12.1 Spatial heterogeneity of the soil moisture content and its impact on the surface flux densities and the atmospheric boundary layer.

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1. INTRODUCTION

Wetzel and Chang (1987) and Sivapalan and Woods (1995) studied the impact of the spatial heterogeneity of the soil moisture content on the horizontally averaged flux densities for grid cells containing subareas which local values of the relative saturation differ from the horizontally averaged relative saturation of the grid cell.

The main aim of this paper is to show that introduction of prescribed lateral variations in the soil moisture content might lead to better predictions of the surface flux densities and the near-surface temperature and specific humidity in large-scale atmospheric models. The focus of our study will be on the seasonal cycle. However, we will also investigate the impact of the differences in surface energy flux densities for the temperature and specific humidity in the atmospheric boundary layer and at 2 m, the height at which most SYNOPS observations are taken.

2. THEORY

To describe the spatial variation of the relative saturation within the grid cell of a large-scale atmospheric model, we adopt the Variable Infiltration Capacity (VIC) model described by Wood et al. (1992). Within this framework, the infiltration capacity, denoted by s, is defined as the maximal volume of water that can be stored in the soil column below a surface of unit area. The actual volumetric content, denoted by v is defined as the volume water per unit area that is stored in the soil column. At each point the ratio of the actual volume of water stored per unit area and the maximal volume that can be stored per unit area defines the relative saturation at each point within the grid cell: v/s.

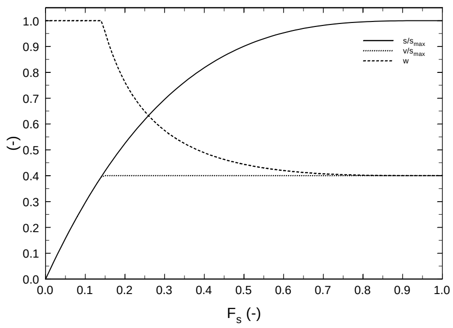

Figure 1: *The cumulative distribution of s/s_{max} for β=0.3 and the accompanying distribution of v/s_{max} for a value of 0.4 times s_{max} for v_{unsat}. Also shown is the distribution of the relative saturation.*

In the VIC model the infiltration capacity is considered as a random variable of which the cumulative distribution, denoted by F_s, can be described using (Sivapalan and Woods 1995):

$$F_s = 1 - \left(1 - \frac{s}{s_{\max}}\right)^{\beta}$$

where s_{max} is the extremum of s within the grid cell, and β is a parameter describing the spatial heterogeneity in s.

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In regions where the long-term precipitation is uniform, the soil column is recharged at the same rate throughout the grid cell. As a result, the volumetric soil moisture content is approximately uniform in the unsaturated part of the grid cell. For subareas within this part we adopt therefore a constant value for v: v_{unsat}. In the saturated areas v is bound by its maximal value and here v equals s (see Figure 1).

a. The distributed approach

Because the relative saturation varies over the grid cell (Figure 1), the soil moisture stress experienced by plants varies over the grid cell. In areas where s is small, the soil is wet and the soil moisture stress is small. In contrast, in areas where s is large, the soil is usually much drier and the plants experience a much larger stress. Because the canopy conductance and the surface flux densities depend non-linearly on the local relative saturation (Ronda et al. 2001), it is not possible to calculate the grid cell averaged surface flux densities analytically from the distribution of the relative saturation.

In this study we use a numerical approach and impose a subgrid over our grid cell. For that purpose, we divide the values of s which range from 0 to s_{max} into a number of intervals and assign the average value of s, denoted as s_i of each interval to a subarea. Then, for each subarea we calculate the relative saturation, denoted by w_i. For each subarea the calculated relative saturation is used to compute the soil moisture. Then, the surface energy budget is solved for each subarea separately to obtain the surface temperature and the resulting latent heat flux density and sensible heat flux density. Afterwards, the subarea flux densities, denoted by f_i are aggregated to obtain the averaged grid cell flux density.

b. The bulk approach

In the bulk approach the grid cell is taken as one homogeneous patch with the relative saturation given by $W/W_{max,}$, the ratio of the total volume of water stored in the soil column of the grid cell and the maximal volume of water that can be stored in the soil column of the grid cell. Consequently, the surface flux densities of the grid cell can directly be calculated when the meteorological conditions at the reference are known.

