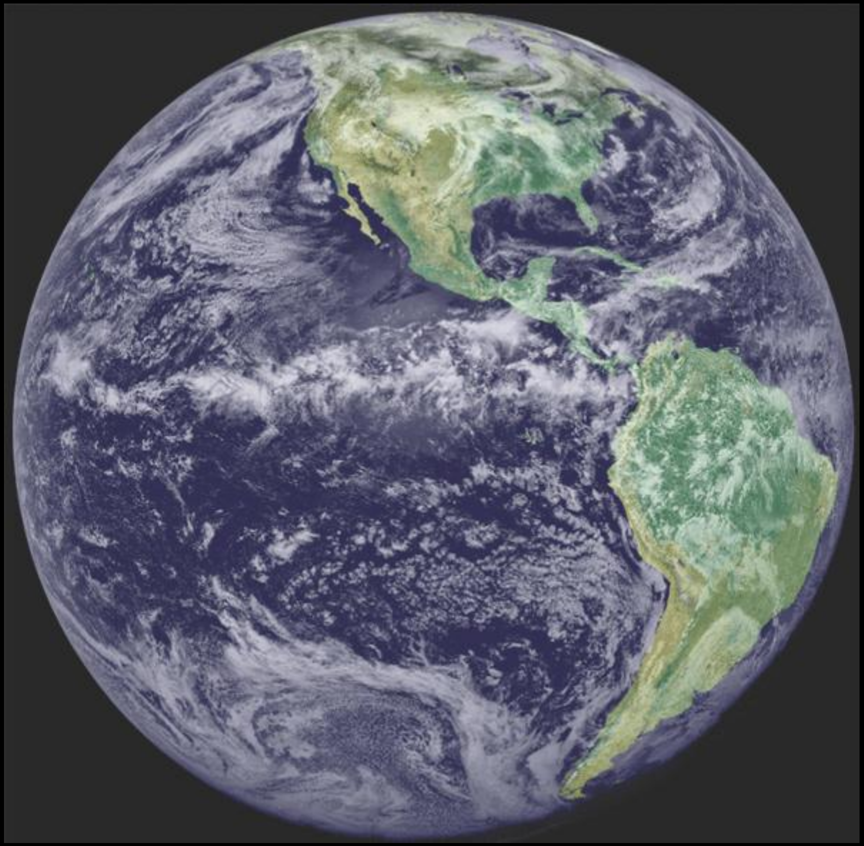
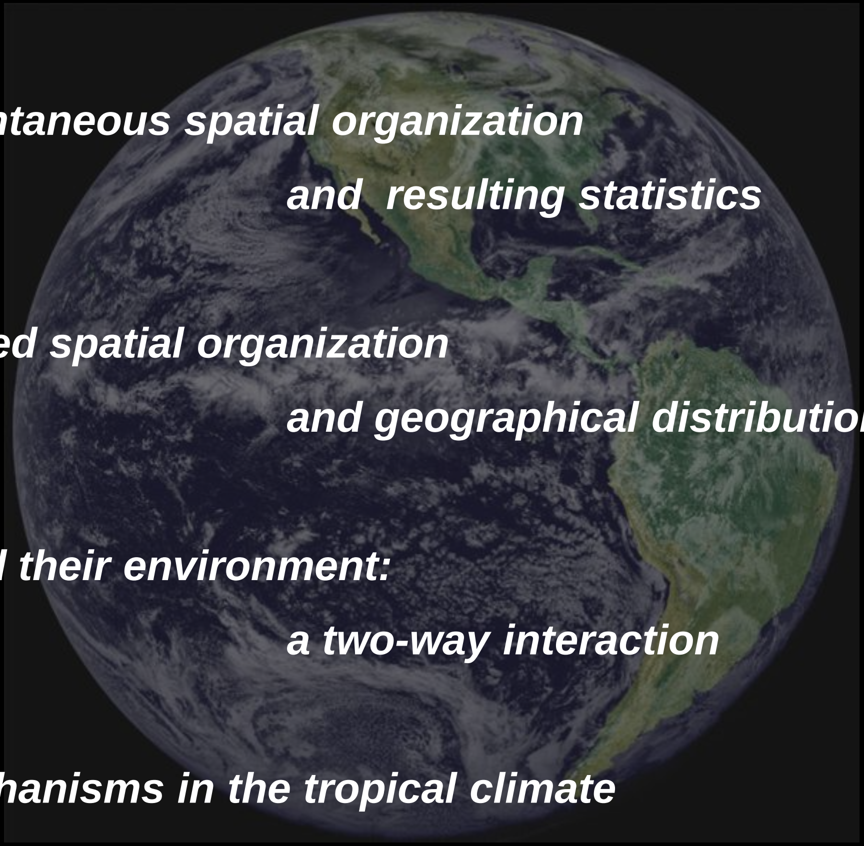


Tropical and subtropical cloud systems

Outline



A. *Cloud spontaneous spatial organization*
and resulting statistics

B. *Cloud forced spatial organization*
and geographical distribution

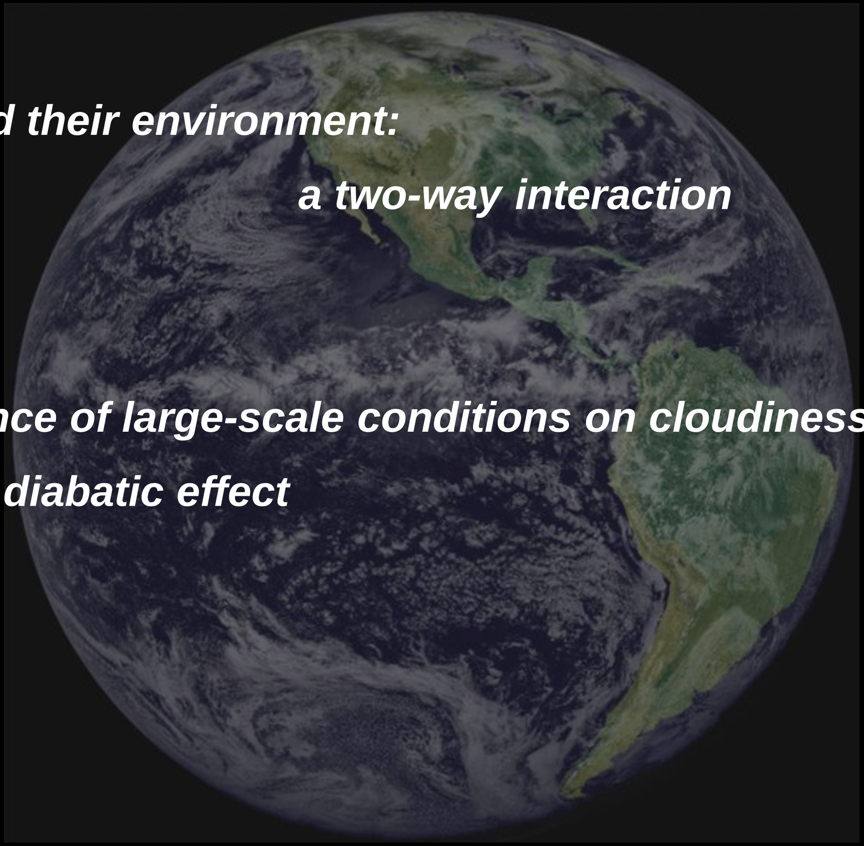
Lecture 1

C. *Clouds and their environment:*
a two-way interaction

D. *Cloud mechanisms in the tropical climate*

Lecture 2

Outline



C. *Clouds and their environment:*

a two-way interaction

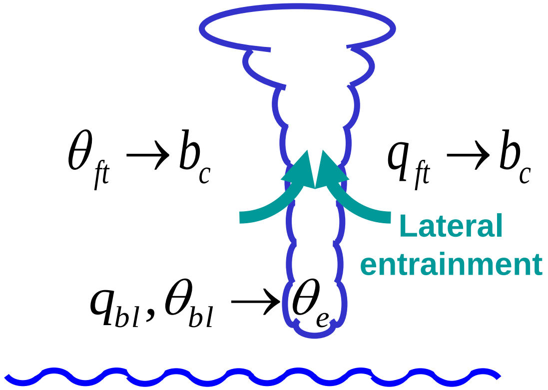
1. *Influence of large-scale conditions on cloudiness*
2. *Cloud diabatic effect*

Main sensitivities

Convective clouds are sensitive to the moisture and temperature stratification …

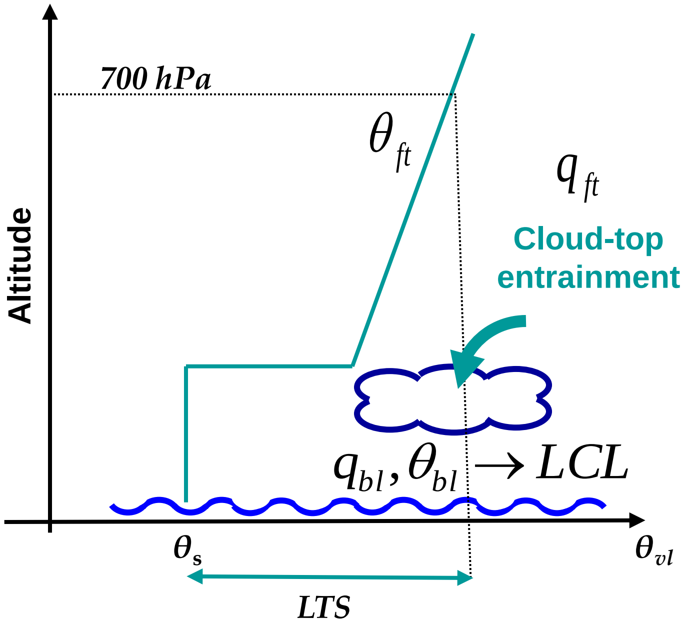
- **the moister, the more clouds;**
- **the smaller potential temperature lapse rate, the more convective clouds.**

… through cloud buoyancy.

