenic*) or otherwise. Its dominant theme will be the contribution of the oceans to these processes. Observable changes to the climate due to anthropogenic inputs could be expected to have already occurred; Chapters 6 and 7 will show that recent climatic shifts are, by the year 2002, increasingly likely to be largely attributable to these inputs. We will see that numerical climate models are an important means towards this recognition. The stability of the ocean's thermal and freshwater responses to change, its absorption of a significant proportion of anthropogenically derived compounds from the atmosphere, and its coupling with the atmosphere, form important pieces of the puzzle of climatic change.

To understand how the ocean affects, and is affected by, the climate we need to briefly consider the climate system as a whole. It is a complex, many-faceted system; Fig. 1.1 illustrates its major constituents and interactions. There are five components: the atmosphere, the ocean, the cryosphere (ice sheets, transient snow and sea-ice), the biosphere, and the geosphere (the solid earth). The system

Fig. 1.1. A schematic diagram of the climate system. [From Bigg, 1992d]



is driven by short wavelength, principally visible and ultra-violet, solar radiation, with longer wavelength, infra-red, radiation being emitted into space to maintain thermal equilibrium. In the rest of this chapter we will briefly examine the different components of the climate system, including the timescales on which it operates and the principles behind the numerical models that increasingly are required to study it. However, we will begin with a short discussion of the basic energy source driving the climate – radiation from the Sun.

1.1 Solar radiation

The interior of the Sun, where the nuclear reactions occur that ultimately lead to life on Earth, is incredibly hot, at a temperature of several million degrees Celsius. However, the *electromagnetic radiation* (see Appendix A) that provides the energy for the climate system is derived from the outer layers of the Sun. The greatest amount of radiation comes from the *photosphere*, a layer some 300 km thick in the solar atmosphere. This varies in temperature from $10\,000\text{K}^1$ at the bottom to 5000K at the top. Outside the photosphere are much less dense regions – the *chromosphere* and *corona*. While these outer regions are at much higher temperatures, up to millions of degrees in the corona, their low density means that they radiate relatively little energy. Most of this is at very short, X-ray and gamma-ray wavelengths which affect the upper atmospheres of the planets (see §3.7 – Carbon 14) but do not penetrate into the lower atmosphere.

The Sun appears to us as (almost) a *black body*. That is, the *spectrum* and total energy of electromagnetic radiation emitted from the Sun (as from all surfaces, and indeed molecules) is a function of its temperature. The total energy flux, E , emitted by a black body follows the *Stefan–Boltzmann Law*:

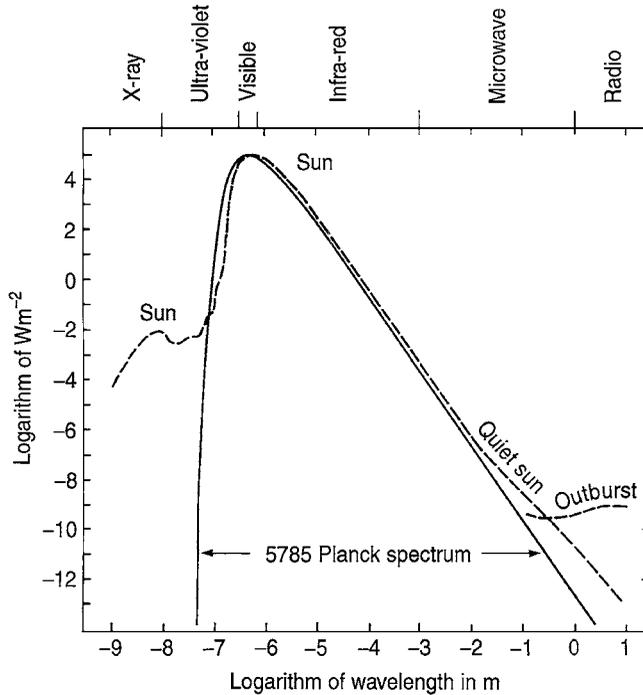
$$E = \sigma T^4 \quad (1.1)$$

¹ The absolute scale of temperature is in degrees Kelvin (K). In this scale 0K is the coldest possible temperature when all molecular motion has stopped. The freezing point of water, 0°C , is 273.16K in this scale. Note, however, that a change of 1K is equivalent to a change of 1°C .

