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Application of the concept of blending height to the calculation of surface fluxes in a mesoscale model

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Summary. The concept of blending height is implemented into the nonhydrostatic mesoscale transport and flow model METRAS for the calculation of grid averaged subgrid scale surface fluxes of momentum, heat and moisture over flat terrain with different land-use categories. Two-dimensional simulations including diurnal variations of the surface temperature and humidity are carried out for different horizontal resolutions of the model. Fine and coarse horizontal grid sizes are applied. The differences between the surface fluxes calculated with fine and coarse horizontal grid sizes are small for averages over the simulation domain, if convection is absent. However, for convective conditions the domain averaged surface fluxes are depending on the turbulence parameterization. If convection is inadequatly parameterized, the domain averaged surface fluxes vary with model resolution. In contrast, the application of the so-called "bottom-up/top-down"-parameterization results in mean surface fluxes independent on model resolution.

Anwendung des Konzepts der Blendhöhe zur Berechnung von bodennahen Flüssen in einem mesoskaligen Modell

Zusammenfassung. Das Konzept der sogenannten Blendhöhe zur Berechnung von gittergemittelten Werten der bodennahen subskaligen Impuls-, Wärme- und Feuchteflüsse über heterogenen Böden wird im nichthydrostatischen mesoskaligen Transport- und Strömungsmodell METRAS angewendet. Zweidimensionale Simulationen werden unter Berücksichtigung des Tagesganges der Bodentemperatur und -feuchte für unterschiedliche Horizontalgitterweiten durchgeführt. Es werden sowohl feine als auch grobe Horizontalgitterweiten verwendet. Wenn keine Konvektion auftritt, ergeben sich im Mittel über das Simulationsgebiet nur geringfügige Unterschiede zwischen den mit feinen und mit groben Horizontalgitterweiten berechneten bodennahen Flüssen. Unter konvektiven Bedingungen hängen die über das Modellgebiet gemittelten bodennahen Flüsse von der Turbulenzparameterisierung ab. Ist die Konvektion unzureichend parameterisiert, variieren die gebietsgemittelten bodennahen Flüsse mit der Modellauflösung. Die Anwendung der sogenannten "bottom-up/top-down"-Parameterisierung führt dagegen zu gebietsgemittelten bodennahen Flüssen, die von der Modellauflösung unabhängig sind.

1 Introduction

In heterogeneous terrain, it is necessary to parameterize the effect of subgrid scale surface inhomogeneities on the surface energy and momentum fluxes. Quite often, subgrid scale heterogeneity is ignored in mesoscale models. Instead, a homogeneous surface is assumed and the grid averaged surface fluxes are calculated by application of the surface-layer-similarity theory. Effective roughness lengths may be used to calculate surface fluxes over heterogeneous terrain, with the surface flux of heat depending on mean surface temperatures (e.g. WOOD and MASON 1990, SCHLÜNZEN et al. 1994). However, this parameterization breaks down, if the stratification is stable on average within a grid element, but unstable over small patches of a hot surface type, e.g. over ice with open water spots. In this case, strong upward heat fluxes against the mean temperature gradient may occur. This phenomenon is called *Schmidt's paradox* (LETTAU 1979).

As an alternative to the use of effective roughness lengths, grid averaged surface fluxes can be calculated as averages over subgrid scale surface fluxes. For this purpose, the parameterization of CLAUSSEN (1991) can be applied, in which the subgrid scale fluxes are functions of subgrid scale surface temperatures, humidities and roughness lengths. The subgrid scale surface fluxes are additionally depending on effective roughness lengths and on the blending height. The blending height, a scale height first introduced by WIERINGA (1976), can be defined as that height at which the flow is approximately in equilibrium with the local surface and also independent of horizontal position (MASON 1988). Thus, above that height effects of roughness changes will not be recognizable individually, but an overall stress and velocity profile will exist, representing the roughness of a larger area.

Up to now, the concept of blending height has been used in various models for calculation of surface fluxes (e.g. STÖSSEL and CLAUSSEN 1993, CLAUSSEN 1995, GRÖTZNER et al. 1996). The objective of the present work is to test the concept of blending height as modified by CLAUSSEN (1991), when it is used in a high-resolution mesoscale model.

