Hydrostatic Stability

We saw earlier that in the radiative equilibrium profile, temperature in the lower troposphere decreases with height much faster than what is observed. It turns out that the radiative equilibrium profile is *hydrostatically unstable* in the lower troposphere. This gives rise to convection, which imposes its own temperature structure. In the following, we will look at the stability of a fluid in hydrostatic balance.

1. The incompressible fluid case

First consider the hydrostatic stability in an incompressible fluid ($\partial\rho/\partial p = 0$). Water is a reasonable example. Suppose the fluid is horizontally homogeneous but the density changes with height. As a thought experiment, it doesn't matter but in reality you could generate density gradient by having water with different temperatures or by adding different amounts of salt to the water; salty water is denser. These will turn out to be very important for ocean circulation.

Our strategy is to introduce a small vertical displacement to a parcel and see whether the parcel will bounce back (in which case it will be stable) or drift away (in which case it will be unstable). Let us start with the case $d\rho/dz < 0$, i.e. density decreases with height, and the fluid is at rest and in hydrostatic equilibrium. If a fluid parcel is displaced upward by the distance Δz, its density will be greater than its soundings by $\Delta z\, d\rho/dz$, and it will experience downward buoyancy[1] acceleration $g\Delta z\, d\rho/dz/\rho$, so that

$$\frac{d^2\Delta z}{dt^2} = \frac{gd\rho}{\rho dz}\Delta z$$

Such a situation is stable, and the parcel will oscillate around its original point with a frequency of $\left(\frac{-gd\rho}{\rho dz}\right)^{1/2}$, which is called the buoyancy frequency. Conversely, the fluid is unstable to infinitesimal perturbations if $d\rho/dz > 0$. The situation with $d\rho/dz = 0$ is called neutral.

If the density gradient is due to temperature gradient, so that $\rho = \rho_{ref}(1-\alpha(T-T_{ref}))$, where α is the volume thermal expansion coefficient, the stability criterion becomes $dT/dz = 0$.

To be more precise, dT/dz needs to be slightly negative for convection to occur. For a fluid bounded by two plates, this was worked out by Rayleigh (1916): convection occurs when the Rayleigh number exceeds a threshold (~1000). The Rayleigh number, defined as
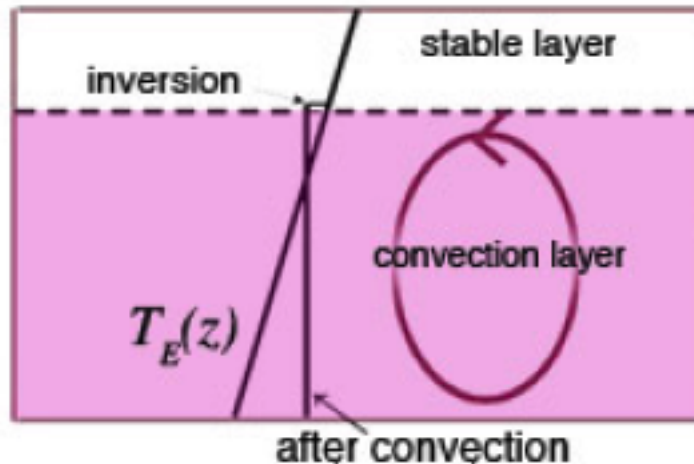
$$R_a = \frac{\Delta Tg\alpha H^3}{\nu\kappa}$$

where H is the distance between the two plates, and the denominator is the product of the molecular diffusion coefficients of momentum and heat. The Rayleigh number is a measure of the relative importance of the convective and molecular heat transport. For

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[1] Buoyancy is known to Archimedes, 260BC.

At this link (http://paoc.mit.edu/labweb/experiments.htm, go to GFDII), you can find a movie of convection in a water tank. An often overlooked region is the entrainment zone.

In the atmosphere, density decreases with height. Should I infer that it is always hydrostatically stable?

The answer is of course no. The atmosphere is not incompressible at all. When an air parcel is displaced vertically, it quickly adjusts to the pressure of its new environment to maintain mechanical equilibrium. This causes its temperature and density to change according to the ideal gas law. How do we then analyze the hydrostatic stability of the atmosphere?

2. Extension to compressible fluid

We shall consider movement of the air parcel to be fast, so that effects from thermal conduction and radiation can be neglected. In this case, there is no heat exchange between this parcel and its environment. This is called adiabatic. Variations of density as a parcel adjusts to a new pressure adiabatically can be derived from the ideal gas law and the first law of thermodynamics, the latter may be written as

$$dh - vdp = \delta q$$

where h is the enthalpy, and for ideal gas, $dh = c_p dT$, where c_p is the specific heat at constant pressure. We use this form of the first law because we want to relate temperature to pressure. For adiabatic processes, $\delta q = 0$ so we have

$$c_p dT - vdp = 0 \qquad (*)$$

Divide both sides by T and use the ideal gas law pv=RT, we have

$$c_p d\ln T - Rd\ln p = 0$$

Integration leads to

$$T^{c_p} p^{-R} = const$$

This leads to a conserved quantity for dry air in adiabatic processes called the potential temperature:

$$\theta = T \left(\frac{p}{p_{ref}} \right)^{-R/c_p}$$

By convention, a value of 10^5 Pa is used for p_{ref}. For ideal gases with diatomic molecules, according to statistical mechanics, the internal energy is $5/2RT$ so the specific heat at constant volume is $c_v=5/2R$, and the specific heat at constant pressure is $c_p=7/2R$ (because $c_pT=h=u+pv=c_vT+RT$).

