U.S. Department of Commerce
National Oceanic and Atmospheric Administration
National Weather Service
National Centers for Environmental Prediction
5830 University Research Court
College Park, MD 20740-3818

Office Note 497
https://doi.org/10.25923/9qej-k604

The Surface Layer Parameterization in the NMM Models

Zavisa Janjic

NOAA/NWS/NCEP Environmental Modeling Center

College Park, Maryland

May 2019*

* published posthumously, manuscript first drafted in February 2013

1. Introductory overview

Since sufficient vertical resolution has been reached over the operationally used integration domains, the well established and almost universally accepted similarity theory (Monin and Obukhov, 1954) has been adopted replacing the early approach used in the Eta model (Janjic, 1990). The first attempt to apply the similarity theory with the step-mountains was made by Lobocki (1993). The parameterization described here was developed by Janjic (1996a, 1996b).

2. The similarity theory

a. Notation and definitions

In order to calculate the surface fluxes, the similarity theory requires that boundary conditions be prescribed at two levels in the air, z_1 and z_2. The turbulent fluxes are assumed to be constant between these two levels. The values of the relevant variables at the lowest model level height z_{lm} above the surface are usually used as the upper boundary conditions.

The turbulent fluxes for momentum M, heat H, buoyancy H_v and humidity E are, respectively, traditionally written in the flux-gradient form

$$M = -<u'w'> = K_M \frac{dU}{dz}, \; H = -<\Theta'w'> = K_H \frac{d\Theta}{dz},$$

(1)

$$H_v = -<\Theta_v'w'> = K_H \frac{d\Theta_v}{dz}, \; E = -<q'w'> = K_H \frac{dq}{dz}.$$

Eqs.(1) can be rewritten as

$$\frac{dU}{dz} = \frac{M}{K_M}, \; \frac{d\Theta}{dz} = \frac{H}{K_H},$$

(2)

$$\frac{d\Theta_v}{dz} = \frac{H_v}{K_H}, \; \frac{dq}{dz} = \frac{E}{K_H}.$$

In (1) and (2), $<...>$ denotes ensemble averaging, the primes indicate the deviations from the mean values, K_M and K_H are the exchange coefficients for momentum and heat, respectively, and the subscript v indicates that the virtual potential temperature is used. The scales defined by

$$u_* = M^{1/2}\,,\ \Theta_* = \frac{H}{u_*}\,,\ \Theta_{v*} = \frac{H_v}{u_*}\,,\ E_* = \frac{E}{u_*} \tag{3}$$

are also traditionally used.

b. Finite difference form of the fluxes in the surface layer

If F is one of the fluxes (2), S is the corresponding variable that is being transported by turbulence, U, Θ, Θ_v or q, and K_F is the exchange coefficient K_M or K_H, after integration of a formula of the form (2) from height z_1 to height z_2 taking into account that the fluxes are constant with height, one obtains

$$S_2 - S_1 = F\int_{z_1}^{z_2} \frac{dz}{K_F}\ . \tag{4}$$

Defining the bulk exchange coefficient K_{Fbulk} by

$$\frac{z_2 - z_1}{K_{Fbulk}} = \int_{z_1}^{z_2} \frac{dz}{K_F}, \tag{5}$$

one may write the fluxes (1) in the finite difference form

$$F = K_{Fbulk}\frac{S_2 - S_1}{z_2 - z_1}. \tag{6}$$

c. Vertical profiles and the Obukhov length

According to the similarity theory, within the surface layer,

$$\frac{\partial S}{\partial z} = \frac{S_*}{kz}\varphi_F(\zeta) \tag{7}$$

Here,

$$S_* = \frac{F}{M^{1/2}} = \frac{F}{u_*} \tag{8}$$

is a scale of type (3), φ_F are empirically or otherwise determined functions,

$$\zeta = \frac{z}{L} \tag{9}$$

is the nondimensional combination of geometric height and the Obukhov length scale

$$L = \frac{M^{3/2}}{k\beta g H_v}, \tag{10}$$

where

$$\beta = \frac{1}{\Theta_{lm}},$$

Θ_{lm} is the potential temperature at the lowest model level, k is the von Karman constant and g is gravity.

Note that if ζ tends to zero, $\varphi_F(0) = const$ (typically close to 1), and the integration of (7) with respect to z leads to the familiar log profile of S between the heights z_1 and z_2. As can be seen from the definitions (9) and (10), ζ tends to zero either for neutral stratification, or when the distance from the surface z tends to zero for nonzero L. Thus, when the surface is approached, the vertical profiles of all variables tend to assume the logarithmic form.