Two-dimensional simulations of the flow over a flat, but heterogeneous surface are carried out by use of the nonhydrostatic mesoscale transport and flow model METRAS, developed at the Meteorological Institute of the University of Hamburg. Simulations at high spatial resolution and resolved surface heterogeneity are compared with simulations at low spatial resolution and subgrid scale surface characteristics. Using a grid size that resolves surface char-

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acteristics, the influence of changes in surface characteristics on the flow is calculated directly. For large grid sizes, the concept of blending height is applied to calculate surface fluxes. In this case, the surface characteristics are subgrid scale.

The concept of blending height is summarized in Section 2 and the model briefly described in Section 3. Model initialization and results are presented in Sections 4 and 5. Conclusions considering the application of the concept of blending height to mesoscale models are drawn in Section 6.

2 The concept of blending height

The concept of blending height is based on the theory of local advection. Close to the ground, the flow is in equilibrium with the local surface, i.e. it can be described as a stationary and locally homogeneous flow. At sufficiently large altitudes, the flow becomes independent of local surface characteristics and horizontally homogeneous, representing a regional equilibrium over a larger domain. At the blending height l_b, both assumptions are approximately valid. Formally, l_b can be defined by the following equation for the grid averaged surface stress $<\tau>$:

$$<\tau> = \frac{\kappa^2 U^2(l_b)}{\Delta x} \int_{\Delta x} \frac{1}{\left(\ln \frac{l_b}{z_0(x)}\right)^2} dx . \quad (1)$$

κ denotes the von Kármán constant ($\kappa = 0.4$), $U(l_b)$ the mean wind speed at height l_b, and $z_0(x)$ the roughness length depending on the horizontal coordinate x. For simplicity, equation (1) is written here for only one horizontal coordinate (x). The horizontal grid size is denoted by Δx. The grid averaged surface stress $<\tau>$ is defined by:

$$<\tau> = \frac{1}{\Delta x} \int_{\Delta x} \tau(x)\, dx . \quad (2)$$

However, equation (2) can not be solved to calculate $<\tau>$, since only grid scale quantities are know and $\tau(x)$ is unknown in real applications. Therefore, equation (1) is used and l_b has to be estimated.

CLAUSSEN (1991) showed that the blending height can be calculated from the length scale L_x, which represents the average extension of different surface type patches within the grid cell:

$$\frac{l_b}{L_x}\left(\ln \frac{l_b}{z_{0e}}\right) = c_1 \kappa , \quad (3)$$

where $c_1 = 1.75$ has been found by numerical experiments and z_{0e} is the so-called effective roughness length.

When replacing the integral in (1) by a sum over surface patches of fractions f_j and roughness lengths z_{0j}, then the effective roughness length z_{0e} can be defined by (e.g. WIERINGA 1976):

$$\frac{1}{\left(\ln \frac{l_b}{z_{0e}}\right)^2} = \sum_j \frac{f_j}{\left(\ln \frac{l_b}{z_{0j}}\right)^2} \quad (4)$$

To compute l_b and z_{0e}, equations (3) and (4) have to be solved iteratively. Knowing l_b, equation (1) can be solved and the momentum flux $<\tau>$ may be calculated.

The concept of blending height is used with flux aggregation. This means, that the subgrid scale fluxes are summed up within a surface grid cell to obtain grid averaged fluxes:

$$<u_*^2> = \sum_j f_j u_{*j}^2 , \quad (5)$$

$$<\theta_* u_*> = \sum_j f_j \theta_{*j} u_{*j} , \quad (6)$$

$$<q_* u_*> = \sum_j f_j q_{*j} u_{*j} , \quad (7)$$

The subgrid scale fluxes of momentum (ρu^2_{*j}), latent ($\rho L u_{*j} q_{*j}$) and sensible heat ($\rho c_p u_{*j} \theta_{*j}$) are calculated for each land-use type (indicated by the index j) within a surface grid cell seperately. They are depending on wind velocity U, temperature θ and specific humidity q in the surface grid cell:

$$u_{*j} = \sqrt{C_{mj}}\, U , \quad (8)$$

$$\theta_{*j} = \frac{C_{hj}}{\sqrt{C_{mj}}}\left(\theta - \theta_j^s\right), \quad (9)$$

$$q_{*j} = \frac{C_{qj}}{\sqrt{C_{mj}}}\left(q - q_j^s\right). \quad (10)$$

θ_j^s denotes the surface temperature, characteristic for the land-use type j, which is calculated in METRAS from a surface energy budget equation by using the force-restore-method (DEARDORFF 1978). Also, the humidity q_j^s of the subgrid scale surface class j is calculated by use of the force-restore-method.

