A Numerical Sensitivity Study on the Impact of Soil Moisture on Convection-Related Parameters and Convective Precipitation over Complex Terrain

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ABSTRACT

The impact of soil moisture on convection-related parameters and convective precipitation over complex terrain is studied by numerical experiments using the nonhydrostatic Consortium for Small-Scale Modeling (COSMO) model. For 1 day of the Convective and Orographically Induced Precipitation Study (COPS) conducted during summer 2007 in southwestern Germany and eastern France, initial soil moisture is varied from $-50\%$ to $+50\%$ of the reference run in steps of $5\%$. As synoptic-scale forcing is weak on the day under investigation, the triggering of convection is mainly due to soil-atmosphere interactions and boundary layer processes. Whereas a systematic relationship to soil moisture exists for a number of variables (e.g., latent and sensible fluxes at the ground, near-surface temperature, and humidity), a systematic increase of 24-h accumulated precipitation with increasing initial soil moisture is only present in the simulations that are drier than the reference run. The time evolution of convective precipitation can be divided into two regimes with different conditions to initiate and foster convection. Furthermore, the impact of soil moisture is different for the initiation and modification of convective precipitation. The results demonstrate the high sensitivity of numerical weather prediction to initial soil moisture fields.

1. Introduction

Despite the advances in the parameterizations of physical processes and higher grid spacings of numerical weather prediction (NWP) models over the past decades, quantitative precipitation forecasting still remains a challenge. In particular, the forecast of deep moist convection with weak synoptic forcing is still inadequate for many applications. Besides uncertain initial and boundary conditions, inaccuracies of numerical methods and/or the incomplete description of physical processes influence the performance of NWP models. One reason for the uncertainties of the initial state is that soil moisture fields are largely unknown and simple assumptions about the initial state introduce large errors (van Weverberg et al. 2010). Observations from a newly installed soil moisture network (SOMONET, Karlsruhe Institute of Technology; Krauss et al. 2010) in the region of a low mountain range (the Black Forest in southwestern Germany) show that simulated and measured soil moisture values may differ strongly. The mean bias for most of the stations and measurement depths lies around $20\%–30\%$ (Hauck et al. 2011). Soil moisture, however, is an important parameter in the soil–atmosphere system, which influences the soil system’s hydrology as well as the availability of humidity in the planetary boundary layer (PBL) through evaporative processes.

In situations with weak synoptic forcing, the PBL characteristics and their impacts on the triggering and/or dynamics of convective storms depend on the partitioning of available energy (net radiation minus ground heat flux) into sensible and latent heat, which in turn is determined by soil moisture. Previous observational and modeling studies demonstrated that soil moisture strongly contributes to the variability of surface temperature and precipitation via the exchange of water and energy between the land surface and the atmosphere (e.g., Koster et al. 2000). As was stated by Dai et al. (1999), soil moisture may decrease the daily temperature range by increasing daytime surface evaporative cooling. Furthermore, mesoscale circulations generated by land surface wetness inhomogeneities often have a stronger impact than turbulent fluxes on the formation of clouds and subsequent precipitation (e.g., Schädler 1990; Chen and Avissar 1994; Taylor et al. 2007). These circulations...
are generated by surface sensible heat flux gradients that result from spatial variations in surface evapotranspiration, solar irradiance reflection/absorption, and thermal energy storage in the soil (Cheng and Cotton 2004).

Previous studies produced contradictory results for the existence and sign of the soil moisture–precipitation feedback, which may vary spatially and temporally. According to the theory of the soil moisture–rainfall feedback described by Pal and Eltahir (2001), wet soil conditions lead to an increase in the PBL’s total energy (described by the moist static energy). This increase is expected to result in an increase in the convective available potential energy (CAPE). Regional climate simulations on the continental scale reveal that higher values of soil moisture lead to an enhanced amount of evapotranspiration, which is supposed to be indirectly responsible for more intense convective rainfall (Schär et al. 1999). It is assumed that a positive feedback mechanism between initial soil moisture and future rainfall exists: Wet soils are associated with the buildup of shallow boundary layers with high values of low-level moist entropy by concentrating the supply of heat and moisture by the surface fluxes into a comparatively small volume of air (Schär et al. 1999; Pal and Eltahir 2001). This mechanism represents a source of convective instability. Findell and Eltahir (2003) used a 1D PBL model to analyze the impact of the degree of soil saturation on PBL development and the triggering of convection in different atmospheric settings. They found a small positive feedback between soil moisture and rainfall over flat terrain. Gallus and Segal (2000) found an increase in precipitation with increasing soil moisture in the midwestern United States by varying the initial soil moisture content from 60% drier to 30% wetter than the control simulation. In contrast to this, Pan et al. (1996) found that in a humid atmosphere, an increase in soil moisture decreased the rainfall due to insufficient thermal forcing to initiate convection. Rainfall was enhanced when the lower atmosphere was thermally unstable and relatively dry. For the same reason, a reverse relationship between soil moisture content and precipitation was found for a dryline case in western Texas by Martin and Xue (2006). Using a coupled 1D land surface–PBL model, Ek and Holtslag (2004) showed that decreasing soil moisture may actually lead to an increase in PBL clouds in some cases. The stability of the layer into which the PBL is growing is considered to be important for determining the sign of feedback (Findell and Eltahir 2003).