Solar radiation

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Fig. 1.2. The Sun's spectrum, seen from space (broken line). Both scales are logarithmic. For comparison, a Planck spectrum for a temperature of 5785K is shown (solid line). Note the accentuation of long and short wavelength energies emitted by the Sun, particularly during solar flares.



where σ is the Stefan–Boltzmann constant and T is the temperature in degrees Kelvin (a list of constants and their values can be found in Appendix A). The energy density, E_λ or radiant energy per unit wavelength, λ , per unit volume per second, is given by

$$E_\lambda = \frac{8\pi c}{\lambda^5} \left[\frac{1}{e^{hc/\lambda kT} - 1} \right] \quad (1.2)$$

where c is the speed of light, k is Boltzmann's constant and h is Planck's constant. The Sun's spectra, as observed from space (Fig. 1.2), obeys (1.2) for a temperature near 6000K. However, for very small (X-ray) and very long (microwave) wavelengths the solar spectrum is enhanced due to contributions from the outer regions of the solar atmosphere (see §§3.7 and 7.1.1).

The vast majority of the energy that reaches the Earth comes from the ultra-violet through visible to infra-red part of the spectrum. The peak energy is in the visible, near wavelengths that we see as the colour blue. The variation in the amount of energy emitted by the Sun is probably small on non-geological time scales. At the Earth's distance from the Sun this *solar constant* is about 1.38 kWm^{-2} . On very long time scales, comparable with the life of the planet, astrophysicists believe that the Sun's irradiance varies dramatically as the supply of fuel within the Sun changes. We will see in §1.8 that variation in the Earth's orbit can affect the amount of energy reaching the Earth's surface by a few per cent, on time scales of thousands of years. However, over several decades to centuries solar irradiance is thought to vary by significantly less than this. Satellite measurements extend back only to 1978 and these reveal irradiance changes of only 0.08% between *sunspot* maxima (higher) and minima (lower).

This does not, however, preclude larger changes in more active beats of the 11 year solar cycle, or the existence of frequencies in the Sun's behaviour of which we have only a dim perception (see §7.1.1).

1.1.1 *The effective temperature of the Earth*

If the Earth was a sterile planet like the Moon, with no atmosphere, oceans or biosphere what temperature would we expect the surface to possess, given the solar constant, S , at the Earth's astronomical position? If we think of the Earth as a flat disc, viewed from the Sun, then the surface area illuminated by solar radiation is πr^2 , where r is the radius of the Earth. The energy absorbed is thus $(1 - a)S\pi r^2$, where a is the *albedo*, or the proportion of the Sun's energy reflected from the Earth back into space (c. 30%).² For equilibrium between the absorbed solar radiation and the emitted radiation from the whole Earth's surface of area $4\pi r^2$, the Earth's temperature, T_E , will therefore, from (1.1), be

$$T_E = \left[\frac{(1 - a)S}{4} \right]^{0.25} \quad (1.3)$$

Equation (1.3) gives a surface temperature for this hypothetical atmosphere-less planet of 255K, or -18°C , much colder than the Earth's average surface temperature of about 16°C . This effective planetary temperature is more typical of the real atmospheric temperature at a height of about 6 km above the surface. The atmosphere clearly has a significant impact on the distribution of the energy contributing to this effective temperature and will thus be the first component of the climate system to be considered.

1.2 The atmosphere

The atmosphere is a largely homogeneous mixture of gases, both horizontally and vertically, over the height range important for climate: namely the *troposphere* and *stratosphere* (Fig. 1.3). The composition of this apparently stable mixture, air, is shown in Table 1.1. The balance of the dominant constituents of air is thought to have evolved considerably over the lifetime of the planet; for instance, oxygen is likely to have been a product, rather than a necessity, of life (the *Gaia hypothesis*). The climate system's immense natural variability will be a recurring theme of our discussion throughout the book.