The transfer coefficients C_{mj}, C_{hj} and C_{qj}, are approximated by (CLAUSSEN 1991)

$$C_{mj} = \left[\frac{\kappa}{\ln \frac{l_b}{z_{0j}} \frac{\ln(z_p/z_{0e})}{\ln(l_b/z_{0e})} - \psi_m\left(\frac{z_p}{L_j}\right)}\right]^2,$$

$$C_{hj} = \frac{\kappa^2}{\left[\ln \frac{l_b}{z_{0j}} \frac{\ln(z_p/z_{0e})}{\ln(l_b/z_{0e})} - \psi_m\left(\frac{z_p}{L_j}\right)\right]\left[\ln \frac{l_b}{z_{0tj}} \frac{\ln(z_p/z_{0et})}{\ln(l_b/z_{0et})} - \psi_h\left(\frac{z_p}{L_j}\right)\right]},$$

$$C_{qj} = \frac{\kappa^2}{\left[\ln \frac{l_b}{z_{0j}} \frac{\ln(z_p/z_{0e})}{\ln(l_b/z_{0e})} - \psi_m\left(\frac{z_p}{L_j}\right)\right]\left[\ln \frac{l_b}{z_{0qj}} \frac{\ln(z_p/z_{0eq})}{\ln(l_b/z_{0eq})} - \psi_h\left(\frac{z_p}{L_j}\right)\right]}.$$

Here, the roughness lengths of the subgrid scale surface types z_{0j}, z_{0tj}, z_{0qj}, and effective roughness lengths z_{0e}, z_{0et}, and z_{0eq} are used. The subscript t stands for the flux of sensible heat, q for the flux of latent heat. The influence of thermal stratification on the transfer coefficients is included by use of stability functions ψ_m, ψ_h and the Obukhov length L_j. The height z_p is introduced to account for the fact that the wind velocity is basically logarithmic in the surface layer and that U in equation (8) is a grid average value of the lowest model layer Δz. z_p is defined by

$$\ln \frac{z_p}{z_{0e}} = \frac{1}{\Delta z} \int_{z_{0e}}^{\Delta z} \ln \frac{z}{z_{0e}} \, dz \tag{11}$$

For further details on the concept of blending height the reader is referred to CLAUSSEN (1991) and for its realization in METRAS to HERRMANN (1994) and SCHLÜNZEN et al. (1994).

3 The mesoscale model METRAS

For the numerical experiment we use a two-dimensional version of mesoscale transport and flow model METRAS. It is baced on the primitive equations in a nonhydrostatic and anelastic mode. Wind, temperature and humidity are derived from prognostic equations, whereas density and pressure are calculated from diagnostic ones (SCHLÜNZEN 1988, 1990; SCHLÜNZEN et al. 1994). The model METRAS has been successfully applied in the mesoscale-γ- and mesoscale-β-range (e.g. NIEMEIER and SCHLÜNZEN 1993, WU and SCHLÜNZEN 1992) to estimate atmospheric pollutant transport in coastal areas (e.g. SCHLÜNZEN and KRELL 1994, SCHLÜNZEN 1994) and to study tracer transports close to sources (e.g. BIGALKE 1992).

Subgrid scale fluxes are calculated in METRAS from first order closure theory, with optional formulations for the exchange coefficients. In the following, only the turbulence parameterizations that are used in this paper are described in more detail.

