Boundary-layer parameterization for Finnish regulatory dispersion models

A. Karppinen, S. M. Joffre and P. Vaajama

Finnish Meteorological Institute, Air Quality Research, Sahaajankatu 20 E, FIN-00810 Helsinki, Finland

Abstract: This paper presents the boundary-layer parameterization model which is applied as a meteorological pre-processor by Finnish regulatory dispersion models. The parameterization scheme is based on the energy-flux method, which evaluates the turbulent heat and momentum fluxes in the boundary layer. We have compared the model predictions with those of the corresponding scheme of Berkowicz and Prahm, using the same synoptic input data. The comparison shows the basic physical differences between these two models. Some numerical results from these two pre-processors differ substantially, while the results for net radiation are statistically nearly identical.

Keywords: boundary layer, energy flux, parameterization, pre-processor.

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1 Description of the FMI meteorological pre-processor

The boundary-layer (BL) parameterization model of the Finnish Meteorological Institute (FMI) is based on the method originally developed by van Ulden and Holtslag (1985).

1.1 Basic scaling parameters

The first scaling parameter is the friction velocity, which determines the shear production of turbulence kinetic energy at the surface. According to surface-layer similarity theory, the measured wind speed at the height z, $U(z)$, can be written as:

$$u_* = kU(z) / \left[\ln(z / z_0) - \psi_m(z / L) + \psi_m(z_0 / L)\right] \tag{1}$$

where z_0 is the roughness parameter, k is von Karman constant ($k \approx 0.4$), L is the Monin–Obukhov length scale, and the stability function ψ_m is modelled as (van Ulden and Holtslag, 1985):

$$\begin{aligned} \psi_m &= (1 - 16z / L)^{1/4} - 1 && \text{for } L < 0 \\ \psi_m &= -17\left[1 - \exp(-0.29z / L)\right] && \text{for } L > 0 \end{aligned}$$

The turbulent heat flux at the ground surface, H_0, and the friction velocity determine the second parameter, the temperature scale for turbulent heat transfer:

$$\theta_* = -H_0 / \left(\rho c_p u_*\right)$$

where ρ is the air density and c_p is the specific heat capacity of air.

The third BL scaling parameter is the Monin-Obukhov length scale L, defined as

$$L = T_2 u_*^2 / kg\theta_* \tag{3}$$

where T_2 is the air temperature at a height of 2 m and g is the acceleration due to gravity.

1.2 The surface heat flux

The basic parameter governing the stability and turbulence production in the boundary layer is the surface turbulent heat flux, H_0. In a stationary and horizontally homogeneous boundary layer, the balance of the turbulent energy at the ground surface is given by (van Ulden and Holtslag, 1985):

$$H_0 + \lambda E_0 = Q^* - G \tag{4}$$

where λE_0 is the latent heat flux (λ is the latent heat of vaporization of water and E is the evaporation rate, Q^* is the net radiation flux, and G is the conductive heat flux to the ground. The latent heat flux is estimated by applying the linearized saturation curve and the surface energy budget:

$$\lambda E_0 = \alpha\left(\frac{S}{S+1}\left(Q^* - G\right) + \rho c_p u_* \theta_d\right) \tag{5}$$

where S is the slope of the saturation enthalpy curve and θ_d is an empirical temperature scale. For S we use the equation $S = 51371.0 \times \exp(5423.0\,(1/273.16 - 1/T_2)/T_2$, and for θ_d a constant value, $\theta_d = 0.033$ K (van Ulden and Holtslag, 1985). Table 1 shows the parameter α as a function of the surface moisture.

Table 1 Values of the Priestley–Taylor parameter α in the FMI model. Notation: rr is the precipitation for the last 12 hours (mm), S_U is the synoptic code for the state of ground, W_w, W_1 and W_2 are synoptic codes for the present weather, weather of the previous hour and weather of the previous 3 hours, respectively.

