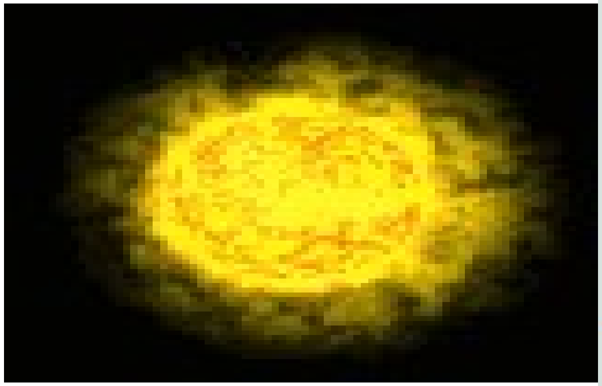
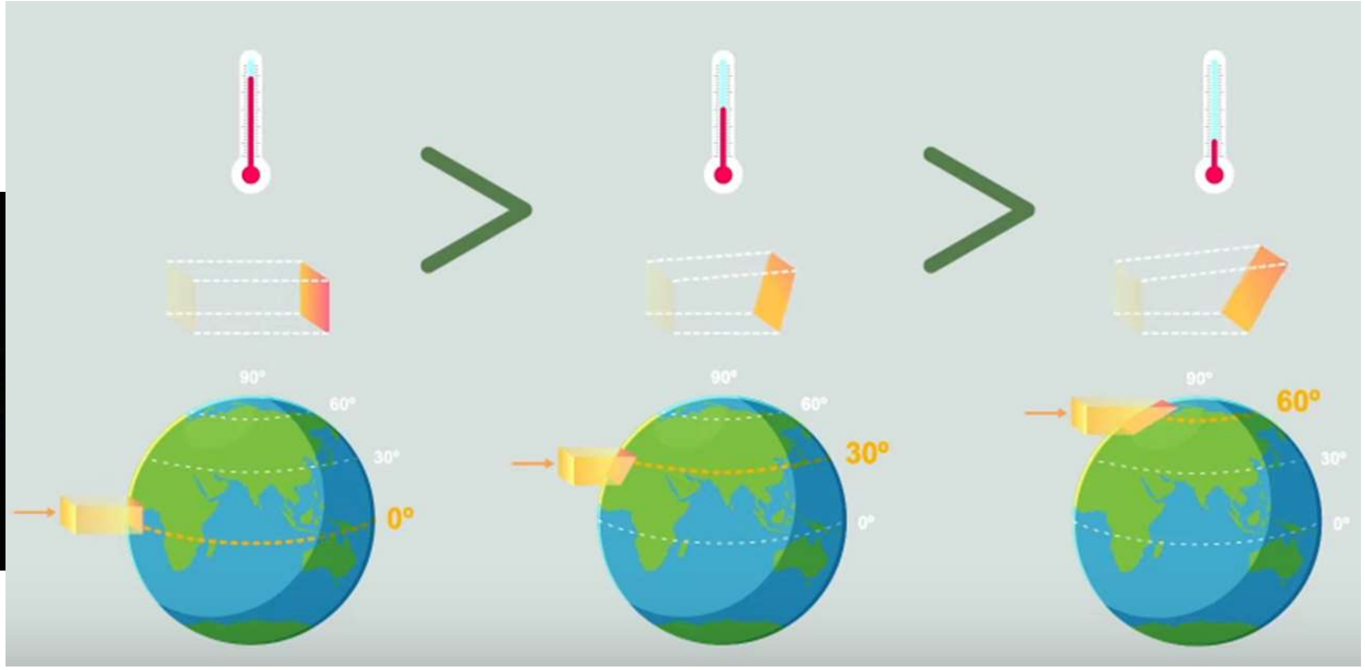
Source of HEAT

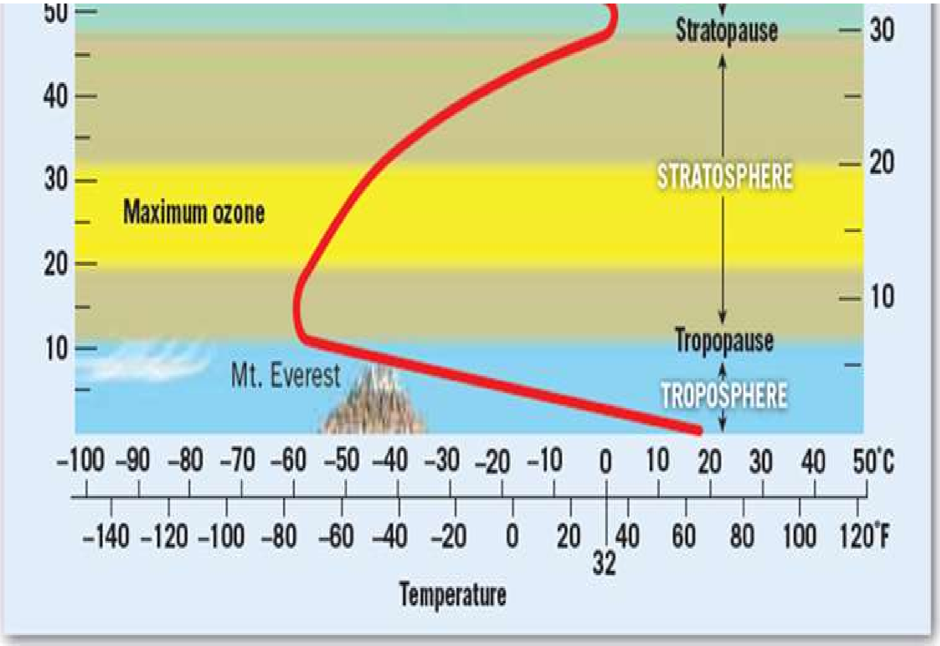
Solar Energy- source of HEAT for our planet





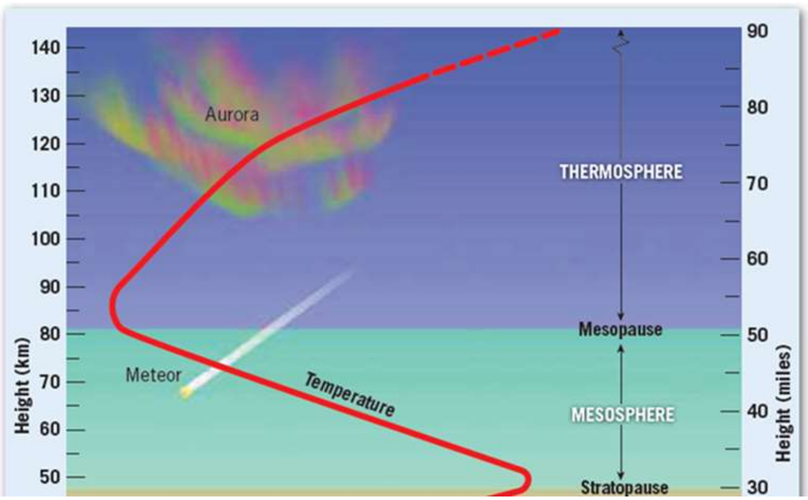
Stratosphere

- 15 to 50 km altitude
- Temperature increases with altitude because ozone (O3) absorbs ultraviolet radiation from the Sun and cause heating in the layer
- Very little vertical mixing occurs in the stratosphere. This is because the stratosphere experiences a temperature inversion, where cold air lies beneath warm air, in contrast to the opposite occurrence in the troposphere
- Top of the stratosphere is known as the stratopause



Mesosphere

- 50 to 90 km altitude
- Temperature decreases with altitude due to decrease in ozone, that leads to abundant vertical mixing
- Top of the mesosphere is known as the mesopause



Thermosphere

- 90 to 500 km altitude
- Contains only a tiny fraction of the atmosphere's mass.
- Temperature increases with altitude above 90 km, and is constant above 200 km.
- This heating is due to absorption of solar radiation (wavelengths less than 0.2 microns) by molecular oxygen and nitrogen
- The highest temperatures in the atmosphere can be found in the thermosphere, about 2000 K can occur.

