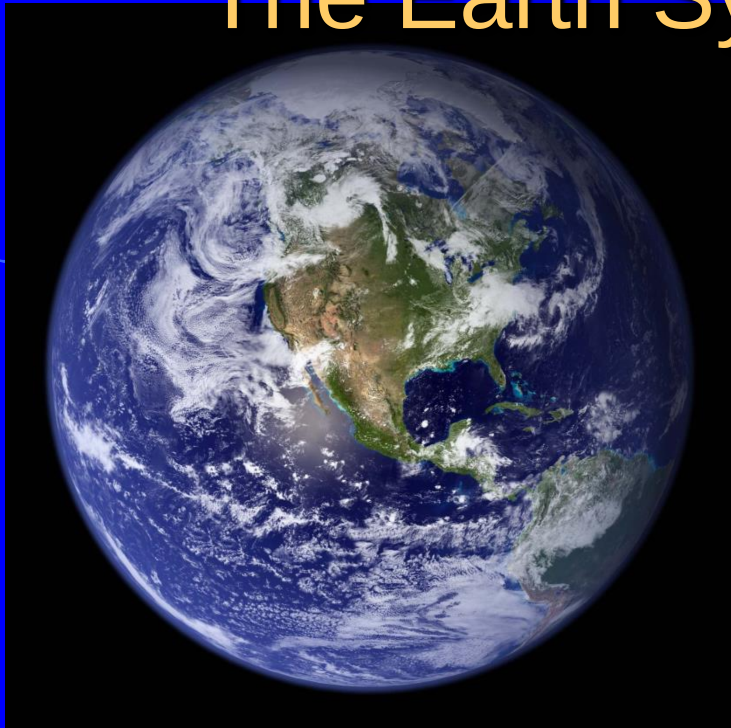
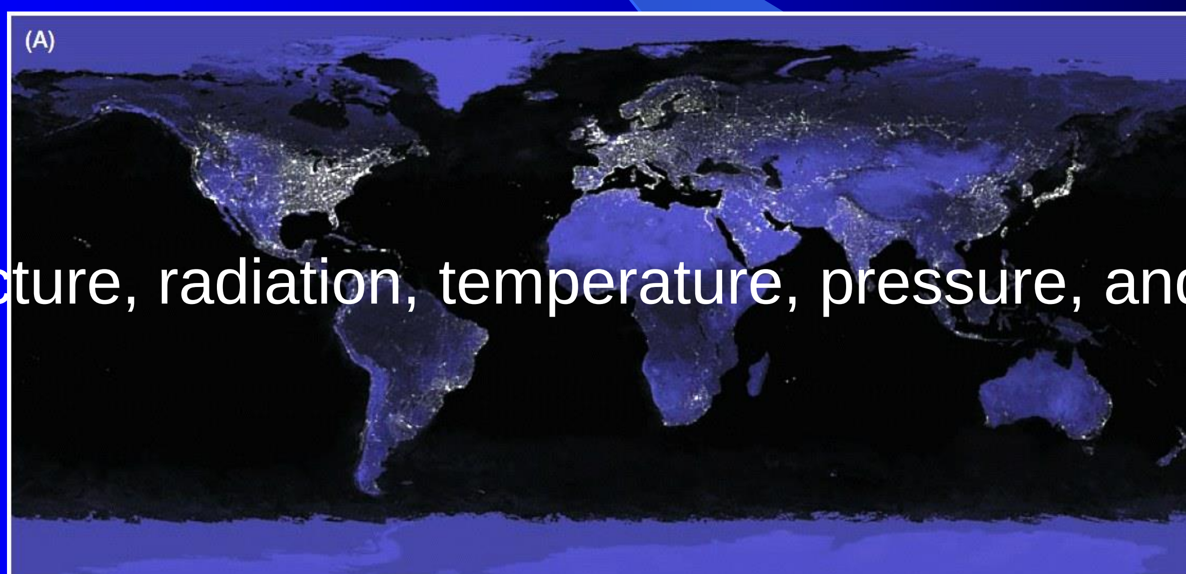
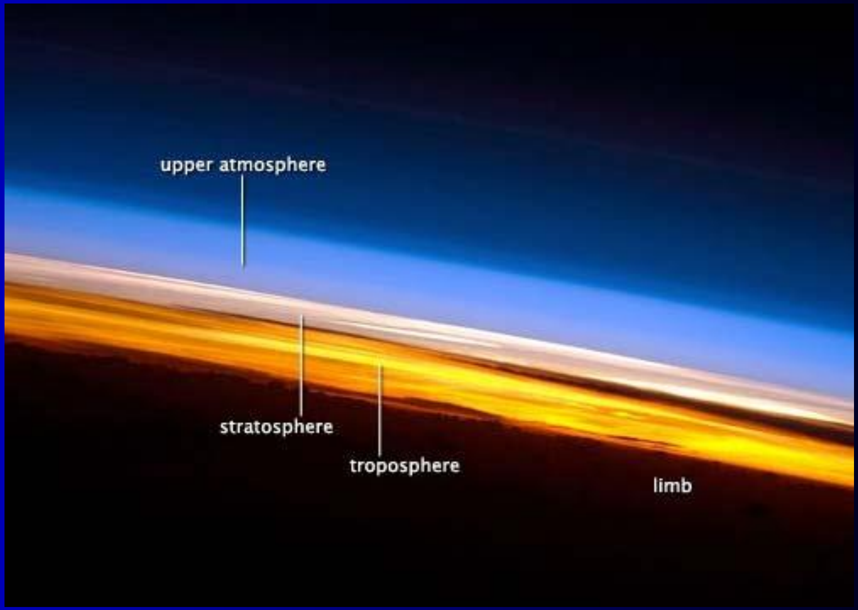
The Earth System - Atmosphere I

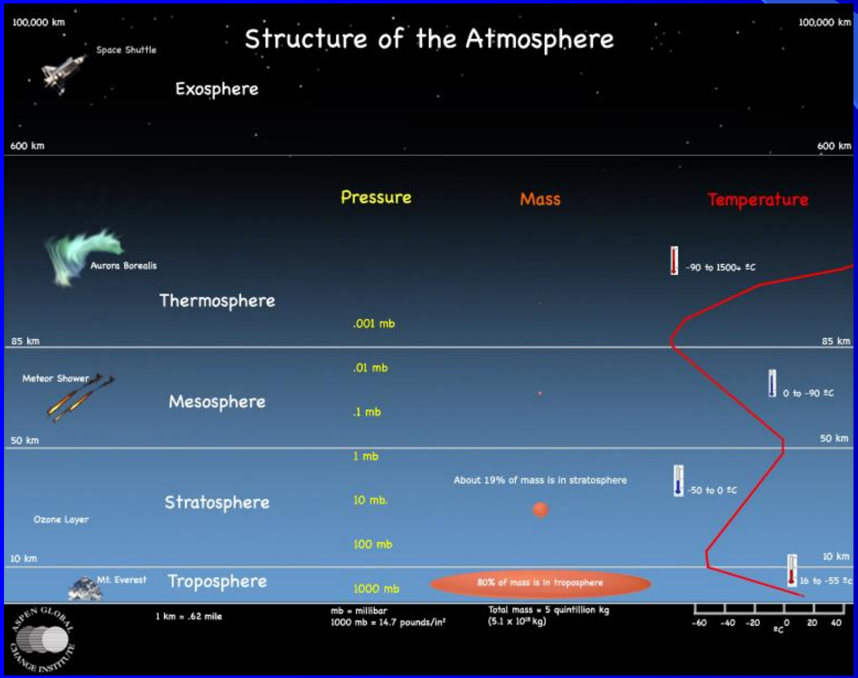


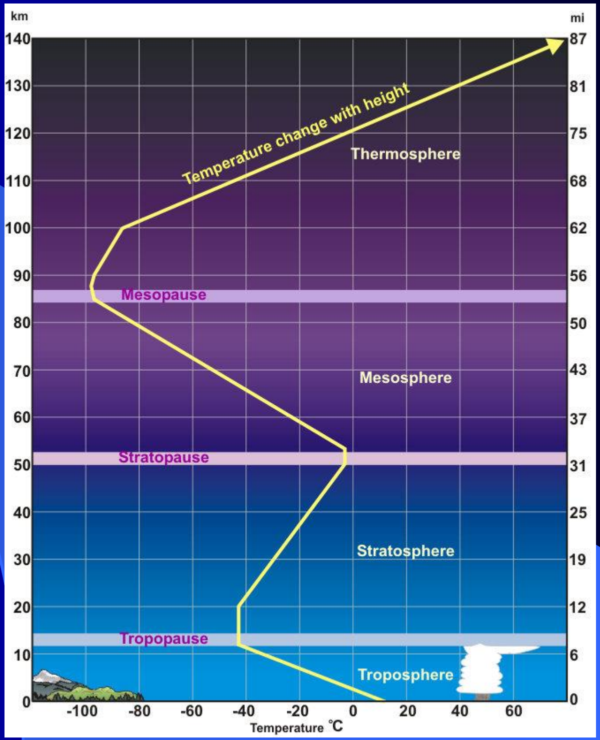


Structure, radiation, temperature, pressure, and water

The Atmosphere







Components of the Earth's Atmosphere

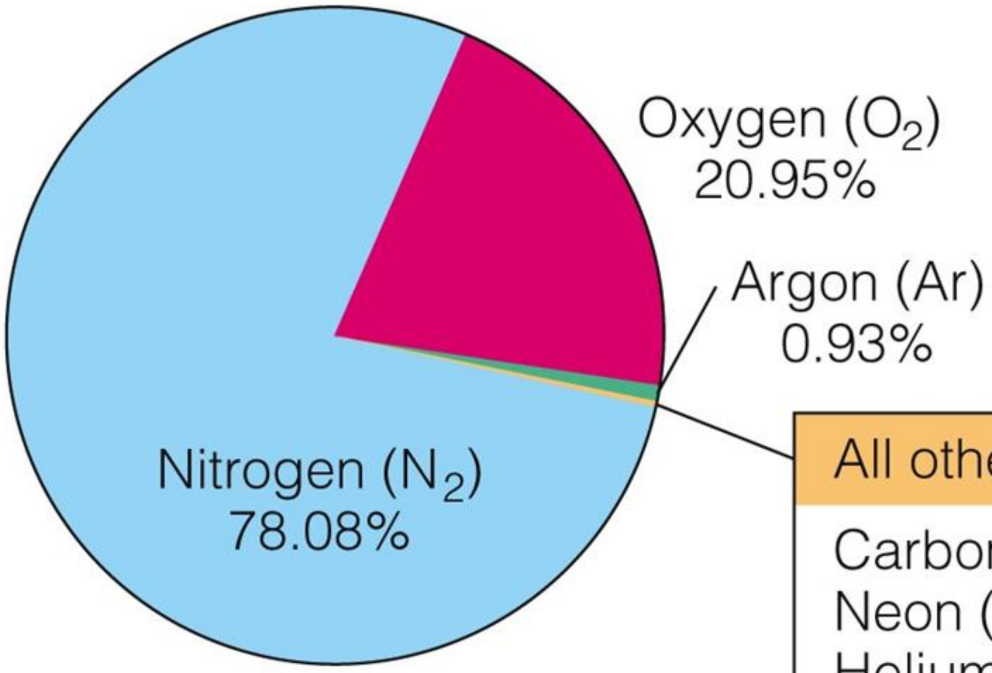
- The major gases of the atmosphere are

  Nitrogen, oxygen and argon
- The concentration of these gases is approximately constant.
- Minor gases in the atmosphere are

  Carbon dioxide, methane, ozone, water vapor and NO_x
- Also present in the atmosphere are

  Aerosols: tiny suspended liquid or solid particles

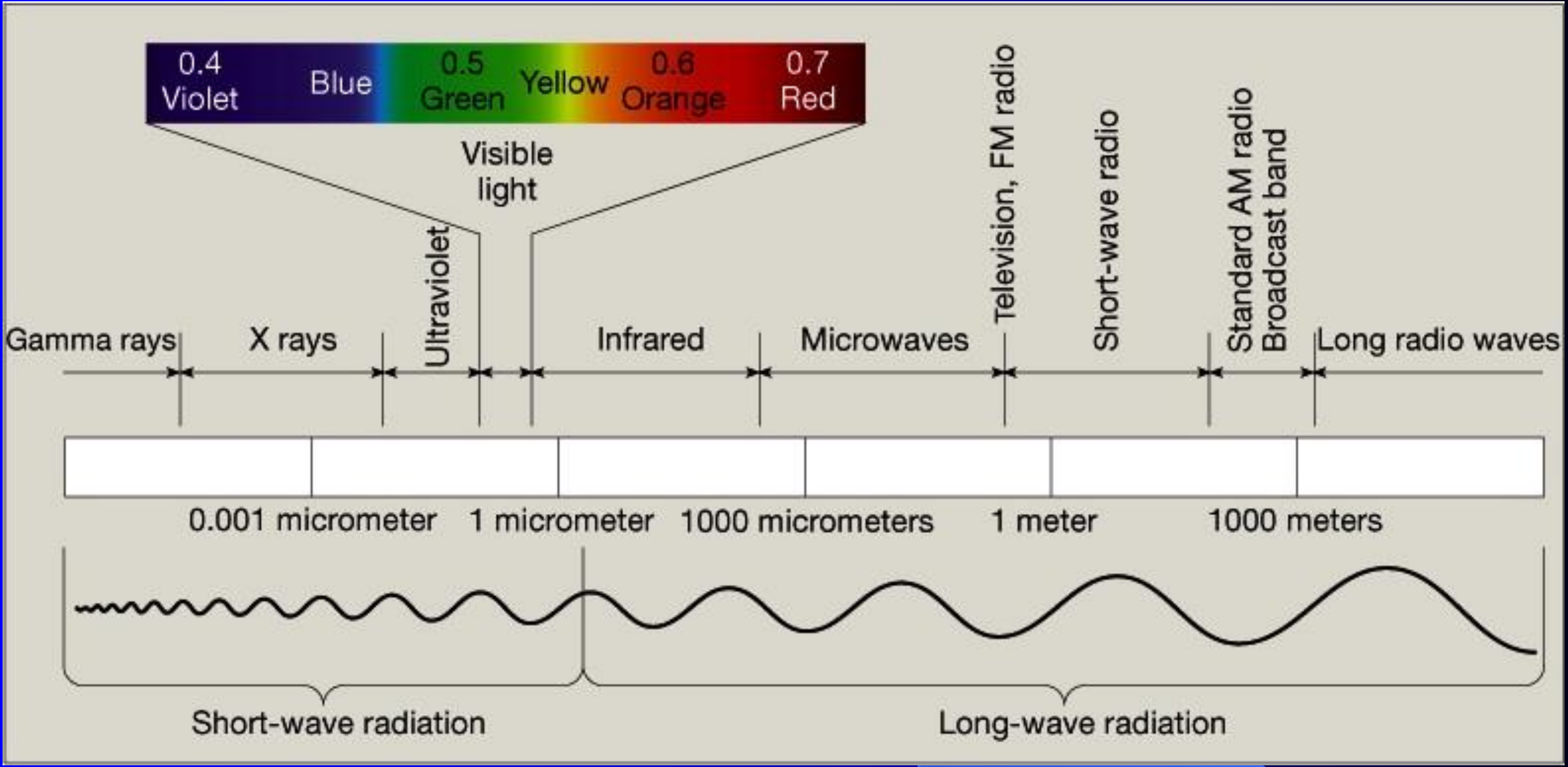
Composition of the Earth's Atmosphere



All other gases, 0.04%	
Carbon dioxide (CO_2)	0.035%
Neon (Ne)	0.0018%
Helium (He)	0.00052%
Methane (CH_4)	0.00014%
Krypton (Kr)	0.00010%
Nitrous oxide (N_2O)	0.00005%
Hydrogen (H_2)	0.00005%
Ozone (O_3)	0.000007%

