2.6 Water vapour in the atmosphere

If we compress a sample of air containing water vapour at a temperature greater than 0°C, then the water vapour pressure e will increase until water begins to condense from the sample (Figure 2.1,AA_1). From this point the water vapour pressure is constant (A_1B), and is referred to as the saturated vapour pressure e_s, which is a function of temperature. Eventually a very large increase of pressure is required to produce a small change in the volume of the liquid (BB_1).

If the temperature is 0.01 °C then we find that the horizontal line A_1B represents a mixture of water vapour, water and ice. In dealing with unsaturated water vapour, it is sufficiently accurate to treat water vapour as a perfect gas with gas constant R' equal to $R_d/0.622$.

When a change of phase occurs (vapour to liquid; liquid to solid; vapour to solid, known as *sublimation*), heat is released or must be absorbed. The amount of heat required to transform one gram of water into vapour at a constant temperature is defined as the *latent heat of evaporation*, and has a value of $2.50\times10^6\,\mathrm{J\,kg^{-1}}$ (at 0 K). The *latent heat of sublimation* (ice to vapour, $2.823\times10^6\,\mathrm{J\,kg^{-1}}$ at 0 K) and the *latent heat of fusion* (ice to water, $0.333\times10^6\,\mathrm{J\,kg^{-1}}$ at 0 K) are similarly defined. The relationship between the equilibrium pressure e_s and the temperature during a phase change is given by the *Clausius–Clapeyron equation*,

$$\frac{de_s}{dT}=\frac{L_{12}}{T(V_2-V_1)} \tag{2.24}$$

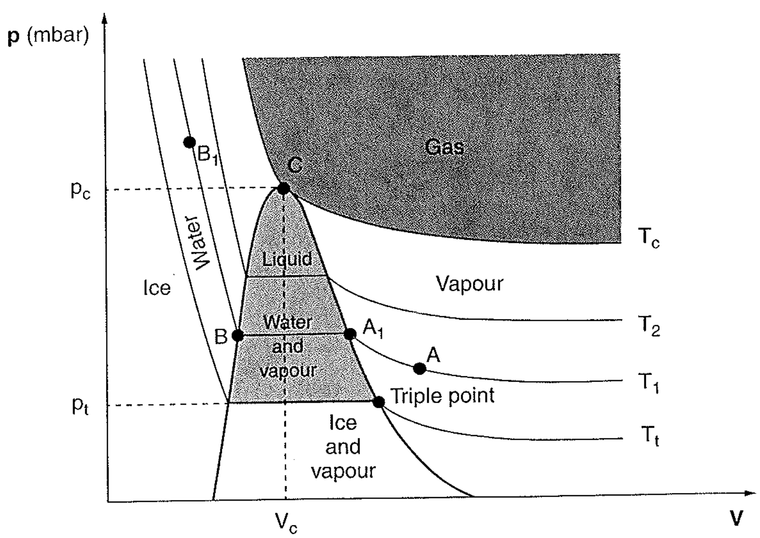

Figure 2.1 Pressure–volume graph for water known as an Amagat–Andrews diagram. Shows phase changes along isotherms in the (p, V) domain. A sample of water vapour is considered at a state corresponding to point A, i.e. at a temperature T_1 and pressure p_1 greater than the triple point temperature T_t and pressure p_t. If the vapour is compressed isothermally, the pressure increases until point A_1 is reached when liquid water and water vapour coexist in equilibrium, i.e. some water vapour has condensed to form liquid water (from Tsonis, 2002)

where L is the latent heat, and the subscripts refer to a phase change from a state defined by 1 to that defined by 2. Considering equation 2.21 for the change of phase from ice to water, it may be shown that it takes about 135 atmospheres pressure to lower the melting point of ice to –1 °C. Therefore in the atmosphere the melting point is strictly 0 °C. However, water can remain as a liquid at temperatures well below 0 °C. In these circumstances a sample of moist air can be supersaturated with respect to ice, but unsaturated with respect to supercooled water. If two surfaces, one of water and one of ice, come into contact with the moist air, then water will evaporate, but ice will grow by deposition of the water vapour. This occurs often in clouds at temperatures below 0 °C, and this is important in the formation of precipitation, as we shall see.

2.7 Atmospheric processes: saturated adiabatic lapse rate

For a sample of moist air in which no evaporation or condensation occurs, equation 2.20 may be used. However, when condensation occurs and the resulting water falls out of the sample, the mass of the sample changes and heat is lost with the fallout of the water. This is known as a *pseudo-adiabatic process*. If all the condensed water remains in the sample, then the process is of course reversible. In the atmosphere conditions are usually such that some, but not all, of the condensed water falls out of any sample of moist air.

