Overview of Physical Parameterizations

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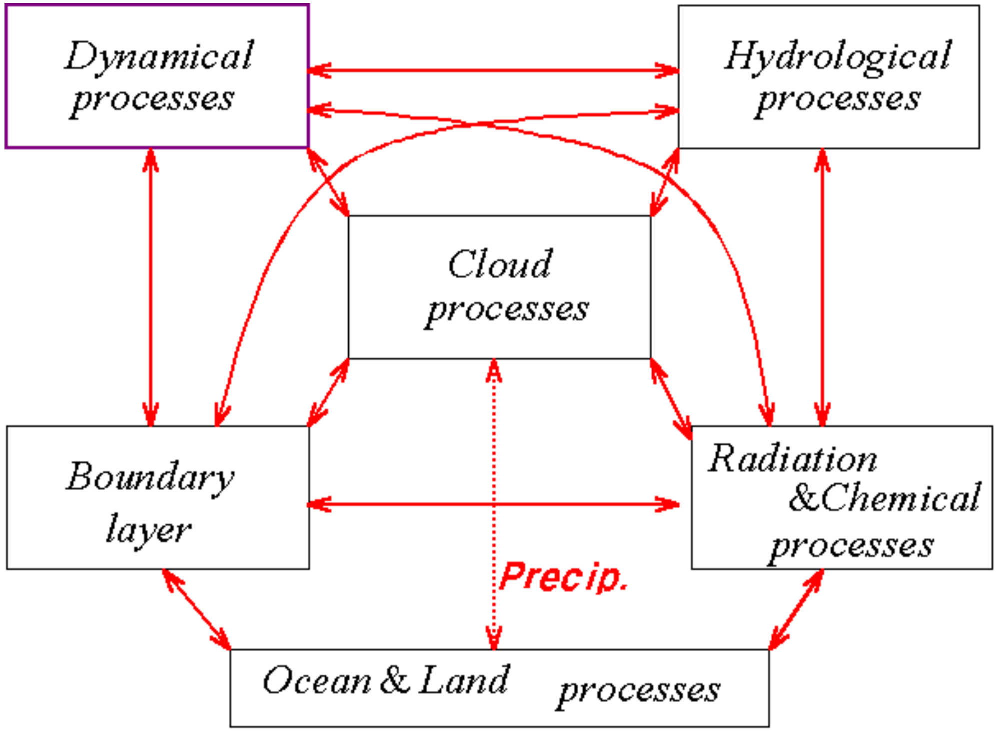
1) Concept



*** Physical process in the atmosphere**

: Specification of heating, moistening and frictional terms in terms of dependent variables of prediction model

→ Each process is a specialized branch of atmospheric sciences.

*** Parameterization**

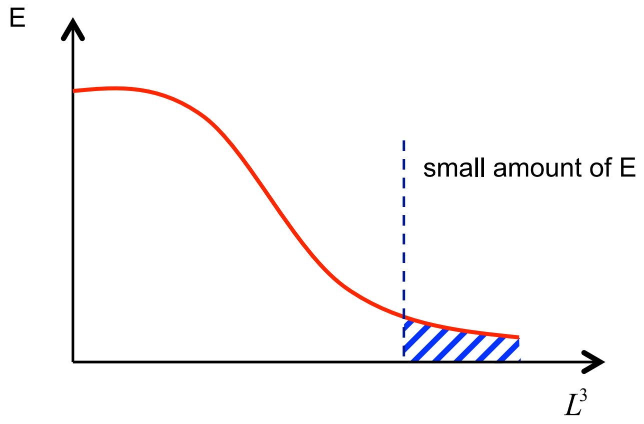
The formulation of physical process in terms of the model variables as parameters, i.e., constants or functional relations.

1) **Concept…**continued

Subgrid scale process

Any numerical model of the atmosphere must use a finite resolution in representing continuum certain physical & dynamical phenomena that are smaller than computational grid.

- Subgrid process (Energy perspective)



$\Delta x \rightarrow 0$, the energy dissipation takes place by molecular viscosity

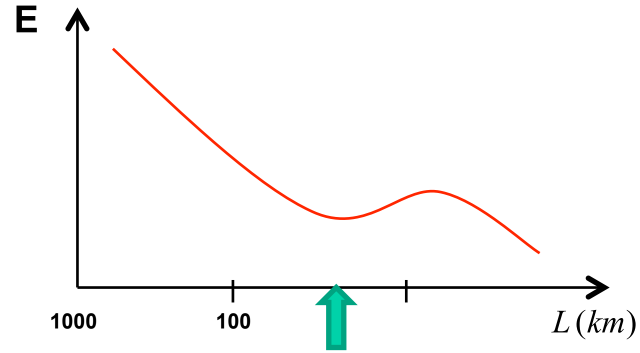
?

Objective of subgrid-scale parameterization

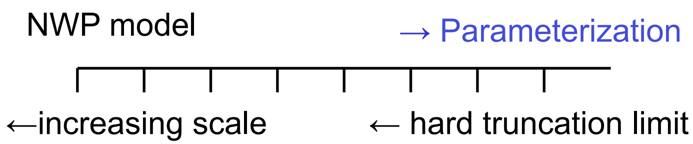
"To design the physical formulation of energy sink, withdrawing the equivalent amount of energy comparable to cascading energy down at the grid scale in an ideal situation."

1) **Concept…**continued

※ Parameterization that are only somewhat smaller than the smallest resolved scales.

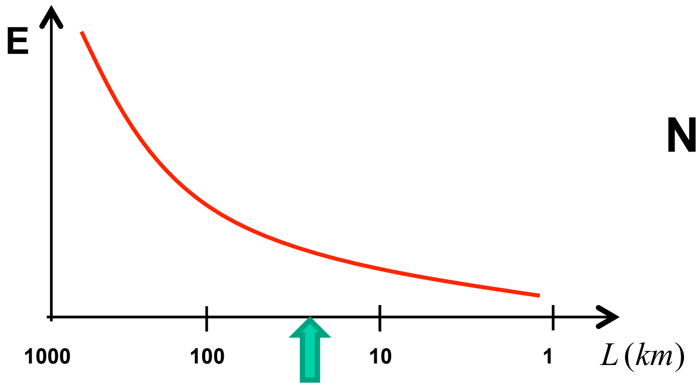


Model



where truncation limit ; spectral gap

Unfortunately, there is no spectral gap



Nature

2) Subgrid scale process & Reynolds averaging

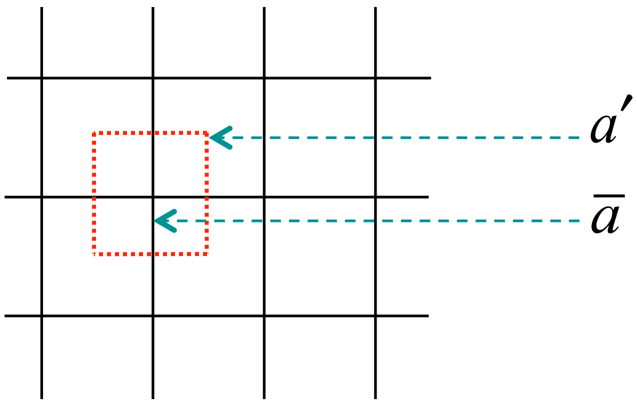
Consider prognostic water vapor equation

$$\frac{\partial \rho q}{\partial t} = -\frac{\partial \rho u q}{\partial x} - \frac{\partial \rho v q}{\partial y} - \frac{\partial \rho w q}{\partial z} + \rho E - \rho C \qquad \cdots(1)$$

In the real atmosphere,

$$u = \overline{u} + u', \qquad q = \overline{q} + q' \qquad \left(\begin{array}{l} ※\, \overline{a} : grid - resolvable \\ a' : subgrid\ scale\ perturbation \end{array} \right)$$

ρ' is neglected



2) Subgrid scale process & Reynolds averaging…continued

*** Rule of Reynolds average :** $\overline{q'} = 0,\ \ \overline{u'\overline{q}} = 0,\ \ \overline{\overline{u}\,\overline{q}} = \overline{u}\,\overline{q}$

then eq.(1) becomes

$$\frac{\partial\rho\overline{q}}{\partial t} = \underbrace{-\frac{\partial\rho\overline{u}\overline{q}}{\partial x} - \frac{\partial\rho\overline{v}\overline{q}}{\partial y} - \frac{\partial\rho\overline{w}\overline{q}}{\partial z}}_{①} \underbrace{- \frac{\partial\rho\overline{u'q'}}{\partial x} - \frac{\partial\rho\overline{v'q'}}{\partial y} - \frac{\partial\rho\overline{w'q'}}{\partial z}}_{②} + \rho E - \rho C \quad \cdots(2)$$

① grid-resolvable advection (dynamical process)
② turbulent transport

*** How to parameterize the effect of turbulent transport**

a) $-\rho\overline{w'q'} = 0$: 0th order closure

b) $-\rho\overline{w'q'} = K\dfrac{\partial\overline{q}}{\partial z}$: 1st order closure (K-theory)

c) obtain a prognostic equation for $\overline{w'q'}$ from (1), (2)

$$\frac{\partial\rho wq}{\partial t} = -\frac{\partial\rho uwq}{\partial x} + \cdots$$

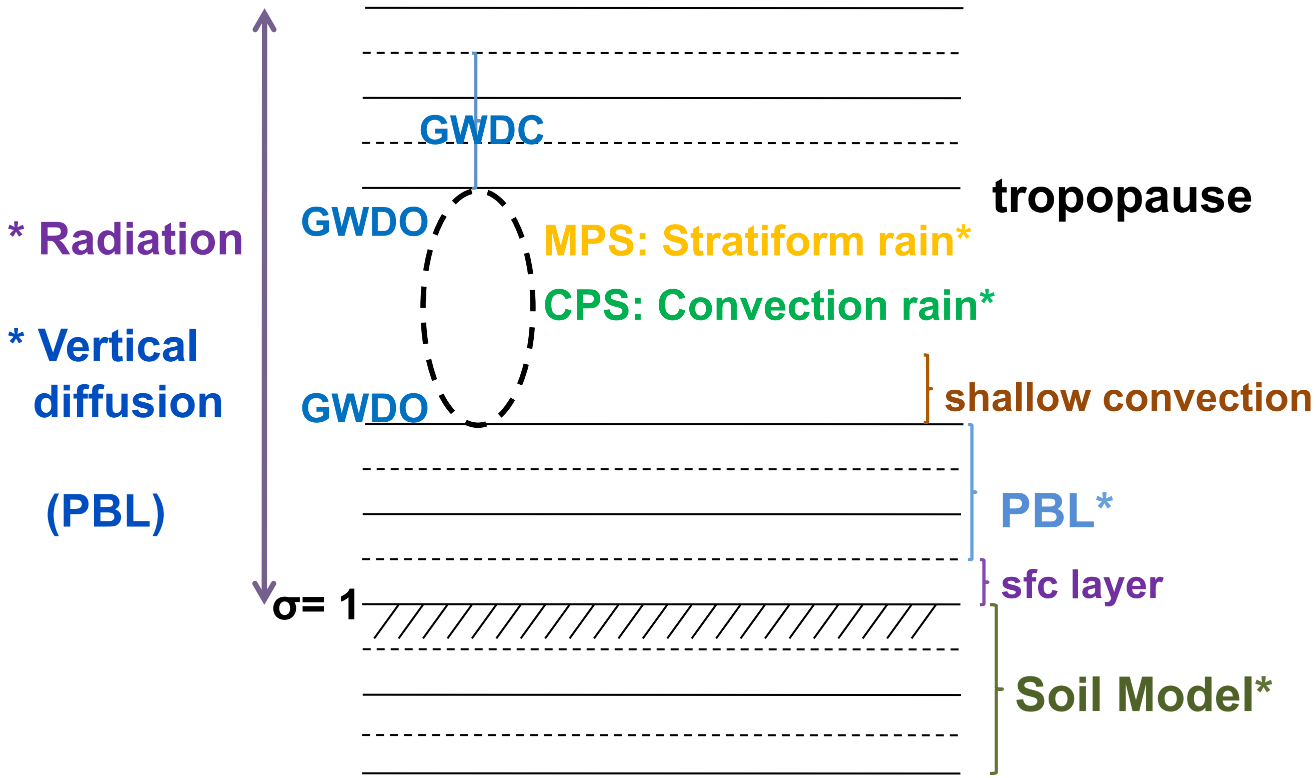
taking Reynolds averaging,

$$\frac{\partial\rho\overline{w'q'}}{\partial t} = \frac{\partial\rho\overline{w'w'q'}}{\partial z}$$

$$-\rho\overline{w'w'q'} = K'\frac{\partial\rho\overline{w'q'}}{\partial z}$$: 2nd order closure

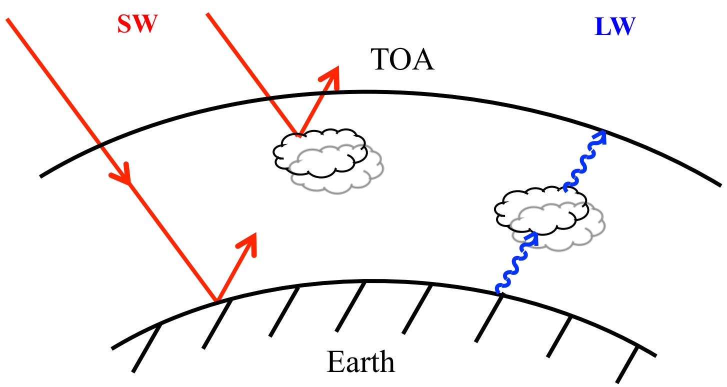
3) Schematic of physics algorithm : In modeled atmosphere : 6* ~9

$$\frac{d\ln\theta}{dt} = \frac{H}{c_p T}, \quad \frac{dq}{dt} = S, \quad \frac{d\vec{u}}{dt} = \nabla_z \vec{\tau}$$



1. Radiation

1.1 Concept



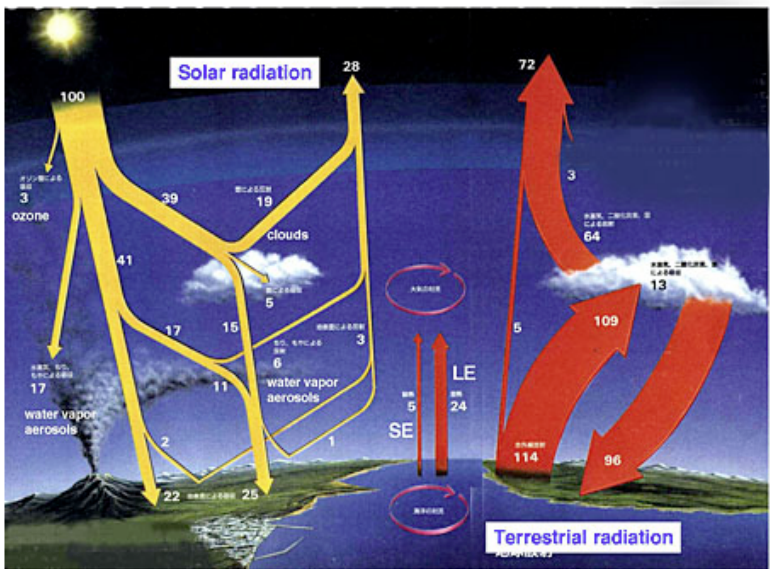
TOA : S = 1360 Wm^{-1}

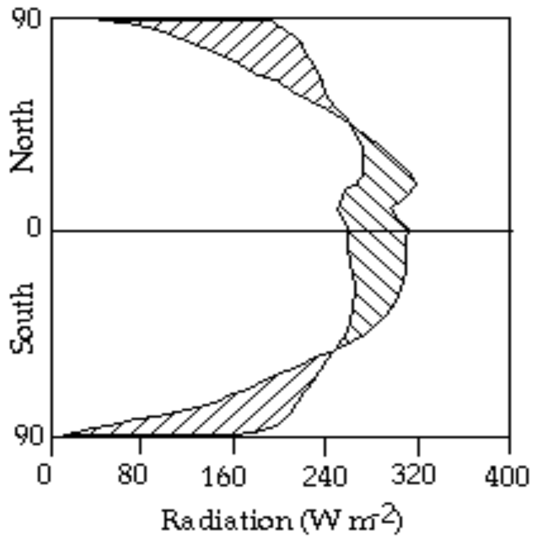
Mean Flux : $\frac{S}{4} = 340\ Wm^{-1}$ → Energy source for Earth

30% : reflected from the atmosphere clouds

→ Back to space by terrestrial infrared radiation

$$\left\{ \begin{array}{l} \text{25\% : absorbed in the atmosphere} \\ \text{45\% : absorbed at the earth surface} \end{array} \right\}$$



