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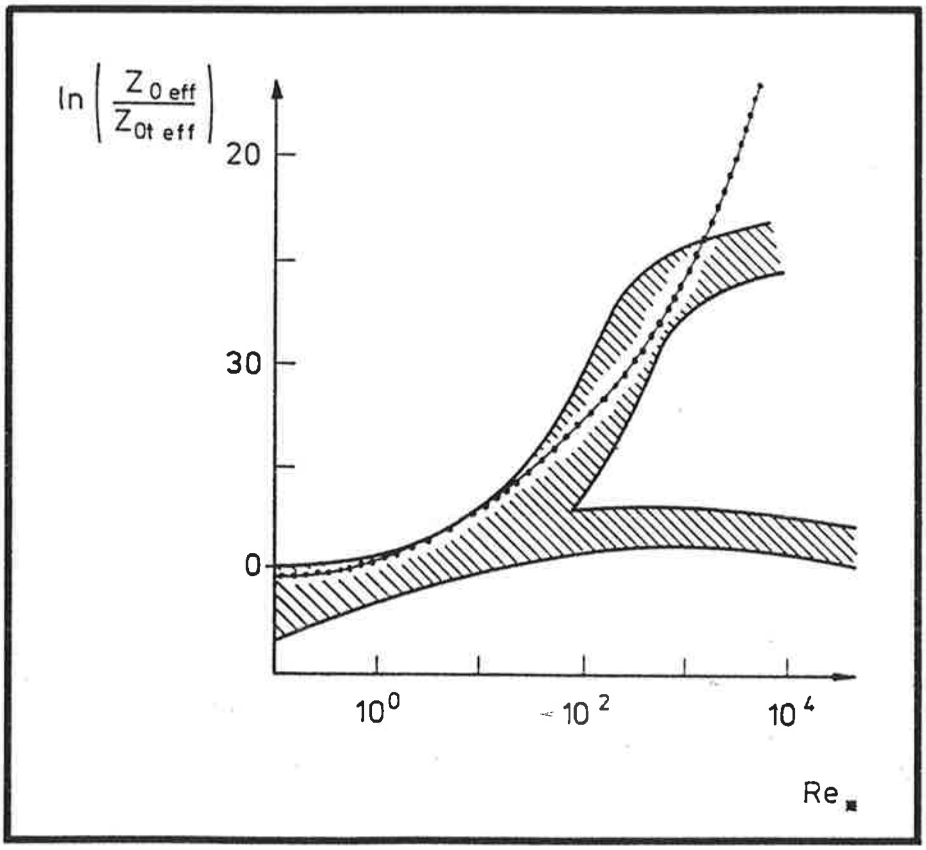

SCALE AGGREGATION IN SEMI-SMOOTH FLOW

by

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Scale aggregation in semi-smooth flow*

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Abstract

Recent work on vertical momentum, heat, and moisture transport over heterogeneous terrain has indicated that small areas of large roughness or obstacles dominate the regional momentum flux, whereas the regional heat and moisture fluxes are determined by the dominant surface cover. Therefore, it is hypothesized that regional momentum flux should be evaluated from an effective roughness length which implicitly accounts for the form drag of obstacles, whereas regional heat and moisture fluxes should be estimated from local surface parameters such as local roughness lengths. This hypothesis compares favorably with other proposals when testing it by using numerical simulations of regional surface fluxes in boundary-layer flow over homogeneous terrain with sparsely distributed roughness elements, i.e. in so-called *semi-smooth flow*.

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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. In heterogeneous terrain this leads to the problem of estimating area averages of surface fluxes and associated roughness lengths, the latter being defined only for homogeneous conditions.

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, 1990,1991a). Implicit in this concept is the assumption that at sufficiently large heights above a heterogeneous surface, the modification of air flow due to changes in surface conditions will not be recognizable individually, and an overall stress or heat flux profile will exist, representing the surface conditions of a large area. Consequtively, 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.

However, the idea of forming averages from the knowledge of homogeneous surfaces yields an aggregated surface roughness which is always smaller than the surface roughness of the roughest surface within a grid domain. This conflicts with the results of Klaassen's (1992) study of average fluxes from heterogeneously vegetated regions. It also conflicts with observations above irregular forests with many clearings (Wieringa, 1992). In this study, it will be argued that this conflict can be attributed to the neglect of form drag due to obstacles such as isolated trees and houses, windbreaks, small hills, or edges of tall vegetation.

