Definitions & Computations

1. CAPE

A. SBCAPE (Surface Based CAPE)

Convective Available Potential Energy (CAPE) represents the vertically integrated positive buoyancy of a parcel experiencing adiabatic ascent. From the Skew-T Log P Diagram and Sounding Analysis Remote Training Module, RTM-230 (NWSTC, 2000) CAPE is expressed mathematically as

$$CAPE = g \int_{Z_{LFC}}^{Z_{EL}} \left(\frac{T_{vp} - T_{ve}}{T_{ve}} \right) dz,$$

where T_{vp} is the virtual temperature of the parcel and T_{ve} is the virtual temperature of the environment, Z_{el} is the height of the equilibrium level, Z_{LFC} is the level of free convection, and g is gravity. The units for CAPE are expressed in J/kg.

Alternate forms of the CAPE equation **do not use virtual temperature**, but use environmental and parcel temperatures in degrees Celsius.

In AWIPS, CAPE is labeled as the "*Positive Energy Above LFC*" on the Skew-T program, and is calculated using **potential temperature**.

The parcel potential temperature and sounding potential temperature are computed at each level between the LFC and the EL. When the parcel temperature is greater than the sounding temperature, the two values are then summed up. Only **one LFC** is computed; thus, double passes of the parcel trace across the environmental temperature trace are not accounted for (e.g., in midlevel inversions). In addition, if there are **small positive areas below the LFC** above the negative areas, then AWIPS skew-T will not compute these values. See the AWIPS validation web site for more details. (A graphical depiction of the positive and negative areas on a sounding resulting from a rising parcel originating at ground level is shown below in Figure 1.)

Note that Parcel Characteristics shown on the AWIPS Skew-T program are derived in different ways depending on the time of the sounding.

The parcel parameters of T and Td are computed one of two ways:

1) If the sounding is a morning sounding (06 UTC - 18 UTC), then the parcel temperature is the computed forecast max temp, and the parcel dewpoint is the lowest 50 mb mean buoyancy in the sounding. This is labeled in the display as a modified parcel, indicating the parcel has the potential to reach these characteristics during the day.

2) If the sounding is a 00 UTC sounding, then the surface temperature and buoyancy are used as the parcel temperature and dewpoint, respectively. This assumes that the atmosphere is pretty much mixed out. This forms the basis of all the parameters on the right side of the table display. Arrays of parcel temperatures, dewpoints, and pressures are computed containing these parcel characteristics.

(Note: In BUFKIT, starting with version 4.02, there are many options for computing and displaying CAPE, based on selectable parcel lifting level, but once the parcel's lifting level is defined, the CAPE computations are similar to AWIPS; i.e., it sums up the positive energy area above LFC to the EL and does not include any negative areas).

However, in BUFKIT, one can account for midlevel inversions and more than two positive/negative areas on the sounding by using the ECAPE manual parcel lifting options (See Section C-Most Unstable CAPE).

Note: Neither AWIPS Skew-T or BUFKIT CAPE computations incorporate a virtual temperature correction to the parcel path and environmental temperature trace to account for the effect of moisture on air density (buoyancy).

NSHARP does incorporate a Virtual Temperature Correction in their thermodynamic computations. See this AWIPS sounding validation web site for information.

This effect will slightly increase the temperature of the parcel and the environment in moist low levels. Also, when lifting the parcel with a virtual temperature correction, it does not exactly follow the moist adiabatic lapse rate since the skew-T moist adiabats have typically been constructed using temperature and not virtual temperature. For more description on how AWIPS calculates CAPE (and other parameters) see the [AWIPS validation web site](#).

For information on the various lifting methods employed on the Interactive Skew-T Program in AWIPS, see the AWIPS User's Manual.

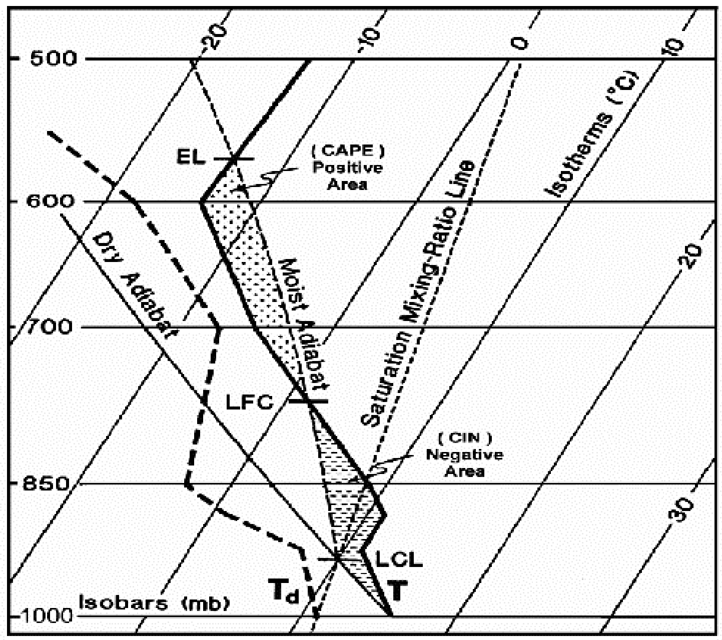


Figure 1. Showing the Positive (CAPE) and Negative (CIN) areas (from NWS/OSF/OTB, 1991).

Strengths

CAPE integrates a substantial portion of the thermodynamic information contained in a sounding. It is proportional to energy available for a rising parcel. CAPE provides an estimate of maximum updraft strength (**W_{max}**) in convective storms by the relationship:

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

CAPE is a fundamental indicator of potential intensity of deep, moist convection. Operationally, CAPE is more popular than indices such as Lifted Index or K Index which uses temperature and dew point from only a few mandatory levels in a sounding.

Research has shown that low-level CAPE may have relevance to tornado production. More CAPE in the lowest levels above the ground suggests stronger potential for large low-level accelerations and enhanced low-level mesocyclone intensification.

A recent study, Davies (2002) showed that stronger supercell tornadoes tend to have more low-level CAPE (0-3 km) than non-tornadic storms (see this graph from Davies, 2002).

Weaknesses

Sensitive to both magnitude of buoyancy and the depth of integration.

In AWIPS, there is no way to quantify layered CAPE, for example, surface to 3 km CAPE.

Surface based computations do not account for layers that are not well mixed, and may grossly underestimate buoyancy in situations where parcels are experiencing elevated ascent.

As in all parcel theory indices, CAPE assumes no mixing with the surrounding environment, and ignores effects of freezing and water loading. If ambient temperature is used instead of virtual temperature to calculate CAPE, lower CAPE values will result.

See this graphic showing two forecast soundings from the Eta and LAPS models from 1200 UTC 25 March 2002 in southern Oklahoma. Both forecast soundings depicted zero net CAPE (Surface-based, most unstable CAPE and mean layer CAPE), which could be misapplied because there is CAPE in the sounding. To compute it, you must manually select (lift) the parcel from above the stable layer (~ 780 mb).

On the other hand, the Eta BUFKIT sounding from the same time and location, because the parcel lifting level is interactively selectable from the most unstable level, came up with an elevated CAPE value of 735 J/kg. (Note: Severe hail occurred in the vicinity of this sounding across portions of southern and southeastern Oklahoma.)

SBCAPE will **overestimate** realized instability when soundings possess shallow moist layers. SBCAPE value alone does not account for effects of vertical distribution of CAPE.

The estimates of maximum updraft strength (W_{max}) based on CAPE are usually twice as high as in observed updrafts because of water loading and mixing effects. In well-organized convective storms, vertical velocity in updrafts are much closer to W_{max}.

Supercells can have strong updrafts even when the static instability, as measured by CAPE, is modest (See McCaul and Weisman, 2001). This is due to vertical shear effects.

The virtual temperature correction can increase low-level CAPE calculations by 20-50 J/kg (see this graph from Davies, 2002).

Definitions & Computations

B. Mean layer CAPE (often abbreviated MLCAPE)

On SPC discussions, this is the CAPE calculated using the lowest **100 mb AGL mean layer** temperature and moisture values (same as in BUFKIT).

In the AWIPS interactive Skew-T, a similar computation is the "Mean Temp Lift" parcel lifting option. However this uses the lowest 50 mb mean potential temperature layer above the surface, so these values will differ slightly from BUFKIT.

Strengths

MLCAPE is more representative of realized buoyancy because it incorporates parcel mixing effects. MLCAPE and low-level lapse rates have been shown to be two good parameters for discrimination of general thunderstorms (i.e., whether convection produces lightning). See Craven et al., 2002.

Soundings taken in proximity of thunderstorms usually possess more than 250 J/kg of MLCAPE.

MLCAPE, when combined with LCL height, has been shown to be a very good discriminator for tornadic supercells (see Craven et al. 2002).

Weaknesses

MLCAPE is likely to underestimate elevated and/or surface-based buoyancy if layers are not well-mixed. MLCAPE has more difficulty in discriminating between general thunderstorms and severe thunderstorms (lots of overlap). See Craven et al. (2002).

