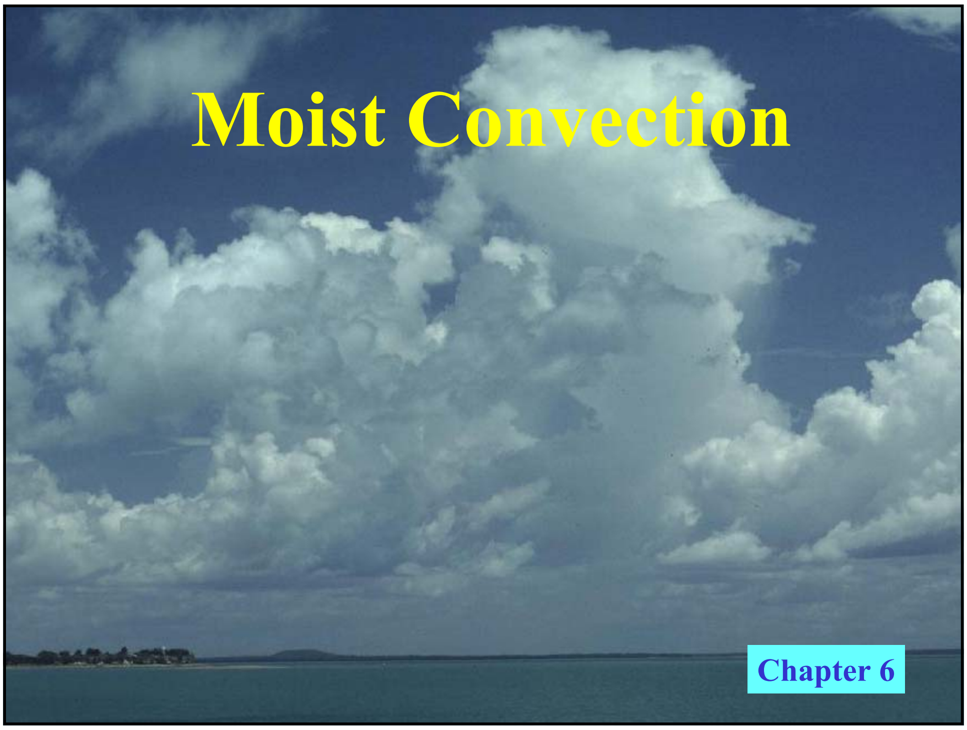

Moist Convection
Chapter 6

Trade Cumuli

Afternoon cumulus over land


## Convectively-driven weather systems

- Deep convection plays an important role in the dynamics of tropical weather systems.
- To make progress in understanding these systems, we must separate the two scales of motion, the large-scale system itself, and the cumulus cloud scale.
- We need to find ways of representing the gross effect of the clouds in terms of variables that describe the large scale itself, a problem referred to as the cumulus parameterization problem.

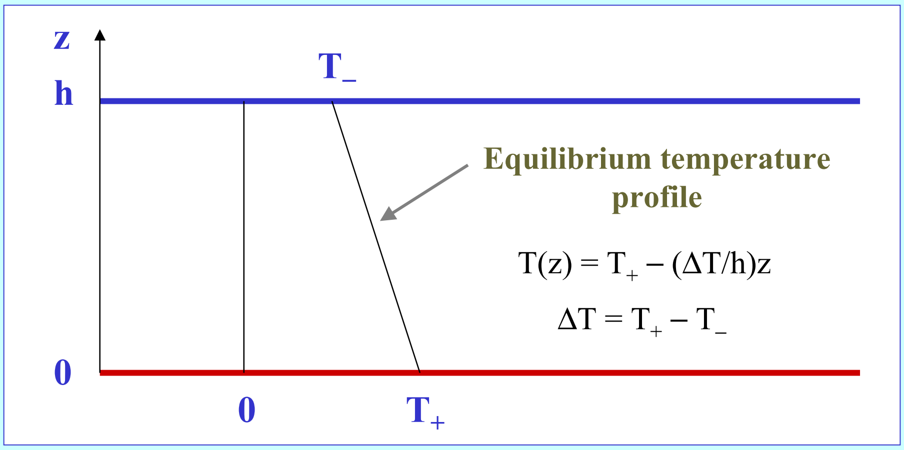
## Understanding convection

- First we consider certain basic aspects of moist convection, including those that distinguish it in a fundamental way from dry convection.
- Then we consider the conditions that lead to convection and the nature of individual clouds, distinguishing between those that precipitate and those that do not.
- Finally we examine the effects of a field of convective clouds on its environment and vice versa.

## Dry versus Moist Convection

### 1. Dry convection

## The classical fluid dynamical problem of convective instability between two horizontal plates



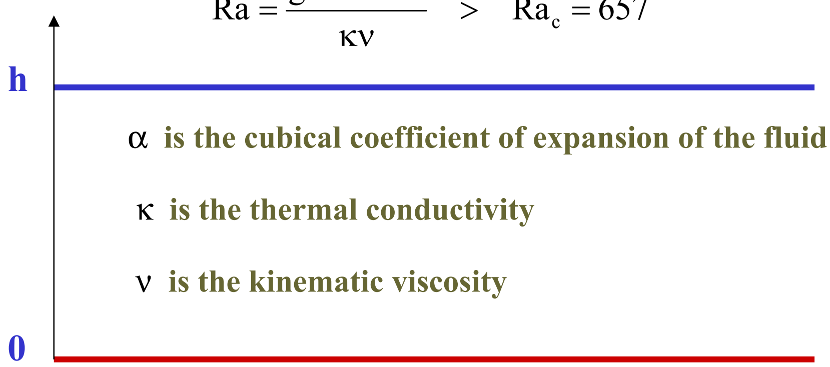

$$T(z) = T_+ - (\Delta T/h)z$$

$$\Delta T = T_+ - T_-$$

**Convectively instability occurs if the Rayleigh number,** Ra, **exceeds a threshold value,** $Ra_c$**.**

## The Rayleigh number criterion

$$Ra = \frac{g\,\alpha\,\Delta T\,h^3}{\kappa\nu} > Ra_c = 657$$



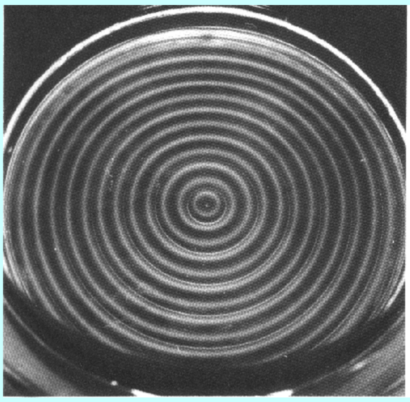

- $\alpha$ **is the cubical coefficient of expansion of the fluid**
- $\kappa$ **is the thermal conductivity**
- $\nu$ **is the kinematic viscosity**

