SURFACE FLUX PARAMETERIZATION SCHEMES: DEVELOPMENTS AND EXPERIENCES AT KNMI

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Abstract

In this paper a summary is given of observations and modelling efforts on surface fluxes, which have taken place at KNMI. Emphasis is put on those aspects that are important for numerical weather and climate models e.g. surface roughness lengths for momentum and heat, surface stomatal resistance for evaporation and related quantities. Comparisons of flux estimates with observations from the Cabauw facilities and from the MESOGERS data are given. We present a discussion of exchange coefficients for momentum and heat in stable and very stable conditions and compare current approaches. New formulae for the exchange coefficients are proposed, which are consistent with observations from the Cabauw facilities.

1. INTRODUCTION

The growing interest in the interaction between the atmosphere and the underlying surface is driven by the realisation that the surface fluxes of momentum, heat and moisture determine to an important degree the steady state of the atmosphere. The atmospheric part of this interaction takes place in the boundary layer with a shallow layer near the surface layer, where Monin-Obukhov theory applies. Monin-Obukhov (MO) similarity theory is a generally accepted framework for describing the surface layer and virtually all numerical models make use of it in one way or the other. The surface fluxes are of crucial importance not only because they influence the steady state of the atmosphere, but also because they determine the mean profiles of the surface layer and the atmospheric boundary layer (see Holtslag and Nieuwstadt, 1986). This is equally true for short term forecasts where computed boundary layer parameters are important forecast products. Examples are wind and temperature forecasts at standard observation height. In addition fields of wind, exchange coefficients, boundary layer height, needed as input for air pollution and ocean wave models are dependent upon the surface fluxes.

The tendency to apply up to date boundary layer theory, can also be seen in the air pollution community (cf., Gryning et al., 1987). Experimental work and modelling efforts at KNMI have attempted to parameterize the surface fluxes of momentum, heat and moisture in terms of routinely measured meteorological parameters (Holtslag and Van Ulden, 1983; Van Ulden and Holtslag, 1985; Holtslag and De Bruin, 1988). The purpose of these studies was to make stability classification with help of MO-theory accessible to air pollution modellers, hydrologists, and wind engineers. It is hoped that the similarity approach will ultimately replace the classical Pasquill- Gifford stability classification. The flux schemes are also applied in short range weather forecast models (e.g. Reiff et al., 1984).

The purpose of this paper is to summarize the results and conclusions from surface flux observations and modelling efforts at KNMI, with emphasis on those aspects that are important for numerical forecast models. Most of the experimental "surface-flux" work was carried out near the 200 m meteorological mast at Cabauw in the centre of the Netherlands. A large number of studies have been based on datasets from this mast, which has the advantage that the datasets and the location are well documented from different points of view (see Monna and Van der Vliet, 1987 for a survey).

Apart from Cabauw measurements, data from the MESOGERS-84 experiment will also be used. The difference between the two datasets is related to the terrain complexity. The Cabauw site is fairly homogeneous, with only small complex terrain effects. The MESOGERS data are from the south of France, which is much less homogeneous with undulating terrain and variable vegetative cover. Real terrain is often inhomogeneous at horizontal scales that are small compared to model grid spacing. One of the purpose of the MESOGERS-experiment was to study the problem of subgrid parameterisation, i.e. to study the surface fluxes at regional scales.

It is clear that the information from these two sites is not sufficient to develop surface flux schemes that have world wide applicability. Surface fluxes depend on many parameters such as vegetation, soil type, soil moisture, snow cover, drainage etc, while these parameters vary only over a limited range for the data discussed here. It is believed, however, that these data can elucidate certain aspects of the land surface parameterization problem.

Section 2 gives a short description of the data used in this paper. In section 3 a number of formulations will be introduced for the surface fluxes, while in section 4 the surface parameters for the approaches are discussed. The latter section also contains a comparison of different flux estimates with our data. Subsequently, section 5 contains a discussion of exchange coefficients. Since, reasonable agreement exists in literature on the exchange coefficients for neutral and unstable conditions, we concentrate ourselves on the coefficients for stable and very stable conditions. Finally, in section 6 a discussion and summary is given.

2. DESCRIPTION OF DATA SETS

In this paper we make use of data gathered almost continuously at the 200 m meteorological mast of Cabauw in the centre of the Netherlands (Driedonks et al, 1978) and of data obtained during the MESOGERS-84 experiment in the south of France (Weill et al. 1988).

The Cabauw site is located in flat terrain consisting mainly of grassland interrupted by narrow ditches. Up to a distance of about 200 m from the mast there are no obstacles or perturbations of any importance; further on we find some scattered trees and houses for most wind directions. For easterly winds the flow is perturbed by tree rows, orchards and a village. Although the terrain looks very uniform, some weak inhomogeneous terrain effects have been observed. They are mainly reflected in perturbed flux-profile relationships below 20 m (cf. Beljaars et al., 1983; Schmid, 1988).

The MESOGERS site, located in the south of France, is much more complex. The terrain is undulating with a maximum amplitude of about 50 m . The

$$-\frac{H}{\rho c_p u_*} = \theta_* = \frac{k(\theta_1 - \theta_s)}{\ln(\frac{z_1}{z_{oH}}) - \psi_H(\frac{z_1}{L}) + \psi_H(\frac{z_{oH}}{L})}, \tag{5}$$

$$-\frac{\lambda E}{\rho \lambda u_*} = q_* = \frac{k(q_1 - q_s)}{\ln(\frac{z_1}{z_{oQ}}) - \psi_Q(\frac{z_1}{L}) + \psi_Q(\frac{z_{oQ}}{L})}, \tag{6}$$

where u_* is the friction velocity, θ_* is the temperature scale, q_* is the humidity scale, k is the Von Kármán constant (k = 0.4), z_1 is the height of the first model level and L is the Obukhov length scale. ψ_M, ψ_H and ψ_Q are dimension-less stability functions of height and Obukhov length (see also section 5).

In Equations (4)-(6), the parameters z_{oM}, z_{oH}, and z_{oQ} are known as the surface roughness lengths for momentum, heat and moisture. From (1)-(6) it can be seen that the exchange coefficients will depend, among other things, on their related roughness lengths. The roughness lengths are often taken as equal, while it is characteristic that the roughness length for momentum might be much larger than the other two. We will return on this issue in section 4.

For the application of (1)-(3) or (4)-(6) we need the values of U_1, θ_1, q_1, θ_s and q_s. These variables are usually represented by separate prediction equations, except for q_s which can be obtained by specifying or calculating the surface relative humidity RH_s, as for example

$$q_s = RH_s\, q_{sat}\,(T_s)\ . \tag{7}$$

Here T_s is the actual surface temperature, which is related to θ_s. Alternatively, q_s can be eliminated for vegetated surfaces, by introducing a stomatal resistance r_c:

$$q_{sat}(T_s) - q_s = \frac{\lambda E}{\rho \lambda}\, r_c\ , \tag{8}$$

which with (3) leads to (e.g. Bell and Dickinson, 1987) orientation of the ridges and valleys is typically north-south with a mean distance between ridges of about 5 km. The second complication is the surface vegetation. The vegetation varies from one field to the other, ranging from bare soil and grassland to vineyards, cornfields and stands of trees. The different parcels have typical dimensions of a few hundred meters (see also Durand et al., 1987 and Beljaars, 1988a).

Both sites have clay soil, which is very efficient in supplying water to the vegetation. Even during the first two weeks of MESOGERS with very dry looking soil (with cracks in the clay), quite large evaporation rates were measured (see section 4).

3. FLUX FORMULATIONS

In weather and climate models the surface fluxes of momentum (τ), sensible heat heat (H) and latent heat (λE) are usually expressed as (e.g. Louis, 1979)

$$\frac{\tau}{\rho} = C_M U_1^2 , \tag{1}$$

$$-\frac{H}{\rho c_p} = C_H U_1 (\theta_1 - \theta_s), \tag{2}$$

$$-\frac{\lambda E}{\rho \lambda} = C_Q U_1 (q_1 - q_s), \tag{3}$$

where ρ is the density, c_p is the specific heat and λ is the latent heat of vapourization. U_1, Θ_1 and q_1, are wind speed, potential temperature and humidity at the first model level in the atmosphere; Θ_s and q_s are the corresponding surface values, and C_M, C_H, C_Q are exchange coefficients for momentum, heat and moisture.

