Topic 1
Properties of clouds

1.1 Cloud types

According to the World Meteorological Organisation (WMO) there are ten basic cloud types. Like plants and animals, clouds are classified by type (genus), species and variety. We will only describe the ten genera here (for more detail on the 14 species and nine varieties see Dunlop, 2008):

1. Cirrus
2. Cirro Stratus
3. Cirro-Cumulus
4. Alto-Stratus
5. Alto-Cumulus
6. Cumulo-Nimbus
7. Strato-Cumulus
8. Stratus
9. Nimbo-Stratus
10. Cumulus

These can be broadly categorised into high, medium and low altitude clouds.

1.1.1 High-altitude clouds

These form near the tropopause at around 10 km (in the mid-latitudes).

Cirrus: are usually associated with fair weather, but could indicate an approaching warm front. They consist of only ice particles and have a feathery or wispy form. They are abbreviated Ci.

Cirro Stratus: are layer clouds which cover most of the sky. They usually herald bad weather. They are abbreviated Cs.

Cirro-Cumulus: are ice clouds usually with irregular attractive patterns. They are a heaped form of cirrus. They are abbreviated Cc.

1.1.2 Medium-altitude clouds

These develop at heights between about 2500 and 5500m.

Alto-Cumulus: Are flattened globules of cloud which are a mixture of ice and super-cooled water. They are white and grey in colour and are often the first sign that thunderstorms will follow. They are abbreviated Ac.

Alto-stratus: Are layer clouds of a dull grey colour. They are often the first sign that steady rain from nimbo-stratus is to follow. They are abbreviated As.

Nimbo-stratus: These are thicker, lower versions of altostratus and are always associated with rainfall or snow. They can be so thick that they make the day appear dark. They are abbreviated Ns.

1.1.3 Low-altitude clouds

These have bases at around 1-2km.

Stratocumulus: These clouds are usually composed of liquid water (but can be mixed-phase in the Arctic) and are grey and white with darker areas inside. Their apperance is rounded and rolled. They are abbreviated Sc.

Stratus: Usually liquid water clouds which are grey in colour. They are the lowest clouds in the sky and in fact form 'fog' if over hills or coasts. They are abbreviated St.

Cumulus: These heaped clouds are often described as 'cauliflower-like' in appearance. They are grey at the base and white at the top. They are abbreviated Cu.

Cumulo-nimbus: These are the classic anvil-headed clouds which can reach as high as 18-20 km in the tropics. They form low down and usually consist of liquid water, but at higher altitudes this water freezes to form water ice. They are abbreviated Cb.

There are other classifications of clouds such as: Lenticular clouds, Nacreous clouds, Noctilucent clouds, Mountain wave clouds, Contrails and distrails and fog/mist. For a none technical introduction to these cloud types see Lloyd (2007); Dunlop (2008).

There is another very important cloud type for the atmospheric physicist, not mentioned so far: the cloud chamber. These are used to investigate important microphysical processes occurring in clouds under controlled conditions. There are several around the world, most notably the large Aerosol Interactions and Dynamics in the Atmosphere (AIDA) chamber in Karlsruhe, Germany; the Cloud project at CERN `http://public.web.cern.ch/public/en/research/CLOUD-en.html`; and of course the Manchester Ice Cloud Chamber (MICC) `http://www.cas.manchester.ac.uk/restools/cloudchamber/`. Throughout this course I will use examples in several of these cloud types.

1.2 Cloud-scale motions

As covered in the meteorology part of the course the momentum equation is:

$$\frac{\partial \mathbf{v}}{\partial t} + \mathbf{v} \cdot \nabla \mathbf{v} = -\frac{1}{\rho}\nabla P - g\hat{\mathbf{k}} + \ldots \tag{1.1}$$

In the meteorology part of the course it was argued that the atmosphere was roughly in hydrostatic balance, which is true on the synoptic scale, and so vertical motion was ignored. On the scale of clouds and smaller convective elements this is not true. The way we often deal with this is by applying perturbation theory to the momentum equation. That is we assume that the pressure and density can be defined by a reference state that is in hydrostatic balance and small perturbations from that reference state: $P = P_0 + \delta P$ and $\rho = \rho_0 + \delta\rho$.

