

## 1. Origin, Composition and Mean Structure of the Atmosphere<sup>1</sup>

When the Earth was formed, about  $4.5 \times 10^9$  years ago, it had no atmosphere. The atmosphere formed later from the volatile substances ejected by volcanoes. While the present atmosphere is roughly 75.5% nitrogen (N), 23.1% oxygen (O) and 1.3% argon (Ar) by mass, the gaseous emissions from volcanoes are roughly 85% water vapour, 10% carbon dioxide ( $CO_2$ ) and a few percent nitrogen (N) and sulphur (S) or sulphur compounds (e.g., sulphur dioxide  $SO_2$  and hydrogen sulphide  $H_2S$ ). Free oxygen is notably absent from the volcanic emissions. How can we explain this discrepancy?

To understand the formation of the present atmosphere we need to consider it as part of a coupled system, comprising the *atmosphere*, the *hydrosphere*, the *biosphere* and the *lithosphere*. The atmosphere comprises everything above the Earth's surface, the hydrosphere encompasses all water substance, the biosphere encompasses all plant and animal life, and the lithosphere covers the Earth's crust. The total mass of volatile material in this coupled system is about 0.025% of the Earth's mass.

### Evolution of the hydrosphere

Water substance is water in all its states and forms, e.g., vapour, cloud and precipitation, and the hydrosphere comprises the total mass of water substance on or above the Earth's surface. However, the atmosphere cannot hold more than a tiny fraction of the water vapour injected by volcanic eruptions. The rest rains out, adding to the hydrosphere. About 97% of all water by mass within the hydrosphere is in the ocean. Of the remainder, about  $\frac{3}{4}$  is in the ice caps of Greenland and the Antarctic,  $\frac{1}{4}$  is in the lakes and ground water, and only  $\frac{1}{3000}$  is in the atmosphere as water vapour. Moreover, the entire mass of the atmosphere is only about 1/300th of the mass of the hydrosphere.

Suppose the rate of release of steam by volcanoes over the past century is representative of the average rate over the Earth's lifetime. Then the present mass of the hydrosphere is smaller, by two orders of magnitude, than the amount of water that has been injected into the atmosphere. What has happened to the water? There are two possibilities:

- (i) water has leaked from the hydrosphere in the deep ocean along seams in the Earth's crust, or
- (ii) large amounts of water have been destroyed as a result of photodissociation by ultraviolet radiation.

(*Dissociation* is the reversible decomposition of a compound. *Photodissociation* is the dissociation of a compound because of the absorption of electromagnetic radiation.)

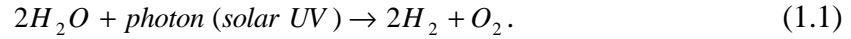
---

<sup>1</sup> This lecture draws heavily on the first chapter from Wallace and Hobbs (1977).

## Atmospheric oxygen and life

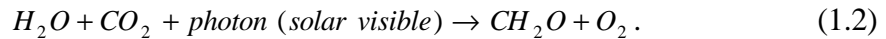
If (ii) is the case, we can explain the origin of atmospheric oxygen. There are two ways in which oxygen could have been formed.

- (a) The first possibility is through the dissociation of water:



This reaction requires ultra-violet solar radiation.

- (b) The second possibility is *photosynthesis*. This is when green plants and phytoplankton manufacture carbohydrates from atmospheric carbon dioxide and water in the presence of sunlight. For example,



It is well established that the photosynthesis reaction (1.2) has produced significant amounts of oxygen on the Earth (much more, in fact, than is now in the atmosphere). However, it is not known whether it has produced enough oxygen to account for the oxidation of materials in the Earth's crust that has occurred since the Earth's formation.

The possible role of the photodissociation reaction is a matter of controversy. The reaction rate depends on other photochemical reaction that compete for the same ultraviolet radiation. Also, the rate of  $O_2$  production depends on the rate at which the  $H_2$  produced in the reaction escapes to space (we discuss this mechanism of escape below). If the escape rate is slower than the production rate (a possibility!), then the  $O_2$  and  $H_2$  will recombine.

The production of  $O_2$  by photosynthesis is closely linked with biological processes, since  $CH_2O$  produced in (1.2) is the basic building block for the carbohydrate molecules that form plant cells. Since  $O_2$  is abundant in the Earth's atmosphere and nearly absent in the atmospheres of Venus and Mars, which are lifeless or nearly lifeless, it is tempting to conclude that most of the oxygen in the Earth's atmosphere was produced by photosynthesis.

It is thought that the crucial first stage in the evolution of single-celled organisms  $4 \times 10^9$  years ago required an  $O_2$ -free environment. Geological evidence indicates that primitive forms of plant life had evolved to the point where they began to release very small amounts of  $O_2$  through photosynthesis between 2 and  $3 \times 10^9$  years ago. It is believed that these early life forms developed in an aqueous environment, far enough below the surface to be protected from the Sun's lethal ultraviolet radiation, but close enough to the surface to obtain visible radiation needed for photosynthesis.

