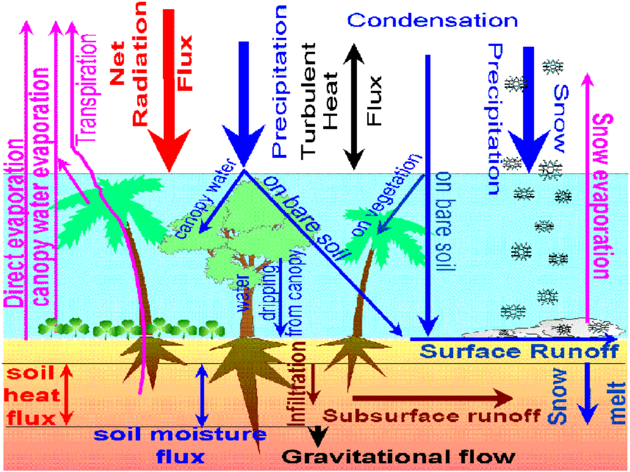
The planetary boundary layer (PBL)

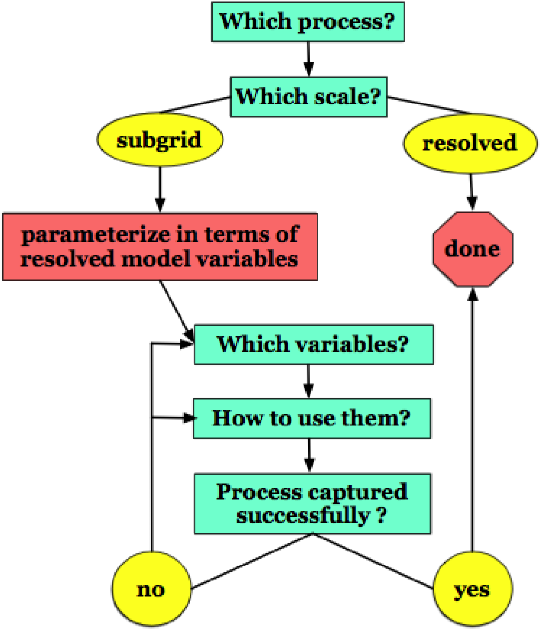


http://www.met.tamu.edu/class/metr452/models/2001/soilmodel.gif

Parameterization flow chart

Take 5 minutes to discuss with your neighbor:

- What are the effects/processes in the PBL that need to be captured in a global model?
- Which variables are available for that?
- How could these variables be used?



Governing equation for the horizontal wind

$$\frac{\partial u}{\partial t} + u\frac{\partial u}{\partial x} + v\frac{\partial u}{\partial y} + w\frac{\partial u}{\partial z} - fv = -\frac{1}{\rho}\frac{\partial p}{\partial x} + \nu\nabla^2 u \qquad (1)$$

ν = kinematic viscosity.

- The spatial scales that are important in the PBL are much smaller than elsewhere in the atmosphere.
- Thus separate the prognostic variables into mean and subgrid-scale terms (e.g., $u = U + u'$).
- Apply the Boussinesq approximation (retain density fluctuations in the buoyancy terms only) and Reynolds averaging to obtain:

$$\begin{aligned}\frac{\partial U}{\partial t} + U\frac{\partial U}{\partial x} + V\frac{\partial U}{\partial y} + W\frac{\partial U}{\partial z} &= fV - \frac{1}{\rho_0}\frac{\partial P}{\partial x} + \nu\nabla^2 U \\ &\quad -\frac{\partial}{\partial x}\overline{u'u'} - \frac{\partial}{\partial y}\overline{u'v'} - \frac{\partial}{\partial z}\overline{u'w'}\end{aligned} \qquad (2)$$

ρ_0 = density of the mean state.