The temperature of the atmosphere varies strongly both in the vertical and with latitude. The latter is due to an imbalance in the radiation received over the Earth's surface throughout the year due to the planet's orbit and *obliquity* (Fig. 1.4 and §1.8). The circulation of both the atmosphere and ocean are ultimately derived from this energy imbalance; they act to counter it, in the ratio of about 3:2 respectively.

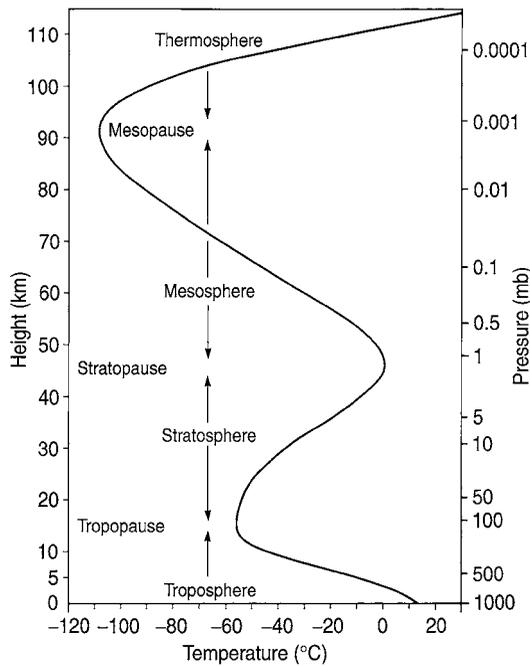
The vertical temperature distribution shown in Fig. 1.3 comes about because the atmosphere is basically heated from two sources: the ground and the upper

² Note that this average planetary albedo assumes that the hypothetical, sterile, Earth has the same net reflectivity as the real Earth-atmosphere system. Thus this albedo is *not* the surface reflectivity (see §1.4).

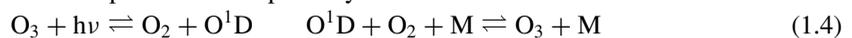
Table 1.1. *The major constituents of the atmosphere*

Gaseous constituent	Molecular form	Proportion (%)
Nitrogen	N ₂	78.1
Oxygen	O ₂	20.9
Argon	Ar	0.93
Water vapour	H ₂ O	variable: 0.1–1
Carbon dioxide	CO ₂	0.037
Methane	CH ₄	0.000175
Nitric oxide	N ₂ O	0.000032
Ozone	O ₃	variable c. 0.000005

Fig. 1.3. Zonal mean vertical profile of temperature during June at 45°N.

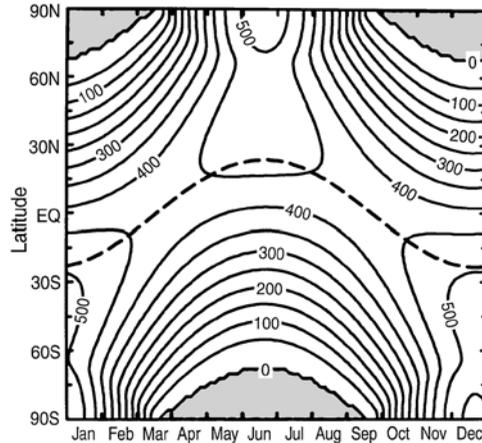


stratosphere (although we will see in the next sub-section that this is a significant simplification). The ground (or ocean surface) is a heat source since some 49% of the incoming solar radiation is absorbed there. There is also an important heat source between 30 and 50 km above the ground, in the *ozone layer*. When ozone, O₃, absorbs a photon of ultra-violet light – denoted $h\nu$ because this is the energy of a photon of frequency ν – the molecule dissociates in the process to form an oxygen molecule and an energetic oxygen atom, O¹D, where one electron is displaced into a higher energy state than in the ordinary oxygen atom (see Appendix B for the Periodic Table of the Elements and a discussion of electron orbitals). This can then react with an oxygen molecule to reform ozone as part of the Chapman cycle:



The air molecule, M (that is, predominantly N₂ or O₂), is necessary in the second reaction in (1.4) as the reaction produces excess energy. This is carried

Fig. 1.4. Contour plot of daily average insolation at the top of the atmosphere as a function of season and latitude. The contour interval is 50 Wm^{-2} . The heavy broken line indicates the latitude of the sub-solar point at noon. [Fig. 2.6 of Hartmann (1994), *Global Physical Climatology*. Reprinted with permission from Academic Press.]



away by M thereby stabilizing O_3 , which would otherwise dissociate. Such a reaction is called *exothermic*. Hence the air becomes warmer, as the temperature of a medium is merely a reflection of the average kinetic energy of its molecules.