If a parcel quickly adjusts so that it has the same pressure as its surroundings, to compare its density to that of its surroundings, we only need to know their T or θ. Therefore if $d\theta/dz>0$, a displaced parcel will experience a restoring force and if $d\theta/dz<0$, it will be hydrostatically unstable (note θ is inversely proportional to ρ given pressure). $d\theta/dz=0$ is neutral.

From Eq. (*), if one assumes that parcel displacement is always in hydrostatic balance, then $dp=-\rho gdz$ and we have

$$c_p dT + gdz = 0$$

From this, one can define a quantity called dry static energy$=c_pT+gz$, which is conserved following hydrostatically balanced adiabatic motions. In practice, it's conserved approximately and can be more convenient to use.

We can of course derive the stability criterion in terms of dT/dz. With the definition of the dry static energy, we have $-dT/dz=g/c_p$ as the stability criterion. $-dT/dz$ is called the lapse rate. If temperature decreases with height faster than this (~10K/km), it is unstable. Expressing the criterion in terms of temperature is only for convenience in comparing with a temperature profile. The potential temperature (or the dry static energy), instead of temperature, is the most relevant quantity here.
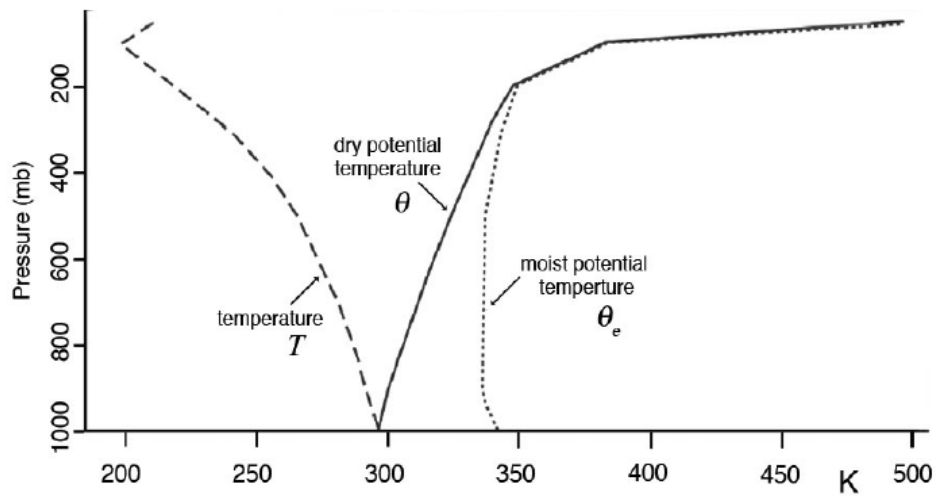
Figure 4.9: Climatological atmospheric temperature T (dashed), potential temperature θ (solid) and equivalent potential temperature θ_e (dotted) as a function of pressure, averaged over the tropical belt $\pm 30°$.

Inspection of the observed temperature profile indicates that the lapse rate is ~5K/km in the lower troposphere.

$$\frac{dT}{dz} = \frac{T(500hPa) - T(1000hPa)}{z(500hPa) - z(1000hPa)} = \frac{(270 - 295)K}{(5.5 - 0.1)km} \sim -5K / km$$

Should we conclude that the troposphere is mostly hydrostatically stable thus free of convection? (We could also look more directly at how potential temperature changes with height)

The atmosphere is indeed mostly stable to dry convection. But the release of latent heat from the condensation of water vapor as the air expands and cools can make it unstable to moist convection.

Before we move to moist convection, let's look at the dry case a bit more. First look at what happens in the stable condition. Like in the incompressible case, a vertically displaced parcel will bounce back and oscillate with the buoyancy frequency. This gives rise to gravity waves, which we will discuss in more detail in later lectures. The only difference here is that unlike in the incompressible case where density/temperature is conserved following adiabatic motion, here it is the potential temperature. Therefore, the buoyancy frequency N is defined by

$$N^2 = g\frac{d\ln\theta}{dz}$$

and is ~1.e-2/s in the troposphere.

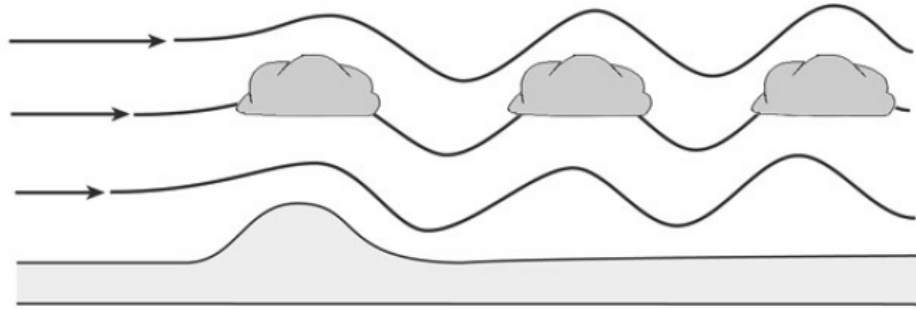Some examples of gravity waves from Marshall and Plumb:

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Figure 4.13: A schematic diagram illustrating the formation of mountain waves (also known as lee waves). The presence of the mountain disturbs the air flow and produces a train of downstream waves (cf., the analogous situation of water in a river flowing over a large submerged rock, producing a downstream surface wave train). Directly over the mountain, a distinct cloud type known as lenticular ("lens-like") cloud is frequently produced. Downstream and aloft, cloud bands may mark the parts of the wave train in which air has been uplifted (and thus cooled to saturation).

Figure 4.14: Looking downwind at a series of lenticular wave clouds in the lee of the Continental Divide of North America. (Photo courtesy of Dale Durran.)

