The possible influence of urban surfaces on rainfall development: a sensitivity study in 2D in the meso-γ-scale

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Abstract

Urban areas can represent a considerable part of the model domain in meso-scale numerical simulations with typical horizontal domain lengths of 20–200 km. This paper addresses the question of the extent of the influence of urban surfaces on the development of convective precipitation in meso-γ-scale numerical models. For this purpose, a spatially variable parameterization scheme for surface sensible heat flux, surface latent heat flux, and roughness is introduced into a meso-scale numerical model. A sensitivity study in 2D is performed to assess the impact of variations of the individual parameters on the development of precipitation. The results indicate that surface conditions should not be neglected and can have considerable influence on convective rainfall. It appears that within a time frame of 4 h it is particularly the sensible heat flux variations that have the most significant impact. © 2000 Elsevier Science B.V. All rights reserved.

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1. Introduction

Urban areas modify boundary layer processes mostly through the production of an urban heat island (Vukovich and Dunn, 1978; Bornstein, 1986; Draxler, 1986; Oke, 1987).
1987) and by increasing turbulence through locally enhanced roughness (Loose and Bornstein, 1977; Stull, 1988; Theurer et al., 1992; Rotach, 1993a,b). Thus, as a consequence, larger scale meteorological processes can also be affected by the presence of urban agglomerations (Huff, 1986; Oke, 1987; Goldreich, 1995).

Although the importance of the lower boundary for meso-scale models has been discussed for some time (see e.g., Pielke, 1984; Pielke et al., 1991), detailed representation of the surface heat and moisture fluxes has only recently been included in meso-scale numerical modelling (Lee et al., 1993; Moussiopoulos, 1994). The principal reasons for this are that the heat and moisture transfers at the surface depend on a variety of soil parameters (i.e., heat capacity, permeability, porosity) as well as vegetation parameters (i.e., type of plant, leaf surface, root density), for which explicit data are mostly not available. A realistic treatment of the lower boundary, however, requires these parameters for the calculation of the energy and moisture transport.

Existing soil models are mostly designed and validated for rural surfaces (e.g., Sievers et al., 1983; Noilhan and Planton, 1989; Lee et al., 1993; Braud et al., 1995). However, in meso-scale numerical models, with grid lengths of 100 m up to a few kilometers, urban areas can represent a considerable fraction of the model domain and should not be neglected. In urban areas, artificial surfaces (concrete, asphalt, stone pavements, slate) and natural surfaces (parks, trees, gardens) vary on a relatively small spatial scale, which makes the definition of representative parameters for meso-scale grid dimensions difficult. Further, the transfer coefficients for heat and momentum are affected by turbulent flow, which is not well-established in inhomogeneous urban environments.

The complexity of the system increases further when precipitation is considered. In the case of intensive rainfall, for example, impermeable and normally dry surfaces can be transformed into temporary lakes, locally enhancing the evaporation. On the other hand, drainage systems transport away much of the rainwater, which is then no longer available for evapo-transpiration processes. Thus, modifications of the precipitation fields due to the presence of urban areas are of particular interest for meteorologists, as well as being crucially needed by hydrologists.

On the synoptic scale Loose and Bornstein (1977) have shown by observation and modelling that urban areas may significantly affect synoptic scale fronts. In cases of weak to moderately developed heat islands, most fronts slow down when moving over the urban area. In cases of pronounced heat islands, the retardation is less pronounced but still observable.

Effects of large cities on meso-scale or convective rainfalls were studied by the Metropolitan Meteorological Experiment (METROMEX), which took place in the 1970s in the US (Changnon et al., 1977; Huff, 1986). The experiment suggested that large urban areas could have an impact on the frequency of convective storms. Enhancement of rainfall frequencies over and downwind of urban areas was one of the results of this vast study. In the vicinity of St. Louis, for example, heavy storm events occur twice as often as in areas without urban agglomerations, and rain bearing cells exposed to the urban environment yield more precipitable water than comparable cells in rural areas (Huff, 1977). Rainfall enhancement is observed for the urban area itself and at a distance of 40 km downwind of the city centre. This increase was attributed mostly to the
modifications of the airflow due to the heat island and the increased roughness. The possibility that the presence of increased giant CCN concentrations originating from the urban area could also play a role in the precipitation processes was not excluded.

The present study is a first step in the development of a surface model for both rural and urban areas, in a meso-γ-scale meteorological model. The sensitivity of meso-scale processes to the lower boundary is assessed, using surface and soil-dependent parameters. Both rural and urban surfaces are considered, allowing on the one hand a study of the influence of urban areas on cloud and rainfall development, and on the other hand an assessment of the relative importance of the individual surface parameters. To achieve this, a three-parameter scheme describing surface roughness and soil-dependent surface sensible and latent heat fluxes is introduced into the Clark model (Clark, 1977; Clark and Grabowski, 1991), as described in Section 2.

Processes leading to precipitation are essentially 3D. Since it is difficult, however, to make extensive sensitivity studies in a 3D framework, for this study the 2D version of the model is used. It is assumed that the 2D simplification provides a good means to assess the relative importance of the response of the atmosphere to different surface parameters that describe the surface conditions. Effects becoming apparent in 2D should also be qualitatively relevant in 3D, even if quantitative differences should be expected. Those surface parameters affecting rainfall in the 2D simulations would then be tested in a 3D framework in further studies.

