Radiation & Radiative Transfer II

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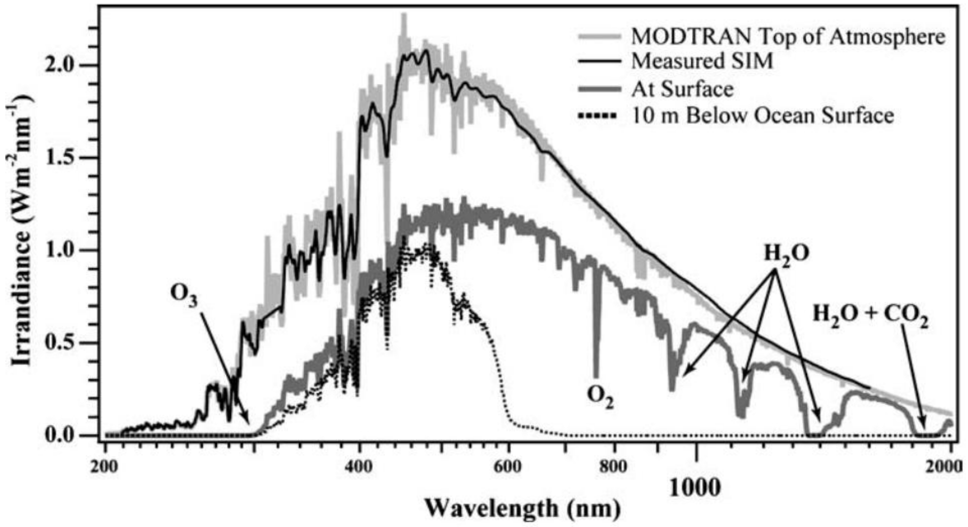
Recap from last lecture

Please go to: responseware.eu
Session ID is: **env320**



$\boldsymbol{Solar\ constant}$: $\boldsymbol{S_0 = 1370\ W\ m^{-2}}$

Planck Function: $B_\lambda(\mathrm{T}) = \frac{2hc^{*2}/\lambda^5}{e^{hc^*/\lambda kT}-1}$ (W m^{-2} sr^{-1} µm^{-1})

Monochromatic intensity of a blackbody

$\mathbf{Wien's\ displacement\ law}$: $\lambda_{max} = \frac{2897}{T}$ (µm)

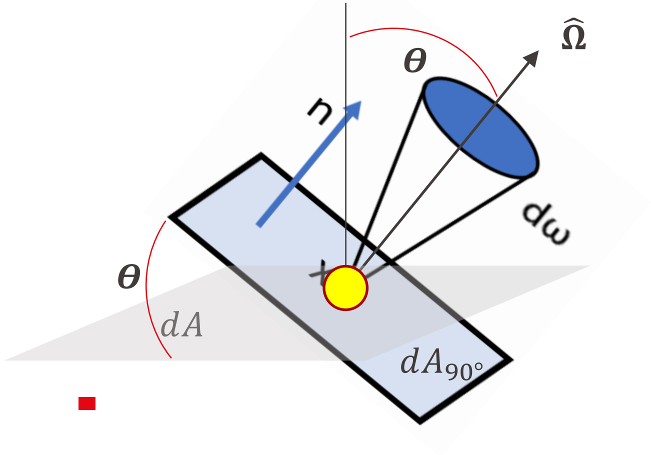
$\mathbf{Stefan-Boltzmann\ law}$: $F_B\ (\mathrm{T}) = \pi \mathrm{B}(\mathrm{T}) = \pi \int B_\lambda(T)d\lambda = \sigma T^4$

Total emitted flux of a blackbody

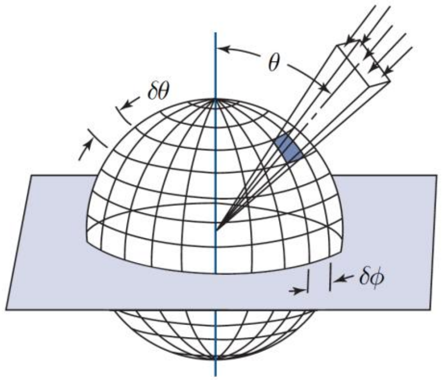
T in K, B(T) in W m^{-2} µm^{-1} sr^{-1}, F_B (T) in W m^{-2}

$$F = \int I(\boldsymbol{\Phi}, \boldsymbol{\theta}) \cos \boldsymbol{\theta}\, d\omega; \quad F \propto \frac{1}{d^2}$$

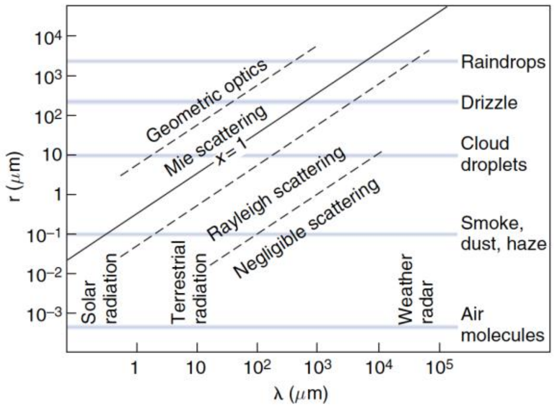
Intensity is directional



Irradiance is hemispheric



Extinction = absorption + scattering

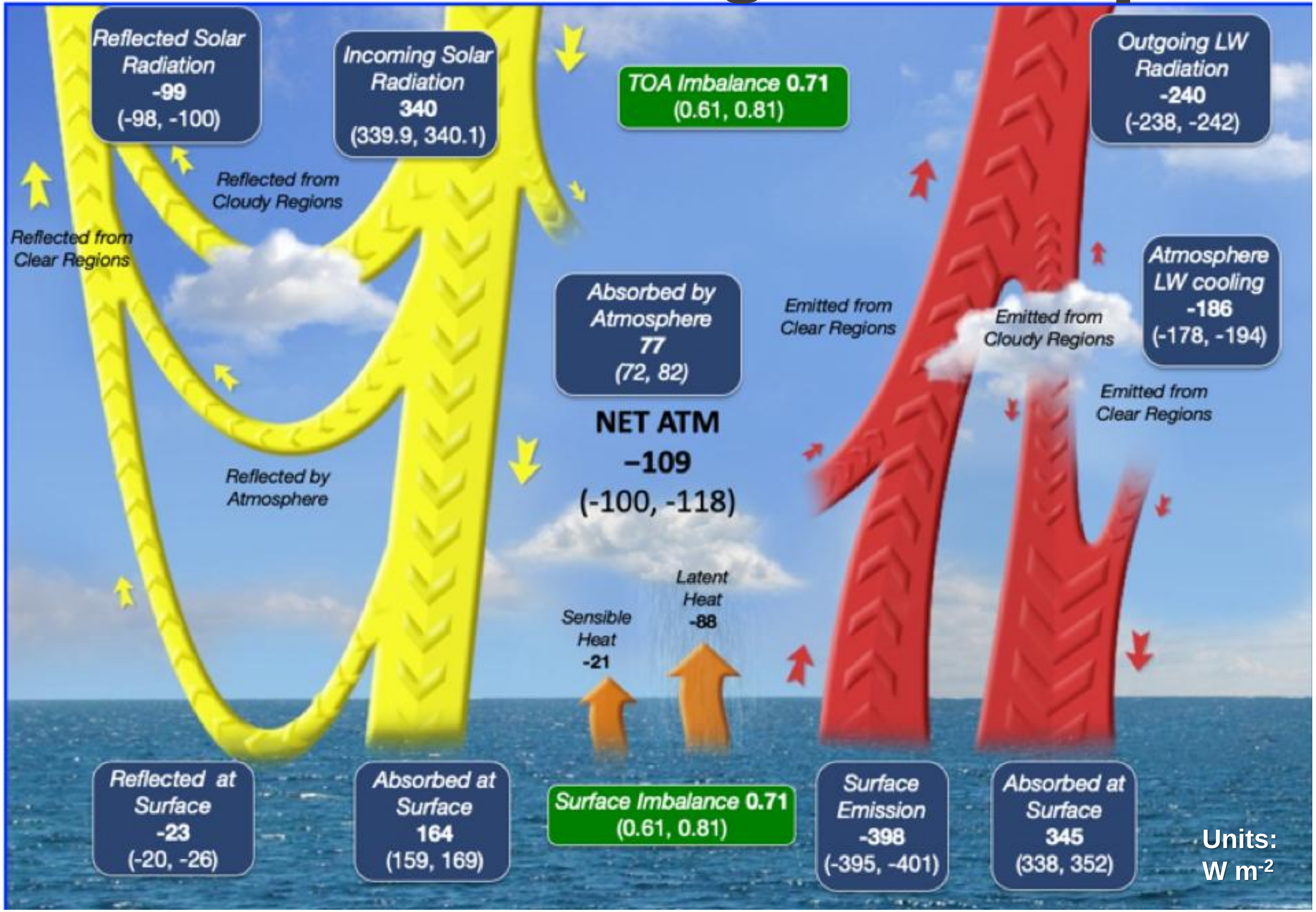


Take home messages from last lecture

- Shortwave radiation is solar radiation, thermal infrared (longwave) is Earth’s radiation.
- Know what is: radiation, intensity, radiance, flux density, irradiance, insolation.
- Remember Planck Function, Wien’s displacement law, Kirchhoff’s law, Stefan-Bolztmann’s law
- Planetary albedo
- Rayleigh scattering: applies to air molecules
- Mie scattering: applies to aerosols and cloud droplets / crystals

Radiative transfer through the atmosphere

Absorption and scattering (reflection)

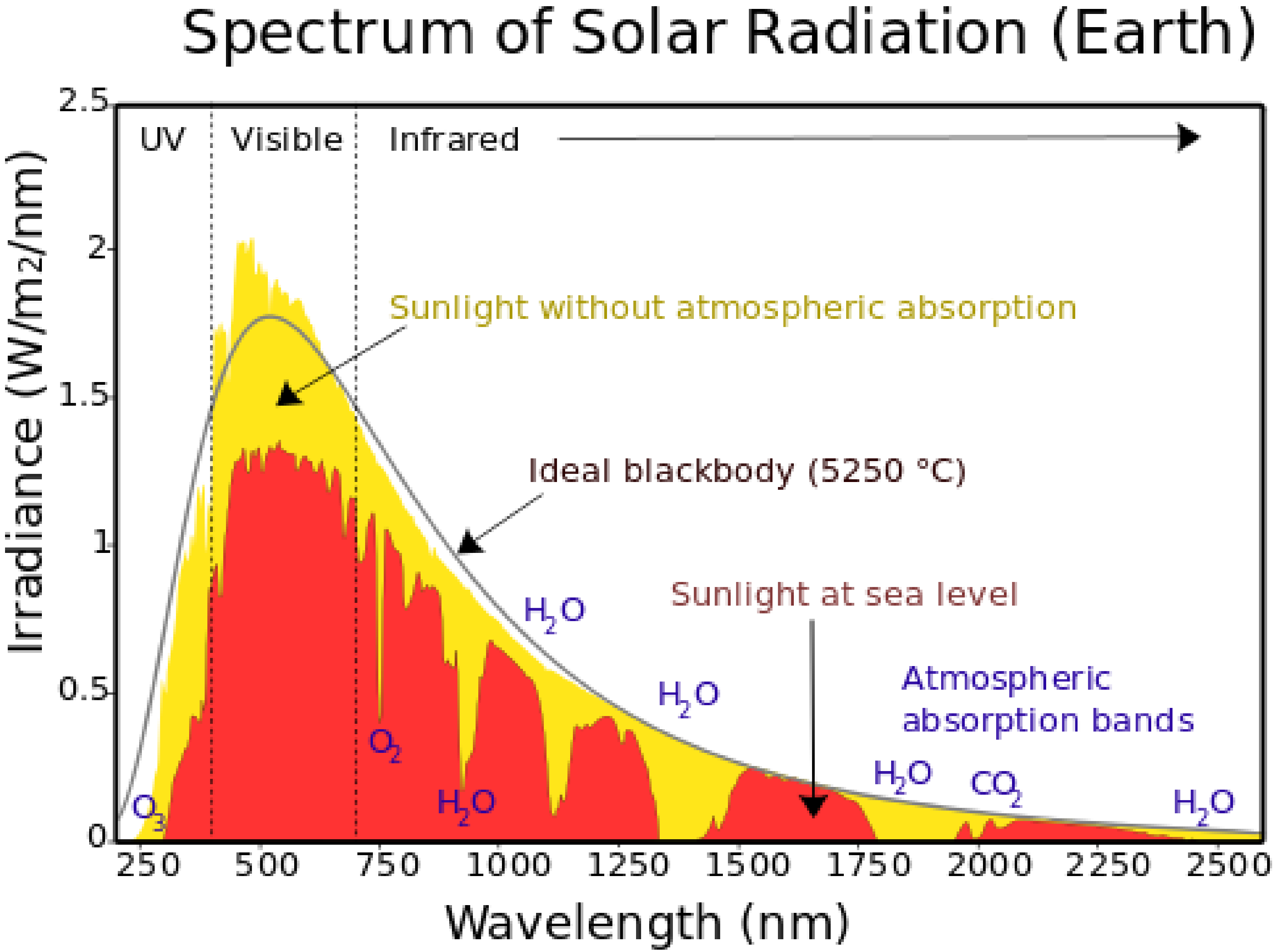


Absorption and re-emission

Surface has a net positive imbalance, atmosphere a net negative, that is globally averaged.
Need sensible and latent heat process to equilibrate.

https://ceres.larc.nasa.gov/science/

Absorption by atmospheric gases



Absorption continua

Lyman-α: Hydrogen electron falls from energy level n=2 (orbital) to n=1 and emits radiation in space (deeper radiation penetration into atmosphere)

Radiation interaction: absorption (scattering, emission) in quantities of photons. Photons contain energy:

$$E = h\tilde{\nu}$$

h Planck constant ($6.626 \cdot 10^{-34}$ Js)

Absorption continua in x-ray and UV range caused by:

- Photoionization (remove electrons from atoms, extreme ultraviolet $\lambda \leq 0.1\ \mu m$, happens in ionosphere)
- Photodissociation (break molecules, ultraviolet $\lambda \leq 0.31\ \mu m$, happens down to stratosphere, O_2 breakup important of O_3 production)

→ energy is converted into kinetic energy (temperature increase of gas).

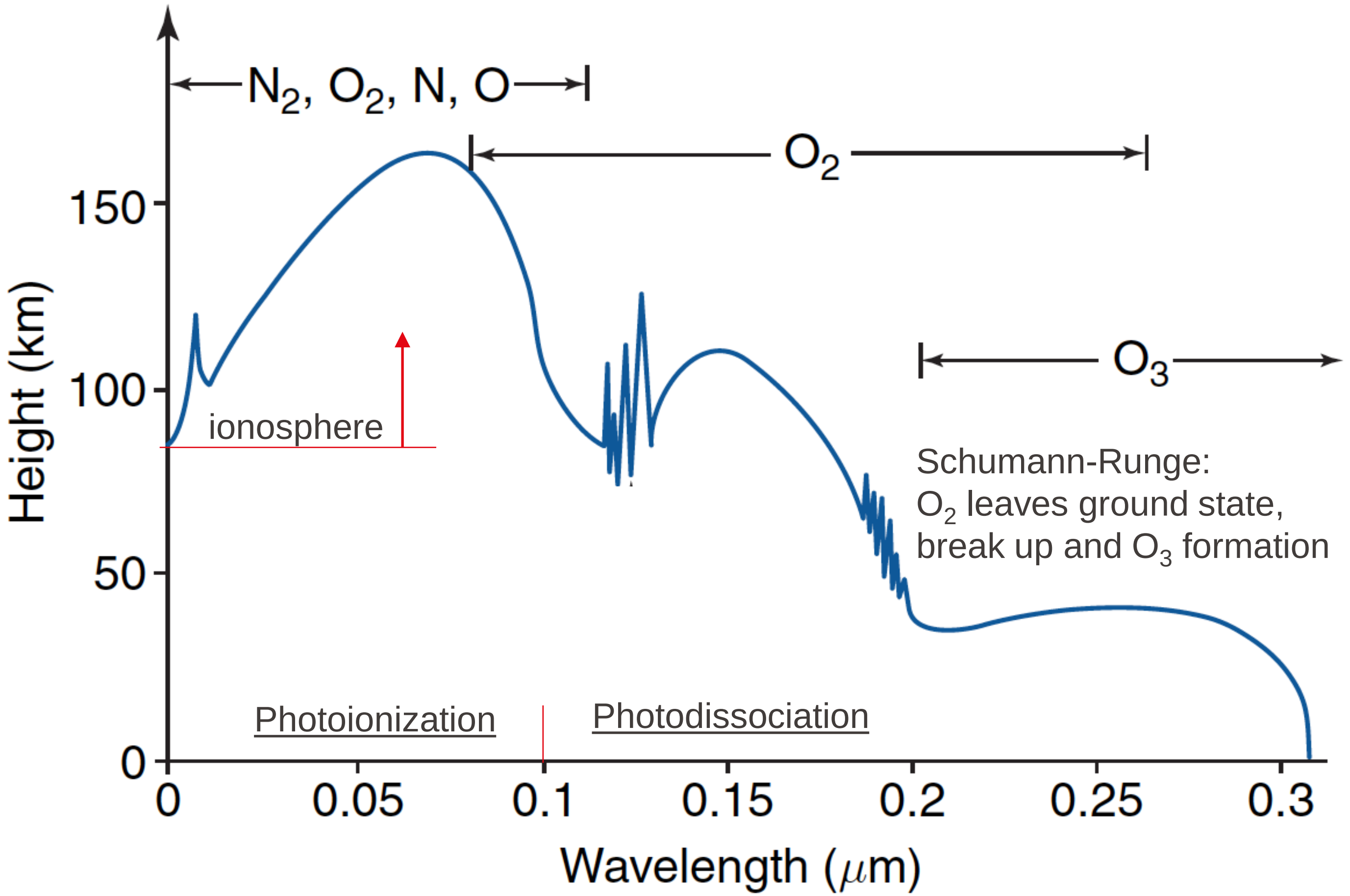
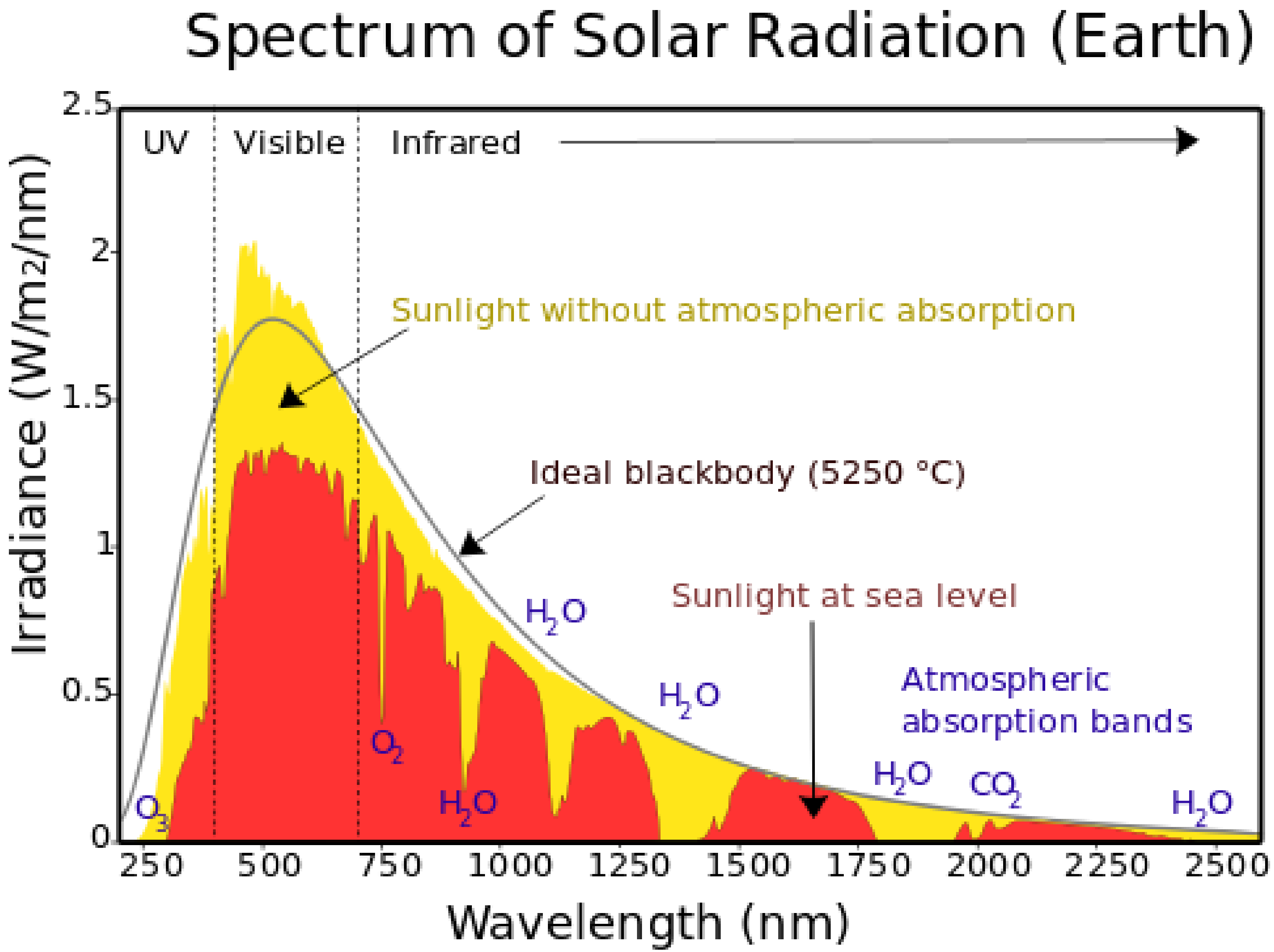


Figure 4.20: depth of penetration of solar UV radiation

Depletion of UV radiation in atmosphere!