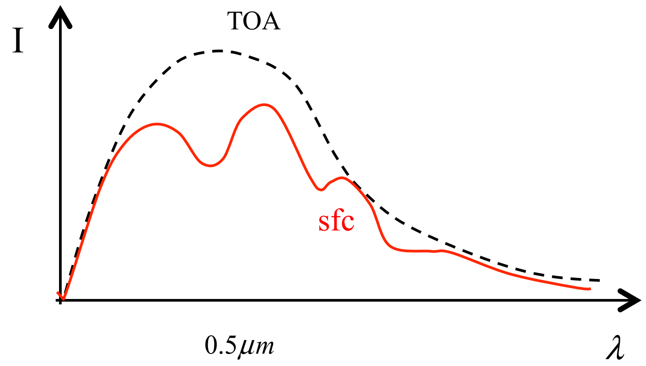


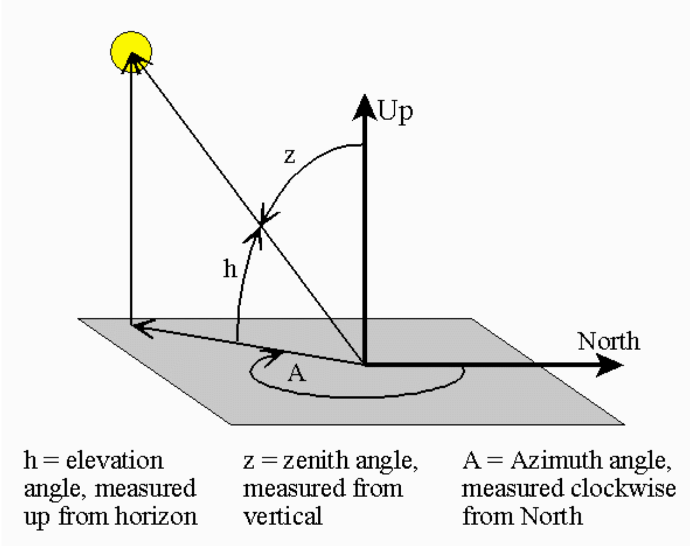
Energy loss

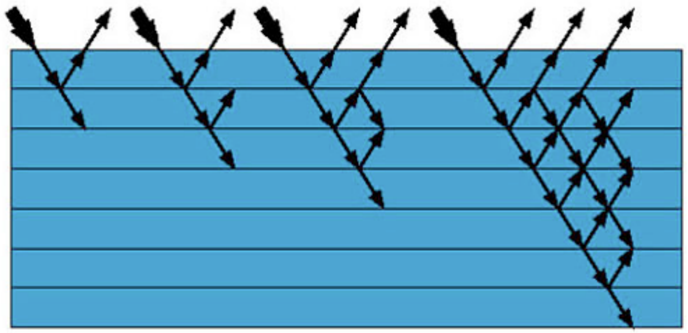
Energy gain

Energy loss

1.2 Solar radiative transfer







- At TOA,

$$F = S\left(\frac{dm}{d}\right)^2 \cos\theta_0 \qquad (\theta_0 \text{ :Zenith angle})$$

(Insolation)

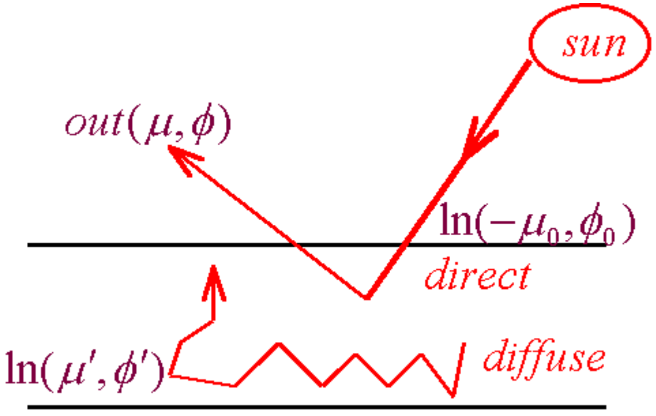
- Basic equations $\quad \mu = \cos\theta$

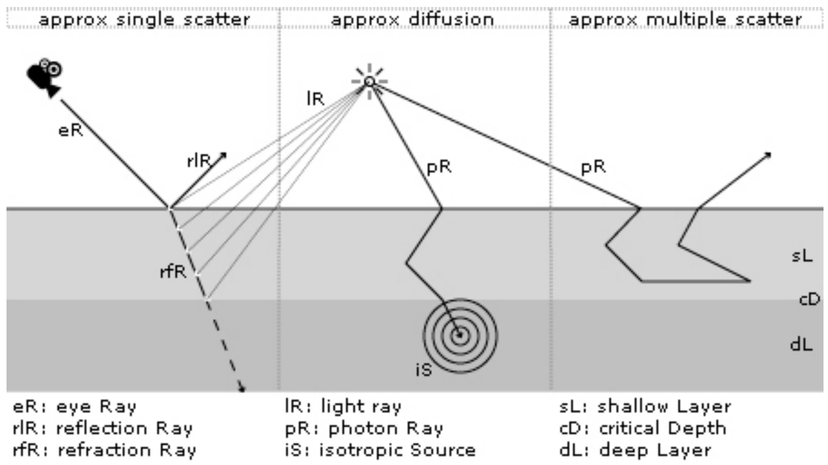
$$\mu \frac{dI(\tau,\mu,\phi)}{d\tau} = I(\tau,\mu,\phi) - J(\tau,\mu,\phi)$$

absorption source emission

$$d\tau = -k_v \rho_a dz \qquad \tau(\text{optical depth}) = \int_z^{z_\infty} k_v(z')\rho_a(z')dz'$$

$$= \int_0^p k_v(p')q(p')\frac{dp'}{g}$$





$$J = J(\tau,\mu,\phi) = \frac{\tilde{\omega}}{4\pi}\int_0^{2\pi}\int_{-1}^{1} I(\tau,\mu',\phi')P(\mu,\phi;\mu',\phi')\,d\mu'\,d\phi \text{[diffuse (multiple) scattering]}$$

$$+\frac{\tilde{\omega}}{4\pi}F_0 P(\mu,\phi;-\mu_0,\phi_0)e^{-\frac{\tau}{\mu_0}} \text{[single(direct) scattering]}$$

P : Scattering phase function :redirects (μ',ϕ') (μ,ϕ)

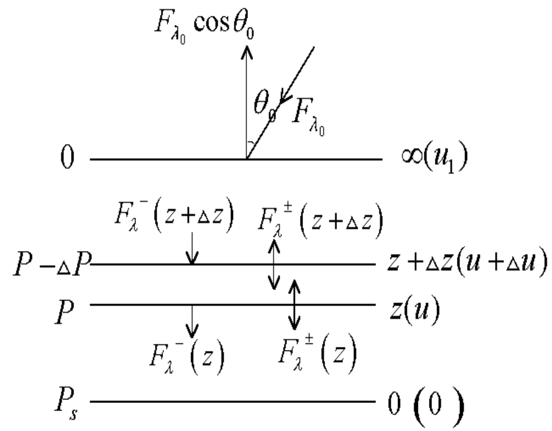
$\tilde{\omega} = \dfrac{\sigma_s}{\sigma_e}$: Scatteing albedo

scattering cross section/extinction(scattering + absorption) cross section

* remove ϕ dependency using $P(\cos\theta)$ function

* P, $\tilde{\omega}$, Albedo depend on λ, particle size & shape.

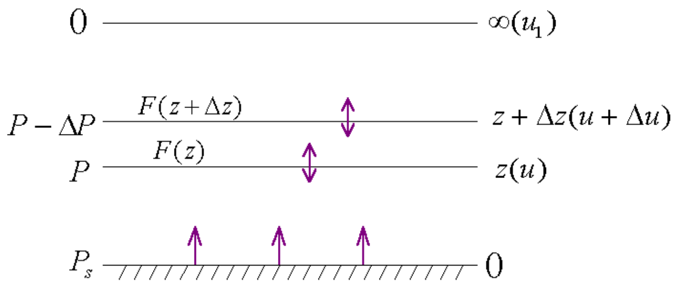
$$P(\cos\phi) = \sum_{l=0}^{N}\tilde{\omega}_l P_l(\cos\phi) \quad \text{: Legendre Polynomial}$$

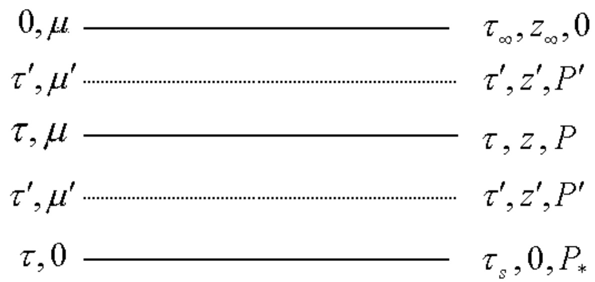


Radiative transfer equation solver.

→ Discrete - ordinates method
Two - Stream and Eddington's approximation
Delta - function adjustment and similarity principle
δ - Four stream approximation

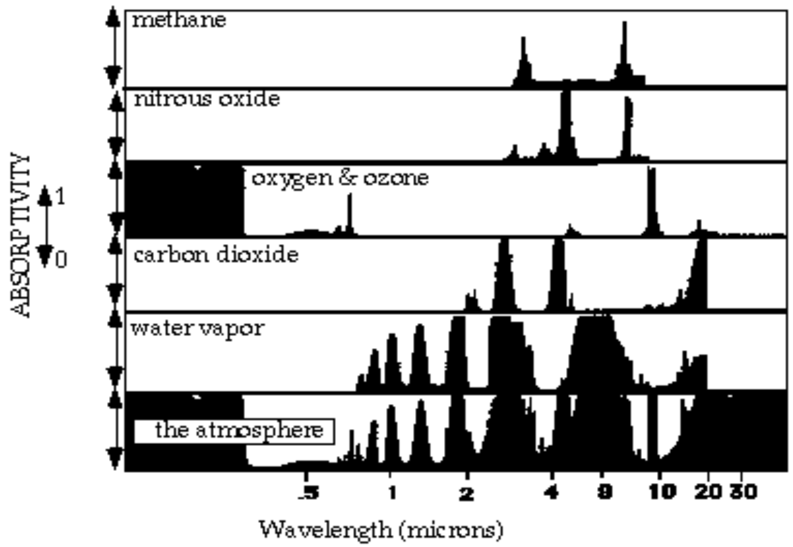
1.3 Terrestrial radiation

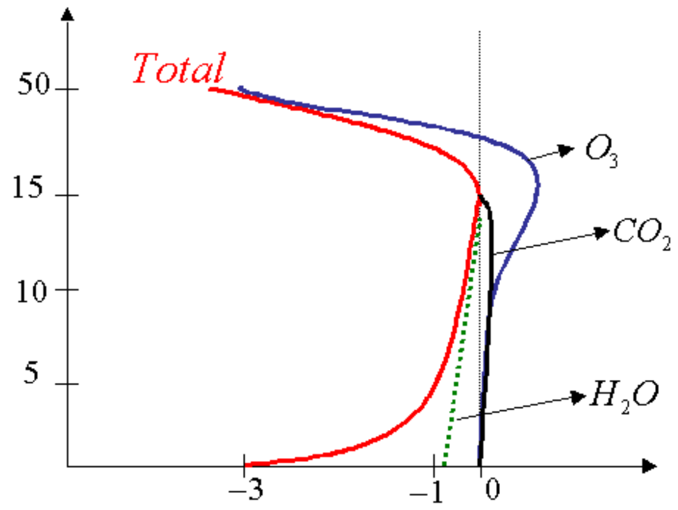




$$F(z) = F^{\uparrow}(z) - F^{\downarrow}(z)$$
$$\Delta F = F(z+\Delta z) - F(z)$$

$$\left.\frac{\partial T}{\partial t}\right|_{IR} = -\frac{1}{c_P\rho}\frac{\Delta F}{\Delta P} = -\frac{g}{c_P}\frac{\Delta F}{\Delta u}$$



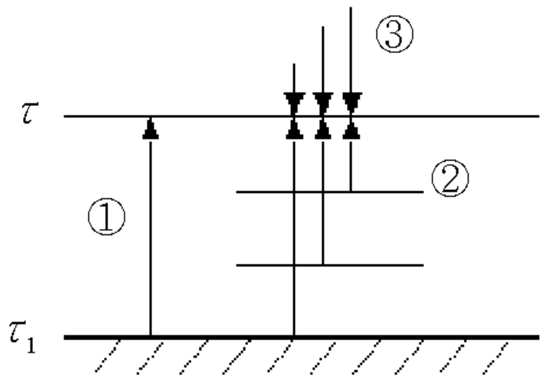


In spectral bands (monochromatic)

$$\uparrow \quad \mu \frac{dI_\nu(\tau,\mu)}{d\tau} = I_\nu(\tau,\mu) - B_\nu(T)$$

$$\downarrow \quad -\mu \frac{dI_\nu(\tau,-\mu)}{d\tau} = I_\nu(\tau,-\mu) - B_\nu(T)$$

B.C. $\begin{cases} SFC\ (\tau=\tau_1),\ I_\nu(\tau,\mu) = B_\nu(T_s) \\ TOP\ (\tau=0),\ I_\nu(0,-\mu) = 0 \end{cases}$



$$F^\uparrow(\tau) = 2\pi B_\nu(T_s)\int_0^1 e^{-\frac{(\tau_1-\tau)}{\mu}}\mu\, d\mu + 2\int_0^1\int_\tau^{\tau_1} \pi B_\nu[T(\tau')]e^{-\frac{(\tau'-\tau)}{\mu}}\, d\tau' d\mu$$

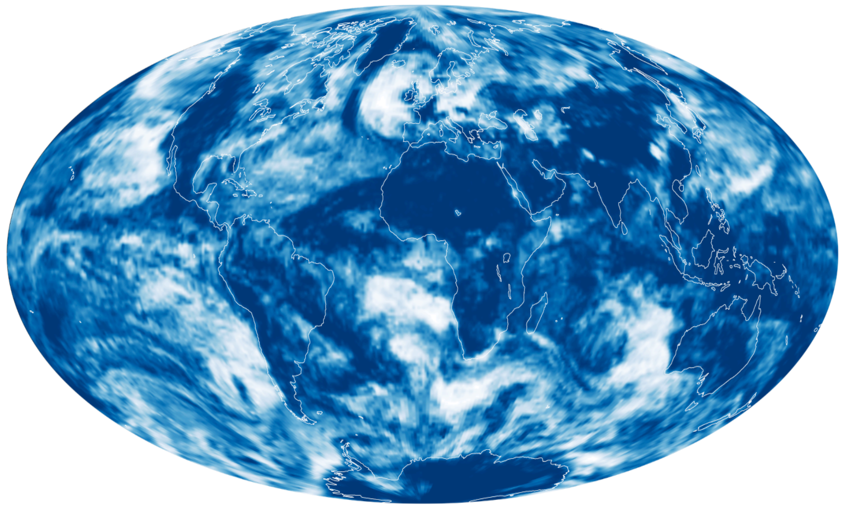
$$F^\downarrow(\tau) = 2\int_0^1\int_0^\tau \pi B_\nu[T(\tau')]e^{-\frac{(\tau-\tau')}{\mu}}\, d\tau' d\mu$$

$$d\tau = -k_\nu \rho dz \qquad \tau_1 = \int_0^{u_1} k_\nu du, \qquad u_1 = \int_0^\infty \rho dz$$

LW is time consuming !

1.4 Cloud fraction

i) Conventional method (Slingo)



$f = f_c + f_l$

f_c : depends on precipitation, p_{top}, p_{bottom}

f_l : depends on RH $= 1 - \left[\dfrac{1 - RH}{1 - RH_0}\right]^{0.5}$

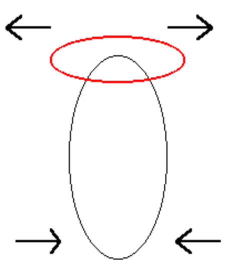
where RH_0 is the critical value of RH,

which is optimized based on observations

ii) Advanced method (Chou)

- inclusion of ice, liquid
- consistent treatment of water substance for both precipitation & radiative properties.

f_c : uses information of detrained water substances from sub-grid scale clouds in convective p arameterizations



detrained

$u_{q_c}, u_{q_l}, k_{q_c}, k_{q_l}$

in radiation code

f_l : use the hydrometeor information from microphysics routine, $q_c, q_s, q_i, \cdots$

iii) Radiation properties

$$\tau_i^c = cwp[a_i + \frac{b_i}{r_{ei}}] f_{ice} \text{ (optical thickness)}$$

$$w_i^c = 1 - c_i - d_i r_{ei} \text{ (co-albedo)}$$

$$g_i^c = e_i - f_i r_{ei} \text{ (asymmetry factor)}$$

$$f_i^c = (g_i)^2$$

a-f : coeff : depends upon band and k-

$$\overline{\tau_c} = \sum_i \tau_i \qquad i : \text{each gas}$$

(The effective optical thickness for each spectral band)

- The long wave cloud emissivity (E_{cld})

$$c_f' = E_{cld} c_f$$

$$E_{cld} = 1 - e^{-Dk_{abc}cwp}$$

where $D = 1.66$: diffusivity factor

$k_{abc} = k_l(1 - f_{ice}) + k_i f_{ice}$: absorptivity coefficient

- For diagnostic microphysics scheme,

- cloud water scale height

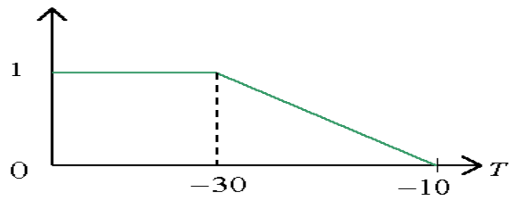
$$h_1 = a \ln(1.0 + \frac{b}{g}\int_{P_T}^{P_S} q dp)$$

- cloud droplet size,

$$r_{ee} \begin{cases} = 10\mu m & \text{over ocean} \\ < 10\mu m & \text{over land} \end{cases} : \text{warm clouds}$$

r_{ei}: $10\mu m$ (low) ~ $30\mu m$: (high) ice clouds

– ice fraction, f_{ice}



- For prognostic microphysics scheme,

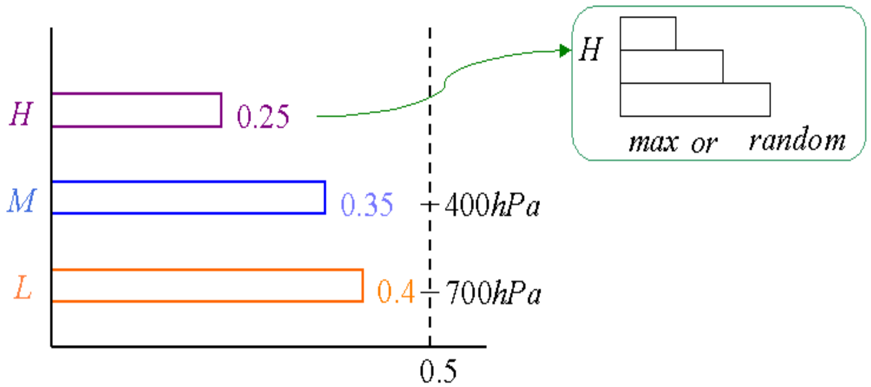
: Broadband radiation : Radiative properties are explicitly computed from prognostic water substances

: Simplified radiation : Dudhia (1989)

$$\alpha_p \text{(absorption coefficient)} = \frac{1.66}{2000}\left(\frac{\pi N_0}{\rho_{rs}^3}\right)^{\frac{1}{4}} \quad m^2 g^{-1} = \begin{cases} 2.34\times10^{-3} \quad m^2 g^{-1} & \text{for snow} \\ 0.33\times10^{-3} \quad m^2 g^{-1} & \text{for rain} \end{cases}$$

$$u_p \text{ (effective water path length)} = (\rho q_{rs})^{\frac{3}{4}} \Delta z \times 1000 \quad gm^{-2} \rightarrow \tau_p \text{(transmission)} = \exp(-\alpha_p u_p)$$

iv) Cloud overlapping



Maximum overlapping : 0.4
Minimum overlapping : 1.0
Random overlapping : $H+(1-H)M+\{1-H-(1-H)M\}L=0.6$

- Computation :

τ is scaled by A_c (cloud cover) at a given layer.

