The Influence of Rain Sensible Heat and Subsurface Energy Transport on the Energy Balance at the Land Surface

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In land surface models, which account for the energy balance at the land surface, subsurface heat transport is an important component that reciprocally influences ground, sensible, and latent heat fluxes and net radiation. In most models, subsurface heat transport parameterizations are commonly simplified for computational efficiency. A simplification made in all models is to disregard the sensible heat of rain, \( H_r \), and convective subsurface heat flow, \( q_{cv} \), i.e., the convective transport of heat through moisture redistribution. These simplifications act to decouple heat transport from moisture transport at the land surface and in the subsurface, which is not realistic. The influence of \( H_r \) and \( q_{cv} \) on the energy balance was studied using a coupled model that integrates a subsurface moisture and energy transport model with a land surface model of the land surface energy balance, showing that all components of the land surface energy balance depend on \( H_r \). The strength of the dependence is related to the rainfall rate and the temperature difference between the rainwater and the soil surface. The rainwater temperature is a parameter rarely measured in the field that introduces uncertainty in the calculations and was approximated using the either air or wet bulb temperatures in different simulations. In addition, it was shown that the lower boundary condition for closing the problem of subsurface heat transport, including convection, has strong implications on the energy balance under dynamic equilibrium conditions. Comparison with measured data from the Meteostation Haarweg, Wageningen, the Netherlands, shows good agreement and further underscores the importance of a more tightly coupled subsurface hydrology–energy balance formulation in land surface models.

**Abbreviations:** CLM, Common Land Model; LSM, land surface model.

Land surface models (LSMs) parameterize the mass, energy, and momentum balances at the land surface. They serve as a lower boundary condition in atmospheric models and upper boundary condition in hydrologic models. Several studies have shown the importance of accurately representing land surface processes in climate models (Bonan et al., 2002; Guo et al., 2006; Koster et al., 2006). Land surface models are also often applied in stand-alone or offline mode in comparison and parameterization studies such as PILPS, as well as in the interpretation of measured data (Henderson-Sellers and Henderson-Sellers, 1995; Shao and Henderson-Sellers, 1996). Land surface models seek to close the energy balance at the land surface, which encompasses net radiation, \( R_{net} \), latent heat flux, \( LE \), sensible heat flux, \( H \), and ground heat, \( G \). Calculation of the individual terms requires a number of simplifying assumptions because the physical processes that are associated with the different energy fluxes are very complicated and cannot be incorporated into LSMs using a first-principles approach. Furthermore, additional terms of the full energy balance equation, such as energy storage in the canopy, are omitted. As an example, two land surface models in current use, TESSEL (Betts et al., 2001) and Noah (Ek et al., 2003), still use a crude parameterization to link the soil heat flux to the surface temperature, without proper representation of the canopy, and also do not take into account energy that is used for photosynthesis (see Jacobs et al. [2008] for the relevance of the various terms). This may lead to an accumulated energy balance error, which gives rise to large inaccuracies of those terms that are determined as a residual, such as ground heat flux.

Some of the simplifications and relative significance of different processes have been studied in detail. For example, Peters-Lidard et al. (1998) studied the effect of the thermal conductivity, \( \lambda \), on \( G \) and found that parameterizations accounting for the dependence of \( \lambda \) on soil moisture strongly influences the land surface energy balance. Bittelli et al. (2008) developed a coupled model of subsurface heat, water vapor, and liquid water transport to simulate evaporation in bare soil. They showed that vapor transport cannot be neglected in calculations of evaporation under dry conditions and demonstrated the importance of the surface resistance formulation. Maxwell and Miller (2005) and Kollet and Maxwell (2008) studied the influence of three-dimensional subsurface moisture transport on the land surface.
energy balance, including the presence of a shallow, free water table. They identified a critical water table depth. If the water table is located above that depth, land surface processes strongly depend on the location of the water table. This dependence also results in spatial structures in the energy variables that are also found in measured data for cases where the water table depths vary spatially (Chen et al., 1997; Maxwell et al., 2007; Kollet and Maxwell 2008).

In field studies, closure of the energy balance is attempted through eddy covariance measurements in conjunction with radiation and ground heat flux monitoring. In these studies, however, relatively large errors of up to 20% are not uncommon, mainly due to fetch uncertainty (e.g., Twine et al., 2000) and instrumental errors (Foken and Wichura, 1996). Recent corrections of ground heat flux measurements, usually performed using heat flux plates and temperature profiles, have been applied and shown to be useful in improving $G$ estimates (Heusinkveld et al., 2004; Liebethal et al., 2005; Jacobs et al., 2008). Additional corrections include incorporation of additional energy storage components of the land surface such as the canopy enthalpy change and photosynthesis flux. These corrections help in closing the energy balance and may alleviate up to 10% of the original error (Meyers and Hollinger, 2004; Jacobs et al., 2008); however, closure of the energy balance still remains a difficult task.

The value of $G$ is important in determining the land surface temperature right at the land surface–atmosphere interface and serves as the upper boundary condition of the subsurface heat transport equations. Inspection of the parameterization of subsurface heat transport reveals that only transport by conduction is commonly considered (Dai et al., 2003). This simplification basically decouples heat transport from subsurface moisture transport, which is described usually by a one-dimensional implementation of Richards’ equation in LSMs. In reality, however, the coupling of heat and moisture transport can be quite strong if flow velocities and temperature differences are large, leading to considerable transport of heat through the process of convection in the shallow subsurface. This process has been used by hydrogeologists to determine groundwater recharge rates from vertical temperature logging with ensuing inverse modeling (e.g., Silliman and Booth, 1993; Ronan et al., 1998). Detailed studies exist on coupled subsurface moisture, heat, and also vapor transport (Philip, 1957; Sophocleous, 1979; Alvenas and Jansson, 1997; Saito et al., 2006), yet some fundamental aspects have not been discussed in the coupling of the terrestrial hydrologic and energy cycles.

