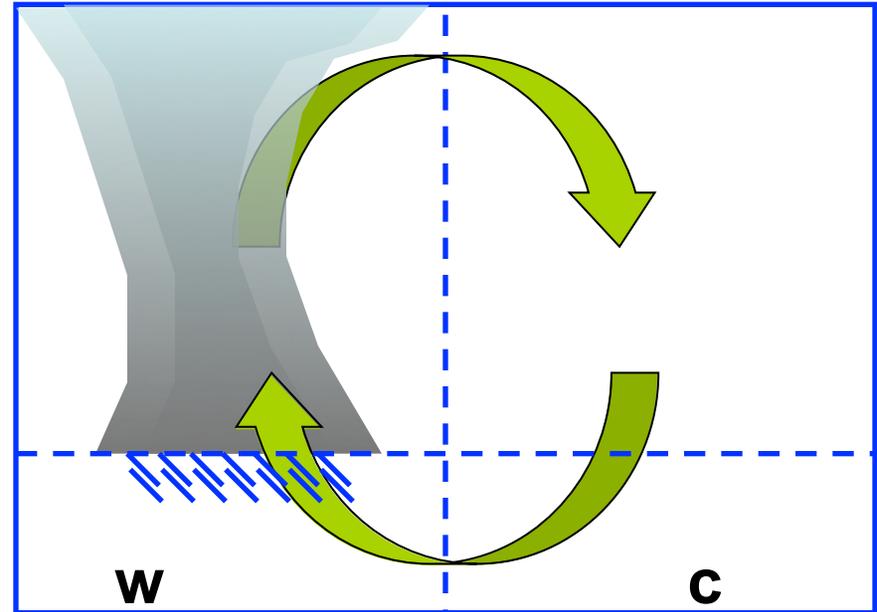


The Walker Circulation model

We will consider a two-layer column model for adjacent columns, one more strongly convecting than the other. We'll label the more strongly convecting box "w"--a warmer surface implying more convection--and the other "c". The constraint we'll invoke is that mass continuity must apply over both columns, i.e., what goes up in one box has to come down in the other.

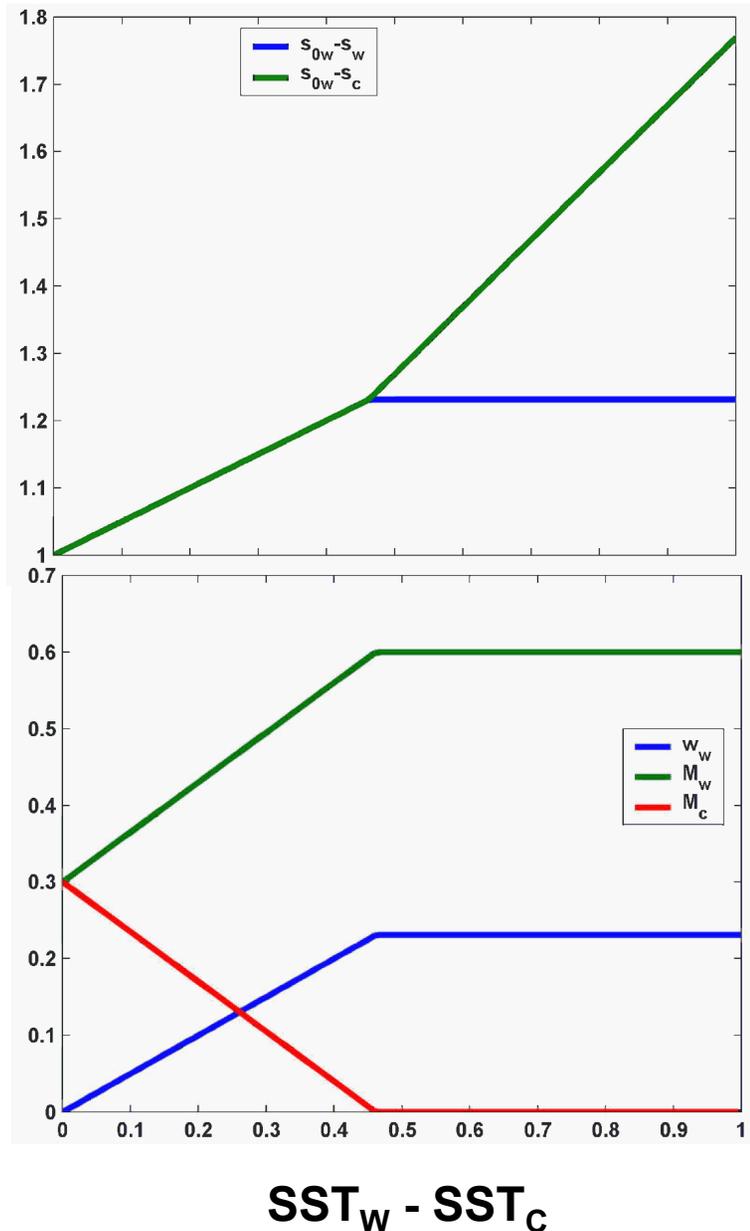


The following solutions (from Emanuel 2005) assume the surface flux can be written as:

$$F_s \propto (\theta_{e0}^* - \theta_{eb})$$

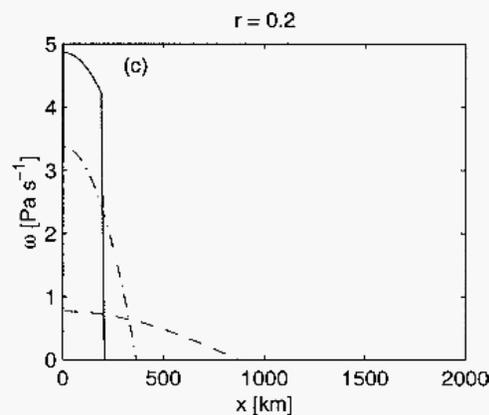
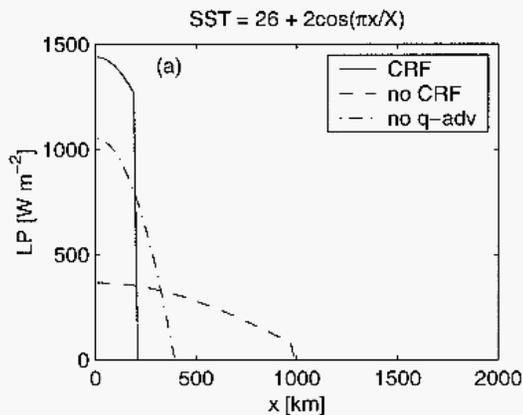
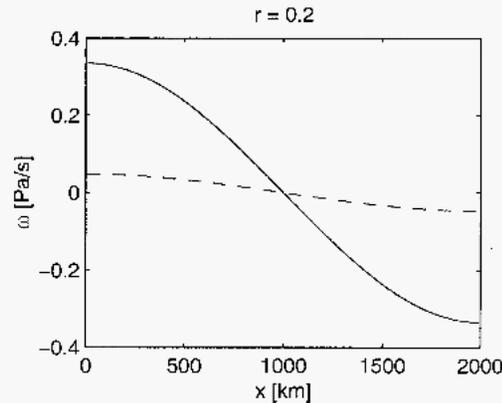
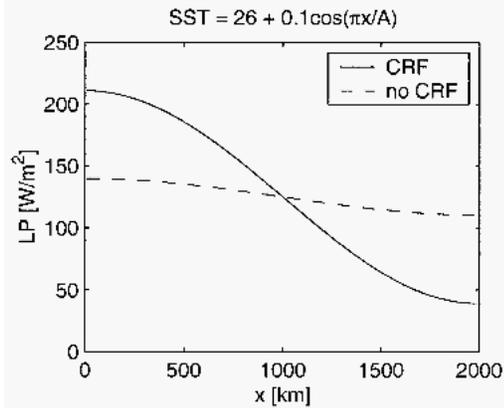
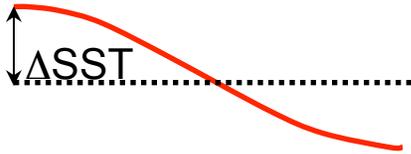
θ_{e0} is proportional to sea surface temperature.

2-Box solution



- Beyond a certain threshold SST difference, the convective updraft mass flux ceases in the cooler box
- When $M_U=0$ in the cooler box, the vertical velocity in the FT (<0) balances radiative cooling

A continuous 2D ($\sim N$ column) model



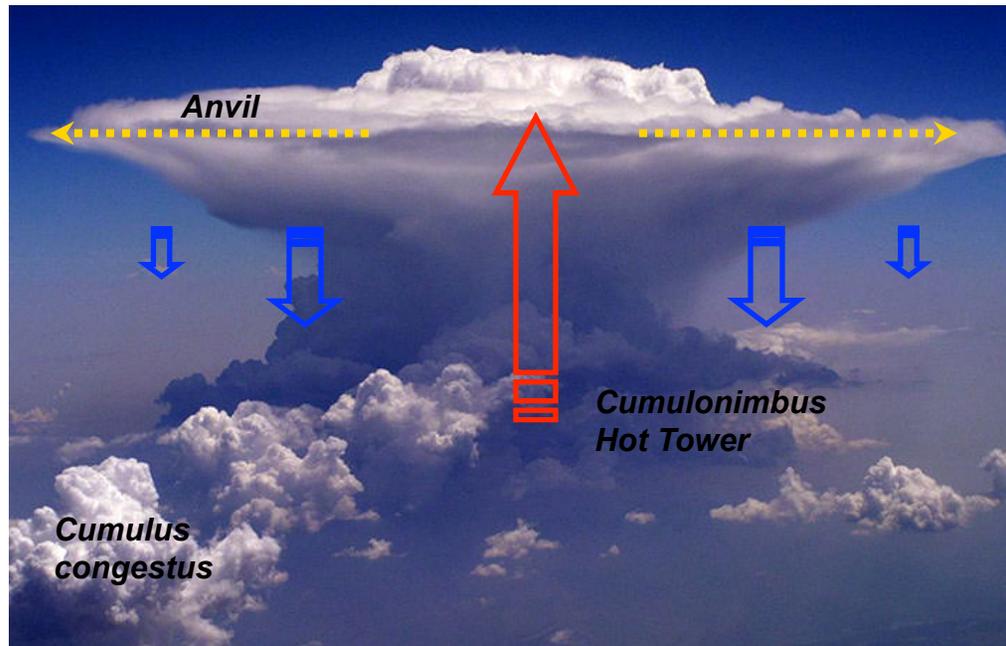
- In this model, SST is varied sinusoidally across a longitudinal domain.
- For a weak SST spatial amplitude, precipitation and vertical velocity are continuous (\sim sinusoidal)
- However, for strong SST spatial amplitude, precipitation restricted to warmest portion of the domain.
- Effects such as interactive clouds and horizontal moisture transport significantly modify the solution.