In order to estimate the regional momentum and heat fluxes, it is hypothesized that only the regional momentum flux is directly affected by the form drag, whereas the regional heat and moisture fluxes are influenced by the dominant surface cover. Therefore, the regional momentum flux is estimated from an effective roughness length which is a measure of form drag exerted by obstacles on the air flow and surface shear stress of surface cover. For computation of regional heat fluxes, local surface conditions, particularly the local roughness lengths, are used.

The new hypothesis is tested against a recent proposal by Beljaars and Holtslag (1991) and an earlier model by Brutsaert (1975). Also Beljaars and Holtslag suppose that the regional momentum flux is dominated by form drag, and the regional heat fluxes, by dominant vegetation cover. They derive, however, an effective roughness length of temperature by assuming that heat fluxes should be constant with height. Consequtively, Beljaars and Holtslag evalutate the regional heat fluxes from effective roughness lengths. The same is valid for Brutsaert's model. Brutsaert's model is based on the eddy renewal concept of the interfacial sublayer underlying a turbulent boundary layer.

The comparison of methods is done by numerical simulations of regional momentum and heat fluxes for homogeneous agricultural terrain with hedgerows set up as shelterbelts. The average distance d between shelterbelts of height h_c is chosen $d/h_c > 15$ which is quite typical for hedgerows in Northern Germany. Moreover, for such large values of d/h_c, the form drag is easily estimated, since wake interference between shelterbelts can be neglected. Such an air flow with negligible wake interference is called *semi-smooth flow* according to Wieringa (1992).

The effective roughness length of a heterogeneous terrain could, in principle, be determined from measurements of wind standard deviations on a routine basis as indicated in Beljaars and Holtslag (1991). Here, in Section 2.2, Arya's (1975) model - with slight modifications - is used to compute the effective roughness length. Arya's model has successfully been applied to the arctic region of pack ice. More sophisticated models of effective roughness length for hilly terrain can be found in Emeis (1990), for example. Effective roughness lengths are not simply prescribed for the numerical simulations being discussed in Section 3, because some of the ideas implicit in the definition of an effective roughness length enter discussion of methods for estimating regional heat and moisture fluxes which will be presented in Section 2.2.

2. Conceptual considerations

2.1 A simple model of effective roughness length

It is suggested that Schlichting's drag partition theory can be used to express the total wind drag τ over a heterogeneous area as the sum of a skin drag S_d and a form drag F_d due to the obstacles:

$$\tau = S_d + F_d \quad . \tag{1}$$

A detailed discussion of Schlichting's theory is found in Marshall (1971) as well as its testing in wind tunnel experiments.

In turbulent flow over rough surfaces, the division between skin drag and form drag is somewhat arbitrary and depends on one's perspective. Here, 'skin drag', or better: surface shear stress, is considered as that portion of the drag associated with roughness elements, whose dimensions are of the order of a few centimeters or less. The effect of these small roughness elements on the surface-layer flow is represented by use of a roughness length z_0 in the usual way. The influence of larger roughness elements is parameterized in terms of an effective roughness length which will be constructed in the following manner.

Following Arya (1975), it is assumed that wakes which originate at obstacles blend at a height l_b such that at heights $z << l_b$ the flow is in equilibrium with the local surface, whereas at $z >> l_b$ the individual wakes will not be recognizable individually, and the flow 'feels' just a rougher surface. At heights $z << l_b$, but not too close to the obstacles, the wind profile can be parameterized by the local roughness length z_0 and local friction velocity u_*. At $z >> l_b$, it is a function of an effective roughness length Z_{0eff} and effective friction velocity U_{*eff}. At $z = l_b$, both conditions are approximately met - in keeping with Mason's (1988) idea of blending height. Consequtively, the ratio of total drag τ and surface shear stress S_{d0} of the open field without any obstacles can be written as

$$\frac{\tau}{S_{d0}} = \left(\frac{ln\frac{l_b}{z_0}}{ln\frac{l_b}{Z_{0eff}}} \right)^2 \quad . \tag{2}$$

Hence

$$ln\frac{Z_{0eff}}{z_0} = ln\frac{l_b}{z_0}\left(1 - (\frac{\tau}{S_{d0}})^{-1/2}\right) \quad . \tag{3}$$

In order to compute the effective roughness length, the form drag F_d and the total surface shear stress S_d have to be known. If the form drag F_d is assumed to be a function of the square of mean wind speed at the mean height h_c of obstacles and an empirically determined drag coefficient c_d, then The ratio of F_d and S_{d0} takes the form (see Arya, 1975):

$$\frac{F_d}{S_{d0}} = \frac{1}{2} c_d \frac{h_c}{d} \left(\frac{1}{\kappa} ln \frac{h_c}{z_0} \right)^2 \quad . \tag{4}$$

κ is the von Kármán constant (here, $\kappa = 0.4$).

