

Principles of Convection I: Buoyancy and CAPE

Produced by The COMET® Program

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Summary

Buoyancy is the upward force that acts on a parcel of air in response to a density difference between that parcel and the surrounding air. This force causes the air parcel to accelerate vertically. Consequently, buoyancy processes are basic to the generation of convective updrafts and downdrafts. Several factors contribute to increase or decrease the buoyancy of air parcels. Increases in potential temperature and water vapor content increase buoyancy, while cloud water and precipitation act to decrease buoyancy.

For large scales of motion, buoyancy and the vertical pressure gradient force tend to balance each other. Such motions are in hydrostatic equilibrium.

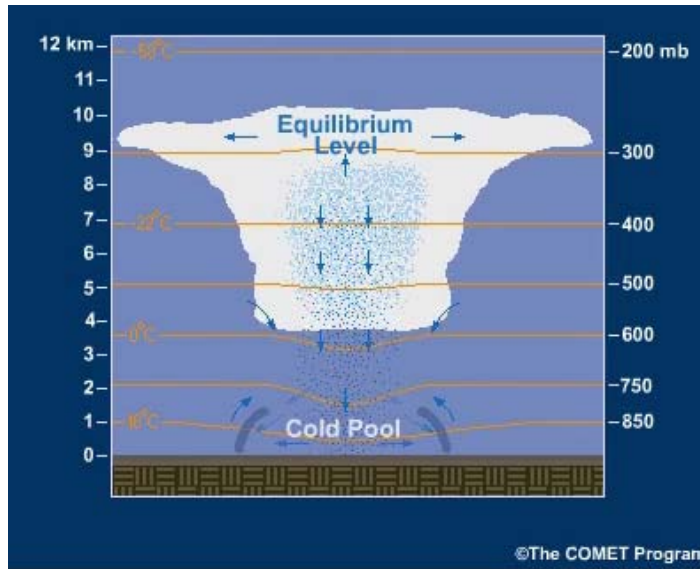
For small scales of motion, as in convection, buoyancy may significantly exceed the vertical pressure gradient force. The resulting vertical accelerations can be much larger than those typically associated with large-scale motions. Such motions are termed non-hydrostatic.

When vertical wind shear is weak, buoyancy is the dominant control on convective updrafts and downdrafts. Stronger shears lead to interactions between the updraft and shear, acting to enhance or suppress vertical acceleration.



All Stages

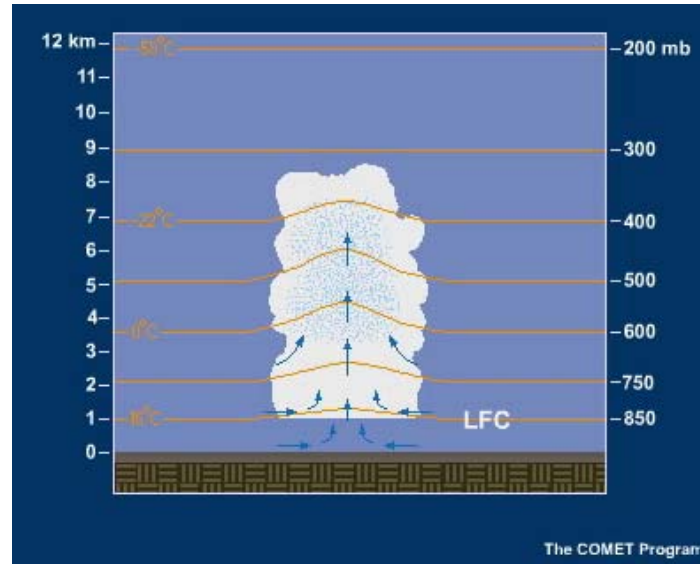
The basic processes associated with buoyancy within a convective storm are illustrated in this schematic of the life cycle of an ordinary cell evolving in an environment with no vertical wind shear.