Whereas the above findings were derived for flat terrain, the influence of soil moisture on convective indices and precipitation over complex terrain has been investigated so far by few authors only. For example, Hohenegger et al. (2009) investigated the soil moisture–precipitation feedback over the Alpine region by simulations with resolved and parameterized convection for one full month. The two systems yielded not only different strengths of feedback, but also different signs. The different feedback signs were found to be related to the presence of a stable layer on top of the PBL. Dry initial soil moisture conditions with 2.2-km grid spacing yield more vigorous thermals, which can break through the stable air barrier more easily, thus leading to a negative soil moisture–precipitation feedback (Hohenegger et al. 2009). By analyzing the sensitivity of soil moisture variations of ±25% with respect to the reference run, Barthlott et al. (2011a) found a considerable but nonsystematic dependence of convective precipitation on soil moisture over complex terrain, independent of the synoptic situation and orographic characteristics. In one of their seven cases (i.e., 12 August 2007), both the increase and reduction in soil moisture led to a lower precipitation amount than in the reference run. This implies that a (local) maximum exists in between. This raises the question: What kind of relationship exists between initial soil moisture and rainfall on the subsequent day in complex terrain? Furthermore, is there an optimal soil moisture content leading to a maximum in convective precipitation for this day? One explanation could be that the reference soil moisture field acts as an upper threshold value, and a further increase inhibits convective activity due to insufficient thermal forcing for convection initiation. The estimation of such a threshold, where the atmosphere does not feel an additional increase in soil moisture, is of considerable interest not only for short-range NWP, but also for climate simulations. The present study investigates this topic by numerical sensitivity experiments for the same day, but with a larger range of soil moisture variations with finer increments (±50% in steps of 5%). In addition, the apparent contradiction between studies on the soil moisture–precipitation relationship summarized in Table 1 will be addressed.

One disadvantage of previous studies investigating the soil moisture–precipitation relationship by numerical experiments is the use of just one drier or wetter soil moisture initialization than that used in the respective reference run. By doing this, only a small part of the possible atmospheric reactions on soil moisture is investigated. This paper addresses these limitations with the aforementioned larger range of soil moisture variations and finer increments. Furthermore, the impact of soil moisture on the full chain of processes starting from the impacts on turbulent fluxes at the ground, near-surface and PBL meteorological variables, evolution of clouds, and subsequent precipitation will be analyzed.
2. Method

The simulations were performed with the nonhydrostatic Consortium for Small-Scale Modeling (COSMO) model (Schättler et al. 2009), which is used for operational regional weather forecasting at the German Weather Service [Deutscher Wetterdienst (DWD)] and several other European weather services. The model employs an Arakawa C grid for horizontal differencing on a rotated latitude/longitude grid. We used model version 4.6 with a grid spacing of 2.8 km and 50 vertical layers, which allows for turning off the parameterization of deep convection. Shallow convection is parameterized using a modified Tiedtke scheme (Tiedtke 1989). A six-class scheme including graupel is used for microphysics, and a two-time-level Runge–Kutta method (Wicker and Skamarock 2002) for time integration is implemented. The turbulence closure is applied using a prognostic equation for the turbulent kinetic energy (TKE; Schättler et al. 2009). The scheme can be classified as Mellor–Yamada level 2.5 (Mellor and Yamada 1974)—that is, the stability functions are explicitly predicted. Initial and hourly boundary data come from the COSMO-EU forecast (i.e., the operational NWP model of the DWD with 7-km grid spacing); the initial time was 0000 UTC for all model runs with an integration time of 24 h. Besides Germany, Switzerland, and Austria, the simulation domain (Fig. 1) contains smaller parts of neighboring countries covering an area of around 1300 × 1200 km² (421 × 461 grid points).

The surface fluxes of momentum, heat, and moisture provide for the coupling between the atmospheric part of the model and the multilayer soil vegetation model TERRA-ML. A TKE-based surface transfer scheme for

![Fig. 1. COSMO simulation area (gray), including the COPS domain (black rectangle).](http://www.journals.ametsoc.org/doi/abs/10.1175/1520-0485-68.12.2971?journalCode=jas)
Table 2. Hydraulic parameters of the different soil types and the percentage of each soil type in the COPS domain. The porosity describes the volume of space in the soil not occupied by particles, whereas the field capacity describes the amount of water that the soil manages to hold after free drainage. The wilting point is defined as the minimal soil moisture content a plant requires not to wilt.

<table>
<thead>
<tr>
<th>Soil type</th>
<th>3 (sand)</th>
<th>4 (sandy loam)</th>
<th>5 (loam)</th>
<th>6 (loamy clay)</th>
<th>7 (clay)</th>
<th>8 (peat)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Porosity</td>
<td>0.364</td>
<td>0.445</td>
<td>0.455</td>
<td>0.475</td>
<td>0.507</td>
<td>0.863</td>
</tr>
<tr>
<td>Field capacity</td>
<td>0.196</td>
<td>0.260</td>
<td>0.340</td>
<td>0.370</td>
<td>0.463</td>
<td>0.763</td>
</tr>
<tr>
<td>Permanent wilting point</td>
<td>0.042</td>
<td>0.100</td>
<td>0.110</td>
<td>0.185</td>
<td>0.257</td>
<td>0.265</td>
</tr>
<tr>
<td>Percentage cover</td>
<td>3</td>
<td>9</td>
<td>61</td>
<td>22</td>
<td>4</td>
<td>1</td>
</tr>
</tbody>
</table>

The weaker the synoptic forcing, the greater is the impact of the interaction processes between soil and atmosphere on the near-surface meteorological variables and the PBL structure. As can be seen from the COSMO-EU analysis displayed in Fig. 3, synoptic-scale ascent is weak over the area under consideration. In this figure, the divergence of the \( \mathbf{Q} \) vector

\[
\mathbf{Q} = -\frac{R}{p} \left( \frac{\partial \mathbf{v}_g}{\partial x} \cdot \nabla T + \frac{\partial \mathbf{v}_g}{\partial y} \cdot \nabla T \right)
\]

is shown, where \( R \) is the gas constant of dry air, \( p \) is pressure, \( \mathbf{v}_g \) is geostrophic wind, and \( T \) is temperature.

