

Day 3:

1. Parameterization of Convective momentum transport
2. Convective microphysics

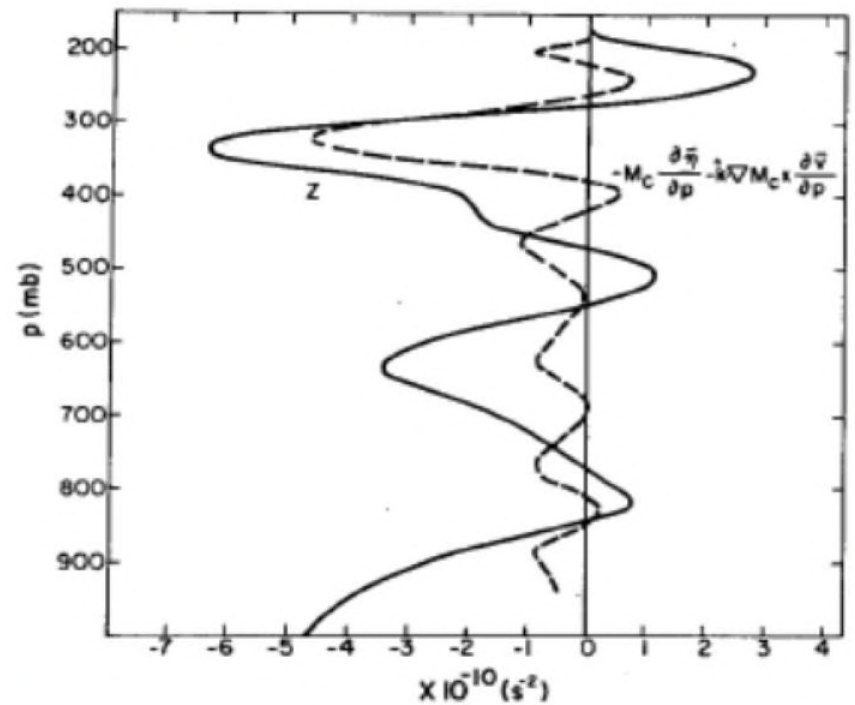
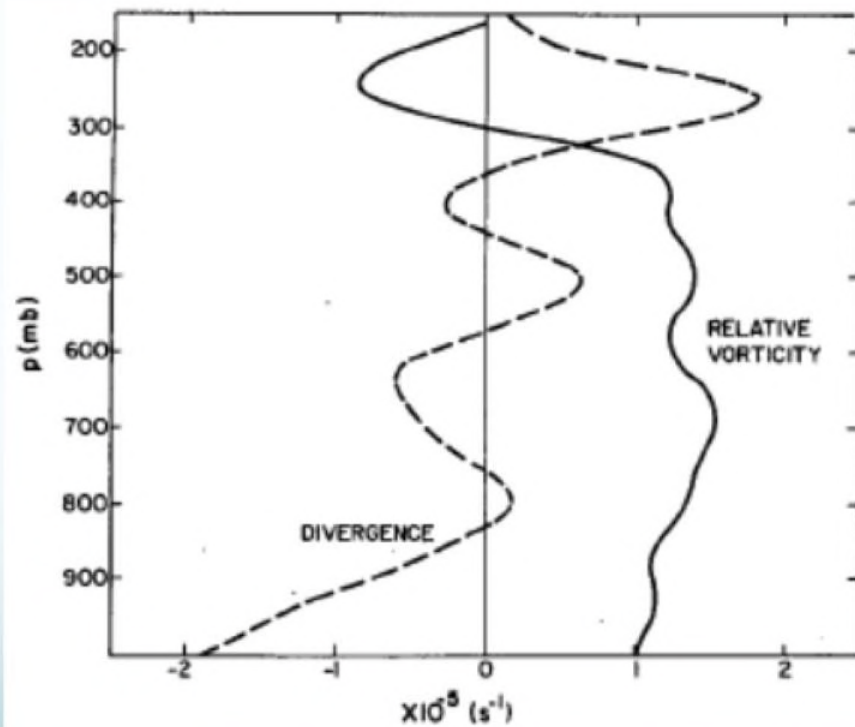
Parameterization of Convective momentum transport

Convection transports not only heat and moisture, but also momentum. In numerical models, often divergence and vorticity were used as predictive variables. Thus in the early days (1970' s and early 80' s), the dynamic effect of convection was studied through vorticity parameterization.

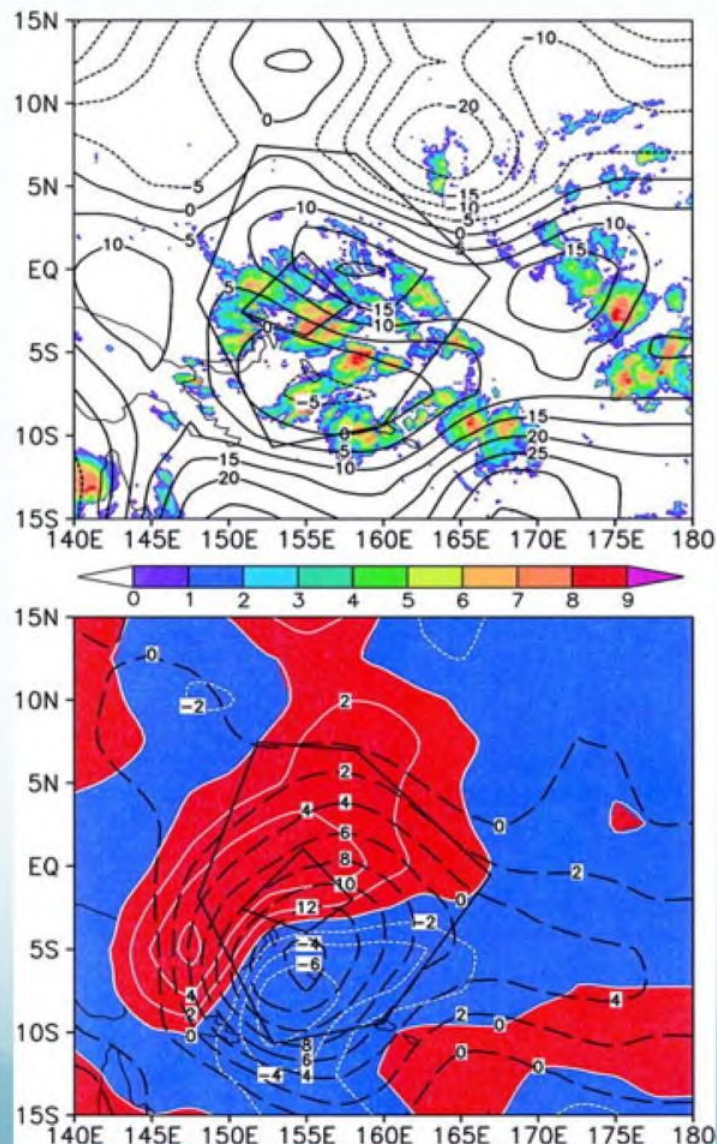
$$\frac{\partial \eta}{\partial t} + \mathbf{v} \cdot \nabla \eta + \eta \nabla \cdot \mathbf{v} + \omega \frac{\partial \eta}{\partial p} + \mathbf{k} \cdot \nabla \omega \times \frac{\partial \mathbf{v}}{\partial p} = 0$$

$$Z \equiv \frac{\partial \bar{\eta}}{\partial t} + \bar{\mathbf{v}} \cdot \nabla \bar{\eta} + \bar{\eta} \nabla \cdot \bar{\mathbf{v}} + \bar{\omega} \frac{\partial \bar{\eta}}{\partial p} + \mathbf{k} \cdot \nabla \bar{\omega} \times \frac{\partial \bar{\mathbf{v}}}{\partial p}$$

$$= -\nabla \cdot \overline{\eta' \mathbf{v}'} - \overline{\omega' \frac{\partial \eta'}{\partial p}} - \mathbf{k} \cdot \nabla \overline{\omega' \times \frac{\partial \mathbf{v}'}{\partial p}}$$



Cho et al. (1979)



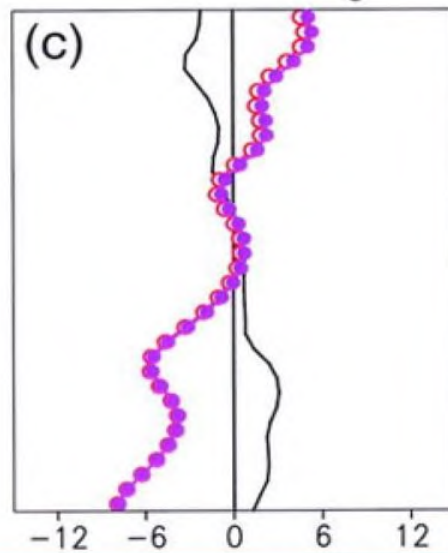
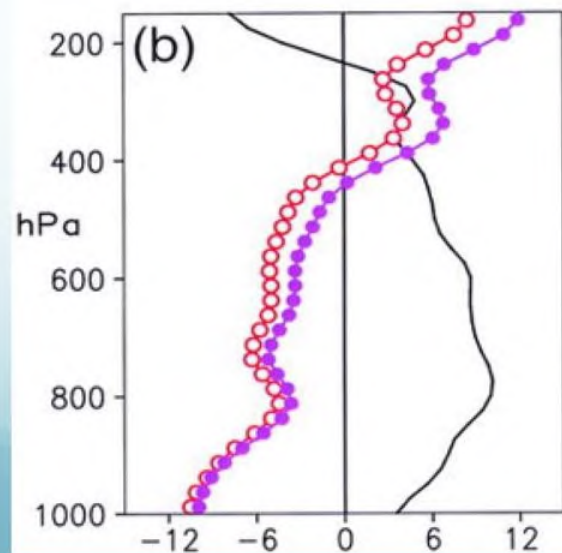
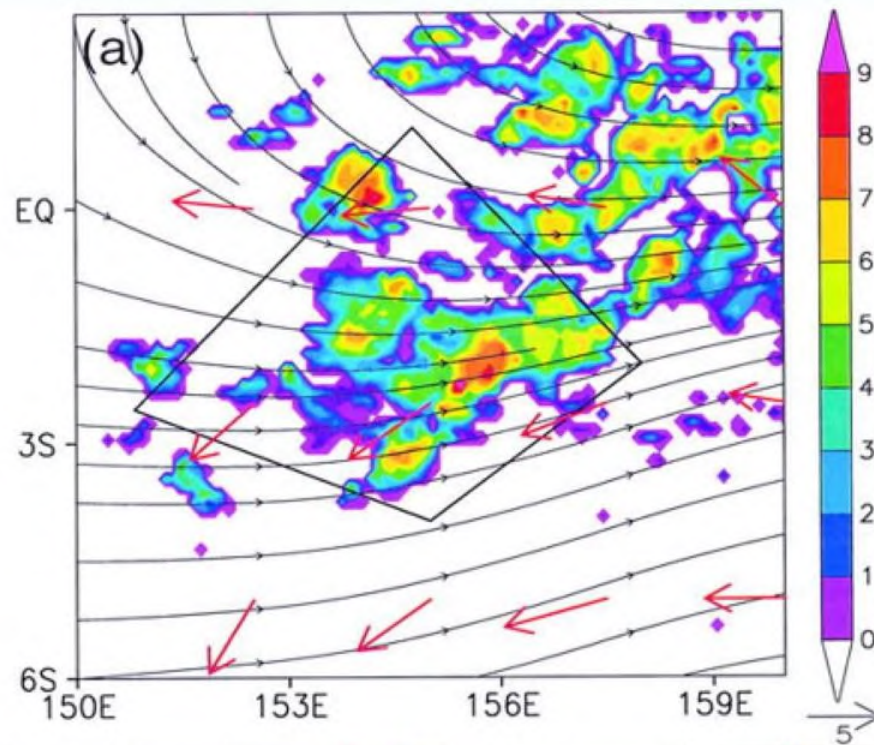
However, later observations found that the vorticity parameterization **cannot explain the observed vorticity dipole structure in convectively active regions**. This led to the effort to parameterize convective momentum transport directly.

$$\mathbf{X} \equiv \frac{\partial \bar{\mathbf{v}}}{\partial t} + \bar{\mathbf{v}} \cdot \nabla \bar{\mathbf{v}} + \bar{\omega} \frac{\partial \bar{\mathbf{v}}}{\partial p} + \nabla \bar{\phi} + f \mathbf{k} \times \bar{\mathbf{v}} = -\nabla \cdot \overline{\mathbf{v}' \mathbf{v}'} - \frac{\partial \overline{\mathbf{v}' \omega'}}{\partial p}$$

$$\square g \frac{\partial M_c (\mathbf{v}_c - \bar{\mathbf{v}})}{\partial p} \square - \frac{\partial M_c (\mathbf{v}_c - \bar{\mathbf{v}})}{\rho \partial z}$$

The difficulty lies in determining the cloud mean momentum.

Unlike the thermodynamic variables such as moist static energy, even for undiluted updrafts, **momentum is not conserved due to the action of convection-induced pressure gradient force.**



An example of convective transport of momentum in a 2-D cloud-resolving model

Mass flux

$$M_c \equiv \rho \sigma_c w_c = \frac{\rho}{N} \sum_i w_i$$

σ_u and w_u are the fractional area and average vertical velocity, respectively, of updrafts. N is the number of grid points in the model domain, and the summation is over all grid points occupied by updrafts.

The eddy momentum flux by convective updrafts is

$$F_c = \frac{\rho}{N} \sum_i w_i (V_i - \bar{V})$$

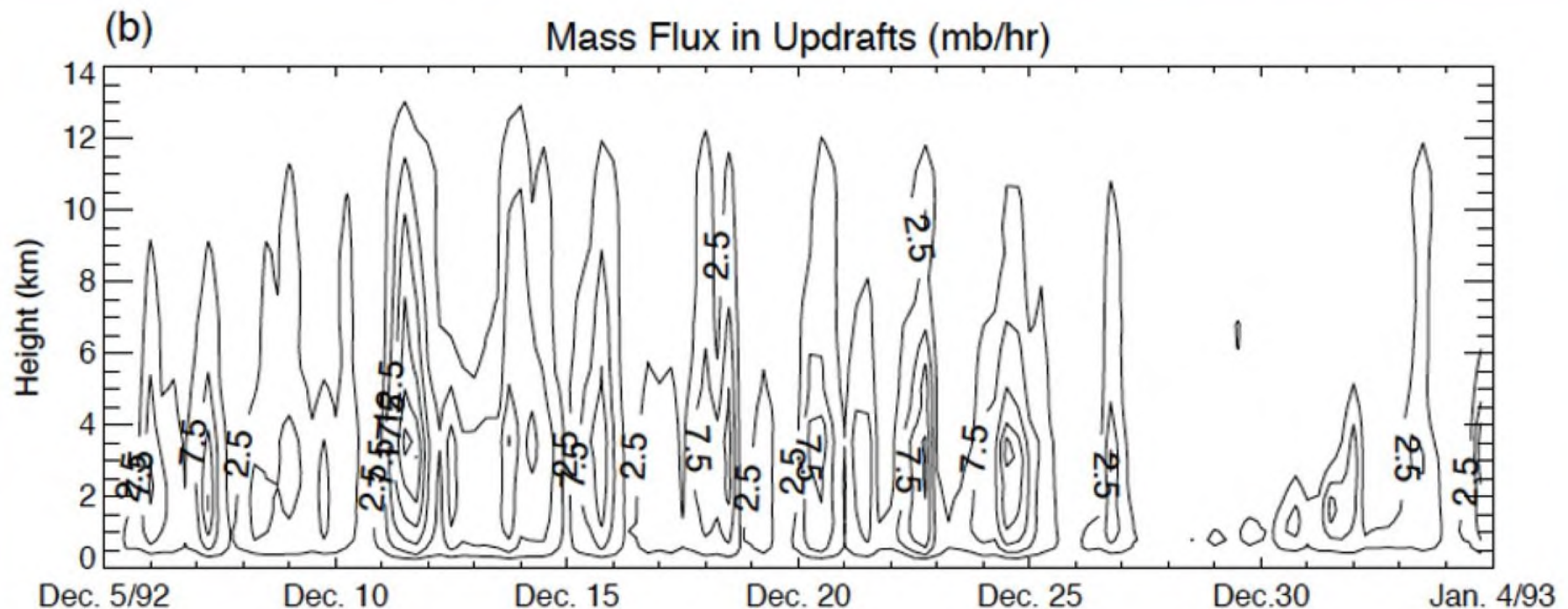
V_i is the horizontal wind at a given point inside updrafts. $\bar{V}$ is the domain averaged wind vector.

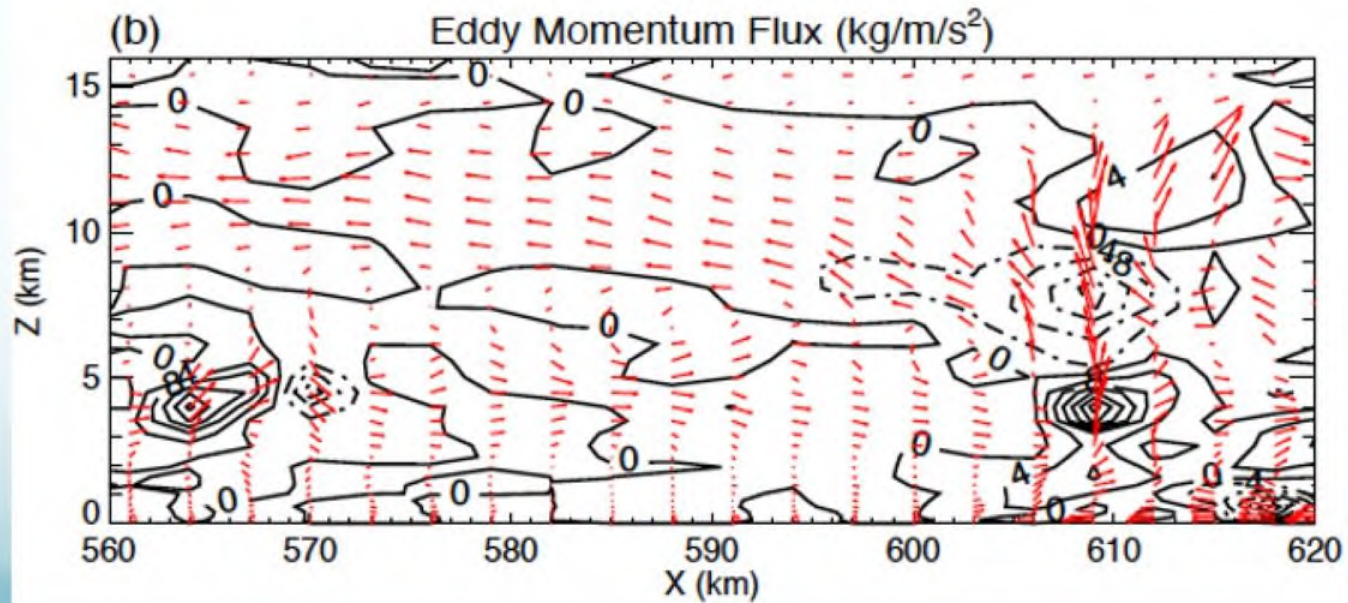
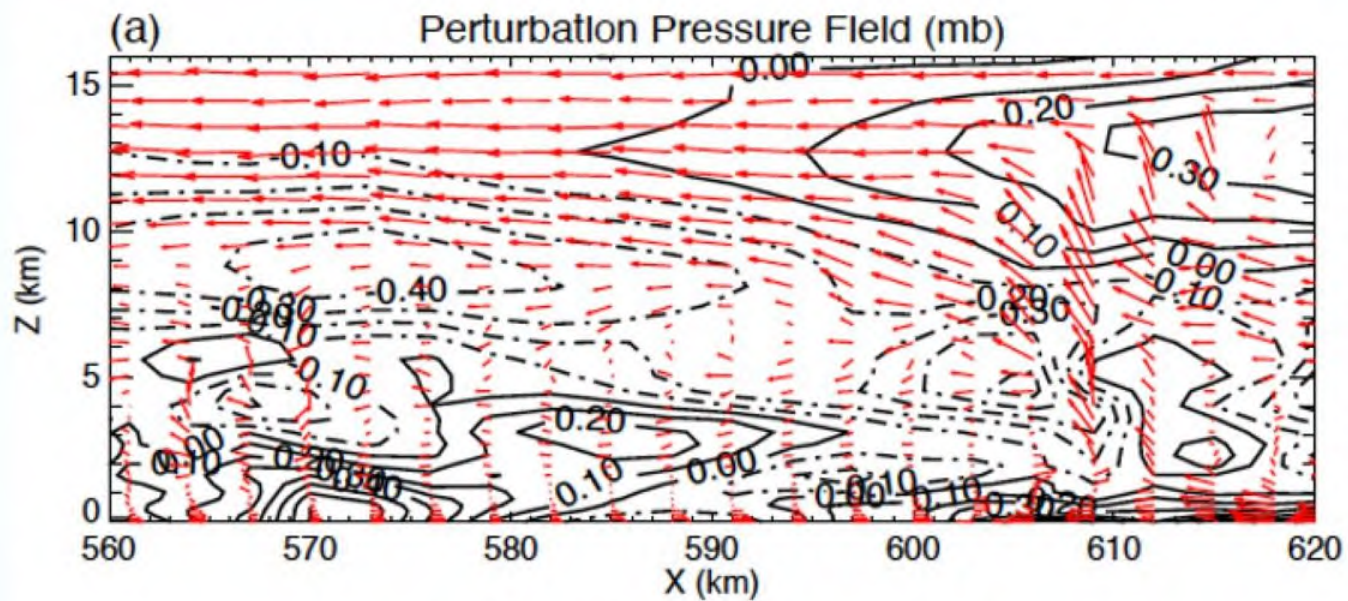
Expressed in terms of updraft mass flux and cloud-mean momentum, the above equation can be formally written as:

