1 Introduction

This chapter gives a quick sketch of some of the material to be covered in this book. We start in Section 1.1 with an outline of some of the more important physical processes that occur in the Earth's atmosphere. To interpret atmospheric observations we need to develop physical and mathematical models; they are briefly discussed in Section 1.2. Two extremely simple models are introduced in Section 1.3: the second of these includes a very basic representation of the greenhouse effect. In Section 1.4 we present a selection of observations of atmospheric processes, together with simple physical explanations for some of them. In Section 1.5 we briefly mention some ideas on weather and climate.

1.1 The atmosphere as a physical system

The Earth's atmosphere is a natural laboratory, in which a wide variety of physical processes takes place. The purpose of this book is to show how basic physical principles can help us model, interpret and predict some of these processes. This section presents a brief overview of the physics involved.

The atmosphere consists of a mixture of ideal gases: although molecular nitrogen and molecular oxygen predominate by volume, the minor constituents carbon dioxide, water vapour and ozone play crucial roles. The forcing of the atmosphere is primarily from the Sun, though interactions with the land and the ocean are also important.

The atmosphere is continually bombarded by solar photons at infra-red, visible and ultra-violet wavelengths. Some solar photons are scattered back to space by atmospheric gases or reflected back to space by clouds or the Earth's surface; some are absorbed by atmospheric molecules (especially water vapour and ozone) or clouds, leading to heating of parts of the atmosphere; and some reach the Earth's surface and heat it. Atmospheric gases (especially carbon dioxide, water vapour and ozone), clouds and the Earth's surface also emit and absorb infra-red photons, leading to further heat transfer between one region and another, or loss of heat to space. Some of these **radiative-transfer** processes are discussed in Chapter 3. Solar photons may also be energetic enough to disrupt molecular chemical bonds, leading to photochemical reactions; see Chapter 6.

The atmosphere is generally close to hydrostatic balance in the vertical, except on small scales; that is, the weight of each horizontal slab of atmosphere is supported by the difference in pressure between its lower and upper surfaces. An alternative statement of this physical fact is that there is a balance between vertical pressure gradients and the gravitational force per unit volume acting on each portion of the atmosphere. On combining the equation describing hydrostatic balance with the ideal gas law we find that, in a hypothetical isothermal atmosphere, the pressure and density would fall exponentially with altitude (see Section 2.3). In the real, non-isothermal, atmosphere the pressure and density variations are usually still close to this exponential form, with an e-folding height of about 7 or 8 km. Gravity thus tends to produce a **density stratification** in the atmosphere.

Given a density stratification of this kind, a small portion of air that is displaced upwards from its equilibrium position will be negatively buoyant compared with its surroundings and hence will fall back towards equilibrium, under gravity; similarly a downward-displaced portion will rise back towards its equilibrium position. Buoyancy therefore acts as a restoring force; the atmosphere is said to be **stably stratified**. The strength of the stability of the stratification varies from one part of the atmosphere to another.

Thermodynamic principles are essential for describing many atmospheric processes. For example, any consideration of the effects of atmospheric heating or cooling will make use of the First Law of Thermodynamics. The concept of entropy (or the closely related quantity, potential temperature) frequently assists interpretation of atmospheric behaviour. Knowledge of changes in phase between vapour, liquid and solid forms of the water in the atmosphere is crucial for an understanding of the formation of rain and snow. Moreover, the associated latent heating and cooling can provide important contributions to heat transfer within the atmosphere–ocean system; for example, evaporation of a droplet of sea water at one location and subsequent condensation of the resulting water vapour into a droplet at another location in the atmosphere transfers heat from the ocean to the atmosphere. The basics of atmospheric thermodynamics are covered in Chapter 2.