Therefore in the $\hat{\mathbf{k}}$ direction, the right hand side of Equation 1.1 becomes:

$$= -\frac{1}{\rho_0 + \delta\rho}\frac{\partial P_0 + \delta P}{\partial z} - g + \ldots \tag{1.2}$$

$$= -\frac{1}{\rho_0 + \delta\rho}\left(\frac{\partial P_0}{\partial z} + \frac{\partial \delta P}{\partial z}\right) - g + \ldots \tag{1.3}$$

but $\frac{\partial P_0}{\partial z} = -\rho_0 g$ (hydrostatic relation), therefore:

$$= -\frac{1}{\rho_0 + \delta\rho}\left(-\rho_0 g + \frac{\partial \delta P}{\partial z}\right) - g + \ldots \tag{1.4}$$

$$= -\frac{1}{\rho_0 + \delta\rho}\frac{\partial \delta P}{\partial z} + \left(\frac{\rho_0}{\rho_0 + \delta\rho} - 1\right) g + \ldots \tag{1.5}$$

$$\cong -\frac{1}{\rho_0}\frac{\partial \delta P}{\partial z} - \frac{\delta\rho_0}{\rho_0} g + \ldots \tag{1.6}$$

In the final equation we can see that the term associated with gravity is equal to the 'Buoyancy' force (which is derived in the Meteorology part of the course also equal to $\frac{\delta\theta}{\theta_0}g$). Hence positive vertical motion can be (and often is) the result of air being less dense than the surroundings ($\delta\rho$ is negative).

Once the air starts to move in the vertical more air will replace it (through continuity). The continuity equation is:

$$\frac{\partial \rho}{\partial t} + \nabla \cdot \rho \mathbf{v} = 0 \tag{1.7}$$

and must be solved simultaneously with the momentum equation.

Under the assumptions above this reduces to:

$$\nabla \cdot \rho_0 \mathbf{v} = 0 \tag{1.8}$$

This means that through continuity of mass, vertical motion will also result in horizontal motion.

The processes can still be adiabatic and hence conserve potential temperature, but the winds will move θ by advection.

$$\frac{d\theta}{dt} = F_{diabatic} \tag{1.9}$$

where $F_{diabatic}$ are processes that change θ (e.g. condensation, radiation and so on). Cloud and precpitation processes result in non-adiabatic processes and will now be discussed.

Thermal convection can therefore be thought of as follows:

Density anomaly Also a potential temperature anomaly that results in a buoyancy force and causes vertical acceleration / motion of the air.

Mass continuity The requirement of mass continuity means that a pressure gradient is set up that results in compensating horizontal and downward motion to replace the air rising in the thermal.

Advection The resulting wind fields advect potential temperature (under adiabatic conditions).

Toroidal circulation is set up due to the compensating downward and horizonal acceleration of air due to the pressure gradient. This causes a lot of mixing between the environmental air and the thermal air. This mixing process is usually referred to as entrainment.

Non-adiabatic processes (such as cloud formation) can significantly alter the dynamics of thermal convection through the release of latent heat.

1.3 Cloud formation mechanisms

You have covered what happens when moist air rises in the companion lectures to this part of the course. When the air is below saturation the air cools while conserving dry potential temperature, θ. When the air becomes saturated with water, the condensation of water from the vapour phase is accompanied by latent heat of vapourisation / condensation, which heats the air slightly.

The majority of clouds all form in basically the same manner when humid air is cooled below its dew point. This occurs when air comes in contact with a cold surface (and cools by conduction) or when it is forced to rise. The first process can occur at night, when the surface cools by radiating heat away to space or it can occur in coastal regions when warm air over land advects over a colder water mass. In the second process, air may be forced to rise in the atmosphere by four different mechanisms:

- Through heating of the ground during the day, which creates bubbles of warm air that break away from the surface as ‘thermals’, in the process known as convection;
- Through being forced to rise over a mountain range or similar barrier, which is known as orographic uplift;
- Through uplift at frontal surfaces in a depression: frontal uplift.
- Through a process known as convergence, where air flows into an area from different directions. When this occurs at the surface, there is only one way in which the accumulation of air can escape (due to continuity) and so it is forced to rise.

