Cloud Microphysics and Properties. Part II: Moisture and Stability

Course on “atmospheric aerosols and clouds with introduction to process oriented modeling”,
Sao Paulo University

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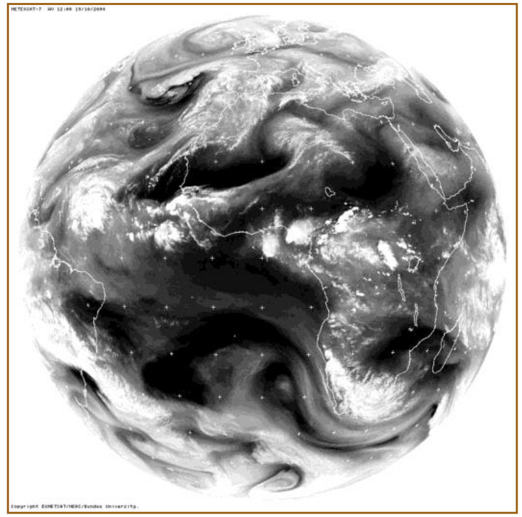
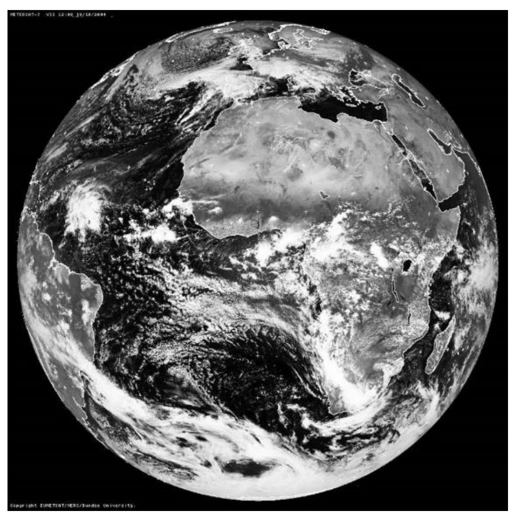
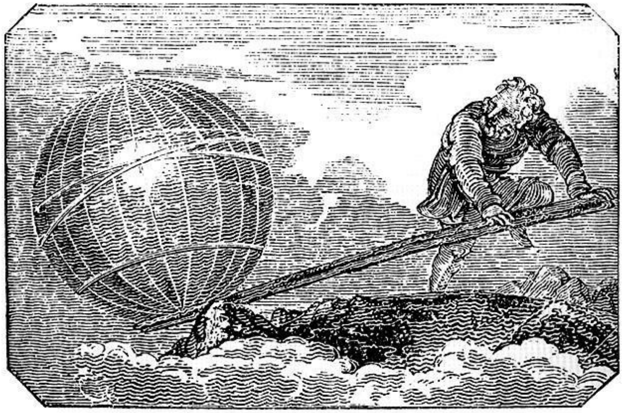


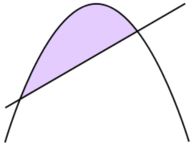


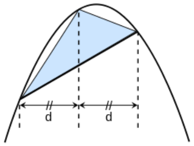
Outline

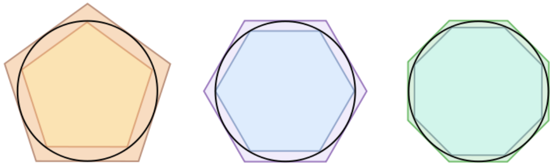
- » Introduction
- » Buoyancy and Moisture
- » Tephigrams
- » Humidity and Temperature
- » Stability
- » Overview of Triggering Mechanisms
- » Summary

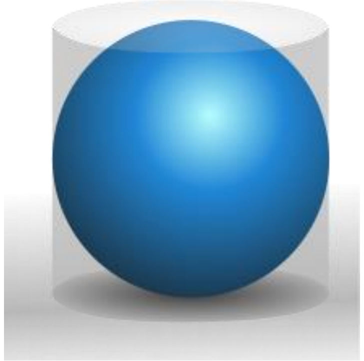
INTRODUCTION

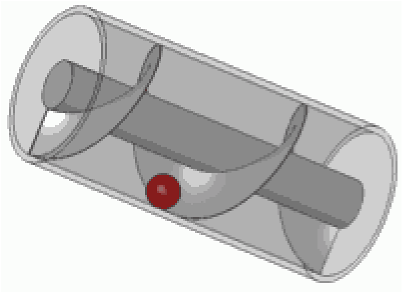




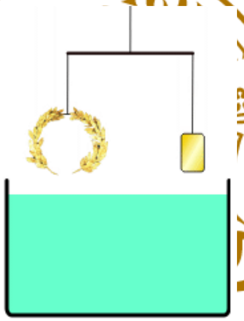




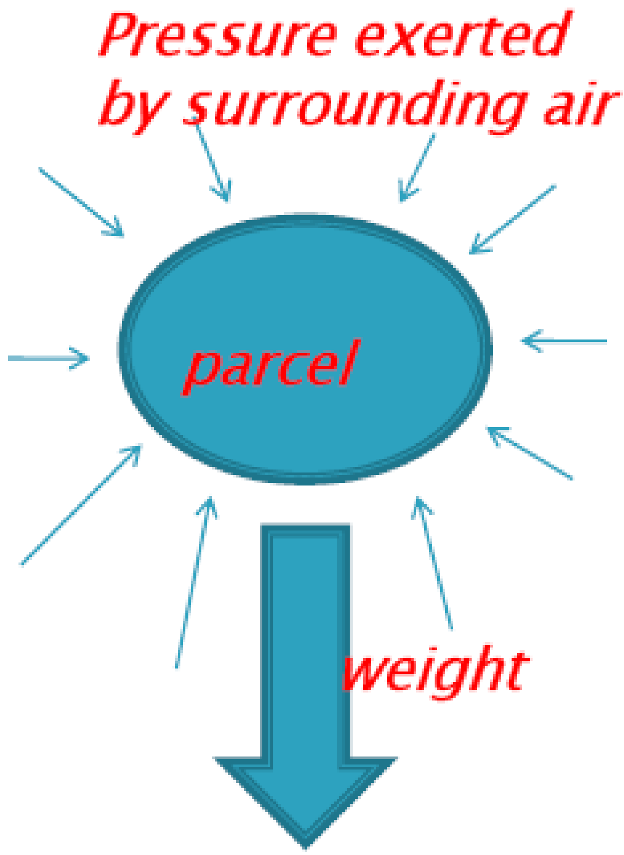




Buoyancy of a parcel is the vector sum of two forces acting on it



Upthrust is the net pressure force exerted on parcel by environment



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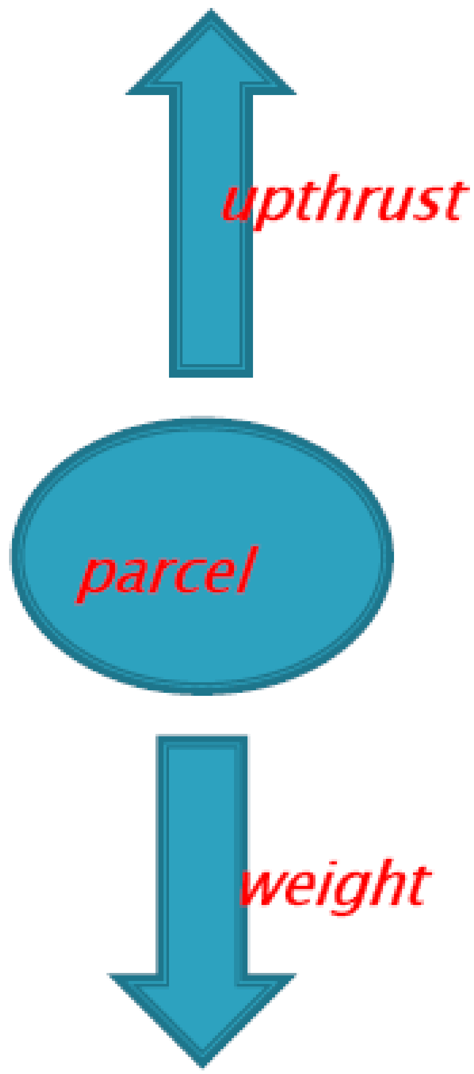


Buoyancy Force

- Generally convection is motion of fluid that transfers heat
  - convection of convective clouds is driven by buoyancy force
  - **Buoyancy force = upthrust acting on parcel – weight of parcel**
- Archimedes principle: upthrust on a body is equal to the weight of the fluid displaced by it
  - Upthrust exerted on parcel (V, T_{parcel}, ρ_{parcel}, p_{parcel}) by environment (T_{env}, ρ_{env}) is $\rho_{env}\, g\, V$, while parcel's weight is $\rho_{parcel}\, g\, V$, so their vector sum gives:

$$\boxed{F_B \equiv -g\frac{\rho'}{\rho_{env}}}$$

- where F_B is buoyancy force per kg of air in parcel and $\rho' = \rho_{parcel} - \rho_{env}$

Buoyancy Force

- Air parcel at rest is 'neutrally buoyant': $F_B = 0$
- +vely buoyant parcel accelerates upward: $F_B > 0$
- -vely buoyant parcel accelerates downward: $F_B < 0$
- Ideal Gas Law: $p = \rho R T$ where $R = R^*/m$ is individual gas const., m is molecular wt., $R^* = 8.3145$ J/mol/K is universal gas const.
  - for dry air, $R = R' = 287$ J kg^{-1} K^{-1}
  - for moist air, $R = R_m = R'(1 + 0.6\, q_v)$
- Parcel approximation: $p_{parcel}(z) \approx p_{env}(z)$
- Neglecting other vertical forces and with $T' = T_{parcel} - T_{env}$

$$\boxed{\frac{Dw}{Dt} \approx F_B \approx g \frac{T'}{T_{env}}}$$

- Warm (cold) parcels may rise (sink), being lighter (denser) than environment, lowering overall gravitational potential energy (PE)

Adiabatic Processes

- An **adiabatic** process: **no energy or mass enters or leaves the system**.
- Many atmospheric processes are adiabatic (or nearly) – especially if involving vertical motion of air.
  - Air is a poor conductor of heat
  - mixing is often slow enough for a body of air to retain its identity distinct from the surrounding air during ascent.
- Near-surface processes are frequently non-adiabatic.