If the first model level is positioned in the surface layer C_M, C_H, C_Q can be derived from surface layer theory, which states that (e.g., Yaglom, 1977)

$$\frac{\tau}{\rho u_*} = u_* = \frac{kU_1}{\ln(\frac{z_1}{z_{oM}}) - \psi_M(\frac{z_1}{L}) + \psi_M(\frac{z_{oM}}{L})} , \tag{4}$$

$$\frac{-\lambda E}{\rho\lambda} = \frac{C_Q\, U_1\, (q_1 - q_{sat}\,(T_s))}{(1 + r_c\, C_Q\, U_1)} \; . \qquad (9)$$

For vegetated surfaces, H and λE can also be represented by the Penman-Monteith equation, which for λE reads as (Monteith, 1981)

$$\lambda E = \frac{S(Q^* - G) + \rho c_p\, (q_{sat}(T_1) - q_1)\,/r_a}{S + (1 + r_c/r_a)\,\gamma} . \qquad (10)$$

Here S is the slope of the saturation specific humidity curve, $\gamma = c_p/\lambda$, Q^* is the net radiation at the surface, G is the soil heat flux, T_1 is the actual temperature in air, and r_a is the so-called aerodynamic resistance. In this context the latter can be written as $1/r_a = C_H\, U_1$, where it is assumed that $C_H = C_Q$.

As a simplification of Eq.(10), Priestley and Taylor (1972) argued that for homogeneous saturated surfaces λE can be estimated by

$$\lambda E = \alpha\, \frac{S}{S+\gamma}\, (Q^* - G), \qquad (11)$$

if there is minimal advection (see also Brutsaert, 1982). In (11) α is a parameter related to surface humidity. De Bruin and Holtslag (1982) verified this relation for daytime conditions above a grass covered surface in the Netherlands (see also De Bruin, 1983).

4. PARAMETER ESTIMATION AND COMPARISON OF FLUXES WITH DATA

4.1 The momentum flux

The relation between the wind speed at the lowest model level and the momentum flux is mainly determined by the roughness length z_{oM}. In neutral conditions, z_{oM} is even the only unknown parameter in Eq. (4) and is usually derived from geographical maps with help of empirical tables, relating different types of surface cover and land use to z_{oM}-values (e.g. Davenport, 1960; Van Dop, 1983; Wieringa, 1981,1986). Such procedures are necessarily inaccurate since we need effective values for an entire grid square. This constitutes a fundamental problem. Formally speaking, the roughness length concept is only applicable to homogeneous terrain, whereas real terrain is very often inhomogeneous at all scales between 100 m and the grid scale. The fundamental question of how to average over inhomogeneous conditions or surfaces has not been answered yet in a satisfactory way (cf. André and Blondin, 1986; Taylor, 1987), but even with efficient tools it would be a very difficult task to go from land use maps to subgrid averaged z_{oM}-values.

At the two experimental sites described in section 2, we have inhomogeneous terrain effects that are of particular importance to the momentum flux problem. It is shown by Beljaars et al. (1983) that the local roughness length of the Cabauw site (of the order of 1 cm) is smaller than the effective roughness length (of the order of 10 cm). The effective roughness length is defined here as the one that is needed in Eq. (4) to derive the correct area averaged momentum flux in a model (cf. Fiedler and Panofsky, 1972). A simple way of estimating an effective z_{oM} for a particular measuring site is to analyse the standard deviation of wind speed or wind direction (see Beljaars, 1987, 1988a, 1988b). The values obtained in this way are representive of an upstream fetch of a few kilometers. For easterly winds at the Cabauw site we find z_{oM} values between 10 and 20 cm, which are an order of magnitude larger than the local value. For MESOGERS we obtain values between 20 and 50 cm.

In Fig.1 we show u_* derived from the wind speed at 10 m height and the effective roughness length in comparison with eddy correlation measurements at 20 m height. The latter is sufficiently far away from the surface that it can be considered to be representive of a larger area. The way u_* has been derived from the wind speed is very similar to what is common practice in forecast models. Equation (4) is used with a prescribed value for z_{oM}; the stability corrections, however, are derived from observed temperature differences. The figure shows a reasonable correspondence, indicating that the effective z_{oM} is needed rather than the local one. Comparable conclusions could be drawn from analysis of MESOGERS data (cf. Beljaars, 1988a).

The conclusion we draw here, is that the momentum flux can be well

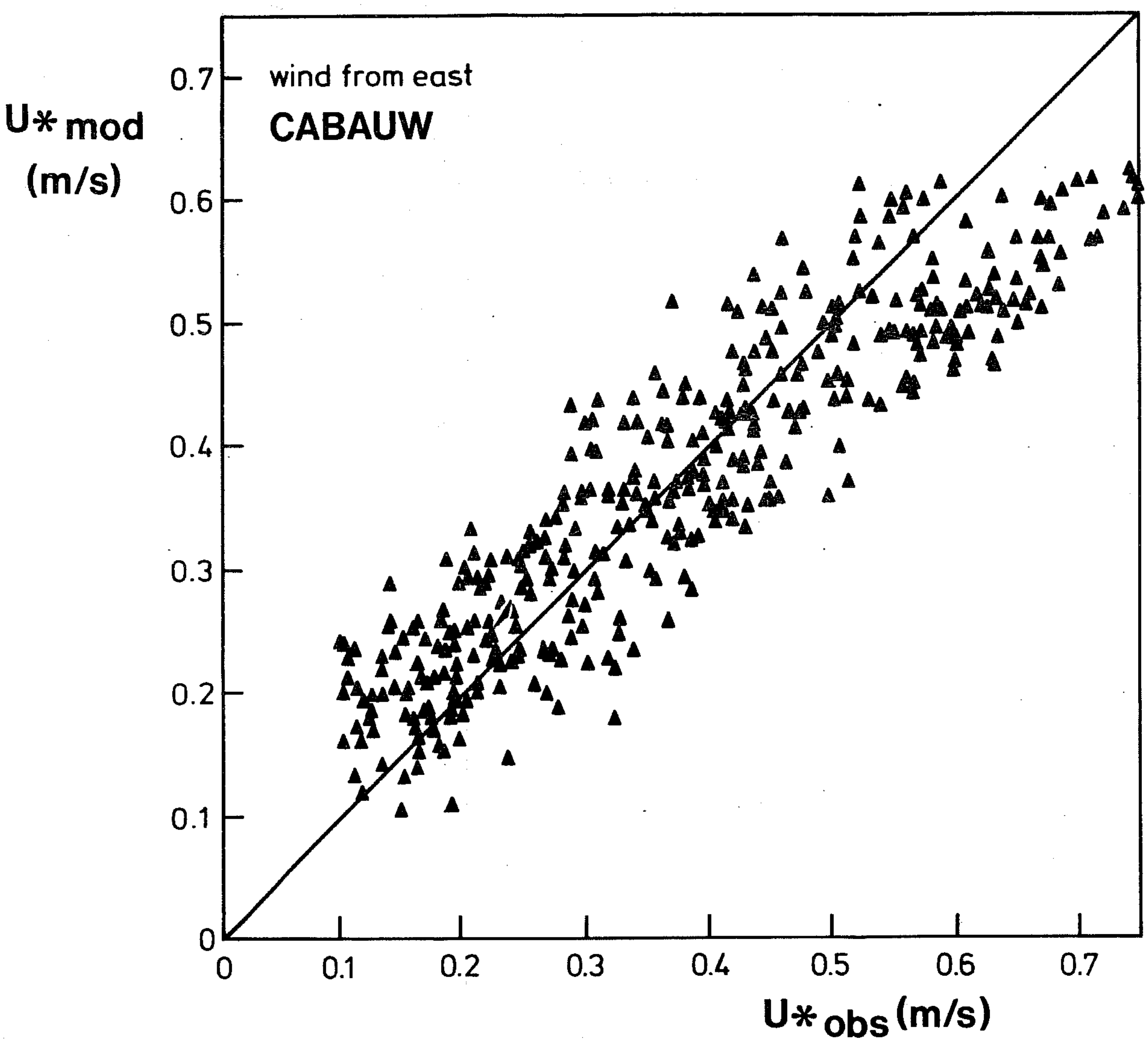

Fig.1 A comparison of friction velocity $u_{*\,mod}$ calculated with Eq.(4) from measurements of 10 m wind speed, Obukhov length and effective roughness length, with u_{*obs} obtained from eddy correlation measurements. Here the Cabauw turbulence data are used (Beljaars, 1982).

predicted, provided that an effective roughness length is known. This effective roughness length should be representative for the area of interest (i.e. a gridsquare in prediction models). For the experimental sites we were able to determine z_{oM} for horizontal scales up to a few kilometers. Extension up to grid scale is necessary for model computations, but we think that the actual methods to do this are fairly inaccurate. Information on horizontal scales of a few kilometers arround wind stations would be helpfull in this context and can be obtained by measuring wind standard deviations on a routine basis (Beljaars, 1987). It should also be noted that it is important to compute the momentum flux as accurately as possible, not only to get the momentum balance correct but also because it is a prerequisite to arrive at accurate estimates of heat and moisture fluxes.

