Convective and meso-scale meteorology

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I Instabilities of the moist atmosphere

(assumed to be known : notions of potential temperature, dry convection, Brünt-Vaissala oscillations)

books:
- Fundamentals of Atmospheric Physics, Salby, Academic Press
- Cloud dynamics, Houze, Academic Press
- Storm and cloud dynamics, Cotton, Bryan and van den Heever, Academic Press

Other books (advanced level):
- Thermodynamics of Atmospheres and Oceans, Curry &Webster
- Atmospheric Convection, Emanuel, Oxford Univ. Press

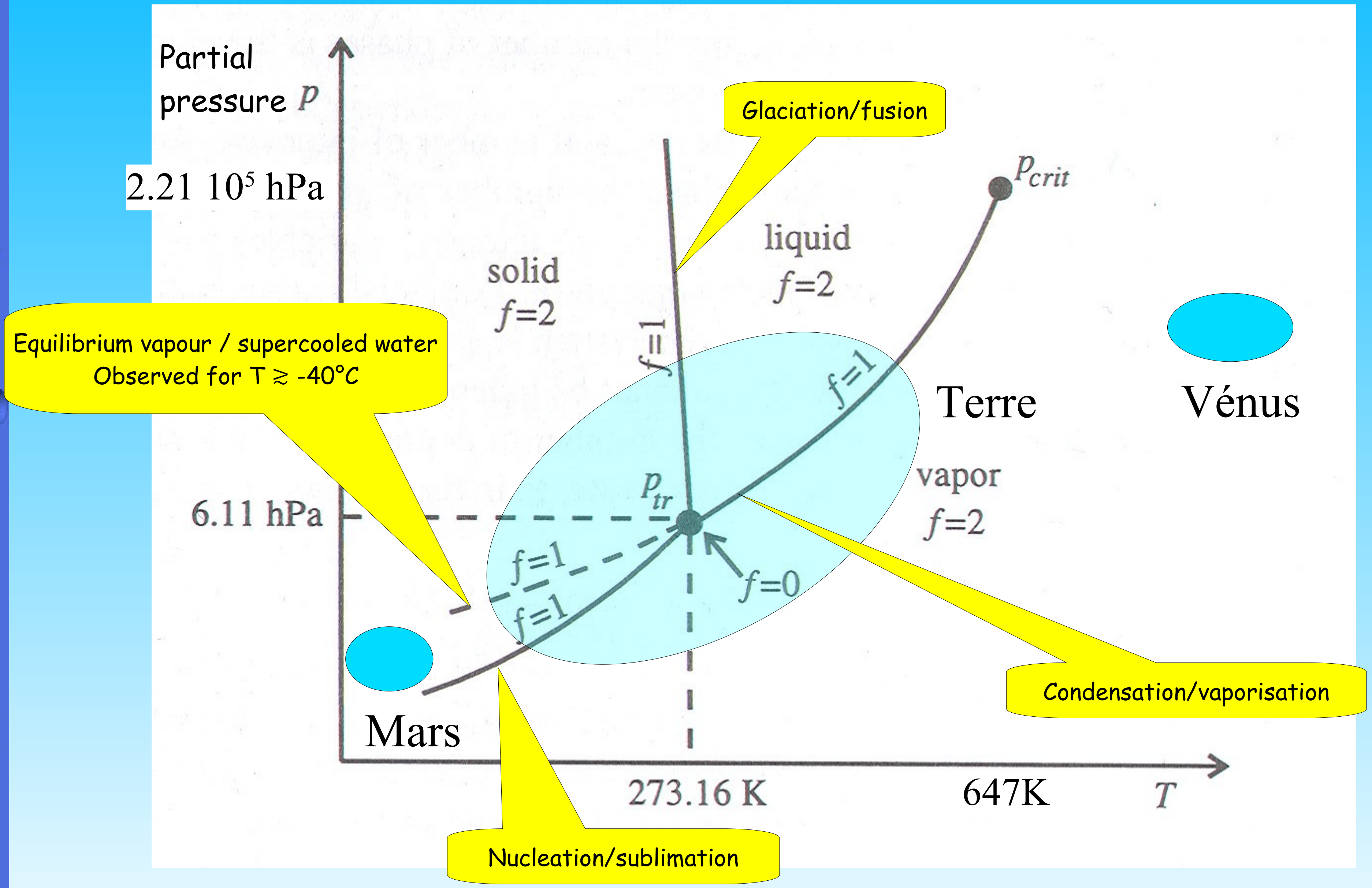
Updated 22/09/2022

Outline

- Introduction
- Thermodynamics of unsaturated moist air. Boundary layer
- Thermodynamics of saturated most air. Conditional and potential convective instabilities
- CAPE. Thermodynamic diagram
- Onset, propagation and organisation of convection

Introduction

Thermodynamic diagram of pure water



The conditions of Earth atmosphere are such that water can be found under its three phases.

Moist air thermodynamics

Two gas phases, dry air(d), water vapour (v), one liquid phase (l) and one ice phase (i)

pressure $p = p_d + e$ (dry air + water vapour pressure denoted as e)

mass mixing ratio $r = \frac{\rho_v}{\rho_d}, r_l = \frac{\rho_l}{\rho_d}, r_i = \frac{\rho_i}{\rho_d}, r_T = r + r_l + r_i$

$M_d = 29\,g$ $\qquad$ $M_v = 18\,g$

$R_d = 287\,J\,kg^{-1}\,K^{-1}$ $\qquad$ $R_v = 461.5\,J\,kg^{-1}\,K^{-1}$

$C_{pd} = 1005.7\,J\,kg^{-1}\,K^{-1}$ $\qquad$ $C_{pv} = 1870\,J\,kg^{-1}\,K^{-1}$

$C_l = 4190\,J\,kg^{-1}\,K^{-1}$ at $T > 0\,°C$ $\qquad$ $C_i = 2106\,J\,kg^{-1}\,K^{-1}$ at $T \approx 0\,°C$

$$\frac{R_d}{R_v} = \epsilon = 0.622 \qquad \frac{C_{pv}}{C_{pd}} = \beta = 1.86 \qquad \kappa = \frac{R_d}{C_{pd}} = 0.285$$

$$r = \frac{e/(R_v T)}{p_d/(R_d T)} = \epsilon \frac{e}{p-e} \qquad p_d = p\frac{\epsilon}{\epsilon + r} \qquad e = p\frac{r}{\epsilon + r} = p_d \frac{r}{\epsilon}$$

saturation pressure e^S, saturating mixing ratio r^S (function of p and T)

relative humidity $\mathrm{H} \equiv \frac{e}{e^S} = \frac{r}{r^S}\left(\frac{1 + r^S/\epsilon}{1 + r/\epsilon}\right)$

specific volume $\alpha \equiv \frac{1}{\rho} = \frac{V_a + V_l + V_i}{m_d + m_v + m_l + m_i} = \alpha_d \left(\frac{1 + r_l(\alpha_l/\alpha_d) + r_i(\alpha_i/\alpha_d)}{1 + r_T}\right) \simeq \frac{\alpha_d}{1 + r_T}$

$$\alpha \simeq \frac{R_d T}{p_d} \frac{1}{1+r_T} = \frac{R_d T}{p} \frac{1+r/\epsilon}{1+r_T}$$

Non saturated air: notion of virtual temperature $\boldsymbol{T_v \equiv T \frac{1+r/\epsilon}{1+r} \simeq T(1+0{,}608\,r)}$

Saturated air: notion of density temperature $\boldsymbol{T_\rho \equiv T \frac{1+r^S/\epsilon}{1+r_T} = T_v \frac{1+r^S}{1+r_T}}$

T_v is the temperature of dry air with the same density as moist air in the unsaturated case : $p = \rho R_d T_v$

T_ρ is the temperature of the dry air with the same density as moist air in the saturated case : $p = \rho R_d T_\rho$

In the tropical regions where r may reach 0,04, the difference between T and T_v may reach 2,5%.

T_v is always larger than T. This is not always true for T_ρ which may be smaller than T when the load in liquid water is high.

In the unsaturated case, $\frac{1}{p}\frac{dp}{dz} = -\frac{g}{R_d T_v}$

Entropy for the moist unsaturated case

We consider here transformations where r is conserved.

Entropies s_d, s_v for the dry part of the air and water vapour:

$s_d = C_{pd} \ln(T/T_0) - R_d \ln(p_d/p_0), \quad s_v = C_{pv} \ln(T/T_0) - R_v \ln(e/p_0)$

Entropy per unit mass of dry air :

$$s = s_d + r\, s_v = (C_{pd} + r\, C_{pv}) \ln(T/T_0) - R_d (1 + r/\epsilon) \ln(p/p_0) + A$$

where we have put in A (calculate it!) a number of constant terms (depending of r).

Defining $s \equiv (C_{pd} + r\, C_{pv}) \ln(\theta/T_0) + A$, the <u>potential temperature</u> is

$$\theta \equiv T \left(\frac{p_0}{p} \right)^{\kappa \frac{1 + r/\epsilon}{(1 + r\beta)}} \simeq T \left(\frac{p_0}{p} \right)^{\kappa (1 - 0{,}24\, r)} .$$

It is conserved for reversible adiabatic unsaturated transformations.

Since r is conserved, T can be replaced by T_v in the above expression.

The <u>virtual potential temperature</u> is defined as

$$\theta_v \equiv T_v \left(\frac{p_0}{p} \right)^{\frac{R_d}{C_{pd}} \frac{1 + r/\epsilon}{(1 + r\, C_{pv}/C_{pd})}} \simeq T_v \left(\frac{p_0}{p} \right)^{\frac{R_d}{C_{pd}} (1 - 0{,}24 r)} .$$

θ_v is conserved in the same conditions as θ.

Comparing θ_v for two parcel is the same, when they are brought to the same pressure as comparing their virtual temperature and hence their density.

<u>The θ_v profile determines stability for moist unsaturated atmosphere.</u>

Thermodynamics of moist saturated air

Potential instability.

Conditional instability.

Moisture condensation

The saturation partial pressure depends on the temperature through the Clausius-Clapeyron law $d\ln(e_S)/dT = L/R_vT^2$

Approximate formula (in hPa)

$e_S^{liquide} = 6{,}112 \exp(17{,}67\ T/(T+243.55))$

$e_S^{glace} = \exp(23{,}33086-6111{,}72784/T + 0{,}15215 \ln(T))$

Exemples of saturating ratios

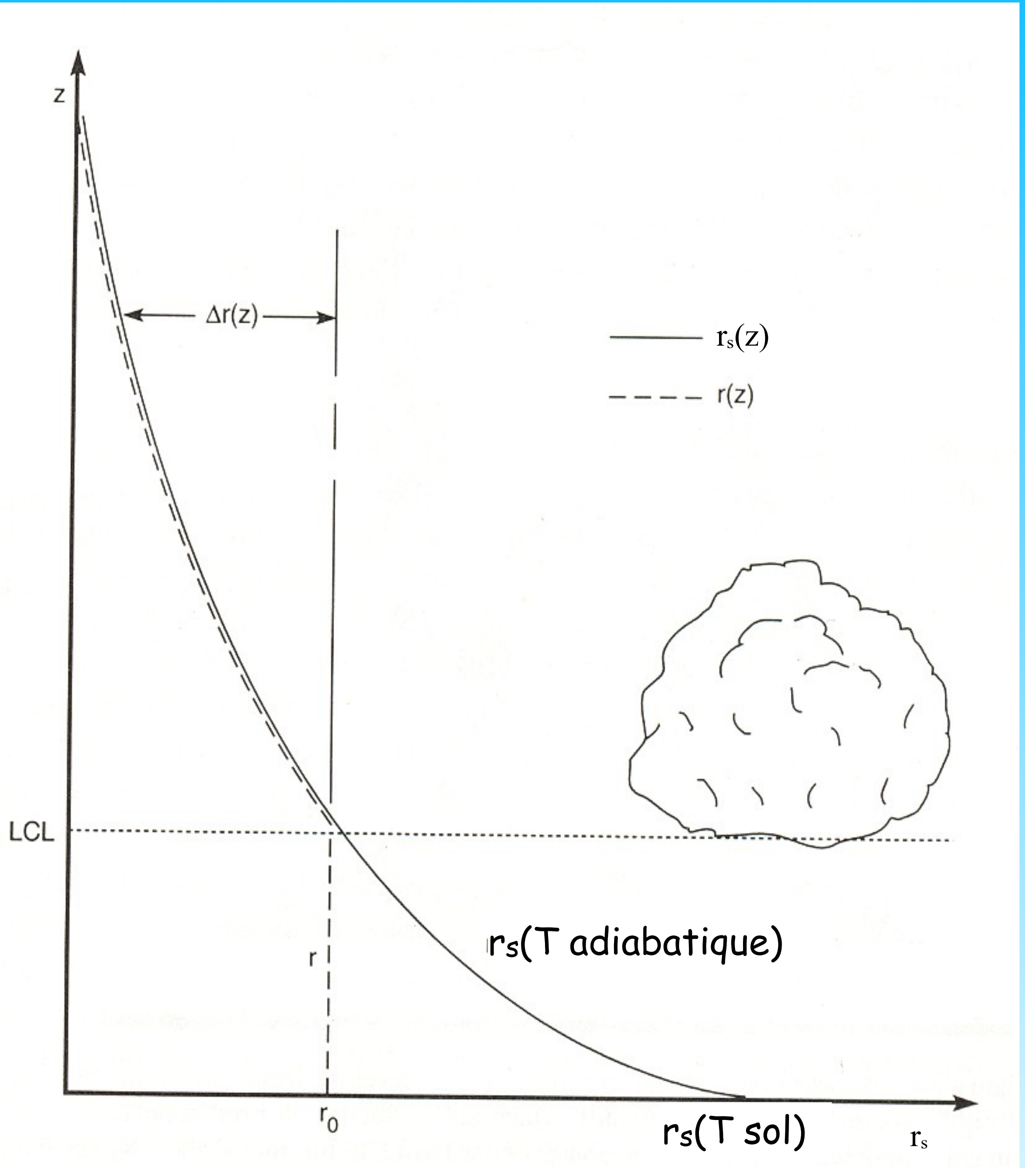
at 1000hPa and T=20°C: r_s = 14,5 g/kg,

at 800 hPa (2000m) and T = 7°C: r_s =7,8 g/kg,

at 500 hPa and T=-30°C r_s =0,47 g/kg,

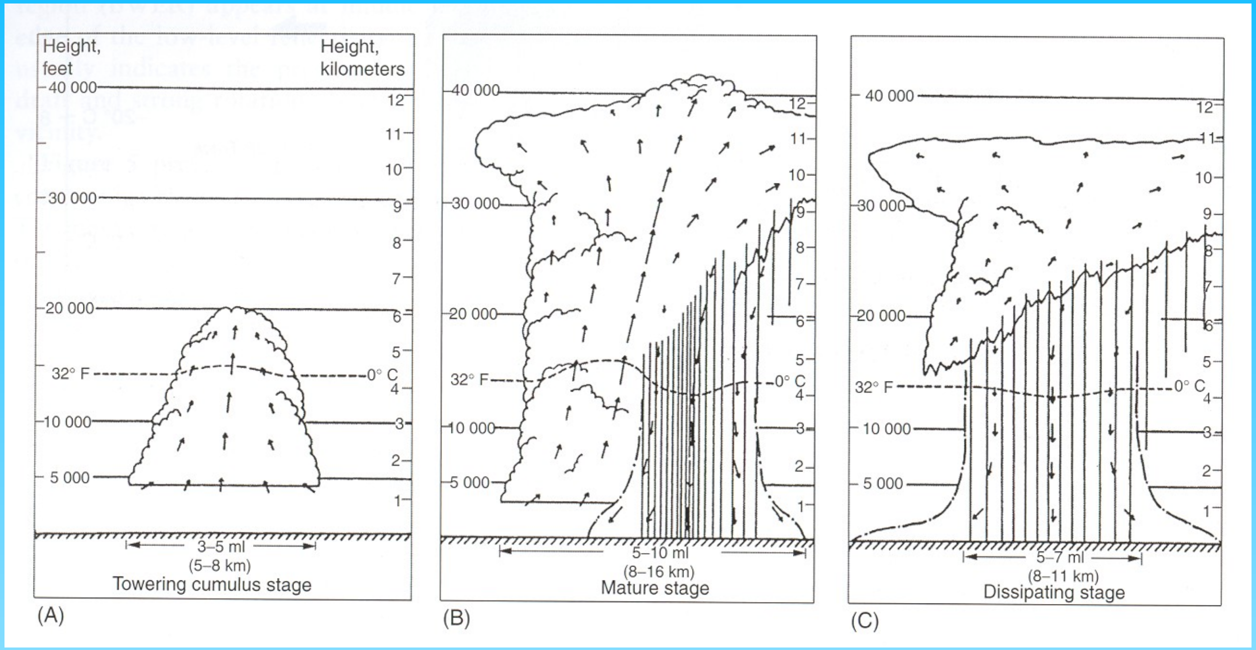
at 100 hPa and T =-80°C r_s =0,003 g/kg,

(the atmospheric water content is divided by approximately 4 orders of magnitude between the ground and 100 hPa in the tropics)



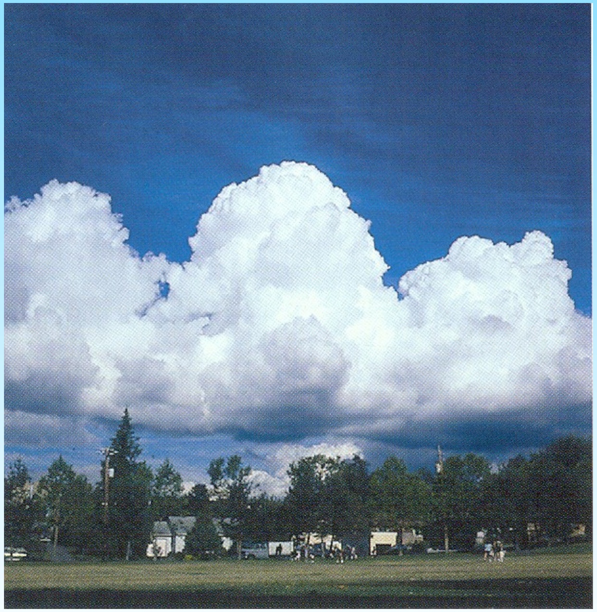
LCL (lifting condensation level): level at which parcels rising from the ground condensate

Formation of convective clouds



cumulus

cumulonimbus

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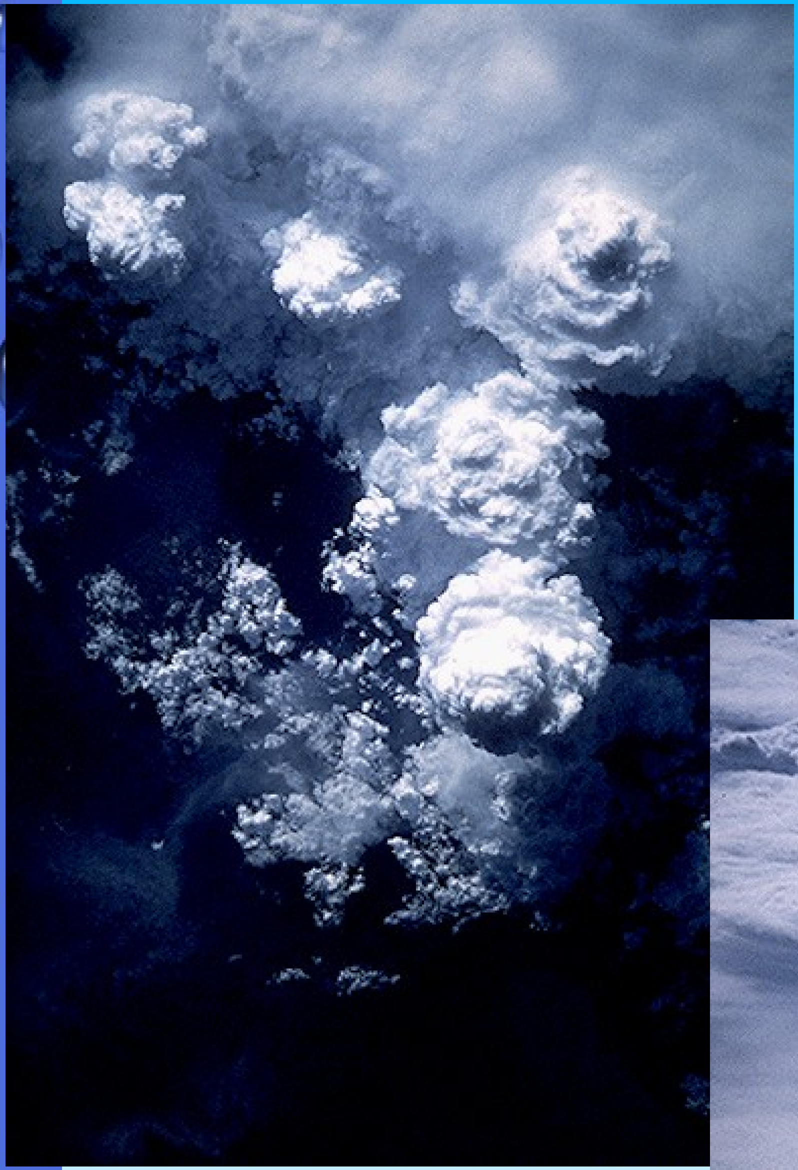
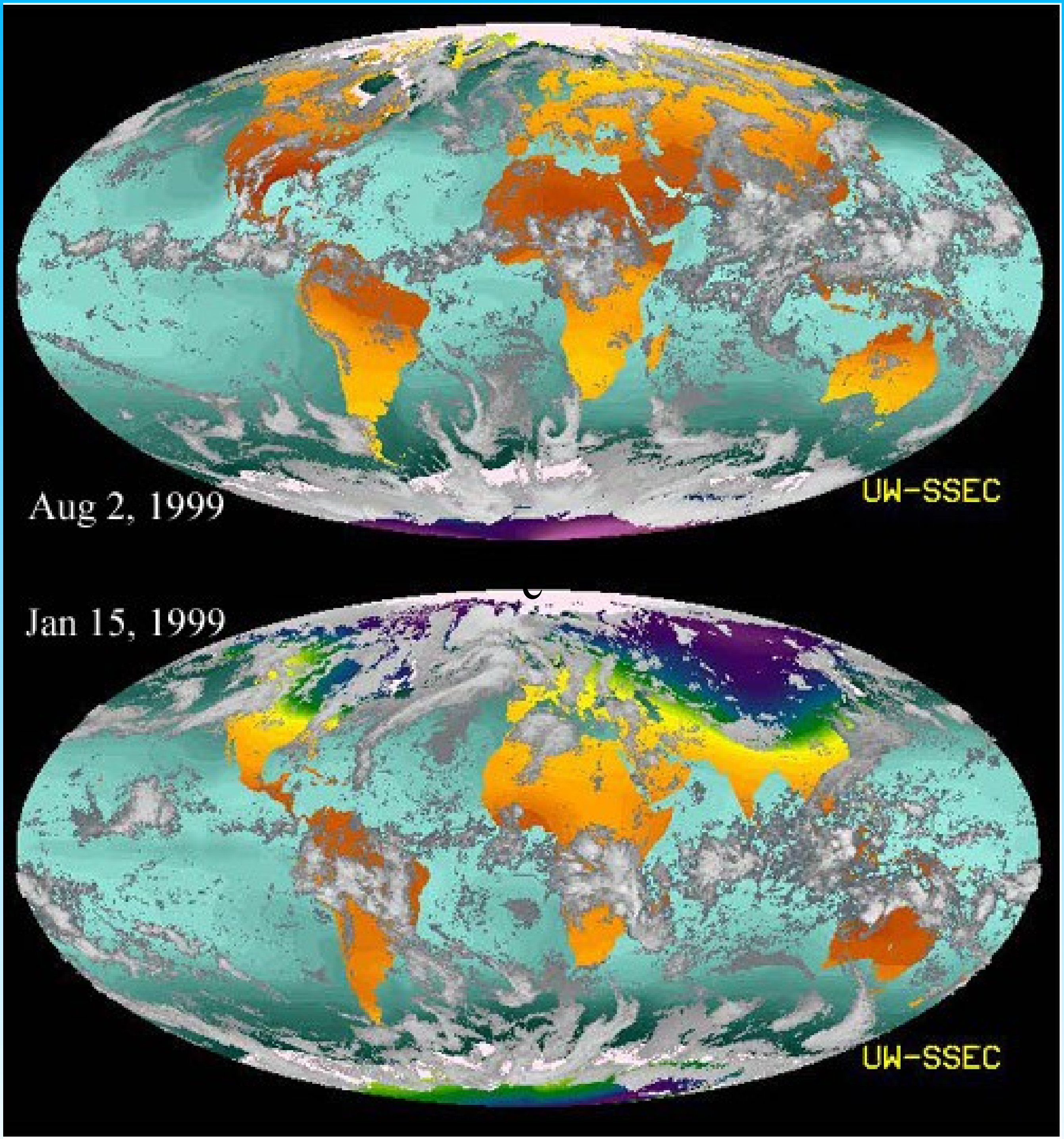


