Radiation and its properties

Textbooks and web sites references for this lecture:

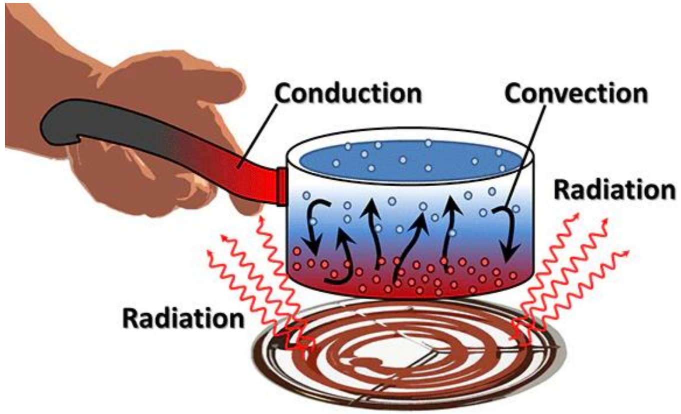
- Robert McIlveen, Fundamentals of Weather and Climate, Chapman & Hall, 1995, ISBN 0-412-41160-1 (§ 2.7)
- Joseph M. Moran e Michael D. Morgan, Meteorology, The Atmosphere and the Science of Weather, Mc Millan College Publishing Company, 1994, ISBN 0-02-383341-6 (§ 2)

Heat transfer processes

- *Conduction* - Where molecules transfer energy by coming into contact with one another.

- *Convection* - Where a fluid moves from one place to another, carrying it's heat energy with it.
  - In atmospheric science, convection is usually associated with vertical movement of the fluid (air or water).
  - Advection is the horizontal component of the classical meaning of convection.

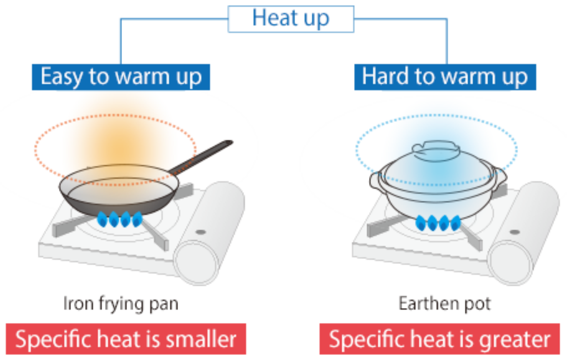
- *Radiation* - The transfer of heat by radiation does not require contact between the bodies exchanging heat, nor does it require a fluid between them.



Other important thermodynamic concepts

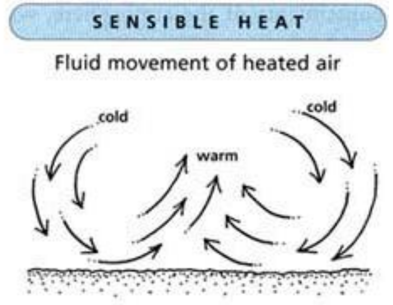
Heat capacity

- amount of heat added to a substance
- change in temperature
- (e.g., water has a higher heat capacity than air)



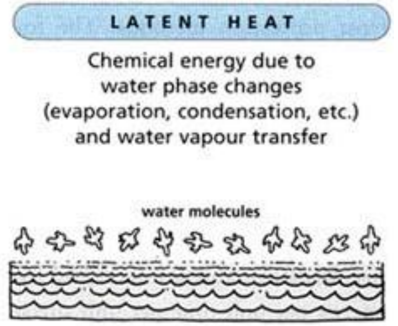
Sensible heat

- The heat that can be measured by a thermometer.

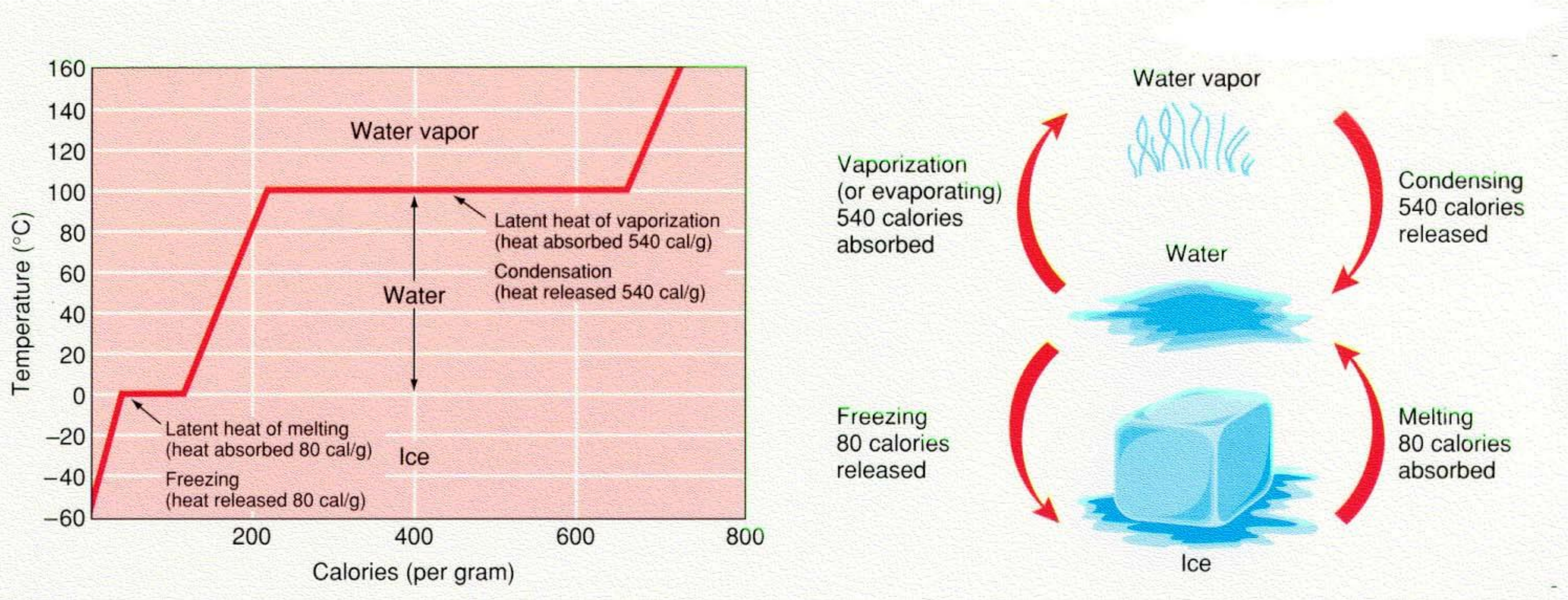


Latent heat

- Heat required to change a substance from one state to another.
- (e.g., heat can be added/removed from a substance without its temperature changing)



Latent heat



Latent heat- heat that goes into a system that causes a change in state rather than an increase in temperature

Energy of a Photon

- $E = h\nu$ or hc/λ

  Where:

  E= energy of a photon

  ν = frequency (1/sec)

  h = Planck's constant (6.63 X 10^{-34} kgm^2/sec)

  c = speed

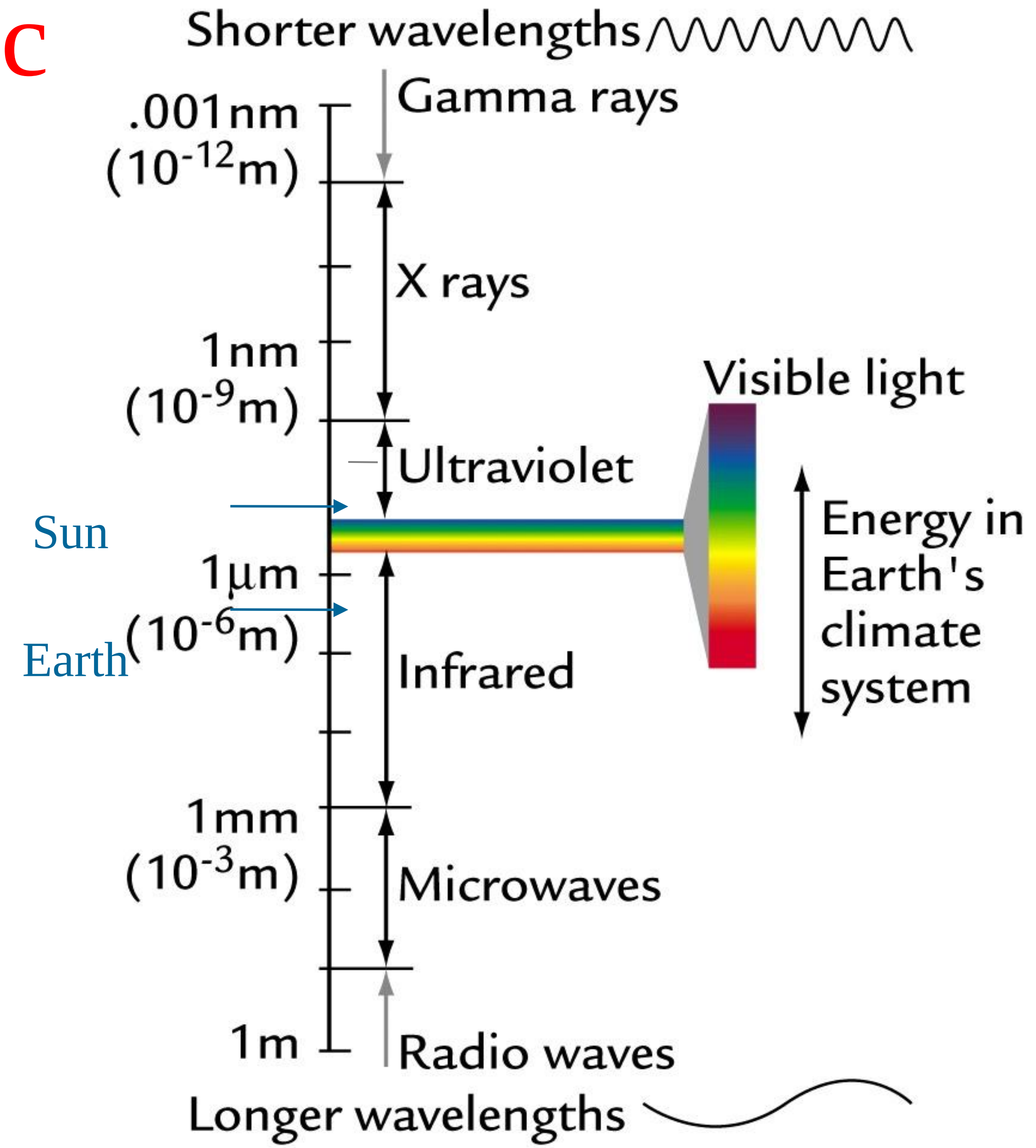
  λ = wavelength

- Thus shortwave & longwave radiation

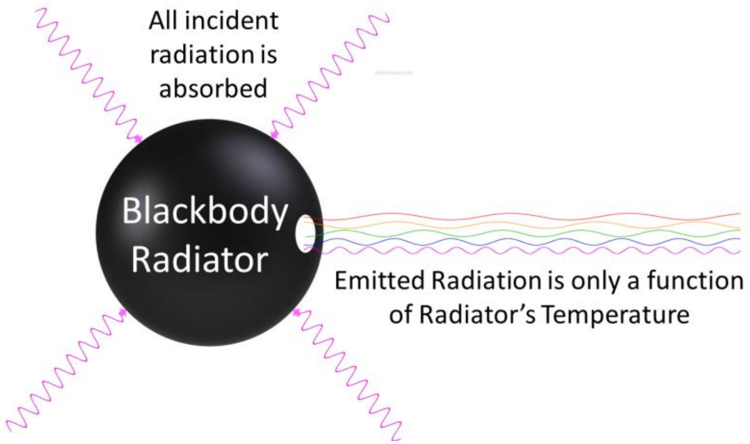
= high energy photons	= lower energy
= high frequency	= lower frequency

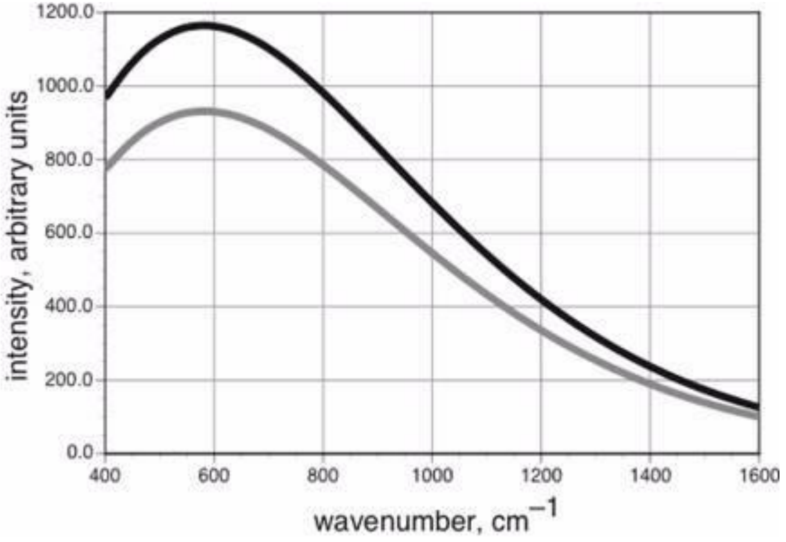
Electromagnetic Spectrum

AM waves ~100 m
Microwaves ~ 1 mm
Infrared ~ 10^{-6} m
Visible light ~ 5 10^{-7} m
UV ~ 1 10^{-7} m



Blackbodies and Graybodies





- A blackbody is a hypothetical object that absorbs all of the radiation that strikes it. It also emits radiation at a maximum rate for its given temperature.
  - Does not have to be **black**!
- A greybody absorbs radiation equally at all wavelengths, but at a certain fraction (absorptivity, emissivity) of the blackbody rate
  - For a real greybody, emissivity is equal for all wavelengths
- The energy emission rate is given by:
  - Planck's law (wavelength dependent emission)
  - Stefan Boltzmann law (total energy)
  - Wien's law (peak emission wavelength)

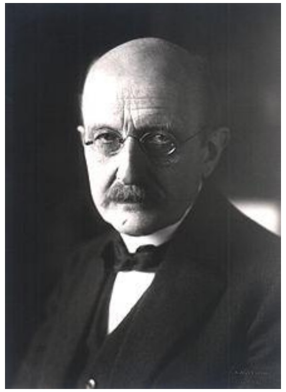
Blackbody Radiation

Planck's Law

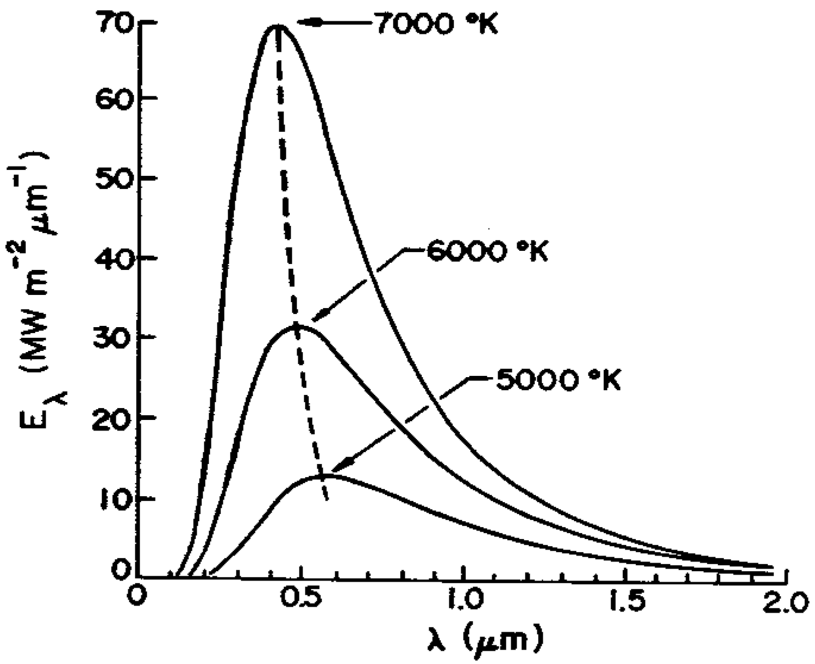
(W m^{-2} m^{-1} = J m^{-2} m^{-1} s^{-1})

$$E_\lambda = \frac{c_1}{\lambda^5[\exp(c_2/\lambda T) - 1]}$$

$$\simeq c_1 \lambda^{-5} \exp(-c_2/\lambda T)$$



Marx Planck
(Kiel 1858 – Göttingen 1947)



