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Atmospheric Thermodynamics and Vertical Stability

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Abstract: ***There are three main types of processes in which be interested: 1) Cooling or warming at constant pressure. 2) Expansion or compression. 3) Mixing. Vertical motion of air masses in the atmosphere are accompanied by changes in pressure. So there are adiabatic expansion with and without condensation condensation. when a region of the atmosphere is unstable, convection may ensue, with vertical ascents and compensating descents of air i.e vertical mixing as well as horizontal mixing so far, air can be forced to rise even in a stable atmosphere, so can be vertical stability, conditional and latent instability, potential or convective instability if the opposite happen, i.e that the humidity increases with height, the layer will become stable with lifting. it is the case of potential or convective stability.***

Keywords: Thermodynamics – Atmosphere – Adiabatic expansion – Mixing – stability

1. Atmospheric Systems

From the thermodynamically point of view, a system is any body of a given mass and composition under study. All the rest of the bodies with which the system may eventually interact are called its surroundings.

In the atmosphere we deal basically with two types of systems: those composed of air, with a humidity that can go from 0 to saturation, and clouds, composed of saturated air and either water droplets or ice crystals. (There are also mixed clouds, with both drops and crystals, but we shall not consider them.)

Unsaturated air can be considered as a mixture of :

a) Dry air : a mixture of gases in constant composition, with mean molecular weight 28.964.
b) Water vapour: its content can be expressed, for instance, b the vapour pressure or, aswe shall do here, its molar N_v, defined as

$$N_v = \frac{number\ \ of\ moles\ of\ water\ \ vapour}{total\ number\ \ of\ moles} \simeq \frac{number\ \ of\ moles\ of\ water\ \ vapour}{number\ \ of\ moles\ of\ dry\ air} \qquad (1)$$

Where the last approximate equality results from the fact that the number of moles of water vapour is always much smaller than that of dry air.

Bout dry air and water vapour in atmospheric conditions, and also their mixture as moist air, can be considered with good approximation as ideal gases. The presence of water vapour alters slightly the average molecular weight of the mixture, but with enough approximation for our purposes, we can assume that the average molecular weight of the mixture is always M=28.9* and that the equation of state the corresponding ideal das law:

$$pV = \frac{mRT}{M} = nRT \qquad (2)$$

Where p= pressure , V= volume, m=mass, R=8.3J/mol.k=universal gas constant, M=28.9=average molecular weight and n= total number of moles≃ number of moles of dry air.

The presence of water vapour will also change slightly other parameters, such as the heat capacity of the air, but we shall ignore these corrections.

When we deal with clouds, the presence of water droplets or ice crystals also alters some parameters. However, the content of condensed water is only of the order of one or a few grams per kilogram of air, and therefore its influence can be neglected, except for one important fact: latent heat is absorbed or a absorbed when water changes its physical state, i.e. during evaporation, condensation, freezing or melting. This important effect must be taken into account in the formulas.

2. First Principle of Thermodynamics, as Applied to Air and Clouds

The general formulation of the First Principle of Thermodynamics is:

$$dU = \delta Q + \delta A$$

Here U = internal energy of the system, Q= heat absorbed by the system, A= work performed upon the system by external forces. U is a state function, i.e. its value is determined (except for an arbitrary additive constant) by the state of the system, therefore the changedU, or ΔU for a finite process, depends only on the initial and final states, and not on the path followed by the process. A and Q are not state functions, so that δA and δQ, or A and Q for finite processes, depend on the path followed by the process.

In the atmosphere, the only work that can be received by the system is expansion work; therefore:

$$\delta A = -p\ dU$$

Introducing (4) into (3), the firs principle becomes

$$dU = \delta Q - p\ dU$$

The first principle can also be expressed in terms of another state function, called the enthalpy H and defined by

$$H = U + pV$$

If we differentiate (6) and introduce (5), we obtain

$$dH = \delta Q + Vdp$$

Which is the alternative expression for the first principle.

