# Atmospheric stability

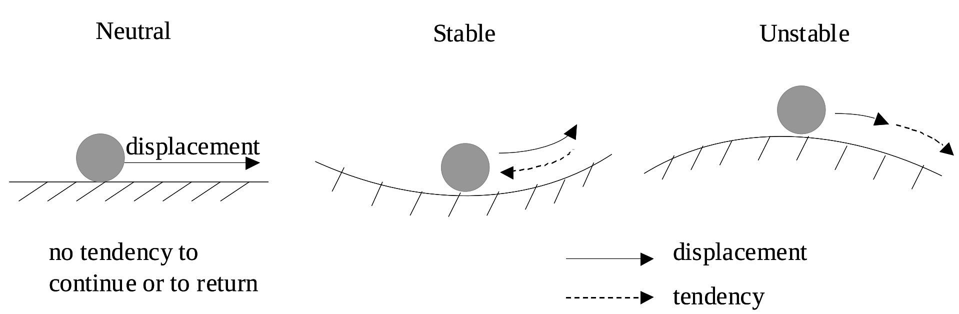
The stability or instability of the atmosphere (or a layer thereof) is the state of the atmosphere with respect to the reaction of a volume or “parcel” of air to a vertical displacement. The stability of the atmosphere determines the likelihood of convective activity, cloud type (stratus or cumulus), likelihood of atmospheric turbulence, the extent of mixing (pollutants etc.).

Stability is classified as:

- stable
- neutral
- unstable, or
- conditionally unstable
  (This latter classification depending on whether the air is saturated with moisture or not)

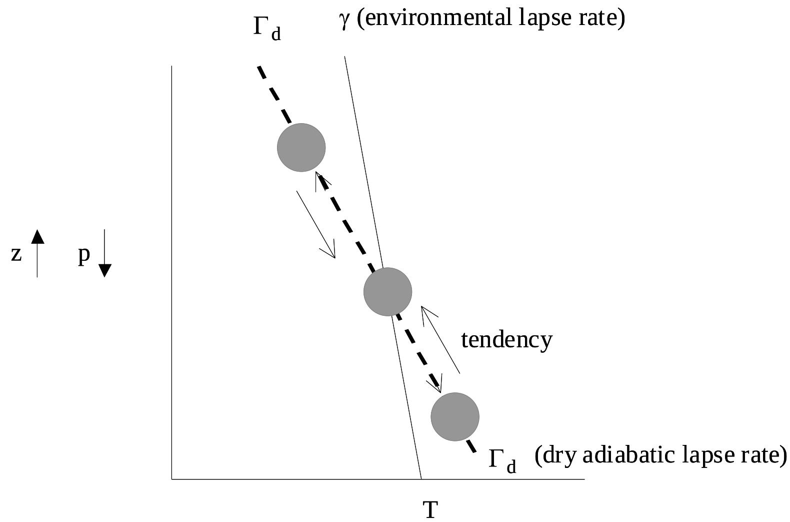
To determine stability classification we must examine the tendency of the atmosphere to resist or enhance an initial displacement.

For example, consider a ball bearing on three different surfaces:




The same thing happens in the atmosphere when we displace a “parcel” or bubble of air in the vertical direction. We can use a thermodynamic diagram to examine this.

When $\gamma < \Gamma_d$

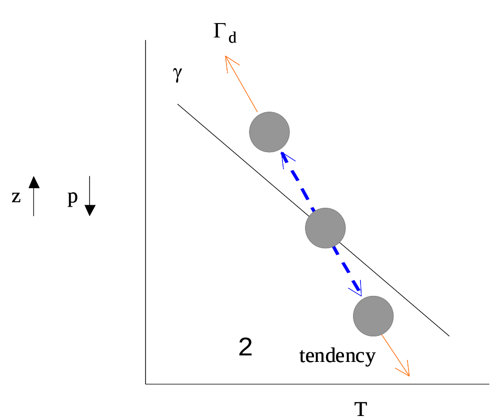



Consider a parcel at a given pressure level. Impose a displacement in the vertical direction (upward or downward).

