Evaluation of the ground heat flux simulated by a multi-layer land surface scheme using high-quality observations at grassland and bare soil

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Abstract

Two parameterisations for the dependence of the soil thermal conductivity on the soil water content are compared, using the multi-layer land surface scheme TERRA of the Consortium for Small-scale Modeling (COSMO) atmospheric model. The simulations were carried out in offline mode with identical atmospheric forcing data from the Meteorological Observatory Lindenberg of the German Meteorological Service (Deutscher Wetterdienst). The results show that the ground heat flux computed by the reference version of TERRA is systematically overestimated under dry conditions. In this version, the thermal conductivity does not depend on the simulated water content of the soil. Since the ground heat flux is part of the surface energy balance it affects the other components such as turbulent heat fluxes and surface temperature. An overestimation of the ground heat flux during daytime leads to an underestimation of the other surface fluxes and to a reduced surface warming, during afternoon and night this behaviour is reversed. The two formulations for soil thermal conductivity, presented by O. JOHANSEN (1975) on the one hand and M.C. McCUMBER and R.A. PIELKE (1981) on the other hand, both reduce the ground heat flux in TERRA under dry conditions, the former yielding good results while the latter is even leading to underestimations. In addition to this, the former is also applied in coupled mode in the climate version of the COSMO model, the COSMO-CLM, for Africa, resulting in improved diurnal cycles of near-surface temperature in dry regions. Furthermore, it is shown with the Lindenberg measurements that the soil temperature and hence the ground heat flux are particularly influenced by the effects of shading of the incoming solar radiation due to the vegetation cover, leading to a significantly reduced solar radiation at the sub-canopy land surface, even under a layer of grass. While TERRA is performing well when being applied to non-vegetated areas, and is further improved by the method of O. JOHANSEN (1975), for future improvements of TERRA the insulating effects by the vegetation should be represented.

Keywords: Land surface processes, Soil thermal characteristics, Regional Climate Modelling, Land-atmosphere interactions

1 Introduction

The ground heat flux is an important component of the surface energy balance in models for numerical weather prediction or climate simulations. It is determined by the soil thermal conductivity and heat capacity which both depend on properties such as soil texture, i.e. fractions of sand, loam and clay particles, and soil water content. This means, that these soil properties affect the amount of energy which is available, for example, to the turbulent heat fluxes and for warming the soil surface, and can therefore have a significant impact on near-surface temperature and humidity, low-level clouds and precipitation. For instance, the soil water content was found to be a key quantity determining the interactions between land surface and atmosphere (e.g. KOSTER et al., 2002; ASHARAF et al., 2012). In many regions of the world there exists a strong relationship between surface moisture deficits and the occurrence probability of hot extremes (MUELLER and SENEVIRATNE, 2012).

One pre-requisite for simulating the coupling of the land surface energy and water cycles correctly is a realistic description of the dependence of the ground heat flux on the soil water content. The thermal conductivity of water is about a factor of 25 larger than that of air. This means, a wet soil has a much larger thermal conductivity than a dry soil (see e.g. HILLEL, 1982). But, in the reference version of the multi-layer land surface scheme TERRA of the Consortium for Small-scale Modeling...
(COSMO) mesoscale atmospheric model (Steppeler et al., 2003), the soil thermal conductivity is kept constant in the vertical throughout the entire soil column and also in time. Its value corresponds to a fixed medium soil wetness (Doms et al., 2011). As a consequence, the thermal conductivity and therefore the simulated ground heat flux are systematically overestimated under dry conditions, leading to an underestimation of the other fluxes of the surface energy balance and to a reduced surface warming during daytime. During afternoon and night this behaviour is reversed.

During the last decades several formulations for representing the soil thermal conductivity and its dependence on the soil water content have been developed. They are applied in remote sensing studies (e.g. Verhoef, 2004; Murray and Verhoef, 2007; Verhoef et al., 2012) or in models for numerical weather prediction (NWP), agricultural meteorology or climate simulations (e.g. Noilhan and Planton, 1989; Viterbo and Beljaars, 1995; Peters-Lidard et al., 1998; Lu et al., 2007; Dharssi et al., 2009). Most of these formulations originate directly or in a modified form from one of the following two key publications: Johansen (1975) and McCumber and Pielke (1981). In this study, we present experiments with these two approaches applied to the multi-layer land surface scheme TERRA in offline mode. Additionally, Johansen (1975) was tested in coupled mode in the climate version of the COSMO model, the COSMO-CLM, for Africa.

2 Model description

The COSMO model (Steppeler et al., 2003; Doms et al., 2011) is a nonhydrostatic limited-area atmospheric prediction model, which is developed and maintained by the COSMO consortium (http://www.cosmo-model.org). It is designed for both operational NWP and various scientific applications on the meso-β and meso-γ scale. Furthermore, the COSMO model was expanded by the CLM community (http://www.clm-community.eu) to become applicable as a regional climate model, called COSMO-CLM (e.g. Rockel et al., 2008).

The COSMO model is based on the primitive thermo-hydrodynamical equations describing compressible flow in a moist atmosphere. The model equations are formulated in rotated geographical coordinates and a generalized terrain following height coordinate. A variety of physical processes, such as radiation, cloud microphysics, convection, or land surface processes, are taken into account by parameterisation schemes. The governing equations are integrated using the mode-splitting approach to split up the equations into a longer model time step for the processes on larger time scales such as advection and the tendencies from the physical parameterisations, and into a short time step for the fast sound wave processes. Several options for a two time-level 2nd and 3rd order Runge-Kutta split-explicit scheme are available (Baldauf et al., 2011).

