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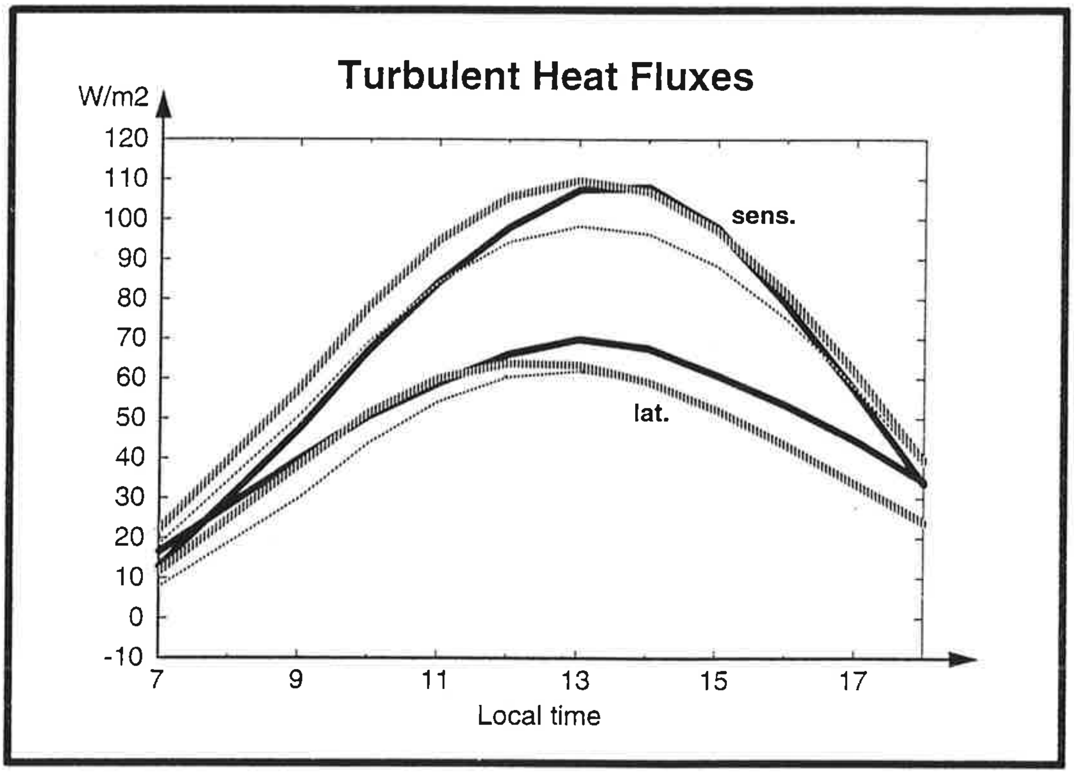

FLUX AGGREGATION AT LARGE SCALES: ON THE LIMITS OF VALIDITY OF THE CONCEPT OF BLENDING HEIGHT

von

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Flux aggregation at large scales: On the limits of validity of the concept of blending height*

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Abstract

The concept of blending height is known to be applicable to estimate areally-averaged surface heat and momentum fluxes over heterogeneous terrain with horizontal scales of surface variations much smaller than 10 km. Here, the performance of this concept is explored beyond this limit of validity. This is accomplished by analysing a three-dimensional meso-scale simulation of land-sea breeze systems which originate at large-scale temperature differences between water and land surfaces and by employing a one-dimensional version of the meso-scale model as a vertical column of a hypothesized macro-scale model.

It appears that areally-averaged surface fluxes can be reproduced reasonably well. This is valid for area-averaged fluxes obtained by the average of surface fluxes on each land type, i.e. for so-called flux aggregation, and for a combination of flux aggregation and so-called parameter aggregation where similar land types are combined into an aggregated land surface. The weak dependence of averaged surface fluxes on secondary, meso-scale motions agrees with earlier theoretical considerations.