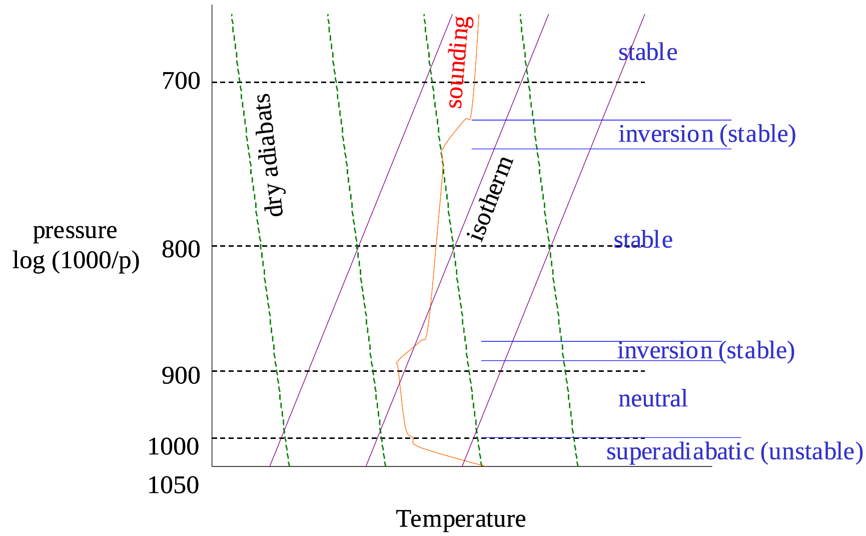
***A displaced parcel will change temperature (if adiabatic) at a rate governed by the Poisson equation ($1^{st}$ law of thermodynamics) or at the adiabatic lapse rate.***

The example above is of an atmosphere that is stable (at least to a dry process with no condensation) because a “parcel” of air displaced upwards (downwards) will cool (warm) at the dry adiabatic lapse rate, become cooler (warmer) than its surroundings and therefore denser (lighter) and will tend to return to its original position.

When $\gamma > \Gamma_d$, called a superadiabatic lapse rate, the displaced parcel wants to keep moving in the direction of the displacement. This is an unstable layer of atmosphere.



Layers of atmosphere can be classified according to their stability or instability.




**Note**: some of these layers may be “conditionally stable” depending on whether the air is saturated with moisture.

## Moist adiabatic processes (saturation adiabats)

So far we have discussed only dry processes in which neither condensation nor evaporation of existing water droplets in a cloud is occurring. But condensation releases large amounts of heat to the atmosphere and evaporation consumes large amounts of heat. The latent heat of evaporation (or condensation), $L=2.5\ 10^6$ J/kg. Approximately six times as much heat is needed to evaporate 1 kg of water as is needed to raise its temperature from 0 $^oC$ to 100 $^oC$, the boiling point.

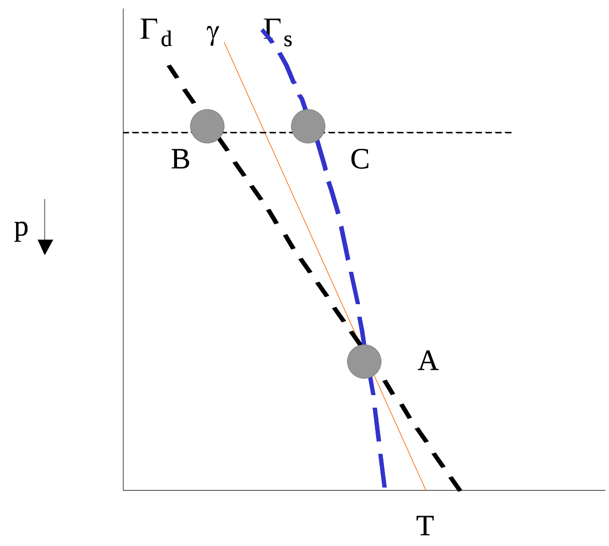
If, during a lifting process, adiabatic cooling is sufficient to bring the air to saturation and, if lifting continues, the air will continue to cool but the release of latent heat will offset the temperature decrease to some extent. Thus, during saturated ascent, the rate of decrease of temperature of a parcel of air will be less than during a dry adiabatic process. The difference in lapse rates between dry and saturated processes will depend on the amount of water vapor available for condensation. Thus, near the earth's surface (high pressure and air density) in warm climates, the saturation adiabatic lapse rate $\Gamma_s$ will be considerably less than $\Gamma_d$ (say, 5 $^oC$/km compared with 9.8 $^oC$/km). On the other hand, high in the atmosphere and at low temperatures where there is little moisture content, the saturation adiabatic lapse rate will approach $\Gamma_d$.

