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## *Description of some convective indices implemented in the COSMO model*

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# *Description of some convective indices implemented in the COSMO model*

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# Contents

# 1 Introduction

Convective indices describe characteristics of the thermodynamical state of the atmosphere with regard to convection and are widely used by forecasters to estimate the potential for convective precipitation. Some of the most popular indices include Convective Available Potential Energy (CAPE) and Convective INhibition (CIN). Most of these stability indices are related to the vertical structure of temperature, humidity and wind and can be calculated from radiosonde measurements or output of NWP models.

The following document describes the algorithms used in the COSMO model (Version 4.10) code to calculate a set of convective indices.

# 2 Parcel theory

A common way to assess the instability of the atmosphere is to apply the parcel method, which evaluates the buoyancy of an air parcel displaced a finite distance under a pseudo-adiabatic or reversible process. The advantages of using the parcel approach includes the ability to evaluate the stability of an unsaturated air mass to the displacement of the saturated parcel, the capability of measuring the total reservoir of potential energy for the convection of a parcel and the ability to assess *conditional instability*, which occurs, when a small displacement is stable but a sufficiently large one is not. The disadvantage of the parcel method is that it cannot account for the reaction of the environment to the parcel displacement. For a detailed description of the parcel theory we refer to [4].

# 3 Convective indices

## 3.1 CAPE and CIN

The potential energy PE estimates the available potential energy of the atmosphere. In the parcel theory framework it is calculated by integrating the buoyancy of a parcel lifted from a starting level $z$ along the vertical parcel trajectory and ending at the equilibrium level $z_{\mathrm{EL}}$. We define the potential energy PE of a parcel using the virtual temperature according to [3]

$$\mathrm{PE}(z) = g \cdot \int_{z}^{z_{\mathrm{EL}}} \frac{T_v^p - T_v^e}{T_v^e} dz \tag{1}$$

Here, $g$ denotes gravitational acceleration, $T_v^{\{e,p\}}$ the virtual temperature $T_v = T(1 + \alpha q_v)$ of the environment and the parcel, respectively, $q_v$ is the specific humidity and $\alpha = (R_v - R_d)/R_d \approx 0.608$. $R_v$ and $R_d$ denote the gas constants of water vapour and dry air, respectively. The ascent of the parcel is calculated dry adiabatically whenever the parcel is unsaturated (i.e. before reaching the condensation level) and pseudo-adiabatically when the parcel is saturated with respect to the environmental air. PE usually consists of a negative part accumulated below the level of free convection (LFC) $z_{\mathrm{LFC}}$, commonly known as Convective INhibition (CIN)

$$\mathrm{CIN}(z) = -g \cdot \int_{z}^{z_{\mathrm{LFC}}} \frac{T_v^p - T_v^e}{T_v^e} dz \tag{2}$$

which is the energy that must be put into a parcel for its lifting until it gets positively buoyant and a positive part accumulated above the level of free convection, known as Convective

Available Potential Energy (CAPE)

$$\text{CAPE}(z) = g \cdot \int_{z_{\text{LFC}}}^{z_{\text{EL}}} \frac{T_v^p - T_v^e}{T_v^e} dz \tag{3}$$

which is the energy released by the parcel once it is positively buoyant and rises without the addition of external energy. The relation PE($z$) = CAPE($z$) - CIN($z$) holds.

Note that CAPE calculated with the parcel method is just an *upper bound* of the atmosphere's available potential energy due to the fact that the parcel cannot react with the environment (see [4]).

Since CAPE and CIN does not only depend on the thermodynamic structure of the sounding, but also on the parcel in question and on what thermodynamic process is applied in the parcel's displacement, there are many ways to calculate CAPE, which makes the comparison of different variants difficult. We adopted two CAPE and CIN methods which are widely used and robust in practice. They differ only by the definition of the starting parcel.

