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Atmospheric Dynamics: Effects of Moisture

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Thanks for slides from Ros Cornforth & Hylke De Vries

Lecture Outline

- Diabatic processes (resulting in heating)
- Saturation of humidity and moist variables
- Conditional instability and moist convection
- Moisture in large-scale waves
  - Moist up – dry down parameterisation
  - Shear instability with moisture
  - African Easterly waves
  - Equatorial waves coupled with convection
- Transport of moisture within cyclones and stormtracks
- Active research areas

Equations of fluid dynamics

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- Only assuming the *continuum hypothesis*

$$\frac{\partial \rho}{\partial t} + \nabla.(\rho \underline{u}) = 0$$ Conservation of mass

$$\frac{\partial (\rho \underline{u})}{\partial t} + \nabla.(\rho \underline{u}\,\underline{u}) = -\rho \nabla \Phi - \nabla p + \rho \underline{B}$$ Conservation of momentum

$$\frac{\partial (\rho s)}{\partial t} + \nabla.(\rho s \underline{u}) = \rho Q$$ Conservation of entropy

ρ=density $\underline{u}$=velocity

Φ=geopotential p=pressure $\underline{B}$=friction + body forces

$s(p.\rho)$=specific entropy Q=entropy source (from diabatic processes)

1. Role of Constituents in Dynamics

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- Trace constituents (gases, aerosol) advected by winds

$$\frac{\partial(\rho q)}{\partial t} + \nabla.(\rho q \underline{u}) = \rho S \qquad \frac{Dq}{Dt} = S$$

Eulerian (flux) form — Lagrangian form

- How do they affect atmospheric dynamics?

1. influence density and pressure (Dalton's Law of partial pressures)

$$p = \sum_i p_i \qquad p_i = \frac{n_i}{n_A} R_* T = \frac{\rho_i}{M_i} R_* T$$

R_* = Universal gas constant

n_A = Avogadro's number

M_i = molecular weight

2. large particles may have different inertia
3. influence heating

Influence on Heating

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- Radiatively active
  - molecules/particles can scatter photons
  - molecules can absorb and re-radiate photons
  - particularly in infrared since photon energies are comparable with molecular energy levels (vibration, rotation)
  - radiative flux convergence $\Rightarrow$ heating of atmosphere

- Phase changes of water
  - Latent heat release from condensation of water vapour
  - Diabatic cooling associated with evaporation of liquid/ice

Radiatively Active Constituents

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- Greenhouse gases

  (water vapour, carbon dioxide, methane, ozone …)

  absorb infrared radiation and re-emit at wavelength corresponding to local temperature
- Clouds – liquid and ice phases
- Ozone
  - photolysis at solar UV wavelength ⇒ heating in stratosphere
- Aerosol
  - Mainly scattering of solar and absorption of IR
  - strongly dependent on aerosol properties (chemical composition, size, shape etc)

Radiative Effects on Dynamics

- Greenhouse gases (aside from water)
  - Long-term radiative forcing of climate
  - Time variation not considered for numerical weather prediction
  - However, climatological profiles needed for retrieval of satellite data
- Stratospheric ozone hole
  - Ozone hole is key example of chemistry coupled with dynamics
  - Rapid ozone loss requires cold temperatures (polar stratospheric clouds) and return of sunlight in spring
  - But less ozone $\Rightarrow$ less heating $\Rightarrow$ colder for longer
- Stratospheric ozone and thermal tide
  - Heating on sunlit side of planet excites thermal tide (diurnal and semi-diurnal periods)
  - Strong signal even in Tropical surface pressure *(Lindzen)*
- Mineral dust
  - Dust parameterisation over West Africa improves skill in forecasts *(Tompkins and Rodwell)*
- *Clouds…*

2. Thermodynamics with Moisture
Recap dry potential temperature

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- Temperature attained by moving an air parcel *adiabatically* to reference pressure, p_0

$$\theta = T\left(\frac{p}{p_o}\right)^{-\frac{R}{c_p}}$$

Specific entropy $s = c_p \ln\left(\frac{\theta}{\theta_o}\right)$

- Adiabatic motions following dry air parcels

$$\frac{Ds}{Dt} = \frac{D\theta}{Dt} = 0$$

- Stable stratification (dry atmosphere)

$$\frac{\partial\theta}{\partial z} > 0 \qquad N^2 = \frac{g}{\theta_0}\frac{\partial\theta}{\partial z}$$

N = oscillation frequency

Effects of moisture loading

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Molecular weight of water differs from “dry air” mix.
⇒ lower density of moist air at fixed *T, p.*

In unsaturated air *virtual potential temperature* θ_v is conserved:

$$\theta_v = \left(\frac{1 + r/\varepsilon}{1 + r}\right) T \left(\frac{p}{p_0}\right)^{-\frac{R_d}{c_{pd}}}$$

r=mixing ratio of water vapour

$\varepsilon = M_w/M_d \approx 0.622$

With heating from below, obtain well-mixed convective BL.

Characterised by uniform *r* and θ_v.

Often parameterised using turbulence closure scheme.

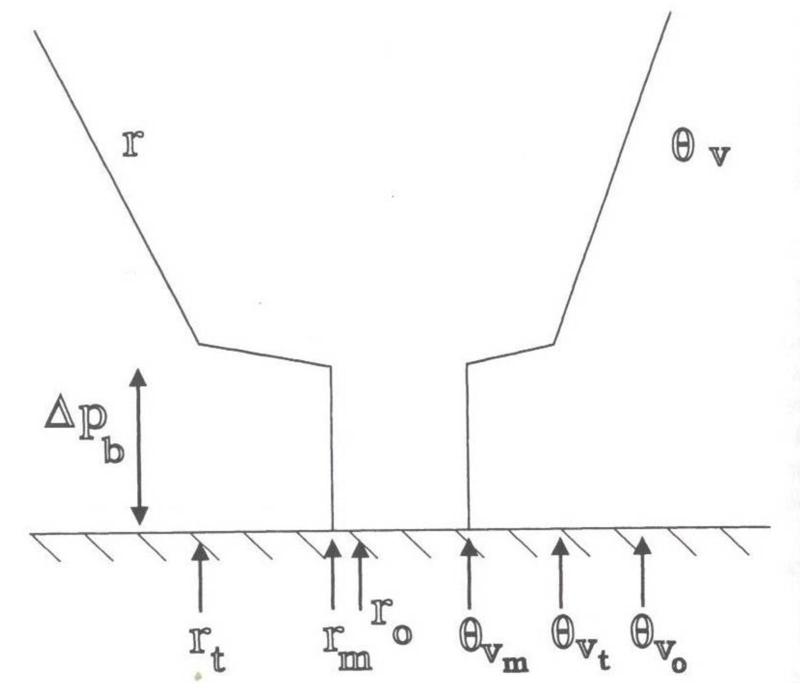

Fig. 13.6 Idealized structure of a cloud-free convective boundary layer.

Saturation Vapour Pressure

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- Gas and liquid can co-exist in equilibrium along a particular curve in pressure-temperature coords.
- Clausius-Clapeyron equation gives its slope:

$$\frac{de_s}{dT} = \frac{Le_s}{RT^2}$$

where L= latent heat of vapourisation

$e_s = e_s(T)$ = *saturation vapour pressure*

- Relative humidity $RH=e/e_s$

Latent Heat Release

- Unsaturated air parcels conserve water vapour mixing ratio

$$r = \frac{\rho_w}{\rho} = 0.622\frac{e}{p}$$

- But if they ascend adiabatically to lower pressure, temperature will fall as well as vapour pressure, *e*

- Clausius-Clapeyron is nonlinear, so eventually $e_s(T) = e = rp/0.622$

- Condensation of vapour results in latent heat release:

$$\delta Q = -L\,dq$$

- For reversible saturated process:

$$Tds = c_V dT + pd\left(\frac{1}{\rho}\right) - Ldq_s$$

Specific humidity

$q=\rho_w/\rho_d = rp/(p-e) \approx r$

Effects of saturation

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Saturation vapour pressure is a function of temperature only: $e=e_s(T)$
Collapses 3 variables (pressure, temperature and humidity) into one conserved variable for *reversible moist adiabatic processes*, *equivalent potential temperature*:

$$\theta_e = T\left(\frac{p_d}{p_0}\right)^{-\frac{R_d}{c_{pd}+r_t c_l}} \exp\left[\frac{Lr}{(c_{pd}+r_t c_l)T}\right]$$

r_t =mixing ratio of total water

c_l =heat capacity of liquid water

Or the *moist static energy* (conserved if also hydrostatic):

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

- Cloudy convective BL or *stratocumulus-topped BL*
- Characterised by uniform h, r_t
- *See Emanuel's book (1994)*

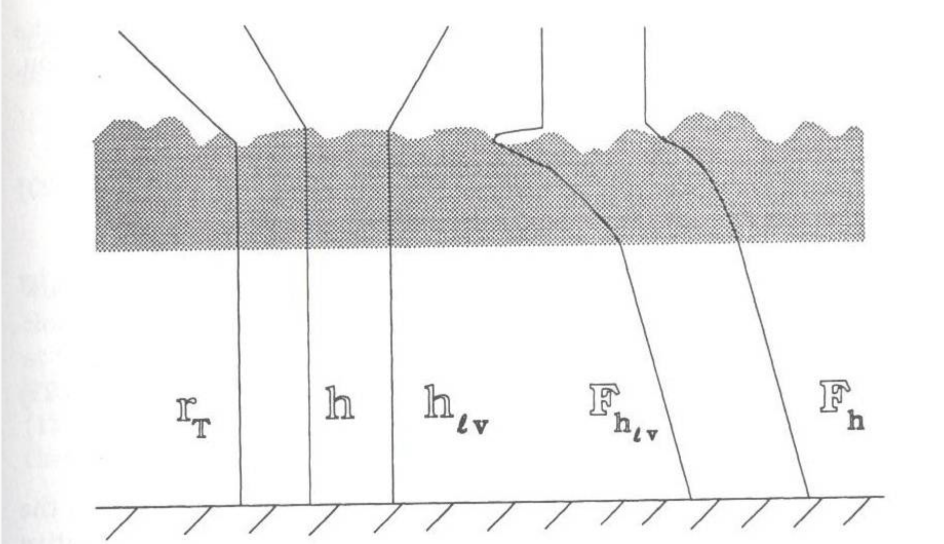

Fig. 13.9 Idealized profiles of conserved variables and turbulent fluxes of h_{lv} and h in a stratocumulus-topped mixed layer.

3. Conditional Instability

$\theta_{es}(p,T) = \Theta$ describes a single curve on a thermodynamic diagram - the *moist adiabat*

$$\frac{\partial \theta_{es}}{\partial z} < \frac{\partial \theta}{\partial z}$$

static stability always lower for saturated air parcels.

$$\frac{\partial \theta_{es}}{\partial z} < 0 < \frac{\partial \theta}{\partial z}$$

conditionally unstable (i.e., only if saturates)

Lifting condensation level (LCL) is pressure that an unsaturated parcel would need to be lifted adiabatically to reach saturation.

Estimate of cloud base.

Conditional instability indicates that sufficient vertical motion could trigger saturation and then moist convective instability.

Deep versus shallow convection

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Shallow convection usually develops a well defined inversion at top.

In Tropics and Subtropics this occurs in regions of large-scale descent.

In regions of deep convection net motion is upwards, occurring entirely in convective updrafts (with partially compensating descent in surrounding clear air).

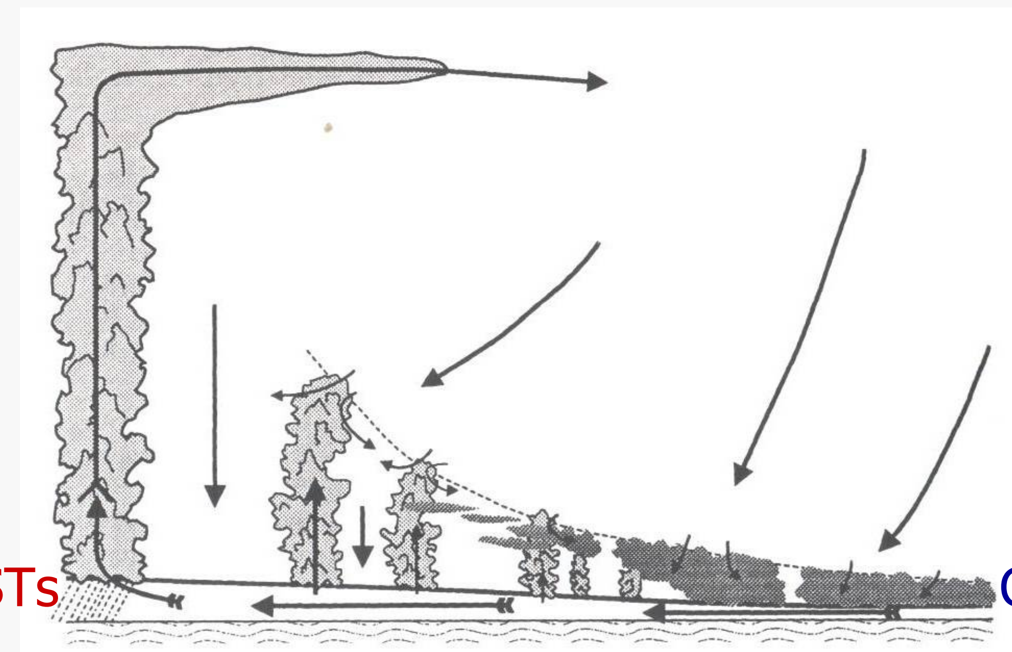

Fig. 14.7 Structure of the tropical atmosphere, showing the various regimes, approximately as a function of sea surface temperature.

What should a parameterisation achieve?

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Modify the resolved scales through:

- Redistribution of temperature (via advection, mixing and heating)
- Redistribution of water (vapour, liquid, ice)
- Redistribution of momentum

Closure assumptions required to represent *statistics* of sub-grid scale fluxes in terms of resolved variables.

If the resolved scales were to depend on the recent history of convective-scale motions then *predictability* of large-scales would be severely limited to hours.

$\Rightarrow$ Convection assumed to be in *quasi-equilibrium* with large-scale environment:

timescale for convective fluxes to respond to environment

<< timescale for modification of environment by convection

4. Moisture in large-scale waves
Moist baroclinic instability

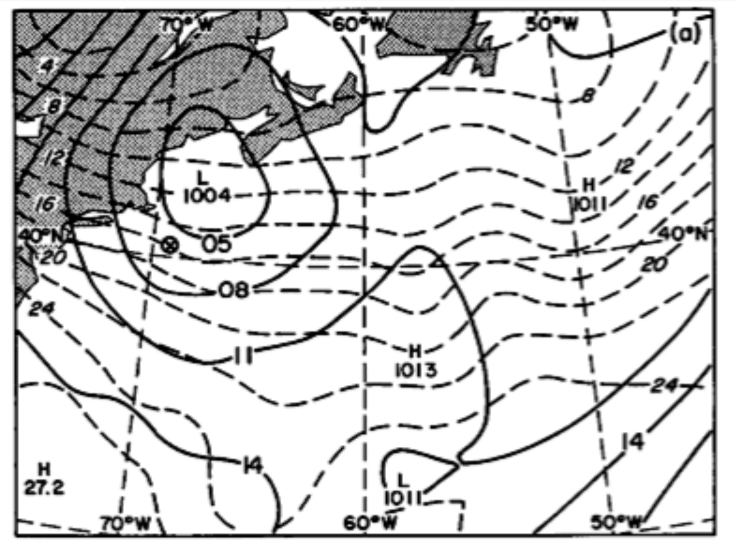
- Why would/should we be interested?
  - Influence of diabatic processes on cyclone development
  - Forecast error and link to those processes
  - Changes in storms in a future climate

- What to expect? Diabatic heating affects wave
  - Propagation
  - Growth
  - Vertical structure
  - Non-modal growth possibilities …

- **Can we obtain ‘moist’ CRWs and retain a simple interpretation of moist baroclinic instability?**