By processes to be discussed, the build up of  $O_2$  in the atmosphere led to the formation of an ozone ( $O_3$ ) layer in the upper atmosphere. This filters out incoming solar radiation in the ultraviolet (UV) part of the spectrum. As the ozone layer developed, less and less UV radiation penetrated to the Earth's surface. This provided an increasingly favourable environment for plant life which was able to spread upward into the uppermost layers of the ocean. As it did so, it gained access to increasing visible radiation.

The picture seems to be:

---

More  $O_2 \rightarrow$  less UV radiation  $\rightarrow$

More access to visible radiation  $\rightarrow$

More abundant plant life  $\rightarrow$  more  $O_2$  production

---

In this way life may have slowly but inexorably worked its way to the surface until it finally emerged on to land some  $4 \times 10^8$  years ago.

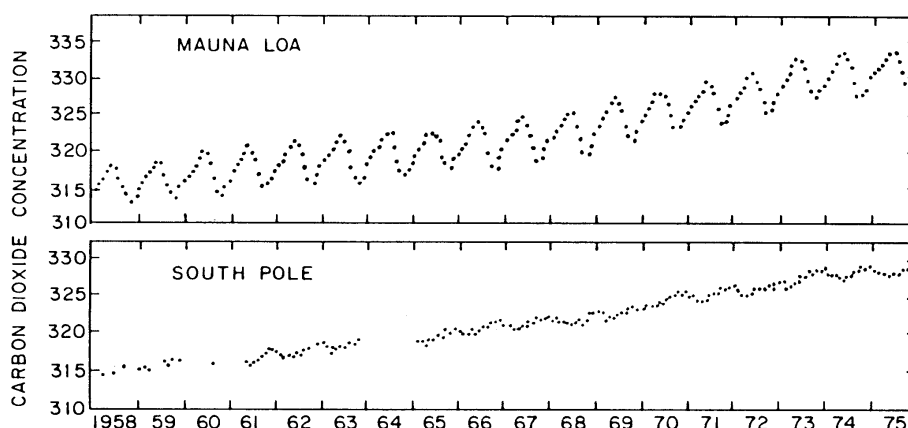
### **Oxygen and carbon budgets**

For each  $O_2$  molecule produced in (1.2), one molecule of carbon is incorporated into an organic compound. Most of these carbon atoms are oxidized again in respiration or in the decay of organic matter (in which bacteria produce energy from food stored in the corpus of dead organisms)



However, one in every  $2 - 3 \times 10^4$  carbon molecules photosynthesized escapes oxidation and is buried or "fossilized". Most of the Earth's unoxidized carbon is contained in shales. Smaller amounts are stored in more concentrated form in fossil fuels (coal, oil, and natural gas). The carbon stored in the biosphere is only a minute fraction of the total storage. The relative amount of carbon stored is as follows: 1 unit in the biosphere (marine), 1 unit in the biosphere (land), 70 units in atmospheric  $CO_2$ , 4000 units in dissolved oceanic  $CO_2$ , 800 units in fossil fuels, 800,000 units in shale, and 2,000,000 units in carbonate rocks.

Burning fossil fuels undoes the work of photosynthesis. People burn in 1 year what it took photosynthesis about  $10^3$  years to produce. Fortunately, photosynthesis has been at work for hundreds of millions of years. The bulk of the organic carbon in the lithosphere is stored in a form that is too dilute for people to exploit.



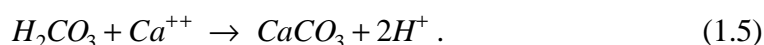
**Figure 1.1.** Carbon dioxide concentrations at Maura Lao, Hawaii, and at the South Pole in recent decades. The units are parts per million. Notice the pronounced annual variation and the increasing trend of the annual mean.

Of the net  $O_2$  that has been produced by plant life (i.e., production by photosynthesis - destruction by decay) during the Earth's history, only about 10% is stored in the atmosphere. Much of the remainder has found its way into oxides (e.g.,  $Fe_2O_3$ ) and carbonate compounds ( $CaCO_3$  and  $MgCO_3$ ) in the Earth's crust. The formation of carbonate compounds is of particular interest since it is the major sink for the vast amount of  $CO_2$  that has been released in volcanic activity.

Carbonates are formed by ion-exchange reactions that take place in certain marine organisms (e.g., the one-celled *foraminifera*). The dissolved  $CO_2$  forms a weak solution of carbonic acid:



Then a sequence of reactions follow with the net result



The  $CaCO_3$  enters into the shells of animals, which are eventually compressed into limestone in the Earth's crust.  $MgCO_3$  is produced by a similar sequence of reactions.

