Physics Algorithms : Overview

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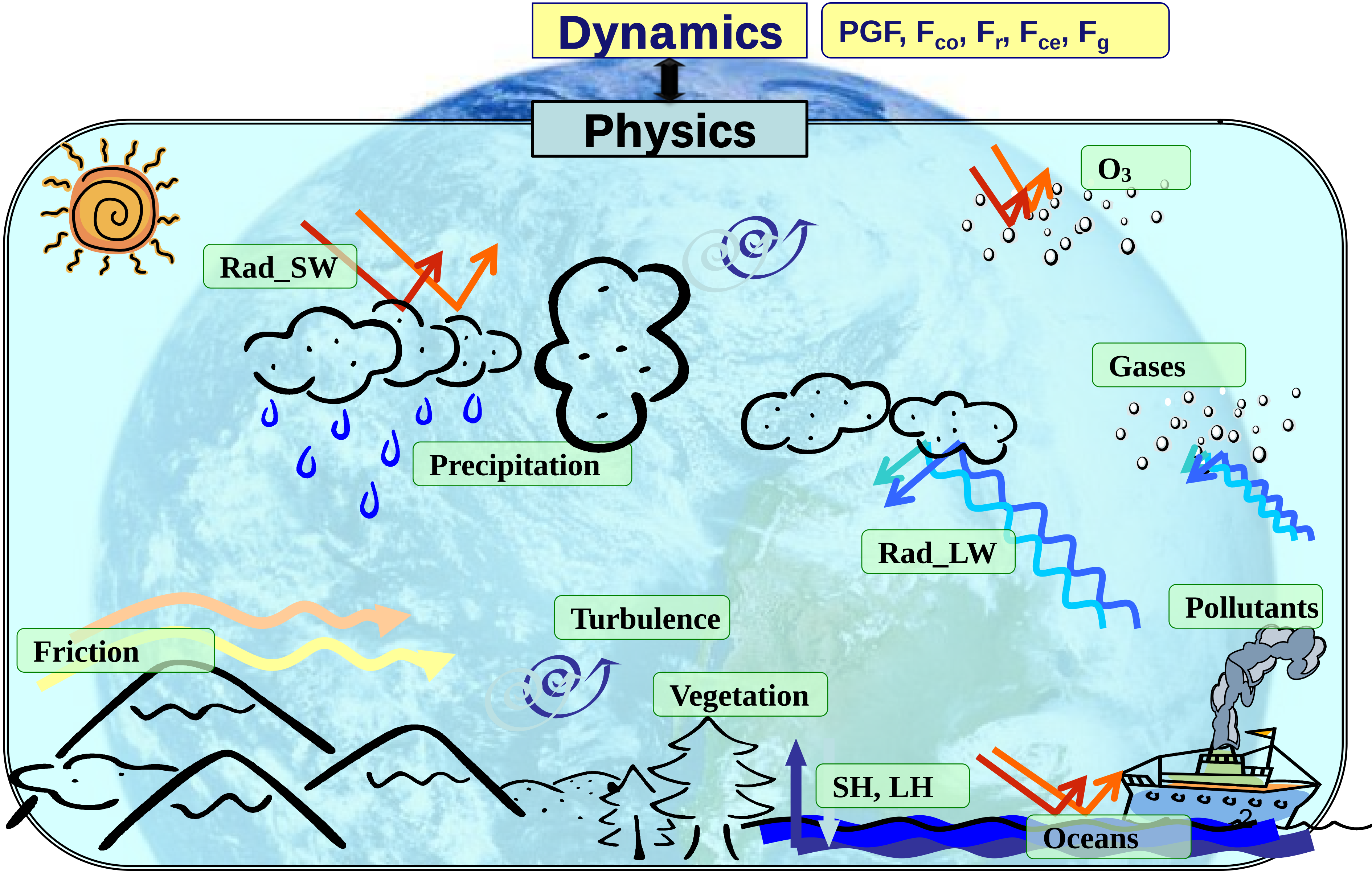
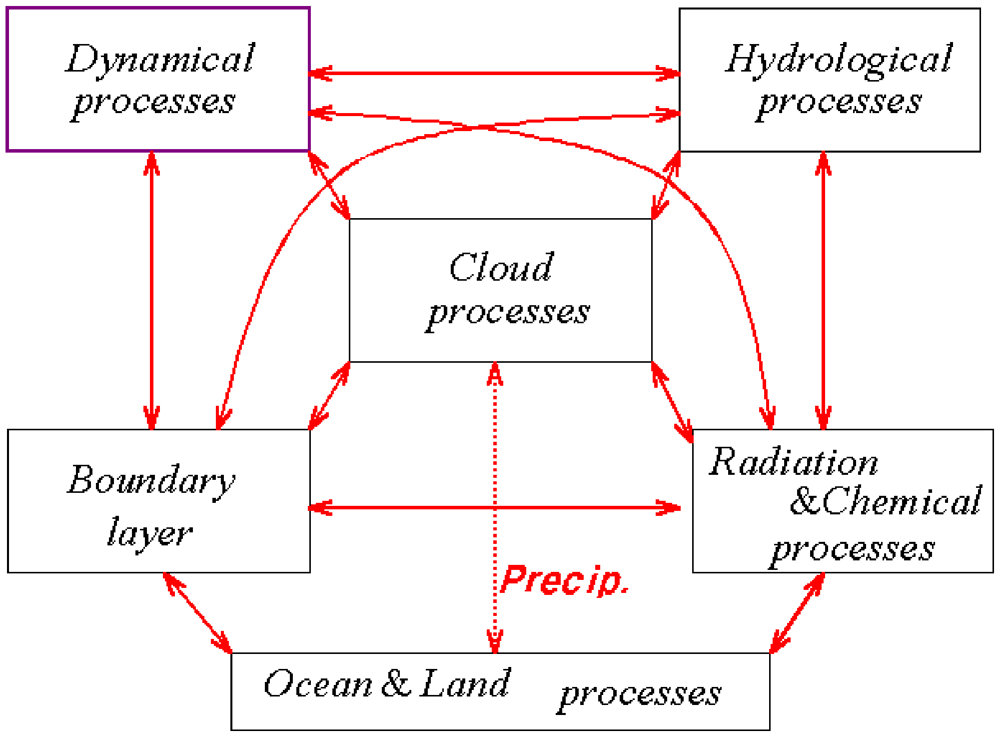
0. Introduction



Schematic configuration of physics



*** 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**

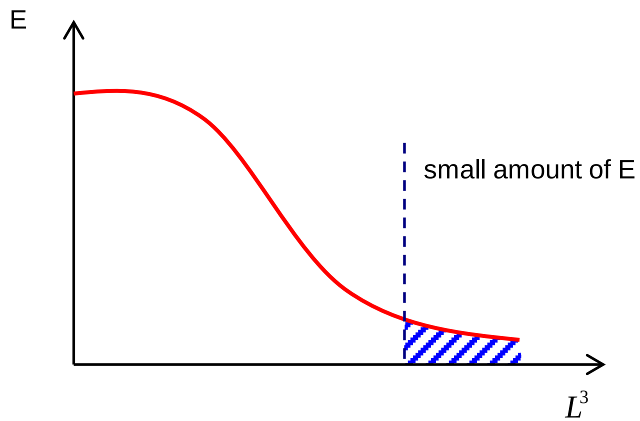
The formulation of physical process in terms of the model variables as parameters. (constants or functional relations)

1) Subgrid scale process

*** 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)

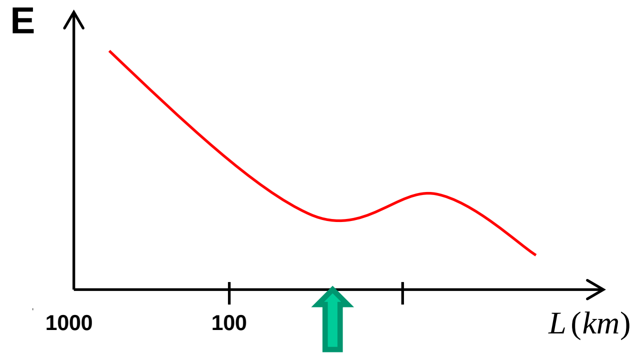


- $\Delta x \rightarrow 0$, the energy dissipation takes place by molecular viscosity (smallest grid size $\gg$ idealized situation)

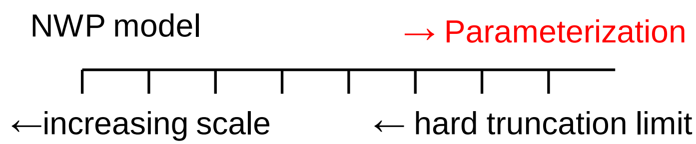
• Objective of subgrid scale parameterization

"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."

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

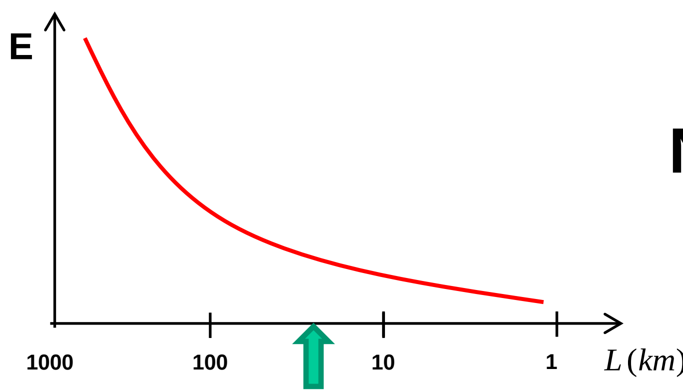


Model



Where truncation limit ; spectral gap

Unfortunately, there is no spectral gap



Nature

2) Subgrid scale process & Reynolds averaging

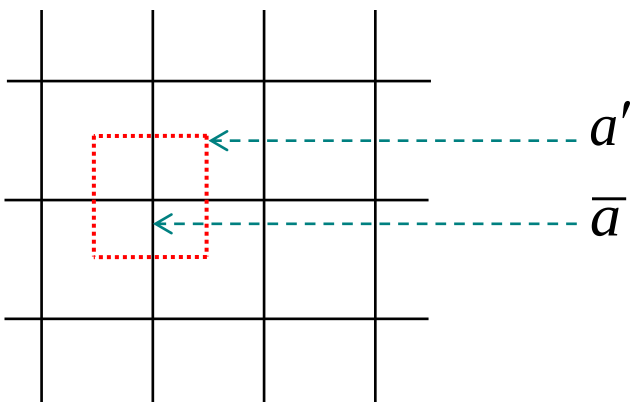
Consider prognostic water vapor equation

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

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



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

then eq.(1) becomes

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

①　　　　　　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)

$$\frac{\partial \rho w q}{\partial t} = -\frac{\partial \rho u w q}{\partial x} + \cdots$$

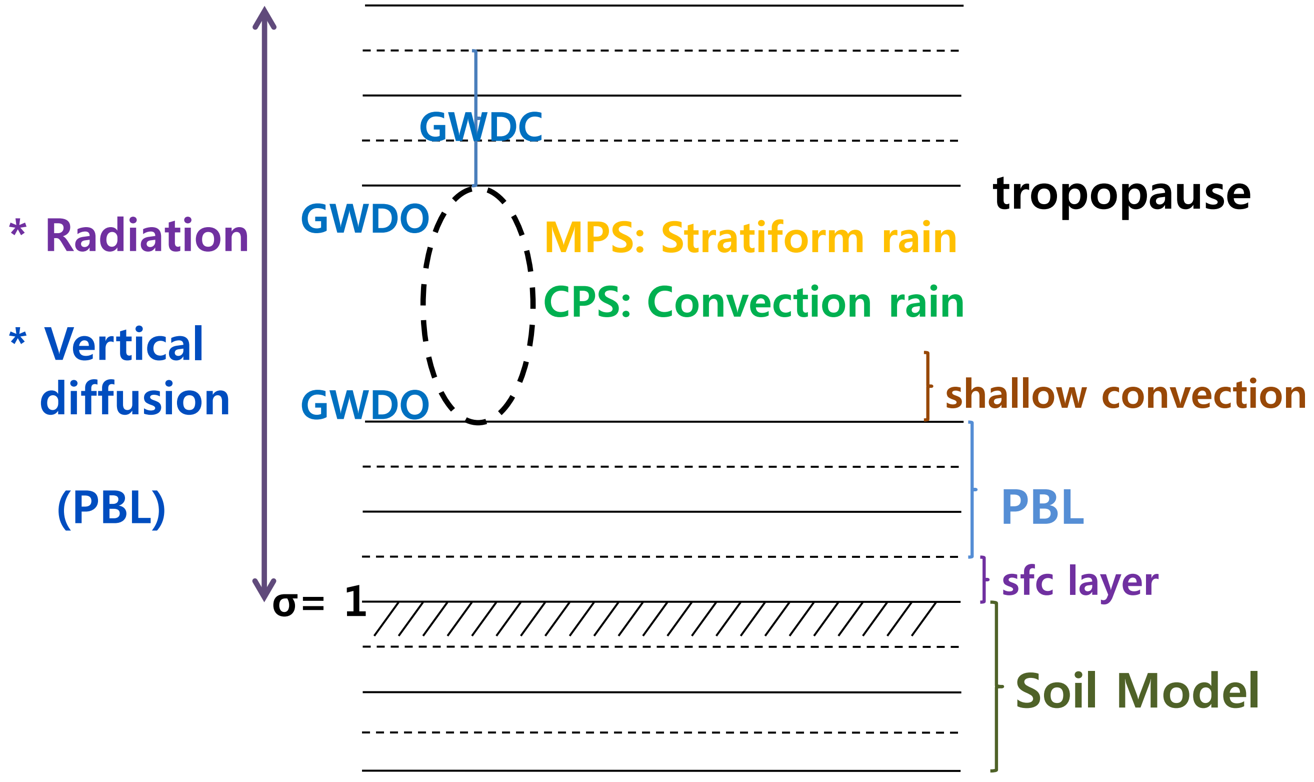
taking Reynolds averaging,

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

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

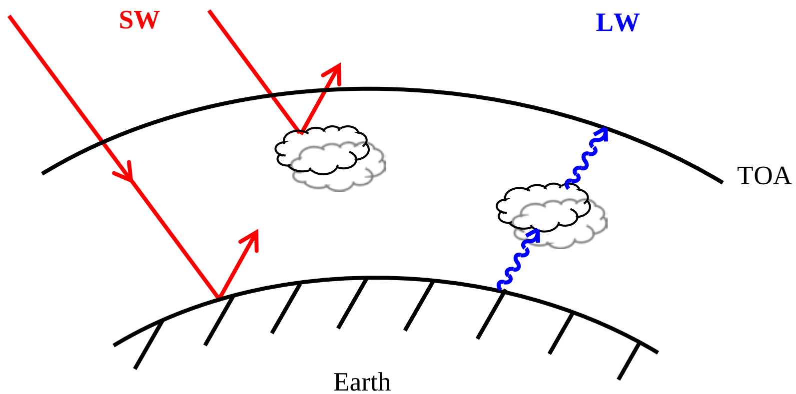
In modeled atmosphere : 7 or more

$$\frac{d\ln\theta}{dt}=\frac{H}{c_pT}\ , \qquad \frac{dq}{dt}=S, \qquad \frac{d\vec{u}}{dt}=\nabla_z\vec{\tau}$$