The physical parameterisation schemes include the multi-layer soil and vegetation model TERRA (Doms et al., 2011). It simulates the energy and water balance at the land surface and in the ground, providing the land surface temperature and humidity as lower boundary conditions for computing the energy and water fluxes between surface and atmosphere. In TERRA, all processes are modelled one-dimensionally in the vertical, no lateral interactions between adjacent soil columns are considered.

The soil temperature is calculated by the heat conduction equation, while the soil water content is predicted by the Richards equation. Both equations are discretized by a multi-layer scheme using the same layer depths for both temperature and water content. This allows to include the freezing and thawing of soil water or ice, respectively. The layer depths can be specified by the user. In operational NWP applications usually 8 soil layers are used. The depths of their lower boundaries are as follows: 0.01 m, 0.03 m, 0.09 m, 0.27 m, 0.81 m, 2.43 m, 7.29 m, and 21.87 m. This layer discretization is also used in the offline simulations presented in this study. The temperature of the lowest layer acts as lower boundary condition of the heat conduction equation. It is usually set to a climatological annual mean value of the near-surface temperature. For instance, in operational NWP applications temperatures from the Climate Research Unit data set (Mitchell and Jones, 2005) can be used. It may be noted here that both soil temperature and water content are defined at the centers of the layers, but conductivities and fluxes at their boundaries. At the interface between surface and atmosphere, the surface energy balance equation is solved, yielding the new surface temperature. In TERRA, the surface temperature is represented by the temperature of the uppermost soil layer. There is no additional temperature of the leaves or the canopy, but this is a very common approach in atmospheric models currently used for operational NWP. The surface energy balance equation takes into account the total surface net radiation, the sensible and latent heat flux, and the ground heat flux. These atmospheric energy fluxes constitute the upper boundary condition of the soil heat conduction equation. The ground heat flux and the different options for the soil thermal conductivity are described in Section 3.

Precipitation reaching the ground is separated into infiltration or surface runoff. Depending on the vegetation fraction, bare soil evaporation and transpiration from the vegetation are computed. The total moisture flux is computed as the area weighted average. Snow is simulated with a single-layer snow model. It takes into account snow ageing with respect to albedo and density.

3 Soil thermal conductivity

In the multi-layer land surface scheme TERRA, the vertical soil temperature profile is computed by the 1-D heat
conduction equation
\[
\frac{\partial T_{so}}{\partial t} = \frac{1}{\rho c} \frac{\partial}{\partial z} \left( \frac{\partial T_{so}}{\partial z} \right) \tag{3.1}
\]
where \( T_{so} \) is the soil temperature, \( \rho c \) the volumetric heat capacity and \( \lambda \) the soil thermal conductivity. In the reference version of TERRA, taken from COSMO model version 4.23, \( \rho c \) is a function of soil water and ice content (Doms et al., 2011). \( \lambda \) varies with soil texture, but it is computed for a medium soil wetness, i.e. the average of field capacity and permanent wilting point, being constant in time (Doms et al., 2011). The scheme was originally designed in this way in order to limit its degree of complexity. At the beginning of the development it appeared that by introducing too many potential feedback loops the scheme may deliver unfavourable results (E. Heise, DWD, pers. comm., 2014). As a consequence, \( \lambda \) has at each grid point a medium value which is constant in the entire vertical soil column and in time. Under dry conditions, this leads to an overestimation of the simulated \( \lambda \) and ground heat flux, while under wet conditions they are underestimated.

### 3.1 Johansen (1975)

The Johansen (1975) parameterisation is described by Peters-Lidard et al. (1998) and was adapted to the TERRA code by Block (2007), accordingly. The soil thermal conductivity is calculated using
\[
\lambda = \lambda_{dry} + K_e(\lambda_{sat} - \lambda_{dry}) \tag{3.2}
\]
where \( \lambda_{dry} \) and \( \lambda_{sat} \) are the thermal conductivities of dry and saturated soils (in W m\(^{-1}\) K\(^{-1}\)), respectively. \( K_e \) is the Kersten number, which is a function of the degree of saturation \( S_r = \theta/\theta_s \) (where \( \theta \) is the soil water content (in m\(^3\) m\(^{-3}\)) and \( \theta_s \) its saturation value) and the phase of the water. For unfrozen soils,
\[
K_e = \log S_r + 1.0 \quad S_r > 0.1, \tag{3.3}
\]
and for frozen soils,
\[
K_e = S_r. \tag{3.4}
\]

The dry thermal conductivity is:
\[
\lambda_{dry} = \frac{0.135 \text{ W m}^{-1} \text{ K}^{-1} \cdot \rho_d + 64.7 \text{ W m}^{-1} \text{ kg K}^{-1}}{2700 \text{ kg m}^{-3} - 0.947 \rho_d} \tag{3.5}
\]
where \( \rho_d \) is the bulk density of the soil (in kg m\(^{-3}\)), and 2700 kg m\(^{-3}\) is the density of soil solids. \( \rho_d \) is obtained from the porosity \( n \) (in m\(^3\) m\(^{-3}\)) by:
\[
\rho_d = (1 - n)2700 \text{ kg m}^{-3}. \tag{3.6}
\]