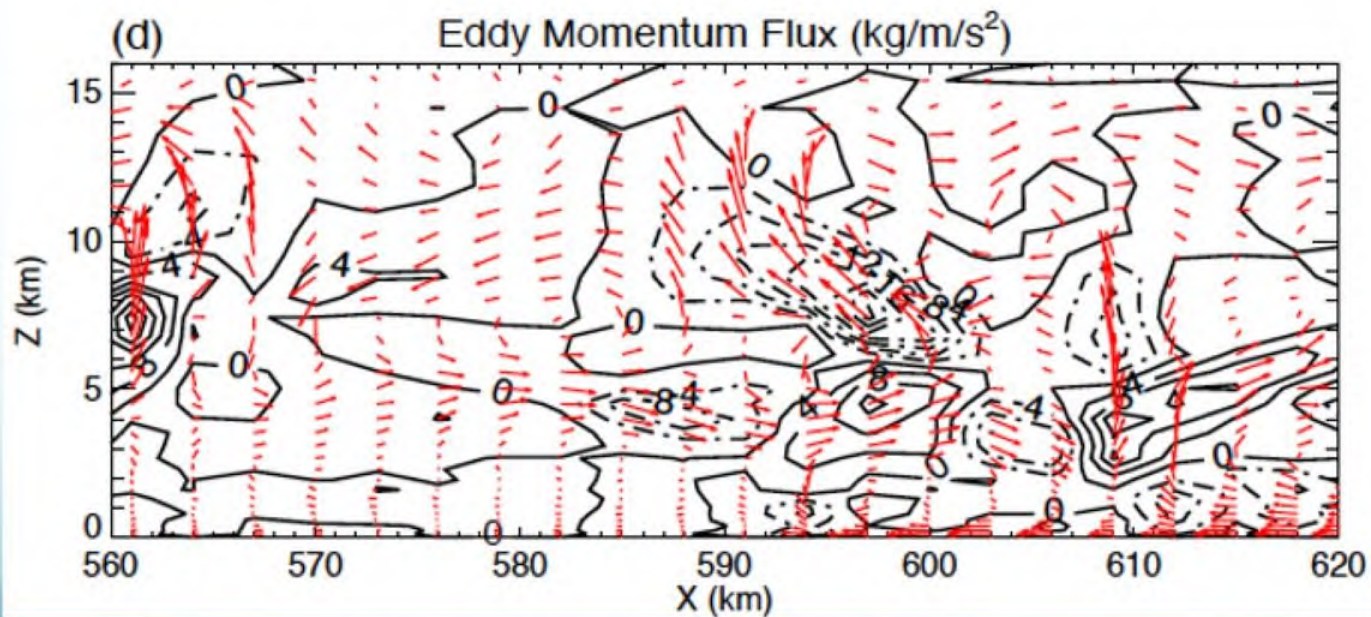
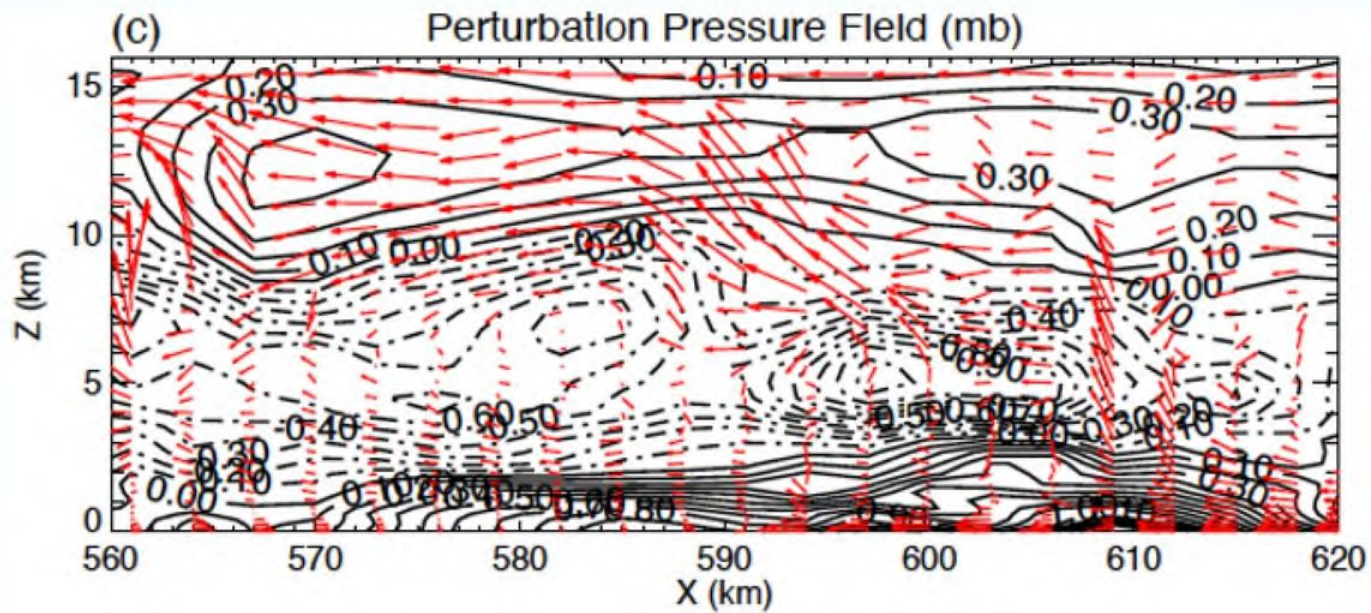
$$F_c = M_c (V_c - \bar{V})$$

where V_u is the mass-flux-weighted mean horizontal momentum of updrafts, defined by

$$V_c = \frac{\rho}{N} \sum_i w_i V_i / M_c.$$







The effect of momentum transport by convective updrafts on the momentum field is

$$X = -\frac{\partial M_c (V_c - \bar{V})}{\rho \partial z}$$

The convection-induced pressure gradient force contributes to the apparent momentum source X through its effect on the cloud-mean momentum V_c . For steady-state clouds, momentum equation in updrafts is approximated by

$$\delta V_c - \varepsilon \bar{V} + \frac{\partial (M_c V_c)}{\rho \partial z} = -\frac{1}{\rho} \sigma_c (\nabla p)_c$$

where ε and δ are the mass entrainment and detrainment at the cloud boundaries

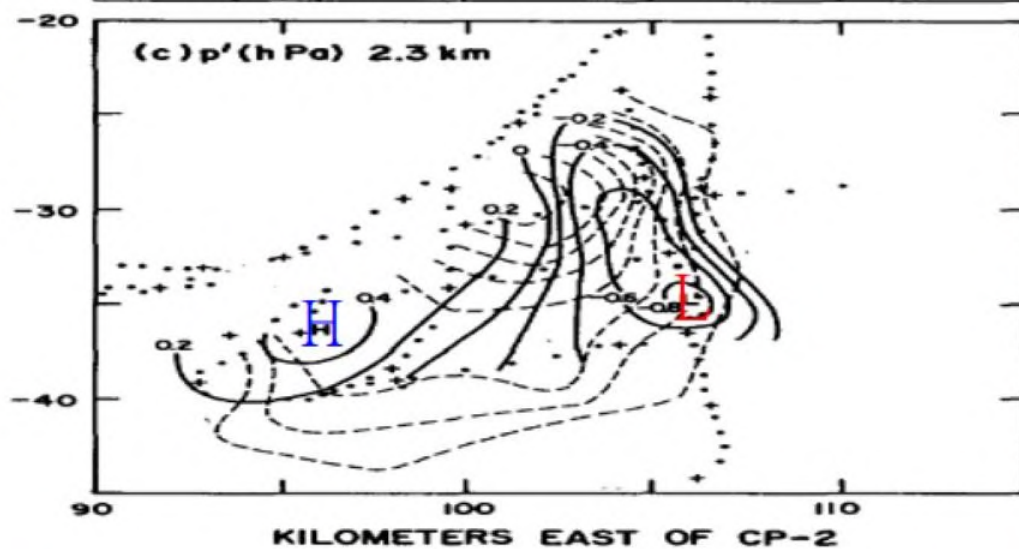
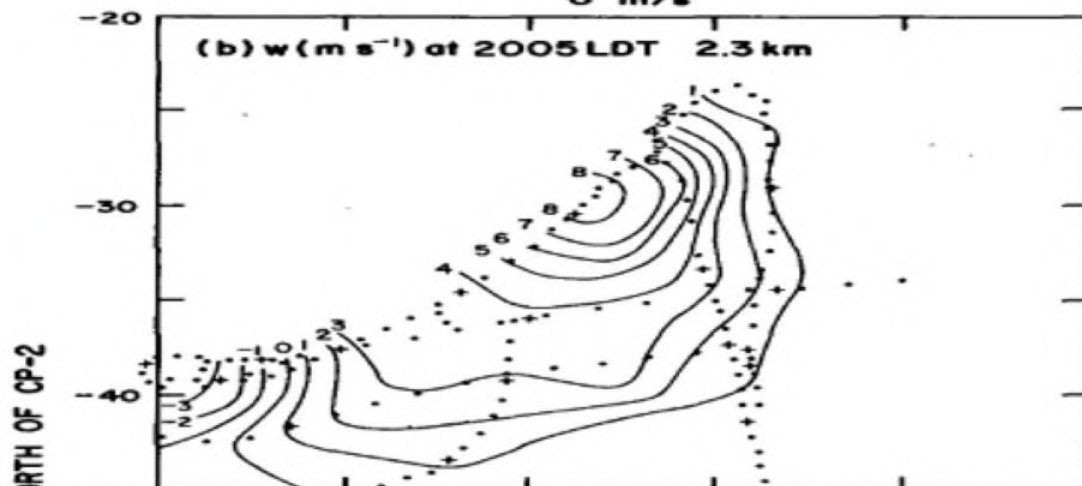
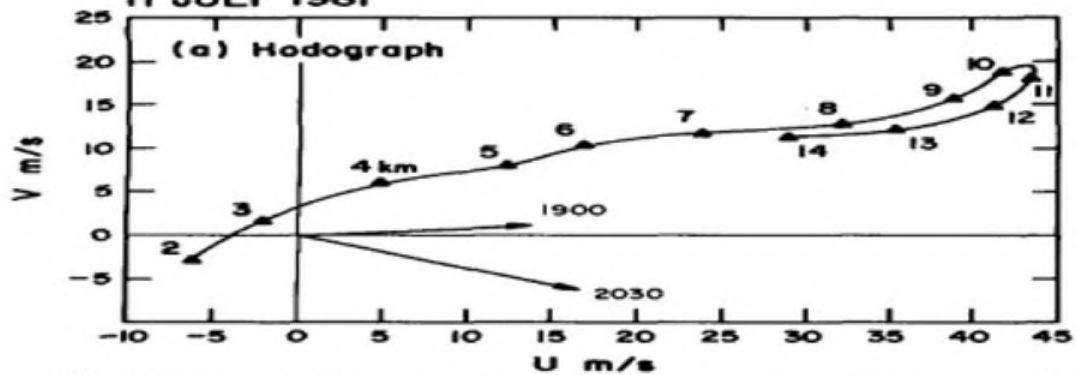
$$\frac{\partial M_c}{\rho \partial z} = \varepsilon - \delta$$

Substituting these into momentum source equation gives:

$$X = M_c \frac{\partial \bar{V}}{\rho \partial z} + \delta(V_c - \bar{V}) + \frac{1}{\rho} \sigma_c (\nabla p)_c$$

The **first term** is the product of the updraft mass flux and the vertical wind shear, and **is often interpreted as the vertical advection by the compensating subsidence** in the cloud environment; **the second represents the horizontal momentum detrainment** from the updrafts into their environment; **the last term is the pressure gradient across the updrafts**. It is this term that complicates the parameterization of convective momentum transport.

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We can establish an equation for this pressure field. The 3-D momentum equation is

$$\rho \frac{\partial \mathbf{v}_3}{\partial t} + \rho \mathbf{v}_3 \cdot \nabla_3 \mathbf{v}_3 + 2\rho \mathbf{\Omega} \times \mathbf{v}_3 = -\nabla p + \mathbf{k} \rho B$$

Note that

$\mathbf{v}_3 \cdot \nabla_3 \mathbf{v}_3 + 2\mathbf{\Omega} \times \mathbf{v}_3 = \nabla_3 \left(\frac{1}{2} \mathbf{v}_3 \cdot \mathbf{v}_3 \right) + \boldsymbol{\eta} \times \mathbf{v}_3$ where $\boldsymbol{\eta} = \nabla_3 \times \mathbf{v}_3 + 2\mathbf{\Omega}$ is 3-D absolute vorticity.

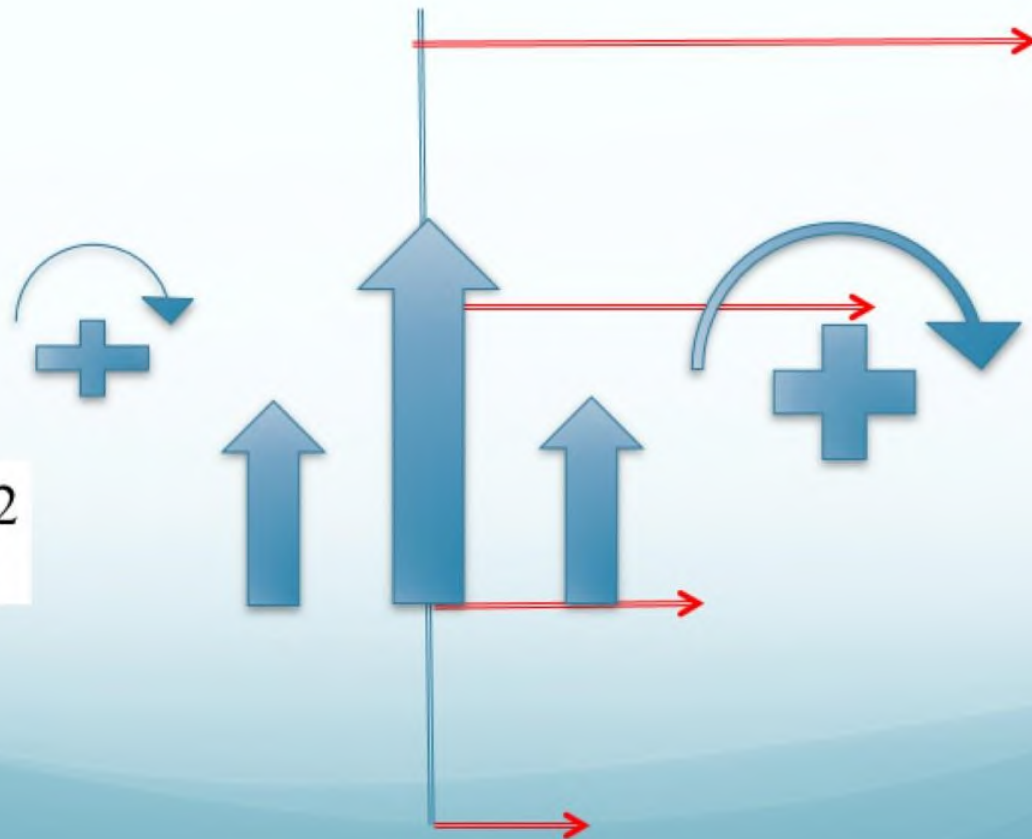
Take the 3-D divergence operation on the 3-D momentum equation and make use of mass continuity equation:

$$\begin{aligned} \nabla_3^2 p &= -\nabla_3 \cdot \left(\frac{\rho}{2} \nabla_3 \mathbf{v}_3 \cdot \mathbf{v}_3 \right) - \nabla_3 \cdot (\rho \boldsymbol{\eta} \times \mathbf{v}_3) + \nabla_3 \cdot (\rho \mathbf{k} B) \\ &= -\nabla_3^2 \left(\frac{\rho}{2} \mathbf{v}_3 \cdot \mathbf{v}_3 \right) + \nabla_3 \cdot \left(\frac{1}{2} \mathbf{v}_3 \cdot \mathbf{v}_3 \nabla_3 \rho \right) + \rho \eta^2 + \mathbf{v}_3 \cdot (\nabla_3 \times \rho \boldsymbol{\eta}) + \nabla_3 \cdot (\rho \mathbf{k} B) \end{aligned}$$

$$\nabla_3^2 p \propto -p \quad \Longrightarrow \quad p \propto -\eta^2$$

Consider an updraft in a vertical shear environment. Upshear of the updraft, the horizontal vorticity is less than that on the downshear side of the updraft. Thus, high pressure is located upshear and low pressure is located downshear, creating a pressure gradient across the updraft.

$$p \propto -\eta^2$$



$$\nabla_3^2(\nabla p) = \nabla \left\{ -2\rho \left[\frac{\partial w}{\partial x} \frac{\partial u}{\partial z} + \frac{\partial w}{\partial y} \frac{\partial v}{\partial z} + \frac{\partial v}{\partial x} \frac{\partial u}{\partial y} \right] - \rho \left[\left(\frac{\partial u}{\partial x} \right)^2 + \left(\frac{\partial v}{\partial y} \right)^2 + \left(\frac{\partial w}{\partial z} \right)^2 \right] \right. \\ \left. + w^2 \rho \frac{\partial}{\partial z} \left(\frac{1}{\rho} \frac{\partial \rho}{\partial z} \right) \right\}$$

We formally applied a gradient operator on both sides of the pressure equation, because the mean pressure over the convective area does not affect the momentum transport. For a waveform horizontal variation in perturbation pressure, the left-hand side of the above equation is proportional to $-\nabla p$. Thus averaging the rhs over the convective area gives a measure of the forcing to the pressure gradient field.

$$\nabla_3^2(\nabla p) = \nabla \left\{ -2\rho \frac{\partial w}{\partial x} \frac{\partial \bar{u}}{\partial z} - 2\rho \frac{\partial w}{\partial y} \frac{\partial \bar{v}}{\partial z} - 2\rho \frac{\partial w}{\partial x} \frac{\partial u'}{\partial z} - 2\rho \frac{\partial w}{\partial y} \frac{\partial v'}{\partial z} \right. \\ \left. - \rho \left[\left(\frac{\partial u}{\partial x} \right)^2 + \left(\frac{\partial v}{\partial y} \right)^2 + \left(\frac{\partial w}{\partial z} \right)^2 \right] + w^2 \rho \frac{\partial}{\partial z} \left(\frac{1}{\rho} \frac{\partial \rho}{\partial z} \right) \right\}$$

The first two terms on the rhs are the linear forcing, proportional to the vertical shear of the mean wind. The remainder represents nonlinear forcing, of which the first term is related to the perturbation wind shear, the terms inside the square brackets are related to the divergent flow and the third term is related to the atmospheric stratification.

Let's consider the linear forcing only:

$$\nabla_3^2(\nabla p) = \nabla \left\{ -2\rho \frac{\partial w}{\partial x} \frac{\partial \bar{u}}{\partial z} - 2\rho \frac{\partial w}{\partial y} \frac{\partial \bar{v}}{\partial z} \right\}$$

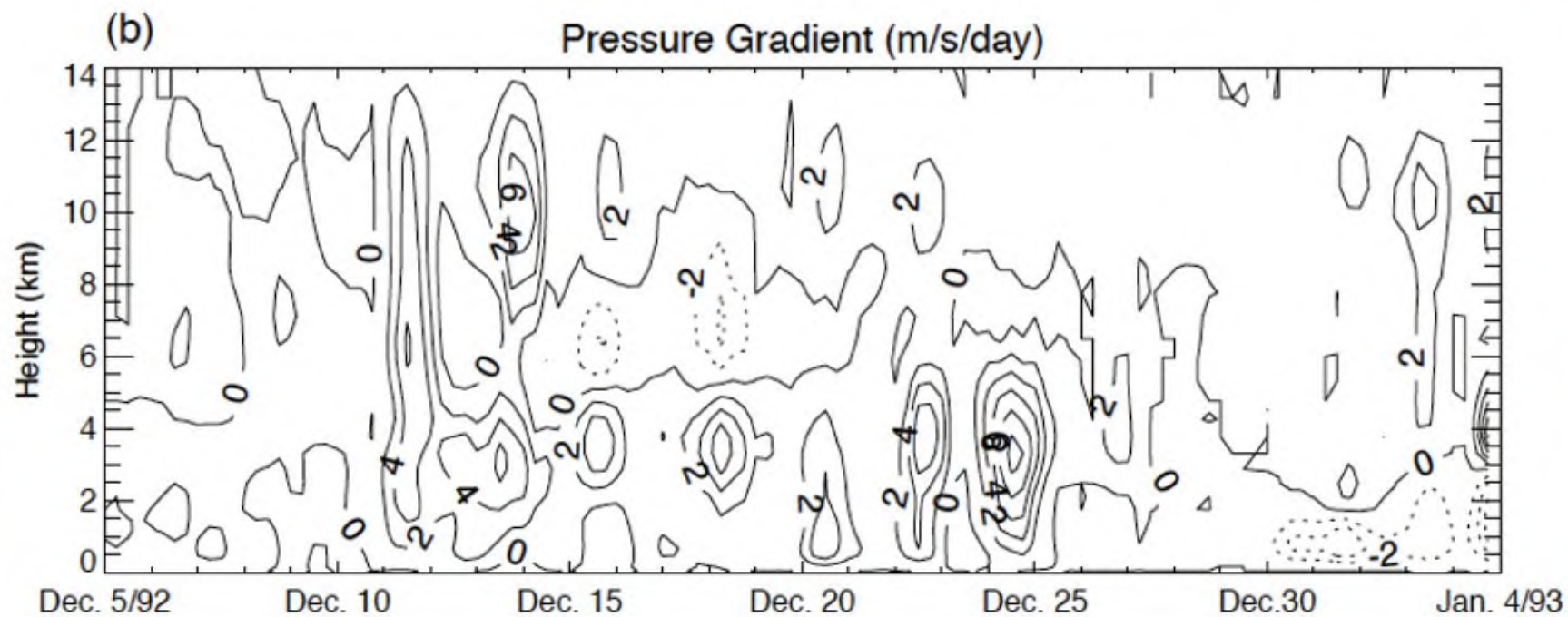
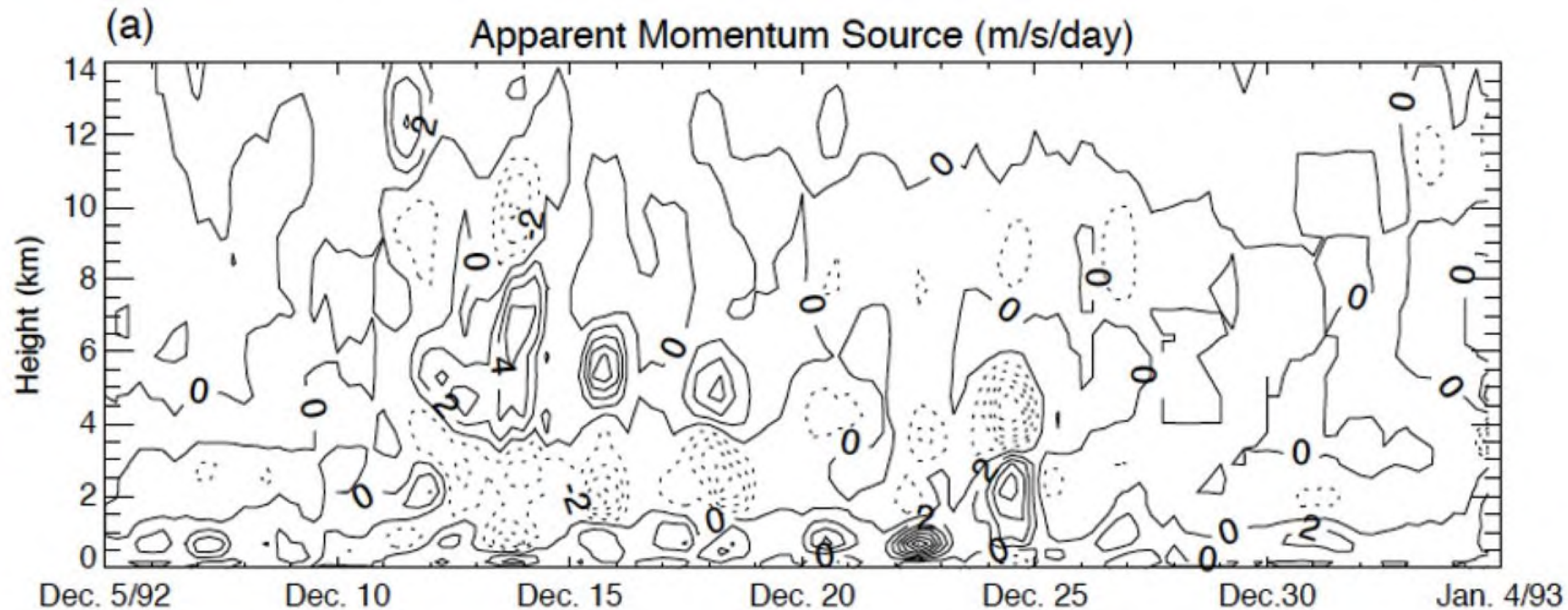
$$\frac{\partial p}{\partial x} \propto -\rho w \frac{\partial \bar{u}}{\partial z}$$

$$\frac{\sigma_c}{\rho} \left(\frac{\partial p}{\partial x} \right)_c \propto -\sigma_c \rho w_c \frac{\partial \bar{u}}{\rho \partial z} = -M_c \frac{\partial \bar{u}}{\rho \partial z}$$

$$\frac{\sigma_c}{\rho} \left(\frac{\partial p}{\partial x} \right)_c = -\alpha M_c \frac{\partial \bar{u}}{\rho \partial z}$$