Including the process of convection raises an important question at the land surface boundary, namely the impact of the sensible heat of rainfall on the different components of the energy balance. If the temperature of the rainwater differs from the temperature of the soil surface, precipitation generates a sensible heat flux at the land surface. The significance of this additional sensible heat flux has only been studied previously in coupled ocean–atmosphere simulations in a tropical environment by Gosnell et al. (1995). They found that the sensible heat of rain cannot be neglected during many rainfall events. Deru and Kirkpatrick (2002) modeled the coupled heat and moisture transfer from buildings including the sensible heat of rain. They approached the subsurface heat transport problem in a concise but empirical manner without analyzing feedbacks with the energy balance at the land surface. They found that convection can have a significant impact on subsurface heat transport.

In addition, to close the problem of subsurface heat transport, a lower boundary condition at the bottom of the simulated soil column needs to be defined that is commonly unknown because of the lack of field data. In practice, two approaches have been applied in LSMs without much reflection on the impact on the energy balance at the land surface. For example, in the Common Land Model (CLM), a no-flow temperature boundary condition is applied at the bottom of the soil column (Dai et al., 2003). In contrast, in the Community Noah Land Surface Model, the lower boundary consists of a Dirichlet condition with a constant temperature defined at a certain depth below the land surface (Mahrt and Ek, 1984; Peters-Lidard et al., 1997). Because the soil column implemented in LSMs is relatively short (on the order of $10^0$ m), the lower boundary will strongly influence the energy fluxes at the land surface.

In this study, the influence on the energy balance of subsurface heat transport by convection, i.e., moisture movement, and the lower boundary condition for temperature was studied. Numerical simulations were performed on a one-dimensional soil column, including the energy balance at the land surface and coupled heat and fluid flow in the subsurface. The effect of the sensible heat of rain was included, applying two approximations of the rainwater temperature to account for uncertainty because of the lack of measured data. The simulations were forced by a measured atmospheric time series from the Netherlands. Additionally, the influence of the lower temperature boundary condition on the energy balance at the land surface was quantified by simulating dynamic equilibrium conditions using varying Dirichlet and a no-flow condition. Comparison with measured soil temperature profiles shows that the applied model is a useful tool to study coupled subsurface moisture and heat transport influenced by land surface processes.

**Theory**

**Subsurface Heat Transport**

In discussion of the theory, it is important to distinguish clearly between the ground heat flux and subsurface heat transport. In LSMs, the ground heat flux, $G$, is one of the major components of the energy balance at the land surface in addition to net radiation, $R_{\text{net}}$, sensible heat, $H$, and latent heat, LE:

\[ R_{\text{net}} = H + LE + G \]  

[1]

In CLM and other LSMs, $G$ is estimated as the residual of Eq. [1], with $R_{\text{net}}$, $H$, and LE obtained from physical parameterizations. Thus,

\[ G = R_{\text{net}} - H - LE \]  

[2]

Commonly, LSMs apply the ground heat flux right at the land surface, i.e., the immediate interface between the soil and the adjacent air that constitutes the lower boundary of the atmosphere and the upper boundary of the subsurface. Thus, $G$ is part of the boundary condition of the problem of subsurface heat transport that is commonly expressed in LSMs as
\[
\frac{\partial}{\partial t}\left[ \theta C_w T + (1 - \phi) C_s T \right] = \nabla \cdot \lambda \nabla T + q_T \tag{3}
\]

where \( T \) is temperature [K], \( \theta \) is the volumetric moisture content \([L^3 T^{-3}]\), \( t \) is time \([T]\), \( \rho_w \) is the density of water \([M L^{-3}]\), \( C_w \) is the specific heat capacity of water \([J M^{-1} K^{-1}]\), \( C_s \) is the heat capacity of the solid \([M L^{-1} T^{-2} K^{-1}]\), \( \phi \) is the volumetric porosity \([L^3 L^{-3}]\), \( \lambda \) is the thermal conductivity \([W M^{-1} L T^{-1}]\), and \( q_T \) is a general thermal sink–source term \([W L^{-3}]\).

Thus, changes in subsurface heat storage are assumed to be solely caused by conductive heat transfer \( q_{cd} = \lambda \nabla T \) and source–sinks. The coupling with soil moisture occurs only through changes in the effective heat capacity in the first term on the left-hand side.

A more complete formulation includes the transport by convection, \( q_{cv} \), which is expressed in the second term on the right-hand side:

\[
\frac{\partial}{\partial t}\left[ \theta C_w T + (1 - \phi) C_s T \right] = \nabla \cdot \lambda \nabla T - \nabla \cdot C_w T q + q_T \tag{4}
\]

where \( q \) is moisture mass flux \([M L^{-2} T^{-1}]\). This formulation incorporates the transport of heat by moisture, which is more realistic and more closely couples moisture and heat transport than Eq. [3]. In this study, the major assumptions applied in Eq. [4] are that \( \rho_w \) is constant under the studied temperature conditions, fluxes due to mechanical dispersion are negligible, and \( \lambda \) does not depend on the water content in our case. The latter assumption was applied to clearly extract the effects under investigation, i.e., convection, the sensible heat of rain, and the lower boundary condition. Note that Eq. [4] is physically not exhaustive. For a rigorous discussion of the thermodynamics of soil moisture, see, e.g., Bear and Nitaq (1995).

As mentioned above, at the ground surface, the upper boundary condition consists of the sum of all incoming thermal fluxes originating from land surface processes. In LSMs, these are assumed to be equal to \( G \) following Eq. [2]. Thus at the land surface at \( z = 0 \),

\[
\lambda \nabla T = G \tag{5}
\]

An additional thermal flux may originate, however, from infiltrating water due to rainfall and canopy throughfall. This results in additional terms in the equation of the upper boundary condition if the temperature of the infiltrating water, \( T_{rain} \), is different from the temperature of the soil surface, \( T_s \). Thus, at the land surface at \( z = 0 \),

\[
\lambda \nabla T + C_w T q = G + H_1 \tag{6}
\]

with

\[
H_1 = \rho_w C_w (T_{rain} - T_s) q_{inf} \tag{7}
\]

where \( q_{inf} \) is the infiltration flux stemming from precipitation and canopy throughfall \([L T^{-1}]\). We refer to \( H_1 \) as the sensible heat of rain. In the energy balance equation (Eq. [1]), \( H_1 \) is part of the sensible heat flux component, \( H \), which involves transfer across the land surface interface caused by a moisture movement, \( H_s \), and a temperature gradient, \( H_t \) (Deru and Kirkpatrick 2002):

\[
H = H_s + H_t \tag{8}
\]

There exist no accessible data on rainwater temperature to our knowledge. Therefore \( T_{rain} \) was treated as an uncertain input parameter in this study and assumed equal to the wet-bulb temperature, \( T_{rain} = T_{wb} \), or the air temperature at the reference height, \( T_{rain} = T_{air} \). An approximate equation for the wet-bulb temperature was obtained from Jensen et al. (1990).

