

## **-Section 1: Basics**

### ***Scales of motion***

This course is concerned with the way in which the ideas of fluid dynamics can be used to understand atmospheric phenomena; in particular we shall apply physical insights such as Newton's laws of motion and the laws of thermodynamics to explain aspects of the motion systems which we see in the atmosphere.

For example, when we look at the sky we become aware that atmospheric motions are not totally random, but are arranged into some kind of characteristic types or patterns. Thus we are all familiar with the cloud patterns which might appear on a sunny afternoon over land, having rather flat bases and bubbly tops a bit like cabbages (namely cumulus clouds). Watching these we might readily deduce from the size of their shadows that they have sizes around a few hundred meters, ranging from tens of meters up to a km or two and that they form and dissipate in a matter of a few tens of minutes. Although no two cumulus clouds are exactly the same, they have a certain similarity in shape size and timescale, that makes us think that we ought to be able to explain why such shapes and sizes exist. We observe motions taking place at other scales too. At the smallest scale there are the kinds of eddying motion that causes flags to flutter and which raise swirls of dust, where the scale is only a few meters (or maybe centimetres) up to a few tens of meters and the timescale is from fractions of a second to a few seconds. Looking at the largest scale, say with time-lapse pictures from geostationary satellites, we see vast swirling motions filling a large fraction of a hemisphere and persisting for days or weeks. There are a host of identifiable patterns in between these extremes. Atmospheric motion has discernable patterns at different scales.

Some typical horizontal scales are

- Planetary scale waves: few 10,000km
- Cyclones and anticyclones: few 1000km
- Fronts in cyclones: 1000km x 100km
- Hurricanes: 100km x 100km
- Cloud clusters and squalls 10s of km
- Cumulonimbus 10km
- Cumulus few 100m to few km

We may expect that the scale may affect what processes we need to take into account. For instance for the largest scale motion lasting several days the fact that the earth is rotating (and hence so is the frame of reference in which we usually contemplate the systems) might be expected to matter, whereas for a cloud lasting 20 minutes the rotation of the frame in that period will be pretty tiny and might be expected not to significantly affect what happens.

### ***Composition***

Air is a mixture of gases. In the bulk of the atmosphere the air currents stir this mixture to a uniform composition much faster than diffusive separation under gravity

can separate the constituents out<sup>1</sup>. Some constituents have sources and sinks which produce noticeable changes. In particular the processes of evaporation and condensation of water vapour are quite rapid. It is convenient therefore to consider air as a mixture of dry air and water vapour. The composition of dry air is fairly constant and is as appears in the table.

Table 1

gas	% by volume	% by mass
Nitrogen	78.09	75.51
Oxygen	20.95	23.14
Argon	0.93	1.3
Carbon dioxide	(var) ~ 0.3	~ 0.5
Neon	$180 \cdot 10^{-5}$	$120 \cdot 10^{-5}$
Helium	$52 \cdot 10^{-5}$	$8 \cdot 10^{-5}$
Krypton	$10 \cdot 10^{-5}$	$29 \cdot 10^{-5}$
Hydrogen	$5.0 \cdot 10^{-5}$	$0.35 \cdot 10^{-5}$
Xenon	$0.8 \cdot 10^{-5}$	$3.6 \cdot 10^{-5}$
Ozone	var ~ $0.1 \cdot 10^{-5}$	~ $0.17 \cdot 10^{-5}$

*var* indicates that the quantities of this constituent are variable.

Above about 80km oxygen begins to be dissociated by UV radiation from the sun. At still greater heights nitrogen also dissociates.

### Mixing ratios

The ratio of some importance is the mass of a constituent in an air sample to the total mass of the sample. This ratio is usually called the *mass mixing ratio*. There is an analogous quantity in terms of the number of molecules. As the volume of a gas at a standard pressure and temperature is proportional to the number of molecules (and independent of the substance) this ratio has come to be known as the *volume mixing ratio*. Thus the volume mixing ratio is defined as the number of molecules of a constituent in an air sample to the total number of dry air molecules in the sample. Note that the volume mixing ratio is also called the *mole fraction*, which is in many ways a better term, but “volume mixing ratio” is too firmly entrenched for it to be abandoned.