Definitions & Computations

C. Most Unstable CAPE (often abbreviated MUCAPE)

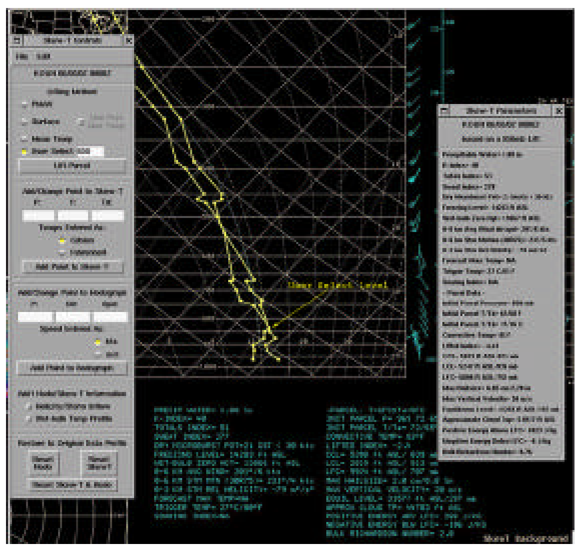
In SPC discussions, this is CAPE calculated by using a parcel from a pressure level which results in the **most unstable CAPE** value possible in the lowest 300 mb AGL.

(Note: this is different than PMAX on AWIPS Skew-T, which uses the most unstable parcel in the lowest 50 mb, based on highest wet bulb temperature.)

For information on the various lifting methods employed on the Interactive Skew-T Program in AWIPS, see the AWIPS User's Manual.

The best way to assess elevated instability using AWIPS skew-T program is to visually inspect the sounding and pick the parcel level where the most CAPE results above the surface (**note: use the "User Select" option for lifting method**). Usually, this is between 900 and 700 mb.

See this example of selecting a parcel on an AWIPS skew-T below.



Strengths

MUCAPE is the best sounding measure for elevated buoyancy and assessing potential for elevated convection.

In BUFKIT, one can interactively compute CAPE from any level. See this BUFKIT sounding example of a parcel lifted from 764 mb and the resulting CAPE of 957 J/kg.

Note the following differences in the characteristics of the air parcels used to compute CAPE for various model analysis packages:

1) ETA "surface" CAPE (from Volume Browser) uses the highest $_{e}$ in the 70 mb just above the surface. (Note in BUFKIT, the average temperature and dew point in the lowest 100 mb is the default CAPE, but parcel lifting levels can be modified more easily than in AWIPS).

The ETA "boundary layer" CAPE uses the highest $_{e}$ averaged for one of six 30-mb-boundary layers located in the 180 mb just above the surface (for example, if the surface was at 1000 mb, the first layer would be 1000-970, the second 970-940, the third 940-910, etc.).

2) In the RUC, the air parcel used to compute CAPE is from the maximum 40 mb average $_{w}$ in the lowest 300 mb above the surface.

3) In the AVN/MRF, the "surface" CAPE uses the avg. surface parameters for each grid box. The "most unstable" CAPE uses the highest $_{e}$ from the boundary layer at each grid point.

4) In LAPS, the "surface parcel" uses the latest LAPS surface temperature, dew point and elevation to lift the parcel at each grid point. LAPS CAPE is a new positive energy, so any negative energy is subtracted from the positive energy.

Weaknesses

Compared to Surface-based CAPE, PMAX-based CAPE will occasionally result in slightly lower CAPE values when most unstable parcel is at the surface.

"Tall-thin" CAPE is more susceptable to waterloading than "short and fat" CAPE. For example, tropical storms, which develop in soundings characterized by high ELs, and tall-thin CAPEs, are not as likely to be as deep (in terms of convective growth) as shallow-topped cool season supercell storms, where representative soundings indicate short-fat CAPE.

Definitions & Computations

2. CIN

Defined as Convective INhibition energy (CIN), a measure of the "negative area" on a sounding between the surface and the LFC.

CIN is the amount of work required to lift a parcel through a layer that is warmer than the parcel. The parcel must be forced upward sufficiently to overcome the negative buoyancy. This negative area is often referred to as a "lid" or "cap". The formula for CIN (from RTM-230) is very similar to CAPE:

$$CIN = g \int_{Z_{SFC}}^{Z_{LFC}} \left(\frac{T_{vp} - T_{ve}}{T_{ve}} \right) dz,$$

where Z_{SFC} is the height of the surface and all other variables are the same as in the CAPE calculation. The units for CIN are the same as for CAPE, in J/kg. **Other computations use temperature (or potential temperature) instead of virtual temperature**. The larger the CIN value, the more stable the layer of air is between the surface and LFC, the more difficult it will be to lift a parcel of air to its level of free convection.

In order to get the parcel of air to reach the LFC, the forcing mechanism has to have sufficient strength to push the boundary layer air through the negatively buoyant layer.

See Figure 1 for a depiction of positive and negative areas on a sounding.

Note: In AWIPS, the negative area encompasses the negative energy below the LFC. Similar to AWIPS CAPE calculations, the potential temperatures of both the sounding and the parcels are computed at each level and summed up for **both** 1) below the LCL (to the ground), and 2) above the LCL (to the LFC). The negative area then includes both potential temperature calculations. Note: as in AWIPS CAPE calculations, **actual negative area in a Skew-T might be underestimated in AWIPS** because it only computes one LFC. Once a parcel reaches the LFC, even for just a short time, any negative areas will not be accounted for.

Note: If a sounding contains an LFC beneath a midlevel inversion, the AWIPS CIN computations will not reflect this energy. If a sounding has a very small positive area below any CIN area, the AWIPS CIN computation may erroneously display no CIN (see this sounding for an example).

Note: Similar to CAPE, neither AWIPS or BUFKIT incorporates a Virtual Temperature correction to parcel trajectory. Doswell and Rasmussen (1994) mention CIN in their conclusions regarding the virtual temperature correction with respect to calculating CAPE. They point out that CIN values are relatively small in convective situations and that an average virtual temperature change of 1 K can affect CIN by about 35 J/kg which would be significant if the CIN was approximately -100 J/kg. Consequently, they suggest making the virtual temperature correct to CAPE calculations as well as CIN. CIN is also briefly discussed in the COMET CD-ROM module "Anticipating Convective Storm Structure and Evolution" (1996) in relation to CAPE and convection.

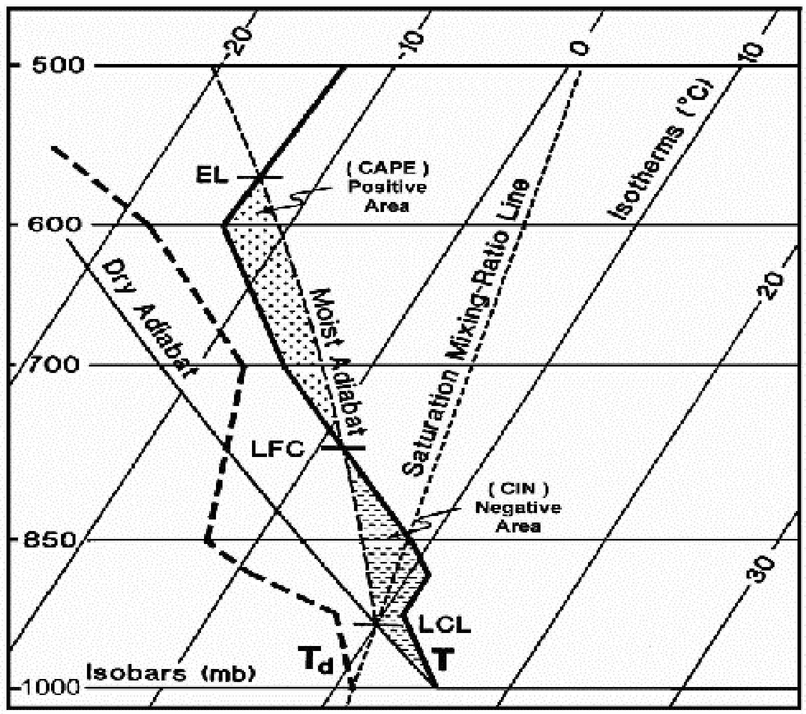


Figure 1. Showing the Positive (CAPE) and Negative (CIN) areas (from NWS/OSF/OTB, 1991).

Strengths

Assessing Convective Inhibition (CIN) is important to diagnosing the potential for deep convection. Generally speaking, the larger the value of CIN, the more difficult it will be for a parcel of air to reach the LFC. This statement is most applicable for a surface based parcel. For parcels that are **not** surface based (e.g., elevated convection), an appreciable amount of low-level CIN can be present, but parcels can still become positively buoyant if forced ascent occurs **above** the stable layer and some CAPE is present above the stable layer (in many of these elevated convection cases, the level above which lifting occurs is from 850 to 700 mb). In these cases, CIN above the LFC is usually minimal. In cases where CIN is large, storms and supercells are less likely to produce tornadoes (Grant, 1995).