We shall now apply equation (7) to the systems of our interest: first to clear air, and then to clouds.

Let us consider one mole of air. We define the molar heat capacity of air at constant pressure, C_p, as the heat necessary to increase the temperature of one mole of air by one degree at constant pressure. Now let us assume that we heat one mole of air at constant pressure (dp =0), producing a temperature dT. Then (7) gives.

$$(\mathrm{dH})_{\mathrm{p=const}} = (\delta Q)_{\mathrm{p=const}} = \mathrm{C_p\ dT}$$

But it can be shown in thermodynamics that the enthalpy (as also the internal energy) of ideal gases is a function of only the temperature. Then the value of dH does not depend on whether the pressure is kept constant or not. Therefore we can write, for any (infinitesimal) process:

$$dH = C_p \mathrm{dT}$$

Introducing (9) into (7) we obtain

$$C_p\ dT = \delta Q + Vdp$$

This is the expression that we need to study the atmospheric processes of air.

Now let us consider one mole of saturated air containing water droplets (a cloud of water droplets; the case of ice crystals would be similar), and let us repeat the same arguments as before. When we consider a process at constant pressure, the heat received by air includes now two terms: one due to the increase in temperature, as before, and another one due to the latent heat absorbed by water if any evaporation takes place. We define the molar latent heat of vaporization, L_v as the heat necessary to evaporate one mole of water. Then the new term that we must consider will be of the form $L_v\ dN_v$ where dN_v the increase is in molar fraction of water vapour in the air, i.e. approximately the number of moles of water evaporated in each mole of air. Thus. Instead of Equation (8), we have now:

$$(\mathrm{dH})_{\mathrm{p=const}} = (\delta Q)_{\mathrm{p=const}} = \mathrm{C_p\ dT} + L_v\ dN_v$$

We repeat now the previous argument about the enthalpy, to show that we can write, for any process:

$$\mathrm{dH} = \mathrm{C_p\ dT} + L_v\ dN_v$$

and, introducing this expression into (7) :

$$\mathrm{C_p\ dT} + L_v\ dN_v = \delta Q + Vdp$$

This is the formula that we need in order to study atmospheric processes in a water cloud (a similar expression would be valid for an ice cloud, with appropriate substitutions for $\mathrm{C_p}$ and L_v). dN_v can be expressed in terms of T and p as independent variables. According to the definition of N_v and to the gas laws,

$$N_v = \frac{e}{p}$$

Where e is the partial vapour pressure of water an p the total pressure. Differentiating:

$$dN_v = \frac{de}{p} - \frac{e}{p^2}dp$$

This differential will only appear the cases when the air is saturated with water vapour, so that dN_v results from the evaporation or condensation of water. In that case, de is linked to the variation of temperatures by the Clausius-Clapeyron equation (cf. Ch.II,§ 4):

$$dN_v = \frac{L_v e}{RT^2 p}dT - \frac{e}{p^2}dp$$

Which can be subtituted into (13) in order to have the formal expressed in terms of the variations of T and p.

Main Processes in the Atmosphere

There are three main types of processes in which we shall be interested:
1) Cooling or warming at constant pressure
2) Expansion or compression
3) Mixing

Cooling at constant pressure will refer to a process at a constant level (e.g.at ground), where the pressure variations are small egnored in this context. Expansion and compression are linked with vertical motins in the atmosphere, as know that pressure varies with height. We shall consider them in turn.