1. Radiation

1.1 Concept



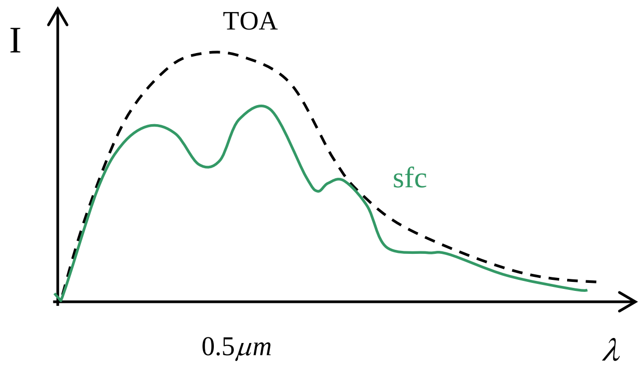
TOA : S = 1360 Wm^{-1} , $Mean\ Flux : \frac{S}{4} = 340 Wm^{-1}$ → Energy source for Earth

30% : reflected from the atmosphere clouds

$\left\{ \begin{array}{l} \text{25\% : absorbed in the atmosphere} \\ \text{45\% : absorbed at the earth surface} \end{array} \right\}$ → Back to space by terrestrial infrared radiation

At low latitude : energy gain
At high latitude : energy loss

1.2 Solar radiative transfer



- At TOA,

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

(Insolation)

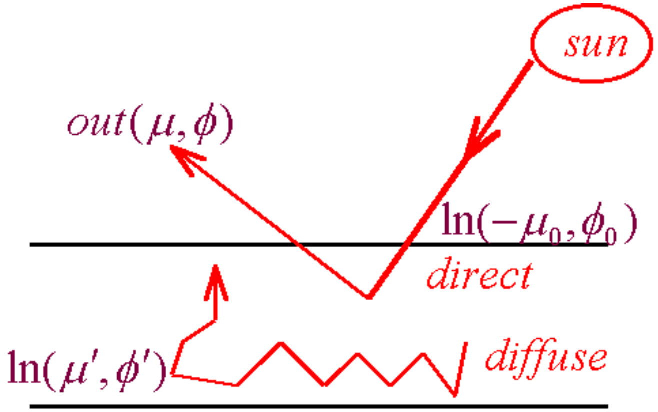
- Basic equations

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

absorption source emission

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

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



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

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

P : Scattering phase function :redirects $(\mu',\phi') \rightarrow (\mu,\phi)$

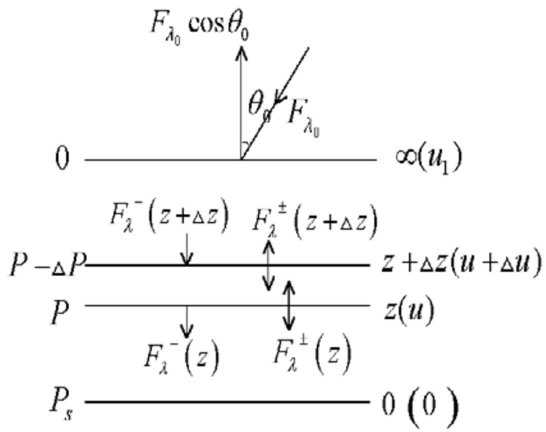
$\tilde{\omega} = \frac{\sigma_s}{\sigma_e}$: Scatteing albedo

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.

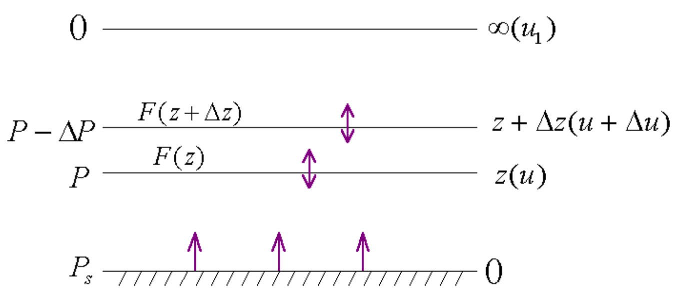
$$P(\cos\phi) = \sum_{l=0}^{N} \tilde{\omega}_l P_l(\cos\phi) \quad \text{: Legendre Polynomial}$$

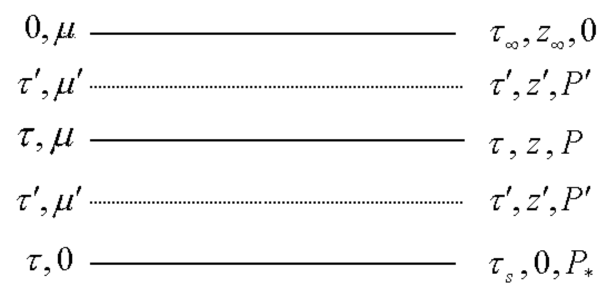


Radiative transfer equation solver.

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

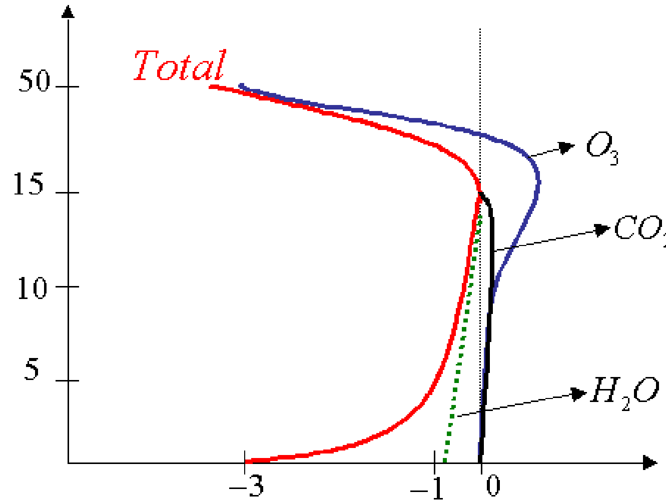
1.3 Terrestrial radiation





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

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



In spectral bands (monochromatic)

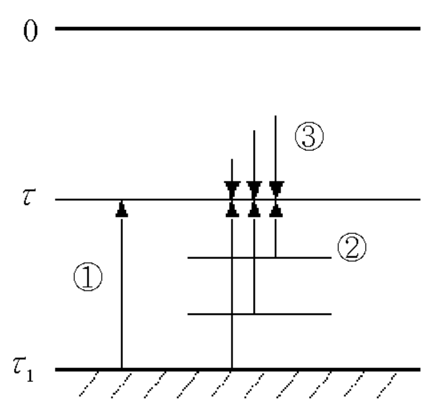
$$\uparrow \quad \mu \frac{dI_\nu(\tau,\mu)}{d\tau} = I_\nu(\tau,\mu) - B_\nu(T)$$

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

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

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

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



$$d\tau = -k_\nu \rho dz$$

$$\tau_1 = \int_0^{u_1} k_\nu du, \quad u_1 = \int_0^\infty \rho dz$$

1.4 Cloud fraction

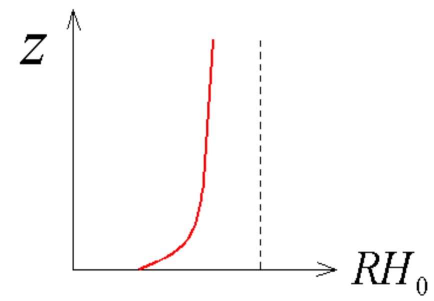
1) Conventional method

$$f = f_c + f_l$$

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

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

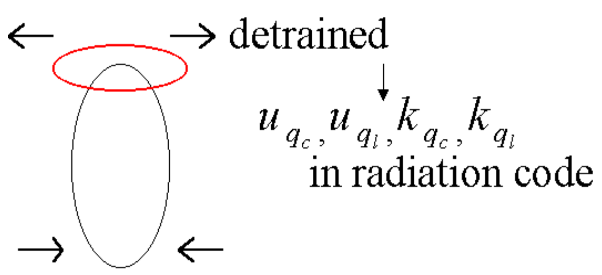
where RH_0 is the critical value of RH which is optimized based on observations. (Slingo's method)



2) Advanced method

- 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 parameterizations



f_l $\qquad$ $q_c, q_s, q_i, \cdots$

i) with diagnostic microphysics (CCM3)

- cloud water scale height r_l

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

- cloud droplet size,

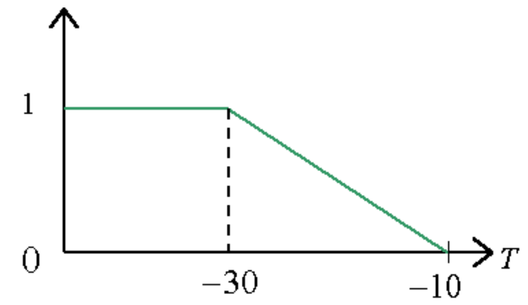
$$r_{ee} \begin{cases} = 10\mu m & \text{over ocean} \\ < 10\mu m & \text{over land} \end{cases}$$

$\cdots$ warm cloud

r_{ei}: $10\mu m \sim 30\mu m$
(low) (high)

$\cdots$ ice cloud

f_{ice} : ice fraction



- The radiation properties of ice cloud in the short wave spectral region :

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

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

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

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

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})

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

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

※ $D = 1.66$:diffusivity factor

k_{abc} : LW absorptivity coefficient.

$= k_l (1 - f_{ice}) + k_i f_{ice}$

ii) with prognostic microphysics (MM5, WRF)

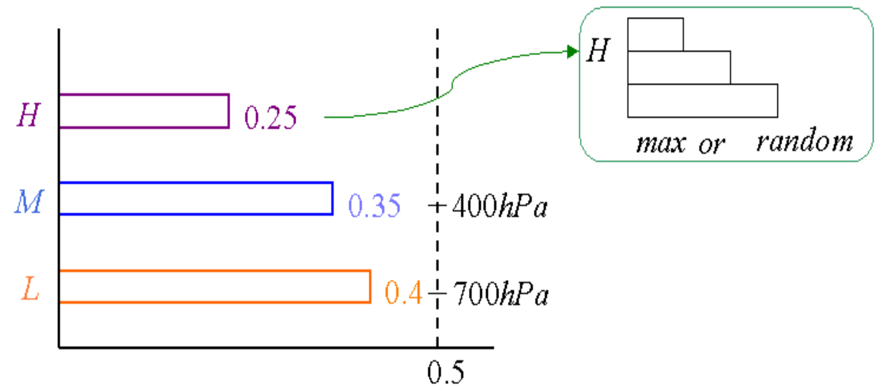
α_p (absorption coefficient)

$$= \frac{1.66}{2000} \left(\frac{\pi N_0}{\rho_{rs}{}^3}\right)^{\frac{1}{4}} \quad m^2 g^{-1} = \begin{bmatrix} 2.34 \times 10^{-3} \quad m^2 g^{-1} \text{ for snow} \\ 0.33 \times 10^{-3} \quad m^2 g^{-1} \text{ for rain} \end{bmatrix}$$

- u_p (effective water path length)

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

iii) Cloud overlapping



Max. 0.4
Min. 1.0
Random : $H + (1-H)M + \{1 - H - (1-H)M\}L = 0.6$

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

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

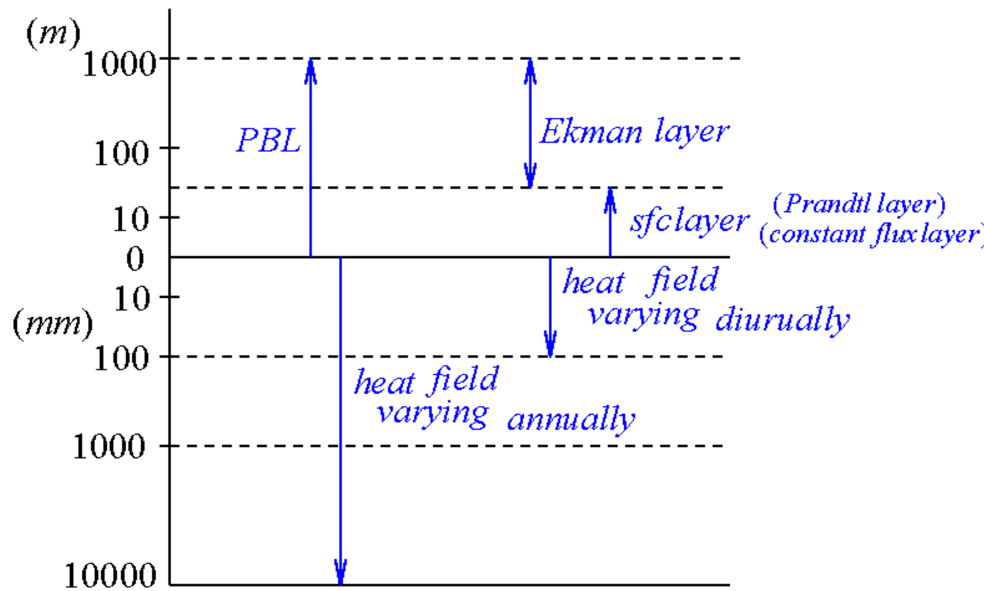
※ In very high resolution $A_c = 0$ or 1 ← in WRF

