MULTI-LEVEL BOUNDARY LAYER SCHEMES

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This paper presents an updated version of some sections of earlier reviews by André (1981, 1983) and Sommeria (1984).

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1. INTRODUCTION

The Planetary Boundary Layer (PBL) is usually defined as the lowest part of the atmosphere where thermal and dynamical influences of the underlying surface (water or land) can be felt. The depth of the PBL varies by more than an order of magnitude between extreme conditions, its average maximum depth being around 1 km. The main physical feature of the PBL, which is also very often used as an alternative definition, is that it is most of the time in a turbulent regime due to the many instabilities which are induced by shear or convection close to the surface. In this paper, we shall concentrate on the turbulent PBL and on the description of turbulent fluxes using multi-level schemes. An alternative way to describe the turbulent PBL fluxes, namely the bulk approach, is described in the following paper (Driedonks, 1985).

The larger scale atmospheric flow determines, to a significant extent, the properties of the PBL. For given surface conditions it is, for example, possible to describe the microstructure of the PBL by using upper boundary conditions and pressure field deduced from larger scales. This kind of approach has been intensively studied during the last ten years. Driedonks and Tennekes (1981) call it "interpretative", in the sense that in such a method the feedback of the PBL on the large scale flow is not taken into account. This approach has nevertheless given rise to many useful concepts and has led to the development of many detailed models (see e.g. Zeman, 1981, for review).

On the other hand, the parameterization problem is unfortunately very often considered from a very restrictive, and possibly inadequate, point of view : given a larger scale flow, as described by a numerical model, what are the minimum PBL constraints which one has to introduce in order to further integrate the larger scale model ? Since the overall energetic budget must be balanced, one would then be inclined to answer that a crude description of sources (sensible and latent heat fluxes) and sinks (dissipation of kinetic energy through drag and momentum fluxes) may be enough. Although this approach may be called "interactive" (Driedonks and Tennekes, 1981), it is clear that it does not always lead to satisfactory answers.

This latter method has been used in the past mainly for two reasons : time-saving and the idea that global energetic balance would be enough to insure proper time evolution. This last concept must be studied more carefully. Based on the fact that it would take of the order of one week for surface evaporation fluxes to totally recycle atmospheric water vapor, which represents the largest reservoir of energy in the atmosphere, it is often assumed that PBL sources and sinks are really important only on the medium and long term. It should however be kept in mind that only approximately two days would be necessary for PBL kinetic energy dissipation (acting at a rate of 2 to 3 Wm^{-2}) to completely slow down atmospheric circulation (whose total kinetic energy is of the order of 100 Whm^{-2}). This shows that a detailed representation of PBL processes is required ever for short term evolution (see e.g. Anthes, 1978), specially for small and meso-scale motions which are known to be very dependent on the detailed structure of the PBL.

The large scale flow as well as the PBL have the same lower boundary, i.e. the earth's surface. Part of the parameterization problem is then to control and/or predict the surface characteristics, like its temperature and humidity as well as its roughness. We shall assume here that these properties can be known for example from the solution of a surface energy budget (see section 2 for some general indications, and the papers by Sommeria, 1985, and Sellers, 1985, for more details). The general idea on which the parameterization techniques are based is that it is possible to deduce the turbulent fluxes at the surface (momentum and heat) from the knowledge of the surface characteristics and flow properties, i.e. that the turbulence is approximately in stationary equilibrium.

Once the surface fluxes have been determined, it still remains necessary to gain some information about the way they will be vertically distributed. The meteorological variables respond indeed to the flux divergence and not to the flux itself. The altitude chosen for the lowest level in the large-scale model is then obviously of prime importance. We shall assume here that this lowest level lies within a few tens of meters from the surface, and we shall develop adapted methods to vertically distribute turbulent forcing.