Variable substances	
Water vapor (H_2O)	1-4%
Aerosols	variable

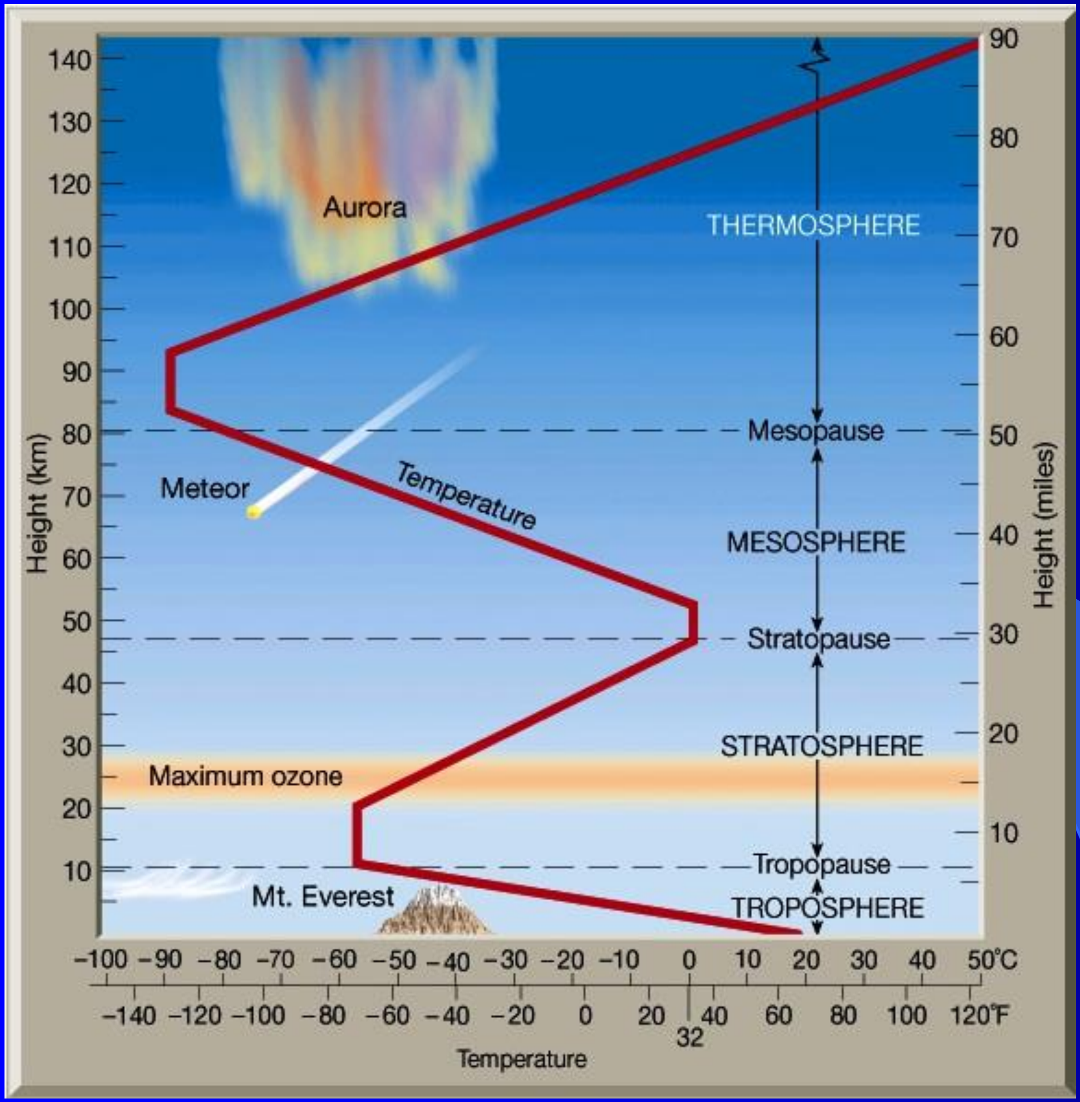
The Electromagnetic Spectrum

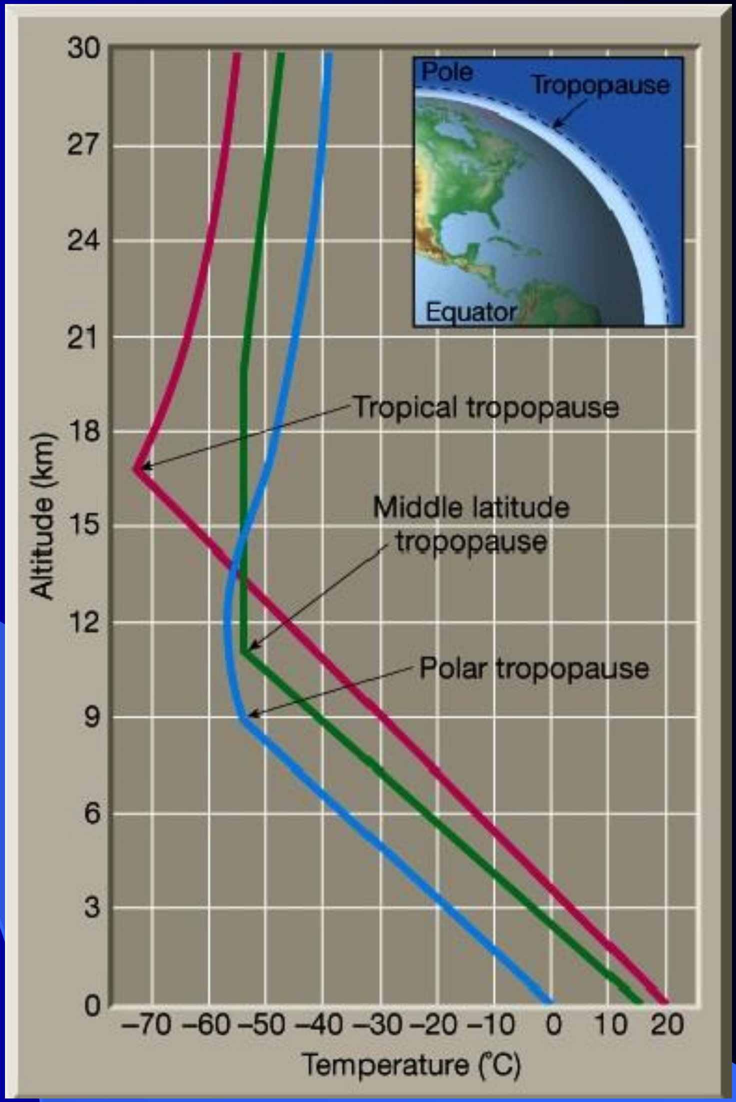


Wavelength and frequency are inversely related: Frequency = 1/wavelength

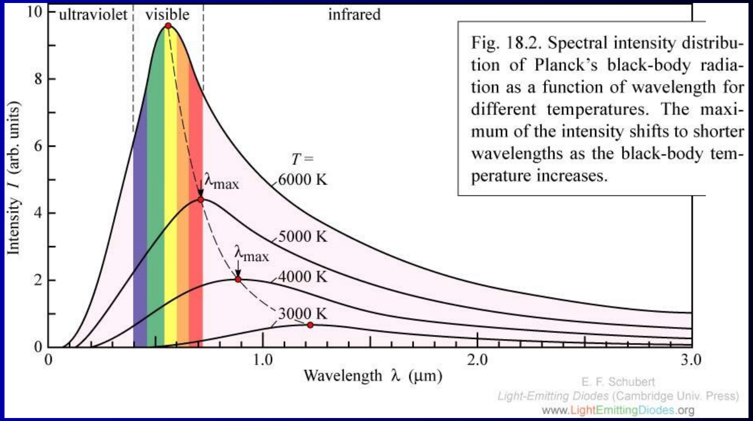
Photon = light particle: Energy of photon = Planck’s constant x frequency

Layers of the Earth's atmosphere and variations in the height of the tropopause as a function of latitude





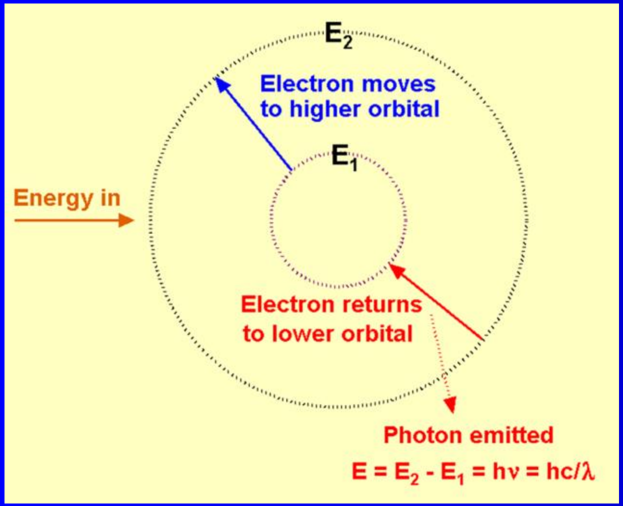
Types of Spectra

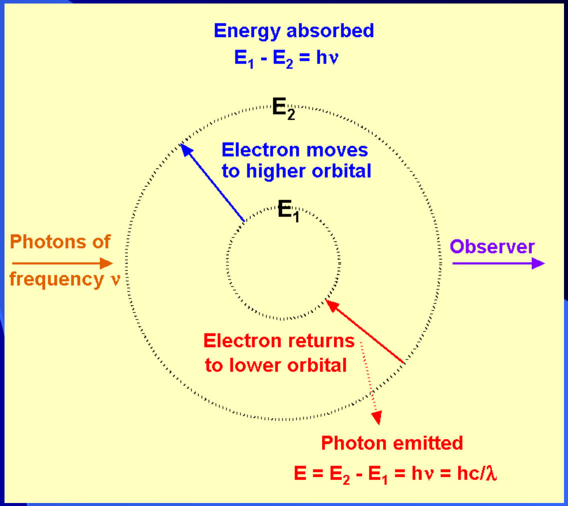


A **continuous spectrum** is emitted by solids, liquids and compressed gases. A **perfect radiator** emits the maximum amount of energy at all wavelengths as a function of temperature.

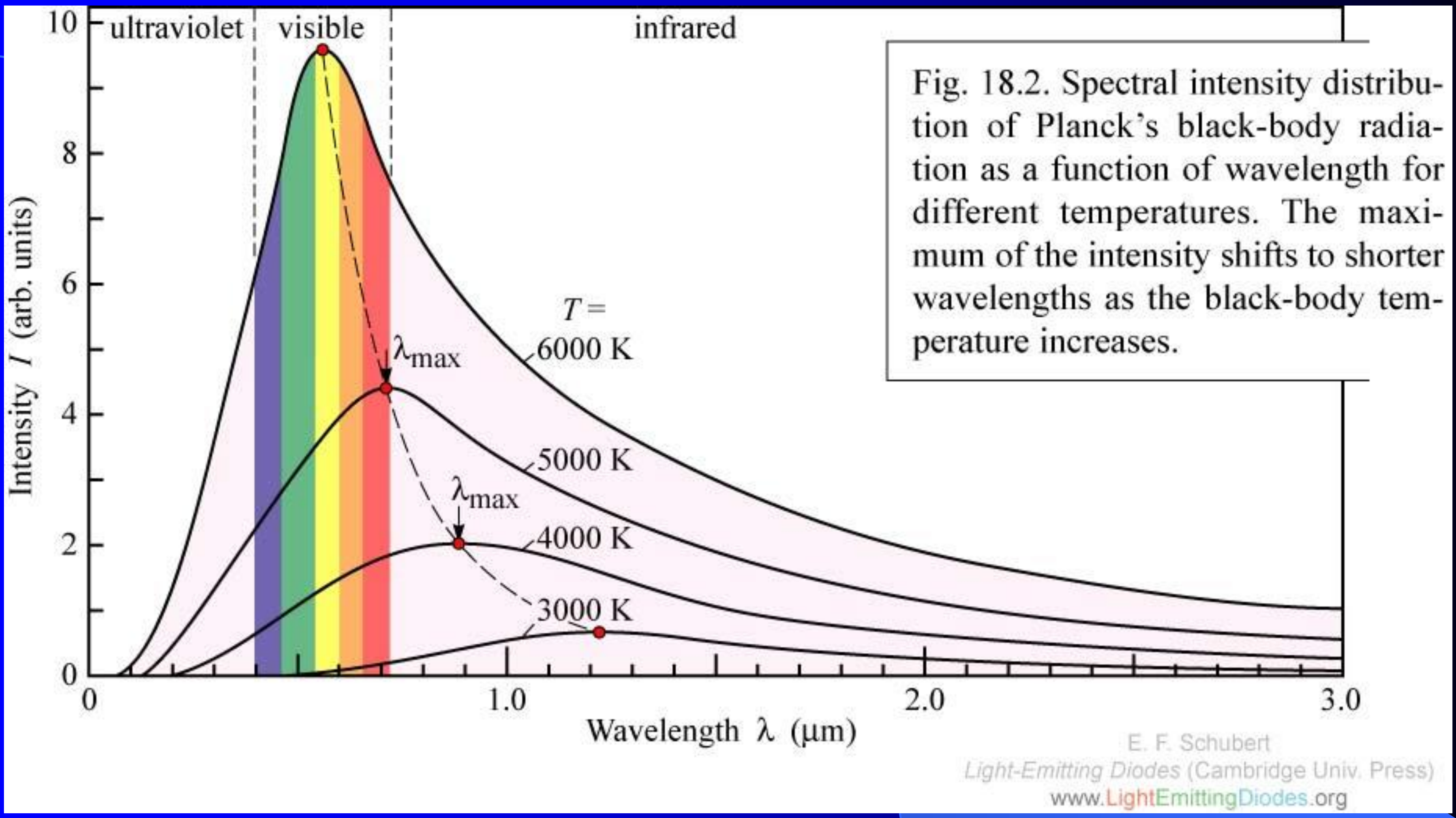
An **emission spectrum** occurs when energy applied to the atom causes an electron to move from a lower orbital to a higher orbital. The electron returns to a lower orbital and emits energy corresponding to the energy difference between the two orbitals (E_2 - E_1). Planck's constant is *h*, ν is frequency, and λ is wavelength.

An **absorption spectrum** occurs when photons that have exactly the right energy to move an electron from one orbital to another interact with the atom. When the electron returns to the lower orbital, the emitted photon can travel in any direction. Thus, the observer notices a decrease in the number of photons of this energy (wavelength). Planck's constant is *h*, ν is frequency, and λ is wavelength.





Continuous spectrum



Equations describing *Perfect Radiators (Black bodies)*

Stefan-Boltzmann Law: $E = \sigma T^4$

E = total energy radiated per unit area of a black body

$\sigma = 5.670373 \times 10^{-8}\ \mathrm{J\ s^{-1}\ m^{-2}\ K^{-4}} = 5.670373 \times 10^{-8}\ \mathrm{W\ m^{-2}\ K^{-4}}$

T = Temperature (in K)

Wien's Displacement Law : $\lambda_{max} = a/T$

λ_{max} = wavelength of the peak of the emission of a black body

$a = 2.8977685 \times 10^{-3}$ m K = 2,897,768 nm K = 2900 μ m

T = Temperature (in K)

Stratospheric Ozone Depletion

- One of a number of global-scale problems caused by anthropogenic perturbations to planetary chemical processes
- Results from large-scale industrial manufacture and release of synthetic compounds (chlorofluorocarbons, CFCs) in quantities that can interfere with the basic functions of the Earth's atmosphere
- Unanticipated - like acid rain, global warming, etc., effects of CFCs were not expected… only appreciated in hindsight

Ozone Formation

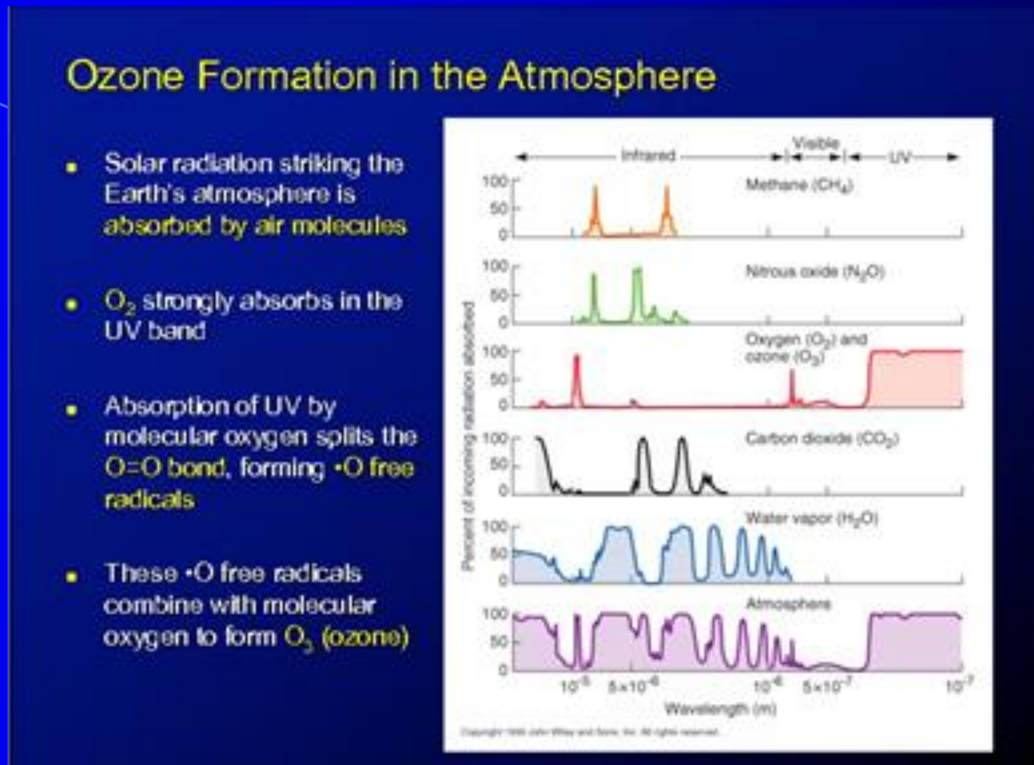
- In the stratosphere, UV intensity is sufficient to form O_3

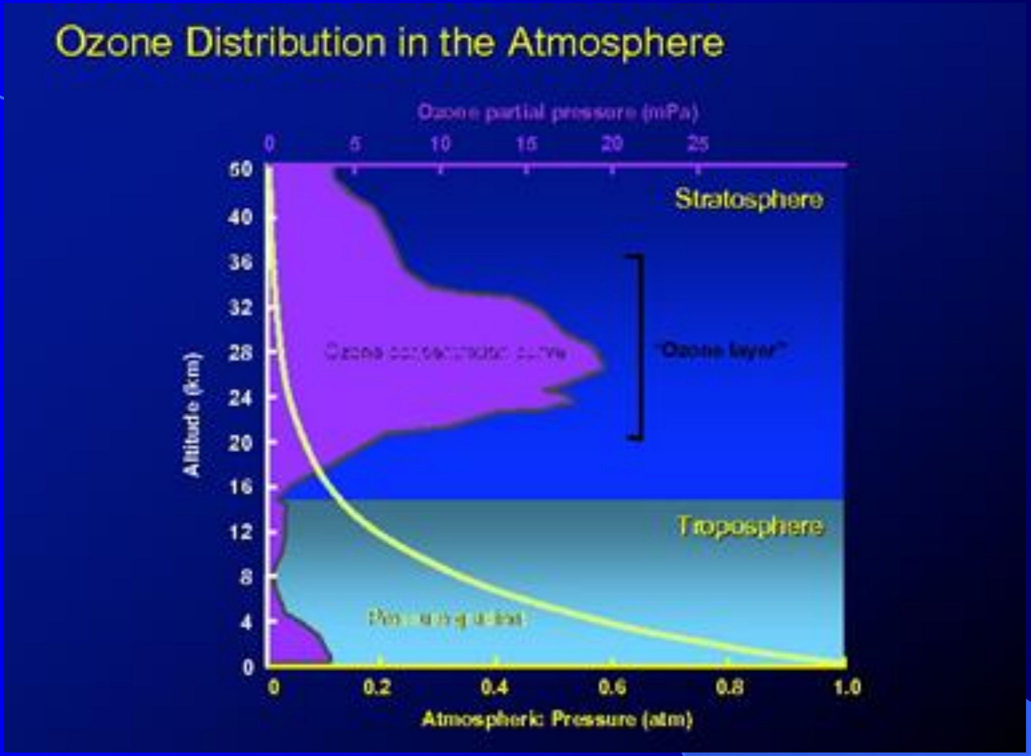
$O_2(g) + UV = 2 \cdot O$
$\cdot O + O_2(g) = O_3(g)$

- UV radiation can also destroy ozone molecules...