2. Land-surface processes

2.1 Concept

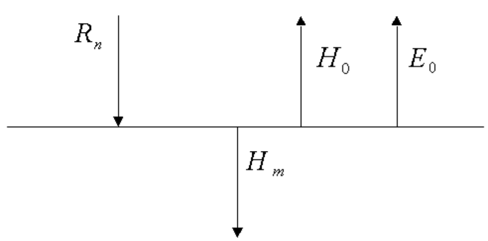
1) Surface layer

: compute surface fluxes and
update surface temperature and humidity
by solving soil model and surface energy
budget



2) Surface energy budget

$$C_g \frac{\partial T_s}{\partial t} = R_n - H_m - H_0 - E_0$$



2.2 Surface layer

1) Bulk method

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

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

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

$$C_D, C_H = fn(R_i, Z_0, \cdots)$$

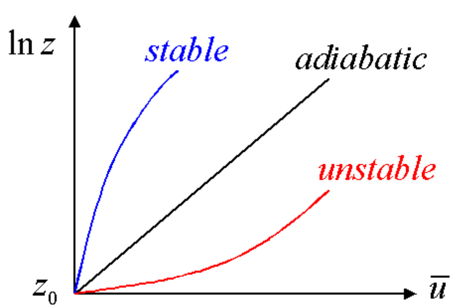
$$\frac{\partial T}{\partial t} \begin{cases} = \dfrac{\partial}{\partial z}\dfrac{H_0}{C_P} + \dfrac{T}{\theta}\rho k_z \dfrac{\partial \theta}{\partial z} & \text{at surface layer} \\ = \dfrac{T}{\theta}\rho k_z \dfrac{\partial \theta}{\partial z} & \text{above sfc} \end{cases}$$

where $k_z = fn(R_i, S)$ ← wind shear

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



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

※ **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 $L = u_*^2\overline{\theta}/(kg\theta_*) = -\frac{\rho C_p \theta_0 u_*^3}{kgH_0}$ $\quad Ri = \frac{g}{\overline{\theta}}\frac{\partial \theta}{\partial z} \Big/ (\frac{\partial u}{\partial z})^{-2}$ $\quad \frac{h_s}{L} = \frac{\phi_m^2(hs/L)}{\phi_t(hs/L)} Ri$

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

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

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

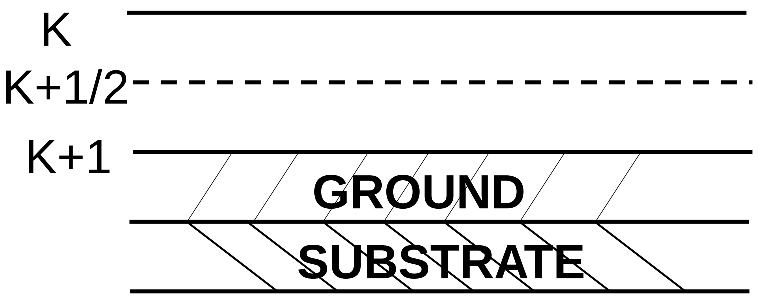
$$E_0 = -\rho L \overline{q'w'} = -\rho L C_q U \Delta q$$

2.3 Soil model

1) Slab model

$$\frac{\partial T_s}{\partial t} = \lambda_T (R_n - LE - H) - \frac{2\pi}{\tau}(T_s - T_2)$$

$$\frac{\partial T_m}{\partial t} = \frac{1}{\tau}(T_s - T_m)$$

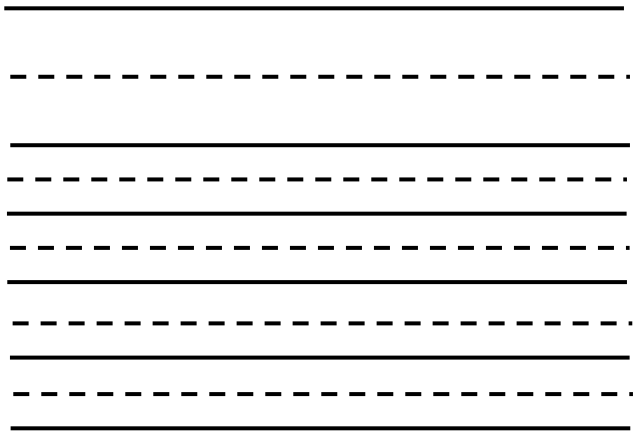


2) Multi-layer model

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

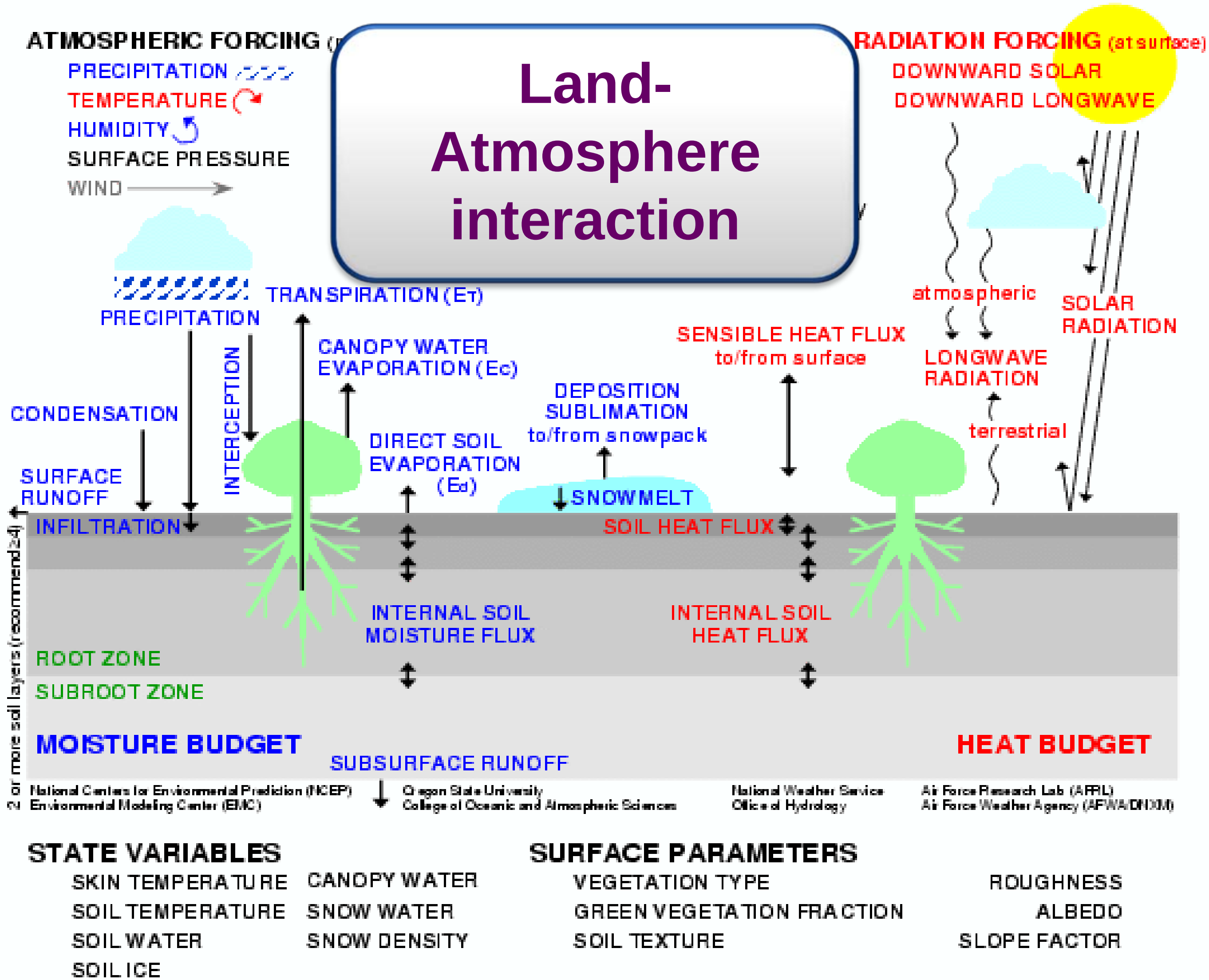
$$(\rho C)_i \frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left(\lambda T \frac{\partial T}{\partial z}\right)$$

$$\frac{\partial \eta}{\partial t} = \frac{\partial}{\partial z}\left(D \frac{\partial \eta}{\partial z}\right) + \frac{\partial K}{\partial z} + F_\eta$$



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

2.4. NOAH soil model (Chen and Dudhia 2001)



Method : Penman Approach

A combination of the energy balance and bulk transfer formula

$$Evap: = E_{dir} + E_c + E_t$$

i) E_{dir} (direct evaporation)

: direct evaporation from the bare soil, using the soil water flux at the surface

$$E_{dir} = (1 - \sigma_f)\left[-D(\Theta)_0 - K(\Theta)_0\right]$$

vegetation fraction diffusivity conductivity

ii) E_c (Canopy reevaporation)

: when rain falls to the ground, the leaves first intercept the rain up to a S. The canopy water then reevaporates and the excess drips to the ground canopy water amount, n = 0.5

$$E_c = \sigma_f \left(\frac{C^*}{S}\right)^n E_p$$

canopy capacity (=0.002m)

iii) E_t (Transpiration)

: process whereby extracts water from the root zone and release it to the atmosphere from the leaf stomata during photosynthesis

$$E_t = \sigma_f \left(1 - \frac{C^*}{S}\right)^n \overline{\beta} E_{tp} \,, \qquad \beta_i = \frac{\Theta_i - \Theta_{wilt}}{\Theta_{fc} - \Theta_{wilt}}$$

Saturation ratio

iv) E_p (Potential evaporation) and E_{tp} (potential evapotranspiration)

E_p : Maximum evaporation when soil is saturated

E_{tp} : Maximum evaporation when plant has not water stress

$$R_{net} = G + H(T_s) + LE_p$$

$$LE_p = \frac{[R_{net} - G]\Delta + (1+\gamma) LE_a}{\Delta + 1 + \gamma}$$

$$LE_{tp} = \frac{[R_{net} - G]\Delta + (1+\gamma) LE_a}{\Delta + (1+\gamma)(1 + C_h V r_s)}$$

where $\gamma = \dfrac{4\sigma T_a^3}{\rho_a c_p C_h U}$, $\Delta = \dfrac{L}{C_p} \dfrac{dq_s}{dT}\Big|_{T_a}$ $\qquad E_a = \rho_a C_h U [q_s(T_a) - q_a]$

drying power of air

v) Soil hydrology : Darcy's law

$$\frac{\partial \Theta}{\partial t} = \frac{\partial}{\partial z}\left[D(\Theta) \frac{\partial \Theta}{\partial z} \right] + \frac{\partial K(\Theta)}{\partial z}$$

Precipitation and direct evaporation form the water flux at the surface while deep layer drainage provides the bottom boundary condition. When precipitation rate exceeds the rate at which the soil can transport the water downward, runoff is assumed.

vi) Canopy budget

$$\frac{dC^*}{dt} = \sigma_f \, \mathrm{Pr}\, ecip - E_c$$

vi) Soil thermodynamics

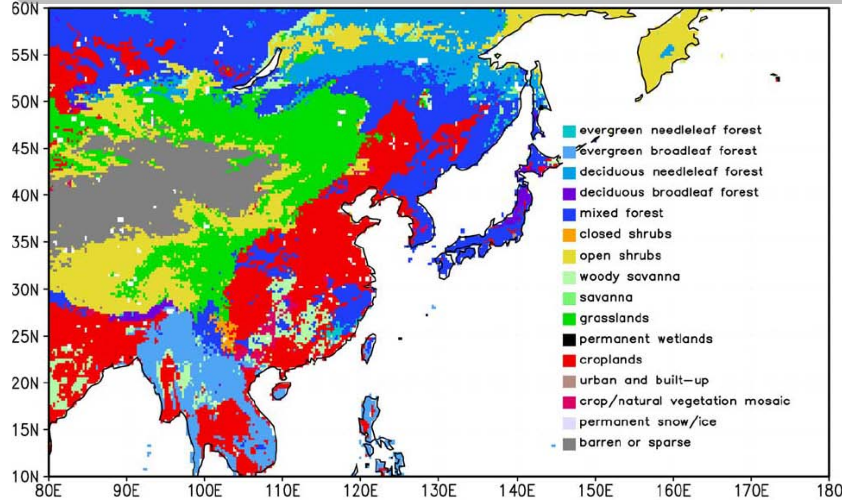
$$C(\Theta)\frac{\partial T}{\partial t} = \frac{\partial}{\partial z}\left[K_T(\Theta) \frac{\partial T}{\partial z} \right]$$

heat capacity with the ground heat flux as the surface boundary condition, and prescribed deep soil temperature

2.5. Vegetation type → z0, albedo

Vegetation types

MODIS (satellite)

