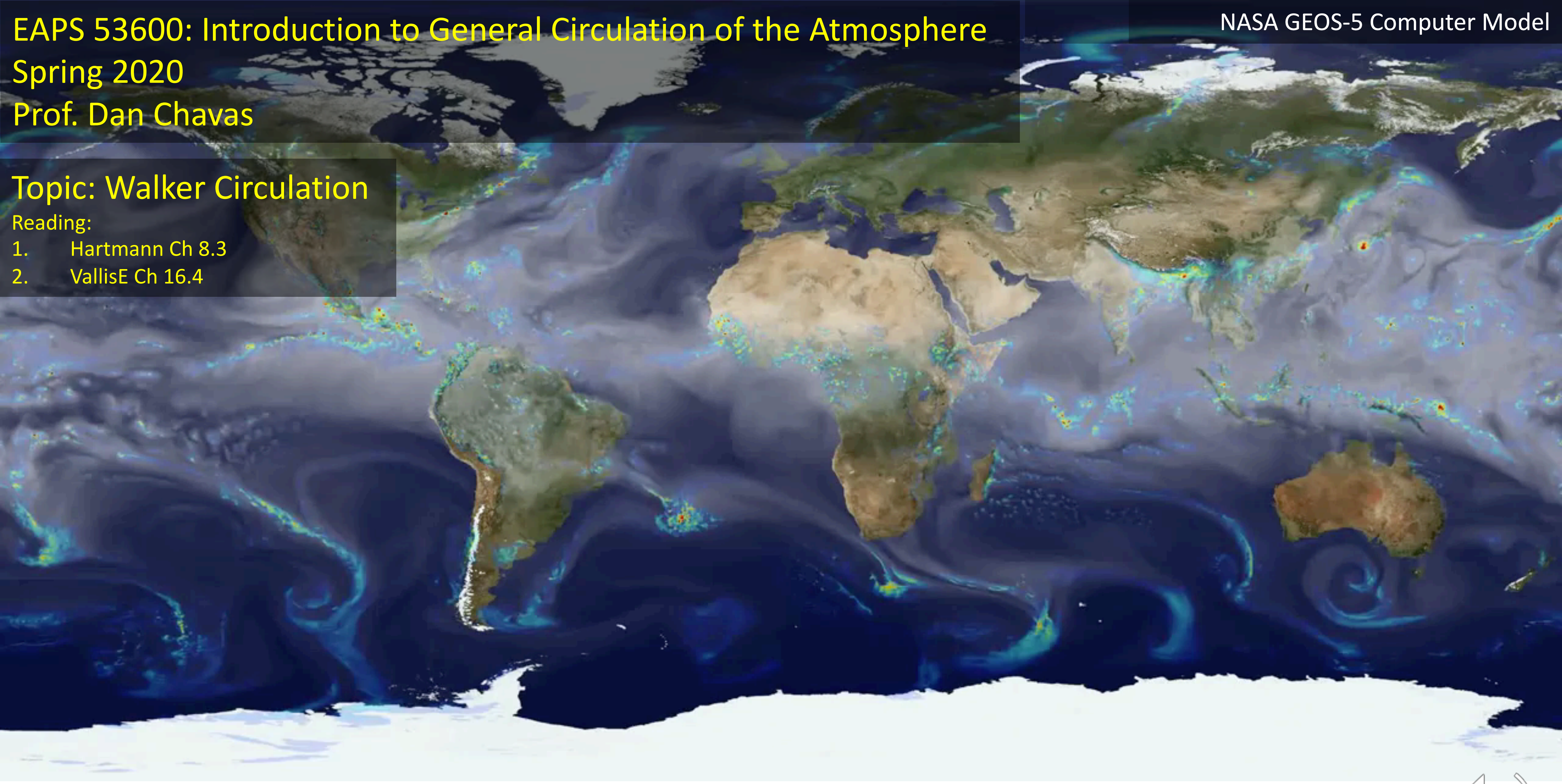
EAPS 53600: Introduction to General Circulation of the Atmosphere
Spring 2020
Prof. Dan Chavas

Topic: Walker Circulation

Reading:
1. Hartmann Ch 8.3
2. VallisE Ch 16.4



White: total precipitable water (brigher white = more water vapor in column)
Colors: precipitation rate ($0 - 15\frac{mm}{hr}$, red=highest)

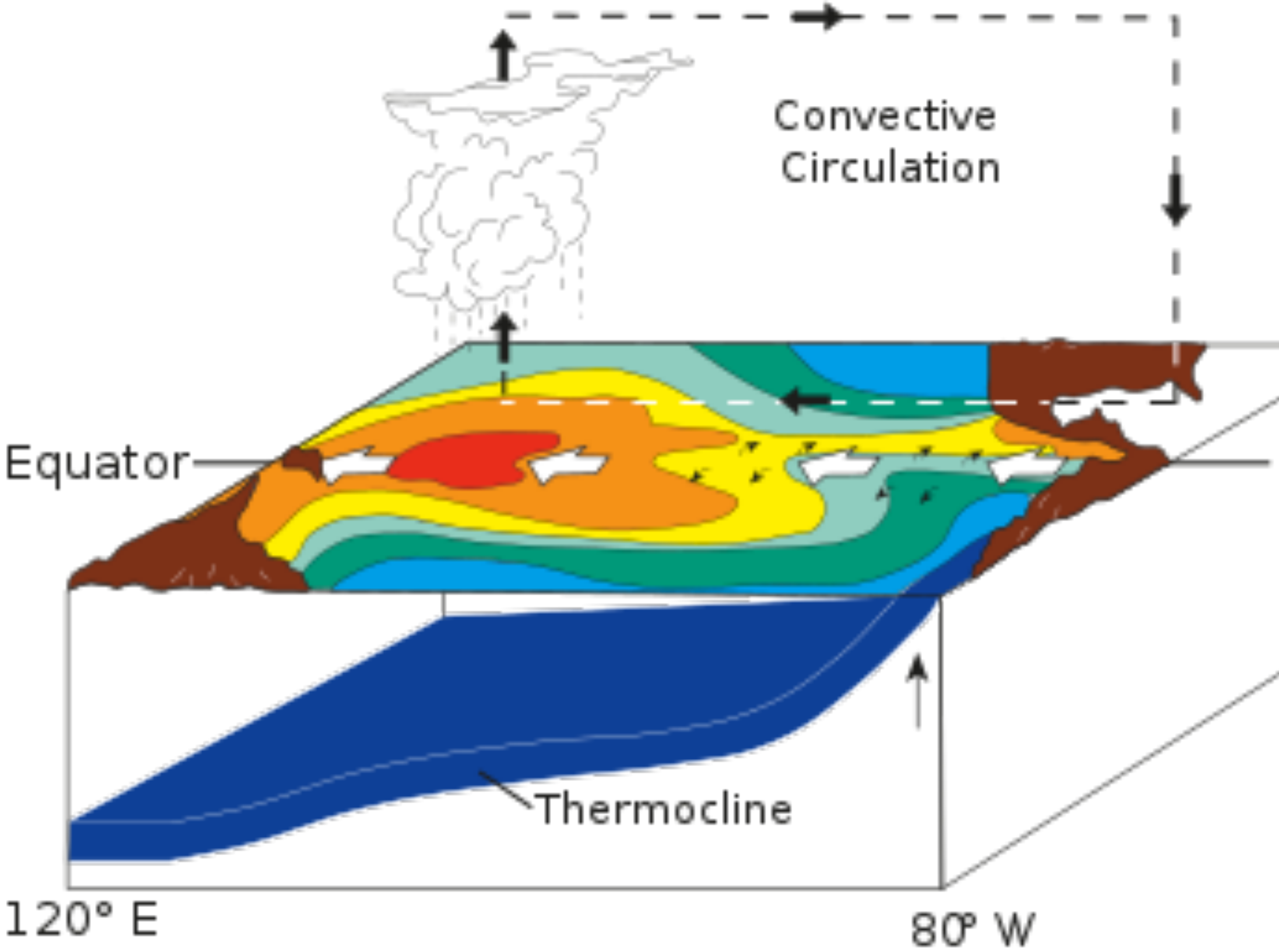
Source: https://svs.gsfc.nasa.gov/cgi-bin/details.cgi?aid=30017

Learning outcomes for today:

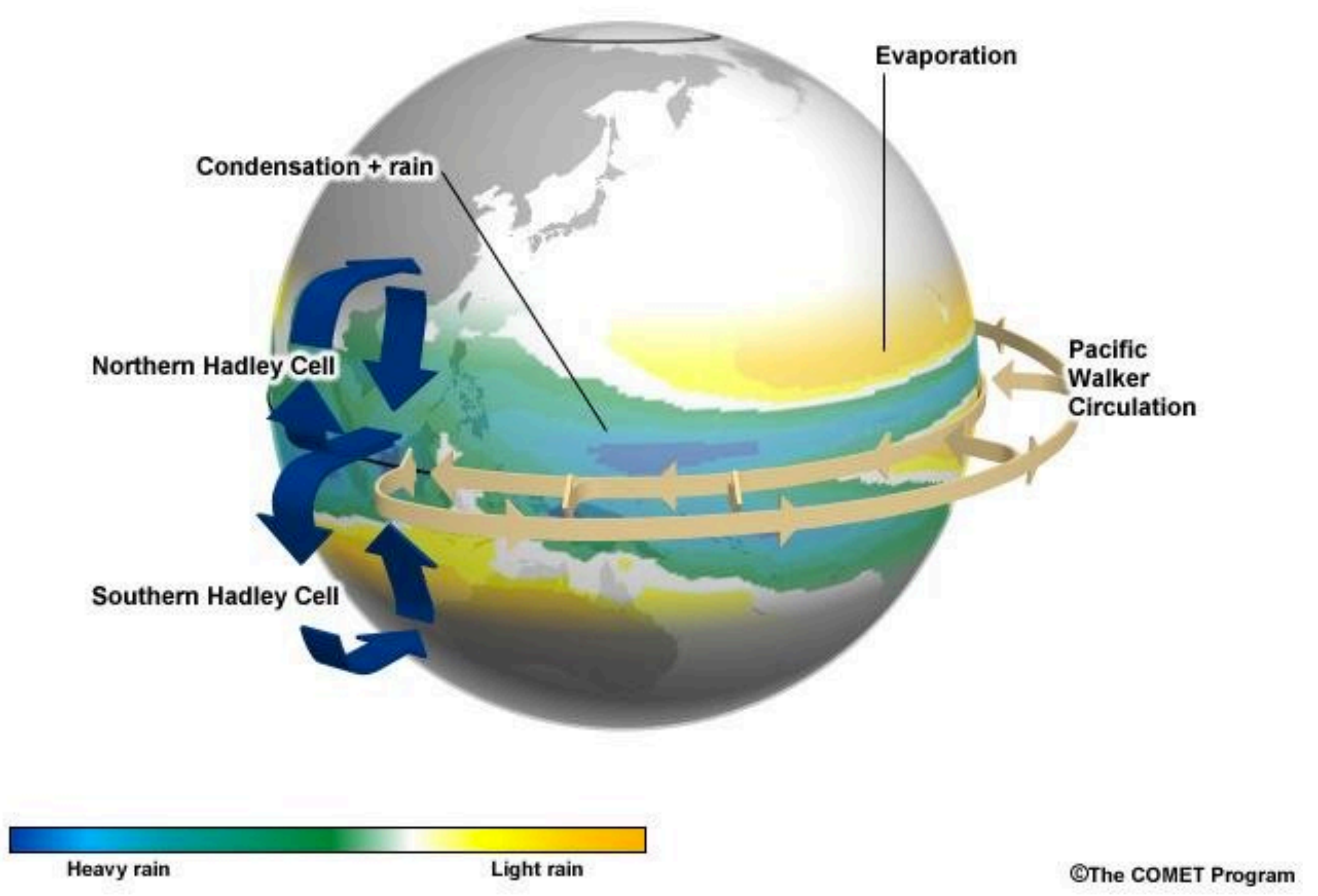
- **Describe what the Walker circulation is and why it changes with ENSO**
- **Explain the vertical structure of temperature and moisture in the tropics**
- **Explain how temperature and moisture vary horizontally in the tropics**
- **Explain how the Walker circulation can be understood conceptually using a 2-layer, 2-column model**

Basics

Walker Circulation: air rises where it is relatively warm – over the West Pacific Warm Pool