Most models depict CIN as a convective forecast parameter, displayable via the AWIPS Volume Browser. There are differences in both individual model forecast CIN computations and CIN values derived from the AWIPS interactive Skew-T.

(See the AWIPS validation web site for more information on how AWIPS computes CIN.) The differences in CIN calculations are a result of what parcel level is lifted. (See above section on CAPE)

A nowcasting version of CIN is also available as a product derived from the GOES-8 Sounder. GOES derived products (including CIN) are heavily based on the ETA model forecast soundings as a first guess and are only produced in cloud-free regions.

One can relate CIN to a vertical velocity, w_{lift}, or the estimated amount of lifting required to overcome the negative area by the following expression:

$$W_{lift} = \sqrt{2 * CIN}$$

Weaknesses

It is often quite difficult to assess how much lifting will overcome the negative energy (CIN). Normally a parcel will need to be lifted by some external process in order to reach its LFC. Mesoscale sources such as boundaries are the usual mechanisms which supply sufficient lifting.

CIN calculations in AWIPS might **overestimate** the amount of CIN in a sounding because the AWIPS calculation **does not** apply a virtual temperature correction. This is especially likely when the CIN area is quite moist. An average virtual temperature change of 1 K can affect CIN by about 35 J/kg which would be significant if the CIN was approximately -100 J/kg (from COMET AWIPS validation page).

CIN is sensitive to changes in boundary layer values. A change in the surface dew point or the mean mixing ratio in the boundary layer will change the value of CIN. When selecting the start point for lifting a parcel, be sure to accurately reflect the boundary layer conditions at the time when you expect convection to begin.

As in all parcel theory indices, **CIN assumes no mixing** with the surrounding environment, and ignores water loading. The value of CIN will vary **depending on the parcel chosen to lift**. In cases of elevated instability surface based CIN may be quite misleading. As a result the operational use of CIN is far from easy. However, for surface-based convection, given an adequate forcing mechanism, the **probability of deep convection increases** when CIN decreases below 50 to 70 J/kg, **but it is quite difficult to determine an exact threshold value below which convection will (or will not) occur**.

For example, see this tornado proximity sounding from Davies (2001) as an example where significant CIN (150 J/kg) remained in close proximity to a large tornado-bearing storm.

Total CIN below the effective LFC may be grossly underestimated in AWIPS skew-T sounding output (up to 70 J/kg) due to computational restrictions in the AWIPS skew-T program for using only one LFC.

The virtual temperature correction can increase CIN by 20-50 J/kg.

Definitions & Computations

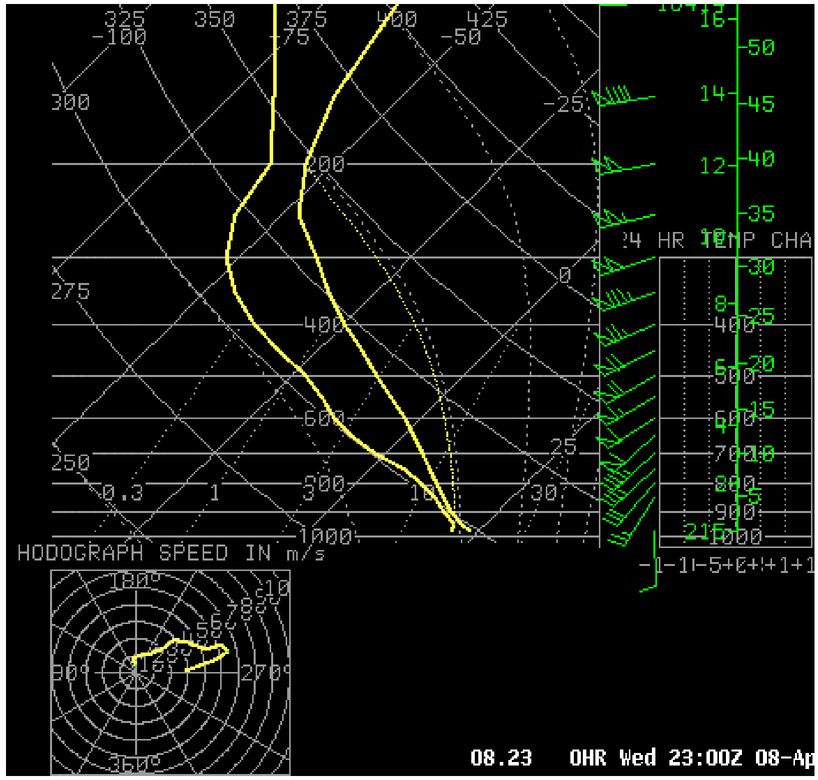
3. LCL

The Lifting Condensation Level (LCL) is the height at which a parcel becomes saturated when lifted dry adiabatically (see Figure 1). The LCL is commonly used to estimate the level of a **cloud base** from surface based convection. The computed LCL using a **Mean 100 mb Layer (MLLCL)** from the surface has been shown to have the **highest correlation** to measured cloud base (Craven et al. 2002).

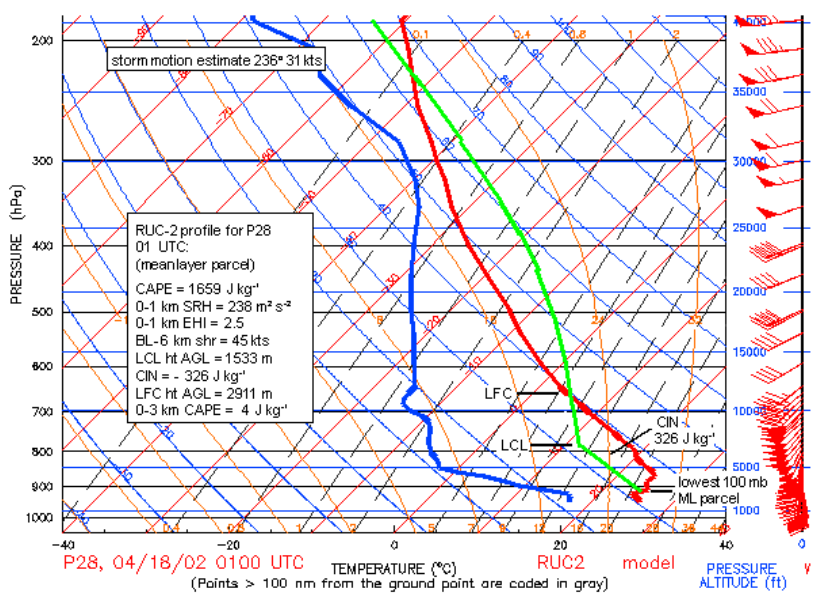
Representative parcels for determining the LCL and associated stability are **dependent on temperature and dew point mixing proportions** in the boundary layer.

The SPC uses a mean 100 mb layer parcel to compute LCL height.

The illustration below is an example of a sounding with a relatively low (~ 1935 ft AGL) LCL. This LAPS sounding was from 2300 UTC on April 8, 1998 at a point near Birmingham, AL (BMX). This is a representative tornado proximity sounding for the destructive tornado-bearing storm that struck near Birmingham later that evening.



The sounding below is an example of a sounding with a relatively high LCL. Several supercell storms which developed in the environment characterized by this sounding did not produce tornadoes. Note strong vertical shear (45 kts from 0-6 km) was present in this sounding. See this web site for details on this case.



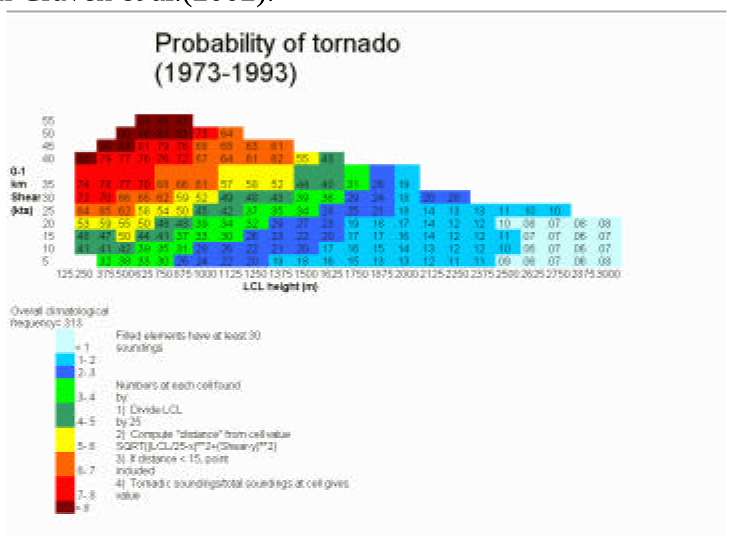
Strengths

Recent research has related the LCL to the amount of low-level relative humidity which can affect cooling through evaporation of rain in the downdraft portion of supercell storms (See Markowski et al. 2002). The higher the LCL is in the near-storm environment, the drier the boundary layer will be. Lower LCL heights and thus, lower cloud bases, are associated with greater amounts of boundary layer moisture and appear to indicate a higher frequency of significant tornado events (See again Craven et al. 2002).