Upon integration of (7) from height z_1 to height z_2 within the surface layer,

$$S_2 - S_1 = \int_{z_1}^{z_2} \frac{S_*}{kz} \varphi_F(z) dz. \tag{11}$$

Note that the assumption about constant fluxes implies that the fluxes F and the scales S_* can be freely moved in and out of the integrand on the rhs of (11). Thus, (11) can be rewritten as

$$S_2 - S_1 = \frac{S_*}{k} \int_{z_1}^{z_2} \varphi_F(z) \frac{L}{z} \frac{dz}{L},$$

or utilizing (8) and (9)

$$S_2 - S_1 = \frac{F}{ku_*} \int_{z_1}^{z_2} \varphi_F(z) \frac{d\zeta}{\zeta}.$$

The neutral case, when L tends to infinity and ζ tends to zero is a singular point and requires a special treatment. Therefore,

$$S_2 - S_1 = \frac{F}{ku_*}\int_{z_1}^{z_2}[\varphi_F(z) - \varphi_F(0)]\frac{d\zeta}{\zeta} + \varphi_F(0)\frac{d\zeta}{\zeta}$$

and

$$S_2 - S_1 = \frac{F}{ku_*}\int_{z_1}^{z_2}[\varphi_F(z) - \varphi_F(0)]\frac{d\zeta}{\zeta} + \frac{F}{ku_*}\int_{z_1}^{z_2}\varphi_F(0)d(\ln z). \tag{12}$$

Since the function $\varphi_F(z)$ is known, the integrals on the rhs of (12) can be evaluated, yielding

$$\Phi_F = \Psi_F(\zeta_2) - \Psi_F(\zeta_1) + \varphi_F(0)\ln\left(\frac{z_2}{z_1}\right). \tag{13}$$

Thus, (13) can be rewritten in the form

$$S_2 - S_1 = \frac{F}{ku_*}\Phi_F\,. \tag{14}$$

The integral functions Ψ_F in (13) [and therefore Φ_F in (14)] are known either in the analytical, or in the tabular form, and several sets of such functions based on different measurements are in use.

In particular, for $F = M$, $S = U$, $\Phi_F = \Phi_M$, from (14)

$$U_2 - U_1 = \frac{M}{ku_*}\Phi_M\,,$$

and

$$U_2 - U_1 = \frac{u_*}{k}\Phi_M\,. \tag{15}$$

If z_1, z_2, U_1, U_2 and the Obukhov length L are known, u_*, and consequently the momentum flux M, can be readily obtained from (15). After that, any other flux F can be computed from (14).

d. Computation of Obukhov length

The problem is that L is typically not known, and that (14) is highly implicit. The Obukhov length L is sometimes (c.f., e.g., Lobocki, 1993) estimated by establishing an approximate relationship of the form

$$L = L\left(Ri_{bulk}, z_2 - z_{S_1}, z_2 - z_{S_2}, ..., \ln\frac{z_2}{z_{S_1}}, \ln\frac{z_2}{z_{S_2}}, ... \right),$$

where Ri_{bulk} is the bulk Richardson number computed using finite differences, and z_{S_1}, z_{S_2}, … are the heights at which the lower boundary conditions are specified for the variables S_1, S_2 ... . However, this approach becomes impractical for more than one z_S.

Here, the iterative approach is chosen for its flexibility with respect to various modifications and refinements (e.g., various stability functions, using more than one z_S). With this approach, the first guess momentum flux M^0 is used together with the first guess flux H_v^0 in order to specify the first guess Obukhov length L^0. The first guess fluxes can be computed using the first guess bulk exchange coefficients K_{Mbulk}^0 and K_{Hbulk}^0 in the finite difference bulk formula (6). In addition, note that substituting (6) into (14) one obtains

$$S_2 - S_1 = K_{Fbulk} \frac{S_2 - S_1}{z_2 - z_1} \frac{\Phi_F}{ku_*}$$

and solving this equation for u_*,

$$M^{1/2} = u_* = \frac{\Phi_F}{k} \frac{K_{Fbulk}}{z_2 - z_1}. \tag{16}$$

Thus, given the first guess Obukhov length L_0, the improved, first iteration bulk exchange coefficients are obtained from

$$\frac{K_{Fbulk}^1}{z_2 - z_1} = \frac{k}{\Phi_F} u_*^0,$$

where, as before, the superscript 0 denotes the first guess values, and the superscript 1 denotes the value after the first iteration. Then the procedure is repeated until the convergence is reached, i.e., until the fluxes do not change any more with new iterations. Good first guess bulk exchange coefficients can be obtained by using the values at the end of the iterative procedure in the previous model time step. The described procedure is used in the model and can be summarized as follows

Momentum flux:

$$M^i = K^i_{Mbulk} \frac{U_2 - U_1}{z_2 - z_1}$$

(Over water, the difference $U_2 - U_1$ is limited to $35\, ms^{-1}$),

Buoyancy flux:

$$H^i_v = K^i_{Hbulk} \frac{\Theta_{v2} - \Theta_{v1}}{z_2 - z_{\Theta 1}},$$

Obukhov length:

$$L^i = \frac{\left(M^i\right)^{3/2}}{k\beta g H^i_v}$$

Integral stability functions for momentum and buoyancy:

$$\Phi^i_M = \Phi_M\left(L^i\right),\ \Phi^i_H = \Phi_H\left(L^i\right),$$

Updated bulk exchange coefficients:

$$\frac{K^{i+1}_{Mbulk}}{z_2 - z_1} = \frac{k\left(M^i\right)^{1/2}}{\Phi^i_M},$$