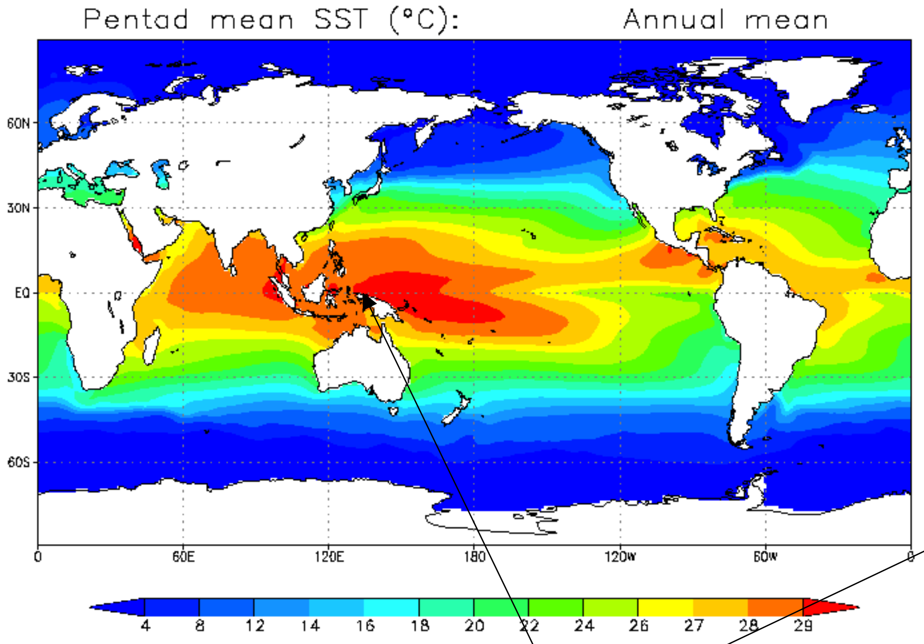


https://en.wikipedia.org/wiki/El_Ni%C3%B1o%E2%80%93Southern_Oscillation

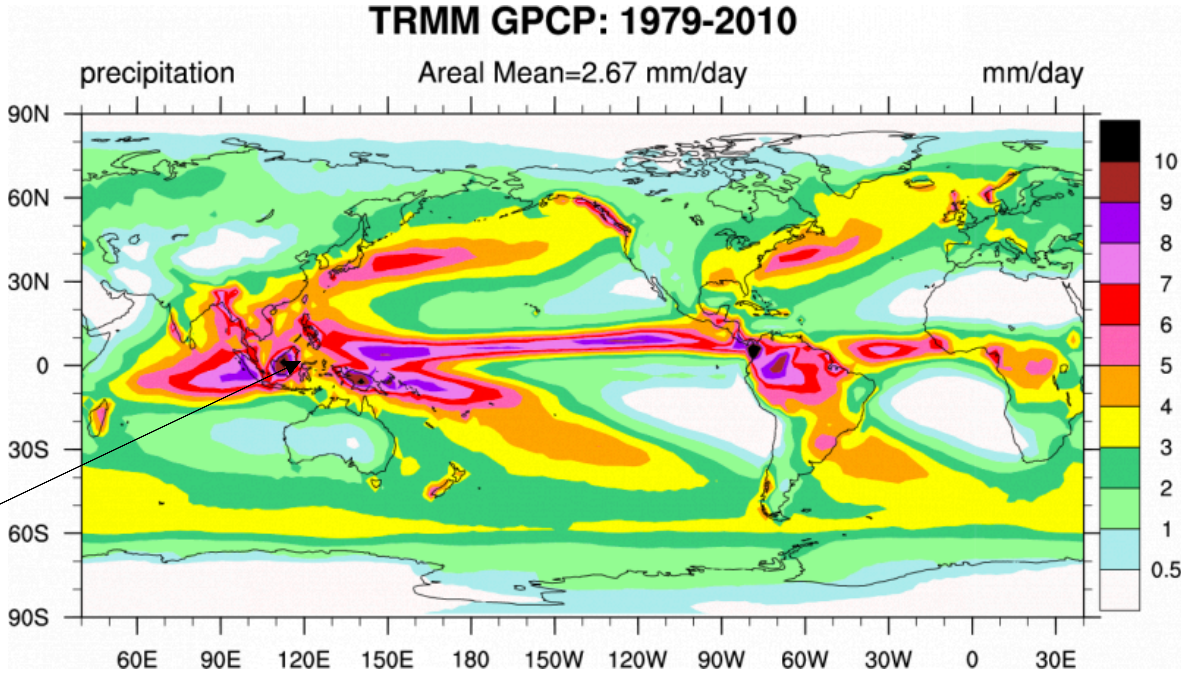


https://www.meted.ucar.edu/training_module.php?id=738#.XoMcaZNKgWo

Annual-mean sea surface temperature



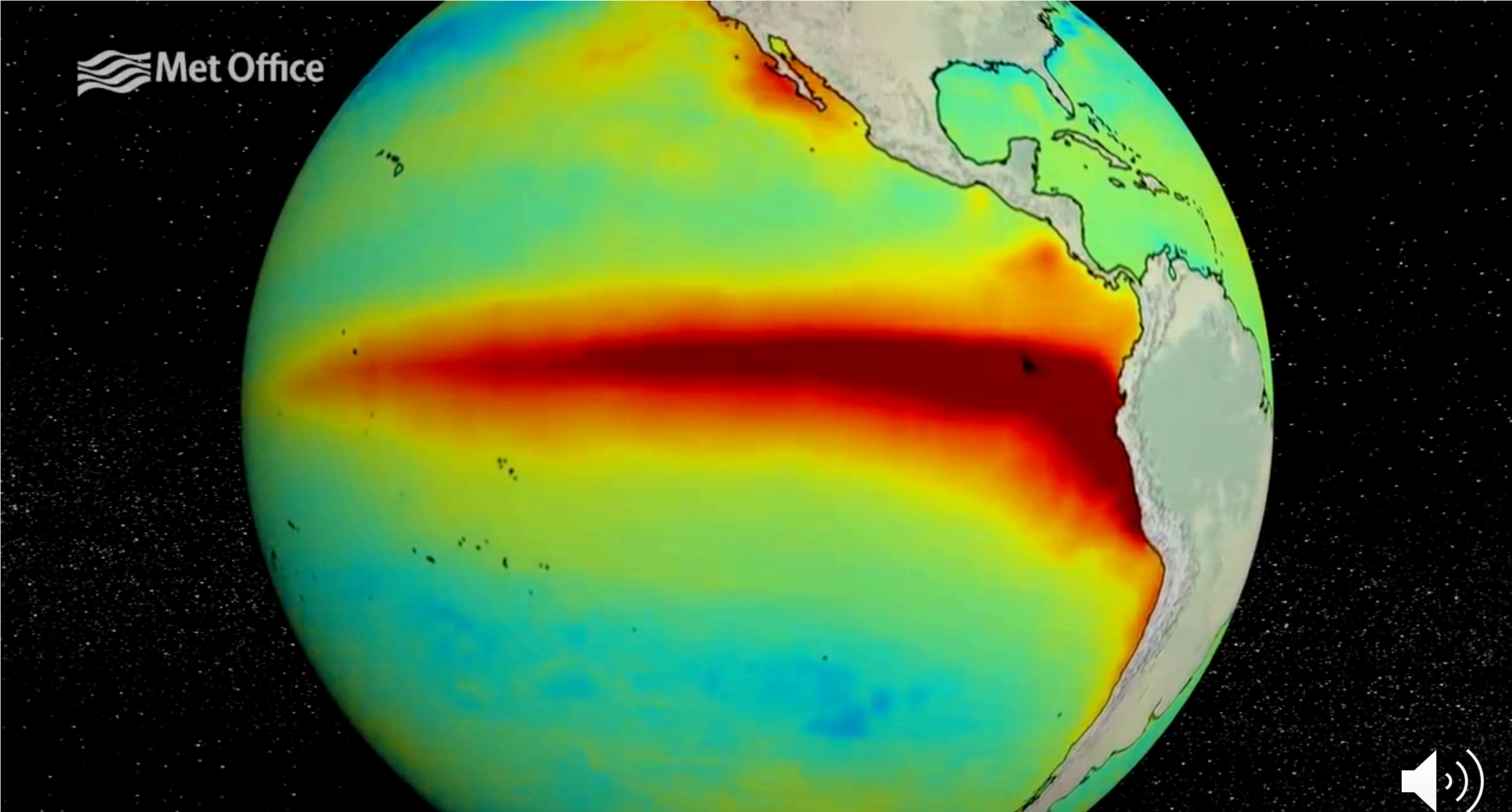
Annual-mean rainfall



"Maritime continent" /
West Pacific Warm Pool

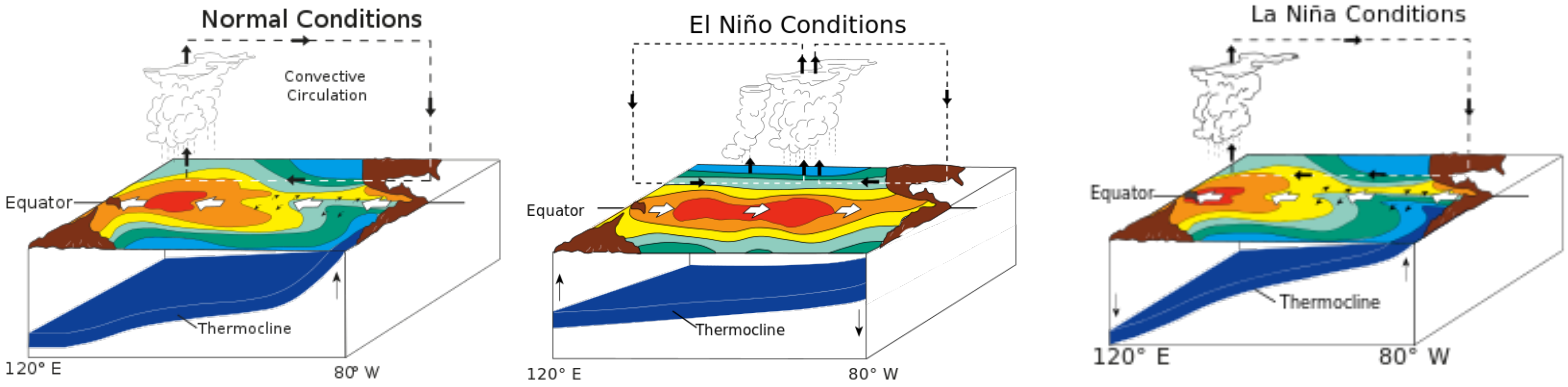
Variability in the walker circulation: El Nino Southern Oscillation (ENSO)

https://www.youtube.com/watch?v=WPA-KpldDVc



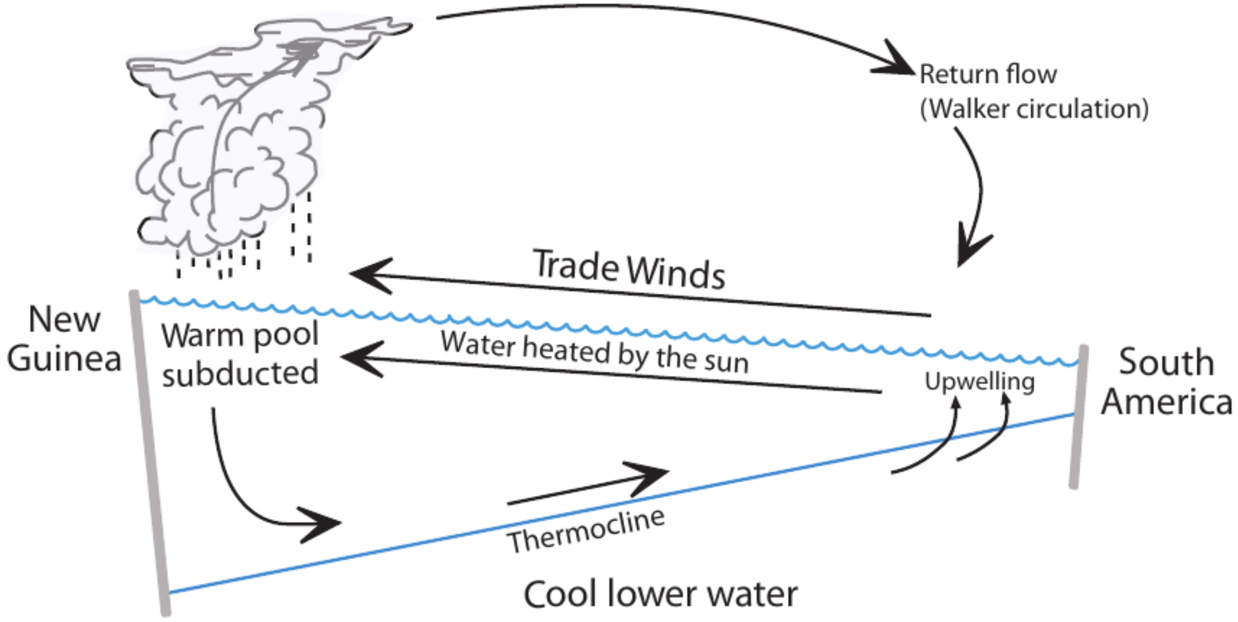
https://www.youtube.com/watch?v=WPA-KpldDVc