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1. Introduction

In numerical models of atmospheric flow it is necessary to consider the properties of boundary-layer flow as averaged over the grid size of the model. Recently, the so-called concept of blending height has become a useful approach to the parameterization of areally-averaged surface fluxes over heterogeneous terrain (e.g. Wieringa, 1986, Mason, 1988, Claussen, 1990, 1991). Implicit in this concept is the assumption that regional momentum and energy fluxes should be estimated at the blending height, which, according to Mason (1988) is defined as a scale-height at which the flow changes from equilibrium with the local surface to independence of horizontal position. Far above the blending height, modifications of air flow due to changes in surface conditions will not be recognizable individually, but an overall stress or heat flux profile will exist, representing the surface conditions of a large area.

The concept of blending height is derived from the theory of local advection in the atmospheric surface layer. Hence the concept of blending height should be applicable to variations in surface conditions at scales considerably smaller than 10 km, i.e. for so-called disorganized or type A landscapes (see Shuttleworth, 1988). At these scales, the concept of blending height has been tested by micro-scale models (e.g. Mason 1988, Claussen 1991).

In flow over terrain inhomogeneities at scales larger than 10 km, i.e. over so-called organized or type B landscapes, blending takes place essentially above the surface layer where Coriolis effects must not be ignored. Moreover, in type B landscapes, secondary circulations may develop which mix momentum and energy throughout the planetary boundary layer efficiently and which presumably affect surface fluxes.

Here, the limits of validity of the concept of blending height are explored by applying this concept to a situation in which it should certainly fail: to heterogeneous terrain from which, under weak synoptic forcing, strong and persistent land-sea breeze systems are triggered. Therefore, areally-averaged surface heat and momentum fluxes are diagnosed from a three-dimensional meso-scale simulation of land-sea breeze systems. A one-dimensional version of the meso-scale model is then employed as a vertical column of a hypothesized macro-scale model into which flux averaging according the concept of blending height is implemented.

Before the numerical experiment is described in Section 3, the concept of blending height is briefly summarized in Section 2. Moreover, it is shown how the concept of blending height is related to so-called *parameter aggregation* and *flux aggregation*.

2. The concept of blending height

In heterogeneous terrain, Wieringa (1986) suggests averaging momentum fluxes at a blending height. He interprets the blending height as a height above which modifications of air flow due to changes in surface conditions will not be recognizable individually, but an overall stress or heat flux profile will exist, representing the surface conditions of a large area. Mason (1988) more explicitly defines the blending height l_b as a scale height at which the flow is approximately in equilibrium with the local surface and also independent of horizontal position. Using the latter definition, the momentum flux $-(\overline{uw})$ on average over a heterogeneous surface is

$$[-(\overline{uw})] = \sum_i f_i(-(\overline{uw})_i) = \kappa^2 U^2(l_b) \sum_i \frac{f_i}{(ln\frac{l_b}{z_0^i})^2} \quad , \tag{2.1}$$

where [...] denote a horizontal average, f_i is the fractional area covered by a patch $'i'$ with the roughness length z_0^i. U is the mean wind speed and κ is the von Kármán constant (here, $\kappa = 0.4$). From Equation 2.1, also an aggregated roughness length z_{0a} can be defined by

$$\frac{1}{(ln\frac{l_b}{z_{0a}})^2} = \sum_i \frac{f_i}{(ln\frac{l_b}{z_0^i})^2} \tag{2.2}$$

Mason (1988) provides a heuristic model which indicates that

$$\frac{l_b}{L_c}\left(ln\frac{l_b}{z_0}\right)^2 \sim 2\kappa^2 \quad , \tag{2.3}$$

where L_c is the horizontal scale of roughness variations, and from Equation 2.3 one can conclude that $l_b/L_c \sim O(10^{-2})$. Claussen (1990) deduces the blending height from numerical simulations of air flow over a surface with randomly varying roughness. He finds that the sum of errors due to the assumptions of horizontal homogeneity and equilibrium with the local surface attains a minimum at a height which is roughly as large as the diffusion height scale l_d

$$\frac{l_d}{L_c}\left(ln\frac{l_d}{z_0}\right) \sim c_1\kappa \qquad , \tag{2.4}$$

where the constant c_1 should be $O(1)$. Claussen (1991) finds $c_1 = 1.75$. Using either estimate of blending height, Equation 2.3 or 2.4, one obtains reasonably accurate estimates of an aggregated roughness length. Differences between estimates are small particularly when considering the inaccuracy in determining L_c (see also last paragraph of the Appendix). From simulations of air flow over randomly varying surface conditions, Claussen (1991) infers that L_c is the length scale at which on average the surface conditions change over a larger fetch.