The  $H^+$  ions released in (1.5) react with metallic oxides in the Earth's crust, from which they steal an oxygen atom to form another water molecule. An oxygen atom from the atmosphere eventually replaces the stolen one. Therefore,  $O_2$  is removed from the atmosphere during the formation of carbonates and is given back when the carbonates dissolve. It has been suggested that foraminifera and other carbonate producing sea animals have the ability to regulate the amount of  $O_2$  in the atmosphere.

The widespread occurrence of the fossils of sea animals in limestone deposits points to the important role of ion exchange reactions in sea water in removing  $CO_2$  from the Earth's atmosphere.

Measurements indicate that the rate of removal of  $CO_2$  from the Earth's atmosphere is not large enough to compensate for the increasing rate of input due to the burning of fossil fuels. See Fig. 1.1.

### **Other atmospheric constituents**

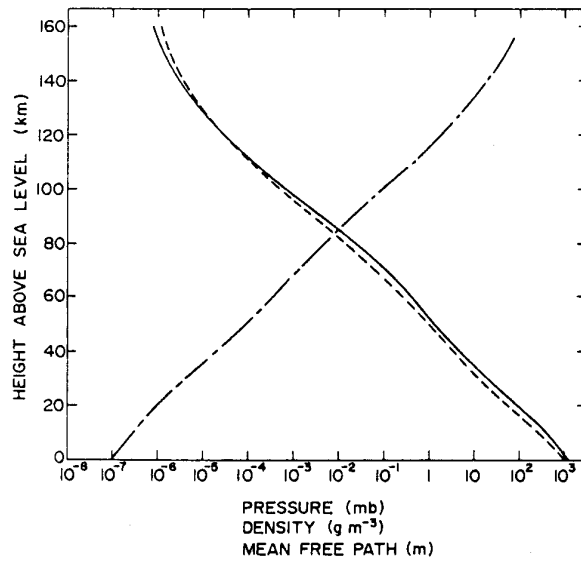
Molecular nitrogen accounts for nearly three quarters of the mass of the atmosphere. A small fraction (perhaps 20%) of the nitrogen released into the atmosphere has entered into nitrates in the Earth's crust. Atmospheric  $N_2$  is *fixed* (chemically combined with other substances) and removed from the atmosphere in three main ways: (a) fixation by biological micro-organisms, mainly on land; (b) fixation by lightning and other ionizing processes in the atmosphere; and (c) industrial fixation, mainly to make artificial fertilizers. However, because it is chemically inert and has a low solubility in water ( $\approx 1/70$  that of  $CO_2$ ), most of the  $N_2$  released by volcanoes has remained in the atmosphere. Indeed, because of the nearly complete removal of water vapour and  $CO_2$  it has become the dominant gaseous constituent.

Sulphur and its compounds hydrogen sulphide ( $H_2S$ ) and sulphur dioxide ( $SO_2$ ), which are released into the atmosphere by volcanic emissions, are quickly oxidized to sulphur trioxide ( $SO_3$ ) which dissolves in cloud droplets to form a dilute solution of sulphuric acid ( $H_2SO_4$ ). The same process is responsible for *acid rain*, which occurs downwind of major industrial centres where sulphur-bearing coal and oil are burned. After "raining out", the sulphate ions combine with metal ions to form sulphates within the Earth's crust. Sulphur dioxide may also react with ammonia in the presence of liquid water to produce ammonium sulphate.

Of the other trace gases, Argon is the most abundant. It is chemically inert. About 99.7% of argon is the product radioactively decaying potassium ( $^{40}K$ ) with the solid Earth.  $^{40}K$  decays very slowly, the half life being  $1.3 \times 10^9$  years, which about  $\frac{1}{3}$  the age of the Earth! Atmospheric helium is also mainly a by-product of radioactive decay.

### **Vertical profiles of pressure and density**

The vertical variability of pressure and density is much larger than the horizontal and temporal variability of these quantities. For this reason, it is useful to define a "standard atmosphere" which represents a horizontal and time averaged structure that is a function only of height (see Fig. 1.2). Within the lowest 100 km,  $\ln p(z)$  decreases almost linearly with height, i.e.,



**Figure 1.2.** Vertical profile of pressure, density and the mean free path. (From Wallace and Hobbs, 1977.)

where  $z$  measures height above mean sea level and  $B$  is a constant. (The reason for this will be explored later.) Consequently,

$$p(z) = p(0)\exp(-z/H_s).$$

$H_s$  is called the *scale height* of the atmosphere; it is an "e-folding" decay scale for  $p$ . This means that the pressure drops off by a factor of  $e$  (2.718) every  $H_s$  in the atmosphere.

The mean surface pressure,  $p(0)$ , is 1013.25 mb, or. 1013.25 hPa (or hectopascals). Note that 1 hPa = 100 Pa and 1 Pa (pascal) represents a force of 1 newton acting over 1 square metre. In the meteorological literature, 1 hPa is often referred to as a millibar (1/1000th of a *bar*, the old unit of pressure).