Thermodynamic diagrams such as the skew T-log p diagram include a series of lines representing saturation adiabatic processes (saturation adiabats).

- Check the Skew T-log p diagrams for $\Gamma_s$ lines

## Conditional instability

Since $\Gamma_s < \Gamma_d$, our analysis of the stability of a layer of atmosphere will depend on whether or not the air is saturated (whether or not cloud in present) because, under certain conditions, opposite conclusions would be drawn with regard to stability or instability. Suppose the observed lapse rate is intermediate between $\Gamma_d$ and $\Gamma_s$.



Here, the sounding is shown by the solid line $\gamma$.

If the layer is dry and no condensation occurs, as before, lifting a parcel from point A will result in dry adiabatic cooling to point B. At this point, the air is colder than its surroundings (given by the sounding temperature at the same pressure level), denser, and would want to sink back to its starting point. Our conclusion would be that the layer of atmosphere is stable.

If the layer is saturated and lifting causes condensation, the parcel will follow the moist adiabatic to point C and will be warmer, and less dense, than the surrounding air. It will want to continue in free buoyant ascent. Thu, is moist, the layer is unstable.

This situation, in which the nature of the stability or instability is dependent upon the status of the moisture in the atmosphere (unsaturated or saturated), is called conditional instability.

In general the troposphere is rather stable to dry processes (except in the lower when the sun heats the surface) and would be generally inactive except for the instability caused by the release of latent heat in clouds.

In general we can classify a layer of atmosphere according to the relationship between the observed lapse rate $\gamma$ and the two process lapse rate $\Gamma_d$ and $\Gamma_s$. Thus:

$\gamma > \Gamma_d$ unstable (absolutely)

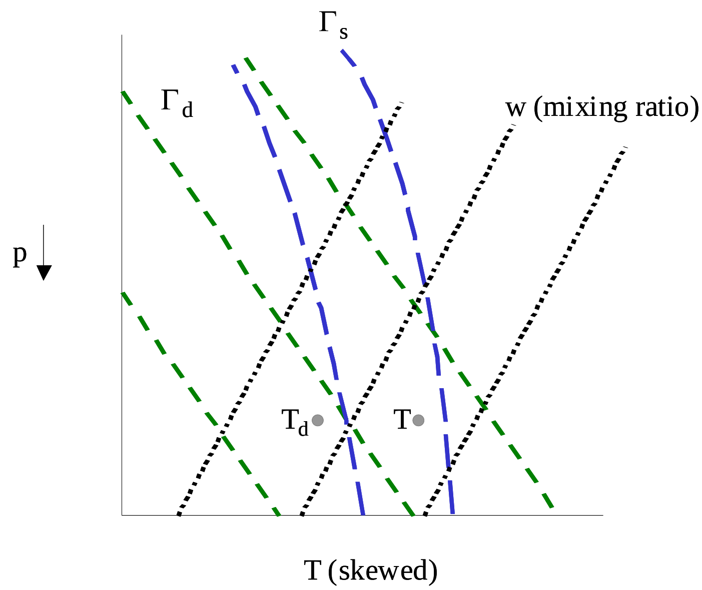
$\Gamma_d > \gamma > \Gamma_s$ conditionally unstable

$\gamma < \Gamma_s$ stable (absolutely)

If $\gamma < 0$, i.e., temperature increases with height, this is called an inversion and is an extreme vase of stability. Pollution episodes are commonly associated with inversions, or at least, inversions aloft which trap pollutants in the lowest layers of the atmosphere.

