Convection Parameters

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1 Introduction

There are many parameters used for the identification of weather situations which are likely to produce thunderstorms with severe weather. They all follow simplified conceptual models of the conditions that cause convection. Thresholds have been defined for most of the parameters to transform them to different levels of warnings. However, false alarm rates are generally high and probability of detection can be low.

2 Parameters

Here is a commented list of some convection parameters. The author is pretty sure that though most of the widely used parameters are included, many may not. The list only reflects my current knowledge. Most of the parameters are based on a single aspect linked to convection. However, some of the more modern ones combine those parameters very successfully. I group them into 5 groups, starting with the simplest temperature-only based indices, those which include humidity and then those related to wind. The fourth and the fifth group contain some advanced parameters and finally thematic ones related to specific perils like hail, downdrafts and lightening.

2.1 Simple temperature-based Indices

1. Lapse Rate - γ
   γ (the negative vertical temperature gradient $-\frac{dT}{dz}$) provides information about the stability of an atmospheric layer. An ascending air parcel with a cooling rate lower than γ eventually becomes warmer than its environment and accelerates. The faster the temperature decreases with height the more unstable is the atmosphere. The following stability criteria apply:

$$\begin{array}{rcrcll} & & \gamma & \geq & 9.8\frac{^\circ C}{km} & : \text{absolute unstable atmosphere} \\ & & \gamma & = & 9.8\frac{^\circ C}{km} & : \text{dry adiabatic lapse rate} \\ 9.8\frac{^\circ C}{km} & > & \gamma & \leq & 6\frac{^\circ C}{km} & : \text{conditionally unstable} \\ 6\frac{^\circ C}{km} & \geq & \gamma & \geq & 5.5\frac{^\circ C}{km} & : \text{moist adiabatic} \\ 5.5\frac{^\circ C}{km} & > & \gamma & & & : \text{stable.} \end{array} \tag{1}$$

   In case of conditional instability, values of $CAPE$ can become positive and values of LI can become negative, meaning that thunderstorms can develop. Unstable conditions cannot last for long.

Sometimes the largest cooling rate of an atmospheric column is used to evaluate its stability. Furthermore the lapse rate of the lowest 500m above ground γ_{0-500} is used to estimate the stability of the boundary layer. A threshold of 10 to 11 K/km characterizes unstable conditions in this case. However, the surface air temperature is a very uncertain variable in models and so values of γ_{0-500} can become very unrealistic. Another variable is the lapse rate directly above the boundary layer, i.e. $\gamma_{2000-4000m}$. A temperature decrease of 8K/km is assumed to be a threshold separating stable and unstable conditions.

2. Vertical Totals - VT
The vertical totals index (see Miller, 1972) is the temperature difference between the 850hPa level and the 500hPa level

$$VT = T_{850} - T_{500}. \tag{2}$$

It is therefore a measure of the average vertical temperature gradient of a layer of the atmosphere starting approximately at the top of the atmospheric boundary layer and extending through half of the air mass. The stronger the vertical temperature gradient the more likely are thunderstorms. Usually a threshold of $26K$ is assumed to best separate thunderstorm prone weather from weather that cannot produce thunderstorms. $VT = 26K$ equals approximately a vertical temperature gradient of 0.65K/100m, i.e. the average observed lapse rate.

3. Boyden Index - $BoydI$
The Boyden Index (Boyden, 1963) was originally designed to assess the thunderstorm risk at frontal passages. Explicitly it takes only the temperature in 700 hPa into account. However, the height difference of the 1000 and 700 hPa level, i.e. the thickness of this layer is proportional to its mean temperature as well. The index is defined as

$$BoydI = 0.1(z_{700} - z_{1000}) - T700. - 200 \tag{3}$$

and should not be used outside frontal passages e.g. for air mass thunderstorms.

4. CAP
CAP is a stable region in the lower troposphere that impedes convection from the boundary layer. The formal definition is the maximum temperature difference of a lifted parcel and its environment. CAP exists mainly in morning soundings during summer season. CAP can only be used to tell something about the resistance against convection of low level air parcels. It is related to CIN. The following thresholds apply:

$$\begin{aligned} CAP \leq 0K: &\quad \text{no } CAP \\ 0K < CAP \leq 2K: &\quad \text{weak } CAP \\ 2K < CAP \leq 4K: &\quad \text{moderate } CAP \\ 4K < CAP: &\quad \text{strong } CAP. \end{aligned} \tag{4}$$

If there is a weak CAP and sufficient $CAPE$ storms are likely to occur soon. If there is a strong CAP strong forcing is necessary to overcome it and start convection even if there is sufficient $CAPE$.

2.2 Temperature and Humidity based Indices

1. Cross Totals - CT
The cross totals index (Miller, 1972) overcomes the disadvantage of VT not to include humidity. Thunderstorms cannot develop if there is not enough humidity even in case of strong vertical temperature gradients. Thus, the cross totals index is defined as the difference of the dew-point temperature in $850hPa$ and the temperature in $500hPa$
$$CT = T_{d,850} - T_{500}. \tag{5}$$
It therefore increases with increasing humidity in lower levels of the atmosphere. The chance for thunderstorms to occur increases if CT is above $20K$ and heavy showers and tornadoes are supposed to be likely if CT is above $29K$.

2. Total Totals TT
The totals totals index (Miller, 1972) is the sum of VT and CT. It therefore increases with increasing humidity in the lower levels of the atmosphere and increasing vertical temperature gradients
$$TT = VT + CT = T_{850} - T_{500} + T_{d,850} - T_{500} = T_{850} + T_{d,850} - 2T_{500}. \tag{6}$$
The following three thresholds are usually applied:
$$TT \begin{cases} \geq 44K & \text{thunderstorms possible} \\ \geq 50K & \text{severe thunderstorms possible} \\ \geq 55K & \text{severe thunderstorms likely.} \end{cases} \tag{7}$$

3. Modified Total Totals - TT_m
The modified TT uses the average of the surface ($2m$), the $925hPa$ and the $850hPa$ observations instead of the $850hPa$ observations alone in order to better describe the state of the boundary layer.
$$TT_m = (T_{2m} + T_{925} + T_{850})/3 + (T_{d,2m} + T_{d,925} + T_{d,850})/3 - 2T_{500}. \tag{8}$$
The threshold for thunderstorms to occur is usually set to $57K$.