In the parameterization of DUNST (1982) the exchange coefficient depends on the friction velocity $u_* = \sqrt{\langle u_*^2 \rangle}$ and the local gradient Richardson number Ri:

$$K_M = \kappa u_* z_a \exp\left(-\frac{3\pi}{\sqrt{15}} \frac{z - z_a}{z_i}\right) \times \frac{\sin(\pi z/z_i)}{\sin(\pi z_a/H_n)}$$

$$\times \left[1 - A\left(\left(\frac{\sqrt{15}\, z_i}{2\pi z}\right)^{0.134} Ri - \left(\frac{\sqrt{15}\, z_i}{2\pi z}\right)^{0.268} \frac{Ri^2}{2}\right)\right] \tag{12}$$

For stable stratification we use $z_i = H_n(1-5.1Ri)$ and $A = 2.7 + 9.3Ri$, for unstable stratification we apply $z_i = H_n(1-1.5Ri)$ and $A = 0.79 + 0.4Ri$. The formulations contain the empirical constants $z_a = 10$m and $H_n = 1000$ m. The parameterization by DUNST works well for stable, neutral and slightly unstable stratification. It will be used for these conditions throughout this study. With the empirical constants used here, it is insufficient for strong convective conditions (LÜPKES and SCHLÜNZEN 1996). Thus, for convective conditions, we also test the "bottom-up/top-down" parameterization used by RAASCH (1988).

The "bottom-up/top-down" parameterization of RAASCH (1988) is a nonlocal closure approach, based on WYNGAARD and BROST (1984). The exchange coefficient for the vertical flux of sensible heat is as

$$K = \min(K^b, K^t), \tag{13}$$

where K^b and K^t are the exchange coefficients for turbulent fluxes from bottom to the top of the convective boundary layer and vice versa. The turbulent fluxes at the top of the convective boundary layer are caused by entrainment of warmer air from outside of the boundary layer due to convective motions, overshooting the inversion height z_i. They are denoted as "top-down" by WYNGAARD and BROST (1984), in contrast to "bottom-up" for the surface fluxes. Assuming a linear profile of the flux of sensible heat, RAASCH calculates K^b and K^t from

$$K^b = \frac{w_* z_i}{g_b}\left(1 - \frac{z}{z_i}\right),$$

$$K^t = \frac{w_* z}{g_t}.$$

w_* is calculated following DEARDORFF (1970):

$$w_* = \left(-\frac{g}{\theta} z_i u_* \theta_*\right)^{\frac{1}{3}},$$

with $g = 9.81$ m s^{-2}. The dimensionless gradient functions g_b and g_t are derived from numerical experiments and are expressed as:

$$g_b = 0.4\left(\frac{z}{z_i}\right)^{-\frac{3}{2}},$$

$$g_t = 0.7\left(1 - \frac{z}{z_i}\right)^{-2}.$$

RAASCH's exchange coefficient results in an intensive vertical mixing in case of convection.

4 Initialization

Seven surface types (water, mudflats, sand, gras, bushes, mixed forest, urban areas) are distributed randomly on stripes of 1 km width within a simulation domain of 48 km. Simulations of diurnal cycles are carried out for three horizontal grid sizes: 1 km, 2 km, 4 km. The according time steps of the model are calculated during the run and are in the order of 30 s for the 1 km grid size and 60 s for the 4 km grid size. For the grid size of 1 km the surface characteristics are resolved by the grid. Therefore, the concept of blending height has no effect in this case. For the other grid sizes (2 km and 4 km) the surface characteristics are not resolved, and their influence on the surface fluxes is included in the model by application of the concept of blending height. Additional to the above, also grid sizes of 250 m and 16 km are applied in some simulations. With the used land-use characteristics, the blending height calculated from equations (3) and (4) is in the range between 85 m and 110 m.

The vertical grid size is 20 m in the lowest 100 m, increasing continiously with height by 20% of the vertical depth of the layer below (maximum grid size 924 m). The top of the model domain is at a height of 6500 m. The initial potential temperature is 288 K at the surface, increasing by 0.2 K/100 m with height. Thus, stable stratification is assumed. The initial humidity is 73% at the surface, decreasing to 30% at model top with a value of 70% at a height of 1 km.

All case studies are performed for two large scale wind velocities. The strong wind case is initialized with a geostrophic wind of 10 m s^{-1}. For the weak wind case, a geostrophic wind of 2 m s^{-1} is prescribed. The wind direction within the boundary layer is almost perpendicular to the stripes. All simulations are carried out for a cloudless day at the end of June. A simulation period of three days is chosen to focus on a fully developed boundary layer.

5 Model results

In the following, only results of the third day of the simulations are interpreted. Due to intense solar heating, the temperature in the boundary layer increases during the simulation period and the vertical profiles differ slightly for the various model resolutions at the beginning of the third day.