Stage 1

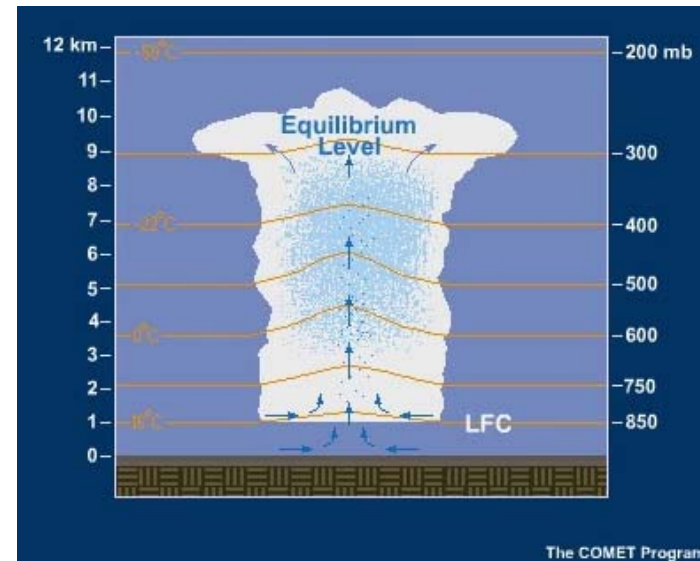
In an unstable environment, once air is lifted to its level of free convection (LFC), it will continue to rise as long as it is warmer than the air around it.

Level of Free Convection (LFC): The height at which a lifted parcel of air would first become warmer (less dense) than the surrounding air. Once the parcel reaches the LFC and becomes less dense than the surrounding air, it continues to rise freely until it becomes as cool (as dense) as the surrounding air. This upper level is the Equilibrium Level (EL).



Stage 2

Eventually, the parcel reaches an equilibrium level (EL) at which its temperature equals that of the surrounding environment. Since the parcel carries a certain amount of upward momentum, some additional ascent occurs beyond the EL. This additional rise forms the overshooting top above the anvil. Finally, because the parcels are now cooler than their surrounding environment, they sink back toward the EL. Subsequently, the parcels may oscillate vertically about the EL, dampening with time. As the process repeats, air parcels accumulate at this level and spread laterally, creating the cloud anvil.

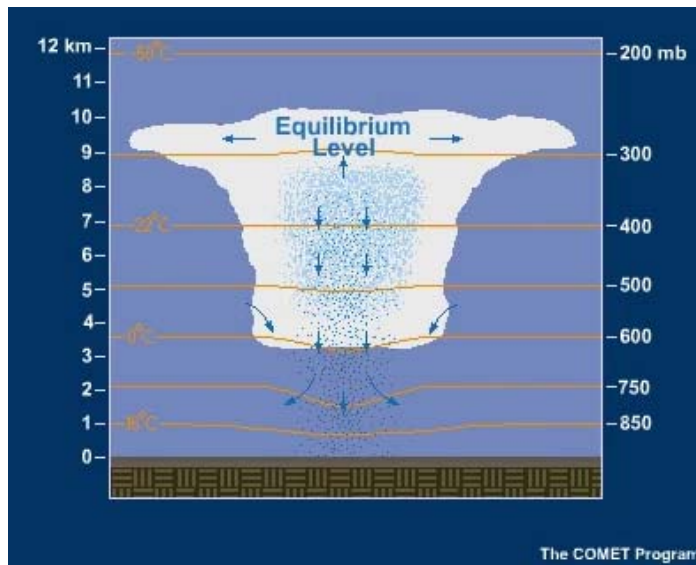


While all this is going on, moisture is condensing in the rising updraft air. The weight of the condensed moisture will eventually be too much for the updraft to hold aloft. Subsequently,

Stage 3

Because precipitation loading can subtract significantly from positive buoyancy, the initial effect of precipitation is to drag air downward, as shown in the animation, and initially, it is the most significant contributor to the strength of a downdraft.

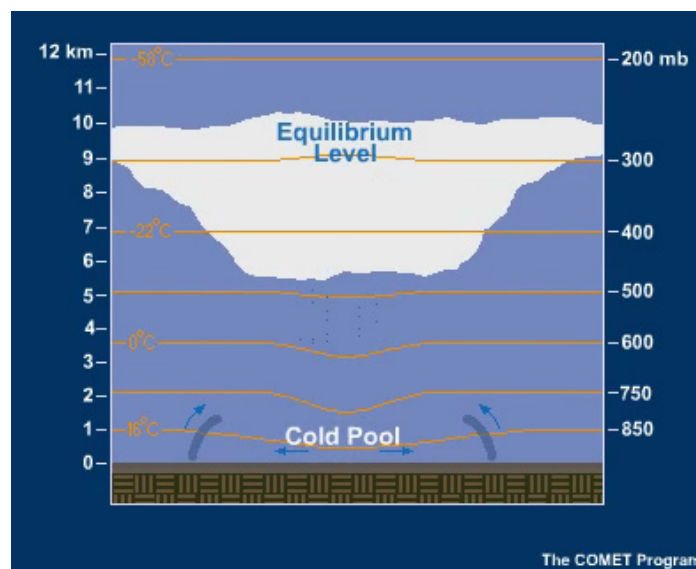
Other process contributing to the cooling of the downdraft include (1) entrainment of drier air at mid-levels and (2) evaporation of rain as it falls below the cloud base. Both of these processes make the downdraft colder than the surrounding air, further enhancing its downward acceleration.