4. K Index - K
The K index is developed by George (1960). Like the VT it is based on the vertical temperature gradient between $850hPa$ and $500hPa$. Higher humidity in $850hPa$, expressed by higher $T_{d,850}$ increases K. Furthermore, lower humidity in higher levels, expressed by the dew-point depression in $700hPa$ decreases the chance of thunderstorms to occur resulting in
$$K = T_{850} - T_{500} + T_{d,850} - (T_{700} - T_{d,700}). \tag{9}$$
When all temperatures are provided in $°C$ the following thresholds are usually applied:
$$K \begin{cases} 0K \text{ to } 20K & 0\% \text{ to } 20\% \text{ thunderstorm probability} \\ 21K \text{ to } 25K & 20\% \text{ to } 40\% \text{ thunderstorm probability} \\ 26K \text{ to } 30K & 40\% \text{ to } 60\% \text{ thunderstorm probability} \\ 31K \text{ to } 35K & 60\% \text{ to } 80\% \text{ thunderstorm probability} \\ 36K \text{ to } 40K & 80\% \text{ to } 90\% \text{ thunderstorm probability} \\ > 40K & 90\% \text{ to nearly } 100\% \text{ thunderstorm probability.} \end{cases} \tag{10}$$

5. Modified K Index - K_{mod}
As an improvement of the K index the $850hPa$ temperatures are replaced by the $\ln p$ weighted temperatures of the surface ($2m$) and $850hPa$ by Charba (1977)

$$K = \overline{T} - T_{500} + \overline{T_d} - (T_{700} - T_{d,700}). \tag{11}$$

This should be an improvement since now the K index contains also information of the layer below $850hPa$.

6. Humidity Index HI
It is defined by Litynska et al. (1976) closely to the K index as

$$HI = (T_{850} - T_{d,850}) + (T_{700} - T_{d,700}) + (T_{500} - T_{d,500}) \tag{12}$$

A value of 30K is suggested as a threshold for thunderstorms with thunderstorms more likely for smaller values.

7. S Index - SI
SI uses the same variables as K but in different proportions. It is defined as

$$SI = TT - (T_{700} - T_{d,700}) - \begin{cases} 0 & \text{if} \quad & VT & > & 25 \\ 2 & \text{if} \quad 25 & \geq & VT & \geq & 22 \\ 6 & \text{if} \quad 22 & > & VT. \end{cases} \tag{13}$$

SI is introduced by the German Military Geophysical Office (Reymann et al., 1998) and is considered useful from April to September. It is very similar to the K index. In fact it can also be written as $K - T_{500} - \zeta$ where ζ is the third term on the right hand side of eq. (13) which penalizes in case of low values of VT, i.e. low vertical temperature gradients.

8. Rackliff Index - RaI
RaI compares the potential wet bulb temperature at $900hPa$ to the dry bulb potential temperature at $500hPa$

$$RaI = \Theta_{w,900} - \Theta_{500}. \tag{14}$$

Therefore it provides information about the latent instability of an air mass at $900hPa$, a non-standard height.

9. Modified Jefferson Index $JEFF$
As the K also $JEFF$ (Jefferson, 1963a,b, 1966) uses the dew-point temperature at $850hPa$ and the dew-point depression at $700hPa$. However, instead of the temperature difference between the $850hPa$ and the $500hPa$ level it uses the potential wet-bulb temperature at $850hPa$

$$JEFF = 1.6\Theta_{w,850} + T_{d,850} - -0.5(T_{700} - T_{d,700}) - 8. \tag{15}$$

This Index is a further development of the Rackliff Index. It uses standard altitudes and is less temperature sensitive. This index is usually applied to air mass thunderstorms without any dynamical trigger. A usual threshold for $JEFF$ is $29°C$. In polar air masses a somewhat smaller threshold of $27°C$ applies.

10. Adedokun Index - AI
AI (defined by Adedokun, 1981, 1982) is the difference of wet bulb temperature at $850hPa$ (or alternatively $2m$) and the saturated wet bulb temperature at $500hPa$

$$AI = \Theta_{w,850} - \Theta_{ws,500}. \tag{16}$$

11. Bradbury Index - $BradI$
This index defined by Bradbury (1977) is also called Potential Wet-Bulb Index since it assesses potential instability by wet-bulb temperatures

$$BradI = \Theta_{w,500} - \Theta_{w,850}. \tag{17}$$

This index is also called Pickup Index after Pickup (1982). The lower the index the stronger is potential instability. This is why this index is sometimes called potential instability index, which should not be confused with the PII introduced by Van Delden (2001). A threshold of -2K in summer and 3K in other seasons is sometimes suggested.

12. Potential Instability Index - PII
This index is defined by Van Delden (2001) and measures potential instability of the atmospheric layer between 925 and $500hPa$

$$PII = \frac{\Theta_{e,925} - \Theta_{e,500}.}{z_{500} - z_{950}} \tag{18}$$

PII is a discretized version of the more generic Convective Potential.

13. Ko Index - Ko
Ko (Andersson et al., 1989) describes the potential instability between lower and higher levels of the atmosphere. It is thus based on the equivalent potential temperatures Θ_e as

$$Ko = 0.5(\Theta_{e,500} + \Theta_{e,700}) - 0.5(\Theta_{e,850} + \Theta_{e,1000}). \tag{19}$$

It is smallest if dry and cold air lies above warm and humid air. Usual thresholds are

$$\begin{array}{rl} Ko > 5K & \text{no thunderstorms possible} \\ 5K \geq Ko > 3K & \text{thunderstorms possible} \\ 2K \geq Ko & \text{severe thunderstorms possible.} \end{array} \tag{20}$$

14. Convective Instability Index - CI
CI is an estimate of potential instability of the lower to medium troposphere based on Θ_e. It compares Θ_e of the surface and near-surface layer to the one in $500hPa$

$$CI = 0.5 * (\Theta_{e,2m} + \Theta_{e,925}) - \Theta_{e,500}. \tag{21}$$

A usual threshold is 5°.

15. Lifted Index - LI
LI is a measure of stability of the atmosphere between $z = 0$ and $500hPa$. Defined by Galway (1956) it is the temperature difference of a parcel that is lifted from the surface to its LCL dry adiabatically and further pseudo adiabatically to $500hPa$.

