# Section 4

# Parameterization of important atmospheric and surface processes, effects of different parameterizations

# Development of neural network ensemble stochastic convection parameterizations for climate models using CRM simulated data

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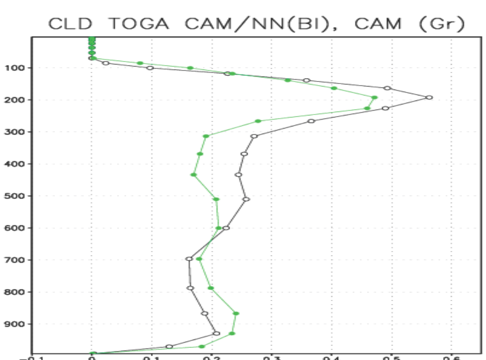
A novel approach based on the neural network (NN) technique has been formulated and used for development of NN ensemble stochastic convection parameterizations for climate and NWP models. This fast NN convection parameterization is built based on direct learning cloud physics from Cloud Resolving Model SAM (System for Atmospheric Modeling, Khairoutdinov and Randall, 2003) simulated data. SAM simulations have been initialized with and forced by 120-day long TOGA-COARE data. SAM data simulated over the TOGA-COARE location have been averaged to produce hourly and horizontally, 256 km x 256 km, means. The data was projected onto a GCM space of atmospheric states to implicitly define a stochastic convection parameterization. That is only a subset of relevant SAM variables available in a climate model (NCAR CAM) is selected for creating an NN training data set.

An ensemble of NNs have been trained and their accuracy is estimated vs. SAM simulated data. This NN ensemble represents the stochastic convection parameterization. The inherent uncertainty of the stochastic convection parameterization is indicated and estimated.

Validation of the NN stochastic convection parameterization in NCAR CAM has been done in a diagnostic mode (CAM/NN). Actually, CAM inputs have been used, at every time step and grid point for calculations of the NN convection parameterization to produce its outputs as a diagnostic product. Parallel decadal CAM and CAM/NN simulations have been produced for 1990-2001 for winters (NDJF) excluding the TOGA-COARE 1992-93 winter. The CAM/NN run includes bias corrections or model calibration. SAM-CAM bias corrections are consistently applied to NN inputs and outputs to account for differences between SAM and CAM or their different “virtual realities”. SAM-CAM bias corrections are calculated for the TOGA-COARE point (-2 S, 155 E) and time averaged for the TOGA-COARE winter. The point bias is applied at every time step and grid point throughout the decadal CAM/NN diagnostic run. The CAM/NN and CAM runs are compared for consistency, and with the NCEP reanalysis



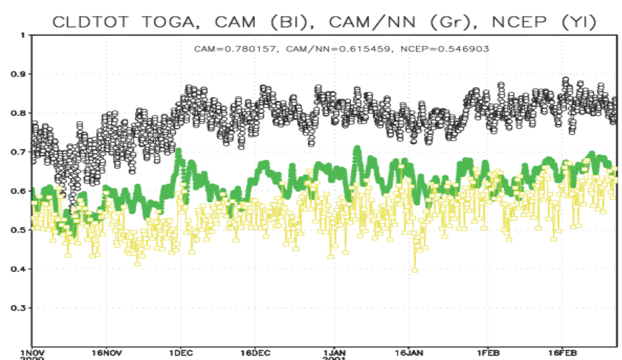




Fig. 1 (left) Vertical profiles of decadal mean CLD, in fractions, for CAM/NN and CAM

Fig. 2 (right) Time series of decadal mean total cloudiness (in fractions) for the TOGA-COARE location for the CAM run (black) and CAM/NN (green) runs, and for the NCEP reanalysis (yellow).

The decadal mean CLD profiles for CAM/NN and CAM (Fig. 1) are consistent and close to each other including the maximum at 200 hPa. The time series of the decadal mean total CLD (Fig. 2) for the CAM run show measurably higher magnitudes, with the mean of 0.78, compared to those of the time series for the CAM/NN run, with the mean of 0.61. The time series of the NCEP reanalysis show lower magnitudes, with the mean of 0.54, which are significantly closer to those of the time series for CAM/NN.