The set-up of the simulations is presented in Section 3. The impact of the individual surface parameters on meso-scale processes is assessed by analysing the rainfall development (Section 4). Conclusions are drawn in Section 5.

2. The Clark model

The Clark model (e.g., Clark, 1977; Clark and Farley, 1984) represents a finite-difference approximation to the anelastic, non-hydrostatic equations of motion. It expands the system variables around profiles of an idealized atmosphere with constant stability (denoted as overbar), deviations from the hydrostatic equilibrium (denoted with a prime), and space- and time-dependent perturbations (denoted as double prime). The Navier–Stokes equation is approximated as:

$$
\rho \frac{dv}{dt} = -2 \bar{\rho} \vec{\Omega} \times \vec{v} - \nabla \bar{p}^r + k \bar{\rho} g \left( \frac{\Theta^r}{\gamma p} - \frac{q_r'}{\gamma} + \varepsilon (q_v'' - q_c - q_R - q_l) \right) + \frac{\partial \tau_{ij}}{\partial x_j}
$$

(1)

where $\rho$ is the density of air, $\vec{v}$ is the 3D wind vector, $k$ is the unit vector in the vertical direction, $\vec{\Omega}$ is the angular rotation vector of the earth, $p$ is the atmospheric pressure, $g$ is the gravitational constant, $\Theta$ is the potential temperature, $\varepsilon$ is $R_v/R_d - 1$ with $R_d$ and $R_v$ being the gas constants of dry and humid air, $\gamma$ is the ratio of the heat capacities at constant pressure and constant volume ($c_p/c_v$), $q_v$ is the water vapour mixing ratio, $q_c$, $q_R$, and $q_l$ are the mixing ratios of cloud water, rainwater and ice, respectively, and $\tau_{ij}$ is the stress tensor describing the subgrid-scale turbulent processes.
Heat and moisture development are expressed by

\[ \frac{d\Theta}{dt} = \frac{\rho L}{\rho C_p} C_d + \nabla \cdot \left( \rho K_H \nabla \Theta \right) \]

\[ \frac{d\theta}{dt} = -\rho C_d + \nabla \cdot \left( \rho K_H \nabla \theta \right) \]

where \( L \) is the latent heat released from condensation processes, \( T \) is the temperature, \( C \) gives the rate of all phase transition of atmospheric water, and \( \Theta \) is a non-dimensional normalized potential temperature defined as \( \Theta = (\Theta' + \Theta'')/\Theta' \). The budget equations of the mixing ratios of cloud water, \( q_c \), rainwater, \( q_r \), and ice, \( q_i \), are given in detail in Bruintjes et al. (1994). \( K_H \) is the Richardson number dependent variable eddy mixing coefficient used for heat and moisture transport (Clark, 1979).

The microphysics in the model includes warm rain parameterization based on the Kessler (1969) scheme, and ice parameterization based on the work of Koenig and Murray (1976). Since this model has been described elsewhere (e.g., Clark, 1977; Clark and Farley, 1984), only the basic features and modifications are described here.

### 2.1. The surface parameters

The parameters used to characterize the surface in this study act on the sensible and latent heat fluxes as well as on the surface roughness. The surface forcing of the heat and moisture fluxes are described by adding the following terms to the subgrid-scale mixing terms \( \rho K_H \partial \Theta' / \partial z \) and \( \rho K_H \partial \theta / \partial z \) in Eqs. 2 and 3, respectively.

\[ \frac{1}{\rho C_p} \mu(x) S_0 e^{-\alpha f(Z, \varphi, h)} \]

\[ \frac{1}{L \Theta} B(x) S_0 e^{-\alpha f(Z, \varphi, h)} \]

where the surface sensible and latent heat fluxes depend on (i) \( \mu \), a factor describing the conversion from incoming solar radiation to sensible heat flux depending on the soil type; (ii) \( B \), the Bowen ratio; (iii) \( f \), a geometric function describing the reduction of the solar energy \( S_0 = 1395 \text{ W m}^{-2} \) by variation of the angle of the sun to the surface as a function of the zenith angle \( Z \), latitude \( \varphi \), and topography gradients in the East direction \( h \) (see Clark and Gall, 1982). The surface heat fluxes are distributed vertically using an exponentially decaying function, where \( \alpha \) is the height at which the surface sensible heat flux has dropped to 1/e of its initial value. Using an exponential vertical distribution function as opposed to a linear distribution function results in most parts of the fluxes distributed in low levels, but small amounts of heat being transported even above the boundary layer.

The surface shear stress \( \tau_0 \) is a function of the drag coefficient \( C_D \) and of the fluid velocity \( U \)

\[ \tau_0 = \rho C_D U^2 \]
The relationship between the drag coefficient \( C_D \) and the surface roughness is given by:

\[
C_D = \kappa^2 / \left( \ln \left( \frac{z_{\text{surf}}}{z_0(x)} \right) \right)^2 f_m \left( R_i, \frac{z}{z_0} \right)
\]

where \( z_{\text{surf}} \) is the height of the first grid point above the surface, and \( \kappa \) the von Kármán constant. \( C_D \) is applied to the stress tensor components \( \tau_{13} \) and \( \tau_{23} \). The function \( f_m \) represents the influence of atmospheric stability, and is specified as:

\[
f_m = 1 - \frac{2 a_j R_i}{1 + 3 a_j b_j C_u |1 + (z/z_0)| |R_i|} \quad R_i < 0
\]

\[
f_m = \frac{1}{1 + \left( \frac{2 a_j R_i}{1 + a_j |R_i|} \right)} \quad R_i > 0
\]

where all three coefficients \( a_j, b_j, \) and \( c_j \) are set to 5, and \( R_i \) is the local Richardson number (see Clark et al., 1996).

