Numerical Weather Prediction
Parametrization of diabatic processes

Convection I: an overview

Peter Bechtold

Convection

- Lectures:
  - An overview (only about 5 simple principles to remember)
  - Parametrisation of convection
  - The ECMWF mass-flux parametrisation and Tracer transport
  - Forecasting of Convection
- Exercises

Convection

- Aim of Lectures:

The aim of the lecture is only to give a rough overview of convection in the context of the general circulation. The student is not expected to be able to directly write a new convection code- the development and full validation of a new convection scheme takes years. The best exercise is to start with an existing code, run some offline examples on Soundings and dig in line by line ….. The trend is toward explicit representation of deep convection in limited area NWP (no need for parameterization), but for global we are not there yet, and will need parameterizations for the next decade

- Offline convection Code:

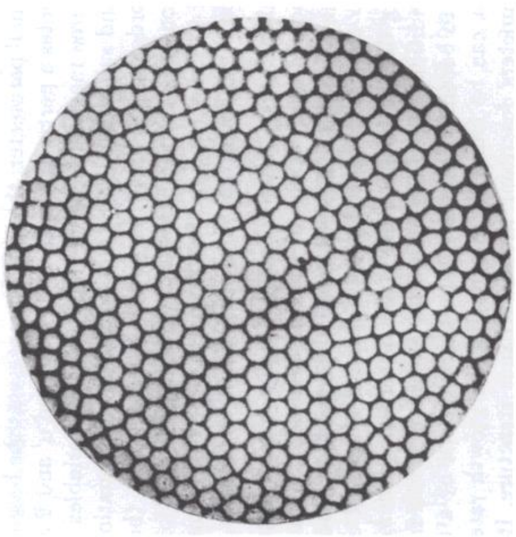
Can be obtained from peter.bechtold@ecmwf.int

Convection Parametrisation and Dynamics - Text Books

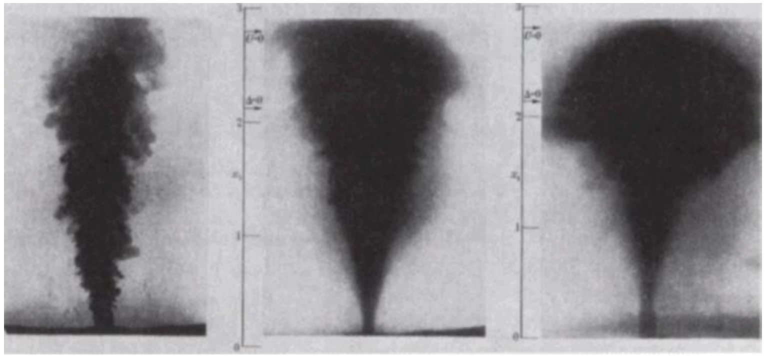
- Emanuel, 1994: Atmospheric convection, *OUP*
- Houze R., 1993: Coud dynamics, *AP*
- Holton, 2004: An introduction to Dynamic Meteorology, *AP*
- Bluestein, 1993: Synoptic-Dynamic meteorology in midlatitudes, Vol II. *OUP*
- Peixoto and Ort, 1992: The physics of climate. *American Institute of Physics*
- Emanuel and Raymond, 1993: The representation of cumulus convection in numerical models. *AMS Meteor*. *Monogr*.
- Smith, 1997: The physics and parametrization of moist atmospheric convection. *Kluwer*
- Dufour et v. Mieghem: Thermodynamique de l'Atmosphère, 1975: *Institut Royal météorologique de Belgique*
- Anbaum, 2010: Thermal Physics of the atmosphere. J *Wiley Publishers*

AP=Academic Press; OUP=Oxford University Press

Convection=heat the bottom&cool the top



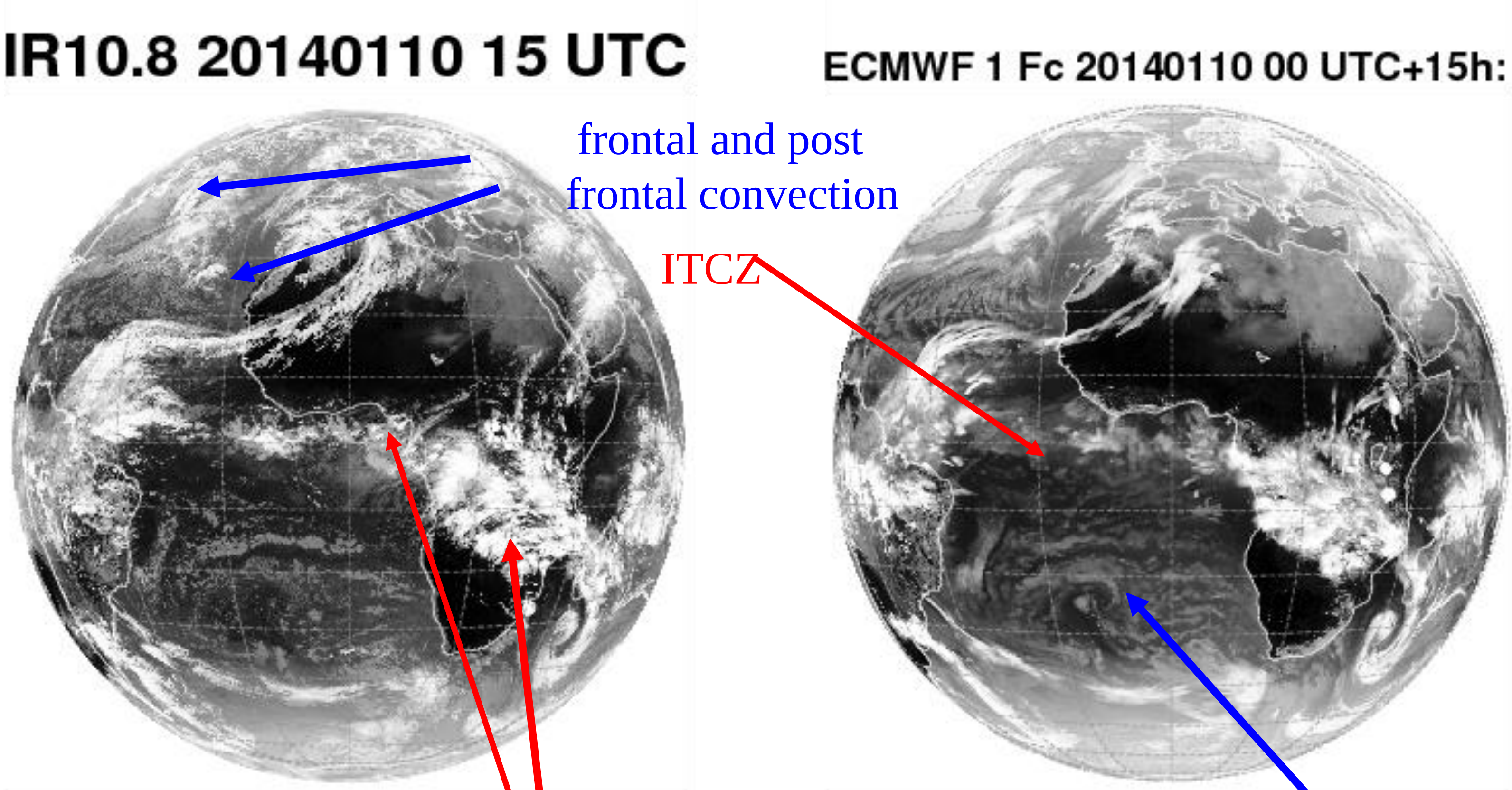
Rayleigh-Benard cellular convection



Classic plume experiment

Pre-frontal deep convection July 2010 near Baden-Baden Germany

Moist convection : Global



African Squall lines
diurnal cycle

Sc convection

Outline

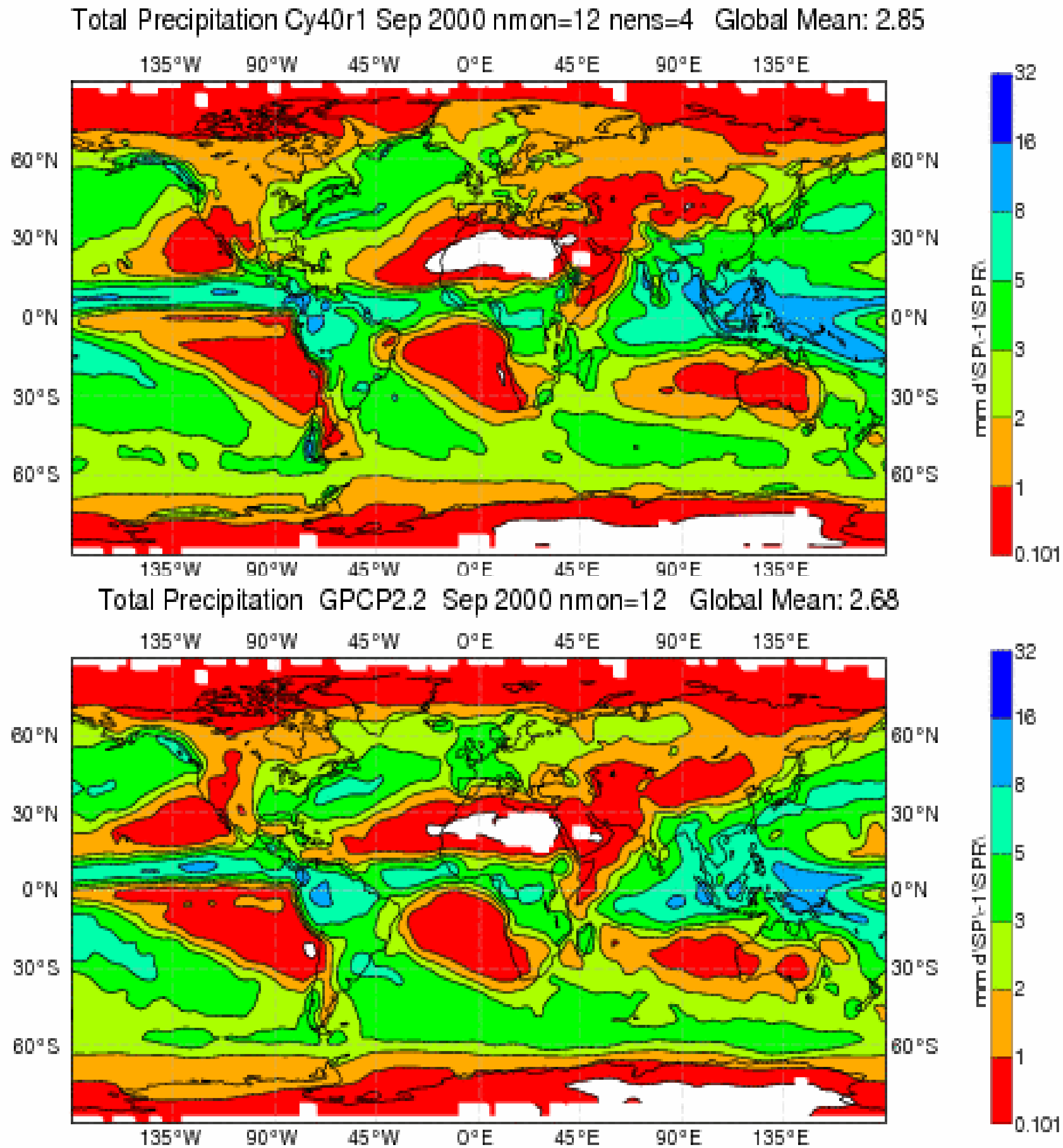
General:

- Convection and tropical circulations
- Tropical waves
- Middle latitude Convection

Useful concepts and tools:

- Buoyancy
- Convective Available Potential Energy
- Soundings and thermodynamic diagrams
- Convective quasi-equilibrium
- Large-scale observational budgets

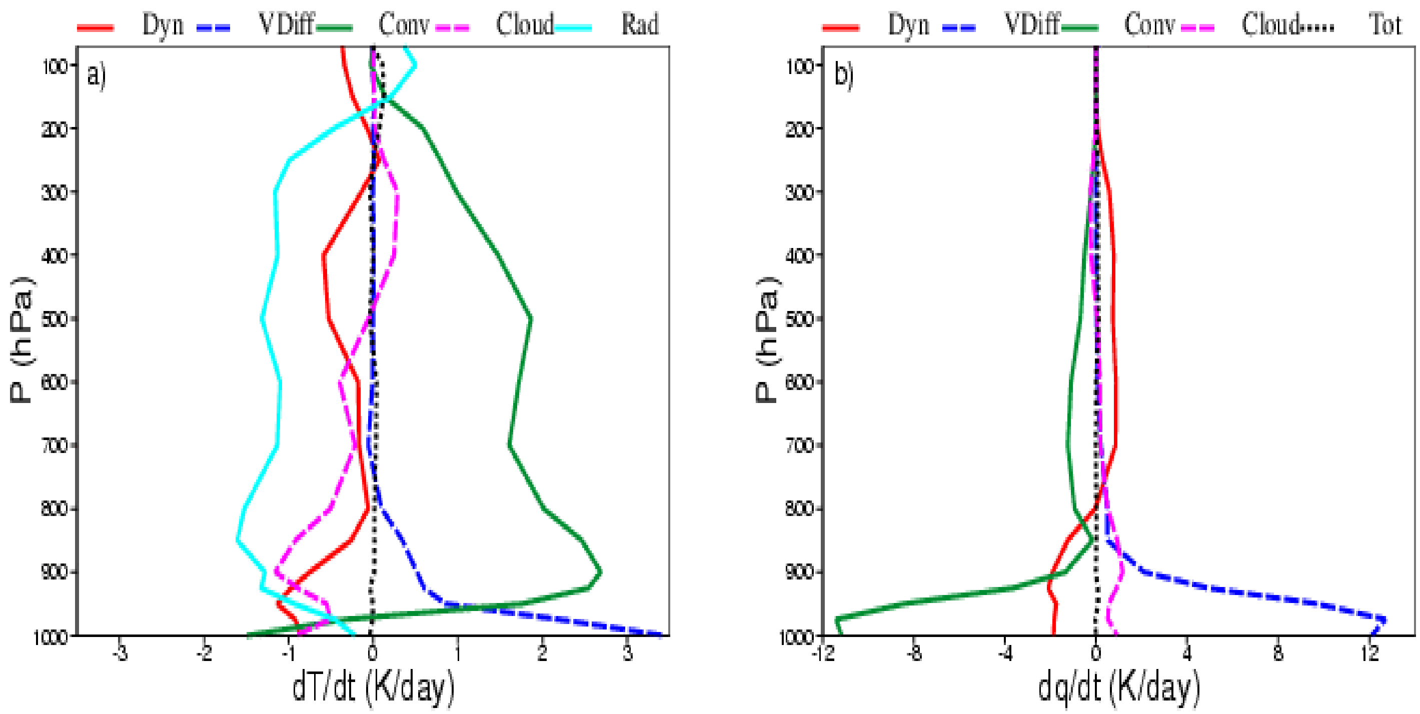
It's raining again… 2000/2001 annual precipitation rate from IFS Cy40r1 (2014) GPCP2.2 dataset



about 2.7-2.8 mm/day is falling globally, but most i.e. 5-7 mm/day in the Tropics

ECMWF

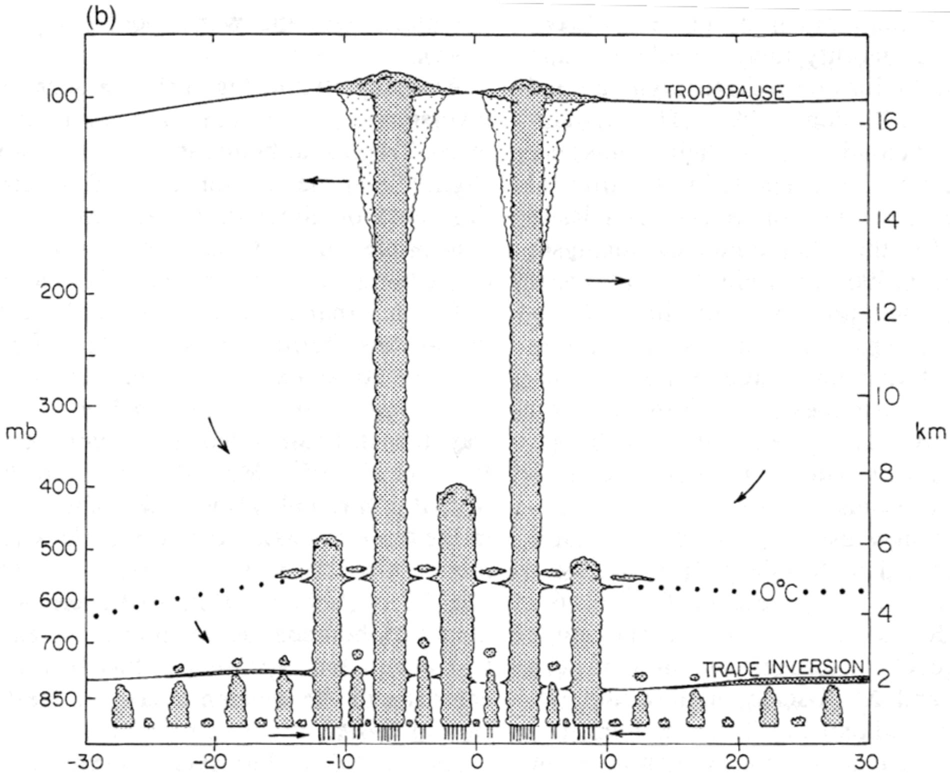
Model Tendencies – Tropical Equilibria

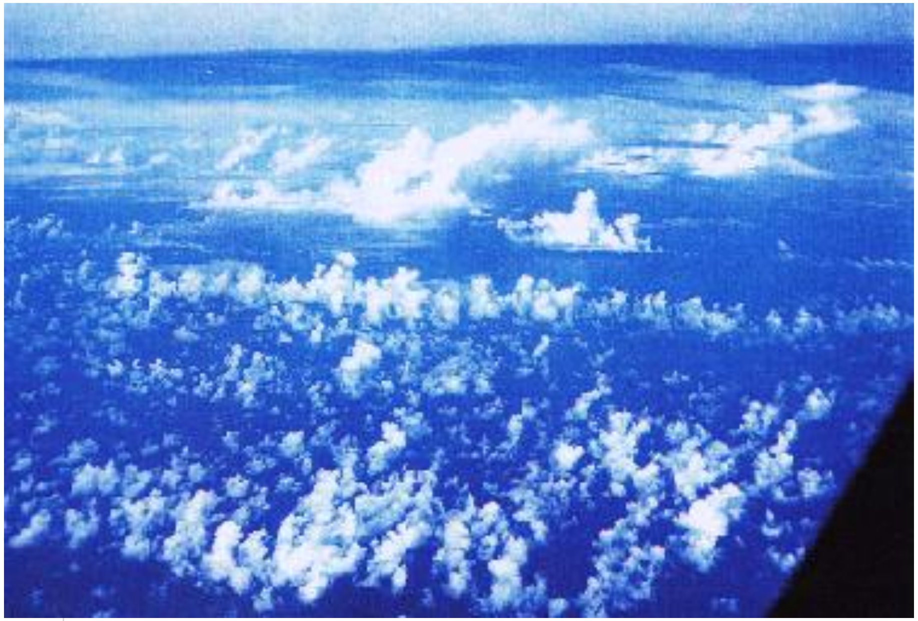


Nevertheless, the driving force for atmospheric dynamics and convection is the radiation

