数值模式内物理参数化
个人的看法

潘华陆

The purpose of these lectures

- Give you a flavor of how complicated the phenomena we try to parameterize are and that the only way to make the forecasts better is to build physically based schemes
- Give you a personalized view of the problems and the solutions.

What is the purpose of a parameterization package?

- Because of the limited computer resource, we can not hope to simulate the atmosphere to the molecular level. Some important physics (e.g. heating) are needed to make the model simulation closer to the real atmosphere.
- Convection and turbulence are two mechanism that have been considered necessary for the maintenance of the atmospheric circulation. Over the years we learned that complications of the earth surface and the complexities of the land also are important. In addition, cloud and radiation also make important contributions. While the model resolution increase seemed to give us better handle of the circulation of the large scale, errors made in the parameterization schemes continue to hamper our ability to simulate the atmosphere.

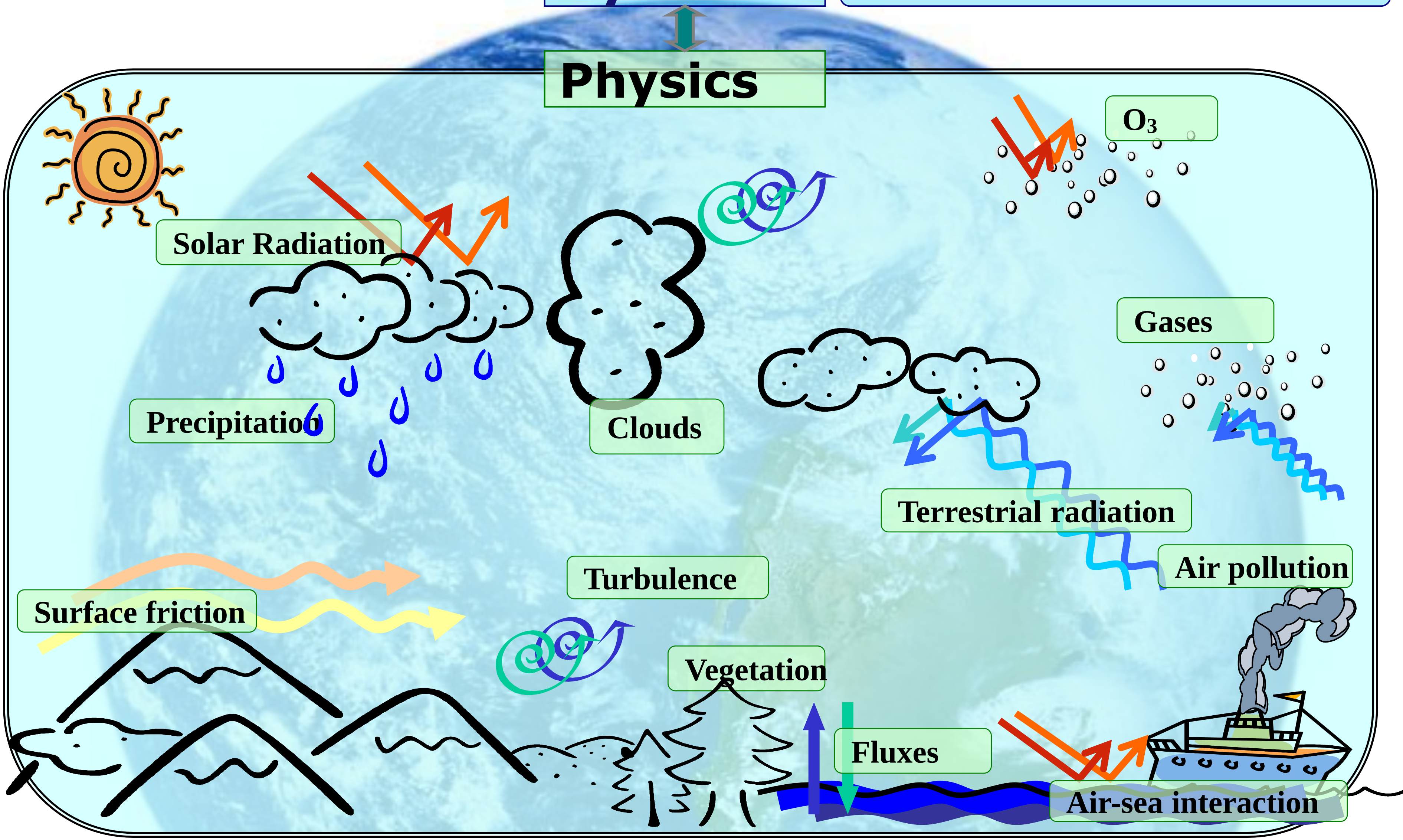
Parameterization for weather and climate models

- Traditionally, climate modelers worry about time mean of the simulation. The idea is that, if we can get the time mean right, we don't need to know how it gets there. Over time, we are learning that how the model gets there is still important. In that, we mean the weather. Climate is still the average of the weather. It is important to get the weather statistics right for climate models as well.
- Weather modelers, on the other hand, worry more about getting the timing and intensity of the weather events and less about the time mean. In that sense, weather modelers traditionally worry less about cloud-radiation feedbacks and more about the evolution of the diurnal boundary layer (as an example).
- We need to learn to do both.

Numerical modeling

Dynamics

PGF, Gravity, Coriolis force, advection



What is physical parameterization?

- Moist and dry turbulent mixing in the atmosphere act mostly in scales smaller than all climate models and most of the weather models.
- Representing sub-grid scale effects of turbulence is commonly referred as physical parameterization.
- Other form of heating (clear sky radiation, microphysics for saturated grid) are not strictly parameterized. But the effect due to fractional clouds need to be parameterized.

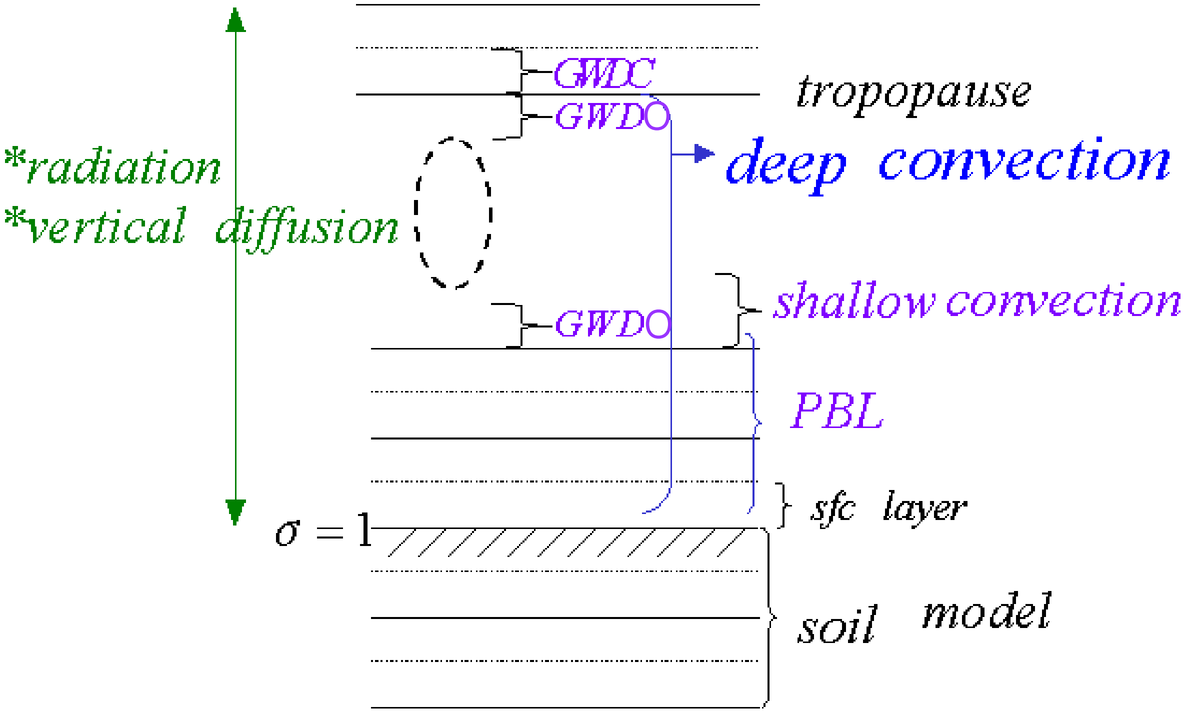
What is parameterization?