Negative LI reflect instability. The higher the absolute value of the negative LI the more unstable is the atmosphere. LI is usually very good correlated to $SBCAPE$. However, LI has the advantage not to depend on the hight of EL which is usually a bad estimate in model output.The following thresholds are often applied:

$$\begin{array}{rcccrcl} & & LI & \geq & 0K & : & \text{stable atmosphere - no thunderstorms possible} \\ 0K & > & LI & \geq & -2K & : & \text{thunderstorms possible} \\ -2K & > & LI & \geq & -6K & : & \text{thunderstorms likely} \\ -6K & > & LI & & & : & \text{severe thunderstorms likely.} \end{array} \tag{22}$$

16. Showalter Index - $ShowI$
$ShowI$ is defined by Showalter (1947) alike LI but for an air parcel lifted from $850hPa$ rather than the surface. Thresholds are

$$\begin{array}{rcccrcl} -3K & < & ShowI & \leq & 3K & : & \text{thunderstorms possible} \\ & & ShowI & \leq & -3K & : & \text{severe thunderstorms possible.} \end{array} \tag{23}$$

17. Deep Convective Index - DCI
The DCI (Barlow, 1993) is a combination of temperature and humidity in $850hPa$ as well as the LI

$$DCI = T_{850} + T_{d,850} - LI. \tag{24}$$

Thunderstorms are supposed to be possible if DCI is higher than $30°C$.

18. Thomson Index - $ThomI$
The Thomson Index is based on the K index and a version of the lifted index for the case where not a surface-based air parcel is used but the average over the lowest $50hPa$ above ground

$$ThomI = K - LI_{50hPa}. \tag{25}$$

It should be an improvement of K since it contains information about the layer below $850hPa$.

19. Theta-E Index TEI
This index provides information about elevated convection. It is the value of the steepest negative gradient of the equivalent potential temperature Θ_e, i.e. the largest cooling and drying with height. If an air parcel is lifted in a layer with a steep cooling and drying it may become warmer than its environment leading to elevated convection. The condition for this, however, is lifting at the lower boundary of this layer. Thresholds are

$$\begin{array}{rl} TEI < 5 & \text{not favorable} \\ 5 \leq TEI < 9 & \text{potential} \\ 9 \leq TEI & \text{very high potential.} \end{array} \tag{26}$$

20. Yonetani Index - $YonI$
This index is defined by Yonetani and meant to combine conditional instability

with low-level moisture in order to detect air mass thunderstorms. Yonetani (1979) defined it as

$$YonI = \begin{cases} 0.966\Gamma_L + 2.41(\Gamma_U - \Gamma_W) + 9.66RH - 15 & \text{if} \quad RH > 0.57 \\ 0.966\Gamma_L + 2.41(\Gamma_U - \Gamma_W) + 9.66RH - 16.5 & \text{if} \quad RH \leq 0.57 \end{cases} \tag{27}$$

and modified it later (Yonetani, 1990) to

$$YonI_{mod} = \begin{cases} 0.964\Gamma_L + 2.46(\Gamma_U - \Gamma_W) + 9.64RH - 13 & \text{if} \quad RH > 0.57 \\ 0.964\Gamma_L + 2.46(\Gamma_U - \Gamma_W) + 9.64RH - 14.5 & \text{if} \quad RH \leq 0.57. \end{cases} \tag{28}$$

The variables are

$$\begin{aligned} \Gamma_L &= \text{lapse rate between 900 and } 850hPa \\ \Gamma_U &= \text{lapse rate between 850 and } 500hPa \\ \Gamma_W &= \text{pseudoadiabatic lapse rate at } 850hPa \\ RH &= \text{Relative Humidity in } [0,1]. \end{aligned} \tag{29}$$

2.3 Wind-related Parameters

1. Vertical Shear - VS

   Vertical shear is the change of wind with height. This change is a vector. Usually, however, only the magnitude is of interest. Many different measures of shear in the atmosphere exist. Some of them are discussed here.

   Higher values of vertical shear usually lead to more organized and persistent thunderstorms.

   (a) Boundary Layer to 6 km Vertical Shear - VS_{BL-6km}
   Supercells are usually associated with vertical shear values of 65 to $75\frac{km}{h}$.

   (b) Effective Bulk Vertical Shear - VS_{eff}
   This is the VS within a storm and thus the difference of the wind speed at the LPL and EL. Sometimes only the lowest 40 to 60% of this range is used. Values of 45 to $75\frac{km}{h}$ and above increase the likelihood of supercells.

   (c) Surface-1-km Vertical Shear - VS_1
   SV_1 magnitudes higher than $30\frac{km}{h}$ tend to favor supercell tornadoes.

   (d) Surface-3-km Vertical Shear - VS_3
   This shear is often used to get a criterion for longevity of convection. The larger the shear the longer lasting is the convection. Strong shear also supports high values of helicity. Thresholds are

$$\begin{array}{rl} 0-3 & \text{weak} \\ 4-5 & \text{moderate} \\ 6-8 & \text{large} \\ >8 & \text{severe.} \end{array} \tag{30}$$

(e) Deep Layer Shear - DLS

Deep layer shear is defined as the vertical wind shear between the ground level ($10m$) and $6km$ (usually density weighted in order to approximate mass flux). The following thresholds apply

$$\begin{array}{rcccll} & & DLS & > & 40knots & : \text{ if storms develop then supercells} \\ & & & & & \quad \text{are likely} \\ 40knots & \geq & DLS & \geq & 30knots & : \text{ supercells possible if environment} \\ & & & & & \quad \text{is very unstable} \\ 20knots & \geq & DLS & \geq & 15knots & : \text{ minimum for organized convection} \\ & & & & & \quad \text{with mid level winds of 25 knots.} \end{array} \tag{31}$$

2. Storm Relative Winds - SRW

Storm relative winds are wind speeds and directions relative to the movement of the storm. The movement of storms can be deducted from radar observations while the average environment wind speeds follow from radiosondes or model output. Three kinds of SRW are usually used to a) identify sustained supercells, b) distinguish between tornadic and non-tornadic supercells and c) to classify supercells into high precipitation, classical and low precipitation.

(a) Surface-2km Storm Relative winds - SRW_2

SRW_2 is meant to represent low-level storm inflow. The majority of sustained supercells have values of at least $30\frac{km}{h}$.