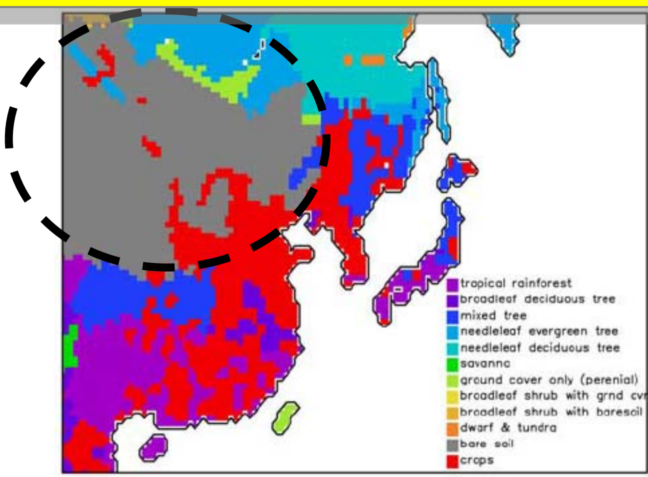


13 type data set
1 degree

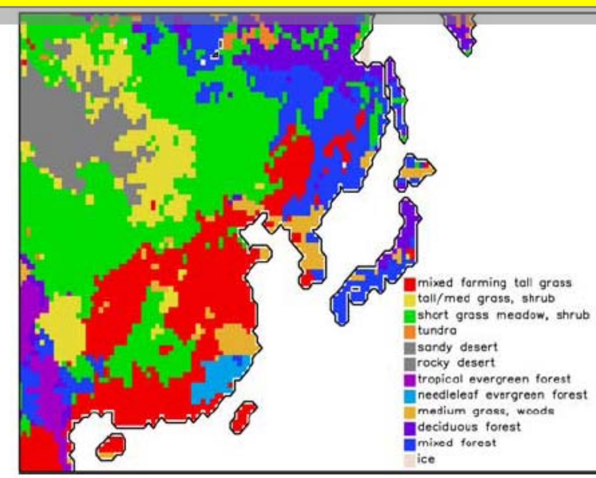
12 type data set
20 min

The Simple Biosphere model (SiB)

quite broad desert region



United States Geological Survey's (USGS)

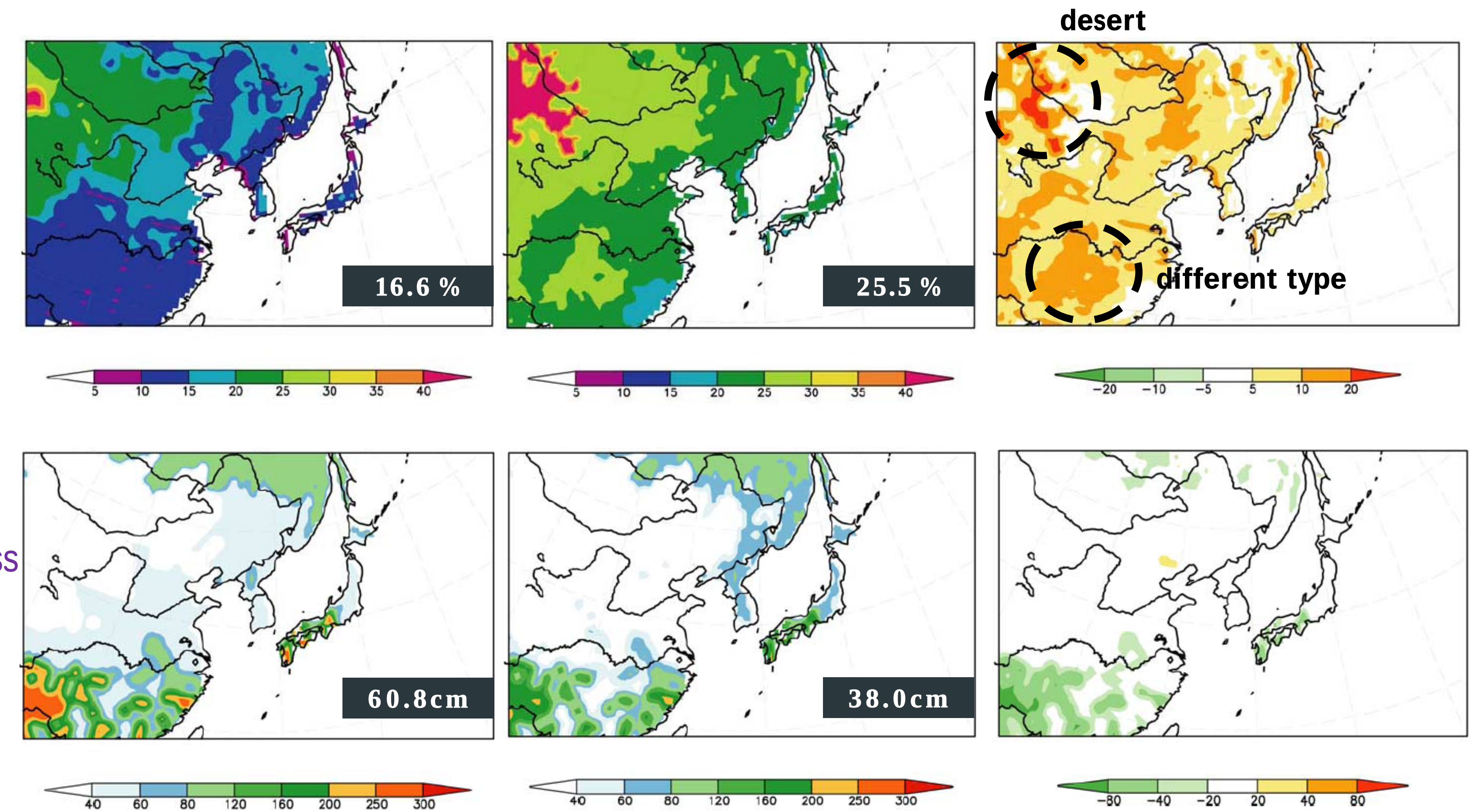


Albedo and Roughness length (z0)

	Sib	USGS	USGS-Sib

Albedo

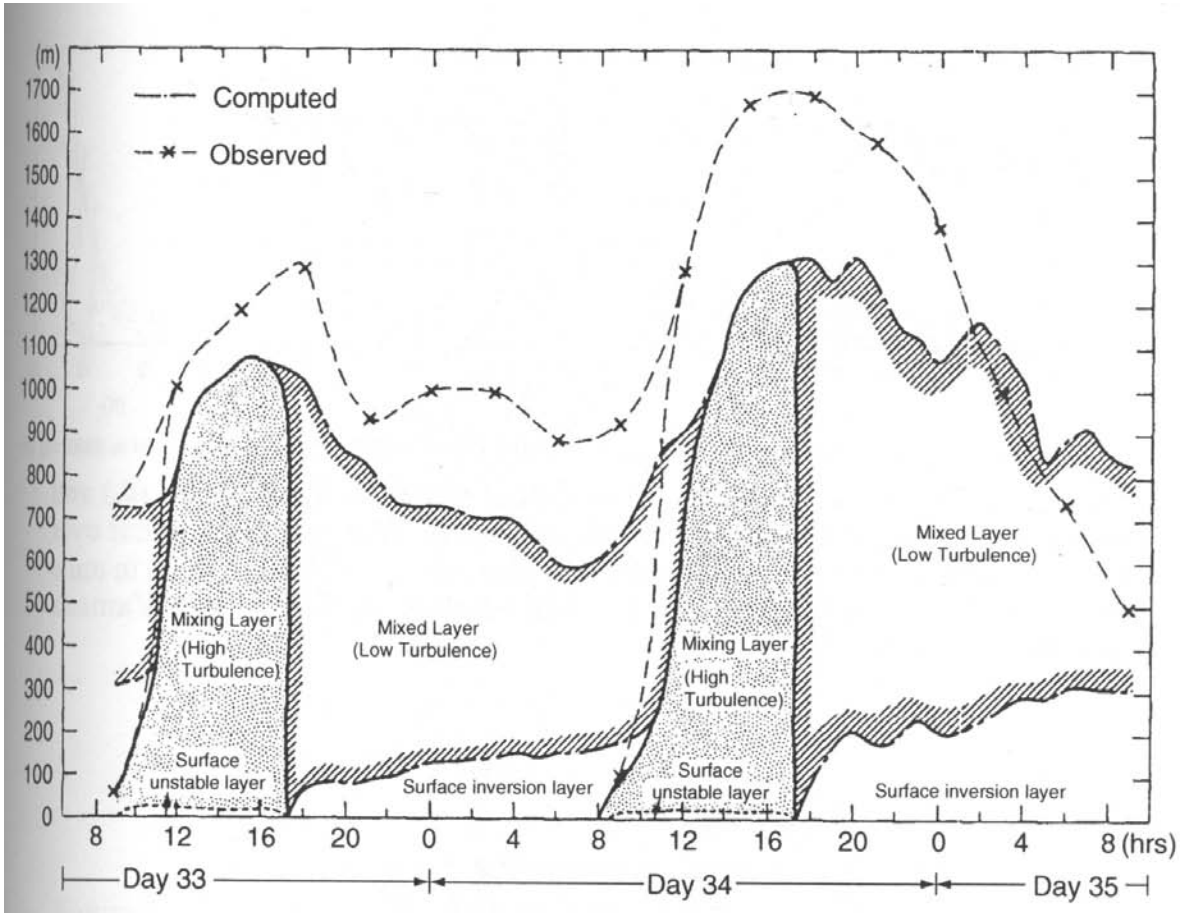
Roughness length



3. Vertical diffusion (PBL)

Purpose

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



Yamada and Mellor(1975)

Planetary Boundary Layer Structure

Daytime profiles

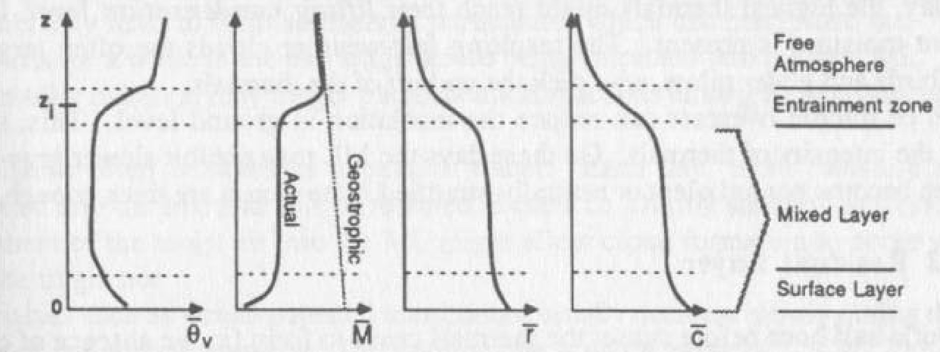


Fig. 1.9 Typical daytime profiles of mean virtual potential temperature $\overline{\theta_v}$, wind speed $\overline{M}$ (where $\overline{M}^2 = \overline{u}^2 + \overline{v}^2$), water vapor mixing ratio $\overline{r}$, and pollutant concentration $\overline{c}$.

Daytime flux profiles

Nighttime flux profiles

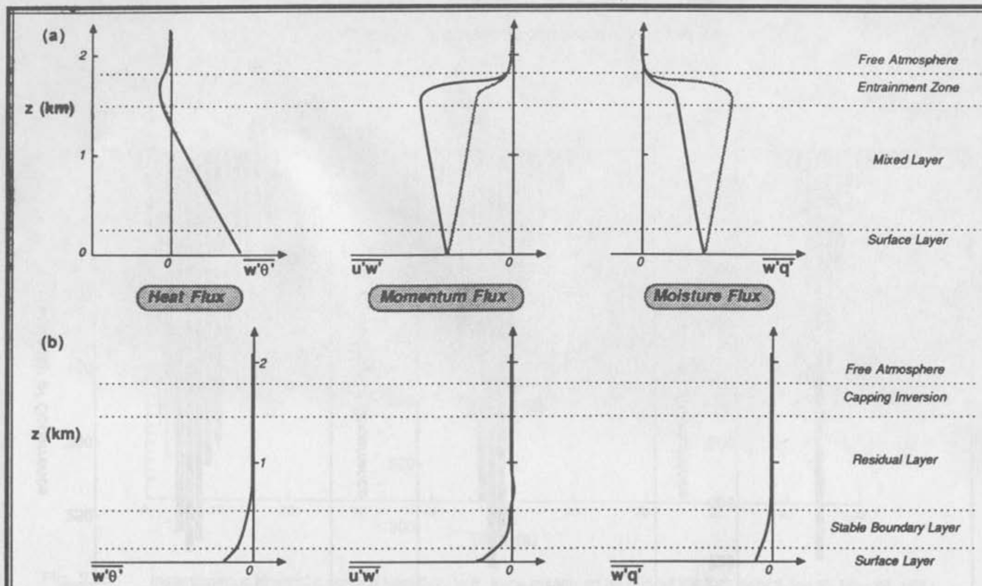


Fig. 2.15 Idealized vertical soundings of turbulent fluxes for the (a) convective mixed layer and (b) stable boundary layer.