Adiabatic Processes:

- Ascent of dry convective plumes
- Large scale lifting/subsidence

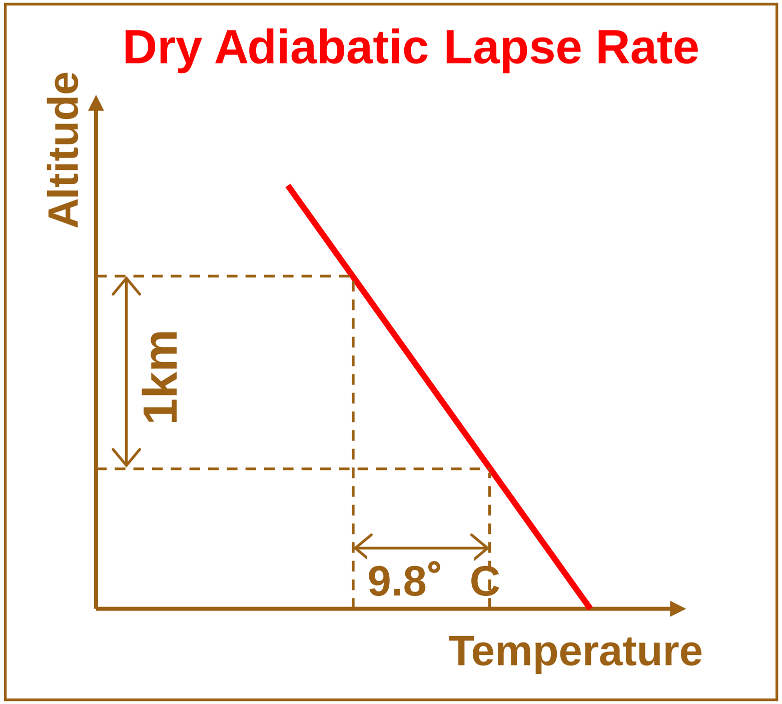
Non-Adiabatic Processes:

- Radiative heating/cooling
- Surface heating/cooling (conduction/convection/latent)
- Removal of water from atmosphere by precipitation
- Water added from evaporation of precipitation falling from above
- Condensation or evaporation in an undilute airmass is “pseudoadiabatic”

Temperature decreases both during ascent of parcels and in the environment with increasing height

Lapse Rate

- **Lapse Rate** is -1 × vertical gradient of temperature.
- Decrease in temperature with height of **dry air** due to a decrease in its pressure is the **Dry Adiabatic Lapse Rate** = g/c_p = 9.8° C/km.
  - From 1st Law of Thermodynamics, parcel approximation and hydrostatic balance:



$dQ = c_p\, dTparcel - \alpha$
$dpparcel$ where $\alpha = 1/\rho$

$$dQ = 0 \;\Rightarrow\; c_p\, dT_{parcel}/dz \approx (R'T_{env}/p_{env})dp_{env}/dz \approx -g$$

Unsaturated adiabatic parcel

Potential temperature, θ, is constant in unsaturated adiabatic parcels

» θ is temperature a parcel would have if compressed adiabatically to 100 kPa

» dQ =0, and integrating 1st Law of Thermodynamics yields

$$\theta \equiv T\left(\frac{100\ kPa}{p}\right)^{k}$$

$$k = \frac{R'}{c_p} \approx 0.286$$

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

unsat. ad. parcel

$$F_B \approx g\frac{T'}{T_{env}} = g\frac{\theta'}{\theta_{env}}$$

» θ (and q_v) mixes linearly:

– θ of mixture of 2 parcels is weighted-average (by mass) of their initial values of θ

– Vertical mixing within PBL is driven by solar heating of surface and **makes θ** (and q_v) **vertically uniform**

» Proof with1D model: consider long sequence of mixing events among a stack of parcels, with each event occurring after displacement of two parcels adiabatically from different levels to the same level.

BUOYANCY AND MOISTURE

Water vapour condenses to form the mass of clouds, which consist of either cloud-droplets or ice crystals

Atmospheric Water

- Troposphere contains nearly all of the water in the atmosphere
  - Mostly it is water vapour
  - Some is cloud-liquid or cloud-ice too
- Orders of magnitude of (vapour) mixing ratio:
  - ~1 g kg^{-1} in middle troposphere
  - ~10 g kg^{-1} in lower troposphere

Sources and Sinks of Vapour

Sources:

- Evaporation from surface: requires sunlight, conduction from surface to provide latent heat

- Evaporation of precipitation: falling from above: latent heat supplied by cooling of air

Sinks:

- Precipitation: snow, graupel/hail, rain

- Condensation at surface: dew, frost

» Most water in atmosphere above a certain location is **not** from local evaporation there, but was advected from some other remote place

Effects from Moisture on Buoyancy

Atmospheric water influences the dynamics of troposphere

- Convective processes driven by changes in buoyancy from latent heat of its phase changes
- Latent warming and cooling of air during condensation and evaporation promotes ascent and descent respectively

» Also outside clouds, slight effects on buoyancy:

- mixture of humid air is slightly less dense than dry air at the same temperature and pressure
- molecular weight of water = 18 g mol^{-1}
- mean molecular weight of dry air ≈ 29 g mol^{-1}

$\rho_{\text{pure water vapour}} \approx 0.6\ \rho_{\text{air}}$

Latent Heat

Latent heat of evaporation of water is $L_v \approx 2.5$ MJ kg^{-1}

» Compare with specific heat capacity of dry air:

$C_p \approx 1004$ J kg^{-1} K^{-1}

Exercise: evaporate 10 gram of liquid water into 1 m^3 of air:

latent heat absorbed ≈ 25,000 J

air is cooled by ≈ 20 K

Conversely, condensation onto cloud-droplets causes latent heat to be released and air is warmed. Latent heats of condensation and evaporation are equal.

Latent heats of melting and freezing, $L_m \approx 0.3$ MJ kg^{-1}

Latent heat of sublimation and vapour growth $L_s \approx 2.5 + 0.3$ MJ kg^{-1}

TEPHIGRAMS

Tephigram is a meteorological chart for analysis of vertical structure of atmosphere

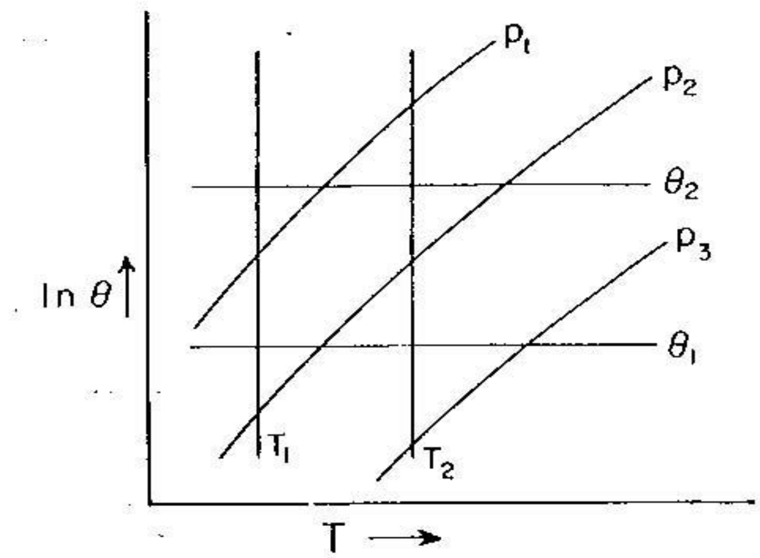
- how favourable environment is for convective clouds

A Tephigram

» Physical state of an air parcel is defined uniquely by two '*state variables*' (see the ideal gas law):

– *p* and *T*

– or *T* and θ

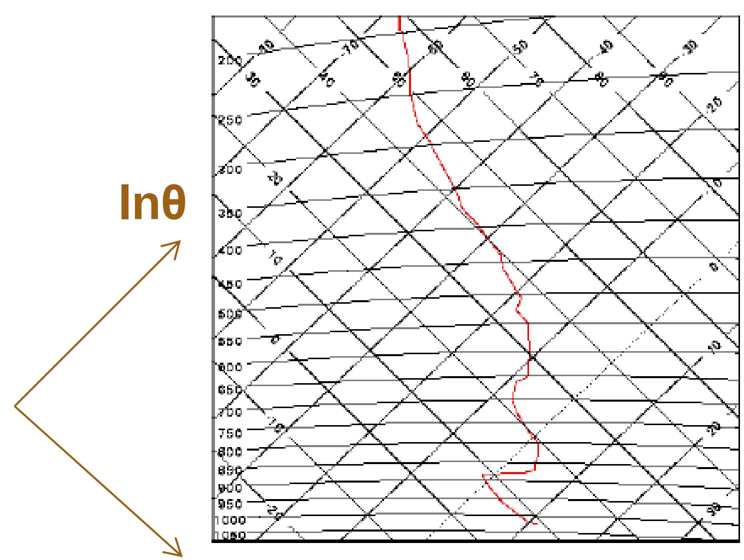


» *A graph of T* vs ln(*θ) has contours of constant pressure that are almost straight lines.

» Rotation of axes by 45º creates a **tephigram**.

» Meteorologists use it to ascertain vertical structure of atmosphere

– E.g. to predict intensity of convective clouds



Hemsby, 28 May 1997 1200UTC

Non-adiabatic reversible transfer of heat, *dQ*, into parcel (*T*), changes entropy by $d\varphi = dQ/T$

$$d\varphi = (c_v dT + p d\alpha)/T = (c_p dT - \alpha dp)/T = c_p\left[\frac{dT}{T} - k\frac{dp}{p}\right] = c_p d\theta/\theta$$

1st Law

$$\Longrightarrow \boxed{\varphi = c_p \ln\theta + constant}$$

$$dQ = 0 \Rightarrow \varphi = constant \Rightarrow \theta = constant$$

Unsaturated adiabatic motions follow isentropic surfaces

» Potential temperature is always a measure of entropy, φ

» In a cyclic process, total heat added to non-adiabatic parcel is

$$\boxed{\oint dQ = \oint T d\varphi = \oint c_p T \mathrm{dln}\,\theta = area\ on\ tephigram \times c_p}$$

» **Area on tephigram always represents energy (/c_p)**

» Radiosondes measure vertical soundings through atmosphere:

– Temperature, pressure, wind-velocity (GPS)

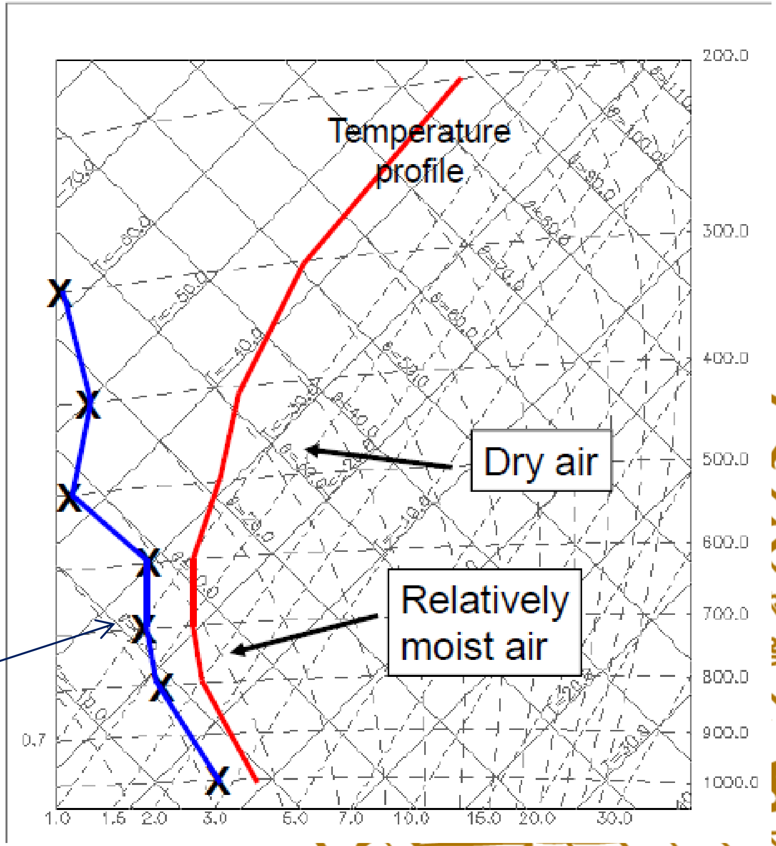
– Dew-point temperature

– Measurements transmitted by radio

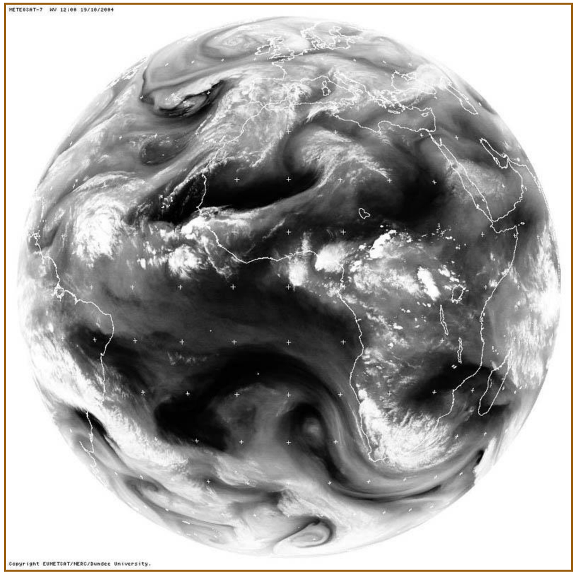
» Dew-point temperature (T_d) is temperature to which air must be cooled to become saturated