(b) 4-6-km Storm Relative Winds - SRW_{4-6}

SRW_{4-6} is used to distinguish between tornadic and non-tornadic supercells. Tornadic supercells have values larger than $30\frac{km}{h}$.

(c) Anvil Level/9-11-km Storm Relative Winds - SRW_A

SRW_A is based on the storm relative wind speeds at anvil height, usually approximated as 9 to $11km$ They are used to discriminate the precipitation from a supercell by

$$\begin{array}{rcccll} & & SR_A & \leq & 40\frac{km}{h} & : \text{ high precipitation supercell} \\ 40\frac{km}{h} & < & SR_A & \leq & 110\frac{km}{h} & : \text{ classic supercell} \\ 110\frac{km}{h} & < & SR_A & \geq & & : \text{ low precipitation supercell.} \end{array} \tag{32}$$

3. Helicity

Helicity is the extent to which corkscrew-like motion occurs. It is generally defined as the following integral over the volume V

$$H = \int \vec{u} \cdot (\nabla \times \vec{u}) dV. \tag{33}$$

H is a conservative variable for an incompressible Euler (viscosity free) fluid.

Here we look at the vertical integral of the horizontal wind $\vec{v_h}$ and vorticity $\vec{\zeta_h}$ only. In this case helicity is the vertical integral of the transfer of vorticity from the environment into a rising air parcel in convective motion. It is therefore defined as

$$H = \int \vec{v_h} \cdot \vec{\zeta_h} \, dz = \int \vec{v_h} \cdot \nabla \times \vec{v_h} \, dz. \tag{34}$$

H is zero if the horizontal wind direction does not change with z, since in this case $\vec{v_h}$ and $\nabla \times \vec{v_h}$ are perpendicular and the scalar product vanishes. H is positive if the wind veers (changes clockwise with z) and negative if it backs. The units of H are $m^2/s^2 = J/kg$ and thus H is an energy density.

4. Storm Relative Helicity - SRH
   SRH provides a measure for the wind change with height including its magnitude and direction in relation to the storm movement. It is introduced by Davies-Jones et al. in 1990. Usually the lowest $3km$ of the atmosphere are used. It is defined as

$$SRH = -\int_0^h \vec{k} \cdot (\vec{v}(z) - \vec{c}) \times \frac{\partial \vec{v}(z)}{\partial z} dz \tag{35}$$

with

$$\begin{aligned} \vec{k} &= \text{vertical unit vector} \\ \vec{c} &= \text{moving velocity of cell} \\ \vec{v}(z) &= \text{wind vector.} \end{aligned} \tag{36}$$

SRH is not uniquely calculated since there exist different opportunities to estimate $\vec{c}$ from observations. One way to calculate SRH (Markowski and Richardson, 2010) is

$$SRH = \sum_{n=1}^{N-1} [(u_{n+1} - c_x)(v_n - c_y) - (u_n - c_x)(v_{n+1} - c_y)] \tag{37}$$

with the wind speed components u and v, the components of the cell moving velocity c_x and c_y and the height levels $n = 1$ to N.

Usually one might say that

$$\begin{array}{rclcl} & SRH & < & 150\frac{m^2}{s^2} & : \text{minor rotation} \\ 150\frac{m^2}{s^2} \le & SRH & < & 300\frac{m^2}{s^2} & : \text{thunderstorms with rotation} \\ 300\frac{m^2}{s^2} \le & SRH & < & 450\frac{m^2}{s^2} & : \text{supercells with rotation possible} \\ 450\frac{m^2}{s^2} \le & SRH & & & : \text{supercells with tornadoes possible.} \end{array} \tag{38}$$

SRH does not contain any information about the energy of convection. This is why the energy helicity index EHI is defined which combines both, SRH and $CAPE$.

5. Storm Relative Directional Shear $SRDS$
   $SRDS$ is a difference of differences. Storm motion has to be estimated first. Then the difference between the windspeed at the bottom of convection minus the storm motion and at 3km altitude minus the storm motion is built. The following thresholds apply:

$$\begin{array}{rl} SRDS \le 30 : & \text{weak} \\ 30 < SRDS \le 60 : & \text{some} \\ 60 < SRDS \le 90 : & \text{moderate} \\ 90 \le SRDS : & \text{strong} \end{array} \tag{39}$$

Strong $SRDS$ is important to produce a favorable environment for tornadoes.

6. Vorticity Generating Parameter VGP
The vorticity generating parameter is introduced by Rassmussen et al. in 1998 . It is meant as a proxy for the tilting of horizontal vorticity. It is defined as the product of the square root of $CAPE$ times the 3-km wind shear, thus

$$VGP = \sqrt{CAPE}\ VS_3. \tag{40}$$

Supercell tornadoes are rare if $VGP < 0.3$ and likely if it is ≤ 0.6.

7. Storm Motion STM
This is the expected speed (vector) with which storms move. Its magnitude is usually estimated as a fraction (e.g. 75%) of the wind speed average of the lowest $6km$ of the atmosphere. Storms move slower than the environmental wind since they have higher mass due to the liquid/frozen water content. Vertical wind shear increases STM. The direction in which the storm moves is usually about 30° to the right of the average wind speed, i.e. the storm veers. This method is sometimes called 75%30° rule. Other values like 80% and 25° are also common. However, this approach misses any mesoscale dynamics which can lead to large deviations from reality.

The Bunkers' method (Bunkers et al., 2000) uses the vector which is 7.5m/s to the right of the bulk shear vector between the 0-500m average wind and the 0-6000m average wind.

This parameter can be used to estimate also the direction in which tornadoes move. It is also important in order to calculate the storm relative helicity.

8. Hodograph Length HL

Drawing wind vectors of different altitudes all from the same origin and connecting all tips of the arrows creates a hodograph as a visualization of direction and magnitude of wind shear with altitude. the length of the hodograph can be used as a combined index of wind shear direction and magnitude. It does, however, not include information of whether the wind is veering or backing with height. $HL3$ is the hodograph length created from wind vectors between 0 and 3 km height while $HL6$ is build from wind vectors between 0 and 6 km height.