Cooling

Dew and frost form as a result of condensation or sublimation of water vapour on solid surfaces on the ground, which cool during the night, by radiation, to temperatures below those corresponding to saturation. The process can be understood on the diagram of

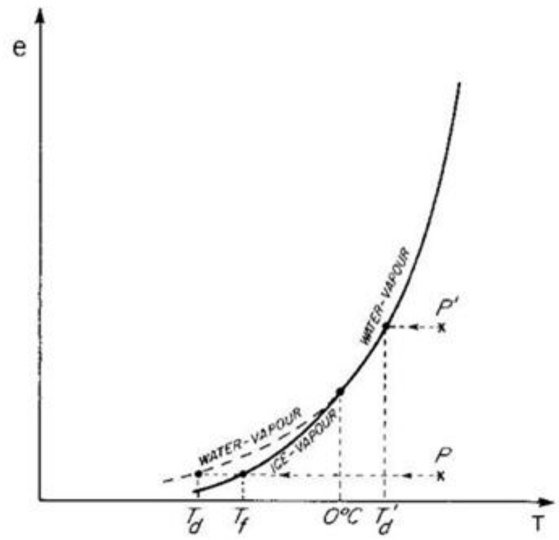


Fig during adiabatic ascent, for dry air, is 9.8 K\km, and only slightly less (up to a few %) for moist, unsaturated air, For saturated ascent, this rate varies with T and p, but always less than the previous value, and can be as low as 3 K\km. We shall call β=-dT\dz the temperature lapse rate. For dry air.

Horizontal mixing

If two different adjacent air masses mix, the process occurs essentially at constant pressure. if no condensation take place , the result is the air whit temperature and a molar ratio which are the weighted average (weighted over the number of moles) of the values for the two air masses

It may happen that two unsaturated air masses produce condensation when mixing. this occur if one of masses is hot and humid , while the other is cold , and it is due to the curvature of the vapour pressure curve . figure 5 illustrates the process . it can be shown that the representative point of mixture line joining the two mixing air masses . the figure shows that this may lead to super saturation ,with the subsequent condensation . this can produce a fog , called mixing fog jet aircraft trails are a case of mixing fog , where the hot and humid gasses from the motor exhausts mix with the outer cold (and therefore dry)air .

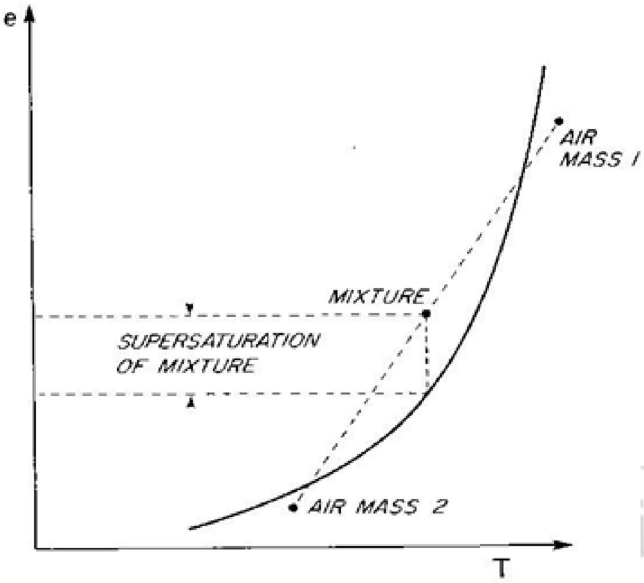


Figure 5: *Mixing fog. T*= temperature; *e*= water vapour pressure. Due to the curvature of the saturation vapour pressure curve, mixing can lead to supersaturation

Vertical mixing

When a region of the atmosphere is unstable , convection may ensue , with vertical ascent and compensating descents of air . this may lead to a thorough mixing of whole layer , particularly in areas with strong insolation.