α	Synoptic measurements
1.0	night-time OR ($S_U > 4$ or $S_U = 2$) OR ($49 < W_w < 100$)
0.9	$19 < W_w < 30$ AND none of the above holds
0.8	$4 < W_2 < 10$ AND none of the above holds
0.7	$4 < W_1 < 10$ AND none of the above holds
0.6	rr > 5.0 AND none of the above holds
0.5	none of the above holds

Using the equations above, the temperature scale θ_* can be written as

$$\theta_* = \left(\frac{\alpha S}{S+1} - 1\right)\left(Q^* - G\right) / \rho c_p u_* + \alpha\theta_d \tag{6}$$

1.3 The surface radiation budget

Equation 4 includes the fluxes Q^* and G, which are not routinely observed. In order to parameterize the radiation budget, we decompose the net radiation as:

$$Q^* = Q_i^* + \Delta Q^* \tag{7}$$

where Q_i^* is the radiation flux in an isothermal atmosphere and ΔQ^* is the deviation from this in the real atmosphere. The deviation is evaluated in terms of the air temperature T_t at a reference height $z_r = 50$ m, and the surface radiation temperature T_0:

$$\Delta Q^* = 4\sigma T_2^3 (T_r - T_0) \tag{8}$$

where σ is the Stefan–Boltzmann constant (5.67×10^{-8} W m^{-2} K^{-4}).

The net radiation for an isothermal atmosphere, Q_i^* is evaluated from the net shortwave radiation K^*, L_{net} and L_c the incoming longwave radiation from clouds. L_{net} is the difference of the incoming longwave radiation L_{in} from the atmosphere and the outgoing longwave radiation L_{out} ($L_{net} = L_{in} - L_{out}$):

$$Q_i^* = K^* + L_{net} + L_c \tag{9}$$

At high latitudes, the net radiation at the surface correlates better with sunshine duration than with cloud cover. The net shortwave radiation K^* is therefore modelled as a function of the incoming shortwave radiation flux at the surface and the albedo:

$$K^* = (1 - r) R_S(K_R) \tag{10}$$

where r is the surface albedo and R_S is a function of the observed hourly sunshine duration and K_R.

The clear-sky radiation part (K_R) can be estimated as a function of the solar elevation angle ϕ:

$$K_R = I_0\big(0.987 - 0.0909(\sin\phi + 0.118)\big)\sin\phi \tag{11}$$

where I_0 is the solar constant (1372 W m^{-2}).

The effect of clouds is described by a regression equation, based on data of hourly solar radiation and sunshine from the Jokioinen observatory in Southern Finland. The equation is

$$R_s = K_R \frac{0.66R_{SS}^2 - 0.92R_{SS} - 0.043}{R_{SS}^2 - 1.19R_{SS} - 0.15} - 0.001K_R^2 \frac{0.33R_{SS}^2 - 0.38R_{SS} - 0.007}{R_{SS}^2 - 1.61R_{SS} - 0.073} \tag{12}$$

where R_{SS} is the observed hourly sunshine duration. The empirical constants refer to a ten-year period (1965–1974).

The outgoing longwave radiation L_{up} from the surface is a function of the surface radiation temperature T_0. In an isothermal atmosphere, T_0 can be replaced by a temperature at a reference height T_r, and L_{net} can therefore be written as

$$L_{net} = \sigma T_r^4 \left(c T_r^2 - 1\right) \tag{13}$$

where $c = 9.35 \times 10^{-6}$ is an empirical constant.

The contribution to the radiation budget due to the incoming longwave radiation from clouds is parameterized by a regression equation. The radiation from clouds L'_c is defined by

$$L_c' = C_c \sigma T_c^4 \left(1 - c T_2^2\right) \tag{14}$$

where the cloud-base temperature T_c is evaluated assuming an adiabatic lapse-rate between the surface and the cloud base: $T_c = T_2 + \Gamma_d z_c$, where $\Gamma_d = -0.01$ °C/m and $Z_c$is the cloud-base height, derived from observations. The total amount of dominant clouds, C_c, is the combined amount of low and medium-altitude clouds.