4.2 The sensible heat flux

For the estimation of the sensible heat flux H, in principle the roughness length for heat z_{oH} is needed, as can be seen from Eq. (5). In weather and climate models it often is assumed that $z_{oH} = z_{oM}$. This means that no distinction is made between the surface skin temperature θ_s and the temperature θ_o at the height of the roughness length for momentum. To get an impression of the validity of this assumption, we have compared $\theta_o - \theta_s$ with θ_* in Fig.2 using the Cabauw data. Here we have assumed that the skin temperature is equal to the observed surface radiation temperature. Furthermore, θ_* is derived from profile measurements (temperatures at 10 m and 0.6 m, wind at 10 m; see Beljaars, 1982), and θ_o is derived by extrapolating the surface layer temperature profile downward to $z = z_{oM}$.

From the comparison in Fig.2 it is seen that the temperature difference can vary considerably from +6 K in stable conditions to -6 K in very unstable conditions. This temperature difference can be related to z_{oM} and z_{OH}, by

$$\frac{\theta_o - \theta_s}{\theta_*} = \frac{1}{k} \ln\left(\frac{z_{oM}}{z_{oH}}\right) . \tag{12}$$

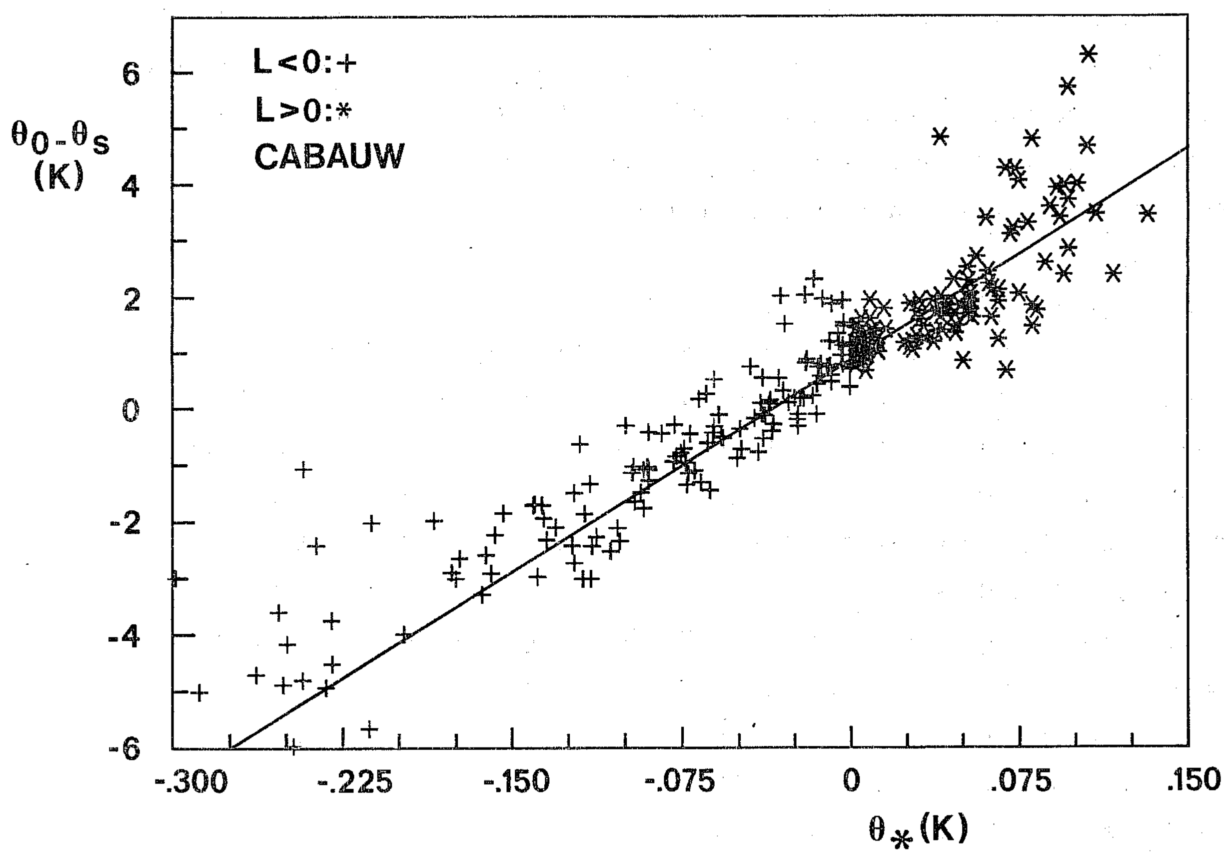

Fig.2 The relationship between the near-surface temperature difference ($\theta_o - \theta_s$) and the turbulent temperature scale θ_* (see text) for Cabauw data (June, 1986). The straight line is an "eye"-fit of the data points.

This ratio is known as B^{-1} in literature (Garratt and Hicks, 1973). Brutsaert (1982) concludes that typically $B^{-1} \approx 6$ for homogeneous grass covered surfaces, which means that $z_{oH} \approx z_{oM}/10$. Our findings of Fig.2, however, indicate $B^{-1} \approx 25$, as represented by the solid line. This means that $z_{oH} = 4.5\ 10^{-5}\ z_{oM}$, which is substantially smaller than the above value. Also for MESOGERS the parameter B^{-1} turns out to be much larger than found in literature (see Beljaars, 1988a). We think that this discrepancy is related to inhomogeneous terrain effects, which will be discussed in more detail in section 6. This will have important consequences for the estimation of the surface temperature and for the derivation of the sensible heat flux. The data of Fig.2 are consistent with the findings of Holtslag and Van Ulden (1983) and Holtslag and De Bruin (1988). In addition the latter authors showed that B^{-1} might be much larger for small wind speed cases in nighttime conditions.

4.3 The latent heat flux

For the estimation of the latent heat flux λE, we have to know z_{oQ} (see Eq.(6)). This quantity can in principle be derived from $(q_o - q_s)/ q_*$ in a similar way as above for B^{-1}, where q_o is the humidity at z_{oM}. However, no direct observations of q_s are available and general q_s is not equal to the saturation value at temperature T_s. This means that we have to specify two quantities to estimate λE, namely z_{oQ} and q_s. Because of the analogy between heat and moisture transfer we take $z_{oQ} = z_{oM}$ (Brutsaert, 1982).

Instead of specifying q_s, we will use the Penman-Monteith equation (10) and the Priestley-Taylor formula (11), which means that r_c and α have to be known. Fig. 3, taken from De Bruin and Holtslag (1982), shows a variation of r_c and α for typical summertime conditions in the Netherlands. Also a relatively dry period is indicated. For the "normal" period representive values are $r_c = 60$ s/m and $\alpha = 1.1$. For the dry period we have $r_c = 160$ s/m and $\alpha = 0.8$ (as shown by dashed lines in Fig.3).

We have tested the Penman-Monteith equation (10) with $r_c = 60$ s/m and $z_{oQ} = z_{oH} = 4.5\ 10^{-5}\ z_{oM}$ (section 4.2), both for independent Cabauw data and the MESOGERS data (Figs. 4a and 4b). In Figs. 5a and 5b a similar

