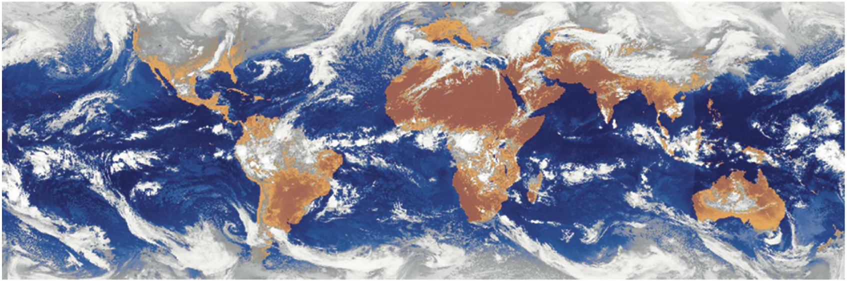
Atmospheric Convection

Phil Austin

July 30, 2007



source: Bony et al., 2006

Outline

1. Satellite/reanalysis views of tropical clouds (MODIS, ISCCP, Bony et al.)
2. Basics: Moist thermodynamics, buoyancy, CAPE, mixing diagrams, conditional/slice instability
3. Impact of clouds on large scale fields (Q_1, Q_2, mass flux models)
4. Equilibrium coupling of shallow and deep convection: one cell model
5. Entrainment, detrainment, buoyancy sorting
6. What controls convective cloud top height?

References

General material:

- Atmospheric Convection, Kerry A. Emanuel, 1994: Canada, US, used
- ECMWF training – Convection II, Bechtold, Jacob, Gregory, Khain
- Dave Randall's General Circulation text (Chapter 6)

Articles:

- Stevens, B., 2005: Atmospheric Moist Convection
- References tbd

and of interest:

- Kerry Emanuel's tropical meteorology course
- Roland Madden MJO lecture

Walker circulation Webster and Chang, 1988

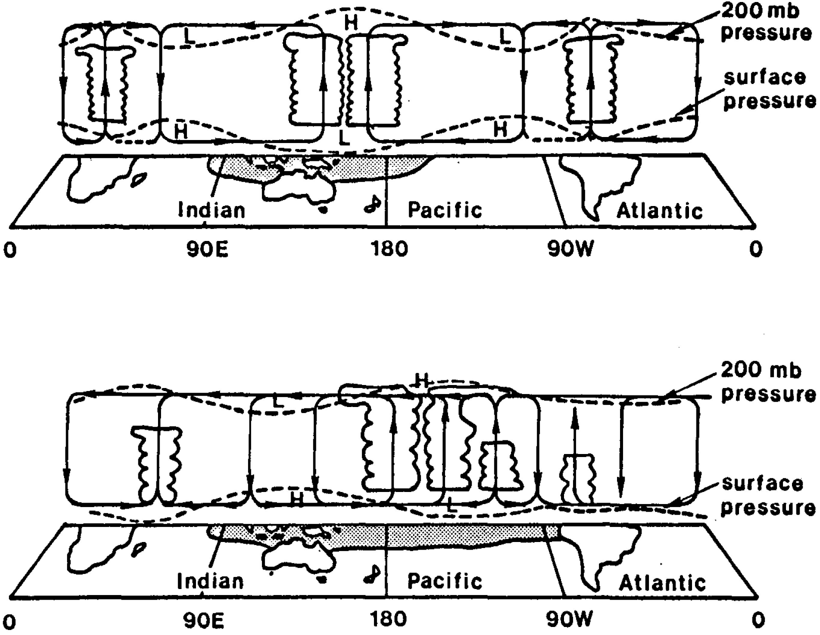
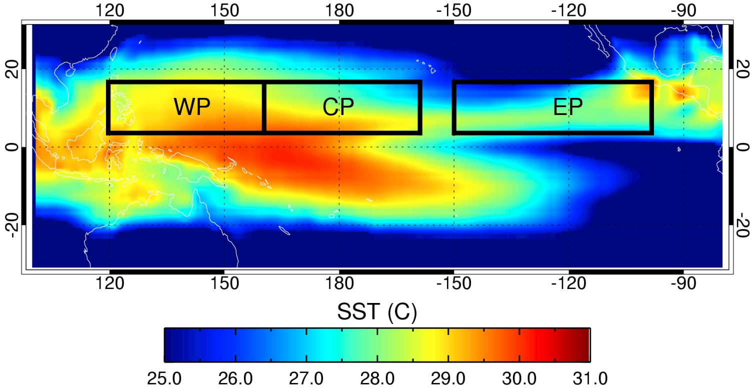
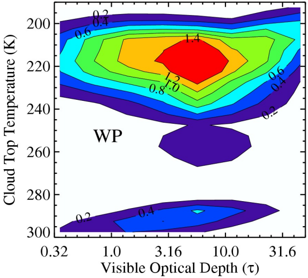


FIG. 1. Schematic view of the Walker Circulation along the equator during El Niño (lower

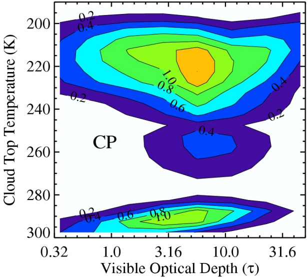
Tropical SST: Sep 2003-Aug. 2005 Kubar, Hartmann, Wood, 2007

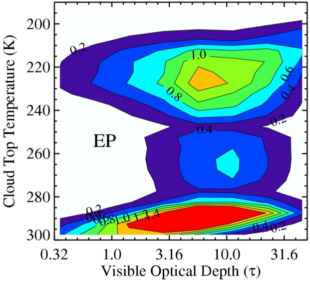


Cloud histograms, 2003-2005



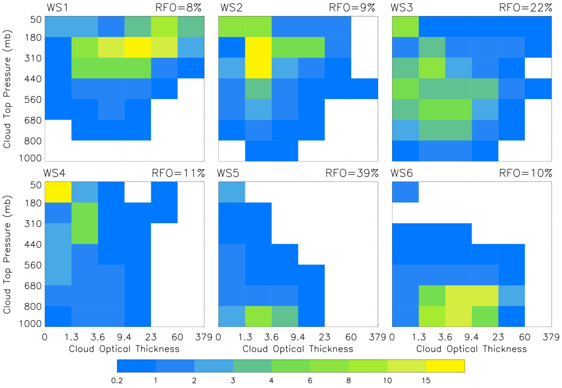
Optical depth/cloud top temperature histograms in the Western, Central and Eastern Pacific (Kubar et al., 2007)





ISCCP: 1983-2004 (Rossow et al., 2005)

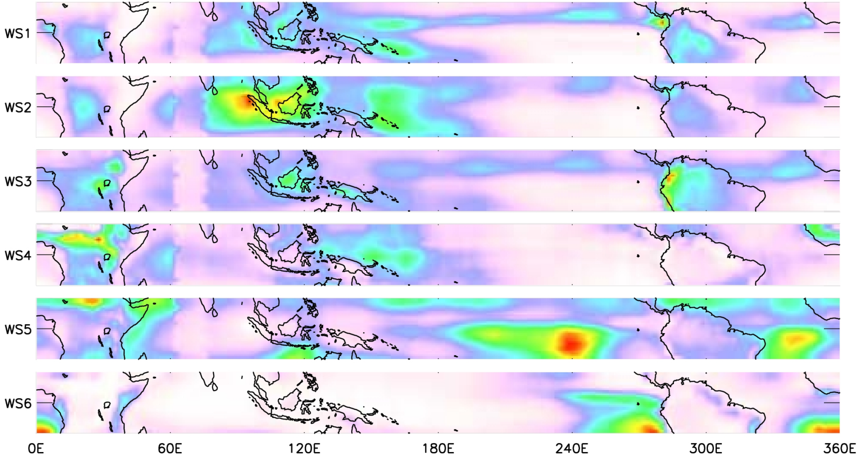
20 year tropical cloud climatology, six “weather states”