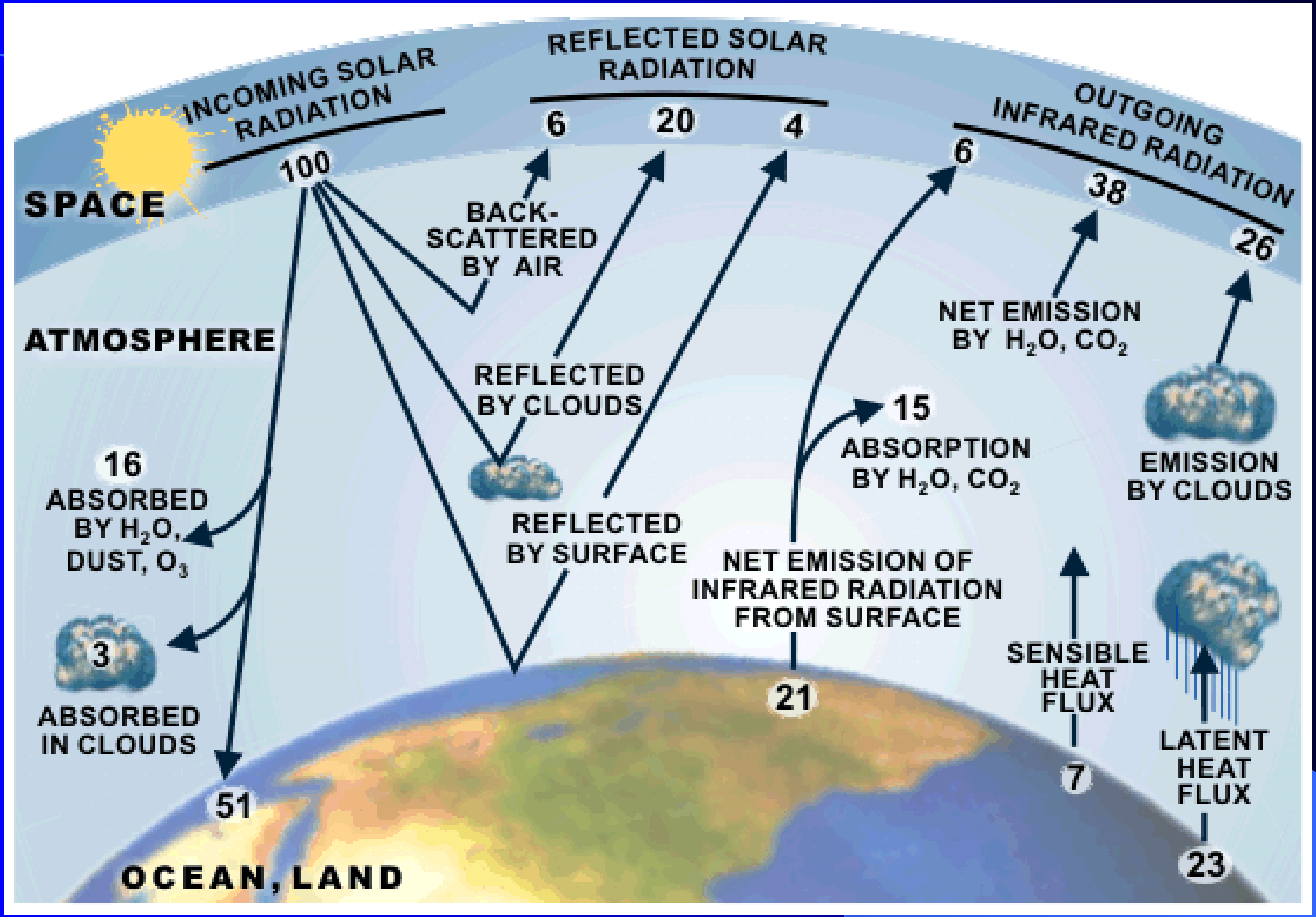
$O_3(g) + UV = \cdot O + O_2(g)$
$O_3(g) + \cdot O = 2O_2(g)$

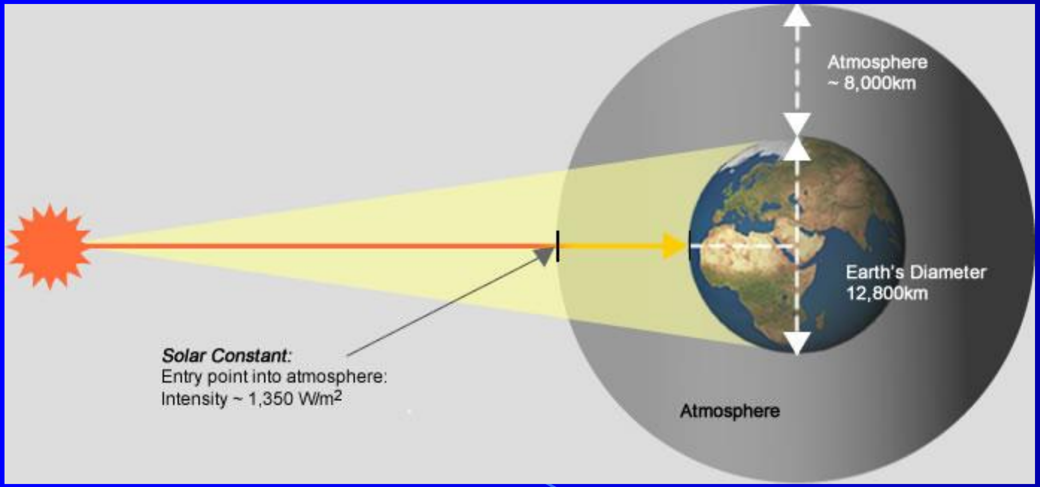
- The net effect (normally) is to maintain a low equilibrium concentration of ozone in the middle stratosphere.


Ozone Formation in the Atmosphere
Solar radiation striking the Earth's atmosphere is absorbed by air molecules
O_2 strongly absorbs in the UV band
Absorption of UV by molecular oxygen splits the O=O bond, forming •O free radicals
These •O free radicals combine with molecular oxygen to form O_3 (ozone)
Infrared
Visible
UV
Methane (CH_4)
Nitrous oxide (N_2O)
Oxygen (O_2) and ozone (O_3)
Carbon dioxide (CO_2)
Water vapor (H_2O)
Atmosphere
Percent of incoming radiation absorbed
100
50
0
10^{-5}
5×10^{-6}
10^{-6}
5×10^{-7}
10^{-7}
Wavelength (m)


Ozone Distribution in the Atmosphere
Ozone partial pressure (mPa)
0
5
10
15
20
25
50
40
36
32
28
24
20
16
12
8
4
0
Altitude (km)
Stratosphere
"Ozone layer"
Troposphere
0
0.2
0.4
0.6
0.8
1.0
Atmospheric Pressure (atm)

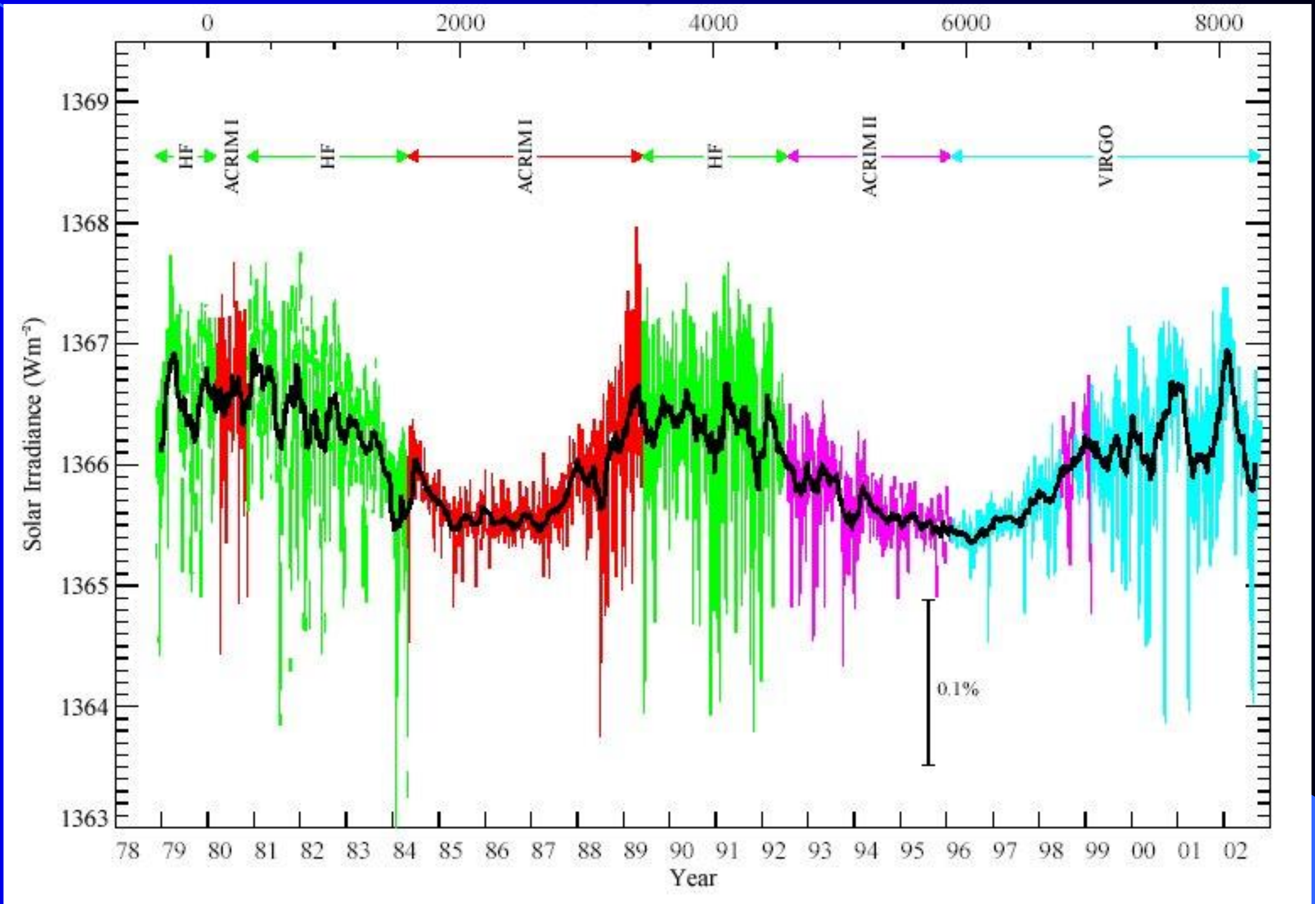
Absorption, emission, and reflection of incoming short-wave radiation from the sun and outgoing long-wave radiation from Earth

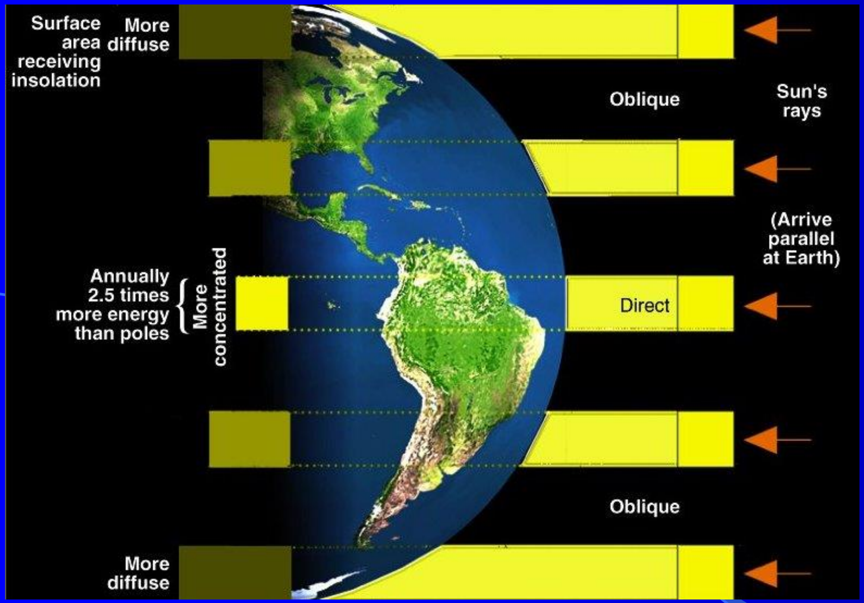




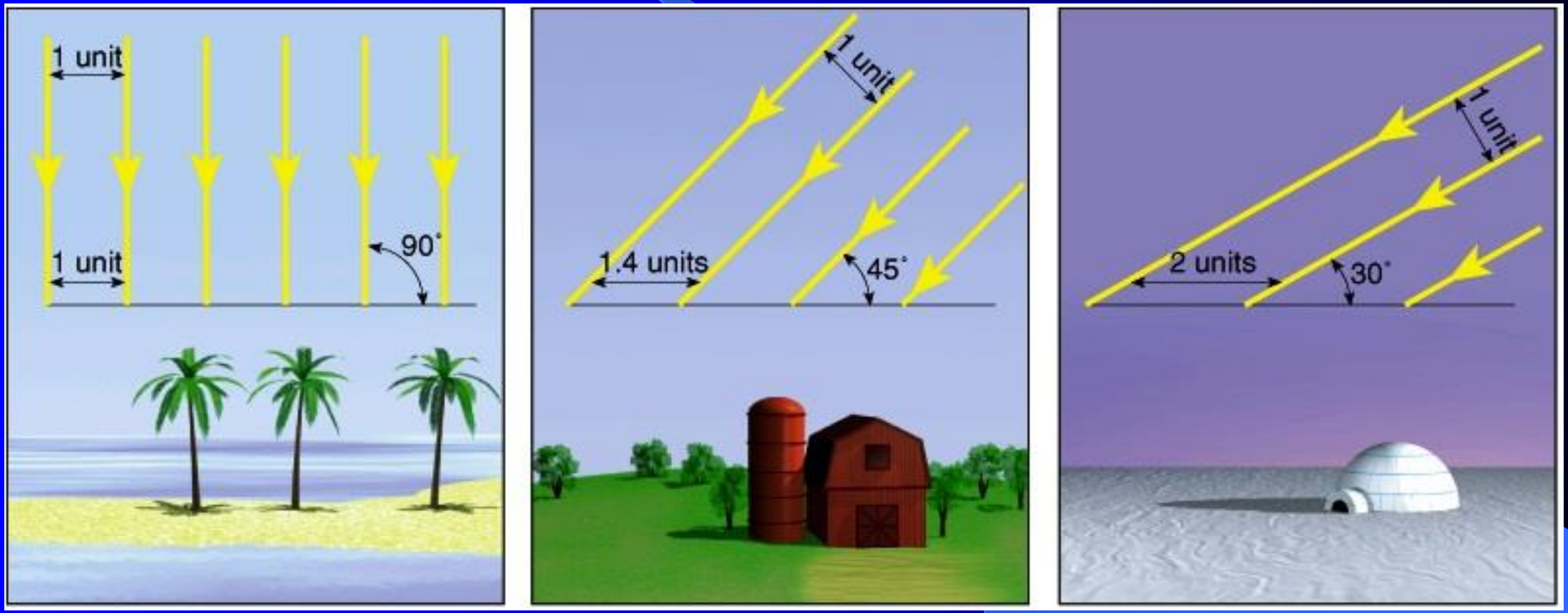
Solar constant – amount of energy received per unit area for a surface that is oriented perpendicular to the sun's rays at a distance equal to the average distance of the Earth from the sun.

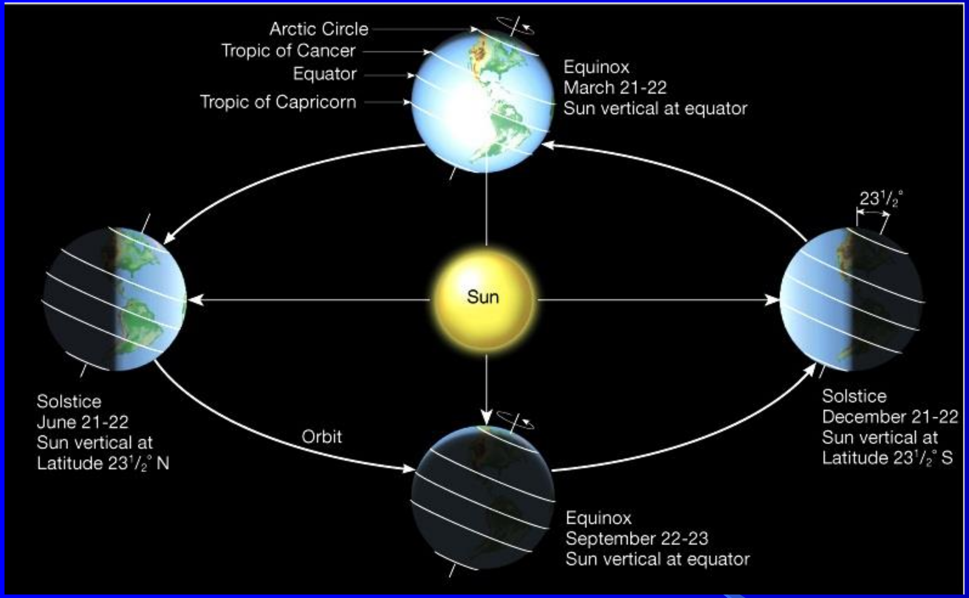
Energy output from the sun varies with time and therefore the solar constant is not constant. These variations tend to be cyclical. The one shown here has an approximately 10 year cycle.





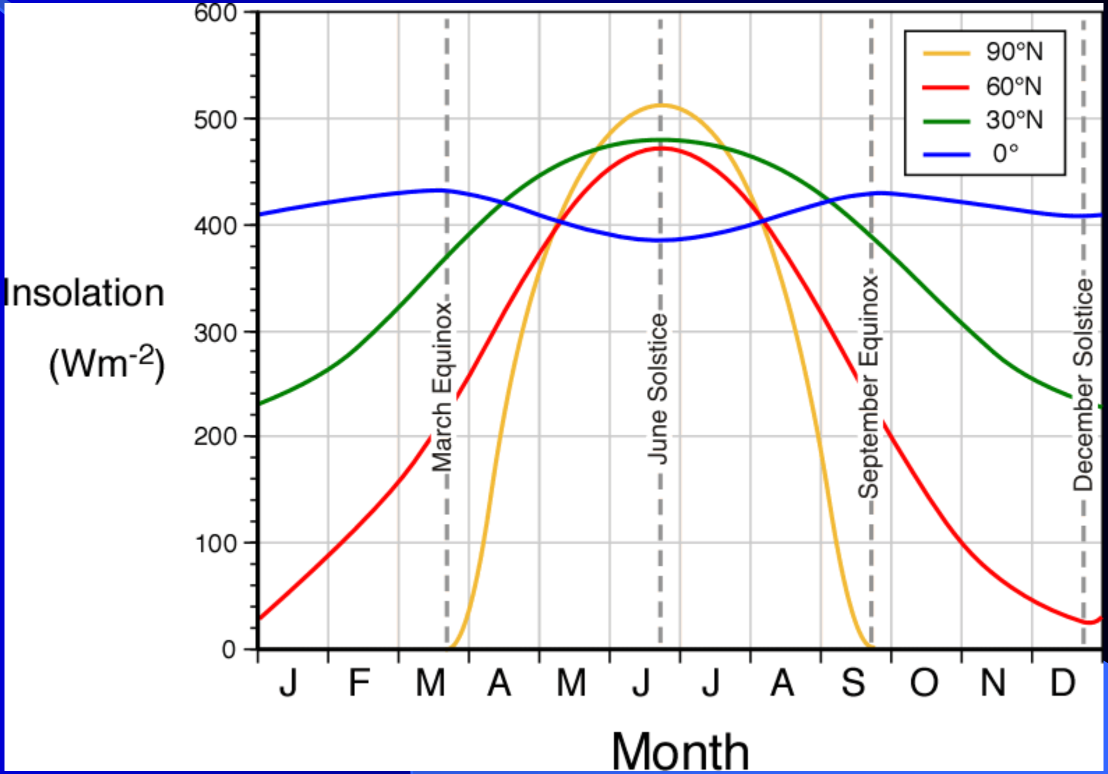
Insolation – Incoming Solar Radiation – energy absorbed per unit area at the Earth's surface. The energy absorbed varies with latitude because the angle of the sun's rays changes.