Measure of relative concentration: mixing ratio (mole fraction) C_X [mol mol^{-1}]

$$C_X = \frac{\text{\# moles of X}}{\text{mole of air}}$$

remains constant when air density changes

$\Rightarrow$ stable measure of atmospheric composition

SPECIES	MIXING RATIO (dry air) [mol mol^{-1}]
Nitrogen (N_2)	0.78
Oxygen (O_2)	0.21
Argon (Ar)	0.0093
Carbon dioxide (CO_2)	400×10^{-6}
Neon (Ne)	18×10^{-6}
Ozone (O_3)	$(0.01\text{-}10) \times 10^{-6}$
Helium (He)	5.2×10^{-6}
Methane (CH_4)	1.8×10^{-6}
Krypton (Kr)	1.1×10^{-6}

Trace gases

Air also contains variable H_2O vapor (10^{-6}-10^{-2} mol mol^{-1}), aerosol particles, and many trace gases at < 1×10^{-6} mol mol^{-1}

Trace gas concentration units:

1 ppm = 1 μmol mol^{-1} = 1×10^{-6} mol mol^{-1}

1 ppb = 1 nmol mol^{-1} = 1×10^{-9} mol mol^{-1}

1 ppt = 1 pmol mol^{-1} = 1×10^{-12} mol mol^{-1}

Conversion from Mixing Ratio to $\mu g\, m^{-3}$

Atmospheric species mass concentrations are sometimes expressed in terms of mass per volume, most frequently as $\mu g\, m^{-3}$. Given a concentration m_i, in $\mu g\, m^{-3}$, the molar concentration of species i, in $mol\, m^{-3}$, is

$$c_i = \frac{10^{-6}\, m_i}{M_i}$$

where M_i is the molecular weight of species i.

Noting that the total molar concentration of air at pressure p and temperature T is $c = p/RT$, then

$$\text{Mixing ratio of } i \text{ in ppm} = \frac{RT}{p M_i} \times \text{concentration of } i \text{ in } \mu g\, m^{-3}$$

If T is in K and p in Pa (see Table A.6 for the value of the molar gas constant, R)

$$\boxed{\text{Mixing ratio of } i \text{ in ppm} = \frac{8.314\, T}{p M_i} \times \text{concentration of } i \text{ in } \mu g\, m^{-3}}$$

As an example, let us determine the concentration in $\mu g\, m^{-3}$ for O_3 corresponding to a mixing ratio of 120 ppb at $p = 1$ atm and $T = 298$ K. The 120 ppb corresponds to 0.12 ppm. Then

$$\begin{aligned}\text{Concentration in } \mu g\, m^{-3} &= \frac{p M_i}{8.314\, T} \times \text{Mixing ratio in ppm} \\ &= \frac{(1.013 \times 10^5)(48)}{8.314(298)} \times 0.12 \\ &= 235.6\ \mu g\, m^{-3}\end{aligned}$$

Now let us convert 365 $\mu g\, m^{-3}$ of SO_2 to ppm at the same temperature and pressure:

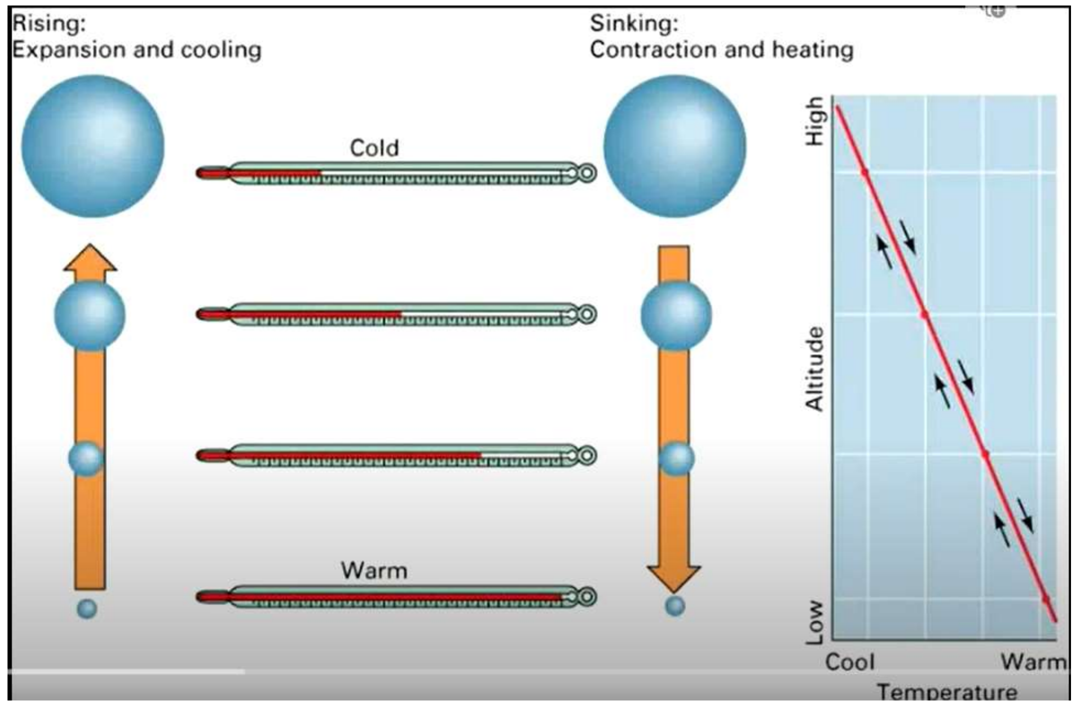
$$\begin{aligned}\text{Mixing ratio in ppm} &= \frac{(8.314)(298)}{(1.013 \times 10^5)(64)} \times 365 \\ &= 0.136\ \text{ppm}\end{aligned}$$

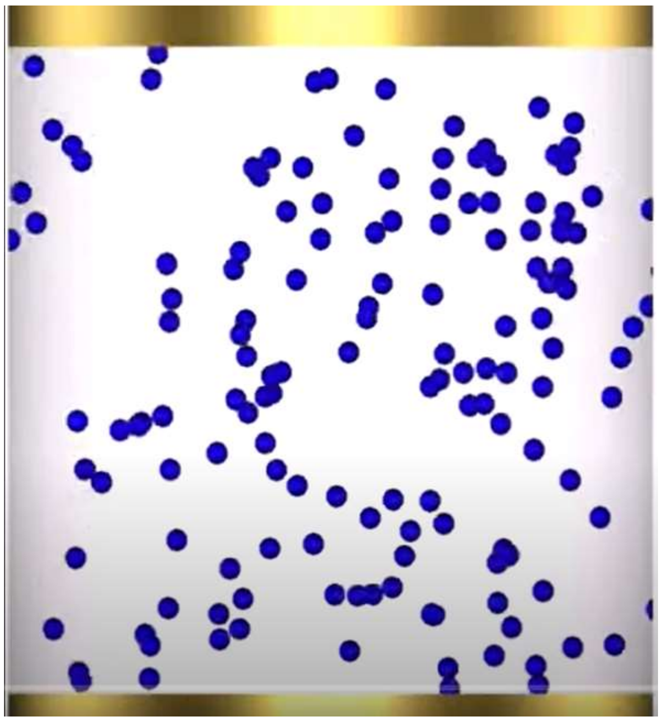
Units	Numerical Value
L-atm/mol-K	0.08206
J/mol-K*	8.314
cal/mol-K	1.987
m^3-Pa/mol-K*	8.314

