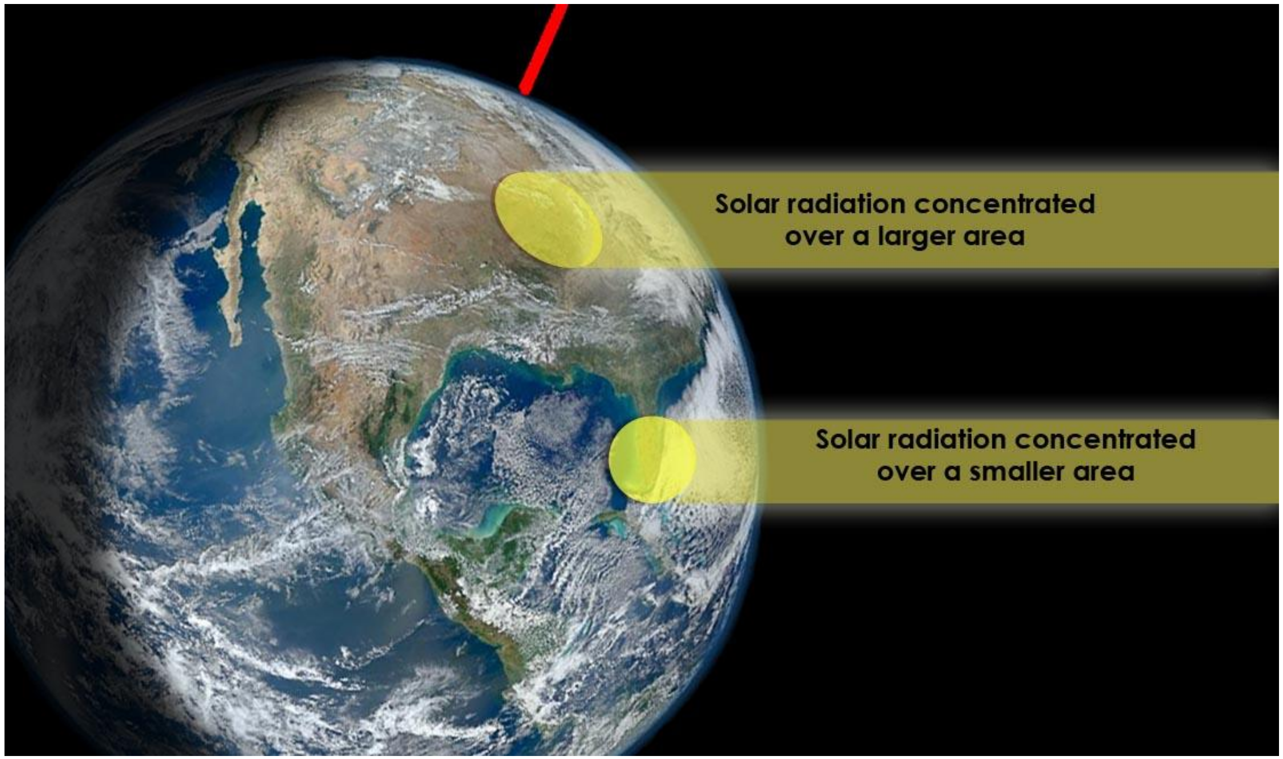
Lesson 8:
Solar Radiation at the Top of the Atmosphere



Course: Laboratory of Atmospheric Remote Sensing
Laurea Magistrale in Atmospheric Science and Technology

Content

- Energy from the Sun
- Solar constant, total solar irradiance, mean irradiance
- Thermal infrared radiation
- Equilibrium temperature of the earth
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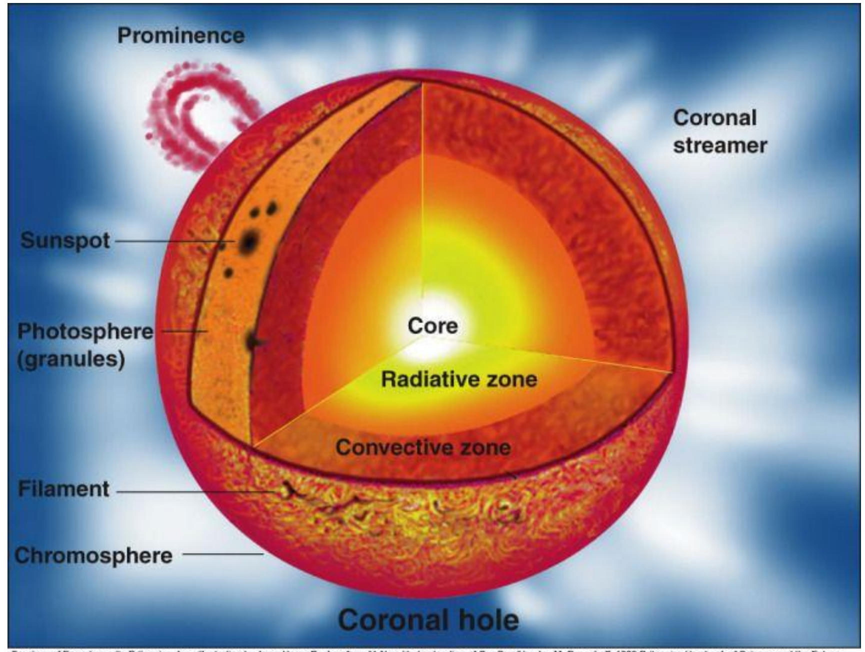
- Terrestrial Radiation
- Annual mean net (downward) shortwave radiation, annual mean net (outgoing) longwave radiation, net radiation imbalance at the TOA
- Single atmospheric layer above the surface: grey atmosphere, black/white atmosphere, selective atmosphere
- Two-layers atmosphere above the surface
- Earth-atmosphere energy balance: sensible and latent heat fluxes

Reading material:
Liou K. N., An introduction to atmospheric radiation. Chps. 2-4.

Energy from the Sun (1)



Courtesy of Encyclopaedia Britannica, Inc.; illustration by Anne Hoyer Becker; from "A New Understanding of Our Sun," by Jay M. Pasachoff, 1989 Britannica Yearbook of Science and the Future

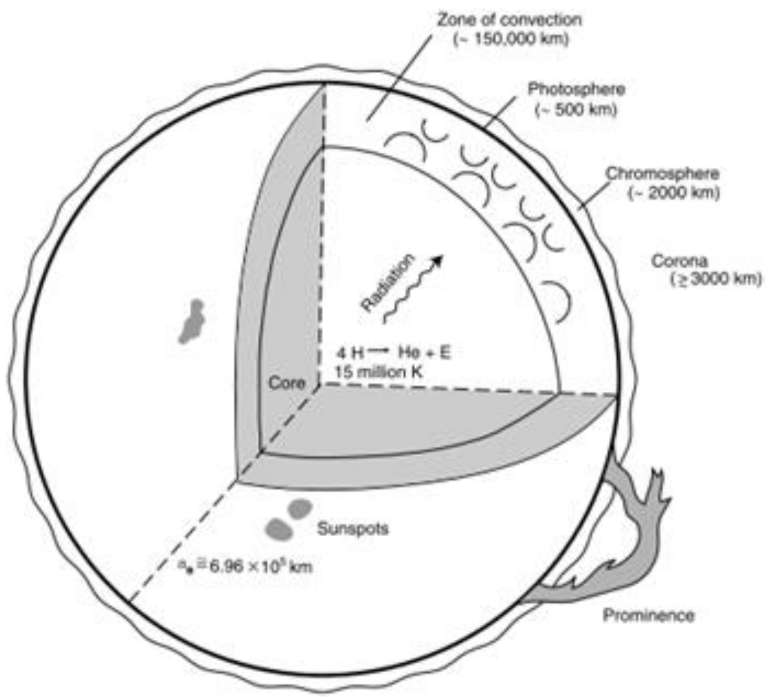
- $R_s \cong 6{,}96{\times}10^5$ km
- $M_s \cong 1{,}99{\times}10^{30} kg$
- Main constituents: hydrogen (90%), helium (10%), small amount of heavier elements including oxygen, carbon, nitrogen, neon, iron, silicon, magnesium, sulfur, and calcium.
- T_s decreases from a central value of $\cong 5{\times}10^6 K$ to $\cong 5800\ K$at the surface
- $\rho_s \cong 1{,}40\ g\ cm^{-3}$

Approximately 90% of the sun's mass is contained in the inner half of its radius.