In atmospheric physics we use the usual macroscopic definitions of the temperature and pressure of a gas. From the kinetic theory of gases, these have well-known interpretations in terms of the mean kinetic energy of molecules and the mean transfer of momentum by molecules, respectively. When considering dynamical processes – that is, the response of atmospheric motions to applied forces – we can average other physical quantities such as density and velocity over many molecules and regard the atmosphere as a continuous fluid; individual molecular motions need not be taken into account. It is clear from the most cursory weather observations that the resulting bulk fluid motion of the atmosphere is still very complex. However, when the motion is viewed on a large scale (say hundreds of kilometres in horizontal extent), some simplifying features appear. In particular, Coriolis forces play significant roles: these forces result from the rotation of the Earth and tend to deflect a moving portion of air to the right of its motion in the Northern Hemisphere and to the left in the Southern Hemisphere. A near balance between Coriolis forces and horizontal pressure gradient forces leads to wind motions that circulate along isobars (surfaces of constant pressure) at a given height. The sense of the circulation is anticlockwise around low-pressure regions and clockwise around high-pressure regions in the Northern Hemisphere and vice versa in the Southern Hemisphere. The basic principles of atmospheric fluid dynamics are introduced in Chapter 4.

An important feature of the buoyancy restoring effect in a stably stratified atmosphere is that it can support fluid-dynamical waves, known as **gravity waves**,[1] in which the fluid pressure, density, temperature and velocity fluctuate together. These waves may propagate, allowing one part of the atmosphere to 'communicate' over great distances with other parts, without a corresponding transport of mass. The Coriolis force can also act as a restoring force, giving rise to further types of fluid wave motion. In particular, on large scales we find **Rossby waves**, which depend crucially on the rotation and the spherical geometry of the Earth and are associated with many observed large-scale disturbances in the troposphere and the stratosphere. As in many other branches of physics, the study of wave motions is an essential part of atmospheric physics; see Chapter 5.

As noted above, solar radiation can initiate photochemical reactions by dissociating atmospheric molecules. A host of other types of chemical reaction between atmospheric molecules, both natural and man-made, can also occur. Atmospheric chemistry is a large and highly complex subject; in this book we focus on one small but significant branch of the subject, namely the chemistry of stratospheric ozone. This provides a good example of physical principles in operation and is highly topical, with direct application to the Antarctic ozone hole and global depletion of ozone; see Chapter 6. It also demonstrates the importance of atmospheric transport processes, by which the winds blow chemical species from one part of the globe to another.

No study of the atmosphere can make progress without suitable observations, and all observational techniques rely to some extent on physical principles. One important type of observational technique is that of **remote sounding**, which depends on the detection of electromagnetic radiation emitted, scattered or transmitted by the atmosphere. In Chapter 7 we describe several examples of remote sounding, looking both at space-borne and at ground-based methods.

Climate change is a topic of great current concern. An understanding of how the Earth's climate has changed in the past, what determines its current state, and how it will change in the future depends on a detailed knowledge of a wide variety of physical processes, some of which are touched upon in this book. An outline of some of the more important physical concepts and processes associated with climate change is given in Chapter 8.

1.2 Atmospheric models

Unlike laboratory physicists, atmospheric researchers cannot perform controlled experiments on the large-scale atmosphere. The standard 'scientific method', of observing phenomena, formulating hypotheses, testing them by experiment, then formulating revised hypotheses and so on, cannot be applied directly. Instead, after an atmospheric phenomenon is discovered, perhaps by sifting through a great deal of data, we develop

[1] Not to be confused with the gravitational waves of General Relativity!

models, which incorporate representations of those processes that we hypothesise are most important for causing the phenomenon. Models act as surrogate atmospheres, on which 'thought experiments' can be performed. These models are usually formulated in terms of mathematical equations, and the 'experiments' are performed by solving these equations (perhaps by computer) under various conditions and interpreting the solutions in terms of physical behaviour. Occasionally a laboratory apparatus may provide a useful atmospheric model. The performance of the model (and thus the appropriateness of the hypothesised set of processes) is judged by comparing the model's behaviour with that of the atmosphere.

Normally, a **hierarchy** of models is used, starting with simple 'back-of-the-envelope' models and progressing through models of intermediate complexity to the highly complex 'general circulation models' which require large computer resources. The models considered in this book are mostly of the simpler type, although the more complex ones are briefly discussed in Chapter 9. Since the simpler models can usually be investigated analytically and their workings fully explored, they can provide a basic 'physical intuition', which can then be applied to the interpretation of the more complex models. Because of their very simplicity, however, they cannot usually be expected to give accurate simulations of observed atmospheric behaviour.