The heating during the day results in convective up- and downward motions from noon to the early evening. The vertical wind velocity is shown as an example for the 1 km grid size and the weak wind case, modelled by application of DUNST's parameterization (Fig. 1) and of RAASCH's

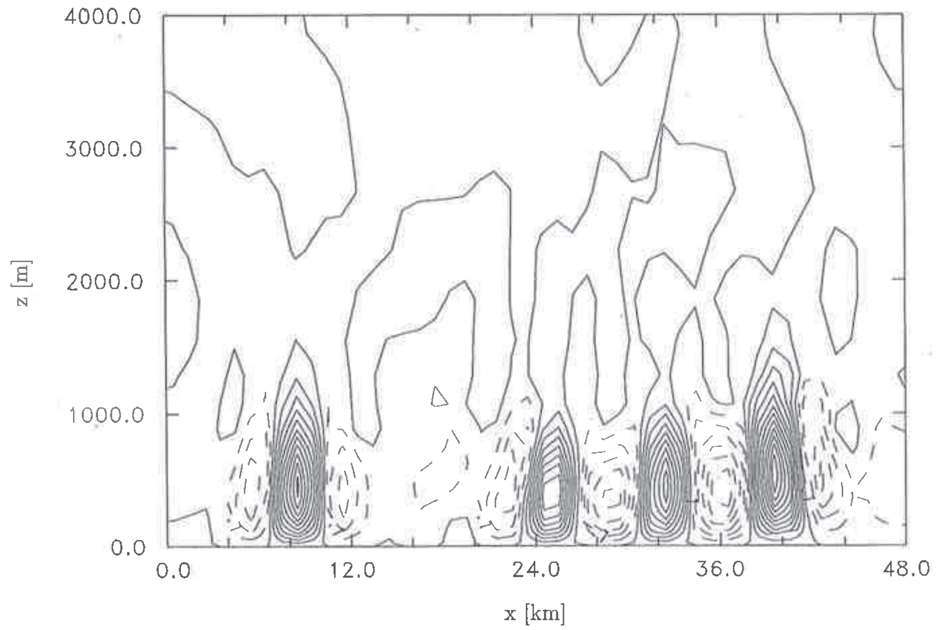

Fig. 1. Vertical wind velocity at 2:00 p.m. of the third day in the case of weak wind (2 ms^{-1}) by use of DUNST's parameterization. The grid size is 1 km. Values range from −1.0 to 1.2 ms^{-1} with an isoline increment of 0.1 ms^{-1}.

Abb. 1. Vertikalwindgeschwindigkeit um 14:00 Uhr des dritten Tages im Fall schwachen Windes (2 ms^{-1}) unter Benutzung von DUNSTs Parameterisierung. Die Gitterweite beträgt 1 km. Die Werte liegen zwischen −1,0 und 1,2 ms^{-1}, mit einem Isolinienabstand von 0,1 ms^{-1}.

