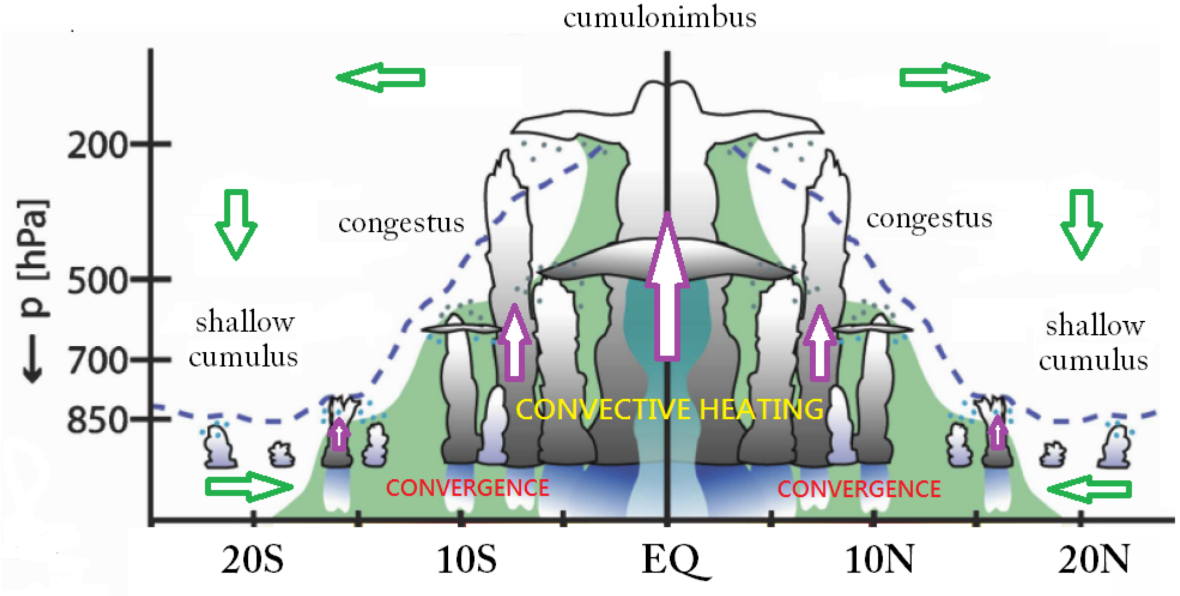
A useful thermodynamic framework for the diagnosis of tropical climate variability

余嘉裕

中央大學 大氣科學系

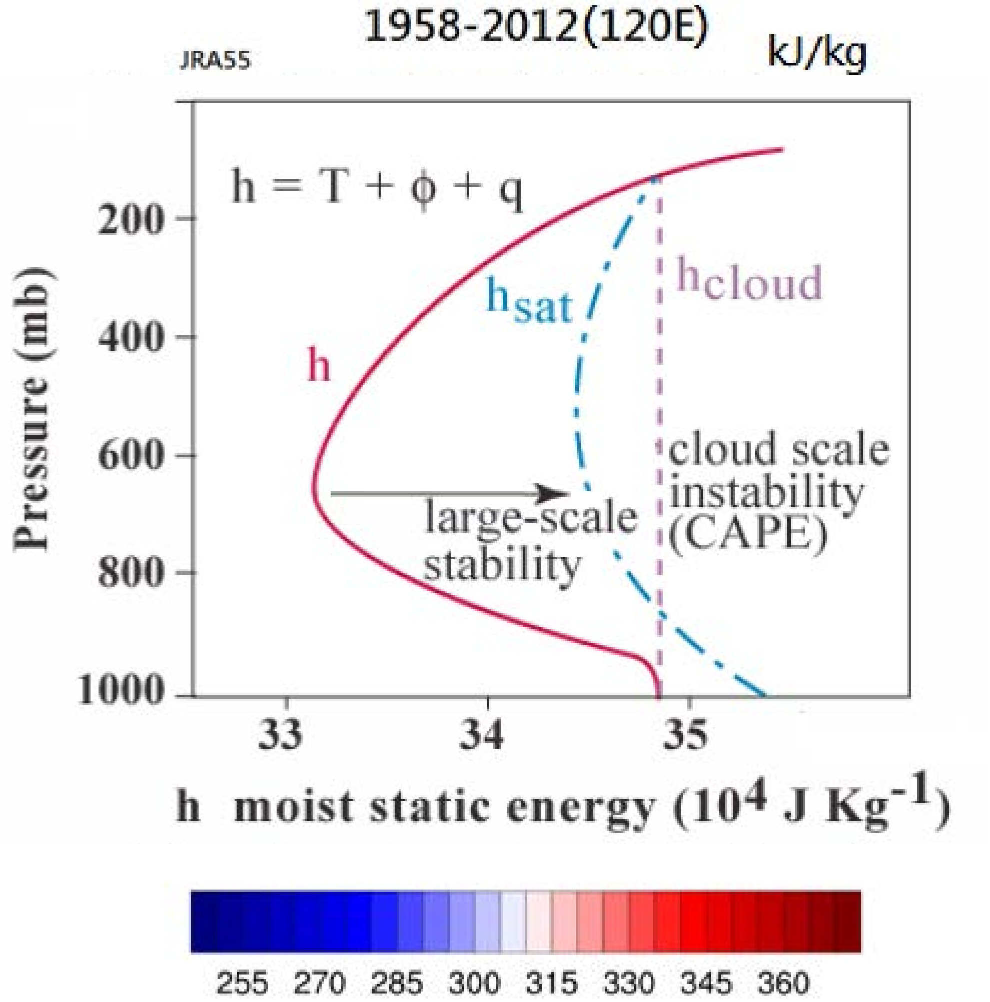
(October 5, 2016 at CWB)

Motivation



In the tropics, large-scale flows introduce instability into atmospheric column while convection activities act to release the column instability through a quick convective overturning, with the latter typically lasting only a few hours—a mutual balance must occur between the two very different scales.

MSE profile & CAPE



Moist Static Energy Budget

$$(\partial_t + D_{adv})S + \omega \partial_p S = Q_c + g \partial_p F_T \qquad (1)$$

$$(\partial_t + D_{adv})q + \omega \partial_p q = Q_q + g \partial_p F_q \qquad (2)$$

where F_T and F_q (in w/m2) represents heat and moisture fluxes, respectively; Q_c and Q_q represent "convective heating" and "moisture sink", respectively. $S = T + gz$ is the "dry static energy" and D_{adv} is the mean flow advection. T and q are in *J/kg* , and $\kappa = R / C_p$.

- (1)+(2) yields the following moist static energy equation:

$$(\partial_t + D_{adv})h + \omega\partial_p h = Q_c + Q_q + g\partial_p F_T + g\partial_p F_q \qquad (3)$$

where $h=T+gz+q$ is the "moist static energy" in meteorology or the "moist entropy" in physics.

- Vertically integrating (3) from the top of atmosphere (P_T) to surface (P_0), one gets the moist static energy budget:

$$\boxed{\left\langle (\partial_t + D_{adv})h \right\rangle + \left\langle \omega\partial_p h \right\rangle = F^{net} \qquad (4)}$$

where

$$\left\langle (\ \ ) \right\rangle = g^{-1}\int_{p_T}^{p_0} (\ \ )\, dp$$

$$F^{net} = SW^{net} + LW^{net} + H + E$$

- Vertically integrating (2), one gets the moisture budget:

$$\boxed{\left\langle (\partial_t + D_{adv})q \right\rangle + \left\langle \omega \partial_p q \right\rangle = E - P} \quad (5)$$

- Vertical integrating (1), one gets the dry static energy budget:

$$\boxed{\left\langle (\partial_t + D_{adv})S \right\rangle + \left\langle \omega \partial_p S \right\rangle = SW^{net} + LW^{net} + H + P} \quad (6)$$

- In deriving (4), we've employed the following convective closure:

$$\int_{p_T}^{p_0} Q_c \frac{dp}{g} + \int_{p_T}^{p_0} Q_q \frac{dp}{g} = 0$$

- Under convective quasi-equilibrium constraints (Yu and Neelin 1997; Yu et al. 1998; Neelin and Zeng 2000), the vertical advection of moist static energy, moisture, and dry static energy by convection can be approximated by

$$\langle \omega \partial_p h \rangle \approx \frac{\Delta p}{g} M \, \nabla \cdot \mathbf{V_T}$$

$$-\langle \omega \partial_p q \rangle \approx \frac{\Delta p}{g} M_q \, \nabla \cdot \mathbf{V_T}$$

$$\langle \omega \partial_p S \rangle \approx \frac{\Delta p}{g} M_s \, \nabla \cdot \mathbf{V_T}$$

where Δp is the depth of troposphere and $\mathbf{V_T}$ is the baroclinic wind.

- M , Mq, Ms (in J/Kg) denotes respectively the “gross moist stability” (GMS), “gross moisture stratification”, and “gross dry stability” of the tropical atmosphere, which are defined as

$$M = \Delta p^{-1} \int_{p_T}^{p_0} (\partial_p h)\, \Omega(p)\, dp$$

$$M_q = -\Delta p^{-1} \int_{p_T}^{p_0} (\partial_p q)\, \Omega(p)\, dp$$

$$M_s = \Delta p^{-1} \int_{p_T}^{p_0} (\partial_p S)\, \Omega(p)\, dp$$

where $\Omega(p)$ describes the vertical structure of convection.