WS1=most convectively active, WS6=least convectively active

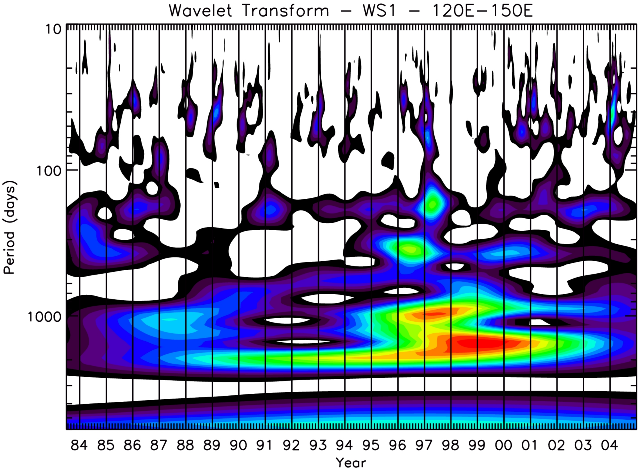
ISCCP: 1983-2004 (Rossow et al., 2005)

Geographic distribution of weather states:

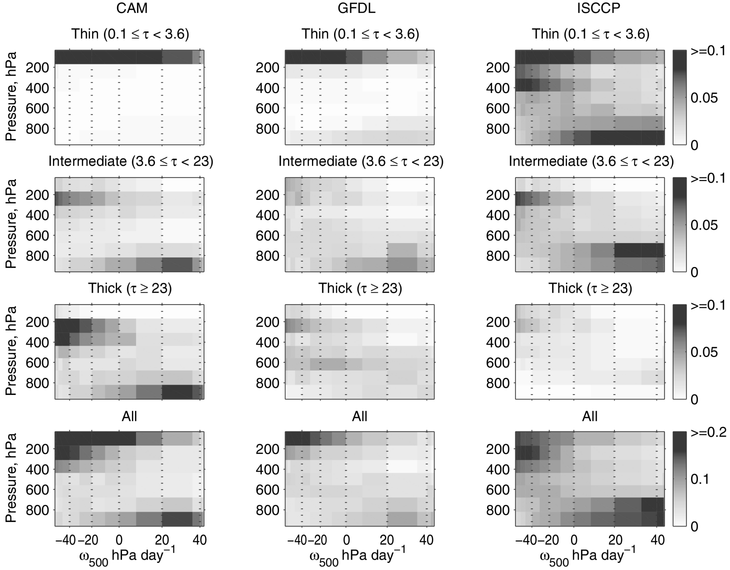


ISCCP: 1983-2004 (Rossow et al., 2005)

Wavelet analysis of most convective active weather state (WS1)

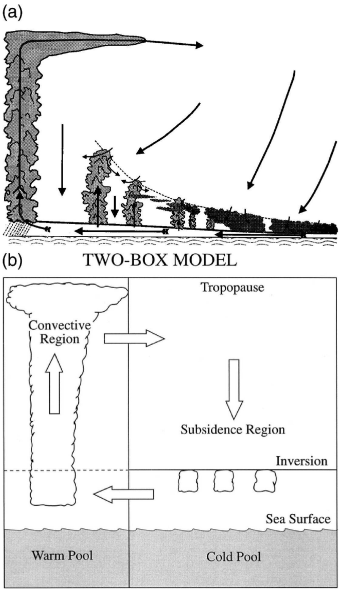


ISCCP simulator intercomparison (Wyant et al., 2006)



Missing middle cloud compensated by excess high cloud.

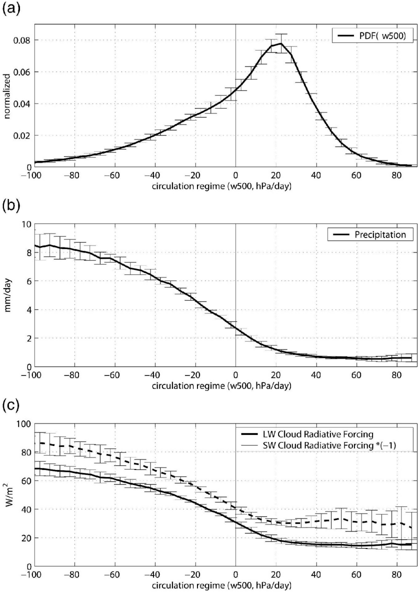
Simple two-box model Bony et al., 2006



A variety of one, two and three cell models idealize the observed circulation:

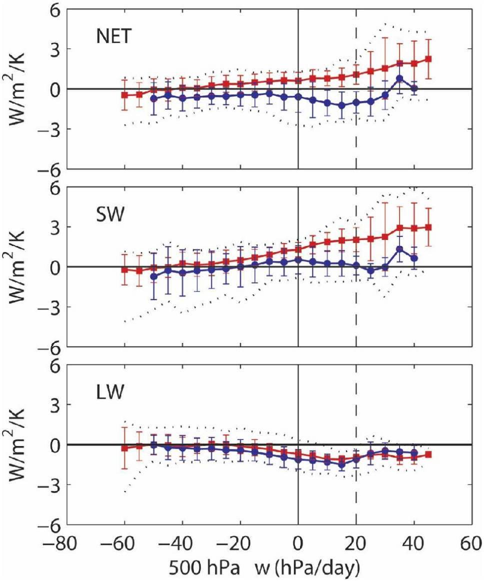
(e.g. Sarachik (1978), Betts and Ridgway (1987), Pierrehumbert (1995), Miller (1997), Larson et al. (1999))

Conditional sampling on ω



ERA-40, GPCP and ERBE data for 1985-1989, 30° N - 30° S (Bony et al., 2006)

Feedback uncertainties in GCMs (Bony et al., 2006)



Sensitivity of tropical (30° N - 30°) cloud radiative forcings for 15 AR4 coupled models (Bony et al., 2006)

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Some thermodynamics: static energy h

$$\text{first law:} \quad \frac{du}{dt} = q - p\frac{d\alpha}{dt} \tag{1}$$

$$\text{enthalpy:} \quad k = u + p\alpha = c_p T \tag{2}$$

$$\text{first law:} \quad \frac{dk}{dt} = q + \alpha\frac{dp}{dt} \tag{3}$$

Use the hydrostatic approximation[1]

$$\frac{dp}{dt} = -\rho g \frac{dz}{dt} \Rightarrow \partial_z p = -\rho g \tag{4}$$

$$\text{(moist) static energy:} \quad h = k + gz \tag{5}$$

$$\text{first law:} \quad \frac{dh}{dt} = q \tag{6}$$

[1] see Madden and Robitaille, 1970, Betts, 1974

moist, liquid and dry static energy (following Emanuel 1994, chap. 4)

$$\text{Total enthalpy:} \quad K = m_d k = m_d k_d + m_v k_v + m_l k_l \qquad (7)$$

Introduce the enthalpy of evaporation (latent heat):

$$l_v = k_v - k_l \qquad (8)$$

and rearrange (7) to get

$$k = (c_{pd} + r_t c_l)T + l_v r_v$$

where the vapor and liquid mixing ratios are $r_v = m_v/m_c$,
$r_l = m_l/m_d$ and $r_t = r_v + r_l$.
If r_t is constant then (3) becomes:

$$dq = (c_{pd} + r_t c_l)dT + d(l_v r_v) - \alpha_d dp \qquad (9)$$

So for an adiabatic process in a hydrostatic atmosphere $dq = 0$ and the moist static energy

$$h = (c_{pd} + r_t c_l)T + l_v r_v + (1 + r_t)gz$$

and liquid water static energy

$$h_l = (c_{pd} + r_t c_{pv})T - (l_v r_l) + (1 + r_t)gz$$

are both conserved.
If $r_l = 0$ then h_l reduces to the dry static energy

$$h_d = (c_{pd} + r_v c_{pv})T + (1 + r_v)gz$$

and if the parcel is saturated with vapor mixing ratio r_s then the saturated moist static energy is

$$h_s = (c_{pd} + r_s c_{pv})T + l_v r_s + (1 + r_s)gz$$

Approximate entropy

Dividing (9) by T gives:

$$ds = \frac{qdt}{T} \approx \frac{c_p}{T} dT + d\left(\frac{l_v}{T} r_v\right) - \frac{R_d}{p_d} dp_d$$