- Using a few parameters to represent the effect.
- Selecting and tuning of the parameters.
- The end?
- The challenge is to build a simple set of equations (the scheme) to represent the sub-grid scale phenomenon as closely as we can.
- The better the scheme, the fewer the tunable parameters should matter.
- Convergence issue: when grid size is small enough, can the scheme behave properly?

Concept

1) concept

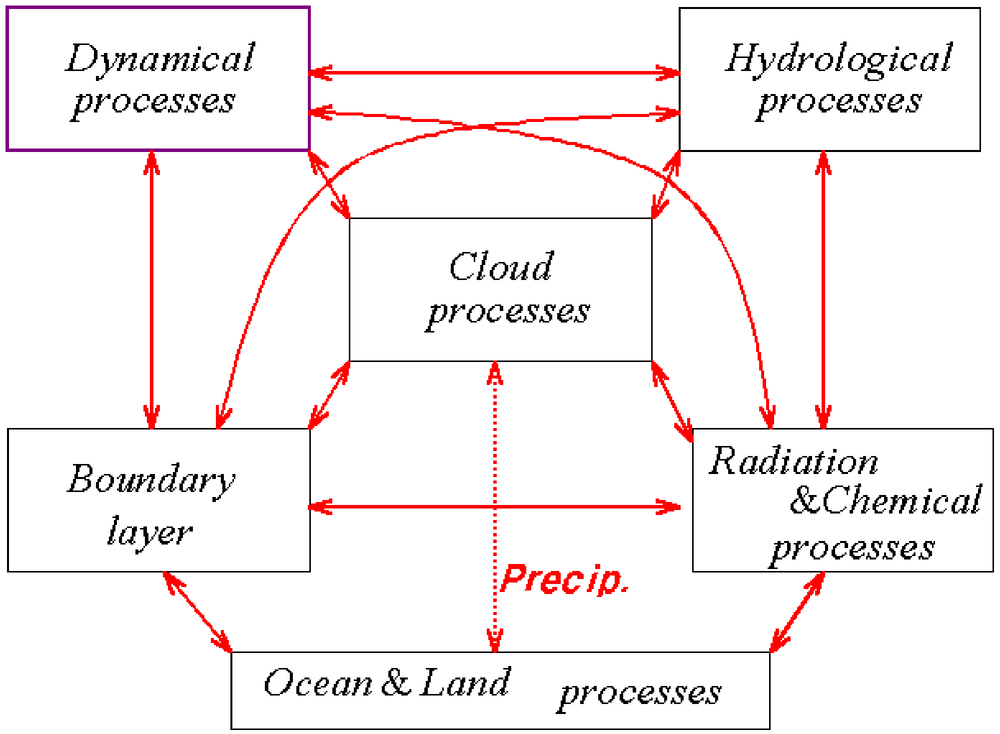
$$\frac{d\ln\theta}{dt} = \frac{H}{c_p T}\ , \qquad \frac{dq}{dt} = S$$



Engineer vs physicist

- Engineers have to solve real life problems which are complicated and not-well-understood. Practical solutions are necessary evils. Use simple equations with a few parameters to adjust to the current solution is a practical way to move forward.
- In modeling the atmosphere, using a few local observations to form a parameterization tends not to work because the same turbulence acts differently over different regions: land-ocean, tropics-midlatitudes, etc. A new consensus is forming that we need to formulate the schemes based on better understanding of the phenomenon. So physical understanding becomes more important in formulating the scheme.

Schematic configuration



⋆ 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.

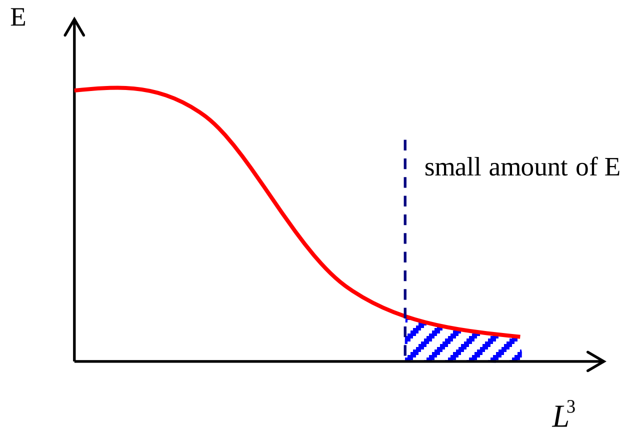
* Parameterization

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

* 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)

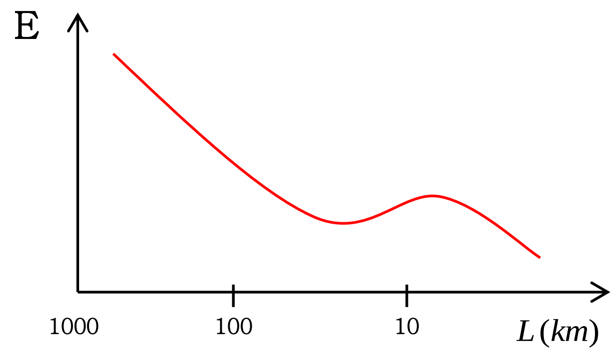


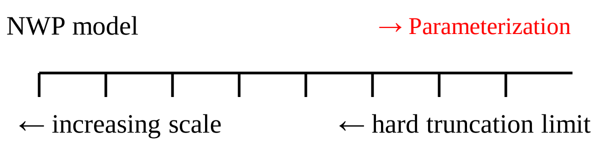
- $\Delta x \to 0,$ the energy dissipation takes place by molecular viscosity (smallest grid size idealized situation)

• 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."

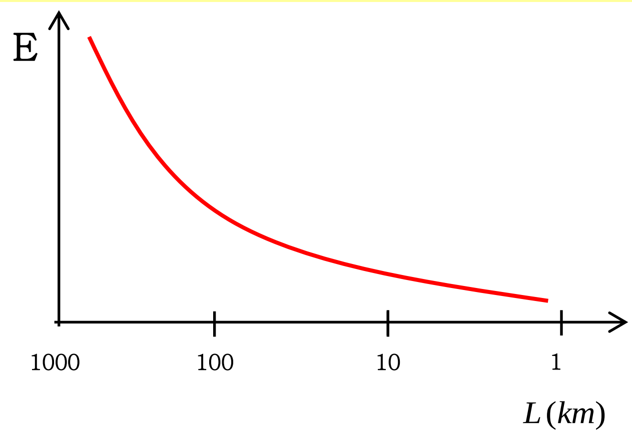
※ Parameterization that are only somewhat smaller than the smallest resolved scales.
If the real atmosphere was like that,





Where truncation limit ; spectral gap

Unfortunately, there is no spectral gap



2) Subgrid scale process & Reynolds averaging

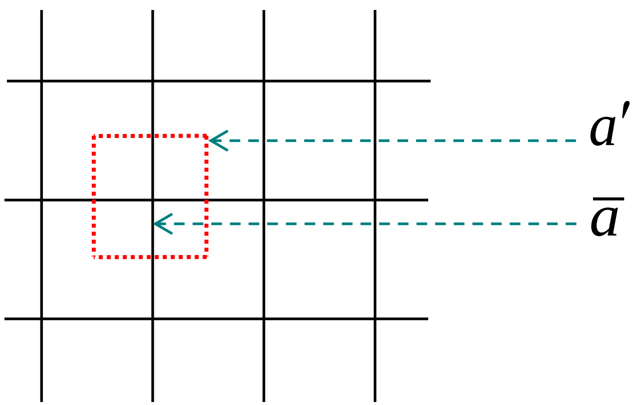
Consider prognostic water vapor equation

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

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 : $\overline{q'} = 0, \ \ \overline{u'\overline{q}} = 0, \ \ \overline{\overline{u}\,\overline{q}} = \overline{u}\,\overline{q}$

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,

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

$$-\rho\overline{w'w'q'} = K'\frac{\partial\overline{\rho w'q'}}{\partial z}$$: 2nd order closure (HW: 4-1)

Parameterizations are personal

- Since we may have different understandings of how the phenomenon we are parameterizing behaves, parameterization schemes are personal. I will primarily talk about how I view the various schemes. I hope my example will give you an idea how you might make a better scheme!!