Energy from the Sun (2)

Solar energy is believed to be generated by the steady conversion of four hydrogen atoms to one helium atom in fusion reactions.

The amount of energy released in nuclear fusion causes a reduction of the sun's mass.



According to Einstein's law relating mass and energy and converting the energy radiated by the sun, we find that almost 5 million tons of mass per second are radiated by the sun in the form of electromagnetic energy.

Energy from the Sun (3)

1. As a result of the extremely high temperatures in the deep interior of the sun, collisions between atoms are sufficiently violent to eject many electrons from their orbits.

2. The energy emitted by nuclear fusion in the form of photons can pass through the inner part of the sun without being absorbed by the electrons. However, closer to the sun's surface, the temperature decreases and the heavier atoms such as iron begin to recapture their outer electrons.

3. The flow of photons coming from the interior is blocked by the appearance of the absorbing atoms, that cause the temperature to drop sharply at some depth below the surface. Thus, the outer region of the sun consists of a layer of relatively cool gas resting on top of a hotter interior.

4. As a consequence, the gas at the bottom of the cool outer layer is heated by the hot gas in the interior, so it undergoes expansion and rises toward the surface.

5. Once it reaches the surface, the hot gas loses its heat to space, cools, and descends into the interior.

Energy from the Sun (4)

The entire outer layer breaks up into ascending columns of heated gas and descending columns of cooler gas.

The region in which this large-scale upward and downward movement of gases occurs is called the **zone of convection**, which extends from a depth of about 150,000 km to the surface of the sun.

Below this depth, energy is transported within the sun by means of **electromagnetic radiation**.
Near the surface, because of the blocking of radiant energy, energy is transferred by **convection** and **electromagnetic radiation**.
Above the surface, energy transport is again by means of **electromagnetic radiation**.

Solar constant (1)

Is the distribution of electromagnetic radiation emitted by the sun, as a function of the wavelength incident on the top of the earth's atmosphere.

It is defined as the flux of solar energy (energy per unit time) across a surface of unit area normal to the solar beam at the mean distance between the sun and the earth and denotes the amount of total solar energy reaching the top of the atmosphere (TOA).

$$\boxed{TSI \cong 1360 W\ m^{-2}}$$

Total solar irradiance (TSI) or Solar constant

Solar constant (2)

Although historically it is called the *solar constant*, S is variable over time according to the temporal trend of the average and instantaneous activity of the solar activity.
For this reason, the most suitable scientific term to describe it is **Total Solar Irradiation (TSI).**
This name also suggests to us that we are talking about an irradiance.

The value has been obtained for many years from measurements made with instruments housed on satellites that rotate around the Earth and with surface measurements (which cover the pre-satellite period), recalibrated on the satellite ones.
The present value was obtained with observations made by the Total Irradiance Monitor (TIM) instrument installed on the satellite (SOURCE)

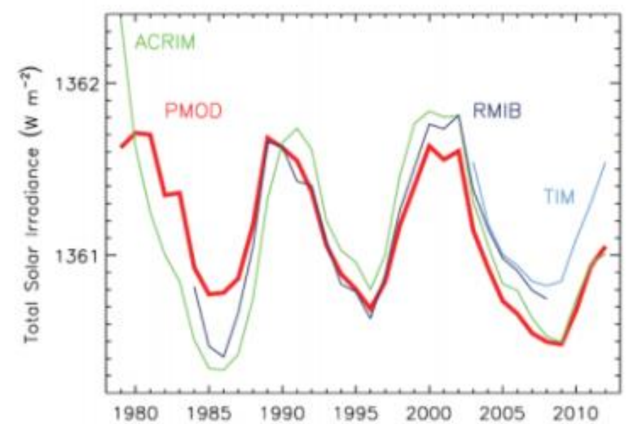


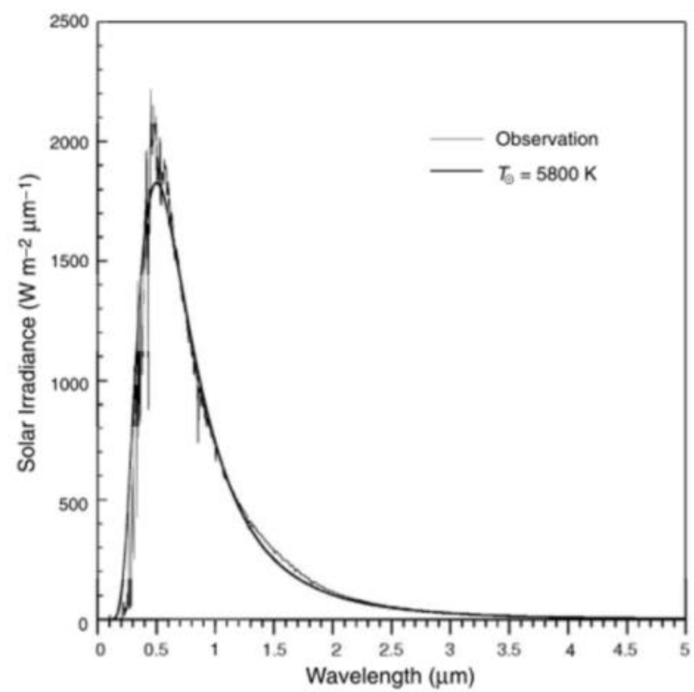
Figure 8.10 | Annual average composites of measured total solar irradiance: The Active Cavity Radiometer Irradiance Monitor (ACRIM) (Willson and Mordvinov, 2003), the Physikalisch-Meteorologisches Observatorium Davos (PMOD) (Frohlich, 2006) and the Royal Meteorological Institute of Belgium (RMIB) (Dewitte et al., 2004).These composites are standardized to the annual average (2003–2012) Total Irradiance Monitor (TIM) (Kopp and Lean, 2011) measurements that are also shown.

$$\boxed{TSI = 1360.8 \pm 0.5\, W\, m^{-2}}$$

Solar constant (3)