The spatially varying parameters used to characterize the surface are the conversion rate \( \mu \), the Bowen ratio \( B \), and the roughness length \( z_0 \). Although these parameters are time invariant, the surface forcings change with time, since the equations also contain time-dependent functions, such as the elevation angle of the sun in the case of the heat fluxes, and the wind speed in the case of the surface shear stress. For this sensitivity study effects of cloud cover, which would introduce random reductions of the surface fluxes in time and space, have not been considered in order to concentrate on the spatial variability of the surface parameterization.

3. Set-up of the simulations

3.1. The model domain

The model domain describes a 200-km west–east cross-section including an urban area (Paris, France), which is situated between 50 and 90 km from the western boundary (Fig. 1). The horizontal grid resolution is constant and set to 500 m. The vertical co-ordinate is variable, resolving the boundary layer with a resolution as fine as 40 m and the higher troposphere with an average spacing of 250 m. The domain height extends to 23 km. Using a time step of 10 s, 250 min are simulated for each run.

3.2. The profile

The simulations are initialized with one profile of temperature, dew point, and winds (Fig. 2), which is distributed homogeneously over the whole domain. The profile is based on the midday sounding from Trappes, situated about 30 km west of Paris, on June 27, 1990. The sounding is unstable with a convective available potential energy (CAPE) of 2140 J/kg. Calculation of two stability indices indicates a high likelihood for the outbreak of storms on that day: the Rackliff (1962) index is 31 (threshold for
Fig. 1. $yz$-cross-sections of the surface $\mu$ (a), $B$ (b) and $z_0$ (c) are shown by the shaded area. The topography is outlined by the solid line. The length of the domain in indicated in km on the $x$-axis, the $y$-axis on the left-hand side gives the scale for the conversion rate, the Bowen ratio, and the roughness lengths $z_0$ in m. The $y$-axis on the right-hand side gives the scale for the topography in m.

Increased likelihood of thunderstorms is 27, and the Boyden (1963) index is 96 (threshold 94). The profile is moist up to a height of about 5 km, then topped by a comparatively drier layer. The total precipitable water in the air column from cloud base to cloud top is 40.7 mm.

The winds on June 27, 1990 were southwesterly in the lower troposphere, and westerly throughout the rest of the atmosphere, showing steadily increasing wind speeds from 1.5 m/s at the surface to 25 m/s at 9 km. For the 2D study, the observed wind profile has been modified to account for the restricted westerly flow. The wind speeds have been slightly reduced, and increase steadily from 0.5 m/s at the surface to a maximum of 22 m/s at 9 km.

The model simulations are started 2 h prior to the sounding, and therefore, the temperature and dew point values of the midday profile have been slightly cooled in the boundary layer to account for the earlier simulation time.
3.3. Initialization of soil parameters and the surface

A high resolution land-use data set was provided by the Institut Géographique National de France for the area of Île de France, a region of approximately 130 km$^2$ in northern France. From this data set, 10 different soil classes are constructed, and each is then assigned typical values for $\mu$, $B$, and $z_0$. In the case of the Bowen ratio ($B$) and the roughness length ($z_0$), these are extracted from standard references such as Oke (1987) or Stull (1988). The estimation of the conversion factor from incoming solar radiation to sensible heat flux, $\mu$, is based on results of Clark and Gall (1982) and Thielen and Gadian (1996).

For the 2D studies, non-averaged west–east cross-sections intersecting the city centre of Paris have been extracted from the complete data set. Fig. 1 shows the cross-sections of the three surface parameters across the whole model domain. The urban area is situated roughly between 50 and 90 km, and is surrounded by less urbanized or rural areas. Fig. 1a and b illustrate that for this cross-section, up to 37% of the incoming solar radiation is converted to sensible heat flux over the urban area, while over rural areas...
this conversion can be up to 20% lower. The height $\alpha$ (Eqs. 4a and 4b), at which the surface sensible fluxes have decreased to $1/e$ of their initial values is set to 500 m. The value of the surface roughness length, shown in Fig. 1c, increases from 20 cm in the rural areas to maximum values of nearly 2 m in the city centre. On the outflow boundary, no detailed surface data are available from 120 km onwards, and therefore a constant, representative of the rural environment, is applied.

The ranges of $\mu$, $B$, and $z_0$ for the different surfaces are shown in Table 1.

For the following model studies, we assume that each surface parameter can have two different configurations.

(a) The surface parameter is constant throughout the length of the domain. This is denoted with a C. The constant values have been chosen more or less arbitrarily, but are targeted to represent the threshold value between rural and urban conditions. Values used are $\mu = 0.28$, $B = 0.8$, and $z_0 = 0.2$ m.

(b) The surface parameter varies spatially as illustrated in Fig. 1a–c by the shaded curve. It describes a rural environment including an extended urban area of about 40 km in diameter. This is the so-called ‘urban’ set-up and is denoted with a U.

### 3.4. Boundary conditions and filters

The model is set up with open lateral boundary conditions, and free-slip upper boundary conditions, using an 8-km deep Rayleigh friction absorber layer to dampen reflection of waves at the artificial upper boundary. The surface is treated as a no-slip boundary. A Robert–Asselin time filter is applied to prevent the development of numerical instabilities due to the time-centred differencing schemes.

