The Influence of Soil and Vegetation Parameters on Atmospheric Variables Relevant for Convection in the Sahel

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ABSTRACT

A key issue in modeling the Sahelian climate is to correctly predict the energy fluxes between the land surface and the atmosphere. A problem faced by land surface models in the Sahel is the horizontal heterogeneity of soil and vegetation properties in the region, where measured data are scarce. Experiments have been designed to evaluate a land surface model both in offline mode and coupled to the Advanced Regional Prediction System (ARPS), a mesoscale atmospheric model. For the evaluation in offline mode, an observational dataset of 58 days from the Hydrological and Atmospheric Pilot Experiment in the Sahel (HAPEX-Sahel) is gathered to interpret the results. For the evaluation in the coupled mode, boundary layer development is simulated for 4 individual days. The model is able to reproduce the observations close to measurement errors. Sensitivity experiments are conducted to identify the most important parameters that affect the simulation of the convective available potential energy (CAPE) and the equivalent potential temperature ($\theta_e$), two variables closely linked to convection and rainfall in the Sahel. During the selected cases, model results depend significantly on the soil moisture conditions. When bare soil evaporation is dominant, the soil parameters have the most impact on CAPE and $\theta_e$, whereas in dry soil conditions, the vegetation parameters gain in importance. Differences due to the change of one parameter can reach 400 J kg$^{-1}$ for CAPE and 4 K for $\theta_e$, which is large enough to affect local convection.

1. Introduction

Mesoscale climate models have been coupled to land surface models in the Sahel in order to account for realistic representations of surface fluxes. The magnitude and partitioning of these fluxes affect the properties of the boundary layer and hence have an effect on deep convection and rainfall. Numerous studies have demonstrated the importance of this relationship for the Sahel region (Charney 1975; Clark et al. 2001; Taylor and Clark 2001; Taylor et al. 2002; Xue and Shukla 1993; Zeng et al. 1999). Therefore, there has been a large effort to develop highly complex land surface models to represent more realistically the physical processes at the atmosphere–surface interface. A problem faced by these land surface models in the Sahel is the lack of measurements of soil and vegetation parameters in the area. Bad estimations for these parameters can lead to unrealistic simulated fluxes with feedback to other meteorological parameters.

In this study, our focus is on understanding how soil and vegetation parameters affect meteorological conditions relevant for convection in the Sahel. First, a one-dimensional soil–vegetation–atmosphere transfer (SVAT) model, developed by DeRidder and Schayes (1997), is evaluated in “offline” mode. Although the importance of extensive field measurements to evaluate these kinds of models is recognized, detailed measurements apart from the Hydrological and Atmospheric
Pilot Experiment in the Sahel (HAPEX-Sahel; Gout- 
orbe et al. 1997b) are scarce for this region. For this 
study, a dataset of 58 days [day of year (DOY) 235–292] 
was gathered from the intensive observation period of 
1992. The observation period includes a significant part 
of the rainy season, with intense rainfall events, fol- 
lowed by an elongated dry period. The evaluation is 
performed on several variables: net radiation, and sen-
sible, latent, and soil heat fluxes.

Although offline testing of SVAT models may be 
useful, the results must be viewed with caution. By im-
posing meteorological forcing inputs on the land sur-
face model, the land surface is not able to evolve as it 
would if coupled to an atmospheric model. The lack of 
atmospheric feedback can lead to erroneous results in 
model behavior (Margulis and Entekhabi 2001). There-
fore, the SVAT model of De Ridder and Schayes is 
coupled to a one-dimensional version of the Advanced 
Regional Prediction System (ARPS), a mesoscale me-
teorological model developed at the University of 
Oklahoma (Xue et al. 2000, 2001). The coupled model 
is used to evaluate the development of the boundary 
layer during four days of the HAPEX-Sahel observa-
tion period. Radio sounding data for these four days, 
representing different conditions, are used to initialize 
and evaluate the model.

The land surface fluxes that define the development 
of the boundary layer have an important effect on the 
mobility state and stability of the atmosphere and 
ence on convection and rainfall. Rainfall in the Sahel 
is dominated by the passage of organized mesoscale 
convective systems from July to September. Taylor et 
al. (1997) demonstrated a strong feedback mechanism 
between the land surface and rainfall during HAPEX-
-Sahel, especially a few days after a rainfall event. Sur-
face-induced preciprom anomalies were large enough to 
influence rainfall at convective length scales, so a cor-
rect simulation of surface fluxes is crucial in this region. 
With our one-dimensional atmospheric model, rainfall 
is not simulated, but an evaluation can be made for two 
variables that are closely linked to moist atmospheric 
instability, that is, the convective available potential en-
ergy (CAPE) and the equivalent potential temperature 
($\theta_e$). Monkam (2002) has demonstrated the similarity of 
patterns of rainfall and CAPE in West Africa. So the 
ability of the coupled model to simulate $\theta_e$ and CAPE 
is relevant in relation to atmospheric modeling in this re-
region.

Subsequently, the sensitivity of the modeled $\theta_e$ and 
CAPE to the soil and vegetation parameters of the 
 coupled model is tested. It is recognized that soil hy-
drology and vegetation play an important role by alter-
ing the partitioning of land surface fluxes. In a review 
article, Pielke (2001) states that soil and vegetation dy-
namics are as much a part of the climate system as 
are the atmospheric variables. In semiarid areas in 
the United States this topic has been the subject of 
modeling studies for some decades, as the models 
evolved from simple, coarse-resolution models to 
highly complex models with a high resolution (Grasso 
2000; Lanicci et al. 1987; Shaw et al. 1997; Trier et al. 
2004; Ziegler et al. 1995). The main focus of these stud-
ies was the influence of the soil moisture distribution on 
the atmospheric environment, and more particular the 
formation of a dryline and the evolution of the bound-
ary layer. Small differences in initial soil moisture dis-
tribution were found to have a large impact on the 
thermodynamic stability.