Convective clouds above Brazil
(pictures taken aboard a space shuttle)

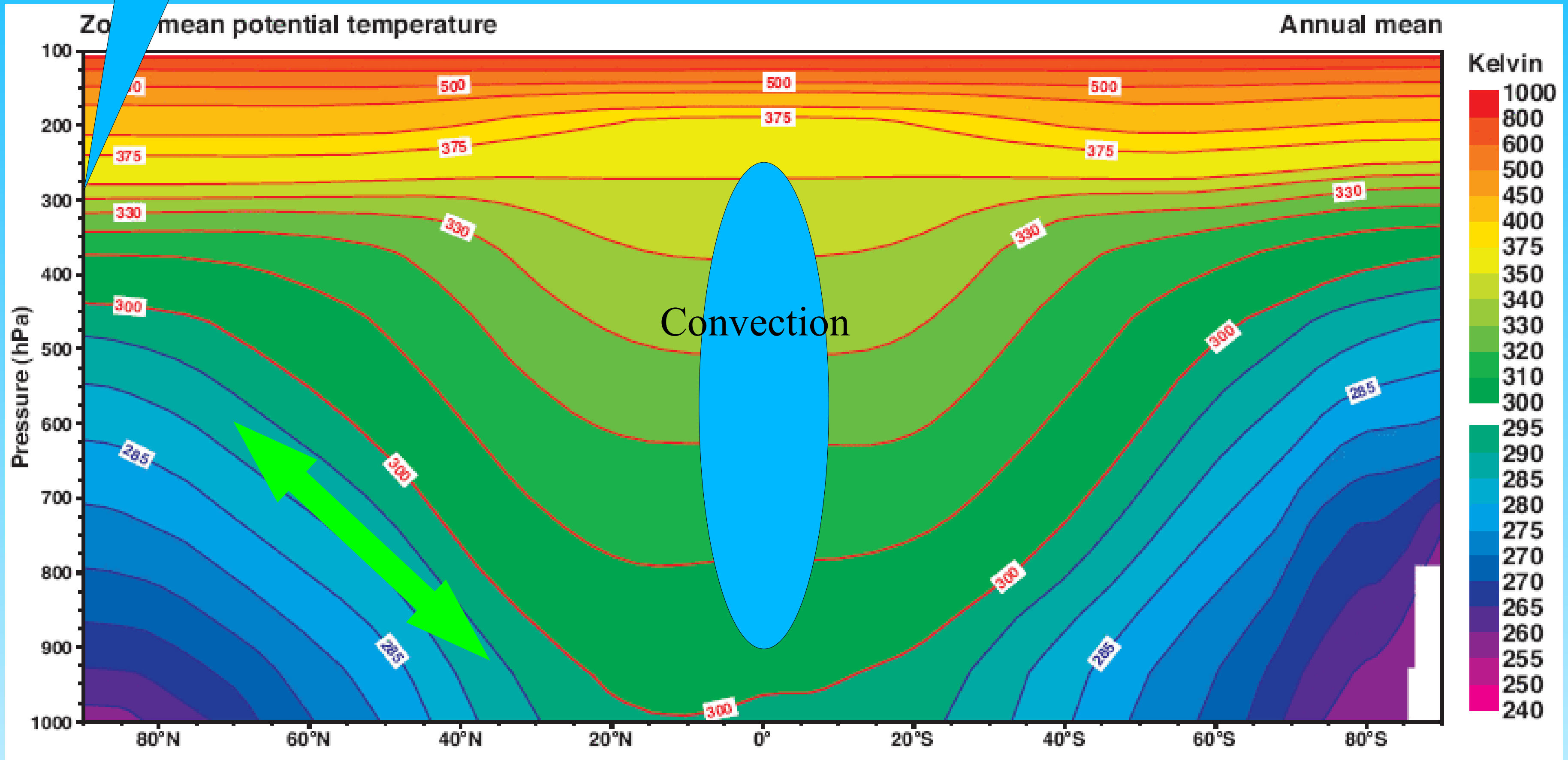
Cloud cover
ISSCP data

comparison
January-July



Beware of error in
ERA-40 atlas.
200 HPa should be there

ECMWF ERA-40 atlas



In the mid and high latitudes, poleward isentropic motion is accompanied by upward motion. Hence the cloud bands.

In the tropic, vertical upward motion needs heating. Convection organizes as compact clusters.

Moist saturated air thermodynamics – Basic laws

Equilibrium of temperature T, and free energy g at constant pressure between the two phases, with $g=u+p\alpha-Ts=h-Ts,\quad h=u+p\alpha,\quad u$ et h being only function of T for a perfect gas and with $p=e^S$ for the water vapour phase.
The latent heat is $L=h_v^S-h_l=T\left(s_v^S-s_l\right)$

Kirchhoff law

$$dL=dT\left[\left(\frac{\partial h_v}{\partial T}\right)_p-\left(\frac{\partial h_l}{\partial T}\right)_p\right]+dp\left[\cancel{\left(\frac{\partial h_v}{\partial p}\right)_T}-\left(\frac{\partial h_l}{\partial p}\right)_T\right]$$

Perfect gas law

$$=dT\left[C_{pv}-C_l\right]+dp\left[\cancel{-\alpha_l}-\cancel{p\left(\frac{\partial \alpha_l}{\partial p}\right)_T}\right]$$

Negligible specific volume of the condensed phase

$$\boldsymbol{\frac{dL}{dT}=C_{pv}-C_l}$$

vaporization $L_0=2{,}5\times10^6$ J kg^{-1} à 0°C.

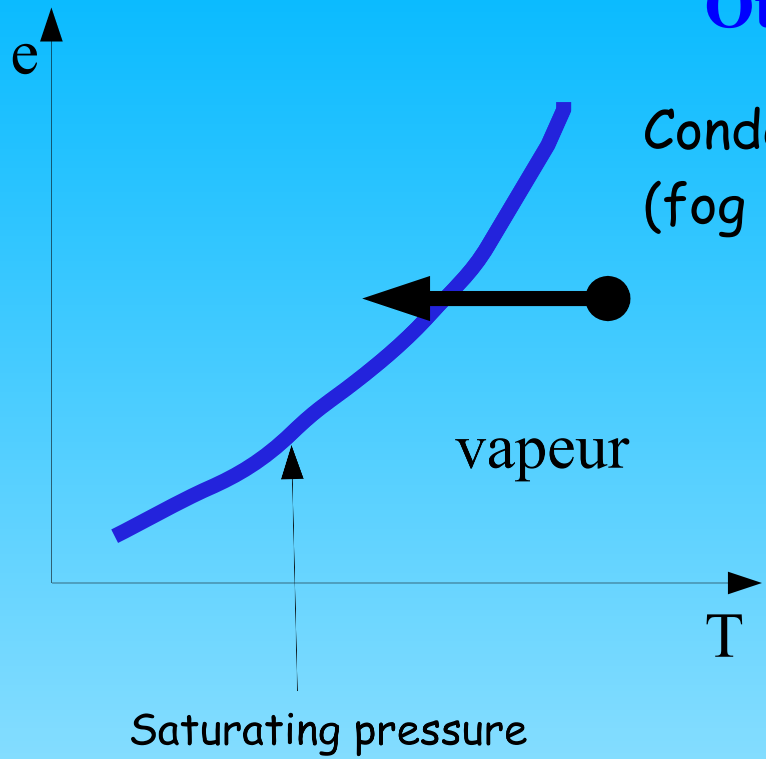
Clausius-Clapeyron law

For a variation of the equilibrium of the two phases: $d\,g_v=d\,g_l$
Using the definition of g and the first law of thermodynamics $T\,ds=du+p\,d\alpha$

$$-s_v\,dT+\alpha_v\,de^S=-s_l\,dT+\alpha_l\,de^S$$

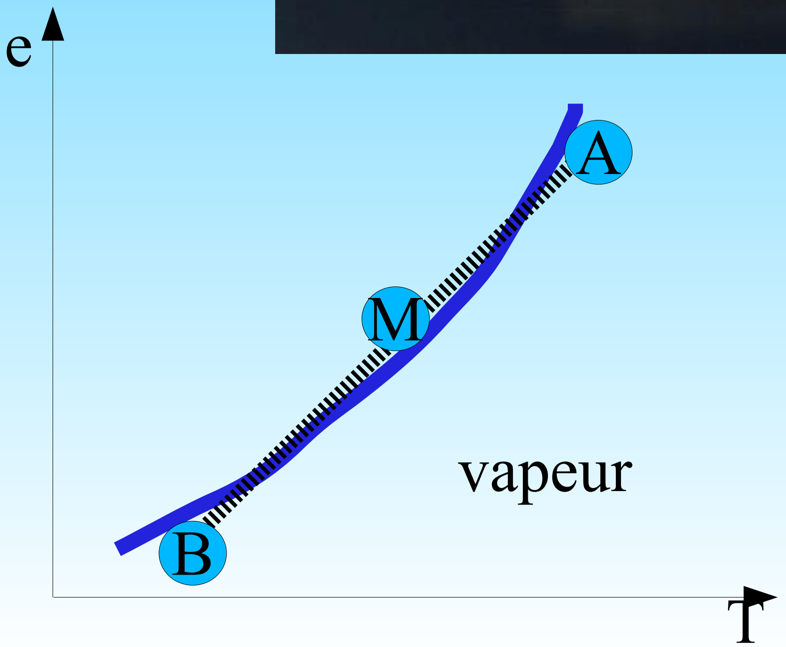
$$\frac{\boldsymbol{de^S}}{\boldsymbol{dT}}=\frac{s_v-s_l}{\alpha_v-\alpha_l}=\frac{\boldsymbol{L}}{\boldsymbol{T\left(\alpha_v-\alpha_l\right)}}\approx\frac{L\,e^S}{R_v\,T^2}$$

Other forms of condensation



Condensation by isobaric cooling
(fog and dew point)

Condensation by mixing of warm moist air (A) with cold dry air (B) (generation of contrails and fog above lakes)



Other types of clouds

Altitude clouds

Alto-cumulus

Contain liquid droplets between 2000 and 6000 m in mid-latitudes. Cluster into compact herds. They are often, during summer, precursors of late afternoon and evening developments of deep convection.

Cirrus

Composed of ice, rarely opaque.
Are formed above 6000m in mid-latitudes.
They are often precursors of a warm front.
In the tropics are formed as remains of anvils or by in situ condensation of rising air, up to the tropopause.

Other types of clouds

Low stratiform clouds

Strato-cumulus

Composed by water droplets, opaque or very opaque, base under 2000m, associated with weak precipitations

Nimbo-stratus

Very opaque low clouds, undefined base, associated with persistent precipitations, snow by cold weather

Stratus

Low clouds with small opacity, undefined base under 2000m or at the ground (fog)

Equivalent potential temperature

For a parcel of humid air, the entropy per unit mass of dry air is

$s = s_d + r\, s_v + r_l\, s_l$ with r_l and r_v the mixing ratios for liquid and vapour water

$s_d = C_{pd} \ln(T/T_0) - R_d \ln(p_d / p_0)$ for the dry air,

$s_v = C_{pv} \ln(T/T_0) - R_v \ln(e / p_0)$ for water vapour,

$s_l = C_l \ln(T/T_0)$ for liquid water.

Using $L = T(s_v^S - s_l)$ and $\mathrm{H} = e/e^S$,and after a few manipulations (using $r_T = r_l + r_V$):

$$s = s_d + r\, s_v + r_l\, s_l = s_d + r(s_v - s_v^S) + r(s_v^S - s_l) + r_T\, s_l = s_d + r_T\, s_l + \frac{L r}{T} + r(s_v - s_v^S)$$

$$= (C_{pd} + r_T C_l) \ln(T/T_0) - R_d \ln(p_d/p_0) + \frac{L r}{T} - r R_v \ln(\mathrm{H})$$

The equivalent potential temperature θ_e can be defined such that

$$s = (C_{pd} + r_T C_l) \ln(\theta_e / T_0)$$

hence $\boldsymbol{\theta_e = T \left(\frac{p_0}{p_d}\right)^{R_d/(C_{pd} + r_T C_l)} (\mathrm{H})^{-r R_v/(C_{pd} + r_T C_l)} \exp\left(\frac{L r}{(C_{pd} + r_T C_l) T}\right)}$

This quantity is conserved under both saturated and non saturated moist adiabatic transforms where condensates are carried aloft.

For a saturated parcel, $\theta_e = T \left(\frac{p_0}{p_d}\right)^{R_d/(C_{pd} + r_T C_l)} \exp\left(\frac{L r^S}{(C_{pd} + r_T C_l) T}\right)$

function of (T, p_d, r_T). case,

Since $r = r_T$ in the unsaturated, θ_e is always a function of (T, p_d, r_T)

Potential instability under the presence of moisture

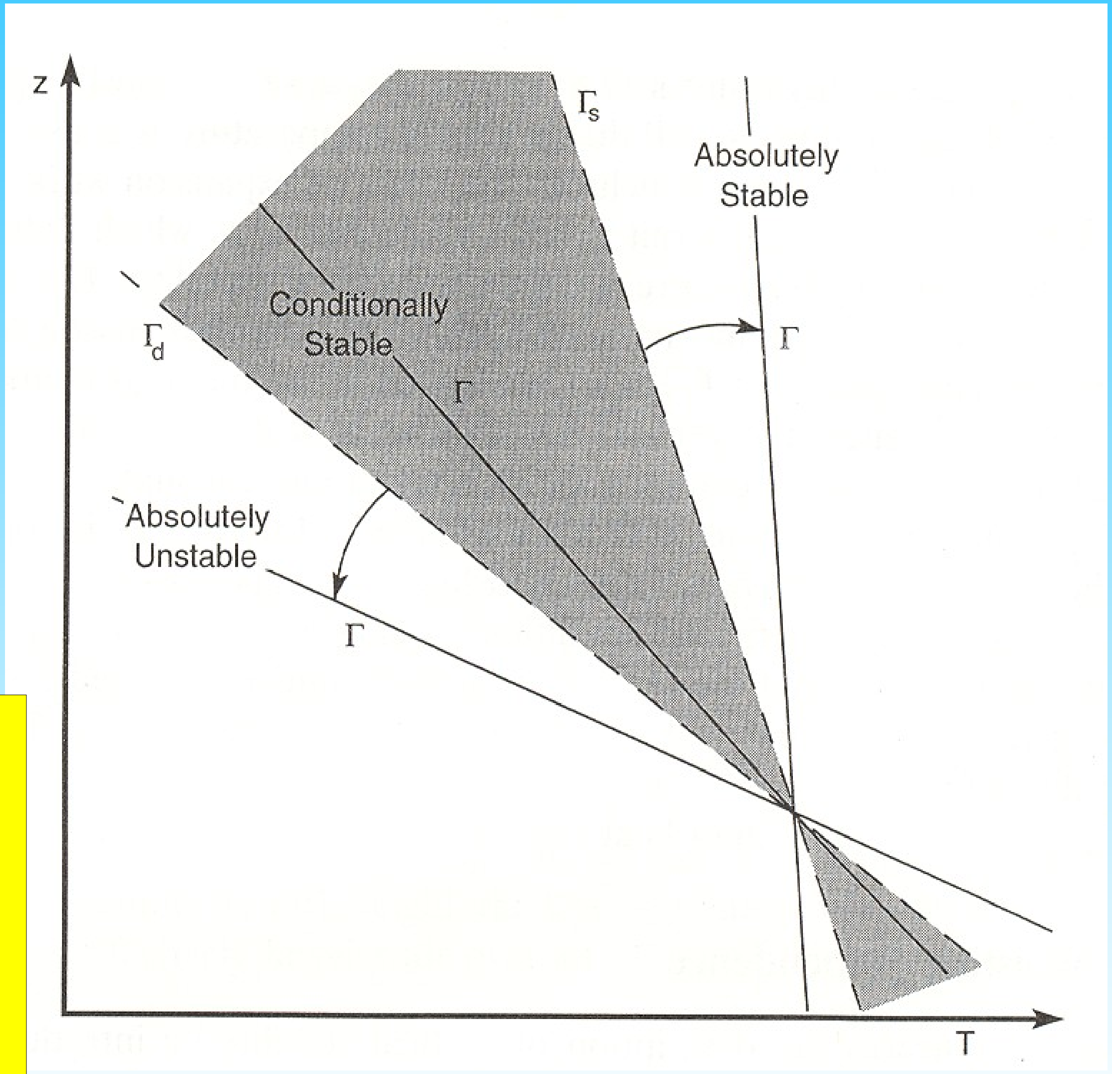
The conserved quantity for a moist saturated adiabatic is the equivalent potential temperature

$$\theta_e(T,p) \approx \theta \exp \frac{L r^S(T,P)}{C_p T}$$

$$\frac{\partial \theta_e}{\partial T} = \frac{\theta_e}{T}\left(1 - \frac{L r^S}{C_p T}\right) > 0$$

Γ_d : dry adiabatic (constant θ)

Γ_s: moist saturated adiabatic (constant θ_e)



Instability conditions compared to that of dry air .
The instability for unsaturated air where d θ_e/dz < 0 is potential because it does not show up until the air is saturated.

Simplification: we neglect the effect of water vapour on air density (virtual temperature effect), hence

$T_1 > T_2 \iff \rho_1 < \rho_2$

For saturated air

$\Theta_{e1} > \Theta_{e2} \iff T_1 > T_2$

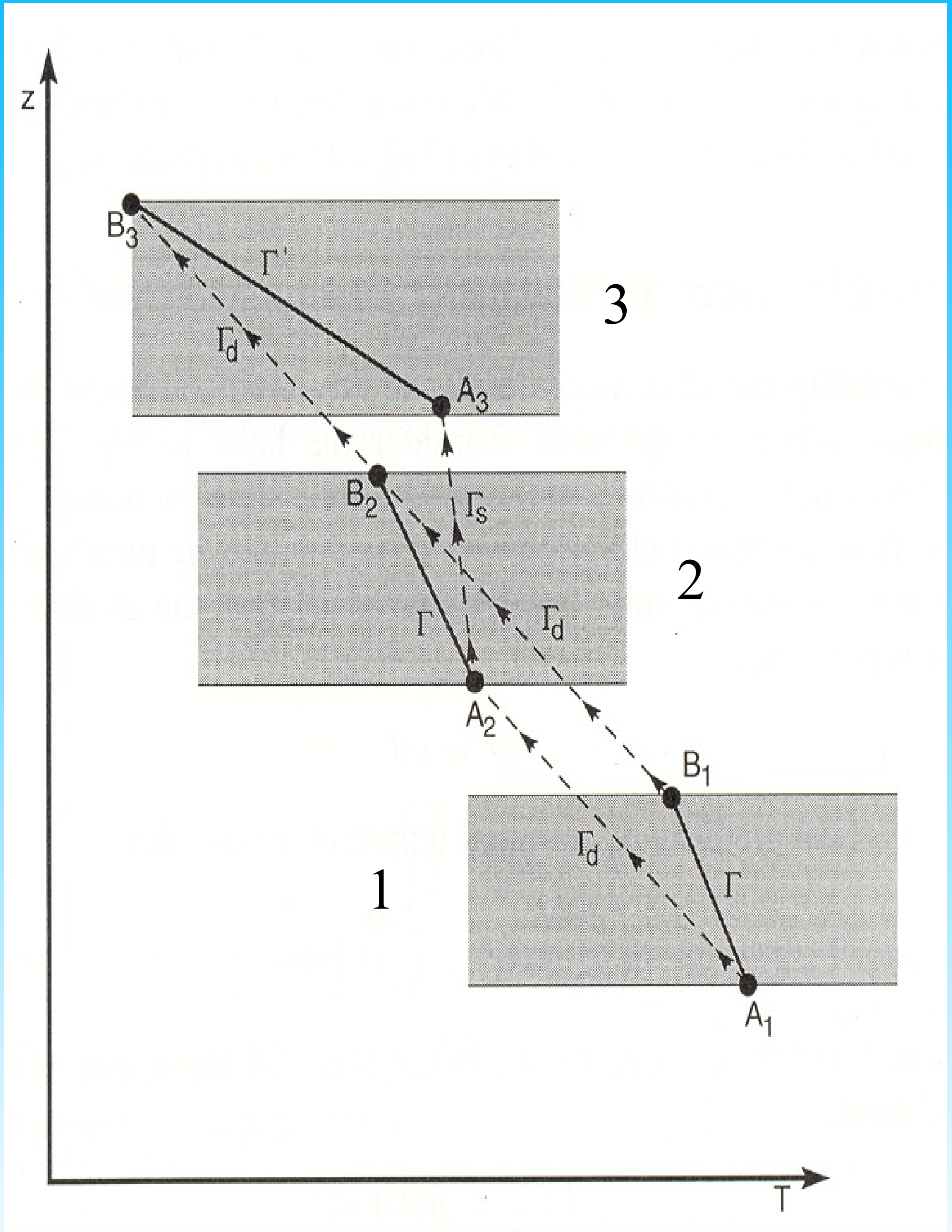
Potential instability

Potential instability appears when

$\frac{d\theta}{dz} > 0$ but $\frac{d\theta_e}{dz} < 0$

It is realised when a potentially unstable layer is lifted,for instance by crossing some orographic zone ou due to frontal transport. as soon as the first bottom layer gets saturated, convection is initiated.