Absorption by atmospheric gases

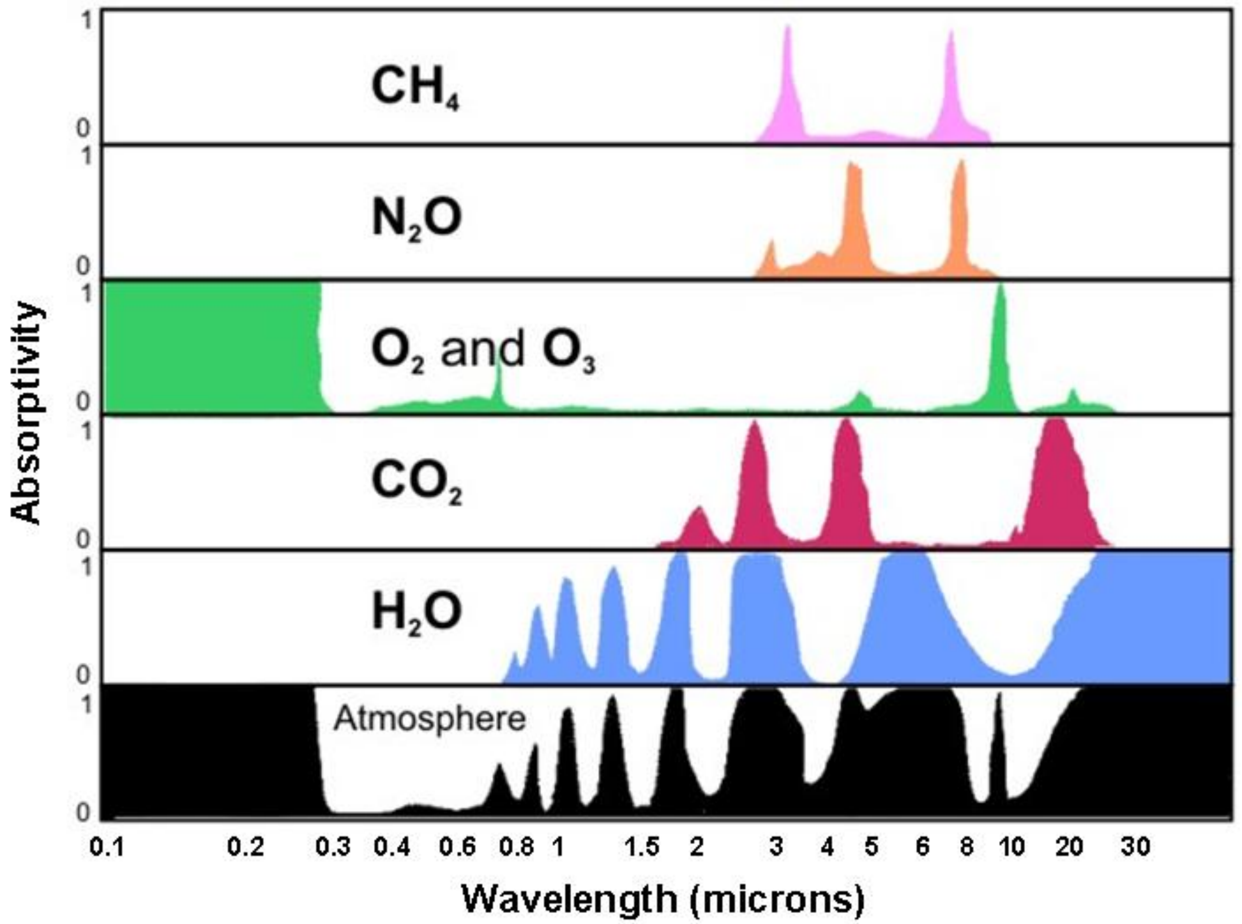


Absorption by gases

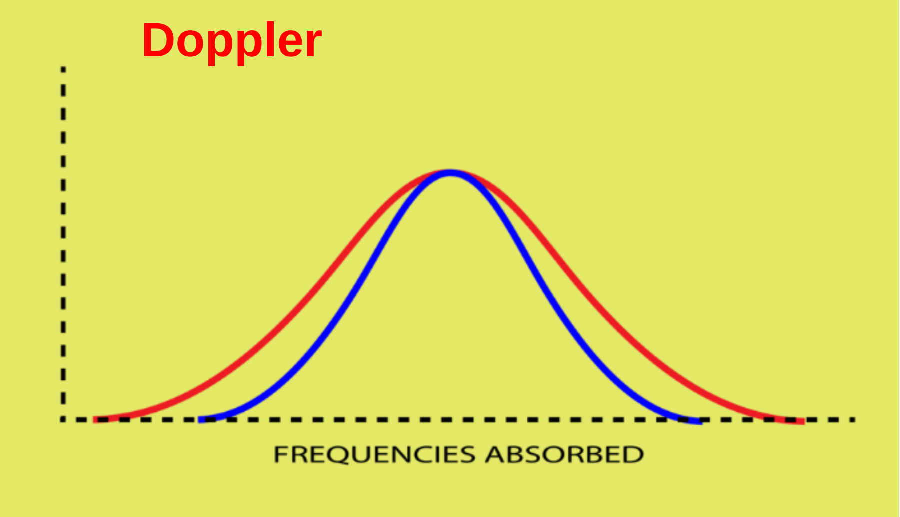
Absorption lines, mostly in visible and infrared spectrum, caused by internal energy state of gas molecule, only discrete energy transitions are possible.

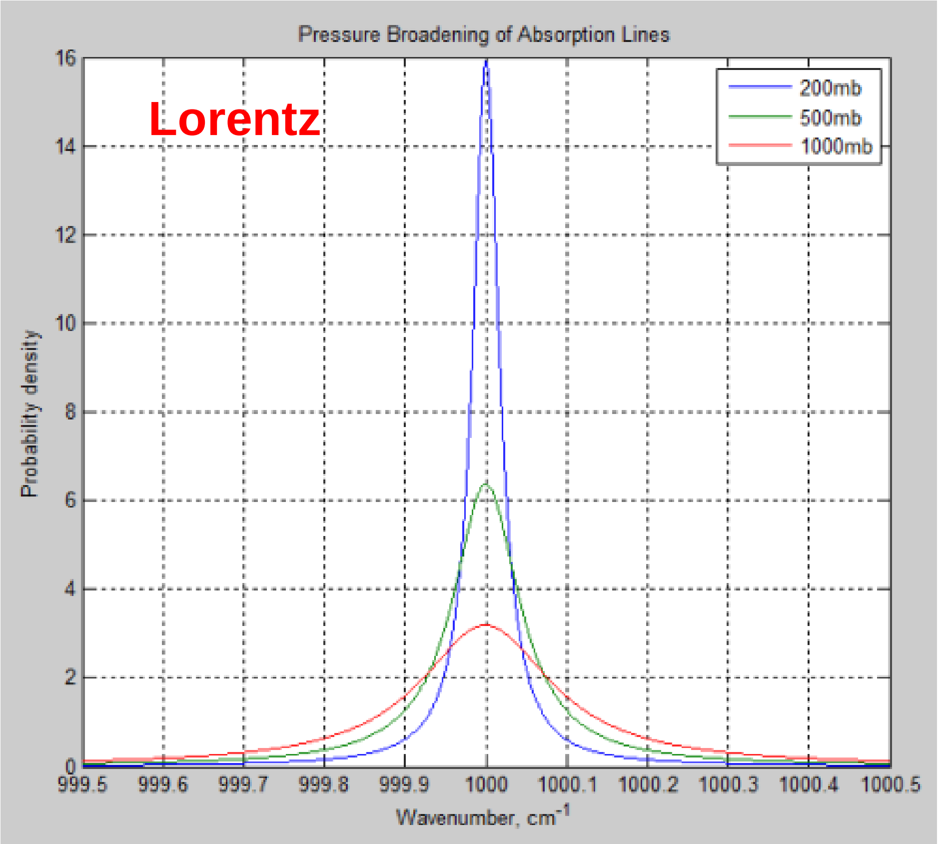
$$E = E_0 + E_v + E_r + E_t$$

E_0 Energy level of orbits
E_v vibrational energy level
E_r rotational energy level
E_t translational energy (random motion)
.



Line Broadening





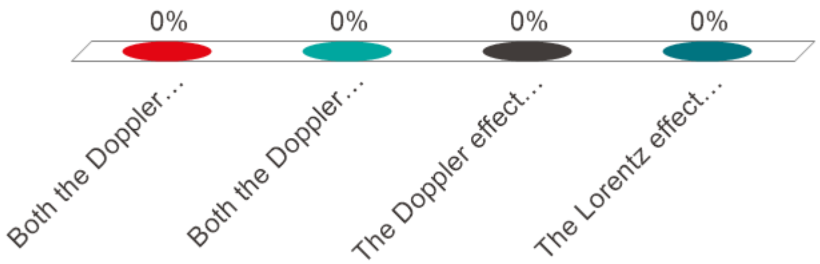
As a beam of light passes through an atomic vapor, the atoms will absorb a certain range of the beam's frequencies. The range depends on the velocity distribution – that is, on the temperature. **Atoms at lower temperatures (blue curve)** absorb fewer frequencies than at **higher temperatures (red curve).** This phenomenon, called **Doppler broadening (temperature effect)**, can be measured spectroscopically and thus can serve as a thermometer.

In addition there is the Lorentz effect (pressure effect): molecules collide with each other and change energy levels. Depends on density and temperature.

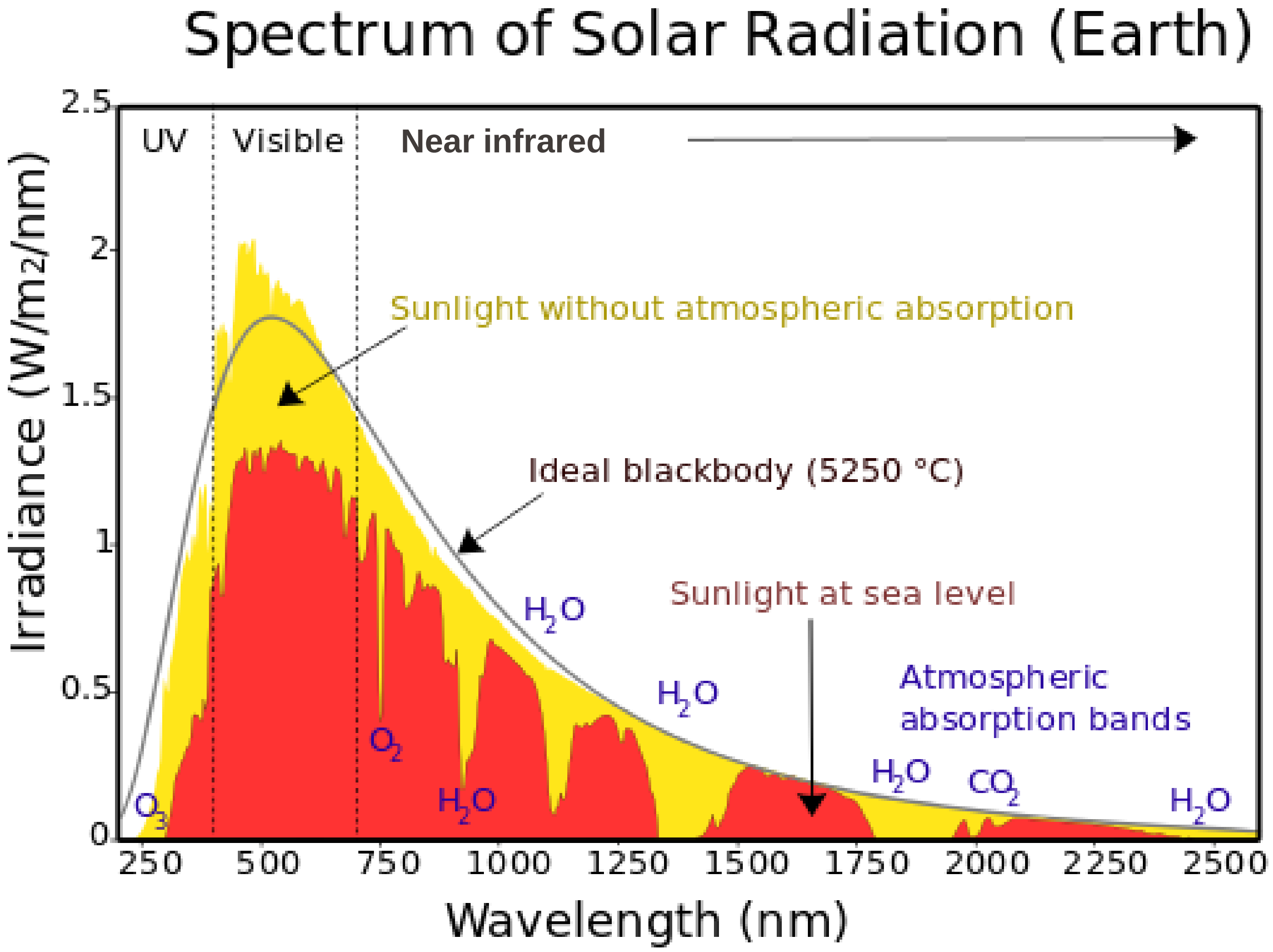
https://www.nist.gov/image/dopplerbroadeningpng

Relevance of Doppler and Lorentz effects

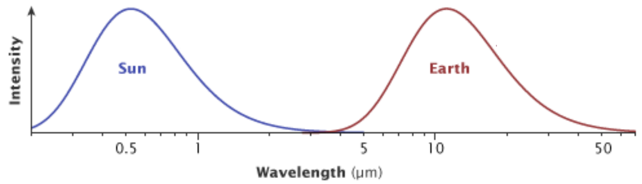
A. Both the Doppler and Lorentz effects are larger in the lower atmosphere.
B. Both the Doppler and Lorentz effects are larger in the higher atmosphere.
C. The Doppler effect is more important in the upper atmosphere.
D. The Lorentz effect is more important in the upper atmosphere.



Absorption by gases



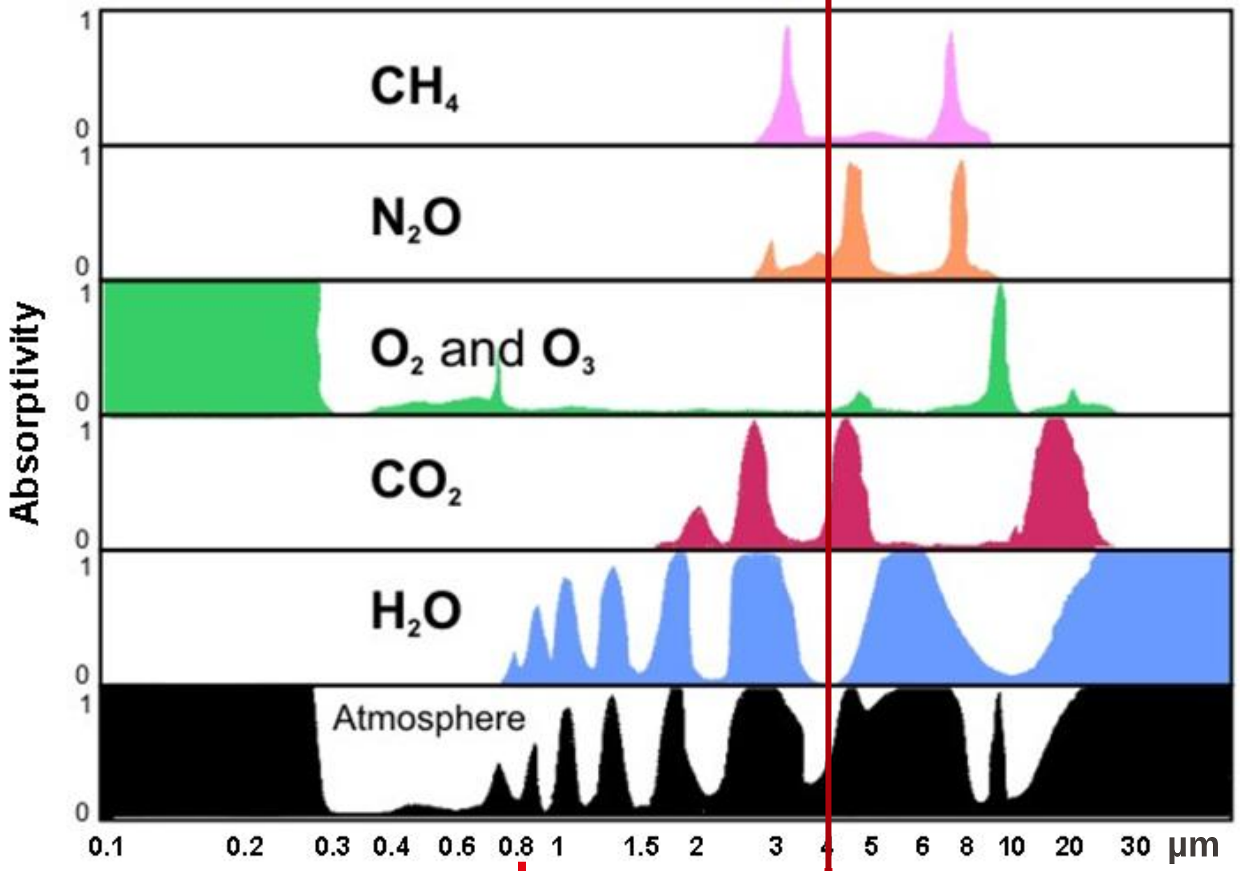
And here?



- https://www.e-education.psu.edu/meteo300/node/686

Absorption by gases

Strong absorption by greenhouse gases!



UV-visible | near infrared | thermal infrared

terrestrial radiation

Greenhouse gases

Fundamental consideration:

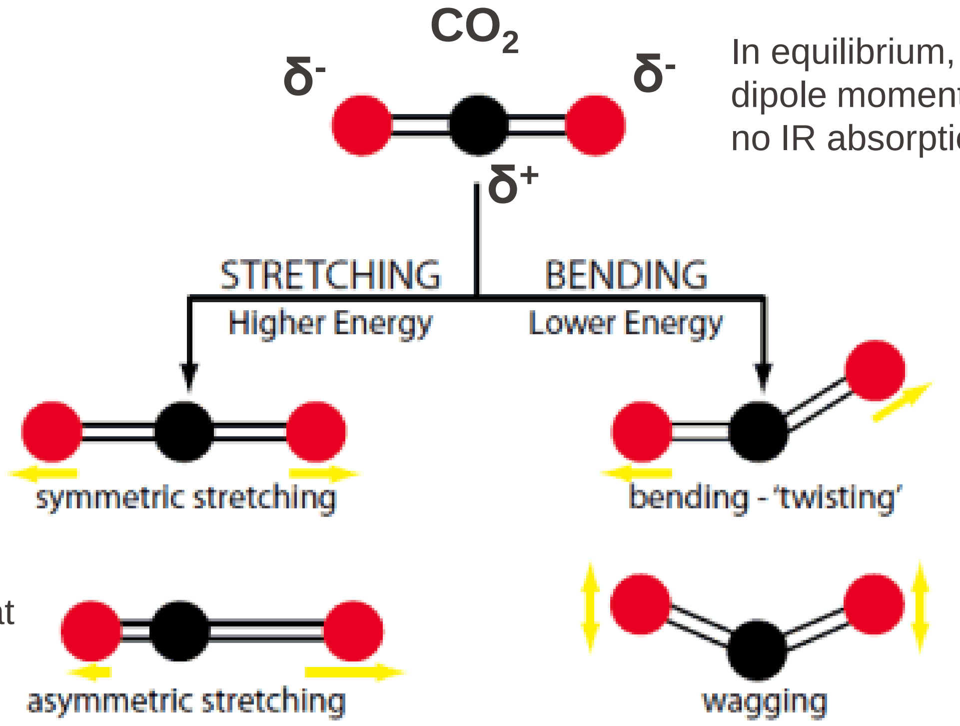
1. **Temperature** is a measure of the average energy of **molecular motion** in matter: to and fro translation, intramolecular vibration, and rotation. The sum of these motions’ energies can be described as the “thermal energy”. Thermal energy and, hence, **temperature can change as** various forms of energy, including **electromagnetic energy**, interact with the sample and **change the average energy of motion**.
2. **Molecular vibrations** (and some energetic rotations) have energy level spacings that **correspond to energies in the IR region** of the electromagnetic spectrum. Thus IR radiation absorbed by molecules causes **increased vibration**.
3. **Collisions** between these energized molecules in the atmosphere transfer energy among all the molecules, which increases the average thermal energy and, hence, **raises the temperature**.