Substituting into

$$X = M_c \frac{\partial \bar{V}}{\rho \partial z} + \delta(V_c - \bar{V}) + \frac{1}{\rho} \sigma_c (\nabla p)_c \longrightarrow X = (1 - \alpha) M_c \frac{\partial \bar{V}}{\rho \partial z} + \delta(V_c - \bar{V})$$



Model Cloud Wind Field

$$\zeta(r, \phi, z) = \zeta_c + \zeta_1(r, z) \sin(\phi - \phi_\zeta)$$

$$\delta(r, \phi, z) = \delta_c + \delta_1(r, z) \cos(\phi - \phi_\delta)$$

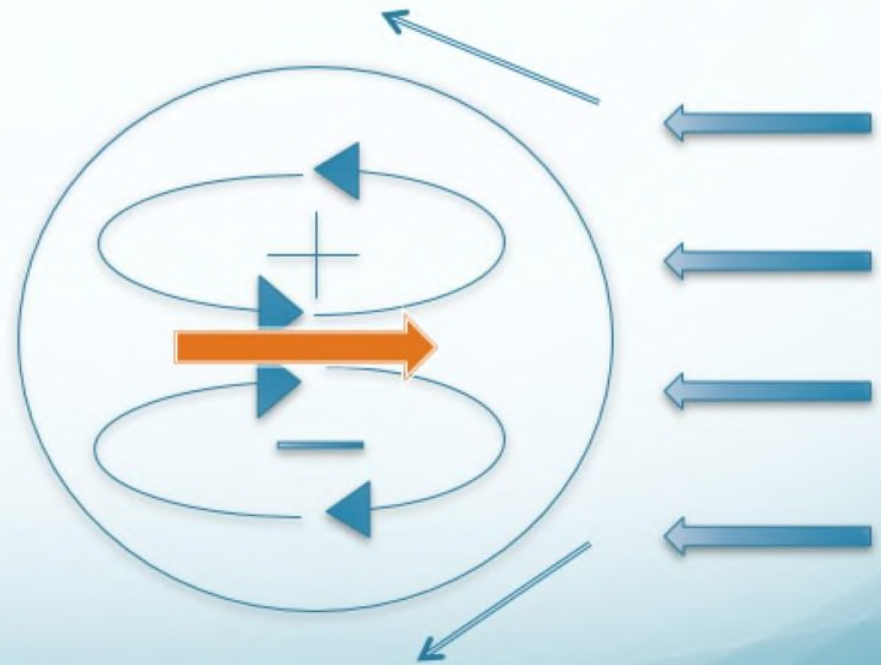
$$\zeta_1(r, z) = C(z)k^2 J_1(kr)$$

$$\delta_1(r, z) = -E(z)k^2 J_1(kr)$$



Divergence produces radial winds

vorticity dipole in updraft
produces cloud mean wind



$$\begin{aligned}
\mathbf{V}' &= \left(EkJ_1'(kr) \cos(\phi - \phi_\delta) - \frac{CJ_1(kr)}{r} \cos(\phi - \phi_\zeta) \right) \hat{r} \\
&\quad - \left(\frac{EJ_1(kr)}{r} \sin(\phi - \phi_\delta) - CkJ_1'(kr) \sin(\phi - \phi_\zeta) \right) \hat{\phi} + \frac{\delta_c}{2} r \hat{r} + \mathbf{V}_0 \quad \text{for } r \leq R \\
\mathbf{V}' &= \frac{R^2}{2r^2} J_1'(kR) \{ (Ck \cos(\phi - \phi_\zeta) + Ek \cos(\phi - \phi_\delta)) \hat{r} \\
&\quad + (Ck \sin(\phi - \phi_\zeta) + Ek \sin(\phi - \phi_\delta)) \hat{\phi} \} + \frac{\delta_c}{2} \frac{R^2}{r} \hat{r} \quad \text{for } r > R.
\end{aligned}$$

Fourier-Bessel Expansion of Cloud Pressure Field

$$G = \frac{p}{\bar{\rho}} + \frac{1}{2} \mathbf{V}_3 \cdot \mathbf{V}_3. \quad (12)$$

The equation for G can be derived readily from (8):

$$\nabla_3^2 G + 2 \frac{\partial \bar{\rho}}{\bar{\rho} \partial z} \frac{\partial G}{\partial z} + \frac{\partial^2 \bar{\rho}}{\bar{\rho} \partial z^2} G = f, \quad (13)$$

where

$$\begin{aligned} f = & \frac{\partial \bar{\rho}}{\bar{\rho} \partial z} \frac{\partial \frac{1}{2} \mathbf{V}_3 \cdot \mathbf{V}_3}{\partial z} + \frac{\partial^2 \bar{\rho}}{\bar{\rho} \partial z^2} \frac{1}{2} \mathbf{V}_3 \cdot \mathbf{V}_3 \\ & - \frac{1}{\bar{\rho}} \nabla_3 \cdot (\bar{\rho} \boldsymbol{\xi} \times \mathbf{V}_3) + \frac{\partial(\bar{\rho} B)}{\bar{\rho} \partial z}. \end{aligned} \quad (14)$$

$$G(r, \phi, z) = \sum_{n=0}^{\infty} (G_n(r, z) \cos(n\phi) + G_n^*(r, z) \sin(n\phi))$$

where

$$G_n(r, z) = \int_0^{\infty} H_k^n(z) J_n(kr) dk$$

and

$$G_n^*(r, z) = \int_0^{\infty} H_k^{n*}(z) J_n(kr) dk$$

$$\frac{d^2 H_k^n}{dz^2} + 2 \frac{\partial \bar{\rho}}{\bar{\rho} \partial z} \frac{d H_k^n}{dz} + \frac{\partial^2 \bar{\rho}}{\bar{\rho} \partial z^2} H_k^n - k^2 H_k^n = f_k^n \quad (19)$$

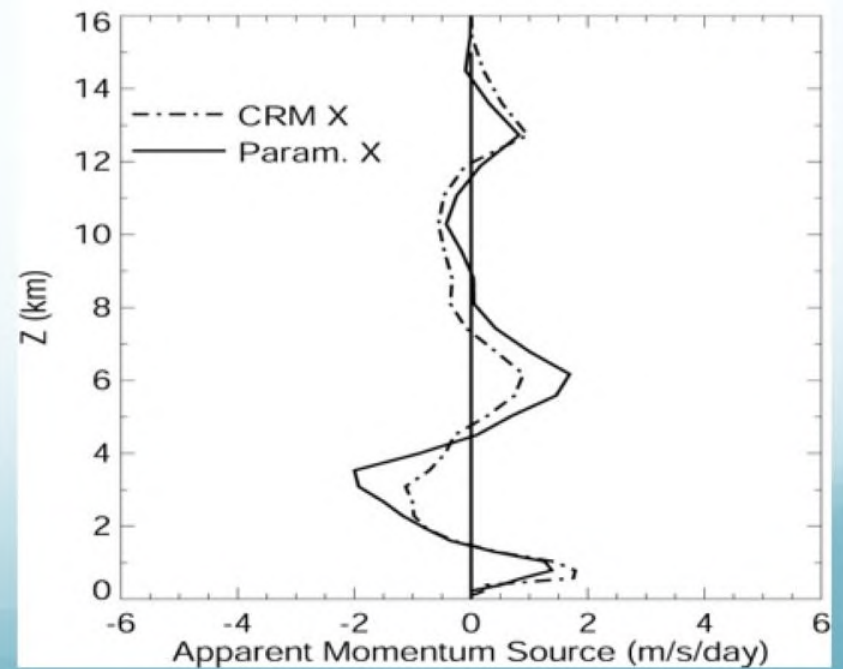
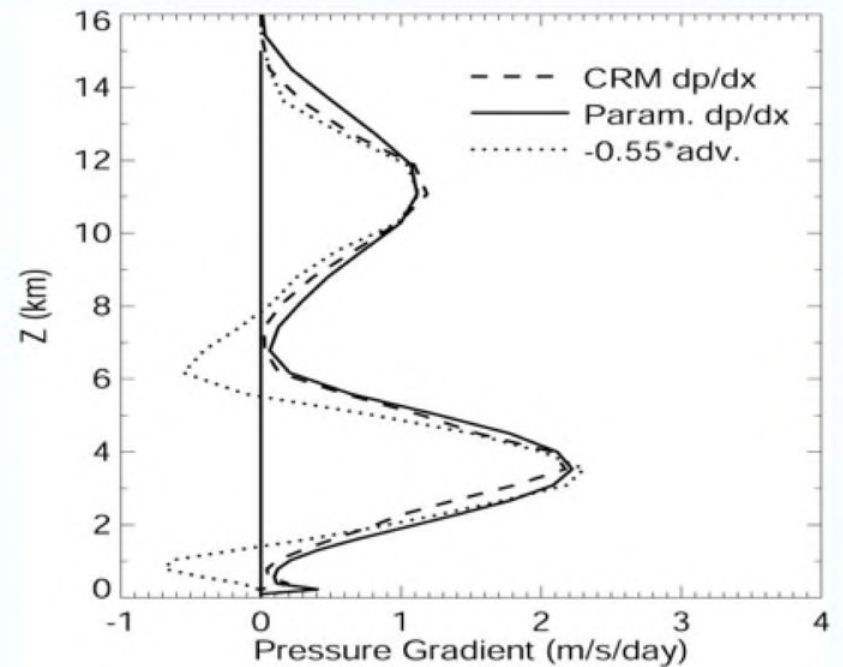
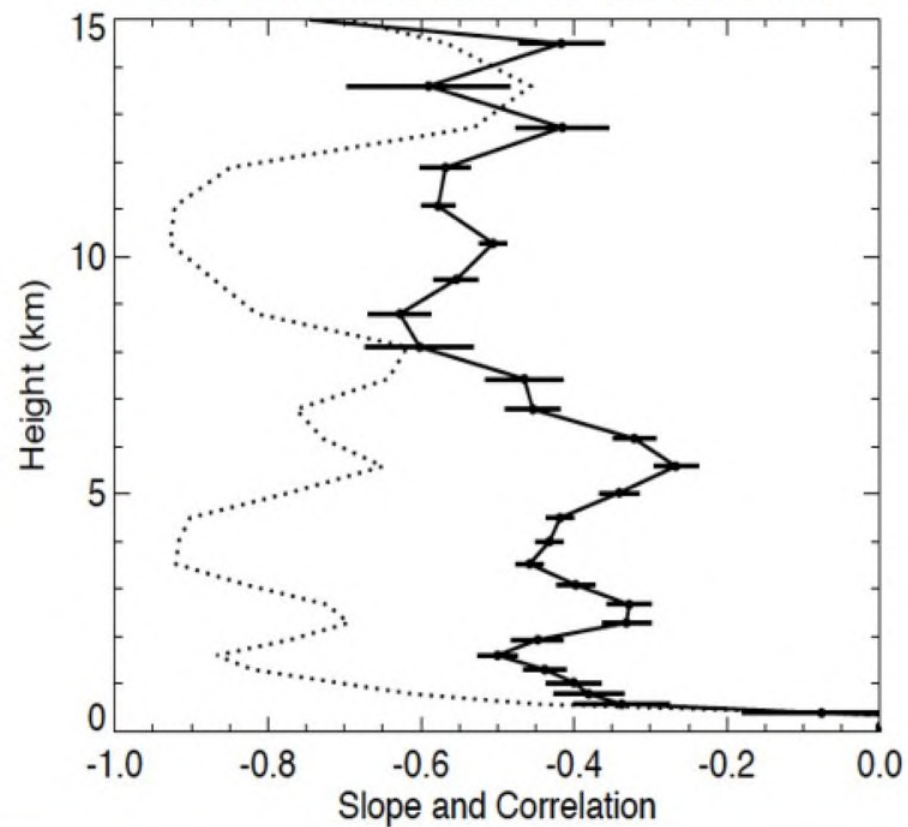
$$\frac{d^2 H_k^{n*}}{dz^2} + 2 \frac{\partial \bar{\rho}}{\bar{\rho} \partial z} \frac{d H_k^{n*}}{dz} + \frac{\partial^2 \bar{\rho}}{\bar{\rho} \partial z^2} H_k^{n*} - k^2 H_k^{n*} = f_k^{n*} \quad (20)$$

where

$$f_k^n = \frac{k}{\pi} \int_0^\infty r J_n(kr) \left(\int_0^{2\pi} f \cos(n\phi) d\phi \right) dr \quad (21a)$$

$$f_k^{n*} = \frac{k}{\pi} \int_0^\infty r J_n(kr) \left(\int_0^{2\pi} f \sin(n\phi) d\phi \right) dr. \quad (21b)$$

Linear Regression between Pressure Gradient Force and Vertical Advection



For linear forcing, we can formally derive an equation for the pressure gradient force. Assume w is a sine wave:

$$w = w_0 \cos kx \cos ly \sin mz \text{ for } -\frac{\pi}{2} < kx, ly < \frac{\pi}{2}$$

$$\nabla_3^2(\nabla p) = \nabla \left\{ -2\rho \frac{\partial w}{\partial x} \frac{\partial \bar{u}}{\partial z} - 2\rho \frac{\partial w}{\partial y} \frac{\partial \bar{v}}{\partial z} \right\}$$

$$p = \frac{2\rho}{k^2 + l^2 + m^2} \left(\frac{\partial \bar{u}}{\partial z} k \sin kx \cos ly + \frac{\partial \bar{v}}{\partial z} l \cos kx \sin ly \right) w_0 \sin mz$$

$$\frac{1}{\rho} \frac{\partial p}{\partial x} = \frac{2w_0}{k^2 + l^2 + m^2} \left(\frac{\partial \bar{u}}{\partial z} k^2 \cos kx \cos ly - \frac{\partial \bar{v}}{\partial z} kl \sin kx \sin ly \right) \sin mz$$

$$\frac{1}{\rho} \frac{\partial p}{\partial y} = \frac{2w_0}{k^2 + l^2 + m^2} \left(-\frac{\partial \bar{u}}{\partial z} kl \sin kx \sin ly + \frac{\partial \bar{v}}{\partial z} l^2 \cos kx \cos ly \right) \sin mz$$

Averaging over cloud area, we get

$$\frac{\sigma_c}{\rho} \left(\frac{\partial p}{\partial x} \right)_c = - \frac{\sigma_c 2k^2}{k^2 + l^2 + m^2} \frac{\partial \bar{u}}{\partial z} w_c = -\gamma_x M_c \frac{\partial \bar{u}}{\rho \partial z}$$

$$\frac{\sigma_c}{\rho} \left(\frac{\partial p}{\partial y} \right)_c = - \frac{\sigma_c 2l^2}{k^2 + l^2 + m^2} \frac{\partial \bar{v}}{\partial z} w_c = -\gamma_y M_c \frac{\partial \bar{v}}{\rho \partial z}$$

Thus, convection-induced pressure gradient depends on **the horizontal as well as vertical scales of convection updrafts**, in addition to mass flux and wind shear.

For shallow convection, m is large, thus pressure gradient force is small. Similarly, horizontally smaller clouds have relatively larger pressure gradient force.

For individual convective cells, $k \sim l \sim m \sim 10$ km

$$\frac{\sigma_c}{\rho} \left(\frac{\partial p}{\partial x} \right)_c = - \frac{\sigma_c 2k^2}{k^2 + l^2 + m^2} \frac{\partial \bar{u}}{\partial z} w_c = -0.67 M_c \frac{\partial \bar{u}}{\rho \partial z}$$

$$\frac{\sigma_c}{\rho} \left(\frac{\partial p}{\partial y} \right)_c = - \frac{\sigma_c 2l^2}{k^2 + l^2 + m^2} \frac{\partial \bar{v}}{\partial z} w_c = -0.67 M_c \frac{\partial \bar{v}}{\rho \partial z}$$

For squall lines in y-direction, $l=0$, $k \sim m$

$$\frac{\sigma_c}{\rho} \left(\frac{\partial p}{\partial x} \right)_c = - \frac{\sigma_c 2k^2}{k^2 + m^2} \frac{\partial \bar{u}}{\partial z} w_c = -M_c \frac{\partial \bar{u}}{\rho \partial z}$$

$$\frac{\sigma_c}{\rho} \left(\frac{\partial p}{\partial y} \right)_c = 0$$

An example from 3-D CRM simulation

Convection-induced pressure perturbation

$P_{XLU}, P_{XLD}, P_{XLE}$ Linear forcing

$P_{XNU}, P_{XND}, P_{XNE}$ Nonlinear perturbation wind shear forcing

$$\nabla_3^2 \left(\frac{\partial p}{\partial x} \right) = \frac{\partial}{\partial x} \left\{ -2\rho \frac{\partial w}{\partial x} \frac{\partial \bar{u}}{\partial z} - 2\rho \frac{\partial w}{\partial y} \frac{\partial \bar{v}}{\partial z} - 2\rho \frac{\partial w}{\partial x} \frac{\partial u'}{\partial z} - 2\rho \frac{\partial w}{\partial y} \frac{\partial v'}{\partial z} - \rho \left[\left(\frac{\partial u}{\partial x} \right)^2 + \left(\frac{\partial v}{\partial y} \right)^2 + \left(\frac{\partial w}{\partial z} \right)^2 \right] + w^2 \rho \frac{\partial}{\partial z} \left(\frac{1}{\rho} \frac{\partial \rho}{\partial z} \right) \right\}$$

S_{XU}, S_{XD}, S_{XE} Nonlinear stratification forcing
 D_{XU}, D_{XD}, D_{XE} Nonlinear divergence forcing

$P_{yLU}, P_{yLD}, P_{yLE}$

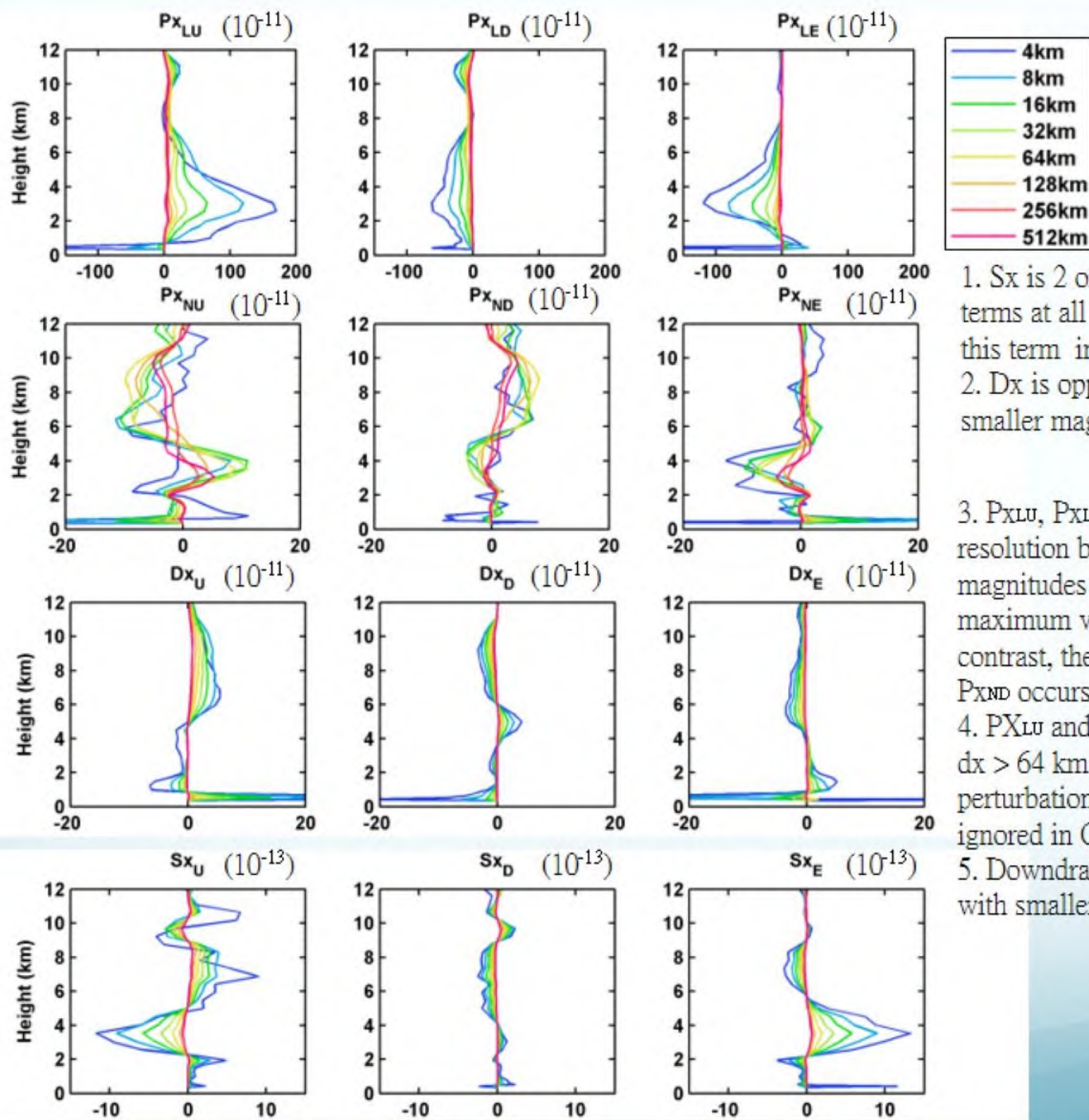
$P_{yNU}, P_{yND}, P_{yNE}$

D_{yU}, D_{yD}, D_{yE}

$$\nabla_3^2 \left(\frac{\partial p}{\partial y} \right) = \frac{\partial}{\partial y} \left\{ -2\rho \frac{\partial w}{\partial x} \frac{\partial \bar{u}}{\partial z} - 2\rho \frac{\partial w}{\partial y} \frac{\partial \bar{v}}{\partial z} - 2\rho \frac{\partial w}{\partial x} \frac{\partial u'}{\partial z} - 2\rho \frac{\partial w}{\partial y} \frac{\partial v'}{\partial z} - \rho \left[\left(\frac{\partial u}{\partial x} \right)^2 + \left(\frac{\partial v}{\partial y} \right)^2 + \left(\frac{\partial w}{\partial z} \right)^2 \right] + w^2 \rho \frac{\partial}{\partial z} \left(\frac{1}{\rho} \frac{\partial \rho}{\partial z} \right) \right\}$$

S_{yU}, S_{yD}, S_{yE}

Subscript U,D, E represent updraft, downdraft, and environment, respectively



1. Sx is 2 order smaller than the other three terms at all the scales. It is ok to ignore this term in VMT scheme.

2. Dx is opposite in sign to Px_{NU} , but with smaller magnitude.

3. Px_{LU} , Px_{LD} , and Px_{LE} increase with resolution because resolved u and v magnitudes increase with resolution. The maximum value occurs at $dx = 4$ km. In contrast, the maximum value of Px_{NU} and Px_{ND} occurs at $dx = 16$ -64 km.