- Physically, M measures the net stability of moist atmosphere or, alternatively, the effectiveness of exporting moist static energy in presence of convection.
- Mq measures the effectiveness of precipitation or, alternatively, the available moisture for condensation in presence of convection.
- Ms measure the net stability of atmosphere without moisture effect.
- The relation between M, Mq , Ms, and NGMS:

$$\boxed{NGMS = \frac{\langle \omega \partial_p h \rangle}{-\langle \omega \partial_p q \rangle}}$$

$$M = M_s - M_q \; ; \quad NGMS = M / M_q$$ (Raymond et al. 2009)

- In tropics, warming $\Rightarrow M_s \uparrow$; moistening $\Rightarrow M_q \uparrow$

Definitions of Key Variables

(1) $\mathbf{V_T}$: baroclinic wind, defined as

$$\mathbf{V_T}(x,y,p,t)=\frac{\mathbf{V}(x,y,p,t)-\mathbf{V}_0(x,y,t)}{A^{+}(p)}$$

(2) $\Omega(p)$: vertical structure of convection, defined as

$$\Omega(p)=\int_{p}^{p_0}(A^{+}-\overline{A^{+}})\,dp'$$

where

$$A^{+}(p)=\kappa\int_{p}^{p_0}A(p)\,d\ln p$$

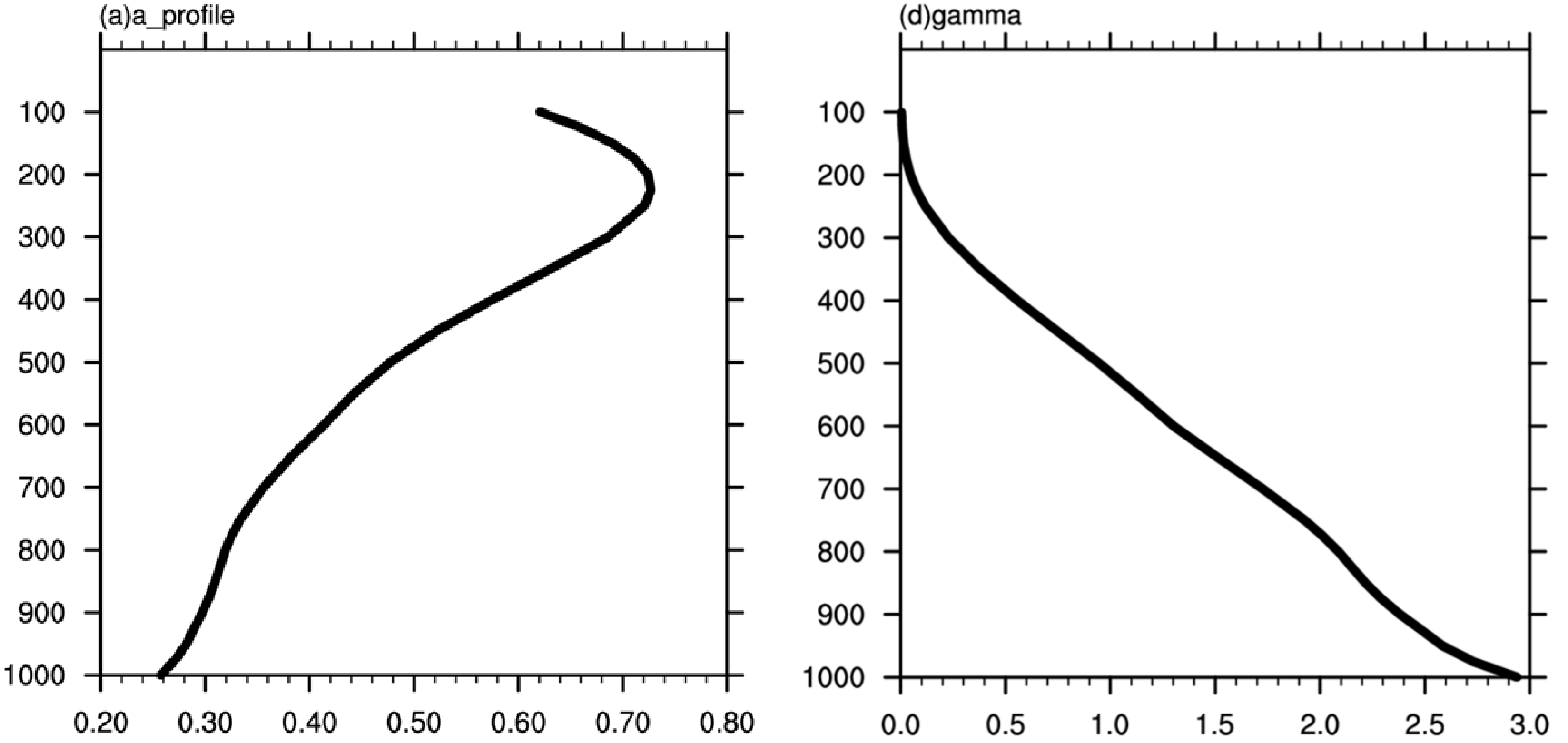
Temp. perturbation subject to moist adiabatic process

$$A(p)=\frac{1}{(1+\gamma)}\exp\left[-\kappa\int_{p}^{p_b}\frac{1}{(1+\gamma)}\,d\ln p\right]$$

and $\gamma = dq_{sat}/dT$ is the fractional change of water vapor per unit *T* change.

A(p) and γ (p) Profiles

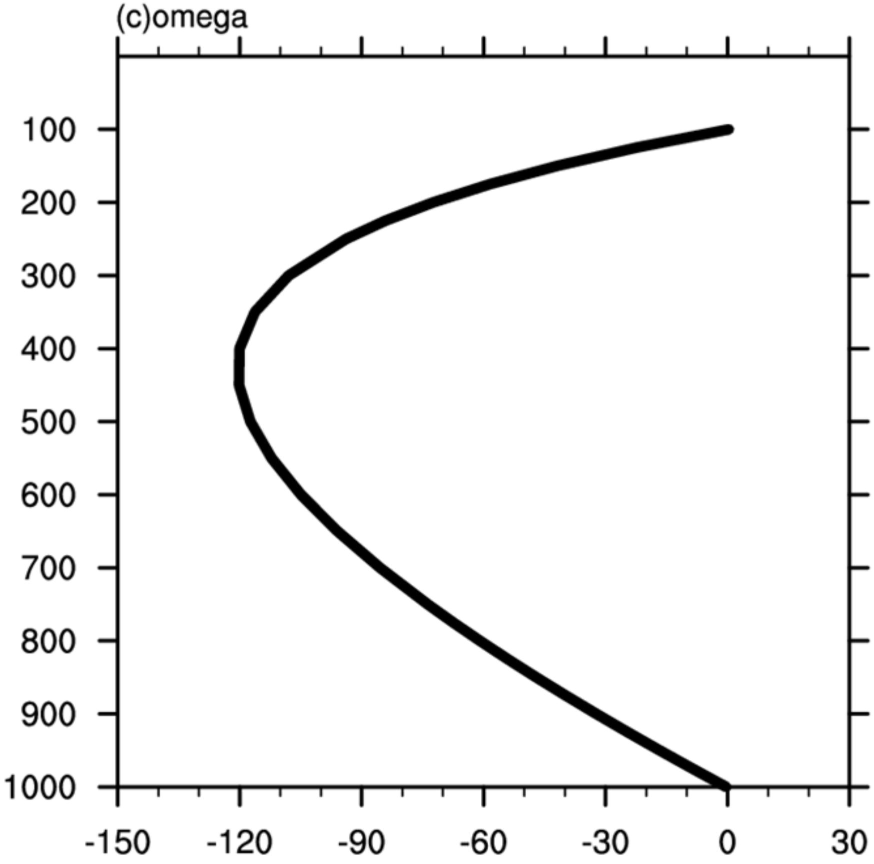
(ERA-interim climatology)



Vertical Structure of Convection

(P_T =100hPa)