The density curve in Fig. 1.2 lies close to the  $p$  curve and has very nearly the same slope. It follows that

$$\mathbf{r}(z) = \mathbf{r}(0)\exp(-z/H_s) ,$$

also. Later in the course we will find that half the mass of the atmosphere lies below the 500 mb level, which has a mean height of about 5.5 km, less than 1/1000<sup>th</sup> of the Earth's radius. Indeed, 99% of the mass of the atmosphere lies within the lowest 30 km above sea level.

## Atmospheric composition as a function of height

If there are no sources or sinks, the ratios of various gaseous constituents at a particular level are determined by two competing physical processes: molecular diffusion and mixing due to fluid motions.

Diffusion by random molecular motions tends to produce an atmosphere in which the mean molecular weight of the mixture of gases gradually decreases with height. The density of each constituent decreases exponentially with height, but the scale height  $H$  is inversely proportional to its molecular weight. Effectively, each constituent behaves as if it, alone, were present. Clearly, at the highest levels, only the lightest gases (hydrogen and helium) are present in appreciable amounts. The reasons for this behaviour will be discussed later.

In contrast to molecular diffusion, the mixing due to the motions of macroscale air parcels does not depend on molecular weight. Therefore, where this process dominates, atmospheric composition tends to be independent of height.

The relative effectiveness of molecular diffusion increases in proportion to

- (i) the root mean square velocity of the random molecular motions, and
- (ii) the mean free path between collisions.

Consider the following analogy to illustrate (ii); compare the diffusion process with the process of separation of the sexes when they seek out their respective toilets during the break in a concert. The more crowded the foyer (i.e., the smaller the "mean free path"), the slower the rate of diffusive separation. In the mixing by fluid motions, the analogue of the mean free path is the "mixing length". This depends on the spectrum of scales of motion present in the atmosphere.

The factor that influences the relative effectiveness of molecular diffusion compared with mixing by fluid motion is the increase of the mean free path with height.

In the lower atmosphere the mean free path is so short that the time required for the vertical separation of the heavier and lighter constituents by molecular diffusion is many orders of magnitude longer than the time required for turbulent fluid motions to homogenize them. The two processes become comparable at a height of about 100 km. Above this height, molecular diffusion becomes increasingly dominant. The transition level is called the *turbopause*. The region below is called the *homosphere*, the region above the *heterosphere*.

The composition of the lower part of the heterosphere is strongly influenced by the photodissociation of diatomic oxygen. This gives rise to large numbers of free oxygen atoms.

At higher levels, there is a noticeable increase in the relative abundance of the lighter constituents. The heaviest major constituent, diatomic nitrogen decreases most rapidly with height. Around 500 km the atmosphere is predominantly atomic oxygen, with only traces of diatomic nitrogen and the very light constituents, helium and hydrogen. Above 1000 km, helium and hydrogen are the dominant species.

Note that the atmosphere is a remarkably thin film of gas. The radius of the Earth is  $a_e = 6.37 \times 10^6 m = 6370 km$ , but the height of the turbopause is  $\sim 100 km$ , and the height of the tropopause only  $\sim 10 km$

### Escape of the lighter constituents

Above 500 km the mean free path between collisions is so long that collisions are relatively rare. Under these circumstances it is possible for some of the faster moving molecules to escape the Earth's gravitational field into space. The escape velocity  $V_e$  is independent of the molecular species, and depends only on the height  $z$ . For a particle of mass  $m$  to escape, its kinetic energy  $\frac{1}{2}mv^2$  must exceed the gravitational potential  $GM_e m/r$ , where  $M_e$  is the mass of the Earth,  $r$  is the radius from the centre of the Earth and  $G$  is the gravitational constant. Therefore, the escape velocity  $V_e = \sqrt{2GM_e/r}$ . At a height of 500 km above the Earth's surface,  $r = (6.37 + 0.5) \times 10^6 m$ , whereupon

$$V_e = \sqrt{[2 \times 6.67 \times 10^{-11} \times 5.97 \times 10^{24} \div (687 \times 10^6)]},$$

or approximately 11km/sec.

The most probable velocity of any molecular species is

$$V_0 = \sqrt{[2kT/(Mm_H)]}, \quad (1.6)$$

where  $k$  is Boltzmann's constant ( $1.38 \times 10^{-23} JK^{-1}$ ),  $T$  is the absolute temperature,  $M$  is the molecular weight, and  $m_H$  is the mass of hydrogen atom ( $1.67 \times 10^{-27} kg$ ). One of the the assumptions on which Eq. (1.6) rests is an equipartition of kinetic energy among the various molecular constituents. It follows that the mean velocity of a lighter constituent is higher than that of a heavier one. However, the individual molecules of a gas species have a range of velocities scattered about  $V_0$ . The kinetic theory of gases predicts that only about 2% of the molecules have velocities greater than  $2V_0$  and only one molecule in  $10^4$  has a velocity greater than  $3V_0$ .