The contribution of high clouds to the surface radiation budget is considered negligible, particularly when there are several cloud layers. It has been assumed above that, in the presence of low clouds, the boundary layer is nearly neutrally stratified and, in contrast to the shortwave radiation, the longwave radiation is not a function of the temperature difference $T_r - T_0$.

To account for the annual cycle in albedo, the regression method was applied separately for winter and summer. The following regression equations were obtained for Finnish conditions, based upon the ten-year period 1965–1974 from the Jokioinen radiation records.

For the period during which the ground is not covered by snow,

$$L_c = 0.64 L_c' \tag{15a}$$

and, for the snow-covered period,

$$L_c = 0.56 L_c' \tag{15b}$$

The ground heat flux G is evaluated as a function of the difference between the air temperature at the reference height and the surface radiation temperature:

$$G = -A_G (T_r - T_0) \tag{16}$$

where A_G is an empirical constant (5 W m^{-2} K^{-1}).

There is no routine observations of the radiation temperature T_0 However, the value of ($Q^* - G$) can be computed from Equations 4–6, simultaneously with Equation 16 involving the other BL parameters.

For an unstable boundary layer ($L < 0$), we use the fact that ΔQ^* is strongly correlated with Q^*:

$$\Delta Q^* = -A_H Q^* \tag{17}$$

where A_H is an empirical heating coefficient estimated from 10-year radiation measurements at Jokioinen observatory.

From Equations 7, 8, 16 and 17 it follows that

$$Q^* - G = \frac{Q_i^*}{1 + A_H}\left(1 - \frac{A_G A_H}{4\sigma T_2^3}\right) \tag{18}$$

For a stable boundary layer ($L < 0$) it follows from Equations 7 and 8 that

$$Q^* - G = Q_i^* + \left(4\sigma T_2^3 + A_G\right)\left(T_r - T_0\right) \tag{19a}$$

The temperature difference, which is strongly affected by wind speed, is estimated from the potential temperature profile equation:

$$T_r - T_0 = \theta_*\left(30 + 5 z_r / (kL)\right) - \Gamma_d z_r \tag{19b}$$

where $z_{0h} = 0.03$ cm, for the temperature roughness height for heat, and $z_r = 50$ m have been used.

This completes the parameterization for the difference ($Q^* - G$). We have derived an estimate for the temperature scale θ_* as a function of the friction velocity u_* and the Monin–Obukhov length L. Using Equations 1 and 3, the parameters L, u_*, and θ_* are solved by an iterative procedure.

2 Comparison of the FMI and Berkowicz and Prahm models

2.1 *Net radiation estimates*

The FMI method divides the net radiation into three parts: the shortwave radiation from the Sun, blackbody radiation from clouds and the ground, and the longwave radiation of the isothermal atmosphere. The shortwave radiation is estimated by a regression equation, which uses observed hourly sunshine duration as the regression model variable. The radiation from clouds is modelled by another regression equation, which uses the total cloudiness and the cloud height as parameters.

The method of Berkowicz and Prahm (1982) (BP) uses two regression models, one for daytime and one for night-time, which apply synoptic measurements of cloudiness as the most important variable.

The net radiation estimates of these two schemes are shown in Figure 1. The difference increases as the solar elevation increases under daytime clear sky conditions and as temperature decreases under night-time conditions.