**Implementation in COSMO and fieldextra** The parcel ascent is performed using a parcel with given start temperature and humidity. It is then lifted dry adiabatically from model level to model level, preserving the potential temperature at the starting level

$$\theta = T \left( \frac{p_0}{p} \right)^{R_d/c_p} \tag{4}$$

until it reaches saturation at the lifted condensation level (LCL) $z_{\text{LCL}}$. Then it is further lifted pseudo-adiabatically, preserving the equivalent potential temperature at the lifting condensation level

$$\theta_e = T \left( \frac{p_0}{p} \right)^{R_d/c_p} \exp \left( \frac{L_v r_s}{c_p T} \right) \tag{5}$$

until it reaches the level of free convection $z_{\text{LFC}}$ where $T^e = T^p$ and buoyancy gets positive. The parcel continues its ascent pseudo-adiabatically until it reaches the equilibrium level $z_{\text{EL}}$. $p_0 = 1000$ hPa, $r_s$ is the saturation mixing ratio $r_s = R_d/R_v \cdot e_s/(p - e_s)$, $e_s$ is the saturation water vapour pressure, $c_p$ is the heat capacity at constant pressure of dry air and $L_v$ is the latent heat of vapourization. All constants are defined as in the COSMO code and are listed in the Appendix for convenience. The main algorithm is as follows

- set start temperature, humidity and start level kstart for the ascent
- calculate potential temperature ($\theta$) for start parcel
- loop over all levels
- if below LCL, perform dry ascent ($\theta$ const) to next level
- if at LCL, calculate equivalent potential temperature ($\theta_e$)
- if above or at LCL, perform wet ascent ($\theta_e$ const) to next level
- calculate buoyancy $T_v^p - T_v^e$
- check for LFC (buoyancy changes from negative to positive)
- accumulate CAPE or CIN (dependent on sign of buoyancy)
- check for 500 hPa level and store $T_v^p - T_v^e$ (used for SI and SLI)
- end loop over all levels

**Remarks**

- The saturation water vapour pressure is approximated using Teten's formula (also used in the COSMO code)
$$e_s(T) = b_1 \exp\left(b_{2w}\frac{T-b_3}{T-b_{4w}}\right)$$
with $b_1$=610.78 Pa, $b_{2w}$=17.27, $b_3$=273.16 K and $b_{4w}$=35.86 K

- Since the temperature cannot be derived analytically from the equivalent potential temperature, it is calculated numerically using the Newton iteration method
$$\begin{aligned} T^{(k+1)} = & \quad T^{(k)} - \tfrac{f(T^{(k)})}{f'(T^{(k)})} \\ f(T) = & \quad T(\tfrac{p_0}{p})^{R_d/c_p} \exp(\tfrac{L_v r_s}{c_p T}) - \theta_e \end{aligned}$$
where $k$ is the iteration index. There are usually less than 10 iterations needed for an accuracy of 0.003 K.

- Tests showed that very often a LFC is already found within the boundary layer even if significant capping inversions with a substantial amount of CIN are present above, implying that there are more than one LFC. Since CIN is defined as the part of PE below the LFC no CIN is present in these cases. To handle these situations in a meteorological meaningful way a check is done if the CIN within the capping inversion is larger than half (empirical value) the CAPE in the layer below. If this is the case, the LFC is set to the upper boundary of the CIN region (i.e. the second LFC) and CIN is defined as the CIN of the capping inversion.

- Since CIN is defined as the part of PE below the LFC, all contributions to CIN above the LFC are subtracted at the end of the ascent.

- In unstable situations (no LFC is present) or CAPE=0, CIN is set to "undefined value"

#### 3.1.1 CAPE and CIN based on the most unstable parcel

The parcel with the largest CAPE value of all starting parcels within a certain layer above the earth's surface is chosen. By definition, the CAPE value of this parcel is an upper bound of all parcels starting within the given layer. This has the advantage that the CAPE value is independent of large vertical gradients of temperature and humidity often occurring in the surface layer. Therefore this variant of CAPE is a robust measure also during night when shallow surface-based inversions are present. CIN is calculated using the same starting parcel. For a graphical representation of an ascent using the most unstable parcel see Figure 1 (black line).

**Implementation in COSMO and fieldextra** To calculate the CAPE and CIN using the most unstable parcel (CAPE_MU and CIN_MU), an ascent is performed for each model level starting at the surface and ending at 300 hPa (adjustable parameter) above the surface. For each ascent the CAPE value is calculated and the parcel with the maximum value is taken as the starting parcel. CIN_MU is the CIN corresponding to this starting parcel. Since several ascents need to be performed for each grid point, these indices are computationally more expensive than the mean layer CAPE/CIN and the Showalter Index/Surface Lifted Index.