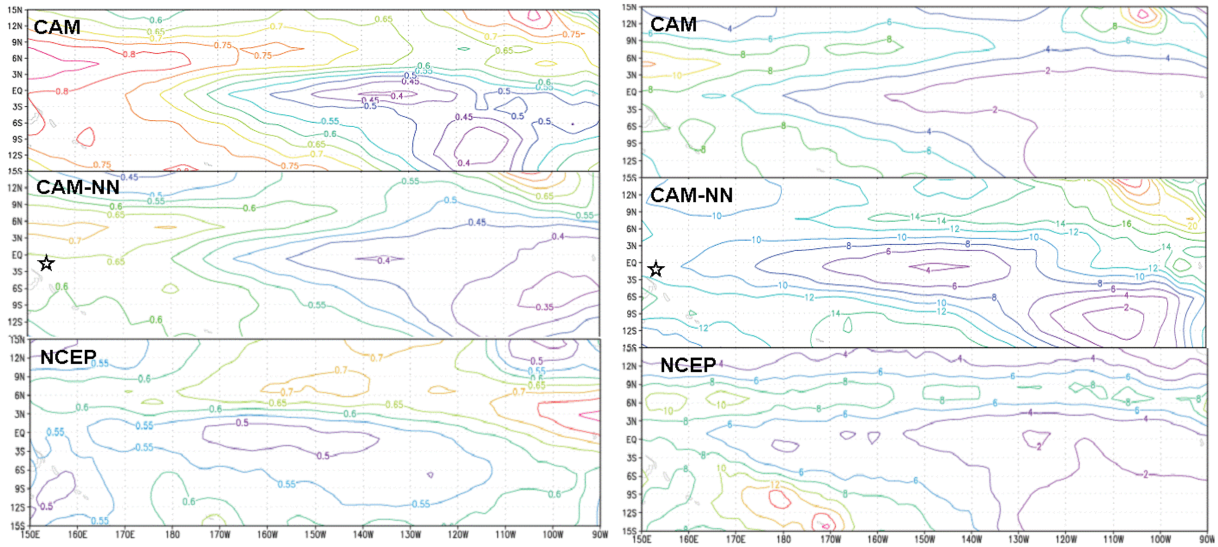



Fig. 3 Decadal mean winter (NDJF) distributions for the Tropical Pacific region for CLD (left panels), in fractions, and precipitation (right panels), in mm/day, for: NCEP reanalysis (the bottom panels), the CAM/NN with the SAM-CAM bias corrections run (the middle panels), and the CAM run (the upper panels). The TOGA-CORE location, for which the NN convection parameterization and bias correction have been calculated for the TOGA-COARE winter only, is shown by a star in the middle panels.

The patterns of both CLD and precipitation for the CAM/NN run are generally consistent with those of the CAM run, and with those of the NCEP reanalysis. The CLD magnitudes for the CAM/NN run are measurably closer to those of the NCEP reanalysis than those of the CAM run. Compared the NCEP reanalysis, the PREC magnitudes for the CAM/NN run are closer to those of the CAM run for 15º S to around the Equator, but overestimated for the area north of 6º N.

The CLD decadal time series for the TOGA-COARE location and the CLD distribution for the Tropical Pacific region are consistent in the sense that for both the CAM run shows measurably higher magnitudes compared to those of the CAM/NN run, the later being much closer to those of the NCEP reanalysis. These results obtained for the decadal CAM/NN simulation seem to be positive and encouraging.

**Conclusions: (**1) A novel NN approach has been formulated and used for development of NN ensemble stochastic convection parameterizations for climate models. **(**2) Developed NN ensemble stochastic convection parameterizations have been tested in parallel decadal CAM/NN and CAM runs. The obtained results are positive and encouraging.

**Future plans:** Developing NN ensemble stochastic convection parameterizations using SAM simulations driven by CAM forcing for longer periods, different geographic locations, and diverse weather conditions. It will allow us to develop a NN ensemble stochastic convection parameterization that can be used globally.

**Acknowledgements:** Marat Khairoutdinov (SUNY – Stony Brook) for providing SAM.

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**Modification of the Kain-Fritsch convective parameterization scheme in the Meso Scale Model**

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The Japan Meteorological Agency has been operating the Meso Scale Model (MSM) – a nonhydrostatic model with a horizontal grid size of 5 km – to support the mitigation of meteorological disasters since March 2006. The MSM adopts the Kain-Fritsch convective parameterization scheme (KF scheme: Kain and Fritsch 1990; Kain 2004) along with a cloud microphysics scheme to estimate sub-grid scale convection. Previously, the KF scheme in the MSM often predicted false precipitation along coastlines due to its oversensitivity to topography and differences in roughness between sea and land in conditions with highly moist incoming air (Fig. 1 (b)) compared to the results of actual observation (Fig. 1 (a)). Narita (2010) showed that a higher mixing rate between sub-grid convective updrafts and their environmental air was effective in suppressing this type of false precipitation prediction. We have now modified the KF scheme to introduce the variable radius of the sub-grid updraft in order to increase the mixing rate.