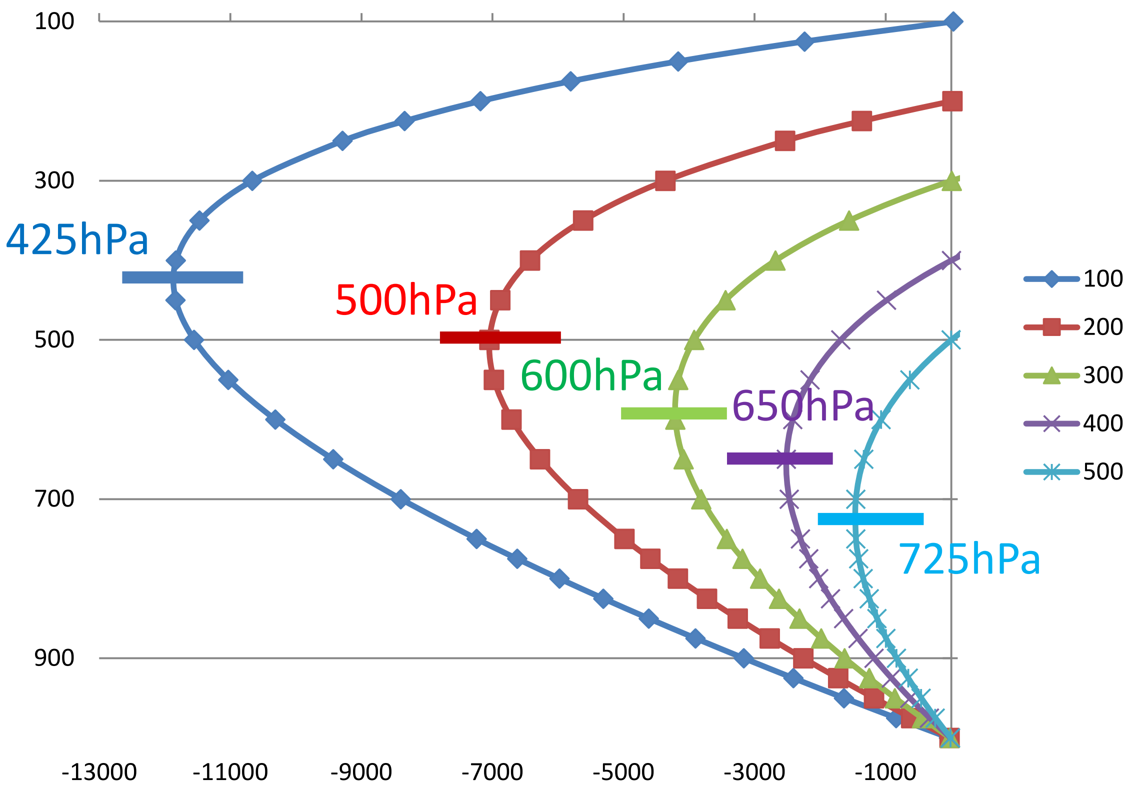
$\Omega(p)$



Vertical Structure of Convection

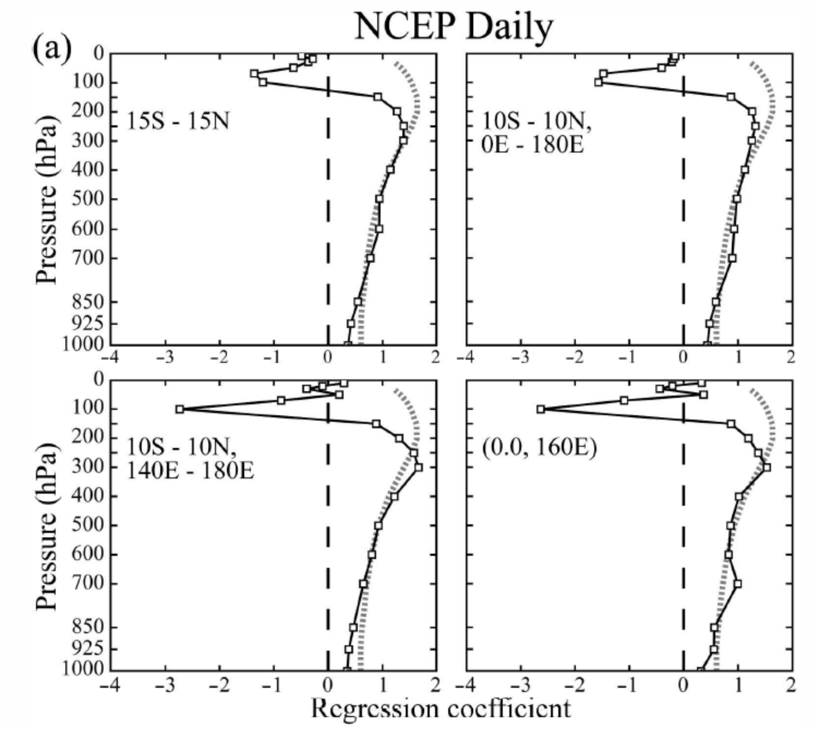
(Varying P_T)

$\Omega(p)$

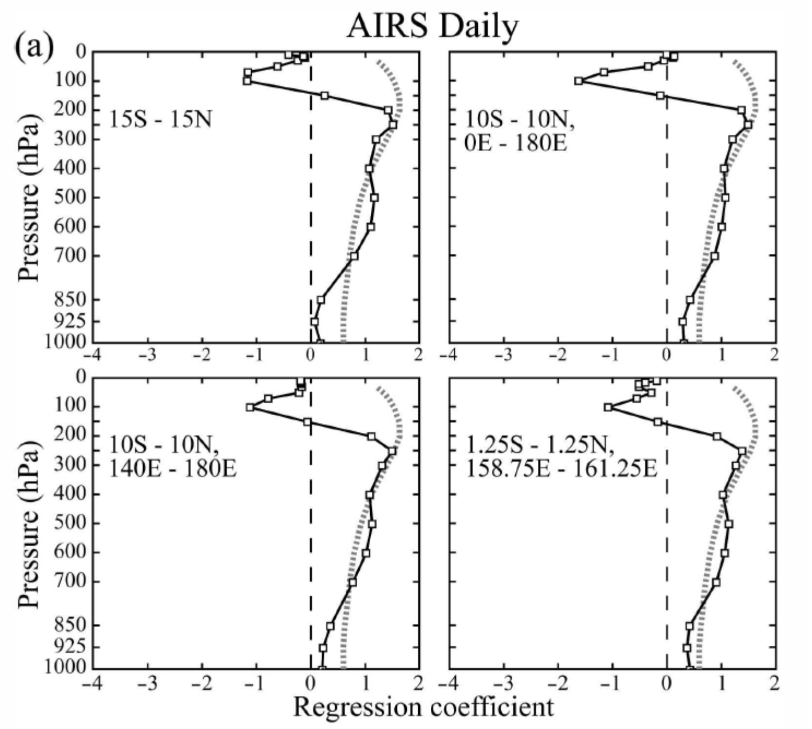


How Closeness to Moist Adiabat?

(sounding observations)



Reanalysis soundings

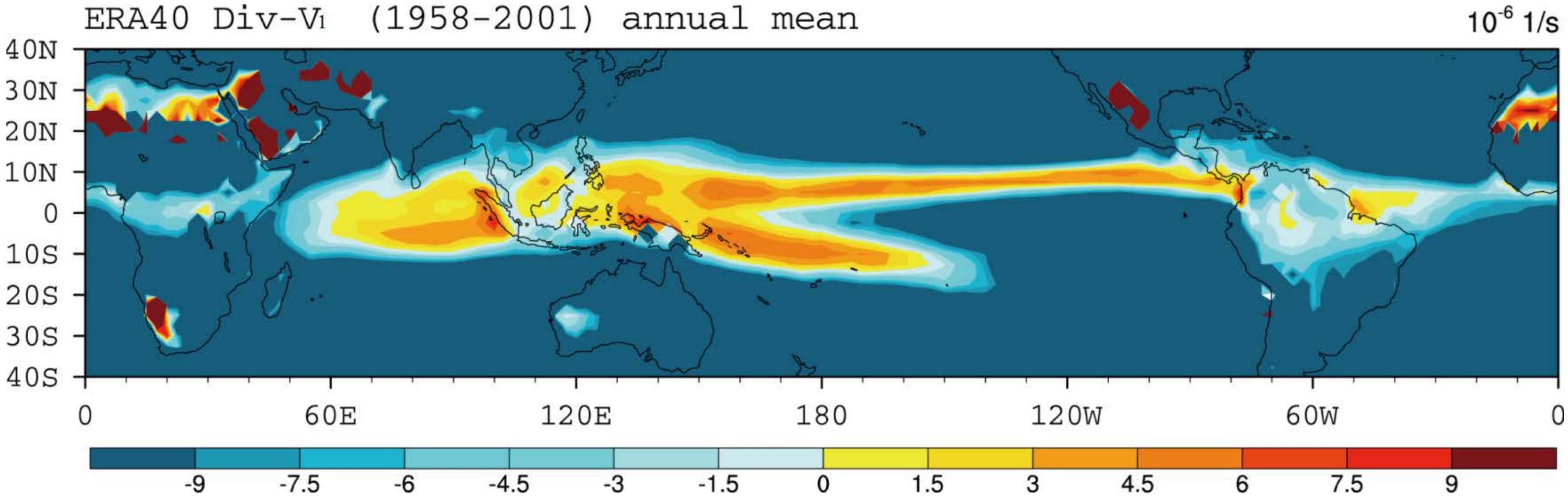


Satellite soundings

(Adopted from Holloway and Neelin 2007)