$V/V_0$	Fraction
1	0.5
2	0.02
3	$10^{-4}$
4	$10^{-6}$
6	$10^{-20}$
10	$10^{-50}$
15	$10^{-90}$

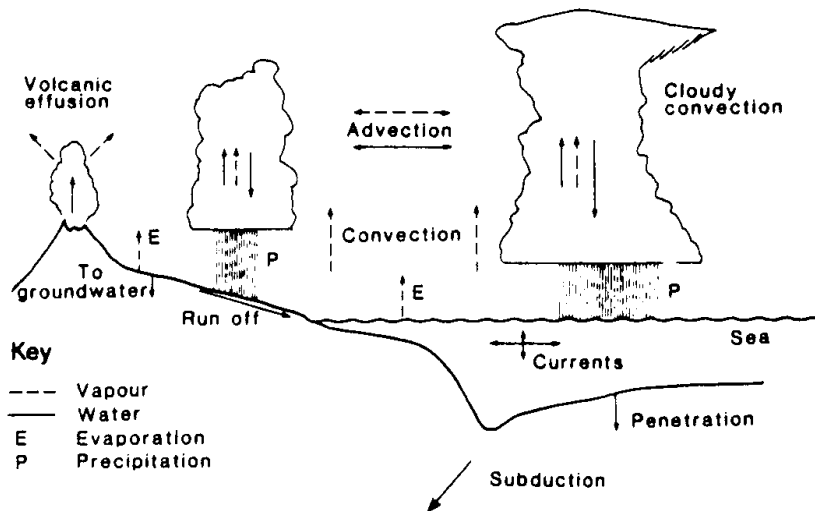
**Table 1.1.** Fraction of gas molecules with velocities  $V$  greater than various multiples of the most probable velocity  $V_0$ . (From Wallace and Hobbs, 1977.)

In the Earth's atmosphere temperatures at the base of the "escape region", sometimes called the exosphere) are on the order of 600K. Then Eq. (1.6) gives for a hydrogen atom ( $M = 1$ ),  $V_0 = 3 \text{ km/sec}$ . Therefore, according to Table 1.1, for each collision near 500km, the probability of escape (i.e.,  $V > V_0$ ) is slightly greater than  $10^{-6}$ . Corresponding time period required for the escape of all the hydrogen from the Earth's atmosphere is much less than the lifetime of the Earth. This accounts for the relative absence of free hydrogen in the atmosphere despite its continual production due to the dissociation of water. For atomic oxygen ( $M = 16$ ),  $V_0 \approx 0.8 \text{ km/sec}$ . The probability of escape is so low that the loss over the lifetime of the Earth is negligible.

### **Variable constituents: water vapour and ozone**

Unlike the other gases, the concentrations of water vapour and ozone are highly variable in space and time. Although these gases are present only as trace constituents, they are very important in the atmospheric energy balance.

Water substance is very unevenly distributed through the atmosphere, and almost entirely confined to low levels, and especially in the tropics. The main source of atmospheric water vapour is evaporation from the Earth's surface, especially over the tropical oceans. The main sink is condensation that occurs in clouds. The concentration of water vapour is largest near the surface, and decreases very rapidly with height to very small values above 10km. For example, the specific humidity is