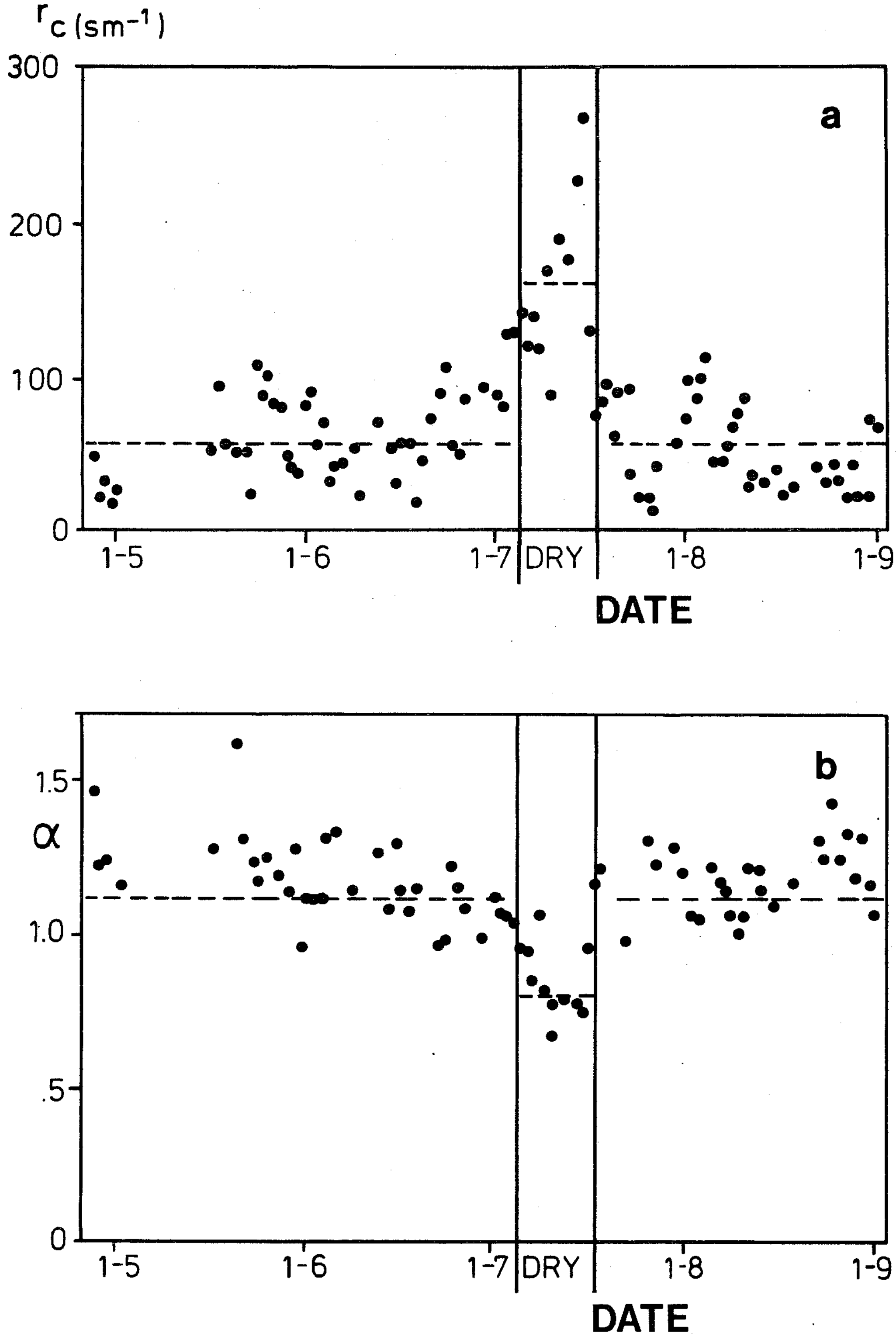

Fig.3 Variation of stomatal resistance r_c (Fig. 3a) and Priestley-Taylor parameter α (Fig. 3b) for the summer of 1977 in Cabauw (De Bruin and Holtslag, 1982).

comparison is given for the Priestley-Taylor formula (11). The observed values of λE are derived with a Bowen-ratio method, where the Bowen-ratio $Bo = H/\lambda E$ is estimated from turbulence measurements (MESOGERS) and profile measurements (Cabauw). As such, λE follows from observations of $(Q^* - G)$ and Bo with

$$\lambda E = \frac{(Q^* - G)}{1 + Bo} .$$

The models also use observations of $Q^* - G$. Therefore the comparisons of Figs. 4 and 5 show how well the fluxes are modelled with Eqs. (10) and (11) for given values of $Q^* - G$.

The results of Figs. 4 and 5 indicate that the estimates of λE for Cabauw and MESOGERS are reasonably good, both for the Penman-Monteith and the Priestley-Taylor approach. For the MESOGERS data the latter is somewhat surprising, since these data include cases of dry air advection. Similar results as shown here for λE, can be given for the sensible heat flux H. Generally, the relatively good estimates for λE (and H) can be explained by the fact that both the Cabauw and MESOGERS surroundings are well provided with water.

5. A COMPARISON OF EXCHANGE COEFFICIENTS FOR MOMENTUM AND HEAT IN STABLE AND VERY STABLE CONDITIONS

In weather and climate models the exchange coefficients for momentum and heat are usually specified as (e.g. Louis, 1979)

$$C_M = C_{MN}\, F_M\, (Ri_o,\ z_1/z_{oM}) \qquad (13)$$

$$C_H = C_{HN}\, F_H\, (Ri_o,\ z_1/z_{oH}) \qquad (14)$$

$$C_Q = C_{QN}\, F_Q\, (Ri_o,\ z_1/z_{oQ}) . \qquad (15)$$

Here C_{MN}, C_{HN}, and C_{QN} are the neutral exchange coefficients, which follow from (1) - (6) for neutral stability ($L = \infty$); F_M, F_H and F_Q are dimensionless functions of the specified quantities. Ri_o is the bulk

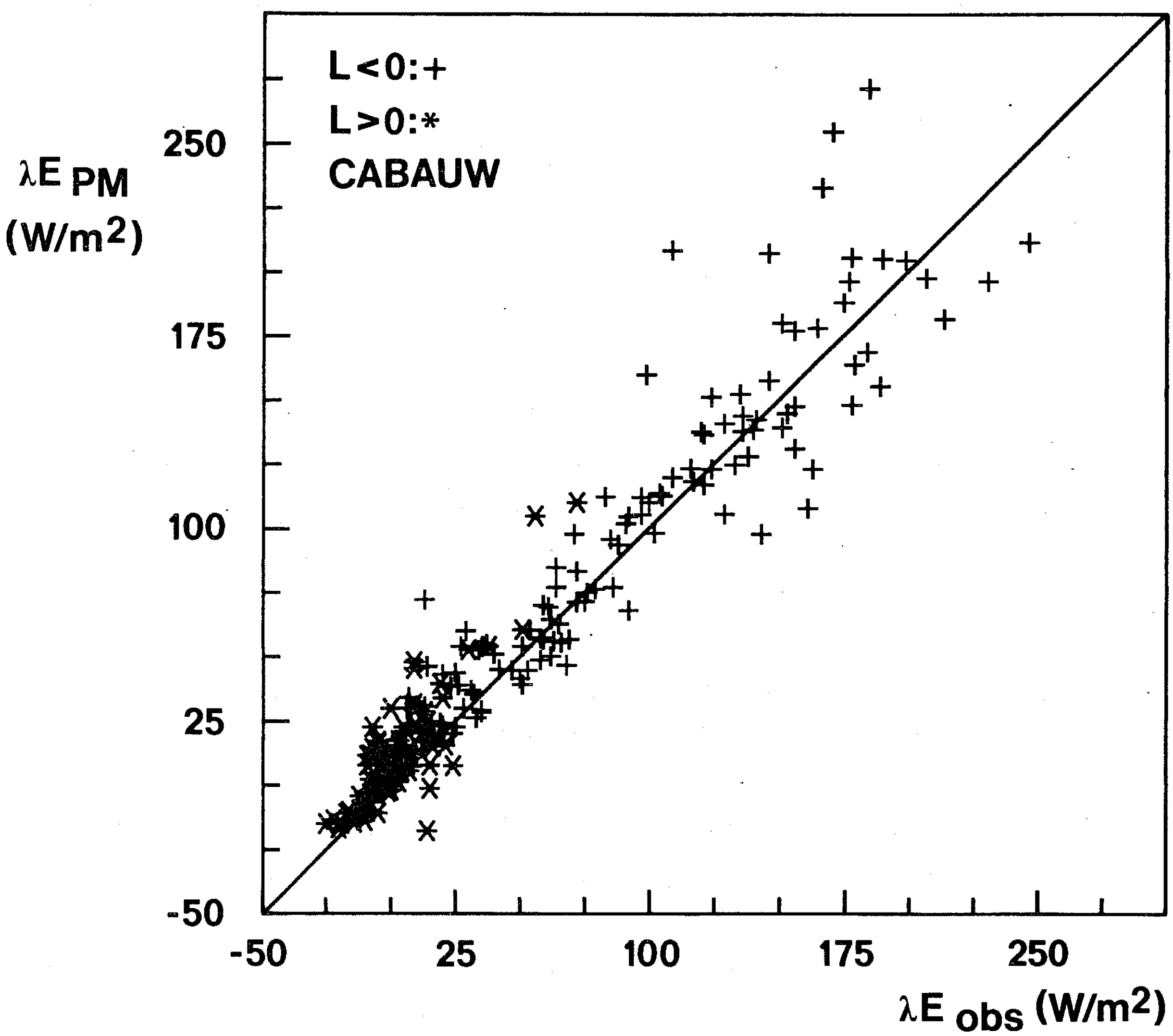

Fig.4a Estimates of latent heat flux λE_{PM} with Eq.(10) and λE derived from observations with Bowen's ratio method (λE_{obs}). The Cabauw data of June, 1986 are used.

