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SENSITIVITY OF MESOSCALE RAINFALL TO FOREST COVER AND AGGREGATION OF SURFACE FLUXES FROM PARTIALLY WET AREAS

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SUMMARY

Results of the final phase of the Alliance project "Mesoscale modelling with spatially varying surface conditions" are described. A meso-ß scale model was used to model a frontal intrusion in south west France during HAPEX-MOBILHY. The skill of the model to reproduce the observed variation in temperature, humidity and windspeed over the domain is acceptable, although there were errors in the timing and positioning of the front. A stable boundary layer was both observed and modelled over the forested area. The associated negative sensible heat flux provided the energy to sustain evaporation from the wet forest canopy under conditions of low radiation. A large windshear over the stably stratified boundary layer provided the required turbulent kinetic energy to maintain the downward transport of sensible heat. Sensitivity experiments showed that forest can increase the local rainfall by up to 30% when compared with a bare soil domain. Half of this increase is from positive feedback of the intercepted water which re-evaporates. The high roughness length of the forest accounts for the rest of the increase in rainfall and for the accompanying increase in soil moisture.

The area mean fluxes of evaporation from the model were compared to the fluxes from a one dimensional model to assess aggregation methods. Predictions of total evaporation flux were accurate using simple averages of the surface parameters. The partition of evaporation into transpiration, evaporation of intercepted water and evaporation from bare soil requires more complex aggregation methods.

FINANCE

Visits:	2 months in 1991	= £2200
	1 week in 1991	= £300
	1 month in 1992	= £1100
	1 week in 1992	= £300
	4 return flights	= £640
	Total expenditure	= £4540
Awards:	1991	= £1600
	1992	= £1800
	Total award	= £3400
	Expenditure-Awards	= £1140

INTRODUCTION

Experiments with atmospheric models have shown that forest can influence the climate. This influence may occur on a range of scales, from small local scales (de Bruin and Jacobs, 1989), through mesoscales (André *et al.*, 1989) to global scales (Rowntree, 1988). On all these scales, physiognomic characteristics of the forests such as surface conductance, aerodynamic roughness and albedo play an important role in determining the forest-atmosphere interaction. On a global scale, forests may also affect the hydrological balance of regions by increasing the rainfall (Sud et al., 1988).

Stewart (1977) was the first to demonstrate experimentally Rutter's (1967) suggestion that forest, when its surface is wet, can draw energy from the overlying air to sustain their evaporation of intercepted rainfall. Such wet canopy evaporation is sustained by regional scale advection of energy since the available energy in the form of net radiation is generally low under rainy conditions. The present study examines the possibility of modelling the behaviour of wet forest sufficiently accurately on the meso-ß scale to reproduce the experimental evidence of regional scale advection.

The increase of rainfall in the presence of forests at the meso-scale has not been studied extensively. André *et al.* (1989) modelled an increase in rainfall downwind of the Landes forest in south-west France and Meher-Homji (1980) showed some evidence of a decrease in rainfall after deforestation in India. Two characteristics of the forest can cause an increase in rainfall; The high evaporation rates of intercepted water during rainfall can give a positive feedback, increasing the subsequent rainfall, and the high roughness lengths can result in enhanced water vapour convergence in the boundary layer which also increases the rainfall. The present paper also attempts to show how important a role these two effects have on the rainfall at the meso-β scale.

Numerical weather prediction models calculate the surface fluxes of heat and momentum on a grid-average basis. For heterogeneous land-surfaces, it is necessary to account for sub-grid variations of surface fluxes within the present land-surface schemes, which have been developed for homogeneous surfaces. Shuttleworth (1988) suggests that the aggregation methods for model parameters depend on whether the spatial scale of the variation is smaller or greater than 10km. When the spatial scale is below 10km, the atmospheric turbulence in the planetary boundary layer tends to blend out the effects of the variation and the atmosphere effectively presents to the surface uniform variables of wind speed, humidity and temperature which can then be used to calculate regional surface fluxes (Mason(1988), Claussen(1991), Blyth *et al.* (1993)). Where the variation is at scales above 10km this blending does not occur. This paper concerns the latter case and quantifies errors incurred when mean meteorological variables are used to calculate mean surface fluxes in the presence of large scale heterogeneity. This is achieved by comparing the results of a detailed, three-dimensional (3-D) numerical model, where the heterogeneity is explicitly modelled, with the results of a simple, one-dimensional (1-D) model, where the heterogeneity is allowed for within the surface scheme.

The initial phase of the study is reported elsewhere (Blyth *et al.*, 1991). A meso-scale model was used to simulate a frontal intrusion that had been monitored during HAPEX-Mobilhy (André *et al.*, 1986). A comparison was made of the model output and the observations made on June 5th, 1986 which showed that, despite an error in the timing and position of the front, the model was able to reproduce the salient features of this mesoscale event.

The results of the model were subsequently used to investigate the sensitivity of the front to the land-surface cover (Blyth *et al.*, 1993b) and to look at the aggregation problem outlined above (Noilhan *et al.*, 1993). The results of these investigations are reported here.

MODEL LAND SURFACE SCHEME

The land surface scheme is described in full by Noilhan and Planton (1989). Only those elements which relate to the interception of rainfall will be given here. The interception capacity of the vegetative canopy, W_{rmax}, is given by

$$W_{rmax} = 0.2 * LAI * veg \tag{1}$$

where LAI is the leaf area index and *veg* is the fraction of the grid covered by vegetation. A canopy interception reservoir, W_r is filled by precipitation, P and depleted by evaporation, E_r:

$$\delta W_r/\delta t = P * veg - E_r \tag{2}$$

where δt is the time step. If the input to the reservoir is greater than the capacity, the extra rainfall is added to the throughfall. Following Deardorff (1978) the interception reservoir covers only a fraction $veg*\delta$ of the gridsquare:

$$\delta = (W_r/W_{rmax})^{2/3} \tag{3}$$

Evaporation of intercepted rainfall from the reservoir, E_r is parameterized as:

$$E_r = veg\ \delta\ (q_{sat}(T_s)-q_a)/r_a \tag{4}$$

where $q_{sat}(T_s)$ is the saturated humidity at the surface temperature, T_s, q_a is the specific humidity at the lowest model layer and r_a the aerodynamic resistance to water vapour transport. Further details of the land surface scheme can be found in Noilhan and Planton (1989). Mahfouf and Jacquemin (1989) describe a detailed validation of the interception parameterization over agricultural crops in the HAPEX square.