(17)

$$\frac{K^{i+1}_{Hbulk}}{z_2 - z_{\Theta 1}} = \frac{k\left(M^i\right)^{1/2}}{\Phi^i_H},$$

where i denotes the iteration count. Even though there is no formal guarantee, as a rule, this procedure converges, and the fluxes are computed with satisfactory accuracy typically with no more than 3 iterations.

e. Stability functions limits and Beljaars correction

Different stability functions φ_F (and therefore Ψ_F) can be used over land and over water. The range of non-vanishing turbulent fluxes on the stable side can be extended by imposing an upper limit on ζ. Whenever ζ reaches or exceeds the prescribed upper limit value, it is set to this prescribed value. This may be needed in order to warm the land surface by the air overnight and thus prevent unrealistic plummeting of the skin temperature. On the other hand, note that this procedure can lead to unrealistically large fluxes in the case of strong stability.

On the other side of the stability range, the Beljaars (1994) correction is applied in order to avoid the singularity in the case of free convection. With this correction, a fraction of the surface buoyancy flux H_v is converted into the kinetic energy of the near surface wind induced by the large eddies so that the friction velocity u_*, and therefore the Obukhov length (10) remain nonzero. The fraction of the surface buoyancy flux converted into the kinetic energy is assumed to be

$$U_B^2 = (\gamma w_*)^2 \tag{18}$$

where γ is an empirical constant, and the scale w_* is defined by

$$w_* = (\beta g H_v H_{PBL})^{1/3},$$

where H_{PBL} is the PBL height (or a constant on the order of typical PBL height). Beljaars (1994) suggested the value $\gamma = 1.2$. In practical applications, the only modification required is to add U_B calculated from (18) to the wind speed at the upper boundary of the surface layer.

Since the stability functions are known only for a certain stability range, a lower boundary is imposed on ζ as well. This value is reached if the Beljaars correction is not sufficient to prevent ζ to do so.

3. The lower boundary condition

a. The problem

The similarity theory requires that boundary conditions be prescribed at two levels in the air, z_1 and z_2. The values of the relevant variables at the lowest model level z_{lm} above the surface are usually used as the upper boundary conditions. However, the definition of the lower boundary condition is not straightforward.

As already pointed out, independently of stratification, the profiles of the relevant atmospheric variables tend to assume the logarithmic form as the lower boundary is approached. Since the log function has a singularity for $z = 0$, it is traditionally assumed that the log profile ends at some small but finite height z_0 above the surface, and that the variable considered $(U, \Theta, \ldots)$ takes on its lower boundary value at this height. This is justified by the assumption that the values of the relevant variables in the thin layer of the air adjacent to the surface take on the surface values. This situation is schematically represented in Fig. 1a. The height z_0, is called the roughness height or roughness length. As can be seen, from (13) and (14), with all other parameters the same, the variation of z_0 can significantly affect the fluxes.

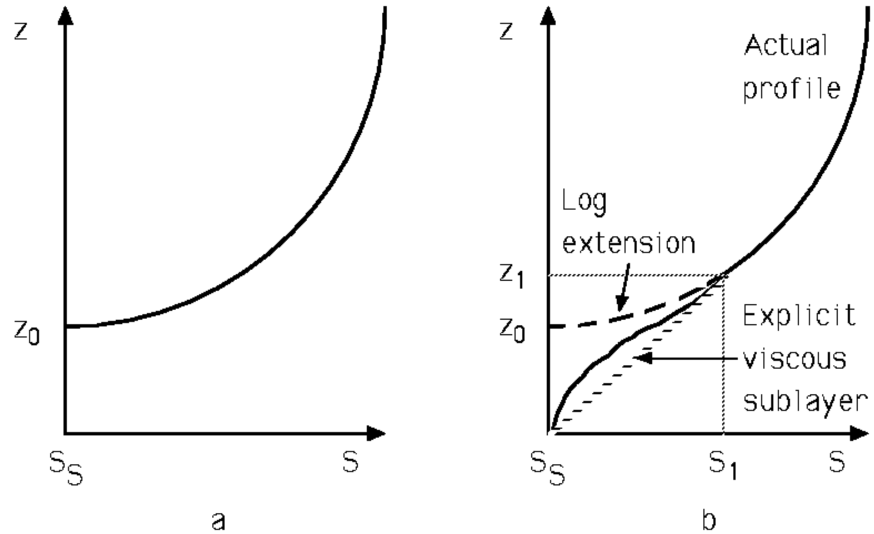

Fig. 1. Logarithmic profile ending (a), and logarithmic profile with viscous sublayer ending (b).