Relatively low LCLs suggest greater low-level relative humidity near the ground and thus, more unstable air originating in the Rear Flank Downdraft (RFD), which researchers have claimed is critical to tornadogenesis (Markowski et al. 2002). Lesser values of boundary layer relative humidity (from high LCLs) might increase stability in Rear-Flank Downdrafts (RFDs) and decrease tornado potential. Rasmussen and Blanchard (1998) showed that LCLs in tornadic supercell soundings were significantly lower (Median value was approximately **800 meters** AGL with no occurrences above 1500 meters AGL) than LCLs in nontornadic supercell soundings.

METAR temperature dew point depressions ($T\text{-}T_d$) are a decent proxy to the local LCL height in a well-mixed boundary layer, so this parameter can be analyzed hourly on the mesoscale. **$T\text{-}T_d$ spreads** at the surface ranging from **0 to 22 F** correspond to LCL heights less than 1500 m AGL in a well-mixed boundary layer and **12 F** spreads correspond to 800 m.

A combination of LCL height (using mean 100 mb layer parcel) and 0 to 1 km shear has been shown to be highly correlated to significant tornado occurrence. See figure below from Craven et al.(2002).



The graphed data from Craven et al. (2002) show a strong signal between significant tornadoes (F2 or greater) and significant hail/wind. Significant tornadoes tend to occur with relatively high 0-1 km shear and relatively low LCL height (e.g. less than 1500 m AGL). On the other hand, storms that produce big hail (greater than 2") and/or wind gusts 65 knots or greater, but no strong or violent tornadoes, tend to possess weaker low-level shear and higher cloud bases.

LCL height is **not** affected by the virtual temperature correction.

Weaknesses

Major variations can occur in small time and space scales with LCL. Actual LCL heights in tornadic storms may be considerably lower so RFD approximations by surface or model data are quite crude at times. LCL computations **suffer the same limitations** as that of CAPE and CIN calculations in terms of parcel origination levels. **Be aware of the level where the saturated parcel originated. The MLLCL may be the best approximation to actual cloud base.**

Definitions & Computations

4. LFC

The LFC, Level of Free Convection, is the height at which a parcel lifted dry adiabatically to saturation at the LCL and moist adiabatically above the LCL would first become warmer (less dense) than the surrounding air. At the LFC, the parcel experiences positive buoyancy and starts to accelerate upward without further need for forced lifting (See Figure 1 for the graphical procedure).

Strengths

Low-level CAPE and CIN are related to the height of the LFC (see Figure 1). **Lower** LFC heights imply **more** low-level CAPE and thus, can be correlated to **increasing tornadic likelihood in supercells** because of the associated potential for stronger low-level vertical accelerations (see this graph of LFC height from Davies, 2002, study of supercell storms, Rasmussen, 2001 and more cases from Davies, 2002).

In addition, **higher** LFCs tend to imply **more** CIN, and **lower** tornado probability.

Weaknesses

A relatively low LFC height, by itself, does not say anything about the depth of CAPE or total CAPE. Total CAPE and of course, shear, must also be assessed for severe potential. CIN may be a better indicator of whether a storm is surface based and thus, have a higher tornado potential. The virtual temperature correction can lower the effective LFC by 200-500 m (see this figure from Davies, 2002).

Definitions & Computations

5. Shear

A. Bulk Shear

Bulk shear is calculated by computing the magnitude of the shear vector between two layers, such as the surface or boundary layer (ex. 0-500 m AGL mean wind) and a representative middle layer (such as 6 km AGL). In BUFKIT, the bulk shear is labeled "Shear layer difference" and can be plotted on the overview screen (See this BUFKIT overview example).

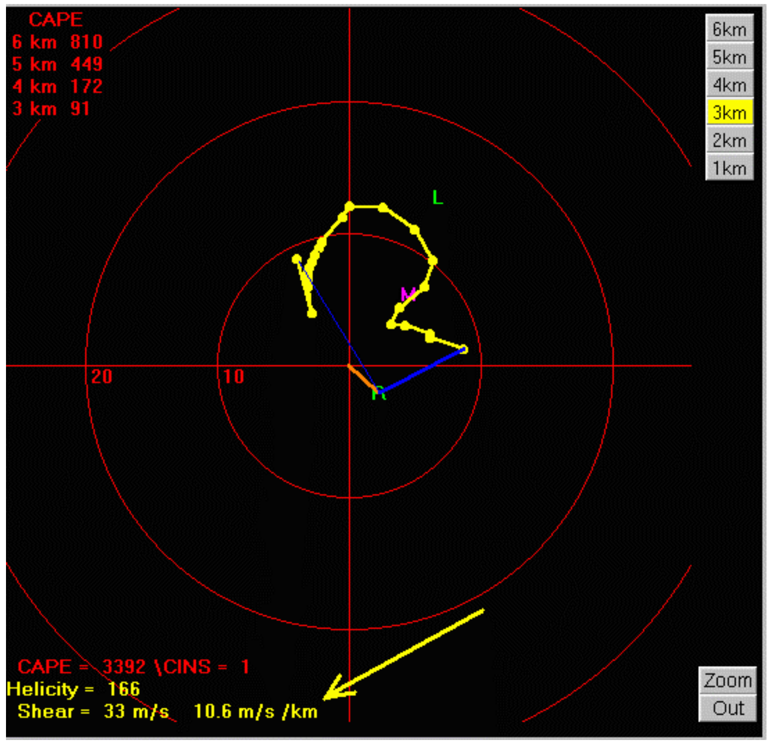
B. Mean Shear

Mean shear is defined as the length of the hodograph divided by the depth over which the hodograph was measured. This quantity is computable in the BUFKIT overview screen by selecting the button labeled "Shear (length of hodo)" and clicking on units of (m/s)/km.

The value of shear shown below which appears in the lower left of the hodograph graphic in BUFKIT is actually hodograph length [in m/s and in (m/s)/km]. The ending point that determines the length of the hodograph computations is selectable in kilometer increments from 1 to 6 km. The default is 4 km. (Note: CAPE values are also displayable in layer integral amounts as well).

Note: Bulk shear and mean shear are not computable from an AWIPS skew-T display; however, you can plot the forecast values on plan view from any of the models available. (See this AWIPS D2D example of 0-6 km bulk wind shear vector from a 36 hr ETA forecast)

Note: Typical AWIPS skew-T shear values derived from model soundings will be based on only 2 or 3 data points in the first kilometer due to the vertical remapping procedures employed in AWIPS (50 mb vertical resolution). Thus, AWIPS model soundings may show unrealistic shear values. If at all possible, use the BUFR files or a sounding with native model resolution in displaying vertical wind shear on soundings, especially when computing mean shear in the lowest 1 or 2 kilometers.



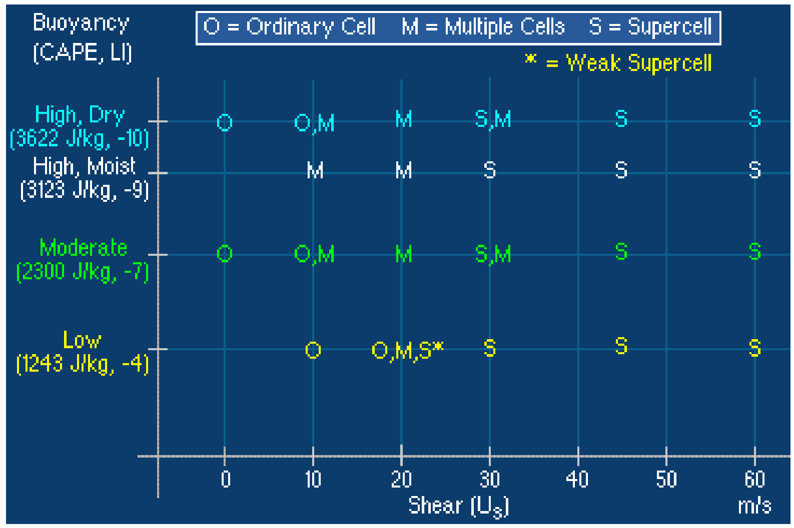
Strengths

Shear is the **most important** parameter for convective storm **organization and persistence**. Increasing vertical shear (for a given amount of thermodynamic instability) often results in greater convective storm organization, and longevity.

From observations and numerical modeling simulations, bulk shear, mean shear, and/or hodograph length have all been used to help quantify the amount of vertical wind shear capable of producing the dynamic pressure perturbations and resulting midlevel rotation in supercells.