- Flux for each of $A_c, (1-A_c) \rightarrow$ summation

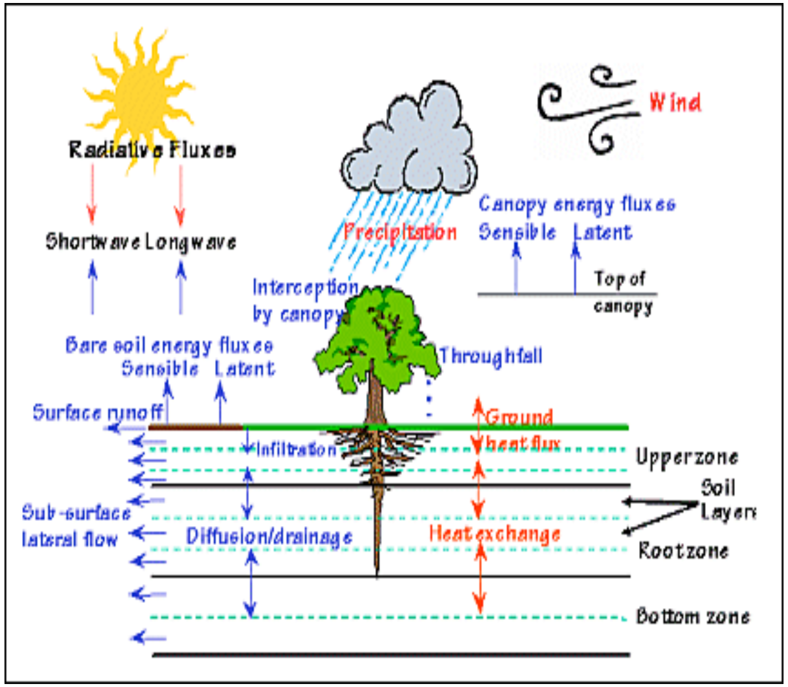
- Issues : $A_c = 0$ or 1 ← in WRF versus partial cloudiness in GFS

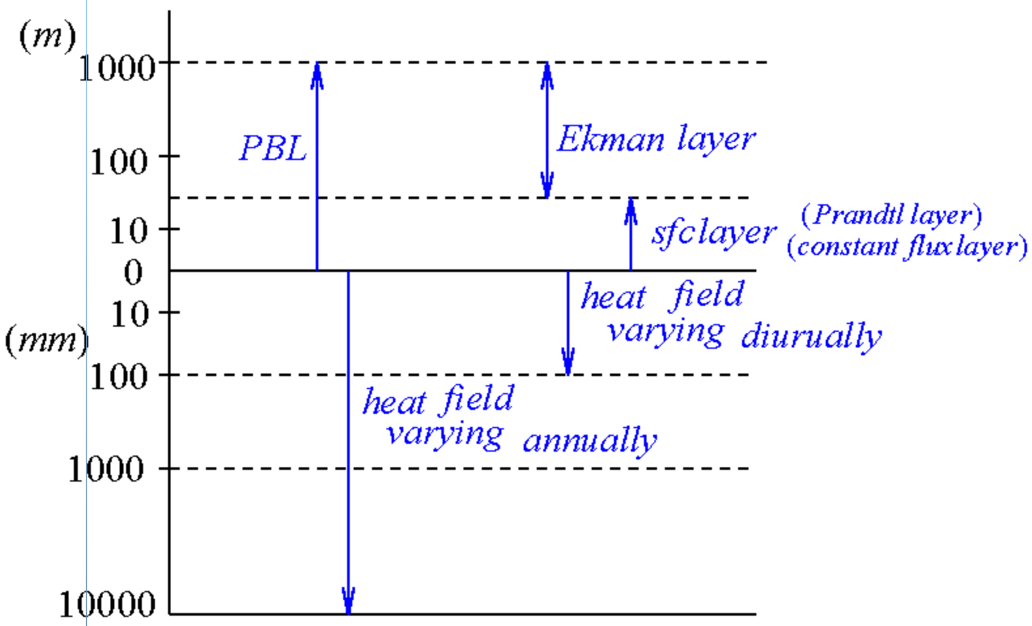
Radiation ← *reflect SW, LW* — Cloud

Radiation → *enhance or reduce cloud activity* — Cloud

2. Land-surface processes

2.1 Concept : Surface layer + soil model





Atmospheric surface layer : 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 atmospheric models, it is defined the height of the lowest model level.

2.2 Surface layer parameterization

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 schem e itself. Sea surface temperature can be predicted by the surface energy budget and mixed layer mixing

1) Bulk method

$$H_0 = \rho C_p C_H \left|\vec{V}_a\right| \Delta T$$

$$E_0 = \rho L C_H \left|\vec{V}_a\right| \Delta q M_a$$

$$\vec{\tau}_0 = \rho C_D \left|V_a\right| \vec{V}_a$$

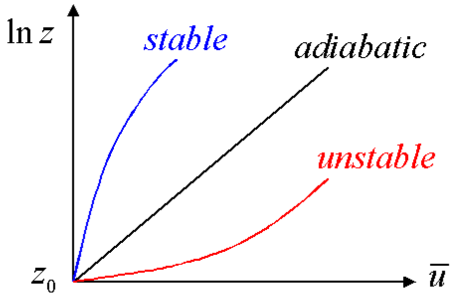
C_D, C_H : prescribed

$= 0.01$ over land, $= 0.001$ over water

2) Monin-Obukov similarity

$$\frac{k_z}{u_*}\frac{\partial u}{\partial z} = \phi_m(z/L), \quad \frac{k_z}{u_*}\frac{\partial \theta}{\partial z} = \phi_t(z/L)$$

Integrate, $F_m = \int_{z_0}^{h_s} \frac{dz}{z}\phi_m dz = \ln(\frac{h_s}{z_0}) - \psi_m(h_s, z_0, L)$



$$\vec{u} = C_u[\ln\frac{z}{z_0} + \Phi]$$

Φ : curving factor

※ **Profile function** : ϕ_m and ϕ_t

Dyer and Hicks formula for similarity
(Businger formula : complex)

- unstable (L < 0)

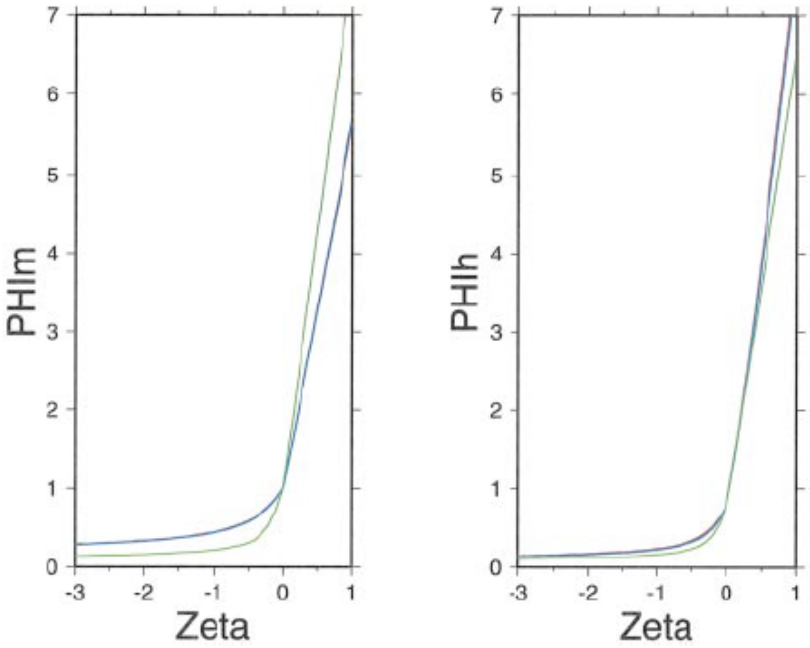
$$\phi_m = (1 - 16\frac{0.1h}{L})^{-\frac{1}{4}} \quad \text{for } u, v$$

$$\phi_t = (1 - 16\frac{0.1h}{L})^{-\frac{1}{2}} \quad \text{for } \theta, q$$

- stable (L > 0)

$$\phi_m = \phi_t = (1 + 5\frac{0.1h}{L})$$

where $L = u_*^2\overline{\theta} / (kg\theta_*) = -\dfrac{\rho C_p \theta_0 u_*^3}{kgH_0}$



Kantha (2003)

$$\frac{h_s}{L} = \frac{\phi_m^2(hs/L)}{\phi_t(hs/L)} Ri = \zeta \text{ (Zeta)}$$

where $Ri = \dfrac{g}{\overline{\theta}}\dfrac{\partial\theta}{\partial z} \Big/ (\dfrac{\partial u}{\partial z})^{-2}$

※ **Useful relation** :

Given the F_m, F_H $\quad C_D = k^2 / F_m{}^2, C_Q = C_H = k^2 / (F_m F_t) \quad u_* = kU / F_m$

$$\tau_0 = \rho k_m \frac{du}{dz} = -\overline{u'w'} = \rho C_p U^2$$

$$H_0 = -\rho C_p k_h \frac{d\theta}{dz} = \rho C_p \overline{\theta' w'} = -\rho C_p C_H U \Delta\theta$$

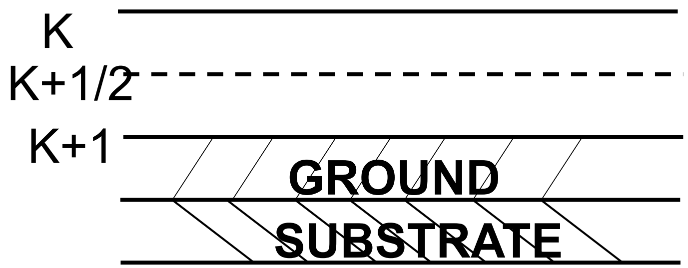
$$E_0 = -\rho L \overline{q'w'} = -\rho L C_q U \Delta q$$

2.3 Soil model

1) Slab model : force-restore method

$$\frac{\partial T_s}{\partial t} = \lambda_T (R_n - LE - H) - \frac{2\pi}{\tau}(T_s - T_a)$$

$$\frac{\partial T_m}{\partial t} = \frac{1}{\tau}(T_s - T_m)$$ **: substrate T, daily mean**



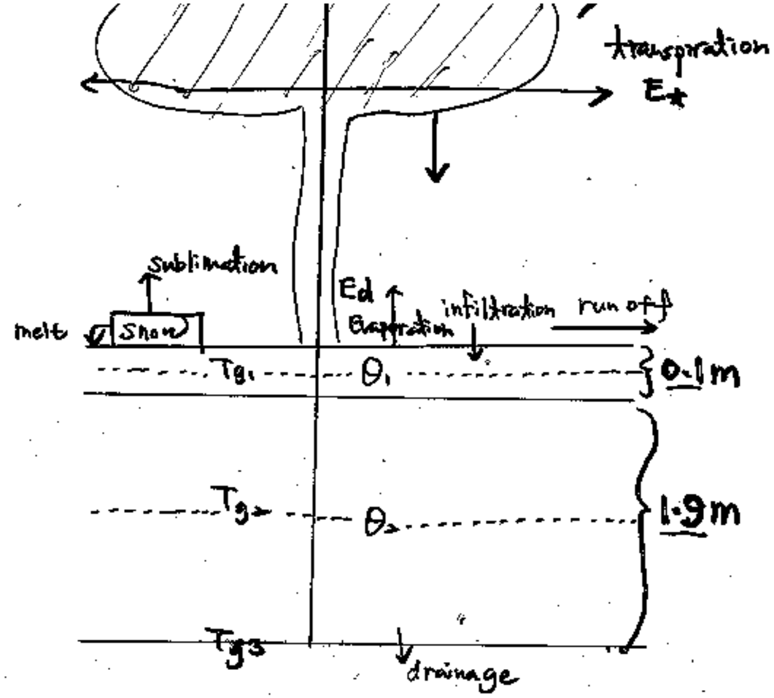
2) Multi-layer model : OSU method

$$\frac{\partial T_s}{\partial t} = \lambda_T (R_n - LE - H)$$ **: surface T**

$$(\rho C)_i \frac{\partial T_g}{\partial t} = \frac{\partial}{\partial z}\left(\lambda T_g \frac{\partial T_g}{\partial z}\right)$$ **: soil T**

$$\frac{\partial \Theta}{\partial t} = \frac{\partial}{\partial z}\left(D \frac{\partial \Theta}{\partial z}\right) + K\frac{\partial \Theta}{\partial z} + F_\Theta$$ **: soil moisture**

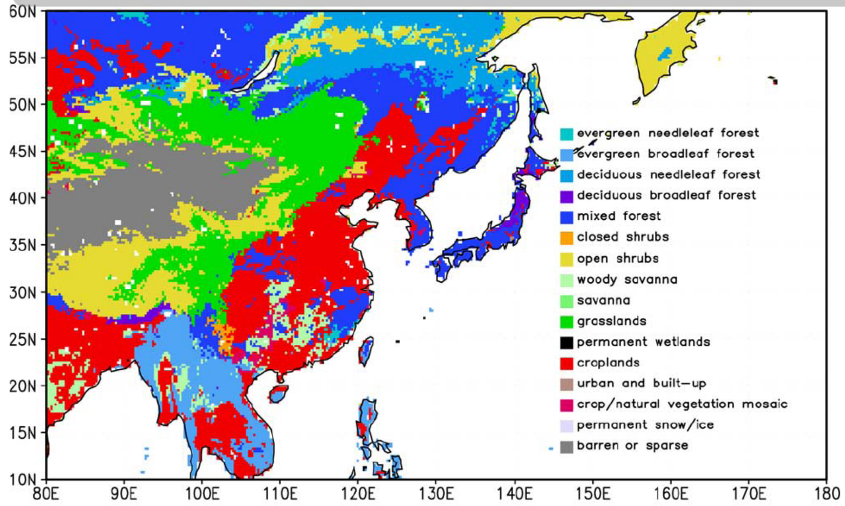
- NOAH, SIB, PLACE, VIC, CLM, etc



2.5. Vegetation type → z_0, Albedo

Vegetation types

MODIS (satellite)

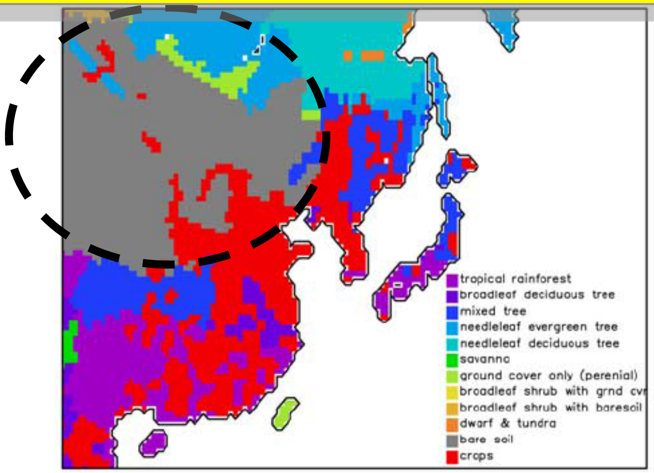


13 type data set
1 degree

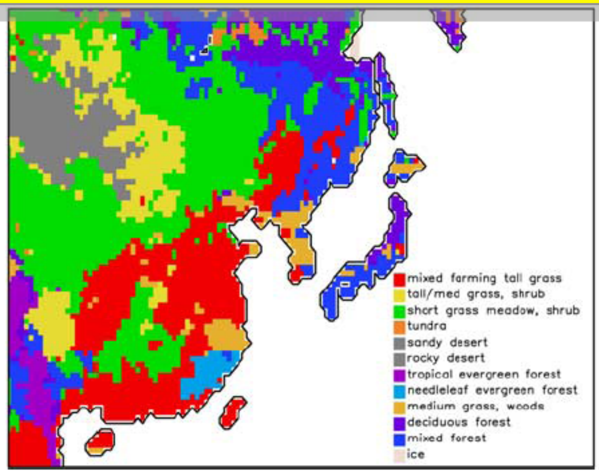
12 type data set
20 min

The Simple Biosphere model (SiB)