El Nino / La Nina: a **“sloshing” of the upper Pacific Ocean**

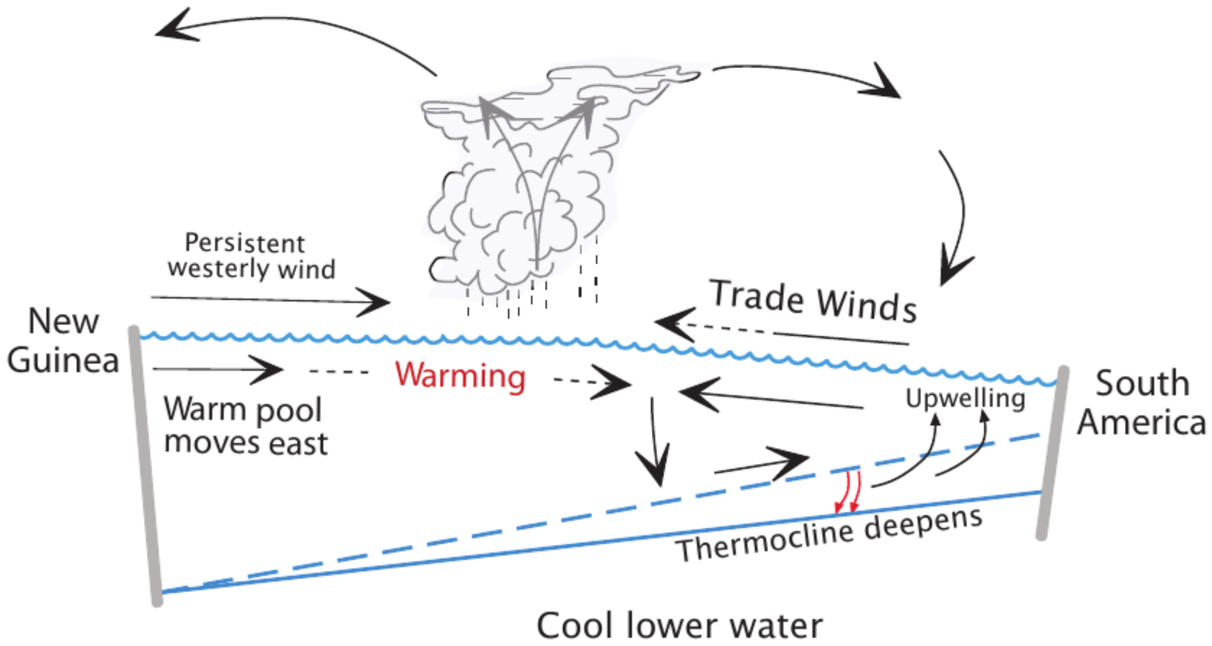


https://en.wikipedia.org/wiki/El_Ni%C3%B1o%E2%80%93Southern_Oscillation

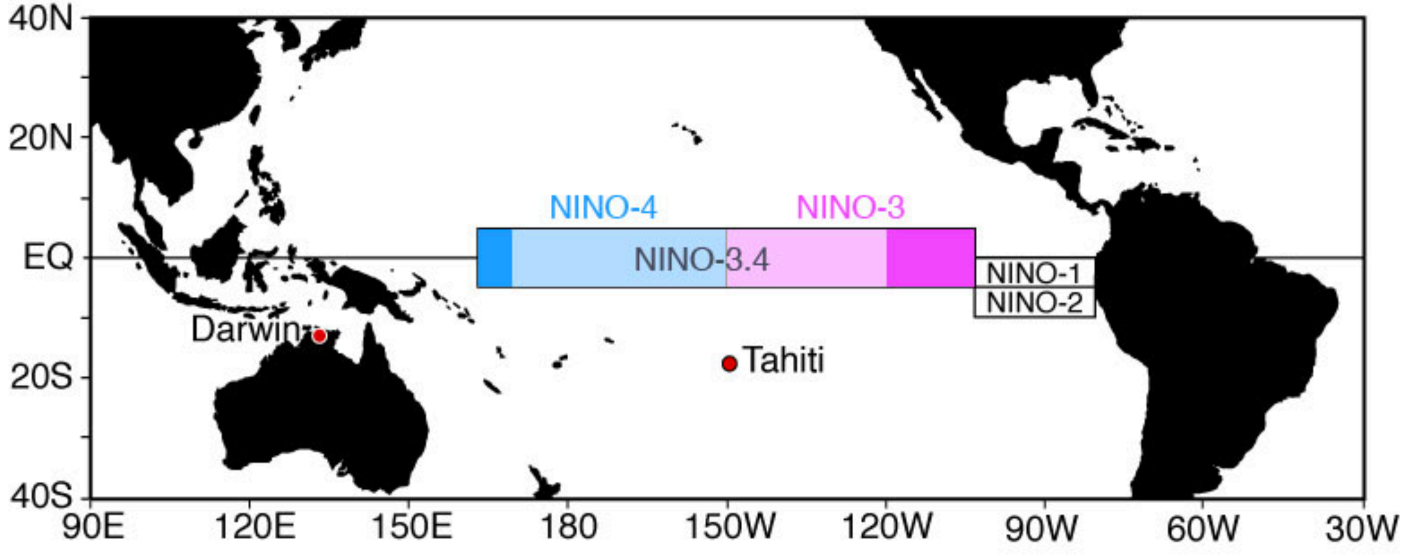
Fig. 16.10:



Return flow
(Walker circulation)
Trade Winds
New Guinea
Warm pool subducted
Water heated by the sun
Upwelling
South America
Thermocline
Cool lower water

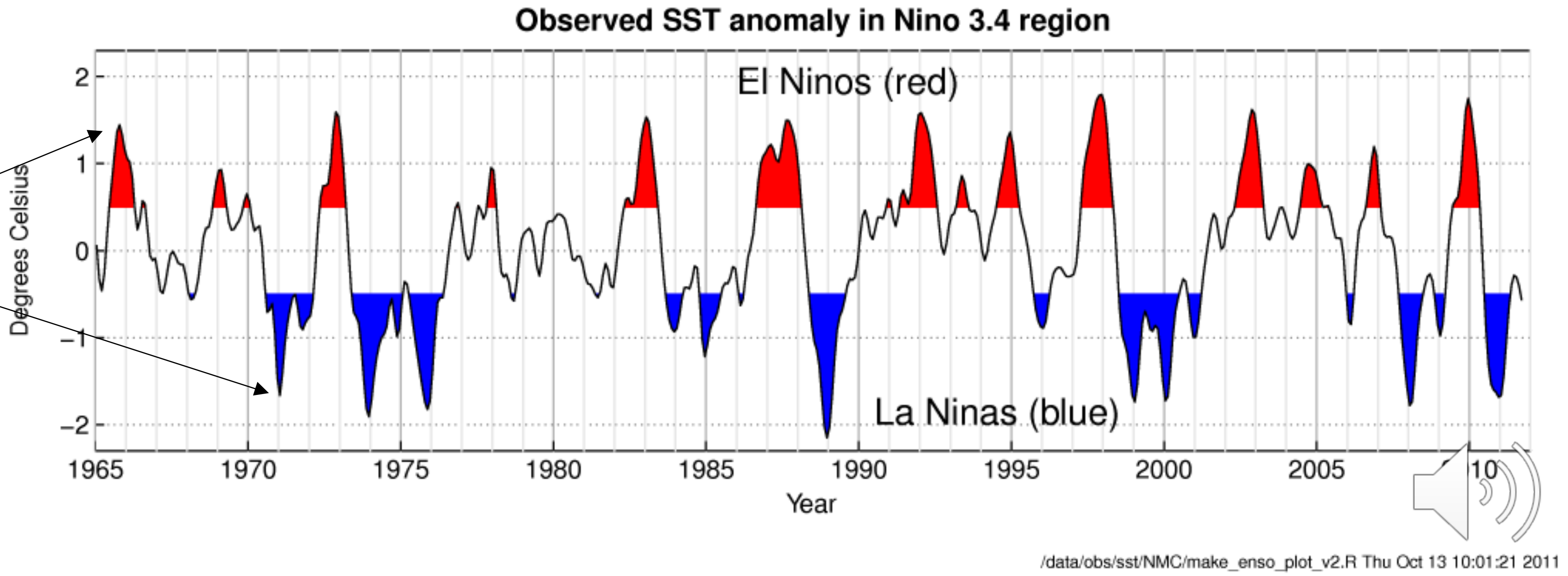


Persistent westerly wind
Trade Winds
New Guinea
Warming
Warm pool moves east
Upwelling
South America
Thermocline deepens
Cool lower water



Duration: $\tau \sim 6 - 12\ months$
Frequency: 2-7 years (non-periodic)

Typically peaks in November



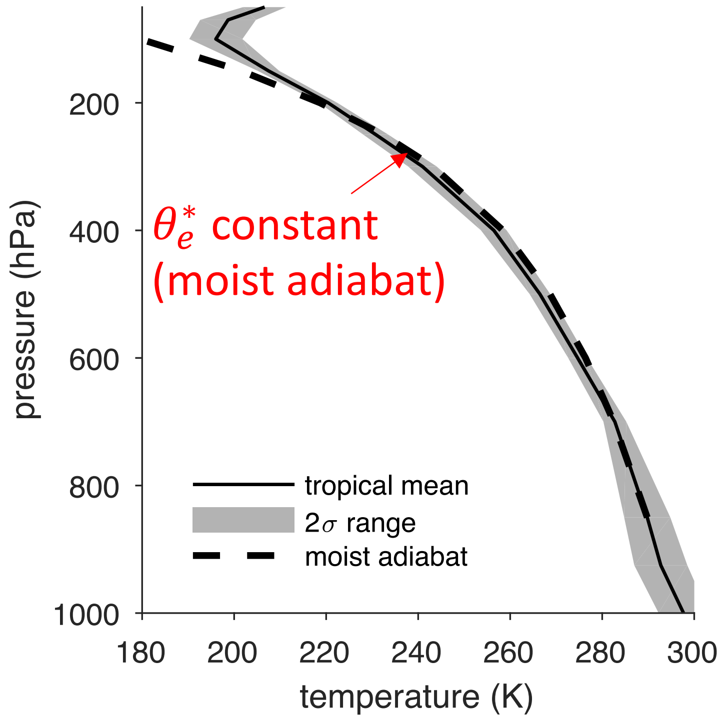
https://afriquetopnews.com/manual_el_nino_diagram_chart.php

A primer on the temperature and moisture in the tropics

1. Free-tropospheric temperature profile in the tropics
2. Moisture profile in the tropics
3. Horizontal temperature variations – free troposphere vs. boundary layer
4. How everything is linked to sea surface temperatures

The temperature profile in the tropics is nearly moist adiabatic

This means that the tropics are nearly neutral to saturated ascent



$*$ = saturated

$$\theta_e \approx \theta^{\frac{L_v r_v}{C_p T}} \quad [K]$$

$$\theta_e^* \approx \theta^{\frac{L_v r_v^*}{C_p T}} \quad [K]$$

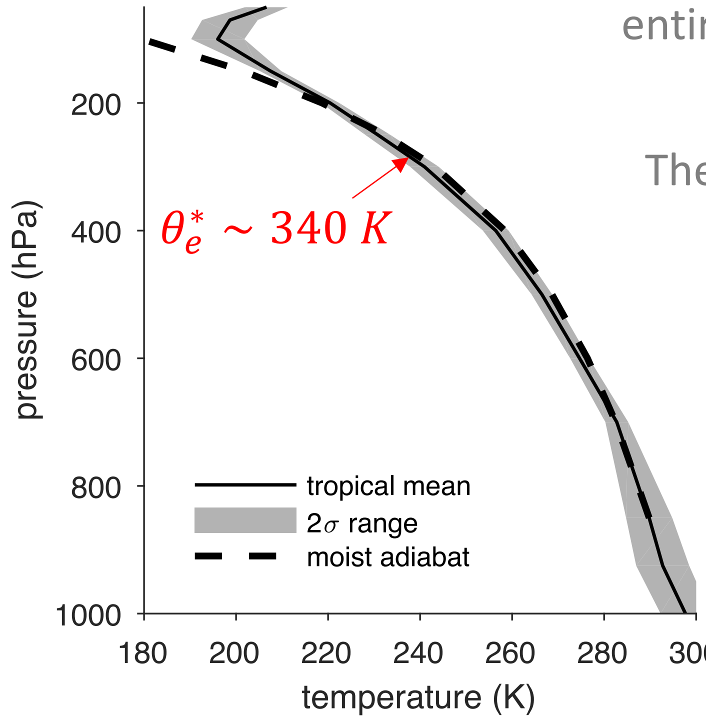
r_v^*: saturation mixing ratio

Figure 3.2: Mean temperature profile of the atmosphere for the tropical region (20°S-20°N) for the years 1981-2010 according to the NCEP-DOE reanalysis (solid) and temperature of a pseudoadiabatic parcel ascent lifted from saturation at 850 hPa and initialised at the tropical mean temperature (dashed). Shading represents the $\pm 2\sigma$ range of monthly temperatures for all months and all gridpoints in the tropical belt.