2. DETERMINATION OF THE SURFACE FLUX

As stated above it is first necessary to know the surface characteristics before one can determine the surface fluxes. Among those, the roughness length z_0 is very often supposed to be a physical property of the underlying surface. It is for example possible to find in the litterature tables giving values of z_0 depending on the type of surface. A well known exception concerns the sea where the roughness length depends on the wind velocity. In such a case the Charnock's (1955) formula :

$$z_0 = M \frac{u_*^2}{g} \quad ; \quad M \sim 0.035 - 0.050 \qquad (1)$$

where $u_* = (-\overline{u'w'}_{sfc})^{1/2}$ is the friction velocity, is of rather general use. It should nevertheless be underlined that the roughness length is not a truly physical quantity, but that it may also depend on the vertical resolution of the model (André and Blondin 1985).

2.1. Surface temperature and humidity

When the underlying sŭrface is the sea, most of the time it is possible to prescribe its temperature T_S and the humidity just above as

$$q_S = q_{sat}(T_S) \tag{2}$$

Such a procedure does not allow obviously for coupling of the atmosphere with the ocean, but this could be of little importance as long as one is not interested with climatic studies.

The land surface temperature and humidity are more difficult to determine. One usually relies on the solution of the surface energy balance

$$R_N = H_S + H_L + H_{GO} \tag{3}$$

where R_N is the net radiative flux reaching the surface and H_S, H_L and H_{GO} are respectively the sensible and latent heat fluxes in the atmosphere and the heat flux penetrating in the ground. The surface atmosphere temperature T_S enters the radiative flux, through the long-wave emission $\varepsilon_G \sigma T_S^4$ leaving the surface, and the atmospheric heat fluxes H_S and H_L, see below. If the other terms in (3) are known, it is then possible to compute T_S. Such a method is nevertheless difficult to apply, mainly because the ground heat flux H_{GO} has to be determined through the (expensive) solution of the conductivity equation :

$$\rho_G c_G \frac{\partial T_G}{\partial t}(z,t) = -\frac{\partial}{\partial z} H_G(z,t) \tag{4a}$$

$$H_G(z,t) = -\mu_G \frac{\partial T_G}{\partial z}(z,t) \tag{4b}$$

where the heat capacity $\rho_G c_G$ and heat diffusivity μ_G of the soil have to be known in order to compute the temperature T_G and heat flux H_G in the ground. H_{GO} is then given by

$$H_{GO} = H_G(0,t) = -\mu_G \frac{\partial T_G}{\partial z}(0,t) \tag{5}$$

It appears to be necessary to consider at least 6 levels in the ground to obtain an accurate solution of (4).

In order to avoid such complexities, Bhumralkar (1975) proposed to integrate vertically (4a) across a small thickness δ just below the surface and to use for H_G (δ ,t) the exact solution corresponding to the 24-hour oscillation. When δ goes to zero it comes

$$\frac{\partial T_s}{\partial t} = -\frac{2}{\rho_G c_G}\left(\frac{\pi}{\nu_G \tau_1}\right)^{1/2}\left(R_N - H_s - H_L\right) - \frac{2\pi}{\tau_1}\left(T_s - \overline{T}_G\right) \qquad (6)$$

where $\nu_G = \frac{\mu_G}{\rho_G c_G}$, τ_1 = 24h, and where $\overline{T_G}$ is the mean soil temperature equal, for example, to the ground temperature averaged over the previous 24 hours.

The humidity at the surface may be determined by simply assuming it is in a constant ratio with $q_{sat}(T_s)$, or by solving the water balance in one or more layers of soil (e.g. Deardorff, 1977). This kind of method is described for example by Carson (1981). More precise and refined methods are now being developed, see Sellers (1985).

From now on we shall assume that the surface properties are known and we shall turn to the problem of computing the turbulent surface fluxes.