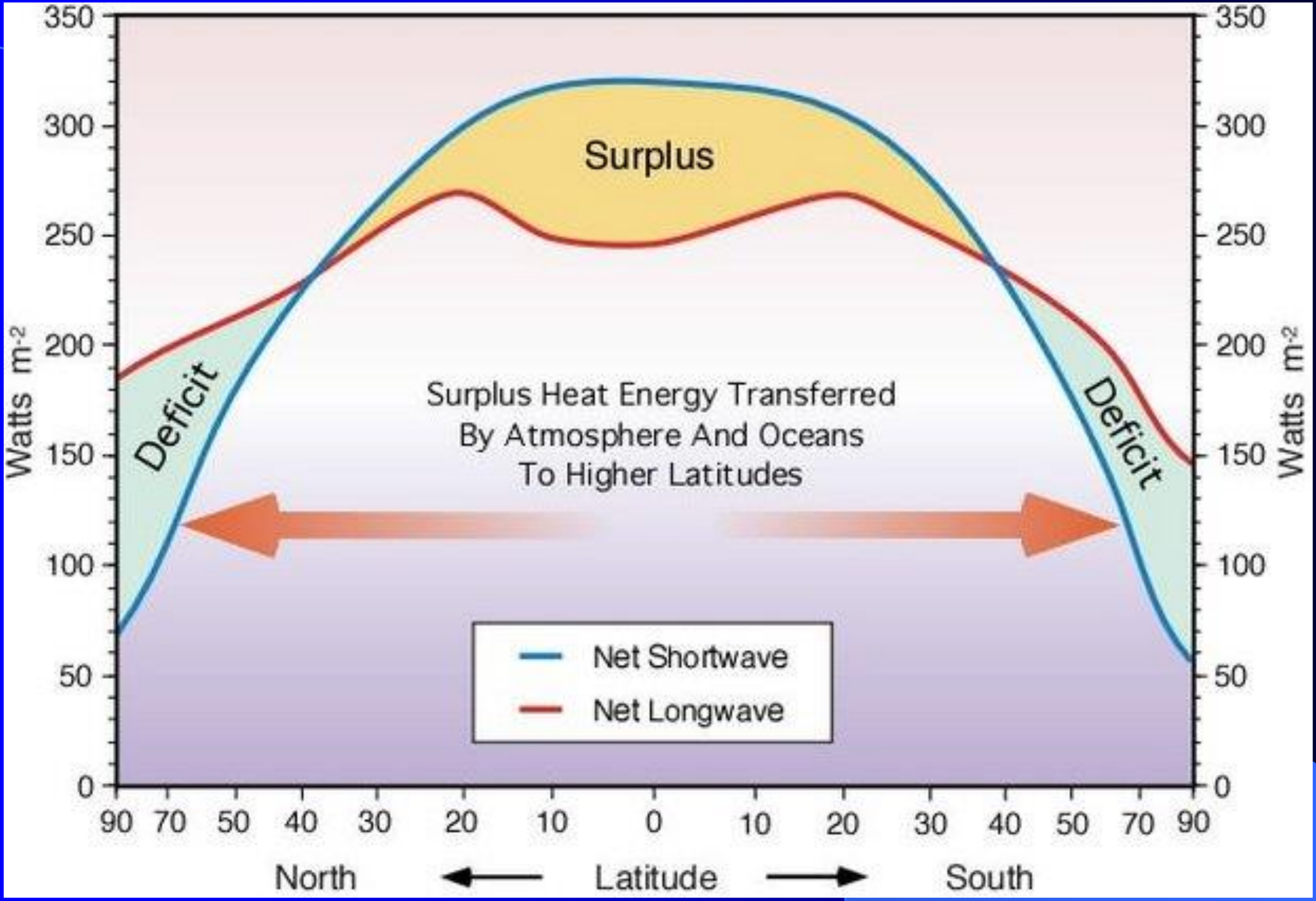


The Earth's axis is tilted at 23 ½ degrees with respect to the plane of the ecliptic. As the Earth travels around the sun the illuminated surface changes. This gives rise to the seasons.

Variation in insolation at various latitudes as a function of time of year. Variations are much greater at higher latitudes than at lower latitudes.



Variation in incoming short-wave radiation and outgoing long-wave radiation as a function of latitude. At high latitudes there is an energy deficit while at low latitudes there is an energy surplus.



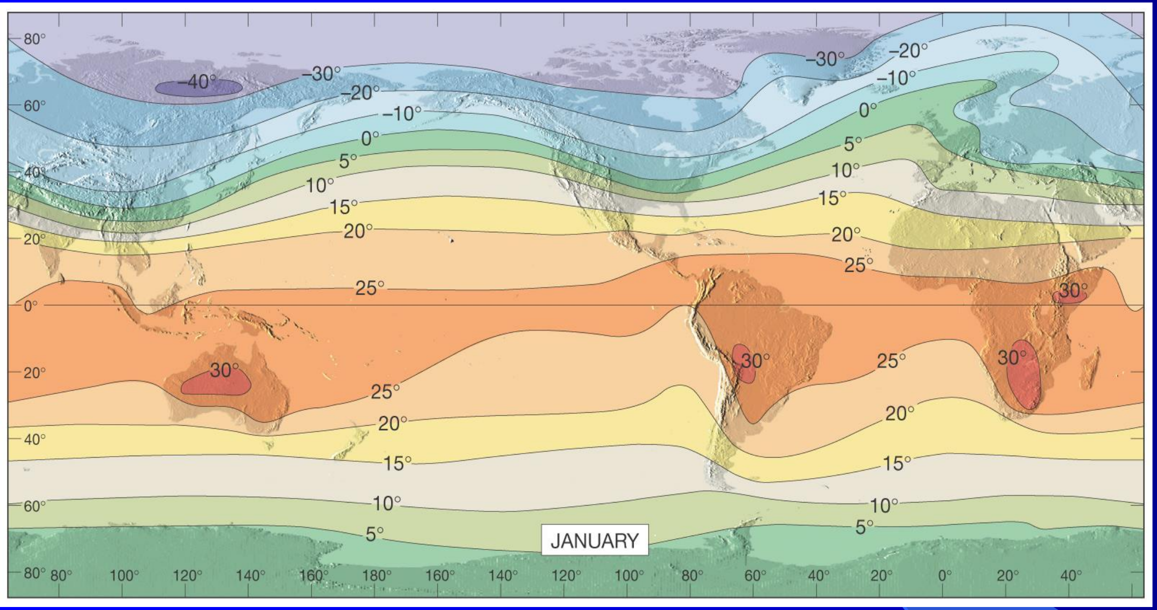
Heat Balance

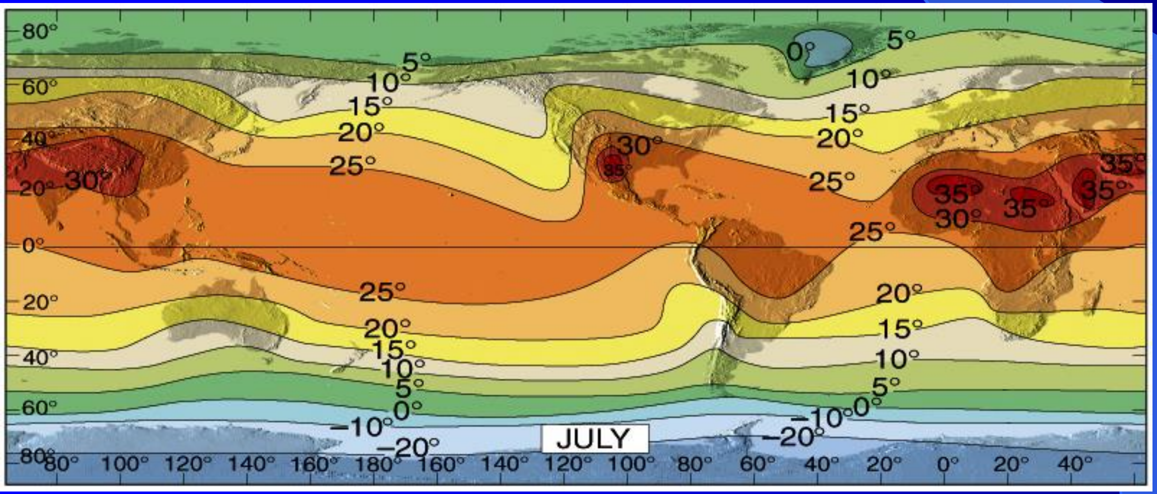
- Earth's average T constant. Total heat inflow = total heat outflow
- Average T across Earth's surface remains constant. Net transfer of heat from areas of surplus to areas of deficit

Heat Transport

- Simple thermal convection
- Large scale eddies produced by Earth's rotation
- Ocean currents
- Heat transport by water vapor.

Global and Regional Temperature Variations



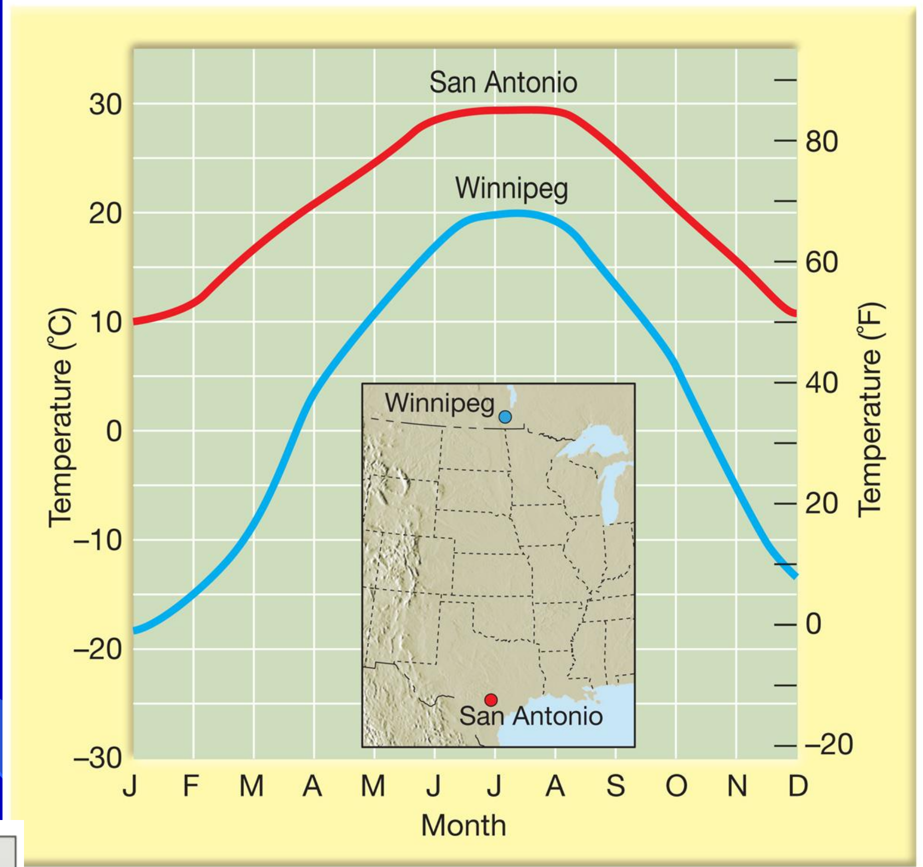


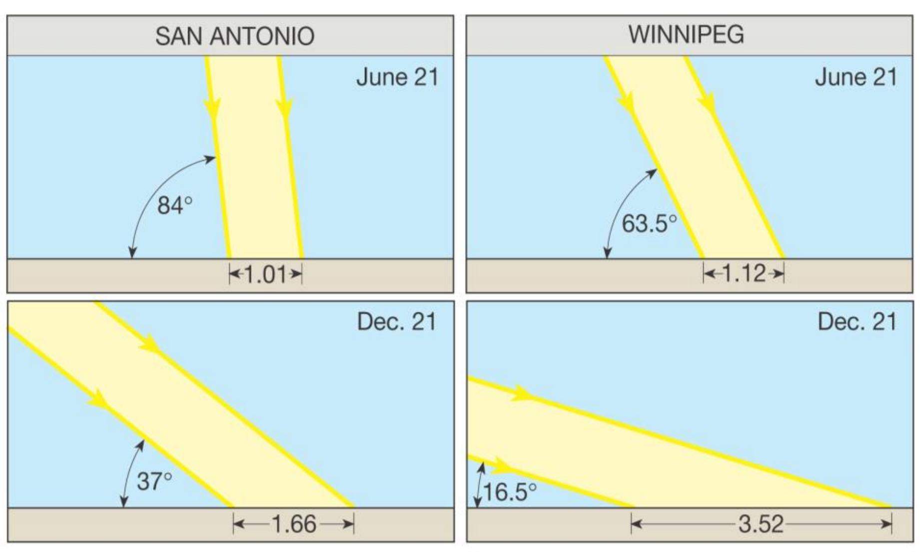
The mean daily temperatures vary with the position of the sun in the sky. In January the sun is directly overhead at 23 degrees South latitude and the maximum temperature is found south of the equator. The reverse is true in July. Note that the isotherms move more over the continents than the ocean. This is because of the higher heat capacity of water relative to land.

Factors that control temperature changes

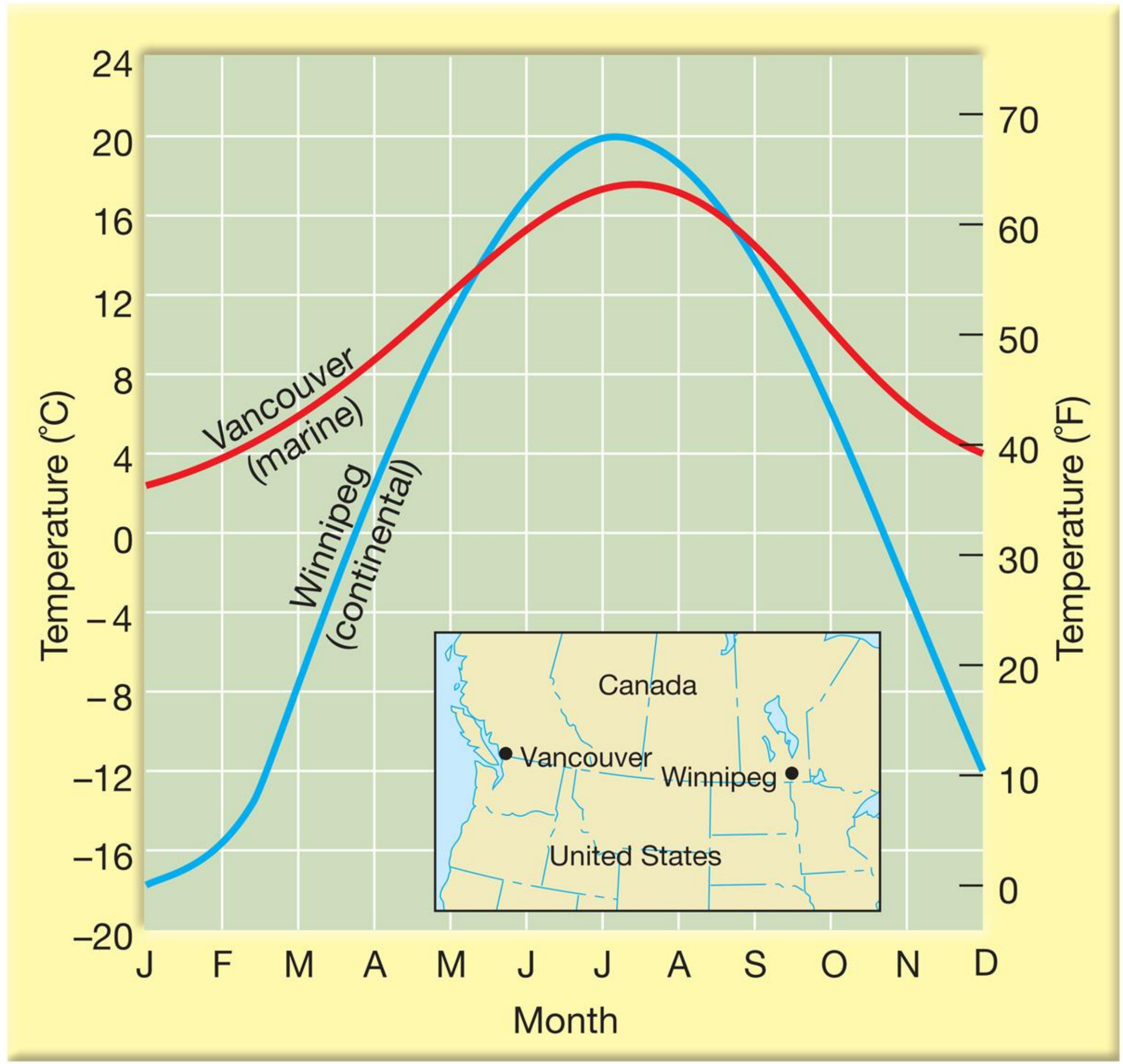
1. Latitude (angle of Sun)
2. Differential heating of land and water
3. Ocean Currents
4. Altitude
5. Geographic position
6. Cloud cover & albedo

Latitude

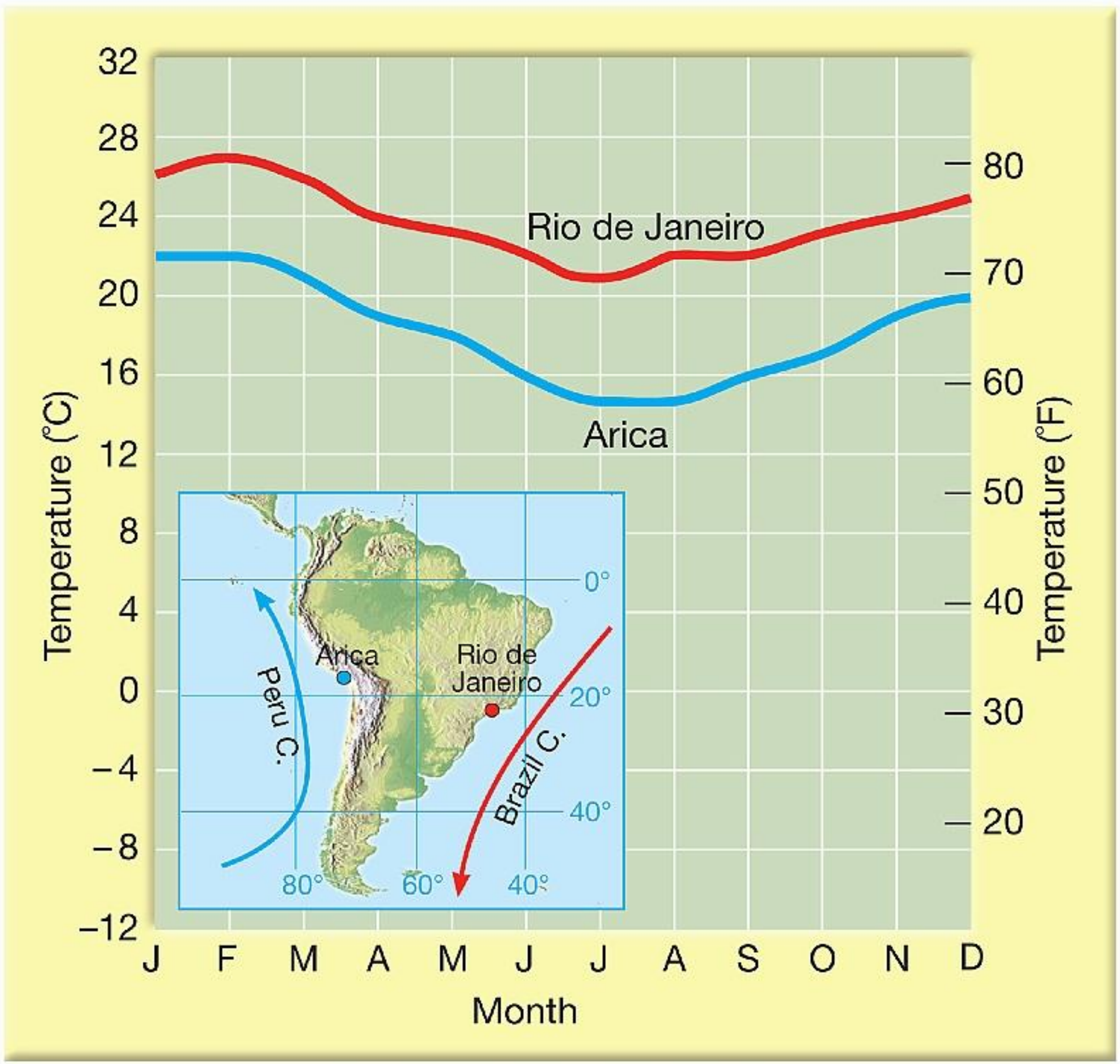

San Antonio
Winnipeg
Temperature (°C)
30
20
10
0
–10
–20
–30
Temperature (°F)
80
60
40
20
0
–20
J F M A M J J A S O N D
Month


SAN ANTONIO
WINNIPEG
June 21
84°
1.01
June 21
63.5°
1.12
Dec. 21
37°
1.66
Dec. 21
16.5°
3.52

