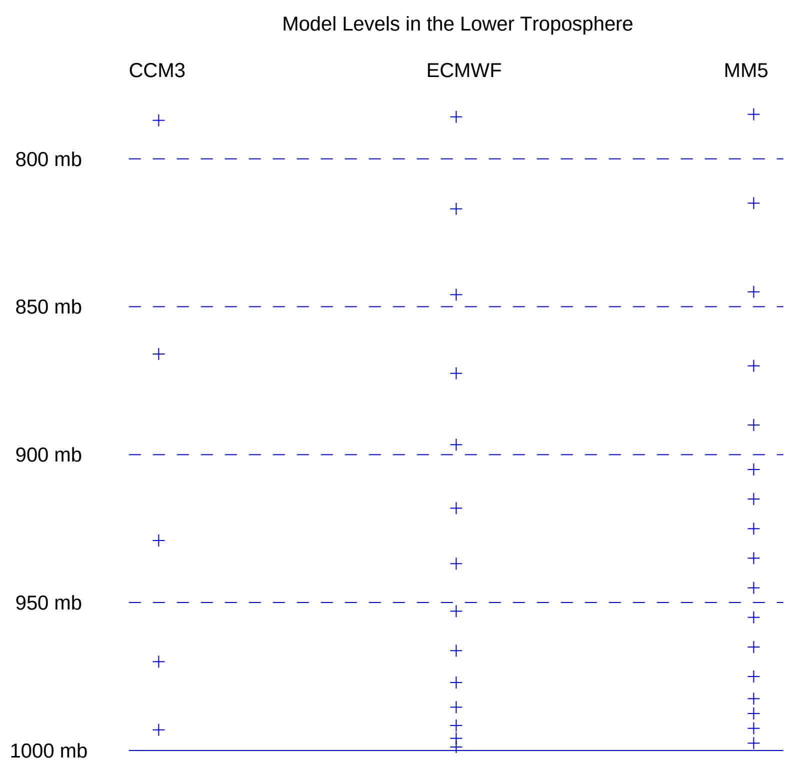
Lecture 8. Parameterization of BL Turbulence I

In the next two lectures we will summarize several approaches to parameterization of BL vertical turbulent transports that are commonly used in large-scale forecast and climate models. In such models, the horizontal grid resolution is insufficient to resolve the most energetic BL turbulent eddies, which might be tens of meters to 1-2 km across. Furthermore, while the lowest one or two model levels are usually taken to be 100 m or less from the ground to resolve stable BLs, the vertical grid spacing at a height of 1 km is typically 100-500 m, so the vertical structure of the BL can be at best coarsely resolved. Table 1 shows the distribution of thermodynamic gridpoints in the lowest 20% of the atmosphere for three representative models- the NCAR Community Climate Model version 3 (CCM3, 18 levels overall), the ECMWF operational forecast model (60 levels overall), and the MM5 mesoscale model as used for real-time forecasting in the Pacific Northwest (37 levels overall).

Three parameterization approaches are popular. In order of simplicity, they are:

1. Mixed layer models
2. 'Local' closures based on eddy diffusivity
3. 'Nonlocal' closures

Horizontal turbulent fluxes are invariably neglected as they are very small compared to advection by the mean wind. We will reserve discussion of parameterization of cloudy boundary layers for later.



Mixed Layer Models

Mixed layer models (MLMs) assume that $\overline{u}$, $\overline{v}$, and $\overline{\theta}$ in the BL are uniform ('well-mixed'). They are most applicable to convective BLs and represent stable BLs rather poorly. However, they are relatively simple to add moist physics too, and do not require a fine vertical grid to work. They are. mainly used by researchers and teachers as a conceptual tool for understanding the impacts of different physical processes on BL turbulence. However, at least one GCM (the CSU/UCLA GCM) uses a mixed layer model to describe the properties and depth of its lowest grid layer. For simplicity, we will consider a case with no horizontal advection or mean vertical motion, no thermal wind, and no diabatic effects above the surface. We will assume that the surface momentum and buoyancy fluxes are given (in general, these will depend upon the mixed layer variables, but we needn't explicitly worry about this now). We let h be the mixed layer top, at which there may be jumps in the winds and potential temperature, denoted by Δ. Turbulence in the mixed layer entrains free-tropospheric air from just above the mixed layer, causing h to rise at the **entrainment rate** w_e.