The saturated thermal conductivity is estimated using
\[
\lambda_{sat} = \lambda_s^{1-q} \lambda_q^{p-x_u} \lambda_w \tag{3.7}
\]
with \( \lambda_s \) the thermal conductivity of the soil solids, \( \lambda_q = 2.2 \text{ W m}^{-1} \text{ K}^{-1} \) that of ice and \( \lambda_w = 0.57 \text{ W m}^{-1} \text{ K}^{-1} \) that of water. \( x_u \) denotes the unfrozen volume fraction. \( \lambda_s \) is found from
\[
\lambda_s = \lambda_q^{2/3} (\frac{q}{q+1}) \tag{3.8}
\]
where \( q \) is the quartz content, \( \lambda_q = 7.7 \text{ W m}^{-1} \text{ K}^{-1} \) is the thermal conductivity of quartz and \( \lambda_s \) is that of other minerals (2.0 W m\(^{-1}\) K\(^{-1}\) for \( q > 0.2 \) and 3.0 W m\(^{-1}\) K\(^{-1}\) otherwise). Following Peters-Lidard et al. (1998) it is assumed that the quartz content is equal to the sand content, the latter is given by Block (2007).

### 3.2 McCumber and Pielke (1981)

In the McCumber and Pielke (1981) parameterisation the soil thermal conductivity is calculated using
\[
\lambda = 418 \exp[\left(\frac{-pF}{2.7}\right)] \text{ W m}^{-1} \text{ K}^{-1} \quad pF \leq 5.1, \tag{3.9}
\]
\[
\lambda = 0.171 \text{ W m}^{-1} \text{ K}^{-1} \quad pF > 5.1, \tag{3.9}
\]
where
\[
pF = \log \left| \psi(\theta) \right| \tag{3.10}
\]
and \( \psi(\theta) \) is the matric potential (in cm) at the soil water content \( \theta \).

### 4 Experiments and observational data

#### 4.1 Lindenberg/Falkenberg

In order to analyse the dependence of the soil thermal conductivity on the soil water content, experiments with the multi-layer land surface scheme TERRA in offline mode were carried out. This methodology is described e.g. by Chen et al. (1997) or Schulz et al. (1998). For this purpose, TERRA was forced with a set of atmospheric observations, which are downward shortwave and longwave radiation, total precipitation, near-surface wind speed, air temperature, and specific humidity. For this study, observations from the boundary layer field site Falkenberg were used, providing the atmospheric forcing variables as mentioned before, as well as several quantities for model validation such as total surface net radiation, temperature, water content, and heat flux at different depths of the soil.

Falkenberg is a site at the Meteorological Observatory Lindenberg – Richard-Alßmann-Observatory – of the German Meteorological Service (Deutscher Wetterdienst), located about 5 km south of the observatory. It is a grass land site representative for farmland surfaces in the heterogeneous rural landscape typical for large parts of northern central Europe (Neisser et al., 2002; Beyrich et al., 2006). It is in continuous operation since 1998 with a main focus on the near-surface boundary layer and soil processes. Due to the high quality standard, the continuous availability, and the wide variety
of the in-situ boundary layer measurements, Falkenberg has become a reference layer site in the Coordinated Energy and Water Cycle Observations Project (CEOP) (https://www.eol.ucar.edu/field_projects/ceop), a WMO project for building up an integrated global observing system for the energy and water cycle (see also Koike, 2004). The continuous soil measurements at Falkenberg are carried out at the southern part of the site. The soil temperature is measured by a single profile of PT-100 resistance probes (1/10 DIN Class B, tolerance (in K) of ±0.03 ± 0.0005 · T (with T in °C)), installed at 12 levels at depths between 5 cm and 150 cm. The soil heat flux is derived at a depth of 5 cm and 10 cm as an average of 6 or 4 heat flux plates, respectively (Rimco CN3/HP3, McVan Instruments, Australia; calibration accuracy of ±5 percent specified by the manufacturer), to account for small scale heterogeneities directly under the grass surface. At each depth the sensors are installed in two spatially separated groups located 5 m west and east from the central soil temperature profile, respectively. The heat flux plates which are embedded in the soil create a discontinuous thermal conductivity in the soil. This causes a flow distortion error, affecting the heat flux measurements, which is corrected for by the Philip (1961) method. In order to compare the measurements of soil temperature and heat flux with the model simulations (see Section 5.1), the measurements are interpolated and extrapolated, respectively, to the model layers of the soil model. Temperature values are needed at the layer centers, heat flux values at the interfaces between adjacent layers. A detailed description of the measurement conditions, instrumentation, data acquisition and the comprehensive quality control procedures is given by Beyrich and Adam (2007).

The offline simulations were carried out for the full year of 2010. This year was characterised by a wet spring with frequent rain events, and a dry period in June/July. In the second half of July rain events started again. The main land surface parameter values needed to appropriately describe the Falkenberg site within TERRA are given in Table 1. They were based on measurements and estimates adapted to the site conditions present in 2010. The annual cycles of vegetation ratio and leaf area index (LAI) were prescribed by a sinusoidal fit for the transitions between their minimum and maximum values in spring and autumn, while the roughness length was kept constant in time. The predominant vegetation species are perennial ryegrass (Lolium perenne) and red fescue (Festuca rubra). For these grass species an albedo value of 0.18 was used. The soil texture especially prevailing at the radiation measurement spot is dominated by sandy pale soil (Eutric Podzoluvisol) (see Hierold et al., 1997) according to the FAO soil classification (FAO, 1988), for which an albedo value of 0.2 was used. The prescribed albedo values are in good agreement with experimental findings for this site described by Beyrich and Adam (2007). Although the meadow of the site is mowed several times a year to keep the vegetation height below 20 cm, the impact of the surrounding crop fields on the 10-m wind speed was represented by a slightly increased roughness length $z_0$ of 0.03 m on annual average.