To close the problem of subsurface heat transport, a lower boundary condition is required. Common approaches include the application of a constant temperature at a certain depth, \( z = z_{bc} \) (e.g., Noah):

\[
T_{bc} = T_{const} \tag{9}
\]

or a no-flow boundary condition (e.g., CLM):

\[
\lambda \frac{\partial T}{\partial z} = 0 \tag{10}
\]

Subsurface Moisture Transport

The subsurface heat transport equation is coupled with a moisture transport equation (i.e., Richards’ equation) mainly through \( \theta \) in the heat storage term and \( q \) in the convection term:

\[
q = -k(z) k_r(\psi) \frac{\partial \rho_w \psi}{\partial z} \nabla (\psi - \rho_w g z) \tag{11a}
\]

where \( k(z) \) is the intrinsic saturated permeability of the porous medium \([L^2]\), \( z \) is the vertical space coordinate \([L]\), \( \psi \) is the pressure \([M L^{-1} T^{-2}]\), \( k_r(\psi) \) is the relative permeability as a function of pressure (dimensionless), \( \mu_w \) is the dynamic viscosity \([M L^{-1} T^{-1}]\), and \( g \) is the gravitational constant \([L T^{-2}]\).

The functional relationship of \( k_r(\psi) \) is expressed via the van Genuchten function:

\[
k_r(\psi) = \frac{[1 - (\alpha \psi)^{1 - n}]^{1/(1 - n)}}{[1 + (\alpha \psi)^{1 - n}]^{1/(1 - n)}} \tag{11b}
\]

where \( \alpha \) \([L^2 T^{-1}]\) and \( n \) (dimensionless) are soil-specific parameters.

The mass balance of subsurface soil moisture under variably saturated conditions is

\[
\phi \frac{\partial S_w(\psi)}{\partial t} = \nabla \cdot q + q_{gs} \tag{12}
\]

where \( S_w \) is the relative saturation, \( S_w = \theta/\phi \) (dimensionless), and \( q_{gs} \) is a general source–sink term \([T^{-1}]\). Note that \( q_{gs} \) includes fluxes from transpiration that are applied differentially with depth based on the parameterization of the vertical root distribution, which decreases exponentially with depth.

The problem of variably saturated subsurface flow is closed using a Neumann boundary condition at the land surface, \( z = 0 \):

\[
-k(z) k_r(\psi) \frac{\partial \rho_w}{\partial z} (\psi - \rho_w g z) = EI \tag{13}
\]

where EI represents exfiltration and infiltration fluxes due to evaporation from the soil and precipitation, respectively \([T^{-1}]\).
In this study, a constant-pressure boundary condition, $\psi = \psi_{bc}$, is used at the bottom to reflect a free water table at a certain depth $z = z_{bc}$ below the land surface.

The choice of different combinations of boundary conditions for coupled heat and fluid flow has direct physical implications requiring careful consideration that are discussed in detail below.

**Materials and Methods**

**Numerical Implementation of ParFlowE**, **Coupling with CLM, and Verification**

A new parallel, numerical model, ParFlowE, that simulates coupled moisture and heat transport based on Eq. [4] has been developed. This model is based on the original version of ParFlow, which simulates variably saturated groundwater flow and integrated overland flow only. The parallel and numerical implementation of ParFlow has been discussed in detail before (Ashby and Falgout, 1996; Jones and Woodward, 2001; Kollet and Maxwell, 2006) and is not repeated here. In ParFlowE, coupled three-dimensional subsurface heat transport was implemented in identical fashion using a cell-centered finite difference scheme in space and an implicit backward Euler differencing scheme in time. Upwinding was used with the convection term to guarantee numerical stability, albeit at the cost of accuracy. The solution algorithms used in ParFlow were fully exploited in ParFlowE by extending the solution vector of the Newton–Krylov method to two dimensions. Note that additional infrastructure for a multidimensional solution vector was implemented for this application but also to facilitate relatively straightforward implementation of, e.g., multiphase flow in the future. It is important to mention that functional relationships were implemented in ParFlowE to relate density and viscosity to temperature and pressure, and thermal conductivity to saturation. These parameters were assumed constant in the present study, however, in order to extract information on the influence of convection and the sensible heat of rain. Therefore, these relationships were not used in the simulations.

The ParFlow model was verified using an analytical solution for conductive–convective heat transport with a temperature-dependent sink term along a semi-infinite, one-dimensional rod. Figure 1 shows the comparison of the analytical and numerical solutions using ParFlowE in dimensionless form, which exhibit excellent agreement. The discrepancies near the right boundary stem from the fact that a constant-temperature boundary condition was used in the simulation with ParFlowE instead of the semi-infinite conditions used in the analytical solution.

To study the influence of convection, $q_{cv} = C_w T q$, and the sensible heat of rain, $H_l$, a second version called ParFlowEnc (ParFlowE with no convection) was implemented that neglects both of these processes and is based on Eq. [3] and [5]. The results were then juxtaposed with simulations using ParFlowE based on Eq. [4] and [6] accounting for $q_{cv}$ and $H_l$.

The Common Land Model (CLM; Dai et al., 2001), which accounts for the mass and energy balance at the land surface, provides the upper boundary condition (for energy and mass) for ParFlowE. Therefore the CLM was implemented modularly into ParFlowE.

In previous studies, the CLM has already been coupled to ParFlow in a modular fashion by replacing the one-dimensional hydrology in the CLM based on isolated columns with the three-dimensional, variably saturated, subsurface flow with integrated overland flow of ParFlow (Maxwell and Miller, 2005; Kollet and Maxwell, 2008). Following this original approach, ParFlowE and ParFlowEnc were coupled with the CLM. In addition to replacement of the subsurface hydrology, the one-dimensional subsurface heat transport in the CLM was replaced in a similar fashion with the three-dimensional heat transport equation including the process of convection of ParFlowE. (Note that only one-dimensional single column experiments were performed in this study.) To reflect the modularity of the applied approach, the newly developed coupled model is named ParFlowE[CLM] and ParFlowEnc[CLM].