Hoskins et al. (1978) demonstrated that the traditional form of the quasigeostrophic omega equation can be rewritten using the \( \mathbf{Q} \) vector, and that regions of upward (downward) vertical motion are associated with \( \mathbf{Q} \) vector convergence (divergence). The COPS domain is located north of a weak mid-European ridge in weak southwestern winds with values of \( \mathbf{V} \cdot \mathbf{Q} \) near zero. During the course of the day, the ridge moves slowly to the east (not shown), but the upper trough over the British Isles and its associated regions with significant upward vertical motion remain far away from the region of interest. Consequently, the triggering of convection on that day is primarily controlled by soil–atmosphere interactions and boundary layer processes.

At first, a reference run was performed with soil moisture values coming from the 7-km COSMO-EU forecast. Then, sensitivity studies were conducted where the initial volumetric water content (VWC) was varied in steps of 5% ranging from −50% to +50% with respect to the reference run. These percentage changes are defined not in terms of absolute volumetric water content, but relative to the amount that exists in the reference run. The change in soil moisture is applied to the entire model domain indicated in Fig. 1 and to all soil model levels. This keeps mesoscale disturbances that may develop in tight soil moisture gradient belts outside of the area of interest. As soil moisture is bounded by porosity and wilting point, changes to its mean value tend to also change the shape of the probability density function (PDF; Fig. 4a). As soil moisture is increased, the dominant value of the distribution shifts to higher values with lower frequencies of occurrence. In accordance with Western et al. (2002) or Jones and Brunsell (2009), a positive (negative) skewness occurs when the mean approaches the lower (upper) boundary. As is obvious from Figs. 4b and 4c, not only the mean soil moisture content VWC, but also the variances \( \sigma_{VWC}^2 \) and the variance-to-mean ratio \( \text{vmr} = \sigma_{VWC}^2 / \langle VWC \rangle \) of the initial fields increase with soil moisture. Accordingly, the relative soil water content RWC, defined as

\[
\text{RWC} = \frac{\text{VWC} - \text{WP}}{\text{PO} - \text{WP}},
\]

where WP is the wilting point and PO is porosity, increases with increasing initial soil moisture until the entire soil in the COPS domain is almost saturated. The mean VMC of the uppermost level (0–1 cm) of the reference run is 33% by volume, which is the same as the
monthly mean during August 2007 as observed by the soil moisture network (Krauss et al. 2010). The mean minimum and maximum values of the measurements for that month are 25% and 41% by volume with a standard deviation of 5% by volume. Thus, considering the standard deviation, the very dry (17% by volume) and very humid (48% by volume) conditions of our simulations lie close to the mean extreme values.

The convective indices CAPE and convective inhibition (CIN) were calculated in steps of 30 min in the entire model domain by lifting a parcel that reflects the mean values of the temperature and moisture in the lowest 50 hPa. Additionally, the lifting condensation level (LCL) was determined by Henning’s formula: \( \text{LCL} = 125 \times \left( T_0 - T_{d0} \right) \), with \( T_0 \) and \( T_{d0} \) being the near-surface temperature and dewpoint temperature, respectively. A method using vertical profiles of equivalent potential temperature \( \theta_e \) and the one of a hypothetically saturated atmosphere after Holton (2004, p. 294) was applied to derive the level of free convection (LFC). The latter variables had to be computed independently of CAPE and CIN because they are not model output variables in the COSMO

FIG. 2. (a) Orography of the COPS domain and (b) COSMO soil types: 3 (sand), 4 (sandy loam), 5 (loam), 6 (loamy clay), 7 (clay), 8 (peat), and 9 (water). The black rectangle in (a) indicates the area for averaging or summation of meteorological variables.
model so far. The value of $\theta_v$, which is proportional to the moist static energy, was calculated using the numerical algorithm of Bolton (1980).

3. Observations and reference run

To give an overview about the actual convection event on 12 August 2007, this section briefly describes the evolution of convective precipitation based on measurements and the results of the reference run. The 24-h accumulated precipitation as simulated with unchanged soil moisture indicates a number of convective cells and their tracks oriented from southwest to northeast (Fig. 5a). The major part of simulated precipitation occurs in the northern part and north of the Vosges mountains, in the northern Rhine valley, and the northwestern Black Forest. The maximum accumulated precipitation of the reference run is 30 mm. The surface station network reveals a similar highest precipitation amount of 27 mm at the northeastern edge of the southern Black Forest. Deep convection also did occur west of the Black Forest and in the Swabian Jura. In those regions, however, almost no precipitation was simulated by the model. As far as the timing of convective precipitation is concerned, the model does not simulate the convective showers observed during the morning hours (Fig. 5b). Note that the time series are divided by their respective maximum values in order to allow a comparison of point measurements with gridded model data. The showers occurring during the afternoon are simulated by the model, but their relative rain amount is smaller. The point in time of the simulated maximum precipitation (2300 UTC) is identical to the one of the measurements. Even if not all cells are simulated at the right place, it can be stated that the model captures the convective activity of the day under investigation rather well.