A classical case

- The surface layer parameterization using the similarity profile function is nearly universally adopted.
- Based on scaling argument, dimensionless profile functions can be formulated using local measurements.
- Problem is the applicability to situations not measured: under strong wind conditions in the typhoon environment.
- Boundary layer researcher have, for years, try to apply the same method to the boundary layer with little success.

Surface layer

1) Bulk method

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

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

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

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

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

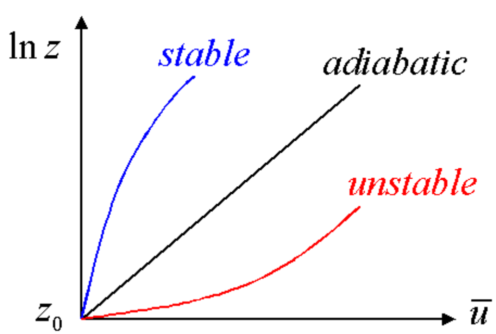
where $k_z = fn(R_i, S)$

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)

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

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

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

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

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

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

Applying the surface layer formulation to the interface

- Over ocean:
  - $H = \rho\, c_p\, C_h\, U\, (T_s - T_a)$
  - $LE = \rho\, L\, C_h\, U\, (q_s(T_s) - q_a)$
- Over land, the situation is much more complicated. The early effort to parameterize the heat flux is to apply the same formula as for ocean with an additional parameter (β) for evaporation
  - $LE = \beta\, \rho\, L\, C_h\, U\, (q_s(T_s) - q_a)$

Heat exchange over land

- Since the heat storage for land surface is much smaller than for ocean, the incoming and outgoing heat fluxes are nearly balanced. So we often use a surface heat budget to calculate the ‘skin’ temperature Ts.
- The problem with using a single parameter β to handle evaporation over many different situations proved impossible.

Land heat flux

- In fact, evaporation for most of the land area comes in the form of transpiration from vegetations. Trees are living entities and 'know' when to conserve water. So the newer generation of surface models focuses on the transpiration from different vegetations. This is a more reasonable way to deal with the problem.
- Transpiration also can draw water from a deeper layer of the soil than surface evaporation. So soil moisture initiation becomes an issue.
- Most schemes handle the transpiration in the form of a resistance – stomatal resistance. Lots of our lack of understanding of the way leaves transpire can be hidden in this resistance formula, a parameterization.
- The only way to make advances is to obtain better understanding of the biology of the transpiration process.
- Convergence issue. The approach I follow uses an additional quantity called potential evaporation as the upper limit of transpiration. This is the maximum evaporation that can be attained given the same conditions other than the fact that the surface is saturated. This will result in a lower skin temperature and the resulting potential evaporation should be similar to the so-called pan-evaporation. In this approach, we specify the actual evaporation based on a resistance approach modifying the potential evaporation. The surface energy balance (budget) is then used to determine the actual skin temperature and the rest of the heat fluxes.

Land heat fluxes - complications

- In a given grid area, we all know how complicated the underlying surface can be : tarmac road surfaces, parking lots, buildings, slight slopes of the surface.
- Soil moisture can have sources from fallen precipitation (runoff problem), underground reservoir, and snow melt. It can have sinks from evaporation and runoffs.
- Differential heating can also lead to enhanced turbulence.

Heat flux over oceans

- It is not that simple as well. The surface roughness parameter is used in the similarity profile functions. Over land, we have some idea on how to measure this – larger for trees than for grasses, etc. For ocean, the waves bring in complications. In strong wind situation, waves can break and sea spray can form. How do we deal with that?
- The classical way is the use of the Charnock formula to deduce roughness length based on the frictional velocity. This works for wind stress but roughness length for heat and moisture is quite different. TOGA COARE data have been used to derive some of these quantities over ocean. Most of the observations are for wind speeds lower than 20 m/s. We are facing the problem of applying the method to hurricane forecast these days.

How much complication is necessary

- We can certainly imagine modeling at the leave and pebble resolution. Is that necessary? The question is the importance of the resolution to the modeling of the effect on the atmosphere.
- With an additional dimension in the vertical, we must consider mixing the in-homogeneities through the boundary layer.
- The increase in vertical resolution is still not compatible with the increases in horizontal resolution. How accurate can and do we decide on the depth of the PBL?
- Given the uncertainties in the parameterization of the other physical effects, we should try to match the level of complications amongst all schemes.
- Not knowing the cloud cover accurately probably has as big an impact to the surface flux as all the complications in the surface scheme.

Why is the similarity profile function successful?

- I believe part of the reason of the success of the similarity profile function approach is the spatial and temporal scale of the phenomenon. Because the mixing in the surface layer happened within minutes, it is fast compared to the weather which are dominated by the synoptic scale disturbances (in mid-latitudes and in the tropics). As the phenomenon life time is longer, the details of the phenomenon becomes more important. The life cycle of a cumulus cluster is on the order of 10-20 hours and the details of the interaction of the cloud to the large-scale becomes important.

Boundary layer and turbulence parameterization

- The traditional approach is to apply the Reynolds' averaging method to the flux terms and select an order to truncate the expansion. So for first order closure, we are reduced to a specification of the coefficient of diffusivity.
- Another popular approach for GCM is the mix-layered approach. Recognize that over large portion of the ocean, the boundary layer is nearly well-mixed, a single column can be used to represent the entire PBL. This is a zero-order closure.

2.4 Vertical diffusion (PBL)

2.1 Local vertical diffusion

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

Local Richardson number

2.2 Nonlocal PBL

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

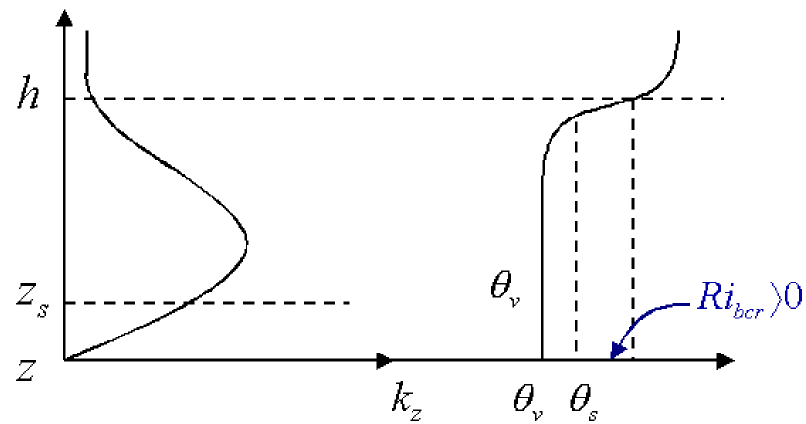
$$k_{zm} = k w_s z\left(1 - \frac{z}{h}\right)^p$$

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

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

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

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



2.3 TKE (Turbulent Kinetic Energy)

TKE eqn.

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

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

(Mellor & Yamada, 1982)

PBL development over land

- The PBl over land has a typical diurnal behavior. During the day, as the surface warms up, boundary layer depth increases. When heating reaches maximum, PBL growth also ceases. Modeling this behavior using coarse vertical resolution models has turned out difficult using the traditional K closure method.
- The problem is that the determine of the coefficient of diffusivity K is typically done with the local Richardson number. During fast growth of the PBL, most models fail to capture the growth.
- Large-eddy simulation studies using high resolution to explicitly resolve the larger eddies showed that most of the mixing in the turbulent PBL is done by the largest scale eddies (eddies with the scale of the PBL depth).