In summary, the module CLM calculates the mass and energy balances at the land surface that lead to $G$, $H_l$, $q$, and $E$. Those are passed to the subsurface moisture and heat transport algorithm of ParFlowE[CLM] and ParFlowEnc[CLM] as boundary conditions and sink terms based on Eq. [6], [12], and [13], respectively, and are used in the ensuing computations of the subsurface moisture and temperature fields. The temperature and moisture values of the top model layers that are required to calculate the mass and energy balances for the next time step are then passed back to the CLM module.

Because simulations using ParFlowE[CLM] were to be used to assess whether convection and $H_l$ are important, the numerical implementation of the boundary conditions that include the atmospheric forcing term, $G$, at the top needed to be verified in ParFlowE[CLM] and ParFlowEnc[CLM]. This was done by performing a 1-yr simulation using both the original heat transport formulation in CLM and the formulation in ParFlowEnc[CLM]. Both codes are basically identical if the upper temperature boundary condition is implemented correctly in ParFlowEnc[CLM]. In this boundary verification exercise, the same atmospheric time series was utilized including long- and shortwave radiation, air temperature, precipitation, wind speed, barometric pressure, and specific humidity in 1-h time steps. Figure 2 shows the temperature results of the surface layer (Layer 1) and two deeper layers (Layers 2 and 3) for the entire year. The results are basically identical, as expected. Yet small discrepancies exist close to the freezing
point between the results using the original heat transport formulation in CLM and the results using ParFlowE[CLM] without convection. This is due to the fact that ParFlowE does not yet accurately account for freezing processes. Therefore, the months with freezing temperatures have not been included in the ensuing analysis if not explicitly stated otherwise.

**Modeling Rationale, Data, and Model Setup**

The modeling rationale was based on the hypothesis that subsurface heat transport by convection, \( q_{cv} \), and the influence of the sensible heat of rain, \( H_{s} \), cannot be neglected under certain conditions. This was tested by comparing simulations based on the Eq. [3] and the corresponding upper boundary condition defined by Eq. [5] with simulations based on Eq. [4] and the corresponding boundary condition defined by Eq. [6] and [7]. The former was done using ParFlowE[CLM] without convection and the latter was done using ParFlowE[CLM] with convection. That both models differ in the additional terms in Eq. [4] and the boundary condition of Eq. [6], and basically provide the same result as the CLM when convection is neglected was shown above (Fig. 2). Thus differences in the results can be directly attributed to the influence of \( q_{cv} \) and \( H_{s} \).

The model setup was based on the Meteostation Haarweg, Wageningen, the Netherlands (http://www.met.wau.nl/haarweg-data/; verified 20 Aug. 2009) of the Meteorology and Air Quality Section, Wageningen University and Research Center (51°58′ N, 5°38′ W, altitude 7 m above sea level). The Meteostation Haarweg served only the purpose of providing a real-world climatological background and input data set instead of using a purely synthetic data. This long-term observation site is situated in grassland and covers various meteorological variables. Hourly averaged observations of incoming long- and shortwave radiation, air temperature and humidity at a 1.5-m height, precipitation, wind speed at a 10-m height, and barometric pressure during 2002 were used to force the simulation. In addition, manual observations of the water table depth were used. Additional data monitored at the site are soil temperature and soil heat flux, which were used in the analysis of the simulation results. Soil temperature is continuously monitored at 5-, 10-, and 20-cm depths under bare soil, and at 5-, 10-, 20-, 50-, and 100-cm depths under grass. Soil heat flux is measured at the 5-cm depth under bare soil and grass. The soil is mainly clay, originating from floodplain deposits of the Rhine River. The site can be considered as well watered. The climatology is characterized by a rainfall surplus (difference between precipitation and evapotranspiration) of some 410 mm yr\(^{-1}\) (based on actual evaporation for the years 1994–2005). On average, the surplus occurs in the months September to May and a small deficit occurs in the other months. The annual mean air temperature is 10.6°C (Jacobs et al., 2007). For a more detailed description of the test site, see Jacobs et al. (2008).

Table 1 contains the model input parameters, which were derived from literature values and physical reasoning (Schaap and Leij, 1998). Note that the bold values constitute the reference run. The average water table depth, wtd, of 0.85 m was calculated from manually measured data from the site and was held constant in the simulations. The vertical discretization of the one-dimensional column was 0.025 m and a time step of \( t = 3600 \) s was used. The length of the column was 6 m with a constant-pressure boundary condition at the bottom of 50501.724 Pa to simulate \( \text{wtd} = 0.85 \) m. Thus, water could leave the domain at the bottom following the Darcy relationship at the boundary. The temperature boundary condition at the bottom was of Dirichlet type to simulate a constant \( \text{T}_{bc} \) at a certain aquifer depth. A reference value of \( \text{T}_{bc} = 283 \) K was obtained from the literature (Fest et al., 2007), which is consistent with the value provided by Jacobs et al. (2007) and was varied in the simulations. In the application of the lower temperature boundary condition, a shallow no-flow condition was also applied, which is physically unreasonable, although part of the subsurface heat transport parameterization in the CLM.

The land cover type was grass (bare soil conditions were not considered at this stage). The CLM uses the International Geosphere–Biosphere Program classification scheme (Dai et al., 2003). In the simulations, the default parameters in the CLM for grassland were applied.

The atmospheric time series used to force the model in hour time steps consisted of observations during 2002, which constituted a relatively wet year (the data can be downloaded from http://www.met.wau.nl/haarweg-data/). Hourly precipitation and temperature are shown in Fig. 3. The model was run for successive years using the same forcing to obtain a dynamic

**Table 1. Summary of input parameters for the one-dimensional column simulation.** The parameter values of the reference simulation are indicated in italics.