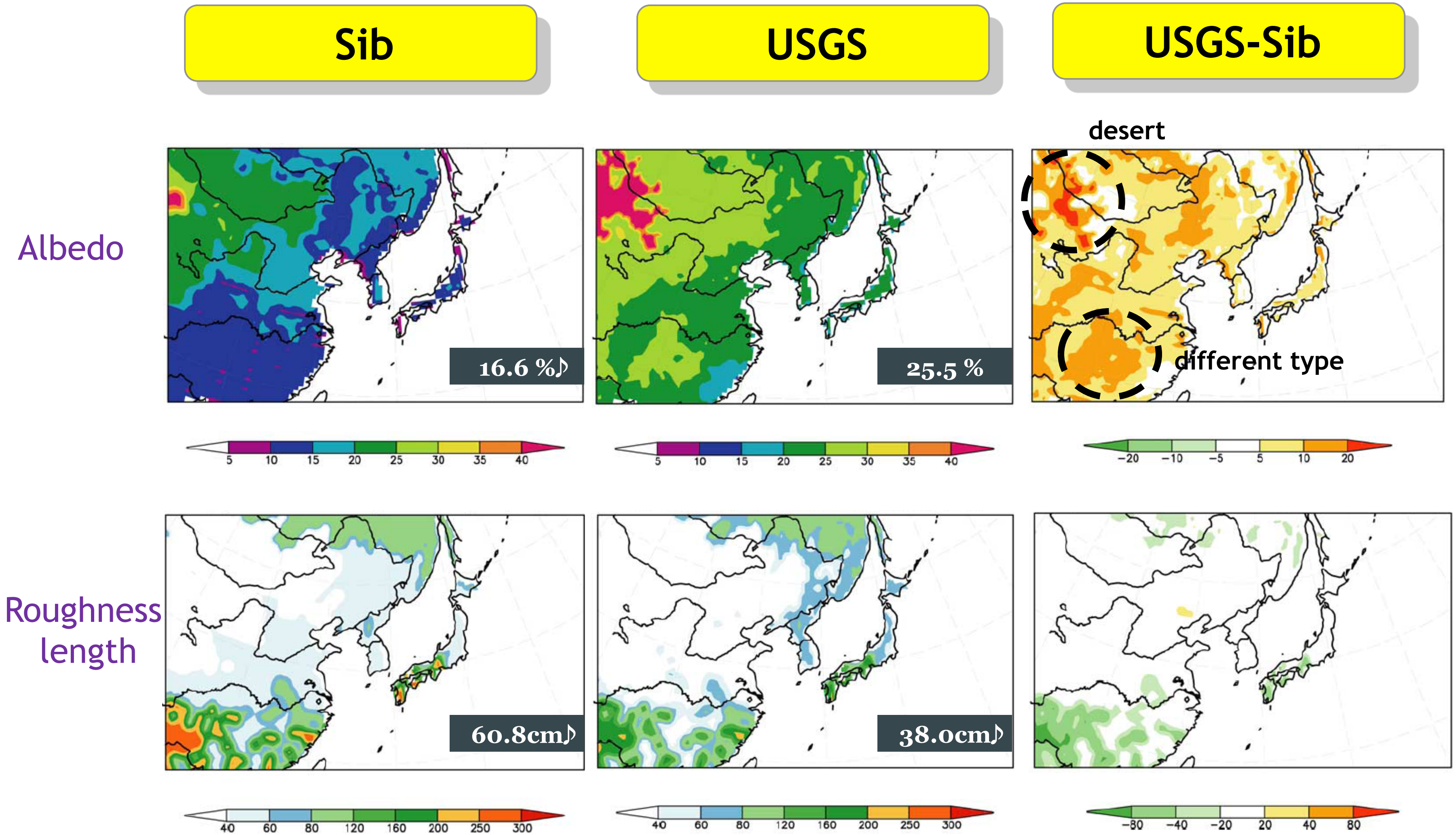
quite broad
desert region



United States Geological Survey's (USGS)



Albedo and Roughness length (z_0)

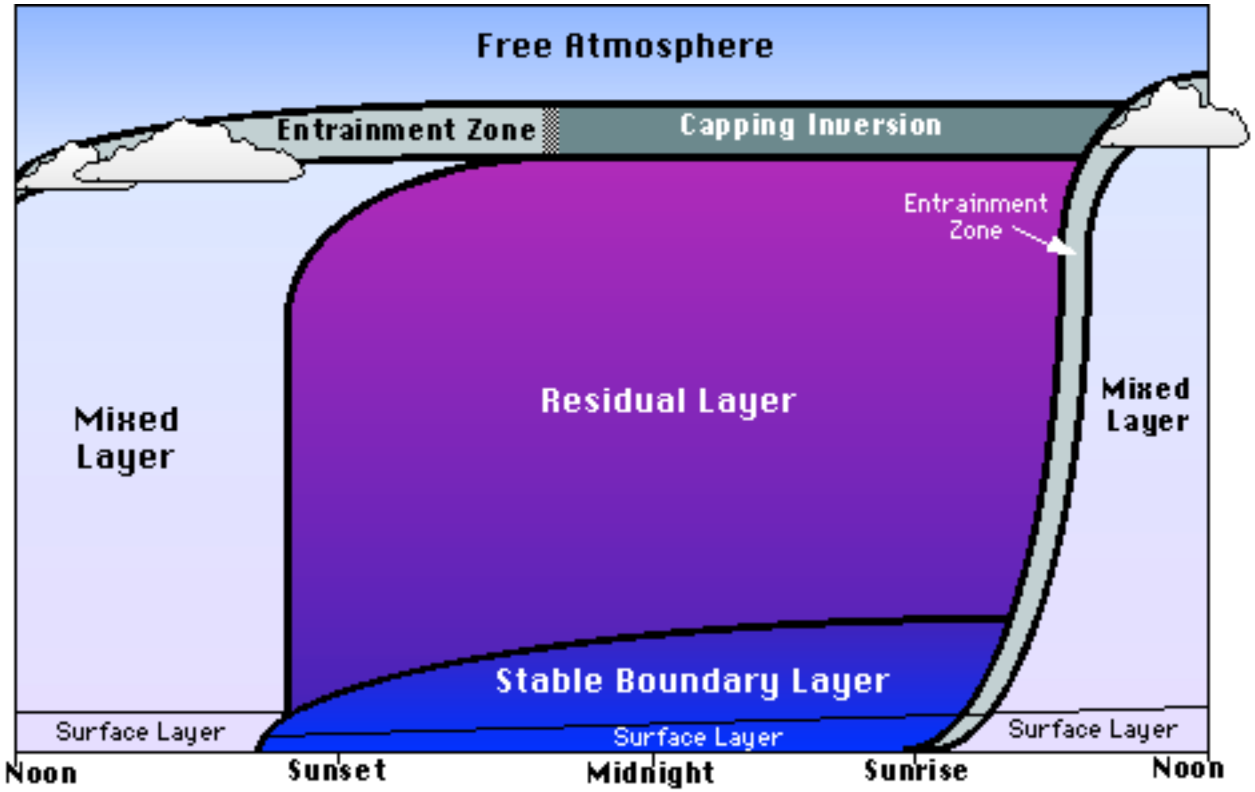


Kang and Hong (2008)

3. Vertical diffusion (PBL)

3.1 Concept

- computes the parameterized effects of vertical turbulent eddy diffusion of momentum, water vapor and sensible heat fluxes



Structure of Planetary Boundary Layer
adapted from Stull, 1988.

3.2 Planetary Boundary Layer Structure : schematic

Daytime profiles

Daytime flux profiles

Nighttime flux profiles

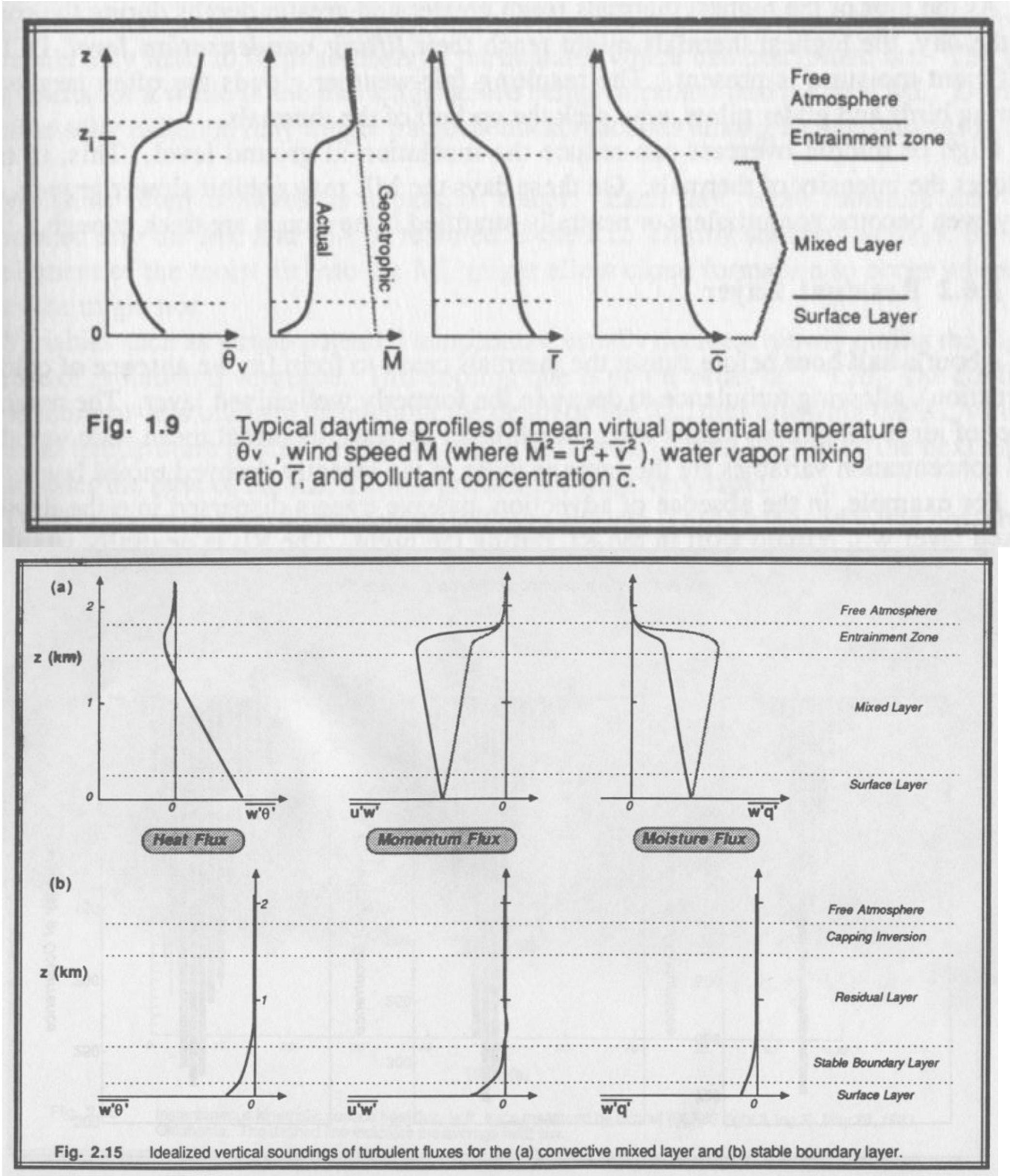


Fig. 1.9 Typical daytime profiles of mean virtual potential temperature $\overline{\theta}_v$, wind speed $\overline{M}$ (where $\overline{M}^2 = \overline{u}^2 + \overline{v}^2$), water vapor mixing ratio $\overline{r}$, and pollutant concentration $\overline{c}$.

Fig. 2.15 Idealized vertical soundings of turbulent fluxes for the (a) convective mixed layer and (b) stable boundary layer.

Stull (1988)

3.3 Classifications : how to determine, k_c

i) Local diffusion (Louis 1979)

$$\frac{\partial c}{\partial t} = \frac{\partial}{\partial z}(-\overline{wc}) = \frac{\partial}{\partial z}(k_c \frac{\partial c}{\partial z}) \quad k_c: \text{ diffusivity}, \quad k_m, k_t = l^2 f_{m,t}(Ri)\left|\frac{\partial U}{\partial z}\right|$$

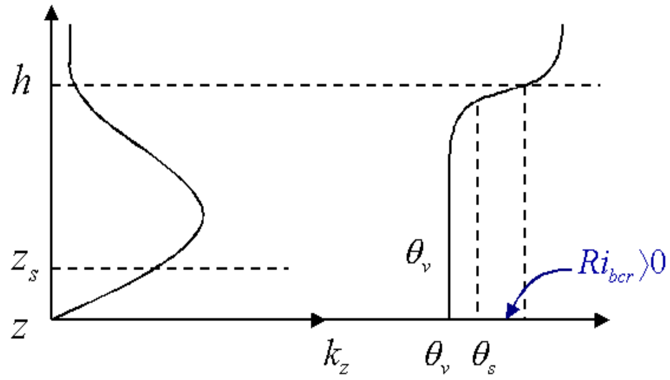
Local Richardson number

ii) Nonlocal diffusion (Troen and Mahrt 1986)

$$\frac{\partial c}{\partial t} = \frac{\partial}{\partial z}(-\overline{wc}) = \frac{\partial}{\partial z}(k_c(\frac{\partial c}{\partial z} - \gamma_c))$$

$$k_{zm} = k w_s z(1-\frac{z}{h})^p, \; h = R_{ibcr}\frac{\theta_m}{g}\frac{U^2(h)}{(\theta_v(h)-\theta_s)}$$

$$\theta_s = \theta_{va} + \theta_T \; (= b\frac{\overline{(\theta_v' w')_0}}{w_s}), \; w_s = u_* \phi_m^{-1}$$



iii) Eddy mass-flux diffusion (Siebesma and Teixeria 2000)

$$\frac{\partial c}{\partial t} = \frac{\partial}{\partial z}(-\overline{wc}) = -\frac{\partial}{\partial z}[-k_c \frac{\partial \overline{c}}{\partial z} + M(c_u - \overline{c})]$$

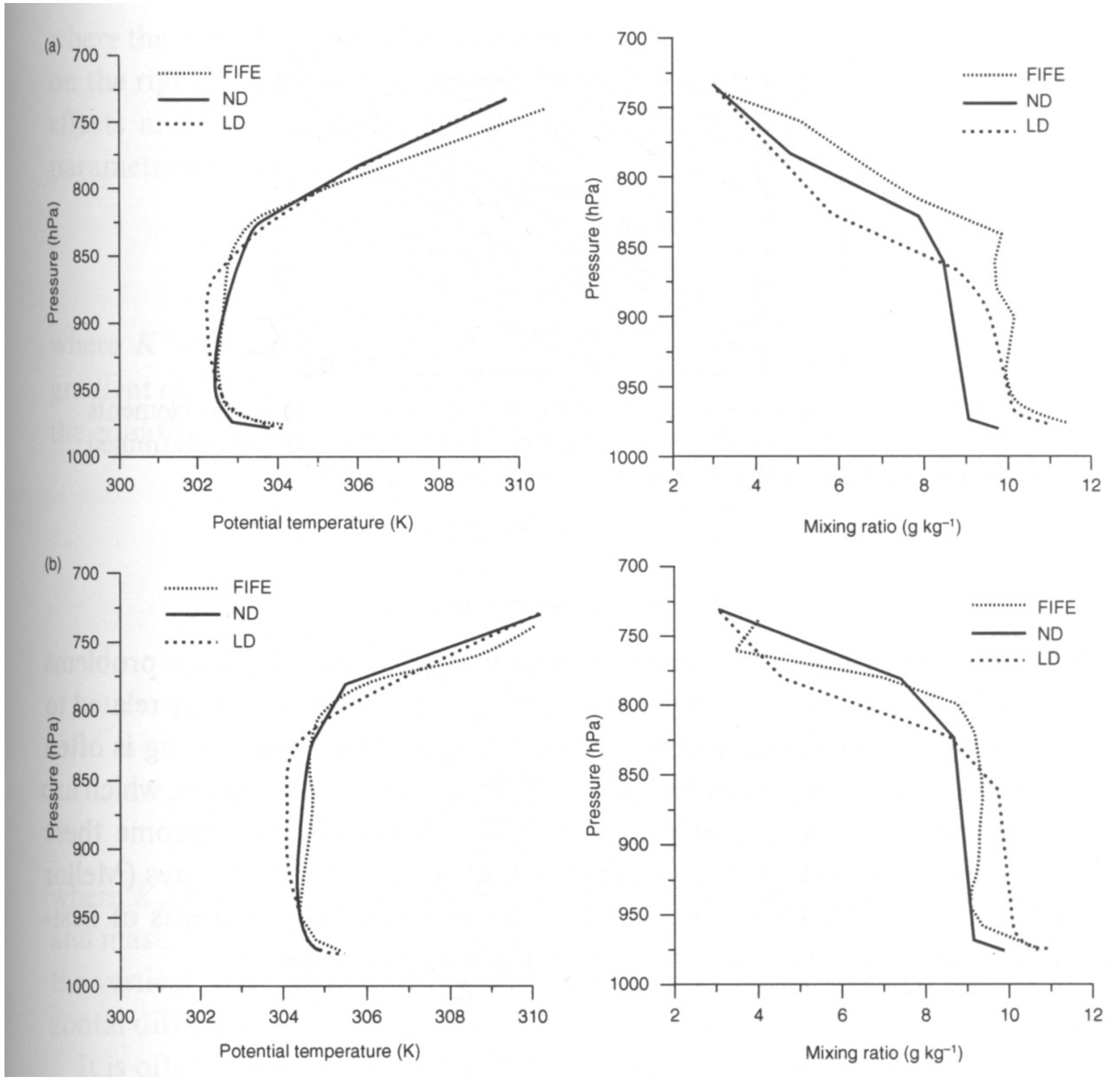
small eddies strong updrafts

iv) TKE (Turbulent Kinetic Energy) diffusion (Mellor and Yamada 1982)