1.15 Common Units of Pressure

Unit	Average Air Pressure at Sea Level
pascal (Pa)	101,325
kilopascal (kPa)	101.325
atmosphere (atm)	1 (exactly)
millimeters of mercury (mm Hg)	760 (exactly)
inches of mercury (in Hg)	29.92
torr (torr)	760 (exactly)
pounds per square inch (psi, lbs./in^2)	14.7

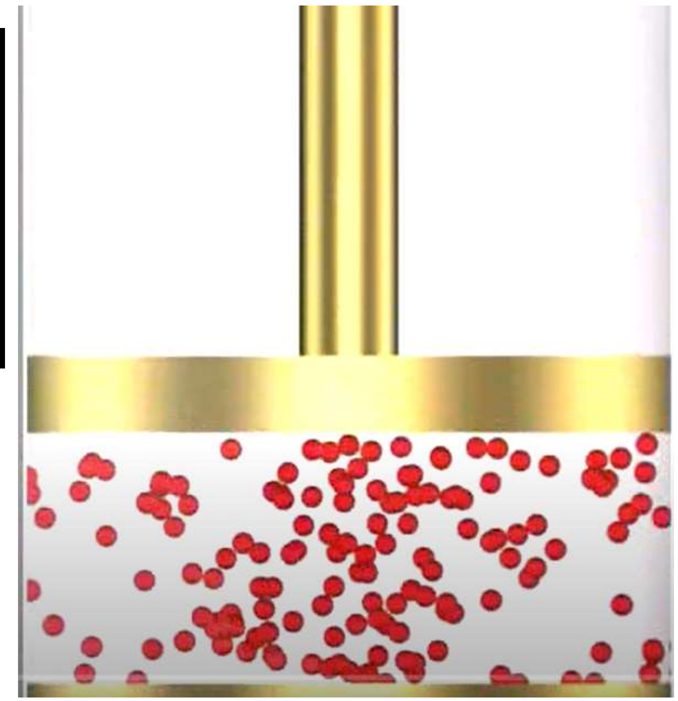
Temperature of an Air Parcel in the atmosphere changes via. Adiabatic expansion/contraction





Adiabatic Processes = heating by compression or cooling by expansion

Pressure increase results in more Kinetic energy that results in more temperature

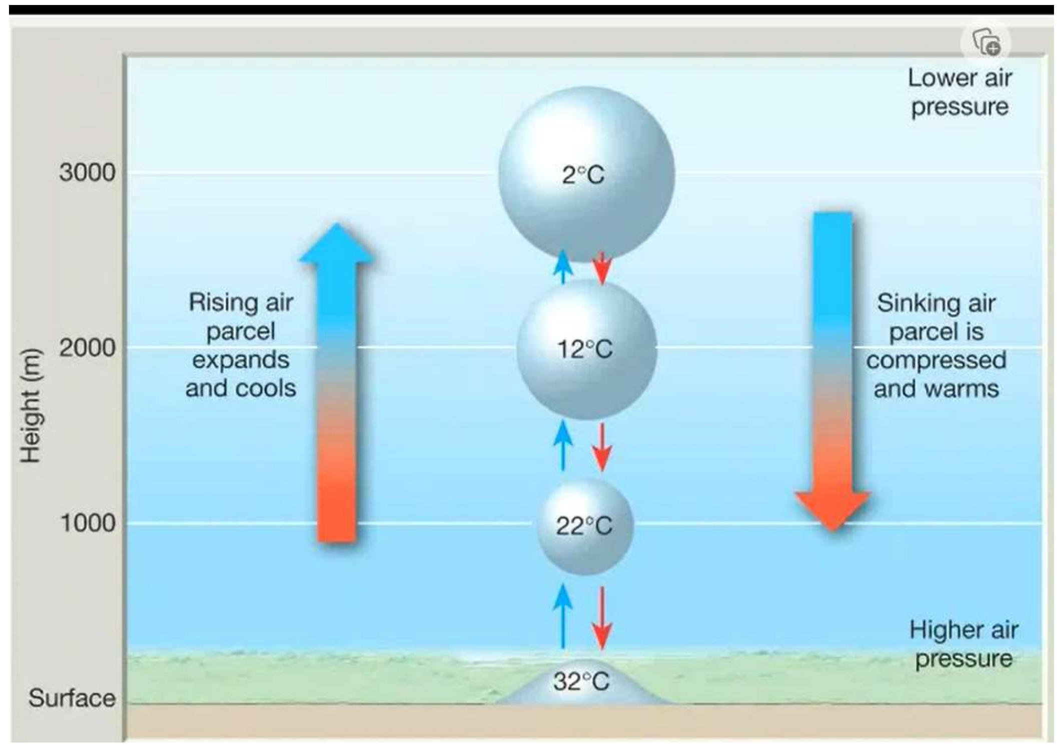


Dry Adiabatic Lapse Rate

Lapse rate means how much gases expand as a function of temperature (or can be said how temperature changes with altitude)

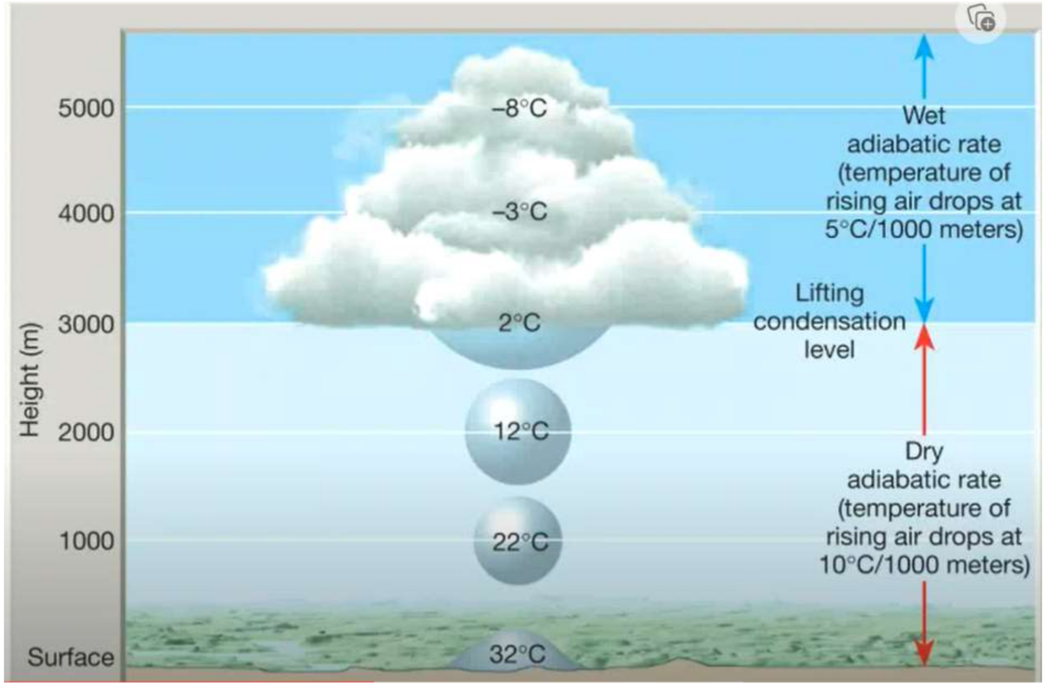
Units of lapse rate °C per 1 km

Dry Adiabatic Lapse Rate = 10°/1,000 m



MOIST/WET/SATURATED Adiabatic lapse rate

- If the air parcel is lifted high enough it gets colder and reaches a saturation level meaning no longer it can hold the water vapor in gaseous state and thus condensation happens (gaseous water vapor to liquid)
- The process of condensation happens when temperature reaches to dew point temperature
- The condensation mechanism releases energy, therefore the rate of cooling decreases (4 to 9 °C per Km, depending on water vapor content of air parcel)