From Stull (1988)

* Classifications :

3.1 Local vertical diffusion

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

Local Richardson number

3.2 Nonlocal PBL

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

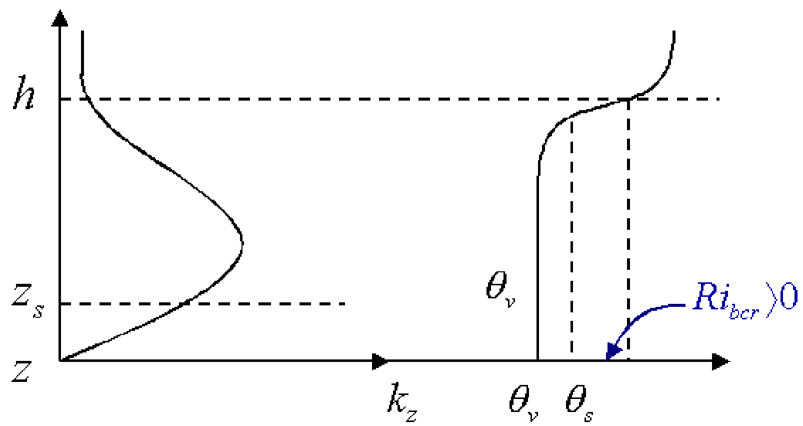
$$k_{zm}=kw_s z(1-\frac{z}{h})^p$$

$$h=R_{ibcr}\frac{\theta_m}{g}\frac{U^2(h)}{(\theta_v(h)-\theta_s)}$$

$$\theta_s=\theta_{va}+\theta_T \ (=b\frac{\overline{(\theta_v{'}w')_0}}{w_s})$$

$$p_r=\left[\frac{\phi_t}{\phi_m}+bk\frac{0.1h}{h}\right]$$

$$w_s=u_*\phi_m^{-1}$$



3.3 TKE (Turbulent Kinetic Energy)

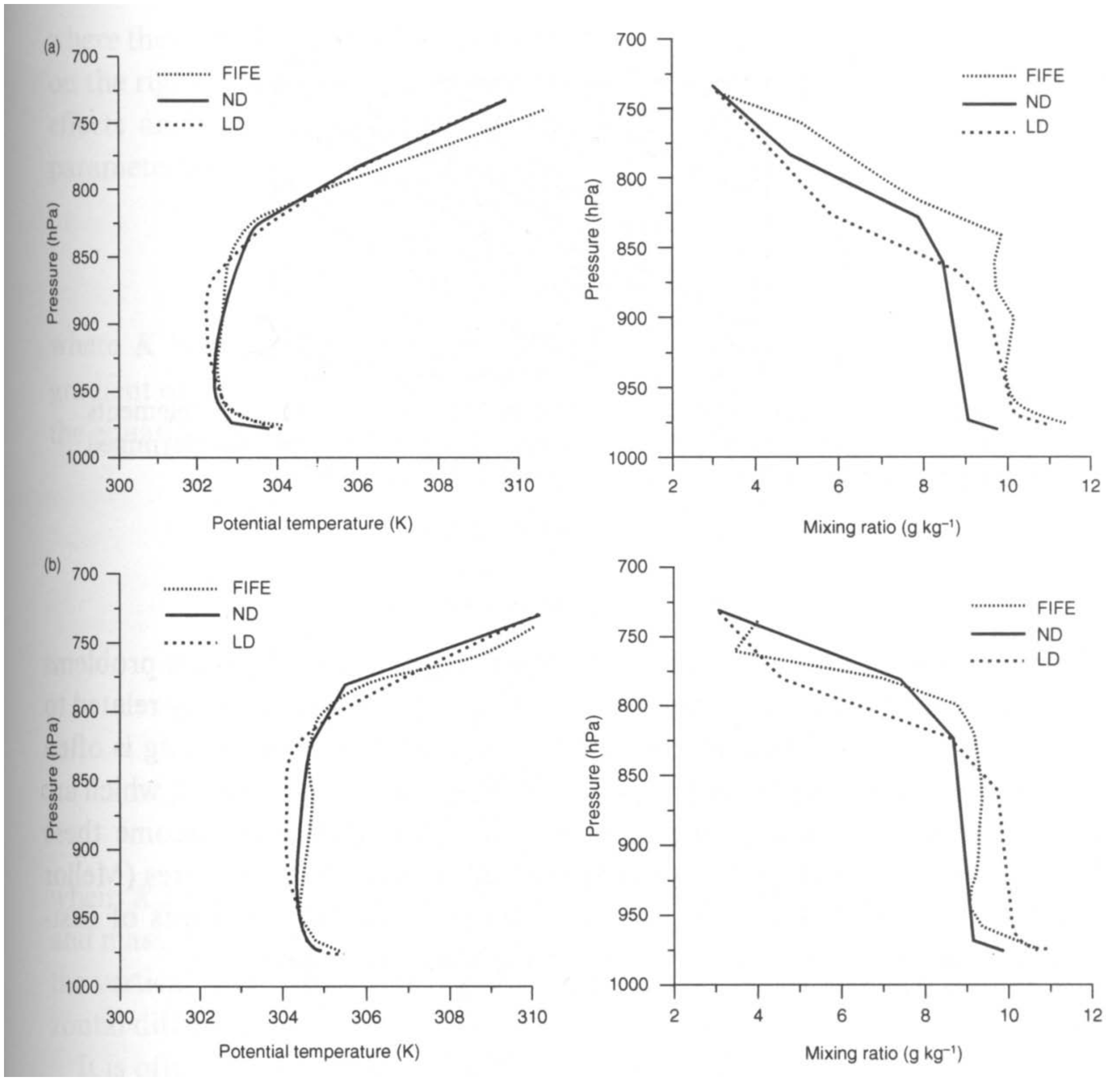
TKE eqn.

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

$$\rightarrow \overline{u_i u_j} \rightarrow k_z = fn(\overline{e_{ij}})$$

(Mellor & Yamada, 1982)

From Hong and Pan (1996)



Local scheme typically produces unstable mixed layer

4. Gravity Wave Drag

*** GWDO : GWD induced by orography**

*** GWDC : GWD induced by convection**

4.1 Concept

This scheme 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.

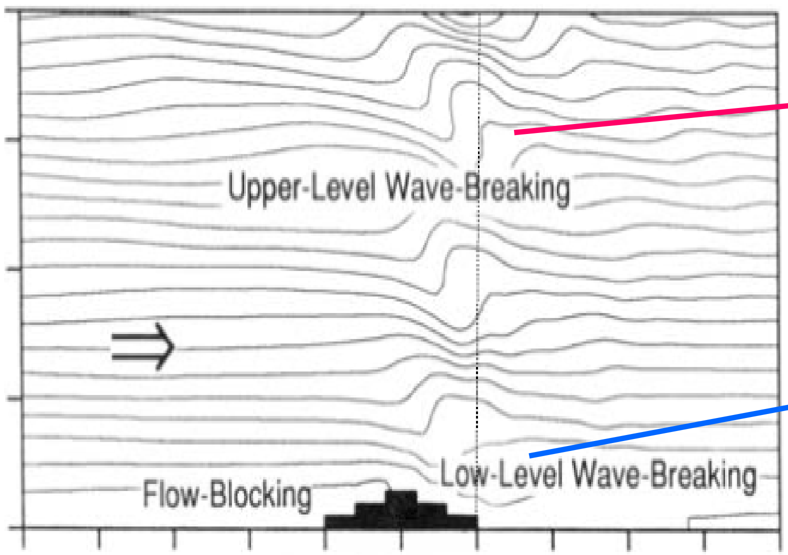
- Wave drag amount at reference level :

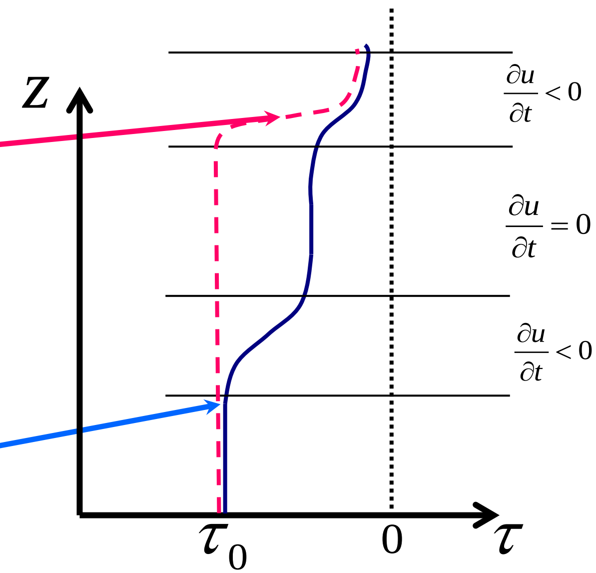
$$\tau_{GWD} = -E\frac{m}{\Delta x}\frac{\rho_0 U_0^3}{N_0}\frac{Fr^2}{Fr^2 + C_G / OC} \qquad E \equiv (OA+2) C_E Fr_0 / Fr_c$$

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.**

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





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

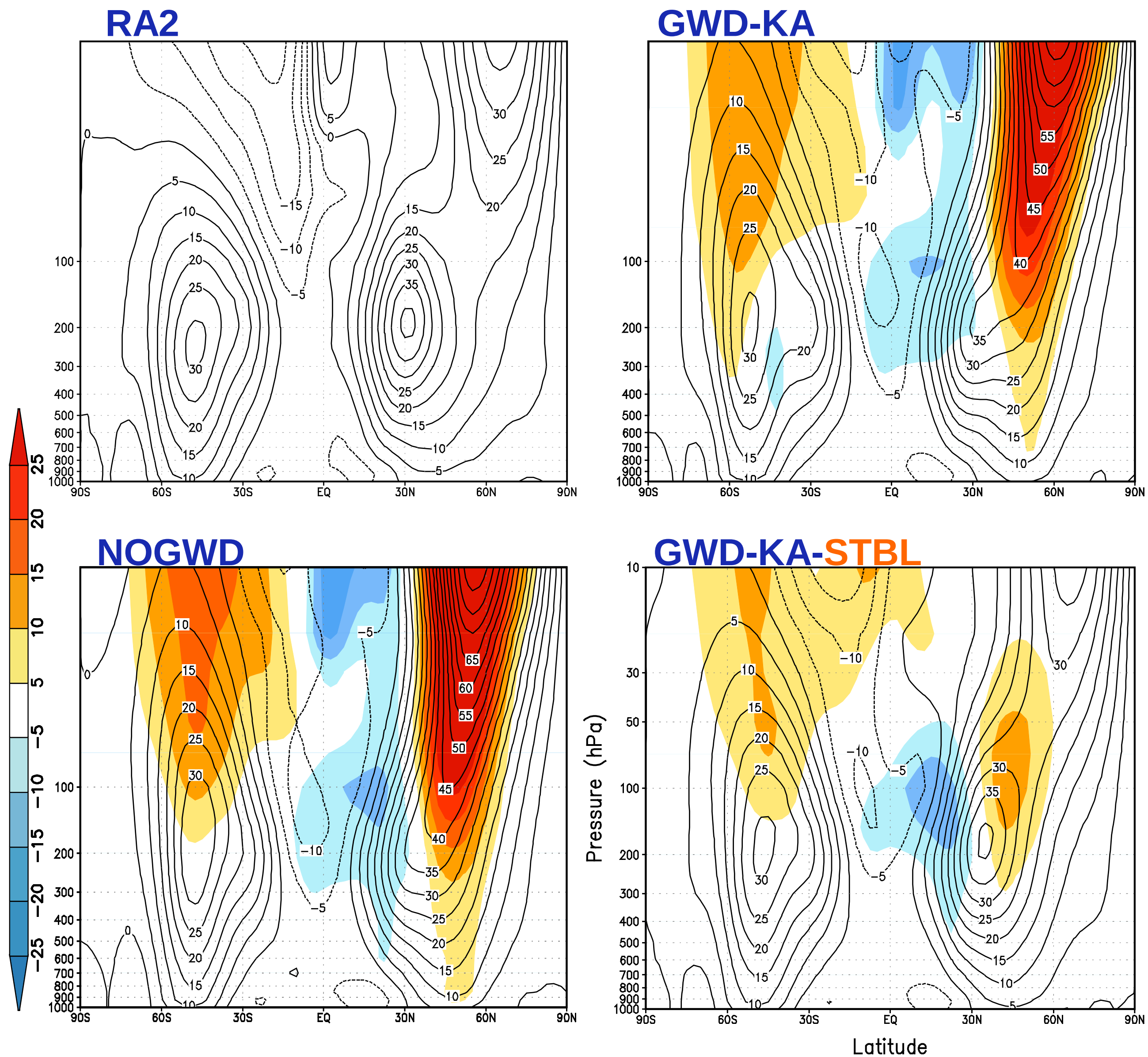
$$\frac{\partial v}{\partial t} = -\frac{1}{\rho}\frac{\partial \tau_y}{\partial z}$$

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)

4.3 Impact of GWDO

Zonal-averaged zonal wind (96/97 DJF)



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

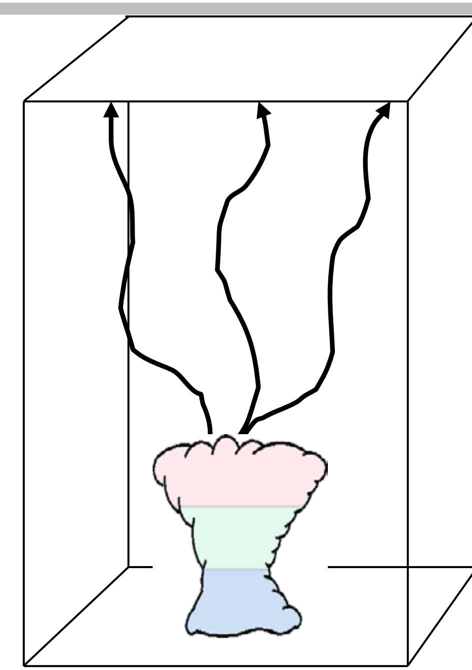
Kim and Arakawa
→Improves upper level jets
→Improves the sea level pressure

(Kim and Hong, 2009)

Parameterizations of Convective GW Drag (CWDC)