### 4.2 Africa

In addition to the offline experiments, the effects of a variable soil thermal conductivity $\lambda$ depending on the soil water content are investigated in 3-D coupled model mode. In contrast to the offline set-up, here in coupled mode, the land surface can feedback to the atmosphere (see e.g. Schulz et al., 2001). Therefore, it can be analysed whether changes at the land surface may have an impact on near-surface atmospheric variables, such as for instance temperature. A perfect test-bed for this kind of study would be a place which is completely dry and without vegetation, such as the Sahara desert in Africa. This is for two reasons: firstly, the model soil can be expected to be very dry, therefore the values of $\lambda$ computed by the two different approaches will be much smaller than the one in the reference version of TERRA, and secondly, the shading due to vegetation, which is perturbing the ground heat flux, is excluded.

For this study, COSMO-CLM was used, a version of the COSMO model designed for regional climate simulations. Here, COSMO-CLM simulations over Africa are presented, using the Coordinated Regional climate Downscaling Experiment (CORDEX; Giorgi et al., 2009) Africa domain at a horizontal resolution of 0.44 ° with 35 atmospheric layers (cf. Panitz et al., 2014; Kothe et al., 2014b). The COSMO-CLM model version 4.8_clm18 was used. Lateral boundary conditions were provided by ERA-Interim (Dee et al., 2011) from the European Centre for Medium-range Weather Forecasts for the simulation period 2008–2010. Two COSMO-CLM simulations were carried out, the first one with the unmodified model as reference, and the second one using the Johansen (1975) approach for calculating $\lambda$. The McCumber and Pielke (1981) approach was not tested here, because the offline experiments (Section 5.1) show that it tends to deliver too extreme values for $\lambda$.

For validation, a high-resolution gridded dataset of daily maximum and minimum 2-m temperatures for Africa for the period 2008–2010 is used. This dataset was created by Krähenmann et al. (2013) using the

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**Table 1:** Main land surface parameter values in TERRA representing the Falkenberg site. $\sigma_{\text{min}}$ and $\sigma_{\text{max}}$ are the minimum and maximum vegetation ratio, representing winter and summer conditions for grass. $\text{LAI}_{\text{min}}$ and $\text{LAI}_{\text{max}}$ are the respective values but for the leaf area index. The transitions between these values in spring and autumn are prescribed by a sinusoidal fit. $\alpha_z$, $\alpha_{\text{gr}}$, and $\alpha_{\text{bs}}$ are the surface albedos for grass and bare soil, and $z_{\text{gr}}$ and $z_{\text{bs}}$ are the turbulent roughness lengths for grass and bare soil, respectively.

| Parameter | $\sigma_{\text{min}}$ | $\sigma_{\text{max}}$ | $\text{LAI}_{\text{min}}$ | $\text{LAI}_{\text{max}}$ | $\alpha_z$ | $\alpha_{\text{gr}}$ | $z_{\text{gr}}$ (m) | $z_{\text{bs}}$ (m) |
|-----------|----------------------|----------------------|----------------------|----------------------|----------|----------------------|-------------------|-------------------|
| Value     | 0.55 0.80 0.5         | 2.5                  | 0.18 0.2             | 0.03 0.03             |          |                      |                   |                   |
regression-kriging regression-kriging (RKRK) gridding algorithm. A well-known dataset which is widely used in climate studies is the one by the Climate Research Unit (CRU, Mitchell and Jones, 2005). But, here the RKRK data appear to be favourable for the following reasons: RKRK uses significantly more observing stations in Africa than CRU (338 versus 228), it applies satellite-derived predictors to compensate for missing temperature information in data sparse regions, and it has daily data coverage. The root mean square error of the dataset with respect to the original point observations is about 1.5 K in western Africa and 2.3 K on average over all Africa.