2.2. The bulk transfer coefficients

The turbulent surface flux ($F_x = \overline{w'x'}_{sfc}$) of a conservative quantity x can be determined from the difference between $\overline{x_1}$ and $\overline{x}_{sfc}$, where $\overline{x}_{sfc}$ is the surface value and $\overline{x}_1$ the value in the atmospheric flow given as $\overline{x_1}$ = $\overline{x}(z_1)$, z_1 being the altitude of the lowest level available in the larger scale model. By using a "drag" or "bulk transfer" coefficient C_x it comes :

$$F_x = \overline{w'x'}_{sfc} = -C_x \left|\underline{V}\right|_1 \left(\overline{x}_1 - \overline{x}_{sfc}\right) \qquad (7)$$

where $|\underline{V}|$ is the wind velocity. Eq.(7) is no more than a mathematical definition of C_x, although it bears resemblances with the well-known Boussinesq or Prandtl gradient-diffusion-type approximation. The question is now of course to estimate C_x, possibly as a function of z_1, atmospheric stability, roughness length. It is useful to specialize Eq. (7) to the three quantities of interest for meteorological purposes

$$\left|\overline{\underline{V}'w'}\right|_{sfc} = -u_*^2 = -C_D V_1^2 \qquad (8)$$

$$\overline{w'\theta'}_{sfc} = \frac{H_s}{\rho C_p} = -u_*\theta_* = -C_H |\underline{V}|_1 (\bar{\theta}_1 - T_s) \qquad (9)$$

$$\overline{w'q'}_{sfc} = \frac{H_L}{\rho L} = -u_* q_* = -C_q |\underline{V}|_1 (\bar{q}_1 - q_s) \qquad (10)$$

where ρ is the mean density of air at the surface, C_p its heat capacity at constant pressure and L the latent heat of condensation. Eq. (8) does not give any information about the direction of the surface stress. It must thus be completed by the statement that $\overline{\underline{V}'w'}_{sfc}$ is oriented along the wind direction close to the surface, i.e. in the present case at level z_1.

The simplest way to use (8) - (10) is to prescribe the value of the bulk transfer coefficients C_D, C_H and C_q. Many choices can be made, some of which are reviewed by Carson (1981). Let's simply say that the crudest, but still satisfactory, choice seems to be

$$C_D = C_H = C_Q \qquad (11)$$

with C_D of the order of $3\text{-}5 \times 10^{-3}$ over land and $1\text{-}2 \times 10^{-3}$ over sea. Some attempts have been made to include the effect of surface topography (Gates et al., 1971), or atmospheric stability (Arakawa, 1972) on a rather empirical basis. We shall not comment on these proposals but turn now to the similarity theories of the PBL, which provide an adequate framework to deal with these problems.

2.3. The Monin-Obukhov similarity theory

The Monin-Obukhov similarity theory takes full advantage of the fact that small-scale turbulence close to the earth's surface is in stationary equilibrium with respect to larger scale forcing. Above a horizontal surface the flow depends then only upon a few governing parameters, which are the friction velocity u_*, square-root of the downdard surface momentum flux $\overline{u'w'}_{sfc}$ (see Eq. (8)), the surface buoyancy flux B_{sfc},

$$B_{sfc} = \overline{w'\theta'}_{sfc} + 0.608\, T_s\, \overline{w'q'}_{sfc} \quad , \qquad (12)$$

and the buoyancy parameter $\beta = g/T_s$.

The Monin-Obukhov similarity theory predicts that (see e.g. Businger, 1973)

$$\frac{\partial \bar{u}}{\partial z} = \frac{u_*}{kz}\, \phi_u (z/L) \qquad (13)$$

$$\frac{\partial \bar{\theta}}{\partial z} = \frac{\theta_*}{kz}\, \phi_\theta (z/L) \qquad (14)$$

$$\frac{\partial \bar{q}}{\partial z} = \frac{q_*}{kz}\, \phi_q (z/L) \qquad (15)$$

where L is the Monin-Obukhov length

$$L = - \frac{u_*^3}{k \beta B_{sfc}} \qquad (16)$$

and the three "universal" functions ϕ_u, ϕ_θ and ϕ_q , which describe the influence of stability, have to be determined from experiment. These are now fairly satisfactorily known, with the possible exception of very stable situations, and have been critically reviewed by Yaglom (1977). As an example, the functions given by Businger et al. (1971) from the Kansas experiment are :