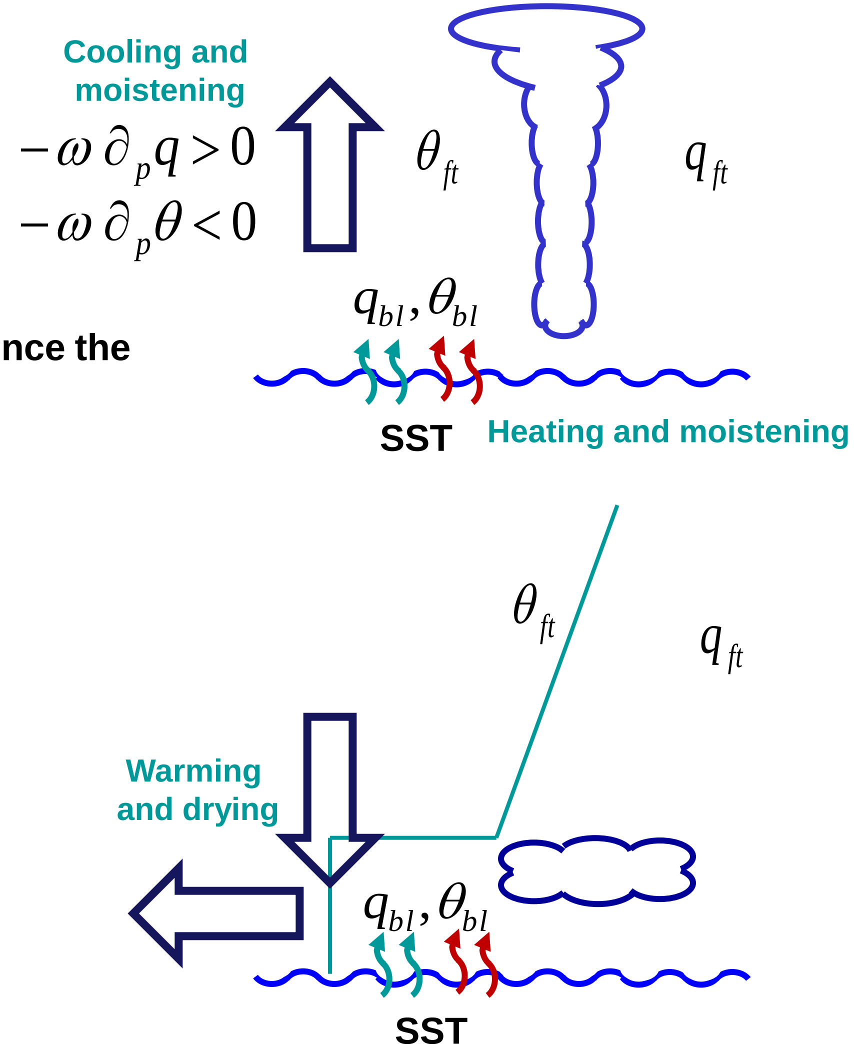


Stratus are sensitive to the moisture and temperature stratification at boundary layer top …

- **the moister, the more clouds;**
- **the larger the temperature jump, the more clouds.**



Main sensitivities

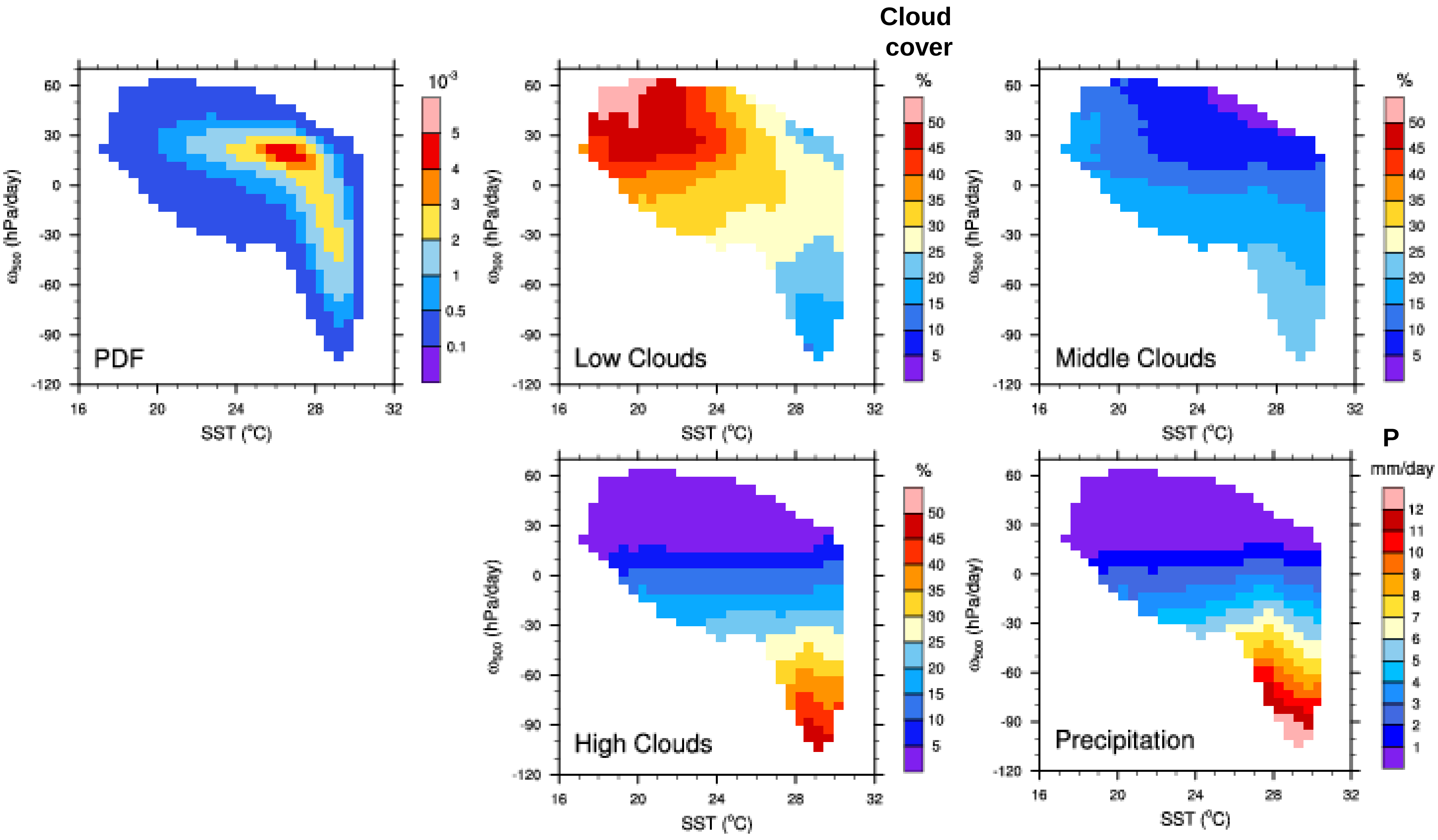


Clouds are sensitive to all the processes that influence the moisture and temperature stratification …

- **surface fluxes (SST);**
- **vertical advection (ω)**
- **horizontal advection.**

