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Effect of soil hydraulic parameters on the local convective precipitation

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Abstract

The impact of soil hydraulic parameters upon the formation of local convective precipitation is investigated by using the Penn State-NCAR MM5 Modeling System. Four soil hydraulic parameter effects are considered. These comprise the effect of two different soil datasets (US vs. Hungarian), the effects of different parameterizations of field capacity and the wilting point soil moisture contents and the subgrid-scale variability effect of soil moisture content, respectively. Precipitation from 6 convective events were simulated and verified on a spatial scale of 18 km by comparison with observed precipitation fields. The latter were analysed from a total of 657 precipitation gauges using ordinary block kriging. It is demonstrated that the simulated convective precipitation is sensitive to the used soil dataset. The rank order correlation coefficient of $R_s = 0.62$ (explained variance 38 %) between observed precipitation and simulations using the Hungarian soil dataset (control run) is higher than those using the US soil dataset ($R_s = 0.53$, explained variance 29 %). The difference of about 10 % explained variance is highly significant ($p < 0.01$). Simulations with alternatively field capacities result in a correlation of $R_s = 0.60$, which is not significantly different from the control run. Simulations with alternatively wilting points, however, result in $R_s = 0.56$, which is significantly lower than the correlation for the control run ($p < 0.05$). Finally, the application of an alternatively spatial distribution of soil moisture contents results again in a lower correlation of $R_s = 0.54$ ($p < 0.01$). Amongst the four soil effects, the soil dataset effect and the subgridscale variability of the soil moisture were greater than the parameterization effects. The results suggest that regional soil datasets should be preferred in the mesoscale models. It is also obvious that high-resolution global soil data sets in the climate models should be constructed extremely carefully because of the sensitivities obtained.

Zusammenfassung

Den Einfluss hydraulischer Eigenschaften des Bodens auf lokale, konvektive, Niederschläge wurde mit dem Penn State-NCAR MM5 Modellsystem untersucht. Vier Effekte zufolge der hydraulischen Eigenschaften des Bodens wurden untersucht. Dazu gehören der Effekt durch Verwendung von zwei unterschiedlichen Bodendatensätzen (US vs. Ungarisch), die Effekte unterschiedlicher Parameterisierungen der Feldkapazität und der Bodenfeuchte des Welkpunktes sowie der Effekt infolge der subskaligen Variabilität der Bodenfeuchte. Niederschlagsfelder von 6 konvektiven Ereignissen wurden simuliert und auf einer räumlichen Skala von 18 km durch Vergleich mit beobachteten Niederschlagsfeldern verifiziert. Letztere wurden aus 657 Punktmessungen mittels *Ordinary Block-Kriging* analysiert. Die simulierten, konvektiven, Niederschäge reagierten sensitiv auf die verwendeten Bodendaten. Im Vergleich mit dem beobachteten Niederschlag ist der Rangkorrelationskoeffizient von $R_s = 0.62$ (erklärte Varianz 38 %) des mit dem ungarischen Bodendatensatz simulierten Niederschlags (Kontrolllauf) höher als der des mit dem US Bodendatensatz simulierten Niederschlags ($R_s = 0.53$, erklärte Varianz 29 %). Die Differenz von ca. 10 % erklärter Varianz ist hoch signifikant ($p < 0.01$). Simulationen mit alternativen Feldkapazitäten ergaben eine Korrelation von $R_S = 0.60$, die sich nicht signifikant vom Kontrolllauf unterscheidet. Hingegen resultierten Simulationen mit alternativen Welkpunkten in einer Korrelation von $R_s = 0.56$, die signifikant unter der des Kontrolllaufes liegt ($p < 0.05$). Abschließend resultierte auch die Verwendung alternativer räumlicher Verteilungen des Bodenfeuchtegehalts in einer geringeren Korrelation von $R_s = 0.54$ ($p < 0.01$). Von den vier Effekten erwiesen sich die unterschiedlichen Bodendatensätze sowie die subskalige Variabilität der Bodenfeuchte als bedeutender als die Verwendung alternativer Parameterisierungen. Die Ergebnisse weisen darauf hin, dass regionalen Bodendatensätzen in mesoskaligen Modellen der Vorzug gegeben werden sollte. Zufolge der gezeigten Sensitivitäten sollten auch hochauflösende Bodendatensätze für Klimamodelle mit Sorgfalt erstellt werden.

1 Introduction

Soil hydraulic characteristics determine all basic elements of the terrestrial hydrological cycle. Surface runoff is highly sensitive to soil texture and the parameterization of soil water conductivity function (WARRACH-SAGI et al., 2008). Similar sensitivity is also found for evapotranspiration (e.g. EK and CUENCA, 1994; CUENCA et al., 1996; MÖLDERS et al., 2005) and precipitation (e.g. PAN et al., 1996; KOSTER et al., 2003). For all these processes, the soil-precipitation relation-

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ship is of high importance (HOHENEGGER et al., 2009). This relationship is highly nonlinear and its analysis in the treatment of convection is unavoidable. Precipitation formation in convective clouds is affected by characteristics of both the atmosphere and the land surface. The atmospheric influence is significant via triggering (FINDELL and ELTAHIR, 2003; JUANG et al., 2007), instability (ELTAHIR and PAL, 1996) and the interaction between cloud and environmental air (BETTS, 1992; EK and HOLTSLAG 2004). The land surface effect can also be significant in some cases. This effect is mostly regulated by hydraulic properties of the soil-vegetation system (HORVÁTH et al., 2009). There have been some observational (TAYLOR and LEBEL, 1998; TAYLOR and ELLIS, 2006; KRAUSS et al., 2008) and modeling (e.g. PAN et al., 1996; EMORI, 1998; HORVÁTH et al., 2007; GAO et al., 2008) efforts to quantify the soil hydraulic - convective precipitation relationship, but a comprehensive study on this subject is still lacking.

