Physics of the atmosphere

LECTURE 2
RADIATIVE BUDGET

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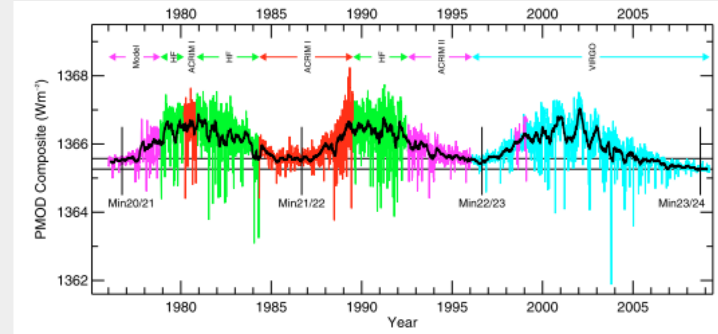
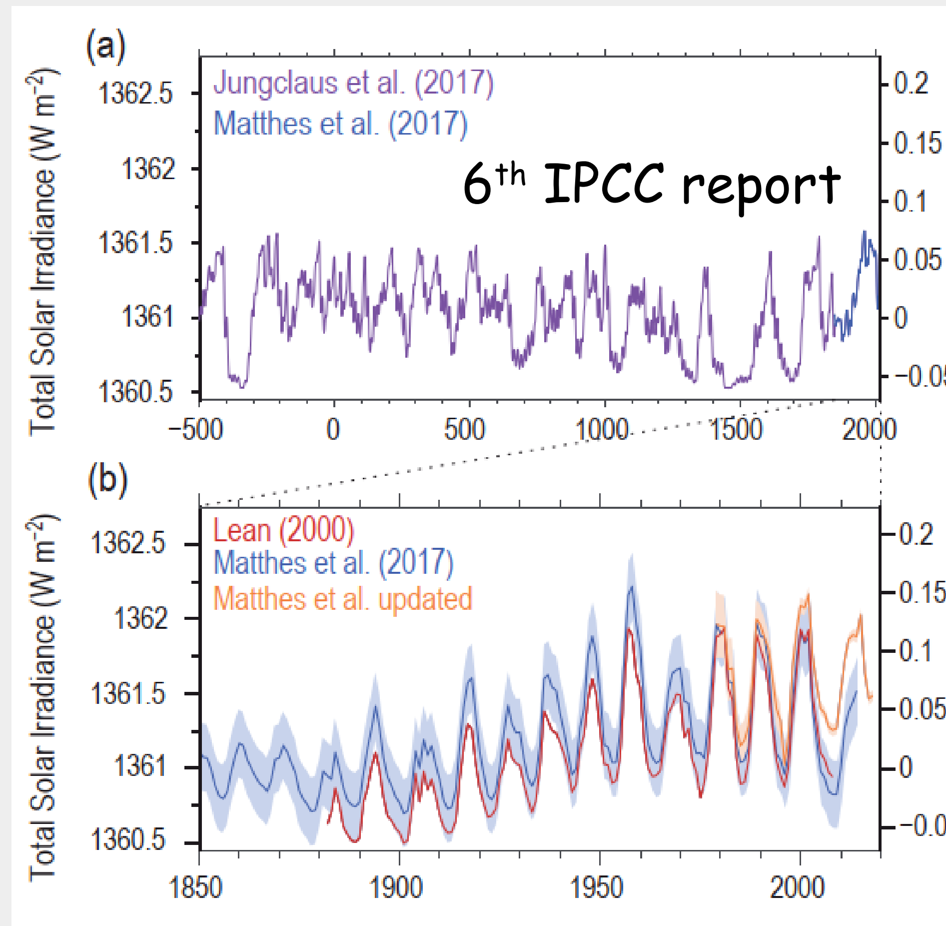
I.1 Incoming solar radiation and terrestrial outgoing radiation

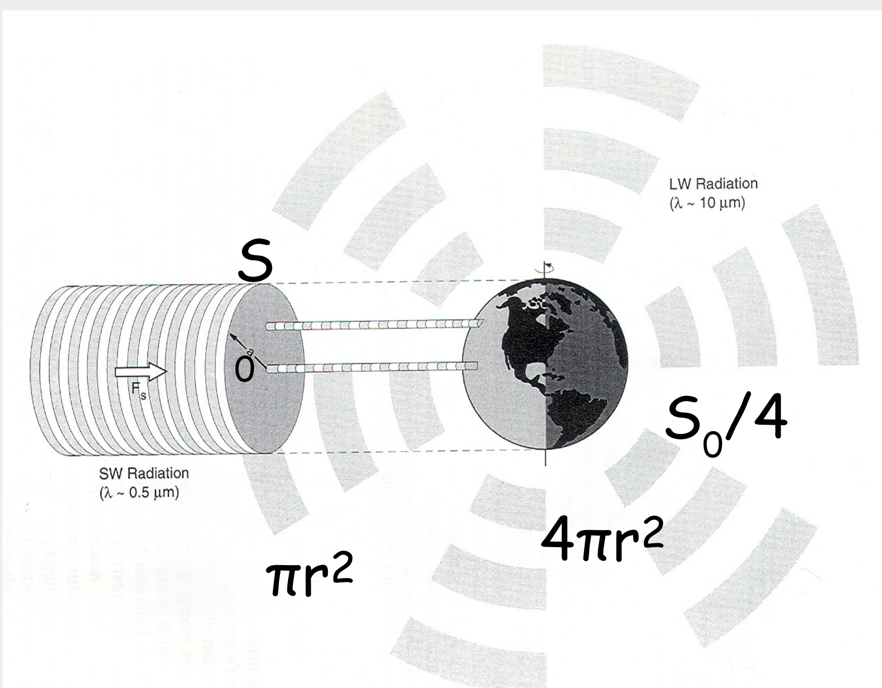
Incoming solar radiation
$S_0 \approx 1361\ W\ m^{-2}$

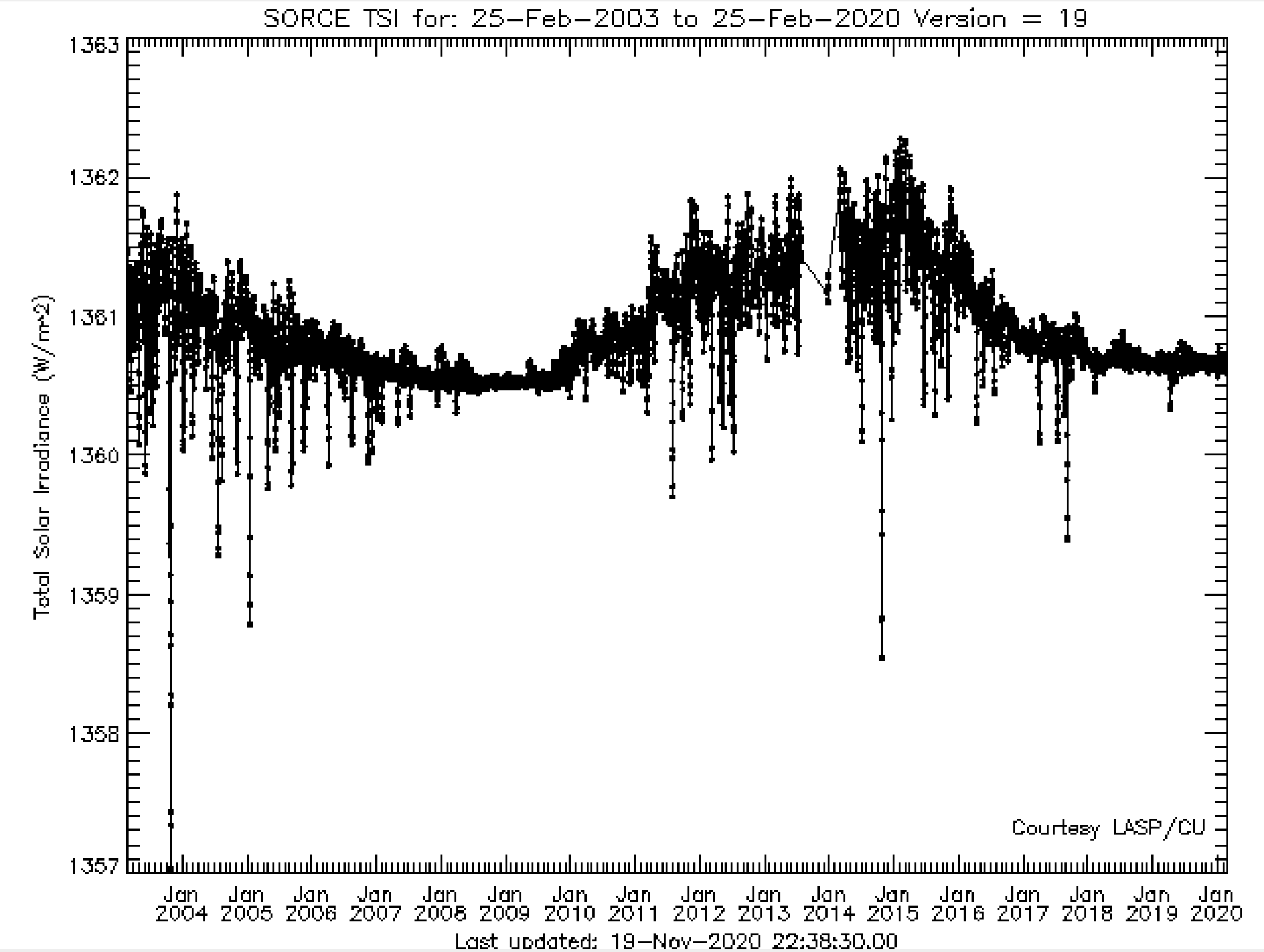
Modulation of the solar constant over satellite era
(Froelich, 2009)





The Earth receives the solar flux et radiates back the energy to space.

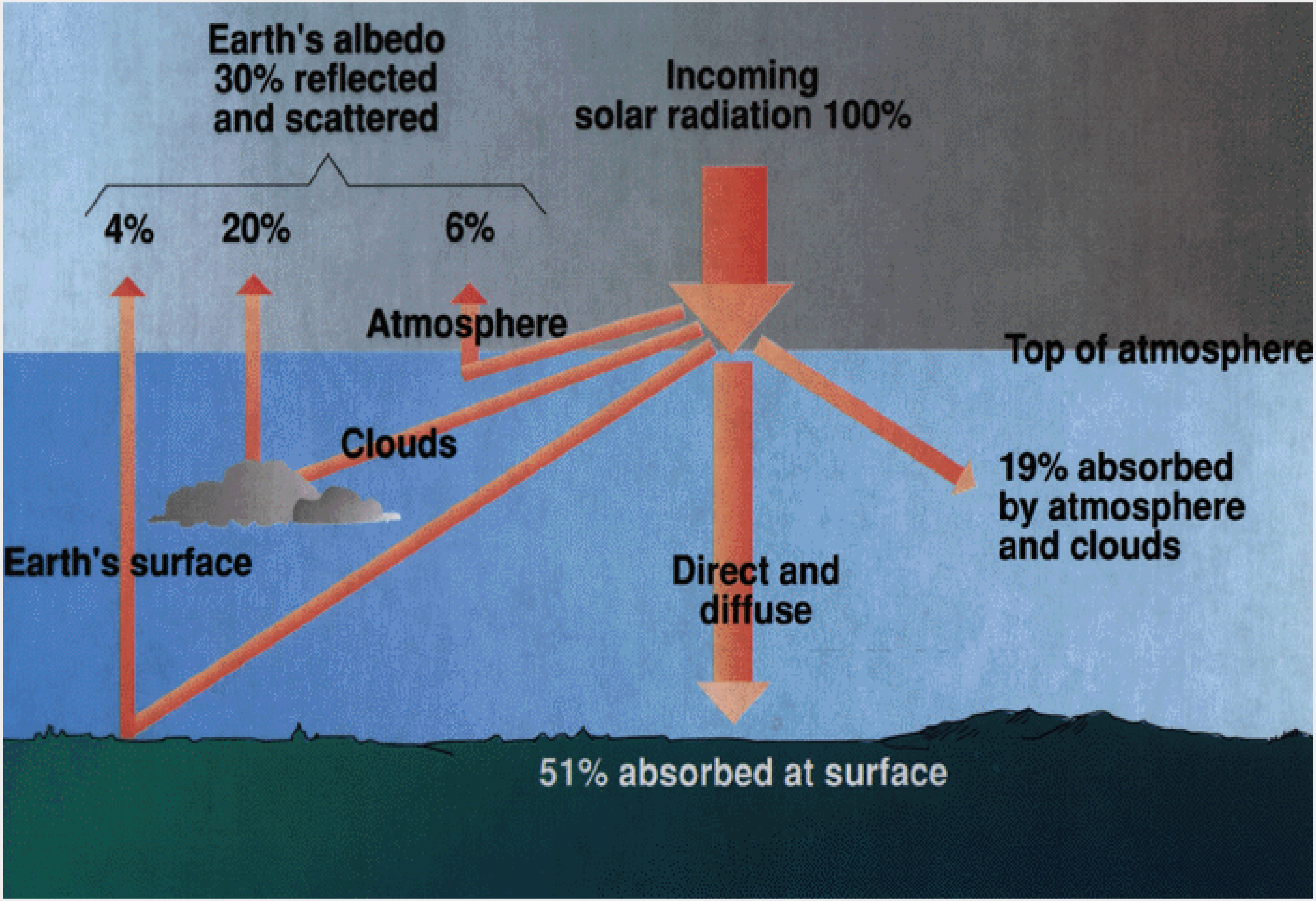




Solar constant during the mission SORCE

To retain: It is difficult to perform an absolute measurement of the solar constant. It has shown no trend over the satellite era. Historical variations over the last 2500 years are much less than the effect of a CO_2 doubling that is equivalent to + 16 W/m2.

I.2 Budget of the solar incoming radiation



I Introduction

II Interaction of radiation with matter

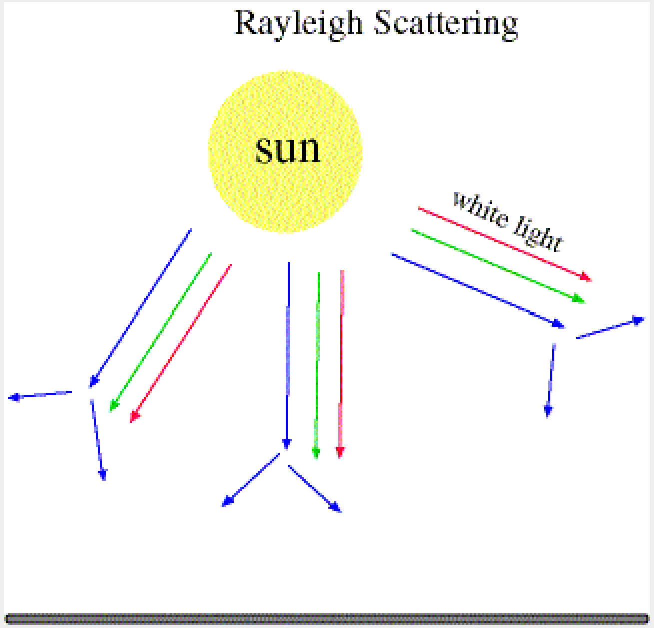
III Radiative budget of the Earth

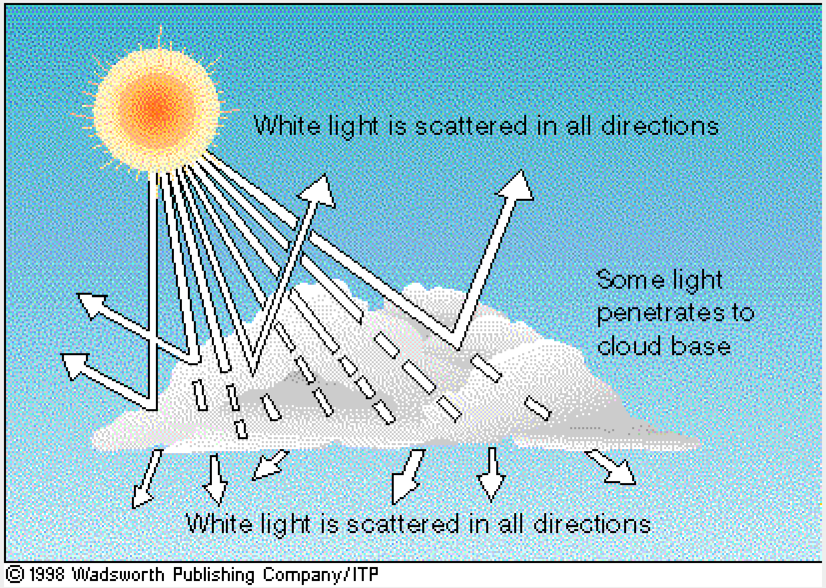
IV Greenhouse effect

V Climatology of the radiative budget

VI Climate sensitivity

Atmospheric diffusion

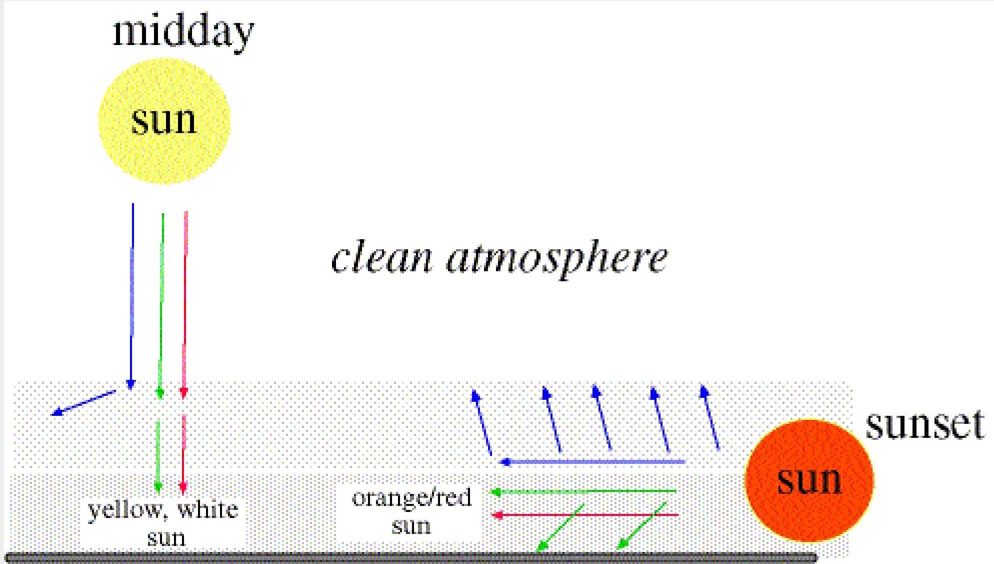


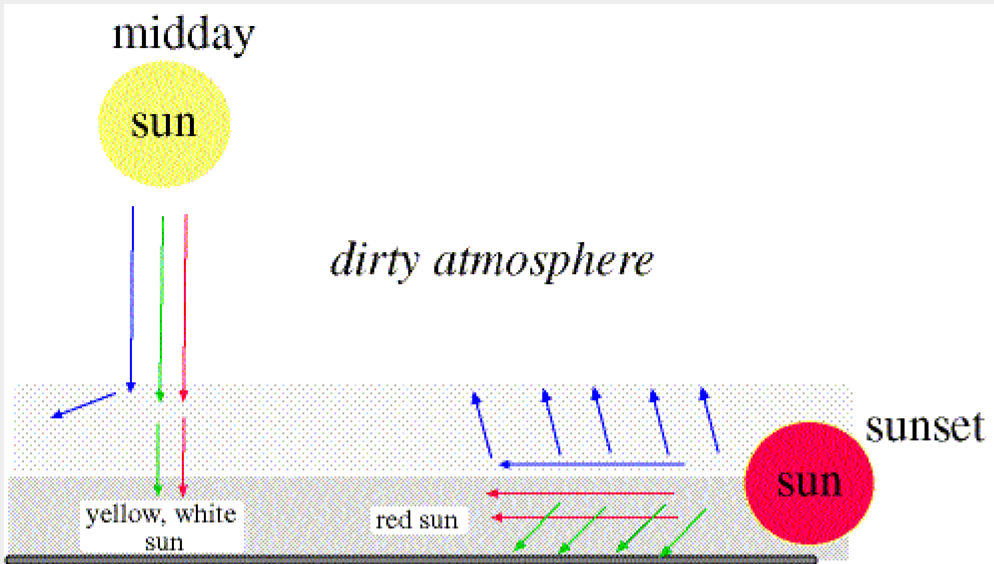


Rayleigh diffusion for the small parcels (gas molecules) such that a << λ.
The diffused power varies as $\sim 1/\lambda^4$

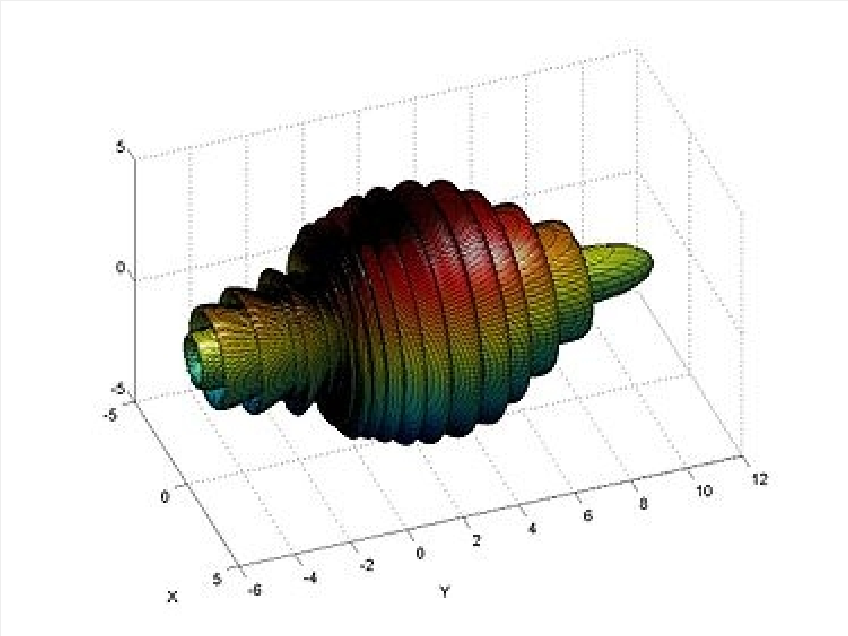
Lorenz-Mie diffusion for large parcels (droplets) such that a > λ.
The diffused power is independent of λ.
(a: size of diffusing parcels)

Sky and cloud colour

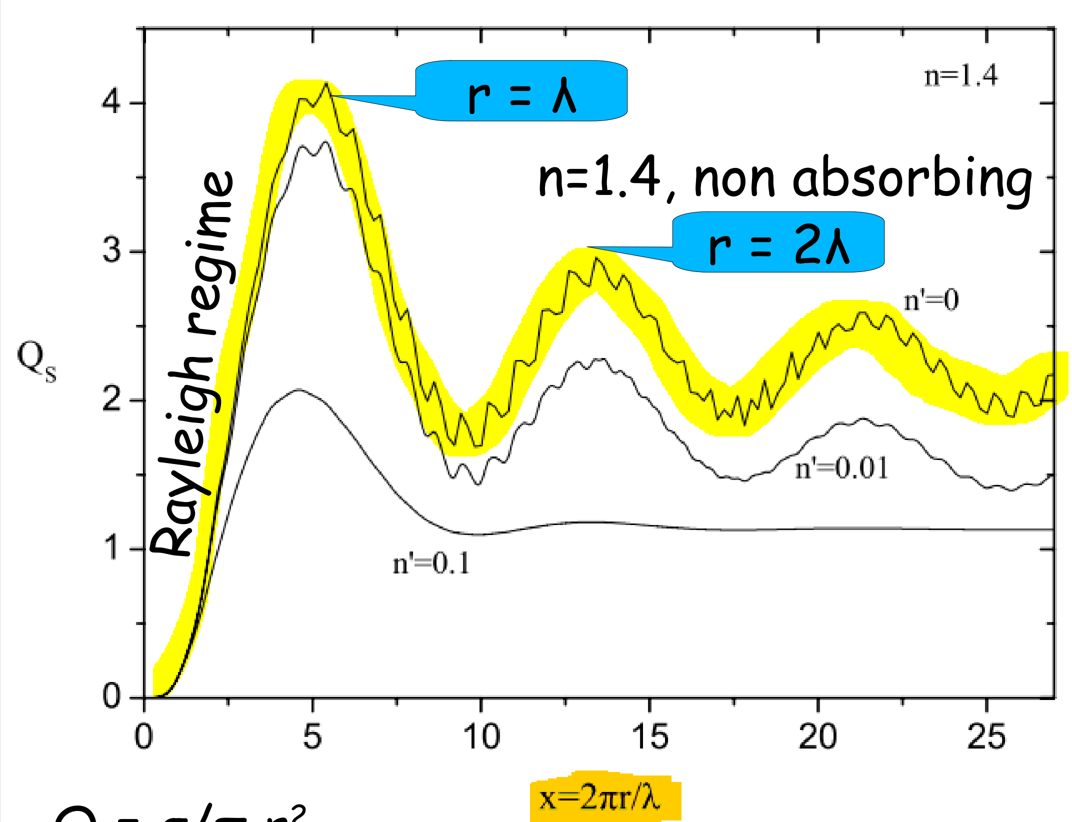




MIE DIFFUSION



Mie diffusion diagram at λ =0.633 μm for a beam coming from left and a 2 μm radius spherical particle.



