Parameterizing large-scale dynamics using the weak temperature gradient approximation

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In the tropics, our picture of the dynamics should start from a single column

$\partial_t\theta$ + wS = Q_c + Q_R,
$\partial_t q$ + w$\partial_z q$ = Q_q

with S / $\partial_z\theta$, Q_q=(e-c)+ ∂_z w'q'

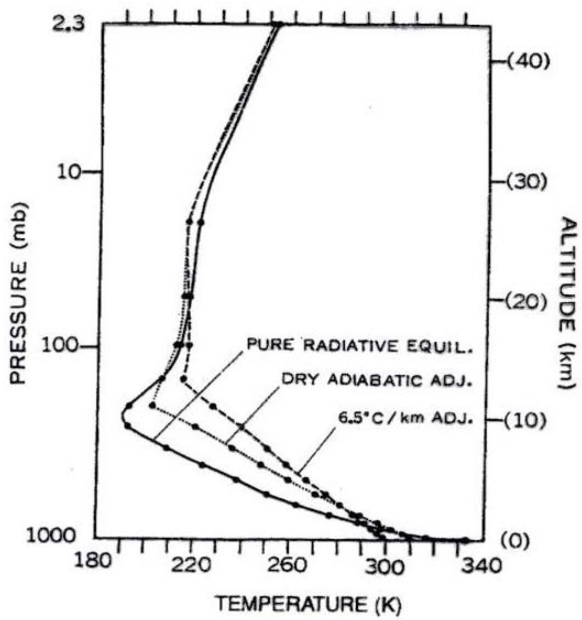
where we have neglected horizontal advection terms u_h¢$r_h(\theta$,q)

Single-column models represent this system with parameterized physics. Cloud-resolving models on small domains, if we focus only on aggregate statistics, do also.

There are a number of ways of using such models: e.g., to study radiative-convective equilibrium states, or to force them to simulate observations.

I propose a way which is idealized, like RCE, but goes a step further by incorporating a representation of large-scale dynamics. I'll argue that this is a useful way of characterizing a given model physics.

The simplest case is "Radiative-convective equilibrium", in which Q_R=-Q_c, w=0; no dynamics.
Example 1: global mean climate theory.



Manabe and Strickler 1964

Example 2: the idea of RCE is built into our dry theories of tropical atmosphere dynamics. Axisymmetric models of the Hadley circulation (Schneider 1977; Schneider and Lindzen 1977; Held and Hou 1980; etc.) formulate the heating as relaxation back to a prescribed RCE temperature profile:

$$\partial_t\theta + \partial_y(v\theta) = [\Theta(y) - \theta]/\tau$$

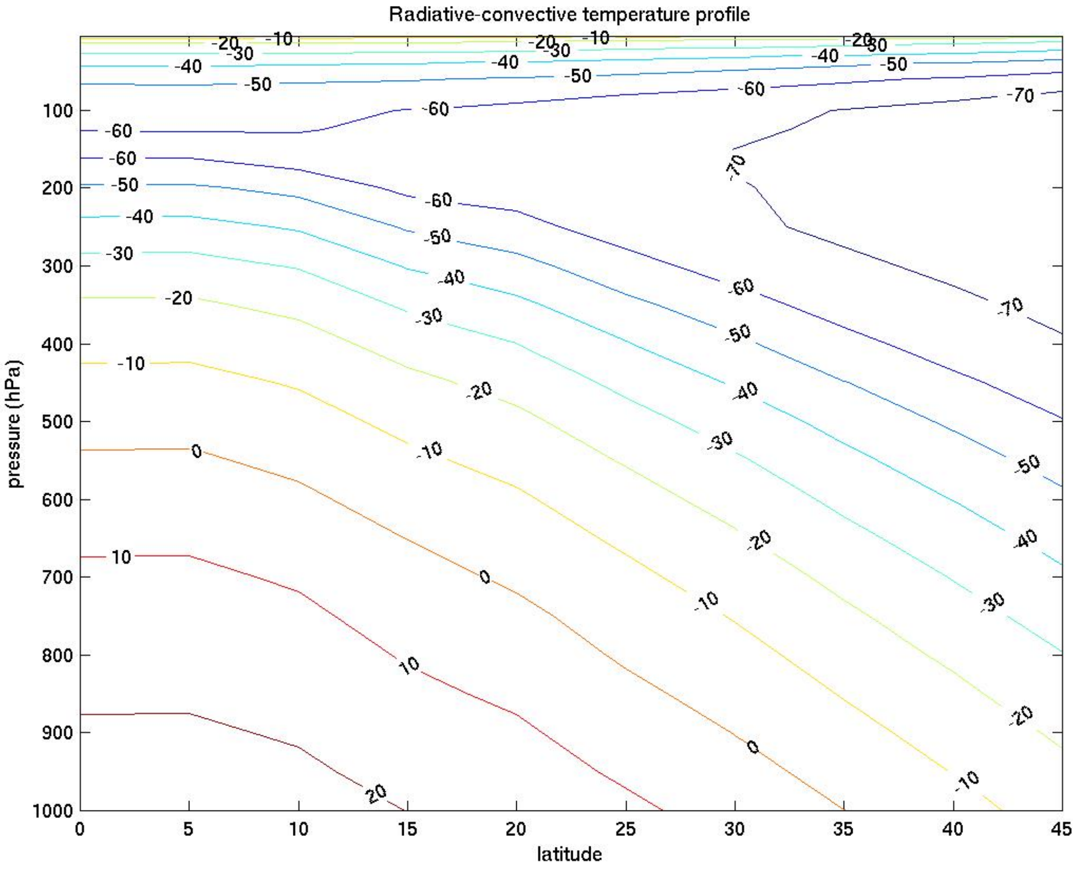
$\Theta(y)$ is set by latitudinally varying insolation, and implicitly depends on moist physics.

RCE Calculations with Emanuel single-column model

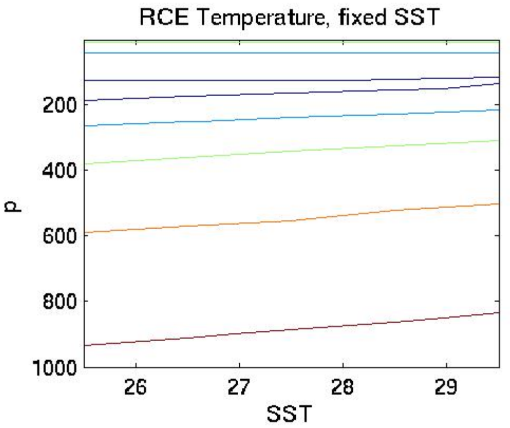
(Renno et al. 1994, *JGR*, **99**, 14429-14441)

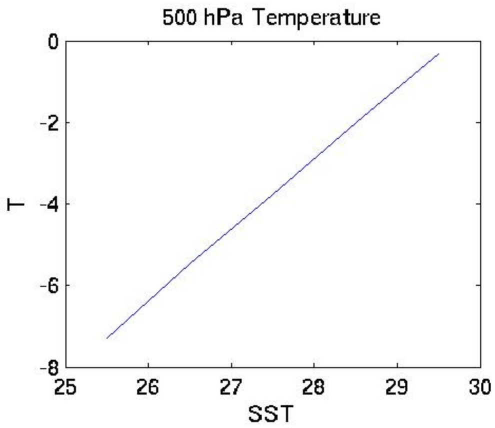
- Convective and (clear-sky) radiation parameterizations
- Slab ocean mixed layer
- Annual average insolation (function of latitude)
- CO2, Ozone at reasonable values
- Surface wind speed=7 m/s
- Surface albedo tuned to give reasonable climate near equator