In the example, the layer stay unsaturated between position 1 and 2 but gets just saturated in A2. During the motion from 2 to 3. The bottom of the layer is saturated all the way while the top stay unsaturated. Intermediate points are first unsaturated then become saturated somewhere between 2 and 3. As a result, between 2 and 3, the layer becomes unstable, starting from the bottom when it saturates.



Complementary note: Simplified calculation of the saturated moist gradient

In a saturated adiabatic transform, and for a unit mass of dry air:

$$\left(C_p + r^s C_{pv} + r_l C_{pl}\right) dT + L\, d r^s - R_d T\, d\log p_d - r^s R_v\, d\log e^S = 0\,.$$

(neglected terms in green)

Using the ideal gas law,

$$R_d T\, d\log p_d = \frac{1}{\rho_d}\, dp = -g\, dz\,.$$

Then we need to write the variation of r^S as a function of T et p :

$$d r^s = \left(\frac{\partial r^s}{\partial T}\right) dT + \left(\frac{\partial r^s}{\partial p}\right) dp\,.$$

Using once again the hydrostatic law, we obtain :

$$\left(C_p + L\left(\frac{\partial r^s}{\partial T}\right)\right) dT = -g\left(1 - \rho L\left(\frac{\partial r^s}{\partial p}\right)\right) dz\,,$$

hence

$$\Gamma_S = \Gamma_d \frac{1 - \rho L\left(\frac{\partial r^s}{\partial p}\right)}{1 + \frac{L}{C_p}\left(\frac{\partial r^s}{\partial T}\right)} \approx \Gamma_d \frac{1 + \frac{L r^S}{R_d T}}{1 + \frac{L^2 r^S}{R_v T^2}}$$

Saturation equivalent potential temperature

This temperature is defined for unsaturated ambiant air and is the equivalent temperature for saturated air at the same temperature and pressure as the ambiant air

$$\theta_e^* = \theta_e(T, p_d, r^S(T, p_d)) = T\left(\frac{p_0}{p_d}\right)^{R_d/(C_{pd}+r_T C_l)} \exp\left(\frac{L\, r^S(T, p_d)}{(C_{pd}+r_T C_l)\, T}\right)$$

This temperature is only function of T and p_d

For a saturated ambiant air, it is identical to θ_e

This temperature determines the onset condition of deep convection. The comparison between unsaturated ambiant air and a rising saturated parcel conserving θ_e cannot be done on θ_e because the moisture contribution to this quantity is different for the ambiant air and the rising parcel. If the ambiant air is brought to the same saturation conditions as the rising parcel, while preserving its pressure and temperature, it is guaranteed that an equality between θ_e^* of the ambiant air and θ_e of the rising parcel leads to an equality of temperatures and that an inequality of ambiant θ_e^* and cloud parcel θ_e leads to an inequality of temperatures of the same type because $\frac{\partial \theta_e^*}{\partial T} > 0$ (check it!).

We neglect here the effects of a lower density of water vapour with respect to dry air.
We neglect also the effect of removing the precipitations from the rising air parcel.
Such effects are less important than those related to latent heat within a convective cloud.
They can be taken into account in a more complete theory (Emanuel's book)

Conditional instability

When an air parcel is displaced verticaly, it first rises along a dry adiabatic and hence reaches its condensation level (LCL).
It then continue to rise as a saturated parcel following a pseudo-adiabatic path.
Next it meets its neutral buoyancy level (LFC) when its temperature θ_e equals θ_e^* of the ambiant air.

<u>At this time, the parcel temperature equals that of the ambiant air.</u>

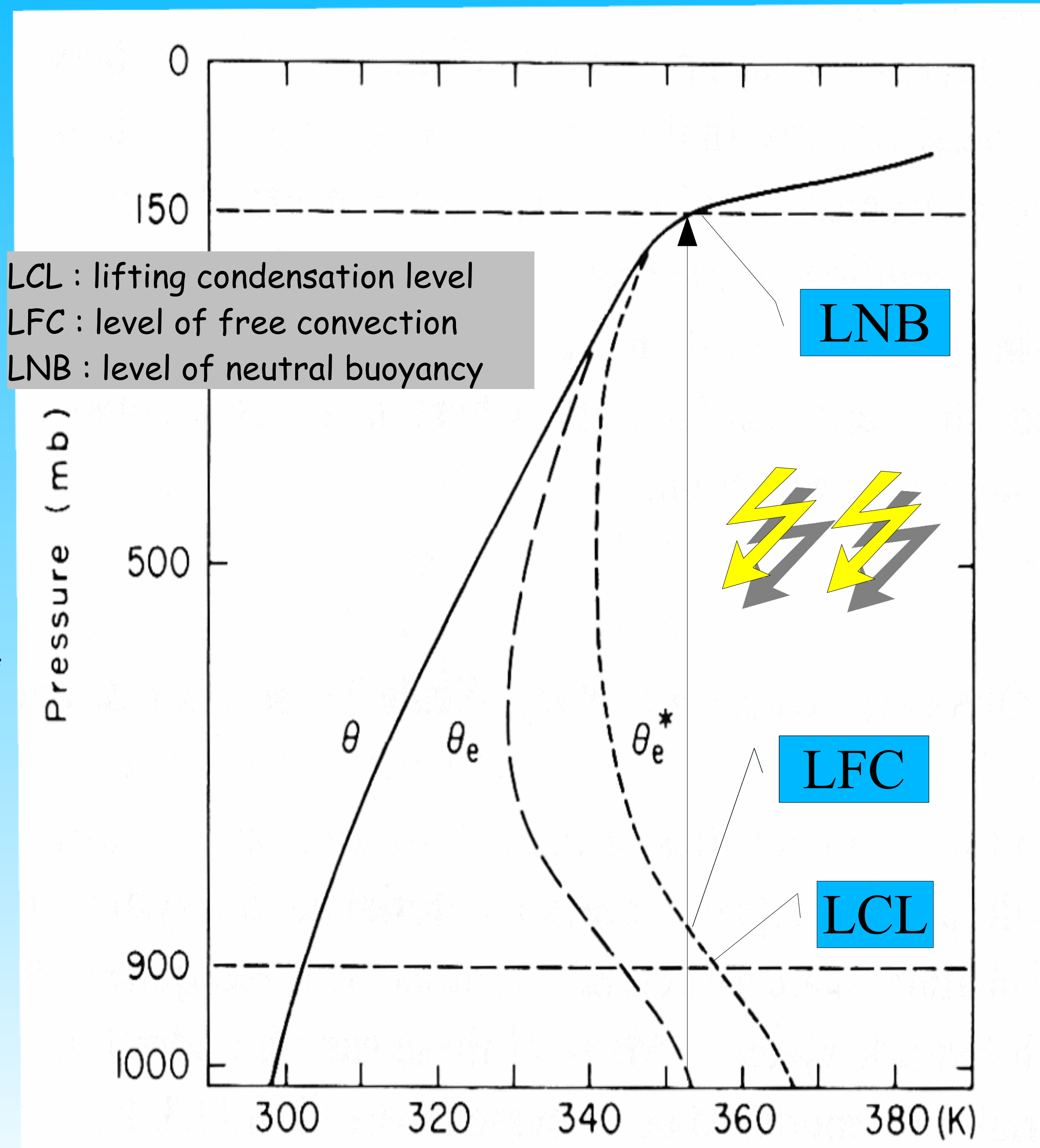
The ascent continues if $\frac{d\,\theta_e^*}{d\,z}<0$

Hence it is the profile θ_e^* and non that of θ_e which determines the stability since an inequality of the saturated temperature leads, at the same pressure, to an inequality of the same type on the temperatures since

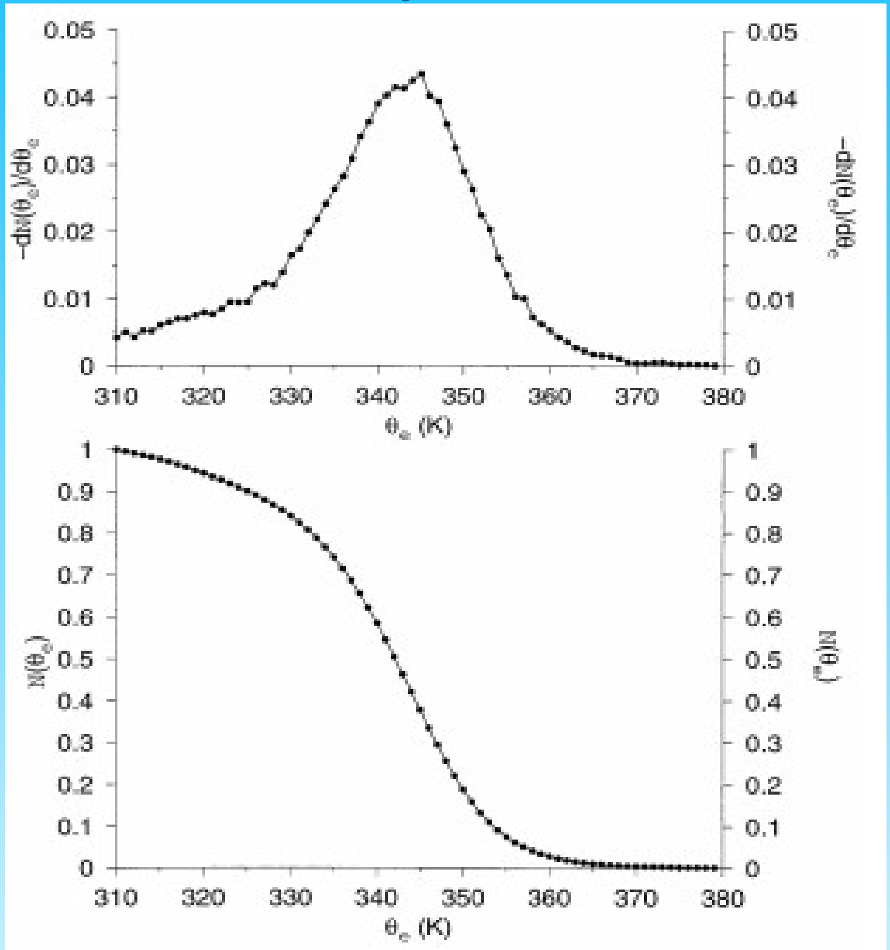
$$\frac{\partial\,\theta_e^*}{\partial\,T}>0$$

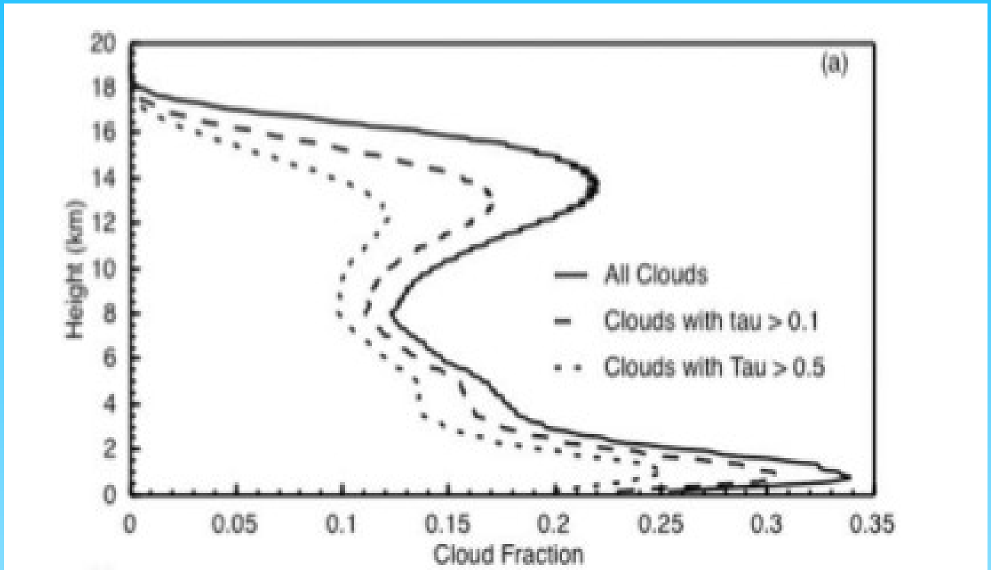
Notice: effect of moisture upon density is neglected.

Typical convective situation in the tropical region



12 to 13 km





Observed cloud fraction (CALIOP lidar) in the tropical region

(Fu et al., GRL 2007)

Probability density and cumulated probability of θ_{ep} in tropical region

0,5% of the clouds reach the tropical tropopause (17,5 km, 100 hPa, T=200K, θ=380K)

TO RETAIN

- Inside an idealized cloud, the ascending motion of a non diluted parcel is described by an adiabatic path (when condensates are transported) or a pseudo-adiabatic when condensates precipitate.
-
- Thermodyamical variables built to describe these transformations generalize the dry air potential temperature.
-
- The instability condition of dry air generalizes to moist **saturated** air by replacing the potential temperature by the equivalent potential temperature.
-
- In moist unsaturated air, the instability depends of the capacity of air to become saturated and acquire positive buoyancy. In the case of parcel motion within undisturbed air, the condition is that a finite perturbation transport the parcel from the ground to the free convective level.

Potential temperatures and usage

- Θ_v : virtual potential temperature, depends on *r, T, p*
Conservation : reversible adiabatic path of moist unsaturated air
Usage : Stability of moist unsaturated air
Application : Mixed boundary layer

- Θ_e : equivalent potential température, depends on *r, T, p* under non saturated conditions ($H<1$) and of r_T, *T, p* under saturated conditions ($H=1$)
Conservation : reversible adiabatic path of moist air, either unsaturated or saturated with condensates transported with the ascending parcel
Usage : Stability of moist saturated air, potentiel instability
Application : Destabilisation of a moist air layer

- Θ^*_e : pseudo-equivalent saturation potential temperature depends of T and p_d
Conservation : non applicable, defined for the ambiant air among convective region
Usage and application: stability of the atmosphere with respect to deep convection, conditional instability

Potential temperatures and usage (continued)

- Θ_l : liquid potential temperature (see TD), depends on *T, p, r_T, r_l* (and r_i if generalized to account for ice)

Conservation : same as Θ_e

Application : temperature and buoyancy of air detrained from the cloud at a given level after evaporation of condensates, instability of the bottom of a stratified cloud layer (TD)

Potential temperatures and mixing (important)

- The potential temperature is the temperature of the parcel at the reference pressure after adiabatic compression. Therefore, the enthalpy in this state is $H = Cp\ \Theta$. As the enthalpy is an additive quantity, the enthalpy of a mixture is the linear combination of the enthalpies of the components. The potential temperature of a mixture is then also the linear combination of the potential temperatures of the components.
- In the moist air Θe is the temperature of the dry air obtained after total water extraction by adiabatic decompression and recompression to the referece pressure. The enthalpy is then $H = Cp\ \Theta e$. For the same reasons as above Θe is an additive quantity.
- The liquid potential temperature Θ_l is also additive
- The pseudo-equivalent saturation potential temperature Θ^*_e is NOT additive

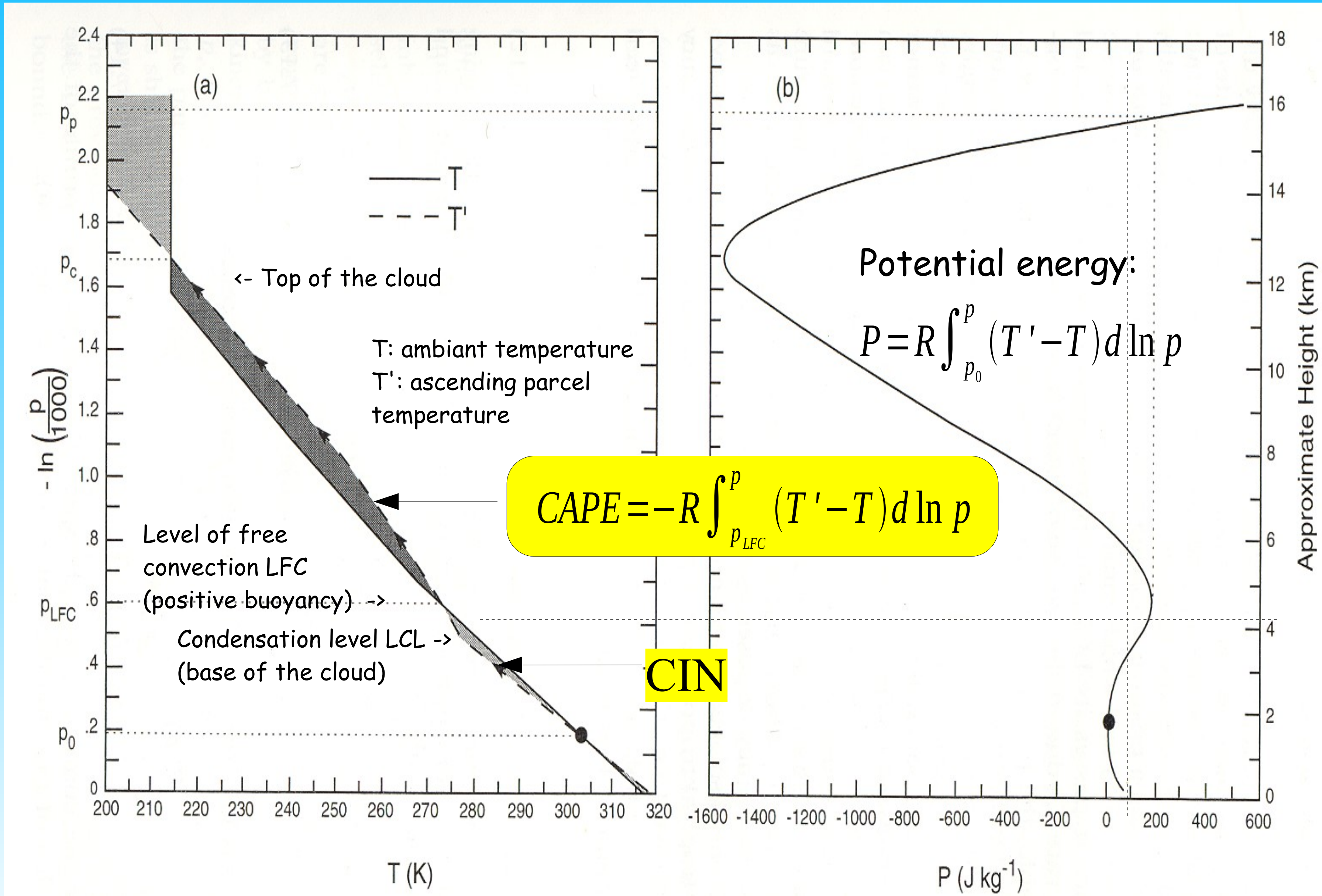
Notice: Conditional instability is associated to the motion of a single parcel which must reach its level of free buoyancy to rise by itself ;
Potential instability is associated to the motion of a layer which becomes unstable after getting lifted and saturated under the effect of a constrain (orography, sea breeze, ...)

Limitations :
In a real cloud, most of the ascending parcels are diluted, entrainment plays an important role. However, especially in big convective systems, a small amount of parcels are undiluted and determine the altitude reached by top cloud and its anvil (controversial)
Supersaturation with respect to ice can be important in the mixed and iced regions of the cloud.

CAPE

Thermodynamic diagram.

History of an ascending air parcel within a cloud



Calculation of the CAPE

The ambient air is in hydrostatic equilibrim $0 = \frac{-1}{\rho_a}\frac{\partial p}{\partial z} - g$

The equation for the vertical motion of the ascending parcel is $\frac{dw}{dt} = \frac{-1}{\rho}\frac{\partial p}{\partial z} - g$

The densities and temperature of the ambient air and the ascending parcel differ but the pressures at the same level are the samesince the pressure of the moving parcel equilibrates very fast as long as its speed is negligible with respect to sound speed.