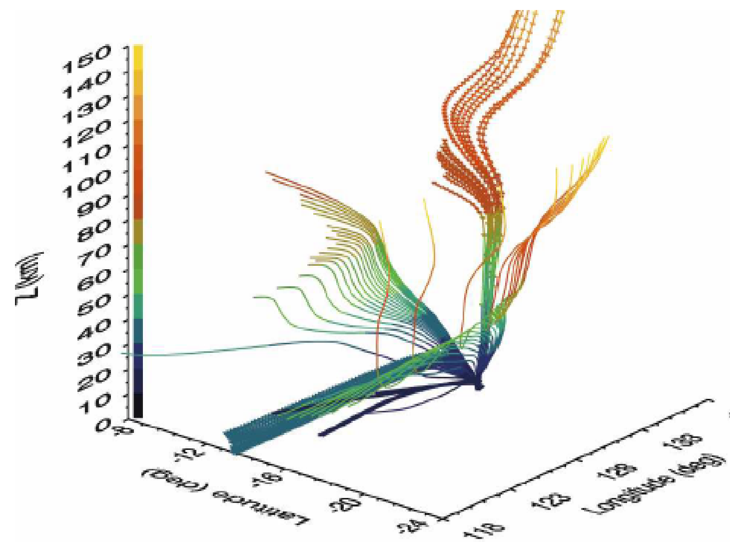
Columnar CGWD parameterizations

- **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



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Cloud-top GW Momentum Flux Spectrum

Song and Chun (2005, JAS)

▪ Cloud-top GW momentum flux spectrum

$$M_{ct}(c,\varphi,z_{ct}) = \mathrm{sgn}[c-U_{ct}(\varphi)]\rho_{ct}\frac{2(2\pi)^3}{A_h L_t}\left(\frac{g}{c_p T_{ct} N_q^2}\right)^2 \underbrace{\frac{N_{ct}|X|^2}{|c-U_{ct}(\varphi)|}}_{\text{WFRF}}\underbrace{\Theta(c,\varphi)}_{\text{Source}}$$

- ✓ c : phase speed (-100 m/s ~ 100 m/s, dc = 2 m/s)
- ✓ **φ : wave-propagation direction ($0^\circ \le \varphi < 180^\circ$)**

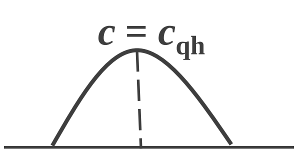
  [$\varphi = 0^\circ, 90^\circ$ in Song et al. 2007; Chun et al. 2008; Song and Chun 2008,
  $\varphi = 45^\circ, 135^\circ$ in Choi and Chun 2011]

▪ Wave-filtering-and-resonance-factor (WFRF)

- ✓ Wave filtering by the vertical propagation condition
- ✓ Resonance between the vertical harmonics consisting of convective source and natural wave modes with the vertical wave numbers given by the dispersion relation of internal GWs

▪ Convective source spectrum

$$\Theta(c,\varphi) = q_0^2\left(\frac{\delta_h\delta_t}{32\pi^{3/2}}\right)^2\frac{1}{\sqrt{1+(c-c_{qh})^2/c_0^2}},$$

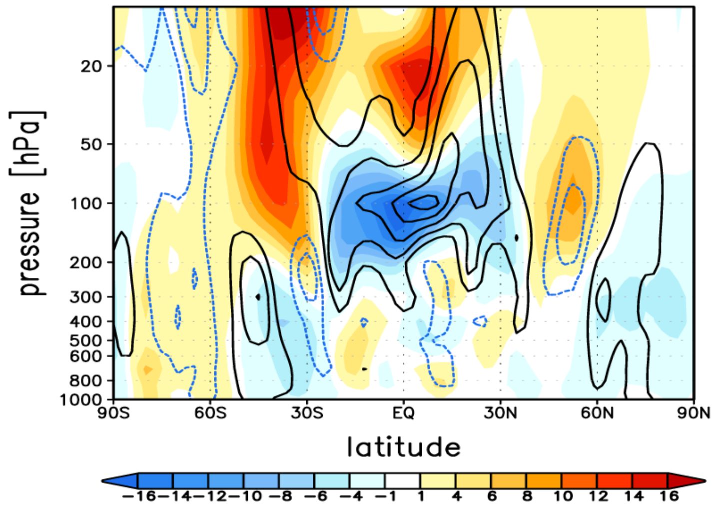


- ✓ δ_h, δ_t : spatial and time scales of the convective source (= 5 km, 20 min), $c_o = \delta_h / \delta_t$
- ✓ **c_{qh} : moving speed of the convective source ($=c_{qx}\cos\varphi + c_{qy}\sin\varphi$)**

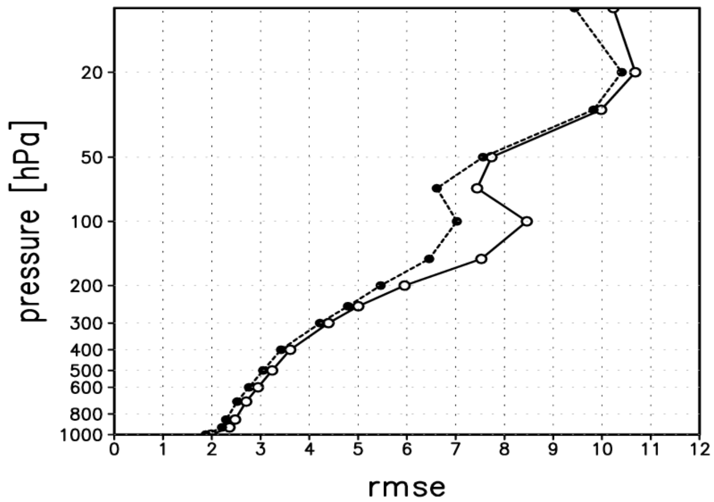
  [$c_{qh}(\varphi) = (\bar{u}_{CL} - \bar{u}_{LLJ})\cos\varphi + (\bar{v}_{CL} - \bar{v}_{LLJ})\sin\varphi$ (Corfidi et al. 1996) in the original parameterizations
  $c_{qh}(\varphi) = \bar{u}_{700}\cos\varphi + \bar{v}_{700}\sin\varphi$ in Choi and Chun (2011)]

Improvement by GWDC

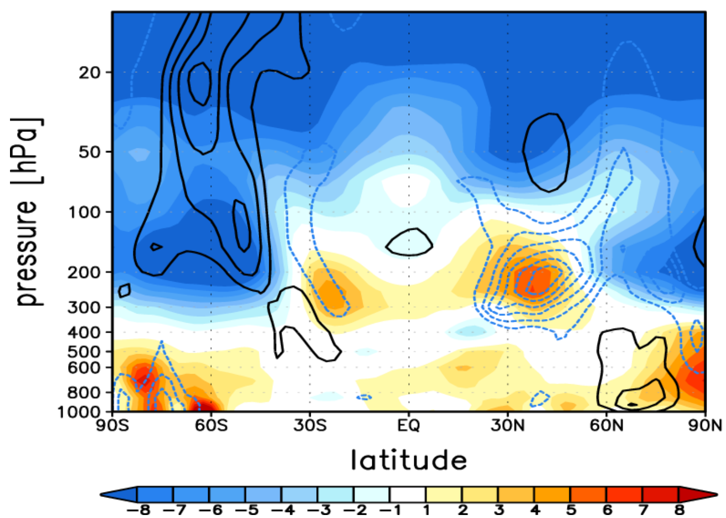
a) SAS_CB98 Zonal wind difference



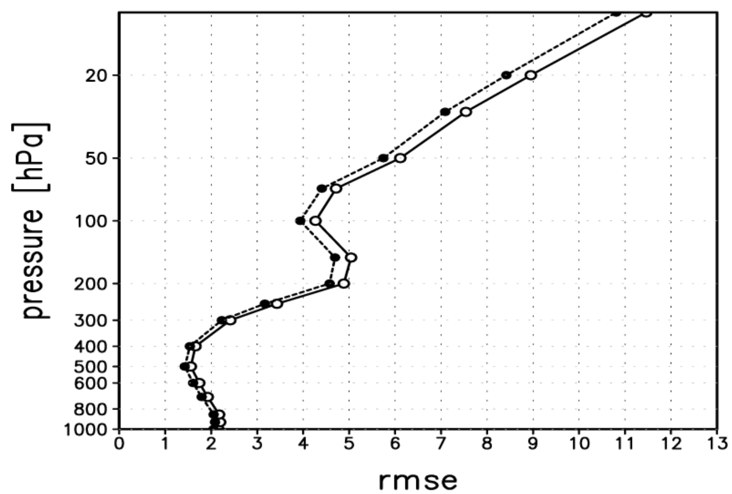
c)



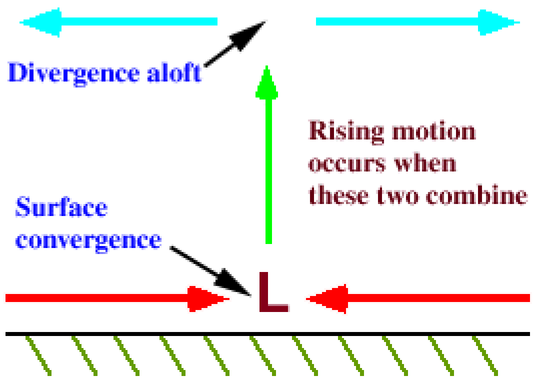
a) SAS_CB98 Temperature difference

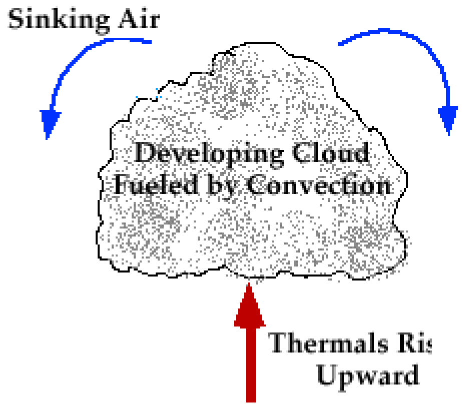


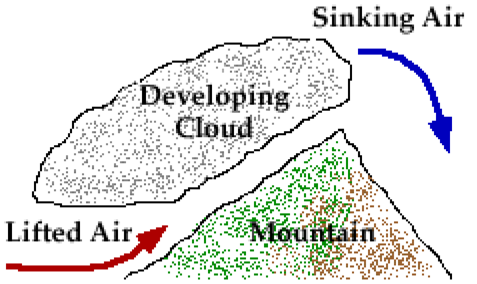
b)



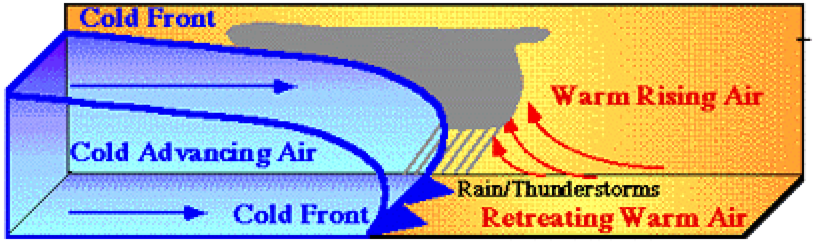
* Precipitation Processes: Introduction



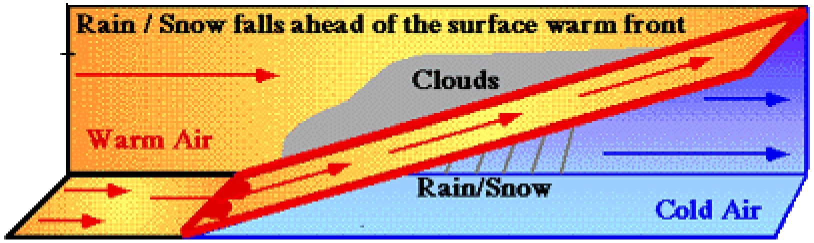




Cold front



Warm front



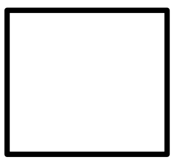
Precipitation algorithms (CPS and MPS)

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

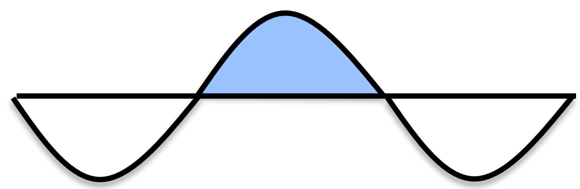
In modeled atomsphere, RH < 1

But generate clouds by releasing CAPE→ requires parameterized precip. process

Deep convection : 1~10km



$\overline{RH} < 1$



$\triangle x \rightarrow 0$, more grid-resolvable precipitation

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

- Grid-resolvable, explicit, large-scale, cloud, microphysics: Supersaturation → clouds
- Subgrid scale, implicit, cumulus parameterization, parameterized convection, deep convection: Convectively unstable → clouds

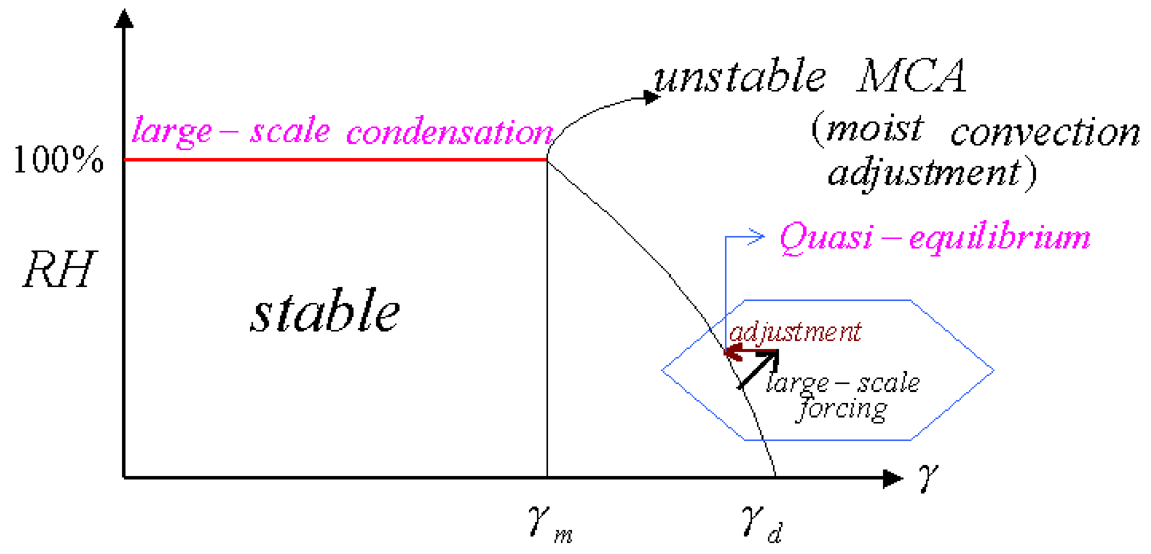
5. Deep Convection

Parameterized convection
Cumulus convection
Subgridscale precipitation
Implicit precipitation