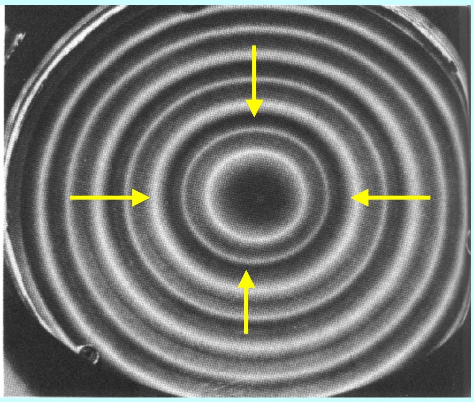
**The Rayleigh number is ratio of the gross buoyancy force that drives the overturning motion to the two diffusive processes that retard or prevent it.**

## The nature of the instability

- For $Ra < Ra_c = 657$, the equilibrium temperature gradient is stable (Lord Rayleigh, 1916).
- For $Ra > Ra_c$, small perturbations to the equilibrium are unstable and overturning motions occur.
- If Ra is only slightly larger than $Ra_c$, the motion is organized in regular cells, typically in horizontal rolls.
- As $Ra - Ra_c$ increases, the cells first take on a hexagonal planform and later become more and more irregular and finally turbulent (Krishnamurti, 1970a,b).
- The turbulent convective regime is normally the case in the atmosphere.

## Circular buoyancy-driven convection cells



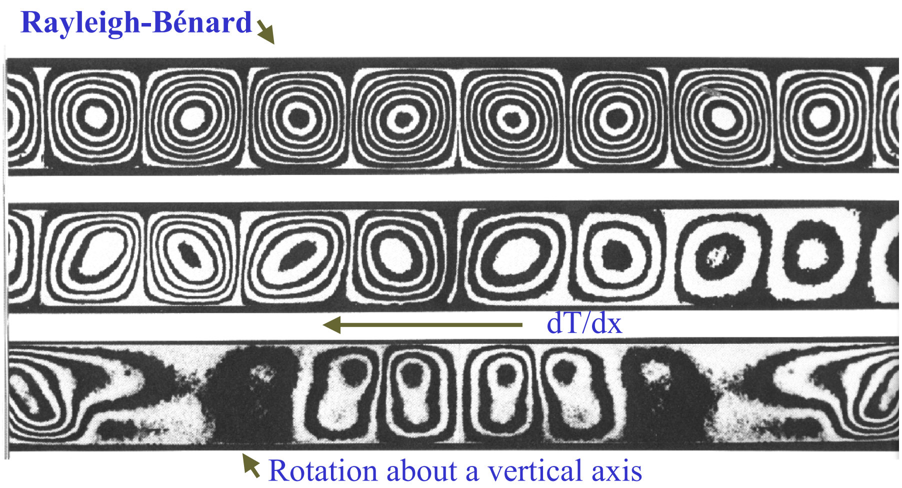


$Ra = 2.9Ra_c$

Uniformly-heated base plate

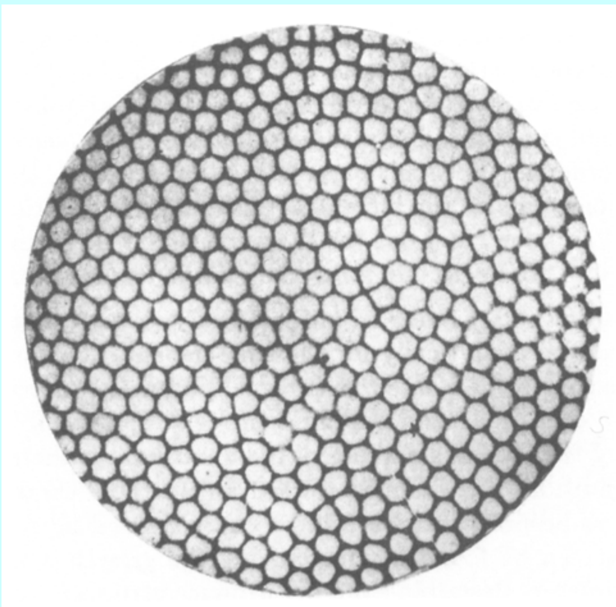
Base plate is hotter at the rim than at the centre

## Buoyancy-driven convection rolls

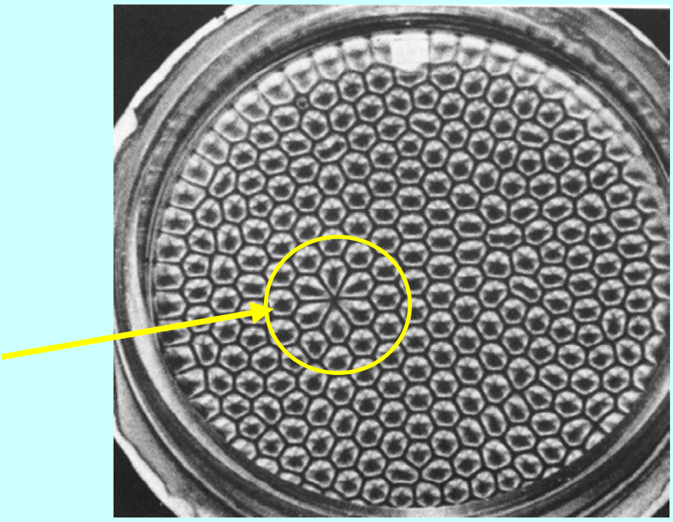


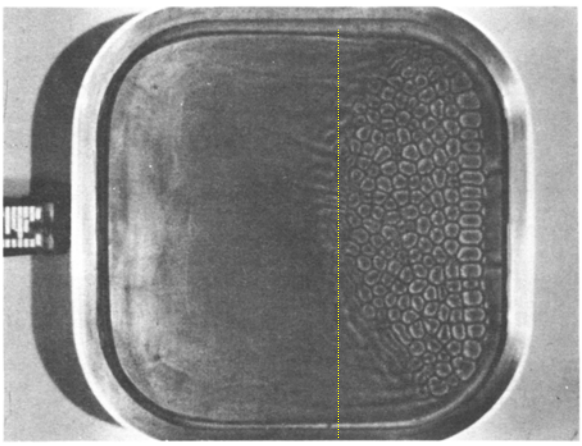

**Differential interferograms show side views of convective instability of silicone oil in a rectangular box of relative dimensions 10:4:1 heated from below.**