TKE equation : $$\frac{\partial \overline{u_i u_j}}{\partial t} + u_j \frac{\partial \overline{u_i u_j}}{\partial x_j} = -\frac{\partial}{\partial x_k}[\overline{u_i u_j u_k} + \frac{1}{\rho} \cdots]$$

$$\overline{u_i u_j} \Longrightarrow k_z = \text{fn (TKE)}$$

3.4 Local versus nonlocal



Local scheme typically produces unstable mixed layer

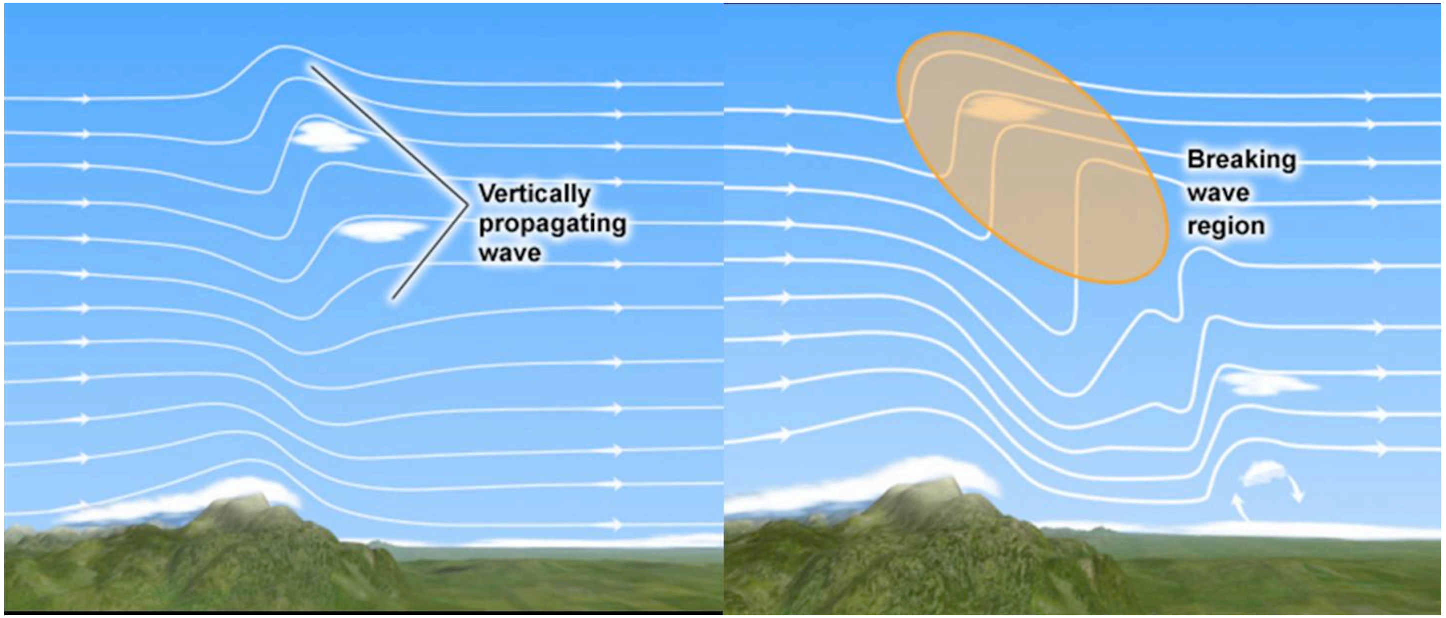
Hong and Pan (1996)

4. Gravity Wave Drag

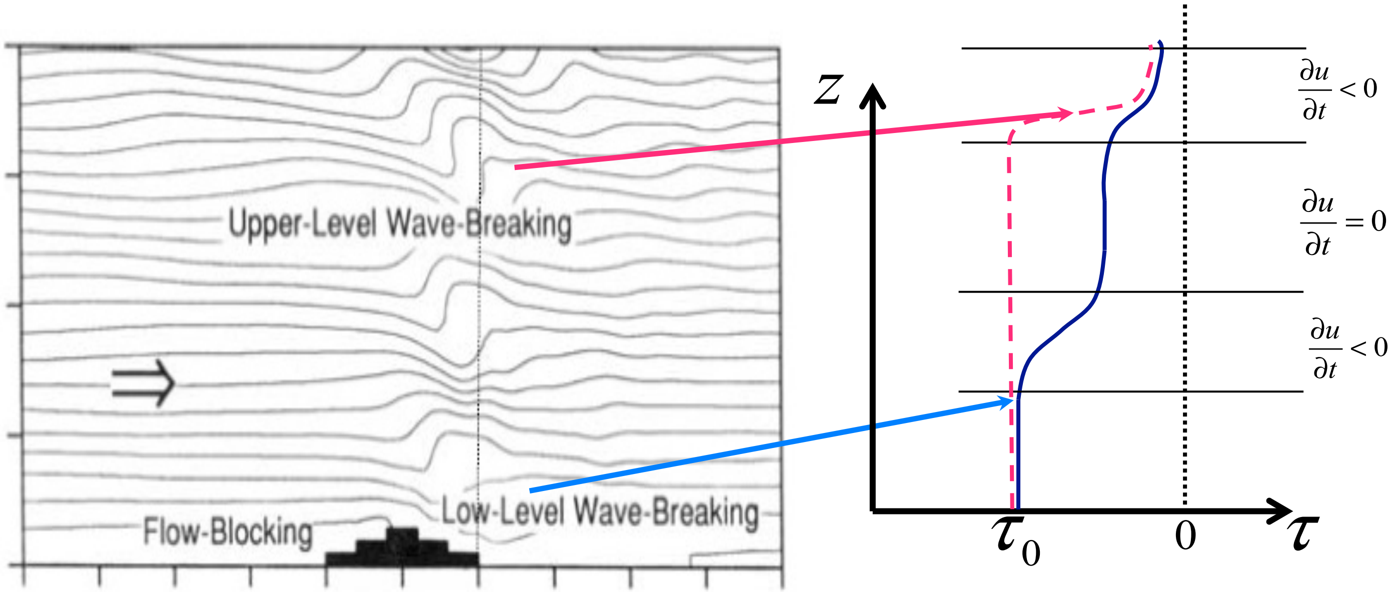
- GWDO : GWD induced by sub-grid scale orography
- GWDC : GWD induced by precipitating deep convection

4.1 Concept

This scheme (GWDO) includes the effect of mountain induced gravity wave drag from sub-grid scale orography including convective breaking, shear breaking and the presence of critical levels. Effects are strong in the presence of strong vertical wind shear and thermally stable layer.



4.2 Enhanced lower tropospheric gravity wave drag (Kim and Arakawa 1995)



$$\frac{\partial u}{\partial t} = -\frac{1}{\rho}\frac{\partial \tau_x}{\partial z}$$

$$\frac{\partial v}{\partial t} = -\frac{1}{\rho}\frac{\partial \tau_y}{\partial z}$$

Stress at reference level : $\tau_0 = -E\frac{1}{\Delta x}\frac{\rho_0 U_0^3}{N_0}\frac{Fr^2}{Fr^2 + 0.5/OC}, \quad U_0 = \frac{1}{h}\int_{k=1}^{k=kpbl} U dz$

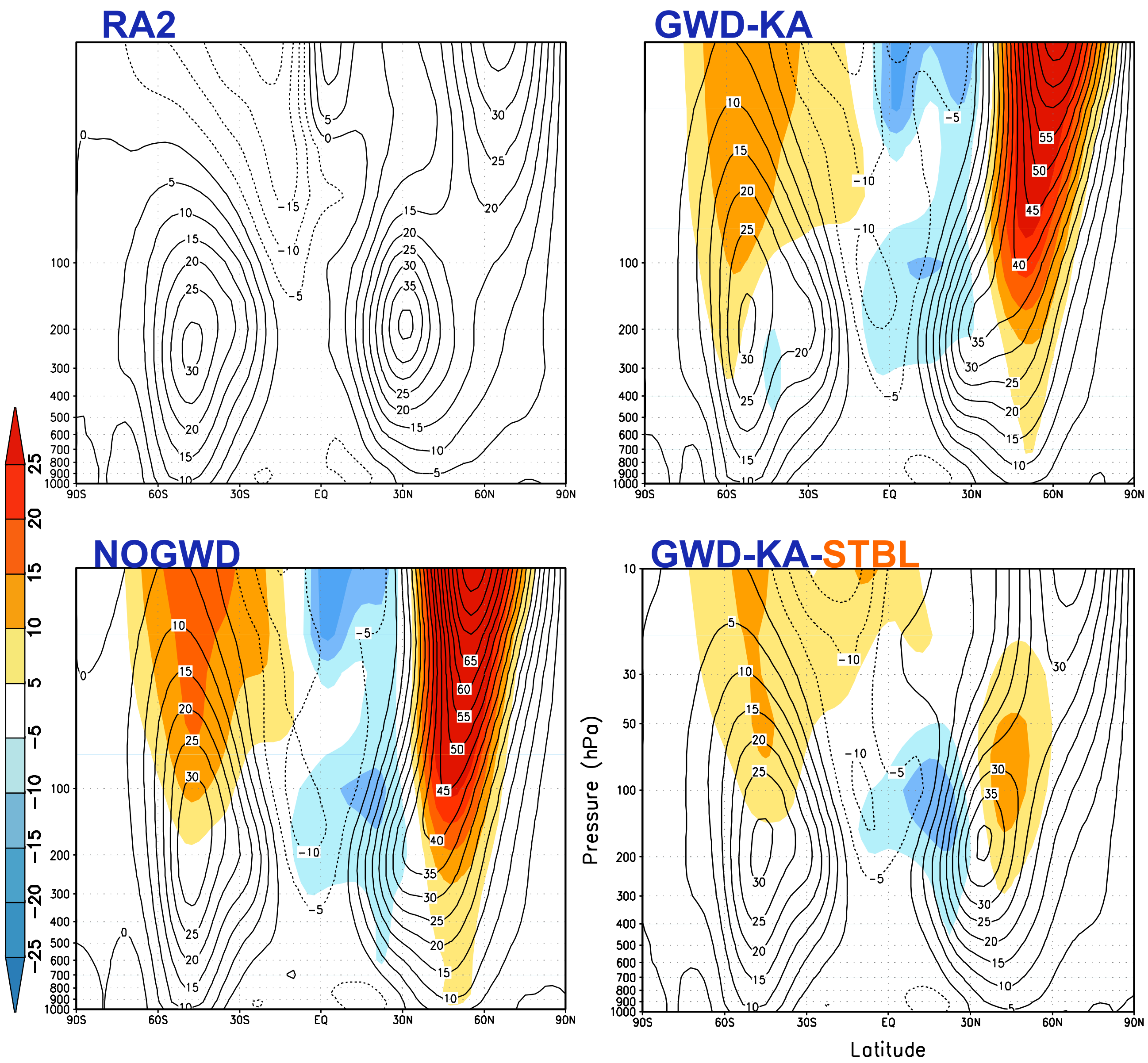
Reference level (KA95) : Max (2, KPBL)

Conventional : the conventional Ri number-based wave- breaking mechanism using the saturation hypothesis, which works mainly in the upper atmosphere

Advanced: the new orographic statistics-based wave- breaking mechanism using half-theory (Scorer parameter ~ BVF**2 / U**2) and half-empiricism obtained from mesoscale mountain wave simulations, which works mainly in the lower atmosphere. **Flow blocking is also introduced recently.**

4.3 Impact of GWDO

Zonal-averaged zonal wind (96/97 DJF)

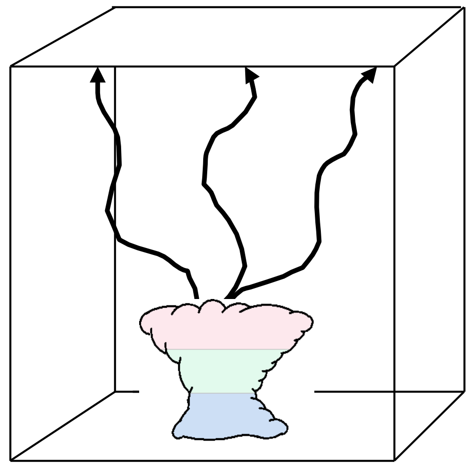


Contour : Zonal averaged zonal wind
Shaded: Deviations from the RA2

Kim and Arakawa
→Improves upper level jets
→Improves the sea level pressure

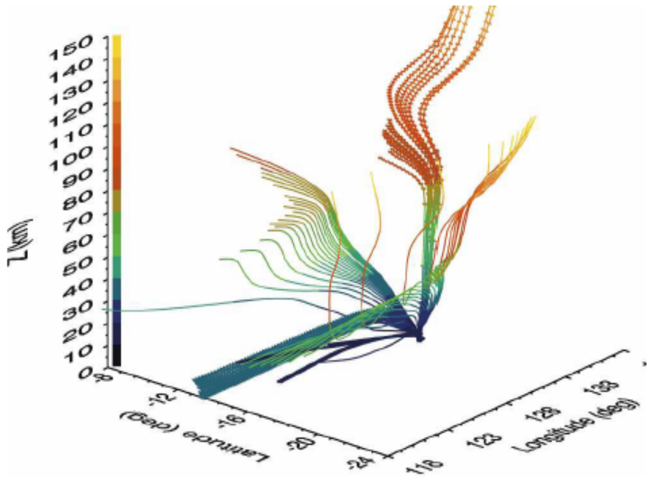
(Kim and Hong, 2009)

4.4 Convective GWD (CWDC)



Columnar CGWD parameterizations

- **Chun and Baik (1998, 2002):** The momentum flux spectrum for the CGWD parameterization was first analytically formulated
- **Chun et al. (2008):** A nonlinear source effect was included in the CGWD parameterization of Song and Chun (2005) that had taken account of a diabatic source alone



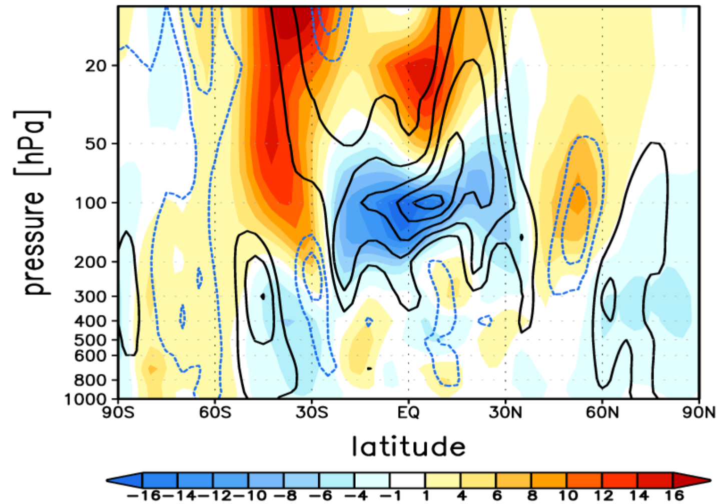
Ray-based CGWD parameterization

- **Song and Chun (2008):** GW propagation properties were explicitly calculated and a three-dimensional propagation of GWs was realistically represented
- **Choi and Chun (2011):** Two free parameters, the moving speed of the convective source and wave-propagation direction, were determined

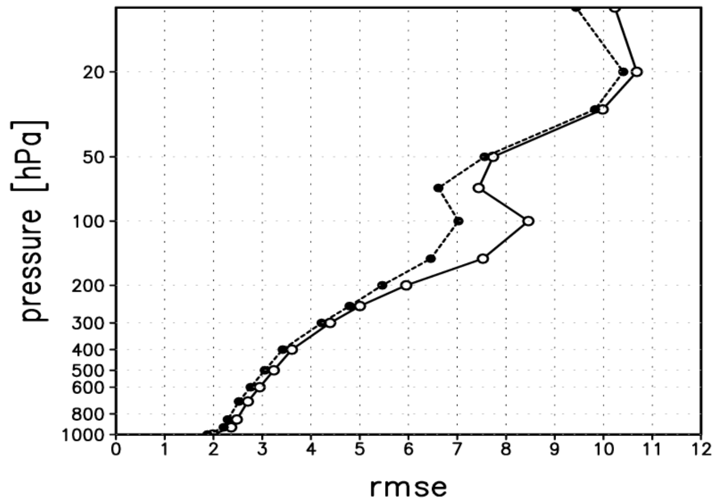
Prof. Chun, Hye-Young, chunhy@yonsei.ac.kr