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**



5.2 Kuo scheme (1965)

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

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

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

- Modified Kuo

- Krishnamurti el al. (1980, 1983)

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

- Anthes (AK : 1977)

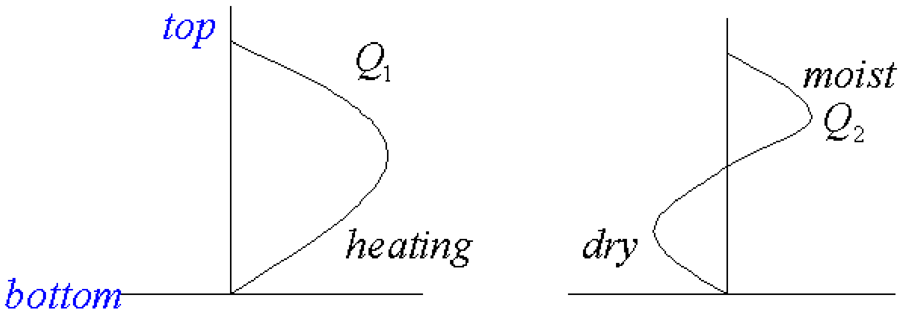
$$b = \left(\frac{1-RH}{1-RH_c}\right)^n ,$$ Check convective instability

- **Heating and moistening profiles**

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

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

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

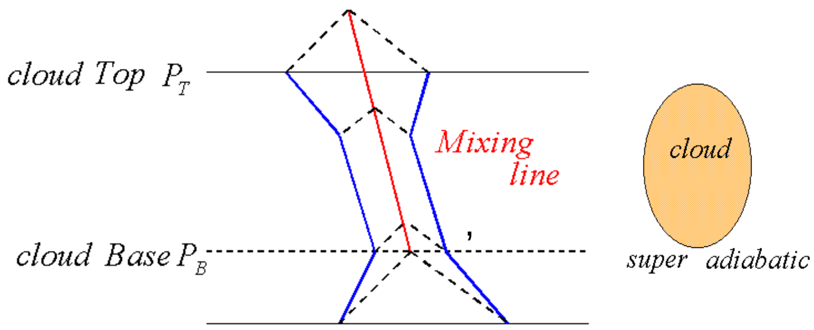


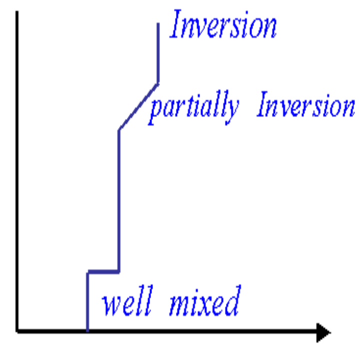
- **Remarks :**

Kuo scheme produces NPS. Also, warm bias in the lower troposphere

5.3 Betts-Miller scheme (1986)

- Adjust toward reference profiles that are based on observational evidence of convective equilibrium





- **Reference profile**

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

$$\frac{\partial q}{\partial p} = \beta \left(\frac{\partial q}{\partial p_*}\right)_M \qquad \beta = \frac{\partial p^*}{\partial p}$$

p^* : saturation pressure (= 1.2 for example)

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

M : Mixing line

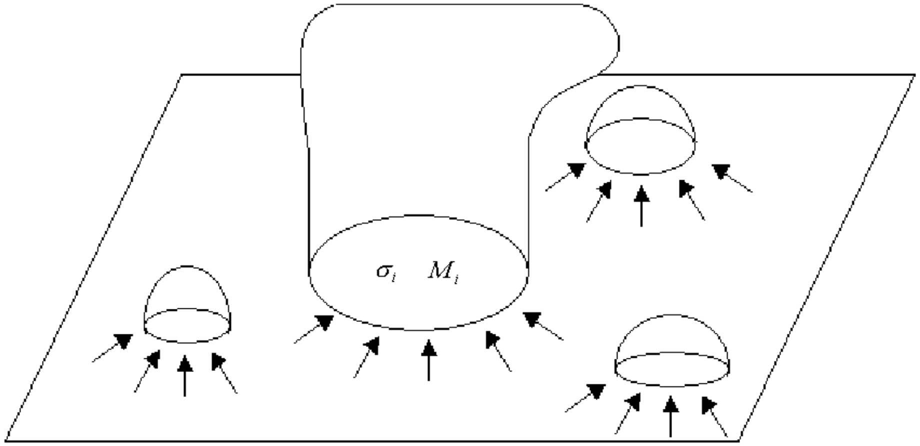
Energy Constraints : $\int_{P_B}^{P_{T+1}} C_P \left(T_R - \overline{T}\right) dp = \int_{P_B}^{P_{T+1}} (q_R - \overline{q}) dp = 0$

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

- mass flux approach, 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

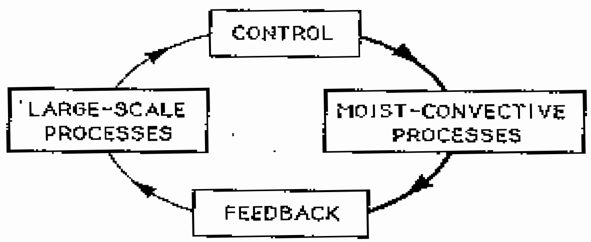
σ_i : fractional area covered by ith cloud

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

environment

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

: net mass flux/unit large-scale horizontal area



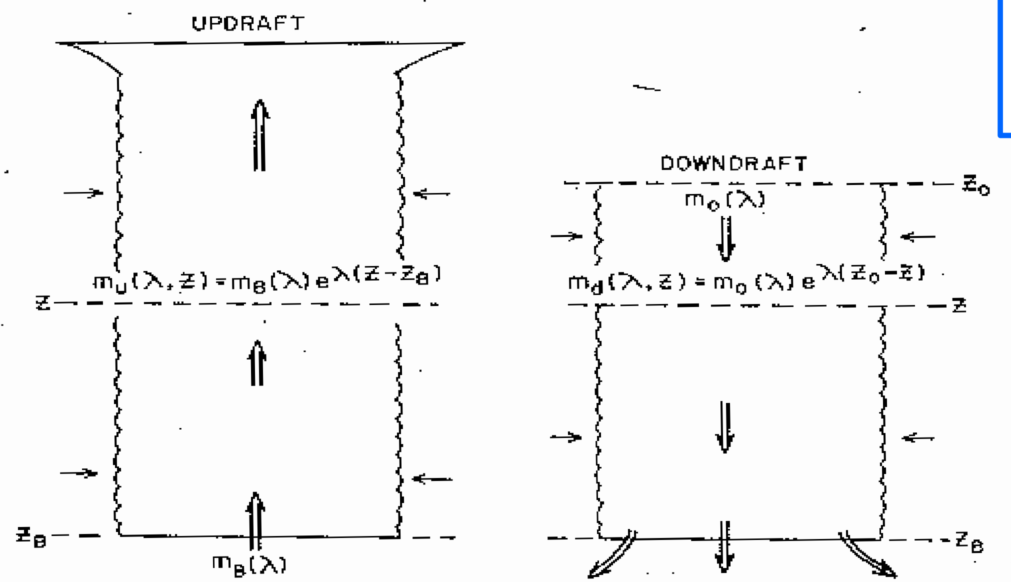


FIG. 4.10. Model for updraft and downdraft of cloud type λ (from Johnson 1976).

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

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

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

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

$S_i : C_pT + gz$ of i[th] cloud

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

Ci : condensation in the ith cloud

E : evaporation of liquid water in the environment

Qr : Radiational heating

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

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

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

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

Detraining clouds

detrainiment, entrainment

Adiabatic warming due to hypothetical subsidence between the clouds

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

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

Spectral cloud ensemble

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

subensemble

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

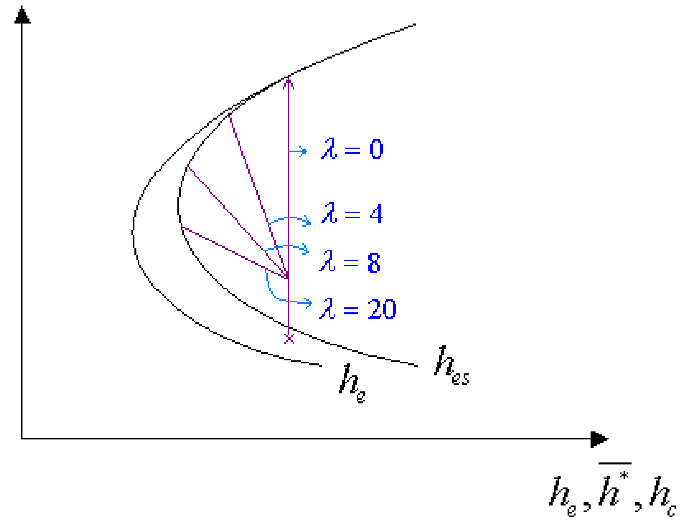
$$= \int_0^{\lambda_{\max}} m_B(\lambda)\eta(z,\lambda)d\lambda$$

Mass flux at cloud base

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

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

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



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

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

Large−scale forcing
>0 : destabilized

Adjustment
<0 : stabilization

Kernel : Cloud scheme kinetic energy

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

$$\Rightarrow m_B$$

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

*** AS type mass flux schemes**

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

1.1 Purpose

more vigorous vertical mixing of q and T in the same way as the vertical diffusion with the enhanced turbulence diffusivity within clouds. 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.

1.2 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)

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

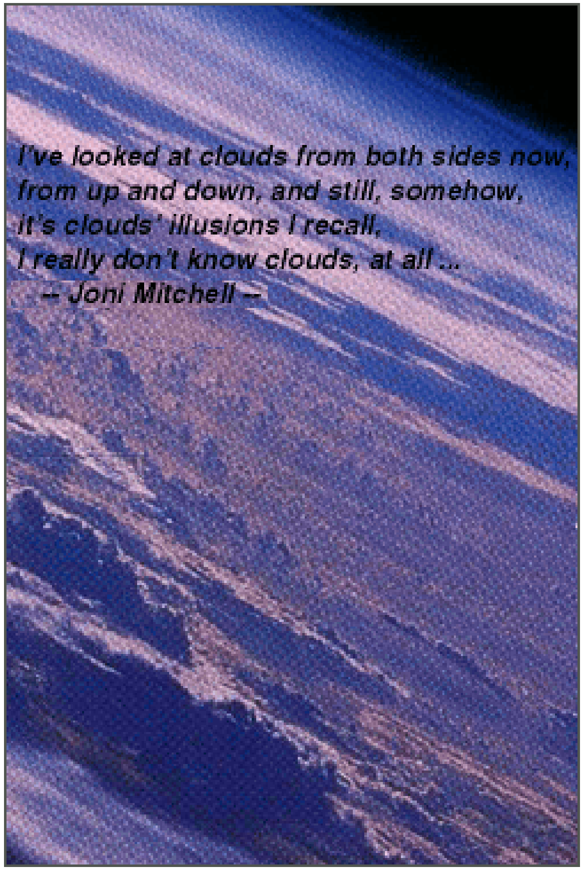
$$\frac{\partial q}{\partial t} = \frac{1}{\rho}\frac{\partial}{\partial z}\left(\rho K\frac{\partial q}{\partial z}\right)$$

Han and Pan (2011)

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

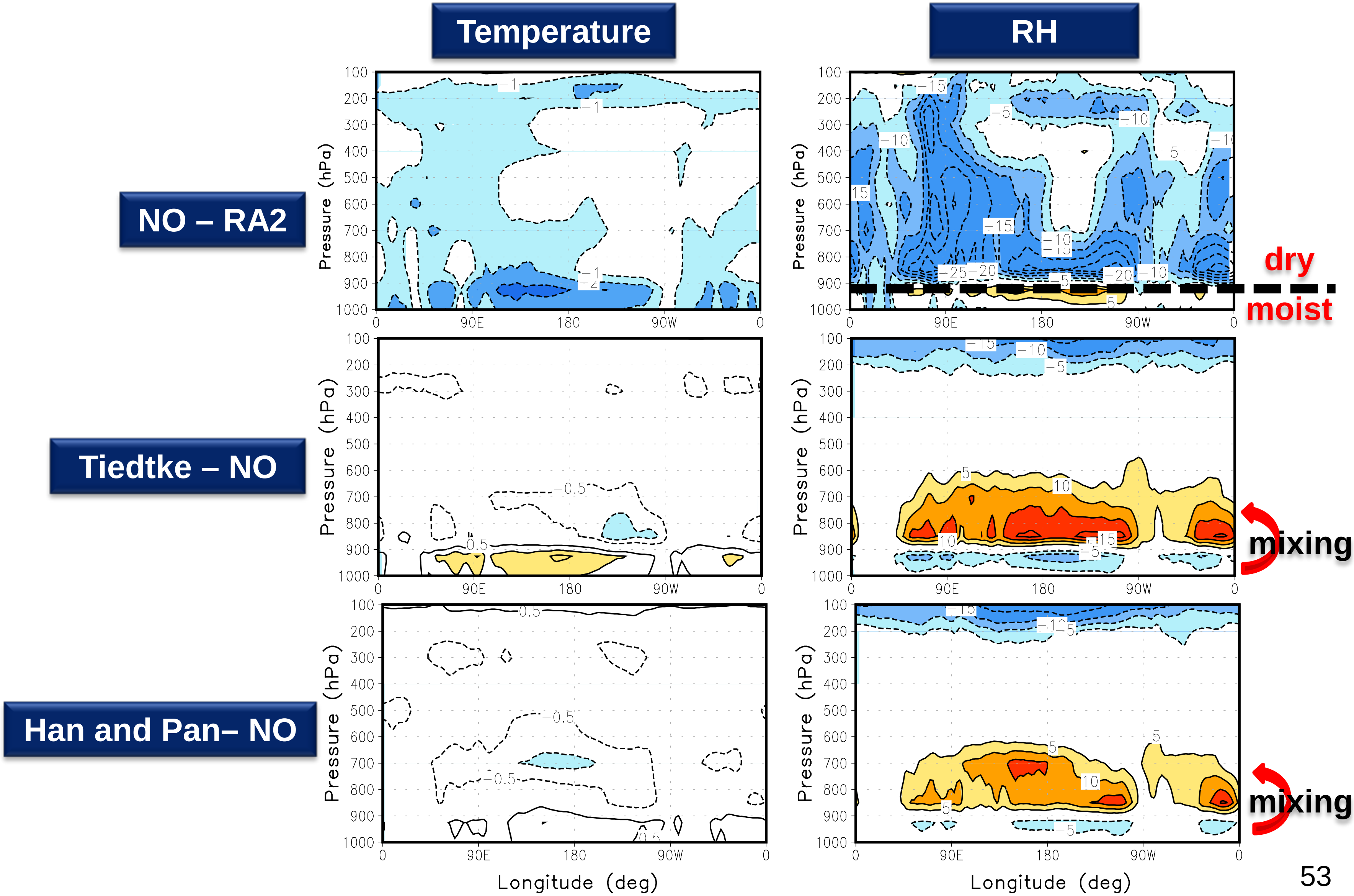
$$\frac{\partial(\eta s)}{\partial z} = (\varepsilon\bar{s} - \delta s)\eta$$