The aim of this study is to analyze the soil hydraulic parameters – convective precipitation relationship. Soil hydraulic parameters are highly uncertain. They depend strongly upon soil texture and the soil database used. The former dependence is well known (e.g. MIHAILOVIC et al., 1992), the latter less so (OSBORNE et al., 2004). At the same time, the latter dependence seems to be reasonable, knowing that particle-size limits and soil textural classes are different in different soil taxonomies (e.g. FILEP and FERENCZ, 1999). This is remarkably well shown by NEMES et al. (2005) comparing Hungarian, European and USA soil characteristics. Some notes on this subject can also be found in the work of HORVÁTH et al. (2009). Besides this, there is also large uncertainty in the field capacity (θ_f) and wilting point (θ_w) soil moisture contents. They depend not only on soil texture, but also on the parameterization of soil hydraulic functions (soil moisture potential (Ψ) and hydraulic conductivity (K)) and the criteria applied in their estimation procedure (ÁCS, 2005). Lastly, the impact of microscale variability of soil moisture content (θ) (BELL et al., 1980) upon evapotranspiration should be also considered as accurately and simply as possible (e.g. ÁCS and SZÁSZ, 2002). To date we do not know which soil hydraulic characteristics are the most relevant regarding convective precipitation formation. Our aim is to establish a "sensitivity hierarchy", if any exists.

This aim is pursued by performing sensitivity studies and evaluating the results against observed data using statistical tools. Precipitation fields are estimated from rain-gauge data using kriging interpolation. The agreement between different precipitation fields (simulated versus observed and simulated versus simulated) was evaluated by calculating Spearman's rank order correlation coefficients. Significance tests are also carried out to estimate how important the obtained differences are. Model runs are performed for six one-day case studies during the summer in 2006 and 2007 in Hungary. For the simulations the Penn State - NCAR MM5 modeling system was applied.

2 Theory

2.1 The mesoscale weather prediction modell

Numerical simulations were performed by Version 3 of the Penn State - NCAR MM5 (Fifth-generation Mesoscale Model) modeling system (DUDHIA, 1993). The model applies a terrain-following sigma coordinate system. The predictive variables are: pressure perturbation, the three momentum components, temperature, specific humidity and the mixing ratio of the different types of hydrometeors (cloud water, cloud ice, rain, snow and graupel particles). The partial differential equation system is solved by using a relaxation lateral boundary condition and a radiation upper boundary condition. Model runs were performed using a horizontal resolution of 6 x 6 km, and 26 vertical levels.

Grell's scheme (GRELL et al., 1994) was applied for the parameterization of convection based on the rate of destabilization or on quasi-equilibrium. A simple single-cloud scheme with updraft and downdraft fluxes and compensating motion determining the heating/moistening profile was used. The formation of cloud and precipitation elements is simulated with an explicit bulk microphysical scheme (REISNER et al. 1998) with five different types of hydrometeors: cloud water, cloud ice, rain, snow and graupel particles. The collision coalescence processes between different types of hydrometeors, furthermore the diffusion of vapor, the freezing of liquid elements and the melting of ice particles are simulated. The equation of conservation is not only solved for the mixing ratios of hydrometeors but for the concentration of cloud ice as well.

The planetary boundary layer (PBL) is described by the local mixing PBL scheme predicting turbulent kinetic energy based on JANJIĆ's (1990, 1994) work. Compared with other non-local or high-order closure schemes, this PBL scheme proved to be more efficient, because it needs less computer capacity. Exchange coefficients are needed in order to calculate vertical fluxes with an implicit diffusion scheme. These are calculated by the land surface scheme presented briefly below.

2.2 The land-surface model

Land-surface processes are simulated by the Noah LSM (the successor of the Oregon State University Land-surface Model). The model consists of a multilayer soil model (MAHRT and PAN, 1984), a single-layer snow model (CHEN and DUDHIA, 2001) and a canopy model (PAN and MAHRT, 1987). Evapotranspiration of the soil-plant system is estimated as the sum of transpiration and evaporation from the soil and the wetted parts of the vegetation. Both transpiration and soil evaporation depend on θ_f and θ_w. Soil hydraulic functions are parameterized after CAMPBELL (1974) as follows:

$$\Psi(\theta) = \Psi_S \cdot \left(\frac{\theta}{\theta_S}\right)^{-b} \tag{2.1}$$

and

$$K(\theta) = K_S \cdot \left(\frac{\theta}{\theta_S}\right)^{2b+3}, \tag{2.2}$$

where Ψ_S is the saturated soil moisture potential [m water column], K_S is the saturated hydraulic conductivity [m s^{-1}], θ_S is the saturated soil moisture content [m^3 m^{-3}] and b is the pore size distribution index. In calculating evapotranspiration the inhomogeneous areal distribution of θ is supposed. This is simply taken into account by specifying θ_f and θ_w (see section 3.2.2). A detailed description of the model can be found in CHEN and DUDHIA's (2001) work.

2.3 Statistical methods

The irregularly spaced precipitation gauge data were analyzed on a regular verification grid with a spatial resolution of about 18 km. Ordinary block kriging was used for the statistical analysis. This method estimates the areal precipitation from a weighted mean of observations. The statistically optimal determination of the weighting factors follows from a linear system of equations containing spatial auto-correlation functions (RUBEL and HANTEL, 2001; RUBEL and BRUGGER, 2009).

The Wilcoxon signed rank significance test (WILCOXON, 1945; WILKS, 1995) was applied to show whether simulated precipitation field differences were significant. The Wilcoxon test is a non-parametric test so there is no restriction on the distribution of the data. The only requirement is that the number of samples above and over the median have to be roughly the same, which is usually fulfilled in the case of precipitation. The null hypothesis in both analyzes is that the two sample spreads have the same mean. If the null hypothesis is rejected, it means that the two sample spreads are significantly different.