Figure 4.15: Atmospheric gravity waves formed in the lee of Jan Mayen island (only 50 km long situated 375 miles north-northeast of Iceland), observed in February 2000. The wind is blowing from the WSW. A volcano — called Beerenberg — forms the north end of the island and rises to a height of over 2 km. Note the similarity between the atmospheric wake and that formed on a water surface by a ship, superimposed on the figure.

3. Hydrostatic stability of a moist atmosphere

Before we discuss the hydrostatic stability of a moist atmosphere, let's first introduce how we might describe moist air

Specific humidity

$$q = \frac{\rho_v}{\rho} = \frac{m_v}{m}$$

Mixing ratio

$$r = \frac{m_v}{m_d} = \frac{q}{1-q} \sim q$$

Without condensation, r and q are conserved. Despite this convenience, for discussions of chemical equilibrium, it is the absolute concentration that matters.

Now in order to calculate the density, one needs to know the specific humidity (or mixing ratio) in addition to the pressure and temperature. According to the Gibbs-Dalton law of partial pressures, vapor and dry air each behave as if the other were absent. We have

$$p_d = \rho_d R_d T$$
$$e = \rho_v R_v T$$

(1)

where e is vapor pressure,

Add together the two equations in (1), we have

$$p = (\rho_d R_d + \rho_v R_v)T$$
$$= \rho R T$$
$$R = (1-q)R_d + qR_v$$
$$= R_d \left[1 + \left(\frac{1}{\varepsilon} - 1 \right) q \right]$$
$$\sim R_d (1 + 0.61q)$$

where $\varepsilon = \dfrac{R_d}{R_v} = \dfrac{M_v}{M_d} \approx 0.622$. Thus, for unsaturated air at the same pressure and

temperature, the one with more water vapor is lighter because of the lower molecular weight of water vapor. One often defines the virtual temperature $T_v = T(1 + 0.61q)$ to absorb this dependence into temperature. In the tropics near the surface, q can reach 20g/kg, so its density is the same as a dry parcel that is 3-4K higher in temperature.

When an air parcel with water vapor is cooled, water vapor may condense if the saturation vapor pressure is lower than the vapor pressure. The saturation vapor pressure is described by the Clausius-Clapeyron equation, which can be derived assuming the condensed phase and the vapor phase are in thermal, mechanical, chemical equilibrium.

In this case, the Gibbs free energy, temperature, and pressure are the same for the two phases:

$$dG = -s_v dT + v_v de_s$$
$$dG = -s_c dT + v_c de_s$$

where e is the saturation vapor pressure. From this, we have
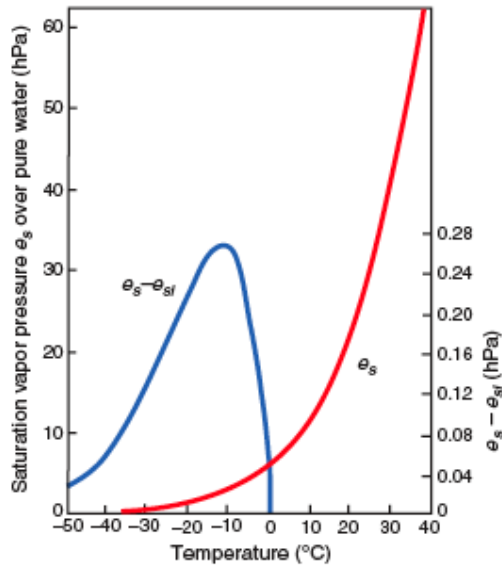
$$\frac{de_s}{dT} = \frac{L}{T \Delta v}$$

where L=TΔs is the latent heat for the phase transition, and Δv is the change in specific volume from the phase transition. For phase transition between vapor and a condensed phase, the specific volume of the condensed phase is small, so $\Delta v = R_v T / e_s$ and we have

$$\left(\frac{d \ln e_s}{dT} \right) = \frac{L}{R_v T^2}$$

Assuming L to be constant (roughly true), we have

$$e_s = e_s(T_0) \exp \left[-\frac{L}{R_v} \left(\frac{1}{T} - \frac{1}{T_0} \right) \right]$$

Note it is only a function of T, and does not depend on P. The saturation vapor pressure over water, and the difference between the saturation vapor pressure over water and ice are shown here:

A useful formula for saturation vapor pressure over liquid water in practice that is accurate to within 0.3% between -35C and 35C is

$$e_s = 6.112 \exp\left(17.67 \frac{T - 273.15}{T - 29.65}\right)$$

(Bolton, 1980, Monthly Weather Review, 108, 1046-1053)

The saturation vapor pressure roughly decreases by a factor 2 every 10 degree decrease in the temperature. It is also useful to define relative humidity RH=e/e*, where e* is the saturation vapor pressure[2].

Let's now examine the thermodynamic behavior of an air parcel accompanying vertical motion. Starting with unsaturated air. In this case, the relative concentration (e.g. the mixing ratio) is conserved, and as pressure decreases with height, the vapor pressure also decreases. However, the saturation vapor pressure decreases much faster because of the decrease in temperature due to adiabatic expansion. For example, if a parcel is moved up 1km unsaturated, its temperature drops by ~10K so its saturation vapor pressure drops by about a factor of 2. While its pressure also drops, it's only by ~10% so its saturation mixing ratio (or saturation specific humidity) still decreases by about a factor of 2. At some point, the saturation vapor pressure becomes lower than the vapor pressure and condensation occurs. This is defined as the lift condensation level (LCL), which marks the base of clouds.