Radiosonde observations between 1100 and 1130 UTC show stable conditions in the Rhine valley and weak instability in the Murg valley ($\text{CAPE} = 70 \text{ J kg}^{-1} \text{ CIN} = 5 \text{ J kg}^{-1}$). This agrees quite well with the domain-averaged values of the reference run at 1200 UTC ($\text{CAPE} = 84 \text{ J kg}^{-1} \text{ CIN} = 11 \text{ J kg}^{-1}$), which is also reflected by the stable mean temperature stratification $\frac{dT}{dz} = -0.59 \text{ K (100 m)}^{-1}$ between 1 and 5 km AGL. More information about the day under investigation is given by Bennett et al. (2011), who focus on the convection observed by Doppler-on-Wheels radars over the northern Black Forest using the Weather Research and Forecasting (WRF) model at 700-m horizontal grid spacing. Hagen et al. (2011) investigate the role of the wind profile and low-level convergence for convection initiation over the COPS domain.

4. Results of the sensitivity experiments

In the following, mostly mean or spatially integrated values over the entire COPS domain (Fig. 2a) will be analyzed. This approach assures that the results are not restricted to small-scale processes, which can significantly vary in space and time. The effects of orographic structures, different soil types, and land use, however, are all incorporated internally.

a. Precipitation and clouds

At first, we analyze the 24-h accumulated precipitation and the 24-h mean cloud cover, both spatially
integrated over the COPS domain (Fig. 6). For drier soil moisture conditions, the daily precipitation amount increases with the increase in initial soil moisture. The maximum amount of precipitation is simulated in the reference run. The further increase in soil moisture leads to slightly lower precipitation amounts fluctuating around 80%–90% of the value of the reference run. The maximum precipitation amount in the COPS domain is reached for a grid point in the $1_{25\%}$ run. The overall trend shows a positive relationship of maximum precipitation to soil moisture up to the increase of 25%. Higher soil moisture initializations do not lead to a further systematic increase or decrease. Concerning the percentage coverage of the COPS domain with rain, the maximum is observed at 15% soil moisture reduction. Higher soil moisture contents lead to less area covered with precipitation. Despite the highest maximum precipitation amount at increased soil moisture, the lower percentage of areas with precipitation in total leads to a reduced accumulated precipitation.

Whereas both total and high cloud cover remain more or less constant for all sensitivity runs, midlevel cloud cover decreases with increasing initial soil moisture. The increase in the low-level cloud cover with soil moisture can be explained as clouds developing in/at the top of the PBL due to the more humid boundary layers in the wet experiments. For an increase in soil moisture larger than 20%, the sensitivity to soil moisture is reduced and all mean cloud covers remain more or less constant. The increase in the amount of maximum precipitation with low-level cloud cover could be explained by the fact that shallow convection conditions the atmosphere for

**Fig. 4. PDF of (a) soil moisture, (b) mean volumetric and relative soil moisture content, and (c) statistical variability for all 21 model runs. The brighter the grayscale value in (a), the higher the respective increase and decrease in initial soil moisture compared to the reference run (in steps of 5%).**
Further deep convection by moistening the middle troposphere and preventing the evaporation of convective turrets through the entrainment of dry air in agreement with findings of Zehnder et al. (2006) or Kirshbaum (2011).

Examining the time evolution of simulated precipitation in Fig. 7 reveals that the first occurrence of convective precipitation is between 0900 and 1000 UTC in all model runs. In the wet simulations, however, first showers are comparatively weak. When taking a threshold of 30-min accumulated rain amount of 50 L for the whole COPS domain, there is a clear time shift of up to 4 h between dry and wet simulations (not shown). The precipitation sum of the reference run gradually increases until 1900 UTC. Later on, convective activity is enhanced and the accumulated precipitation increases significantly until the end of the forecast period reaching a value of approximately 12 000 L. Because of the limitation of the forecast time to 24 h, it cannot be investigated whether the precipitation after 2400 UTC is significantly different than before midnight (i.e., whether the total precipitation of the rain event is sensitive to the run length). However, the precipitation between 2300 and 2400 UTC is already decreasing (Fig. 5b), so the effect of the run length on the total precipitation is probably small.