The more comprehensive models bring together many of the physical principles introduced in this book and allow for interactions between them. Some of these interactions may involve complex feedbacks, however, so it may be difficult to establish causal relationships among the various processes that are involved.

Historically, a major use of comprehensive atmospheric models has been for weather forecasting. With the development of supercomputers, reliable longer-term climate forecasting is also becoming a feasible proposition, although the models used for this purpose are still inadequate in some respects. Complex models are also used for **data assimilation**, by providing a dynamically self-consistent means of interpolating, in space and time, between sparse data points from a variety of sources. Data assimilation aims to provide an accurate estimate of the time-evolving, three-dimensional state of the atmosphere, and is nowadays a vital component of the weather-forecasting process.

1.3 Two simple atmospheric models

It is a basic observational fact that the Earth's mean surface temperature is about 288 K. In this section we consider whether this can be explained in simple terms, given the input of solar radiation and some elementary atmospheric physics. We consider two models; the first turns out to be seriously defective, but the second, which includes a simple representation of the **greenhouse effect**, gives a surface temperature in reasonable agreement with observations. Both models assume that radiation is the only heat-transfer process. To quantify this process we introduce the **irradiance**, i.e. the power per unit area, associated with any given stream of radiation: see Section 3.2.1 for a more precise definition.

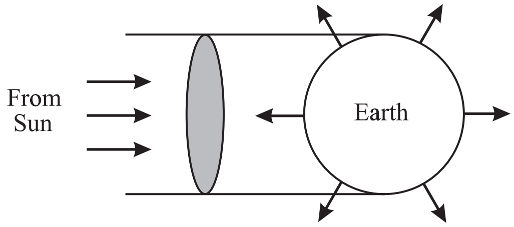

Fig. 1.1 Illustrating the calculation of the temperature of the Earth, ignoring any absorption of radiation by the atmosphere. The parallel arrows indicate solar radiation, confined within a tube of cross-sectional area πa^2. The radial arrows indicate outgoing thermal radiation from the total surface area $4\pi a^2$ of the Earth.

1.3.1 A model with a non-absorbing atmosphere

The solar power per unit area at the Earth's mean distance from the Sun (the **total solar irradiance**, TSI, formerly called the **solar constant**) is $F_s = 1370\,\mathrm{W\,m^{-2}}$. The solar beam is essentially parallel at the Earth, so the power that is intercepted by the Earth is contained in a tube of cross-sectional area πa^2, where a is the Earth's radius; see Figure 1.1. The total solar energy received per unit time is therefore $F_s \pi a^2$.

We assume that the Earth–atmosphere system has a **planetary albedo** A equal to 0.3; that is, 30% of the incoming solar radiation is reflected back to space without being absorbed: this is close to the observed mean value. The Earth therefore reflects $0.3F_s\pi a^2$ of the incoming solar power back to space.

If the Earth is assumed to emit as a **black body** at a uniform absolute temperature T then, by the Stefan–Boltzmann law,

$$\text{Power emitted per unit area} = \sigma T^4, \tag{1.1}$$

where σ is the Stefan–Boltzmann constant.[2] However, power is emitted in all directions from a total surface area $4\pi a^2$, so the total power emitted is $4\pi a^2 \sigma T^4$. We assume in the present model that all of this power is transmitted to space, with none absorbed by the atmosphere. Then, assuming that the Earth is in thermal equilibrium, the incoming and outgoing power must balance, so

$$(1 - A)F_s\pi a^2 = 4\pi a^2 \sigma T^4. \tag{1.2}$$

On substituting the values of A and F_s into this, we obtain $T \approx 255\,\mathrm{K}$. The temperature obtained from this calculation is called the **effective emitting temperature** of the Earth: see equation (3.36). Its value is significantly lower than the observed mean surface temperature of about 288 K. The present model is clearly lacking in some vital ingredient; we find in Section 1.3.2 that inclusion of the radiation-trapping effect of the atmosphere (the 'greenhouse effect') leads to a surface temperature that is much closer to reality.

