Boundary layer equilibrium [2005]
– over tropical oceans

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Based on:

Betts, A.K., 1997: Trade Cumulus: Observations and Modeling. Chapter 4 (pp 99-126) in *"The Physics and Parameterization of Moist Atmospheric Convection*, Ed. R. K. Smith, NATO ASI Series C: Vol. **505**, Kluwer Academic Publishers, Dordrecht, 498pp.

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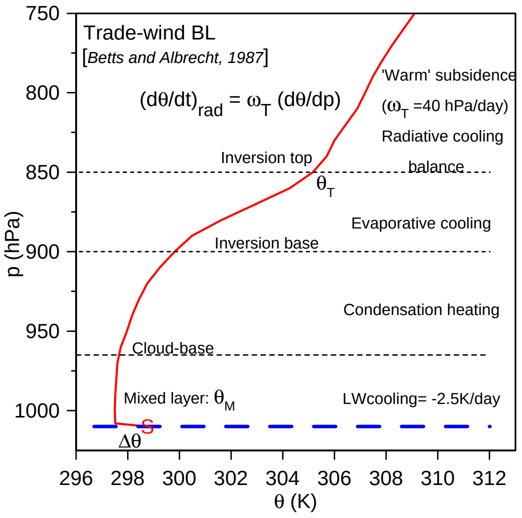
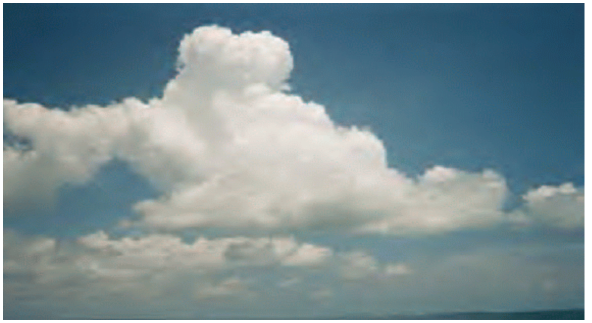




Why is mixed layer cooler than the ocean SST?

LW cooling = -2.5 K/day

Clouds redistribute heat and water and modify radiative balance

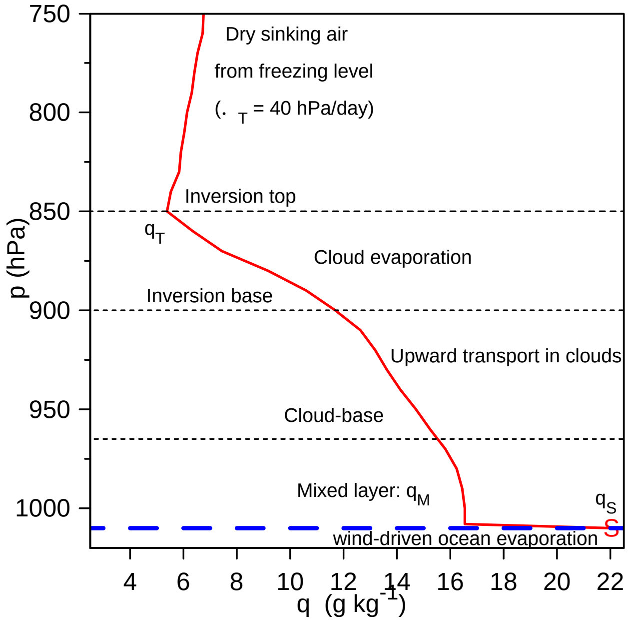
Equilibrium for whole layer:

$$0 = (g/C_p)\,\Delta N_T \quad + \omega_0\,\Delta\theta + \omega_T\,(\theta_T - \theta_M)$$

-40	+10	+30 W m^{-2}
cooling	surface flux	subsidence

Surface velocity scale: $\omega_0 = \rho V_0\, C_D \approx$ 90hPa/day

Subsidence: $\omega_T \approx$ 40 hPa/day



Why is the mixed layer not saturated, as the air blows over ocean? Evaporation from ocean is balanced by subsidence of dry air above.

$$0 = \omega_0\ (q_S(SST) - q_M) + \omega_T\ (q_T - q_M)$$

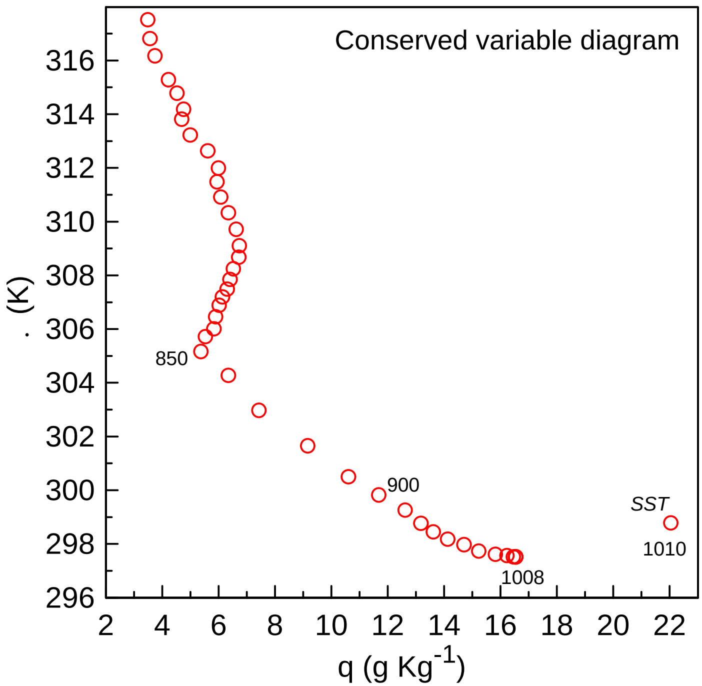
$$q_M = [\omega_0\ q_S(SST) + \omega_T\ q_T]/(\omega_0 + \omega_T)$$

A weighted average

q_M = [90*22 +40*5]/130 = 16.7 g/kg

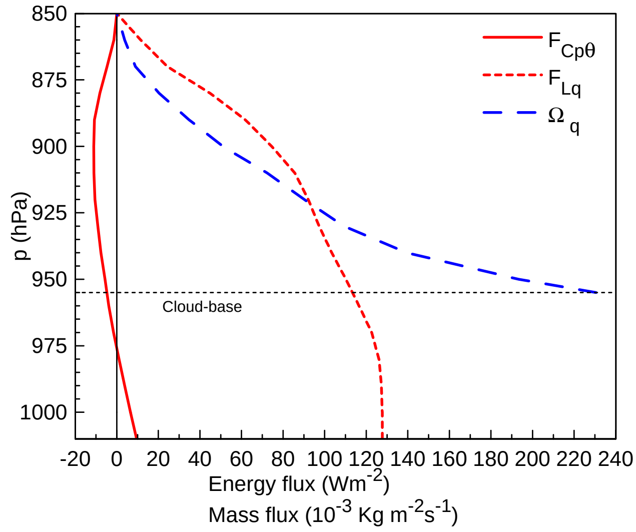
so $\theta_{EM} \approx$ 346K

cloud-base ≈ 960hPa



Can think of the two balances on a ‘conserved parameter’ diagram: “Mixing” of surface point and 850hPa point, modified by radiation.