2.1 Parameter aggregation

Provided that L_c and f_i are known, than the blending height and the aggregated roughness length can be obtained from Equations 2.2 and 2.3 or 2.4. The average momentum flux is finally computed from the aggregated roughness length. The computation of areally-averaged fluxes from aggregated parameters will be called *parameter aggregation* in the following. Formally, an areally-averaged flux $[\Phi]$ is

$$[\Phi] = f(\vec{\psi}_a, ...) \qquad , \tag{2.5a}$$

where the vector of aggregated surface parameters is a function of surface parameters of each land type,

$$\vec{\psi}_a = f(\vec{\psi}_i) \qquad . \tag{2.5b}$$

For example, z_{0a} is given by Equations 2.2, 2.3 or 2.4, but for an aggregated albedo α_a, $\alpha_a = \sum_i f_i\alpha_i$ (see Appendix).

2.2 Flux aggregation

In stratified flow, it has been proposed (e.g. Noilhan and Lacarrère, 1992, Wood and Mason, 1991) to apply the method of parameter aggregation also to estimation of areally averaged heat fluxes, i.e. by defining proper values of aggregated albedo, aggregated leaf area index, or aggregated stomatal resistances. However, parameter aggregation will fail, if surface conditions vary strongly. For example, definition of an aggregated soil heat conductivity is cumbersome in the presence of water and soil. The heat flux into soil is predominantly conductive, whereas the heat flux into water could be influenced by water advection or thermohaline circulation. Likewise, it has been shown (e.g. Claussen 1990, Blyth *et al.* 1993) that an aggregated stomatal resistance is impossible to find if the local resistances vary strongly.

A second complication arises due to the non-linearity of the relationship between turbulent fluxes and vertical mean profiles. For example, the vertical gradient of potential temperature can be positive on average over a larger area, whereas the averaged heat flux is upward, because strong turbulence in small regions of unstable stratification can dominate the averaged heat flux resulting in an averaged heat flux counter to the averaged vertical gradient of potential temperature. This process is quite important in the winterly polar ice zones (e.g. Stössel and Claussen, 1993). In order to circumvent these problems, Claussen (1991) suggests to compute momentum and heat fluxes at the blending height for each land-use type which can be identified in the area under consideration. Consequtively, the averaged surface fluxes are obtained by the average of surface fluxes on each land-use surface weighted by its fractional area f_i. This method is called *flux aggregation* in the following. Formally,

$$[\Phi] = \sum_i f_i \Phi_i \quad , \tag{2.6a}$$

where

$$\Phi_i = f(\vec{\psi}_i, ...) \quad . \tag{2.6b}$$

Fluxes Φ_i also depend on turbulent transfer coefficients which, in turn are functions of some of the components of $\vec{\psi}_i$. The requirement of computing the surface fluxes for each land type at the blending height leads to a revised formulation of turbulent transfer coefficients which differs from the conventional formulation (see Claussen 1991).

3. The experiment

The non-hydrostatic meso-scale model GESIMA (Kapitza and Eppel 1992, Eppel *et al.* 1992) has been used to simulate the land-sea breeze systems over Northern Germany and Southern Danmark. This simulation is described in detail by Jacob (1991). The soil - atmosphere interface of GESIMA is briefly summarized in the Appendix.

The landscape of Northern Germany and Southern Danmark forms a penninsula, approximately 80 kilometers wide, with the North Sea and the Baltic Sea to the West and to the East, respectively. Along the coast of the North Sea, there are numerous islands; moreover, zones with extended tidal flats are found which reach up to fifteen kilometers out into the sea.

The simulation analysed here is this of a summer day (June 23rd) where a weak southerly (i.e. parallel to the penninsula) large-scale wind prevails. Hence land-sea breeze systems develop during the morning between 10 and 11 local time at each coast and at larger islands. During the afternoon, some systems merge into a larger system over the penninsula.