$$\frac{\partial \bar{u}}{\partial t} - f(\bar{v} - v_g) = -\frac{\partial}{\partial z}\overline{u'w'} \qquad (1)$$

$$\frac{\partial \bar{v}}{\partial t} + f(\bar{u} - u_g) = -\frac{\partial}{\partial z}\overline{v'w'} \qquad (2)$$

$$\frac{\partial \bar{\theta}}{\partial t} = -\frac{\partial}{\partial z}\overline{w'\theta'} \qquad (3)$$

$$\frac{\partial h}{\partial t} = w_e \qquad (4)$$

Since the left hand sides of (1-3) are height-independent, the right hand sides must be, too, so the fluxes of u, v, and θ are linear with z (note that this would no longer be the case for a baroclinic BL in which $\mathbf{u}_g$ varied with height, or in the presence of internal sources or sinks of θ) .The fluxes are given at the surface. The entrainment deepening of the BL, in which free-tropospheric air with value $\overline{a} + \Delta a$ of some property a (= u, v, θ) is replaced by BL air with $a = \overline{a}$ at the rate w_e, requires a flux

$$\overline{w'a'}(h) = -w_e \Delta a$$

Thus the right hand side of the mixed layer equation for $\overline{a}$ is just.

$$-\frac{\partial}{\partial z}\overline{w'a'} = -\frac{-w_e \Delta a - \overline{w'a'}(0)}{h}$$

This closes the set of equations (1-4) except for a specification of w_e, called the **entrainment closure**. This is the big assumption in any MLM. For cloud-free unstable to nearly neutral mixed layers, formulas such as that from last time (Moeng and Sullivan 1994) are commonly used:

$$\overline{w'b'}(h) = -w_e \Delta b = -(0.2 w_*^3 + u_*^3)/h, \quad \text{where} \quad \Delta b = g \Delta\theta_v / \theta_{v0}$$

Recall that w_* and u_* are determined by the surface buoyancy and momentum fluxes, respectively, so this closure determines w_e in terms of known variables, enabling (1-4) to be integrated forward

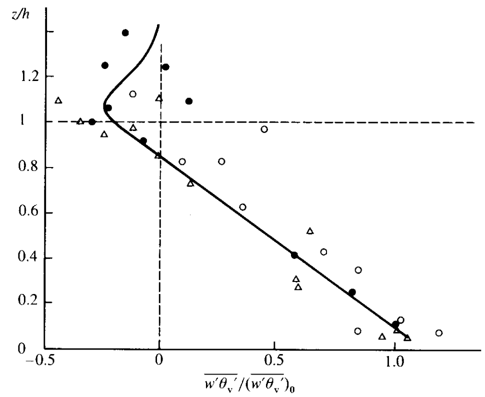


Fig. 6.2 Experimental data on the vertical variation of the virtual heat flux, normalized by its surface value; h is the depth of the mixed layer. Data are for three days from the 1983 ABL experiment; see Stull (1988, Figs. 3.1, 3.2 and 3.3). See also Fig. 6.23 of this volume.

In a convective BL, entrainment buoyancy flux is -0,2B_0

in time.   This type of entrainment closure is well-supported by observational evidence and LES simulations, especially in the purely convective BL, in which the above relation reduces to

$$\overline{w'b'}(h) = -(0.2w_*^3 + u_*^3)/h = -0.2B_0$$

An observational verification of this from data taken in a daytime convective BL over land is shown above. In fact, a classic application of a MLM is to the deepening of a convective boundary layer due to surface heating; we'll look at this when we discuss the diurnal cycle of BLs over land.

Local (*Eddy diffusivity*) *parameterizations* (Garratt 8.7)

In eddy-diffusivity (often called ***K*-theory**) models, the turbulent flux of an adiabatically conserved quantity *a* (such as θ in the absence of saturation, but not temperature *T*, which decreases when an air parcel is adiabatically lifted) is related to its gradient:

$$\overline{w'a'} = -K_a \frac{d\bar{a}}{dz} \tag{5}$$

The key question is how to specify K_a in terms of known quantities. Three approaches are commonly used in mesoscale and global models:

i) **First-order closure**, in which K_a is specified from the vertical shear and static stability, or by prescribing a
ii) **1.5-order closure** or **TKE closure**, in which TKE is predicted with a prognostic energy equation, and K_a is specified using the TKE and some lengthscale.
iii) ***K*-profiles**, in which a specified profile of K_a is applied over a diagnosed turbulent layer depth.