Relate equilibrium structure to convective fluxes: F_q, F_θ [illustration]



Assume ω = 40hPa/day in cloud layer, below cloud-base decreases linearly to zero at surface.
Assume radiative cooling $\partial\theta_{Rad}/\partial t$ = -2.4 K/day

Equilibrium means steady state [assume horizontally homogeneous]

$0 = \partial F_q/\partial p + \omega\, \partial \bar{q}/\partial p$
$0 = \partial F_\theta/\partial p + \omega\, \partial \bar{\theta}/\partial p + \partial\theta_{Rad}/\partial t$

[where F_q and F_θ represent the convective fluxes of total water and 'liquid water potential temperature' above cloud-base]

Integrate to give fluxes from ω, θ and q profiles, and $\partial\theta_{Rad}/\partial t$.
This gives *equilibrium fluxes* [in units of W m^{-2}] *from profiles*

Simple mass-flux model [illustration]
Can couple fluxes with a *mass flux transport model* for shallow convection

$F_q = \Omega_q\,(q_c - q)$ with $q_c = q_B$ a cloud-base value of 16.54 gkg^{-1}

and compute the Ω_q shown in the figure.

Shallow Cumulus

– non-precipitating

– net latent heat release = 0

– but transport heat because condense water, advect it upward and reevaporate it [a “refrigerator”]

– buoyant, because of condensation but still ‘cold’, because of liquid

– conserved variables: $\theta_E = \theta + Lq/C_p$

$\theta_L = \theta - L\ell/C_p$

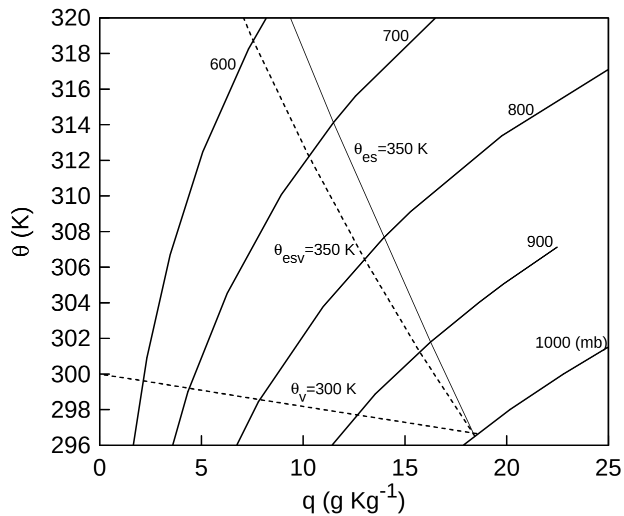
$q_T = q + \ell$

– represent by mass transport of air with sub-cloud properties to higher levels

– equilibrium structure over ocean is balance of convective transports, subsidence, and radiative flux divergence (cooling)

Conserved Variable diagram – 2

– Similar to other thermodynamic diagrams; just θ, q as axes



Dry virtual potential temperature

$\theta_v = \theta(1+.608*q/1000)$

– vapor is less dense

– 'Saturation Points' of equal density

– Slopes 1K every 6g kg^{-1}

[Could use as axis]

Wet virtual potential temperature

– if parcels carry liquid .. Denser; $\Delta\ell$=2 g kg^{-1}/100hPa

– $\theta_v = \theta(1+.608*q/1000 - \ell/1000)$

– line of equal density

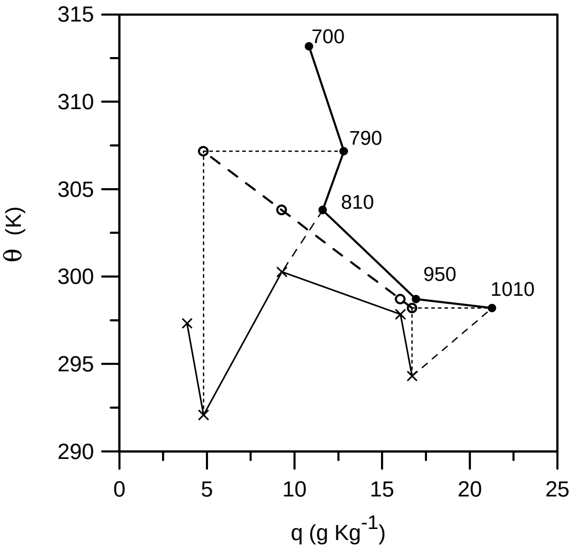
$(\partial\theta/\partial p)_{\theta esv} \approx 0.9\ (\partial\theta/\partial p)_{\theta es}$

Parameterizing shallow convection with a mixing line representation

– parameterize a cloud field: what do these simple diagnostic studies tell us?
– two approaches:

a) parameterize fluxes, and their gradients:
eg with mass flux model; say cloud-base q-flux = surface q flux
[Problem from a ‘climate perspective is that system may drift to either dry or cloudy state]

b) parameterize structure: eg ‘mixed layer’ or ‘mixing line’.



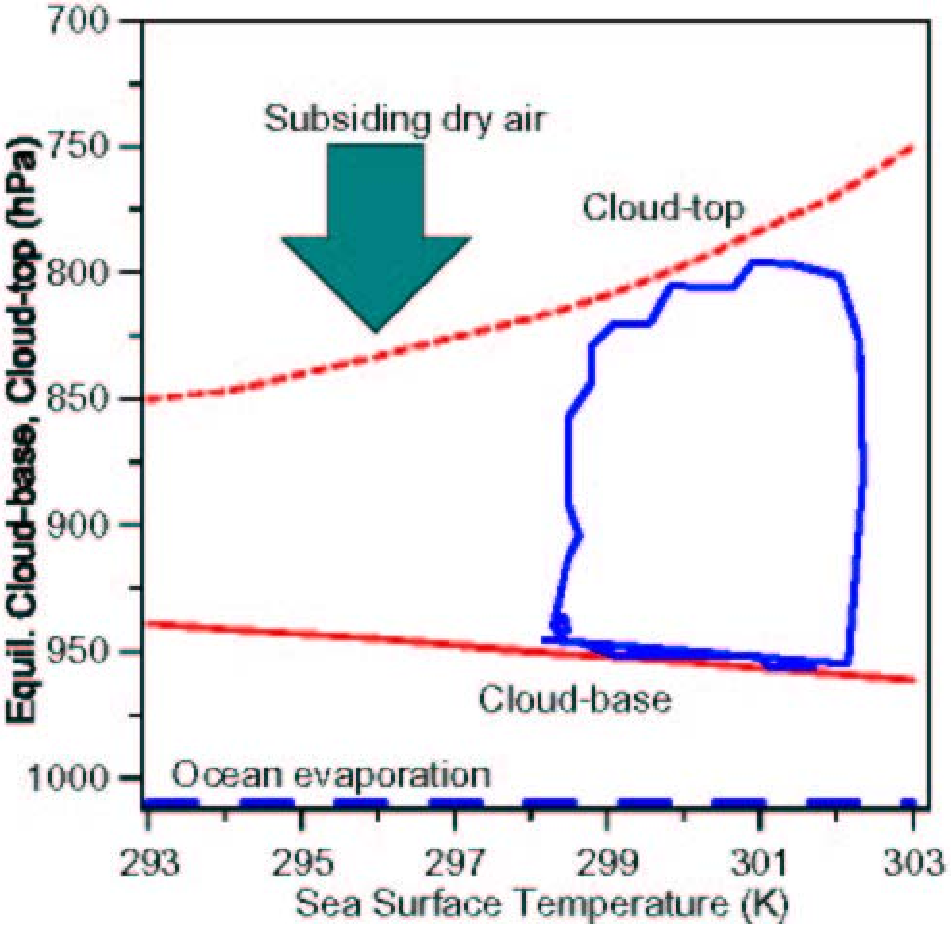
Single mixing line can represent whole BL structure of both clear and cloudy air.