$Q = \sigma/\pi r^2$

The Mie diffusion for a sphere is modulated by the parameter $x = 2\pi r/\lambda$ (r radius).

→ uniform optical regime at large x

For absorbing parcels (imaginary refraction index ≠ 0), the oscillation are damped.

Basic physical principles

The interaction of radiation with matter arises from the nature of matter composed by charged particles in motion.

The classical theory of electromagnetism says that any charged moving parcel emits or absorbs radiation. The quantum theory of electromagnetism says that atoms and molecules emits and absorbs during transitions during states.

Thermal agitation due to the motion of molecules and their internal vibrations (counted within C_p) is also coupled with radiation.

BLACK BODY LAW

The black body law has been established by Planck from the observation of the radiative properties of matter at high temperature. The derivation of this law has needed the revolutionary hypothesis that matter can only absorb or emit radiation at frequency ν by quantas $h\nu$.

For Planck, this hypothesis was at first only an unexplained calculation artefact. It is only after Einstein demonstrated in 1905 the corpuscular nature of light in his paper on the photo-electric effect that it was realized the deep physical meaning of this hypothesis which appeared afterwards as the foundation of quantum physics.

In the Planck theory, the perfect black-body is able to absorb entirely any incoming radiation whatever its frequency.

Quantum theory says that the energies of states of matter are quantified and that the absorbed or emitted photons are associated with transitions between these states. Real body is absorbing and emitting over a given number (which can be very large) of frequencies in the spectral domain. When a large number of atoms are coupled, the peripherical electronic layers degenerate into of continuum of states, generating also continuum of transitions within some energy intervals. In a gas, the rays are widened by the motion of molecules and the ensuing collisions or by Doppler effect. This allows to fill the spectral domain and makes the perfect body law highly relevant for real matter.

The Kirchhoff law says that a real body is able to emit radiation in the same proportion it is able to absorb it.

I Black-body law at temperature T

a) Planck law

Black-body monochromatic radiance per unit wavenumber, unit solid angle unit and unit surface emissivity

$$B_\lambda(T)\,d\lambda = \frac{2hc^2}{\lambda^5\left[\exp\frac{ch}{k\lambda T}-1\right]}\,d\lambda$$

$h = 6{,}63\,10^{-34}\ J\,s$ *Planck constant*

$k = 1{,}38\,10^{-23}\ J\,K^{-1}$ *Boltzmann constant*

b) Stefan law

$$B(T) = \int_0^\infty B_\lambda(T)\,d\lambda = \frac{\sigma}{\pi}T^4$$

$\sigma = 5{,}670\,10^{-8}\ W\,m^{-2}$

By integrating over a half-shere of solid angle

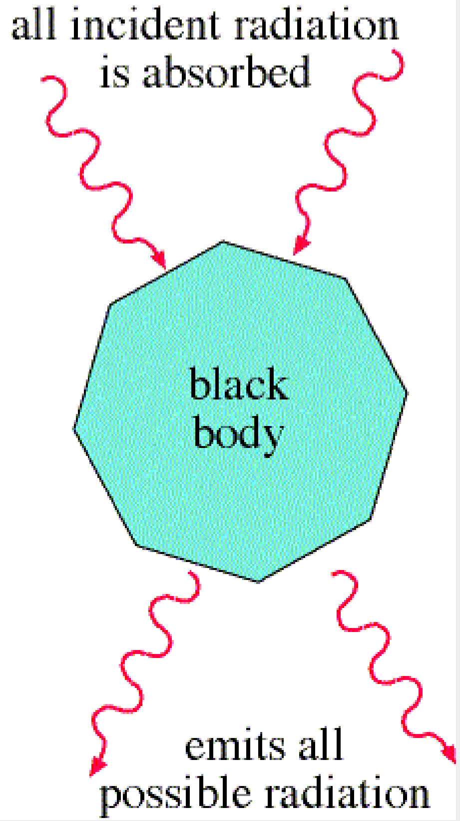
$$\iint B(T)\cos\theta\,d\omega =$$

$$\iint B(T)\sin\theta\cos\theta\,d\theta\,d\phi =$$

$$\pi B(T) = \sigma T^4$$

The black body law is valid for a system under thermodynamic equilibrium:

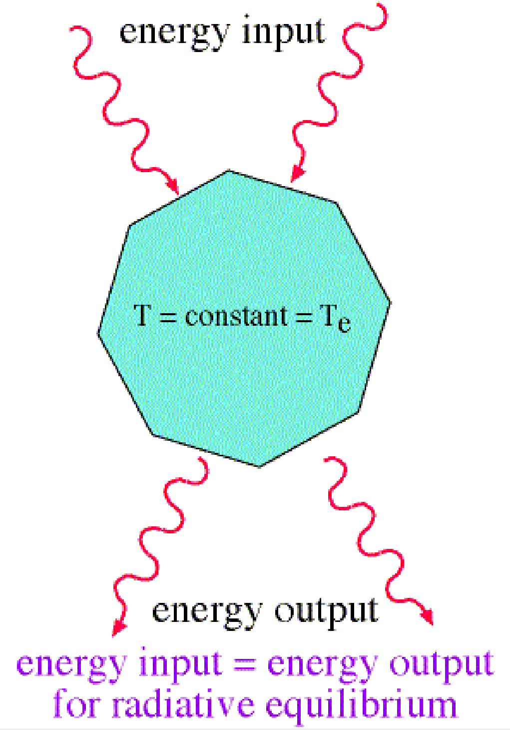
- In statistical physics, the macro-state that maximizes the number of micro-states (entropy)
- A macroscopic state that, under fixed external constraints, does not evolve spontaneously.
- Counter-exemple: laser



c) Wien law

$$\lambda_{max} T = A = 2898\ \mu m\ K$$

at Earth surface	$T = 288\ K$	$\lambda_{max} = 9{,}9\ \mu m$
at the top of the atmosphere	$T = 255\ K$	$\lambda_{max} = 11{,}3\ \mu m$
at the surface of the Sun	$T = 6110\ K$	$\lambda_{max} = 0{,}47\ \mu m$



II Kirchhoff law for the real systems

a_λ *absorption, characteristic property of the material*

For any system at thermal equilibrium, the emission is

$$I_\lambda(T) = a_\lambda B_\lambda(T)$$

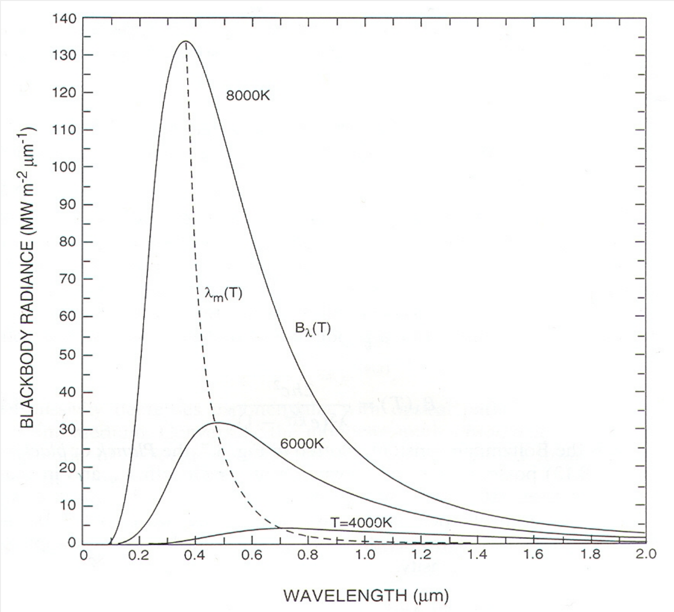
For a black body $a_\lambda = 1$ *et* $I_\lambda(T) = B_\lambda(T)$

For a real system $a_\lambda < 1$ *et* $I_\lambda(T) < B_\lambda(T)$

Any selective absorbant is a selective emettor

Absorptivity and emissivity are the same coefficient.

The emission at a given temperature is determined jointly by the properties of the system described by a_λ ***and the temperature which determines*** $B_\lambda(T)$



Contributions of the molecular vibration modes to the electromagnetic spectrum

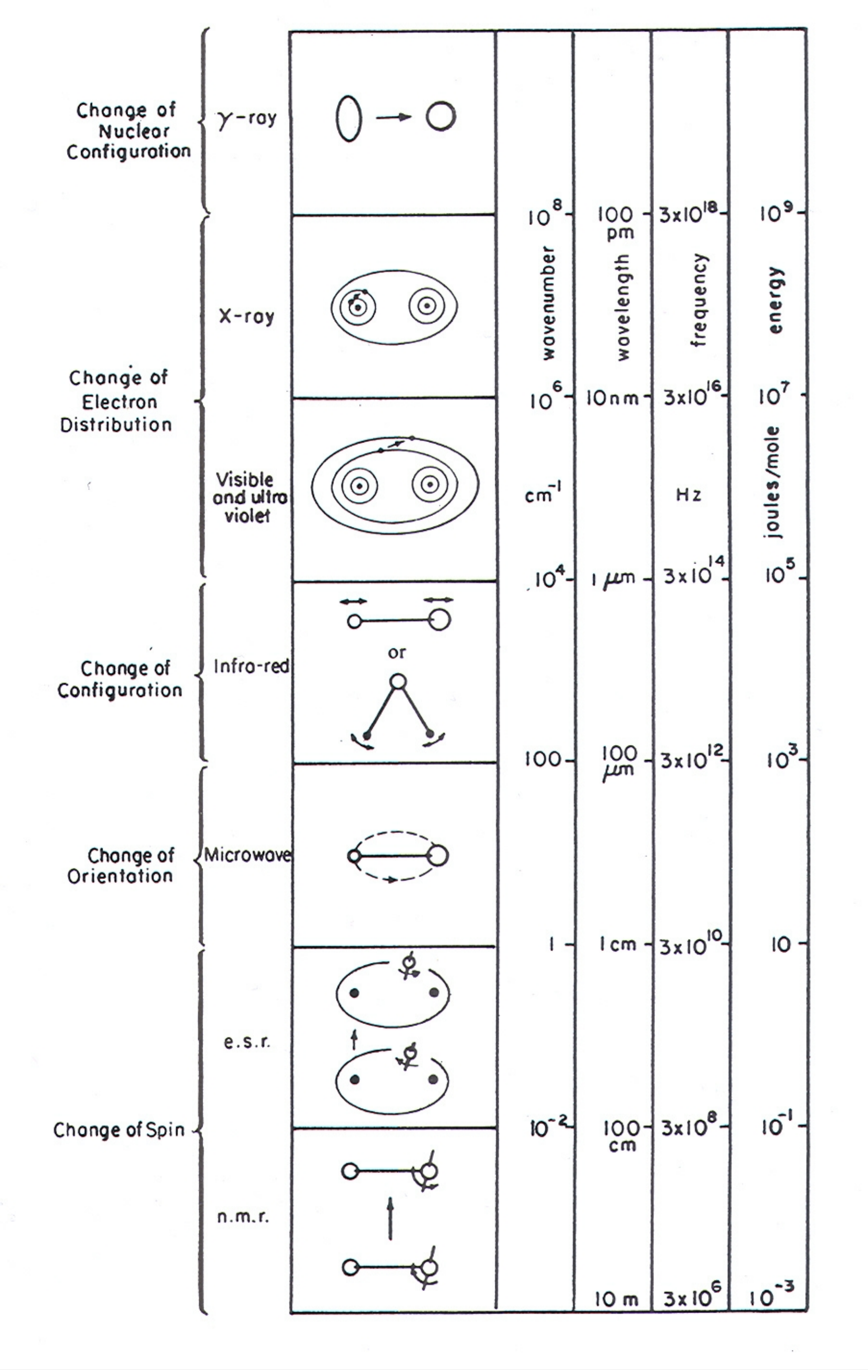
Photodissociation
Ionisation
Electronic transitions

Vibration and rotation
$\lambda > 0.7\ \mu m$

Rotation for

$\lambda > 20 \mu m$

Spin inversion



In the UV to micro-wave domain

The smallest wavenumbers (high frequency) interact with the lighest particles (electrons).

The lowest frequencies interact with the molecular structure (vibration, rotation).

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II Interactions of rayonnement with matter

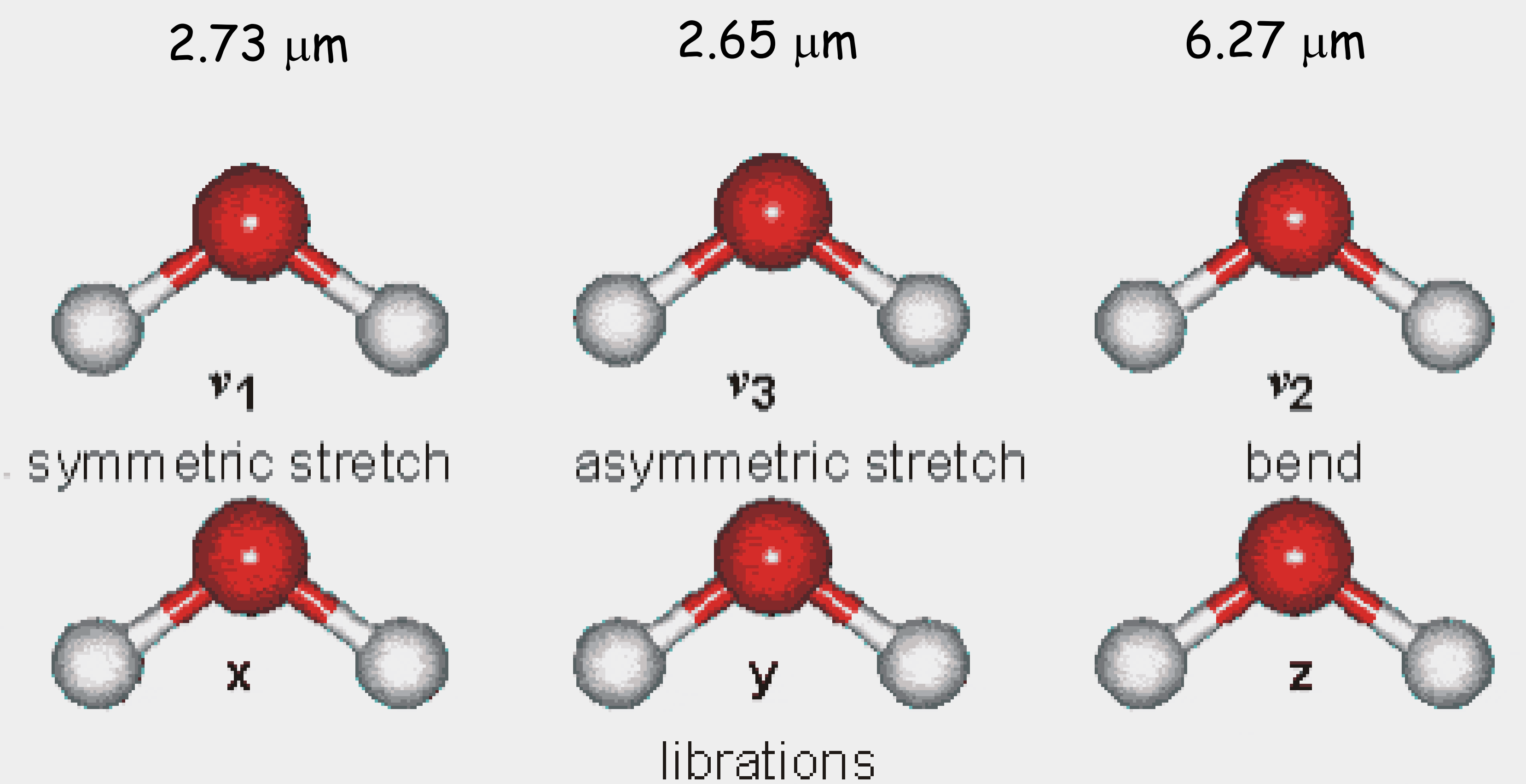
III IR gas absorption

IV Greenhouse effect

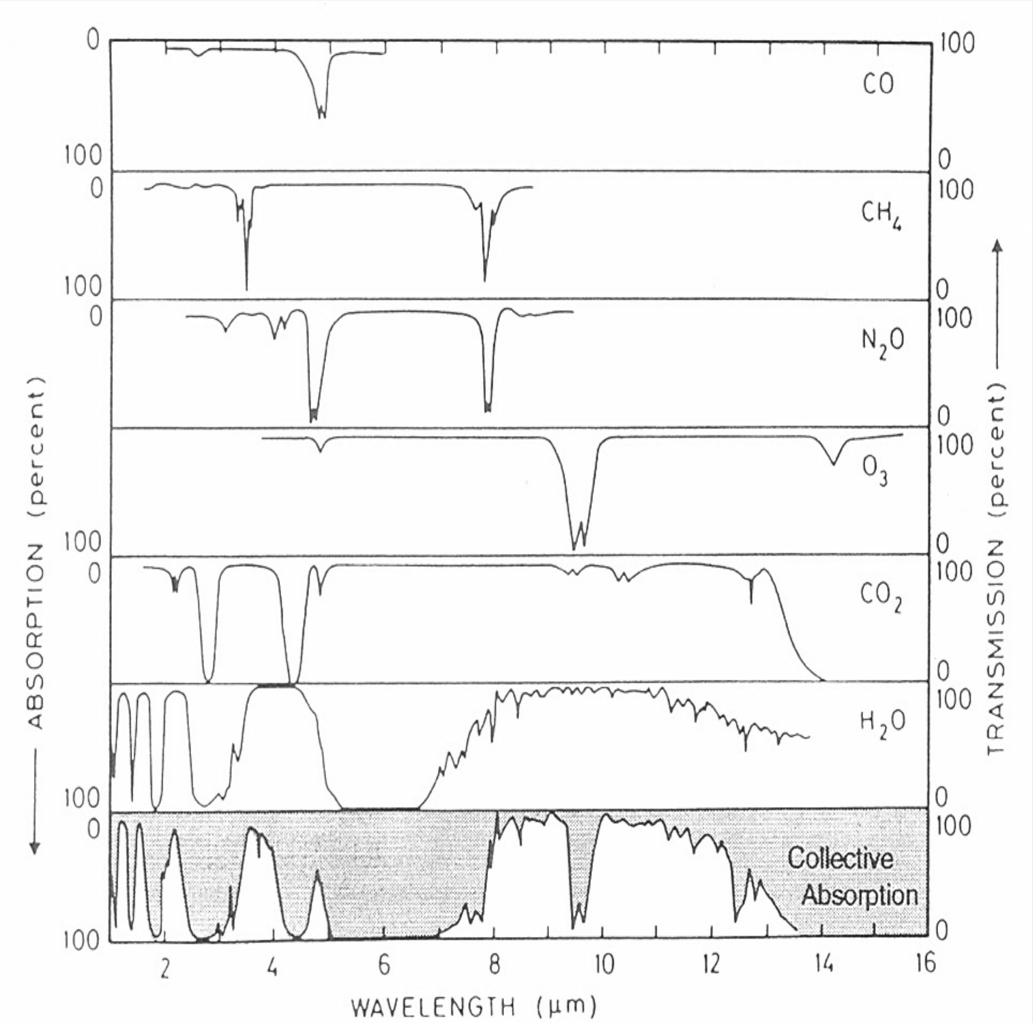
V Climatology of the radiative budget

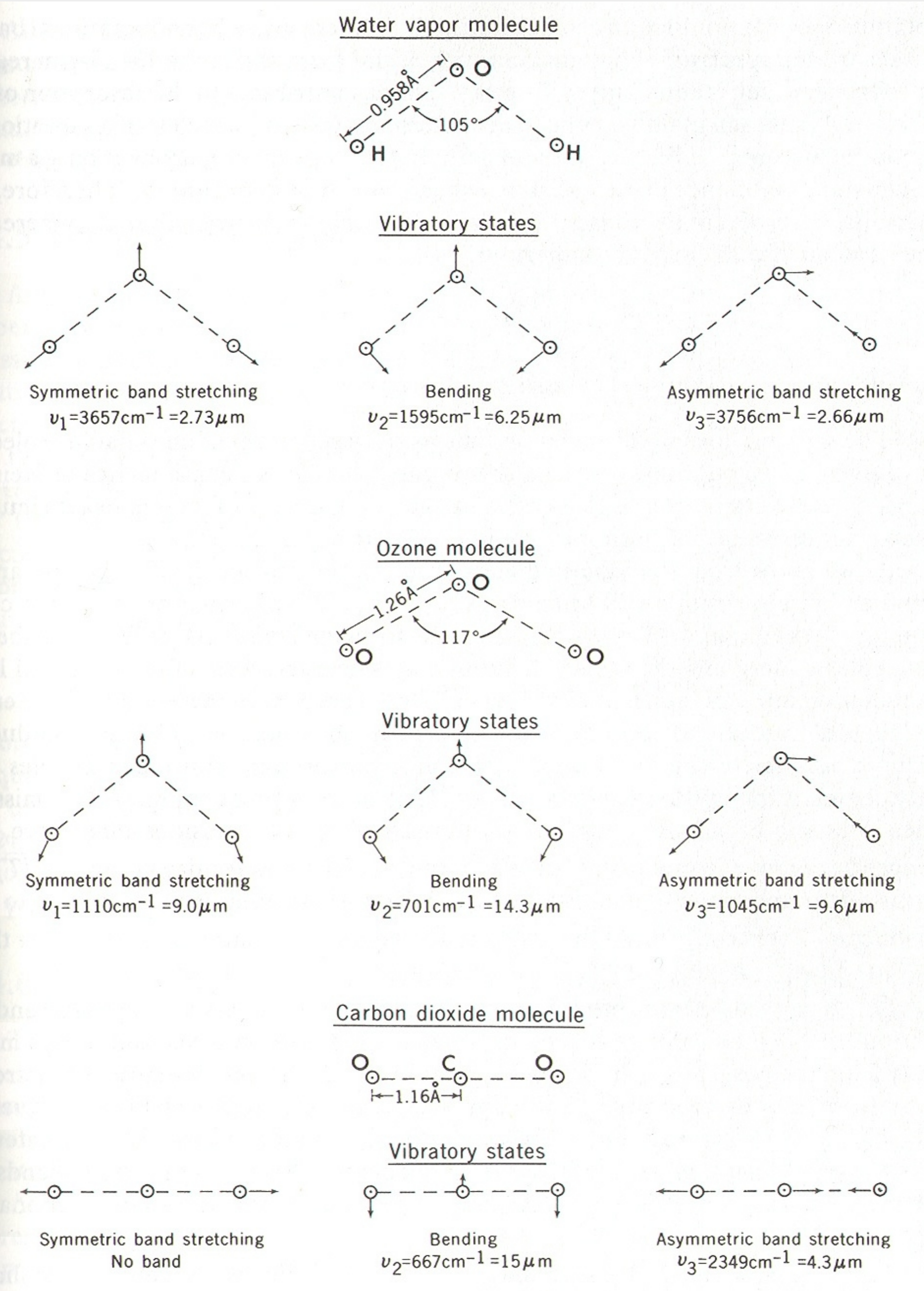
VI Climate sensitivity

Vibration-rotation modes of the H_2O molecule



Leads to numerous transition and absorption lines in the spectrum.