From here on we will drop overbars except on fluxes, so $a(z)$ will refer to an ensemble or horizontal average at level z. The following discussion of theseapproaches is necessarily oversimplified; a lot of work was done in the 1970's on optimal ways to use them. An excellent review of first, 1.5, and second-order closure is in Mellor and Yamada (1982, *Rev. Geophys. Space Phys.*, **20**, 851-875)

First-order closure

We postulate that K_a depends on the vertical shear $s = |d\mathbf{u}/dz|$, the buoyancy frequency N^2, and an eddy mixing lengthscale l. In most models, saturation or cloud fraction is accounted for in the computation of N^2. From the shear and stability one defines a Richardson number Ri = N^2/s^2. Dimensionally,

$$K_a = \text{length}^2/\text{time} = l^2 s F_a(\text{Ri}) \tag{6}$$

One could take the stability dependence in F(Ri) the same as found for the surface layer in Monin-Obhukov theory, e. g. $F(\text{Ri}) = [\phi_a(\zeta)\phi_m(\zeta)]^{-1}$, where ζ depends on Ri as in the surface layer, and $\phi_a = \phi_m$ (if a is momentum) or ϕ_h (if a is a scalar). This is fine in the stable BL. In the convective BL it gives (see notes p. 6.2) $F_m(\text{Ri}) = (1 - 16\text{Ri})^{1/2}$ and $F_m(\text{Ri}) = (1 - 16\text{Ri})^{3/4}$. However, in nearly unsheared convective flows, one would like to obtain a finite K_a independent of s in the limit of small s. This requires $F_a \propto (-\text{Ri})^{1/2}$ so $K_a \propto l^2 s(-\text{Ri})^{1/2} = l^2(-N^2)^{1/2}$. This is consistent with the M-O form for K_m but not for K_h. Thus, we just choose $K_h = K_m$ to obtain:

$$F_{h,m}(\text{Ri}) = \begin{cases} (1 - 16\text{Ri})^{1/2}, & \text{(unstable)} \\ (1 - 5\text{Ri})^2 & \text{(stable)} \end{cases}$$

No turbulent mixing is diagnosed unless Ri < 0.2. Every model has its own form of F(Ri), but most are qualitatively similar to this. Usually, if this form is used within the stable BL, the F's are enhanced near the surface (no Ri = 0.2 cutoff) to account for unresolved flows and waves driven, for instance, by land-sea or hill-valley circulations that can result in spatially and temporally intermittent turbulent mixing.

Many prescriptions for l exist. The only definite constraint is that $l \rightarrow kz$ near the surface to match (6) to the eddy diffusivity in neutral conditions to that observed in a log layer, $K_a = ku_*z$. One commonly used form for l (suggested by Blackadar, 1962) is

$$l = \frac{\lambda}{1 + \lambda/kz}$$

where the 'asymptotic lengthscale' λ is chosen by the user. A typical choice is λ = 50-100 m, or roughly 10% of the boundary layer depth. The exact form of l is less important than it might appear, since typically there will be (i) layers with large K_a, and small gradients (i. e. fairly well mixed layers) in which those small gradients will just double (but still be small) if K_a, is halved, to maintain the same fluxes, (ii) layers with small K_a where physical processes other than turbulence will tend to dictate the vertical profiles of velocity and temperature, and (iii) a surface layer, in which the form of K_a is always chosen to match Monin-Obuhkov theory, and so is on solid observational ground.

1.5-order closure

Now we prognose the TKE $e = q^2/2$ based on the shear and stability profiles. Using the same eddy mixing lengthscale as above, dimensionally

$$K_a = (\text{length})(\text{velocity}) = lq\, S_a(G_M, G_H), \qquad a = M \text{ (momentum) or } H \text{ (heat)}$$

$$G_M = l^2 S^2/q^2, \; G_H = l^2 N^2/q^2$$

Closure assumptions and measurements discussed in Mellor and Yamada dictate the form of S_M and S_H in terms of the nondimensional shear and stratification G_M and G_H. These are complicated algebraic expressions, but are shown in the figure on the next page.