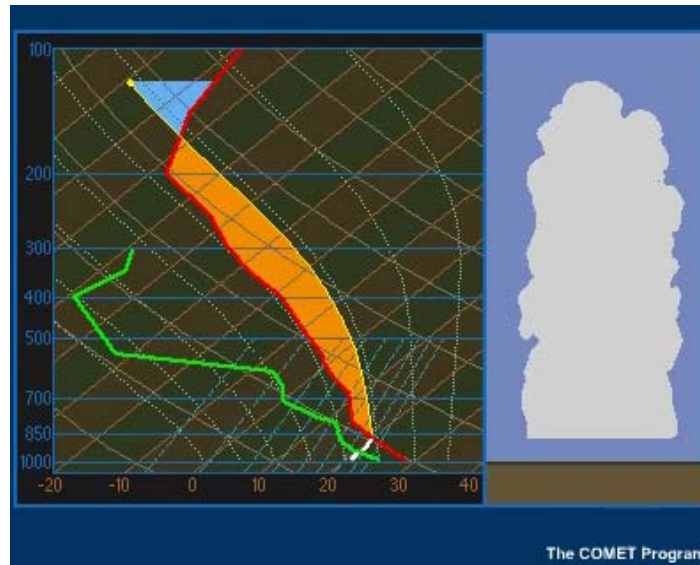
Studies have shown that much of the downdraft air originates from the environmental air entrained at mid-levels in a storm, generally between 3 and 5 km AGL.

Stage 4

When the downdraft reaches the surface it spreads out, forming the cold pool. The downdraft and spreading cold pool represent the final stages in the life cycle of the cell. At this point, the buoyancy has all become negative.



So what makes a parcel of air buoyant? We know that a parcel of air that is warmer than its environment will rise. As an unsaturated air parcel located near the surface rises, its temperature will follow a dry adiabat. Its dewpoint will follow a constant mixing ratio line. Where the dry adiabat crosses the constant mixing ratio line, the air becomes saturated and condensation commences. This is the lifted condensation level, or LCL. As the parcel continues to rise, it follows a moist adiabat. As a result, the rising air parcel becomes warmer and less dense than the surrounding air. So long as the ascent path along the moist adiabat is warmer than the environment, the air parcel will remain less dense than its surroundings and will continue to rise. This condition is referred to as instability. If there is no convective inhibition blocking the ascent of the rising air parcel, as shown in this example, the lifted condensation level becomes the level of free convection, or LFC. In the presence of convective inhibition, the level of free convection may be higher than the lifted condensation level, but never lower. Convective inhibition is discussed later in the module.



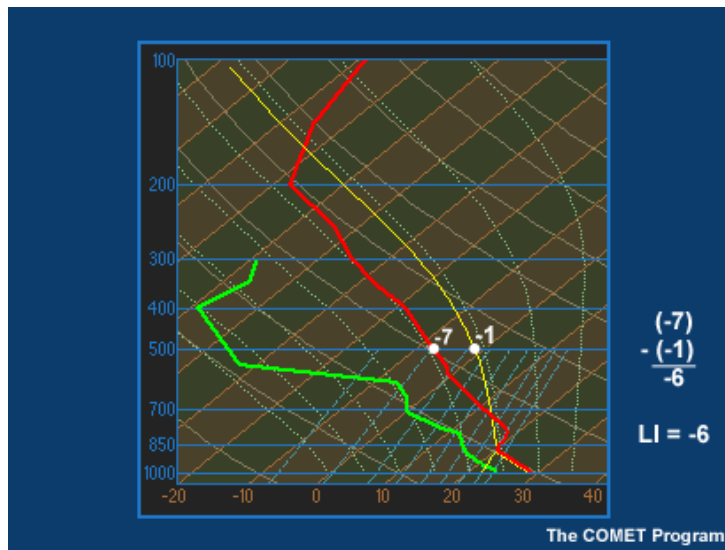
The best prognostic tool that forecasters can use to anticipate the potential strength of buoyancy processes is a thermodynamic diagram such as the skew T-log p diagram. By following appropriate parcel ascent and descent curves, a forecaster can use a skew-T diagram to estimate the potential strength of both updrafts and downdrafts for a convective cell. We explore these methods later in the module.