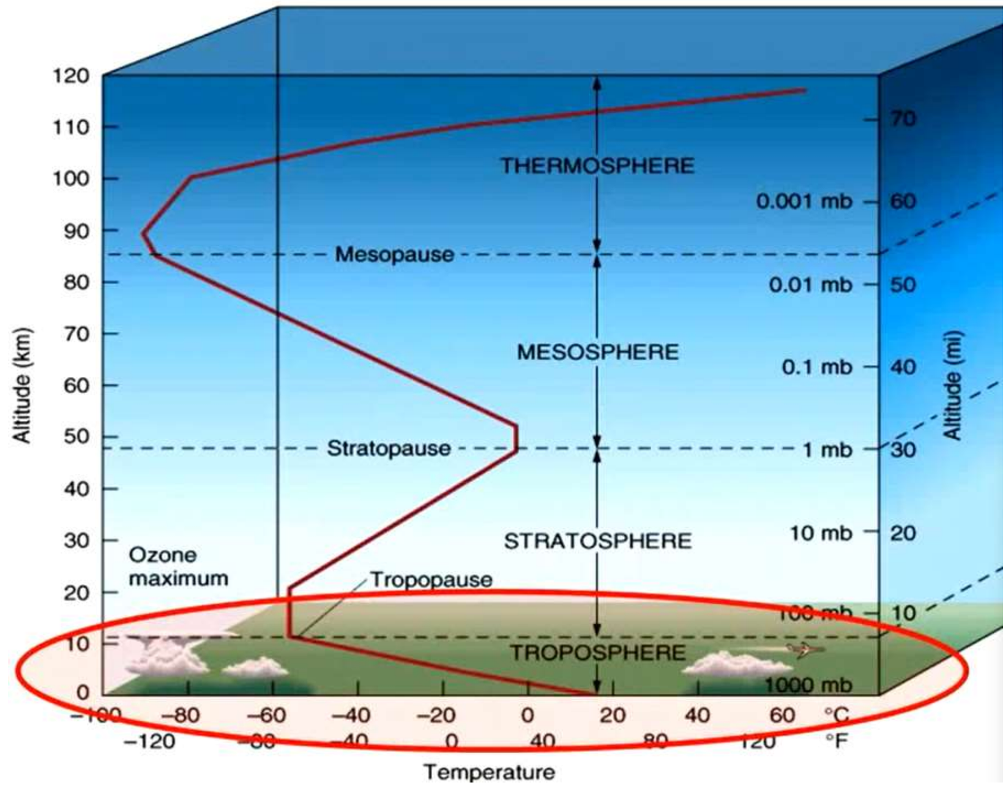


Adiabatic lapse rate is different than Environmental Lapse Rate

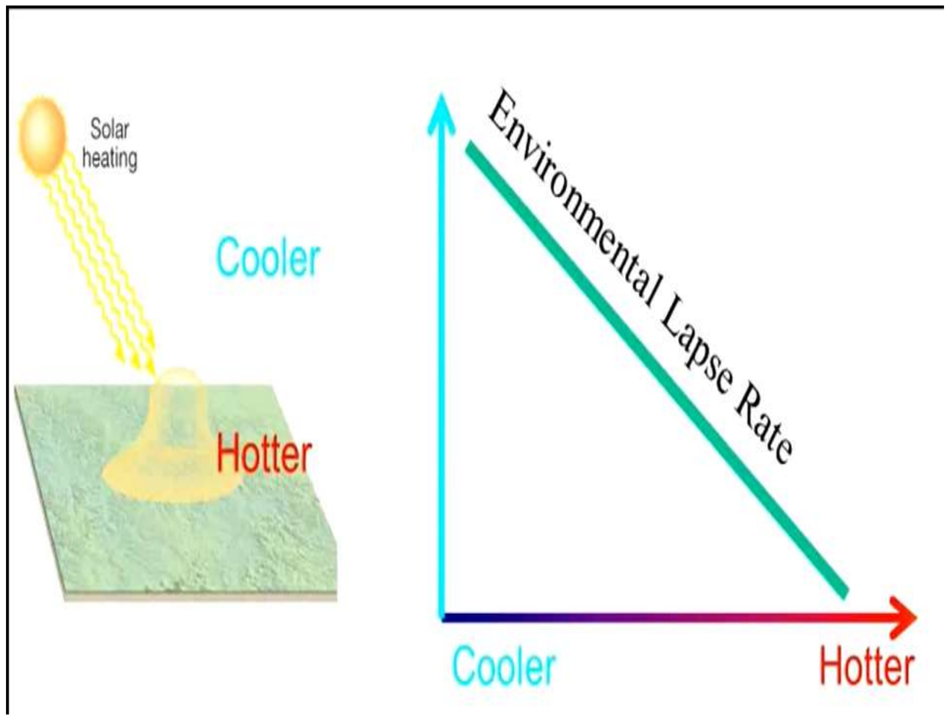
The change in atmospheric temperature with altitude is termed as Environmental Lapse Rate (ELR)

Environmental Lapse Rate (ELR)

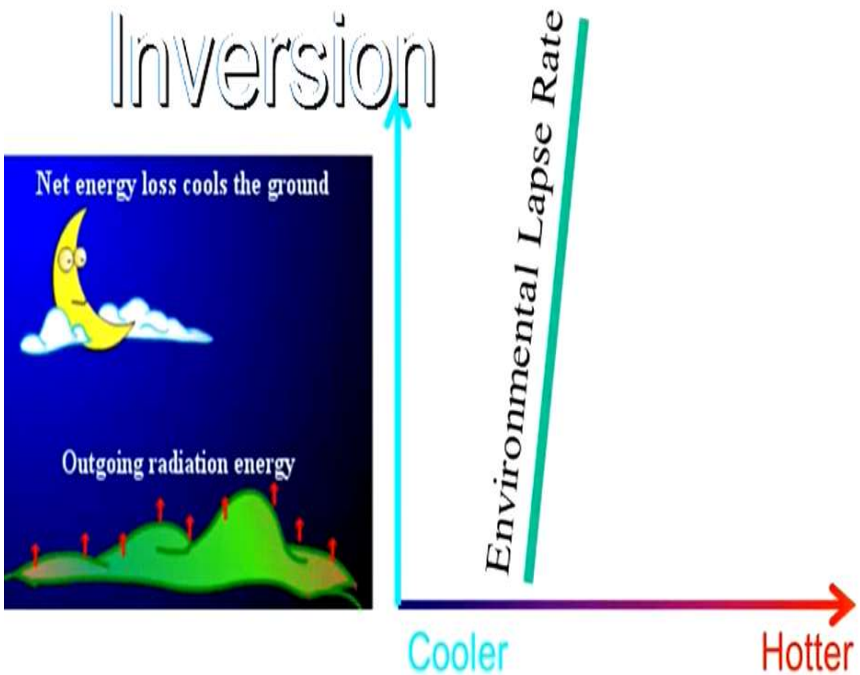
Temperature difference per 1000 m altitude difference