### 3.5. Presentation of the different simulations

The notation for the different runs is a composite of the two letters C or U. The first letter stands for the sensible heat flux parameter $\mu$, the second for the Bowen ratio $B$, and the third for the roughness length $z_0$. For example, CCC denotes the control run where all parameters are kept constant, UCC denotes the simulation where only the sensible heat flux parameter varies while all the other parameters are kept constant. In order to exclude superposition of soil forcing with topography-related effects such as differential slope heating, varying terrain is only considered in a few explicit runs. In these cases, the same configuration as for the control run (CCC) and the urban run

<table>
<thead>
<tr>
<th>Type of surface</th>
<th>$\mu$</th>
<th>$B$</th>
<th>$z_0$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>City centre, tall buildings</td>
<td>0.30–0.40</td>
<td>1.5–3.0</td>
<td>1.0–2.0</td>
</tr>
<tr>
<td>Urbanized areas, villages</td>
<td>0.25–0.30</td>
<td>0.8–1.5</td>
<td>0.5–1.0</td>
</tr>
<tr>
<td>Forest, open grassland</td>
<td>0.20–0.30</td>
<td>0.2–0.8</td>
<td>0.2–0.5</td>
</tr>
<tr>
<td>Lakes, rivers</td>
<td>0.10–0.20</td>
<td>0.1–0.2</td>
<td>0.001–0.2</td>
</tr>
</tbody>
</table>
Table 2  
Description of the different simulations for the three parts including the name of the simulations, the minimum and maximum values for the surface parameters $\mu$, $B$, and $z_0$ as used in the simulations.

<table>
<thead>
<tr>
<th>Simulation</th>
<th>$\mu$</th>
<th>$B$</th>
<th>$z_0$ (m)</th>
</tr>
</thead>
<tbody>
<tr>
<td>Part I</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>CCC</td>
<td>0.28</td>
<td>0.8</td>
<td>0.2</td>
</tr>
<tr>
<td>UUU</td>
<td>0.10–0.40</td>
<td>0.2–2.0</td>
<td>0.2–2.0</td>
</tr>
<tr>
<td>Part II</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>UCC</td>
<td>0.10–0.40</td>
<td>0.8</td>
<td>0.2</td>
</tr>
<tr>
<td>CUC</td>
<td>0.28</td>
<td>0.2–2.0</td>
<td>0.2</td>
</tr>
<tr>
<td>CCU</td>
<td>0.28</td>
<td>0.8</td>
<td>0.2–2.0</td>
</tr>
<tr>
<td>Part III</td>
<td></td>
<td></td>
<td></td>
</tr>
<tr>
<td>TOPC</td>
<td>like CCC but with topography</td>
<td></td>
<td></td>
</tr>
<tr>
<td>TOPU</td>
<td>like UUU but with topography</td>
<td></td>
<td></td>
</tr>
</tbody>
</table>

(UUU) are used combined with topography. They are denoted as TOPC and TOPU, respectively. The different simulations are listed in Table 2.

The presentation of results is divided into three parts. The first part describes a comparison between the control run CCC and the most realistic run UUU, where all surface parameters vary spatially. In Part II, the relative importance of the individual parameters for the atmospheric processes is discussed for the sensible heat flux parameter (UCC), the Bowen ratio (CUC), and the roughness length (CCU). The third part addresses the additional influence of topography by comparing results between TOPC and CCC, and between TOPU and UUU.

4. Results

The spin-up time varies for the different simulations and is the shortest for the “urban” run at 80 min, and the longest for the control run at 95 min. The analysis is therefore focused on integration times of longer than 100 min.

Rainfall data have been calculated at the surface. All other data presented in the following analyses, such as fluxes and vertical velocity, are a mean of data at the surface and the next grid point at 40 m.

4.1. Part Ia: comparison between the control run CCC and the urban run UUU

Fig. 3a shows the accumulated rainfall over the last 3 h of simulation CCC. The maximum amounts to 12.7 mm, with an average rainfall total of 2.3 mm, and a standard deviation of 3.1. The curve shows that there are two local maxima located at around 35 and 165 km, which are of comparable strength, and which vary between 10 and 12 mm. In the centre of the domain rainfall totals are low.

The temporal development of the surface rainfall for CCC is summarized in Fig. 4. The spatial co-ordinate, showing the horizontal length of the domain, is presented on the
Fig. 3. Rainfall totals for CCC (a) and UUU (b), accumulated from 100 to 250 min of simulation and averaged over 10 grid points. The rainfall total is given in mm, the $x$-axis gives the distance from the western boundary in km. The location of the urban surface is indicated by two dashed lines in (b).

As could be expected from earlier studies (e.g., Hill, 1974), the homogeneous heating at the ground produces a superadiabatic lapse rate which destabilizes the air above the ground. Subsequently, first a sub-cloud circulation with regularly spaced up and downdraught regions is initiated prior to the development of clouds (not shown). Within the stronger updraughts of this circulation, small convective clouds with an average width of 2–3 km develop about 100 min throughout the length of the domain. Although most of them contain rainwater, they are small, with an average width of 2–3 km. Except for the rainfall cell at 30 km east (see Fig. 4A), most other cells only produce traces of rainfall ($<0.3$ mm/h) and last for less than 10 min.