Nonlocal diffusion schemes

- By defining the coefficient of diffusivity using the scale length of the largest turbulence (PBL height), this scheme is able to simulate the growth of the boundary layer over land.
- Together with a reasonable land surface scheme that gives realistic sensible and latent heat flux, the combined scheme works well with a reasonable number of vertical layers.

Cloud topped PBL

- Over eastern ocean region, the subsidence from the subtropical High region limits the growth of the PBL. With the ocean underneath, the PBL forms a nearly well-mixed layer. With constant theta and nearly constant specific humidity, the relative humidity will increase with height in the PBL. Cloud can form easily. Cloud-topped PBL present another challenge.
- If cloud is passive, we can simply diagnose the cloud and be done with it. Cloud-top radiation cooling can lead to instability and additional source for turbulence.

Heat flux

MRF PBL

$$\overline{w'\theta'} = -K_h^{surf}\left(\frac{\partial\theta}{\partial z} - \gamma_h\right)$$

$$K_m^{surf} = \kappa w_s z\left(1 - \frac{z}{h}\right)^2$$

$$K_h^{surf} = \mathrm{Pr}^{-1} K_m^{surf}$$

$$\gamma_h = 6.5\frac{\overline{(w'\theta')_0}}{w_s h}$$

Revised model

$$\overline{w'\theta'} = -\left(K_h^{surf} + K_h^{Sc}\right)\frac{\partial\overline{\theta}}{\partial z} + K_h^{surf}\gamma_h$$

$$K_h^{Sc} = 0.85\kappa V_{Sc}\frac{\left(z - z_b\right)^2}{h_b - z_b}\left(1 - \frac{z - z_b}{h_b - z_b}\right)^{1/2}$$

$$V_{Sc}^3 = \frac{g}{\theta_0}(h_b - z_b)\Delta R/(\rho c_p)$$

(Simplified after Lock et al., 2000)

$$-\overline{(w'\theta_v')}_{h_b} = c\frac{\Delta R}{\rho c_p}$$

where $c=0.2$

$if\ c_p\Delta\theta_e / L\Delta q_t > 0.7,$ $C=1.0$

(CTEI condition)

Diagnosing the PBL

- Given that we do not have routinely available PBL depth estimates, the diagnosis is primarily estimates. For instances, for dry soil and clear sky (desert), the PBL depth can reach 4-5 km.
- For eastern ocean, the depth of the PBL is typically 1 km. The depth is a balance of the subsidence from the subtropical high and the surface heating. Cloud becomes an issue.
- Convection responds to the diurnal heating also so that requires connections to the convection scheme.

Counter gradient effect

- Diffusion schemes works on vertical gradient. When a layer is well-mixed, vertical gradient of potential temperature should be very small. So how do we complete the mixing? Hence the birth of the so-called counter gradient effect parameterization.
- More physical is perhaps the idea of an additional parameterization of the large-eddy directly: the mass-flux approach. This approach has been proposed recently and seems to be a more physical way to model the counter-gradient term.

Compatibility issues

- Given how we really don't have a good handle of the behavior of the boundary layer except on the first order (we hope to get the PBL height correct to the 10% level), how is a sophisticated land surface scheme going to help? We should first strive to get the PBL height evolution nearly correct for the entire earth. I am not sure if we have done that yet.
- Schemes need to match in levels of sophistication and accuracy.

Deep convection

- Perhaps the most difficult and the most famous of all parameterizations for the atmosphere is the deep convection scheme.
- Early schemes such as the convective adjustment scheme (forcing a moist adiabat) to the Kuo scheme (adjusting the profile to the moist adiabat over some time period) did not prove successful.
- The Arakawa-Schubert scheme proposed in 1974 stands out as capturing the essence of the cumulus effect and not just some mechanistic adjustment to a moist adiabat.

The Arakawa-Schubert scheme

- The key essence of the AS scheme is the idea that updraft does not really warm the atmospheric column. The updraft simply maintains a balance of latent heat release and adiabatic cooling of the cloud parcels. The actual warming comes from the compensating subsidence due to the updraft. The famous ‘subsidence warming and drying’ is completely different from the other schemes that tries to adjust to a moist adiabat.
- The adjustment to the moist adiabat is actually calculated in a round about way in the closure of the scheme : quasi-equilibrium assumption. This assumption seeks to reduce the cloud work function by the clouds. Cloud work function is very closely linked to the buoyancy and the so-called CAPE. So reducing/removing CAPE is effectively a move toward a moist adiabat.

Some criticism of the AS scheme

(I say this with all the respect for Prof. Arakawa)

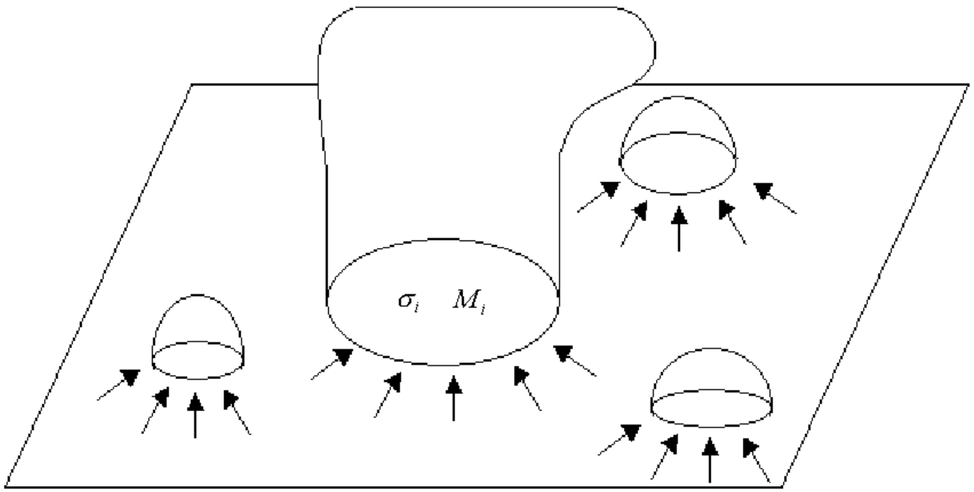
- The AS scheme does not have a trigger to prevent convection from breaking out. As long as there is conditional instability and the cloud work function exceeds some climatological mean for the cloud depth, convection will be invoked. Since most of the tropics is conditionally unstable, the AS scheme will be active most of the time. This scheme was originally designed for a GCM where timing of the onset of convection is not critical. As long as CAPE is removed over a period of time, it was considered acceptable.
- In reality, the below cloud condition plays an important role on the timing of the convection onset. For convection over land, it is apparent that sub-cloud layer structure is quite important to decide if convection will break out or not.
- Even over most of the tropical oceans, conditional instability in an air column does not guarantee onset of convection. Some convergence activities are usually required before convection breaks out.
- While the AS scheme captured some important physics, we found that we can augment the scheme with additional physics to make the scheme work for weather forecast as well. In the end, improving the weather effect due to convection may also help the climate modeling.
- Another problem with the AS scheme is the basic assumption that the area of the updraft is much smaller than the area of the model grid point. This is a good assumption until the model grid size goes below 10 km.