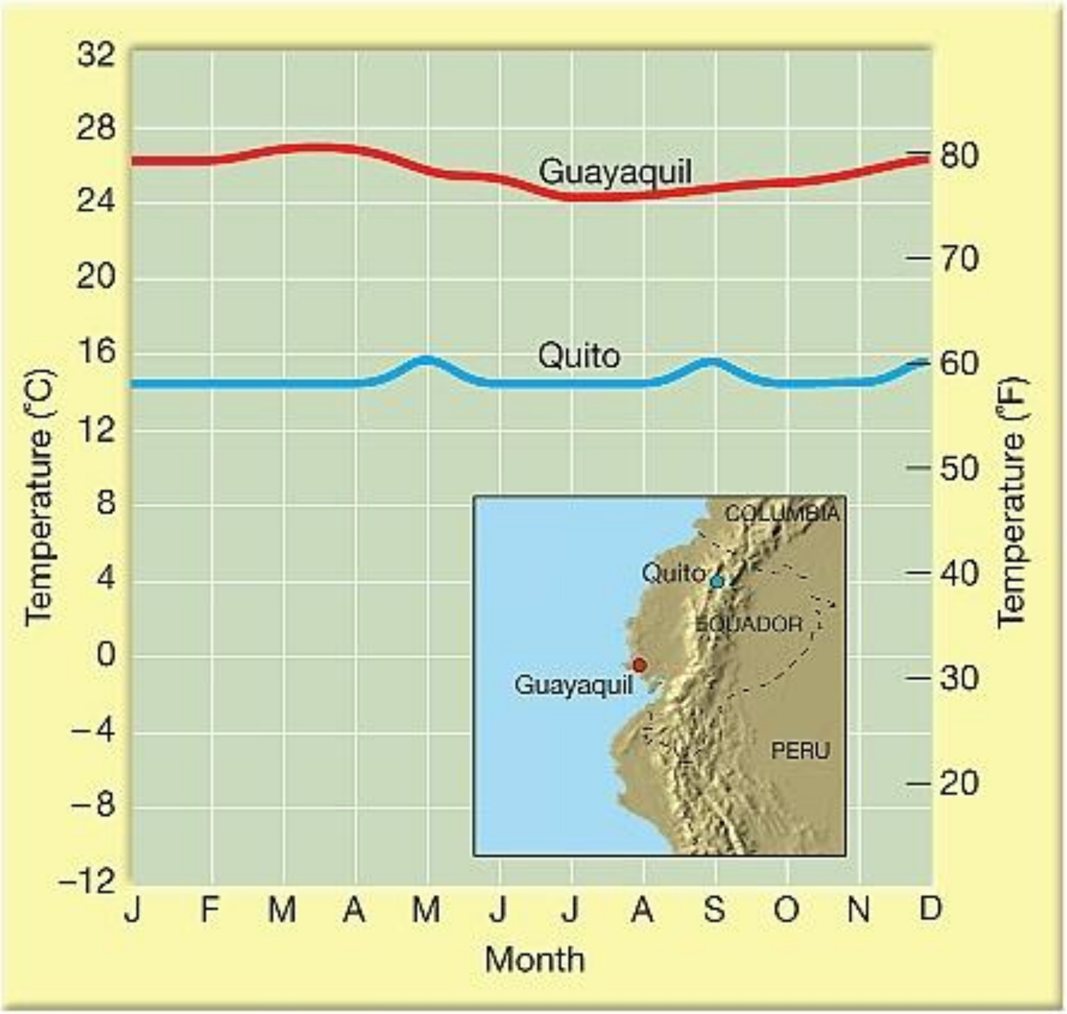
Land and Water



Ocean Currents



Altitude

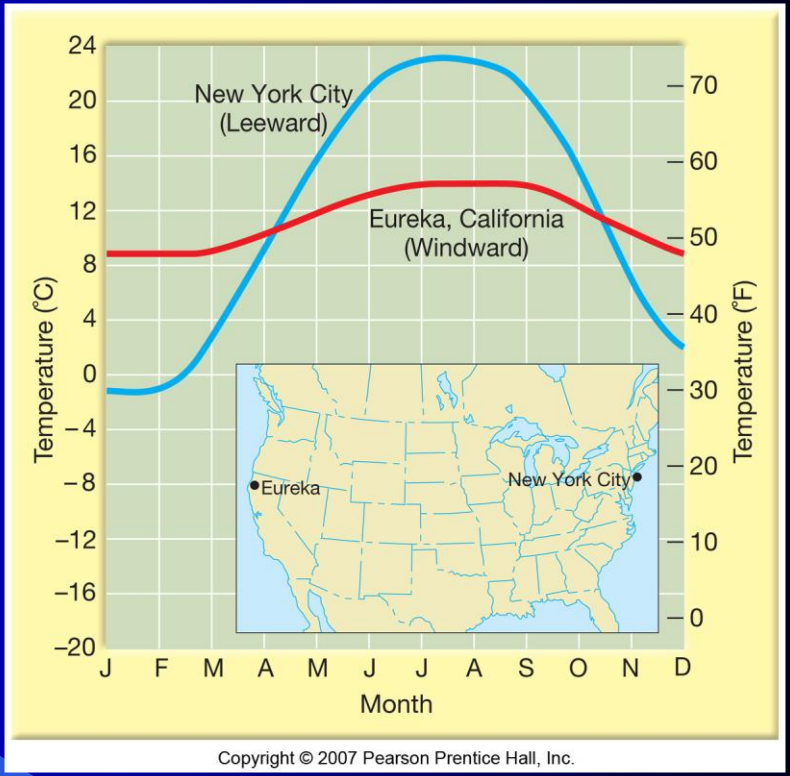


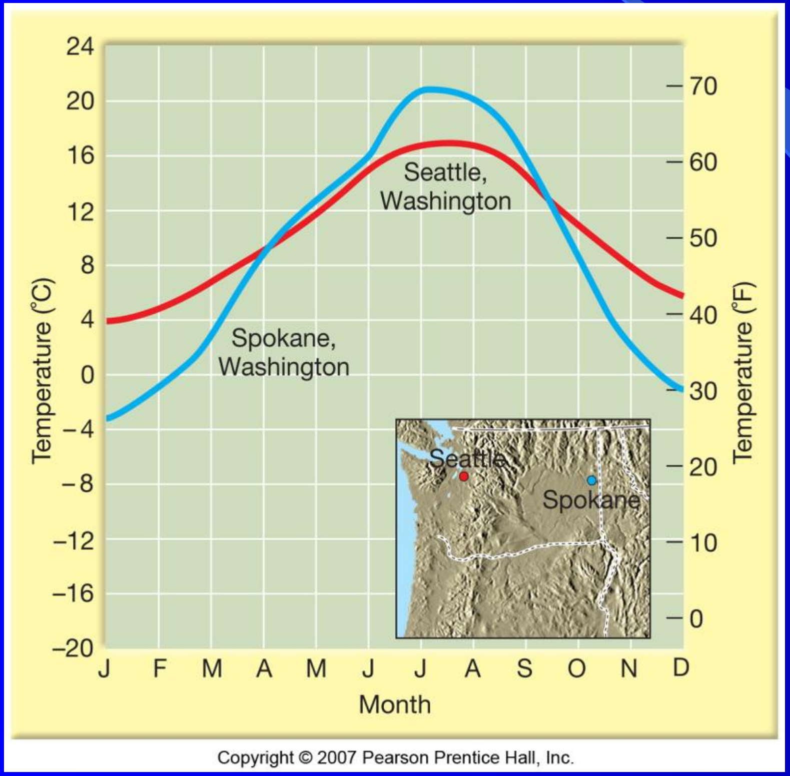
(a) **Temperature graph**

(b) **Quito is high in the Andes**

(c) **Guayaquil is at sea level**

Windward coasts are affected by moderating winds off the sea; Leeward coasts are not.

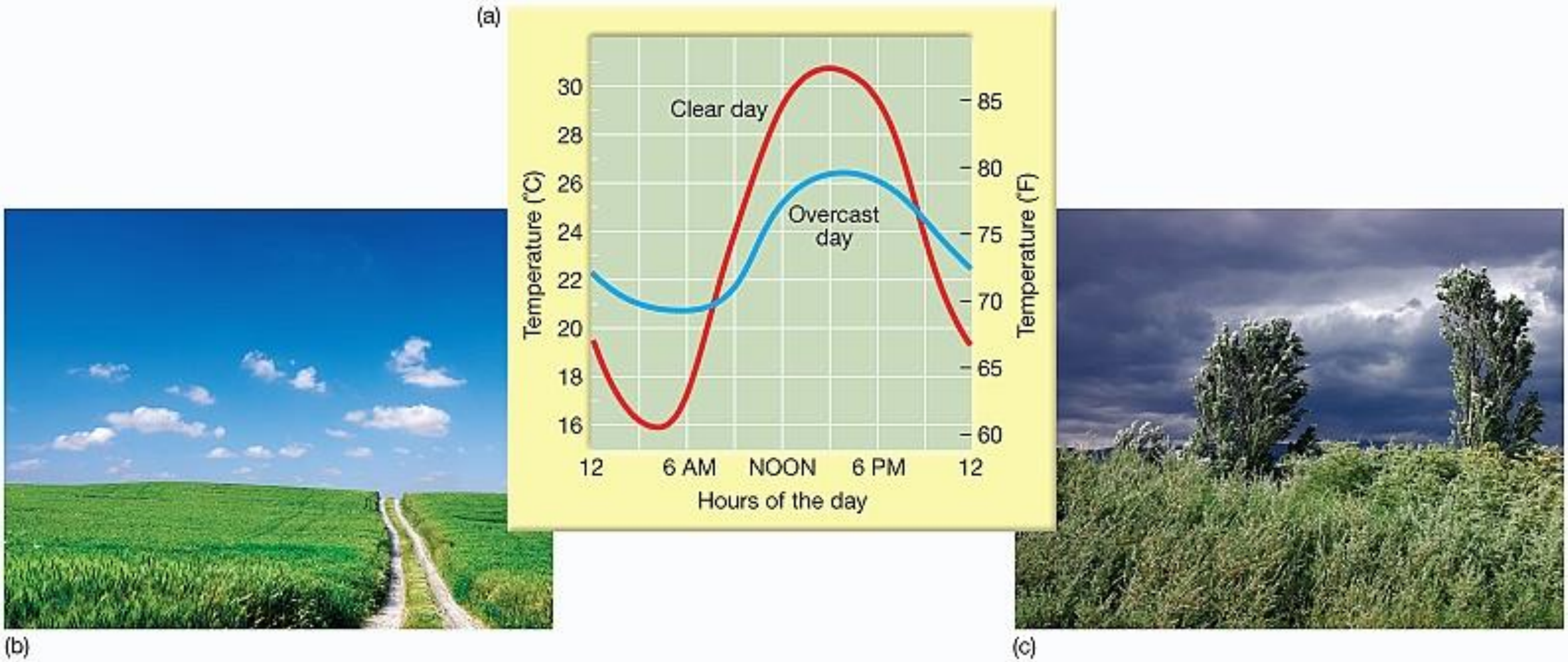




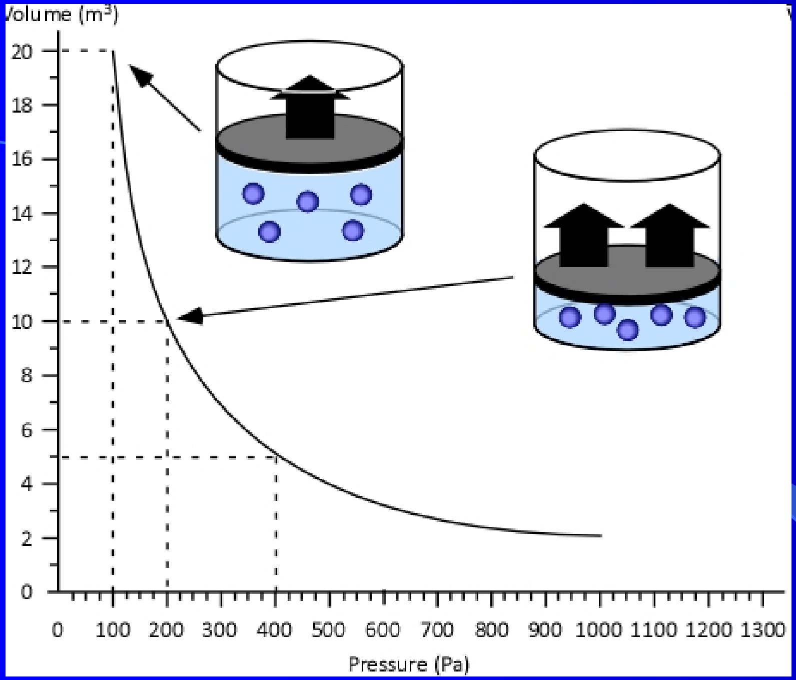
Mountain ranges can act as barriers

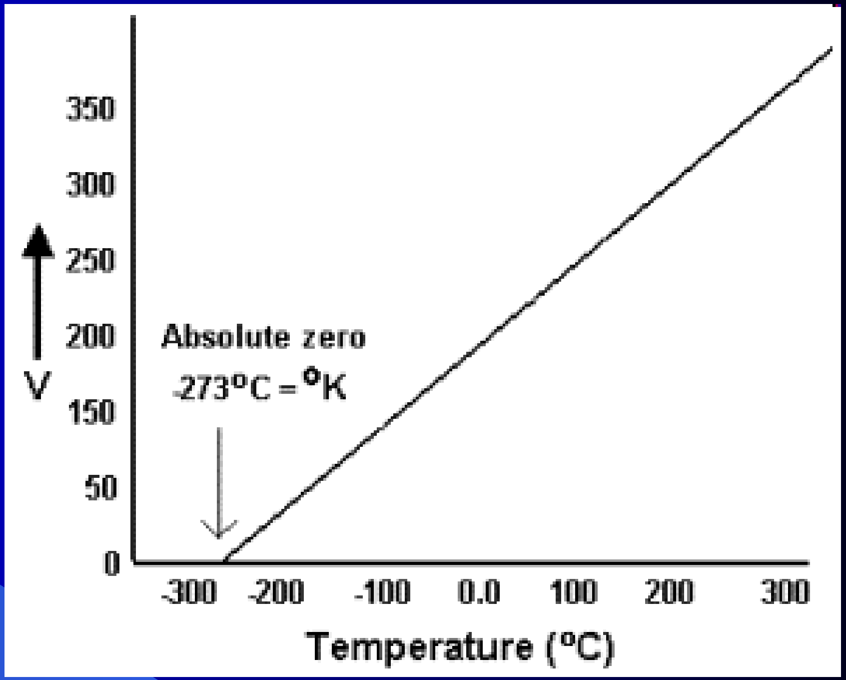
Cloud Cover & Albedo

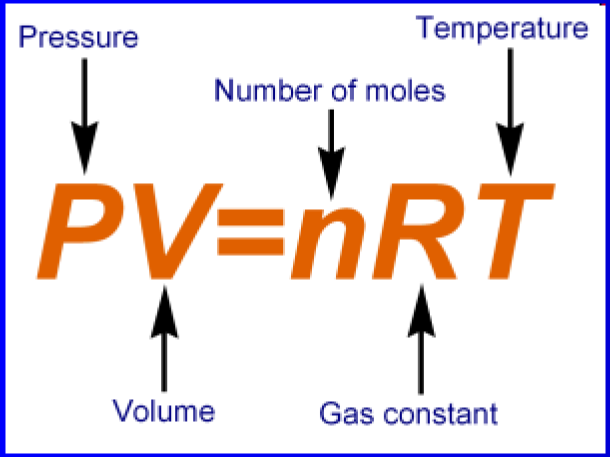
- Clouds both reflect and trap radiation
- Surface **albedo** controls how much incident radiation is absorbed



Ideal Gas Law







P = ρRT or Pα = RT

α = 1/ρ

α= specific volume

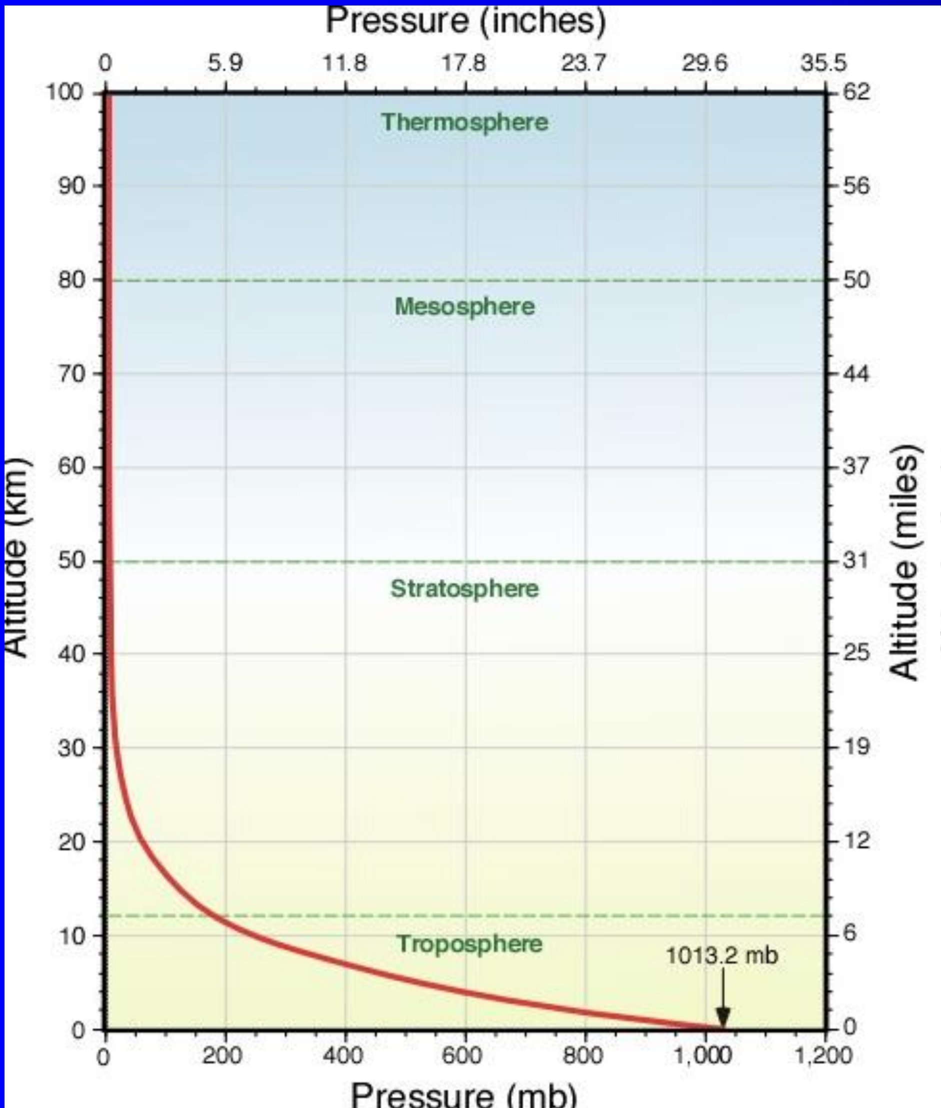
Atmospheric Pressure

Air pressure is the force exerted by the weight of the overlying air, i.e., force/unit area

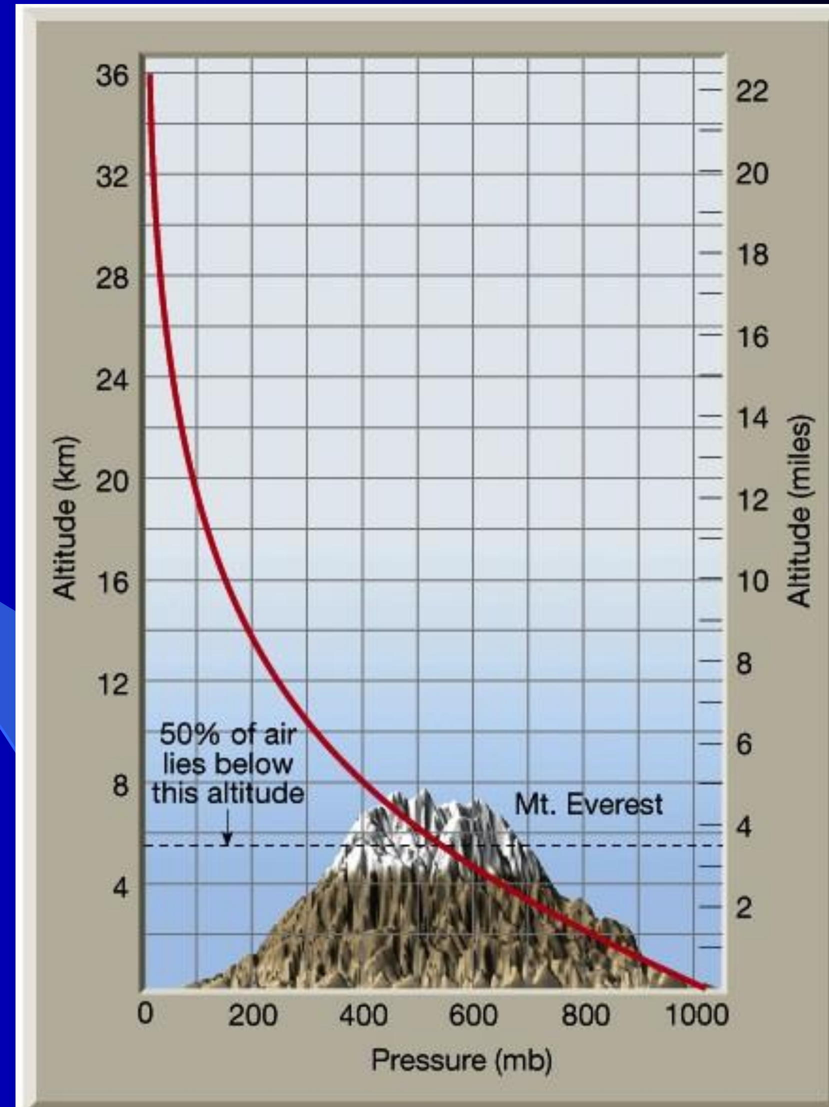
- Air pressure decreases smoothly with altitude
- Because air is highly compressible, 50% of the of the atmosphere lies below 5.5 km, 99% lies below 32 km, the remaining 1% extends from 32 km to 500 km
- Barometers are used to measure air pressure. There are various types of barometers such as mercury barometers or aneroid barometers.