Let us assume that such a thorough mixing has taking place through a layer of a given thickness, say between p1 , and p2 . we want to find out what will be the final vertical distribution of temperature . we start by considering an infinitesimal layer of thickness dz at the pressure p , and we imagine that we bring it to the level of 1000 mb , adiabatically . its temperature will then become equal to its potential temperature θ.,by definition of this parameter (cf . equation (24)) . Now let us imagine that we bring the whole finite layer to the 1000 mb level , infinitesimal layer, and that we mix it thoroughly at that level . According to §7 , the result will be that all this air will acquire the temperature $\bar{\theta}$ is equal to the mass- averaged potential temperature . Now we imagine that we redistribute all the air , infinitesimal layer by infinitesimal layer , to their original situation in the interval of pressure $p_1$to p_2* , in so doing , every parcel of air conserves the value of potential temperature , now equal to $\bar{\theta}$ for all the air. thus the final result is a finite layer between p_1 and p_2) with constant potential temperature $\bar{\theta}$ throughout the whole thickness . this corresponds to a distribution of T according to (24) i,e . the parcel of air with potential temperature $\bar{\theta}$ undergoes if raised through that height , and we saw that this is given by a lapse rate $\beta_d = 9.8\ k/km.$. This is expressed by saying that the result of vertical mixing without condensation is a dry adiabatic layer.

Notice that in this result we are considering the instantaneous final stratification of the layer, rather than the process undergone by rising air. but both will be represented by the final stratification of the layer, rather than the process undergone by rising air . but both will be represented by the same t =f (z)curve , and this will be the lower part of the curve figure 4 , labeled as unsaturated adiabatic .(unsaturated adiabatic and dry adiabatic are to be taken as equivalent names for the same curve , since we neglect the influence of humidity when the air is under saturation) if the mixing is accompanied by condensation , starting at a certain level , the result will be describe , statically , by the total curve in figure 4 that represent the ascent process

Vertical stability. conditional and latent instability

If we have a mass of homogeneous fluid , at uniform T , in a container of laboratory dimensions , and we heat a portion of the lower layer , we create a vertical instability . the heated fluid becomes less dense than its surrounding , thereby experiencing an upwards force (buoyancy) , and starts rising . since in the atmosphere , or a portion there of , is in equilibrium with respect to vertical displacements .

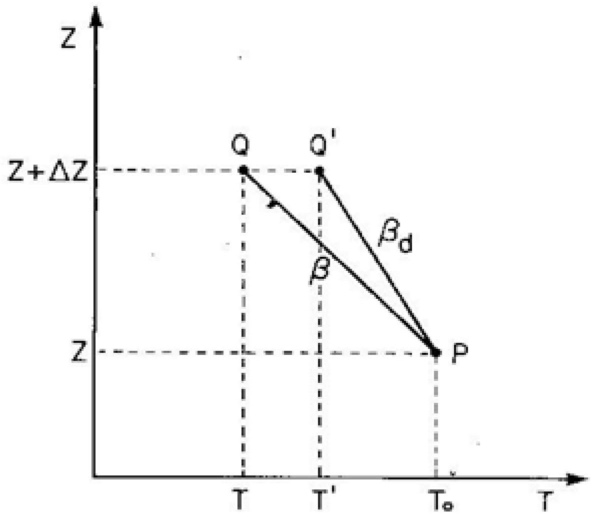


Fig. 6. *Vertical instability.* If the geometric temperature lapse rate β at height z is larger than β_d, an unsaturated parcel rising Δz from P will have at Q' a temperature $T' > T$, where T is the temperature of the surrounding air (Q); it will tend to continue rising.

Now let us consider a level in a non- saturated atmosphere at which β is the geometric lapse rate. By this we mean the vertical temperature stratification not to be confused descends. With the process lapse rate that gives the change of T with z when a parcel rises or descends Let us assume that Let $> \beta_d$. If we image a virtual displacement, we shall have the situation of Figure 6 here P is the representative point in a graph z, T of the atmosphere at level z. At a level $z + \Delta z$ (Δz can be considered here as infinitesimal), the representative point is Q, and the temperature is $T_0 - \beta\Delta z$. A parcel displaced in Δz from P,on the other hand, will follow an adiabat β_d , and at $z + \Delta z$ will have a temperature $T_0 - \beta_d\Delta z > T_0 - \beta\Delta z$. Because the pressure will be equal for parcel and environment, the density $\acute{p}$ of the parcel will be smaller than the density p of the environment, and the parcel will experience positive buoyancy, i.e an upward force. Because this buoyancy is positive, after the virtual displacement the parcel will tend to continue rising. It is clearly an unstable situation. If $> \beta_d$, the case is that of figure 7. Then the buoyancy is negative, so that the displaced parcel will tend to back to the original level z. It is a stable case.