Also in the Sahelian region a number of studies have 
focused on the influence of vegetation and land use 
change on the climate by using coupled climate models 
(Hoffmann and Jackson 2000; Taylor et al. 2002; Xue 
and Shukla 1993). The influence of soil parameters on 
climate simulations is less explored although Osborne 
et al. (2004) found that it can have a significant impact. 
A recent study of Kahan et al. (2006) with an offline 
land surface model also demonstrated the impact of 
some individual soil and vegetation parameters on the 
surface energy and water balance in the Sahel. They 
found that leaf area index, stomatal resistance, and soil 
hydraulic conductivity are crucial to allow enough 
evaporation in this region.

The goal of this study is to provide further under-
standing of the role of soil and vegetation parameters in 
climate modeling in the Sahel. Therefore, the sensitivity 
of atmospheric variables relevant for convection initia-
tion in our coupled model is tested for a broad range of 
soil and vegetation parameters. In this paper, section 2 
and appendix B describe the land surface model and 
the HAPEX-Sahel dataset, used for the offline evalu-
ation. A description of the atmospheric model and the 
setup for the online experiments are also given in this 
section. In section 3 the model runs are evaluated and 
the impact of soil and vegetation parameters is tested. 
Conclusions are given in section 4.

2. Methods

a. Offline evaluation

1) LAND SURFACE MODEL

The land surface model used in this study is a one-di-

and six soil layers. Energy and water budgets are calculated separately for both the soil and the vegetation and weighted by their respective cover fractions. Soil heat and water transfer equations are numerically solved using the Crank–Nicholson iteration technique. Canopy temperature and moisture content (both inside and on the leaves) are solved with prognostic equations. A detailed description of the model is summarized in appendix B. A particular feature of the model is the use of a physically based parameterization of the transpiration process, requiring only one empirical function in the stomatal resistance formulation.

The radiation transfer scheme takes into account both short- and longwave radiation. To realize the radiation scheme's calculations, seven input parameters are needed: the green and dry grass albedo ($\alpha_g, \alpha_d$), the dry soil albedo ($\alpha_s$), the vegetation and soil emissivity ($\varepsilon_v, \varepsilon_s$), the leaf area index (LAI), and the green leaf fraction ($g_v$). The vegetation is further described by five parameters: the displacement height ($d$), the roughness length for momentum ($z_0m$), the minimum stomatal resistance ($r_{st}$), the total plant resistance ($r_p$), and the shape parameter of the root density profile ($\beta$), following the equation of Jackson et al. (1996).

The soil is represented in the model by six soil layers with variable thicknesses (0.005, 0.025, 0.07, 0.3, 0.6, and 1 m), with a total depth of 2 m. Soil water potential and hydraulic conductivity are expressed by the approach of Clapp and Hornberger (1978). Their equations require four input parameters: the saturated soil moisture content ($\theta_{sat}$), the saturated soil water potential ($\psi_{sat}$), the saturated hydraulic conductivity ($K_{sat}$) and the shape parameter of the water retention curve ($b_{ch}$).

The model is forced by meteorological input that is either obtained from measurements or supplied by an atmospheric model. The input consists of the precipitation intensity, the air temperature, density and specific humidity, the wind speed, and the incoming short- and longwave radiation. The output supplied by the model comprises, amongst others, the net radiation, sensible, latent, and soil heat fluxes.

Since soil and vegetation conditions in the Sahel are very particular, the model is adjusted to deal with them.
To take into account the effect of a soil crust, which is present in most parts of the HAPEX-Sahel area, \( K_{\text{sat}} \) of the topsoil layer is lowered by one order of magnitude, following Vandervaere et al. (1997). The growth and wilting of the vegetation during the observation period is implemented by using a time-variable LAI, vegetation cover fraction, green leaf fraction, and displacement height. These vegetation parameters had constant values in the original model of De Ridder and Schayes (1997).

2) FIELD DATA

The study area is situated near the Banizoumbou village in the “East Central Super Site” (13°33'49"N, 2°40'93"E) of the HAPEX-Sahel experiment (Goutorbe et al. 1997b) (Fig. 2). The region is covered mainly by aeolian sand of variable depth with an underlying laterite plateau. The soil in the study area is sandy, with slight fractions of silt and clay, and a horizontal heterogeneous soil surface crust on top (Braud et al. 1997; d’Herbes and Valentin 1997). The local vegetation consists of a sparse herbaceous layer in a fallow savannah area. The grass layer of the regional savannah ecosystem is composed essentially of annual plants with a number of C\(_3\) species (\textit{Mitarcarpus scaber}, \textit{Tribulis terrestris}) growing in competition with C\(_4\) species (\textit{Aristida mutabilis}, \textit{Aristida adscensionis}, \textit{Cenchrus biflorus}; Monteny et al. 1997).

For the meteorological forcing of the SVAT model, use is made of in situ–measured air temperature and humidity at 2 m and wind speed at 10 m. Air density, incoming downward solar radiation, and rainfall are taken from the nearby meteorological station of Banizoumbou. The time resolution of these measurements is 10 min. Incoming longwave radiation data are not available on the site, and are taken from the southern site, about 60 km from the study area. The evolution of the vegetation parameters is deduced from local measurements of vegetation height, LAI, green LAI, and cover fraction. The shape parameter of the root density profile (\( \beta \)) is deduced from herbaceous root profile data from the western supersite.