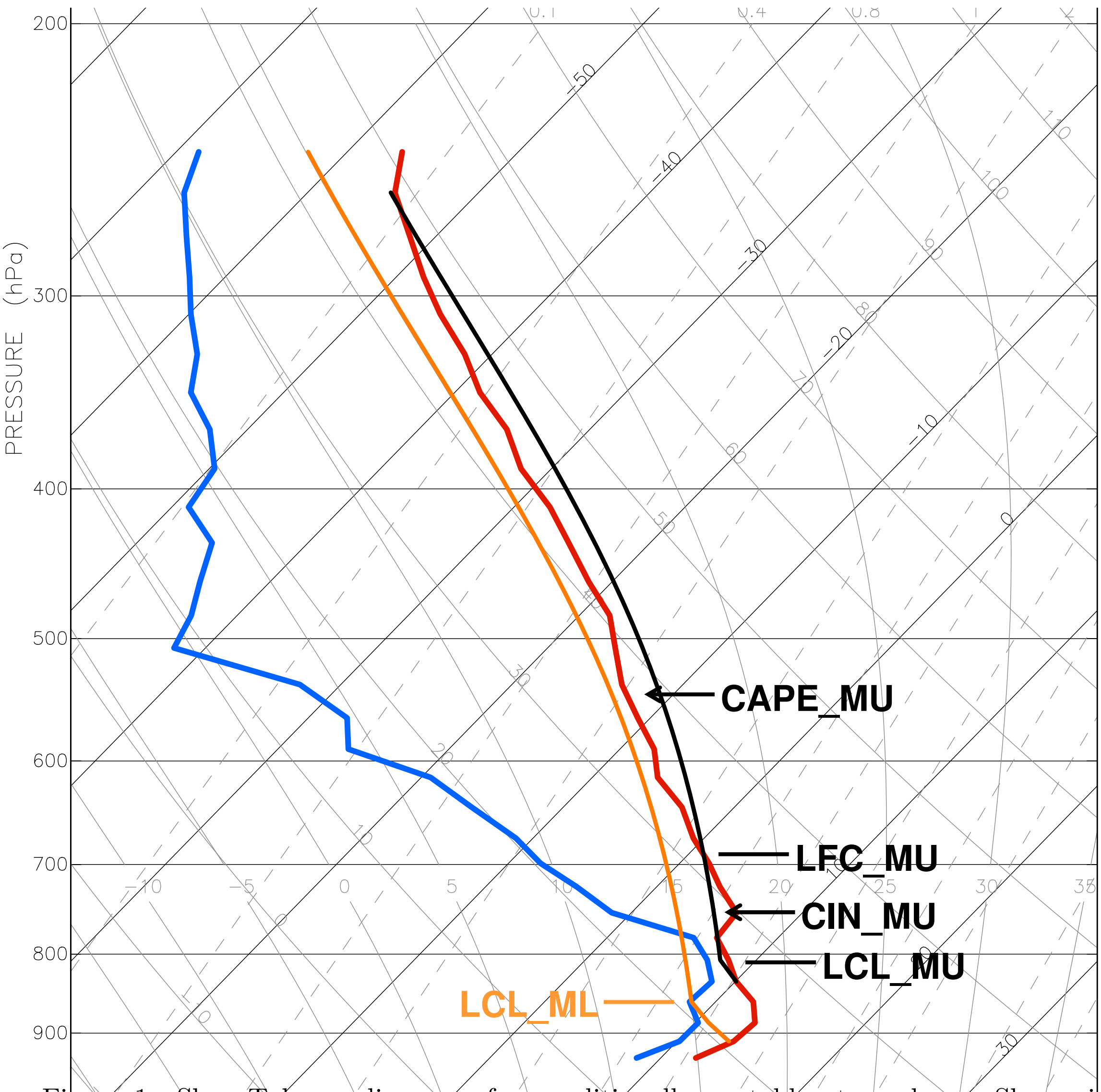


Figure 1: Skew-T log p diagram of a conditionally unstable atmosphere. Shown is the vertical distribution of temperature (red line) and dewpoint temperature (blue line) of the environment and two ascents using the parcel method starting from the most unstable parcel (black line) and a mean layer parcel (orange line). In the mean layer parcel ascent no level of free convection (LFC) is found, therefore CAPE_ML=0 and CIN_ML is undefined. In the most unstable parcel ascent a LFC (LFC_MU) is present, resulting in defined and non-negative CAPE_MU and CIN_MU. The lifting condensation levels are marked with LCL_MU and LCL_ML, respectively.