Figures from Bretherton and Sobel, 2002

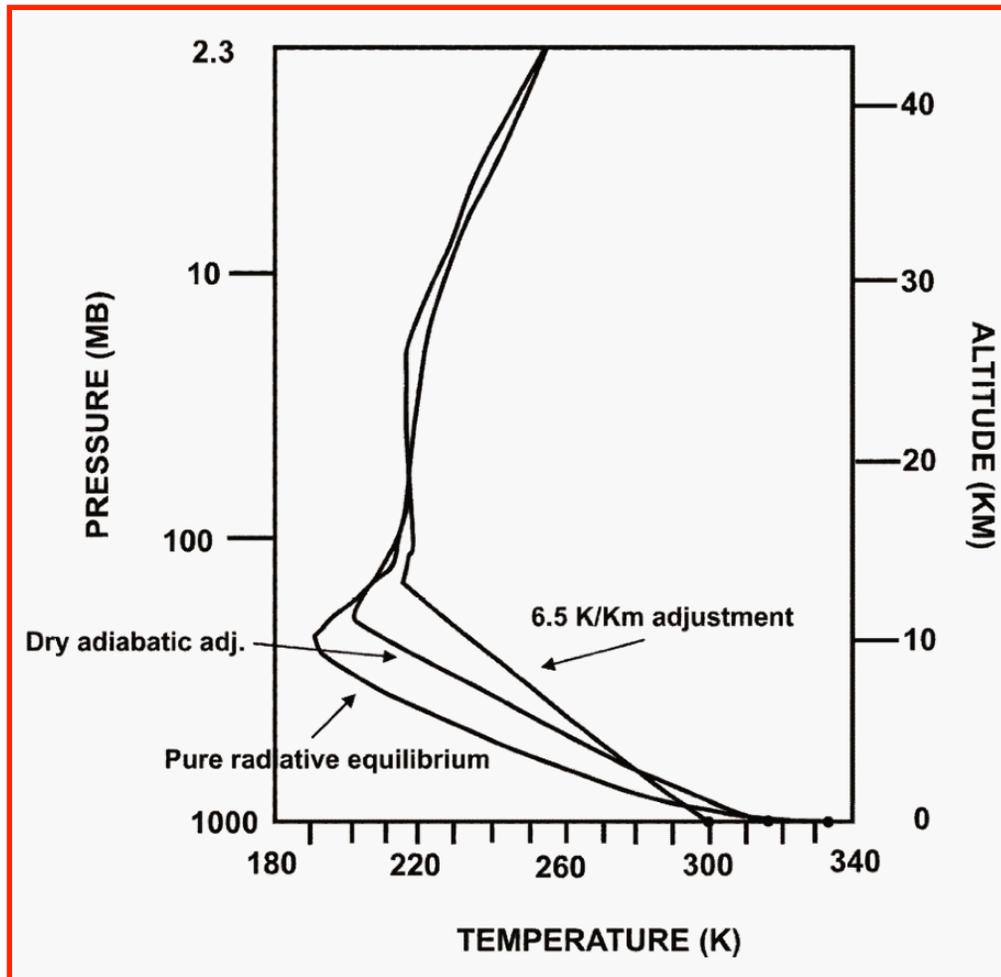
- The big picture perspective of tropical convection
- Basic aspects of tropical cumulus convection

*Goal: Introduce concepts necessary
to understand tropical convection*

Some big picture considerations



Revisiting radiative equilibrium



- Pure radiative equilibrium is in fact **unstable** for conditions in the **troposphere** [but is stable and comparable to observations in the stratosphere].
- Tropospheric γ_d significantly exceeds the observed lapse rate ($\sim 6.5\text{K/km}$)

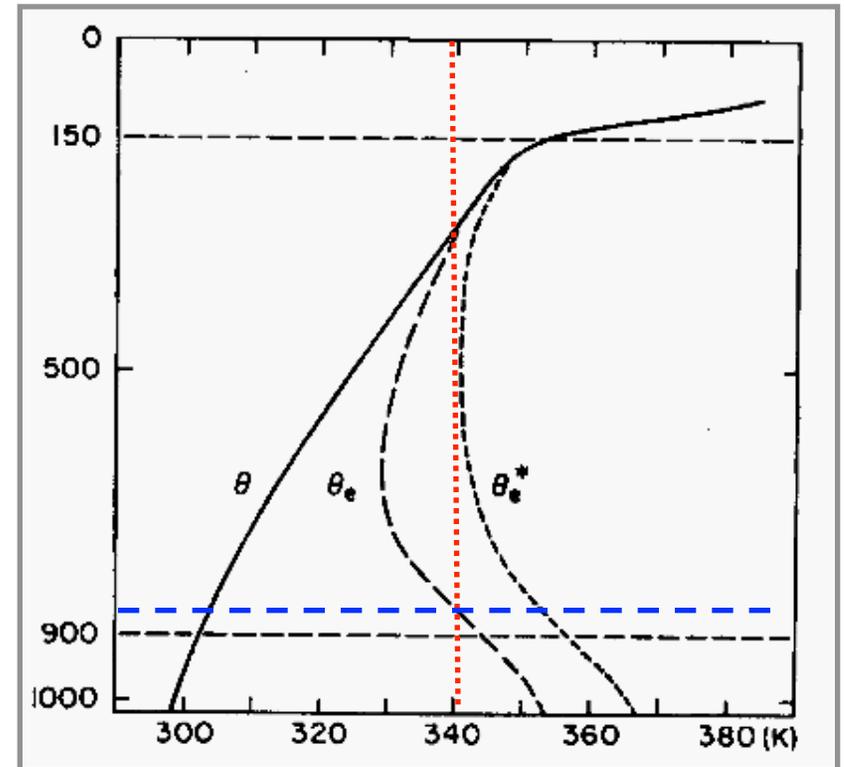
To account for the smaller observed lapse rate, is necessary to account for vertical heat transport via moist convection.

Emanuel, 2005

Convection and large-scale overturning

As we saw in the traditional view of the Hadley circulation, the thermally-direct Hadley cell comprises equatorward low level flow toward the ITCZ with upward motion in the rising branch of the circulation. This circulation effects a net transport of heat poleward away from the equator.

However, is this view consistent with observed thermodynamic profiles?



Consider the thermodynamic equation close to the ITCZ, where $v \approx 0$:

$$\frac{\partial \theta_e}{\partial t} \approx -w \frac{\partial \theta_e}{\partial z}$$

For a large-scale vertical flow ($w > 0$) in the upper troposphere ($\frac{\partial \theta_e}{\partial z} > 0$), this implies a negative—or cooling—tendency in θ_e . This is inconsistent with a thermally-direct circulation transporting heat away from the equator.