## The lifting condensation level (LCL)

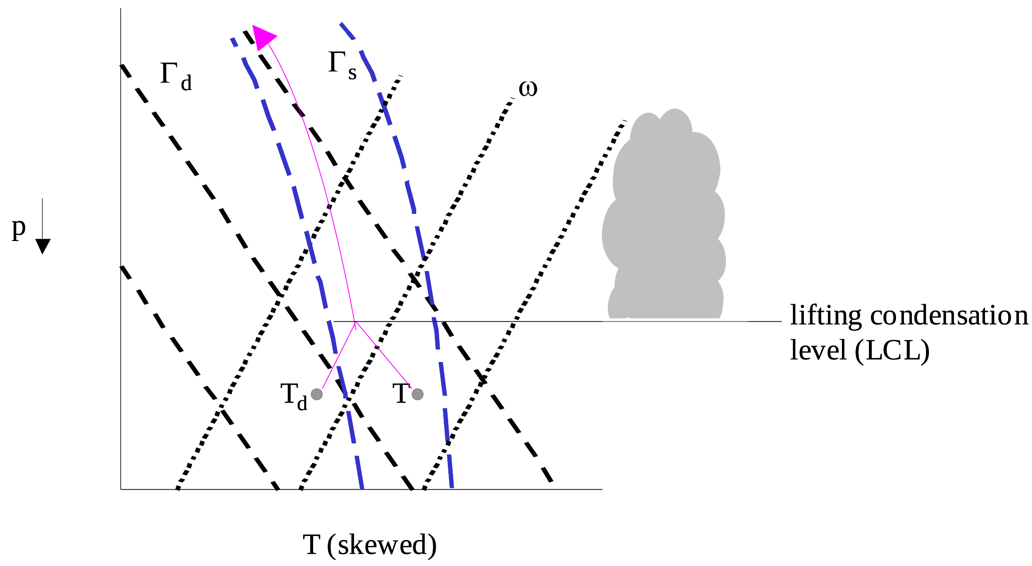
A parcel of air forced to lift in the atmosphere will ultimately cool to the saturation point. During the initial unsaturated part of the ascent no moisture is condensing out and the mixing ratio (w) of the air remains constant (the amount of vapor in grams remains in constant proportion to the amount of dry air in kilograms). On the other hand, other moisture variables such as dew points $T_d$, relative humidity, and vapor pressure all change. For this reason, thermodynamics diagrams include a fifth set of lines representing constant mixing ratio.




Knowing T and p (pint T on the diagram) the mixing ratio lines can be interpolated to give the saturation mixing ratio - the amount of vapor the air can hold in g/kg at saturation.

Knowing $T_d$ and p (point $T_d$ on the diagram) the mixing ratio lines can be interpolated to give at the actual mixing ratio – the amount of vapor actually existing in the air at that pressure level.