The situation near the surface is schematically represented in Fig. 1b. Within the thin layer of air adjacent to the surface there is not enough space for turbulent eddies to develop so that the molecular transport is dominating. This layer is often called viscous or interfacial layer. Recent experience (e.g., Janjic 1994, 1996; Chen et al. 1997) indicates that taking into account the processes in the viscous sublayer can affect significantly the surface fluxes and consequently the PBL evolution and, e.g., moist convection.

The diagram in Fig. 1b also indicates how these processes can be taken into account. The first and perhaps more popular approach is to define different z_0's for different variables, possibly dependent on flow regime. This technique is applied over land following Zilitinkevich (1995).

The second method is to match the log curve by a separate viscous sublayer profile, and to specify the height and the value of the considered variable at the matching point. By doing so, the lower boundary values for the turbulent layer denoted earlier by z_1 and S_1 would be defined. This approach is applied over water following Janjic (1994).

b. Viscous sub-layer over land

The Zilitinkevich formula is based on the following considerations. In the near-surface logarithmic range $z_{0U} << z << L$ and

$$\Theta(z) = \Theta_0 + \frac{H}{ku_*}\varphi_H(0)\ln\left(\frac{z}{z_{0U}}\right). \tag{19}$$

Here z_{0U} is the roughness length for momentum, and Θ_0 is the potential temperature extended logaritmically from the log layer downward to the level z_{0U}. Thus, Θ_0 is an integration constant, not necessarily equal to the surface value Θ_S. The problem is how to relate the roughness length z_{0U} to the roughness length for temperature z_{0T} at which the log profile reaches the surface value of the potential temperature Θ_S.

Near the surface, the momentum flux $M = u_*^2$ consists of two terms, i.e., the pressure term u_{p*}^2 and the viscous term $u_{\nu*}^2$. Close to the roughness elements the velocity scale is u_* and the velocity gradient scale is

$$\frac{u_*}{z_{0U}} \tag{20}$$

Then, the viscous contribution in the momentum flux is

$$u_{\nu*}^2 = \nu \frac{u_*}{z_{0U}}, \tag{21}$$

and the corresponding temperature scale is defined by

$$\Theta_{\nu*} = \frac{H}{k u_{\nu*}} = \frac{H}{k\sqrt{\nu \frac{u_*}{z_{0U}}}}. \tag{22}$$

The temperature increment

$$\Delta\Theta = \Theta_0 - \Theta_S \tag{23}$$

is scaled by $\Theta_{\nu*}$, i.e. the assumption is made that

$$\Delta\Theta = C_Z \Theta_{\nu*} = C_Z \frac{H}{k\sqrt{\nu \frac{u_*}{}}} \tag{24}$$

where C_Z is a constant to be determined empirically. Then, from (19) and (23), assuming that $\varphi_H(0) = 1$ as is often done,

$$\Theta(z) = \Theta_S + \Delta\Theta + \frac{H}{k u_*} \ln\left(\frac{z}{z_{0U}}\right). \tag{25}$$

Substituting (24) into (25), and rearranging one obtains

$$\Theta(z) = \Theta_S + \frac{H}{ku_*}\left[\frac{C_Z u_*}{\sqrt{\nu \frac{u_*}{z_{0U}}}} + \ln\left(\frac{z}{z_{0U}}\right)\right]$$

and

$$\Theta(z) = \Theta_S + \frac{H}{ku_*}\left[C_Z \sqrt{\frac{z_{0U} u_*}{\nu}} + \ln\left(\frac{z}{z_{0U}}\right)\right]. \tag{26}$$

Defining Reynolds number

$$Re = \frac{z_{0U} u_*}{\nu}, \tag{27}$$

(26) can be further rewritten as

$$\Theta(z) = \Theta_S + \frac{H}{ku_*}\left[C_Z Re^{1/2} + \ln\left(\frac{z}{z_{0U}}\right)\right]$$

and

$$\Theta(z) = \Theta_S + \frac{H}{ku_*}\ln\left[\frac{z \exp\left(C_Z Re^{1/2}\right)}{z_{0U}}\right],$$

or, finally,

$$\Theta(z) = \Theta_S + \frac{H}{ku_*}\ln\left[\frac{z}{z_{0T}}\right], \tag{28}$$

where

$$z_{0T} = z_{0U} \exp\left(-C_Z Re^{1/2}\right) = z_{0U} \exp\left(-C_Z \sqrt{\frac{z_{0U} u_*}{\nu}}\right). \tag{29}$$

Formula (29) is in good qualitative agreement with the experimental data by Sun and Mahrt (1995).

c. Viscous sub-layer over water

The viscous sub-layer over water surfaces (Janjic, 1994) is designed using the second of the two described approaches, the explicit one. Two simplifying modeling assumptions are made:

- There are two distinct layers: (i) a thin viscous sublayer immediately above the surface, where the vertical transports are determined entirely by the molecular diffusion (note that the equilibrium profiles within such viscous sublayer must be linear since the molecular diffusivities are assumed to be constant), and (ii) a turbulent layer above it, where the vertical transports are defined entirely by the turbulent fluxes.