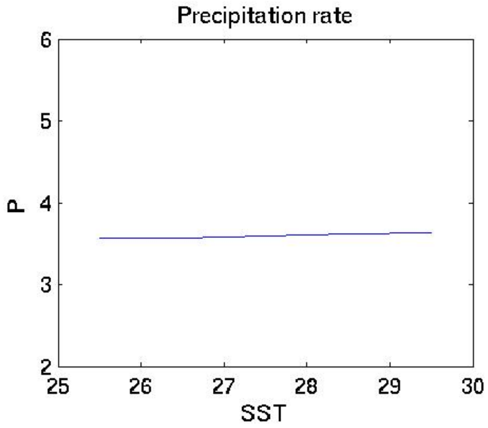
RCE temperature calculated with the Emanuel single-column model



In RCE, P=E, and atmos. temperature increases ~linearly with SST (slope>1)

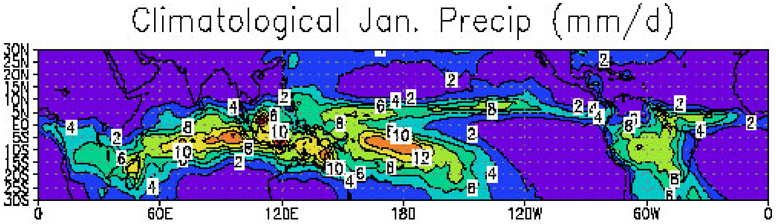


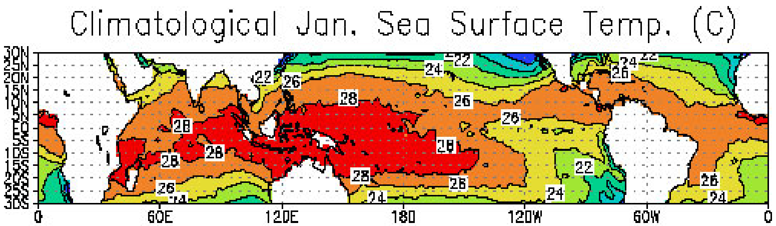


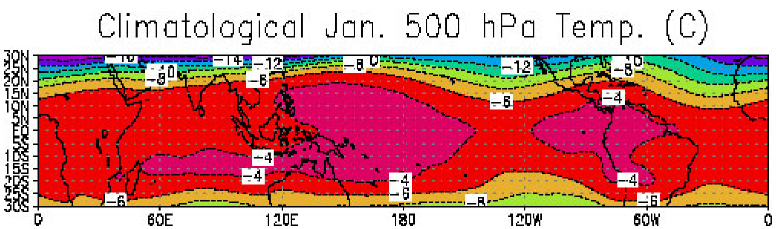


These calculations done with fixed SST

The real tropical atmosphere doesn’t do that







In the tropics, the dominant balance in the θ equation is not

$Q_c + Q_R$ ¼ 0, but instead

$Q_c + Q_R$ ¼ wS,

with θ itself held nearly constant at const. z by large-scale adjustment (cf. Fovell’s talk yesterday).

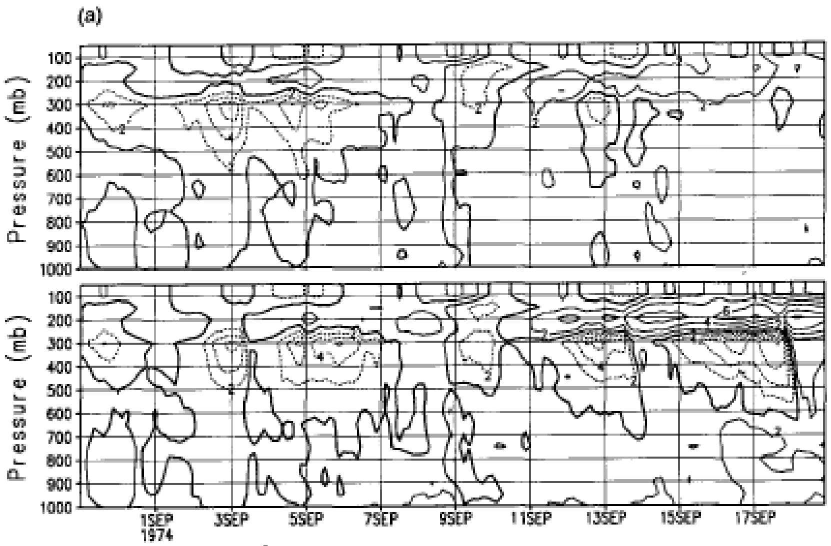
RCE may be a useful theoretical construct, but it is not a good approximation to the state of the atmosphere in general.

To get a little more dynamic range out of an SCM (or CRM) can also impose a nonzero large-scale w. RCE simulations with unrealistically large values of radiative cooling are really just doing this under another name.

Still the model must satisfy wS ¼ Q_c + $Q_{R,}$
S and Q_R can't change that much, so Q_c (i.e., precip) will.

Application of SCMs to testing convection schemes: specify large-scale w from observations

Temperature error with (above) & without (below) downdrafts



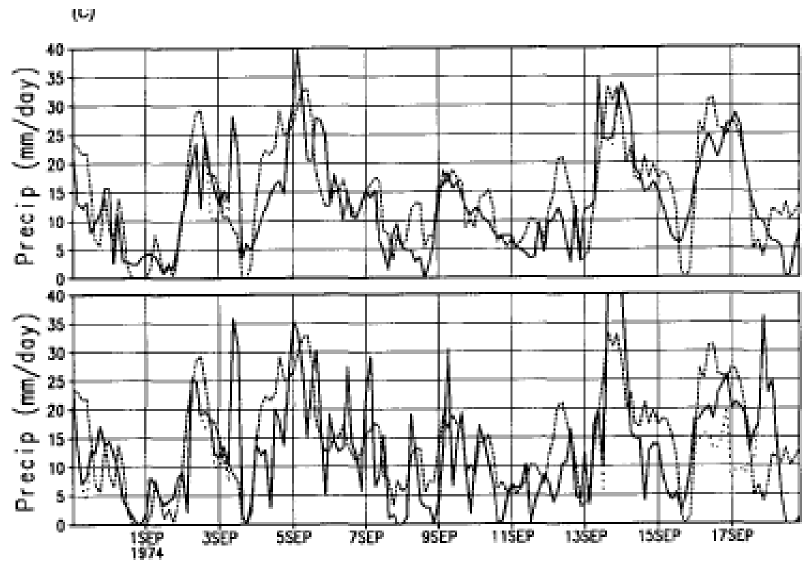
time →

Sud and Walker 1993, *Mon. Wea. Rev.* **121**, 3019-3039
Adding downdrafts to a scheme, see the difference. Forced by time series of advection terms from GATE field experiment. Compare temperature & humidity to obs.

Precipitation is *not* a meaningful variable to compare to obs!

$$\frac{\partial s}{\partial t} = -\mathbf{u}\cdot\nabla s - \omega\frac{\partial s}{\partial p} + Q_c + Q_R$$

$$\frac{\partial q}{\partial t} = -\mathbf{u}\cdot\nabla q - \omega\frac{\partial q}{\partial p} + Q_q$$