Effects of "switching off" diabatic processes

- Case Study: QE-II storm (September, 1978)
  - Massive 60 mb deepening in 24 hr
  - Developing in relatively 'weak' low-level baroclinic zone
  - Storm responded hurricane-like to the heating **(Guyakum,1983)**
- **Kuo et al (1990)** : adiabatic vs full-physics *simulations*



from Kuo etal, 1990, Fig. 1

Initial condition

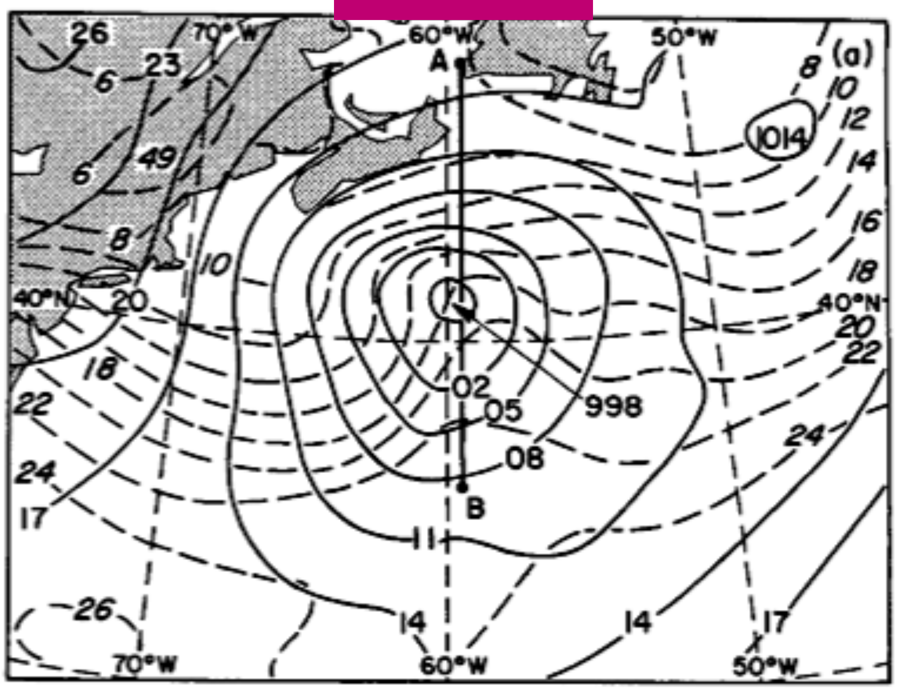
--- = Surface temp,

__ = Surface pressure

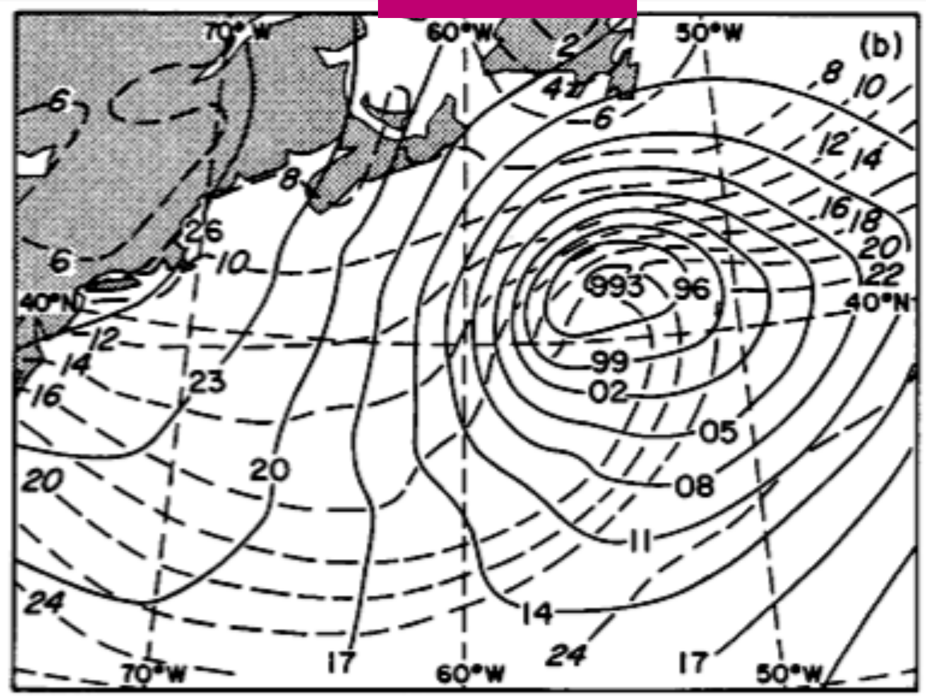
Adiabatic model

- - - - = surface temp   __ = surface pressure

T = 12 HR



T = 24 HR



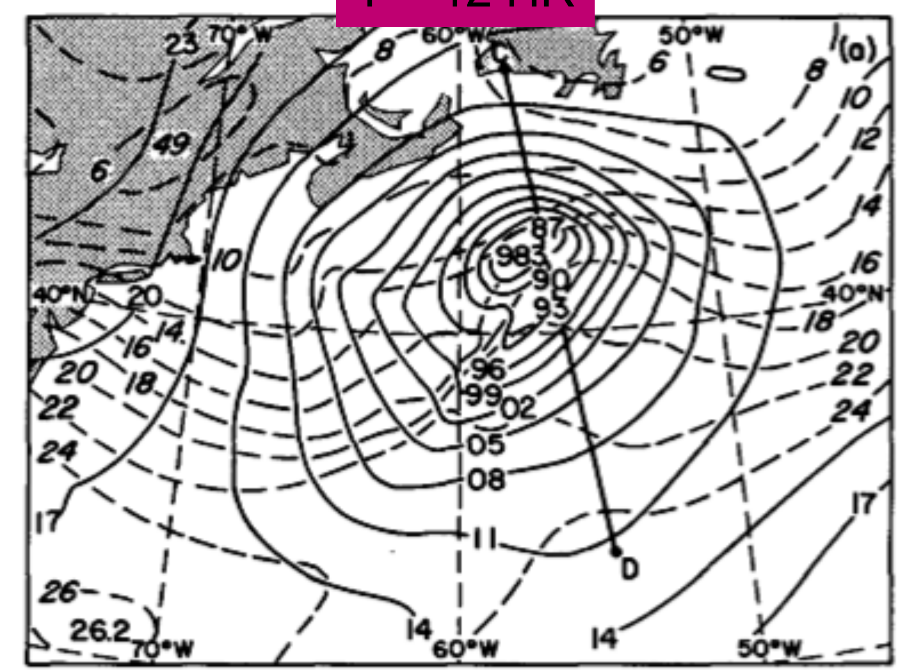
from Kuo etal, 1990, Fig 2

Full-physics model

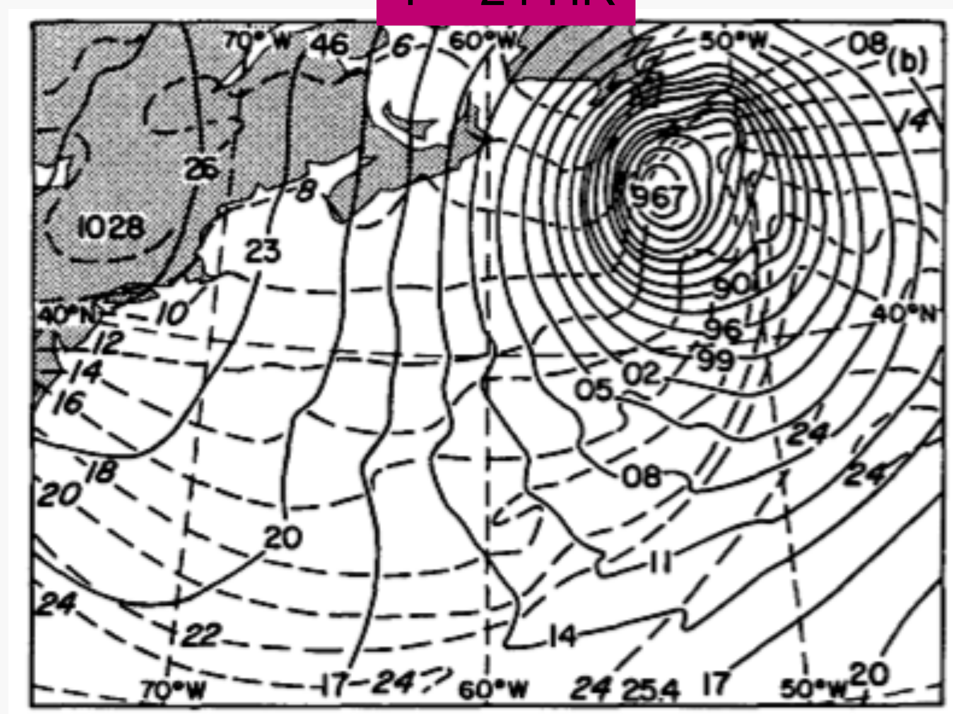
- --- = surface temp
- __ = SLP

full vs adiabatic:
deeper
faster
more intense vertical motion

T = 12 HR



T = 24 HR



from Kuo etal, 1990, Fig 7

Reduced static stability argument

- Recall from dry baroclinic wave theory
  growth rate $\sigma \approx 0.31 \dfrac{f\Lambda}{N}$ where shear $\Lambda = \dfrac{\partial \overline{u}}{\partial z}$
- Saturated regions $N_s^2 = \dfrac{g}{\theta_0} \dfrac{\partial \theta_{es}}{\partial z} < N^2$
- Generally for lower static stability expect faster growth
- But, descent ⇒ adiabatic compression ⇒ T increases
  ⇒ descending air is always below saturation.
- Small amount of ascent often attains saturation
- "Moist-up, dry-down" parameterisations of latent heat release.

Effective static stability

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- Heading only on ascent introduces asymmetry and nonlinearity in the thermodynamic equation.

1. Use $\theta = \theta(p, \theta_{es})$ and conservation of θ_{es} for a saturated process

$$\Rightarrow \quad \frac{D\theta}{Dt} = \frac{Dp}{Dt}\frac{\partial\theta}{\partial p}\Big|_{\theta_{es}} + \frac{D\theta_{es}}{Dt}\frac{\partial\theta}{\partial\theta_{es}}\Big|_{p} = \omega\frac{\partial\theta}{\partial p}\Big|_{\theta_{es}}$$

2. Assume saturated on ascent (ω<0), unsaturated on descent:

$$\frac{\partial\theta}{\partial t} + \underline{V}.\nabla\theta = -\omega\frac{\partial\theta}{\partial p} + \hat{\omega}\frac{\partial\theta}{\partial p}\Big|_{\theta_{es}}$$

=ω when ω<0, otherwise 0

3. O'Gorman (*JAS, 2010)* suggested a means to calculate an *effective static stability* approximating truncated upward velocity by a rescaling of full vertical velocity:

$$\hat{\omega} \approx \lambda\omega + \varepsilon$$

4. Effective static stability as $$-\frac{\partial\theta}{\partial p}\Big|_{eff} = -\frac{\partial\theta}{\partial p} + \lambda\frac{\partial\theta}{\partial p}\Big|_{\theta_{es}}$$

Balanced part of vertical motion

- Moist up – dry down approach requires large-scale vertical velocity
- Back to quasi-geostrophic theory

$$D_g(f+\xi_g) = f_0 \frac{1}{\rho_r}\frac{\partial(\rho_r w)}{\partial z} \qquad D_g b' + N^2 w = B$$

…and use thermal wind balance to eliminate rate of change:

$$N^2\nabla_h^2 w + f_0^2 \frac{\partial}{\partial z}\left(\frac{1}{\rho_r}\frac{\partial(\rho_r w)}{\partial z}\right) = 2\nabla_h.\underline{Q} + f_0\beta\frac{\partial v_g}{\partial z} + \nabla_h^2 B$$

QG omega equation

w obtained by inversion

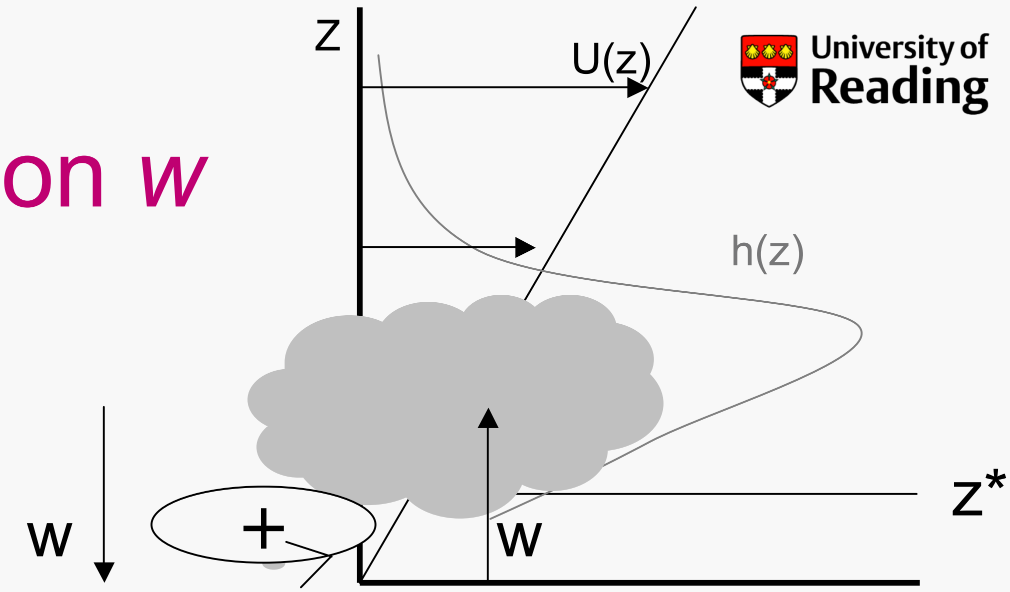
"forcing" of ascent by geostrophic flow and heating (*B*)

(in this form, Hoskins *et al* , *QJRMS,* 1978)

Heating closure on *w*

- Diabatic heating ~ w

$$\dot{\theta} = \epsilon h(z) w^*$$



- Wave-tuned heating by inverting omega eqn

$$\mathsf{L}_\mathsf{w} w' = F + F_h$$

  - w^* = *Dynamically* induced vertical velocity at z* (**Mak, 1982**)
  - w^* = *Total* vertical velocity at z* (**Snyder & Lindzen, 1991**)

  Wave-CISK = “conditional instability of the second kind”

- Diabatic heating profile => PV change following flow

$$\dot{q}(z) \sim \frac{\partial \dot{\theta}}{\partial z} \sim \epsilon \frac{\partial h}{\partial z} w^*$$

Diabatic Effects within Rossby Waves
Wave-CISK parameterisation

Specify heating following Mak [1982], Snyder and Lindzen [1991]

$$\dot{\theta}(k,z,t) = \epsilon h(z) w_*(k,t)$$

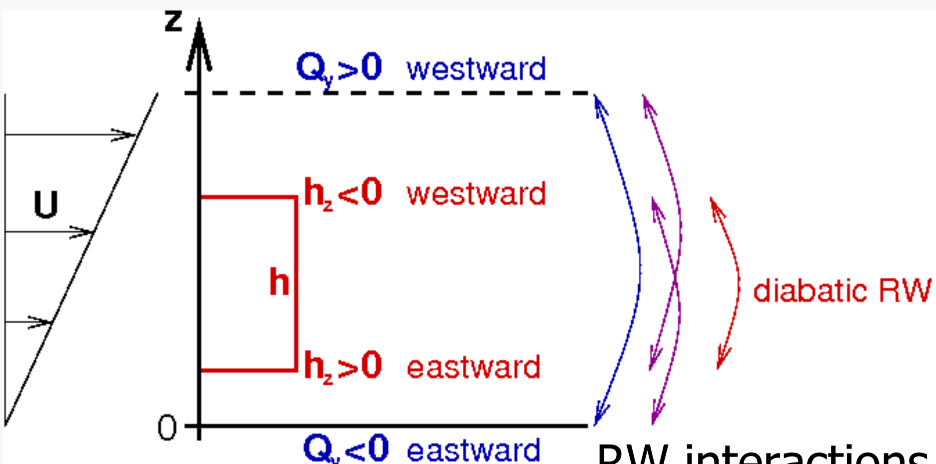
$$\dot{q}(k,z,t) = \epsilon \left[\frac{S^{-1}}{\rho} \frac{\partial}{\partial z} \left(\frac{\rho h(z)}{N^2} \right) \right] w_*(k,t).$$

Heating proportional to *w* **at a particular level z***

Profile specified by h(z) ⇒ diabatic QGPV tendency ∝ ∂h/∂z.

Assuming symmetry – heating on ascent; cooling on descent.

⇒ retains sinusoidal structure of wave.



RW interactions ⇒ mutual growth

Moist Rossby wave phase speed:

$$c = U - \frac{v}{kq}\frac{\partial Q}{\partial y} + \frac{w_*}{kSN^2}\frac{\partial h}{\partial z}$$