- The fluxes must be continuous across the interface between the viscous and turbulent sub-layers, i.e., the fluxes at the top of the viscous sublayer must coincide with the fluxes at the bottom of the turbulent sublayer.

On the other hand, Liu et al. (1979) (hereafter referred to as LKB79), proposed the following profiles in the immediate vicinity of a smooth surface [LKB79, Eq (8)],

$$U_1 - U_S = D_1\left(1 - e^{-z_1\frac{u_*}{D_1\nu}}\right)\frac{M}{u_*} \tag{30}$$

$$\Theta_1 - \Theta_S = D_2\left(1 - e^{-z_1\frac{u_*}{D_2\chi}}\right)\frac{H}{u_*} \tag{31}$$

$$q_1 - q_S = D_3\left(1 - e^{-z_1\frac{u_*}{D_3\lambda}}\right)\frac{E}{u_*}. \tag{32}$$

Here, the subscript S denotes the surface values, the subscript 1 indicates the values at height z_1 above the surface where the molecular diffusivities are still dominant, D_1, D_2 and D_3 are parameters to be discussed later, ν, χ and λ are the molecular diffusivities for momentum, heat and water vapor, respectively, and, as before, M, H and E are the turbulent fluxes above the viscous sublayer.

For a small argument ξ of the exponential function

$$1 - e^{-\xi} \approx \xi \tag{33}$$

Thus, for small

$$\xi = -z_{1U} \frac{u_*}{D_1 \nu} = -z_{1T} \frac{u_*}{D_2 \chi} = -z_{1q} \frac{u_*}{D_3 \lambda} \tag{34}$$

(30)-(32) can be rewritten as

$$\nu \frac{U_1 - U_S}{z_{1U}} = M \ , \tag{35}$$

$$\chi \frac{\Theta_1 - \Theta_S}{z_{1T}} = H \ , \tag{36}$$

$$\lambda \frac{q_1 - q_S}{z_{1q}} = E \tag{37}$$

Here, the heights z_{1U}, z_{1T}, and z_{1q} are defined by (34), i.e.,

$$z_{1U} = \frac{\xi \nu D_1}{u_*} , \tag{38}$$

$$z_{1T} = \frac{\xi \chi D_2}{u_*} , \tag{39}$$

$$z_{1q} = \frac{\xi \lambda D_3}{u_*} . \tag{40}$$

and the values of the respective physical quantities at these heights are indicated by subscript 1 in (35)-(37).

Note that both modeling assumptions will be satisfied if the heights (38)-(40) are chosen to represent the depths of the viscous sublayers for the respective variables. The continuity of the fluxes across the interface between the entirely viscous and the entirely turbulent layers is reflected by (35)-(37). The viscous fluxes on the left hand sides of (35)-(37) are in the finite difference form, but their numerical values coincide with the values obtained from the differential formulas in the equilibrium steady state. As already pointed out, in that case the profiles of the respective variables in the viscous sublayers are linear.

Using the bulk momentum and heat exchange coefficients K_{Mbulk} and K_{Hbulk}, respectively, the turbulent fluxes in the surface layer above the viscous sublayer are represented by

$$M = K_{Mbulk} \frac{U_{lm} - U_1}{\Delta z_U} \tag{41}$$

$$H = K_{Hbulk} \frac{\Theta_{lm} - \Theta_1}{\Delta z_\Theta} \tag{42}$$

$$E = K_{Hbulk} \frac{q_{lm} - q_1}{\Delta z_q} . \tag{43}$$

Here, the subscript lm denotes the variables at the lowest model level and

$$\Delta z = z_{lm} - z_1$$

where z_1 is the height of the viscous sub-layer for the variable considered.

Combining (35)-(37) with (41)-(43), after some algebra, one obtains (Janjic, 1994)

$$U_1 = \frac{U_S + \frac{K_{Mbulk} z_{1U}}{\nu \Delta z_U} U_{lm}}{1 + \frac{K_{Mbulk} z_{1U}}{\nu \Delta z_U}} \tag{44}$$

$$\Theta_1 = \frac{\Theta_S + \frac{K_{Hbulk} z_{1T}}{\nu \Delta z_T} \Theta_{lm}}{1 + \frac{K_{Hbulk} z_{1T}}{\nu \Delta z_T}} \tag{45}$$

$$q_1 = \frac{q_S + \frac{K_{Hbulk} z_{1q}}{\nu \Delta z_q} q_{lm}}{1 + \frac{K_{Hbulk} z_{1q}}{\nu \Delta z_q}} . \tag{46}$$

Thus, the required lower boundary conditions for the turbulent layer are expressed as weighted means of the values at the surface and at the lowest model level. These boundary conditions are defined at the heights (38)-(40). Note that (44)-(46), together with (38)-(40) represent a closed system provided the parameters D_1, D_2, D_3 and ξ are known.