MODELLED HEAT TRANSFER OVER WETTED FOREST

The observed and modelled stable boundary layer structure (Figures 1a and 2a) over the forest is a result of the combination of synoptic weather conditions and the interaction of the forest with the lowest layers of the atmosphere. The wet forest evaporates at a high rate, cooling the surface relative to the warmer overlying air. The radiation received at the surface and heat storage in the forest are not sufficient to maintain these high evaporation rates and energy is drawn from the warmer air passing over the forest. This is shown in Figure 3 where the sensible and latent heat flux profiles are shown for a fully forested grid box at position (25,26), adjacent to the central site. A positive latent heat flux at the surface of about 80 Wm^{-2} is coupled with a negative sensible heat flux bringing heat downwards to the evaporating surface. These evaporation rates are of similar magnitude as those observed by Stewart (1977) for a wet canopy at Thetford Forest, and Gash *et al.*, (1992) who modelled interception for the Landes forest.

The supply of warm air to the evaporating surface can only be maintained if there is downward entrainment of warm air into the stable layer by turbulence. The turbulent kinetic energy (TKE) budget is dominated by two terms, the buoyancy term and the dynamical, wind shear production term. The magnitude of these two terms can be calculated for the gridbox from (Bougeault and Lacarrère, 1989):

$$P_d = c_k k \sqrt{e} \left(\frac{\delta u}{\delta z} \right)^2 \tag{5}$$

$$P_b = \beta \left(\overline{w'\theta'} + T\, \overline{w'q'} \right) \tag{6}$$

in which P_d and P_b are the dynamic, wind shear production, and the buoyancy production term respectively, c_k is a constant, e is the mean eddy kinetic energy, $\delta u/\delta z$ the windspeed gradient and ß is the buoyancy parameter while $\overline{w'\Theta'}$ and $\overline{w'q'}$ are the eddy fluxes of heat and water vapour, T is the temperature and k is von Karman's constant.

Using modelled values for the variables used in (5) and (6) the buoyancy term, P_b, is -0.5 10^{-3} m^2s^2 and the wind shear term, P_d, is 1.2 10^{-3} m^2s^2. The stable layer is therefore a sink for turbulence. The large gradient in wind speed over the forest (Figure 2c) provides the mechanism for the entrainment of warm air. The resulting TKE profile is positive in the lower boundary layer (Fig 3c) which sustains the downward transport of sensible heat to the wet forest canopy. It is interesting to note that this downward transport of sensible heat could only have been modelled with a TKE parameterization of the diffusion coefficients, since a parameterization based on stability would probably reduce the vertical diffusion too much, and result in a much more stable boundary layer.

The present model results provide the first theoretical evidence for the maintenance of wet canopy evaporation through a negative sensible heat flux and show, perhaps more importantly, how local scale processes are linked to mesoscale meteorological phenomena.

SENSITIVITY OF MESOSCALE RAINFALL TO LAND SURFACE COVER

André *et al.* (1989) have shown that forest may increase the rainfall downwind of the forests. They suggested that the primary mechanism for this influence was an increase in humidity of the air downwind the forest, as a result of rapid evaporation of intercepted rainfall. However, an increase in roughness length may also increase the precipitation by a change in the boundary layer water vapour convergence. Note, however, that the increased evaporation is also a roughness effect as it is predominantly the high aerodynamic conductance of forest which gives rise to high evaporation rates from wet canopies.

To quantify the relative importance of these two effects, three sensitivity tests were carried out. In these tests the land surface in the model domain was replaced by a homogeneous surface cover. Forest and bare ground were used to represent surfaces with maximum and minimum potential for feedback, respectively. Also a hypothetical surface was used which had the same roughness as the forest but no interception capacity, causing the evaporation rates to be lower than from the bare soil run. This was intended to isolate the two feedback mechanisms of high roughness and high evaporation rates. The same large scale forcing was used in each case. The values of the surface parameters, W_r and z_0 for the three sensitivity runs and mean values for the control run are given in Table 1.

The spatial distribution of the rainfall for three of the runs (control, forest and bare ground) is shown in Figure 4. There is an area of rainfall on the coast line in both the control run and the forest run. This is likely to be solely a result of increased roughness length of the forested area in that part of the domain as the area over which this precipitation develops is not large enough to create a substantial humidity feedback. This rainfall is not associated with the front which is positioned further to the east. In the eastern frontal region the forested run simulates more rainfall than either the bare soil or control run. Unlike the rainfall on the coast line, this is likely to be caused by a combination of both the humidity feedback and the roughness length.

To quantify these two effects a comparison is made of the water balance between the three sensitivity runs. To assess sensitivity which avoids the sea and the mountain ranges of the Pyrenées and the Massif Central a subdomain of the model was chosen (Fig 5). These exclusions make the results more sensitive to the land surface cover. The daily rainfall, the change in soil moisture and the daily evaporation of the three sensitivity runs is shown in Table 1.

The forest run produces more rainfall (3.8mm) than the forest without interception (3.4mm), although the increase in soil moisture is the same (1.94mm). This implies that the extra rainfall caused by the feedback of the evaporation of intercepted water is merely recycled within the boundary layer and does not enter the soil. The forest without interception produces more rainfall than the bare soil (2.9mm), although the evaporation is equal or less throughout the day. This implies that the high roughness length of the forest plays a significant role in increasing the rainfall. The increase in soil moisture of the bare soil run is 0.7mm less than that of the forest without interception. 0.5mm is due to the difference in rainfall and 0.2mm is due to the high evaporation rates of the bare soil after the rainfall event.

From these runs it can be tentatively concluded that, of the 30% increase in rainfall between the forest run and the bare soil run, about half is recycled through interception and evaporation back to the atmosphere and does not reach the soil. The other half, caused by the increase in roughness length represents a net gain of water to the land from the atmosphere.

AGGREGATION METHODS

To test out aggregation methods to represent this spatial mix of weather types, a one-dimensional (1-D) model is used. The surface fluxes from the 1-D model are compared to the spatially averaged surface fluxes from the 3-D model.

The 1-D model represents the way the area would be modelled in a large scale model, with the advected variables of wind velocity, temperature and humidity and the forcing variables being specified from the 3-D model. The 1-D model has the same physics and boundary conditions as the 3-D model but represents an area of 250km by 240km (see Figure 6). About one third of the area is covered by forest on sandy soil and the rest is crops and sparse grass land on loamy soil. With a relatively homogeneous overlying meteorology, Noilhan and Lacarrère (1992) obtained good results for areal evaporation when they used average logarithms of the roughness length, reciprocals of the minimum surface resistance, albedo, leaf area index, vegetation fraction (veg) and the clay and sand contents of the soils.