<table>
<thead>
<tr>
<th>Parameter</th>
<th>Value</th>
<th>Units</th>
</tr>
</thead>
<tbody>
<tr>
<td>Soil</td>
<td></td>
<td></td>
</tr>
<tr>
<td>van Genuchten parameter ( \alpha )</td>
<td>( 1.0 \times 10^{-4} )</td>
<td>Pa(^{-1})</td>
</tr>
<tr>
<td>van Genuchten parameter ( \beta )</td>
<td>2.0</td>
<td>–</td>
</tr>
<tr>
<td>Porosity, ( \phi )</td>
<td>0.451</td>
<td>–</td>
</tr>
<tr>
<td>Relative residual water content, ( S_{res} )</td>
<td>0.15</td>
<td>–</td>
</tr>
<tr>
<td>Intrinsic saturated permeability, ( k )</td>
<td>( 1.0 \times 10^{-12} )</td>
<td>m(^2)</td>
</tr>
<tr>
<td>Water table depth, ( wtd )</td>
<td>0.85</td>
<td>m</td>
</tr>
<tr>
<td>Thermal</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Specific heat capacity of water, ( C_{w} )</td>
<td>( 4.188 \times 10^{6} )</td>
<td>kg ms(^{-2}) K(^{-1})</td>
</tr>
<tr>
<td>Specific heat capacity of solid, ( C_{s} )</td>
<td>( 1.932 \times 10^{6} )</td>
<td>kg ms(^{-2}) K(^{-1})</td>
</tr>
<tr>
<td>Thermal conductivity, ( \lambda )</td>
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<td>W m(^{-1}) K(^{-1})</td>
</tr>
<tr>
<td>Boundary condition temperature, ( T_{bc} )</td>
<td>280, 283, 286</td>
<td>K</td>
</tr>
<tr>
<td>Rain temperature, ( T_{rain} )</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Air temperature ( T_{air} )</td>
<td></td>
<td></td>
</tr>
<tr>
<td>Wet-bulb temperature ( T_{wbt} )</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>
Results and Discussion

Influence of Convection and Sensible Heat of Rain

We focus on the months with air temperatures above freezing. Soil freezing and the associated processes, such as sublimation, are very complicated processes that are currently not accurately accounted for in ParFlowE. Therefore, to prevent any bias, the months May to November of 2002 are included in the major analysis from the atmospheric time series if not explicitly mentioned otherwise.

First, it was necessary to obtain an estimate of $H_l$. Since the magnitude of $H_l$ is determined by the rainfall rate and the temperature difference between the rainwater and the soil surface, $H_l$ must be determined for individual events. Several important uncertainties need to be considered to assess the impact of $H_l$ on the energy balance at the land surface. In general, the rainwater temperature is not known a priori. There exist no reliable measurements of rainwater temperature at the Meteostation Haarweg. Therefore the air and wet-bulb temperatures were assumed as the upper and lower bounds, respectively. In addition, the water infiltrating at the surface is reduced by, e.g., canopy storage. Thus, soil surface infiltration rates from canopy throughfall as provided by the CLM for grass cover were used in the analysis of the reference run. Figure 4 shows the frequency distribution of the potential sensible heat flux for a bin size of about 1.5 W m$^{-2}$ and $H_l$ was taken as the sensible heat of infiltrating rainwater, $H_{	ext{infl}}$, due to the temperature difference of the infiltrating water and the soil.

Figure 4 demonstrates that $H_l$ values can be quite large with a maximum negative value of approximately $H_{	ext{infl}} = -60$ W m$^{-2}$ during rainfall events. Application of $T_{	ext{rain}} = T_{	ext{air}}$ instead of $T_{	ext{wbt}}$ does not have a significant effect on the frequency distribution of $H_l$. The majority of $H_l$ values are negligibly small, yet a relatively large number are on the order of $H_l \sim 1$ to 10 W m$^{-2}$.

Inspection of Eq. [7] shows that $H_l$ depends on the temperature gradient and the rainfall rate. To demonstrate this dependence, for $T_{	ext{rain}} = T_{	ext{wbt}}$ the instantaneous values of the temperature difference $\Delta T = T_{\text{rain}} - T_{\text{air}}$, $q_{\text{infl}}$, and $H_l$ were plotted in Fig. 5. During the warm months of the year, $\Delta T$ is mainly negative (~3 K) because the rain temperature is generally lower than the surface temperature. This has a cooling effect, which is reflected in the negative $H_l$ values. $H_l$ values correlate well with large $q_{\text{infl}}$ and, because $q_{\text{infl}}$ may vary across orders of magnitude, it has a large influence on the variability and maximum values of $H_l$ in comparison to $\Delta T$.

From Fig. 4 and 5 it is obvious that $H_l = H_{\text{infl}}$ potentially influences the temperature of the topsoil layer and, thus, the energy balance at the land surface during individual rainfall events. As an example, the effect on soil temperatures is shown in Fig. 6 for a single rainfall event. Directly after the rainfall event, the temperature of the topsoil drops by ~1 K for $T_{\text{rain}} = T_{\text{wbt}}$. The effect is less pronounced for $T_{\text{rain}} = T_{\text{air}}$ because the temperature difference between the soil surface and the rainwater is smaller in this case. The different soil temperatures converge after about 9 h after the rain stopped. There were also differences in the soil temperatures at the beginning of 5 August, when temperatures and evapotranspiration reached their maximum values. This was caused by upward redistribution of deeper soil moisture, of varying temperature for the different cases, resulting from upward evapotranspiration fluxes.
This difference in soil temperatures is not clearly seen at a depth of 5 cm below the land surface in the measured data or in the simulations. This is primarily due to the temperature perturbation being damped by the heat storage capacity of the overlying material. More detailed temperature measurements at shallow depths are necessary to corroborate this finding in the field. Additionally, convection has a relatively small influence because of the clay soil, which has relatively small saturated hydraulic conductivity values, which results in relatively small Darcy velocities for a given gradient. In the case of sandy soil, for example, the temperature disturbance due to $H_1$ may be detectable at greater depth.