The KF scheme was originally developed on a model with coarser grid spacing, and it was assumed that the radius of the sub-grid scale updraft was limited to between 1,000 m and 2,000 m as a function of its upward velocity. Accordingly, we implemented modification to adopt a more suitable radius for the sub-grid scale updraft in the 5-km grid size model. The modified radius depends on the lifted condensation level (LCL), i.e., the radius becomes gradually smaller as the LCL decreases. The mixing rate is formulated to be inversely proportional to the radius of the updraft in the scheme, and the modified radius therefore brings about enhanced mixing. This modification causes damping of excessively developed convection, resulting in a reduction of false precipitation (Fig. 1 (c)).

Using the modified KF scheme, we tested the operational MSM and verified precipitation predictions in an experiment conducted between 7$^{th}$ and 26$^{th}$ July, 2009 (during the highly moist rainy season). The statistical verification results for accumulated precipitation every three hours are shown in Fig. 2. The equitable threat score shown in Fig. 2 (a) indicates an improvement in the range between 10 mm/3h and 25 mm/3h, while the bias score indicates a slightly decreasing trend for the predictive frequency of relatively weak precipitation and an increasing one for that of heavy precipitation (Fig. 2 (b)).

The modified KF scheme successfully suppresses false precipitation predictions along the coast and contributes to the improvement of quantitative precipitation forecasts. The modification was incorporated into the operational MSM in November 2010.

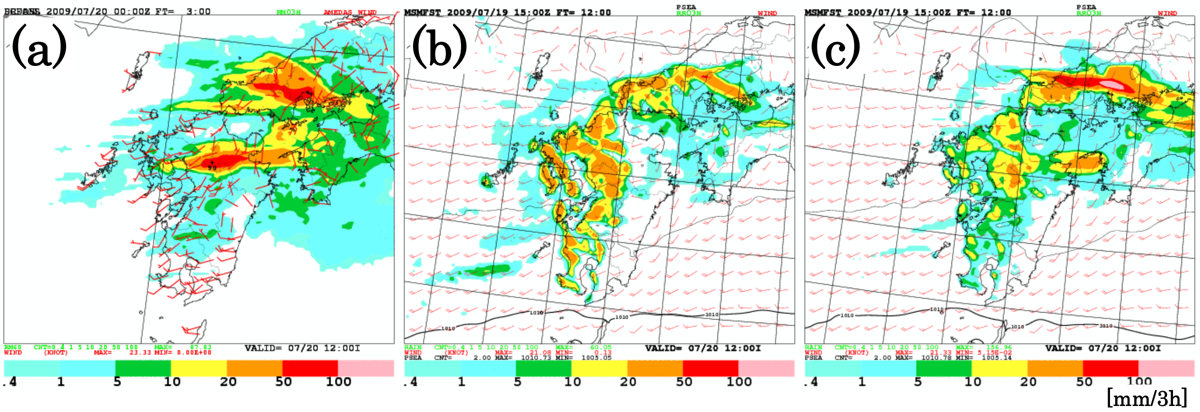


Fig. 1 Improvement of suppression for false precipitation prediction along coastlines: accumulated precipitation [mm/3h] of: (a) observed (radar-raingauge analyzed precipitation); (b) MSM forecasts using the original KF scheme; and (c) MSM forecasts using the modified KF scheme

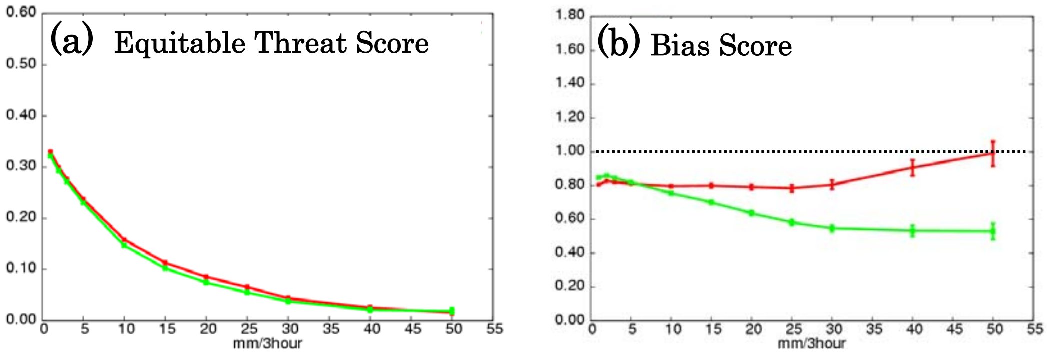


Fig. 2 Statistical verification results for three-hourly accumulated precipitation of MSM forecasts against radar-raingauge analyzed precipitation based on the contingency table in the term between 7th and 26th July, 2009: (a) equitable threat score; (b) bias score. The scores for MSM forecasts using the modified KF scheme are indicated by the red line, while the original ones are indicated by the green line, with error bars showing a 95% confidence interval.