Define the equivalent potential temperature θ_e as

$$c_p \ln \theta_e \equiv s + R_d \ln p_0$$

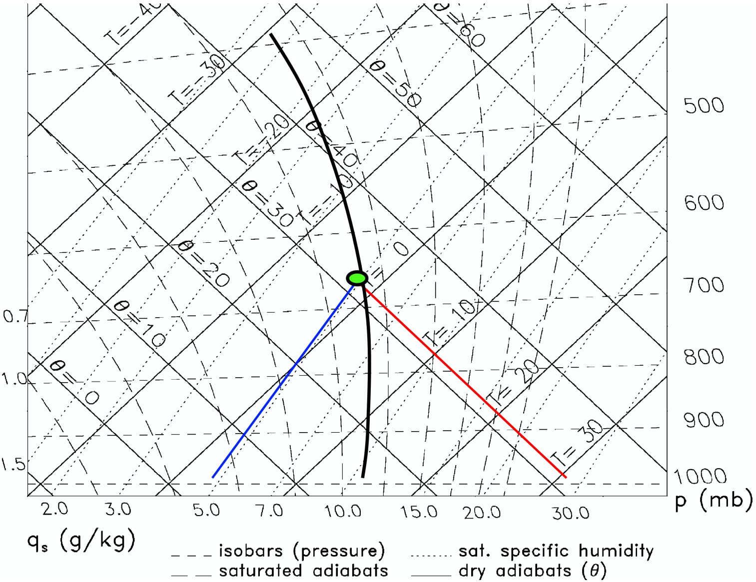
where the reference pressure p_0 is taken to be 100 hPa.
Using this definition:

$$\theta_e = T\left(\frac{p_0}{p_d}\right)^{\frac{R_d}{c_p}} \exp\left[\frac{l_v r_v}{c_p T}\right] = \theta \exp\left[\frac{l_v r_v}{c_p T}\right]$$

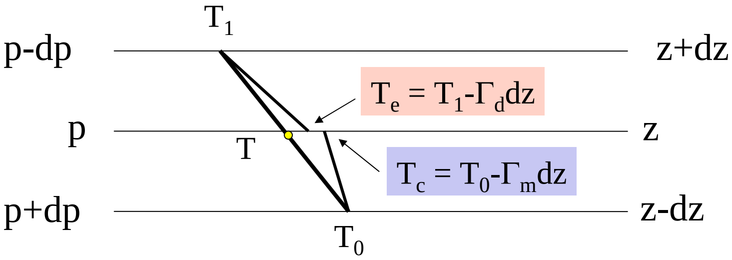
and similarly to the liquid water static energy h_l:

$$\theta_l = \theta \exp\left[\frac{-l_v r_l}{c_p T}\right]$$

Basic tephigram



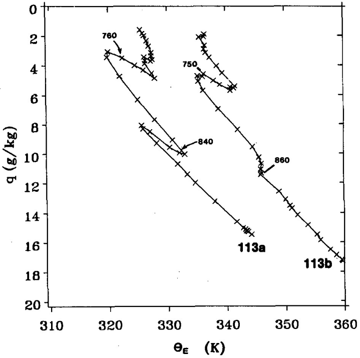
A conditionally unstable atmosphere

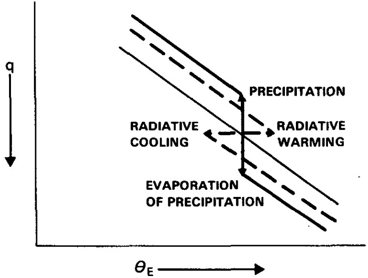


Note that $\Gamma_m < -\frac{dT}{dz} < \Gamma_d$

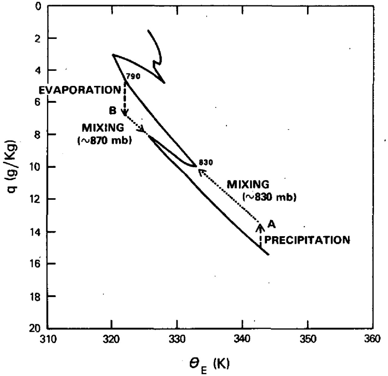
Mixing and evaporation: conserved variables

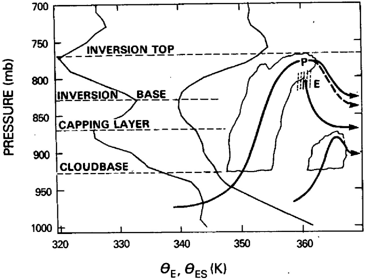
Betts and Albrecht (1987):
FGGE soundings





FGGE soundings, cont.





Buoyancy and CAPE

Define CAPE as the amount of potential energy of a parcel lifted from level i to its level of neutral buoyancy:

$$CAPE_i = \int_i^{LNB} B dz$$

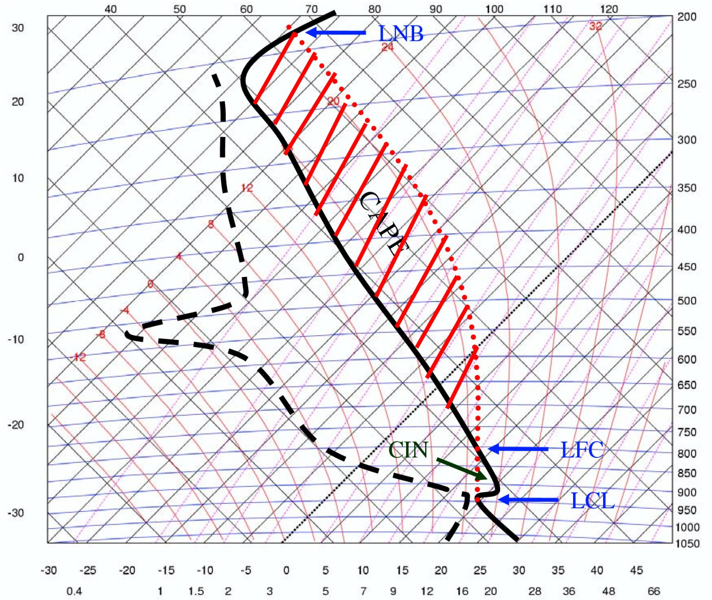
where B is the buoyancy:

$$B = -g\left(\frac{\rho_p - \rho_e}{\rho_p}\right) = g\left(\frac{\alpha_p - \alpha_e}{\alpha_e}\right) = g\left(\frac{T_{vp} - T_{ve}}{T_{ve}}\right)$$

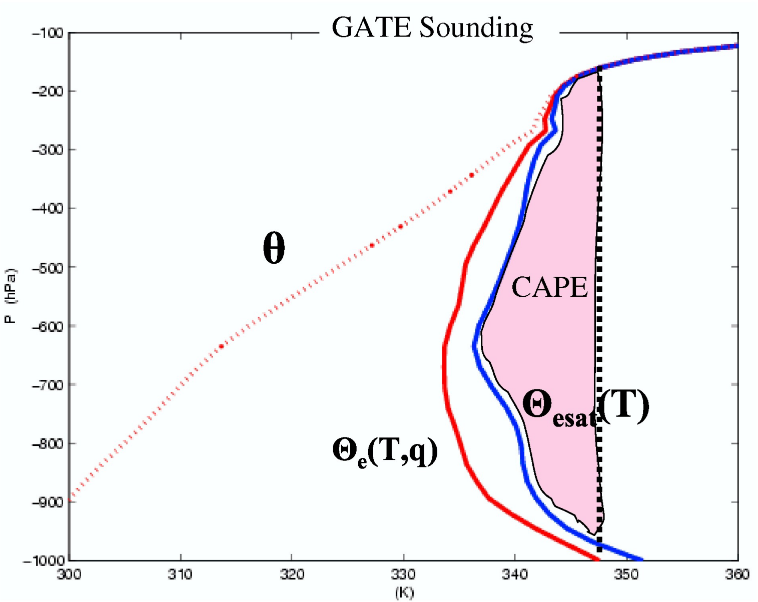
So that, assuming hydrostatic equilibrium:

$$CAPE_i = \int_i^{LNB} g\left(\frac{\alpha_p - \alpha_e}{\alpha_e}\right) dz = \int_{p_n}^{p_i} (\alpha_p - \alpha_e) dp = \int_{p_n}^{p_i} R_d(T_{vp} - T_{ve}) d \ln p$$