De Vries *et al*, *JAS*, *2010*

Moist baroclinic instability

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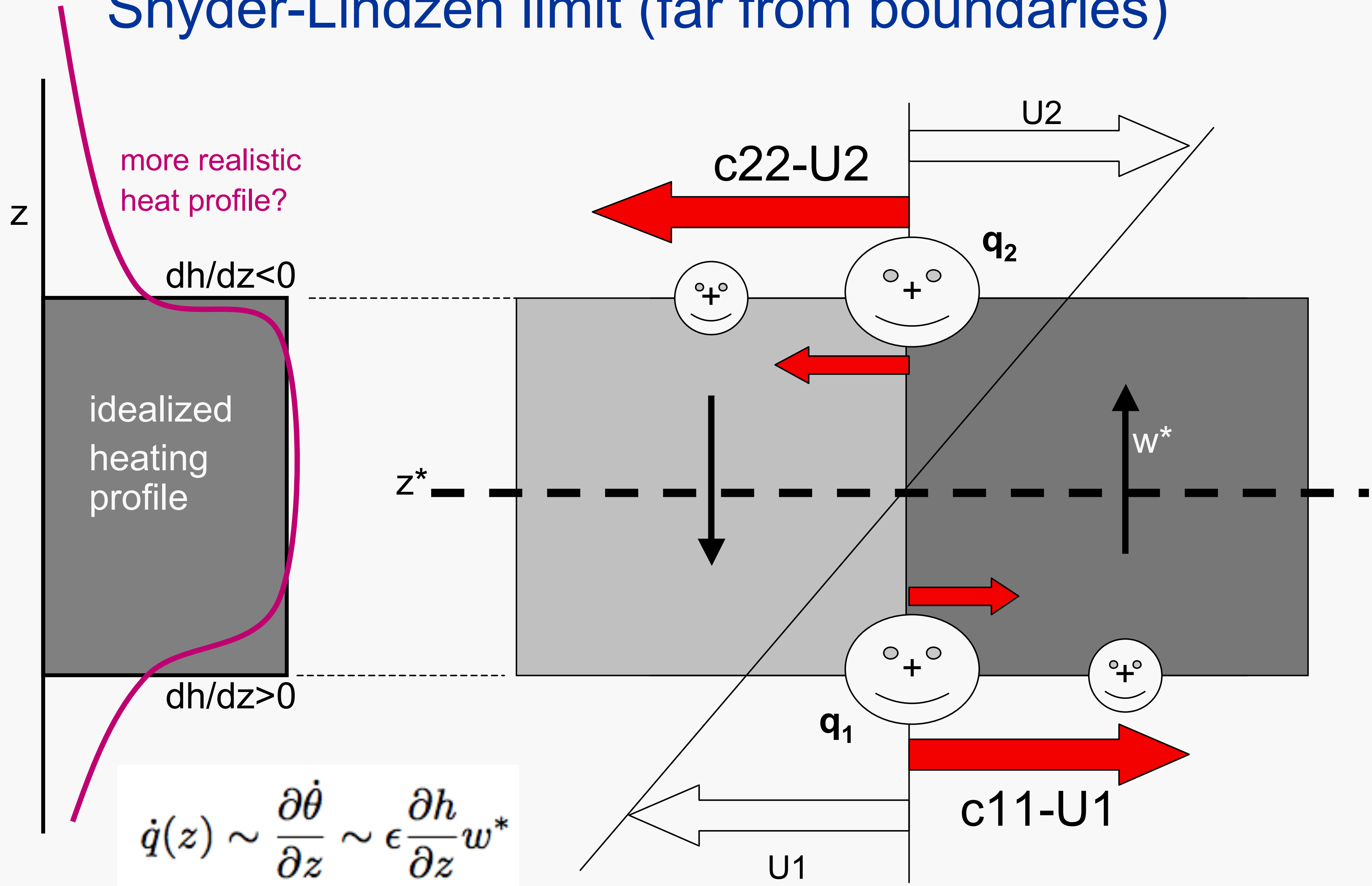
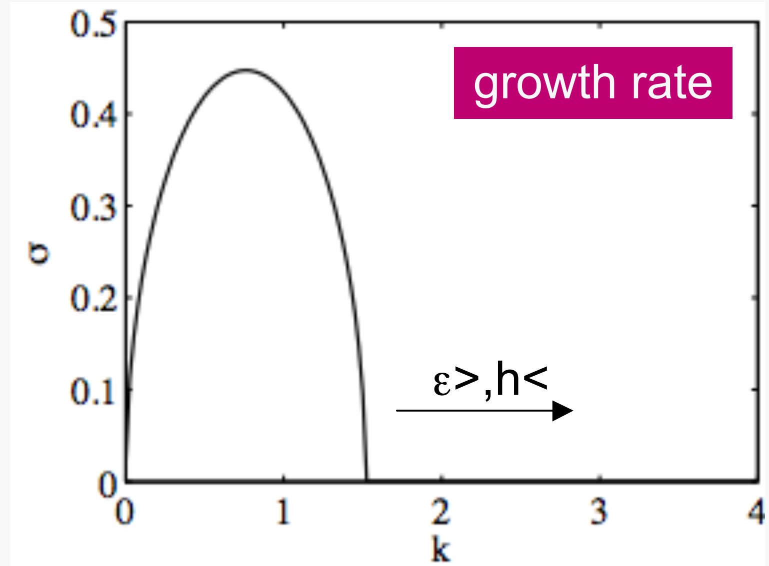
Snyder-Lindzen limit (far from boundaries)



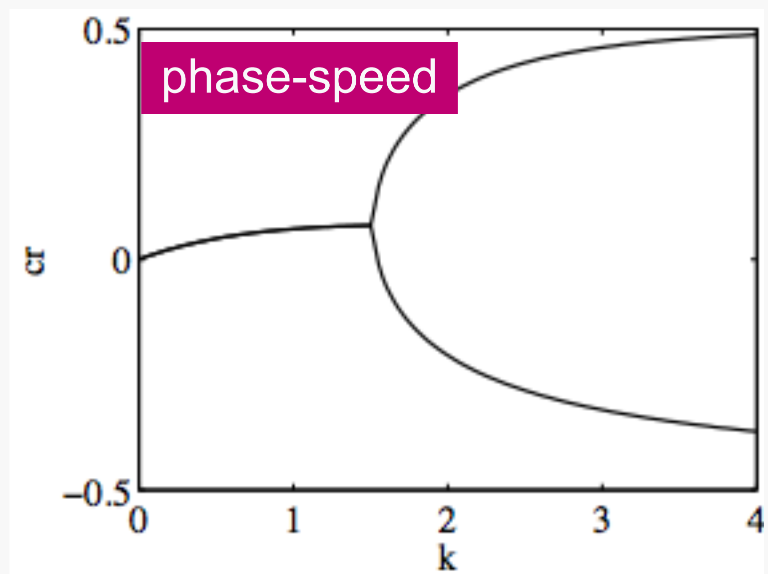
$$\dot{q}(z) \sim \frac{\partial \dot{\theta}}{\partial z} \sim \epsilon \frac{\partial h}{\partial z} w^*$$

Dispersion relation in Snyder-Lindzen limit

- Long-waves unstable and short-waves neutral
- Neutral phase-speeds tend to local flow speed at large k



- (not like a reduced N^2 model)
- Resembles classic Eady model: Also the same interpretation?



Add ground ⇒ 3 wave system

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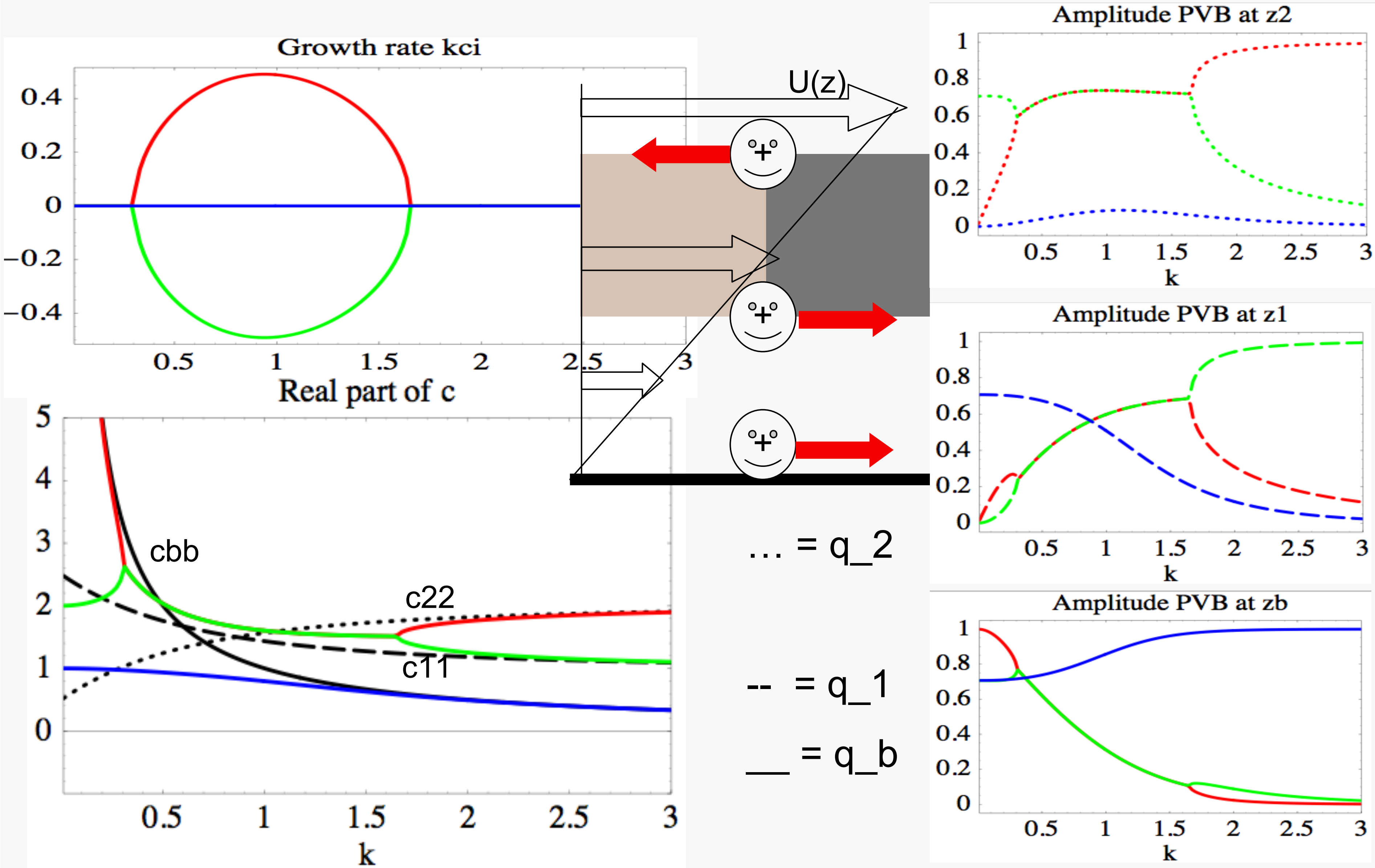
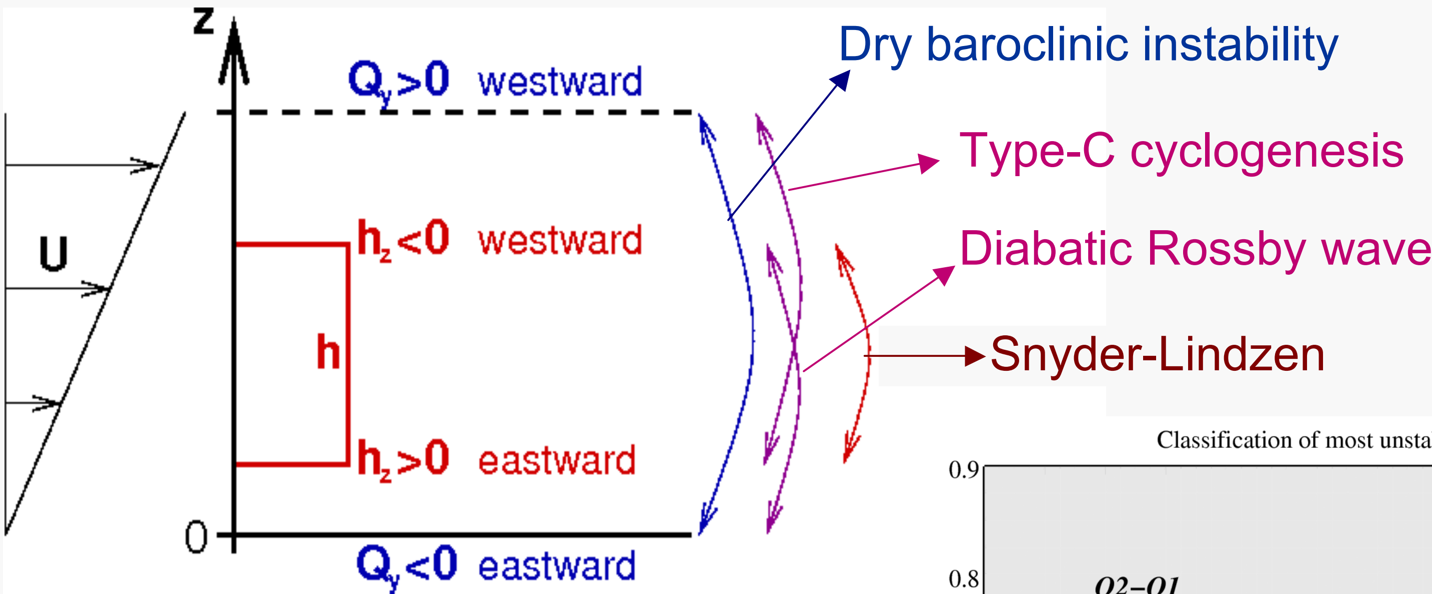


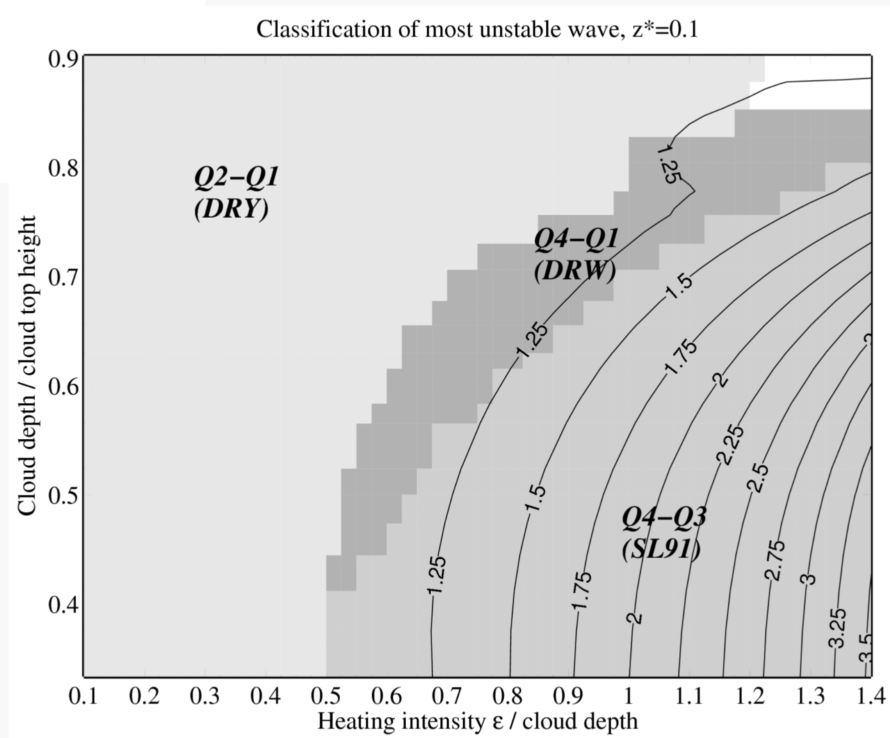
Moist modes and RW coupling



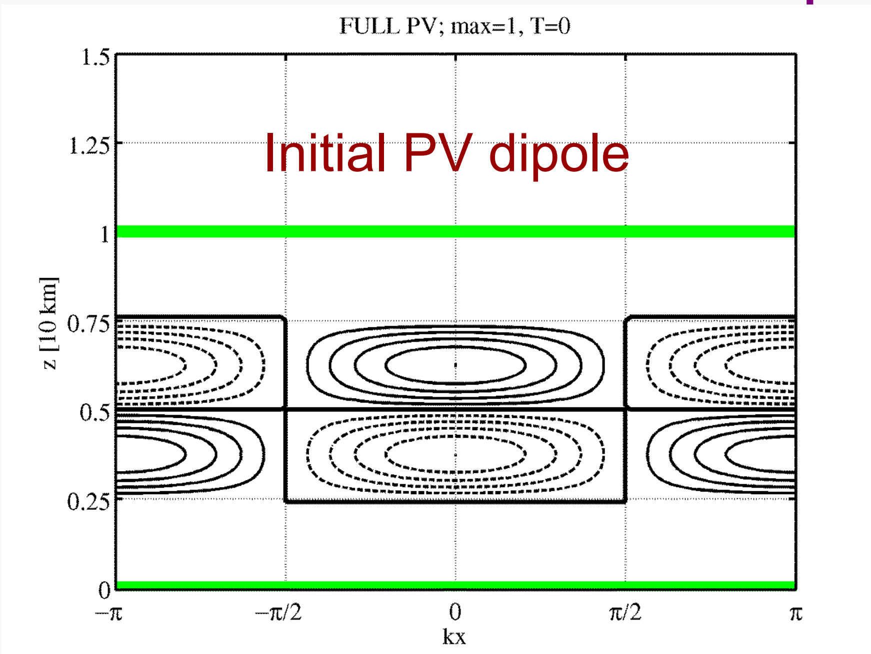
Regime diagram for dominant RW coupling as heating intensity and cloud depth are varied

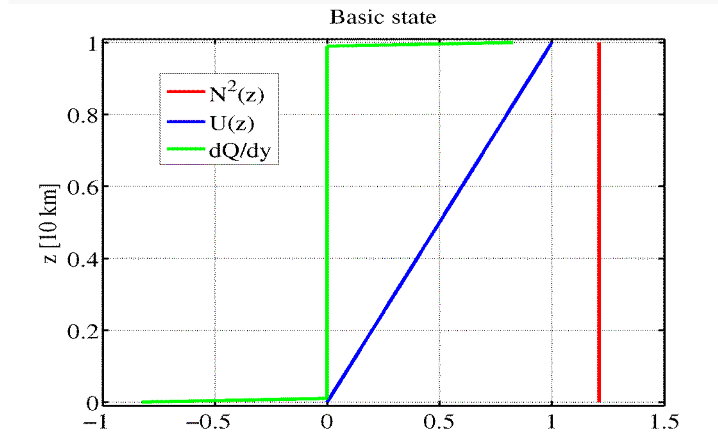
(e.g., Eady basic state)