[2] The concept of a black body is explained in Section 3.1.1; for the moment, all that is required is that the power per unit area emitted by a black body satisfies equation (1.1). The value of σ is given in Appendix A, together with the values of other useful physical constants.

1.3.2 A simple model of the greenhouse effect

We now consider the effect of adding a layer of atmosphere, of uniform temperature T_a, to the model of Section 1.3.1; see Figure 1.2. The atmosphere is assumed to transmit a fraction $\mathcal{T}_{sw}$ of any incident solar (short-wave) radiation and a fraction $\mathcal{T}_{lw}$ of any incident thermal (infra-red, or long-wave) radiation (these fractions are called **transmittances**: see Section 3.4), and to absorb the remainder. We assume that the ground is at temperature T_g.

Taking account of albedo effects and the difference between the area of the emitting surface $4\pi a^2$ and the intercepted cross-sectional area πa^2 of the solar beam (see Section 1.3.1), the mean unreflected incoming solar irradiance at the top of the atmosphere is

$$F_0 = \tfrac{1}{4}(1 - A)F_s, \tag{1.3}$$

or about 240 W m^{-2} with the given values of A and F_s. Of this, an amount $\mathcal{T}_{sw}F_0$ is absorbed by the ground and the remainder $(1 - \mathcal{T}_{sw})F_0$ is absorbed by the atmosphere.

The ground is assumed to emit as a black body. By equation (1.1) it therefore emits an upward irradiance $F_g = \sigma T_g^4$, of which a proportion $\mathcal{T}_{lw}F_g$ reaches the top of the atmosphere, the remainder being absorbed by the atmosphere. The atmosphere is not a black body, but emits irradiances $F_a = (1 - \mathcal{T}_{lw})\sigma T_a^4$ both upwards and downwards, as shown in Figure 1.2. (By Kirchhoff's law, the emittance – the ratio of the actual emitted irradiance to the irradiance that would be emitted by a black body at the same temperature – equals the absorptance $1 - \mathcal{T}_{lw}$; see Section 3.1.1.)

We now assume that the system is in **radiative equilibrium**: that is, energy transfer takes place only by the radiative processes described above, and the associated irradiances are in balance everywhere; we neglect any energy transfers due to non-radiative processes such as fluid motions. Equating irradiances, we have

$$F_0 = F_a + \mathcal{T}_{lw}F_g \tag{1.4a}$$

above the atmosphere, and

$$F_g = F_a + \mathcal{T}_{sw}F_0 \tag{1.4b}$$

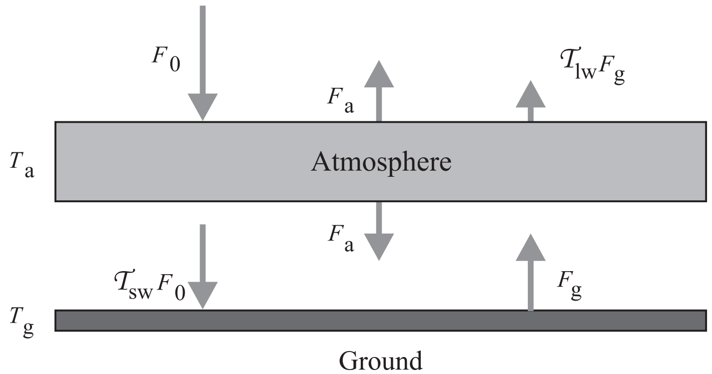

Fig. 1.2 A simple model of the greenhouse effect. The atmosphere is taken to be a layer at temperature T_a and the ground a black body at temperature T_g. Various solar and thermal irradiances are shown.

between the atmosphere and the ground. By eliminating F_a from equations (1.4) we obtain

$$F_g = \sigma T_g^4 = F_0 \frac{1 + \mathcal{T}_{sw}}{1 + \mathcal{T}_{lw}}. \tag{1.5}$$