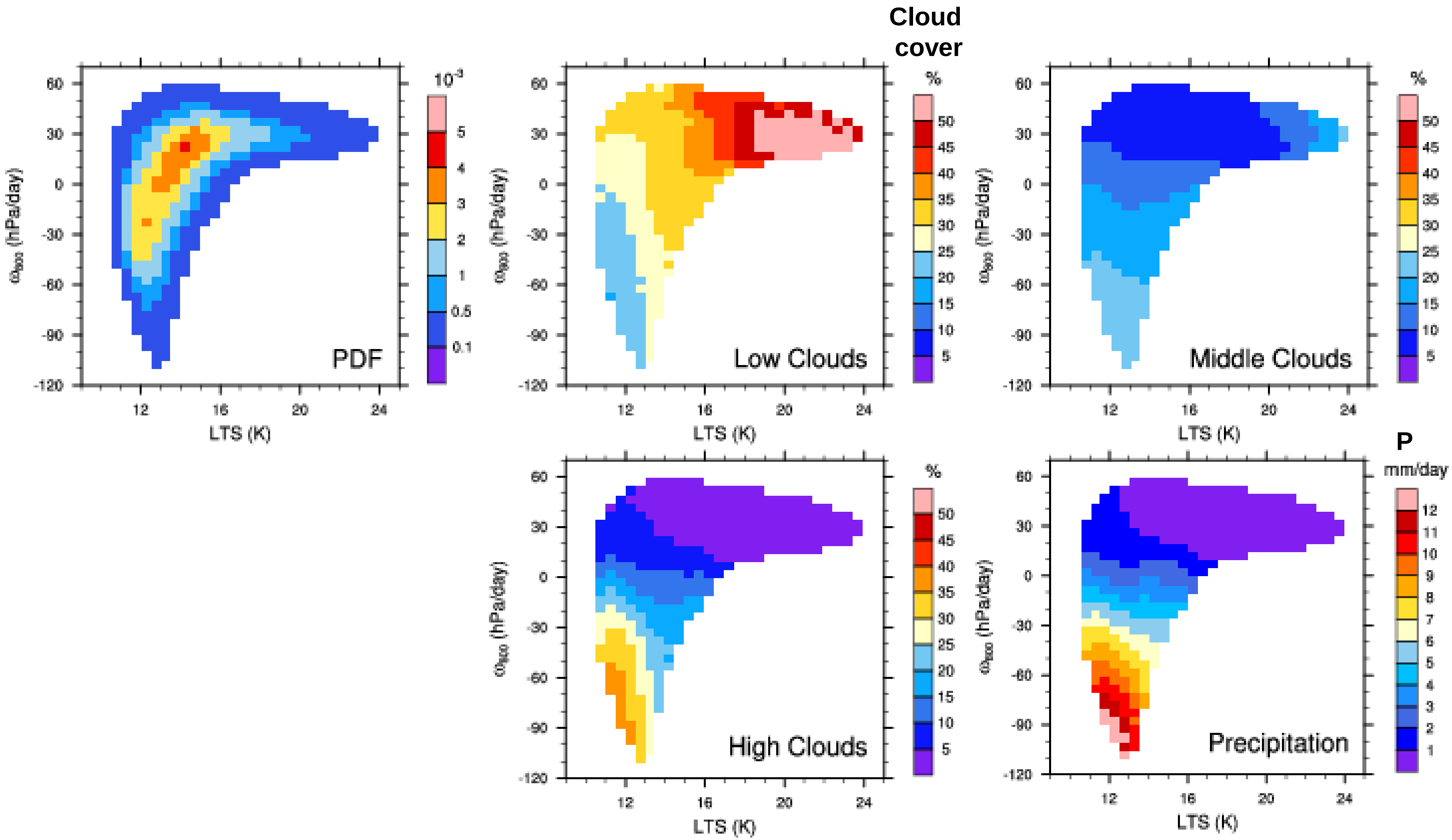
SST and vertical motion

Sensitivity to Sea Surface Temperature and vertical circulation ($\omega(500\ \mathrm{hPa})$)

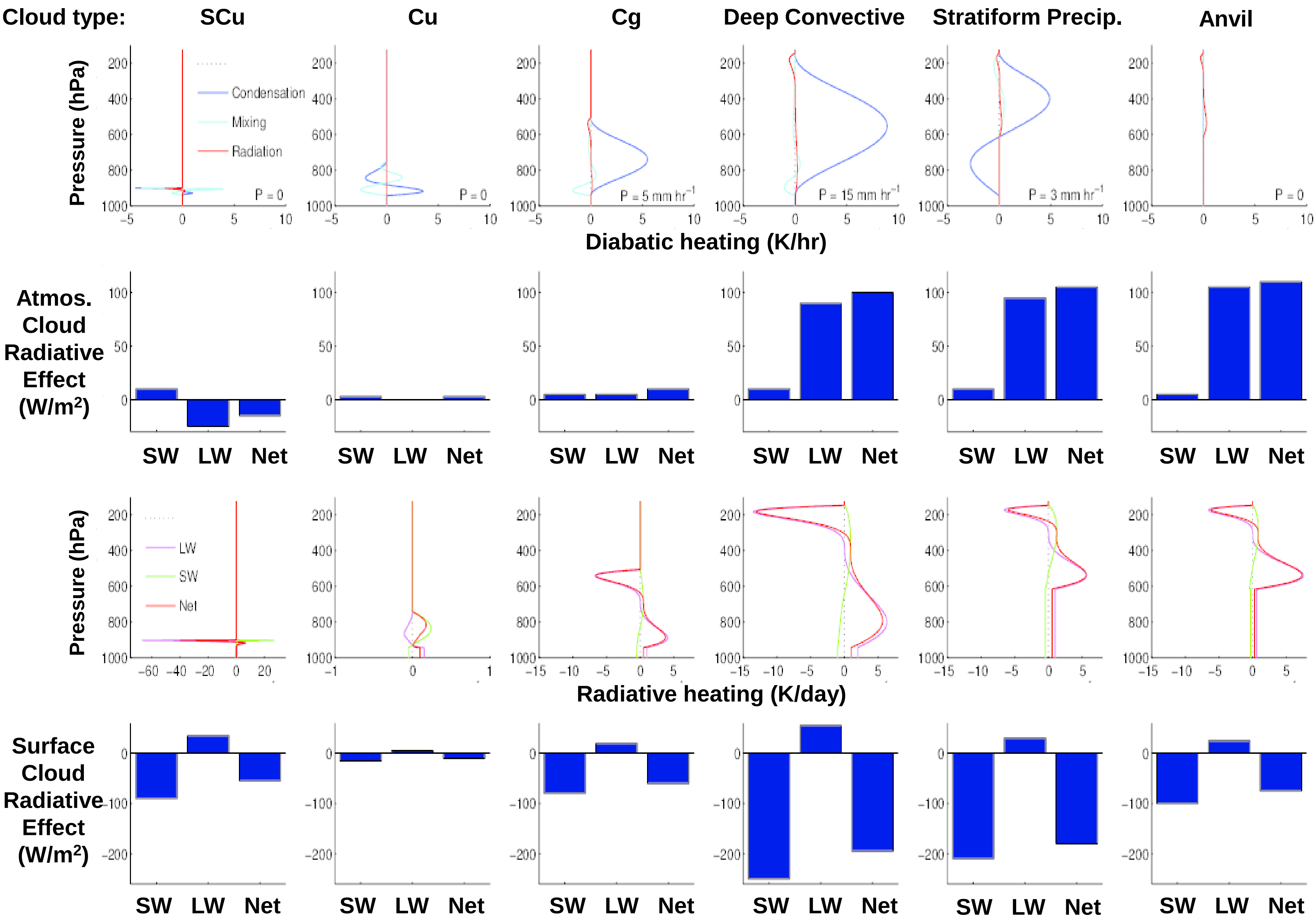


LTS and vertical motion

Sensitivity to Lower Tropospheric Stability and vertical circulation ($\omega(800\ \mathrm{hPa})$)