- Planck's Law describes the rate of energy output of a blackbody as a function of wavelength
- Emission is a very sensitive function of wavelength
- Total emission is a strong function of temperature

Total Blackbody Emission: Stefan-Boltzmann law

- Integrating Planck's Law across all wavelengths, and all directions, we obtain an expression for the total rate of emission of radiant energy from a blackbody:

$$E^* = \sigma T^4 \qquad (W\ m^{-2} = J\ m^{-2}\ s^{-1})$$

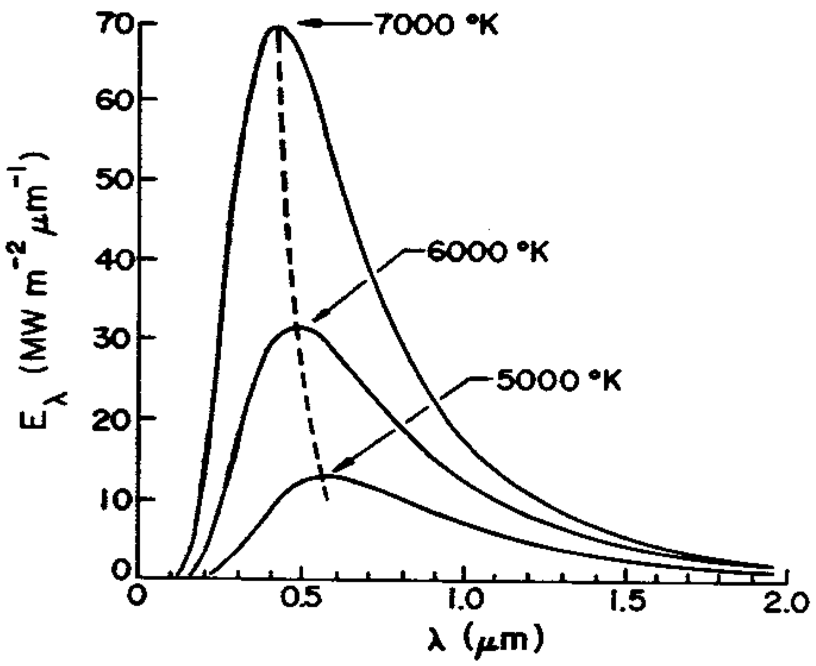
- This is known as the Stefan-Boltzmann Law, and the constant σ is the Stefan-Boltzmann constant: $\sigma = 5.67 \times 10^{-8}\ W\ m^{-2}\ K^{-4}$
- Stefan-Boltzmann says that total emission depends really strongly on temperature!
- This is strictly true only for a blackbody.
  For a **grey body**, **$E = \varepsilon E^*$**, where ε is called the emissivity.
- In general, the emissivity depends on wavelength just as the absorptivity does, for the same reasons: $\varepsilon_\lambda = E_\lambda / E^*_\lambda$

Josef Stefan
(St. Peter 1835 – Vienna 1893)

Ludwig Eduard Boltzmann
(Vienna 1844 – Duino 1906)

Red is Cool, Blue is Hot

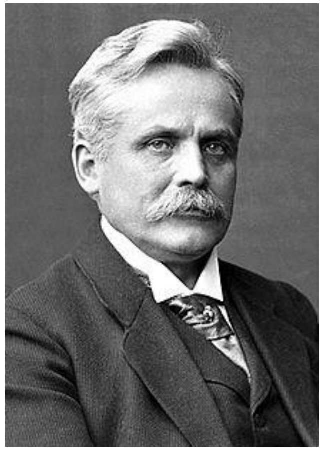
- Take the derivative of the Planck function, set to zero, and solve for wavelength of maximum emission



Wien's Displacement Law

$$\lambda_{max} = \frac{2897}{T}$$

(energy is concentrated at shorter wavelegths for hotter emitters)



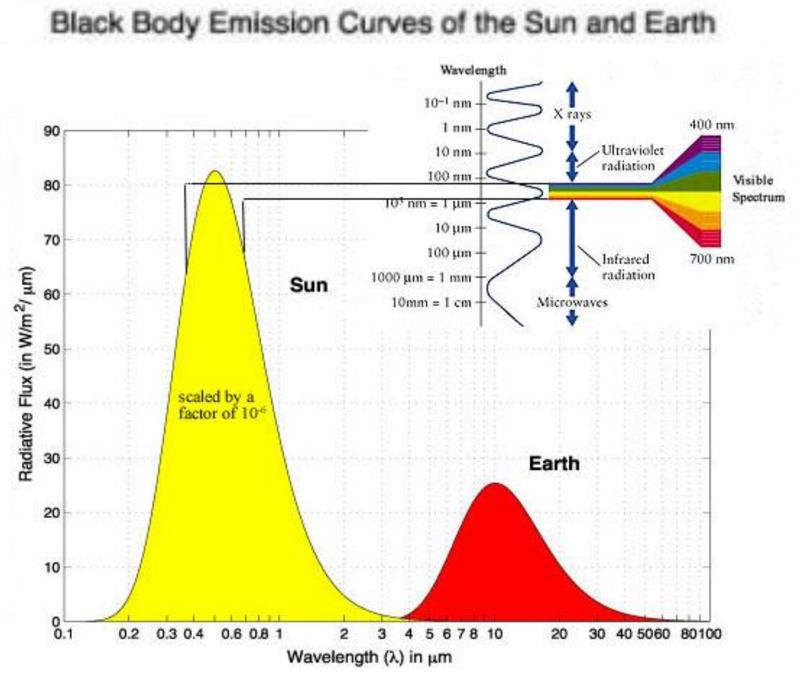
Wilhelm Wien
(Fischhausen 1864 – Monaco di Baviera 1928)

Earth and Sun emission spectrum

$$\lambda_{\max} = \frac{2897}{T}\ \mu m$$

$$\lambda_{sun} = \frac{2897}{6000}\ \mu m = 0.48\ \mu m$$

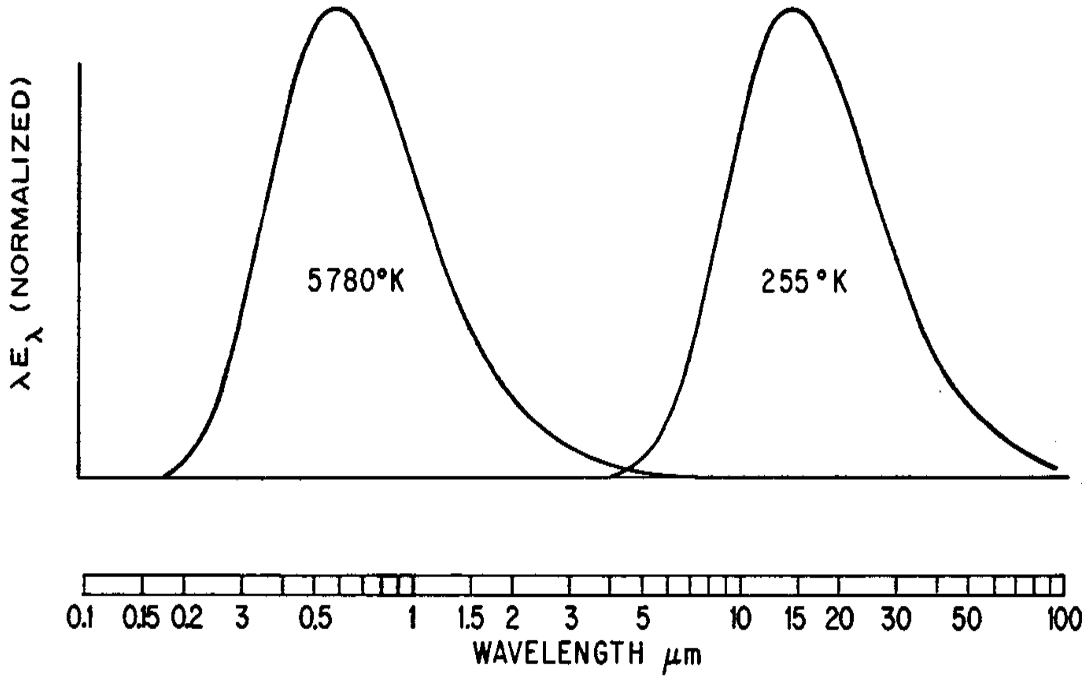
$$\lambda_{earth} = \frac{2897}{255}\ \mu m = 11.4\ \mu m$$



Low temperature moves emission spectrum well into infrared range, that means that mostly heat is radiated away from earth surface. The infrared radiation can be absorbed in air, clouds, or aerosols, causing temperature increase of the atmosphere.

Solar and Planetary Radiation

- Earth receives energy from the sun at many wavelengths, but most is visible or shorter
- Earth emits energy back to space at much longer (thermal) wavelengths
- Because temperatures of the Earth and Sun are so different, it's convenient to divide atmospheric radiation conveniently into solar and planetary



Radiation names

By its source

– **Solar radiation - originating from the sun**

– **Terrestrial radiation - originating from the earth**

By its proper name

– ***ultra violet, visible, near infrared, infrared, microwave, etc….***

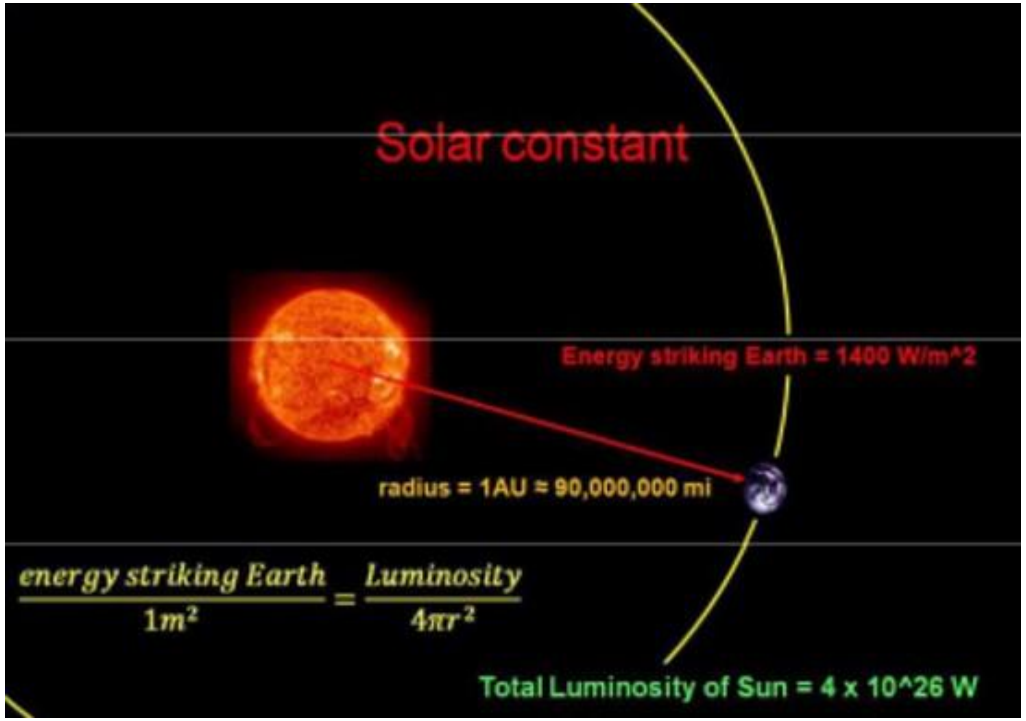
By its wavelength

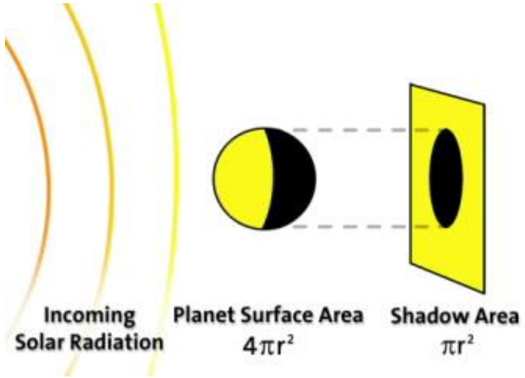
– ***short wave* radiation $\lambda \leq 3$ μm**

– ***long wave* radiation $\lambda > 3$ μm**

Solar constant

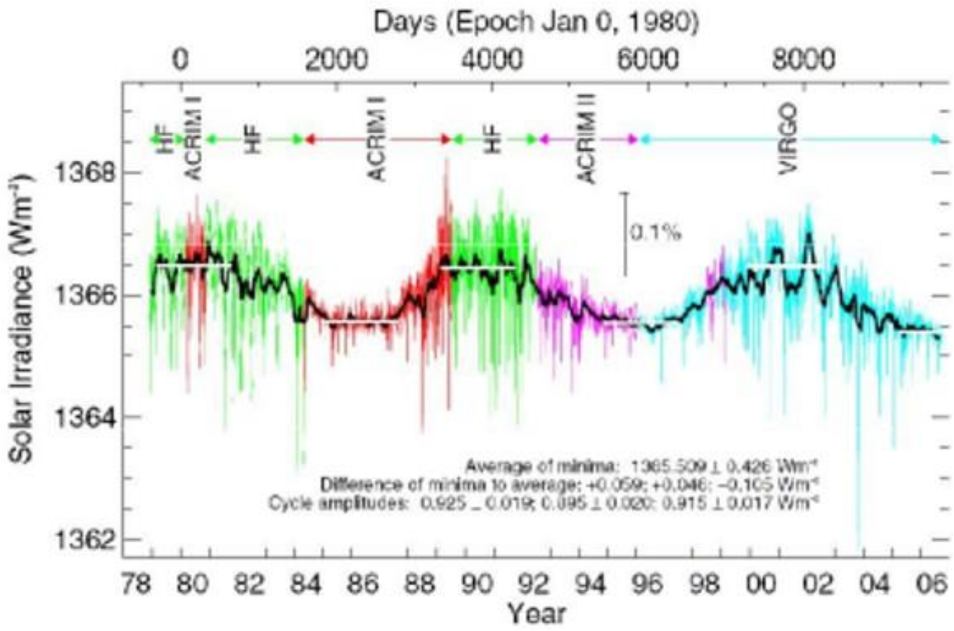
- Data:
  - $P_☼ = 4\ 10^{26}$ W (solar power);
  - $T_☼ = 6000$ K (solar surface temperature);
  - $R_☼ = 6.96\ 10^5$ km = 0.00465 AU (solar radius);
  - $R_{orbit} = 1.496\ 10^8$ km = 1 AU Earth-Sun average distance);
  - $\alpha = 0.31$ (Earth albedo)
  - $\sigma = 5.67\ 10^{-8}\ W\ m^{-2}\ K^{-4}$
- Solar radiation at Sun surface: $\sigma T^4_☼ = 7.348\ 10^7\ W/m^2$
- Solar power: $P_☼ = \sigma T^4_☼\ 4\ \pi\ R_☼^2 = 4.47\ 10^{26}$ W
- Radiation on Earth TOA: $R_T = P_☼ / 4\ \pi\ R_{orbit}^2 = 1370$ W/m^2
- Received radiation: $R'_r = R_T / (1 - \alpha) = 945\ W/m^2$
- Absorbed radiation: Earth surface receives solar radiation at different angles: the received radiation is equivalent to a cicle perpendicular to the Sun, having same R_e radius. Thus: $R_{abs} = R'_r\ / 4 = 236.3\ W/m^2$





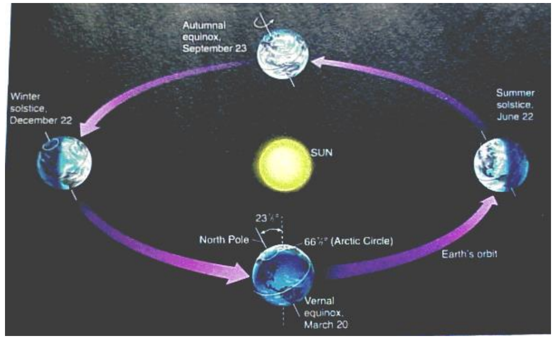
Solar constant is not constant

- Solar radiation varies in time (see next slides)