### 3.1.2 CAPE and CIN based on a mean layer parcel

In this case, the temperature and humidity of the starting parcel is calculated from that representative for a shallow surface layer (mean temperature and humidity of that layer). Like the CAPE/CIN using the most unstable parcel this results in a robust calculation without being too much dependent on the exact structure of the surface layer which may be not representative for the real atmosphere. CIN is calculated using the same string parcel. For a graphical representation of an ascent using the most unstable parcel see Figure 1 (orange line).

**Implementation in COSMO and fieldextra** To calculate the CAPE and CIN using a mean layer parcel (CAPE_ML and CIN_ML) a mean temperature and humidity from the lowest 50 hPa (adjustable parameter) above the surface are used for the "initialization" of the parcel. Only one ascent has to be performed per grid point.

### 3.1.3 CAPE_3KM on a mean layer parcel

CAPE_3KM is similar to the CAPE based on a mean layer parcel. The starting point for the parcel ascent is the same, whereas the ending point of the ascent is different. In case of CAPE based on a mean layer parcel, the ascent is performed until the equilibrium level $z_{\mathrm{EL}}$ is reached. For the CAPE_3KM the ascent is stopped at a height of 3 km above the ground. This particular CAPE is helpful for the forecasters. High values of CAPE_3KM tend to promote a stretching of the air columns and thus the development of tornadoes when high vorticity near the ground and high humidity occur simultaneously. Stretching is considered as significant when CAPE_3KM is greater than 150 J/kg (see [2]).

**Implementation in COSMO and fieldextra** The CAPE_3KM is computed exactly in the same way as the CAPE_ML excepting the fact that the ascent is stopped at 3 km above the ground.

## 3.2 Level of Free Convection

The Level of Free Convection LFC is the altitude at which the buoyancy is first positive, i.e. at which the lifted air parcel first becomes warmer than the surrounding air (see [4]). The PE of a lifted parcel is thus negative under the LFC (CIN) and we have to put some energy in the parcel in order to lift it. In opposite the PE of a lifted parcel is positive above the LFC (CAPE) and free convection occurs. The determination of the LFC is straightforward by computing the buoyancy at each level during the ascent and by searching the level at which the buoyancy sign changes.

**Implementation in COSMO and fieldextra** LFC is computed using a mean layer parcel (i.e. using mean values from the lowest 50 hPa above the surface as starting values for the parcel ascent).
The buoyancy is computed for each level:

$$Buo(k) = T_v^p(k) - T_v^e(k)$$

Then a sign change is searched: we test that $Buo(k) > 0$ and $Buo(k-1) < 0$. If it is the case $k$ is the index corresponding to the LFC. The LFC is then computed as the height above the ground corresponding to this index $k$.

### 3.3 Lifting Condensation Level

Adiabatic expansion of moist air will always ultimately lead to saturation in the earth's atmosphere (see [4]). The LCL is the height at which a parcel, upon dry adiabatic lifting, will first achieve saturation (see [6]). It is often associated with the cloud base. At this point the parcel is just saturated with no cloud liquid water (see [12]). It is computed by comparing the current specific humidity with the specific humidity at saturation under the same conditions (pressure, temperature).

**Implementation in COSMO and fieldextra** We compute the LCL based on a mean layer parcel. We thus use a mean temperature and humidity from the lowest 50 hPa above the surface as starting values. Saturation humidity is computed for each level and the current specific humidity is compared to it. If the current specific humidity is greater or equal to the saturation value the LCL is found.

### 3.4 Showalter Index (SI)

The Showalter Index SI [9] is defined by the difference between the environmental temperature at 500 hPa and the 500 hPa temperature of an air parcel lifted dry adiabatically from 850 hPa to its condensation level and pseudo-adiabatically thereafter.

**Implementation in COSMO and fieldextra** For the calculation of the Showalter Index, an ascent is performed with a starting parcel at 850 hPa. SI is then equal to the temperature difference between the environment and the lifted parcel at 500 hPa. Only one ascent has to be performed per grid point. If the 850 hPa surface is below the topography, SI is set to "undefined value".

### 3.5 Surface Lifted Index (SLI)

The Surface Lifted Index SLI is defined by the difference between the environmental temperature at 500 hPa and the 500 hPa temperature of an air parcel lifted dry adiabatically from the surface to its condensation level and pseudo-adiabatically thereafter.

**Implementation in COSMO and fieldextra** For the calculation of the Surface Lifted Index, an ascent is performed with a starting parcel at the first model level. SLI is then equal to the temperature difference between the environment and the lifted parcel at 500 hPa. Only one ascent has to be performed per grid point.