$$\phi_u(\zeta) = \begin{cases} (1-15\zeta)^{-1/4} & \text{for } \zeta \leq 0 \quad (B_{sfc} \geq 0) \\ (1+4.7\zeta) & \text{for } \zeta \geq 0 \quad (B_{sfc} \leq 0) \end{cases} \qquad (17)$$

$$\phi_\theta(\zeta) = \phi_q(\zeta) = \begin{cases} 0.74\,(1-9\zeta)^{-1/2} & \text{for } \zeta \leq 0 \quad (B_{sfc} \geq 0) \\ 0.74 + 4.7\zeta & \text{for } \zeta \geq 0 \quad (B_{sfc} \leq 0) \end{cases}$$

(18)

It should be noted that although originally Businger et al. (1971) proposed to use $k = 0.35$ for the Karman's constant, there is now growing evidence that $k = 0.41$ is a better value and that the numerical values appearing in Eqs. (17) and (18) may have to be slightly changed (see e.g. Högström, 1985).

By integrating $\phi_{\bullet}(z/L)$ with respect to z, it is possible to derive expressions relating the differences $\bar{u}_1$, $\bar{\theta}_1 - T_s$, $\bar{q}_1 - q_s$ to the turbulent surface fluxes $-u^2_*$, $\overline{w'\theta'}_{sfc}$, and $\overline{w'q'}_{sfc}$. These integrated expressions are compatible with Eqs. (8) to (10), provided $c_{\bullet}$ be chosen according to

$$C_{\bullet} = k^2 \Big/ \left\{ \int_{z_0}^{z_1} \frac{\phi_{\bullet}(z/L)}{z} \, dz \right\}^2 \qquad (19)$$

In the simple case of a neutrally-stratified PBL, i.e. for $|L| = \infty$ (see Eqs. (12) and (16)), $\phi_u(z/L)$ is equal to 1 and Eq. (19) reduces to

$$C_D = \left\{ \frac{k}{\mathrm{Ln}\,(z_1/z_0)} \right\}^2 \qquad (20)$$

Such a formulation is often used (see e.g. Manabe, 1969) in large scale models. If the lowest level is at $z_1 \sim 30$ m and if one selects $z_0 \sim 1$cm as a typical value over land, Eq. (20) leads to a value of approximately 2.5×10^{-3} for C_D, in good agreement with the above quoted values. When stability increases, Eqs. (17) and (19) show that C_D decreases, eventually going to zero for the case of very strong stable stratification.

Expressions like (19) can be solved for C_D and for surface fluxes (in term of wind, temperature, and humidity at the first level of the model) by iterative methods (e.g. Clarke, 1970). More efficient methods have been proposed (see e.g. Deardorff, 1972, or Louis, 1979), in which C_D is diagnostically expressed as a function of the gradient Richardson number R_i

$$R_i = \frac{\beta \, z_1 \, (\overline{\Theta}_1 - T_s)}{\overline{u}_1^2} \qquad (21)$$

An example of such a formulation, which corresponds to

$$C_D = \left[\frac{k}{\ln(z_1/z_0)}\right]^2 \times \begin{cases} \left\{\dfrac{1}{1+b R_i}\right\}^2 & \text{for } R_i \geq 0 \\ 1 - \dfrac{2b \, R_i}{1 + 2bc\left[\dfrac{k}{\ln(z_1/z_0)}\right]^2 \left[\dfrac{z_1}{z_0}|R_i|\right]^{1/2}} & \text{for } R_i \leq 0 \end{cases} \qquad (22)$$

with b = 4.7 and c = 7.4,

is shown in fig. 1. It can be seen that stability decreases very efficiently the turbulent drag and completely inhibits it for Richardson numbers greater than 0.4. Louis et al.(1981) have shown how modifications of the constant b appearing in Eq.(22) lead to an increase or a decrease of the momentum drag for stable situations, and can be used to "tune" the surface-layer parameterization in a numerical model.