- https://www.acs.org/content/acs/en/climatescience/greenhousegases/properties.html

Greenhouse gases

4. For molecular vibrations **to absorb IR** energy, the **vibrational motions must change the dipole moment** of the molecule. **All molecules with three or more atoms meet this criterion and are IR absorbers.**
5. While the Earth's (dry) atmosphere is predominantly composed of non-IR absorbers, N_2 (78%), O_2 (21%), and Ar (~0.9%), **the 0.1% of remaining trace gases contains many species that absorb IR**.
6. Earth's atmosphere is not dry and can contain several % of water vapor. **Water vapor is the main greenhouse gas.**
7. The presence of "natural" (i.e. not influenced by human action) water vapor, CO_2, CH_4, N_2O and other greenhouse gases are **responsible for the natural greenhouse effect.**

In equilibrium, no dipole moment, no IR absorption.



molecular vibrations of CO_2

In equilibrium, no dipole moment, no IR absorption.

Absorption at 4 µm

Absorption at 15 µm

Dipoles means that a molecule has negative and positive charges, and the averaged centers of the charges do not co-incide. This leads to a dipole moment, capable of IR absorption. O_2 and N_2 are not dipoles.

http://www.chm.bris.ac.uk/motm/CO2/?C=S;O=A

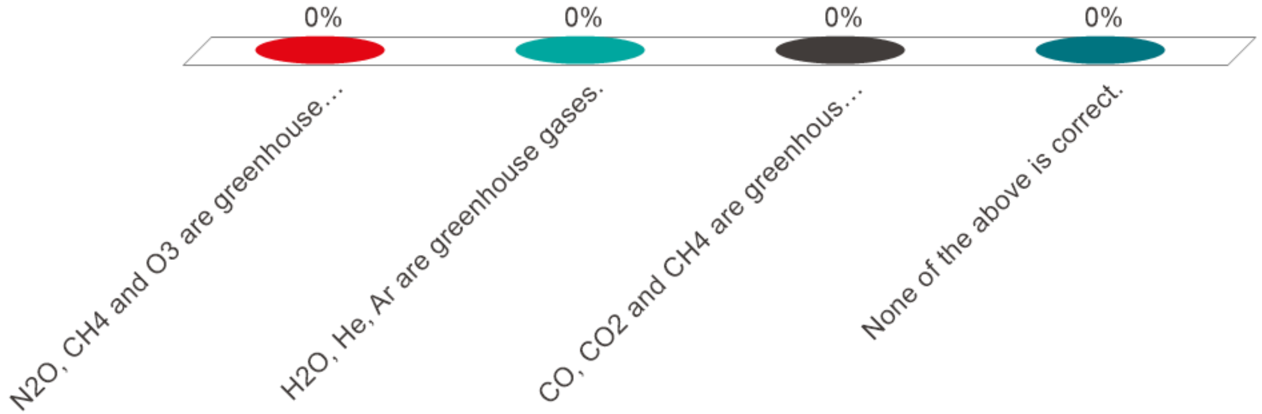
Which statement is correct?

A. N_2O, CH_4 and O_3 are greenhouse gases.

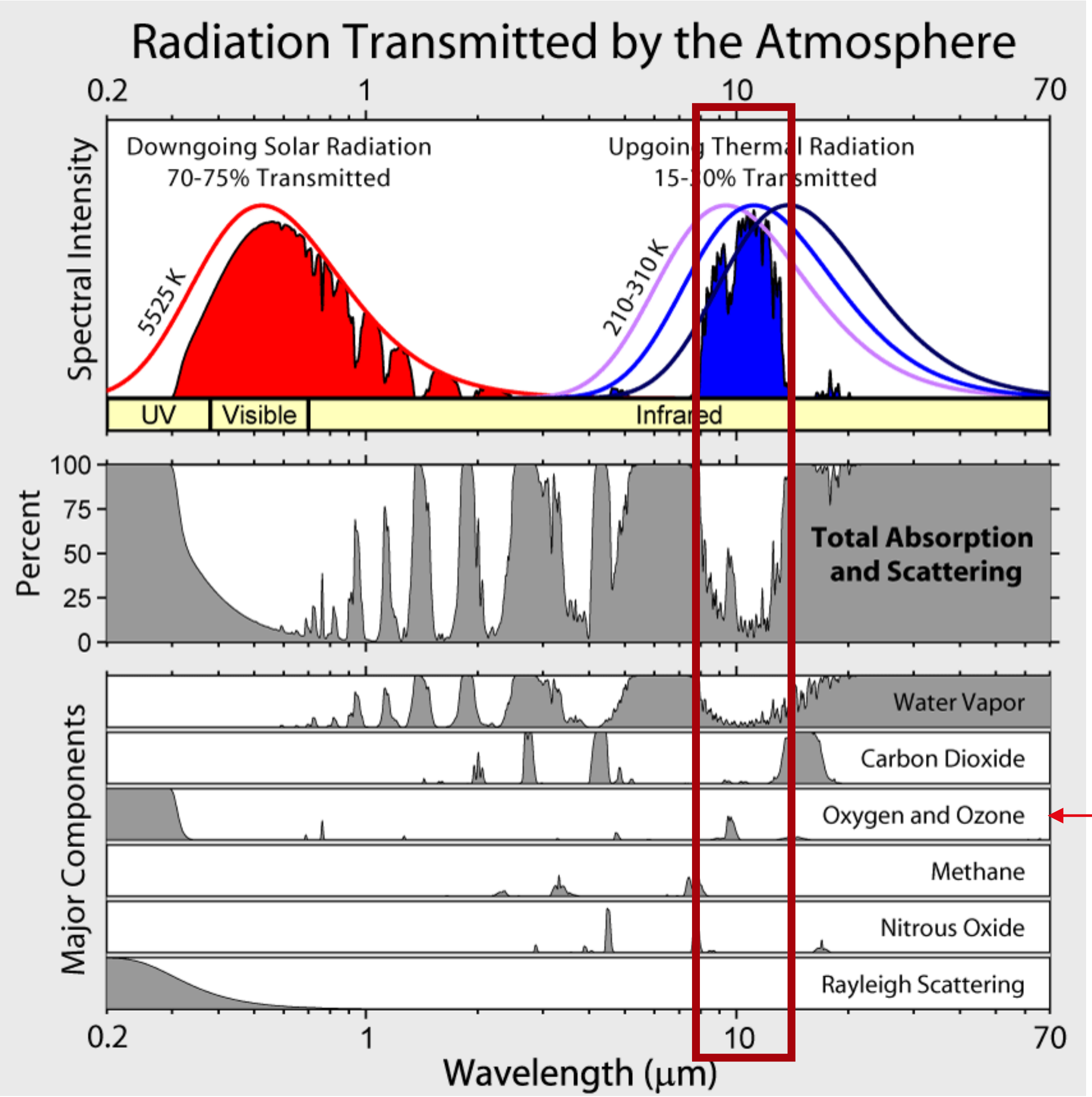
B. H_2O, He, Ar are greenhouse gases.

C. CO, CO_2 and CH_4 are greenhouse gases.

D. None of the above is correct.



Atmospheric «windows»



The **IR window** is between 8-9 and 10 - 12 µm, roughly.

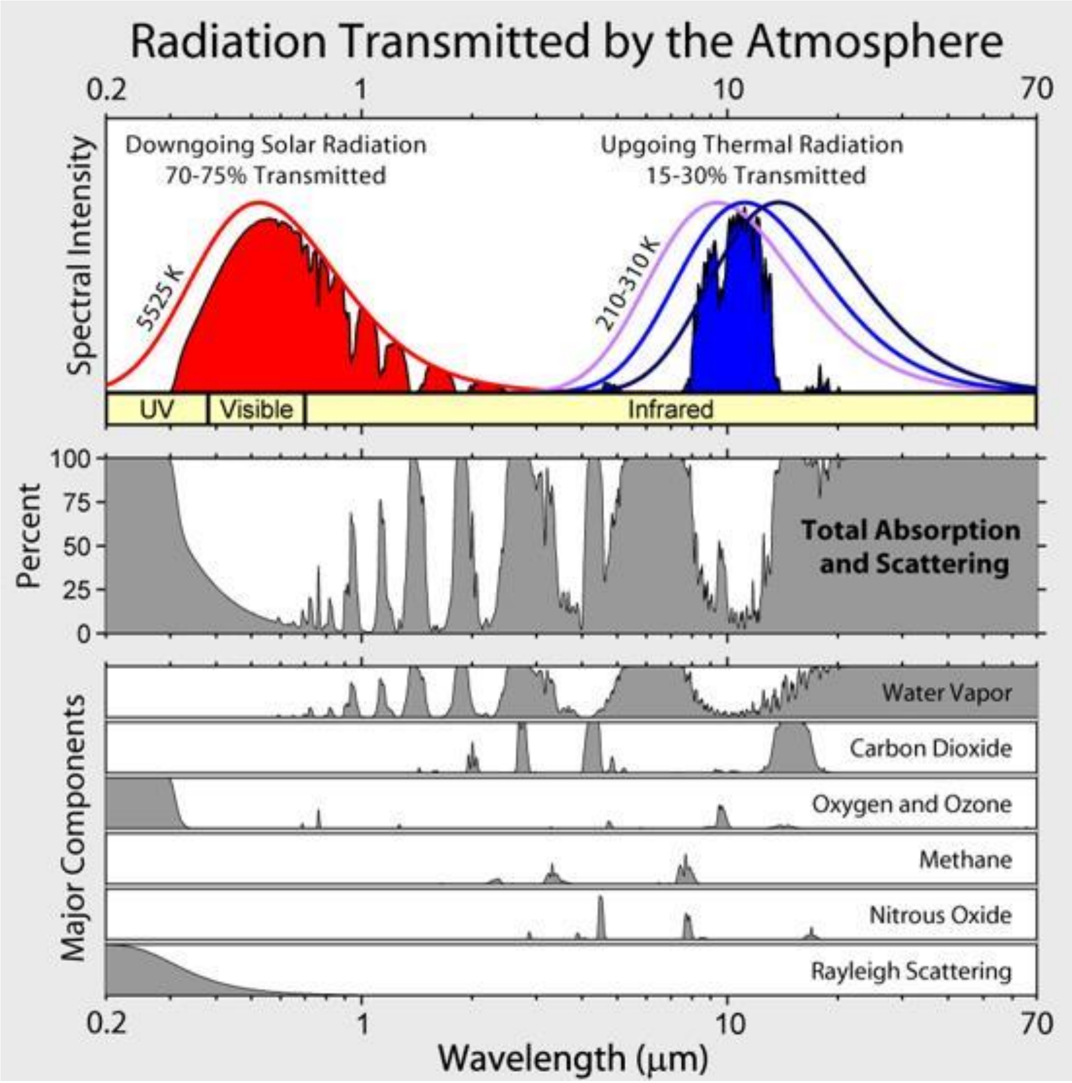
The result of the presence of greenhouse gases is that only a small part of the longwave surface emission makes it to space.

Different greenhouse gases work together to close "atmospheric windows" in the longwave part of the spectrum.

Ozone absorbs at 9.6 µm.

Source: Wikipedia

Can you find other atmospheric windows?

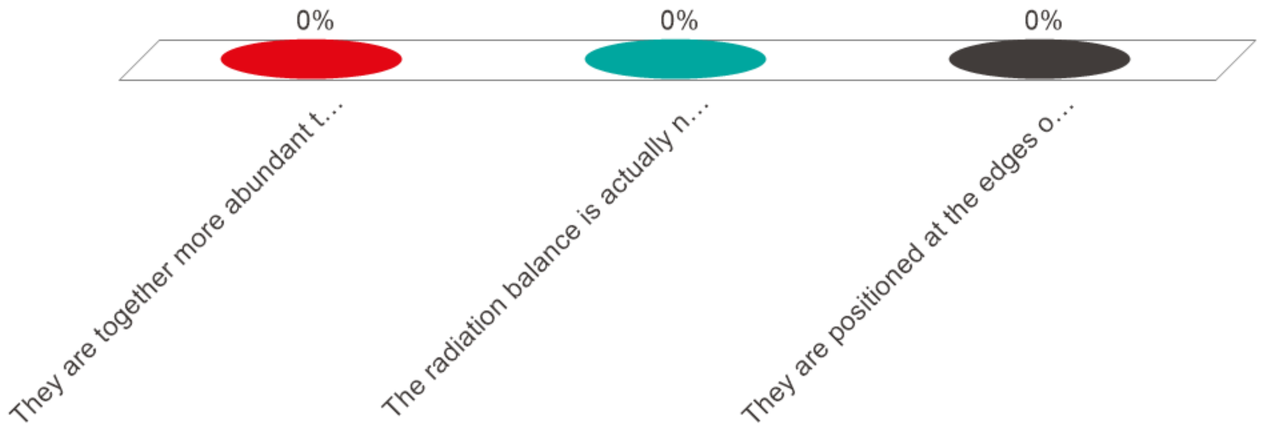


Why is Earth's radiation balance so sensitive to CO_2, N_2O, and CH_4?

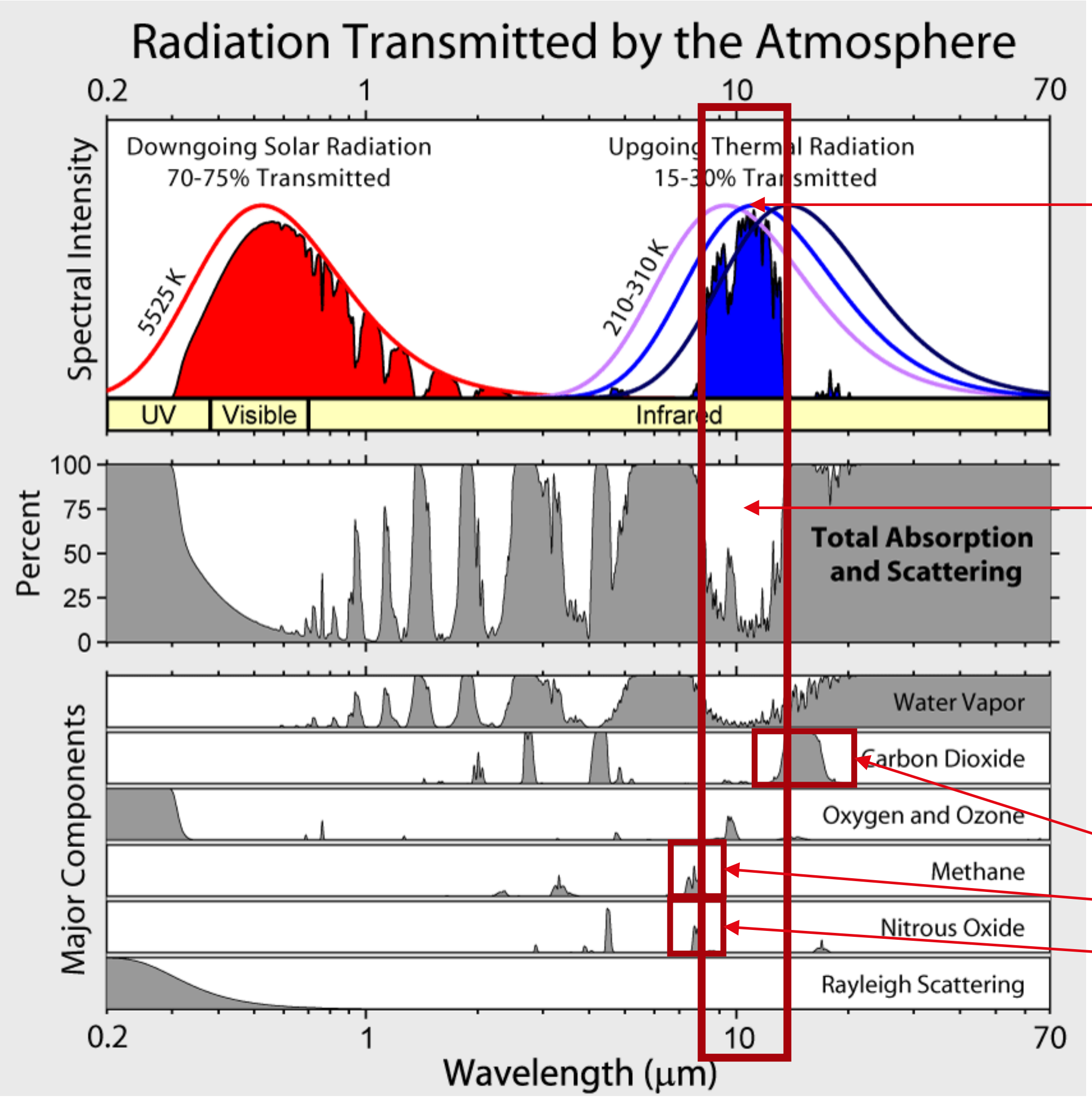
A. They are together more abundant than H2O.

B. The radiation balance is actually not sensitive to them.

C. They are positioned at the edges of the atmospheric window.



Atmospheric «windows»



maximum of terrestrial radiation

8-12 µm

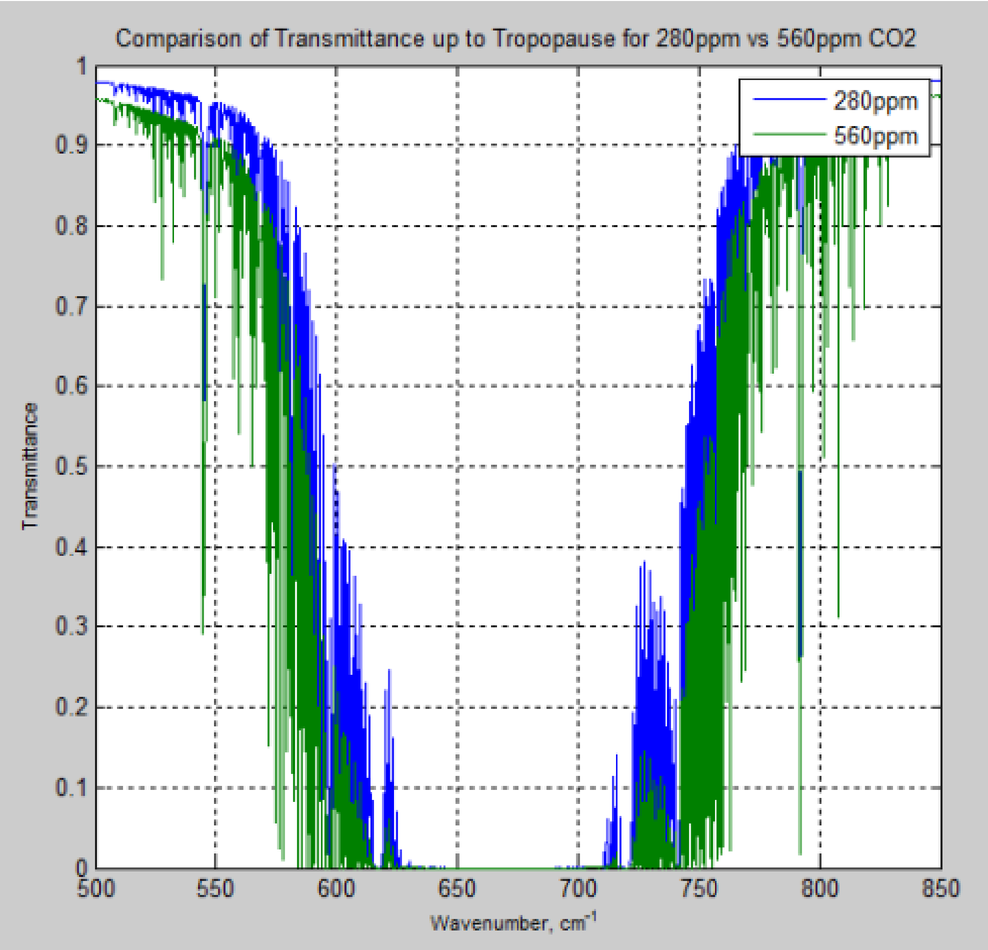
almost transparent
only O_3 absorbs (9.6 µm)

CO_2, CH_4 and N_2O have absorption bands on the edges of the atmospheric window, changing their concentrations can have a large impact

Source: Wikipedia

EPFL

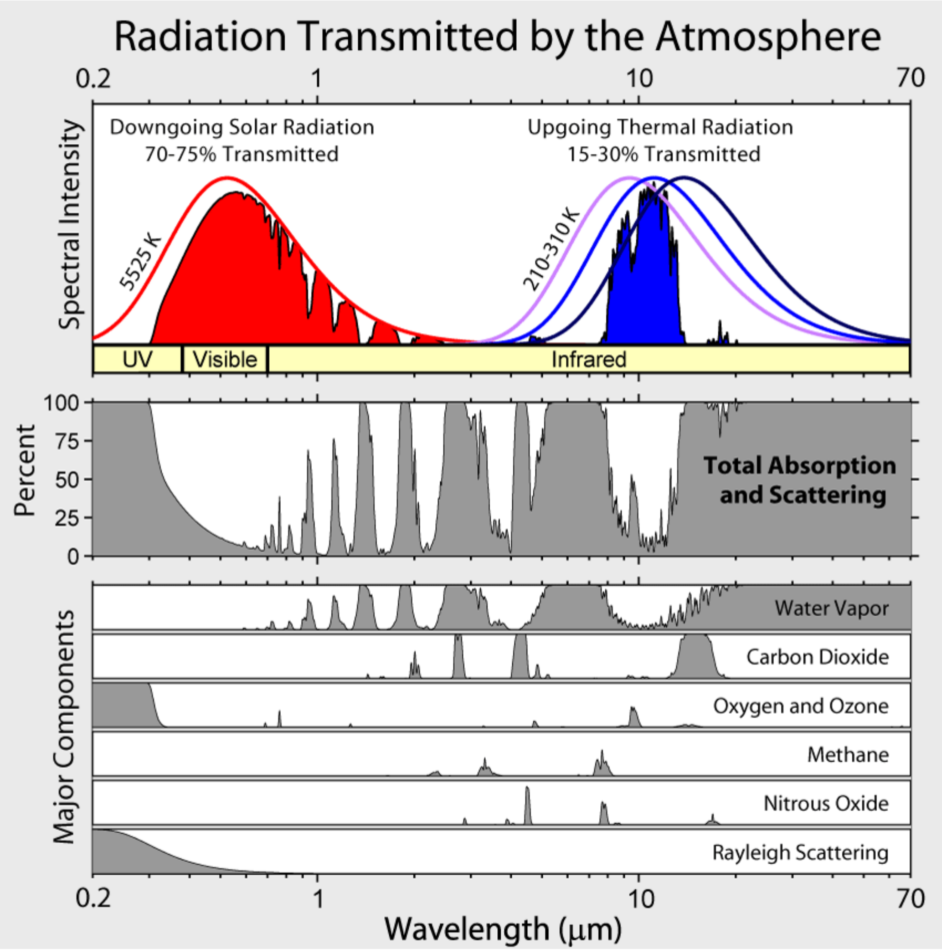
What happens if we «pump» more CO_2 into the atmosphere?