Compared to the reference run, the model runs with reduced soil moisture reveal larger precipitation amounts until 1900–2000 UTC, probably due to a higher thermal forcing for convection initiation. During the same time period, less convective precipitation is simulated in the runs with increased soil moisture. Whereas the wetter simulations and the reference run all show a gradual increase in accumulated precipitation with time, convective activity is reduced in the dry runs after a few hours. During the late afternoon and early evening, convection starts again also in the drier runs. The very dry simulations, however, have a significantly lower amount of rain than the rest of the sensitivity runs. With similar rates of increase and starting time as in the reference run, the precipitation amount of the wet simulations increase after around 1900 UTC. A possible explanation for the different characteristic behaviors of simulated precipitation could be that through large thermal forcing around noon, convective cells are initiated in the dry runs. Because of the lack of sufficient low-level humidity, convection has a short lifetime only. In contrast to this, the lack of adequate thermal forcing in the wet runs could be responsible for the lower precipitation amounts during the afternoon. This hypothesis will now be analyzed with near-surface meteorological variables and convective indices.
b. Energy balance components and near-surface meteorological variables

Under weak synoptic forcing with minor advection, the temperature and humidity supply of the PBL are determined mainly by the components of the surface energy balance. We therefore analyze at first the time series of domain-averaged net radiation $Q$, sensible heat flux $H$, latent heat flux $V$, and Bowen ratio $b$ (Fig. 8). The influence of different initial soil moisture on the net radiation is small: only the maximum values at noon in the drier simulations are slightly higher than the ones from the reference run or the wet simulations. In contrast to findings by van Weverberg et al. (2010), decreasing the soil moisture does not lead to a decrease in net radiation, probably because the albedo effect is not that effective during summertime with a high vegetation cover. The sensible and latent heat fluxes reveal a systematic response to soil moisture: As expected, sensible heat shows a negative and latent heat a positive relationship to soil moisture. The deviations from the reference run of the wet simulations are considerably smaller than the ones of the dry simulations. For example, the largest latent heat flux of the run with 50% soil moisture decrease (increase) is 213 W m$^{-2}$ smaller (50 W m$^{-2}$ larger) than in the reference run. This confirms the findings by Koster et al. (2004) that when the soil is relatively dry, the availability of soil moisture controls evaporation whereas when the soil is close to saturation, evaporation is controlled by the available net radiation. Only in the driest simulations does the Bowen ratio exceed unity during daytime, indicating that for most of the sensitivity experiments, the available energy (net radiation minus ground heat flux) is transformed mainly into latent heat. In agreement with Eltahir (1998), the Bowen ratio during day time decreases with increasing soil moisture.

The response of $H$ and $V$ to soil moisture variations has a significant impact on the near-surface meteorological variables, such as the 2-m temperature ($T_{2m}$), specific humidity ($Q_{V2m}$), relative humidity ($RH_{2m}$), and $\theta_e$, displayed in Fig. 9. In agreement with, for instance, Dai et al. (1999) and Pal and Eltahir (2001), soil moisture exhibits a positive relationship with near-surface humidity and a negative relationship with temperature. Again, larger deviations from the reference run occur for reduced soil moisture. Whereas specific humidity slightly increases during the course of the day for the wet and reference simulations, most of the dry simulations have lower humidity values at the end of the forecast period than at initialization time. The reason for this is the strong reduction of specific humidity between 0900 and 1400 UTC in the dry simulations probably caused by strong entrainment of drier air from above the PBL. After 1800 UTC, specific humidity either remains constant with time or increases slightly again. Relative humidity follows the temperature curves inversely with again smaller values after 24-h integration time. A systematic response to soil moisture is also present for $\theta_e$, being influenced by both temperature and humidity: lower (higher) soil moisture contents lead to lower (higher) $\theta_e$ values exhibiting a clear diurnal evolution, with maximum values occurring around noon and during the early afternoon. Low-level equivalent potential temperature is an important parameter for the potential of deep convection since it is correlated with CAPE (Machado et al. 2002; Parker 2002; Kalthoff et al. 2011). It is worth noting that there is a phase shift between individual maximum values of $\theta_e$: whereas in the driest run, the maximum is reached around noon, a plateau with constant values over a time period of 2–4 h evolves for the simulations with higher soil moisture contents (Fig. 9b). As will be discussed later, this has important implications for the occurrence of deep convection on the day under investigation.
To distinguish between dry and humid boundary layers, the humidity index $H_{\text{I low}}$ was calculated in the form used by Findell and Eltahir (2003) as the sum of the dewpoint depressions at 950 and 850 hPa:

$$H_{\text{I low}} = (T_{950} - T_{d,950}) + (T_{850} - T_{d,850}),$$

with $T_d$ being the dewpoint temperature. In the simulations by Findell and Eltahir (2003), a value larger than 2980 JOURNAL OF THE ATMOSPHERIC SCIENCES VOLUME 68

Fig. 8. Domain-averaged energy balance components: (a) net radiation, (b) sensible heat flux, (c) latent heat flux, and (d) Bowen ratio.

Fig. 9. Domain-averaged near-surface (a) relative humidity, (b) equivalent potential temperature, (c) temperature, and (d) specific humidity.
15 K indicated that there is not enough low-level humidity to allow for rainfall or shallow clouds. In this study, HI\textsubscript{low} shows a more humid PBL in the wet experiments with a systematic response to initial soil moisture (Fig. 10a). The diurnal cycle exhibits maximum values between 1400 and 1600 UTC. The fact that convective precipitation is simulated despite HI\textsubscript{low} being larger than 15 K in the driest runs indicates the impact of strong thermal forcing in these runs (Fig. 10b). In general, the precipitation amount decreases with increasing values of HI\textsubscript{low}.