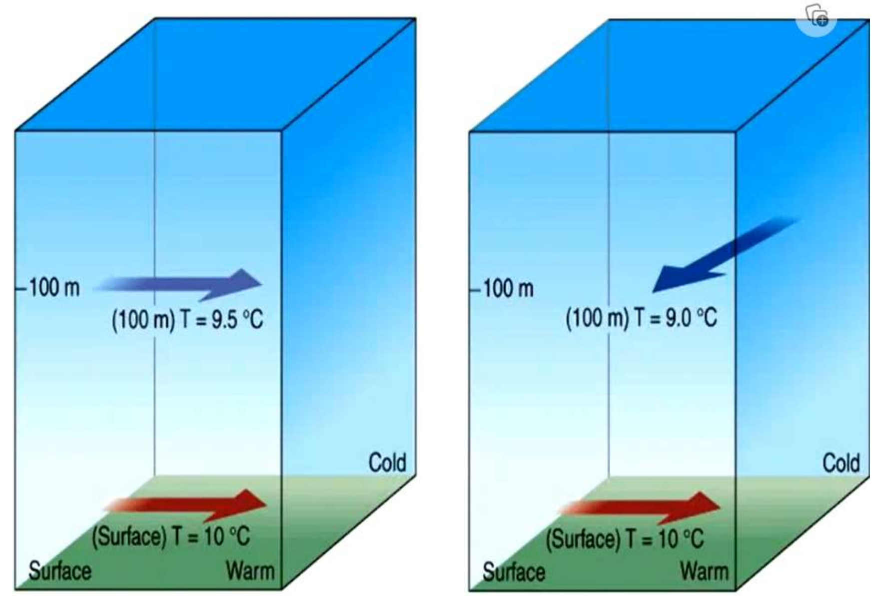
Factors affecting ELR



Factors affecting ELR

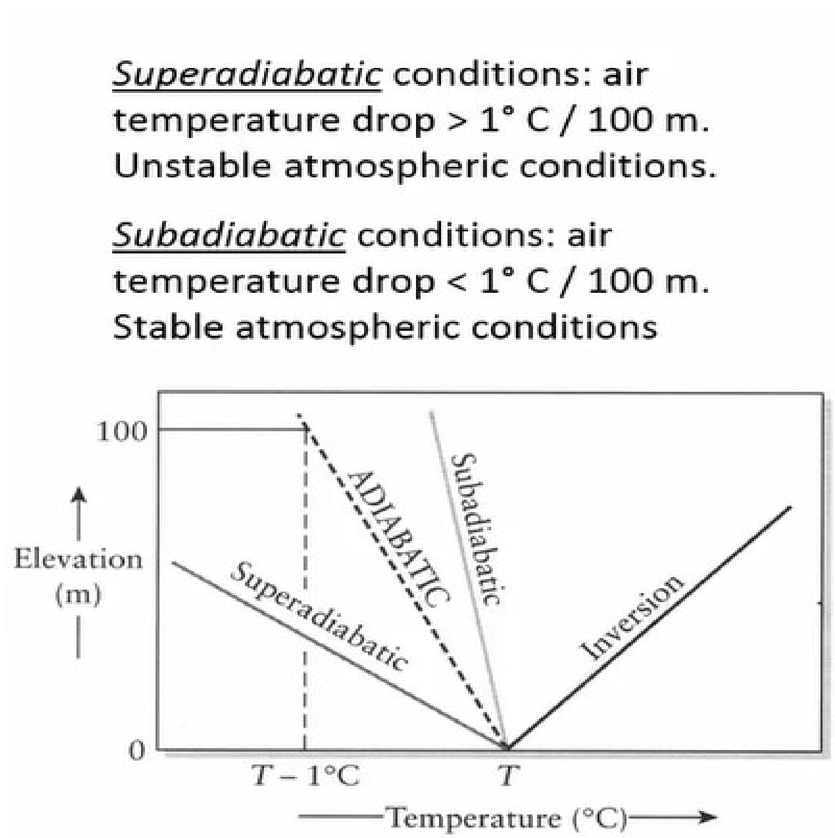


Hot air mass sits above the cold air mass
(mostly happens in wintertime)



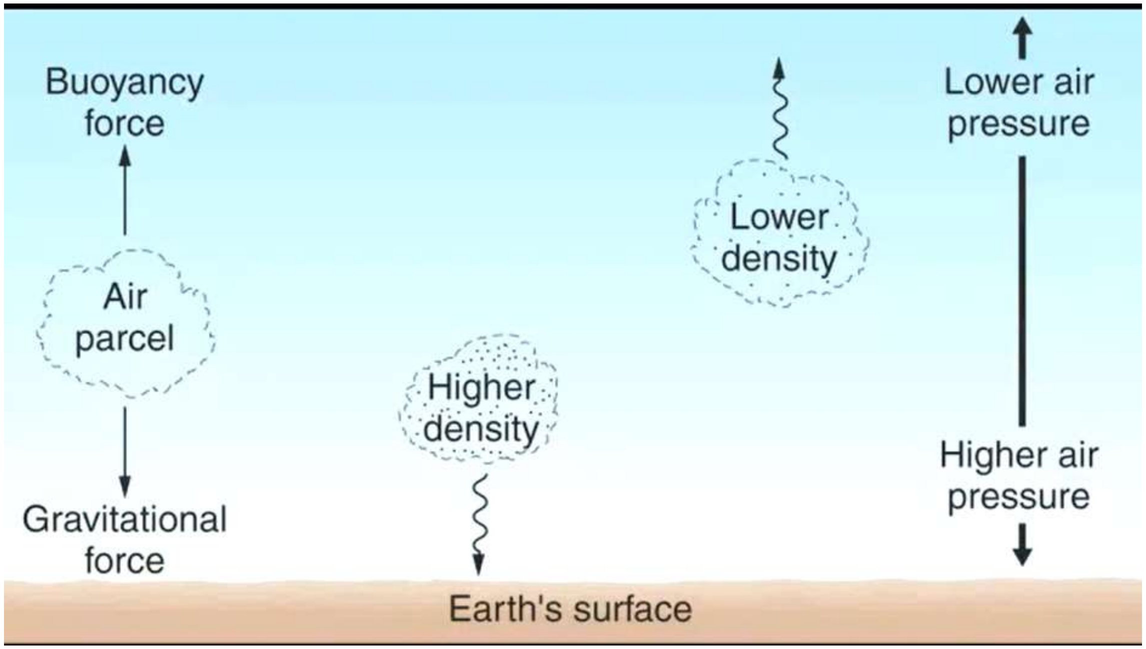
Advection = horizontal transport of air

Environmental Lapse rate classifications



- Superadiabatic refers to cooling more than expected (ELR>ALR)
- Subadiabatic referes to cooling lesser than expected (ELR<ALR)
- Inversion is an opposite phenomenon that refers to increase in temperature with elevation

What makes an Air Parcel rise or sink?



- Density of air parcel determines sinking or rising.
- If density of air parcel is lower then surrounding air it will RISE, while greater then it will SINK.
- Temperature determines the density of air parcel, air parcel cooler than surrounding have higher density while hot air parcel have lower density

Atmospheric Stability

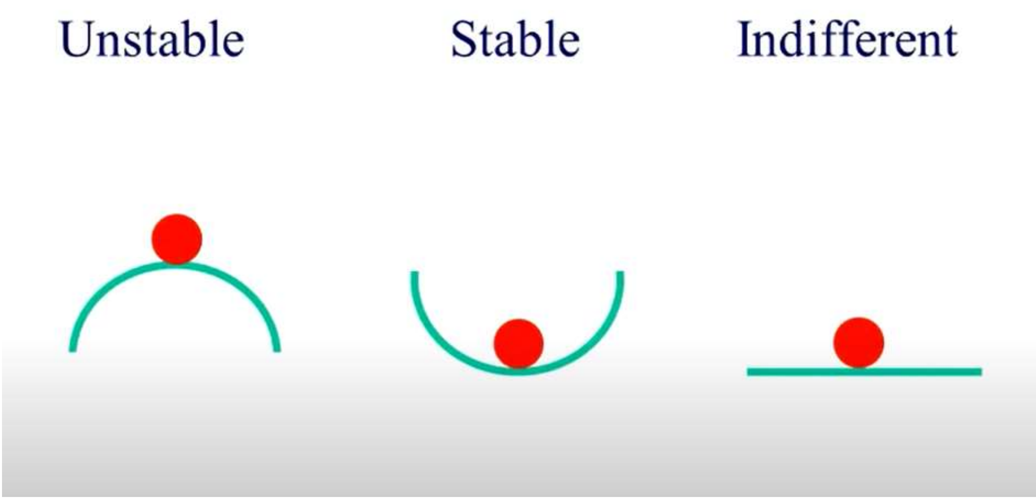
Tendency of atmosphere to resist or enhance further free vertical motion of air parcel which is forced initially is called atmospheric stability

Types of equilibrium :

- **Stable equilibrium** : after a small displacement the body returns to its **original** equilibrium position.