Surface energy fluxes are obtained from measurements by a micrometeorological station, which used the
Bowen ratio method (Monteny et al. 1997). Net radiation is measured at 9 m and the soil heat flux at 0.025-m depth (mean of three sensors). Latent and sensible heat fluxes are calculated from the gradient of water vapor pressures and temperatures at 0.2 and 1.2 m above the canopy. The time resolution of the surface flux measurements is 20 min. Kahan et al. (2006) report that the root-mean-square error of the daily mean fluxes is 9.94 W m$^{-2}$.

3) PARAMETER IDENTIFICATION

The evaluation period [day of the year (DOY) 235–292] includes a significant part of the rainy season, followed by a long period of drought. Figure 3 shows the rainfall amounts during the evaluation period. A detailed list of the model parameters, along with their sources, is presented in Table 1. As mentioned before, the parameters are chosen prior to the model runs and no calibration is performed.

The albedo parameters ($\alpha_g$, $\alpha_d$, $\alpha_s$) were measured over herbaceous vegetation in the Sahel by Allen et al. (1994). Vegetation and soil emissivity values ($\varepsilon_v$, $\varepsilon_s$) were reported by Braud et al. (1997) for a nearby study area. Measurements of stomatal resistances of Sahelian vegetation were made by Hanan and Prince (1997). The roughness length for momentum ($z_{0m}$) is deduced from local vegetation data, obtained by Monteny et al. (1997). Since no in situ measurements for the total plant resistance ($R_p$) are available, the value reported by Braud et al. (1997) for fallow savannah is retained. Root density profiles of herbaceous vegetation were measured at the Western Super Site by Hanan et al. (1997). The shape parameter of the root density profile ($\beta$) is deduced from these data.

As mentioned before, the soil hydraulic properties of the model are the Clapp and Hornberger parameters for sandy soils, with addition of a soil crust. Initial soil temperatures and soil moisture contents were implemented from in situ–measured data by Monteny et al. (1997). Since the maximum measured soil moisture content in the area, measured by Braud et al. (1997), seemed to be less than the Clapp and Hornberger maximum value, relative values were transferred to the model, which is a commonly used method (B. van den Hurk 2005, personal communication).

b. Online model runs

1) ATMOSPHERIC MODEL

The model used for the simulation of the boundary layer development is the Advanced Regional Predictive System, version 4.5.2 (Xue et al. 2000, 2001). ARPS is a nonhydrostatic mesoscale atmospheric model including conservation equations for momentum, heat, mass, water (vapor, liquid, and ice), subgrid-scale turbulent kinetic energy and the state equation of moist air. The model contains detailed parameterizations for cloud

<table>
<thead>
<tr>
<th>Vegetation and surface parameters</th>
<th>Value</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Green grass albedo ($\varepsilon_v$)</td>
<td>0.18</td>
<td>Allen et al. (1994)</td>
</tr>
<tr>
<td>Dry grass albedo ($\alpha_d$)</td>
<td>0.24</td>
<td>Allen et al. (1994)</td>
</tr>
<tr>
<td>Dry soil albedo ($\alpha_s$)</td>
<td>0.25</td>
<td>Allen et al. (1994)</td>
</tr>
<tr>
<td>Vegetation emissivity ($\varepsilon_v$)</td>
<td>0.96</td>
<td>Braud et al. (1997)</td>
</tr>
<tr>
<td>Soil emissivity ($\varepsilon_s$)</td>
<td>0.97</td>
<td>Braud et al. (1997)</td>
</tr>
<tr>
<td>Minimum stomatal resistance ($r_{st, sm}$, s m$^{-1}$)</td>
<td>125</td>
<td>Hanan and Prince (1997)</td>
</tr>
<tr>
<td>Roughness length for momentum ($z_{0m}$, m)</td>
<td>0.38</td>
<td>Monteny et al. (1997)</td>
</tr>
<tr>
<td>Roughness length for heat ($z_{0h}$, m)</td>
<td>0.0038</td>
<td>Braud et al. (1997)</td>
</tr>
<tr>
<td>Total plant resistance ($R_p$, s)</td>
<td>$9.4 \times 10^8$</td>
<td>Braud et al. (1997)</td>
</tr>
<tr>
<td>Root distribution coefficient ($\beta$)</td>
<td>0.952</td>
<td>Hanan et al. (1997)</td>
</tr>
</tbody>
</table>

<table>
<thead>
<tr>
<th>Soil parameters</th>
<th>Value</th>
<th>Source</th>
</tr>
</thead>
<tbody>
<tr>
<td>Saturated soil moisture content ($\theta_s$, m$^3$ m$^{-3}$)</td>
<td>0.395</td>
<td>Clapp and Hornberger (1978)</td>
</tr>
<tr>
<td>Saturated soil water potential ($\psi_s$, m)</td>
<td>0.121</td>
<td>Clapp and Hornberger (1978)</td>
</tr>
<tr>
<td>Saturated hydraulic conductivity crust ($K_{sat1}$, m s$^{-1}$)</td>
<td>$1.76 \times 10^{-5}$</td>
<td>Vandervaere et al. (1997)</td>
</tr>
<tr>
<td>Saturated hydraulic conductivity ($K_{sat2}$, m s$^{-1}$)</td>
<td>$1.76 \times 10^{-4}$</td>
<td>Clapp and Hornberger (1978)</td>
</tr>
<tr>
<td>Water retention curve exponent ($\beta_{sw}$)</td>
<td>4.05</td>
<td>Clapp and Hornberger (1978)</td>
</tr>
</tbody>
</table>
microphysics, cumulus convection, and radiation transfers. The simulations are performed with ARPS in a one-dimensional mode. In the vertical there are 100 levels extending to over 20 km, with the first level at 10-m height and 28 levels in the lowest 1 km.