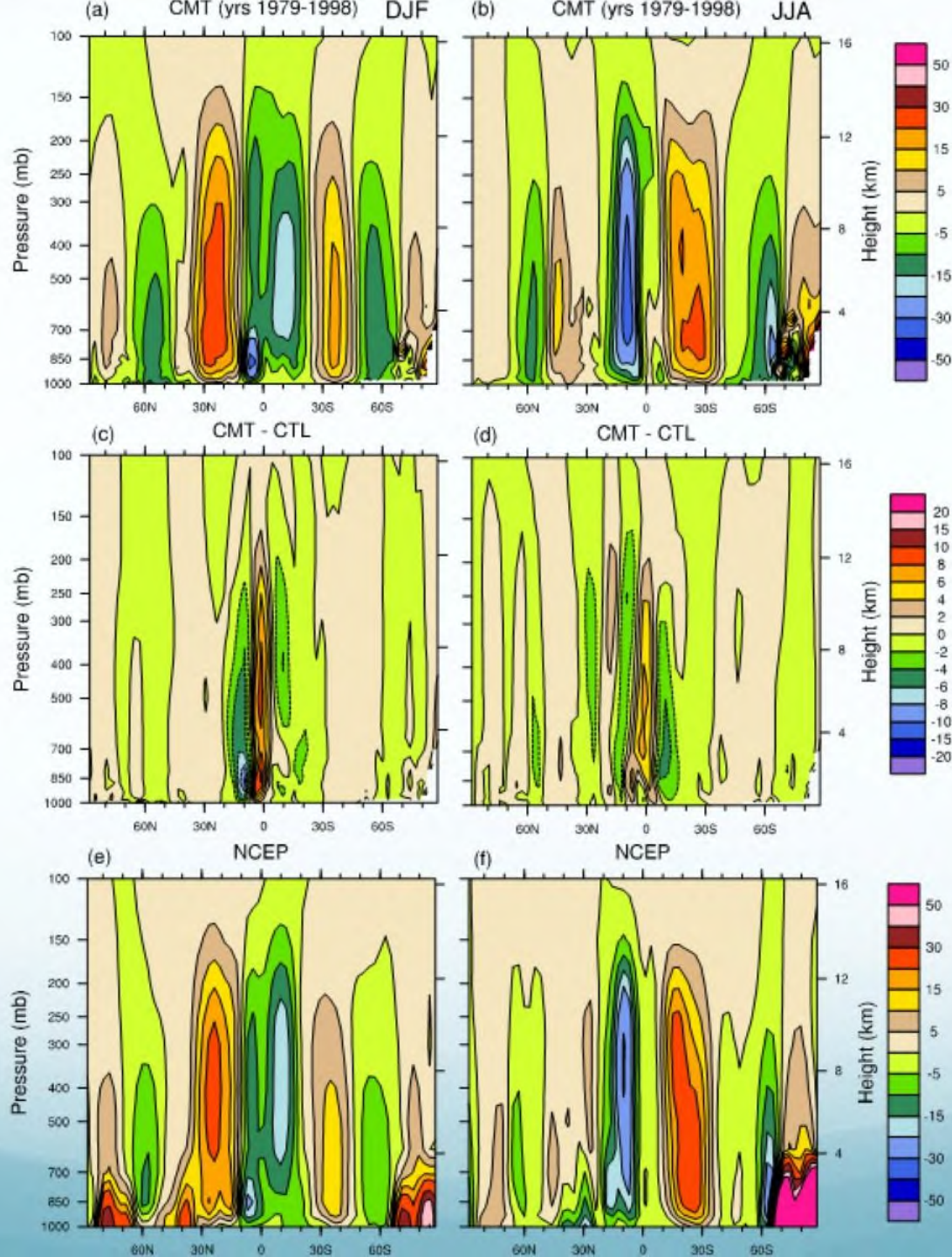
4. Px_{LU} and Px_{NU} became comparable at $dx > 64$ km. Should nonlinear perturbation wind shear forcing (Px_{NU}) be ignored in CMT scheme in GCM model?

5. Downdraft has opposite sign to updraft with smaller magnitude for all four terms.

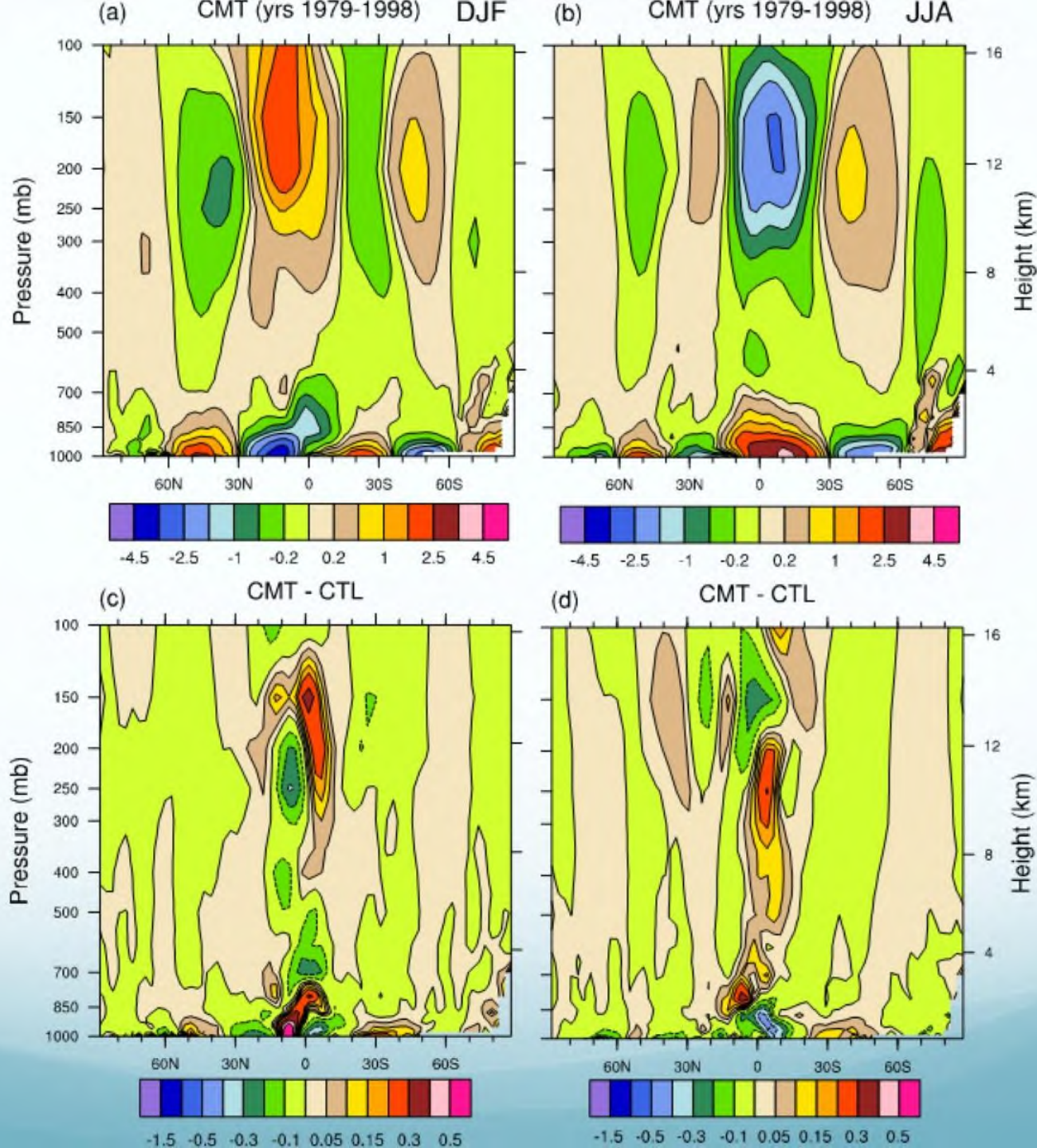
GCM Experiments

20 year simulation using NCAR CCM3

Vertical
velocity



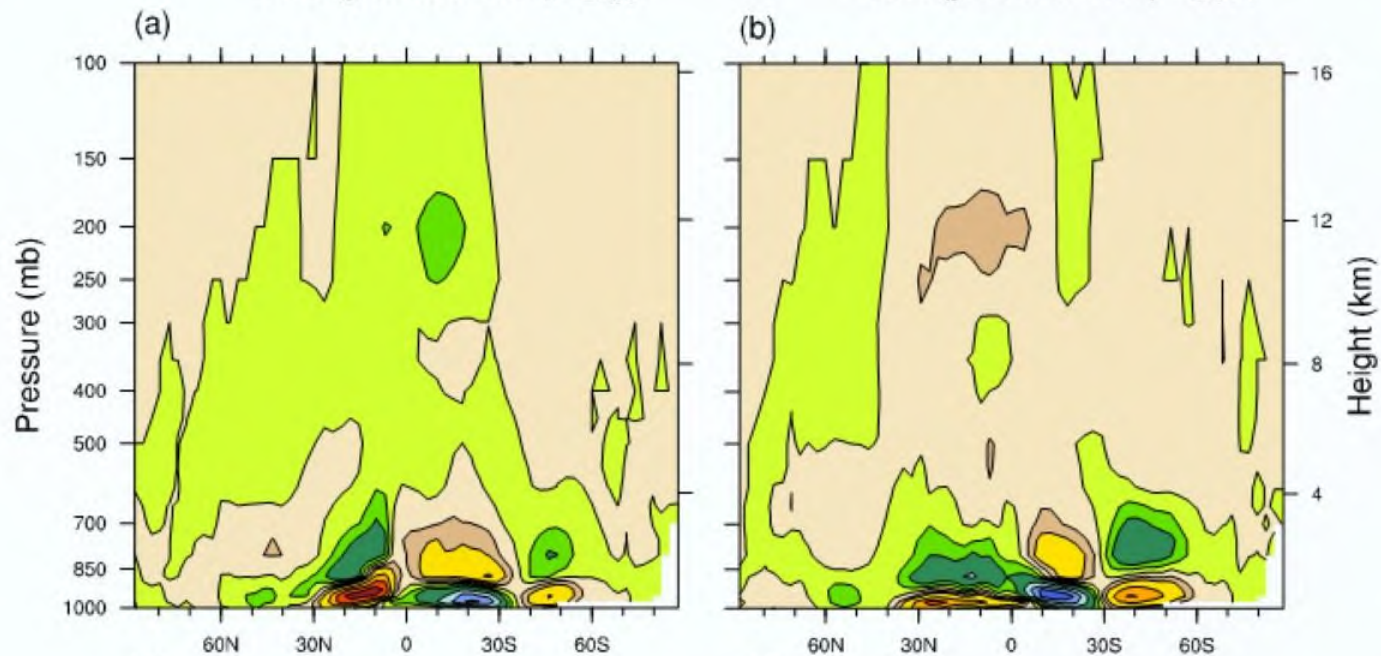
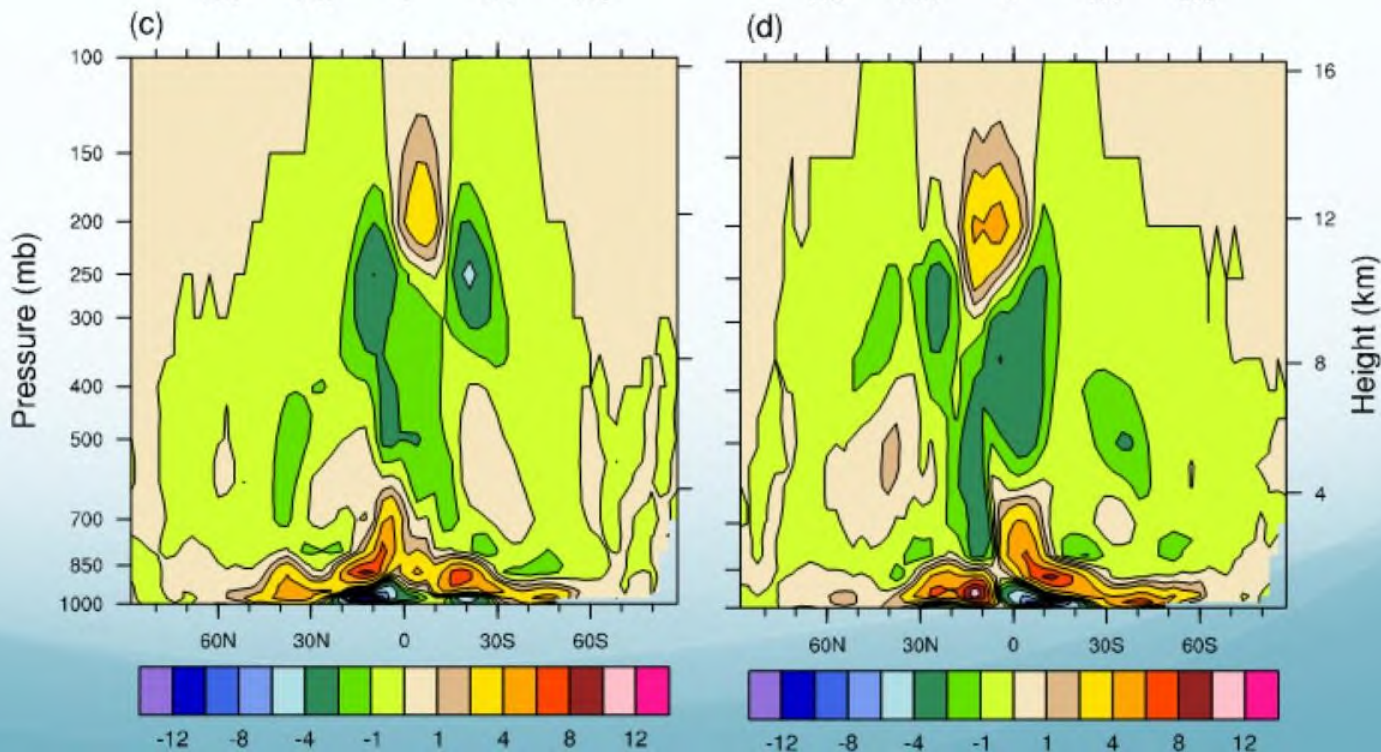
Meridional wind

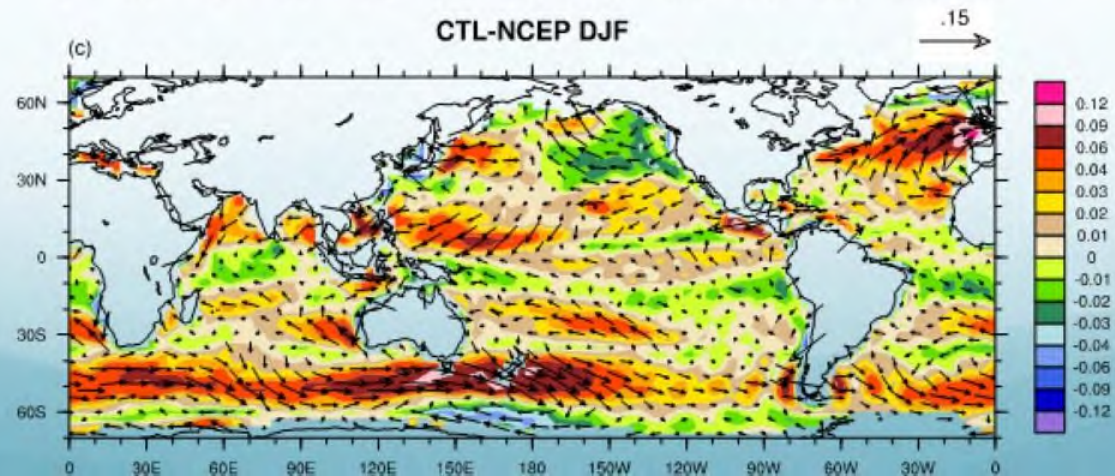
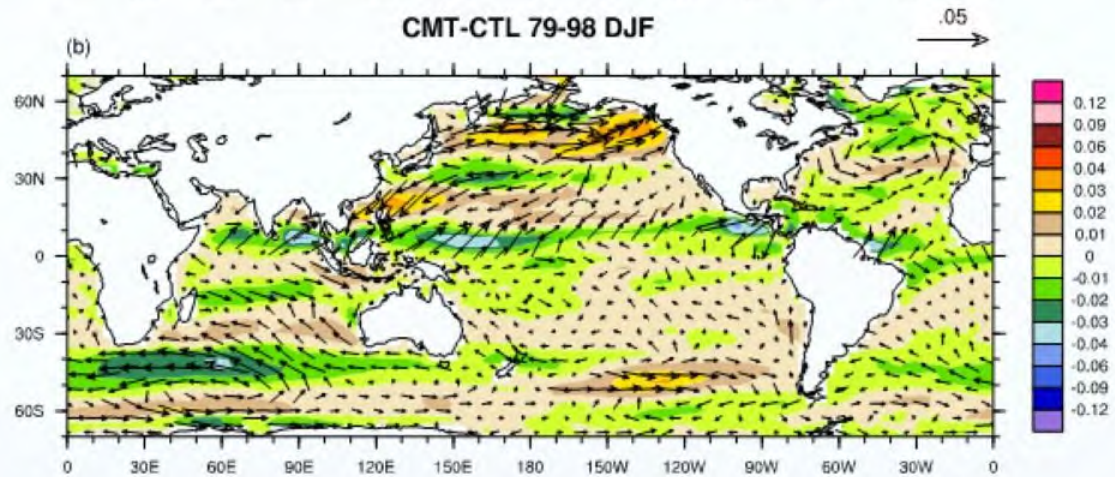
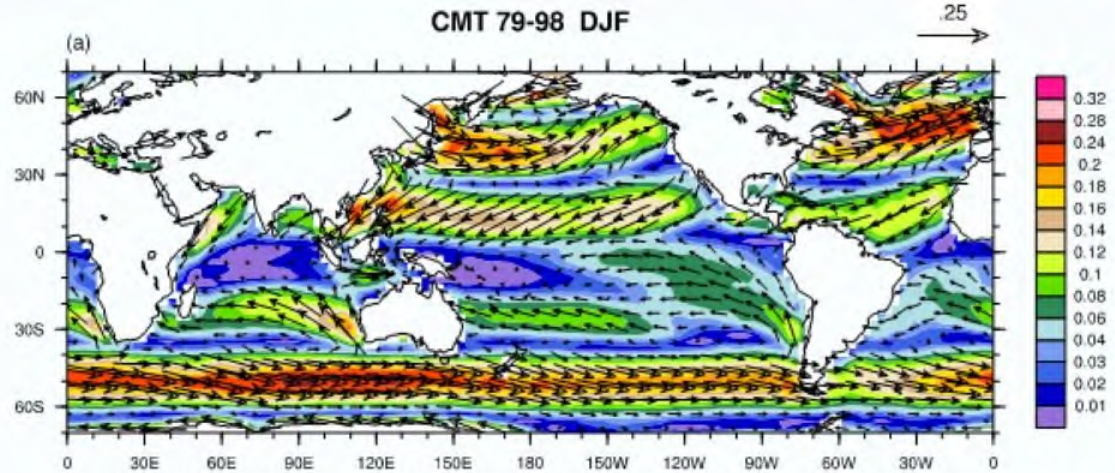


CMT (yrs 1979-1998) DJF

CMT (yrs 1979-1998) JJA

meridional

Zonal apparent
momentum source



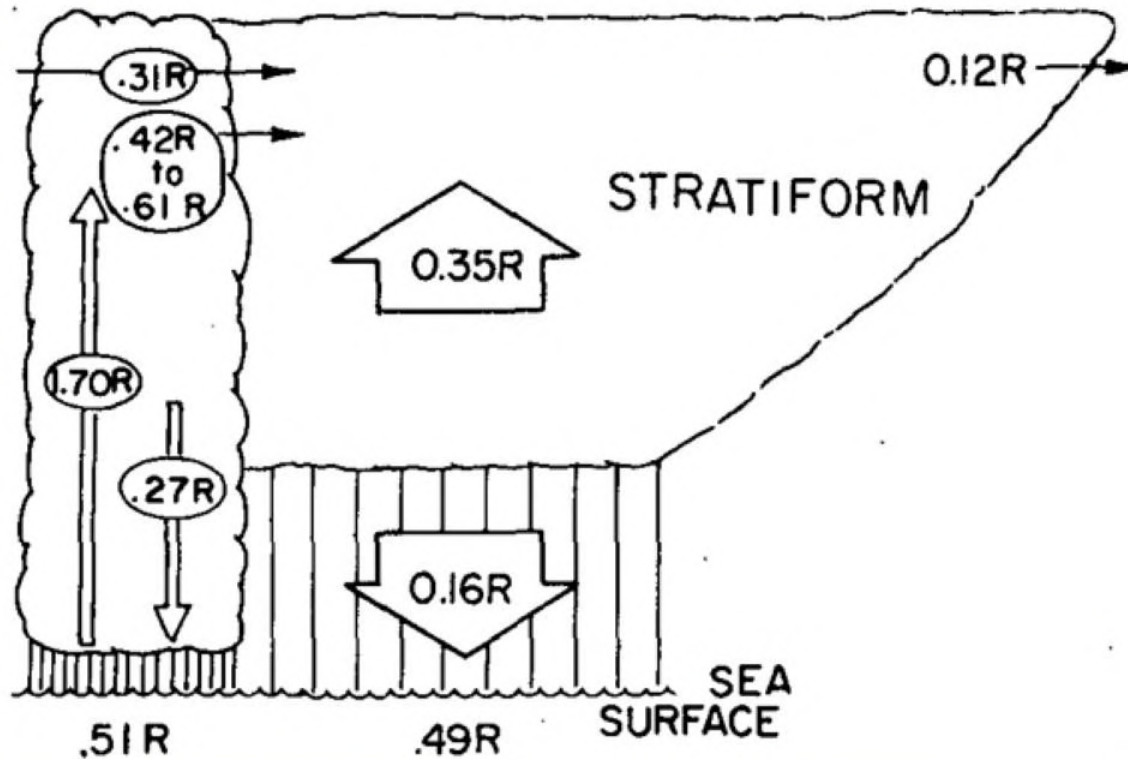
Microphysics Parameterization for Convection in GCMs

Why convective microphysics?