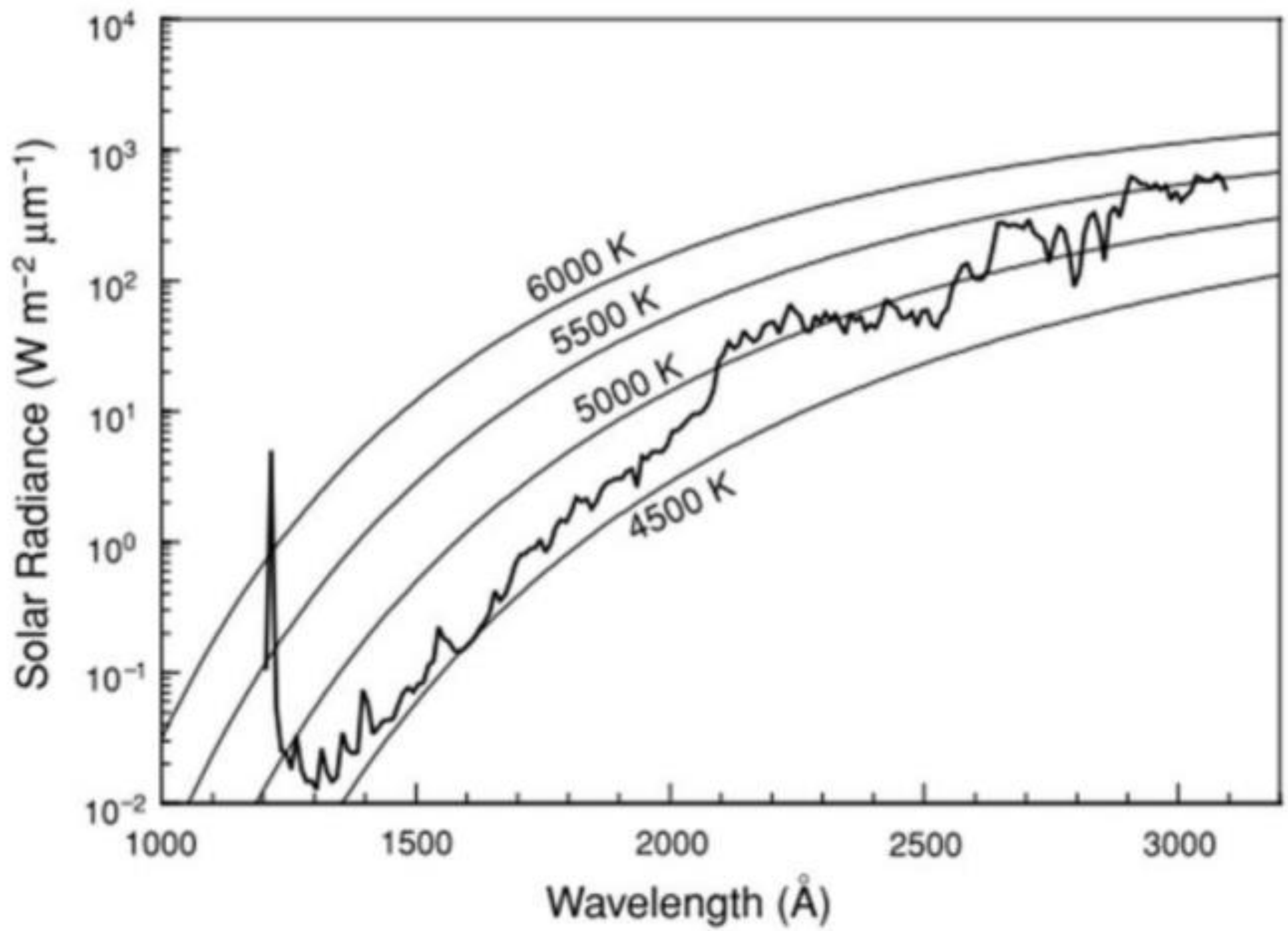
The Stefan-Boltzmann law suggests that, if the Sun were a black body, to obtain this value of the solar constant, it would have to be at a temperature of about 5800 K.
Wien's law also tells us that the maximum emission should be around 500 nm, which corresponds to yellow-green light.
In reality, the solar spectrum in some spectral regions also differs substantially from that of a black body at 5800 K.

The reasons for the discrepancies are based on the physical mechanisms underlying the solar emission.
In fact, the solar radiation that reaches the Earth is produced by a layer called the photosphere which is located at about 5800K, but part of the radiation is absorbed by the outer layers, especially in the UV, and emissions from more internal and warmer layers are not totally absorbed and they reach the Earth.



Solar constant (4)

Distortions of the spectrum compared to that of a black body at 5800 K are found mainly in UV.
Note the emission of radiation at λ=128 nm (Lyman α) produced by the de-excitation of the hydrogen atom.



Total solar irradiance (1)

By identifying the Sun with a black body at 5800 K the emitted irradiance will be equal to

$$F_{\odot} = \sigma T^4 = 5.67x10^{-8}x5800^4 = 6.41x10^7 W m^{-2}$$

Considering the radius of the photosphere equal to $R_S = 7x10^5 km = 7x10^8 m$ we obtain a radiative flux emitted isotropically by the Sun equal to

$$f_{\odot} = 4\pi R_S^4 \cong 4x10^{27} W$$

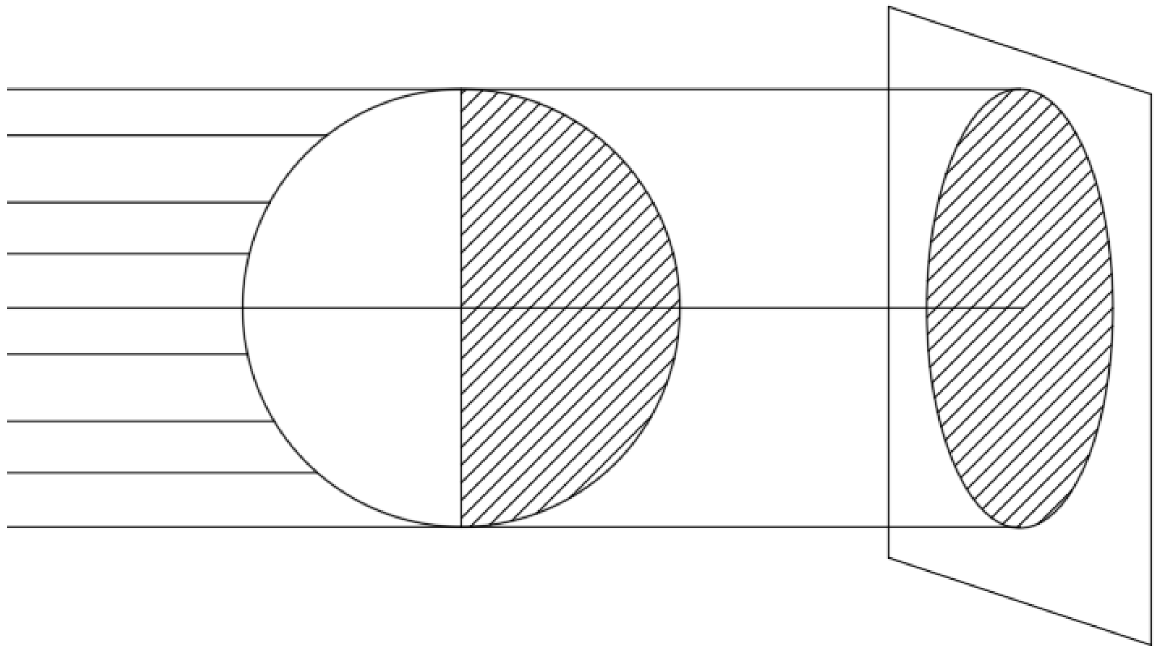
If we assume that there is no absorption in the space between the Sun and Earth (we neglect the effect of the presence of Mercury and Venus and other interplanetary matter), the same flux will cross a sphere with a radius equal to the average distance between the Sun and the Earth ($R_{S-E} = 1.5x10^{11} m$).
The irradiance that crosses the surface of this sphere will be:

$$\boxed{I_{S-E} = \frac{f_{\odot}}{4\pi R_{S-E}^4} = 1390\ W\ m^{-2} \approx TSI = S}$$

Total solar irradiance (2)

TSI is the power emitted by the Sun that hits a unit surface at the distance $1,5 \times 10^{11}$ m (1 astronomic unit).

How much is intercepted by a unit area on the Earth planet?



Shadow produced by the planet $A_S = \pi r_E^2$

In the dark side the insolation is zero.
In the illuminated side the inclination of the surface change from 0° up to 90°.

Mean irradiance

$f = S\pi r_E^2$ mean intercepted flux

$\bar{Q} = \frac{S\pi r_E^2}{4\pi r_E^2}$ ⟵ Area of a sphere with radius r_E

$\bar{Q} = \frac{S}{4} = 340\ W\ m^{-2}$

f can be computed also as $\int_{emi} S cos\theta dA = S \int_{2\pi} \int_{\pi/2} r_E^2 cos\theta d\Omega = \pi r_E^2 S$

Thermal infrared radiation

- The earth–atmosphere system reflects about 30% of the incoming solar radiation at the TOA.

- A large portion of the incoming solar radiation is absorbed by the earth's surface (70% ocean, 30% land.)

The global equilibrium temperature of the earth–atmosphere system remains relatively constant. Consequently, <u>radiant energy emitted from the sun that is absorbed in the earth–atmosphere system must be re-emitted to space so that an equilibrium energy state can be maintained</u>.
This emitted radiation is defined **thermal infrared radiation**.

Equilibrium temperature of the earth–atmosphere system (1)

$\bar{r}$ = global albedo of the earth–atmosphere system
r_E= earth's radius
S = solar constant
T_E = equilibrium temperature of the earth– atmosphere system

$$F_{in} = \frac{S}{4}(1 - \bar{r}) = 238\, W\ m^{-2}$$

Radiative equilibrium $\rightarrow\ F_{in} = F_{out}$

$$\frac{S}{4}(1 - \bar{r}) = \sigma T_E^4 \quad \Rightarrow \quad \boxed{T_E = \sqrt[4]{(1 - \bar{r})S/4\sigma} \cong 255\ K = -18\ °C}$$

From Planck's and Wien's displacement laws, the radiance emitted from the earth and the atmosphere is smaller than that of solar radiation, whereas the wavelength for the intensity peak of the earth's radiation field is longer than that of solar radiation.