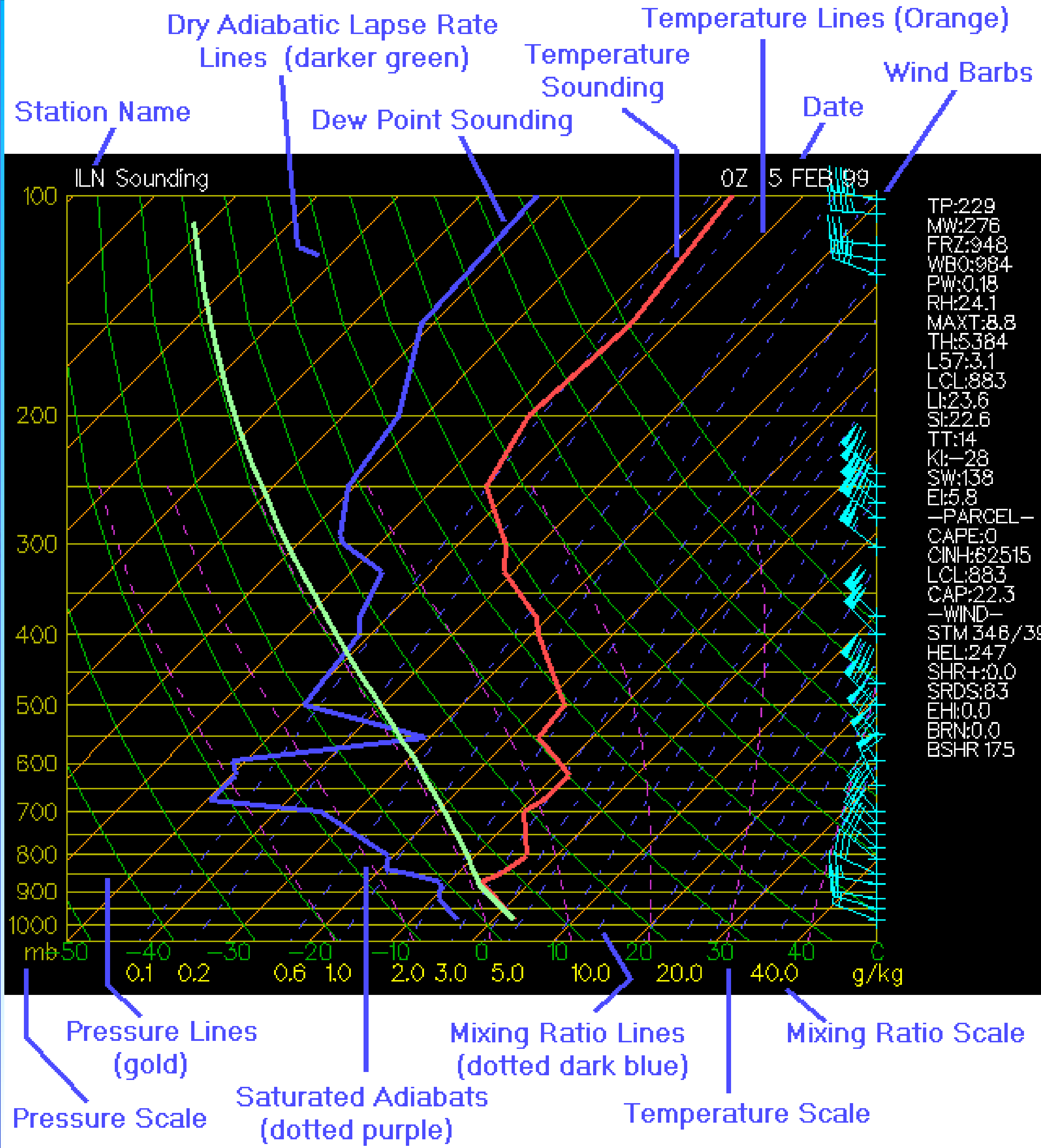
Combining these two equations, and using the perfect gas law, we get

$$\frac{dw}{dt} = \frac{\partial p}{\partial z}\left(\frac{1}{\rho_a} - \frac{1}{\rho}\right) = -\rho_a g\left(\frac{1}{\rho_a} - \frac{1}{\rho}\right) = -g\left(\frac{\rho - \rho_a}{\rho}\right) = g\left(\frac{T - T_a}{T_a}\right)$$

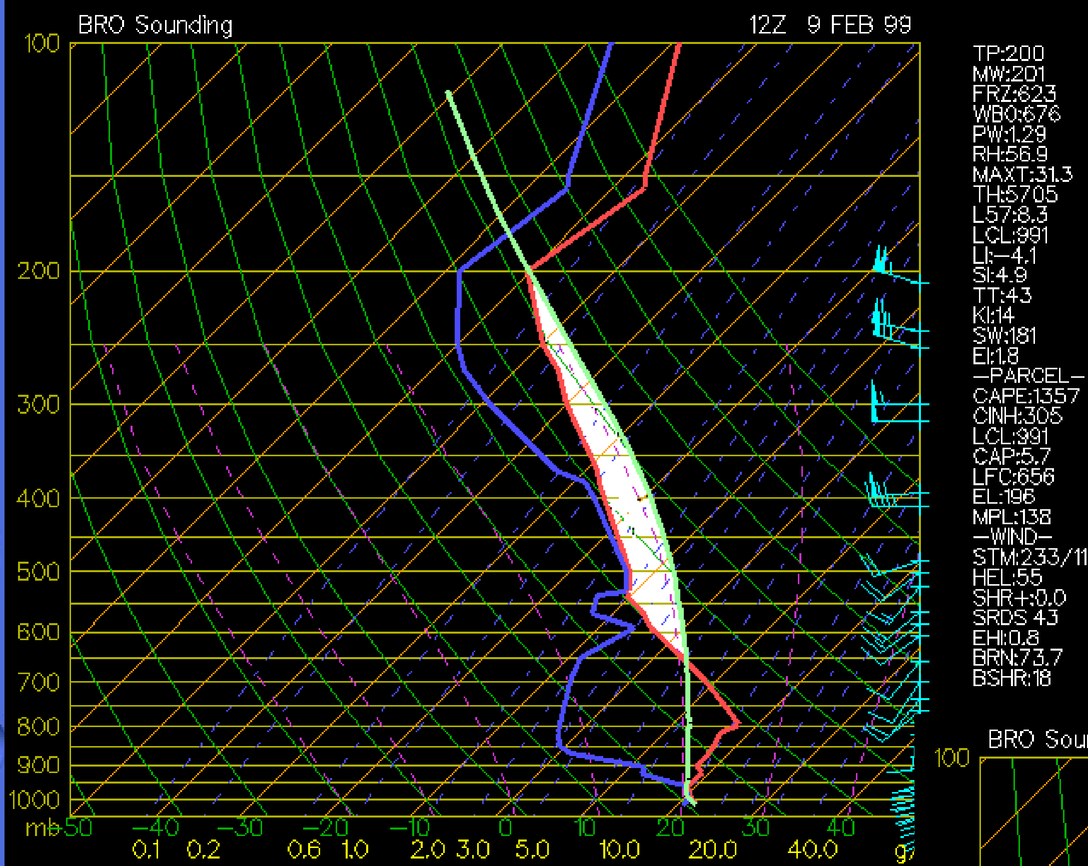
The work performed by the buoyancy between two levels (z_1, p_1) and (z_2, p_2) is therefore

$$W = \int_{z_1}^{z_2} g\left(\frac{T - T_a}{T_a}\right) dz = \int_{p_2}^{p_1}\left(\frac{T - T_a}{\rho_a T_a}\right) dp = R\int_{p_2}^{p_1}\left(\frac{T - T_a}{p}\right) dp = R\int_{p_2}^{p_1}(T - T_a)\, d\, lnp$$

The CAPE is W calculated between the LFC and the LNB.
The CIN is W calculated between the ground and the LFC.

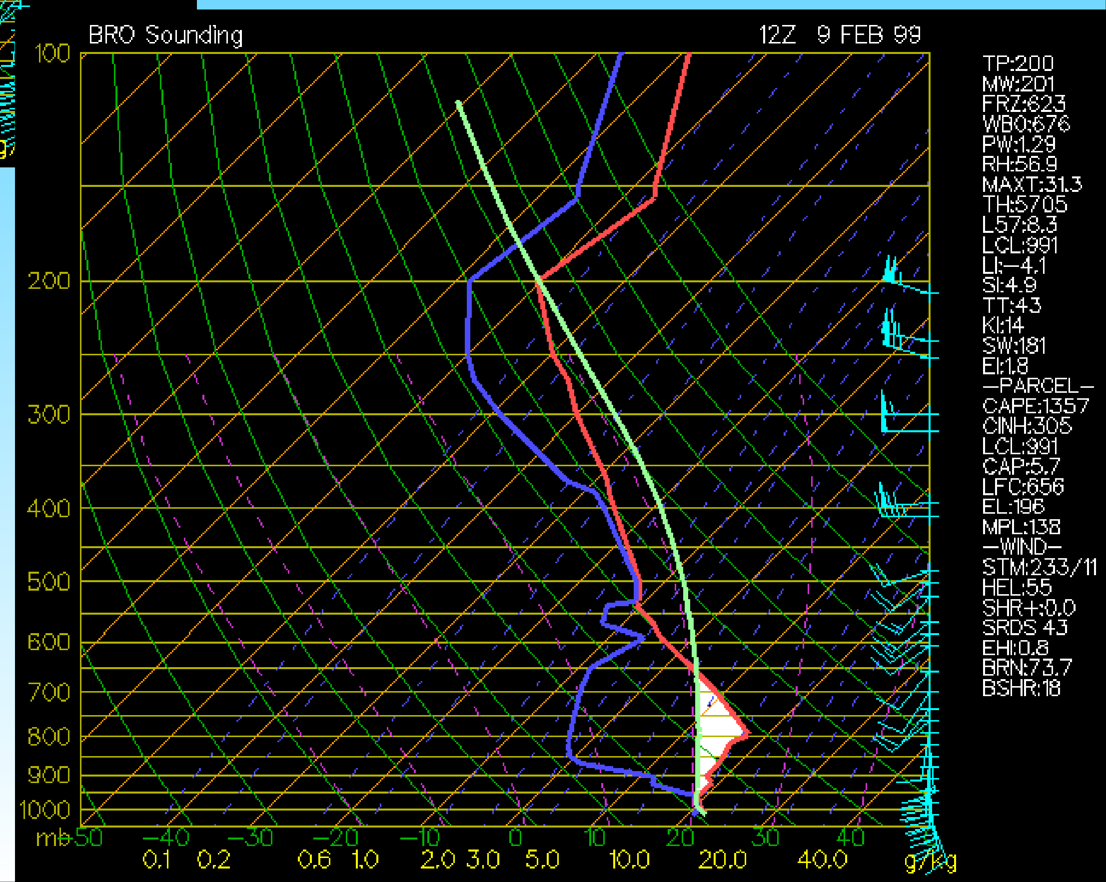


Meteorological diagram to display a sounding


BRO Sounding
12Z 9 FEB 99
TP:200
MW:201
FRZ:623
WB0:676
PW:1.29
RH:56.9
MAXT:31.3
TH:5705
L57:8.3
LCL:991
LI:−4.1
SI:4.9
TT:43
KI:14
SW:181
EI:1.8
—PARCEL—
CAPE:1357
CINH:305
LCL:991
CAP:5.7
LFC:656
EL:196
MPL:138
—WIND—
STM:233/11
HEL:55
SHR+:0.0
SRDS 43
EHI:0.8
BRN:73.7
BSHR:18

CAPE

CIN


BRO Sounding
12Z 9 FEB 99
TP:200
MW:201
FRZ:623
WB0:676
PW:1.29
RH:56.9
MAXT:31.3
TH:5705
L57:8.3
LCL:991
LI:−4.1
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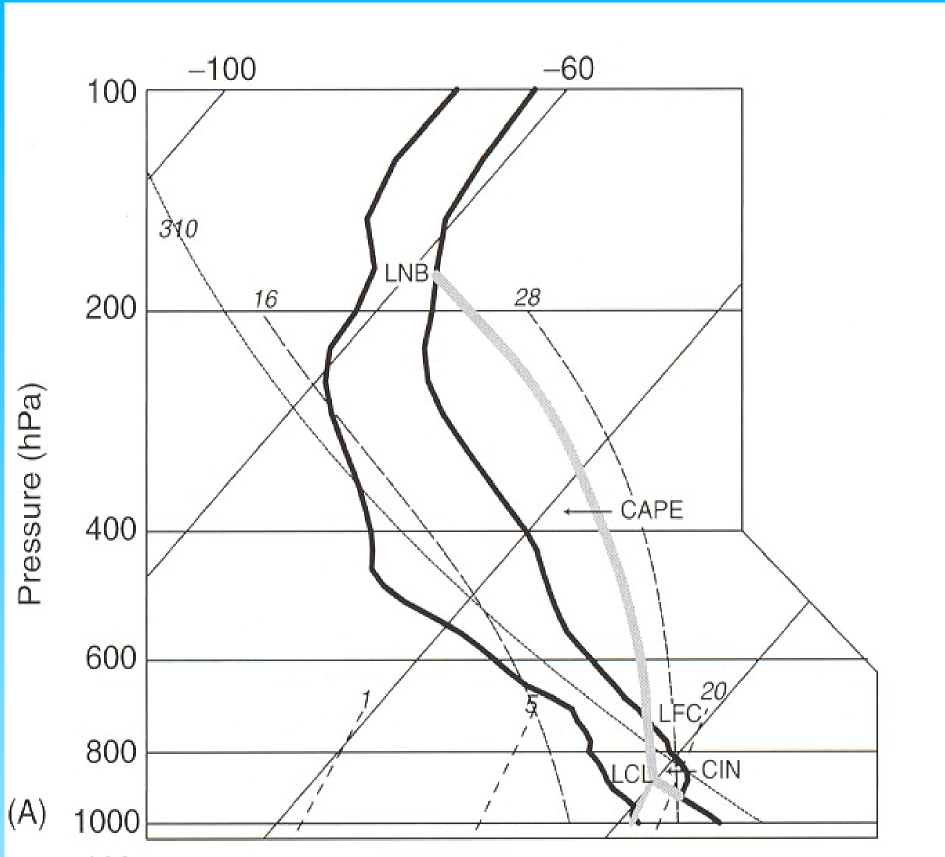
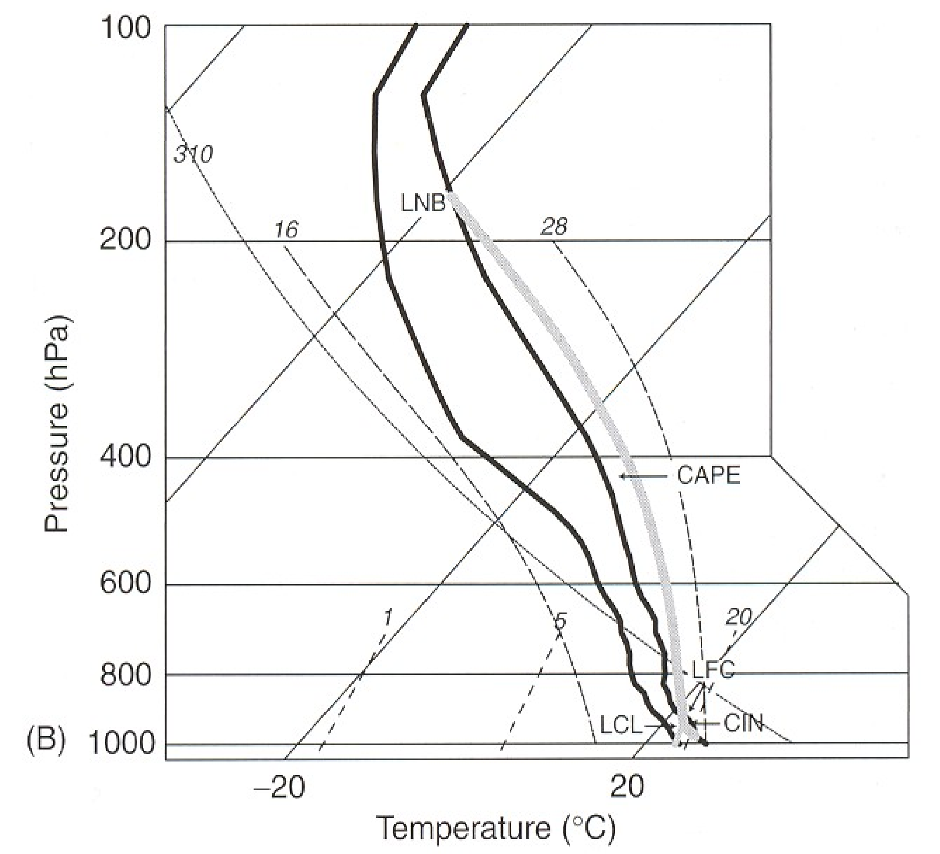
TT:43
KI:14
SW:181
EI:1.8
—PARCEL—
CAPE:1357
CINH:305
LCL:991
CAP:5.7
LFC:656
EL:196
MPL:138
—WIND—
STM:233/11
HEL:55
SHR+:0.0
SRDS 43
EHI:0.8
BRN:73.7
BSHR:18

CAPE and meteorological diagram



Subtropical summer conditions: Dry profile and moist boundary layer
Large CAPE (3000-4000 J kg^{-1}) but large CIN inhibiting the onset.

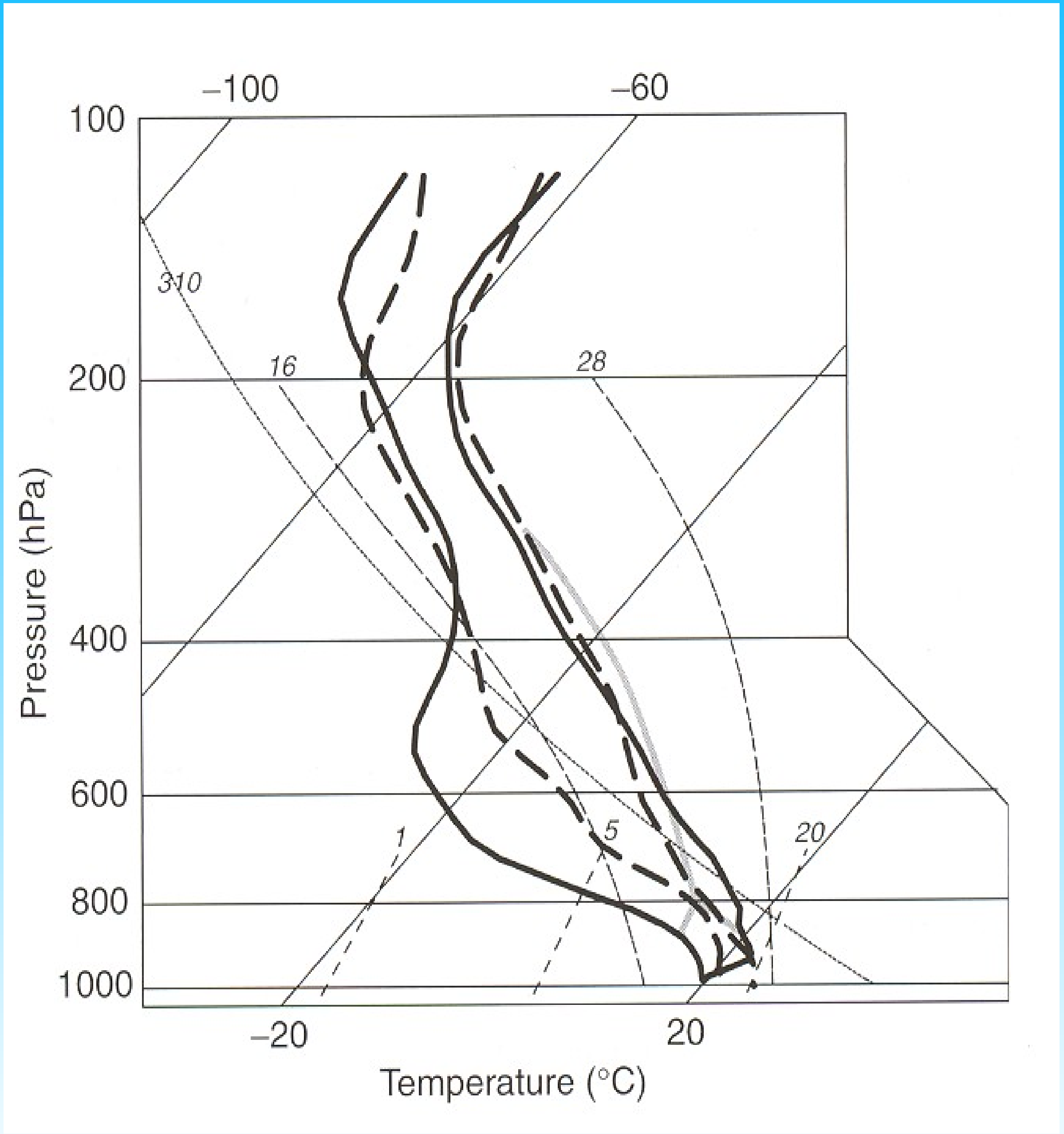
Tropical oceanic region close to moist adiabatic conditions : Moderated CAPE (1000-2000 J kg^{-1}) but small CIN favouring the onset



Example of evolution towards an unstable situation (potential instability) during the day

Full: beginning
Dash: end

Saturation and destabilisation of the layer 800-870 hPa

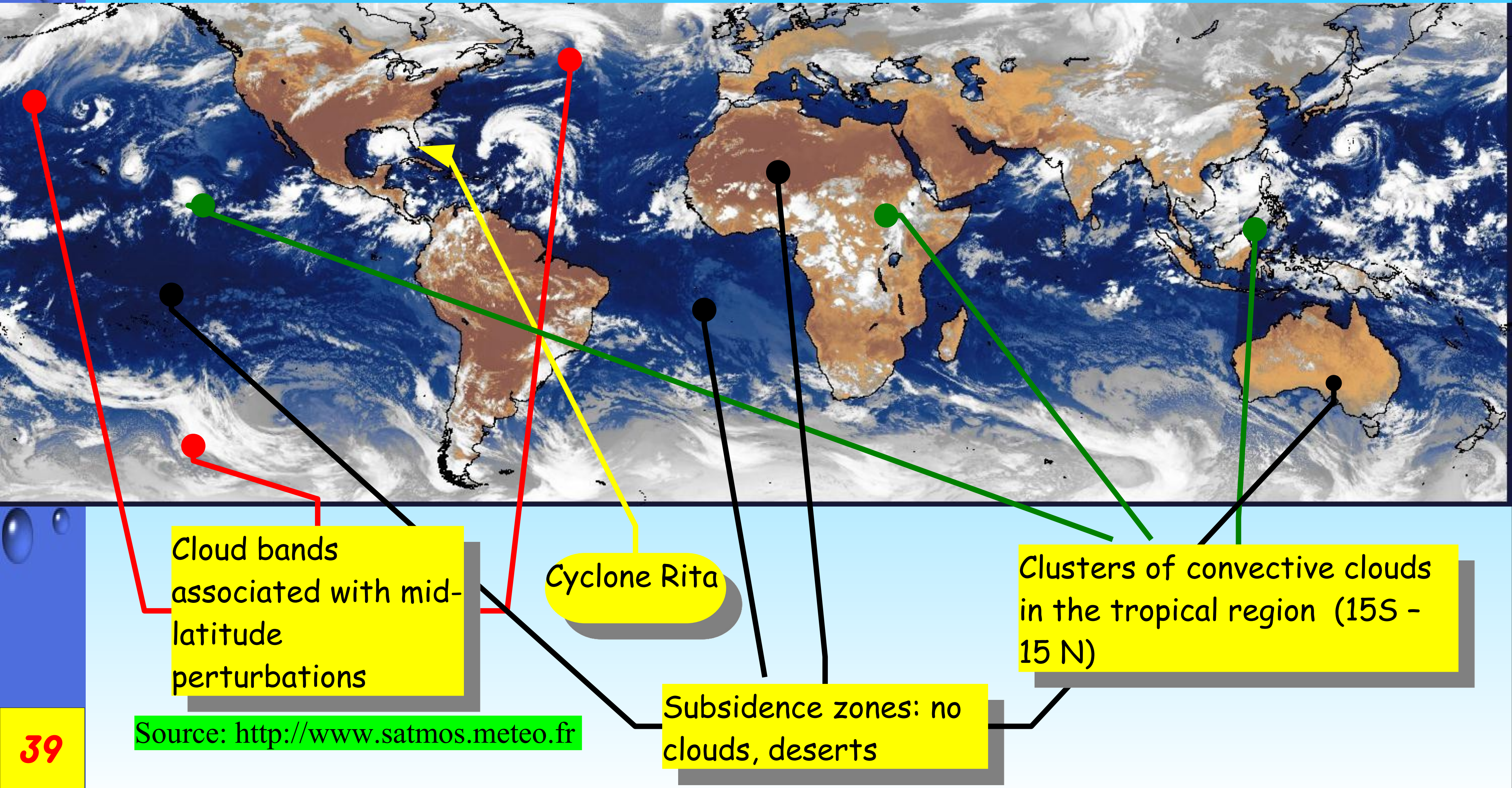


Cloud distribution
(mostly tropical)

Large-scale organisation of clouds

IR false color composite image, obtained par combined data from 5 Geostationary satellites 22/09/2005 18:00TU