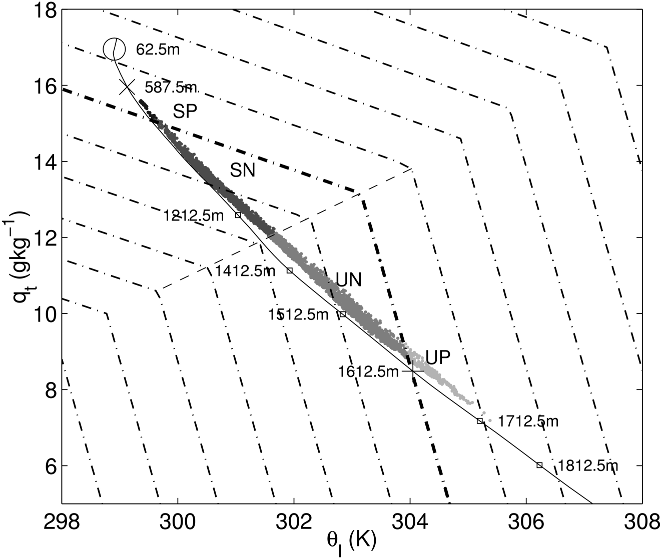
CAPE example source: Bechtold, Jakob and Gregory, 2006



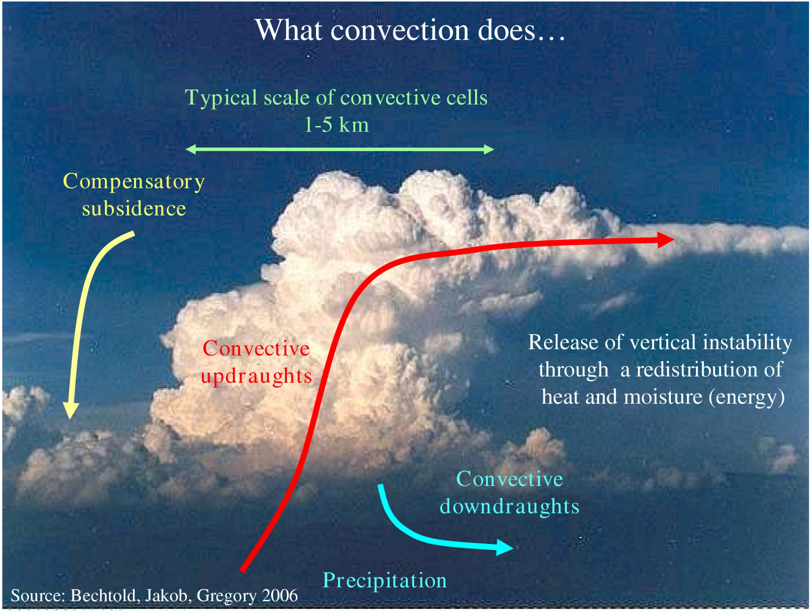
CAPE in θ_e, q coordinates, source: Bechtold, Jakob and Gregory, 2006



A slice through a modeled cloud at 1612 m, in θ_l, q_t coordiates



A reminder


What convection does…
Typical scale of convective cells
1-5 km
Compensatory
subsidence
Convective
updraughts
Release of vertical instability
through a redistribution of
heat and moisture (energy)
Convective
downdraughts
Precipitation
Source: Bechtold, Jakob, Gregory 2006

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Slice method (Bjerknes, 1938, Randall, 2006)

Divide a domain into N vertical columns of fractional area σ_i with vertical velocity w_i and static energy h_i. Then

$$\sum_{i=1}^{N} \sigma_i = 1; \; \sum_{i=1}^{N} \sigma_i w_i = \overline{w}; \; \sum_{i=1}^{N} \sigma_i h_i = \overline{h}$$

$$\text{static energy flux: } F_h = \rho\overline{wh} - \rho\overline{w}\overline{h} = \sum_{i=1}^{N} \rho\sigma_i (w_i - \overline{w})(h_i - \overline{h})$$

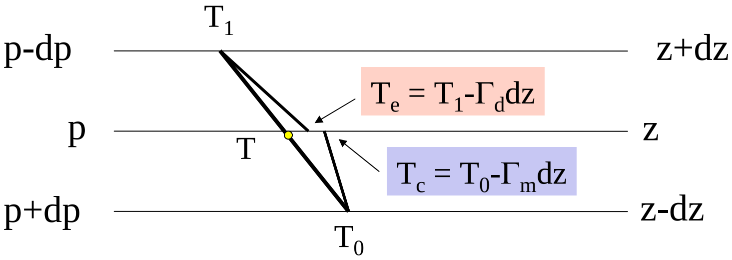
The fraction of columns of convective cloud (c)/environment (e):

$$\sigma_c = \sum_{\{cloudy\}} \sigma_i \text{ and } \sigma_e = 1 - \sigma_c$$

Conditional averages:

$$w_c = \frac{\sum_{\{cloudy\}} \sigma_i w_i}{\sigma_c} \text{ and } h_c = \frac{\sum_{\{cloudy\}} \sigma_i h_i}{\sigma_c}$$

A conditionally unstable atmosphere



Note that $\Gamma_m < -\dfrac{dT}{dz} < \Gamma_d$

Slice method ...

$$\sigma_c w_c + \sigma_e w_e = \overline{w} \tag{10a}$$

$$\sigma_c h_c + \sigma_e h_e = \overline{h} \tag{10b}$$

Since $w_e < 0$ both T_c and T_e are increasing:

$$\begin{aligned} \partial_t T_c &= w_c(\Gamma - \Gamma_m) > 0 \\ \partial_t T_e &= w_e(\Gamma - \Gamma_d) > 0 \end{aligned}$$

and using (10a):

$$\begin{aligned} w_c &= \overline{w} + (1 - \sigma_c)(w_c - w_e) \\ w_e &= \overline{w} - \sigma_c(w_c - w_e) \end{aligned}$$

Slice method ...