Spatial pattern of $\nabla \bullet \mathbf{V}_T$

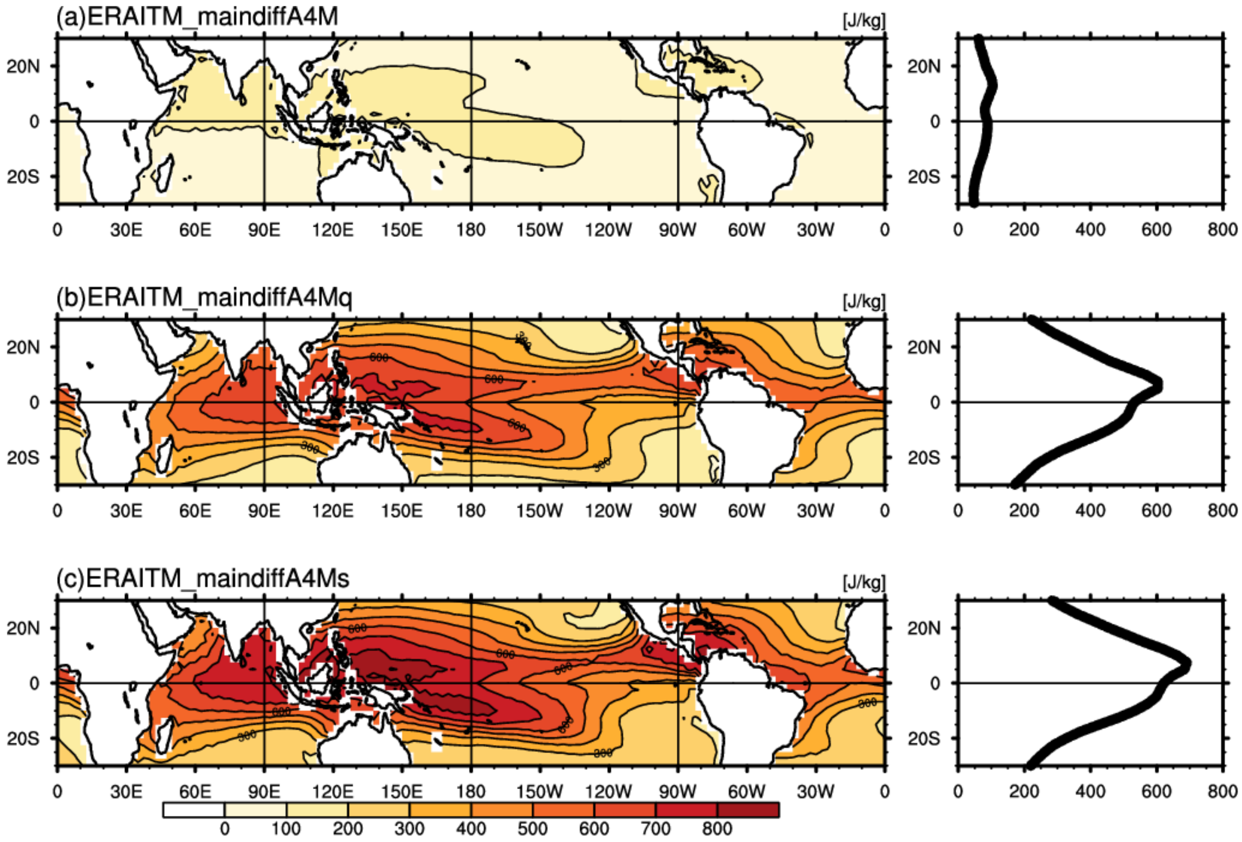
$$\mathbf{V}_T = (\mathbf{V}_{150} - \mathbf{V}_{1000}) \ /A^{+}$$



$\nabla \bullet \mathbf{V}_T$ > 0 implies upper (lower) level divergence (convergence) and upward transport of MSE.

Climatology of M, Mq and Ms

(ERA-interim)



(a)ERAITM_maindiffA4M

(b)ERAITM_maindiffA4Mq

(c)ERAITM_maindiffA4Ms

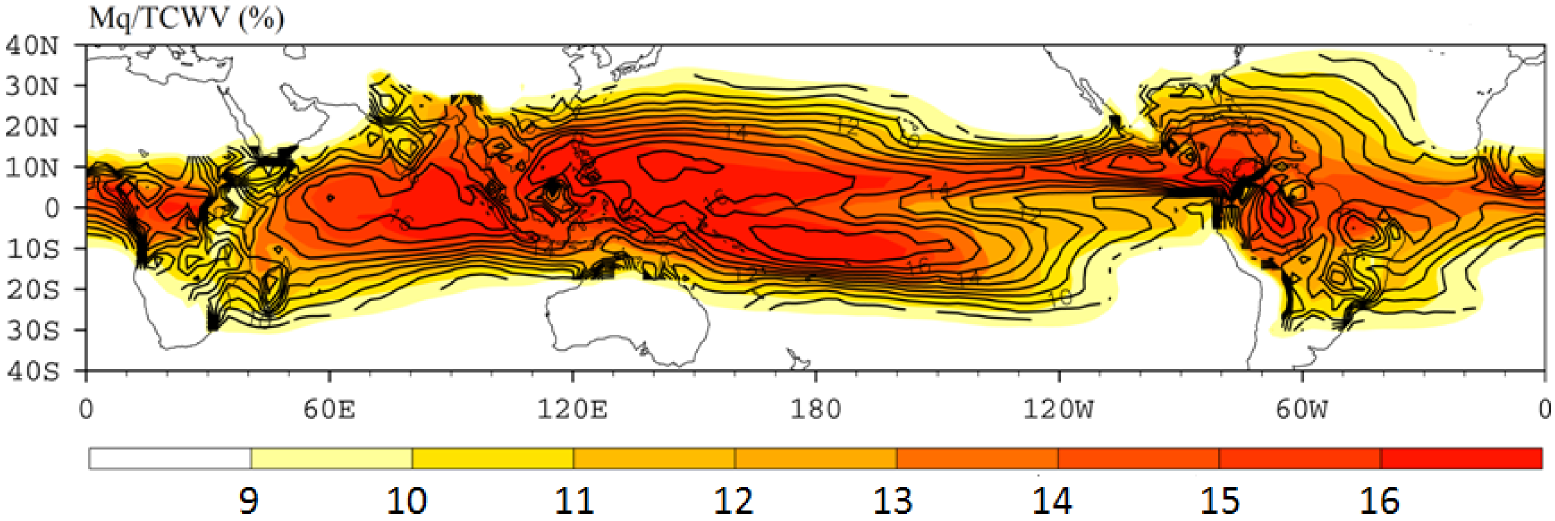
M (20N~20S)
: 50~100 J/kg

Mq (20N~20S)
: 300~600 J/kg

Ms (20N~20S)
: 350~700 J/kg

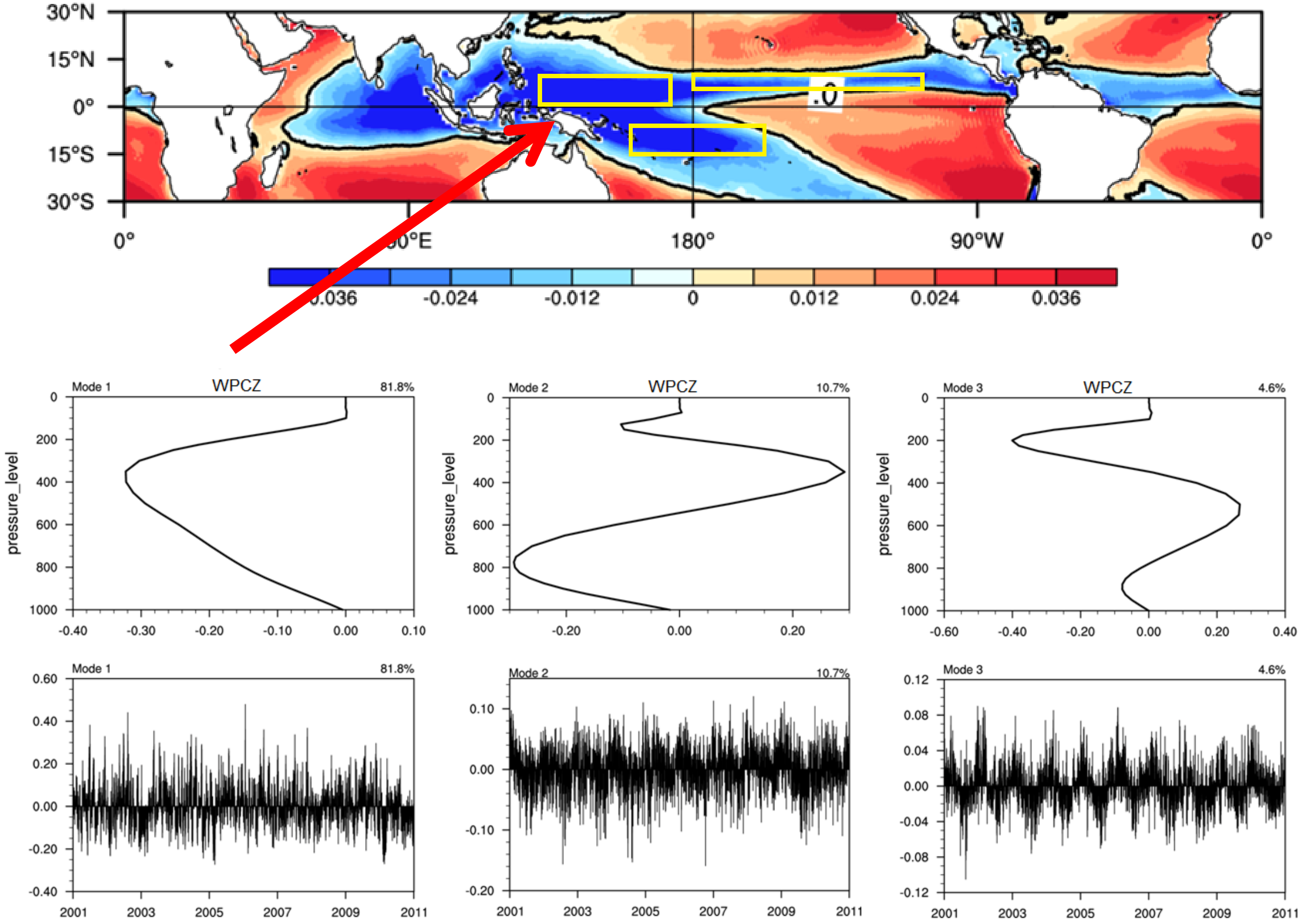
Precipitation Efficiency

(Mq/TCWV)