Above the boundary layer, for Temperature there is on average radiative-convective equilibrium; and convective-dynamic equilibrium over the large-scale disturbance, whereas for moisture there is roughly an equilibrium between dynamical transport (moistening) and convective drying. - *Global Budgets are very* similar

Distribution of convective clouds



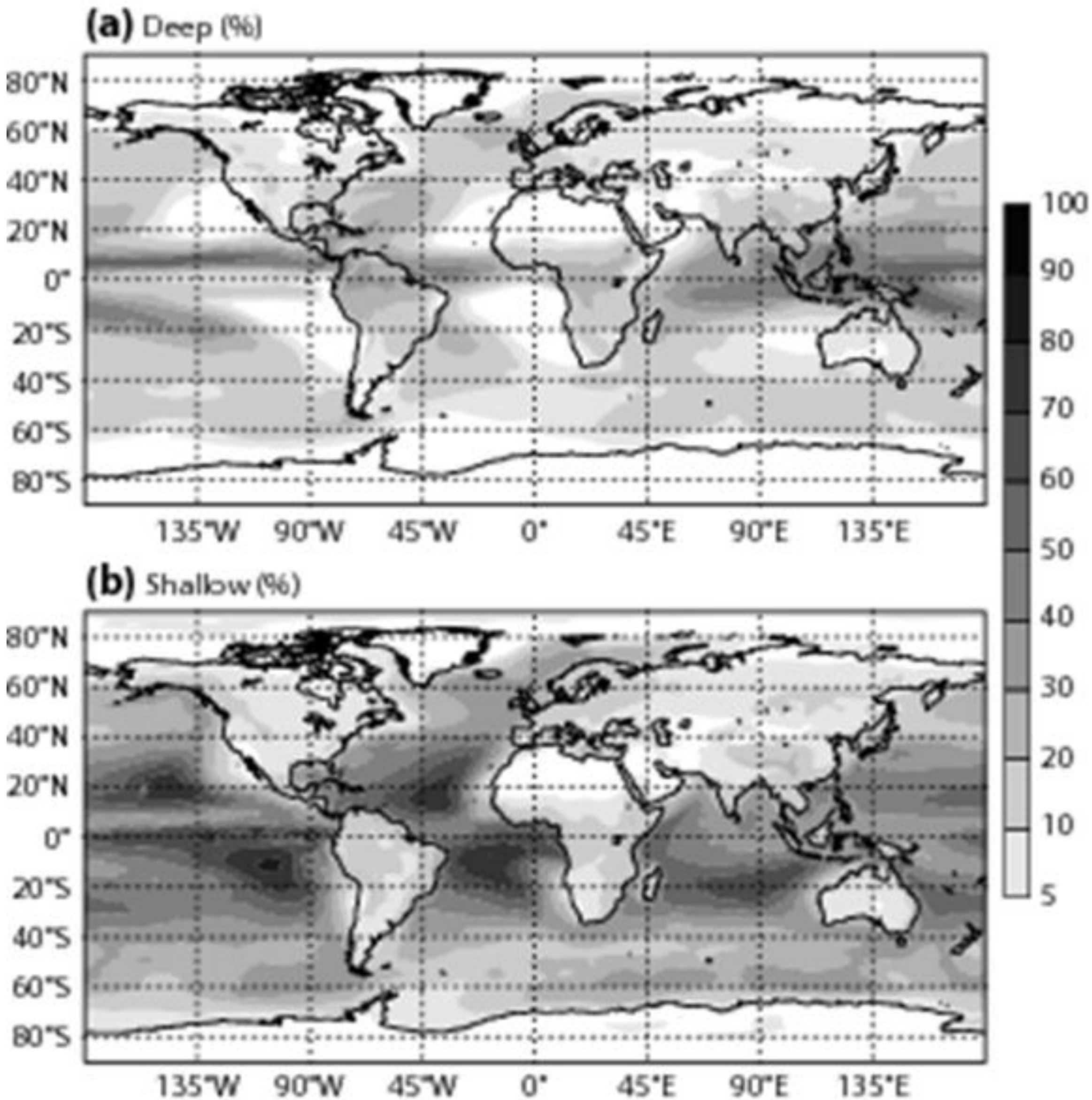


Johnson et al., 1999, JCL

Tri-modal: Shallow cumulus, Congestus attaining the melting level, Deep penetrating convection

Distribution of deep and shallow IFS Cy40r1 (2014)

Deep type including congestus



Convection and tropical circulations (1)

ITCZ and the Hadley meridional circulation: the role of trade-wind cumuli and deep tropical towers

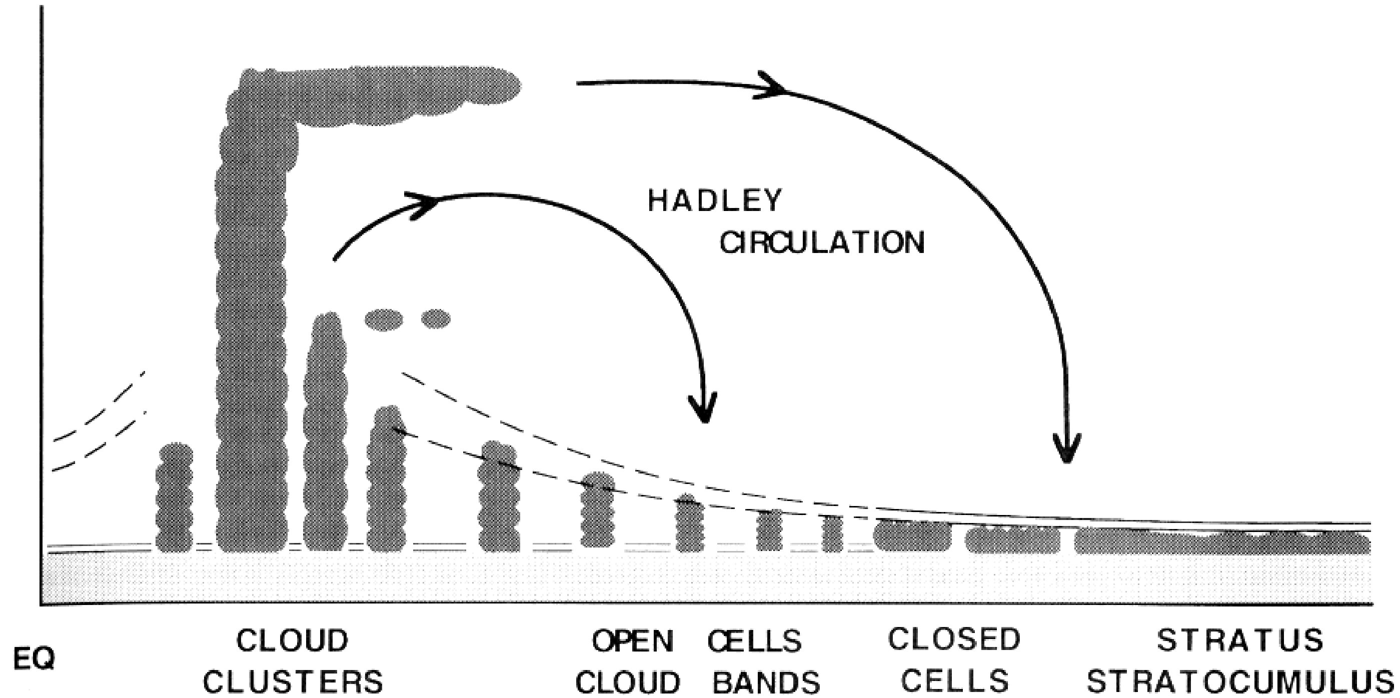
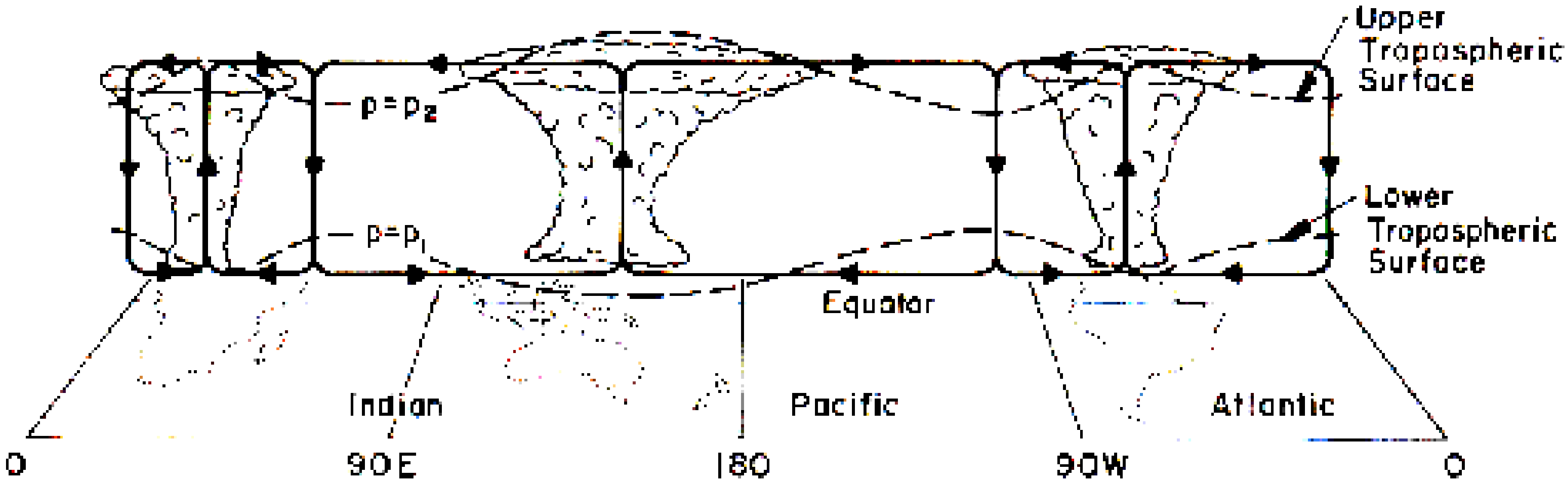


Figure 13 Schematic NE–SW cross section over the northeastern Pacific, summarizing typical observed cloud regimes. From right to left, the sea surface temperature increases and subsidence decreases. The stippled area is the PBL, the top of which is shown by the continuous and discontinuous double-stroked lines. The dashed lines above the cumulus clouds show an inversion layer, which is principally the trade wind inversion. (Redrawn from Arakawa, 1975.)

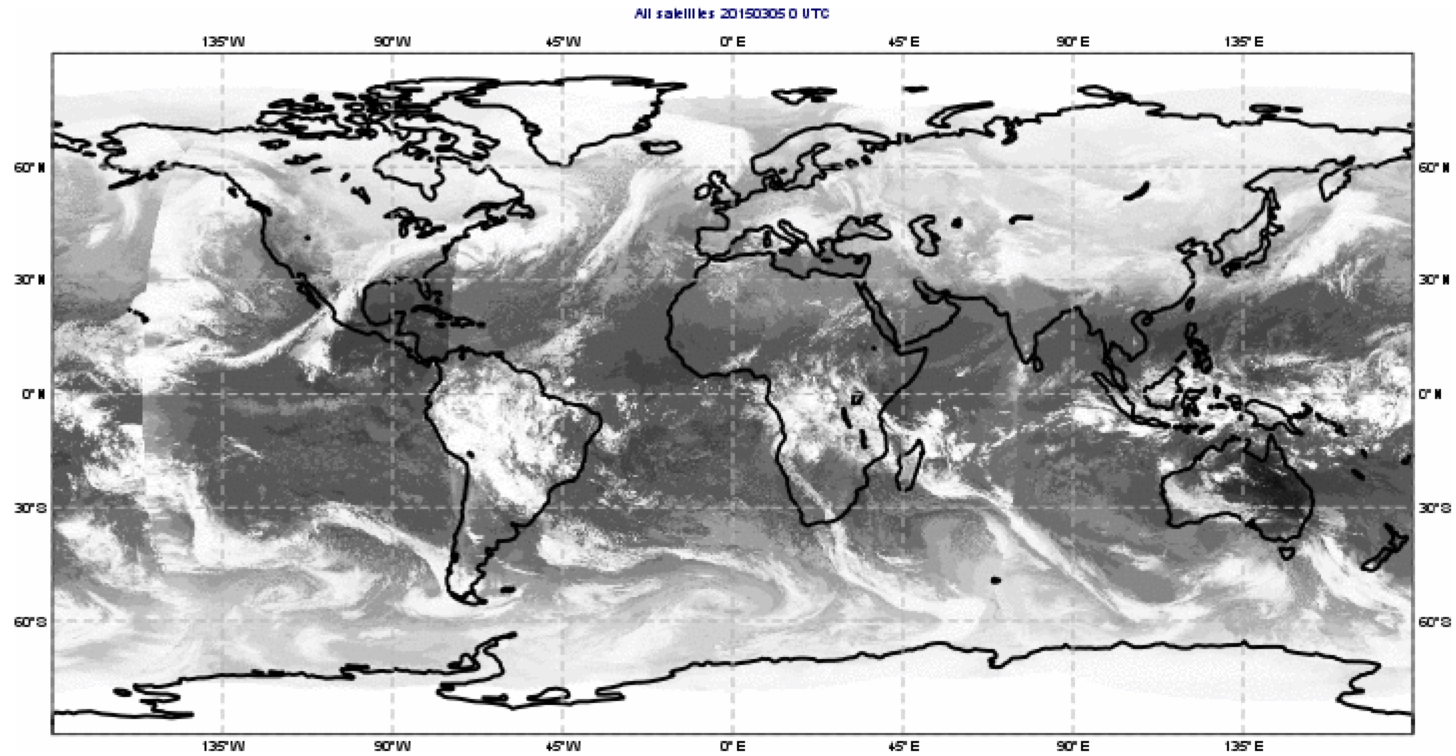
Convection and tropical circulations (2)

The Walker zonal Circulation



From Salby (1996)

Rossby, Kelvin, MJO and African easterly Waves



Analytical: solve shallow water equations (see Lecture Note)

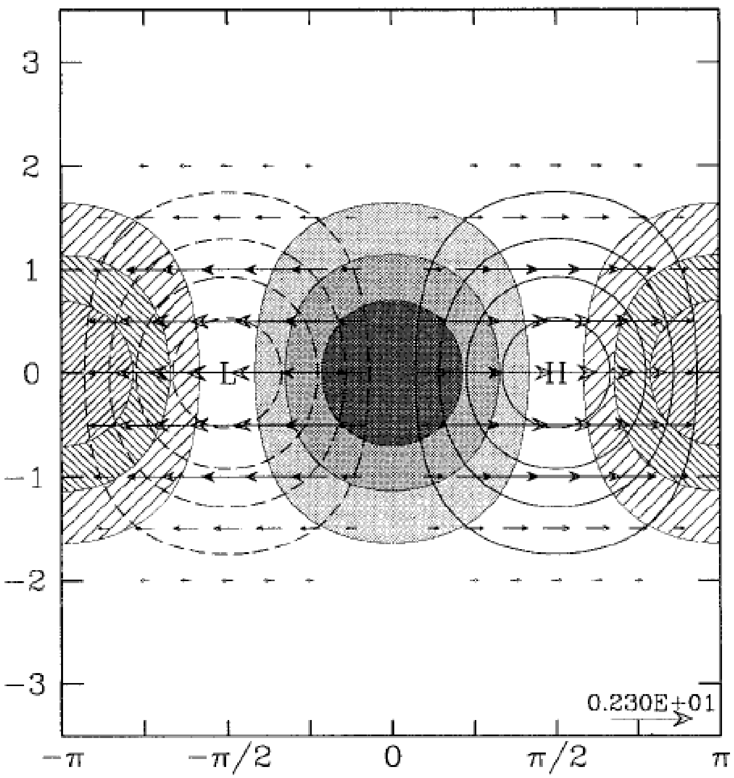
$$u = u_0 \, f(y) \, e^{i(kx - \omega t)}; \quad f(y) = e^{-y^2/2}$$

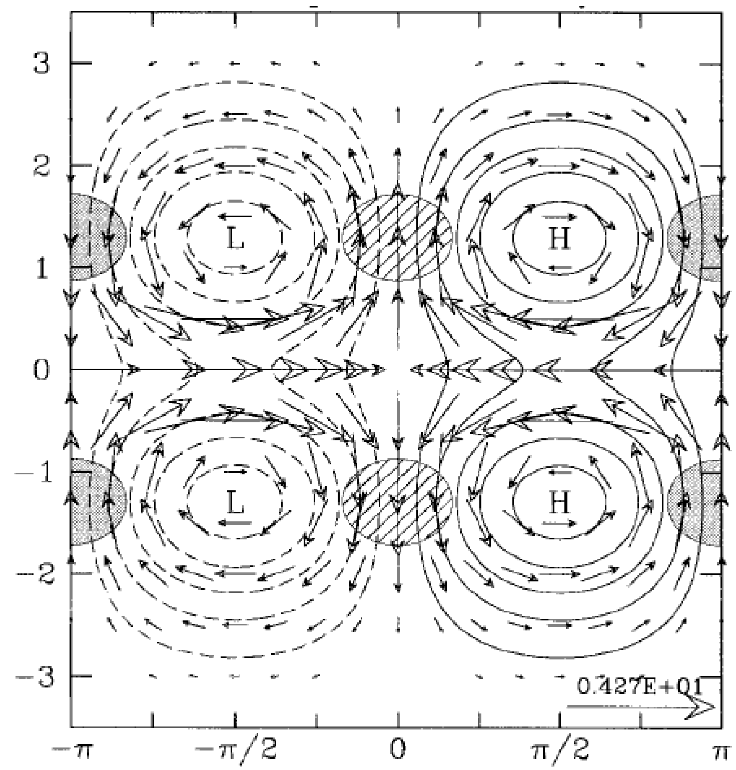
$$v = \hat{v}(y) \, f(y) \, e^{i(kx - \omega t)}; \quad \hat{v}(y) = Hermite\ Polynomials$$