The event-based nature of $H_1$ can be further characterized by plotting the absolute ratios of convective vs. conductive heat fluxes, $q_{cv}/q_{cd}$, for the shallow subsurface. Figure 7 does this and reveals that relatively few events lead to $q_{cv}/q_{cd} > 0.1$, with a maximum value of about 0.5. They occur mainly from May to August, when temperature differences between the rainwater and the soil surface are largest during strong rainfall events. It is important to note that Fig. 7 was derived from fluxes between the first and second model layer, i.e., at a depth of 2.5 cm below the land surface, because of the finite difference discretization. Thus, $q_{cv}$ at this depth is reduced by moisture and energy storage in the top model layer.

The question remains whether $H_1$ due to superposition of individual rainfall events during a certain time period influences the energy balance at the land surface.
balance equation (Eq. [2]), the influence of $H_l$ gets traced back to the ground heat flux through the coupling of the different energy components in the CLM and the upper boundary condition of the subsurface heat transport.

For practical purposes, it is important to analyze on which time scales consideration of heat convection influences estimates of average energy fluxes. Therefore, hourly, daily, and weekly differences of $\Delta G$, $\Delta H$, $\Delta LE$, and $R_{net}$ were calculated between the simulations including and neglecting the process of convection. The results for the months with air temperatures above freezing are shown in Fig. 9. The simulations including convection were performed using $T_{rain} = T_{wbt}$ and $T_{rain} = T_{air}$. The largest influence of $H_l$ due to precipitation was observed for LE and $H$. Maximum differences for hourly fluxes were up to 100 W m$^{-2}$ during and immediately after rainfall and evapotranspiration events.

In October and November, $E_{av},$ was also significantly affected by $H_l$. because of relatively small $R_{net}$ values and the dominance of $G$. From Fig. 9 it is obvious that $H_l$ must be taken into account if energy balances are calculated on an hourly basis. Negative and positive differences in the individual components of the energy balance are averaged across longer time periods and become insignificant for averaging periods longer than 1 wk for $H$ and LE (monthly data are not shown here). This is due to the noisy behavior of the differences, i.e., it appears that neglecting $H_l$ may not introduce a significant bias in the results. In the case of $G$, however, Fig. 8 implies that convection influences the monthly averages when ground heat flux is a significant component of the energy balance. The apparent impact is quite large because the monthly mean $G$ estimates are relatively small. Figure 8 also reveals that, for example, a decrease in $H$ and $G$ results in an equivalent increase in LE because $R_{net}$ is partitioned into $H$, $G$, and LE following energy conservation in the CLM.

The results have important implications for the analysis of measured data from eddy covariance stations and suggest that data up to a day after rainfall events should be discarded if conventional analysis techniques are to be used that neglect $H_l$. The simulations were performed, however, for a test site where strong rainfall events are relatively rare. In cases with strong convective rainfall events with large temperature differences between the air and the land surface, the impact of $H_l$ can be expected to influence average estimates of the different energy components on the order of $\sim 1\%$ for as long as one week.

A yearly energy balance and also water balance shows that convection, rain sensible heat, and the lower temperature boundary condition have only minor effects on the different components (Table 2). Changes in the energy balance were negligible and changes in the water balance were on the order of 1 to 2%. It should be noted that the difference of precipitation and total evapotranspiration, $P - ET$, corresponds to the yearly average groundwater recharge in this case. This is because the change in soil moisture storage during the year is zero due to the spin-up procedure performed in the analysis and there is no surface runoff (all water potentially ponding at the surface infiltrated during the simulations). It is also remarkable that the estimated groundwater recharge value of about 375 mm is relatively close to the water surplus value of some 410 mm without calibration of the model to the data.

The large instantaneous differences in $H$ and LE shown in Fig. 9 are due to the parameterizations in the CLM. Relatively small changes in the energy regime at the land surface may result in large changes in energy fluxes. For example, the temperature gradient between the ground surface and the canopy (in this case, grass) in the calculation of the sensible heat flux may change, which will feed back into the closure of the energy balance at the land surface. This is illustrated by

$$H_g = \sigma G \rho_s c_p C_{soil} u_s (T_a - T_s)$$

where $\sigma$ is the vegetation fraction (dimensionless), $\rho_s$ is the density of air [M L$^{-3}$], $c_p$ is the specific heat of dry air [L$^2$ T$^{-2}$ K$^{-1}$], $C_{soil}$ is the transfer coefficient between the canopy air and the underlying ground, $u_s$ is the wind velocity within the canopy [L T$^{-1}$], $T_s$ is the temperature at the land surface [K], and $T_a$ is the air temperature within the canopy [K].

In turn, changes in $H_g$ will result in changes in $T_s$ that are calculated following an iterative procedure in the CLM. Additionally, humidity at the ground surface and thus evaporation, $E_g$, are influenced by changes in $T_s$ following

$$E_g = \sigma G \rho_s C_{soil} u_s (q_s - q_a)$$

where $q_s$ is the surface air specific humidity [M M$^{-1}$], and $q_a$ is the air specific humidity within the canopy [M M$^{-1}$]. The ground surface air specific humidity, $q_s$, is directly dependent on the $T_s$ and is calculated using

$$q_s = q_{sat} \exp \left( \frac{\psi_s}{RT_s} \right)$$

where $q_{sat}$ is the saturated specific humidity [M M$^{-1}$], $\psi_s$ is the matric potential at the ground surface [M L$^{-1}$ T$^{-2}$], and $R$ is the gas constant [L$^2$ T$^{-2}$ K$^{-1}$]. The value of $E_g$ will again influence $q_a$ reciprocally in the iterative procedure in the CLM and through

$$q_a = \frac{c_A q_{sat} + c_V q_{sat} + c_G q_g}{c_A + c_V + c_G}$$

![Fig. 8. Cumulative ground heat fluxes, $G_{cum,n}$, normalized by ground heat fluxes neglecting convection, $G_{cum,nc}$, for the simulations using rainwater temperatures, $T_{rain}$, taken as the air temperature, $T_{air}$, and the wet-bulb temperature, $T_{wbt}$.](image-url)
where $q_{atm}$ is the atmospheric specific humidity [M M$^{-1}$]; $q_{f, sat}$ is the leaf specific humidity [M M$^{-1}$], and $c_A$, $c_V$, and $c_G$ are exchange coefficients for the atmosphere, vegetation, and ground surface, respectively. This will lead to changes in the humidity gradient in the calculation of the potential transpiration of the wet foliage ($E_{f,pot}$) and may even reverse the gradient following

$$E_{f,pot} = ho_b r_b^{-1} (q_{f, sat} - q_a)$$  \[18\]

where $r_b$ is the leaf boundary resistance [T L$^{-1}$].