5 Results of numerical experiments

5.1 Offline mode

The approaches by Johansen (1975) and McCumber and Pielke (1981) (henceforth J75 and MP81, respectively) were applied to the multi-layer land surface scheme TERRA in offline mode, using the methodology as described in Section 4.1. Fig. 1a) shows the diurnal cycles of the surface solar (or shortwave) radiation balance, as measured at Falkenberg in July 2010 and as simulated by the reference version of TERRA (henceforth TERRAref). The model agrees very well with the observation. This is also the case in the TERRA experiments using J75 and MP81 (not shown). The first half of the month is dominated by clear-sky conditions with a solar net radiation of 600–700 W m⁻², while in the second half the radiative forcing is significantly reduced during two short periods around 17 July (Julian day 198) and 23 July (Julian day 204). These interruptions of the irradiation are caused by clouds. Fig. 1c) illustrates that also the sum of the surface sensible and latent heat fluxes is very well simulated by TERRAref. Here, measurements are shown which were adjusted for energy balance closure while maintaining the Bowen ratio (cf. Twine et al., 2000). On the other hand, the surface total net radiation, i.e. the sum of the solar and thermal (or longwave) net radiation, is overestimated by TERRAref during daytime and underestimated during nighttime, respectively, by about 50 W m⁻² (Fig. 1b). Together with Fig. 1a) this means that the upward thermal radiation is underestimated during daytime and overestimated during nighttime, which is indicating a surface cold bias during the day and a warm bias during the night. This is actually found, as shown in Fig. 1d). In TERRA, the surface temperature is represented by the temperature of the uppermost soil layer. It has a layer depth of 0.01 m which means that it exhibits a “thermal inertia”. Therefore, the simulated surface temperature is not able to represent the full diurnal temperature range of a skin temperature “seen” by the atmosphere. This is discussed in detail e.g. by Schulz et al. (1998). In Fig. 1d) the simulated surface temperature is compared to the observed brightness temperature, which is derived from the measured...
upward thermal radiation over the grass, hence representing partly ground and partly grass temperature, and therefore exhibiting less thermal inertia. As mentioned before, a vegetation temperature, in literature also called canopy temperature, is not represented by TERRA. The result shown by Fig. 1d) indicates that this is a systematic model shortcoming which prevents the model to represent the full diurnal surface temperature range over vegetated areas. The differences between the observed and simulated maximum and minimum temperatures, respectively, amount to about 4 K. Beside the reasons mentioned here, also the simulated ground heat flux contributes to the underestimated diurnal surface temperature range and the overestimated total net radiation, because it is overestimate as well (shown below, Fig. 5).

The two cloudy periods around 17 and 23 July mentioned in Fig. 1a) are associated with two rain events on these respective days which increase the soil wetness rapidly. This is seen in Fig. 2 which compares the soil water content of the four uppermost soil layers at Falkenberg during July 2010, as simulated by TERRA_ref, and by J75. After a very dry period at the beginning, the sudden moistening due to the two rain events in the second half of the month is clearly visible in all four layers. The simulated soil water content is very similar in both model versions. This is also the case for MP81, but for the sake of clarity of the Figures this is not shown. For layer three and four, also measurements of the soil water content are available, revealing that timing as well as amplitude of the rain-induced increases and peaks are captured by TERRA very well.

Fig. 3 illustrates the soil thermal conductivity \( \lambda \) at the three uppermost soil layer interfaces during July 2010 as simulated by TERRA_ref with constant \( \lambda \) (being about \( 1.45 \text{ W m}^{-1}\text{K}^{-1} \) for sand as at Falkenberg), and by J75 and MP81 with variable \( \lambda \) values depending on soil water content. They are computed in two steps: In the first step, the current values of the simulated soil water content are used to calculate \( \lambda \) for each layer, based on one of the two approaches, and in the second step, the values of \( \lambda \) are interpolated from the layer centers to the interfaces between adjacent layers. This is necessary because the computation of the ground heat flux between adjacent layers requires the thermal conductivity at the layer boundaries. This means that the \( \lambda \) values at the depths of 1 cm, 3 cm and 9 cm, shown in Fig. 3, correspond to the layer interfaces between the four uppermost layers of TERRA, and they are computed from the simulated values of soil water content presented in Fig. 2. As Fig. 3 shows, changes in soil moisture are directly reflected in the thermal conductivity. During the dry period at the beginning of the month the \( \lambda \) values computed by J75 and MP81 are very low, while during the two rain events in the second half of the month increasing soil moisture results in increasing values of thermal conductivity. Here, the variable \( \lambda \) values tend to oscillate around the constant mean value of \( \lambda \) of TERRA_ref. MP81 shows more extreme values than J75, yielding lower \( \lambda \) values under dry conditions and higher values under wet conditions.

**Figure 2:** Temporal evolution of the soil water content at a depth of a) 0–1 cm, b) 1–3 cm, c) 3–9 cm and d) 9–27 cm at Falkenberg during July 2010 as simulated by the reference version of TERRA, and by **Johansen** (1975). For the two lower layers also measurements are available.
Figure 3: Temporal evolution of the soil thermal conductivity $\lambda$ at a depth of a) 1 cm, b) 3 cm and c) 9 cm at Falkenberg during July 2010 as simulated by the reference version of TERRA with constant $\lambda$, and by Johansen (1975) and McCumber and Pielke (1981) with variable $\lambda$ depending on soil water content.

At Falkenberg, direct measurements of $\lambda$ are currently not available. In principle, it would be possible to derive $\lambda$ from other quantities which are observed. But, this is a technical exercise which is beyond the scope of this article. Nevertheless, the approaches by J75 and/or MP81 were evaluated and compared to measurements by several authors, for instance, Peters-Lidard et al. (1998) or Lu et al. (2007).

Fig. 4 compares the vertical profiles of $\lambda$ for the three TERRA versions shown in Fig. 3. The curves were computed for the mean soil moisture profiles at Falkenberg on 1–16 July 2010 (Julian day 182–197) in the three TERRA offline simulations. During the dry period in summer, the top soil layers had dried out, leading to lower values of $\lambda$ in J75 and MP81. In contrast, the bottom layers are still wet, leading to higher values of $\lambda$. It can be noted, again, that MP81 delivers more extreme values than J75. Furthermore, it is illustrated that the value of the constant $\lambda$ (and hence the ground heat flux) of TERRAref appears to be overestimated under dry conditions, and underestimated under wet conditions.