Unsaturated air: find T, T_d at p by drawing lines of constant θ and q
Cloudy air: find T, T_d [for total water] at p by drawing lines of constant θ_{es} and q

Often useful to have a model for structure; eg to compute radiation

[example: Betts and Ridgway,1989]

Climate equilibrium in the Tropics



Shallow Trade-wind cumulus flowing into deep precipitating tropical convergence zone

Trade cumulus balance between ocean evaporation and sinking of dry air

Tropical Climate equilibrium

[Betts and Ridgway, JAS 1988,1989]

- **Consider subsiding branches of tropical circulation, like the Trades.**

  [Moisture evaporated here flows into the convergence zones and tropical disturbances where it is precipitated]

- **Energy balance closures give radiative-convective equilibrium; but there are several important timescales**

1. Subcloud layer thermal balance [one-day timescale]

$$H + H_B = \Delta N_B$$

Where radiative cooling of sub-cloud layer,

$\Delta N_B \approx 10 Wm^{-2}$ [-2.5K/day * 50hPa]

Cloud-base flux: $H_B \approx - 0.2\ H$ [surface flux]

Giving $H = \Delta N_B/1.2 \approx 8\ Wm^{-2}$

So radiative cooling of sub-cloud layer gives small sea-air temperature difference and small sensible heat flux.

[Bowen ratio over tropical oceans is small.]

2. CBL budgets [1 to 2-day timescale]

Heat

$$H \quad + \quad (C_p/g)\omega_T(\theta_T-\theta_M) \quad = \quad \Delta N_T$$

surface cooling — subsidence warming — CBL radiative

Water

$$\lambda E + (L/g)\omega_T(q_T-q_M) \quad = \quad 0$$

surface — subsidence drying

λE linked to subsidence ω_T

Given ω_T, $(\Delta N_T\text{-}H)$ gives θ_T and CBL-top

3. Tropospheric energy balance [10-day timescale]

The atmospheric energy balance averaged over the tropics can be written

$H + \lambda E = \Delta N_{TR}$ + atmospheric export from tropics

Where radiative cooling of troposphere $\Delta N_{TR} \approx 150\ Wm^{-2}$

So surface evaporation

$\lambda E \approx 150 - 8 \approx 142\ Wm^{-2}$

The mechanism is that the radiative cooling drives the subsiding branch, bringing dry air into the CBL, which balances evaporation locally; and the moisture flows into the convergence zones, condensing and releasing latent heat which balance the radiative cooling.

4. Ocean mixed layer and SST equilibrium [>100day timescale]

$H + \lambda E = N +$ oceanic export from tropics

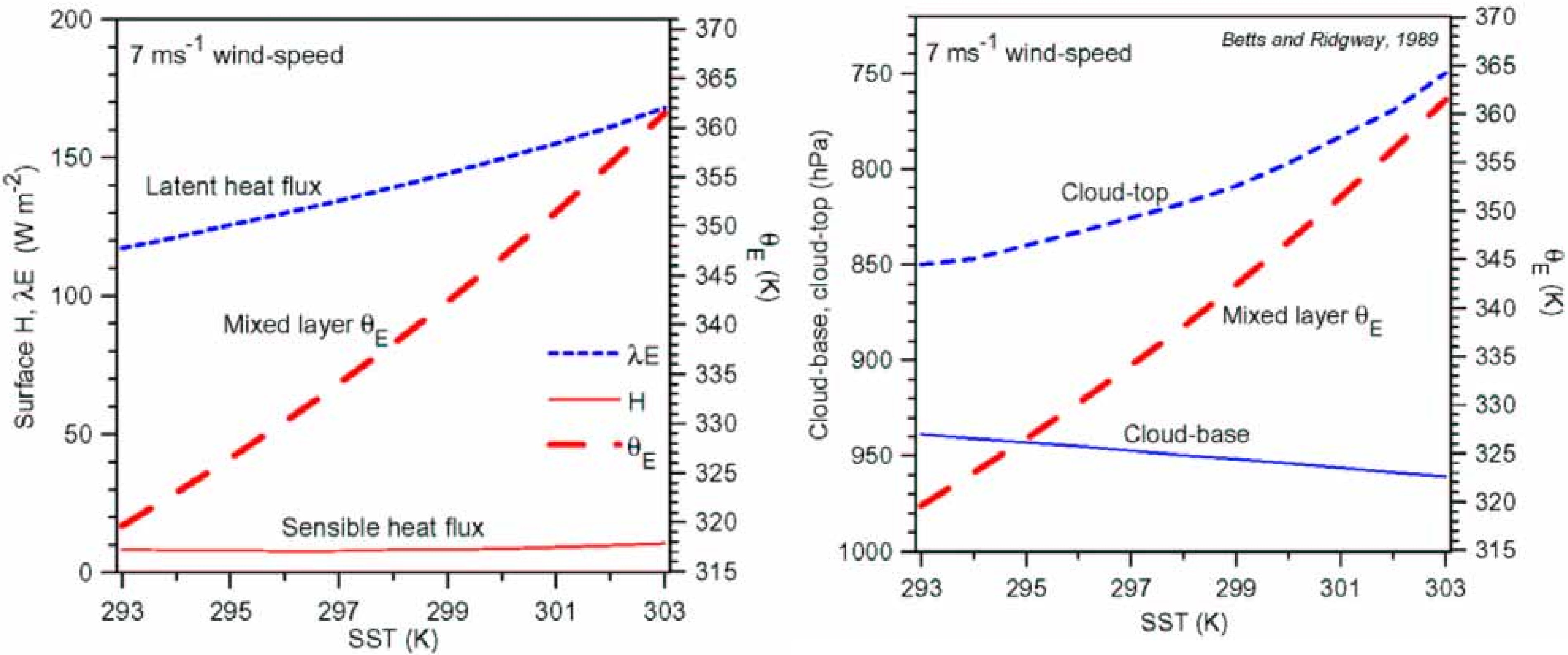
Where N is the net incoming radiative flux at the surface [shortwave + longwave]

This is the long-time-scale equilibrium that controls SST. [The big terms are the shortwave heating and the evaporation, but the downward longwave flux depends on water vapor ‘greenhouse effect’]

Solve the coupled system using 1 and 2 [plus 3, 4].