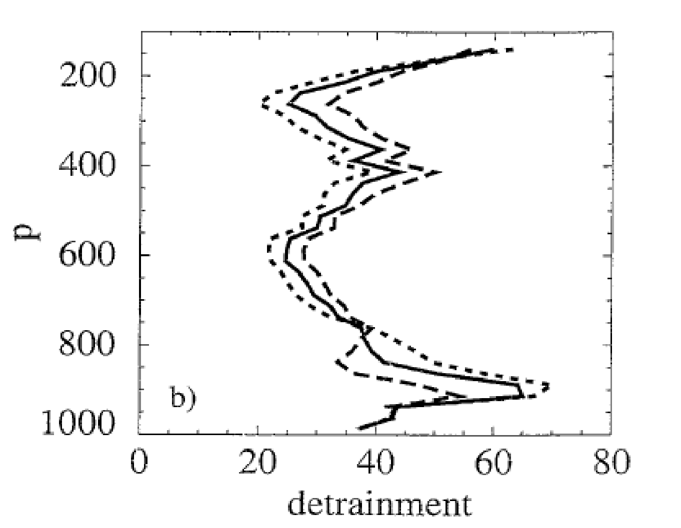
(GOES-10 (135O), GOES-12 (75O), METEOSAT-7 (OE), METEOSAT-5 (63E), MTSAT (140E))



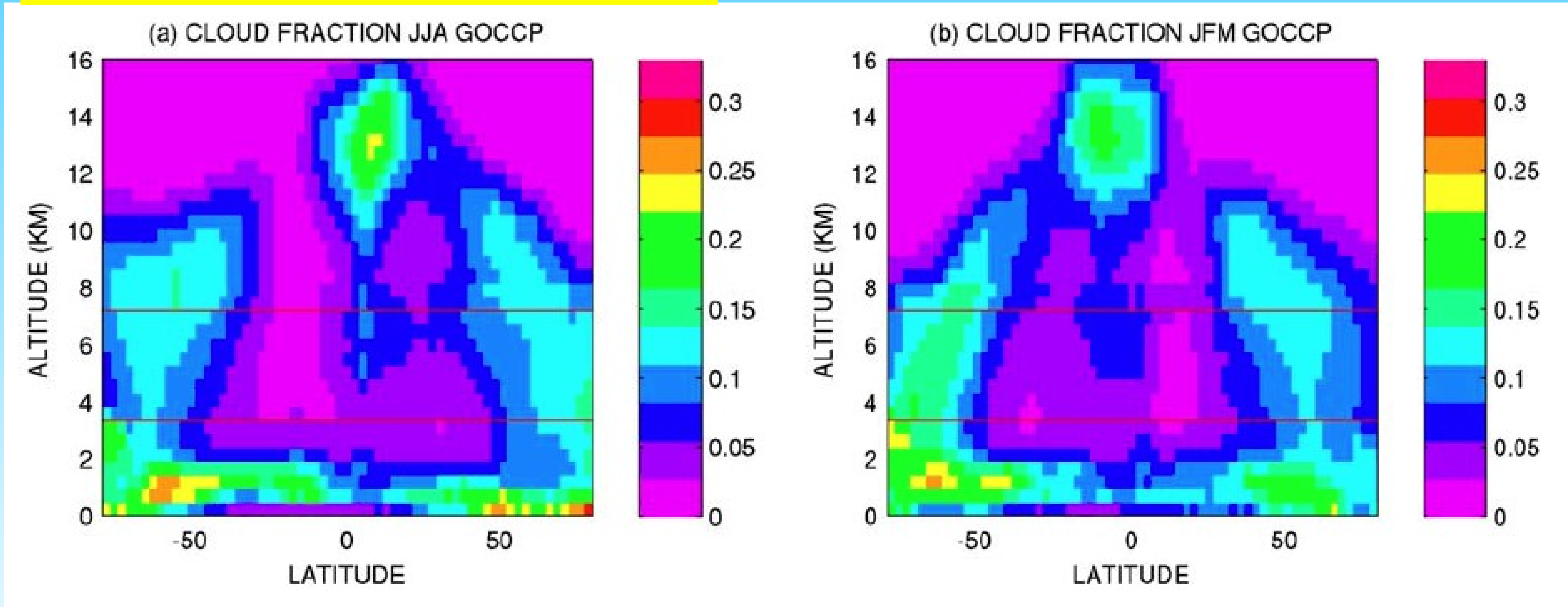
Source: http://www.satmos.meteo.fr

Observed distribution of clouds

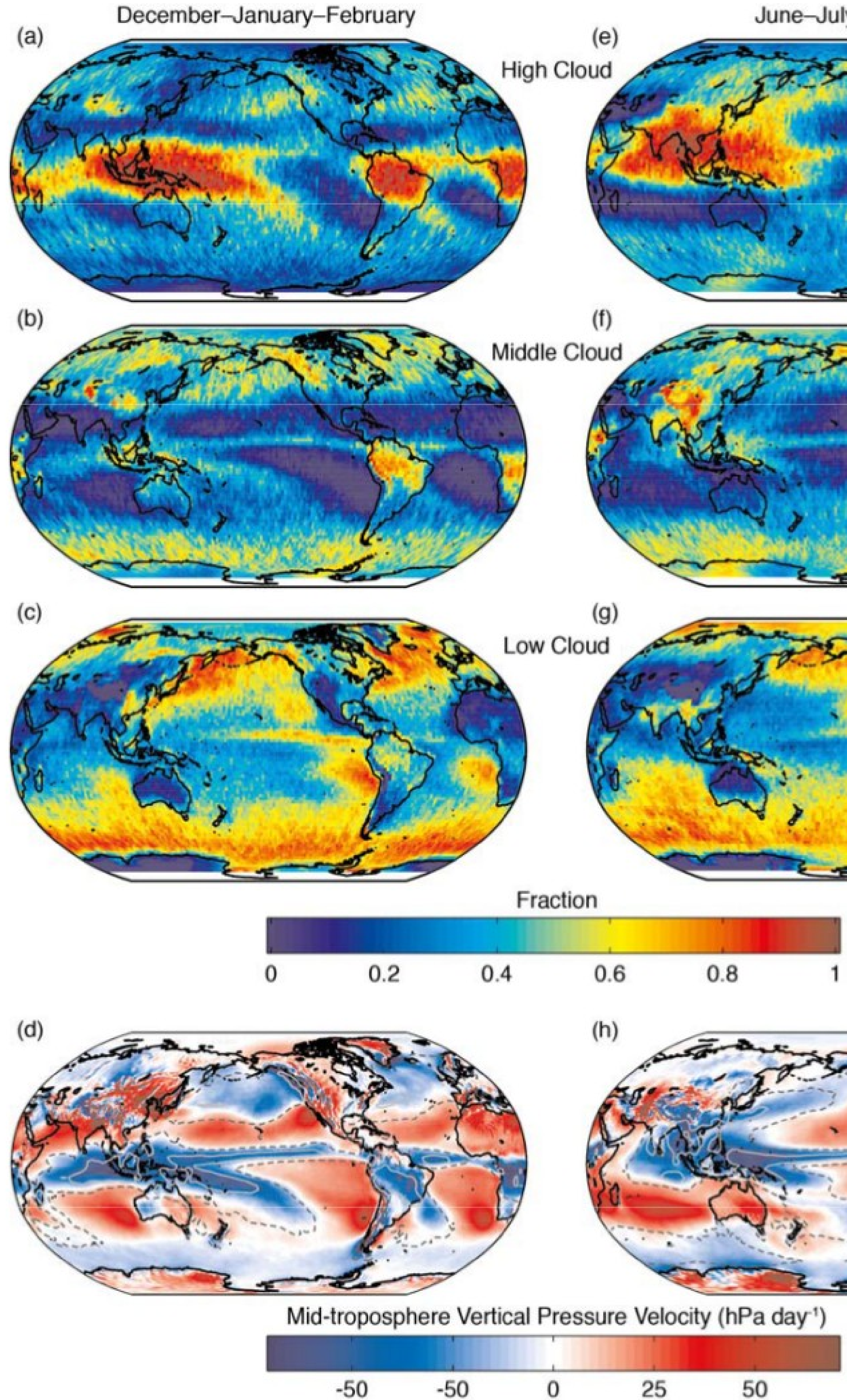
Detrainement of clouds measured during TOGA-COARE (1992-1993).
Johnson et al., J. Climate, 1999



CALIPSO lidar (launched 2006)



Chepfer et al., JGR 2010

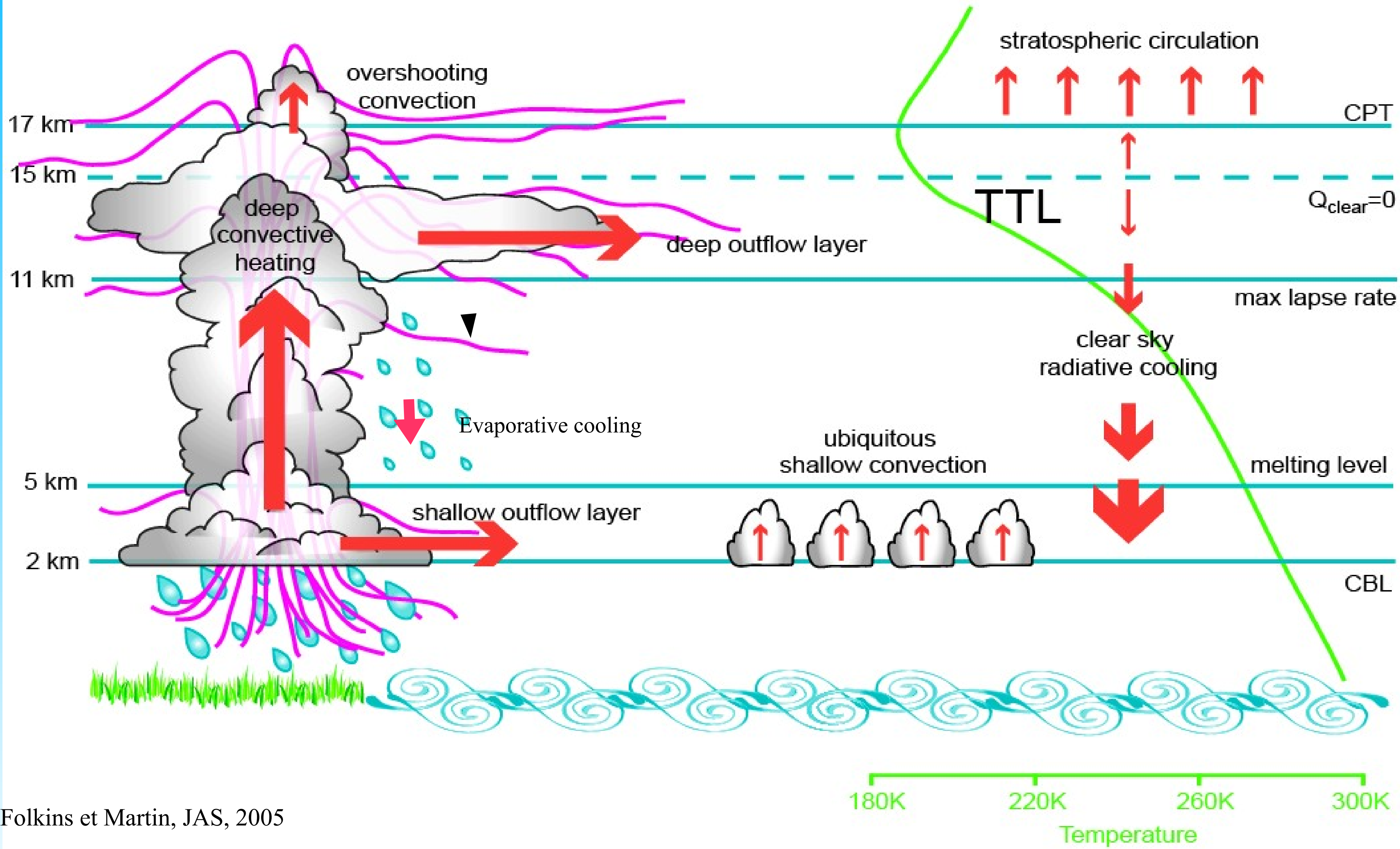


Cloud cover from space lidar and radar (CALIOP + CLOUDSAT)

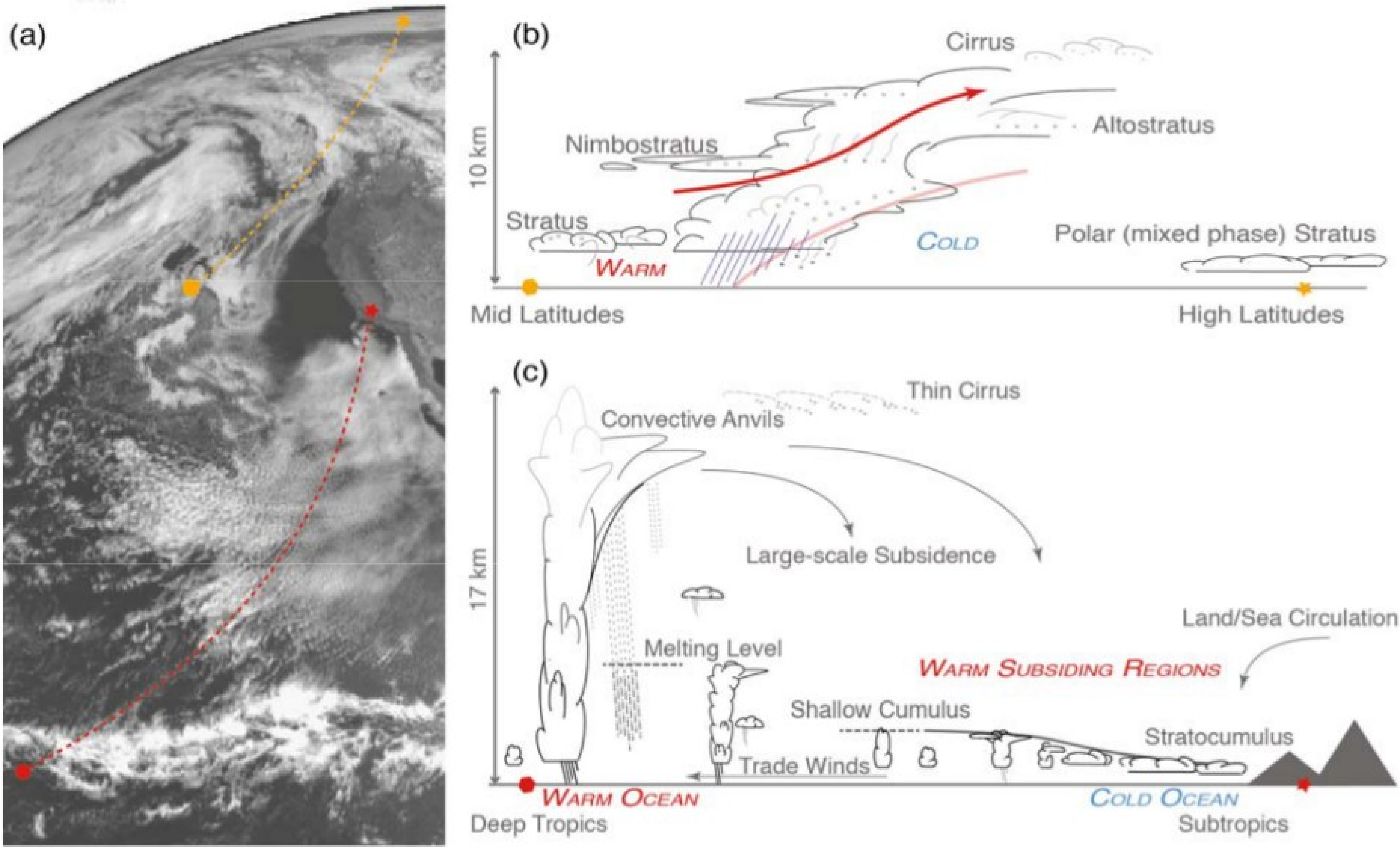
Vertical velocity (ERA-I)

ICCP 2013 report

Tropical Tropopause Layer and Deep Convection



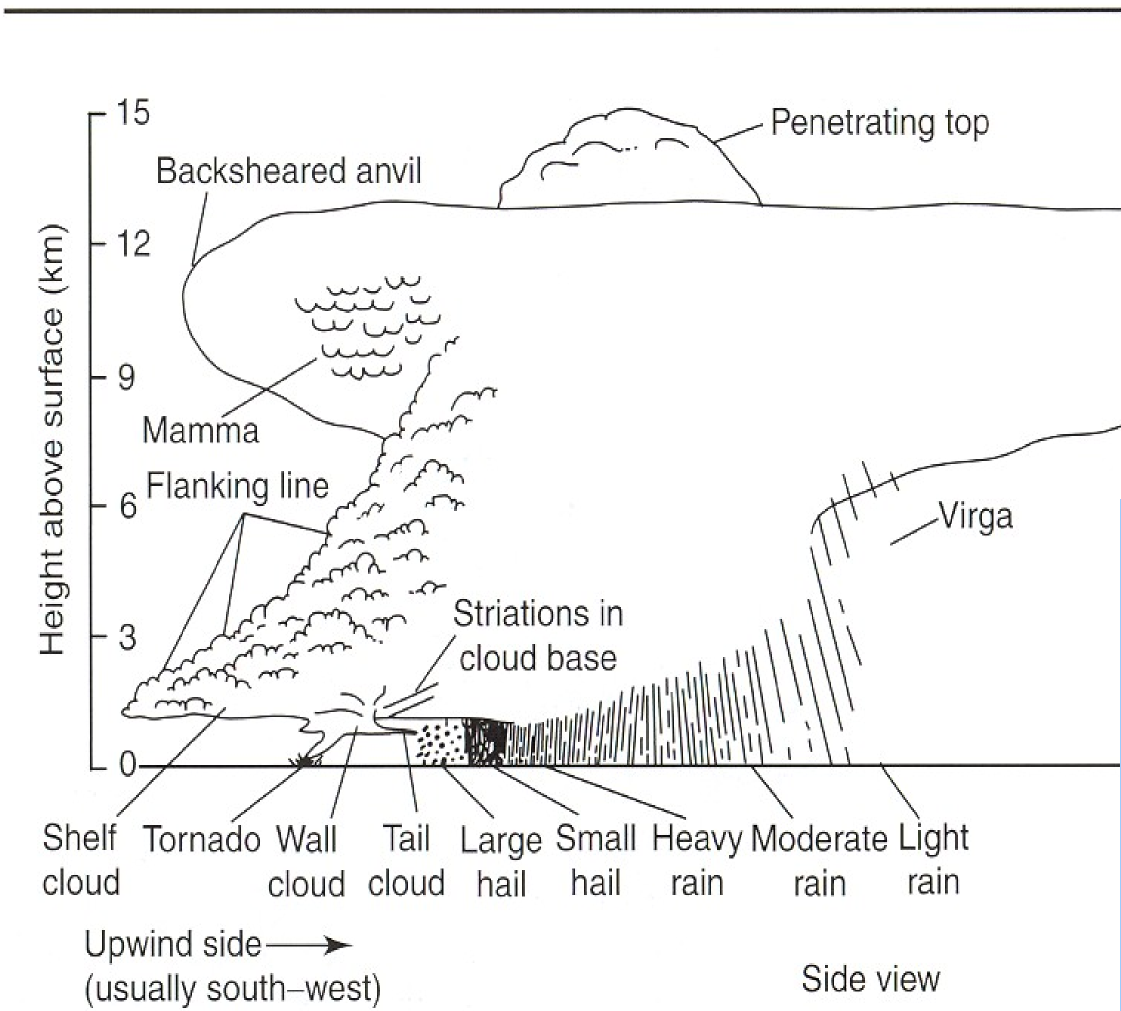
Folkins et Martin, JAS, 2005

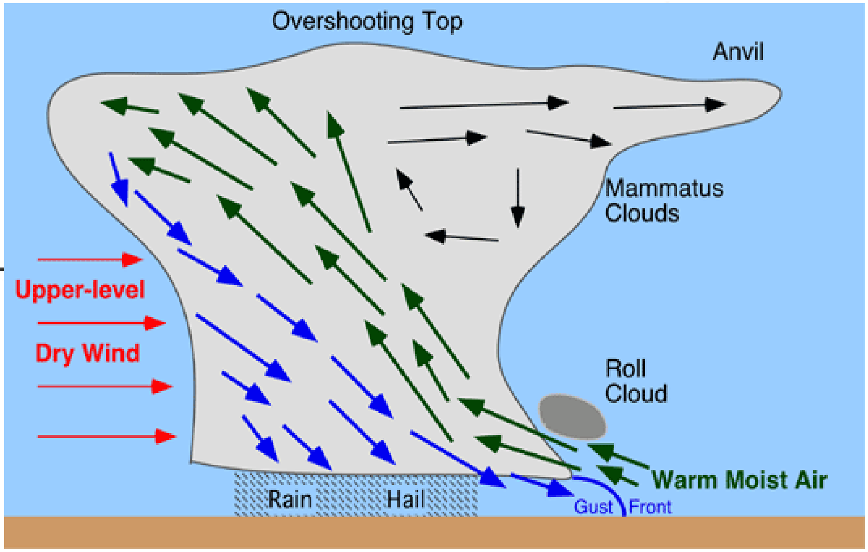


(a)
(b)
Cirrus
Altostratus
Nimbostratus
10 km
Stratus
Warm
Cold
Polar (mixed phase) Stratus
Mid Latitudes
High Latitudes
(c)
Thin Cirrus
Convective Anvils
Large-scale Subsidence
17 km
Land/Sea Circulation
Melting Level
Warm Subsiding Regions
Shallow Cumulus
Stratocumulus
Trade Winds
Warm Ocean
Cold Ocean
Deep Tropics
Subtropics

Onset, propagation and organisation of convection (from local instability to mesoscale organisation)

Structure of a convective supercell (generating tornadoes and hailstones)