As a final remark concerning the use of Monin-Obukhov similarity theory to retrieve surface fluxes, we shall emphasize the fact that it can only be used in models where the diurnal variation is included. It has indeed been validated experimentally only in the case where mean gradients and turbulent fluxes appearing in Eqs.(13) to (16) are computed by averaging over 10 mn or so. On the other hand, the surface heat flux averaged over one diurnal period is dominated by the relatively short afternoon period of unstable temperature gradient near the surface, while the corresponding averaged temperature gradient is dominated instead by large positive (stable) values at night. As a result, the diurnally time-averaged surface flux is counter to the corresponding time-averaged vertical gradient of

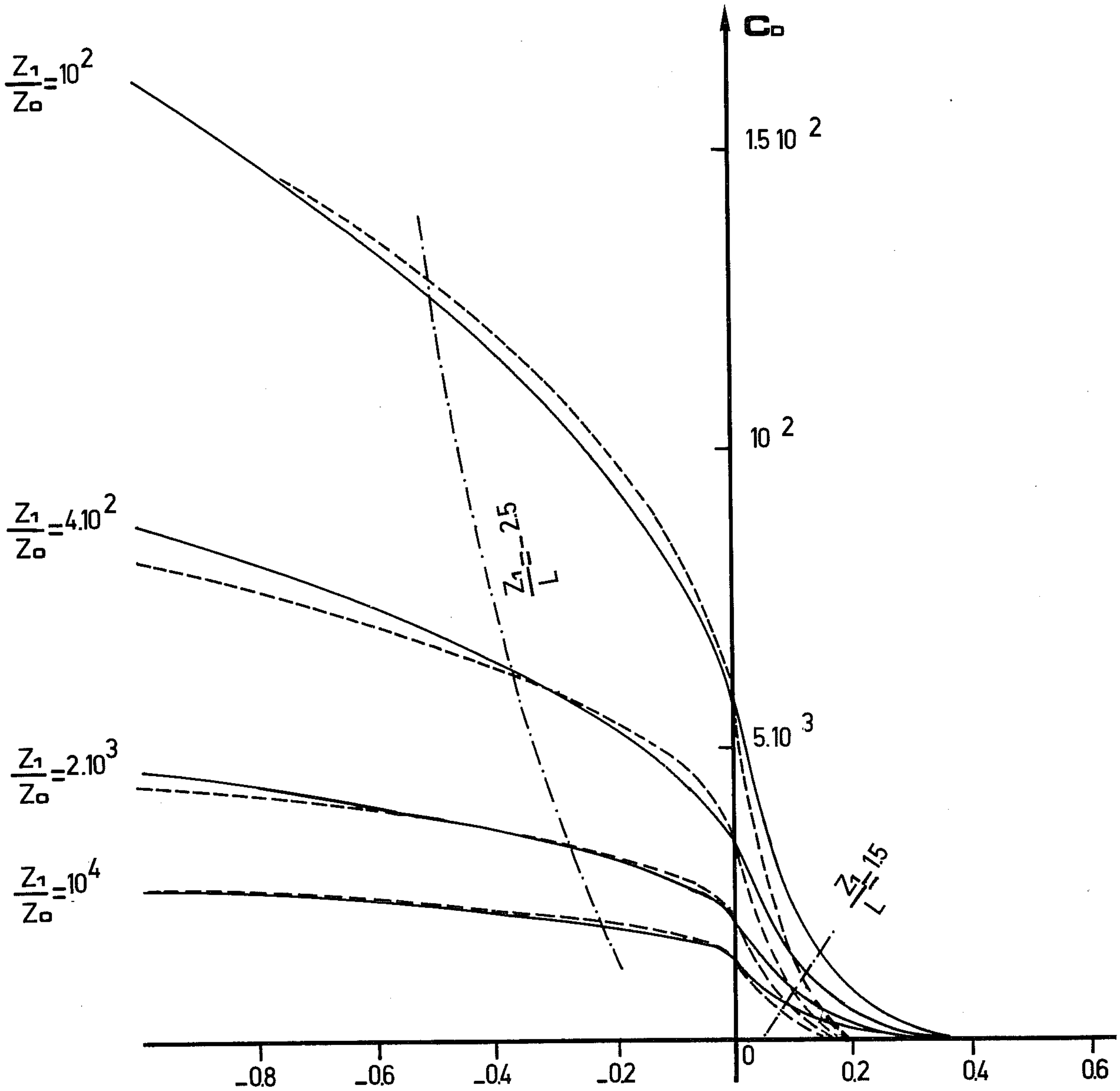

Figure 1 : Drag coefficient C_D as a function of the Richardson number R_i in the surface boundary layer following Louis (1979).

potential temperature. This counter-gradient relationship, which has been documented by Mahrt et al. (1985), prevents use of the usual Monin-Obukhov similarity theory in models without diurnal variation.

3. PARAMETERIZATION OF BOUNDARY-LAYER TURBULENT EXCHANGE

Once the turbulent surface fluxes have been determined, it is still necessary to estimate the value of the turbulent fluxes inside the PBL, in order to effectively compute their divergence, which is the quantity to be included in the rate equations in the large scale model.

It can always be assumed, at least as a definition of the eddy-diffusivity coefficient, that the turbulent fluxes $\overline{w'x'}$ can be parameterized using the Boussinesq formula, which states that they are proportional to the vertical derivative of the corresponding mean quantity $\overline{x}$:

$$\overline{w'x'} = - K_x \frac{\partial \overline{x}}{\partial z} \tag{23}$$

where eddy diffusivities K_x are by many orders of magnitude greater than molecular diffusivities.