Notice that with $\Delta z < 0$ the same conclusions would be reached. If the atmosphere is saturated, the same type of argument can be made, but using β_s (rather than β_d) for the process lapse rate of the parcel.

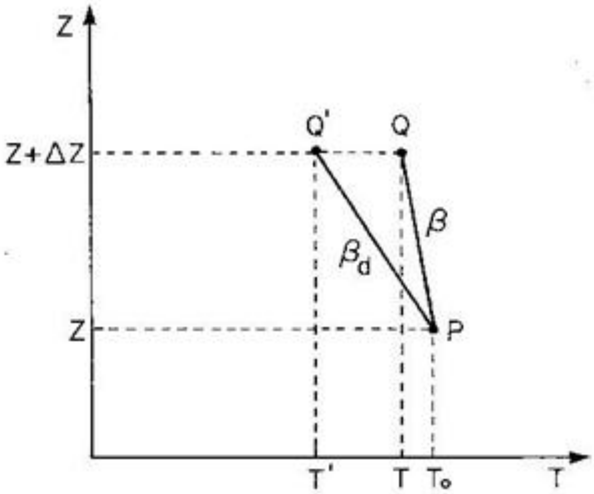


Fig 7. *Vertical stability*. If the geometric temperature lapse rate β at height z is smaller than β_d, an unsaturated parcel rising Δz from P will have at Q' a temperature $T' < T$, where T is the temperature of the surrounding air (Q); it will tend to sink back to z.

Thus we have the following conditions:

$$\left.\begin{array}{r} \text{Unsaturated} \quad \beta \underset{s}{\overset{u}{\lesseqgtr}} \beta_d \\ \\ \text{Saturated} \quad \beta \underset{s}{\overset{u}{\lesseqgtr}} \beta_s \end{array}\right\}$$

Where the inequality signs correspond to the unstable case (u) and the lower one to stability (s) . If the equality sign holds, the equilibrium is indifferent: the displaced parcel remains where it is left. All these cases are summarized schematically in Figure 8. If the lapse rate of the atmosphere at P (i.e. at level z and pressurep) lies to the left of βd, the atmosphere is absolutely unstable (at that level) ; if it lies to the right of β_s, it is absolutely stable . If it lies between β_d, and β_s it is said to be conditionally unstable, meaning that it is stable if not saturated , but unstable if saturated .

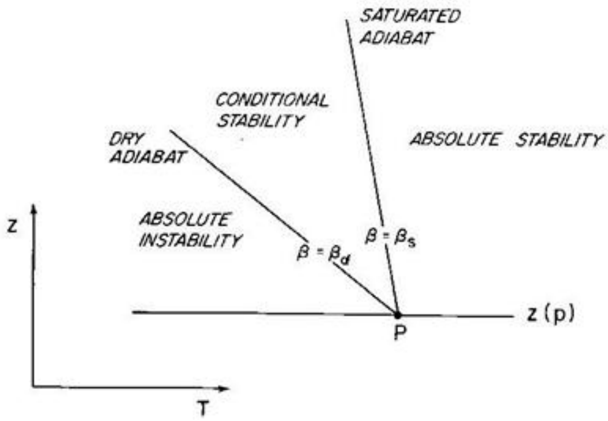
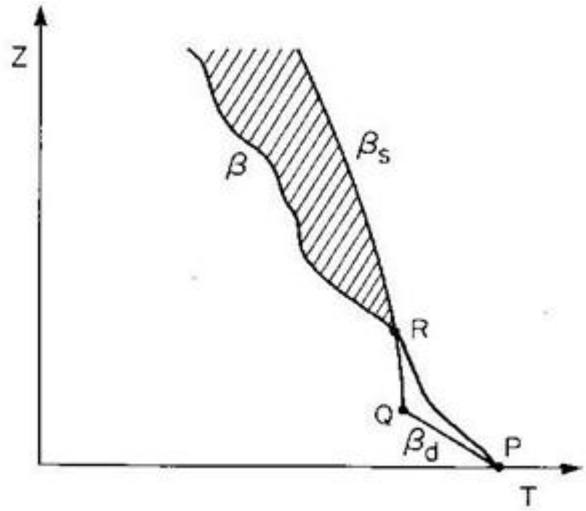