2.4 Advanced Parameters

1. Energy Index - EI
Defined by Darkow (1968) EI is the difference of the non-kinetic energy in $500hPa$ and a certain lower level, usually $850hPa$ or $925hPa$

$$EI = h_{500} - h_{850}. \tag{41}$$

The non-kinetic energy is sometimes also called static energy. For an atmosphere which does not contain liquid water or ice it is given by

$$h = c_pT + gz + L_vr \tag{42}$$

with

$$\begin{array}{rcl} c_p &=& \text{specific heat of air for constant pressure} \\ T &=& \text{absolute temperature in } K \\ g &=& \text{earth acceleration} \\ z &=& \text{geopotential height} \\ L_v &=& \text{latent heat of evaporation} \\ r &=& \text{mixing ratio of water vapor.} \end{array} \tag{43}$$

Some terms should be altered if there is liquid water and/or ice. The EI is positive if warm and humid air lays above cold and dry air which is a stable situation. The following thresholds are applied

$$\begin{array}{rl} EI \geq 0 & \text{No activity expected} \\ 0 > EI \geq -2 & \text{isolated severe thunderstorms possible} \\ -2 > EI & \text{severe thunderstorms probable, tornadoes possible.} \end{array} \tag{44}$$

Note that the EI only provides a potential for thunderstorms. In case there is a strong CAP which cannot be broken through, nothing might happen.

2. Lifted Condensation Level - LCL
   The LCL is the height in which a parcel lifted from the surface becomes saturated. It is therefore a good approximation of the cloud base height in case of forced ascend. LCL is given by

$$LCL = \frac{T_{2m} - T_{d,2m}}{\frac{g}{c_p} - \frac{gT_{d,2m}^2}{\varepsilon L_v T_{2m}}} \tag{45}$$

   with the dew-point temperature T_d. In this form LCL is not the absolute height of the condensation level but the height above ground.

3. Mixed Layer Lifted Condensation Level - LCL_{ML}
   It is defined as the LCL but not for lifting from the surface but from the mixed layer (usually approximated by the average over the lowest $100hPa$).

4. Level of Free Convection - LFC
   The LFC is the height in which a parcel lifted from the surface becomes lighter than the surrounding air and starts free ascent. In order to calculate it, a parcel has to be lifted dry-adiabatically from the ground to the LCL and pseudo-adiabatically upwards from the LCL until its density is lower than the density of the surrounding air.

   The lower the LFC the more likely convection happens. Experience shows that tornadoes seems to be more likely for supercells with LFC less than $2000m$ above ground. And thunderstorms are more likely to be initiated and maintained if LFC is below $3000m$.

5. Equilibrium Level - EL (Level of Neutral Buoyancy LNB)
   An air parcel rising adiabatically from the ground beyond its LFC becomes lighter than the surrounding air and accelerates its ascent. The higher it rises, however, the weaker is the acceleration and at the EL it vanishes. Inertia lets the air parcel rise above the EL, however, with negative buoyancy. EL is primarily used to estimate the height of the anvil of deep convection.

6. $\Delta L_1 = LFC - LCL$
The difference between LFC and LCL contains information about the probability of deep convection. If LFC is much higher than LCL chances are good that convection stays shallow. If on the other hand the difference between LFC and LCL is small sudden deep convection can occur.

7. $\Delta L_2 = EL - LFC$
The difference between EL and LFC provides a measure of deepness of convection if convection occurs.

8. Convective Available Potential Energy - $CAPE$
Different types of $CAPE$ are defined for different purposes. They all have in common that they are a measure of energy available for convection in case of no entrainment and thus are the integral over buoyancy. The major variables therefore are the virtual potential temperatures of an air parcel $\Theta_{v,p}$ and of the environmental air $\Theta_{v,e}$. Usually the following thresholds apply for $CAPE$

$$\begin{array}{rcccrcl} 0\frac{J}{kg} & < & CAPE & \leq & 1000\frac{J}{kg} & : & \text{weak instability} \\ 1000\frac{J}{kg} & < & CAPE & \leq & 2500\frac{J}{kg} & : & \text{moderate instability} \\ 2500\frac{J}{kg} & < & CAPE & \leq & 4000\frac{J}{kg} & : & \text{strong instability} \\ 4000\frac{J}{kg} & < & CAPE & & & : & \text{extreme instability.} \end{array} \tag{46}$$

   (a) Surface Based $CAPE$ - $SBCAPE$
   $SBCAPE$ is defined as the integral

$$SBCAPE = g \int_{LFC}^{EL} \frac{\Theta_{v,p} - \Theta_{v,e}}{\Theta_{v,e}} dz. \tag{47}$$

   It is the energy that is available for convection of an air parcel that is lifted from the surface to its LFC.

   (b) Normalized $CAPE$ - $NCAPE$
   $NCAPE$ is an energy density of a vertical column. Since the energy release described by $SBCAPE$ happens between the LFC and EL, $NCAPE$ is defined as

$$NCAPE = \frac{SBCAPE}{EL - LFC}. \tag{48}$$

   $NCAPE$ is therefore a measure of acceleration (mass specific energy divided by length). High values of $NCAPE$ represent fast development.

   (c) Mixed Layer $CAPE$ - $MLCAPE$
   $MLCAPE$ uses not only parcels rising from ground level but from the whole mixed layer which is usually approximated as the lowest approximately $100mb$ layer. Different approximations for $MLCAPE$ are used. Usually the

following thresholds apply

$$\begin{array}{rclcl} 0\frac{J}{kg} & < & MLCAPE & \leq & 1000\frac{J}{kg} & : & \text{weak instability} \\ 1000\frac{J}{kg} & < & MLCAPE & \leq & 2500\frac{J}{kg} & : & \text{moderate instability} \\ 2500\frac{J}{kg} & < & MLCAPE & \leq & 3500\frac{J}{kg} & : & \text{strong instability} \\ 3500\frac{J}{kg} & < & MLCAPE & & & : & \text{extreme instability} \end{array} \tag{49}$$

(d) Maximum Unstable $CAPE$ - $MUCAPE$
$MUCAPE$ is the $CAPE$ of the most unstable parcel within the lowest $300hPa$. Note that each parcel has its own LCL, LFC and EL.