### 3.6 $\text{SWISS}_{00}$ Index (SWISS00)

The $\text{SWISS}_{00}$ Index was especially designed for northern Switzerland during night conditions [5]. It is based on a combination of the Showalter Index, the wind shear between 3 and 6 km a.s.l. and the dewpoint depression on 600 hPa. This definition is similar to that of the SWEAT Index developed for the USA.

$$\mathrm{SWISS}_{00} = \mathrm{SI} + 0.4\mathrm{WSh}_{\text{3-6km}} + 0.1(T - T_d)_{600\mathrm{hPa}}$$

where SI is the Showalter Index defined above and $\mathrm{WSh}_{\text{3-6km}}$ is the wind shear between 3 and 6 km.

**Implementation in COSMO and fieldextra** The $\mathrm{SWISS}_{00}$ Index is calculated as defined in section 3.6. The vertical wind shear is approximated by

$$\mathrm{WSh}_{\text{3-6km}} = \sqrt{u^2 + v^2}_{6\mathrm{km}} - \sqrt{u^2 + v^2}_{3\mathrm{km}}$$

## 3.7 $\mathrm{SWISS}_{12}$ Index (SWISS12)

The $\mathrm{SWISS}_{12}$ Index was especially designed for northern Switzerland during day conditions [5]. It is based on a combination of the Surface Lifted Index, the wind shear between ground and 3 km a.s.l. and the dewpoint depression on 650 hPa. This definition is similar to that of the SWEAT Index developed for the USA.

$$\mathrm{SWISS}_{12} = \mathrm{SLI} - 0.3\mathrm{WSh}_{\text{0-3km}} + 0.3(T - T_d)_{650\mathrm{hPa}}$$

where SLI is the Surface Lifted Index defined above and $\mathrm{WSh}_{\text{0-3km}}$ is the wind shear between surface and 3 km.

**Implementation in COSMO and fieldextra** The $\mathrm{SWISS}_{12}$ Index is calculated as defined in section 3.7. The vertical wind shear is approximated by

$$\mathrm{WSh}_{\text{0-3km}} = \sqrt{u^2 + v^2}_{3\mathrm{km}} - \sqrt{u^2 + v^2}_{\mathrm{z(kdim)}}$$

z(kdim) is the height of the lowest model level.

## 3.8 Heat Index

The Heat Index (HI) is an index that combines air temperature and relative humidity in an attempt to determine the human-perceived equivalent temperature, how hot it feels, termed the felt air temperature. The human body normally cools itself by perspiration, or sweating, which evaporates and carries heat away from the body. However, when the relative humidity is high, the evaporation rate is reduced, so heat is removed from the body at a lower rate causing it to retain more heat than it would in dry air. Measurements have been taken based on subjective descriptions of how hot subjects feel for a given temperature and humidity, allowing an index to be made which create a correspondence between a temperature and humidity combination and a higher temperature in drier air.

The HI derives from a model comprising a collection of equations developed by Steadman [11]. It is the result of extensive bio-meteorological studies. The model is reduced by an iterative procedure to a relationship between temperature and relative humidity versus apparent temperature, Steadman developed a table based on this relationship.
The equation used commonly results from a multiple regression analysis performed on the data from Steadman's table [7] and is given by:

$$\begin{aligned} HI &= -42.379 + 2.0490T + 10.1433R - 0.2248TR - 6.8378 * 10^{-3}T^2 \\ &\quad - 5.4817 * 10^{-2}R^2 + 1.2287 * 10^{-3}T^2R + 8.5282 * 10^{-4}TR^2 - 1.99 * 10^{-6}T^2R^2 \end{aligned}$$

where $T$ is the ambient dry bulb temperature (°F) and $R$ is the relative humidity (%). This formula is only meaningful for temperatures above 80°F (corresponding to 26.7°C) and relative humidities above 40%.

The meaning of a Heat Index value can be summarized as follows:

| Heat Index | Effects |
|---|---|
| <70°F | no effect for most of the people |
| 71-79 °F | discomfort for half of the people |
| 80-90°F | caution: fatigue is possible. Discomfort for all the people |
| 91-105°F | extreme caution: sunstroke, heat cramps, heat exhaustion possible |
| 106-130°F | danger: sunstroke, heat cramps, heat exhaustion likely |
| >130°F | extreme danger: heat strokes and sunstroke likely |

**Implementation in fieldextra** The Heat Index is computed with the formula given in section 3.8 and for temperatures above 80°F (corresponds to 26.7°C) and relative humidities above 40%. For all other combinations of $T$ and $R$ the HI is flagged as undefined.