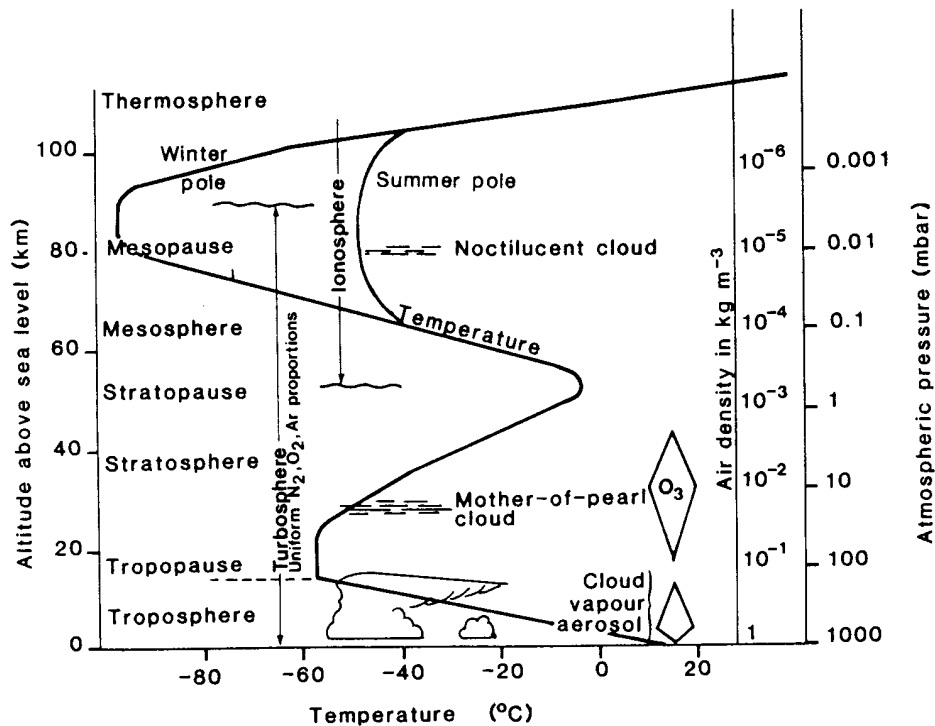


**Figure 1.3.** Schematic of the hydrological cycle emphasising the atmospheric branch (from McIlveen 1991).

defined as  $q = \frac{\text{mass water vapour}}{\text{total mass}}$ , and is usually expressed in  $g/kg$ . The specific humidity of the whole atmosphere  $\sim 0.3 g/kg$ , but the specific humidity in the tropics (typically) exceeds  $3 g/kg$ . It follows that cloud (and hence precipitation) is almost entirely confined to the troposphere.

If all the water vapour in the atmosphere were condensed, it would be equivalent to about  $3 cm$  of liquid water at the Earth's surface. On the other hand, were the ocean water distributed evenly across the earth, it would be  $2.8 km$  deep. Note that on average, about  $1 m$  of rain falls annually, yet there is only  $3 cm$  of precipitable water in the atmosphere at any instant. The typical "life time" of a molecule of water vapour in the atmosphere is only on the order of a week. This rapid recycling of water molecules in the atmosphere is a feature of the *hydrological cycle*. (See Fig. 1.3.)

Ozone is generated by photochemical reactions in the layer between  $20$  and  $60 km$ . It is generated also in polluted air over cities, but in much smaller quantities (fortunately because ozone is quite poisonous). At the Earth's surface, ozone is rapidly destroyed. It reacts with plants and dissolves in water. At levels between  $10$  and  $25 km$ , it is a stable chemical species with a lifetime on the order of months. In the long-term average, there is a slow downward flux of ozone from the high level source region to the sink at the Earth's surface. We discuss the production (and destruction) of ozone later in the course.



**Figure 1.4.** Vertical profile of temperature. (From McIlveen 1991.)

### The vertical temperature structure

The vertical structure of temperature for the standard atmosphere is shown in Fig. 1.4. This profile is characteristic of conditions in middle latitudes. The profile can be divided into four distinct layers: the troposphere, stratosphere, mesosphere, and thermosphere. The tops of these layers are called the *tropopause*, *stratopause*, *mesopause*, and *thermopause*, respectively.

The troposphere has more than 80% of the mass of the atmosphere. It has virtually all of the water vapour, clouds and precipitation as well. Indeed, virtually all weather-producing systems are confined to the troposphere. Occasionally, the tops of large thunderstorms may penetrate into the lower stratosphere.

The troposphere is characterized by strong vertical mixing. For example, in clear air it is not unusual for an air molecule to traverse the whole depth of the troposphere in a few days. In the updraughts of large thunderstorms, particles can travel from near to the ground to the tropopause in 10 minutes or so.

The tropopause marks an abrupt change in certain trace quantities. Water vapour decreases sharply, while ozone concentration often increases by an order of

magnitude. These strong gradients arise because there is little mixing between dry, ozone-rich stratospheric air and relatively moist, ozone-poor tropospheric air. The lack of mixing is a consequence of the high static stability of the stratosphere. Because of this, matter injected into the stratosphere, for example from volcanic eruptions, nuclear explosions or through overshooting thunderstorm tops, tends to remain there for long periods. Indeed, the residence time of particles is typically measured in years compared with a few days to a few weeks. Thus the stratosphere in the troposphere serves as a "reservoir" for certain types of atmospheric pollution.

The pressure at the stratopause is about 1 mb. At the Earth's surface it is about 1000mb. Therefore, 99.9% of the atmosphere's mass lies in the troposphere and stratosphere. Of the mass that remains, 99% lies in the mesosphere and only 1% in the thermosphere.

The mesosphere, like the troposphere, is a region in which the temperature decreases with height. Therefore, vertical air motions are not inhibited by large stability. During summer there is sometimes enough lifting to produce cloud layers in the upper mesosphere over parts of the polar regions. These clouds may be visible from the ground at twilight when the lower atmosphere is in darkness. They are called *noctilucent clouds*.