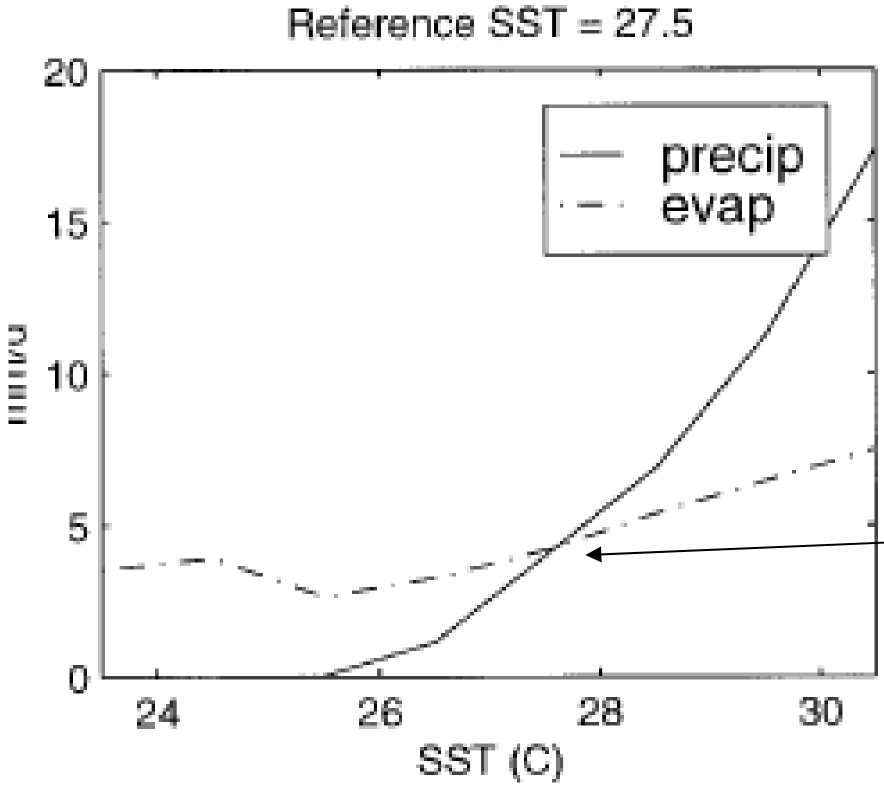


Dominant balance in s
equation is $\omega\partial s/\partial p$ ¼ $Q_c + Q_R$,
variations Q_c'À Q_R', and $\partial s/\partial p$ ¼ const,
so Q_c»ω, and P»sQ_c dp. Thus if ω given,
P is too, almost independently of model physics
(which is what we're trying to test)

Weak Temperature Gradient Methodology for an SCM

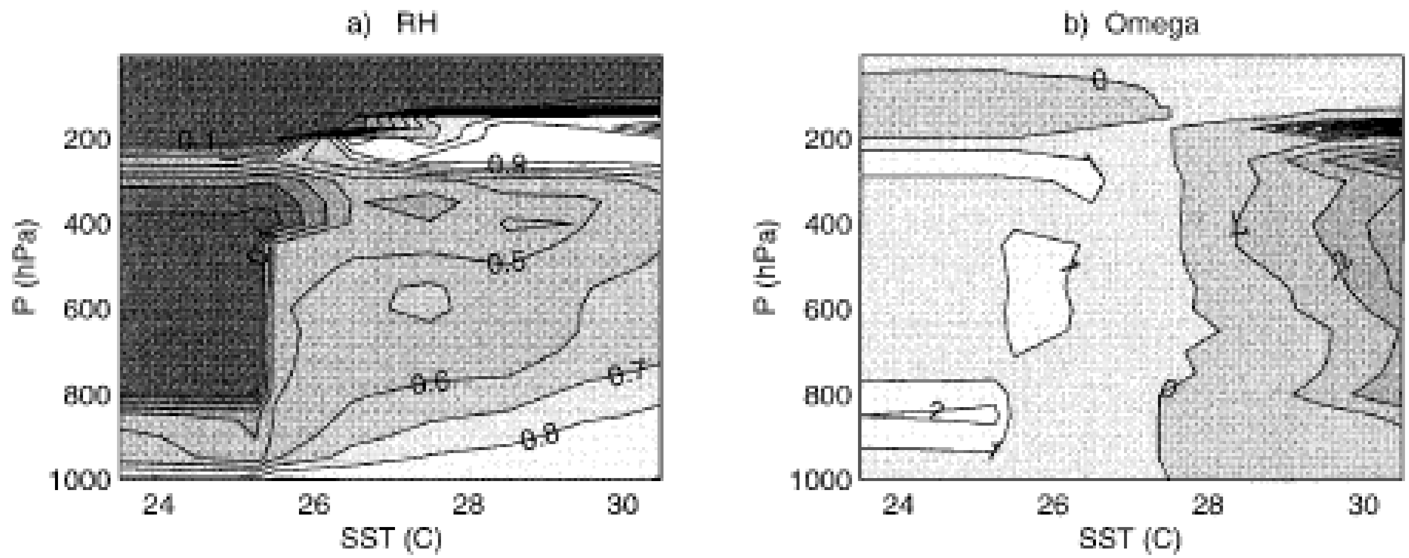
- Choose a tropospheric θ profile. It can be obtained by running the SCM to RCE. Then change the model:
- Above a nominal PBL, keep θ fixed (don't time-step it). In the PBL, allow θ to change as normal.
- Compute large-scale w in the free troposphere from $wS=Q$, with Q determined by the model physics.
- Use the w so determined in the moisture equation. This implies an interactively computed moisture convergence, so $P \neq E$.
- Need w in the PBL; interpolate from the value at PBL top down to zero at the surface.
- Now change the SST (or some other input parameter), and see how precip, w, q, etc. change.
- Horizontal moisture advection can be added in various ways if desired.