[7] does not mention at which height the temperature and the relative humidity should be considered. But this index attempts to determine the human-perceived equivalent temperature so the typical height is about 1.5-2 m. Thus, we choose to consider them at 2 meters height because of realism and simplicity. These outputs from the COSMO model are thus used:
- 2m temperature
- 2m relative humidity.

### 3.9 Bulk Richardson number and height of the planetary boundary layer (PBL)

The stability of the atmosphere plays a key role in the transport of air masses as well as in the transport of pollutants or humidity. The vertical transport is prevented in case of a stable stratified atmosphere whereas it is enhanced under unstable, or convective, conditions [12]. The dynamic stability is defined as the ability of an air mass to resist or recover from finite perturbations of a steady state. The perturbations are mechanically or thermally initiated [4].

The boundary layer is defined as the lowest part of the troposphere, whose behavior is directly influenced by the surface forcing with a response time of less than an hour [12]. This layer is turbulent, showing rapid fluctuations of the physical quantities, while the rest of the atmosphere (free atmosphere) is usually non turbulent or intermittently turbulent ([6]). Its thickness varies during the day cycle and during the year. It is generally thin at night and in the cool season and thicker during the day and in the warm season [12].

The characterization of the atmospheric stability and of the boundary layer height is crucial for a good comprehension of our atmosphere [8] and applications in air quality.

The following document describes the algorithms used in the COSMO model code to calculate the Bulk Richardson Number (BRN) and the height of the Planetary Boundary Layer (HPBL).

### 3.9.1 Bulk Richardson Number

The gradient Richardson Number is a dimensionless number relating the buoyant consumption term to the mechanical production term of the TKE budget equation [10].

$$Ri = \frac{(g/\theta_v)(\partial\theta_v/\partial z)}{(\partial u/\partial z)^2 + (\partial v/\partial z)^2} \tag{6}$$

Where $\theta_v$ is the virtual potential temperature in $K$, $g$ is the acceleration due to gravity in $m/s^2$ , $z$ is the height in $m$, $u$ is the zonal wind in $m/s$ and $v$ is the meridional wind in $m/s$.

The Bulk Richardson Number BRN is an approximation of the gradient Richardson Number formed by approximating local gradients by finite difference across layers [12]. When approximating $\partial\theta_v/\partial z$ by $\Delta\theta_v/\Delta z$ and $\partial u/\partial z$ and $\partial v/\partial z$ by $\Delta u/\Delta z$ and $\Delta v/\Delta z$ respectively, we obtain the BRN:

$$BRN = \frac{g\Delta\theta_v\Delta z}{\theta_v[(\Delta u)^2 + (\Delta v)^2]} \tag{7}$$

We compute it between the ground and a height z:

$$BRN(z) = \frac{g(\theta_v(z) - \theta_{v_{ground}})(z - z_{ground})}{\theta_v(z)[(u(z) - u_{ground})^2 + (v - v_{ground})^2]} \tag{8}$$

Some authors [6] consider that $\theta_{v_{ground}} = \theta_v(z_{0,h})$ where $z_{0,h}$ is the energy roughness length [1]. In NWP models we consider however the surface values or the 2 meters values for the temperature related quantities and the 10 meters values for the winds.
According to [12], the bulk form of the Richardson Number, BRN, is used most frequently in meteorology because of the discrete nature of rawinsonde measurements and of numerical models.

[6] suggests this interpretation for the BRN:

| BRN | Flow Type | Level of Turbulence due to Buoyancy | Level of Turbulence due to Shear |
|---|---|---|---|
| $< 0$, large | turbulent | large | small |
| $< 0$, small | turbulent | small | large |
| $> 0$, small | turbulent | none | large |
| $> 0$, large | laminar | none | small |

**Implementation in the COSMO model** The BRN is computed according to equation (8). We consider no velocity at the ground, i.e. $u_{ground} = 0$ and $v_{ground} = 0$. The virtual

potential temperature at the denominator should be the mean virtual potential temperature for the whole layer. We thus have $\overline{\theta_v(z)} = [\sum_{k=1}^{n} \theta_v(k)]/(n+1)$, where n is the number of levels between the ground and the level $k$, instead of $\theta_v(z)$.
We thus obtain:

$$BRN(z) = \frac{g(\theta_v(z) - \theta_{v2m})(z - z_{ground})}{\overline{\theta_v(z)}[u(z)^2 + v(z)^2]} \tag{9}$$

### 3.9.2 Height of the Planetary Boundary Layer

The height of the planetary boundary layer HPBL is a fundamental parameter characterizing the structure of the lowest troposphere [8]. It can be derived from profiles or parameterized. The second approach is used in NWP models. The parameterization is based on prognostic or diagnostic equations. The prognostic equations are more complicated and diagnostic methods are thus preferred.