Equilibrium temperature of the earth–atmosphere system (2)

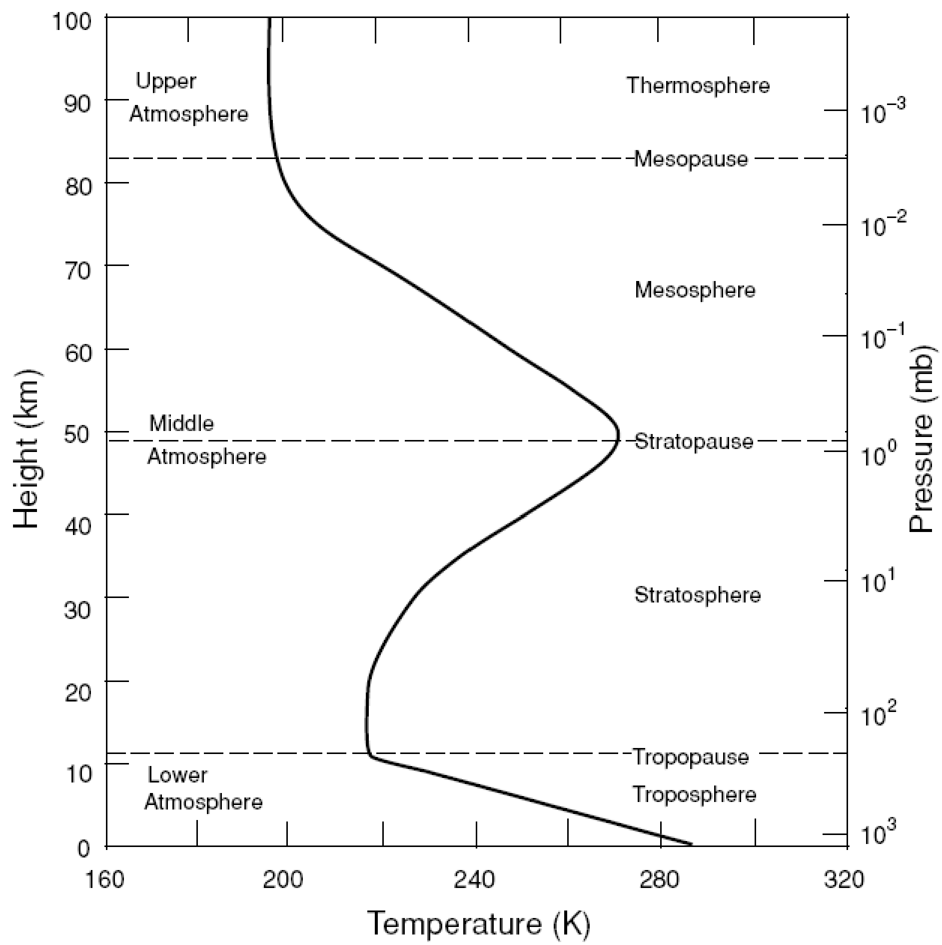


Figure 3.1 Vertical temperature profile after the U.S. Standard Atmosphere and definitions of atmospheric nomenclature.

Terrestrial Radiation (1)

The energy emitted from the earth–atmosphere system is defined as **thermal IR** or **terrestrial radiation**.

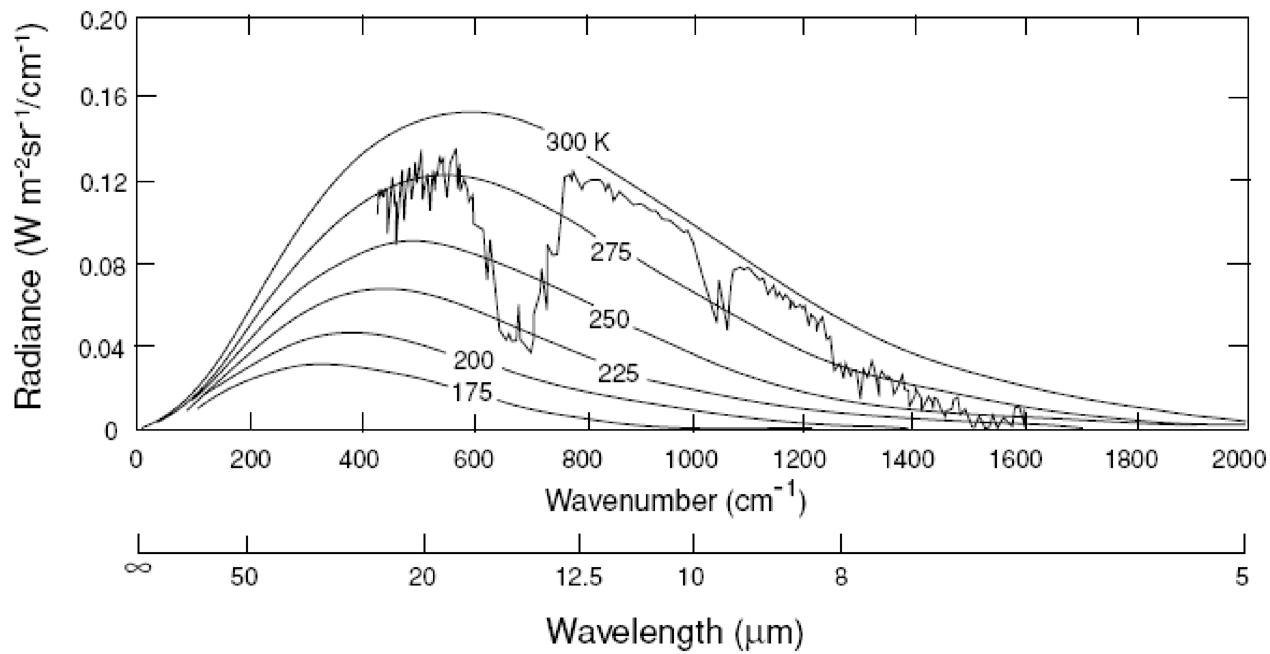


Figure 4.1 Theoretical Planck radiance curves for a number of the earth's atmospheric temperatures as a function of wavenumber and wavelength. Also shown is a thermal infrared emission spectrum observed from the Nimbus 4 satellite based on an infrared interferometer spectrometer.

The spectral distribution of radiance emitted by a blackbody source at various temperatures in the earth's atmosphere.

Terrestrial Radiation (2)

$T_S = 288\ K$ mean climatological surface temperature

$\boxed{255\ \mathrm{K} = T_E < T_S = 288\ K}$ $\Rightarrow$ greenhouse effect

The average temperature of the earth's surface is 288 K (15 ° C), so the surface cannot be considered responsible for the radiative emission that compensates for the absorption of solar radiation.

In fact, the surface is surrounded by a gaseous layer (the atmosphere) which almost completely absorbs the radiation emitted by the surface, and emits radiation both towards the surface and towards space.

Since the emission of radiation towards space due to the surface is only 15% of what is absorbed by solar radiation, 85% is emitted from the atmosphere.

Terrestrial Radiation (3)

The impact of the atmosphere on the radiation balance produces the so-called **greenhouse effect**: the interaction of the atmosphere with solar radiation and with that emitted from the surface and the emission of atmospheric radiation towards the surface and towards space result in the maintenance of a surface temperature 33 °C higher than that which would be obtained without the presence of the atmosphere.

We will conceptually demonstrate the action of the greenhouse effect with a simplified model of the Space-Atmosphere-Surface system, in which the atmosphere is described by a single layer with different emissivity in the visible (ε_{vis}) and in the thermal infrared (ε_{irr}), the treated surface as a black body on the whole spectrum and space is limited by the Top Of Atmosphere (TOA) in which solar radiation enters and terrestrial radiation exits.