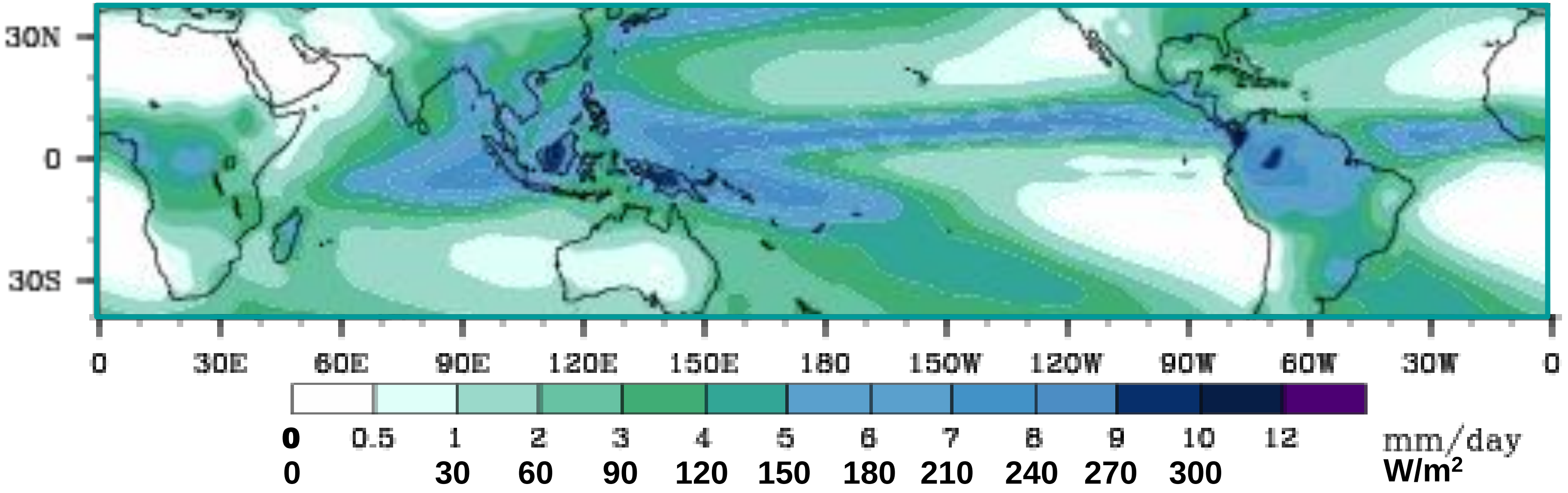


Diabatic heating by type of clouds



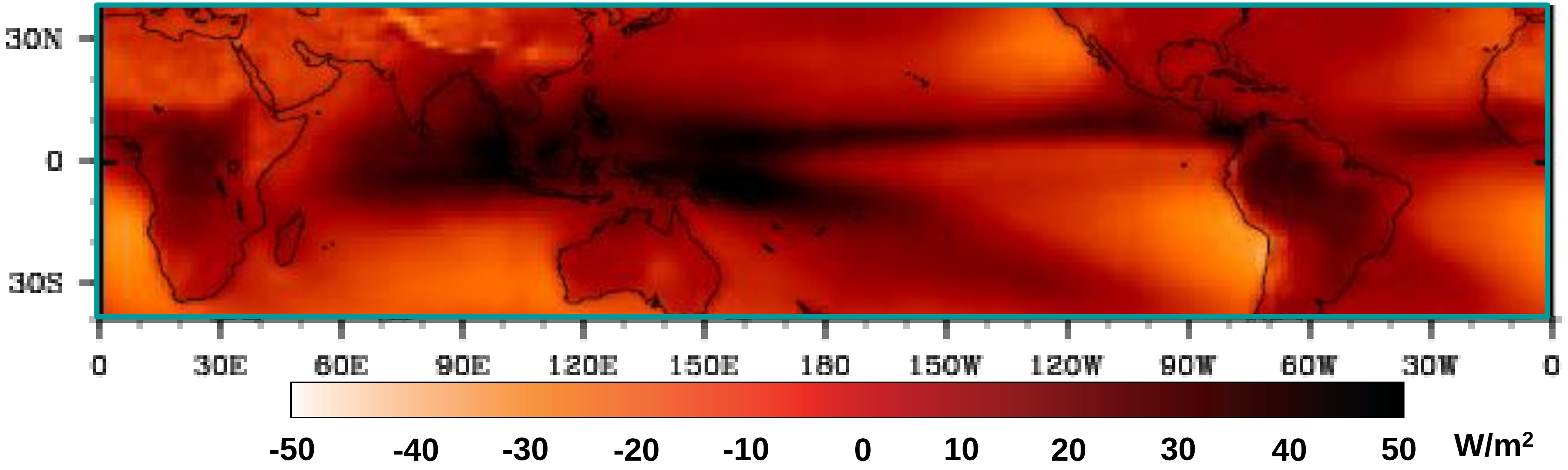
Geographical distribution of diabatic heating

Atmospheric latent heating

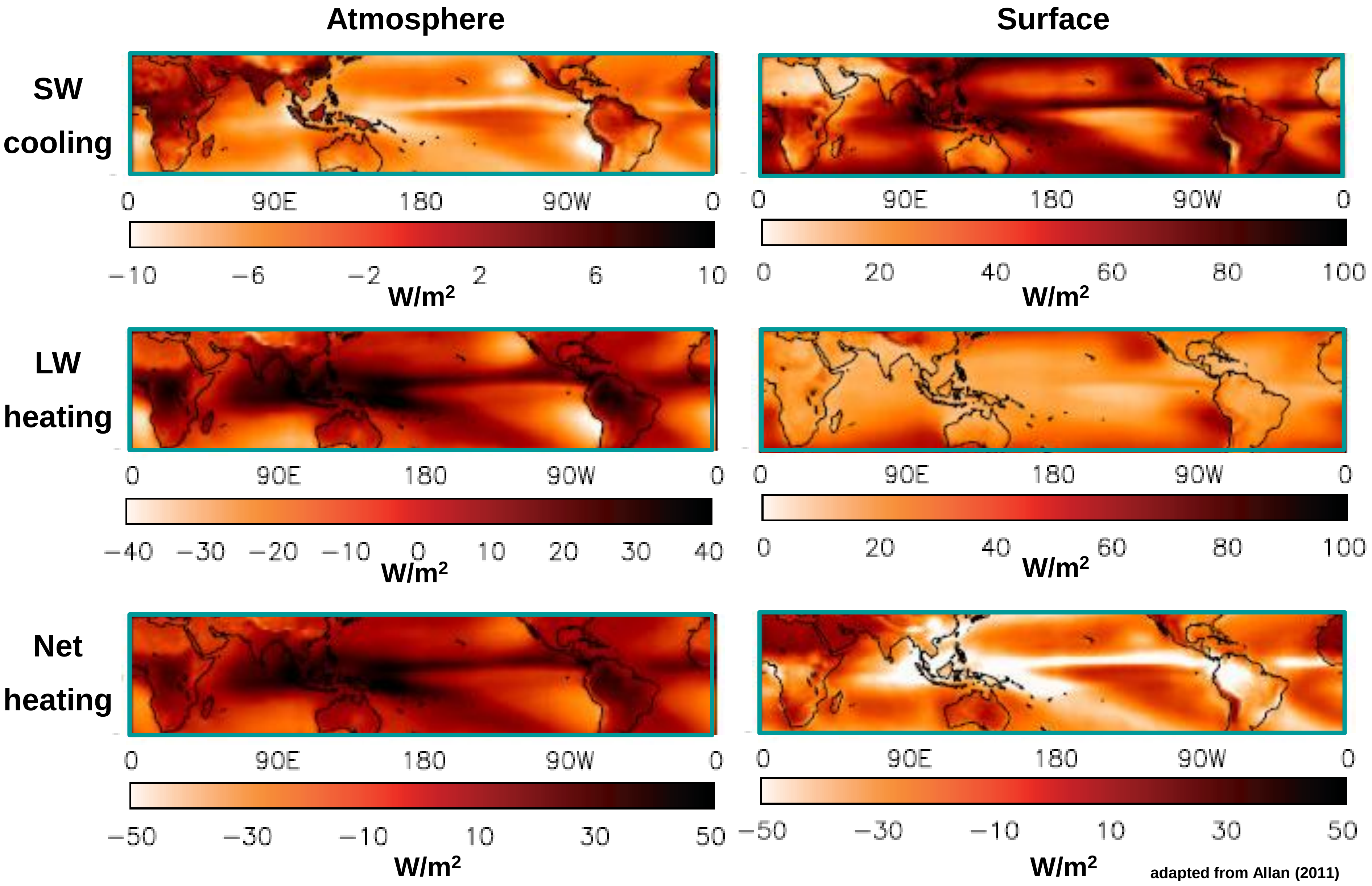


Atmospheric net radiative heating

from Allan (2011)

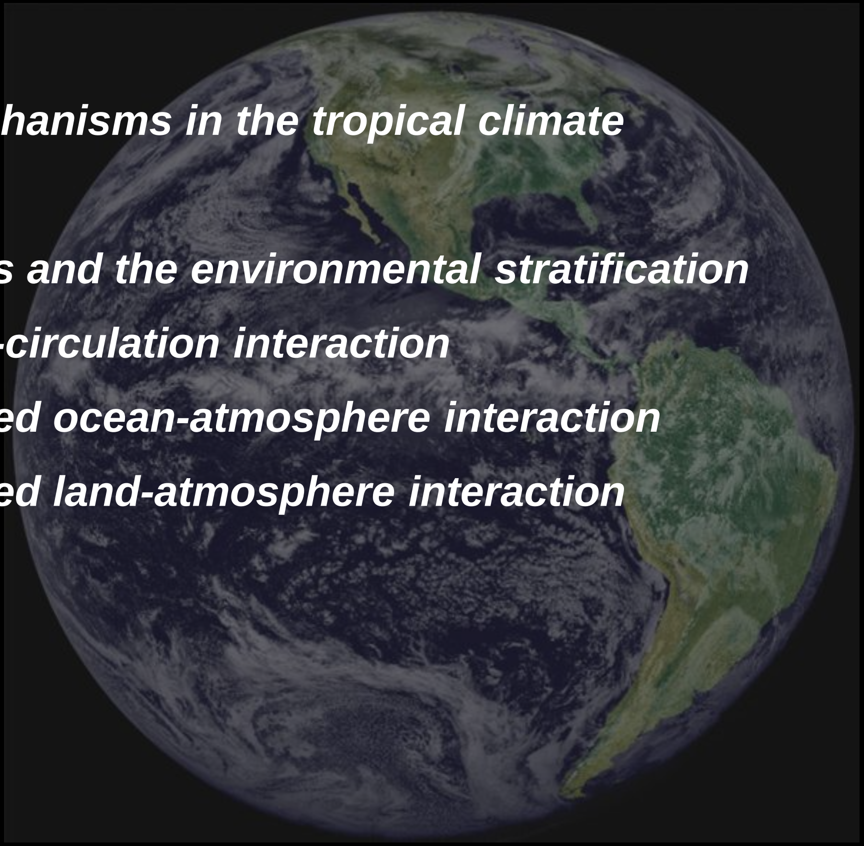


Radiative heating



adapted from Allan (2011)

Outline



D. Cloud mechanisms in the tropical climate

1. *Clouds and the environmental stratification*
2. *Cloud-circulation interaction*
3. *Coupled ocean-atmosphere interaction*
4. *Coupled land-atmosphere interaction*