The meteorological variables of rainfall, radiation temperature, humidity and wind speed vary considerably across the domain on June 5th, however (see previous section). Equation 3, to quantify δ, was developed for a meso-scale model and assumes that uniform rain covers the grid square. This assumption does not hold for large scale models however and it is necessary to account for sub-grid variability of rainfall distribution (Dolman and Gregory, 1992). In this study, the assumption is made that uniform rain covers a fraction, S, of the grid square of the 1-D model. S is calculated at each time step from the 3-D model and defined as the fraction of the domain where rainfall is greater than 0.1mm. The calculation of δ is modified accordingly.

$$\delta = S\left(\frac{W_r}{W_{rmax}}\right)^{\frac{2}{3}} \tag{7}$$

The spatial variation of the humidity gradient across the 1-D model domain affects the calculations of the evaporation from wet and dry vegetation due to the coincidence of the low humidity gradient with evaporation of intercepted water and the high humidity gradient with transpiration. The resulting overestimated evaporation of intercepted water has already been compensated for by (7). It is possible to compensate for the resulting underestimation of transpiration by reducing the minimum surface resistance from $58sm^{-1}$ to $25sm^{-1}$.

RESULTS

Figure 7 shows the diurnal cycle of the area-average evaporation fluxes from the 3-D model and from the 1-D model using the aggregation methods described in the previous section, summarised in Table 2. The discrepancy between the evaporation calculated in run 'a' with the other runs demonstrates that the evaporation flux is largely dependent on the specification of the correct area average incoming radiation and rainfall (Shuttleworth, 1988), and is relatively insensitive to their spatial variation.

The partition of evaporation into transpiration, evaporation of intercepted water and evaporation from bare soil is more sensitive to heterogeneity. Figure 8 shows the diurnal cycle of predicted transpiration for the 3-D model and the 1-D model simulations. In run 'b', low intensity rain falls over the whole domain resulting in high evaporation of intercepted water, low evaporation from bare soil and negligible transpiration (Fig. 8). The reduction of the fractional wetted area achieved using Equation 7 (run "c") had the effect of bringing the rate of evaporation of intercepted water down to levels given by the 3D model. The transpiration rate is increased but is still only half that given by the 3D model (Fig. 8), the bare soil evaporation is increased to the rate given by the 3-D model due to the increase in the humidity gradient.

If the effect of the variation in humidity gradient on transpiration is compensated for (run "d"), the rate of transpiration is increased to that given by the 3-D model (Fig. 8). The bare soil evaporation is slightly reduced and the evaporation of intercepted water is slightly reduced.

DISCUSSION

The present study provides evidence of the mechanism for maintaining negative sensible heat flux over forest, which is capable of sustaining wet canopy evaporation when the radiation is low. The main condition needed for this mechanism to occur is a strong wind shear within a stably stratified boundary layer. The positive, dynamical contribution in the turbulent kinetic energy budget then dominates the negative, buoyancy contribution, and a positive turbulent kinetic energy is obtained, providing the mechanism for downward transport of heat and upward evaporation. PERIDOT was able to model this effect because the vertical diffusion parameterization is dependent on the turbulent kinetic energy. This mechanism will occur when the size of the forest exceeds a critical limit, otherwise a different boundary layer structure, relevant to the neighbouring vegetation would be observed. The present model run gives an estimate of at least 10 km for this size, this being the size of a model gridbox. How much smaller the required size can be, is not clear and will require other model runs with a smaller grid size.

The results of the sensitivity experiments show that the land surface has a considerable effect on the rainfall in a frontal situation. The total rainfall, when the surface was covered by forest, was 30% higher than when the surface was all bare soil. It was demonstrated that the higher evaporation rates from a wet forest during a rainfall event accounts for about half of that increase but does not contribute to an increase in soil moisture. The convergence of water vapour in the boundary layer due to the high roughness of the forest accounts for the other 50% of the increase in rainfall which increases the soil moisture. Thus, the increased rainfall recycling from a positive humidity feedback will influence the regional rainfall dynamics and patterns only and can be important for regional weather prediction models. However, the roughness effect can influence the long term water balance of a region by providing a net gain of water by the land from the atmosphere.

Prediction of evaporation by the one dimensional model is accurate if the correct area-average radiation and rainfall are specified, irrespective of the variation of the meteorological and surface conditions.

However, the correct partition of the evaporation into transpiration, evaporation of intercepted water and evaporation of the bare soil requires knowledge of the fractional area covered by rainfall and the changes to transpiration due to variation of the vertical gradient of the humidity in the wet and dry areas. The water balance, and therefore the long term energy balance, depends on the source of the evaporated water. For instance, transpiration takes water from the deep soil, while the bare soil evaporation takes water from the top soil. It is therefore important for long term climate predictions to model the correct balance between these evaporation types. This was achieved using simple methods to relate the rainfall predicted by the 3-D model to the evaporation of intercepted water, and by calibrating the transpiration by changing the minimum surface resistance. To be able to use this method, the spatial distribution of the rainfall and the effective minimum surface resistance within the 1-D model domain are required.

ACKNOWLEDGEMENTS
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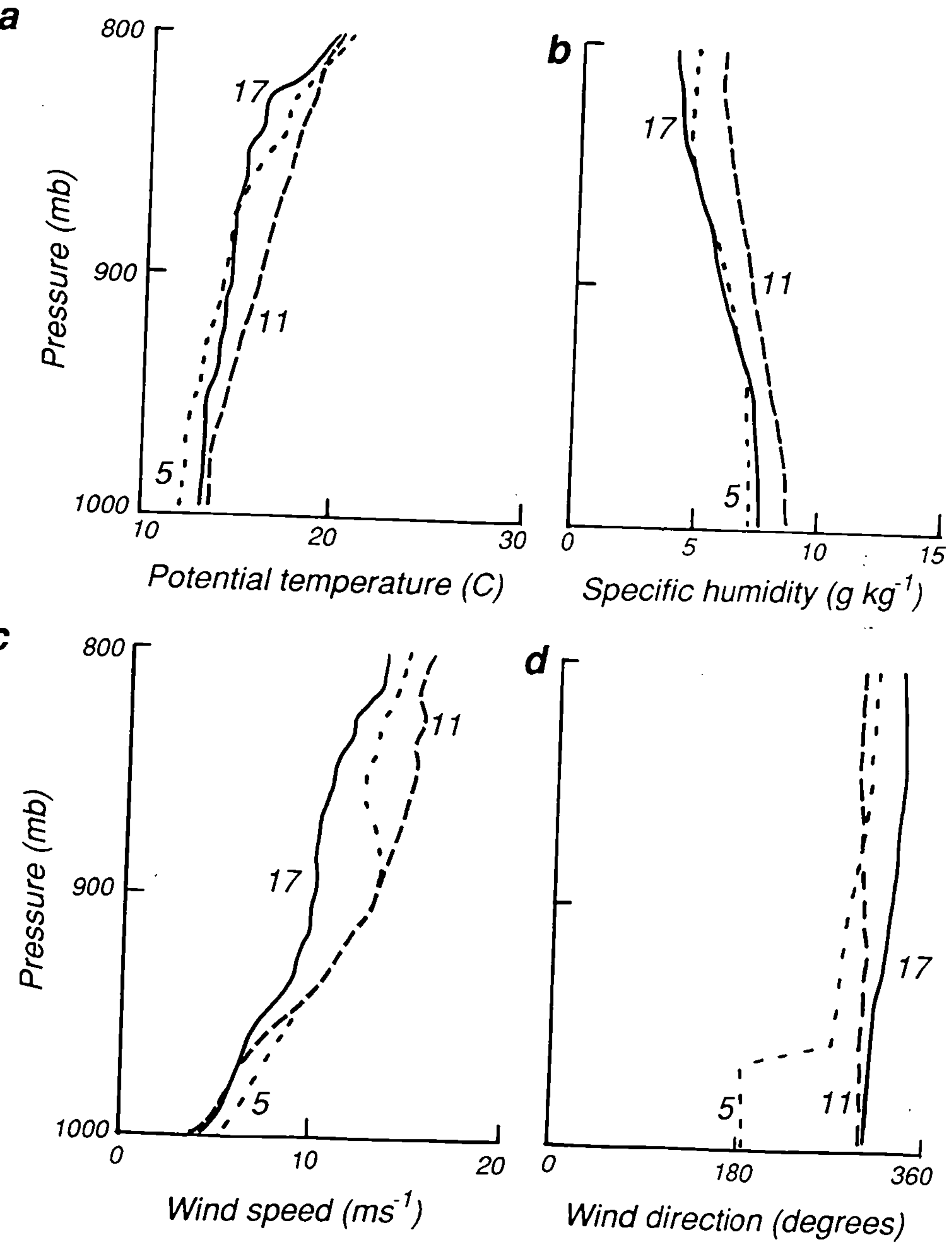
TABLE 1. **Water balance components of sensitivity experiments.**