Governing equation for the horizontal wind II

- Correlation terms like $\overline{u'w'}$: eddy fluxes or Reynolds stresses
- Neglect viscous effects for large Reynolds number ($UL/\nu \gg 1$, $L =$ length scale). Assume x, y scales $\gg$ z-scales:

$$\frac{\partial U}{\partial t} + U\frac{\partial U}{\partial x} + V\frac{\partial U}{\partial y} + W\frac{\partial U}{\partial z} - fV = -\frac{1}{\rho_0}\frac{\partial P}{\partial x} - \frac{\partial}{\partial z}\overline{u'w'} \qquad (3)$$

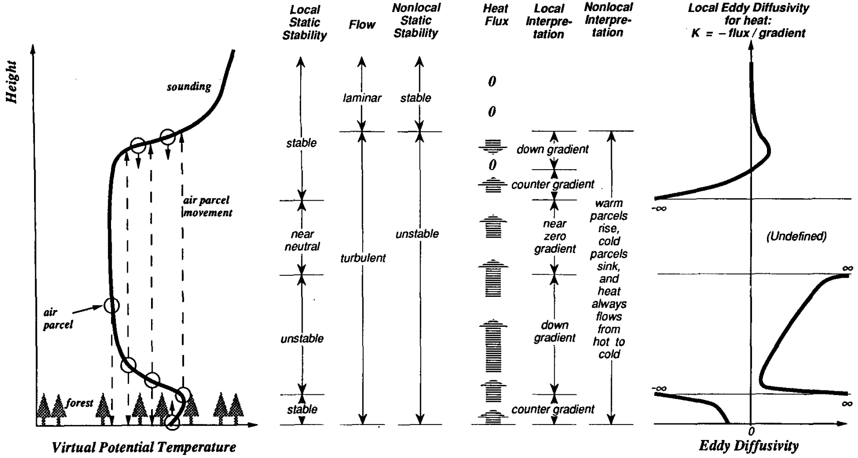
- There are different ways to obtain $\overline{u'w'}$:
- First-order closure or K-theory:

$$\boxed{\overline{u'w'} = -K_M \frac{\partial U}{\partial z}} \qquad (4)$$

K_M = eddy diffusion coefficient for momentum.

- Assuming K_M = constant is the easiest, but is insufficient to simulate many observed states of the PBL.

Local vs. non-local interpretations of PBL processes



Stull, BAMS, 1991

K_M as a function of mixing length and stability

- The Richardson number Ri is a dynamic stability measure for turbulence
- $Ri = \frac{g\partial\theta/\partial z}{(\partial U/\partial z)^2}$ = dimensionless ratio of buoyant suppression of turbulence to shear generation of turbulence
- Instead of a constant K_M, obtain K_M as a function of shear and Ri.

$$K_M = l^2 F_M \left| \frac{dU}{dz} \right| \tag{5}$$

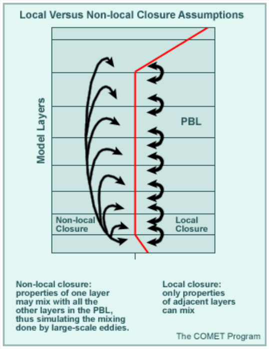
F_M = function of Ri, l = turbulence length scale:

$$\frac{1}{l} = \frac{1}{\kappa z} + \frac{1}{\lambda} \tag{6}$$

κ = von Karman constant (=0.4), z = height.

- λ = asymptotic value such that l is limited by the PBL height.

Limitations to mixing length approach



The mixing length approach

- Is valid only in boundary layers not having large eddies
- Only allows for downgradient transport (unless $K_M < 0$ is allowed)
- Neglects vertical transport of perturbation velocities

1.5-order local closure schemes

- Obtaining K_M from a prognostic equation for E is referred to as 1.5-order closure scheme:

$$\boxed{K_M = 0.52 l_M \sqrt{E}} \tag{7}$$

where

$$E = \frac{1}{2}(\overline{u'^2} + \overline{v'^2} + \overline{w'^2}) \tag{8}$$

and l_M is an empirical mixing length.