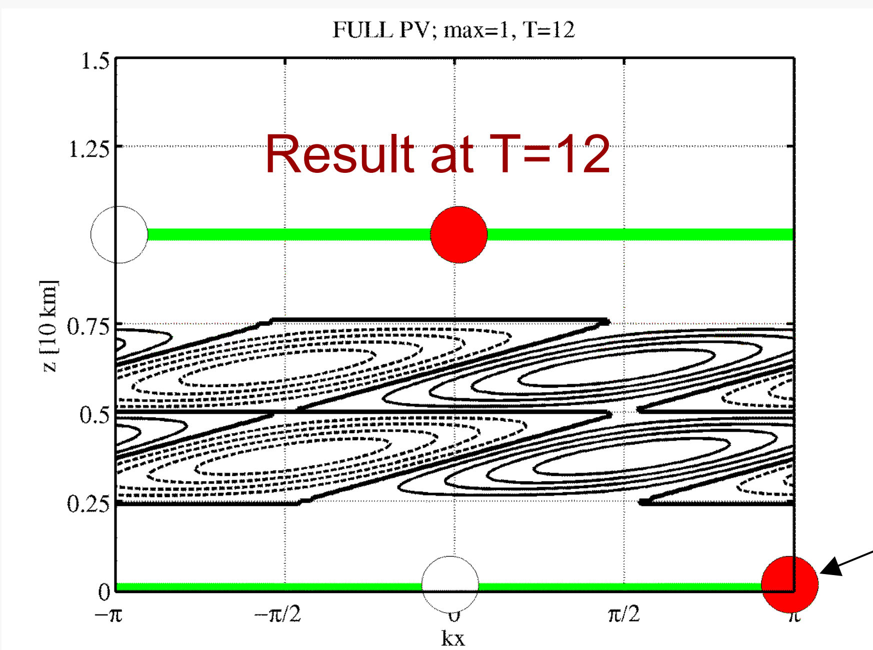
De Vries *et al, JAS, 2010*



Baroclinic initial value problem: Eady model







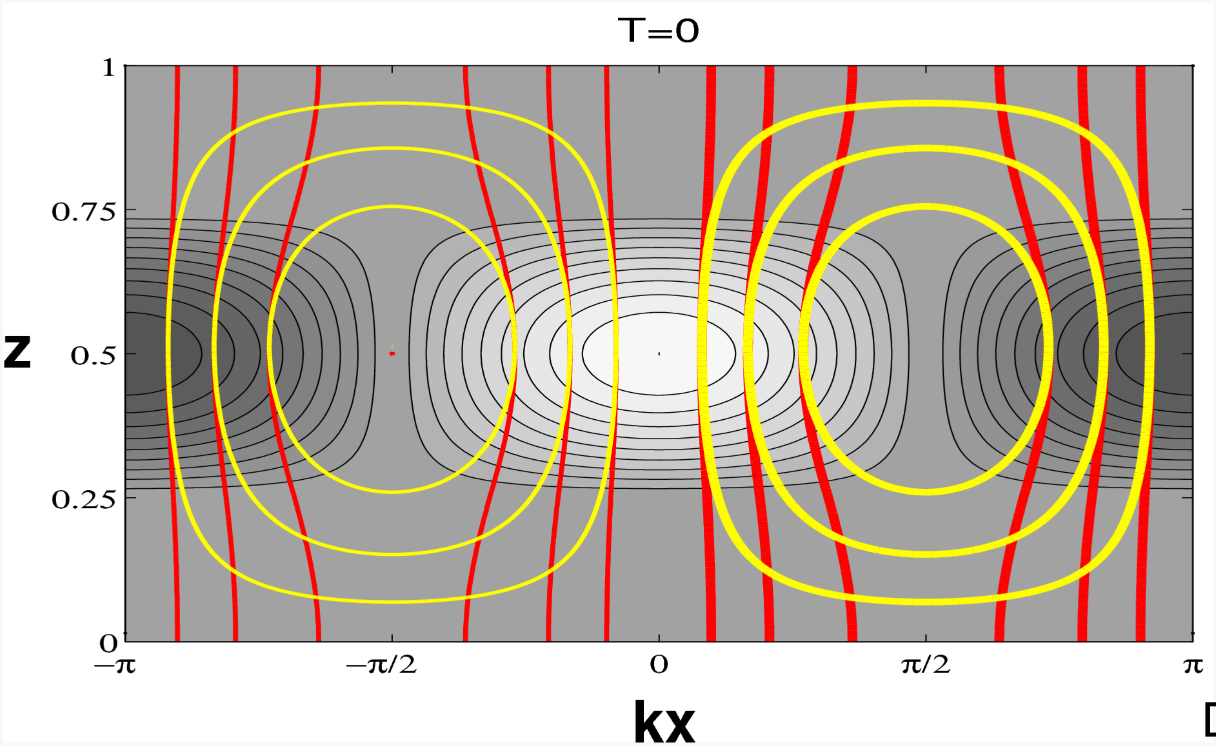
- Uniform vertical shear and static stability
- f-plane (no interior PV gradient)
- Rigid lid

Red = +ve boundary PV anomaly

White = -ve boundary PV anomaly

Moist baroclinic initial value problem

- Example here for "Green model" (Eady model + Q_y=β)
- Constant shear, rigid lid and meridional PV gradient everywhere
- Simple interior PV wave as initial condition



Shading = PV

Red = v

Yellow = w

De Vries et al, JAS, 2010

Full solution moist QG

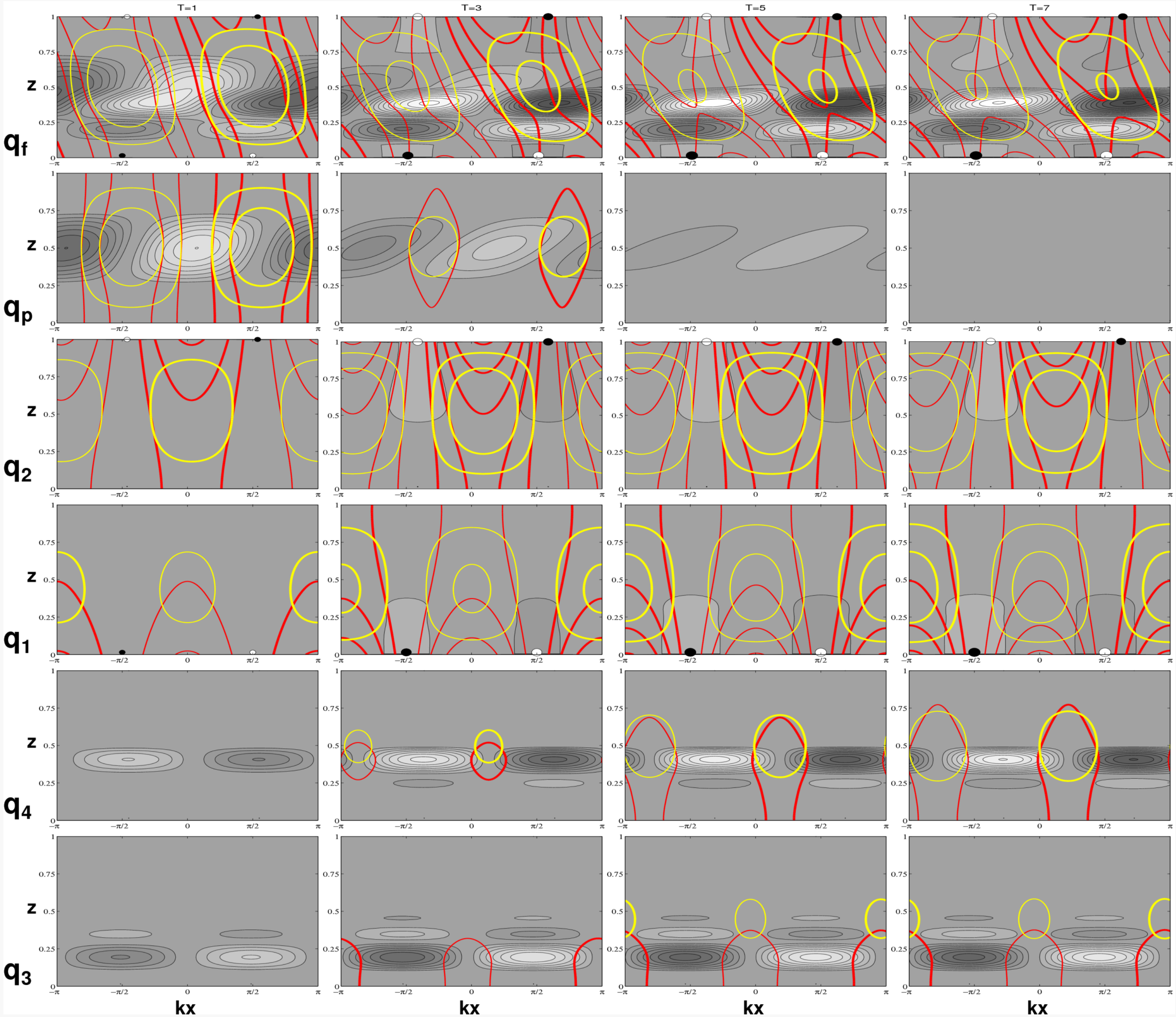
Passive PV

Upper CRW dry

Lower CRW dry

Upper CRW moist

Lower CRW moist



De Vries et al, JAS, 2010

Diabatic Effects - asymmetrical approach

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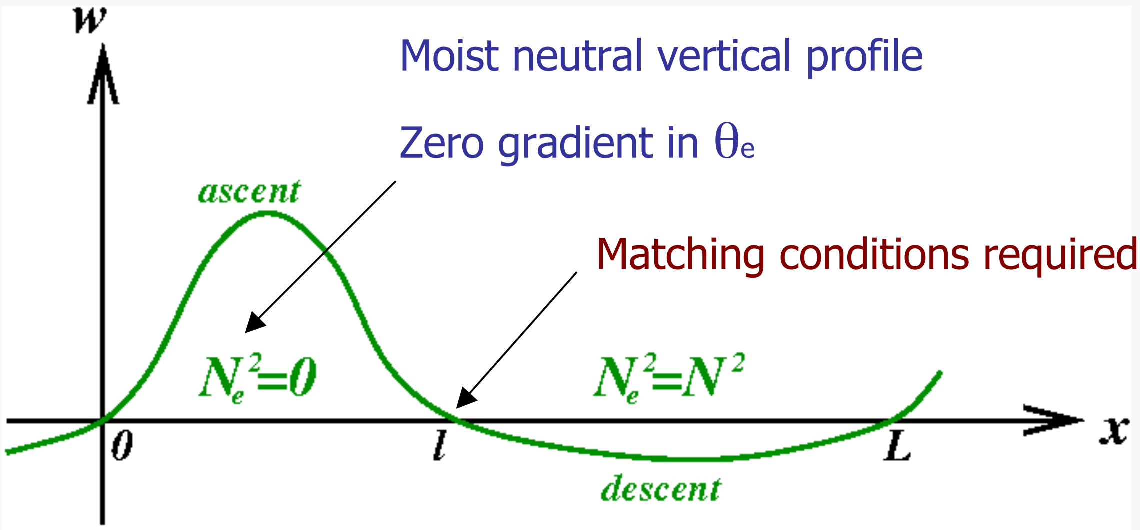
Assumption: saturated ascent; unsaturated descent

+ ascent/descent regions have different static stability *and* widths.

Emanuel, Fantini and Thorpe [1987] found NM solutions in Eady model.

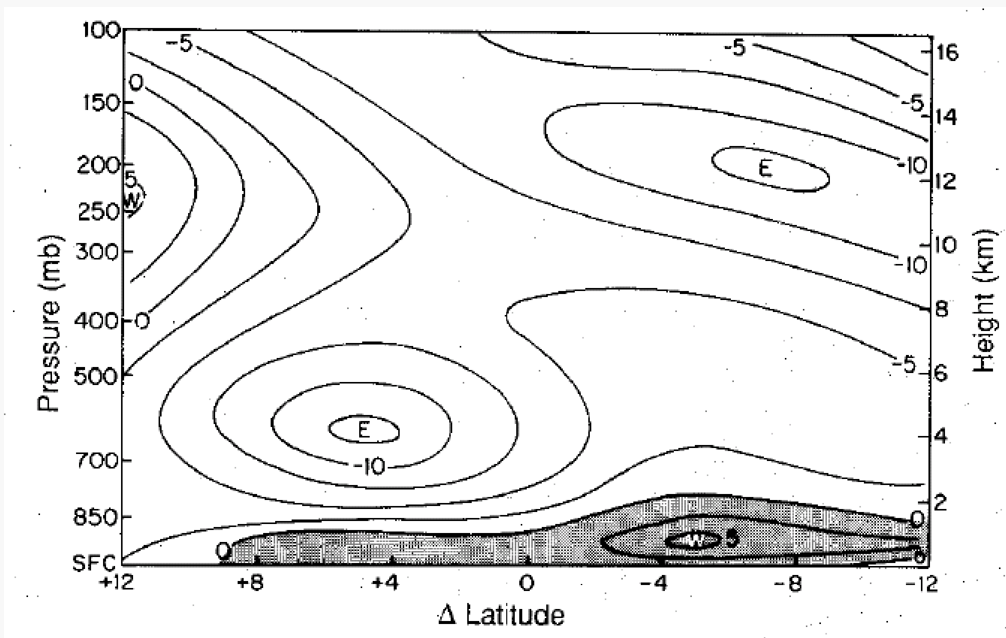


Excites other wavenumbers ⇒ generalisation is difficult

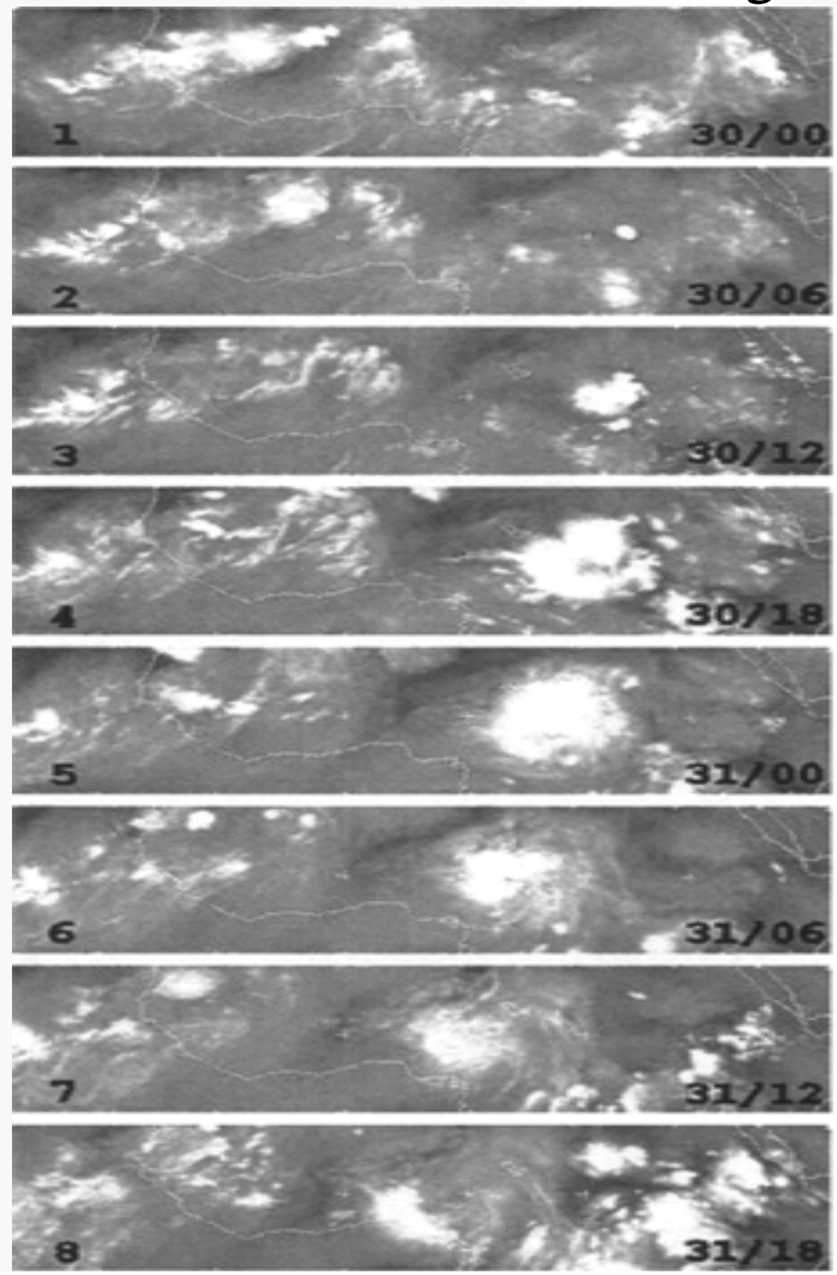
African Easterly Waves

Waves propagating across West Africa along African Easterly Jet

Crucial to rainfall across region



Mean zonal wind (30W-10E)
Reed et al MWR, 1977



Time-average across N. Africa

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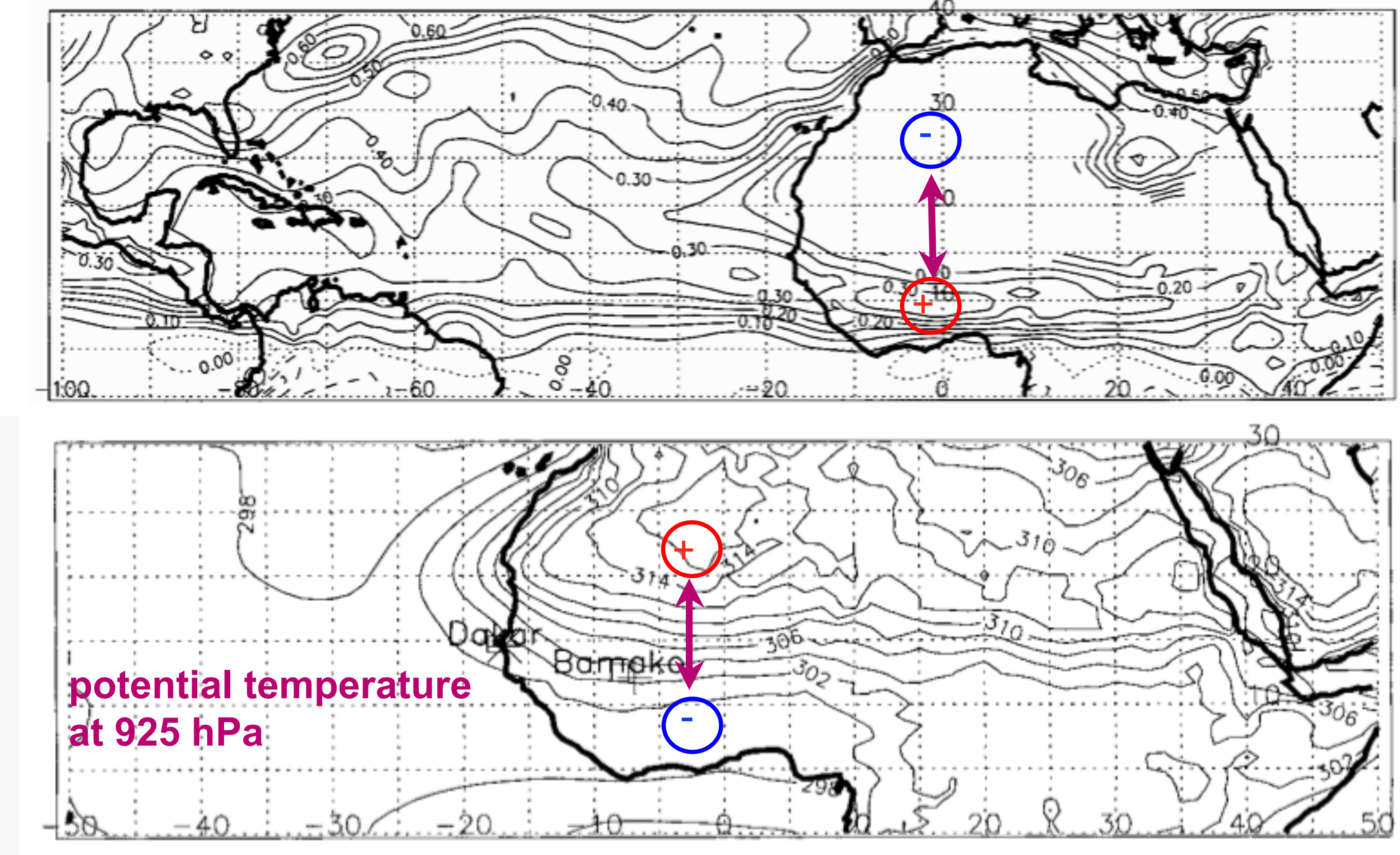
PV at 315K