4.5 Improvement by GWDC (Jeon et al. 2010, APJAS)

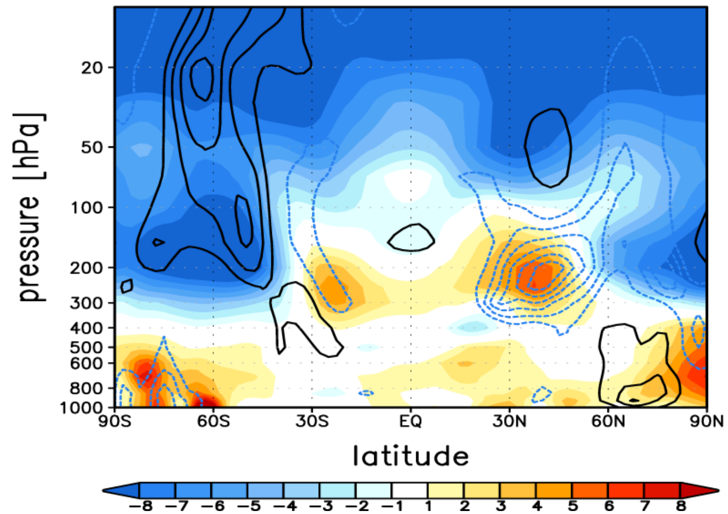
a) SAS_CB98 Zonal wind difference



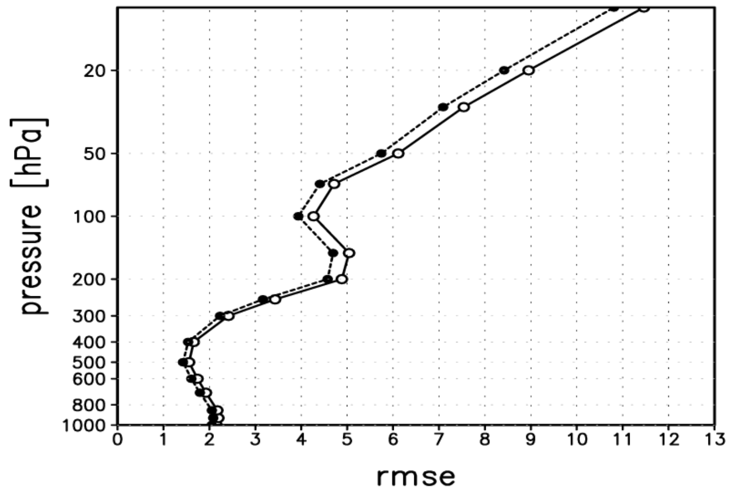
c)



a) SAS_CB98 Temperature difference



b)



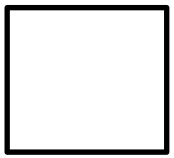
Precipitation Processes

Concept : precipitation algorithms (CPS and MPS)

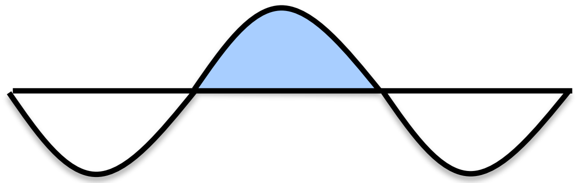
- In real atmosphere, dynamical motion -> RH > 1 => clouds form -> produces rain
- In modeled atmosphere, RH < 1

  But generate clouds by releasing CAPE→ requires parameterized process

  Deep convection : 1~10km



$\overline{RH} < 1$



$RH = 1$

$\Delta x \Longrightarrow 0$, more grid-resolvable precipitation

Thus, we need the cumulus parameterization scheme to account for releasing conditional instability due to subgrid scale motion

- Grid-resolvable (MPS): Supersaturation → clouds
- Subgrid scale (CPS) : CAPE removal → clouds

5. Deep Convection

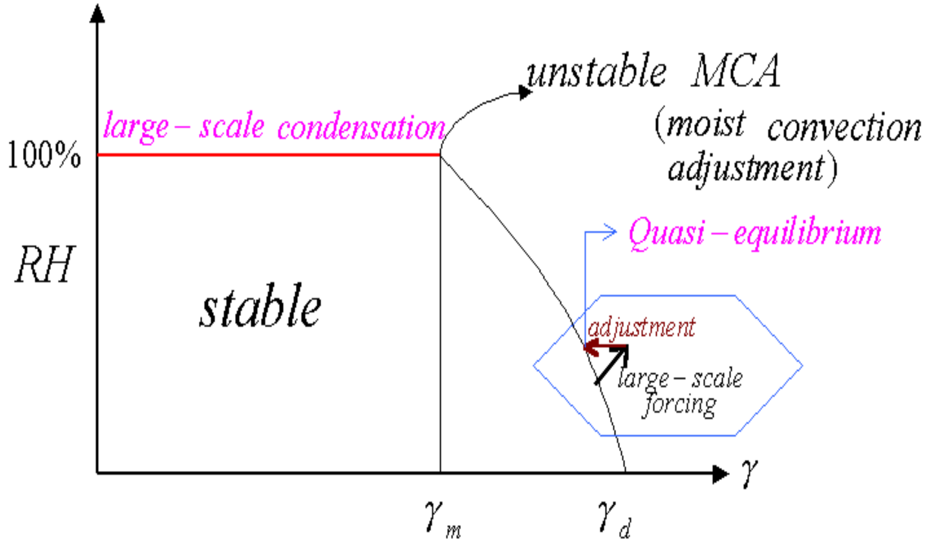
Parameterized convection
Cumulus convection
Subgridscale precipitation
Implicit precipitation

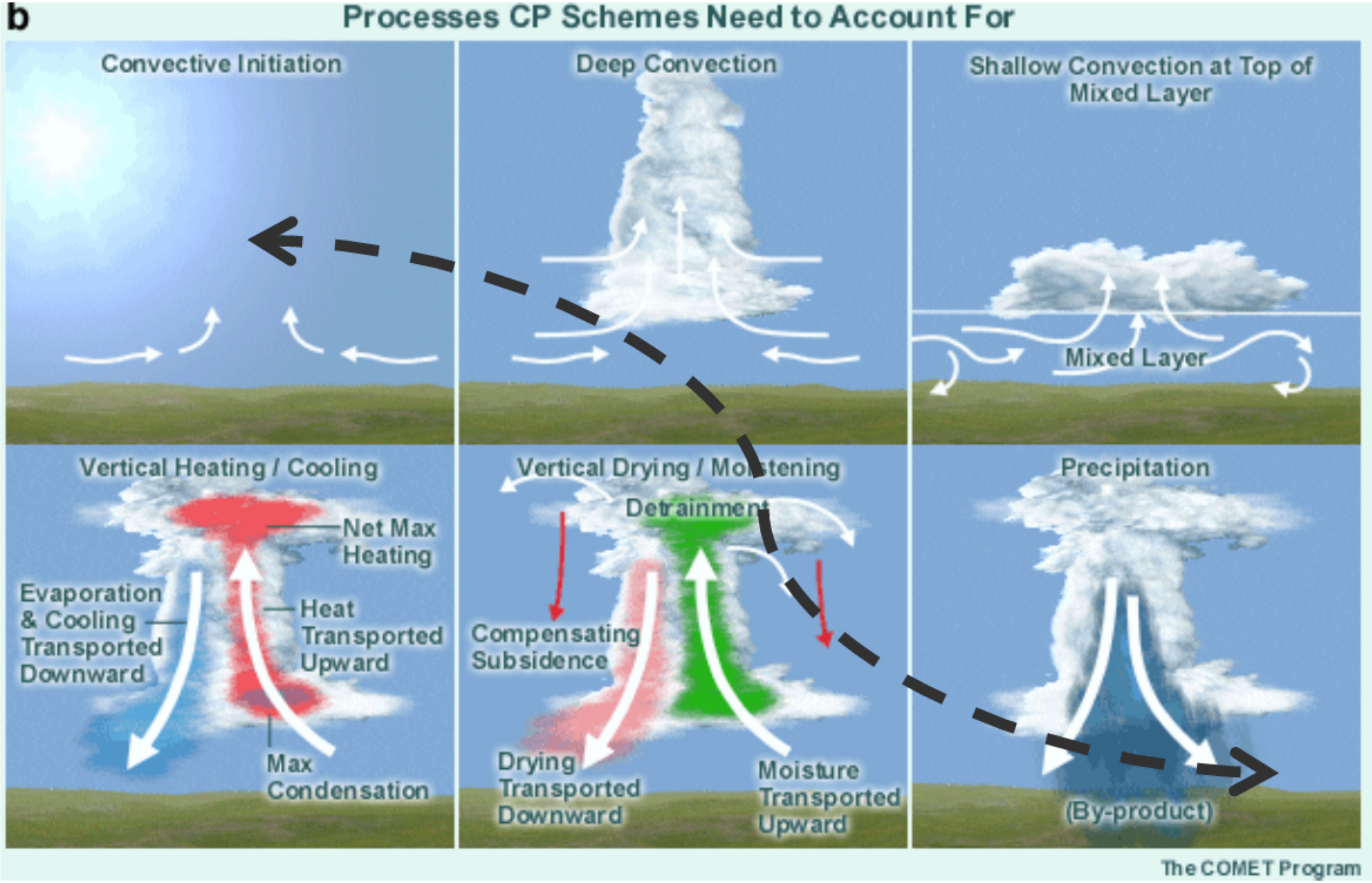
Jul 25, 2010, Seoul

5.1 Concept

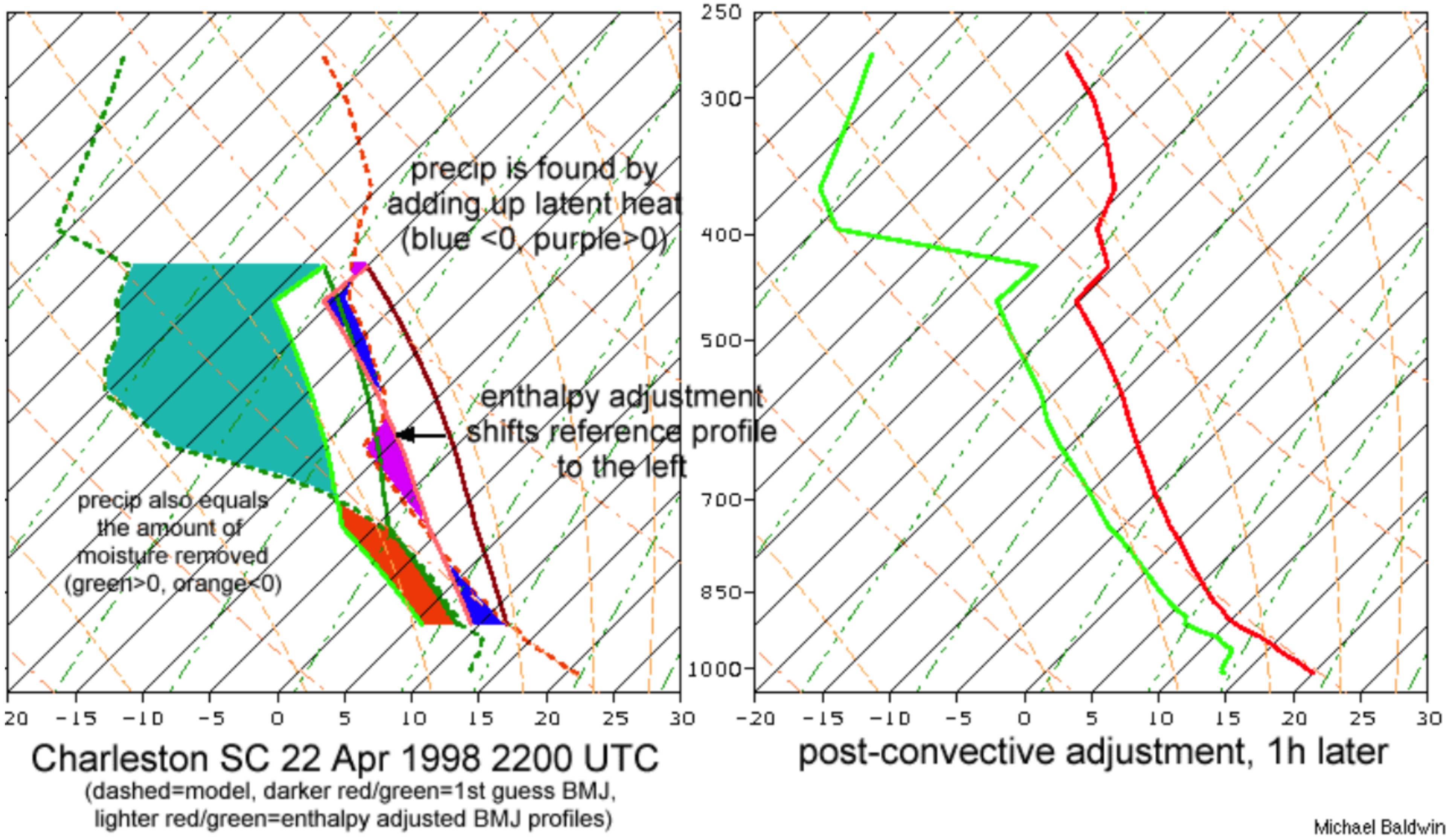
- represents deep precipitating convection and feedback to large-scale
- must formulate the collective effects of subgridscale clouds in terms of the prognostic variable of grid scale

- **Closure assumption**





CPS does not consider the detailed evolution of convection !



CPS consider changes in profile before and after convection

5.2 Kuo scheme (1965)

: Cloud is formed in proportional to column-integrated moisture convergence

$$M_t = -\frac{1}{g}\int_0^{P_s} \nabla \cdot (vq)dp + F_{g_s}$$

$$\int_0^{P_s} \frac{\partial q}{\partial t}dp = gbM_t\,,\quad \text{b : moistening factor}$$

$$\int_0^{P_s} \theta_c dp = gL(1-b)M_t \qquad \frac{\theta_c}{\pi} = \frac{\theta_a - \theta}{\tau}$$

- Modified Kuo scheme

Krishnamurti el al. (1980, 1983) : proportional to

vertical advection of moisture

$$M_t = -\frac{1}{g}\left\{\int_0^{P_s} \boldsymbol{\omega}\ \frac{\partial q}{\partial p}dp + F_{q_s}\right\}$$

Anthes (AK : 1977) : a revised moistening factor and parcel buoyancy

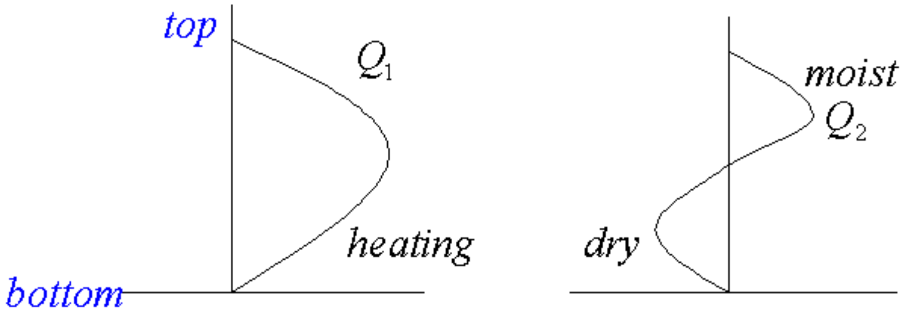
$$b = \left(\frac{1-RH}{1-RH_c}\right)^n$$

- **Heating and moistening profiles**

$$\frac{d\theta}{dt} = \frac{1}{\pi}\left[gL(1-b)M_t Q_1 + Q_r \right]$$

$$\frac{dq}{dt} = -g(1-b)M_t Q_2$$

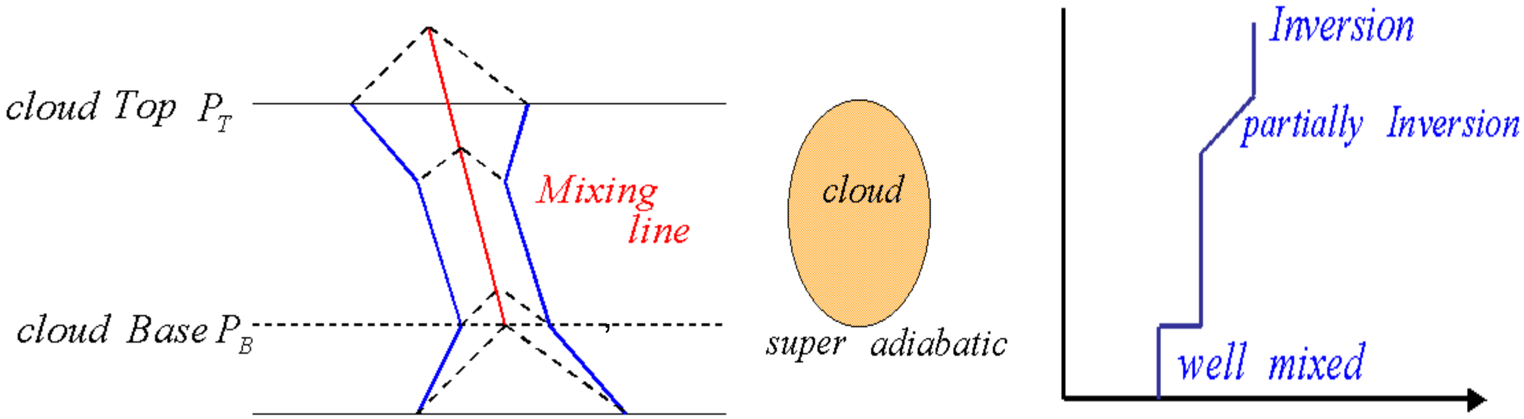
$$\int_0^{P_s} Q_1 dp = \int_0^{P_s} Q_2 dp = 1$$