The agreement between simulated and observed precipitation fields was estimated using the rank order correlation. Like all ranked correlations, this estimation is also non-parametric giving the freedom of not relying on the distribution of the samples. Since the probability density function of 24 h sum of precipitation can be generally characterized as lognormal or gamma, the use of Spearman's correlation estimation is valid. The null-hypothesis is that the two compared samples do not differ significantly. In the presented statistical analyses the significance level denotes the probability at which we can accept the null-hypothesis.

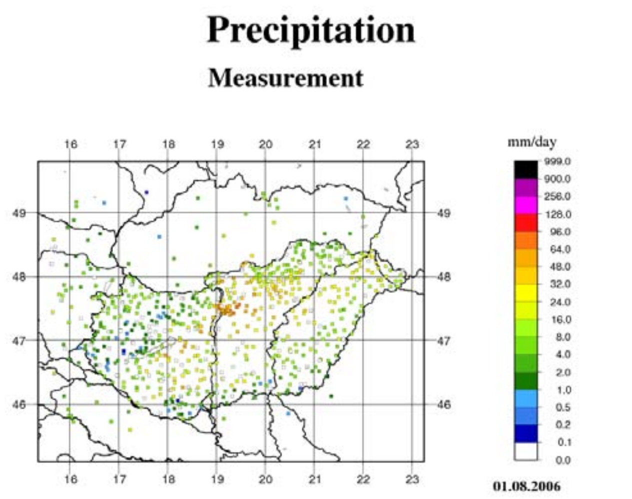

Figure 1: Station positions and measured precipitations together with the model domain on 1st August 2006.

Fisher's r to z transformation (FISHER, 1915, 1921) was applied to show whether the differences between the rank order correlation coefficients were significant. The transformation "normalizes" the Spearman coefficients, as it would have been calculated from a normally distributed sample resulting in the "z" value. Afterwards the level of significance is determined from "z" by defining the value corresponding to the function of normal distribution (ZELEN and SEVERO, 1972).

3 Data

Strong local convective storms with developed land-surface/air interactions were studied. The following 6 days during the period 2006–2007 were selected: 1st August 2006, 7th August 2006, 5th May 2007, 22th May 2007, 1st June 2007 and 20th August 2007.

3.1 Initializations

The initial conditions for the MM5 model run were obtained from the ECMWF analysis and the forecast. For all days 00 UTC data were used. For the top level of the MM5 model the 100 hPa layer was chosen. Initial soil data (temperature and soil moisture) were also taken from the ECMWF analysis.

3.2 Basic land-surface parameters

In this study, the USGS-24 category land use dataset was applied. In the Carpathian Basin, the prevailing vegetation type is type 2, "Dryland Cropland and Pasture". Towards mountain regions the "Grassland" type increases. In the Carpathian mountains, "Deciduous Broadleaf Forest" prevails, but there is also the "Mixed Forest" type. In all runs, that areal distribution of soil texture was

Table 1: USA and Hungarian soil hydraulic parameters as used in the Noah LSM. b = pore size distribution index, Ψ_S = saturated soil water retention, K_S = saturated soil water conductivity, θ_S = saturated soil moisture content.

Soil texture	b		Ψ_S (m)		K_S (m s^{-1})		θ_S (m^3 m^{-3})	
	USA	HU	USA	HU	USA	HU	USA	HU
1) Sand	2.79	3.02	0.069	0.06	$4.60 \cdot 10^{-5}$	$3.26 \cdot 10^{-5}$	0.339	0.507
2) Loamy sand	4.26	3.9	0.036	0.126	$1.41 \cdot 10^{-5}$	$2.52 \cdot 10^{-5}$	0.421	0.598
3) Sandy loam	4.74	3.99	0.141	0.143	$5.23 \cdot 10^{-6}$	$1.14 \cdot 10^{-5}$	0.434	0.476
4) Silt Loam	5.33	4.18	0.759	0.182	$2.81 \cdot 10^{-6}$	$2.73 \cdot 10^{-6}$	0.476	0.487
5) Silt	5.33	3.54	0.759	0.223	$2.81 \cdot 10^{-6}$	$2.00 \cdot 10^{-6}$	0.476	0.496
6) Loam	5.25	4.2	0.355	0.224	$3.38 \cdot 10^{-6}$	$4.58 \cdot 10^{-6}$	0.439	0.468
7) Sandy Clay Loam	6.66	4.21	0.135	0.132	$4.45 \cdot 10^{-6}$	$7.98 \cdot 10^{-6}$	0.404	0.439
8) Silty Clay Loam	8.72	5.04	0.617	0.234	$2.04 \cdot 10^{-6}$	$6.20 \cdot 10^{-7}$	0.464	0.491
9) Clay Loam	8.17	4.74	0.263	0.207	$2.45 \cdot 10^{-6}$	$3.05 \cdot 10^{-6}$	0.465	0.58
10) Sandy Clay	10.73	3.58	0.098	0.89	$7.22 \cdot 10^{-6}$	$4.58 \cdot 10^{-6}$	0.406	0.5
11) Silty Clay	10.39	4.06	0.324	0.324	$1.34 \cdot 10^{-6}$	$1.05 \cdot 10^{-6}$	0.468	0.453
12) Clay	11.55	6.21	0.468	0.228	$9.74 \cdot 10^{-7}$	$8.00 \cdot 10^{-7}$	0.468	0.541

Table 2: A brief description of the conditions used in the runs in this study.