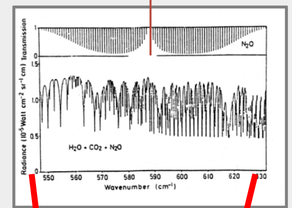


Absorption by the atmospheric molecules

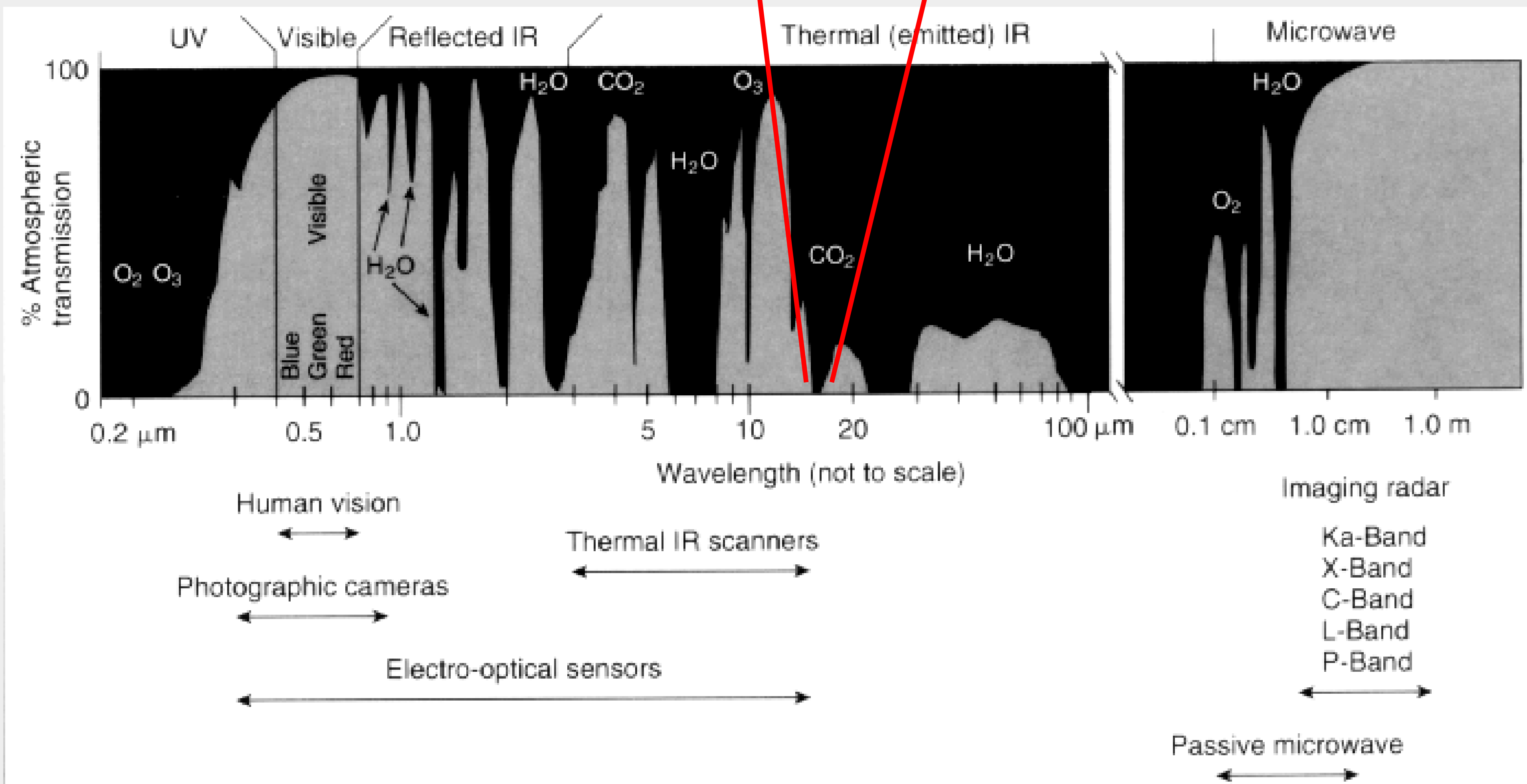
Transmission = 1 - Absorption

In the IR:
Gas diffusion can be neglected

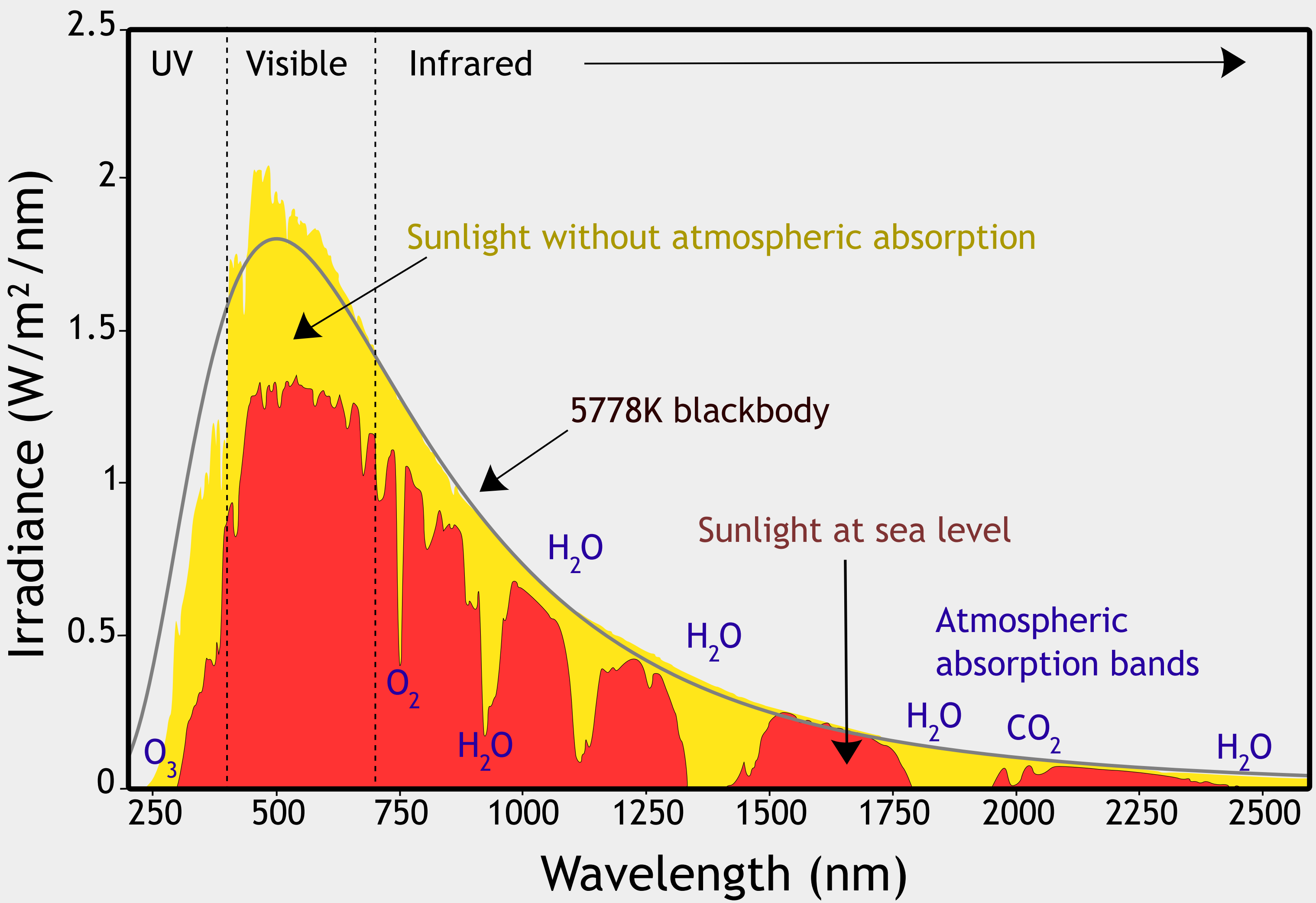
17μm



The absorption bands are composed of a multitude of individual rays. .

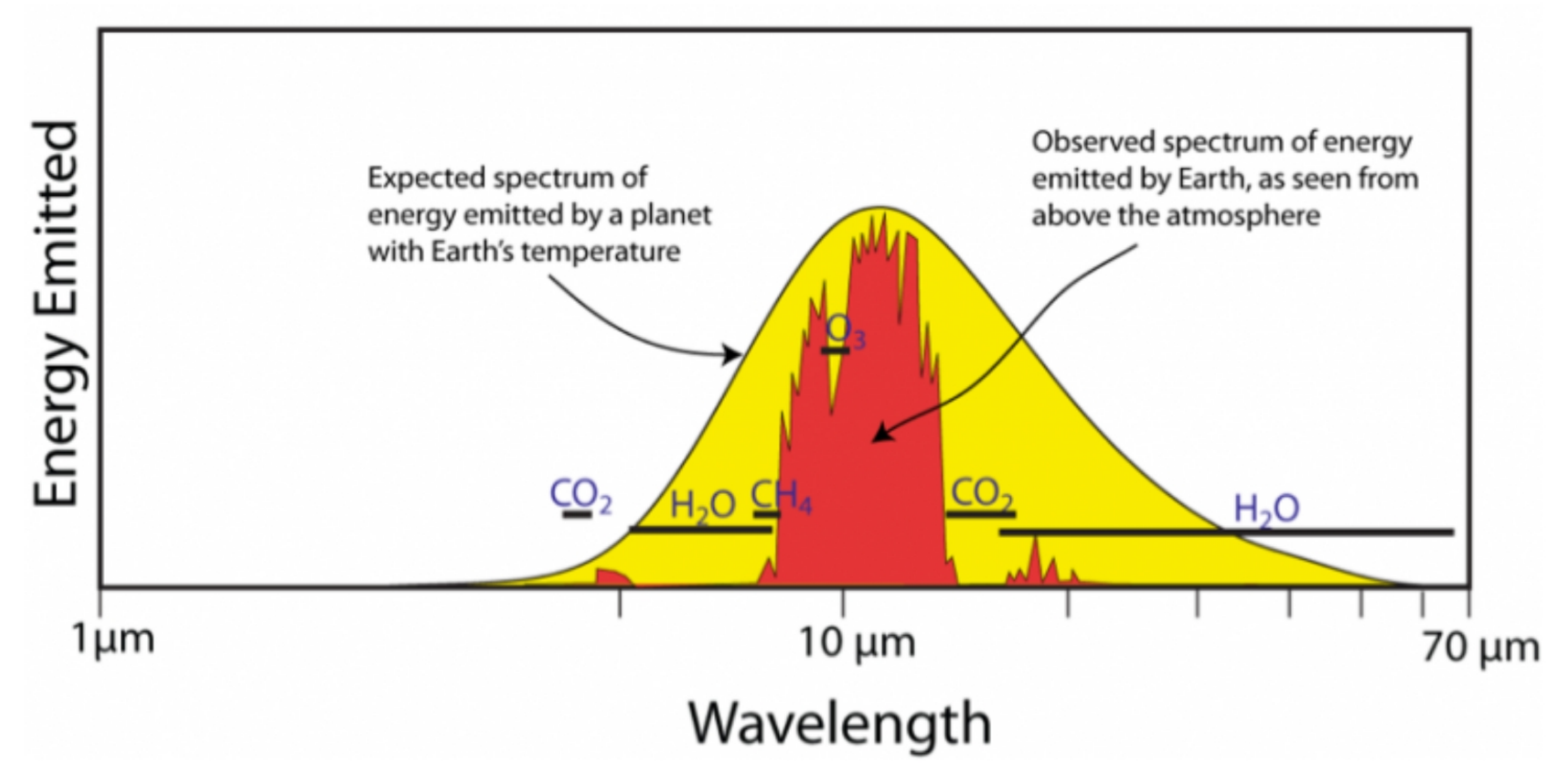


Spectrum of Solar Radiation (Earth)



Total absorption of the non reflected flux : ~ 1/3

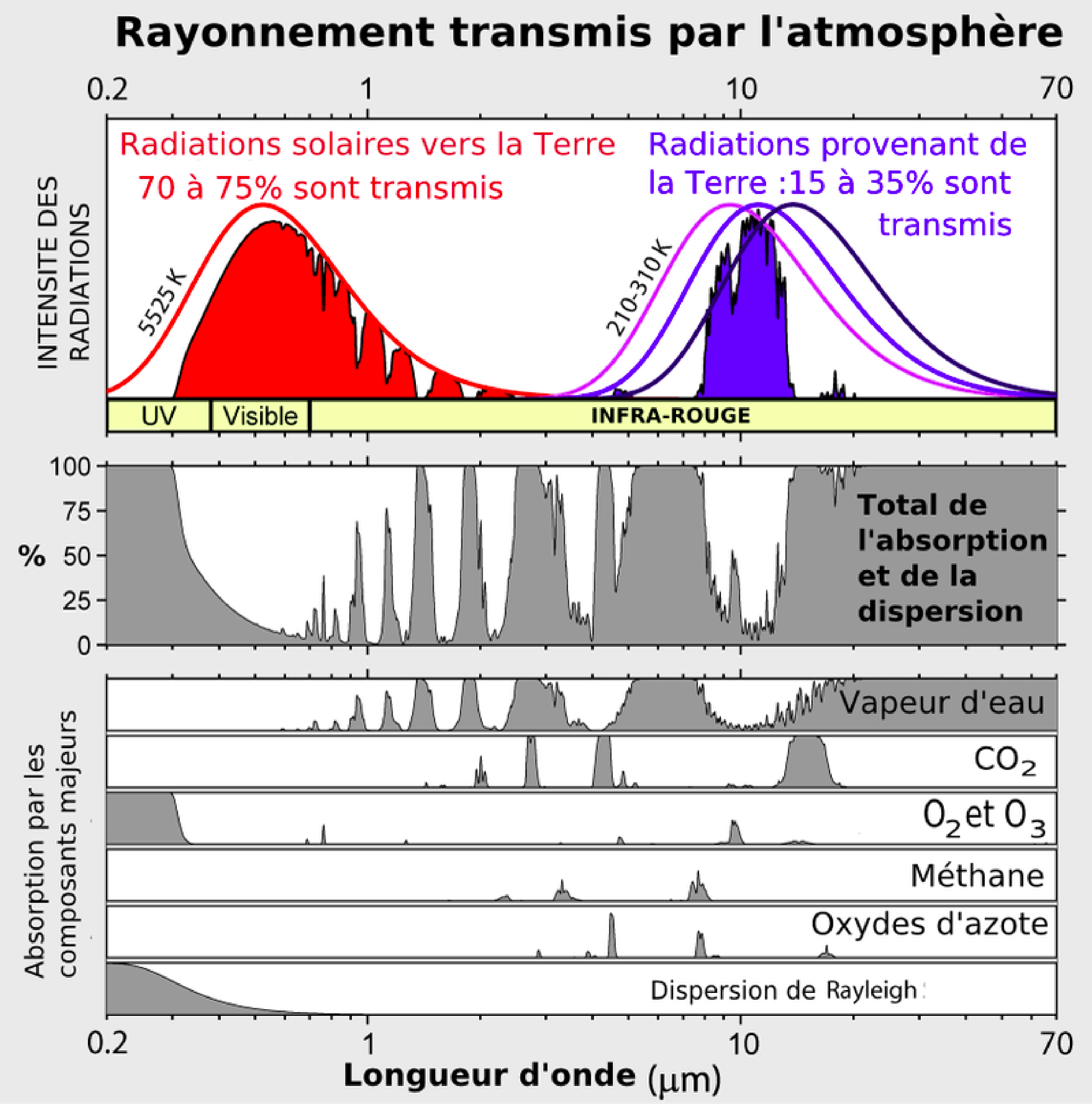
(gas and clouds: 79 W m^{-2} from 240 W m^{-2})

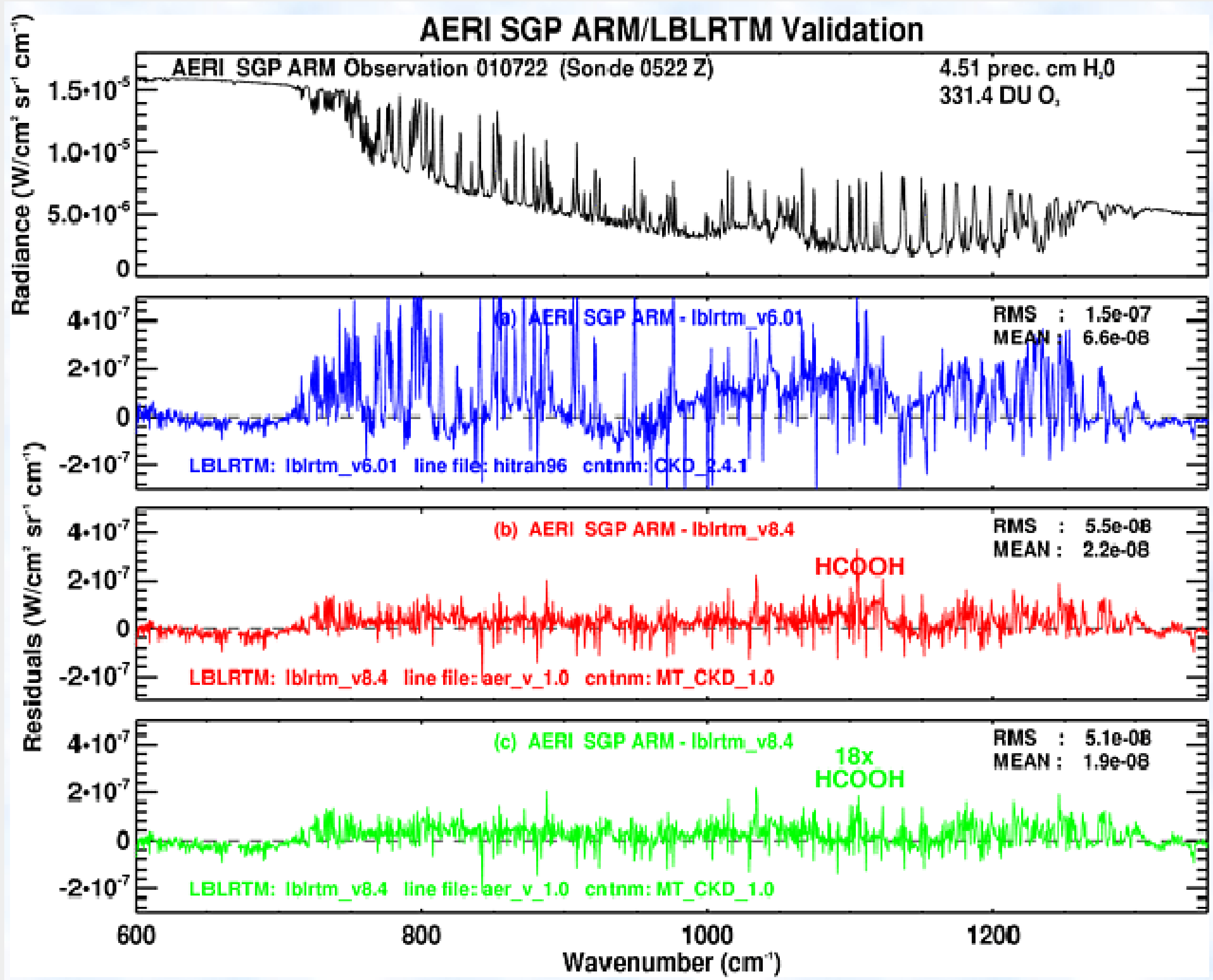


Energy Emitted
Expected spectrum of
energy emitted by a planet
with Earth's temperature
Observed spectrum of energy
emitted by Earth, as seen from
above the atmosphere
O_3
CO_2
H_2O
CH_4
CO_2
H_2O
1 µm
10 µm
70 µm
Wavelength

Absorption of the atmosphere
and comparison of the terrestrial and solar black-body spectra

- disjoint domains
- The atmosphere is transparent for a large part of the visible radiations, window around 10 μm in the IR in the absence of clouds





Residue with HITRAN96

Residue with HITRAN2000 + modelisation of the continuum

Comparison of a line by line radiative model with spectroscpic measurements of the descending thermal flux. Clear sky case of the ARM Great Plains station (Utah).

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Without green house effect

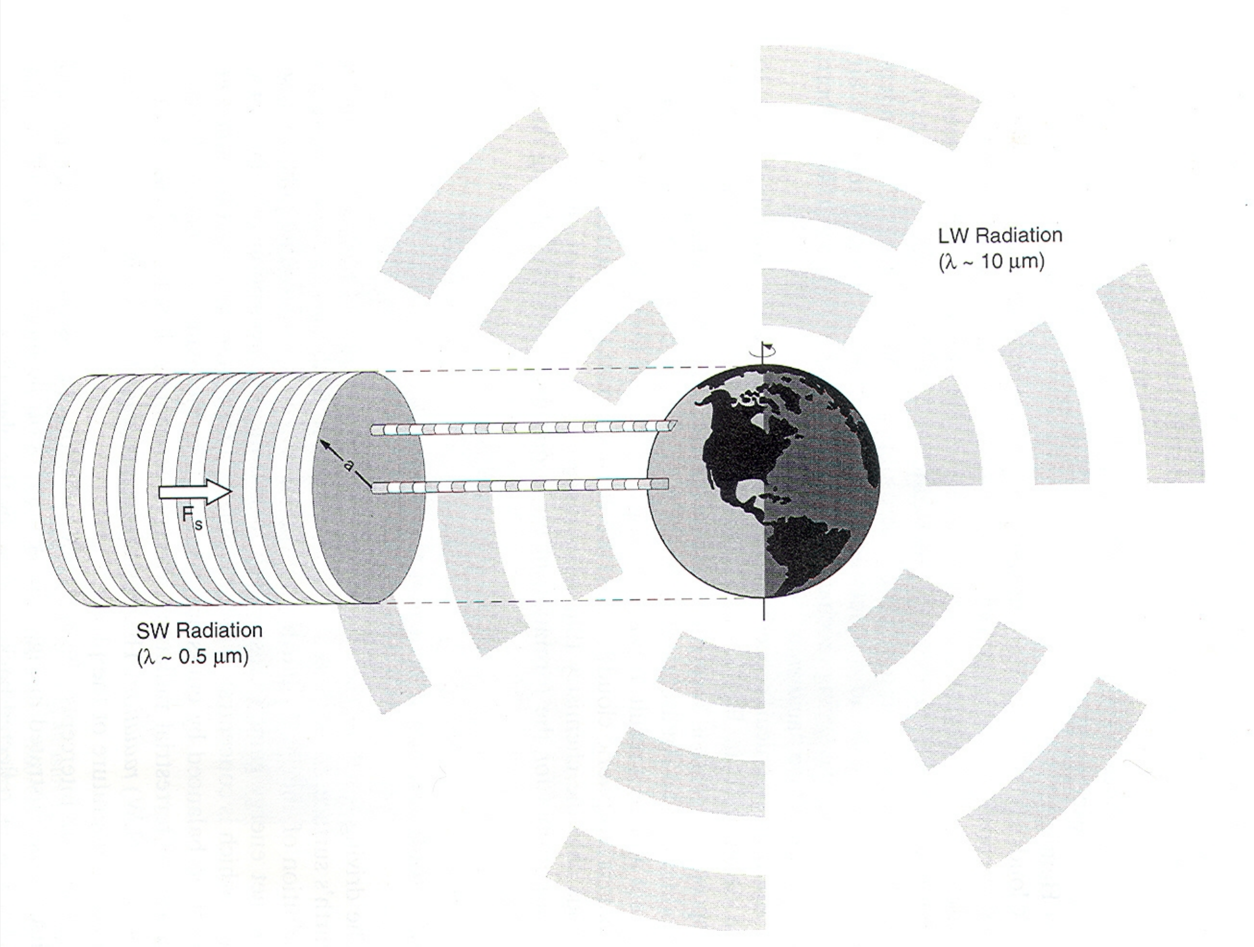
Albedo $\alpha = 0{,}3$

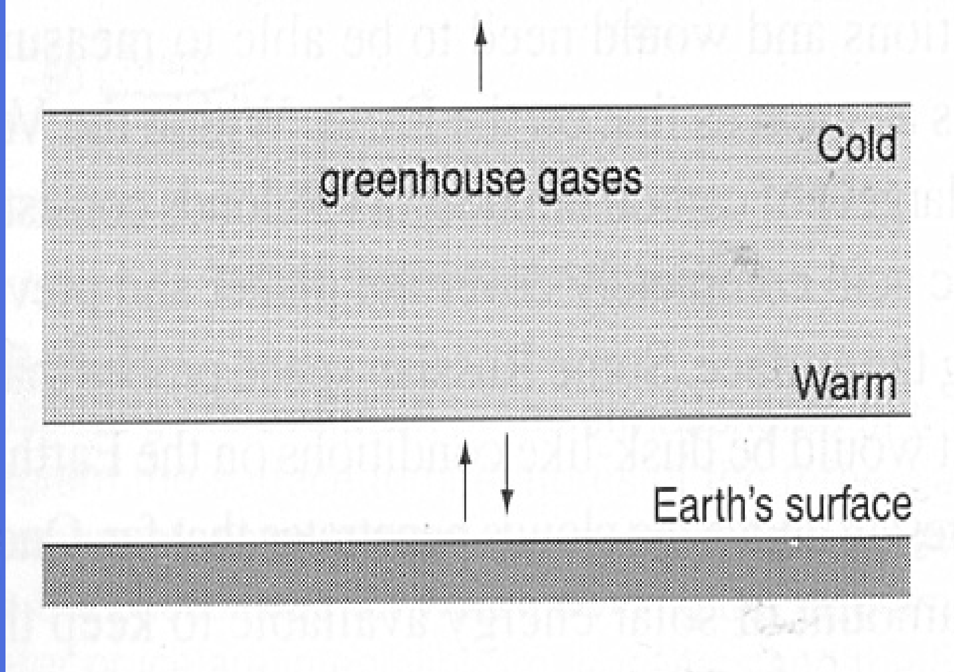
$$\frac{1}{4} S_0 (1-\alpha) = \sigma T_e^4$$