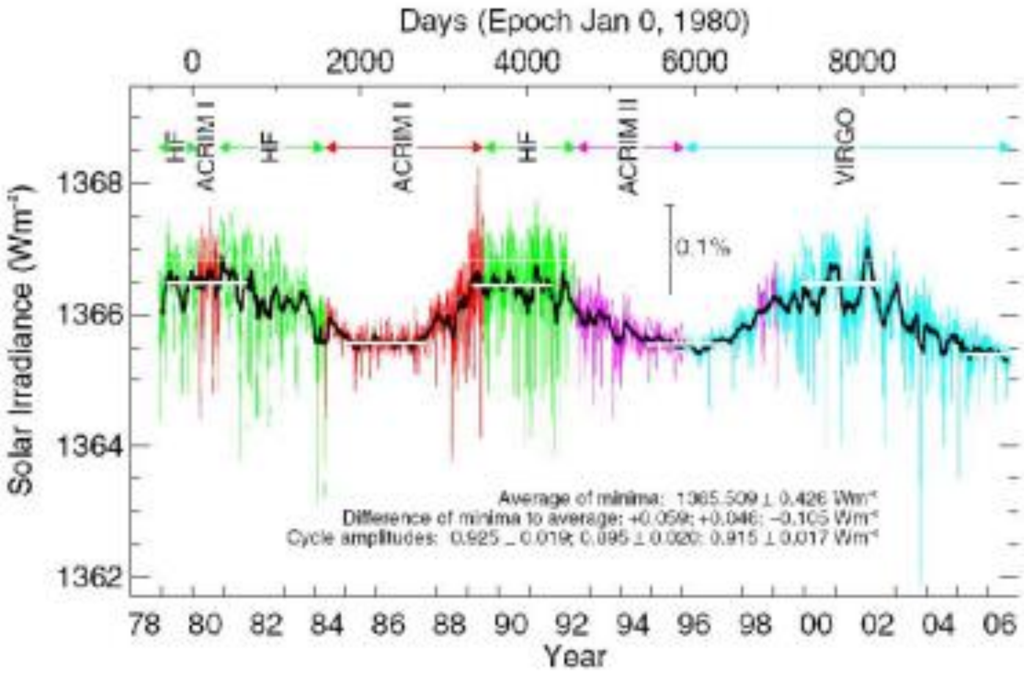


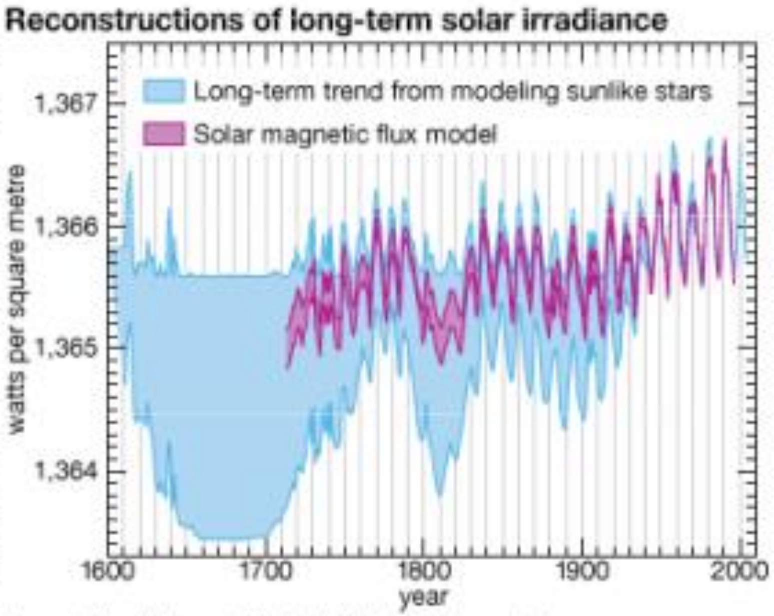
- Mean variations are small but individual variations can be 1% of solar radiation

- Earth-Sun distance varies in time (and so solar radiation)
- Eccentricity of solar orbit: 3% → Earth-Sun distance can vary by 3%
  - Minimum distance:
    - $R_{orbit,min}$=0.085 R_{orbit}=1.272 10^8 km
    - $F_{0,min}=\sigma T^4_{☼} (R_{☼}/R_{orbit,min})^2$=1897 Wm^{-2}
  - Maximum distance:
    - $R_{orbit,max}$=1.015 R_{orbit}=1.518 10^8 km
    - $F_{0,max}=\sigma T^4_{☼} (R_{☼}/R_{orbit,max})^2$=1351 Wm^{-2}
- Thus distance can alter significantly the quantity of radiation at the TOA



Variations of solar constant



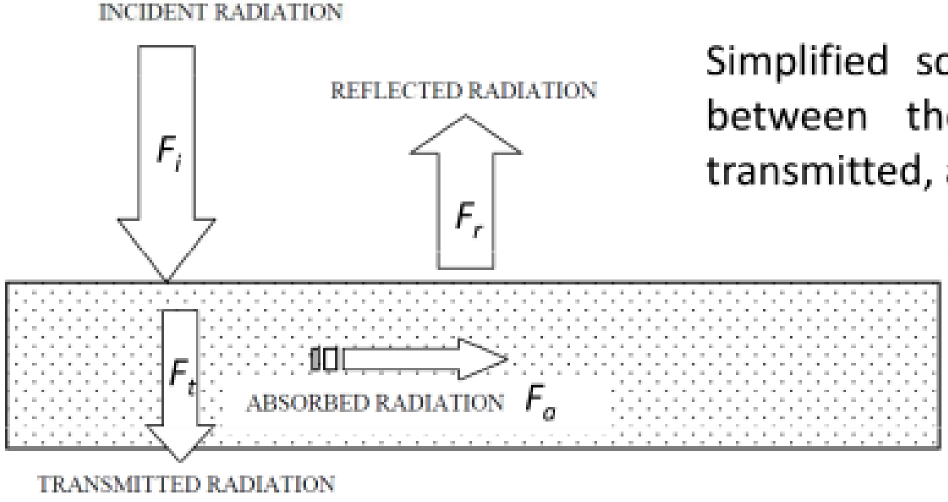


Source: Climate Change 2007: The Physical Science Basis, Summary for Policymakers, Intergovernmental Panel on Climate Change

Flux ≅ Irridiance

Measurements of solar constant over last 30 years indicate variation of ≈0.03%
Model based reconstruction of long term irradiance suggests slight increase by ≈0.01%. A regular oscillation of ≈0.01% is observed in an eleven year cycle!

Energy conservation



Simplified scheme of the balance between the incident, reflected, transmitted, and absorbed radiation

Box Model

Kirchhoff's law

$$F_i = F_r + F_a + F_t \qquad \frac{F_r}{F_i} + \frac{F_a}{F_i} + \frac{F_t}{F_i} = 1$$

$$\alpha + \varepsilon + o = 1$$

Albedo, Absorption, Opacity

Efficiency factors

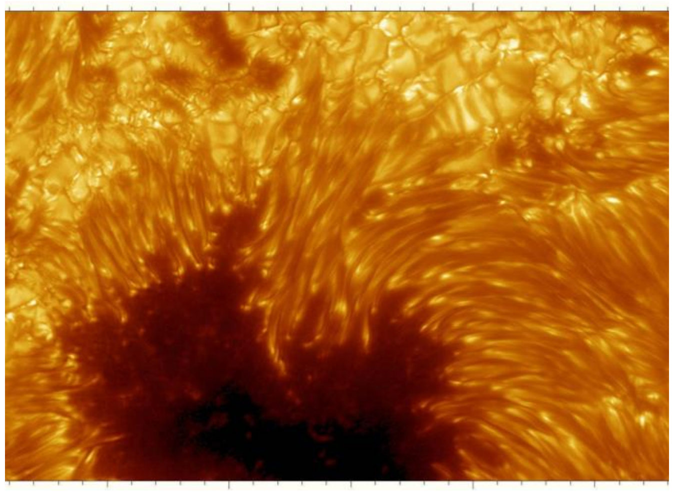
ε: emissivity (=absorptivity)
α: albedo
o: opacity (=1-transmittivity)

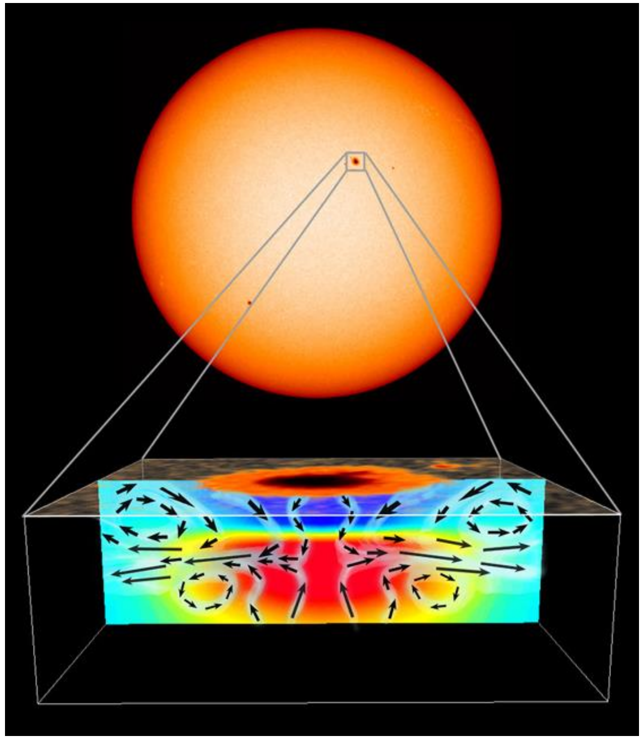
Black body: ε=1, α=0
Opaque body: o=0

The incident, absorbed, reflected, and transmitted flux depends sensitively on the wavelength λ of the radiation!

Sunspots

- Sunspots = dark areas on the sun surrounded by brighter faculae (plages)
  - Disturbance in the magnetic field that inhibits convection
  - Dark spots = lower temperature than normal (convection from below)
  - Faculae (or plages) = higher temperature than normal
  - Also, more energy from solar flares during sunspots





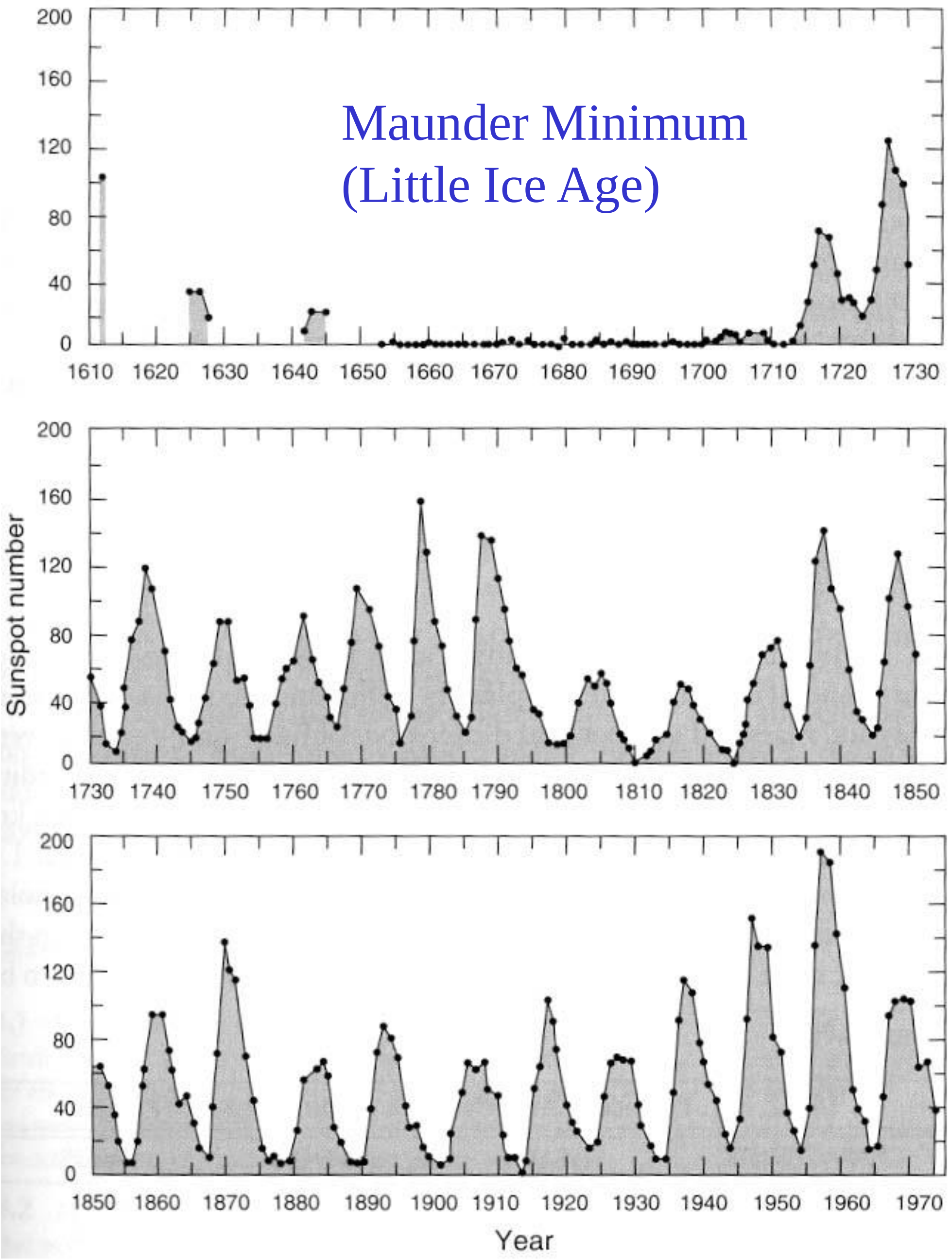
- Net effect:

  More sunspots = more solar radiation (0.1-0.2% variation)

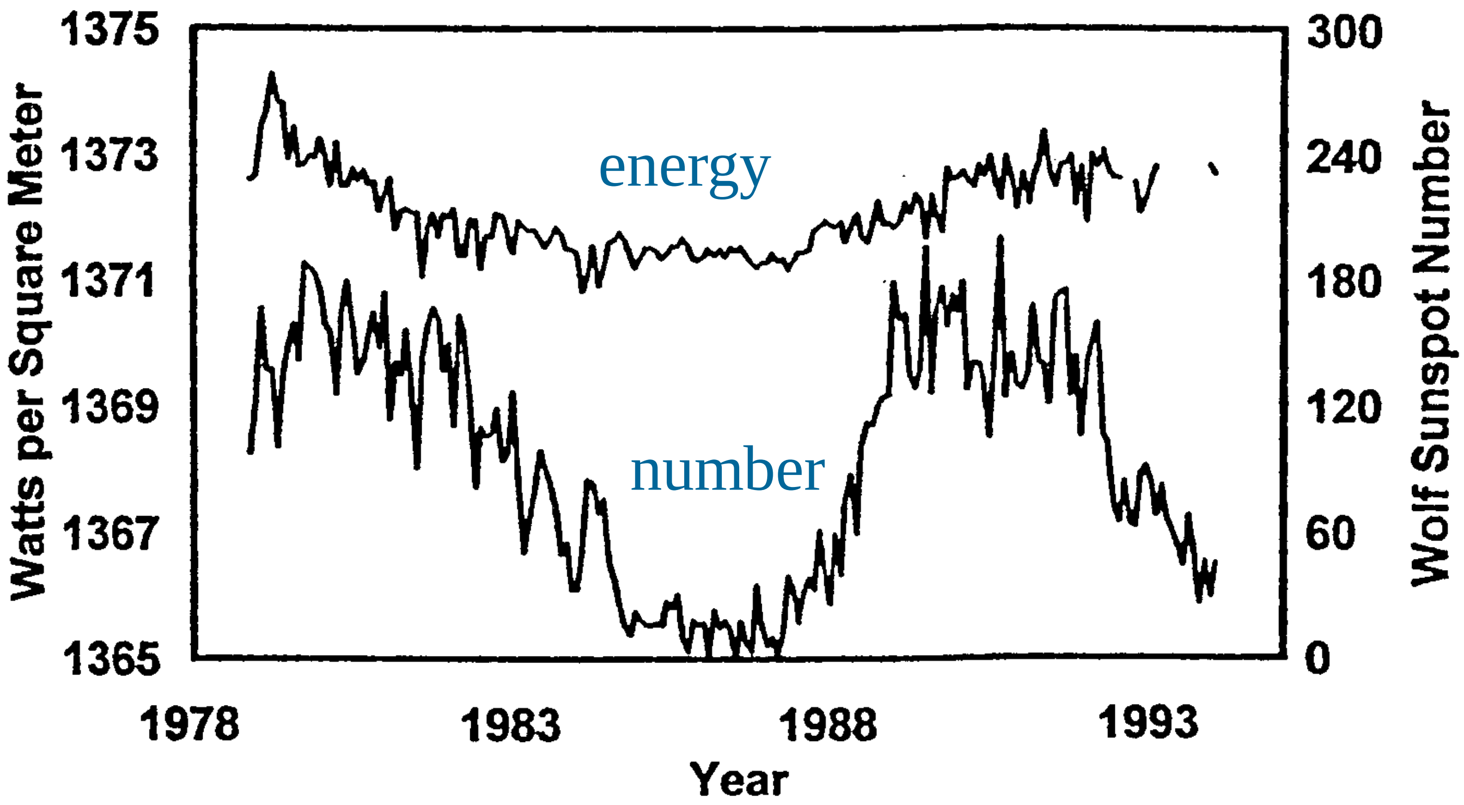
Sunspots

More sunspots ↑ More radiation ↑

Edward Walter Maunder
(1851 –1928)