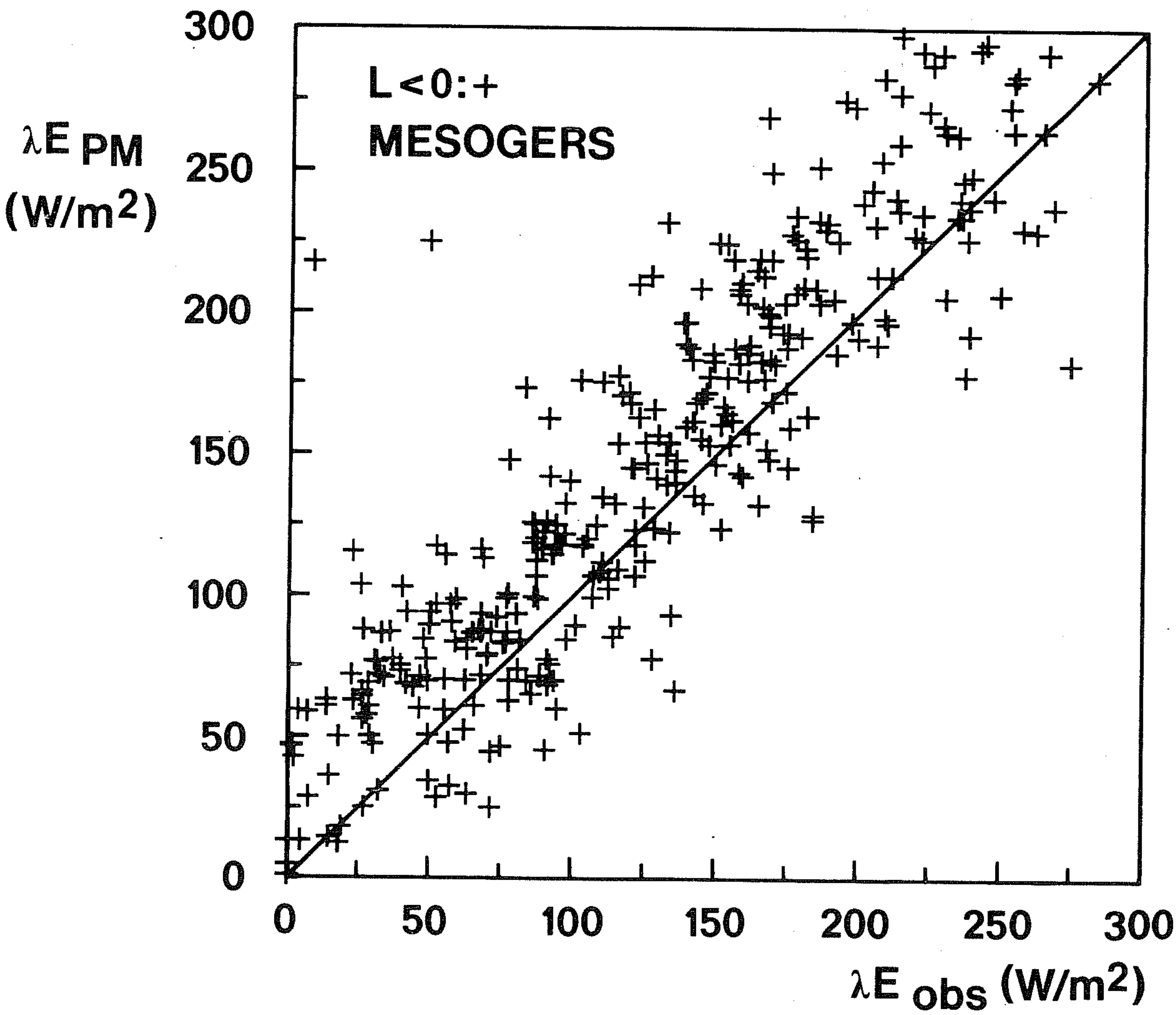

Fig.4b As Fig. 4a for the MESOGERS data (daytime only).

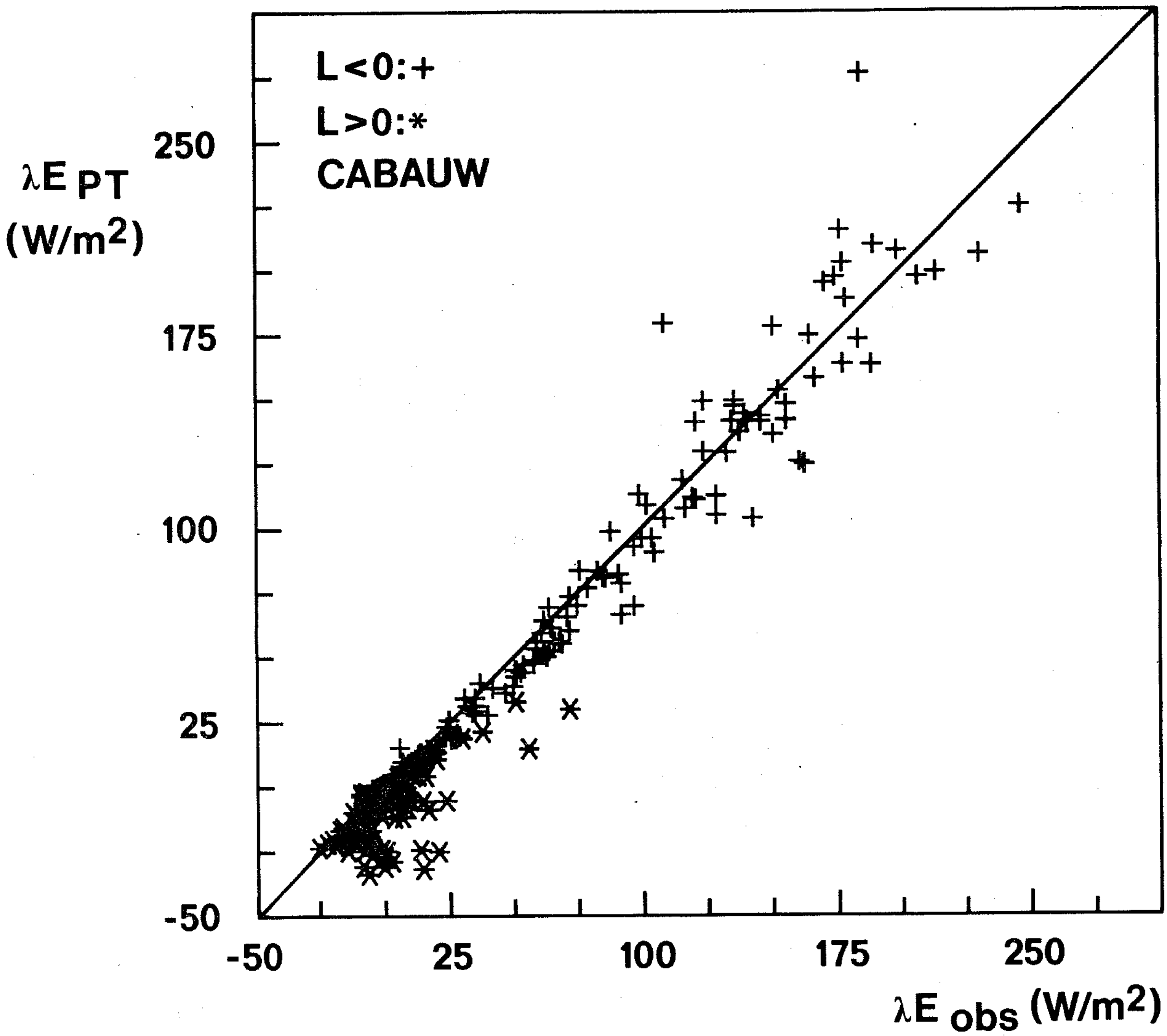

Fig.5a Estimates of latent heat flux λE_{PT} with Eq.(11) and λE derived from observations with Bowen's ratio method (λE_{obs}). The Cabauw data of June, 1986 are used.