As the model is one-dimensional, we choose situations where the advection terms are likely to be small. The four days used here—21 August, 10 and 18 September, and 8 October 1992 (DOY 234, 254, 262, 282)—were selected by evaluating the stationarity of the time series of the available radio soundings. Atmospheric profiles of temperature and humidity are almost in steady state above the atmospheric boundary layer for these days, so the influence of advection is minimal. During these selected periods, the flow close to the surface is predominantly southwesterly, comparatively moist, and topped by an easterly flow of dry continental air (Goutorbe et al. 1997a).

The land surface model is coupled to ARPS for the model runs, having the same characteristics and setup as in the “offline” simulations. Except for the LAI and vegetation cover fraction, which were deduced for the modeled days from all available HAPEX-Sahel measurements in fallow savannah area. The initial soil moisture contents are taken from the output of the offline simulations. The initial atmospheric temperature, humidity, and wind speed are based on the HAPEX-Sahel radio sounding ascents for 0500 LT on 21 August and 8 October 1992 and for 0700 LT on 10 and 18 September 1992 (Dolman et al. 1997a). The radio soundings were launched near the village of Hamdallay, Niger (13°33′N, 2°24′E; Fig. 2). The coupled model simulates the development of the boundary layer from the initial time until 1700 LT, when the output was evaluated with HAPEX-Sahel radio sounding ascents (Dolman et al. 1997b).

2) SENSITIVITY EXPERIMENTS

To investigate if the soil and vegetation parameterizations in the coupled model have a significant influence on the model results, sensitivity experiments were carried out. In relation to the previous offline sensitivity experiments of Kahan et al. (2006), it is interesting to see which parameters have an important impact on the model results and if this impact is constant during the evaluation period. Estimating the measurement errors and possible range for each parameter is the first part of the analysis. Afterward, a series of simulations are performed in which the parameter under study is adapted to the upper or lower boundary of uncertainty (Table 2, where the names for each sensitivity experiment are also shown).

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The uncertainty in their measurements of LAI and vegetation cover fraction during the HAPEX-Sahel campaign were reported by Monteny et al. (1997). Stomatal resistances of the fallow savannah area in the

<table>
<thead>
<tr>
<th>Date</th>
<th>LAI</th>
<th>cov</th>
<th>$R_p$</th>
<th>$R_{st}$</th>
<th>$z_{om}$</th>
<th>$z_{oh}$</th>
<th>$K_{sat}$</th>
<th>$b_{ch}$</th>
</tr>
</thead>
<tbody>
<tr>
<td>21 Aug 1992</td>
<td>+</td>
<td>1.35</td>
<td>0.38</td>
<td>1.41E+09</td>
<td>188</td>
<td>0.08</td>
<td>0.00004</td>
<td>8.80E-04</td>
</tr>
<tr>
<td></td>
<td>-</td>
<td>0.90</td>
<td>0.30</td>
<td>9.40E+08</td>
<td>125</td>
<td>0.04</td>
<td>0.0004</td>
<td>1.76E-04</td>
</tr>
<tr>
<td>18 Sep 1992</td>
<td>+</td>
<td>2.10</td>
<td>0.68</td>
<td>1.41E+09</td>
<td>188</td>
<td>0.08</td>
<td>0.00004</td>
<td>8.80E-04</td>
</tr>
<tr>
<td></td>
<td>-</td>
<td>1.40</td>
<td>0.54</td>
<td>9.40E+08</td>
<td>125</td>
<td>0.04</td>
<td>0.0004</td>
<td>1.76E-04</td>
</tr>
<tr>
<td>10 Sep 1992</td>
<td>+</td>
<td>2.28</td>
<td>0.64</td>
<td>1.41E+09</td>
<td>188</td>
<td>0.08</td>
<td>0.00004</td>
<td>8.80E-04</td>
</tr>
<tr>
<td></td>
<td>-</td>
<td>1.52</td>
<td>0.51</td>
<td>9.40E+08</td>
<td>125</td>
<td>0.04</td>
<td>0.0004</td>
<td>1.76E-04</td>
</tr>
<tr>
<td>8 Oct 1992</td>
<td>+</td>
<td>1.61</td>
<td>0.74</td>
<td>1.41E+09</td>
<td>188</td>
<td>0.08</td>
<td>0.00004</td>
<td>8.80E-04</td>
</tr>
<tr>
<td></td>
<td>-</td>
<td>1.07</td>
<td>0.59</td>
<td>9.40E+08</td>
<td>125</td>
<td>0.04</td>
<td>0.0004</td>
<td>1.76E-04</td>
</tr>
</tbody>
</table>

Table 2. Identification of the sensitivity experiments.
Sahel show a lot of variation (Hanan and Prince 1997), which is captured in our parameter range. In their simulations, Goutorbe et al. (1997a) and Braud et al. (1997) used a value of 0.07 m for \( z_{\text{om}} \) and measurements in arid and semiarid areas demonstrated the variability of this parameter (Visser et al. 2005). The sensitivity of \( z_{\text{om}} \) and \( z_{\text{mh}} \) is tested by changing these parameters by respectively 100% and one order of magnitude that corresponds to a realistic variability for the Sahel (Verhoef et al. 1997). For simulations over Europe, De Ridder and Schayes (1997) used a total plant resistance of 0.5 \( \times \) 10\(^{9} \) s, so variations of \( r_p \) in the Sahel should be inside the parameter range used in our experiments.

Soil hydraulic properties in the Sahel show a large horizontal heterogeneity and 11 main types of surface soils are distinguished in a 1° square around Niamey, Niger (Cuenca et al. 1997; d’Herbès and Valentin 1997). However, measurements of these soil properties are scarce in this region. Braud et al. (1997) derived soil parameters from field measurements in their study. They differ most importantly from the Clapp and Hornberger parameters by the saturated hydraulic conductivity and the shape parameter for the water retention curve. So the sensitivity of the model to these parameters is investigated by varying them to values close to the measured ones.