The transmission of radiation through the atmosphere is reduced with higher mixing ratios of CO_2.

Why is transmittance not the full story for the Greenhouse effect?

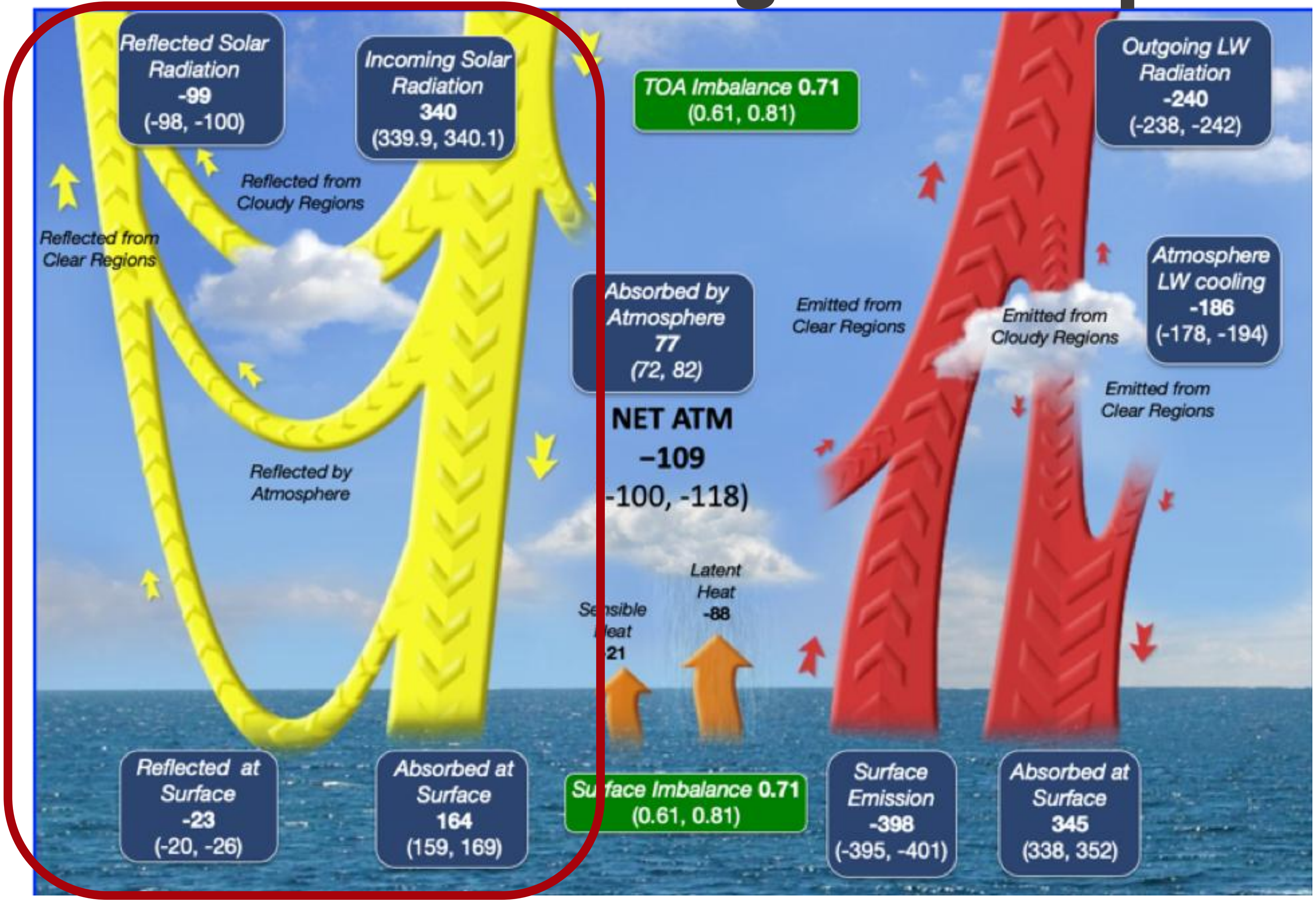
Greenhouse effect



1. Solar radiation is transmitted through Earth's nearly transparent atmosphere (transparent to shortwave radiation).
2. Earth's surface absorbs the solar radiation and warms.
3. The Earth emits thermal IR radiation (blackbody radiation).
4. The atmosphere is much less transparent to thermal infrared radiation and absorbs it.
5. The absorbed radiation leads to warming of the atmosphere, which in turn **emits** thermal IR radiation in all directions, also partly downwards.
6. So the net thermal IR flux from the Earth (as blackbody) to space is greatly recuded. This diminishes the radiative cooling of the Earth's surface and leads to surface warming. (This explains why the outgoing terrestrial radiation flux is much larger than the incoming solar radiation flux at the Earth's surface.)

Radiative transfer through the atmosphere

Absorption and scattering



https://ceres.larc.nasa.gov/science/

Quantifying atmospheric extinction

Beer-Lambert-Bouguer Law for extinction

$$dI_\lambda = -I_\lambda \rho r k_\lambda ds$$

ρ = density of air
r = mass of gas per unit mass of air
k_λ = mass extinction coefficient (m^2 kg^{-1})

Integrated through all atmosphere (top = ∞):

$$I_\lambda(z) = I_\infty \, e^{\frac{-\tau\lambda}{\cos\theta}}$$

Optical depth: $\tau_\lambda = \int_z^\infty \rho r k_\lambda dz$

Transmissitivity: $T = e^{\frac{-\tau\lambda}{\cos\theta}}$

Absorptivity: $\alpha = 1 - T = 1 - e^{\frac{-\tau\lambda}{\cos\theta}}$
in absence of scattering

Extinction is proportional to:
- to the radiation intensity
- path length *ds* (mass and gas properties)

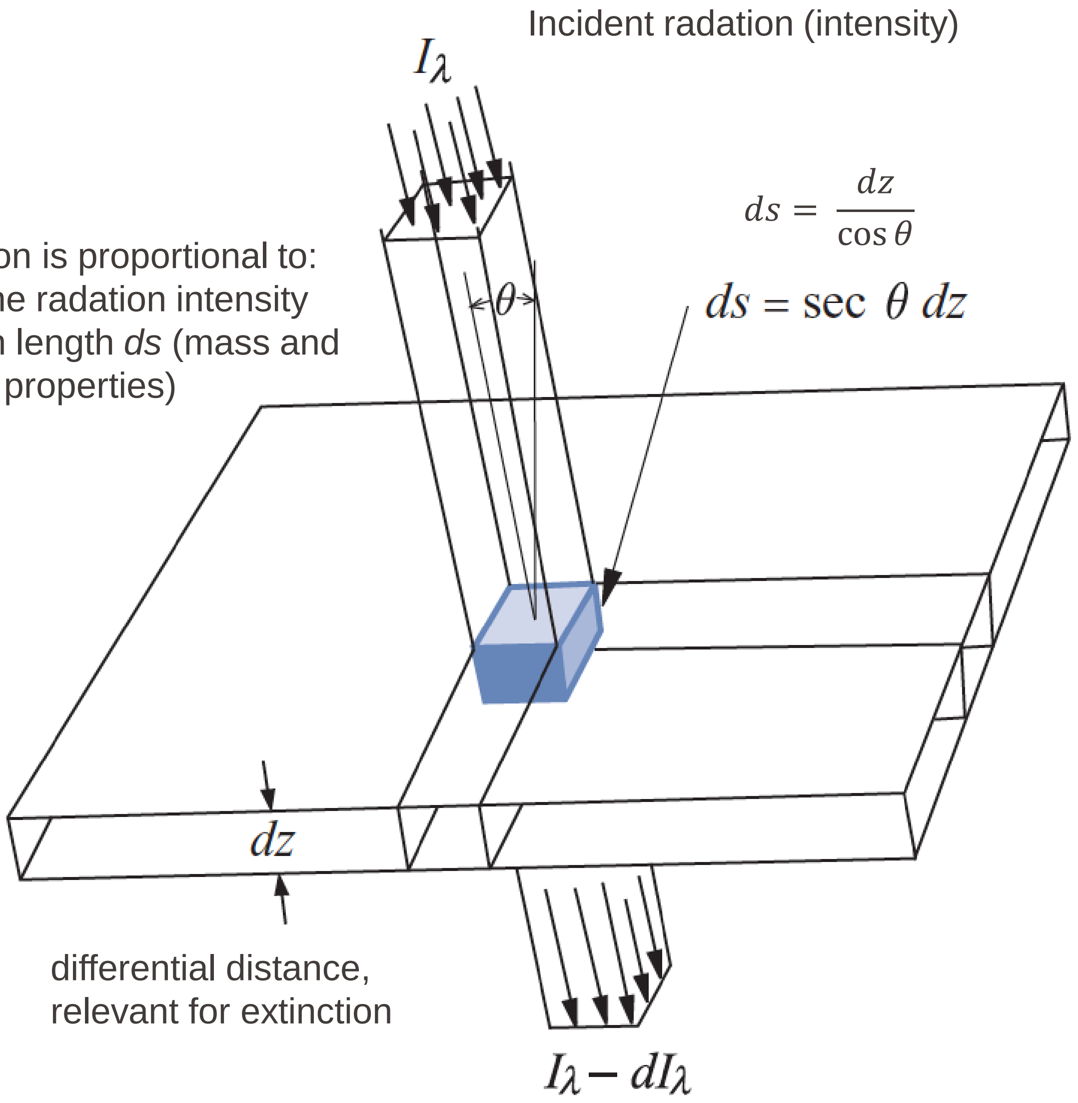


Figure 4.10

Optical depth of the atmosphere

Optical depth (dimensionless): measure of depleted incident (downward) radiation through a layer of atmosphere

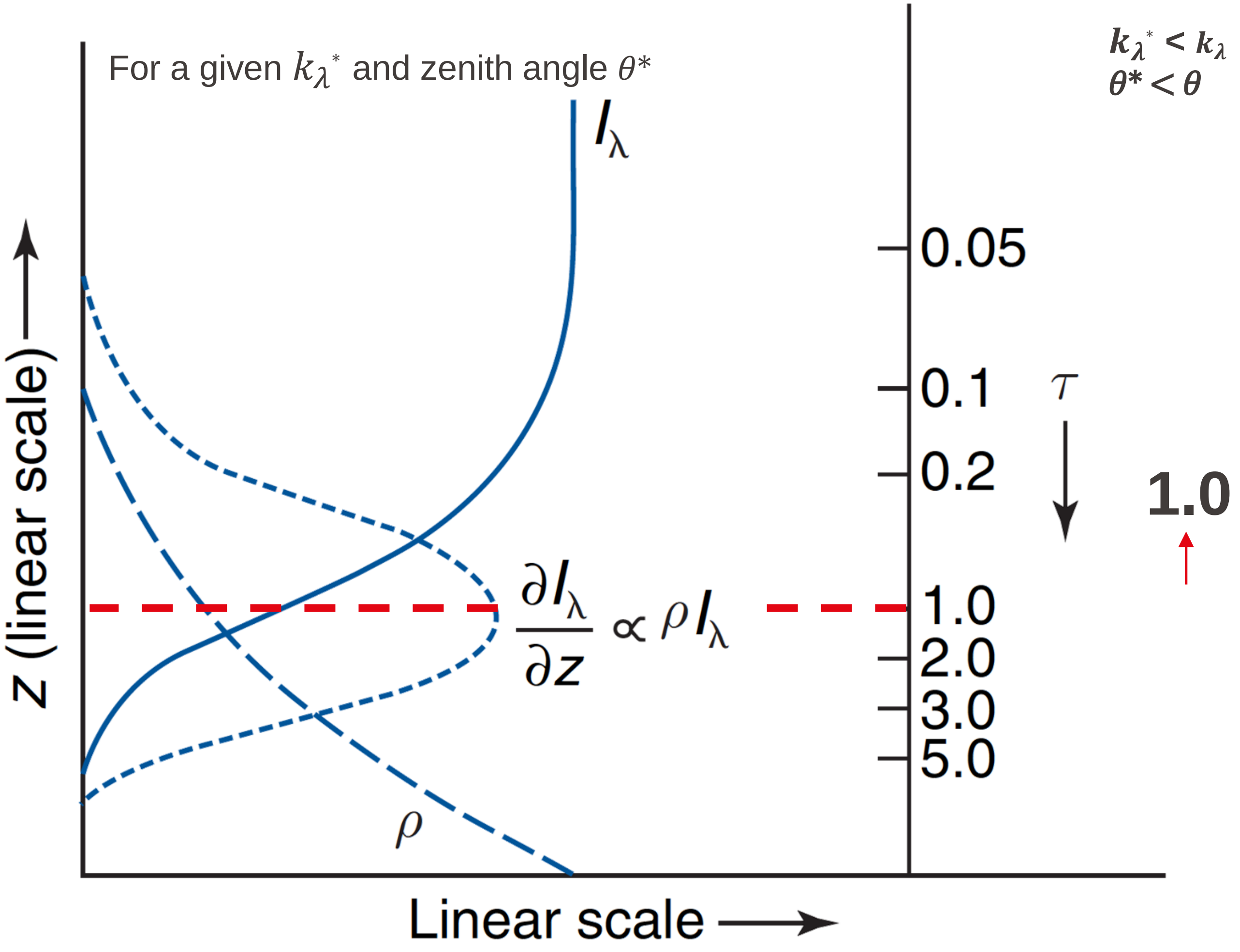
$$\tau_\lambda = \int_z^\infty \rho r k_\lambda dz$$

ρ = density of air
r = mass of absorbing gas per unit mass of air
k_λ = mass absorption coefficient ($m^2\ kg^{-1}$)

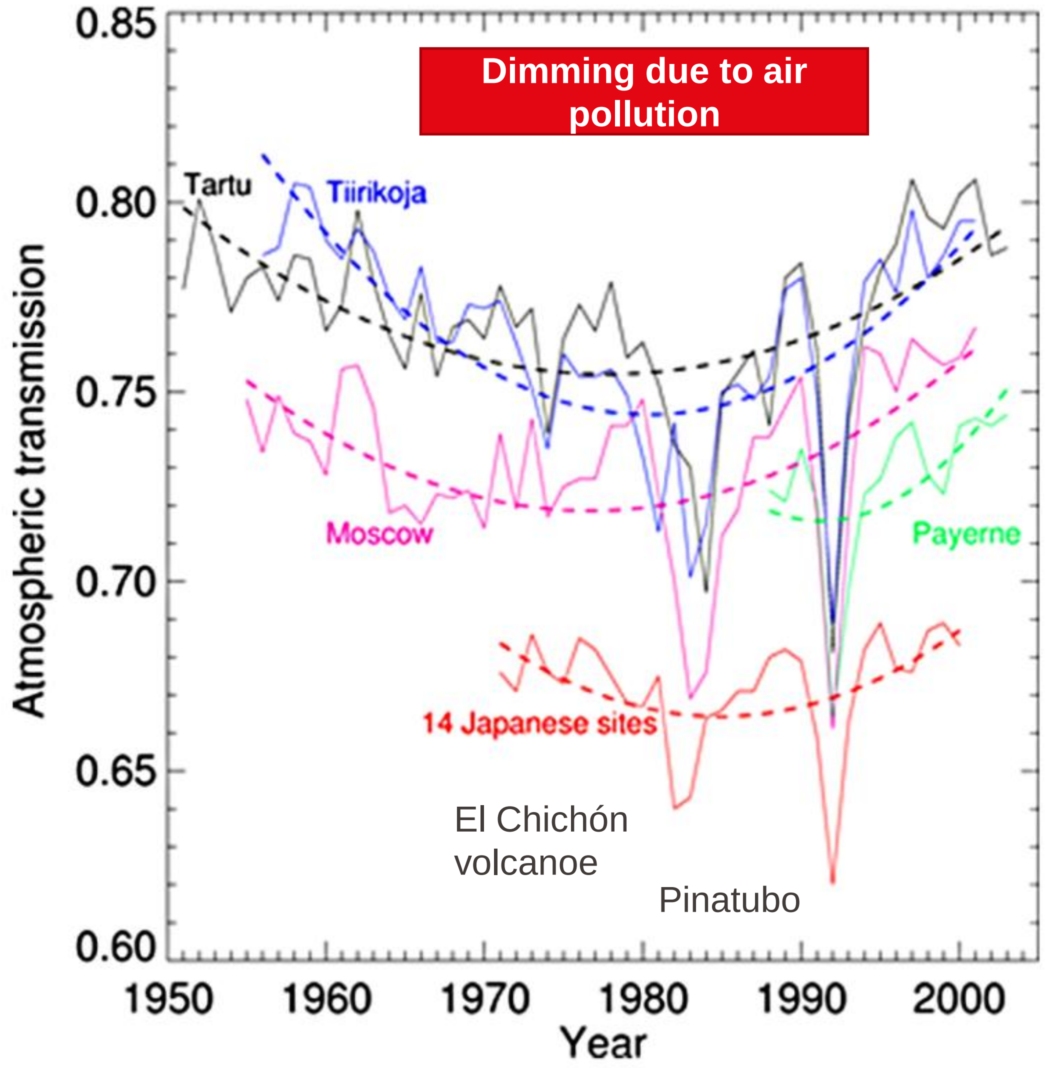
r and k_λ are independent of height (why?)

In the atmosphere, maximum extinction happens at a height at which optical depth is approximately one. This is because of decreasing air density with height and depleted radiation with depth.

If solar zenith angle > 0°, $\tau_\lambda = 1$ higher up in atmosphere.



Aerosol-radiation interaction



Aerosols prevent solar radiation from reaching the Earth's surface, they have a «dimming» effect.