Measuring air pressure

- ***Units*** – Pascals, millibars, mm of mercury, atmospheres
- One pascal (Pa) = 1 N/m^2
- One millibar (mb) = 1×10^3 $dynes/cm^2$ = 100 Pa
- One atmosphere (Atm) = 101,325 N/m^{2} = 1012.35 mb



Pressure (inches)
0
5.9
11.8
17.8
23.7
29.6
35.5
Thermosphere
Mesosphere
Stratosphere
Troposphere
1013.2 mb
Altitude (km)
100
90
80
70
60
50
40
30
20
10
0
Altitude (miles)
62
56
50
44
37
31
25
19
12
6
0
0
200
400
600
800
1,000
1,200
Pressure (mb)



Altitude (km)
36
32
28
24
20
16
12
8
4
Altitude (miles)
22
20
18
16
14
12
10
8
6
4
2
50% of air
lies below
this altitude
Mt. Everest
0
200
400
600
800
1000
Pressure (mb)

Table 6–1 U.S. standard atmosphere

Height (KM)	Pressure (MB)	Temperature (°C)
50.0	0.798	–2
40.0	2.87	–22
35.0	5.75	–36
30.0	11.97	–46
25.0	25.49	–51
20.0	55.29	–56
18.0	75.65	–56
16.0	103.5	–56
14.0	141.7	–56
12.0	194.0	–56
10.0	265.0	–50
9.0	308.0	–43
8.0	356.5	–37
7.0	411.0	–30
6.0	472.2	–24
5.0	540.4	–17
4.0	616.6	–11
3.5	657.8	–8
3.0	701.2	–4
2.5	746.9	–1
2.0	795.0	2
1.5	845.6	5
1.0	898.8	9
0.5	954.6	12
0	1013.2	15

Calculation of Atmospheric Pressure

The basis of the calculation is the hydrostatic equation

$$P = \rho g h$$

P = pressure, ρ = density, g = 9.8 m s^{-2}, h = height (or depth)

$$P = P_o \, e^{-\frac{gh}{RT}}$$

P = pressure
P_o = initial pressure
g = acceleration of gravity (9.8 m s^{-2})
h = height (m)
R = gas constant (287 J kg^{-1} K^{-1})
T = temperature (K)

Types of thermodynamic processes

- Isothermal – occur at constant temperature
- Isobaric – occur at constant pressure
- Adiabatic – occur at constant heat, i.e. there is no heat exchange with the surrounding environment.

Sounding curve – the measured variation of atmospheric temperature with altitude. Determined worldwide at 12 midnight and 12 noon GMT.

Lapse rate – change in temperature as a function of change in elevation. There are various lapse rates – environmental, dry adiabatic, wet adiabatic.

Heat and Temperature Changes

For solids and liquids

$\Delta H = cm\Delta T$

c = specific heat, m = mass, T in ^{o}C

For gases

$\Delta H = Cv\Delta T + p\Delta\alpha$ or $\Delta H = Cp\Delta T - \alpha\Delta p$

In the atmosphere processes are adiabatic

$0 = Cp\Delta T - \alpha\Delta p$

$\Delta T = (\alpha/Cp)\Delta p$

Substituting for $\Delta p = -(Pg/RT)\Delta h$ and $\alpha = (RT)/P$

$\Delta T/\Delta h = - g/Cp = 0.98\ ^{o}C/100$ m for dry air

$Cp = Cv + R$

For dry air

$Cp = 1003$ J/kg K
$Cv = 717$ J/kg K
$R = 287$ J/kg K

Process and Sounding Curves

- Process curve represents the variation with time of pressure and temperature for a single air parcel
- Sounding curve represents the measured temperatures and pressures of different parcels of air at various heights in the atmosphere at one particular time.

Lapse Rates

- Environmental lapse rate = $\gamma = -(\Delta T/\Delta h)_{observed}$
- Adiabatic lapse rate = $\Gamma = -(\Delta T/\Delta h)_{adiabatic\ process}$

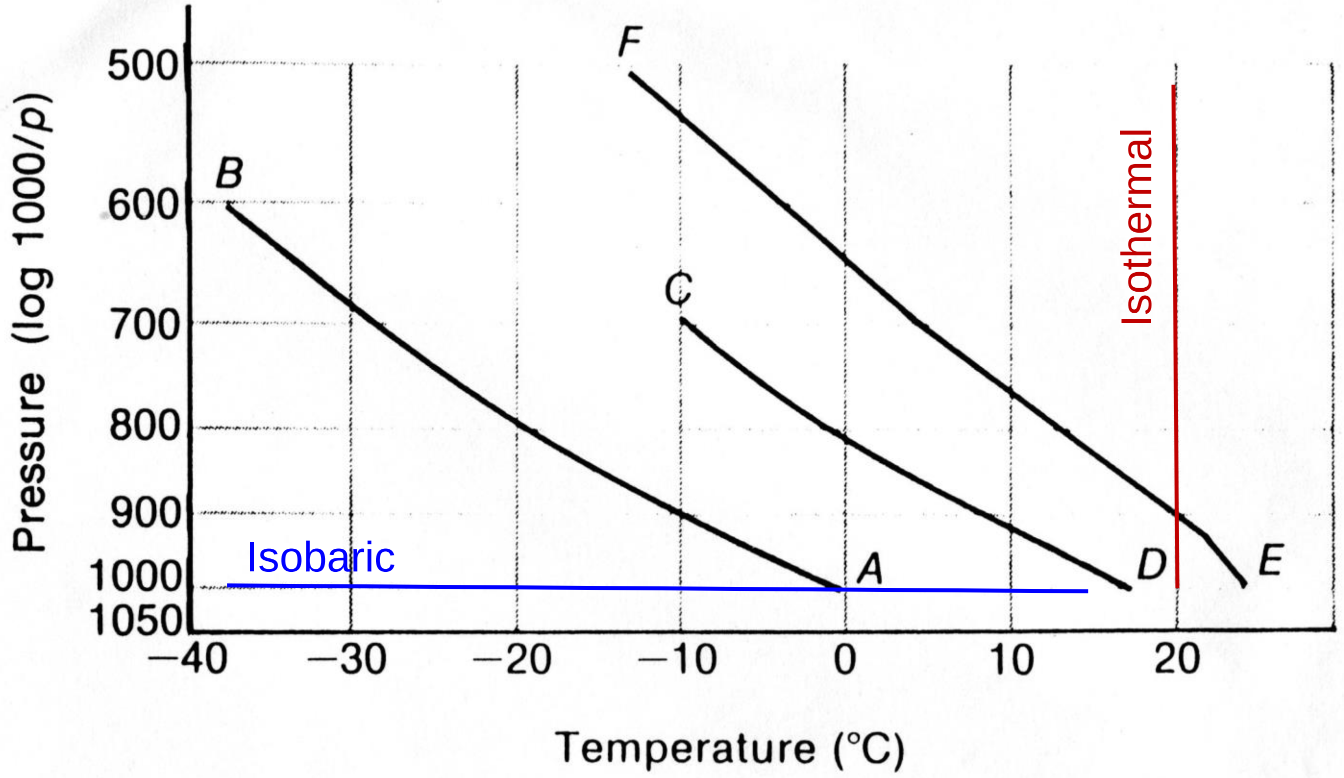


Figure 4.5 Thermodynamic diagram: *AB* and *CD* are process curves representing the change in temperature with adiabatic change in pressure; *EF* is the sounding curve representing the variation of temperature with pressure in the air above Washington, D.C., 7:00 P.M., July 1, 1957.

Potential temperature – the temperature an air parcel would have if it was adiabatically moved to 1000 mb.

Atmospheric Stability

Types of equilibrium:

- Stable – if air parcel is displaced forces arise that cause it to return to its original position

  Stable equilibrium: $\gamma < \Gamma$

- Neutral – no force arises from the displacement

  Neutral equilibrium: $\gamma = \Gamma$

- Unstable – the displacement gives rise to forces that tend to increase the displacement

  Unstable equilibrium: $\gamma > \Gamma$

- Inversion – temperature increases with height

  Types – radiation, subtropical, frontal

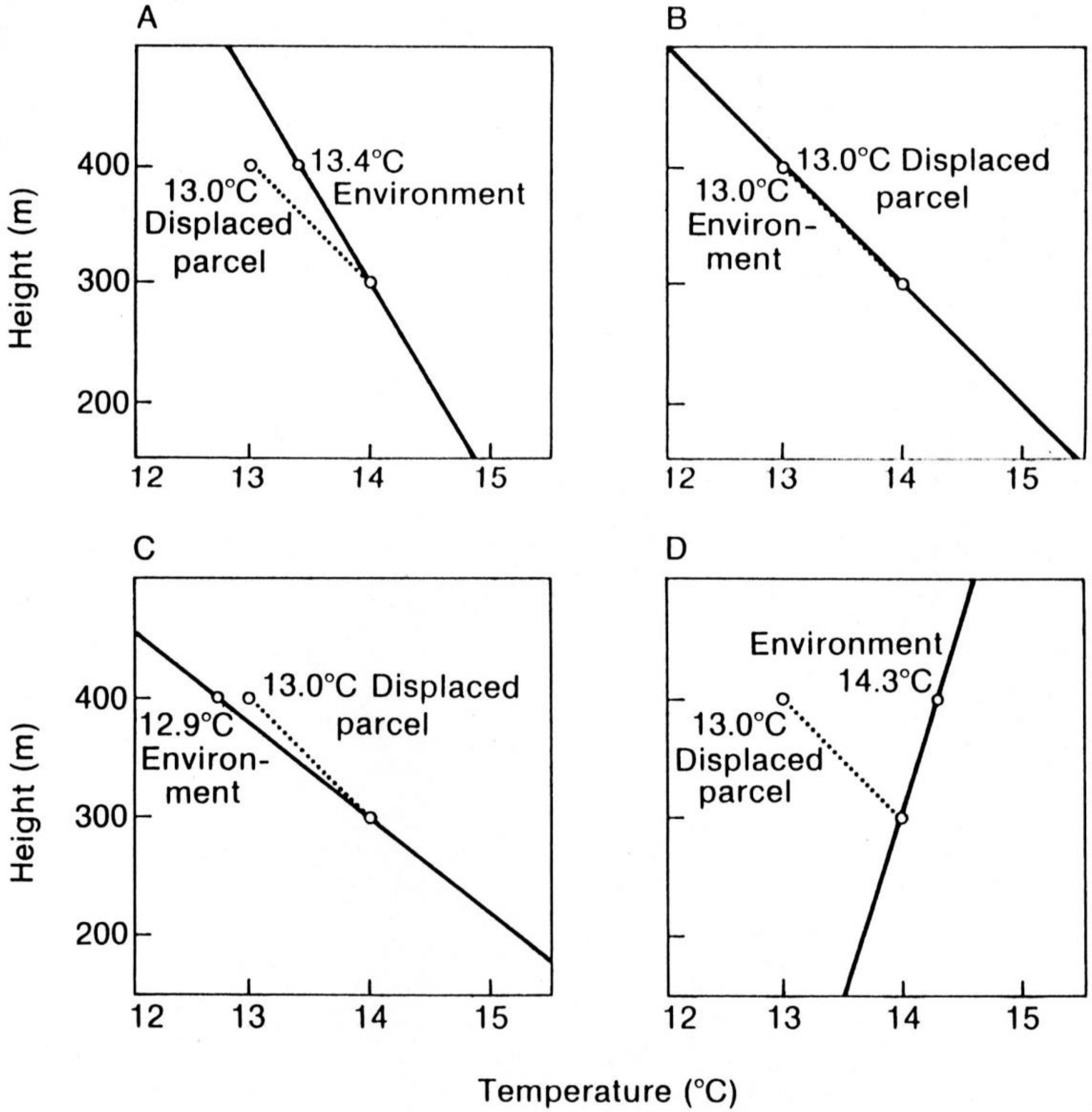
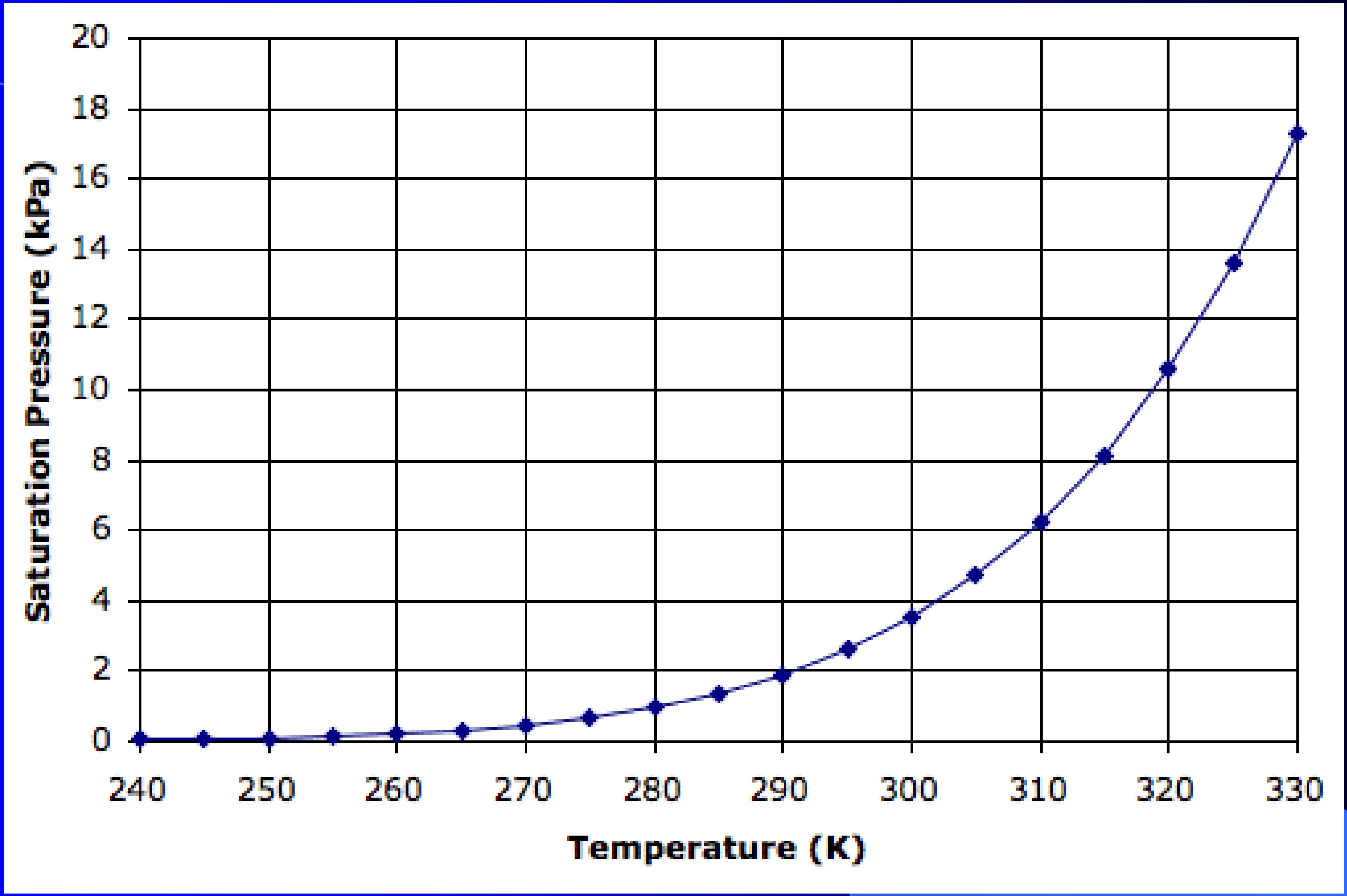


Figure 4.7 Examples of stability. The solid lines are sounding curves and the dotted lines process curves for four schematic cases: (A) normal lapse rate (slightly stable); (B) adiabatic lapse rate (neutral equilibrium); (C) superadiabatic lapse rate (unstable); (D) inversion (very stable).