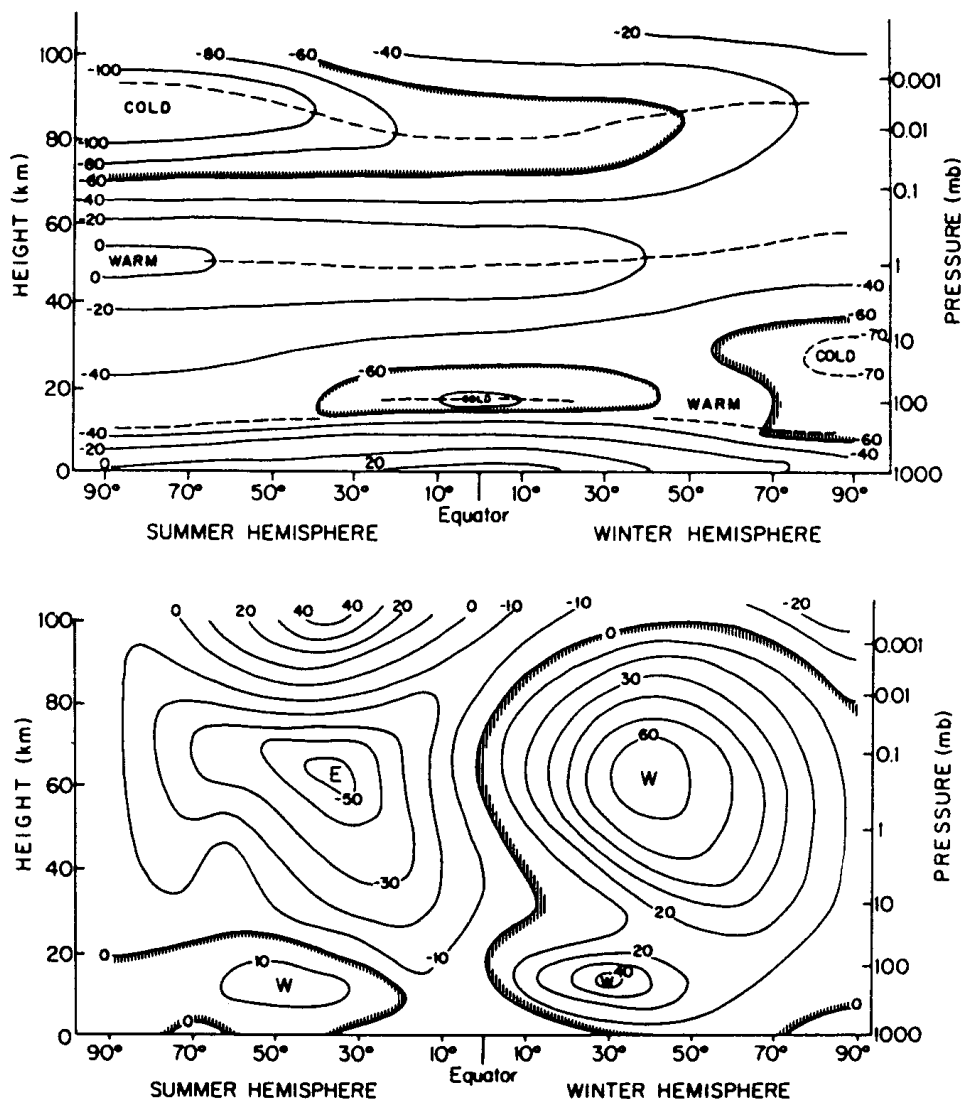
The thermosphere extends several hundred kilometres above the mesopause. The temperature structure there depends strongly on the amount of solar activity with temperatures ranging between 500K and 2000K. The thermopause is a level of transition to a more or less isothermal profile. Above 500km molecular collisions are so infrequent that temperature, as such, is difficult to define. The physical processes responsible for the vertical temperature structure will be discussed later.

### **Temporal and latitudinal variability**

The mean temperature profile shown in Fig. 1.4 is present at nearly all latitudes and in all seasons. However, there is considerable latitudinal and seasonal variability as shown in Fig. 1.5a.

Temperature decreases with latitude in the troposphere. The latitudinal gradient is about twice as steep in the winter hemisphere as in the summer hemisphere. The tropopause is much higher and colder over the tropics than over the polar regions.

In the lower stratosphere the latitudinal distribution is rather complicated. The summer hemisphere has a cold equator and a warm pole. The winter hemisphere is cold at both equator and pole with a warmer region in middle latitudes. The cold pool of stratospheric air over the winter pole is very variable. On occasions it disappears for a period of a few weeks during midwinter. During these so-called "sudden warmings", the stratospheric temperatures over individual stations have been observed to rise by as much as 70 deg in one week.



**Figure 1.5.** Meridional cross-section of longitudinally average (a) temperature and (b) zonal wind. (From Wallace and Hobbs, 1977.)