- Precipitation in convection is generated through microphysical processes
- Convection-cloud interaction; anvil clouds are from convective detrainment
- Aerosol-convection-cloud interaction

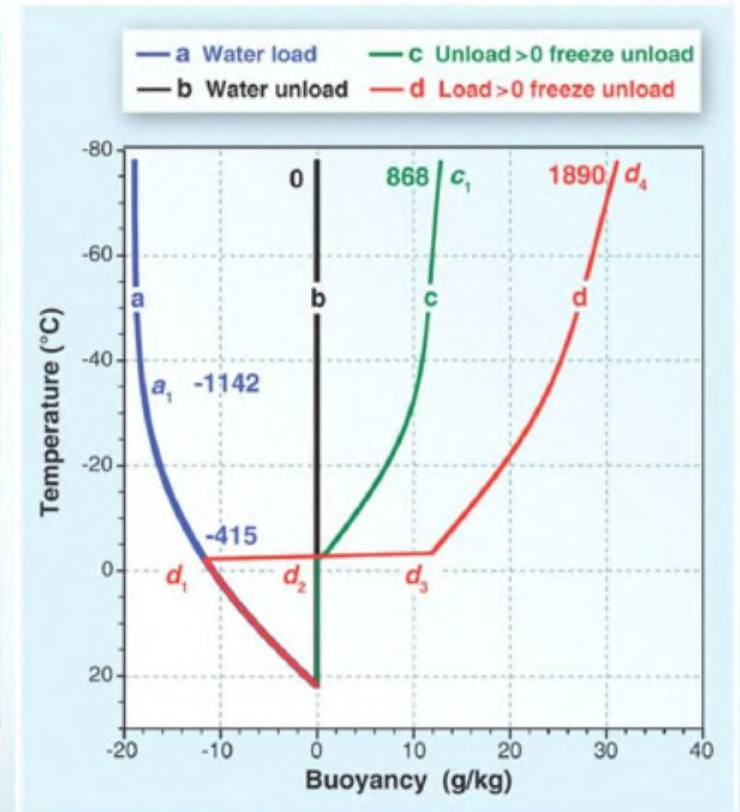
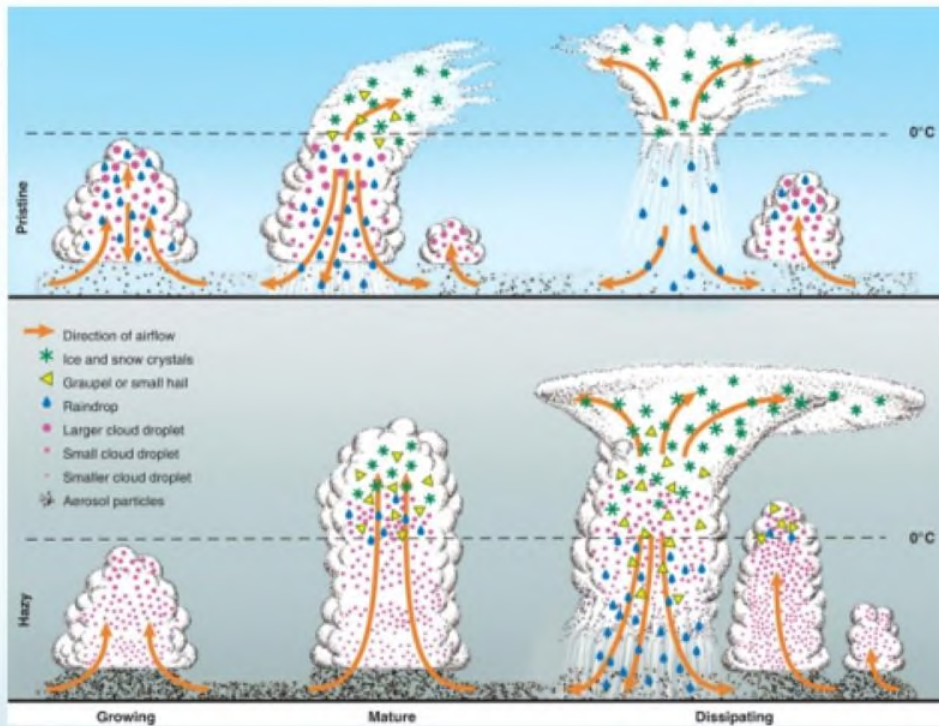


Convection-cloud interaction



Gamache and Houze, 1983, JAS

Aerosol-convection-cloud interaction



D Rosenfeld et al. Science 2008;321:1309-1313

Recall that the equation for cloud water content is given by:

$$\frac{\partial M_u q_l}{\rho \partial z} = -\delta_u q_l + c_u - R_r$$

In early convection parameterization schemes, the conversion of cloud water to rainwater is based on simple intuition.

In Arakawa and Schubert scheme

$$R_r = c_0 M_u q_l$$

In Emanuel's scheme, $q_l = (1 - \varepsilon) q_{la}$

q_{la} is the adiabatic liquid water mixing ratio

In Tiedtke (1989) scheme

$$R_r = \frac{M_u}{\rho} K(p) q_l$$

$$K(p) = \begin{cases} 0 & \text{for } p_B - p \leq \Delta p_{crit} \\ 6 \times 10^{-4} m^{-1} & \text{for } p_B - p > \Delta p_{crit} \end{cases}$$

where Δp_{crit} is set to 150 mb over ocean and 300 mb over land.

Single moment parameterization

$$\frac{\partial M_u q_l}{\rho \partial z} = -\delta_u q_l + c_{ul} - G_{pl}$$

$$\frac{\partial M_u q_i}{\rho \partial z} = -\delta_u q_i + c_{ui} - G_{pi}$$

where q_l and q_i are cloud liquid and ice mixing ratio, respectively
The conversion terms G_{pl} and G_{pi} , which represent sinks to cloud water and ice due to conversion to rain and snow, are parameterized through sophisticated cloud microphysical processes

$$G_{pl} = f_c (Q_{aut} + Q_{rac1} + Q_{sac1} + Q_{fho} + Q_{fhe} - Q_{mlt})$$

$$G_{pi} = f_c (Q_{agg} + Q_{saci} - Q_{fho} - Q_{fhe} + Q_{mlt})$$

where f_c is cloud fraction, aut is for autoconversion, rac1 and sac1 for accretion of cloud droplets by raindrops and snow, fho and fhe for homogeneous and heterogeneous freezing, mlt for melting of ice crystals, agg for aggregation of ice crystals, and saci for accretion of ice crystals by snow, respectively.

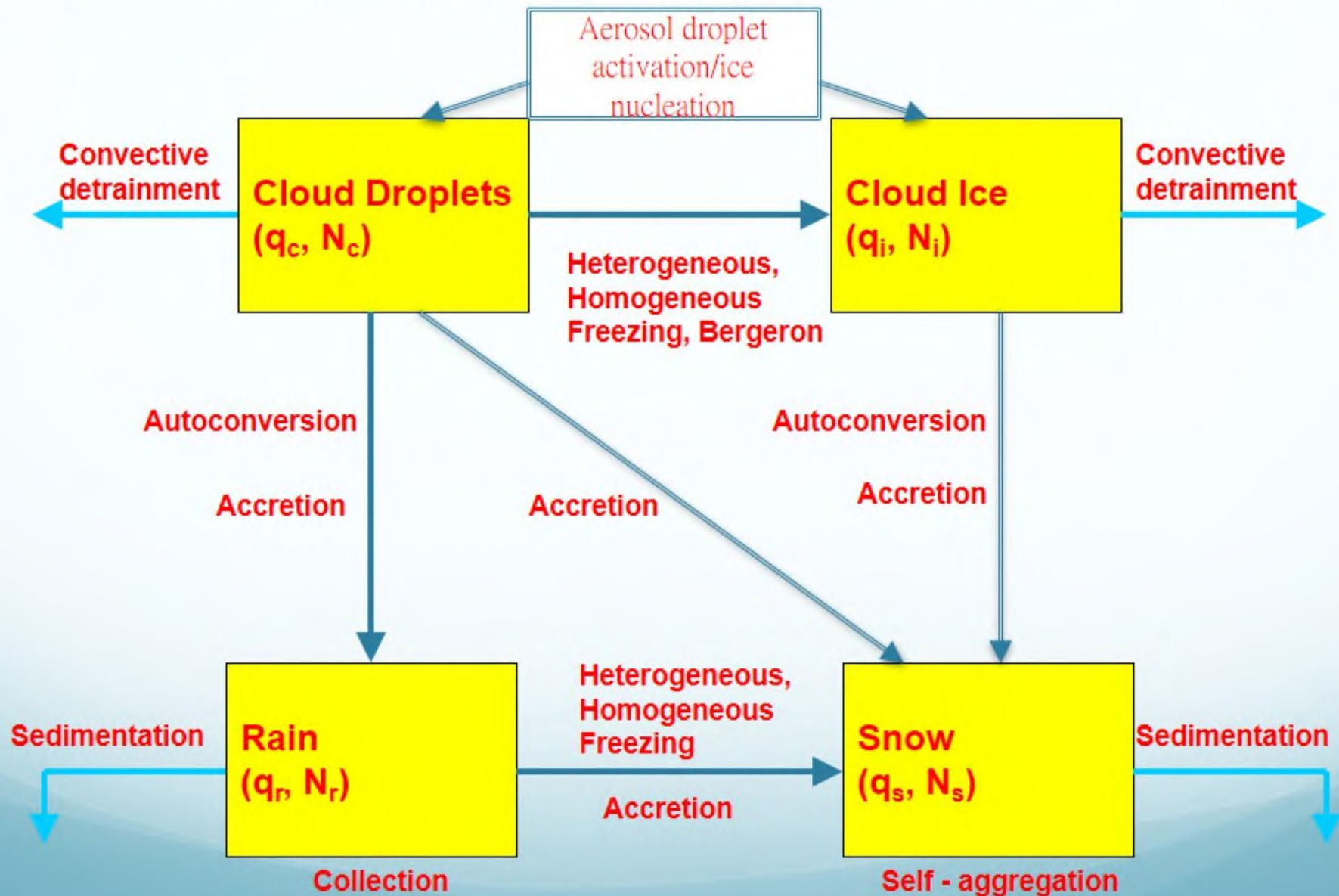
Two moment convective microphysics Scheme

- Two-moment (mass and number concentration), 4-species (ice, liquid water, snow and rain)
- Based on the Morrison-Gettelman microphysics scheme for large-scale clouds, with heavy modifications to fit convective clouds
- No subgrid variability is considered
- Ice nucleation of Liu and Penner (2005), Liu et al. (2007), dependent on vertical velocity and nuclei number

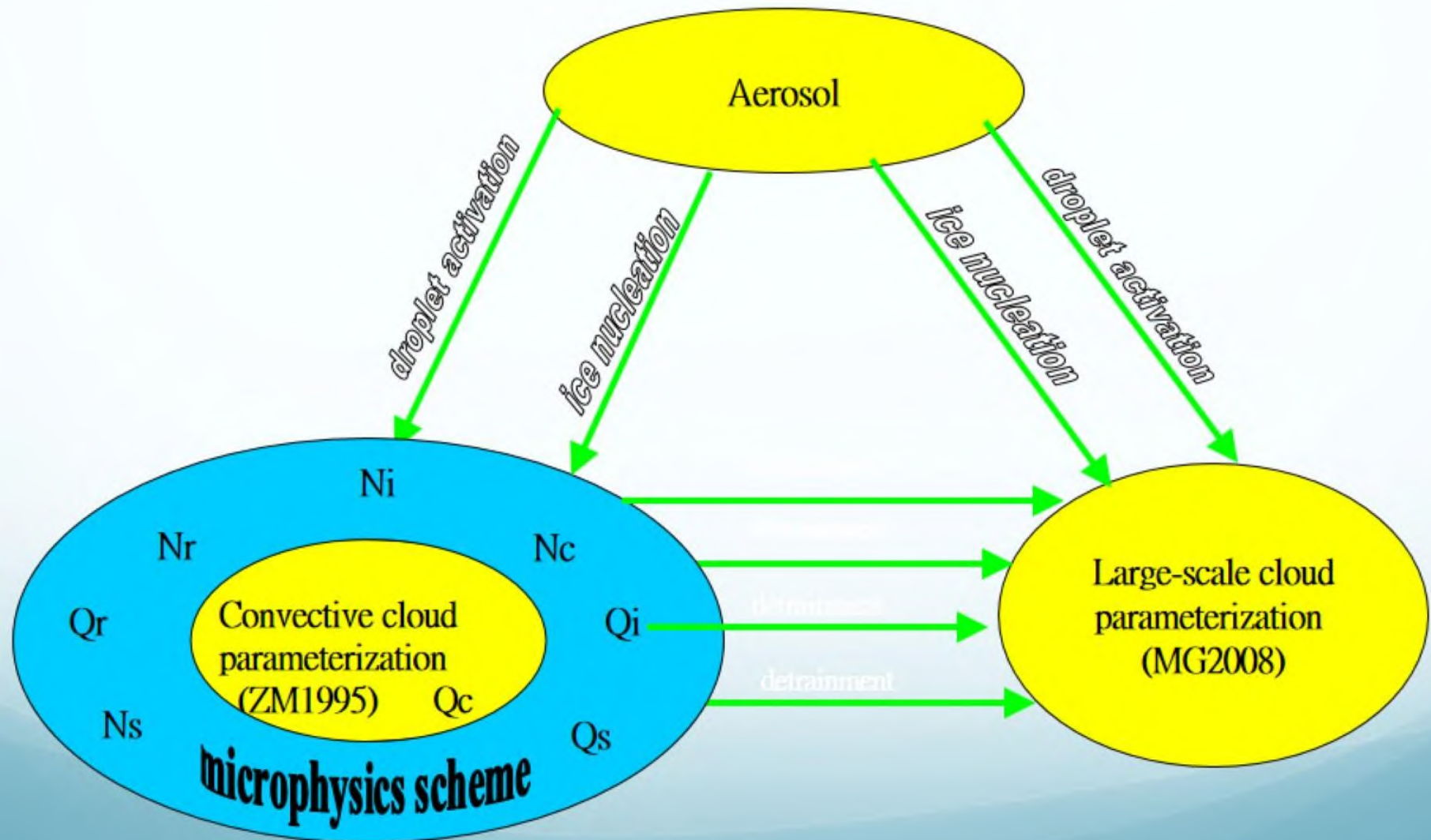
Two moment convective microphysics Scheme

- Contact freezing and Immersion freezing both depending on aerosol number
- Bergeron-Findeisen process: liquid water becomes ice in one time step if IWC > 0.5 mg/kg in mixed phase regime
- Updraft velocity considered in rain and snow budget equations in addition to terminal velocities
- All equations are diagnostic, same as for other thermodynamic fields in a steady-state 1-D cloud model

Two-moment microphysics scheme for convective clouds



Implementation of the microphysics scheme in the CAM5



$$\frac{\partial}{\partial z}(M_u q_x) = -D_u q_x + \sigma_u S_x^q$$

$$\frac{\partial}{\partial z}(M_u N_x) = -D_u N_x + \sigma_u S_x^N$$



where x refers to c for cloud water, i for cloud ice, s for snow and r for rain, $\sigma = M_u/w_u$ is convective cloud fraction. S_x is the source/sink per unit cloud area.

$$\frac{\partial K_u}{\partial z} = -\frac{v_u}{M_u}(1 + \beta C_d)K_u + \frac{1}{f(1 + \gamma)}g \frac{T_{v,u} - T_{v,e}}{T_{v,e}}$$

$$K_u = \frac{w_u^2}{2}$$

$$S_c^q = P_{cond}^{q_c} - P_{auto}^{q_c} - P_{accr}^{q_c} - P_{accs}^{q_c} - P_{fhet}^{q_c} - P_{fhm}^{q_c} - P_{Berg}^{q_c}$$

$$S_c^N = P_{nuc}^{N_c} - P_{auto}^{N_c} - P_{accr}^{N_c} - P_{accs}^{N_c} - P_{fhet}^{N_c} - P_{fhm}^{N_c}$$

$$S_i^q = P_{cond}^{q_i} - P_{auto}^{q_i} - P_{accs}^{q_i} + P_{fhet}^{q_c} + P_{fhm}^{q_c} + P_{Berg}^{q_c}$$

$$S_i^N = P_{nuc}^{N_i} - P_{auto}^{N_i} - P_{accs}^{N_i} + P_{fhet}^{N_c} + P_{fhm}^{N_c}$$

$$S_r^q = P_{auto}^{q_c} + P_{accr}^{q_c} - P_{accs}^{q_r} - P_{fhet}^{q_r} - P_{fhm}^{q_r} - P_{fallout}^{q_r}$$

$$S_r^N = P_{auto}^{N_c} + P_{accr}^{N_c} - P_{accs}^{N_r} - P_{fhet}^{N_r} - P_{fhm}^{N_r} - P_{fallout}^{N_r} + P_{aggr}^{N_r}$$

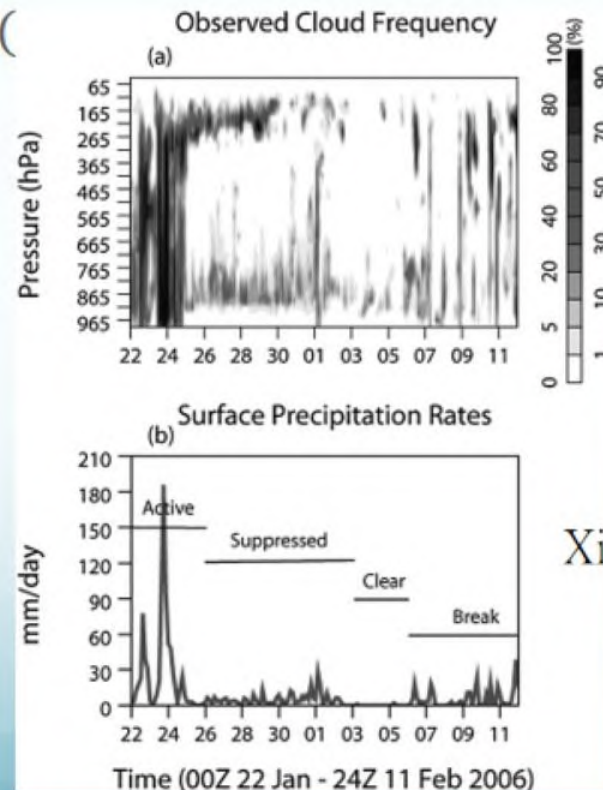
$$S_s^q = P_{auto}^{q_i} + P_{accs}^{q_i} + P_{accs}^{q_r} + P_{accs}^{q_c} + P_{fhet}^{q_r} + P_{fhm}^{q_r} - P_{fallout}^{q_s}$$

$$S_s^N = P_{auto}^{N_i} + P_{accs}^{N_i} + P_{accs}^{N_r} + P_{accs}^{N_c} + P_{fhet}^{N_r} + P_{fhm}^{N_r} - P_{fallout}^{N_s} + P_{aggs}^{N_s}$$

Song and Zhang (JGR, 2011)

TWP-ICE SCM Simulation

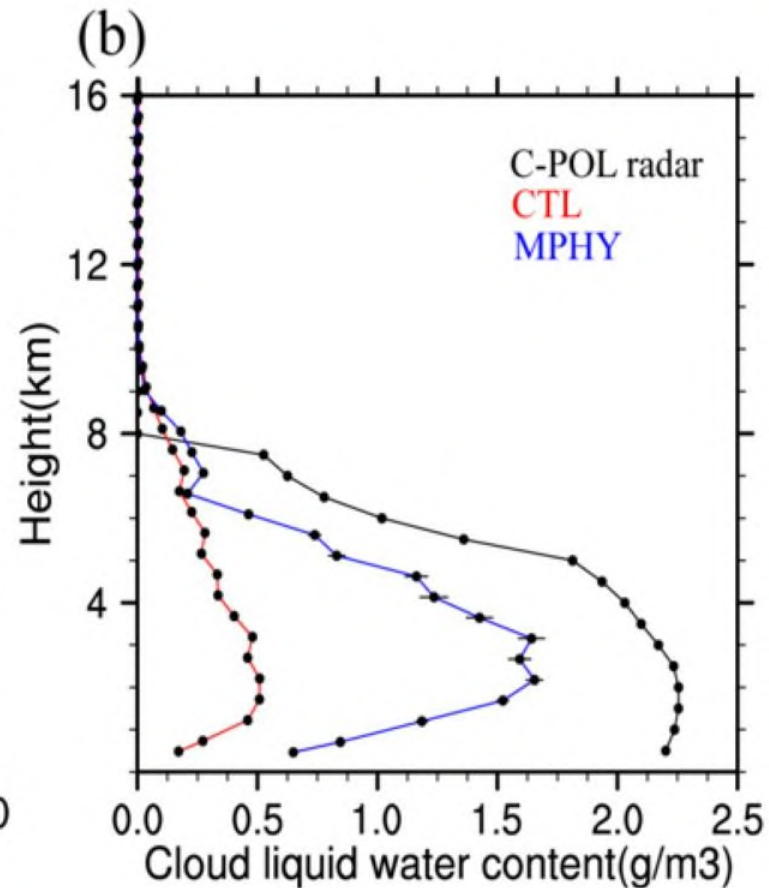
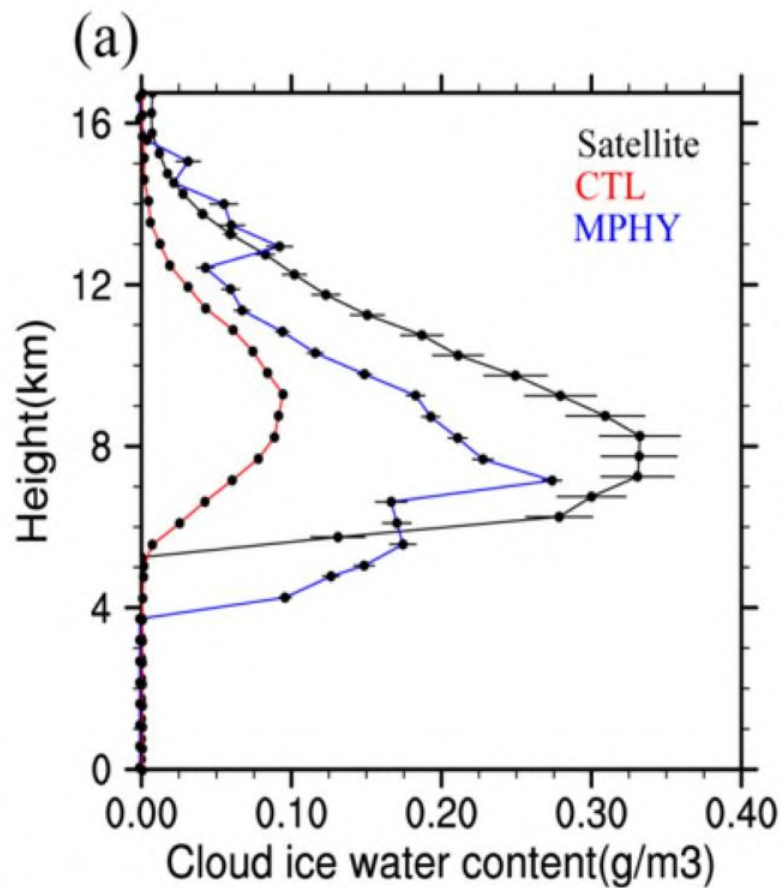
- Jan. 19-Feb. 12, 2006 at Darwin, Australia
- NCAR CAM3.5 SCM
- Large-scale Forcing Data by DC