In Fig. 5, the diurnal cycles of the ground heat flux as measured at Falkenberg in July 2010 are compared to the simulations with TERRAref and J75, at soil depths of 3 cm and 9 cm. The amplitudes of the diurnal cycles are considerably overestimated in TERRAref during the entire month. In contrast, they are substantially reduced in the J75 experiment during the dry period (Julian day 182–197) and match the measurements better. When the soil gets moistened after the rain events, the soil thermal conductivity in J75 increases (cf. Fig. 3), which allows for enhanced amplitudes of the ground heat flux (Julian day 198–204). At the end of the month, the conductivity by J75 exceeds the one of the reference, which is also reflected by the ground heat flux (Julian day 205–212). The comparison of Figs. 3b) and c) with Figs. 5a) and b) shows that the evolution of the soil thermal conductivity from low to high values during the month is associated with a corresponding increase of the diurnal amplitudes of the ground heat flux. Fig. 6 is similar to Fig. 5, but shows MP81, instead. Here, the diurnal amplitudes of the ground heat flux are even more reduced during the dry period, leading to an underestimation, and vice versa during the wet period. This is explained by the more extreme values of $\lambda$ computed by MP81 when compared to J75 (see Fig. 3).

The behaviour of the soil temperature at the depths of 6 cm and 18 cm illustrated by Fig. 7 generally corresponds to the ground heat flux shown in Fig. 5. The amplitudes of the diurnal cycles are overestimated in TERRAref during the entire month, in particular dur-
Figure 5: Diurnal cycles of the ground heat flux as measured at Falkenberg in July 2010 compared to the results of the reference version of TERRA, and of Johansen (1975), at a soil depth of a) 3 cm and b) 9 cm.

Figure 6: Same as Fig. 5 but for McCumber and Pielke (1981).

Figure 7: Diurnal cycles of the soil temperature as measured at Falkenberg in July 2010 compared to the results of the reference version of TERRA, and of Johansen (1975), at a soil depth of a) 6 cm and b) 18 cm.

Due to these deficiencies, MP81 is not further considered in this study.

An additional uncertainty in the surface energy balance and therefore also in the ground heat flux is due to the insulating effects by the vegetation with respect to the sub-canopy land surface (cf. e.g. Deardorff, 1978). In the presence of a vegetation cover, the energy balance at the land surface, meaning at the interface between atmosphere and soil, is considerably modified compared to the case without vegetation. In particular, a part of the incoming solar radiation is reflected, scattered or absorbed by the vegetation, the latter creating an additional heat storage at the canopy level. As a consequence, less solar radiation reaches the ground below the vegetation, causing a shading of the sub-canopy land surface. The heat stored by the canopy changes the canopy temperature, modifying the thermal radiation balance and the turbulent heat fluxes at canopy and ground level. In TERRA, the insulating effects by the vegetation, including the shading, are not represented, as the incoming solar radiation is directly used in the surface energy balance equation. Strictly speaking, due to this the model is only applicable to bare soils, but not appropriate for vegetated areas. In addition to the grass land site, there exists another site at Falkenberg, where measurements of...
soil temperature under bare soil are available. It is a spot with an area of about 1 m², where the grass is permanently removed and therefore insulating effects by vegetation are completely avoided. These additional data enable a comparison of temperatures under grassland and bare soil.

In order to illustrate the insulating effects by vegetation, Fig. 9 shows the diurnal cycles of the soil temperature at a depth of 6 cm, but here for bare soil instead of grass-covered soil (compare Fig. 7a). The main obvious difference in the bare soil case is that the diurnal amplitudes of the measured soil temperature are much larger than in the grass-covered case. For bare soil, the measurement and the model results are in particularly good agreement. From the comparison of Figs. 7a) and 9 it can also be seen that here the simulated soil temperatures for grass-covered and for bare soil are similar, though not identical. This appears to be surprising, but it can be explained by the fact that the land surface parameter values which are recommended for these two land use types at the Falkenberg site are very similar, in particular for the albedo, or the roughness length (see Section 4.1, Table 1). Therefore, the solar radiative and atmospheric forcings are very comparable as well, resulting in these similar soil temperature simulations.

For better visibility, three days are selected from Fig. 7a), i.e. 2–4 July 2010 (Julian day 183–185), and they are depicted in Fig. 10a). The diurnal temperature range at a depth of 6 cm is about 2 K lower in J75 when compared to TERRAref, but the amplitudes of the diurnal cycles of the soil temperature are still considerably overestimated for grass land by both model versions. In contrast, when comparing the TERRA simulations with the measurement under bare soil, J75 turns out to be very good (Fig. 10b). A good performance of J75, using different observational data, was also found by Peters-Lidard et al. (1998) and Lu et al. (2007). In order to compare TERRAref and J75 in a more objective way, some statistical measures computed for their simulations of the soil temperature at a depth of 6 cm (as shown in Figs. 7a) and 9) are given in Table 2. The most significant result here is that J75 reduces the root mean square error (RMSE) for both, grass land and bare soil. This indicates that J75 is systematically better able to describe the thermal processes in the soil than TERRAref because it is more physics-based. The bias behaves differently, but it is less suited than the RMSE to assess whether J75 can significantly better represent the diurnal cycles of the soil temperature.

In order to quantify the insulating effects by vegetation, Figs. 11a) and 11b) compare the diurnal temperature range (DTR) of the measured and simulated soil temperatures, for both, grass land and bare soil. The DTR is the difference between the daily maximum and minimum temperature. Here, the DTR was computed from the diurnal cycles of the Falkenberg soil temperatures at a depth of 6 cm in July 2010, as shown in Figs. 7a) and 9, respectively. The average values of the DTR for July 2010 are presented in Table 3. For the measurements, the average DTR is reduced by the presence of the grass cover from 11.62 °C to 5.84 °C, this is almost exactly a factor of 2. This means that the insulation and in particular the shading by vegetation, even by grass, can have a substantial effect on the ground heat flux and the surface energy balance.