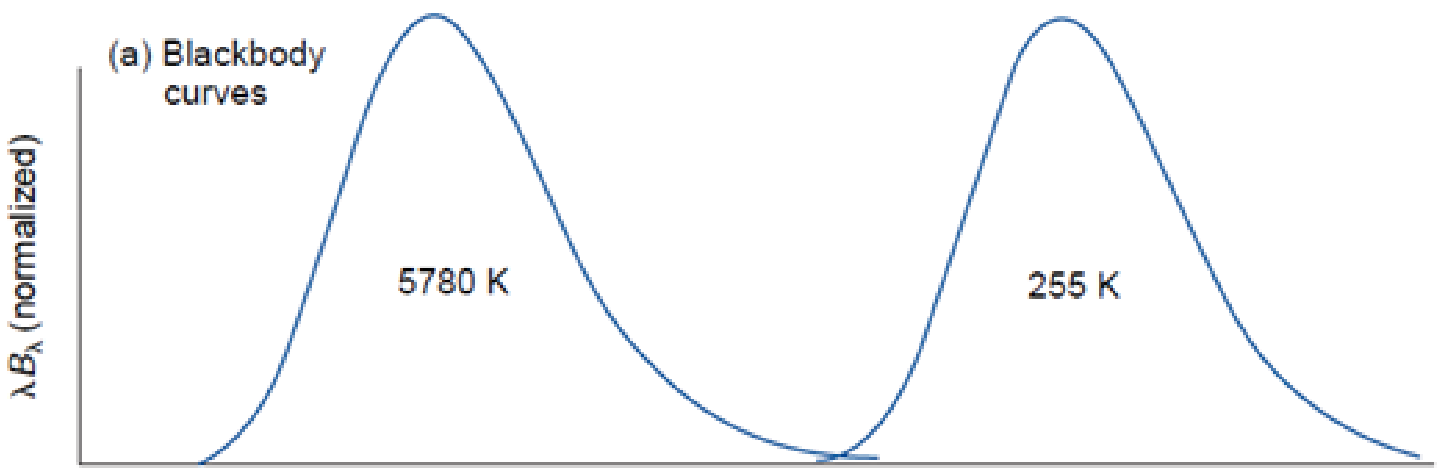
Terrestrial Radiation (4)

$T_S = 288\ K$ mean climatological surface temperature

$255\ \mathrm{K} = T_E < T_S = 288\ K \quad \Rightarrow$ greenhouse effect

For Kirchhoff law: $\varepsilon_\lambda = \alpha_\lambda$

Atmosphere with different emissivity in the shortwave and longwave.



Terrestrial Radiation (5)

Solar radiation is also called **shortwave radiation** because solar energy is concentrated in shorter wavelengths with its peak at about 0.5 μm.

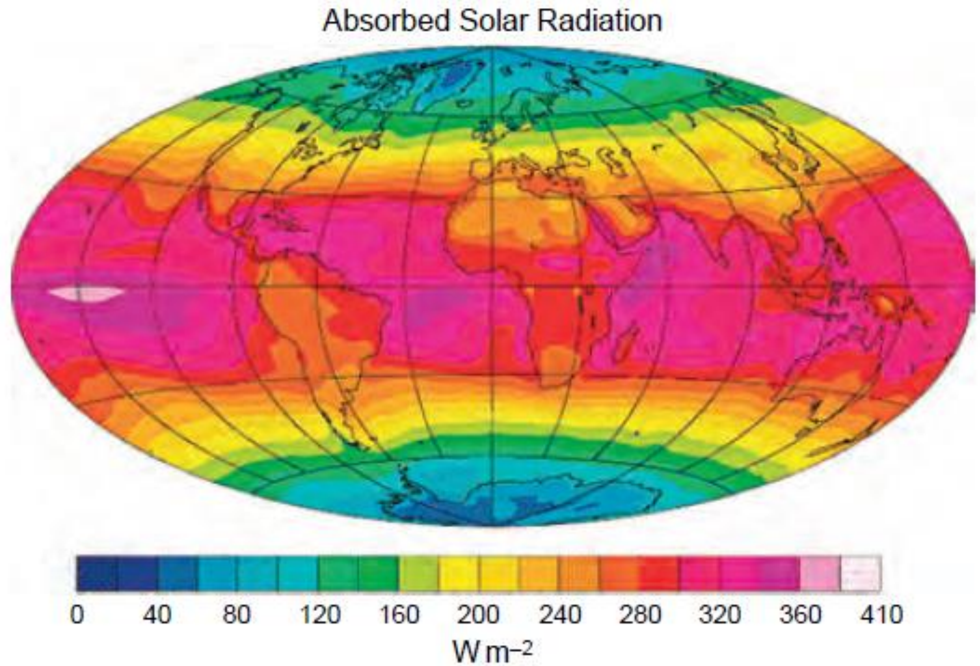
Thermal infrared radiation emitted from the earth and the atmosphere is also referred to as **outgoing longwave radiation** (OLR) because its maximum energy is in the longer wavelengths at about 10 μm.
The solar and infrared spectra are separated into two spectral ranges above and below about 5 μm, and the overlap between them is relatively small.

Terrestrial Radiation (6)

In a clear atmosphere without clouds or aerosols about 50% of solar energy transmits through the atmosphere and is absorbed by the earth's surface.
In contrast, energy emitted from the earth is largely absorbed by carbon dioxide, water vapor, ozone, and other trace gases in the atmosphere.

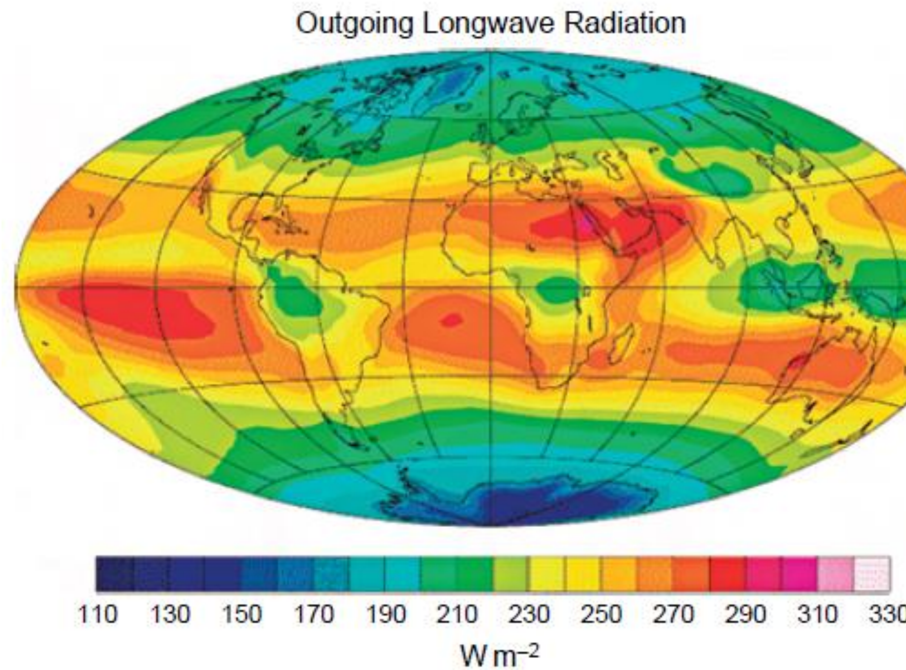
The trapping of thermal infrared radiation by atmospheric gases is typical of the atmosphere and is therefore called the **atmospheric effect**.
It is also referred to as the **greenhouse effect**.