The lifted index (LI) is a simple parameter used to characterize the amount of instability in a given environment. It is calculated by lifting a representative parcel of air from the surface along the dry adiabat to its LCL, then along the moist adiabat to a level aloft, commonly 500 mb. Then we subtract the temperature of the lifted parcel from the observed temperature at that level.

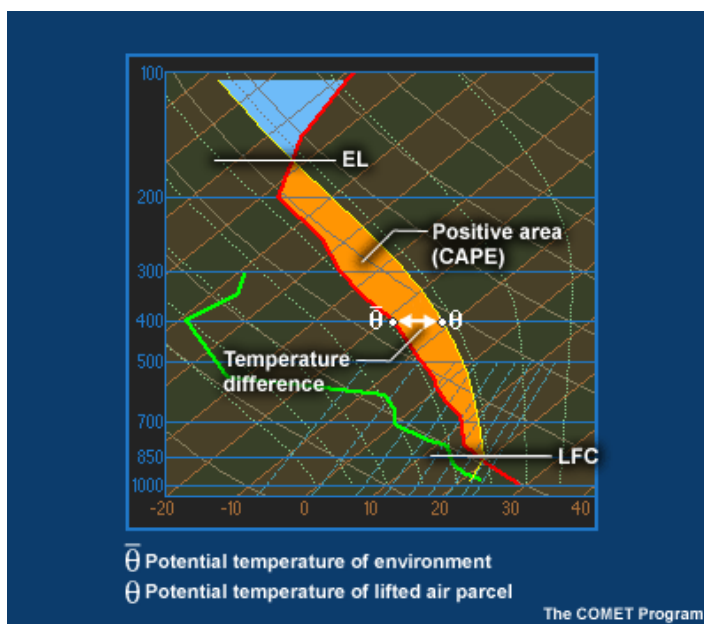
When the lifted parcel is cooler than the environment at 500 mb, the LI is positive, we have negatively buoyant conditions, and the parcel will sink.

Conversely, when the lifted parcel is warmer than the environment at 500 mb, the LI is negative, we have positively buoyant conditions, and the parcel will continue to rise. LIs less than zero thus suggest the potential for convection, with LIs less than -4 increasing the potential for severe convection. LIs of -10 to -12 are not that uncommon in the spring and summer months over the central plains of the U.S.

The LI is an easy-to-calculate estimate of the convective potential. However, since it is calculated at only one level, there are many situations in which it may not properly represent the true convective potential. For this reason, LIs should not be used without also having reference to the full sounding.



A more complete estimate of buoyant energy can be computed by determining the temperature difference between the ascent path and the environment at all levels from the LFC up through the equilibrium level. When we sum up all of these temperature differences, we arrive at a number equal to the positive area between the two curves, as shown in this figure. This area provides a measure of the integrated effects of the potential temperature differences between the rising parcel and its environment. The quantitative measure of this positive area is CAPE, or convective available potential energy. CAPE is measured in units of joules per kilogram (J/kg). Most sounding programs calculate CAPE.



When calculating CAPE, we normally lift a parcel that reflects the mean values of the temperature and moisture in the lowest 50 to 100 mb. This layer represents the average heat and moisture conditions fueling convective storms.

The environmental CAPE for convective storms is often in the range of 1000–2000 J/kg. However, values higher than 5000 J/kg sometimes occur.

Remember, you should always use CAPE in conjunction with a full sounding so that you can monitor the strength of low-level inversions and the position of the LFC, or other factors related to the vertical distribution of CAPE that could also modulate convection.

One reason why CAPE is such a useful parameter for forecasters is that CAPE is directly related to the maximum potential speed of the updraft, or Wmax, as shown by this equation. This equation is derived from a simplification of the vertical momentum equation by ignoring mixing, water loading, and pressure effects. In reality, these simplifications mean that Wmax often overestimates the actual updraft value by a factor of 2.