The surface shear stress S_d on the surface between obstacles is smaller than the stress S_{d0} on the surface without any obstacles because of the reduction of wind speed downstream of each obstacle. According to Arya (1975):

$$\frac{S_d}{S_{d0}} = \left(1 - m \frac{h_c}{d} \right) \quad . \tag{5}$$

where m is an empirical parameter which should depend on the porosity and the shape of the obstacles considered.

The blending height is supposed to vary with the distance between obstacles. The dispersion of wakes generated at obstacles is presumably governed by the same processes leading to the development of internal boundary layers downstream of a step change in surface roughness. Hence

$$\frac{l_b}{z_0} = a \left(\frac{d}{z_0} \right)^{4/5} \quad . \tag{6}$$

Here, $a = 0.35$ is chosen. Arya (1975) proposes a larger value referring to Elliott's (1958) model of internal boundary layers. However, it seems that Elliott's model overestimates the depth of internal boundary layers (e.g. Claussen, 1991b, p. 47).

Wieringa (1976, 1992) suggests that the blending height depends on the height of obstacles rather than on the distance between them. He proposes

$$l_b \simeq 2h_c \quad . \tag{7}$$

There are no experimental tests whether l_b follows Equation 6 or 7. However, for many applications, e.g. $z_0 \sim O(0.1\text{m})$, $h_c \sim O(1\text{–}10\text{m})$, $d \sim O(100\text{m})$, use of either Equation 6 or 7 yields a similar estimate of Z_{0eff}, because, if $l_b/z_0 >> 1$, then small variations in l_b affect Z_{0eff} only little.

2.2 Regional heat fluxes

2.2.1 Brutsaert's model

Brutsaert (1975, 1979) deduces from his theory that the ratio of roughness lengths for velocity and scalar admixtures, respectively, is approximately constant only over porous surfaces such as dense vegetation. Over heterogeneous terrain with bluff roughness elements, this ratio should vary with the friction velocity and the roughness length itself. Brutsaert does not distinguish between local and effective roughness length. In the context of this study, his theory is applied to effective roughness lengths. Hence (see Brutsaert 1975, 1979)

$$ln\left(\frac{Z_{0eff}}{Z_{0teff}}\right) = 2.3 \tag{8a}$$

$$ln\left(\frac{Z_{0eff}}{Z_{0teff}}\right) = \kappa\left(7.3\left(\frac{U_{*eff}Z_{0eff}}{\nu}\right)^{1/4}Pr^{1/2} - 5\right) \tag{8b}$$

where Z_{0teff} is the effective roughness length of temperature, ν is the molecular viscosity, and Pr is the molecular Prandtl number (Pr=0.71). The same equations are valid for the effective roughness length of humidity Z_{0qeff} with the exception that the molecular Schmidt number Sc (Sc=0.6) should be used instead of Pr.

Brutsaert's theory is validated mainly by laboratory experiments as indicated by Figure 1. It should be noted, however, that Equation 8b becomes unrealistic for large Reynolds numbers $Re_* = U_{*eff}Z_{0eff}/\nu$, say $Re_* > 10^3$.

2.2.2 Beljaars and Holtslag's model

Beljaars and Holtslag (1991) derive an effective roughness length of temperature by assuming that the sensible heat flux is not directly affected by form drag and, therefore, should be constant with height. Furthermore, Beljaars and Holtslag argue that the temperature profile above the blending height l_b can be parameterized by an effective temperature scale Θ_{*eff} and Z_{0teff} - the same conceptual picture as applied to the velocity profile. This leads to:

$$ln\left(\frac{l_b}{Z_{0teff}}\right) = \frac{ln(l_b/z_0)\ ln(l_b/z_{0t})}{ln(l_b/Z_{0eff})} \quad . \tag{9}$$

From Equation 9 it is seen that Z_{0teff} varies with the distance between obstacles, but it is independent of friction velocity. Both statements are at variance with Brutsaert's model. Nevertheless, Beljaars and Holtslag's values of Z_{0teff} are confirmed by measurements near Caubauw and Les Gers (MESOGERS '84).