- Absolute humidity
  - The actual amount of water vapor present in the atmosphere
  - Can be measured in terms of density of water vapor or partial pressure (P_{H2O})
- Relative humidity
  - When the number of molecules that evaporate equals the number that condense, the vapor is saturated, this is the dew point temperature
  - The amount of water vapor in under-saturated air is the relative humidity
  - This is the ratio of the actual vapor pressure to the saturation vapor pressure
  - Temperature exerts a strong control on water vapor capacity of air
- Dewpoint
  - Temperature at which the air, without changing its moisture content, would saturated with water vapor

Variation in saturation vapor pressure as a function of temperature. Warm air can hold more water vapor than cold air.



$$\text{Mixing Ratio} = w = \frac{mass\ of\ water\ vapor}{mass\ of\ dry\ air + water\ vapor}$$

$$\text{Relative Humidity (\%)} = \frac{w}{w_s} \cdot 100 = \frac{measured\ mixing\ ratio}{saturated\ mixing\ ratio} \cdot 100$$

w = amount of water vapor present determined from the dewpoint temperature
w_s = amount of water vapor the air could hold at the measured temperature

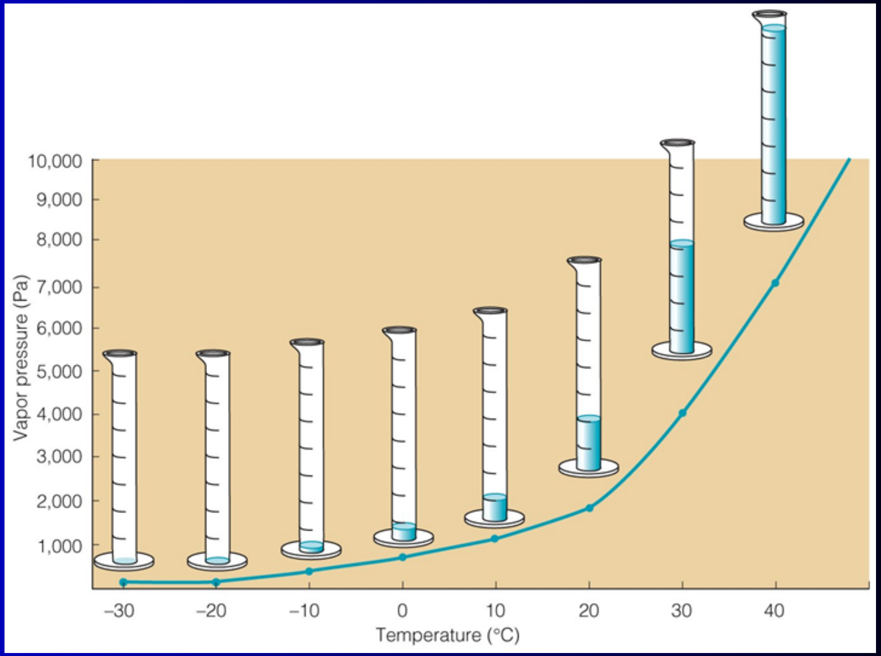
Air Temp = 20°C Dewpoint Temp = 10°C

$$\text{RH} = \frac{7.76}{15.0} \cdot 100 = 52\%$$

SATURATION MIXING RATIO (gm kg^{-1}) OVER WATER AT 1000 MB

	Temperature									
	0	1	2	3	4	5	6	7	8	9
40	49.8									
30	27.7	29.4	31.2	33.1	35.1	32.3	39.5	41.9	44.4	47.0
20	15.0	15.9	17.0	18.1	19.2	20.4	21.7	23.1	24.6	26.1
10	7.76	8.31	8.88	9.49	10.14	10.83	11.56	12.34	13.16	14.03
0	3.84	4.13	4.44	4.77	5.12	5.50	5.89	6.32	6.77	7.25
−0	3.84	3.57	3.31	3.08	2.85	2.64	2.45	2.27	2.10	1.94
−10	1.794	1.658	1.529	1.410	1.300	1.197	1.110	1.013	0.931	0.855
−20	0.785	0.720	0.659	0.604	0.552	0.505	0.461	0.421	0.384	0.350
−30	0.318	0.289	0.263	0.239	0.217	0.196	0.178	0.161	0.145	0.131

Relative humidity can be changed either by the addition of water vapor or by a change in temperature. A fog can form when the air becomes saturated in water vapor.



Types of Fogs:

- Warm frontal fog – warm rain falling through cold air saturates the colder air (addition of water vapor).
- Steam fog – cold air blowing over warm water becomes saturated (addition of water vapor).
- Radiation fog – cooling of air near ground level on cold clear night with light winds leads to saturation (decrease in temperature).
- Advection fog – warm air carried over a colder surface resulting in saturation (decrease in temperature).

Saturation Adiabatic Process

Saturation Adiabatic lapse rate (Γs) – lapse rate when condensation is occurring. Energy released by the condensation of water vapor is added to the parcel. This results in a smaller lapse rate.

Latent heat of vaporization – heat required to go from liquid to vapor

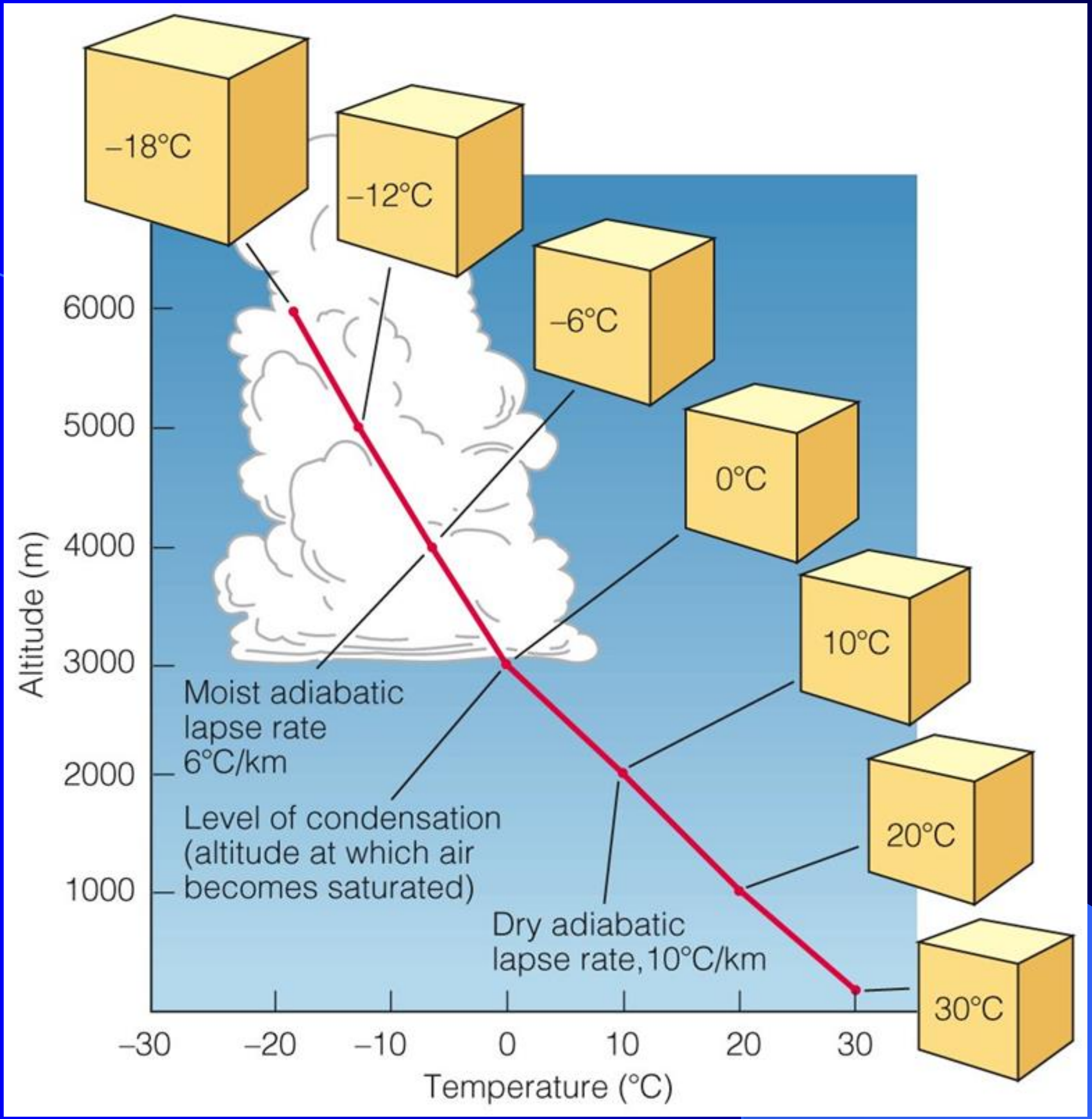
$L = (596 – 0.56T)$ cal g^{-1} or $L = (2493664 – 2343T) =$ J kg^{-1}

T in degrees Celsius

Latent heat of sublimation – heat required to go from solid to vapor = 676 calories gm^{-1} = 2828384 J kg^{-1}

Saturation adiabatic lapse rate varies as a function of temperature. At low temperatures it approaches the dry adiabatic lapse rate (because at low T there is very little water vapor in the air).

Wet-bulb potential temperature – the temperature at which a saturation adiabat intersects the 1000 mb isobar.


–18°C
–12°C
–6°C
0°C
10°C
20°C
30°C
6000
5000
4000
3000
2000
1000
Altitude (m)
Moist adiabatic lapse rate 6°C/km
Level of condensation (altitude at which air becomes saturated)
Dry adiabatic lapse rate, 10°C/km
–30
–20
–10
0
10
20
30
Temperature (°C)

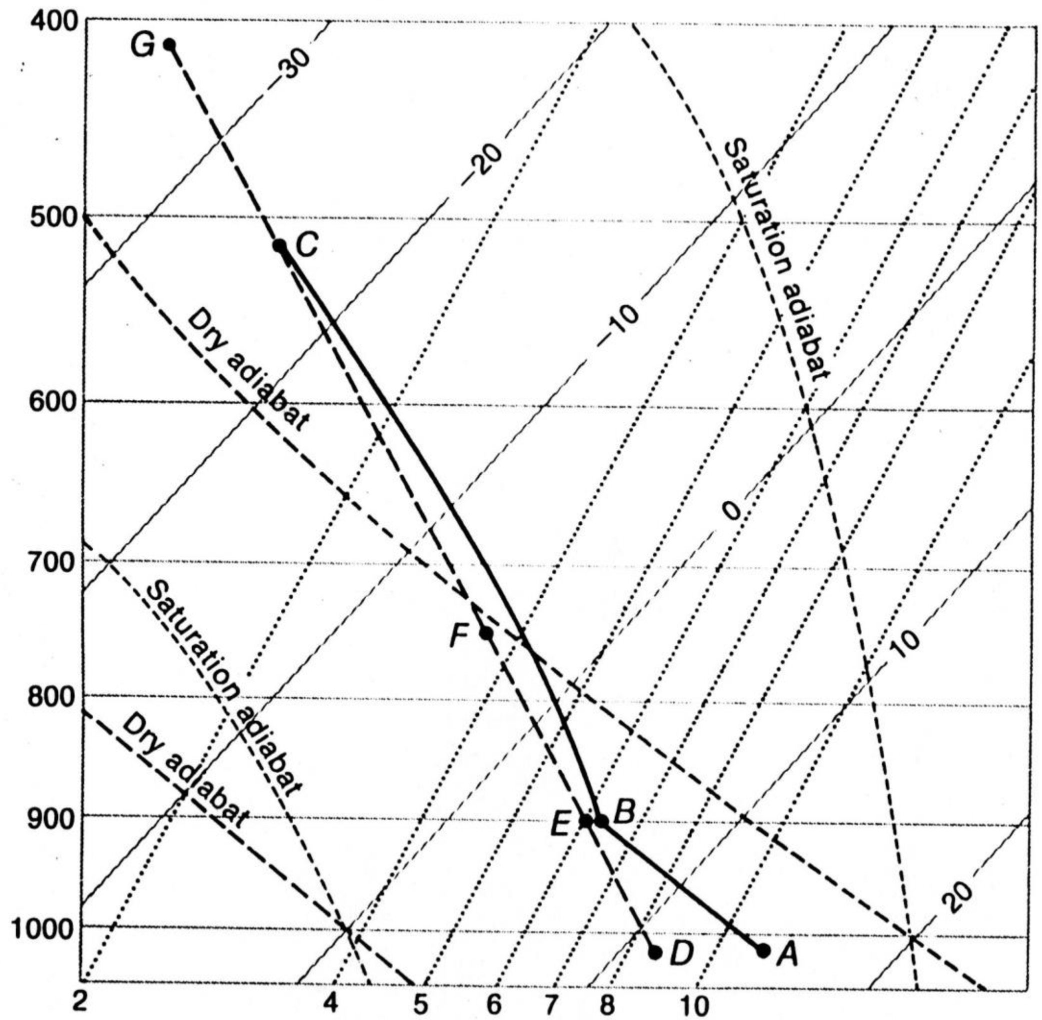


Figure 5.9 Thermodynamic diagram (Skew T–Log P diagram), showing saturation adiabats (*curved dashed lines*) and saturation mixing ratio lines (*dotted lines*). The process that a parcel with an initial temperature of 15°C, a pressure of 1013 mb, and a mixing ratio of $6.1 \cdot 10^{-3}$ would experience if lifted adiabatically is represented by curve *ABC*.

Stability of Cloudy Air

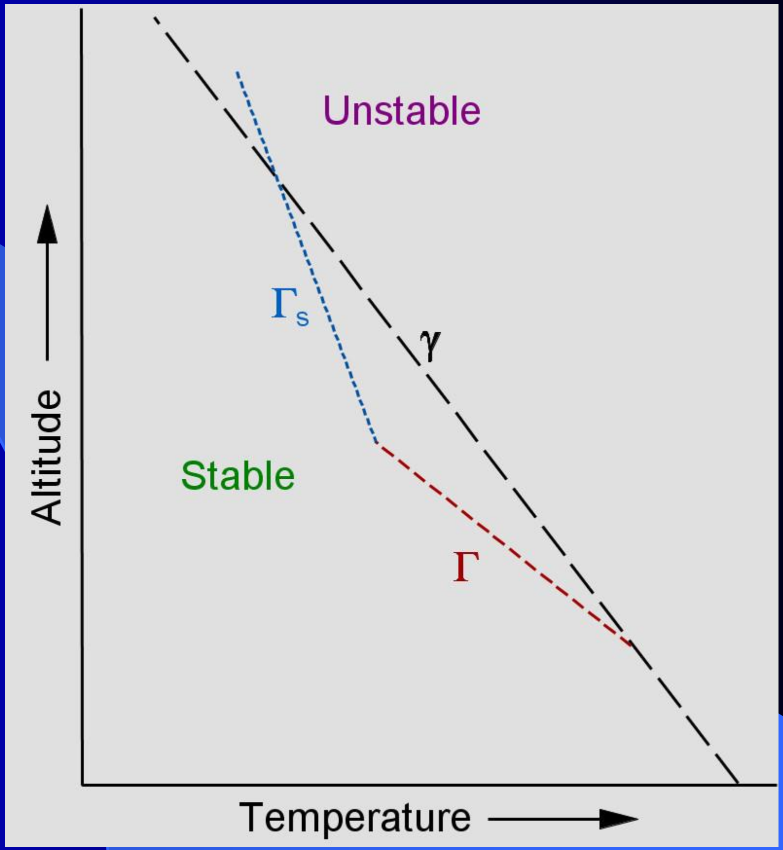
$\gamma < \Gamma s \rightarrow$ stable

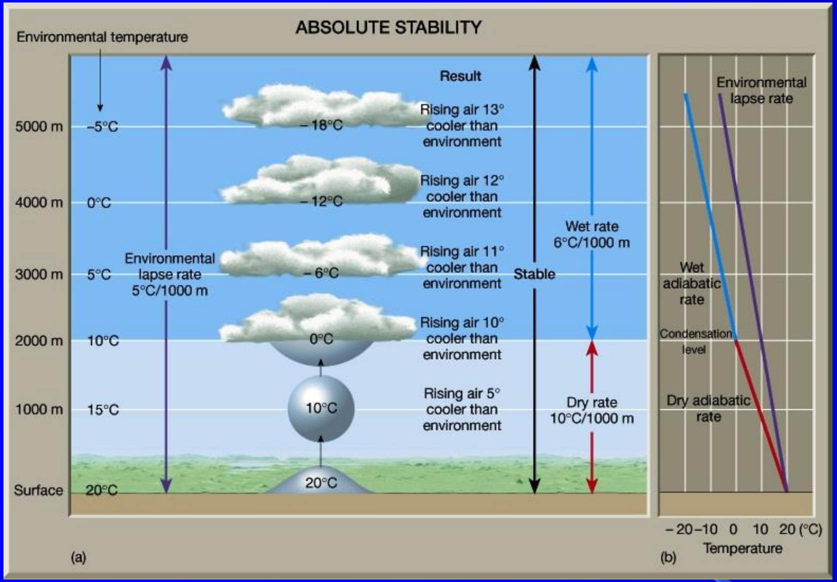
$\gamma = \Gamma s \rightarrow$ neutral

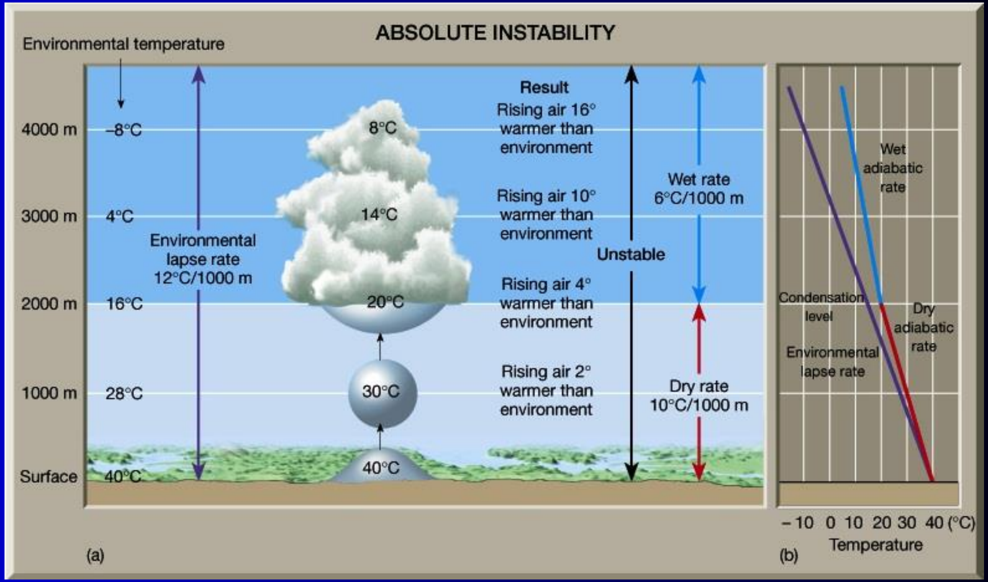
$\gamma > \Gamma s \rightarrow$ unstable

$\Gamma > \gamma < \Gamma s \rightarrow$ conditional instability

Consider the figure to the right. The adiabatic lapse rate is greater than the environmental lapse rate so the atmosphere is stable. If the air parcel is mechanically forced to rise it eventually becomes saturated and cools at the wet adiabatic lapse rate. The wet adiabatic lapse rate is less than the environmental lapse rate. If the air continues to rise mechanically (for example being pushed up the side of a mountain) a point is reached where the air parcel becomes warmer than the surrounding air and the atmosphere becomes unstable.