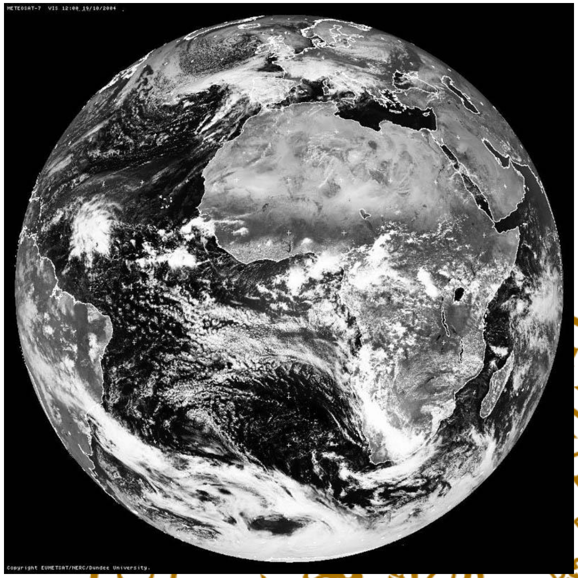
– Measure of moisture in air

– T_d is plotted on tephigram



HUMIDITY AND TEMPERATURE





Vapour

- » Water vapour has a variable mass concentration in the atmosphere
  - – 0 to a few per cent (by mass) of air globally.
- » The most important GHG.
- » Water's phase changes and latent heating/cooling alters buoyancy of air, driving the dynamics of weather systems such as thunderstorms, mid-latitude cyclones ...
- » Emission/absorption of longwave radiation by clouds affects climate
- » Saturated air is a mixture of air, vapour and any condensate
  - – condensation warms this air, making it less dense than unsaturated, drier air at the same pressure and boosting its buoyancy.
- » Vapour's contribution to total pressure = (partial) vapour pressure, e

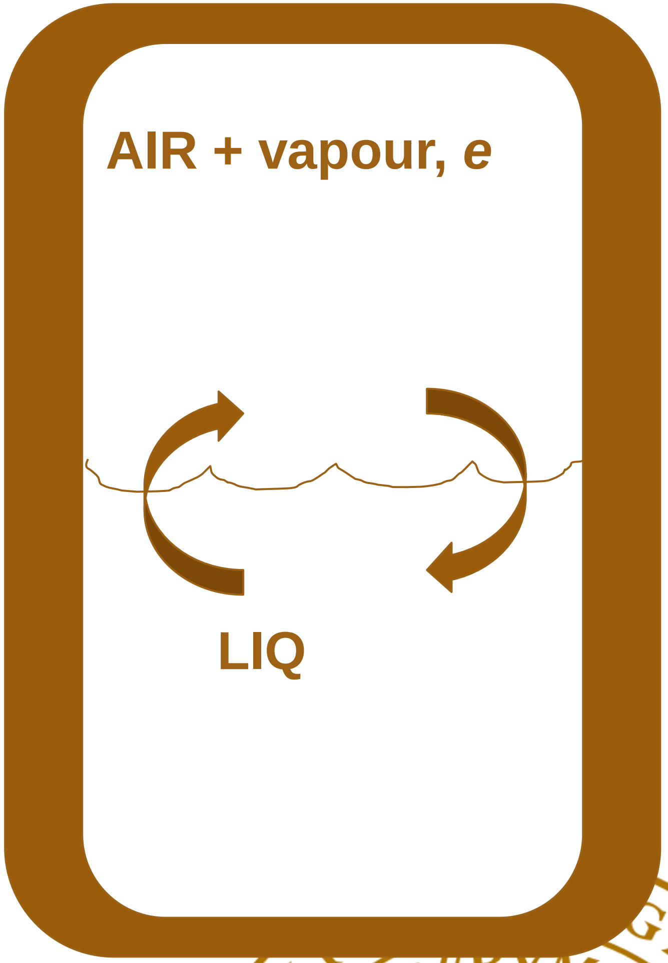
Saturation

» Air is ‘***saturated***’ when there are equal rates of gain and loss of water molecules between liquid and vapour in a closed container

» Saturation is when $e = e_s$

» **condensation** can start in air brought to saturation, if it has aerosol particles (sites for condensation)

- – There would be no condensation in pure air without aerosol (unless huge *e*)
- – Troposphere always has aerosols everywhere



Closed container in lab

Saturated vapour pressure, e_s

» Hypothetical value of *e* at which saturation would occur

» Function only of temperature:

$$\boxed{e_s(T) \approx A \exp(-B/T)}$$

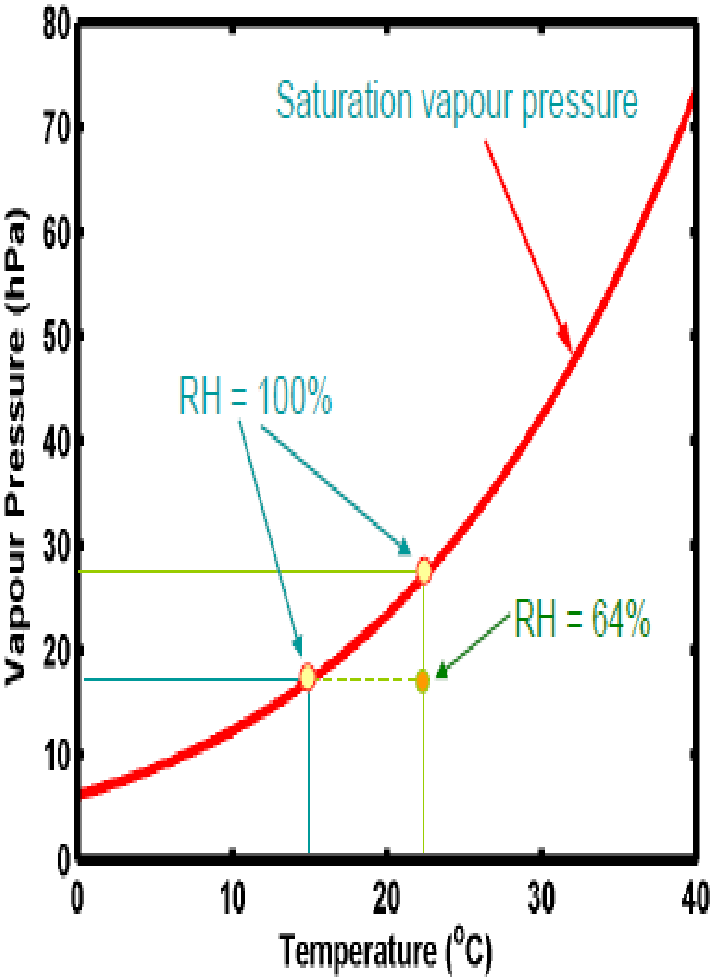
A = 2.53 10^{11} Pa and B = 5420 K, *T* is in K

» Relative humidity (RH) is most useful measure of moisture

– $RH = e / e_s$

» Same amount of water vapour in air will give different *RH* at different temperatures.

» *RH* governs condensation or evaporation, so affects how humid it **feels**



Saturation

» Saturation vapour pressure (w.r.t. liquid), $e_s(T)$, is hypothetical value of e when saturation is reached over liquid in closed container

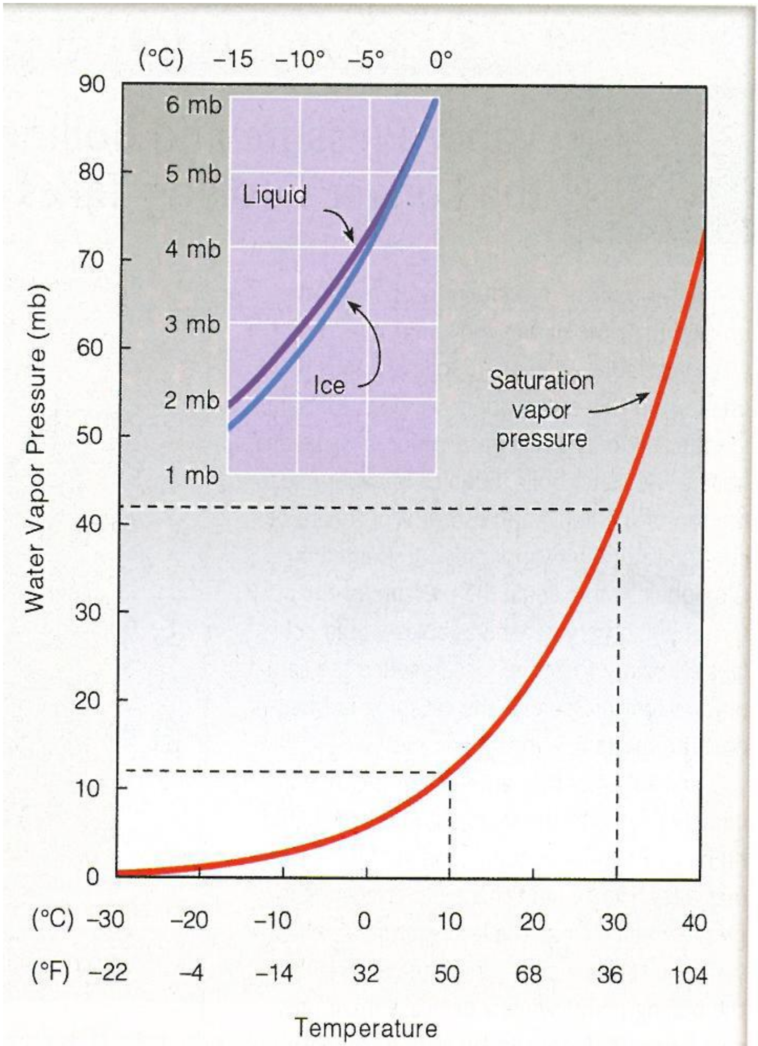
– Fluxes of evaporating and condensing molecules are equal

– Steady state

– sat. m.r.: $q_s = \epsilon \, e_s / p$

– Tephigram: contours of constant q_s are straight slanted lines; the one thru dew-point temp. (at same p) equals q

– Condensation on droplets if $e > e_s$



» 'sat. vap. pres. w. r. t. ice', $e_{s,i}(T)$, defined similarly except for ice in container

Condensation and evaporation for liquid drops in moist air

Condensation/evaporation

$$\frac{dr}{dt} \propto \frac{s_w}{r}$$

RH (%) = s_w + 100%

- At surface of drop, air is always saturated and vapour pressure is e_s
- In ambient air, any vapour pressure, e, is possible (related to RH)
- Vapour molecules diffuse in opposite direction to gradient of vapour density (or *e*)
- RH > 100 % (s_w > 0) in ambient air ➡ net condensation of vapour onto a liquid drop
- RH < 100 % (s_w < 0) in ambient air ➡ evaporation of drop