To determine the evolution of q, we use the TKE equation

$$\frac{\partial}{\partial t}\left(\frac{q^2}{2}\right) = S + B + T - \varepsilon. \tag{7}$$

We model the shear and buoyancy production terms using eddy diffusion to find the fluxes:

Mellor and Yamada, 1982

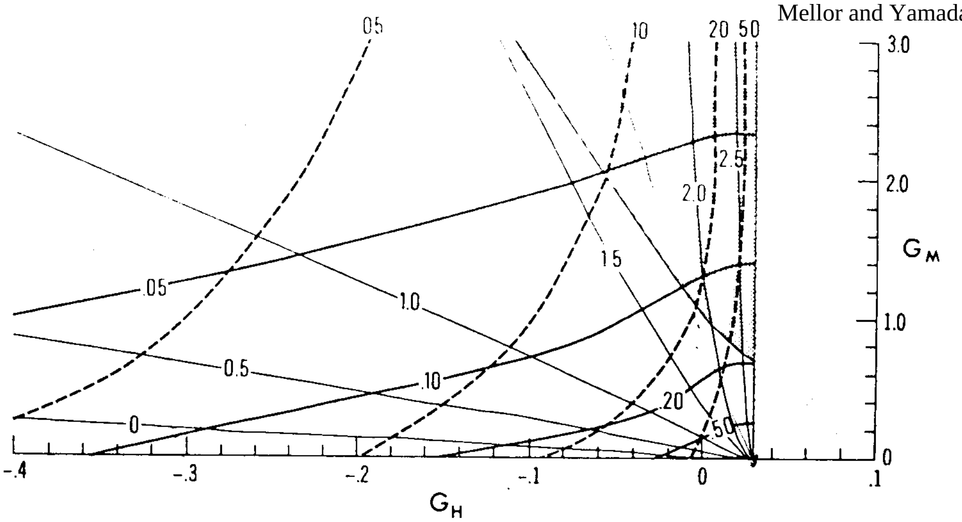


Fig. 3. The stability functions $S_M(G_H, G_M)$ and $S_H(G_H, G_M)$. The heavy solid lines are contours of S_M, whereas the dashed lines are contours of S_H. The lighter solid lines are contours of $(P_s + P_b)/\varepsilon$. One could also draw lines of constant $R_i = G_H/G_M$, which are radial lines on this diagram. The shaded portion is where $\langle w^2\rangle/q^2 \lesssim 0.12$.

Stability functions for TKE closure.

$$S = -\overline{u'w'}\frac{\partial \bar{u}}{\partial z} - \overline{v'w'}\frac{\partial \bar{v}}{\partial z} = lq\, S_m|d\mathbf{u}/dz|^2$$

$$B = \overline{w'b'} = -lq\, S_h N^2$$

We model the transport term by neglecting pressure correlations and using eddy diffusion to model the flux of TKE:

$$\overline{w'p'} \approx 0$$

$$\overline{w'e'} = -K_q \frac{\partial}{\partial z}\left(\frac{q^2}{2}\right) = -lqS_q \frac{\partial}{\partial z}\left(\frac{q^2}{2}\right) \quad (S_q \text{ is often taken to be 0.2})$$

$$T = -\frac{\partial}{\partial z}\left(\overline{w'e'} + \frac{1}{\rho_0}\overline{w'p'}\right) = \frac{\partial}{\partial z}\left(lqS_q \frac{\partial}{\partial z}\left(\frac{q^2}{2}\right)\right)$$

Lastly, the dissipation term is modelled in terms of characteristic turbulent velocity and length-scales. While the lengthscale in ε is related to the master lengthscale, it is necessary to intoduce a scaling factor to get the TKE to have the right magnitude:

$$\varepsilon = q^3/(lB_1),, \quad B_1 \approx 15.$$

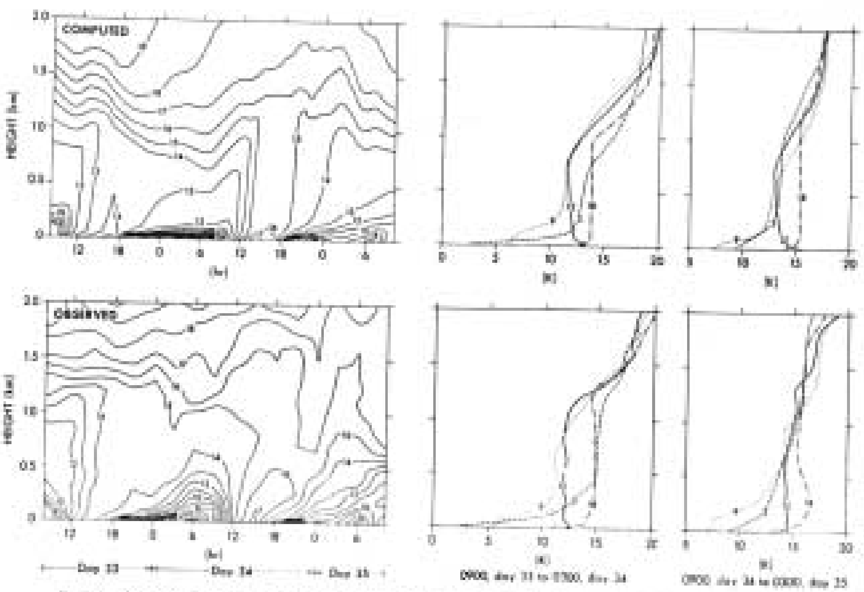


Fig. 11. Observed and calculated atmospheric boundary layer and vertical and temporal variations of mean virtual potential temperature −273°K. Units are degrees Kelvin.