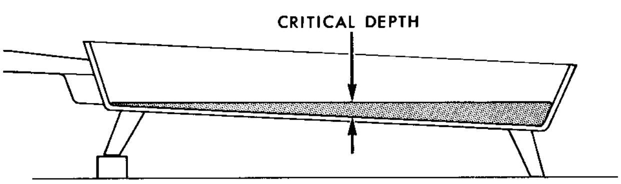
## Bénard convection – hexagonal cells



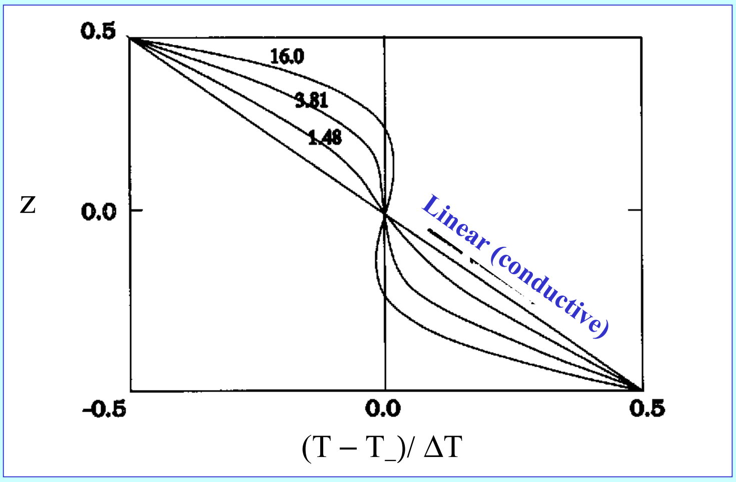
## Imperfections in a hexagonal Bénard convection pattern





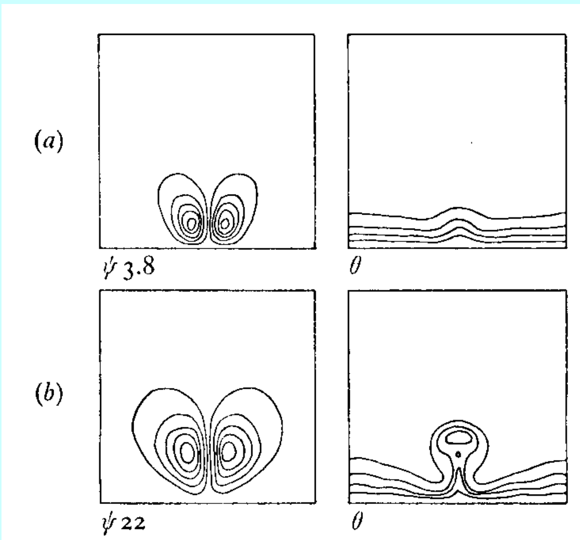


## Temperature profiles as a function of Ra




**As the Rayleigh number increases above $Ra_c$, the vertical profile of the horizontally-averaged temperature departs significantly from the linear equilibrium profile resulting from conduction only.**

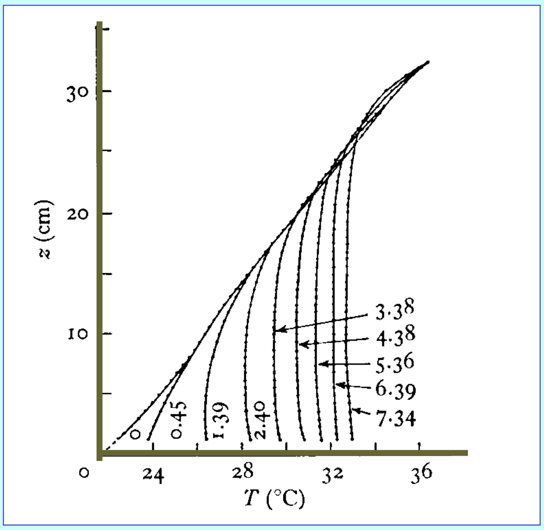
## Penetrative convection



## The formation of plumes or thermals rising from a heated surface

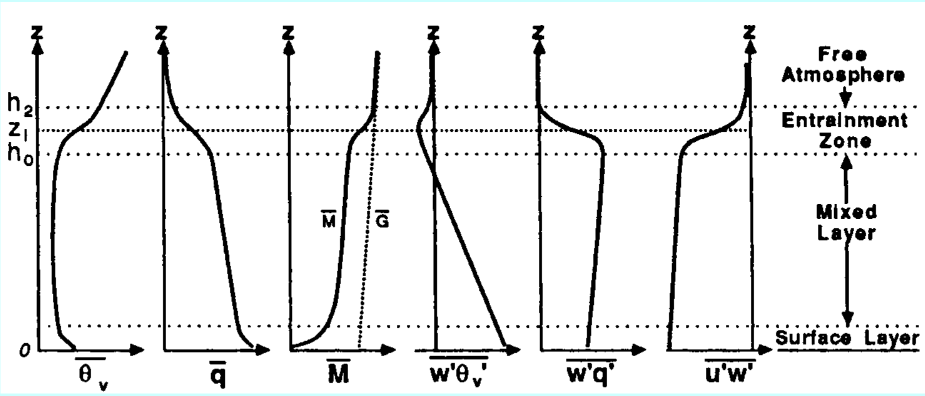
Higher heating rate

**In the turbulent convection regime, the flux of heat from heated boundary is intermittent rather than steady and is accomplished by the formation of thermals**

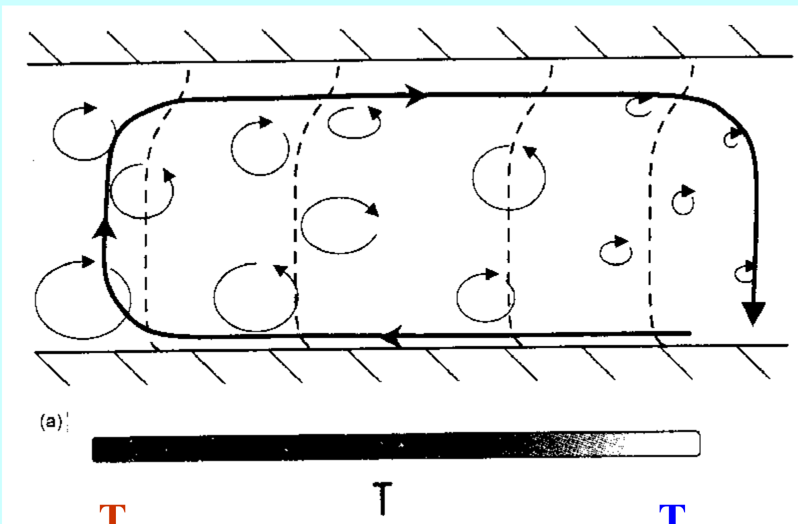



**Vertical profiles of temperature in a laboratory tank, set up initially with a linear stable temperature gradient and heated from below. the profile labels give the time in minutes. (From Deardorff, Willis and Lilly, 1969).**