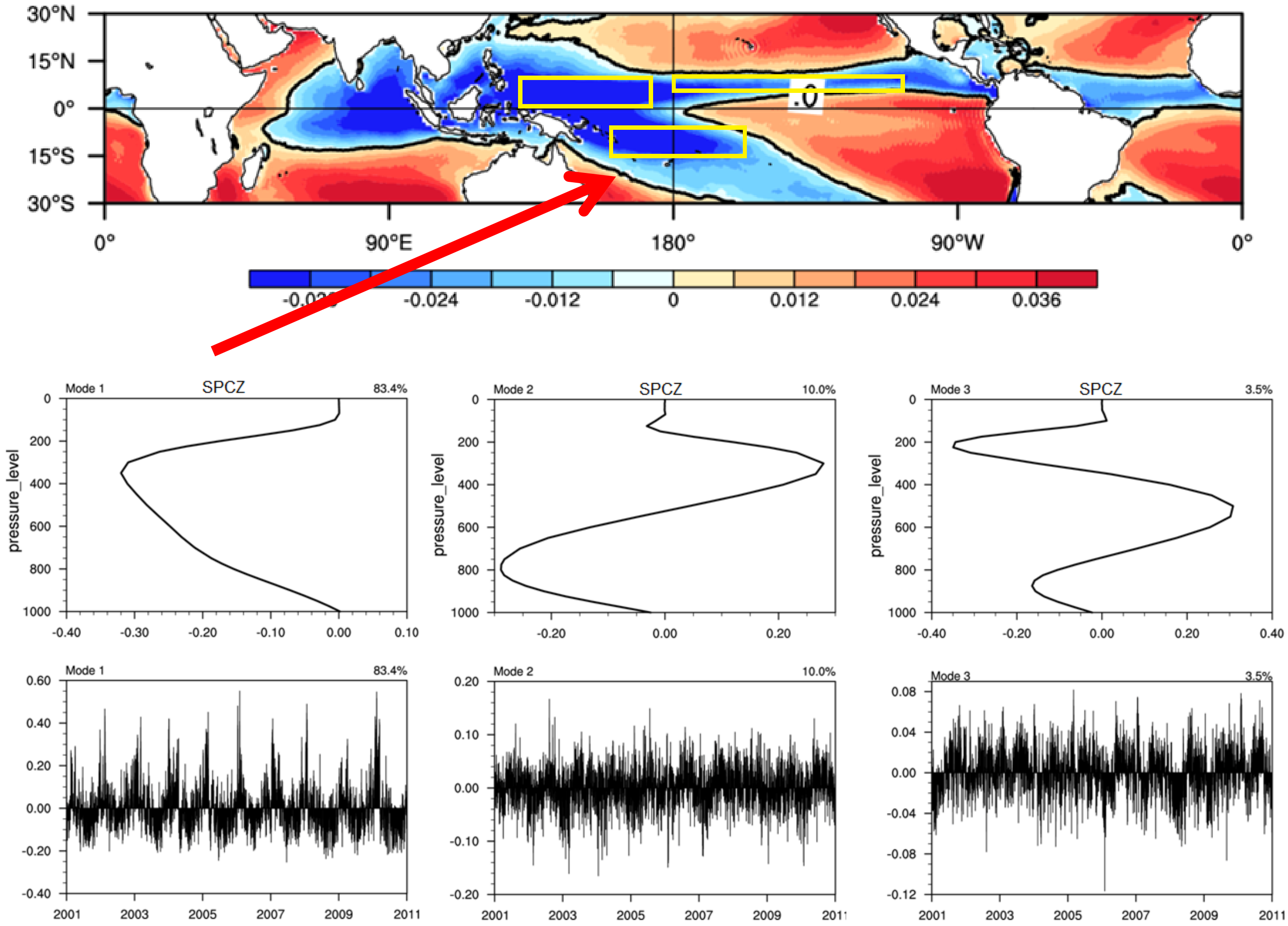
1. Physically, the ratio Mq/TCWV (in unit of %) measures the fraction of water vapor available for precipitation in the presence of convection.
2. Over the tropical convergence zone, only 12~18 % of the total column water vapor (TCWV) can be condensed to form precipitation.

Vertical Modes of Convection



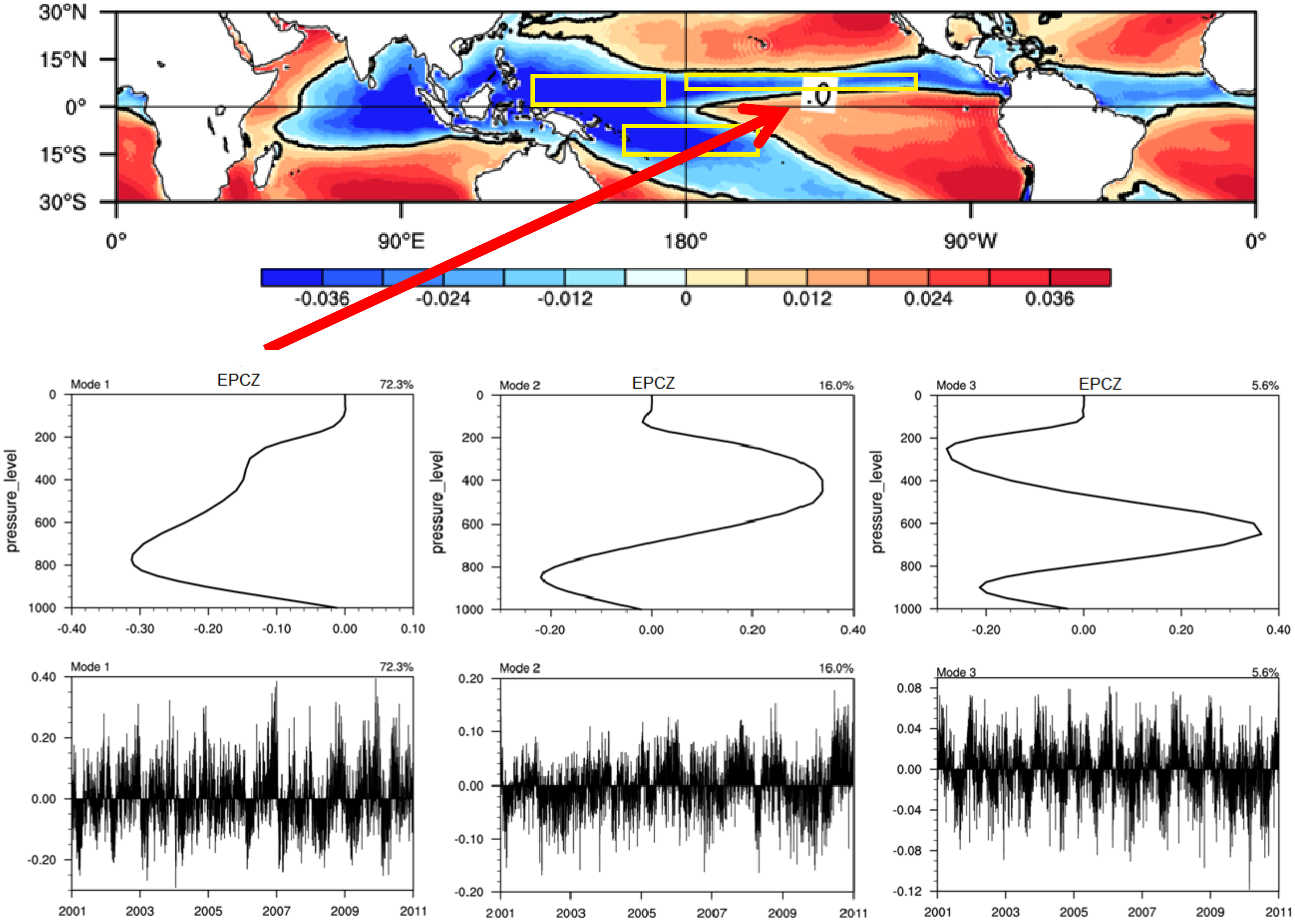
(from ERA-interim 10-year daily omega data with 0.5 x 0.5 resolution)