Using this equation, a relatively high CAPE of 2000 J/kg would indicate a Wmax of 63.2 m/s, or, allowing for the overestimate, a more probable Wmax of 31.6 m/s, or about 65 knots.

$$W_{max} = \sqrt{2 \text{ CAPE}}$$

When CAPE = 2000:

$$\begin{aligned} W_{max} &= \sqrt{2 (2000)} \\ &= \sqrt{4000} \\ &= 63.2 \text{ m/s} \end{aligned}$$

or, allowing for an overestimate of 2x
= 31.6 m/s (~65 kt)

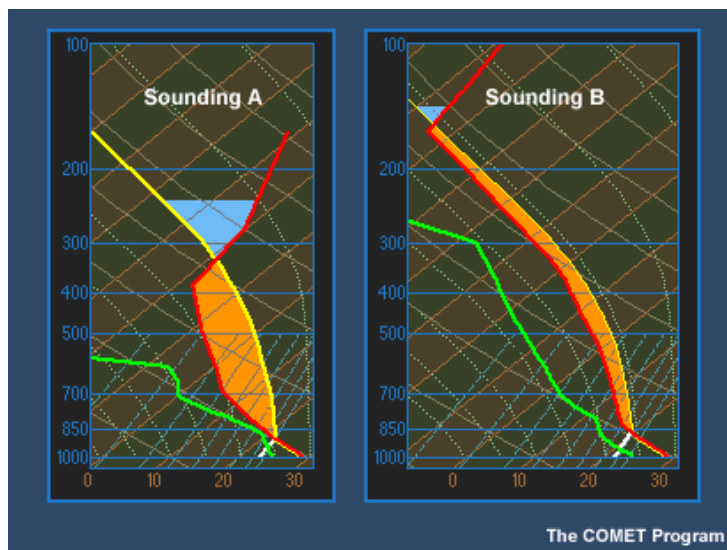
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Effects of Entrainment and Water Loading – 1

The net effect of CAPE on a rising parcel is influenced significantly by the vertical distribution of the buoyancy and moisture in the sounding. These influences arise from the effects of both entrainment (mixing) and water loading.

For instance, the two soundings available here display identical magnitudes of CAPE, but Sounding A has the CAPE concentrated in the lower half of the profile while Sounding B has the CAPE spread over a much deeper layer. All other things being equal, Sounding A would

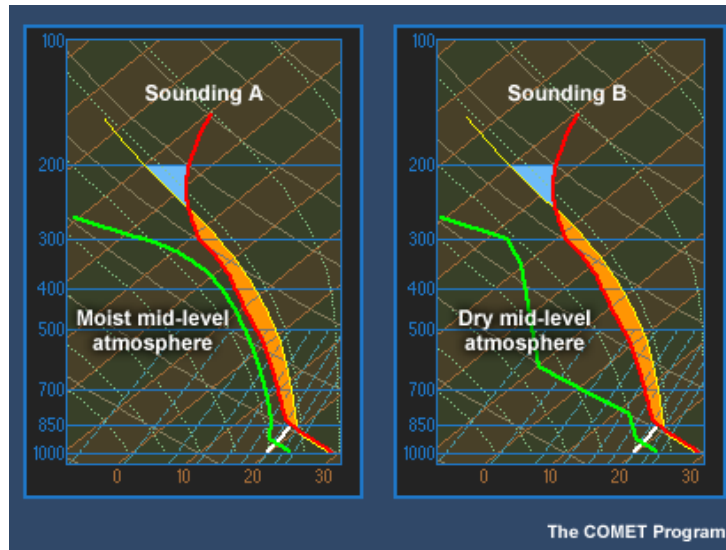
produce a stronger updraft than Sounding B, especially at lower levels. The primary reason for this is that a surface air parcel with Sounding A would accelerate more quickly, thereby not allowing as much time for the entrainment of dry, stable mid-level air that could reduce its net buoyancy. In addition, the stronger low-level updraft of Sounding A would carry more precipitation aloft thus reducing the amount of precipitation falling back into the lowest portion of the updraft. This reduces the downward drag of water loading in the bottom portion of the cloud, helping to maintain a strong updraft. Although the negative effects of entrainment with Sounding B will be minimal, the enhanced water loading effects will significantly reduce the maximum updraft strength compared with Sounding A.