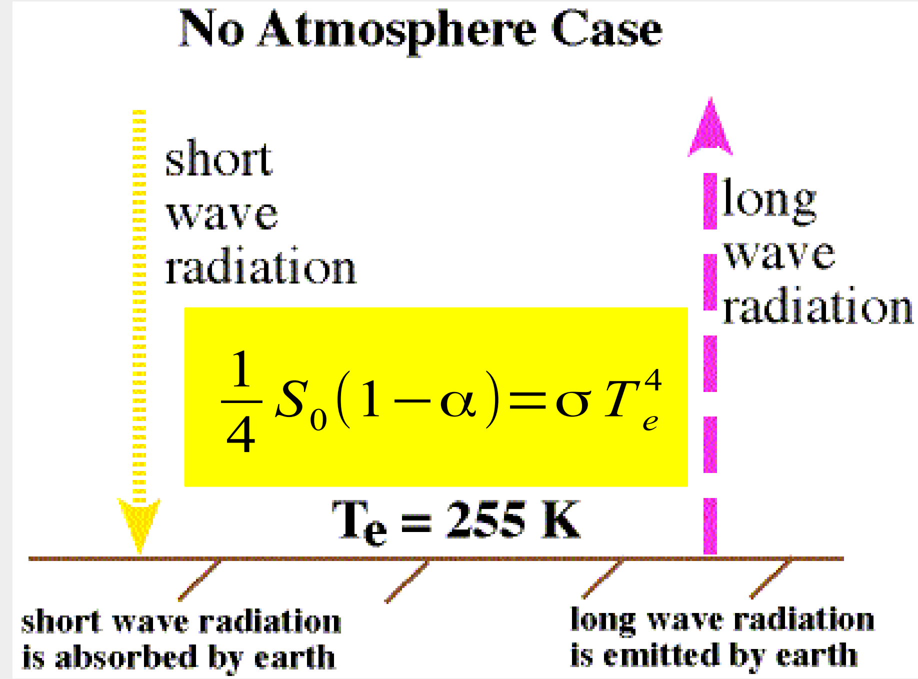
Surface temperature T_e = 255 K

- Incoming solar radiation S_0 = 1367 W m^{-2}

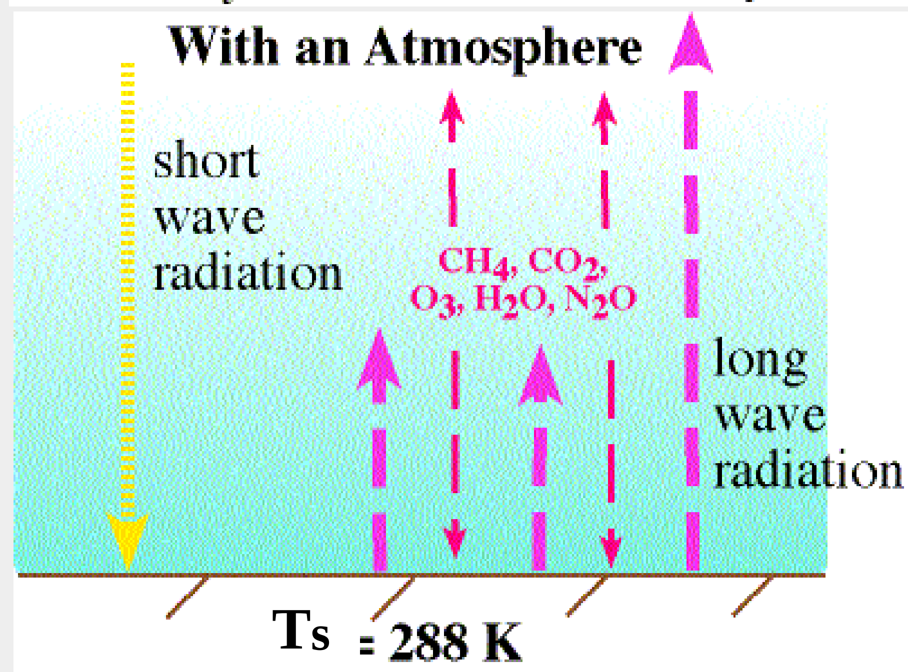
We assume implicitly here the existence of processes that redistribute the heat at the surface.

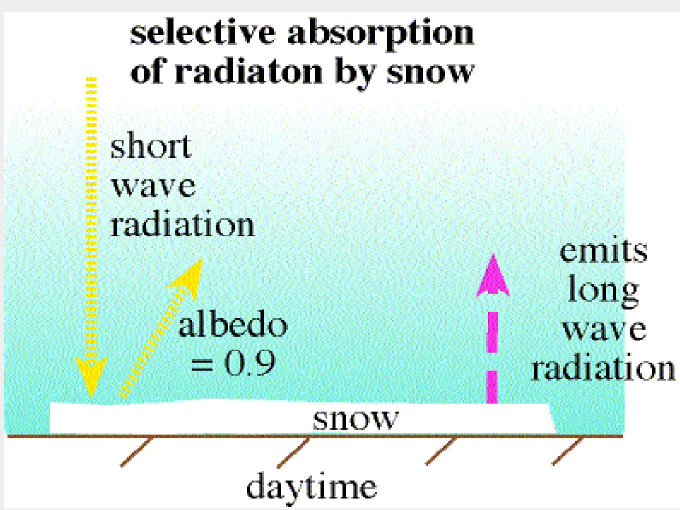


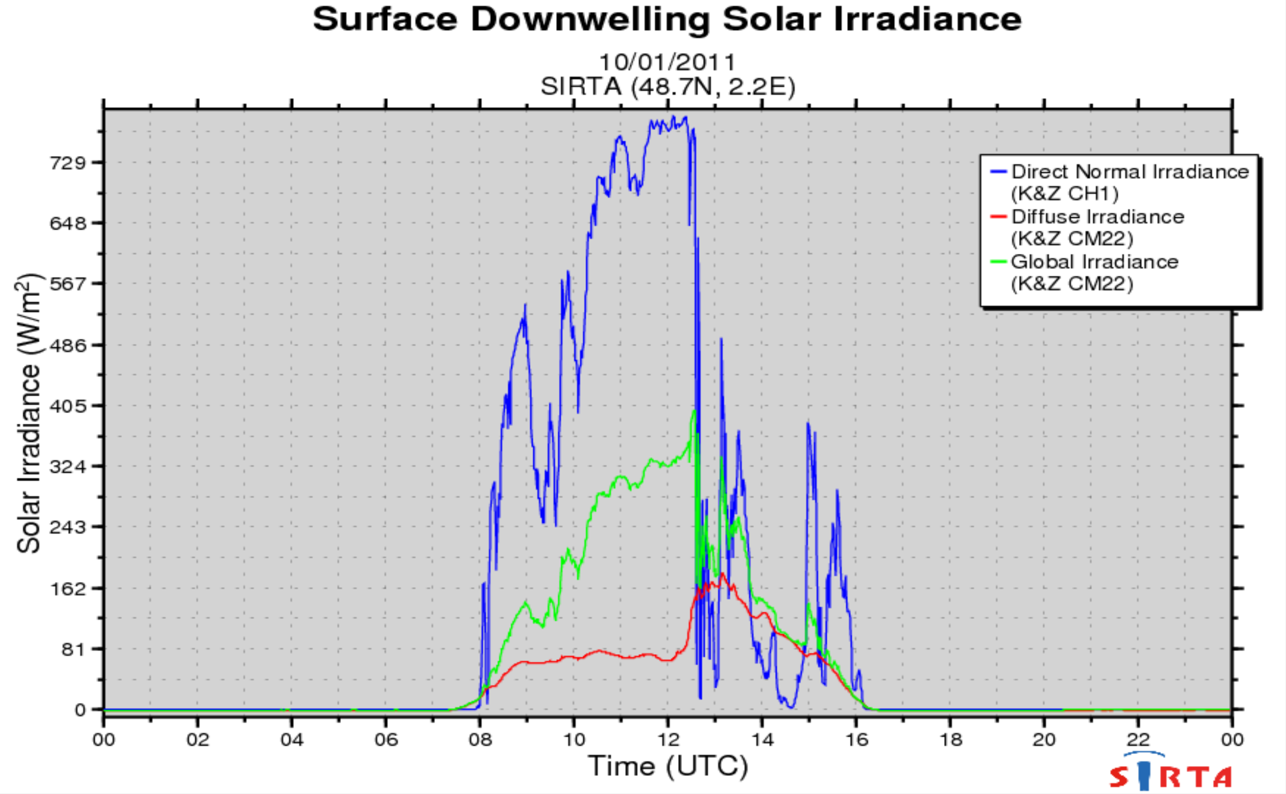




Green house effect

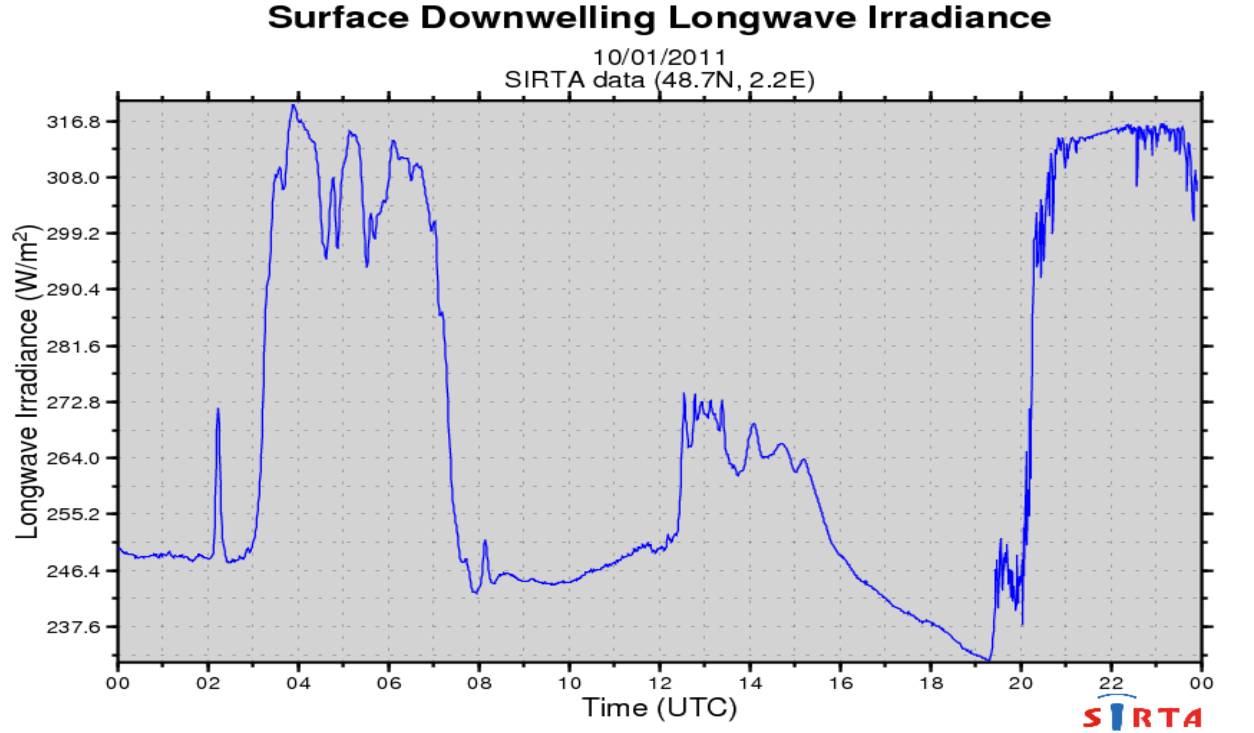




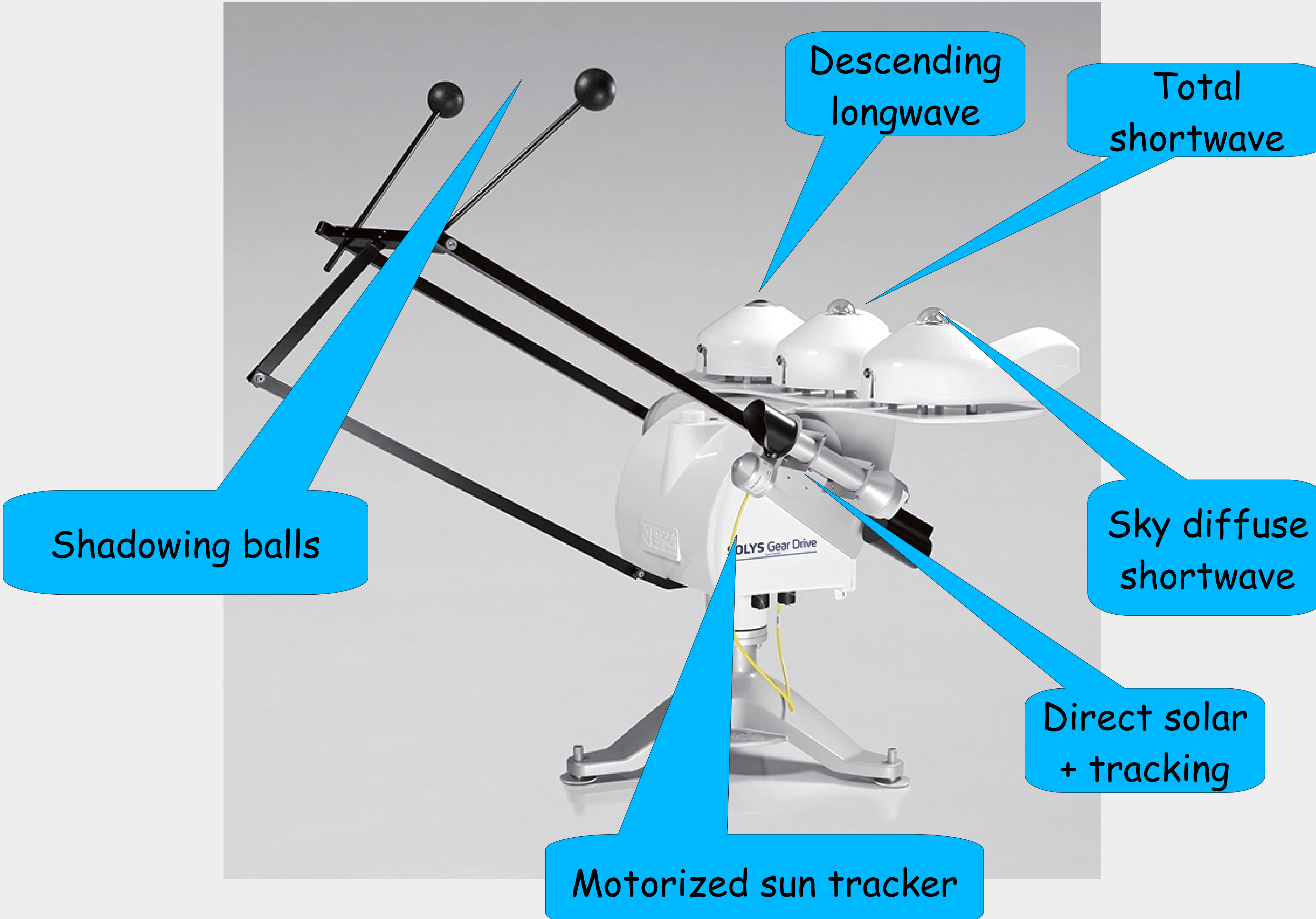


Solar irradiance as measured at IPSL SIRTA observatory

Descending infra-red flux as measured at SIRTA

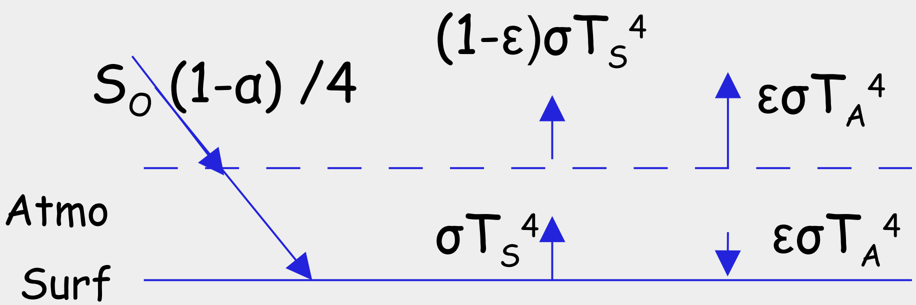


A composite instrument measuring all components of downward radiative flux at the surface



First model: Greenhouse effect for a one layer isothermal atmosphere

Let us take al atmosphere which is transparent to the solar incoming radiation and which behaves as a grey isothermal body for the long waves



ε absorption - emissivity of the atmosphere

T_A temperature of the atmosphere

T_S surface temperature

Budget at the top of the atmosphere

$$\frac{S_0}{4}(1-\alpha)-(1-\epsilon)\sigma T_S^4-\epsilon\sigma T_A^4=0$$

Budget of the atmosphere

$$\epsilon\sigma T_S^4-2\epsilon\sigma T_A^4=0$$

Surface budget

$$\frac{S_0}{4}(1-\alpha)+\epsilon\sigma T_A^4-\sigma T_S^4=0$$

$$T_S=\left(\frac{S_0}{2\sigma}\frac{(1-\alpha)}{(2-\epsilon)}\right)^{1/4}=T_e\left(\frac{2}{2-\epsilon}\right)^{1/4}$$

For the Earth: T_e=255°K, Ts=303°K =+30°C,

The difference is due to the "geeenhouse" effect : absorption of the emitted long-wave radiation by the atmosphere and re-emission towards the surface which is heated by the atmosphere.

Sensitivity of climate and retroactions

Simple case of a one-layer atmosphere $\sigma T_S^4 = \frac{2\phi}{2-\epsilon}$

with absorption $\epsilon = \epsilon_{CO2} + \epsilon_{H2O}$ and $\phi = 241$ W m^{-2}, the incoming solar flux. ϵ_{CO2} is fixed and depends of the emissions while ϵ_{H2O} dépends on the temperature. We write $\delta\epsilon = \delta\epsilon_{CO2} + \frac{d\epsilon_{H2O}}{dT_S}\delta T_S$

Hence

$$\frac{2\phi}{(2-\epsilon)^2}\left(\delta\epsilon_{CO2} + \frac{d\epsilon_{H2O}}{dT_S}\delta T_S\right) = 4\sigma T_S^3 \delta T_S$$

$$\frac{2\phi}{(2-\epsilon)^2}\delta\epsilon_{CO2} = \left(4\sigma T_S^3 - \frac{2\phi}{(2-\epsilon)^2}\frac{d\epsilon_{H2O}}{dT_S}\right)\delta T_S$$

Doing a more detailed and realistic calculation :

3.7 W/m^2 for a doubling of CO2

3.2 W/m^2/K

1.5 W/m^2/K with Γ and RH kept constant.

The sensitivity factor then decreases from 3.2 W/m^2/K to 1,7 W/m^2/K, and the resulting heating increases from 1,2 to 2,2 °C.

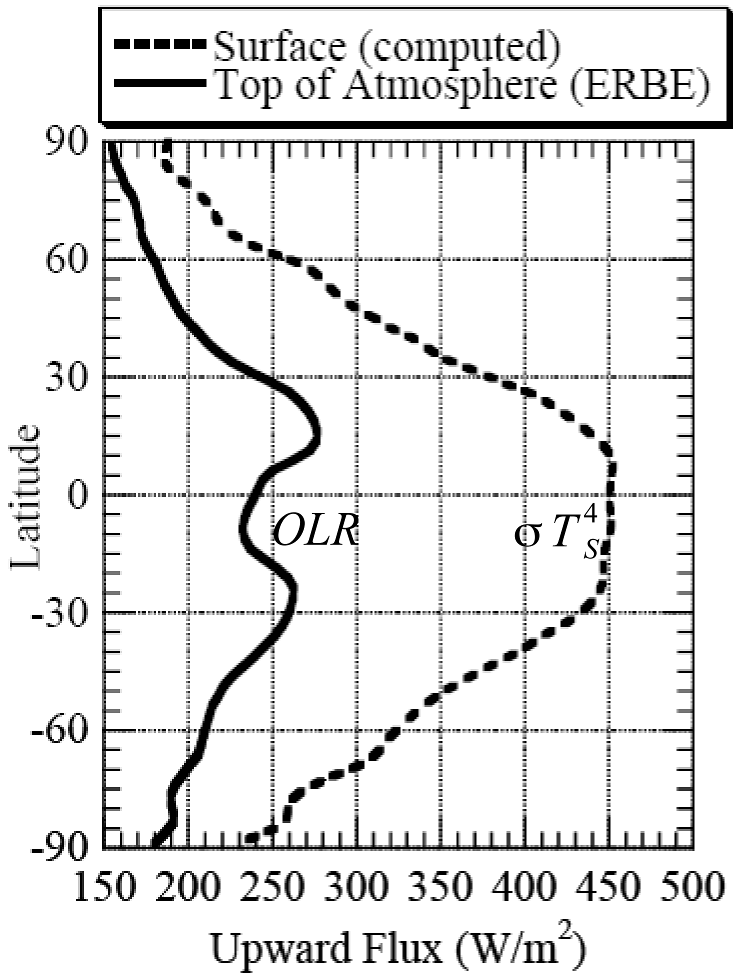


Figure 3.7: The Earth's observed zonal-mean OLR for January, 1986. The observations were taken by satellite instruments during the Earth Radiation Budget Experiment (ERBE), and are averaged along latitude circles. The figure also shows the radiation that would be emitted to space by the surface (σT_s^4) if the atmosphere were transparent to infrared radiation.