4.4. Arakawa-Schubert (1972)

- 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

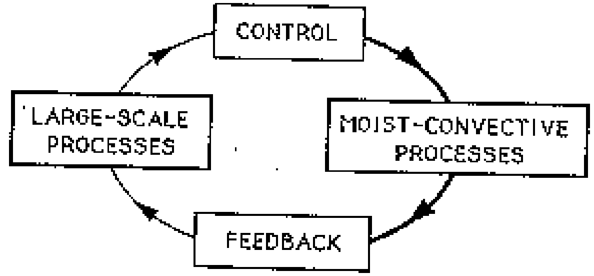
σ_i : fractional area covered by ith cloud

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

environment

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

: net mass flux/unit large-scale horizontal area



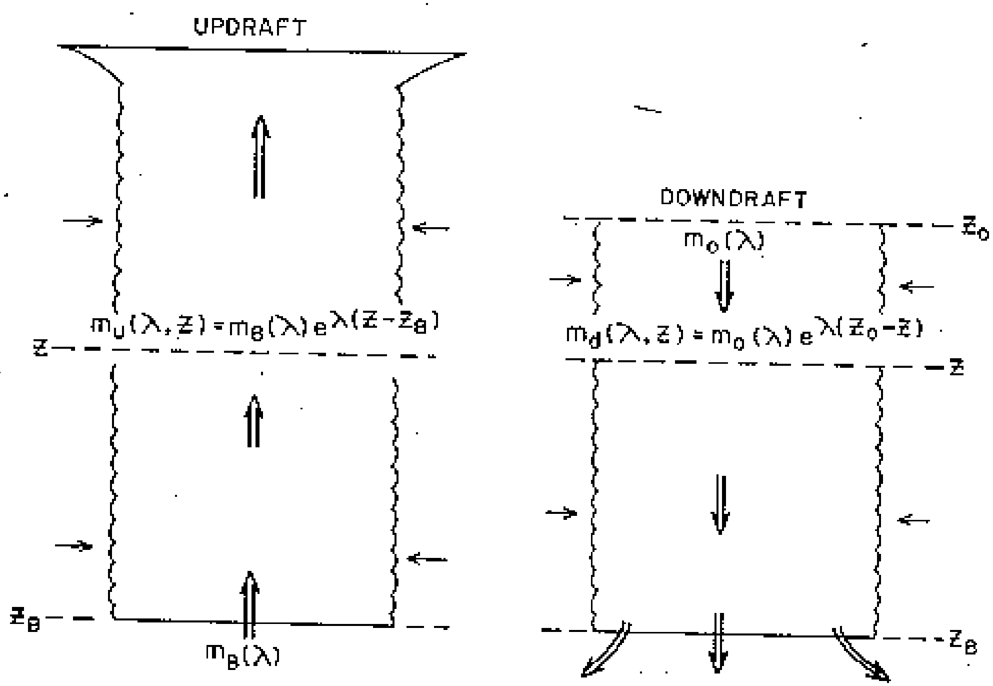


FIG. 4.10. Model for updraft and downdraft of cloud type λ (from Johnson 1976).

Large-scale flux across grid box

Exchange of S between environment and clouds

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

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

$S_i : C_pT + gz$ 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 $\sigma_c \ll 1$, $\bar{s} \approx \tilde{s}$

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

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

Adiabatic warming due to hypothetical subsidence between the clouds

Detraining clouds

detrainiment, entrainment

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

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

- Spectral cloud ensemble

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

subensemble

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

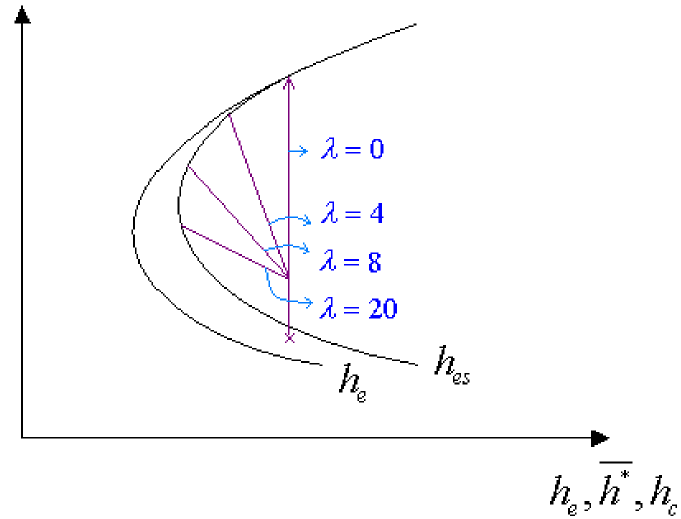
$$= \int_0^{\lambda_{\max}} \underline{m_B(\lambda)}\eta(z,\lambda)d\lambda$$

Mass flux at cloud base

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

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

$$\eta(z,\lambda) = e^{\lambda(z-zB)}$$; 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

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

Large-scale forcing
>0 : destabilized

Adjustment
<0 : stabilization

Kernel : Cloud scheme kinetic energy

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

=> m_B

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

Cumulus momentum mixing

- The original AS paper never mentioned momentum mixing. The problem has always been that momentum is not conserved for the updraft parcel. As the parcel expands, work is done. I later found out that the RAS always coded the momentum mixing (without consideration of the pressure gradient effect). It turns out that the cumulus momentum mixing has tremendous importance to the typhoon modeling. Without some cumulus momentum mixing, typhoon formation in the model behaves very strangely.
- Using LES study, people came up with a way to parameterize the pressure gradient effect. However, we use the large-scale gradient to model this and not sure if that is the right physics. As a result, we have to tune the strength of the momentum mixing in order to get the best performance for typhoon forecasts.

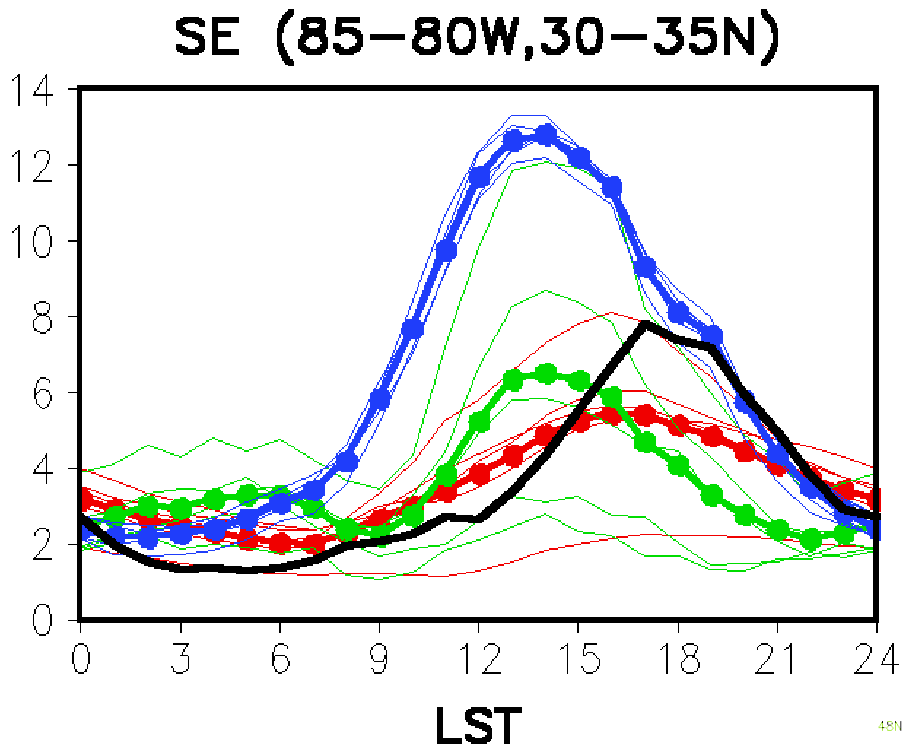
Can we make it better?

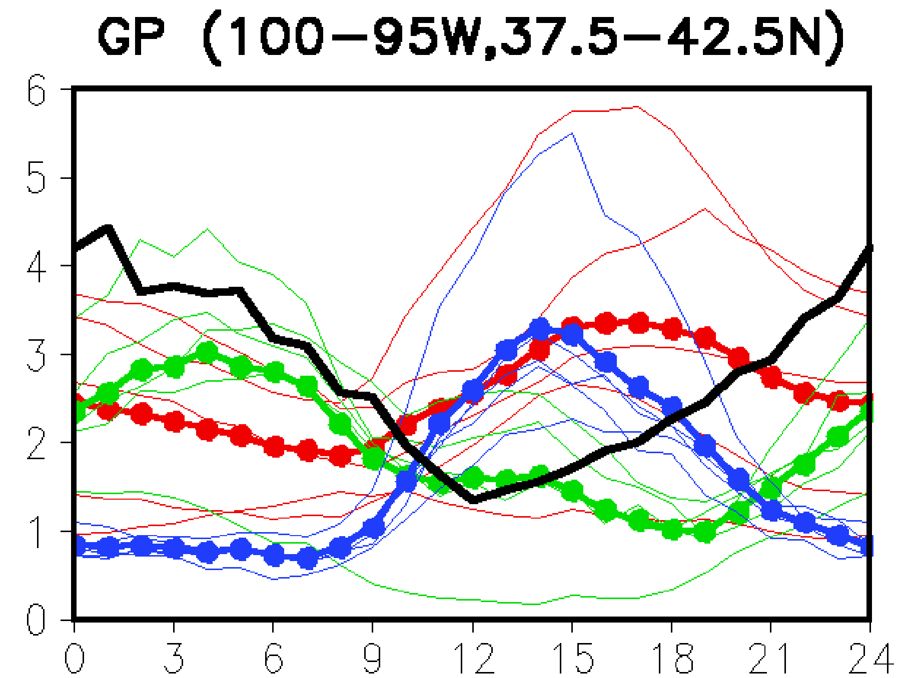
- Trigger. The current trigger is extremely simple.
- Cloud top determination. There is an assumption made about the moist adiabat using the environmental variables.
- Closure adjustment.
- Momentum mixing.
- Downdraft interaction with ground.
- What happens when the grid size gets smaller and smaller?