Remarks : Kuo scheme produces excessive point maximum precipitation. Also, warm bias in the lower troposphere. This scheme take into account of large-scale mixing as well as turbulent mixing

5.3 Betts-Miller scheme (1986)

: adjust toward reference profiles that are based on observational evidence of convective equilibrium



Reference profile to be adjusted : originally based on tropical cyclones

$$\theta_R'(P) = \bar{\theta}(P_B) + \beta\ M_\theta (P - P_B)$$

$$\frac{\partial q}{\partial p} = \beta \left(\frac{\partial q}{\partial p_*} \right)_M \quad \beta = \frac{\partial p^*}{\partial p}$$ p^* : saturation pressure (= 1.2 for example)

$$M_\theta = 0.85 \left(\frac{\partial \theta^*}{\partial P^*} \right)_M$$ M : Mixing line

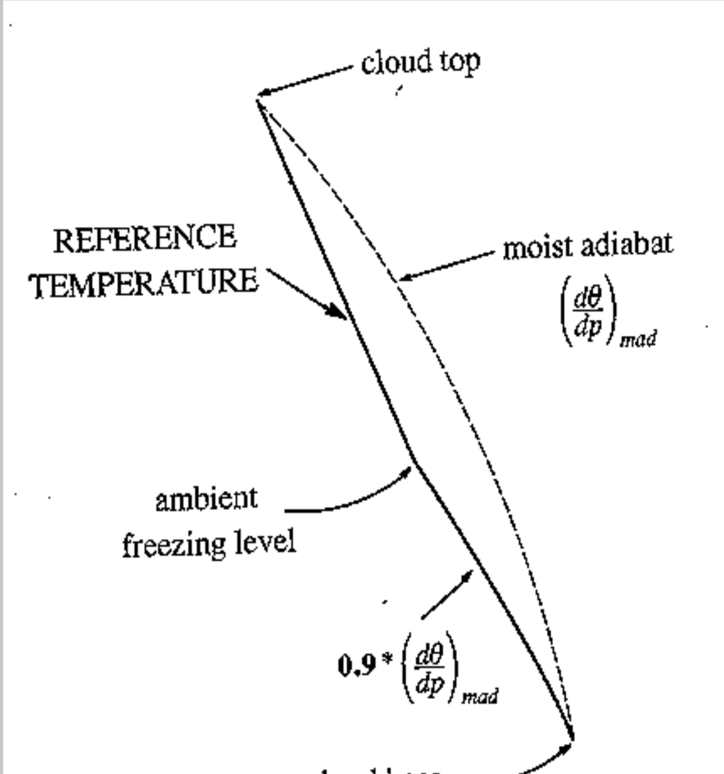
Energy Constraints : $$\int_{P_B}^{P_{T+1}} C_P \left(T_R - \bar{T}\right) dp = \int_{P_B}^{P_{T+1}} \left(q_R - \bar{q}\right) dp = 0$$

- Convective tendencies & Precipitation

$$\left(\frac{\partial \overline{T}}{\partial t}\right)_{Cu} = \frac{T_R - \overline{T}}{\tau}$$

$$\left(\frac{\partial \overline{q}}{\partial t}\right)_{Cu} = \frac{q_R - \overline{q}}{\tau}$$

$$\Pr ecip = \int_{P_0}^{P_T} \left(\frac{q_R - \overline{q}}{\tau}\right)\frac{dP}{g} = -\frac{C_P}{L}\int_{P_0}^{P_T}\left(\frac{T_R - \overline{T}}{\tau}\right)\frac{dp}{g}$$



Betts (1986)

- Adjustment scheme

: -Manabe (hard adjustment : toward a moist adiabat)
-Kuo, BM (soft adjustment : toward reference profiles

- Remarks :

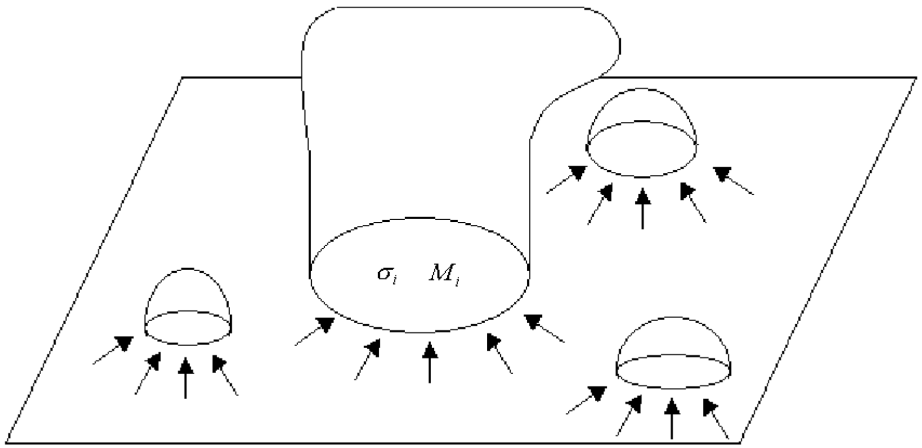
- Uncertainty in reference profiles → 2×CO2 → similar climate ???

5.4 Mass-flux schemes : Arakawa-Schubert (1974)

i) Concept

-- Mass flux approach, cloud ensemble, quasi-equilibrium
-- Theoretical frame work for CPS

- Area is large enough so that cloud ensemble can be a statistical entity
- Area is small enough so that cloud environment is approximately uniform horizontally



M_i : vertical mass flux through ith cloud

σ_i : fractional area covered by ith cloud

$M_c \equiv \sum_i M_i$: total vertical mass flux

environment

$$\rho M = M_c + \tilde{M}$$

: net mass flux/unit large-scale horizontal area

ii) Quasi-equilibrium : cloud forcing ~ large-scale adjustment

: CPS computes the warming (cooling) in the grid box due to adiabatic descent (ascent), rather than computing latent heat releaser in cloud models

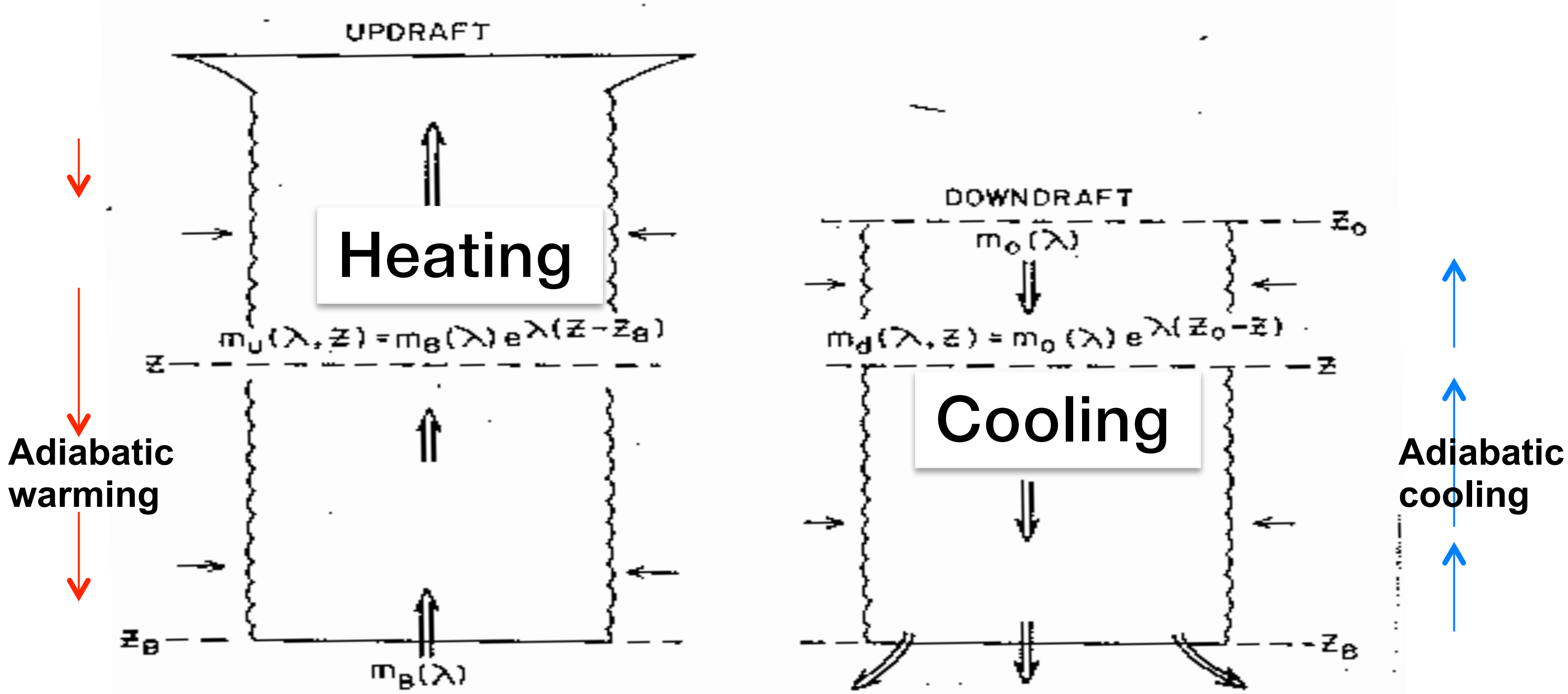


FIG. 4.10. Model for updraft and downdraft of cloud type λ (from Johnson 1976).

iii) Energy budget equations

Large-scale flux across grid box　　　　Exchange of S between environment and clouds

$$\frac{\partial}{\partial t}\rho\left(1-\sigma_c\right)\tilde{s} = -\vec{\nabla}\cdot\overline{\left(\rho\tilde{V}S\right)} - \frac{\partial}{\partial Z}\left(\tilde{M}\tilde{S}\right) - \sum_i\left(\frac{\partial M_i}{\partial Z}+\rho\frac{\partial\sigma_i}{\partial t}\right)S_{i_b} - LE + \tilde{Q}_R$$

$$\frac{\partial}{\partial t}\rho\sum\sigma_i s_i = -\frac{\partial}{\partial z}\left(\sum_i M_i s_i\right) + \sum_i\left(\frac{\partial M_i}{\partial z}+\rho\frac{\partial\sigma_i}{\partial t}\right)S_{ib} + \sum_x\left(LC_i + Q_{Ri}\right)$$

$S_i : C_pT + gz$ (dry static energy) of i^th cloud

$S_{ib} : C_PT + gz$ of the air entraining into or detraining from the i^th cloud

C_i : condensation in the i^th cloud

E : evaporation of liquid water in the environment

Q_r : Radiative heating

- Entrainment : $\dfrac{\partial M_i}{\partial z} + \rho\dfrac{\partial \sigma_i}{\partial t} > 0,\quad S_{ib} = \tilde{S}$

- Detrainment : $\dfrac{\partial M_i}{\partial z} + \rho\dfrac{\partial \sigma_i}{\partial t} < 0,\quad S_{ib} = S_i$

iv) Approximation

- Assume $\sigma_c \ll 1$, $\bar{s} \simeq \tilde{s}$

$$\frac{\partial}{\partial t}\rho\bar{s} = -\nabla\cdot(\rho\bar{v}\bar{s}) - \frac{\partial}{\partial z}(\rho\bar{w}\bar{s}) - \overline{\nabla\cdot(\rho\bar{v}s - \rho\bar{v}\bar{s})}$$

$$+\underline{M_c\frac{\partial\bar{s}}{\partial z}} - \sum_{dc}(\frac{\partial M_i}{\partial z} + \rho\frac{\partial\sigma_i}{\partial t})\underline{(\delta_i - \bar{s})} - LE + \theta_R$$

detraininent, entrainment

Detraining clouds

Adiabatic warming due to hypothetical subsidence between the clouds

$$\frac{\partial}{\partial t}\rho\bar{q} = -\nabla\cdot(\rho\bar{v}\bar{q}) - \frac{\partial}{\partial z}(\rho\bar{w}\bar{q}) - \overline{\nabla\cdot(\rho\bar{v}q - \rho\bar{v}\bar{q})}$$

$$+\underline{M_c\frac{\partial\bar{q}}{\partial z}} - \sum_{dc}(\frac{\partial M_i}{\partial z} + \rho\frac{\partial\sigma_i}{\partial t})(q_i - \bar{q}) - E$$

- Spectral cloud ensemble :

$$M_c(z) = \int_0^{\lambda_{max}} \underline{m(z,\lambda)d\lambda}$$

Sub-ensemble

mass flux of between λ and $d\lambda+\lambda$

$$= \int_0^{\lambda_{max}} \underline{m_B(\lambda)}\eta(z,\lambda)d\lambda$$

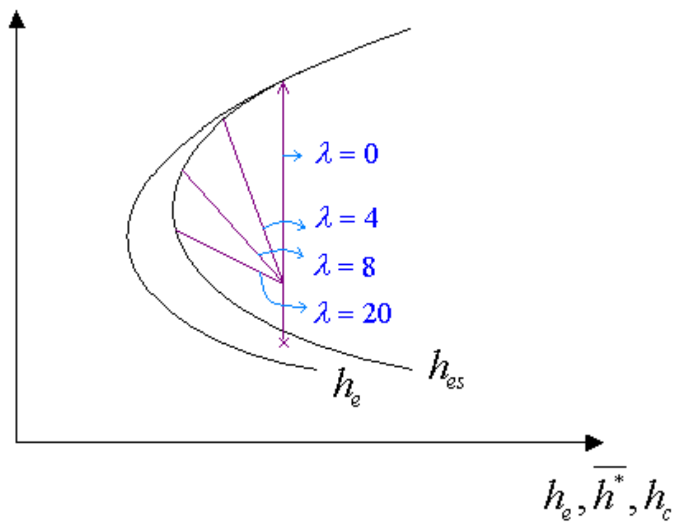
Mass flux at cloud base

$$\eta(z,\lambda) \equiv \frac{m(z,\lambda)}{m_B(\lambda)}$$; normalized subensemble mass flux

v) Closure

$$\frac{\partial m(z,\lambda)}{\partial z} = \mu(z,\lambda)\eta(z,\lambda)$$

$$\eta(z,\lambda) = e^{\lambda(z-zB)} \quad ; \text{ mass flux profile}$$



- Cloud work function

$$A(\lambda) = \int_{z_B}^{z_D(\lambda)} \eta(z,\lambda) g \frac{T_c(z,\lambda) - \overline{T}(z)}{\overline{T}} dz$$

- Q-G equilibrium

$$\frac{dA(\lambda)}{dt} = \left.\frac{dA(\lambda)}{dt}\right|_{LS} + \left.\frac{dA(\lambda)}{dt}\right|_{C} \;;\; 0$$

Large-scale forcing
>0 : destabilized

Adjustment
<0 : stabilization

Kernel : Cloud scheme kinetic energy

$$K_{ij} = \frac{A_i' - A_i}{(m_B \Delta t)} \qquad \sum_{j}^{i_{\max}} K_{ij}(m_B \Delta t)_j + F_i = 0 \quad \Rightarrow m_B$$

$$\longrightarrow \text{compute } \frac{\partial \bar{s}}{\partial t}, \frac{\partial \bar{q}}{\partial t} \text{ with } \eta, m_B$$

5.5 Other schemes

* AS type mass flux scheme

Grell scheme (1993) : removes lateral mixing to find the deepest cloud

Simplified AS (SAS, Han and Pan 2011): revised cloud physics from the Grell

Relaxed AS (RAS, Moorthi and Suarez 1992): linearized profile function

* Other mass flux schemes : Low-level control convective schemes (Stensrud 2007)

Kain and Fritsch (1993) : CAPE based sophisticated convective plume model

Emanuel (1991) : Stochastic mixing cloud model

Tiedtke (1989) : Large-scale moisture convergence (KUO) based mass flux

Gregory-Rowntree (1990): Parcel buoyancy based turbulence in cloud model

6. Shallow Convection

6.1 Concept

more vigorous vertical mixing of q and T above the mixed layer top. With the enhanced vertical eddy transport between LCL and inversion level, this process does not allow the excess moisture trapped near the surface in synoptically inactive regions (non-precipitating convection).

➔ Cooling and moistening above LCL and heating and warming below.

