Surface layer parameterization in *WRF*

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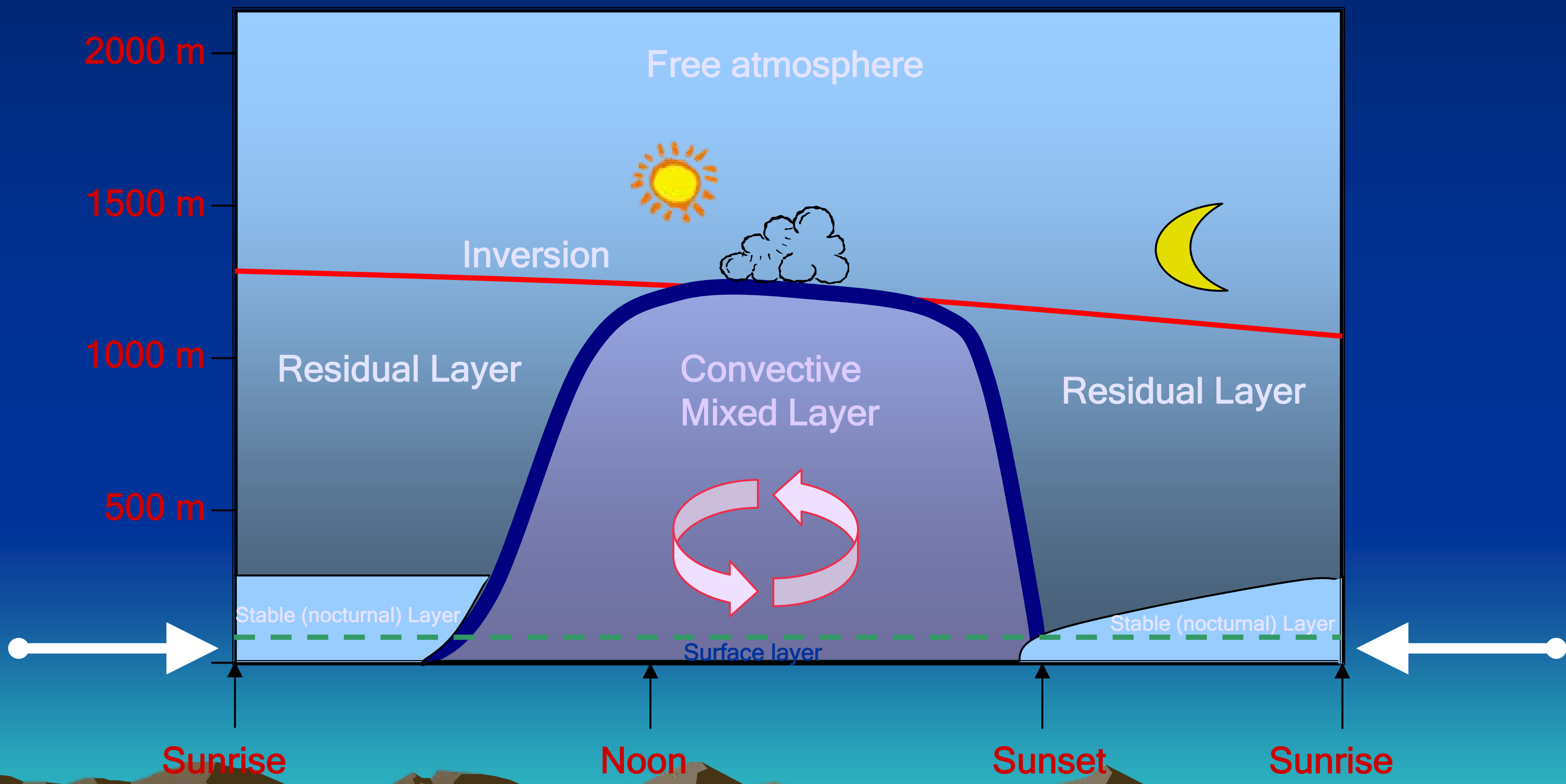
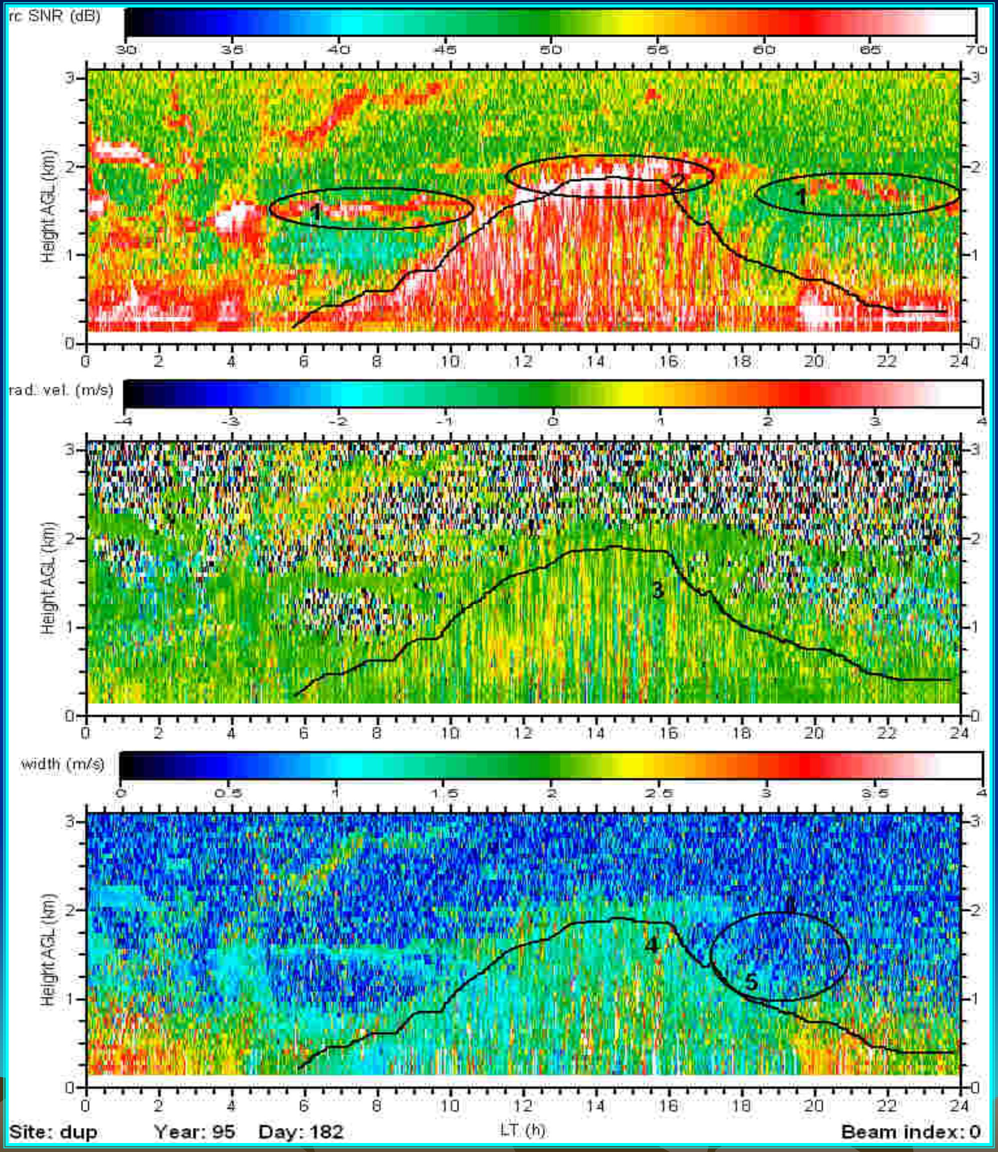
Surface Boundary Layer:

The atmospheric surface layer is the lowest part of the atmospheric boundary layer (typically about a tenth of the height of the BL) where mechanical (shear) generation of turbulence exceeds buoyant generation or consumption. Turbulent fluxes and stress are nearly constant with height in this layer.



Adapted from "Introduction to Boundary Layer Meteorology" - R.B. Stull, 1988.

Convective Boundary Layer as seen by a Radar Wind Profiler:



Dupont, TN
915-MHz WP
Lat 36.28 N
Lon 86.52 W
Alt 155 m

Closure problem:

In the set of equation for turbulent flow the number of unknowns is larger than the number of equations, therefore there are unknown turbulence terms which must be parameterized as a function of known quantities and parameters.

Much of the problem in numerical modeling of the turbulent atmosphere is related to the numerical representation (or parameterization as a function of known quantities and parameters) of these fluxes. This problem is known as closure problem.

$$\frac{\overline{P}}{\Re} = \overline{\rho}\,\overline{T_v}$$

$$\frac{\partial \overline{U_j}}{\partial x_j} = 0$$

$$\frac{\partial \overline{U}}{\partial t} + \overline{U_j}\frac{\partial \overline{U}}{\partial x_j} = -f_c(\overline{V_g} - \overline{V}) - \frac{\partial(\overline{u_j'u'})}{\partial x_j}$$

$$\frac{\partial \overline{V}}{\partial t} + \overline{U_j}\frac{\partial \overline{V}}{\partial x_j} = +f_c(\overline{U_g} - \overline{U}) - \frac{\partial(\overline{u_j'v'})}{\partial x_j}$$

$$\frac{\partial \overline{q_T}}{\partial t} + \overline{U_j}\frac{\partial \overline{q_T}}{\partial x_j} = +S_{qT}/\overline{\rho_{air}} - \frac{\partial(\overline{u_j'q_T'})}{\partial x_j}$$

$$\frac{\partial \overline{\theta}}{\partial t} + \overline{U_j}\frac{\partial \overline{\theta}}{\partial x_j} = -\frac{1}{\overline{\rho}C_p}\left[L_v E + \frac{\partial \overline{Q_j^*}}{\partial x_j}\right] - \frac{\partial(\overline{u_j'\theta'})}{\partial x_j}$$

$$\frac{\partial \overline{C}}{\partial t} + \overline{U_j}\frac{\partial \overline{C}}{\partial x_j} = +S_c - \frac{\partial(\overline{u_j'c'})}{\partial x_j}$$

from “**Introduction to Boundary Layer Meteorology”** R.B. Stull, 1988.

Local and *non-local* closure:

Closure can be *local* and *non-local*.

- For *local closure*, an unknown quantity in any point in space is parameterized by values and/or gradients of known quantities at the same point.

- For *non-local closure*, an unknown quantity at one point in space is parameterized by values and/or gradients of known quantities at many points in space.

Use of *first-order closure* schemes for evaluating turbulent fluxes is common in many boundary layer, mesoscale, and general circulation models of the atmosphere.

Local closure – first order:

If we let X be any variable, then one possible first order closure approximation for the flux $\overline{X'w'}$ is:

$$\overline{X'w'} = -K \frac{\partial \overline{X}}{\partial z} \qquad (**)$$

Where different K are associated with different variables.
K_m is for momentum;
K_H is for heat;
K_E is for moisture.

Some experimental evidence suggests: $K_H = K_E = 1.35\ K_m$

Despite of the complexity of the Earth's surface, widely used parameterizations of the turbulent exchange in the surface layer generally remain rather simple.

(**) These relationships between local fluxes and local gradients were introduced first by:
Boussinesq, J., 1877: Essai sur la theorie des eaux courants, *Mem. Pres. Par div. Savants a l'Academie Sci.*, Paris, 23, 1–680.

Friction velocity

When the turbulence is generated by wind shear near the ground, the magnitude of the surface Reynold’s stress is an important scaling parameter in the similarity theory. The total vertical flux of horizontal momentum, measured near the surface is:

$$\tau_{xz} = -\rho\overline{u'w'} \quad and \quad \tau_{yz} = -\rho\overline{v'w'}$$

$$\Rightarrow |\tau| = \left[\tau_{xz}{}^{2} + \tau_{yz}{}^{2}\right]^{1/2}$$

Based on this relationship, a velocity scale *u** is defined as:

$$u_*^2 = \left[\overline{u'w'}^{2} + \overline{v'w'}^{2}\right]^{1/2} = |\tau| / \rho$$

Within a Surface layer, (also known as Prandtl layer, or constant flux layer, even if this last term is inaccurate) in terms of the first-order turbulence closure we can write:

$$\tau = -\rho\,\overline{u'w'} = \rho\,u_*^2 \qquad \Rightarrow \qquad \frac{d\tau}{dz} = \frac{d}{dz}K_m\frac{d\bar{u}}{dz} = 0 \qquad or \qquad \frac{d\bar{u}}{dz} = \frac{const}{K_m}$$

The constant is simply a turbulent flux at the surface, $\tau\,(z=0)$.