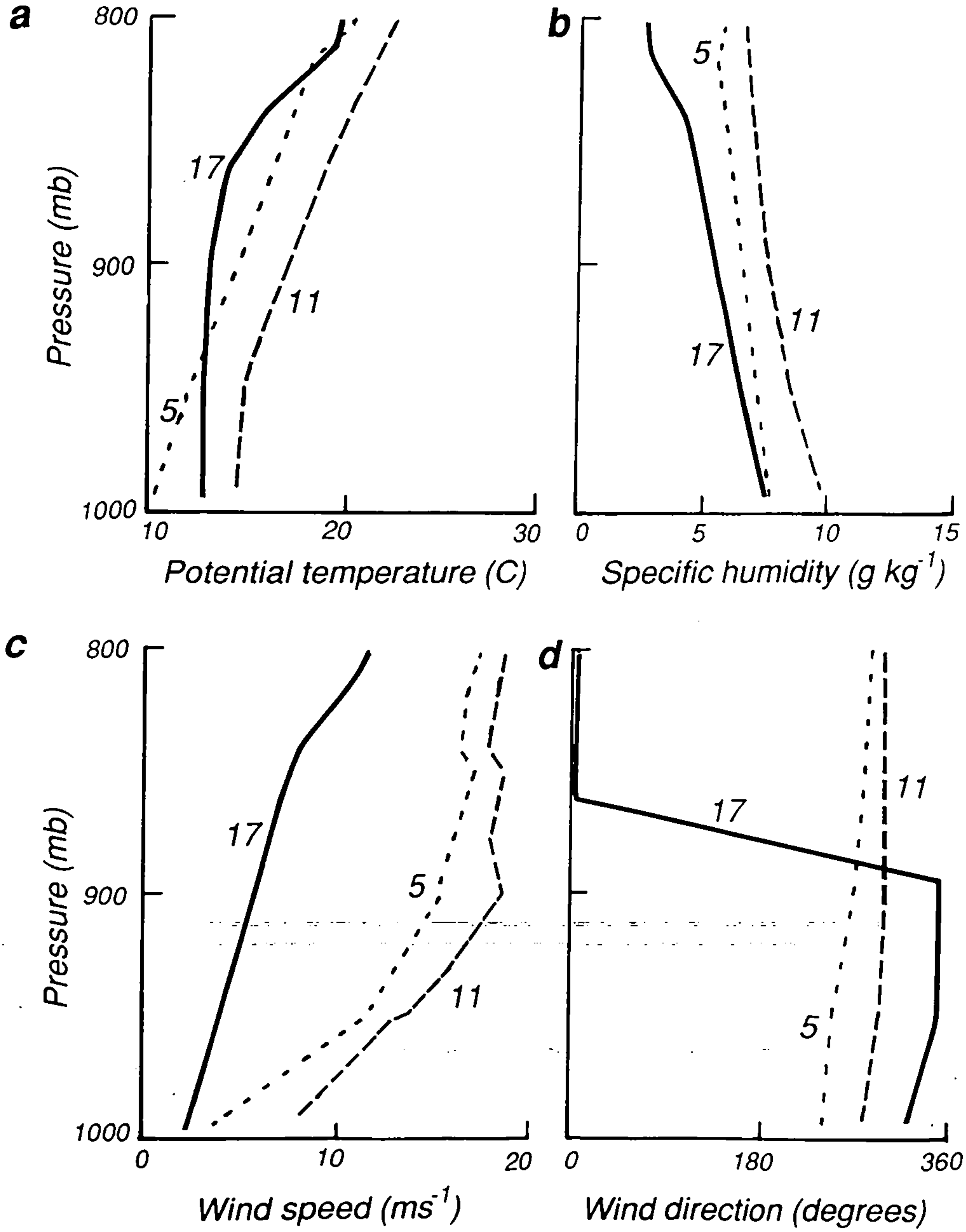
experiment	z_0(m)	W_{rmax}	P (mm)	Δm	E_{tot}
control	.276	.17	3.3	1.42	1.89
forest	1.0	.46	3.8	1.93	1.83
bare soil	0.01	0.0	2.9	1.22	1.68
forest (W_r=0)	1.0	0.0	3.4	1.94	1.46