ECMWF

The Kelvin wave

The n=1 Rossby wave





V=0, eastward moving ~18 m/s

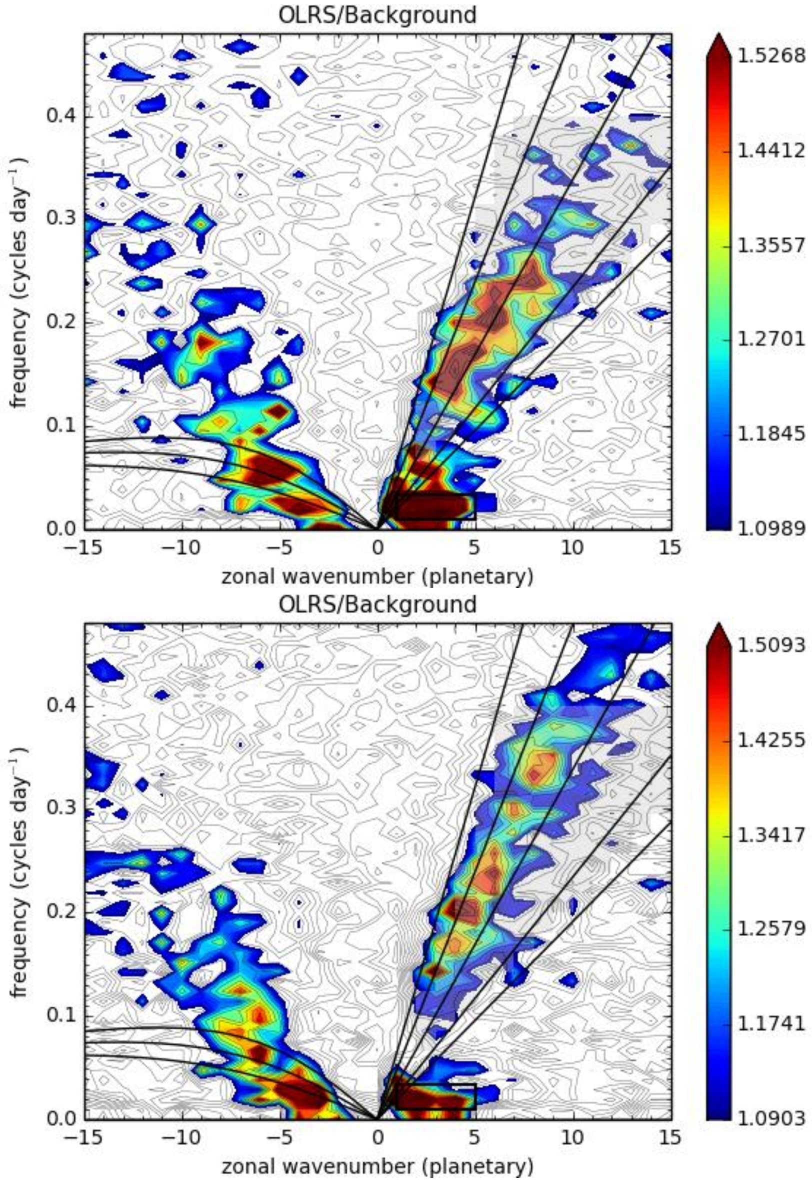
sym. around equator

westward moving ~5 m/s

sym. around equator

OLR anomaly shaded, winds max at equator

Wavenumber frequency Diagrams of OLR



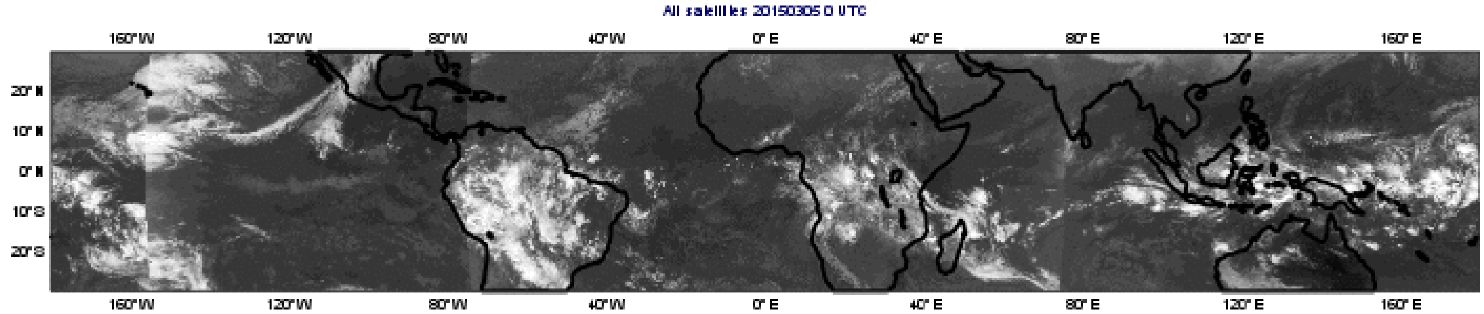
ECMWF Analysis

Cy40r1 (2014)

software courtesy Michael Herman (New Mexico Institute)

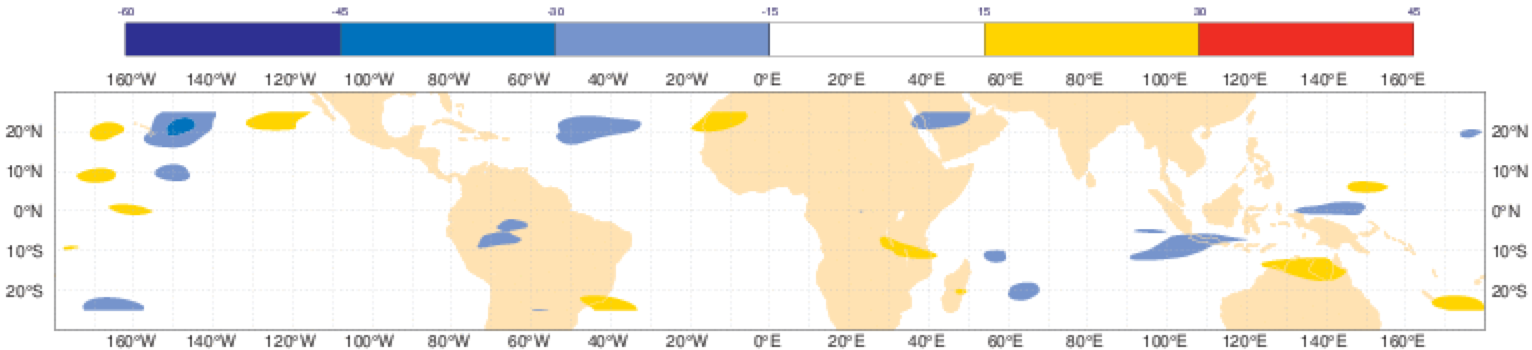
(all spectra have been divided by their own= smoothed background)

Rossby & MJO using OLR filtering 5.3.2015-16.3 2015

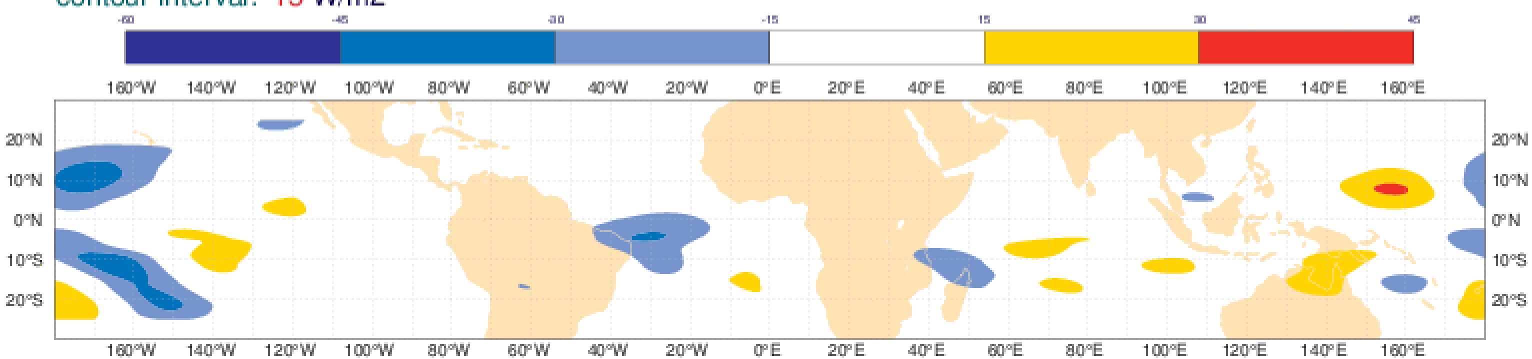


Real time monitoring of kelvin waves OLR (ECMWF) 20150305
contour interval: 15 W/m2

Forecast base time 2015 03 09



Real time monitoring of rossby waves OLR (ECMWF) 20150305
contour interval: 15 W/m2



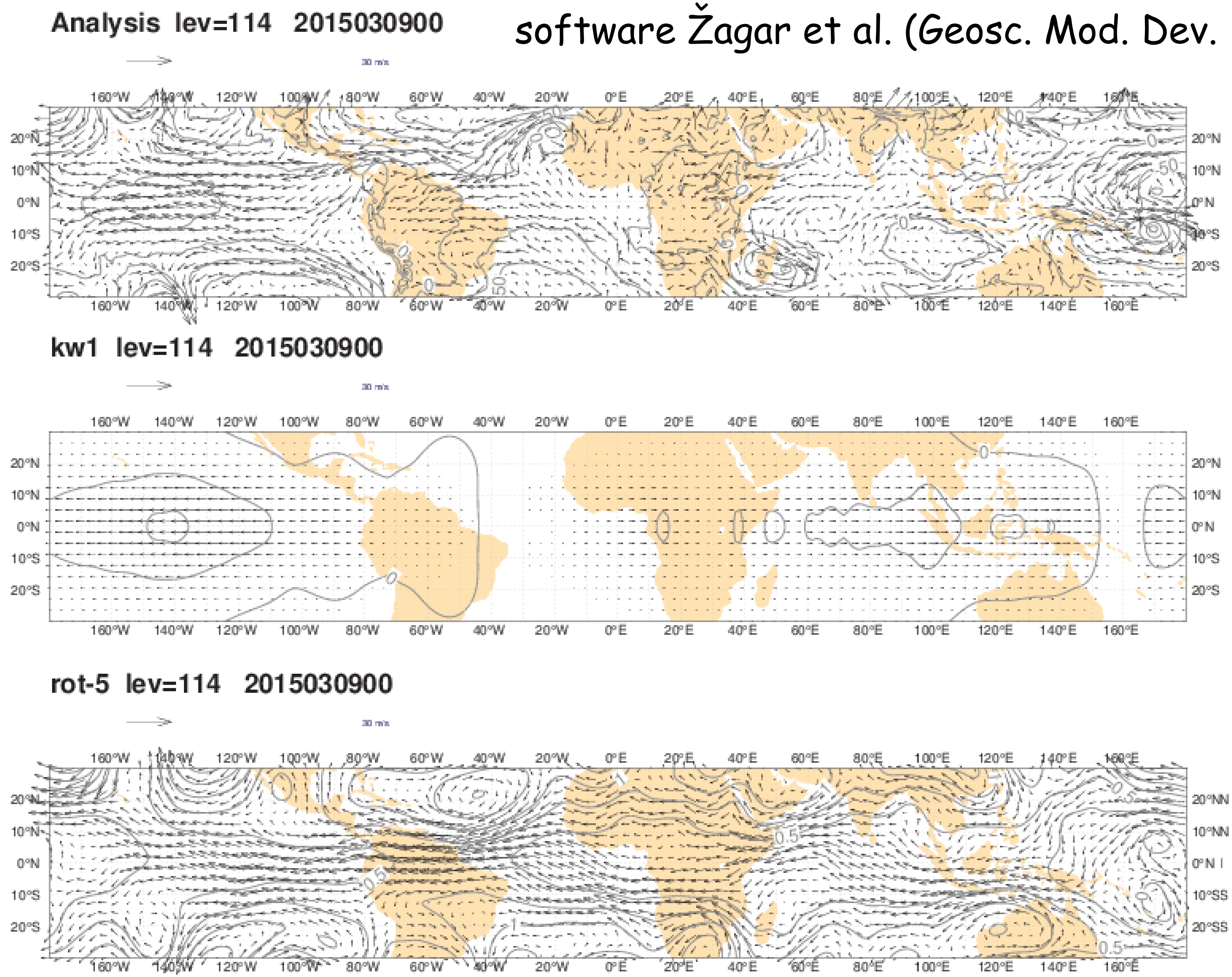
software courtesy M Herman following Wheeler and Weickman (MWR 2001)

ECMWF

Normal mode projection and filtering

software Žagar et al. (Geosc. Mod. Dev. 2015)