Which can be combined to give the rate of increase of convection:

$$\partial_t(T_c - T_e) = w_c(\Gamma - \Gamma_m) - w_e(\Gamma - \Gamma_d) =$$
$$\overline{w}(\Gamma_d - \Gamma_m) + (w_c - w_e)\left[(1 - \sigma_c)(\Gamma - \Gamma_m) + \sigma_c(\Gamma - \Gamma_d)\right]$$

so that convection is favored for a rapidly ascending narrow updraft and a wide sinking environment ($\sigma_c \to 0$):

Energy and moisture tendencies (Bechtold, Jacob and Gregory, 2006)

Given the dry static energy: $h_d = c_p T + gz$ and the specific humidity q decomposed into $\phi = \overline{\phi} + \phi'$:

$$\partial_t \overline{h_d} = \underbrace{-\overline{\mathbf{V}}_H \cdot \nabla \overline{h_d} - \overline{w}\partial_z \overline{h_d}}_{I} + \overbrace{\underbrace{L(\overline{c} - \overline{e})}_{II} - \underbrace{\partial_z \overline{w' h'_d}}_{III}}^{Q1} + \underbrace{c_p Q_R}_{IV} \quad (11)$$

$$\partial_t \overline{q} = \underbrace{-\overline{\mathbf{V}}_H \cdot \nabla \overline{q} - \overline{w}\partial_z \overline{q}}_{I} - \overbrace{\left(\underbrace{(\overline{c} - \overline{e})}_{II} + \underbrace{\partial_z \overline{w' q'}}_{III} \right)}^{Q2} \quad (12)$$

where

I=resolved scale transport

II=large-scale condensation/evaporation

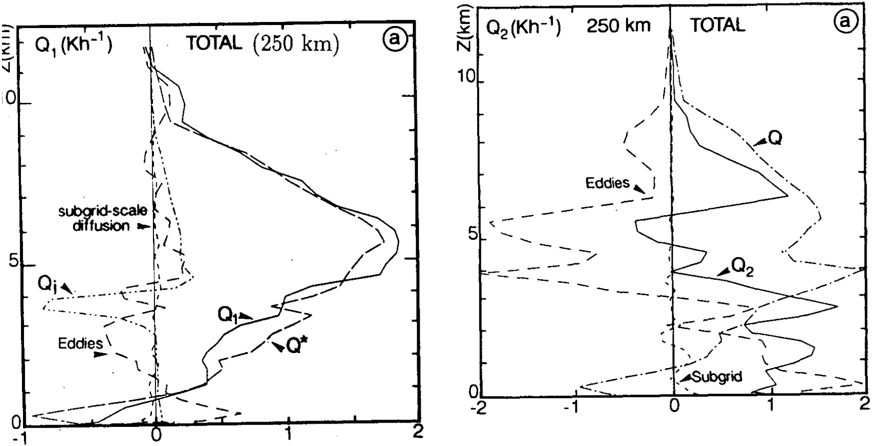
III=subgrid-scale transport (turbulence + convection)

IV=radiation

Q_1=apparent heat source, Q_2=apparent moisture sink

Q_1 and Q_2 from a CRM

(Caniaux, Redelsperger, Lafore, J. Atmos. Sci., 1994)



where $Q^* = L(\overline{c} - \overline{e}), Q = (\overline{c} - \overline{e}),\ Q_1 = \frac{1}{c_p}(Q_* - \partial_z \overline{w' h'_d})$

and $Q_2 = Q + \partial_z \overline{w' q'}$

What can we say about the two eddy terms $\partial_z \overline{w' h'_d}$ and $\partial_z \overline{w' q'}$?

Mass flux approximation (BJG, 2006)

We can use a simple mass flux approximation to get some physical insight into Q_1 and Q_2.
Recall (10): if $\sigma_c \ll 1$ then $h_e \approx \overline{h}$ and

$$\overline{h} = \sigma h_c + (1 - \sigma) h_e \tag{13}$$

$$\overline{w'h'} = \overline{wh} - \overline{w}\overline{h} = \sigma(1 - \sigma)(\overline{w_c} - \overline{w_e})(\overline{h_c} - \overline{h_e}) \tag{14}$$

and since $\overline{w_c} \gg \overline{w_e}$

$$F_h = \rho\overline{w'h'} = \rho\sigma w_c(\overline{h_c} - \overline{h}) = M_c(\overline{h_c} - \overline{h})$$

where $M_c = \rho\sigma w_c$ is the convective mass flux.

Mass flux continued

How does the cloud ensemble M_c depend on height? Try a simple entraining/detraining plume:

$$\frac{\partial M_c}{\partial z} = \epsilon - \delta$$

$$\frac{\partial (M_c \overline{h_{dc}})}{\partial z} = \epsilon \overline{h_d} - \delta \overline{h_{dc}} + Lc$$

so that the apparent heat source Q_1:

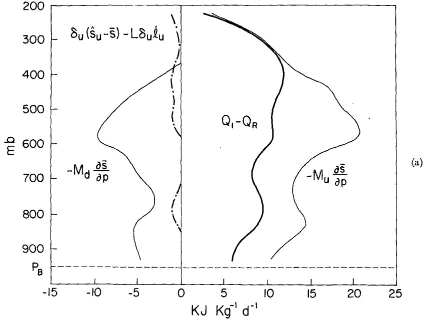
$$Q_1 = L(\overline{c} - \overline{e}) - \partial_z \overline{w' h'_d} = L(\overline{c} - \overline{e}) - \partial_z (M_c (\overline{h_{dc}} - \overline{h_d}))$$

$$Q_1 = \underbrace{M_c \frac{\partial \overline{h_d}}{\partial z}}_{I} + \underbrace{\delta(\overline{h_{cd}} - \overline{h_d})}_{II} - \underbrace{Le}_{III}$$

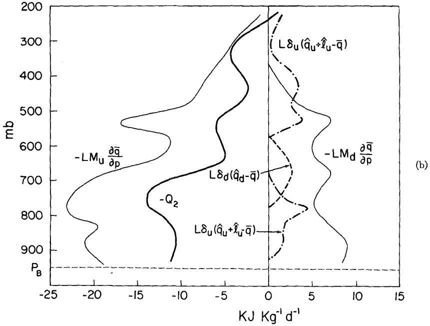
where term I represents the warming of the environment due to *compensating subsidence*, *II* is detrainment and *III* is evaporation of cloud and precipitation.

Q1 and Q2 diagnosed with plume model (Nitta, 1977)

static energy terms:

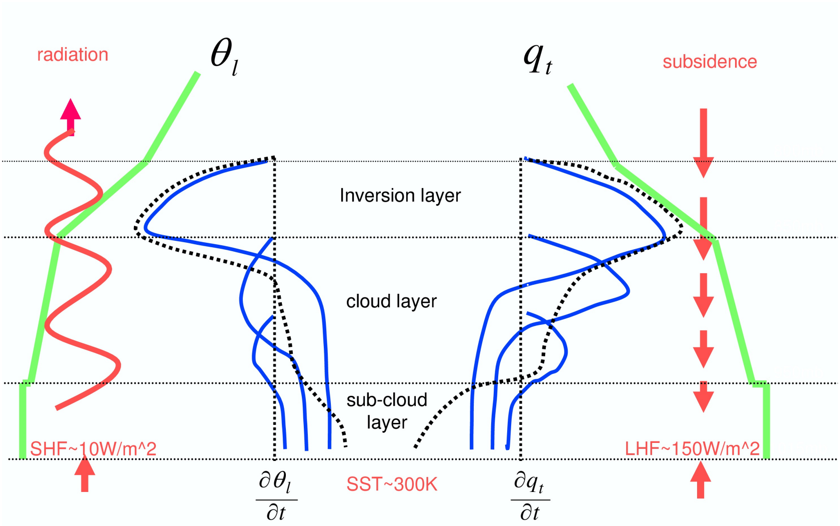


specific humidity terms:



Small clouds cool and moisten at cloud top, large clouds moisten and heat through compensating subsidence. (Note that this model include downdrafts).

Heating/moistening for 4 cloud sizes



Note that the in the tropics the boundary layer fluxes, subsidence and radiation are all tightly coupled.