Vertical Modes of Convection



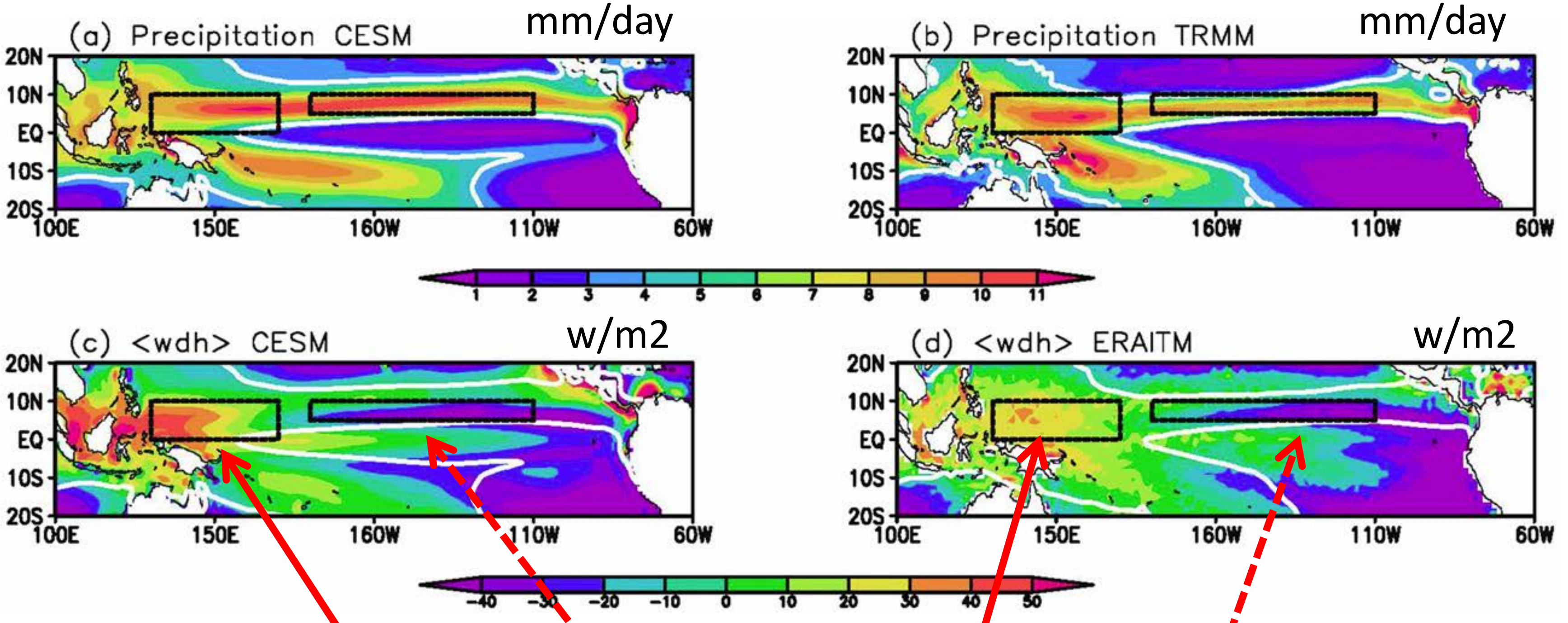
(from ERA-interim 10-year daily omega data with 0.5 x 0.5 resolution)

Vertical Modes of Convection



(from ERA-interim 10-year daily omega data with 0.5 x 0.5 resolution)

Vertical MSE transport in Tropics



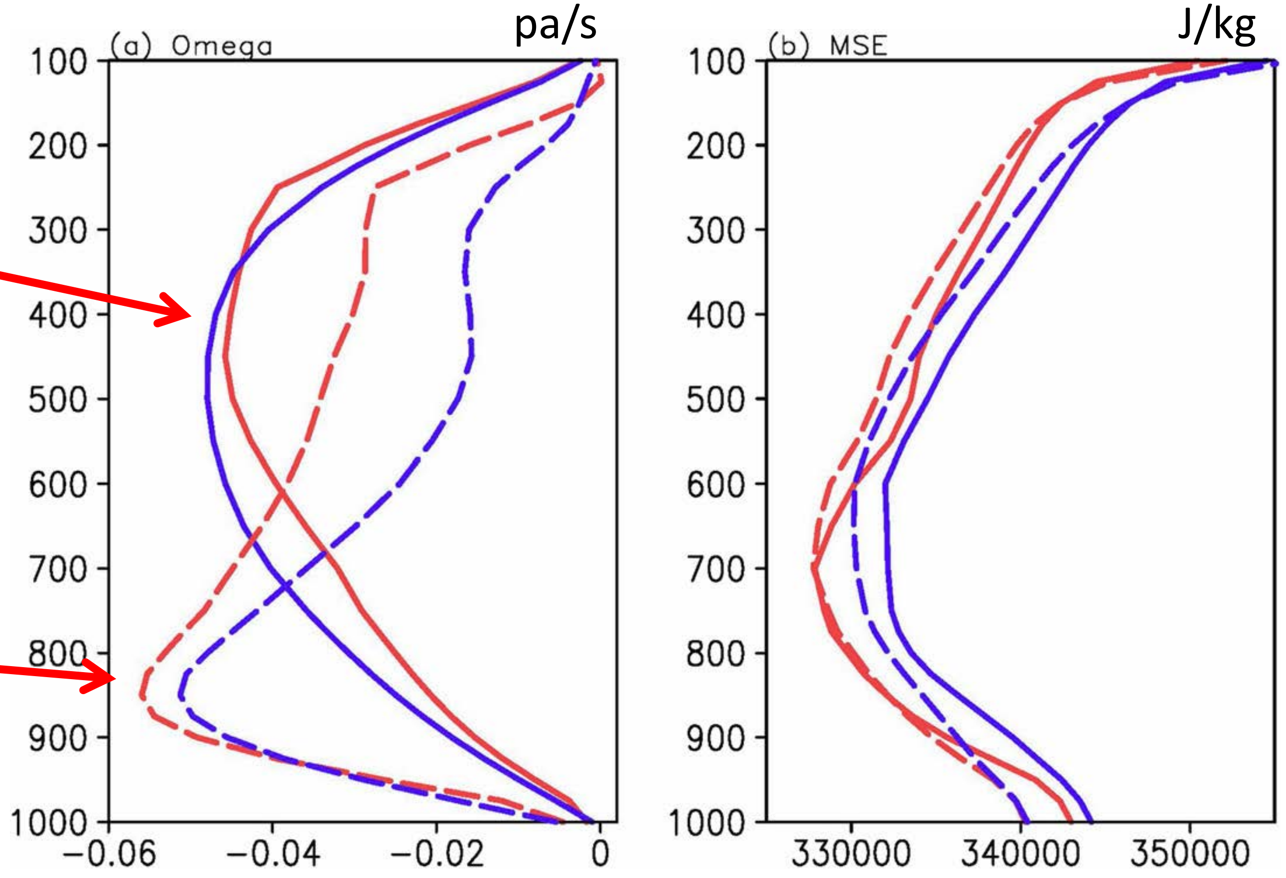
Negative vertical MSE transport over the eastern Pacific ITCZ