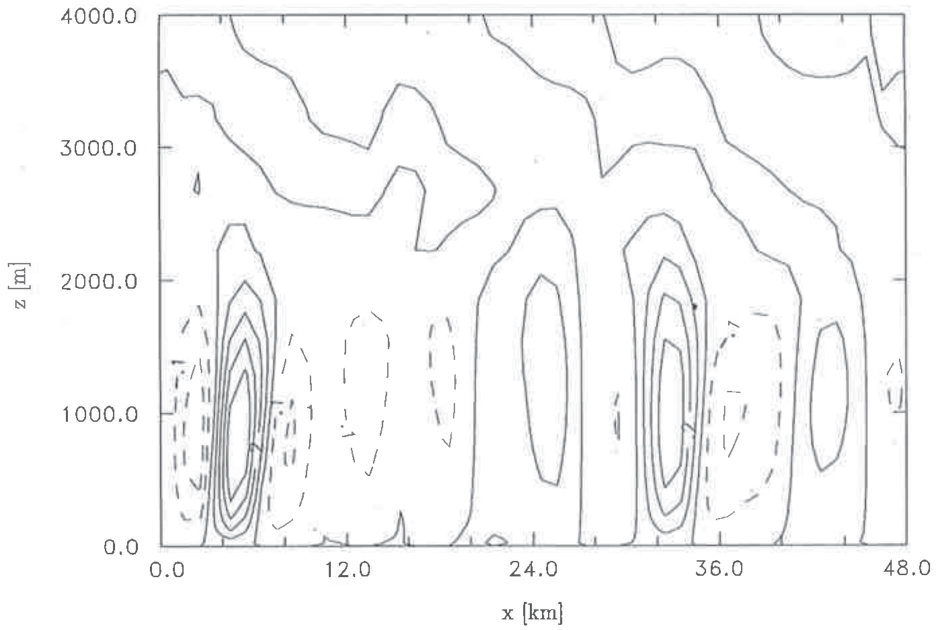

Fig. 2. Same as Fig. 1, for RAASCH's parameterization. Values range from −0.2 to 0.5 ms^{-1} with an isoline increment of 0.1 ms^{-1}.

Abb. 2. Wie Abb. 1, für RAASCHs Parameterisierung. Die Werte liegen zwischen −0,2 und 0,5 ms^{-1}, mit einem Isolinienabstand von 0,1 ms^{-1}.

parameterization (Fig. 2). For all model runs 5 min mean surface fluxes of sensible heat, averaged over the entire model domain, are calculated. The results are shown in Fig. 3 to 5.

Using DUNST's K_M parameterization, differences between the domain averaged surface fluxes of sensible heat are below 50 W m^{-2}. This is true for the strong wind case (Fig. 3) and for the weak wind case (Fig. 4). The differences in the early afternoon are caused by convective motions that are not well parameterized by DUNST's parameterization (LÜPKES and SCHLÜNZEN 1996). Due to the insufficient subgrid scale mixing, up- and downwind areas develop in the model and are advected with the mean wind. The typical height of the simulated structures is 1.5 km and the typical diameter of the upwind cells variies between 6 km for a grid size of 2 km and 1 km for a grid size of 250 m. The maximum vertical wind within the upwind cells occurs between 1:00 p.m. and 3:00 p.m., with values between 0.7 m s^{-1} for a grid size of 2 km and 4 m s^{-1} for a grid size of 250 m. If these cells were realistic, their aspect ratio should not depend on the grid size, which is the case here. We have to conclude from this and from other case studies, that the parameterization of DUNST (1982) is insufficient for strong convective conditions, if the grid is coarse and subgrid scale convection has to be parameterized. However, the parameterization works well for slightly unstable, neutral and stable stratification.

Using the "bottom-up/top-down" turbulence parameterization as applied by RAASCH (1988), the subgrid scale convection is parameterized: There is little change of vertical winds with model resolution. The vertical winds are below 0.5 m s^{-1}. This is true for all used grid sizes. Convection is mainly treated as a subgrid scale phenomenon, intense up- and downdrafts do not occur. The differences between the domain averaged surface fluxes of sensible heat are below 15 W m^{-2} (13%), even for the weak wind case (Fig. 5).

In none of the simulations, the flow did show a clear transition from heterogeneity to homogeneity with increasing height — an assumption implicitly included in the