Distance drops fall before totally evaporating, below cloud (RH = 80%, 280 K)

Initial radius	Distance fallen
1 μm	2 μm
3 μm	0.17 mm
10 μm	2.1 cm
30 μm	1.69 m
0.1 mm	208 m
0.15 mm	1.05 km

Measures of humidity

- **(Saturation) Mixing Ratio** (kg/kg), q = Ratio of mass of water vapour to original mass of dry air, and if

$$\boxed{q_{(s)} = \frac{\epsilon e_{(s)}}{p - e_{(s)}} \approx \frac{\epsilon e_{(s)}}{p}}$$

$\epsilon = 0.622$

- **Vapour density** (kg m^{-3}), ρ_v = mass of water vapour per unit volume of moist air, such that:

$$e = \rho_v R_v T$$

$R_v = 461.5$ J kg^{-1} K^{-1} is individual gas constant for water vapour

- **Dew point depression** (^{o}C or K) = difference between temperature and dew point temperature, T_d
  - The smaller the depression, the more humid the air

$$\boxed{e_s(T_d) = e}$$

- **Specific Humidity** (kg/kg) ratio of mass of water vapour to total mass of moist air

- **Supersaturation**: percentage excess of vapour pressure beyond saturated value

$$s_w(\%) = 100\left(\frac{e}{e_s} - 1\right)$$

» Concept in cloud physics: moist air is regarded as mixture of dry air with water vapour

» Vapour mixing ratio = $q_v = \dfrac{\text{mass of vapour}}{\text{mass of dry air it is mixed with}}$

$\approx \rho_v / \rho = (e / R_v T)/(p/R' T) = \epsilon\, e/p$

$\boldsymbol{\epsilon = \dfrac{R'}{R_v} = 0.622}$

» Ideal gas law for dry air ($p = \rho R' T$) and for vapour cpt ($e = \rho_v R_v T$)

» Often we omit to write “vapour”, or omit the subscript ‘v’

– “mixing ratio” means the vapour mixing ratio

– Vapour density is ρ_v = mass of vapour per unit volume of air

» Cloud-liquid mixing ratio = $q_L = \dfrac{\text{mass of cloud}-\text{droplets}}{\text{mass of dry air}} = \dfrac{LWC}{\rho}$

» Liquid water content is LWC = mass of cloud-liquid per unit vol. of air

» Total water mixing ratio is $q_T = q_v + q_L$

Unsaturated yet moist adiabatic parcels

» During dry adiabatic ascent of a parcel:

- both the vapour mixing ratio and the potential temperature (q and θ) are constant

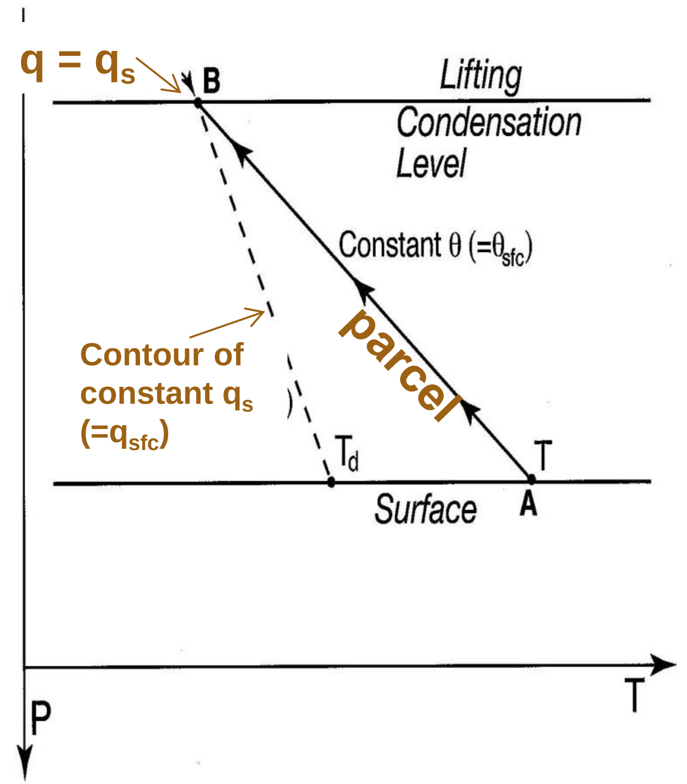
- temperature, saturated vapour pressure and saturated mixing ratio (T, $e_s = e_s(T)$ and q_s) all decrease.

» At the moment when saturation is just reached (e.g. LCL):

$$q_s(p, T) = q \Leftrightarrow T = T_c$$

ONSET OF CONDENSATION

» The temperature, T_c, is called the isentropic condensation temperature

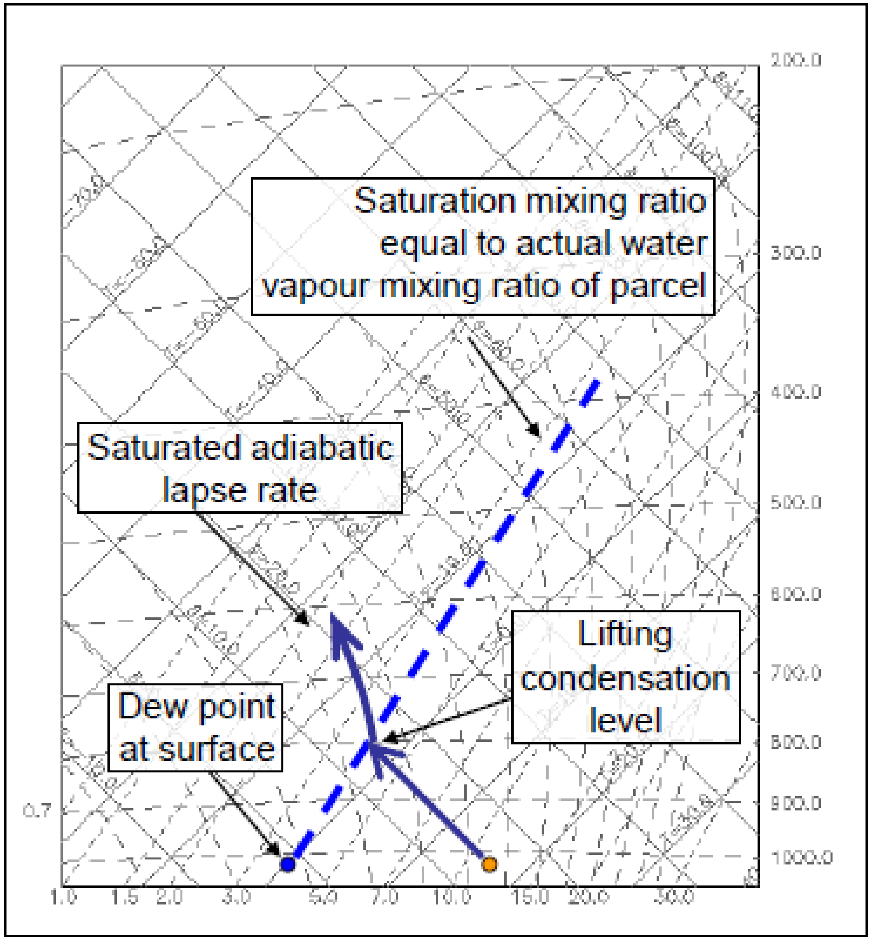


Influence from condensation on air parcel lifted from surface

pressure	Temp.	θ		Relative humidity (%)	Liquid water (g/kg)
700	-9	20		100	1.2
800	-3	15		100	-
900	7	15		60	-
1000	15	15	Air parcel	36	-

How the effect looks on tephigram:

- Initially, an adiabatic parcel is unsaturated (RH < 100%)
  - Pressure decreases and parcel expands, cooling
  - Parcel follows '**dry adiabatic lapse rate**', Γ_d, of 10 K/km
  - e_s decreases due to cooling, so RH increases
- When temperature reaches dew-point, there is **saturation**
  - $e = e_s$ and RH = 100% in parcel
  - Lifting condensation level (LCL)
- Ascent above LCL involves **condensation** onto cloud-droplets from aerosols
  - parcel follows '**saturated adiabatic lapse rate**', Γ_s,



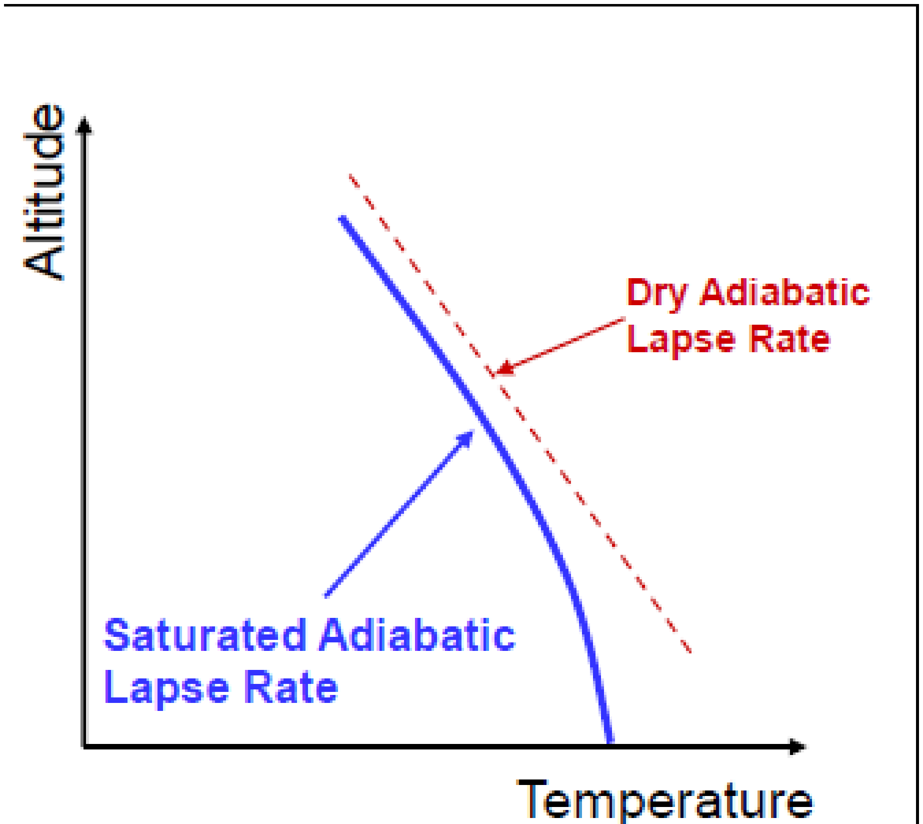
Saturated adiabatic lapse rate, Γ_s

» Condensation releases latent heat

- Ascending saturated air cools more slowly than if it were unsaturated

» **Saturated adiabatic lapse rate has no single fixed value**

- increases as temperature decreases, from 4 K /km for very warm tropical air to 9 K/km at -40ºC.