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# Vertical resolution dependency of boundary layer schemes


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## 1 Introduction

The performance of physical processes in numerical weather prediction (NWP) model is significantly affected by vertical resolution because physical processes in the current NWP models are vertical one dimensional models.

In the atmospheric boundary layer, turbulence transports momentum, heat and moisture. The turbulence is mainly driven by surface flux, i.e. exchange of momentum, heat and moisture between the atmosphere and the surface. In addition, at the top of stratocumulus, which is often generated at the top of the mixed layer, longwave radiation cooling also drives turbulence inside stratocumulus. One of the main roles of boundary layer (or turbulent) schemes in the NWP models is to represent the transport by turbulence.

Boundary layer schemes should also be sensitive to vertical resolution, especially in stratocumulus at the top of which sharp discontinuity and considerably large subsidence can be seen. Lenderink and Holtslag (2000) pointed out that boundary layer schemes based on prognostic or diagnostic turbulent kinetic energy (TKE) can give artificial mixing, which they call "numerical entrainment", due to insufficient vertical resolution.

On the basis of that, with the aim of increasing the vertical resolution of the operational NWP models, the sensitivity of the improved Mellor-Yamada level 3 model (MYNN3) (Nakanishi and Niino 2009) employed in JMA's operational mesoscale model (Saito et al. 2007) was examined using a single column model (SCM) in the Unified Model (UM) of the UK Met Office[1] for typical boundary layer situations. The configurations used in GABLS2[2] for diurnal changes of cloud-free boundary layer and EUROCS (Duynkerke et al. 2004) for diurnal changes of stratocumulus-capped boundary layer were adopted as test cases.

## 2 Configurations of the experiments

### 2.1 Boundary layer scheme

The MYNN3 assessed in the experiments predicts four covariances of turbulent fluctuation: TKE, self-covariance of liquid water potential temperature ($\overline{\theta_l'^2}$), water content ($\overline{q_w'^2}$), and their covariance ($\overline{\theta_l' q_w'}$). Buoyancy flux appearing in the production term of TKE is evaluated using the bi-normal probability distribution function (PDF) to incorporate the effect of subgrid condensation (Sommeria and Deardorff 1977). The width of the bi-normal PDF is determined from the prognostic covariances in the MYNN3. Flux plays an important role in generating turbulence inside clouds.

### 2.2 Vertical resolution

In order to examine the dependency on vertical resolution, three sets of vertical layer assignments were prepared. The grid spacings are regulated by an arithmetic sequence written as

$$\Delta z_n = \Delta z_1 + a(n-1), \tag{1}$$

with the common difference $a$ and the first grid spacing increment $\Delta z_1$. The height of the $n$-th layer $z_n$ is determined as $z_{n+1} = z_n + \Delta z_n$.

In Table 1, the parameters necessary to establish the vertical layer assignments ($z_1$, $\Delta z_1$, $a$) in the three configurations and the number of layers up to some typical height are shown.

### 2.3 Test cases

#### 1) GABLS2 (diurnal change of cloud-free boundary layer)

The simulation starts with a fully developed mixed layer at midday. The diurnal change is realized through forced ground temperature. The maximum and minimum ground temperatures are about 292 K and 272 K, respectively. During the daytime, the ground temperature changes in a form of a sine curve, while its decrease during the night is represented by a linear function.

#### 2) EUROCS (stratocumulus-capped boundary layer)

The initial condition displays a sharp inversion at a height of around 600m above the surface. The average jump across the inversion is given by $\Delta\theta_l = 12$ K and $\Delta q_w = -3.0$ g kg$^{-1}$. The long radiative upward flux is imposed as a function of the total water path, and a simple shortwave radiation scheme is employed. The forced horizontal divergence in the original configuration is instructed as $1 \times 10^{-5}$ s$^{-1}$, which leads to the subsidence rate $w = -1 \times 10^{-5} z$ m s$^{-1}$ with the height above the surface $z$. The forced surface fluxes of heat and moisture are also given.