The interaction of the updraft with an environment characterized by strong vertical shear of the horizontal wind permits some storms to develop nonhydrostatic vertical pressure gradients that can be as influential in developing updrafts as the buoyancy effects (Weisman and Klemp, 1984).

From the COMET CD-ROM, " A Convective Storm Matrix", the figure below shows hodograph length versus CAPE in simulations of various storm types.



Operationally, lower-bound thresholds of bulk shear (0 to 6 km) of **15-20 m/s** and mean shear values around **.001 s^{-1}** can be used as a first approximation to help determine potential supercell environments. Note: additional factors (e.g., buoyancy distributions, mesoscale variations, etc.) should be considered as well because they can significantly modulate the character of severe storm environments.

Rasmussen and Blanchard (1998) found that mean shear in the lowest 4 km AGL was able to distinguish (to a degree) between supercells that produced significant tornadoes and those that only produced large hail. Recent and ongoing research has focused on mean shear in the lowest kilometer above the ground and has found even more distinguishing signals (See section on 0-1 km SRH). SPC typically uses 20 kts of shear in the lowest 1 km AGL as a lower bound threshold for a significant tornadic supercell.

Other research such as Craven et al. (2002) and Markowski et al. (2002) using proximity soundings have found that the 0-1 km layer shear is the primary distinguishing kinematic parameter that separates supercells that produce significant tornadoes from those that do not. See this figure from Craven et al. (2002) which shows a remarkable lower threshold of 10 m/s (20 kts) in the statistical distribution.

Also, see Markowski et al. (2002) study of RUC model proximity soundings which showed a statistically significant difference in the lowest 1 km layer.

Observations of mature derecho environments (Evans and Doswell, 2001) suggested that bulk shear in the lowest 2 km was predominately greater than 15 m/s when combined with high CAPE.

Weaknesses

Bulk shear (surface to 6 km) has limited utility in distinguishing between supercells that produce significant tornadoes and those that do not (see Rasmussen and Blanchard, 1998).

Hodograph length is more sensitive to vertical resolution and noise in the observations. Computations using numerous model sounding layers often yield unrealistic high values of shear and should be smoothed.

Definitions & Computations

6. Storm-Relative Helicity (SRH)

Storm-Relative Helicity (SRH) is proportional to streamwise vorticity and storm-relative winds and takes into account storm motion. The mathematical expression for SRH, as defined Davies-Jones et al. (1990) is

$$SRH = \int_0^h (V - C) \cdot w dz$$

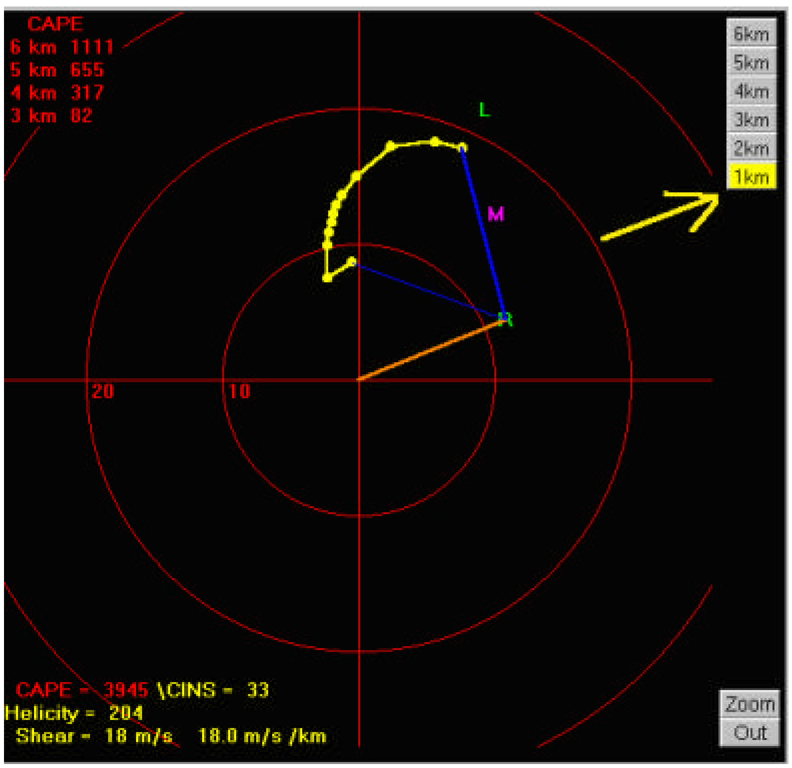
where V is the horizontal velocity (ground-relative vector wind), C is the storm motion, and w is the horizontal vorticity vector. The integration is over the inflow layer of the storm from 0 km to some depth h (typically 1 to 3 km).

SRH is computed in AWIPS skew-T soundings from 0 to 3 km and it uses a default storm motion of 30R75 (30 degrees to the right and 70 % of the 0-6 km density weighted average wind). See this example (SRH represents twice the area swept out by the storm-relative wind vectors in the lowest 3 kilometers).

In BUFKIT, the SRH is labeled "Helicity". The storm motion vector used in the helicity computations incorporate the Bunker's Storm Motion Technique (See this COMET web site), which has proven physically and statistically superior to the storm motion 30R75 used in AWIPS skew-T calculations.

Note: The RUC and Eta model output of Helicity incorporates the Bunker's Storm Motion, displayable from the Volume Browser. However, the LAPS model uses a slightly different storm motion for its SRH calculations. LAPS storm motions are typically 25 degrees to the left of the Bunker's Storm Motion and thus, often result in considerably less SRH values than the Eta or RUC (see this web site for more details). This AWIPS hodograph composite shows the slight differences in storm motion computations and resulting SRH for three models (Eta-yellow, RUC-blue, LAPS-red) for the same forecast sounding.

Note: In version 4 of BUFKIT, there are now options to manually integrate SRH at various layers above the ground, such as from 0 to 1 km. See the hodograph example below of 0-1 km SRH.

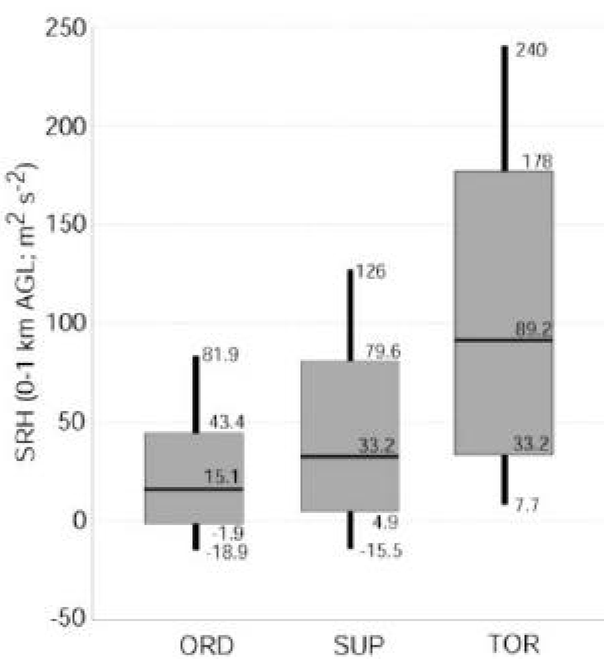


Strengths

Research and operations have found some correlations between increasing SRH values (from the surface to the lowest 3 kilometers) and tornado intensity [Johns et al. (1990), Davies-Jones et al. (1990), and Kerr and Darkow (1996)]. Observed 0-3 km mean SRH using Kerr and Darkow's proximity sounding study showed the following SRH values for various intervals of F scale: Mean 0-3 km SRH was 66 m^2s^{-2} for FO, 140 m^2s^{-2} for F1 tornadoes, 196 m^2s^{-2} for F2, 226 m^2s^{-2} for F3 tornadoes, and 249 m^2s^{-2} for F4 tornadoes. (Note: No F5 tornadoes were in their study).

However, operational experience has shown that current or projected 0-3 km SRH values exceeding 100 m^2s^{-2} often reflect a potential for supercells. The higher the SRH, the greater the potential for supercells.

Recent research (Rasmussen, 2001) has found discrimination ability between 0-1 km SRH and supercells that produce significant tornadoes (F2 or greater). See the graphic below which shows a box and whiskers graph of 0-1 km SRH for soundings associated with supercells with significant (F2 or greater) tornadoes labeled "TOR", supercells without significant tornadoes (only large hail), labeled "SUP", and nonsupercell thunderstorms (only lightning was reported near the sounding), labeled "ORD". The gray boxes denote the 25th to 75th percentiles of the data set, with the heavy horizontal bar at the median value. Vertical lines (whiskers) extend to the 10th and 90th percentiles (as in Rasmussen and Blanchard, 1998).



On the AWIPS hodograph plot, one can estimate the SRH for any storm motion using the lines of constant helicity.

Weaknesses

SRH is very sensitive to changes in the horizontal wind vector and storm motion and thus, to use it effectively in mesoscale analysis, the parameter inputs must be updated frequently by METARS, profilers, VAD winds, ACARS, or other data sources.