Dimensional analysis suggests that $K_m = l \cdot u_s$ is a combination of length, l, and velocity, u_s scales.

von Karman proposed $l = k \cdot z$ and $u_s = u^*$ on the basis of laboratory experiments with a well-established layer of constant turbulent fluxes. Here, z is the height above the surface.

The "constant" $k = (0.40 +/- 0.01)$ is known as the von Karman constant.

Integration of the previous eq. gives an expression for the logarithmic velocity profile in the surface layer

$$\overline{u(z)} = \frac{u_*}{k} \ln \frac{z}{z_0}$$

where z_0 is surface roughness.

Monin and Obukhov suggested a universal stability correction of the previous eq. in the following form

$$\overline{u(z)} = \frac{u_*}{k}\left(\ln \frac{z}{z_0} - \Psi\left(z/L \right) \right)$$

where $L = \frac{u_*^3 \overline{\vartheta_v}}{kg\left(\overline{w'\vartheta_v'}\right)}$ is the Monin-Obukhov length scale.

In the surface layer Monin-Obukov similarity theory can be used to describe the logarithmic wind profile.

The *log wind profile* is a semi-empirical relationship used to describe the vertical distribution of horizontal wind speeds above the ground within the atmospheric surface layer. The equation to estimate the wind speed (u) at height z (meters) above the ground is:

$$u_z = \frac{u_*}{k}\left[\ln\left(\frac{z-d}{z_0}\right) + \Psi\left(z/L\right)\right]$$

Where:

u_* is the friction velocity (m s^{-1}),
κ is von Karman's constant (~0.40),
d is the zero plane displacement,
z_0 is the surface roughness (in meters),
Ψ is a stability term and
L is the Monin-Obukov stability parameter.

$$u_z = \frac{u_*}{k}\left[\ln\left(\frac{z-d}{z_0}\right) + \Psi\left(z/L\right)\right]$$

Zero-plane displacement (d) is the height in meters above the ground at which zero wind speed is achieved as a result of flow obstacles such as trees or buildings. It is generally approximated as *2/3* of the average height of the obstacles.

- For example, if estimating winds over a forest canopy of height $h = 30\ m$, the zero-plane displacement would be $d = 20\ m$.

Roughness length (z_0) is a corrective measure to account for the effect of the roughness of a surface on wind flow, and is between *1/10* and *1/30* of the average height of the roughness elements on the ground.

- Over smooth, open water, expect a value around *0.0002 m*,
- over flat, open grassland $z_0 \approx 0.03\ m$,
- cropland $z_0 \approx 0.1\text{-}0.25\ m$,
- and brush or forest $z_0 \approx 0.5\text{-}1.0\ m$
- (values above *1 m* are rare and indicate excessively rough terrain).

Friction velocity (u^*) is the layer-averaged value.

Ψ(z/L) is an empirical function, which is not defined in the theory.

- Under *neutral stability conditions*, $z/L = 0$ and Ψ drops out.
- In *stable conditions* $z/L > 0$ and $\Psi < 0$.
- In *unstable conditions* $z/L < 0$ and $\Psi > 0$.

$$u_z = \frac{u_*}{k}\left[\ln\left(\frac{z-d}{z_0}\right) + \Psi\left(z/L\right)\right]$$

Empirical essence of *Ψ(z/L)* has resulted in a great variety of possible forms of it. However, historically first expressions proposed by Businger et al. (1971), Dyer (1974) and Webb (1970) still remain the most popular.

The function *Ψ(z/L)* is the correction to the logarithmic wind profile resulting from the deviation from neutral stratification.