In the absence of an absorbing atmosphere, we would have $\mathcal{T}_{sw} = \mathcal{T}_{lw} = 1$, so F_g would equal F_0, giving $T_g \approx 255\,\mathrm{K}$, as in Section 1.3.1. Taking rough values for the Earth's atmosphere to be $\mathcal{T}_{sw} = 0.9$ (strong transmittance and weak absorption of solar radiation) and $\mathcal{T}_{lw} = 0.2$ (weak transmittance and strong absorption of thermal radiation), we obtain a surface temperature of $T_g \approx 286\,\mathrm{K}$, which is quite close to the observed mean value of 288 K. This close agreement is somewhat fortuitous, however, since in reality non-radiative processes also contribute significantly to the energy balance.

We can also find the atmospheric emission from equations (1.4):

$$F_a = (1 - \mathcal{T}_{lw})\sigma T_a^4 = F_0 \frac{1 - \mathcal{T}_{sw}\mathcal{T}_{lw}}{1 + \mathcal{T}_{lw}} \tag{1.6}$$

and this gives the temperature of the model atmosphere, $T_a \approx 245\,\mathrm{K}$.

This model provides a simple example of the **greenhouse effect**: the raised surface temperature is due to the fact that there is less absorption (greater transmittance) for solar radiation than there is for thermal radiation. Thus the atmosphere readily transmits solar radiation but tends to trap thermal radiation.[3] Atmospheric gases that absorb and emit infra-red radiation but allow solar radiation to pass through relatively unscathed are called **greenhouse gases**.

One way to quantify the 'greenhouse effect' of an absorbing gas is in terms of the amount $F_g - F_0$ by which it reduces the outgoing irradiance from its surface value: in the case discussed above this reduction is $140\,\mathrm{W\,m^{-2}}$. This equals the difference between the amount $(1 - \mathcal{T}_{lw})\sigma T_g^4 = 304\,\mathrm{W\,m^{-2}}$ of the thermal emission from the 'warm' surface that is absorbed by the 'cool' atmosphere and the smaller amount $F_a = (1 - \mathcal{T}_{lw})\sigma T_a^4 = 164\,\mathrm{W\,m^{-2}}$ that the atmosphere re-emits upwards. Since the atmosphere is in equilibrium, it also equals the difference between the downward emission F_a from the atmosphere and the small proportion $(1 - \mathcal{T}_{sw})F_0 = 24\,\mathrm{W\,m^{-2}}$ of the solar irradiance that it absorbs.

1.4 Some atmospheric observations

In this section we present a selection of examples of basic atmospheric observations and give some indication of their physical explanation. Further details are given in later chapters of this book.

[3] The term 'greenhouse effect' is a misnomer, however, since the elevated temperature in a greenhouse does not primarily depend on the similar radiative properties of glass, but rather on the suppression of convective heat loss.

1.4.1 The mean temperature and wind fields

Figure 1.3 shows a typical example of the vertical structure of the temperature in the lowest 100 km of the atmosphere. The atmosphere is conventionally divided into layers in the vertical direction, according to the variation of temperature with height. The layer from the ground up to about 15 km altitude, in which the temperature decreases with height, is called the **troposphere** and is bounded above by the **tropopause**. The layer from the tropopause to about 50 km altitude, in which the temperature rises with altitude, is called the **stratosphere** and is bounded above by the **stratopause**. The layer from the stratopause to about 85–90 km, in which the temperature again falls with altitude, is called the **mesosphere** and is bounded above by the **mesopause**. Above the mesopause is the **thermosphere**, in which the temperature again rises with altitude.

The troposphere is also called the **lower atmosphere**. It is here that most 'weather' phenomena, such as cyclones, fronts, hurricanes, rain, snow, thunder and lightning, occur.

The stratosphere and mesosphere together are called the **middle atmosphere**. A notable feature of the stratosphere is that it contains the bulk of the ozone molecules in the atmosphere; see Figure 1.4. The neighbourhood of the ozone maximum in the lower stratosphere is loosely known as the **ozone layer**. The production of ozone (O_3) molecules occurs through photochemical processes involving the absorption of solar ultra-violet photons by molecular oxygen (O_2) in the stratosphere, three O_2 molecules eventually forming two O_3 molecules. The equilibrium profile of ozone depends also on chemical ozone-destruction processes and on the transport of ozone by the winds (see Chapter 6).