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[2] That saturation vapor pressure over ice is lower than that over water gives rise to the Bergeron process. As will be discussed a bit later, the freezing of liquid water droplets to ice requires freezing nuclei, which are scarce, so liquid water droplets can exist well below 0C (but above about -40C) and called supercooled water. Once ice starts to form, however, because of the lower saturation vapor pressure over ice, ice crystals will grow at the expense of the liquid water droplets, known as the Bergeron process.
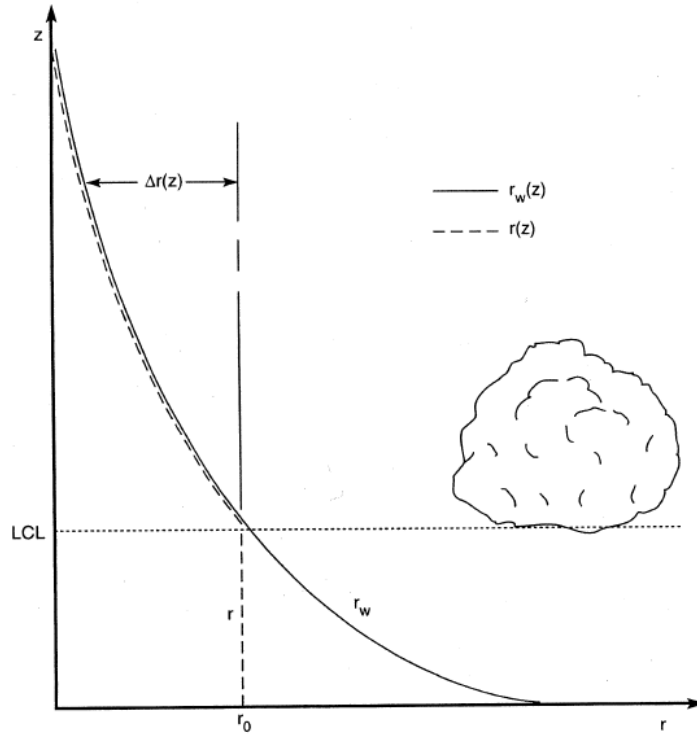
Figure 5.3 Vertical profiles of mixing ratio r and saturation mixing ratio r_w for an ascending air parcel below and above the *lifting condensation level* (LCL).

As air continues to rise above the LCL, temperature continues to drop, and more and more vapor will condense and release latent heat during the process. If temperature drops below 0C, ice will form, adding latent heat of fusion. If all condensates are retained in the parcel, when the parcel starts to move downward, the reverse to what we just described will happen. This is called a reversible saturated adiabatic process. In reality, condensates will fall under gravity. The speed that condensates fall is set by the balance between gravity and the aerodynamic drag, and is called the terminal velocity. As the drag is proportional to the surface area but gravity force is proportional to the volume, larger particles have larger terminal velocities. An mm-sized raindrop has a terminal velocity of a few m/s, hailstones can fall at a speed of above 10m/s (ever got caught in a hail storm?). Cloud droplets are small (a few microns) so that its terminal velocity is only ~1cm/s and is in general considered moving with the air parcel. If we assume as the other extreme that the condensates are removed instantaneously (called the pseudo-adiabatic process), then when air moves downwards, there is nothing to evaporate/sublimate and the air will follow a dry adiabatic (θ=const). This gives rise to the Chinook (or "snow eater" in Indian) phenomena when there is downslope wind in North America.

Including water vapor in the system, we have from the first law of thermodynamics, in an approximate form:

$$c_p dT + d(Lq) - \frac{dp}{\rho} = 0 \tag{2}$$

Divide both sides by T and use the ideal gas law, we have

$$c_p d\ln T + \frac{1}{T}d(Lq) - Rd\ln p = 0$$

$$d\ln\theta = -\frac{1}{Tc_p}d(Lq) \approx -d\left(\frac{Lq}{Tc_p}\right)$$

In the last step, we made an approximation[3] based on the fact that the fractional change in T is small compared to the fractional change in q. This allows us to define the equivalent potential temperature

$$\theta_e = \theta\exp\left(\frac{Lq}{c_pT}\right)$$

which is approximately conserved in adiabatic processes with or without condensation. Note that although θ_e is conserved in adiabatic processes with or without condensation, it can only be used to compare the densities (or temperatures) of two parcels at the same pressure when both are saturated. Therefore, to use θ_e in an analogous matter as θ in a dry atmosphere (i.e. $d\theta_e/dz<0$ implies instability), the whole layer in question must be saturated. This is called potential instability and can happen when an initially unsaturated layer is lifted so that the whole layer saturates.

A somewhat simpler conserved quantity can be derived if the motion is strictly in hydrostatic balance, in which case Eq. (2) becomes

$$c_p dT + d(Lq) + gdz = 0 \qquad\qquad (3)$$

And we can define moist static energy $c_pT+gz+Lq$, which is conserved in adiabatic, hydrostatic motions. In practice, it is conserved to a good approximation.

Now we shall estimate how temperature will change with height for a saturated parcel. Its utility will become apparent. In saturated air, q is equal to saturation specific humidity q^*, and one can use Eq. (3) to get the lapse rate. The basic behavior is of course simply a reduced lapse rate because of the latent heat release. The following shows how one might quantify this (Going from the first equation to the second one, we have used the hydrostatic relation).

$$c_p dT + L\frac{\partial q^*}{\partial T}dT + L\frac{\partial q^*}{\partial p}dp + gdz = 0$$

$$\left(c_p + L\frac{\partial q^*}{\partial T}\right)dT + g\left(1 - \rho L\frac{\partial q^*}{\partial p}\right)dz = 0$$

$$-\frac{dT}{dz} = \frac{g}{c_p}\frac{1 - \rho L\dfrac{\partial q^*}{\partial p}}{1 + \dfrac{L}{c_p}\dfrac{\partial q^*}{\partial T}}$$

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[3] For a derivation without this approximation, see the book by Emanuel, Atmospheric convection (1994), section 4.5. The resulting equivalent potential temperature has a more complicated form but will be exactly conserved for reversible adiabatic motions.