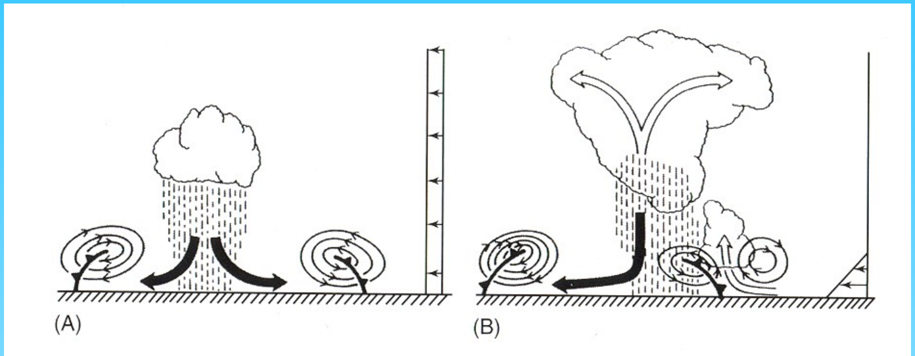


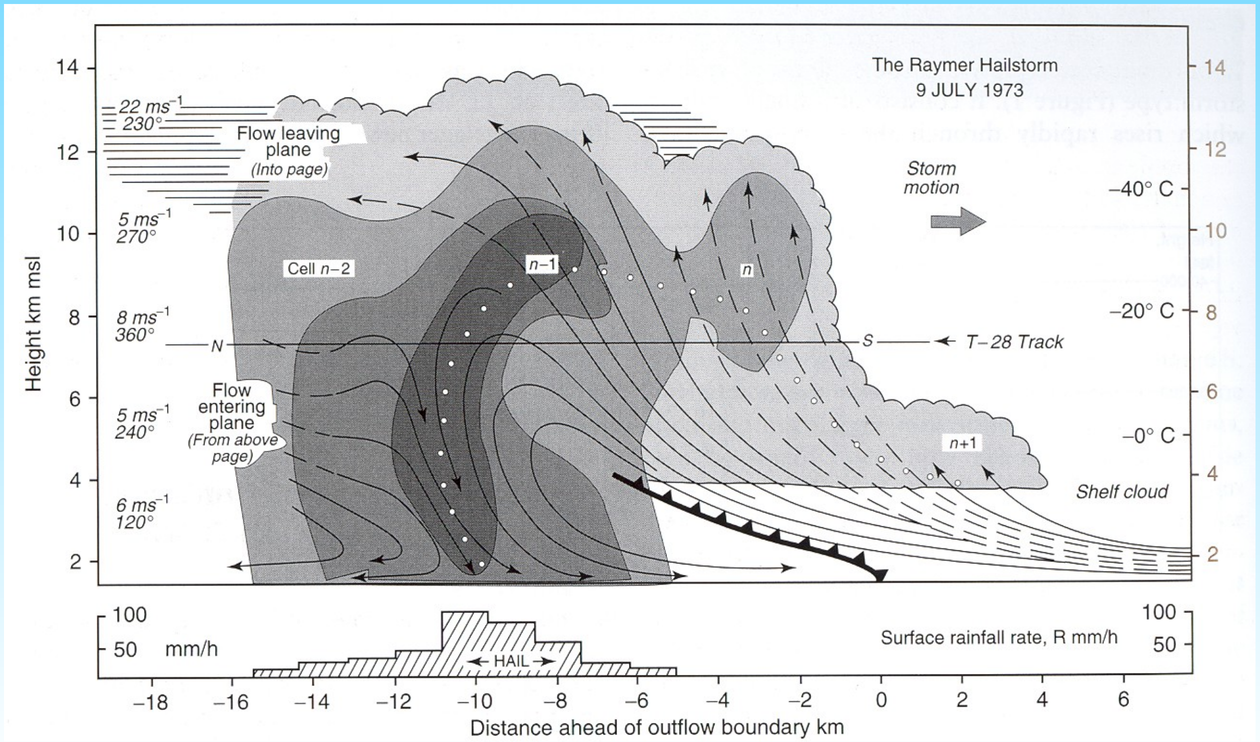
Fronts de cumulonumbus

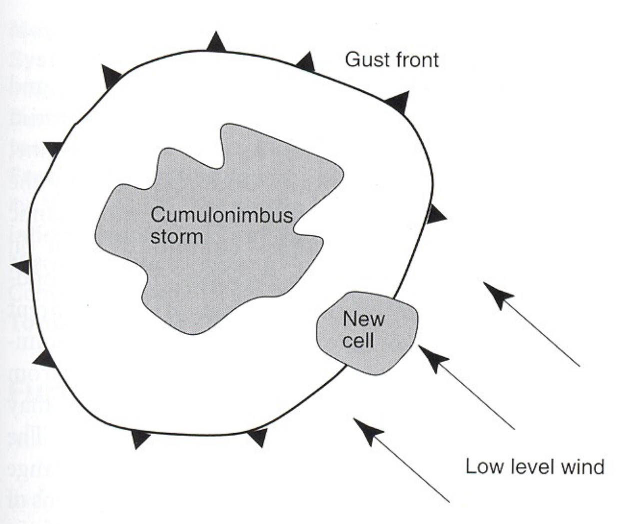
Jospin Tornado
2011, Missouri
150 victims

CAPE does matter

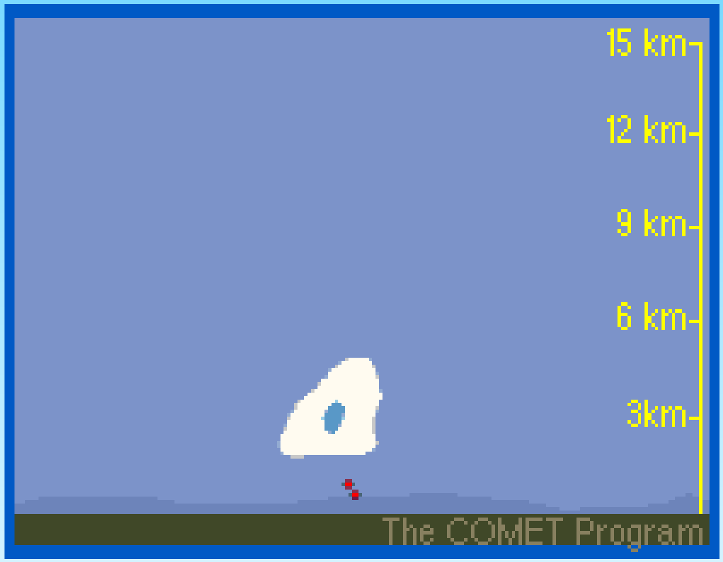
Propagation / regeneration of a convective cell under the presence of a vertical wind shear combined with gravity currents.

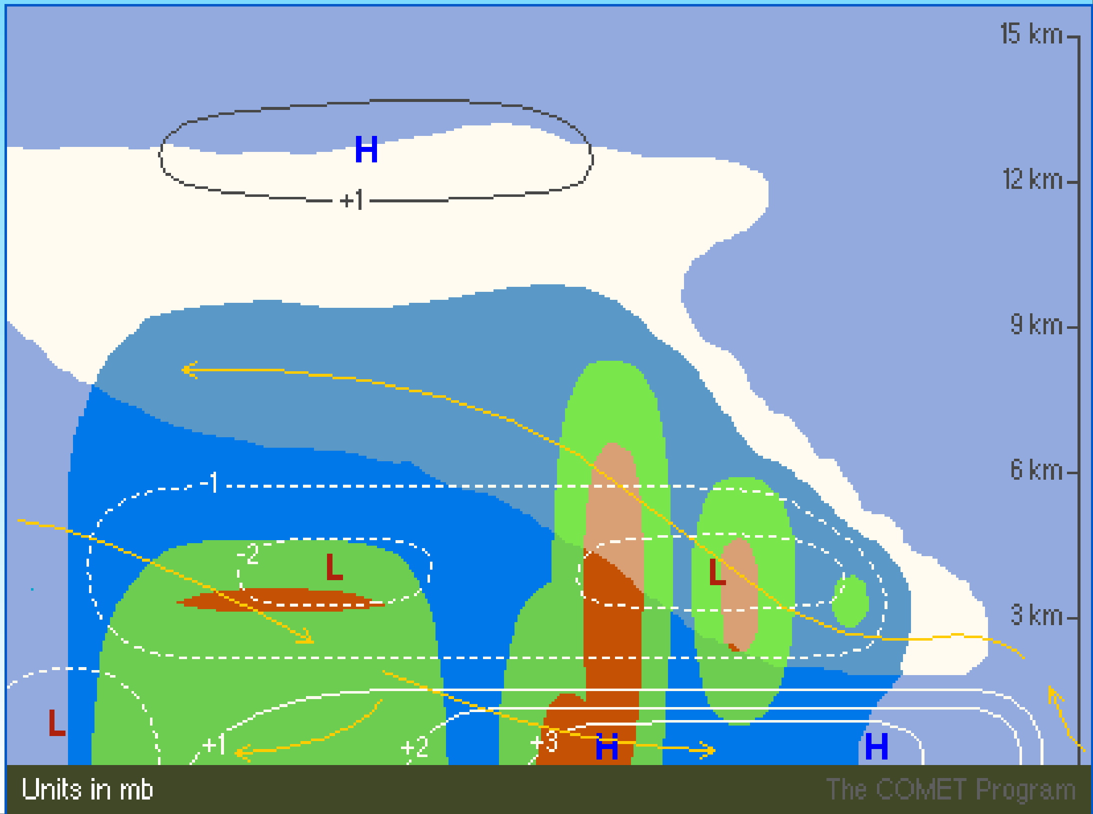


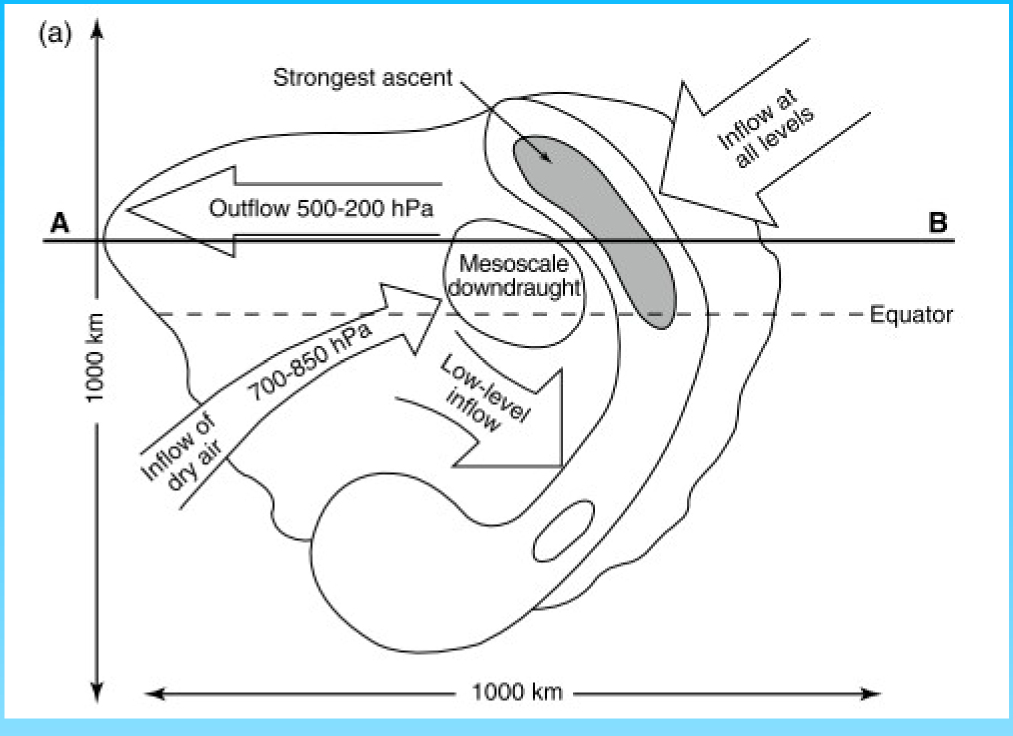





Shear Vector


15 km
12 km
9 km
6 km
3km
The COMET Program


15 km
12 km
9 km
6 km
3 km
H
+1
-1
-2
L
L
L
+1
+2
+3
H
H
Units in mb
The COMET Program



Conceptual model of a large tropical mesoscale system

Front low level inflow gets into the convective ascent and generates high altitude outflow. Back mid-level inflow gets cooled by evaporation of stratiform precipitations and generates low level outflow.

Houze, 2004

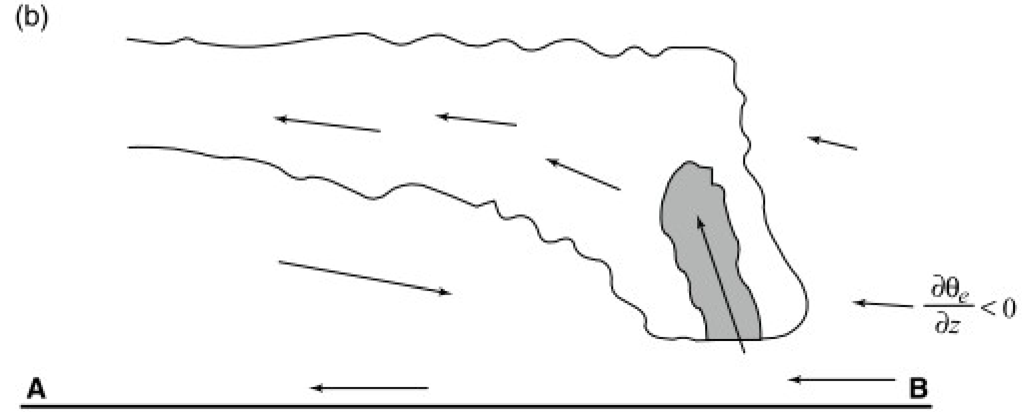


Figure 10. Conceptual model of a supercluster, which is a large mesoscale convective system of the type that occurs over the western tropical Pacific. (a) Plan view and (b) zonal vertical cross section along line AB. Note the depth of the inflow layer at B. From *Moncrieff and Klinker* [1997].

2D scheme of a mesoscale system according to Moncrieff. Stratification (θ_e) is preserved by the meso-scale motion, i.e.

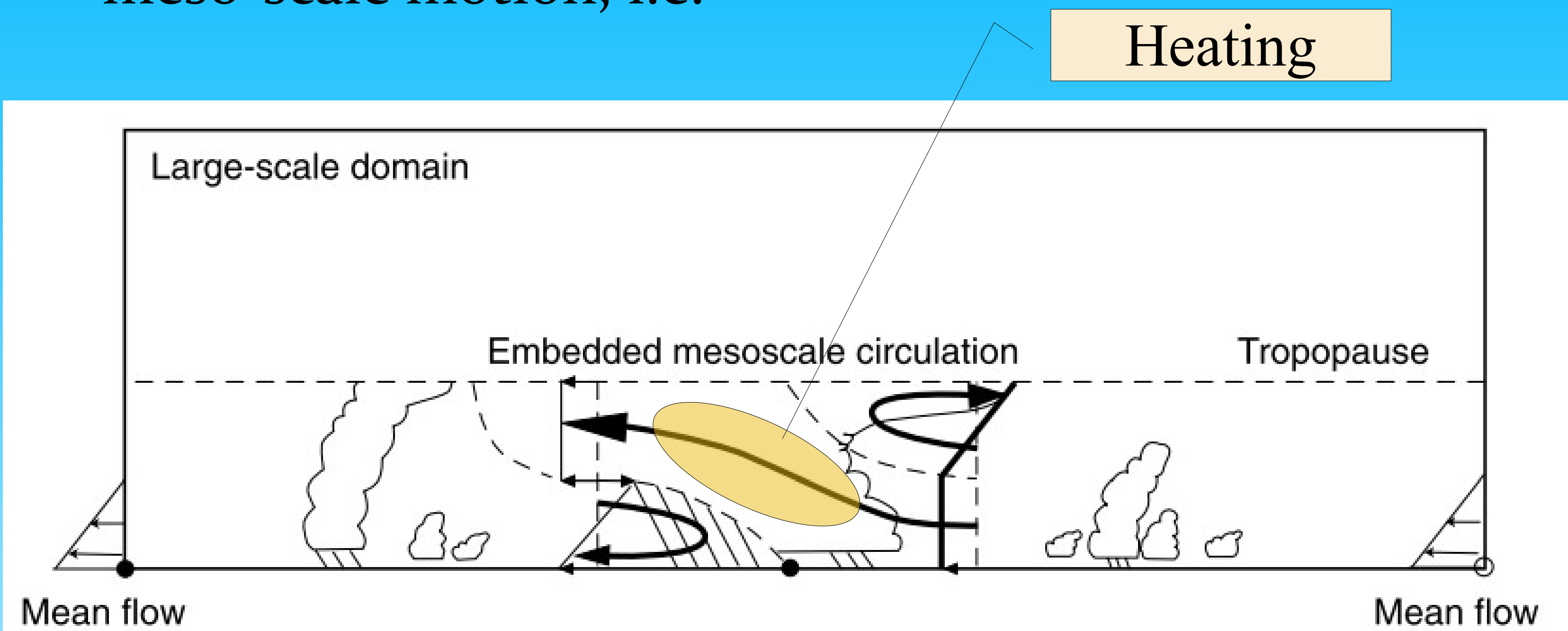
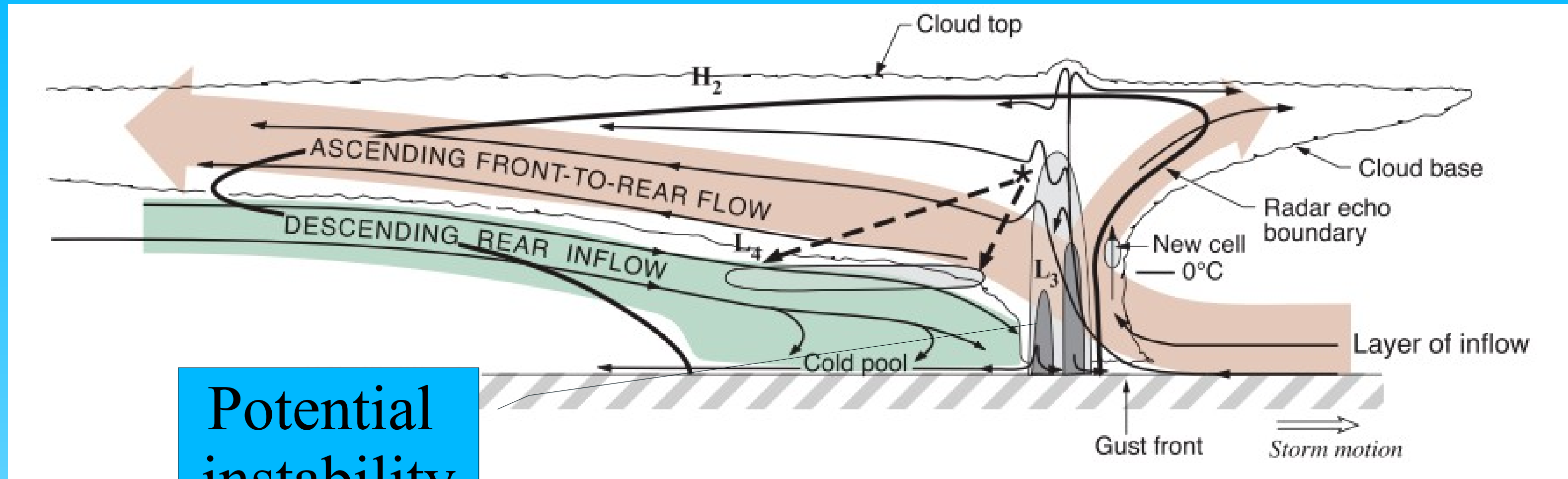


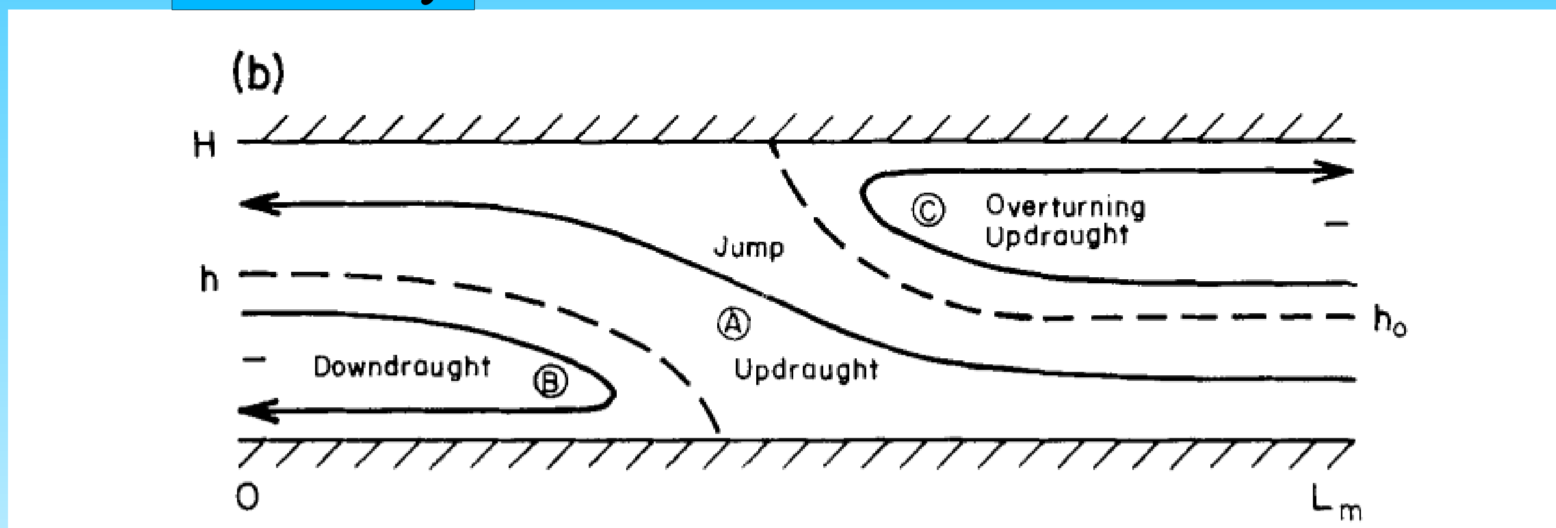
Figure 6. Schematic diagram showing the airflow relative to a two-dimensional, steady state mesoscale convective system in a large-scale environment of given wind shear. The environmental air entering the updraft is potentially unstable, and there is a pressure decrease across the system from right to left at middle levels. The streamlines are those required by conservation of mass, momentum, entropy, and vorticity. Adapted from *Moncrieff* [1992].

Houze, 2004

One can show that this structure develops as a stationary gravity wave in response to a tropospheric heating within a shear flow. Q : how does it develop ? One possibility is symmetric instability see later in this course.