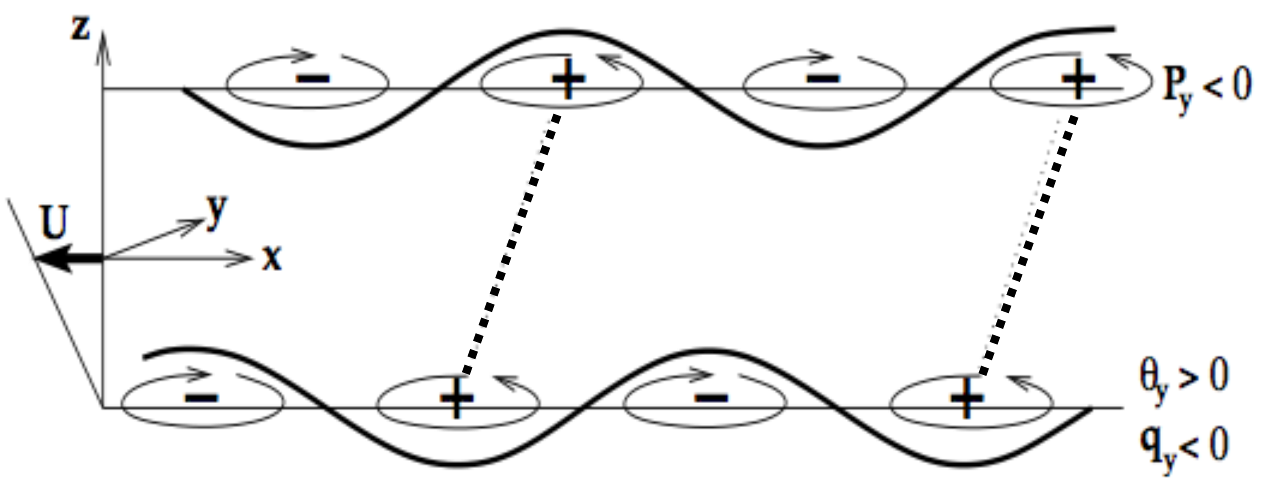


Pytharoulis and Thorncroft (1999)

Horizontal cross-sections of Ertel PV at 315 K and potential temperature averaged over August 1995

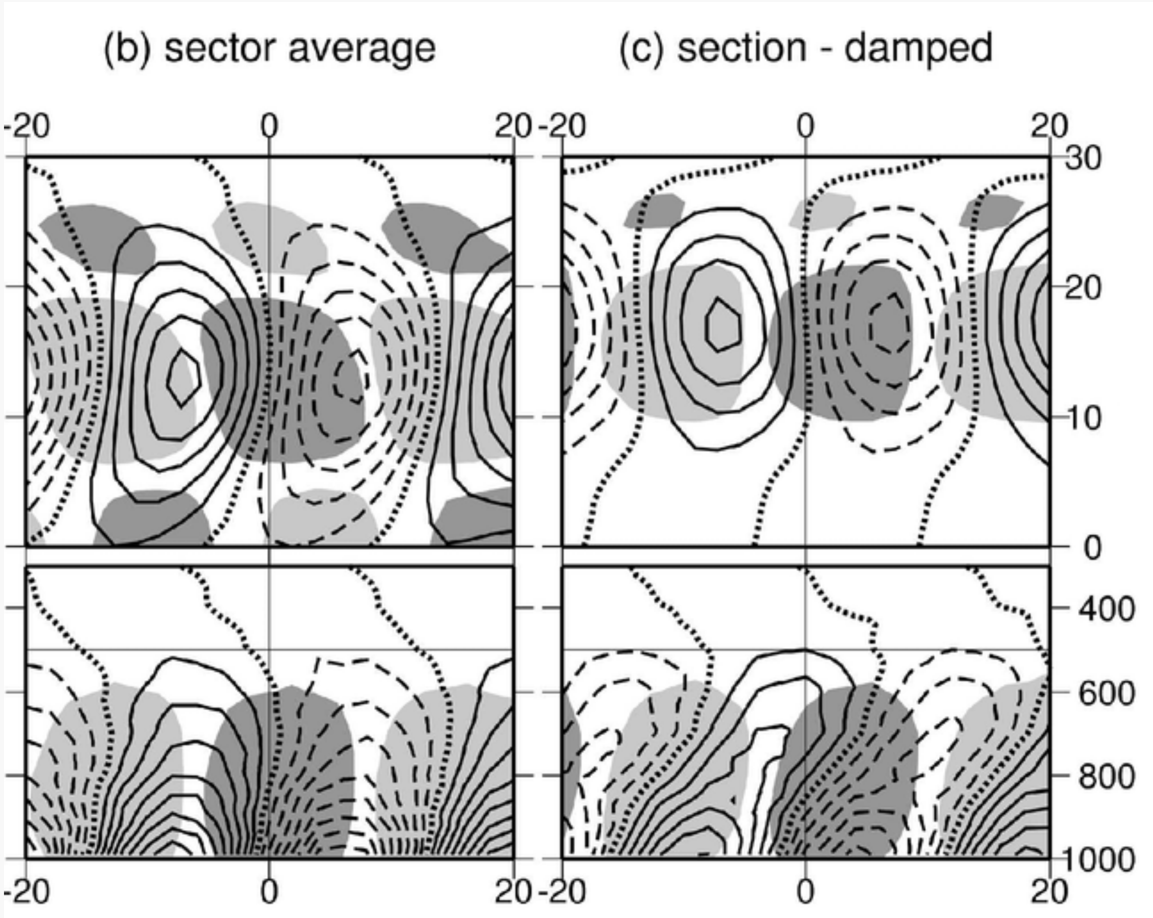
Schematic of AEW mechanism

Vertical section along African Easterly Jet
Phase-locking between CRWs: eastward tilt with height



Mutual baroclinic growth is possible for dry dynamical structure

Dry normal modes



- Obtained by

Hall, *JAS,* 2006

- Dry PE model
- Contours = v

Short-range forecasts in W. Africa

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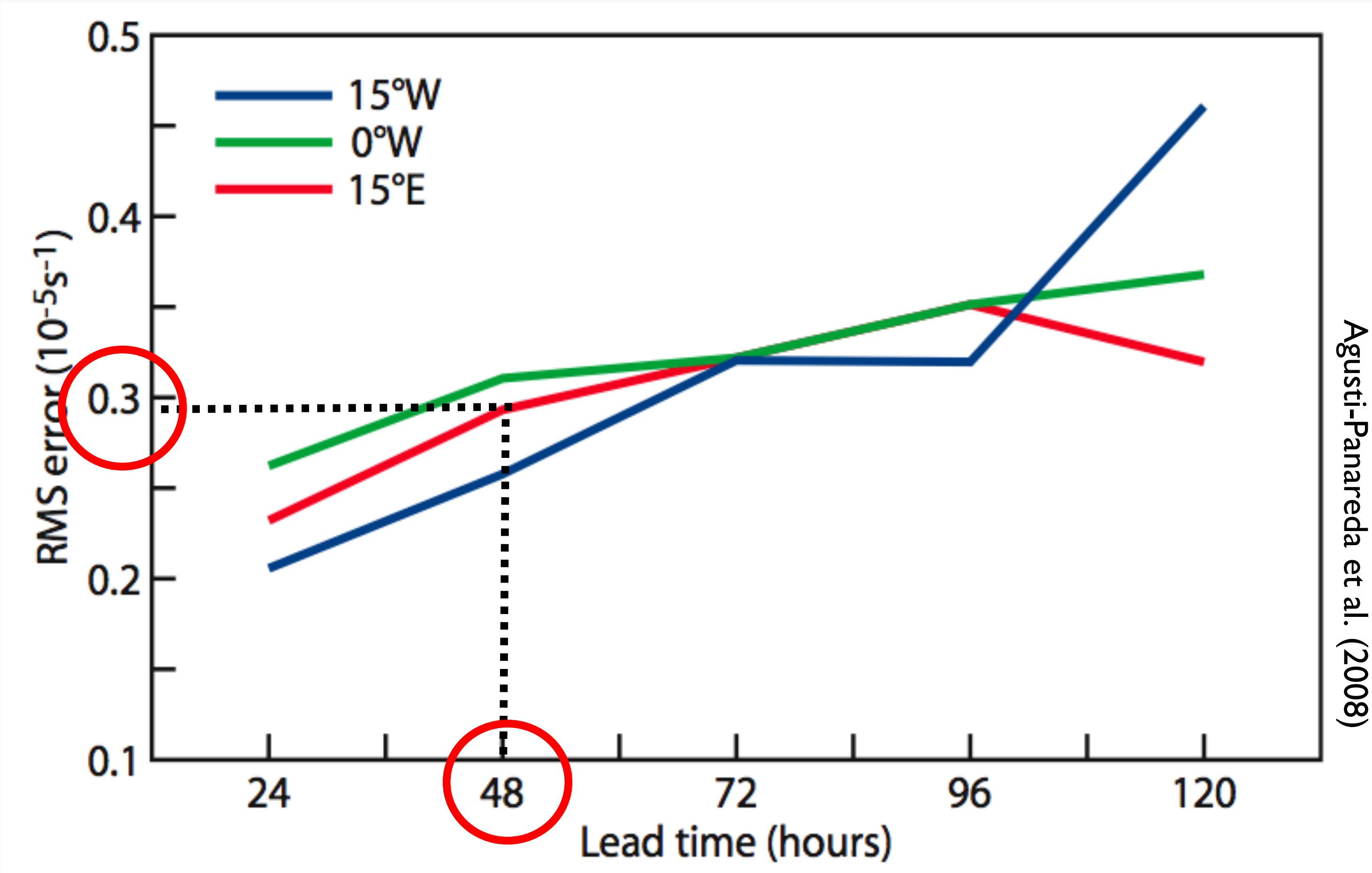
RMS error of 700 mb curvature vorticity from ECMWF forecasts at different lead times, averaged over boxes centred on 15W, 0W and 15E



Models have little skill beyond about 2 days - initial conditions and model formulations

Sensitivity to the basic state: Moist singular vectors

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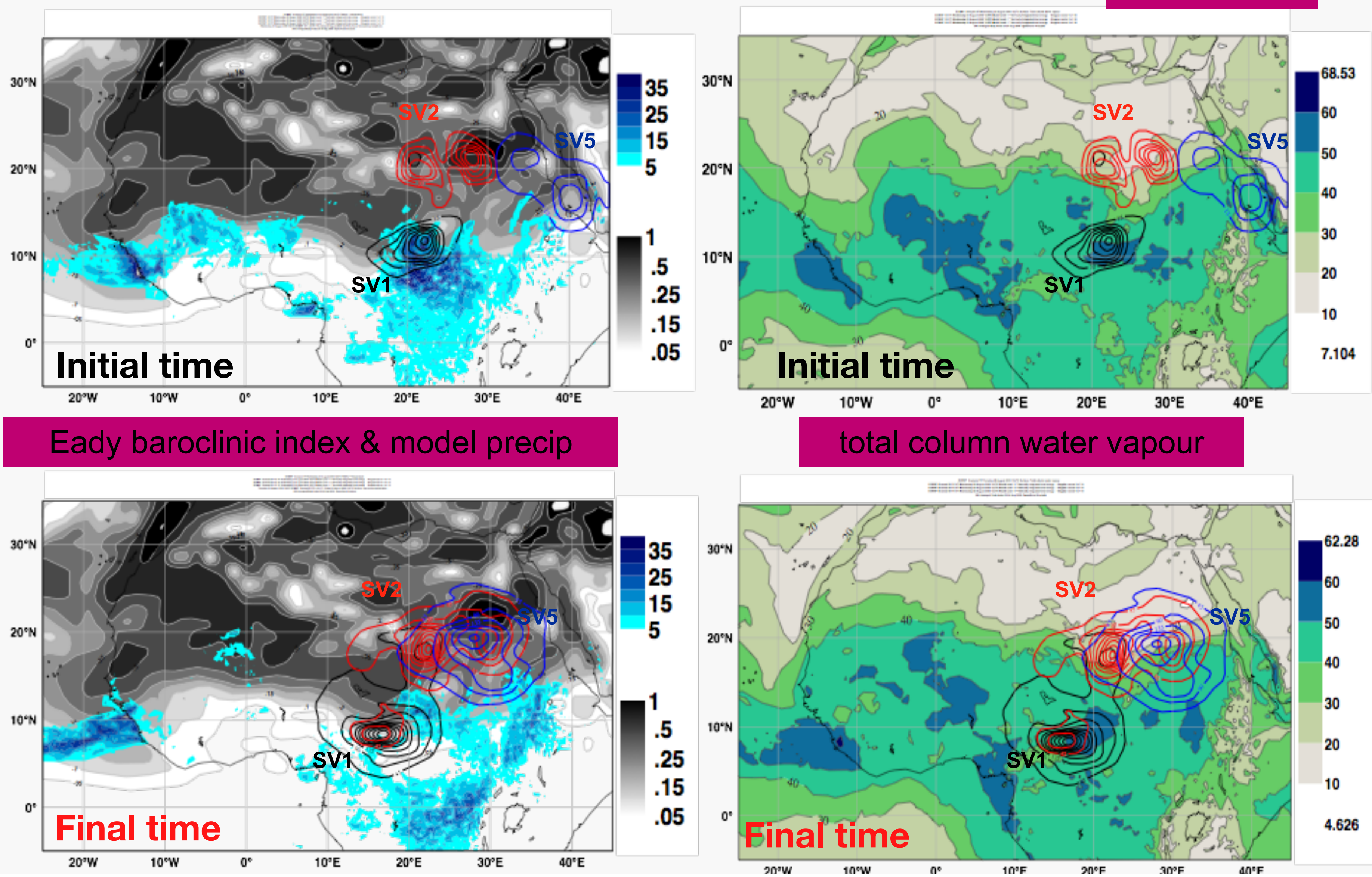
SVs 1,2,5



Eady baroclinic index & model precip

total column water vapour

Major difficulties

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- Rain across West Africa is associated with passage of AEWs
- Precipitation breaks result from lack of AEWs
- Forecast skill of these large-scale waves is surprisingly poor
  - All NWP models result in decay of waves within 2 days!