(e) 3kmCAPE
$3kmCAPE$ characterizes the $CAPE$ of the lowest $3km$ and is defined as

$$3kmCAPE = \begin{cases} g \int\limits_{LFC}^{min(EL,3km)} \frac{\Theta_{v,p}-\Theta_{v,e}}{\Theta_{v,e}} dz & \text{if } LFC < 3km \\ 0 & \text{if } LFC > 3km \text{ or no } LFC \end{cases} \tag{50}$$

High values of 3kmCAPE indicate high accelerations in low layers. Convection is more likely to start explosively in this case.

(f) Integrated CAPE
Integrated CAPE ore (ICAPE) is a column specific variable rather than a mass specific one. Thus it is provided in J/m^2 instead of J/kg. ICAPE can be defined as the sum over CAPE $\cdot dp/g$ along the column. It seems that there is not much experience with this variable which is defined by Mapes (1993).

9. Convective Inhibition - CIN
CIN is the energy needed to lift an air parcel from the surface to its LFC

$$CIN = g \int\limits_{0}^{LFC} \frac{\Theta_{v,p} - \Theta_{v,e}}{\Theta_{v,e}} dz. \tag{51}$$

If CIN is very low or if there is no CIN at all, convection starts in an early state of instability and only minor convection may happen. If it is too strong, then it cannot be overcome and no thunderstorms can be triggered except in case of forced lifting e.g. by mountains or large-scale dynamic forcing like confluence. The following thresholds usually apply:

$$\begin{array}{rclclcl} & & CIN & \leq & 15\frac{J}{kg} & : & \text{only minor cumuli develop} \\ 15\frac{J}{kg} & < & CIN & \leq & 50\frac{J}{kg} & : & \text{single cell thunderstorms possible} \\ 50\frac{J}{kg} & < & CIN & \leq & 200\frac{J}{kg} & : & \text{multi cell thunderstorms possible} \\ 200\frac{J}{kg} & < & CIN & & & : & \text{stability of stratification too high to overcome} \\ & & & & & & \text{no thunderstorms develop.} \end{array} \tag{52}$$

10. Upward Vertical Velocity UVV
    The upward vertical velocity represents the maximum vertical wind speed. According to parcel theory it is defined as

$$UVV = \sqrt{2CAPE}. \tag{53}$$

    The following thresholds in m/s are used:

$$\begin{array}{rl} UVV \leq 40: & \text{regular updraft} \\ 40 < UVV \leq 60: & \text{strong updraft} \\ 60 < UVV \leq 80: & \text{very strong updraft} \\ 81 < UVV: & \text{extreme updraft.} \end{array} \tag{54}$$

    Real updraft should be slower due to water and ice in the air and entrainment. Thus, the higher the water content the lower is w compared to UVV. Wind shear enables updrafts to persist longer. Rotating cells can have considerably larger updraft speeds. The potential of hail increases which increasing UVV. Large hail requires very strong to extreme updraft pertained by wind shear.

11. Lifted Parcel Level - LPL
    This is the height of the most unstable air parcel used in $MUCAPE$. It allows to identify the layer with highest $CAPE$.

12. Liquid Water Content - LWC
    High liquid water content is a product of strong condensation which released high amounts of energy. On the other hand, if liquid water falls into layers with relative humidity below 100% the water starts evaporating and by that cooling its environment. This leads to negative buoyancy and finally to downdraft. Thus the liquid water content can be used for the estimation of downdraft probability and strength.

13. Wet Bulb Zero Level - WBZ
    This is the height in which the wet bulb temperature is zero, sometimes also called $WB0$. Furthermore the name freezing level FRZ is used occasionally. WBZ is an important parameter with respect to hail and downdrafts since hail starts melting at that altitude and that melting cools the air. The higher the WBZ the less likely hail reaches the ground but the more likely downdrafts occur. The cooling itself lowers the WBZ).

14. Water Vapor Convergence - WVC
    This parameter describes the water vapor convergence in the lower troposphere. It is defined as

$$WVC = -\int_{p_0}^{p} \nabla \cdot (q\vec{v_h}) \frac{dp}{g} \tag{55}$$

    with $p_0 = 925hPa$ and $p = 700hPa$ if orography is higher than $550m$ and $p_0 = 1000hPa$ and $p = 850hPa$ otherwise. q is the specific humidity. A threshold for convection due to water vapor convergence is $0.1 \cdot 10^{-4} gm^2/s$.

15. Craven Parameter C
The Craven Parameter C is the simple product of 100mb $MLCAPE$ and the deep layer shear ($0-6km$, magnitude of the vector difference). The majority of significant severe events (hail ($d > 5cm$), winds ($v > 120km/h$), tornadoes $\geq$ F2) occur when the product exceeds $20,000m^3/s^3$. Some authors say that extreme events are likely if it exceeds $10,000m^3/s^3$

The index is formulated as follows:

$$C = MLCAPE[J/kg] \cdot DLS[m/s] \tag{56}$$

16. Energy Helicity Index - EHI
EHI is defined by Hart and Korotky (1991) as the normalized product of storm relative helicity and $CAPE$

$$EHI = \frac{CAPE \cdot SRH}{160,000}. \tag{57}$$

It therefore combines a measure of static instability with dynamics. It is a useful estimate of tornado risk with the following thresholds:

$$\begin{array}{rl} EHI > 1: & \text{Potential for supercells} \\ 5 \geq EHI > 1: & \text{up to F3 tornadoes possible} \\ EHI > 5: & \text{up to F5 tornadoes possible.} \end{array} \tag{58}$$

Note that tornadoes are also possible if EHI is small due to small $CAPE$ as long as SRH is high. Furthermore a sounding may not be representative since helicity may have high spatial variability.

17. Stability and Wind Shear Index for thunderstorms in Switzerland - $SWISS$
Two indices are developed in Switzerland (Huntrieser et al., 1997) in order to forecast thunderstorm possibility based on 00UTC and 12UTC observations:

    (a) $SWISS_{00}$

    $$SWISS_{00} = ShowI + 0.4VS_{3-6km} + 0.1(T_{600} - T_{d,600}) \tag{59}$$

    where VS_{3-6km} is the vertical wind shear between $3km$ and $6km$ above ground.

    (b) $SWISS_{12}$

    $$SWISS_{12} = LI - 0.1VS_{0-3km} + 0.1(T_{650} - T_{d,650}) \tag{60}$$

    where VS_{0-3km} is the vertical wind shear between $10m$ and $3km$ above ground.