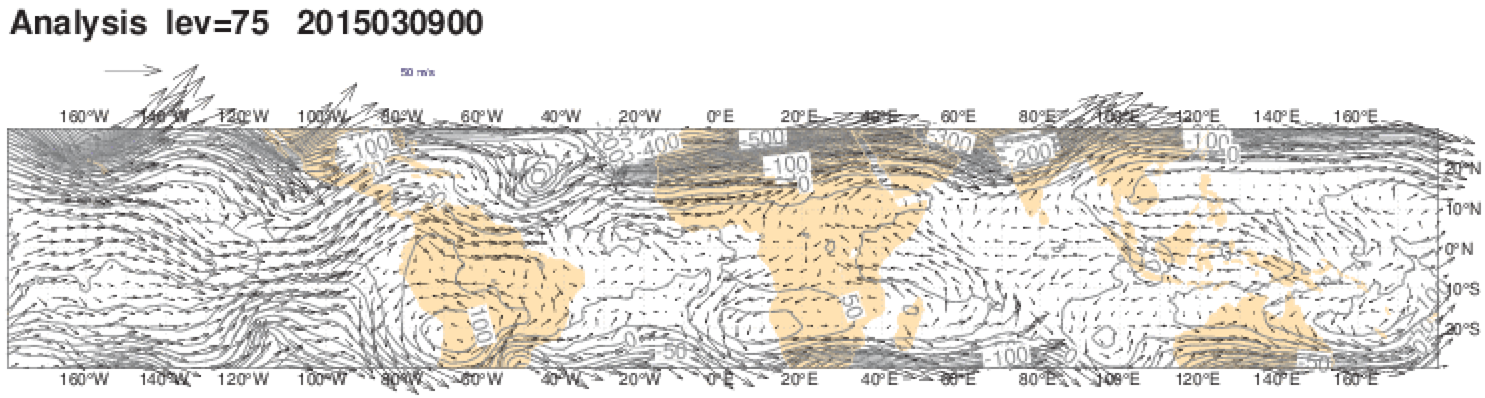
850 hPa

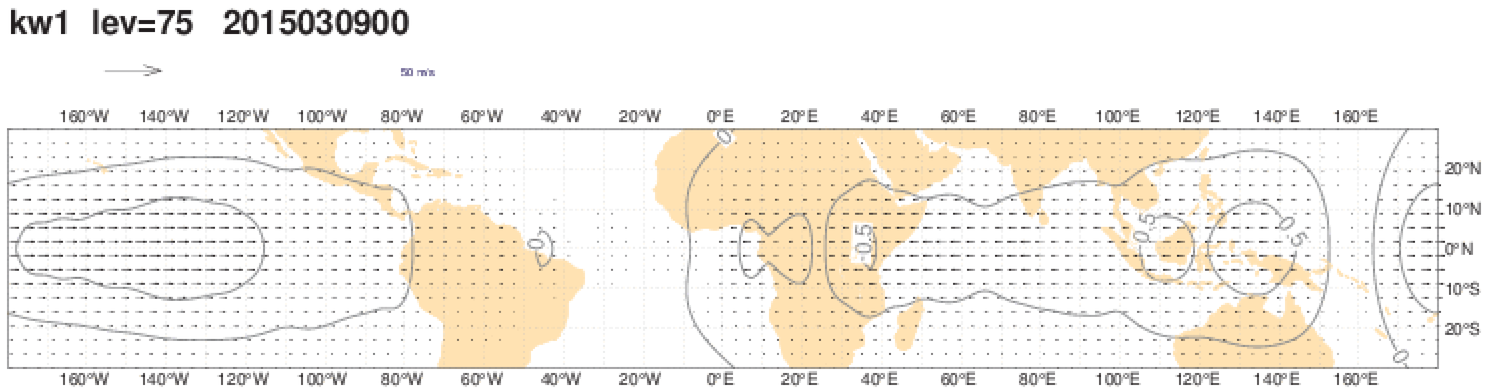


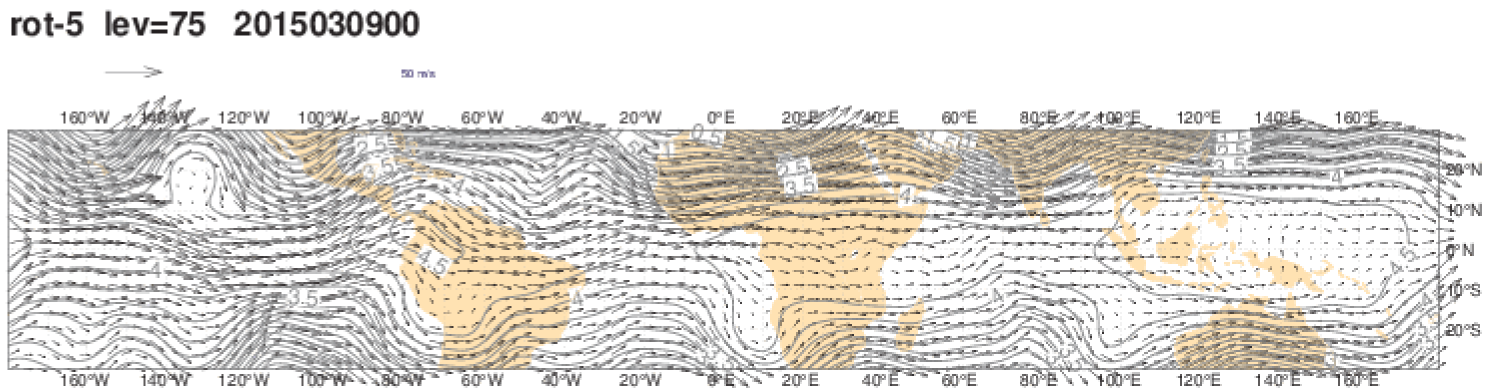
ECMWF

Normal mode projection and filtering

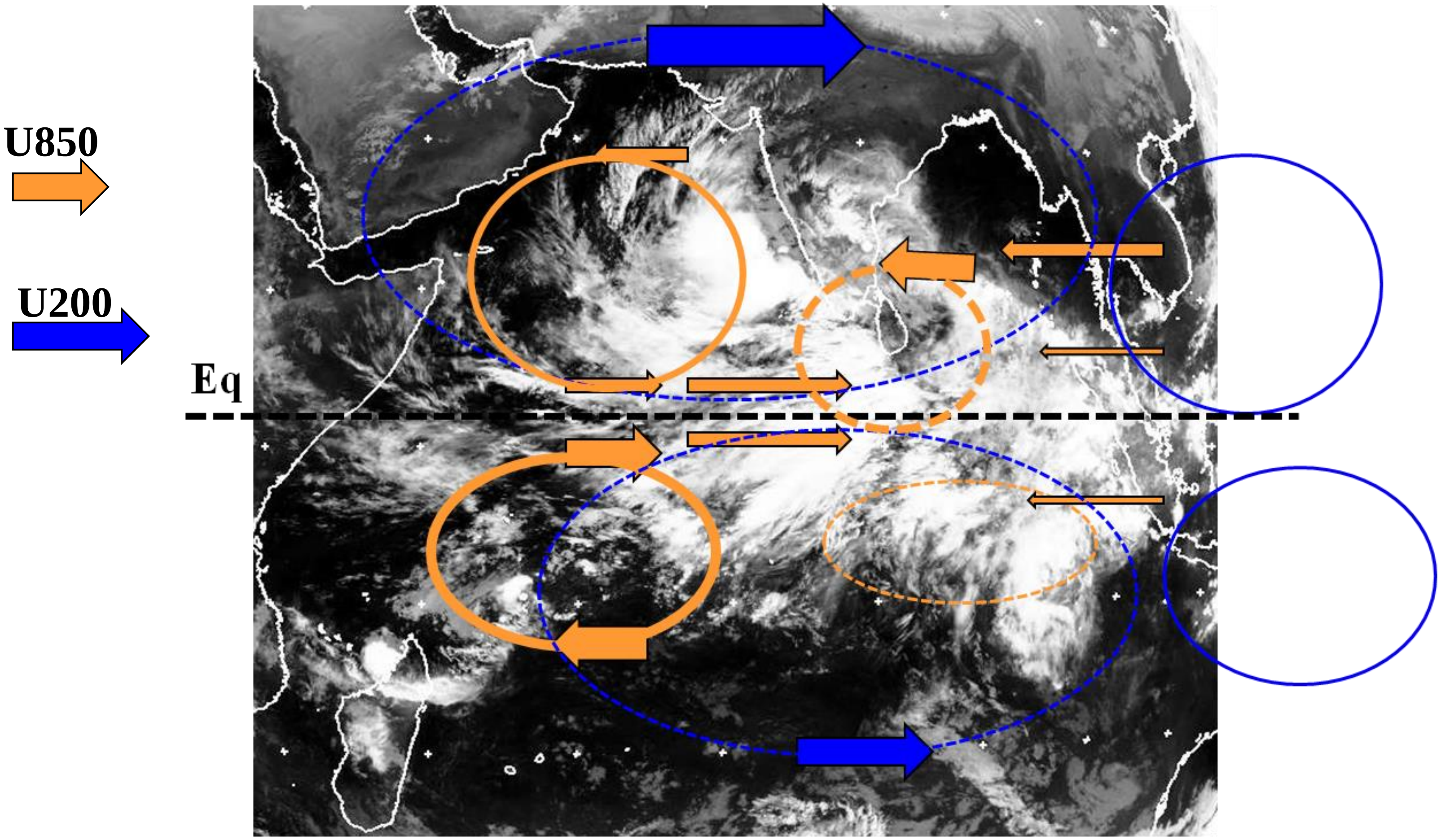
200 hPa





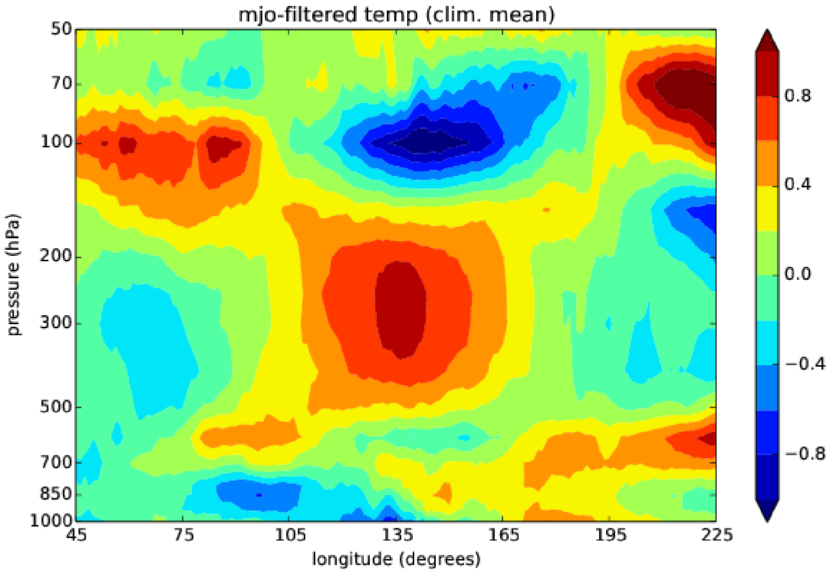


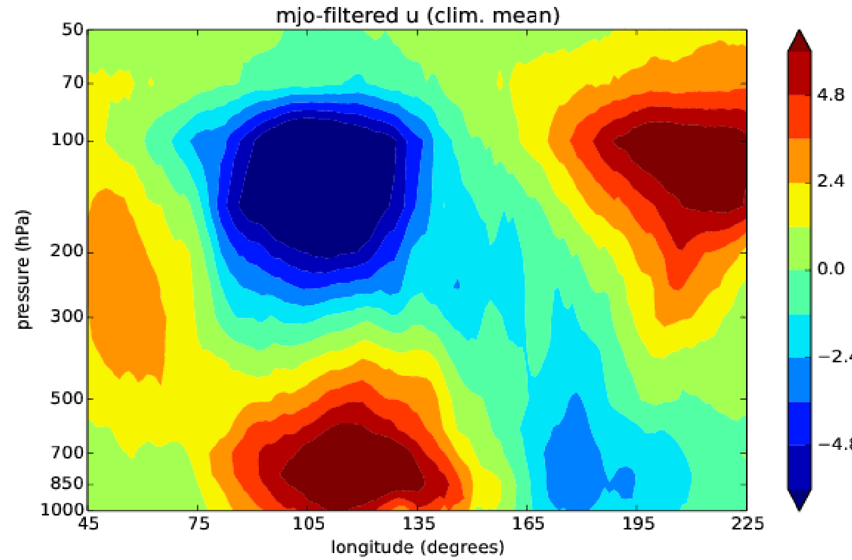
The MJO over Indian Ocean

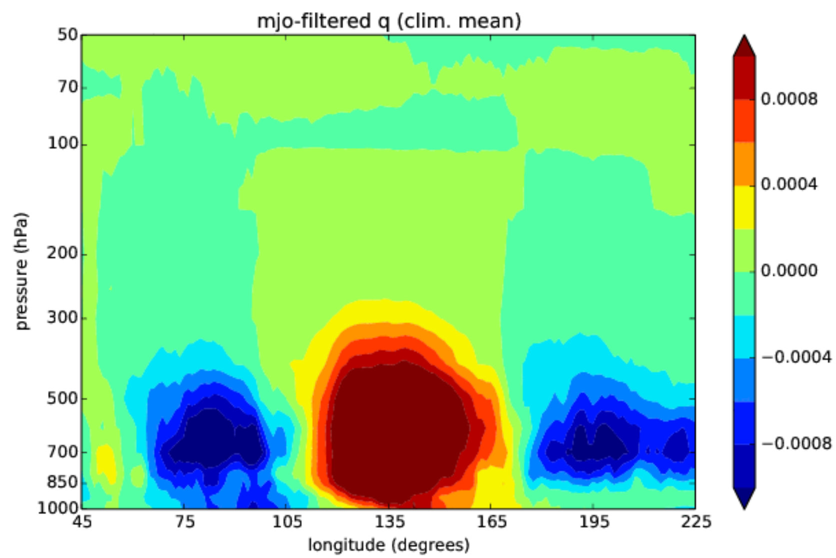


27 November 2011: Meteosat 7 + ECMWF Analysis

MJO composite
vertical structure of T.U.q anomalies





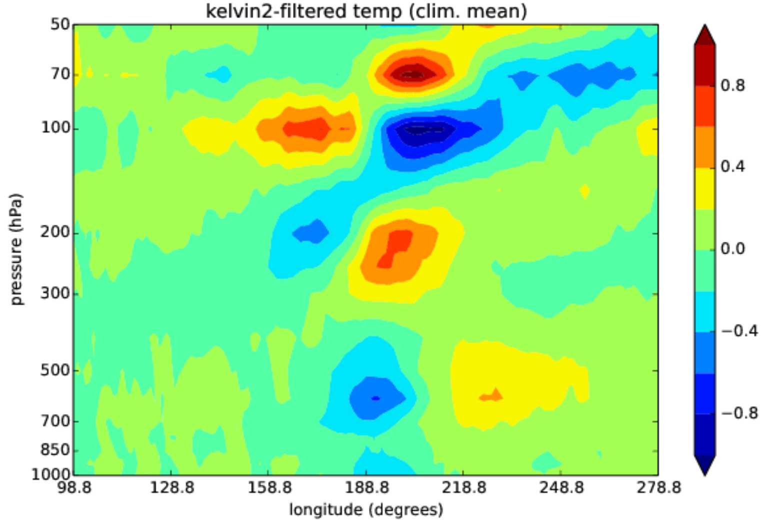


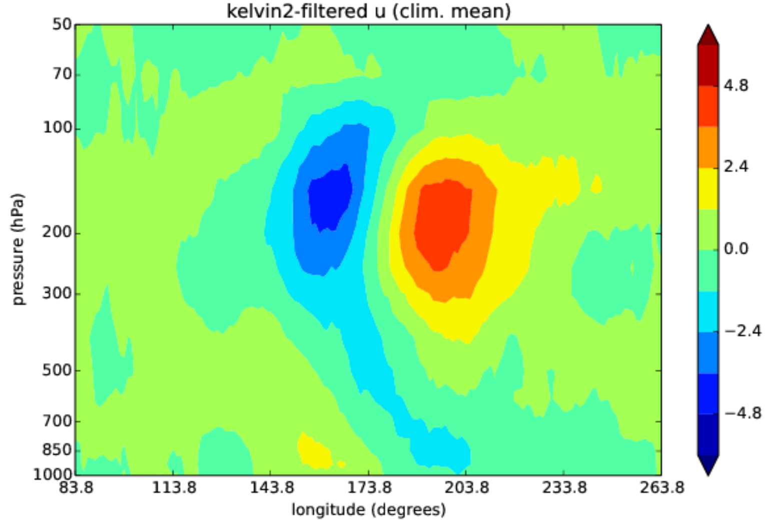
Note: tilted baroclinic structure

Software courtesy Michael Hermann

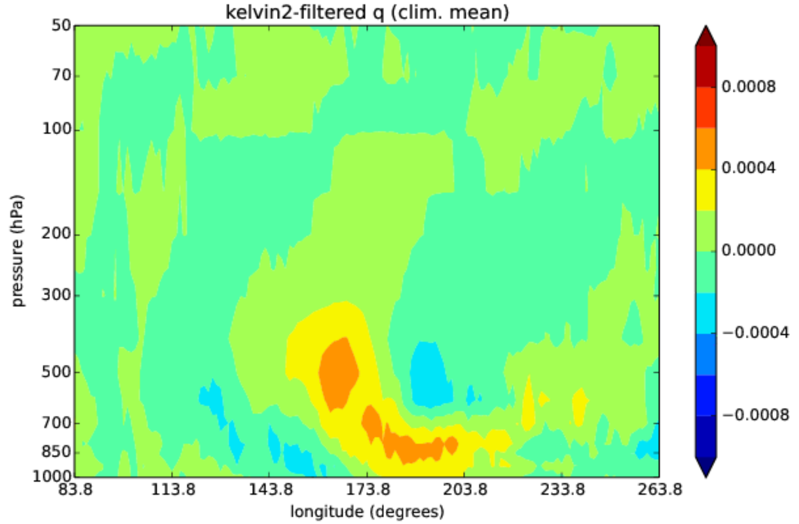
ECMWF

Kelvin composite
vertical structure of T,U,q anomalies



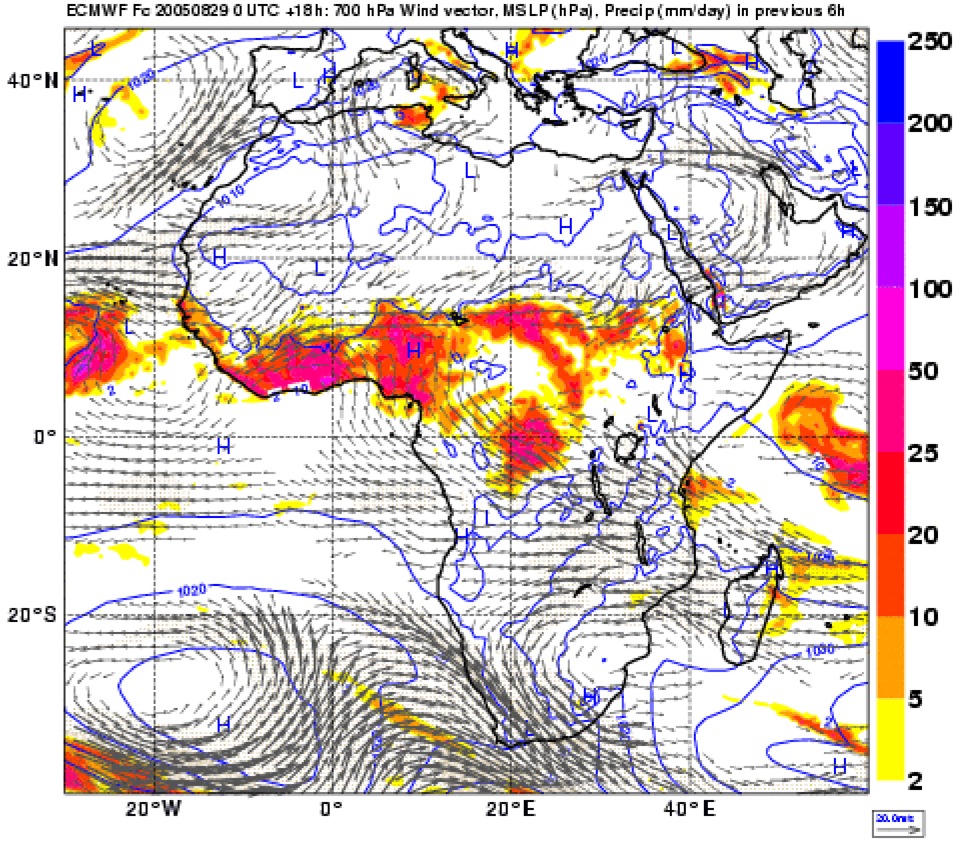


Note: The propagation to the stratosphere and different tilt of T-anomalies in tropos and stratosphere

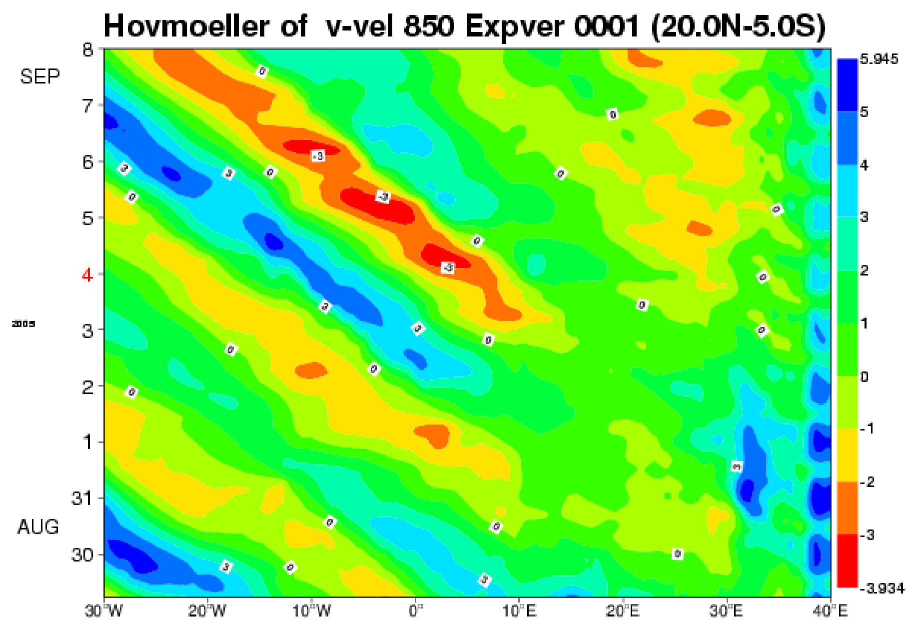


African Easterly waves

700 hPa wind, MSLP, and precip



Hovmoeller diagrams as an easy way to plot waves (propagation + amplitude)



Summary: the weather and thermal equilibria

- Suppose we have a series of fine day with an anticyclone, the temperature above the boundary-layer barely changes, Why?

$$\frac{d\theta}{dt} \approx 0 \quad \Rightarrow w\frac{d\theta}{dz} = \frac{d\theta}{dt}\Big|_{rad} = -\frac{2K}{86400s} \Rightarrow$$

w ~ -0.5 cm/s subsidence

~0.5 K/100 m

- But what happens when it is raining 100 mm/day ?