Some variations on this terminology may be found. In particular meteorologists use the term *mixing ratio* to mean the ratio of the mass of water vapour contained in an air sample to the mass of *dry* air with which it is mixed. Likewise some authors define mass mixing ratio of a pollutant as the ratio of the mass of pollutant to the total mass of air excluding the pollution. For pollutants which appear in only tiny quantities the difference between the two definitions will be negligible. (Mass mixing ratios of

<sup>1</sup> Viscosities for the gases of which air is composed are such that timescales for molecular diffusion over distances of 100m in the lower atmosphere are about 25 years, or 2500years for scales of 1km and 250,000 years for the depth of the troposphere

water vapour rarely exceed 40g/kg. So the difference between the two definitions of mass mixing ratio does not exceed 4%.)

For minor constituents the mixing ratios will often be expressed in parts per million by volume (denoted as `ppmv') or analogously by mass. Thus the mixing ratio of hydrogen in dry air is 50 ppmv or 3.5 ppmm

The principal variable constituent of air is water vapour. This has maximum mixing ratio near the surface in the tropics, where mass mixing ratios of 4‰ (40 parts per thousand or 40g/kg) are common. In middle latitudes mass mixing ratios of water near the surface are typically a few parts per thousand. The mixing of water decreases rapidly with height in the lowest 10 km of the atmosphere.

### The Gas Law

To a very good approximation air behaves like a perfect gas. As shown in most thermodynamic textbooks, for such a gas the pressure,  $p$ , volume,  $V$ , and absolute temperature,  $T$ , are related by

$$pV = nRT$$

**Eq 1**

where  $n$  is the number of moles present and  $R$  is a universal constant (called *the universal gas constant*.  $R$  has the value  $8.3143 \text{ J K}^{-1} \text{ mol}^{-1}$ ).

Note that many meteorology books use the symbol  $R^*$  for the universal gas constant (and use  $R$  for what we call  $R_d$  below).

The above equation is true for a gas of a single constituent. It turns out to be true also for a gas comprising a mixture of constituents. This is a consequence of Dalton's law of partial pressures, which states that each gas exerts a pressure force independent of the presence of the other gases. Thus the total pressure,  $p = \sum_i p_i$ , where  $p_i$  denotes the pressure of the  $i^{\text{th}}$  gas, for which the number of moles present we will denote by,  $n_i$ . Thus

$$pV = \sum_i n_i RT$$

**Eq 2**

Now each mole of a gas contains the same number of molecules (Avagadro's number). So if  $n$  is the total number of moles present, we must have  $n = \sum_i n_i$  and

Eq 2 reduces to Eq 1.

A difficulty with equation Eq 1 arises because it contains a volume, whereas it is more convenient to have equations containing variables defined at a point. Therefore it is usually rewritten in terms of density  $\rho$ . If the total mass of gas in the sample is  $M$  and if an appropriately defined average molecular weight for the mixture is  $\bar{m}$ , then  $M = n\bar{m}$ , so that Eq 1 can be written

$$p = \frac{R}{\bar{m}} \rho T .$$

We may set  $R_d = R/\bar{m}$ , giving

$$p = R_d \rho T$$

**Eq 3**

Eq 3 is the *gas equation* the mixture.  $R_d$  is simply called the *gas constant* for that particular mixture. Clearly it varies according to composition.

As we have seen, the biggest variation in composition comes from alterations in the moisture content. In some contexts this will obviously have to be taken into account, but for the purposes of this course it will be sufficiently accurate to treat the air as if it contained no water vapour, i.e. it is dry. For dry air  $\bar{m}$  is usually denoted by  $m_d$ . This is found experimentally to have a value of 28.96kg/kmole, giving a value of  $R_d$  for dry air of 287 J kg<sup>-1</sup> K<sup>-1</sup>.

### **Average temperature variation with height**

As part of the meteorological observing system balloons measuring temperatures, humidities, winds and pressures are released from several hundred stations around the world. These give information to a height of about 30km, usually twice per day. In addition a few stations occasionally launch rockets which give similar information from 20km to 80 or 0km. Figure 1 shows an example of the temperature structure measured by these techniques. The global observing network also relies heavily on satellites which can obtain similar information (with better horizontal coverage but poorer vertical resolution).