Cloud moistening, drying, and restratification

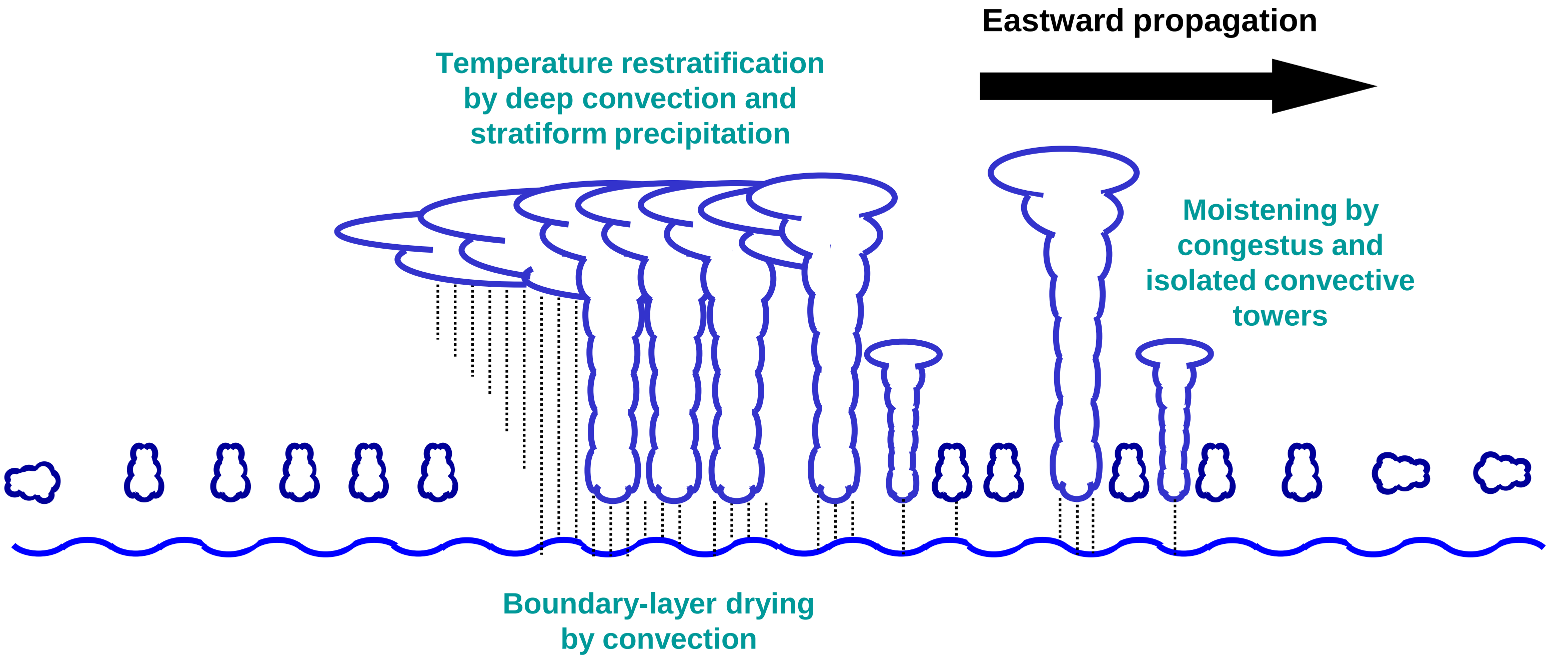
- **Clouds are sensitive to the moisture and temperature stratification …**

- **Clouds modity the stratification:**

  convective clouds moisten in their upper part, and deep one dry the subcloud layer

  clouds change the temperature stratification through diabatic heating:

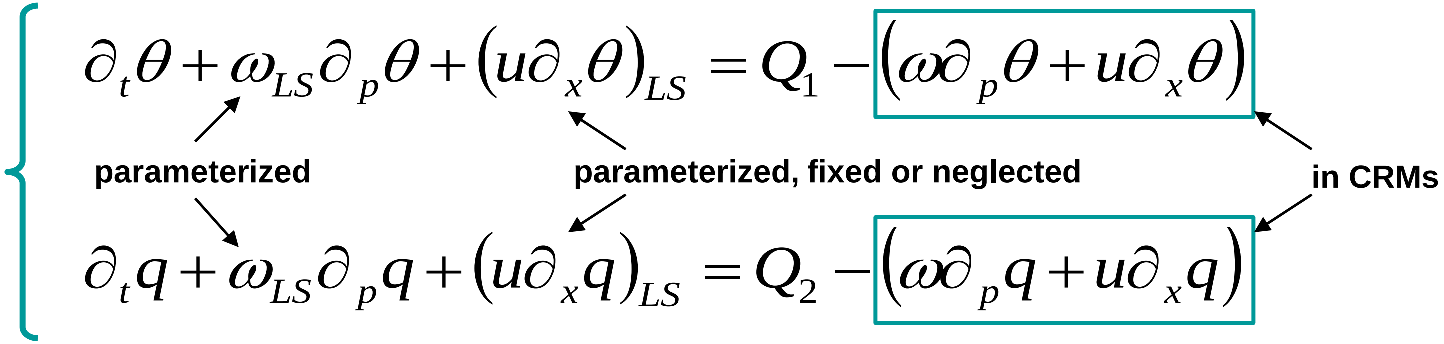
- **Best example : the shallow –to-deep-to-stratiform transition**



A simplified approach: parameterized large-scale *ω*

Hypthesis: parameterizing *ω* in a single-column model or CRM captures the main interaction between physics and dynamics.

Energy and water budgets are simplified:



$$\begin{cases} \partial_t\theta + \omega_{LS}\partial_p\theta + \left(u\partial_x\theta\right)_{LS} = Q_1 - \boxed{\left(\omega\partial_p\theta + u\partial_x\theta\right)} \\ \partial_t q + \omega_{LS}\partial_p q + \left(u\partial_x q\right)_{LS} = Q_2 - \boxed{\left(\omega\partial_p q + u\partial_x q\right)} \end{cases}$$

parameterized

parameterized, fixed or neglected

in CRMs

2 approaches:

- **Weak Temperature Gradient (WTG) approximation, based on the observation that large-scale free-tropospheric gradients of temperature are small in the tropics;**
- **Damped Gravity-Wave (DGW) , based on a simplified gravity wave equation.**

Weak Temperature Gradient approximation

Bretherton and Sobel (2000)

- **Strict: the free_tropospheric temperature is vertically uniform**

$$\theta = \theta_r(p) \quad \partial_x \theta_{LS} = 0$$

$$\underset{0}{\cancel{\partial_t \theta}} + \omega_{LS} \partial_p \theta_r + \underset{0}{\cancel{(u \partial_x \theta)_{LS}}} = Q_1 \quad \Rightarrow \quad \boxed{\omega_{LS} = \frac{Q_1}{\partial_p \theta_r}}$$

heating ⇒ ascent

cooling ⇒ descent

- **Relaxed:**

$$\boxed{\omega_{LS} \partial_p \theta_{LS} = -\frac{\theta_{LS} - \theta_r}{\tau}}$$

warmer than reference ⇒ ascent

cooler than reference ⇒ descent

- **In the boundary layer: uniform divergence over a nominal boundary layer depth**

Damped gravity wave approximation

Kuang (2008)

Linearized, non-rotating primitive equations:

$$\left\{\begin{aligned} \partial_t u &= -\partial_x \Phi - \varepsilon u \\ \partial_x u + \partial_p \omega &= 0 \\ \partial_p \Phi &= -\frac{RT_v}{p} \end{aligned}\right. \quad \Longrightarrow \quad \partial_p\left[(\partial_t + \varepsilon)\partial_p \omega\right] = -\frac{R}{p}\partial_x^2 T_v \qquad \text{(1)}$$

Looking for a wave solution: $(\omega, T_v) = (\omega_{LS}, T_{vLS})e^{ikx}$

$$\Longrightarrow \quad \partial_p\left[(\partial_t + \varepsilon)\partial_p \omega_{LS}\right] = \frac{R}{p}k^2 T_{vLS}$$

Assuming the variations of *ω* are slow, and a base state $(\omega, T_v) = (0, T_{vr})$**:**

$$\boxed{\partial_p\left[\varepsilon \partial_p \omega_{LS}\right] = \frac{R}{p}k^2 (T_{vLS} - T_{vr})}$$

warmer than reference $\Rightarrow$ **increase of convergence with *p***

cooler than reference $\Rightarrow$ **decrease of convergence with *p***

Note: **The Weak Pressure Gradient (WPG) approximation is a similar approach up to (1), but with Boussinesq equations**

Romps (2012)

General case: pressure gradients and diabatic heating

Using the hydrostatic approximation:

$$dp = -\rho g dz = -\frac{gp}{RT_v} dz \Rightarrow \ln\left(\frac{p}{p_t}\right) = \frac{g}{R}\int_z^{z_t} \frac{dz}{T_v}$$

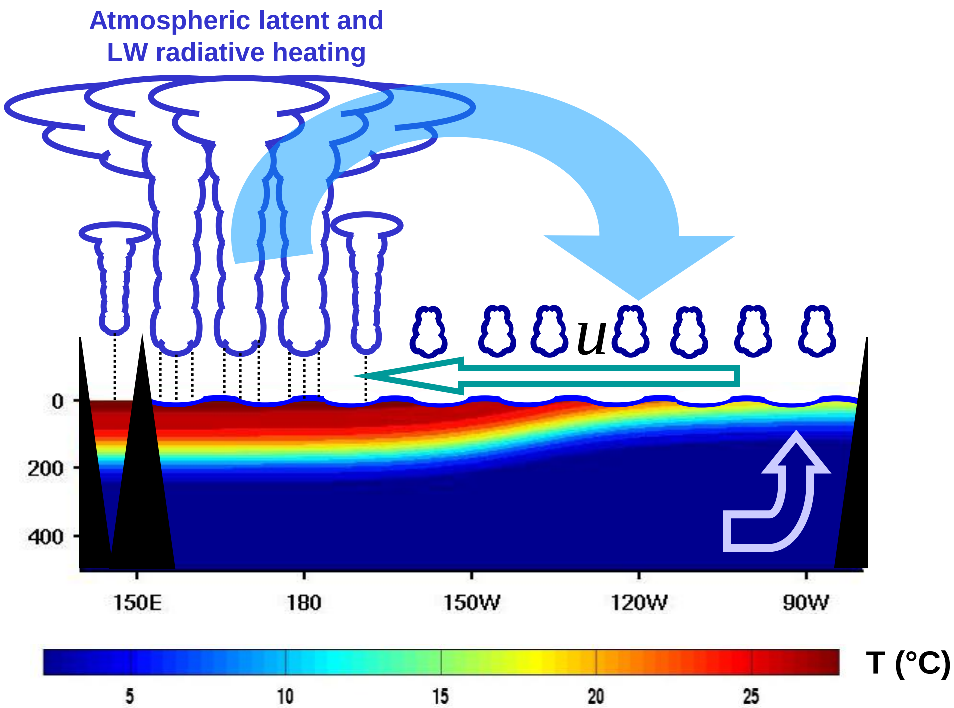
We can write the effect of clouds on the pressure gradients as follow:

$$\partial_t^{cl} \nabla p = -p \frac{g}{R} \int_z^{z_t} \frac{\partial_t^{cl} \nabla T_v}{{T_v}^2} dz$$