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


I've looked at clouds from both sides now,
from up and down, and still, somehow,
it's clouds' illusions I recall,
I really don't know clouds, at all ...
-- Joni Mitchell --

1.3 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 Purpose

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

7.2 Classification according to the complexity in microphysics

1) Diagnostic : condensation, evaporation of falling precipitation

2) Bulk microphysics :

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

3) 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 Gunn and Marshall(1958), respectively. The size distributions for both rain and snow are formulated according to an inverse-exponential distribution and its formula for rain can be expressed by

$$N_R(D_R) = N_{0R}\exp(-\lambda_R D_R) \qquad \text{A1}$$

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,

$$\lambda_R = \left(\frac{\pi\rho_w N_{0R}}{\rho q_R}\right)^{1/4} \qquad \text{A2}$$

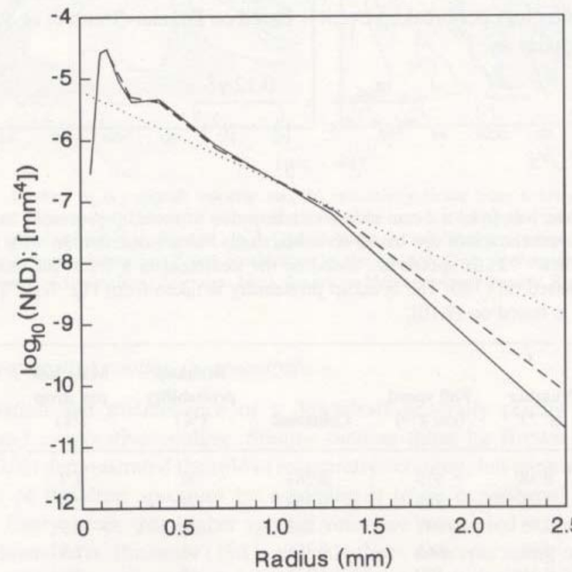


FIG. 9.21. Evolution of a drop spectrum in a subsaturated rainshaft including the effects of coalescence and breakup. The spectrum at the top of the rainshaft is Marshall–Palmer with a rainfall rate of 110 mm h^{-1}. (······), 0 km; (- - -), 1 km; (–), 2 km; R = 110 mm h^{-1}; t = 24 min. From Tzivion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

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

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

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

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)

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

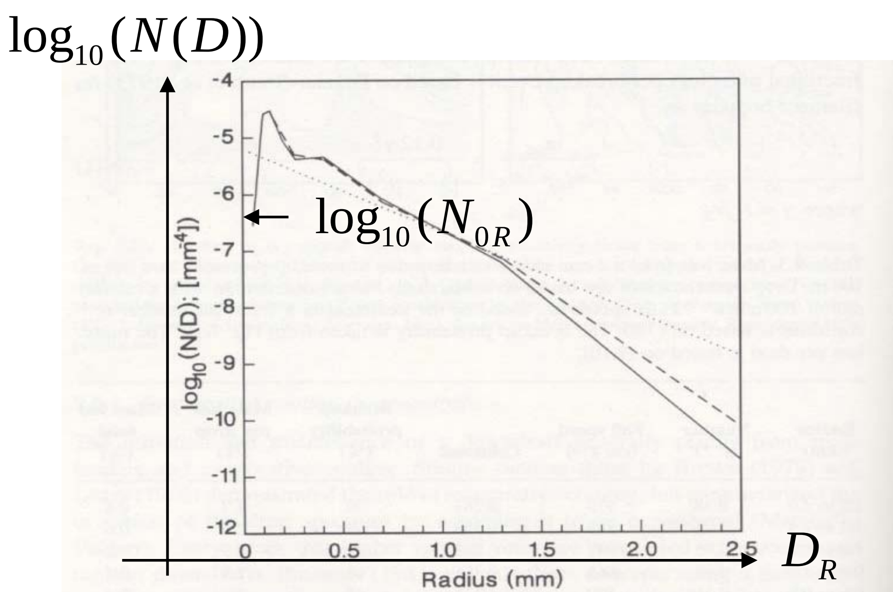


Fig. 9.21. Evolution of a drop spectrum in a subsaturated rainshaft including the effects of coalescence and breakup. The spectrum at the top of the rainshaft is Marshall–Palmer with a rainfall rate of 110 mm h^{-1}. (······), 0 km; (- - -), 1 km; (–), 2 km; R = 110 mm h^{-1}; t = 24 min. From Tzivion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

Single moment scheme

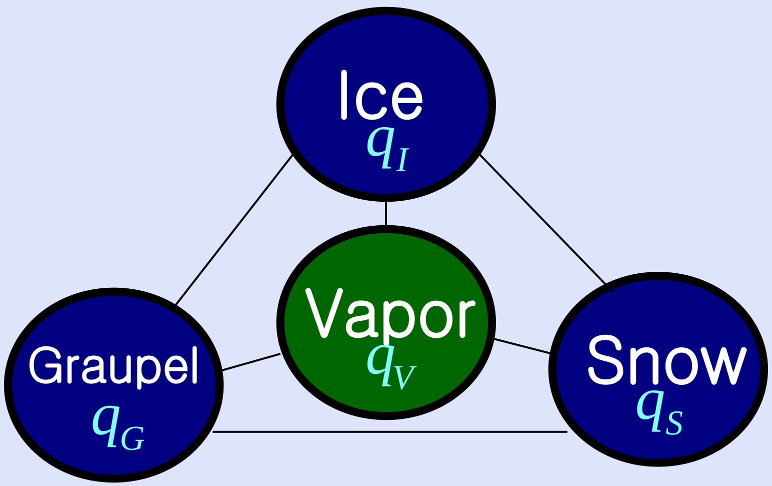
$$dN_{DR} = N_{0R}\exp(-\lambda_R D_R)dD_R$$

Double moment scheme

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

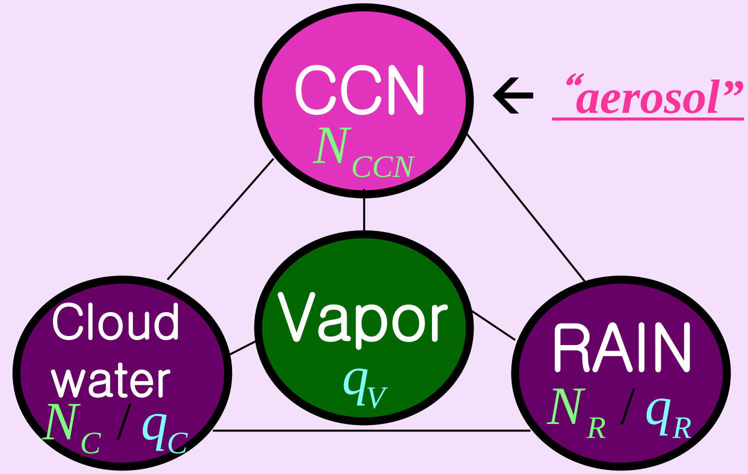
WSM versus WDM

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)

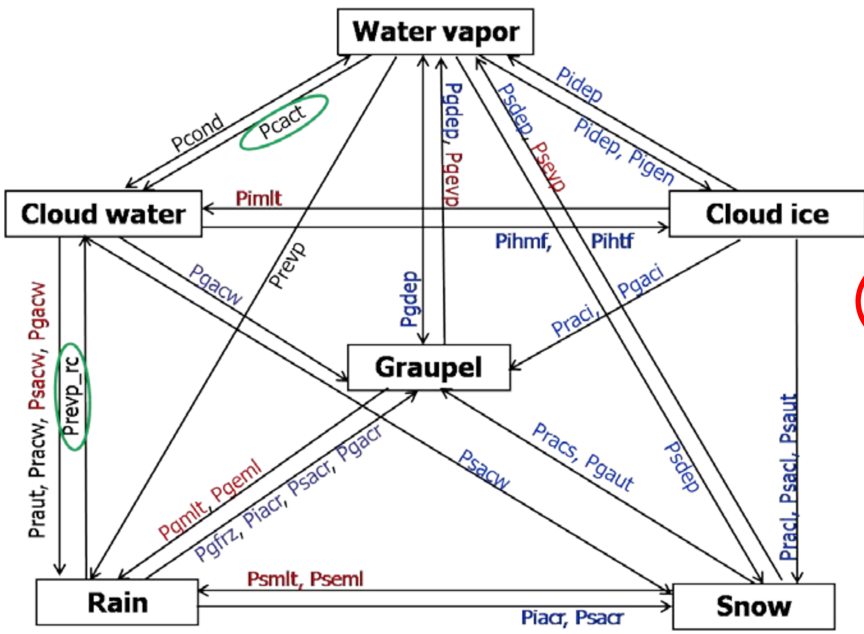
N: Cloud water, Rain, CCN

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

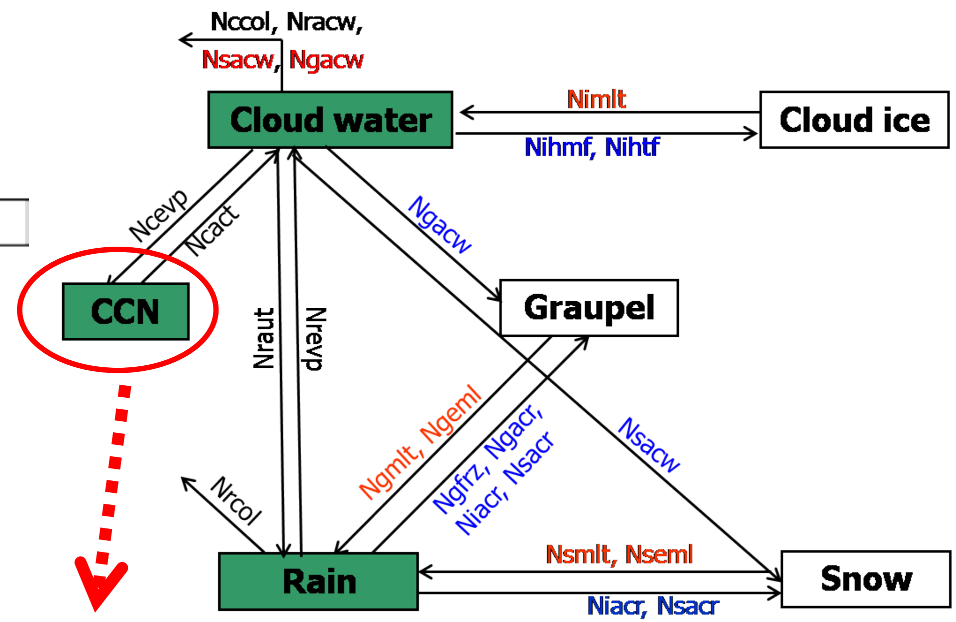
(Lim and Hong, 2010)

Flowchart of the source/sink terms (S_X) in the WDM6

Mixing ratio (q)



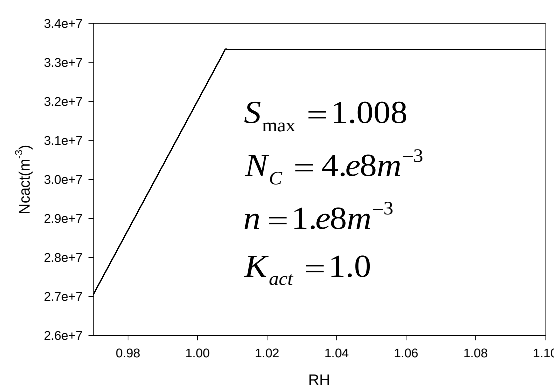
Number concentration (N)



[Source/Sink terms for CCN number concentration]

$$S_{CCN} = -Ncact + Ncevp$$

$$Ncact = \frac{\max\left\{0, (n+N_C)\min\left[1, (S/S_{\max})^{k_{act}}\right] - N_C\right\}}{\Delta t}$$



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)

Subgrid-scale parameterization for physics may be necessary even at 1 km or smaller since the finite model grid cannot resolve all the nature explicitly

```
! 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
```

** Warm rain processes (Hong and Lim 2010)

*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]

$D_R \geq 100$ 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

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

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

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Thank you !