- **Unstable equilibrium** : after a small displacement the body **does not return** to the original equilibrium position and moves to a new equilibrium position.

- **Neutral equilibrium** : after a displacement the body **remains** in the displaced position.



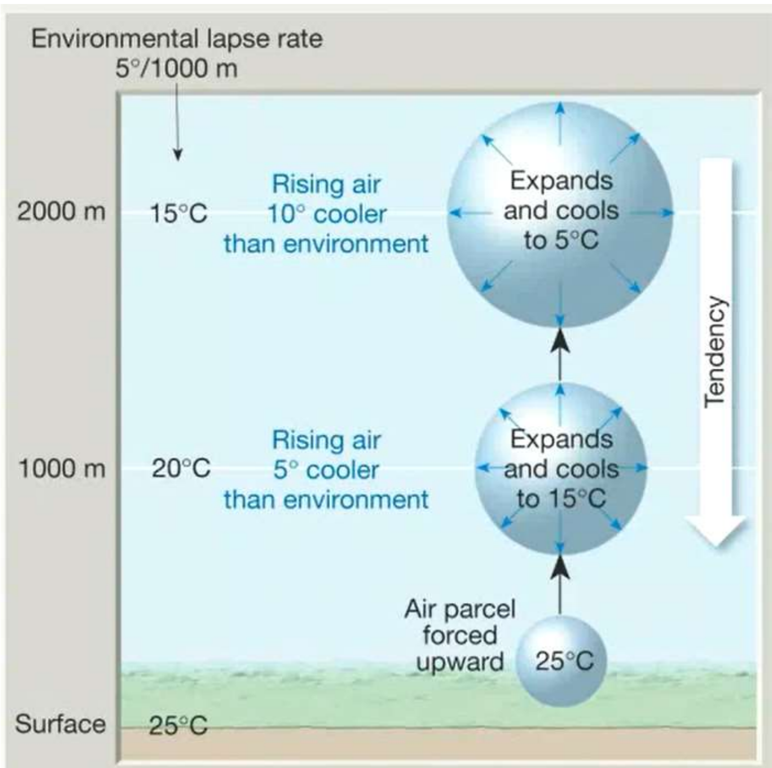
Assessing Atmospheric Stability

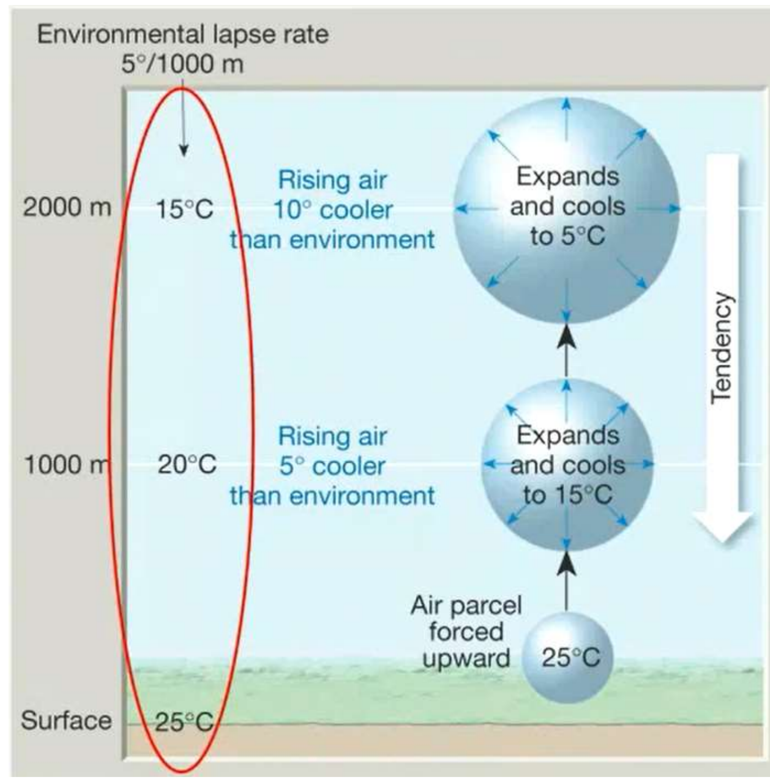
- The bottom line -
  - To determine whether or not a parcel will rise or sink in the atmosphere, one must compare the parcels temperature (T_p) with that of the environment (T_e) at some altitude:
  - if $T_p > T_e$ what will the parcel do?
  - if $T_p = T_e$ what will the parcel do?
  - if $T_p < T_e$ what will the parcel do?

- Cooler air packet released in atmosphere has lower temperature than prevailing conditions (atmosphere) then air packet descends

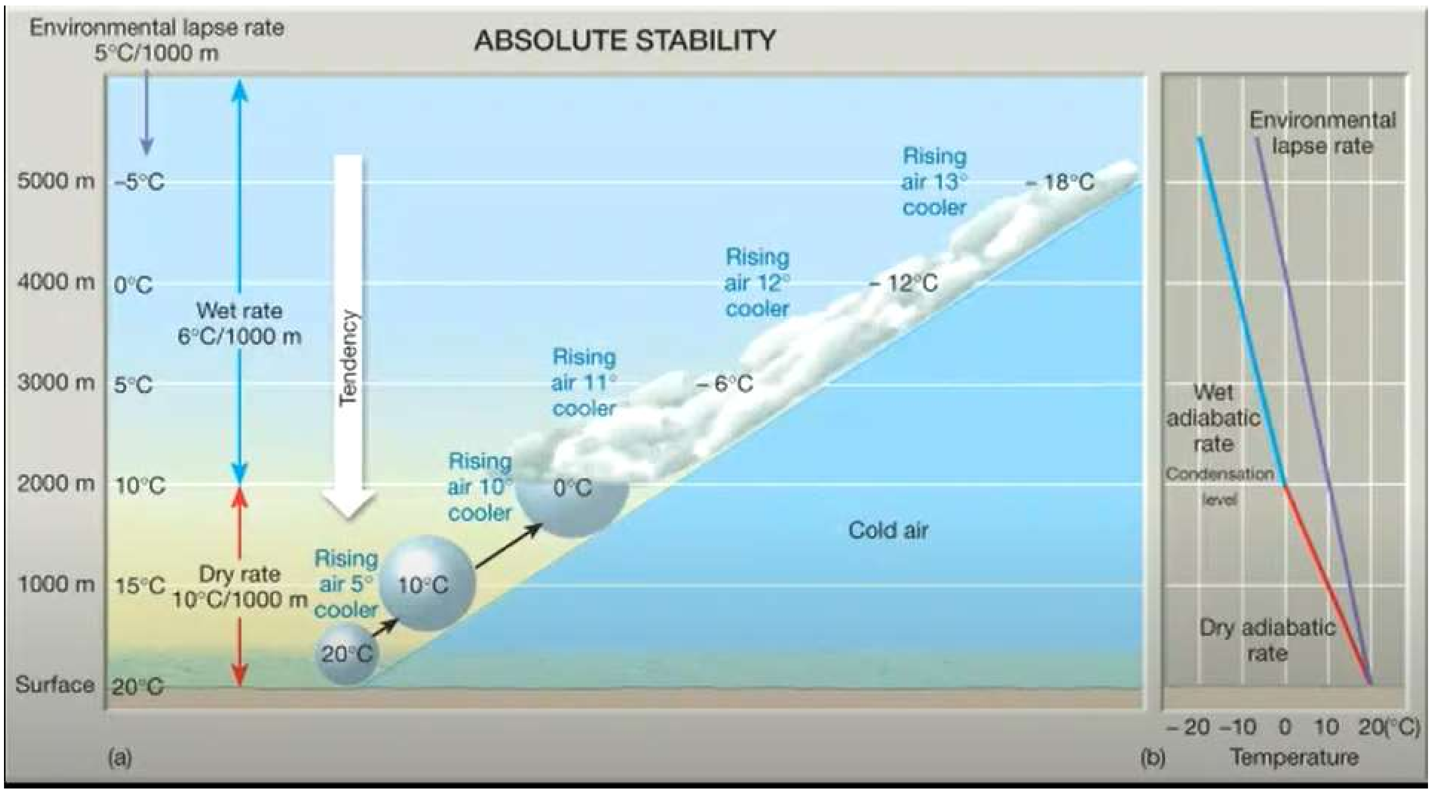
- Air packet released in atmosphere having higher temperature than surrounding temperature, air packet rises.