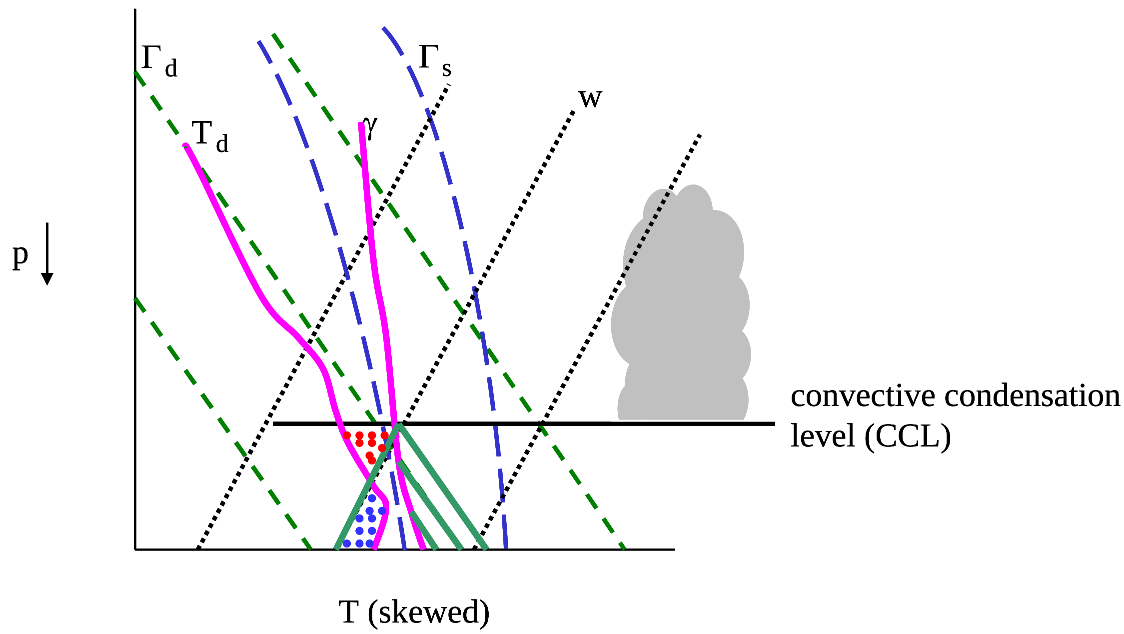
Because mixing ratio remains constant, we can observe the change in dew point during a lifting process, the increase in relative humidity, and the approach to saturation. On ascent T changes at the rate $\Gamma_d$ and $T_d$ changes at a rate determined by w = constant. This process scan be examined graphically:




The lifting condensation level is the level at which T reaches $T_d$ (which itself has decreases but at a lesser rate). The air is now saturated and continued lifting causes condensation in the form of a cloud. Above the LCL, a parcel of air will follow the saturation adiabat.

## The convective condensation level (CCL)

The convective condensation level is obtained when the surface is heated (e.g. solar heating during the day) and a convectively mixed layer (of constant $\theta$ and w) is created until saturation is achieved at the top. This assumes sufficient moisture in the air. In California summers, the air is too dry and the Pacific high pressure makes the air too stable for sufficient development of a deep cloud convective layer. Again, it is best to represent this process graphically:




The solid line (labeled $\gamma$) is the original sounding, say, at sunrise. Solar heating progressively heats the surface and warms the lowest layer of air. The convective mixing generates a layer of constant potential temperature and constant mixing ratio until, eventually, the air is saturated at the top.

If the layer above the CCL is conditionally unstable, the cumulus cloud could develop substantially.

## Stability indices

As a simple guide to the likelihood of convective activity, a number of indices have been developed to indicate the general level of instability in the atmosphere. The index used most commonly is the Lifted Index (LI).

To calculate the Lifted Index:

1) Determine the mean mixing ratio and potential temperature in the lowest 1 km of the atmosphere.
2) Find the LCL for this mean w and $\theta$.
3) Follow a saturation adiabat from this level to 500 mb and find $T'_{500}$ of the lifted parcel.

4) Subtract $T'_{500}$ from the observed 500 mb temperature $T_{500}$ to obtain

$$LI = T_{500} - T'_{500}$$
$$(T_{sounding} - T_{parcel})_{500mb}$$

As a rule-of-thumb:

| | |
|---|---|
| • $LI > 2$ | no activity expected |
| • $0 < LI < 2$ | RW probable, isolated T possible |
| • $-2 < LI < 0$ | T probable |
| • $-4 < LI < -2$ | severe T possible |
| • $LI < -4$ | severe T probable, tornadoes possible |

However, there is some geographic variability. For example, in the mountainous west, orographic lifting can produce thunderstorms at larger(move positive) value of LI.

In addition, a triggering mechanism is needed to initiate convective activity (frontal lifting, surface heating, orographic lifting etc.), and there needs to be sufficient water vapor in the atmosphere.