c. Condensation levels and convective indices

The probability and intensity of convective precipitation can be characterized by a number of convective indices. Here, we analyze CAPE, CIN, and the height of the LCL, LFC, and the equilibrium level (EL). Higher soil moisture results in a smaller dewpoint depression by the reduced near-surface temperature and increased humidity (Fig. 9). As a consequence, a decrease in the height of the cloud base is expected. The results show a negative relationship of LCL and LFC to soil moisture (i.e., they are lower in the wet experiments; Fig. 11). The EL is almost not affected by increased soil moisture, but when initial soil moisture is reduced, EL can be lowered by a maximum of about 4 km. Whereas the LCL shows a diurnal cycle with only one maximum during the early afternoon, the time series of the LFC is more structured. After 0800 UTC, there is a maximum around 1400 UTC with two adjacent minima around 1000 and 1800 UTC, where convection is more likely to occur. Starting from very low levels during the morning hours, the EL quickly rises to values above 7 km AGL after 0800 UTC. In the wet experiments and the reference run, the EL remains more or less constant between 9 and 10 km AGL, whereas the dry simulations show a decrease in EL after around 1700 UTC. The LFCs between individual sensitivity runs before 0730 UTC are close together. Afterward, the spread of the ensembles becomes larger and the aforementioned systematic relationship to soil moisture becomes visible.

After the rise of the EL around 0800 UTC, CAPE also increases, reaching its maximum during the late afternoon at 1800 UTC in all sensitivity runs independent of initial soil moisture. In the dry experiments, there is a systematic relationship of CAPE to soil moisture: the lower the initial soil moisture, the lower the CAPE. A systematic increase in CAPE with increasing soil moisture is only present until 1800 UTC. Later, CAPE is affected by precipitation but remains comparatively high until the end of the forecast period. The largest values of CIN are not reached until 2200–2300 UTC, when precipitation cools the near-surface temperature via evaporation. As far as the sensitivity to initial soil moisture is concerned, CIN responds after 0900 UTC with increased domain-averaged values for drier runs and decreased values for the wetter ones. However, the response is not systematic through the entire range of 50% throughout the whole day. The diurnal cycle of CIN is linked with the LFC (e.g., the relative CIN maxima at 1330 and 2200 UTC correspond to higher levels of the LFC).

The precipitation simulated in the evening can be explained by the still large values of CAPE. This raises the question of the origin of high CAPE values in the afternoon. On days with weak synoptic forcing, the diurnal cycle of convection over land is dominated by the destabilization of the lower levels of the atmosphere by daytime surface heating. A linear relationship between CAPE and near-surface $\theta_e$ was proposed by Mapes (1993) and Williams and Renno (1993), which was confirmed with measurements by, among others, Machado et al. (2002), Kohler et al. (2010), and Kalthoff et al. (2011). As is obvious from Figs. 9 and 11, the temporal evolution of CAPE and near-surface $\theta_e$ is connected:
large values of $\theta_e$ correspond to high values of CAPE and vice versa. In agreement with findings by, for instance, Machado et al. (2002), low values of CAPE show a strongly reduced correlation with $\theta_e$ (not shown). Furthermore, a quasi-linear relationship is found for $\theta_e$ values larger than $50^\circ$–$55^\circ$C with correlation coefficients between 0.8 and 0.9. A closer look at the time evolution is provided in Fig. 12, where all domain-averaged values of both variables during the day are displayed against each other. Before noon, $\theta_e$ increases by $6^\circ$–$10^\circ$C whereas the increase in CAPE is still small. The higher the initial soil moisture, the larger the increase in CAPE during the next hours. For example in the reference run, $\theta_e$ remains more or less constant between 1400 and 1730 UTC, whereas CAPE grows by around 200 J kg$^{-1}$. This increase can be explained by the lowering of the LFC and a constant or even rising EL during that time. In contrast to findings by Guichard et al. (2004), for a given $\theta_e$ the corresponding CAPE is higher during the afternoon/evening than during the morning (i.e., the time evolution is counterclockwise). The same applies to the time evolution of CIN, where high values of $\theta_e$ correspond to lower CIN values whose maximum values are reached during the evening and night.

In general, the potential for the evolution of thunderstorms is increased when CAPE values are high and CIN is small. Hence, the occurrence of such optimal conditions and its response to soil moisture variations are of particular interest. For each sensitivity study, the number of grid points where CAPE $> 600$ J kg$^{-1}$ and CIN $\leq 10$ J kg$^{-1}$ were counted. The 24-h accumulated number of grid points increases with initial soil moisture (Fig. 13a), indicating a higher storm potential in the wet simulations. Whereas in the very dry simulations, these optimal combinations are very rare, the number of grid points strongly increases in the range from $-20\%$ to $+20\%$ soil moisture initialization. When analyzing the time evolution of the occurrence of these optimal conditions, again the increase with soil moisture is apparent (Fig. 13b). The fact that convective precipitation is simulated also in the drier runs after 0900–1000 UTC (see Fig. 7) indicates the existence of a mechanism allowing convective inhibition to be overcome. In addition, an important finding is the fact that different soil moisture initializations do not influence the time of day at which most of these optimal combinations occur.

d. Low-level convergence and midlevel moisture

In this section, two additional parameters controlling the triggering and evolution of convective clouds are analyzed: (i) low-level convergence zones and (ii) midlevel moisture conditions.
Several studies emphasized the role of lifting by low-level convergence zones for the initiation of convection (e.g., Raymond and Wilkening 1982; Wilson and Schreiber 1986; Kalthoff et al. 2009; Barthlott et al. 2010, 2011b). These convergence zones can be generated, for example, by differential heating due to the orography and surface moisture gradients or by mountain-induced flow modifications. They are important for reducing entrainment into the PBL (leading to a low-level \( \theta_e \) maximum) or for allowing strong updrafts to directly overcome convective inhibition (Garcia-Carreras et al. 2011). To examine the existence and strength of low-level convergence in our simulations, the divergence of the 10-m wind field was calculated, and only the convergent contributions were accumulated in the COPS domain (Fig. 14a).