$$\int_{surf}^{10km} Cp\frac{dT}{dt}\rho_{air}\, dz = L_v \rho_{water}\, Pr(m/s)$$

$$c_p = 1005\, J/kg/K; \quad \rho_{water} = 1000 kg/m3; \quad L_v = 2.5x10^6 \text{ J/kg}$$

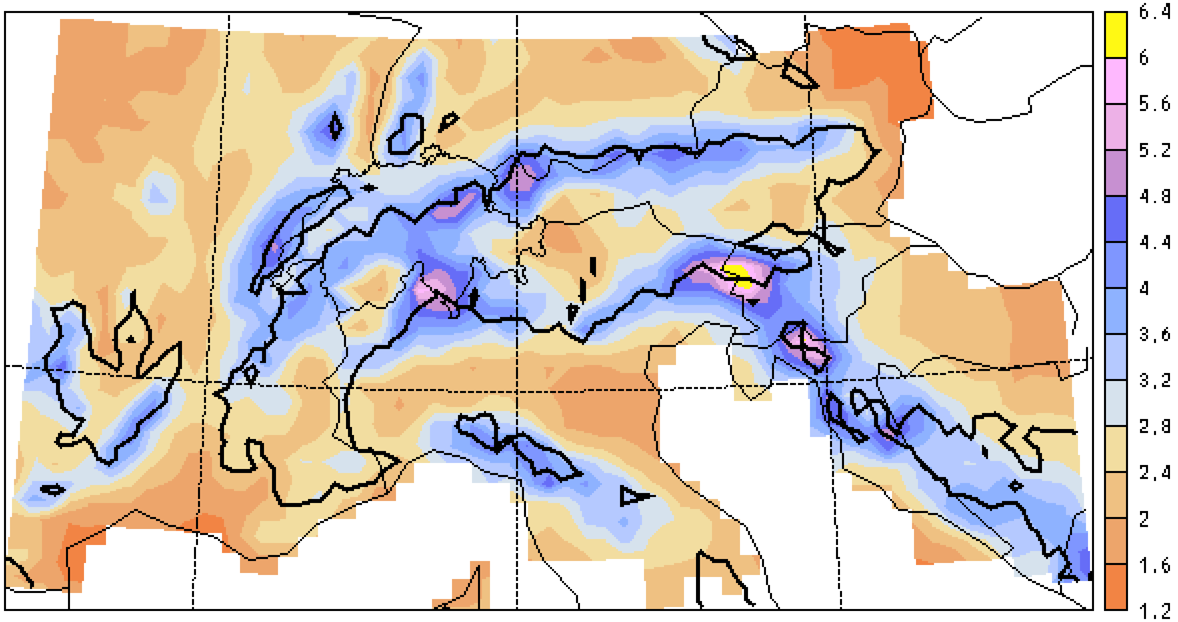
$$Pr = 100\frac{mm}{day} = 1.147\, m/s\, x10^{-6}$$

100 mm/day precipitation heats the atmospheric column by 2867 W/m2 or by 25 K/day on average. This heating must be compensated by uplifting of w ~ 10 cm/s ➔ heavy precip/convection requires large-scale perturbation.

Midlatitude Convection (1)

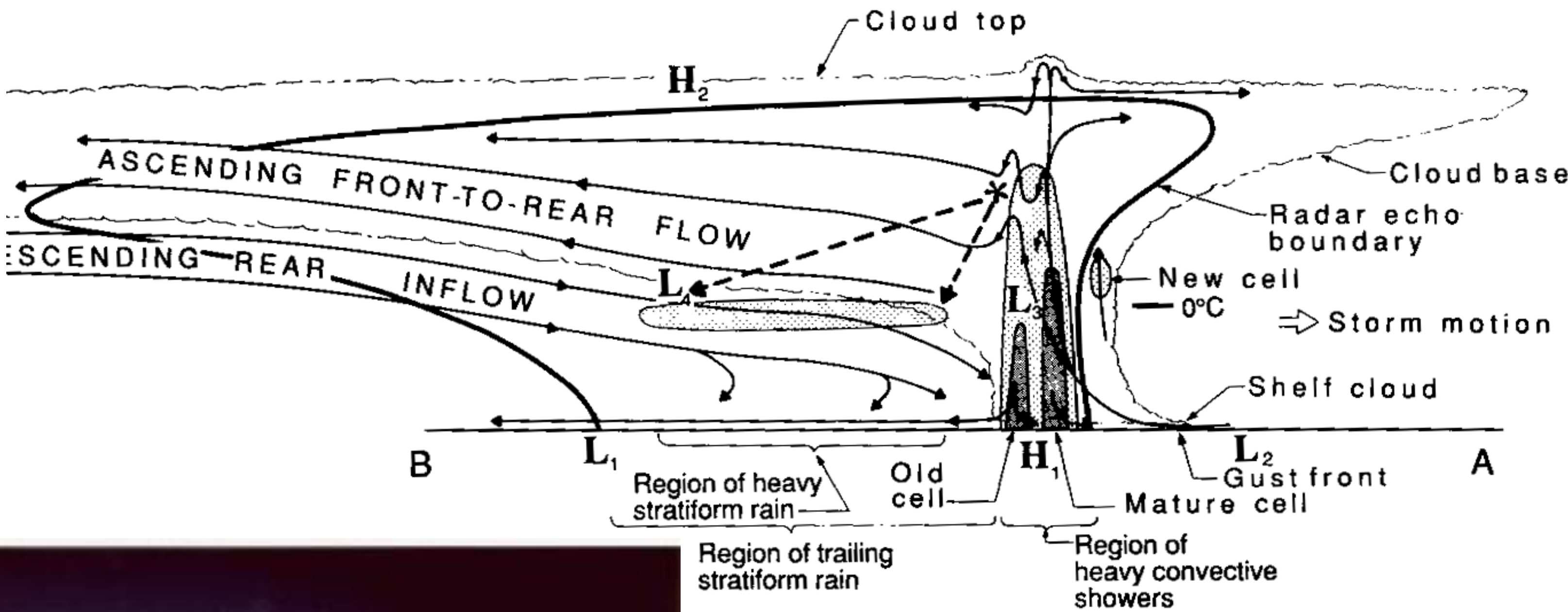
Europe climatology (Frei and Schär, 1998)

In Europe most intense precipitation is associated with orography, especially around the Mediterranean, associated with strong large-scale forcing and mesoscale convective systems



Midlatitude Convection (2)

Squall line system conceptual and observed



- Distinctive convective and trailing stratiform regions with characteristic inflow (Houze et al. 1989)

Supercell over Central US, Mai 1998, flight level 11800 m

Midlatitude Convection (3)

European MCSs (Morel and Sénési, 2001)

Density Map of Triggering ….. over Orography

1978 C. MOREL and S. SENESI

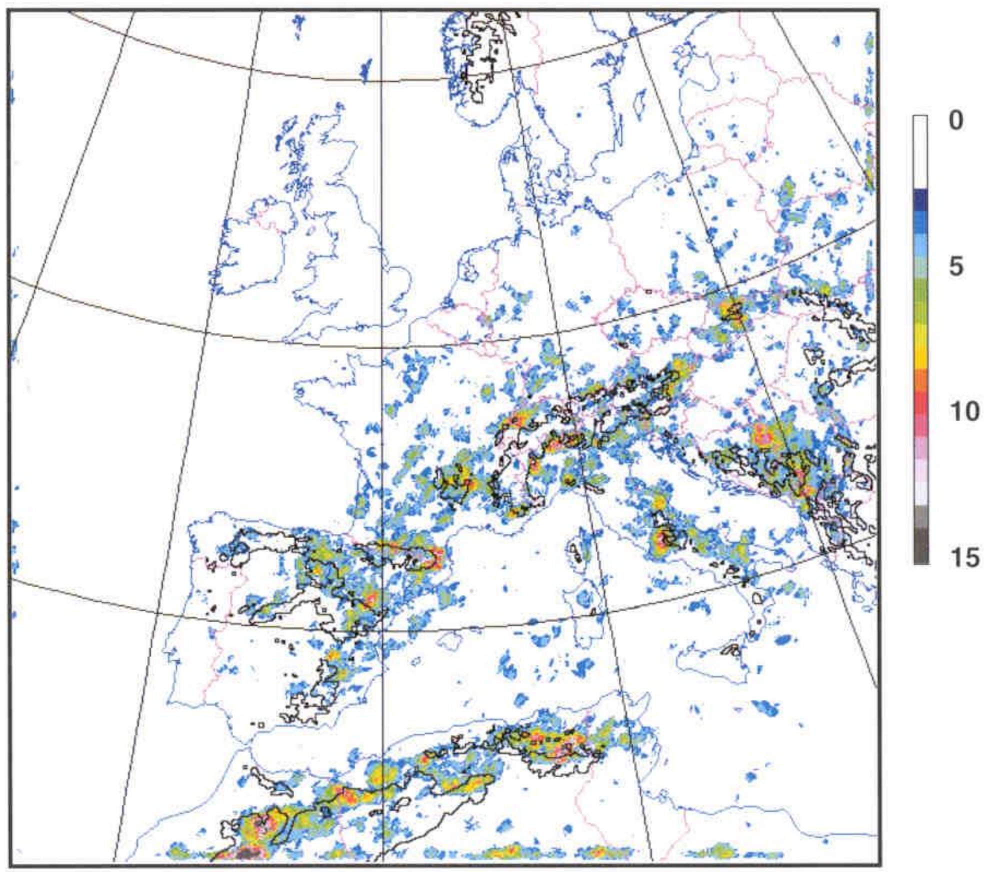


Figure 2. Density map of mesoscale convective systems (MCSs) triggering for the 4813 trajectories beginning normally (in number of MCSs triggering over each pixel). The black solid line is the 1000-metre elevation contour.

Midlatitude Convection (4)

European MCSs (Morel and Sénési, 2001)

Time of Trigger and mean propagation

CLIMATOLOGY OF EUROPEAN MCSs. II 1991

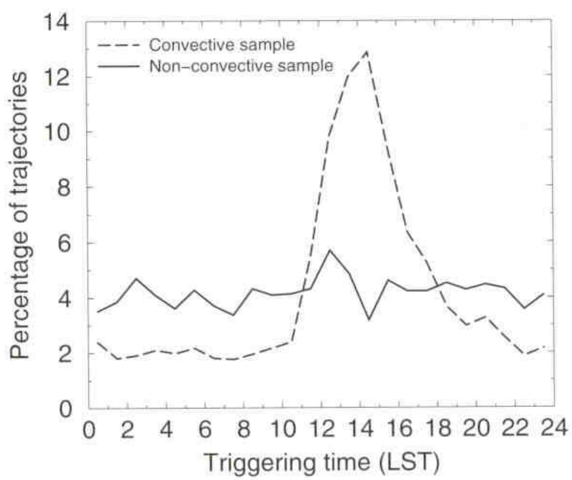


Figure 15. Distribution of the triggering time for 2723 'simple' trajectories of the convective sample (dashed line) and for the 2105 'simple' trajectories of the non-convective sample (solid line), in Local Solar Time (LST).

CLIMATOLOGY OF EUROPEAN MCSs. II

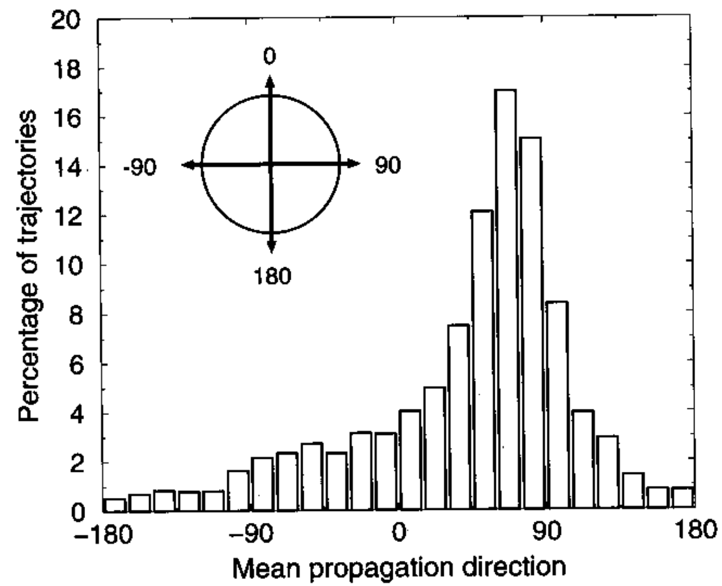
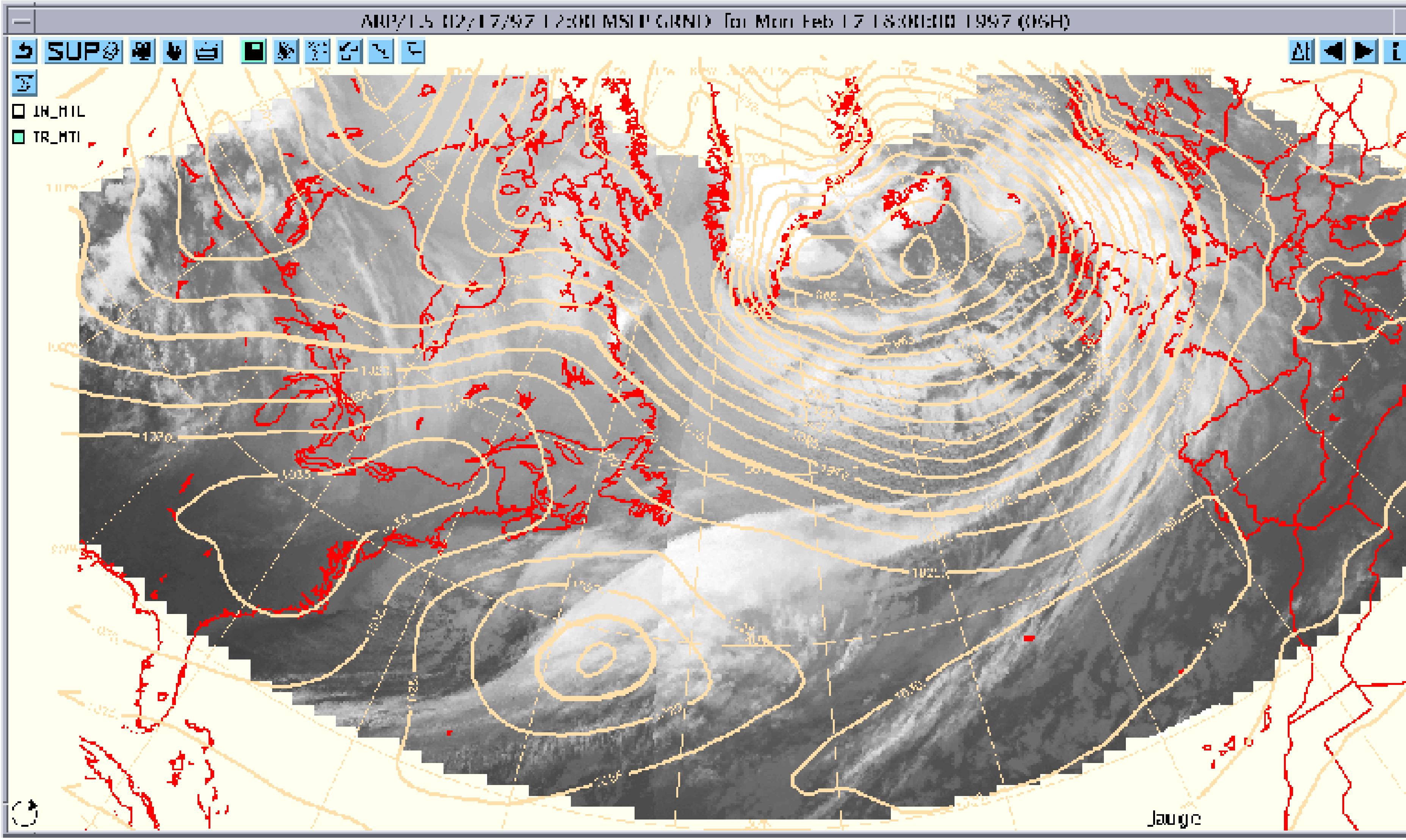


Figure 5. Distribution of the mean propagation direction for the whole database.