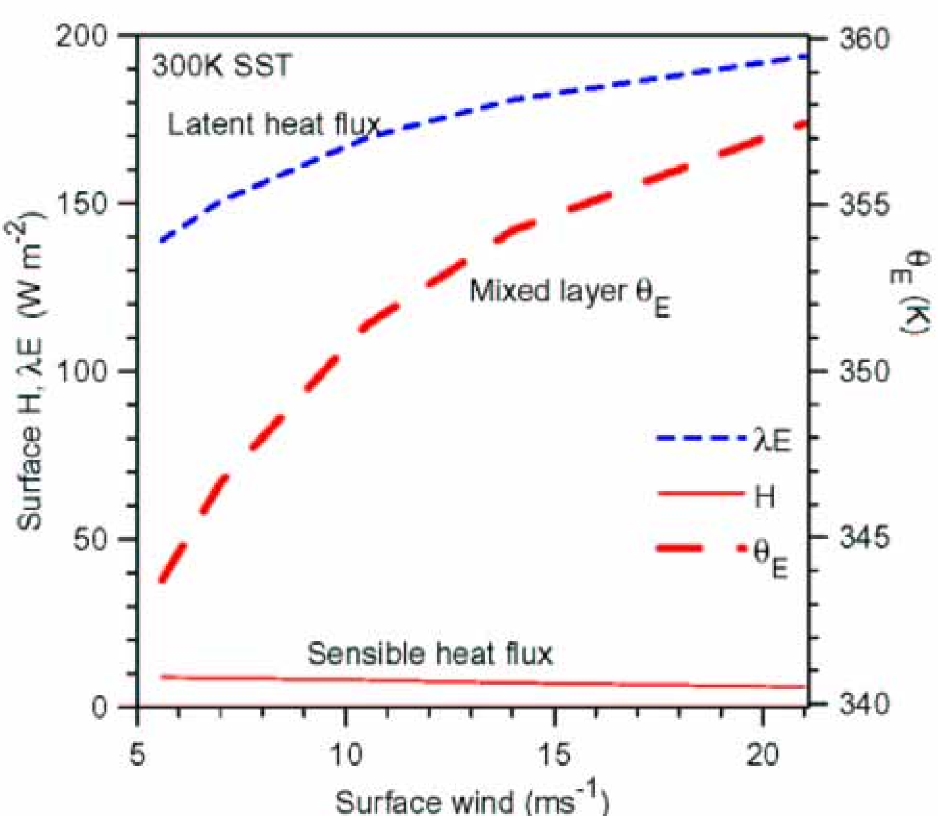
[Betts and Ridgway, 1989]

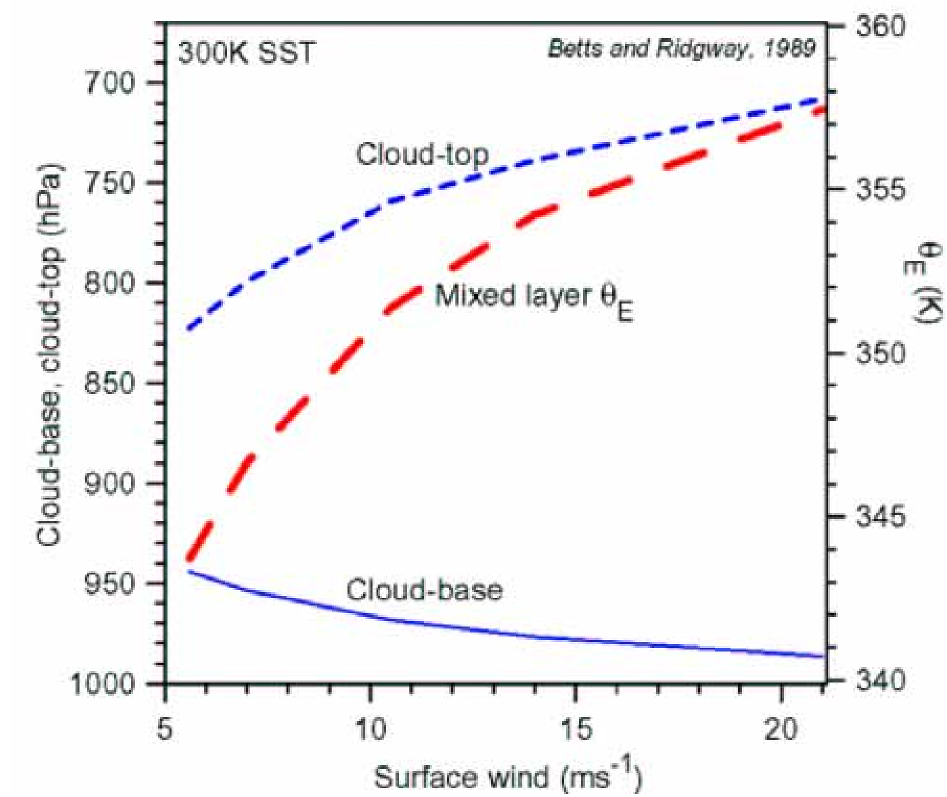
Vary SST with fixed wind-speed



- Evaporation increases with SST *Uses 1,2,3*
- θ_E increases with SST [and cloud-base descends a little]