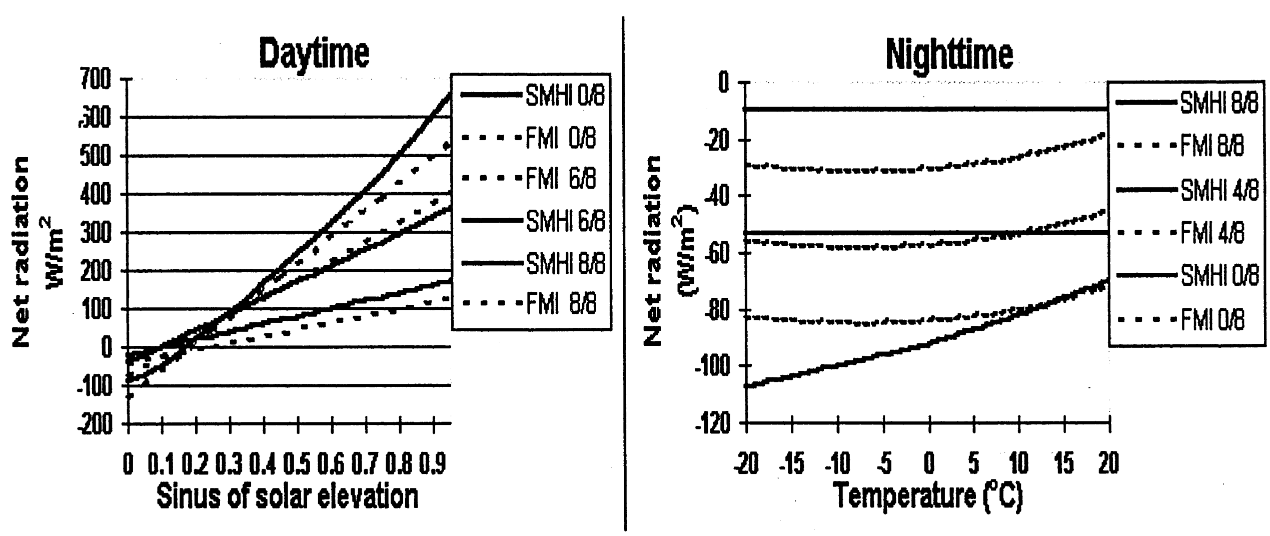

Figure 1 The net radiation estimates computed by the FMI and SMHI models. The key indicates the model and the cloudiness in octas.

2.2 *Partitioning of the energy*

The methods have one basic difference in their energy partitioning schemes. The BP model evaluates the resistances r_a and r_c and the humidity deficit in a fairly complicated way. The FMI model uses the modified Priestley–Taylor model (van Ulden and Holtslag, 1985), which divides the evaporation into two components, one correlated with ($Q^* - G$) and a non-correlated component. In the FMI model, only two empirical parameters have therefore to be evaluated.

The surface and soil moisture are also estimated differently in the models. The FMI model uses synoptic weather codes and the rain amount to estimate the surface moisture, and the Priestley–Taylor parameter α. The BP model uses the accumulated net radiation as a measure of the soil moisture. The estimates are therefore not directly comparable.

2.3 *Statistical comparison of the outputs of the two models*

We have compared the predictions of these two models, using identical synoptic data as input values.

The Danish OML meteorological pre-processor (Olesen and Brown, 1992) was used in the comparison. The basis of this model is essentially identical with the model used in the SMHI in Sweden. We selected a 10-year period from 1983 to 1993 for the comparison; however, the preliminary results presented here cover only the year 1983. The meteorological data were collected from southern Finland, and were interpolated from 3-hourly measurements to hourly values.

Figure 2 shows the quantile-quantile distributions of the net radiation, as estimated by the OML and FMI models. The distributions agree very well, except for a slight difference for net radiation values in the range from 200 to 400 W m^{-2}. Figure 3 presents the turbulent heat flux estimates of these two models; there are substantial differences between them.

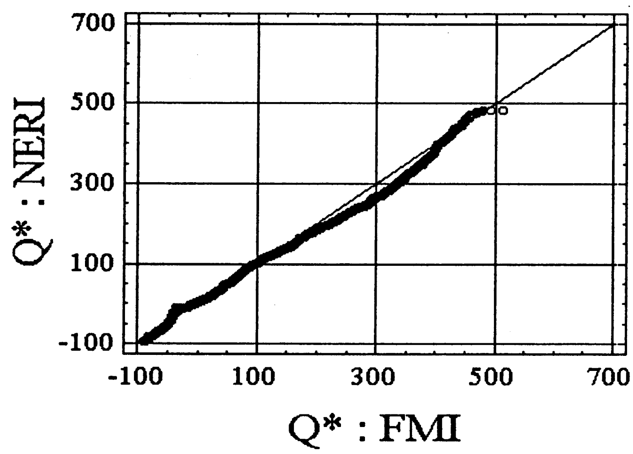

Figure 2 The quantile–quantile plot of the net radiation estimates of the FMI and OML models.