The three-dimensional meso-scale simulation has been initialized by letting one-dimensional vertical profiles of temperature and velocity adjust to homogeneous terrain. Using this one-dimensional profile, the three-dimensional simulation was started. It turned out that it takes only a few hours for the flow to adapt to heterogenous terrain. Hence model results of the first six hours of the day (i.e. the adaptation phase and a little bit more) will not be presented.

Using the one-dimensional version of GESIMA, the land-sea breeze simulation has been recomputed. The one-dimensional model has been initialized with the same vertical profiles of temperature and velocity as used for the three-dimensional simulation. After the initialization phase, the soil - atmosphere interface was changed to take into account the flux aggregation process. Hence momentum and energy fluxes Φ_i were computed at the blending height for each land type. Then, the fluxes were averaged weighted by the fractional area f_i of each land type to arrive at an areally-averaged surface flux.

In Figure 1, the surface fluxes of sensible and latent heat are shown. The full lines indicate the areally-averaged heat fluxes obtained by the three-dimensional model. The hatched lines depict the surface heat fluxes [Φ] of the one-dimensional model.

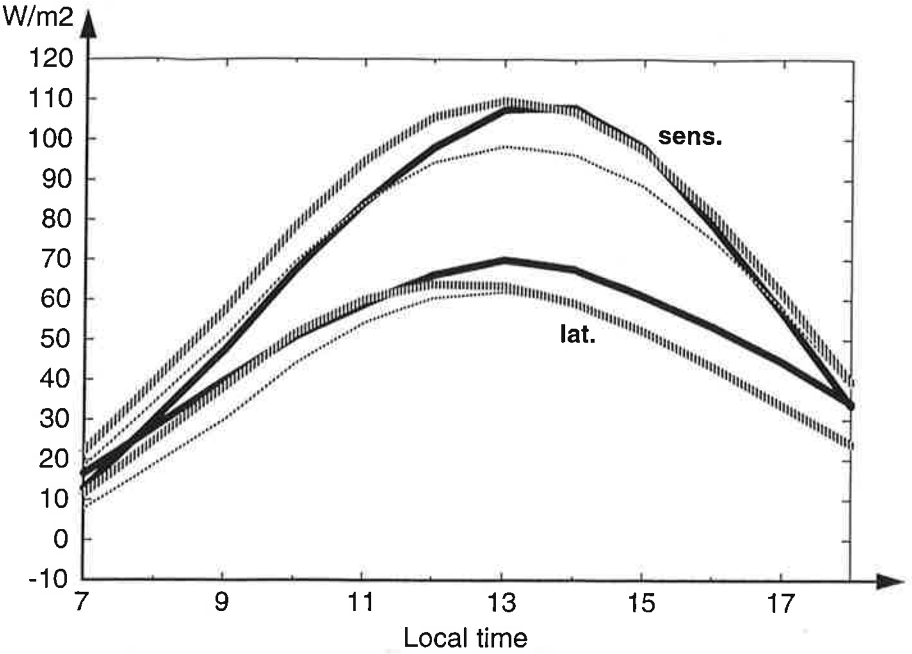

Fig.1: *Sensible and latent heat flux as function of local time. Sensible heat fluxes are labeled 'sens.', and latent heat fluxes, 'lat.'. Full lines indicate areally-averaged heat fluxes from the three-dimensional meso-scale model. Hatched lines are heat fluxes from the one-dimensional model using flux aggregation, and dotted lines, using a combination of flux aggregation and parameter aggregation.*

The dotted lines represent the surface heat fluxes obtained by a combination of flux aggregation and parameter aggregation. From Table 1 in the Appendix, it is obvious that there are three land types which strongly differ by their parameters: land, tidal flats, and water. For water, an infinitely large heat capacity is assumed, i.e. the water temperature is taken constant. Tidal flats and the other land types differ in their water field capacity and capillarity. It is assumed that tidal flats are always completely wet due to frequent inundation (e.g. Claussen 1988), whereas for other land types low (perhaps somewhat unrealisticly low) values of capillarity are assumed to obtain very dry land surfaces. Hence for all land types, except tidal flats, parameters have been aggregated. (The aggregated parameters are listed in Table 1 in the Appendix.) The surface heat and momentum fluxes over the 'aggregated land surface' are evaluated from the vector of aggregated parameters. Here,

$$\Phi_l = f(\vec{\psi_a}, ...)$$

with

$$\vec{\psi_a} = f(\vec{\psi_i}) \qquad i = 2, ..., 9, \quad i \neq 7 \qquad .$$

The areally-averaged surface fluxes for the entire flow domain are obtained using flux aggregation:

$$[\Phi] = f_w \Phi_w + f_t \Phi_t + f_l \Phi_l \qquad ,$$

where the indices w,t, and l stand for water, tidal flats, and aggregated land surface, respectively.