Small and short-lasting cells are formed continuously after 100 min throughout the simulation. But the airflow is increasingly dominated by the cloud convection and downdraughts caused by the precipitation, which results in convective cells, which are organized on larger space and time scales than during the first 100 min of simulation. Larger cells occur at 100 min (a), 170 min (b), and 210 min (c), at an average spacing of...
Fig. 4. Distance and time-plot of the rainfall intensity for the control simulation CCC. The distance is shown on the x-axis in km, and the simulation time on the y-axis in min. The rainfall intensity is given in mm/h. A lower cut-off level of 1 mm/h is chosen allowing a better presentation of the isolated precipitation cells. The contour lines are drawn in 1, 5, 10, 20, 30 and ≥ 50 g/kg. The grey shades are defined by the grey scale next to the diagram.

60–80 km. The larger rainfall fields average 15–20 km in width, and last between 30 and 60 min. At some locations, repeated formation of rainfall is observed, e.g., around 30 km (A) and 160 km (C). To some extent, re-occurring rainfalls are also simulated at 110 km (B), but those cells are weak and do not result in increased rainfall totals (Fig. 3a).

The total accumulated rainfall for simulation UUU is shown in Fig. 3b. Obviously, the rainfall patterns are strongly influenced by the variable surface parameterization scheme, and switching on of all surface parameters results in rainfall enhancement and re-organization of the rainfall fields. As compared to CCC, the rainfall is focused over the urban surface and at a distance of 60–80 km downwind of it. Unlike in CCC, hardly any rainfall is simulated west of the "urban" area. The maximum rainfall, located over the eastern part of the "urban" area, is 13 mm greater than in the control run, and amounts to 26 mm. The average rainfall is with 6.3 mm more than twice as high as in CCC, and also the standard deviation is significantly higher with 7.6.

Fig. 5 shows the surface sensible heat fluxes (solid) and the surface latent heat fluxes (dotted) at noon, spatially averaged over 5 km for simulation UUU (and also UCC for the sensible heat fluxes and CUC for the latent heat fluxes). Since no topography is considered, they are obviously closely related to the input parameters given in Fig. 1.
Fig. 5. Surface sensible (solid line) and surface latent heat fluxes (dashed line) at noon for simulation UUU. The distance from the western boundary is given on the x-axis in kilometers, the strength of the heat fluxes in W m⁻². The data is averaged over 10 grid points.

The surface latent heat fluxes show the largest variability, and range from 100 W m⁻² over the urban area to 575 W m⁻² over rural parts. In comparison, the surface sensible heat fluxes range only between 200 and 350 W m⁻². Towards the outflow boundary of the domain, where the surface parameters do not vary spatially, the fluxes remain at a constant value of 330 W m⁻² for the latent heat fluxes and 275 W m⁻² for the sensible heat fluxes. These are the same values for simulation CCC across the whole domain. The urban area is clearly characterized by an increase in surface sensible heat flux of 100 W m⁻² and a reduction in surface latent heat flux of an average of 200 W m⁻². The average value of the surface sensible heat fluxes amounts to 270 W m⁻² over the rural areas and to 324 W m⁻² over the urban area. In contrast, the average surface latent heat fluxes amount to 350 W m⁻² over the rural areas and to 170 W m⁻² over the urban area. The overall average of the surface latent heat fluxes is 35 W m⁻² higher than that of the surface sensible heat fluxes. Comparing with data from the literature (e.g., Stull, 1988), it appears that the variation and magnitude of the surface sensible heat fluxes agree well with mid-June conditions, while the surface latent heat fluxes may be slightly overestimated over some rural areas.

The increased surface sensible heat flux over the urban area produces increasingly high-reaching convection, resulting in the development of rainwater within 80 min at this location. This is illustrated in Fig. 6, which shows the temporal development of the
Fig. 6. Distance and time-plot of the rainfall intensity for the simulation UUU. The two dashed lines indicate the approximate limits of the urban surface to the West and the East. Axes and grey scale same as Fig. 3.

The precipitation becomes spatially focused at two locations, and re-occurring rainfalls appear over the urban surface forcing area (A), indicated roughly by the two dashed lines in Fig. 6, and about 60 km downwind of it (B). Repeated formation of rain cells occurs particularly at the eastern and leeward part over the urban area, for instance at 90, 120, 150, and 180 min, resulting in higher accumulated rainfall over this region (26 mm) than elsewhere in the domain (Fig. 3). Over the western (windward) parts, the rainfall is comparatively low. At 70 km downwind of the urban area, a long-lasting cell (> 100 min) with several regions of increased convective rainfall is formed at 80 min. Towards later stages of the simulation (around 200 min), another smaller and shorter-lasting cell is triggered at the same location.

Fig. 7 shows the vertical wind averaged over the last 3 h of the simulation, from 100 to 240 min, for simulation CCC (a) and UUU (b). Spatially, the data have been averaged over 4-km strips. Descending winds have a negative sign. The location of the urban area is marked by the two dashed lines in (b).
In CCC the average vertical winds fluctuate around zero across the domain, showing only a light periodicity in magnitude at a distance of 40 to 50 km. The three local maxima coincide approximately with the location of the three regions of repeated rainfall (Fig. 4).

Unlike in the control run, in simulation UUU the average winds are mostly ascending and less fluctuating. Two particular regions can be distinguished.