Precipitation in WTG simulations with Emanuel’s model

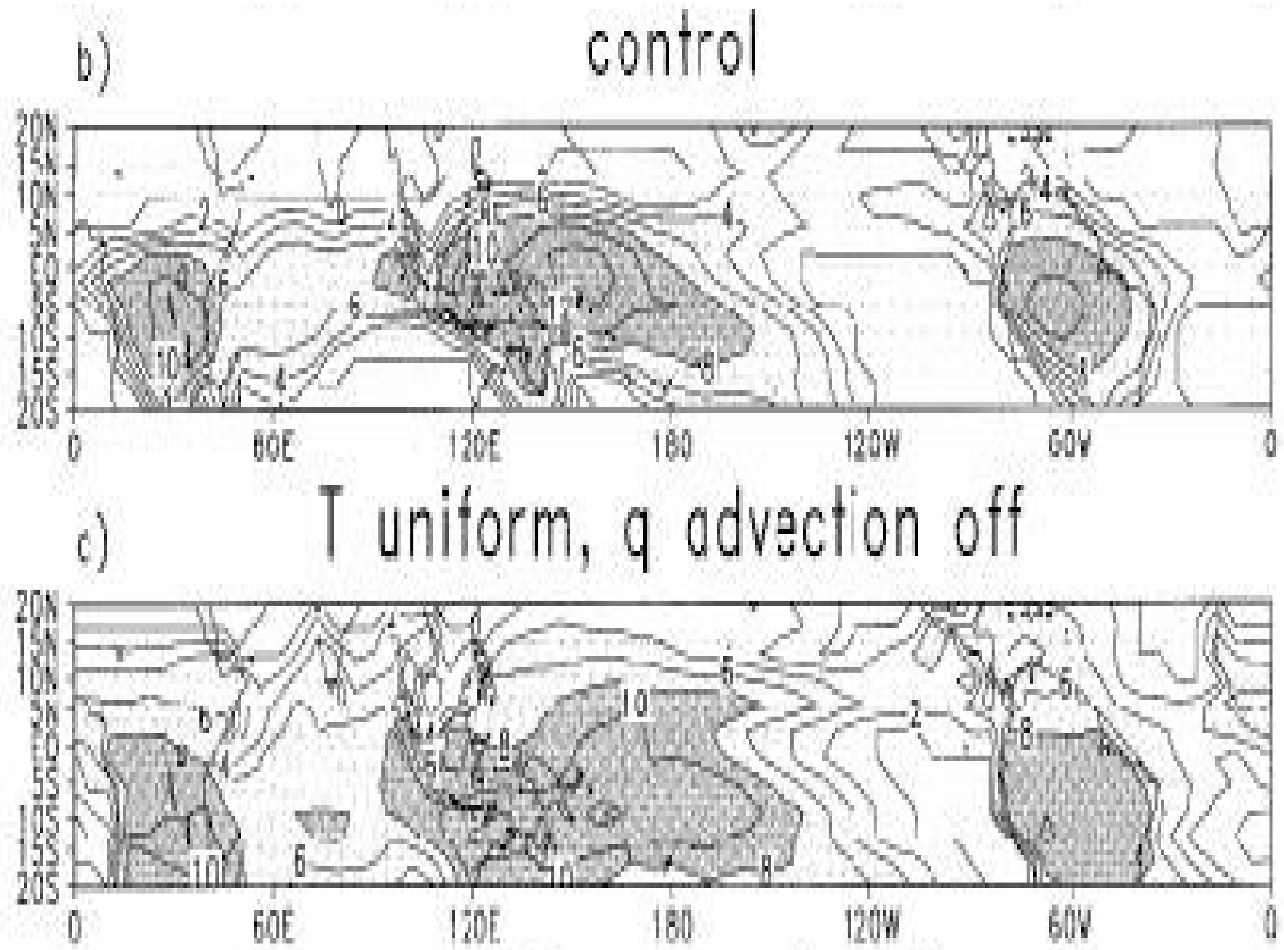


The RCE solution is “stable”

Sobel and Bretherton 2000


a) RH
b) Omega
P (hPa)
200
400
600
800
1000
SST (C)
24
26
28
30
0.1
0.9
0.5
0.6
0.7
0.8
0
2

WTG simulations with the QTCM



Claim: the WTG simulation strategy forces the SCM physics in a way fundamentally similar to the way the quasi-steady component of large-scale dynamics does, e.g., in a GCM.

It cannot straightforwardly incorporate the influences of transients, esp. gravity waves. This makes it very difficult to do observational case studies with WTG.

The WTG SCM formulation, together with a physics package, forms a mean field theory for tropical precipitation. Tropical mean θ is determined by an average tropical SST. Local precip depends on the difference between local SST and that average. However it does not depend on SST gradients; there is no spatial scale. It is also fundamentally thermodynamic. These features distinguish it from, e.g., Lindzen and Nigam (1987).

The first theory of this type is that of Neelin and Held (1987), using the moist static energy budget:

$$P = P_{RCE} + (\Delta s/\Delta m)(E + Q_R)$$

Where Δs and Δm are the dry and gross moist stabilities (cf. Frierson’s lecture) and E is surface latent heat flux (neglecting sensible).

There are other ways of parameterizing large-scale dynamics. E.g. Mapes (2004); Bergman and Sardeshmukh (2004) propose:

$$dw/dt = \tau^{-1}(w-Q/S)$$

In the steady-state limit, this produces an apparent WTG balance, $wS=Q$. However, it does not constrain θ. The θ profile will adjust to the SST in the same way it does under RCE. This method is more suited to study of transients, rather than quasi-steady climate balances.

BS04 also developed a fancier parameterization of wave dynamics for an SCM.

WTG simulations with the Goddard Cumulus Ensemble Model
(W.-K. Tao & associates, NASA GSFC)

- 2D cloud resolving, 1km horiz. grid spacing
- Rain, snow, graupel
- Cloud ice and cloud water
- Cloud-Radiation interactions
- Anelastic, non-hydrostatic
- 512 km x 41 vertical levels; model top at 22km
- Periodic lateral boundary conditions
- Lower BC is ocean with uniform SST
- Domain-averaged horizontal wind field relaxed towards 5 m/s easterlies with zero shear with 1 hr relaxation timescale

Perez, Sobel, Gu, Shie, Tao, and Johnson, submitted to J. Atmos. Sci.
See also Raymond and Zeng (2005)

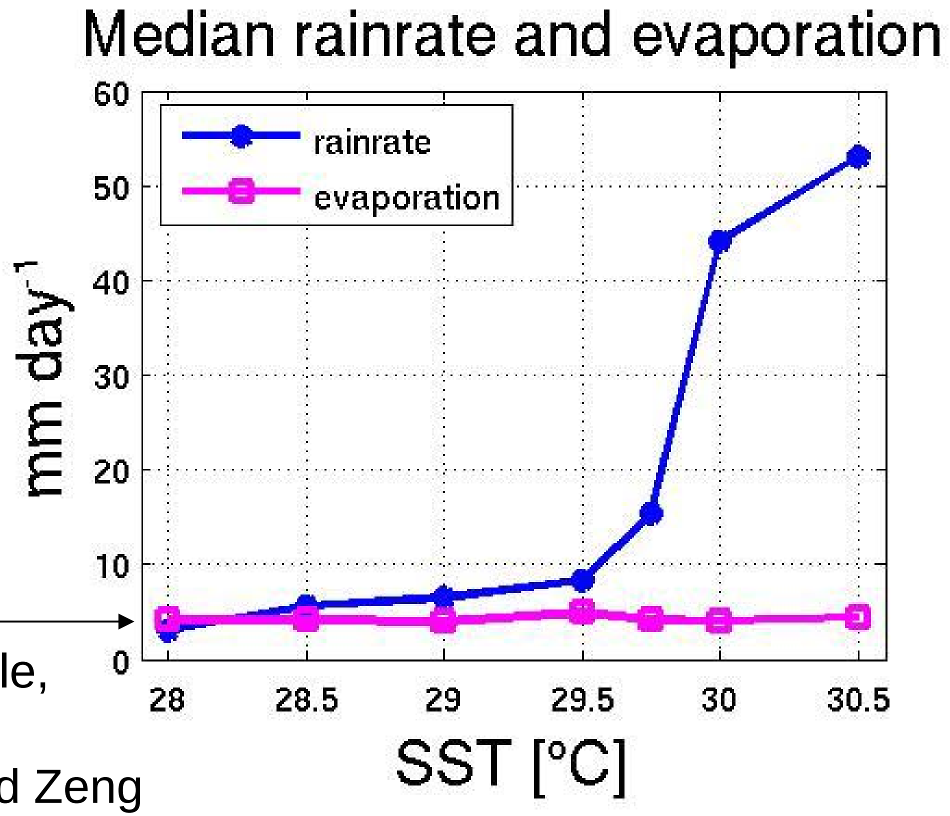
Procedure

- Apply relaxation on RHS of temp equation:

  $d\theta/dt = Q_c + Q_R + (\Theta - \langle\theta\rangle)/\tau$, where d/dt includes only advection by resolved flow in CRM; <T> is horiz. mean
- Then last term represents large scale advection:

  $w = (\Theta - \langle\theta\rangle)/(\tau S)$; in steady state then

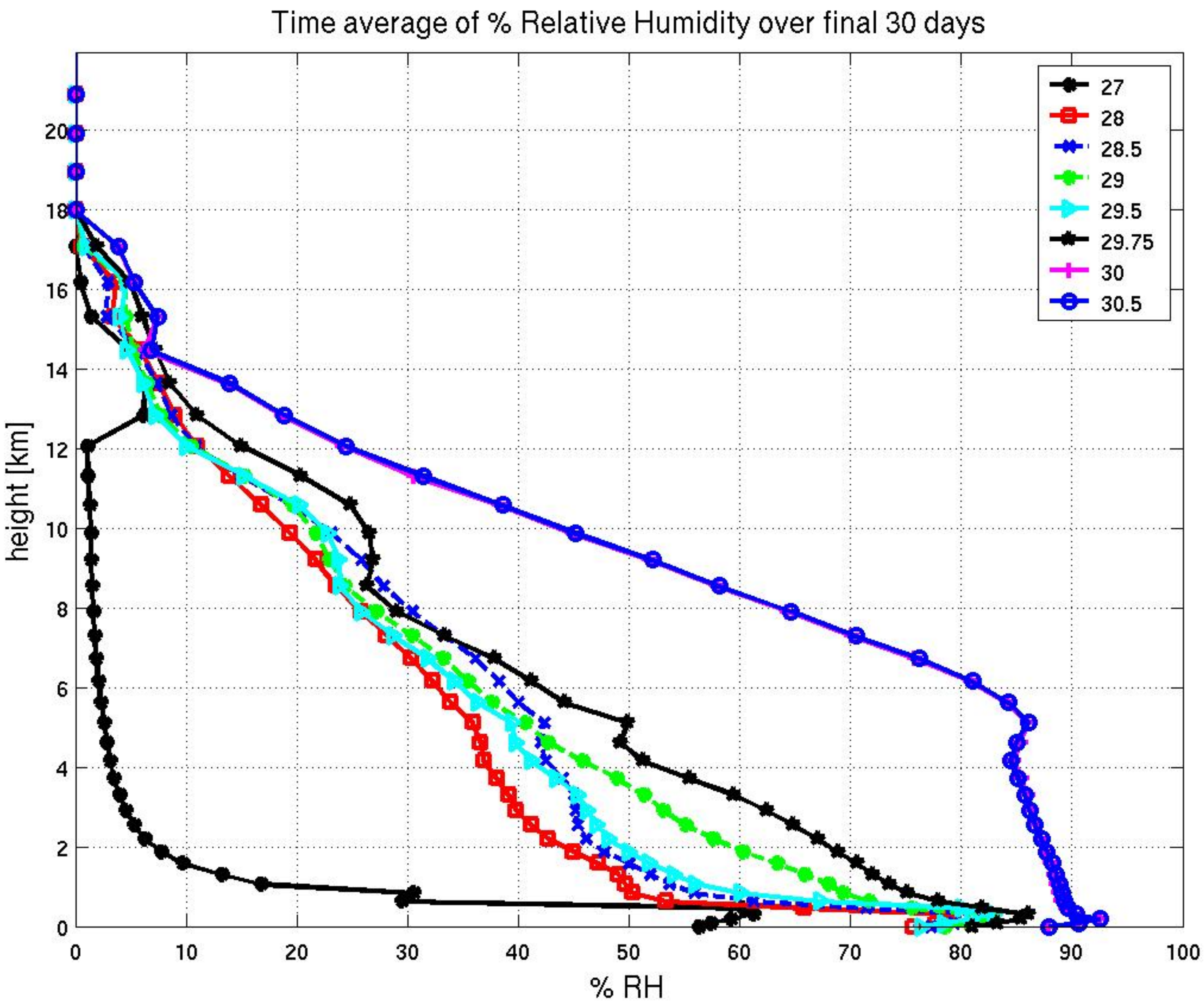
  $wS = Q_c + Q_R$, as appropriate
- Use resulting w to advect moisture, *determining moisture convergence, and thus domain-average precip, interactively*



RCE is still (almost) stable, in contrast to Raymond and Zeng (2005)

Relative Humidity

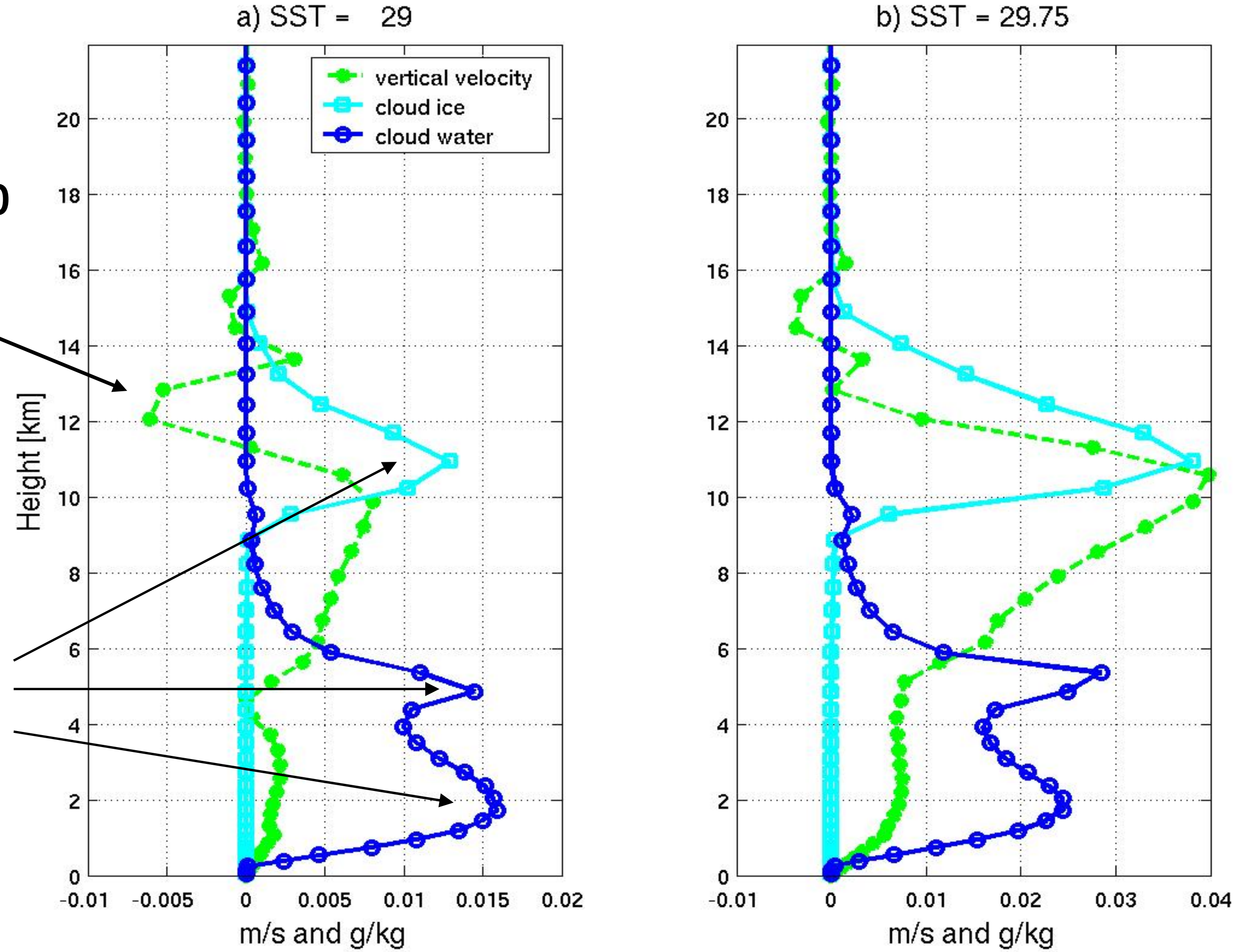
- **Two clusters**:
  1. Moist BL (RH>80%) then drier
  2. RH>80% until freezing level
- Bifurcation at SST ~ 29.75 C