3. Results and discussion
   a. Offline evaluation

   Figure 4 compares the sensible and latent heat fluxes calculated by the land surface model with the observed fluxes during the evaluation period (DOY 235–292). Measurements were interrupted during DOY 237 and during a few hours of DOY 243. The overall performance of the model is satisfactory, with the exception of DOYs 247–250 and 252–256 where sensible heat is overestimated and latent heat is underestimated. These are some dry days after a rainfall event in the wet period indicating that drying out of the soil may occur too fast. Sensible heat fluxes are well predicted for most of the time during the rainy season. In the dry season, starting at DOY 265, the model predictions during daytime are correct. During the night, however, the observations show a lot of variation, which is not captured by the model. The observed variation is probably due to instability of the Bowen ratio method during the night when differences in air temperature and vapor pressure are low. Chanzy (1996) also reports this problem in the HAPEX-Sahel database (http://www.ird.fr/hapex). Modeled and observed latent heat fluxes agree reasonably well during the wet period, with a slight underestimation for some days. The decrease of latent heat fluxes going into the dry season is captured by the model, although there are also problems during night time, when the deviations are large.

   For the comparison between observed and simulated fluxes, mean daily root-mean-square errors (RMSEs) and biases are calculated. The results for all evaluation variables are provided in Table 3. As mentioned before, the RMSEs of the mean daily measured fluxes are around 10 W m\(^{-2} \) (Kahan et al. 2006). Our results are close to this value, except for a larger RMSE of latent heat fluxes. The deviations are the largest in the days following a rainfall event, as can be seen in Fig. 4. The drying of the topsoil layer in the model occurs a little bit too fast, probably due to the Clapp and Hornberger parameters for sandy soils, which are not identical to the soil characteristics in the Sahel region. The parameter values are retained as the overall performance of the model is satisfactory and they form a consistent set of parameters. Changing one of them might result in a physically unrealistic combination of parameter values.

   Relative errors are on the order of 10%–20% of observed mean daily fluxes for sensible, latent, and soil heat fluxes, and less than 10% for net radiation. The mean biases for sensible and ground heat fluxes show a good correspondence between the model and the observations. However, the mean bias for latent heat fluxes and to a lesser extent for net radiation depicts an overestimation of modeled fluxes. As Table 3 shows that the bias is small during daytime, the overestimation is influenced by the nighttime fluxes, when the Bowen ratio method is not so accurate.

   When coupled to an atmospheric model, the accuracy of the predicted fluxes during daytime is the most important in relation to the temperature and moisture state of the atmosphere. During the day, the fluxes are also much larger than during nighttime. In this context, the model performs reasonably well. These results suggest that the Clapp and Hornberger soil parameters can be applied successfully in a region with such large spatial variability of soil properties as the Sahel, even on a plot scale (Vandervaere et al. 1997). The implementation of a soil surface crust was necessary to reduce the bare soil evaporation after rainfall events. Previous simulations without a soil crust (not shown here) overestimated the latent heat fluxes by 50% shortly after rainfall events and underestimated them during the dry season. The reduction of \( K_{\text{sat}} \), which mimics the soil crust, has a significant effect on latent heat fluxes in the wet season, which are now in good agreement with the observations.

   The objective of this evaluation phase was to assess how our modified land surface model would fare in the
Sahelian environment. The simulation has shown that the surface fluxes during daytime are well simulated for most of the time during the 2-month period. In a next phase, the model is coupled to a one-dimensional atmospheric model without further modification.

**b. Online evaluation**

The structure of the boundary layer at a particular location responds to the surface forcing of land cover types with a typical length scale of 10 km, depending on

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**TABLE 3. RMSEs and biases (W m\(^{-2}\)) for the offline validation run for daily mean, daytime mean [0900–1700 local time (LT)], and nighttime mean (1700–0900 LT) fluxes.**

<table>
<thead>
<tr>
<th>Model variable</th>
<th>RMSE daily</th>
<th>Daytime</th>
<th>Nighttime</th>
<th>Bias daily</th>
<th>Daytime</th>
<th>Nighttime</th>
</tr>
</thead>
<tbody>
<tr>
<td>(H)</td>
<td>9.34</td>
<td>29.23</td>
<td>7.23</td>
<td>-1.05</td>
<td>4.99</td>
<td>-3.75</td>
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<td>12.22</td>
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<td>(G)</td>
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<td>11.26</td>
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<td>-20.05</td>
<td>10.46</td>
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wind speed and atmospheric stability (Samuelsson and Tjernström 1999). In the HAPEX-Sahel area, a number of land cover types will be present in the “footprint” of the boundary layer profile, whereas in our simulations only the fallow savannah is represented. Nevertheless, the simulations are valuable as the fallow savannah is the dominant land cover near the station of Hamdallay and the fluxes are representative of the total area average flux, with errors less than 15% (Dolman et al. 1997a).

Figure 5 shows the observed and modeled dewpoint and air temperatures at the end of the simulation (1700 LT). In all cases, the profiles agree reasonably well with the observations. However, the moisture content of the boundary layer is clearly underestimated on 8 October, in dry soil conditions (last rainfall on 15 September). In the other case of dry soil conditions, 21 August (last rainfall on 15 August), the humidity of the boundary layer is underestimated to a lesser extent. The boundary layer development in wet soil conditions (10 and 18 September) is better simulated by the model. The model does not reproduce the sharp vertical gradients around the inversion, for example, on 18 September, which is partly due to the coarse vertical model resolution.