Cumulus convection

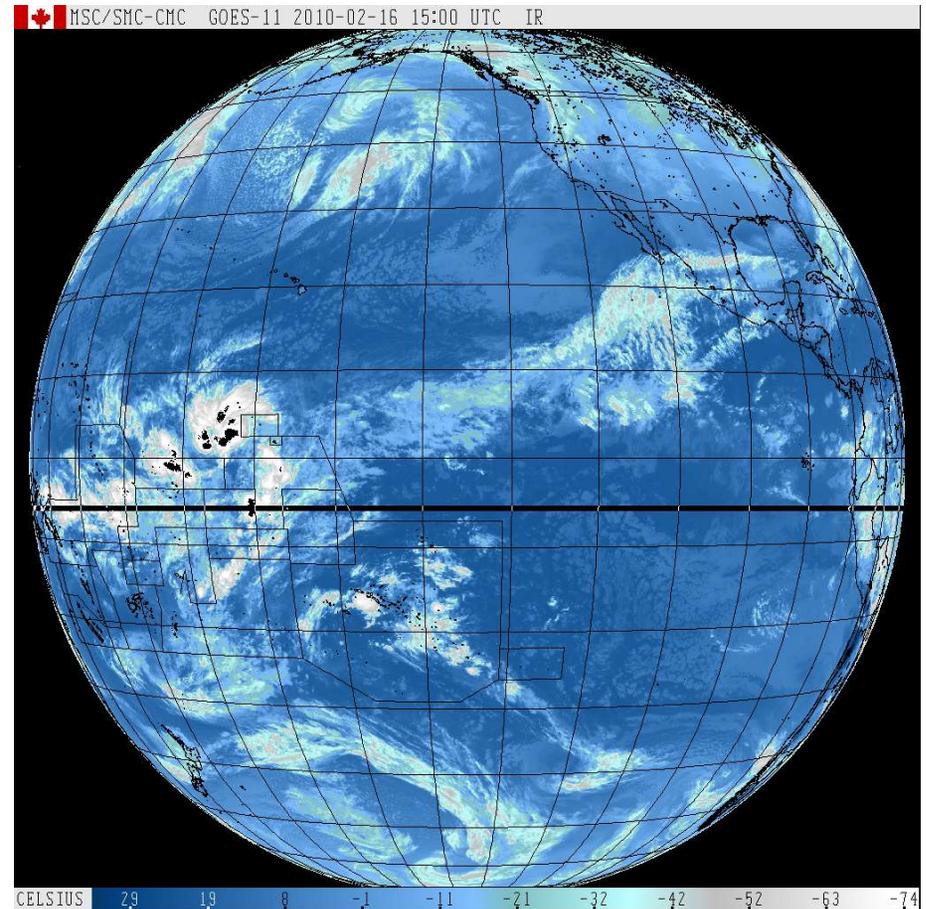
A resolution of the inconsistency between the poleward heat transport and adiabatic cooling in the upper branch of the Hadley circulation is the (pseudoadiabatic) delivery of relatively warm parcels from the surface to upper troposphere through the cores of large cumulonimbus clouds.

- Thus, *cumulus convection*, from the Latin terms meaning “heap” and “to carry together,” represents the process by which an air parcel less dense than its surroundings rises and forms a convective cell/cloud.
- Newton’s 2nd Law: Positive buoyancy of the parcel, i.e., the parcel is less dense than the environment, generates upward acceleration known as a convective updraft



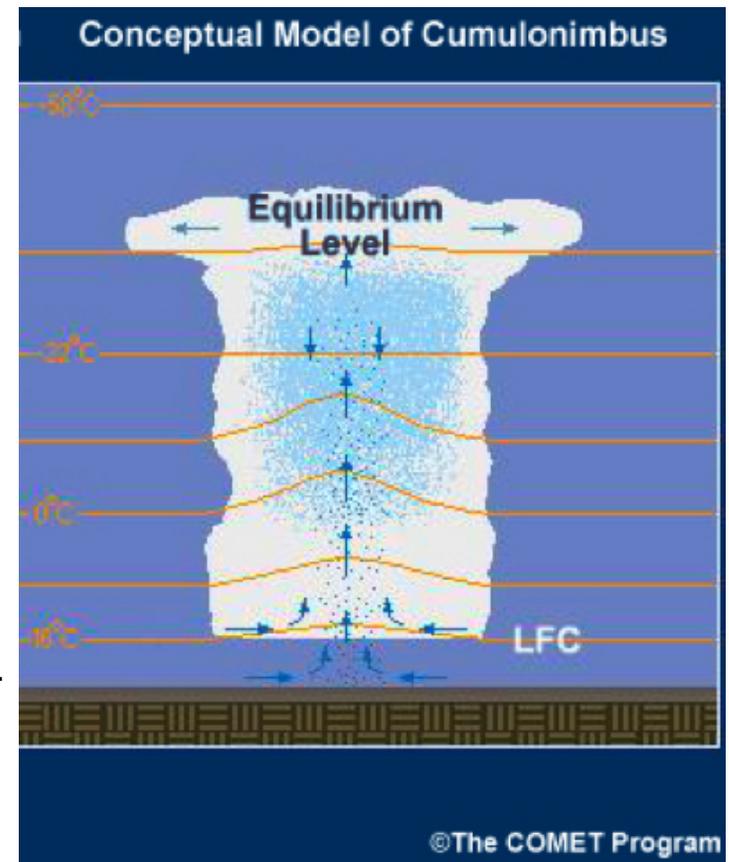
Snapshot of tropical convection

- “Snapshot” of Pacific cloudiness: relatively few “hot towers” over the oceans [Note: much of the cloudiness here represents spreading high clouds from areas of intense convection.]
- Riehl and Malkus (1958) estimated only 1500-5000 hot towers are needed at any one time to accomplish vertical heat transport
- Bjerknes (1938) demonstrated that an arrangement with vanishingly small convective fraction promotes narrow fast updrafts and broad slow downdrafts.