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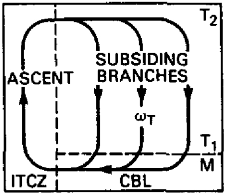
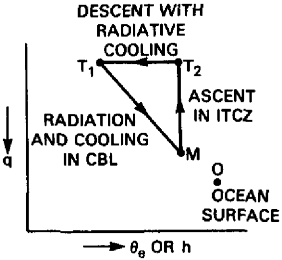
One cell model (Betts and Ridgway, 1988)



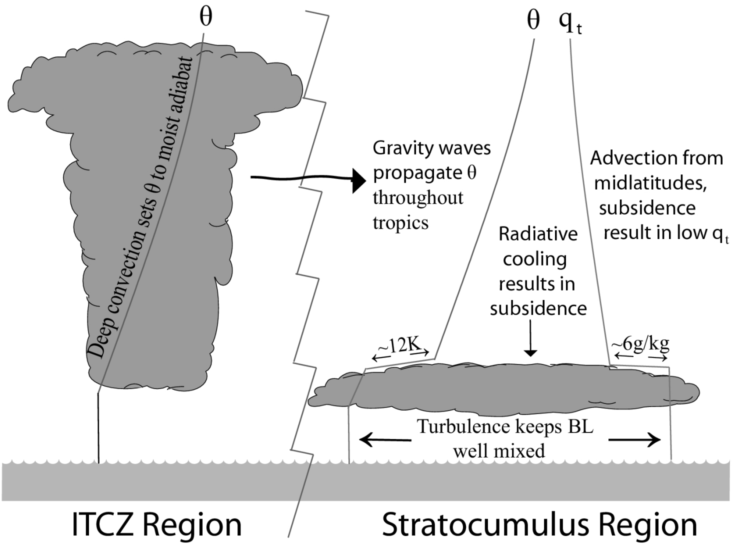


Some constraints:

- Free tropospheric temperature is horizontally uniform
- Convection is in equilbrium with large scale forcing
- Subsidence balances radiative cooling in the descending branch

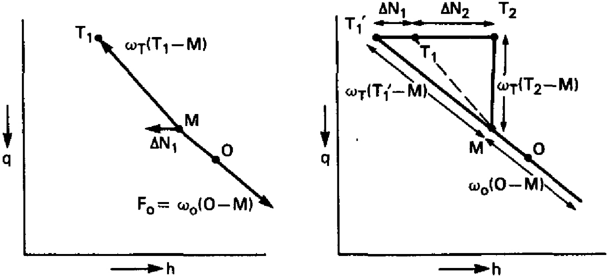
$$w \frac{d\theta}{dz} = Q_R$$

WTG approximation (Caldwell and Bretherton (2007))



One cell model, continued

Betts and Ridgway 1988: balances for the boundary layer (left) and entire troposphere (right)



Need models of column radiation and cloud/humidity profiles get ΔN.

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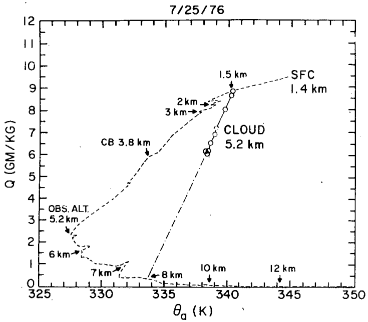
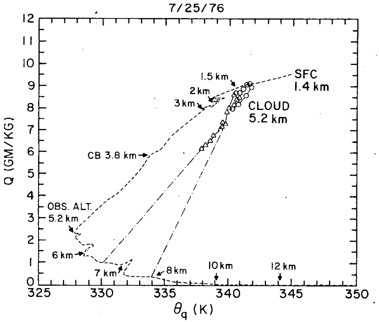
Evidence for buoyancy sorting (Paluch, 1979)

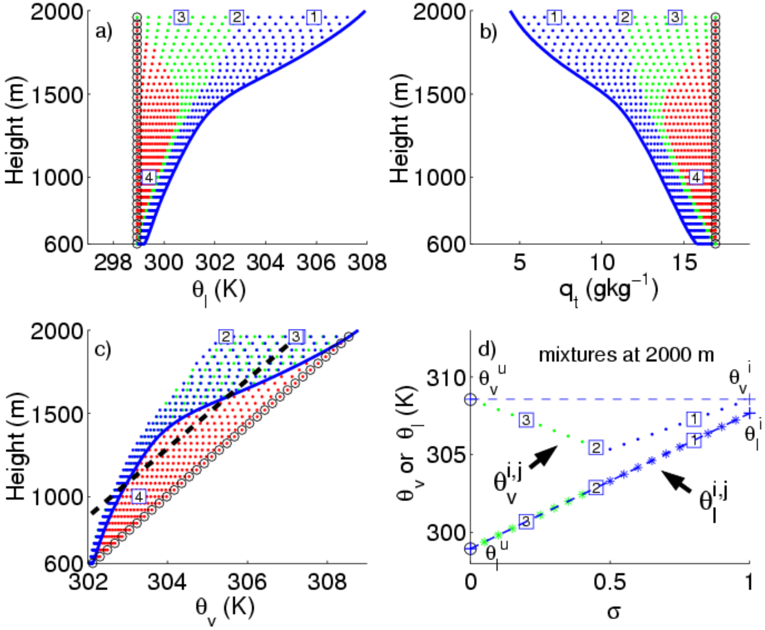
In-cloud observations appear to be formed by mixing between two distinct levels