Annual mean net (downward) shortwave radiation



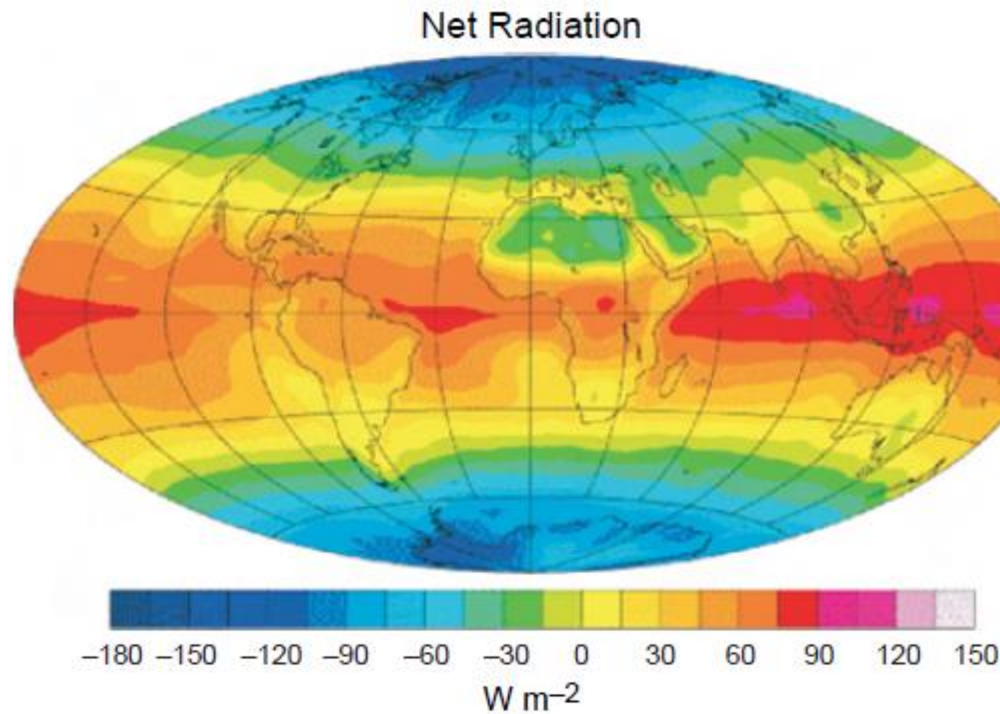
- At the tropics ≈300 W m^2: the sun is nearly directly overhead at midday throughout the year.
- Within the tropics: highest values over cloud-free regions of the oceans, where annual-mean local albedos range as low as 0.1, lowest values over the deserts where albedos are ≈ 0.2 and locally range as high as 0.85.
- Polar regions ≈100 W m^2 : winters are dark and the continuous summer daylight is offset by the high solar zenith angles and the high albedo of ice.

Annual mean net (outgoing) longwave radiation



- Gentler equator-to-pole gradient and more regional variability within the tropics.
- The regions of low OLR over Indonesia and parts of the tropical continents reflect the prevalence of deep convective clouds with high, cold tops.
- Highest annual mean OLR: deserts and equatorial dry zones over the tropical Pacific, where the atmosphere is relatively dry and cloud free.

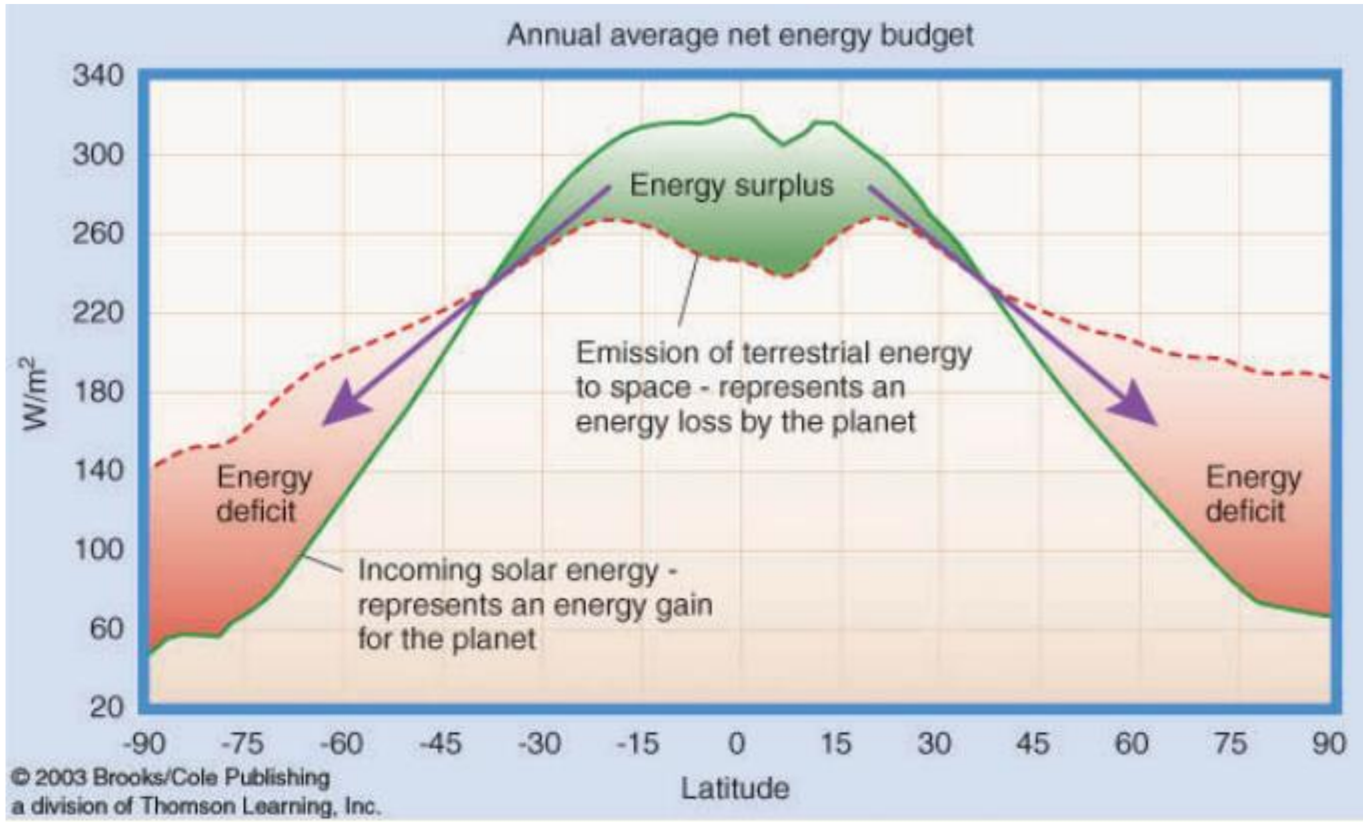
Net radiation imbalance at the TOA (1)



- ❑ surplus of incoming solar radiation over outgoing longwave radiation in low latitudes
- ❑ deficit in high latitudes

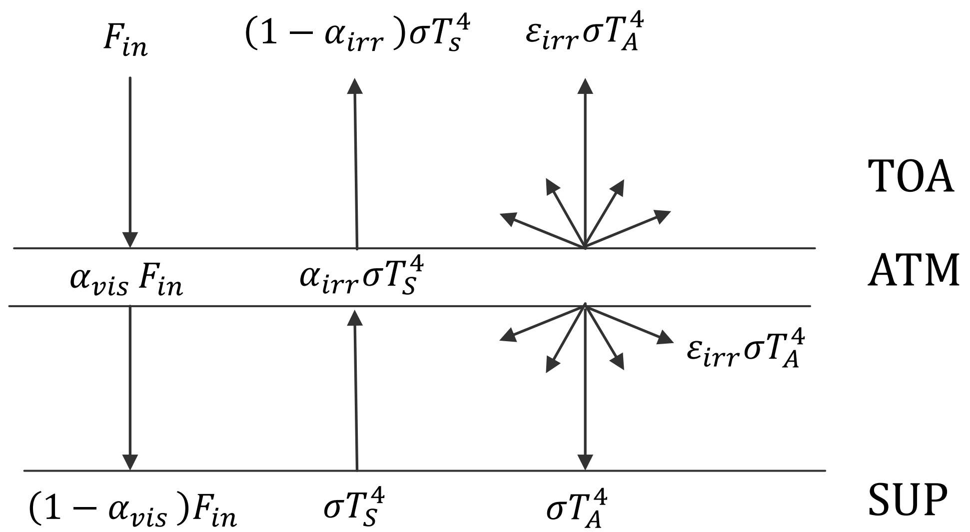
❗ over some of the world's hottest desert regions, the outgoing longwave radiation exceeds absorbed solar radiation

Net radiation imbalance at the TOA (2)



Zonally averaged radiation balance in the atmosphere.