Nocturnal precipitation maximum over the great plains

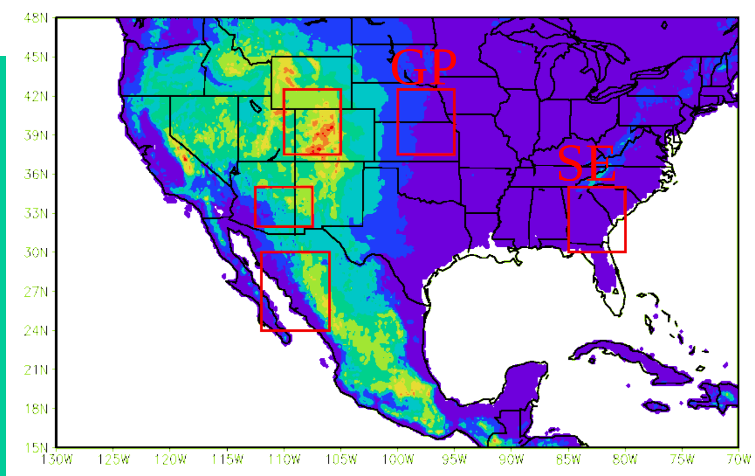
- Lee et al (2008) examined the diurnal convection signals over North America for several models and found the GFS to exhibit nocturnal convection signal similar to observations.
- Five AMIP runs for northern summer were made for each model.
- The GFS convection signal over the Southeast US is too early compared to observations.

Diurnal Cycle of Rainfall – Ensemble Mean and Spread





Obs(HPD)
GFDL
NCEP
NASA
(spread) (ensemble mean)

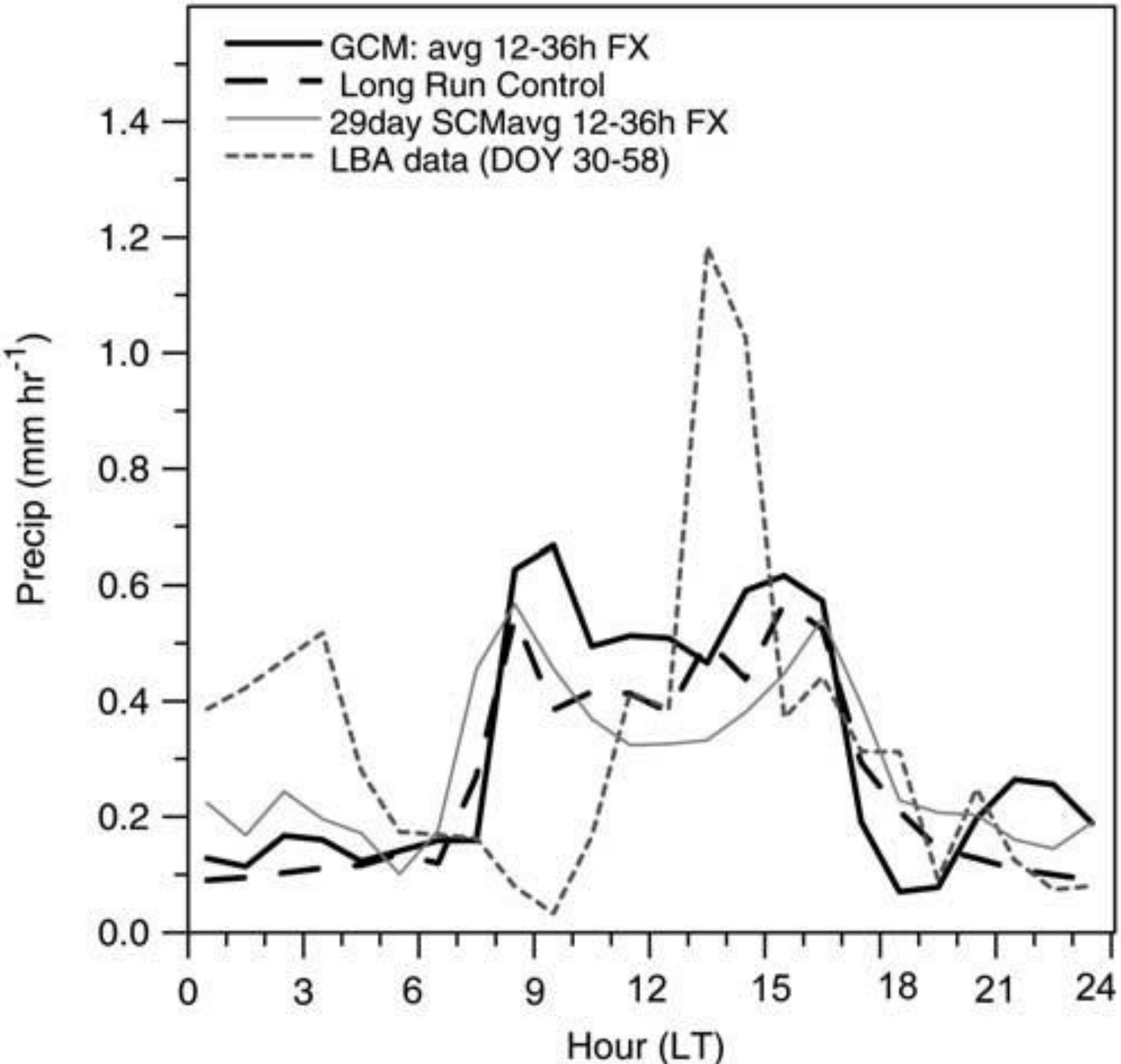


Primary reason for the nocturnal signal

- Trigger function was able to choose parcels above the inversion to initiate convection.
- Weakening of the nocturnal jet and turbulent mixing actually reduces convection during the day.

Diurnal signal over Rondonia, Brazil

- A. Betts compared diurnal precipitation signal over Amazon region for February and found a significant afternoon maximum while some of the models do not produce it.
- Three-hourly GFS forecasts (to 48 hours) for the day 30-58 period for 2010 for the similar area were averaged to examine the diurnal signal.
- The GFS forecasts do exhibit early afternoon maximum.