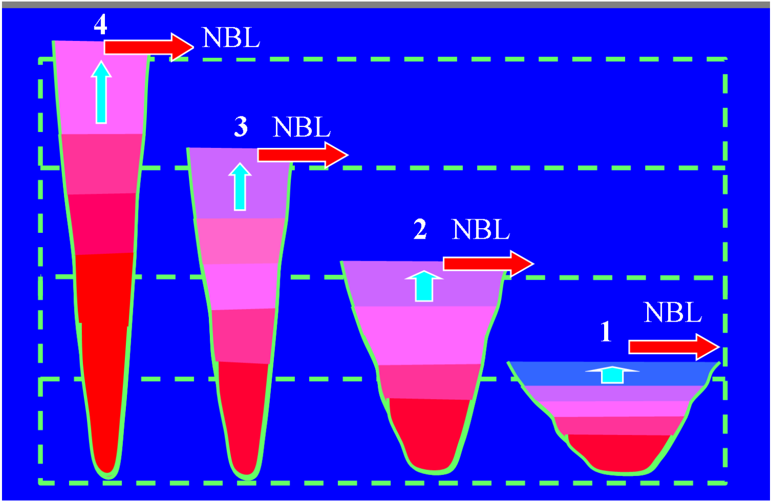


In fact, the cloud parcels are moving to their level of neutral buoyancy

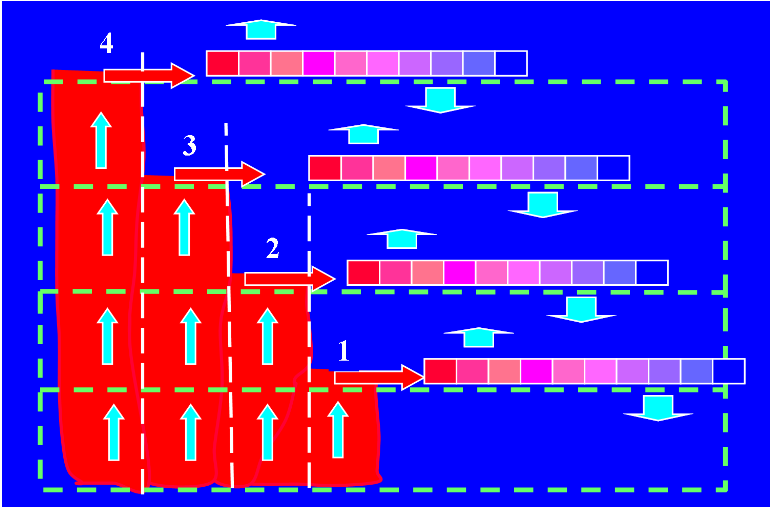
Buoyancy sorting in shallow clouds



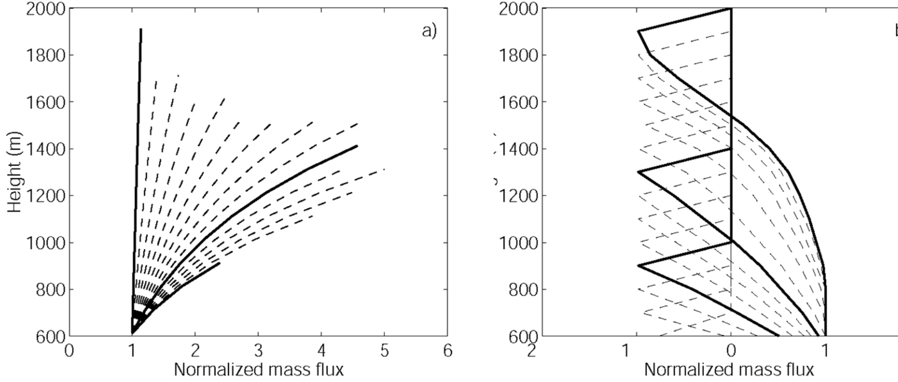
Spectral entraining plumes



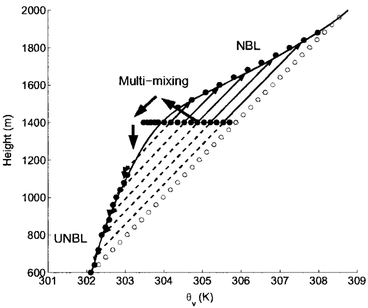
Episodic mixing/buoyancy sorting

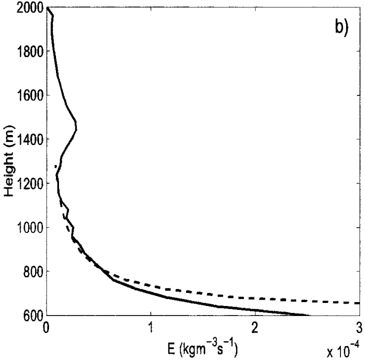


Buoyancy sorting vs. entraining plume

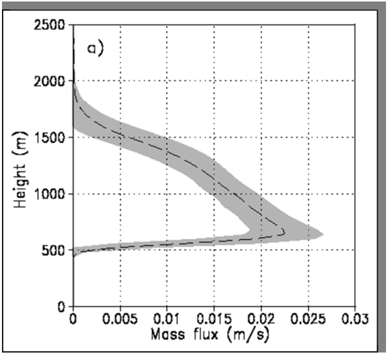


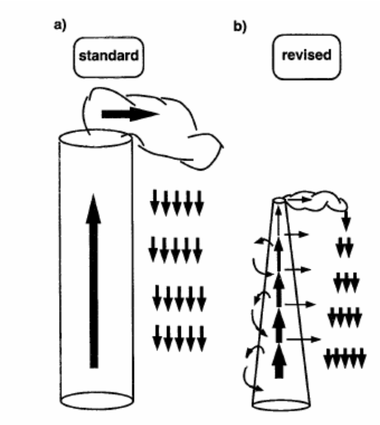
Detrainment and the cloud size distribution Zhao and Austin, 2003





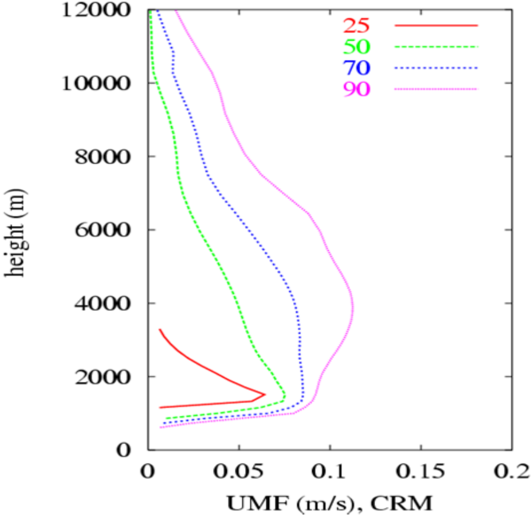
Mass flux decreases with height Siebesma, 2005





$$\frac{1}{M}\frac{\partial M}{\partial z} = \epsilon - \delta$$

and CRMs indicate the mass flux is sensitive to relative humidity



(Derbyshire et al. 2004)

Outline

1. Satellite/reanalysis views of tropical clouds (MODIS, ISCCP, Bony et al.)
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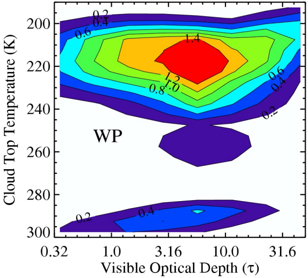
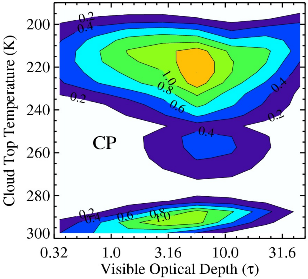
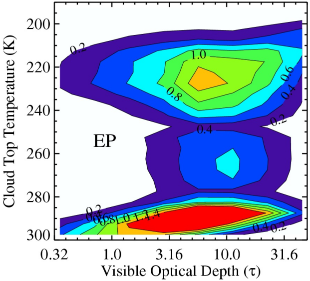
5. Entrainment, detrainment, buoyancy sorting
6. ⇒ What controls convective cloud top height?

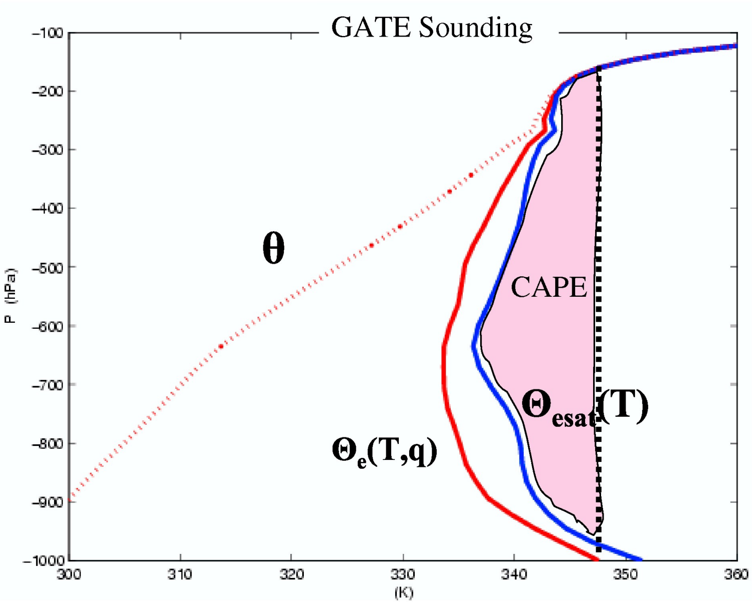
Why do clouds detrain before they hit the tropopause?



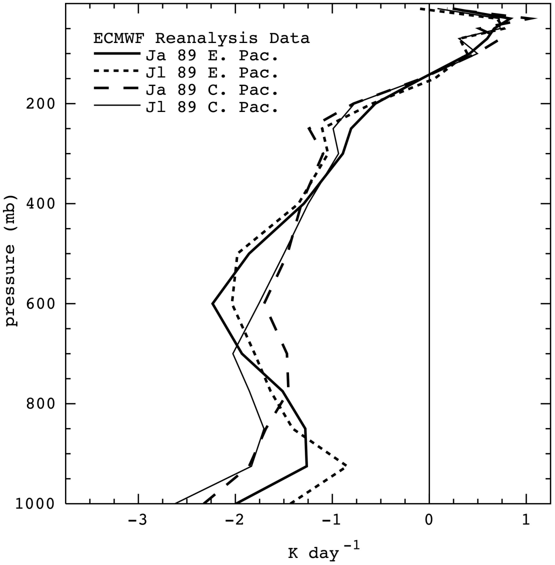




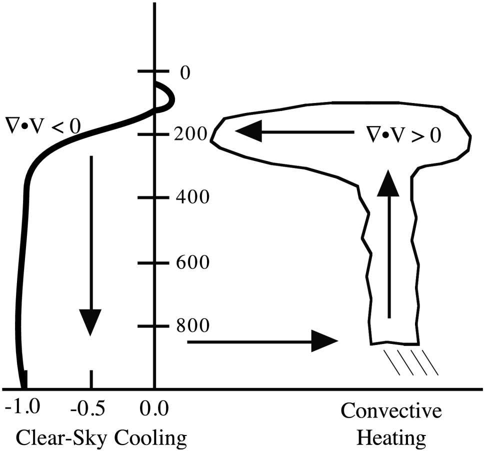
Is cloud top determined by sub-cloud θ_e?