TABLE 2. **Description of one-dimensional runs.**

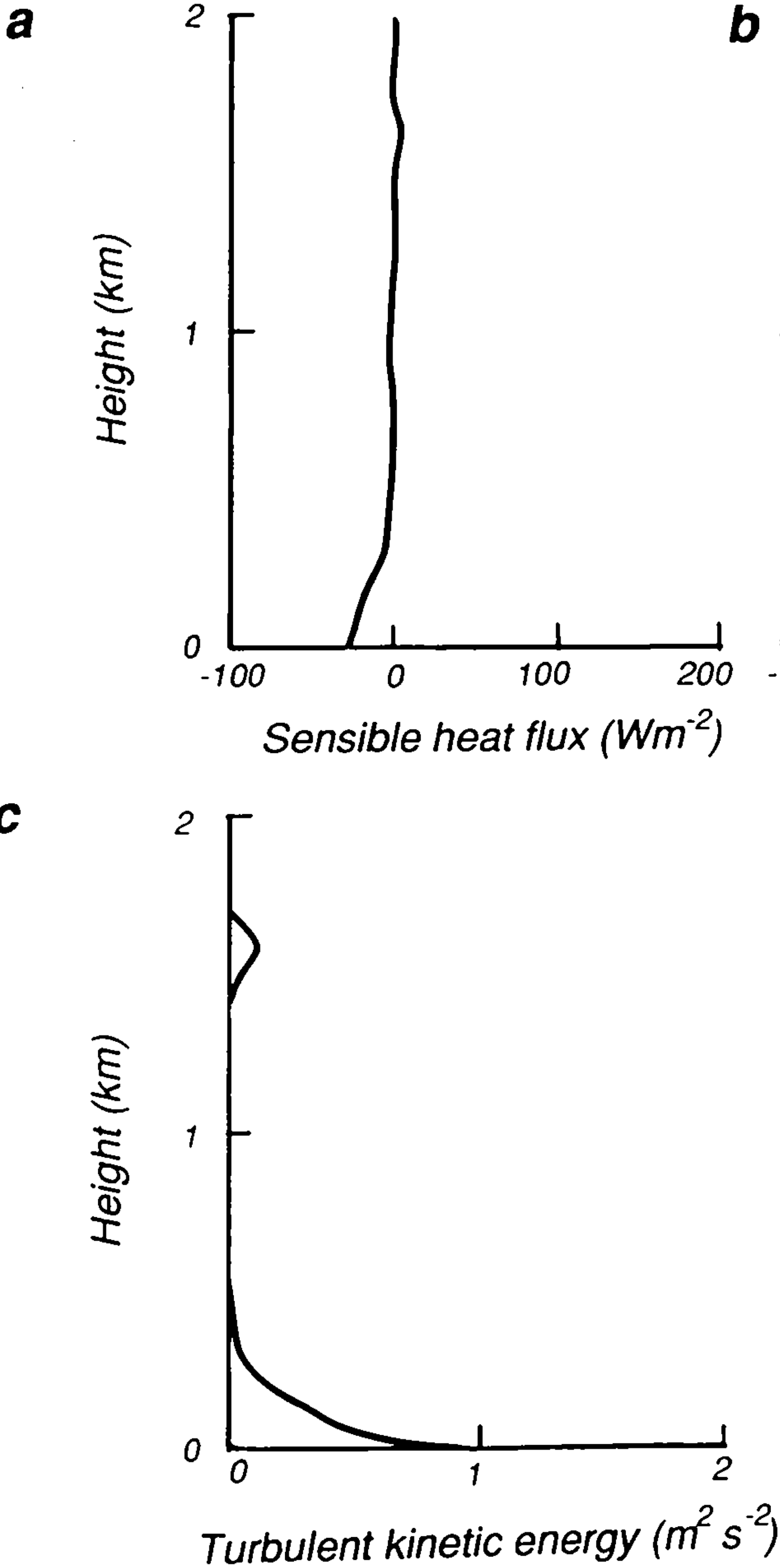
a	Rainfall and radiation calculated in code
b	Area-average rainfall and radiation imposed from 3-D model.
c	Same as b but using Equation 7 instead of Equation 3
d	Same as c but using a minimum surface resistance of 25 sm^{-1} instead of 58 sm^{-1}



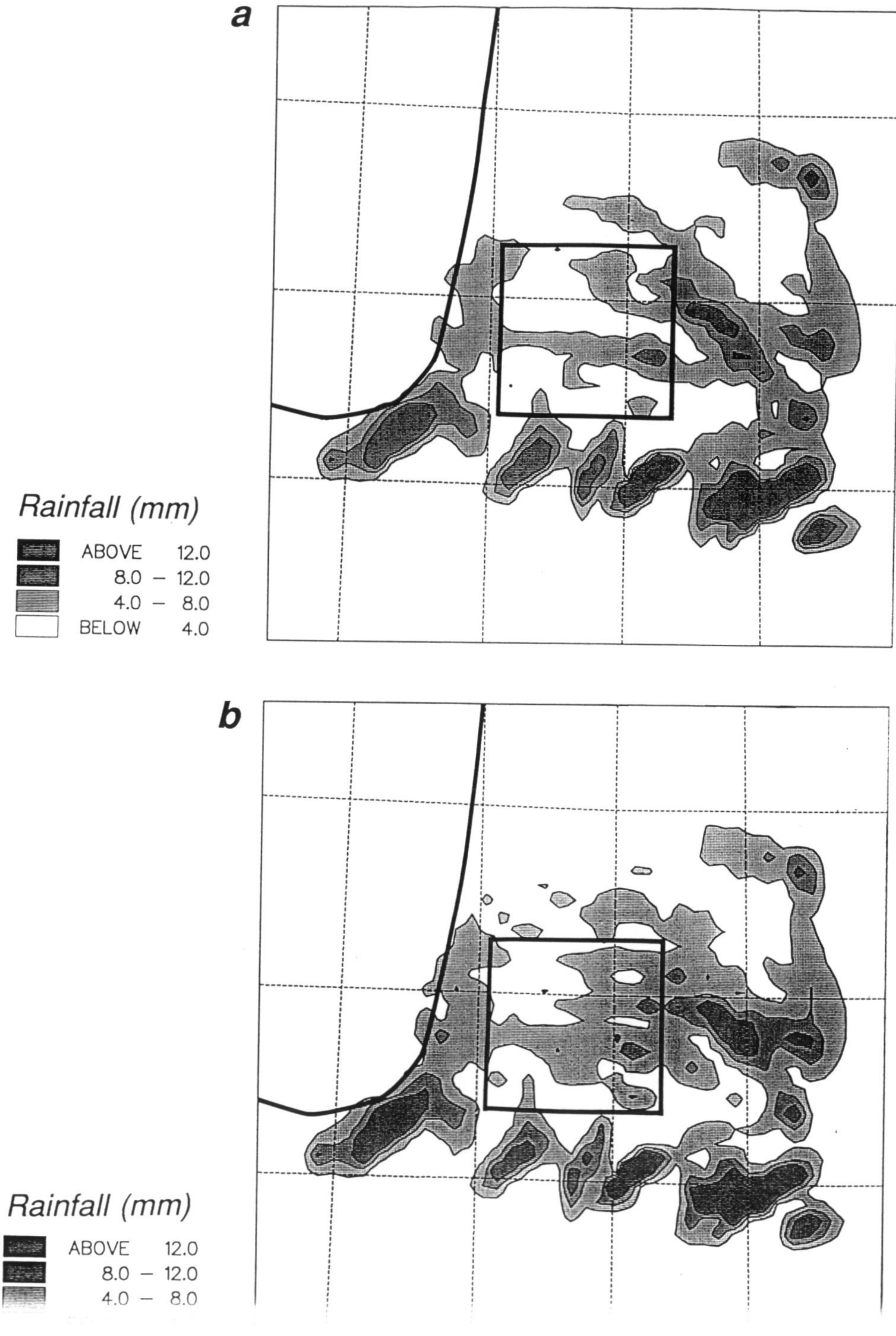
1. Observed profiles at central site at 05:00, 11:00 and 17:00 UT of
(a) temperature, (b) humidity, (c) wind speed and (d) wind direction