Fig. 8 Condition of vertical stability. The diagram summarizes the stability conditions for both unsaturated and saturated air (see text). T = temperature; z = height.

So far, we have considered only infinitesimal displacements. Air can be forced to rise even in a stable atmosphere (for instance, if wind blows against a mountain slope) and it may eventually follow p process curve such as is indicated in Figure 9. Here the curve β indicates the geometric stratification of the atmosphere. The parcel considered follows β_d until saturation at Q (lifting condensation level) and then follows the saturated adiabat β_s . At R (level of free convection), the representative point of the parcel crosses to the right of the curve β, and the buoyancy becomes positive. From R on, the parcel is accelerated upwards. This region, indicated in the figure by the shaded area, is the region of latent instability. In this type of instability, the parcel may actually become mixed to a large extent with the environment, so that the conclusions can be qualitative.

The study of the vertical stability is very important in weather forecasting instability.

Fig 9. Latent instability. T =temperature; z=height. β indicates the temperature stratification of the atmosphere. A parcel rising from *P* will follow an unsaturated adiabat (lapse rate β_d) to *Q* (lifting condensation level) and then a saturated adiabat (lapse rate β_s). The parcel must be forced to rise up to R (level of free convection); from there on, it is accelerated by a positive buoyancy. The shaded area indicates the a saturated adiabat (lapse rate β_s). the parcel must be forced to rise up to R (level of free convection); from there on, it is accelerated by a positive buoyancy. Te shaded area indicates the latent instability.



Is associated with weather perturbations Thus, if conditions are such strong vertical matins occur, 'convective' clouds are formed: first cumuli, and when these develop further, cumulonimbi, i.e. huge clouds with thickness of the order of 10 km, whose tops become glaciated when their temperatures fall below – 40˚C. Showers and hail develop in these cloud, and the strong vertical air currents in them can reach velocities of several tens of meters per second.

Potential or Convective Instability:

In certain cases (for instance, at approach of a cold front), whole layers of atmosphere can experience a lifting. It is important to consider how this will affect the stability conditions.

Let us consider first the case when saturation does not occur. AB is a layer of geometric lapse rate β (Figure 10). As there is no saturation, lifting will bring A to $\acute{A}$ and B to $\acute{B}$ along process dry adiabats β_d. The thickness of $\acute{A}\acute{B}$ (i.e. of the layer after lifting)will lager than that of AB, because the pressure decreases with height and therefore the layer expands. the geometric rate has become $\acute{\beta}$, and it is clear that, while remaining $< \beta_d$, it will be $\acute{\beta} > \beta$. The layer has become less stable. If at the initial level air is converging laterally towards the zone of lifting, $\Delta\acute{z}$ will become still larger as compared with Δz , simply because of mass conservation (the layer starches vertically to allow space for the converging air); the result will be a still larger in stabilization.

Therefore: both ascent of a layer and horizontal convergence decrease the stability. Reciprocally, descent (or subsidence) and divergence, increase the stability.