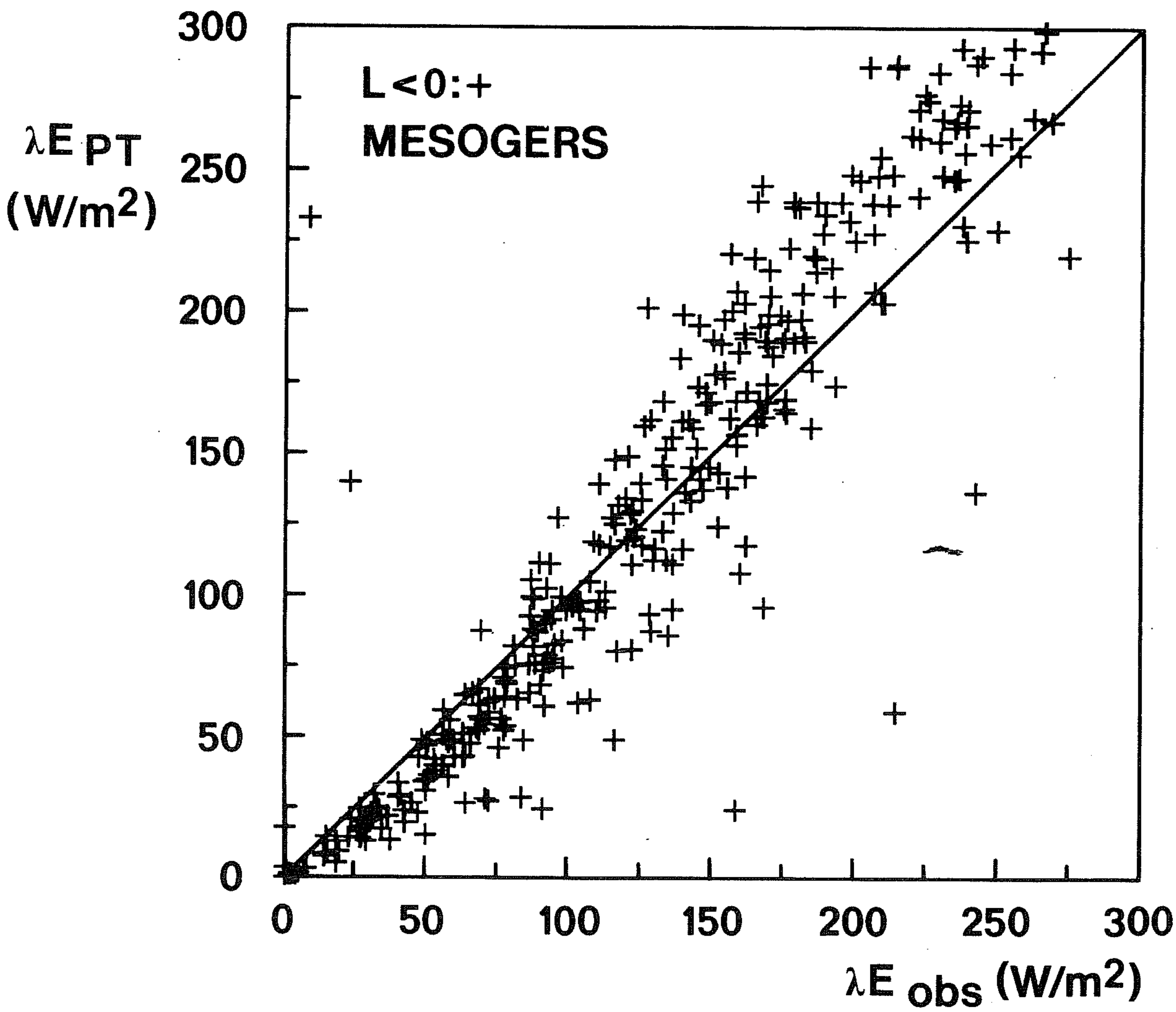

Fig.5b As Fig.5a for the MESOGERS data (daytime only).

Richardson-number between the surface and the first model layer and represents the influence of stability as does L in Eqs. (4) - (6). The advantage of (13) - (15) is that no iteration is needed to compute the surface fluxes, once Ri_o is known. Usually Ri_o of the preceeding time step is taken.

Louis (1979) and Louis et al. (1982) present several formulations for F_M, F_H and F_Q, where it is assumed that $F_Q = F_H$ and $z_{oQ} = z_{oH} = z_{oM}$. In that case Ri_o can be written as

$$Ri_o = \frac{g}{\theta} \frac{(\theta_1 - \theta_0)}{U_1{}^2} z_1, \tag{16}$$

where θ_1, θ_o and U_1 have the same meaning as in Eqs. (1), (3) and (12). For unstable conditions the proposals by the above authors are similar in both papers, whereas for stable conditions the formulae are quite different, especially for F_M. This is illustrated in Figs. 6 and 7. For comparison also two other formulations are given in these figures. One of the formulations is based on Eqs. (1), (2), (4) and (5), where

$$\psi_M = -5 \frac{z}{L} \tag{17}$$

and

$$\psi_H = \psi_M \tag{18}$$

is taken, as originally proposed by Webb (1970). It is seen that Eqs. (17) and (18) result in exchange coefficients which vanish at $Ri_o = 0.2$. Louis (1979) noted that this produces unrealistic results within a weather prediction model, since in such cases "the ground becomes energetically disconnected from the atmosphere and starts cooling by radiation at a faster rate than is actually observed. As such unrealistic nighttime ground temperatures are produced". We can explain this by the fact that (17) is only valid for $z/L \leq 0.5$, as has been confirmed by observations (Hicks, 1976; Holtslag, 1984).

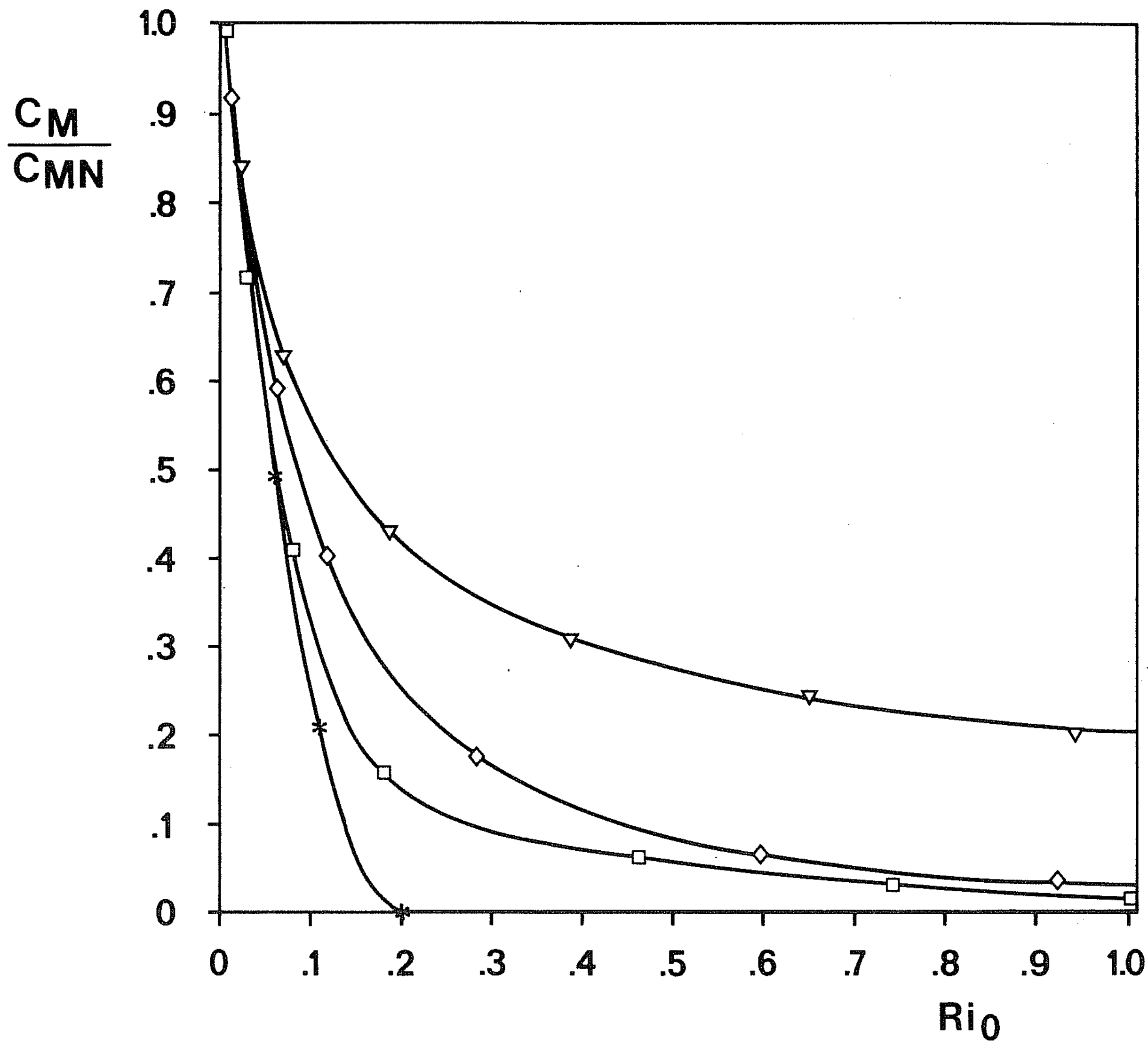

Fig.6 Normalised drag coefficient for momentum C_M/C_{MN} as a function of Richardson number Ri_o for stable conditions according to

1) Louis (1979) : marked with "rhombs"
2) Louis et al. (1982): "triangles"
3) the results of Webb (1970): "stars"
4) the present results, where $z_1/z_{oM} = 400$ is used: "squares"

Hicks (1976) analysed stable wind profiles above flat homogeneous terrain, and showed that (17) is not longer applicable. Holtslag (1984) used the results of Hicks in combination with an effective roughness lenth, for the description of stable wind profiles in Cabauw up to $z/L = 10$. Recently, Holtslag and De Bruin (1988) showed that a similar profile function is a good descriptor for nighttime temperature profiles up to $z/L \approx 7$. They confirmed the equality of ψ_H and ψ_M and propose

$$- \psi_M = a \frac{z}{L} + b \left(\frac{z}{L} - \frac{c}{d}\right) \exp \left(- d \frac{z}{L}\right) + \frac{bc}{d} , \qquad (19)$$

where $a = 0.7$, $b = 0.75$, $c = 5$ and $d = 0.35$.