- The prognostic equation for E is given as:

$$\frac{dE}{dt} = -\underbrace{\overline{u'w'}\frac{\partial U}{\partial z}}_{I} - \underbrace{\overline{v'w'}\frac{\partial V}{\partial z}}_{I} + \underbrace{\frac{g}{\overline{\theta}}\overline{w'\theta'}}_{II} - \frac{\partial}{\partial z}\left(\underbrace{\overline{E'w'}}_{III} + \underbrace{\frac{\overline{p'w'}}{\rho}}_{IV}\right) - \underbrace{\epsilon}_{V}$$

with 6 unknowns: $\overline{u'w'}, \overline{v'w'}, \overline{w'\theta'}, \overline{E'w'}, \frac{\overline{p'w'}}{\rho}, \epsilon$

1.5-order local closure schemes

- Closure for the fluxes ($\overline{u'w'}, \overline{v'w'}, \overline{w'\theta'}$) can be obtained as in eq. (4) with K_M depending additionally on the mean E
- Closure for the transport/diffusion terms:

$$\overline{E'w'} + \frac{\overline{p'w'}}{\rho} \propto \frac{l_1}{\sqrt{E}} \frac{\partial E}{\partial z} \tag{9}$$

- Closure for the dissipation term:

$$\epsilon = \frac{E^{3/2}}{l_2} \tag{10}$$

where l_x are empirical length-scale parameters chosen to best match observations

Turbulent kinetic energy budget (Fig. 5.4 Stull, 1988)

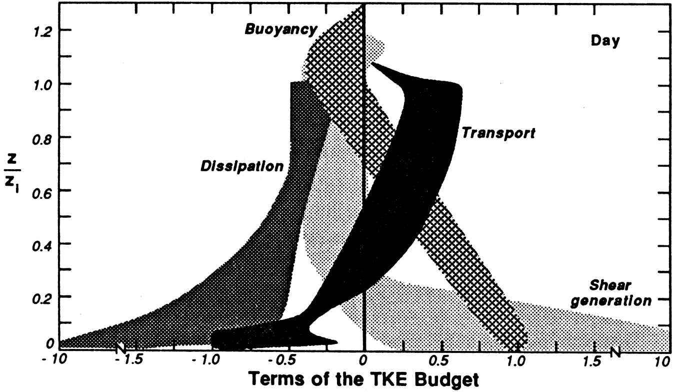


Fig. 5.4 Normalized terms in the turbulence kinetic energy equation. The shaded areas indicate ranges of values. All terms are divided by w_*^3 / z_i, which is on the order of $6 \times 10^{-3} m^2 s^{-3}$. Based on data and models from Deardorff (1974), André et al. (1978), Therry and Lacarrere (1983), Lenschow (1974), Pennell and LeMone (1974), Zhou, et al. (1985) and Chou, et al. (1986).

TKE in stratocumulus clouds at night

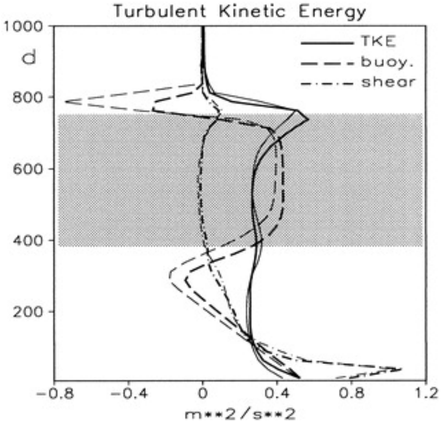


Figure : left: cloud resolving models; right: single column model (SCM)

Lenderink and Holtslag, MWR, 2000

Second-order closure

- Here the eddy covariance terms are obtained from prognostic equations, e.g.:

$$\begin{aligned}\frac{\partial \overline{u'w'}}{\partial t} &= -\overline{u'w'}\frac{\partial W}{\partial z} - \overline{w'w'}\frac{\partial U}{\partial z} - \frac{\partial \overline{u'w'w'}}{\partial z} + \frac{g}{\overline{\theta}}\overline{u'\theta'} \\ &+ \overline{\left(\frac{p'}{p}\right)\left(\frac{\partial w'}{\partial x} + \frac{\partial u'}{\partial z}\right)} - 2\epsilon \end{aligned} \tag{11}$$