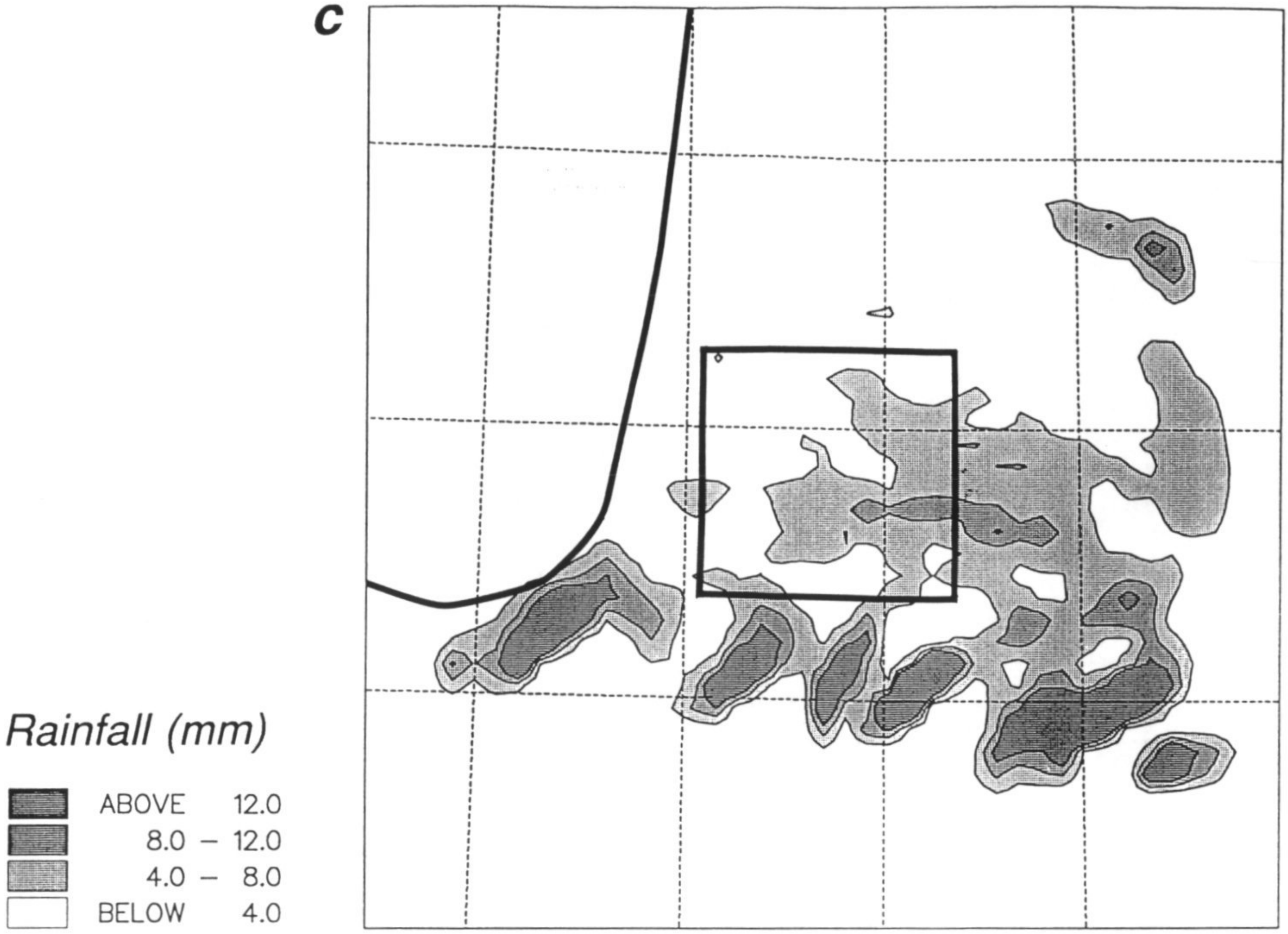


2. Modelled profiles at position (24,25) at 05:00, 11:00 and 17:00 UT of (a) temperature, (b) humidity, (c) wind speed and (d) wind direction