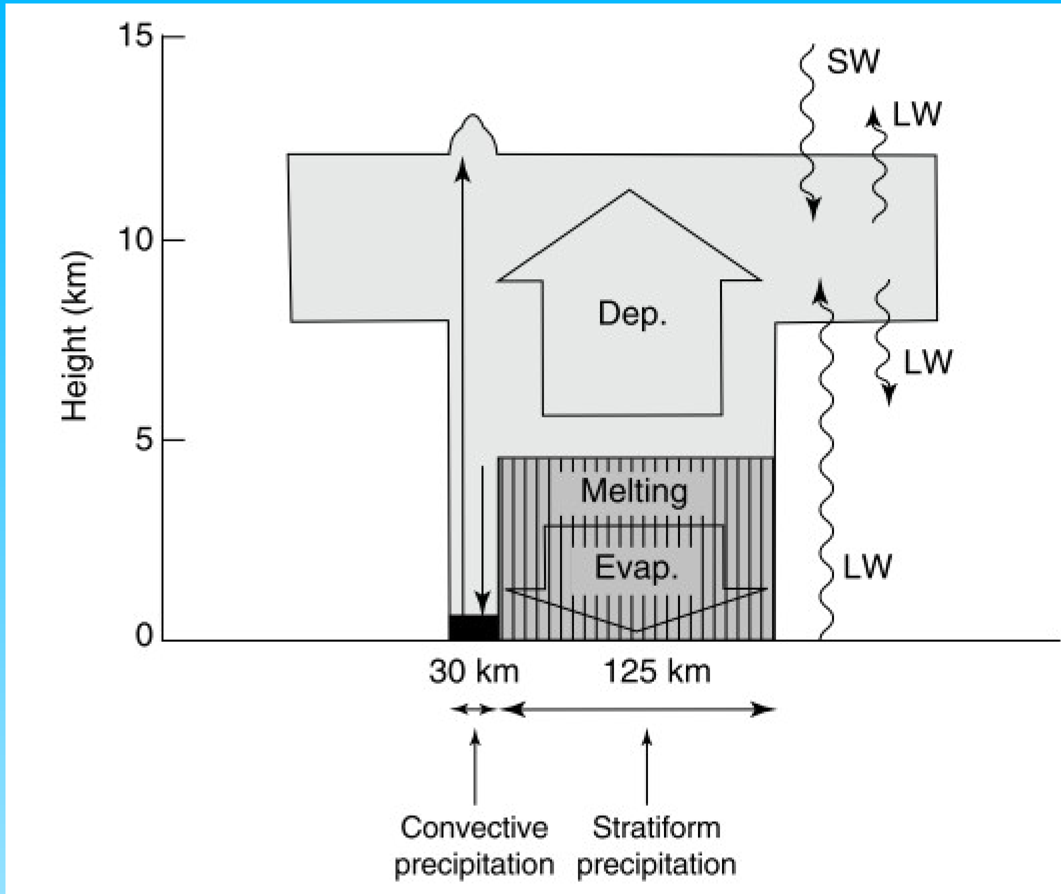


Potential instability



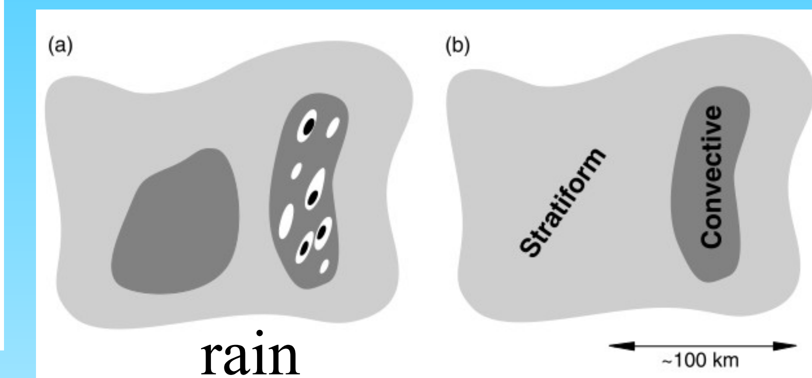
Moncrieff 1992
Houze 2018

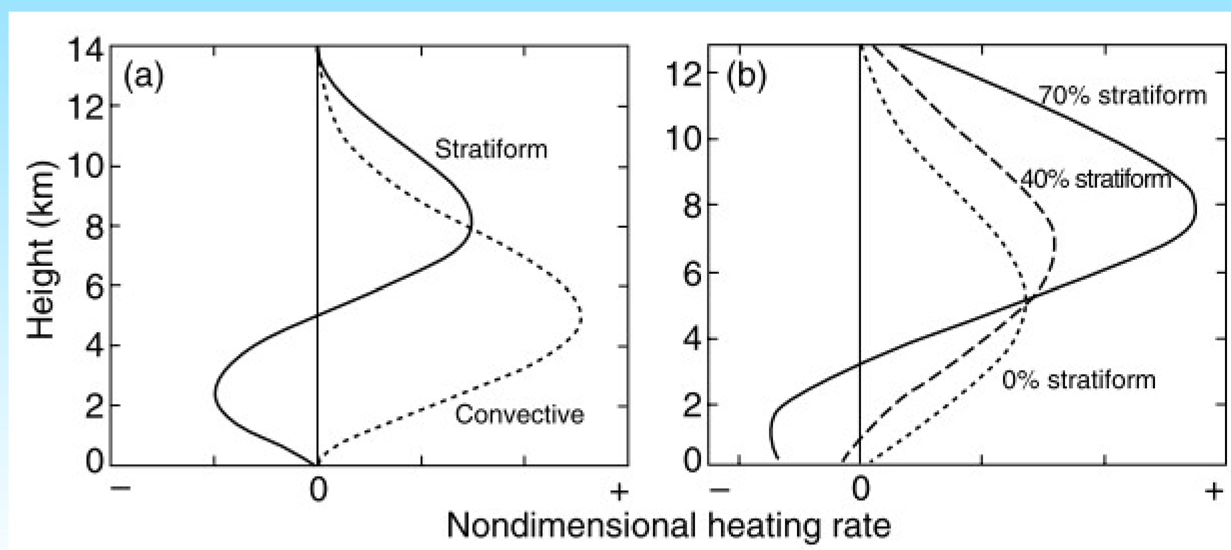
Stratified circulation and instability are combined : schematic structure of a squall line where instability induce mixing and reduce stratification at the front while a stratified circulation is established in the back.



Houze, 2004

Conceptuel scheme of a meso-scale system. Large contribution of the stratiform region.
The vertical shape of induced heating (condensation) du depends on the stratiform part.





THE QUESTION OF DILUTION

Riehl and Malkus, 1958 : The most intense convective towers are undiluted

Romps and Kuang, 2010 : The most intense convective towers are largely diluted

Dauhut et al., 2017 : dilution of 0.5 in the most intense phase

Hector the Convector
https://www.youtube.com/watch?v=xjPumywGaAU

https://www.youtube.com/watch?v=02Josm7WWb8

ENTRAINMENT FROM THE SIMULATION OF HECTOR

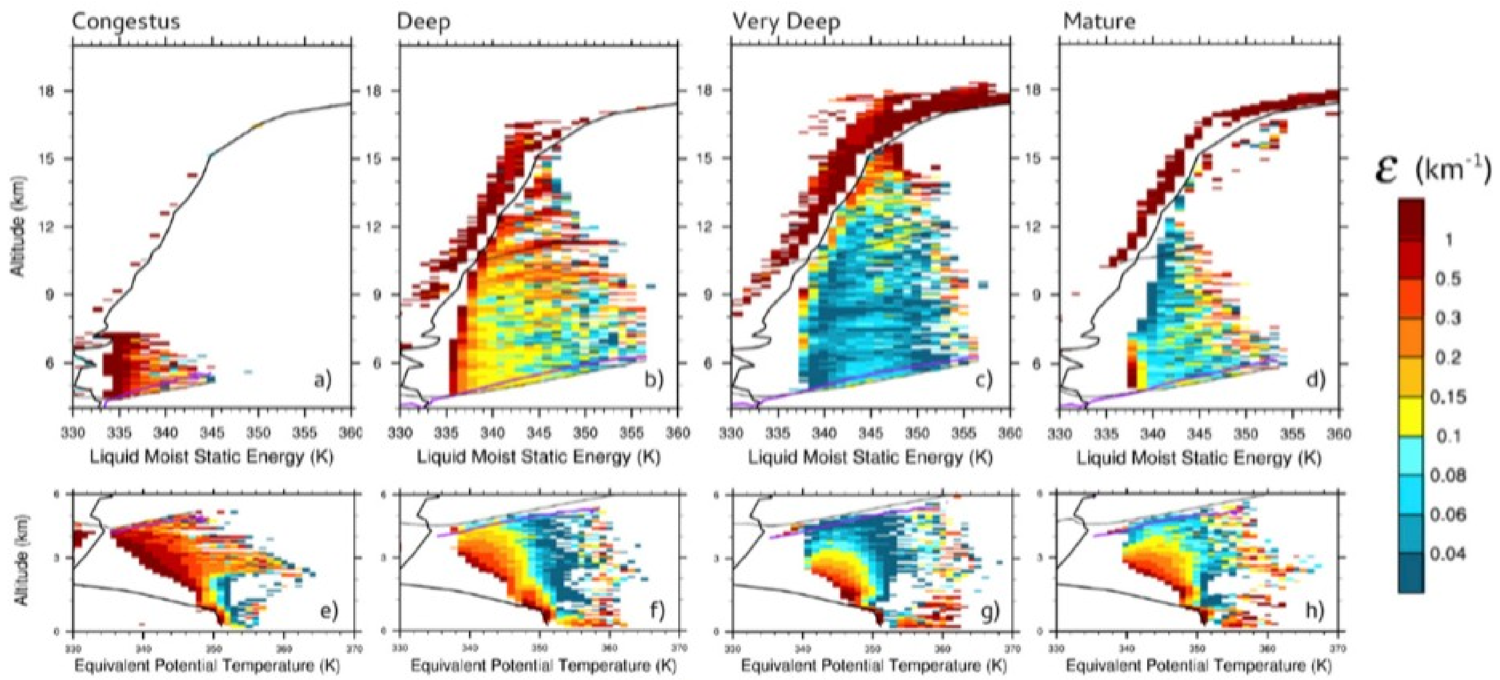


FIG. 9. The entrainment rate for the same time periods as in Fig. 4: during the (a),(e) congestus convection phase, (b),(f) deep convection phase, (c),(g) very deep convection phase, and (d),(h) mature convection phase. (top) Values above the freezing level computed with ξ_l as the isentropic coordinate; above the purple line, the error in ε due to liquid precipitation is less than 0.02 km^{-1}. (bottom) Values below the freezing level computed with θ_e as the isentropic coordinate; below the purple line, the error in ε due to ice processes is less than 0.02 km^{-1}. The black lines are as in Fig. 2. The wide white areas without entrainment values correspond to bins where $\tilde{w}$ or ε is negative.

$$\varepsilon(\phi^{\mathrm{env}} - \phi) = \frac{\tilde{\dot{\phi}}}{\tilde{w}} = \frac{\langle \rho \dot{\phi} \rangle}{\langle \rho w \rangle}.$$

Dauhut et al, 2017, JAS

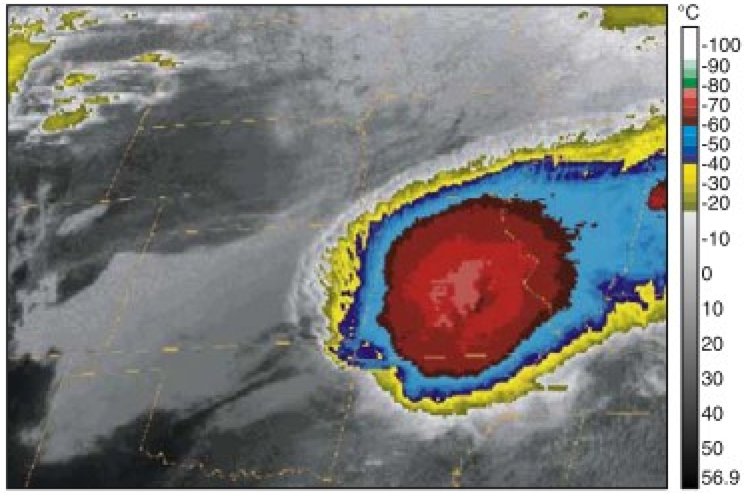
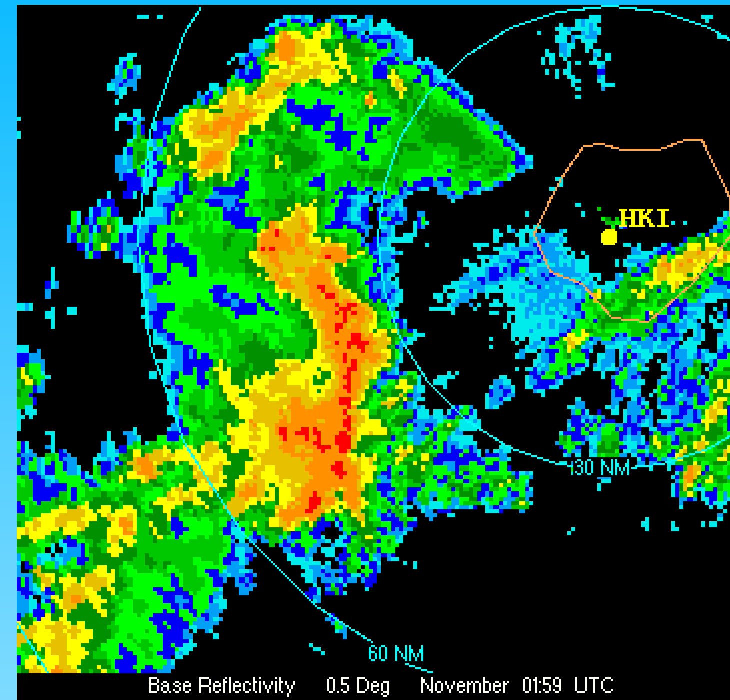
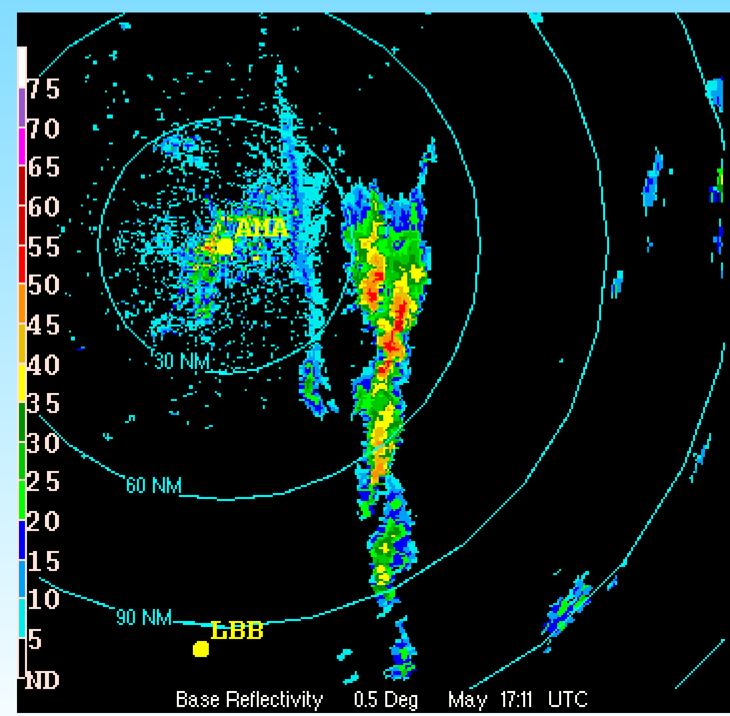


Figure 1. Infrared satellite image of a mesoscale convective system over Missouri. Courtesy of J. Moore, St. Louis University, St. Louis, Missouri.

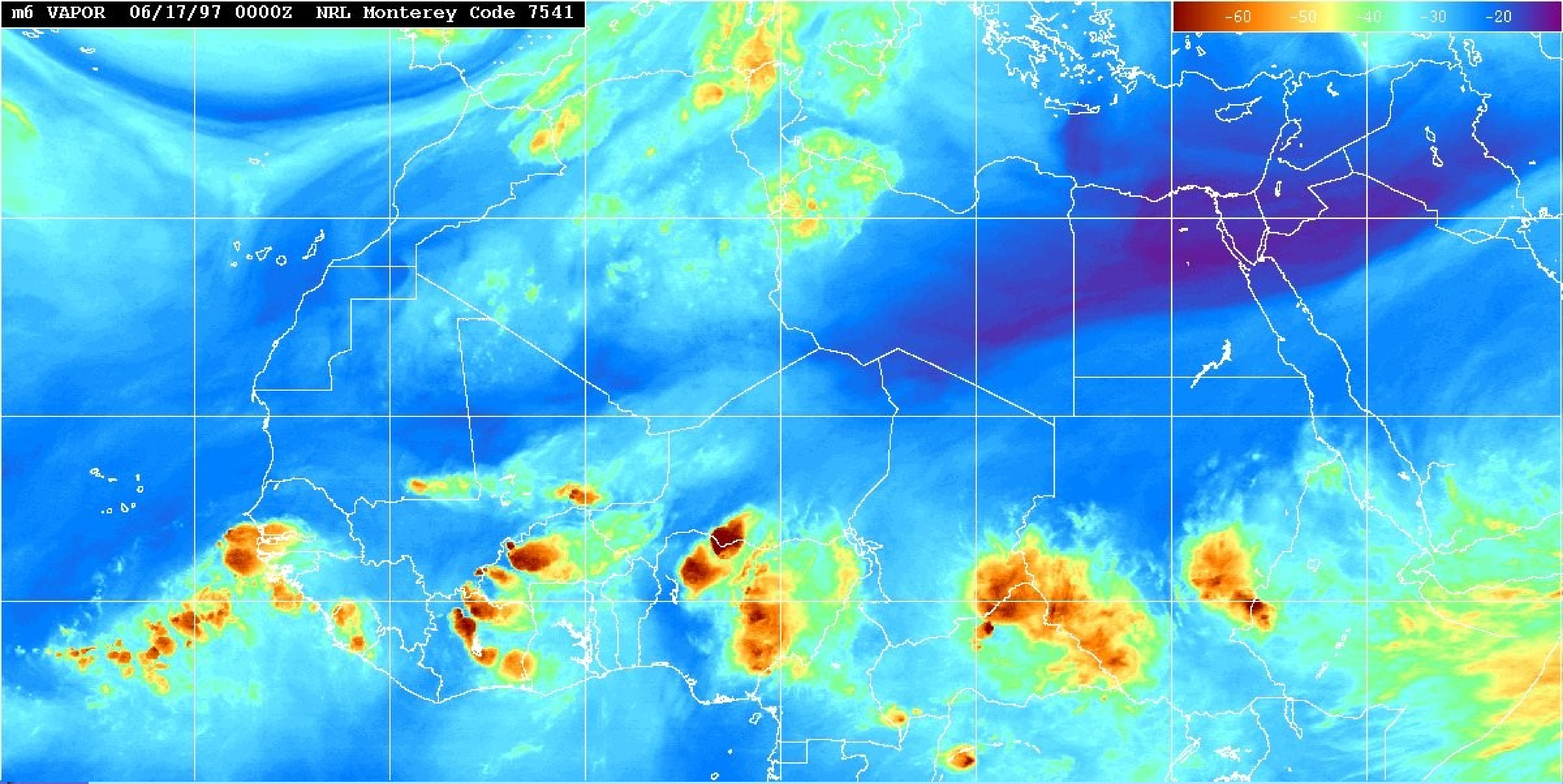
Houze,RG2004





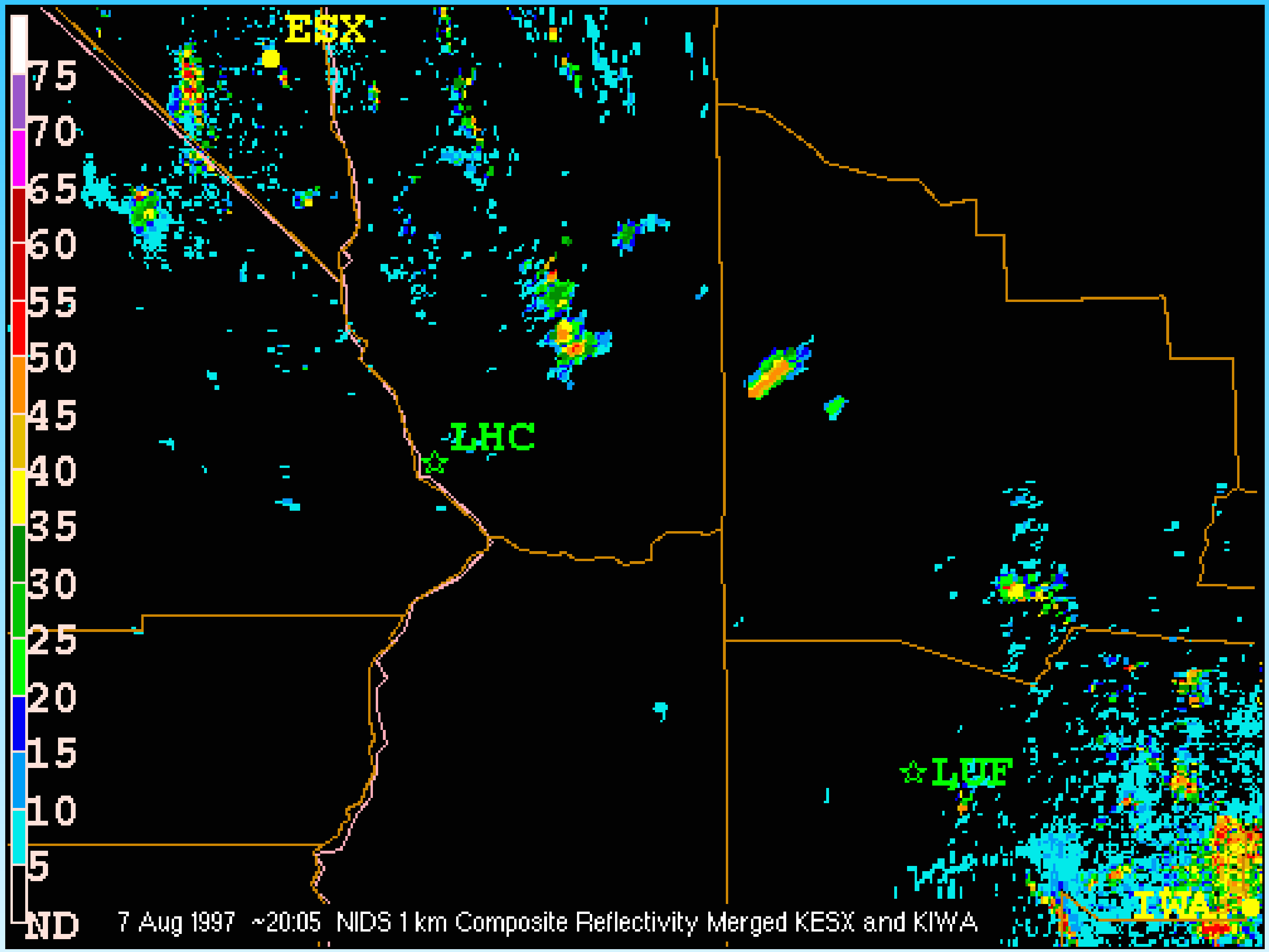
Meso-scale convective system : intense and persistent precipitations over a range of more than 100 km.