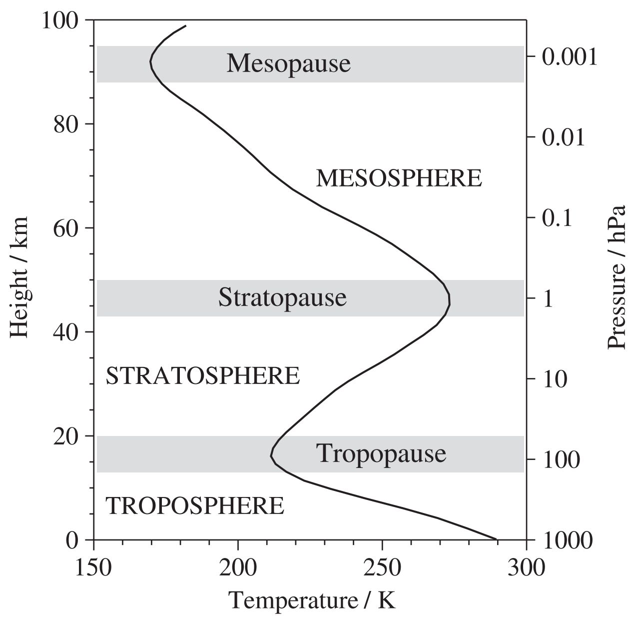

Fig. 1.3 Typical vertical structure of atmospheric temperature (K) in the lowest 100 km of the atmosphere. Based on data from Fleming *et al.* (1990).

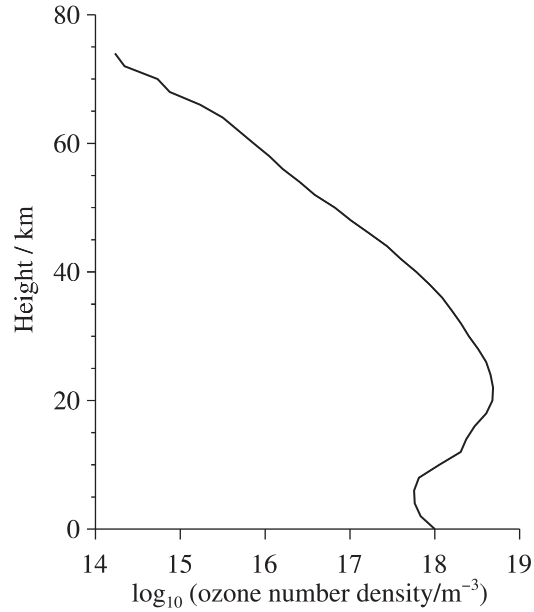

Fig. 1.4 Typical vertical structure of the mean midlatitude ozone number density (molecules m^{-3}). Based on data from US Standard Atmosphere (1976).

Above the middle atmosphere is the **upper atmosphere**, where effects of ionisation become dominant in determining the atmospheric structure and the air becomes so rarefied that the assumption that it can be treated as a continuous fluid starts to break down. In this book we concentrate on the physics of the lower and middle atmospheres.

From hydrostatic balance, the pressure at any level in the atmosphere is proportional to the mass of air above that level. From the pressure axis in Figure 1.3 it follows that approximately 90% of the atmospheric mass is in the troposphere, a little under 10% in the stratosphere and only about 0.1% in the mesosphere and above.[4] Despite their relatively low mass, the stratosphere and mesosphere are not insignificant, however. For example, ozone in the stratosphere absorbs ultra-violet solar radiation, thereby protecting the biosphere from potentially damaging effects.

Figure 1.3 is not representative of all latitudes and seasons. A more comprehensive plot, of the zonal-mean (i.e. longitudinally averaged) temperature averaged over several Januaries, as a function of height and latitude, is given in Figure 1.5. It will be seen that although the general shape of the vertical variation of temperature in midlatitudes is roughly in accord with that in Figure 1.3, there are significant latitudinal variations of the heights and magnitudes of the temperature extrema. For example, the equatorial tropopause is at a greater altitude and colder than that at higher latitudes, the summer stratopause is lower

[4] In this book we use the unit hPa for pressure (1 hPa $= 10^2$ Pa). This is equivalent to the millibar, which was formerly in common use in meteorology.