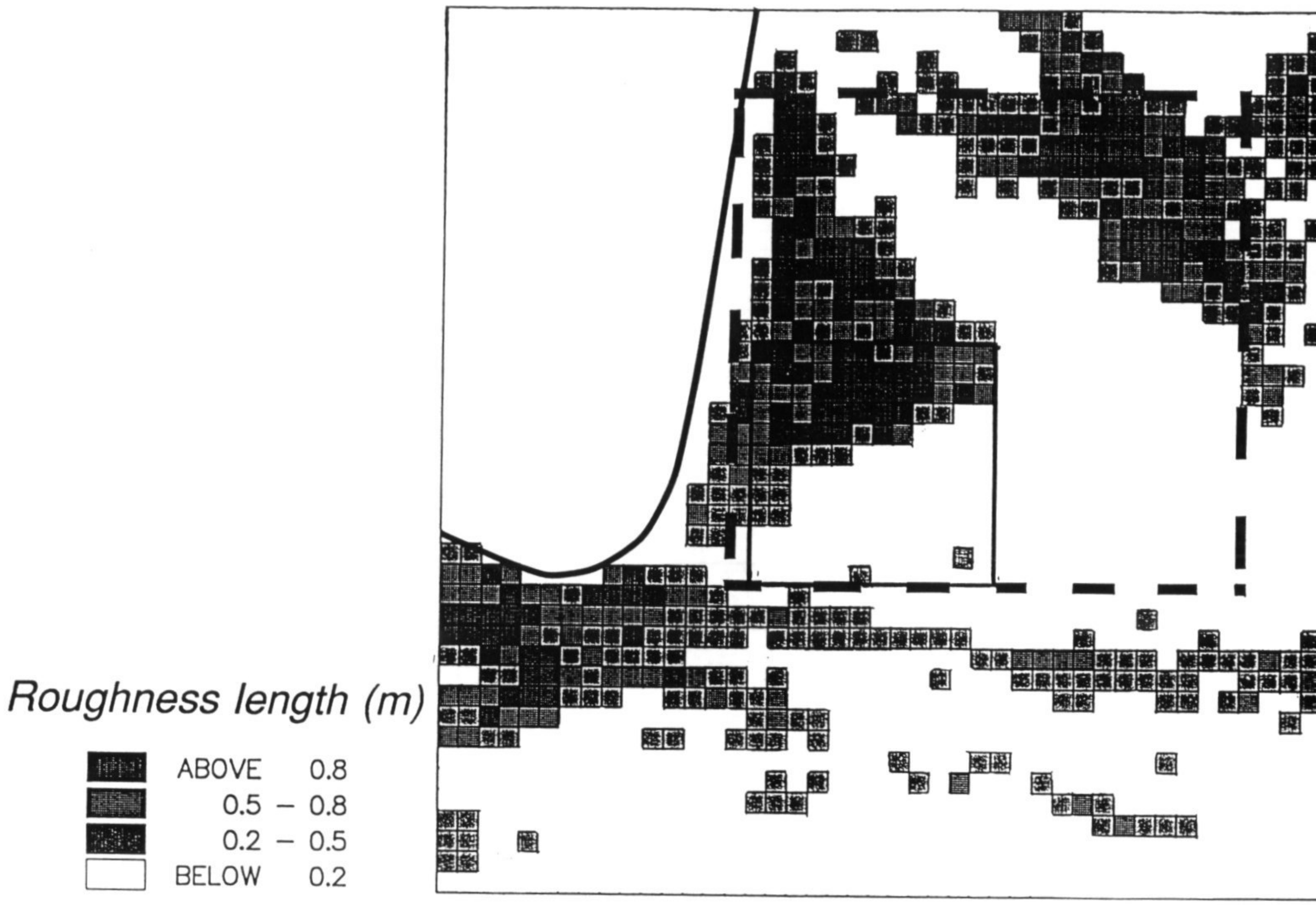


3. Modelled profiles at position (24,25) at 15:00 UT of
(a) sensible heat flux, (b) latent heat flux and (c) turbulent kinetic energy

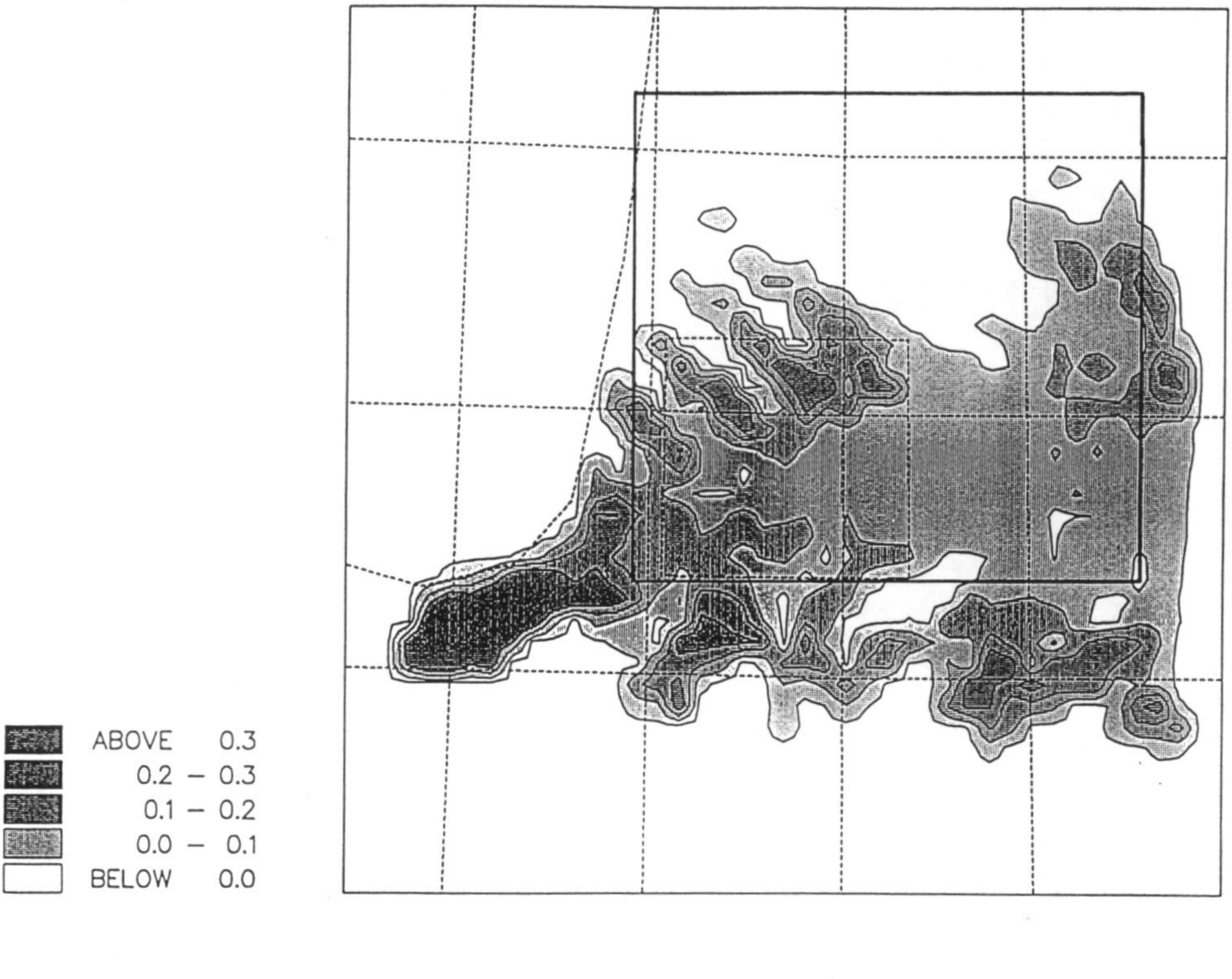

a
Rainfall (mm)
ABOVE 12.0
8.0 – 12.0
4.0 – 8.0
BELOW 4.0
b
Rainfall (mm)
ABOVE 12.0
8.0 – 12.0
4.0 – 8.0