The reactions in equation (1.4) are only part of the full Chapman cycle, which also contains reactions involving photo-dissociation of O_2 , and reactions between the excited oxygen atoms themselves and O_2 . Some of these are likewise exothermic, adding to the energy which is transferred, via chemical reaction, from solar radiation to the middle atmosphere. There are many other reactions involving ozone, some of which will be discussed in §7.2.1.

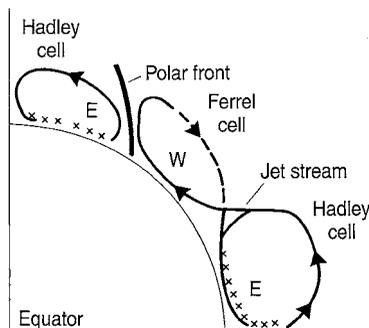
The lower atmosphere is therefore heated both from above and from below. Between these regions is a zone, in the lower stratosphere, where the energy from these heated regions only weakly penetrates. This is strongly stratified, which means that there are large vertical gradients in the concentrations of trace constituents of the air and the *potential temperature* (see Appendix C). The tropopause, at the bottom of the stratosphere where the gradients are greatest, resists penetration by cloud convection, or even diffusion. The well-mixed region below this, the troposphere, is the part of the atmosphere that we will be largely concerned with, because of its direct interaction with the oceans.

The strong heating of the surface at the equator (Fig. 1.4) makes the air less dense, forcing it to rise. Air flows towards this region of rising air, which tends to be concentrated in a narrow band around the globe known as the Inter-Tropical Convergence Zone, or ITCZ. Aloft, the rising air moves polewards to compensate for the surface flow. In the late seventeenth century, when Halley first proposed this mechanism for driving the atmospheric circulation (modified 50 years later by Hadley) it was believed that this *Hadley cell* extended to the polar regions. This seemed logical, as polar air is cold, and so relatively dense, and should therefore flow towards the low pressure regions of the tropics in order to transfer heat from the equator to the poles and so maintain the Earth's thermal equilibrium. By the nineteenth century this idea was seen to be too simplistic. The *Coriolis force*, due to the solid Earth and moving atmosphere revolving at slightly different rates (see §2.5.2), gives this converging near-surface wind a

The atmosphere

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Fig. 1.5. Schematic cross-section of the zonal mean circulation in the troposphere. The dotted upper region of the Ferrel cell indicates that it is a less constant feature. The crosses (and E) show where the surface flow has an easterly component, while W shows where the surface flow is predominantly westerly.



westward component, resulting in the observed easterly³ Trade winds. Ferrel therefore proposed an intermediate ‘Ferrel cell’ in mid-latitudes. Modern observations support this, as shown schematically in Fig. 1.5, a zonal cross-section of the tropospheric flow.

The general circulation of the lower troposphere is shown in Fig. 1.6, and the sea level pressure field for northern winter in Fig. 1.7. The ascending air of the equatorial region is shown by the low pressure. To replace this, easterly winds flow equatorwards driven by high pressure in the sub-tropics, where the air in the tropical Hadley cell descends (Fig. 1.5). This latter, relatively calm, zone has strong westerly winds on its poleward side, which, in turn, lie equatorward of another region of low pressure near 60° of latitude. This region of sub-polar low pressure forms the ascending branch of the polar Hadley cell of Fig. 1.5, with easterly winds at the surface due to polar high pressure. The most vigorous part of this system is where the tropical Hadley cell meets the mid-latitude Ferrel cell. Here the upper level convergence of air produces an extremely strong westerly *jet-stream* in the upper troposphere (Fig. 1.5). This often has a secondary maximum over the mid-latitudes, above the polar front (Fig. 1.5). This polar front jet-stream steers the transient pressure systems that we experience on the ground in the mid-latitudes. The latter systems are a significant mechanism in the redistribution of heat from equator to pole.