The viscous sub-layer over water is assumed to operate in three different regimes: (i) smooth and transitional, (ii) rough and (iii) rough with spray, depending on the roughness Reynolds number

$$Re = \frac{z_0 u_*}{\nu}. \tag{47}$$

Here,

$$z_0 = \max\left(0.018\frac{u_*^2}{g}, 1.59\times 10^{-5}\right). \tag{48}$$

When the Reynolds number exceeds a prescribed value Re_r the flow ceases to be smooth and the rough regime is entered. In the rough regime the momentum is transported also by pressure forces on the roughness elements so that (30) loses validity (LKB79). Consequently the viscous sub-layer for momentum is turned off. However, for heat and moisture, the viscous
sublayer is still operating until the rough regime with spray is reached at a critical value Re_s when the viscous sublayer collapses completely. In the rough regime with spray the breaking waves and the spray are assumed to provide much more efficient way of exchange of heat and moisture between the ocean and the air than that that can be accomplished by the molecular viscosity. Note that instead in terms of Re, the boundaries between the regimes can be also defined in terms of u_*, since Re is a monotonous function of u_*.

For the parameters D_1, D_2 and D_3 appearing in (30)-(32), LKB79 suggest [Eq (11)]

$$D_1 = G Re^{1/4}, \tag{49}$$

$$D_2 = G Re^{1/4} Pr^{1/2}, \tag{50}$$

$$D_3 = G Re^{1/4} Sc^{1/2}, \tag{51}$$

where $Pr = \nu/\chi$ is the Prandtl number, $Sc = \nu/\lambda$ is the Schmidt number, and G is a constant, but different for different regimes. With these definitions, and the definition (47), (38)-(40) take the form

$$z_{1U} = \xi\nu \frac{G\left(\frac{z_0 u_*}{\nu}\right)^{1/4}}{u_*} \tag{52}$$

$$z_{1T} = \xi\chi \frac{G\left(\frac{z_0 u_*}{\nu}\right)^{1/4} Pr^{1/2}}{u_*} \tag{53}$$

$$z_{1q} = \xi\lambda \frac{G\left(\frac{z_0 u_*}{\nu}\right)^{1/4} Sc^{1/2}}{u_*}. \tag{54}$$

For the smooth regime LKB79 used the value of G which was close to 30. When the flow ceases to be smooth they suggest the value of about 10 which fits best the Mangarella et al. (1973) data. These two values are also applied in the Eta model for the corresponding regimes. The Prandtl number and the Schmidt number were assumed to be the same, i.e., $Pr = Sc = 0.71$, and the molecular viscosity for momentum was $\nu = 0.000015$. The molecular diffusion coefficients for heat and moisture, χ and λ , are determined by ν , Pr and Sc .

Concerning the values of u_* at which the transitions between the different regimes occur, they are $u_{*r} = 0.225\, ms^{-1}$ and $u_{*s} = 0.70\, ms^{-1}$, respectively. These values qualitatively agree with the measurements. The parameter ξ is a tuning parameter in (52)-(54).

In practical implementation, the viscous sublayer variables are iterated together with the Obukhov length. As the first guess, u_* from the previous time step is used in (48) to compute the first guess z_0 . With the first guess depths z_{1U}, z_{1T} and z_{1q} calculated from (52)-(54), the first guess lower boundary conditions for the surface layer U_1 , Θ_1 and q_1 can now be obtained from (44)-(45) using K_{Mbulk} and K_{Hbulk} from the previous time step. However, in order to prevent the two-grid-interval oscillation, the average values of U_1 , Θ_1 and q_1 from the first guess and the previous time step are used. Analogous procedure is repeated in the subsequent iteration steps.

4. Illustration of performance

For historical reasons, and for easier comparison with experimental data, the surface fluxes are sometimes cast in the form

$$M = C_D |U_{10}| U_{10}$$

$$H = C_H |U_{10}| (\Theta_{10} - \Theta_S) .$$

Here, C_D and C_H are the “bulk aerodynamic coefficients” (C_D is the “drag coefficient”), the subscript 10 denotes the values at 10 m height, and S indicates the value at the lower boundary.

The equivalent bulk aerodynamic coefficients C_D and C_H over land obtained with the described surface layer scheme are shown, respectively, in Figs. 2 and 3 for the neutral stratification (empty squares) and the difference $(\Theta_{10} - \Theta_S)$ taking on the values -10^0 (diamonds) and 10^0 (filled squares) as functions of the wind speed at 10 m . The computations were made with $z_{0U} = 0.1\,m$ and synthetic stability functions of Lobocki (1993).

The analogous results for the bulk aerodynamic coefficients C_D and C_H over water surfaces are shown, respectively, in Figs. 4 and 5 for the neutral stratification (empty squares) and the difference $(\Theta_{10} - \Theta_S)$ taking on the values -5^0 (diamonds) and 5^0 (filled squares) as functions of the wind speed at $10\,m$. In the case of the stable stratification, the parts of the plots corresponding to the values of the bulk Richardson number exceeding 0.19 are missing.