Virtual temperature evolution observed during two days of the Wangara expt. (top) and modelled with a TKE closure (bottom)

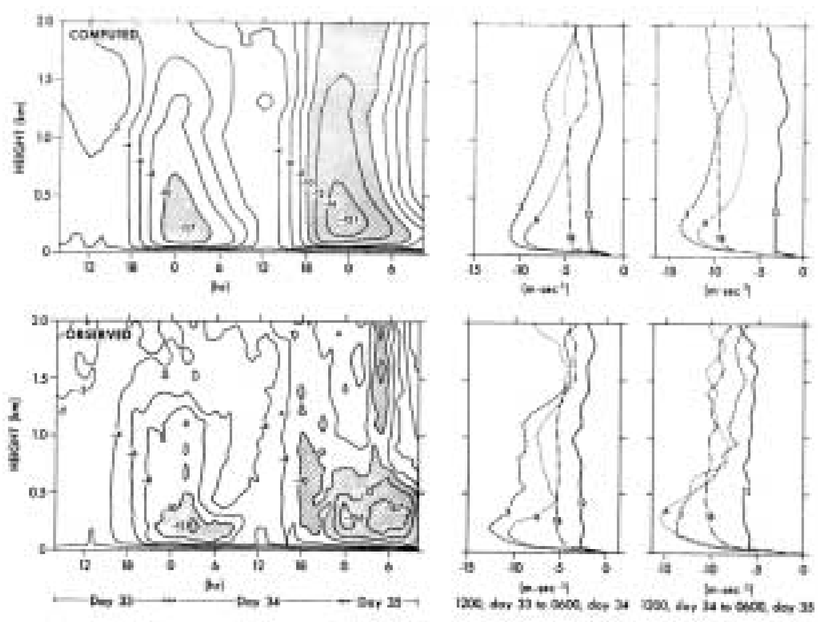


Fig. 12. Observed and calculated atmospheric boundary layer and vertical and temporal variations of the eastward mean wind component. Units are meters per second.

Same for *u* velocity.

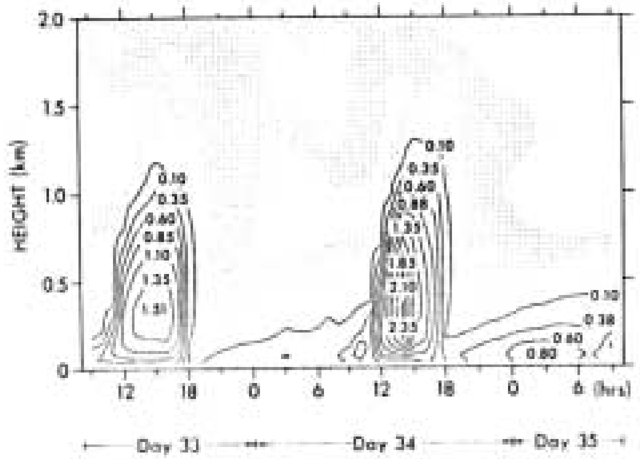


Fig. 14. Time and space variation of computed q^2 (twice the turbulent kinetic energy); units are square meters per second. The stippled areas indicate regions where $10^{-4} < q^2 < 10^{-2}$ $m^2 s^{-2}$.

Modelled TKE profile for simulation on previous page

With these forms for all the terms in the TKE equation, it can now be integrated forward in time. The basic improvement in using TKE vs. 1st order closure is that there is TKE ‘transport’ (through eddy diffusion) and storage. In the surface layer, where storage and transport are negligible compared to local shear and buoyancy production of TKE, the latter must balance dissipation, and, one finds that $\varepsilon = S + B$, so

$$q^3/(lB_1) = lq\, S_m|d\mathbf{u}/dz|^2 - lq\, S_h N^2$$

$$q^2 = B_1 l^2\{S_m|d\mathbf{u}/dz|^2 - S_h N^2\}$$

Hence, q can be eliminated in place of the local shear and stratification, and we recover the first order closure method.