Observed tendencies in surface solar radiation

	1950s-1980s	1980s-2000	after 2000
USA	-6	5	8
Europe	-3	2	3
China/Mongolia	-7	3	-4
Japan	-5	8	0
India	-3	-8	-10

Units in W m^{-2}

- Fuzzi et al., 2015, doi:10.5194/acp-15-8217-2015

[Wild, 2021, https://doi.org/10.1175/BAMS-D-11-00074.1](https://doi.org/10.1175/BAMS-D-11-00074.1)

Global view of aerosol optical depth

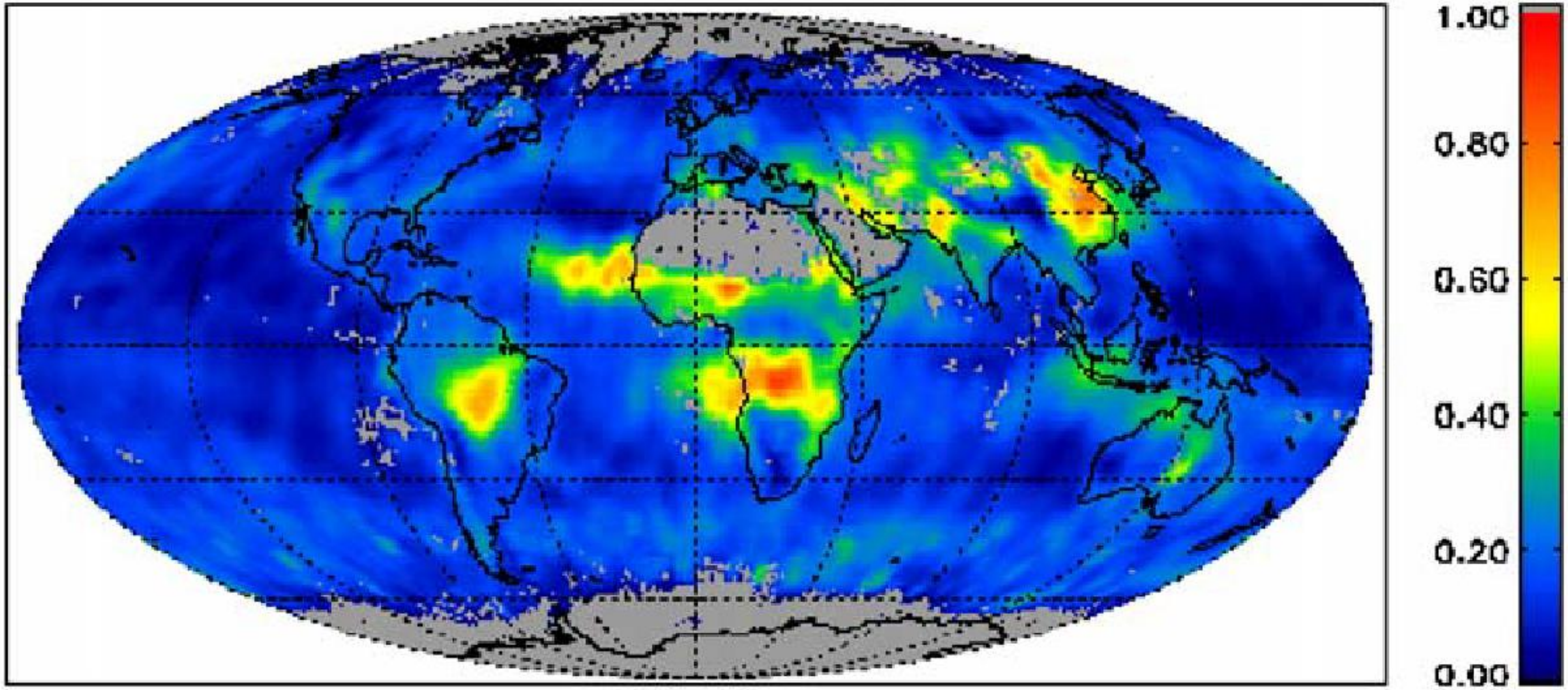
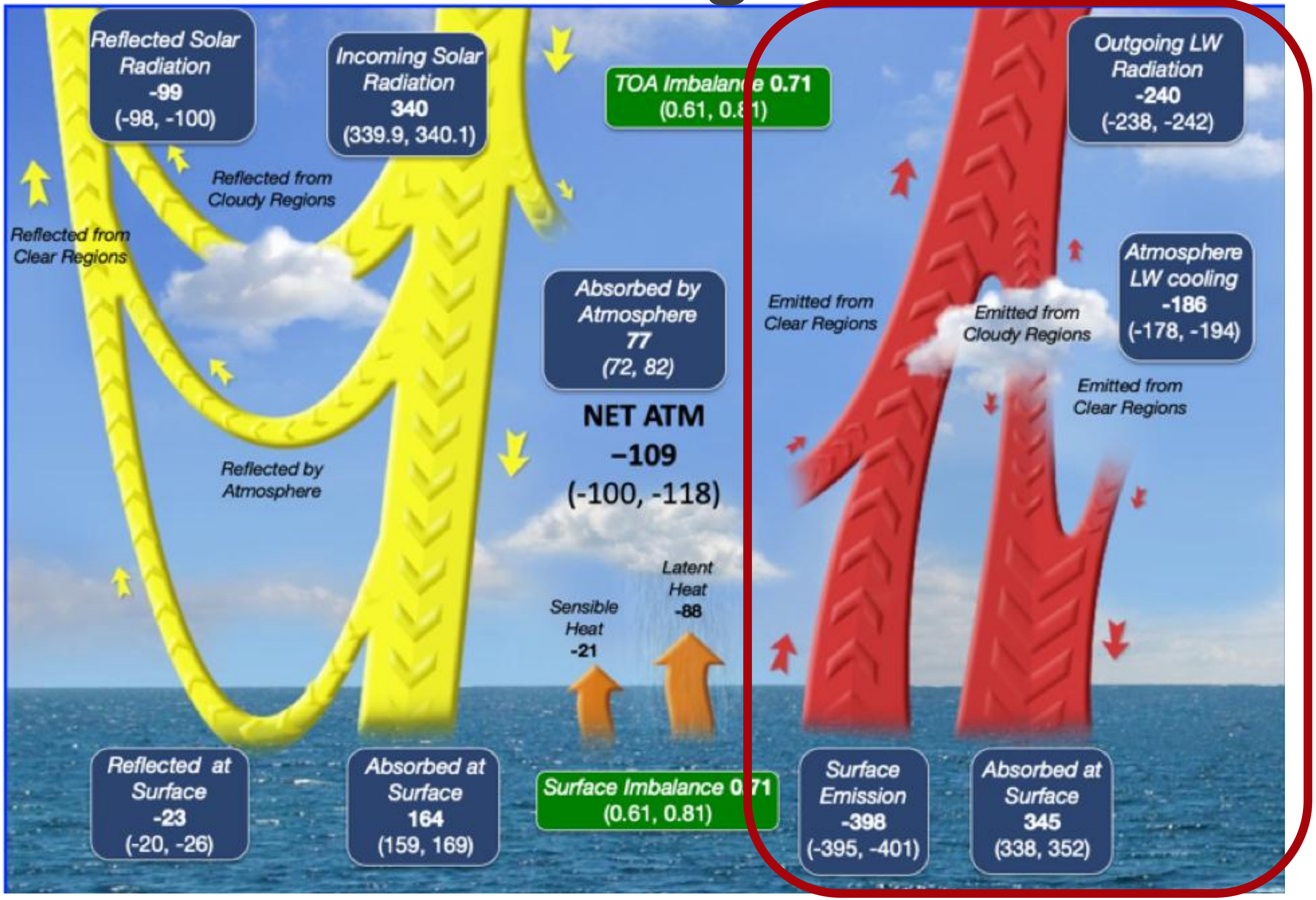


Figure 2. A global 8-days composite of MODIS aerosol optical thickness, τ_{aer}, at 0.55 μm

doi:10.1029/2002JD003273

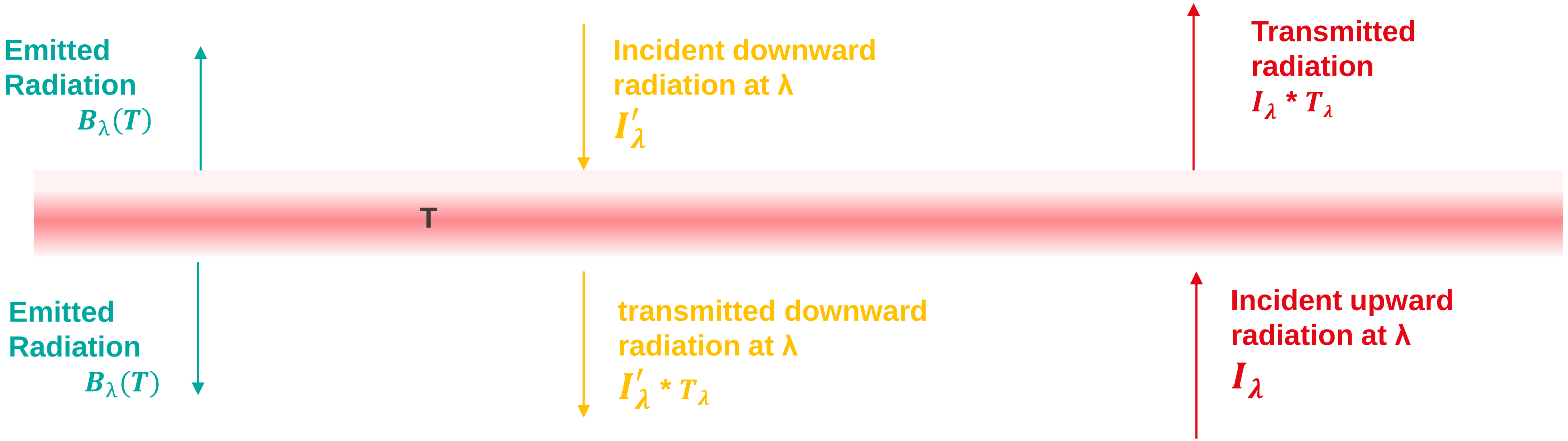
Radiative transfer through the atmosphere



Absorption and re-emission (no scattering)

https://ceres.larc.nasa.gov/science/

Consider a thin layer of atmosphere



Longwave radiative transfer: Schwarzschild

Longwave: Absorption + Emission
(Absorption = Extinction, simplification: no scattering because wavelength is large compared to particles/molecules)

$dI_\lambda(\boldsymbol{absorption})$= -$I_\lambda k_\lambda \rho r ds = -I_\lambda \alpha_\lambda$

α_λ absorptivity of the layer

I_λ monochromatic intensity
k_λ mass absorption coefficient
ρ air density
r mixing ratio of absorbant
ds path length

+

$dI_\lambda(\boldsymbol{emission})$= $B_\lambda(T)\varepsilon_\lambda$

$B_\lambda(T)$ Planck function
ε_λ emissivity of gas

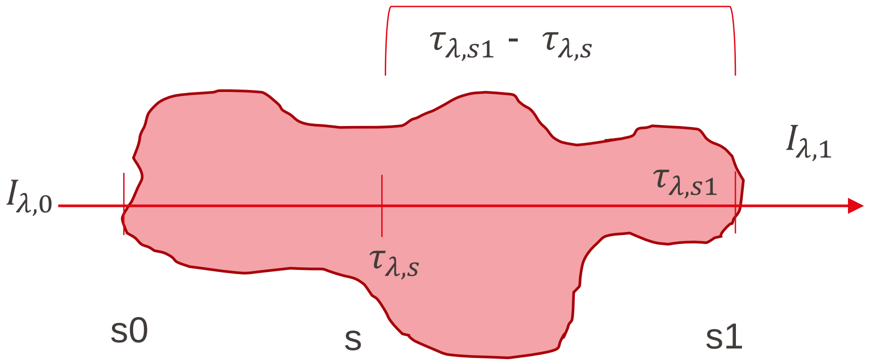
Kirchhoff
$\varepsilon_\lambda = \alpha_\lambda$

Schwarzschild equation:

$$dI_\lambda = -(I_\lambda - B_\lambda(T))k_\lambda \rho r ds$$

Schwarzschild equation integrated over a path:

$$I_\lambda(s_1) = \underbrace{I_{\lambda 0} e^{-\tau_\lambda(s_1,0)}}_{\text{1. attenuation}} + \underbrace{\int_0^{s_1} k_\lambda \rho r B_\lambda[T_s] e^{-\tau_\lambda(s_1,s)} ds}_{\text{2. emission and absorption}}$$



1: monochromatic intensity from s = 0 reaching s1, depletion along path

2: monochromatic intensity emitted by gas along path s = 0 to s = s1 reaching s1

$e^{-\tau_\lambda}$ denotes transmissitivity

Influence of optical depth

Schwarzschild equation:

$$dI_\lambda = -\left(I_\lambda - B_\lambda(T)\right)k_\lambda \rho r ds \;\rightarrow\; I_\lambda(s_1) = I_{\lambda 0}e^{-\tau_\lambda(s_1,0)} + \int_0^{s_1} k_\lambda \rho r B_\lambda[T_s] e^{-\tau_\lambda(s_1,s)}ds$$

- $\tau_\lambda = 1, \qquad |B_\lambda(T) - I_\lambda| = {}^1/_e$
- $\tau_\lambda \ll 1$, emissitivity of atmosphere is so small that radiation does not escape Earth
- $\tau_\lambda \gg 1$, absorption by atmosphere so large that radiation does not escape Earth

- Result depends strongly on wavelength, in absence of absorption bands, atmospheric ***windows*** «open» and IR escapes directly to space

Radiative temperature tendency

Defined as the temperature change over time due to absorption and emission of radiation in an atmospheric layer.
Also called radiative heating or cooling, or flux divergence.

$$\frac{dT}{dt} = - \frac{dF(z)}{dz} * \frac{1}{\rho c_p}$$

$F = F\uparrow - F\downarrow$, F is net flux
$F(z)$ radiative flux density as function of altitude
c_p heat capacity of molecules

Using Schwarzschild's equation
to calculate flux divergence:

$$dI_\lambda = -(I_\lambda - B_\lambda(T))k_\lambda \rho r ds \quad , dF = \int dI(\boldsymbol{\Phi}, \boldsymbol{\Theta}) \cos \boldsymbol{\Theta} \, d\omega$$

$$\left(\frac{dT}{dt}\right)_\lambda = -\frac{1}{\rho c_p}\int_{4\pi} \frac{dI_\lambda}{ds} d\omega = -\frac{1}{c_p}\int_{4\pi} k_\lambda r\,(I_\lambda - B_\lambda)\, d\omega \text{ (°C/day)}$$

daytime dominated by shortwave, nighttime by longwave
Cold, clear sky night vs warmer, cloudy night

apply to longwave only:
- 2.5°C/day
apply to shortwave only:
+0.5°C/day

Upward flux is defined as negative
Integration over a sphere because of diffuse radiation emitted by the atmospheric layer

With the divergence equation we can calculate atmospheric heating and cooling rates for greenhouse gases (*cooling to space approximation*).

Heating and cooling from radiative transfer (gases only)

Nobel prize 2021!

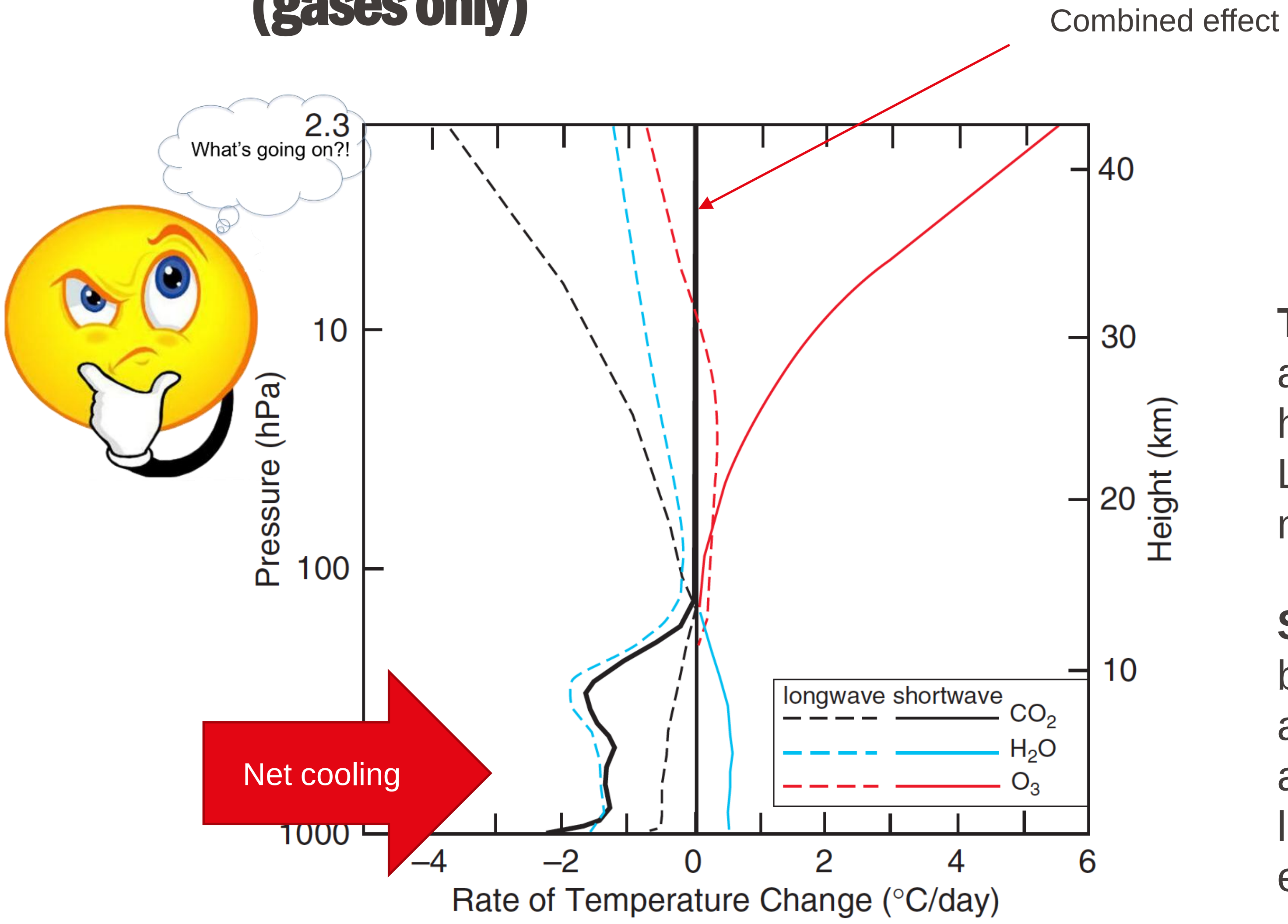


Fig. 4.29

Climate model from the 60s (Manabe and Strickler (1964)

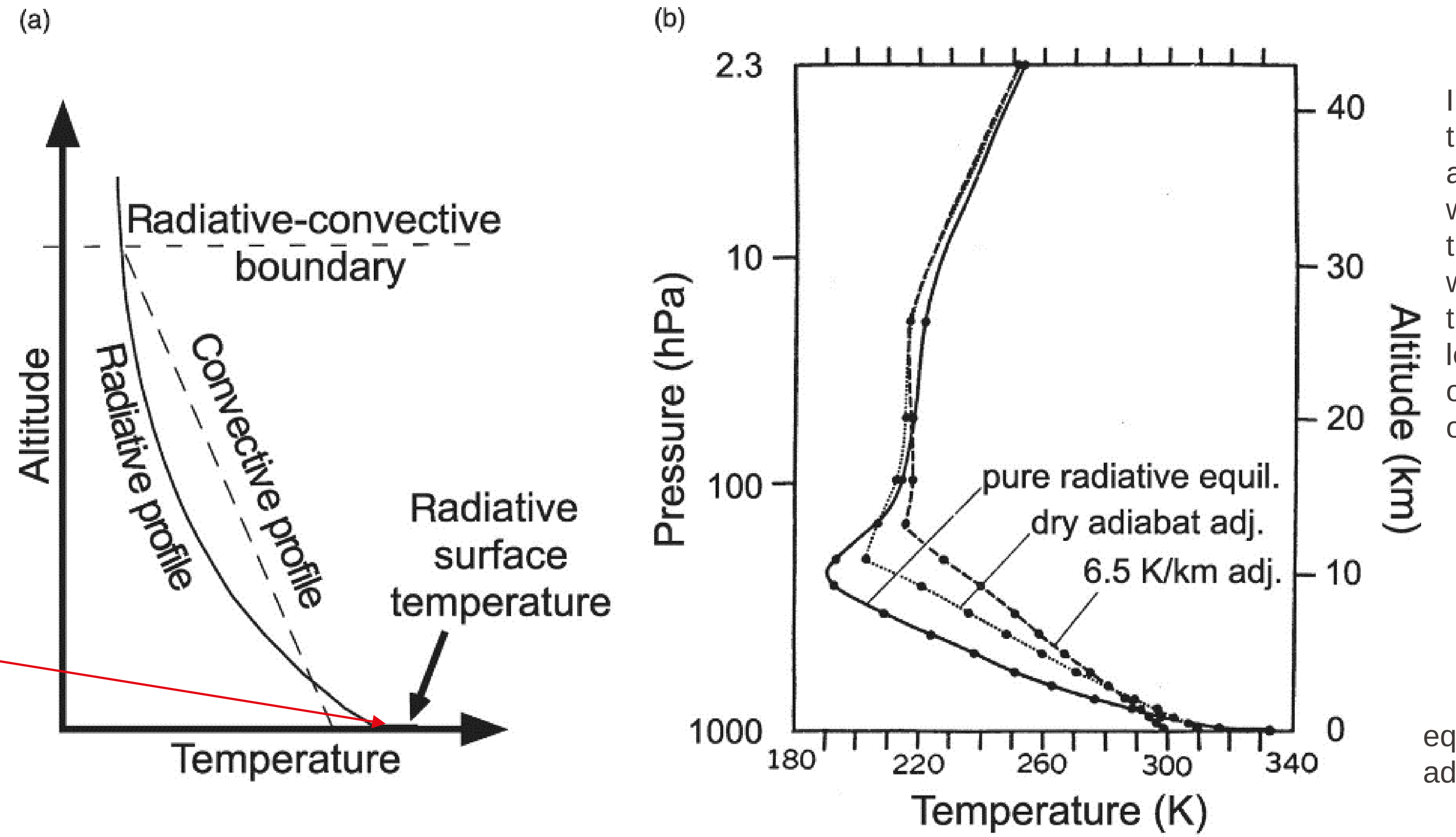
- Longwave & shortwave
- Average Earth cloudiness assumption
- 3 greenhouse gases

Troposphere: Cooling due to H_2O, mainly, and CO_2. Effect of H_2O decreases with height along with smaller mixing ratio. Longwave CO_2 effect and shortwave H_2O nearly cancel each other.