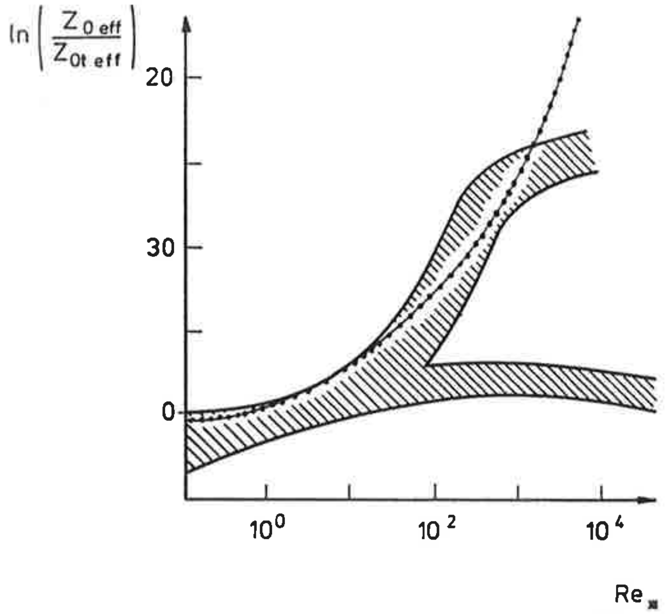

Figure 1: Ratio of effective roughness lengths of velocity and temperature as function of friction Reynolds number. The dotted line is computed from Brutsaert's model (Equation 8b). The hatched area indicate experimental data for arrays of bluff obstacles (upper branch) and for more fibrous material such as vegetation (lower branch). This figure is redrawn with modification from Hicks (1985).

2.2.3 A new concept

An alternative view on Beljaars and Holtslag's conceptual model is the following hypothesis. Recent work on the aerodynamics of shelter (see McNaughton, 1988) has shown that there exists a zone of reduced turbulent transport of heat and vapor behind obstacles up to distances x of approximately $x/h_c \sim 8$. Beyond that, further downstream, lies a wake region of increased turbulent transport. Hence it seems unlikely that turbulent heat fluxes remain constant with height. On the other hand, in boundary-layer flow without wake interference between obstacles, i.e. $d/h_c >> 15$, the horizontal averages of surface heat fluxes should be little affected, because the reduction of wind speed downstream and upstream of obstacles covers only a small area in comparison with the area between obstacles. Therefore, it is hypothesized that the regional heat fluxes should be determined from local parameters of the dominant surface cover, i.e. from local roughness lengths z_0, z_{0t}, z_{0q} of velocity, temperature, and humidity, respectively. Only the regional momentum flux is supposed to be a function of the effective roughness length Z_{0eff}.

At the first glance, Beljaars and Holtslag's model and the new proposal resemble each other. However, the latter relaxes the assumption of height independent heat and moisture fluxes. Moreover, and perhaps more important, the latter concept is easily applicable to heterogeneous terrain with both obstacles and patchy surface variations as shown in Claussen and Klaassen (1992). For these terrain conditions, the regional heat fluxes are estimated from an weighted average of local heat fluxes for each land type by considering aggregated roughness lengths as outlined in Claussen (1991a).

3. Numerical simulations

3.1 Terrain conditions

Regional momentum and heat fluxes are simulated for a homogeneous agricultural terrain with sparsely distributed hedgerows. The average distance between hedgerows is chosen $d = 370$m which is a typical value observed in Northern Germany (e.g. Timm, 1984). The height h_c of hedgerows is taken as $h_c = 6$m, hence $d/h_c \simeq 62$.

In order to check the sensitivity of the regional fluxes to variations in effective roughness length, d is variied from $d = 370$m to $d = 230$m, i.e. to $d/h_c \simeq 38$, a typical value registered in Northern Germany approximately forty years ago - before the industrialization of agriculture.

For the evaluation of the effective roughness length, the following parameters are chosen. The local roughness length z_0 of the agricultural terrain is prescribed as $z_0 = 0.03$m. For the drag coefficient, $c_d = 0.8$ is chosen in keeping with Gross (1987). Also, this value is close to Marshall's (1971) measurements of unobstructed form drag coefficients of slender elements. From wind tunnel experiments, Arya (1975) specifies $m = 20$ for prismatic, solid obstacles. Here, the same value of m is taken as a first guess. (It turns out that the ratio Z_{0eff}/z_0 is not very sensitive to variations of m for large values of d/h_c.)

From Equation 6, the blending height is evaluated as $l_b \simeq 13$m for $d/h_c \simeq 38$ and $l_b \simeq 20$m for $d/h_c \simeq 62$. Hence $Z_{0eff} \simeq 0.24$m for $d/h_c \simeq 38$ and $Z_{0eff} \simeq 0.16$m for $d/h_c \simeq 62$ from Equation 3, 4, 5. Using Wieringa's (1992) estimate of $l_b = 2h_c = 12$m, $Z_{0eff} \simeq 0.23$m for $d/h_c \simeq 38$ and $Z_{0eff} \simeq 0.14$m for $d/h_c \simeq 62$. As mentioned in Section 2.1, this is quite small a difference.