$-\rho L \dfrac{\partial q^*}{\partial p}$ is positive but tends to be small (<0.1), while the value of the denominator can be quite large (3-4) near the surface. Given T and p, one can evaluate the partial derivatives of q* with respect to T and p with the aid of the Clausius-Clapeyron equation and the definition of specific humidity. This is called the moist (or wet) adiabat lapse rate. Note it is a function of T and approaches the dry one at low temperatures. A schematic is shown here:
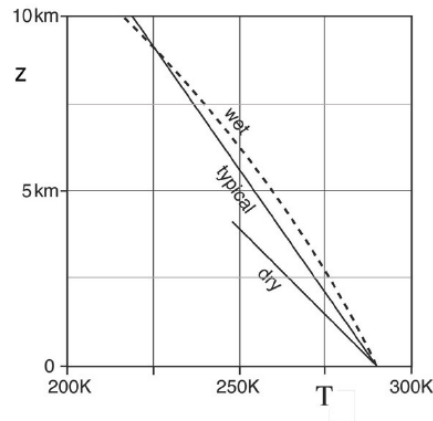

Figure 4.19: A schematic of tropospheric temperature profiles showing the dry adiabat, a typical wet adiabat, and a typical observed profile. Note that the dry adiabatic ascent of a parcel is typically cooler than the surroundings at all levels, whereas the wet adiabat is warmer up to about 10 km. The wet and dry lapse rates are close to one another in the upper troposphere, where the atmosphere is rather dry.

Below (Salby Fig. 7.6) shows is a simple representation of the tropical troposphere (although the LCL level shown is higher than a typical tropical condition), and air is unsaturated everywhere. What's the hydrostatic stability of this atmosphere?
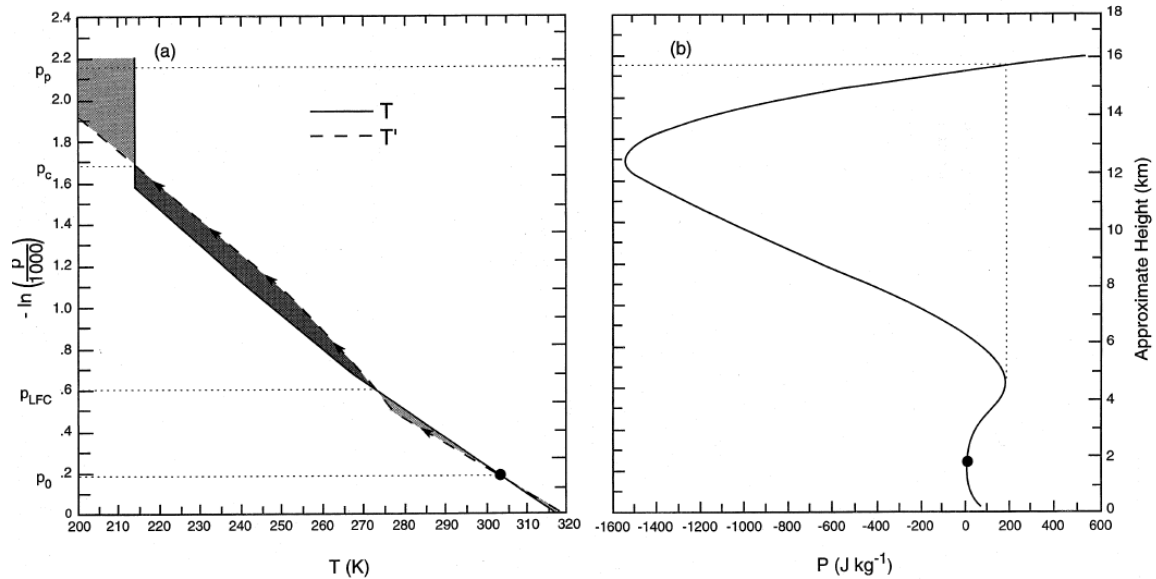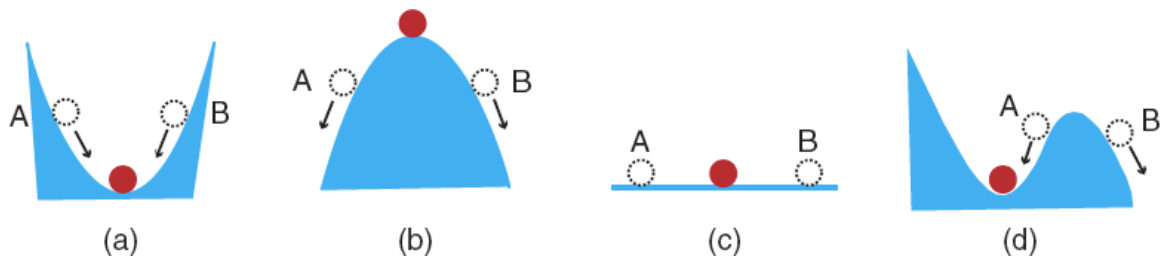
Figure 7.6 (a) Environmental temperature (solid line), under conditionally unstable conditions representative of the tropical troposphere ($\Gamma \cong 8.5\ \mathrm{K\,km^{-1}}$) and stratosphere, and parcel temperature (dashed line) displaced from the undisturbed level $p_0 = 800$ mb. Parcel temperature decreases with height along a dry adiabat below the lifting condensation level (LCL) and along a saturated adiabat above the LCL. It crosses the profile of environmental temperature at the *level of free convection* p_{LFC} and again at the crossing level p_c above the tropopause. (b) Potential energy of the displaced parcel, which is proportional to minus the cumulative area $A(p)$ under the parcel's temperature profile and above the environmental temperature profile that is shaded in (a).