## 3 Results

### 3.1 GABLS2

Fig. 1 shows vertical profiles of potential temperatures in the GABLS2 experiments with a number of vertical layer assignments. In the case of the fully developed mixed layer shown in Fig. 1 (left), while grid configuration (A) generates a mixed layer that is a little too shallow

Table. 1: Specification of vertical layer assignments prepared for the experiments (#: number)

| Grid configuration | (A) | (B) | (C) |
|---|---|---|---|
| Height of the lowest layer $z_1$[m] | 20 | 5 | |
| First grid spacing increment $\Delta z_1$ [m] | 60 | 16.6 | 11 |
| common difference $a$[m] | 40 | 6.6 | 1 |
| # of layers up to 1,000 m | 8 | 17 | 37 |
| # of layers up to 2,000 m | 10 | 23 | 55 |
| # of layers up to 3,000 m | 13 | 30 | 69 |

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at a glance as it does not have sufficient grid spacing to resolve the mixed layer adequately, the heights of the mixed layers for the grid configuration (B) and (C) are mostly identical. A similar feature can also be seen in the stable boundary layer as presented in Fig. 1 (right). These results imply that the vertical layer assignment with 30 layers below a height of 3,000m is enough to represent mixed and stable layers with no clouds.

### 3.2 EUROCS

While no significant differences are seen between grid configurations (B) and (C) in GABLS2, the test case of a boundary layer capped by stratocumulus reveals another dependency on vertical resolution. Vertical profiles of potential temperature with the three vertical layer assignments, including an LES result obtained from the project web site[3], are displayed in Fig. 2, which shows that coarser vertical resolution result in higher inversion. This is not just a problem in terms of the height of the inversion; the higher inversion height implies that too much air is entrained across the inversion layer. Since this test case imposes subsidence, as mentioned above (and it is often true in layers with stratocumulus), the inversion would be simply be lowered by the subsidence. However, the artificial vertical gradients of potential temperature that appear due to coarse resolution generate numerical, i.e. not genuine vertical mixing. Lenderink and Holtslag (2000) call this "numerical entrainment". In order to reduce the numerical entrainment, higher vertical resolution is necessary as shown in Fig. 2.

## 4 Conclusion and Discussion

The GABLS2 experiments confirmed that extremely high vertical resolution is not necessary to properly represent generation of mixed and stable boundary layers in diurnal changes. However, the model run with EUROCS, the case with stratocumulus, revealed that higher vertical resolution is desirable to reduce the false entrainment when TKE-based closure is adopted.

Since the computational resources available in operational NWP systems are fairly limited, it is impossible to secure as many vertical layers as needed even though the experiments showed that higher resolution is desirable. However, the results of the experiments provide an important guide in determining the assignment of vertical layers. They suggest that grid configuration (B) is enough for cloud-free boundary layer, and that it behaves better than a coarser one even in cloudy layers.

JMA plans to enhance vertical resolution in the global and mesoscale models. In addition to assisting with the examination of how vertical layers should be placed, single column models are expected to be useful in determining the performance of physical processes with enhanced vertical resolution.


## Acknowledgment

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[3]http://www.phys.uu.nl/~wwwimau/research/atm_dyn/EUROCS_PART_I/eurocs.html

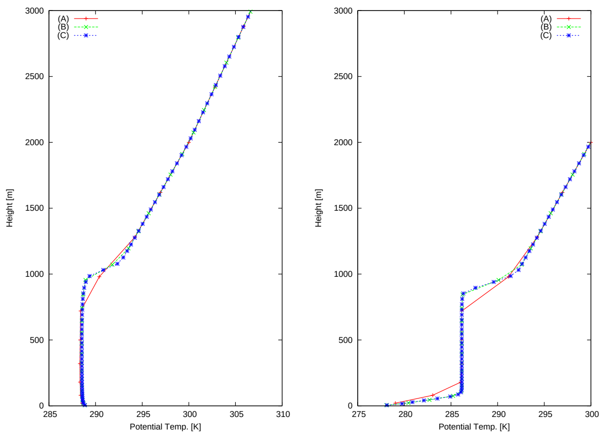


Fig. 1: Vertical profiles of potential temperature in the GABLS2 experiments. Left: when the mixed layer is fully developed; right: when the stable layer is evolved. Each line corresponds to grid configurations (A), (B) and (C) shown in Table 1.

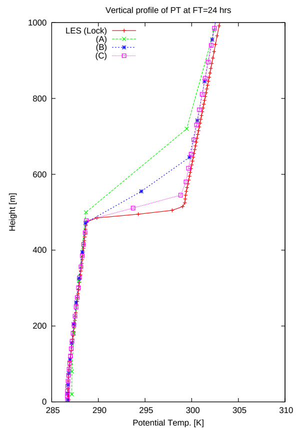


Fig. 2: Vertical profiles of potential temperature in the EUROCS experiments (24 hours after the initial time). Each line corresponds to LES (by A. Lock, obtained from the project web site), the single column model with grid configurations (A), (B) and (C) shown in Table 1.