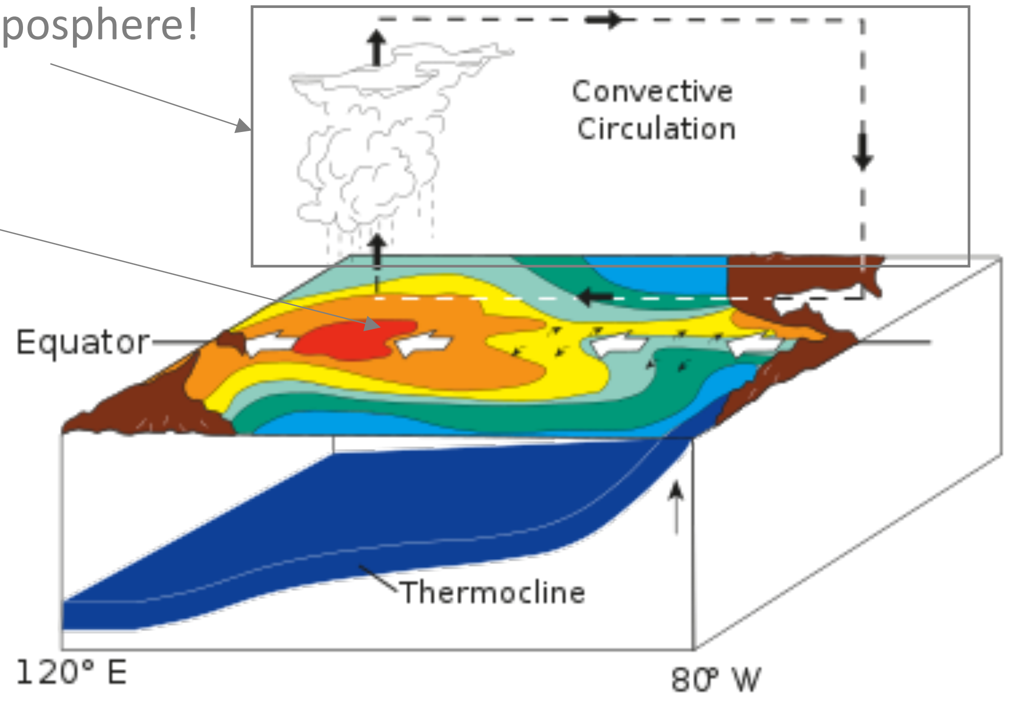
Source: Martin Singh (Monash)

But what determines *which* moist adiabat?

...sets the temperature profile of the entire tropical free troposphere!



The θ_e of boundary layer air here...

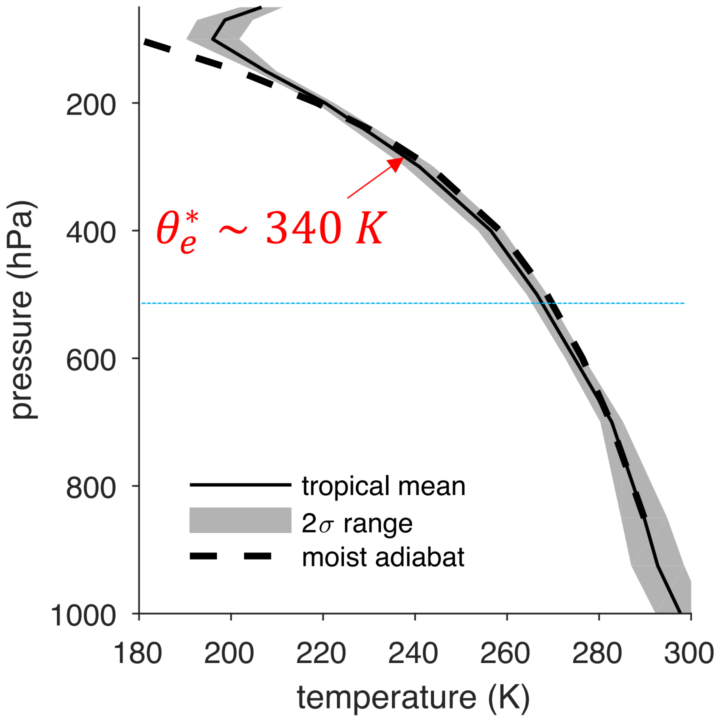


Air enters the free-troposphere primarily via **deep convective clouds**.
Deep convection is how the boundary layer **communicates** to the free troposphere.

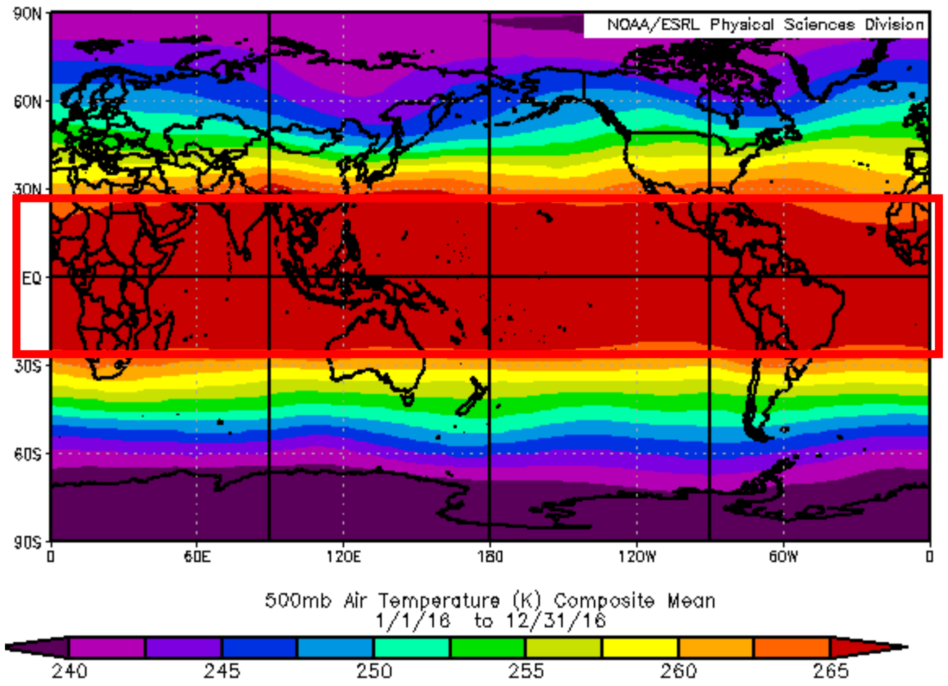
Thus, the properties of the **boundary layer air in regions of deep convection** set the **free-tropospheric temperature profile** across the **entire tropics**.

Tropical free-tropospheric temperatures cannot vary much horizontally because the Coriolis force is weak

(the "Weak Temperature Gradient" approximation)



500 hPa temperature (2016 mean)



Source: NCEP/NCAR reanalysis
(make this yourself at www.esrl.noaa.gov/psd/data/composites/day)

Physically: **gravity waves**, especially Kelvin waves, move fast and so will rapidly **smooth out horizontal pressure gradients** – and thus **temperature gradients** too, via hydrostatic balance.

What about *moisture*? Air in the tropics is driest in the middle of the free troposphere

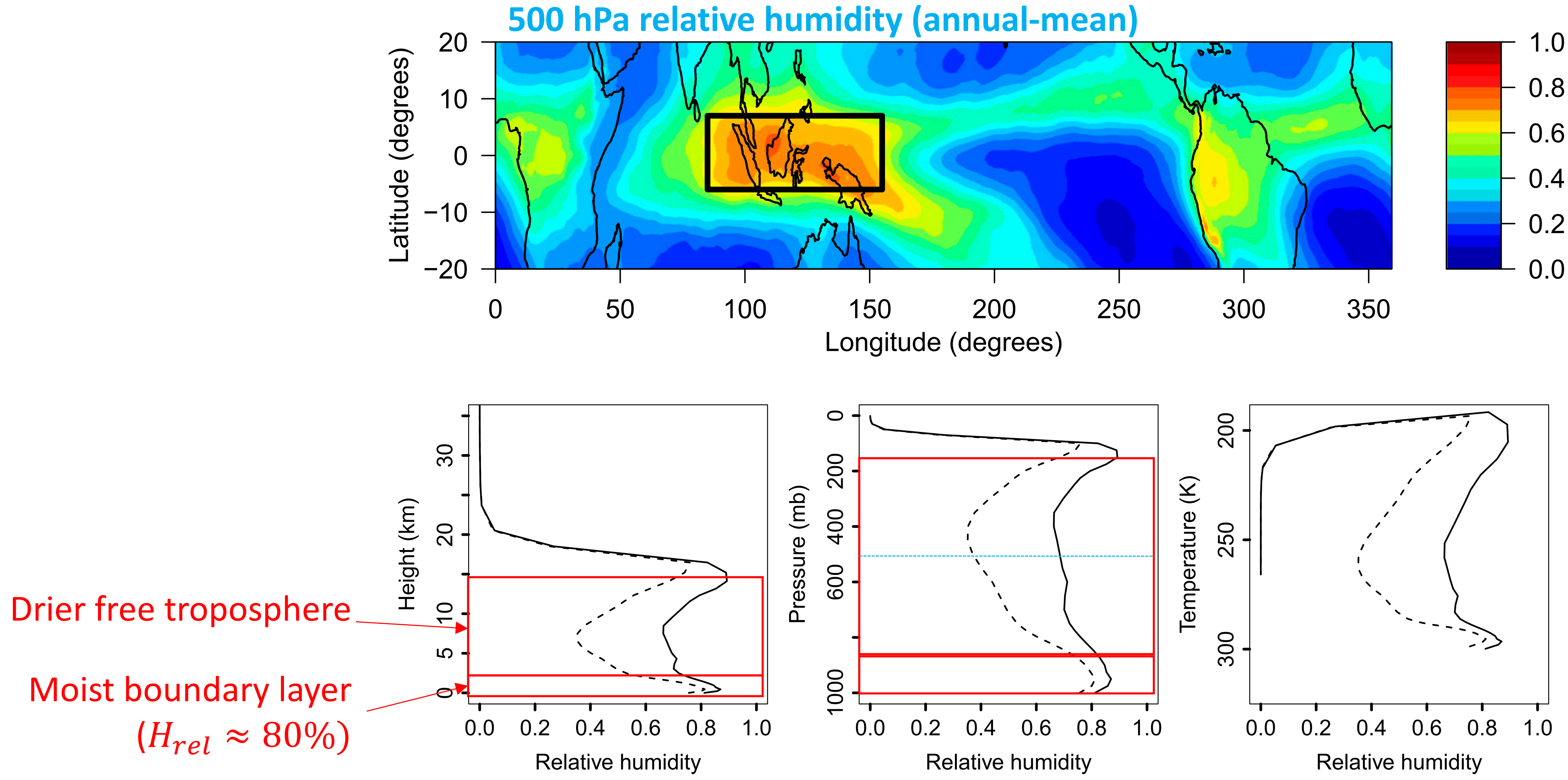
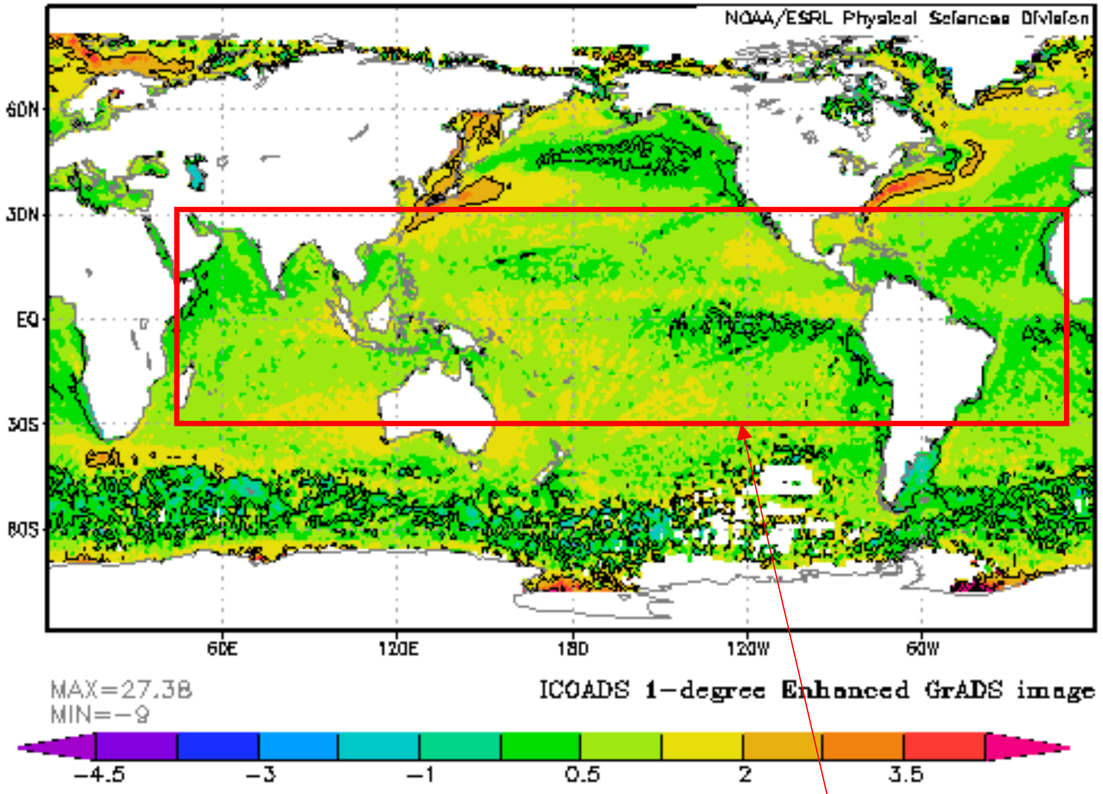