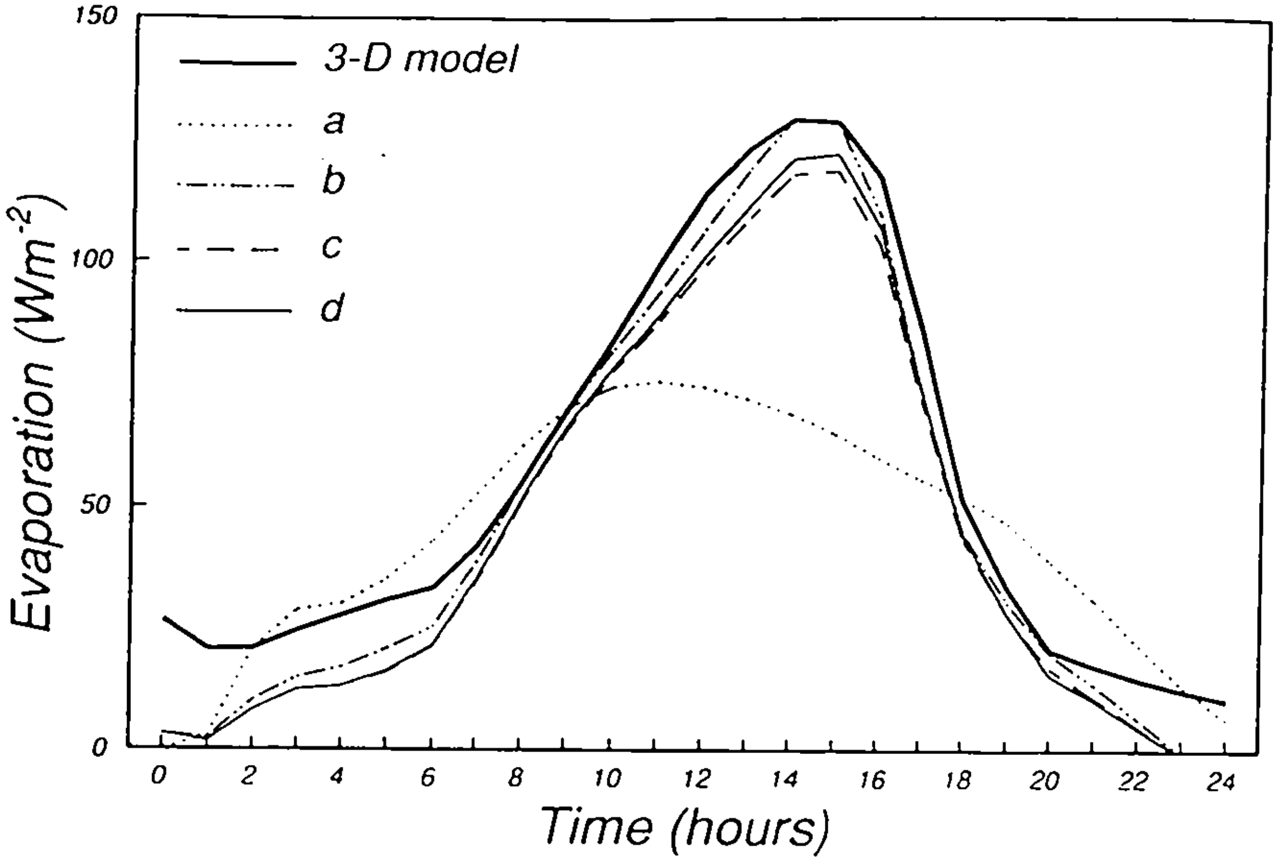
4. Contours of cumulative, modelled rainfall in mm for (a) the control run, (b) the forest run and (c) the bare soil run



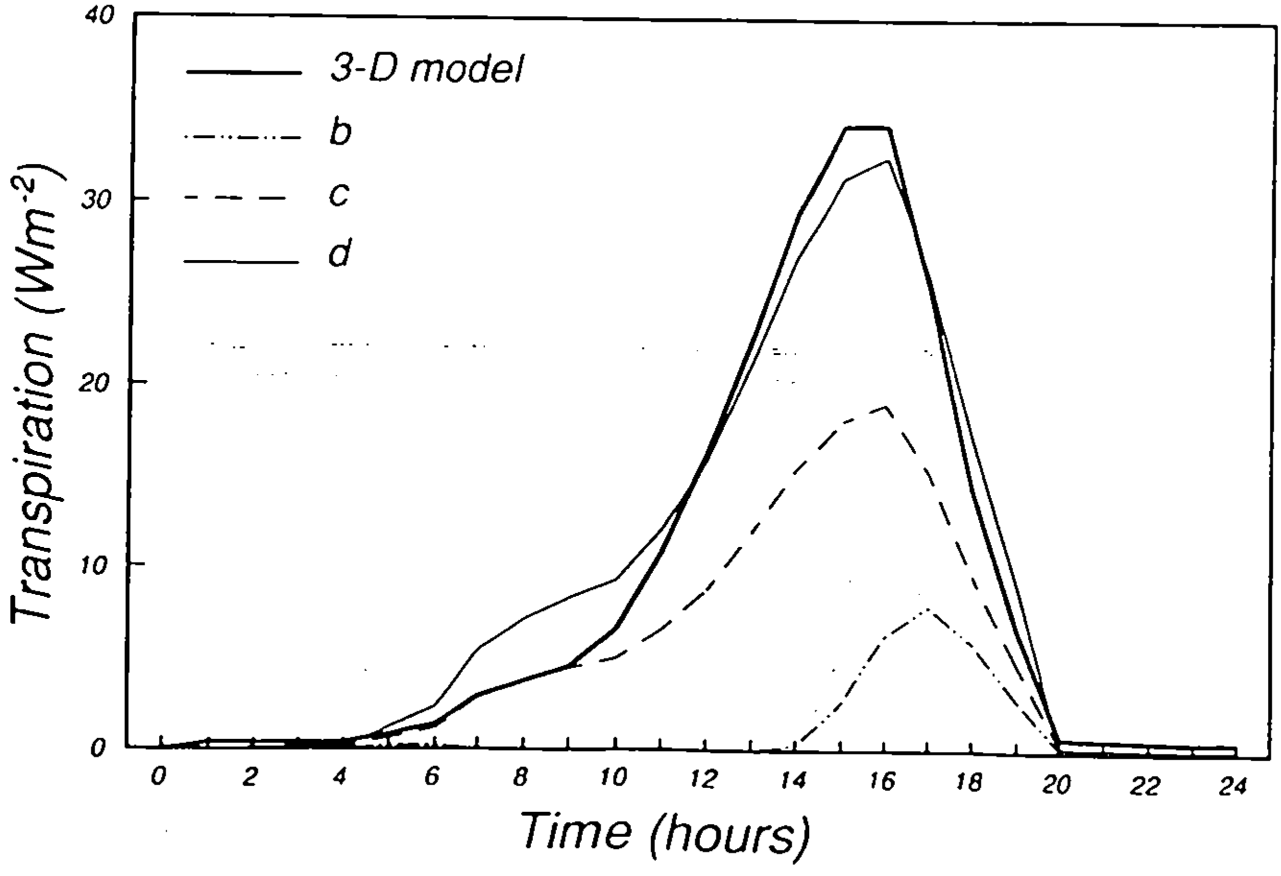
5. Values of roughness length in m in the model. The heavy, dashed line square is the position of the subdomain



6. Modelled fields of wetted vegetation at 15:00 hrs



7. Diurnal variations of modelled evaporation flux.



8. Diurnal variations of modelled transpiration.

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