It is quite obvious from Figure 1 that both, flux aggregation and the combination of flux aggregation and parameter aggregation yield reasonable results. Differences between peak values are within approximately $\pm 10\%$.

The friction velocity has been computed from the three-dimensional model in two different ways. Firstly, the cube of local friction velocities are averaged to obtain $[u_*] = ((\int_{X,Y} u_*^3 \, dxdy)/(XY))^{1/3}$ where X and Y denote the horizontal extent of the flow domain - shown as thin line in Figure 2. Secondly, a friction velocity is derived from the vector average of shear stress - thick line in Figure 2. While the latter represents the momentum loss on average over the flow domain, the former is a measure of dissipation of kinetic energy within the surface-layer. In homogeneous terrain, both friction velocities are the same. Here, they differ due to the onset of land-sea breeze systems, i.e. due to the onset of subgrid-scale motions. Obviously, the aggregation concept (flux aggregation or the combination of flux aggregation and parameter aggregation give almost the same friction velocities) is capable of estimating the areally-averaged momentum loss, but it yields a poor estimate of kinetic energy dissipation.

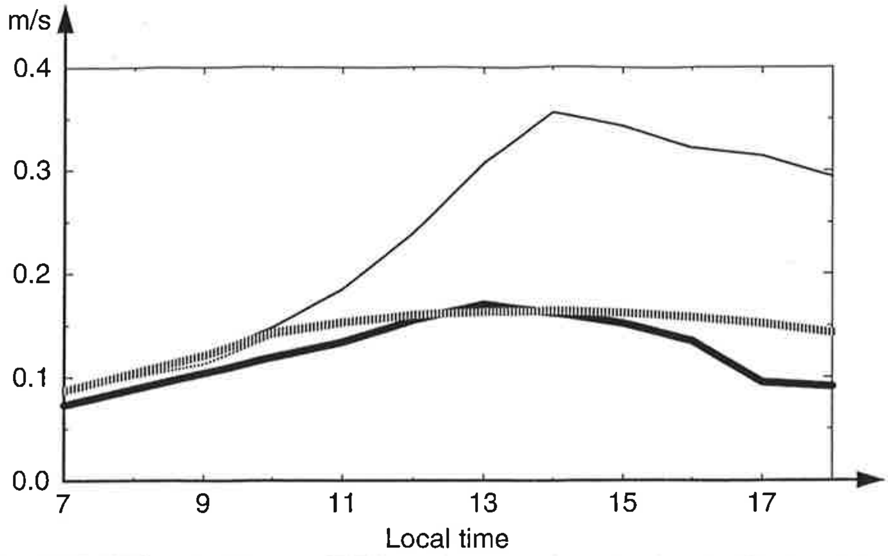

Fig.2: *Friction velocity as function of local time. The full line indicate the areally-averaged momentum flux from the three-dimensional meso-scale model, the thin line, the areally-averaged friction velocity (see text). The hatched line is the friction velocity from the one-dimensional model using flux aggregation, and the dotted line, using a combination of flux aggregation and parameter aggregation. (The dotted line is mostly hidden under the hatched line.)*

Although the areally-averaged surface fluxes of the three-dimensional model are reasonably well represented by the one-dimensional model, the one-dimensional model fails to reproduce the boundary-layer structure over the heterogenous terrain. It yields too cold and too shallow a boundary layer. The average temperatures differ by 3 K, and the height of the temperature inversion is located at 1300 m instead of 1750 m as in the three-dimensional model. These differences can be blamed on the effective vertical mixing due to land-sea breeze systems, an effect not represented in the one-dimensional model. However, this does not seem to be related to the problem of estimating areally-averaged surface fluxes and, hence, will not be analysed here.