European (midlatitude) MCSs essentially form over orography (convective inhibition –see later- offset by uplift) and then propagate with the midtropospheric flow (from SW to NE)

Midlatitude Convection (5)

along the main cold frontal band and in the cold core of the main depression – 17/02/97 during FASTEX



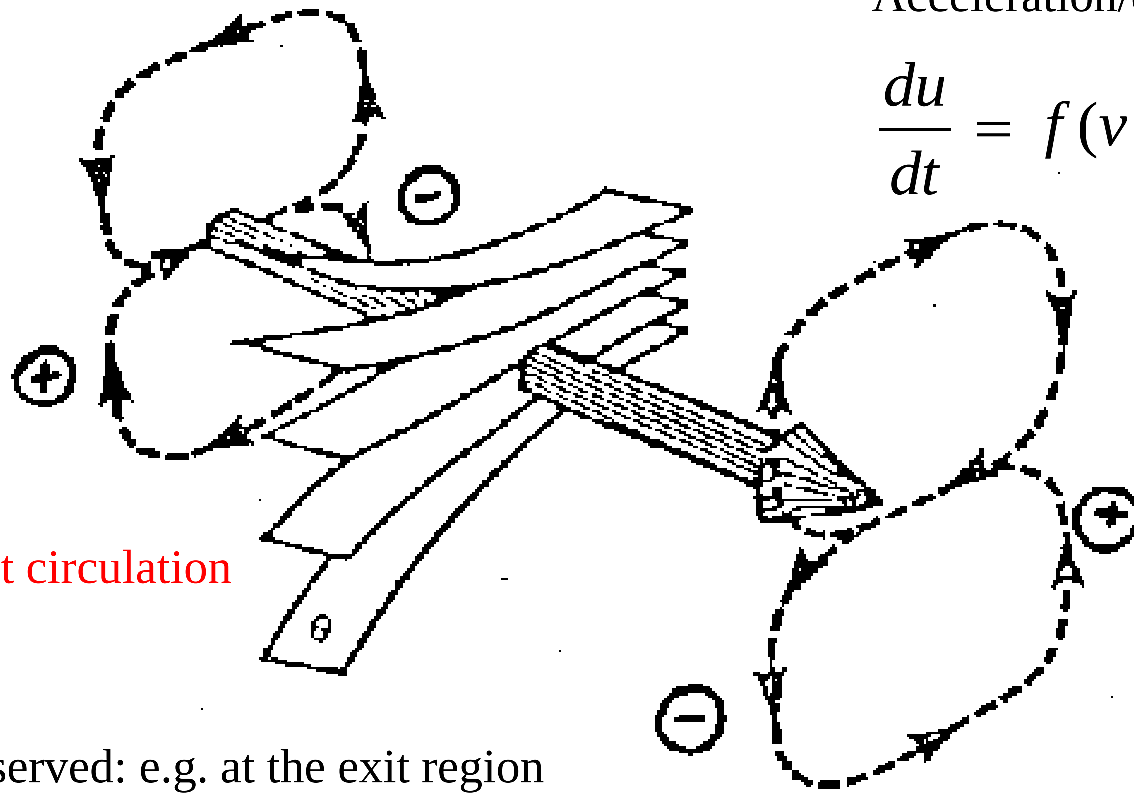
ECMWF

Midlatitude Convection (6)

Forcing of ageostrophic circulations/convection in the right entrance and left exit side of upper-level Jet

Acceleration/deceleration of Jet

$$\frac{du}{dt} = f\,(v - v_g) \equiv fv_a$$



Thermally indirect circulation

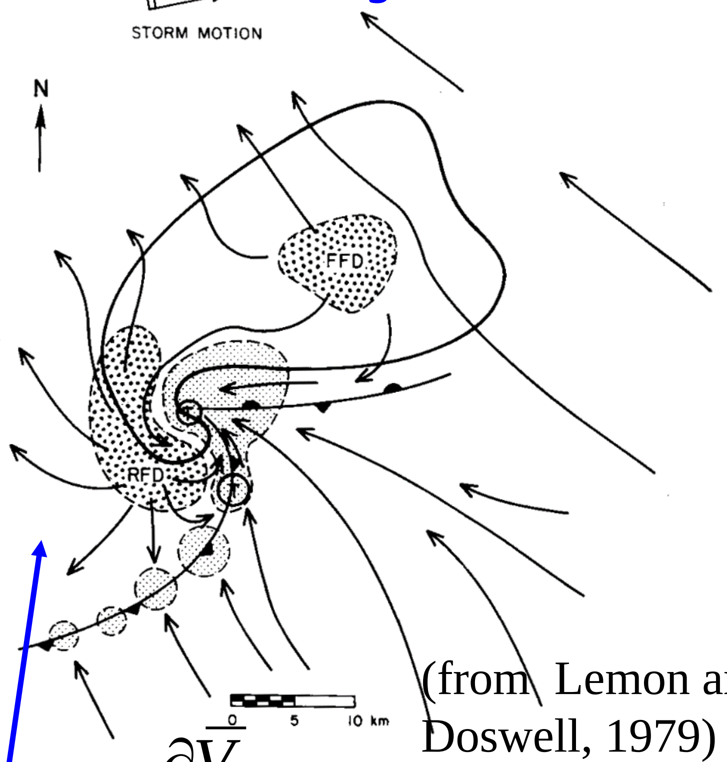
Thermally direct circulation

Total energy is conserved: e.g. at the exit region where the Jet decelerates kinetic energy is converted in potential energy

Midlatitude Convection (7)
Tornadic Storms

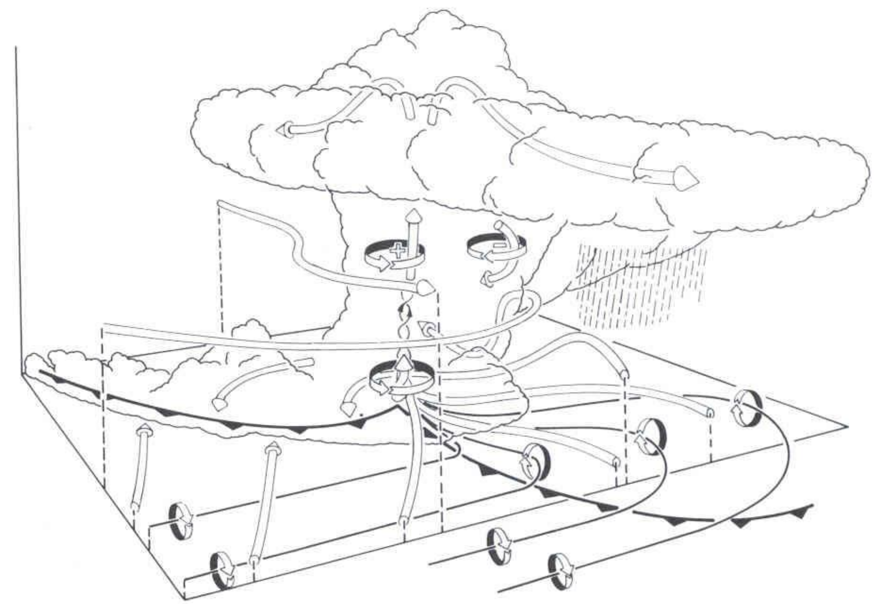
Conversion of horizontal vorticity at front in vertical vorticity by tilting in updraft

Importance of wind shear: Interaction of updraught with environm. Shear creates Rear Flank Downdraught



(from Lemon and Doswell, 1979)

$$P_L' \approx \frac{\partial \overline{V}}{\partial z} \bullet \nabla_z w'$$



(from J Klempp 1987)

A useful quantity in estimating the storm intensity is the"bulk" Richardson number R=CAPE/S^2

Summary: effects and cause of convection

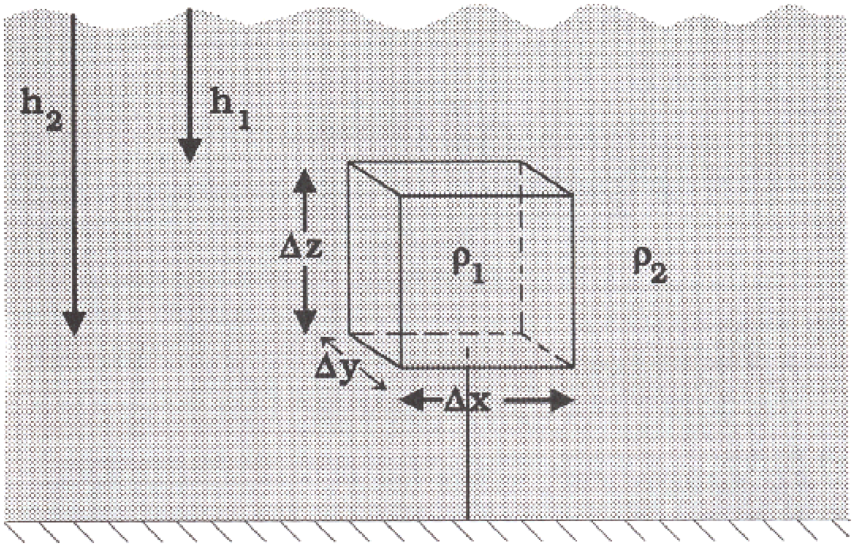
- Convection transports heat, water vapor, momentum … and chemical constituents upwards …. Water vapor then condenses and falls out -> net convective heating/drying

- Deep Convection (precipitating convection) stabilizes the environment, an approximate picture (not true for diurnal cycle convection!) is to consider it as reacting to the large-scale environment (e.g. tropical waves, mid-latitude frontal systems) =“quasi-equilibrium”; shallow convection redistributes moisture and heat

- The effect of convection (local heat source) is fundamentally different in the middle latitudes and the Tropics. In the Tropics the Rossby radius of deformation $R=N\,H/f$ (N=Brunt Vaisala Freq, f=Coriolis parameter, H=tropopause height) is infinite, and therefore the effects are not locally bounded, but spread globally via gravity waves – “throwing a stone in a lake”

Buoyancy (1)- Archimedes said 'Eureka!'

Body in a fluid



Assume fluid to be in hydrostatic equlibrium $\frac{dp_2}{dz} = -\rho_2 g$

$$\rho_2 = const. \longrightarrow p_2 = \rho_2 g h$$

Forces:

Top $F_{top} = -\rho_2 g h_1 \Delta x \Delta y$

Bottom $F_{bot} = \rho_2 g h_2 \Delta x \Delta y$

Gravity $F_{grav} = -\rho_1 g \Delta x \Delta y \Delta z$

Net Force: $F = F_{top} + F_{bot} + F_{grav} = \rho_2 g (h_2 - h_1) \Delta x \Delta y - \rho_1 g \Delta x \Delta y \Delta z = g(\rho_2 - \rho_1) \Delta x \Delta y \Delta z$

Acceleration: $A = F \big/ M_{body} = F \big/ \rho_1 \Delta x \Delta y \Delta z = g \frac{(\rho_2 - \rho_1)}{\rho_1}$

Emanuel, 1994

Buoyancy (2)

Vertical momentum equation:

$$\frac{dw}{dt} = -\frac{1}{\rho}\frac{\partial p}{\partial z} - g$$

$$p = \overline{p} + p' \qquad \rho = \overline{\rho} + \rho' \qquad \frac{\partial \overline{p}}{\partial z} = -\overline{\rho} g$$

$$\frac{dw}{dt} = -\frac{1}{\overline{\rho} + \rho'}\frac{\partial(\overline{p} + p')}{\partial z} - g$$

$$\frac{1}{\overline{\rho} + \rho'} = \frac{1}{\overline{\rho}}\left(\frac{1}{1 + \rho'/\overline{\rho}}\right) = \frac{1}{\overline{\rho}}\left[1 - \frac{\rho'}{\overline{\rho}} + \left(\frac{\rho'}{\overline{\rho}}\right)^2 + \ldots\right]$$

$\rho' << \overline{\rho} \longrightarrow$ Neglect second order terms

Buoyancy (3)

$$\frac{dw}{dt} = -\underbrace{\frac{1}{\overline{\rho}}\frac{\partial \overline{p}}{\partial z}}_{= g} - \frac{1}{\overline{\rho}}\frac{\partial p'}{\partial z} - g + \frac{\rho'}{\overline{\rho}}\underbrace{\frac{1}{\overline{\rho}}\frac{\partial \overline{p}}{\partial z}}_{= -g} + \cancel{\frac{\rho'}{\overline{\rho}}\frac{1}{\overline{\rho}}\frac{\partial p'}{\partial z}}$$

$$\boxed{\frac{dw}{dt} = -\frac{1}{\overline{\rho}}\frac{\partial p'}{dz} - \underbrace{\frac{\rho'}{\overline{\rho}}g}_{=}}$$

B - buoyancy acceleration

Buoyancy (4) T and P and Contributions

Buoyancy acceleration: $$B = -\frac{\rho'}{\overline{\rho}} g$$

Dry air: $$\rho = \frac{p}{RT} \rightarrow \rho' = \frac{p'}{R\overline{T}} - \frac{\overline{p}T'}{R\overline{T}^2} \rightarrow \frac{\rho'}{\overline{\rho}} = \frac{p'}{\overline{p}} - \frac{T'}{\overline{T}}$$

Often (but not always): $$\frac{p'}{\overline{p}} << \frac{T'}{\overline{T}} \text{ and } B \approx g\frac{T'}{\overline{T}}$$

Then $$\frac{dw}{dt} \approx g\frac{T'}{\overline{T}} - \frac{1}{\overline{\rho}}\frac{\partial p'}{\partial z}$$

Hence $$T' > 0 \text{ (warm parcel)} \rightarrow \frac{dw}{dt} > 0 \rightarrow \text{upward accelaration (downward deceleration)}$$

$$T' < 0 \text{ (cold parcel)} \rightarrow \frac{dw}{dt} < 0 \rightarrow \text{upward deceleration (downward acceleration)}$$

Buoyancy (5) moist atmosphere

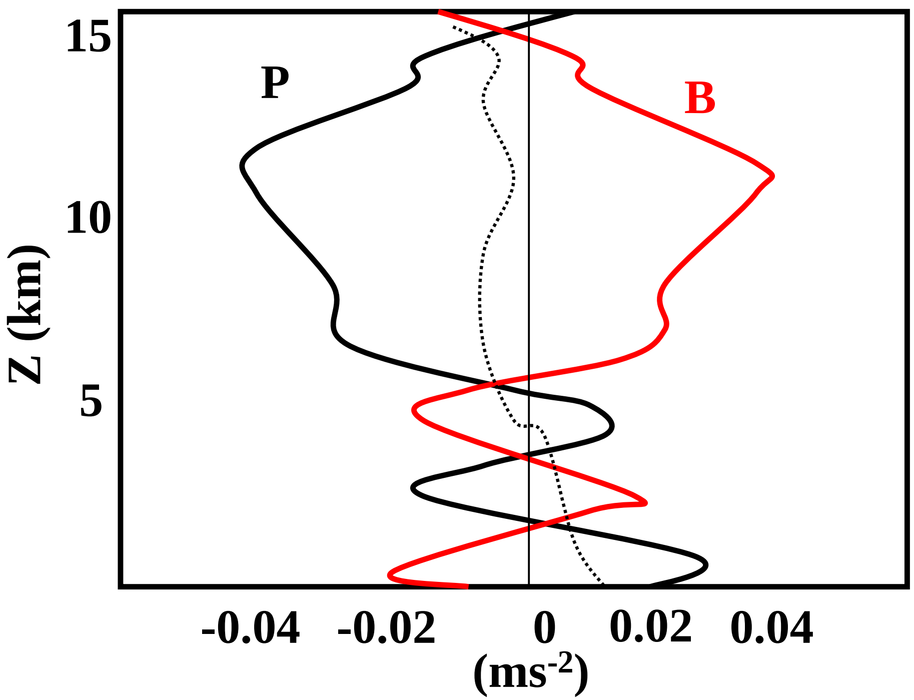
effects of humidity and condensate need to be taken into account

$$B = -g\frac{\rho'}{\bar{\rho}} \approx -g\left(\frac{T'}{\bar{T}} + 0.608q' - q_l\right)$$

In general all 3 terms are important. 1 K perturbation in T is equivalent to 5 g/kg perturbation in water vapor or 3 g/kg in condensate

Non-hydrostat. Pressure gradient effects

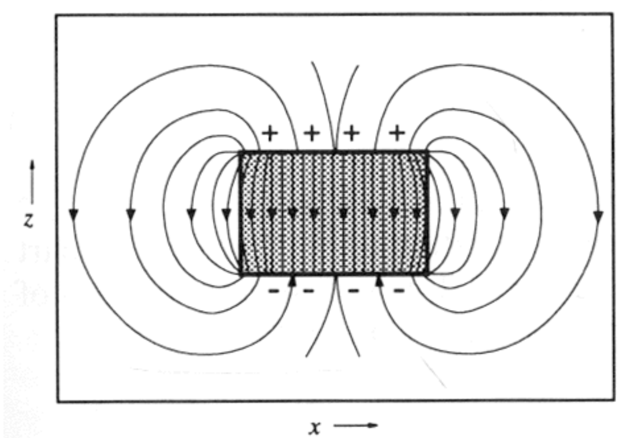
$$\frac{dw}{dt} = -\frac{1}{\rho}\frac{\partial p'}{\partial z} - \frac{\rho'}{\overline{\rho}}g$$



CRM analysis of the terms

Guichard and Gregory

Physics:



Vector field of the buoyancy pressure-gradient force for a uniformly buoyant parcel of finite dimensions in the x-z-plane. (Houze, 1993, Textbook)

Convective Available Potential Energy (CAPE)

Definition:

$$CAPE = \int \vec{F} \cdot d\vec{l} = \int_{base}^{top} B dz$$

$$CAPE \approx \int_{base}^{top} g \frac{T_{cld} - T_{env}}{T_{env}} dz$$

CAPE represents the amount of potential energy of a parcel lifted to its level of neutral buoyancy. This energy can potentially be released as kinetic energy in convection.