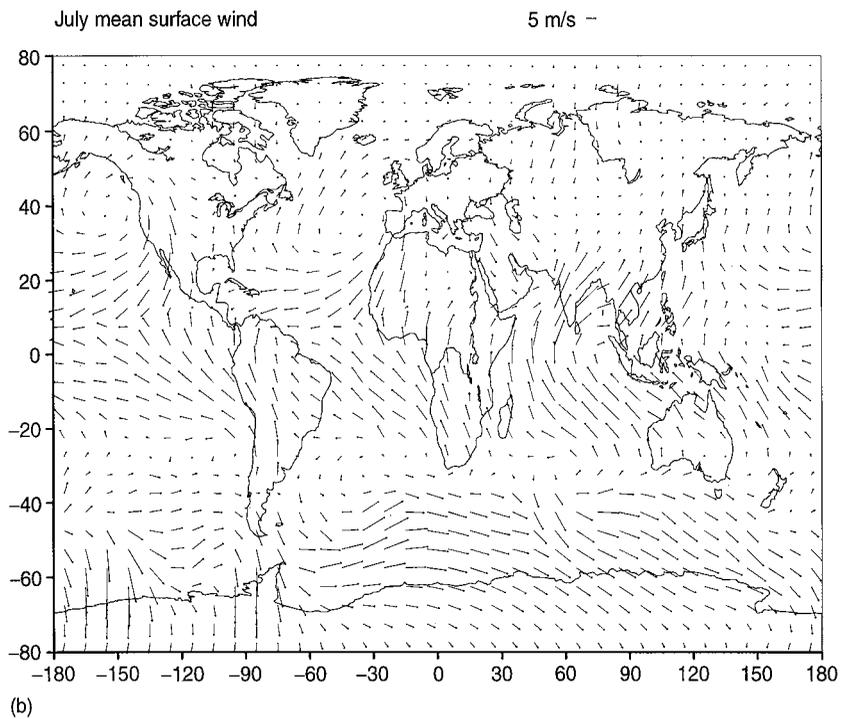
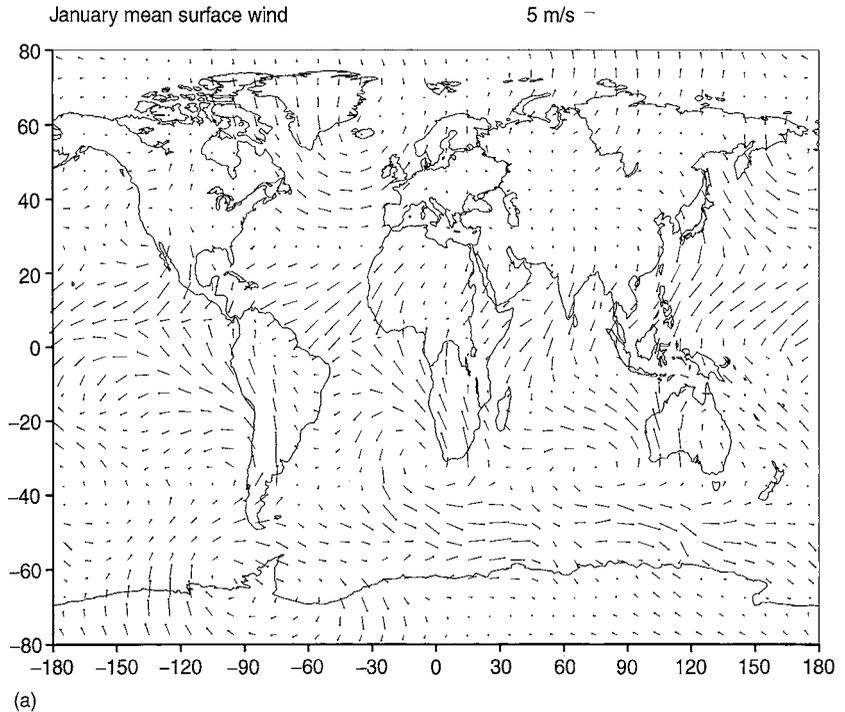
The mainly zonally symmetric structure of the general circulation is mostly due to the latitudinal distribution of the solar radiation received by the Earth; the distribution of land and sea over the Earth’s surface distorts the zonality. Some aspects of this latter interaction will be discussed in Chapters 2 and 5.