4. Discussion and Conclusion

The performance of the concept of blending height is explored beyond its limit of validity. I.e., this concept has been applied to estimate areally-averaged surface fluxes over heterogeneous terrain with surface inhomogeneities at scales considerably larger than 10km. Specifically, a three-dimensional meso-scale simulation of land-sea breeze systems which originate at large-scale temperature differences between water and land surfaces has been analysed, and a one-dimensional version of the meso-scale model has been employed as a vertical column of a hypothesized macro-scale model. It appears that areally-averaged surface fluxes can be reproduced reasonably well by the one-dimensional model.

It is hard to answer the question why the agreement between three-dimensional and one-dimensional simulations is rather good or, alternatively, why the limits of validity of the concept of blending height can easily be stretched. One explanation could be that the surface heat fluxes are basically determined by the net available energy at the earth surface. The net available energy, in turn, is mostly determined by the net incoming solar radiation and atmospheric radiation which are rather homogeneously distributed in cases of cloudless sky or randomly scattered clouds - as it is the case for the simulation analysed here. Since radiative fluxes depend linearly on albedo and emissivity, the average short-wave and long-wave radiative input is simply proportional to the weighted sum of albedo and emissivity, respectively.

Secondly, it seems that land-sea breeze systems, i.e. coherent motions, hardly affect surface fluxes. This, however, agrees with theoretical investigations of the convective boundary layer (CBL) over wavy terrain with variable heat flux by Schumann (1991). Schumann finds that the mixing from the surface into the CBL is controlled by small-scale turbulence which is much less affected by coherent motions. In particular, turbulent diffusivity at the surface appears to be approximately independent of coherent motion. This is not valid if friction velocity (or the cube of friction velocity which is a measure of surface-layer dissipation of kinetic energy) is simply averaged. The (scalar) friction velocity increases strongly with the onset of coherent motions.

Schumann (1991) also states that the terrain inhomogeneities have a strong impact on vertical mixing within the mixed layer of the CBL. This is also supported by this study.

However, it is not of direct relevance to the present study and, therefore, it has not yet been analysed in detail.

The concept of blending height can be applied to flux aggregation and to parameter aggregation. For the former, the averaged surface fluxes are estimated from surface fluxes which are computed for each land type at the blending height, for the latter, from a set of aggregated parameters which could simply be the weighted sum of parameters for each land type - as for aggregated albedo - or could be computed from a seperate model - as for aggregated roughness. It has been argued earlier that parameter aggregation is feasible only if surface conditions do not vary drastically. Hence a combination of parameter aggregation and flux aggregation has been undertaken here. Parameter aggregation is done only for land surfaces which appear similar, and flux aggregation is done for water, tidal flats, and 'aggregated land'. From the model results it is obvious that the computationly cheaper combination of flux aggregation and parameter aggregation yields as fair a representation of areally-averaged surface fluxes as the more expensive flux aggregation.

This study has been undertaken to get some idea on the performance of flux aggregation over terrain with larger-scale variations of surface conditions using the concept of blending height. A somewhat extreme situation in which this concept was expected to fail has been chosen for analysis to outline and to address the problem. Since this is a study of a single case, more systematic studies should follow up. Perhaps two-dimensional meso-scale simulations would suffice as three-dimensional simulations always are quite elaborate.

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Appendix

At the interface between atmosphere and vegetation, it is assumed that no energy and no mass is stored. Hence for the energy fluxes,

$$Q_{rad} + Q_{lat} + Q_{sens} = Q_{ground} \qquad (A.1)$$

The fluxes are considered positive when directed upward. Q_{rad} is the sum of the energy flux densities (dimension: W/m^2) due to short-wave and long-wave radiation:

$$Q_{rad} = -S_o(1-\alpha) - L_a + \epsilon\sigma T_G^4 \qquad (A.2)$$

where α and ϵ are the albedo and emissivity of the surface. The global radiation is indicated by S_o and the atmospheric radiation by L_a. S_o and L_a are parameterized according to Kasten and Czeplak (1980). T_G is the temperature at the air/ground interface.