$$\frac{dw}{dt} = w\frac{dw}{dz} = \frac{1}{2}\frac{dw^2}{dz} \approx g\frac{T'}{\overline{T}}$$

$$w^2(z) = 2\int_0^z g\frac{T'}{\overline{T}}dz = 2 \cdot CAPE$$

$$w = \sqrt{2 \cdot CAPE}$$

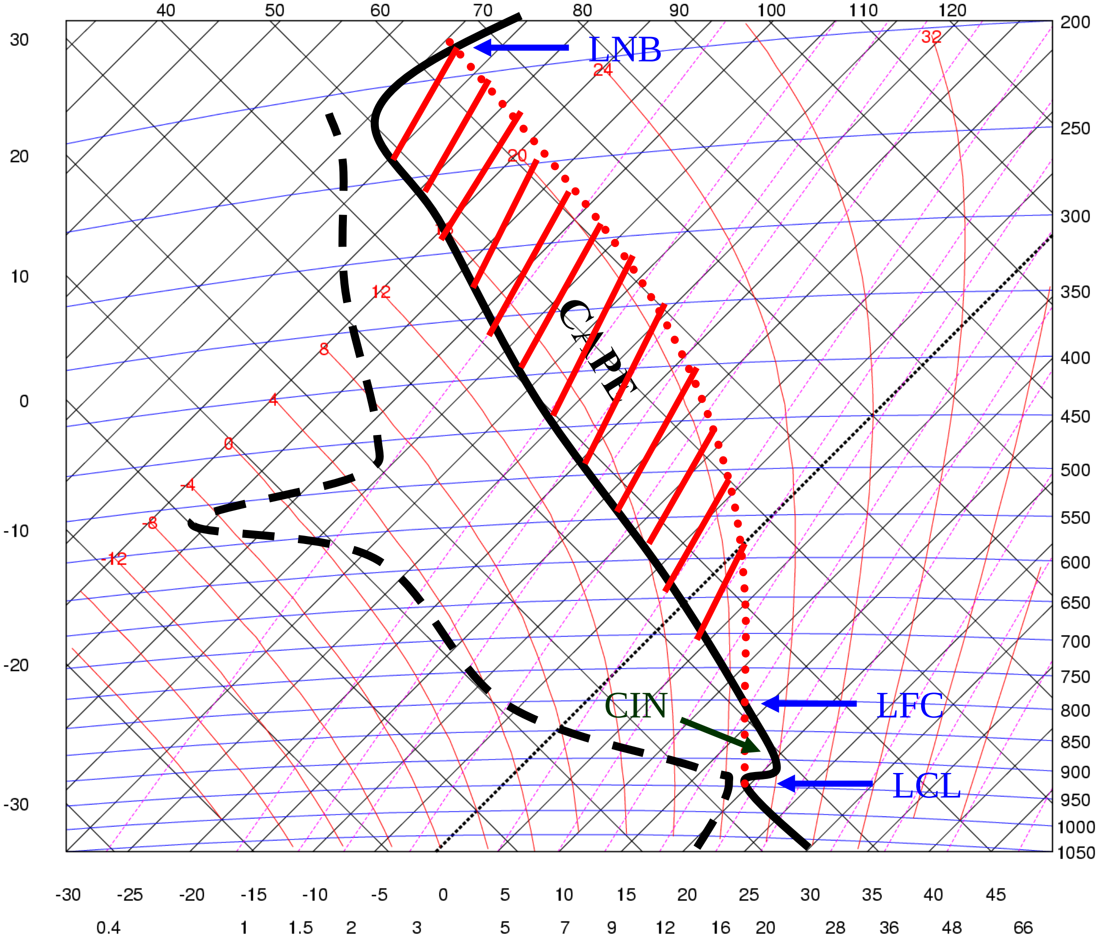
Example:

$$T' = 5K, \quad \overline{T} = 250K, \quad \text{Cloud depth} = 10\text{km}$$

$$w \approx 60 ms^{-1}$$

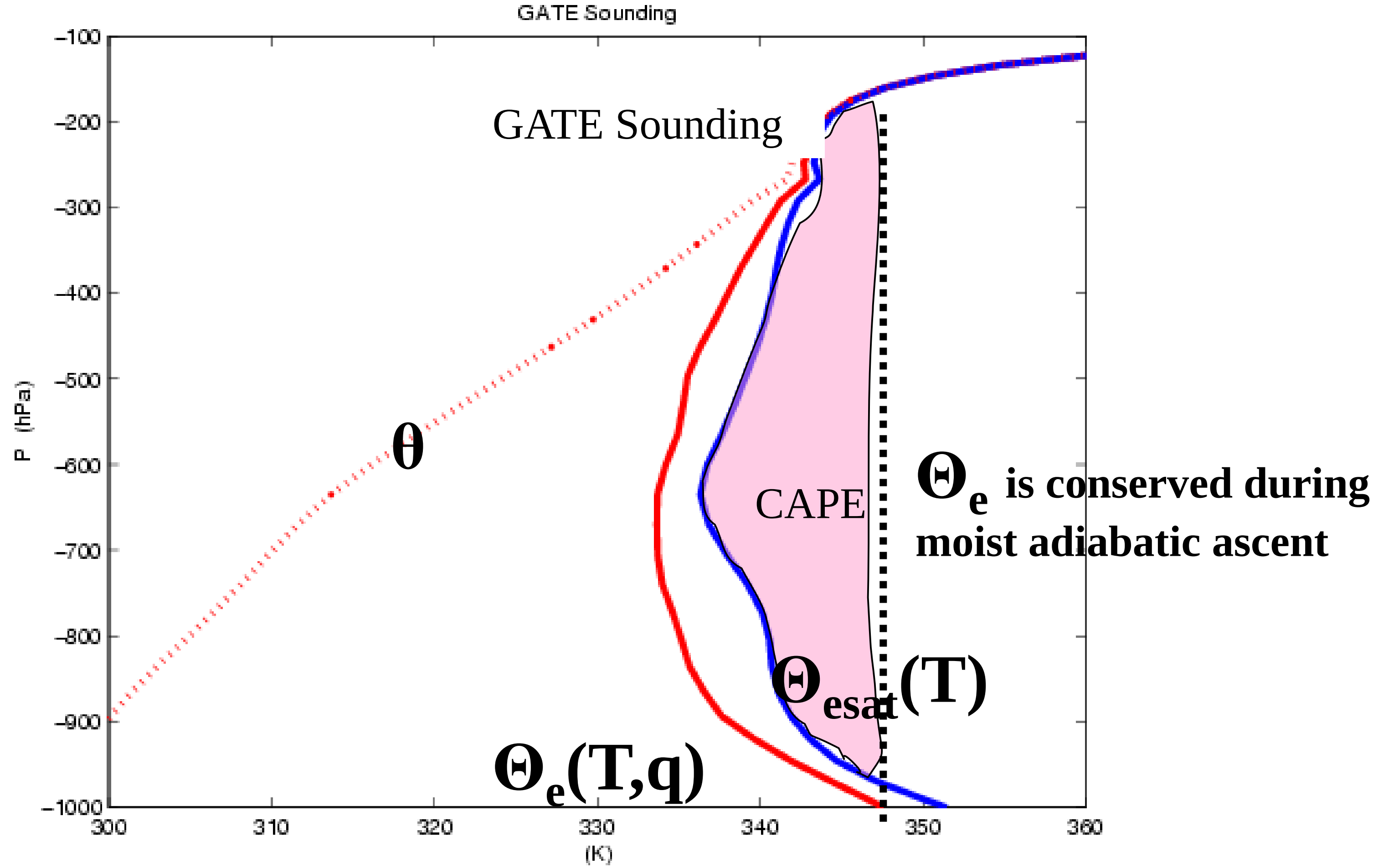
Much larger than observed - what's going on ?

Convection in thermodynamic diagrams (1)
using Tephigram/Emagram



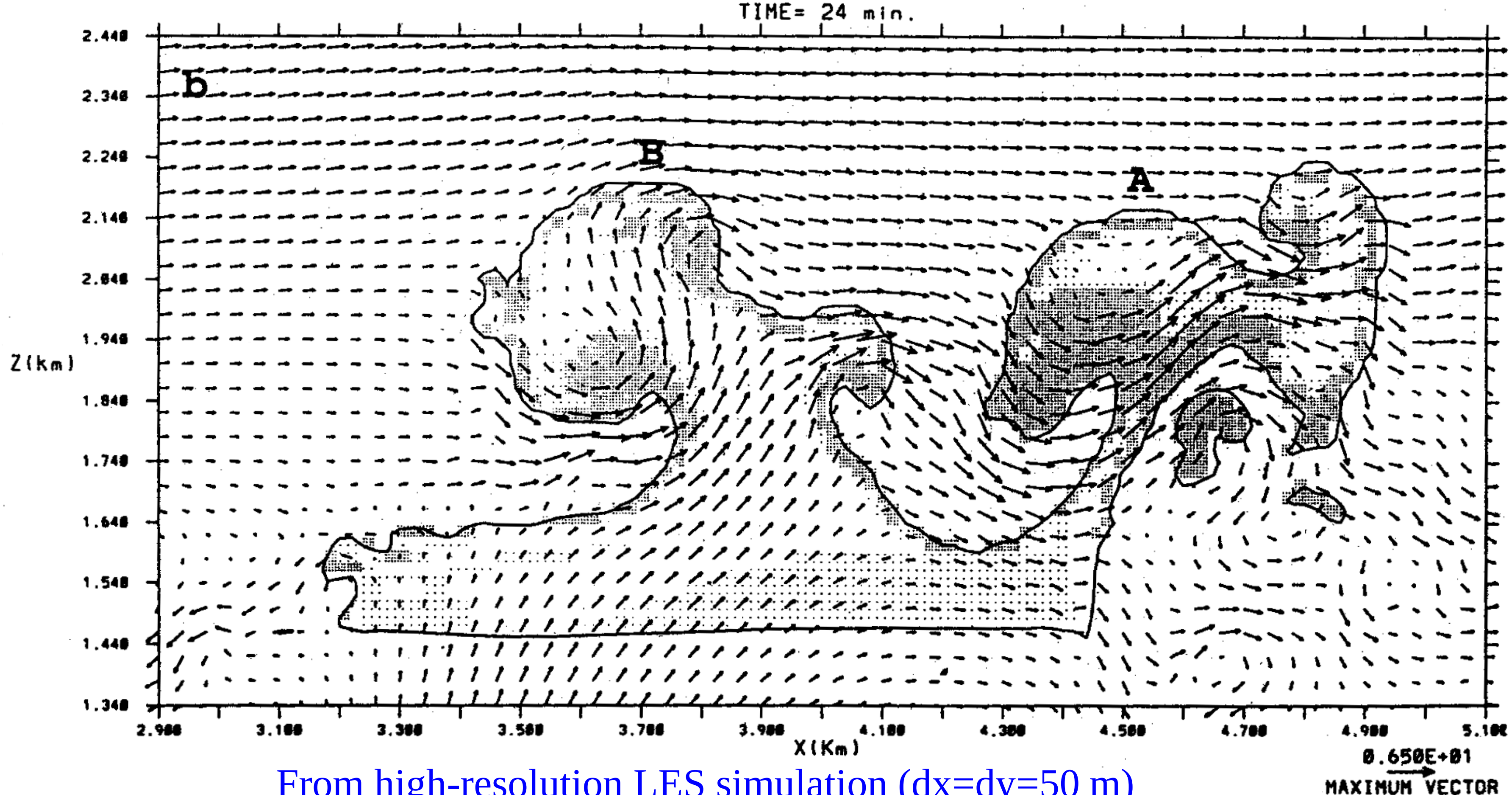
Idealised Profile

Convection in thermodynamic diagrams (2)
using equivalent Potential Temperature and saturated equivalent Potential Temperature



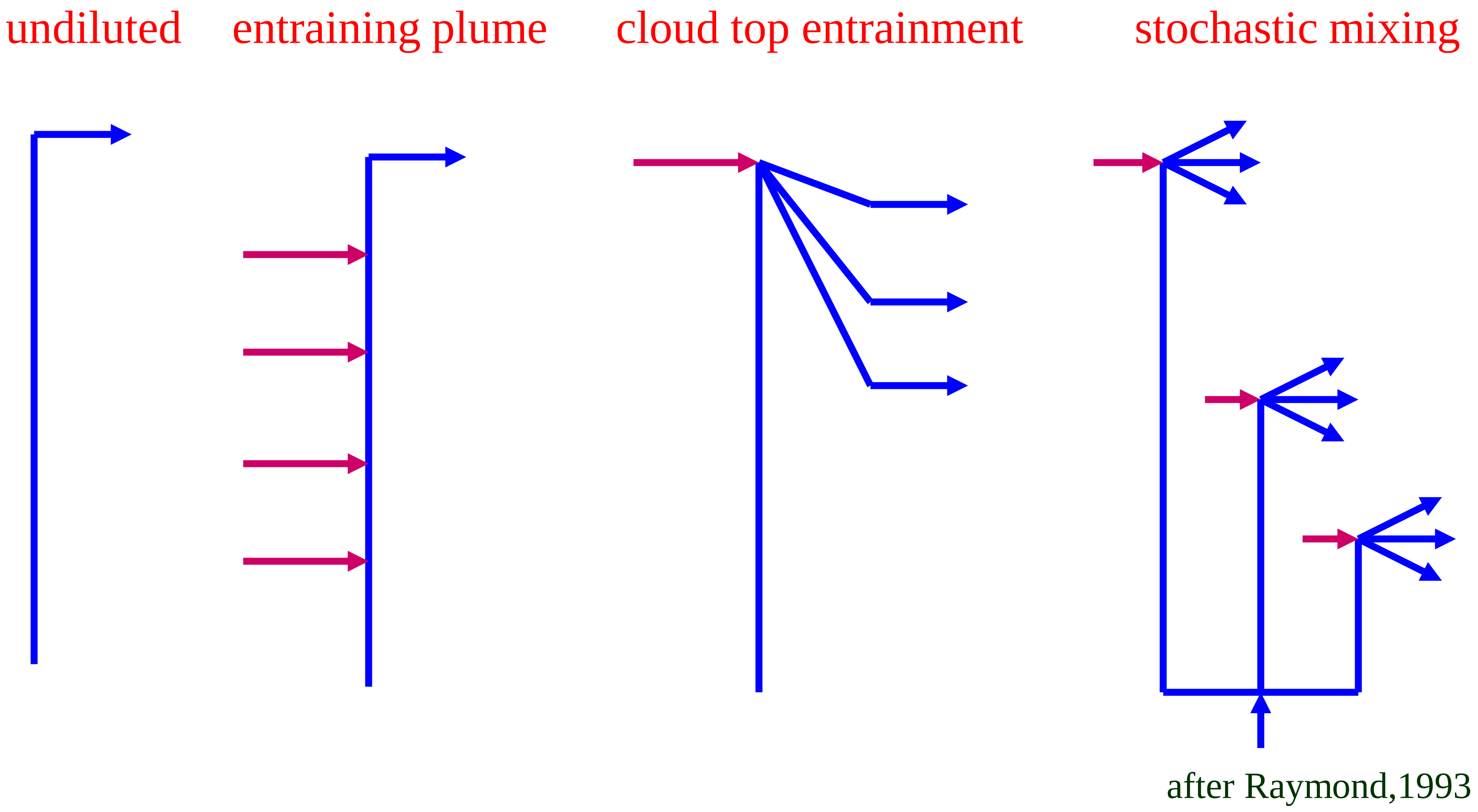
Note that no CAPE is available for parcels ascending above 900 hPa and that the tropical atmosphere is stable above 600 hPa (**θ_e** increases) – downdrafts often originate at the minimum level of **θ_e** in the mid-troposphere.

Mixing and 3D flow
subcloud and cloud-layer Circulations

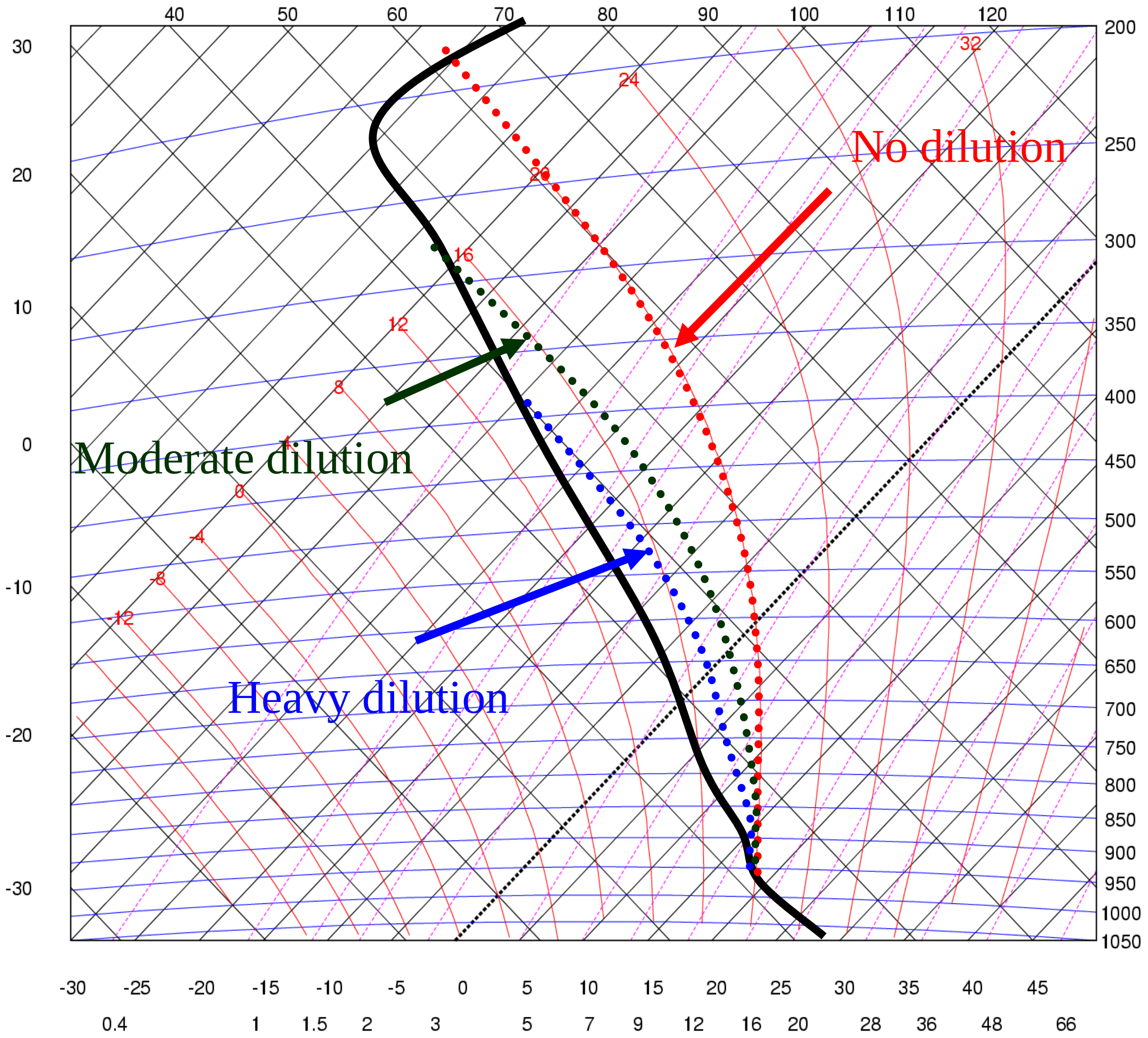


From high-resolution LES simulation (dx=dy=50 m)
Vaillancourt, You, Grabowski, JAS 1997

Mixing models



Effect of mixing on parcel ascent



Large-scale effects of convection (1) Q_1 and Q_2

Thermodynamic equation (dry static energy) :

$$\frac{\partial s}{\partial t} + \nabla \vec{v}_h s + \frac{\partial \omega s}{\partial p} = Q_R + L(c - e)$$

why use s and not T

s $=C_pT+gz$

ds/dz= C_pdT/dz+g

If dT/dz=-g/C_p (dry adiabatic lapse rate), then ds=0

Define averaging operator over area A such that:

$$\overline{\Phi} = \frac{1}{A}\int_A \Phi dA \quad \text{and} \quad \Phi = \overline{\Phi} + \Phi'$$

Apply to thermodynamic equation, neglect horizontal second order terms, use averaged continuity equation:

$$\underbrace{\frac{\partial \overline{s}}{\partial t} + \overline{\vec{v}}_h \nabla \overline{s} + \overline{\omega}\frac{\partial \overline{s}}{\partial p}}_{\text{“large-scale observable” terms}} = \overline{Q}_R + \underbrace{L(\overline{c} - \overline{e}) - \frac{\partial \overline{\omega' s'}}{\partial p}}_{\text{“sub-grid” terms}}$$

In convective regions these terms will be dominated by convection

ECMWF

Large-scale effects of convection (2) Q_1 and Q_2

Define: $$Q_1 \equiv Q_R + L(\bar{c} - \bar{e}) - \frac{\partial \overline{\omega' s'}}{\partial p}$$ Apparent heat source

Analogous: $$Q_2 \equiv L(\bar{c} - \bar{e}) + L\frac{\partial \overline{\omega' q'}}{\partial p}$$ Apparent moisture sink

$$\vec{Q}_3 \equiv \frac{\partial \overline{\omega' \vec{v}_h'}}{\partial p}$$ Apparent momentum source

This quantity can be derived from observations of the "large-scale" terms on the l.h.s. of the area-averaged equations and describe the influence of the "sub-grid" processes on the atmosphere.