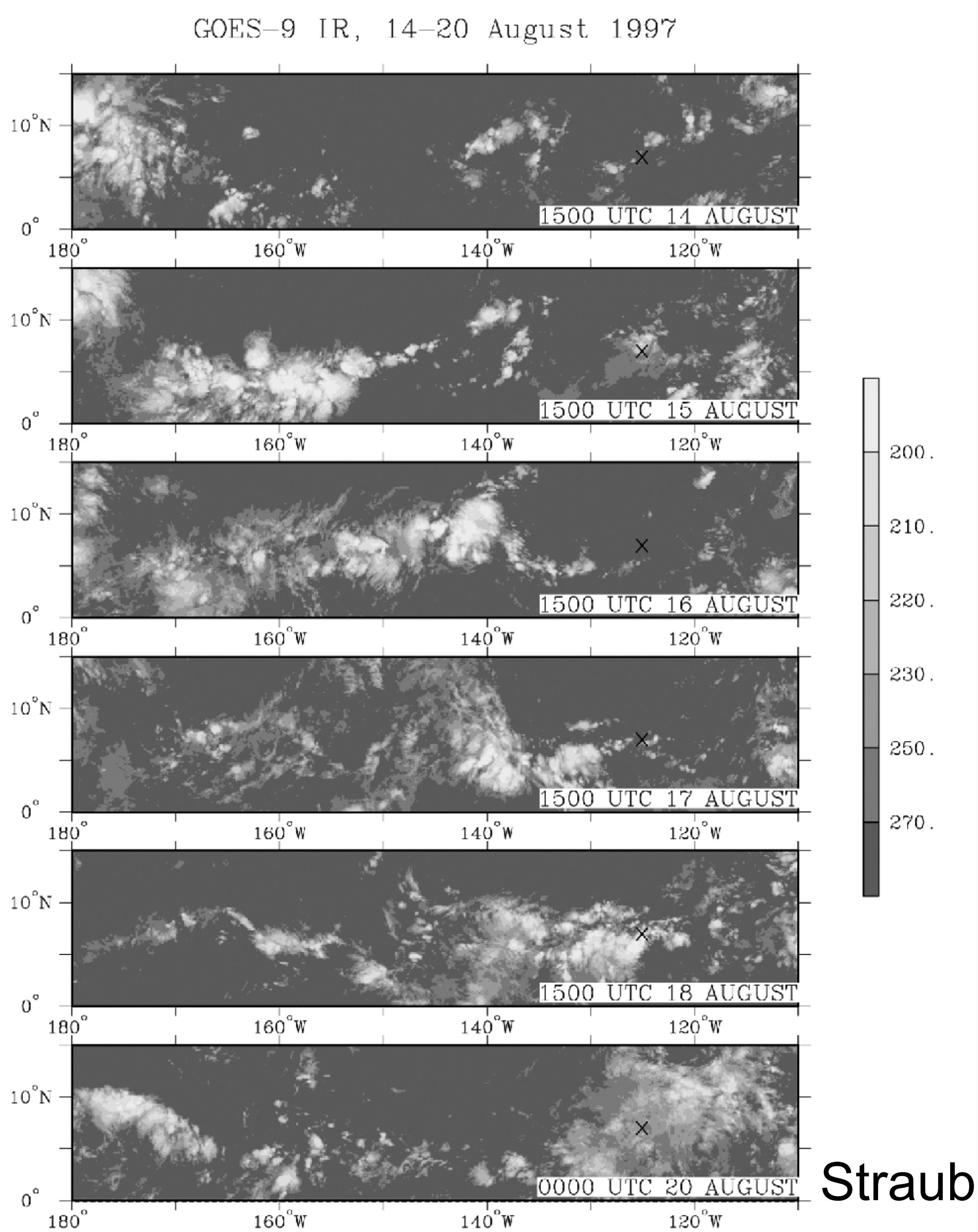
- Cannot diagnose what is wrong in models because understanding of moist influences on AEWs is lacking

- Convection (and associated precip and heating) occurs in moist air masses from South

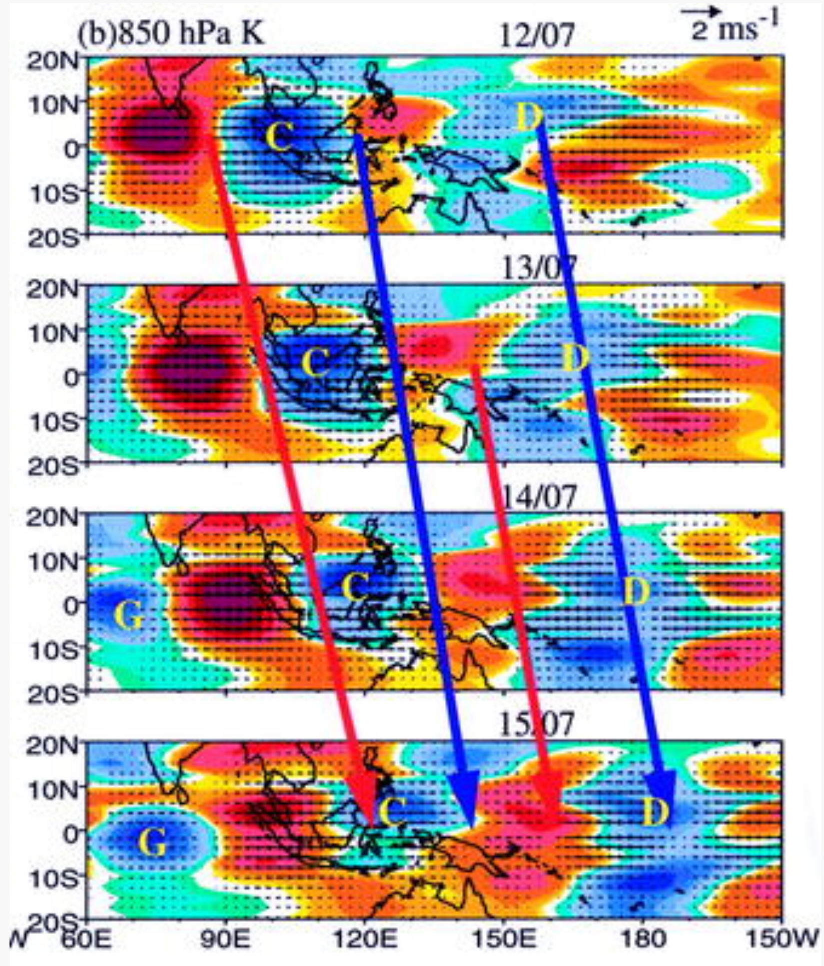
  …but this would be descending sector of dry wave!

- Clearly moist-up dry-down cannot work – new approach needed.

Equatorial Waves: Observations



Straub and Kiladis (2003)



Yang et al (2003)

Equatorial wave theory - recap

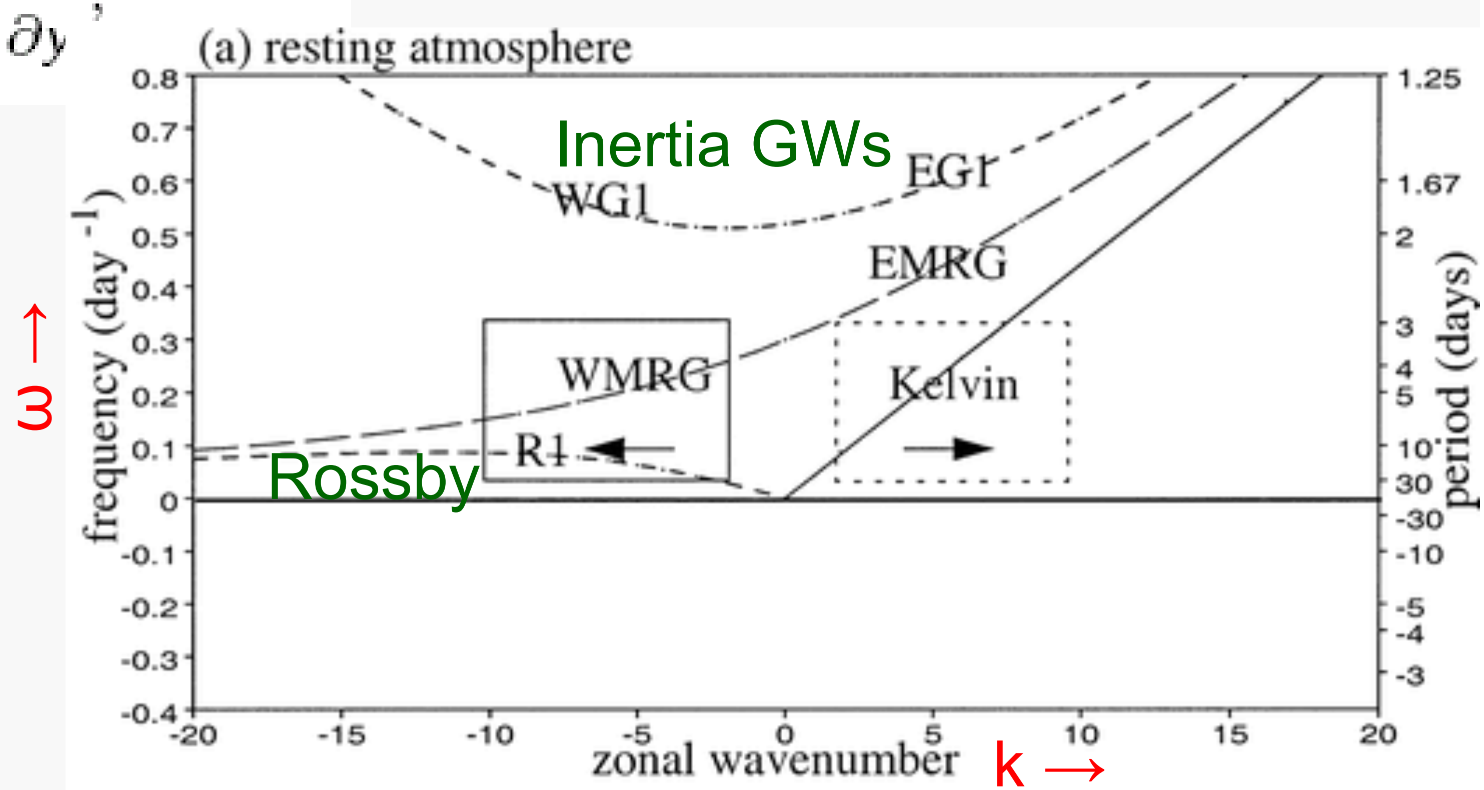
$$\frac{\partial \eta}{\partial t} + H\left(\frac{\partial u}{\partial x} + \frac{\partial v}{\partial y}\right) = 0,$$

$$\frac{\partial u}{\partial t} - fv = -g\frac{\partial \eta}{\partial x},$$

$$\frac{\partial v}{\partial t} + fu = -g\frac{\partial \eta}{\partial y},$$

Eqns linearised about a state at rest.

Solutions below for dry atmosphere.



MRG = mixed Rossby-gravity

Aka “Yanai wave”

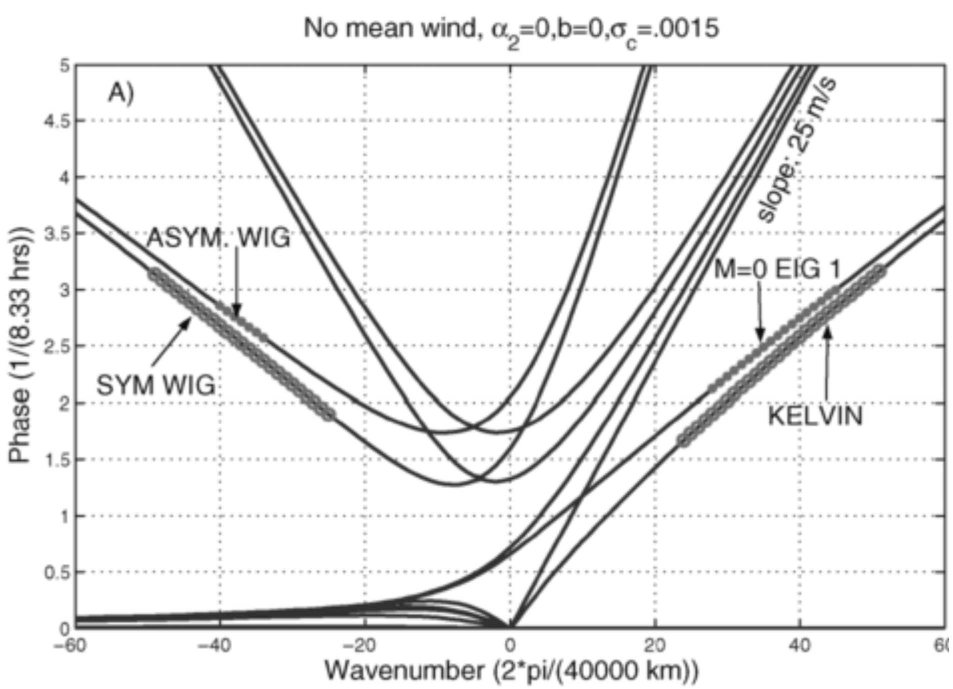
Moist equatorial waves

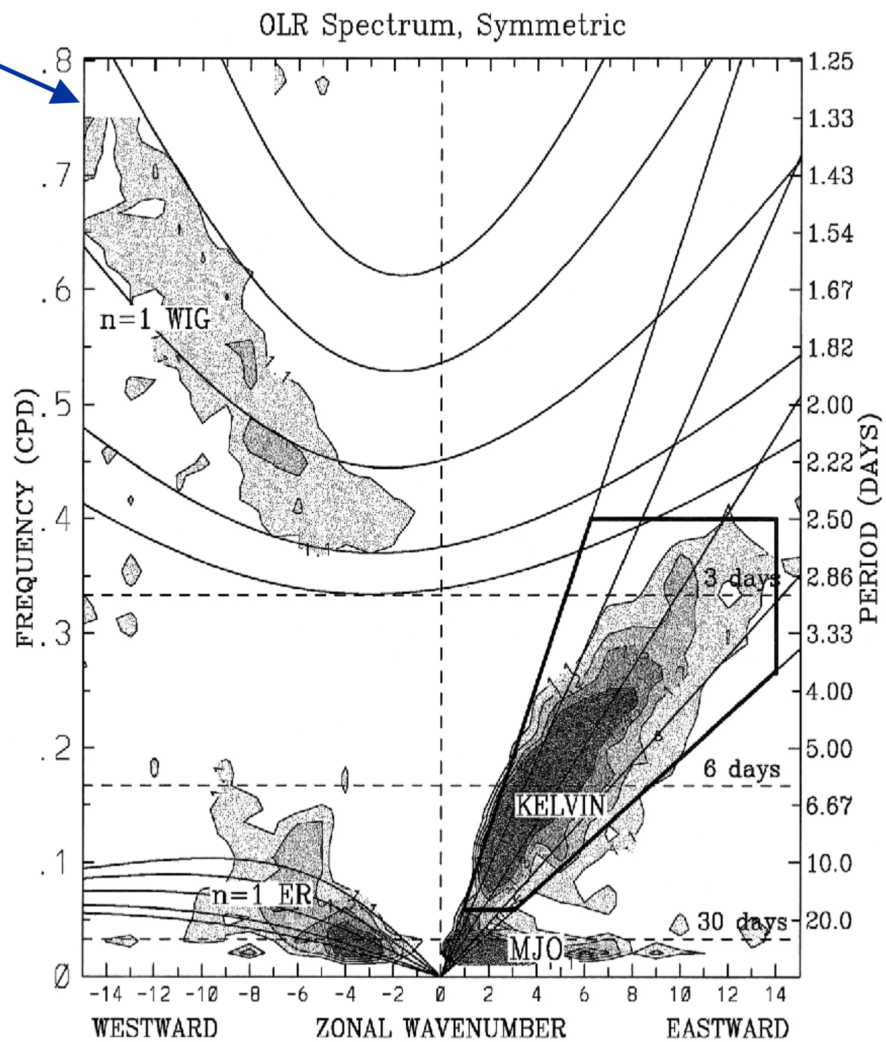
Cold cloud tops (satellite obs) Fourier analysed with dry dispersion relations overlain (equivalent depth fitted)

(Wheeler and Kiladis, *JAS, 1999)*

Semi-analytic results for moist waves

(Majda et al, *JAS, 2003)*





Latest theoretical approach

- Khouider and Majda, *JAS, 2006, 2007, 2008*
- Project equatorial β-plane primitive equations onto first two baroclinic "modes" of vertical structure

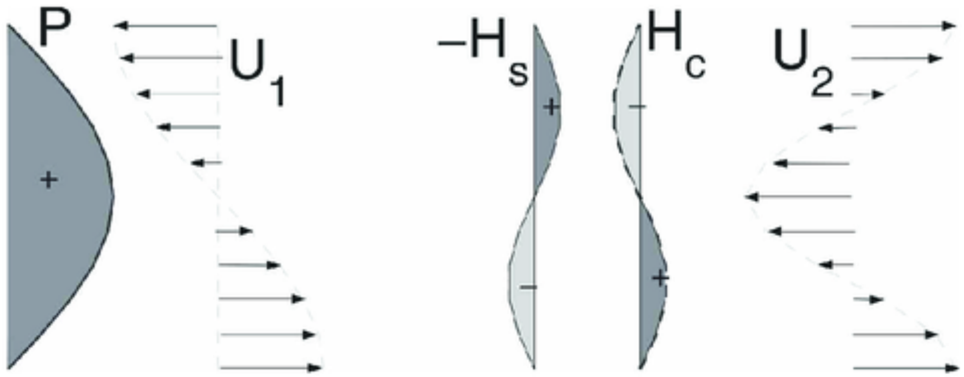


FIG. 1. Sketch of the two-baroclinic-mode model convective parameterization with three cloud types.

