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Principles of Convection

Atmospheric convection is the exchange of heat energy through vertical means, driven by buoyancy forces. Buoyancy is the vertical force exerted on an air parcel due to density differences. If a parcel is less dense than its environment, it will have an upward (positive) buoyant force, while a parcel that is denser than its environment will have a downward (negative) buoyant force. Therefore, an increase in water vapor content and warm temperatures correlate with positive buoyant force, and cool temperatures, low water vapor content, and cloud droplets contribute to negative buoyant force. If the upward buoyant force is greater than the vertical pressure gradient force, hydrostatic equilibrium will no longer exist and the air parcel will rise. Thus, buoyancy is the driving force behind convection.

Skew-T's

Skew-t diagrams are used to determine convective potential. The following are methods used:

Convective Available Potential Energy (CAPE)

If an unsaturated air parcel is lifted, the temperature will follow the dry adiabat, while the dewpoint temperature follows the constant mixing ratio. Where the constant mixing ratio crosses the adiabat on the sounding, this is the LCL and the parcel is now saturated. If the parcel is lifted further until it equals the environment temperature, this is the level of free convection (LFC) and the parcel will rise on its own, following its moist adiabat. As long as this adiabat is to the right of the temperature curve, the parcel will continue rising. In other words, the parcel is warmer than its environment and thus has an upward buoyant force. This parcel will continue rising until its temperature equals the environment temperature, where it will no longer have upward buoyancy. This level is called the equilibrium level (EL). CAPE is the total integrated positive buoyant force between the LFC and the EL, and can be shown on a sounding as the area enclosed between the moist adiabat and the environmental temperature curve running from the LFC to the EL. (see fig 1) It is measured in (J/kg). Thus, the higher the CAPE, the stronger potential there is for convection.

Using CAPE, the Wmax equation can determine updraft strength potential. (see fig2)

However, CAPE distribution must also be considered. (see fig.'s 3 and 4) Concentrated CAPE in the lower levels will produce stronger updrafts, as the parcel will accelerate faster, with slower updrafts occurring when CAPE is more evenly distributed. Stronger updrafts will also allow for less precipitation falling into it, reducing drag and maintaining the updraft for longer amounts of time. Also, water content affects updraft and downdraft strength, with a moister atmosphere causing more latent heat release and thus stronger updrafts, while a dry atmosphere absorbs latent heat, causing stronger downdrafts and weaker updrafts.

Convective Inhibition (CIN)

Contrasting to CAPE, if the moist adiabat traced from the LFC is cooler than the environmental temperature, this area is called convective inhibition (CIN). This is where the parcel will be denser than its environment (negative buoyancy). If CIN is between two areas of CAPE, it's called a capping inversion. To overcome CIN, a parcel must either be lifted to its LFC by synoptic means (such as a cold front), moistened, and/or heated until it is no longer denser than its environment. (see fig 5). This is most effective if all three occur at once; for example, a cold front pushing through during daytime heating with advection of moist air at the surface.

Lifted Index (LI)

A parcel from the surface is (hypothetically) lifted adiabatically to 500 mb. The temperature difference between the environment temperature and the parcel temperature is the lifted index (LI). Thus, if the LI is negative, there is potential convective instability due to the parcel being warmer than the surrounding environment. The potential increases as the departures become larger, with strong potential beginning around -6 degrees C or K. If the LI is positive, there will be positive buoyancy with much less potential for convection.

Downdrafts and Cold Pools

Precipitation falling from a convective cell cools the air nearby due to evaporation and latent heat absorption, as well as downward drag. This causes cool, dense air to flow downward in downdrafts, forming cold pools at the surface. (see fig 6) The downdraft strength and the cold pool determine convective wind strength and the ability to trigger new cells.

The downdraft originates near the level of minimum web bulb temperature, and this is determined on a sounding (fig 7). The parcel then follows the moist adiabat to the surface, as long as there is evaporation throughout the path downward. If not, it follows the dry adiabat and is weaker. Thus drier mid-levels cause stronger downdrafts, with moist profiles causing weaker downdrafts. The area integrated between the downdraft path and the temperature curve measures downdraft CAPE (DCAPE).

The strength of a cold pool determines the strength of the outward pushing downdraft winds. A deeper cold pool that’s colder relative to the environment will produce stronger winds. Also, the further away this cold pool moves from the parent convective cell, the more it will cut off warmer, less dense air, weakening the cell. (fig 8)

Wind Shear

Vertical wind shear is a description of how horizontal wind velocity changes with height. Vertical wind shear is vital in helping forecast convective storm type, convective initiation along cold pools, supercell probabilities, and storm motion. Two ways of expressing this quantitatively are the Bulk Richardson number and the storm relative environmental helicity (SHEH). The Bulk Richardson number is a ratio between one-half the square of the difference between the 6-km density- weighted mean wind speed and a 500-m mean surface layer wind speed, while SHEH is the integrated streamwise vorticity in the lower levels of a storm. These values can be visualized on a hodograph. Hodographs are an essential tool for vertical wind shear evaluation.