Negative $E_{f,pot}$ values (i.e., inward fluxes by definition) or a dry canopy fraction $L_d = 0$ lead to zero transpiration from dry foliage, however, following Eq. [19] in the CLM:

$$E_{st} = \sigma_f L_{SAI} b (E_{f,pot} L_d r_b^{-1} + E_{f,pot})$$  \[19\]

where $\sigma_f$ is the vegetation fraction (dimensionless), $L_{SAI}$ is the stem plus leaf area index [L$^2$ L$^{-2}$]; $L_d$ is the dry canopy fraction (dimensionless), $r_s$ is the leaf stomatal resistance [T L$^{-1}$], and $\delta$ is a step function that is 1 for positive arguments and 0 for zero and negative arguments.

Equations [14–19] demonstrate a two-way feedback cascade in the CLM that reciprocally relates the energy and moisture state of the ground surface with the energy and moisture state within the canopy. Because some of the parameterizations are threshold formulations (e.g., Eq. [19]), even small changes in the energy regime may act as a switch, turning on and off certain processes in the simulation and resulting in large changes in some of the components of the energy balance. The question remains whether this type of parameterization accurately reflects the natural system. Answering this question clearly is beyond the scope of this study.

Influence of the Lower Temperature Boundary Condition

The lower temperature boundary condition closes the problem of subsurface heat transport. In the CLM, this boundary condition is defined as a no-flow condition (Eq. [7]), which means that energy in form of heat cannot leave the domain through the bottom. The question is whether this type of boundary condition is realistic at shallow depth below the land surface.

Near the land surface, the geothermal gradient is perturbed due to diurnal and seasonal temperature fluctuations. The depth of perturbation depends on the amplitude and associated frequency of the temperature fluctuation as well as the thermal conductivity of the subsurface material and convection. A no-flow boundary condition would mean that either $\partial T/\partial z = 0$ or $\lambda = 0$. The latter is not realistic,
because there generally does not exist a natural thermal insulator in the subsurface. In the case of the former, the temperature profile must be constant or exhibit a maximum–minimum at a certain depth below the land surface, which theoretically can only happen in the case of a temperature trend at the upper boundary, i.e., the land surface. Therefore, the assumption of a shallow no-flow temperature boundary condition on the order of 10⁰ m in the CLM is unrealistic in our opinion. Below, the influence of the type of boundary condition (no flow vs. constant temperature) on the land surface energy balance is discussed under spin-up conditions, that is, the mass and energy balances do not change during the simulation period.

Figure 10 shows the cumulative plots of $G$ for the reference simulation with $T_{\text{lb}} = 283 \pm 3$ K and a no-flow boundary condition at 6-m depth, as well as the data from the heat flux plate measurements. In all cases, $T_{\text{rain}} = T_{\text{wbt}}$. Comparison with the data shows that the best match was obtained using $T_{\text{lb}} = 286$ K, which is slightly higher than the value provided in the literature for that region (Fest et al., 2007). Note that in 2002, the measured average yearly temperature was approximately $T_{\text{av}} \approx 283$ K. Thus, the difference $T_{\text{lb}} - T_{\text{av}}$ provides a sense of the ambient temperature gradient between the lower boundary and the atmosphere during the entire spin-up period. This ambient gradient is then reflected in the cumulative ground heat fluxes, which are positive for $T_{\text{lb}} - T_{\text{av}} < 0$, negative for $T_{\text{lb}} - T_{\text{av}} > 0$, and about zero for $T_{\text{lb}} - T_{\text{av}} = 0$. This is intuitive in terms of energy conservation under spin-up or dynamic equilibrium conditions and also follows directly from Table 2. Little attention has been given, however, to the fact that the choice of the lower temperature boundary condition will predetermine $G_{\text{cum}}$ and, thus, also $E_{\text{avl,cum}}$ since $E_{\text{avl}} = R_{\text{net}} - G$. In our simulations, the lower boundary condition has little influence on $E_{\text{avl,cum}}$ because $G$ is only a relatively small fraction of the entire energy balance of the system under investigation. In the case of natural systems with little or no vegetation and a relatively large impact of $G$, for example arid regions, the choice of the lower boundary will have a much more significant impact on $E_{\text{avl}}$.

In case of a no-flow condition at the bottom, $G_{\text{cum}} = 0$ under spin-up condition is guaranteed. This again is reasonably intuitive, because under dynamic equilibrium conditions the energy balance must be equal to zero. Thus, the energy flux in and out of the subsurface occurring across the land surface must be zero for a one-dimensional soil column with a no-flow condition at the bottom. This is shown in Fig. 10.

The sensitivity analysis of the lower temperature boundary condition has implications for long term simulations considering climate change or, more generally, a temperature trend at the land surface (Maxwell and Kollet, 2008). In these simulations the subsurface may serve as a pseudo infinite energy source or sink depending on the direction of the gradient. This behavior has to be accounted for accurately in the simulations to prevent potential biases in the results.

Comparison with Measured Temperature Profiles

To demonstrate the usefulness of the simulations, a comparison with field data was performed without a true calibration based on some objective function. The comparison included varying $T_{\text{bc}}$ by 3 K around $T_{\text{bc}} = 283$ K, the value provided in the literature (Fest et al., 2007), and excluding and including convection with $T_{\text{rain}} = T_{\text{wbt}}$.