- Then closure needs to be found for the triple correlation terms:

$$\overline{u'w'w'} = -\frac{L}{\sqrt{E}}\left[2\frac{\partial \overline{u'w'}}{\partial z} + \frac{\partial \overline{w'w'}}{\partial x}\right] \tag{12}$$

where $L =$ empirical length scale

- A second-order closure for momentum only implies: 6 additional equations and 10 unknowns (triple correlation terms), which is too CPU time consuming for GCMs

Higher-order closure

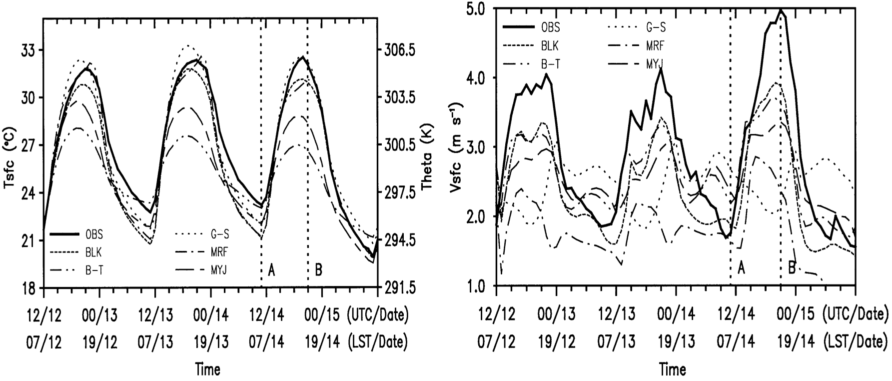
Order of Closure	Correlation Triangle of Unknowns
Zero	$\overline{U}$
	$\overline{V}$ $\overline{W}$
First	$\overline{u'^2}$
	$\overline{u'v'}$ $\overline{u'w'}$
	$\overline{v'^2}$ $\overline{v'w'}$ $\overline{w'^2}$
Second	$\overline{u'^3}$
	$\overline{u'^2v'}$ $\overline{u'^2w'}$
	$\overline{u'v'^2}$ $\overline{u'v'w'}$ $\overline{u'w'^2}$
	$\overline{v'^3}$ $\overline{v'^2w'}$ $\overline{v'w'^2}$ $\overline{w'^3}$

Stull, 1988, Table 6.2

Equations and unknowns for the momentum equations:

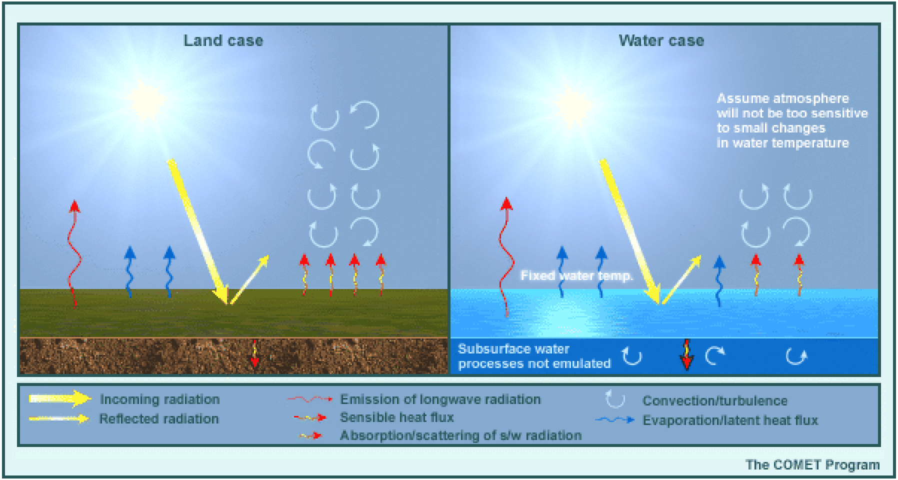
- first moments ($\overline{U_i}$): 3 equations with 6 unknowns ($\overline{u'v'}$, $\overline{u'w'}$, $\overline{v'w'}$, $\overline{u'^2}$, $\overline{v'^2}$, $\overline{w'^2}$)
- second moments ($\overline{u'v'}$): 6 equations with 10 unknowns
- third moments ($\overline{u'v'w'}$): 10 equations with 15 unknowns