To evaluate the accuracy of the model simulations, RMSEs and mean biases are calculated for the lowest 5000 m of the profile, where surface fluxes play a role. In addition, the height of the measured and modeled boundary layer was calculated by looking at the maximum in the derivative of the potential temperature with respect to the height. Furthermore, a comparison is made between mean values of the potential temperature from 300 to 1300 m (below the inversion) and 3000

FIG. 5. Observed (dashed lines) and modeled (solid lines) atmospheric profiles of dewpoint and air temperature at 1700 LT.
to 4000 m (above the inversion), which can be seen as a measure for inversion strength. Lapse rates below and above the inversion are also calculated in these regions.

Another method to intercompare the model simulations for the different cases is the hit rate HR (percent), which considers a desired model accuracy \( A \): 

\[
HR = \frac{100}{n} \sum_{i=1}^{n} m_t
\]

with \( m_t = 1 \) if difference (model, observations) < \( A \) and 0 if difference (model, observations) ≥ \( A \), where \( n \) is the number of comparison data (55 data points for each variable). The desired model accuracy \( A \) is taken as 2 K for air and dewpoint temperature and 1.5 g kg\(^{-1}\) for specific humidity. These values are chosen as they cause significant variations between the sensitivity experiments, discussed in section 3c, thereby facilitating the evaluation of these sensitivity runs. The hit rate ranges from 100% if all model results are within \( A \) of the observations to 0% if none are and is a reliable overall measure for describing model performance (Schlünzen and Katzfey 2003).

Table 4 summarizes the RMSE, bias, and hit rate results for the modeled profiles at 1700 LT of the selected cases. The overall performance of the model is satisfactory, as the mean bias and root-mean-square errors are mostly low. The air temperature is relatively well captured by the model, as root-mean-square errors are below 2 K. The model has slightly more problems with simulating the dewpoint temperature, especially on 8 October, which can also be seen in Fig. 5. The specific humidity is well predicted in the first three cases, but shows the same trend as the dewpoint temperature on 8 October.

Table 5 gives an overview of the resulting boundary layer heights and the potential temperature differences between 300–1300 and 3000–4000 m (\( \Delta \theta \)) (K) for the model runs compared to the observations.

Table 6 lists the lapse rates below the inversion (lapse rate 1) and above the inversion (lapse rate 2) (K km\(^{-1}\)) of the modeled variables compared to the observations.
rates below and above the inversion are given. The lapse rates below the inversion (lapse rate 1) compare well for all cases except for some slight deviations on 10 September. Above the inversion (lapse rate 2) the lapse rate for $T_d$ differs largely on 21 August and 10 September, which can also be seen in Fig. 5. As the variations around 4000 m in the observed profile on 10 September were not observed in the initial profile, they are probably due to advection of air.

Overall, the coupled model is able to reproduce the most important characteristics of boundary layer development in the Sahel. Although the surface was not dealt with in detail, the transfer of soil moisture and heat to the atmosphere is well simulated in most cases, except for the case of 8 October, in very dry soil conditions. As the simulations were one-dimensional, the advection that may have occurred during the simulated days most probably also plays a role in the deviation of the model results.

c. Impact of soil and vegetation parameters

The goal of our sensitivity experiments is to investigate how sensitive model results of CAPE and $\theta_e$ are to variations in model parameters and if this variation changes throughout the season. The values of CAPE and $\theta_e$ are calculated for an air parcel at a height of 500 m, well within the boundary layer. As CAPE is a result of the total temperature profile, even minor advection can play an important role in deviations between modeled and observed CAPE. However, the comparison between different sensitivity integrations is still valuable, as they all started from the same initial profile. This problem does not occur for $\theta_e$, which is more robust for comparison with the observation data.

Figures 6 and 7 show the results of the sensitivity studies for, respectively, $\theta_e$ and CAPE. The original model (reference) is able to simulate $\theta_e$ within a range of 2 K and CAPE within a range of 200 J kg$^{-1}$. The graphs of both variables differ slightly: $\theta_e$ is overestimated by all experiments on 21 August and 10 September, while CAPE is clearly overestimated on 21 August and 8 October. The most striking feature of both figures is the fact that the sensitivity of the model differs significantly throughout the season. On 21 August, the variation between the sensitivity experiments is rather small, due to the low vegetation cover and the limited
water content of the soil, which inhibits large variations in moisture flux to the atmosphere. When the soil is wet, on 10 and 18 September, the soil parameters $K_{\text{sat}}$ and $b_{\text{ch}}$ have a significant influence on modeled $\theta_e$ and CAPE as a large part of the evaporation during these days comes from the soil. The situation is totally different on 8 October, when the soil is very dry, and all the evaporation is due to vegetation transpiration. In this case, the model is very sensitive to changes in the vegetation parameterization, whereas the soil parameters play a minor role.

These results are in agreement with observations during HAPEX-Sahel. Taylor et al. (1997) found that in the first days after rainfall events, the latent heat flux was especially due to bare soil evaporation, which is at that moment 3 times higher than the vegetation transpiration. After two or three days, the soil evaporation drops to nearly zero and hence the influence of soil parameters will diminish. Taylor and Clark (2001) reported differences in specific humidity of 1 g kg$^{-1}$ in the boundary layer due to differences in vegetation and soil structure. This is as large as differences due to antecedent rainfall and leads to estimated differences in CAPE at convective length scales of 600 J kg$^{-1}$ (Taylor et al. 1997). Our study confirms that a change of one parameter can already lead to maximum differences of CAPE of around 400 J kg$^{-1}$. In comparison with typical total CAPE values of 1000 to 2000 J kg$^{-1}$, such contrasts represent a significant local perturbation. These values show that the land surface can modulate the convection in the Sahel as Taylor and Clark (2001) also found in their study.

The sensitivity of the model is not evenly distributed over all the investigated parameters. The roughness lengths for momentum and heat have minor impacts on model results, which can be clearly seen in the graphs of $\theta_e$. This is probably caused by the low surface wind speed on the selected days, which is around 2 m s$^{-1}$. Of all vegetation parameters, the total plant resistance has the largest influence on the model output. Variations in LAI, vegetation cover, and stomatal resistance also have a large impact, but the magnitude depends on the situation. For the soil parameters, the model is most sensitive to the Clapp and Hornberger exponent.