Many studies such as Johns et al. (1993) and Edwards and Thompson (2000) indicate a wide spectrum of SRH values associated with any single tornadic event. (See this graphic from Edwards and Thompson's study for an example of the data scatter associated with CAPE and 0-3 km SRH.)

In AWIPS model calculations of 0-1 km shear (or SRH), there are typically insufficient model layers in the vertical to adequately sample the layer. In BUFKIT, the native resolution of the model is retained.

Research has found that the signal in 0-3 km SRH for tornadic supercells is not as strong as the signal in 0-1 km SRH.

Due to differences in storm motion calculations, model derived SRH can vary.

Definitions & Computations

7. Energy Helicity Index (EHI)

The EHI (Hart and Korotky, 1991), (Davies, 1993) is used operationally for supercell and tornado forecasting. EHI is defined as

$$EHI = \frac{CAPE * SRH}{1.6x10^5},$$

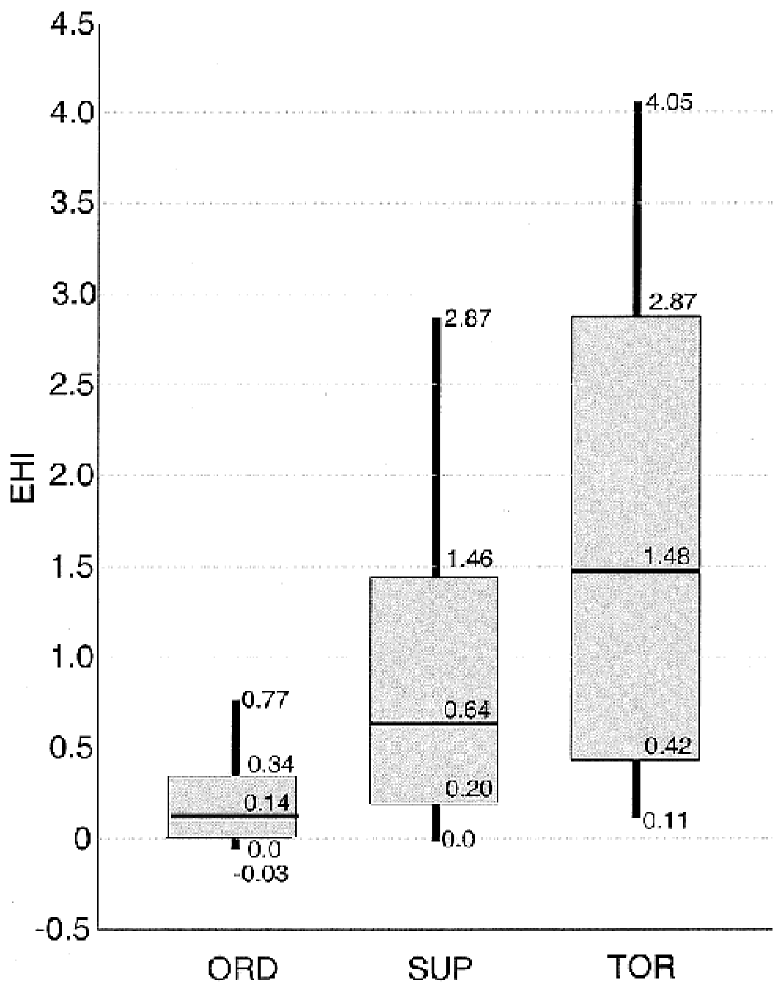
where SRH is Storm-Relative Helicity from 0 to 3 km, and CAPE is integrated positive area from the LFC to the Equilibrium Level (EL).

EHI has also been computed using a 0-1 km SRH.

Strengths

EHI is one of the effective discriminators for significant tornadoes associated with supercells (see http://members.cox.net/jondavies1/2002cases/2002cases.htm).

Increasing values of EHI from 1.0 to 3.0 and higher correspond to increasing probability of tornadic supercells. See the figure below from Rasmussen and Blanchard, (1998) which suggests that **values of EHI around 3.0 or greater indicate a higher likelihood for significant tornadoes.**



See also this AWIPS case example, a 12-hr forecast of EHI from the ETA valid 00 UTC 10 Oct 2001. In this event, the EHI accurately "bulls eyed" the area of where tornadic thunderstorms subsequently developed over west central Oklahoma and eastern Nebraska. Note, the EHI forecast fields for this event also indicated a high probability of tornadoes over portions of KS (EHI values were predicted from 1.0 to 3.0) but no tornadoes occurred in this region. Thus, forecasters can expect a relatively high false alarm rate with this single parameter.

Low-level EHI, measuring the SRH and CAPE below 3 km, may be a better predictor of significant tornadic supercells (see these cases from Davies, 2002) illustrating the usefulness of low-level EHI (and VGP).

Weaknesses

Some overlap in observed values exist between storm "classes", making **standard EHI not always a good parameter for discriminating between storms that produce tornadoes and those that do not**.

Also, as was observed in the representative example from 9 Oct, 2001, EHI values greater than 3.0 (as was forecast in KS) **do not always correlate to tornadic supercells**. High CAPE can over inflate EHI and render it not as effective. The CIN (or lack of surface-based CAPE) can also wreck the EHI forecast.

In low CAPE environments and high shear, EHI might underestimate tornado potential. Since EHI is derived from CAPE and shear, it inherits the same limitations and uncertainties from computations of those parameters.

Definitions & Computations

8. Vorticity Generation Parameter (VGP)

VGP relates the physical concept of the rate of tilting of horizontal vorticity to vertical vorticity. The equation used by Rasmussen and Blanchard (1998) is:

$$VGP = S * \sqrt{CAPE} \; ,$$

where S is the mean shear (hodograph length divided by depth over which the hodo was measured 0-4 km in their study). Mean shear is assumed to be proportional to the horizontal vorticity vector and CAPE 1/2 proportional to the vertical component of velocity. In Rasmussen and Blanchard's study (1998), the CAPE in VGP used a parcel with the virtual temperature correction and a uniformly mixed $_e$ in the lowest 1000 m AGL.

Strengths

VGP has been shown to have discriminating ability between supercells and nonsupercells (See figure 14 of Rasmussen and Blanchard, 1998).

See this graph from Davies (2002) which suggests some limits to the parameter space of low level CAPE and VGP.

Weaknesses

VGP is not as good by itself at discriminating between storms with significant tornadoes. See Rasmussen and Blanchard's (1998) figure 13)

May underestimate tornado potential in low CAPE environments.

As with EHI, since VGP is derived from CAPE and shear, it inherits the same limitations and uncertainties from computations of those parameters.

Definitions & Computations

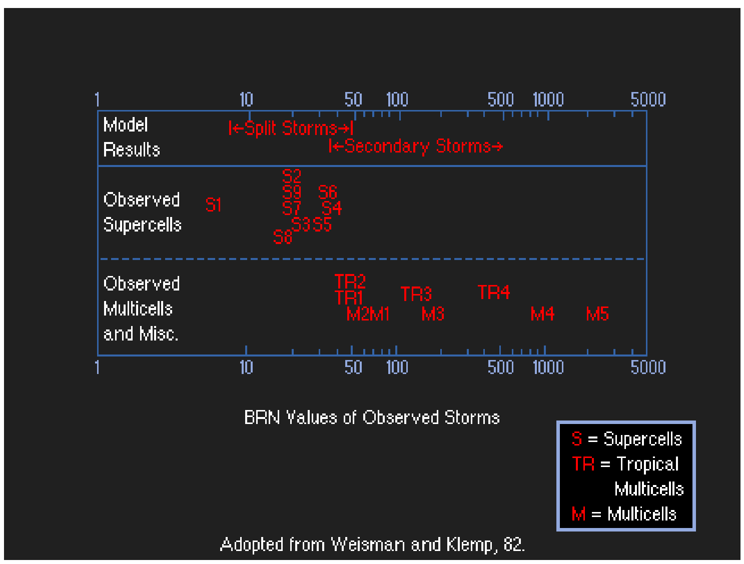
9. Bulk Richardson Number (BRN)

BRN is a rough measure of the buoyancy to shear ratio. The equation for BRN (Weisman and Klemp, 1982) is:

$$BRN = \frac{CAPE}{\frac{1}{2}\overline{U}^2},$$

where CAPE is the integrated positive area resulting from surface parcel ascent from the LFC to the EL, U is the bulk shear determined by subtracting the density-weighted mean wind vector in the lowest half-kilometer layer from the density-weighted mean wind vector in the lowest six kilometer layer.

The image below from COMET CD-ROM, "Anticipating Convective Storm Structure and Evolution" (COMET, 1996) shows various values of BRN for observed and model simulated storm types.



Weisman and Klemp (1982,1984) determined that environments with BRN less than 50 favored the development of supercells, while BRN greater than 35 favored multicells. The overlap area (BRN in the range between 35 and 50) suggested a condition where both supercells and multicells were possible at the same time.