FIG. 1. (top) Map of annual-mean relative humidity at 500 hPa in the tropics from ERA-Interim during the year 2013. (bottom) Mean profiles of relative humidity for the Indo-Pacific warm pool (solid; averaged over the black box in the top panel) and the entire tropical domain (dashed, 20°S–20°N) as a function of (left) height, (center) pressure, and (right) temperature.

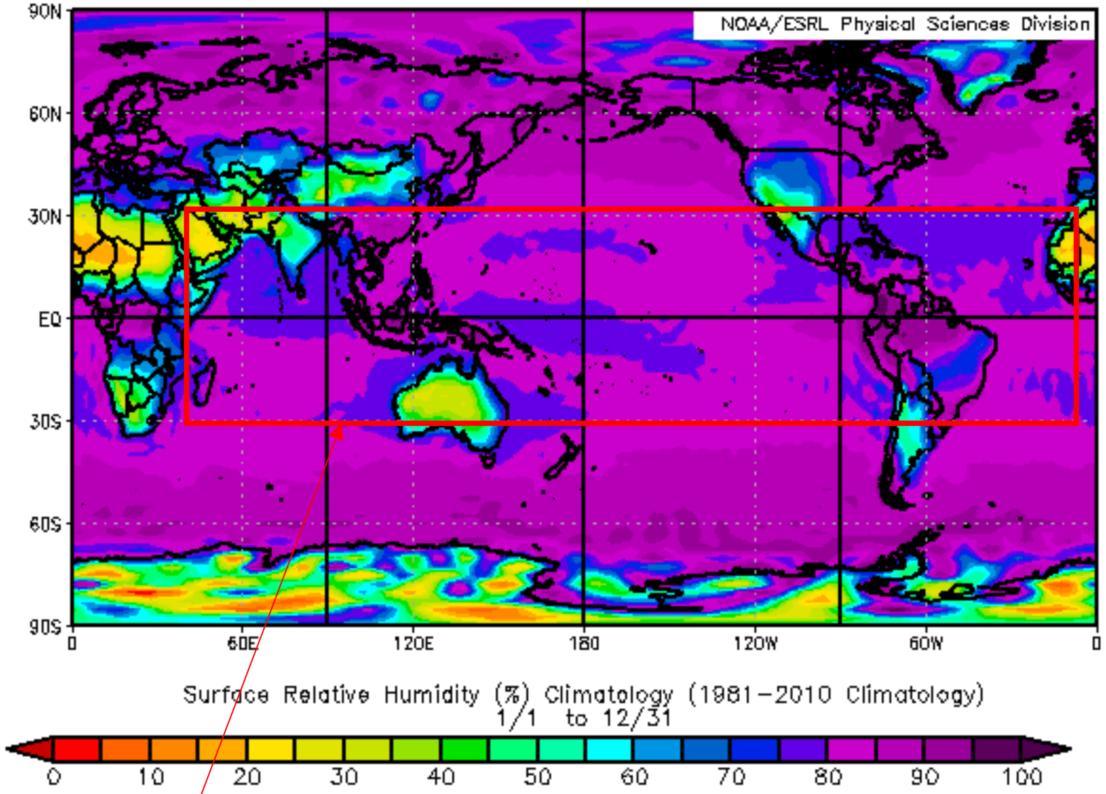
Romps (2014, J. Clim.)
"An Analytical Model for Tropical Relative Humidity"

Boundary layer θ_e (or moist static energy) is strongly controlled by sea surface temperatures

$T_{sst} - T_{2m}$ [K] (1981-2010 mean)



2m H_{rel} [%] (1981-2010 mean)



Source: ICOADS
(make this yourself at https://www.esrl.noaa.gov/psd/cgi-bin/db_search/SearchMenus.pl)

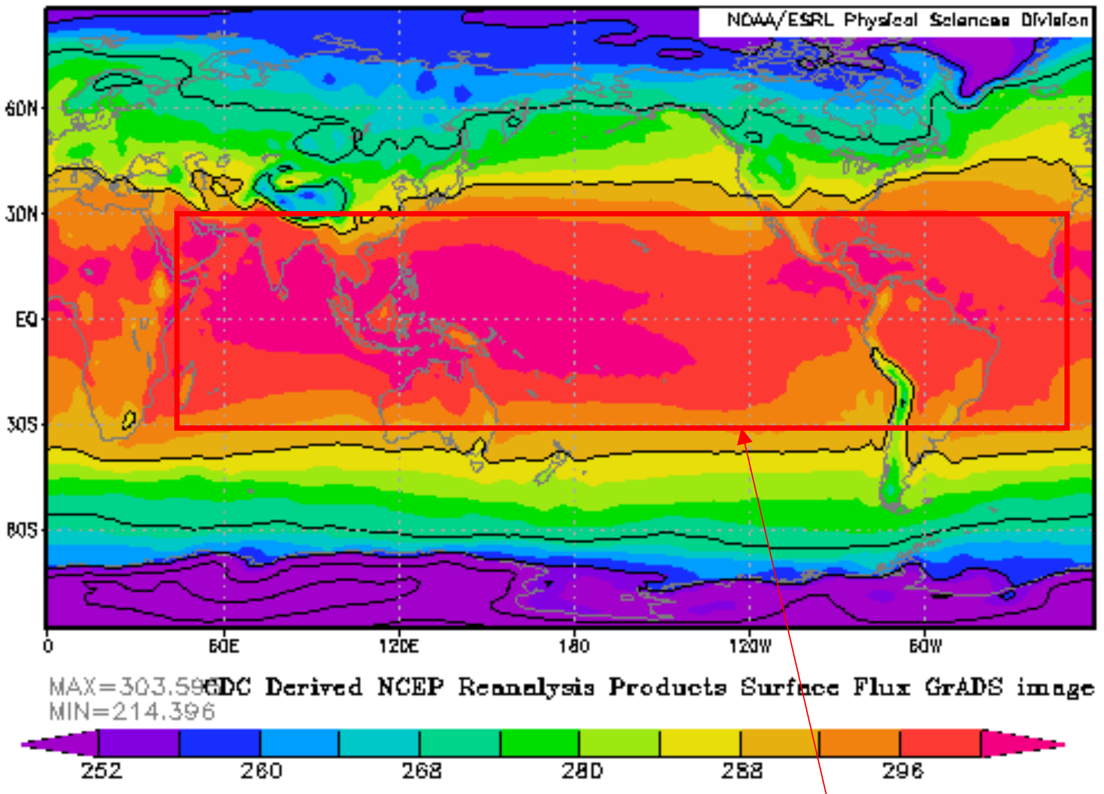
T_{2m} only slightly cooler (~1-2K) than SST

$H_{rel,2m}$ ~80% (remarkably constant over global oceans)

Source: NCEP/NCAR reanalysis
(make this yourself at https://www.esrl.noaa.gov/psd/cgi-bin/db_search/SearchMenus.pl)

Boundary layer θ_e (or moist static energy) is strongly controlled by sea surface temperatures

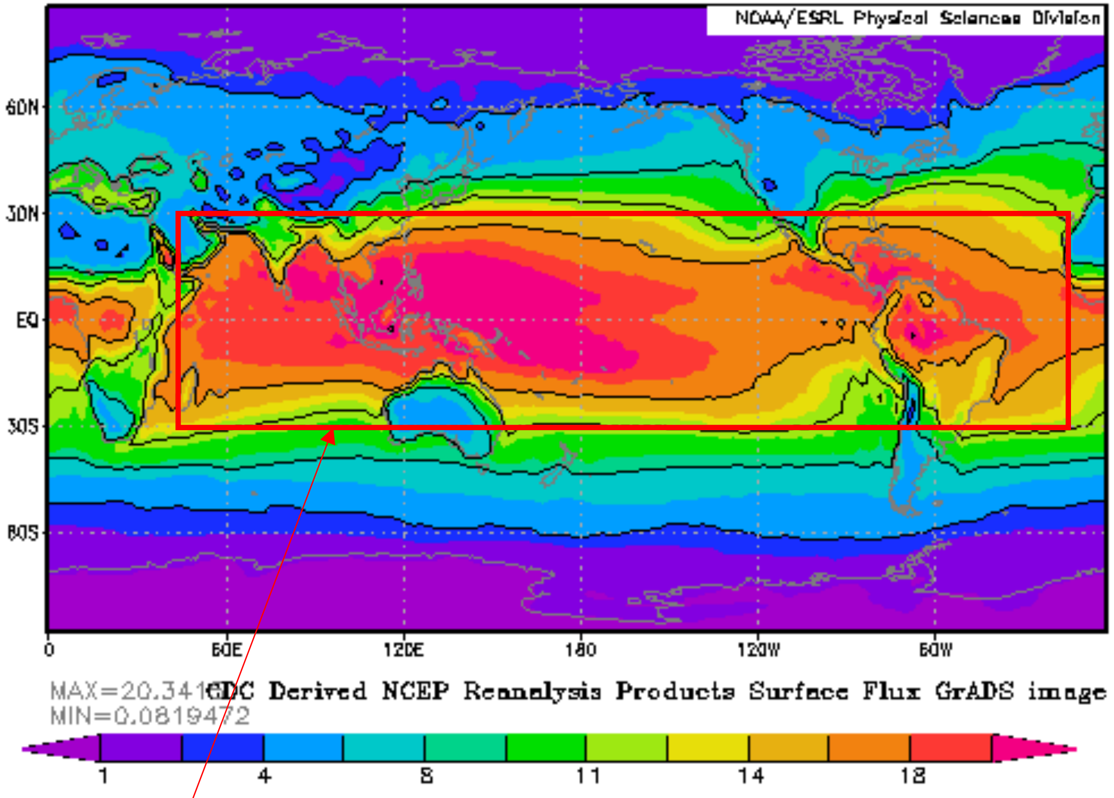
T_{2m} [K] (1981-2010 mean)