In Beljaars and Holtslag's model, the effective roughness length of temperature, Z_{0teff}, is (see Equation 9) $Z_{0teff} \simeq 3.8\ 10^{-5}$m, i.e. $Z_{0eff}/Z_{0teff} \simeq 6.3\ 10^3$, for $d/h_c \simeq 38$ and $Z_{0teff} \simeq 1.5\ 10^{-4}$m, i.e. $Z_{0eff}/Z_{0teff} \simeq 1.1\ 10^3$, for $d/h_c \simeq 62$. The value of $Z_{0eff}/Z_{0teff} \simeq 6.3\ 10^3$ is very close to the observed values near Cabauw and Les Gers reported by Beljaars and Holtslag (1991).

3.2 Boundary and initial conditions

The one-dimensional version of a non-hydrostatic meso-scale model is used to simulate the vertical transport of momentum, heat, and moisture within the atmospheric boundary layer. The model as well as its testing is described in detail in Kapitza and Eppel (1992) and Mengelkamp (1991). The treatment of the atmosphere-soil-vegetation interface is briefly summarized in the Appendix.

The initial and boundary conditions of the model are the following: The daily course of solar radiation is computed for a latitiude of $\phi = 54^o$N at a July 21st. No clouds are assumed. The large scale wind and potential temperature at a height of 6000m are held constant at 14m/s and 322.16K, repectively. The temperature within the soil at a depth of 1m is a constant value of 286K. The inital potential temperature at the surface is 283K, the initial soil water content is 70% of the critical field capacity.

Five days have been simulated for each application in which an almost steady daily cycle of surface fluxes is achieved. Here, only results from the last day will be shown.

3.3 Results

3.3.1 Regional friction velocity

In Figure 2, the effective and local friction velocities, i.e. the square root of regional and local momentum flux, are depicted. Thick curves indicate results when using the present proposal, hatched curves, Beljaars and Holtslag's (1991), and thin curves, Brutsaert's (1975), but with the restriction that $ln(Z_{0eff}/Z_{0teff}) \leq 13$ in Equation 8b. Open circles refer to the friction velocity of the open field without any obstacles. Open triangles top down indicate effective friction velocities over terrain with $d/h_c \simeq 62$, solid triangles top down, with $d/h_c \simeq 38$. Open triangles top up are used for local friction velocities over terrain with $d/h_c \simeq 62$, solid triangles top up, with $d/h_c \simeq 38$.

It is clearly seen that the effective friction velocity increases with increasing effective roughness lengths. The local friction velocity, however, decreases, because the wind speed near the surface decreases as the region becomes rougher. For example, at 12:00 local time, the wind speed on average over the lowest 40m above ground decreases from 8.5m/s over open terrain, i.e. for $Z_{0eff} = z_0 = 0.03$m, to 6.8m/s for $Z_{0eff} =$

0.16m, and to 6.3m/s for $Z_{0eff} = 0.24$m. Differences between the present model, Beljaars and Holtslag's model, and Brutsaert's model are much smaller than differences between friction velocities for different effective roughness lengths. There is, however, one exception.

It turns out that Brutsaert's model yields unrealistically large ratios of effective roughness lengths. For example, at 12:00 local time, $ln(Z_{0eff}/Z_{0teff}) = 22.9$ for $Z_{0eff} = 0.24$m and $ln(Z_{0eff}/Z_{0teff}) = 20.3$ for $Z_{0eff} = 0.16$m (compare with Figure 1). Thus, when using Brutsaert's model the ratio of roughness lengths were not allowed to exceed a value of $ln(Z_{0eff}/Z_{0teff}) = 13$ in keeping with measurements indicated in Figure 1. Without this restriction, effective friction velocities computed from Brutsaert's model deviate more strongly from Beljaars and Holtslag's model and the present proposal. As an example, the thin curve with solid squares gives the effective friction velocity of Brutsaert's original model for $Z_{0eff} = 0.24$m.