Strengths

BRN can be used to provide an estimate of rotation potential in storms without considering storm motion. Indicates a higher likelihood of supercells when value is between 10 and 50.

The BRN value is operationally displayable on both AWIPS observed and model skew-T soundings and on BUFKIT model sounding programs.

The denominator of the BRN equation, known as BRN shear, has been shown in some studies to have the ability to indicate the likelihood of a convective storm to develop low-level mesocyclones (see Stensrud et al., 1997).

Based on mesoscale model data, Stensrud et al. (1997) found that BRN shears of 40-100 m^2s^{-2} indicated a likelihood for storms to develop low-level mesos.

Weaknesses

Operational day-to-day utility is limited due to sensitivity of CAPE value in the numerator of the BRN equation. For large values of CAPE greater than 4000 J/kg, Stensrud et al. (1997) found that BRN was large regardless of the value of the denominator, which is known as BRN shear.

Another limitation of BRN is that is does not take into account the detailed aspects of the low-level curvature, which has been shown to be significant in supercell dynamics (Weisman and Rotunno 1999). In low buoyancy environments, shear-induced pressure forces, which are related in part to the shear from low-level curvature, can be the dominant factor in controlling updraft strength. Conversely, when bulk shear is weak, low-level buoyancy (and lapse rates) can dominate updraft rotation (McCaul and Weisman, 1999).

Look at this AWIPS example from MPX on 0000 UTC 10 May 2000. The skew-T computation of BRN showed a value of only 8, but tornadoes occurred only 20 miles and 1 hour away.

Definitions & Computations

10. Temperature Lapse Rates

The unsaturated Adiabatic Lapse Rate (ALR_d) is defined by Hess (1979) as:

$$ALR_d = -\left({}^{dt}\!/_{dz}\right) = {}^{g}\!/_{C_p}$$

For $g = 9.8 \times 10^2$ cm s^{-2}, $C_p = 1.00$ J gm^{-1} K^{-1} , and $ALR_d = 9.8$ Kkm^{-1} or 9.8 C km^{-1}.

The saturated Adiabatic Lapse Rate (ALR_s) is always less than ALR_d, but approaches ALR_d as pressure increases or temperature decreases. ALR_s ranges from 3.3 C km^{-1} at 500 mb and +20 C to 9.2 C km^{-1} at 1000 mb and -30 C. (See equation 7.3 from Hess, 1979).

Note: In order to take into account the effect of water vapor on the density, one may think of ALR_d and ALR_s as lapse rates of virtual temperature.

Lapse rates are used to assess convective instability and are sometimes displayed (as in BUFKIT example below) in tabular format (note ALR_s greater than 8.0 C/km are highlighted in red in the BUFKIT table).

Lapse Rate °C/km

500	7.6	7.3	7.0	7.0	6.8	6.6	6.2	7.0	7.8
550	7.6	7.2	6.8	6.8	6.5	6.1	5.3	6.2	
600	7.9	7.4	6.9	7.0	6.7	6.1	4.4		
650	8.6	8.1	7.6	7.9	7.9	7.9			
700	8.7	8.2	7.5	7.9	7.9				
750	9.0	8.3	7.4	7.9					
800	9.4	8.6	6.8						
850	10.7	10.5							
900	11.0								
	950	900	850	800	750	700	650	600	550

Strengths

Determination of parcel static stability , and associated stability criteria (using the parcel method), can be found by comparing the observed or forecast temperature lapse rate with ALR_d (see page 13 of RTM-230).

Diagnosis of large mid-trospospheric ALR_s (such as the layer between 700 to 500 mb) have been used as an effective tool for diagnosing synoptic scale effects to potential severe storm development (see Doswell et al., 1985). Steep mid-trospospheric ALR_s in the presence of abundant low-level moisture create high values of large scale convective instability.

Several stability indices have been developed over the years which estimate low-level lapse rates (such as the layer between 850 to 500 mb). The Total Totals (TT) Index (see RTM-230 pg. 18) uses the temperature difference between 850 and 500 mb temps (Vertical Totals) in its computation. Steep lapse rates affect the ability for the environment to be able to transfer momentum.

In BUFKIT, various temperature lapse rates are output on a common table of lapse rates (see graphic above) in C/km at various model layers, with certain values color-coded to highlight the higher ALR_s.

Convective instability results from a combination of sufficient moisture at some level in the lower or middle troposphere, and a LR generally greater than the moist adiabatic lapse rate (which depends on pressure and temperature - see Hess, 1979) above the level of free convection.

Low-level lapse rates (i.e. from the surface to 3 km) have been used operationally to assess and forecast strength of low-level vertical accelerations due to diabatic heating effects.

Large sub-cloud temperature lapse rates (at ALR_d or even superadiabatic) can enhance dry microburst potential (See this WDTB pulse storm downburst web site).

Weaknesses

Assessments of environmental lapse rates by themselves are insufficient to determine parcel buoyancies. Actual parcel instability leading to deep, moist convection is primarily associated with vertical parcel displacements. Thus, the key to the possibility for growth of convective storms is the presence of CAPE, not the environmental lapse rates alone (Doswell, 2001).

Steep lapse rates may signify very dry air aloft which may actually inhibit the development of deep, moist convection in some situations.

Definitions & Computations

11. Storm-Relative Flow (SR-flow)

SR-flow is determined by subtracting the parent storm motion vector from the environmental wind vector. Vectors on a hodograph represents wind flow that the storm experiences at various levels as the storm moves through the environment (see figure below). The storm-relative flow vectors are plotted on both BUFKIT (blue lines drawn from the tip of the storm motion vector to the surface and to 3 km) and AWIPS hodograph (dotted lines) displays.

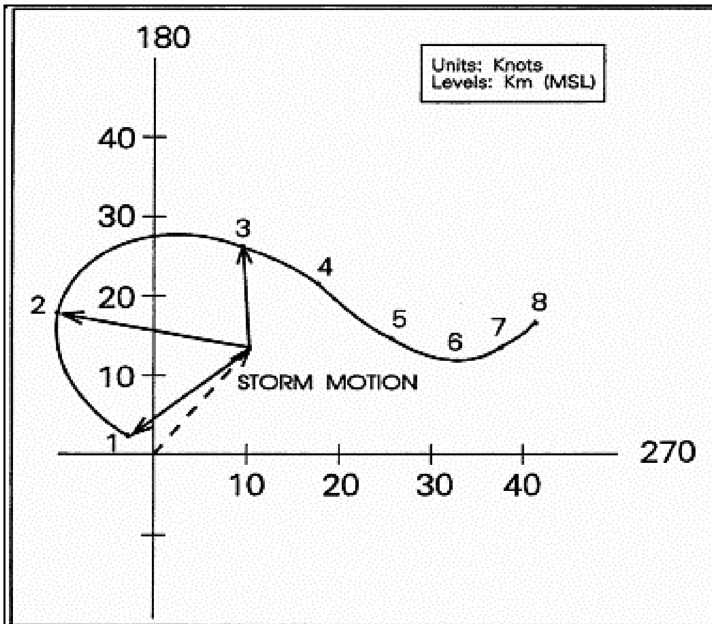


Figure 2. A hodograph showing storm relative wind "inflow" vectors and storm motion. (from NWSTC RTM-230).

SR-flow is related to precipitation distribution with a storm as increasing SR-flow carries precipitation away from the updraft summits of well-organized storms (such as supercells) thereby diminishing the potential for significant water-loading (OTB, 1993).

Strengths

SR-flow is more physically significant in producing a particular storm structure than ground relative winds. **Strong storm-relative flow can produce updraft rotation and tilting**. One can qualitatively assess the amount of SRH by looking at the amount of area swept out on hodograph by the storm-relative flow vectors.

Thompson (1998) found that supercells were more likely to produce tornadoes when midlevel (~ 500 mb) **storm-relative winds were greater than 8-10 m s^{-1}**.

Evans and Doswell (2001) found 0-2 km **system-relative flow stronger in derecho events** than in non-derecho events. This was likely due to faster forward speed and low-level convergence in derecho events.

Near-ground (0-1 km) storm-relative flow (speed) may also be crucial to tornadogenesis (Markowski et al., 2002).

SR-flow significantly influences hail growth because it determines hail trajectories across the updraft.

Weaknesses

Storm-relative flow requires an estimate of storm motion, which is often difficult to determine from observations and especially, in forecasts. It can be difficult to determine the appropriate layer in which SR-flow effects are greatest in a storm.

Multiple storm motions can occur simultaneously with storm systems making storm-relative flow estimates difficult with multicell systems.

Most of the differences in storm-relative flow between tornadic storm and non-tornadic storms reside in the lowest kilometer or so above the ground, where observations of environmental winds on a sub-mesobeta time and space scales are sparse.

Storm-relative flow was not a statistically significant tornado discriminator when RUC proximity soundings were analyzed (see Markowski et al., 2002).