Sunspot number and radiation

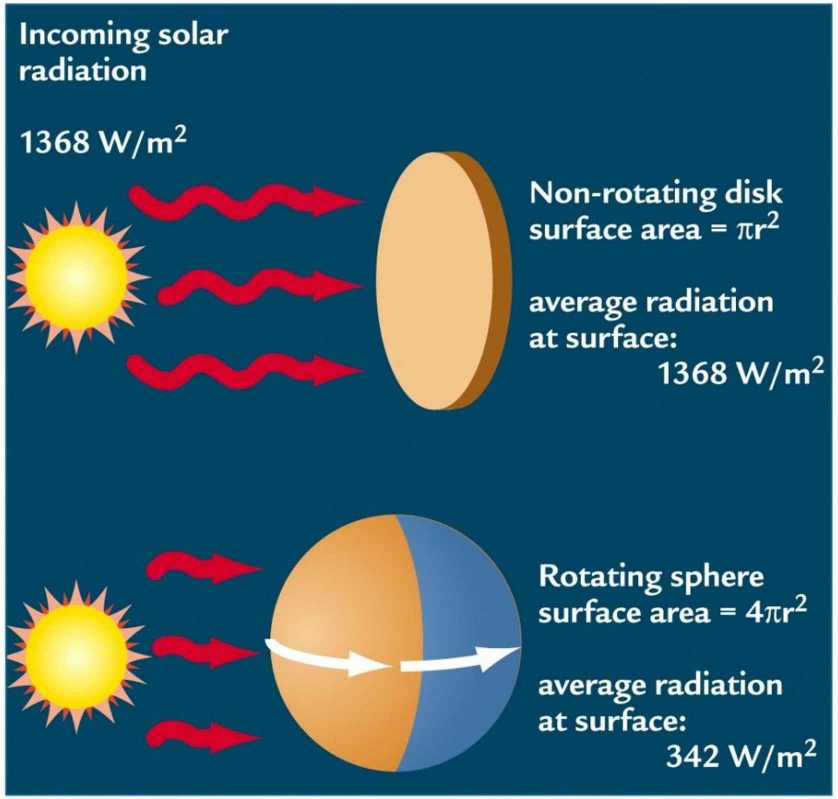


Satellite Data

—— Solar Irradiance = Upper Curve —— Wolf Numbers = Lower Curve

1.5/1373 = .001 or 0.1%

Solar Radiation on Earth



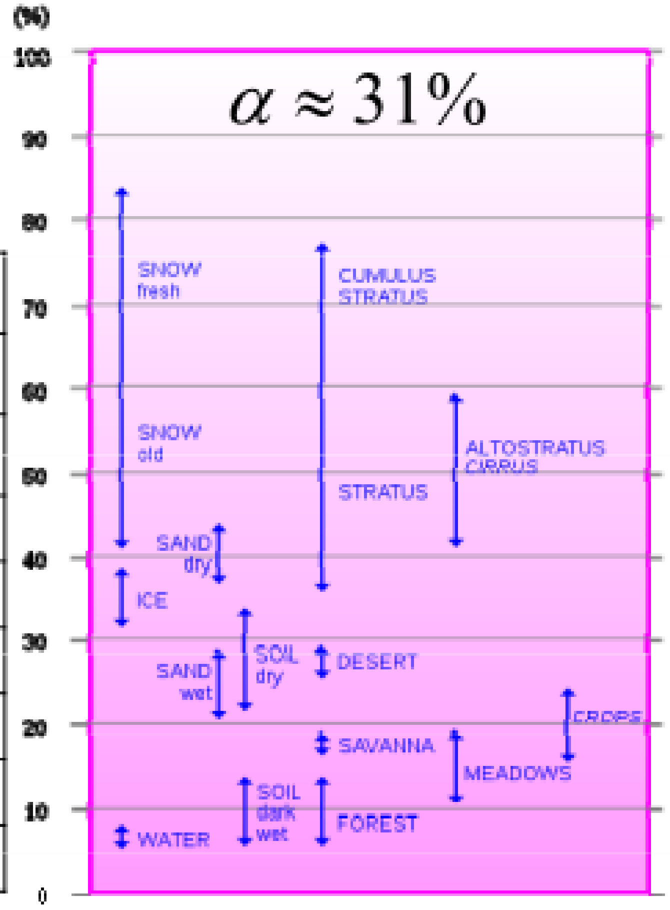
$$R = \frac{S\pi r^2}{4\pi r^2} = \frac{1369}{4} \approx 342\frac{W}{m^2}$$

Average solar radiation at Earth's surface

Albedo

The ratio of reflected to incident solar energy is called the Albedo α

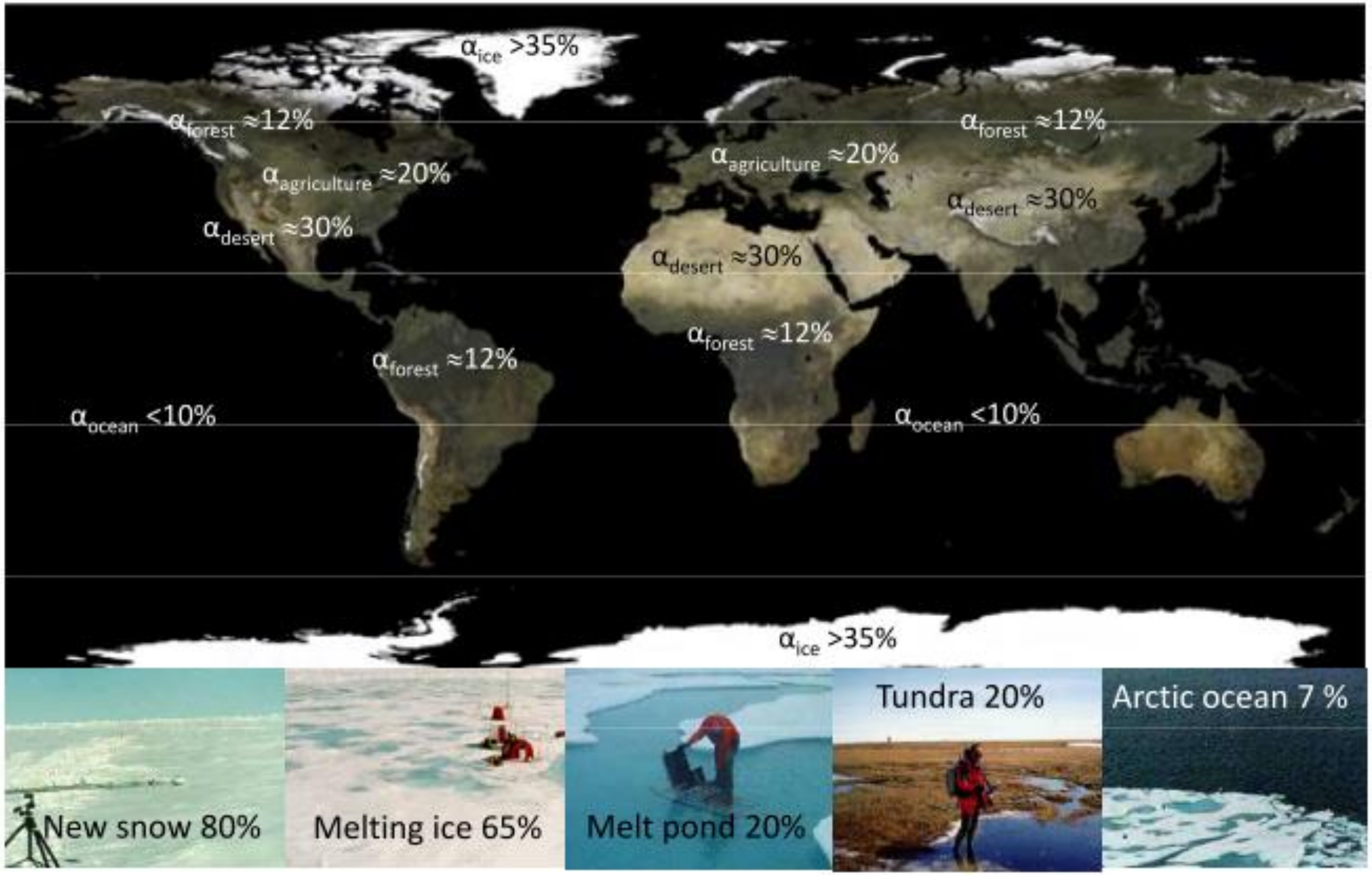
At present cloud and climate conditions:



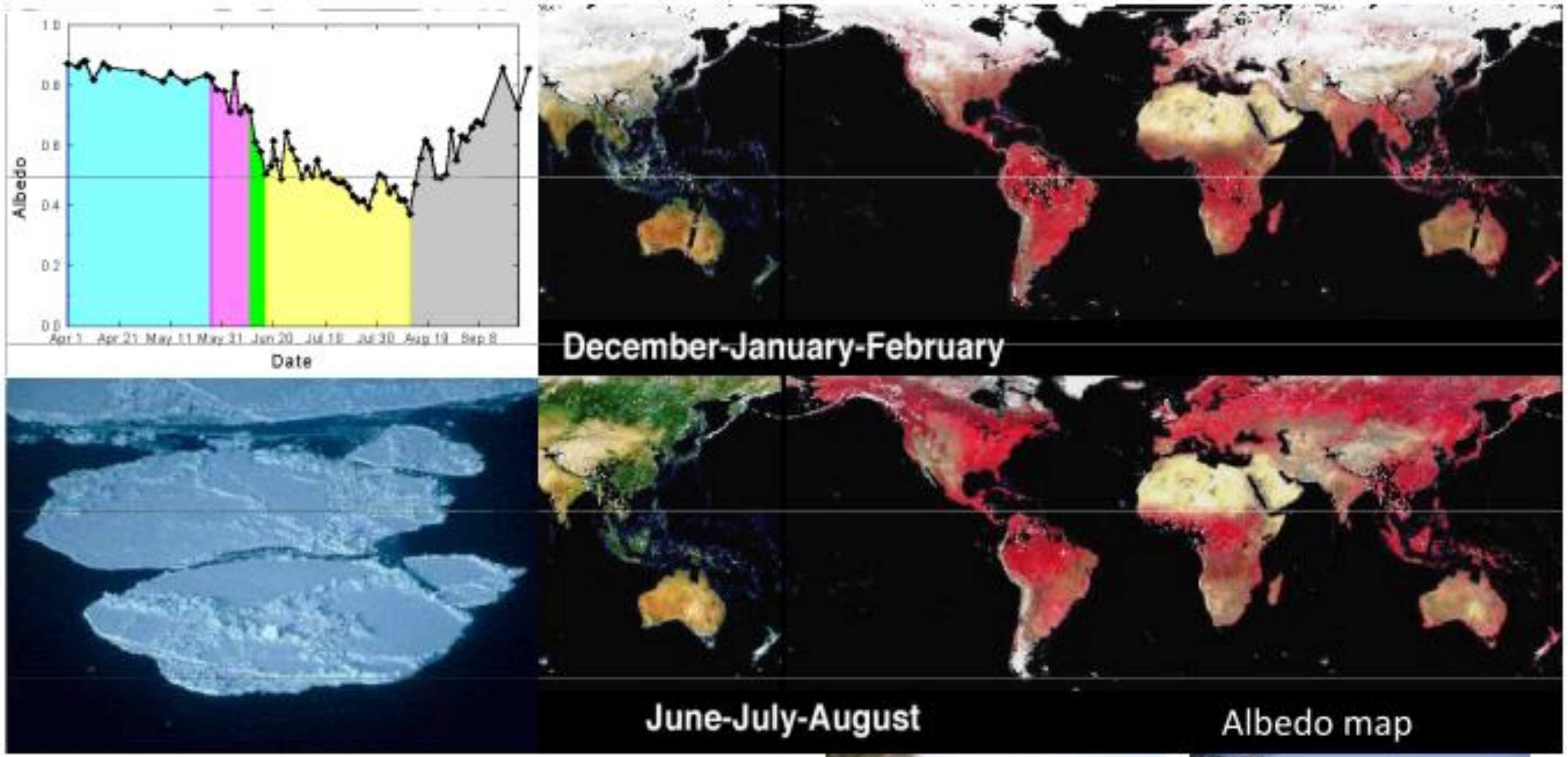
The Albedo depends on the nature and characteristics of the reflecting surface, a light surface has a high Albedo (maximum 1 or 100%), a dark surface has a small Albedo (minimum 0 or 0%).

Surface	Albedo
Asphalt	4-12%
Forest	8-18%
Bare soil	17%
Green grass	25%
Desert sand	40%
New concrete	55%
Ocean Ice	50-70%
Fresh snow	80-90%

Albedo of earth



Seasonal albedo variations



Seasonal changes depends primarily on large area snow and ice formation!

Albedo feedback processes

- Snow has a high Albedo, average over Antarctica is about 80%.
- Snow melt lowers the Albedo, more sunlight is absorbed and temperature increases accelerating melting process.
- If snow forms, the Albedo increases, which results into further cooling because more light is reflected and less light is absorbed.

- Deforestation for generating agricultural land or grassland increases Albedo from ~10 to ~25%, more sunlight is reflected decreasing temperature but also evaporation, cloud formation and precipitation, increasing aridity.

Tradition and experience

Traditional German village with dark slate walls which helps by low Albedo to absorb energy and keep the houses warm in moderate summers and cold winter times.

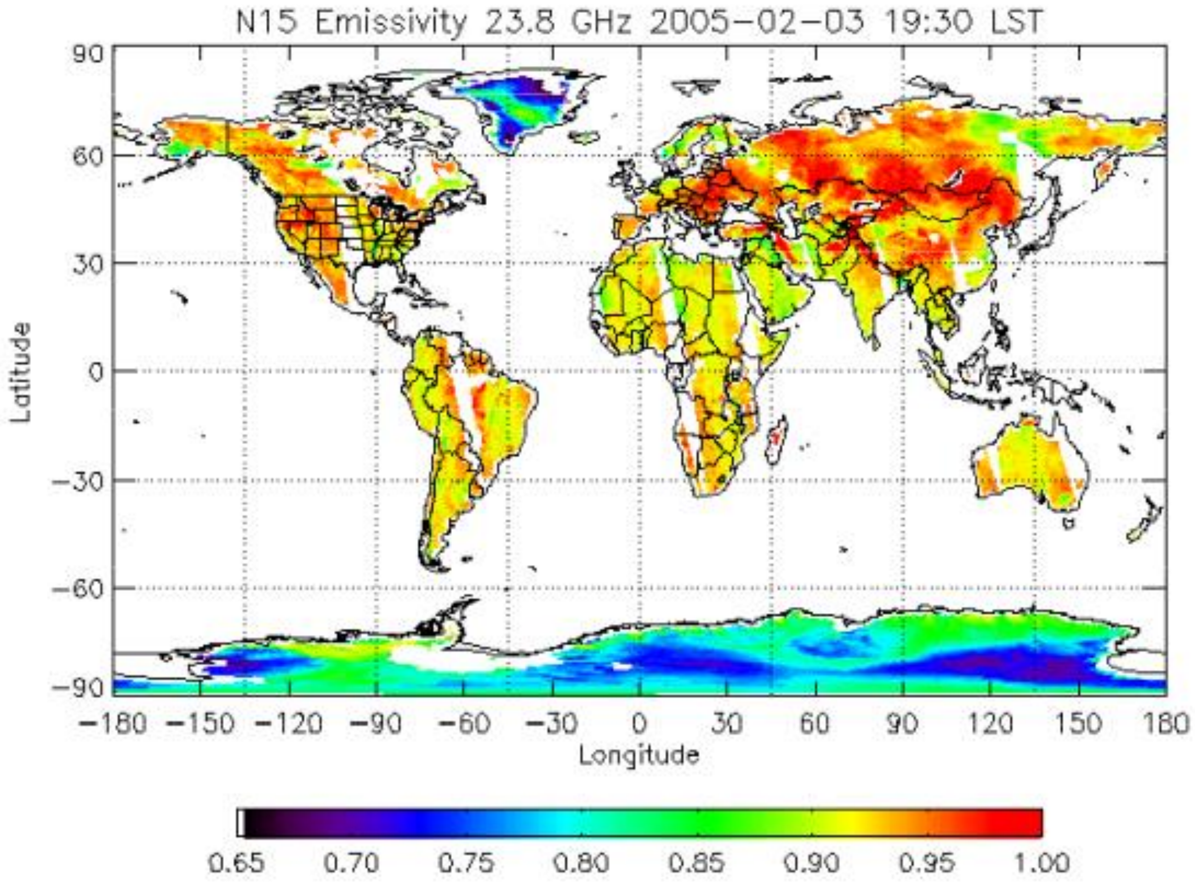
Traditional Greek (Mediterranean) village with chalked walls with high Albedo to reflect solar energy and minimize absorption to keep houses cool in hot summer months.