There is also another way that clouds can form in cloud chambers. Usually the method is to either hold the chamber at 1000 hPa in a pressure vessel, with a liquid or ice coating on the walls of the chamber. Then we suck the air out of the chamber using a fast vacuum pump to around 750 hPa, typically. Another way would be to introduce water vapour using a boiler (a kettle also works) until the air saturates with water vapour. Both these methods have their pros and cons.

1.4 Some definitions

Cloud formation occurs when air becomes saturated with water vapour. The Clausius-Clapeyron equation governs the equilibrium vapour pressure of water vapour, which you will have covered in thermodynamics (see `http://en.wikipedia.org/wiki/Clausius%E2%80%93Clapeyron_relation`):

$$\frac{de}{dT} = \frac{L_v}{T\Delta V} \tag{1.10}$$

where e is the pressure of water vapour, T is the temperature, L_v the latent heat of vapourisation and $\Delta V \cong \frac{R_v T}{e}$ the change in specific volume from the initial phase to the final phase. R_v is the specific gas constant for water vapour. Solving this equation yields for the vapour pressure over a liquid surface:

$$e_{sw} = 610.7 \exp\left(\frac{L_v}{R_v}\left[\frac{1}{273.15} - \frac{1}{T}\right]\right) \tag{1.11}$$

and for over an ice surface:

$$e_{si} = 610.7 \exp\left(\frac{L_s}{R_v}\left[\frac{1}{273.15} - \frac{1}{T}\right]\right) \tag{1.12}$$

where here L_s is the latent heat of sublimation (phase change of ice to vapour).

Mixing ratio has already been defined, but we define saturated vapour mixing ratio

$$w_s(T,P) = \frac{\epsilon e_s(T)}{P} \tag{1.13}$$

which can be defined for either liquid water or ice with the corresponding saturation vapour pressure (this comes from the ideal gas law applied to both water vapour and air).

Previously relative humidity was defined as e/e_s. We will stay with this definition, but also call this saturation ratio. We will refer to saturation ratio over liquid s_l and over ice s_i.

The supersaturation is not considered in these first few lectures, but we will cover it later. It is defined as:

$$S_l = s_l - 1 \tag{1.14}$$

$$S_i = s_i - 1 \tag{1.15}$$

I will use these terms throughout the lectures.

Important note for this course: from this point forward many of the examples in this course will use parcel theory as the equations of motion are more the subject of PhD in atmospheric physics. A type of cloud where parcel theory can prove useful is stratocumulus or a cloud chamber.

1.5 Adiabatic liquid water mixing ratio

We first consider liquid only clouds here. Once clouds start to form in an expanding air parcel the temperature change no longer conserves dry potential temperature, θ since heat is added by the latent heat of vapourisation.

In the accompanying lectures you have derived an expression for the pseudo-adiabatic lapse rate. This was derived assuming that liquid water is removed from the parcel immediately (and so it did not change the heat capacity of the air). In practice the change in heat capacity of the air due to liquid water is negligible for most applications.

It can be shown that during a moist adiabatic process where liquid water alters the temperature of the air such that the quantity:

$$\theta_{q,sat} = T\left(\frac{100kPa}{P}\right)^{R'/(c_p+c_w\times Q)} \exp\left[\frac{w_s L_v}{T(c_p+c_w Q)}\right] \tag{1.16}$$

is conserved. Those interested in a derivation can look at Rogers and Yau (1989). Here, c_w is the heat capacity of liquid water and Q is the total water mixing ratio, which is conserved in an adiabatic parcel.

The adiabatic liquid water mixing ratio (kg/kg) can be defined as the difference between the saturated vapour mixing ratio at the point of interest and the saturated vapour mixing ratio at cloud base (why is this?):

$$ALMR = w_s(T_1, P_1) - w_s(T_2, P_2) \tag{1.17}$$

where T_1 and P_1 are the cloud base temperature and pressure and T_2 and P_2 are the temperature and pressure where the adiabatic liquid water mixing ratio is to be evaluated; $w_s(T, P)$ is the saturation vapour mixing ratio, which has been previously defined.