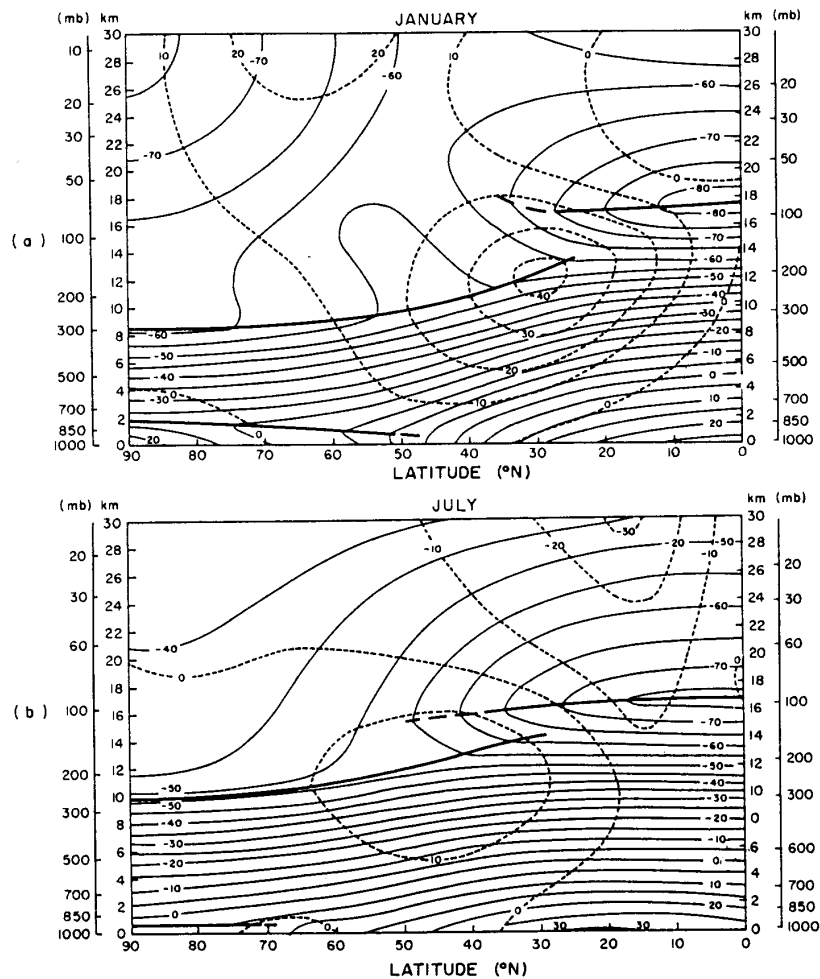
At the stratopause there is a monotonic temperature gradient between the warm summer pole and the cold winter pole. At the mesopause, the situation is exactly opposite: the summer pole is cold, the winter pole is warm.

Temperature has a pronounced diurnal variability in certain regions of the atmosphere. The strongest variability is observed in the upper thermosphere. Here, day to night temperature differences are on the order of several hundred degrees. There are also important (but much smaller) diurnal variations around the stratopause level. These give rise to strong tidal motions in the Earth's upper atmosphere. The

tidal motions are manifest as regular oscillations in surface pressure which are prominent in the tropics.

In the middle and upper troposphere, the day to night changes are typically less than a degree. However, in the lowest few kilometres over land the changes are somewhat larger. At the Earth's surface, over land, the diurnal range is typically of the order of 10 deg. It may exceed 20 deg over high altitude desert regions.

At any particular time, the temperature distribution within the troposphere shows large deviations from the seasonally and longitudinally-averaged cross-section shown in Fig. 1.5a. Much of the equator to pole temperature gradient is often concentrated in narrow zones, called frontal zones. We will discuss these in more detail later.



**Figure 1.6.** Meridional cross-section of longitudinally average temperature and zonal wind. (From Wallace and Hobbs, 1977.)

## Wind structure in the Earth's Atmosphere

The broadscale features of the atmospheric circulation are shown in Fig. 1.5b. This is a latitude-height cross-section of the zonal wind component (zonal means east-west component), averaged with respect to longitude, during the Northern Hemisphere winter. When so averaged, the zonal component is generally an order of magnitude larger than the meridional (i.e., north-south) component. Prominent features are cores of strong westerly winds in middle latitudes at an altitude of 10 km.

However, the strongest zonal winds occur in the mesosphere at an altitude of 60 km. Again there are two jet cores in middle latitudes, the stronger in the winter hemisphere is westerly (from the west); the weaker in the summer hemisphere is easterly. During the equinoxes, these jets undergo dramatic reversals as the latitudinal temperature gradient reverses. Certain important features of the longitudinally-averaged zonal wind field do not show up explicitly in Fig. 1.5b. For example, the "sudden warming" phenomenon described earlier is accompanied by large changes in the longitudinally-averaged zonal wind at high latitudes in the winter stratosphere. The midwinter warmings are accompanied by a pronounced weakening of the westerlies at stratospheric levels. Sometimes the westerlies disappear altogether. These changes in the stratosphere have little effect on the wind structure in the troposphere.

Figure 1.6 shows the temperature structure in the troposphere and lower stratosphere in more detail. Note the prominent break in the tropopause in middle latitudes. The break is further polewards in summer than in winter, but is not so sharp in summer. The position of these so-called jet streams corresponds closely to the tropopause breaks, shown in Fig. 1.6. The tropospheric jet in the winter hemisphere is much stronger than the one in the summer hemisphere. We shall show later that the strength of the jets is related to the magnitude of the horizontal temperature gradient.