Emissivity

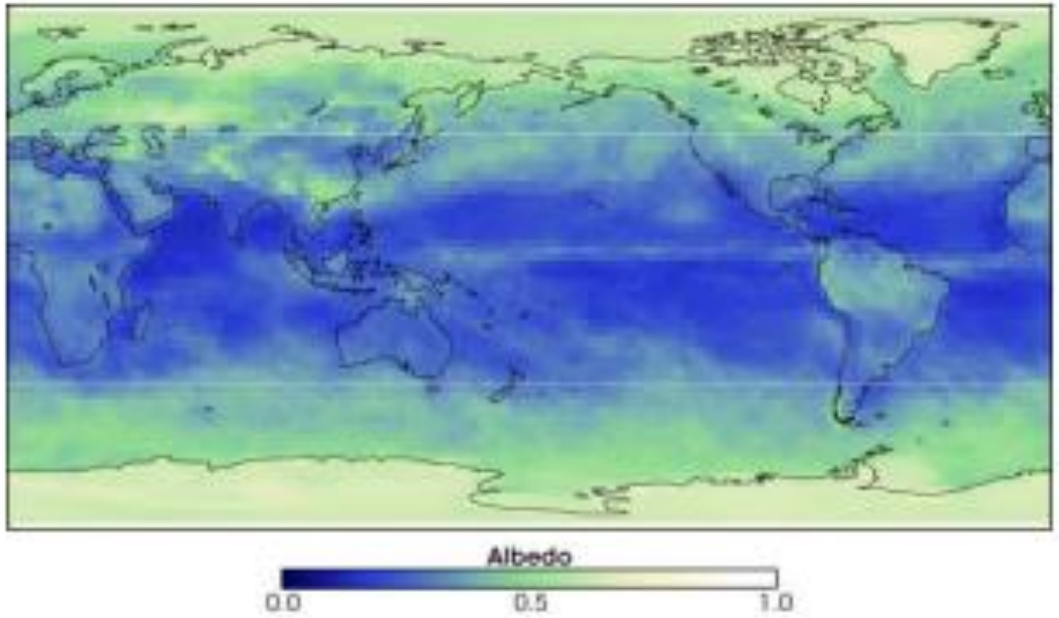
Emissivity in IR range $\varepsilon_\lambda > 0.8$
(λ=4μm – 100μm)

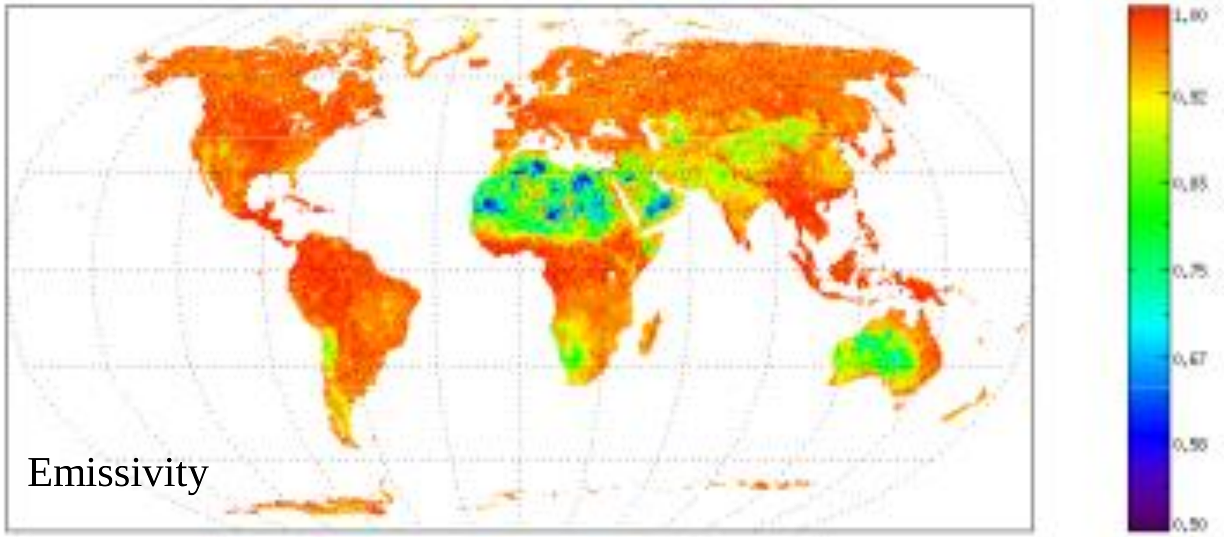
For λ=8μm – 12μm

Surface	Emissivity
Water	**0.993-0.998**
Ice	**0.98**
Green grass	**0.975-0.986**
Sand	**0.949-0.962**
Snow	**0.969-0.997**
Granite	**0.898**
White paint	**0.25-0.31**
Black paint, tar	**0.9-1.0**

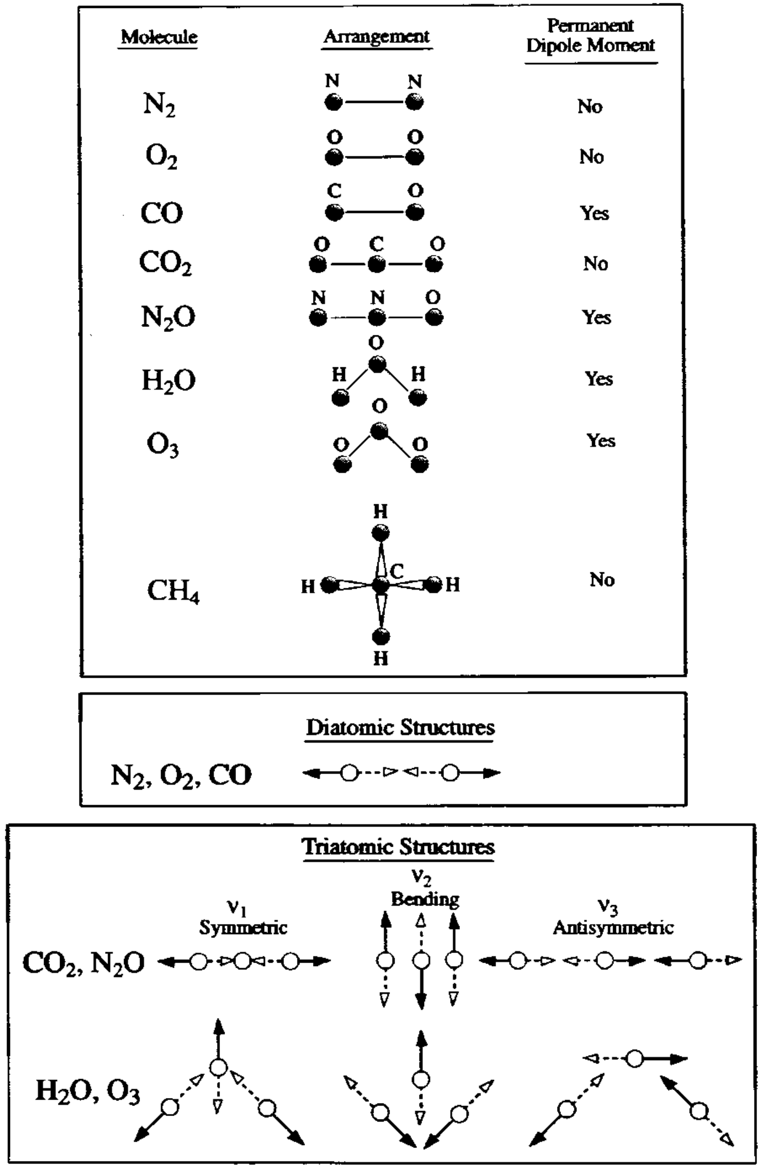


Earth albedo and emissivity





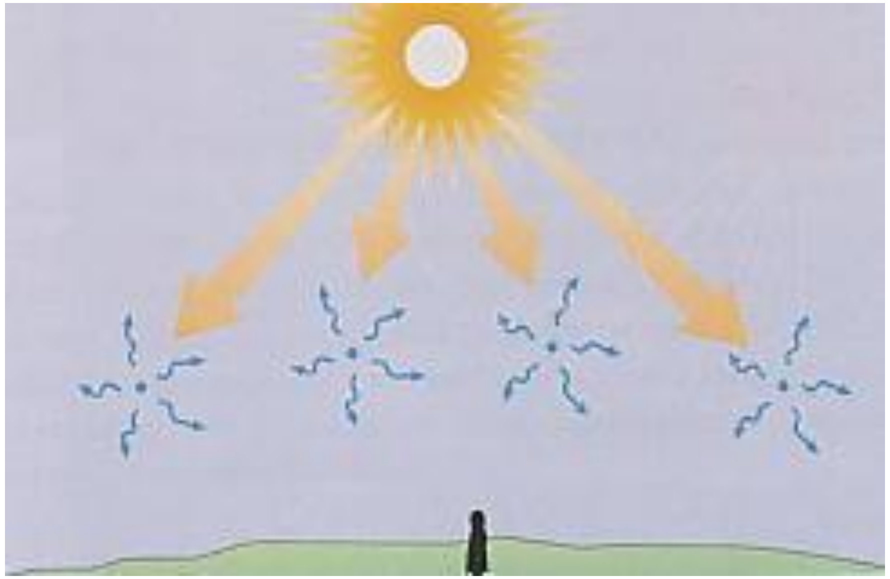
Molecular Absorbers/Emitters



- Molecules of gas in the atmosphere interact with photons of electromagnetic radiation
- Different kinds of molecular transitions can absorb/emit very different wavelengths of radiation
- Some molecules are able to interact much more with photons than others
- Different molecular structures produce wavelength-dependent absorptivity/emissivity

Scattering: Why is the day sky blue?

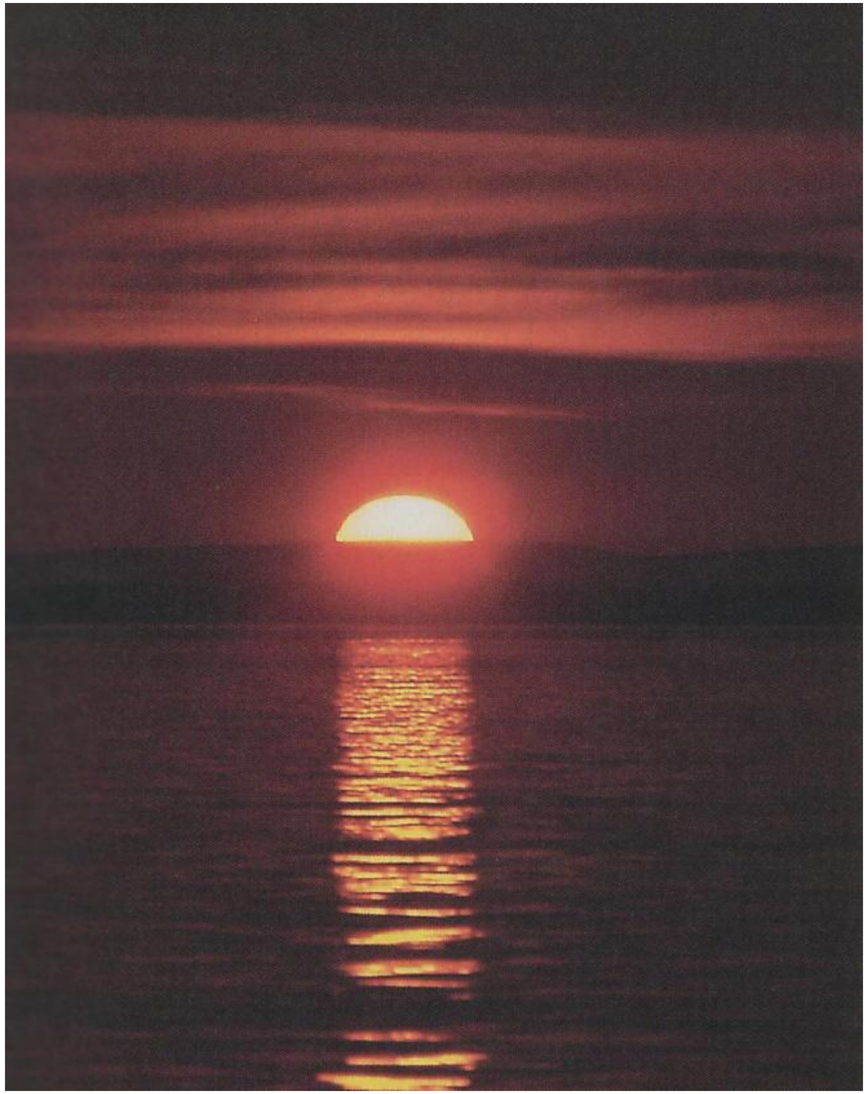
- Sunlight is scattered by air molecules
- Air molecules are much smaller than the light's λ
- Shorter wavelengths (green, blue, violet) scattered more efficiently
- Unless we are looking directly at the sun, we are viewing light scattered by the atmosphere, so the color we see is dominated by short visible wavelengths
  - blue dominates over violet because our eyes are more sensitive to blue light



$$I_0 \sim I/\lambda^4 (1+\cos^2(theta))$$

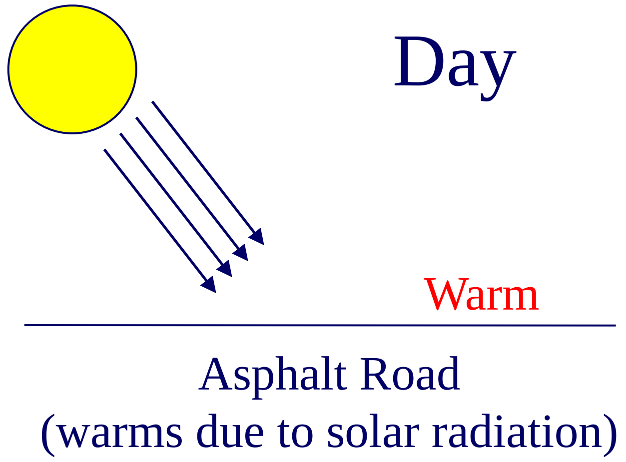
Why are sunsets red?