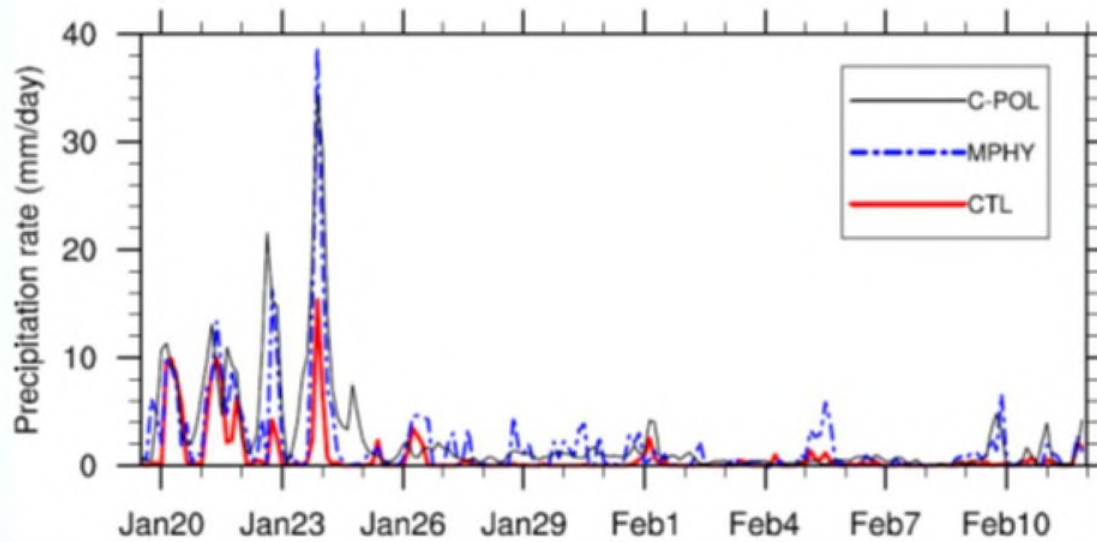
Xie et al. 2010

Convective cloud ice during active monsoon period

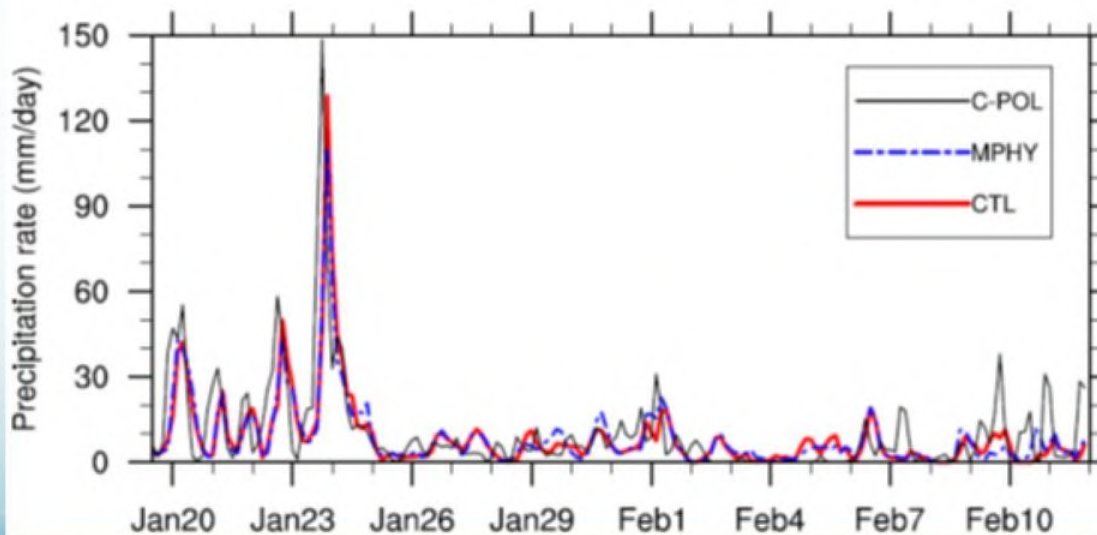
TWP-ICE

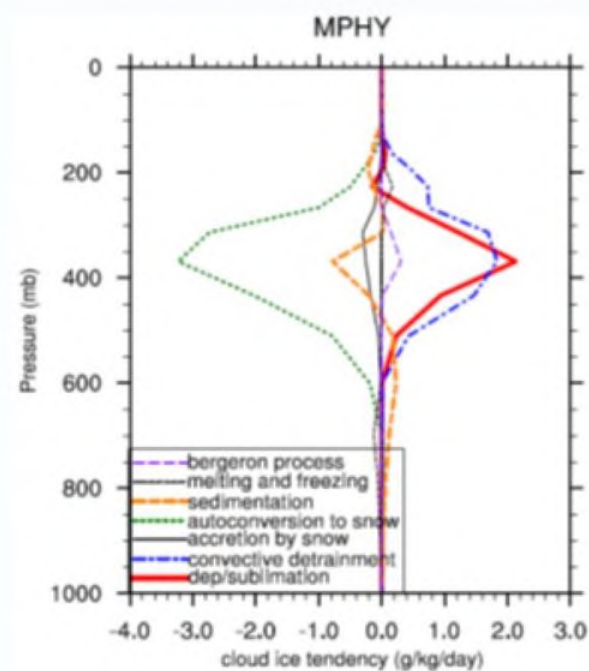
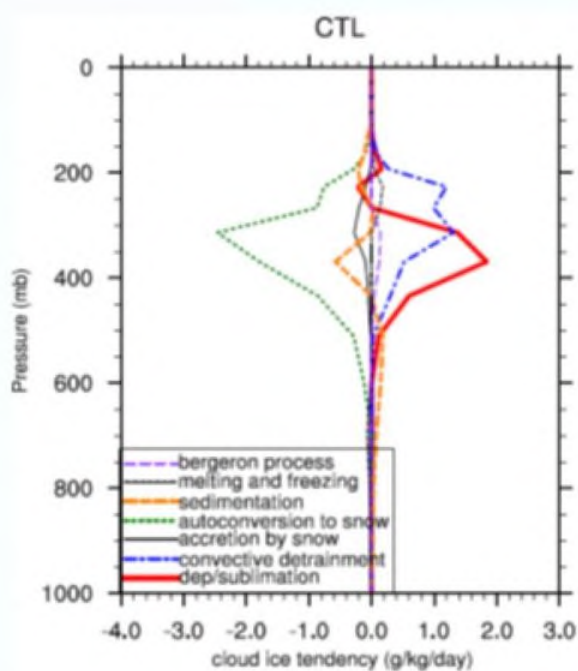


Stratiform precipitation rate

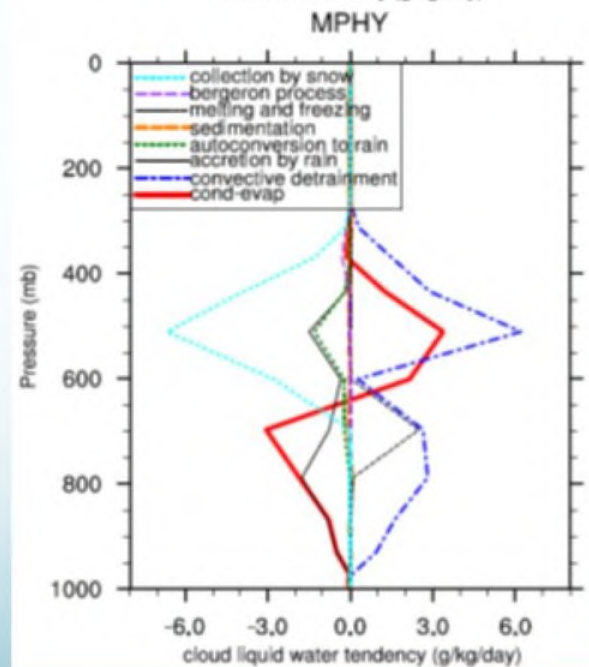
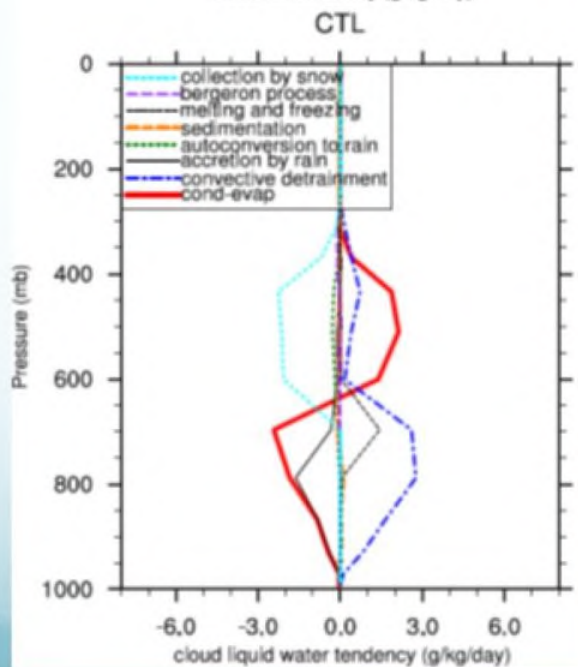


Convective precipitation rate





Large-scale
cloud ice
budget



Results from CAM5

CAM5 Experiments

- Two 6-yr NCAR Community Atmosphere Model v5 (CAM5) simulations are conducted:

CTL: standard CAM5

MPHY: with convective microphysics scheme

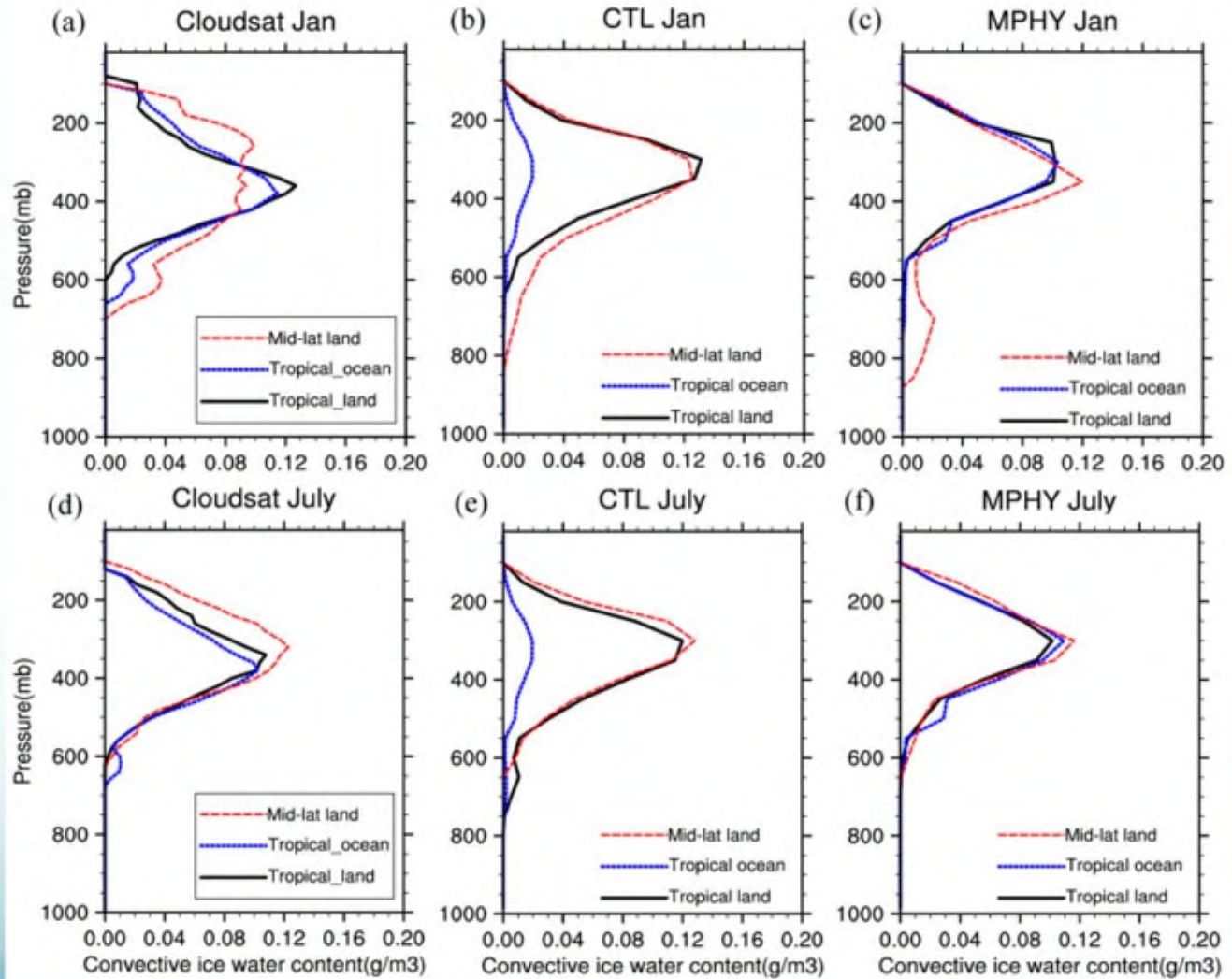
For aerosol-convection expt: 1xaerosol and 10xaerosol runs

- Resolution: 1.9° X 2.5° , 30 levels
- Forcing: climatological aerosol and SST data.
The prescribed aerosol dataset has 12 species including sulfate, sea salt, dust, black carbon, etc.
- Results from the last 5-year simulations are presented here

Evaluation of the microphysics scheme with observations

Ice Water Content (IWC)

Jan



July

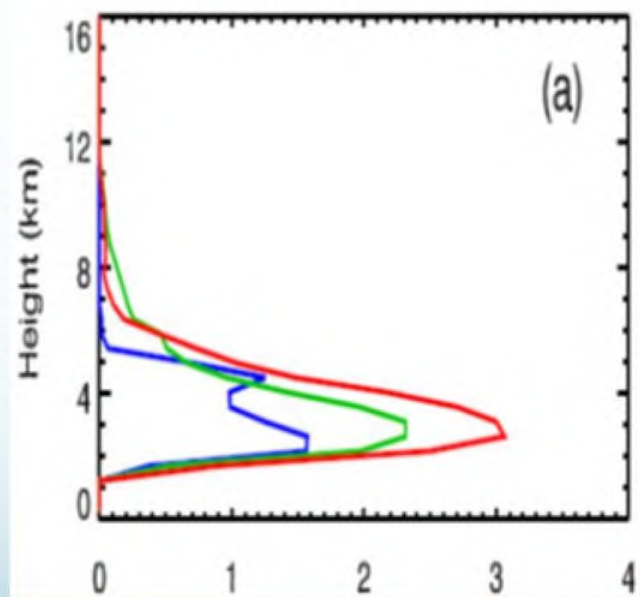
CloudSat

CTL

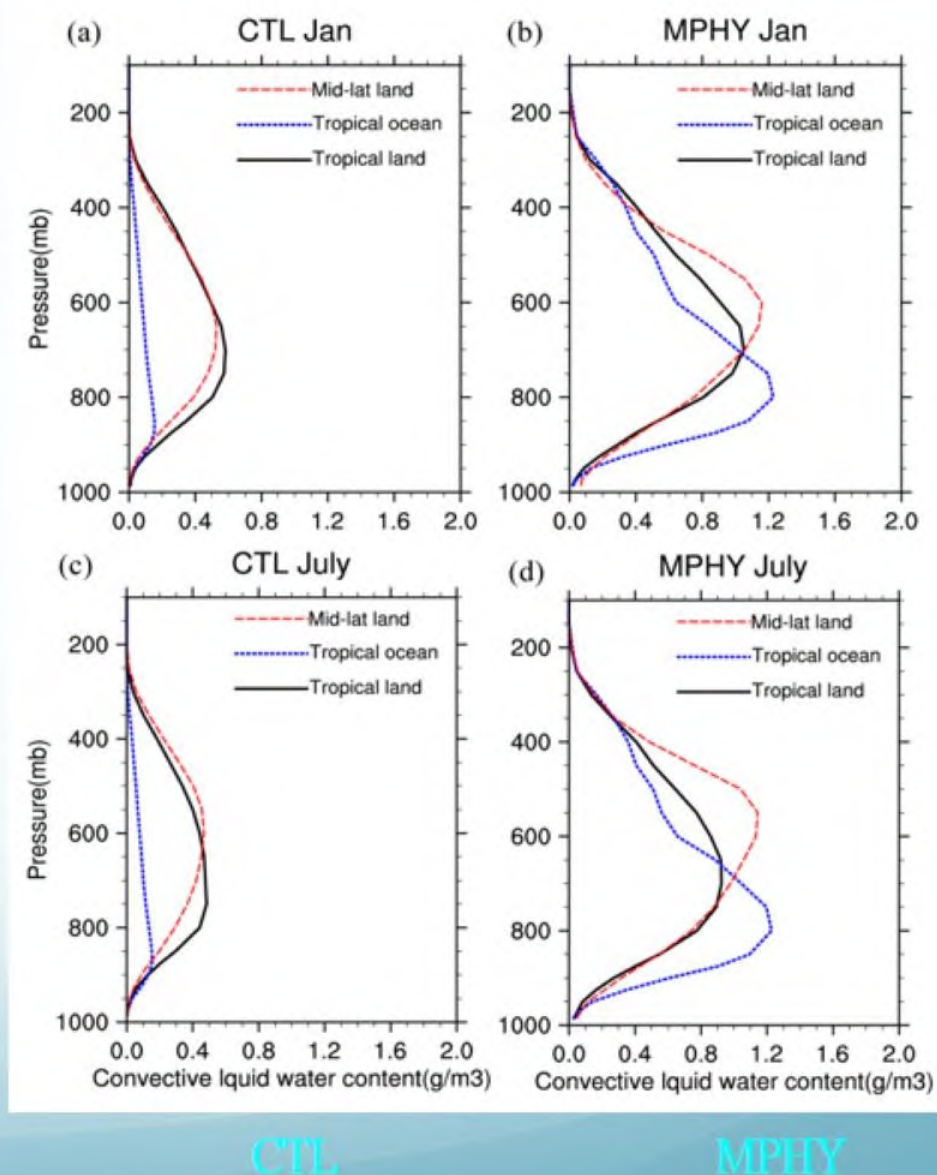
MPHY

Cloud Liquid Water Content (LWC)

Statistics of aircraft observations show the characteristic value for dense cumulus congestus and cumulonimbus is 0.5 to 3 g m^{-3} (Borovikov et al. 1963)

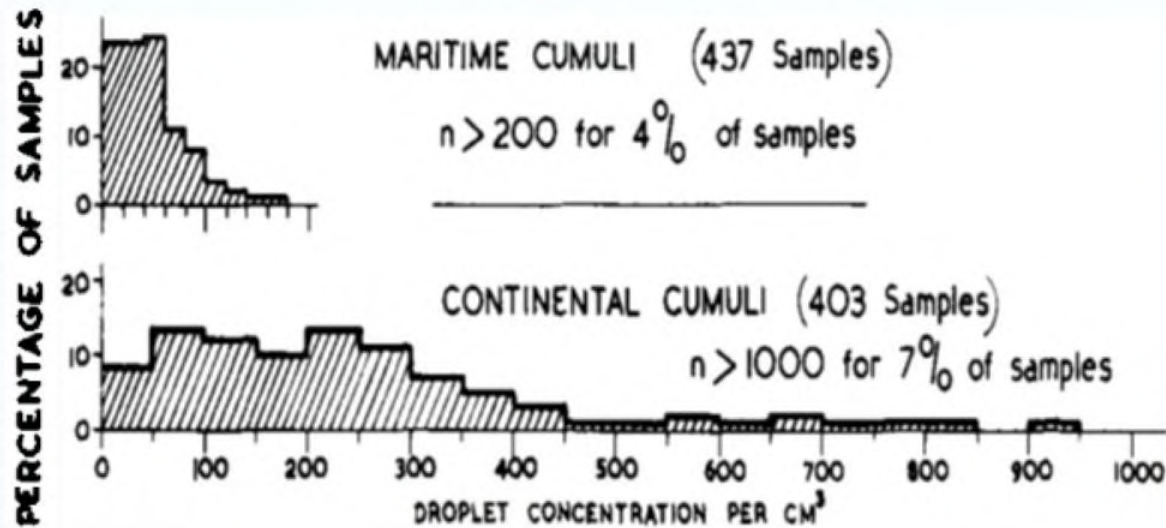


LWC simulated by WRF model under low (blue), intermediate (green) and high (red) CCN, Li et al. 2008, *J. Geophys. Res.*



Cloud Droplet Number Concentration

OBS



Squires, 1958, *Tellus*

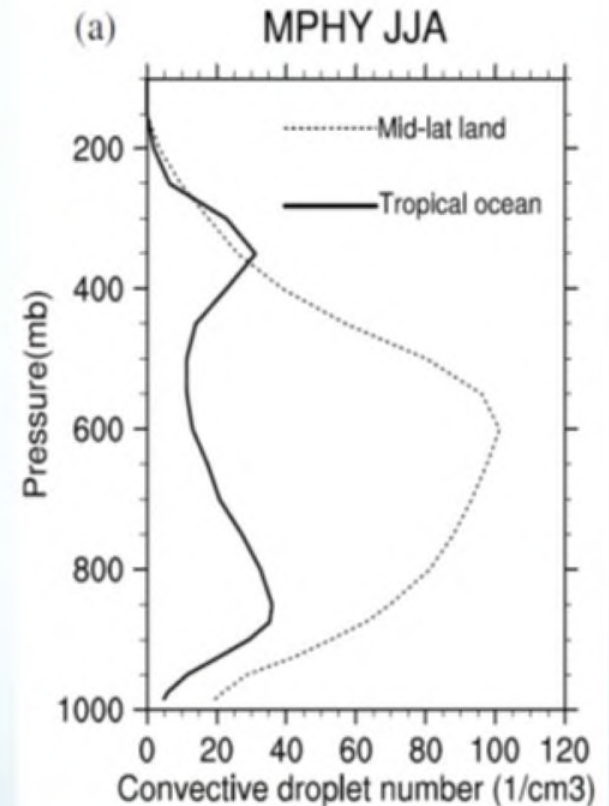
Maritime cumuli:

20~60 per cm³ (Wood et al., 2011)

Continental cumuli:

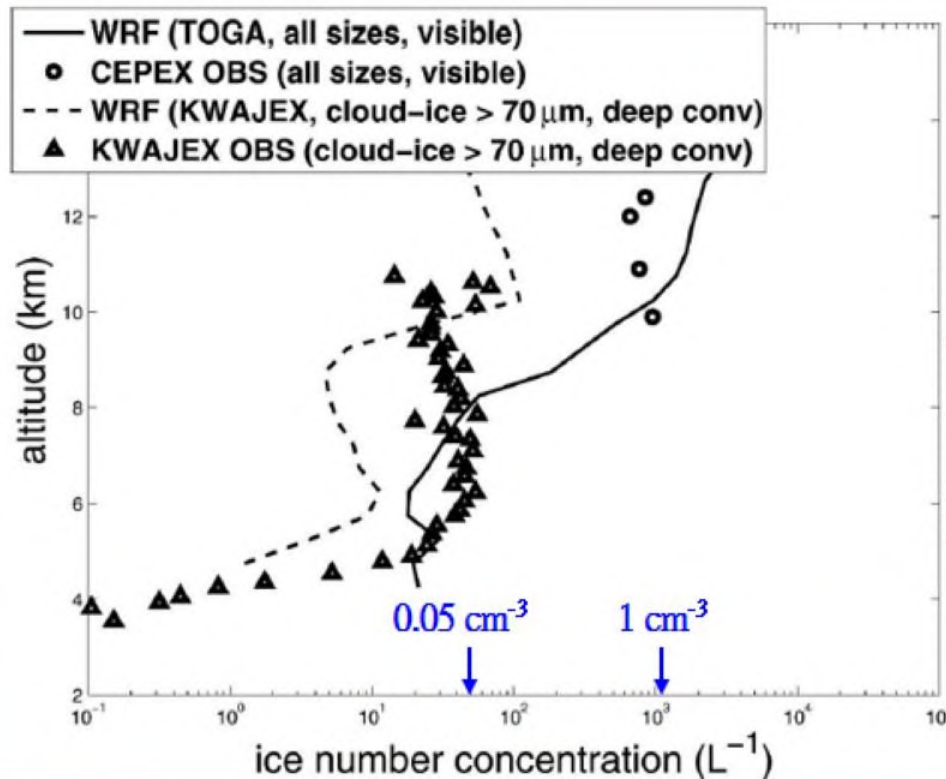
50~300 per cm³ or even higher

CAM5

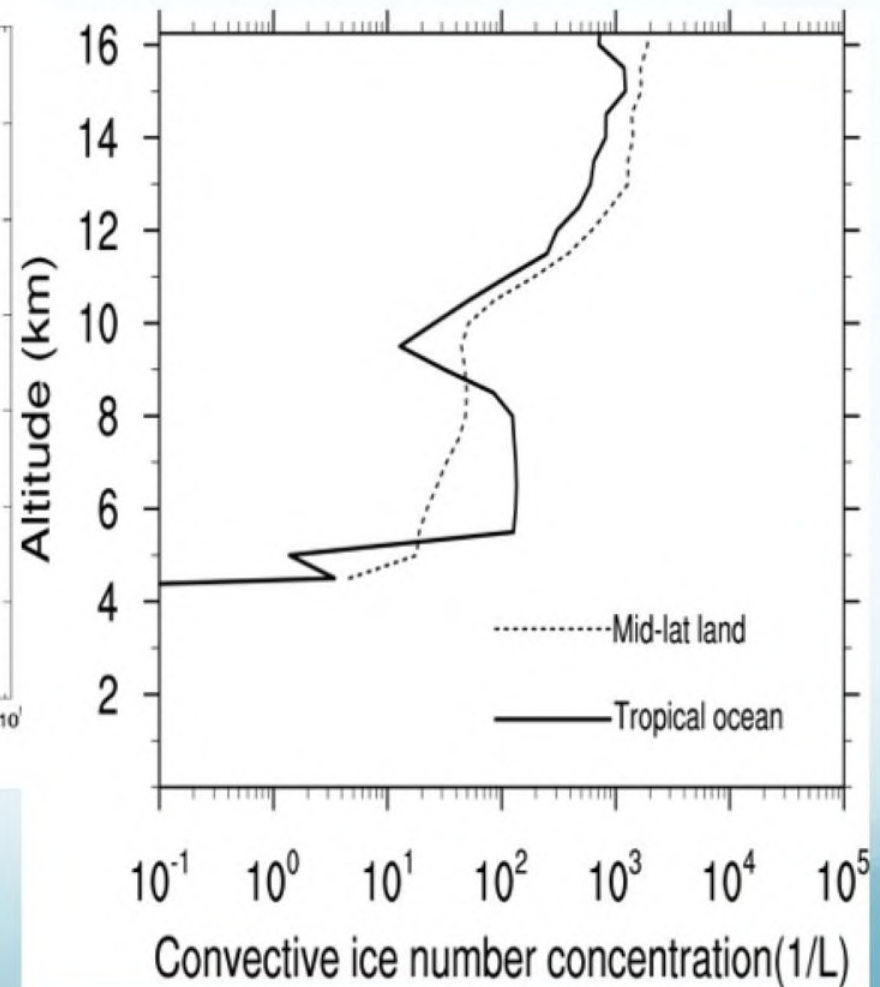


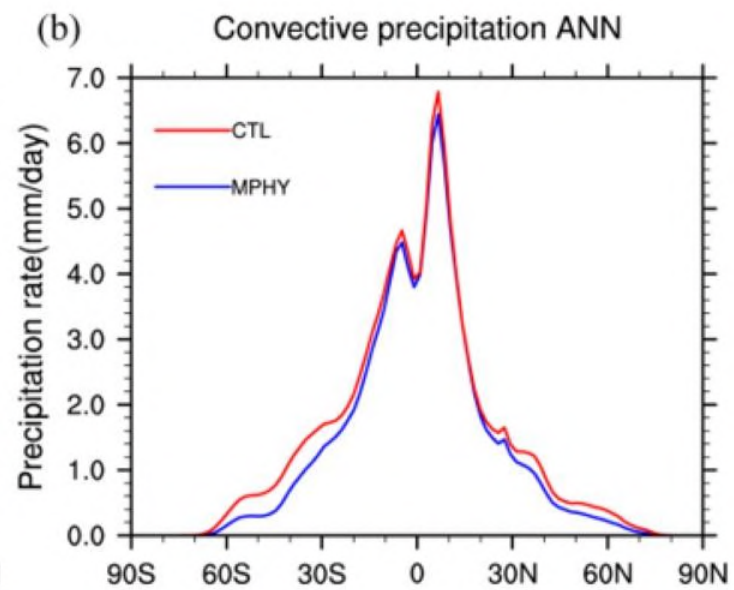
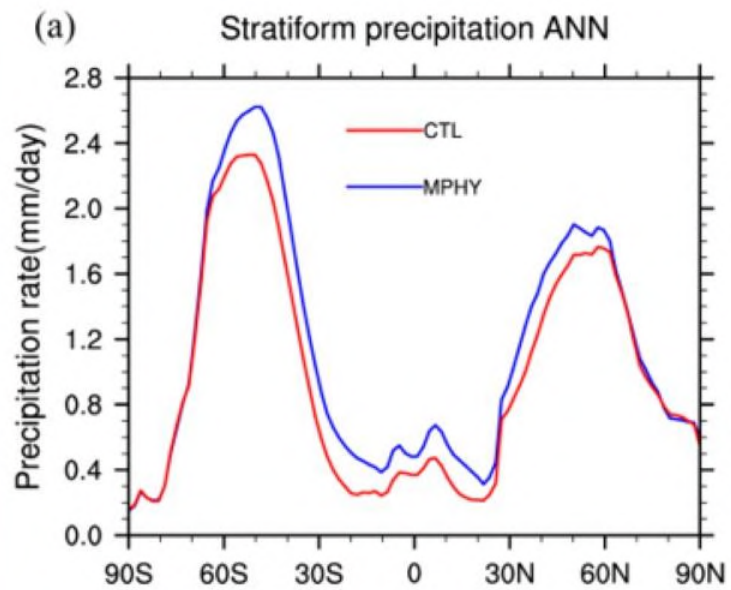
Cloud Ice Crystal Number Concentration