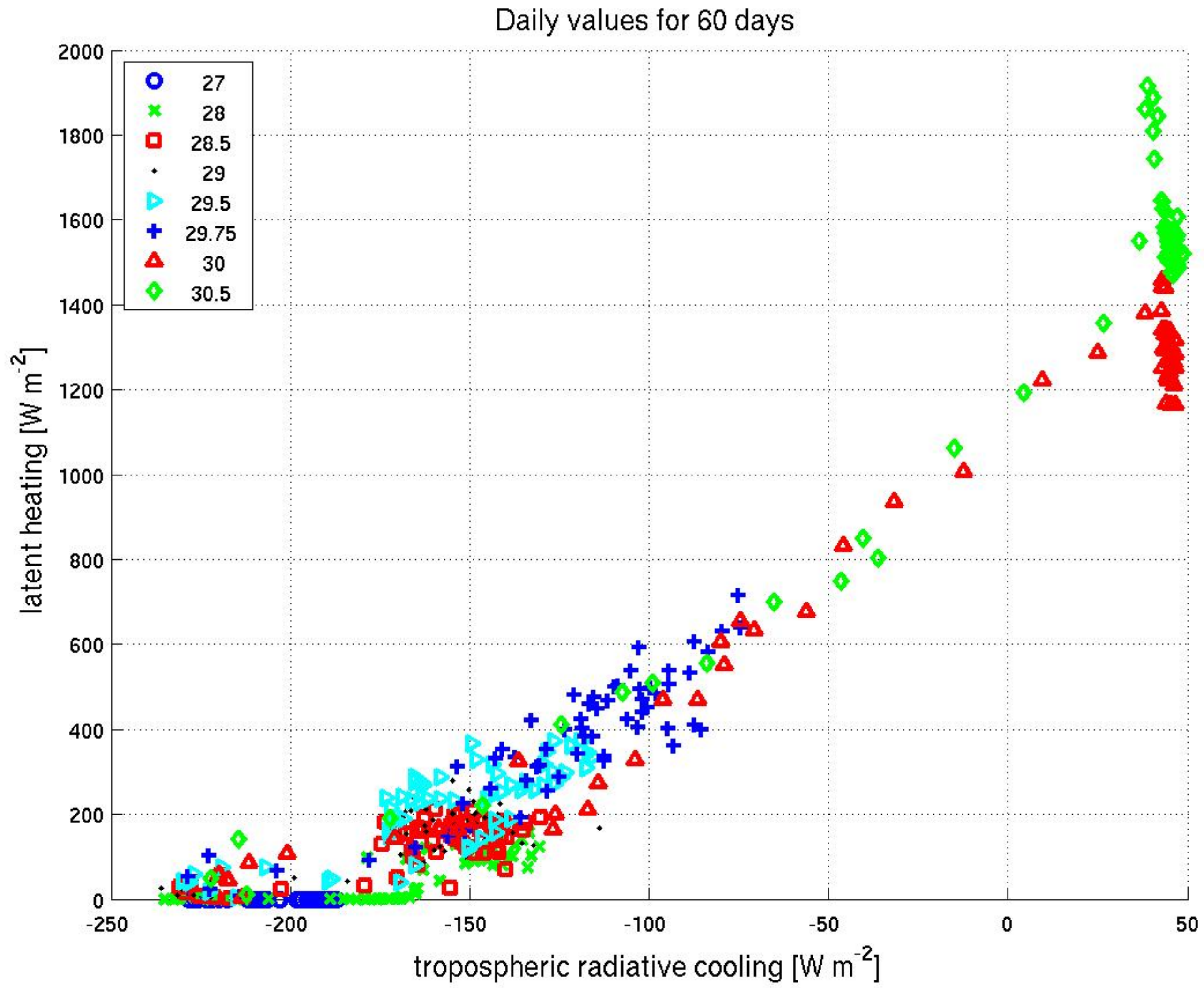
Clouds and WTG vertical velocity

Cloud-top radiative cooling

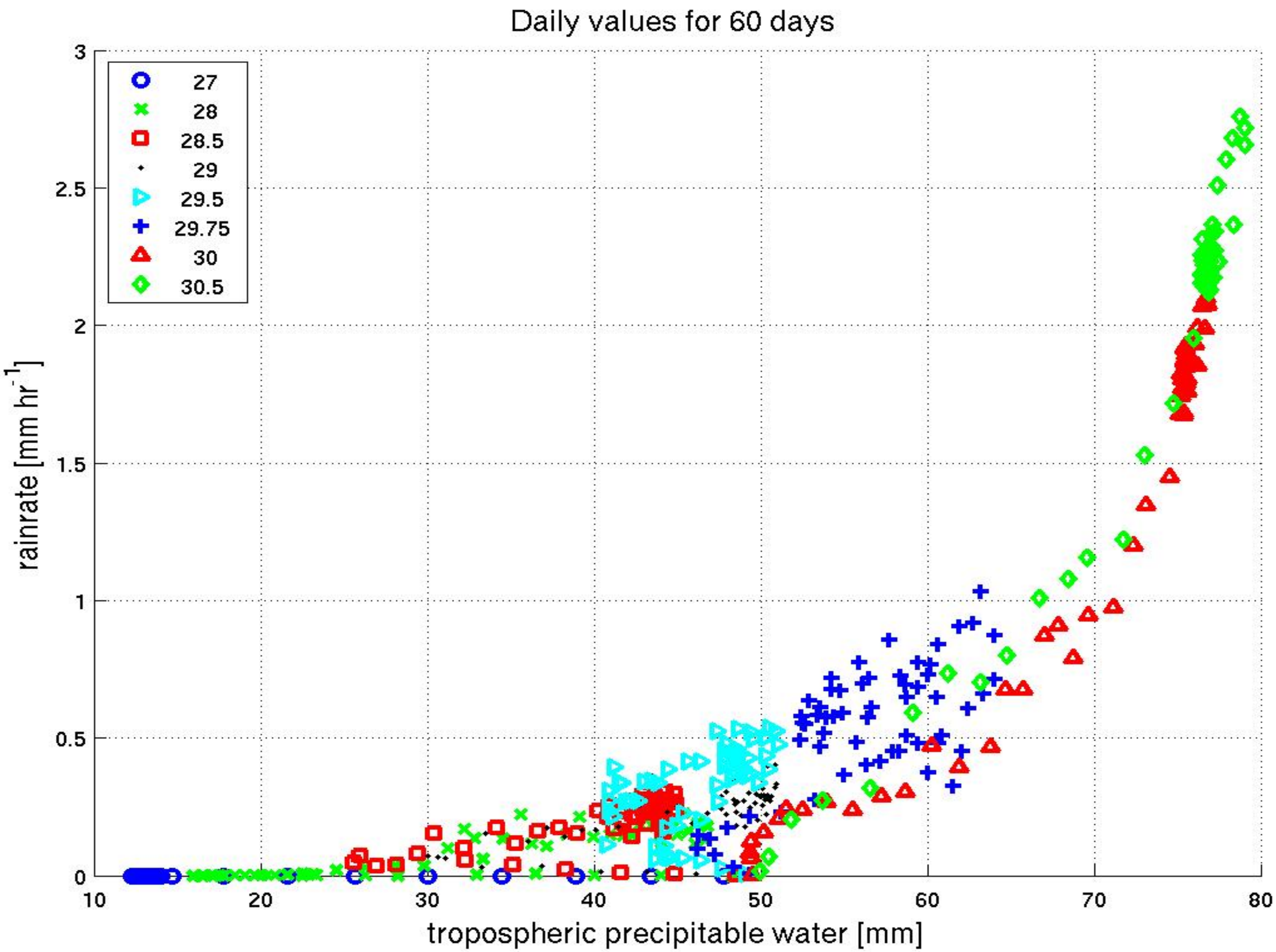
Trimodal cloud distribution (Johnson et al., 1999)