The Bulk Richardson Number method is the standard way to derive HPBL from model outputs [8]. [10] considers it as a robust and fairly accurate method, which is particularly suited when the vertical resolution of the meteorological fields (temperature, winds) is limited.

This method consists in calculating the BRN and then in searching the height at which the BRN reaches a critical value, the critical Bulk Richardson Number. This level is the top of the boundary layer [10]. In the literature, one finds critical values between say 0.2 and 1. The method is however not very sensitive to this critical value. Values between 0.1 and 0.4 are generally admitted [10].
We have to notice that the diagnostic methods usually perform well under convective conditions, whereas their ability to predict the HPBL under stable conditions is poorer [8]. It can happen that the BRN profile never crosses the critical Bulk Richardson Number. In those cases it is thus impossible to find a HPBL with this method.

**Implementation in the COSMO model** There is no consensus in the literature about the critical Bulk Richardson Number. We decide to take into account the current stability conditions for the choice of the critical value. If the conditions are stable, we use a critical value $Ri_{B,cr}$ of 0.33 [14], whereas under convective conditions we use the value prescribed by [13] $Ri_{B,cr} = 0.22$.

The stability assessment is carried out in a simple way by computing the coefficient of the linear regression between the virtual potential temperature and the height in the 4 first model levels. If the coefficient is positive, the conditions are stable and if it is negative the conditions are known as convective. It is given by:

$$\beta = \frac{\sum_{i=1}^{4} z\theta_v - 1/4 \sum_{i=1}^{4} z \sum_{i=1}^{4} \theta_v}{\sum_{i=1}^{4} z^2 - 1/4 \sum_{i=1}^{4} z \sum_{i=1}^{4} z} \tag{10}$$

We then scan the levels starting at the bottom model level until a level satisfying this condition is reached: $BRN(z) > Ri_{B,cr}$. This level is the top of the PBL. If no such level is found a missing value is returned.

We currently have not set some minimal and maximal threshold heights. It could be however useful when using this HPBL for certain applications (dispersion models for example).

### 3.10 Supercell Detection Index

The **Supercell Detection Index (SDI)** was devised by Wicker et al. [15] to help forecasters to detect the mesocyclone of a supercell from high resolution forecast models. At each horizontal grid point (i,j), the first Supercell Detection Index $\mathrm{SDI}_1$ is defined as the product

$$\mathrm{SDI}_{1,ij} := \rho_{ij}\overline{\zeta_{ij}} \tag{11}$$

of the velocity-vorticity correlation

$$\rho_{ij} := \frac{< w'\zeta' >}{(< w'^2 >_{ij} < \zeta'^2 >_{ij})^{1/2}}$$

and the vertically averaged vorticity

$$\zeta_{ij} := (\nabla \times \mathbf{v})_{\mathbf{z}}.$$

Here, $< \ldots >$ denotes the average, taken over a sliding 3D slab of extensions 20 km * 20 km * [1.5 ... 5.5] km, and the overbar on $\zeta$ denotes a vertical average, taken in the height range [1.5 ... 5.5] km.
In contrast to other quantities that are used to predict tornados, like for instance the Storm Relative Environmental Helicity (SREH), or the near surface wind shear, the $\mathrm{SDI}_1$ directly attempts at detecting the stream shape of a supercell in the model. That is the reason, why the $\mathrm{SDI}_1$ can only be applied if the model resolution is sufficiently high. Hence it cannot be used for COSMO-7, since the resolution there is too low. The following threshold values of the $\mathrm{SDI}_1$ for the detection of supercells are given in [15]:

| | |
|---|---|
| $\lvert\mathrm{SDI}_1\rvert = 0.0003$ 1/s: | minimal threshold for supercells |
| $\lvert\mathrm{SDI}_1\rvert > 0.003$ 1/s: | significant signal for supercells |