(a) Over the urban region, the airflow is ascending west, windward of the urban area, and descending in the east, on the leeward side. This airflow pattern is mostly a consequence of the urban heat island. The buoyant air rises over the heated surface, causing convergent winds in the lower boundary layer. The inflowing air becomes then transported upwards to cloud base. Depending on the strength of the updraughts and the local cloud environment, conversion from cloud water to rain and ice is initiated, which eventually falls out in the compensating downdraught region over the leeward parts of the urban area (Fig. 6).

(b) The second region corresponds to the area of increased convective activity downwind of the city centre, and is characterized by large variations of $\bar{w}$. The urban area represents an obstacle to the air flow because it initiates the rising of buoyant air and it enhances the turbulence of the flow due to increased surface roughness. The large variations of $w$ are at least partly induced by the wake structure of the air flow.
downwind of the obstacle. It is likely that within a 3D framework, these waves would be less apparent.

4.2. Part Ib: comparison between the control run UUU and observational data

The initialization profile was taken on June 27, 1990, a day of important convective activity in the Paris region. During the afternoon of June 26, 29°C was measured in the city centre, while the temperatures in the rural surroundings were about 3°C cooler. During the night, moist air from the southwest was advected into the area, and thunderstorms developed and caused heavy rainfall over Paris.

Fig. 8 shows the isohyets of the rainfall totals for the night of 26/27 June (a) and for the afternoon of June 27 (b) as deduced from gauge data. The dots show the positions of the rain gauges, and the dashed lines indicate the boundaries of Paris, the dense urban area typical of large city centres, and of Seine-Saint-Denis a less urbanized county Northeast of Paris.

Fig. 8a shows two areas of high rainfall totals of 30 mm, one centred over Paris, and one about 25 km downwind of it, affecting the western parts of Seine-Saint-Denis. During the afternoon of June 27, another storm approached the area, again from the southwest. The rainfall was higher than in the previous storm and produced totals of 50 mm (Fig. 8b). Again, two regions of equally intense rainfalls were observed, one over the northeastern (leeward) side of the city centre, and the second one about 20 km downwind of it, causing serious flooding in the area of Seine-Saint-Denis (Thielen and Creutin, 1997). Over the southwestern and windward side of the city centre of Paris, a rainfall of lower than 20 mm was observed.

Qualitatively, the analysis of the two storms agrees with simulation UUU on (a) the tendency of low rainfalls over the windward part of the city centre, and (b) the occurrence of two equally strong local rainfalls, one located over the leeward side of the city, and one at a considerable distance downwind of it.

Quantitatively, the magnitude of the simulated rainfall totals of 26 mm in UUU are lower than the observed ones, in the case of the last storm, by a factor 2. Although at a few occasions the simulated maximum rainfall intensities in UUU exceed 100 mm/h, the rainfall totals remain low because of the short duration of the cells. There are several factors which could cause the underestimation of the rainfall, e.g., the parameterization of the microphysics, in particular of the ice phase, which plays an important role in high-reaching convective storms in our latitudes, or the lack of directional wind shear which is an important feature of convective storms (Browning and Ludlam 1962), to name a few.

Larger discrepancies become apparent when comparing the distances between the urban area and the increased rainfall downwind of it. In both storms, the observed distance on June 27 is only half the distance simulated. The observations on June 27 are supported by Escourrou (1991), who also found for Paris a typical distance between two areas of increased rainfall of 20–30 km. It is likely that the overestimation of this distance is also an effect of the 2D simulations, since the air flow is not able to go around obstacles as would be possible in a 3D framework, having consequences on the wake structure downwind of the perturbation.
Compared to the observations from METROMEX, the results for UUU are also in qualitative agreement. It was concluded that large urban areas have impact on the frequency of convective storms (Huff and Vogel, 1973; Chagnon, 1978). Storms tended to occur twice as often over urban agglomerations relative to comparable rural areas. Heavy rain bearing cells exposed to the urban environment yielded 70% more precipitable water than comparable cells in rural regions (Huff, 1977). Rainfall enhancement was observed at an average distance of 40 km downwind of the city centre, a larger distance than observed for the Paris region. It is likely that this is due to the larger extension of urbanized areas in the US relative to Europe, and Paris in particular.

The relative importance of the individual parameters is now studied by switching on only one parameter at the time.

4.3. Part II: relative importance of the three individual surface parameters

4.3.1. Influence of the conversion rate $\mu$ (UCC)

Fig. 9a shows the rainfall totals for simulation UCC (solid line). For comparison, the data for UUU are also plotted (dotted line).

Variation of only the conversion rate $\mu$ results in similar precipitation patterns as in UUU where all the three parameters vary spatially. As compared to UUU, the maximum rainfall over the urban area (20.6 mm) is shifted slightly to the west, and rainfall east of the urban surface is higher. For example, at 125 km, no rainfall is simulated for UUU, while an average of 10 mm is simulated for UCC. Particularly towards later stages of the simulation, after 200 min, the rainfall fields become slightly larger and with more individual convective cells embedded than in UUU, and overall, a higher number of small and short-lasting rain cells are simulated.

The few differences between the runs UCC and UUU seem to indicate that with the present set-up the surface sensible heat flux is the dominant forcing of the rainwater development. This is not surprising because the underlying configuration is based on an unstable environment, weak winds and wind shears, and no topography, leaving positive buoyancy as the dominant forcing to trigger convection.

These results support the observations that urban heat islands can influence convective rainfall development. Additional simulations that are not presented in this paper indicate that the stronger the heat island intensity, the more pronounced the effects are in the vicinity of the forcing area, while weak heat islands mostly favour increased rainfalls at some distance downwind of the heated surface.