Effects of Entrainment and Water Loading – 2

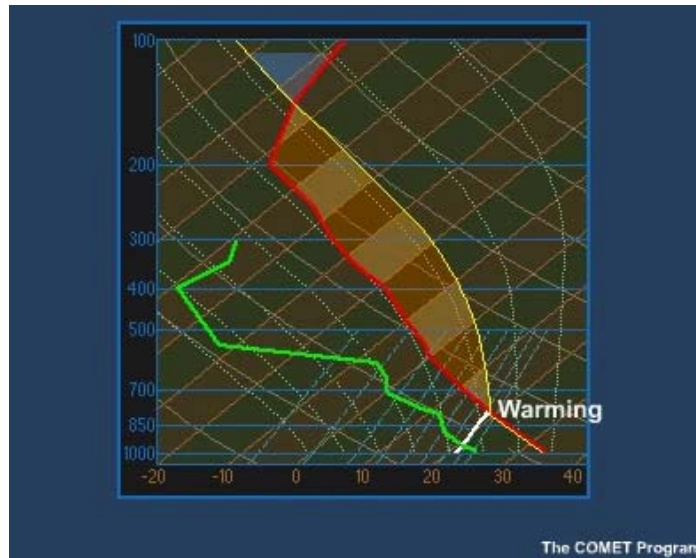
In contrast to the pair of soundings with equal CAPE on the previous page, soundings A and B on this page have identical vertical profiles of CAPE, but Sounding B is drier above the surface. Sounding A would produce a stronger updraft than Sounding B, but Sounding B would produce a much stronger downdraft.

The entrainment of dry mid-level air into humid air reduces buoyancy, primarily through cooling. Thus it will act to weaken an updraft and strengthen a downdraft.



Until now, we have ignored the common occurrence of a capping inversion, the strength of which is measured by Convective Inhibition (CIN). This lid can prevent near-surface parcels from reaching their level of free convection. As a result, this lid can prevent storms from forming, even with high instability aloft. In the presence of a capping inversion, some additional mechanism is required to initiate convection. In the absence of mesoscale lifting, there are three common mechanisms for overcoming a capping inversion:

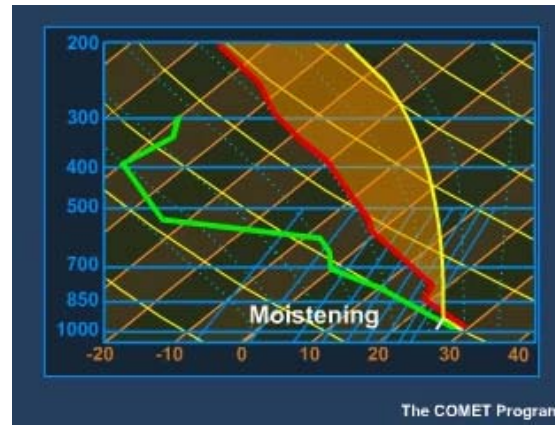
1. Heating (shown here)
2. Moistening
3. Synoptic-scale lifting



In the case of heating, the presence of a lid can prevent convection while the surface temperature climbs. As this figure shows, the instability, measured by CAPE, continues to grow until daytime heating eliminates the capping inversion. The ensuing convection can be much stronger than if it had occurred earlier, before significant heating had occurred. For example, an analysis of data from the central U.S. determined that the stronger the lid, the less likely convection was. However, if storms did form, they were more likely to be severe if the CIN was high.

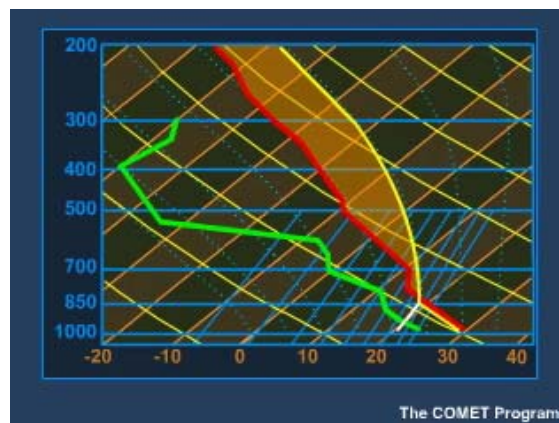
Another way to overcome a capping inversion is to add moisture to the lower atmosphere. This animation shows how moistening the lower atmosphere increases the dewpoint of a low-level air parcel. On the skew-T diagram, this moistening shifts the ascent path far enough to the right that rising air parcel remains buoyant throughout its ascent, eliminating convective inhibition.

Moistening in the lower atmosphere can occur through either low-level advection of moister air into the region or locally through evaporation from a local moisture source such as a lake or irrigated field. Low-level moist advection can produce large changes in a short time to overcome relatively large amounts of CIN as illustrated in the figure. Advection can be important even when the air is moist initially. Evaporation is more effective when the conditions are dry initially or only a small change is needed to overcome the CIN.