The outgoing long-wave radiation to space is reduced by the presence of the atmosphere

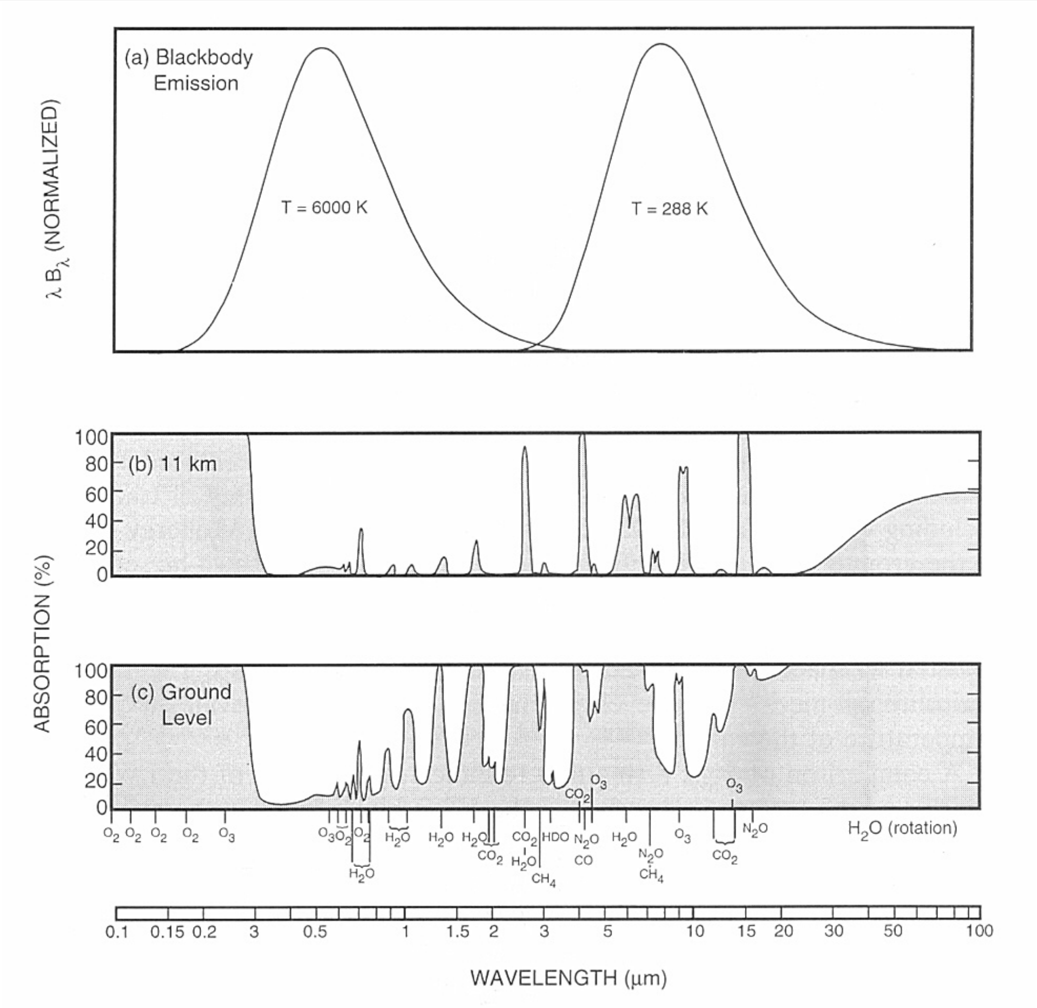
Greenhouse effect

$$G = \sigma T_S^4 - OLR$$

Perrehumbert, CUP, 2010

THE GRAY MODEL APPROXIMATION

The incoming visible spectral domain and the infrared emitted spectral domain are well separated with no overlap. The *gray model* is introduced as an approximation that ignores the detailed spectroscopic properties and consider only two blocks: the *shortwave* corresponding to the incoming radiation and the longwave corresponding to the thermal infrared.



The absorption/emissivity is then described by two parameters only a_{SW} and a_{LW}. In the same way, scattering or reflection, if accounted, has a single coefficient for shortwave.

We abandon here the one-layer simplification and consider here a still gray but startified atmosphere.

IR radiative transfer in a gray stratified atmosphere

Propagation along a given direction
Absorption small column of air

$$dI = -\kappa I \rho dz$$

where κ is the absorption coefficient per unit of mass and length.
Hence, between O and A, the flux weakens as

$$I = I_0 \exp\left(-\int_O^A \kappa \rho dz\right) = t I_0$$

where t is the transmission.
We define the *optical thickness* τ between O and A .

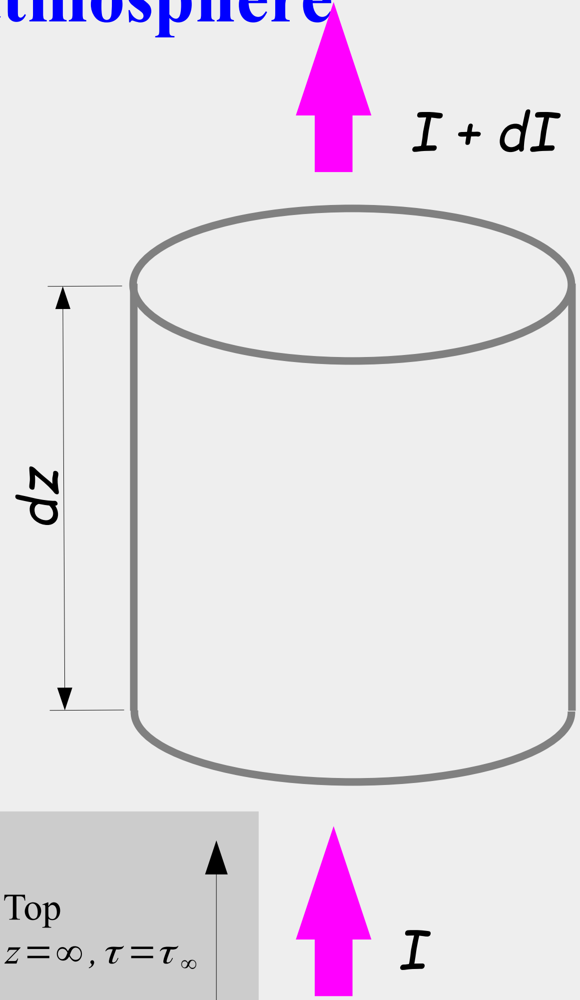
$$\tau = \int_O^A \kappa \rho dz \text{ hence } I = I_0 e^{-\tau}$$

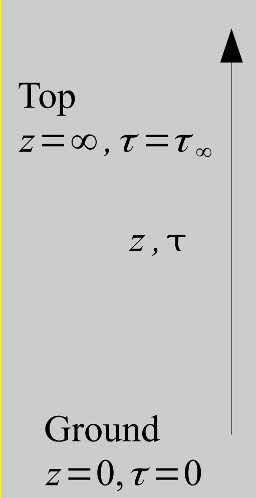
The emission is the product of the black body emissivity by the absorption of the column $E = \kappa \rho dz \times B(T)$ with $B = \pi^{-1} \sigma T^4$

The radiative transfert equation is thus

$$dI = -I \kappa \rho dz + B \kappa \rho dz$$

$$\frac{dI}{d\tau} = -I + B$$





RADIATIVE TRANSFER IN THE ATMOSPHERE

We consider now the ascending and descending fluxes $F^\wedge$ and $F^\wedge$

$$F^\wedge = \int_{\text{upper hemisphere}} I(\theta)\cos\theta \, d\omega$$

$$F^\vee = \int_{\text{lower hemisphere}} I(\theta)\cos\theta \, d\omega$$

It can be shown (admitted or see Salby) that the above 1D radiative transfer law is valid if I is replaced by $F^\wedge$ or $F^\vee$, dz by $5/3\, dz$ and B par πB.
(the argument is based on the ray profile and the fact they are saturated at the center and are transmetting by the aisles). Therefore :

$$\frac{dF^\wedge}{d\tau} = -F^\wedge + \pi B$$

$$\frac{dF^\vee}{d\tau} = F^\vee - \pi B$$

with

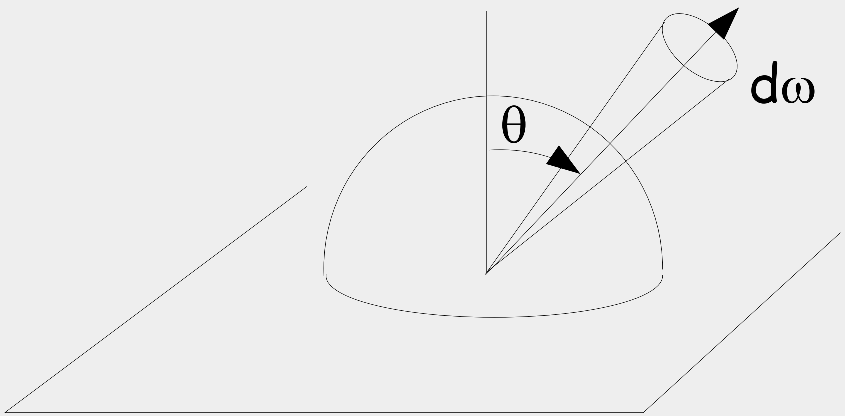
$$\tau = \int_0^z \frac{5}{3} k \rho \, dz'$$

For a stationary, purely radiative, regime

$F^\wedge - F^\vee = \phi$, outgoing IR flux,

is also equal to the ingoing solar flux $S_0(1-A)/4$.

In non stationary regime, $\frac{dF^\wedge}{dz} - \frac{dF^\vee}{dz} = -\rho C_p \left[\frac{dT}{dt}\right]_{rad}$



We define also the *optical depth* χ, which, by convention, is counted from the top of the atmosphere :

$$\chi = -\int_\infty^z \frac{5}{3} k \rho \, dz'$$

By definition

$$\chi_S = \tau_\infty$$
$$\chi = \chi_S - \tau$$

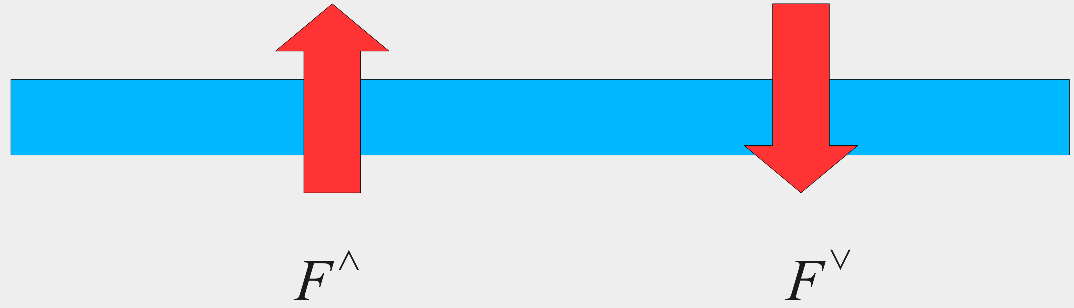
Black body law

$$B(T)=\pi^{-1}\sigma T^4$$

Opical thickness

$$d\tau=\kappa\rho\, dl$$

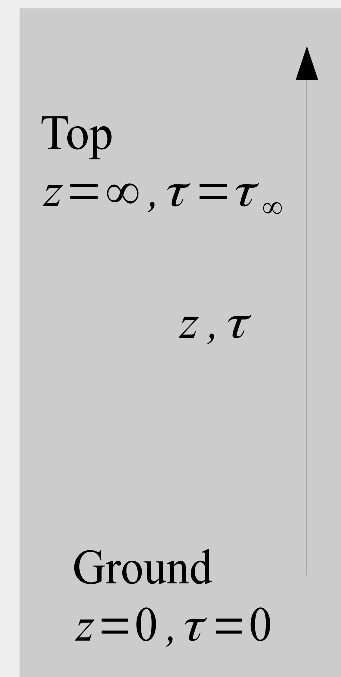
Absorption (m²/kg)
Depends on ν, T, p
(here we discard the dependency in ν)



Equation for upward and downward fluxes (diffusion is neglected for the thermal IR) using τ as vertical coordinate and applying the Kirchhoff law

$$\frac{dF^{\wedge}}{d\tau}=-F^{\wedge}+\pi B(T,\nu)$$

$$\frac{dF^{\vee}}{d\tau}=F^{\vee}-\pi B(T,\nu)$$



Solution

$$F^{\wedge}(\tau)=\sigma T_S^4 e^{-\tau}+\int_0^{\tau}\sigma T^4(\tau')e^{-\tau+\tau'}d\tau'$$

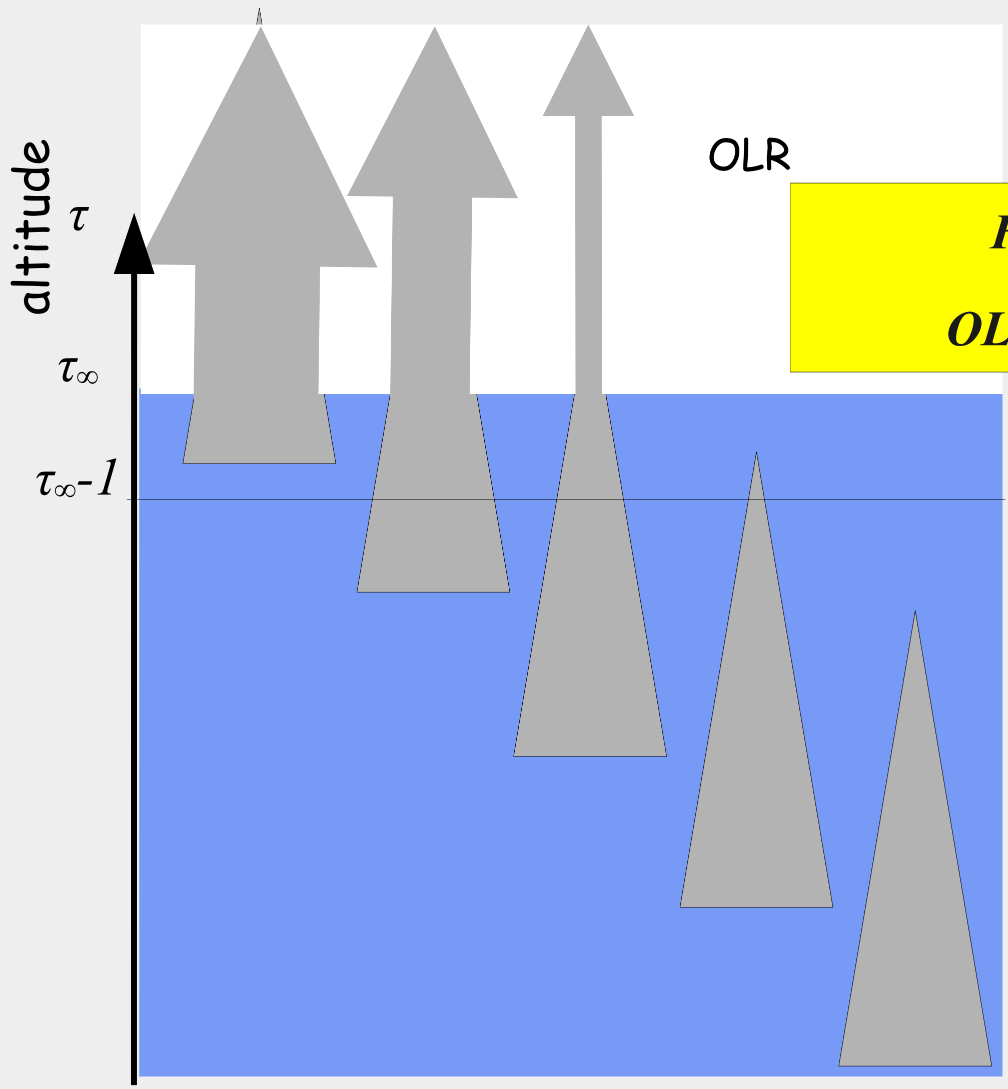
$$F^{\vee}(\tau)=\int_{\tau}^{\tau_{\infty}}\sigma T^4(\tau')e^{-\tau'+\tau}d\tau'$$

Outgoing flux at the top of the atmosphere

$$\boldsymbol{OLR=\sigma T_S^4 e^{-\tau_{\infty}}+\int_0^{\tau_{\infty}}\sigma T^4(\tau')e^{-\tau_{\infty}+\tau'}d\tau'}$$

Pour la Terre $\tau_{\infty}\approx 4$, pour Vénus $\tau_{\infty}\approx 80$

Emission altitude



$$F^{\wedge} = \sigma T_S^4 e^{-\tau} + \int_0^{\tau} \sigma T^4(\tau') e^{-\tau+\tau'} d\tau'$$

$$OLR = \sigma T_S^4 e^{-\tau_\infty} + \int_0^{\tau_\infty} \sigma T^4(\tau') e^{-\tau_\infty+\tau'} d\tau'$$

The upward IR radiation is emitted by the lowest layers is mostly absorbed. The exiting radiation originates from layers of sufficiently small optical depth not to be absorbed by the layers above. By convention, we define the emission level as a layer of unit optical depth :

$$\chi = \tau - \tau_\infty = 1$$

Case of an isothermal atmosphere at the same temperature as the ground

$$OLR = \sigma T_S^4 e^{-\tau_\infty} + \int_0^{\tau_\infty} \sigma T_S^4 e^{-\tau_\infty + \tau'} d\tau' = \sigma T_S^4$$

In this case, the outgoing IR emission to the space is the same as the one that would occur in the absence of absorption in the atmosphere

There is no greenhouse effect for an isothermal atmosphere at the same temperature as the ground

In the real atmosphere, the temperature decreases with altitude in the troposhere but a conseuence of this result is that the concentration of absorbing gases near the ground does not matter. In particular, we should not be mislead by the domination of H_2O in such layers.

Case of an atmosphere with a temperature profile $T = T_S(p/p_S)^{\gamma}$ and uniformly absorbing in the IR domain (gray hypothesis)

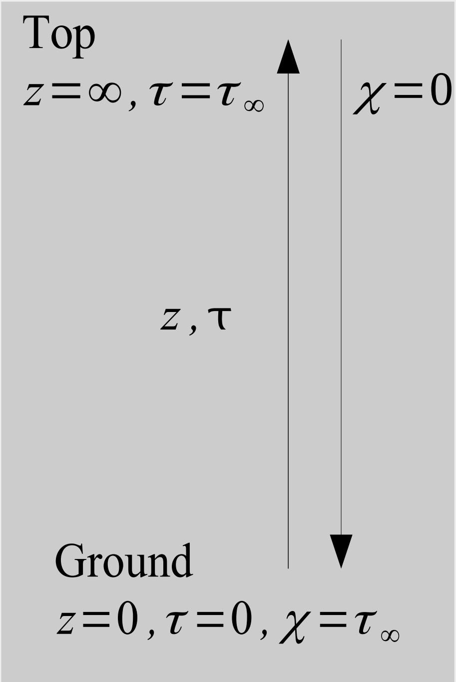
Using the hydrostatic equation $d\tau = \frac{-\kappa}{g} dp$

this, with $\chi = \tau_\infty - \tau$: $T = T_S \left(\frac{\chi}{\tau_\infty}\right)^{\gamma}$

$$OLR = \sigma T_S^4 e^{-\tau_\infty} + \sigma T_S^4 \int_0^{\tau_\infty} \left(\frac{\chi}{\tau_\infty}\right)^{4\gamma} e^{-\chi} d\chi$$

$$OLR = \sigma T_S^4 \left(e^{-\tau_\infty} + \left(\Gamma(1+4\gamma, 0) - \Gamma(1+4\gamma, \tau_\infty)\right) \tau_\infty^{-4\gamma}\right)$$

For τ_∞ large: $\boldsymbol{T_S = \left(\frac{S_0}{4\sigma}(1-\alpha)\right)^{1/4} \Gamma(1+4\gamma, 0)^{-1/4} \tau_\infty^{\gamma}}$



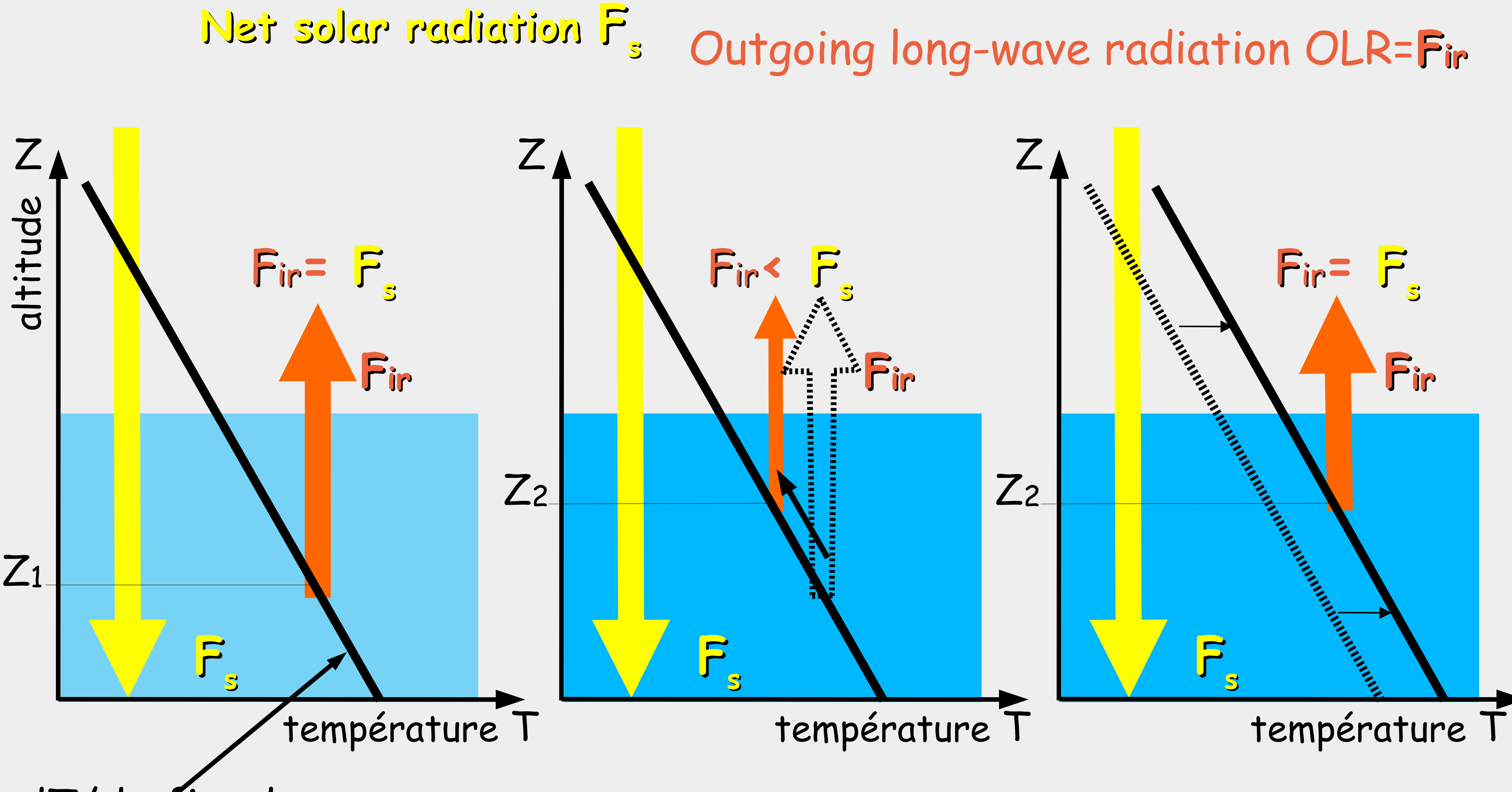
OLR/σ — Greenhouse effect

The greenhouse effect grows with the the total optical thickness and depends of the temperature profile.

If the absorption varies as $\kappa = \kappa_S (p/p_S)^m$ (widening of the rays as a function of pressure),

Then replace γ par $\gamma/(1+m)$ in the last formula.