18. Enhanced Stretching Potential - ESP
ESP is developed by Jon Davies (2005) as a guidline for the development of non-supercell/nonmesocyclone tornadoes. It is thus a useful tool in environments with small SRH and high LCL. The tornadoes that develop under these conditions are

usually weak (F0-F1) but can be strong (F2-F3) occasionally. ESP is computed as

$$ESP = \begin{cases} \left| (\gamma_{0-2km} - 7°C) \frac{\min(MLCAPE_{3km}, 100J/kg)}{50J/kg} \right| & \text{if } SBCAPE \geq 500J/kg \\ 0 & \text{else.} \end{cases} \tag{61}$$

According to Jon Davies, who did not publish this index, it should be used with care. However, in cases of lines of strong wind shift boundaries, non-supercell tornadoes may occur. And this is in line with high values of the ESP.

19. Bulk Richardson Number - BRN
The Bulk Richardson Number is defined by Weisman and Klemp (1982) as the ratio of $CAPE$ to the squared wind shear

$$BRN = \frac{CAPE}{0.5(v_1 - v_2)^2}. \tag{62}$$

v_1 and v_2 are wind speeds in different levels, usually $500m$ and $6km$. BRN is usually used to distinguish different kinds of thunderstorms:

$$\begin{array}{rcccl} & & BRN & < 10 & : \text{thunderstorms unlikely} \\ 10 & \leq & BRN & < 45 & : \text{supercells possible} \\ 45 & \leq & BRN & & : \text{single cells and multi cells possible} \end{array} \tag{63}$$

Since this index is a ratio of two energies it should always be interpreted together with the absolute values of the energies. $CAPE$ characterizes vertical instability while the wind shear tells more about the character of storms. Low wind shear makes multi-cells unlikely and single pulses likely, if enough $CAPE$ is available. Directional wind shear counteracts multi-cell development while speed shear supports it. Supercells are most likely when BRN is in its tens.

20. Dynamic State Index - DSI
This very interesting index is developed by Peter Nevir from Freie Universität Berlin. It measures the distance from hydrostatic and geostrophic equilibrium. I had no time to include it here.

21. Divergence of Q Vector
The divergence of the Q vector is sometimes used as a measure of instability.

2.5 Thematic Parameters

1. Severe Weather Threat Index $SWEAT$
$SWEAT$ is defined by Miller (1972). It consists of four terms

$$SWEAT = a + b + c + d \tag{64}$$

with the humidity term

$$a = 12\, T_{d850} \tag{65}$$

the instability term

$$b = 20 \begin{cases} (TT - 49) & \text{if } TT > 49°C \\ 0 & \text{else} \end{cases} \tag{66}$$

the shear term

$$c = 2v_{850} - v_{500} \tag{67}$$

where the wind speeds have to be provided in knots. The last term is a veering term in case there is veering

$$d = \left\{ \begin{array}{ll} 125\,(\sin(dd_{500} - dd_{850}) + 0.2) & \text{if } dd_{500} - dd_{850} > 0 \\ 0 & \text{else} \end{array} \right) . \tag{68}$$

dd_p means the wind direction in hight level p. $dd_{500} - dd_{850}$ is the veering between $500hPa$ and $850hPa$. If the wind direction in $500hPa$ is $200°$ and in $850hPa$ is $220°$ the difference is $-20°$ and thus the wind is not veering. Further conditions that are applied by some authors are

$$\begin{array}{rcccl} 210° & \leq & dd_{500} & \leq & 310° \\ 130° & \leq & dd_{850} & \leq & 250° \\ d_{500} & > & d_{850} & & \\ v_{500} & > & 15knots & \approx & 28\frac{km}{h} \\ v_{850} & > & 15knots & \approx & 28\frac{km}{h} \end{array} \tag{69}$$

With the last term the $SWEAT$ includes dynamical effects. Veering in 850 to $500hPa$ suggests warm air advection which due to the omega equation implies upward motion. The following thresholds apply:

$$\begin{array}{rl} 150 < SWEAT \leq 300: & \text{Slight severe} \\ 300 < SWEAT \leq 400: & \text{Severe possible} \\ 400 < SWEAT: & \text{Tornadic possible.} \end{array} \tag{70}$$

2. Downdraft $CAPE$ - $DCAPE$
   $DCAPE$ characterizes the available potential energy of downdraft. The integral is from the surface to the level of free sink LFS

   $$DCAPE = g \int_0^{LFS} \frac{\Theta_{v,p} - \Theta_{v,e}}{\Theta_{v,e}} dz \tag{71}$$

   $\Theta_{v,p}$ is usually approximated by the wet bulb potential temperature Θ_w of the sinking parcel, which is the temperature it achieves if rain water or ice is evaporated. The LFS is usually approximated as the height in which the wet bulb temperature is zero (WBZ). The square root of $2 \cdot DCAPE$ is the gain in downdraft velocity from LFS to the ground due to evaporation within the descending air parcel in case of no entrainment and detrainment of air.

3. Supercell Composite Parameter - SCP
   SCP combines $CAPE$, SRH and VS as

   $$SCP = \frac{MUCAPE}{1000J/kg} \frac{SRH_{0-3km}}{100m^2/s^2} \frac{DLS^2}{40m^2/s^2} \tag{72}$$

   Usually values greater than 1 strongly favor supercells, while non-supercell storms are generally associated with SCP values less than 1. A detailed discussion can be found at Thompson et al. (2003).They found Mean SCP values of 4 in case of supercells and 0.2 in case of nonsupercells.

4. Significant Tornado Parameter - STP
One definition of STP is

$$STP = \frac{DLS}{20m/s} \frac{SRH_{0-1km}}{100m^2/s^2} \frac{MLCAPE}{1000J/kg} \frac{2000 - LCL_{ML}}{1500m} \tag{73}$$

while NOAA provides

$$STP = \frac{DLS}{20m/s} \frac{SRH_{0-1km}}{100m^2/s^2} \frac{SBCAPE}{1500J/kg} \frac{2000 - LCL_{SB}}{1500m} \frac{100 + SBCIN}{150J/kg} \tag{74}$$

with the conditions that

- the LCL_{SB} term is set to 1 if LCL_{SB} is less than $1000m$,
- the $SBCIN$ term is set to 1 if $SBCIN > -50J/kg$, and
- the DLS term is capped at 1.5 and set to zero if $DLS < 12.5m/s$.