Using equations 2.1, 2.10 and 2.24, it may be shown that for saturated air which is lifted slightly,

$$\frac{\partial T}{\partial Z} = \gamma_s = \frac{\gamma_d \left(1 + \dfrac{L x_s}{R_d T}\right)}{1 + \dfrac{L^2 x_s}{R' C_p T^2}} \tag{2.25}$$

where γ_s is the *saturated adiabatic lapse rate* (SALR) and x_s is the *saturated humidity mixing ratio*, that is, the mass of water vapour present in the moist air measured per gram of dry air when the moist air is saturated. Although γ_s varies with temperature and pressure, a typical value in the atmosphere is –5.0 °C km^{-1}.

2.8 Stability and convection in the atmosphere

Moist air can become saturated, and hence produce precipitation, by movement upwards in the atmosphere. Consider a sample, referred to as a parcel, of moist air, for which the pressure in the parcel is the same as that of its environment. Assuming that the parcel can move vertically without disturbing the environment and does not mix with its environment, it can be shown from the equations of motion for the atmosphere that

$$\frac{\mathrm{d}w}{\mathrm{d}t} = -\frac{1}{\rho}\frac{\partial p}{\partial Z} - g = \frac{(T - T')g}{T'} \tag{2.26}$$

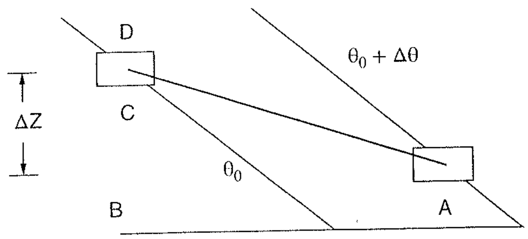

Figure 2.2 Sloping convection: an N–S cross-section with isentropic surfaces (surfaces of constant potential temperature θ) indicated, comparing the potential energy released when a parcel of air moves vertically, horizontally or intermediately

where w is the vertical velocity of the parcel, t is time, T is the temperature of the parcel, and T' is the temperature of the environment. Making the further assumption that the movement is adiabatic,

$$\frac{\mathrm{d}w}{\mathrm{d}t} = \frac{\left(T_0 - T_0'\right)g}{T_0'} + \frac{\left(\gamma - \gamma_a\right)gZ}{T_0'} \tag{2.27}$$

where T_0 and T_0' are the initial temperature of the parcel and the environment respectively, γ is the environmental lapse rate, Z is the vertical coordinate, and γ_a is the appropriate adiabatic lapse rate, being γ_d if the parcel is unsaturated and γ_s if the parcel is saturated. If the temperature is constant in the horizontal then $T_0 = T_0'$ and

$$\frac{\mathrm{d}w}{\mathrm{d}t} = \frac{\left(\gamma - \gamma_a\right)gZ}{T_0'} \tag{2.28}$$

The atmosphere is regarded as stable, neutral or unstable when $\mathrm{d}w/\mathrm{d}t$ is <0, 0 or >0, respectively. If $\gamma_s < \gamma < \gamma_d$ then the atmosphere is *conditionally unstable*, whereas if $\gamma < \gamma_s$ the atmosphere is *absolutely stable* and if $\gamma > \gamma_d$ the atmosphere is *absolutely unstable*. Hence if moist air is lifted by some means it may become saturated and hence unstable, and may then continue to rise without any external force being applied.

This release of instability in the atmosphere, or convection, is of course much more complex than this; it involves entrainment of environmental air into the rising air, and the effect of descending air on the rising air. The variability in predictability of deep convection has been studied by Done et al. (2012). Indeed, convection is not necessarily a vertical process. Consider the exchange of a parcel of air from A to B in Figure 2.2. The potential energy (energy arising from moving a parcel to a greater height against gravity) of the system does not change, and movement from B to C requires energy to raise the centre of gravity of the parcel upwards, as in the movement of the parcel of air discussed so far.

However, movement from A to D could cause the centre of gravity of the parcel to be lowered, causing energy to be released. A full explanation of this type of convection, known as *sloping* or *baroclinic* convection, is beyond the scope of this book (for a review see for example Hide and Mason, 1975; Emanuel, 1994), but it provides the basic mechanism for the development of large scale atmospheric weather systems which produce most of the precipitation in mid-latitudes. Recent numerical model studies have identified the details of this type of convection (Fantini et al., 2012).