	Data samples	Definition of θ_f	Definition of θ_w	Homogenity of θ
Run 1	VÁRALLYAY (1973) FODOR and RAJKAI (2005) NEMES (2002)	pF = 2.3	pF = 4.2	inhomogeneous
Run 2	COSBY et al. (1984)	K = 0.5 mm day^{-1}	pF = 4.2	inhomogeneous
Run 3	VÁRALLYAY (1973) FODOR and RAJKAI (2005) NEMES (2002)	K = 0.5 mm day^{-1}	pF = 4.2	inhomogeneous
Run 4	VÁRALLYAY (1973) FODOR and RAJKAI (2005) NEMES (2002)	pF = 2.3	$\theta_w = 5 \cdot \varrho_b \cdot$hy	inhomogeneous
Run 5	VÁRALLAY (1973) FODOR and RAJKAI (2005) NEMES (2002)	pF = 2.3	pF = 4.2	homogeneous

used which was originally implemented in the MM5. So, in Hungary there are six textural classes from the twelve presented in Tab. 1. These are as follows: loamy sand, sandy loam, loam, sandy clay loam, clay loam and clay. Nevertheless, we used two soil datasets: the well known one from USA and a Hungarian. Since they are quite different, they shall be briefly compared below.

3.2.1 Soil data sets

The USA dataset (hereafter US dataset) is based on the data gathered from about 1400 samples in the USA (COSBY et al., 1984). The Hungarian dataset (hereafter HU dataset) comprises soil data from 576 samples collected in Hungary (VÁRALLYAY, 1973; NEMES, 2002; FODOR and RAJKAI, 2005). In the two datasets different particle-size limits and different methods for determining soil texture classes are used. In the US dataset, USDA particle-size classification, while in the HU dataset by ISSS (**I**nternational **S**oil **S**cience **S**ociety) the accepted Atterberg's classification is used. They differ only in silt particle-size limits. Silt particle-size range is 0.002–0.05 mm for USDA and 0.002–0.02 mm for Atterberg's classifcation. Besides this, in the HU dataset a simple, empirically based method (FILEP and FERENCZ, 1999) is applied for determining soil textural classes. Because of these differences, the soil textural triangles for US and HU soils are also quite different. This is presented in Fig. 2. One basic difference is that the sandy clay and silty clay textural classes do not exist in the HU dataset. It is also interesting that the overlap between the Hungarian clay loam and the USA clay is quite large. Note the shift between the USA and the Hungarian loam. The best overlap is obtained for sand. Of course, there are also noticable differences in the distribution of the soil samples across the soil textural triangle. This is presented in Fig. 3. The Hungarian samples belong mainly to the silt loam and silty clay loam textural classes. In spite of this, the USA samples belong mainly to the sand and loamy sand textural classes. It has to be mentioned that in the HU dataset there are only a few sandy clay loam and sand samples. These soil data set differences together with all other methodological, geological, pedological and climate differences produce also differences in the soil hydraulic properties. These properties (saturated soil moisture content θ_S, saturated

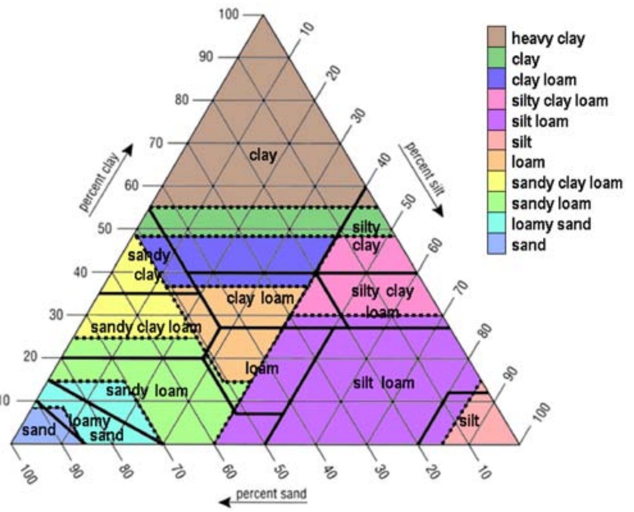

Figure 2: Soil textural triangle used in the United States and in Hungary. Hungarian soil texture classes are denoted with colours, while the US with their names. Soil texture classes separated by solid lines refer to US, while by dashed lines to HU dataset.

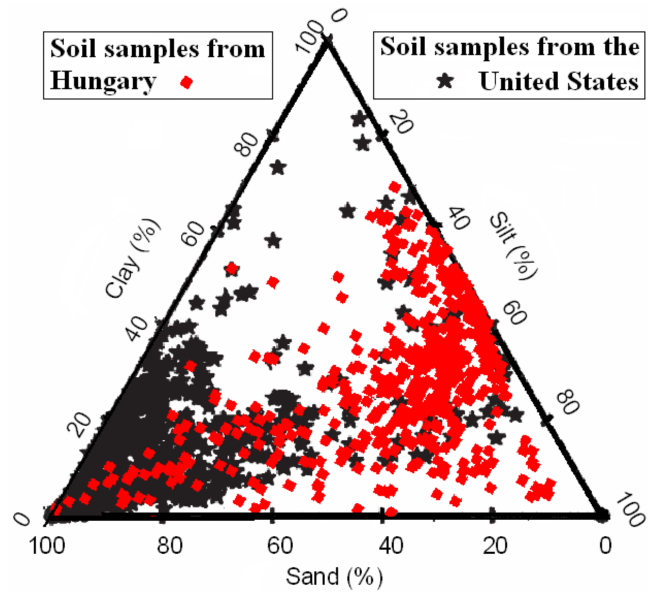

Figure 3: Distribution of soil samples across the soil textural triangle.

soil moisture potential Ψ_S, saturated water conductivity K_S and pore size distribution index b) for all 12 soil textures separately for the US and HU datasets are given in Table 1. It must be said immediately that it is impossible to say what is the relative contribution of any the aforementioned effects in creating the differences. The systematic difference between the US and HU θ_S and b values is obvious. Note that soil hydraulic properties for the sandy clay and silty clay textural classes are also given. They were estimated after FODOR and RAJKAI's (2005) algorithm defining sand, silt and clay fractions as input. Because of these differences there are also systematic differences between the US and HU $\Psi(\theta/\theta_S)$ and $K(\theta/\theta_S)$ values. For the same θ/θ_S values Ψ^{HU} is systematically smaller, while K^{HU} is systematically larger than the corresponding Ψ^{US} and K^{US} values. Since the differences between the US and HU datasets are so large, in the following they will be referred to as US and HU soils.