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# Turbulence length scale formulated as a function of moist Brunt-Väisälä frequency

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In the scales simulated by present NWP models all the 3D turbulent eddies are sub-grid. Still the higher vertical resolution of such models allows physically realistic parametrization of the most energetic eddies at every model level. This is usually done through the concept of turbulence length scale, becoming the only free parameter closing the turbulence scheme. It is then quite evident that the quality of a whole turbulence parametrization is strongly related to the way the length scale is formulated. In the world of NWP models, aiming to deliver realistic simulation for any specific weather type, a unique and general formulation of mixing length is always of great advantage. In this sense the formulation proposed by Bougeault and Lacarrère (1989) (hereafter BL89) seems to deliver a very attractive solution for mesoscale models by offering physically sound results regardless of any specific atmospheric stratification. In their approach the length scale $L$ is related to the two characteristic lengths $l_{up}$ and $l_{down}$ obtained as a distance that a parcel originating at a given level $z$ with an initial kinetic energy equal to the mean TKE of the layer $e(z)$ can travel upwards and downwards before being stopped by buoyancy effects. This can be mathematically expressed as:

$$\int_{z}^{z+l_{up}} \frac{g}{\bar{\theta}_v}\left(\bar{\theta}_v(z') - \bar{\theta}_v(z)\right) dz' \quad = e(z) \tag{1}$$

$$\int_{z+l_{down}}^{z} \frac{g}{\bar{\theta}_v}\left(\bar{\theta}_v(z) - \bar{\theta}_v(z')\right) dz' \quad = e(z), \tag{2}$$

with $\bar{\theta}_v$ and $g$ being virtual potential temperature and gravity constant. In this respect the method offers a non-local length scale affected not only by stability at a given level but also influenced by remote stable zones. Another interesting feature of this method is that when (1) and (2) are evaluated by a second-order accuracy algorithm the length scales in uniformly stratified atmosphere converge toward the well known Deardorff (1980) length scale given as

$$l_{up} = l_{down} = \sqrt{\frac{2e}{N^2}} \quad . \tag{3}$$

This can be then interpreted that the BL89 length scale is in a way a non-local generalization of the Deardorff mixing length (Cuxart et al., 2000).

In the framework of the newly developed TOUCANS (Third Order moments scheme with Unified Condensation Accounting and N-dependent Solver) turbulence scheme the moist processes are treated through modified Richardson number $Ri''$. For consistency reasons it is then desirable to formulate the length scale in terms of (moist) Brunt-Väisälä frequency $N_v^2$ related to the $Ri''$ through the shear $S^2$: $N_v^2 = Ri''S^2$. Keeping the same discrete formulation the equations (1) and (2) can be formally rewritten to desirable form using directly $N_v^2$:

$$\int_{z}^{z+l_{up}} N_v^2(z'-z)dz' \quad = e(z) \tag{4}$$

$$\int_{z+l_{down}}^{z} N_v^2(z-z')dz' \quad = e(z). \tag{5}$$

This reformulation, on top of being consistent with the $Ri''$ concept, also allows to replace the costly and complicated second-order accurate evaluation of the path within the last model level that a particle is reaching by the desired Deardorff relation (3) using the $N_v^2$ of the actual level and the hypothesis that the remaining energy is fully spent by that displacement. This potentially makes the new method attractive also for the linearized formulation in the TL/AD models. In addition this new formulation of BL89 length scale reduces almost twice the computational cost, provided the $Ri''$ is already evaluated earlier in the scheme.

Part of this work was performed in the framework of the CzechGlobe project. Supports of the RC LACE Consortium as well as those of Météo-France are gratefully acknowledged.

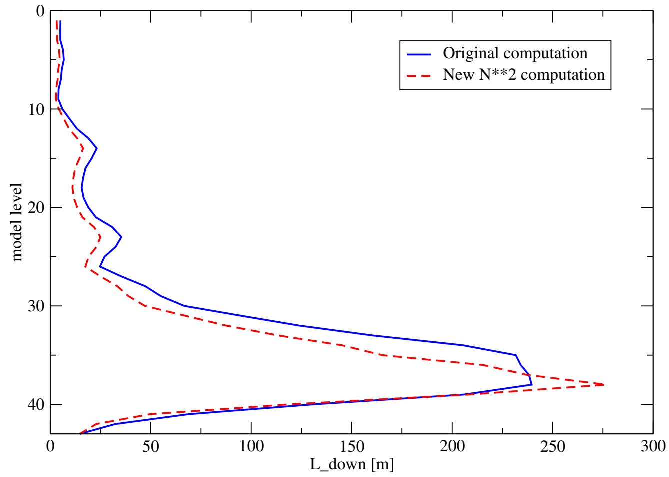


Figure 1: The downward displacement $l_{down}$ evaluated for a randomly chosen model column by original BL89 method (blue full line) and the new method relating length scale directly to the moist Brunt-Väisälä frequency $N_v^2$ (dashed red curve). Although the two methods share a nearly equivalent discrete formulation the difference between the two curves originates in the way how the $N_v^2$ is being evaluated. In the latter case $N_v^2$ is directly derived from the $Ri''$, while the former formulation is evaluated independently of this key parameter of the TOUCANS turbulence scheme.