Greenhouse effect in a stratified atmosphere



dT/dz fixed
by convection

GHG (CO_2) increases, Z_e increases, T_e decreases: Smaller outgoing long-wave radiation

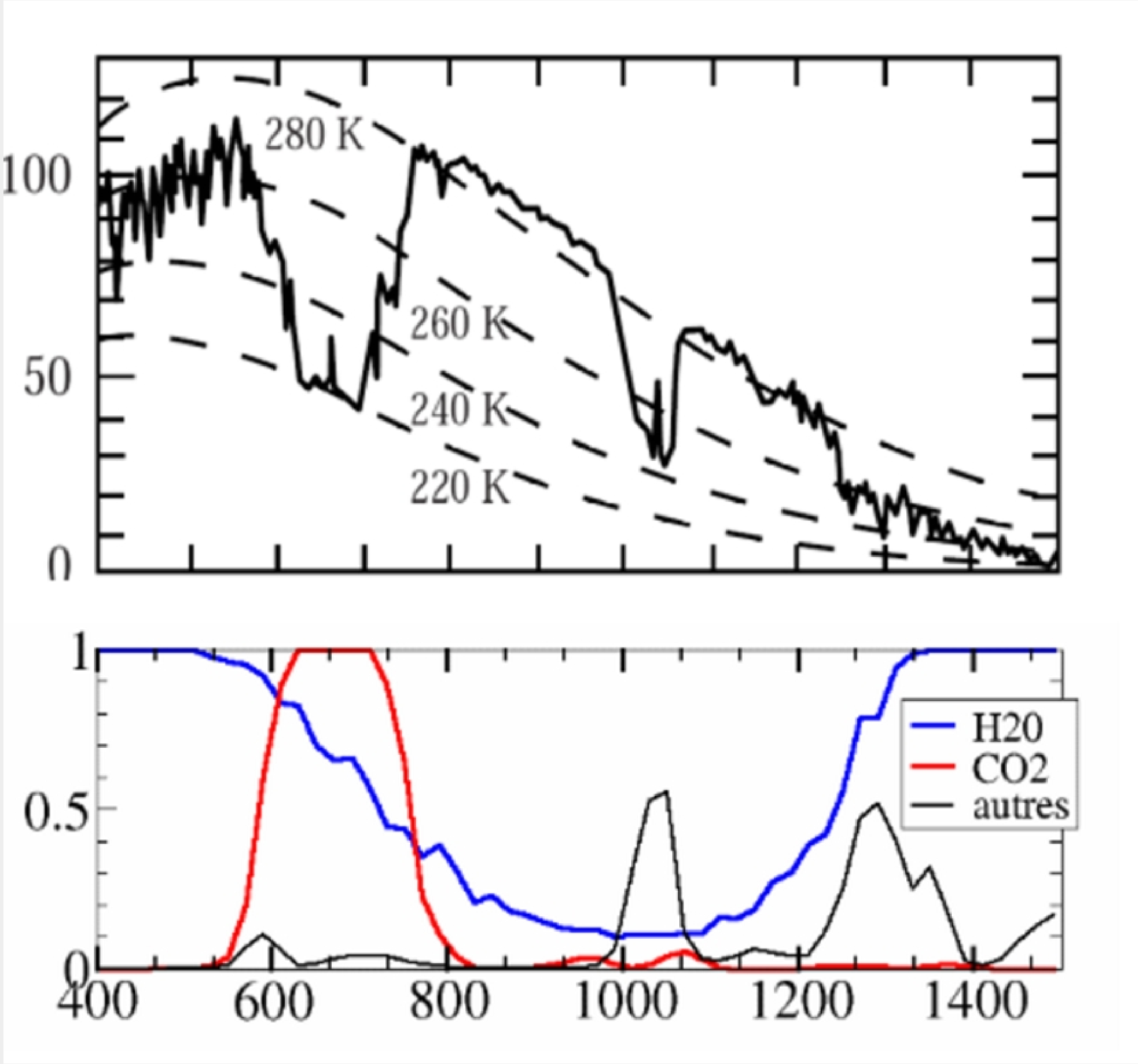
T(z) increases:
Return to equilibrium

Doubling of CO2 :
+150m, -1K, -4W/m2

Dufresnes, 2010

Emission spectrum of the Earth observed from space

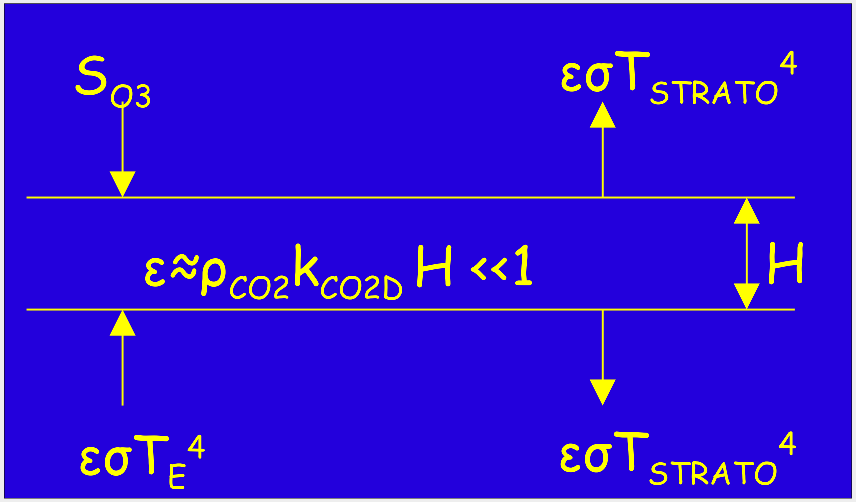
Total atmospheric absorption



Unit : cm^{-1}

Cooling of the stratosphere when CO_2 concentration is rising

The stratosphere is close to a radiative equilibrium where the shortwave absorption (mainly by ozone) is compensated by the long-wave emission upwards towards space and downwards towards the troposphere (mainly by CO_2)



At equilibrium

$$S_{O3} + \varepsilon\sigma T_E^4 = 2\varepsilon\sigma T_{STRATO}^4$$

hence

$$T_{STRATO} = \left(\frac{\frac{S_{03}}{\varepsilon} + \sigma T_E^4}{2\sigma} \right)^{1/4}$$

If the concentration in CO_2 increases, then $\varepsilon \approx \rho_{CO2} k_{CO2} H$ increases.

Assuming :

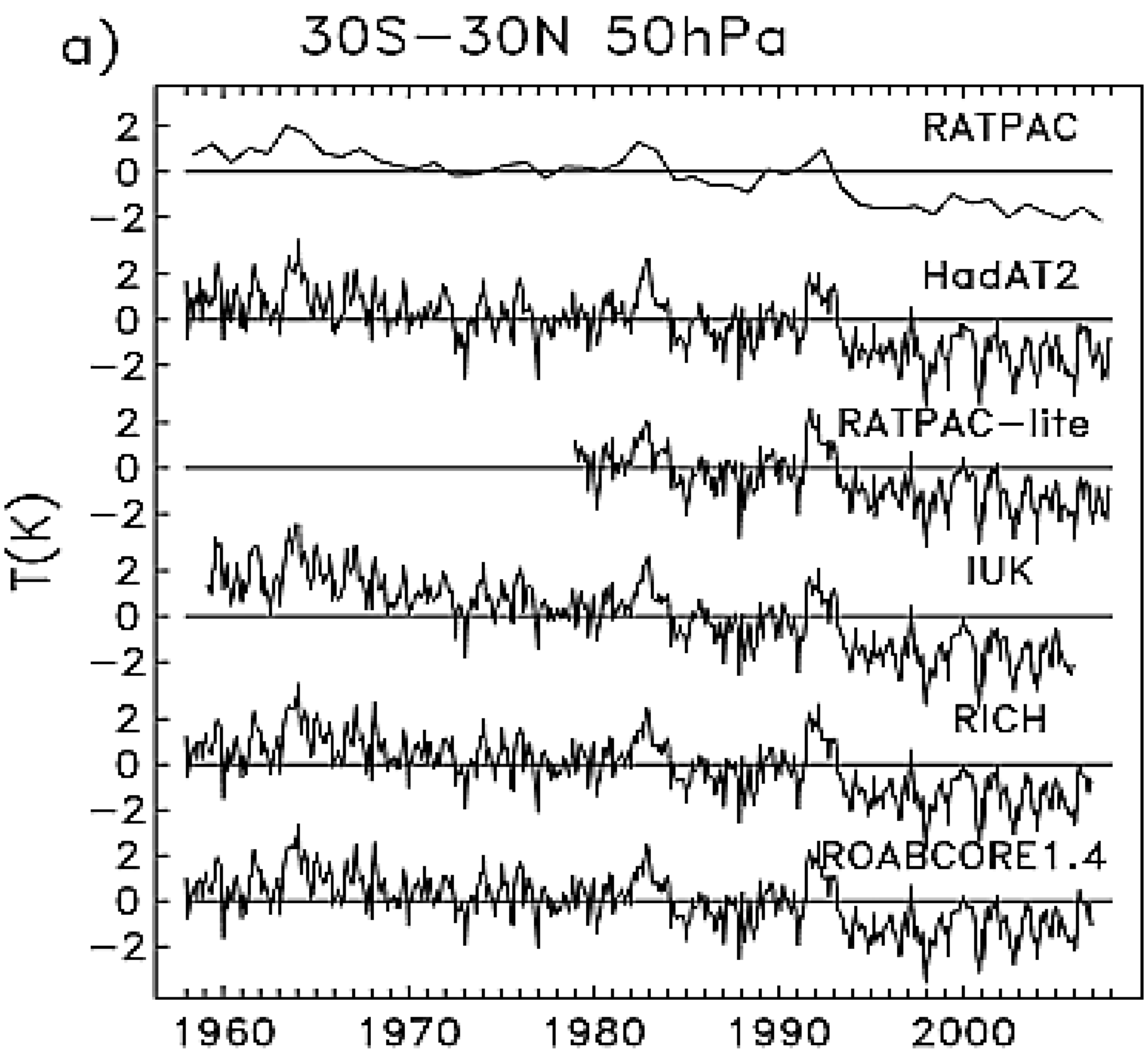
(1) that the concentration in ozone does not change,

(2) that the planetary albedo does not change (hence T_E is preserved)

then T_{STRATO} decreases.

Greenhouse effect = Warming of the lower layers and cooling of the upper layers

Randel - 2008JD010421



This looks at first sight as a clear confirmation of global warming but it can be shown that the stratospheric temperature decay during recent decades is mostly due to the ozone hole.

RADIATIVE-CONVECTIVE EQUILIBRIUM

The radiative convective equilibrium is a steady state of the atmospheric column where the temperature slope is determined by convection in the stratosphere and by radiation in the troposphere.

Conditions:

In the stratosphere, at any level, the upward longwave radiative flux equals the total downward longwave + shortwave flux.

In the troposphere, the radiative fluxes do not balance. The upward longwave flux is smaller than the total downward radiative flux. The difference is compensated by the upward convective flux (latent+sensible heat).

See how to solve the problem for the simplest gray atmosphere in Gray_RCE.pdf and in the RCE climlab notebook.

Greenhouse effect in an atmosphere in pure radiative equilibrium

If $\psi = F^{\wedge} + F^{\vee}$ et $\phi = F^{\wedge} - F^{\vee}$

then $\dfrac{d\psi}{d\chi} = \phi$ et $\dfrac{d\phi}{d\chi} = \psi - 2\pi B$.

In stationary regime: $\dfrac{d\phi}{d\chi} = 0$,

hence $\psi = 2\pi B$ avec $B = \dfrac{\phi}{2\pi}\chi + cste$

at the top of the atmosphere , $F^{\vee} = 0$
(neglecting the incoming long wave flux).

$\rightarrow \psi(\chi=0) = \phi$ et $B = \dfrac{\phi}{2\pi}(\chi+1)$.

Similarly :

$$F^{\wedge} = \frac{\phi}{2}(\chi+2) \text{ et } F^{\vee} = \frac{\phi}{2}\chi,$$

and the temperature of the atmosphere is

$$T = \left(\frac{\phi}{2\sigma}(\chi+1)\right)^{1/4}$$

At ground level $(\chi = \chi_S)$

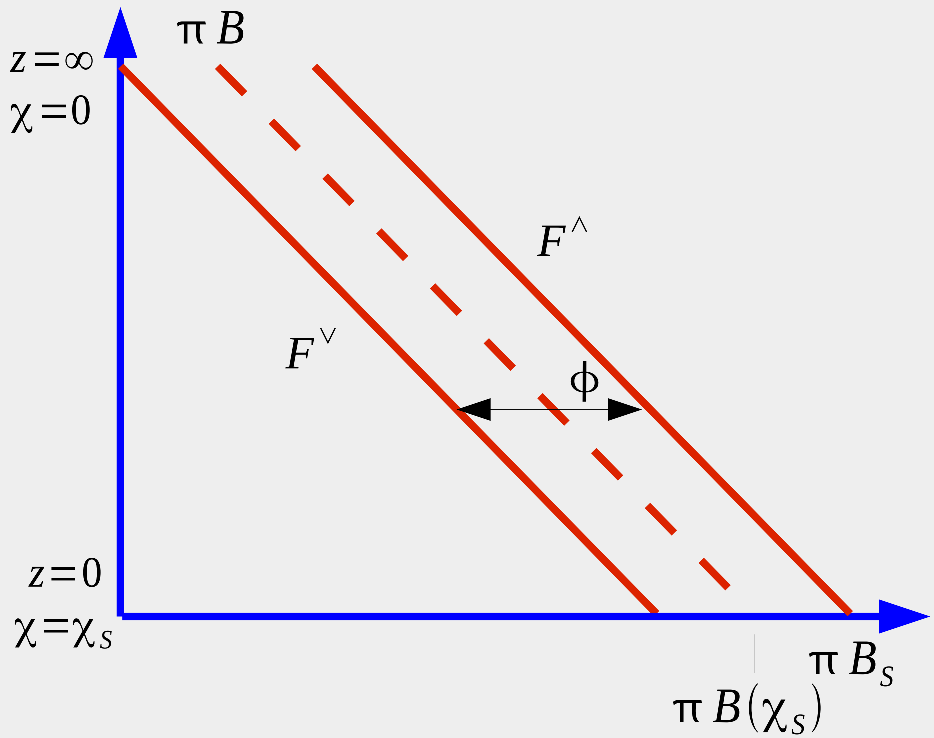
$$F^{\wedge}(\chi_S) = \frac{\phi}{2}(\chi_S+2) = \pi B(\chi_S) + \frac{\phi}{2} = \pi B_S$$

où $\pi B_S = \sigma T_S^4$ is the ground emission.

Greenhouse effect

$$\pi B_S = \frac{\phi}{2}(\chi_S+2) = \sigma T_S^4$$

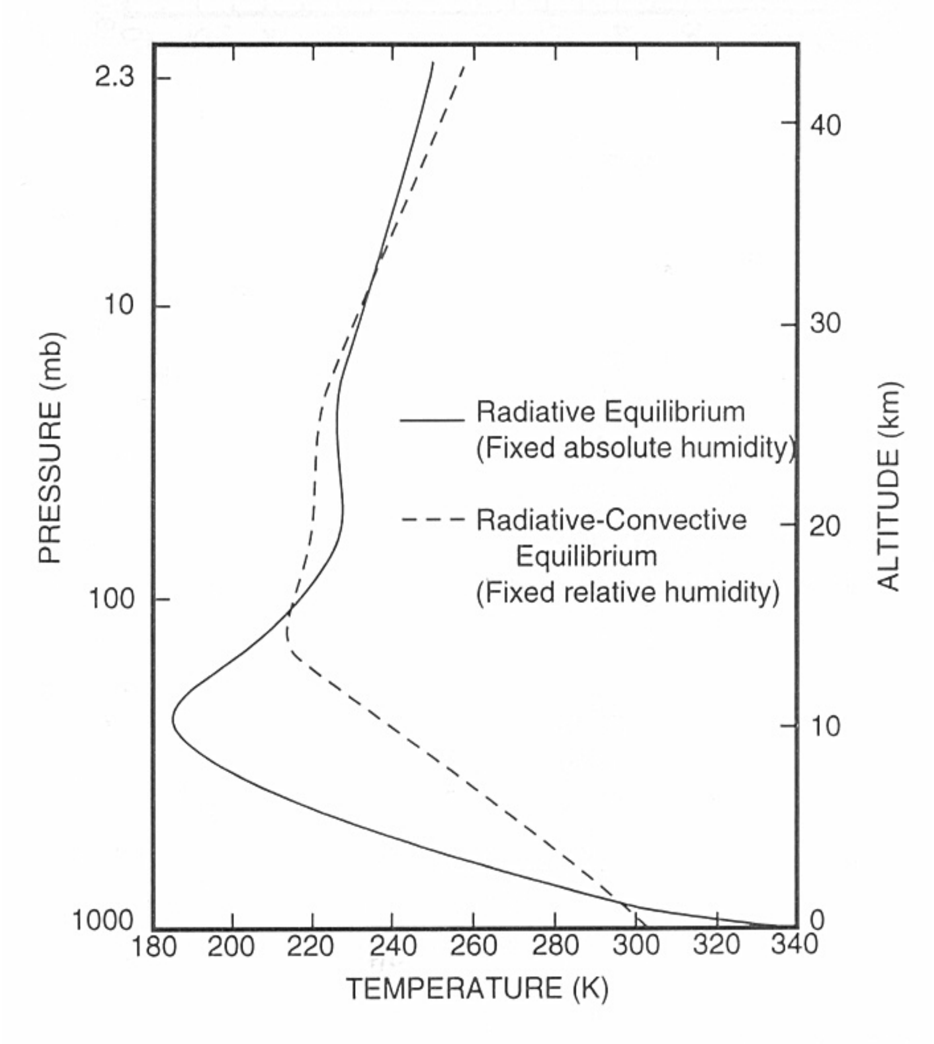
$\chi_S = 0$: no greenhouse effect.
Earth $\chi_S = 4 \rightarrow T_S \approx 336\,K$
for $z \rightarrow \infty$, T tends to $T_e = 215\,K$,
which is the skin temperature
Notice
- the ground temperature is larger than the surfcae air temperature.
- the skin temperature is smaller than 255K, the Earth black-body temperature



Radiative convective equilibrium calculated by including CO_2, H_2O and ozone

The stratosphere is (almost) in radiative equilibrium (basically: UV absorption by O_3 and IR emission IR by CO_2). In the whole troposphere, the radiative budget produces a net cooling. It is compensated by heating due to convective transport.

In the radiative-convective model, the stratosphere is assumed in radiative equilibrium and the troposphere is assumed to have a constant lapse rate dT/dz. The convective adjustment is made such that the upward IR flux is continuous at the tropopause.



From Manabe & Strickler, JAS, 1966

Iterative solution of the RCE by convective adjustment

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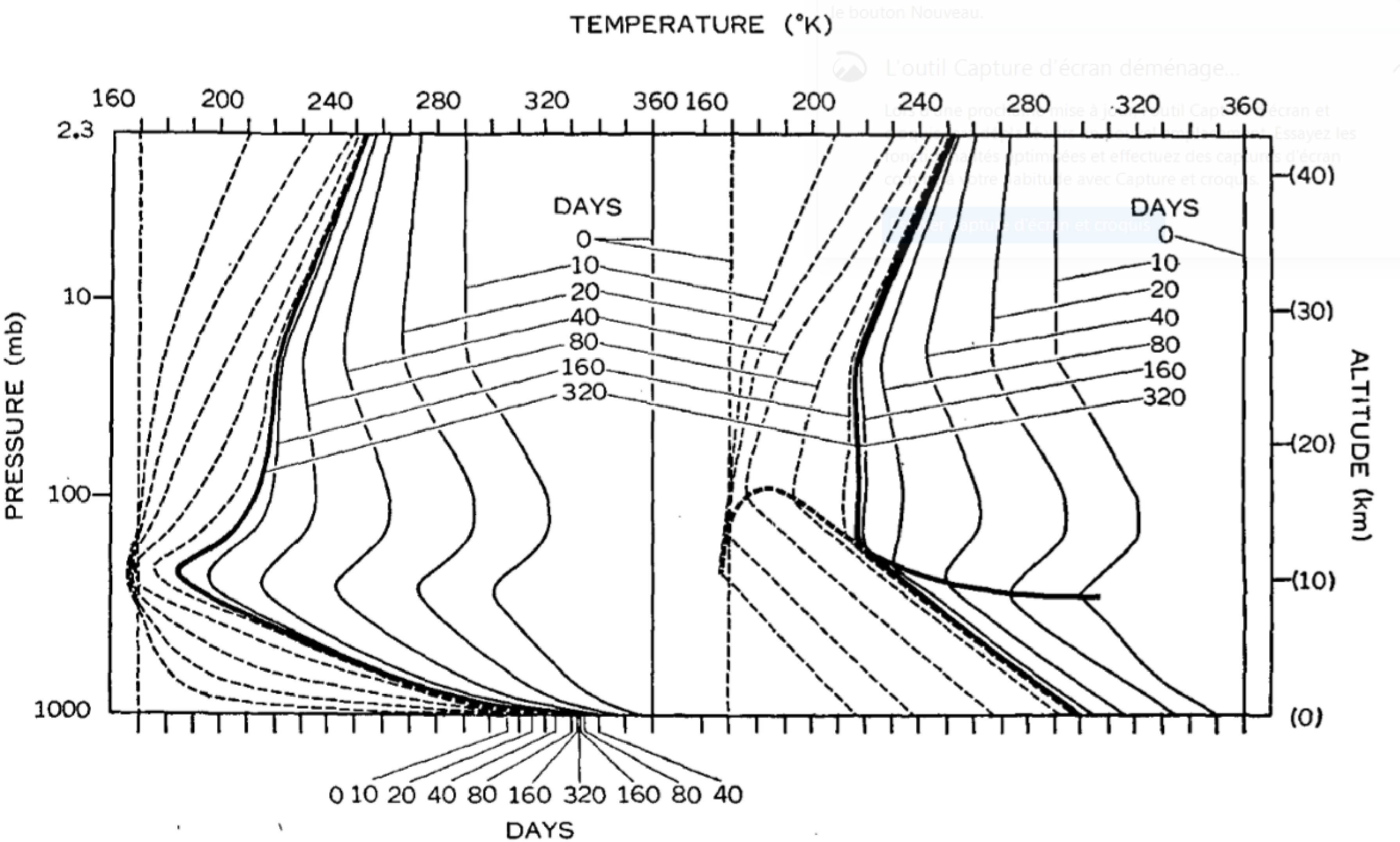


FIG. 1. The left and right hand sides of the figure, respectively, show the approach to states of pure radiative and thermal equilibrium. The solid and dashed lines show the approach from a warm and cold isothermal atmosphere.

HINTS ABOUT CONVECTIVE ADJUSTMENT
OR A SIMPLE CONVECTIVE PARAMETERIZATION

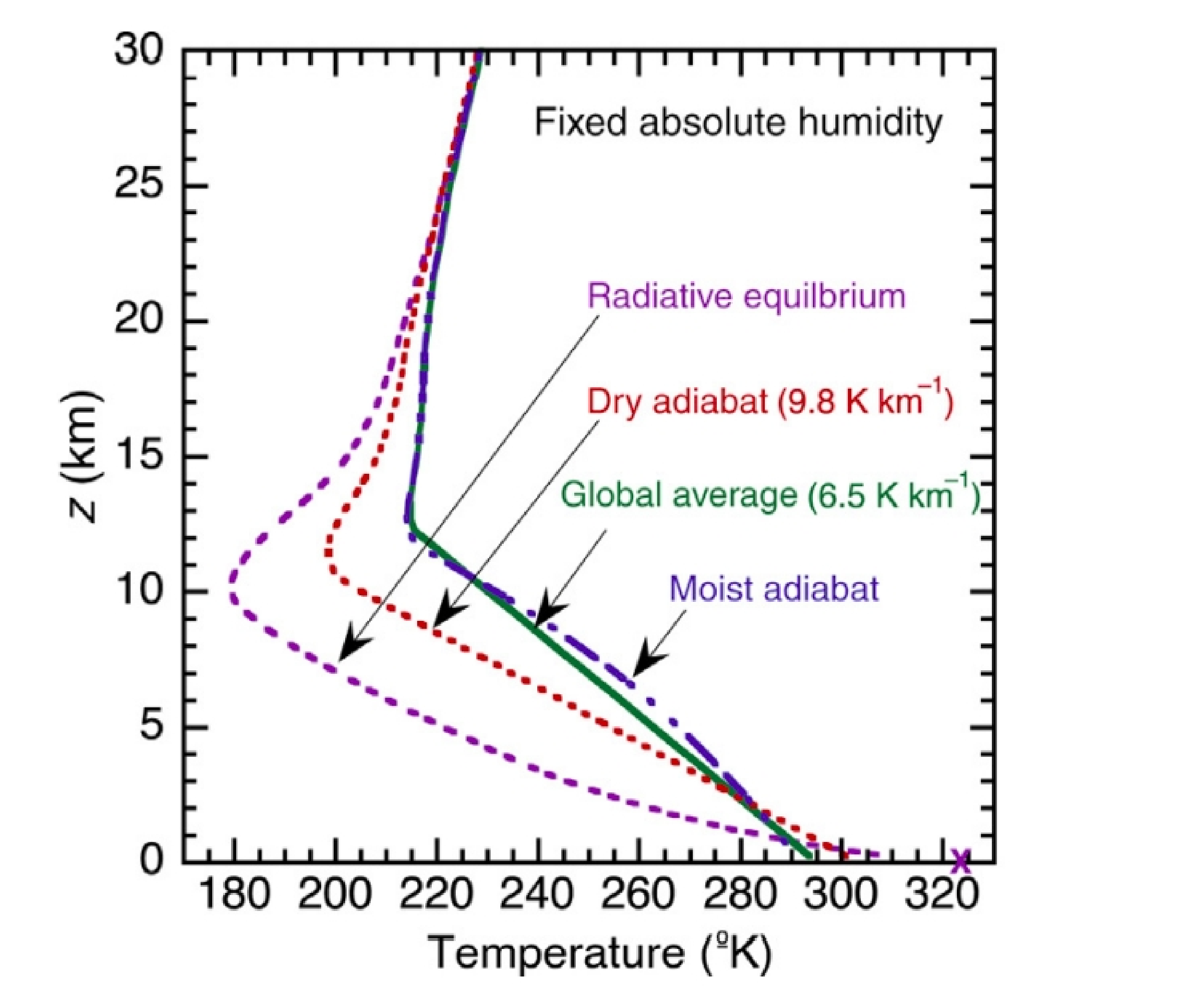
The tropospheric temperature profile is fixed as an equilibrium between convection which relaxes the profile towards the neutral profile of uniform equivalent potential temperature and the destabilisation effect of radiative exchanges which relaxes towards an unstable profile. The result is an average $\Gamma = dT/dz = -6.5$ K per km.