3. RESULTS

a. Offline results

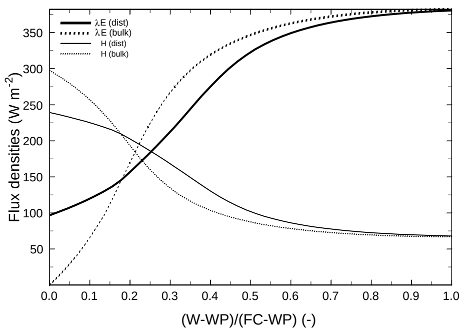

Figure 2: The latent heat flux densities (thick lines) and the sensible heat flux densities (thin lines) as a function of the grid cell canopy water status status for β=0.3.

Figure 2 shows how the latent and sensible heat flux densities depend on the grid cell vegetation moisture status defined by (W-WP)/(FC-WP) where WP is the permanent wilting point, the value of the soil moisture content below which plant wilt and FC is the field capactity. It appears that the distributed approach leads to a weaker variation of the heat flux densties as a function of the grid cell vegetation moisture status. As in Wetzel and Chang (1987) in dry conditions the distributed approach predicts a larger latent heat flux density than the bulk approach. In contrast, in wet conditions, the distributed approach gives a lower prediction of the latent heat flux density than the bulk approach.

b. Impact on seasonal cycle

Next, we study whether the differences in estimated flux densities lead to a weaker seasonal cycle of the latent heat flux density and a resulting enhanced latent heat flux density during the dry season. Therefore, both the distributed and the bulk model are forced with observed data, instead of using a constant climatological forcing. In this section, we use forcings that are obtained in a dry climate during the SEBEX experiment (Gash et al. 1991).

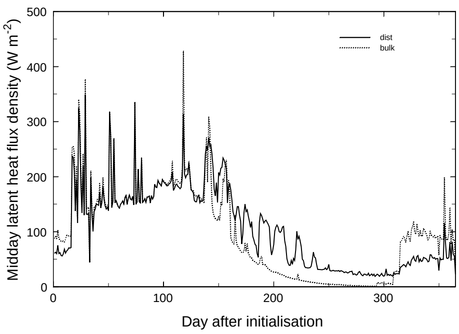

Figure 3: Time series of the midday latent heat flux density for the dry climate run using a distributed (solid) and a bulk approach (dotted).

Figure 3 shows a time series of the midday latent heat flux density, the latent heat flux density averaged from 12:00 UTC till 14:00 UTC. At the beginning of the wet period, from day 1 to 20 and from 320 onwards, the bulk approach gives larger estimates of the latent heat flux density than the distributed approach. At the beginning of the wet period, the midday latent heat flux density predicted by the bulk approach is higher than the latent heat flux. However, as the soil dries down, from day 140 till day 320, the latent heat flux estimated with the bulk approach drops sharply. In contrast, the latent heat flux density predicted by the distributed approach reduces more gradually. At the beginning of the dry period, the midday latent heat flux density predicted by the distributed approach is about twice as large as the latent heat flux density predicted with the bulk model. At the end of the dry season, the bulk approach predicts, in contrast to the distributed approach, even a vanishing of the latent heat flux density, as shown in Figure 3 .

c. Inclusion of boundary layer feedback

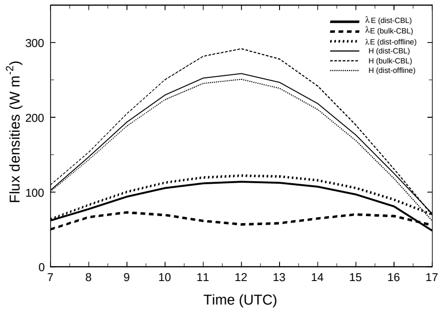

Figure 4: Diurnal variation of the sensible heat flux density (thin) and the latent heat flux density (thick) as calculated using the distributed approach (solid), the bulk approach (dashed) and the distributed approach forced in the surface layer (dotted): results are given for a grid cell relative saturation of 0.37.