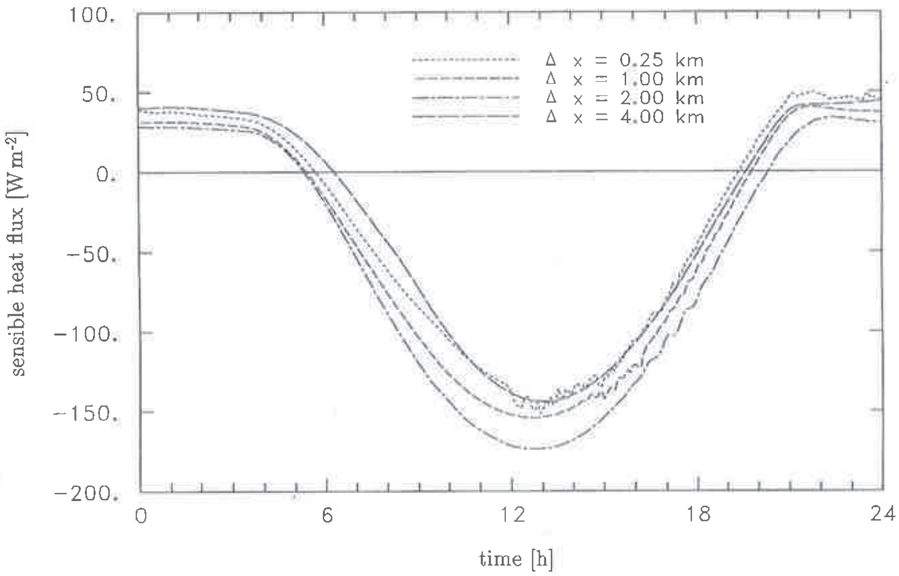

Fig. 3. Mean surface fluxes of sensible heat, calculated by use of DUNST's parameterization for the strong wind case (10 ms^{-1}).

Abb. 3. Mittlere Bodenflüsse fühlbarer Wärme, berechnet mit DUNSTs Parameterisierung im Fall starken Windes (10 ms^{-1}).

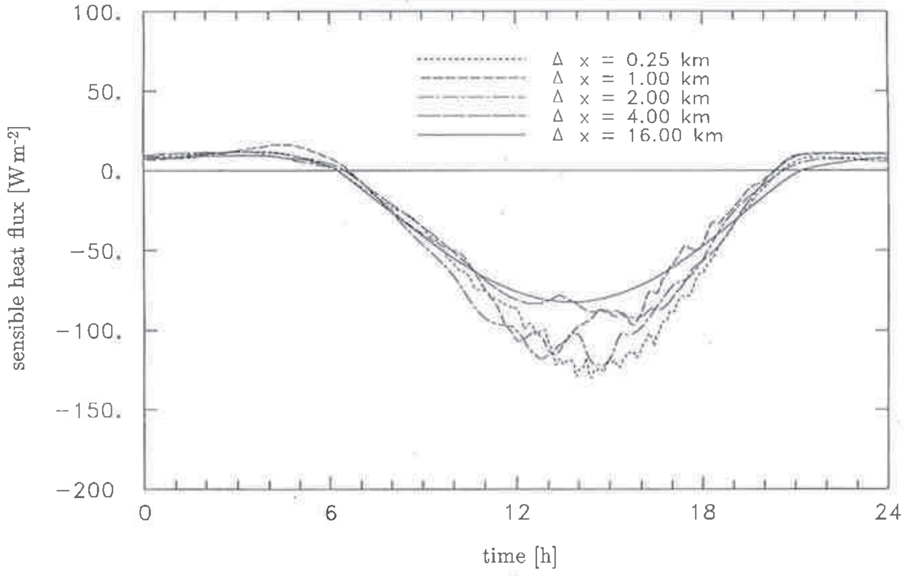

Fig. 4. Mean surface fluxes of sensible heat, calculated by use of DUNST's parameterization for the weak wind case (2 ms^{-1}).

Abb. 4. Mittlere Bodenflüsse fühlbarer Wärme, berechnet mit DUNSTs Parameterisierung im Fall schwachen Windes (2 ms^{-1}).

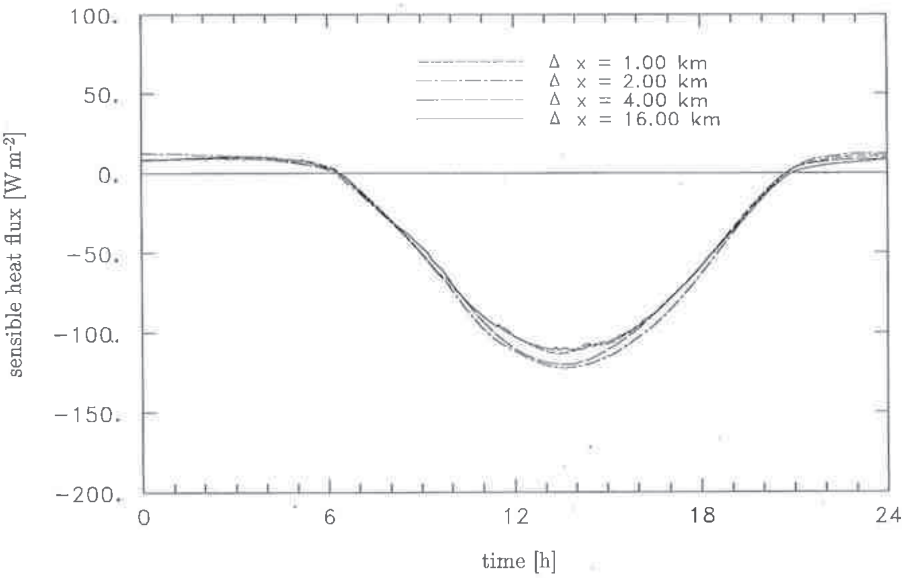

Fig. 5. Mean surface fluxes of sensible heat, calculated by use of RAASCH's parameterization for the weak wind case (2 ms^{-1}).