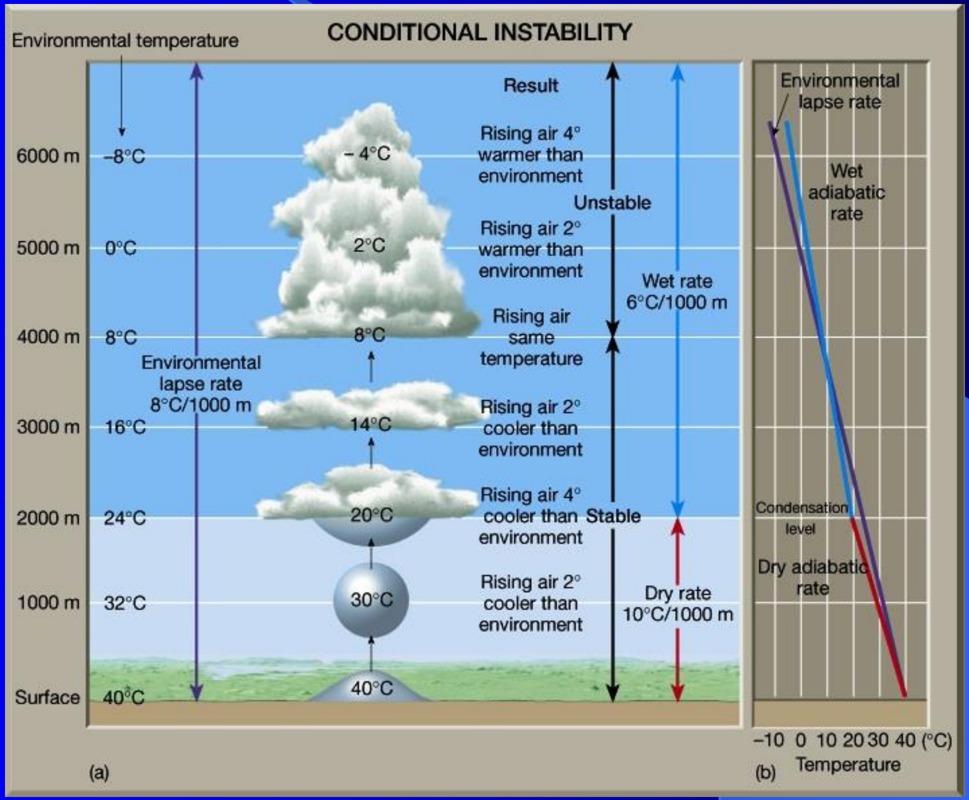






$\Gamma > \Gamma_s > \gamma$

$\Gamma < \Gamma_s < \gamma$



$\Gamma > \gamma > \Gamma_s$

Clouds and cloud formation

- Clouds are visible aggregates of minute water droplets, ice crystals, or both

- They form when air rises and becomes saturated with moisture in response to adiabatic cooling and condensation

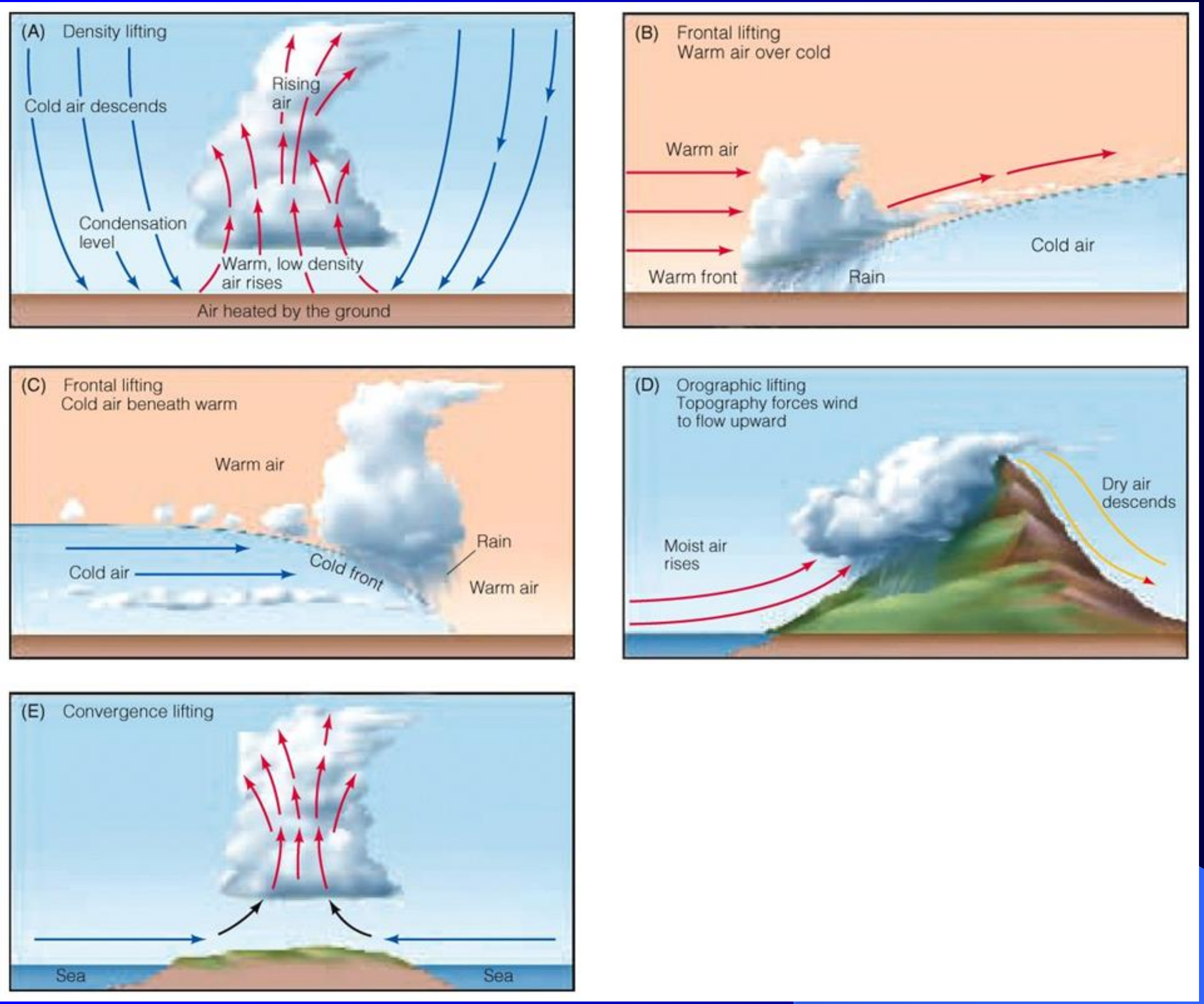
- There are four principal reasons for the upward movement of air, which in turn leads to the formation of clouds

1. Density lifting - Warm, low-density air rises convectively

2. Frontal lifting - Two flowing air masses of different density meet, one forcing the other up

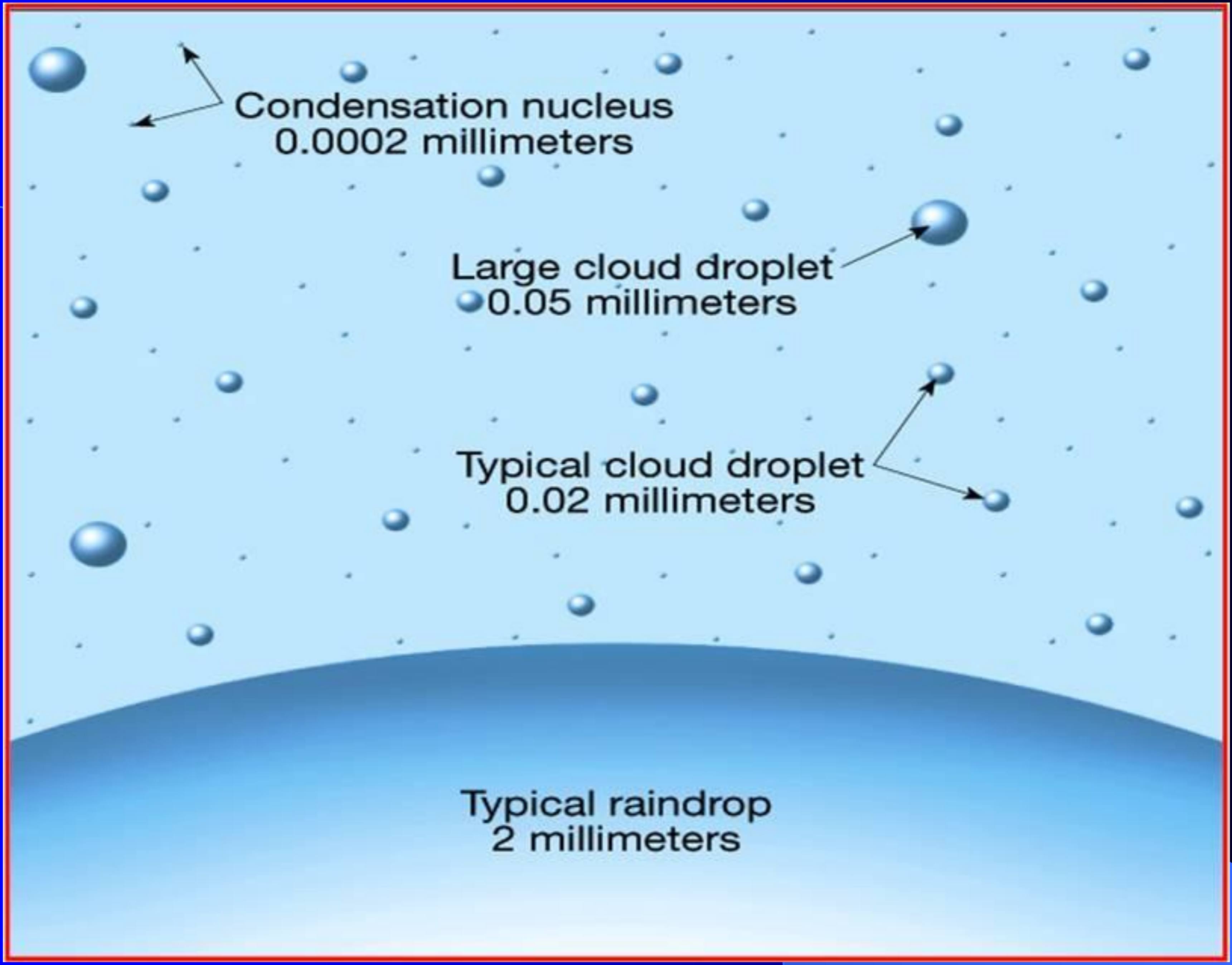
3. Orographic lifting - Flowing air is forced upward due to terrain

4. Convergence lifting - Flowing air masses converge and are both forced upward

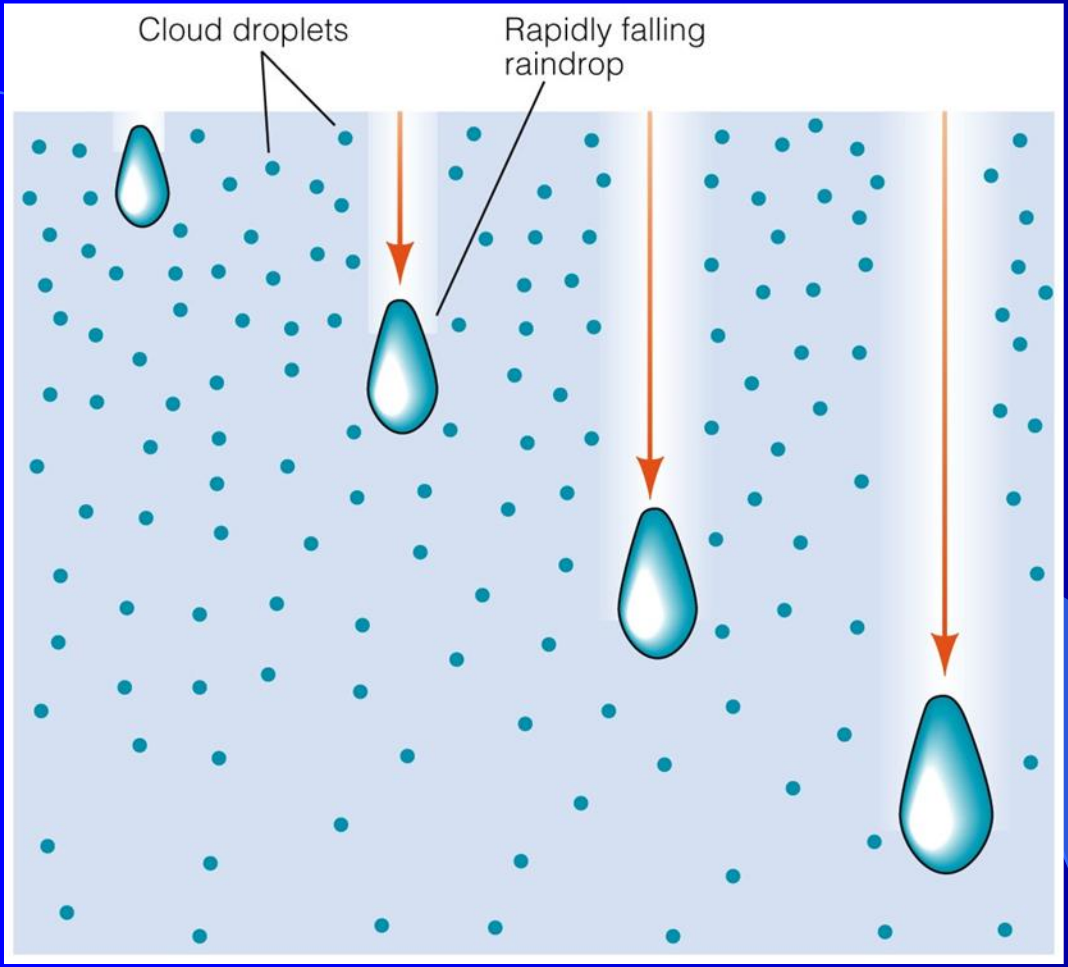


(A) Density lifting
Cold air descends
Rising air
Condensation level
Warm, low density air rises
Air heated by the ground
(B) Frontal lifting
Warm air over cold
Warm air
Cold air
Warm front
Rain
(C) Frontal lifting
Cold air beneath warm
Warm air
Cold air
Cold front
Rain
Warm air
(D) Orographic lifting
Topography forces wind to flow upward
Moist air rises
Dry air descends
(E) Convergence lifting
Sea
Sea

- When the dew point is reached one of two things happens
  - Water condenses
  - Ice crystals form
- These nucleation processes require energy to form a new surface
  - Nucleation sites may be
    - The ground, aerosols


Condensation nucleus
0.0002 millimeters
Large cloud droplet
0.05 millimeters
Typical cloud droplet
0.02 millimeters
Typical raindrop
2 millimeters

Collision-coalescence (warm cloud)

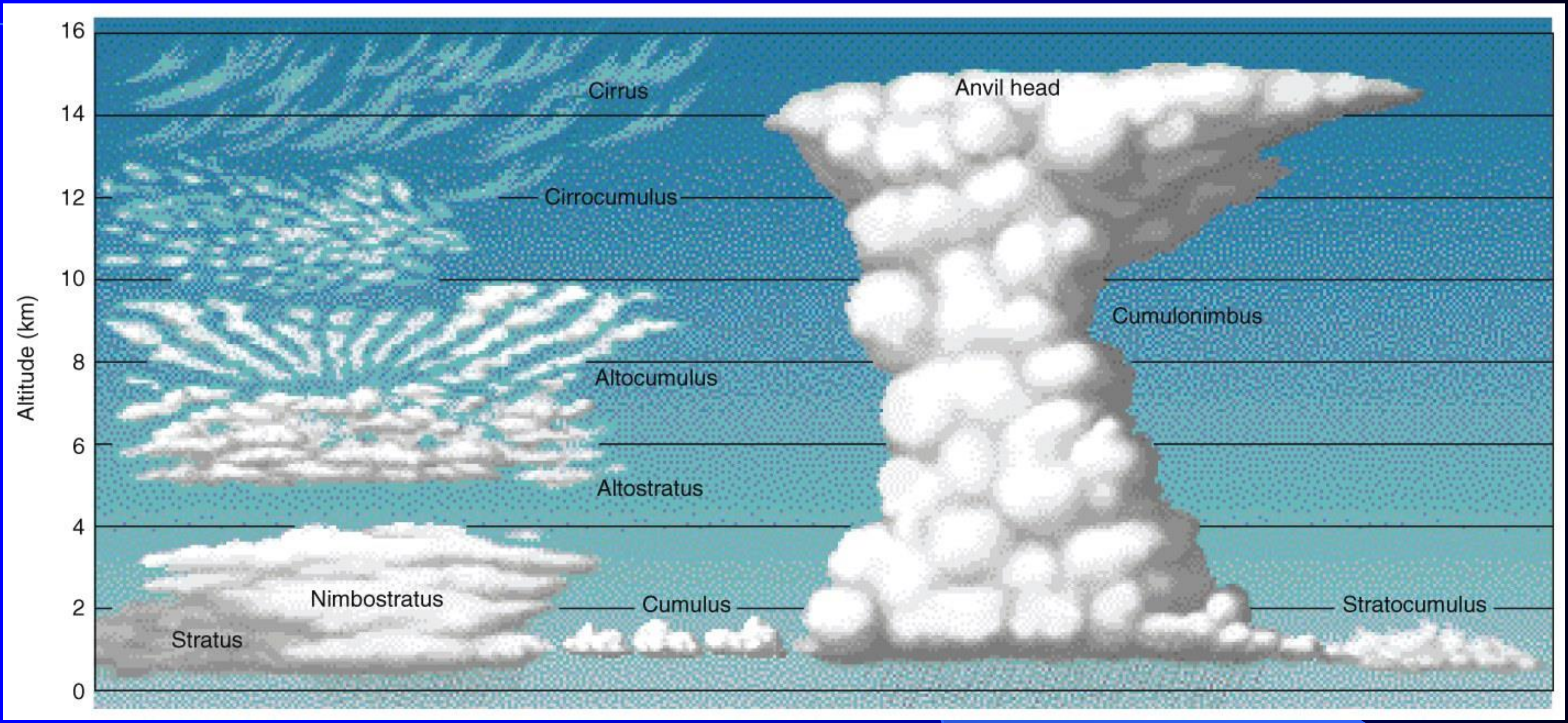


Bergeron process

Table 5–4 Types of precipitation

Type	Approximate Size	State of Water	Description
Mist	0.005 to 0.05 mm	Liquid	Droplets large enough to be felt on the face when air is moving 1 meter/second. Associated with stratus clouds.
Drizzle	Less than 0.5 mm	Liquid	Small uniform drops that fall from stratus clouds, generally for several hours.
Rain	0.5 to 5 mm	Liquid	Generally produced by nimbostratus or cumulonimbus clouds. When heavy, size can be highly variable from one place to another.
Sleet	0.5 to 5 mm	Solid	Small, spherical to lumpy ice particles that form when raindrops freeze while falling through a layer of subfreezing air. Because the ice particles are small, any damage is generally minor. Sleet can make travel hazardous.
Glaze	Layers 1 mm to 2 cm thick	Solid	Produced when supercooled raindrops freeze on contact with solid objects. Glaze can form a thick coating of ice having sufficient weight to seriously damage trees and power lines.
Rime	Variable accumulations	Solid	Deposits usually consisting of ice feathers that point into the wind. These delicate frostlike accumulations form as supercooled cloud or fog droplets encounter objects and freeze on contact.
Snow	1 mm to 2 cm	Solid	The crystalline nature of snow allows it to assume many shapes, including six-sided crystals, plates, and needles. Produced in supercooled clouds where water vapor is deposited as ice crystals that remain frozen during their descent.
Hail	5 mm to 10 cm or larger	Solid	Precipitation in the form of hard, rounded pellets or irregular lumps of ice. Produced in large convective, cumulonimbus clouds, where frozen ice particles and supercooled water coexist.
Graupel	2 mm to 5 mm	Solid	Sometimes called "soft hail," graupel forms as rime collects on snow crystals to produce irregular masses of "soft" ice. Because these particles are softer than hailstones, they normally flatten out upon impact.

- Clouds are classified on the basis of shape, appearance, and height
  - Cumulus: puffy individual clouds, where the flat base marks the condensation level
  - Stratus: sheets of cloud cover spread laterally rather than vertically
  - Cirrus: highest of the clouds, wispy feathers composed of ice crystals
  - Nimbus: rain, as in cumulonimbus



16
14
12
10
8
6
4
2
0
Altitude (km)
Cirrus
Cirrocumulus
Altocumulus
Altostratus
Nimbostratus
Stratus
Cumulus
Anvil head
Cumulonimbus
Stratocumulus