Positive vertical MSE transport over the western Pacific ITCZ

Profiles of MSE & Omega

WP ITCZ

EP ITCZ



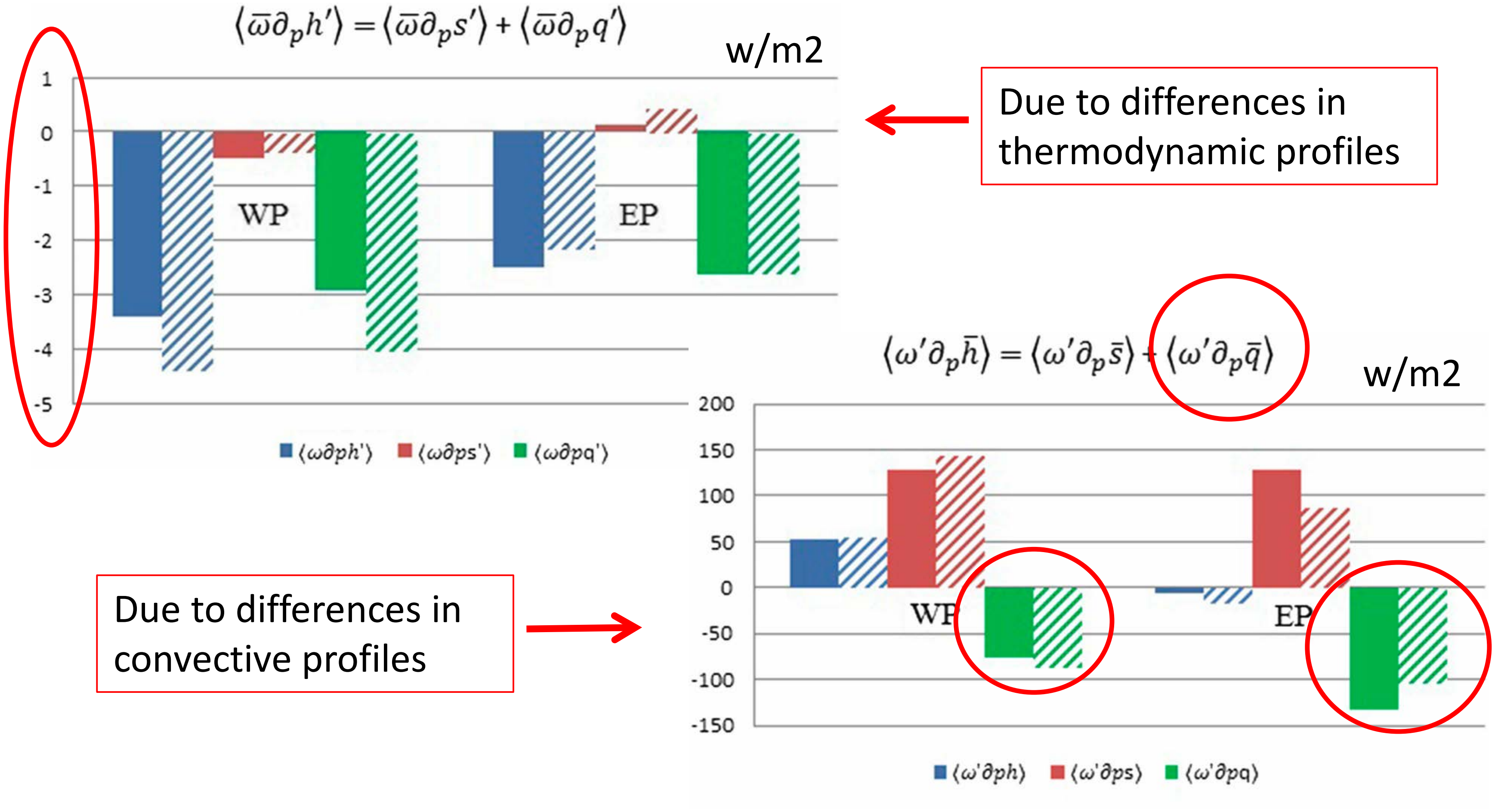
Solid curves: western Pacific ITCZ

Dash curves: eastern Pacific ITCZ

Blue for ERA-interim

Red for CESM

Decomposition of MSE transport



$$\langle\bar{\omega}\partial_p h'\rangle = \langle\bar{\omega}\partial_p s'\rangle + \langle\bar{\omega}\partial_p q'\rangle$$

w/m2

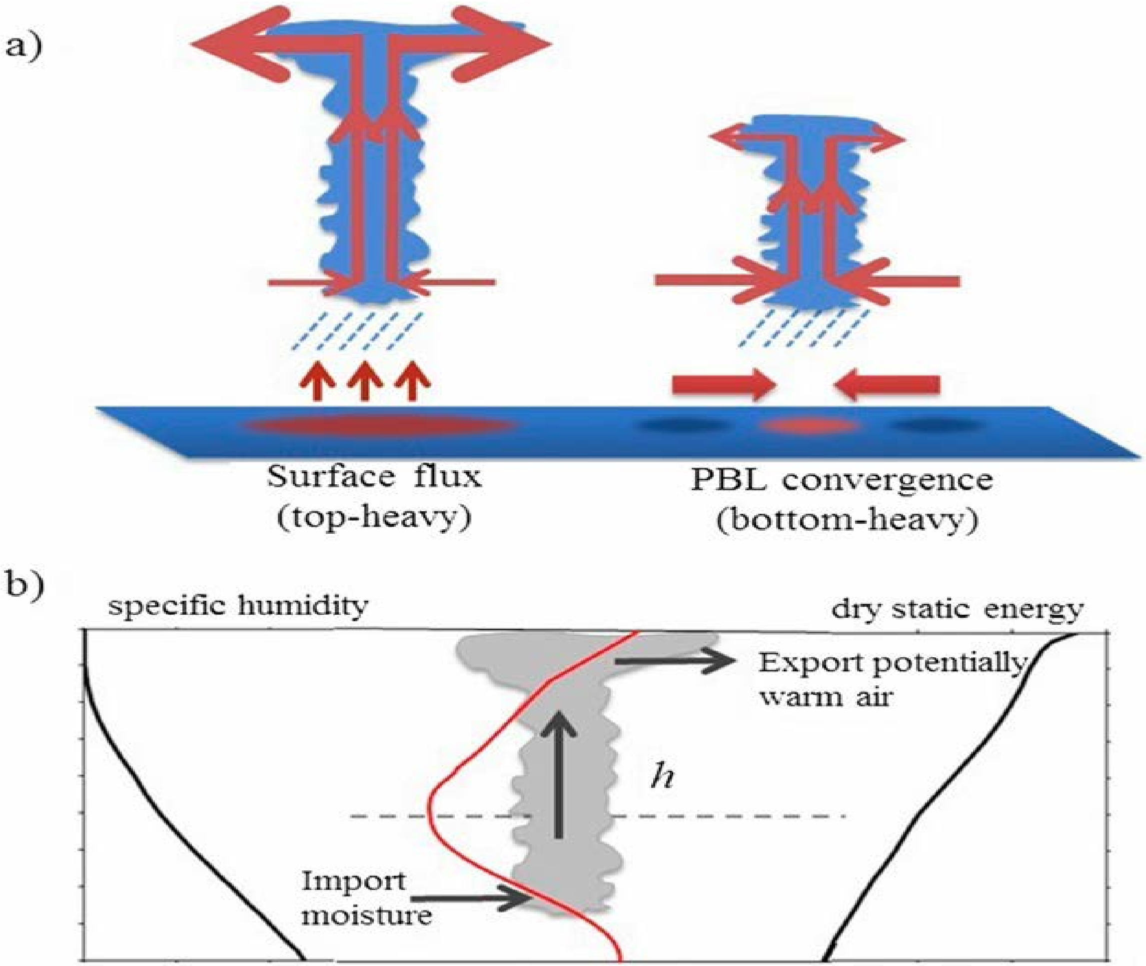
$$\langle\omega'\partial_p \bar{h}\rangle = \langle\omega'\partial_p \bar{s}\rangle + \langle\omega'\partial_p \bar{q}\rangle$$

w/m2

Due to differences in thermodynamic profiles

Due to differences in convective profiles

How vertical structure of convection impacts MSE transport ?



- A simple decomposition of the MSE budget equation shows that the sign of vertical MSE advection is determined mainly by the vertical moisture transport, which is very sensitive to the structure of convection.
- For a top-heavy (bottom-heavy) structure of convection, such as the case occurring in the western Pacific (eastern Pacific) ITCZ, the value of vertical MSE advection tends to be positive (negative), implying an export (import) of the column MSE and a stabilization (destabilization) of the atmosphere.

Why MJO has slow phase speed?

- Gross moist stability has the following equivalence for equivalent depth and moist Kelvin-like wave phase speed,

$$h_e = M / g \quad ; \quad C = (M)^{1/2}$$

- In the tropics, GMS: 50~100 J/kg within 20°N~20°S, corresponding to an equivalent depth of 5~10 m and a moist Kelvin wave phase speed of 7~10 m/s, consistent with the observed MJO phase speed over the tropical Indo-Pacific region.

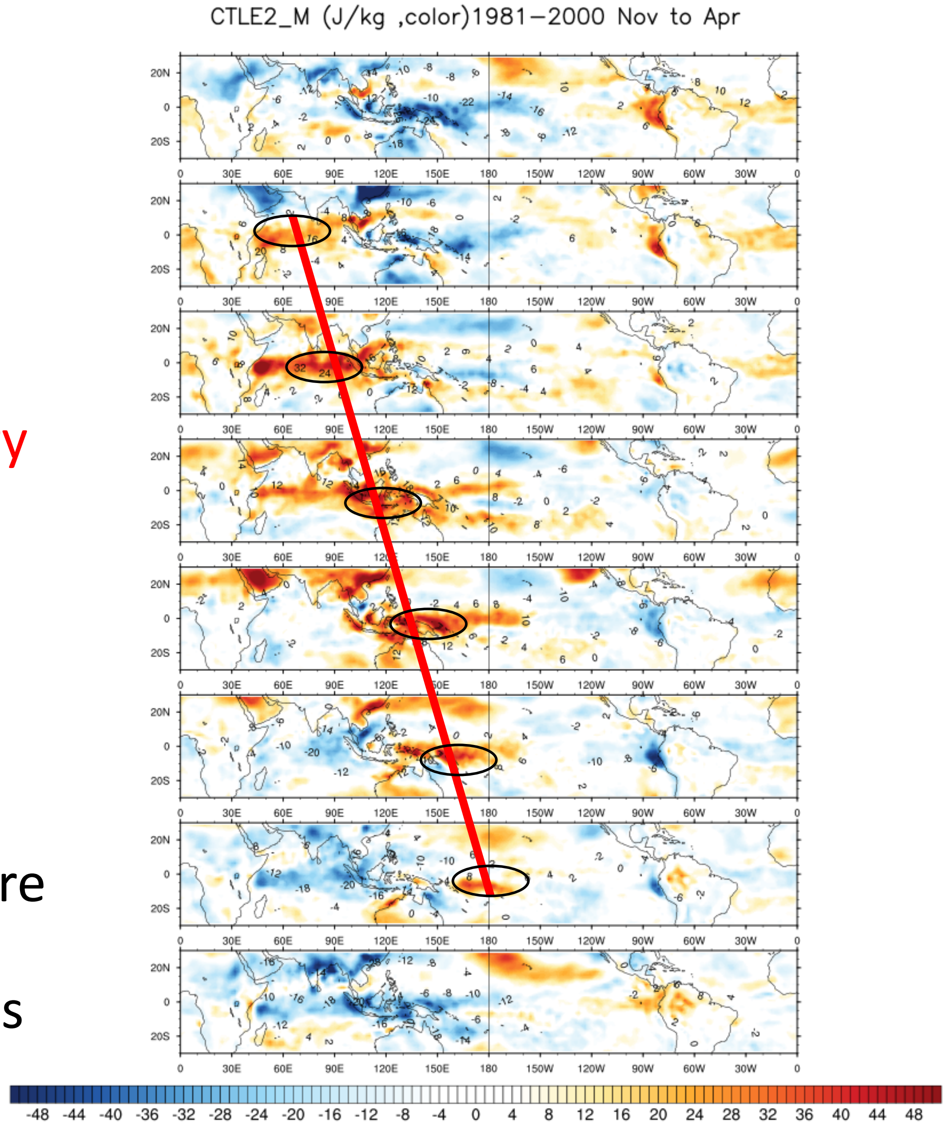
When MJO starts to propagate?

$$(\ )' = \tilde{(\ )}e^{ik(x-ct)}$$

$$C = (M)^{1/2}$$

(1) For negative M, C is pure imaginary, one gets a stationary unstable mode; during which, the atmospheric column undergoes an import of MSE.