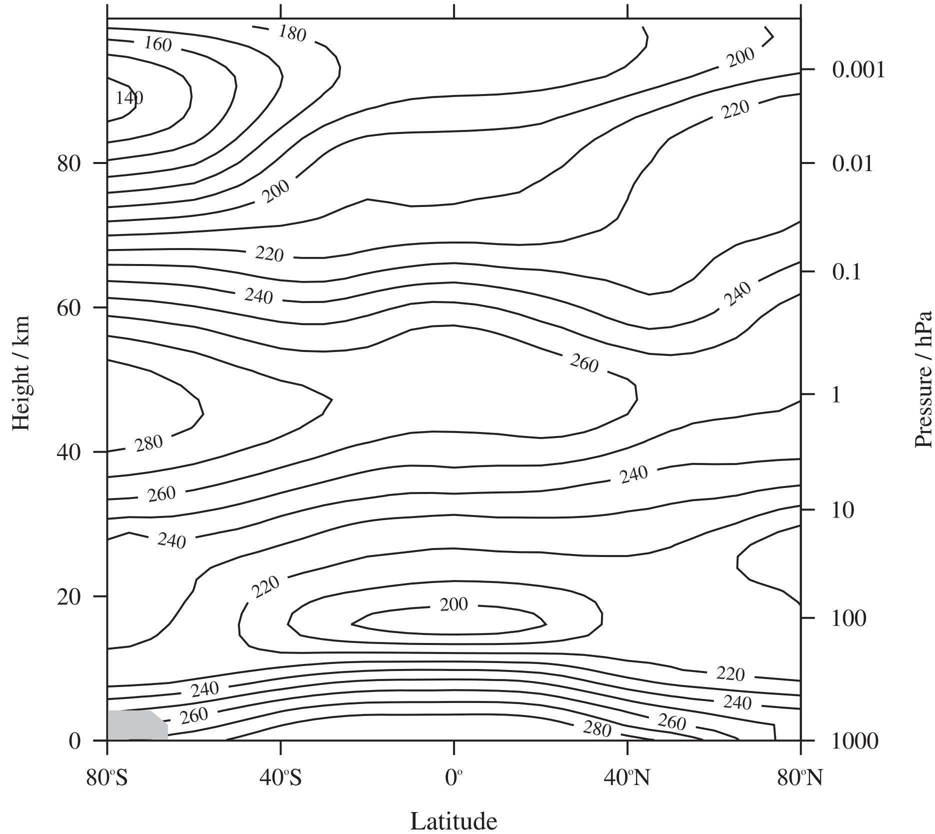

Fig. 1.5 **Zonal-mean temperature (K) for January, from the CIRA (COSPAR International Reference Atmosphere) dataset. A small region at low levels over Antarctica is omitted. Based on data from Fleming *et al.* (1990).**

and warmer than the winter stratopause and the summer mesopause is extremely cold. (In fact the lowest natural terrestrial temperatures are found there.)

Some of these temperature features can be crudely explained in terms of simple physical mechanisms. For example, the warm stratopause can be attributed to the ozone distribution, which peaks in the stratosphere; absorption of solar radiation by the ozone leads to heating of the upper stratosphere and, since in equilibrium this heating is balanced mainly by infrared cooling from carbon dioxide, there must be a local temperature maximum, so that heat can radiate to cooler regions.

If radiative control of temperature were to continue down to the ground, the temperature in the troposphere would decrease much more rapidly with height than is observed, and this temperature profile (and its associated density profile) would be statically unstable. Such a temperature profile could not persist, but might be expected to give rise to convective activity that would modify the temperature profile, causing it to decrease less rapidly with height until it was just statically stable again. This process appears to occur in the moist tropical troposphere, where the temperature decrease with altitude is fairly close to the