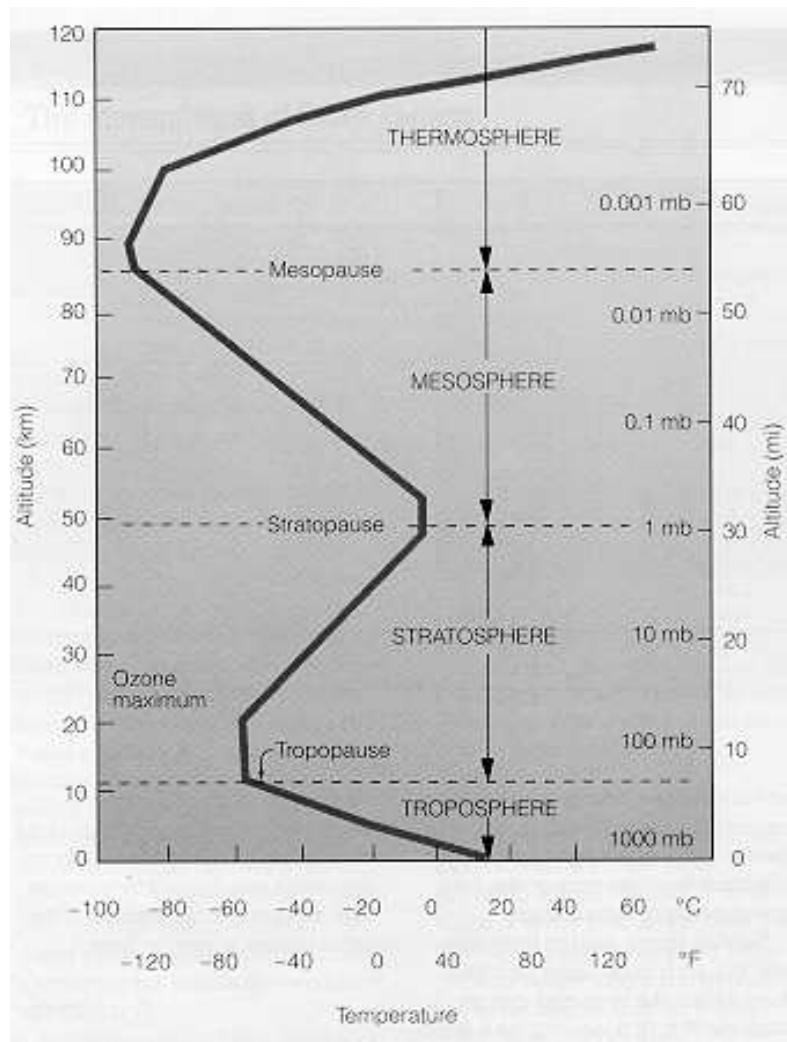
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**Figure 1 : Temperature structure from balloon and radiosonde**

Certain features of the distribution shown in Figure 1 are usually to be found in such plots. In particular, there is a region at the bottom where the temperature decreases sharply with height. The depth of this layer varies with time and place but is typically 10km in middle latitudes but may be 16~km in the tropics. Above that there is a sharp transition to region where the temperature is fairly constant with height or increasing. This continues to around 50km where a region of decrease begins which continues up to about 80km. At this point the temperature begins to increase again and it continues to increase until at great heights the density is sufficiently low for the term "temperature" to begin to become ambiguous.

The different regimes defined by the sign of the temperature gradient are given names as follows. The lower portion in which the temperature decreases with height is called the *troposphere*, the region above in which the temperature is constant or increasing is called the *stratosphere*. The transition point is called the *tropopause*. 'Tropo' is from the Greek for 'mix' and indicates that this region of the atmosphere is comparatively well mixed compared with the stratosphere, 'strato' being derived from the word for 'layered'. The stratosphere is bounded by the stratopause at around 50km, above which is the next region of temperature decrease, known as the *mesosphere*. Above this, starting at about 80km, is the *thermosphere*, the transition region being known as the *mesopause*.

There are several *standard atmospheres* which are idealisations based on this structure which are used for engineering and similar applications (say in aviation) which usually represent the temperature structure in the "spheres" as linear functions of height, joined at the "pauses". An example is shown in Figure 2



**Figure 2: Standard atmosphere**

The reason for the overall shape of the temperature profile is related to the radiation budget of the atmosphere. To a first approximation the atmosphere is everywhere cooling because of the loss of infra-red radiation. This is offset by absorption of visible and near visible radiation from the sun. This is absorbed mainly at the surface, so there is a local temperature maximum there. Some UV radiation from the Sun is also absorbed by ozone in the stratosphere and that is responsible for the existence of the high temperatures near the stratopause. Short wave radiation is also absorbed at great heights in the thermosphere.