Inspection of the diurnal cycle reveals that the maximum values occur around 1400 UTC when thermally driven secondary circulations are fully developed. The maxima in the wet simulations occur up to 2.5 h later than in the dry simulations, indicating a slight time shift as a result of increased initial soil moisture. This suggests that the simulated convergence between 1000 and 1800 UTC is primarily due to thermally induced circulation systems for which large sensible heat flux gradients are necessary. As is obvious from the domain maximum updraft velocities at 500 hPa (Fig. 14b), the initiation of the early afternoon precipitation (Fig. 7) is related to the low-level convergence, especially in the dry simulations for which CIN is high (see Figs. 11 and 13). It should also be mentioned that the convergence is certainly also affected by the precipitation itself and its associated cold pools. However, because of the comparatively small amount of rain in the afternoon and after visual inspection of the wind field, this influence is considered to be small. After their respective maxima, the accumulated convergences decrease until 1800 UTC when the main precipitation events build up. The more intense convective activity is reflected by the larger maximum updraft velocities and also higher precipitation rates (Fig. 7). As a consequence, the further increase in convergence after 1800 UTC is considered to be related mainly to downdrafts and cold pool formation.

The relevance of lower tropospheric lifting for the initiation of deep convection can be estimated by the simulated maximum upward vertical velocity between the surface and the LFC \( w_{\text{max}} \) and the required value to overcome convective inhibition \( w_{\text{CIN}} = \sqrt{2} \times \text{CIN} \) as follows:

\[
    w_{\text{diff}} = w_{\text{max}} - w_{\text{CIN}}.
\]
If \( w_{\text{diff}} \) is positive, CAPE can be released and convective clouds can form if the entrainment of drier environmental air into the rising parcel is not too strong (Adler et al. 2011). As is obvious from Fig. 14c, the number of grid points with positive values of \( w_{\text{diff}} \) is closely connected with the accumulated convergence of the 10-m wind field. This indicates the high importance of boundary layer convergence features for the initiation of deep convection considering the high values of CIN present in these simulations. The fact that fewer grid points in the wet simulations show a positive \( w_{\text{diff}} \) than in the dry simulations during the afternoon is concordant with the smaller but more intense convective cells found in the runs with increased soil moisture (Fig. 6).

The other parameter to be analyzed in this section is the midlevel moisture. Its importance for deep convection has been acknowledged by a number of authors (e.g., Zehnder et al. 2006, 2009; Wu et al. 2009; Hohenegger et al. 2009; Kirshbaum 2011; Adler et al. 2011). A humid middle troposphere is necessary for preventing the evaporation of convective turrets through the entrainment of dry air. Chaboureau et al. (2004) suggest a normalized saturation deficit (NSD) to quantify the occurrence of deep clouds especially when using mean values of larger domains:

\[
\text{NSD} = \frac{r_{\text{sat}} - r}{\sigma_{r_{\text{sat}}} - r}. \tag{5}
\]

This parameter relates the saturation deficit—defined as the mean difference between the saturated water vapor mixing ratio \( r_{\text{sat}} \) and the water vapor mixing ratio \( r \)—to the domain variability of water vapor, which is obtained by using the standard deviation of the saturation deficit \( \sigma_{r_{\text{sat}}} - r \). By examining the temporal evolution of NSD, Chaboureau et al. (2004) found a good agreement between the contour of \( \text{NSD} = 2 \) and the cloud condensate profile of 0.01 g kg\(^{-1}\). Figure 15 illustrates the time evolution of domain-averaged NSD and relative and specific humidity at the 700-hPa level. After 1700–1800 UTC, the NSD is below the threshold of 2, indicating suitable conditions for persisting deep convection. This is in agreement with the occurrence of the major components of convective precipitation on that day (Figs. 7 and 14).

The response to different soil moisture initializations is rather weak at 700 hPa, reflected by the comparatively low spread between individual sensitivity runs, indicating that soil moisture significantly affects meteorological variables of the PBL only. The increase (decrease) in relative humidity (NSD) in the course of the day is due to weak but more or less constant humidity and cold-air advection (not shown), resulting in an increase in specific humidity of around 2.5 g kg\(^{-1}\) until 2000 UTC and moistening through shallow convection. The comparatively dry conditions in the PBL and midtroposphere around noon (Figs. 9d, 10a, and 15c) apparently inhibit the convective cells from having a longer lifetime.

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**Fig. 14.** (a) Domain-accumulated positive convergence of the 10-m wind, (b) domain maximum updraft velocities at 500 hPa, and (c) number of grid points with \( w_{\text{diff}} > 0 \) m s\(^{-1}\).

**Fig. 15.** (a) 700-hPa specific humidity, (b) relative humidity, and (c) normalized saturation deficit.
In addition, weak synoptic-scale lifting with values of 2–4 cm s\(^{-1}\) at 700 hPa between 1530 and 1800 UTC (not shown) may have had an impact on the stronger rain intensities during the late afternoon.