STABLE CONDITIONS



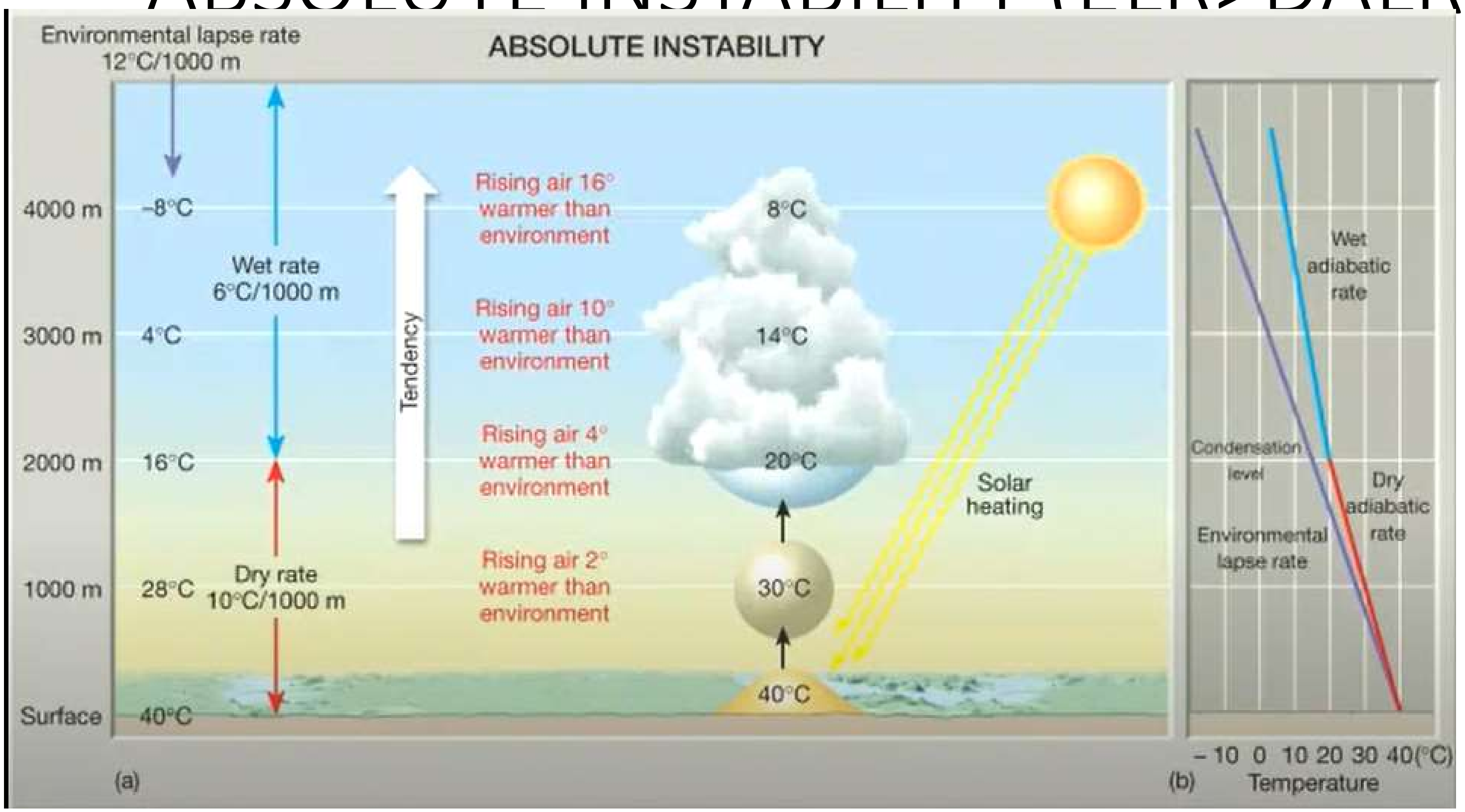


ABSOLUTE STABILITY (DALR>MALR>ELR)



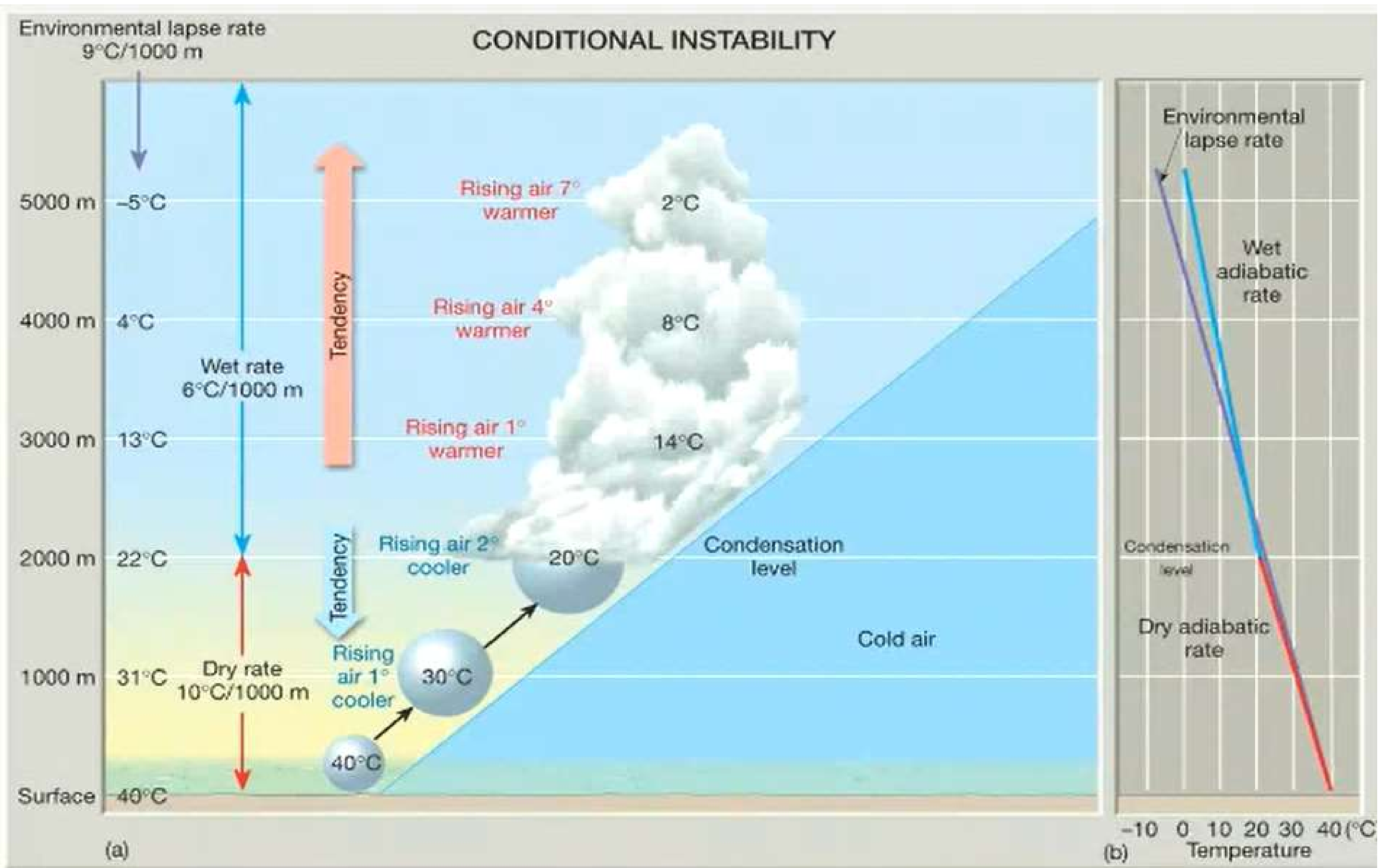
- Also called as Subadiabatic condition (ELR<ALR)
- The natural tendency of air parcel will be to sink downwards if the upward force is discontinued at any elevation.