» Derived from 1st Law, as for Γ_d:

$$dQ = L\, dq_s = c_p\, dT - \alpha dp = c_p dT + g\, dz$$

$$\Rightarrow \Gamma_s = -dT/dz \approx \frac{\Gamma_d}{1+(L/c_p)(dq_s/dT)} < \Gamma_d$$

Saturated adiabatic parcels:

» During saturated ascent, T continues to decrease, and both q_v and $q_s(p, T)$ decrease together

$$q_v \approx q_s(p, T)$$

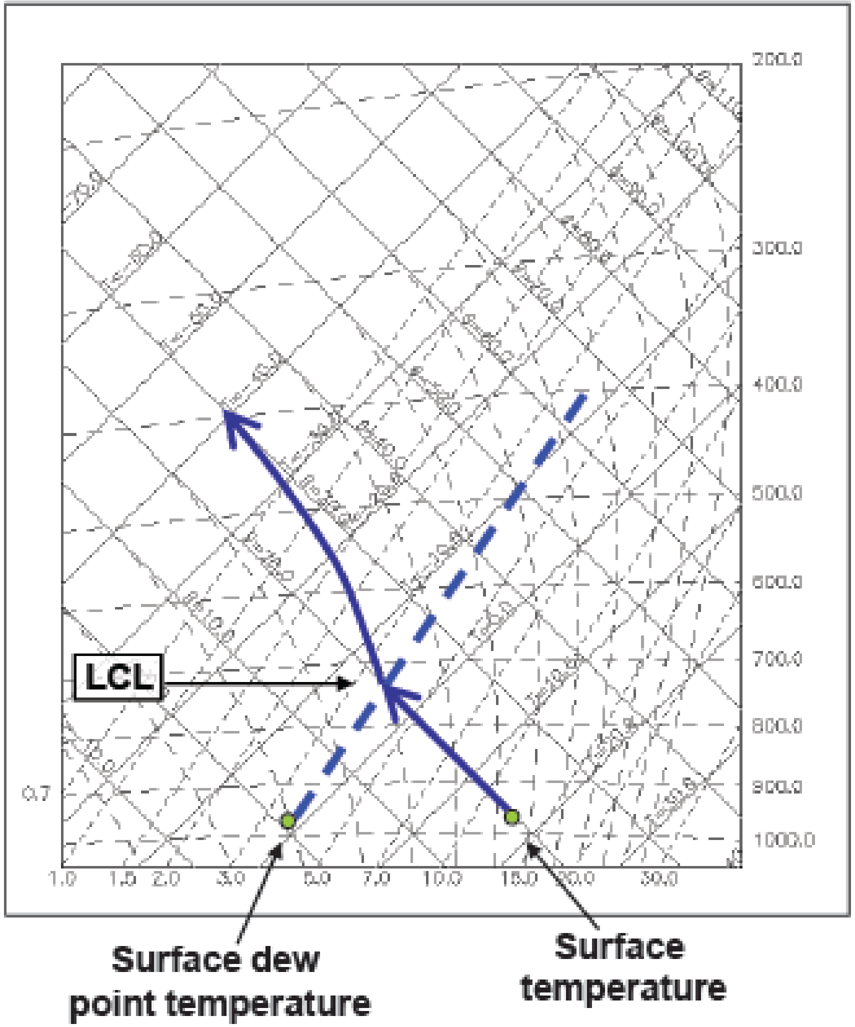
» As adiabatic parcel is closed, the total water mixing ratio, $q_T = q_v + \chi$ is constant, so cloud-water mixing ratio, χ, increases steadily with height from condensation

$$q_v + \chi = \text{constant}$$

during SATURATED ASCENT

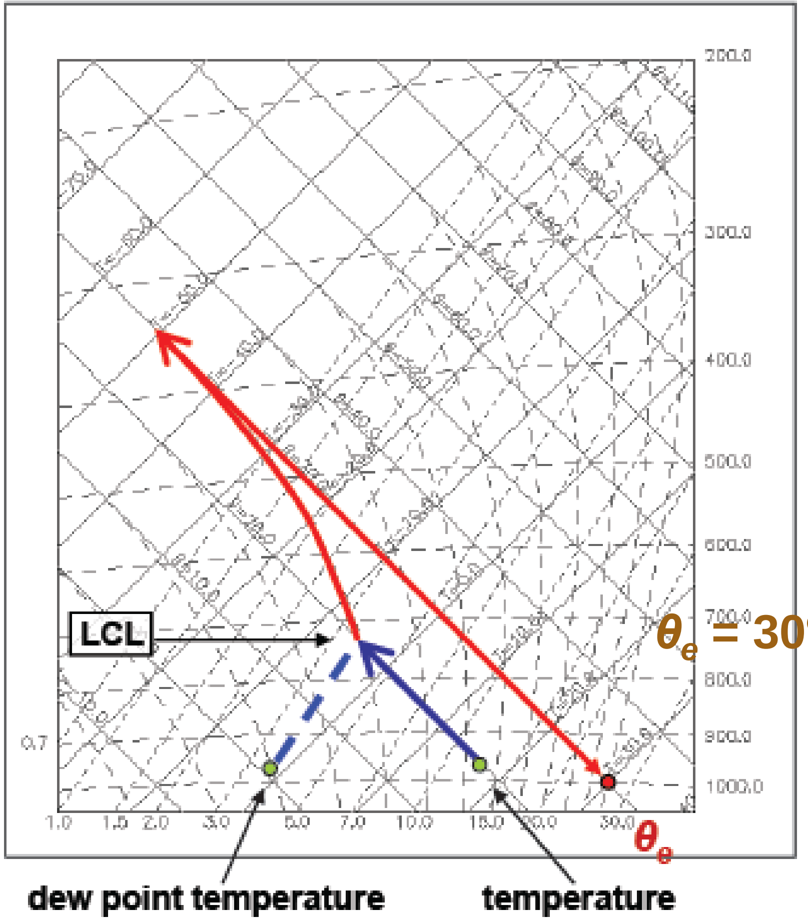
Deducing cloud-base from surface observations using a tephigram

- Follow the dry adiabatic lapse rate from the surface temperature and the constant humidity line from the dew point temperature,
  - lifting condensation level (LCL) is at their intersection.
- This is a *Normand construction*.
- the base of a convective cloud is always close to the LCL



Equivalent potential temperature

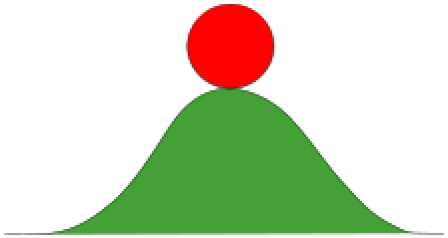
- A metric of temperature representing the possible effects of latent heating if the parcel of moist air were to move up and down
  - Unsaturated ascent to LCL
  - Ascent along saturated adiabat to such cold temperatures all moisture condenses (and removed)
  - θ_e is final temperature of parcel after descent by a dry adiabat to bring parcel back to 1000 mb.

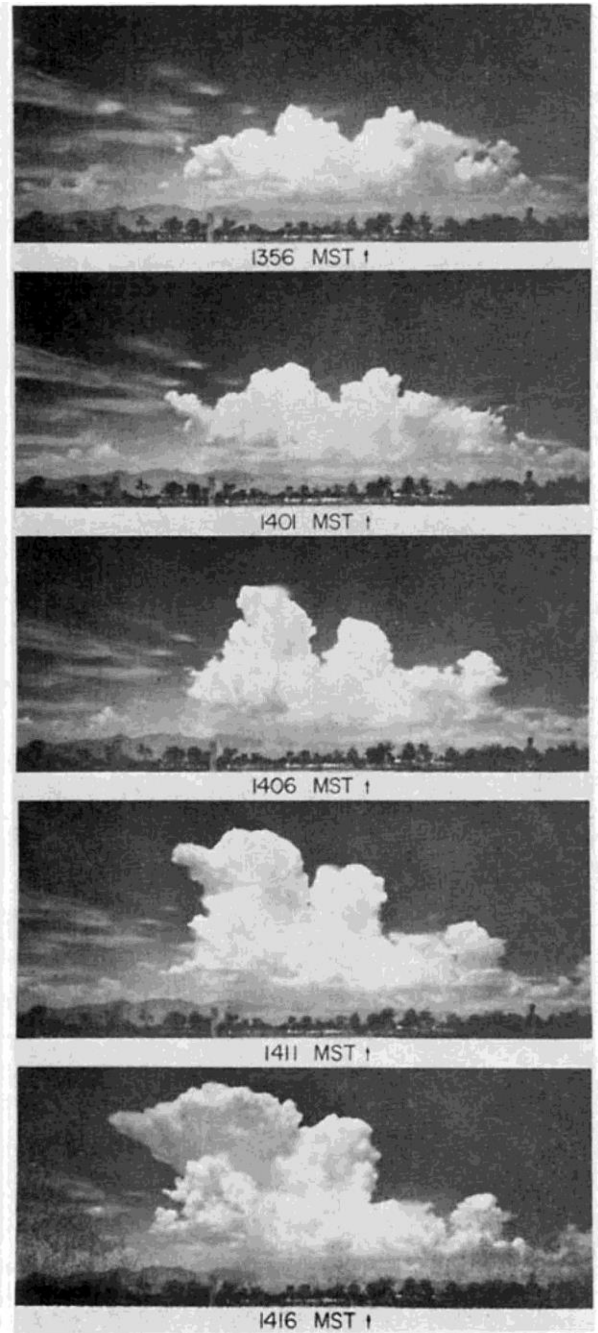


For saturated air ($q_v = q_s(p,T)$):

$$\theta_e = \theta \exp\left[Lq_v/c_pT\right] \approx \theta \times (1 + Lq_v/c_pT)$$

Replace T by T_c (LCL) for unsaturated air.

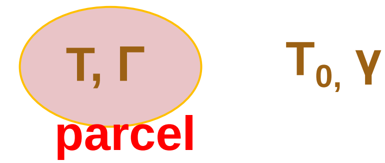




STABILITY

Vertical profile of environmental temperature (and surface moisture) determines lifted parcel’s buoyancy and stability

Lapse rates and convection

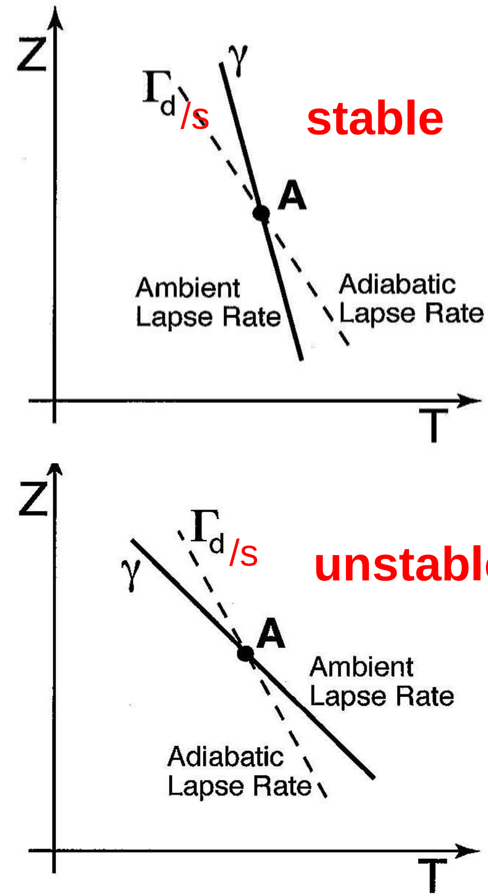


- Lapse rate of environment: $\gamma = -dT_0/dz$

- Parcel initially at rest ($F_B = 0$, at A) is displaced vertically by δz

- Convection starts immediately ($F_B\,\delta z > 0$) if γ exceeds the lapse rate in parcel, which is Γ_s (parcel always saturated) or Γ_d (parcel always unsaturated)
  - $\Gamma_s = -dT/dz =$ 3-7 K/km: convective clouds
  - $\Gamma_d = -dT/dz =$ 10 K/km: dry convection