Little water vapor above 200 hPa

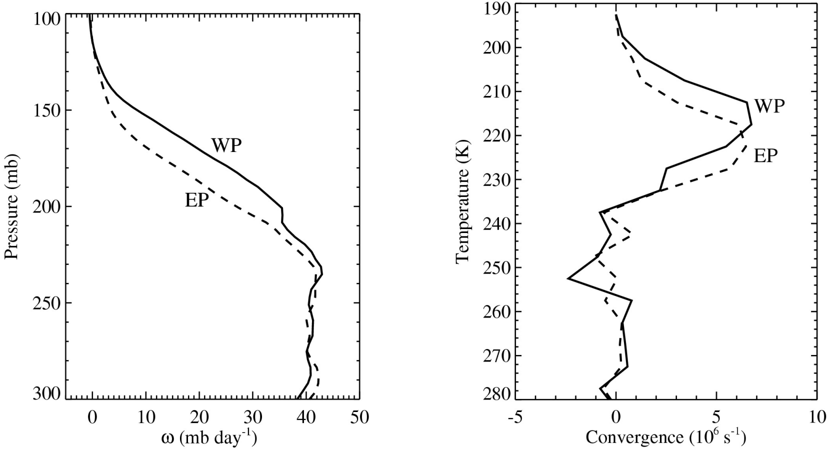


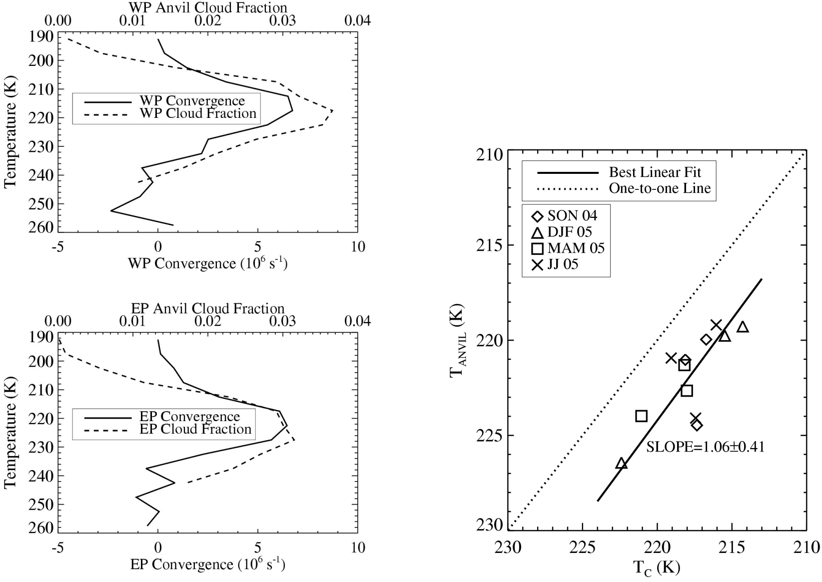
Larsen and Hartmann, 2001

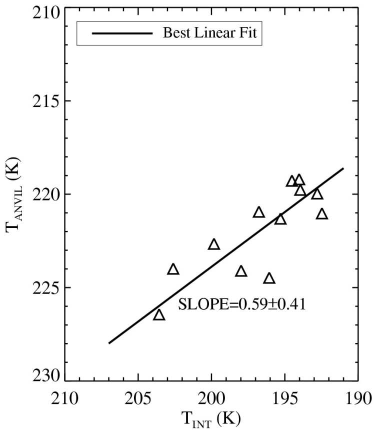


Can mass convergence at 200 hPa promote detrainment?

Kubar, Hartmann and Wood (2007)




WP Anvil Cloud Fraction
0.00
0.01
0.02
0.03
0.04
190
200
210
220
230
240
250
260
Temperature (K)
WP Convergence
WP Cloud Fraction
-5
0
5
10
WP Convergence (10^6 s^{-1})
EP Anvil Cloud Fraction
EP Convergence
EP Cloud Fraction
EP Convergence (10^6 s^{-1})
Best Linear Fit
One-to-one Line
SON 04
DJF 05
MAM 05
JJ 05
SLOPE=1.06±0.41
T_{ANVIL} (K)
T_C (K)
210
215
220
225
230



c.f. poor correlation between anvil temperature and adiabatic cloud top

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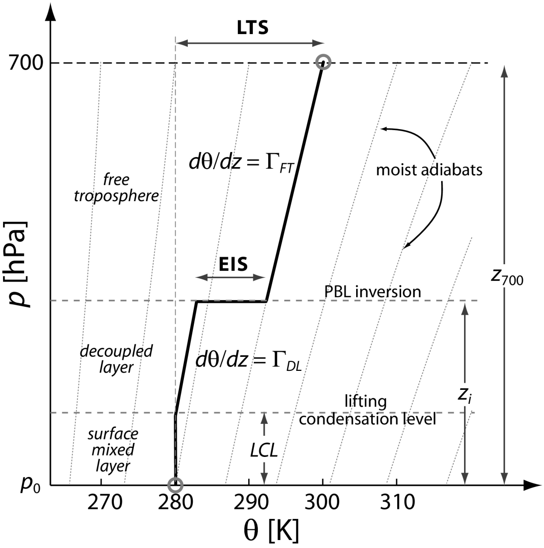
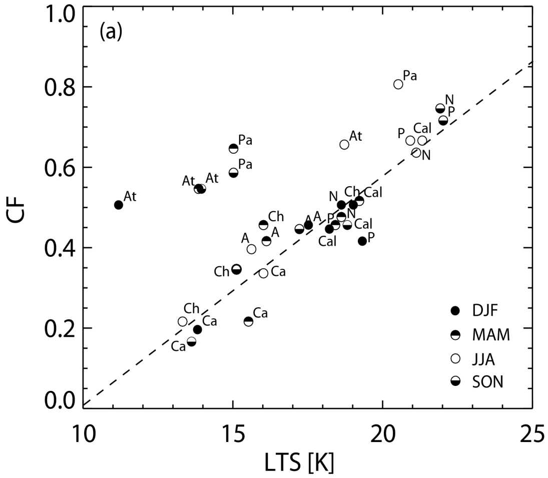
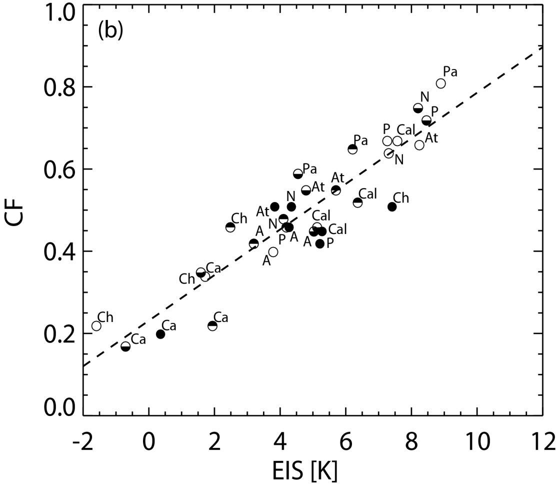
Linking cloud fraction to inversion strength (Wood and Bretherton, 2006)




(a)
CF
LTS [K]
1.0
0.8
0.6
0.4
0.2
0.0
10
15
20
25
DJF
MAM
JJA
SON
Pa
At
P
Cal
N
Ch
A
Ca



(b)
CF
EIS [K]
1.0
0.8
0.6
0.4
0.2
0.0
-2
0
2
4
6
8
10
12
Pa
N
P
P
Cal
At
Pa
N
Pa
At
At
N
Cal
Ch
At
Ch
N
Cal
Cal
A
P
A
A
P
A
Ch
Ca
Ch
Ca
Ca
Ca