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# Forecasting precipitation caused by slantwise convection

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The occurrence of convective snowstorms has been and continues to be a significant forecasting problem for operational meteorologists. This paper presents a case study of a convective winter snowfall that occurred on December 7, 2009 at Moscow region (Russia) and produced 11 cm of snow or 11 mm of water per 24 hours. This snowfall occurred in high pressure region. National Meteorological Service forecasted light precipitation (1 – 1.5 mm per 24 hours). Numerical weather prediction models from the UKMO and NCEP forecasted 1.5 and 2.5 mm, correspondingly. Operational isentropic analysis was used in the course of the case study. Conditional Symmetric Instability (CSI) can be described as instability that arises from the release of latent heat during the slantwise ascent of air parcel [2] ( Fig,1).

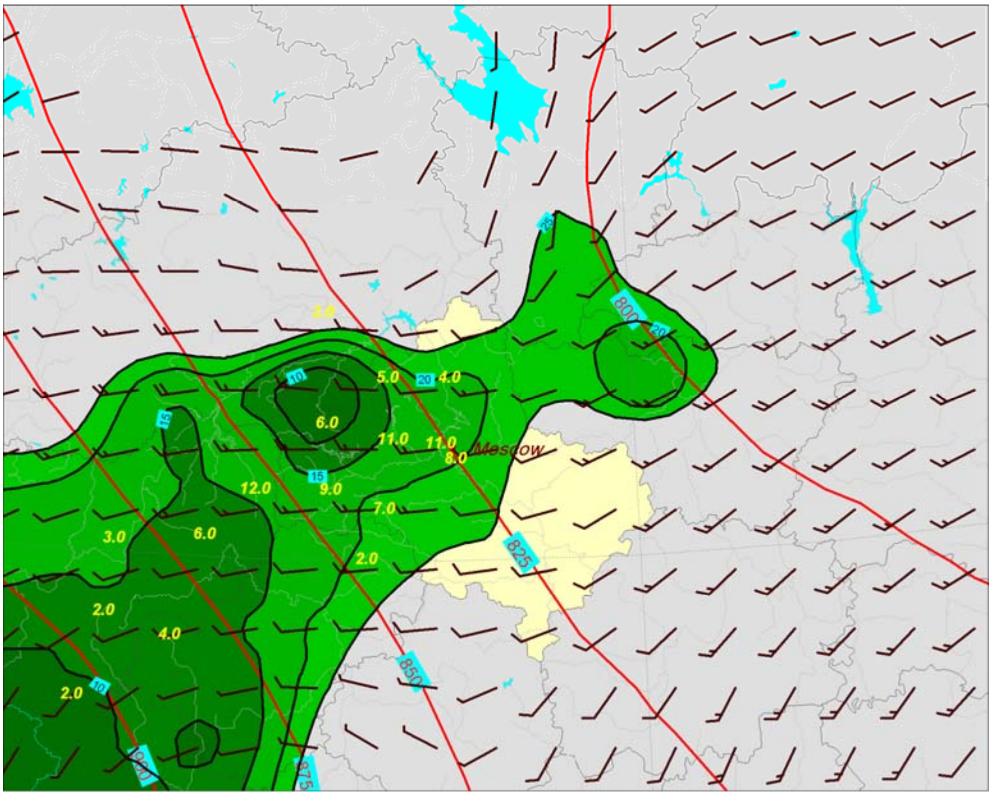


Fig. 1. Pressure on 285°K isentropic level (red), wind and region (green) where the deviation from condensation level ≤25 hPa. Yellow figures - daily precipitation according to obs data.

Method of diagnosing CSI is through the analysis of equivalent potential vorticity (EPV) or moist potential vorticity (MPV) [3]. $EPV = - g\ \eta \cdot \nabla\ \theta_e$, where $\eta$ - absolute vorticity, $\theta_e$ – equivalent potential temperature. If EPV < 0 and the atmosphere is 'saturated' ( RH > 80%), then CSI is present [2] (Fig. 2).