Stratosphere: Equilibrium due to warming by O_3 (UV absorption, and a little bit of IR absorption at 9.6 µm) and cooling by H_2O and CO_2 (dominant cooling agent through longwave emission). Model assumes equilibrium in the stratosphere.

Convective - radiative equilibrium

Because the density of the atmosphere increases close to the surface, the layers become optically thicker and physically thinner. Hence, removing heat near the surface becomes difficult.
If only radiative equilibrium were responsible for the energy balance, Earth's surface would be very hot!

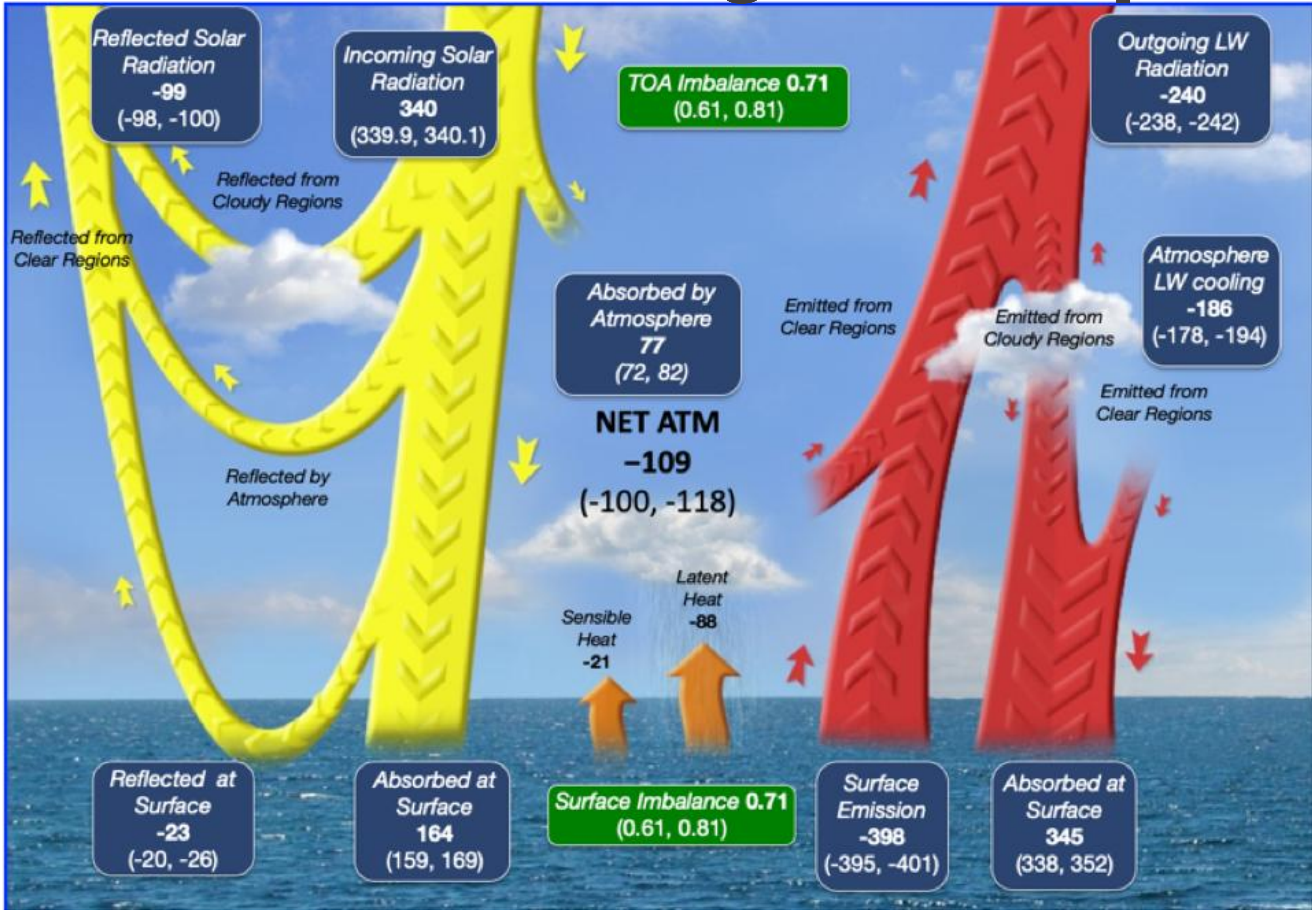


In fact, convective transport of sensible and latent heat of water vapor cools the surface and warms the troposphere and leads to the observed lapse rate of -6.5 K / km.

equil. = equilibrium
adj. = adjustment

Catling, D., & Kasting, J. (2017). Principles of Planetary Atmospheres. In *Atmospheric Evolution on Inhabited and Lifeless Worlds* (pp. 1-168). Cambridge: Cambridge University Press. Wallace and Hobbs, 2006, Ch. 4.3.6

Radiative transfer through the atmosphere



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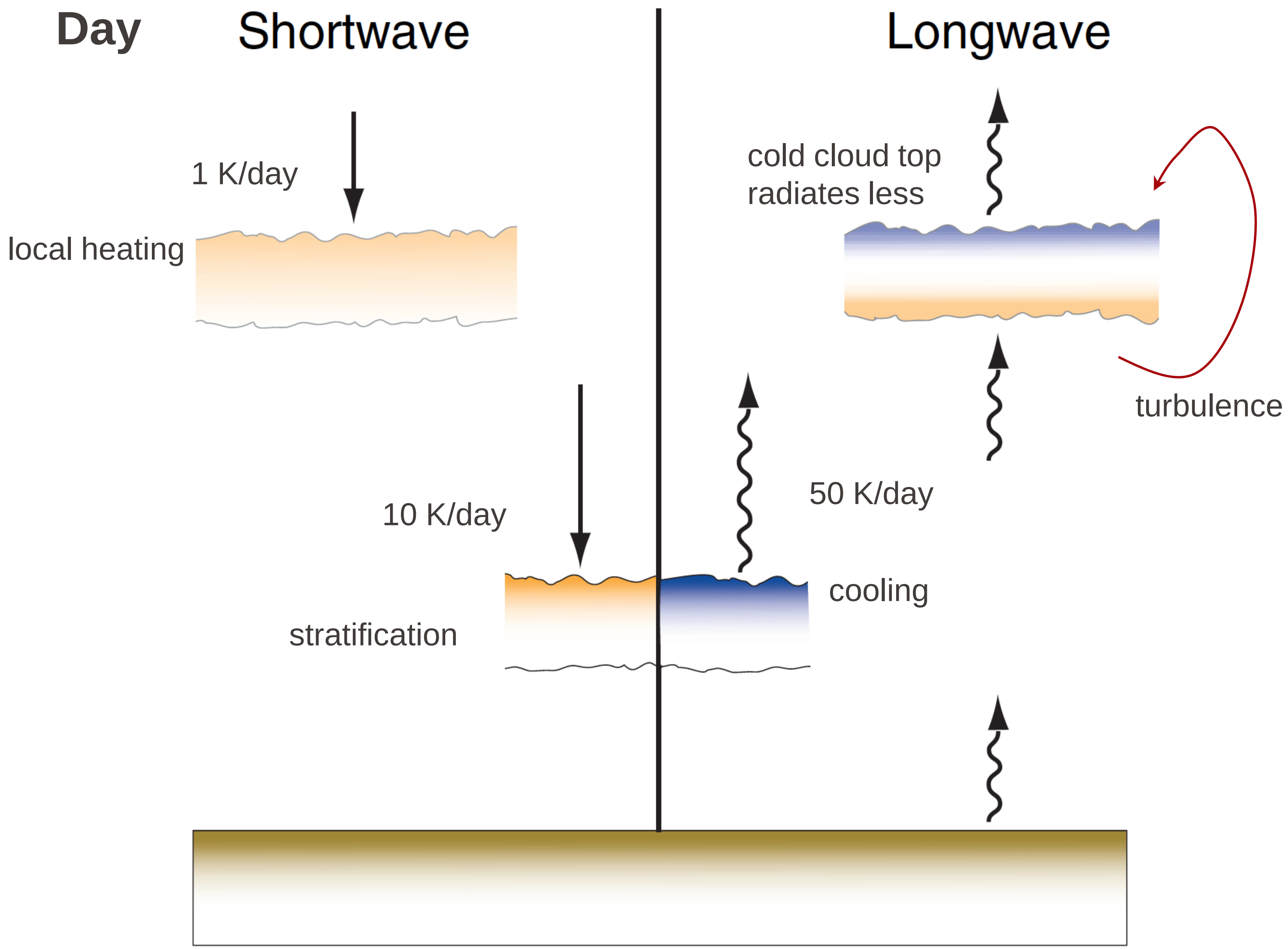
Cooling and heating in cloud layers



Clouds modify the heating / cooling significantly.

Longwave cooling is stronger than shortwave heating averaged over a day.

Longwave cooling and heating typically promotes vertical instability.

Shortwave heating of cloud tops leads to a more stable stratification.

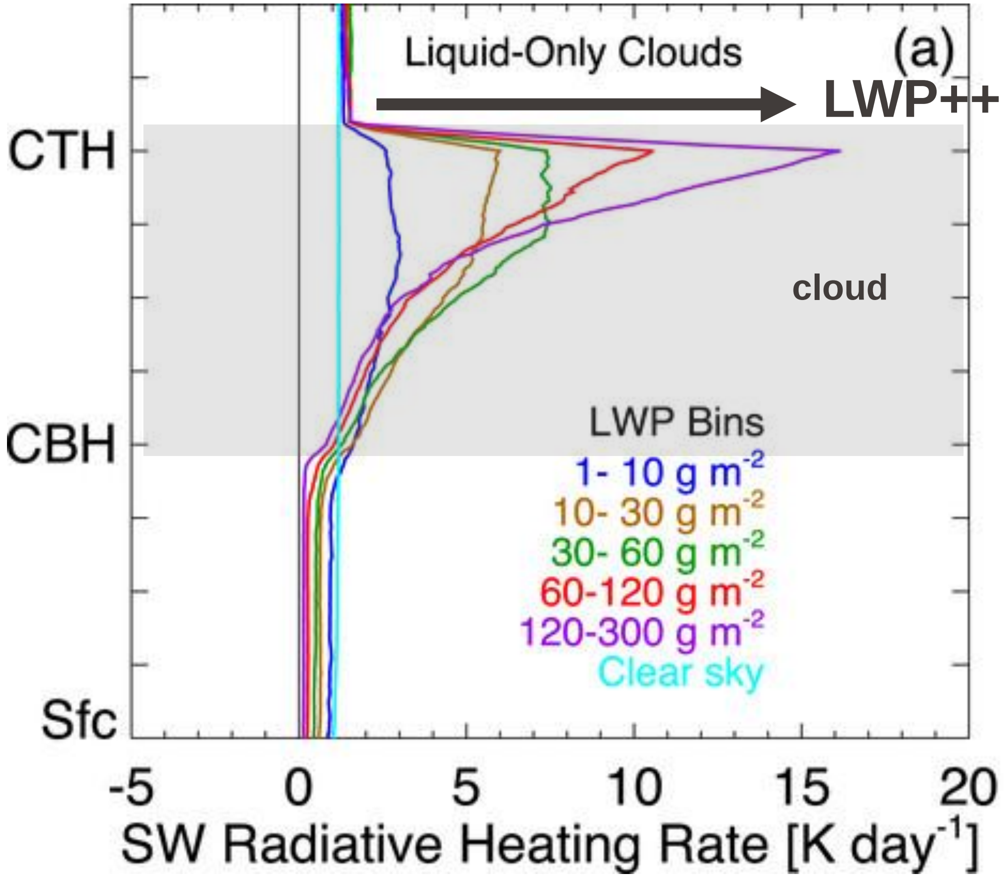
Cloud top heating and cooling observations

Data from 2 years of observation at Utqiagvik, Alaska, USA; only liquid clouds (no ice crystals)
LWP = liquid water path, amount of liquid water in a cloud

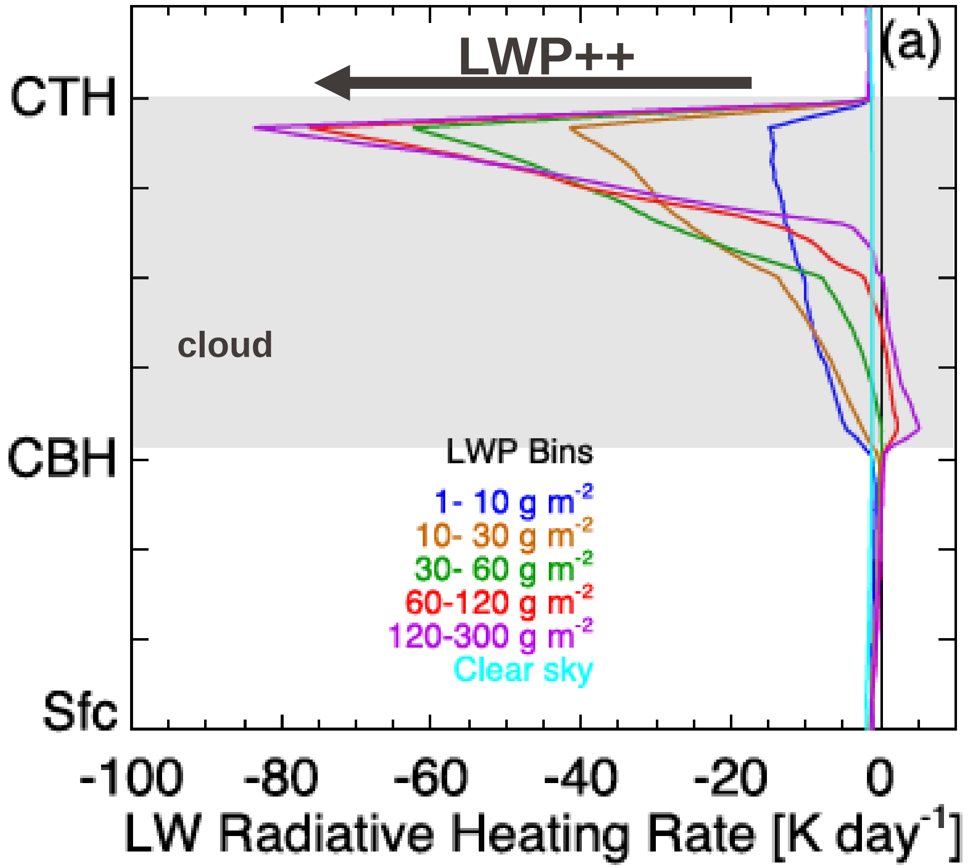
Cloud top height

Cloud bottom height

surface



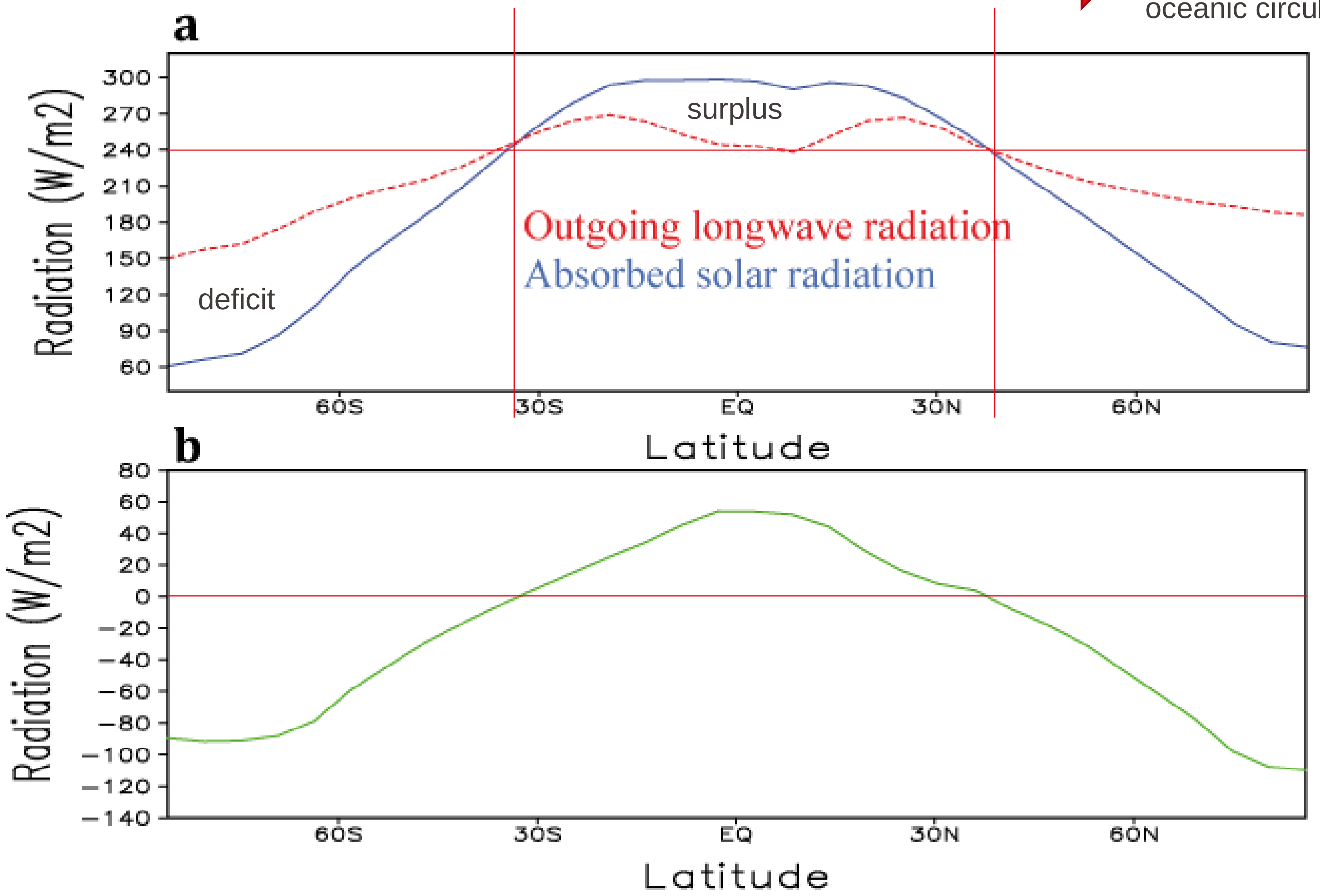
shortwave



longwave

- (Turner et al., 2018)