3.2.2 Determination of θ_f and θ_w values

Field capacity and wilting point soil moisture contents shall be determined in two steps. In the first step, θ_f and θ_w will be determined assuming no subgrid scale variability of θ. These values will be denoted as θ_f^h and θ_w^h. In the second step, the effect of subgrid scale variability of θ upon evapotranspiration will be taken into account via calculating the θ_f^{inh} and θ_w^{inh} values.

Determination of θ_f^h and θ_w^h values

θ_f^h values are determined in different ways for the US and HU datasets. For the US dataset, θ_f^h is calculated from Eq.(2.2) using the following criterion:

$$K(\theta_f^h) = 0.5 \quad mm \cdot day^{-1}. \tag{3.1}$$

This criterion was also used in the work of CHEN and DUDHIA (2001). For HU dataset, θ_f^h is determined by measurements (VÁRALLYAY, 1973) using criterion

$$pF = log_{10}[\Psi(\theta_f^h)] = 2.3. \tag{3.2}$$

VÁRALLYAY (1973) developed a survey system for determination of soil moisture retention curves in the whole suction range (pF 0, 0.4, 1, 1.5, 2, 2.3, 2.7, 3.4, 4.2 and 6.1). Suction is measured by a special apparatus, while soil moisture content expressed in [m^3m^{-3}] by the gravimetric method. θ_S value is obtained by measuring θ for pF 0.

As in the former case, θ_w^h values are separately determined for the US and HU datasets. For the US dataset, θ_w^h is calculated from Eq.(2.1) using criterion

$$pF = log_{10}[\Psi(\theta_w^h)] = 4.2. \tag{3.3}$$

This criterion was also used in the work of CHEN and DUDHIA (2001). For the HU dataset, θ_w^h is determined in two ways. In the first one, θ_w^h is obtained by VÁRALLYAY's (1973) measurements using the same criterion as in Eq. (3.3).

θ_w^h can also be estimated after STEFANOVITS et al. (1999) as follows:

$$\theta_w^h = 5 \cdot \varrho_b \cdot hy, \tag{3.4}$$

where ϱ_b is the bulk density of the soil and hy is the hygroscopic coefficient. hy is determined by VÁRALLYAY's measurements, while ϱ_b is also measured in the laboratory.

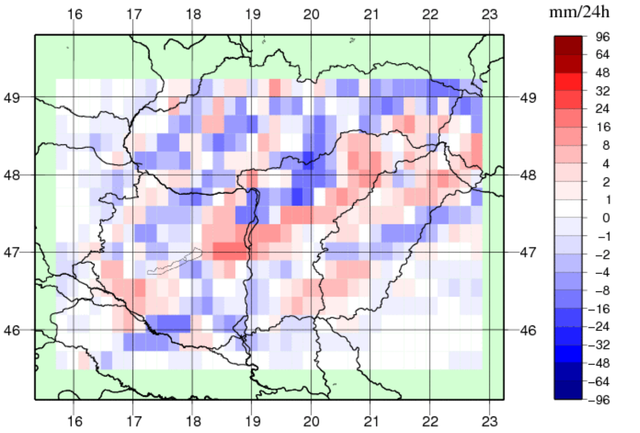

Figure 4: Areal distribution of 24-hours accumulated precipitation differences on 1st August, 2006 obtained using the conditions in Run 1 and Run 2 (Run 1–Run 2). Conditions are presented in Tab. 2.

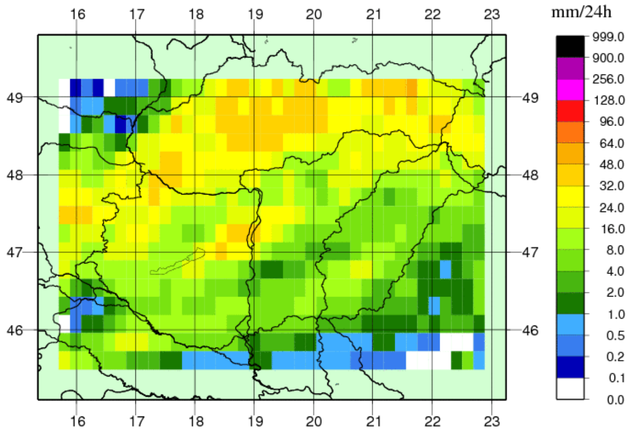

Figure 5: Areal distribution of 24-hours accumulated precipitation on 1st August, 2006 obtained using the conditions in Run 1. Conditions are presented in Tab. 2.

Determination of θ_f^{inh} and θ_w^{inh} values

Evapotranspiration from a land-surface with a homogeneous (E^h) and inhomogeneous (E^{inh}) distribution of θ is quite different (e.g. RONDA et al., 2002; WETZEL and CHANG, 1987). In the soil moisture range between θ_f and θ_w, the slope $\partial E^h(\theta)/\partial\theta$ is larger than the slope $\partial E^{inh}(\theta)/\partial\theta$. So, E^h is lower than E^{inh} in dry conditions (in the vicinity of θ_w) and, vice versa, E^h is larger than E^{inh} in wet conditions (in the vicinity of θ_f). Similarly it is obvious that θ_f^{inh} - $\theta_w^{inh} \gg \theta_f^h$ - θ_w^h (e.g. ÁCS, 2003). To obtain these relationships, both deterministic and statistical-deterministic evapotranspiration models have to be run. However, statistical-deterministic evapotranspiration models are not computationally efficient because Monte-Carlo simulations are needed for generating a randomly variable θ. This difficulty can be avoided using simple parameterizations for θ_f^{inh} and θ_w^{inh}. We used those formulae which agree well with the simulation results obtained by ÁCS and SZÁSZ (2002). These formulae were presented by CHEN and DUDHIA (2001),

$$\theta_f^{inh} = \frac{\theta_S}{3} + \frac{2}{3} \cdot \theta_f^h \tag{3.5}$$

and

$$\theta_w^{inh} = \frac{1}{2} \cdot \theta_w^h . \tag{3.6}$$

3.3 Observed Precipitation

In order to verify the simulated precipitation fields, daily accumulated data were analyzed from high resolution precipitation gauge observations. The gauge data