### Table 2: Bias and root mean square error (RMSE) of the soil temperature at a depth of 6 cm simulated by the reference version of TERRA, and by Johansen (1975). Bias and RMSE are computed with respect to the measurements at Falkenberg in July 2010. The results for grass land correspond to Fig. 7a), those for bare soil to Fig. 9.

|                | Grass land site | Bare soil site |
|----------------|-----------------|----------------|
|                | Bias (K)        | RMSE (K)       | Bias (K)       | RMSE (K)       |
| Reference      | 1.71            | 3.49           | −0.52          | 1.51           |
| Johansen       | 1.08            | 2.94           | −1.15          | 1.42           |
Figure 10: a) Same as Fig. 7a), diurnal cycles of the soil temperature at a depth of 6 cm under grass, but for better visibility only 3 days are shown: 2–4 July 2010 (Julian day 183–185). For better comparison, the three curves were detrended. b) Same as Fig. 10a) but under bare soil.

Table 3: Average diurnal temperature range of the soil temperature at a depth of 6 cm as measured at Falkenberg in July 2010 compared to the results of the reference version of TERRA, and of Johansen (1975). The values for grass land versus bare soil correspond to the Figs. 11a) and 11b), respectively.

|                  | Grass land site | Bare soil site |
|------------------|----------------|----------------|
| Measurement      | 5.84           | 11.62          |
| Reference        | 11.91          | 11.96          |
| Johansen         | 11.36          | 11.42          |

As one consequence of a missing representation of the insulation by vegetation in TERRA, the DTR at a depth of 6 cm is considerably overestimated under grass land by TERRARef and also by J75 during most of July 2010 (Fig. 11a). In contrast, when analysing a case when no vegetation is present, namely at the Falkenberg bare soil site, both model versions agree much better with the measurement (Fig. 11b). In a more detailed view, J75 turns out to be more favourable than TERRARef. During the dry period (Julian day 182–197) the DTR is reduced by J75 by about 2 K compared to TERRARef, and during the moister days (Julian day 200–204) the DTR is increased by J75 by up to 2 K. In both cases, J75 agrees generally better with the measurement than TERRARef. During the wet period (beyond Julian day 205) both model versions and the measurement become similar, however. The better performance of J75 is also reflected by the DTR averaged over the entire month which is closer to the measurement as well (cf. Table 3).

5.2 Climate mode

In this section, the effects of a variable soil thermal conductivity depending on the soil water content are investigated in climate mode, using COSMO-CLM as described in Section 4.2. Fig. 12 shows the observed average diurnal 2-m temperature range (ADTR2m) for Africa as mean over the period 2008 – 2010. This data set was created by Krähenmann et al. (2013). High values are found in arid regions, e.g. up to about 18 °C in the Sahara desert, and low values around 5 °C are found in the tropics. The high diurnal variations in the Sahara can be explained by the prevailing clear-sky conditions, which allow for a pronounced heating during daytime and for an efficient cooling during nighttime, and by the absence of vegetation, as well. In the tropics, the climate is dominated by the inter-tropical convergence zone which is characterised by daily convective rainfall...
and a dense cloud cover. In combination with the evaporative cooling due to evapotranspiration from the tropical rain forest, this allows only for low diurnal variations of the near-surface temperature.

Fig. 13 illustrates the difference of the ADTR2m between COSMO-CLM employing TERRAref and the observations. In contrast to the offline simulations presented in Section 5.1, in coupled mode the atmospheric variables are not part of a forcing, therefore e.g. the 2-m temperature can respond to changes at the land surface. The ADTR2m in the model is underestimated in arid or semi-arid regions by up to 4 K, and overestimated in parts of the tropics and southern Africa. This behaviour in the dry, non-vegetated regions can be explained by the overestimation of the ground heat flux in TERRAref, found in Section 5.1. An overestimated ground heat flux consumes too much energy in the morning and during daytime which is missing for warming up the surface. In the afternoon and during nighttime, an overestimated ground heat flux returns too much energy from deeper soil layers back to the surface, acting against an efficient cooling of the surface during nighttime. Consequently, this leads to an underestimated ADTR2m in TERRAref.

Fig. 13 also shows that the ADTR2m is overestimated in large parts of southern Africa, in particular in a region along the west coast which includes the Namib desert. Following the argument before, this would not be expected in a dry region. This overestimation appears to be caused by a combination of deficiencies in the COSMO-CLM simulations. Panitz et al. (2014) found that the shortwave surface net radiation is overestimated in wide areas of southern Africa, especially for austral winter, and particularly obvious for a narrow region along the west coast. This behaviour is associated with an underestimation of the cloud fraction, documented by Kothe et al. (2014b) and Pfeifroth et al. (2012). Furthermore, the precipitation over the East of southern Africa due to the transport of moisture from the Indian Ocean during austral summer is markedly underestimated (Panitz et al., 2014). During austral winter, precipitation is underestimated over central Africa as well. In addition to the results by Kothe et al. (2014b), there are hints for deficiencies in the simulated cloud cover diurnal cycles, with an underestimated cloud fraction over southern Africa particularly during daytime in summer, leading to an overestimation of the maximum surface temperatures and consequently also of the ADTR2m (U. Pfeifroth, Goethe University Frankfurt, pers. comm., 2012).