Convective vertical velocity can be found using slantwise convective available potential energy (SCAPE): $\omega_c^2$ = SCAPE.

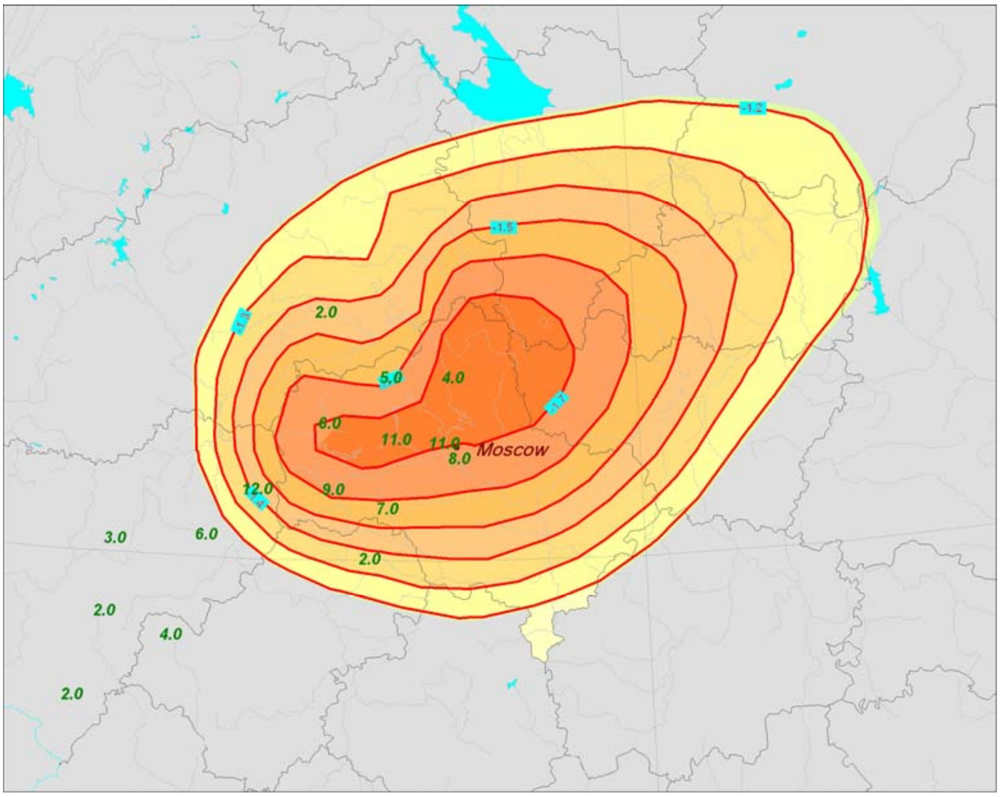


Fig. 2. EPV on 285°K isentropic level on 7.12.2009 00 UTC, predict on 18 hours.

To define SCAPE by [1] it is necessary to calculate additional slantwise correction ($\Delta$T):

$$\Delta T = \frac{1}{2}\,\frac{T\upsilon}{g}\,\frac{f}{\Omega_a}\,\frac{d\,[(\upsilon - \upsilon 0)^2]}{d\,z}$$

where $T_\upsilon$ – mean virtual temperature, $f$– Coriolis parameter, $\upsilon$ and $\upsilon_0$ – wind speed on height z and $z_1$, g – gravitational acceleration, $\Omega_a$ - absolute vorticity, z – height. Further one has to add $\Delta$T to the temperature of the lifted parcel [1]. $\Delta T_{925\text{-}850} = 264.4 \cdot (12 - 2)^2 / (2 \cdot 9.8 \cdot 650) = 2.03$. SCAPE = 76 J $kg^{-1}$, $\omega_k = 8.71$ m/s. For convective precipitation intensivity calculation it is necessary to find mean vertical velocity in Cb cloud using the following empirical formula: $\omega = 0.33\ 10^{-4}\ (m\ \omega_c\ h)$, where m – convective-instable layer depth (hPa), h – convection power (hPa). Using m = 85, h = 250 and $\omega_c = 8.71$, we find: $\omega = 6.1$ cm/s = -23.8 hPa/h. Convective precipitation intensity: $I_c = 1.5\ \Delta q_{850} + 3.0\ \Delta q_{700} = 1.8$ mm/h.
In accordance with [1], time scale for convective precipitation caused by slantwise convection equals approx. $1/f$ and makes 2 - 4 hours. So quantity of precipitation caused by slantwise convection in Moscow region made 6 -8 mm and amount of precipitation – both convective and continuous – made approx. 9 mm per 12 hours.

References

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