Source: ICOADS
(make this yourself at https://www.esrl.noaa.gov/psd/cgi-bin/db_search/SearchMenus.pl)

T_{2m} only slightly cooler (~1-2K) than SST

2m specific humidity q_v [g/kg] (1981-2010 mean)



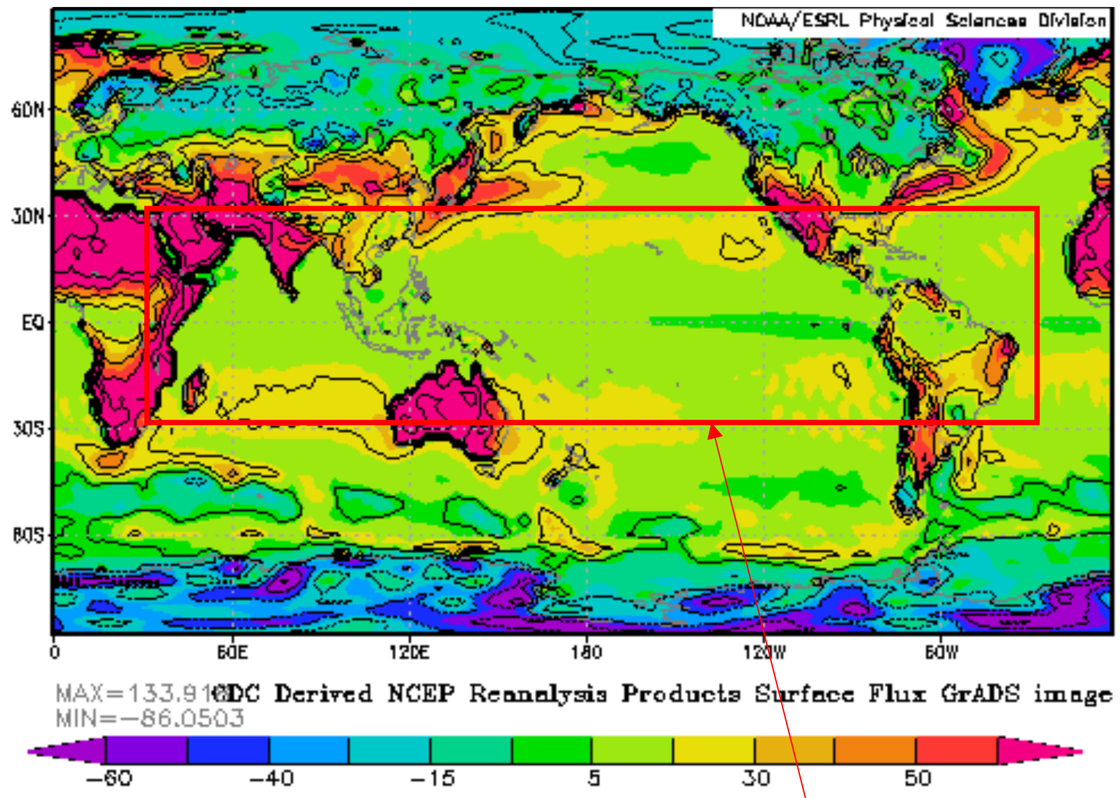
$q_{v,2m}$ follows T_{2m} since RH~80% nearly constant

Source: NCEP/NCAR reanalysis
(make this yourself at https://www.esrl.noaa.gov/psd/cgi-bin/db_search/SearchMenus.pl)

Temp + moisture (+pressure/altitude) define key moist thermodynamic variables: $\theta_e \approx \theta^{\frac{L_v r_v}{C_p T}}$ $[K]$ $h = gz + C_pT + L_v r_v$ $\left[\frac{J}{kg}\right]$

Note: weak temperature gradient approximation does <u>not</u> apply in the <u>boundary layer</u>

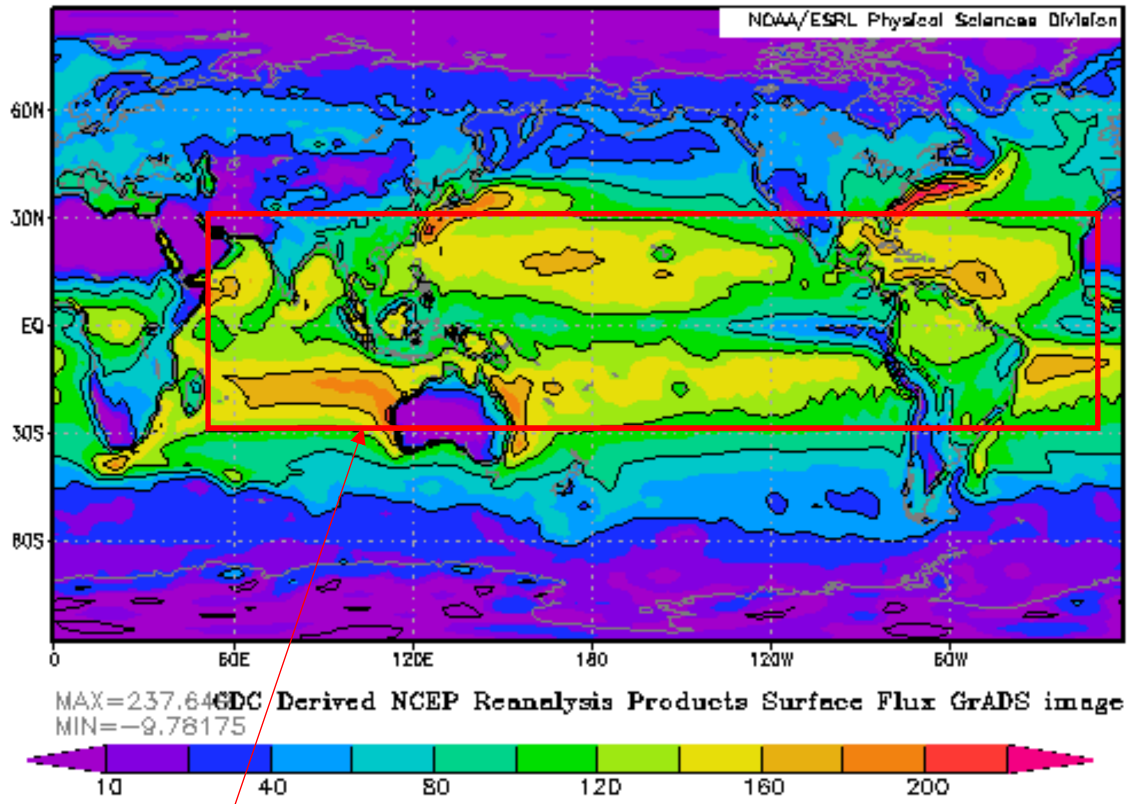
Upward <u>sensible</u> heat flux from surface [W/m²] (1981-2010 mean)



Source: NCEP/NCAR reanalysis
(make this yourself at https://www.esrl.noaa.gov/psd/cgi-bin/db_search/SearchMenus.pl)

$F_{SH,up}$ ~0-30 W/m²

Upward <u>latent</u> heat flux from surface [W/m²] (1981-2010 mean)

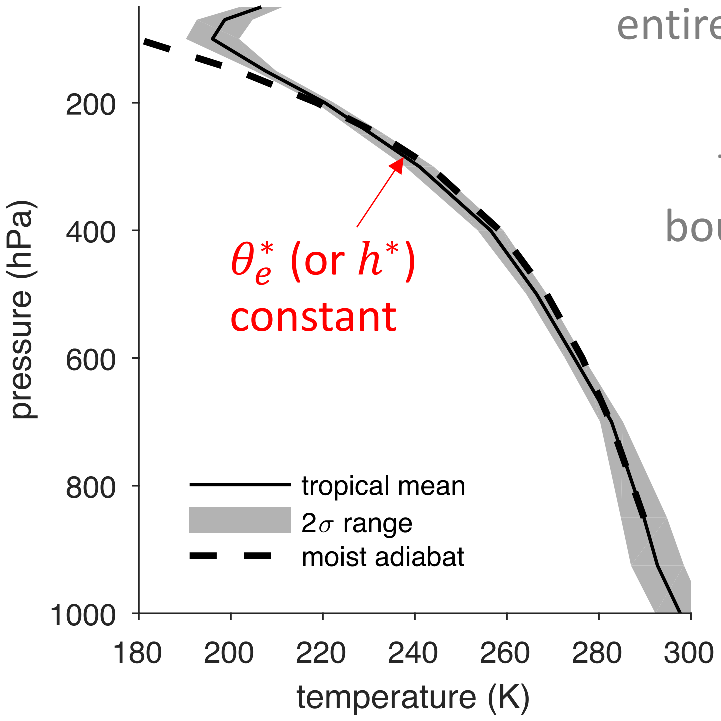


Source: NCEP/NCAR reanalysis
(make this yourself at www.esrl.noaa.gov/psd/data/composites/day)

$F_{LH,up}$ ~20-200 W/m²

Large heat fluxes from surface to boundary layer maintain air temperatures close to SST (and thus their gradients too)
(also surface friction slows horizontal motions)

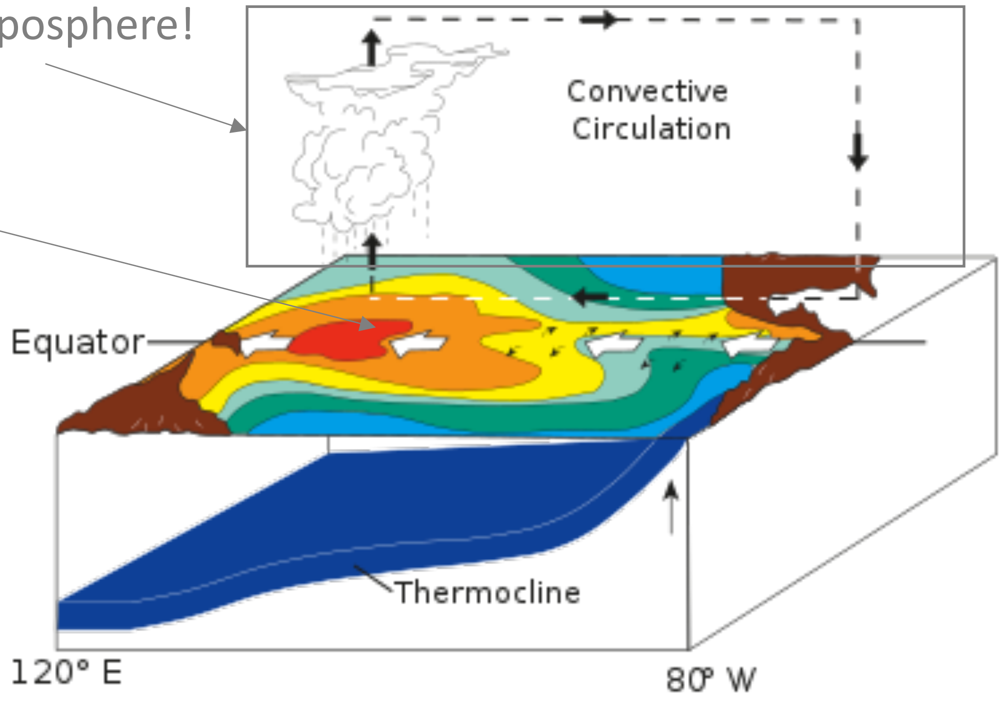
The behavior of the tropical atmosphere is strongly controlled by the SST distribution



θ_e^* (or h^*) constant

...sets the temperature profile of the entire tropical free troposphere!

The θ_e (or h) of boundary layer air here...