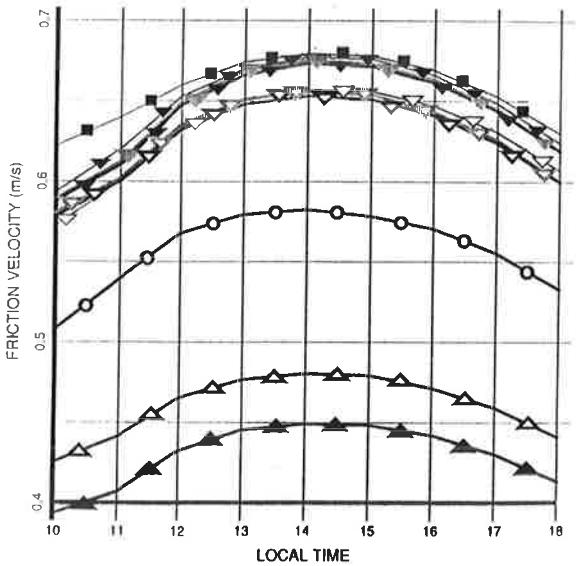

Figure 2: Variation with local time of local and regional friction velocity. Thick lines: present model, hatched lines: Beljaars and Holtslag's (1991) model, thin lines: Brutsaert's (1975, 1979) model, but with the restriction that $ln(Z_{0eff}/Z_{0teff}) \leq 13$. Open circles: open field without obstacles, open triangles down: effective friction velocity for $Z_{0eff} = 0.16$m, solid triangles down: effective friction velocity for $Z_{0eff} = 0.24$m, open triangles up: local friction velocity for $Z_{0eff} = 0.16$m, solid triangles up: local friction velocity for $Z_{0eff} = 0.24$m, solid squares: effective friction velocity for $Z_{0eff} = 0.24$m using Brutsaert's original model (Equation 8).

3.3.2 Regional latent heat flux

In Figure 3, the regional latent heat flux is plotted. There is a slight increase in the latent heat flux as the effective roughness length increases from $Z_{0eff} = z_0$ to $Z_{0eff} = 0.24$m as predicted by Beljaars and Holtslag's and by the present model. Brutsaert's model yields the opposite trend which seems a little bit strange, because one would expect a monotonous change from open field to terrain with increasing effective roughness.

The relative differences in regional latent heat flux are much smaller than in regional momentum flux. (Therefore, a much smaller time interval had to be chosen for the graphics in order to demonstrate some differences.) It seems that differences among models are larger than differences due to changes in effective roughness - except for the present proposal and Beljaars and Holtslag's proposal which yield almost identical result. As for the effective friction velocity, results when using Brutsaert's original model are somewhat off the other results (see thin curve with solid squares).

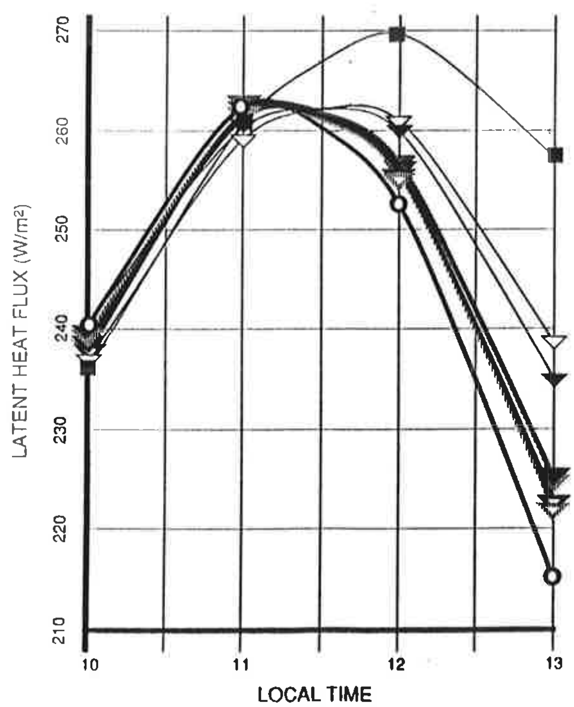

Figure 3: Same as Figure 2, except for regional latent heat flux.

3.3.3 Regional sensible heat flux

In Figure 4, the regional sensible heat flux is depicted. Obviously, the regional sensible heat flux decreases slightly with increasing effective roughness. The opposite trend is observed when Brutsaert's model is used. But as for the regional latent heat flux, these differences are small in comparison with differences in regional momentum flux. Only Brutsaert's model yields noticably large changes which are even stronger, if Brutsaert's original model is not corrected for unrealistically large values of Z_{0eff}/Z_{0teff}.