## Typical profiles of quantities in a convective boundary layer




| | | | | | |
|---|---|---|---|---|---|
| mean virtual potential temperature | mean specific humidity | mean wind speed | buoyancy flux | specific humidity flux | momentum flux |




$T_+$ $T_-$

← boundary temperature gradient

From Emanuel et al., 1994

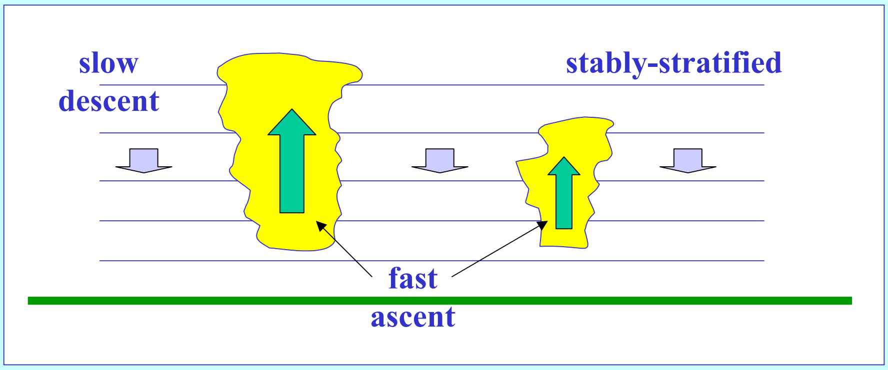
## Dry versus Moist Convection

# 2. Moist convection

## Moist convection

- Let us consider the differences between moist convection and dry convection.
- In dry convection, the convective elements (or eddies) have horizontal and vertical scales that are comparable in size.
- The upward and downward motions are also comparable in strength.
- In moist convection, the regions of ascent occupy a much smaller area than the regions of descent
- The updraughts are much stronger than the downdraughts, except in certain organized precipitating cloud systems.
- Thus, in moist convection there is a strong bias towards kinetic energy production in regions of ascent.

## Moist convection




- Another feature of moist convection that distinguishes it from dry convection is the presence and dynamical influence of condensate.

## Asymmetry of up- and downdraughts

- In moist convection, instability is released in relatively small, isolated regions where water is condensed and evaporated.
- Surrounding regions remain statically-stable and are therefore capable of supporting gravity waves.
- This renders the problem inherently nonlinear, since the static stability of the cloud environment is a function of the vertical velocity therein.

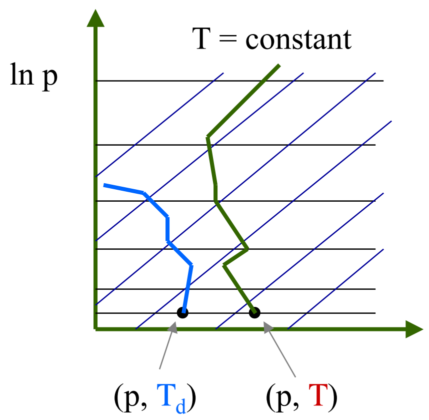
## The conditional nature of moist convection

- An important consequence of phase change and the accompanying release of latent heat is the conditional nature of moist instability.
- A finite amplitude displacement of air to its level of free convection (LFC) is necessary for instability.
- This contributes also to the fundamental nonlinearity of convection.
- In middle latitudes, displacements of air parcels to their LFC can require a great deal of work against the stable stratification.
- Then the problem of when and where conditional instability will actually be released can be particularly difficult.

## Conditional instability

- We usually assess the instability of the atmosphere to convection with the help of an aerological diagram.
- Data on temperature (T), dew-point temperature ($T_d$) and pressure (p) obtained from a radiosonde sounding are plotted on the diagram.
- The two points (p, T) and (p, $T_d$) at a particular pressure uniquely characterize the state of a sample of moist unsaturated air.
- Thus the complete state of the atmosphere is characterized by the two curves on the diagram.

## Aerological diagram with plotted sounding



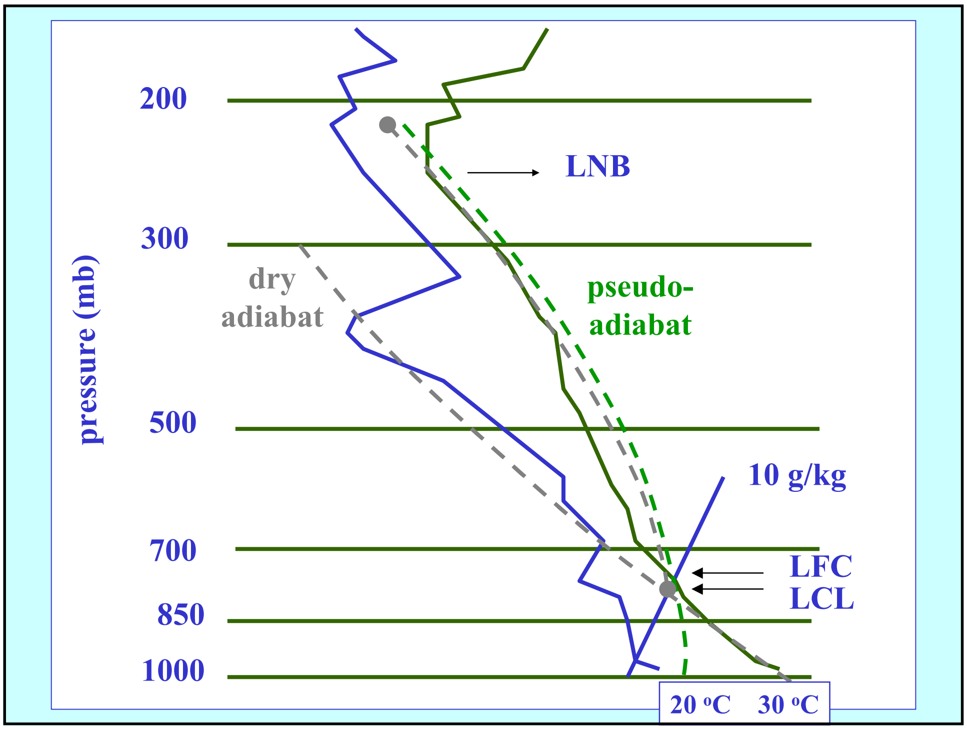

**The effect of the water vapour on density is often taken into account in the equation of state through the definition of the virtual temperature:**

$$T_v = T(1 + r/\varepsilon)/(1+\varepsilon) \approx T(1 + 0.61\,r)$$

**Then, the density of a sample of moist air is characterized by its pressure and its virtual temperature, i.e.**

$$\rho = \frac{p}{RT_v}$$

**Moist air** $(r > 0)$ **has a larger virtual temperature than dry air** $(r = 0)$ **=> the presence of moisture decreases the density of air --- important when considering the buoyancy of an air parcel!**