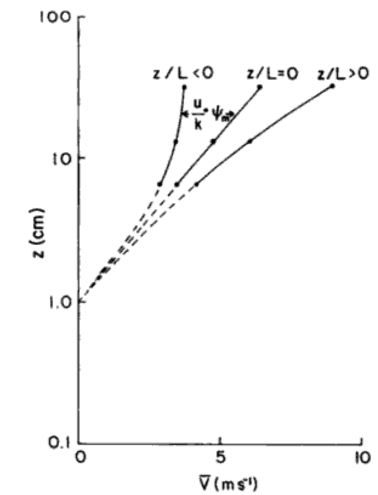


Fig. 7-4. Schematic illustration of the procedure used to compute the wind profile near the ground from observations of mean wind speed at three levels, along with the knowledge of the stability as measured by z/L. The difference between the logarithmic wind profile and the actual wind profile at any level is given by $(u_*/k)\ \psi_M$ [from Eq. (7-27); $\psi_M < 0$ when $z/L > 0$, $\psi_M > 0$ when $z/L < 0$].

from "**Mesoscale Meteorological Modeling**" R.A. Pielke, 2002.

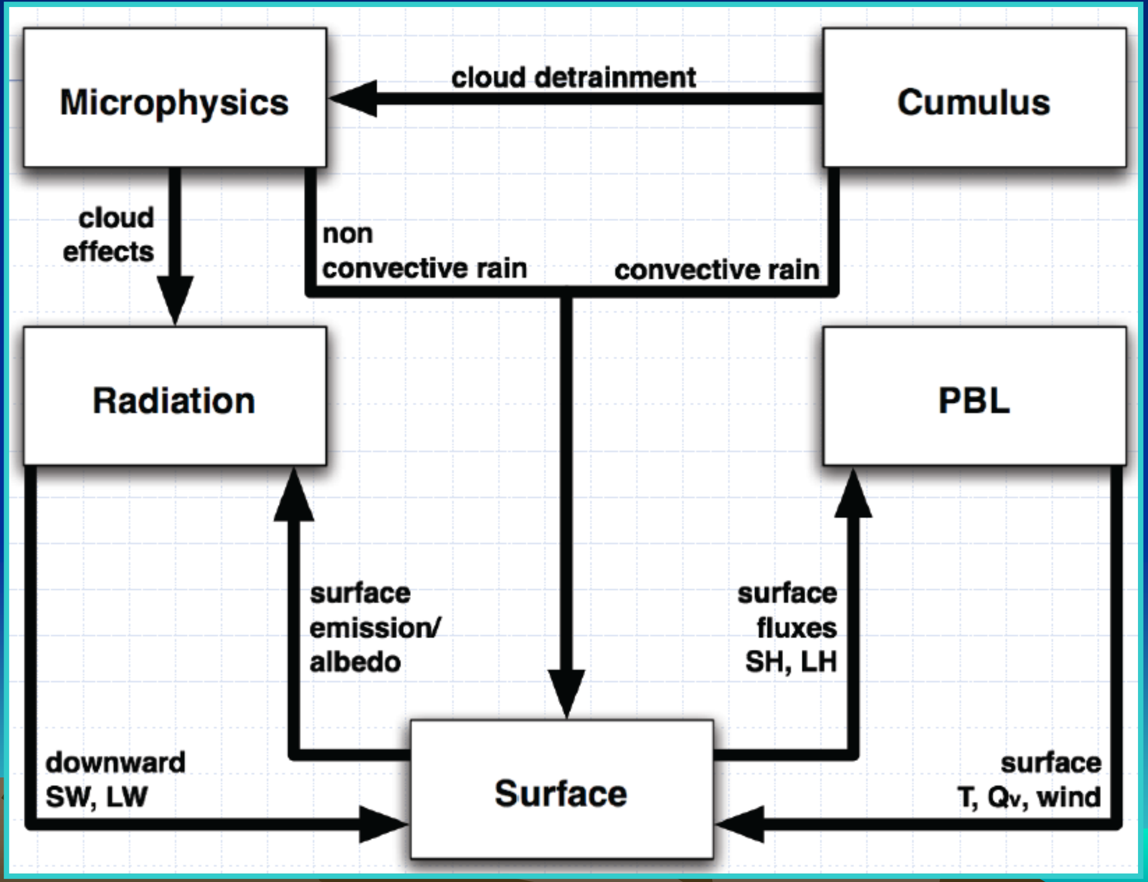
WRF Surface Layer parameterization

The surface layer schemes calculate friction velocities and exchange coefficients that enable the calculation of surface heat and moisture fluxes by the land-surface models. These fluxes provide a lower boundary condition for the vertical transport done in the PBL Schemes.