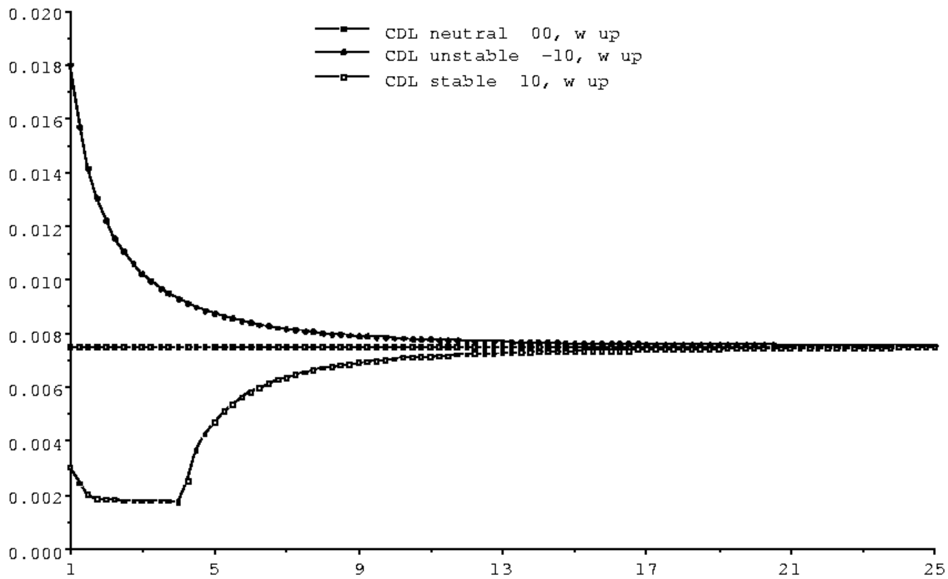

Fig. 2. C_D over land for $z_{0U} = 0.10$ m as function of wind speed.

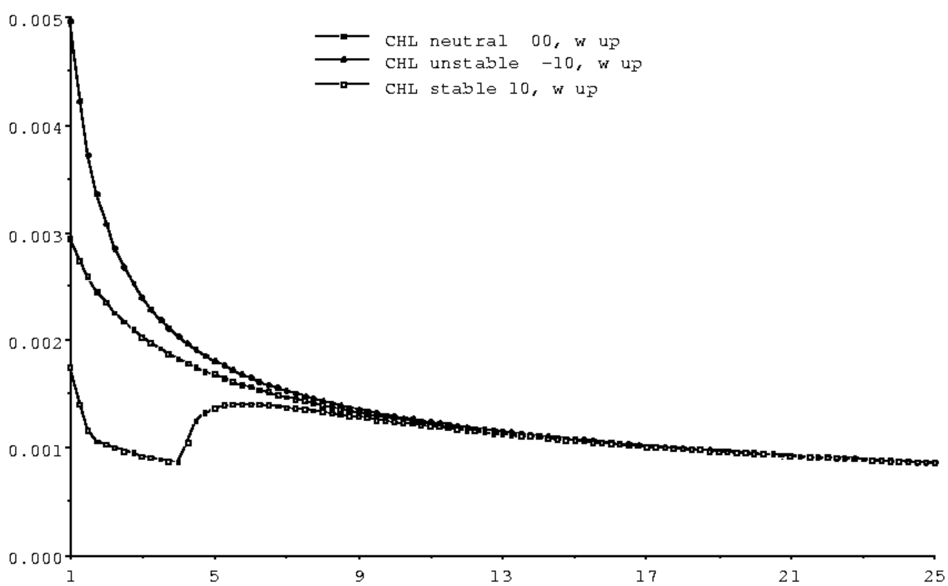

Fig. 3. C_H over land for $z_{0U} = 0.10$ m as function of wind speed.

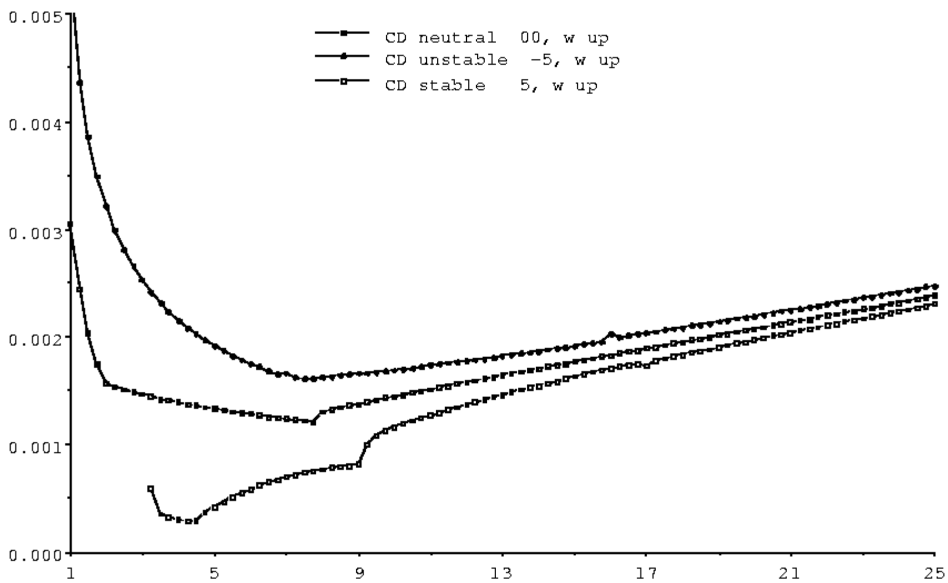

Fig. 4. C_D over water as function of wind speed.