	1	3	5	7	9	11	13	15	17	19	Mean (1-9)
Run 1-2	3	3	3	1	1	0	0	0	0	1	2.2
Run 1-3	2	1	1	0	0	1	1	1	1	1	0.8
Run 1-4	4	3	3	2	0	0	0	0	0	0	2.4
Run 1-5	3	3	2	0	0	0	1	1	1	1	1.6

Thresholds (mm/day)

Figure 6: The number of those days when the 24-hours precipitation fields for different tests and thresholds were strongly significantly different (1% level) using the Wilcoxon signed rank test.

were provided by the Hungarian Meteorological Service (HMS) and comprise measurements from automatic gauges (29 synoptic and 68 on-line climate stations) and manual gauges (560 off-line stations). The location of the Hungarian precipitation stations as well as the synoptic stations in the neighbouring countries are depicted in Fig. 1. At all stations, the types of rain gauge similar to the traditional unshielded Hellmann gauge are operated. Therefore, the systematic underestimation of liquid precipitation, mainly caused by wind-induced, evaporation and wetting losses, is within the range of 0–10 % of the measured precipitation amount. It strongly depends on the wind speed and may be reduced as, for example, demonstrated by RUBEL and HANTEL (1999). Here, this error is neglected, because the differences between observed and simulated precipitation are 1–2 orders of magnitude higher.

To make sure that the observed precipitation represents the same temporal and spatial scale than the simulated precipitation, a 18 x 18 km^2 verification grid was

Table 3: Field capacity soil moisture content values for different soil textures in the runs used in the study.

Soil texture	θ_f (m^3 m^{-3})				
	Run 1	Run 2	Run 3	Run 4	Run 5
1) Sand	0.325	0.236	0.272	0.325	0.234
2) Loamy sand	0.479	0.283	0.348	0.479	0.419
3) Sandy loam	0.379	0.312	0.288	0.379	0.33
4) Silt Loam	0.408	0.36	0.316	0.408	0.368
5) Silt	0.437	0.360	0.312	0.437	0.408
6) Loam	0.406	0.329	0.296	0.406	0.375
7) Sandy Clay Loam	0.354	0.314	0.273	0.354	0.312
8) Silty Clay Loam	0.435	0.387	0.355	0.435	0.407
9) Clay Loam	0.479	0.382	0.387	0.479	0.429
10) Sandy Clay	0.340	0.338	0.33	0.413	0.260
11) Silty Clay	0.340	0.404	0.273	0.358	0.284
12) Clay	0.489	0.412	0.406	0.489	0.463

Table 4: Wilting point soil moisture content values for different soil textures in the runs used in the study.

Soil texture	θ_w (m^3 m^{-3})				
	Run 1	Run 2	Run 3	Run 4	Run 5
1) Sand	0.029	0.01	0.029	0.068	0.058
2) Loamy sand	0.08	0.028	0.08	0.141	0.16
3) Sandy loam	0.064	0.047	0.064	0.045	0.128
4) Silt Loam	0.08	0.084	0.08	0.043	0.16
5) Silt	0.069	0.084	0.069	0.027	0.138
6) Loam	0.088	0.066	0.088	0.044	0.176
7) Sandy Clay Loam	0.061	0.067	0.061	0.032	0.122
8) Silty Clay Loam	0.119	0.12	0.119	0.071	0.238
9) Clay Loam	0.139	0.103	0.139	0.081	0.278
10) Sandy Clay	0.055	0.1	0.055	0.056	0.110
11) Silty Clay	0.113	0.126	0.113	0.032	0.226
12) Clay	0.147	0.138	0.147	0.132	0.294

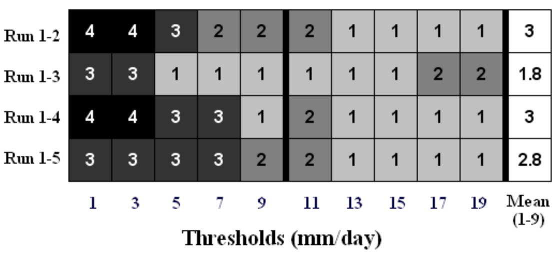

Figure 7: The number of those days when the 24-hours precipitation fields for different tests and thresholds were significantly different (5% level) using the Wilcoxon signed rank test.

defined. On average, about 2 precipitation gauges are located within each grid box. Using them as well as surrounding gauges, areal estimates were derived by applying the statistical interpolation procedure described above.

4 Experimental design

To be able to do a comparative analysis, five runs were performed. The conditions used in the runs are presented in Table 2. θ_f and θ_w values for each run and all 12 soil textures are presented in Table 3 and 4. Run 1 is the reference run. In this run HU soil parameters were used, inhomogeneous areal distribution of θ was supposed, the θ_f and θ_w values were calculated from the pF = 2.3 and pF = 4.2 conditions, respectively. Run 2 differs from Run 1 in that USA soil parameters were used and the θ_f values were calculated from the K = 0.5 mm day^{-1} condition. Run 3 and Run 1 are different only in the parameterization of the θ_f parameter. In Run 3, θ_f was calculated from the K = 0.5 mm day^{-1} condition. Run 4 differs from Run 1 only in the parameterization of the θ_w parameter. In Run 4, θ_w was calculated from the hygroscopic coefficient (see Eq. (3.4)). Run 5 differs from Run 1 only in the assumption of the areal distribution of θ. In Run 5, the homogeneous areal distribution of θ was presumed. This is represented by lower θ_f and higher θ_w values with respect to θ_f and θ_w values supposing inhomogeneous areal distribution of θ.