1.2.1 The greenhouse effect

The vertical profile in Fig. 1.3 shows a decline in temperature of about 6.5°C per kilometre in the troposphere. It can be shown that ascending ‘dry’ air, i.e. air without clouds, changes temperature because of expansional cooling by 9.8°C

³ Confusingly, meteorologists and oceanographers follow different conventions when specifying the direction of fluid flow. Meteorologists use the direction from which the wind has come to describe it, while oceanographers take the direction in which the flow is going. Thus, an easterly wind to a meteorologist is a westward wind to an oceanographer! This unfortunate difference is too entrenched to be easily altered, and this book will use the convention appropriate to the fluid medium being described.

Fig. 1.6. Mean surface wind field in (a) January, and (b) July. [Data from Oort, 1983.]



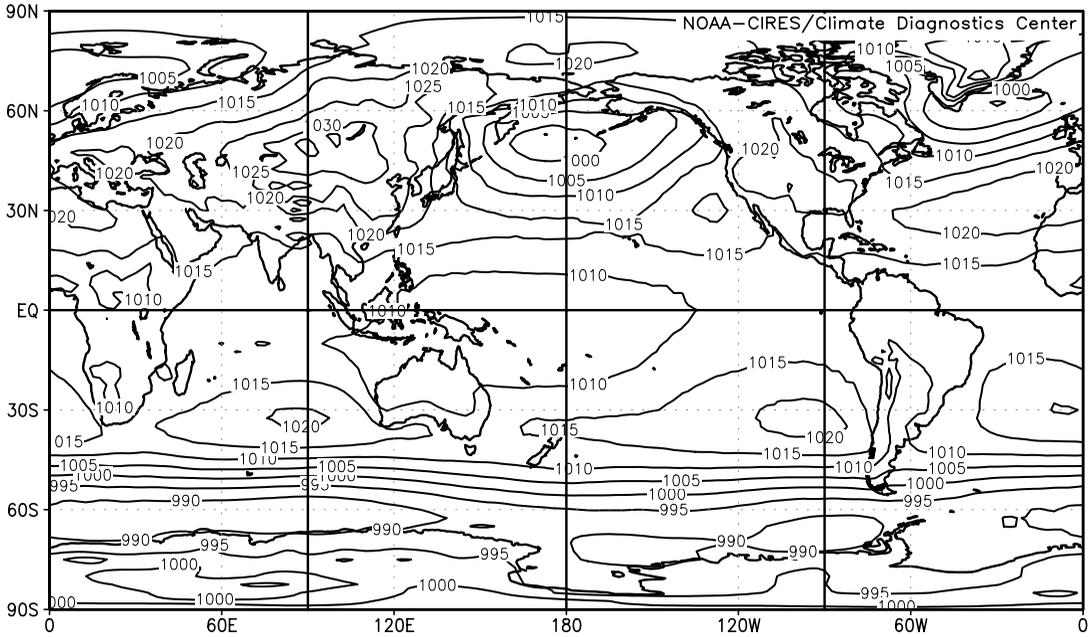


Fig. 1.7. Mean sea level pressure field in January. Contours are every 5 mb. The data is a mean of 17 years of National Meteorological Center model analysis fields.

for each kilometre of *adiabatic* vertical motion (the latter occurs if a parcel of air does not exchange any heat with its surroundings, as is a good approximation in, for example, the formation of *cumulus* clouds – see Appendix C). Within a cloud the decline of temperature with height in vertical motion can approach the typical value of Fig. 1.3, due to the release of *latent heat* upon condensation of water vapour. However, substantially less than half of the troposphere contains cloud at any one time so other processes must be lowering the *environmental lapse rate* below the *dry adiabatic lapse rate*. Diffusion and advection of heat from the stratosphere, the ground, or surrounding air masses is partially responsible but the major reason for the enhancement of tropospheric temperatures is the *greenhouse effect*.