- Each component evolves as in shallow water eqns
- Include an evolution eqn for depth-integrated moisture which couples the components through the heating fields
- Also an evolution equation for boundary layer θ_e

Resulting modal structures

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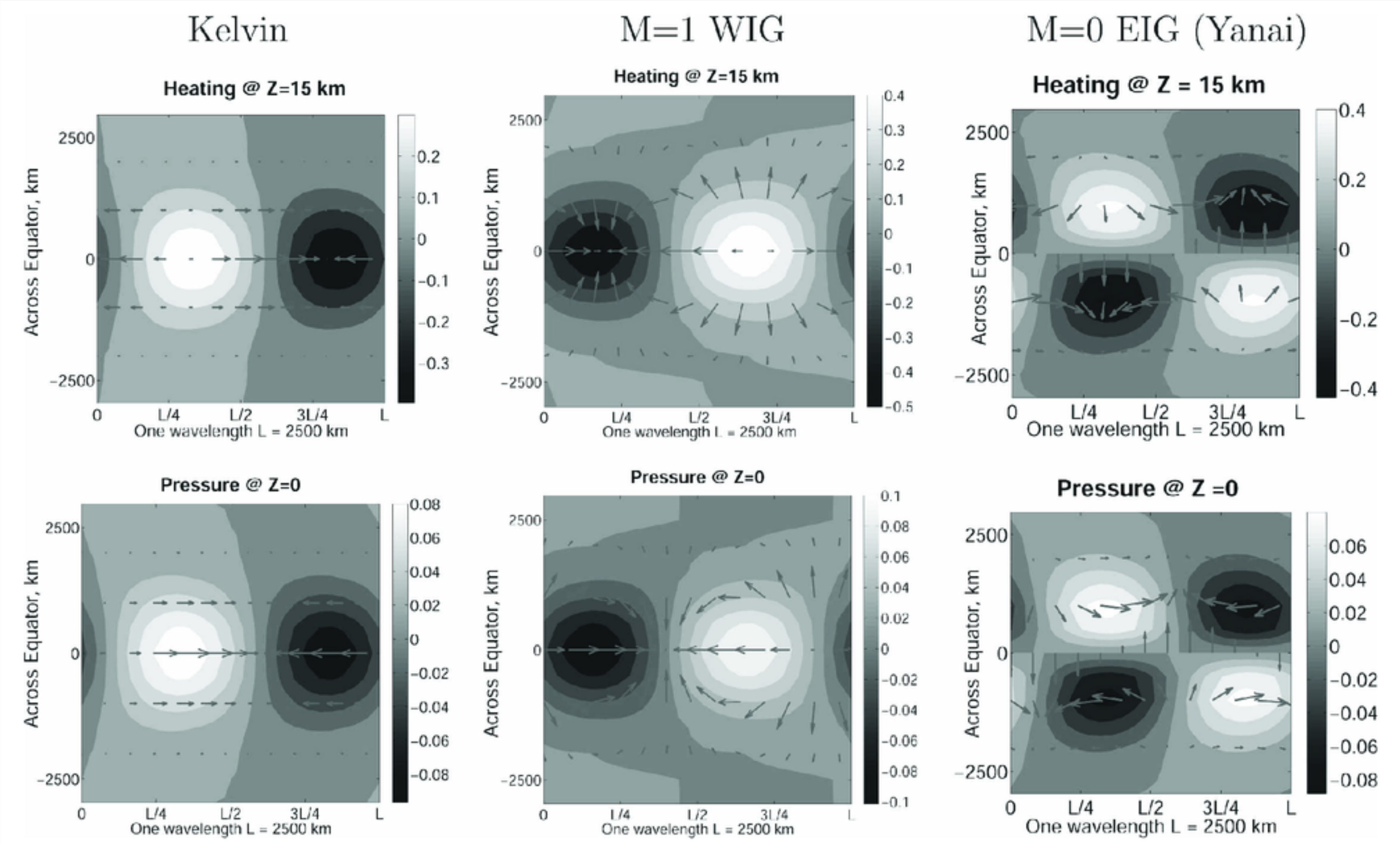
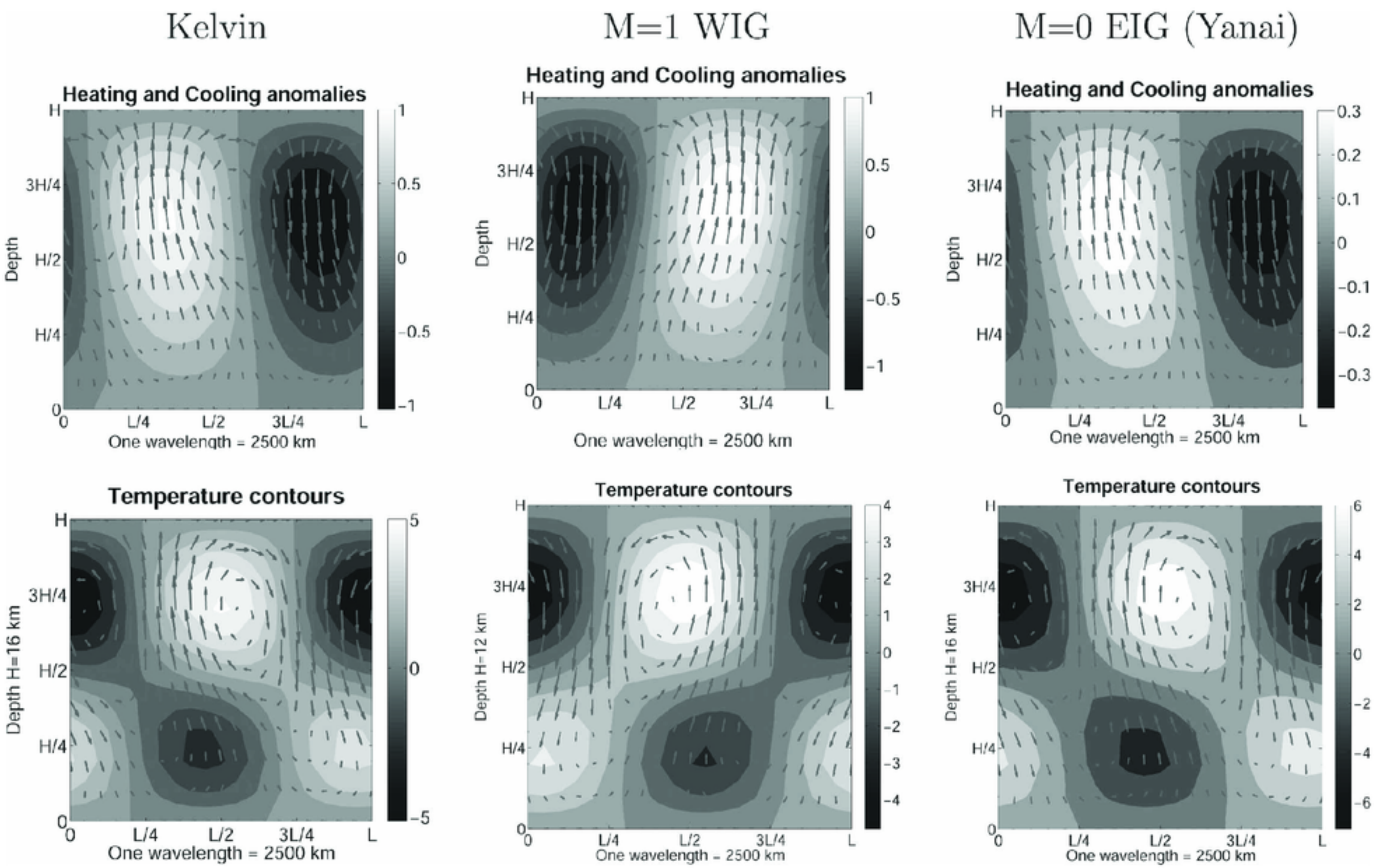


Vertical cross-sections



Mechanisms for moisture coupling

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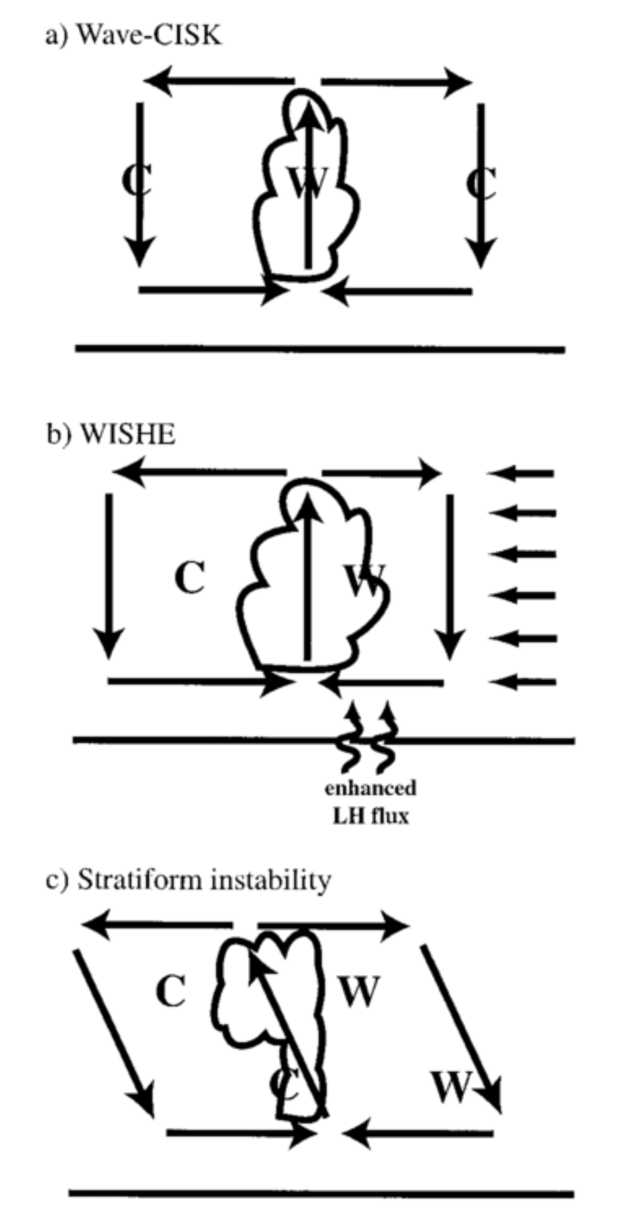
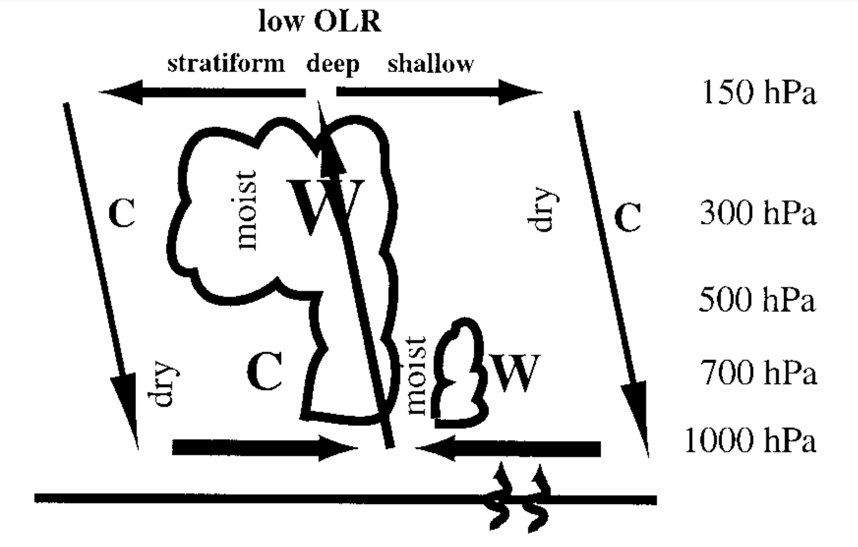
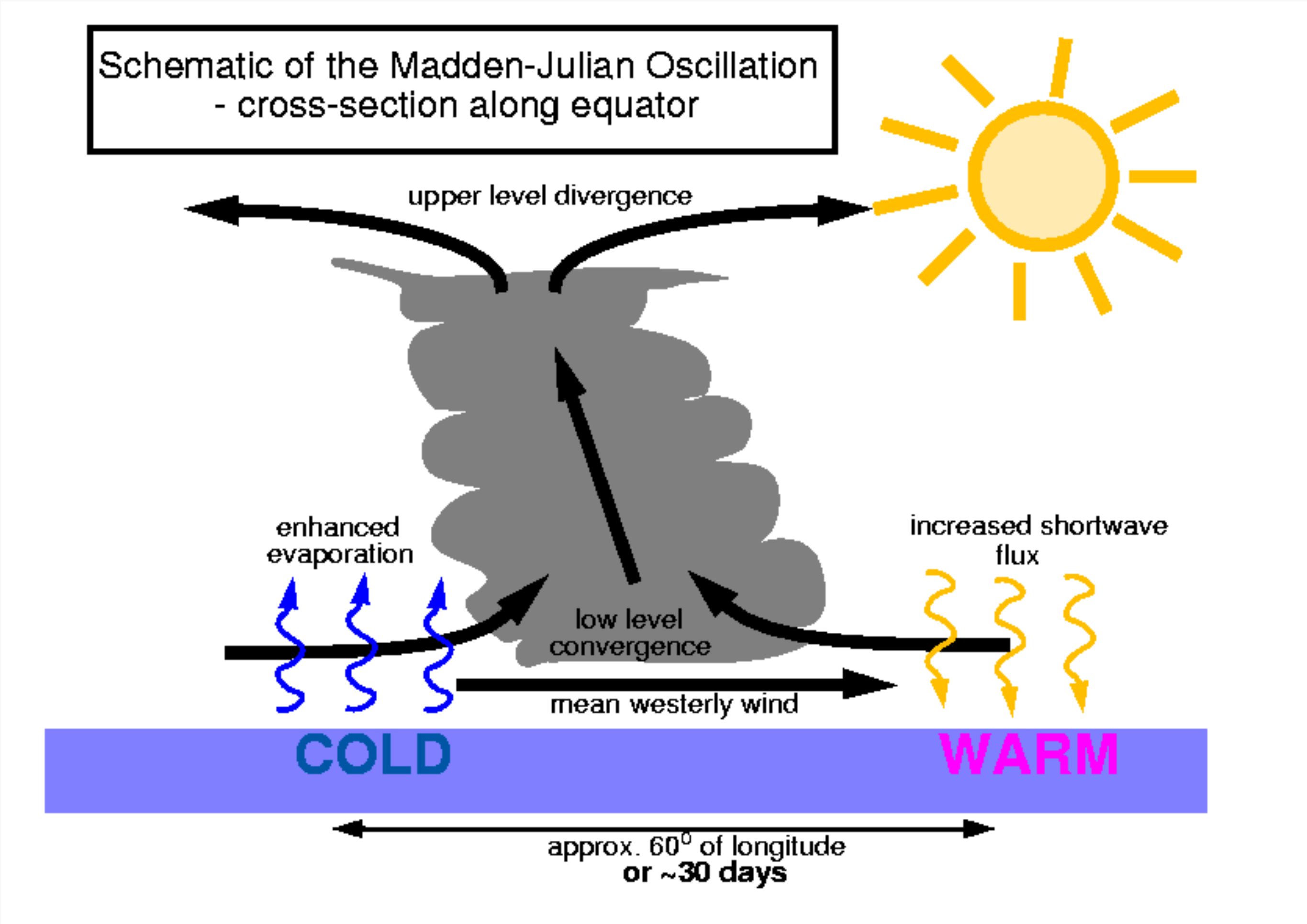


- Straub and Kiladis, *JAS, 2003*
- Argue that moist equatorial waves appear to propagate via a combination of wave-CISK and "stratiform instability" (motivated Khouider model)
- Basis is phasing between variables



Madden-Julian Oscillation



Longer timescale ⇒ coupling with ocean mixed layer temperature

Vitart and Woolnough have shown greater skill in ECMWF forecasts with coupling

5. Moisture transport within cyclones

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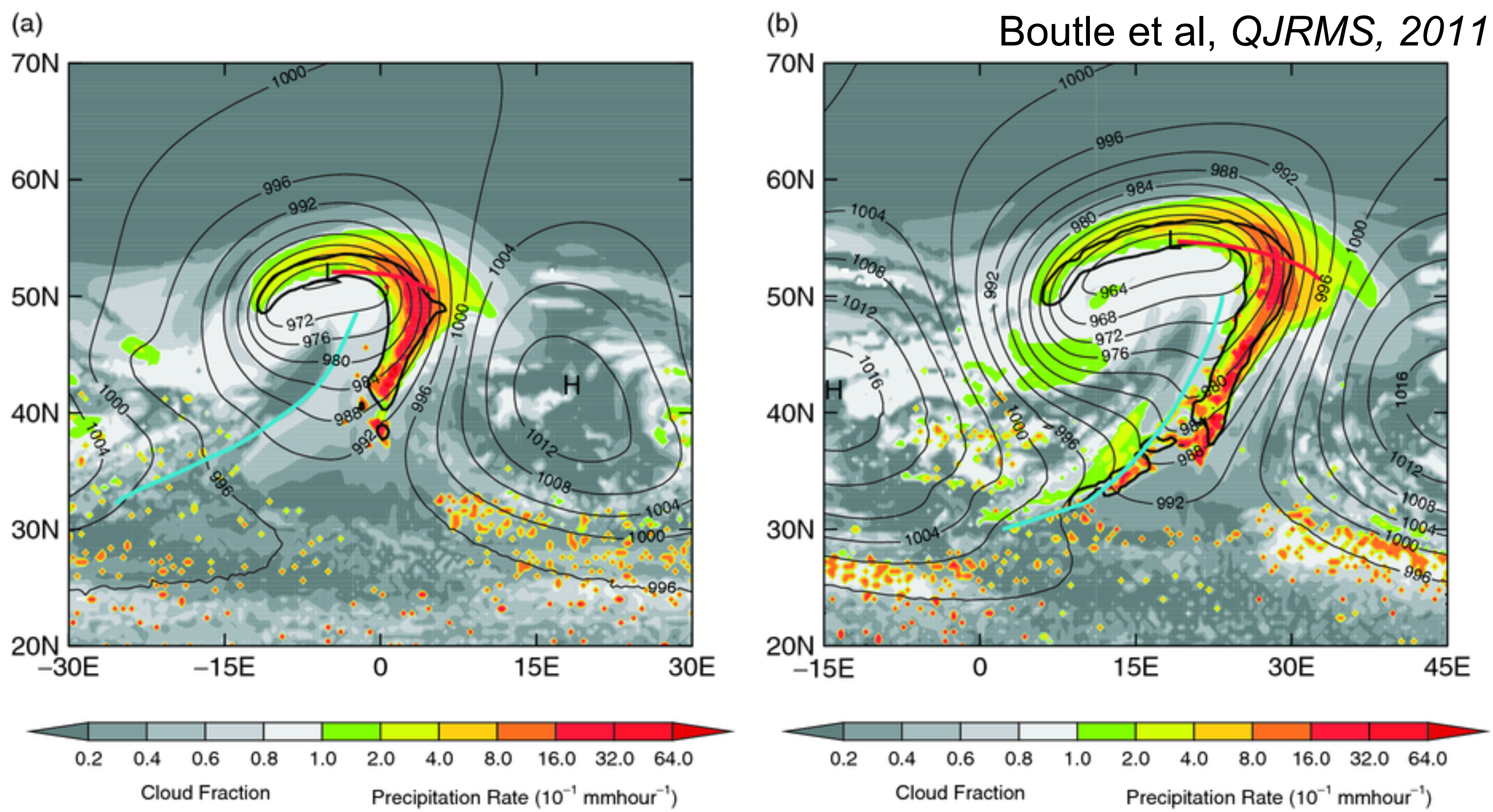
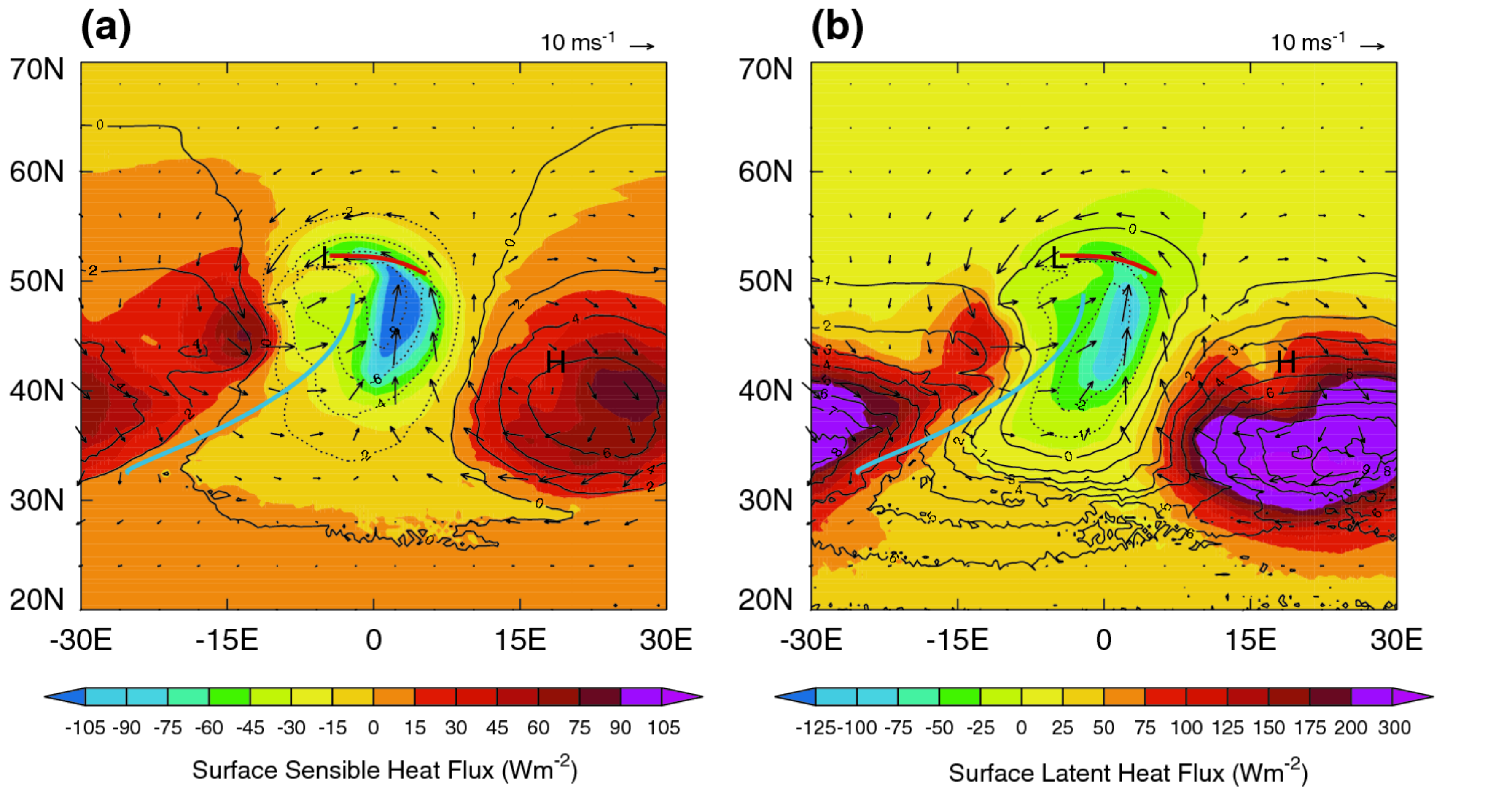
Boutle et al, *QJRMS, 2011*