On the basis of (18) and (19), exchange coefficients for momentum and heat can be derived, which are also shown in Figs. 6 and 7. For momentum these results indicate a much lower exchange coefficient than the one proposed by Louis et al (1982). Our findings are in agreement with the proposals of Carson and Richards (1978), which are used by Bell and Dickinson (1987) for the British Meteorological Office weather prediction models.

The exchange coefficients for heat and momentum, which are based on (18) and (19) can be approximated by

$$F_H = F_M = \frac{1}{1 + 10\, Ri_o\, (1 + 8\, Ri_o)} , \qquad (20)$$

where Ri_o is defined by (16). Here, for simplicity a dependence on z/z_{oM} is neglected, which appears in the original solutions near $Ri_o \approx 0.2$. Note that for heat an additional temperature difference $\theta_o - \theta_s$ can be taken into account by applying (12), which will influence the surface radiation temperature as discussed in section 4.

6. DISCUSSION AND SUMMARY

On the basis of observations in two data sets, a discussion is given of surface parameterizations for numerical weather and climate models. It is shown that an effective roughness length for momentum provides a good estimate of the friction velocity in Cabauw. This

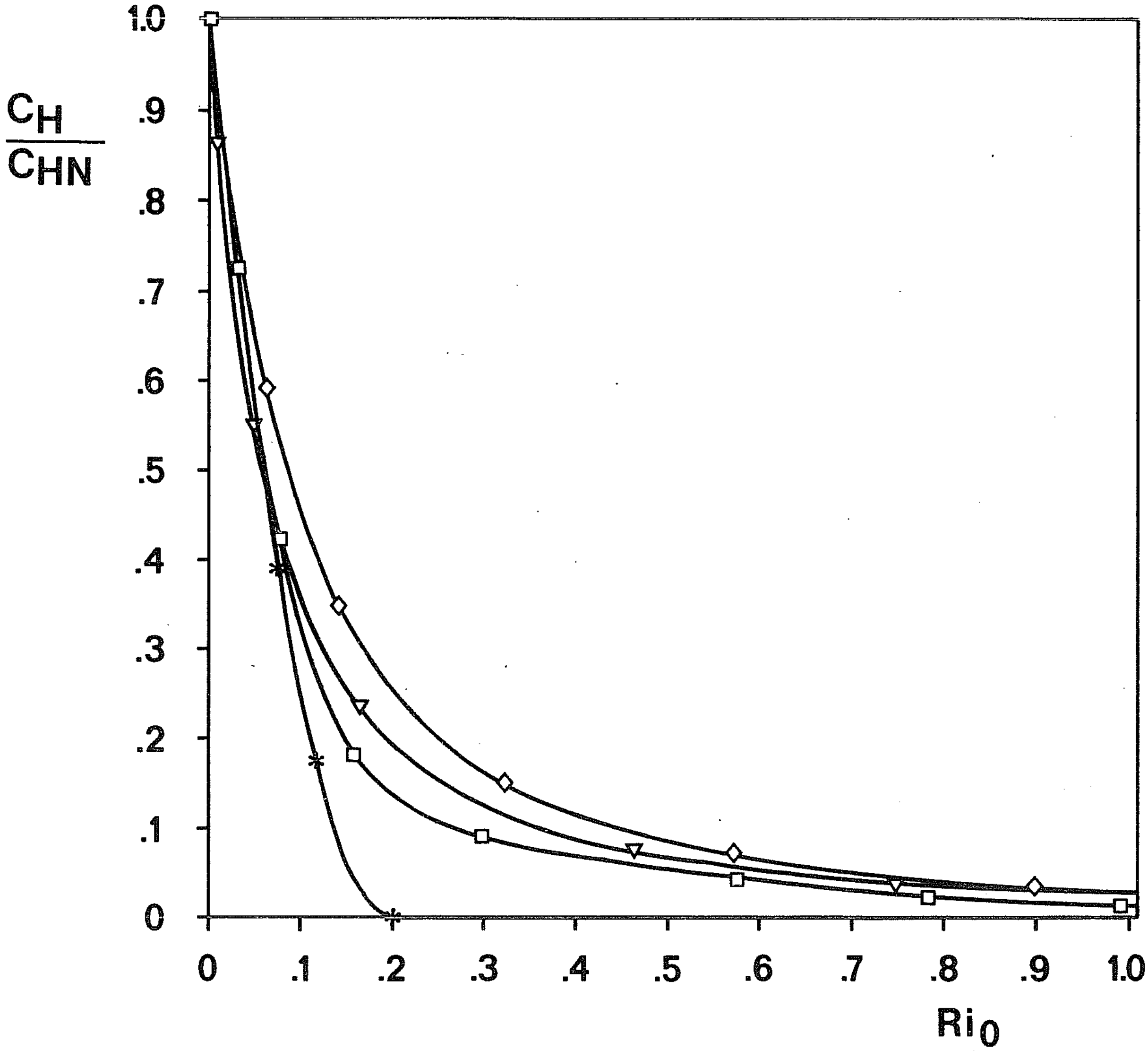

Fig. 7 As Fig. 6 for the normalised drag coefficient for heat C_H/C_{HN}.

effective roughness length is larger than the local value, and represents surface inhomogeneities in the upwind terrain on a scale of a few kilometers. The value of the effective roughness length is mainly determined by the largest roughness elements and by sparsely distributed obstacles as tree rows, houses, etcetera (See also Schmid, 1988). This problem is very similar to the problem of estimating a representive roughness length on the scale of weather and climate models (André and Blondin, 1986; Wieringa, 1986).

The surface roughness length of heat is usually taken equal to the surface roughness length of momentum. This effectively means that no distinction is made, between the surface temperature at the height of the roughness length for momentum and the surface radiation temperature. It is shown, however, that the difference between these two temperatures can be as high as 6 K. Neglecting the difference will influence the estimate for the sensible heat flux and probably the representation of the temperature profile near the surface. Based on our data we obtain roughness lengths for heat (z_{oH}) which are up to 5 orders of magnitude smaller than those for momentum (z_{oM}). This aspect is not confirmed by observations over homogeneous surface conditions, for which typically $z_{oH} \approx z_{oM}/10$ is obtained (Brutsaert, 1982).

We conjecture that this discrepancy is related to the complex terrain aspects of the Cabauw site. The momentum flux is to a large extent concentrated in sparsely distributed obstacles, whereas the heat and moisture fluxes are distributed much more uniformly over the terrain. This also implies that for heat and moisture the open terrain roughness lengths apply, rather than the effective ones. This topic obviously need more investigation; studies are needed that consider this inhomogeneous terrain problem from the smallest scales up to the gridsize scales of numerical models.

For the Cabauw and MESOGERS conditions, good estimates of the latent (and sensible) heat flux are obtained with the Penman-Monteith and Priestley-Taylor equations. These require the specification of a stomatal resistance or a related surface humidity parameter. We have shown that similar values are applicable, both for the Cabauw and

MESOGERS data.

From the comparisons of the flux estimates with observations, we conclude that relatively simple models are very useful. In general, however, this is only the case when the surface parameters are known beforehand or can be estimated with reliable methods. Even with the simple schemes discussed in this paper and using very detailed observations, we find it a difficult task to specify appropriate surface parameters (z_{oM}, z_{oH}, r_c, α).