A number of low concentration, or *trace*, gases in the atmosphere are unresponsive to illumination by short wavelength radiation from the Sun but absorb energy of infra-red wavelengths. The gas molecules do this by increasing their *vibrational* and *rotational* energies, rather than their kinetic energy. How this happens can be illustrated by the water molecule, shown in Fig. 1.8. The bond angle between the hydrogen atoms of an ordinary water molecule is 105° , but if a *photon* of a certain wavelength of infra-red radiation ($6.27 \mu\text{m}$)⁴ collides with the molecule the energy of the photon can be converted into a vibration of the hydrogen bonds, such that the angle between the hydrogen atoms undergoes rapid oscillation of a few degrees. Other forms of oscillation can be excited by wavelengths of 2.66 or 2.74 μm . The *absorption spectra* of H_2O , shown in Fig. 1.9, is more complex than just these three wavelengths, however,

⁴ 1 μm (micrometre, often called a micron) = 10^{-6} m.

Fig. 1.8. Schematic diagram of a water molecule. H represents a hydrogen atom and the central O an oxygen atom. Solid lines show bond positions.

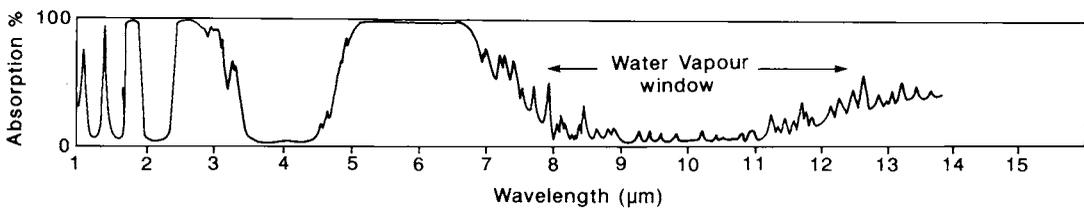
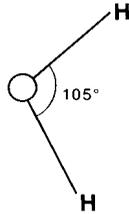


Fig. 1.9. The absorption spectrum of water vapour. Note the region 8–12 μm , known as the 'water vapour window', where there is little absorption of infra-red radiation by the water vapour molecule.

as multiples, or *harmonics*, of the principal absorption wavelengths can also be absorbed. In addition wavelengths which are sums, or differences, of these three (and their harmonics) also show a degree of absorption, although generally of reduced intensity.

Equation (1.2) shows that the wavelength of electromagnetic radiation emitted by an object is inversely related to its temperature. Thus the mean wavelength of the radiation emitted by the Earth's surface, and within the atmosphere itself, will be longer than that of the incoming radiation from the Sun, as the latter has a surface temperature of about 6000K compared to a typical Earth surface temperature of 289K. Fig. 1.10 depicts a typical energy spectrum, seen from the tropopause, of the radiation from the Earth's surface, with the absorption by trace gases shown by shading. There are regions of the spectrum, such as wavelengths shorter than 8 μm and from 15 to 20 μm , where the infra-red radiation is almost totally absorbed by atmospheric gases. It is this absorption, and the associated re-emission of energy, much of which warms the troposphere, that is called the greenhouse effect. This name is a misnomer as the physical mechanism involved in keeping a greenhouse warm is totally different from this radiative physics. There is a small contribution from glass being transparent to solar radiation, but partially reflective to the outgoing infra-red radiation from the air and soil within the greenhouse. However, greenhouses are warm predominantly because the enclosed space eliminates convection, and hence mixing with cooler air.

The principal greenhouse gases, and their relative contribution to the greenhouse effect, are shown in Table 1.2. The percentages shown are not strictly additive because the absorption ranges of the different gases overlap. Table 1.2 also gives the fundamental absorption wavelengths of these molecules, but the complexity of the absorption spectra, with their harmonics and linear combinations of these fundamental wavelengths, must be remembered (see Fig. 1.9). Water vapour is two to three times as important in the total greenhouse effect as carbon dioxide. This fact is often neglected in discussions of greenhouse warming because water vapour is highly variable in concentration, both in space and