The tendency of the pressure gradients resulting from the cloud diabatic effect is proportional to the integral of temperature tendency above, weighted by $1 / {T_v}^2$

Walker circulation

Walker circulation



Very simple model of the surface zonal wind:

$$\partial_t u = 0 = -\frac{1}{\rho_s}\partial_x p_s - \mu u$$

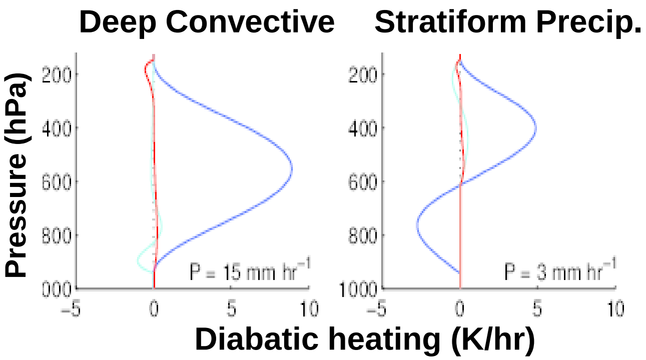
$$u = \frac{g}{\mu} T_{vs} \int_0^{z_t} \frac{\partial_x T_v}{T_v^2} dz$$

$$\partial_t^{cl} \partial_x T_v < 0$$

Latent heat release and LW radiative heating above the warm pool tend to reinforce the easterlies.

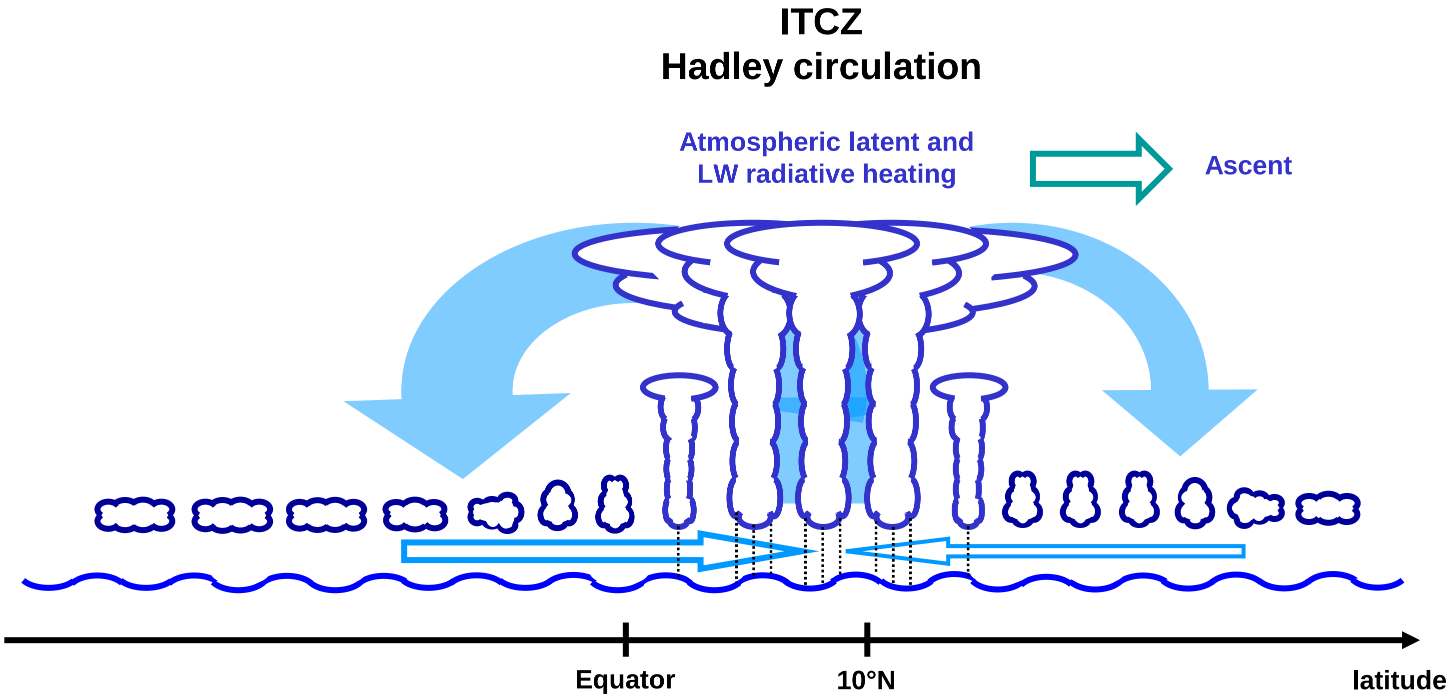
Exercise:

Does the Walker circulation increase or decrease if the fraction of stratiform/convective precipitation increase?



Intertropical convergence zone

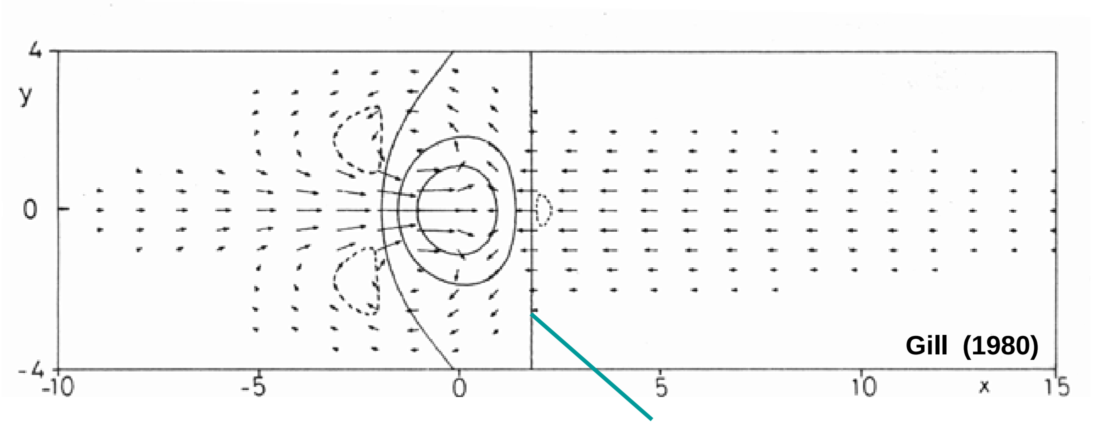
Latent heat release and LW radiative heating in the ITCZ reinforces the Hadley circulation, and more so if the ITCZ is away from the equator.



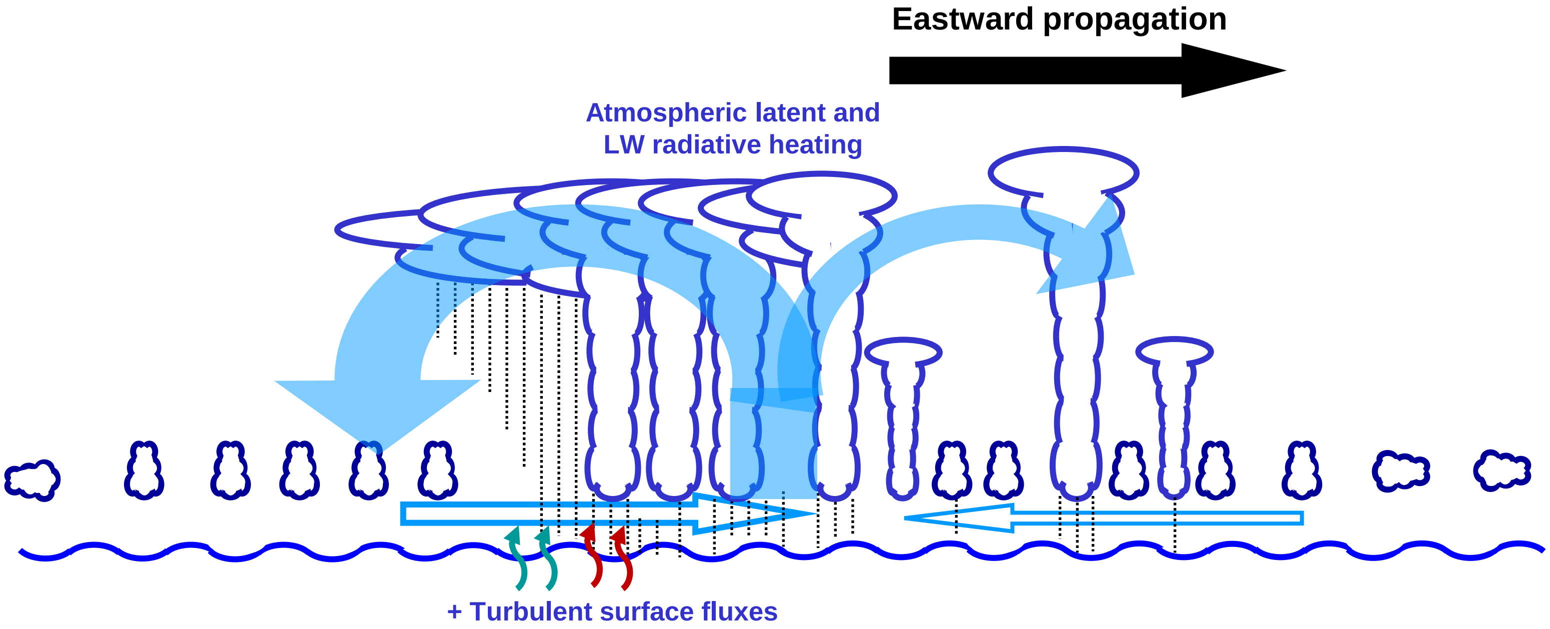
Madden-Julian Oscillation

Linearized shallow-water equations on a β-plane with heating and dissipation:

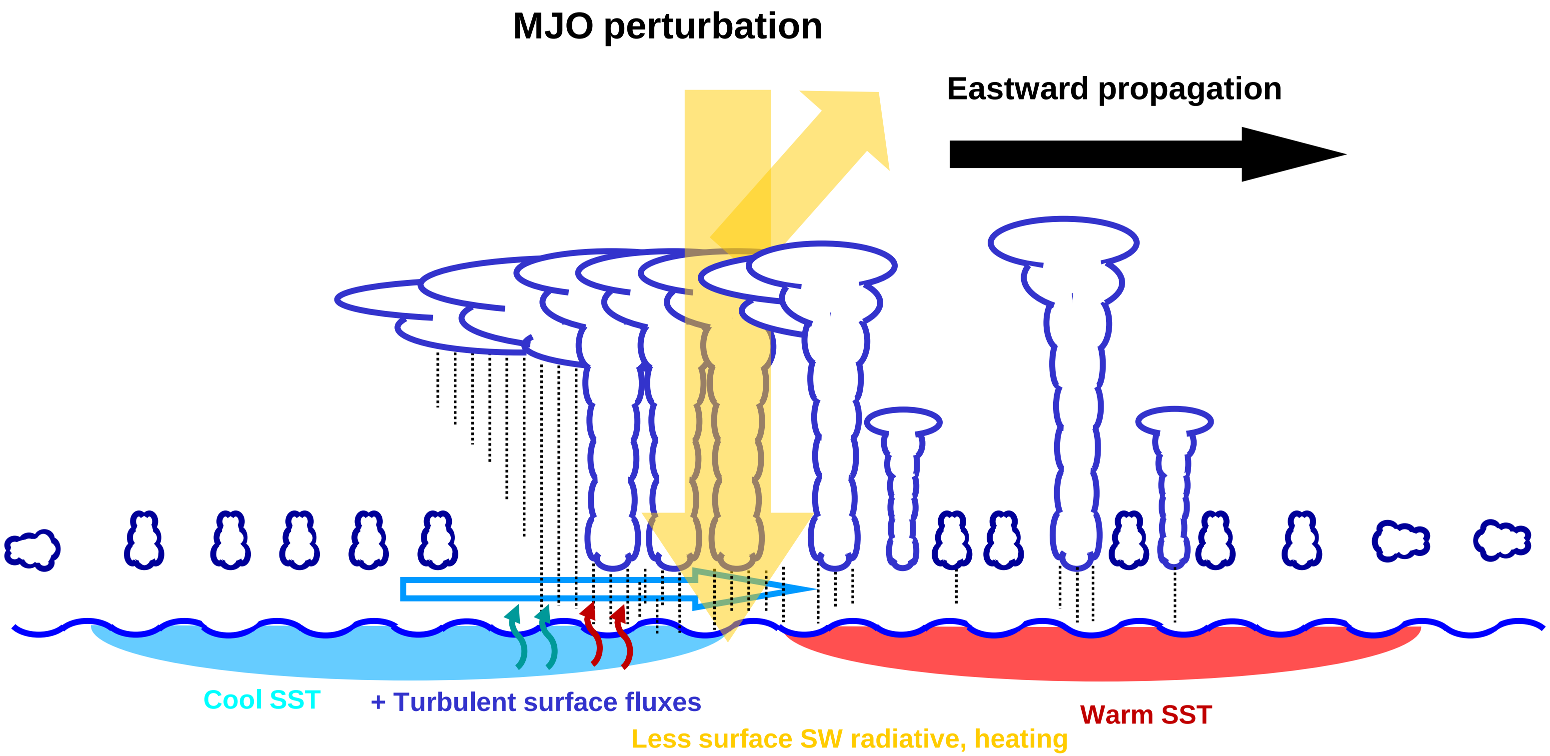
$$\begin{cases} \partial_t u - \beta y v = -\alpha \partial_x T_m - \mu u \\ \partial_t v + \beta y u = -\alpha \partial_y T_m - \mu v \\ \partial_t T_m + \Delta S \left(\partial_x u + \partial_y v \right) = Q - \mu T_m \end{cases}$$