6.2 Classification

- Moist adjustment type : Betts and Miller (1993), Lock et al. (2000), Tiedtke (1983)
- Mass flux type : Kain (2004), Park and Bretherton (2009), Han and Pan (2011)

Tiedtke (1983)

$$\frac{\partial T}{\partial t} = \frac{1}{\rho}\frac{\partial}{\partial z}\left(\rho K\left[\frac{\partial T}{\partial z} + \Gamma\right]\right)$$

$$\frac{\partial q}{\partial t} = \frac{1}{\rho}\frac{\partial}{\partial z}\left(\rho K\frac{\partial q}{\partial z}\right)$$

Han and Pan (2011)

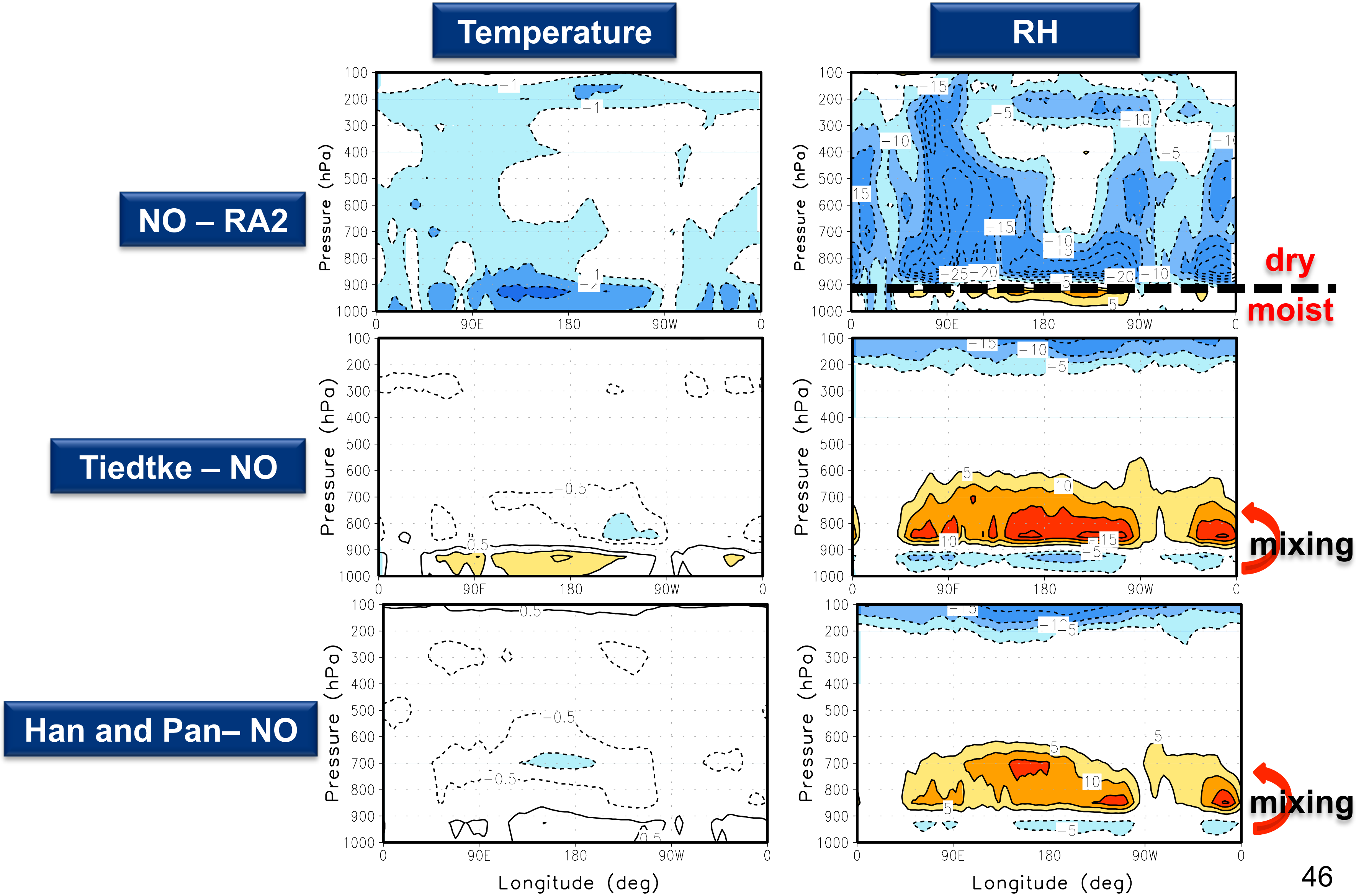
$$\frac{1}{\eta}\frac{\partial \eta}{\partial z} = \varepsilon - \delta$$

$$\frac{\partial(\eta s)}{\partial z} = (\varepsilon\bar{s} - \delta s)\eta$$

$$\frac{\partial[\eta(q_v + q_l)]}{\partial z} = \eta[\varepsilon\bar{q}_v - \delta(q_v + q_l) - r]$$

6.3 Impact of the shallow convection scheme

JJA 1996 simulation in a GCM



7. Non-convective Precipitation

large-scale precipitation
grid-resolvable scale precipitation
explicit moisture scheme
cloud scheme
microphysics scheme

7.1 Concept

- Remove supersaturation after deep and shallow convection, and feedback to large-scale

7.2 Classification : according to the complexity in microphysics

i) Diagnostic : condensation, evaporation of falling precipitation

ii) Bulk microphysics :

hydrometeors with size distribution in inverse-exponential function
- Single moment : predict mixing ratios of hydrometeors
- Double moment : + number concentrations
- Triple moment : + reflectivity

iii) Bin microphysics : divides the particle distribution into a number of finite size or mass categories.

7.3 Precipitate size distributions

Marshall and Palmer(1948) : exponential law
Heymsfield and Platt (1984) : Power law

$$N_R(D_R) = aD_R^{\ b}$$

The rain and snow particles are assumed to follow the size distribution derived by Marshall and Palmer(1948), and G unn and Marshall(1958), respectively. The size distribution s for both rain and snow are formulated according to an in verse-exponential distribution and its formula for rain can be expressed by

$$N_R(D_R) = N_{0R}\exp(-\lambda_R D_R)$$

for rain, where N_{0R} is the intercept parameter of the rain distributions.

The slope parameter of the size distributions for rain λ_R is determined by multiplying (A1) by drop mass (A4) and integrating over all diameters and equating the resulting quantities to the appropriate water contents (= ρq_R). This may be written as,

$$\lambda_R = \left(\frac{\pi\rho_w N_{0R}}{\rho q_R}\right)^{1/4}$$

$$\Gamma(x) = \int_0^\infty t^{x-1}\exp(-t)dt$$

$$\int_0^\infty D_R^{\ 4-1}\exp(-\lambda_R D_R)dD_R$$

$$= \Gamma(4)/\lambda_R^{\ 4}$$

7.4 Bulk Method : 1-Moment versus 2-Moment

Mixing Ratio

(1-moment/ 2-moment scheme)

$$\left(\int \frac{dM(D_R)}{dt} dN_{DR}\right) / \rho = \frac{dq}{dt} (kgkg^{-1}s^{-1})$$

Number concentration

(2-moment scheme)

$$\left(\int \frac{d\,\mathrm{Pr}\,ob(D_R)}{dt} dN_{DR}\right) = \frac{dN}{dt} (m^{-3}s^{-1})$$

$\log_{10}(N(D))$

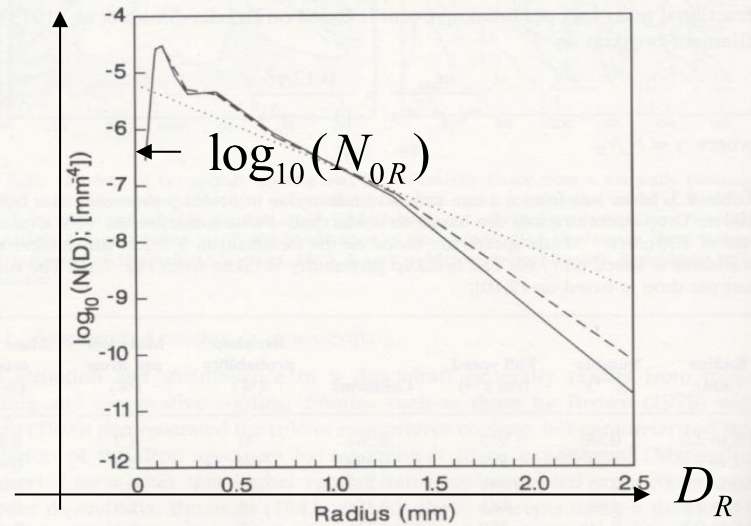


FIG. 9.21. Evolution of a drop spectrum in a subsaturated rainshaft including the effects of coalescence and breakup. The spectrum at the top of the rainshaft is Marshall–Palmer with a rainfall rate of 110 mm h^{-1}. (······), 0 km; (- - -), 1 km; (–), 2 km; R = 110 mm h^{-1}; t = 24 min. From Tzivion et al. (1989). *Journal of Atmospheric Sciences*, **46**, 21. American Meteorological Society. Reproduced with permission

Single moment scheme

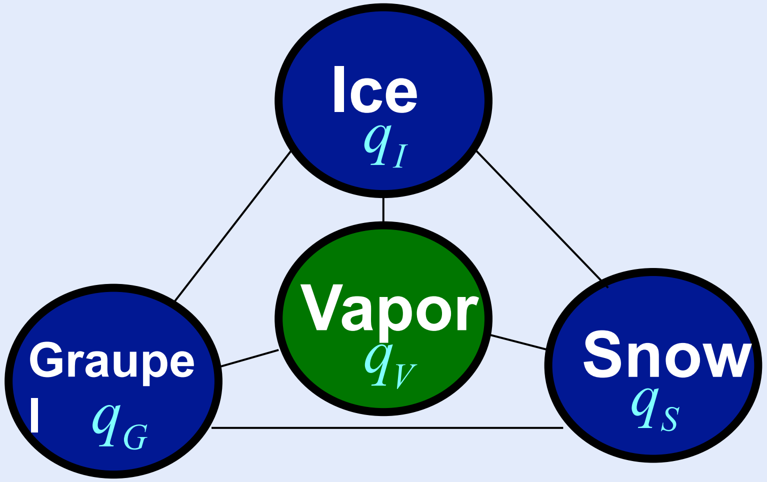
$$dN_{DR} = N_{0R} \exp(-\lambda_R D_R) dD_R$$

Double moment scheme

$$dN_{DR} = N_R \lambda_R^2 (N_R) D_R \exp(-\lambda_R D_R) dD_R$$

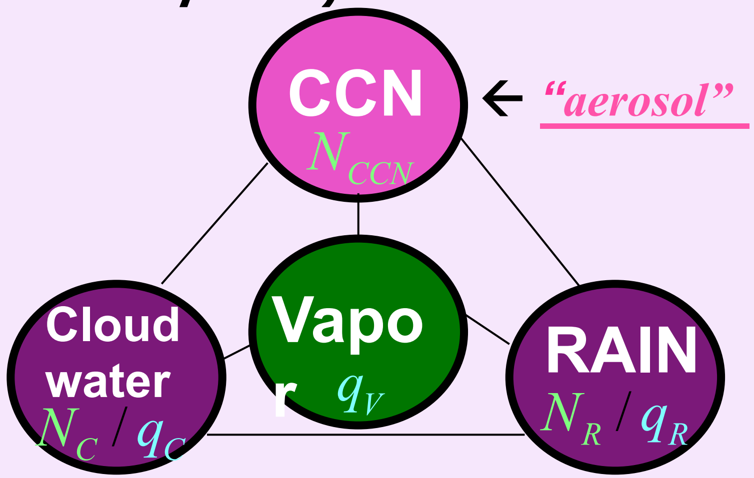
7.5 Bulk Method : 1-Moment (WSM) versus 2-Moment (WDM)

Cold rain processes :
(Hong et al. 2004; Hong and Lim 2006)



q for 4 hydrometeors will be predicted (Single Moment)

Warm rain processes :
(Khairoutdinov and Kogan 2000; Cohardt and Pinty 2000)



N,q for 2 hydrometeors will be predicted (Double Moment)

N: Cloud water, Rain, CCN

q: Cloud water, Rain, Ice, Snow, Graupel, Vapor

(Lim and Hong, 2010)

7.6 WDM6 code

**** Warm rain processes** (Hong and Lim 2010)

```
!  warm rain processes
!
!  - follows the double-moment processes in Lim and Hong
!
!=============================================================
!
      do k = kts, kte
        do i = its, ite
          supsat = max(q(i,k),qmin)-qs(i,k,1)
          satdt = supsat/dtcld
!-------------------------------------------------------------
! praut: auto conversion rate from cloud to rain   [LH 9] [CP 17]
!        (QC->QR)
!-------------------------------------------------------------
          lencon  = 2.7e-2*den(i,k)*qci(i,k,1)*(1.e20/16.*rslopec2(i,k)  &
                   *rslopec2(i,k)-0.4)
          lenconcr = max(1.2*lencon, qcrmin)
          if(avedia(i,k,1).gt.di15) then
            taucon = 3.7/den(i,k)/qci(i,k,1)/(0.5e6*rslopec(i,k)-7.5)
            praut(i,k) = lencon/taucon
            praut(i,k) = min(max(praut(i,k),0.),qci(i,k,1)/dtcld)
!-------------------------------------------------------------
! nraut: auto conversion rate from cloud to rain [LH A6] [CP 18 & 19]
!        (NC->NR)
!-------------------------------------------------------------
            nraut(i,k) = 3.5e9*den(i,k)*praut(i,k)                       &
            if(qrs(i,k,1).gt.lenconcr)
            nraut(i,k) = ncr(i,k,3)/qrs(i,k,1)*praut(i,k)
            nraut(i,k) = min(nraut(i,k),ncr(i,k,2)/dtcld)
          endif
!-------------------------------------------------------------
! pracw: accretion of cloud water by rain     [LH 10] [CP 22 & 23]
!        (QC->QR)
! nracw: accretion of cloud water by rain     [LH A9]
!        (NC->)
!-------------------------------------------------------------
          if(qrs(i,k,1).ge.lenconcr) then
            if(avedia(i,k,2).ge.di100) then
              nracw(i,k) = min(ncrk1*ncr(i,k,2)*ncr(i,k,3)*(rslopec3(i,k)  &
                          + 24.*rslope3(i,k,1)),ncr(i,k,2)/dtcld)
              pracw(i,k) = min(pi/6.*(denr/den(i,k))*ncrk1*ncr(i,k,2)     &
                          *ncr(i,k,3)*rslopec3(i,k)*(2.*rslopec3(i,k)    &
                          + 24.*rslope3(i,k,1)),qci(i,k,1)/dtcld)
            else
              nracw(i,k) = min(ncrk2*ncr(i,k,2)*ncr(i,k,3)*(2.*rslopec3(i,k)  &
                          *rslopec3(i,k)+5040.*rslope3(i,k,1)              &
                          *rslope3(i,k,1)),ncr(i,k,2)/dtcld)
              pracw(i,k) = min(pi/6.*(denr/den(i,k))*ncrk2*ncr(i,k,2)     &
                          *ncr(i,k,3)*rslopec3(i,k)*(6.*rslopec3(i,k)    &
                          *rslopec3(i,k)+5040.*rslope3(i,k,1)*rslope3(i,k,1))  &
                          ,qci(i,k,1)/dtcld)
            endif
          endif
```

***Auto conversion from cloud to rain [C→ R]**

$$\text{Praut [kgkg}^{-1}\text{s}^{-1}] = L/\tau \qquad L = 2.7\times10^{-2}\rho_a q_C\left(\frac{10^{20}}{16\lambda_C^4} - 0.4\right)$$

$$\tau = 3.7\frac{1}{\rho_a q_C}\left(\frac{0.5\times10^6}{\lambda_C} - 7.5\right)^{-1}$$

$$\text{Nraut[m}^{-3}\text{s}^{-1}] = 3.5\times10^9\frac{\rho_a L}{\tau}$$

***Accretion of cloud water by rain [C→ R]**

$D_R \ge 100$ μm

$$\text{Pracw [kgkg}^{-1}\text{s}^{-1}] = \frac{\pi}{6}\frac{\rho_W}{\rho_a}K_1\frac{N_C N_R}{\lambda_C^3}\left\{\frac{2}{\lambda_C^3} + \frac{24}{\lambda_R^3}\right\}$$

$$\text{Nracw[m}^{-3}\text{s}^{-1}] = -K_1 N_C N_R\left\{\frac{1}{\lambda_C^3} + \frac{24}{\lambda_R^3}\right\}$$

$D_R < 100$ μm

$$\text{Pracw [kgkg}^{-1}\text{s}^{-1}] = \frac{\pi}{6}\frac{\rho_W}{\rho_a}K_2\frac{N_C N_R}{\lambda_C^3}\left\{\frac{6}{\lambda_C^6} + \frac{5040}{\lambda_R^6}\right\}$$

$$\text{Nracw[m}^{-3}\text{s}^{-1}] = -K_2 N_C N_R\left\{\frac{2}{\lambda_C^6} + \frac{5040}{\lambda_R^6}\right\}$$

Resolution Dependency

Cut-off horizontal grid length for parameterizations

- PBL : ~50 m (Mirocha, 2008 WRF workshop)
- GWDO : ~ 3 km (hydrostatic approximation)
- GWDC: ~ 3 km (go with CP)
- Cumulus parameterization : ~ 3 km (Shin and Hong 2009), **However, recall the past 20 years**

Cut-off horizontal grid length for Cumulus parameterization

- KMA regional prediction model has been operational without CP even at 80 km until late 1990
- With advances in CP and other physics and initial condition, the cut-off length becomes smaller and smaller
- CP is beneficial even at 4 km (JMA operational model)

Reference : Hong and Dudhia, 2012: Bridging parameterized convection, cloud-resolving, and large-eddies, Bull. Amer. Meteor. Soc., January issue.

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Thanks for your attention !

Modeling is to understand what is happening in nature !