The simplest possible choice is of course to take constant values for K., for example of the order of 10^2 - $10^3 m^2\ s^{-1}$. This is however rather crude and the next step of increased complexity and accuracy is to prescribe a vertical profile K.(z). Along those lines, O'Brien (1970) proposed to fit a cubic profile for $K_u = K_\theta$ between the value K.(z_1), computed at the lowest level z_1 from turbulent surface fluxes and near-surface vertical gradients using Eqs. (8), (13) and (23)

$$K_u(z_1) = k z_1 u_* / \phi_u(z_1/L) \tag{24}$$

and a very small value around the PBL top at z=h. It is in principle possible to either estimate h from the vertical distribution of mean quantities (e.g. as the height of the lowest inversion) or to compute it from a rate equation (see Driedonks, 1985), leading then to partly multi-level and partly bulk methods. Such methods based on the prescription of the eddy diffusivity coefficients have however an important draw back, in the sense that the turbulent transfer is almost independant of the structure and stability of the flow.

3.1. Diagnostic formulation of turbulent fluxes

The simplest way to overcome this difficulty is to make use of the Prandtl method, that is to express the eddy diffusivity as the product of an inverse turbulent time scale by the square of a mixing length

$$K \sim l^2 \left| \partial \underline{v} / \partial z \right| \tag{25}$$

In such a formulation the main question which remains to be solved concerns of course the choice of the mixing length l. From the early proposal by Blackadar (1962) it can assumed that l is of the order of a few tens of meters, i.e. of the size of efficient PBL eddies. A more precise formulation, which has the further advantage of being compatible with Monin-Obukhov similarity theory as expressed by Eq. (24), is

$$l = \frac{kz}{1 + kz / l_\infty} \tag{26}$$

where l_∞ is an "asymptotic" mixing length, usually of the order of 100m. l_∞ needs then to be either prescribed or computed, as will be discussed below.