Q_{lat} and Q_{sens} stand for energy flux densities due to turbulent transports of latent and sensible heat:

$$Q_{sens} = \rho c_p \, C_h \, U_a \, (\Theta_G - \Theta_a) \qquad (A.3)$$

$$Q_{lat} = \rho l_v \, w_e \, C_q \, U_a \, (Q_s(T_G) - Q_a) \qquad (A.4)$$

where ρ is the density of the air, U_a, Θ_a, Q_a are the mean velocity, potential temperature, and specific humidity on the average over the surface-layer grid box, c_p is the specific heat capacity of the air at constant pressure, and l_v is the specific heat of vaporization.

The transfer coefficients C_h, C_q in the Equations A.3, A.4 are evaluated by using conventional boundary-layer similarity theory as in Louis (1979) with the exception that the ratio of local roughness lengths is assumed to be a constant value of $z_0/z_{0t,q} = 10$ over densely vegetated areas and to vary with the friction Reynolds number over terrain with bluff roughness elements (e.g. Brutsaert 1975).

The relative soil moisture w_e is estimated by using a force-restore method (see Deardorff 1978). Here

$$\frac{\partial w_e}{\partial t} = -\frac{E_g - P}{w_k \rho_w} + \frac{\alpha_c}{\rho_w}(1 - w_e) \quad , \qquad (A.5)$$

where E_g and P are evaporation and precipitation, respectively, i.e. the 'forcing' of soil water. ρ_w is the densitiy of water, α_c is the so-called capillaritiy factor which accounts for restoring of water from a ground table. w_k is the critical depth of extracted liquid water that a 0.1 m upper layer of soil is capable of holding before the surface is saturated.

The conductive heat flux into the soil is

$$Q_{ground} = -\lambda \left(\frac{\partial T}{\partial z} \right)_{z=0} \tag{A.6}$$

where λ is the heat conductivity and T the actual temperature of the soil. The soil temperature is computed from the lineare diffusion equation

$$\frac{\partial T}{\partial t} = k \frac{\partial^2 T}{\partial z^2} \quad . \tag{A.7}$$

with k as temperature diffusivity. The surface parameters, used for this study, are listed in Table 1.

Parameter aggregation

The aggregation of parameters is done in the following way. Albedo, emissivity, volumetric soil heat capacity, and field capacity are simply averaged, because these are intensive thermodynamic properties. As a first approximation, the same is done for capillarity. The aggregated soil heat conductivity k_a is obtained by averaging thermal inertia:

$$k_a = \left(\frac{\sum_i f_i c_{s,i} k_i^{1/2}}{\sum_i f_i c_{s,i}} \right)^2 \quad .$$

The rational for this is that the soil heat flux is a linear function of thermal inertia. Hence, for the same thermodynamic forcing, the average soil heat flux is directly proportional to the average thermal inertia.

The aggregated roughness length is evaluated according to Equations 2.2, 2.4. The horizontal scale of surface variation is not easily defined because there is a spectrum of scales. But that does not pose a severe problem. Varying L_c between 1 km to 100 km changes z_{0a} only from 0.26 m to 0.243 m. Here, a value of $z_{0a} = 0.25$m is used as a first, but reasonable guess. Using Mason's (1988) approach, an aggregated roughness length of 0.27m is obtained which would have increased the neutral drag coefficient by approximately 3%.

Table 1:

Allocation of surface parameters to land types. The land types are 1: water and open sea; 2: forests; 3: marshy land; 4: irrigated farmland, moors, sandy soil; 5: hilly terrain with many windbreaks, loamy soil; 6: moderately hilly terrain, sandy soil; 7: tidal flats; 8: heath; 9: urban areas; 10: aggregated land surface. The surface parameters are f_i: fractional cover by land type i; k: soil temperature diffusivity [10^{-6} m^2/s]; c_v: soil heat capacity [10^6 J/m^3/K]; ϵ: emissivity; α: albedo; z_0: roughness length [m]; w_k: field capacity [m]; α_c: capillarity [kg/m^3/s]. $*$ for water, z_0 is computed from the Charnock's formula.