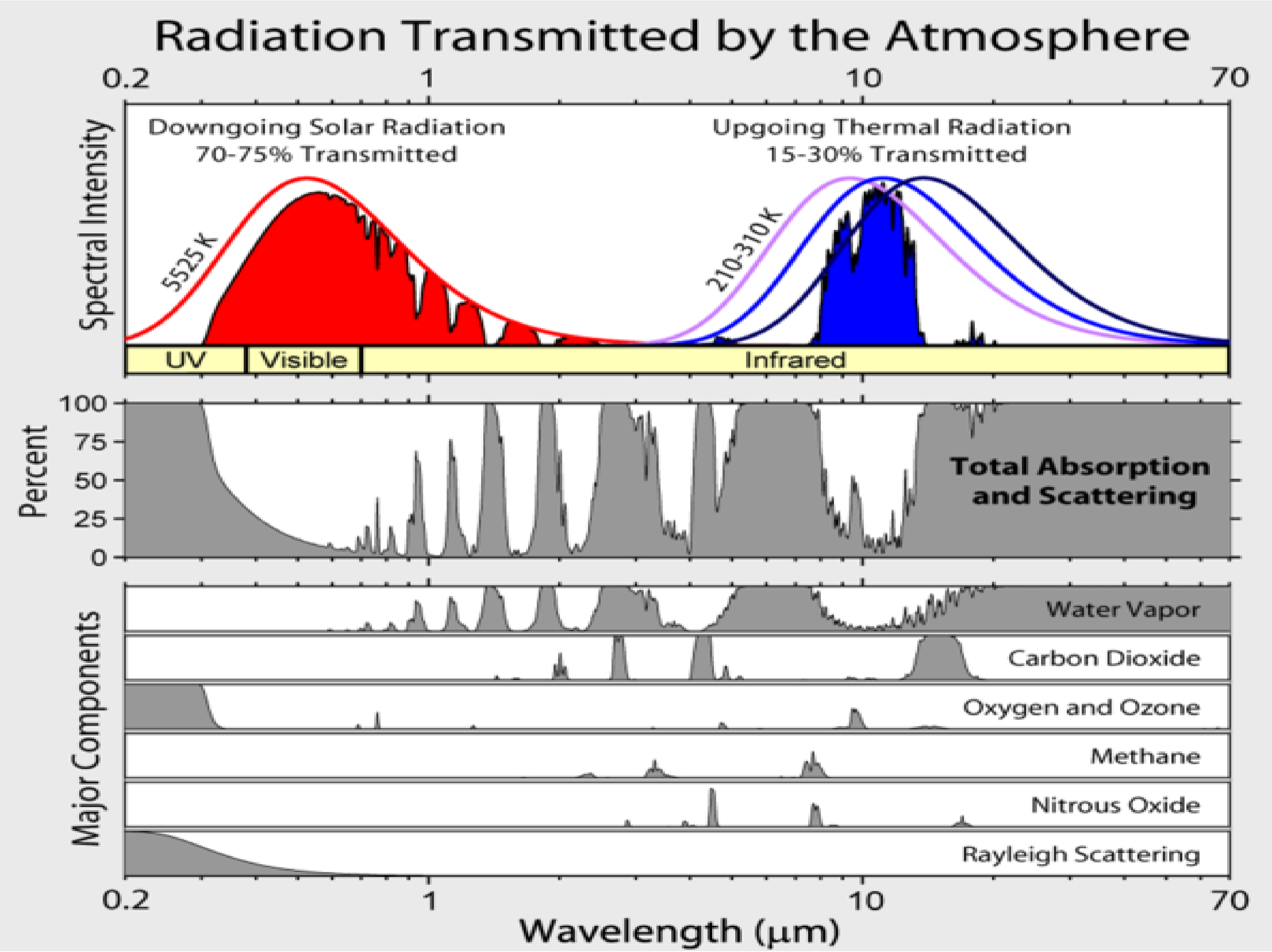
- The sun appears fairly white when it's high in the sky
- Near the horizon, sunlight must penetrate a much greater atmospheric path
  - More scattering
- In a clean atmosphere, scattering by gases removes short visible λ's from the line-of-sight
  - Sun appears orange/yellow because only longer wavelengths make it through
- When particle concentrations are high, the slightly longer yellow λ's are also scattered
  - - Sun appears red/orange



Absorption: Kirchoff's Law

- Objects that are good absorbers are also good emitters
  - Consider an asphalt road
    - » During the day the asphalt absorbs solar radiation and warms
    - » At night the asphalt emits infrared radiation and cools relative to its surroundings




Radiation Transmitted by the Atmosphere
0.2
1
10
70
Spectral Intensity
Downgoing Solar Radiation
70-75% Transmitted
Upgoing Thermal Radiation
15-30% Transmitted
5525 K
210-310 K
UV
Visible
Infrared
Percent
100
75
50
25
0
Total Absorption and Scattering
Major Components
Water Vapor
Carbon Dioxide
Oxygen and Ozone
Methane
Nitrous Oxide
Rayleigh Scattering
Wavelength (μm)

What happens to short wave radiation incident at the top of the atmosphere?

- It can be absorbed by the atmosphere
- It can be reflected by clouds, particles and air molecules back to space
- It can be transmitted to the surface
  - where it can be
    - absorbed
    - reflected back upward into the atmosphere
      - some of this may be absorbed by the atmosphere
      - some may be transmitted through the atmosphere back to space

Atmospheric Windows

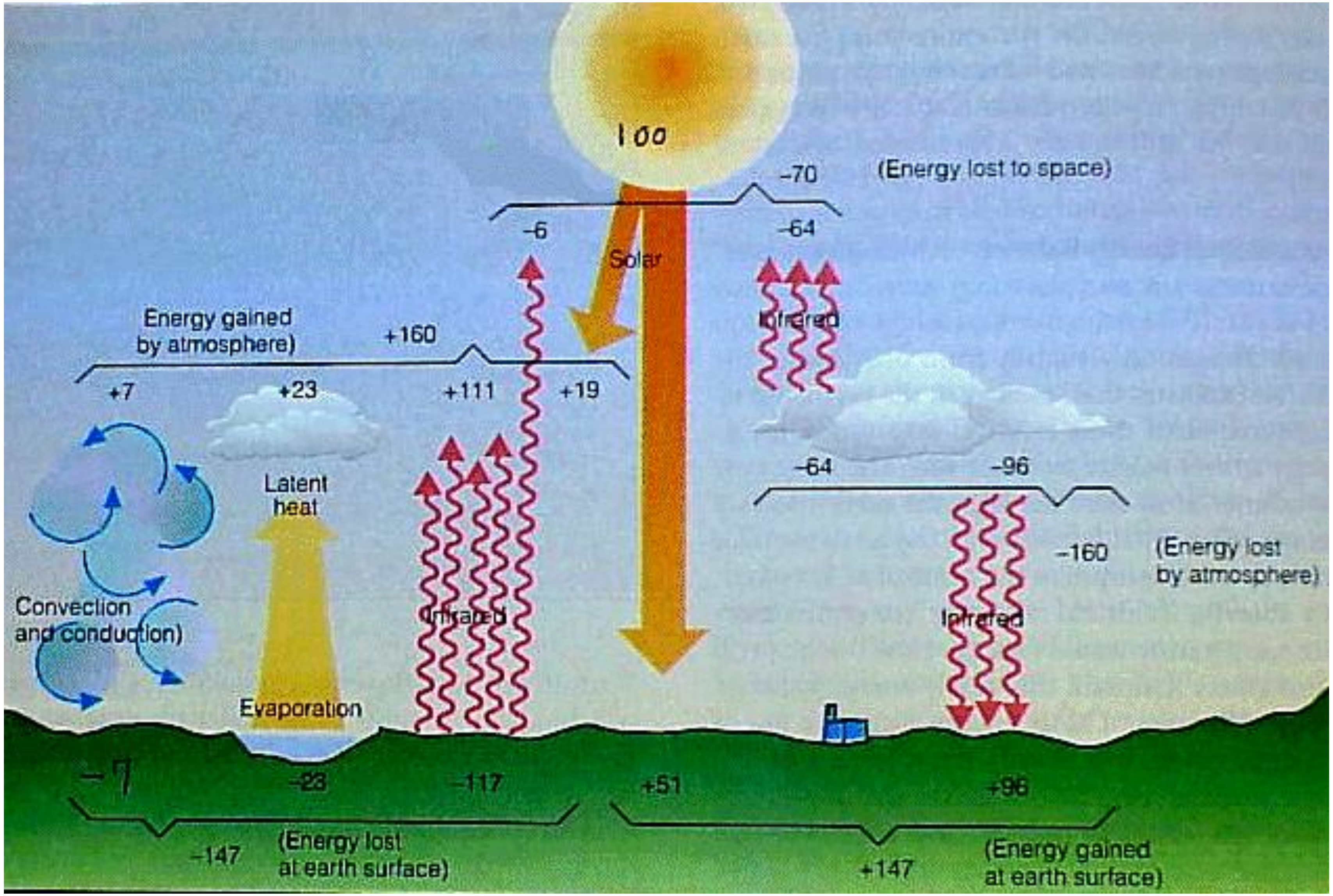
- Portions of the electromagnetic spectrum where atmospheric gases absorb relatively little energy
  - Visible Wavelengths
    - Vision has evolved to use these wavelengths
  - 8-12 micrometers in the terrestrial band

What happens to long wave (terrestrial) radiation?

- **The earth's surface emits LW radiation at temperatures warm relative to the *top of the atmosphere*.**
  - **Some of this radiation escapes directly through the atmosphere to space, thus cooling the planet.**
  - **Some is absorbed by gases and clouds in the atmosphere.**

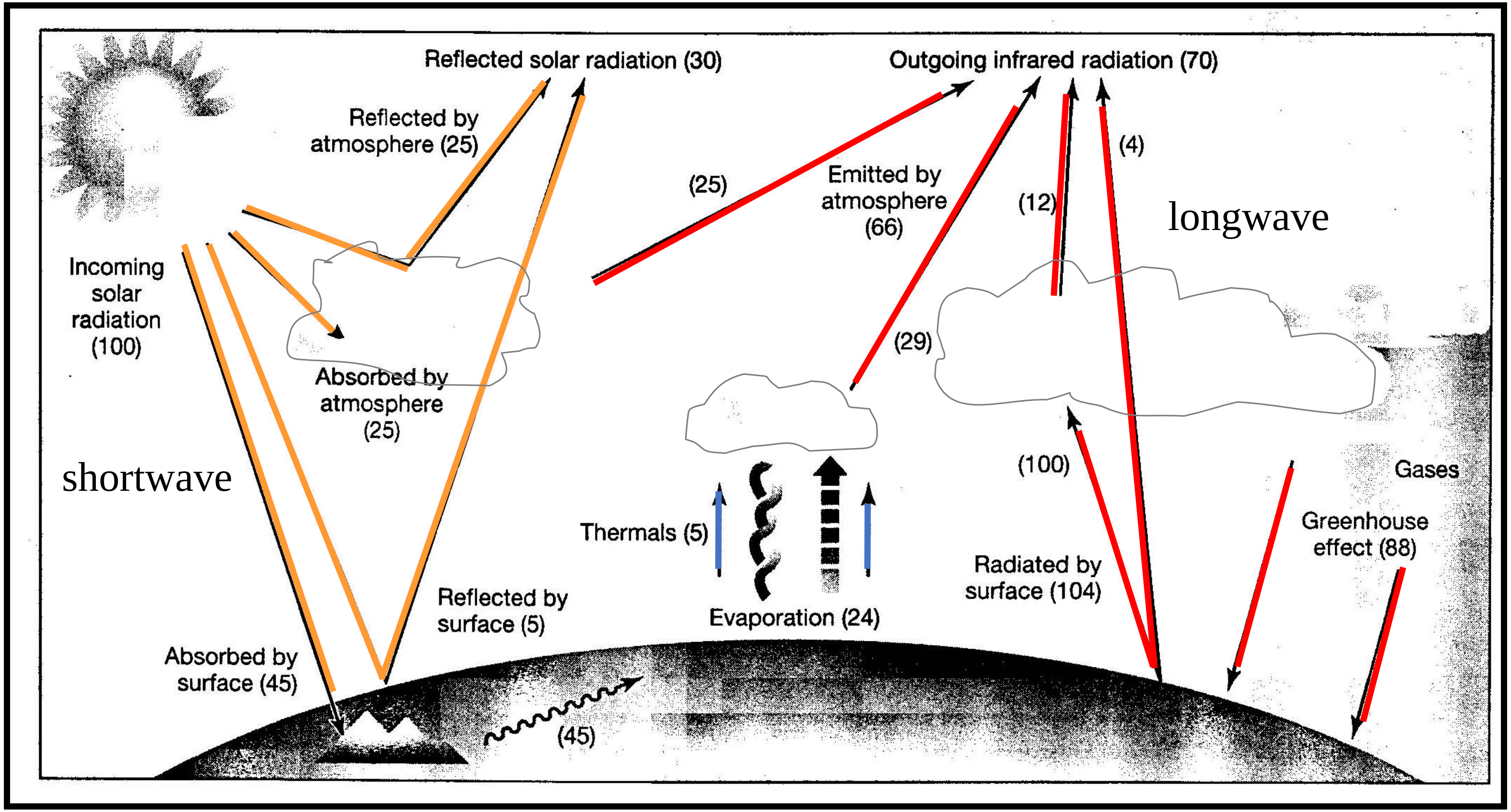
- **The atmospheric gases and clouds emit LW radiation in all directions.**
  - **The atmosphere's LW emission downward "warms" the surface.**
  - **The atmosphere's LW emission upward joins that from the surface escaping to space, thus cooling the planet.**

Annual energy budget



Input: 100 total from Sun
Output: 30 units reflected
Balance: 70 units into system

Earth's Energy Budget



KKC 3-19 100 units of incoming solar radiation

KKC 3-19 100 units of incoming solar radiation

Earth’s Energy Budget

- Incoming short wave - 100 units
- Reflected short wave (30)
  - 25 from atmosphere (clouds and air)
  - 5 from earth’s surface
- Absorbed by atmosphere (25)
  - 25 to heat atmosphere
- Absorbed by Earth (45)
  - 45 to heating the oceans and land (reradiated at 15°C)

Earth's Energy Budget

- Outgoing radiation (long wave)
- Long wave from Earth directly out of system
  - 4
- From Earth to atmosphere
  - 100 – 88 = 12
  - 5 as sensible heat (thermals)
  - 24 as latent heat (evaporation)
- From atmosphere out of system
  - 45 reradiated from Earth
  - 25 converted from short wave by atmosphere

41

45 from Earth

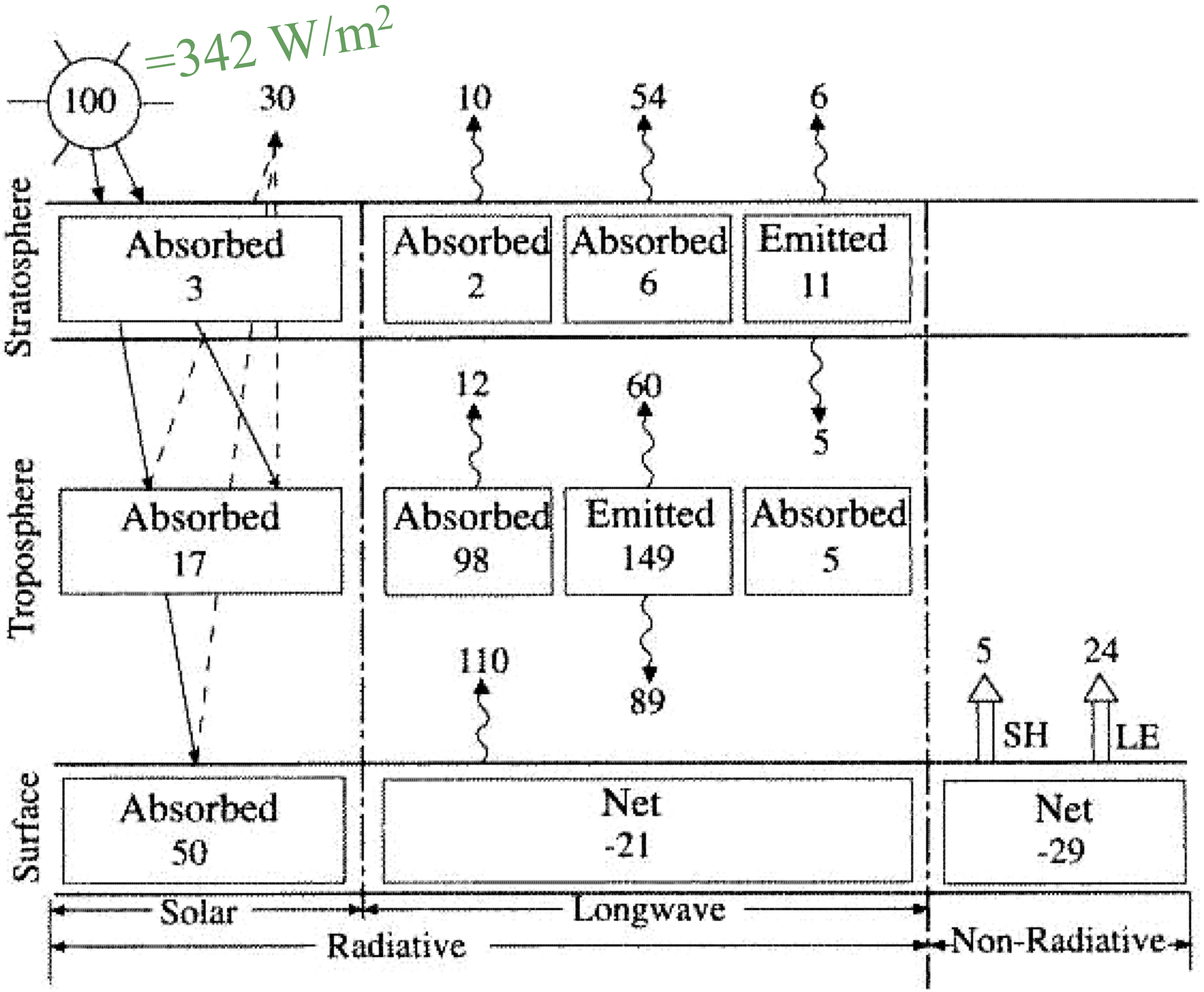
70

Total outgoing long wave = 45 + 25

Radiative imbalance

- The surface gains 30 units of radiation, and the atmosphere loses 30 units of radiation
- Why doesn't the earth's surface heat and the atmosphere would cool indefinitely?
- What other heat transfer processes are there?
  - ✓ Conduction
  - ✓ Convection
  - ✓ Evaporation/condensation (Latent Heat)