4.3.2. Influence of the Bowen ratio $C_U$ (CUC)

Variation of the Bowen ratio produces similar rainfall patterns as control run CCC. This is illustrated in Fig. 9b, which shows the total rainfall for CUC (solid black line).

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Fig. 8. Isohyets of rainfall totals as interpolated by gauge data for (a) the storm during the night from June 26 to 27, and (b) for the storm during the afternoon on June 27. The dots indicate the location of the gauges (climatological network of Météo-France), the dashed lines show the boundaries of the county of Paris and the county of Seine-Saint-Denis.
and CCC (dotted grey line). As compared to CCC, the variable Bowen ratio results in a slight decrease of rainfall over and downwind of the urban surface (95–115 km), and a small increase of water content towards the end of the simulations.

Since the only difference between the two runs consists in the variation of the Bowen ratio, the differences in latent heat fluxes between CCC and CUC are examined next.

Fig. 10 shows the spatial distribution of the latent heat flux 10 m above the surface, averaged from 150 to 180 min at a sampling frequency 10 s, for simulation CCC (dotted line) and simulation CUC (solid line). During this time interval, light convection is present across the domain (Fig. 4). The data are spatially averaged over 10 grid points. As could perhaps be expected, there are no striking differences between the two cases. In CCC, the latent heat flux appears to fluctuate randomly between ±200 W m$^{-2}$ across the domain and no particular region of high or low activity can be distinguished. In CUC, the fluctuations are of the same magnitude as in CCC.

It appears that sharp changes in latent heat fluxes as indicated in Fig. 5 (e.g., at 20, 40 or 90 km), induce locally a slightly higher variability in the data. The biggest difference
Fig. 10. Spatial distribution of the time-averaged subgrid-scale latent heat flux at 10 m above the surface for simulation CCC (top) and simulation CUC (bottom). The distance from the western boundary is given on the x-axis in km, the strength of the heat fluxes in W m⁻². The data are averaged over 10 grid points.

between the two curves can be observed over the urban area and towards the outflow boundary where the time averaged fluxes are lower and less variable than in the control run CCC. These differences in latent heat fluxes, however, are not reflected in the rainfall development within the same time interval (not shown). The reduction of rainfall downwind of the urban area as shown in Fig. 9b, is produced from 200 min onwards only. The rainfall cell developing at 215 min located at 85 km from the inflow boundary in CCC (Fig. 4) is absent in CUC, and the cell at 235 min located at 105 km is much weaker in CUC than in CCC, producing the lower rainfall totals in that area. The slight increase of rainfall towards the outflow boundary in CUC is due to slight differences in rainfall from 170 min onwards, showing a small tendency of higher rainfalls in CUC than in CCC.

Therefore, it seems that sudden changes of the Bowen ratio at the surface may produce local changes in latent heat fluxes. Reduction of surface latent fluxes of 200 W m⁻² from rural to urban areas results in lower average latent heat fluxes in the lower boundary layer. The reduction of average latent heat fluxes, however, does not seem to be coupled to an instantaneous reduction in rainfall. Significant rainfall differences occur only towards the end of the simulations. It therefore seems that under convective conditions, local variations of the surface latent heat flux need more than 3 h to become
effective, and thus, have little effect on rainfall processes within the considered time and length scales.

One should note that the Bowen ratio has been kept constant in time throughout the simulations, which does not take into account the evaporation of rainwater accumulated over impermeable surfaces. In urban boundary layer studies, the latent heat flux is, therefore, an important variable for the urban climate. However, the reaction of convective rainfall processes to spatial variations in the latent heat fluxes seems slow, and in these relatively short simulations, it can be assumed that even a time-dependent Bowen ratio would not significantly modify the results.

4.3.3. Influence of the roughness length (CCU)

In this section, the effects of surface roughness changes on rainfall is assessed. Differences in total rainfall upwind of the urban surface between CCU and CCC are small (Fig. 9c), and only a light increase of rainfall of about 8–10 mm directly over the urban surface is simulated. In CCU, the highest rainfall totals are located between 130 and 150 km, 50–70 km downwind of the urban surface. In fact, with a maximum of 28.8 mm, this represents the highest rainfall total of all simulations. This suggests that a perturbation of the atmospheric flow through increased surface roughness may lead to enhanced rainfall at some distance downwind of the perturbation zone.

Analyses of the temporal development of the rainfalls (not shown) show that the influence of the increased turbulence over the urban surface becomes most effective after 3 h, and downwind of the increased surface roughness. The increased drag over the urban surface results in a slowing down of the air flow, which produces a local convergent wind field. The inflowing air is forced to rise, not as energetically as buoyant air, and after some time reaches the cumulus convection level at some distance downwind of the perturbation. Additional studies with different cut-off levels for $z_0$ (not shown) suggest that there is a relationship between roughness height and rainfall intensities. The higher the roughness height, the higher the rainfall downwind of the perturbation.

The main forcing on convective processes in this case is the increase of subgrid-scale turbulence in the lower levels. Fig. 11 illustrates the differences in Reynolds stress $\overline{\rho u' w'}$ (average between surface and the first grid point above) for the two simulations CCC (dotted line) and CCU (solid line). The Reynolds stress is averaged over the last 150 min of simulation and spatially averaged over 10 grid points.