MJO perturbation



Madden-Julian Oscillation



MJO perturbation

Eastward propagation

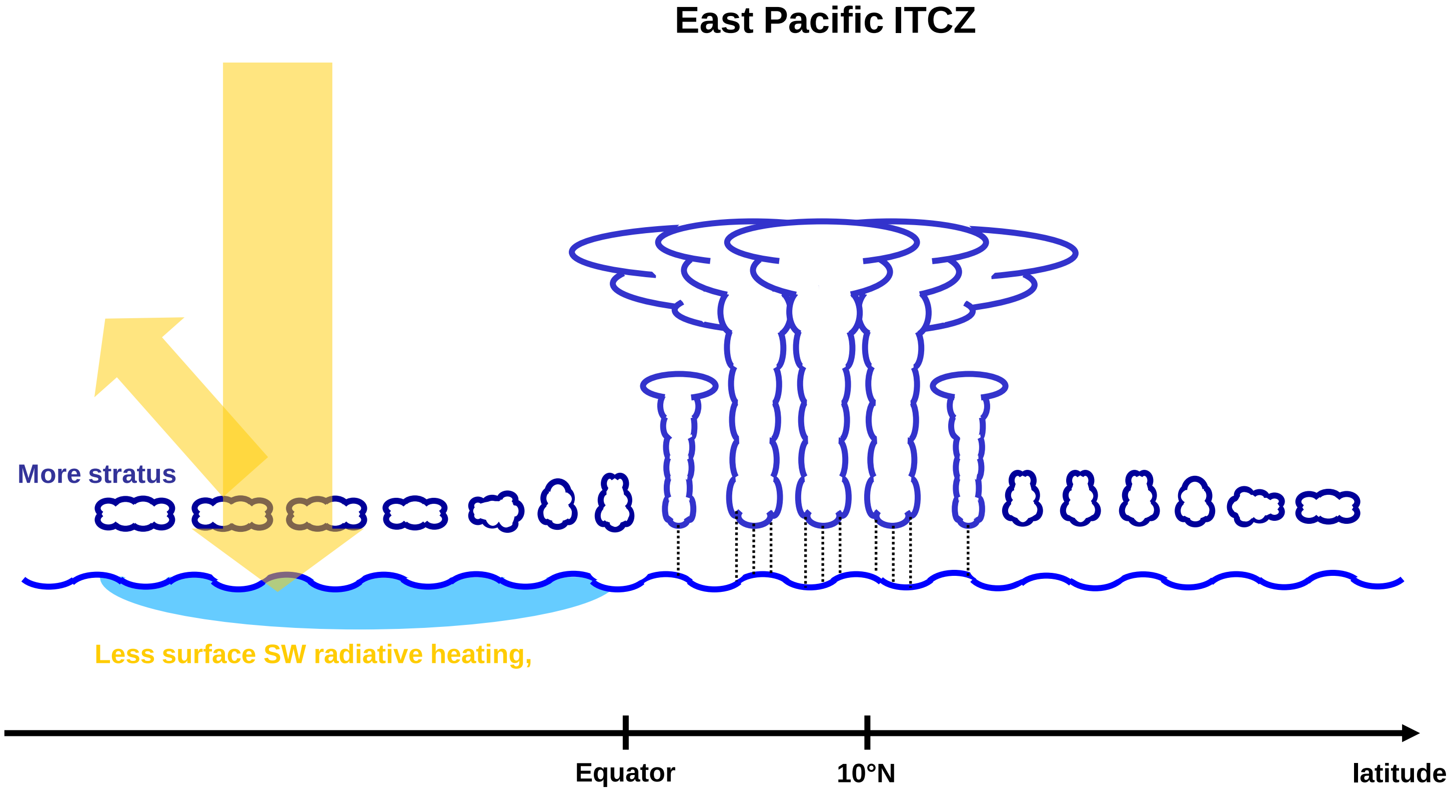
Cool SST

+ Turbulent surface fluxes

Less surface SW radiative, heating

Warm SST

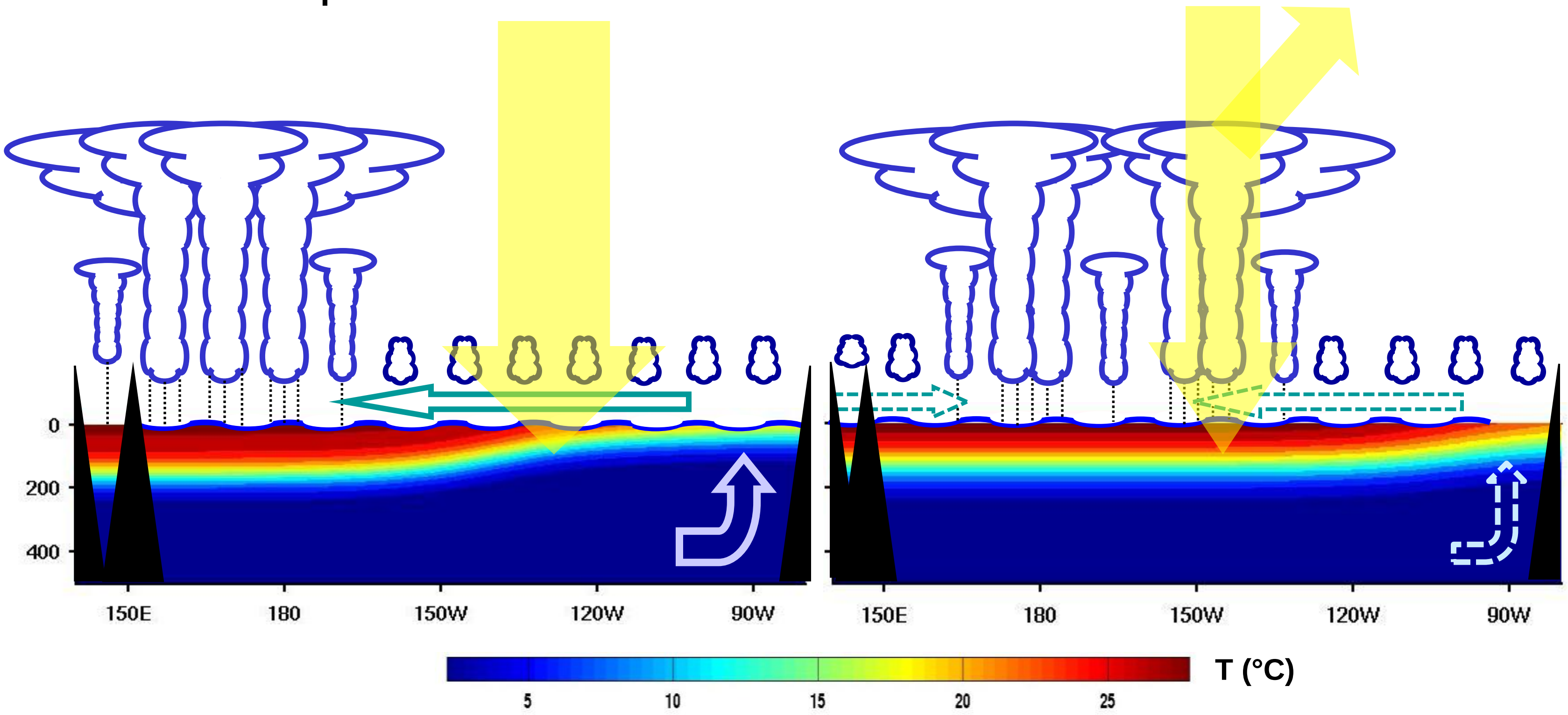
Intertropical convergence zone



El Niño – Southern Oscillation

Normal conditions
in the equatorial Pacific

El Niño conditions



2 feedbacks:

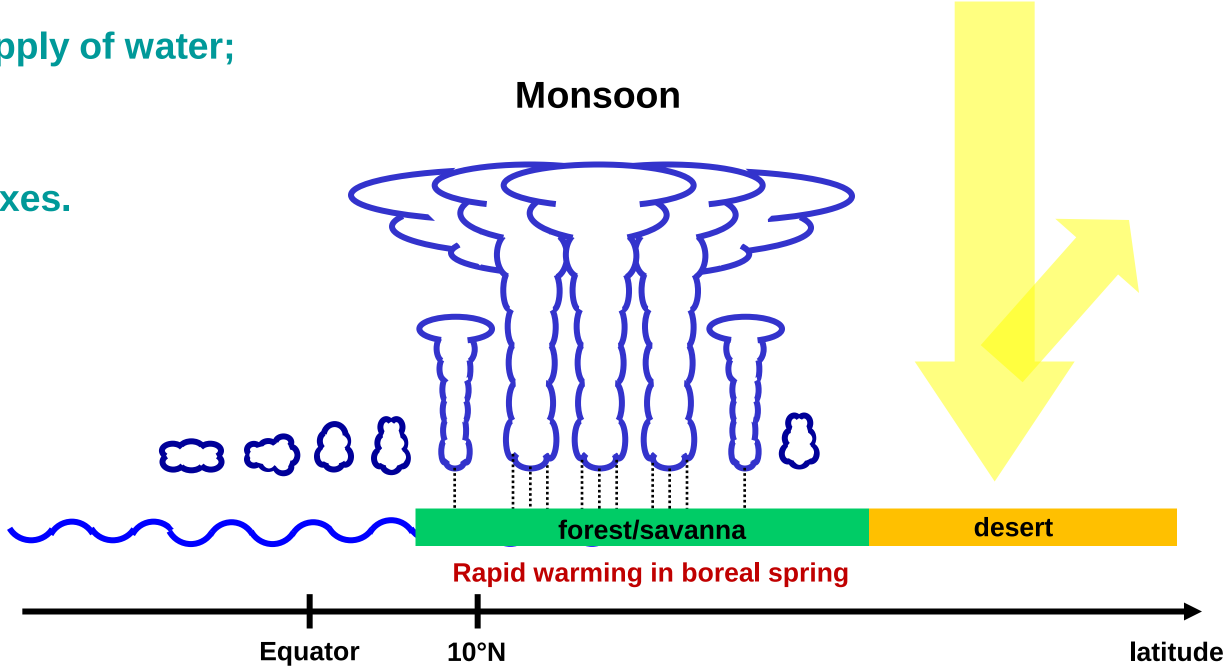
- **Bjerknes feedback: decreased easterlies because of decreased SST gradients (positive)**
- **Surface fluxes feedback: decreased incoming solar radiation due to clouds, increased evaporation (negative)**

Much more diversity of behaviors

Things that matter over land surfaces (compared to oceanic surfaces):

- **Small heat capacity of the surface;**
- **Finite storage and supply of water;**
- **Larger albedo;**
- **Modified turbulent fluxes.**

Often competing effects!



Many questions are still unanswered.

Beyond the interaction between land surfaces and atmosphere, the interaction between continent, ocean and atmosphere is a challenging issue.

See Cathy's talk!

Perspectives

Current questions:

- To understand the interaction between clouds and circulation:

  vertical profiles of latent and radiative heating, of moistening.
- To understand the coupled ocean-atmosphere interaction:

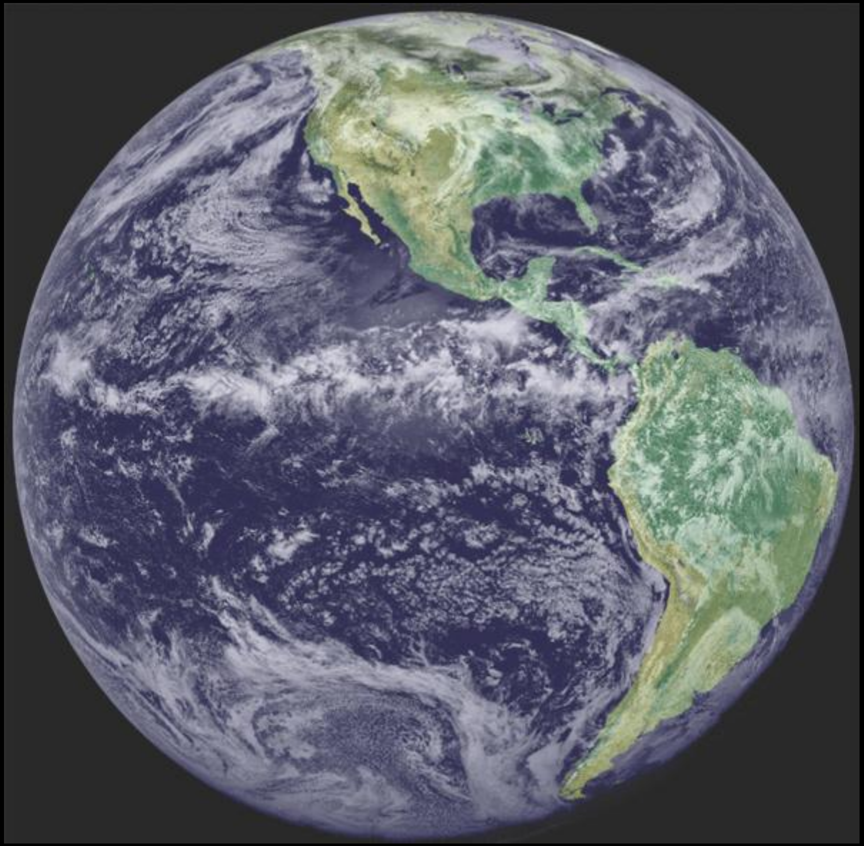
  oceanic mixed-layer response to atmospheric variability and feedbacks

Specific simple but challenging situations to simulate with a GCM:

- Persistence of stratocumulus decks;
- Stratocumulus-cumulus transition;
- Shallow-to-deep convection;
- Large-scale organisation of convection.

Complex systems:

- MJO;
- Continent-ocean-atmosphere systems.



Thank you