The results of the sensitivity studies are summarized in Table 7, where the hit rates of all the experiments are
shown. The hit rates compare well with Figs. 6 and 7. Some parameters like $b_{ch}$, $r_p$, and LAI have a large influence on model results, whereas $z_{0m}$ and $z_{0h}$ have a minor influence. No clear trend is visible from these results: for one simulation day the model can be improved by lowering a parameter while for another day it may be the other way around. By calibrating one parameter, it is impossible to get a correct representation of the profiles for all the simulated cases.

Studies of the sensitivity of coupled atmospheric or climate models to changes in individual parameters are scarce in the Sahel region. The sensitivity of complex coupled models to initial soil moisture and vegetation patterns has been a subject of study in semiarid areas in the United States (Ziegler et al. 1995; Shaw et al. 1997) but the focus was more on patterns and not on single parameters. Most recent studies in the Sahel focus on the effect of land use changes and hence the change of several vegetation parameters at a time (Clark et al. 2001; Taylor et al. 2002; Xue and Shukla 1993). Recently, Kahan et al. (2006) investigated the role of LAI, stomatal resistance, hydraulic conductivity, and thermal diffusivity in offline simulations. In their simulations, spanning a 2-yr period, LAI and stomatal resistance had the most significant effect, which compares well with our results during dry soil conditions. With this study, further insight into the model sensitivity topic in the Sahel area is provided by using online model runs. Large sensitivities to single vegetation or soil parameters are confirmed with the coupled model, whereby the sensitivity depends on the soil moisture conditions.

4. Conclusions

In this paper, a one-dimensional land surface model is applied to a fallow savannah site of the HAPEX-Sahel experiment. Model results are compared with a dataset spanning 58 days, including a part of the rain season, followed by a long period of drought. The model is capable of reproducing observed sensible and latent heat fluxes with an average error around 10–20 W m$^{-2}$, which is close to measurement errors.

Subsequently, the land surface model is coupled to a one-dimensional atmospheric model (ARPS) to test its ability of reproducing observed atmospheric profiles. To this purpose, four different days are selected from the HAPEX-Sahel radio sounding dataset. The selected days represent the different soil moisture and atmospheric conditions during the observation period. Model runs started at 0500 or 0700 LT and lasted till the late afternoon (1700 LT), the moment of maximum boundary layer development.

The profiles simulated by the coupled model are in good agreement with the observations, except for an underestimation of the boundary layer humidity and an overestimation of boundary layer height in very dry conditions. As land surface models of this kind are mostly coupled to GCMs or mesoscale climate models to predict rainfall, the ability of the coupled model to reproduce variables that are closely linked to atmospheric instability is tested. The selected variables are the convective available potential energy and the equivalent potential temperature. The coupled model is able to reproduce late-afternoon values of CAPE within 200 J kg$^{-1}$ and $\theta_e$ within 2 K.

The sensitivity of modeled CAPE and $\theta_e$ to variations of soil and vegetation parameters is furthermore tested in a series of sensitivity experiments. The sensitivity of the coupled model varies for each simulation case. In wet soil conditions, the model is most sensitive to soil parameters as evaporation occurred mainly from the
soil. In dry soil conditions, when moisture fluxes are due to vegetation transpiration, the vegetation parameters have the most influence on model results. The variation of a single parameter never leads to improved model results for all the cases; the effect depends significantly on the case under study. The simulated variations in CAPE and $\theta_v$ (with a maximum around 400 J kg$^{-1}$ and 4 K) are large enough to have an effect on passing mesoscale convective systems. Hence a proper estimation of both soil and vegetation parameters is important in relation to climate modeling studies in the Sahel area.

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APPENDIX A

List of Symbols

- $b_{ch}$: Shape parameter of water retention curve
- CAPE: Convective available potential energy (J kg$^{-1}$)
- $c_p$: Specific heat of air at constant pressure (J kg$^{-1}$ K$^{-1}$)
- $d$: Displacement height (m)
- $D$: Soil hydraulic diffusivity (m$^2$ s$^{-1}$)
- $G$: Ground heat flux (W m$^{-2}$)
- $g_o$: Green leaf fraction
- $H$: Sensible heat flux (W m$^{-2}$)
- $h_r$: Ground surface relative humidity
- HR: Hit rate
- $k$: Von Kármán constant
- $K$: Soil hydraulic conductivity (m s$^{-1}$)
- $K_{sat}$: Saturated hydraulic conductivity (m s$^{-1}$)
- LAI: Leaf area index
- LE: Latent heat flux (W m$^{-2}$)
- $q$, $q_o$, $q_v$: Specific humidity (kg kg$^{-1}$)
- $q_s$, $q_c$: Ground surface and canopy specific humidity (kg kg$^{-1}$)
- $R_n$: Net radiation (W m$^{-2}$)
- $r_{am}$, $r_{ah}$: Aerodynamic resistance for momentum and heat (s m$^{-1}$)
- $r_{as}$, $r_{av}$: Ground surface and canopy aerodynamic resistance for heat (s m$^{-1}$)
- $r_c$: Canopy stomatal resistance (s m$^{-1}$)
- $R_i_b$: Surface layer bulk Richardson number
- $r_p$: Total plant resistance (s)
- $r_{ai}$: Soil-root resistance (s)
- $r_{st}$: Minimum stomatal resistance (s m$^{-1}$)
- $R_w$: Gas constant for water vapor (J kg$^{-1}$ K$^{-1}$)
- $T$, $T_a$: Air temperature (K)
- $T_d$: Dewpoint temperature (K)
- $T_s$, $T_v$: Ground surface and canopy temperature (K)
- $u_a$: Wind speed at reference level (m s$^{-1}$)
- $W_s$: Plant water flux (kg m$^{-2}$ s$^{-1}$)
- $z$: Soil depth (m)
- $z_{0_b}$, $z_{0_m}$: Roughness length for heat (m) and momentum (m)
- $\alpha_s$, $\alpha_d$, $\alpha_g$: Dry soil albedo, dry grass albedo, green grass albedo
- $\beta$: Root distribution coefficient
- $\delta$: Wet vegetation fraction
- $\varepsilon_s$: Soil emissivity
- $\varepsilon_v$: Vegetation emissivity
- $\eta$: Soil moisture content
- $\varphi$: Fraction of the roots in the soil
- $\mu$: Soil thermal conductivity (W m$^{-1}$ K$^{-1}$)
- $\psi_c$, $\psi_v$: Critical leaf water potential (m)
- $\psi_s$, $\psi_c$: Soil and canopy water potential (m)
- $\psi_{sat}$: Saturated soil water potential (m)
- $(\rho c)_s$: Soil volumetric heat capacity (J m$^{-3}$ K$^{-1}$)
- $\theta$: Potential temperature (K)
- $\theta_e$: Equivalent potential temperature (K)
- $\theta_{sat}$: Saturated soil moisture content (m$^3$ m$^{-3}$)