Definitions & Computations

12. Equivalent Potential Temperature

Equivalent potential temperature ($_e$) is a thermodynamic variable related to temperature, moisture and the pseudo-adiabatic process of parcels. It can be used to assess potential convective instability (where $_e$ decreases with height). The mathematical expression for $_e$ is

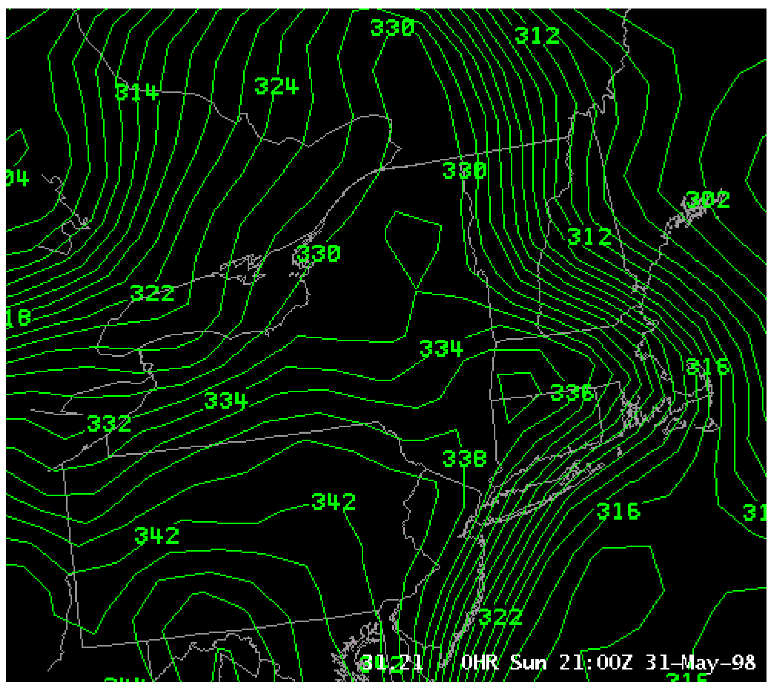
$$q_e = q * e^{\left(Lw_s / C_p T \right)}$$

where   is the potential temperature, L is the latent heat of vaporization (2.5 x 10^6 J kg^{-1} at 0 C), w_s is the saturation mixing ration with respect to water, C_p is the specific heat of water vapor (1005 J K^{-1} kg^{-1}) at constant pressure, and T is the temperature (from Basic Convection 1, OSF/OTB 1991).

See Figure 3 pg. 50 from RTM-230 for a graphical depiction of how to determine equivalent potential temperature on a Skew-T.

The AWIPS Interactive Skew-T can be used to display wet-bulb potential temperature ($_w$), which is approximately equal to $_e$ minus the quantity wL/C_p, where w is the mixing ratio, L is the latent heat, and C_p is the specific heat of water vapor at constant vapor at constant pressure.

Model objective analyses (plan view) of $_e$ are displayed on AWIPS D2D using various models (RUC, ETA, MesoETA, LAPS, MSAS, LAMP, etc.) See graphic below showing an example of surface $_e$ from the MAPS Surface Analysis System (MSAS). (See this web site for more information on MSAS).



An example of surface $_e$ analysis (1900 UTC from 31 May 1998) from the MAPS Surface Analysis System (MSAS). Lines of constant $_e$ (K) are shown.

Another excellent way to display $_e$ so that one can determine potential convective instability is via vertical cross sections and time sections. AWIPS provides nice displays of both time sections and cross sections of $_e$.

In BUFKIT, one can view model vertical time sections of $_e$ (see this example).

Note: There are some differences in how AWIPS and GEMPAK compute $_e$ (see the AWIPS validation web site).

Strengths

The $_e$ of parcels is conservative with respect to dry and moist processes, so it is a useful diagnostic tracer of air trajectories. $_e$ is very sensitive to increases in water vapor content so layers on a sounding where $_e$ (or $_w$) decreases with height are said to be convectively unstable. Convective instability is a relevant parameter in diagnosis of severe weather potential.

Axes of high low-level $_e$ air (ridges) can often be used to assess convective potential of the environment. Cross-sections of $_e$ can be helpful in diagnosing elevated instability (Grant, 1995).

The difference of $_e$ values from a surface maximum to midlevel minimum has been used to estimate downdraft potential in organized convection. Also, max $_e$ differences of 20-30 K (from surface to midlevels ~400 to 700 mb layer) have been correlated to strong downdraft potential in moist microbursts. (See Atkins and Wakimoto, 1991).

Also, see this WDTB Pulse Storm downburst web site).

Weaknesses

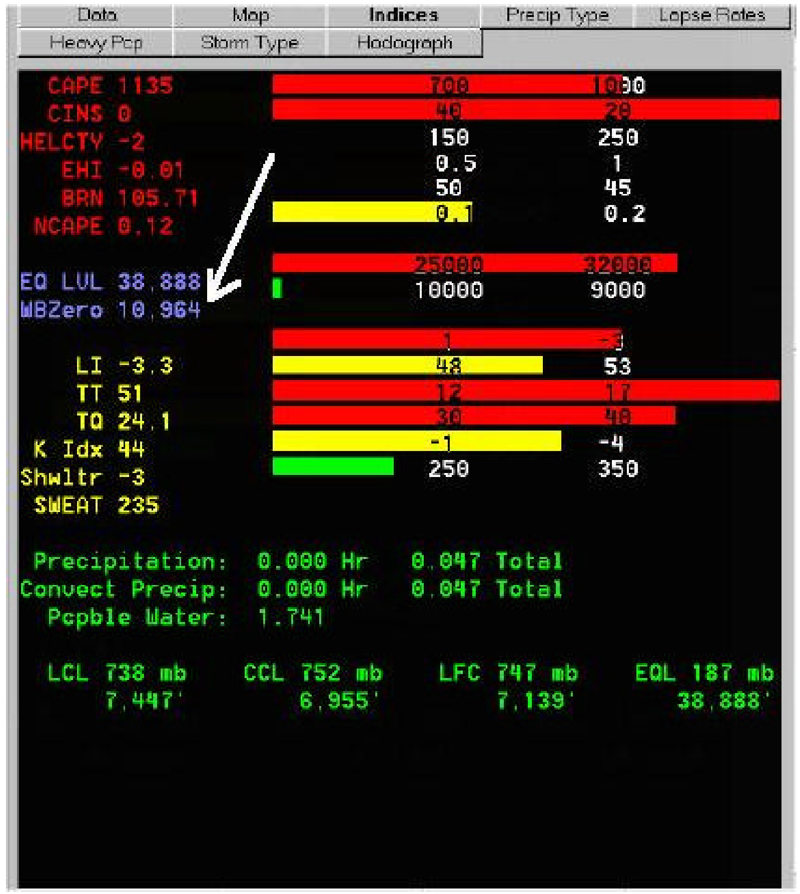
Potential convective instability, as evidenced by decreasing values of $_e$ with height, does not by itself result in deep, moist convection. Just as with steep lapse rates, the parcel has to be lifted (by differential advection, front, etc.) for convection to result.

Another limitation (as with most other parameters) is the difficulty in choosing the right parcel to be lifted.

Definitions & Computations

13. Wet-bulb Zero (WBZ) Height

Usually labeled as WBZ, the Wet-bulb zero is the height at which the wet-bulb temperature is zero. This approximates both the height at which falling hail begins to melt and the height at which the downdraft begins (OSF/OTB, 1993). On BUFKIT, the WBZ height is shown on the Indices screen (in ft AGL) and is also plotted (optionally) on the Skew-T display in red (see figure below).



In AWIPS Skew-T, the WBZ height is displayed on the parameter output in ft Above Sounding Level (ASL).

Strengths

In general, WBZ heights between 7000 ft and 10,500 ft AGL are associated with a potential for large hail at the surface. Higher WBZ heights imply mid- and upper-level stability and imply a large melting zone for falling hail. On the other hand, lower WBZ heights suggest that the lower levels of the atmosphere are too cool and stable to support intense convection.

Weaknesses

WBZ values are only general guidelines for hail potential. Should also analyze the CAPE and CIN in a proximity sounding for updraft potential.

WBZ only partially predicts severe hail potential because it doesn't consider updraft strength or parcel trajectories. Since hailstone growth is related to the residence time a potential hailstone covers across the growth region of a storm, broad, moderate updrafts combined with strong midlevel storm-relative flow (and weaker low-level shear), and higher cloud bases are more likely to produce significant hail fall than storms having strong, but very compact updrafts, and weak midlevel SR-flow. That is why storms with deep mesocyclones often produce hail with high VIL (Vertically Integrated Liquid) values whereas storms with the same VILs in weak SR-flow aloft situations and no mesocyclones do not produce hail at the ground.

In addition, since hail volume increases relative to surface area by a factor of the radius r, large hail can be maintained despite high WBZ heights such as in the presence of a Bounded Weak Echo Region (BWER).