A cumulus “hot tower”

- The schematic on the right illustrates a “mature” cumulonimbus
- Basic elements are as follows:
 - The **lifting condensation level (LCL)**, (not shown) is the level at which an initially unsaturated parcel reaches saturation, i.e., it’s relative humidity reaches 100%.
 - The **level of free convection (LFC)** is the level at which the parcel first becomes warmer, and thus less buoyant, than its environment.
 - The **equilibrium level (EL)** is the level at which the parcel becomes as cool (dense) as the environment. Parcels reaching the EL do not stop abruptly but will oscillate [since N^2 changes sign at this level]. Lateral motions near the cloud top—associated with outflow of the updraft air mass—may form an *anvil* cloud.



Convective available potential energy

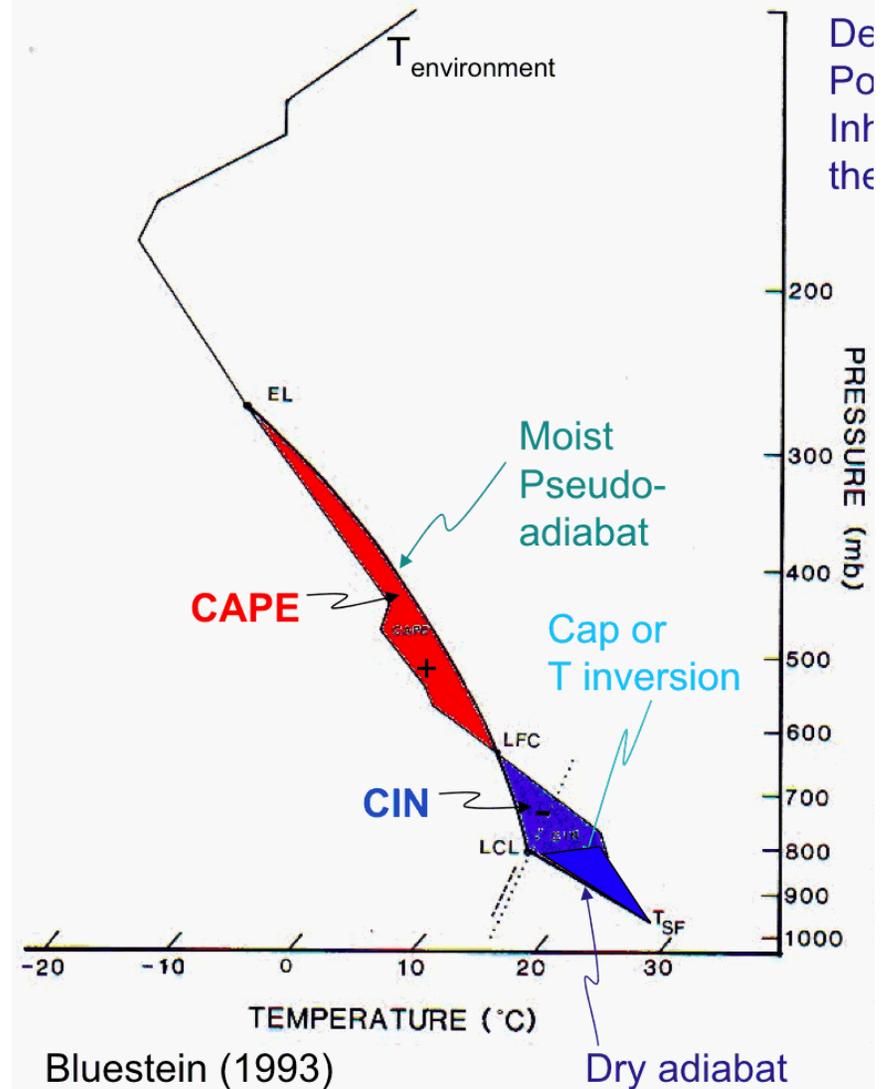
- **Convective available potential energy (CAPE)** is the vertically-integrated buoyancy between the LFC and EL:

$$CAPE = \int_{z_{LFC}}^{z_{EL}} B dz$$

$$= \int_{z_{LFC}}^{z_{EL}} g \frac{\rho_{env} - \rho}{\rho} dz = \int_{z_{LFC}}^{z_{EL}} g \frac{\theta - \theta_{env}}{\theta_{env}} dz$$

- CAPE can be estimated from a thermodynamic diagram as the positive area between the environmental temperature profile and the moist adiabat through the LCL (i.e., the red area on the right).
- Note also the **convective inhibition (CIN)** region (the blue area), effectively the negative CAPE, between the surface and LFC in this profile.

If CIN = 0, the LFC = LCL.



CAPE and convective updrafts

- CAPE and the maximum vertical velocity of a convective updraft are related as:

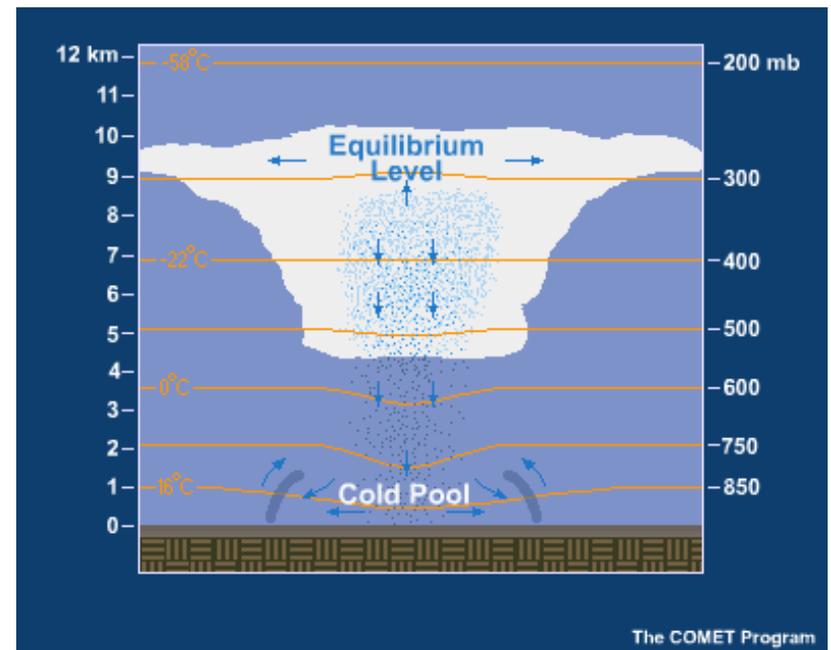
$$w_{\max} = \sqrt{2CAPE}$$

Think: CAPE is dimensionally potential energy (per mass), so this expression follows from energy conservation, setting CAPE equal to the maximum kinetic energy per mass, i.e., $1/2 w_{\max}^2$.

- The actual maximum w attained will generally be less than w_{\max} , up to a factor of 2, since other factors affecting the in-cloud vertical velocity field may come into play.
- What CAPE and updraft velocities are typical?
 - CAPE in thunderstorms is typically 1000-2000 J kg⁻¹ [or m² s⁻²] but can exceed 4000 J kg⁻¹.
 - For CAPE = 1500 J kg⁻¹, $w_{\max} = 55$ m s⁻¹!
 - Tropical oceanic convection, which tends to have small differences between environmental and convective parcel temperature differences, is often characterized by relatively low CAPE and convective updraft velocities.

Precipitation and convective downdrafts

- During ascent, parcels acquire either water or ice droplets associated with the water vapor phase change
 - After these parcels have acquired sufficient mass, they begin to fall [or precipitate]
- The relatively cold—and negatively buoyant—air “dragged” with falling precipitation comprises a *convective downdraft*
 - The downdraft velocity is larger with cooler and/or drier conditions, e.g., the inflow of relatively dry environmental air and evaporation of falling rain both increase downdraft velocities
- A *cold pool* forms beneath the precipitating cloud and spreads out laterally in the ABL.



Moist static energy

A useful very useful thermodynamic quantity is the moist static energy, h , defined as:

$$h = s + L_c q; \quad s = c_p T + \Phi \quad \text{The quantity } s \text{ is the dry static energy.}$$

We can show that h and θ_e are related. Taking the natural logarithm of θ_e , defined for an unsaturated parcel, gives:

$$\ln \theta_e \approx \ln \theta + \frac{L_c q}{c_p T_{LCL}} = \ln T - \frac{R_d}{c_p} (\ln P - \ln P_s) + \frac{L_c q}{c_p T_{LCL}}$$

Taking the differential of the above expression gives:

$$d \ln \theta_e \approx \frac{dT}{T} - \frac{R_d}{c_p} \frac{dP}{P} + d \left(\frac{L_c q}{c_p T_{LCL}} \right) \approx \frac{dT}{T} + \frac{gdz}{c_p T} + \frac{L_c dq}{c_p T_{LCL}}$$

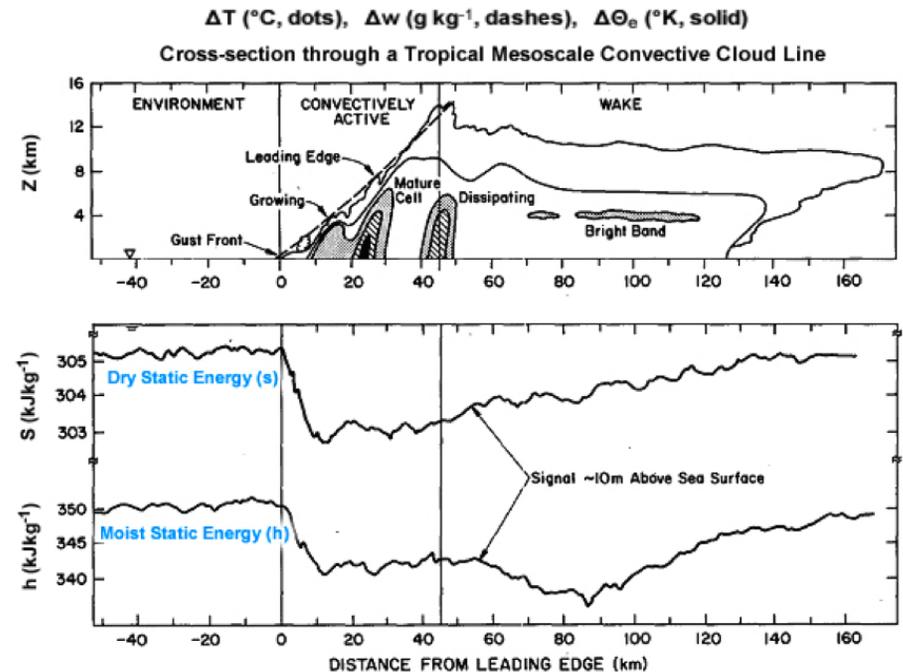
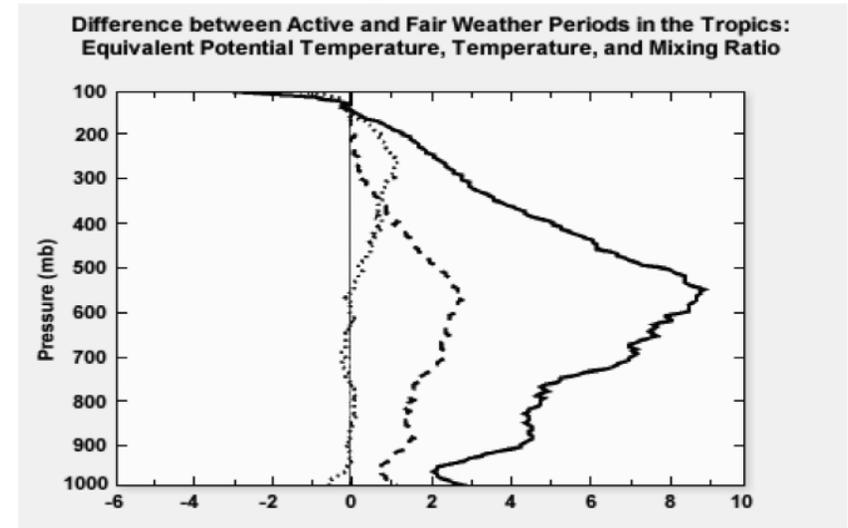
For T [in K] close to T_{LCL} , and using the definition of geopotential, yields:

$$c_p T d \ln \theta_e \approx c_p dT + d\Phi + L_c dq = dh$$

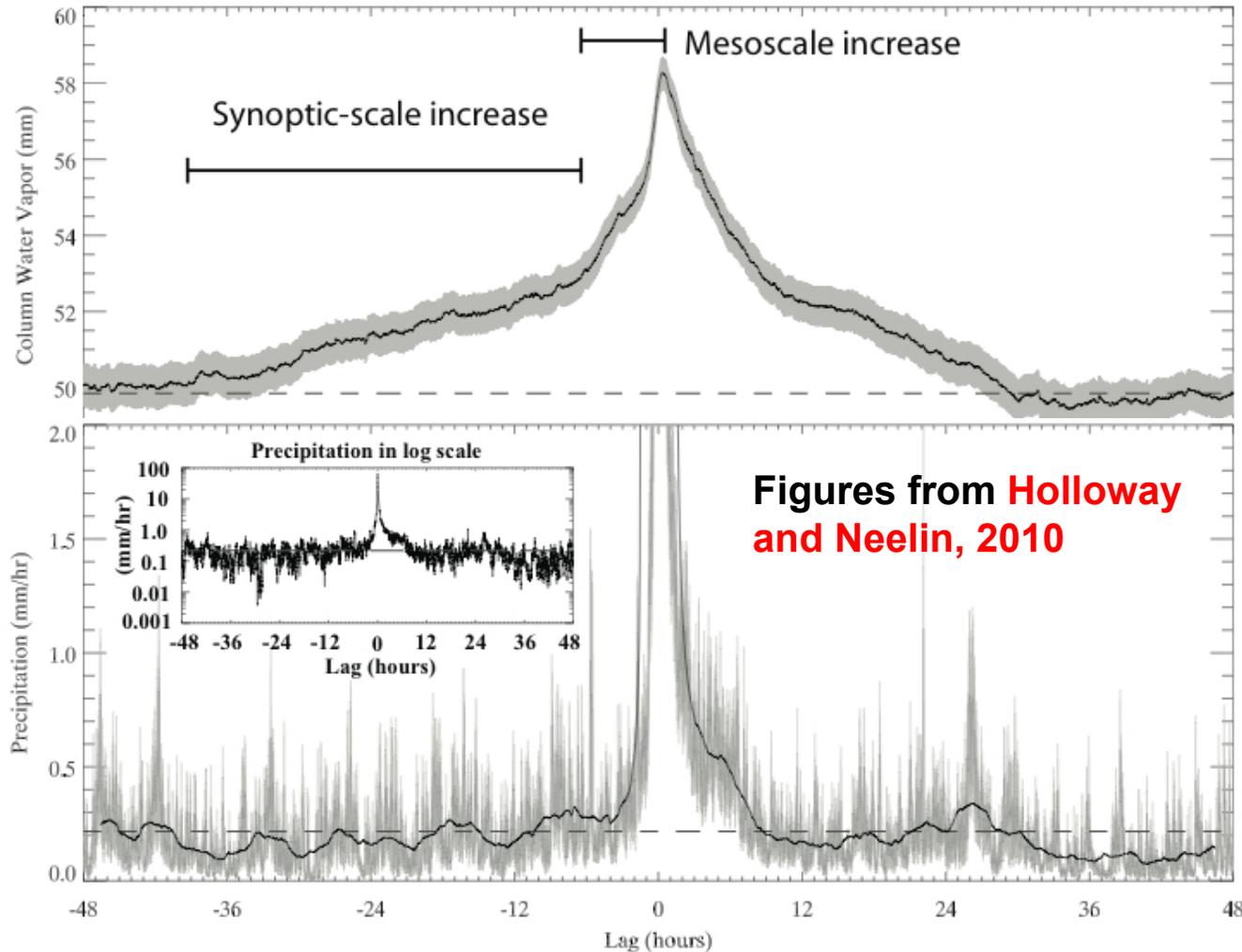
The above relationship implies that moist static energy is approximately conserved when equivalent potential temperature is conserved.

Tropical convective energetics

- The top panel illustrates the vertical profiles of temperature, water vapor mixing ratio, and equivalent potential temperature differences between actively-convecting and nonconvecting tropical conditions.
 - θ_e differences > 0 in the middle troposphere, with differences in water vapor mixing ratio accounting for the bulk of the θ_e differences.
- The bottom panel shows a snapshot of dry and moist static energy for a tropical convective system.
 - The leading edge of the system is at zero, and the system is moving to the west (left).
 - Both s and h drop strongly at the leading edge and slowly recover after dissipation.