Planetary Energy Budget



- Three different balances
- Recycling = greenhouse
- Convective fluxes at surface
- LE > SH
  - LE = latent heat (linked to water vapor flux → water vapor exchanges)
  - SH = sensible heat (linked to temperature → heat exchanges)

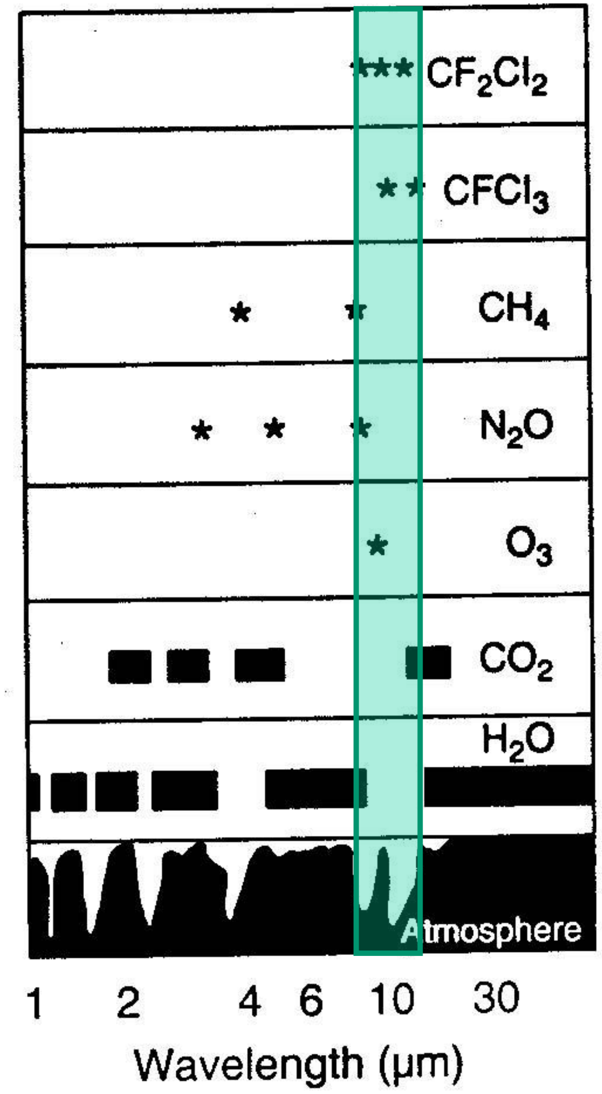
Earth’s Greenhouse Effect

- UV (shorter λ, higher energy that visible light)
  - High energy
  - Breaks bonds

- IR
  - Lower energy (longer wavelength)
  - Increases vibrational and rotational energy of molecules
  - Energy is transmitted to the atmosphere by molecular collisions (kinetic energy)

Atmospheric «window»

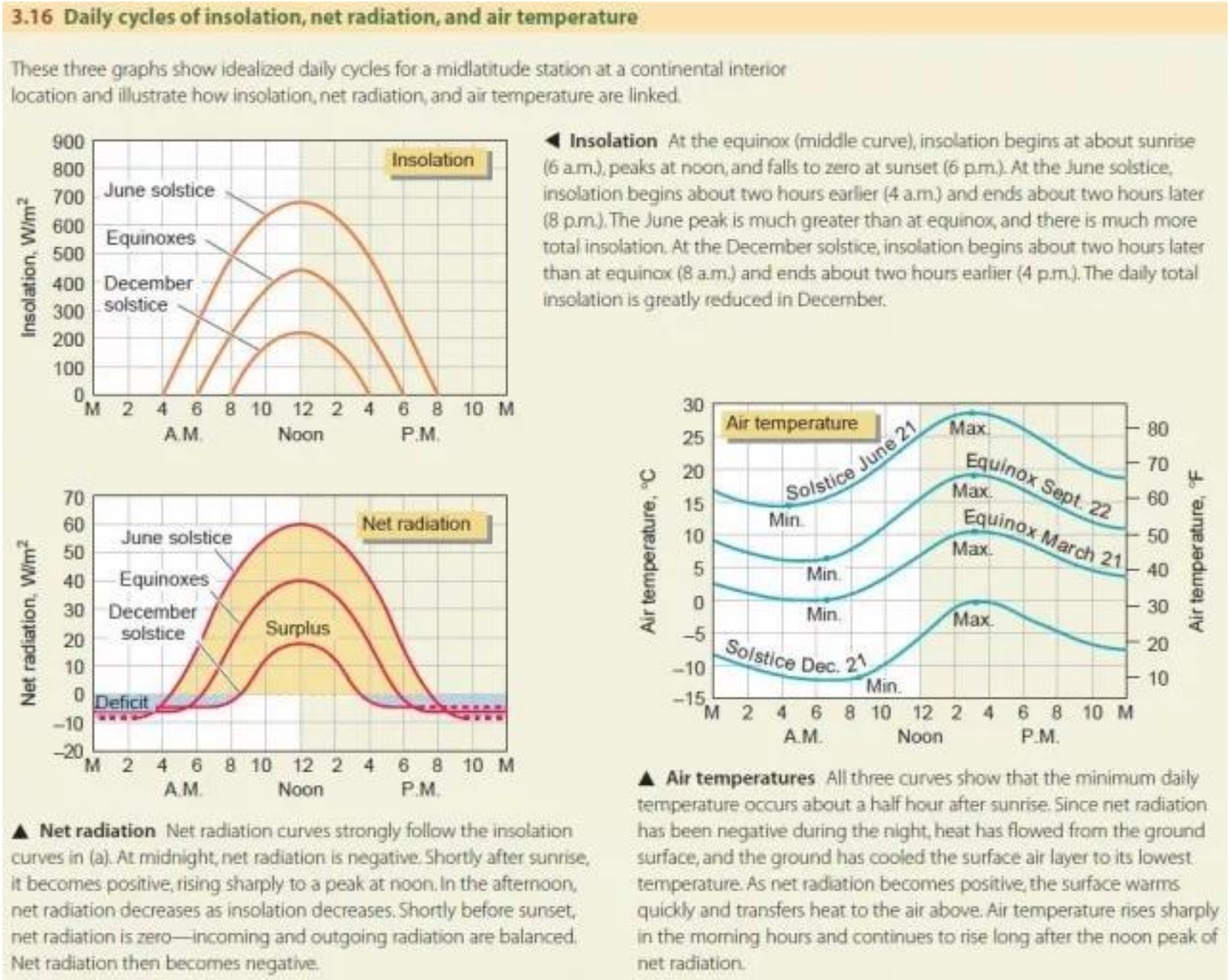
- Greenhouse gases "dirties" the atmospheric window
  - Radiation can escape at wavelengths that are not "dirty"
  - Some radiation can escape at wavelengths that are "dirty", but proportionally less as more gases make window "dirtier"



IPCC 1990

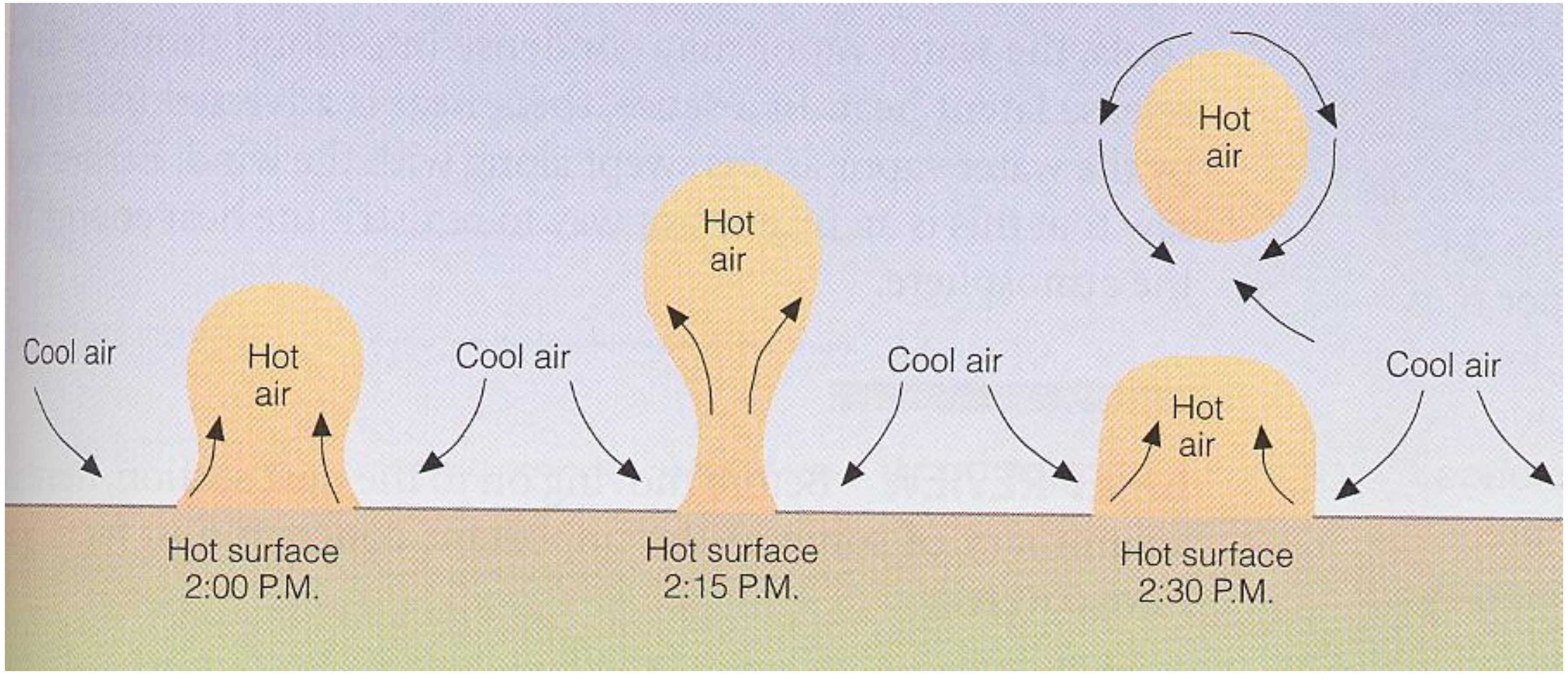
Daily temperature variations

- Each day is like a mini seasonal cycle
  - Sun rays most intense around noon
  - As is the case with the seasons, the maximum temperatures lag the peak incoming solar radiation.
- An understanding of the diurnal cycle in temperature requires an understanding of the different methods of atmospheric heating and cooling:
  - Radiation
  - Conduction
  - Convection



Atmospheric Heating

- Sunlight warms ground
- Ground warms adjacent air by conduction
  - Poor thermal conductivity of air restricts heating to a few cm
- Random motion of “hot” surface air molecules upward leads to heat transfer (diffusion)
- Hot air forms rising air “bubbles” (thermals) leading to convection
  - Mechanical mixing due to wind enhances this mode of heat transport



Vertical temperature profiles - day

- Windy conditions do not allow the formation of a large thermal gradient (superadiabatic) between the atmosphere and the (hot) soil surface, because of the mixing

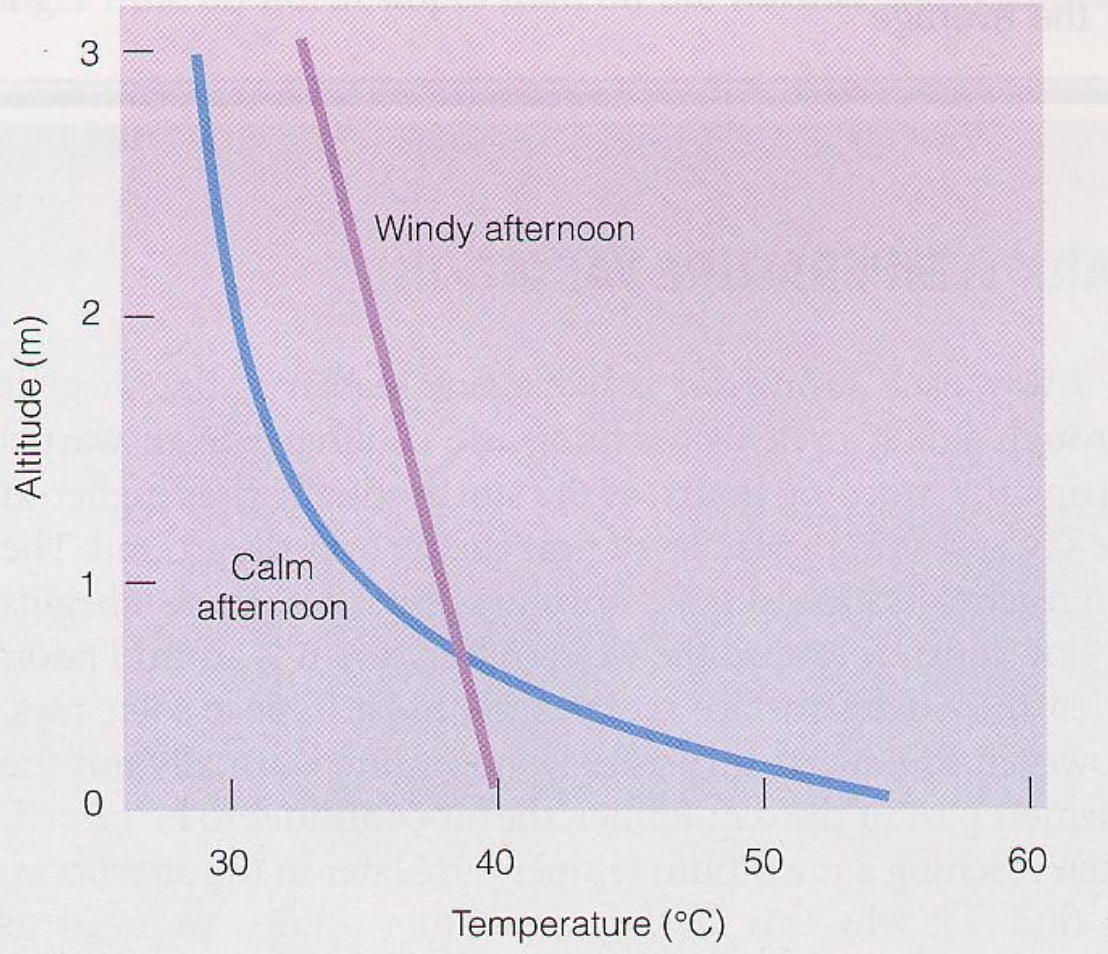


FIGURE 3.13
Vertical temperature profiles above an asphalt surface for a windy and a calm summer afternoon.

Balance of incoming/outgoing energy

- Energy accumulates in the "system" as long as $E_{in} > E_{out}$
- Added energy leads to increasing T
- T_{max} occurs when E_{out} surpasses E_{in}
- T_{max} typically 3-5 p.m. for clear summer skies
  - What factors could change timing?

Solving analitically the Fourier heat conduction equation, by imposing a sinusoidal behavious for temperature forcing at the surface (z=0), the result shows that, between the maximum of the incident radiation (which occurs at 12 a.m., if the sky is not cloudy) and the one of the temperature at z=0 is π/4 (with respect to 2π, the angle lenght of a day, e.g. 1/8 of the daily period). Converting the result in hours, the shift is of 3 hours, which means that the maximum value of T at z=0 occurs at about 3 p.m.