In the experiment with COSMO-CLM using the J75 approach, a considerable improvement, i.e. an increase, of the ADTR2m in large parts of the Sahara and the Sahel by up to 3 K is found (see Fig. 14). Under dry conditions, J75 reduces the ground heat flux compared to TERRAref, which consequently increases the ADTR2m, following the argument in Fig. 13 before. The enhanced representation of the diurnal cycle of the surface temperature might improve as well the simulation of convective systems during the West African Monsoon. Some deficiencies of its representation in COSMO-CLM simulations were documented e.g. by Kothe et al. (2014a). The ADTR2m in southern Africa, which is already overestimated by COSMO-CLM with TERRAref in many regions, as discussed before, is slightly further enhanced by J75. But, this is to be expected due to the dry bias found by Panitz et al. (2014). For an improvement of this model performance, the deficiencies in the simulated cloud cover need to be reduced, but this is beyond the scope of this study.

Figure 12: Observed average diurnal 2-m temperature range [°C] (ADTR2m), i.e. the difference between daily maximum and minimum 2-m temperature, for Africa as mean over the period 2008–2010. The observational data set was created by Krähenmann et al. (2013).

Figure 13: Difference of the ADTR2m [K] between COSMO-CLM using the reference version of TERRA and the observations by Krähenmann et al. (2013) for Africa as mean over the period 2008–2010.
under dry conditions, and high under wet conditions.

Figure 14: Same as Fig. 13 but for COSMO-CLM using the Johansen (1975) approach.

6 Conclusions

Two different approaches for calculating the soil thermal conductivity \( \lambda \) as function of the soil water content were tested in the multi-layer land surface scheme TERRA of the COSMO atmospheric model. In contrast to the reference version of TERRA, which uses a constant \( \lambda \) with respect to the soil moisture, the two approaches by Johansen (1975) and McCumber and Pielke (1981) are able to respond to soil moisture variations, which yields values for \( \lambda \) being variable in time and space, here in particular in the vertical soil column. The values of \( \lambda \), and as well of the simulated ground heat flux, are low under dry conditions, and as well of the simulated ground heat flux, are low under wet conditions.

In the reference version of TERRA, \( \lambda \) is computed assuming a medium soil wetness, i.e. the average between field capacity and permanent wilting point, leading to medium values for \( \lambda \) which are kept constant in time. Under dry conditions, this leads to an overestimation of the simulated \( \lambda \) and ground heat flux, while under wet conditions they are underestimated. In this study, we were focussing on the dry cases. Here, an overestimated ground heat flux consumes too much energy in the morning and during daytime, which is missing for warming up the surface and for the other energy fluxes of the surface energy balance. In the afternoon and during nighttime, an overestimated ground heat flux returns too much energy from deeper soil layers back to the surface, acting against an efficient cooling of the surface and providing too much energy for the surface energy fluxes. In wet cases, this works in the reversed way.

The systematic overestimation of the ground heat flux under dry conditions was found in the reference version of the land surface scheme TERRA in offline mode, as well as in a coupled climate mode. This shows, once more, the potential of a land surface scheme run in offline mode as a tool for studying, analysing and understanding the governing processes in the system, which may allow for understanding the behaviour of the land surface scheme when interactively coupled to an atmospheric model.

The two approaches by Johansen (1975) and McCumber and Pielke (1981) for calculating \( \lambda \) as function of the soil water content were tested in offline mode, and in climate mode with COSMO-CLM (climate mode only for Johansen, 1975). Johansen (1975) leads to an overall improvement of the diurnal cycles of the soil temperatures at different depths (amplitude reduced), as well as at the surface (amplitude increased) under dry conditions. McCumber and Pielke (1981) shows a qualitatively similar behaviour but the effect of the dependence of \( \lambda \) on soil moisture is overestimated, yielding unfavourable results, which confirms the findings of Peters-Lidard et al. (1998). Good improvements were achieved with Johansen (1975) in particular in Africa, where the underestimation of the average diurnal 2-m temperature range under dry conditions by up to 4 K in the reference case was substantially improved. The COSMO model was originally designed for the mid-latitudes, implicitly assuming conditions of a “medium wet” climate. When applying the model to other climate zones, these assumptions may fail, as shown by the Africa simulations with COSMO-CLM in this study.

It was found that the insulation of the sub-canopy land surface by the vegetation can have a substantial effect on the ground heat flux and the surface energy balance. For instance, measurements at Falkenberg have shown that the diurnal temperature range of the soil temperature at a depth of 6 cm can be reduced by a grass cover by a factor of 2, compared to the case when the soil is not covered by vegetation. This effect should not be neglected in land surface schemes for atmospheric models, but, this is the case for the majority of the presently operational models for numerical weather prediction. The insulating effects by vegetation, in particular the shading of the sub-canopy land surface, were already described by e.g. Deardorff (1978), Avissar and Pielke (1989), Schädler et al. (1990), or Oleson et al. (2010), but until now their work is mainly used in models for climate simulations. They introduced, for instance, a shielding factor which is related to the vegetation cover. Furthermore, it was demonstrated e.g. by Schulz et al. (1998) and Vogel et al. (2015) that introducing a separate energy budget for the canopy improves the surface temperature simulated by the soil-vegetation system. In the attempt to further improve the land surface scheme TERRA, this should be considered.

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