In all cases, the model was started at 00 UTC. The model was running for 30 hours to ensure a 6 hour spin-up time. 24-hour accumulated precipitation fields were obtained from +6 and +30 hour accumulated precipitation fields. The domain considered has 115 x 49 grid points. It covers the Carpathian Basin and some parts of the Carpathian mountains (the coordinates of the domain's lower left and upper right points are 45.6° N/15.6° E and 49.3° N/22.8° E, respectively).

5 Validation and comparisons

5.1 Validation results

The model is validated using the results of the reference run (Run 1) and the rain gauge measurements in Hungary. For validation a coarser 18 x 18 km horizontal resolution is used to achieve an approximate number of rain gauges per grid point (RUBEL and RUDOLF, 2001). The relationship between the simulated and the observed daily precipitation values is quantified by using rank order correlation coefficients. All grid point values within the boundary of Hungary for all six days (n=1544 grid points) are taken into account in the analysis. For this point totality, the obtained rank order correlation coefficient (R_s) is 0.52 (Table 5). The simulated values are to some degree overestimated with respect to the measured ones in the range 0 - 15 mm·day^{-1} and underestimated in the range 20 - 35 mm·day^{-1} (not presented).

5.2 Comparisons

Precipitation field differences are analyzed also for the horizontal resolution of 18 x 18 km. The differences were considered from all points of view concerning the soil effects. The possible soil effects are as follows. Comparing the results of Run 1 and 2 (comparison 1), we can analyze how important the soil dataset effect is. Comparing the results of Run 1 and 3 (comparison 2), we get an insight into how important the parameterization of θ_f is. Comparing the results of Run 1 and 4 (comparison 3), we can see how important the parameterization of the θ_w is. And lastly, comparing the results of Run 1 and Run 5 (comparison 4), we can analyze how important the areal distribution of θ is.

Inspecting Table 5, we see that among R_s differences the ${R_s}^{Run1}$ - ${R_s}^{Run2}$ difference is the largest. This R_s difference is strongly significant amongst the considered R_s differences. We could make sure of this by inspecting Table 6. Results referring to significance level are obtained by using Fisher's r to z transformation. Note that these precipitation differences are caused by soil dataset differences (see Table 1). So, for instance, the 24-hours accumulated precipitation differences obtained on 1st August 2006 are presented in Fig. 4. They are unequivocally large. There are both positive and negative differences within a distance of a few tens of kilometres. The largest positive and negative values are about 15-20 mm·day^{-1}. These values are comparable with the 24-hours accumulated precipitation field values obtained by

Table 5: Rank order correlation coefficient R_s depicting the agreement between simulated and analyzed precipitation of different model runs. Results for 6-days (n=1544) and 4-days (n=1029) periods are shown.

	Run 1	Run 2	Run 3	Run 4	Run 5
6-day	0.52	0.48	0.52	0.49	0.48
4-day	0.62	0.53	0.60	0.56	0.54

Table 6: Significance levels (p-values) for 6-days and 4-days simulation periods of the rank order correlation coefficient differences obtained from Tab. 5.

	Run 1-2 (comp. 1)	Run 1-3 (comp. 2)	Run 1-4 (comp. 3)	Run 1-5 (comp. 4)
6 day	0.043	0.413	0.123	0.061
4 day	0.002	0.348	0.034	0.005

Run 1, which are presented in Fig. 5. In Fig. 5, the highest values are about 40 mm·day^{-1}, while the lowest values are about 2 mm·day^{-1}. Note that at many places the relative differences, i.e. precipitation of (Run 1–Run 2)/Run 1, amount to about 50%.

5.3 Significance tests

The sensitivity of convective precipitation to the above mentioned soil effects is also estimated by applying two different significance tests. The Wilcoxon signed rank test was applied to decide whether precipitation differences obtained on the given day were significant or not. In spite of this, Fisher's r to z transformation is applied when the differences between the rank order correlation coefficients are analyzed.

5.3.1 Wilcoxon signed rank test

Precipitation differences are inspected by Wilcoxon signed rank test on the level of 1 and 5 %. This was performed for the model domain. So, Fig. 6 shows the number of the days from the total of six when the precipitation differences were strongly significant. Similarly, Fig. 7 shows the number of the days from the six when the precipitation differences were significant. Inspecting the Figures, the following main characteristics could be emphasized:

- Precipitation differences were strongly significant on four days only once for threshold 1 mm·day^{-1} (Fig. 6). At the same time, precipitation differences were significant on four days four times for thresholds 1 and 3 mm·day^{-1} when comparisons 1 and 3 were carried out (Fig. 7). Strongly significant and significant days appear mainly for lower thresholds between 1 and 9 mm·day^{-1}. In these cases, day numbers vary between 0 and 4. Zero values appear only in the case of the strong significance. For higher thresholds between 11 and 19 mm·day^{-1}, the number of strongly significant days is either 0 or 1, while the number of significant days is either 1 or 2.

- For comparison 3 (θ_w parameterization effect) the mean day value for thresholds 1 - 9 mm·day^{-1} is 2.4. This is also noted in Fig. 6. After this comes comparison 1 (soil dataset effect). In this case, the mean day value for thresholds 1 - 9 mm·day^{-1} is 2.2. For comparison 4 (subgrid-scale variability effect of θ), the mean day value for thresholds 1 - 9 mm·day^{-1} is 1.6, and lastly for comparison 2 (θ_f parameterization effect) the mean day value is only 0.8. Similar results are also obtained for significant differences. There are four cases for comparisons 1 and 3 when the differences were significant on four days. Their mean day value for thresholds 1 - 9 mm·day^{-1} is 3 (Fig. 7). The same for comparison 4 is 2.8 and for comparison 2 it is 1.8. For higher threshold values between 11 and 19 mm·day^{-1}, it is impossible to rank the soil effects.