Hodographs

Since it has been determined that the layer from 0-6 km high is generally most important for thunderstorm forecasting, this is what's used here.

The winds at every 1 km from 0-6 km are first drawn as vectors, and then overlaid on each other. The resulting endpoints are then plotted on a polar grid. The axes represent compass directions, and concentric circles represent wind speeds. (fig 9.)

Since the vectors represent winds at even intervals, the depth represented is equal between each layer, and thus their lengths indicate wind shear at a given layer.

To determine the vertical wind shear magnitude between any two points, a vector is drawn between the two points. The length represents the speed of the vector, with the direction shown on the compass. (fig 10) Therefore, the vector represents wind shear magnitude over the depth of the connected layer. The total shear vector is the vector determined between 0 and 6 km.

Computer programs are used to determine the exact vector, but visual estimation can be done, as explained above.

To determine the mean vertical wind shear vector, the x and y components of each single layer wind shear vector are averaged. This is best done with a computer. However, mean wind shear direction (but not magnitude) can be determined by drawing a line from the surface vector to the 6 km vector.

Since storm motion is generally in the direction of the mean wind, calculating mean wind will indicate storm motion.

For a straight hodograph, the mean wind is simply in the direction of the line on the hodograph (fig11)

For a curved hodograph, each vector is averaged in terms of u and v components by a computer. However, this can be estimated visually by first re orienting the x-y reference frame (if needed), then visually averaging the u and v components using x' and y'. (fig12) This also works for multiple curves.

As seen in the above figures, the hodographs can have counterclockwise curves, clockwise curves, or straight lines. Straight lines indicate speed shear, while curves indicate directional shear. This is extremely important for determining storm processes, as will be explained.

Applications of Wind Shear and Convection

Without wind shear, buoyancy alone would control updrafts and downdrafts. However, the addition of shear strongly influences storm processes and longevity.

Cold Pools and Shear

Cold pools can initiate convection by lifting parcels until they reach the LFC. Without any shear, there are no favored areas for initiation. Also, a cold pool may not generate enough lift to create new convection.

As a cold pool spreads, horizontal positive vorticity is generated on the left edge, while horizontal negative vorticity develops on the right edge (fig13). This is caused by the buoyancy gradient between the cold pool and its environment.

With low level wind shear present, horizontal vorticity develops. This enhances the lift generated on the side of the cold pool where positive and negative vorticity interact. (The "downshear" side) This is due to each side of the vortices creating additive upward motion. On the other side of the cold pool, the additive influence of the vortices only serves to push air across the top of the pool, with minimal net lift. (fig 14) It is important to note that lifting across the downshear side is at a maximum when the vorticity generated by the cold pool equals the vorticity generated by wind shear.

Updrafts and Shear

As an updraft rises, horizontal vorticity is generated along the sides, similar to the edges of a spreading cold pool. Without shear, the vorticity on all sides is balanced and the cloud develops vertically. When shear is present, the horizontal vorticity created by the shear adds to the updraft generated vorticity, and the updraft tilts downshear. (fig 15)

Also, storm tilt results from the updraft blocking the flow across it, causing a pressure gradient. The storm will tilt toward the lower pressure, which is downshear. This allows for longer lasting updrafts, as downdrafts from the top of the storm will flow behind the storm, and not into new updrafts.

Shear and Convective Organization

If a cold pool associated with a cluster of multiple thunderstorm is in a sheared environment, initiation of new convection is favored along the downshear side of the cold pool. Thus, new storms can continuously form along this edge, while the downdrafts from older cells allow the new updrafts to be uninterrupted by downward motion. (fig 16) The longest lived and strongest multicell systems occur in strongly sheared environments, assuming no other forcing mechanisms.

Shear and Supercells

Under strong vertical low level wind shear, an updraft can acquire a rotating couplet, with the vortices rotating clockwise and counterclockwise. Each vortex separates and develops into split, mirror image supercells. Clockwise curved hodographs will causes a dominant right moving supercell, and a counterclockwise hodograph will cause a dominant left moving supercell. (fig 17) CAPE generally needs to be very high, and low level shear needs to be strong for supercell formation. The rotating updrafts allow for the storm to remain highly organized, with new generation of updrafts occurring as old ones die.

Convective Sytems

The interactions between cold pools and shear can cause squall lines, bow echoes, and supercells. The strongest squall lines generally form perpendicularly to the low level shear, especially at low levels. Severe squall lines tend to exist with strong low level shear. As shear deepens, the likelihood for bow echoes and supercells increases. Supercells generally characterize the deepest shear. Shear allows for cold pools to more easily initiate new updrafts as well as deflect downdrafts away from existing updrafts, as explained in the sections above.

Works Cited

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