Annual average flux density



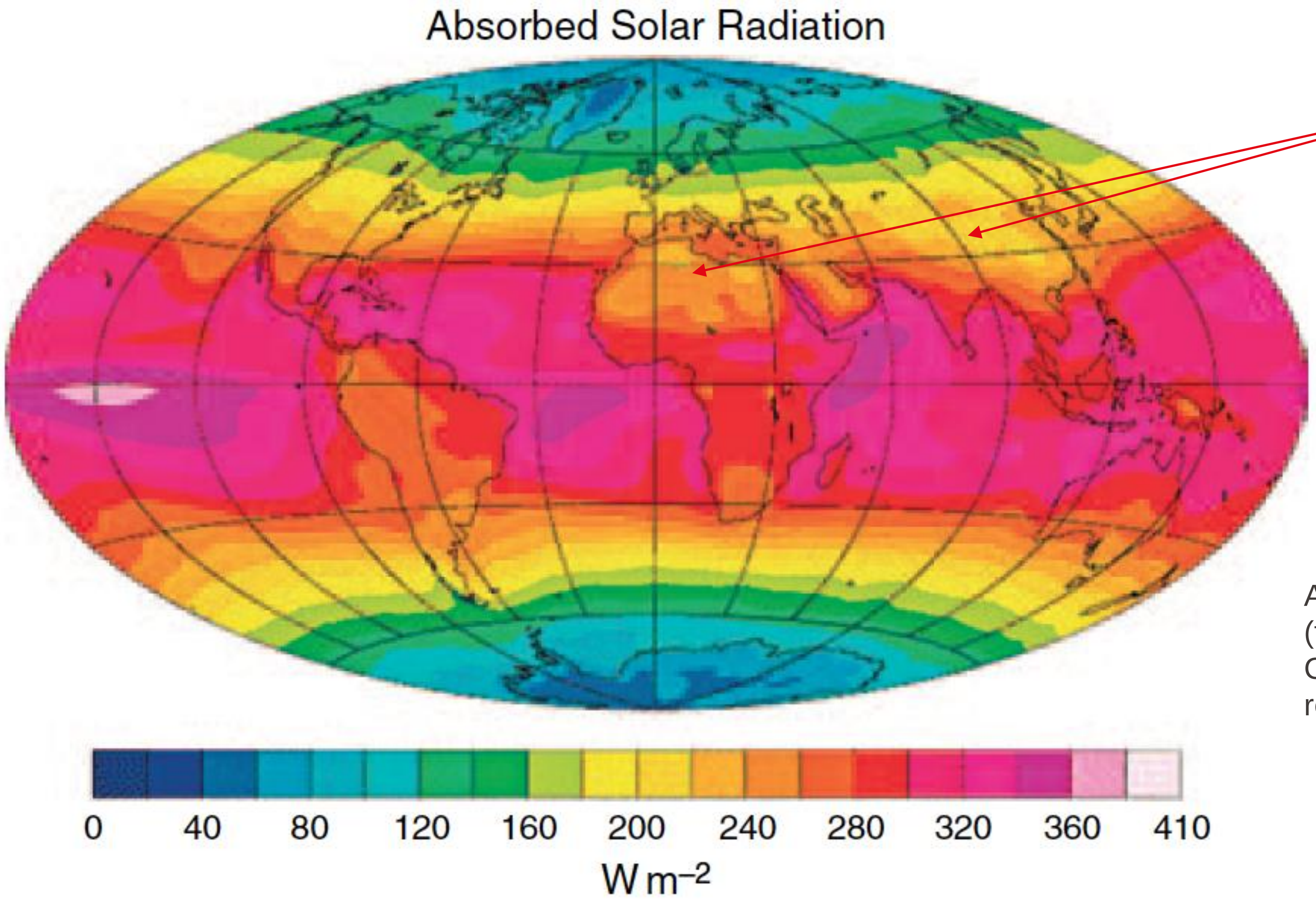
Difference drives atmospheric and oceanic circulation

240 W/m2 averaged absorbed solar radiation, does not look like average, why?

Subtracted the lines from panel a)

http://www.climate.be/textbook/chapter2_node6.html

Radiation Balance

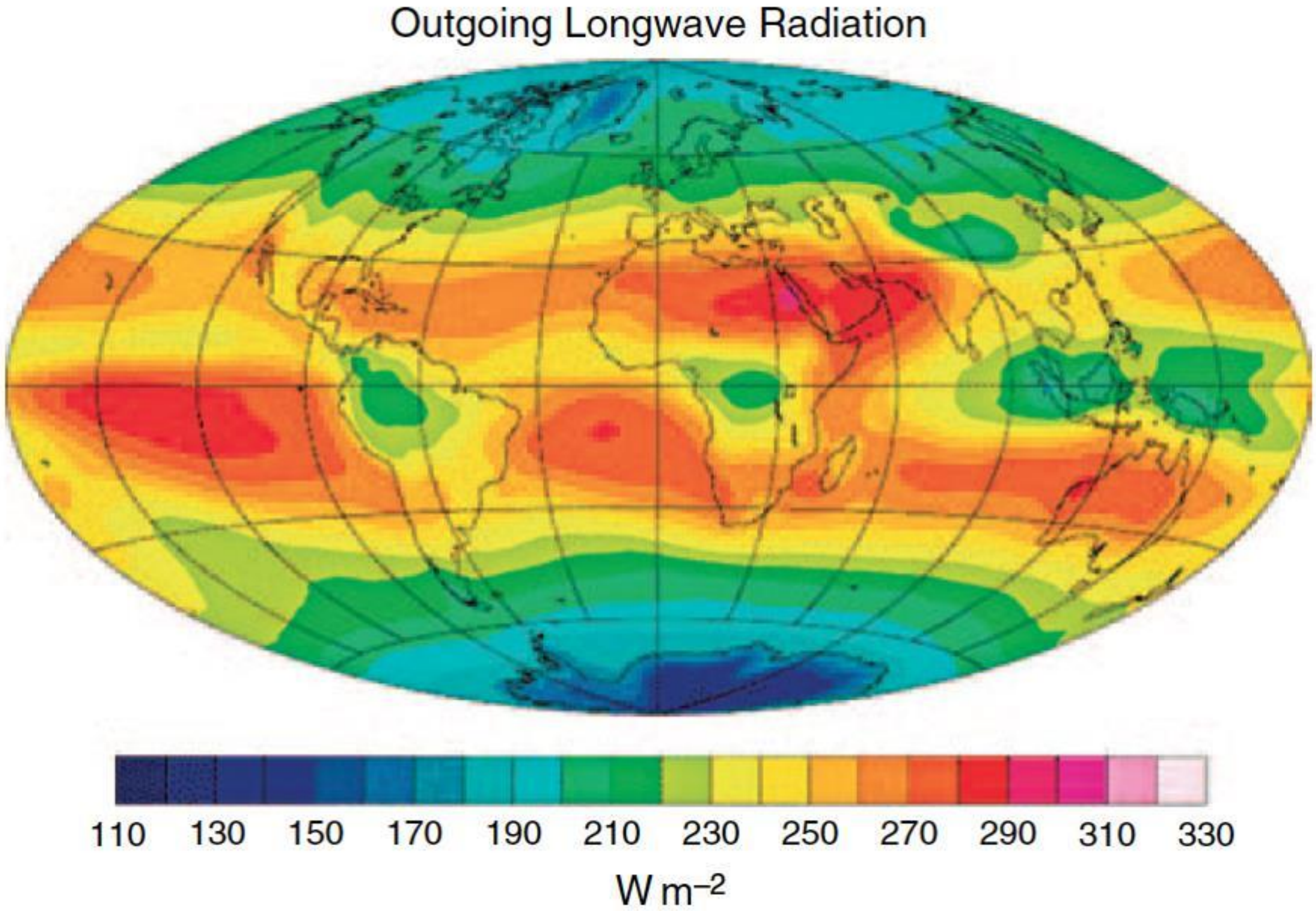


Different.
Why?

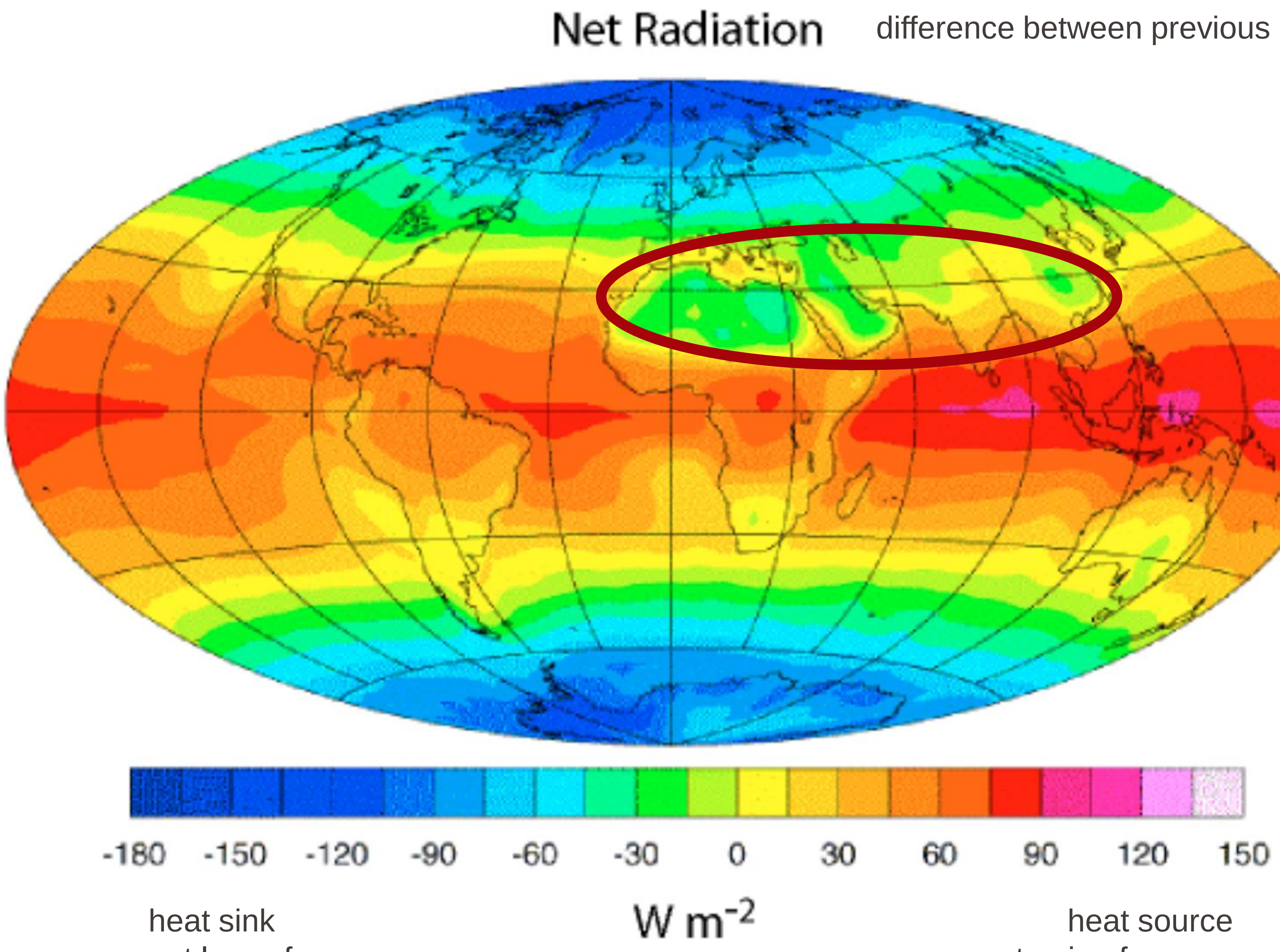
Annual mean net downward shortwave radiation (takes into account: solar declination, local albedo). Clouds reflect more, deserts as well. In polar regions half of the year is dark.

EPFL

Can you indicate regions with high clouds?



Top of atmosphere net radiation



difference between previous plots

heat sink
net loss of energy

heat source
net gain of energy

Sahara desert, Arabian Peninsula and Himalayas are an anomaly. It looses heat at a latitude, where there is normally a net warming. Why?

The difference between the low and high latitudes drives global atmospheric circulation patterns.

Temperature soundings from satellites

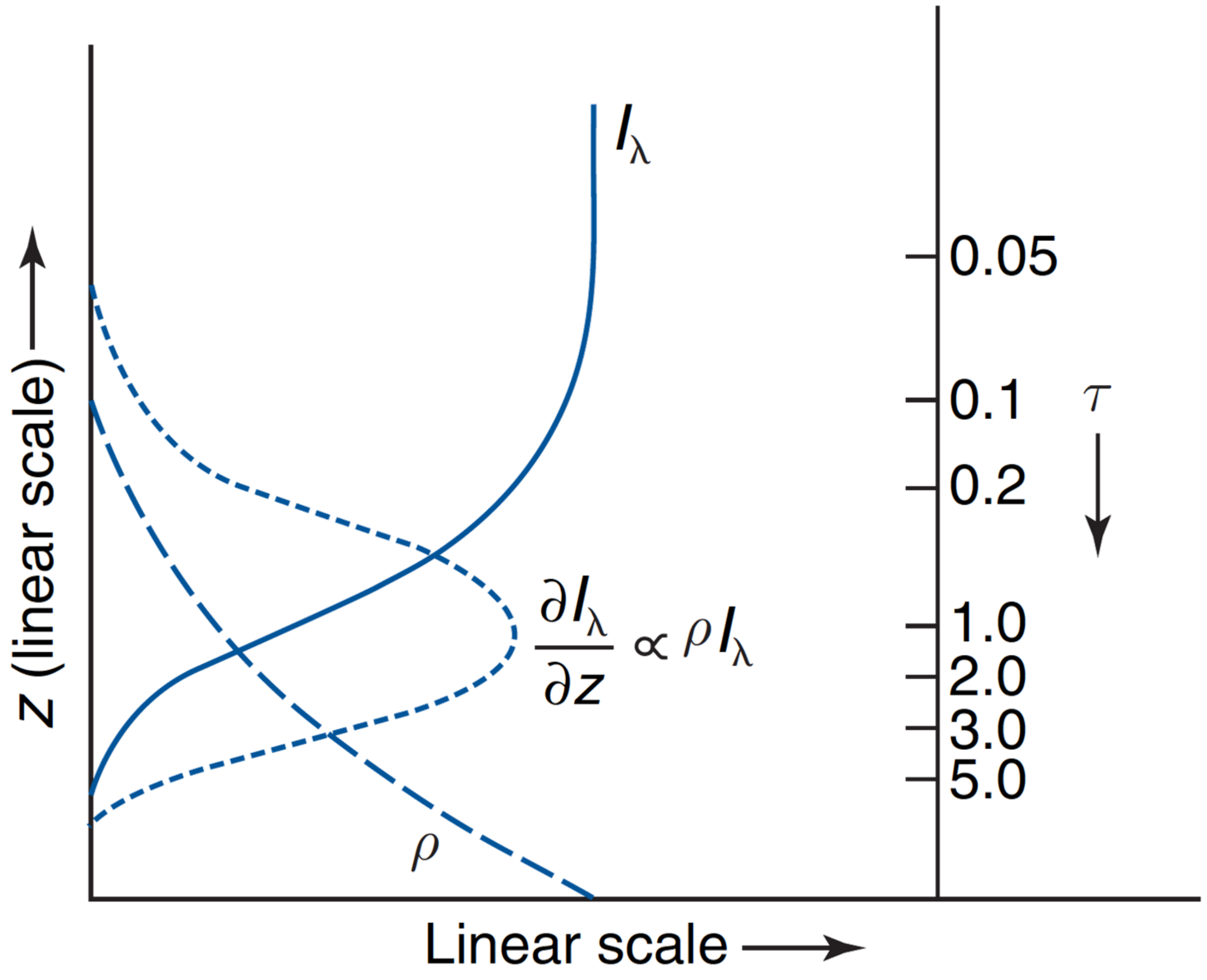
Based on optical depth

$$\tau_\lambda = \int_z^\infty \rho r k_\lambda dz$$

1: Most of the radiation reaching the satellite is emitted at $\tau_\lambda = 1$

2: based on the Planck function (emission as a function of temperature)

$$B(\lambda, \mathrm{T})$$



Earth's emission seen from space

Emission mainly emitted at $\tau_\lambda = 1$
Blue curve shows the temp where $\tau_\lambda = 1$

Infrared interferometer reads Earth «**brightness» temperature**.

In this graph, the radiation emitted by the Earth's surface retrieved at the TOA (e.g on a satellite) is shown. The grey lines show blackbody emissions for different temperatures. Where the "measured" spectrum is close to the curve of a high temperature (called the brightness temperature), atmospheric "windows" are present. Absorption bands show low brightness temperatures.

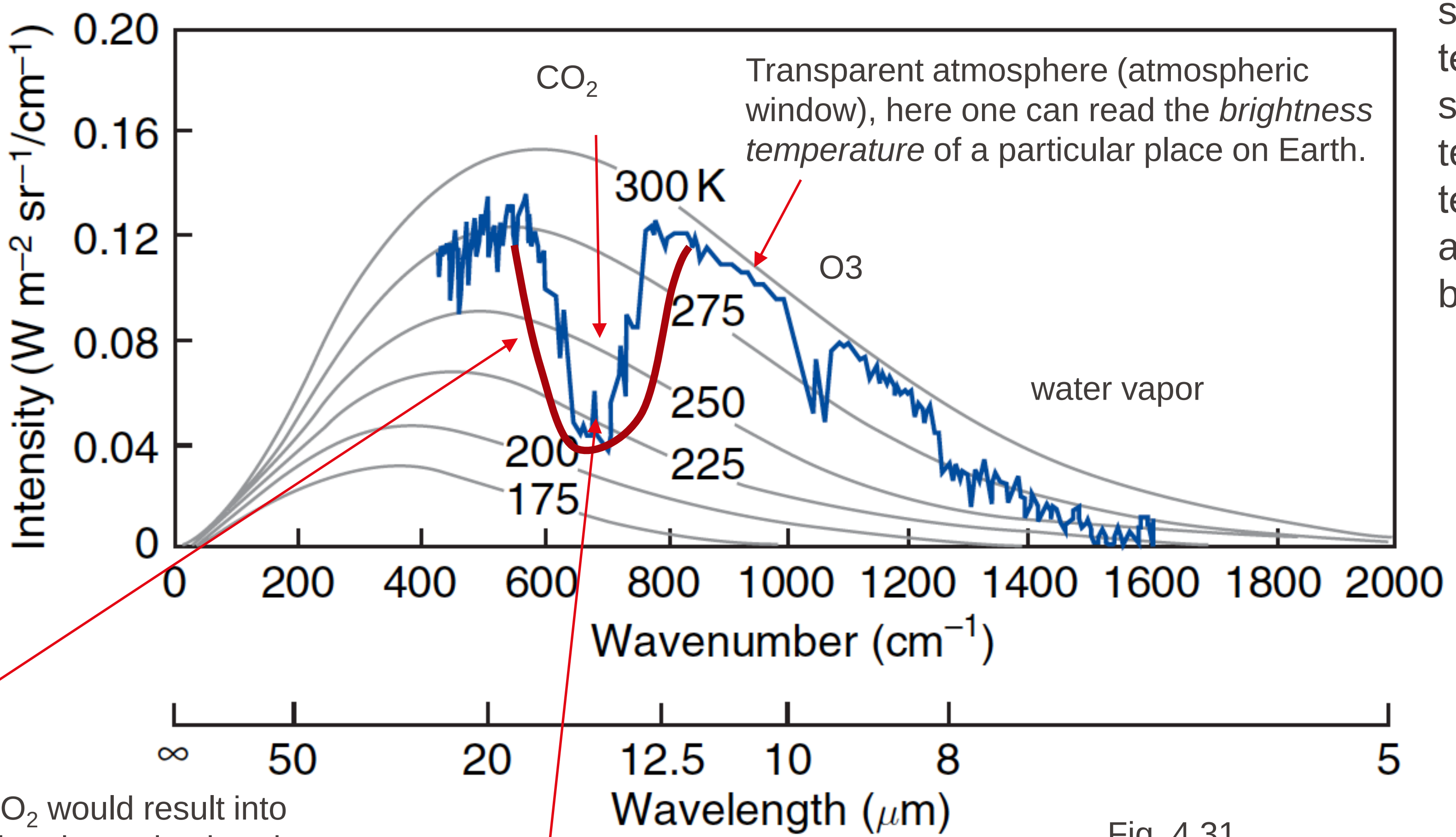
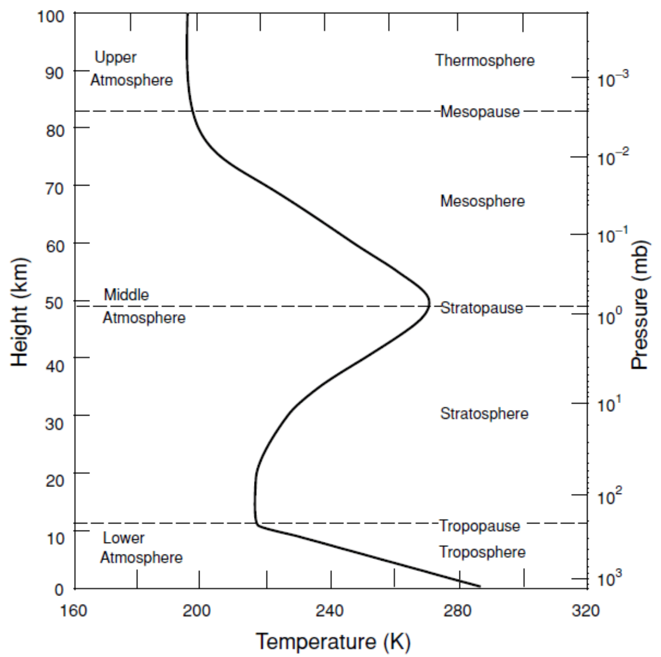