2.9 The growth of precipitation particles

The condensation of water vapour in the atmosphere, brought about by the movement of air upwards, provides water droplets or ice crystals in clouds. Such precipitation particles are denser than the air surrounding them, and therefore they begin to fall at a rate of a few centimetres per second. However, these particles will either evaporate in unsaturated air below the cloud, or be held suspended by vertical currents within the cloud. They will only be able to reach the ground as precipitation if they become large enough to stand evaporative losses and overcome upward air motions.

Jonas (1999) summarizes precipitation microphysics. The droplet concentration is determined by the balance between the rate of supply of vapour by cooling and the rate at which it is removed by deposition. Increases in the supply of vapour, for example with increased vertical air velocity in more vigorous clouds, will result in an increase in the number and density of particles which are activated to become droplets. The concentration of droplets will be reduced by the entrainment of dry air from the environment, and therefore, for example, the droplet concentration in stratocumulus clouds is often higher than in cumulus clouds for similar aerosol particle concentrations, despite the higher updraught in the latter.

The droplet concentration is crucial to the rate of growth of the droplets (see Appendix 2.1) since they compete for the available water. Droplets grow more rapidly by condensation in clouds formed in relatively unpolluted regions than in clouds with high droplet concentrations. The growth rate of droplets by concentration decreases as the radius increases, because the decreasing surface to volume ratio limits the speed with which water can condense. Condensation also tends to produce rather narrow cloud droplet spectra due to the decreasing growth rate as the droplets become larger. Entrainment of dry air into a cloud, which results in partial evaporation of some droplets and the incorporation into the cloud of new nuclei, acts to broaden these narrow spectra.

Over a hundred years ago it was recognized that processes other than condensation were needed to cause the growth of precipitation particles on observed timescales. In 1911 Wegener proposed that rain was formed by the melting of ice particles. The atmosphere normally contains many particles which can act as nuclei for the formation of water droplets. However, there are relatively few particles in the atmosphere which can act as nuclei for the formation of ice crystals except at low temperatures. Consequently clouds rarely become glaciated, i.e. composed mainly of ice crystals rather than water droplets, until their temperature is much less than 0 °C, typically around −15 °C.

In some clouds secondary ice nucleation processes may be active (Mossop, 1978). These can increase the ice particle concentrations significantly. Secondary nucleation is sensitive to temperature and to the cloud droplet spectra. At low temperatures, around −40 °C, nucleation of cloud droplets to form ice crystals will occur even without the presence of a nucleating particle, and therefore at such low temperatures very few clouds contain supercooled drops.

The relatively high saturation mixing ratio over water compared with that over ice, combined with lower concentrations of ice crystals than of water droplets, lead to much more rapid growth of ice crystals by deposition than is the case for droplets. When a cloud of water droplets in air at close to water saturation is cooled and a

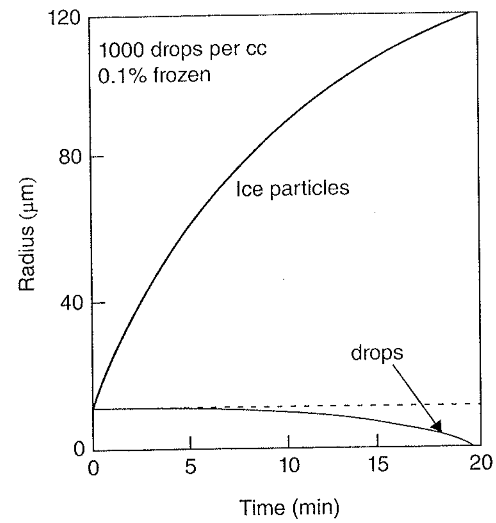

Figure 2.3 Illustrating the growth of ice spheres and decay of water drops in a mixture subject to a constant updraught of 1 m s^{-1}. The air is assumed to be initially saturated with respect to water. The dashed line shows the growth of the droplets in the absence of any ice particles (from Jonas, 1999)

small fraction of the droplets freeze, these ice particles grow very rapidly as shown in Figure 2.3. The rapid growth of the ice crystals at close to water saturation reduces the vapour mixing ratio to a value between those of saturation with respect to ice and water. The result is evaporation of the water droplets.