Vary wind-speed with fixed SST; ω_T





- Evaporation increases with wind

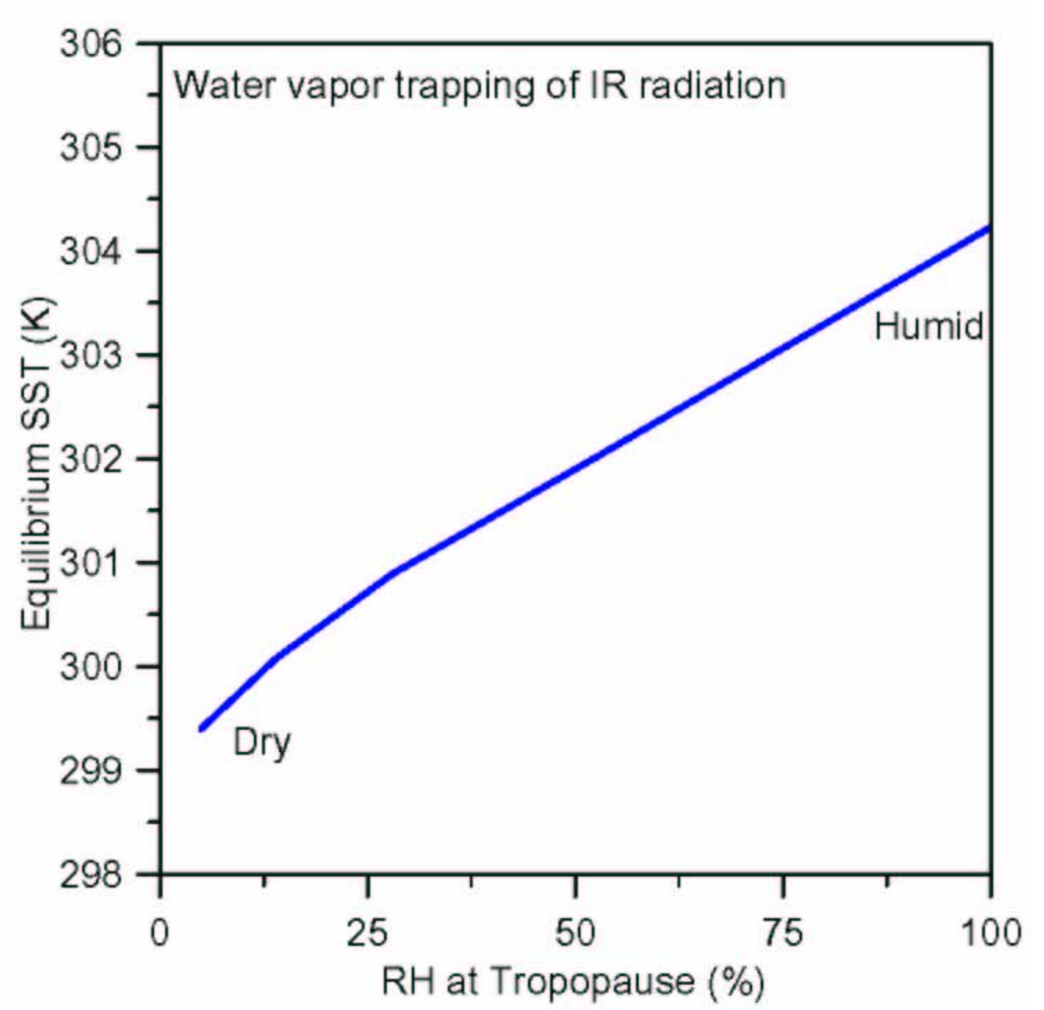
Uses 1,2

- θ_E increases as cloud-base descends, moving towards saturation at SST

SST equilibrium sensitive to LW

- Humid upper troposphere and equilibrium SST increases [greenhouse]

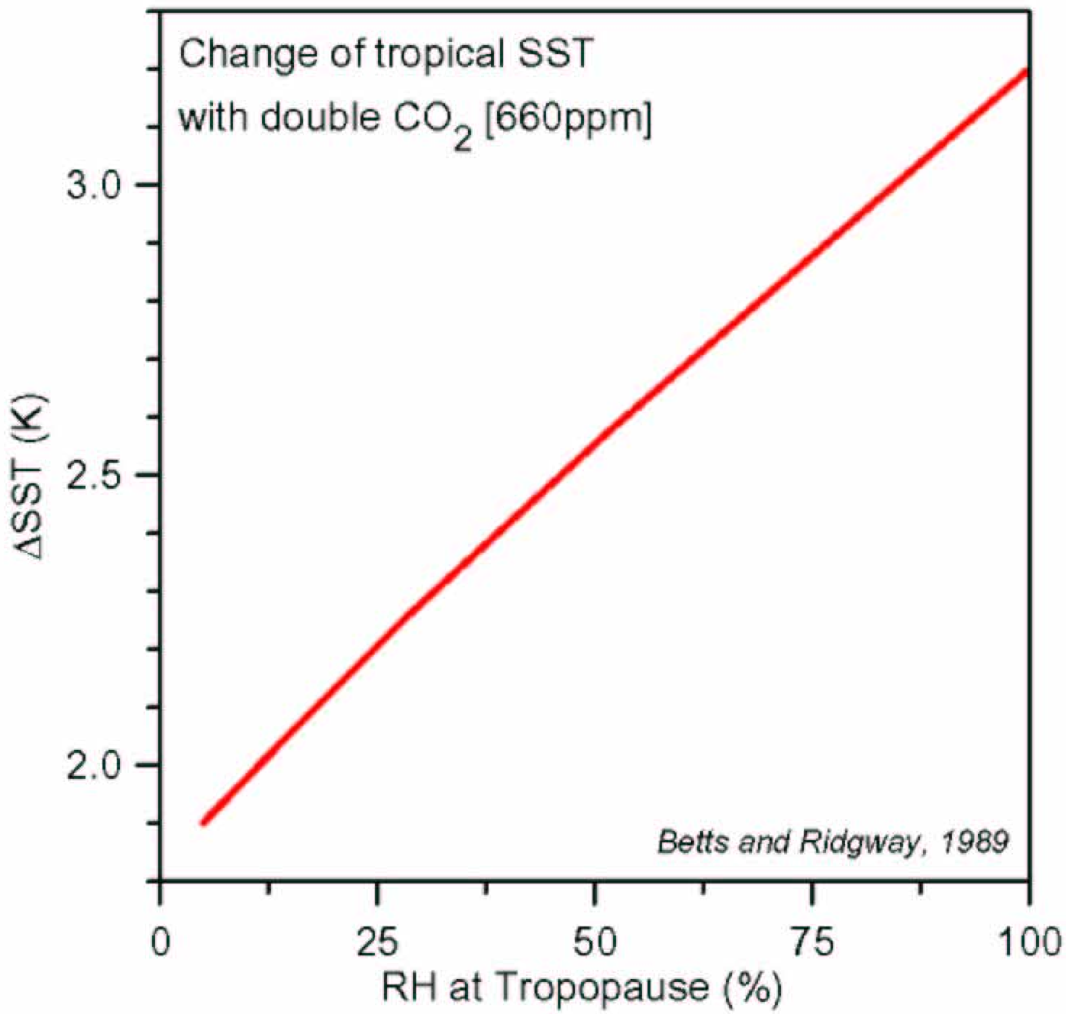
Uses 1,2,3,4



SST sensitive to CO_2

- Tropical climate sensitivity approx 2K for doubling of CO_2
- Sensitivity increases if upper troposphere moist

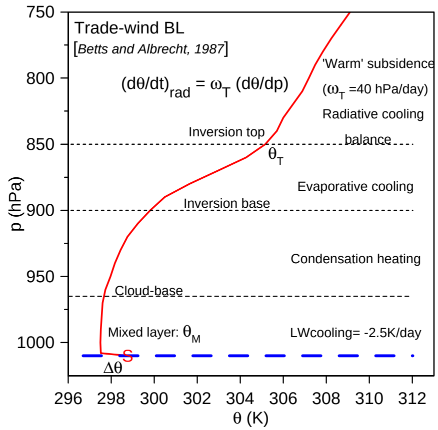
Uses 1,2,3,4

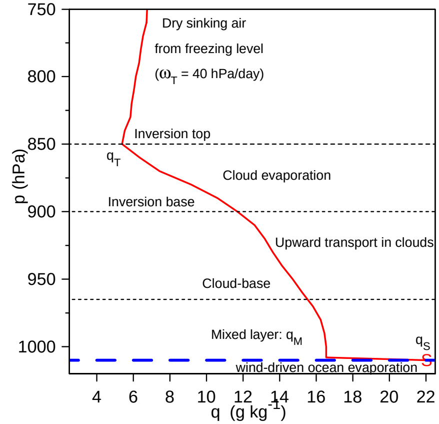


How does ocean BL and land differ?

Radiative cooling
SH, subsidence

Evaporation and
subsidence





Stays a little cooler than ocean and sub-saturated:
surface wind and subsidence control evaporation
[ocean store suns heat; diurnal cycle small]

LAND: what are the essential differences??

Sun heats surface and drives large diurnal cycle; daytime unstable;
cools radiatively at night; at night stable BL
Surface not saturated.. Except inside leaves.
Sun drives evaporation through photosynthesis
[coupled to CO_2 uptake]
Subsidence of dry air still plays key role, averaged over 24hrs.