## Positive and Negative Area
## Convective Inhibition (CIN)

**The positive area (PA)**

$$PA = \frac{1}{2} u_{LNB}^2 - \frac{1}{2} u_{LFC}^2 = \int_{p}^{p_{LFC}} \left(T_{vp} - T_{va}\right) R_d \, d\ln p$$

**The negative area (NA) or convective inhibition (CIN)**

$$NA = CIN = \int_{p_{LFC}}^{p_{parcel}} \left(T_{vp} - T_{va}\right) R_d \, d\ln p$$

## Convective Available Potential Energy - CAPE

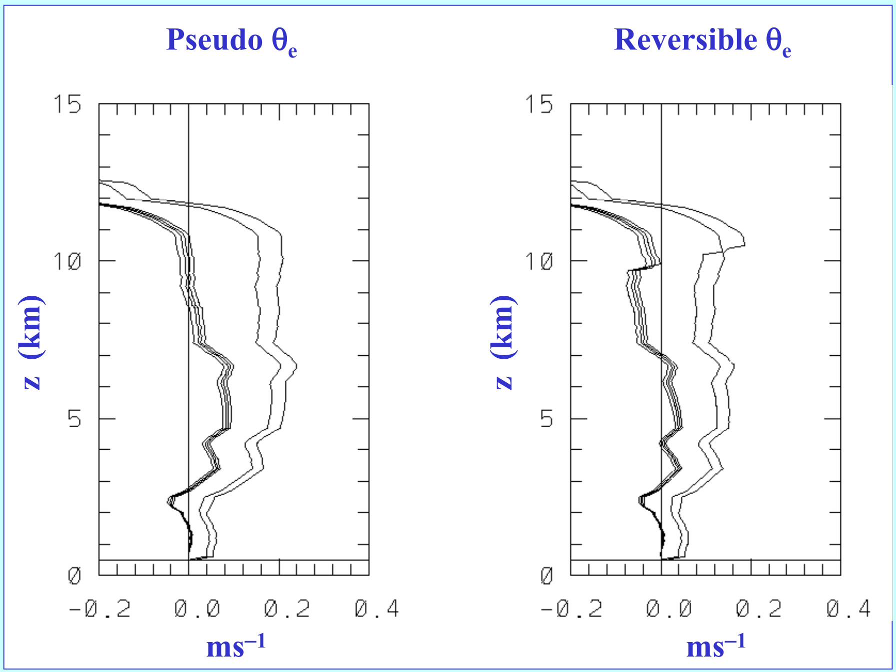
**The convective available potential energy or CAPE is the net amount of energy that can be released by lifting the parcel from its original level to its LNB.**

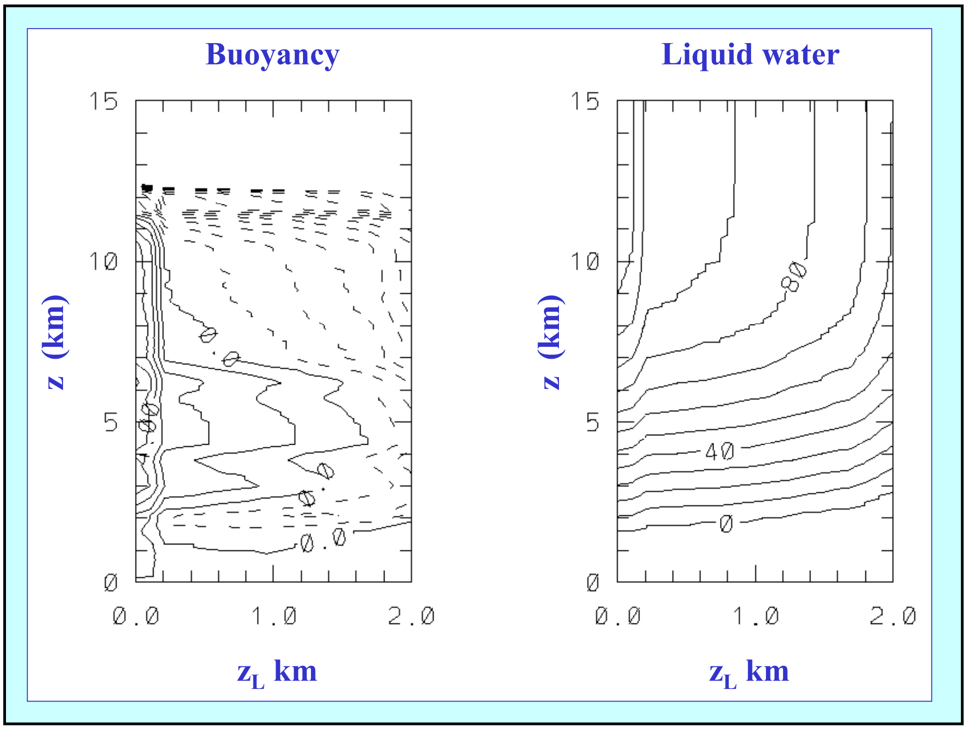
$$CAPE = PA - NA$$

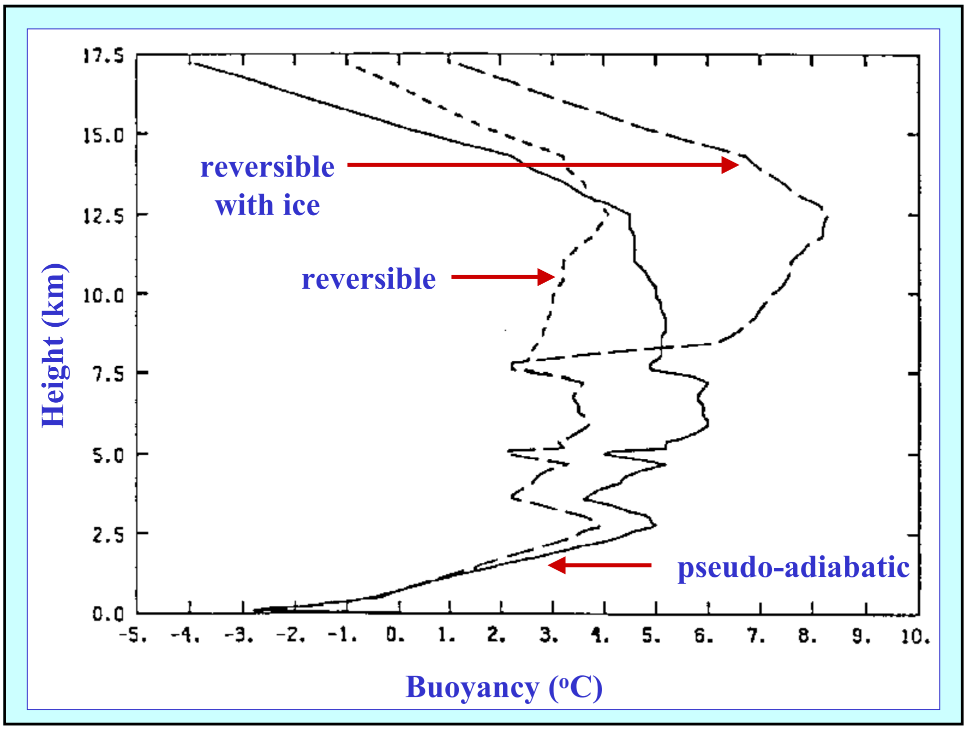
**We can define also the downdraught convective available potential energy (DCAPE)**