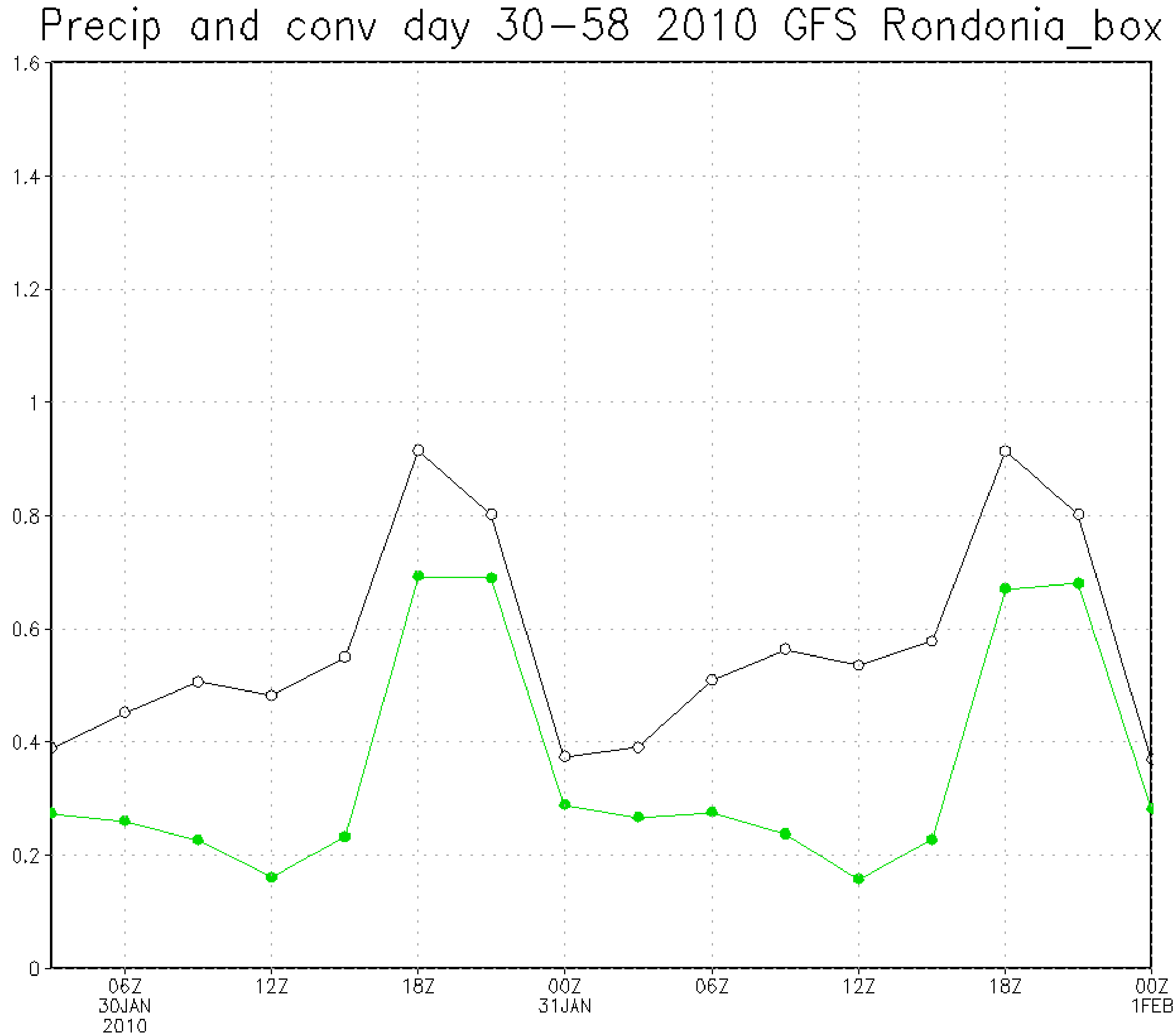

GCM: avg 12-36h FX
Long Run Control
29day SCMavg 12-36h FX
LBA data (DOY 30-58)
Precip (mm hr-1)
Hour (LT)
0.0
0.2
0.4
0.6
0.8
1.0
1.2
1.4
0
3
6
9
12
15
18
21
24