From this, the saturation vapour mixing ratio (Equation 1.13) and the saturation vapour pressure (Equation 1.11) we can calculate the adiabatic liquid water mixing ratio.

$$ALMR \cong \frac{\epsilon e_{sl}(T_1)}{P_1} - \frac{\epsilon e_{sl}(T_2)}{P_2} \tag{1.18}$$

The main problem in solving for $ALMR$ is that when you equate $\theta_{q,sat}$ at cloud base at P_1, T_1 and $\theta_{q,sat}$ within the cloud at P_2 it is not possible to solve analytically for T_2 (the temperature in the cloud). Iteration can be used or the following approximation can be used.

1.6 Solving the adiabatic liquid water mixing ratio equation

Firstly we can approximate Equation 1.16 relation (by using the usual approximation $\exp(x) \cong 1 + x$) as:

$$\theta_{q,sat} \cong T\Pi\left(1 + \frac{w_s(T,P)L_v}{Tc_p}\right) \tag{1.19}$$

where the approximation $\exp(x) \cong 1 + x$ has been applied and $\Pi = \left(\frac{100kPa}{P}\right)^{R/c_p}$ Secondly, substituting Equation 1.13 into Equation 1.19 yields:

$$\theta_{q,sat} \cong T\Pi\left(1 + \frac{\epsilon e_s(T)L_v}{PTc_p}\right) \tag{1.20}$$

or

$$\theta_{q,sat} \cong T\Pi\left(1 + \frac{1.5473 \times 10^3 e_s(T)}{PT}\right) \tag{1.21}$$

The saturation vapour pressure (Equation 1.11) can be approximated to:

$$e_{sw} = 610.7\exp\left(\frac{L_v}{R_v}\left[\frac{1}{273.15} - \frac{1}{T}\right]\right) \tag{1.22}$$

$$\cong 610.7\exp\left(\frac{L_v}{R_v}\left[\frac{T - 273.15}{273.15^2}\right]\right) \tag{1.23}$$

because $273.15^2 \cong 273.15T$.

Thirdly, making the usual approximation to Equation 1.11 (e.g. $\exp(x) \cong 1 + x$) we have that:

$$e_{sw} \cong 610.7\exp\left(\frac{L_v}{R_v}\left[\frac{T - 273.15}{273.15^2}\right]\right) \tag{1.24}$$

$$\cong 610.7\left(1 + \frac{L_v}{R_v}\left[\frac{T - 273.15}{273.15^2}\right]\right) \tag{1.25}$$

and substituting this in Equation 1.20 and rearranging yields:

$$\frac{\theta_{q,sat}}{\Pi_2} \cong T_2 + \frac{\epsilon AL_v}{P_2c_p}\left(1 + \frac{L_v}{R_v}\left[\frac{T_2 - 273.15}{273.15^2}\right]\right) \tag{1.26}$$

where $A = 610.7$, P_2 and T_2 are the pressure and temperature in the cloud and Π_2 is Π evalulated at this pressure. Fouthly Equation 1.26 can be rearranged to a linear equation in T_2:

$$0 \cong T_2 + \frac{\epsilon AL_v}{P_2c_p} - \frac{\theta_{q,sat}}{\Pi_2} - \frac{\epsilon AL_v^2}{P_2c_pR_v \times 273.15} + \frac{\epsilon AL_v^2T_2}{P_2c_pR_v \times 273.15^2} \tag{1.27}$$

which simplifies to:

$$T_2 \cong \frac{\frac{1.7815\times10^7}{P_2} + \frac{\theta_{q,sat}}{\Pi_2}}{1 + \frac{6.868\times10^4}{P_2}} \tag{1.28}$$

where T_2 and P_2 are the temperature and pressure at the point we are interested in. Calculation of the adiabatic liquid water mixing ratio can then be done by substituting T_2 into Equation 1.18.

THE KEY POINTS TO TAKE HOME HERE ARE:

- If a parcel of air ascends and does not saturate the temperature variation with changing pressure will conserve potential temperature, θ.
- Once the parcel of air saturates the temperature variation with changing pressure will conserve $\theta_{q,sat}$.