Now let us consider layer lifting when saturation occurs. Two cases can happen. In the first one, humidity decrease with height the layer, so that the lower part of the layer becomes saturated first. If the layer is AB (Figure 11, both A and will follow fist

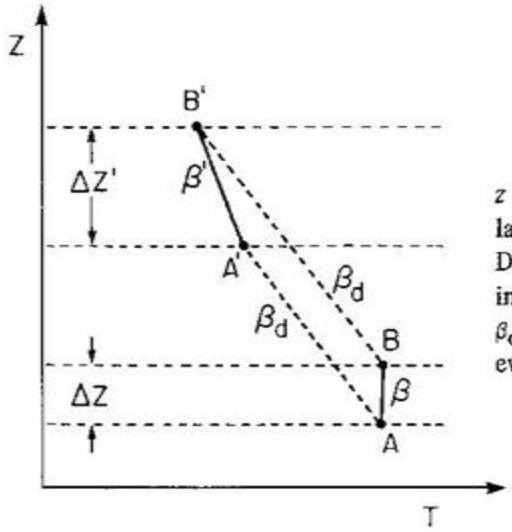
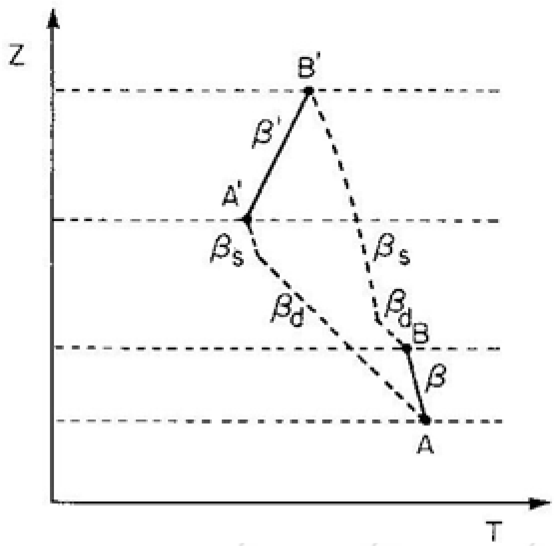


Fig. 10. *Ascent of a layer.* T = temperature; z = height. AB and $A'B'$ indicate the (geometric) lapse rate within a layer before and after rising. Due to the decrease in pressure, the thickness increases ($\Delta z' > \Delta z$) and β tends towards the value β_d. Horizontal convergence will make this effect even more pronounced.

Fig. IV-11. *Potential instability.* T = temperature; z = height. Humidity decreases with height within the layer AB. If the layer rises, its lower portions become saturated earlier and this leads to an unstable lapse rate β' as indicated by $A'B'$.



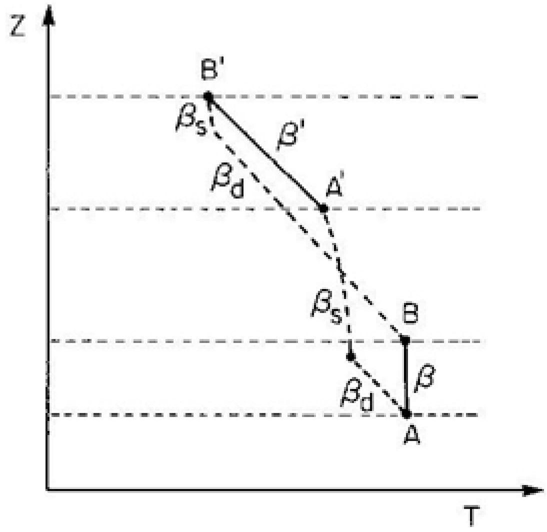


Fig. 12. *Potential stability.* T = temperature; z = height. Humidity increases with height within the layer AB. If the layer rises, its upper portions become saturated earlier and this leads to more stable lapse rates β', as indicated by $A'B'$.

a dry adiabat β_d, and then a saturated adiabat β_s, to the final points $\acute{A}\acute{B}$, but a parcel starting at A will saturate earlier than a parcel starting at B. It is clear that the layer not only loses stability, but can become absolutely unstable. The is called potential or convective instability. If the opposite happens, i.e. that the humidity increases with height. The layer will become more stable with lifting. It is the case of potential or convective stability. This is exemplified in Figure 12.

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