Starting from an initial arbitrary profile. The temperature is progressively adjusted to reach a steady state.

In the adjustment step, the temperature profile is relaxed towards Γ while there is no heat exchanges with the outside of the atmospheric column, so that $\int_{p_{\mathrm{top}}}^{p_{\mathrm{bottom}}} \delta T \, dp$

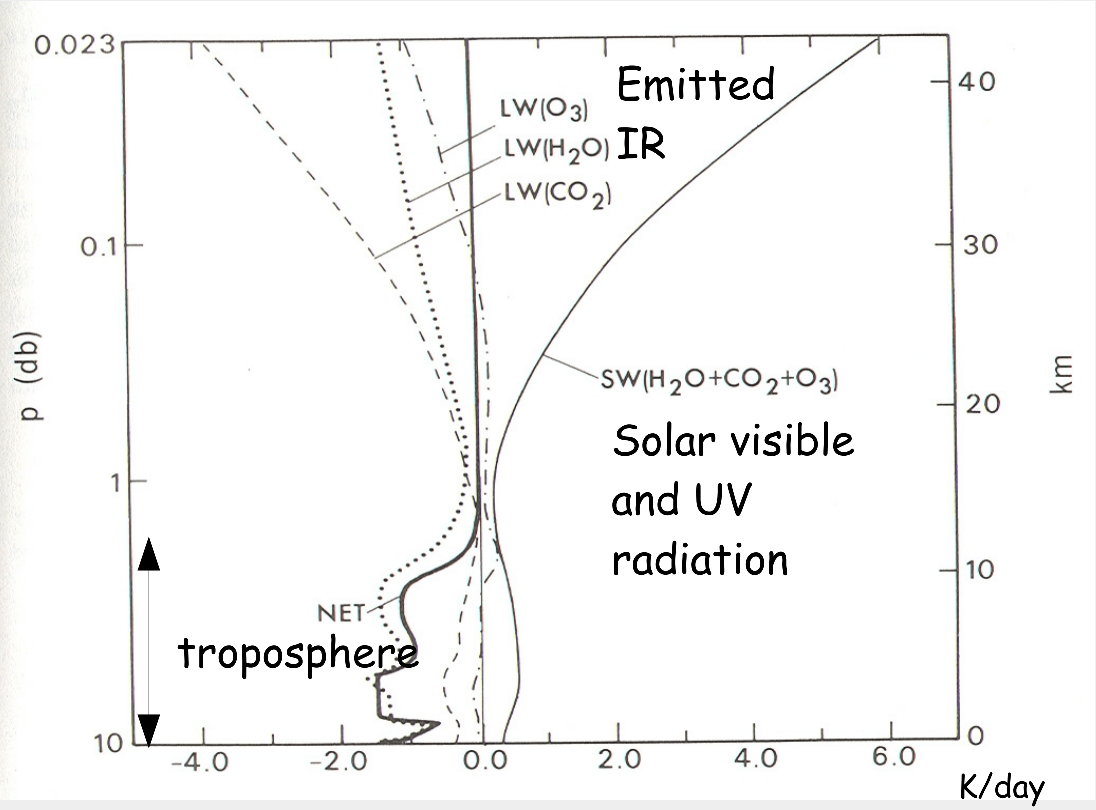
Such a method can be used as a crude parameterization of convection in a dynamical model.

See Manabe, Smagorinsky & Strickler, 1965, in the bibliography.



From Hartmann, Global physical climatology

Contribution of the main GHG to the radiative budget



The temperature profile in the troposphere is not determined by the radiative exchanges. It is governed by the vertical stirring and mixing performed the meteorological perturbations and cloud convection.

The temperature at the surface is determined by the radiative flux at the tropopause rather than by the radiative flux at the surface.

The net radiative flux at the surface (absorbed - emitted) determines the exchanges between the surface and the atmosphere → it controls and limits the evaporation and as such fixes the relative humidity near the surface.

A few considerations on the spectral distribution of CO2 absorption

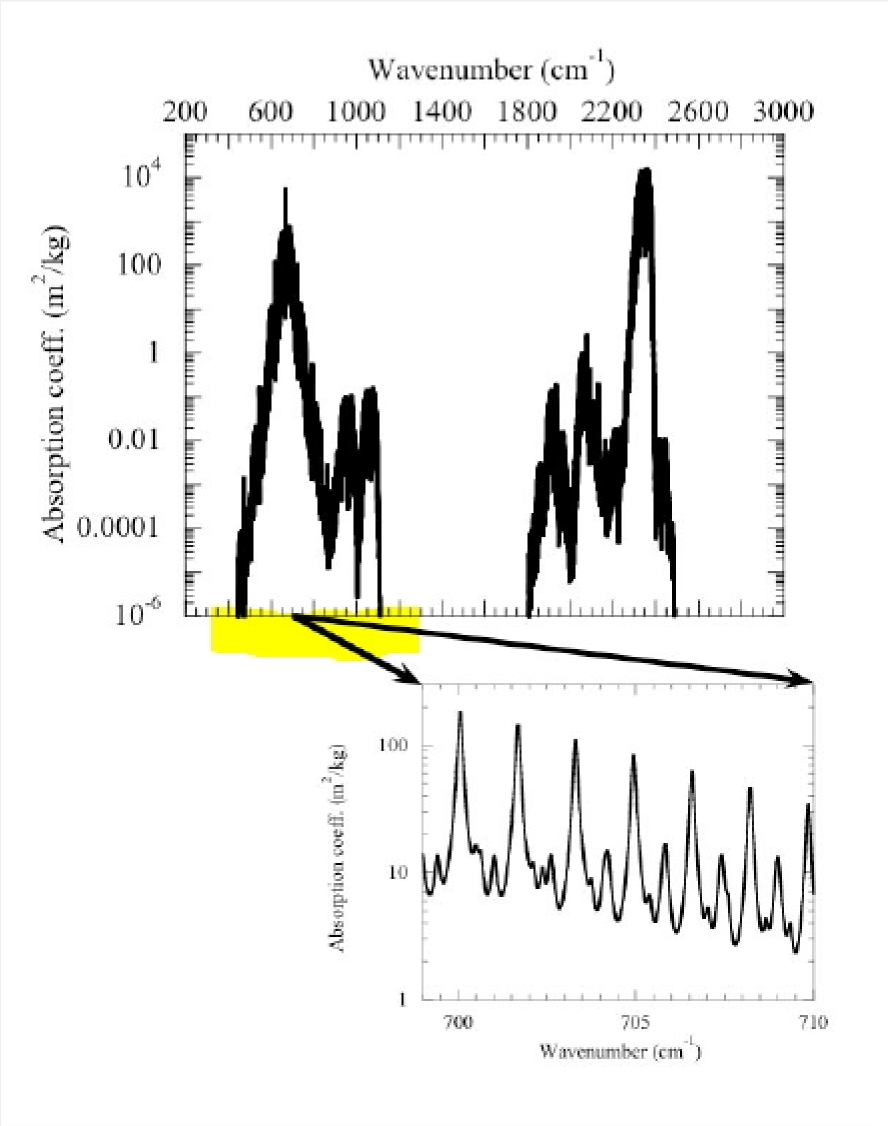


Figure 4.7: The absorption coefficient vs. wavenumber for pure CO_2 at a temperature of $293K$ and pressure of $10^5 Pa$. This graph is not the result of a measurement by a single instrument, but is synthesized from absorption data from a large number of laboratory measurements of spectral features, supplemented by theoretical calculations. The inset shows the detailed wavenumber dependence in a selected spectral region.

Widening of the rays by collision and Doppler effect

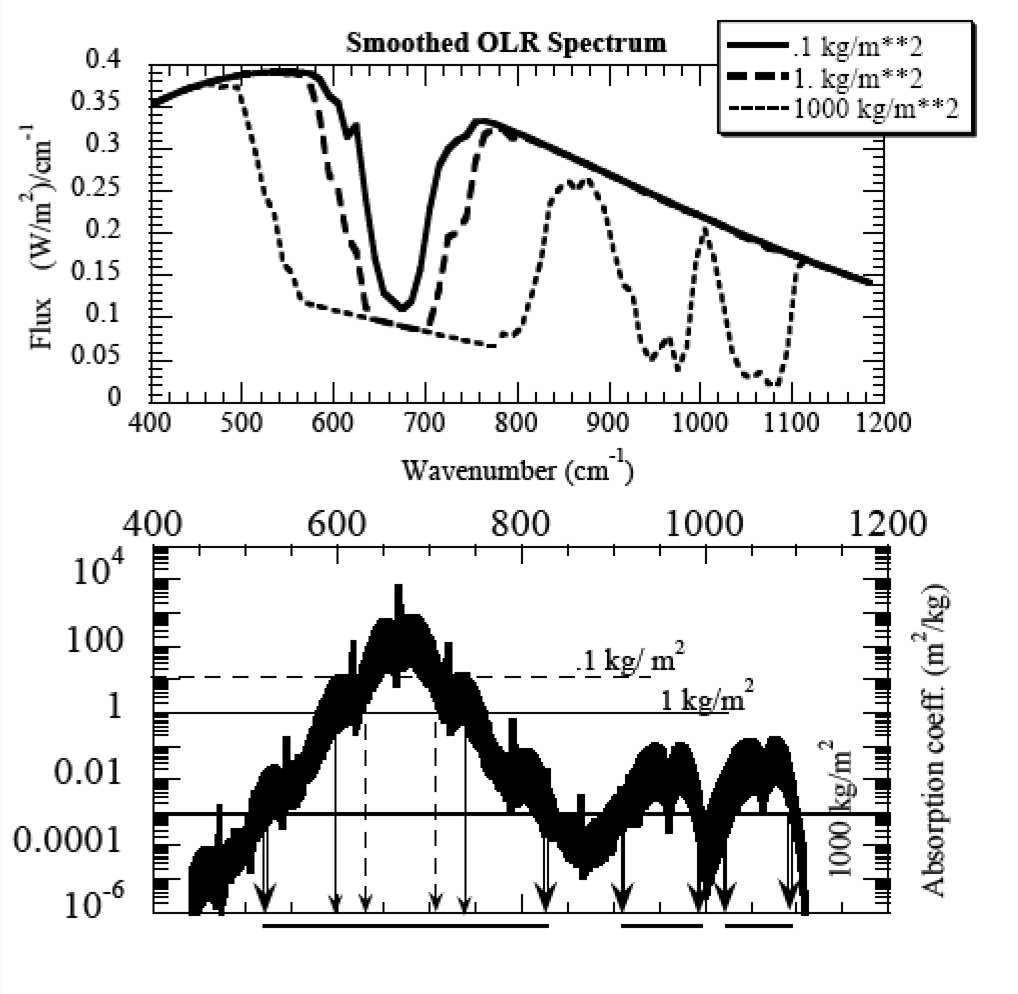
Lorentz and Voigt profiles

→ Absorption bands

$\kappa = 10\, m^2\, kg^{-1}$ leads to an absorption depth of 120m at 1000 hPa for 400 ppmv

Good news :
The growth of CO2 absorption is logarithmic as a function of concentration.

Effect of the stratification and the shape of the bands



Fo 400 ppmv of CO2,with uniform distribution, we get 4kg/m²

Figure 4.12: Lower panel: The absorption coefficient for CO_2 at 1 bar and $300K$, in the wavenumber range of interest for Earthlike and Marslike planets. The horizontal lines show the wavenumber range within which the optical thickness exceeds unity for CO_2 paths of $\frac{1}{10}$, 1 and 1000 kg/m^2. Upper panel: The corresponding OLR for the three path values, computed for the same temperature profiles as in Figure 4.5. The OLR has been averaged over bands of width 10 cm^{-1}.

Note: for minor gases like CFC and CH4, where saturation is not reached, the growth is linear → much larger impact per added molecule.

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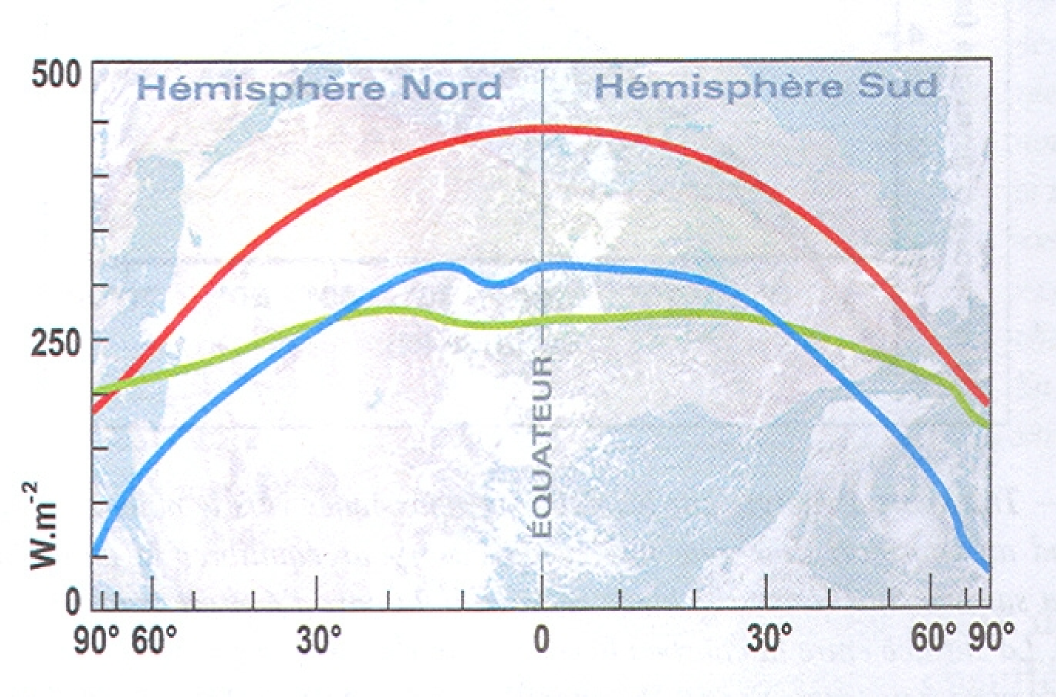
VI Climate sensitivity



Radiative budget of the Earth

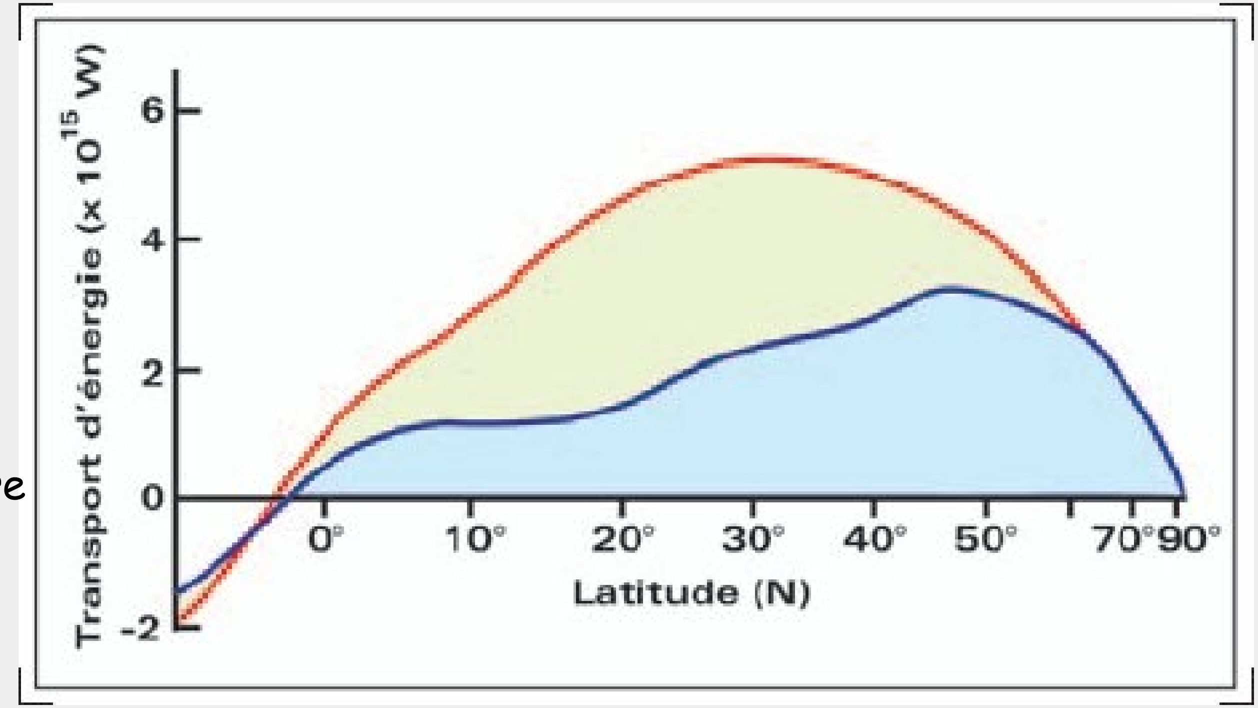
red: incoming radiation at the top of the atmosphere
blue: absorbed solar radiation
green: rIR radiation emmitted to space

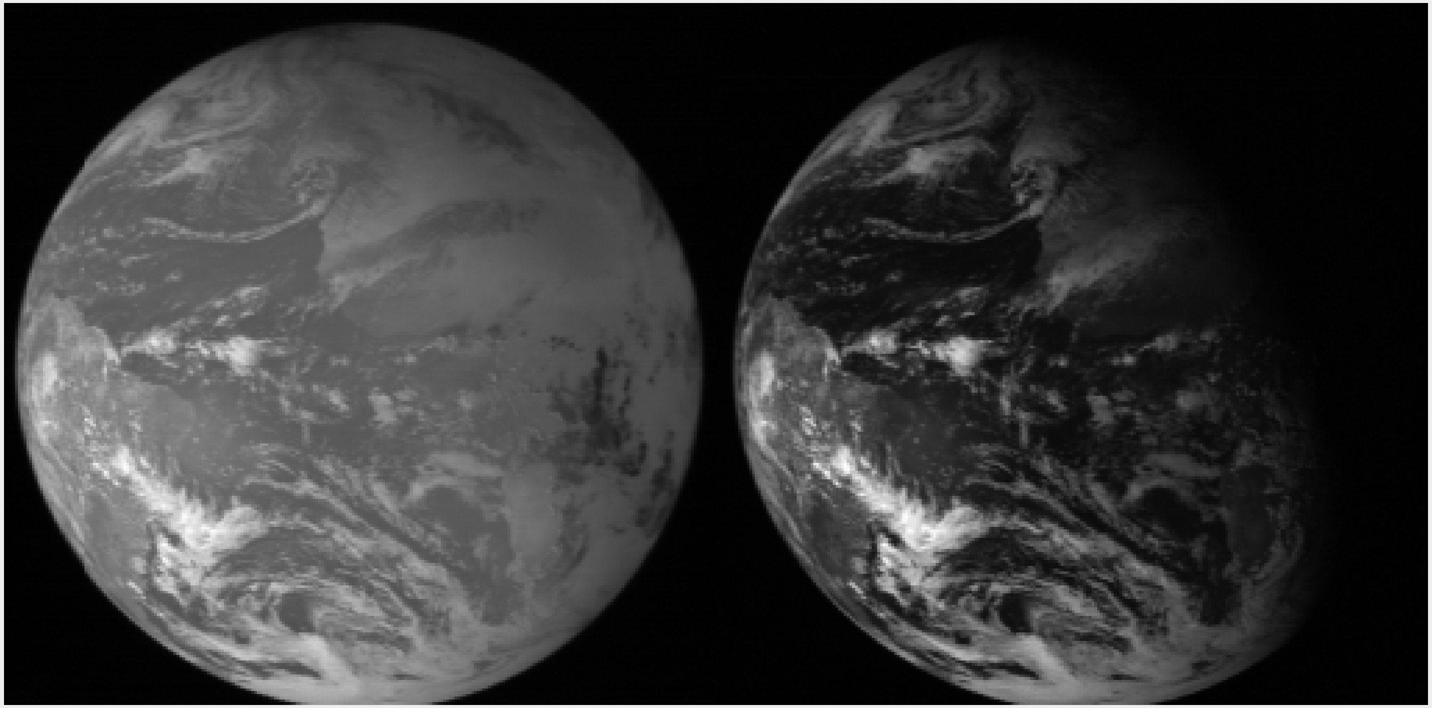
Excess in the low latitudes and deficit in tha high latitudes -> need for a compensating heat transport from low to high latitudes
[Malardel, 2005; Gill, 1982]

Energy transport by the geophysical fluids

red: total transport
blue: transport by the atmosphere
The oceanic transport is shown by the area between the blue and red curves.

The atmosphere and the ocean share almost equally the transport up to 60°.





Total radiative budget (outgoing - incoming)

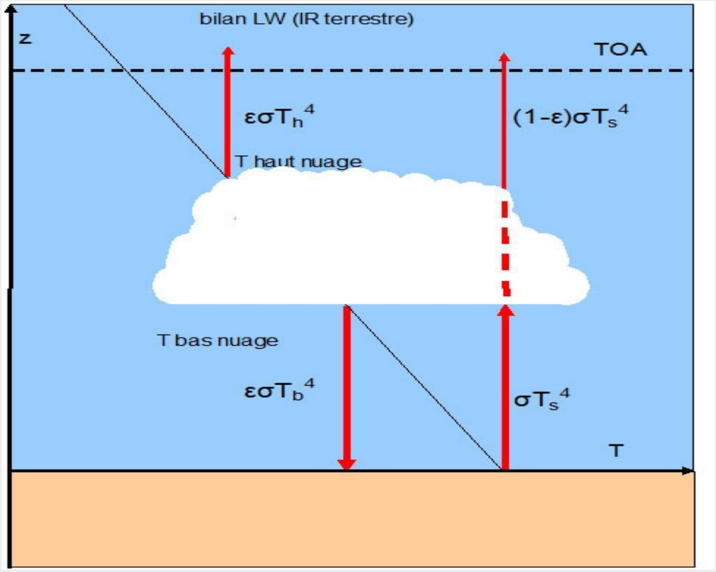
Visible channel

Cloud effect:

During daytime, the dominating effect is to reflect the incoming radiation by the upper surface of clouds (-> cooling effect); during nighttime, only reamains the blocking of IR emission by clouds (-> warming greenhouse effect).