Ice crystals grow by sublimation when they exist in cloud together with supercooled water droplets (section 2.6). As water vapour is removed from the air by this process, the air becomes unsaturated with respect to water, so the droplets evaporate. This continues until either all the droplets have been evaporated or the ice crystals become so large that they fall from the cloud. This process takes from 10 to 30 minutes, and as the ice crystals fall they may melt to form rain (droplets with radii $\geq 20\,\mu m$) which can reach the ground. A theory of this process was derived by Bergeron in 1935, and observations confirming the theory were made by Findeisen in 1939. Hence this process of rain formation became known as the *Wegener–Bergeron–Findeisen process*, sometimes shortened to the *Bergeron process*.

This explains the formation of precipitation in mid-latitudes where clouds usually extend well above the 0 °C level; because the cloud particles normally begin as cloud condensation nuclei, it is referred to as the *cold rain process*. However, it is observed that warm clouds with tops below 0 °C also produce rain. Indeed, such clouds occur in mid-latitudes as well as in the tropics. A *warm rain process* is required. It was discovered that cloud particles normally begin to form on *cloud condensation nuclei* (CCN), which consist of partially or completely soluble aerosol particles. As the cloud particles grow by condensation or sublimation on the CCN, they begin to fall and collect other particles. The type of precipitation which is formed by such collisions depends upon the types of cloud particles present. If the cloud contains only water then rain is formed and the process is known as *coalescence*. However, if only ice crystals exist then snow results and the process is called *aggregation*.

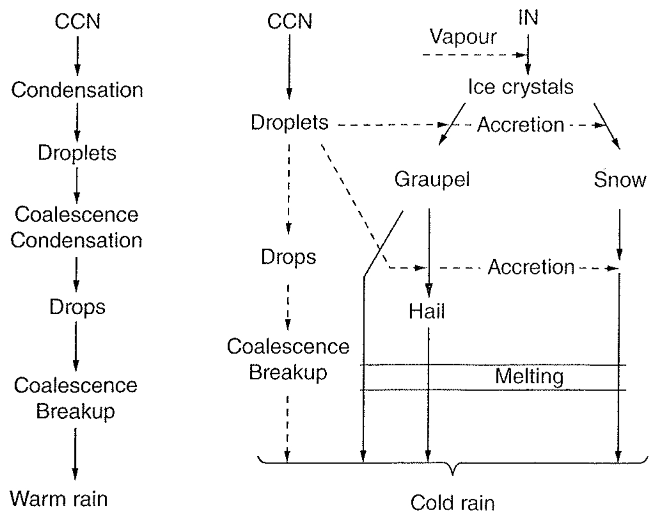

Figure 2.4 The evolution of warm and cold rain starting from cloud condensation nuclei (CCN) and ice nuclei (IN) (from List, 1977)

Owing to the very slow rate of growth of droplets by condensation once they have reached about 20 μm in radius, subsequent growth is mainly by the collision and aggregation process. The collection efficiencies of smaller cloud droplets are relatively low, as is the rate at which droplets encounter one another, owing to the small values of the terminal velocities. However, for larger droplets the process is much more effective, and can lead to the rapid growth of precipitation particles. The distance which large growing droplets fall, relative to the smaller cloud droplets, determines the maximum amount of growth that is possible within a cloud. Turbulent air motions increase the relative path of a fraction of the droplets and this increases the possibility of drizzle formation from shallow clouds such as stratocumulus. Growth by aggregation is also enhanced in clouds with high liquid water contents.

If both water droplets and ice crystals exist then ice and snow pellets (graupel) or hail may form and the process is accretion. Eventually the precipitation particles may reach such a size that they break up, beginning the process again. Detailed descriptions of these processes may be found in Mason (1971). They are sometimes collectively referred to as the *Langmuir process* in recognition of the work of Langmuir in 1948. This process may exist together with the Bergeron process. A summary is given in Figure 2.4.

There have been many reports of ice particle concentrations in convective clouds that are much higher than typical concentrations of ice nuclei (IN) (see for example Mossop et al., 1972; Hobbs and Rangno, 1985; Blyth and Latham, 1993). The large concentrations are true even taking shattering on probe tips into account. The explanation for the large concentrations in most clouds has been found to be the Hallett–Mossop process of splintering during riming (Hallett and Mossop, 1974). Freezing of supercooled raindrops to become instant rimers can be important in short lived clouds (for example Koenig, 1963). Columns of supercooled raindrops have been observed by radar (for example Caylor and Illingworth, 1987; Jameson et al., 1996).