APPENDIX B

Formulation of the SVAT Model

The SVAT model used in this paper computes fluxes of energy and momentum between the land surface and the atmosphere and is of intermediate complexity. A full description of the model can be found in De Ridder and Schayes (1997). It combines a simple turbulence transfer scheme with a rather sophisticated transpiration and soil water scheme. The number of parameters was kept as small as possible while preserving the essential physics of the soil–vegetation–atmosphere processes. The model contains two heat and moisture sources: the vegetation and the soil surface.

The aerodynamic resistances for momentum and heat between the surface and a reference level $z_a$ are calculated as follows:

$$r_{am} = \frac{1}{k^2 u_a} \left[ \ln\left( \frac{z_a - d}{z_{0_m}} \right) \right] F_m^{-1}(R_i_b)$$

$$r_{ah} = \frac{1}{k^2 u_a} \ln\left( \frac{z_a - d}{z_{0_m}} \right) \ln\left( \frac{z_a - d}{z_{0_b}} \right) F_h^{-1}(R_i_b), \quad (B1)$$
where \( k = 0.4 \) is von Kármán’s constant, \( u_c \) is the wind speed at the reference level, \( F_m \) and \( F_v \) are stability corrections (Louis 1979), \( R_{ib} \) is the bulk Richardson number.

The sensible heat flux is written as

\[
H = \rho_s c_p \frac{T_v - T_a}{r_{av}} + \rho_s c_p \frac{T_s - T_a}{r_{as}},
\]

where \( T_v \) and \( T_s \) are respectively the canopy and ground temperature, \( c_p \) is the specific heat of air, and \( r_{av} \) and \( r_{as} \) are the vegetation and ground aerodynamic resistances for heat, calculated by dividing \( r_{sh} \) through the fractional cover.

The latent heat flux is decomposed into the vegetation and ground contribution as follows:

\[
LE = L_w \rho_u \left[ \frac{q_{sat}(T_v) - q_s}{r_{av}} + (1 - \delta) \frac{q_{sat}(T_s) - q_s}{r_{as}} \right. \\
+ \left. \frac{h_g q_{sat}(T_s) - q_s}{r_{as}} \right] + \frac{h_s q_{sat}(T_s) - q_s}{r_{as}},
\]

with \( R_w \) being the gas constant for water vapor.

The water flow through the plants is written as the sum of the transpiration fluxes from each of the soil layers. These fluxes are proportional to the difference in water potential between the leaves and the corresponding soil layer:

\[
W_w = \sum_{i=1}^{6} \rho_u \Phi_i \left( \psi_v + d \right) - \psi_i \left( \frac{r_p}{r_s} \right),
\]

where \( \Phi_i \) is the fraction of the roots in soil layer \( i \), \( \psi_v \), and \( \psi_i \) are the water potentials of the soil layer \( i \), respectively, and \( r_p \) and \( r_s \) are the total plant and soil-root resistance.

The canopy stomatal resistance \( \delta_c \), which appeared in the definition of LE, is computed as follows:

\[
\delta_c = \frac{r_s}{g_s \text{LAI}} \left( 1 - \frac{\psi_v}{\psi_c} \right)^{-1},
\]

where \( g_s \) is the green leaf fraction and \( \psi_c \) is the leaf water potential at which total stomatal closure occurs.

The transfer of heat in the soil is calculated by means of the thermal diffusion equation:

\[
\frac{\partial T}{\partial t} = \frac{1}{(\rho c)_s} \frac{\partial}{\partial z} \left[ \mu(\eta) \frac{\partial T}{\partial z} \right],
\]

where \( z \) is the depth in the soil, \( (\rho c)_s \) is the volumetric heat capacity of the soil, and \( \mu(\eta) \) is the soil thermal conductivity.

The transport of liquid water in the soil is computed by means of Richard’s equation, completed with a transpiration extraction term:

\[
\frac{\partial \eta}{\partial t} = \frac{\partial}{\partial z} \left[ D(\eta) \frac{\partial \eta}{\partial z} - K(\eta) \frac{W(z)}{\rho_w} \right].
\]

The first term on the right-hand side is a diffusion term, where \( D(\eta) \) is the soil hydraulic diffusivity, the second term is the soil hydraulic conductivity \( K(\eta) \) and represents the gravitational flow, and the last term is the integrated transpiration water flux between the ground surface and depth \( z \).

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