Need to understand mean state and diurnal cycle

Coupling of CO_2 and water vapor through the BL

BOREAS Northern Study area [Thompson, Manitoba]

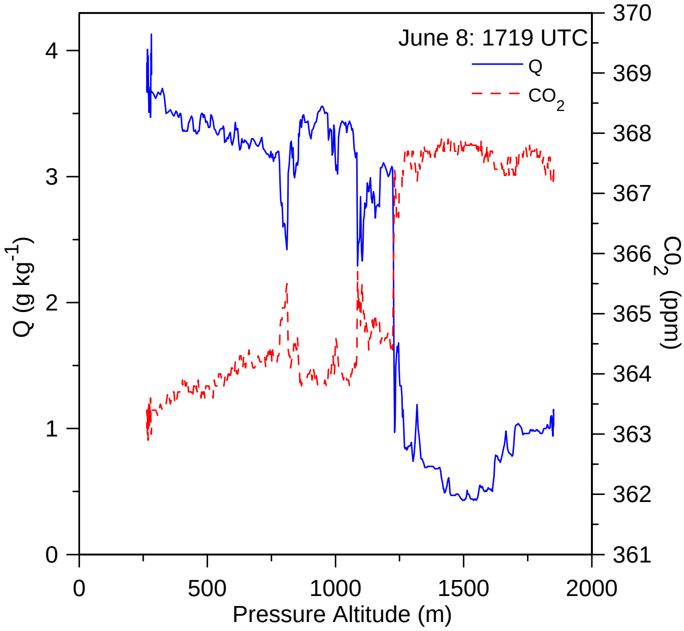


Figure 1 Coupling of CO2 and water vapor profiles of June 8 at 1719 UTC (LST=UTC-6h)

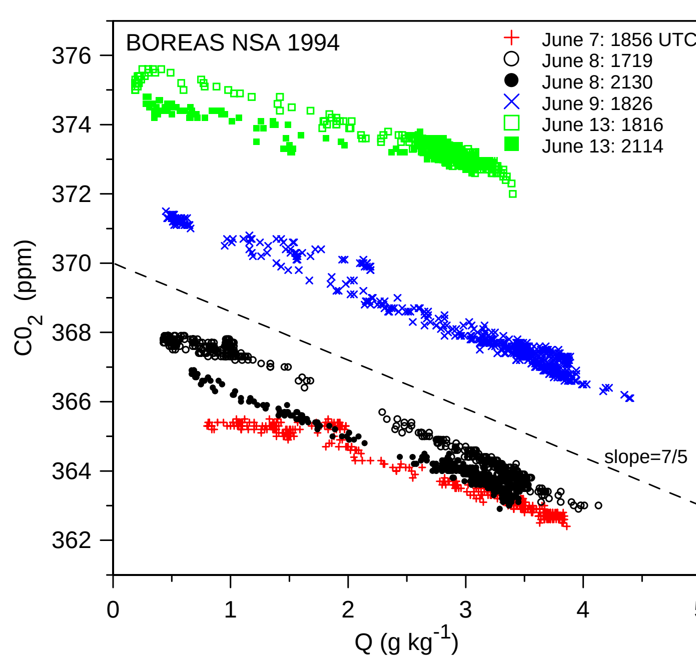


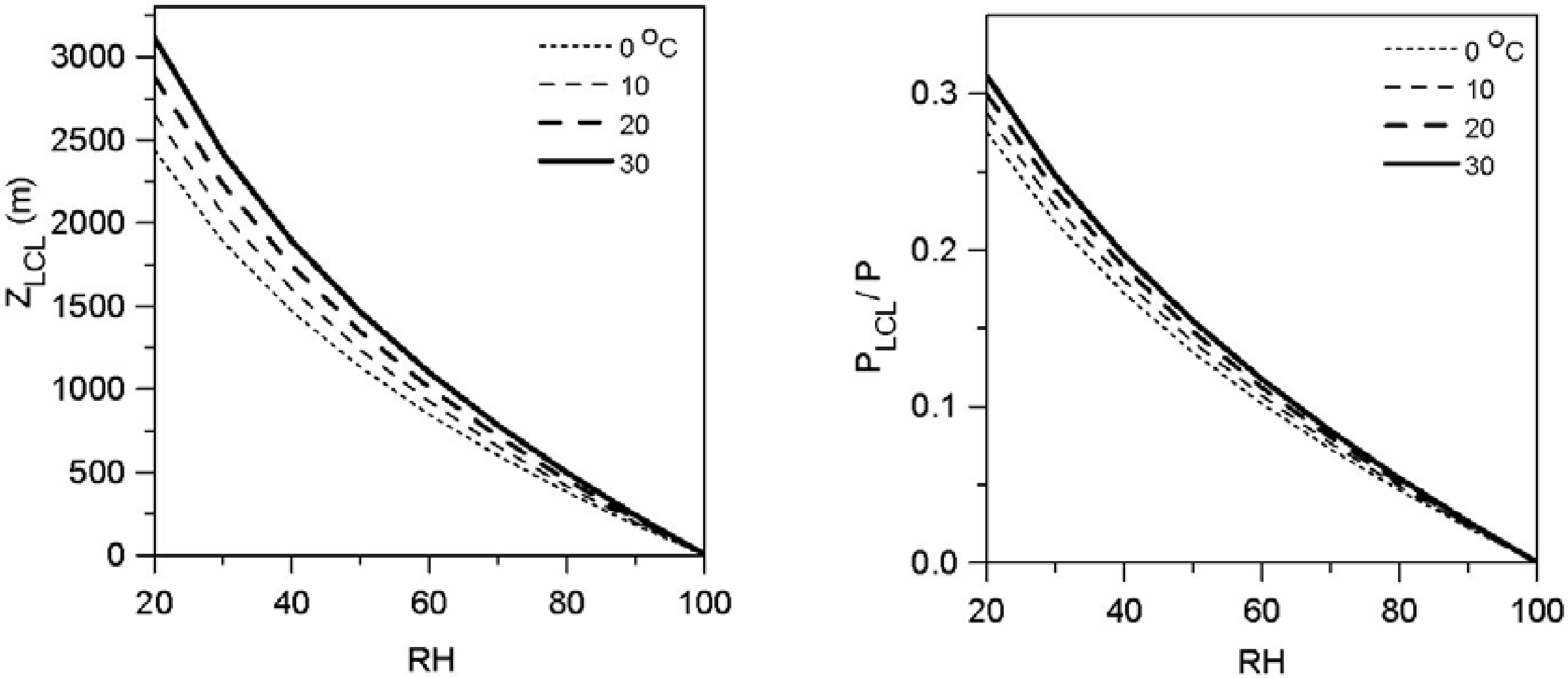
Figure 2. Profiles through the mixed layer on four days in June, showing tight coupling between water vapor and CO_2 structure. Illustrative slope of 7 ppm CO_2 to 5 g kg^{-1} is shown.