Eq. (25) and (26) do not allow for eddy diffusivities taking stability effect into account. As this is of prime importance since stability very efficiently inhibits vertical turbulent transfer, Eq. (25) has to be modified into

$$K \sim l^2 \left| \partial \underline{v} / \partial z \right| f(R_i) \tag{27}$$

where $f(R_i)$ is a function of the local gradient Richardson number R_i evaluated between levels k and k + 1

$$R_i = \frac{\beta\left[\bar{\theta}(z_{k+1}) - \bar{\theta}(z_k)\right]\left[z_{k+1} - z_k\right]}{\left[\bar{u}(z_{k+1}) - \bar{u}(z_k)\right]^2 + \left[\bar{v}(z_{k+1}) - \bar{v}(z_k)\right]^2} \qquad (28)$$

This function has sometime been specified on purely empirical grounds (e.g. Deardorff, 1967) but is more appropriately parameterized either from simplified second-order models of turbulence (Mellor and Yamada, 1974, Sommeria, 1974) or from asymptotic arguments like in Louis (1979) (see the second factor in the r.h.s. of Eq. (22)).

3.2. The eddy kinetic energy formulation and turbulence closure schemes

The methods described in the preceding subsections lead to turbulent mixing which is in stationary equilibrium with the mean flow. It is however known from the recent developments of PBL modelling that prognostic equations for turbulence intensity are necessary if one wants to capture some of its essential features.

These remark leads to the use of an alternative formulation for $K_.$, namely

$$K_{\bullet} = a_{\bullet}\, l\, \bar{e}^{1/2} \qquad (29)$$

where e is the turbulent intensity

$$\bar{e} = (\overline{u'^2} + \overline{v'^2} + \overline{w'^2}) / 2 \qquad (30)$$

predicted from a properly parameterized form of its rate equation like

$$\left\{\frac{\partial}{\partial t} + \bar{u}\frac{\partial}{\partial x} + \bar{v}\frac{\partial}{\partial y} + \bar{w}\frac{\partial}{\partial z}\right\}\bar{e} = \frac{\partial}{\partial z}\left(K_e\frac{\partial \bar{e}}{\partial z}\right) + K_u\left[\left(\frac{\partial \bar{u}}{\partial z}\right)^2 + \left(\frac{\partial \bar{v}}{\partial z}\right)^2\right] - \beta K_\theta \frac{\partial \bar{\theta}}{\partial z} - c_\varepsilon \frac{\bar{e}^{3/2}}{l} \qquad (31)$$

In Eq. (31) the material derivative of $\overline{e}$ along the mean flow changes due to, from right to left, viscous dissipation, production by unstable stratification or destruction by stable stratification , mean shear generation and turbulent transport. The dimensionless constant c_{ε} , a_{θ} , a_u and a_e have to be determined from comparison of model results with some reference experiments. Eq. (31) together with the rate equations for the mean parameters form a closed system if the mixing length l is prescribed.

3.3. The mixing length

It is usually chosen according to Eq. (26), but where l_{∞} can either be prescribed or computed from the vertical distribution of turbulence. A first possible choice for l_{∞} is the altitude of the centre of mass of turbulence (Yamada and Mellor, 1975)

$$l_{\infty} = 0.1 \frac{\int_0^{\infty} \overline{e}^{1/2} z \, dz}{\int_0^{\infty} \overline{e}^{1/2} dz} \tag{32}$$

Such a formulation is adequate whenever turbulence has a continuous vertical distribution, for example decreasing from the surface where it may be produced by shear and convection (see Eq. (31)) to the level of the overlying inversion. However, even in such a simple case, Therry and Lacarrère (1983) have shown that one must distinguish between the mixing length responsible for vertical turbulent transfer, which appears in Eq. (29) and which is related to the vertical size of eddies, and the one responsible for viscous dissipation which appears in the last term of Eq. (31) and which is more probably related to an overall size.