	f_i	k	c_v	ϵ	α	z_0	w_k	α_c
1	0.34	-	-	0.95	0.10	*	-	-
2	0.08	0.70	2.5	0.95	0.15	0.75	0.010	0.008
3	0.10	0.56	2.1	0.95	0.25	0.07	0.010	0.008
4	0.09	0.74	2.9	0.95	0.25	0.17	0.005	0.003
5	0.11	0.73	2.1	0.95	0.25	0.20	0.004	0.002
6	0.15	0.84	2.1	0.95	0.25	0.17	0.003	0.001
7	0.04	0.51	3.5	0.91	0.12	0.0004	-	-
8	0.03	0.70	2.5	0.95	0.15	0.35	0.003	0.001
9	0.06	1.00	2.0	0.90	0.20	0.80	0.003	0.003
10	-	0.75	2.3	0.95	0.23	0.25	0.006	0.004

References

Blyth, E.M., Dolman, A.J., and Wood, N., 1993: Effective resistance to sensible and latent heat flux in heterogeneous terrain. *Quart.J.R.Met.Soc.*, **119**, 423-442.

Brutsaert, W., 1975: A theory for local evaporation (or heat transfer) from rough and smooth surfaces at ground level. *Water Resources Res.*, **11**, 543-550.

Claussen, M., 1988: On the surface energy budget of coastal zones with tidal flats. *Beitr. Phys. Atmosph.*, **61**, 39-49.

Claussen, M., 1990: Area-averaging of surface fluxes in a a neutrally stratified, horizontally inhomogeneous atmospheric boundary layer. *Atmos. Environ.*, **24a**, 1349-1360.

Claussen, M., 1991: Estimation of areally-averaged surface fluxes. *Boundary-Layer Meteorol.*, **54**, 387-410

Deardorff, J.W. 1978: Efficient prediction of ground surface temperature and moisture with inclusion of a layer of vegetation. *Jour. Geophys. Res.* **83**, 1889-1903

Eppel D.P., Kapitza, H., Claussen, M., Jacob, D., Koch, W., Levkov, L., Mengelkamp, H.-T., Werrmann, N., 1993: The non-hydrostatic meso-scale model GESIMA. Part II: Parameterizations and applications. *Beitr. Phys. Atmosph.*, in press.

Jacob, D., 1991: Numerische Simulation der Wolkenbildung in einer Land-Seewind-Zirkulation. *Report 91/E/40* GKSS-Forschungszentrum Geesthacht, Max-Planck-Str., D-2054 Geesthacht. 106p.

Kapitza, H. and Eppel D.P., 1992: The non-hydrostatic meso-scale model GESIMA. Part I: Dynamical equations and tests. *Beitr. Phys. Atmosph.*, **65**, 129-146.

Kasten, F. and G. Czeplak 1980: Solar and terretrial radiation dependent on the amount and type of cloud. *Solar Energy* **24**, 177-189

Louis, J.F., 1979: A parametric model of vertical eddy fluxes in the atmosphere. *Boundary-Layer Meteorol.*, **17**, 187-202.

Mason, P.J., 1988: The formation of areally-averaged roughness lengths. *Quart. J. R. Met. Soc.*, **114**, 399-420.

Noilhan, J., and Lacarrére, P., 1992: GCM gridscale evaporation from mesoscale modelling. in: Proceedings of a workshop held at ECMWF on fine-scale modelling and the development of parameterization schemes. 16.-18.9.1991. 245-274.

Schumann, U., 1991: A simple model of the convective boundary layer over wavy terrain with variable heat flux. *Beitr. Phys. Atmosph.*, **64**, 169-184.

Shuttleworth, J.W., 1988: Macrohydrology - The new challenge for process hydrology. *J. Hydrology*, **100**, 31-56.

Stössel, A. and Claussen, M., 1993: A new atmospheric surface-layer scheme for a large-scale sea-ice model. *Climate Dynamics*, in press.

Wieringa, J., 1986: Roughness-dependent geographical interpolation of surface wind speed averages. *Quart. J. R. Met. Soc.*, **112**, 867-889.

Wood, N. and Mason, P.J., 1991: The influence of stability on effective roughness lengths for momentum and heat transfer. *Q.J.R.Meteorol.Soc.*, **117**, 1025-1056.