$$DCAPE_i = \int_{p_i}^{p_o} R_d (T_{\rho a} - T_{\rho p}) d \ln p$$

**The integrated CAPE (ICAPE) is the vertical mass-weighted integral of CAPE for all parcels with CAPE in a column.**

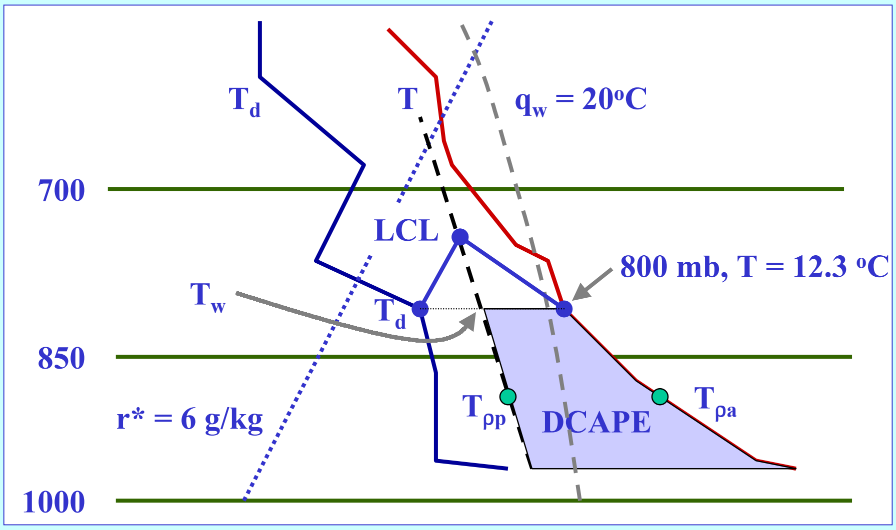



Buoyancy
Liquid water
z (km)
z (km)
zL km
zL km
15
10
5
0
0.0
1.0
2.0
40
80
0


reversible with ice
reversible
pseudo-adiabatic
Height (km)
Buoyancy (oC)
17.5
15.0
12.5
10.0
7.5
5.0
2.5
0.0
-5. -4. -3. -2. -1. 0. 1. 2. 3. 4. 5. 6. 7. 8. 9. 10.

## Downdraught convective available potential energy (DCAPE)

$$DCAPE_i = \int_{p_i}^{p_o} R_d (T_{\rho a} - T_{\rho p}) d\ln p$$




## The pseudo-equivalent potential temperature

- **Approximation:** neglect the liquid water content (set $r_L = 0$).
- Then the approximate forms of $\theta_e$ is a function of state and can be plotted in an aerological diagram.
- Adiabatic processes where the liquid water is ignored are called **pseudo-adiabatic processes**. **They are not reversible!**
- The formula $\theta_e^* \approx \theta \exp (L_v r^*/c_{pd} T)$ is an approximation for **the pseudo-equivalent potential temperature** $\theta_{ep}$ .
- A more accurate formula is:

$$\theta_{ep} = T\left(\frac{p_o}{p}\right)^{0.2854/(1-0.28r)} \exp\left[r(1+0.81r)\left(\frac{3376}{T_{LCL}} - 2.54\right)\right]$$

Temperature at the LCL

## The Lifting Condensation Level Temperature

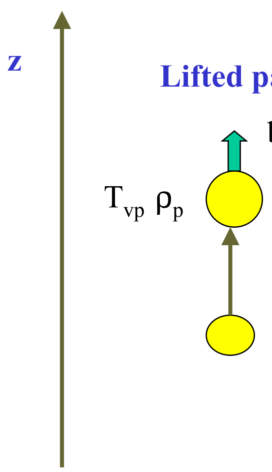
- $T_{LCL}$ **is given accurately (within** 0.1°C**) by the empirical formula**:

$$T_{LCL} = \left[ \frac{1}{T_d - 56} + \frac{\ln(T_K / T_d)}{800} \right]^{-1} + 56$$

$T_K$ **and** $T_d$ **in degrees K**

**[Formula due to Bolton, MWR, 1980]**

## Buoyancy and $\theta_e$




**Lifted parcel**

$$b = -g \frac{(\rho_p - \rho_o)}{\rho_o}$$

**Environment**

$T_o(z)$, $r_o(z)$, $p(z)$, $\rho_o(z)$

**or**

$$b = g \frac{(T_{vp} - T_{vo})}{T_o}$$

**Lifted parcel** **Environment**

$$b = g\frac{(T_{\rho p} - T_{vo})}{T_o}$$

$T_{\rho p}\ \rho_p$ $T_o(z),\ r_o(z),\ p(z),\ \rho_o(z)$

**Below the LCL ($T_{\rho p} = T_{vp}$)**

$$\text{sgn}(b) = \text{sgn}\ \{\ T_p(1+\varepsilon r_p) - T_o(1+\varepsilon r_o) = T_p - T_o + \varepsilon[T_p r_p - T_o r_o]\}$$

$\varepsilon = 0.61$

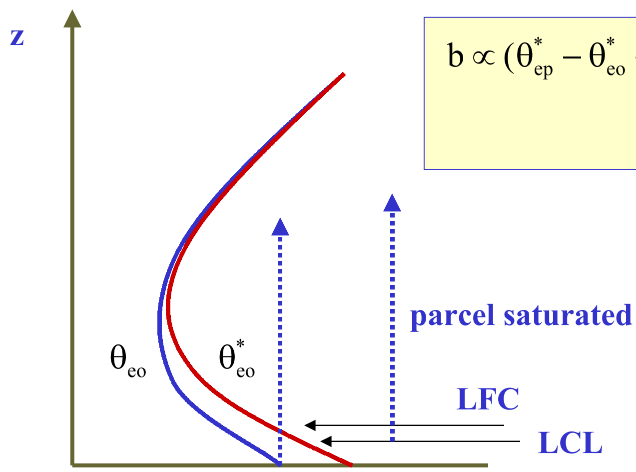
**At the LCL ($T_{\rho p} = T_{vp}$)**

$$\text{sgn}(b) = \text{sgn}\ [T_p(1+\varepsilon r^*(p,T_p)) - T_o(1+\varepsilon r_o)]$$

$$= \text{sgn}\ [T_p(1+\varepsilon r^*(p,T_p) - T_o(1+\varepsilon r^*(p,T_o)) + \varepsilon T_o(r^*(p,T_o) - r_o)]$$

$$\text{sgn}(b) = \text{sgn}\ [T_p(1+\varepsilon r^*(p,T_p) - T_o(1+\varepsilon r^*(p,T_o)) + \varepsilon T_o(r^*(p,T_o) - r_o)]$$