K-profile methods

For specific types of boundary layer, one can use measurements and numerical experiments to specify a profile of eddy diffusivity which matches the observed fluxes and gradients. This can be particularly useful in situations such as stable nocturnal BLs which can be difficult for other methods. Such methods require a diagnosis of BL height h, then specify a profile of K. For instance , Brost and Wyngaard (1978) combined theoretical ideas and observational analysis to proposed the following profile for stable BLs:

$$K_m = ku_* hP(z)/(1 + 5z/L), \quad P(z) = (z/h)(1 - z/h)^{3/2}$$

This method is designed to approach the correct form $ku_* z/\phi_m(z/L)$ in the surface layer, where $z/h << 1$. Similar approaches are have been used for convective BLs. Advantages of the K-profile method are that it is computationally simple and works well even with a coarsely resolved BL, as long as the BL height h can be diagnosed fairly well. On the other hand, it is tuned to specific types of BL,and may work poorly if applied more generally than the situations for which it was tuned.

Comments on local closure schemes

First-order closure is most appropriate for neutral to weakly stable BLs in which little transport of TKE is occuring and the size of the most energetic eddies is a small fraction of the BL depth. In this case, it is reasonable to hope that the local TKE will be dependent on the local shear and stability, and that since the eddies are small, they can be well repressented as a form of diffusion. However, it works tolerably well in convective boundary layers as well, except near the entrainment zone. In an entrainment zone, Transport of TKE into the entrainment zone is required to sustain any turbulence there. Since this is ignored in 1st order closure, there is no way for such a model to deepen by entrainment through an overlying stable layer, as is observed. BL layer growth must instead be by encroachment, i. e. the incorporation of air above the BL which has a buoyancy lower than that within the BL. This does allow a surface-heated convective boundary layer to deepen in a not too unreasonable manner, but creates severe problems for cloud-topped boundary layer modeling. Almost all large-scale models (e. g. CCM3, ECMWF, and MM5) include a first-order closure scheme to handle turbulence that develops above the BL (due to Kelvin-Helmholtz instability or elevated convection, for instance).

1.5-order closure is also widely used, especially in mesoscale models where the timestep is short enough not to present numerical stability issues for the prognostic TKE equation. The Mellor-Yamada and Gayno-Seaman PBL schemes for MM5 are 1.5 order schemes that include the effect of saturation on N^2. The Burk-Thompson scheme for MM5 is a 1.5 order scheme with additional prognostic variables for scalar variances ('Mellor-Yamada Level 3'). The TKE equation in 1.5-order closure allows for some diffusive transport of TKE. This creates a more uniform diffusivity throughout the convective layer, and does permit some entrainment to occur.Quite realistic simulations of the observed diurnal variation of boundary layer temperature and winds have been obtained using this method (see figures on next page). However, getting realistic entrainment rates for clear and cloud-topped convective BLs with this approach requires considerable witchcraft. The BL top tends to get locked to a fixed grid level if there is a significant capping inversion and vertical grid spacing of more than 100 m or so. TKE closure has also proved successful for cloud-topped boundary layers, but again only with grid spacings smaller than is currently feasible for GCMs. Grenier and Bretherton (2001, *MWR*, **129**, 357-377) showed that this method works well for convective BLs even at coarse resolution when combined with an explicit entrainment parameterization at the BL top, implemented as an effective diffusivity.

K-profile methods are widely used in GCM BL parameterizations (e.g. CCM3). For convective boundary layers, a nonlocal contribution is usually also added to the fluxes (see below).

Nonlocal closure schemes

Any eddy diffusivity approach will not be entirely accurate if most of the turbulent fluxes are carried by organized eddies filling the entire boundary layer (such as boundary layer rolls or convection). Consequently, a variety of 'nonlocal' schemes which explicitly model the effects of these boundary layer filling eddies in some way have been proposed. A difficulty with this approach is that the structure of the turbulence depends on the BL stability, baroclinicity, history, moist processes, etc., and no nonlocal parameterization proposed to date has comprehensively addressed the effects of all these processes on the large-eddy structure. Nonlocal schemes are most attractive when the vertical structure and turbulent transports in a specific type of boundary layer (i. e. neutral or convective) must be known to high accuracy. For instance, successful applications include the detailed thermal structure (i. e. deviation from neutral static stability) within a convective boundary layer, or the velocity structure and relation of near-surface wind to geostrophic wind within a near-neutral boundary layer (this is the motivation for the PBL model developed here at UW by Bob Brown's group).