Single atmospheric layer above the surface (1)



ε_{irr} = emissivity in infrared
ε_{vis} = emissivity in visible
α_{irr} = absorbivity in infrared
α_{vis} = absorbivity in visible
T_S = temperature of the surface
T_A = temperature of the layer of atmosphere
σ = Stefan-Boltzmann constant

Single atmospheric layer above the surface (2)

$$\begin{cases} F_{in} = (1-\alpha_{irr})\sigma T_s^4 + \varepsilon_{irr}\sigma T_A^4 & TOA \\ \alpha_{vis} F_{in} + \alpha_{irr}\sigma T_S^4 = 2\varepsilon_{irr}\sigma T_A^4 & ATM \\ (1-\alpha_{vis})F_{in} + \varepsilon_{irr}\sigma T_A^4 = \sigma T_S^4 & SUP \end{cases}$$

3 equations, 2 unknowns

F_{in}, α_{irr} , α_{vis} known

$$\begin{cases} F_{in} = (1-\varepsilon_{irr})\sigma T_s^4 + \varepsilon_{irr}\sigma T_A^4 & TOA \\ \varepsilon_{vis} F_{in} + \varepsilon_{irr}\sigma T_S^4 = 2\varepsilon_{irr}\sigma T_A^4 & ATM \\ (1-\varepsilon_{vis})F_{in} + \varepsilon_{irr}\sigma T_A^4 = \sigma T_S^4 & SUP \end{cases}$$

$$F_{in} - (1-\varepsilon_{irr})\sigma T_s^4 = \varepsilon_{irr}\sigma T_A^4 \qquad TOA$$

$$(1-\varepsilon_{vis})F_{in} + F_{in} - (1-\varepsilon_{irr})\sigma T_s^4 = \sigma T_S^4 \qquad SUP$$

$$\Downarrow$$

$$(2-\varepsilon_{vis})F_{in} = (2-\varepsilon_{irr})\sigma T_s^4 \qquad SUP$$

Single atmospheric layer above the surface (3)

$$(2-\varepsilon_{vis})F_{in} = (2-\varepsilon_{irr})\sigma T_s^4 \qquad SUP$$

$$\frac{(2-\varepsilon_{vis})}{(2-\varepsilon_{irr})}F_{in} = \sigma T_s^4 \qquad SUP$$

$$F_{in} = \sigma T_E^4$$

$$\boxed{T_s = \sqrt[4]{\frac{(2-\varepsilon_{vis})}{(2-\varepsilon_{irr})}T_E^4} = T_E\sqrt[4]{\frac{(2-\varepsilon_{vis})}{(2-\varepsilon_{irr})}}} \qquad SUP$$

$$\sigma T_E^4 - (1-\varepsilon_{irr})\sigma T_s^4 = \varepsilon_{irr}\sigma T_A^4 \qquad TOA$$

$$T_A^4 = \frac{1}{\varepsilon_{irr}}T_E^4 - \frac{(1-\varepsilon_{irr})}{\varepsilon_{irr}}T_s^4 \qquad TOA$$

$$\boxed{T_A = T_E\sqrt[4]{\frac{1}{\varepsilon_{irr}} - \frac{(1-\varepsilon_{irr})}{\varepsilon_{irr}}(\frac{T_s}{T_E})^4}} \qquad TOA$$

Single atmospheric layer above the surface (4)

For **<u>grey atmosphere</u>**:

$\varepsilon_{irr} = \varepsilon_{vis} = \varepsilon$ $\Rightarrow$

$$T_s = T_E$$

$$T_A = T_E \sqrt[4]{\frac{1}{\varepsilon} - \frac{(1-\varepsilon)}{\varepsilon}}$$

There are no effects on the temperature T_A and T_S (no greehouse effect).

For **<u>black and white atmosphere</u>**:

$\varepsilon_{irr} = 1 \quad \varepsilon_{vis} = 0$ $\Rightarrow$

$$T_S = T_E \sqrt[4]{2} = 255 * 1.89 = 303\ K$$

$$T_A = T_E$$

The atmosphere (perfect BB in longwave) emits all the radiation towards the space.

Single atmospheric layer above the surface (5)

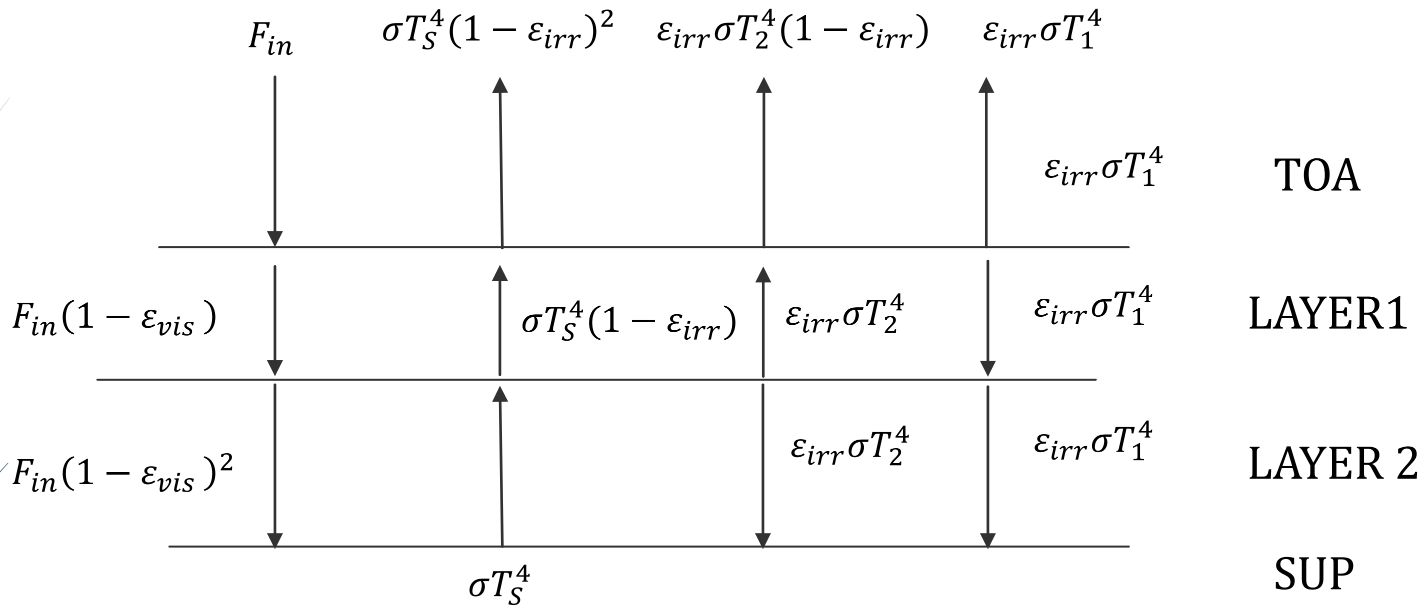
For **selective atmosphere**:

$\alpha_{irr} = 0.8 \quad \alpha_{vis} = 0.2$ $\Rightarrow$ $T_A = T_E \sqrt[4]{\frac{1}{0.8} - \frac{0.2}{0.8}\frac{1.8}{1.2}} = 247\ K < T_E$

$$T_S = T_E \sqrt[4]{\frac{2-0.2}{2-0.8}} = 282\ K \cong \bar{T}_S$$

The atmosphere is colder than T_E.
Part of the outgoing radiation comes from the space.