A majority of significant tornadoes (F2 and larger damage) have been associated with $STP > 1$ while the most non-tornadic supercells have been associated with $STP < 1$.

5. Significant Hail Parameter - $SHIP$
According to NOAA this parameter is designed to distinguish between significant hail (with diameters $\geq 2cm$ and non-significant hail. It is based on 5 parameters and given as

$$SHIP = \frac{-MUCAPE\ q_{MU}\ \gamma_{500-700hPa}\ T_{500hPa}\ DLS}{44 \cdot 10^{-6}} \tag{75}$$

with

$$\begin{aligned} q_{MU} &= \text{water vapor mixing ratio of MU parcel in } g/kg \\ \gamma_{500-700hPa} &= \text{700-500hPa lapse rate in } {}^\circ C/km \\ DLS &= \text{Deep Layer Shear in } m/s. \end{aligned} \tag{76}$$

$SHIP$ values greater than 1 indicate a favorable environment for significant hail. Values larger than 4 are considered very high. Values larger than 1.5 to 2 are usually linked to observed significant hail.

6. Lightning Potential Index LPI
The LPI is designed by NOAA as an empirical estimate of lightning risk. It is defined as

$$LPI = (A + B) \cdot (T_{850} - 272K) \tag{77}$$

limited between 0 and 20000 and with the parameters

$$\begin{aligned} A &= -RH^2 \cdot \left(\frac{d\Theta_e}{dz}\Big|_{600hPa} \right)^2 (min(LI, 0))^2 \\ B &= 0.001 \cdot muCAPE \cdot PW \cdot RH \end{aligned} \tag{78}$$

with the relative humidity RH, the equivalent potential temperature lapse rate $\frac{d\Theta_e}{dz}$ in 600hPa, the lifted index LI, the precipitable water PW, and the maximum unstable CAPE $muCAPE$ of the lowest 3000m above ground. The temperature in 850hPa has to be provided in Kelvin. 272K are subtracted in order to allow thunderstorms with lightning only in case of temperatures at 850hPa larger than -1.15 degrees Celsius. Further lightning indices exist which are not discussed here.

7. Supercell Detection Index - SDI
The supercell detection index SDI is meant to be applied to high resolution models. It is introduced on the basis of Doswell and Burgess (1993) by Droegemeier et al. (1993). Two variables are considered: vertical relative vorticity ζ and vertical wind speed w. These variables need to be available on a fine spatial grid (e.g. 2km horizontal resolution). Then for each location an environment is defined, say 10 grid points in each direction. They are used to define averages of ζ and w. Knowing the environmental average for each grid point allows for the estimation of the local deviation from the average and thus the estimation of the local correlation between ζ and w. The SDI is defined by this correlation multiplied by the column average of ζ ($\bar{\zeta}$) at the location as

$$SDI = \frac{\langle w'\zeta' \rangle}{\left(\langle w' \rangle^2 \langle \zeta' \rangle \right)^{0.5}} \cdot \bar{\zeta}. \tag{79}$$

Values of $SDI = 3 \cdot 10^{-4} s^{-1}$ are a minimum for the development of supercells. The SDI can be seen as a proxy of helicity.

8. Derecho Composite Parameter DCP
The DCP takes four pieces of information into account: 1. the development of a sufficient cold pool by means of $DCAPE$, 2. the degree of organization by wind shear, 3. the mean wind speed to allow downstream development, and 4. the ability to sustain strong storms at the leading edge by means of $MUCAPE$. Evans and Doswell (2001, [?]) proposed the following formula

$$DCP = \frac{DCAPE}{890 \frac{J}{kg}} \frac{Shear_{0-6km}}{20kts} \frac{MeanWind_{0-6km}}{16kts} \frac{MUCAPE}{2000 \frac{J}{kg}}. \tag{80}$$

I calculated DCP for some days on which derechoes occurred in Germany based on data provided by Christoph Gatzen and got no meaningful results. It seams that this parameter is so much tailored to conditions in the US that it can hardly be used in Europe.

9. Conditional Probability of MCS maintenance MMP
Mike Coniglio (NSSL) suggested this parameter. As a probability it can take values between 0 and 1. He proposed the following formula from a fit to observations

$$MMP = 1/\left\{1 + \exp\left[a_0 + a_1\, ms + a_2\, lr + a_3\, MUCAPE + a_4\, mw\right]\right\} \tag{81}$$

with $a_0 = 13$, $a_1 = -4.59 \cdot 10^{-2}$, $a_2 = -1.16$, $a_3 = -6.17 \cdot 10^{-4}$, and $a_4 = -0.17$. The variables are

$$\begin{aligned} ms &= \text{maximum bulk shear between 0-1km and 6-10km above ground } [m/s] \\ lr &= \text{lapse rate in 3 to 8 km in } K/km \\ MUCAPE &= \text{most unstable CAPE in } J/kg \\ mw &= \text{mean wind speed in 3 to 12 } km. \end{aligned} \tag{82}$$

MMP is set to zero for $MUCAPE < 100J/kg$. This parameter has the distinct advantage that it is given as a probability and therefore easy to interpret.

10. Conditional Probability of MCS forward speed > 18 m/s MSP
Also developed by Mike Coniglio (NSSL) this parameter estimates the likelihood that MCS get a forward speed above $18m/s$. MCS with forward speeds above this threshold are much more likely to produce widespread damage than those moving slower. The parameter is calculated with respect to some standardized values. It is defined as

$$MSP = 1/\left\{1 + \exp\left[a_0 + a_1\, d\theta_{xz} + a2\, mw + a3\, lr_z\right]\right\} \tag{83}$$

with $a_0 = -3.46$, $a_1 = 0.447$, $a_2 = 0.1119$, $a_3 = 0.79$ and

$$\begin{aligned} d\theta_{xz} &= \text{standardized values of max. diff. in } \theta_e \text{ between low and mid levels} \\ mw &= 2 - 12km \text{ mean wind speed in } m/s \\ lr_z &= \text{standardized values of 2-6km lapse rate.} \end{aligned} \tag{84}$$

The standardization is to be done with the local history of the respective parameters.

MSP is set to zero for $MUCAPE < 100J/kg$.

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