Over water surfaces, the surface fluxes and surface diagnostic fields are computed in the surface layer scheme itself.

The surface layer scheme handles the fluxes of heat, moisture and momentum from the model surface to the boundary layer above.

It also interacts with the radiation scheme as long/short wave radiation is emitted, absorbed, or scattered from the earth's surface, and with precipitation forcing from the microphysics and convective schemes.



Surface layer options available within WRF

- Similarity theory (MM5)

This scheme uses stability functions from Paulson (1970), Dyer and Hicks (1970), and Webb (1970) to compute surface exchange coefficients for heat, moisture, and momentum. A convective velocity following Beljaars (1994) is used to enhance surface fluxes of heat and moisture. No thermal roughness length parameterization is included in the current version of this scheme. A Charnock relation relates roughness length to friction velocity over water. There are four stability regimes following Zhang and Anthes (1982). This surface layer scheme must be run in conjunction with the MRF or YSU PBL schemes.

- Similarity theory (Eta)

The Eta surface layer scheme (Janjic, 1996, 2002) is based on similarity theory (Monin and Obukhov, 1954). The scheme includes parameterizations of a viscous sub-layer. Over water surfaces, the viscous sub-layer is parameterized explicitly following Janjic (1994). Over land, the effects of the viscous sub-layer are taken into account through variable roughness height for temperature and humidity as proposed by Zilitinkevich (1995). The Beljaars (1994) correction is applied in order to avoid singularities in the case of an unstable surface layer and vanishing wind speed. The surface fluxes are computed by an iterative method. This surface layer scheme must be run in conjunction with the Eta (Mellor-Yamada-Janjic) PBL scheme, and is therefore sometimes referred to as the MYJ surface scheme.

(1/5) Surface layer Similarity theory (MM5) within WRF

- The momentum flux parameterization solves for the friction velocity:

$$\tau = -\rho\overline{u'w'} = \rho u_*^2 \quad \Rightarrow \quad u_* = \left(-\overline{u'w'}\right)^{1/2}$$

This is calculated from:

$$u_* = \frac{kU}{\ln\left(\dfrac{z}{z_0}\right) - \psi_m}$$

- z_0 is specified by land-use category, is the surface roughness.
- k is used = 0.4 in MM5.
- The value of u_* is kept above *0.1 m/s* over land surface.
- The stability parameter Ψ_m is given as a function of the stability parameter:

$$\zeta = z/L$$

(2/5) Surface layer Similarity theory (MM5) within WRF

- For unstable conditions Paulson (1970):

$$\psi_m = 2\ln\left(\frac{1+x}{2}\right) + \ln\left(\frac{1+x^2}{2}\right) - 2\tan^{-1}(x) + \frac{\pi}{2}$$

Where $x = (1-\gamma_1\zeta)^{1/4}$ and Dyer and Hicks (1970) used $\gamma_1 = 16$.

- For stable conditions: $\psi_m = -\gamma_3\zeta$

in general agreement with Webb (1970) and Businger *et al.* (1971) $\gamma_3 = 5$.

The stability is determined using the Bulk Richardson number: $R_{i_B} = gz(\vartheta - \vartheta_0)/\vartheta U^2$
θ_0 being the temperature near the surface (at $z = z_0$).

For $R_{iB} > 0.2$, R_{iB} is set equal to 0.2. The value for ζ is then computed as:

- $R_{iB}\ ln(z/z_0)$ in unstable conditions and
- $R_{iB}\ ln(z/z_0)\ (1.1-5R_{iB})$^$(-1)$ in stable conditions.

This is done to avoid the need to iterate in the solution.

(3/5) Surface layer Similarity theory (MM5) within WRF

• The parameterization for sensible heat flux is similar to that for momentum flux. The characteristic temperature is:

$$\vartheta_* = -\overline{\vartheta' w'} / u_*$$

Is calculated from:

$$\vartheta_* = \frac{k(\vartheta - \vartheta_0)}{\Pr\left[\ln\left(\frac{z}{z_0}\right) - \psi_h\right]}$$

The turbulent Prandtl number P_r is set to 1 in the model, as suggested by Webb(1970).