Cloud and precipitation clearly in ascending parts of cyclone
+ scattered shallow convection elsewhere

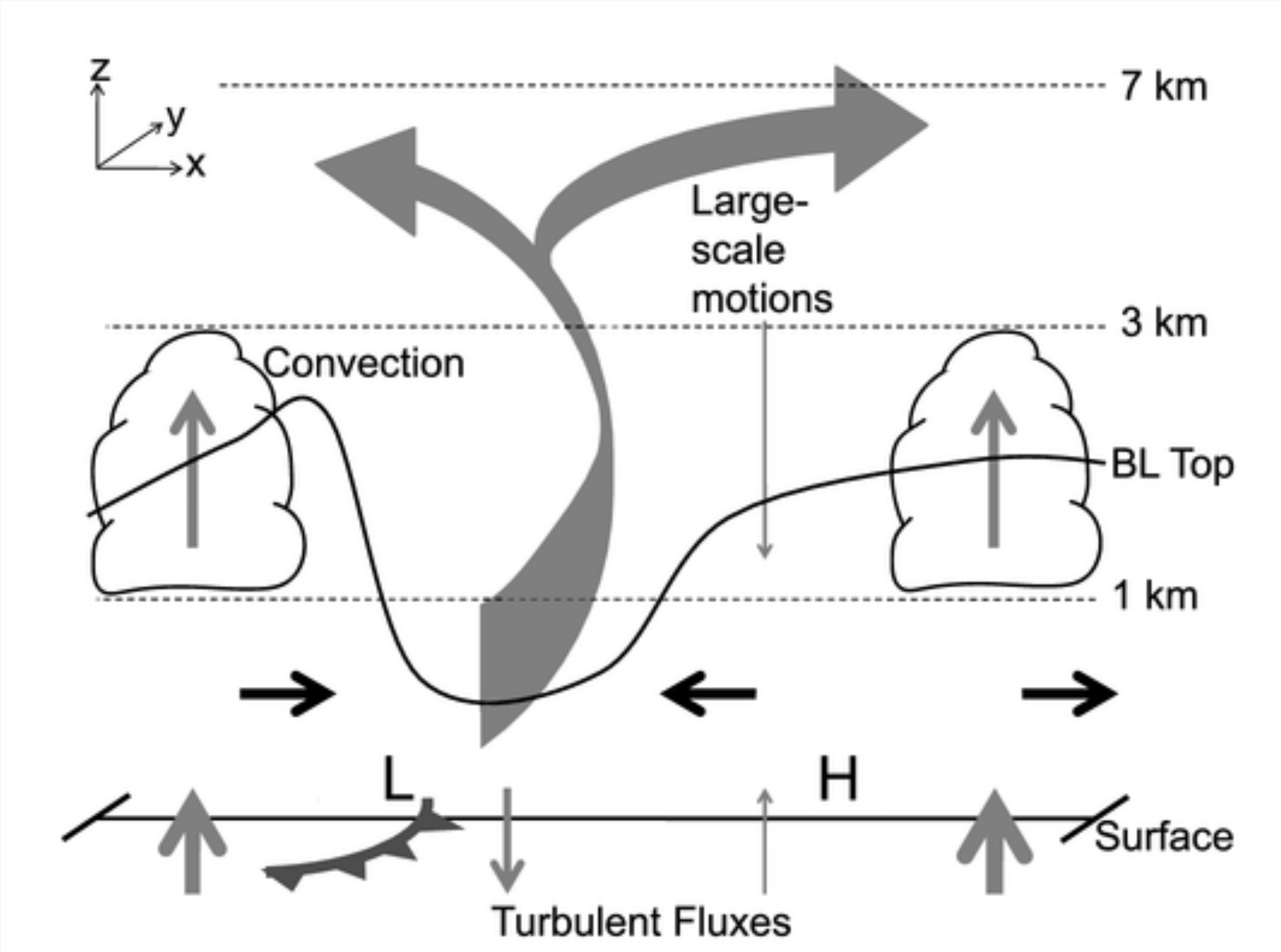
Surface heat fluxes into wave



Heating from ocean/land surface, communicated via BL turbulence.

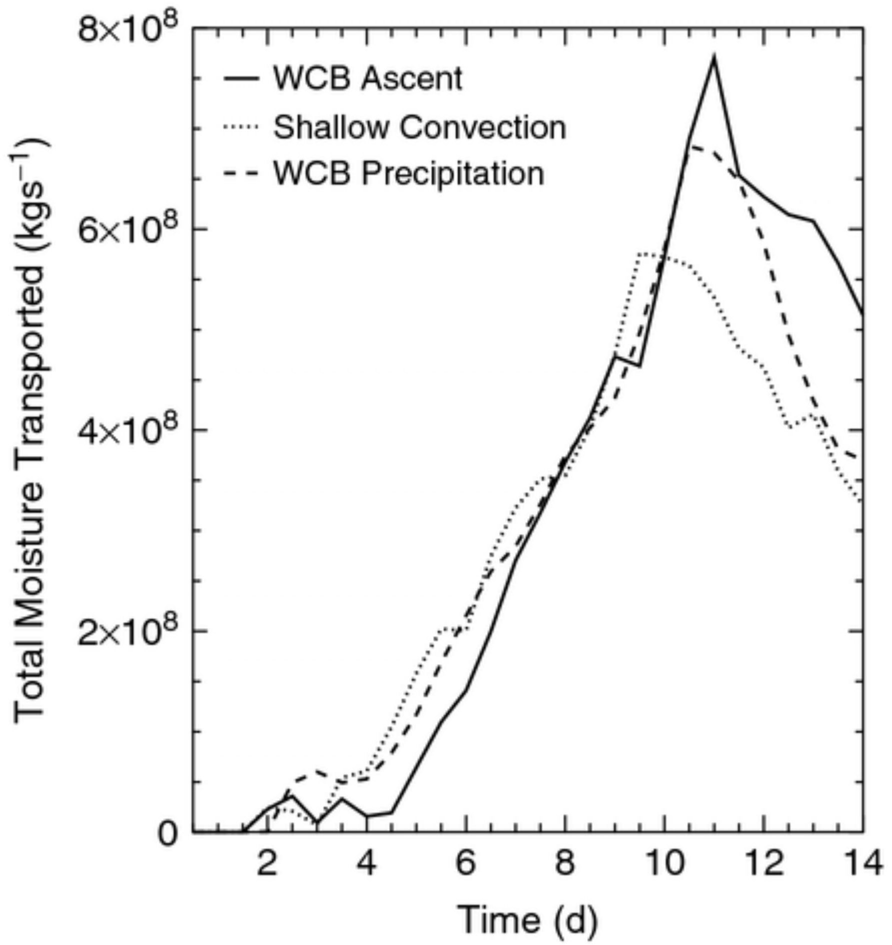
Most input into cold sector of wave – especially moisture

Schematic relating BL dynamics to the cyclone structure



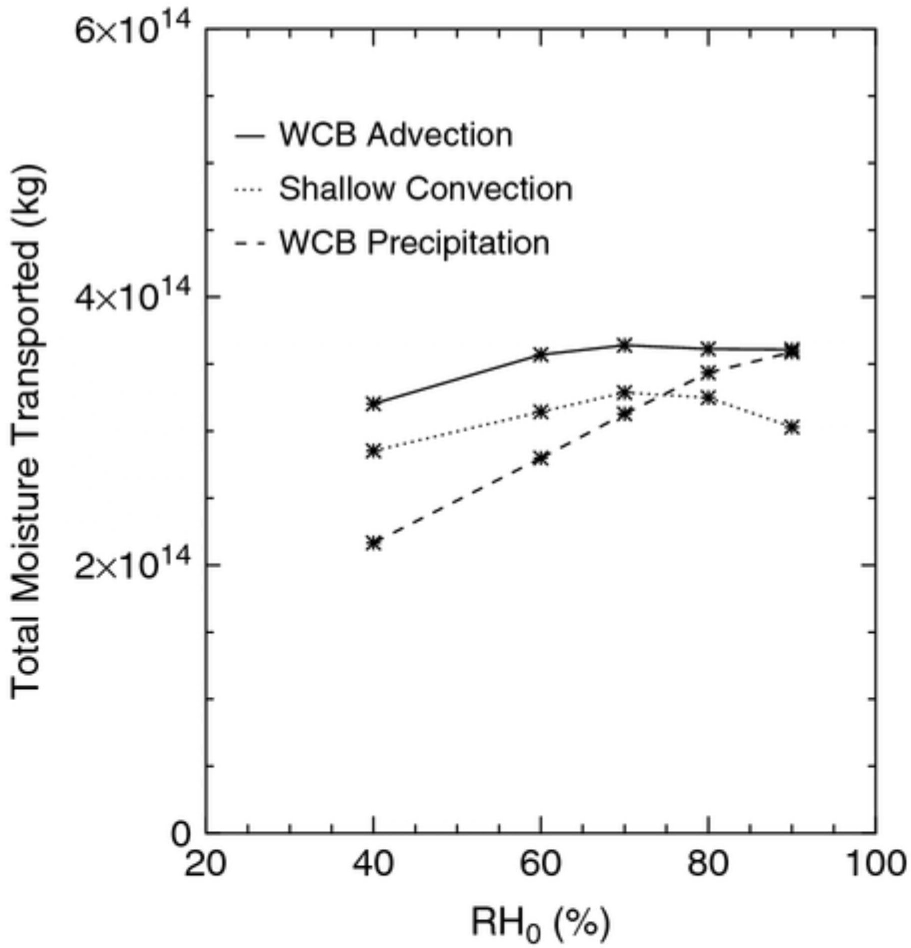
Boutle et al, *BL Dyn, 2010*

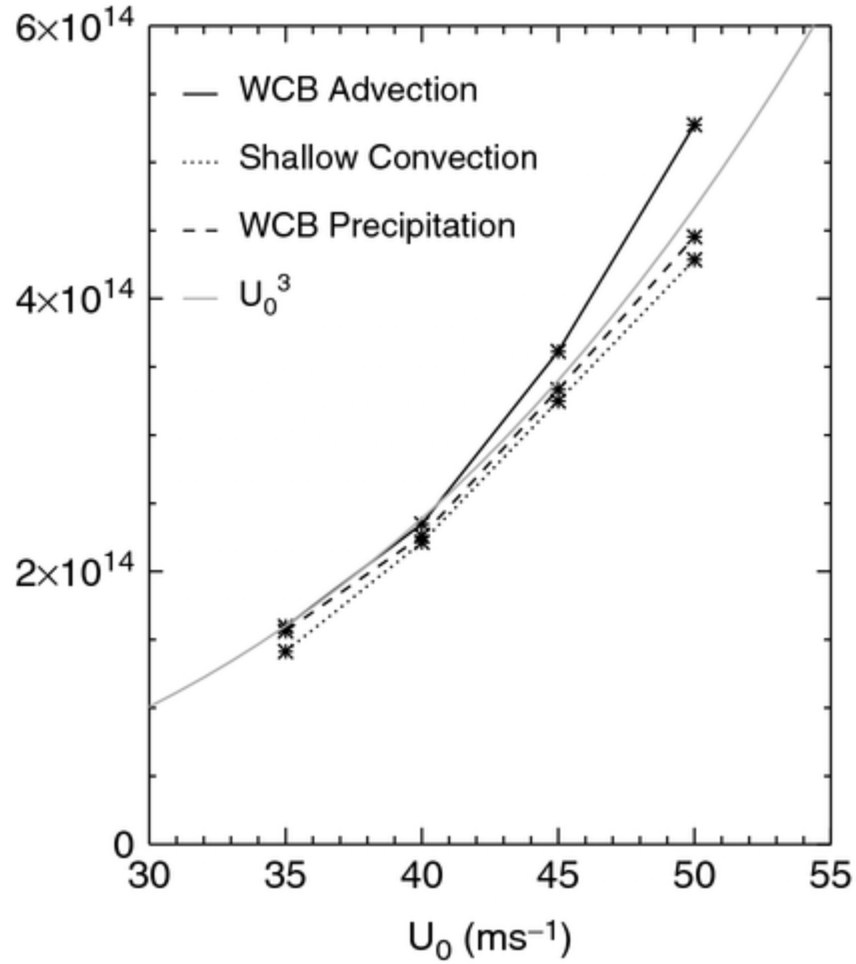
Contributions to moisture transport



- Boutle, Belcher and Plant, *QJRMS, 2011*
- Baroclinic wave life cycles using the Met Office UM
- Moisture transport by shallow convection (in cold sector) matches transport in warm conveyor belt
- WCB is efficient. Almost all moisture transported is precipitated out (ascent + condensation)

Dependence on basic state





Insensitive to initial RH.

Pick-up of moisture within BL is rapid.

Moisture transport by both routes scales with shear of jet

Scaling argument with moisture

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- Growth rate of cyclone and ascent is enhanced by latent heat release but not a great deal
- Due to small scale height of saturation specific humidity (Whitaker and Davis*, JAS, 1994;* De Vries *et al, JAS, 2010)*

$$w \sim 0.5 \frac{f\Lambda}{N}\left(\frac{v_g}{N}\right) \qquad \frac{g}{\theta_0}\frac{\partial\theta}{\partial y} \approx -f\Lambda \qquad r = \frac{0.622 e_s(T) RH}{p}$$

- Boutle *et al (2011)* argue *r ~ Λ* because moisture of air transported from Subtropics is greater for stronger Δθ

- Net result $r'w' \sim \Lambda^3$

Summary of lecture

Transport of moisture and moist scaling

↑

Coupling waves and shear instability

↑

Moist wave phenomena:

Baroclinic and equatorial waves

↑

Convective instability

↑

Variables for saturated atmospheres

↑

Diabatic processes

7. Active research areas

- Moist waves in tropics
  - Need to incorporate effects of shear flow
  - Why do NWP models fail to maintain wave propagation?
- Stormtracks and changes with climate
  - Need better understanding of link with moisture transport
- Diagnosis of model error
  - Use PV tracers as markers of the effects of processes within a model
- High resolution ensemble forecasts
  - Linking mesoscale predictability to skill in precipitation forecasts
  - Limits to predictability imposed by active convection
  - Assimilation of moisture in a balanced fashion?