Daily mean fluxes give model ‘equilibrium climate’ state

- Map model climate state and links between processes using daily means
- Think of seasonal cycle as transition between daily mean states

  + synoptic noise

SMI ⟶ R_{veg} ⟶ RH ⟶ LCL ⟶ LCC



- RH gives LCL [largely independent of T]
- Saturation pressure conserved in adiabatic motion
- Think of RH linked to availability of water

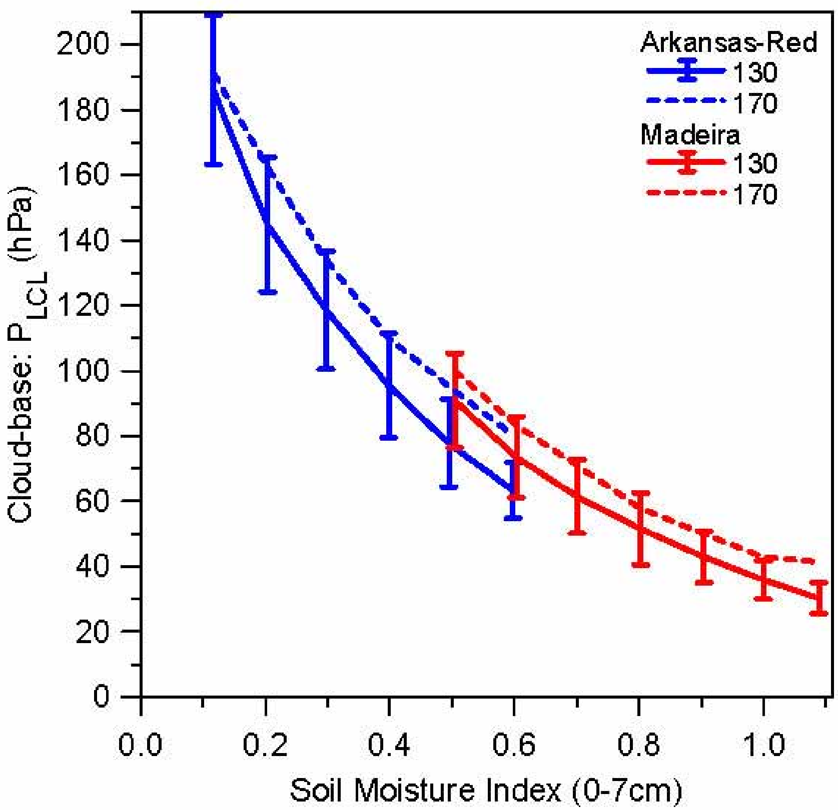
What controls daily mean RH anyway?

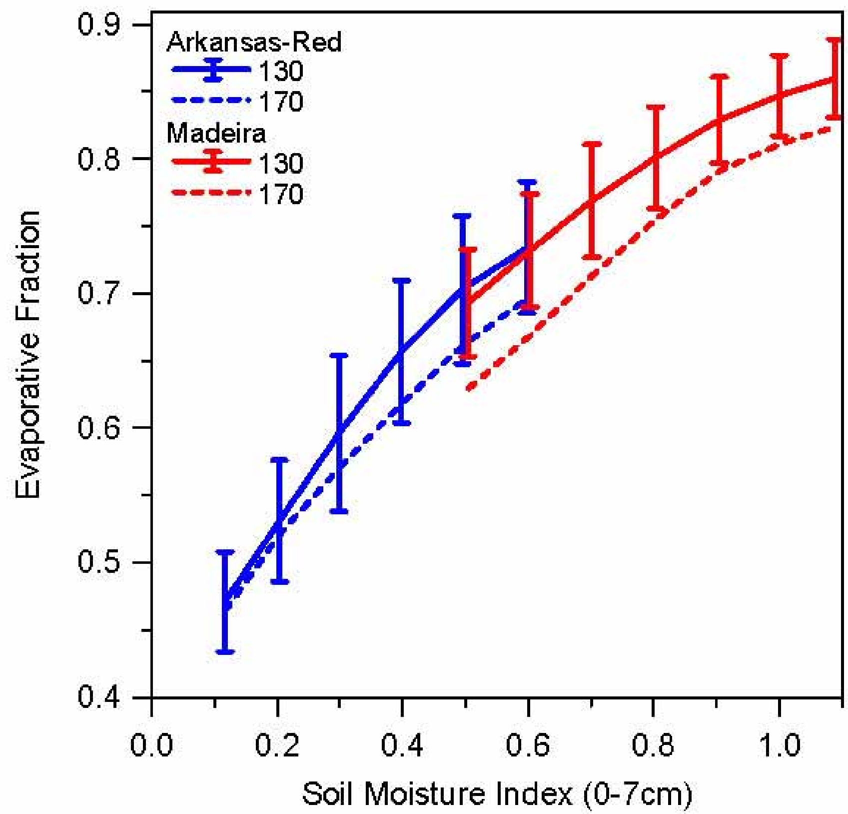
- RH is balance of subsidence velocity and surface conductance
- Subsidence is radiatively driven [40 hPa/day]
  + atmospheric dynamics
- Surface conductance

$$G_s = G_a G_{veg} / (G_a + G_{veg}) \qquad [G_{veg} = 1/R_{veg}]$$

[30 hPa/day for $G_a = 10^{-2}$; $G_{veg} = 5.10^{-3}$ m/s]