- Any convection acts to reduce (super-critical) γ by vertical heat transfer: $\boxed{\gamma \to \Gamma}$



$$F_B \approx \frac{g(T-T_0)}{T_0} = -g\delta z\frac{(\Gamma_d-\gamma)}{T_0}$$

Absolute stability

» Whether parcel is lifted a little or very far, whether it becomes saturated or not, it is always colder (more dense) than environment

– Parcel returns to initial level of neutral buoyancy, as $F_B\ \delta z < 0$

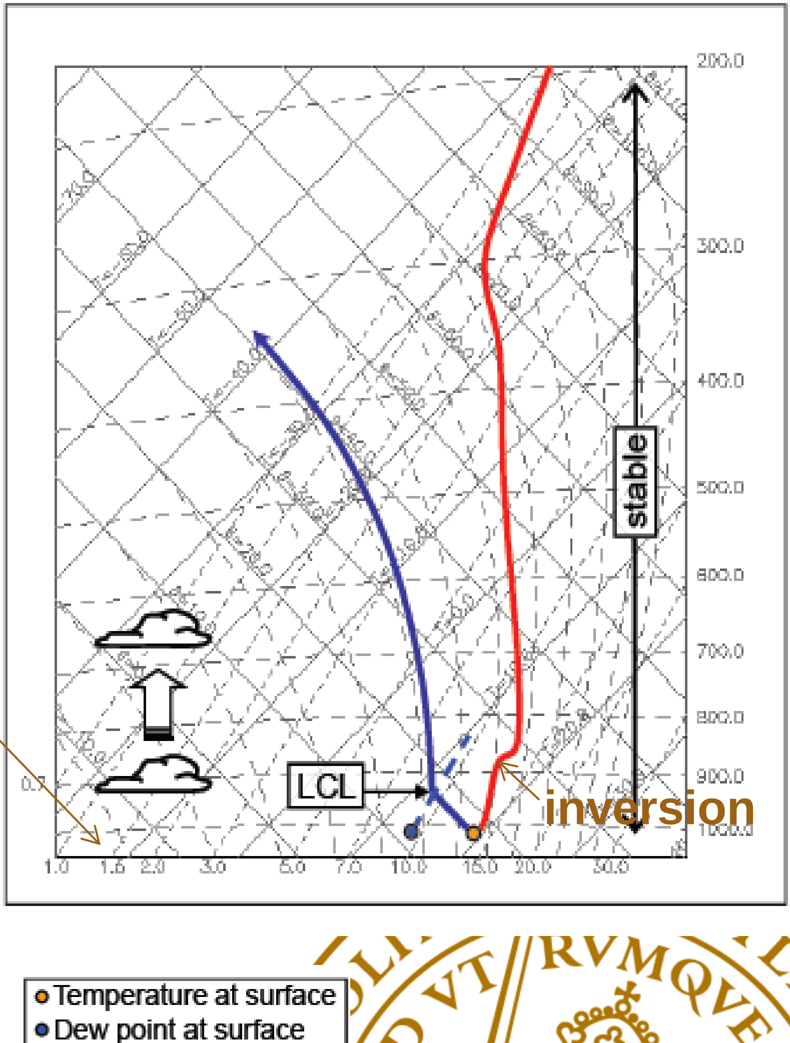
» Any lifting must be forced, needing energy (e.g. ‘**forced convection’)**

– E.g. from mechanical mixing of stable air in strong winds.

» Cloud forms if air lifted above LCL, but is limited to extent of lifting of parcel.

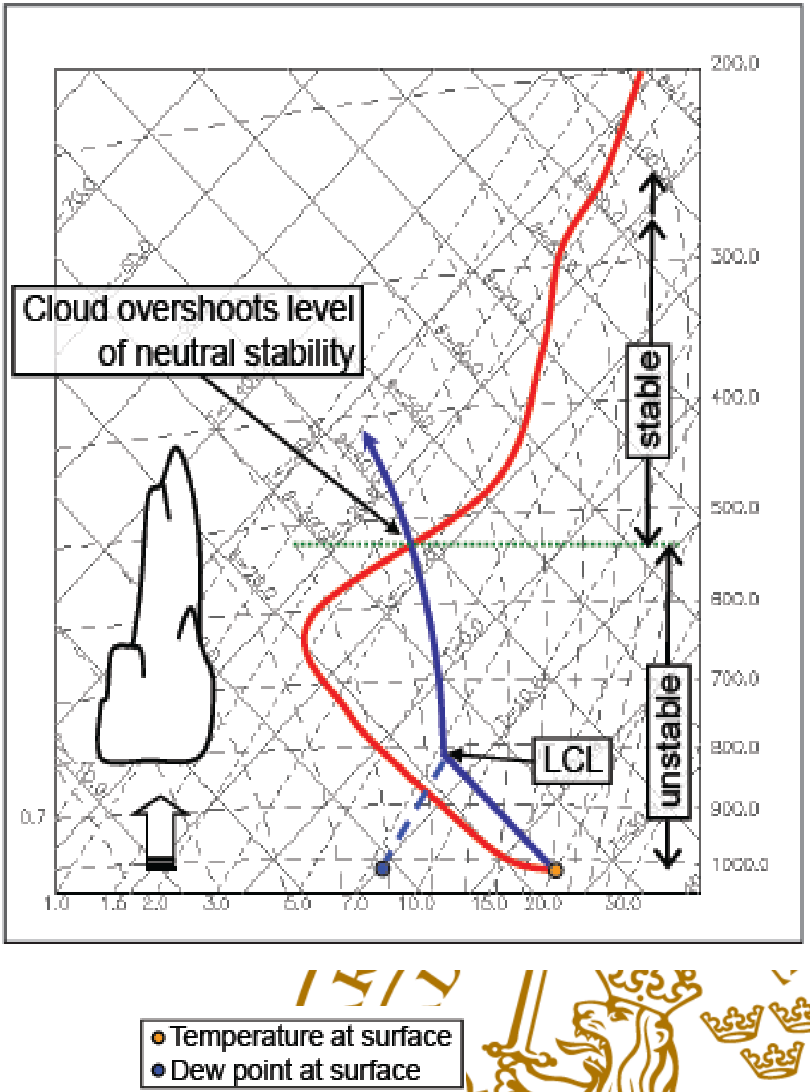
» ‘Inversion’ is where $dT/dz \gg 0$

– stops ascent of buoyant parcels from below, capping convection beneath it.



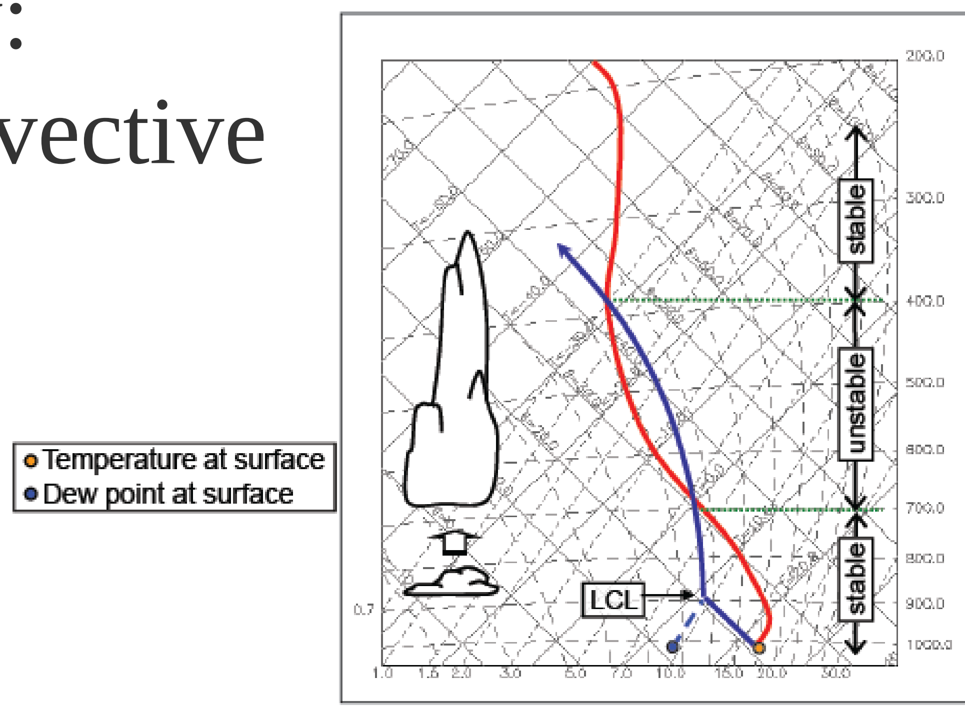
Absolute instability

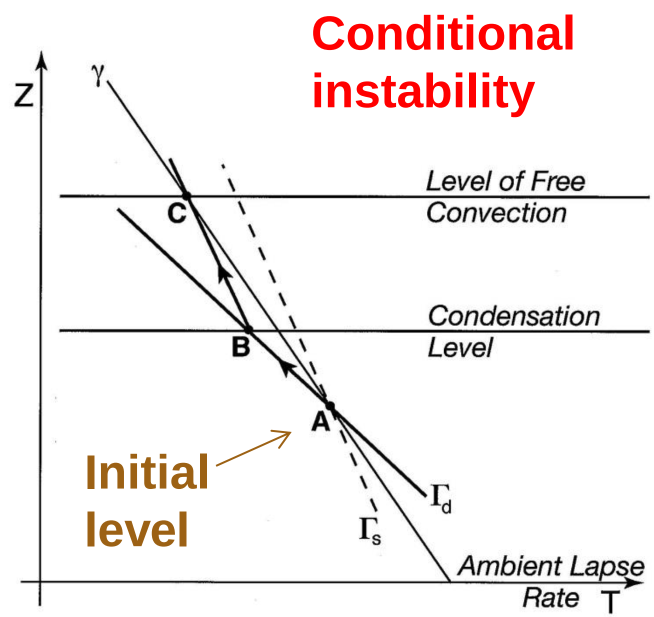
- Whether parcel is lifted a little or very far, it is always warmer (less dense) than environment
  - Parcel tends to move away from initial level, since $F_B\ \delta z > 0$
- Buoyancy force increases above LCL due to warming from release of latent heat from condensation.
- Intense insolation of surface, or advection over a warm surface can form unstable conditions in boundary layer (e.g. *Cu*), causing convection



Conditional instability: the usual cause of convective clouds

- During forced lifting of a parcel through region of ‘static stability’, parcel is initially colder (denser) than environment ($F_B\,\delta z < 0$; line ABC)
  - eventually it becomes warmer (less dense) due to condensation above LCL and LFC
  - Parcel then ascends ($F_B\,\delta z > 0$) and becomes buoyantly unstable





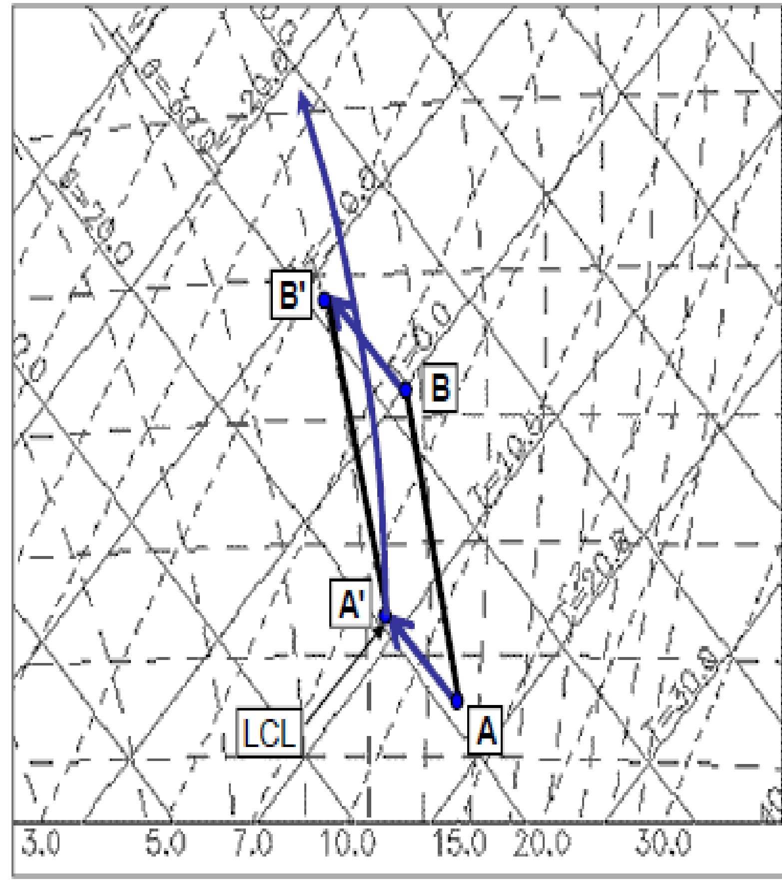
Convective instability of an entire atmospheric layer

- column of air A-B has lapse rate less than the dry adiabatic lapse rate, and is stable.