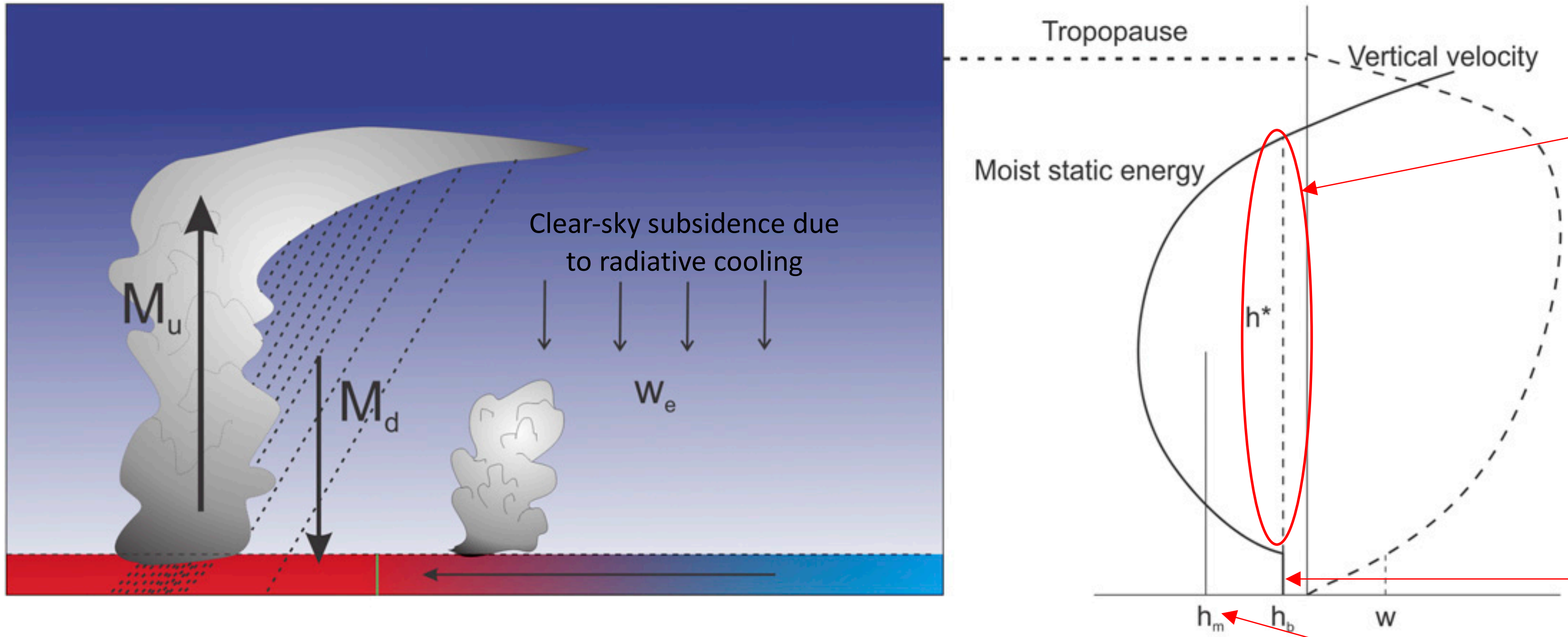
Physical model for the Walker circulation

(or any system with variable SSTs and weak rotation)

Emanuel (2019, JAS)
“Inferences from Simple Models of Slow, Convectively Coupled Processes"

A framework for conceptual understanding of slow, convectively coupled disturbances is developed and applied to several canonical tropical problems, including the water vapor content of an atmosphere in radiative–convective equilibrium, the relationship between convective precipitation and column water vapor, Walker-like circulations, self-aggregation of convection, and the Madden–Julian oscillation. The framework is a synthesis of previous work that developed four key approximations: boundary layer energy quasi equilibrium, conservation of free-tropospheric moist and dry static energies, and the weak temperature gradient approximation. It is demonstrated that essential features of slow, convectively coupled processes can be understood without reference to complex turbulent and microphysical processes, even though accounting for such complexity is essential to quantitatively accurate modeling. In particular, we demonstrate that the robust relationship between column water vapor and precipitation observed over tropical oceans does not necessarily imply direct sensitivity of convection to free-tropospheric moisture. We also show that to destabilize the radiative–convective equilibrium state, feedbacks between radiation and clouds and water vapor must be sufficiently strong relative to the gross moist stability.

Thinking physically about the Walker circulation



(a) (b)

FIG. 1. Illustrating the general conceptual framework for slow, convectively coupled processes. (a) A generic cross section through the tropical atmosphere, showing deep and shallow convection. (b) Characteristic vertical profiles of moist static energy, saturation moist static energy h^* and large-scale vertical velocity are shown. The colors in the subcloud layer represent the magnitude of moist static energy, and the green vertical line separates the deep convectively coupled region at left from the region free of deep convection at right. Deep convective updraft mass fluxes are represented by M_u, downdrafts associated with deep convection by M_d, and the vertical velocity in the clear air by w_e. See text for detailed description.

Constant h^* = moist-adiabatic lapse rate

$$h^* = gz + C_pT + L_vr_v^*$$

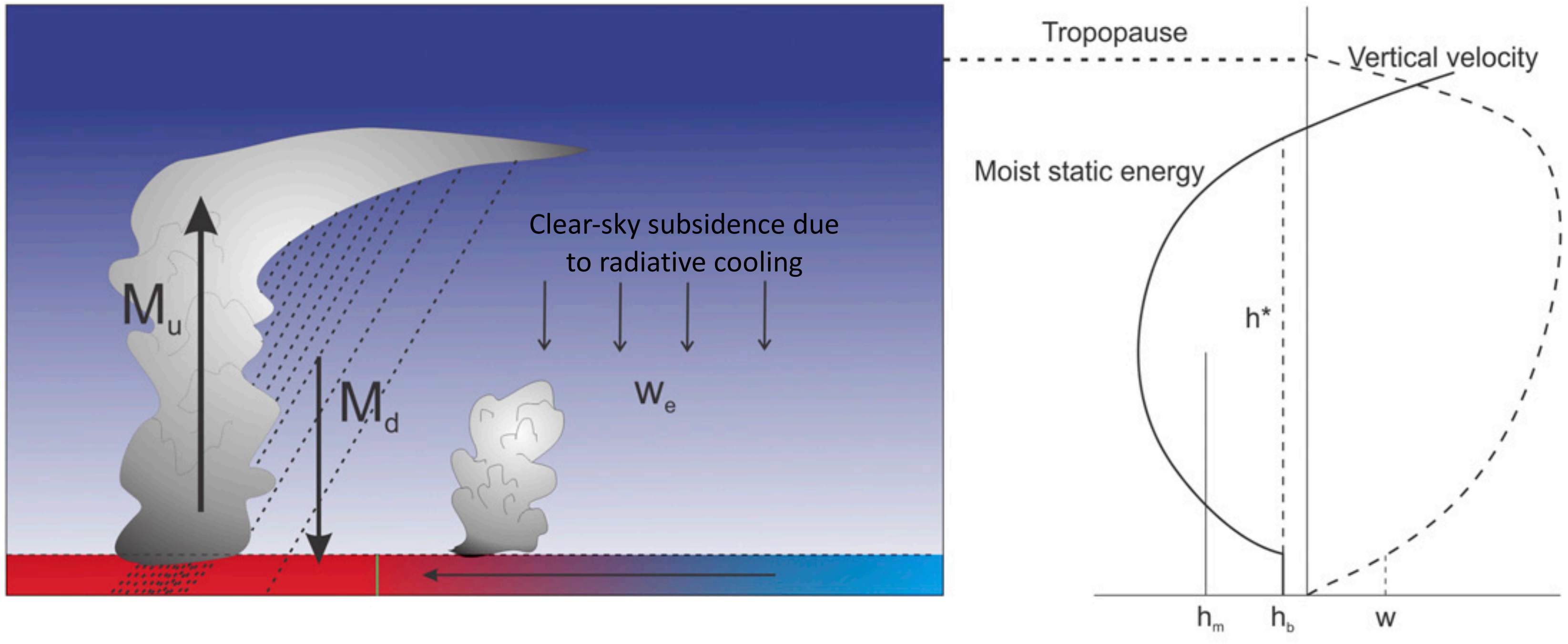
depends on T

$h^* = h_{BL}$

h_m: free tropospheric value of moist static energy.

$h_m < h_{BL}$ because it is drier.

Thinking physically about the Walker circulation

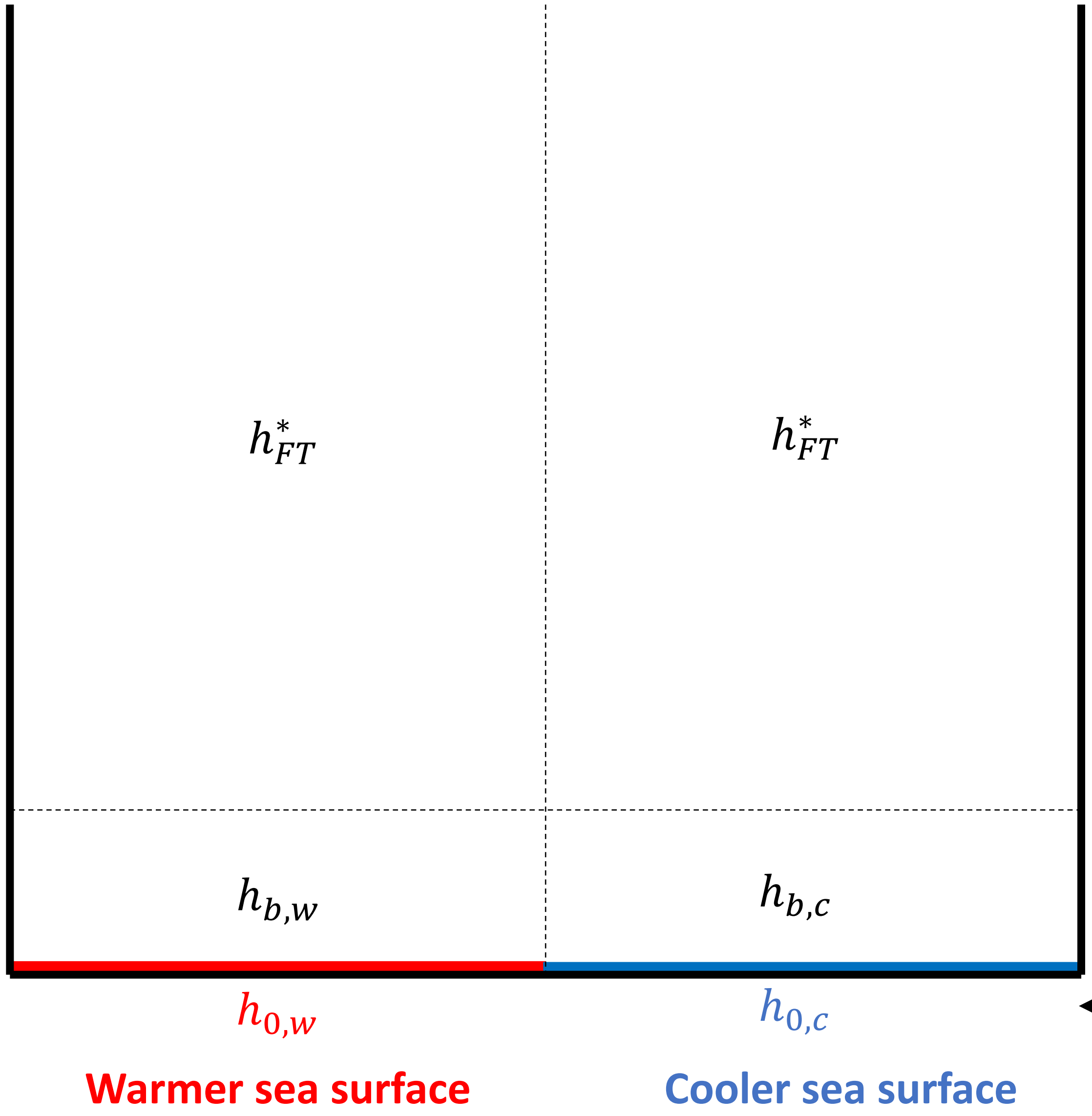


(a) (b)

Thermodynamically, this is no different from the Hadley overturning cell – **air rises where it is warmer/moister** and **sinks where it is cooler/drier**. (This is general for any weakly-rotating system, as you found in the tank lab!)
Dynamically, since the Walker circulation is **zonal**, we don't have to worry about changes in angular momentum.