Overcoming CIN by Synoptic-scale Ascent

One other common mechanism for overcoming convective inhibition is synoptic-scale ascent caused by passage of a short wave or front. These processes act to lift and weaken the inversion layer. On the skew-T diagram, this effectively eliminates the inversion, shifts the inversion to the left, or some combination of the two. In either event, the inversion no longer acts to cap the ascent path of a rising air parcel. Because this process acts fairly slowly on its own, it will be most effective if it coincides with daytime heating and/or moistening of the boundary layer.

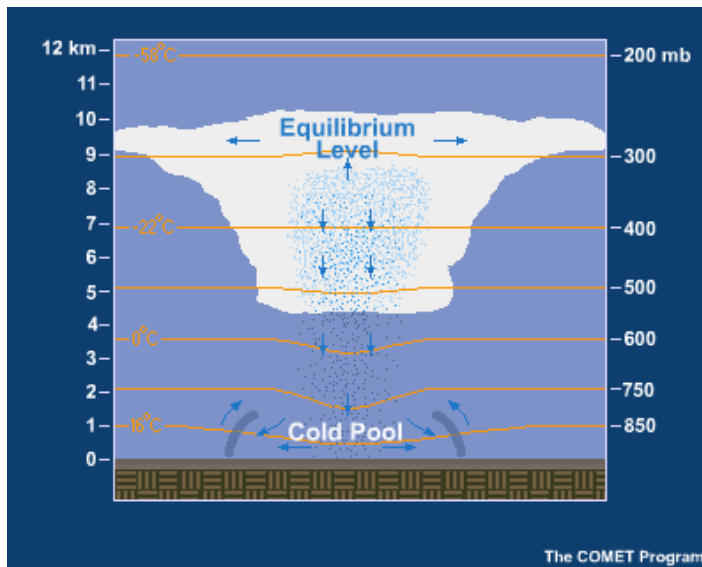


Downdraft Strength: Importance and Contributions - 1

Downdraft strength and the resulting surface cold pool determine the strength of convective winds and the cold pool's ability to trigger new cells. The primary contributors to downdraft strength include precipitation loading and evaporation.

Precipitation loading effectively adds weight to the rising air parcel. The magnitude of precipitation loading depends both on the amount of moisture in the parcel and on the strength of the updraft, which initially suspends the precipitation.

Evaporation makes the falling air parcel colder than the surrounding air. This makes the air more dense and consequently strengthens the downdraft. The effects of evaporation depend on the amount of precipitation that can be evaporated, the dryness of the air it evaporates into, and the precipitation type.



Downdraft Strength: Importance and Contributions – 2

In general, it is difficult to anticipate the net contributions to downdraft strength from precipitation loading. However, we can make some reasonable estimates of the potential contributions from evaporative cooling by again viewing the environmental sounding.

View each of the examples of estimating downdraft potential.

Example 1

Usually the downdraft originates near the level of minimum wet-bulb potential temperature, 3–5 km above the ground. In this example we assume that the downdraft originates at 700 mb and that parcels from this level remain saturated all the way to the surface and do not mix with updraft air.

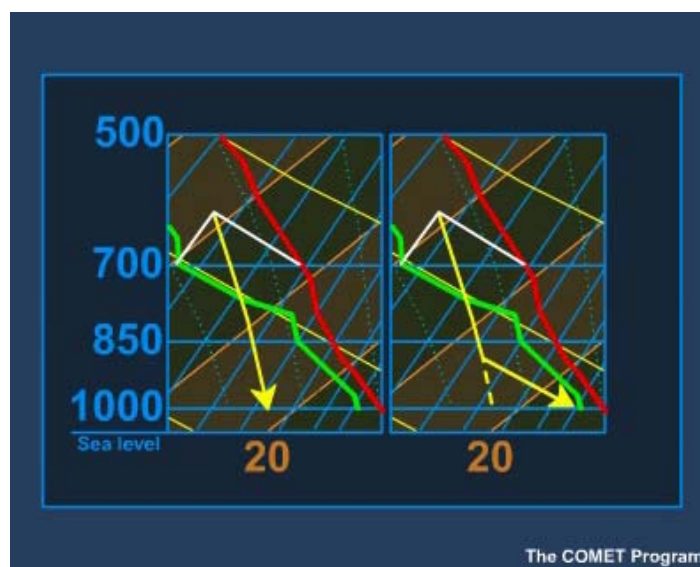
We can estimate the downdraft parcel's path by determining its wet bulb temperature, and then, from that temperature, following a moist adiabat to the surface. On the skew-T diagram, we determine the wet-bulb temperature by lifting the parcel along a dry adiabat to saturation. From that point, it is an easy matter to follow a moist adiabat to the surface. In the example shown here, we can see that the parcel remains cooler than the environment throughout its descent and thus accelerates downward.