- Forced lifting of column cools all of it. If the lower part of column reaches saturation [A'], it starts to cool at the saturated adiabatic lapse rate

  – if this is less than the lapse rate of the column A'-B‘, the column becomes unstable. This type of instability may occur during large scale lifting over frontal surfaces or flow over mountain ranges.

- If a layer will become conditionally (or absolutely) unstable during lifting to saturation and beyond, it is convectively unstable



CONVECTIVE INSTABLITY
⇔ $\partial\theta_e/\partial z < 0$ in a layer

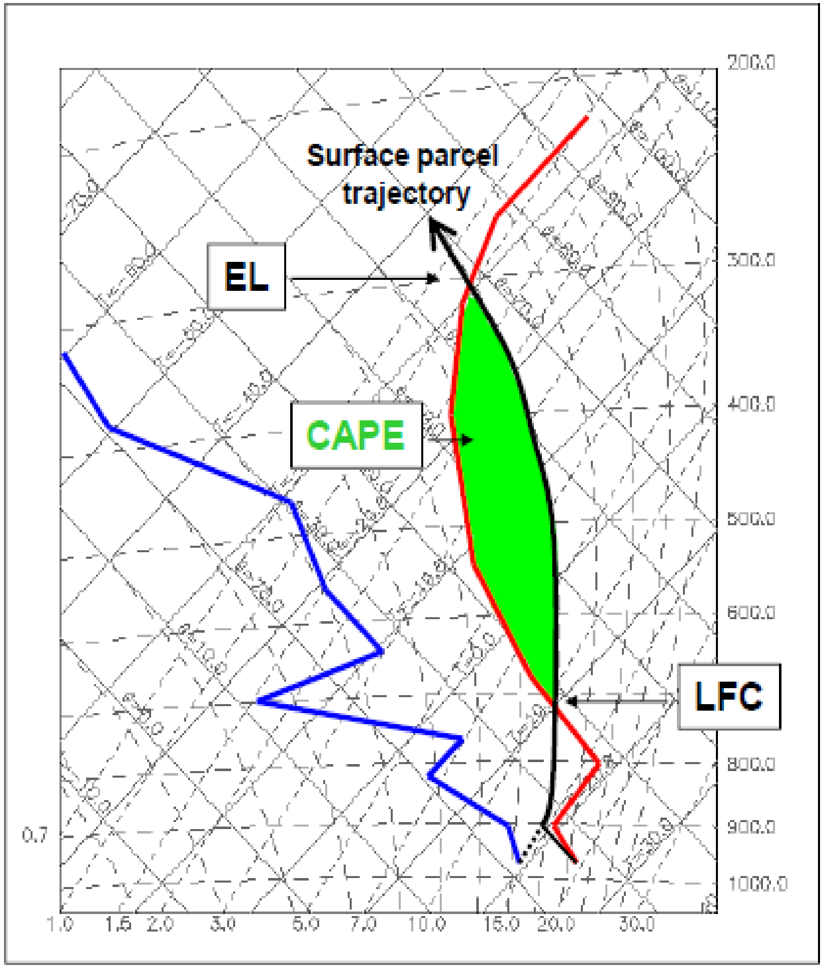
CAPE is a measure of intensity of vertical motions inside convective clouds

- **Convective Available Potential Energy (CAPE)** is work done by buoyancy force acting on a parcel (originally from surface) to lift it from one neutral buoyancy level (‘LFC’, near possible cloud-base) to another (‘EL’, at possible cloud-top), while it is positively buoyant

- **LFC – level of free convection** where a surface parcel becomes warmer than environment.

- **EL – equilibrium level** above LFC where parcel is same temperature as environment again.

- **Area of positive energy between parcel and environment curves from LFC to EL is (proportional to) CAPE**.



CAPE: measure of vigour of convection

- CAPE is energy available to near-surface parcels to convert to kinetic energy (vertical motion) once they have reached the LFC.

- The greater the CAPE, the more vigorous the convection and storms are likely to be

  - The most vigorous convection is usually over land, such as at incursion of warm humid air from the sea at low levels and dry, conditionally unstable, air at upper levels.

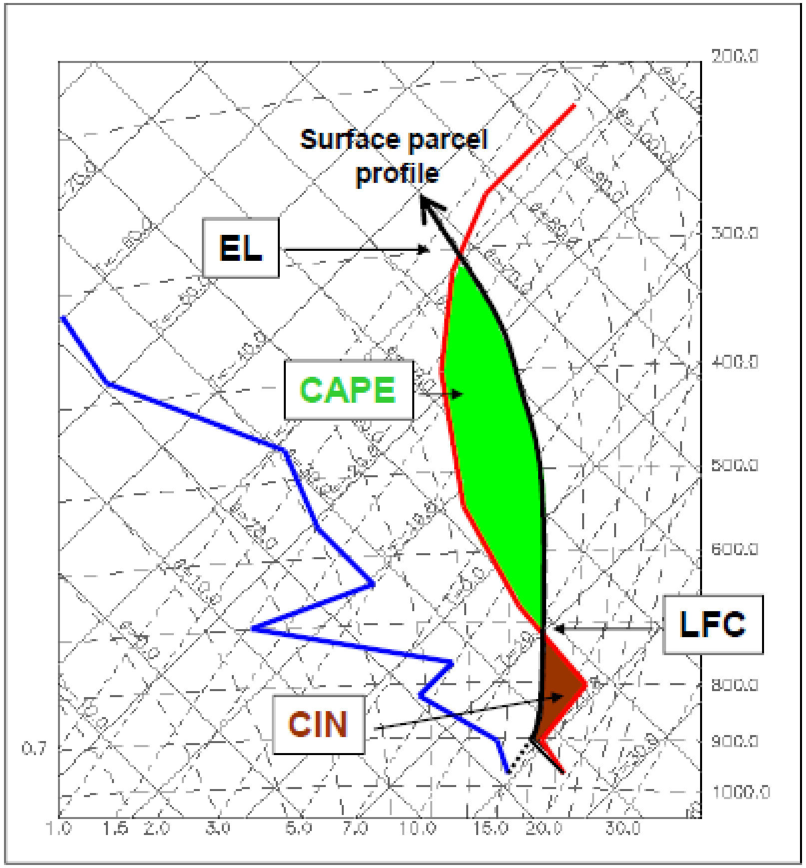
- But CAPE does not say whether any convection will occur !
  - For convection to occur, it needs to be triggered
  - Triggering performed by forced lifting of parcels from surface

- In tropics, lapse rate is conditionally unstable below ~15 km MSL and convectively unstable below 6 km MSL
  - Vertical lifting only sufficient to release instability over a few percent of area of tropics where clouds occur.

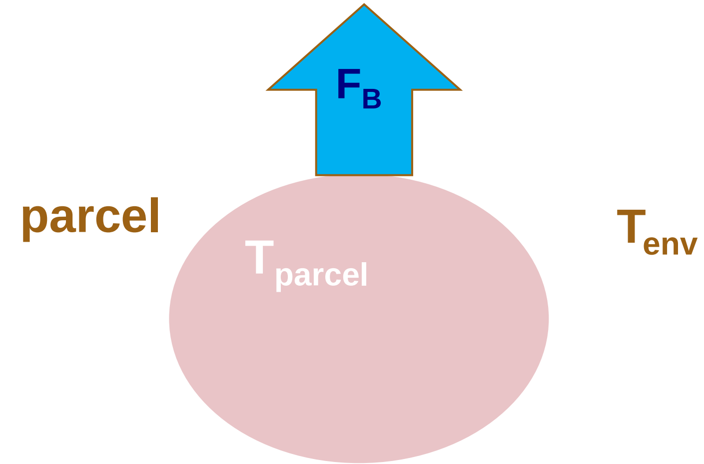
CIN is the strength of the **barrier**, which opposes the forced lifting needed to trigger the onset of any convection

CIN: barrier to onset of convection

- **Convective inhibition** (CIN) is the work done by lifting force (equal and opposite direction to buoyancy force) to raise parcel from surface to LFC where convection starts
  - area below the LFC where environment is warmer than the parcel
- CIN inhibits convection
  - Larger CIN, convection is less likely to occur.



CAPE and CIN



» work = (ave.) force × distance

» CAPE = work done on parcel by buoyancy force (per kg of air in parcel), raising it from LFC to EL

$$= \int_{LFC}^{EL} F_B \, dz$$

= gain of kinetic energy of parcel (per kg) in ascent

$$\approx \tfrac{1}{2} w_{max}^2$$

$$CAPE = g \int_{LFC}^{EL} \frac{T_{parcel} - T_{env}}{T_{env}} dz$$

» CIN defined similarly to CAPE

– except × -1 and integrate from sfc to LFC for CIN

CIN (J/kg)	
Under 50	Weak
50-200	Moderate
Over 200	Strong

CAPE (J/kg)	
Under 500	Weak
500-1500	Moderate
1500-2500	Strong
Over 2500	Extreme

- Strong convection observed when CAPE = 1000 to 3000 J kg^{-1}
  - Prediction: CAPE of 2500 J/kg implies a peak vertical velocity at EL (near max. cloud-top level) of 70 m/s
    - Too fast by 100% due to vertical PPGF, weight of vapour (q_v) and condensate (q_H), mixing of environmental air, level of origin
- To account for weight of vapour, write CAPE using *virtual temperature* (subscript, '*v*') instead of *T*, so $F_B \approx g \dfrac{(T_{v,parcel} - T_{v,env})}{T_{v,env}}$
  - Virtual temperature, T_v, of moist air is temperature that dry air would have at the same density (ρ) and pressure (p)
  - $p = \rho R' T_v$ where R' is for dry air and $T_v \approx T(1 + 0.6 q_v)$
  - Similarly, virtual potential temperature accounts for weight of vapour $\theta_v \approx \theta(1 + 0.6q)$

$$F_B = g\left[\frac{(\theta_{v,parcel} - \theta_{v,env})}{\theta_{v,env}} - \mathrm{q_H}\right]$$

Criteria for instability

T, Γ T_0, γ

parcel

adiabatic parcel that may become saturated

» Absolute stability: $\gamma < \Gamma_s$

$\Leftrightarrow (N^2 >)\ N_s^2 > 0 \ \Leftrightarrow \partial\theta_e/\partial z > 0$