5. Conclusions

Based on a case study from COPS, the numerical simulations presented here demonstrate that soil moisture has a considerable impact on convection-related parameters and convective precipitation over complex terrain. Summarizing the findings concerning the initiation of convection and subsequent precipitation, it can be stated that precipitation in all sensitivity runs starts nearly at the same time (0900–1000 UTC), but the occurrence of the first significant precipitation amounts in the wet simulations is up to 4–5 h later than in the dry simulations. This can be explained partly by the larger thermal forcing and strengthened thermally induced circulations present in the dry simulations. The temporal evolution of convective precipitation during the afternoon can be separated into two periods. Whereas the convective activity before 1800 UTC is comparatively weak, the major rain amount is simulated during the late afternoon/early evening for the following reasons:

- During the afternoon before 1730–1800 UTC, CAPE is still increasing from low values to their respective domain-averaged maxima. Large values of CIN, especially in the dry simulations, impede widespread initiation of deep convection. For the dry simulations, however, there is a higher number of grid points where the maximum upward vertical velocity below the LFC is high enough to overcome CIN, due to stronger thermal forcing and thermally induced circulations with accompanied low-level convergence. This permits more parcels from the PBL to reach their respective LFC, thus leading to the earlier onset of deep convection than in the reference run or the wet simulations. The lower amounts of CAPE together with the drier conditions at 700 hPa are responsible for the short lifetime of the precipitating cells in the dry simulations. In the wet simulations, low-level convergence is smaller and the number of grid points with \(w_{\text{diff}} > 0\) has a lower maximum, which occurs later than in the dry simulations. Overall, the precipitation amounts during the afternoon of the wet simulations are lower than in the reference run or the dry simulations. That means, for the period before 1730 UTC a negative correlation between soil moisture and precipitation is observed (Fig. 7).

- Between 1730 and 1800 UTC, domain-averaged CAPE reaches its maximum in all runs. The maximum number of grid points with high CAPE and low CIN values reflects optimum conditions for convection initiation at that time, too. Particularly, the wet simulations have a larger number of grid points fulfilling this requirement. During the late afternoon, the conditions for deep convection improve because of weak midtropospheric moist and cold advection. The conditions are even better in the wet simulations because higher evapotranspiration leads to higher \(\theta_e\) values in the boundary layer. At that time the triggering by convergence-induced lifting is rather unlikely because of the reduced convergence values and reduced number of grid points with \(w_{\text{diff}} > 0\). However, small convective cells are still present between 1600 and 1800 UTC for most of the sensitivity runs. These cells intensify and persist because of the higher CAPE and a more humid middle troposphere as reflected by the increase in humidity and decrease in NSD below the threshold of 2. A minor but not negligible role is played by the weak synoptic-scale rising of 2–4 cm s\(^{-1}\) during the evening. With respect to the modification of existing cells, convective precipitation increases with soil moisture in the dry simulations whereas a nonsystematic relationship is found in the wet simulations. That means that the impact of soil moisture is different for the initiation and modification of convective precipitation.

Altogether, whereas a systematic relationship exists for a number of variables (e.g., latent and sensible heat fluxes at the ground, near-surface temperature, humidity), the response of convective precipitation to initial soil moisture is more complex: If the soil is rather dry and evapotranspiration is controlled by net radiation, the influence of increasing soil moisture is much weaker and the general response of precipitation to soil moisture is not systematic anymore. For example, a 10% increase in soil moisture with respect to the reference run leads to a rain reduction of 28%, while another 5% increase results in an increase in precipitation by 20%. This could be a reason for the differing soil moisture–precipitation relationships found in previous studies; that is, depending on the position of the reference soil moisture content, for example, the further increase can lead to both an increased or a reduced amount of precipitation. Furthermore, the relative contribution of precipitation during the initiation and modification phase of convection significantly influences the dependence of 24-h accumulated precipitation on the initial soil moisture.

With respect to the reference run, increasing the soil moisture had a much smaller effect on convection-related...
parameters than decreasing it. This finding is consistent with previous studies by Koster et al. (2004), who showed that soil moisture only affects precipitation when soil wetness decreases to less than 30% of saturation. It was further found that no lower threshold for the atmospheric response on soil moisture existed in the analyzed range: each reduction up to −50% affected precipitation and other parameters clearly. After a soil moisture increase of 25% (corresponding to a mean volumetric water content of 41% by volume and a mean relative water content of 85%), however, the general response was very small for the day under investigation.

Complex orography plays an important role for convective processes by its thermal effects (i.e., differential heating of the earth’s surface, leading to slope and valley winds as a prerequisite of mass convergence and associated lifting). In the drier simulations, the strengthened thermally induced circulations are obvious from the stronger low-level convergence and increased number of grid points where CIN can be overcome. In an accompanying work, the role of the orography for convective precipitation has been investigated by simulations for the same day with unchanged soil moisture but removed mountains. The 24-h accumulated precipitation in the COPS domain is reduced by 23% when the Vosges mountains and the Black Forest are removed (not shown).

Even if a single case study is not sufficient for general statements because the behavior depends on the initial atmospheric conditions, too, the results indicate the high impact of initial soil moisture on the forecast of convective precipitation. Considering the mean bias of 20%–30% between measurements and numerical results found in the same region by Hauck et al. (2011), the particular importance of accurate initial soil moisture fields in numerical weather prediction models is emphasized. A potential benefit of a realistic representation of soil moisture in these models is an improvement in quantitative precipitation forecasting.

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