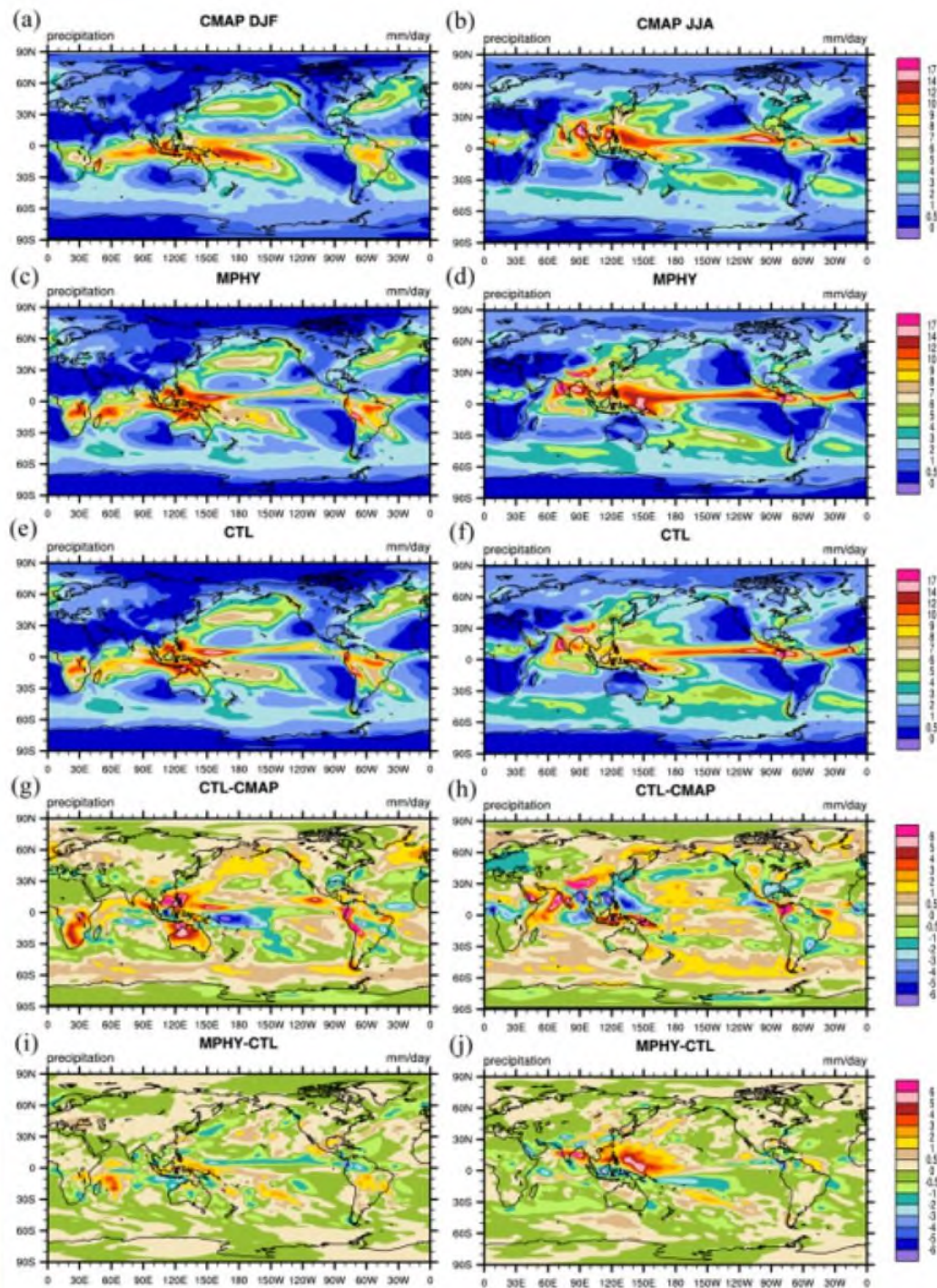
OBS

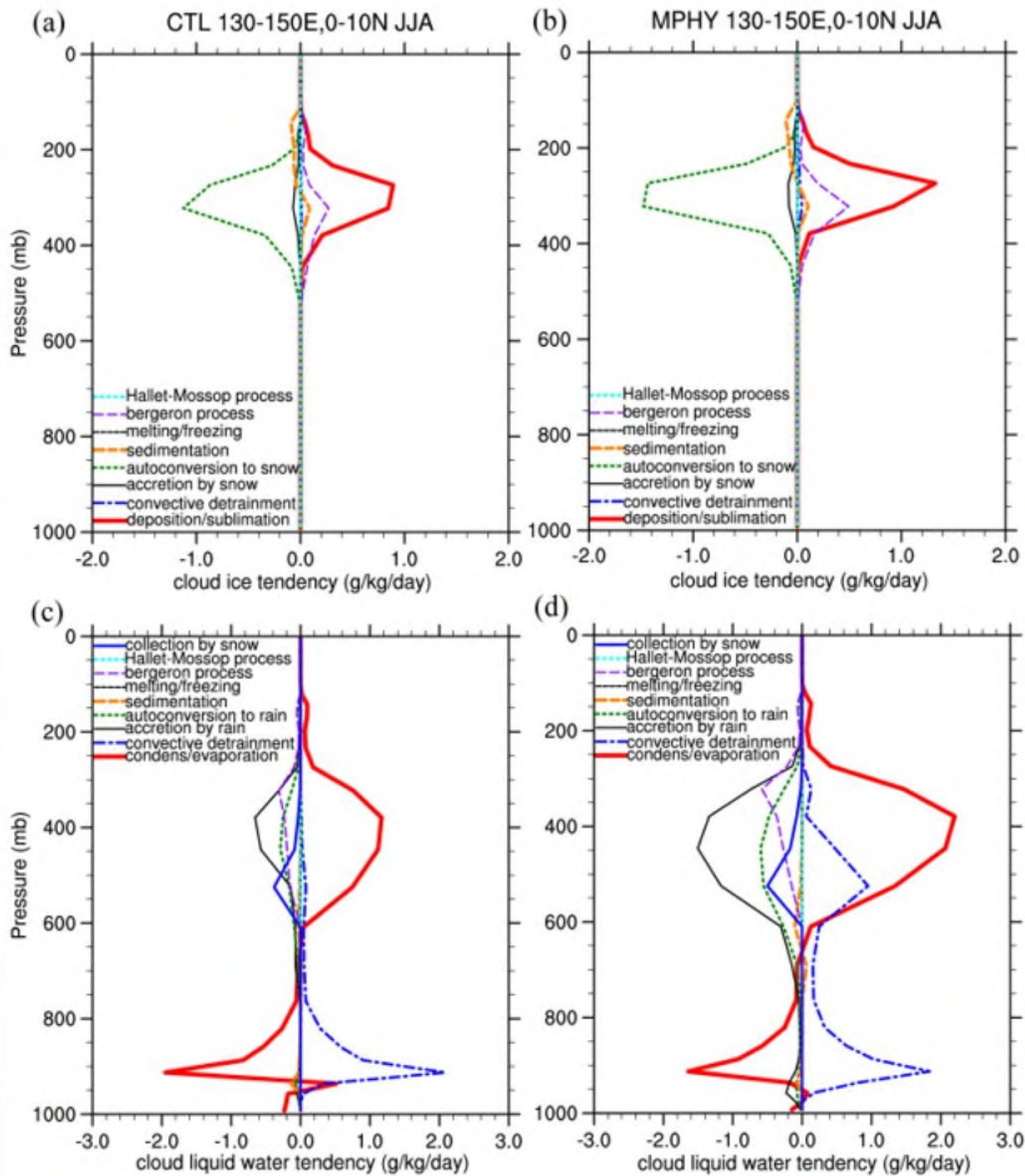


CAM5

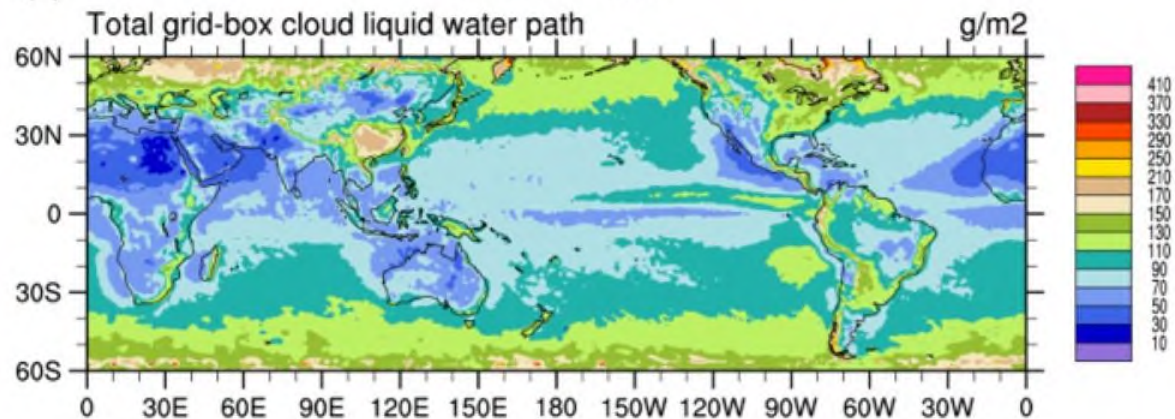




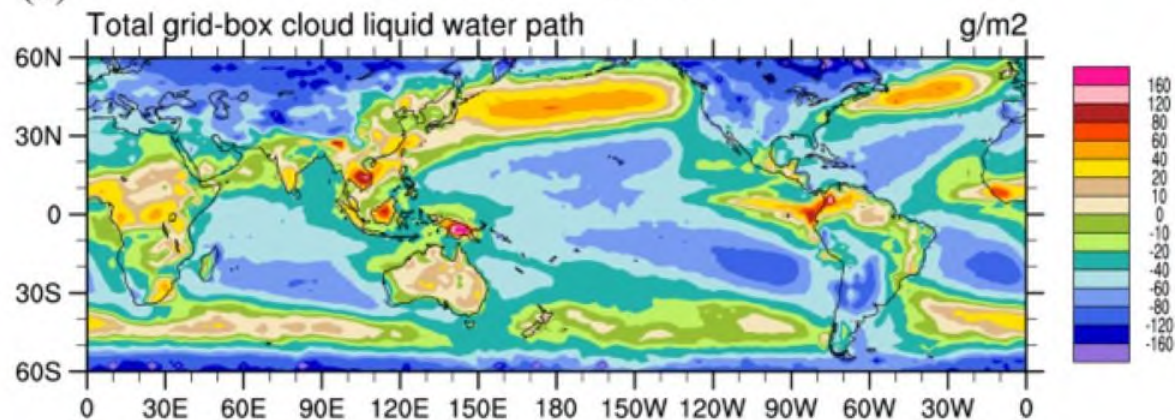




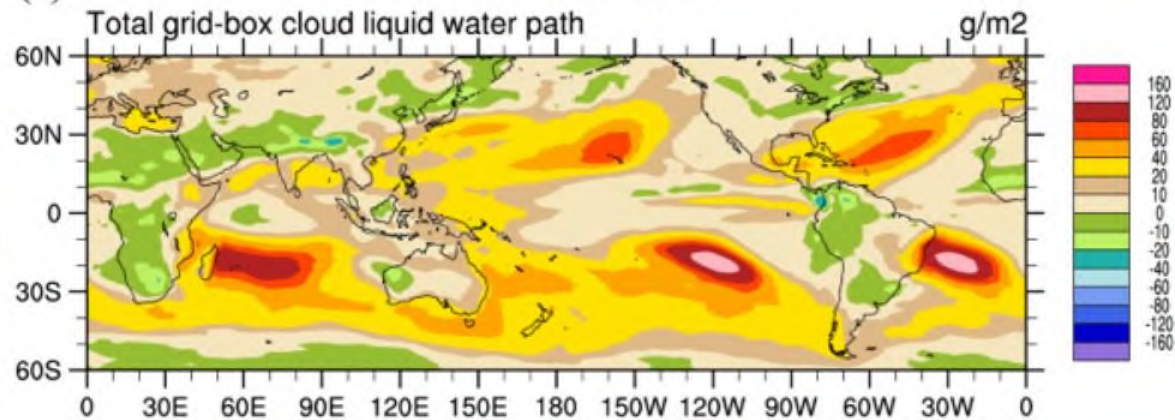
(a) **MODIS ANN**



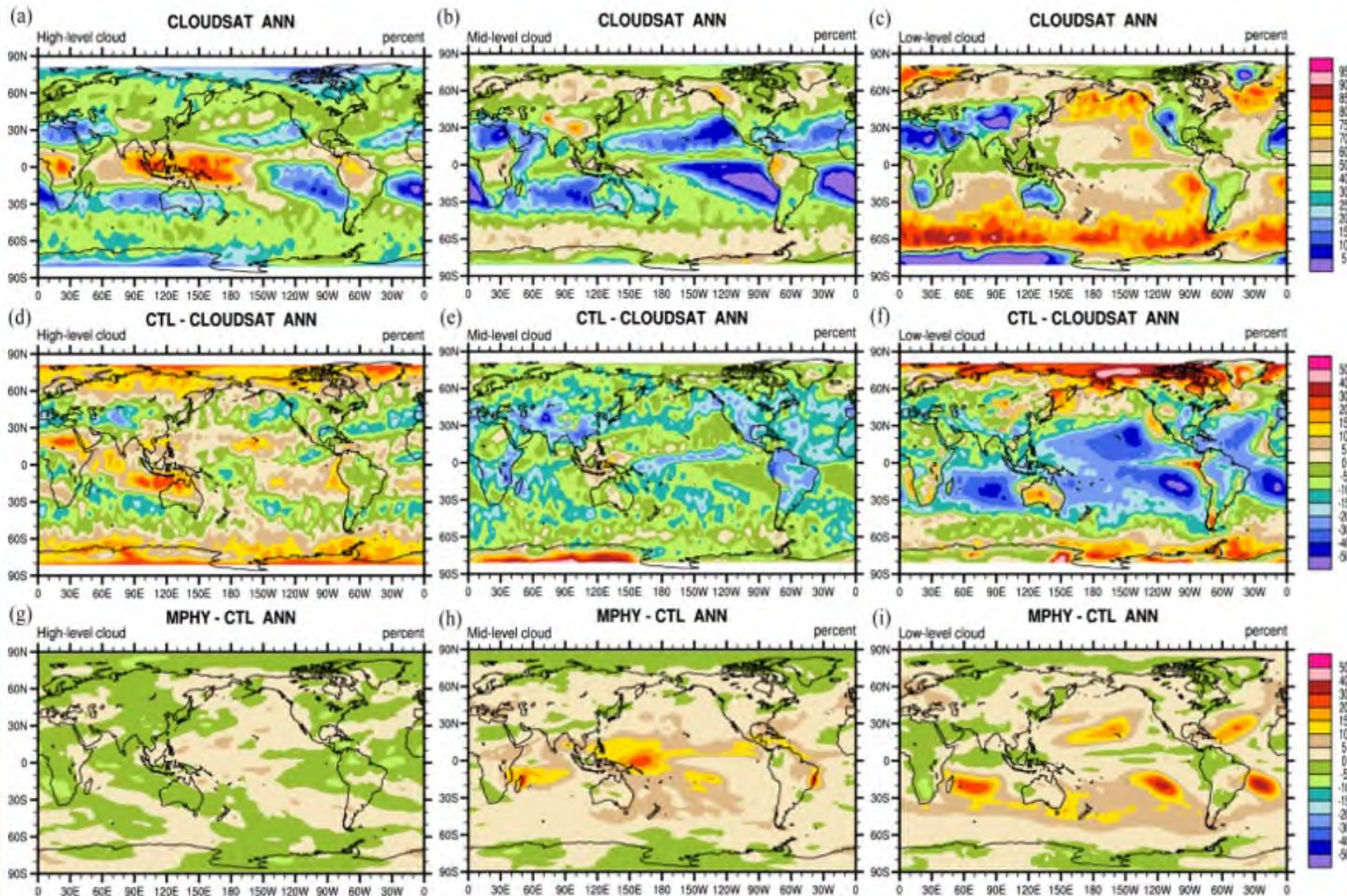
(b) **CTL - MODIS ANN**

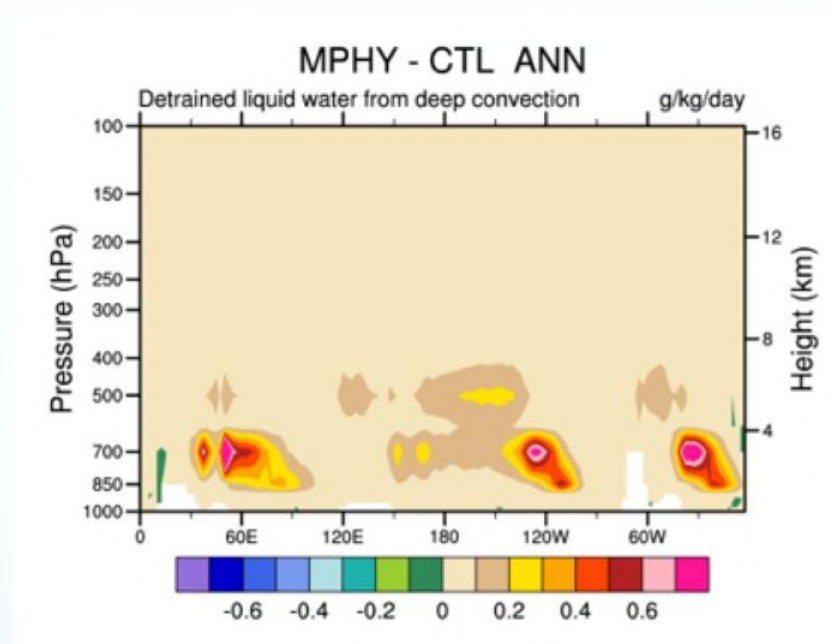


(c) **MPHY - CTL ANN**

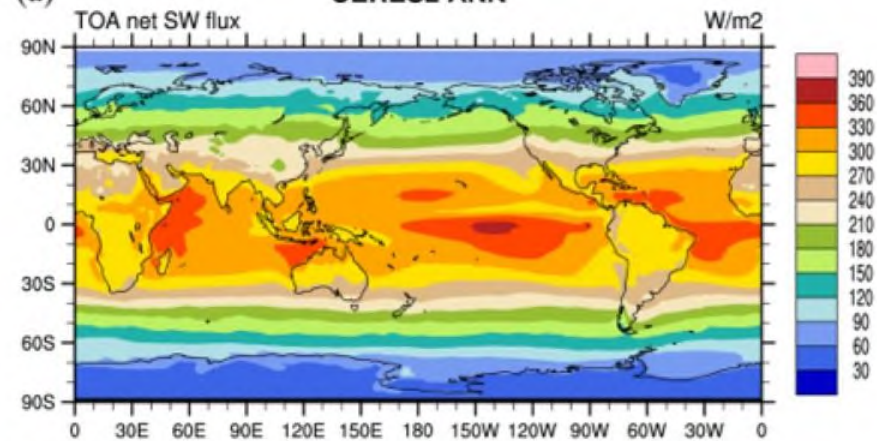


Cloud Amount

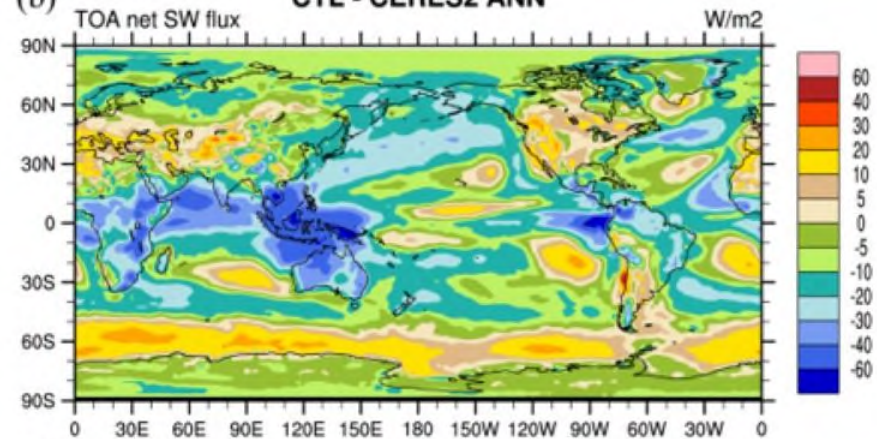




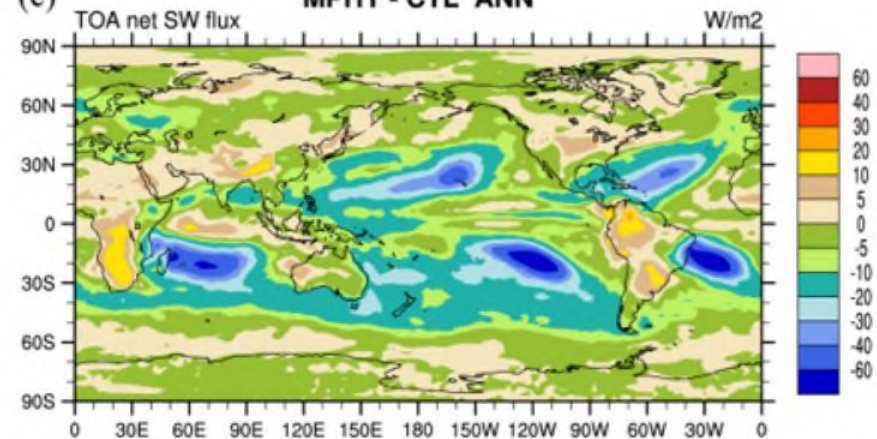
(a) CERES2 ANN



(b) CTL - CERES2 ANN



(c) MPHY - CTL ANN



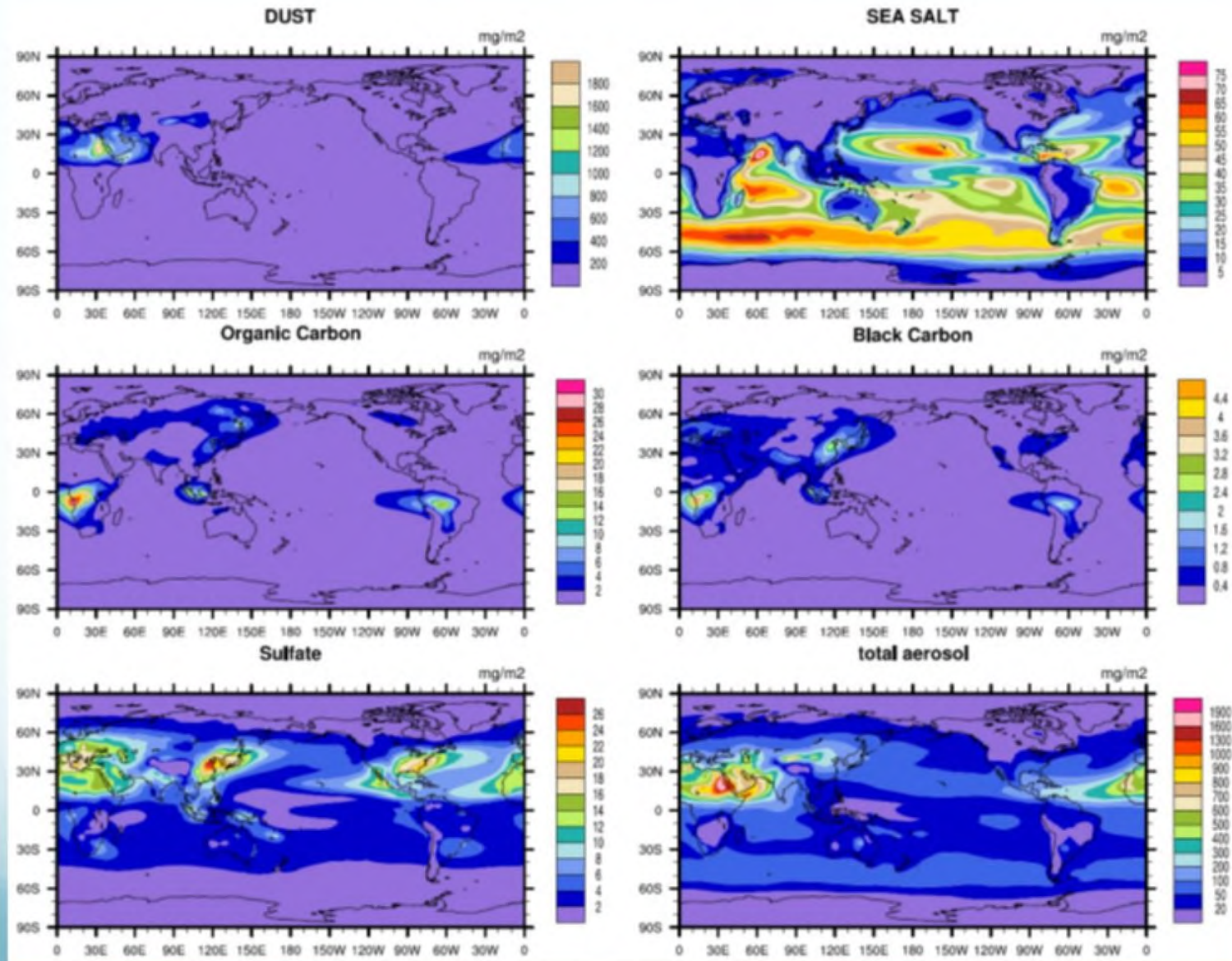
Aerosol Effects on Convection

- 1) 1 x climatological aerosols
- 2) 10 x climatological aerosols to diagnose its one-time-step effect on convection within the 1xaerosol simulation
- 3) 10 x climatological aerosols to predict convection, which feeds back to large-scale fields

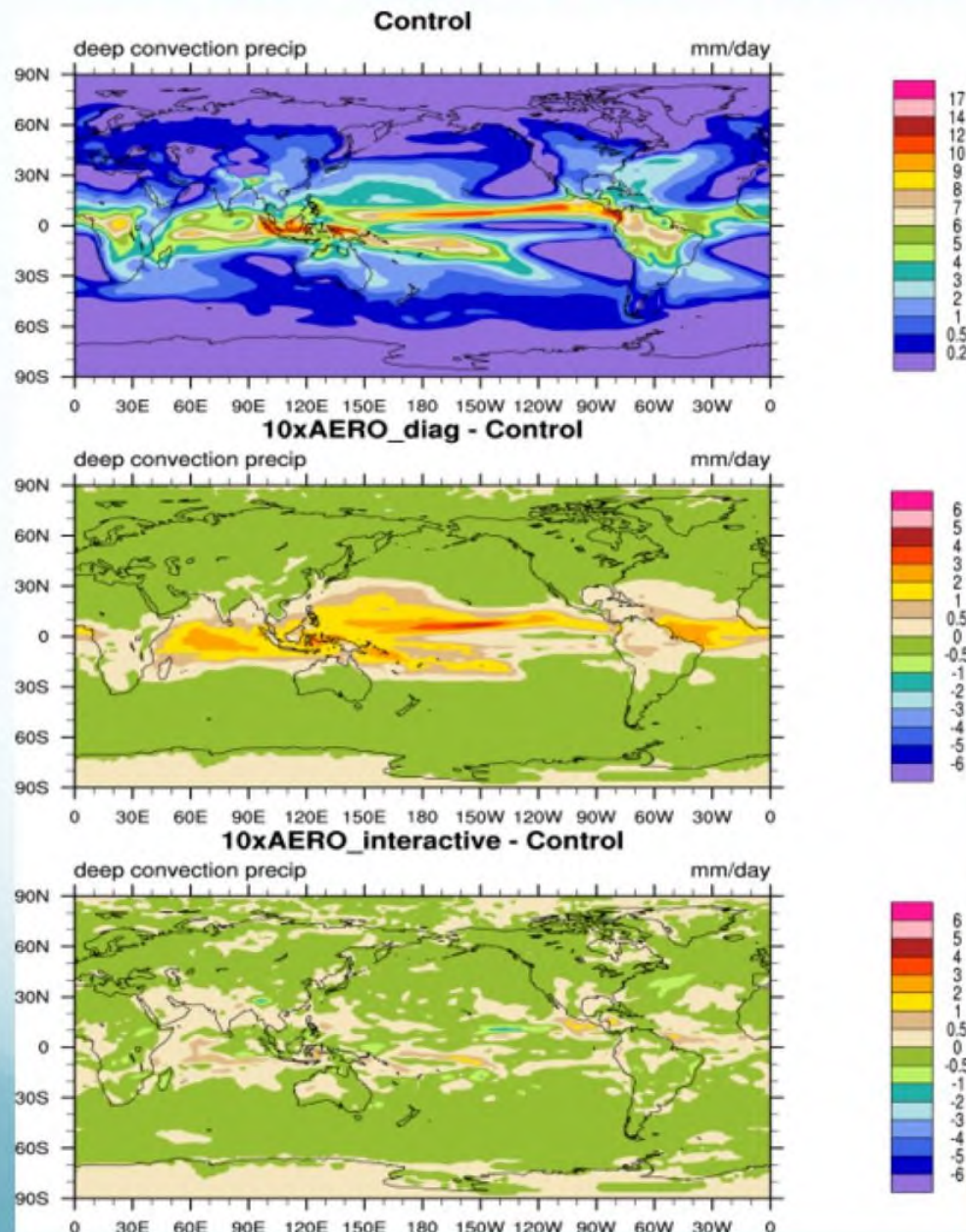
Therefore,

- (2)-(1) gives the **local thermodynamic** effect of aerosols on convection under the **same** large-scale conditions;
- (3)-(1) gives the **total** effect including dynamic feedback.

Global Aerosol Distribution



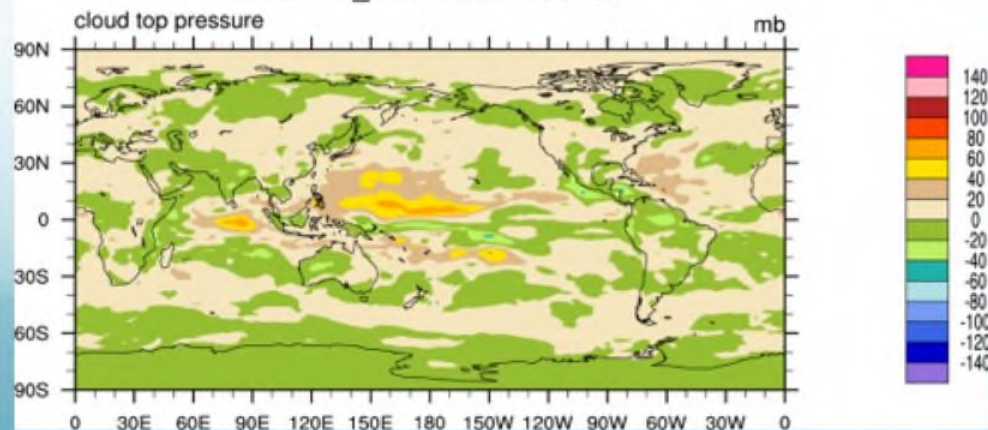
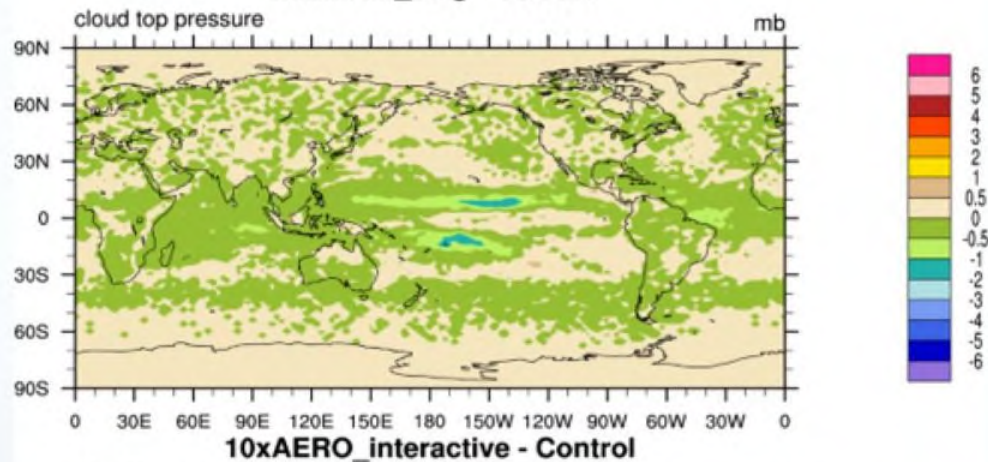
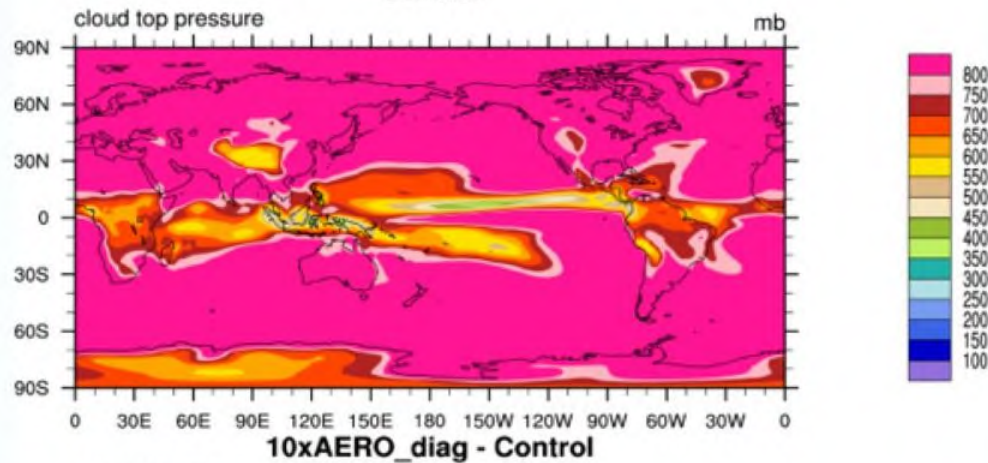
Impacts on global climate simulation



Under the same dynamic and thermodynamic conditions, aerosols invigorate convection.

However, when the invigorated convection is allowed to interact with its environment, the net effect of aerosols is small, and can even change the sign!

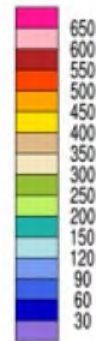
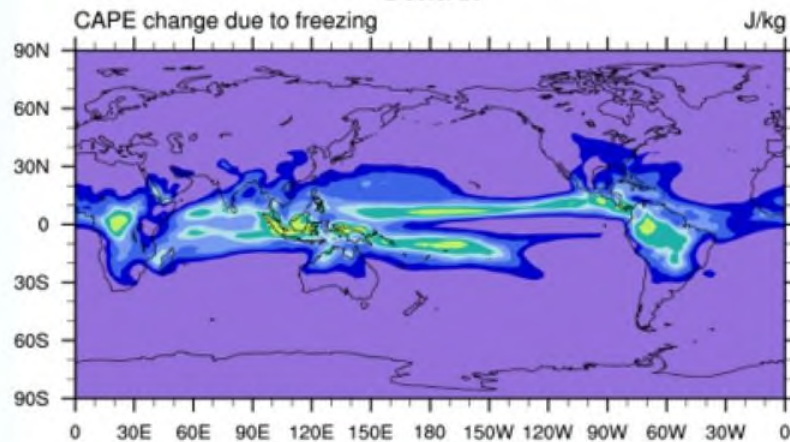
Control



Aerosols make
convective cloud top
higher

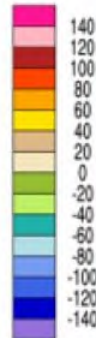
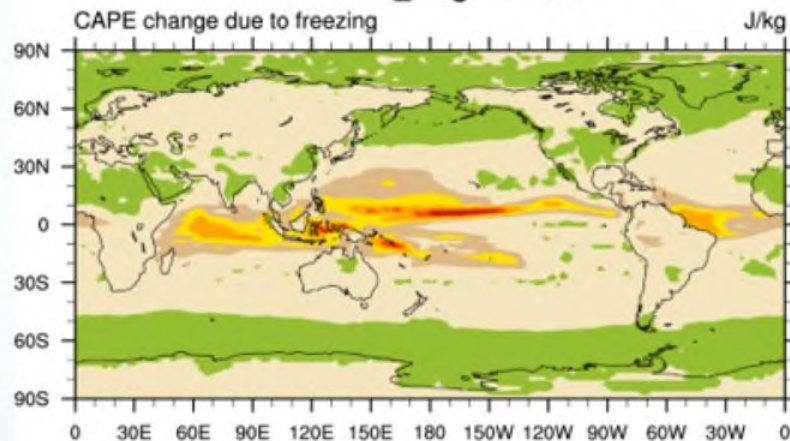
Interaction with
convection environment
is the game changer.

Control



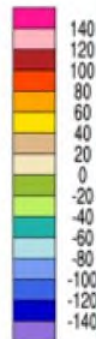
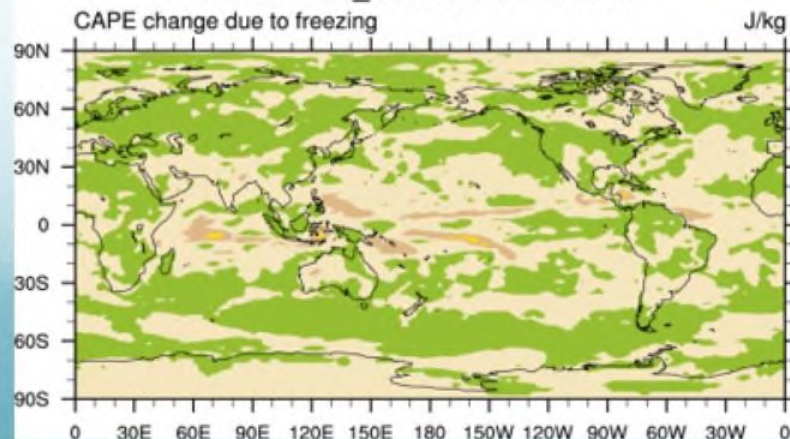
Freezing increases
atmospheric CAPE

10xAERO_diag - Control



More aerosols lead to more
freezing, thus more CAPE

10xAERO_interactive - Control



Interaction with the
environment largely negates
the changes.

Summary

- A new microphysics scheme is introduced in convection parameterization
- It produces a larger amount of cloud IWC and LWC in convective updrafts
- As a result, contribution to large-scale cloud ice and water budget is significantly enhanced, leading to more large-scale precipitation
- Results from CAM5 show improvements in model simulation of cloud microphysical properties, precipitation and radiation

Summary (cont' d)

- The introduction of convective microphysics scheme enables climate models to investigate aerosol-convection-climate interaction.
- For a given environmental condition, aerosols invigorate convection (more CAPE, deeper convection and more convective precipitation)
- However, if the convective environment is allowed to respond to the invigoration of convection, aerosol effect is much reduced and can even change the sign.