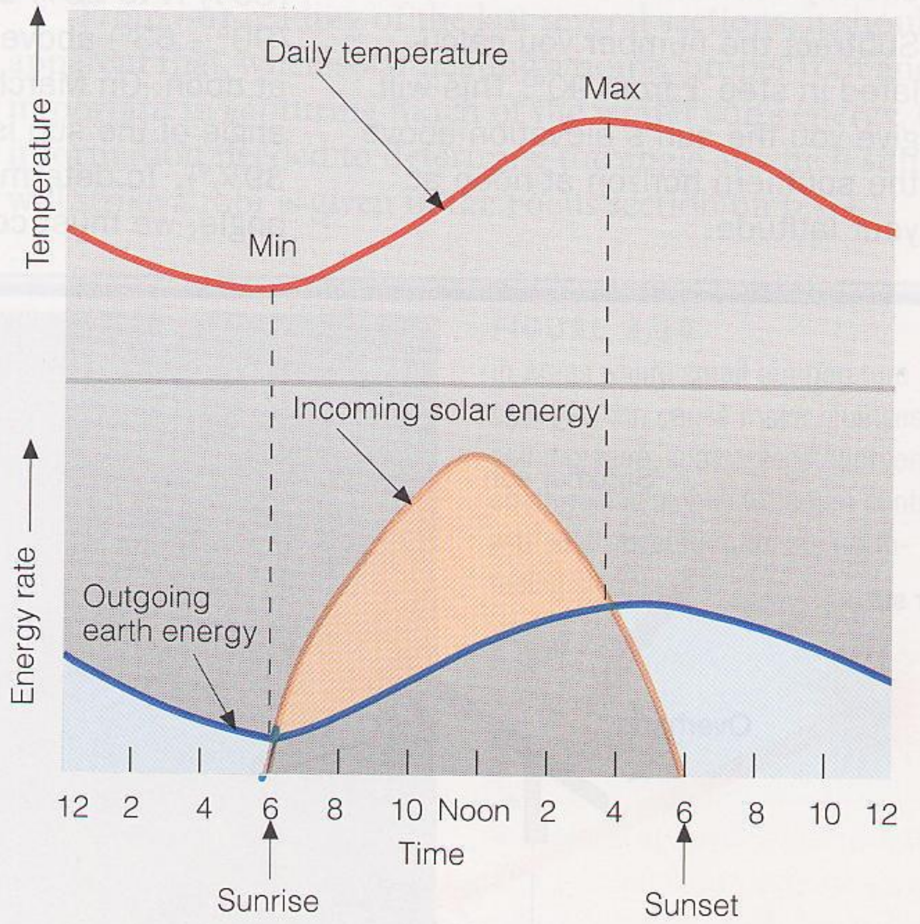


FIGURE 3.14
The daily variation in air temperature is controlled by incoming energy (primarily from the sun) and outgoing energy from the earth's surface. Where incoming energy exceeds outgoing energy (orange shade), the air temperature rises. Where outgoing energy exceeds incoming energy (blue shade), the air temperature falls.

T_{max} Factors

- T_{max} depends on
  - Cloud cover
    - In clear sky conditions, solar radiation is larger
  - Surface type
    - » Absorption characteristics
      - Strong absorbers enhance surface heating
    - » Vegetation/moisture
      - Over wet terrain (or vegetation), available energy is partially used to evaporate water
  - Wind
    - » Strong mixing by wind will mix heated air near ground to higher altitudes

Cooling at Night

- Lower sun angle in evening → $E_{in} < E_{out}$
- During night the ground and air radiate energy to space
  - Ground is better radiator than air, so ground cools faster (Kirchoff's law)
- Ground cools adjacent air layer by conduction
  - Heat transfer from air above is slow due to poor thermal conductivity of air.
  - Why isn't convection as important at night?
    - Because atmosphere is NOT unstable

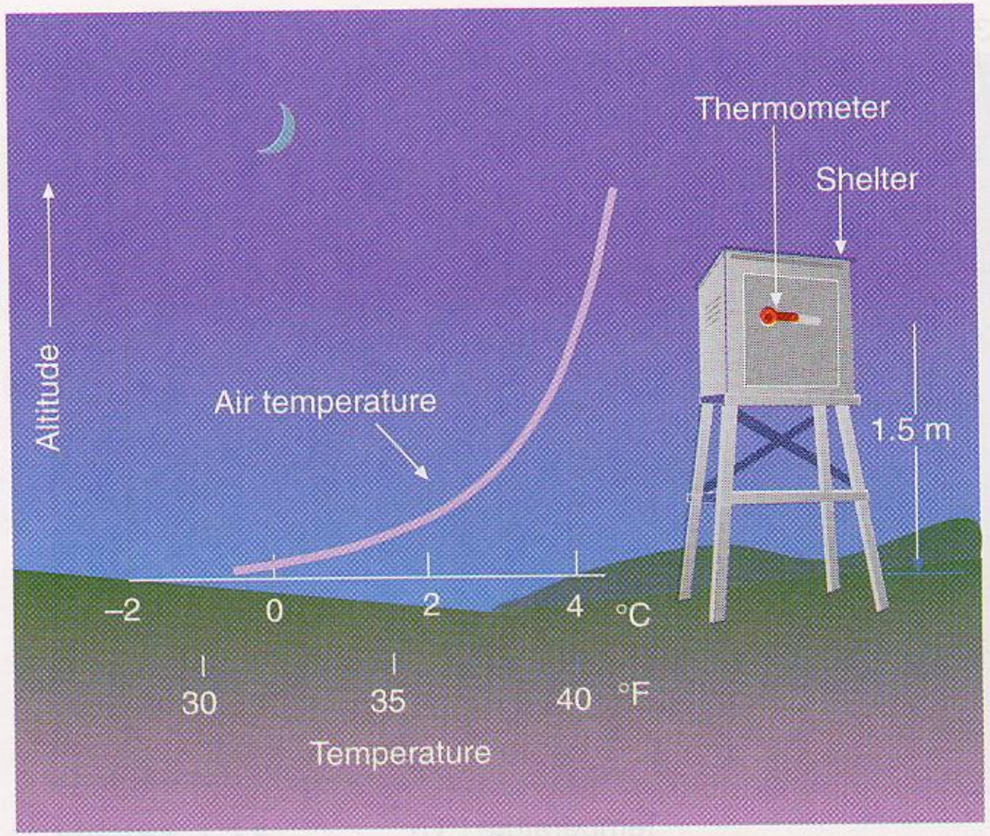


FIGURE 3.15
On a clear, calm night, the air near the surface can be much colder than the air above. The increase in air temperature with increasing height above the surface is called a radiation temperature inversion.

Radiation Inversions

- Radiation inversions
  - Feature increasing T with height above ground
  - Occur on most clear, calm nights (nocturnal inversion)
  - Are strongest when
    - » Winds are calm
    - » Air is dry
      - Less IR absorption
      - Less latent heat release from fog/dew formation
    - » Sky is cloud free
    - » Night is long

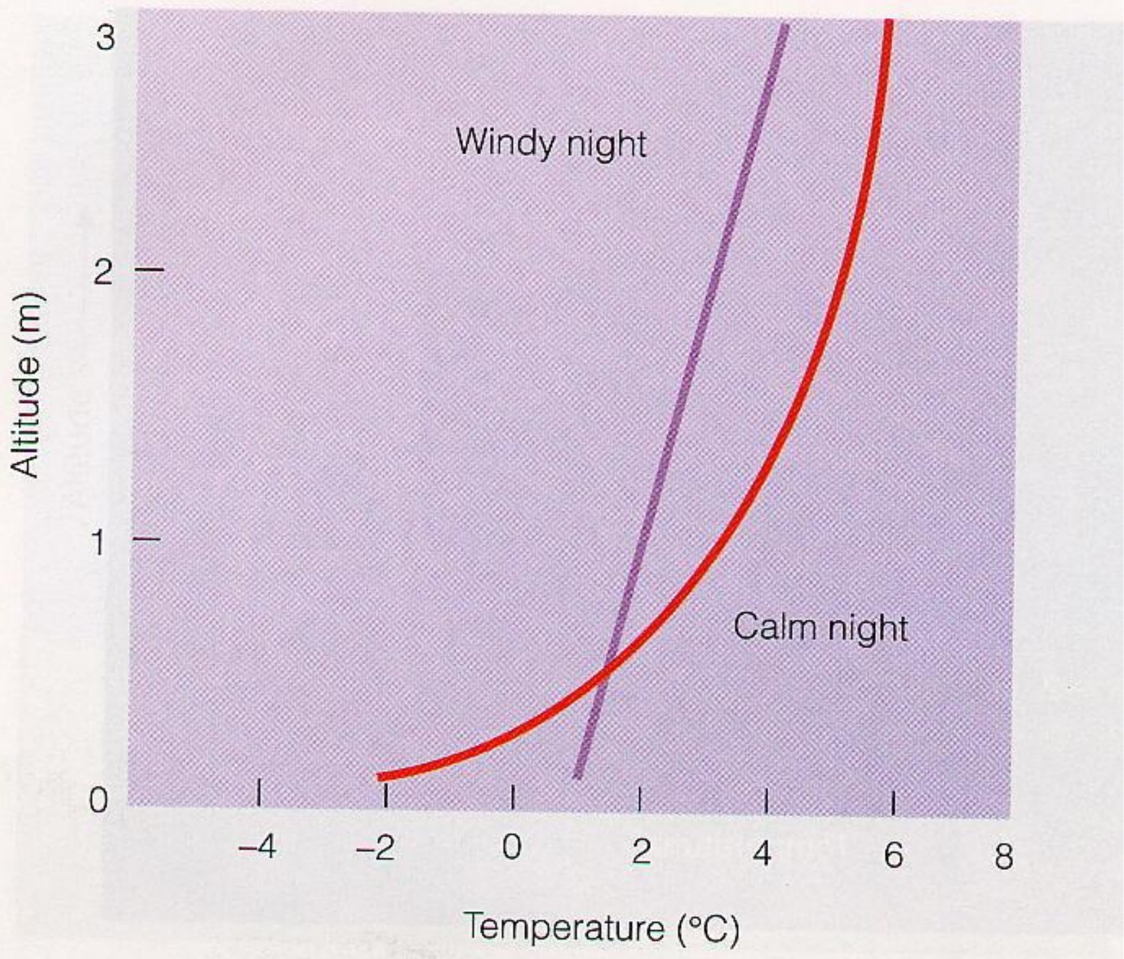


FIGURE 3.16
Vertical temperature profiles just above the ground on a windy night and on a calm night. Notice that the radiation inversion develops better on the calm night.

Cold air pooling

- As air cools, its density increases, leading to:
  - Downslope flow
  - Cold air pooling in low spots/valleys

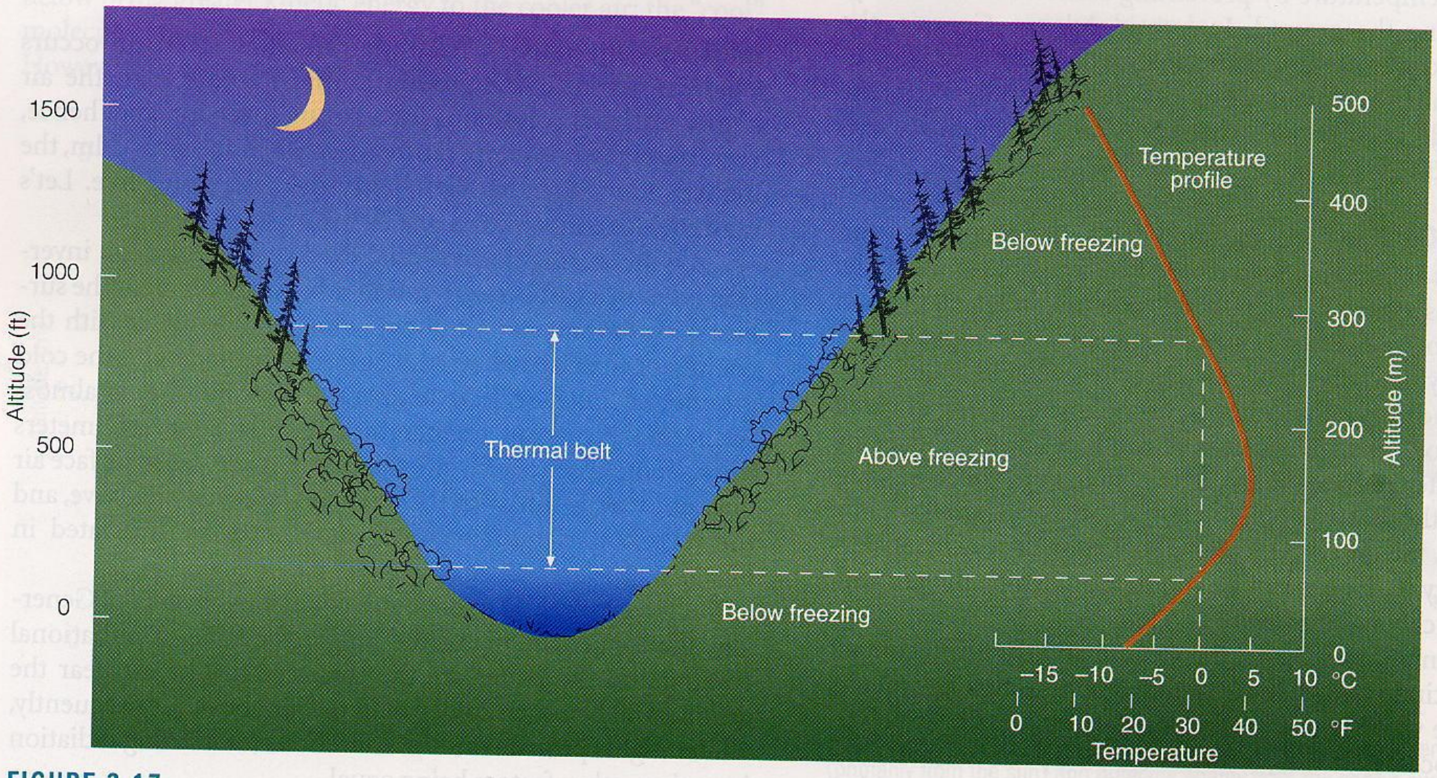
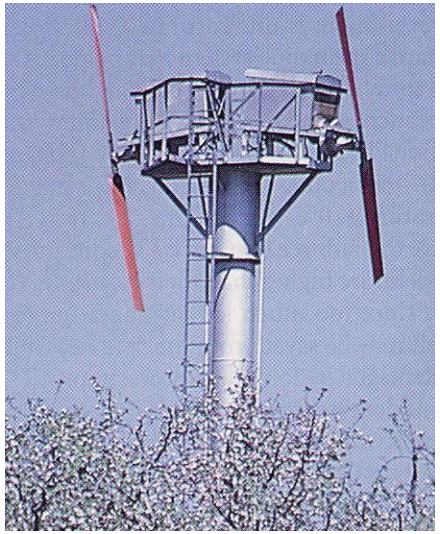


FIGURE 3.17
On cold, clear nights, the settling of cold air into valleys makes them colder than surrounding hillsides. The region along the side of the hill where the air temperature is above freezing is known as a *thermal belt*.

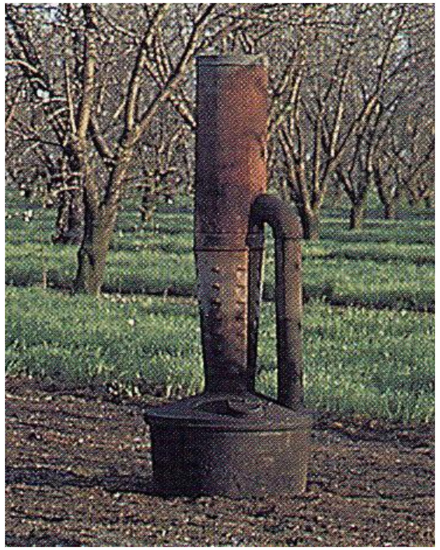
Pozza Tramontana, example of «dolina» (pseudokarst sinkhole), quite frequent in territories with karst

Inversions and agriculture

- Surface cooling can produce sub-freezing temperatures near ground
- Crop damage possible
  - Especially problematic for orchards in “warm” climates
- Mixing down of warmer air aloft can help limit temperature drop near surface



- Big fans and ventilators in order to create mixing and avoid to reach too low temperatures at ground



- Use of fires to warm low atmosphere and create mixing, preventing minima too low