**Since $T_v$ is a monotonic function of $\theta_e$,**




$$b \propto (\theta^*_{ep} - \theta^*_{eo} + \Delta)$$

**small**

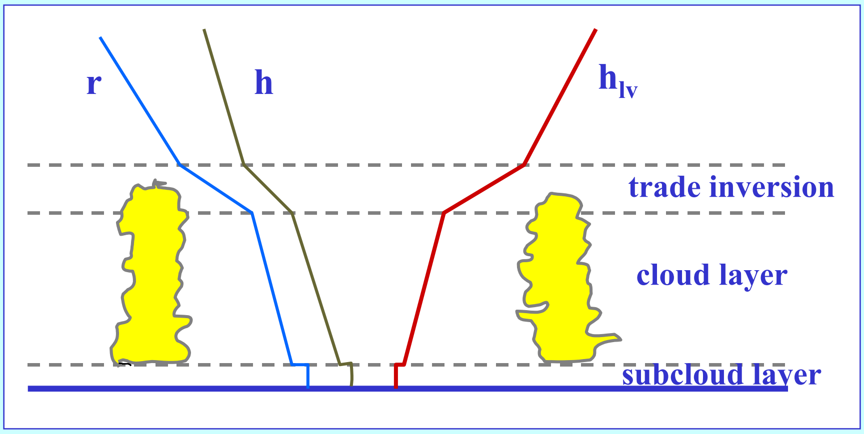
## Shallow convection

- Typically, shallow convection occurs when thermals rising through the convective boundary layer reach their LFC, but when there exists an inversion layer and/or a layer of dry air to limit the vertical penetration of the clouds.
- As the clouds penetrate the inversion, the rapidly reach their LNB; thereafter they become negatively buoyant and decelerate.
- It often happens that the air above the inversion is relatively dry and the clouds rapidly evaporate as a result of mixing with ambient air.
- One can show that this mixing always leads to negative buoyancy in the affected air (see e.g. Emanuel, 1997).

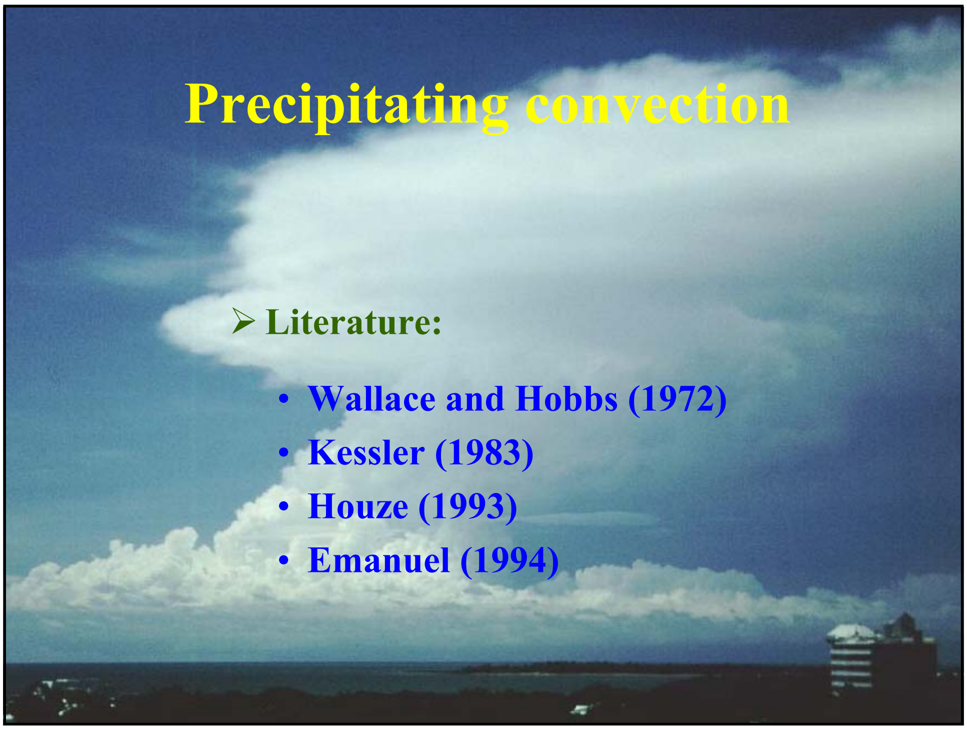
## Shallow convection (cont)

- Shallow clouds transport air with low potential temperature, but rich in moisture, aloft, while the intra-cloud subsidence carries drier air with larger potential temperature into the subcloud layer.
- Thus shallow clouds act effectively to moisten and cool the air aloft and to warm and dry the subcloud layer.
- By definition, shallow clouds do not precipitate and the tiny cloud droplets tend to be carried along with the air.
- The thermodynamic processes within them are better represented by assuming reversible moist ascent rather than pseudo-adiabatic ascent.

Shallow convection in the form of trade-wind cumuli is ubiquitous over the warm tropical oceans



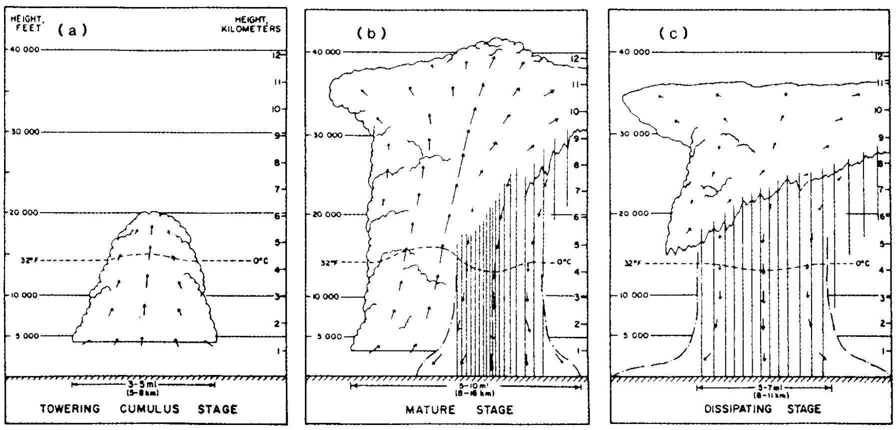

Thermodynamic structure of a trade-cumulus boundary layer



# The common convective shower and thunderstorm

- The Thunderstorm Project (1946-47)
  - Carried out in Florida and Ohio in. Results published by Byers and Braham (1948).
  - Provided a detailed description of the nature of diurnal convection over land (i.e., of air-mass thunderstorms).
  - A typical thunderstorm is a complex of individual convection cells.
  - The evolution can be conceptualized as occurring in three stages: the cumulus stage, the mature stage, the dissipating stage.