To estimate the impact of the boundary layer feedback, we couple the model for the averaged surface flux densities to a convective boundary layer model (Tennekes 1973).

Figures 4 gives the diurnal variation of the surface flux densities, calculated using both coupled runs and the run without boundary layer feedback. It appears that the run without boundary layer feedback gives a slightly higher estimate of the latent heat flux density density than the run where the boundary layer is allowed to feed back on the surface. However, even when the surface is coupled to a convective boundary layer, a bulk approah predicts a considerably lower latent heat flux density when the soil is near the wilting point.

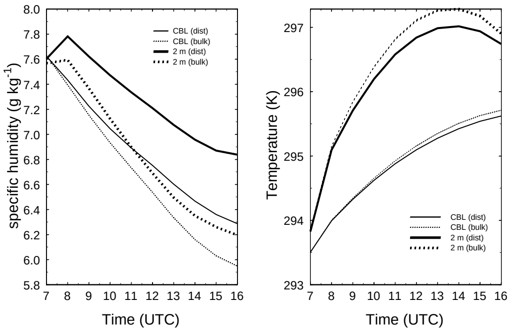

Figure 5: Diurnal variation of the specific humidity (left panel) and the temperature (right panel) at 2 m (thick lines) and in the convective boundary layer (thin lines), simulated using a distributed approach (solid) and a bulk approach (dotted): grid cell relative saturation is 0.37.

Figure 5 shows the specific humidity of the boundary layer and the averaged 2 m specific humidity (left panel). When the distributed approach is used, the boundary layer appears to be about 0.3 g/kg more moist as compared to the bulk approach. Furthermore, the averaged specific humidity near the surface is about 0.7 g/kg larger for the coupled run with the distributed approach than the run with the bulk approach. As result, in areas where our typical distribution describes the spatial variation of the soil moisture content well, using a bulk approach would give a dry bias.

4. CONCLUSION

The distributed approach predicts a weaker seasonal cycle of the latent heat flux density. Even when the boundary layer is allowed to feed back on the surface, a bulk approach gives significantly different estimates of the surface flux densities compared to a distributed approach. In dry conditions the bulk approach typically gives a warm and dry bias as compared to the distributed approach.

We conclude that taking account of the horizontal heterogeneity of the soil moisture content is a prerequisite for a proper representation of the seasonal hydrological cycle within large-scale atmospheric models.

5. REFERENCES

Gash, J.H.C., J.S. Wallace, C.R. Lloyd, A.J. Dolman, M.V.K. Sivakumar, and C. Renard, 1991: Measurements of evaporation from fallow Sahelian savannah at the start of the dry season. *Quart. J. Roy. Meteor. Soc*., **117**, 749-760

Ronda, R.J., De Bruin, H.A.R. and Holtslag, A.A.M., 2001: Representation of the canopy conductance in modelling the surface energy budget for low vegetation. *J. Appl. Meteor*., **40**, 1431-1444

Sivapalan, M., and R.A. Woods, 1995: Evaluation of the effects of general circulation models' subgrid variability and patchniness of rainfall and soil moisture on land surface water balance fluxes. *Hydrol. Proc*., **9**, 697-717

Tennekes, H., 1973: A model for the dynamics of the inversion above a convective boundary layer. *J. Atmos. Sci.,* **30**, 558-567

Wetzel, P.J., and J.-T. Chang, 1987: Concerning the relationship between evapotranspiration and soil moisture. *J. Climate Appl. Meteor.*, **26**, 18-27

Wood, E.F., D.P. Lettenmaier, and V.G. Zartarian, 1992: A land-surface hydrology parameterization with subgrid variability for general circulation models. *J. Geophys. Res.,* **97**, 2717-2728