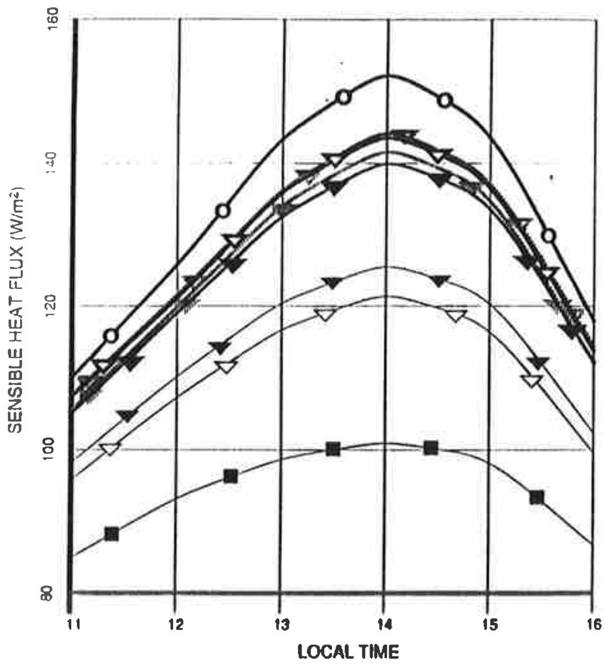

Figure 4: Same as Figure 2, except for regional sensible heat flux.

3.3.4 Discussion

The small changes in regional heat fluxes - in comparison with large changes in regional momentum flux - can be explained in terms of changes in turbulence scales. The latent heat flux and the sensible heat flux are functions of friction velocity u_* and turbulence scales of humidity q_* or temperature Θ_*, such that $Q_{lat} \sim u_* q_*$ and $Q_{sens} \sim u_* \Theta_*$. The local friction velocity decreases with respect to its value u_{*0} in the open field as the regional roughness becomes larger. Therefore, the local values of q_* and Θ_* have to increase above q_{*0} and Θ_{*0}, because the regional heat fluxes remain almost unchanged. This basically means that near the surface, vertical gradients of moisture and of temperature are stronger in air flow above obstacles than above the open field. This, in fact, is observed for shelterbelts (e.g. McNaughton, 1988).

The opposite is valid for effective turbulence scales. While the effective friction velocity increases with increasing regional roughness, the effective scales Q_{*eff} and Θ_{*eff} decrease. The products, i.e. the regional heat fluxes $Q_{lat} \sim U_{*eff} Q_{*eff}$ and $Q_{sens} \sim U_{*eff} \Theta_{*eff}$ vary only slightly, as already suggested by Beljaars and Holtslag (1991). For convenience, their conceptual picture is redrawn with modifications in Figure 5.

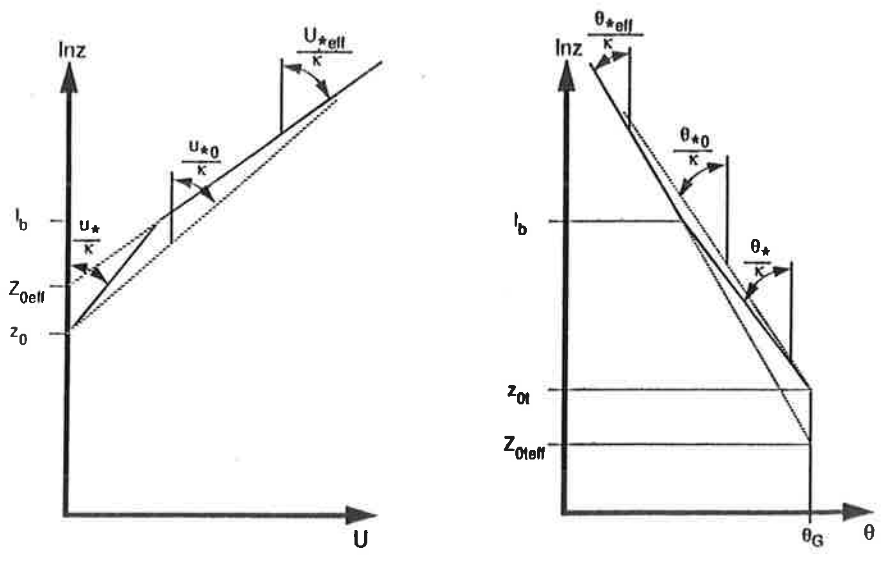

Figure 5: Conceptual picture of near-neutral wind and potential temperature profiles in semi-smooth flow. This Figure is redrawn with modifications from Beljaars and Holtslag (1991).

During daytime, the heat fluxes are often determined by the canopy or vegetation conductance, and changes in roughness length become less relevant for the turbulent heat fluxes. On the other hand, the roughness length of temperature determines the surface temperature and, therefore, the radiative feedback as well as the ground heat flux. It appears that both quantities, surface temperature and ground heat flux, exhibit even less variation with changing effective roughness length Z_{0eff} than the turbulent heat fluxes. Furthermore, the agreement between models is better when considering surface temperature and ground heat flux rather than turbulent heat fluxes. Therefore, plots of the former quantities are not shown.