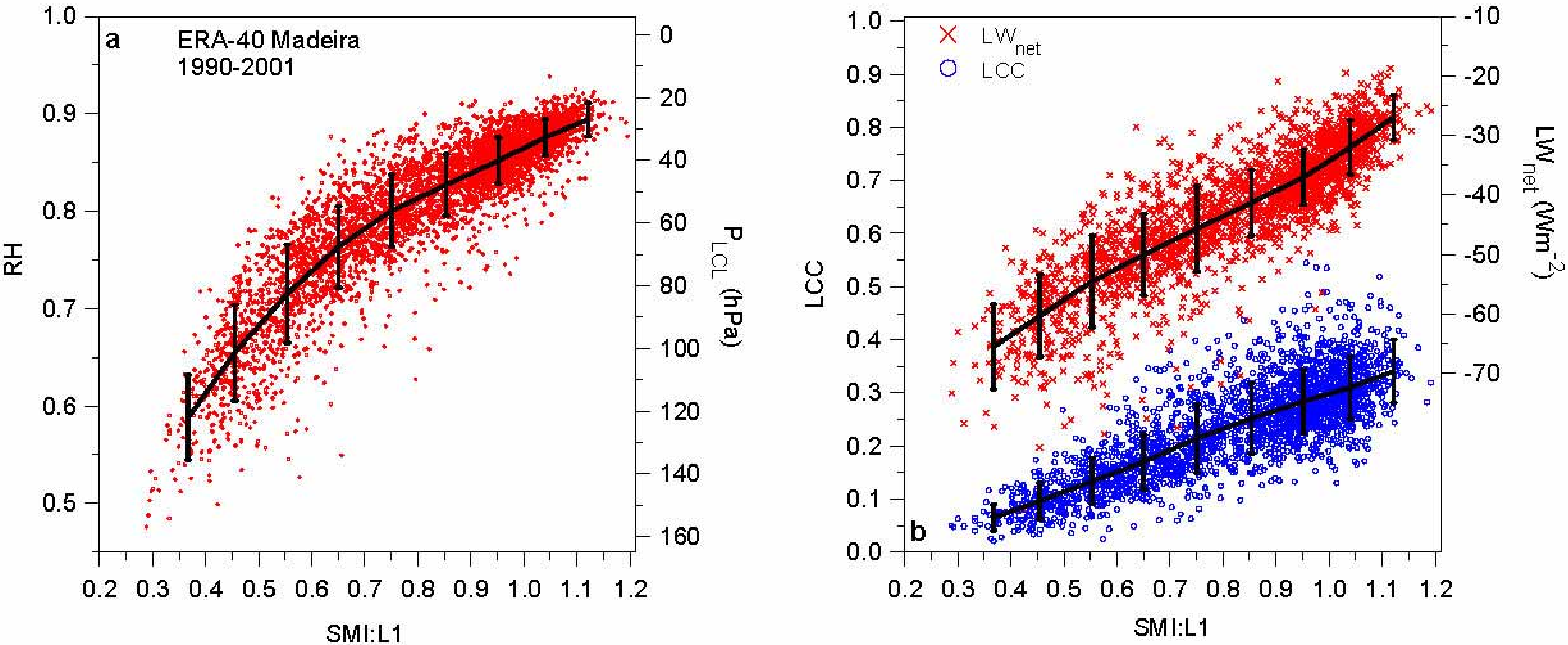
ERA40: soil moisture → LCL and EF





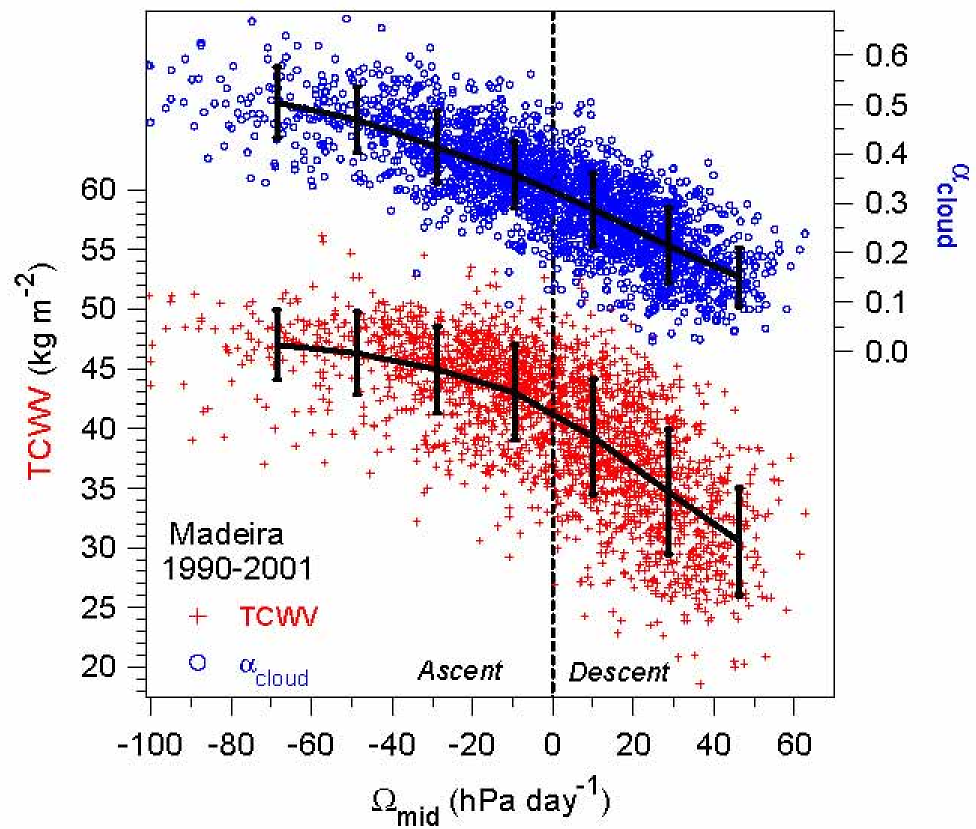
- River basin daily means
- Binned by soil moisture and R_{net}

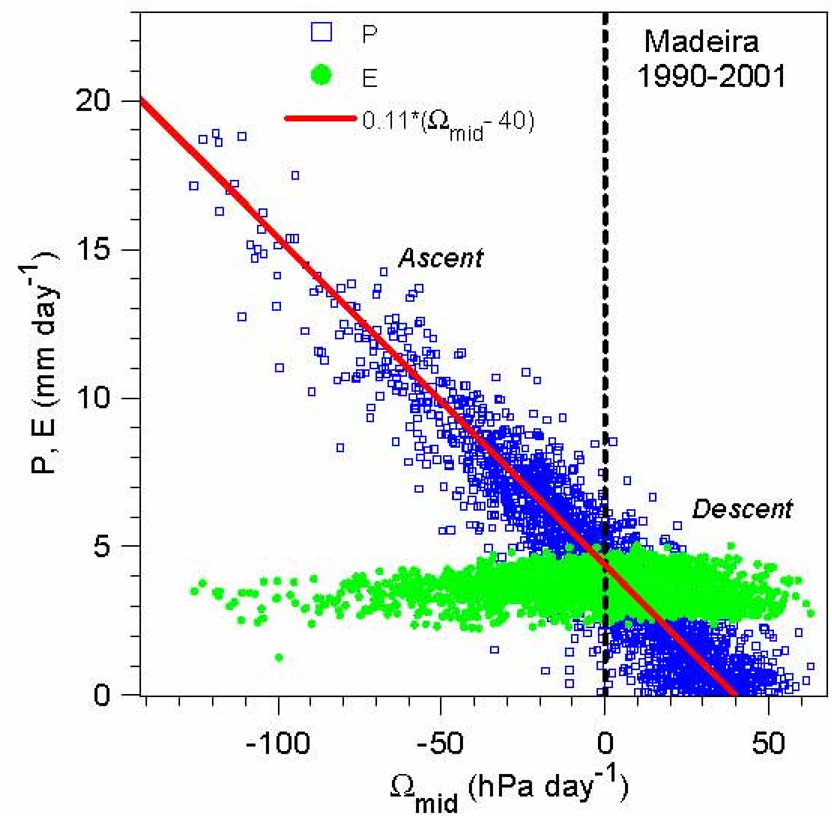
ERA40: Surface 'control'



- Madeira river, SW Amazon
- Soil water ⟶ LCL, LCC and LW_{net}

ERA-40 dynamic link (mid-level omega)





- $\Omega_{mid} \rightarrow$ Cloud albedo, TCWV and Precipitation

Take away these ideas

Ocean equilibrium: balance of radiative cooling, subsidence and surface fluxes

giving a typical tradewind BL with cloud-base 50hPa above surface and a 150hPa deep shallow cumulus layer....

[Solar heating absorbed in deep ocean mixed layer]

Fluxes and BL θ_E go up with SST and wind-speed

Land diurnal cycle driven by solar heating, but *equilibrium* similar to ocean, except a drier mean state because additional ‘vegetative’ resistance to evaporation at surface.

CO_2 and water vapor coupled in BL over vegetation.