For regions of updraft, the product of correlation and vorticity is positive, and thus $\mathrm{SDI}_1 > 0$ holds true. For regions of downdraft $\mathrm{SDI}_1 < 0$.
Since the up- and downdrafts for supercells are coupled to each other, one is, however, rather interested in using the sign information to detect the sense of rotation of the supercells. Wicker et al. [15] thus define the second Supercell Detection Index $\mathrm{SDI}_2$ at grid point (i,j) as

$$\mathrm{SDI}_{2,ij} := \begin{cases} \rho_{ij}\lvert\overline{\zeta_{ij}}\rvert & : \quad w > 0 \\ 0 & : \quad w \leq 0 \end{cases} \tag{12}$$

Thus it holds that $\mathrm{SDI}_2 > 0$ for regions of cyclonic updrafts, and $\mathrm{SDI}_2 < 0$ for regions of anticyclonic updrafts.
In practice, evaluation of the $\mathrm{SDI}_2$ should be sufficient. The threshold values given for $\mathrm{SDI}_1$ are of course also valid for $\mathrm{SDI}_2$.

**Implementation in the COSMO model** Within COSMO, the Supercell Detection Indices $\mathrm{SDI}_1$ and $\mathrm{SDI}_2$ have been implemented by Axel Seifert (DWD) according to (11) and (12), respectively.

# A Table of Constants

| Name | Description | Value | Unit |
|---|---|---|---|
| $g$ | Acceleration due to gravity | 9.80665 | m $s^{-2}$ |
| $R_d$ | Gas constant of dry air | 287.05 | $JK^{-1}kg^{-1}$ |
| $R_v$ | Gas constant of water vapour | 461.51 | $JK^{-1}kg^{-1}$ |
| $c_p$ | Specific heat of dry air at constant pressure | 1005 | $JK^{-1}kg^{-1}$ |
| $L_v$ | Latent heat of vapourization | $2.501 \cdot 10^6$ | $Jkg^{-1}$ |

Table 1: Table of constants

# B Grib parameters of indices

| Name | Unit | GRIB Table | Grib Nr. | Level Type |
|---|---|---|---|---|
| CAPE_MU | 1/s | 201 | 143 | 1 |
| CAPE_ML | J/kg | 201 | 145 | 1 |
| CIN_MU | J/kg | 201 | 144 | 1 |
| CIN_ML | J/kg | 201 | 146 | 1 |
| CAPE_3KM | J/kg | 203 | 137 | 1 |
| LCL_ML | m | 203 | 135 | 1 |
| LFC_ML | m | 203 | 136 | 1 |
| SWISS00 | 1 | 203 | 138 | 1 |
| SWISS12 | 1 | 203 | 139 | 1 |
| BRN | 1 | 203 | 155 | 1 |
| HPBL | m | 203 | 156 | 1 |
| HI | 1 | 250 | 24 | 105 |
| SDI_1 | 1/s | 201 | 141 | 1 |
| SDI_2 | 1/s | 201 | 142 | 1 |

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Issues of the COSMO Technical Reports series are published by the *COnsortium for Small-scale MOdelling* at non-regular intervals. COSMO is a European group for numerical weather prediction with participating meteorological services from Germany (DWD, AWGeophys), Greece (HNMS), Italy (USAM, ARPA-SIMC, ARPA Piemonte), Switzerland (MeteoSwiss), Poland (IMGW) and Romania (NMA). The general goal is to develop, improve and maintain a non-hydrostatic limited area modelling system to be used for both operational and research applications by the members of COSMO. This system is initially based on the COSMO-Model (previously known as LM) of DWD with its corresponding data assimilation system.

The Technical Reports are intended

- for scientific contributions and a documentation of research activities,
- to present and discuss results obtained from the model system,
- to present and discuss verification results and interpretation methods,
- for a documentation of technical changes to the model system,
- to give an overview of new components of the model system.

The purpose of these reports is to communicate results, changes and progress related to the LM model system relatively fast within the COSMO consortium, and also to inform other NWP groups on our current research activities. In this way the discussion on a specific topic can be stimulated at an early stage. In order to publish a report very soon after the completion of the manuscript, we have decided to omit a thorough reviewing procedure and only a rough check is done by the editors and a third reviewer. We apologize for typographical and other errors or inconsistencies which may still be present.

At present, the Technical Reports are available for download from the COSMO web site (www.cosmo-model.org). If required, the member meteorological centres can produce hard-copies by their own for distribution within their service. All members of the consortium will be informed about new issues by email.

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