UNSTABLE CONDITIONS
ABSOLUTE INSTABILITY (ELR>DALR>MALR)



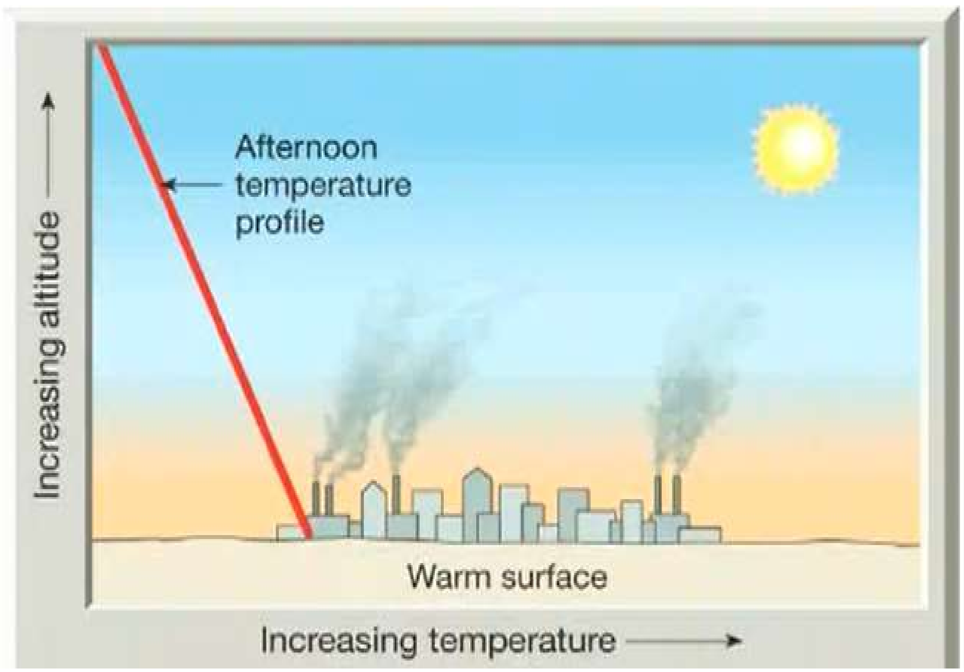
- Also called Superadiabtic conditions (ELR>ALR)
- The natural tendency of air parcel will be to freely RISE upwards if the upward force is discontinued at any elevation.

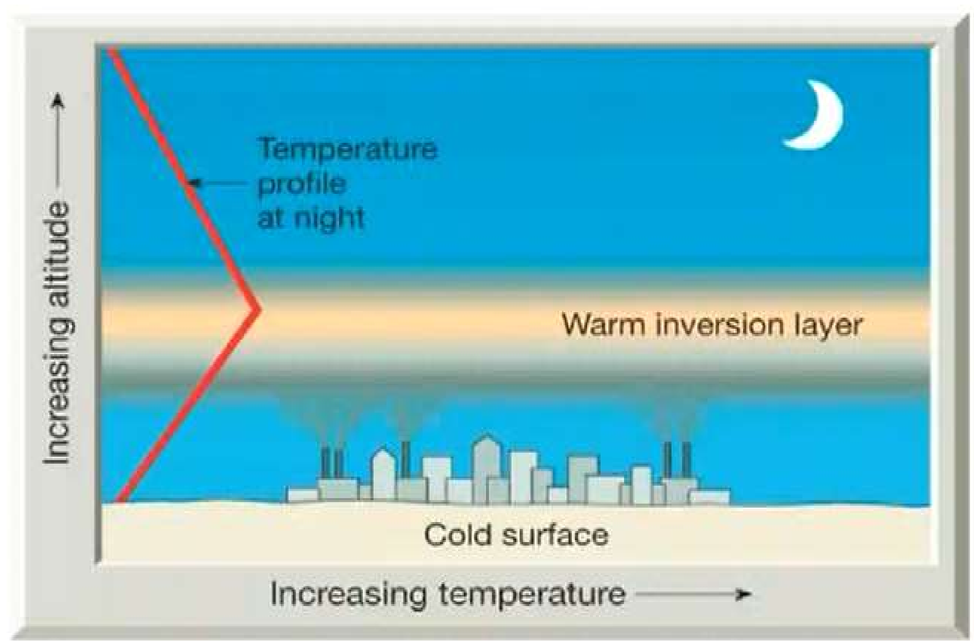
Conditional Instability (DALR>ELR>MALR)



The natural tendency of air parcel will be to sink downwards until the air parcel reaches CONDESATION LEVEL and enters the moist adiabatic lapse rate region from which it will freely RISE upwards

INVERSION- Extreme case of Stability





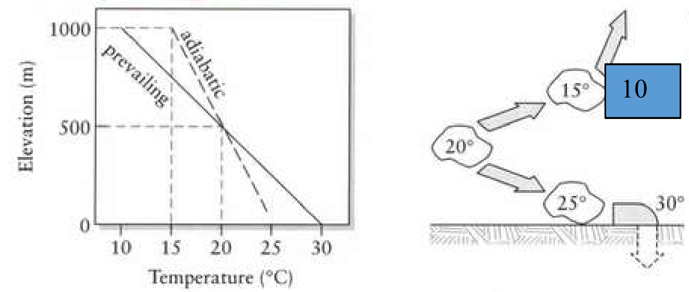
- Imagine a power plant releasing air pollutants at certain height whose temperature is 20 Celsius, and it follows adiabatic lapse rate (1 C per 100 m)

- If the atmospheric conditions are superadiabatic than temperature rate will be higher (say 2 C per 100 m)

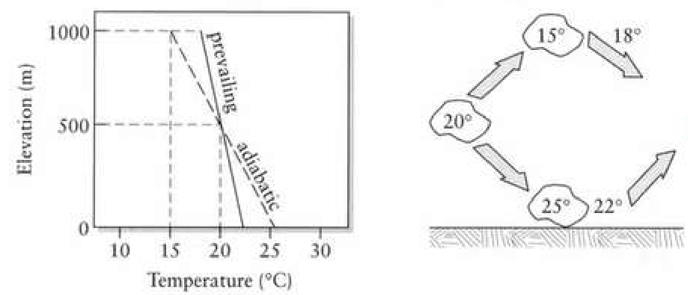
- If the atmospheric conditions are subadiabatic than temperature rate will be lower (say 0.4 C per 100 m)

Vertical Air Movement

The line represents a temperature gradient



A. Superadiabatic conditions (unstable)



B. Subadiabatic conditions (stable)

Example:

- *Unstable conditions:* prevailing temperature at 500 m is 20 °C
  - If a *parcel* of air moves upward 500 m, it will cool by 5°C to 15°C
  - But the prevailing temperature is 10°C, so it will continue rising
- *Stable conditions:*
  - A parcel of air moved upward will be cooler than surrounding air (18°C). It will be forced to descend