Note that:

$$Q_1 - Q_2 - Q_R \equiv -\frac{\partial \overline{\omega' h'}}{\partial p}$$ with $h = s + Lq$ Moist static energy

Large-scale effects of convection (3)

vertical integrals of Q_1 and Q_2

$$\int_{Pt}^{Ps} Q_1 \frac{dp}{g} \equiv \int_{Pt}^{Ps} Q_R \frac{dp}{g} + L\,\mathrm{Pr} + \rho C_p (\overline{w'T'})_{P=Ps} = \int_{Pt}^{Ps} Q_R \frac{dp}{g} + L\,\mathrm{Pr} + HS$$

Surface Precipitation flux

Surface sensible Heat flux

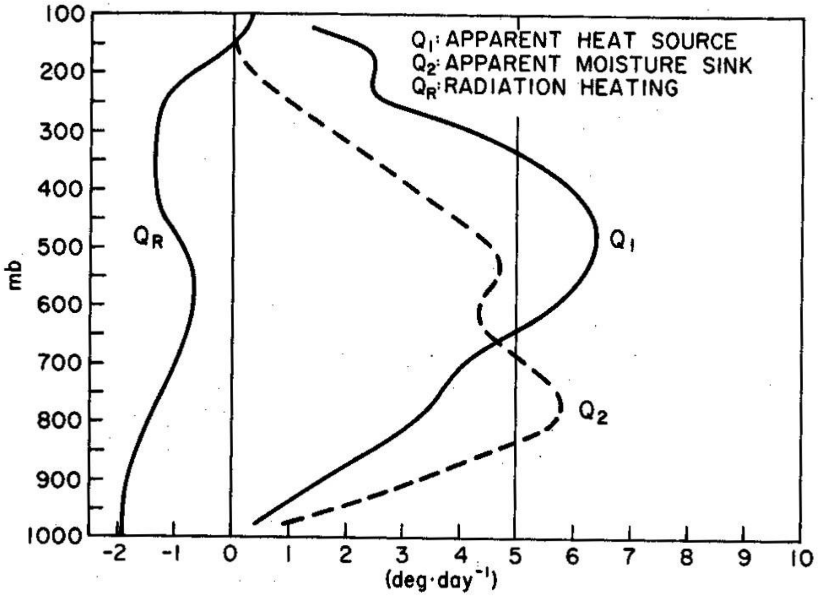
$$\int_{Pt}^{Ps} Q_2 \frac{dp}{g} \equiv L\,\mathrm{Pr} - \rho L (\overline{w'q'})_{P=Ps} = L\,\mathrm{Pr} - HL$$

Surface Precipitation

Surface latent Heat flux

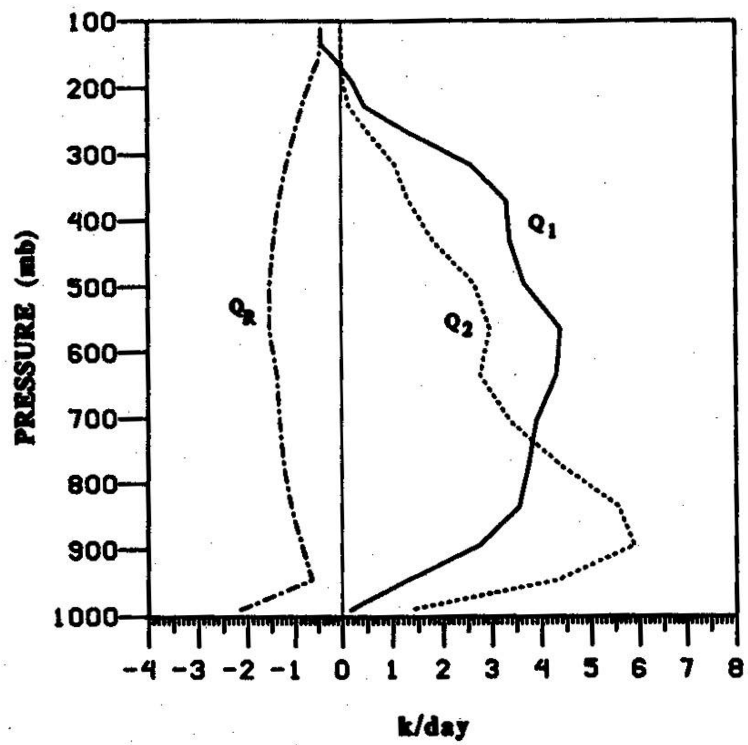
Large-scale effects of convection (3)
Deep convection

Tropical Pacific



Yanai et al., 1973, JAS

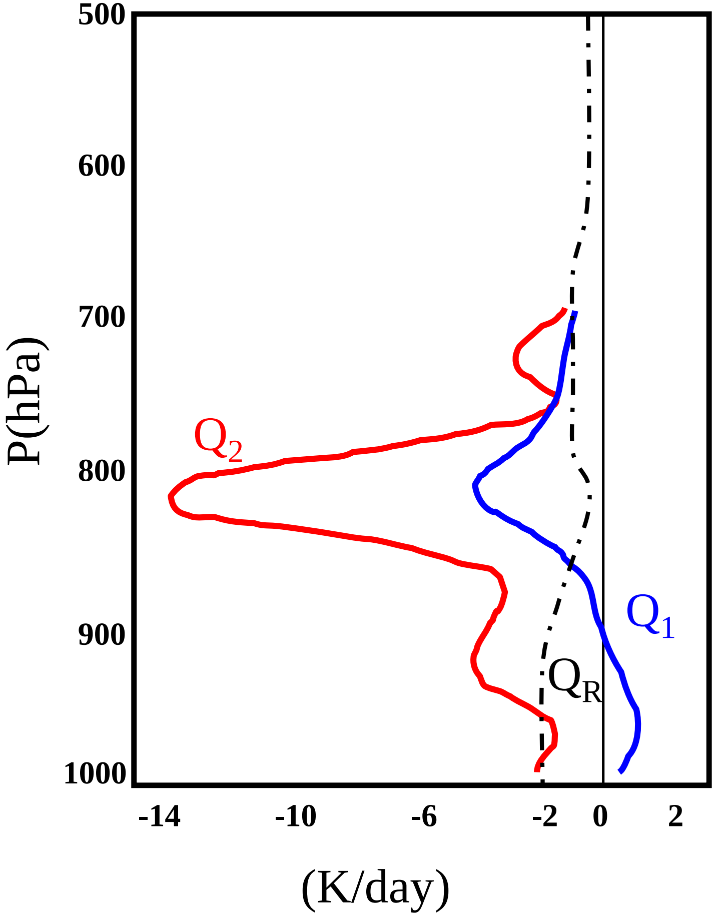
Tropical Atlantic



Yanai and Johnson, 1993

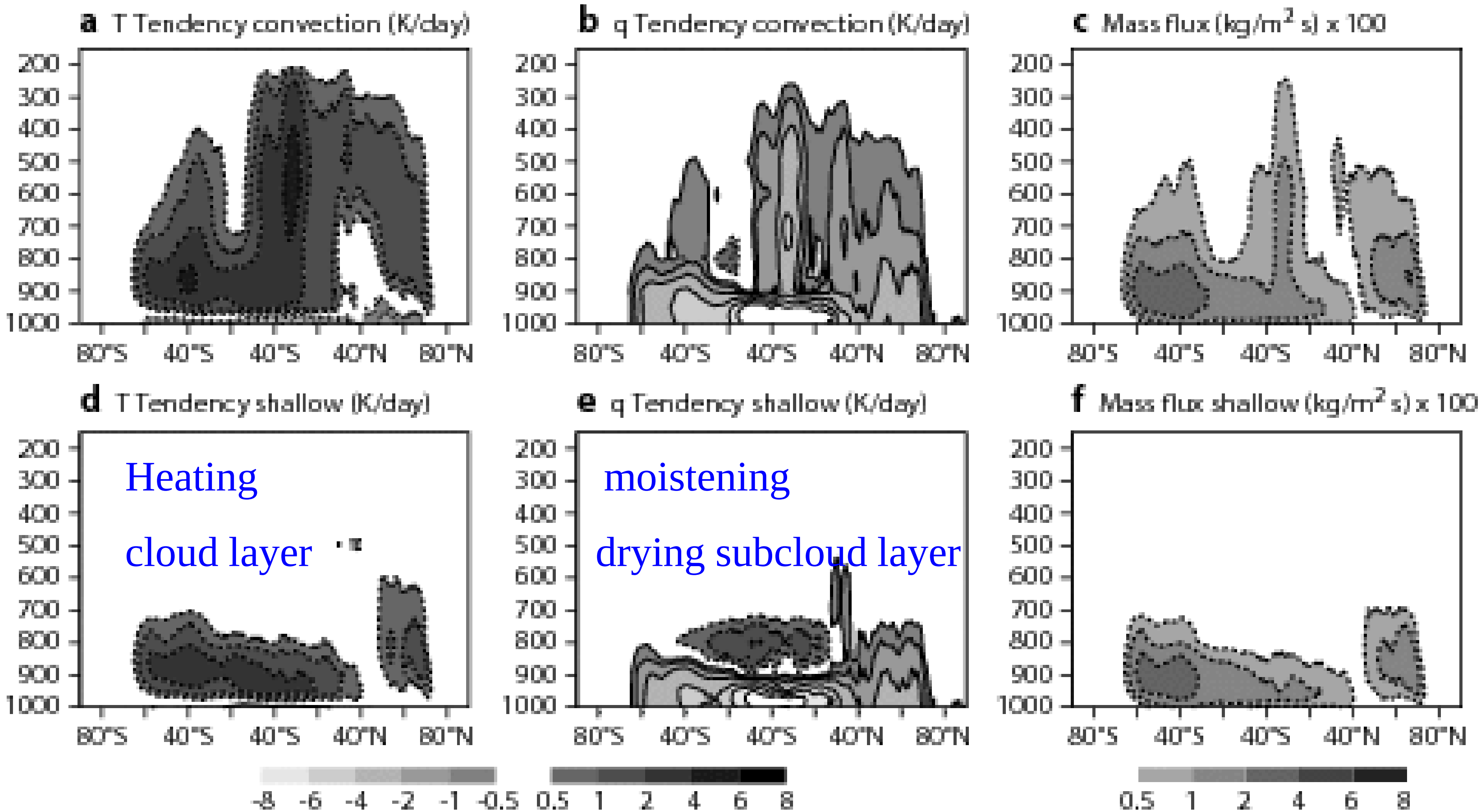
Note the typical tropical maximum of Q1 at 500 hPa, Q2 maximum is lower and typically at 800 hPa

Large-scale effects of convection (5)
Shallow convection



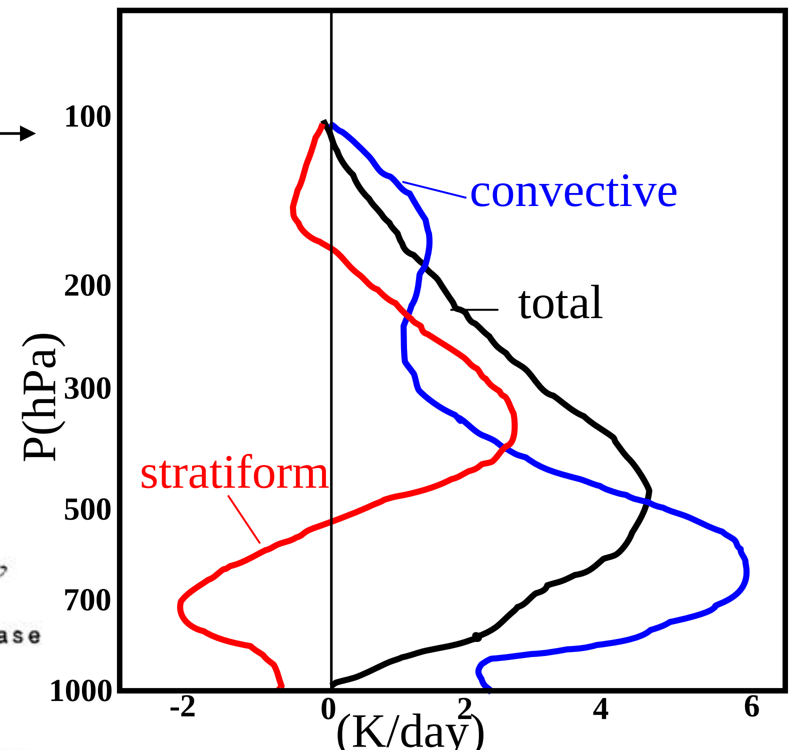
Nitta and Esbensen, 1974, MWR

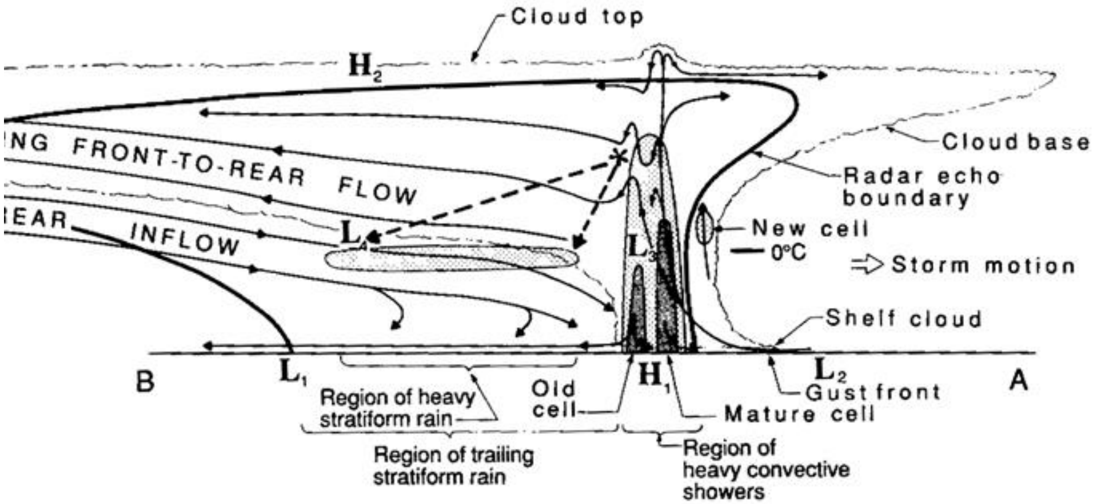
Zonal mean convective tendencies (deep & shallow) July 2013 and mass flux in IFS



Effects of mesoscale organization
The two modes of convective heating

Effects on heating



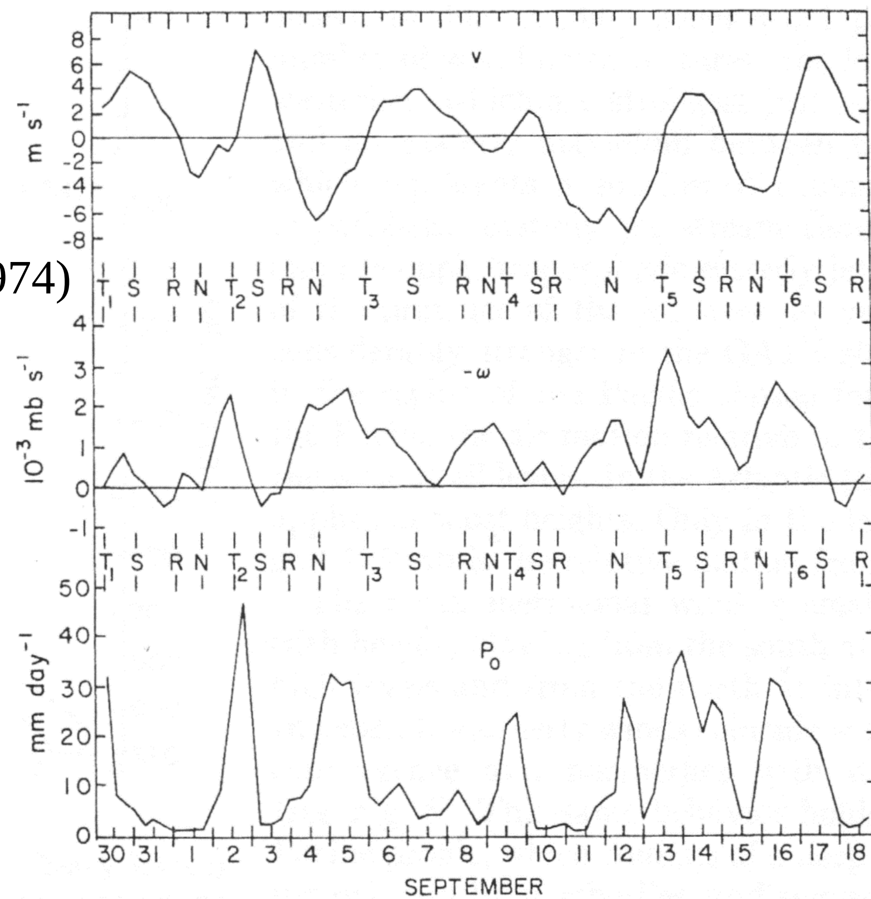


Convective quasi-equilibrium

Arakawa and Schubert (1974) postulated that the level of activity of convection is such that their stabilizing effect balances the destabilization by large-scale processes.

Observational evidence:

GARP Atlantic Tropical Experiment (1974)



v (700 hPa)

–ω (700 hPa)

Precipitation

Thompson et al., JAS, 1979

ECMWF

Summary

- Convection affects the atmosphere through condensation / evaporation and eddy transports
- On large horizontal scales convection is in quasi-equilibrium with the large-scale forcing
- Q1, Q2 and Q3 are quantities that reflect the time and space average effect of convection ("unresolved scale") and stratiform heating/drying ("resolved scale")
- An important parameter for the strength of convection is CAPE
- Shallow convection is present over very large (oceanic) areas, it determines the redistribution of the surface fluxes and the transport of vapor and momentum from the subtropics to the ITCZ