cumulus stage | mature stage | dissipating stage




From Byers and Braham (1948)

## Lifetime of convective clouds

- The lifetime of a convective cell is governed by the time it takes the cloud to grow to the point where precipitation forms and the time it then takes for the precipitation to fall to low levels, i. e.

$$\tau \approx \frac{H}{w_o} + \frac{H}{V_T}$$

depth of convecting layer

typical updraught speed

characteristic fall speed of precipitation

- Typical values reported by Byers and Braham (1948) are: $w_o$ and $V_T \cong 5 - 10 \text{ ms}^{-1}$ and $H \cong 10$ km giving $\tau \cong 0.5$ h to 1 h, in good agreement with observations.

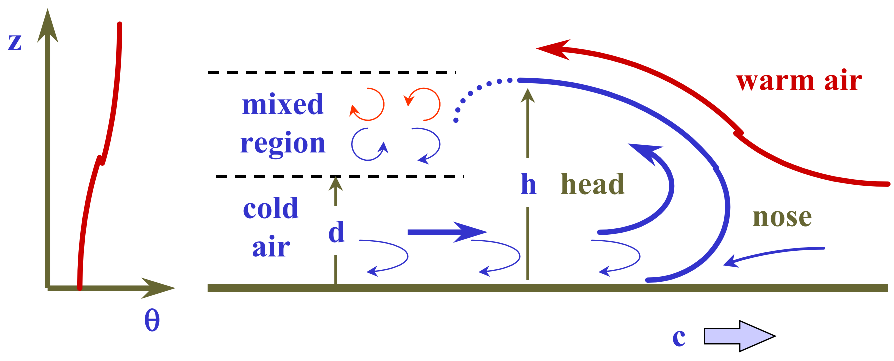
## Thunderstorm characteristics (1)

- The individual updraught and downdraught sampled in the Thunderstorm Project had diameters of 1 - 2 km.
- The updraught magnitudes were mostly in the range 5 - 10 $\text{ms}^{-1}$, with some exceeding 15 $\text{ms}^{-1}$. downdraughts were comparable in magnitude, but usually weaker.
- There is a tendency of individual cells in ordinary thunderstorms to cluster.
- The cell complex may last for several hours.
- It usually translates at a speed corresponding to the average wind in the cloud layer.

## Thunderstorm characteristics (2)

- In conditions favourable to air mass thunderstorms, the vertical wind shear is relatively small.
- Byers and Braham noted a particular side of the existing complex.
- The clumping of individual cells in convective storms is most likely the result of the spreading cold air at the surface.

## Theory of gravity currents




Schematic diagram of a steady gravity current

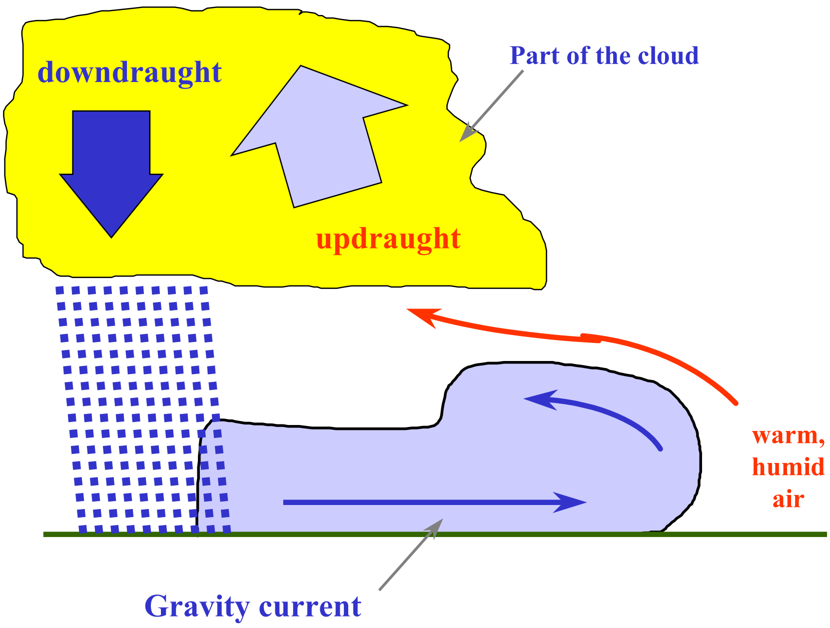
## The gust front

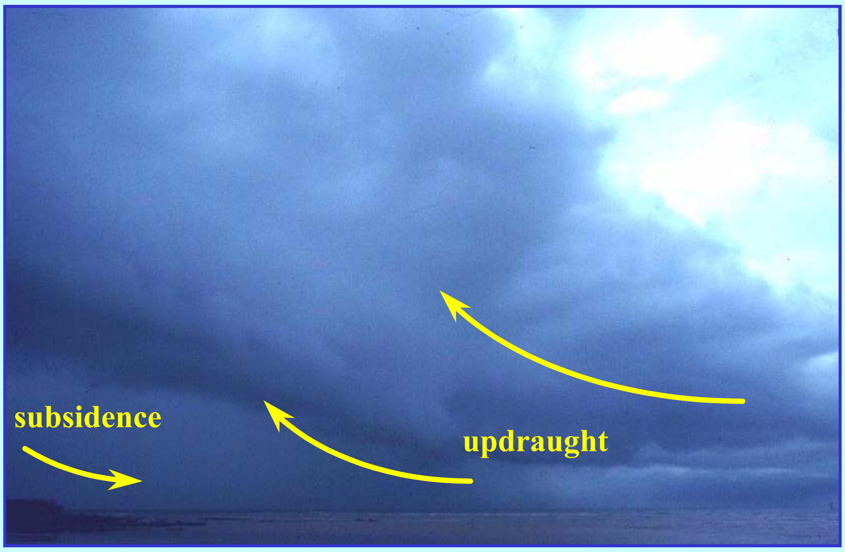
- The gust front gives rise to updraughts along its leading edge with magnitudes comparable to the flow-relative propagation speed of the current,

$$c^2 = \gamma g h \frac{\theta_{vd} - \theta_{va}}{\theta_{va}}$$

$\gamma$ is a dimensional constant of order 1 and h is the depth of the cold air.

**Typically** $c \sim 10 - 15\,\mathrm{ms}^{-1}$






Near-surface outflow from a thunderstorm

Arcus-cloud - Oklahoma

# The End