A variety of shapes (radar images)



Meso-scale organisation : squall line in Africa
Water vapour channel of Meteosat

Squall line : formation of an alignment of convective cells



Echo radar composite

Frequency of MCS

Laing & Fritsch, 1997

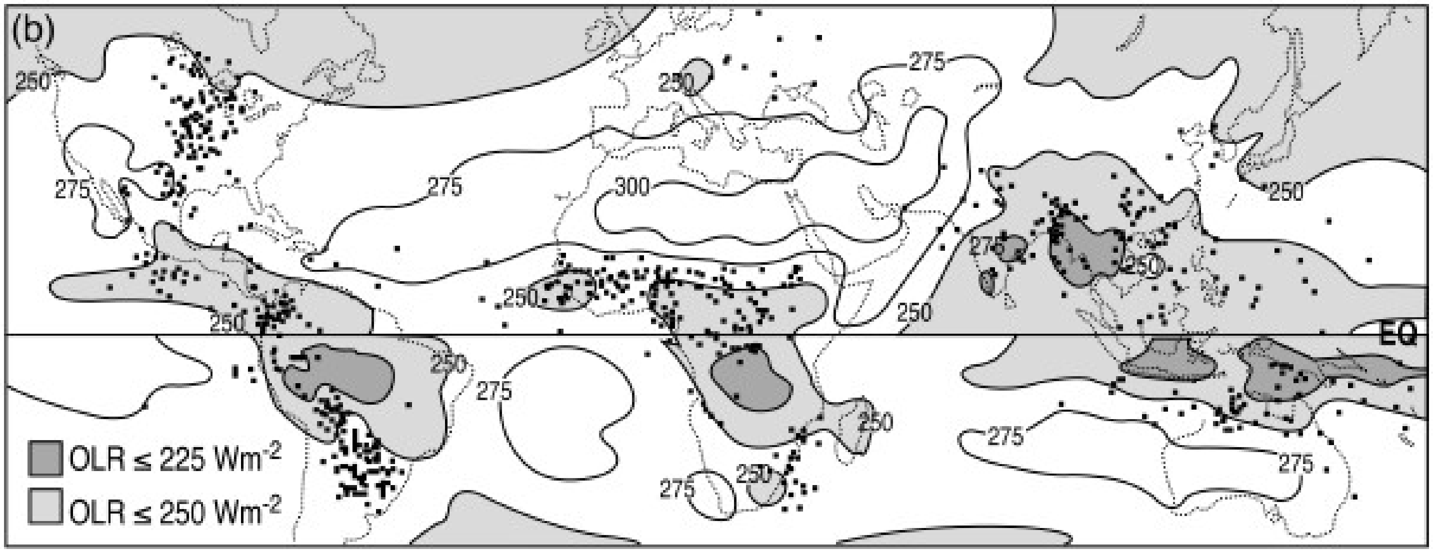


Figure 46. Global distribution of mesoscale convective complexes (dots) and regions of widespread frequent deep convection as inferred by outgoing long-wave radiation

Lightnings

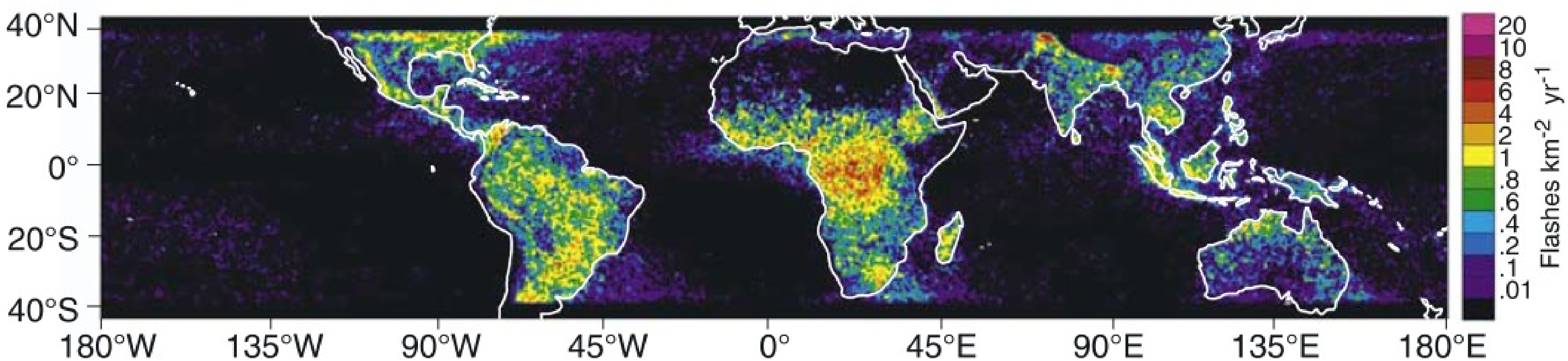


Figure 47. Annual average lightning flash density (flashes per month) for June, July, and August derived from the TRMM Lightning Image Sensor. Courtesy of S. Nesbitt, Colorado State University, Fort Collins.

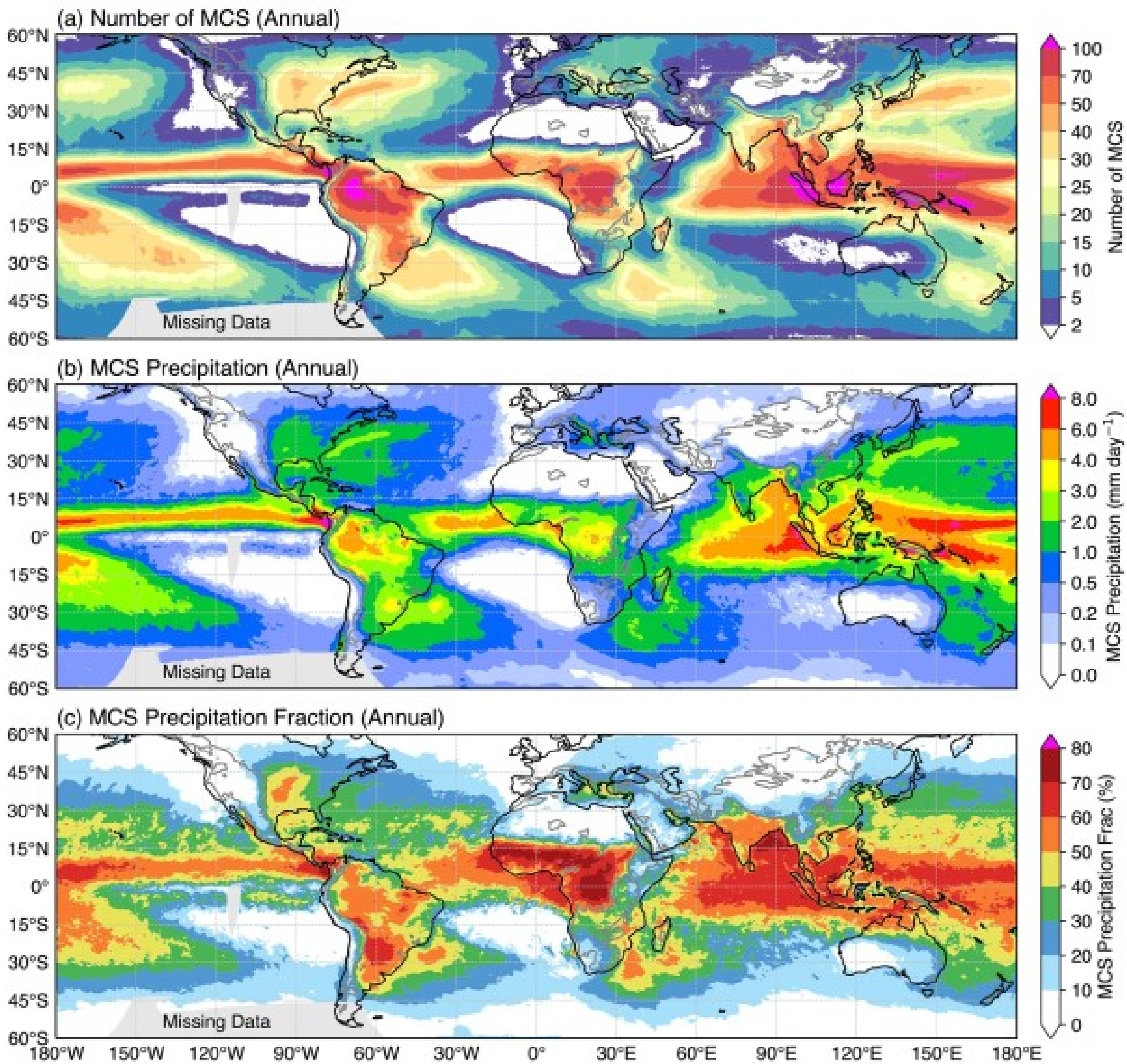
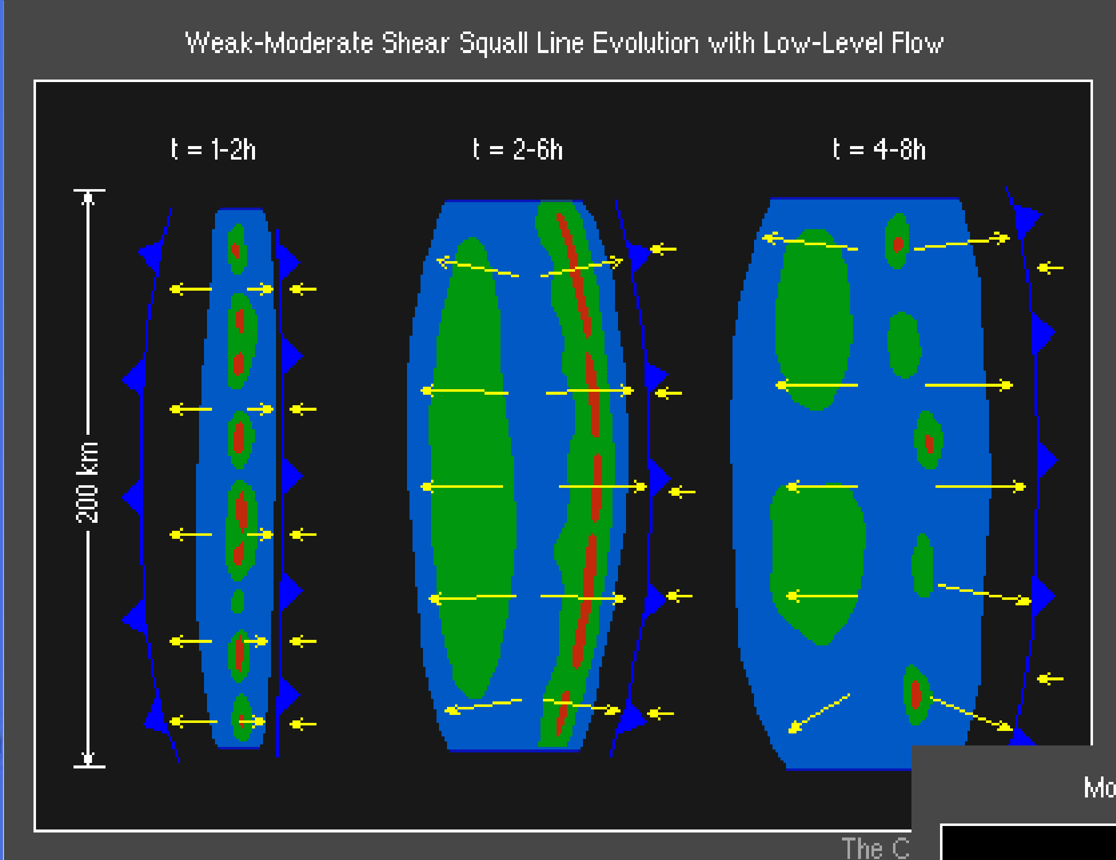


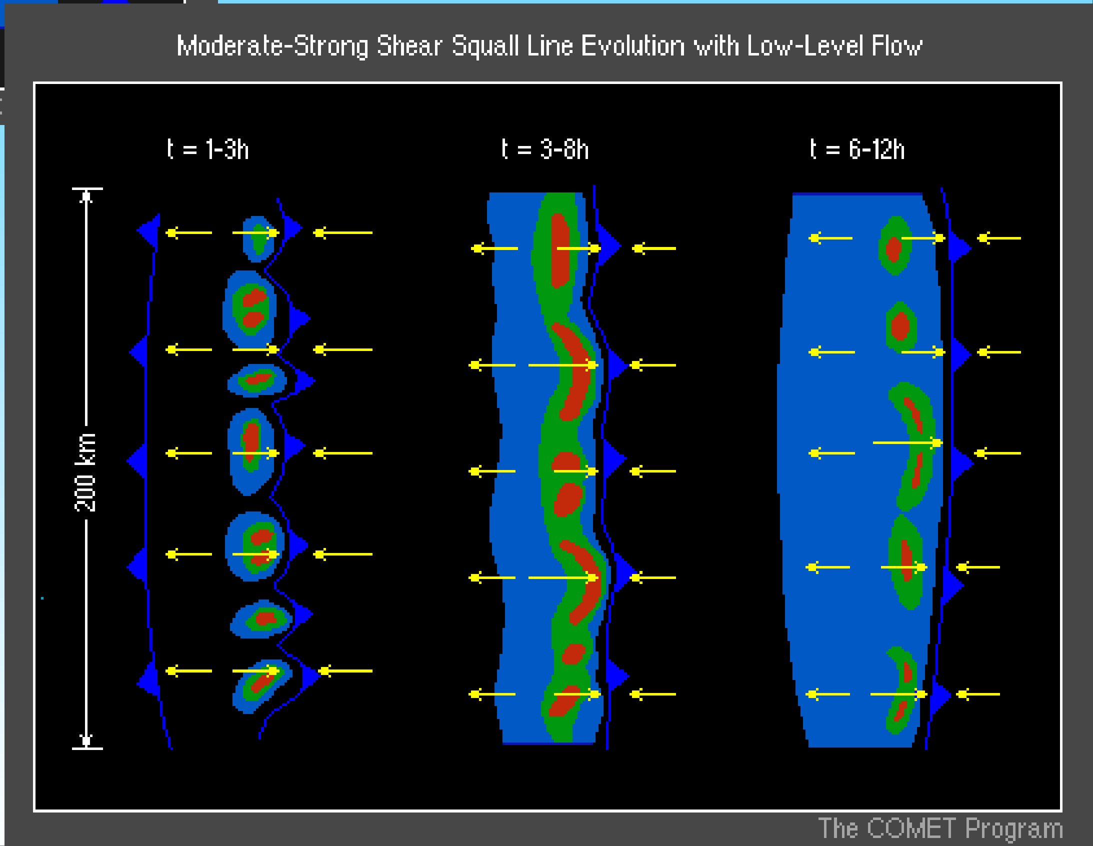
Figure 10. Annual mean global distribution of (a) the number of MCS, (b) MCS precipitation amount, and (c) percentage of MCS precipitation to total precipitation between 2001 and 2019. Dark gray contours show terrains higher than 1,000 m. The gray shaded regions over the Southern Pacific Ocean have frequent (>25%) missing T_b data that affects MCS tracking and is therefore masked out. MCS, mesoscale convective system.

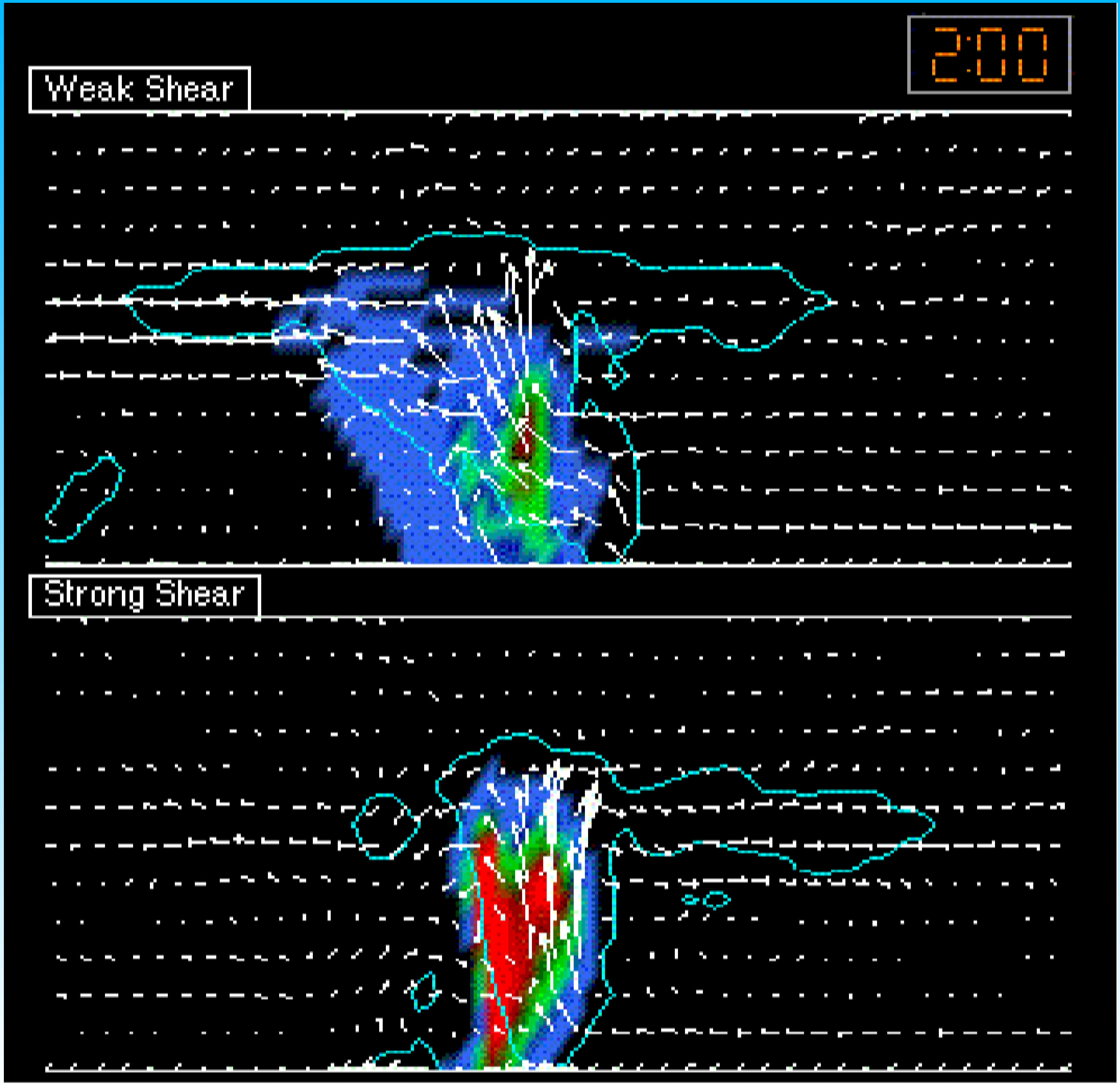
Feng et al., JGR, 2021, 10.1029/ 2020JD034202

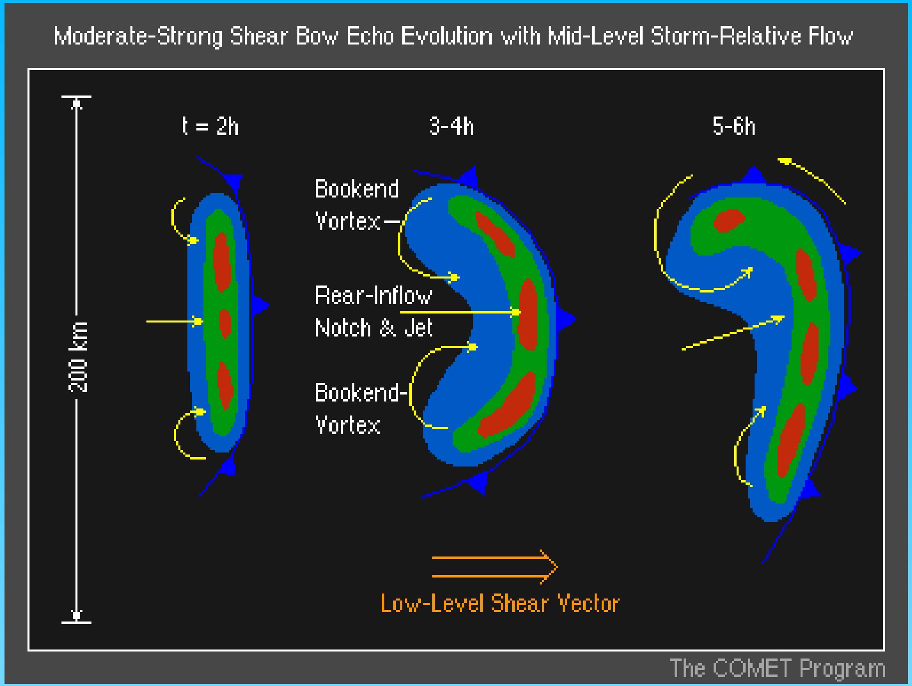


Sqaull line propagating within a moderate shear.
Evolves by getting wider and weaker.

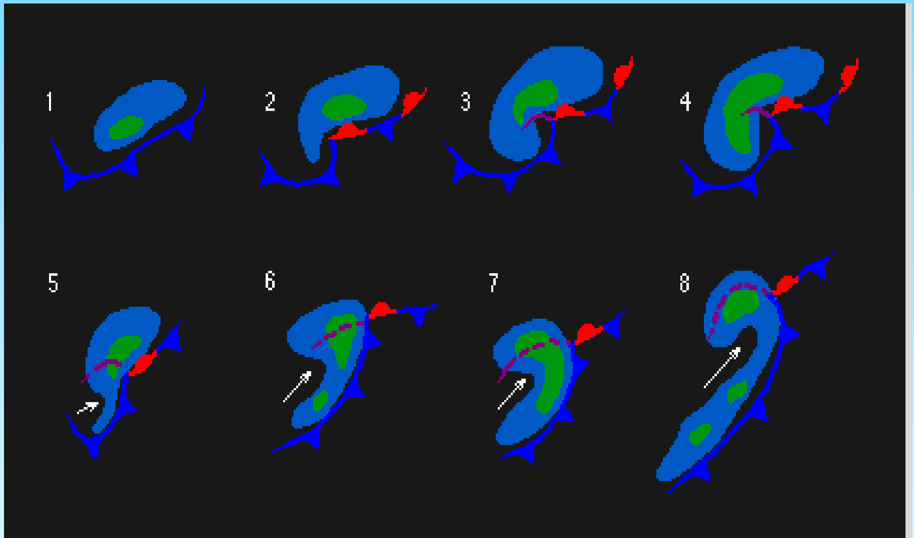
Suall line propagating within a strong shear.
Intense cells, often shapped as arcs.




2:00
Weak Shear
Strong Shear



Squall lines generating meso-scale vortices (possible seeds of tropical cyclones in some locations)



Evolution of an HP supercell to a bow echo. The arrows show the rear downdraft / outflow.
Modified from Moller et al., 1994