Tropospheric radiative cooling vs. latent heating due to precipitation

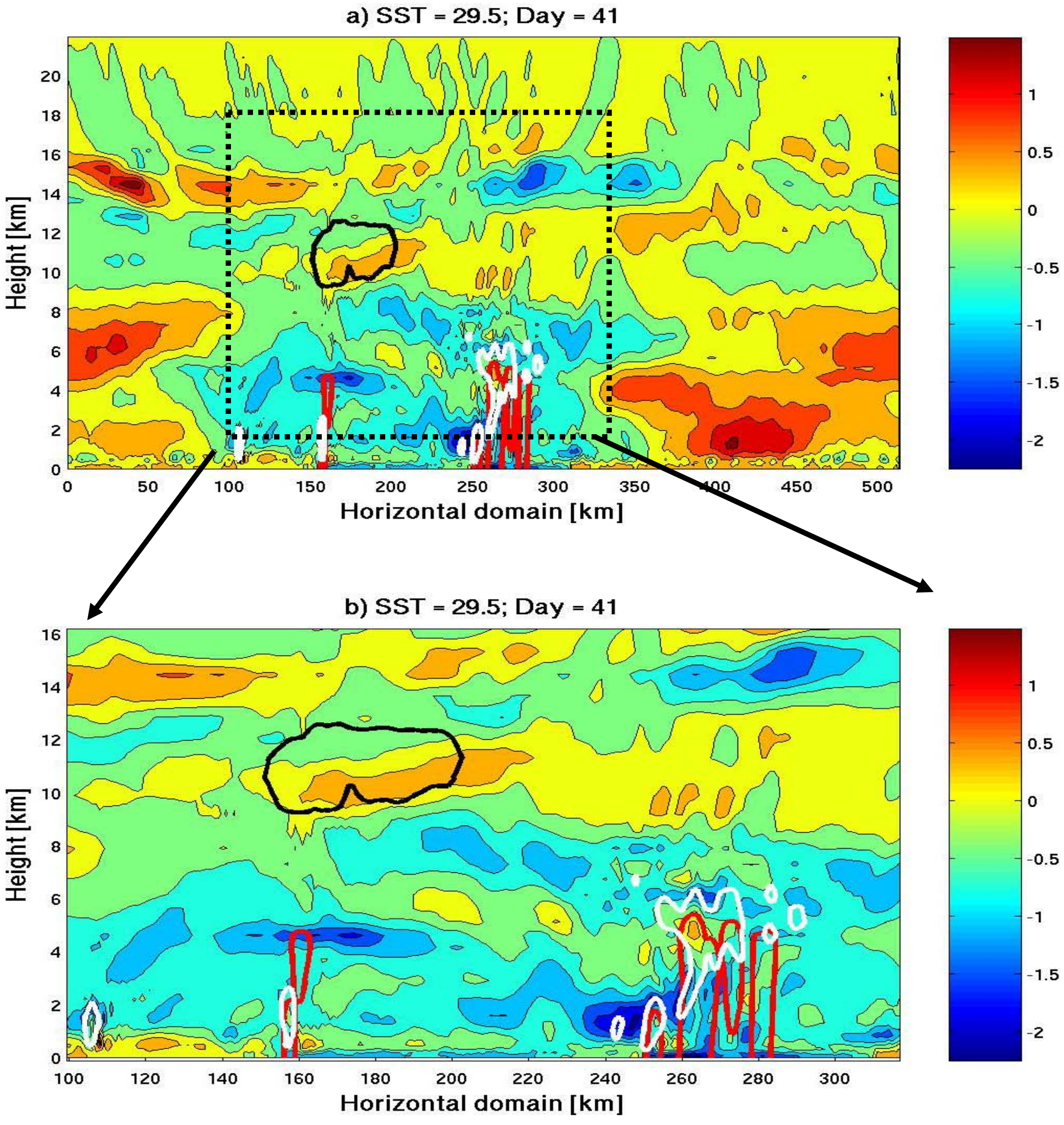


Tropospheric precipitable water vs. rainfall rate



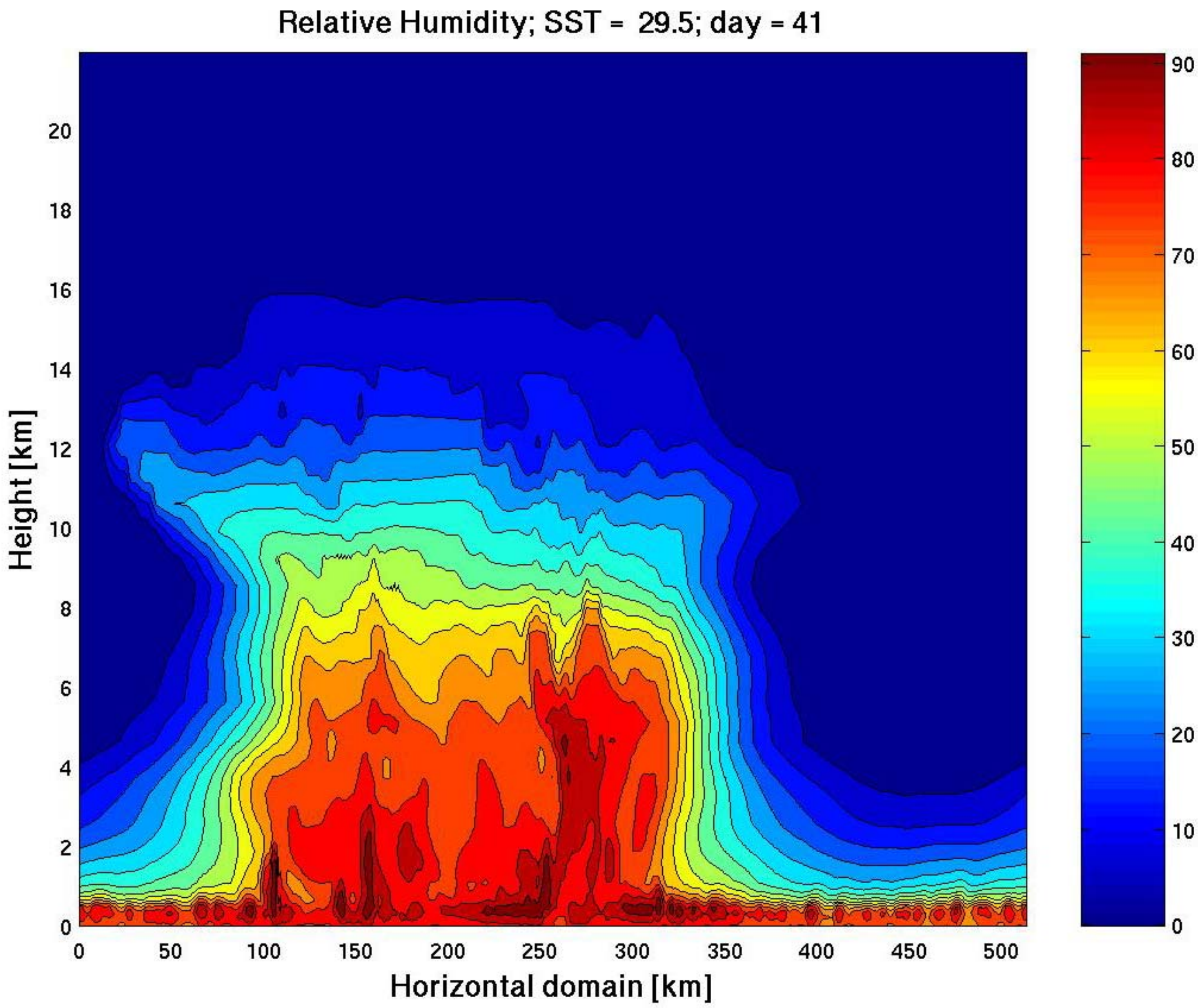
Perturbation potential temperature with minimum contours of

- Rain (red) $2x10^{-4}$ g/kg
- Cloud water (white) $2x10^{-4}$ g/kg
- Cloud ice (black) $1x10^{-4}$ g/kg