» Absolute instability: $\gamma > \Gamma_d$

$\Leftrightarrow (N_s^2 <)\ N^2 < 0 \ \Leftrightarrow \partial\theta/\partial z < 0$

» Conditional instability: $\Gamma_s < \gamma < \Gamma_d$

$\Leftrightarrow CAPE > 0 \Leftrightarrow N_s^2 < 0 < N^2$

$\Leftrightarrow d\theta_{es,\,env}/dz < 0$

$$\theta_{e,s} = \theta \exp\left[\frac{Lw_s(T)}{c_pT}\right]$$

ad. parcel always unsaturated

$$\frac{d^2(\delta z)}{dt^2} = F_B \approx \frac{g(T-T_0)}{T_0} = -g\delta z\frac{(\Gamma_d-\gamma)}{T_0} = -N^2\delta z$$

$$\frac{d\theta}{\theta} = \frac{dT}{T} - k\frac{dp}{p} \Rightarrow (\Gamma_d-\gamma)/T_0 = \frac{1}{\theta}\frac{\partial\theta}{\partial z} \Longrightarrow N^2 = \frac{g}{\theta_0}\frac{\partial\theta_0}{\partial z}$$

instability: $\gamma > \Gamma_d \Leftrightarrow N^2 < 0 \Leftrightarrow \partial\theta/\partial z < 0$

ad. parcel always saturated

$$\frac{d^2\delta z}{dt^2} = -N_s^2\,\delta z$$

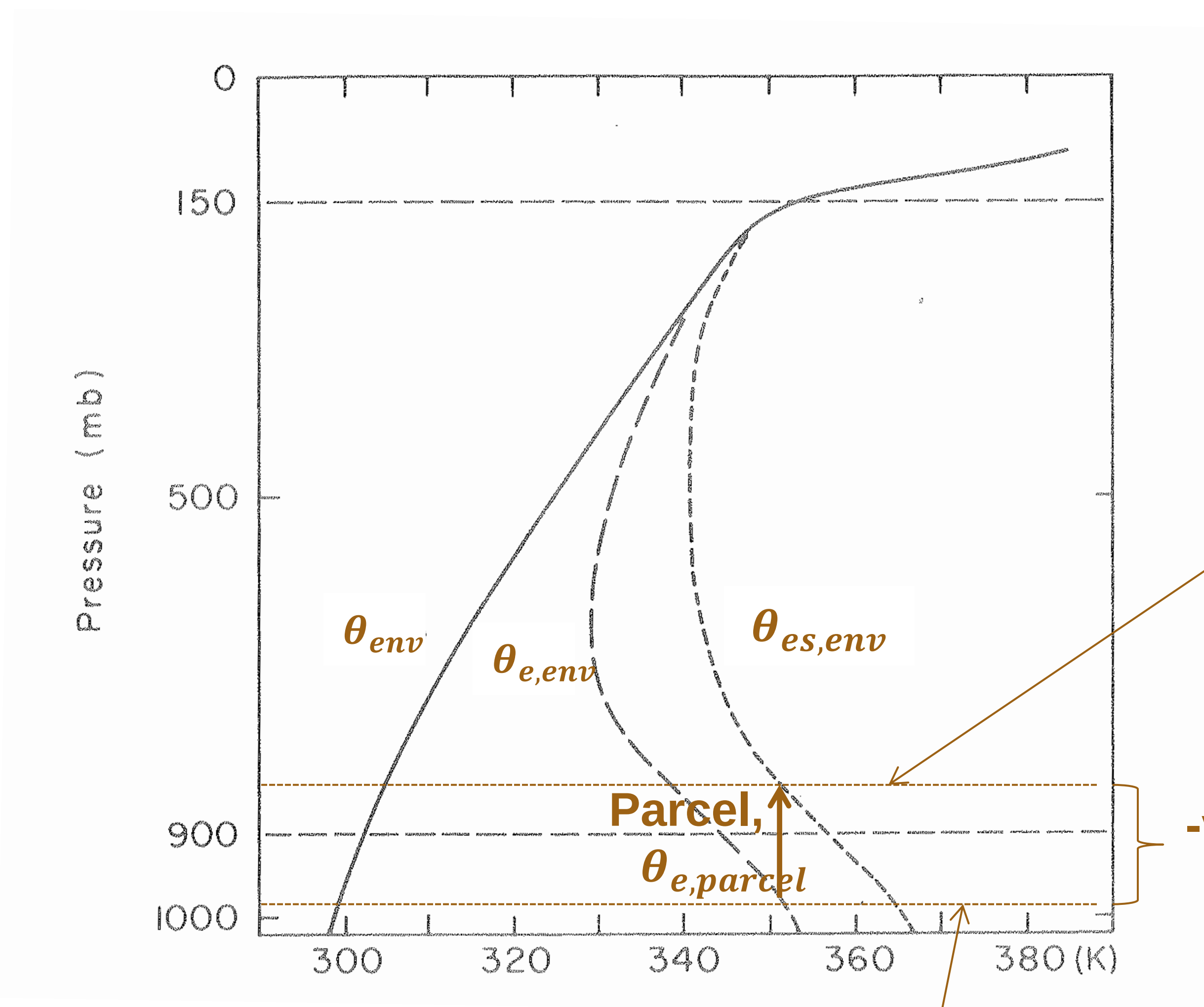
$$where\ N_s^2 \approx \frac{\Gamma_s}{\Gamma_d}\frac{g}{\theta_{e,0}}\frac{\partial\theta_{e,0}}{\partial z}$$

instability: $\gamma > \Gamma_s \Leftrightarrow N_s^2 < 0 \Leftrightarrow \partial\theta_e/\partial z < 0$

Hypothetical equivalent potential temperature of environment with same temperature profile as actual environment but saturated everywhere

OVERVIEW OF TRIGGERING MECHANISMS

Typical tropical sounding: LFC exists for parcel from near surface



In adiabatic parcel, its equivalent potential temperature is constant (whether saturated or not)

$$\theta_{e,parcel} = \theta_{e,env}(z_0)$$

LFC: where parcel reaches neutral buoyancy

$$T' = 0$$

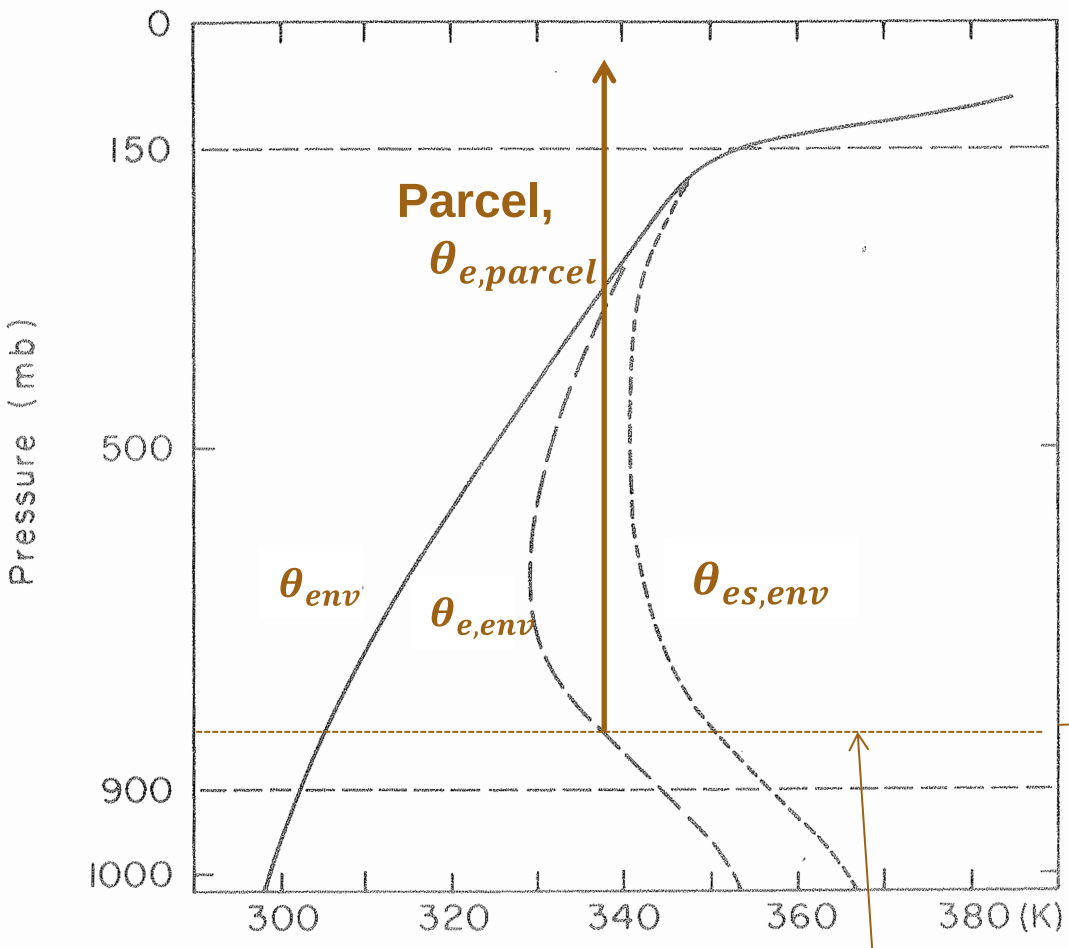
$$\Rightarrow \quad \theta_{e,parcel} = \theta_{es,env}$$

-vely buoyant parcel

$$T' < 0 \quad \Rightarrow \quad F_B < 0$$

Initial level, z_0, of adiabatic parcel

Typical tropical sounding: no LFC exists for another parcel starting higher



Parcel is always -vely buoyant

$$T' < 0 \implies F_B < 0$$

Initial level, z_0, of another adiabatic parcel

Triggers for Convective Clouds

- **Thunderstorms (*Cb*) form in regions of high CAPE, but require a '*trigger mechanism*' to overcome inhibition CIN in profile and lift parcels to LFC.**

- **Only parcels starting close enough to surface have LFC**
  - Most convective clouds have warm low bases (low LCL, LFC)
  - CAPE is defined by adiabatic parcel lifted from surface

- **Possible '*triggering mechanisms*' for thunderstorms, and other deep convection, usually involve forced lifting:**
  - Lifting by orography
  - land-sea breezes at coastlines
  - Gust fronts from one storm initiating a second storm
  - fronts
  - Surface heating or moistening
  - Gravity waves

SUMMARY

Summary

In today's lecture we covered:

- Diverse measures of moisture and heat content
- (equivalent) potential temperature and (total water) mixing ratio are constant in (saturated) unsaturated adiabatic parcels
  - Any ascent always causes cooling inside parcel (1^{st} Law)
  - Cooling slowed by latent heating (condensation) if saturated
- Criteria for buoyant instability were listed
  - Conditional instability (CAPE > 0) causes convective clouds
    - 'condition' = parcels must be lifted to saturation, to LFC
    - convective instability can create conditional instability

FURTHER READING

Highly recommended:

- **Rogers, R. R., and M. K. Yau, 1991: *A short course in cloud physics*. Pergamon Press**

- Wallace, J. M., and P. V. Hobbs, 1977: *Atmospheric science - an introductory survey.*

- Houze, R. A., 1993: *Cloud dynamics.* Academic Press

Optional extra reading:

- Barry, R. G., and R. J. Chorley, 1987: *Atmosphere, Weather and Climate.* Routledge

- Holton’s classic text-book

Questions ?

» Email: Vaughan.Phillips@nateko.lu.se

Extra info