Fig. 4.31

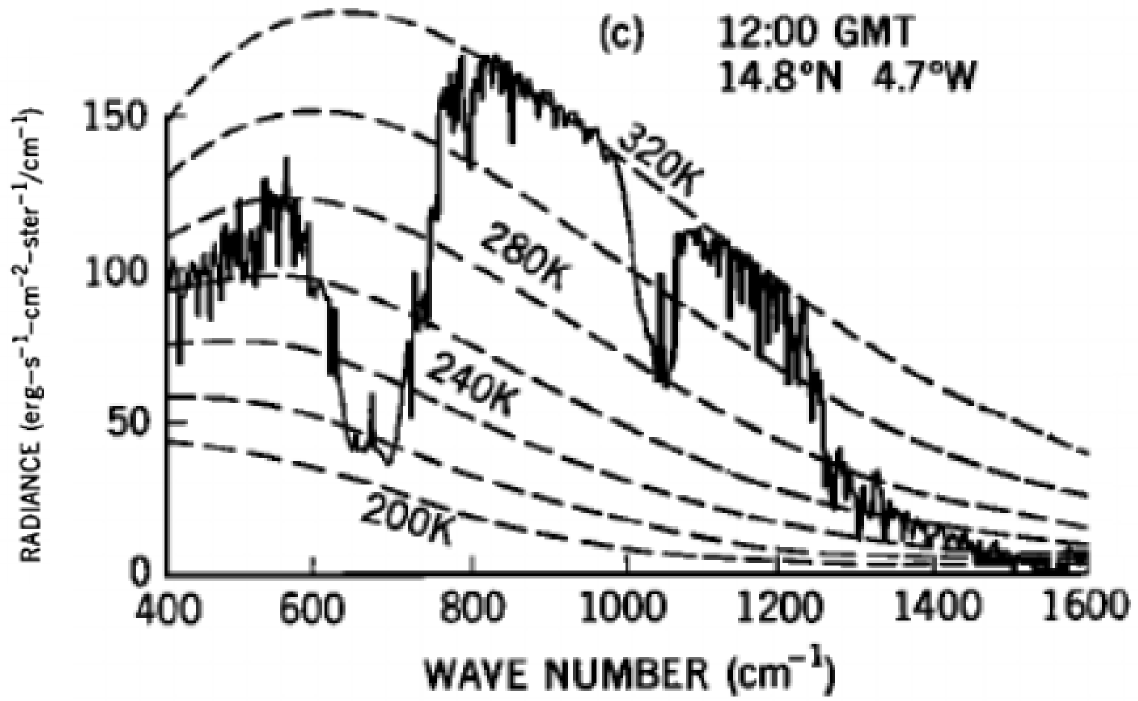
More CO_2 would result into a broader absorption band

Warmer stratosphere, inversion

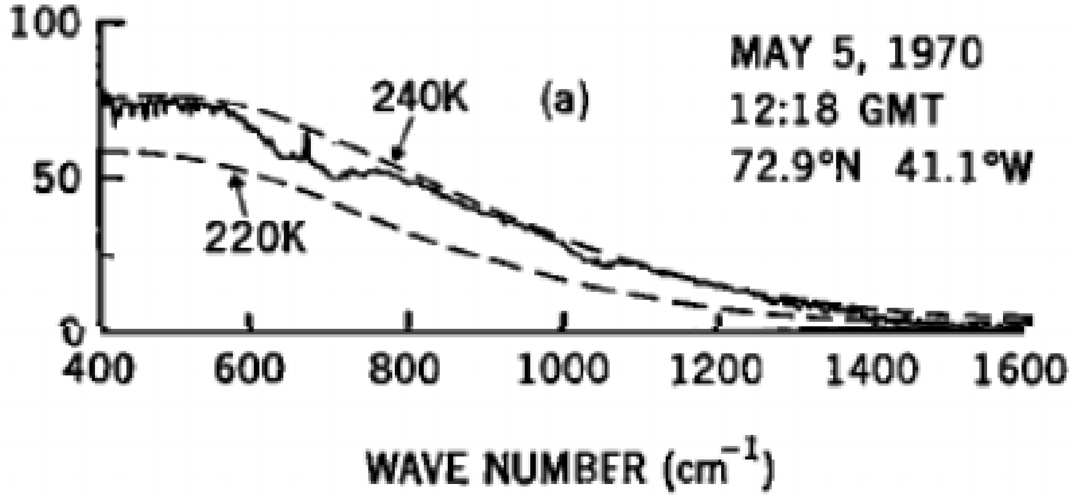


Earth’s emission: different locations

Saharan Desert



Greenland

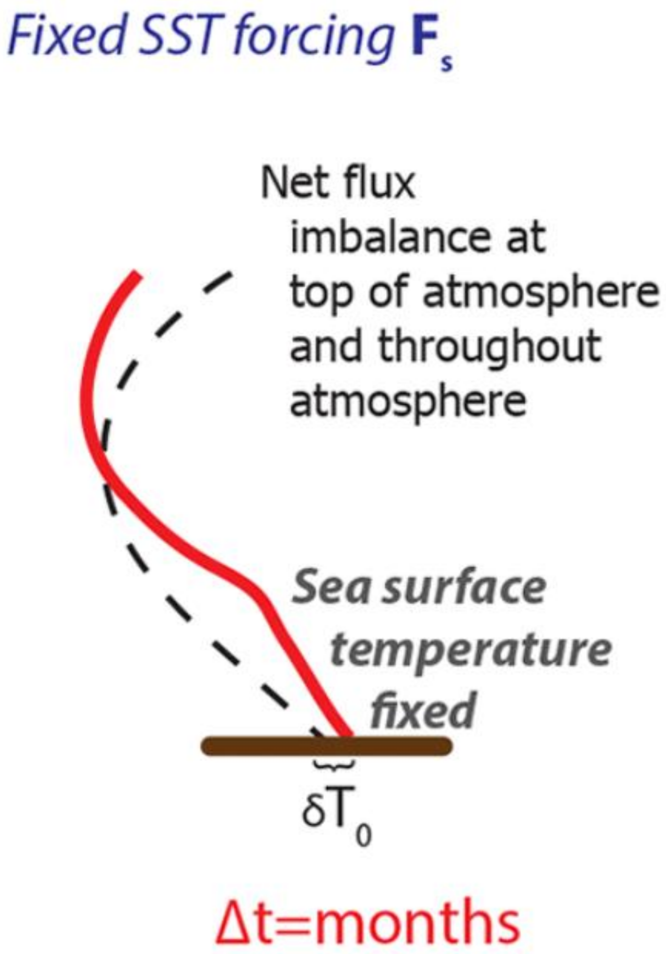


Hanely et al. 1972

Radiative forcing (W m^{-2})

- A forcing that changes the radiation balance of Earth.
- Fundamental definition: Warming of Earth's surface and lower atmosphere is driven by radiative forcing, the difference between the flux of thermal radiant energy from a black surface through a hypothetical, transparent atmosphere, and the flux through an atmosphere with greenhouse gases, particulates and clouds, but with the same surface temperature.
- Climate science definition: Radiative forcing in climate science is commonly compared against the preindustrial time (not hypothetical atmospheres).
- Because greenhouse gases (GHG) and aerosols change the thermal IR radiative flux at the top of atmosphere (TOA, net smaller flux) they induce radiative forcing.

Effective radiative forcing (ERF)

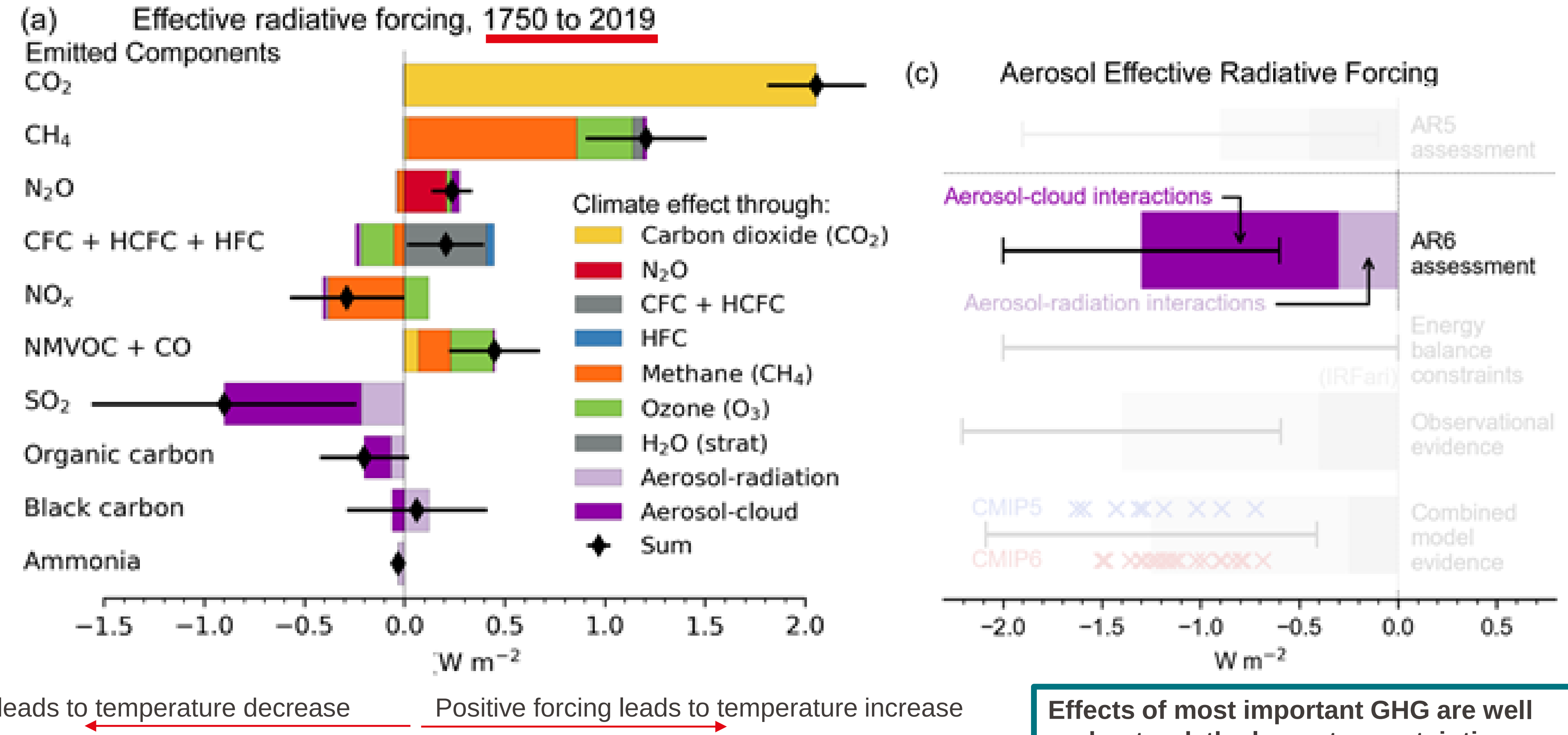


ERF is the ensuing radiative forcing once all rapid adjustments for temperature (including the stratospheric domain), water vapour, surface albedo (snow and ice cover, vegetation), and clouds are taken into account in response to a change in a forcing agent such as increasing GHG concentrations.
Sea surface temperatures and sea ice cover are fixed at climatological values unless otherwise specified. Hence ERF includes both the effects of the forcing agent itself and the rapid adjustments to that agent.

Heinze et al., ESD, 2019, https://doi.org/10.5194/esd-10-379-2019

ERF by component

GHG make up roughly 3.7 W m^{-2} radiative forcing
Compare: to TOA outgoing LW of -240 W m^{-2}

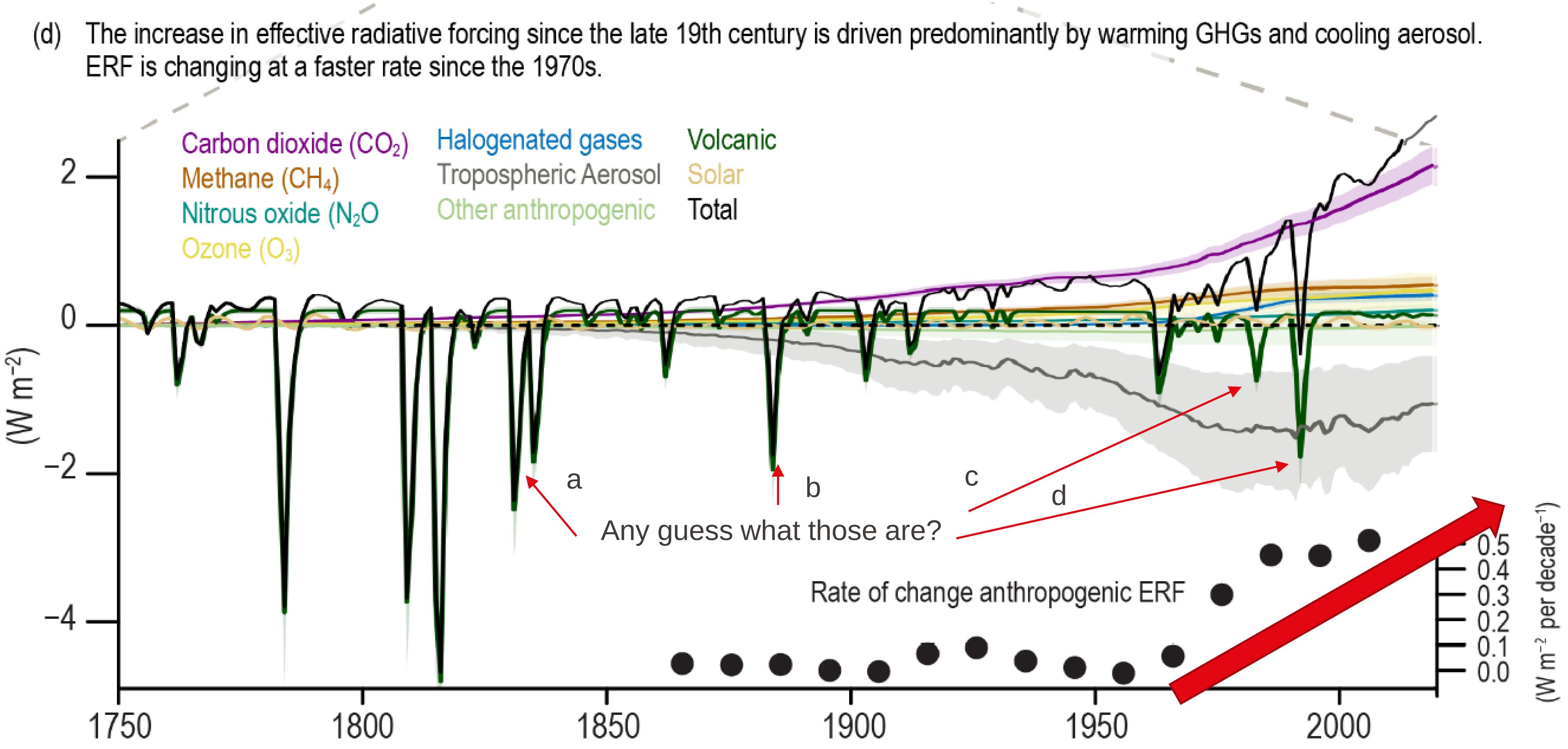


Effects of most important GHG are well understood, the largest uncertainties are with aerosols and clouds.

- IPCC AR6, Fig. TS.15

ERF longterm perspective

a – Kelud 1826, Indonesia
b – Krakatoa 1883, Indonesia
c – El Chichón 1982, Mexico
d – Pinatubo 1991, Philippines

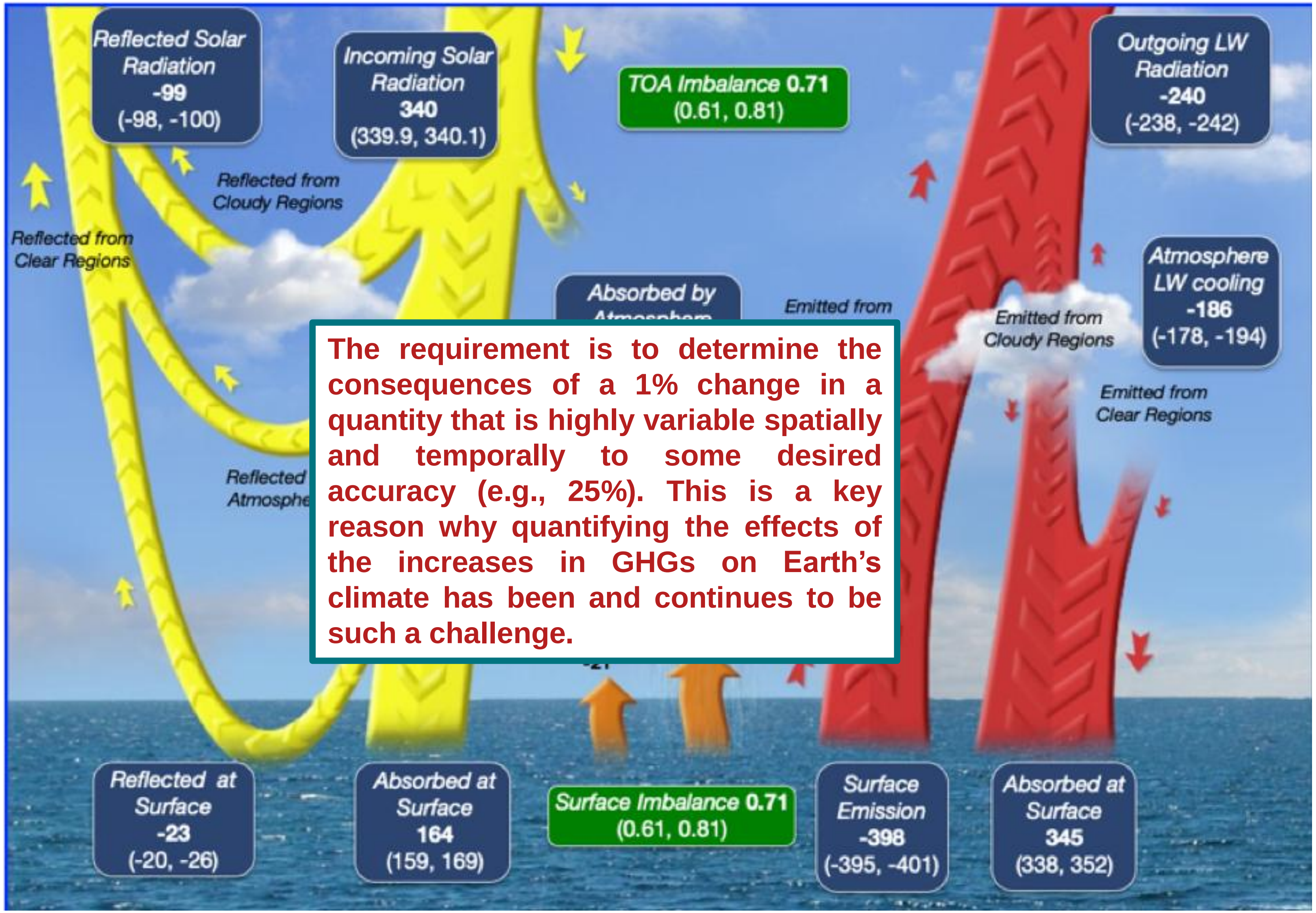


Rate of forcing is increasing as well!

- IPCC AR6, Fig. TS.9

Read here for more on volcanic eruptions

GHG ERF cannot be measured



The requirement is to determine the consequences of a 1% change in a quantity that is highly variable spatially and temporally to some desired accuracy (e.g., 25%). This is a key reason why quantifying the effects of the increases in GHGs on Earth's climate has been and continues to be such a challenge.

- Calculation needs several ingredients:
  - Knowledge of concentration change (measurements) → some more detailed discussion upcoming
  - Greenhouse gas properties: lifetime (reviewed), and effectiveness
  - Models with realistic clouds, temperature and water vapor to estimate how sensitive climate is to ERF
  - **ENV-410 science of climate change**

https://ceres.larc.nasa.gov/science/

Take home messages

- Check out: https://ceres.larc.nasa.gov/science/
- What are atmospheric windows?
- What are key properties of greenhouse gases?
- The warming and cooling effects of clouds and aerosols.
- Unit optical depth as key parameter for temperature profile retrieval.
- Top of atmosphere radiation balance.
- Radiative forcing and the challenge of climate science.