Abb. 5. Mittlere Bodenflüsse fühlbarer Wärme, berechnet mit RAASCHs Parameterisierung im Fall schwachen Windes (2 ms^{-1}).

concept of blending height. With application of Dunst's parameterization, intense vertical wind velocities are present in the whole convective boundary layer. Also the standard deviation of the domain averaged horizontal wind, a measure for the homogeneity of the whole flow, exhibits no significant decrease with increasing height (not shown here). The latter also holds for the diffusion coefficient of RAASCH (1988). The assumption of horizontal homogeneity of the flow at sufficiently large altitudes (see Section 2) is questionable for the studied cases with temporary convection. Nevertheless, the concept of blending height yields reasonable estimates of surface fluxes although its range of validity is exceeded.

6 Discussion and conclusions

Application of the concept of blending height together with RAASCH's "bottom-up/top-down" turbulence parameterization yields area averaged surface fluxes which are almost independent of model resolution. The deviation of the domain averaged surface fluxes of sensible heat remains below 13% for the used grid sizes.

This result is to be expected for an adequate parameterization of mean surface fluxes. After all, the landscape and the meteorological situation remains the same in the model, regardless of the model resolution. The landscape just changes from resolved to subgrid scale.

When using the concept of blending height together with DUNST's turbulence parameterization, the mean surface fluxes vary with model resolution. This discrepancy is attributed to DUNST's parameterization, which seems to be inadequate for strong convective conditions.

No blending height can be determined from vertical profiles of the standard deviation of the mean horizontal velocity due to the absence of a significant transition from a heterogeneous flow to a homogeneous flow with increasing altitude. Using the temperature based definition of the blending height of WIERINGA (1976), we find a blending height slightly higher than that of KALTHOFF et al. (1993). They have found a blending of temperature profiles in their model results, which have been performed for the heterogeneous terrain of the Hildesheimer Börde in the month of June. Compared to our study, the stratification of the atmosphere was more stable and the characteristics of the surface were more homogeneous. Therefore, convection was weaker during the experiment, which explains the different blending heights estimated from horizontal temperature variations. Using a wind-based definition of the blending height, the absence of intense convection might be an important requirement for its estimation.

The question arises whether the actual value of the blending height is of importance for our model results or whether flux aggregation without determination of a blending height would be sufficient for the calculation of surface fluxes in our simulations. To answer this question, the influence of different values of the blending height on the friction velocity is studied. The blending height is calculated from equations (3) and (4) and has values between 85 and 110 m in the model runs. A value of 100 m is used here for the calculation of the friction velocity from equations (5) and (8). They are solved for neutral stratification and with values for the roughness lengths used in the model simulations. As an alternative to the value of 100 m, which is characteristic for the model simulations, we use the height of the lowest grid point ($z_p = 10$ m) as value for the blending height l_b. The comparison of the two friction velocities yields, that the friction velocity, which is calculated with $l_b = z_p$, is about 6% larger than the friction velocity, which is calculated by using the model simulated value of the blending height. Thus in the cases studied here, the precise value of the blending height is of lower importance. Other parameterizations of the model, e.g. the exchange coefficients, can result in larger uncertainties.

We conclude that the range of atmospheric situations in which a blending height can be found from wind profiles and the relevance of the actual value of the blending height for the calculation of surface fluxes has to be further investigated. Special attention should be paid on secondary circulations like convection. Also it should be investigated, whether the concept of blending height gives realistic results, if a sufficient turbulence parameterization is applied and strong up- and downdrafts are developing, i.e. for large eddy simulations. For this purpose, high-resolving numerical experiments and field experiments should be performed.

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