Two-layers atmosphere above the surface (1)



$$\varepsilon_{irr,1} = \varepsilon_{irr,2} = \varepsilon_{irr}$$
$$\varepsilon_{vis,1} = \varepsilon_{vis,2} = \varepsilon_{vis}$$
$$\varepsilon_s = 1$$

$$\begin{cases} F_{in}(1-\varepsilon_{vis})^2 + \varepsilon_{irr}\sigma T_2^4 + \varepsilon_{irr}\sigma T_1^4(1-\varepsilon_{irr}) = \sigma T_S^4 & SUP \\ \varepsilon_{vis} F_{in} + \varepsilon_{irr}\sigma T_S^4(1-\varepsilon_{irr}) + \varepsilon_{irr}{}^2\sigma T_2^4 = 2\varepsilon_{irr}\sigma T_1^4 & LAYER1 \\ \varepsilon_{vis}(1-\varepsilon_{vis})F_{in} + \varepsilon_{irr}\sigma T_S^4 + \varepsilon_{irr}{}^2\sigma T_1^4 = 2\varepsilon_{irr}\sigma T_2^4 & LAYER2 \end{cases}$$

Two-layers atmosphere above the surface (2)

Considering:

$\varepsilon_{irr} = 1$

$\varepsilon_{vis} = 0$

$F_{in} = \sigma T_E^4$

$$\begin{cases} F_{in}(1-\varepsilon_{vis})^2 + \varepsilon_{irr}\sigma T_2^4 + \varepsilon_{irr}\sigma T_1^4(1-\varepsilon_{irr}) = \sigma T_s^4 & SUP \\ \varepsilon_{vis} F_{in} + \varepsilon_{irr}\sigma T_S^4(1-\varepsilon_{irr}) + \varepsilon_{irr}{}^2\sigma T_2^4 = 2\varepsilon_{irr}\sigma T_1^4 & LAYER1 \\ \varepsilon_{vis}(1-\varepsilon_{vis})F_{in} + \varepsilon_{irr}\sigma T_S^4 + \varepsilon_{irr}{}^2\sigma T_1^4 = 2\varepsilon_{irr}\sigma T_2^4 & LAYER2 \end{cases}$$

$$\begin{cases} \sigma T_E^4 + \sigma T_2^4 = \sigma T_s^4 & SUP \\ \sigma T_E^4 + \sigma T_2^4 = 2\sigma T_1^4 & LAYER1 \\ \sigma T_S^4 + \sigma T_1^4 = 2\sigma T_2^4 & LAYER2 \end{cases}$$

$$\begin{cases} T_E^4 + T_2^4 = T_s^4 & SUP \\ T_E^4 + T_2^4 = 2T_1^4 & LAYER1 \\ T_S^4 + T_1^4 = 2T_2^4 & LAYER2 \end{cases}$$

Two-layers atmosphere above the surface (3)

$$\begin{cases} T_E^4 + T_2^4 = T_S^4 & SUP \\ T_E^4 + T_2^4 = 2T_1^4 & LAYER1 \\ T_S^4 + T_1^4 = 2T_2^4 & LAYER2 \end{cases}$$

$$T_E^4 + T_2^4 + T_1^4 = 2T_2^4 \qquad LAYER\ 2$$

$$T_2^4 = T_1^4 + 2T_E^4 \qquad LAYER1$$

$$T_E^4 + T_1^4 + 2T_E^4 = 2T_1^4 \qquad LAYER\ 2 \quad \Rightarrow \quad \boxed{T_1^4 = 3T_E^4}$$

Two-layers atmosphere above the surface (4)

Considering:

$\varepsilon_{irr} = a$

$\varepsilon_{vis} = \mathrm{b}$

$F_{in} = \sigma T_E^4$

We can write the system as:

$$\begin{cases} a(1-a)T_1^4 + aT_2^4 - T_S^4 = -(1-b)T_E^4 \\ a^2T_1^4 - aT_2^4 + aT_S^4 = -b(1-b)T_E^4 \\ \mathrm{a}T_1^4 - a^2T_2^4 - a(1-a)T_S^4 = bT_E^4 \end{cases}$$

$$T_1^4 = \frac{(ab-a)T_E^4 - bT_E^4}{a^2 - 2a}$$

$$T_2^4 = \frac{[(a-1)b^2 + (2a^3+1)b + a^2 + a]T_E^4 + abT_E^4}{-a^2 + 2a}$$

$$T_E^4 = \frac{[(b^2+3a)b + a + 2]T_E^4 + bT_E^4}{2-a}$$

Two-layers atmosphere above the surface (5)

Considering:

$a = 1\ and\ b = 0$ $\quad$ $T_1^4 = T_E^4$ and $T_i^4 = iT_E^4$

Considering:

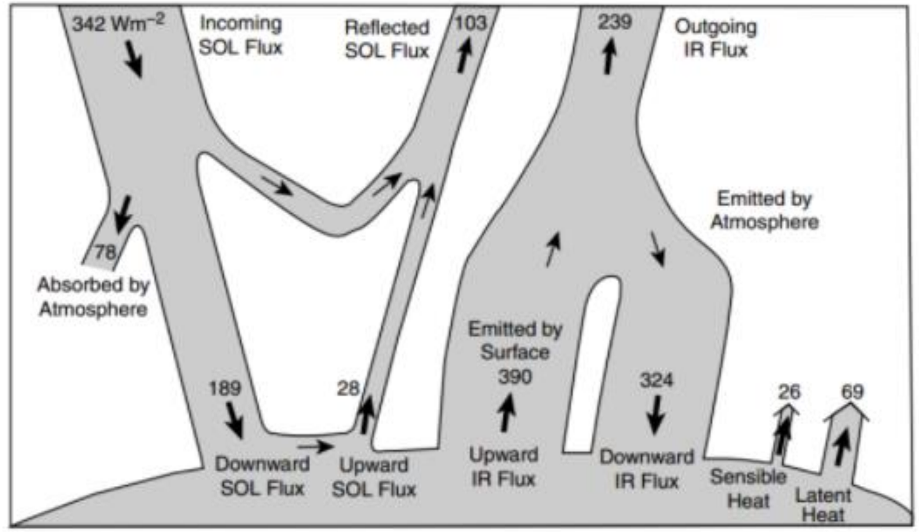
$a \neq 0\ and\ b = 0$

$$T_1^4 = \frac{1}{2-a} T_E^4$$
$$T_2^4 = \frac{1+a}{2-a} T_E^4$$
$$T_S^4 = \frac{2+a}{2-a} T_E^4$$

In this last scenario, the temperature of the surface increases with the number of layers.

This is not realistic since the layers decrease their absorbivity with the altitude.

Earth-atmosphere energy balance (1)



Liou **Figure 8.10** The heat balance of the earth and the atmosphere system. The solar (SOL) constant used is 1366 W m^{-2} so that the incoming solar flux for climatological energy balance is 342 W m^{-2} (round off the decimal point), while the global albedo is taken to be 30%. The "atmosphere" referred to in the graph contains molecules, aerosols, and clouds. The atmospheric thermal infrared (IR) flux is emitted both upward and downward. The upward IR flux from the surface is computed by using a climatological surface temperature of 288 K. At the top of the atmosphere, the energy is balanced by radiative flux exchange. At the surface, however, upward sensible and latent heat fluxes must be introduced to maintain energy balance. Absorption of the solar flux is obtained from the divergence of net solar fluxes at the top and the surface. The width of the shaded area with an arrow is approximately proportional to the flux value.

TOA: 342 (incoming solar) - 103 (reflected solar) - 239 (outgoing IR) = 0

ATMOSPHERE: 78 (solar absorbed) + 390 (incoming surface IR) - 239 (outgoing IR TOA) - 324 (outgoing surface) + 26 (sensible heat) + 69 (latent heat) = 0

SURFACE: 189 (absorbed solar) – 28 (reflected solar) - 390(emitted IR) + 324 (absorbed IR) - 26 (sensible heat) - 69 (latent heat) = 0

Earth-atmosphere energy balance (2)

The interaction of the atmosphere with electromagnetic radiation occurs through absorption, diffusion or emission in the solar and thermal spectrum.

The heating of the atmosphere by the surface is also due to the exchange of sensible and latent heat.

Sensible heat is transferred from the surface to the atmospheric layers in contact with a heat transfer process at the molecular level, but just above the surface the flows occur by vertical transport and mixing of air masses. The flow due to the average vertical transport is inefficient, as the average vertical wind is very weak.

The sensitive heat flow is mainly due to turbulent flows, that is to the vertical transport caused by the turbulent fluctuations of the vertical wind (w') and enthalpy ($c_p T'$)

$$\boxed{Q_s = \rho c_p \overline{w'T'}}$$

Earth-atmosphere energy balance (3)

The **latent heat flow** is caused by the evaporation of the surface water which emits latent heat in the air masses in contact with the surface. Also in this case the flow due to the average vertical speed is completely negligible compared to that due to turbulence

$$\boxed{Q_e = \rho L_v \overline{w'r'}}$$

where r' is the turbulent perturbation of the mixing ratio