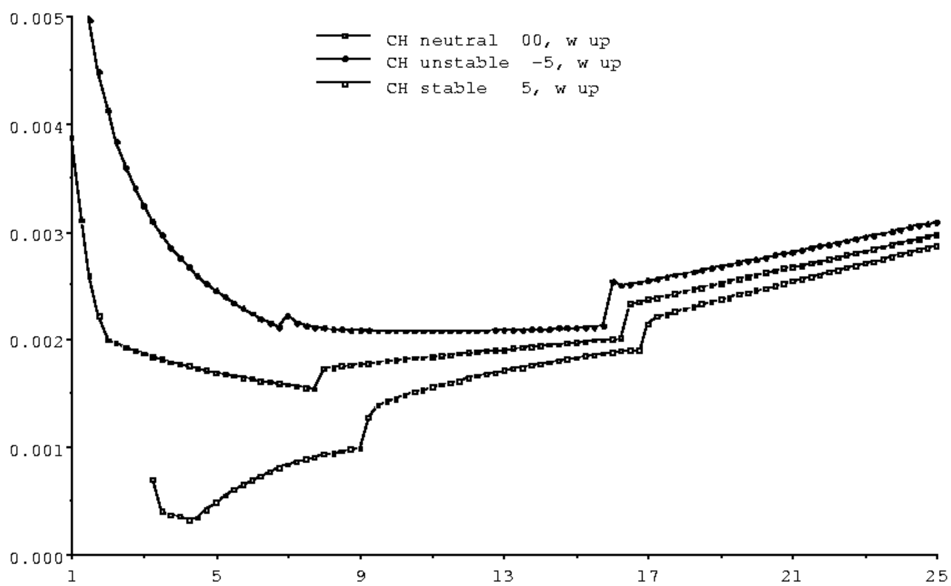

Fig. 5. C_H over water as function of wind speed.

REFERENCES

Beljaars, A.C.M., 1994: The parameterization of surface fluxes in large-scale models under free convection. *Quart. J. Roy. Meteor. Soc.*, **121**, 255-270.

Chen, F., Z. Janjic and K. Mitchell, 1997: Impact of atmospheric surface-layer parameterization in the new land-surface scheme of the NCEP mesoscale Eta model. *Boundary-Layer Meteorology,* **85**, 391-421.

Janjic, Z.I., 1990: The step-mountain coordinate: physical package. *Mon. Wea. Rev.*, **118**, 1429-1443.

Janjic, Z.I., 1994: The step-mountain eta coordinate model: further developments of the convection, viscous sublayer and turbulence closure schemes. *Mon. Wea. Rev.*, **122**, 927-945.

Janjic, Z.I., 1996a: The Surface Layer Parameterization in the NCEP Eta Model. Research Activities in Atmospheric and Oceanic Modelling, WMO, Geneva, CAS/JSC WGNE, 4.16-4.17. [Available from World Meteorological Organization, Case Postale 2300, CH-1211 Geneve 2.]

Janjic, Z.I., 1996b: The surface layer in the NCEP Eta Model. Eleventh Conference on Numerical Weather Prediction, Norfolk, VA, 19-23 August 1996; Amer. Meteor. Soc., Boston, MA, 354-355.

Liu, W.T., K.B. Katsaros and J.A. Businger, 1979: Bulk parametrization of air-sea exchanges of heat and water vapour including the molecular constraints at the interface. *J. Atmos. Sci.,* **36**, 1722-1735.

Lobocki, L., 1993: A procedure for the derivation of surface-layer bulk relationships from simplified second order closure models. *J. Appl. Meteor.*, **32**, 126-138.

Monin, A.S. and A.M. Obukhov, 1954: Basic laws of turbulent mixing in the surface layer of the atmosphere. *Contrib. Geophys. Inst. Acad. Sci. USSR*, (**151**), 163-187 (in Russian).

Sun, J., and L. Mahrt, 1995: Determination of surface fluxes from the surface radiative temperature. *J. Atmos. Sci.*, **52**, 1096-1106.

Zilitinkevich, S.S., 1995: Non-local turbulent transport: pollution dispersion aspects of coherent structure of convective flows. In: Air Pollution III - Volume I. Air Pollution Theory and Simulation (Eds. H. Power, N. Moussiopoulos and C.A. Brebbia). Computational Mechanics Publications, Southampton Boston, 53-60.