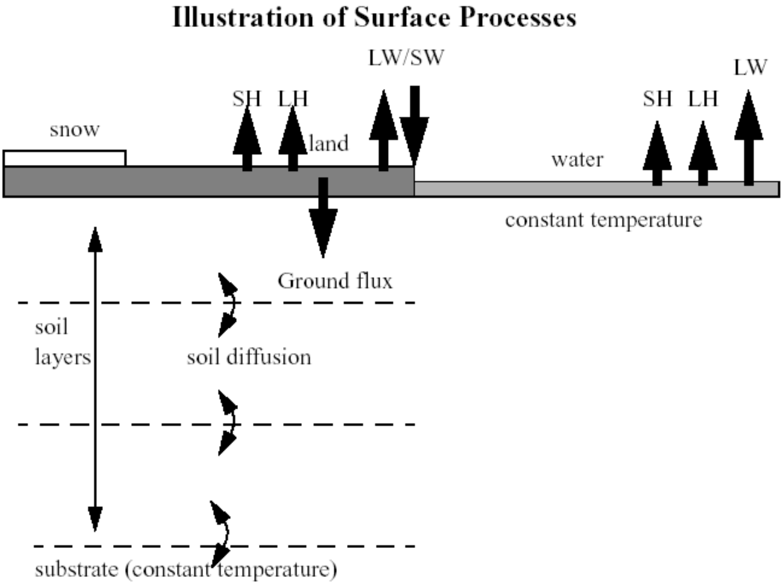
Performance of 5 nonlocal PBL schemes



Zhang and Zheng, 2004

Energy budget land vs. water





www.mmm.ucar.edu/mm5/documents/MM5_tut_Web_notes/MM5/mm5.htm

Surface layer

- The surface layer is defined by the near vertical constancy (variations less than 10%) of the heat, momentum and moisture fluxes.
- It covers the lowest 10% of the atmospheric boundary layer
- The turbulent flux of a variable χ at the surface is obtained from the bulk transfer relation:

$$\boxed{(\overline{w'\chi'})_S = -C_\chi |U_L| (\chi_L - \chi_S)} \tag{13}$$

  C_χ = transfer coefficient. The subscripts L and S denote values at the lowest model level (representing the top of the surface layer) and the surface, respectively, $|U_L|$ = horizontal wind vector at level L.
- Obtain C_χ from Monin-Obukhov similarity theory by integrating the flux-profile relationships over the lowest model layer.
- C_χ depends on wind shear and static stability of the atmosphere above the surface, and the surface roughness. C_χ increases as the atmosphere becomes more unstable.

Surface temperatures over land

- GCMs calculate the land surface T_s from a surface energy balance condition:

$$\boxed{C_L \frac{\partial T_s}{\partial t} = R_{net} + LE + H + G} \tag{14}$$

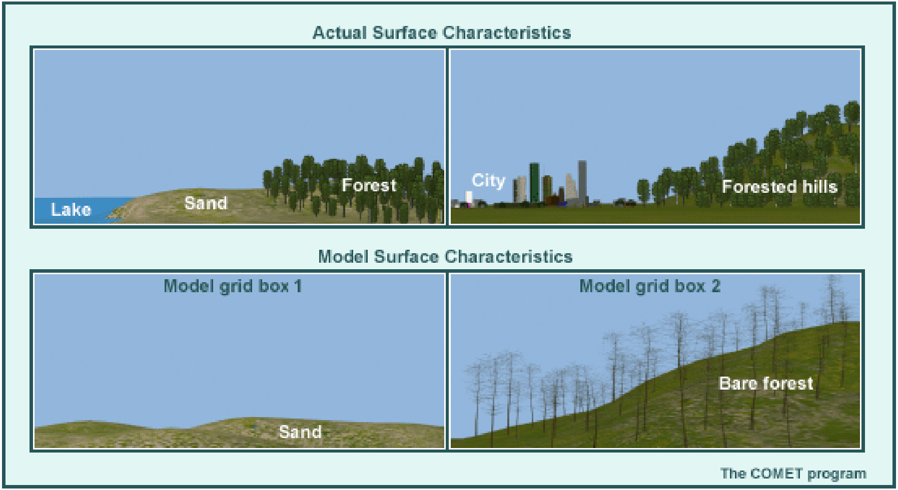
C_L = heat capacity of the layer [J m^{-2} K^{-1}], H = sensible heat flux, LE = latent heat flux, G = ground heat flux, R_{net} = net radiation:

$$R_{net} = (1 - \alpha_s) R^{\downarrow}_{SW} + \epsilon R^{\downarrow}_{LW} - \epsilon \sigma T_s^4 \tag{15}$$