More complicated land surface schemes (e.g. Deardorff, 1978; Sellers et al., 1986) certainly are able to describe the physics in more detail and possible more accurately. However, it is not clear whether all the parameters that specify the land surface in such models, can easily be determined. Our experience in air pollution and wind energy applications is that simple models are often preferable because they need fewer input parameters. The Priestley-Taylor formulations for instance is often preferred over the Pennman-Monteith equation, simply because it needs less input. At present it is not clear in what detail the physics of the surface processes need to be modelled in general circulation models. Pan and Mahrt (1987) present an attractive simple alternative to the more complicated approach of Sellers et al. (1986).

Finally, we have given a comparison of exchange coefficients of momentum and heat for stable and very stable conditions. The comparison is given for coefficients that are in current use and for ones which are based on our findings. It is shown that the revised exchange coefficient of momentum by Louis et al (1982), is significantly larger than the one based on our observations. Our findings are in agreement with the independent observation of Hicks (1976) and the recommendations by Carson and Richards (1978).

In our analysis we assume that all the inhomogeneities can be accounted for by choosing an appropriate effective roughness length. Mahrt (1987) argues that inhomogeneities in stability (which can be due to roughness changes as well), result in enhanced exchange coefficients. For the Cabauw site, however, with a slightly inhomogeneous surface, we feel that the exchange coefficients for homogeneous terrain do quite well as long as we make sure that the area averaged momentum flux is reproduced by the model.

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References

André, J.C. and C. Blondin, 1986: On the effective roughness length for use in numerical three-dimensional models. Bound.-Layer Meteor., 35, 231-245.

Beljaars, A.C.M., 1982: The derivation of fluxes from profiles in perturbed areas. Bound.-Layer Meteor., 24, 35-55.

Beljaars, A.C.M., 1987: The measurement of gustiness at routine wind stations - A review. WMO Instruments and observing methods; report no. 31, Genève. Also as KNMI scientific report WR-87-11, De Bilt.

Beljaars, A.C.M., 1988a: Surface fluxes in moderately complex terrain. AMS, Proc. 8th Symp. Turb. Diff. Conf., 193-196, San Diego, Ca.

Beljaars, A.C.M., 1988b: The observation of wind gustiness at routine wind stations. WMO Techn. Conf. Instr. Obs., 311-316, TECO-1988, Leipzig, DDR.

Beljaars, A.C.M., P. Schotanus and F.T.M. Nieuwstadt, 1983: Surface layer similarity under non-uniform fetch conditions. J. Clim. Appl. Meteor., 22, 1800-1810.

Bell, R.S. and A. Dickinson, 1987: The Meteorological Office operational numerical weather prediction system. Sci. Paper, 41, Meteor. Office, Bracknell.

Brutsaert, W.H., 1982: Evaporation into the atmosphere. D. Reidel Publishing Company, Dordrecht, Holland.

Carson, D.J. and P.J.R. Richards, 1978: Modelling surface turbulent fluxes in stable conditions. Bound.-Layer Meteor., 14, 67-81

Davenport, A.G., 1960: Rationale for determining design wind velocities, J. Struct. Div. Am. Soc. Civ. Eng., 86, 39-68.

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

De Bruin, H.A.R. and A.A.M. Holtslag, 1982: A simple parameterization of the surface fluxes of sensible and latent heat during daytime compared with the Penman-Monteith concept. J. Appl. Meteor., 21, 1610-1621.

De Bruin, H.A.R., 1983: A model for the Priestley-Taylor parameter α J. Clim. Appl. Meteor., 22, 572-578.

Driedonks A.G.M., H. van Dop and W. Kohsiek, 1978: Meteorological Observations on the 213 m mast at Cabauw in the Netherlands. Proc. 4th Symp. Meteor. Observ. and Instrum., April 1978, Denver, U.S.A., AMS Boston, p.41-46.

Durand, P., A. Druilhet, T. Hedde and B. Bénech, 1987: Aircraft observations of the structure of the boundary layer over a rugged hilly region (MESOGERS 84 experiment). Ann. Geophys., 5b, 441-450.

Fiedler, F. and H.A. Panofsky, 1972: The geostrophic drag coefficient and the effective roughness length. Quart. J. Roy. Meteor. Soc., 98, 213-220.

Garratt, J.R. and B.B. Hicks, 1973: Momentum, heat and water vapour transfer to and from natural and artificial surfaces. Quart. J. Roy. Meteor. Soc., 99, 680-687.

Gryning, S.E., A.A.M. Holtslag, J.S. Irwin and B. Sivertsen, 1987: Applied dispersion modelling based on meteorological scaling parameters. Atmos. Environm., 21, 79-89.

Hicks, B.B., 1976: Wind profile relationships from the "Wangara" experiment. Quart. J. Roy. Meteor. Soc., 102, 535-551.

Holtslag, A.A.M., 1984: Estimates of diabatic wind speed profiles from near surface weather observations. Bound.-Layer Meteor., 29 225-250.

Holtslag, A.A.M. and A.P. van Ulden, 1983: A simple scheme for daytime estimates of the surface fluxes from routine weather data. J. Clim. Appl. Meteor., 22, 517-529.

Holtslag, A.A.M. and F.T.M. Nieuwstadt, 1986: Scaling the atmospheric boundary layer. Bound.-Layer Meteor., 36, 201-209.

Holtslag, A.A.M. and H.A.R. de Bruin, 1988: Applied modelling of the nighttime surface energy balance over land. J. Appl. Meteor., 27, 689-704.

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

Louis, J.F., M. Tiedtke and J.F. Geleyn, 1982: A short history of the PBL parameterization at ECMWF. Workshop on Boundary Layer Parameterization, 25-27 nov. 1981, E.C.M.W.F., Reading.

Mahrt, L., 1987: Grid-averaged surface fluxes. Mon. Wea. Rev., 115, 1550-1560.

Monna, W.A.A. and J.G. van der Vliet, 1987: Facilities for research and weather observations on the 213 m tower at Cabauw and at remote locations. Sci. Rep. WR 87-5, KNMI, De Bilt.

Monteith, J.L., 1981: Evaporation and surface temperature. Quart. J. Roy. Meteor. Soc., 107, 1-27.

Pan, H.L. and L. Mahrt, 1987: Interaction between soil hydrology and boundary-layer development. Bound.-Layer Meteor., 38, 185-202.

Priestley, C.H.B. and R.J. Taylor, 1972: On the assessment of surface heat flux and evaporation using large scale parameters. Mon. Wea. Rev., 100, 81-92.

Reiff, J., D. Blaauboer, H.A.R. de Bruin, A.P. van Ulden and G. Cats, 1984: An air mass transformation model for short-range weather forecasting. Mon. Wea. Rev., 112, 393-412.

Schmid, H.P.E., 1988: Spatial scales of sensible heat flux variability; Representativeness of flux measurements and surface layer structure over suburban terrain. Ph.D. thesis University of British Colombia, Vancouver, Canada.

Sellers, P., Y. Mintz, Y.C. Sud and A. Dalcher, 1986: A simple biosphere model (SIB) for use within general circulation models. J. Atmos. Sci., 43, 505-531.

Taylor, P.A., 1987: Comments and further analysis on the effective roughness lengths for use in numerical three-dimensional models. Bound.-Layer Meteor., 39, 403-418.

Van Dop, 1983: Terrain classification and derived meteorological parameters for interregional transport models. Atmos. Environm., 17, 1099-1105.

Van Ulden, A.P. and Holtslag, A.A.M., 1985: Estimation atmospheric boundary layer parameters for diffusion applications. J. Clim. Appl. Meteor., 24, 1196-1207.

Webb, E.K., 1970: Profile relationships: The log-linear range and extension to strong stability. Quart. J. Roy. Meteor. Soc., 96, 67-90.

Weill, A., C. Mazaudier, F. Baudin, C. Klapisz, F. Leca, M. Masmoudi, D. Vidal Madjar, R. Bernard, O. Taconet, B.S. Gera, A. Sauvaget, A. Druilhet, P. Durand, J.Y. Caneil, P. Mery, G. Dubosclard, A.C.M. Beljaars, W.A.A. Monna, J.G. van der Vliet, M. Crochet, D. Thomson, and T. Carlson, 1988: The "MESOGERS 84" experiment: A report. Bound.-Layer Meteor., 42, 251-264.

Wieringa, J., 1981: Estimation of mesoscale and local-scale roughness for atmosperic transport modeling. Air Pollution Modeling and its Application, Plenum, New York, 295-297.

Wieringa, J., 1986: Roughnes-dependent geographical interpolation of surface wind speed averages. Quart. J. Roy. Meteor. Soc., 112, 867-889.

Yaglom, A.M. 1977: Comments on wind and temperature flux-profile relationships. Bound.-Layer Meteor., 11, 89-102.