In both cases, the Reynolds stress fluctuates around zero across the domain, with larger variations toward the western boundary, which is probably an effect of the inflowing air. The most significant difference between the two curves is the much larger variation of the Reynolds stress towards the outflow boundary in CCU, where maximum variations of $\pm 0.8$ N/m$^2$ (from 175 km onwards) are simulated, as opposed to $\pm 0.3$ N/m$^2$ in CCC. This region of large variability coincides with the region of increased convective rainfall, as illustrated in Fig. 9c.

These results suggest that in 2D the precipitation enhancement can also be a function of the differences of the roughness height. In 3D simulations, however, where the air flow has the possibility to go around an obstacle, the effect of roughness changes on the precipitation intensity might be considerably reduced. Based on observational data,
Escourrou (1991) estimated the influence of roughness on the development of precipitation in the region of Paris, and her results qualitatively confirm the simulations. For southwesterly winds, she observed an increase of precipitation from the outskirts of Paris onward, and particularly an increase in the northern parts of the city, downwind of the downtown area. A precipitation enhancement due to roughness of 10% is quoted, but it is not clear if this is a general value or if it refers explicitly to the Paris region.

4.4. Part III: influence of topography (TOPC, TOPU)

Irregular terrain has been added to the simulation set-ups of CCC (denoted as TOPC) and UUU (denoted as TOPU). The rainfall totals for both simulations are shown in Fig. 12 and compared against CCC and UUU, respectively.

In simulation CCC the topography is absent, and the surface forcings are constant across the domain. Therefore, first computational instabilities due to round-off errors need to develop and amplify before the perturbations are important enough to establish convective circulation patterns. Inclusion of topography produces from the very beginning local variations in the pressure, wind and temperature fields through upslope lifting and increased drag. Owing to the generally flat topography, no important temperature differences between sun facing and shadowed slopes can develop on the given time scale.
Consideration of topography results in earlier onset of rainfall development, reduction of small and short-lasting cells, and enhanced rainfall intensities towards the outflow boundary (not shown). Although the topography gradients are weak (1/300 m), it seems that in TOPC enhanced rainfall is simulated just upwind of the slightly elevated terrain 80 km east of Paris (compare Fig. 1 and Fig. 12a).

It is possible that, again, the influence of the surface drag on the atmospheric processes has been overestimated in the 2D framework, contributing to the increased rainfalls downwind of the urban surface, similar to simulation CCU. A comparison study in 3D is envisaged as possible future work to clarify this point.

As compared to UUU, in TOPU cells are slightly smaller and shorter-lasting. The rainwater is again focused over the urban area owing to re-occurring rainfalls. The biggest difference is that the enhanced rainfall downwind of the urban heat island is considerably weaker than in UUU (Fig. 12). This is somewhat contradictory to the findings for TOPC, where the effects of topography resulted in increased rainfall at some distance downwind of the urban surface. Thus, it seems that the modification of the atmospheric processes due to the surface are complex and manifold, and interactions of the different influences make their global effects very difficult to predict. For simulations aimed at reproducing particular case studies, the representation of the
surface can play an important role both for quantity as well as the spatial distribution of the rainfall.

5. Conclusions

In the present study, a spatially variable parameterization scheme for the surface sensible heat flux, the surface latent heat flux, and the roughness is introduced into a meso-scale numerical model. The results indicate that the surface conditions should not be neglected and can have considerable influence on meso-scale processes such as convective rainfall.

On a relatively short time scale of less than 4 h, it is mainly the surface sensible heat fluxes and subsequent buoyancy variations that influence the rainfall development. Inclusion of heat islands result in increased rainfall over and at a certain distance downwind of the heat island. The stronger the heat island intensity, the more pronounced the effects are in the vicinity of the forcing area, while weak heat islands favour increased rainfalls only at some distance downwind of the heated surface. Therefore, in order to produce realistic simulations, a correct initialization of the surface temperature and the surface sensible heat fluxes is most important.

The influences of local changes in the surface latent heat fluxes are comparatively slow acting and do not show significant effects within the 4 h of simulation. This is not surprising in this case, because the basic state of the atmosphere is chosen to be unstable and moist in the lower levels. It also seems that the parameterization of the Bowen ratio should be recalibrated since average values of latent heat fluxes appear low relative to the sensible heat flux in the given rural conditions.

Changes in the roughness length modify the precipitation patterns significantly downwind of the urban surface in these 2D simulations. The effects develop only after 2–3 h of simulation, and seem to be a function of the height of the roughness lengths, although this may well not apply for 3D simulations. However, in combination with a variable sensible heat flux, the buoyancy forces dominate the precipitation development. Similar conclusions could be drawn for those simulations that include topography. Because the topographical gradients are small, the artificially induced heat islands through urbanization remain the dominant surface forcing.

Overall, the results seem to confirm observations, such as those from the METROMEX experiment, that the frequency distribution of rain-bearing cells is enhanced over urban areas, and that precipitation can be enhanced by the presence of urban agglomerations.

For a more realistic assessment of the effect of these parameters on rainfall development, and for a quantitative comparison with observations, the simulations should be performed in three dimensions. 2D simulations represent a limitation for the representation of the dynamical response of the atmosphere to turbulent processes related to cloud formation and precipitation, which are essentially 3D. Further, the time dependency of the soil conditions should not be neglected. A next step will therefore include a soil model to calculate the surface temperatures and moisture contents at each time step dependent on soil conditions.
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References