α_s = surface albedo, $R^{\downarrow}_{SW}$ = downwelling solar radiation, $R^{\downarrow}_{LW}$ = downwelling longwave radiation, ϵ = surface emissivity, σ = Stefan-Boltzmann constant.

- Recently, surface models are being employed within GCMs to evaluate the surface and subsurface T.

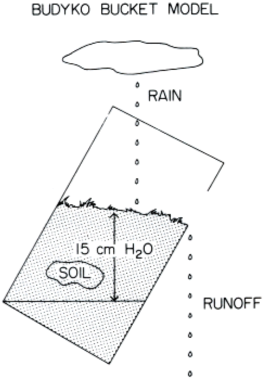
Surface characteristics



Sea surface temperatures (SST)

- Over oceans, T_s is either obtained from interpolated monthly mean observed SSTs or calculated from an interactive ocean model.
- Many atmospheric GCMs (AGCM) use prescribed, observed SSTs for computational efficiency.
- Daily SSTs are obtained by linearly interpolating between mid-month SSTs allowing seasonal variations.
- Drawback: no realistic interaction of the atmosphere with oceans. Surface fluxes are irrelevant to SSTs.
- Constant SST implies that the globally, annually averaged top-of-the-atmosphere (TOA) net radiation may not to be zero.
- GCMs with prescribed SST models are "tuned" to guarantee TOA radiation balance. Any perturbation to the "tuned" model, e.g. a change in the model physics, may cause an imbalance.
- Over sea ice regions, T_s should be evaluated from a sea ice model.

Surface hydrology



www.scopenvironment.org/downloadpubs/scope35/chapter05.html

Easiest method: Budyko's bucket model:

$$\frac{\partial W_s}{\partial t} = P_R - J_{EV} - R + M$$

where:

W_s: Soil water content

P_R: Precipitation rate

J_{EV}: Evaporation rate

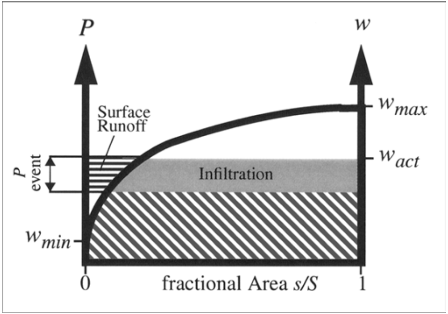
R: Runoff

M: Melt water

- If $W_s > W_{smax} \Rightarrow$ Runoff

W_{smax} = Depth of ground water reservoir

Improved bucket scheme



$w_{min}, w_{max}, w_{act}$ = minimum, maximum and actual soil water capacity; P = precipitation

(Hagemann and Gates, Clim. Dyn., 2003)

- Use a statistical distribution of many subgrid buckets
- Fractional saturation (s/S) of the grid box:

$$\frac{s}{S} = 1 - \left(\frac{w_{max} - w_{act}}{w_{max} - w_{min}}\right)^b$$

b defines shape of the distribution curve of w (depends on model resolution)

- Actual soil water content in the grid box:

$$W_s = w_{min} + \int_{w_{min}}^{w_{act}} \left(1 - \frac{s}{S}\right) dw$$

Take-home messages PBL parameterizations

- Different parameterizations are used for the surface layer and the rest of the boundary layer
- Non-local closure (1.5 order closure or higher) often better capture the boundary layer depth, structure and wind profiles than local schemes but are computationally more expensive
- The surface fluxes are obtained from Monon-Obukhov similarity theory
- Different surface types are considered over land
- Sea surface temperatures can either be obtained from observations or an ocean model