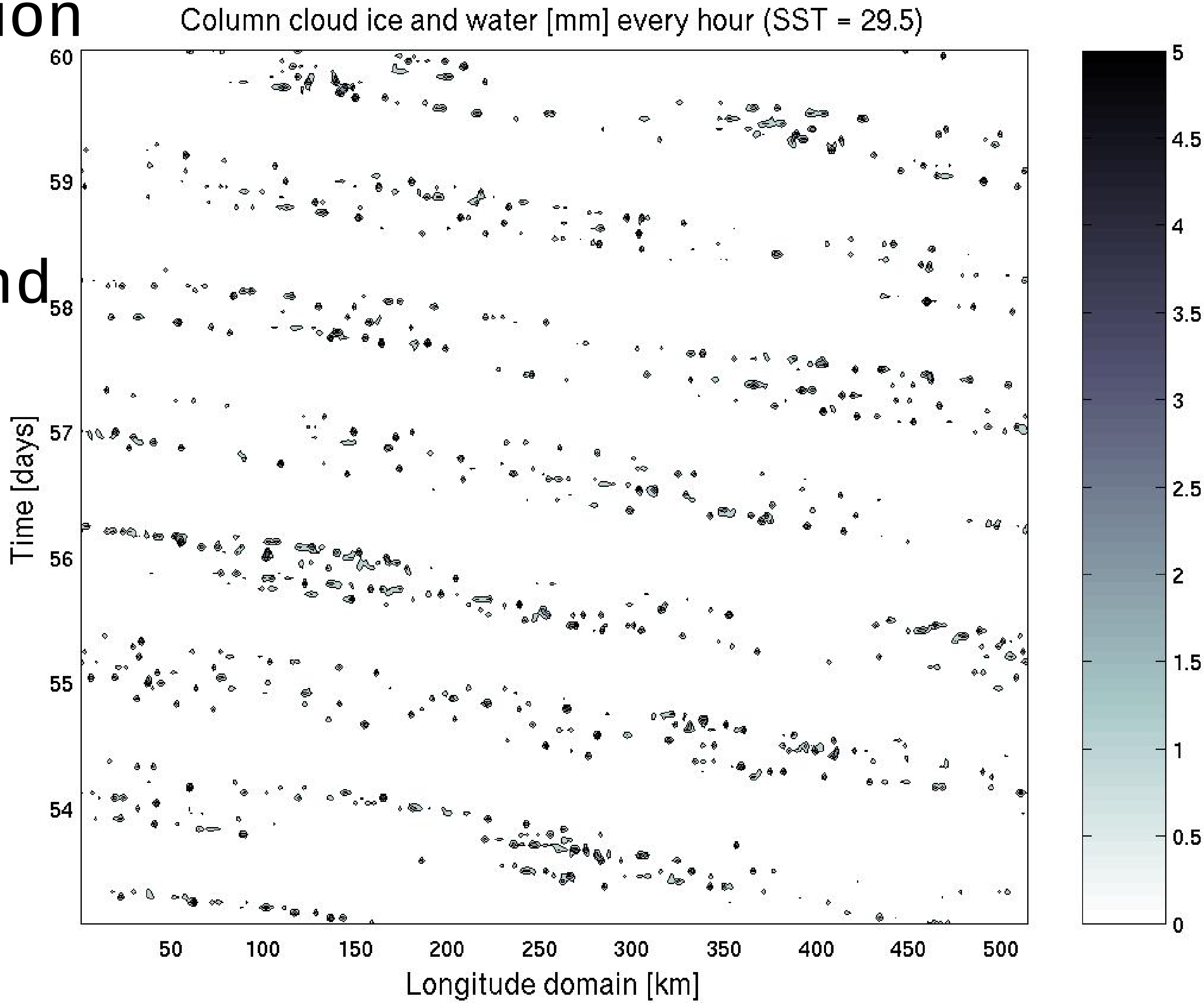


Corresponding relative humidity field

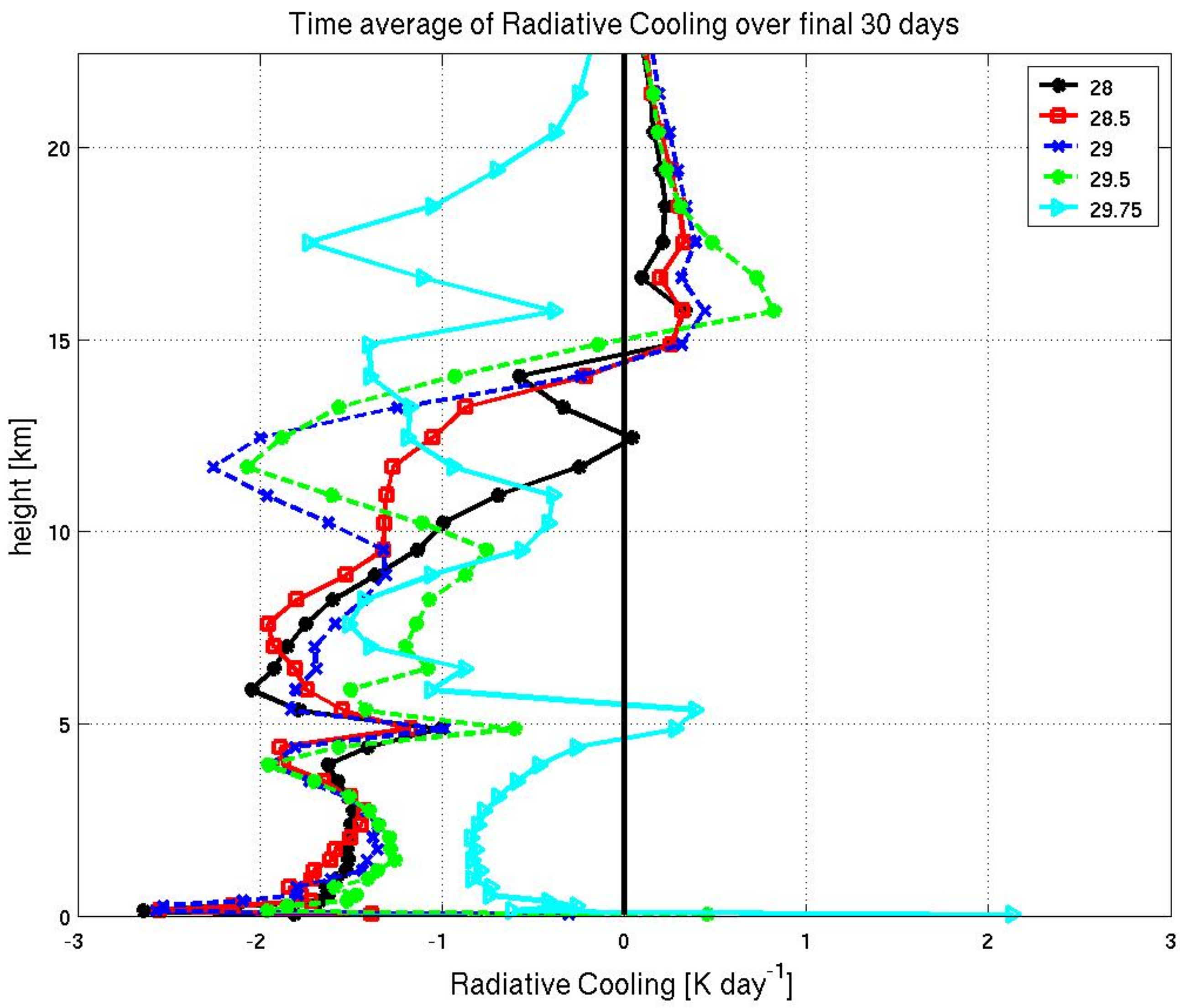
- Even partition of domain into moist and dry
- During month 2, structure is stable and translates westward (right to left).



Cluster propagation speed ¼ mean wind speed



Radiative Cooling



The approximation of the θ equation by $wS=Q$
is a balance approximation: filters gravity waves by
assuming them to be infinitely fast, just like QG etc.

We can formulate explicit equations for the balanced
dynamics (Held and Hoskins 1985; Sobel, Nilsson
and Polvani 2001; Majda and Klein 2003). Versions
of classic models of tropical dynamics such as Gill,
Held-Hou, “mock Walker” etc. can be formulated under
WTG and work more or less fine.

This means when we force an SCM or CRM under WTG,
we have a theory behind the large-scale forcing we’re
applying. We know what phenomena we can capture with
this theory and how well. Not true for imposing large-scale w.