Ψ_h has its own equations.

(4/5) Surface layer Similarity theory (MM5) within WRF

- The parameterization for latent heat flux follows Carlson and Boland (1978)

$$q_* = -\overline{q'w'}/u_*$$

(where *q'* represent fluctuations of humidity from the mean *Q*)

Is calculated from:

$$q_* = \frac{Mk\left(Q - Q_S\left(\vartheta_0\right)\right)}{\ln\left(\frac{ku_*z}{k_a} + \frac{z}{z_l}\right) - \psi_h} \qquad (**)$$

z_l (top is the molecular sublayer) is set to *0.01*.
M is a moisture availability parameter defined by land-use category.
K_a is the background molecular diffusivity set to 2.4x10-5 m2/s.

Eq. (**) is used instead of $q_* = \frac{k\left(Q - Q_0\right)}{\Pr\left[\ln\left(\frac{z}{z_0}\right) - \psi_h\right]}$ to "permit slow diffusion when turbulent transfer = 0".

Equations for *u**, *θ**, and *q** are derived empirically from surface-layer data

(5/5) Surface layer Similarity theory (MM5) within WRF

- For unstable conditions (free convection): $R_{iB} < 0$ and $\left| h/L \right| > 1.5$

$$\psi_m = -1.86\left(\frac{z}{L}\right) - 1.07\left(\frac{z}{L}\right)^2 - 0.249\left(\frac{z}{L}\right)^3$$
$$\psi_h = -3.23\left(\frac{z}{L}\right) - 1.99\left(\frac{z}{L}\right)^2 - 0.474\left(\frac{z}{L}\right)^3$$

(**)

- For unstable conditions (forced convection): $R_{iB} < 0$ and $\left| h/L \right| \le 1.5$

$$\psi_m = \psi_h = 0$$

- For mechanically driven turbulence: $0 \le R_{iB} \le R_{ic} = 0.2$

$$\psi_m = \psi_h = -5\left(\frac{R_{iB}}{1.1 - 5R_{iB}}\right)\ln\left(\frac{z}{z_0}\right)$$

- For stable conditions: $R_{iB} > R_{ic} = 0.2$

$$\psi_m = \psi_h = -10\ln\left(\frac{z}{z_0}\right)$$

h is the height of the PBL

(**) Zhang and Anthes, 1982

Going back to the momentum flux param.:

$$u_* = \frac{kU}{\ln\left(\frac{z}{z_0}\right) - \psi_m}$$

Where:

- For unstable conditions:

$$\psi_m = 2\ln\left(\frac{1+x}{2}\right) + \ln\left(\frac{1+x^2}{2}\right) - 2\tan^{-1}(x) + \frac{\pi}{2}$$

$$x = (1-\gamma_1\zeta)^{1/4}$$

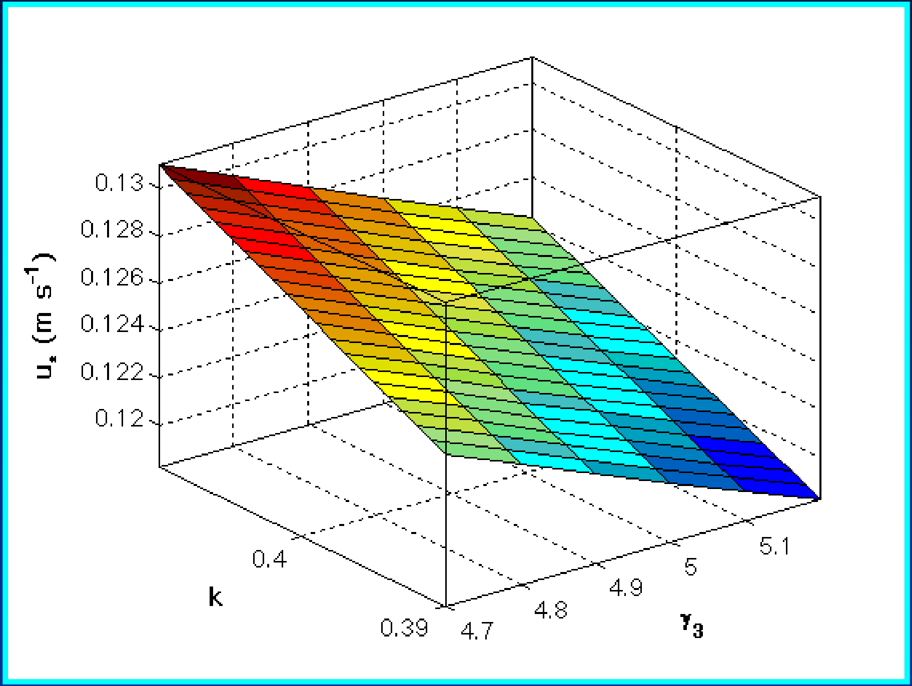
- For stable conditions:

$$\psi_m = -\gamma_3\zeta$$

Source	k	γ_1	γ_3
W70	-	-	5.2
DH70	0.41	16	-
B71	0.35	15	4.7
G77	0.41	-	-
W80	0.41	22	6.9
DB82	0.40	28	-
W82	-	20.3	-
H88	0.40	19	6.0
Z88	0.40	-	-
D74	0.41	16	5

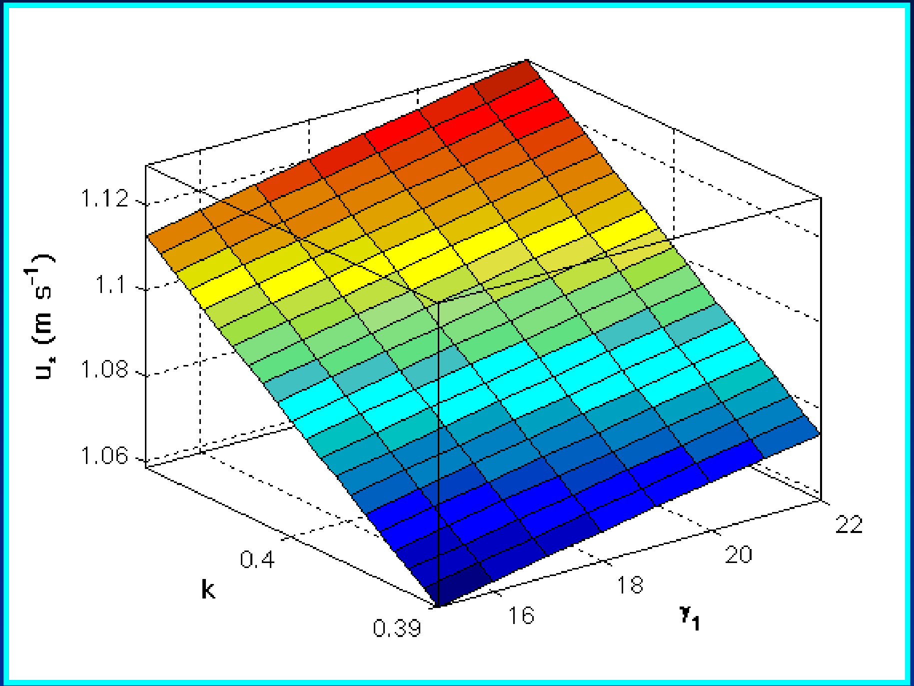
from J. Garrat and R. A. Pielke, 1989: On the sensitivity of Mesoscale Models to surface-layer parameterization constants, *Boundary-Layer Meteorol*., 48, 377-387.

Stable ($Ri_B = 0.22$)



$U = 3$ m/s;
$z = 10$ m;
$z_0 = 0.1$ m;
$g = 9.81$ m/s^2;
$\theta = 296$ K;
$\theta_0 = 290$ K;
$k = 0.39 : 0.41$
$\gamma_3 = 4.7 : 5.2$

Unstable ($Ri_B < 0$)



$U = 10$ m/s;
$z = 10$ m;
$z_0 = 0.1$ m;
$g = 9.81$ m/s^2;
$\theta = 288$ k;
$\theta_0 = 290$ k;
$k = 0.39 : 0.41$
$\gamma_1 = 15 : 22$

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