Figure 11 shows the simulation results that provided the best fit to the data for three monitoring depths below the land surface using $T_{\text{bc}} = 286$ K and $T_{\text{rain}} = T_{\text{wbt}}$. The agreement is good and the quality of the fit increases with increasing depth because subsurface heat transport is represented adequately and high-frequency variations due to processes at the land surface are filtered out; however, $T_{\text{bc}}$ is 3 K larger than the value estimated from long-term measurements and the literature. This might be explained by the applied assumptions, such as $\lambda$ being independent of the moisture content, $\theta$. Additionally, the location of $T_{\text{bc}}$ at 6-m depth might still be too shallow because at this site there are strong surface water–groundwater interactions along drain-
The influence of convection on the RMSEs is relatively minor but The variability at increasing depth is reproduced very well, however, RMSEs and a loss of variability. The latter stems mainly from an uncertainty into the results and was represented by applying the underestimation of the temperature amplitudes in the simulations. The variability at increasing depth is reproduced very well, however, which again suggests that the subsurface heat transport processes including convection are captured accurately. This is also supported by the fact that the RSMEs and VAR values get worse if convection is neglected in all cases. The impact on MEAN is negligible. The influence of convection on the RMSEs is relatively minor but recognizable; however, VAR values are significantly underestimated without incorporating the process of convection.

**Summary and Conclusions**

A newly developed model, ParFlowE coupled with the CLM, was used to study the influence of the sensible heat of rain, $H_l$ and subsurface convective heat transport, $q_{cv}$ on the energy balance at the land surface. The model computes subsurface moisture and energy transport and incorporates a land surface model for energy balance calculations. Because measured data of rainwater temperatures are missing, this parameter introduces uncertainty into the results and was represented by applying the air and wet-bulb temperatures as representative rainwater temperature values.

The influence of $H_l$ and $q_{cv}$ depends on the rainfall rate and the temperature difference between the rainwater and the soil surface and is, thus, clearly event based. Estimates of average energy fluxes are strongly influenced for a time period of up to one week, and in the case of ground heat flux, $G$, up to one month. The latter holds if $G$ is a significant component of the land surface energy balance. The analysis suggests that energy balance closure of observed fluxes cannot be expected if $H_l$ is not included in the analysis (apart from the other storage terms, as discussed in Jacobs et al. [2008]). We expect that the influence of $H_l$ will be even more pronounced in climate regimes with extreme convective rainfall events consisting of very large rainfall rates and large temperature differences between the air and the land surface.

The lower temperature boundary condition that closes the subsurface heat transport problem has a strong influence on $G$ under dynamic equilibrium conditions, i.e., spin-up. Constant-temperature boundary conditions predetermine whether the subsurface acts as an energy source or a sink. If the boundary temperature is smaller than the average air temperature of the applied atmospheric time series, the subsurface acts as a sink and vice versa. This might not be realistic and difficult to alleviate because of a lack of measured data in the deeper subsurface. The strength of the sink–source effect depends on the temperature difference between the average ambient air and the lower temperature boundary during the simulation period. On the other hand, a zero-flux boundary condition assures that the net energy balance of the subsurface is zero independent of the atmospheric time series applied in the simulations. Since this type of boundary condition can only exist if the subsurface temperature profile exhibits a maximum or a minimum at a certain depth, application of this boundary condition is not realistic because this type of condition may exist only for depths larger than $10^{-1} m$, which is generally not considered in land surface energy computations.

The usefulness of the approach was shown by comparing simulations with measured temperature profiles and heat flux plate measurements from the Meteostation Haarweg, Wageningen, the Netherlands. The agreement is good, although no comprehensive calibration has been performed. While incorporation of $H_l$ and $q_{cv}$ improved the curve fits, means, and variances of the simulated time series at different depths below the land surface, the lower boundary condition had a strong impact on the goodness of fit and effectively no influence on the variances. Thus, special care must be taken in defining the correct boundary conditions and associated temperatures.

Because the atmospheric forcing data are geographically specific, not all results can be transferred easily. As a matter of fact, for demonstrating the influence of rain sensible heat and subsurface heat convection, the chosen site is perhaps not optimal, because (i) the soil type of clay results in generally smaller Darcy velocities and thus smaller heat convection during rainfall events; (ii) ground heat flux values are generally small; and (iii) the temperature difference between rainwater and the land surface is also generally quite small. Much larger differences can be expected in regions with more unstructured soils, sparse vegetation, and strong convective storms, e.g., in (semi-)arid regions. Because an effect is clearly detectable on time scales ranging from hours to days under non-optimal conditions, however, it is reasonable to conclude that rain sensible heat and subsurface heat convection cannot be generally neglected in energy balance calculations.

**Table 3. Root mean square error values for the different simulations, as well as simulated and measured yearly average temperature (MEAN$_T$) and variance (VAR$_T$) values for depths of 0.05, 0.5, and 1.0 m, using varying boundary condition temperatures ($T_{bc}$) and neglecting convection (nc) or including convection with rain temperature ($T_{rain}$) taken as the wet-bulb temperature ($T_{wbt}$).**

<table>
<thead>
<tr>
<th>Simulation</th>
<th>RMSE</th>
<th>MEAN$_T$</th>
<th>VAR$_T$</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>0.05 m</td>
<td>0.5 m</td>
<td>1.0 m</td>
</tr>
<tr>
<td>Measured data</td>
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<td>284.8</td>
<td>284.6</td>
</tr>
<tr>
<td>$T_{bc} = 280K$, $T_{rain} = T_{wbt}$</td>
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<td>1.74</td>
<td>1.47</td>
</tr>
<tr>
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<td>2.06</td>
<td>1.88</td>
<td>1.68</td>
</tr>
<tr>
<td>$T_{bc} = 283K$, $T_{rain} = T_{wbt}$</td>
<td>1.95</td>
<td>1.49</td>
<td>1.01</td>
</tr>
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<tr>
<td>$T_{bc} = 286K$, $T_{rain} = T_{wbt}$</td>
<td>1.77</td>
<td>1.13</td>
<td>0.56</td>
</tr>
<tr>
<td>$T_{bc} = 286K$, nc</td>
<td>1.88</td>
<td>1.27</td>
<td>0.59</td>
</tr>
</tbody>
</table>
the future, similar simulations for varying climate and land surface conditions should be performed.

The results from the impact study of the lower temperature boundary condition are much more transferable. Choosing a shallow boundary condition in dynamic equilibrium and long-term simulations will always predetermine trends in the cumulative ground heat fluxes. This is a general conclusion that is valid in all simulations and should be taken under careful consideration in the model setup. Excluding boundary effects encompasses the use of very deep columns in simulations. This should not be particularly difficult to implement nor computationally expensive, because the vertical discretization may increase considerably with increasing depth.

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