In more complicated cases when there are elevated sources of turbulence, e.g. associated to shear layers or to conditionally unstable layers, it has been shown that turbulence properties and efficiency are quite dependent on the location of these different sources (André and Lacarrère, 1985). Bougeault (1985) has accordingly proposed to compute the mixing length from

$$\frac{1}{l} = \frac{1}{kz} + \frac{1}{k(z-z_i)} \qquad (33)$$

where z_i is the height of such a production layer.

4. EXAMPLES OF MULTI LEVEL SCHEMES IMPLEMENTED IN LARGE SCALE MODELS AND PERSPECTIVES

Miyako and Sirutis (1977) have been the first to implement a multi-level scheme based on the use of a rate equation for eddy-kinetic energy in a large scale numerical model. Among their preliminary, but very promising, results one should note that they were able to describe quite satisfactorily turbulent transfer and dissipation inside the boundary layer, but also close to the tropopause, where it is known that, due to strong shear and deep convection, there also exists large dissipation rates of eddy kinetic energy (Chen, 1974). In their simulation, Miyako and Sirutis used only 4 levels in the boundary layer, so that although they apparently described satisfactorily turbulent friction, they were not able to really resolve the boundary layer flow.

This point has been addressed by Anthes et al. (1980), who showed that using a multi-level model with approximately 9 levels inside the boundary layer was "essential to the correct prediction of flow within the PBL, because mass-wind adjustments in the flow over the PBL produce important changes on the pressure gradient within the PBL". Their multi-level scheme (Busch et al.,1976), which does not include a rate equation for eddy kinetic energy like Eq. (31) but uses instead a rate equation for the mixing-length l, was compared to a mixed-layer model, with the conclusion that the two models lead to comparable results under horizontally homogeneous conditions, but that the multi-level model gives better solutions as compared to the mixed-layer model under conditions with rather large horizontal inhomogeneities.

Mailhot and Benoît (1982) studied with more detail the respective accuracies of mixed-layer and multi-level models. Their multi-level model includes rate equations for both the eddy kinetic energy (Eq.(31)) and the mixing length (Busch et al., 1976).They came to the conclusion that approximately 7 levels where necessary inside the boundary layer in order to resolve the flow with accuracy better than 0.6 ms^{-1} for wind and 1.2 to 1.6K for temperature. They also showed, as confirmed later by Manton (1983) and Mahfouf (1984), that boundary-layer parameterization schemes based on a single rate equation for eddy kinetic energy can achieve PBL simulations which are in very good agreement with more detailed PBL models using many more rate equations for turbulent fluxes and variances (see also Therry and Lacarrère, 1983, and André and Lacarrère, 1985). Manton (1983) also confirmed that a poor resolution inside the boundary layer leads to quite important errors in the transfer of heat, moisture and momentum between the boundary layer and the middle levels of the troposphere.

The multi-level boundary-layer schemes, and particularly those based on the use of a rate equation for eddy kinetic energy or a similar quantity, allow for a much better description of turbulent properties of the boundary layer, as is now more and more required with the development of finer resolution numerical weather prediction models. These schemes also present the further advantage of offering various possibilities for rational improvement. Among those, one can look for further progress by either improving the turbulence closure schemes or extending presently used methods to cloudy situations. In the first case it is well-known that PBL modelling now offers many methods which are much more sophisticated, including more and more rate equations for other turbulent quantities like fluxes and variances (e.g. Mellor and Yamada, 1974) and even triple correlations (André et al., 1978). Such a progressive complexification of the closure schemes brings in improvements for the description of turbulent exchange, but at a rate which tends to finally saturate . On the other hand it can be thought that inclusion of cloud processes, and maybe falling rain, would be of prime importance for many meteorological situations. This can be done, at least in principle, by modifying the buoyancy term in Eq. (31) in order to account for the weight of cloud and/or rain drops (Bougeault, 1981 ; Redelsperger and Sommeria, 1981).

It remains of course necessary to perform sensitivity tests and experiments in order to judge by how much such a parameterization scheme really improves the simulation. This, and other related questions, are addressed in the papers by André (1985) and Driedonks (1985).

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