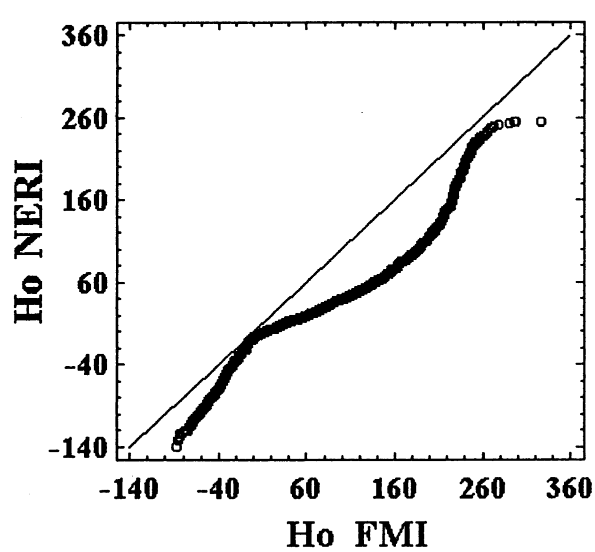

Figure 3 The quantile–quantile plot of estimates of the sensible heat flux.

The ratio of stable to unstable situations as evaluated by the two models is almost the same. In stable conditions, the OML model produces more negative energy flux values than the FMI model. In unstable conditions, the FMI model produces larger turbulent heat flux values than the OML model. The results indicate that the two parameterization schemes divide the available energy between the latent and sensible heat fluxes differently.

Figure 4 shows a comparison of the computed Monin–Obukhov lengths. There is a clear difference between the two curves: the OML model impedes the growth of $1/L$ in stable conditions. This cut-off does not exist in the FMI model, which may produce extremely stable conditions. This is an important difference concerning air quality analyses, as air pollution episodes are often connected with extremely stable situations.

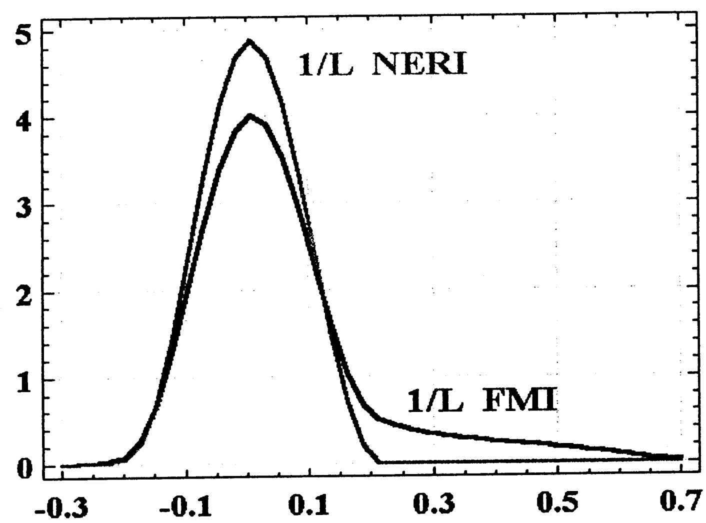

Figure 4 The density distribution of the inverse Monin–Obukhov lengths.

3 Conclusions

We have described the basic mathematical structure of the FMI pre-processor, and compared model predictions with those of the corresponding model applied in Sweden and Denmark. The basic physical ideas of these pre-processors are similar, but the measurements required by the models are somewhat different.

Only one year of synoptic data has been used for the statistical analysis, and the results are therefore preliminary. In future work, the complete data of the 10-year period included in this study will be analysed.

The modelling of the net radiation differs substantially in the two models, but the numerical results are nearly the same. The partitioning schemes for the turbulent sensible heat and the latent heat are clearly different in the models, and the numerical results also differ substantially.

The comparison of the parameterization schemes with measurements is in progress. The measurement campaigns described by Walden *et al.* (1995) and Härkönen *et al.* (1996) have produced data that will be used in these comparisons.

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