Box domain

5S-15S

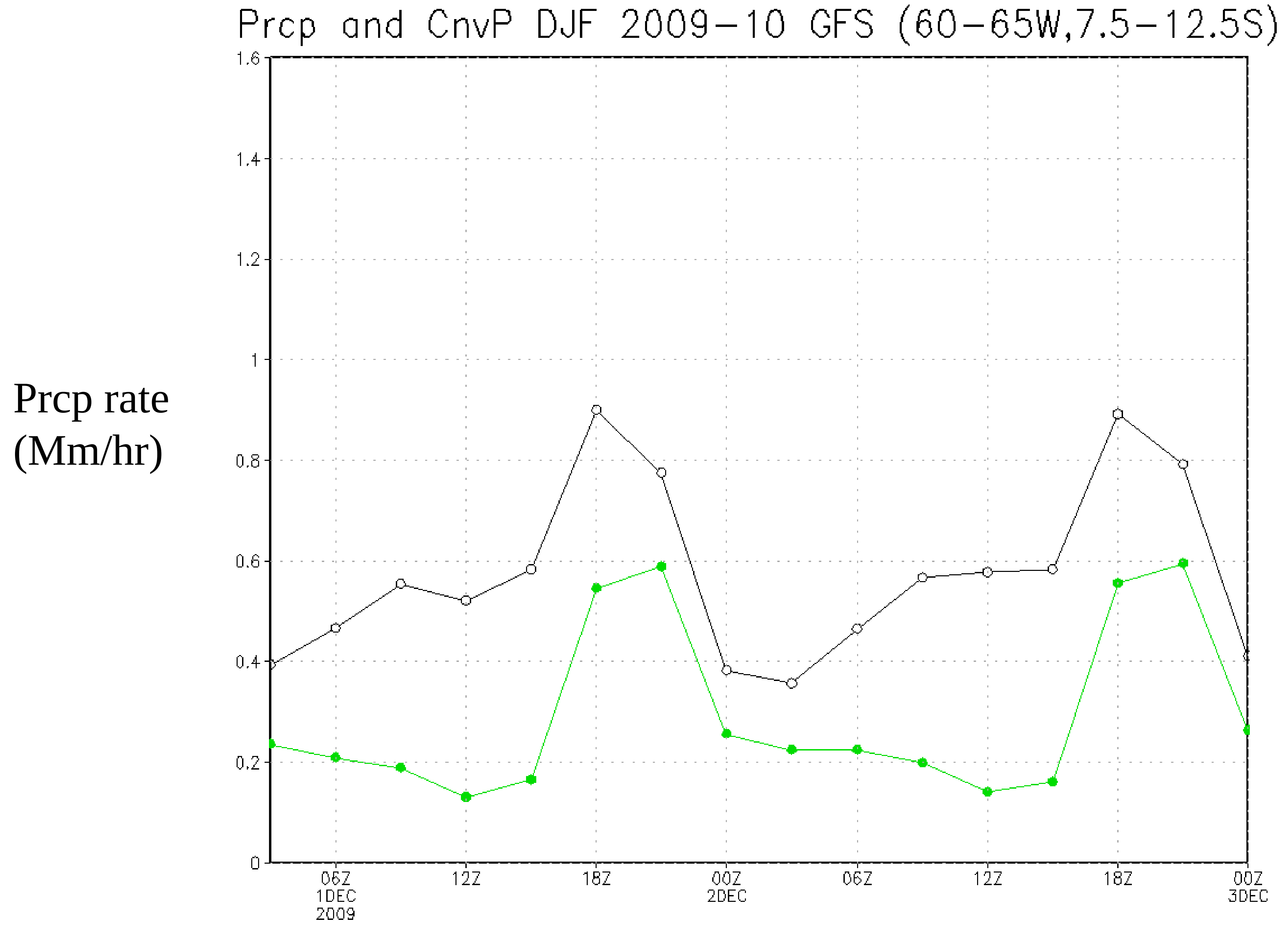
65W-55W

Mm/hr unit



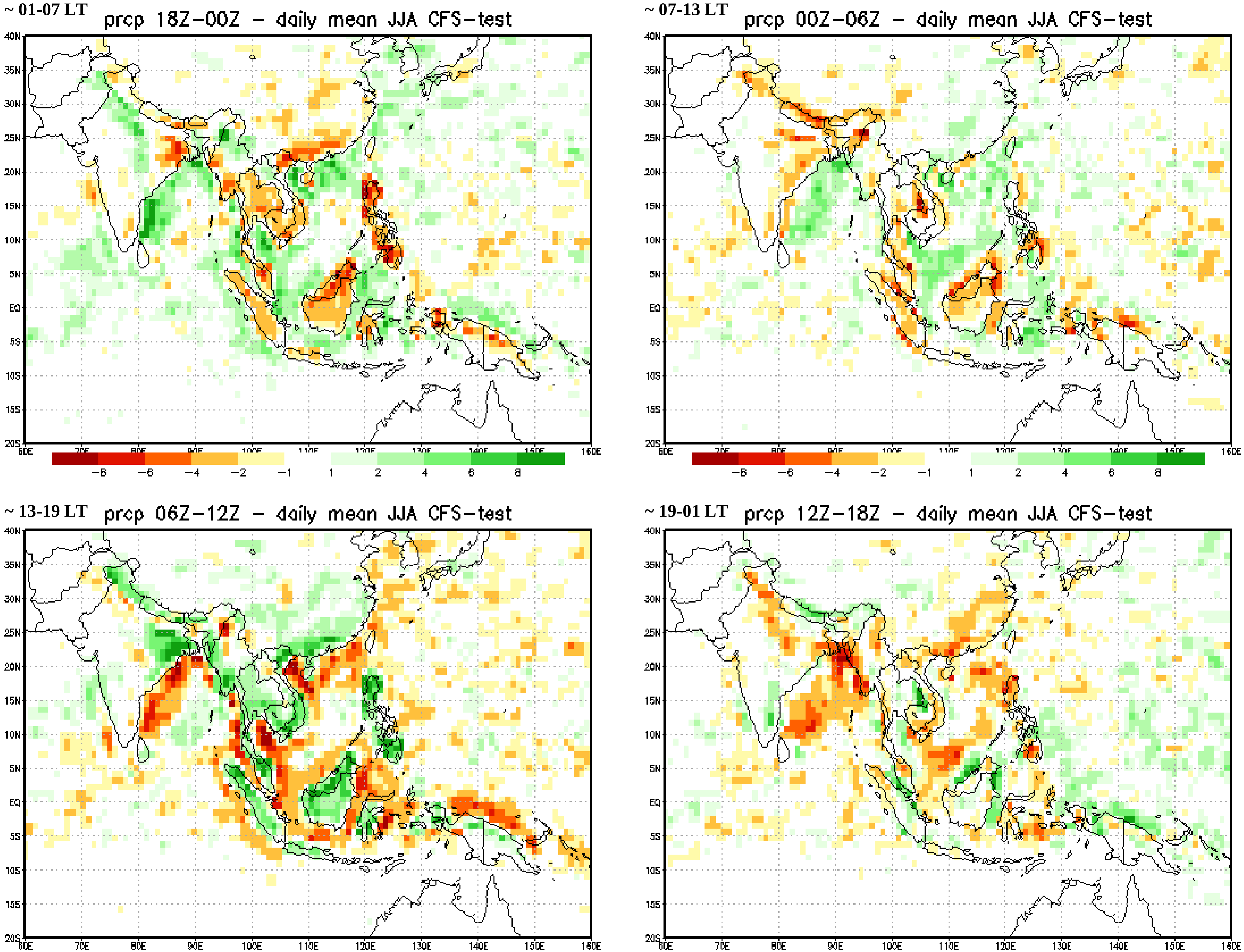
L 0 0 1 2 0 0 1

T 2 8 4 0 2 8 4


Prcp and CnvP DJF 2009-10 GFS (60-65W,7.5-12.5S)
Prcp rate (Mm/hr)
1.6
1.4
1.2
1
0.8
0.6
0.4
0.2
0
06Z 1DEC 2009
12Z
18Z
00Z 2DEC
06Z
12Z
18Z
00Z 3DEC

Maritime continent convection

- J. Slingo has examined several GCMs for the diurnal convection over the maritime continent and found the models missing the signal.
- A four-month coupled run using the CFS v.2 starting 1 May 1981 was made. Four-times daily precip amount was archived and averaged for the Jun-Aug period. Daily mean was subtracted at each grid point to obtain the anomaly from the daily mean (mm/day).
- An afternoon maximum and early morning minimum over land with a reversal of signal off the coast.


~ 01-07 LT
prcp 18Z-00Z - daily mean JJA CFS-test
~ 07-13 LT
prcp 00Z-06Z - daily mean JJA CFS-test
−8
−6
−4
−2
−1
1
2
4
6
8
~ 13-19 LT
prcp 06Z-12Z - daily mean JJA CFS-test
~ 19-01 LT
prcp 12Z-18Z - daily mean JJA CFS-test

Problem with the conventional mass flux schemes

- Most of the mass-flux schemes are based on the original Arakawa-Schubert (1974) assumption that the updraft area is much smaller than the model grid size. This assumption begins to break down when the grid sizes become smaller than 10 km. Since the assumption is fundamental to the parameterization scheme, we are not justified to continue to use such schemes.

Should we go directly to the explicit schemes?

- While the commonly used mass flux schemes should be avoided when σ (the ratio of the updraft area to the grid area) is no longer small, the use of the explicit microphysics scheme is still problematic since the vertical motion in models of grid sizes from 500m to 10 km may not be large enough to smoothly create moist adiabat for the entire grid point. This can and do leads to the so-called grid-point storm when computational instability can lead to excess rainfall and much lower surface pressure for hurricanes.

So when can we stop parameterizing moist convection?

- When $\sigma < .1$, we can safely use the conventional mass flux schemes.
- When $\sigma > .9$, we can most likely use explicit microphysics directly and skip the parameterization.
- When $.1 < \sigma < .9$, we are in no-man's land. We need to parameterize the convection but we can not use the conventional scheme.

Proposing a modification of the A-S scheme

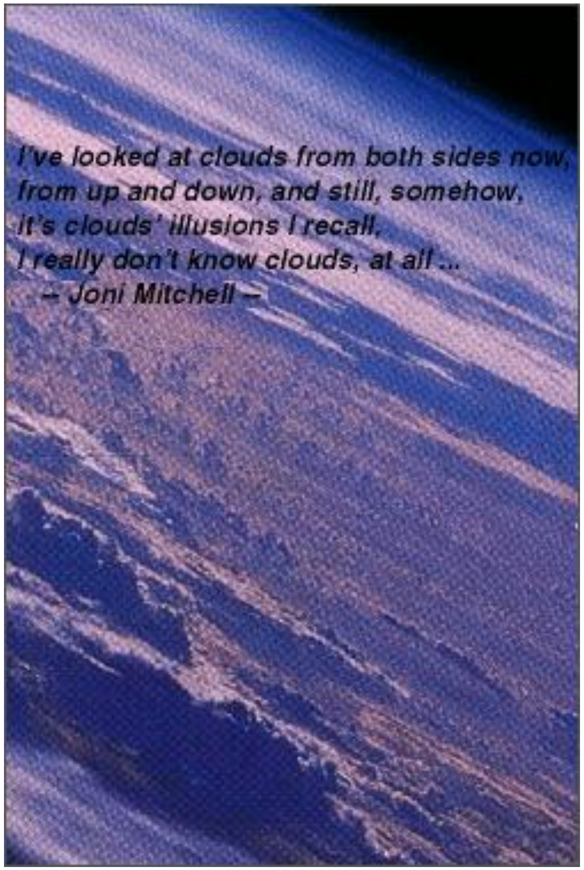
- We have re-derived the A-S scheme removing the assumption that the updraft area be small.
- In doing so, we have arrived at a scheme that can be easily implemented.
- It is similar to the conventional scheme when the updraft area is small.
- Its effect diminishes when the updraft area approaches the grid area (convergence issue).
- It explicitly takes the updraft area into consideration.

Which is the dog and which is the tail

- While accurate calculation of radiation is quite possible, our knowledge and ability to specify cloud fraction is far inferior.
- Observation and modeling of cloud fraction and cloud condensate amount are still quite poor. Yet the effect of cloud on the global climate is not in dispute. We must make better progress with the cloud prediction soon.
- We might say that the tail (clouds) is wagging the dog (radiation) but that is the cruel fact of life.
- People who works on radiation expects the cloud microphysicist to give them the answer. Most microphysics schemes actually only work when the column is saturated so there is no cloud fraction issues (as far as they are concerned).

Cloud fractions

- There are some work in diagnosing the cloud fractions from LES simulations. There are also some work in using a Probability Density Function (PDF) approach to specify cloud fraction.
- As far as I am concerned, this area is wide open for creative minds.


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 --