4. Conclusions

Recent work on vertical momentum, heat, and moisture transport over heterogeneous terrain by Beljaars and Holtslag (1991) has indicated that small areas of large roughness or obstacles dominate the regional momentum flux, whereas the regional heat and moisture fluxes are determined by the dominant surface cover. By assuming that the heat and moisture fluxes are constant with height, Beljaars and Holtslag derive an effective roughness length of temperature - the counterpart of the effective roughness length of velocity which implicitly accounts for the form drag due to obstacles. Their estimated ratio of effective roughness length of velocity and temperature agrees well with observations. Consequtively, Beljaars and Holtslag recommend that regional heat and moisture fluxes should be evaluated from effective roughness lengths.

Here, an alternative view on Beljaars and Holtslag's concept is proposed. Based on observations of air flow in the vicinity of shelterbelts, it is questioned whether vertical transports of heat and moisture are constant with height. On the other hand, it is still believed that the heat and moisture fluxes on the average over a larger area are mainly dominated by surface cover and not directly affected by form drag of obstacles. Relaxing the assumption of height independent heat and moisture fluxes, it is suggested that only the regional momentum flux should be computed from effective roughness length, whereas the regional heat and moisture fluxes should be estimated from local surface parameters, particularly from local roughness lengths.

The new concept is compared with Beljaars and Holtslag's method and with an earlier model by Brutsaert (1975) by using numerical simulations of regional surface fluxes in boundary-layer flow over homogeneous terrain with sparsely distributed hedgerows. It turns out that all concepts yield the same qualitative picture: The regional momentum flux clearly varies with the changing distance between hedgerows, i.e. with the regional roughness of the terrain, whereas the regional heat and moisture fluxes exhibit little variations. Differences between results from Beljaars and Holtslag's and the present method are just marginal, whereas results from Brutsaert's model deviate more strongly from the others. In fact, Brutsaert's model does not seem to be applicable at large friction Reynolds numbers, say $Re_* = U_{*eff} Z_{0eff}/\nu > 10^3$. Unfortunately, these large values of Re_* are commonly observed in atmospheric flow. However with a slight modification, Brutsaert's method leads to results similar to the other methods.

The good agreement between Beljaars and Holtslag's and the new concept is no surprise. Both concepts resemble each other. Implicit in both concepts is the assumption that regional heat fluxes are determined by the characteristics of the dominant surface cover. The advantage of the new concept is that it can easily be used for estimation of surface fluxes in heterogeneous terrain with both obstacles and patchy variations of surface conditions - which is quite important for many applications in large scale numerical models.

Acknowledgements

The author wishes to thank Anton Beljaars, ECMWF, Reading, UK, and Larry Mahrt, Orgeon State University, USA, for discussion.

Appendix

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

$$Q_{rad} + Q_{lat} + Q_{sens} = Q_{ground} \quad . \tag{A.1}$$

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:

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

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:

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

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

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.

w_e is a so-called wetness factor which is assumed to depend on the canopy conductance g_s via (e.g. Deardorff, 1978)

$$w_e = \frac{g_s}{g_s + C_q U_a} \quad . \tag{A.5}$$

For g_s a simple model is used:

$$g_s = g_1 \, g(S_o) \, g(\delta q) \, g(T_a) \, g(m_e) \tag{A.6}$$

where g_1 is the maximum surface conductance which depends on the type of vegetation. The values of g_1 are taken from Wilson *etal.* (1987). For wet leaves g_1 is assumed to become infinitly large. The other terms on the right-hand-side of Equation A.6, account for the reaction of the stomata of plants to solar radiation, specific water vapor deficit δq, actual air temperature T_a, and relative soil moisture m_e which is estimated by using a force-restore method (see Deardorff, 1978). Here, it is assumed (as a very first aproximation) that all plants reveal the same behavior with respect to their stomata, except for the extreme values g_1. The explicit functional dependence of these terms is given in Dolman (1987), who evaluated the evaporation over an oak forest.

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.

The conductive heat flux into the soil is

$$Q_{ground} = -\lambda \left(\frac{\partial T}{\partial z} \right)_{z=0} \tag{A.7}$$

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

$$\frac{\partial T}{\partial t} = k \frac{\partial^2 T}{\partial z^2} \quad . \tag{A.8}$$

with k as temperature diffusivity.

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