Tropical convection and moisture



Figures from **Holloway and Neelin, 2010**

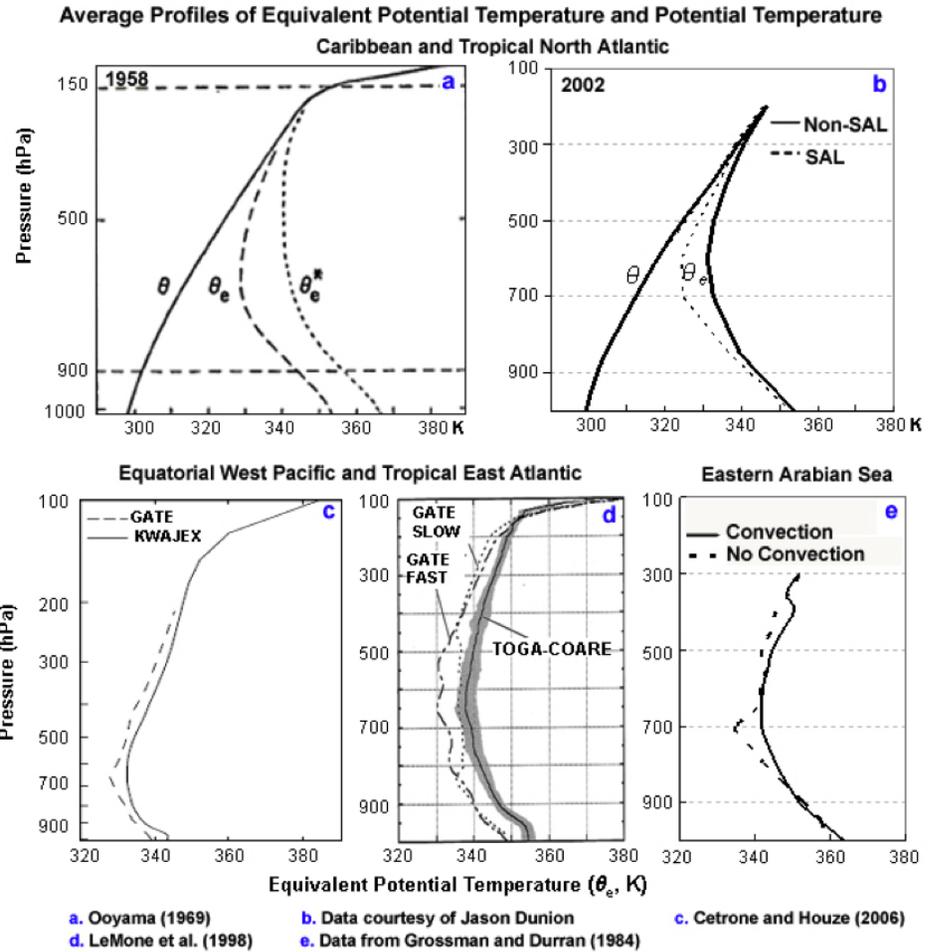
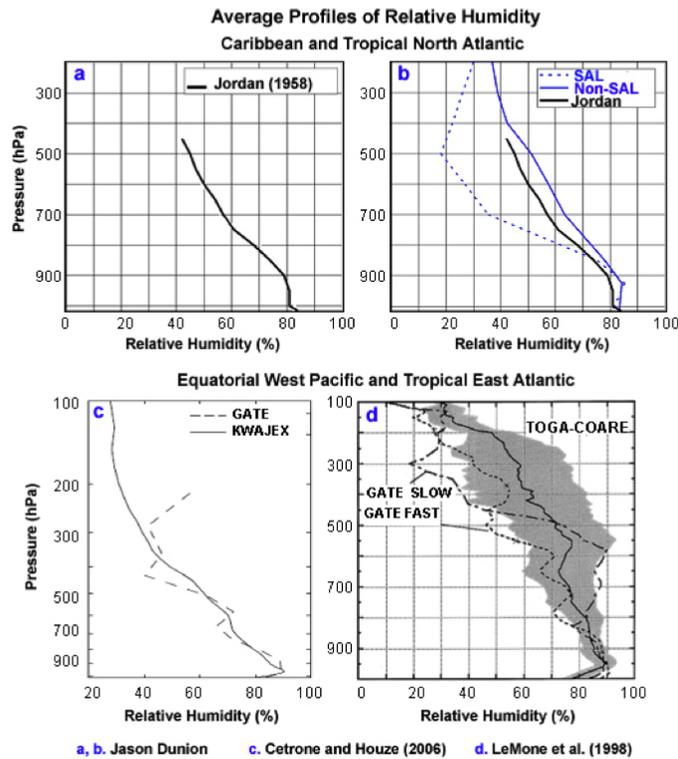
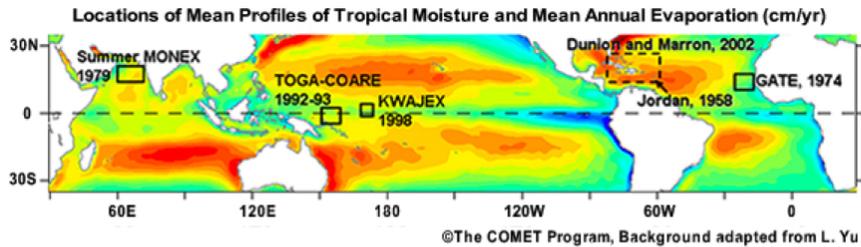
Republic of Nauru, the world's smallest island nation (21 km²).



- Time lead/lag plot for precipitating events at Nauru (0°S, 167°E) in the western Tropical Pacific. A “synoptic”-scale increase in tropospheric moisture is evident roughly 24 hours before the onset of strong convection.

Note the steeper increase ~6 hours before, perhaps indicating organization at smaller (mesoscales) scales.

Variability in conditions conducive to convection



- The tropics is not uniform in terms of conditions favoring deep convection. For the vertical profiles illustrated here, there's a pronounced difference between North Atlantic region air masses originating in the so-called Sahara Air Layer (SAL), which is typically quite dry.