For infinitesimal perturbations (at any height), the air parcel will be unsaturated and the atmosphere is stable to such perturbations. If we lift a parcel from near the surface, it acquire negative buoyancy, but if it has enough kinetic energy to overcome the barrier and reach the level of free convection (LFC), then it becomes positively buoyant and can rise to a much higher level of neutral buoyancy (LNB) that is close to the tropopause. This is an example of finite amplitude instability even though the atmosphere is stable to infinitesimal perturbations. This is termed conditional instability. Because of the rapid increase in stability across the tropopause, much of the convective air will reach their LNB near the tropopause and fan out to produce the anvils. The parcels in fact will rise higher than its LNB because of its non-zero kinetic energy, the resulting penetrative entrainment may be important to the tropopause region.

An illustration of finite amplitude instability is shown in (d).

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A picture over the Amazon (from the cloud appreciation society)

A quantity often used to describe the conditional instability is the convective available potential energy (CAPE). It is related to the maximum vertical velocity that the parcel can achieve if its velocity is zero at its LFC by integrating the buoyancy force upward between its LFC and its LNB.

$$\frac{1}{2}w_{\max}^2 = \int_{LFC}^{LNB} \frac{\rho_e - \rho}{\rho} g\,dz = \int_{LFC}^{LNB} \frac{1 - \dfrac{\rho}{\rho_e}}{\rho} \rho_e g\,dz$$

$$= \int_{LFC}^{LNB} -\frac{1 - \dfrac{T_e}{T}}{\rho} dp = \int_{LFC}^{LNB} -\frac{T - T_e}{\rho T} dp$$

$$= R \int_{LFC}^{LNB} -\frac{T - T_e}{p} dp = R \int_{LNB}^{LFC} (T - T_e)\,d\ln p$$

Typical values for CAPE (depending on the parcel of course) are hundreds of J/kg. A similarly defined value for the area with negative buoyancy is called the convective inhibition.

The existence of conditional instability is special to a moist atmosphere. There is no analog in a dry atmosphere. A consequence is that when the barrier is sufficiently high, one can build up the instability then release it all at once when the barrier is eroded. This

is a factor behind severe thunderstorms in central U.S., which can have CAPE values of a few thousands of J/kg. Such situations however are rare. More generally, CAPE is removed as it is generated (by warming and moistening of the surface and cooling (radiative or adiabatic) of the atmosphere), a situation called quasi-equilibrium.

Note that we have made a number of assumptions here. Here are two of the more important ones: we assumed the motion is hydrostatic, which is not true for small convective elements; we also assumed that there is no mixing with the environment (the parcel rises like a ping-pong ball). Both assumptions make the estimated vertical velocity too high. In general, the effect of mixing is smaller for bigger clouds/air parcels.
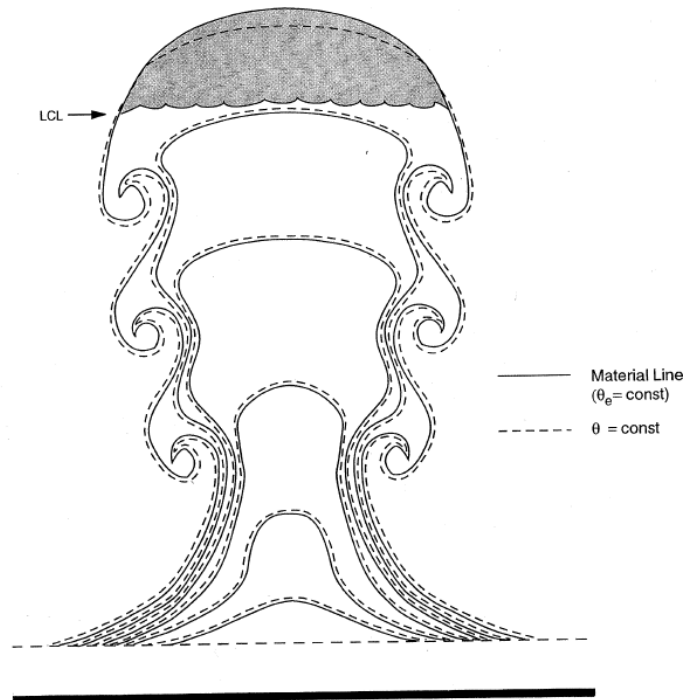

Figure 5.4 An ascending plume of moist air that develops from surface air that has become positively buoyant. Material lines (solid), which are defined by a fixed collection of air parcels, remain coincident with isentropes (θ=const) below the LCL, but they drift to higher θ and higher altitude above the LCL, where air warms through latent heat release. Although they depart from isentropes, material lines remain coincident with pseudo-isentropes (θ_e = const) above the LCL.

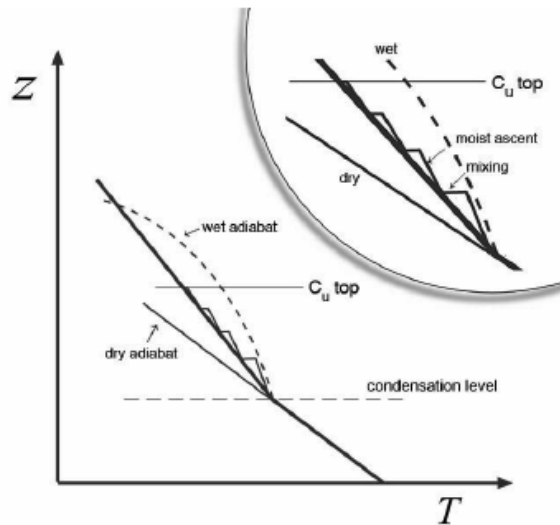The following figure illustrates the effect of strong lateral entrainment.