DAPE, a downdraft equivalent of CAPE, can be calculated by integrating the negative temperature perturbations over the depth of the downdraft path.



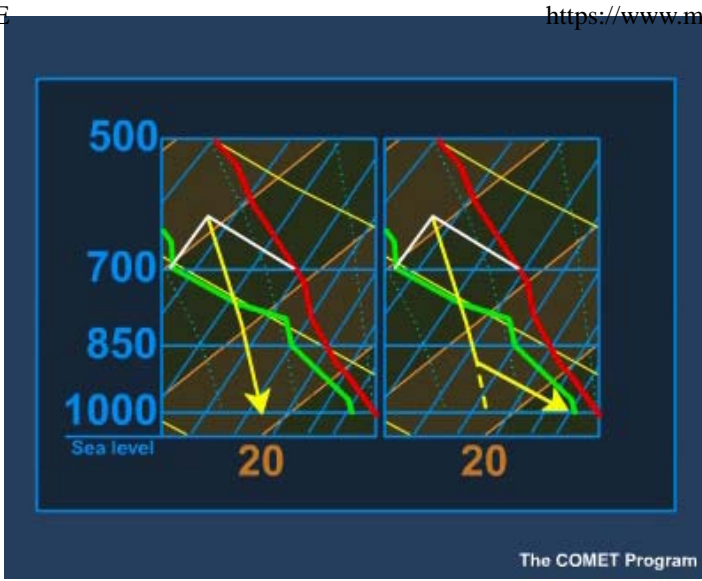
Example 2

Wet-bulb potential temperature gives us an estimate of the MAXIMUM cooling the parcel would experience. Cooling may be less if the parcel does not experience evaporation all the way to the surface. It would then follow the dry adiabat, rather than the moist adiabat.



Example 3

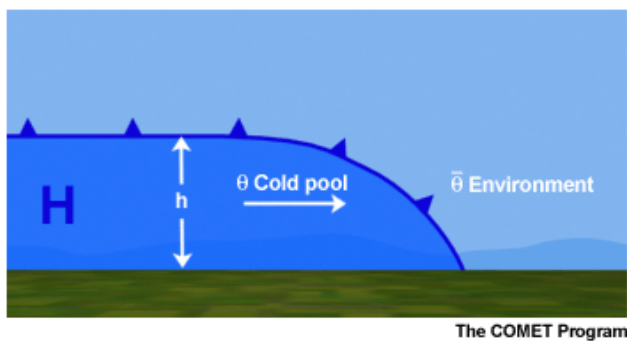
If the mid-level air was more moist, the wet-bulb potential temperature would be higher, and the potential cooling and downdraft strength would be less. This would also be the case if downdraft air mixed with warmer, moist updraft air.



Besides contributing directly to downdraft strength, evaporative processes also contribute to the strength of the surface cold pool, which largely determines the strength of the surface outflow winds beneath the storm.

The strength of a cold pool and its resulting wind gusts are related to the cold pool's theoretical speed of propagation. This propagation speed, in turn, depends on two things: (1) the depth of the cold pool and (2) its temperature relative to the environment. Deeper cold pools and greater temperature differences contribute to higher wind speeds.

The propagation speed of the cold pool and leading gust front affect the longevity of the storm or convective system. If the cold pool moves too quickly and outruns the storm, it can effectively cut off the supply of warm unstable air that had been fueling the storm.



The key points to remember are:

- When vertical wind shear is weak, buoyancy processes are the dominant control on convective updrafts and downdrafts
- Positive contributions to buoyant energy come from increases in potential temperature and water vapor
- Updrafts dominate during the early portion of an ordinary cell's life cycle; downdrafts dominate in the later stages
- Thermodynamic diagrams are essential tools for estimating vertical buoyancy distribution, which has important implications for both updraft and downdraft strength
- The Lifted Index (LI) is a simple quantitative estimate of buoyancy in the midlevels of many convective environments
- CAPE provides a quantitative estimate of buoyant energy, especially when calculated using an appropriate low-level average of both moisture and temperature
- CAPE can be used to estimate updraft strength (W_{max})
- A capping inversion or lid, measured by CIN, can either prevent convective storm development entirely or delay initiation until maximum heating is reached
- Downdraft and near-surface outflow strength depend on both precipitation loading and evaporation processes
- In general, drier mid-levels are associated with stronger downdrafts