5.3.2 Fisher's r to z transformation

To place the soil effects in order, the rank order correlation coefficient differences were also analyzed by Fisher's r to z transformation. To do this we decided to analyze somewhat larger precipitation differences. To achieve this goal only those days were chosen which showed at least one strongly significant difference in comparison 1 when the Wilcoxon signed rank test was used. Days obtained applying this criterion were 1st August 2006, 7th August 2006, 22nd May 2007 and 1st June 2007. These rank order correlation coefficients are also presented in Table 5. It could be seen that the obtained R_s values are somewhat larger. The ${R_s}^{Run1}$ - ${R_s}^{Run2}$ difference is the largest amongst the R_s differences. Note that the ${R_s}^{Run1}$ - ${R_s}^{Run3}$ difference is henceforward the smallest. The corresponding significance levels could be seen in Table 6. We see that the obtained differences are strongly significant for comparison 1 and comparison 4 and significant for comparison 3. R_s difference for comparison 2 is not significant at all. These results practically confirm the results obtained by the Wilcoxon signed rank test regarding the ordering of the soil effects for thresholds 1 - 9 mm·day^{-1}.

6 Summary and conclusions

The quantification of the impact of soil hydraulic properties upon local convective precipitation is a relatively unexplored area. There are only a few studies where this subject is only partly touched. GAO et al. (2008) investigated the sensitivity of accumulated precipitation in June to topography, landuse and soil texture using regional high-resolution datasets for these factors. Soil parameters were locally determined but according to COSBY's (1984) method. A more detailed treatment of the impact of the soil parameters was not given. As in the former case, the OSBORNE et al. (2004) study served also climatological purposes. In this study, two soil datasets were used to investigate the produced effects on climate in the tropical regions. The datasets used differed considerably for tropical soils; the produced precipitation differences were observable, nevertheless, they did not seem to be large. To the best of our knowledge, our study is the only one so far where the impact of the soil upon convective precipitation is systematically studied for mesoscale modeling purposes. We used the Penn State - NCAR MM5 modeling system and like OSBORNE et al. (2004) we also used two soil datasets. The USA one, which could be treated as a global scale dataset, and a Hungarian one which is a regional dataset representative for the Carpathian Basin. The Hungarian dataset was constructed by the **R**esearch **I**nstitute for **S**oil **S**cience and **A**gricultural **C**hemistry (RISSAC); the soil parameter values were determined by laboratory measurements using VÁRALLYAY's (1973) method. The basic differences between the two datasets are also presented. Amongst atmospheric variables only the accumulated daily precipitation is analyzed. Model runs were performed on six days in the summer period of 2006 and 2007 using a horizontal resolution of 6 x 6 km. The model domain covered the Carpathian Basin together with the surroundig mountains. The results were analyzed on a courser grid with a horizontal resolution of 18 x 18 km to account for the spatial resolution of the analyzed precipitation fields, which were obtained from rain gauge data using ordinary block kriging.

In the reference run, the obtained rank order correlation coefficient was $R_s = 0.52$. In the sensitivity studies, four soil effects were considered. As the first effect the soil dataset effect (comparison 1) is noted. We talk about this when one soil dataset is replaced by an another. The second and the third effect (comparison 2 and 3) was the parameterization effect of θ_f and θ_w, respectively. The fourth effect (comparison 4) was the subgrid-scale variability effect of θ. We demonstrated that convective precipitation is sensitive to the soil dataset effect. The R_s difference produced by this effect was significant for the 6-day (p=0.043) and strongly significant for the 4-day (p=0.002) simulations. This finding confirms also the results of MÖLDERS et al. (2005) where among others the importance of the pore-size distribution index b is emphasized. It is also shown that the subgrid-scale variability effect of θ is approximately as large as the soil dataset effect. The 4-day R_s difference produced by this effect was also strongly significant (p=0.005). It could be mentioned that this result does not agree with MÖLDERS' (2001) result. According to MÖLDERS (2001), the subgrid-scale variability of θ is not strong. In our opinion, this contradiction is only virtual. It could be explained by the fact that GESIMA (**Ge**esthacht's **Si**mulation **M**odel of the **A**tmosphere) and MM5 have fundamentally different soil hydraulic representations. For instance, in GESIMA field capacity values are defined according to land-use types, while in MM5 according to soil textures. Further, in GESIMA soil-wetness is simulated by a force-restore method, while in MM5 by using the Richards equation. Lastly, we showed that the aforementioned two effects are much larger than the parameterization effects of θ_f and θ_w.

It has to be said that though the soil dataset effect is large, this effect is much smaller than the cumulus convection parameterization effect (e.g. YANG and TUNG, 2003; PARK and LEE, 2002). This could be checked comparing our results and the results obtained by PARK and LEE (2002). In both cases, the areas studied are approximately equally large, but over the Korean peninsula heavy rain cases were simulated. In both cases the MM5 model is applied using GRELL et al.'s convection (1994) and REISNER et al.'s (1998) cloud microphysic schemes. Comparing our results presented in Fig. 4 with PARK and LEE's results presented in Fig. 2, it is obvious that precipitation formation is much more sensitive to the parameterization of cumulus convection than to the treatment of soil.

The results suggest that domain-specific soil parameters should be preferred in the mesoscale models. The same is also suggested by MÖLDERS (2001). It is also obvious that high-resolution global soil datasets have to be carefully constructed because of the sensitivity registered. These results should be treated as preliminary results. They refer to only six days. Further tests are needed for extracting more general conclusions.

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