Figure 4.24: In cumulus convection buoyant air parcels ascend along a moist adiabat, but repeatedly mix with ambient fluid, reducing their buoyancy. Cumulus top is expected to occur at the level at which the wet adiabat first runs parallel to the environmental curve, as shown in the inset at top right of figure.

Convective adjustment

When the atmosphere becomes hydrostatically unstable, convection develops. It releases the instability and drives the atmosphere towards a neutral state. For a dry atmosphere (or water tank), the adjusted state is that of a roughly constant θ (or density), and it does so through turbulent mixing; parcels mix so they all have the same property. Moist convection will also drive the atmosphere back to a neutral state, i.e. that with a moist adiabatic temperature profile so air lifted from near surface will no longer be buoyant. However, the time it takes for moist convection to completely mix is long. While the ascent of clouds is fast (10m/s) and it only takes ½ hr to go across the troposphere, the descent of clear air (if one view this as the returning leg) is very slow (<cm/s) and can take a month to go across the troposphere. The much more efficient way to restore a neutral state is through the spreading of gravity waves. We will talk more about gravity waves later, but the basic idea is that the compensating subsidence in an atmosphere that is stable to dry convection warms the atmosphere.
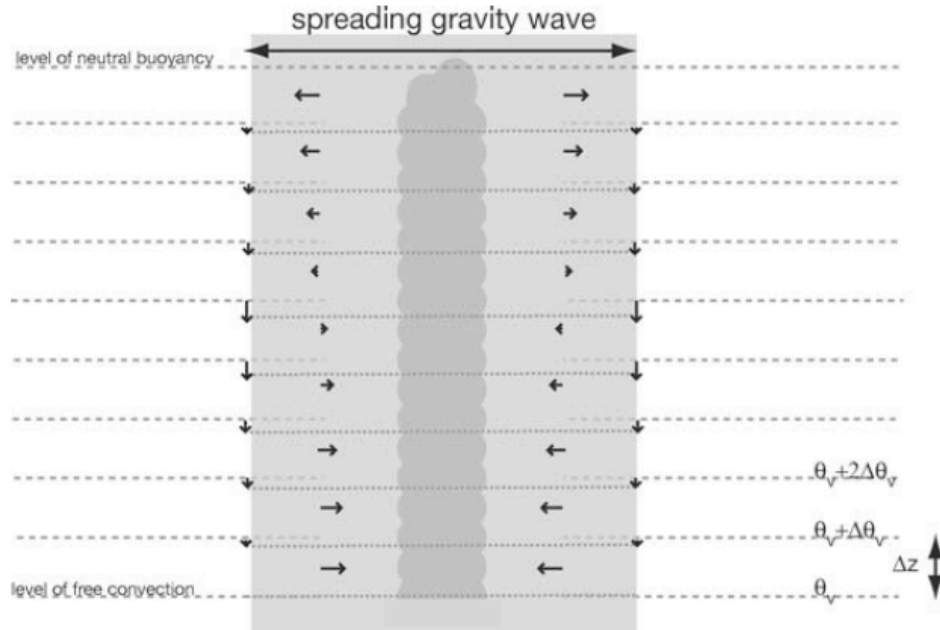
Figure 10 Schematic of adjustment owing to cumulus heating following Bretherton & Smolarkiewicz (1989). Here the transient adjustment to the moist-adiabatic lapse rate of the convecting tower from Figure 3 is shown to be accomplished by a spreading gravity wave front. As the subsidence wave moves away from the region of convection, it is associated with a downward displacement of isentropes, which equilibrates the buoyancy (θ_v) profile.

Radiative-convective equilibrium

How does the atmosphere stay convective (at least in some parts)? It is through the continuous radiative forcing (and surface fluxes) that the atmosphere is driven towards a hydrostatically unstable state. The actual atmospheric state is thus a balance between these two processes. This is called the radiative-convective equilibrium. Convective adjustment is fast (hrs) and radiative adjustment is slow (days) so the radiative-convective equilibrium profile is close to a convectively neutral (or a moist adiabat) profile (convection wins). But this shouldn't obscure the fact that radiation is fundamentally responsible for driving convection.