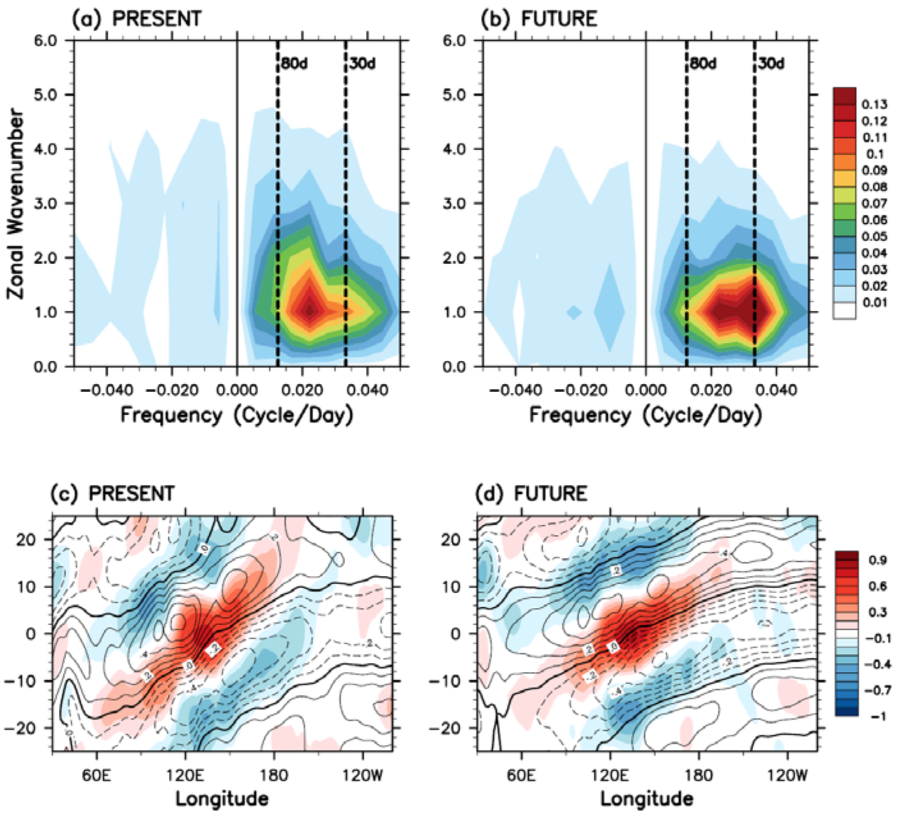
(2) For positive M, C is real, one gets a propagating deep convective mode with phase speed determined by the square root of M; during which, the atmospheric column undergoes an export of MSE.


CTLE2_M (J/kg ,color)1981−2000 Nov to Apr

(ECHAM5-SIT control run)

Why faster MJO in warming world?

(ECHAM5-SIT/RCP8.5)



shadings: u’
contours: p’

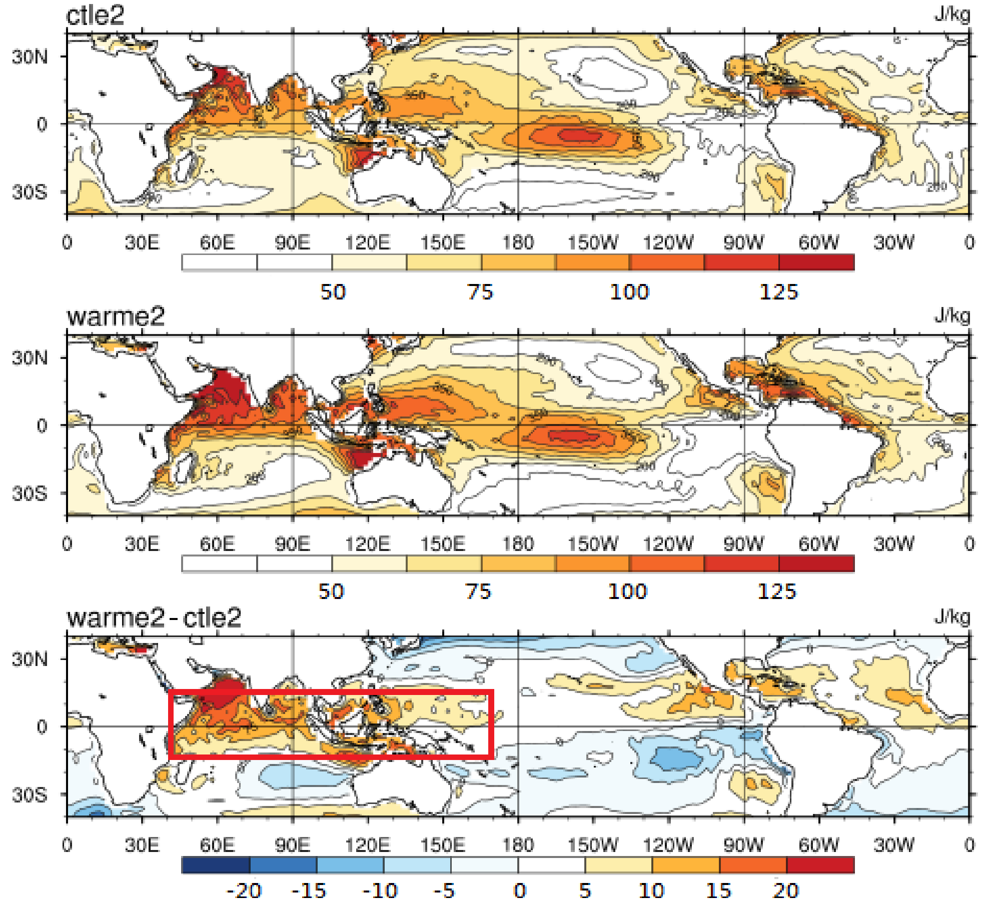
Phase speed:
9~10 m/s

Phase speed:
11~12 m/s

Under global warming, MJO remains wavenumber one structure, but with a larger amplitude and a greater eastward-propagating phase speed.

(courtesy of June Chang)

GMS in present and warming climate



Present climate

Warming climate

Warming - Present

GMS ↑ 15~20 J/kg

- GMS (M) increases from 50~100J/kg to 65~120J/kg, corresponding to about 20% increase in moist Kelvin wave phase speed, consistent with model results.
- Under global warming , enhanced deep convection associated with MJO occurs, resulting in an increased upward transport of moist static energy and a more stable troposphere, in turn, producing a faster MJO cycle (because of larger GMS) as shown in the ECHAM5-SIT/RCP8.5 simulation.

References

Yu, J.-Y., and J. D. Neelin, 1997: Analytic approximations for moist convectively adjusted regions. *J. Atmos. Sci.*, **54**, 1054-1063.

Yu, J.-Y., C. Chou, and J. D. Neelin, 1998: Estimating the gross most stability of the tropical atmosphere. *J. Atmos. Sci.*, **55**, 1354-1372.

Holloway, C. E. and J. D. Neelin, 2007: The convective cold top and quasi equilibrium. *J. Atmos. Sci.*, **64**, 1467-1487. doi:10.1175/JAS3907.1.

Raymond, D. J., S. L. Sessions, A. H. Sobel, and Ž. Fuchs, 2009: The mechanics of gross moist stability. *Journal of Advances In Modeling Earth System*, **1***,* doi:10.3894/JAMES.2009.1.9.

Chen, C.-A., J.-Y. Yu, and C. Chou, 2016: Impacts of vertical structure of convection in global warming: The role of shallow convection. *J. Climate,* **29**, 4665-4684, doi:http://dx.doi.org/10.1175/JCLI-D-15-0563.1

Bui, H. X., J.-Y. Yu, and C. Chou, 2016: Impacts of vertical structure of large-scale vertical motion in tropical climate: moist static energy framework. *J. Atmos. Sci.,* **73**, doi:10.1175/JAS-D-16-0031.1.

Thanks for Listening！