Hence, the Walker cell is largely driven *thermodynamically* (i.e. by zonal variations in moist static energy)

A two-layer, two-column model for the Walker circulation

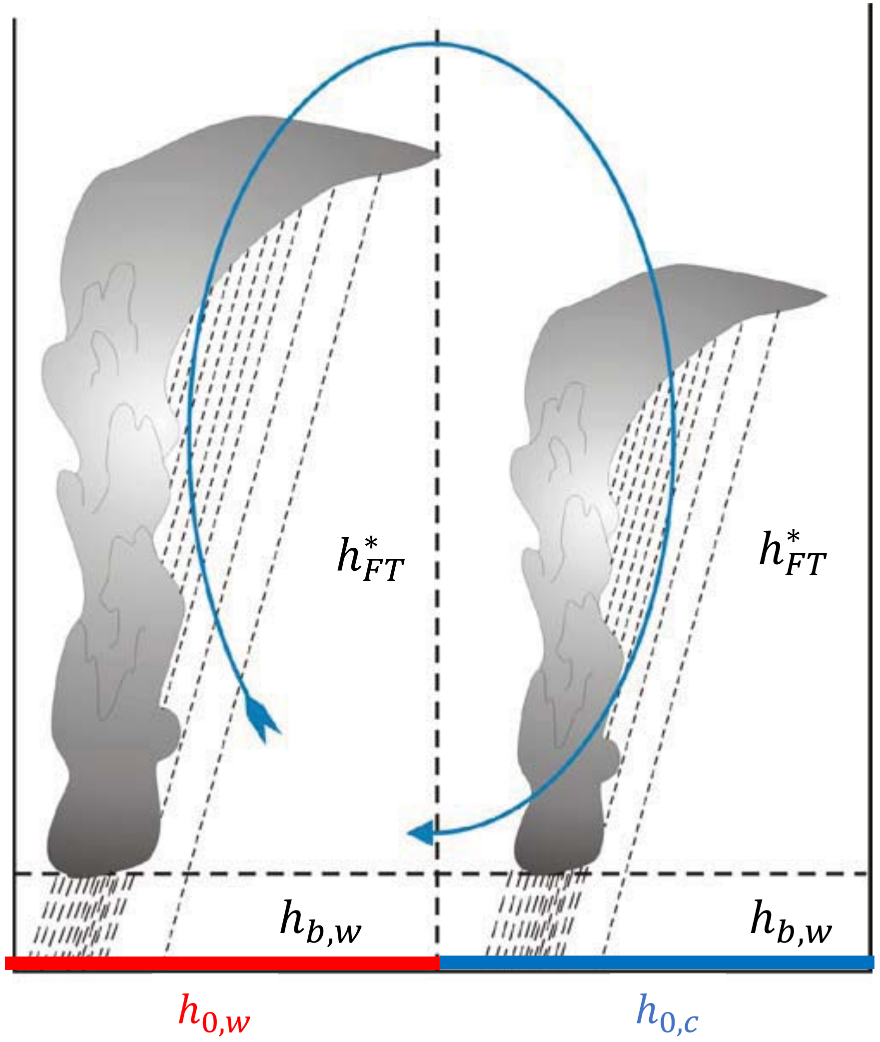


$\leftarrow$ $h_0 = gz + C_pT_{sst} + L_vr^*_v(T_{sst})$

Sea surface acts like infinite v[illegible] of “air” saturated at sea surface temperature

Source: MIT OCW 12.811 2011 (Emanuel)

Weak circulation: deep convection in both columns



Source: MIT OCW 12.811 2011 (Emanuel)

1) $h^*_{FT} = h_{b,w}$
Free tropospheric lapse rate is set by the warmest h_b (which occurs over the warmest SST).
This will then be true in both FT boxes – (weak temperature gradient approx).

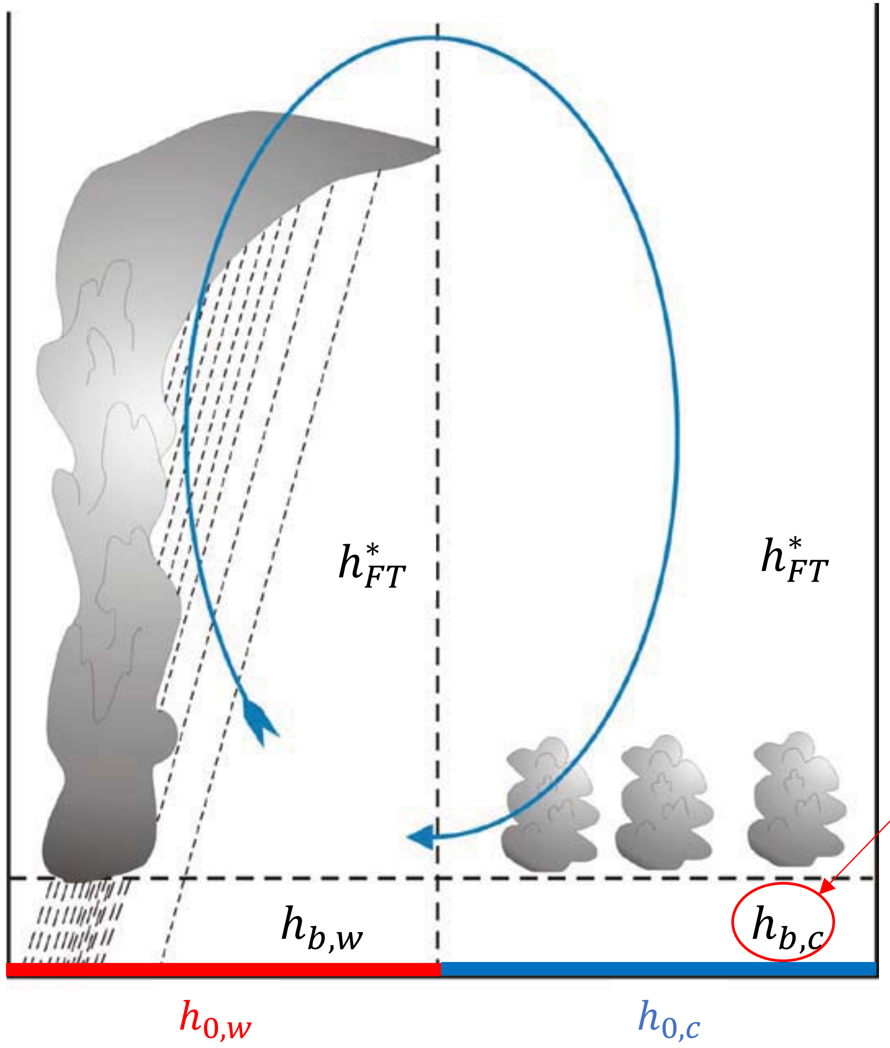
2) $h_{b,c} = h^*_{FT}$
The cold column is also deeply convecting, so $h_{b,c}$ must match h^*_{FT}

(1) + (2): $h_{b,c} = h_{b,w}$
Moist static energy in boundary layer must *also* the same in both columns!

Deep convection is weaker in the **cold column**, but not shut off.
This is because the atmosphere in the cold box is a *bit* more stable relative to the colder ocean surface.

Strong circulation: deep convection only in warm column; circulation suppresses it in cold column

Simplest understanding: this will occur if the SST gradient is sufficiently strong

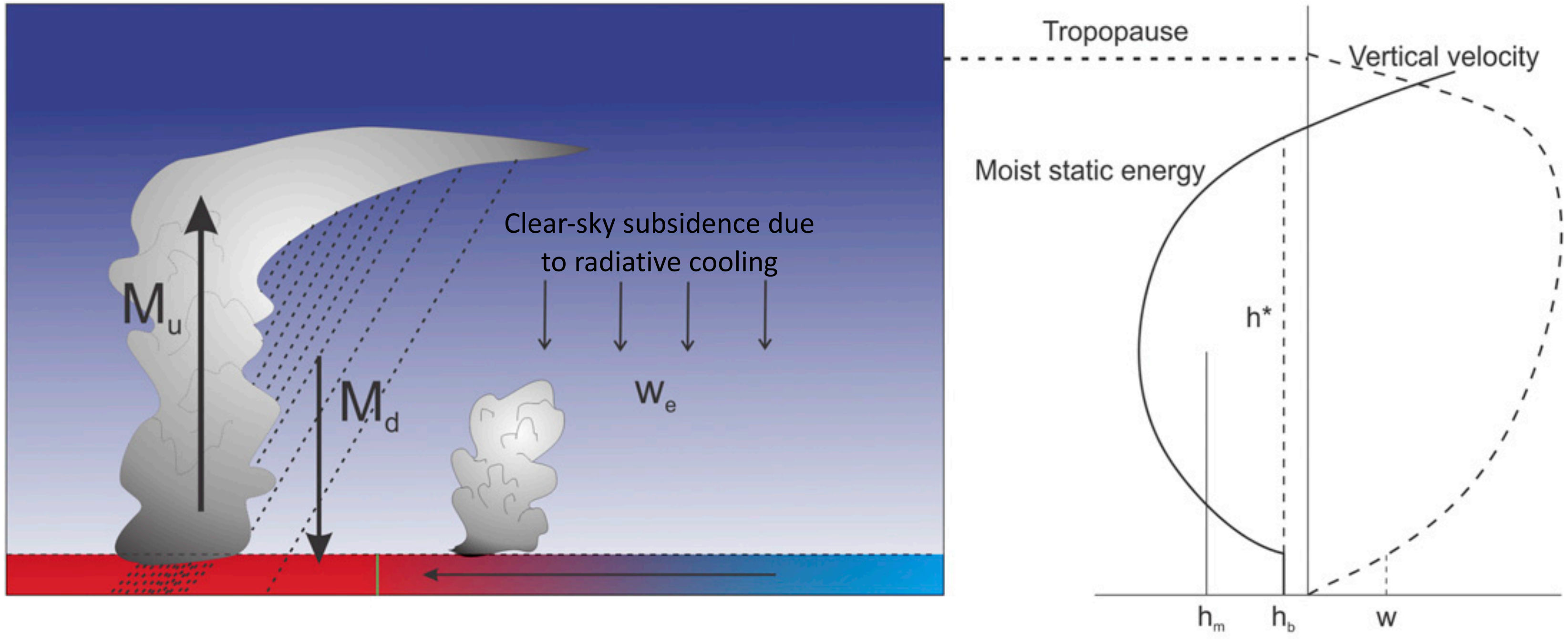


$h_{b,c}$ now varies independently. When deep convection is shut off, the BL can no longer communicate with the FT. So it doesn't "feel" h^*_{FT}.

Source: MIT OCW 12.811 2011 (Emanuel)

This is like the **real Walker cell** over the modern-day Pacific ocean:
Deep convection over the west Pacific warm pool, only shallow convection over the rest of the Pacific.

Can we model the transition between the two regimes?



(a) (b)

Budget of BL moist static energy:

$$d\left(\frac{\partial h_b}{\partial t}+\mathbf{V}_h\cdot\nabla h_b\right)=F_h-(M_d+w_e)(h_b-h_m)-\dot{Q}_b d, \quad (2)$$

where d is the depth of the boundary layer, h_b is the moist static energy of the boundary layer (which is assumed to be well mixed in the vertical), $\mathbf{V}_h$ is the large-scale horizontal velocity in the boundary layer, F_h is the surface enthalpy flux, h_m is a characteristic value of moist static energy in the free troposphere (see Fig. 1b), and $\dot{Q}_b$ is the radiative cooling of the boundary layer. In

"Boundary layer quasi-equilibrium": $\boxed{0=F_h-(M_d+w_e)(h_b-h_m)}$

The **moist static energy of the boundary layer** is assumed **steady**. Its value is set by a **balance between**:

Source: **surface heat fluxes F_h** (latent+sensible)

Sink: downward advection of **low-h air from the free troposphere** by convective downdrafts and radiative-subsidence.

For more, you will need to go to the paper.

It is too much for this class unfortunately.

Emanuel (2019, JAS)

"Inferences from Simple Models of Slow, Convectively Coupled Processes"

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Now go to Blackboard to answer a few questions about this topic!