Using 1D radiative convective equilibrium models, one can already ask a variety of questions about the climate. One is the runaway greenhouse effect.
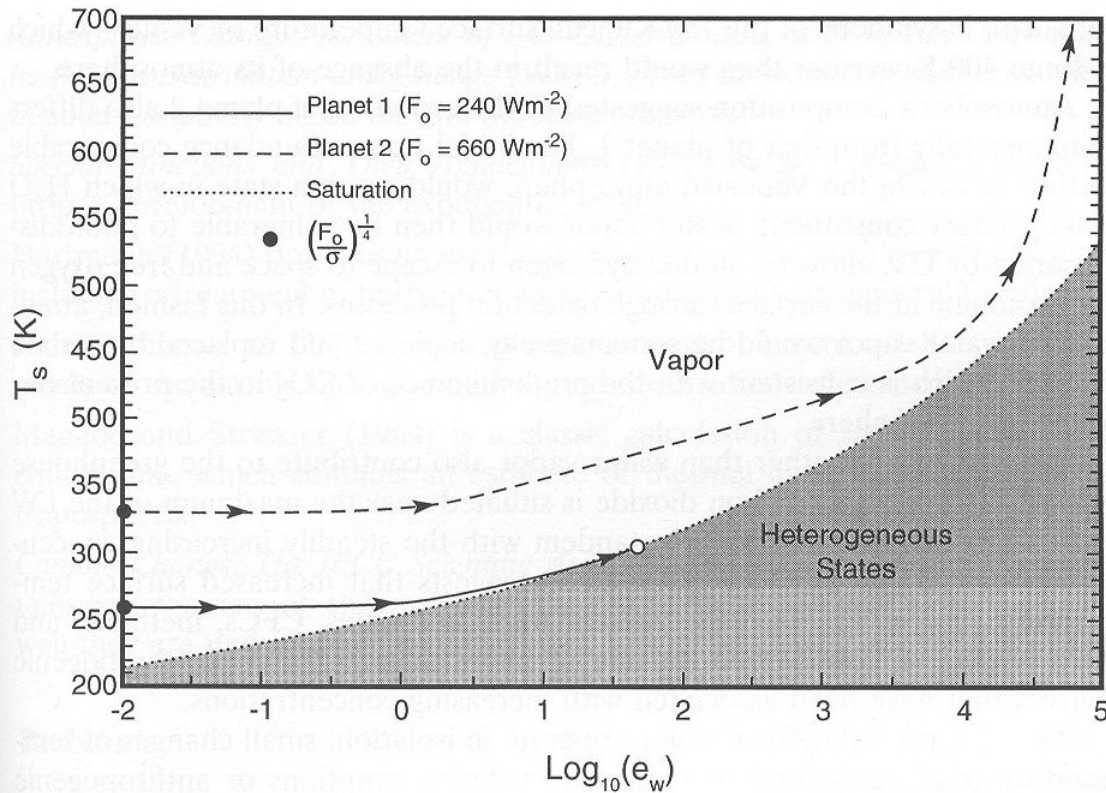
Figure 8.30 Surface temperature under radiative–convective equilibrium, as a function of saturation vapor pressure for water (which is symbolic of atmospheric optical depth to LW radiation τ_s). Radiative–convective equilibrium curves shown for conditions representative of Earth (planet 1) and Venus (planet 2), which characterize atmospheric evolution from an initial state when all water resided at the surface in condensed phase. Positive feedback between temperature and water vapor drives the state of a planet's atmosphere to greater temperature and humidity. For planet 1 (solid line), the radiative–equilibrium curve eventually encounters the saturation curve (dotted line), where positive feedback ceases. However, the radiative equilibrium curve for planet 2 (dashed line) never encounters the saturation curve, so it continues to evolve through positive feedback until all water has been incorporated into the atmosphere. The calculation of radiative–convective equilibrium presumes a constant albedo and relates optical depth to saturation vapor pressure as $\tau_s = 4[e_w(T)/e_w(288\ K)]$.

Because of the large greenhouse effect of water vapor, the more water vapor that you add to the atmosphere, the higher the surface temperature will be. This increases the saturation vapor pressure and allows more water vapor in the atmosphere (as the planet continues to outgas) and providing an even stronger greenhouse effect. For Earth, the increase in water vapor required to increase the surface temperature (through the greenhouse effect) by a certain amount is greater than the increase in saturation vapor pressure so it's stable at the point the two curves cross. (In other words, one can get vapor pressure greater than the saturation vapor pressure.) However, for a greater solar constant, the increase in water vapor required to increase the surface temperature (through the greenhouse effect) can be smaller than the increase in saturation vapor pressure. One then gets into a runaway greenhouse process. This is believed to have happened to Venus, resulting in the lose of its water.

A few words on cloud microphysics.

A cloud microphysicist will be extremely unhappy with the above description. We have assumed that condensation occurs whenever the vapor is supersaturated. However, there are many complications to the story. The saturation vapor pressure is for equilibrium with a plane surface. Growth of a cloud droplet however requires an increase in surface energy. This limits the growth of cloud droplets, and it requires a supersaturation of 300%-400% for cloud droplets to grow by homogeneous nucleation. In the atmosphere, this is alleviated by the presence of cloud condensation nuclei (CCN), which are wettable particles that allow the droplets to grow from a finite size so that only 0.1% supersaturation is needed. Some of the particles are soluble in water and the solution further lows the supersaturation needed. Sulfate particles, e.g., make good CCNs. With more CCNs, one expect to have smaller and more cloud droplets given the same amount of condensed water. Continental air tends to have more CCNs. An important natural production of sulfate aerosols is from the dimethylsulphide (DMS) emission from phytoplankton. It was hypothesized that such emission, by increasing the number of CCN, may modify Earth's albedo, and thus play an important role in the radiation budget. Anthropogenic production of sulfate aerosols is also significant and is believed to have increased. The cloud droplets can grow efficiently by diffusion when they are small (<20micron), but as condensational growth is slow when the droplets become large, in which case, collisional growth becomes dominant.

The issue of condensation nuclei becomes even more severe for ice and homogeneous nucleation is not favored at temperatures above -36C, so cloud droplets can often be supercooled (stay liquid below 0C). This is why airplanes going through such a region encounter icing problems. Once formed, they can grow through condensation of vapor, called deposition. Because the saturation vapor pressure with respect to ice is lower than that with respect to water, ice particles grow faster than droplets. Particles can also grow through collision with supercooled droplets, or riming. Last, ice particles can grow through coagulation, which is favored at temperatures above -5C.

A cloud microphysicist will most likely remain very unhappy with the above sketchy discussion. The book by R. R. Rogers, A short course in cloud physics, has more information.