Radiative effect of the different type of clouds

Clouds diffuse and absorb the solar incoming radiation.
Clouds absorb the upward infrared radiation and emit towards the sky and the ground as black-bodies at the temperature of their top and basis



LW radiative budget

Th: cloud top temperature
Tb: cloud bottom temperature
ϵ : cloud emissivity

Cloud radiative forcing at the top of the atmosphere

CRFir = Lwclear – LWcloud
LWclair = σTs^4
CRFir = $\epsilon \sigma (Ts^4 - Th^4)$

Low thick clouds (Cumulus, Stratus) : large IR emission and large albedo -> cooling effect
Thin cirrus : low albedo but significant absorption in the IR, low emission due to their low temperature -> warming effect
Thick convective clouds: large albédo and low IR emission -> neutral or small effect

Clouds and shortwave radiations

For a particle of radius r the incident solar flux I_0 is reduced by $\pi r^2 k_e I_0$ where k_e is the total extinction by scattering and absorption $k_e = k_s + k_a$

The optical depth of a layer of depth h is

$$\tau = \pi h \int_0^{\infty} k_e(r) r^2 n(r) dr$$

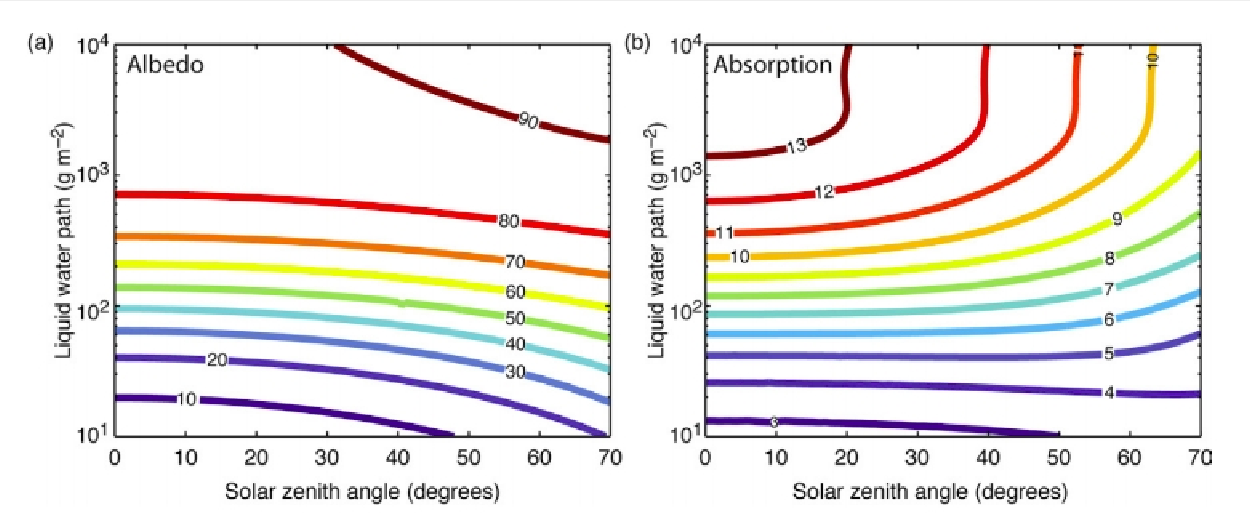
where $n(r)$ is the number density of particles of radius r per cubic meter.

It is found that for $r \gg$ wavelength, $k_e = 2$ and if the particle distribution peaks at $\bar{r}$, then $\tau \approx 2\pi h \bar{r}^2 N$ where $N = \int_0^{\infty} n(r) dr$ is the total number density of the particles

For a water cloud, the total mass of liquid water is $LWC = \frac{4}{3} \bar{r}^3 \rho_L N h$

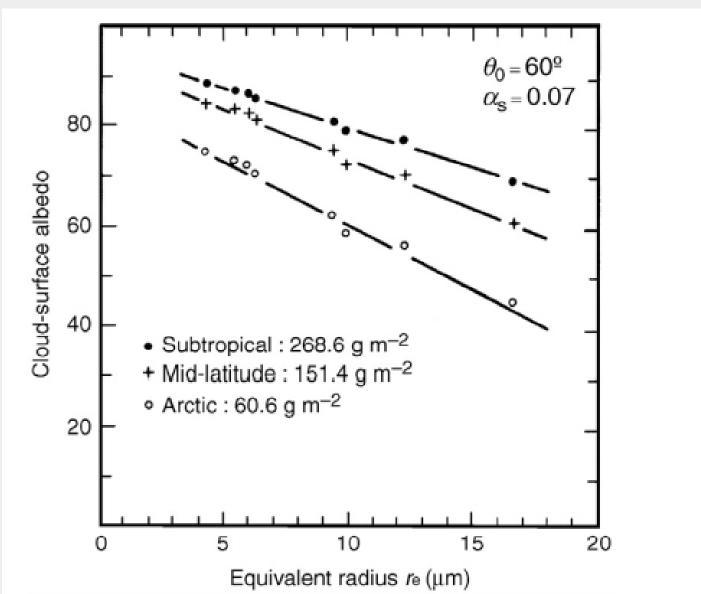
where ρ_L is the density of liquid water.

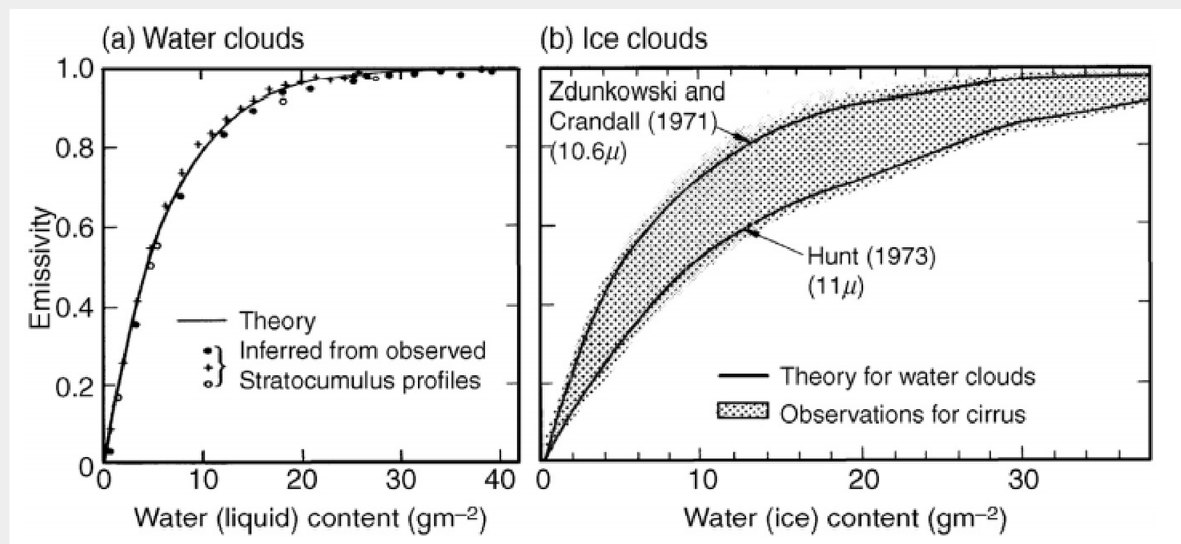
Therefore $\tau = \frac{3}{2} \frac{\text{LWC}}{\rho_L \bar{r}}$. For fixed liquid water τ varies in inverse proportion of the mean radius because small particles expose a larger total surface to radiation than big particles.



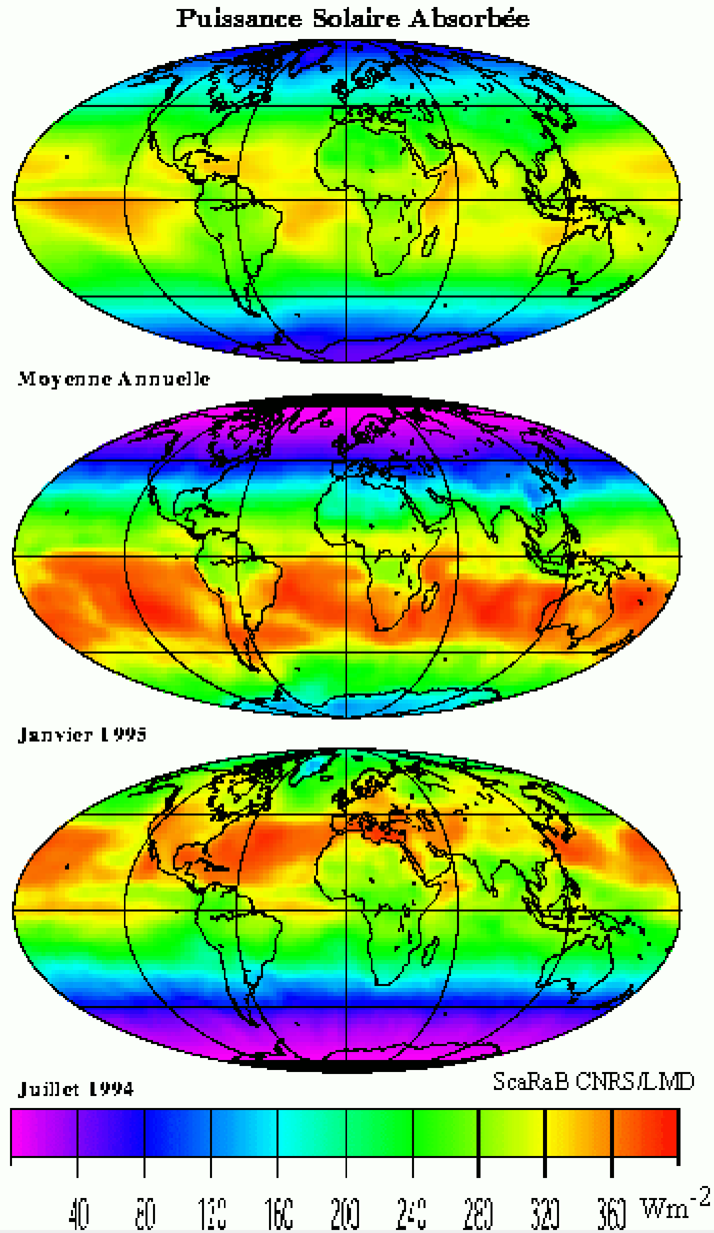
Here $\bar{r}$ = 14 μm

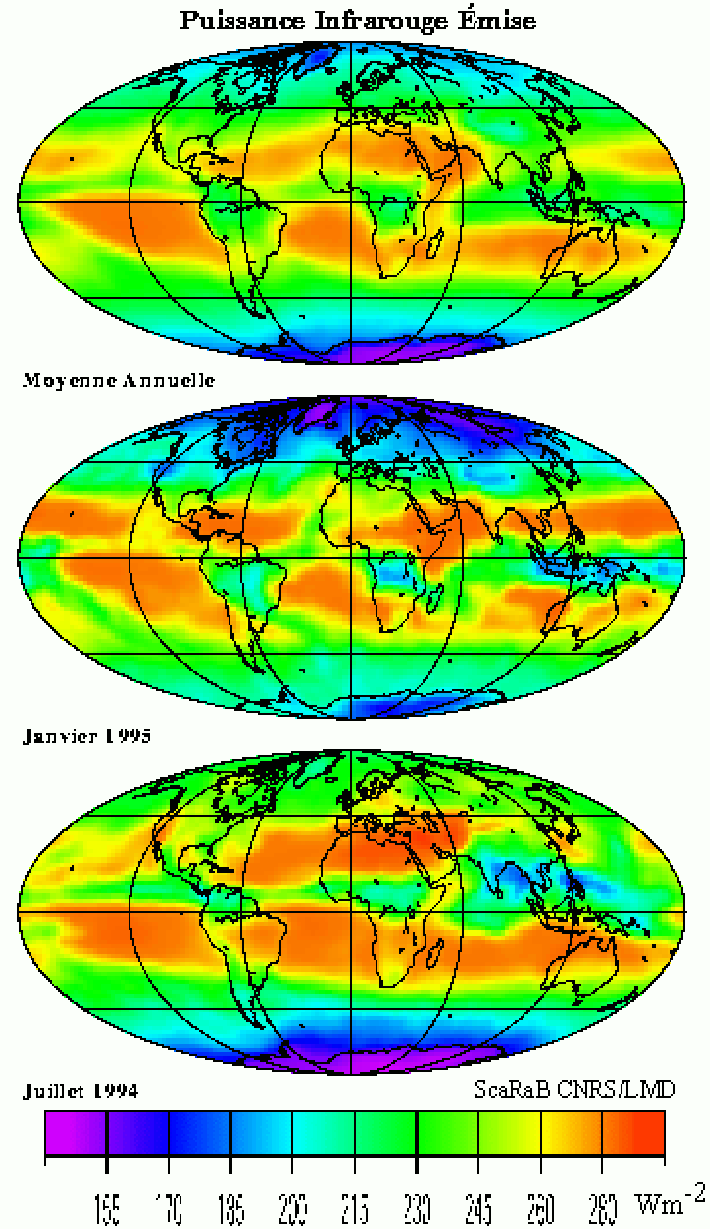
From Hartmann





Infrared absorption (emissivity) by clouds depends essentially on the LWC (or IWC)

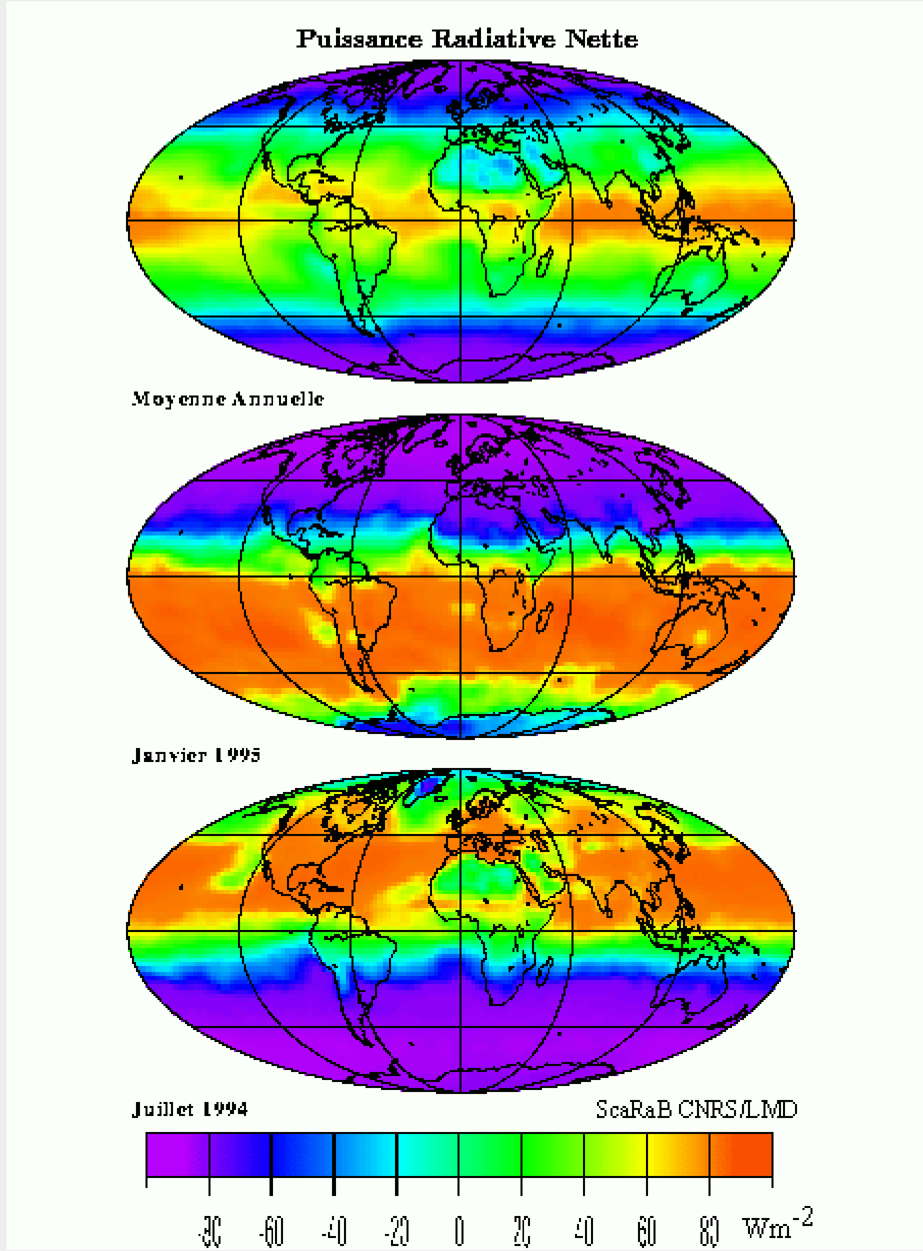




Radiative budget of the Earth from space

Directly measured from a radiometer (ScaRaB,LMD) on board a satellite

Notice, in the tropics, the behaviour of convective zones (small absorption and emission), the ocean (stron absorption and emission) and the deserts (small absorption and strong emission)



Radiative budget of the Earth from space (cont'd)

Notice the compensations between LW and SW in the tropics

Evaluation of the radiative forcing of clouds

- <u>LW radiative forcing of clouds</u>

$$CRFir = - LWclear + LWcloud$$

LWcloud: IR outgoing flux for a cloudy sky as measured by a satellite
LWclear: IR outgoing flux for a clear sky as measured by a satellite or as calculated with radiative model.

- <u>SW radiative forcing of clouds</u>

$$\begin{aligned} CRFsw &= SW_{cloud} - SW_{clear} \\ &= [\alpha_{cloud} - \alpha_{clear}]\, E_o \end{aligned}$$

SWcloud: net incoming solar flux in a cloudy sky
SWclair: net incoming solar flux in a clear sky
Eo : Incoming solar flux at the top of the atmosphere
anua : cloudy sky albedo
aclair : clear sky albedo

Cloud radiative effect

(calculated on each pixel: <average of cloudy cases> – <average of clear sky cases>, Positive flux counted in the descending direction)

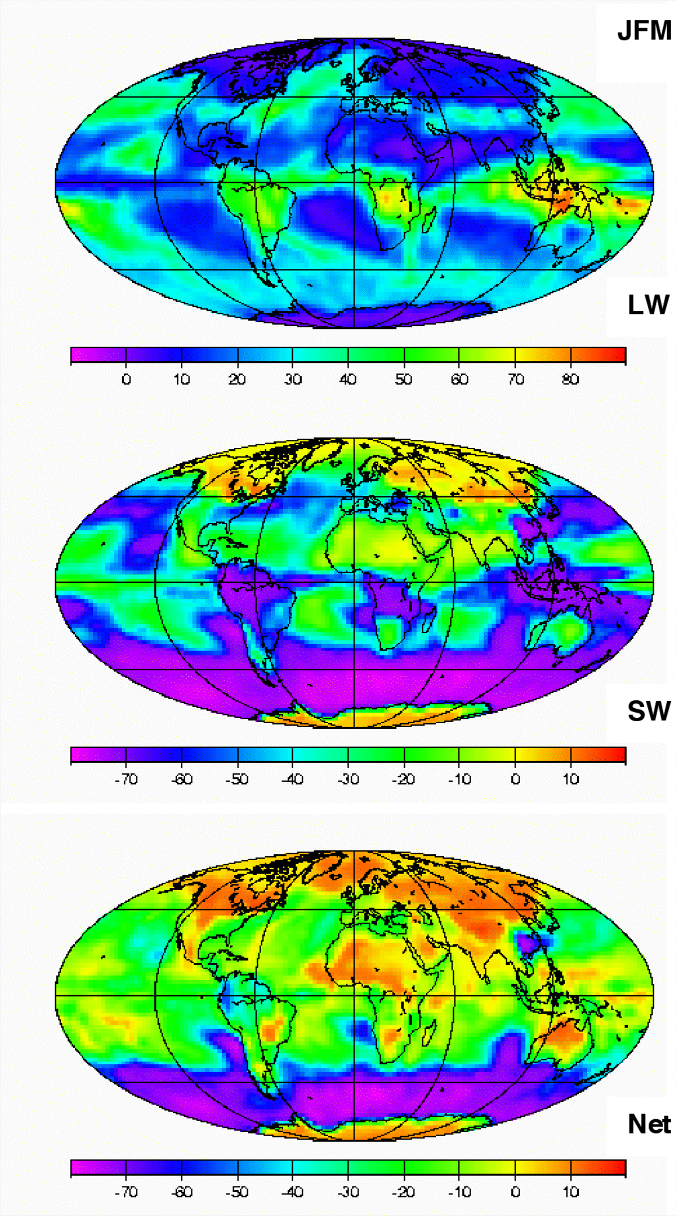
Average in winter 1999 (JFM) in W m^{-2}
data ScaRaB LMD

Long waves (infra-red)

CRF ~ - 20 W/m2

Short waves (visible)

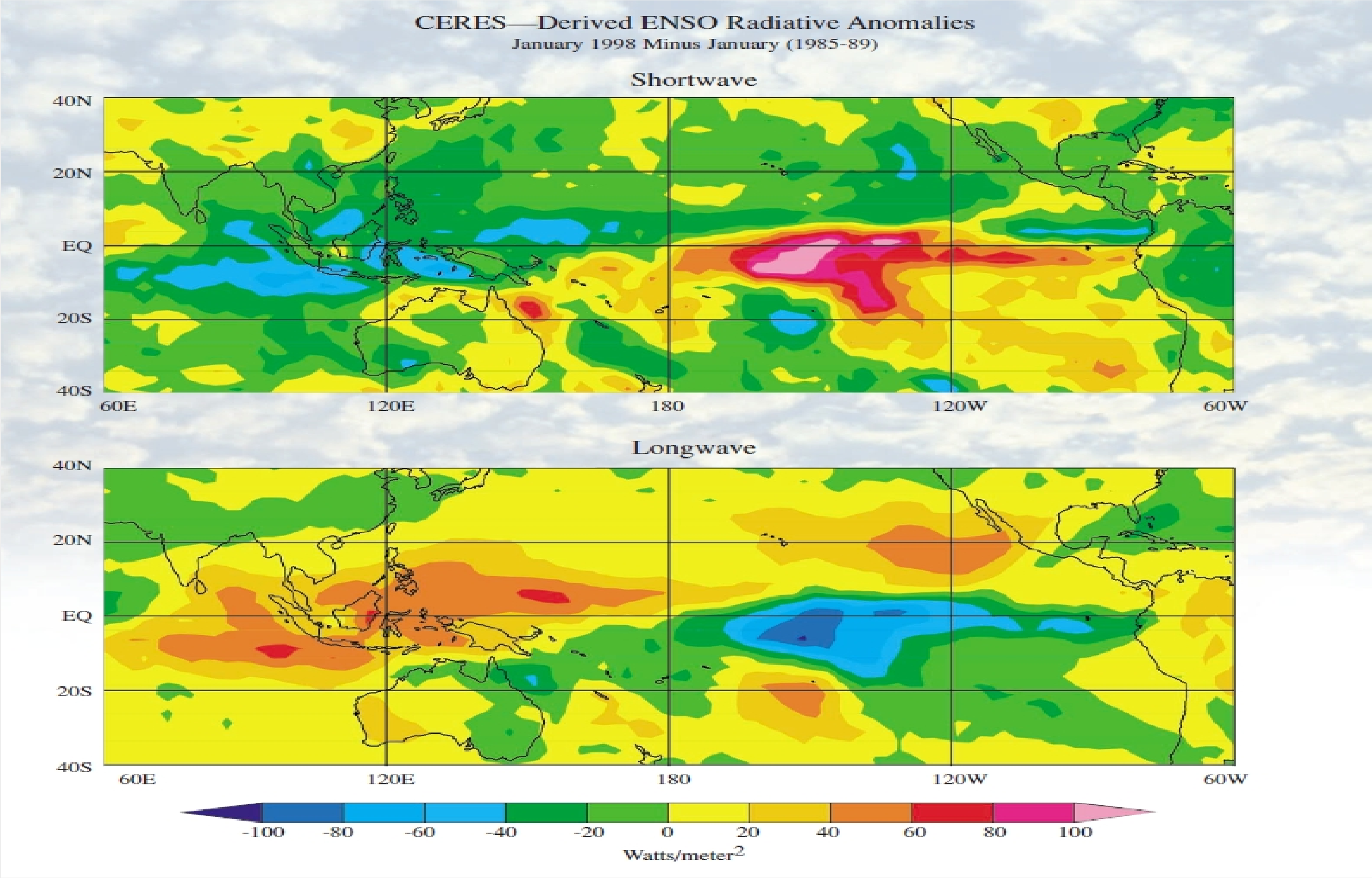
Budget (summing LW and SW)



In the areas of high and cold clouds, the low emissivity of clouds induces a positive effect on the budget (lowering the loss). This effect does not play in the areas of low clouds which emit at a temperature close to the ground temperature.
The reflexion of incoming radition by clouds generates a negative contribution to the budget. The high iced clouds are the more reflective. At mid laitudes the clouds limit the absorption by the ocean (negative role) and limit the reflection above the continents (positive effect). In the total budget the positive and negative effects almost compensate in the tropical region. The negative effects dominate at higher latitudes.

Influence of El Nino on the radiative budget at the top of the atmosphere

Anomalies: January 1998 (El Nino) – mean de January (1985-1989)

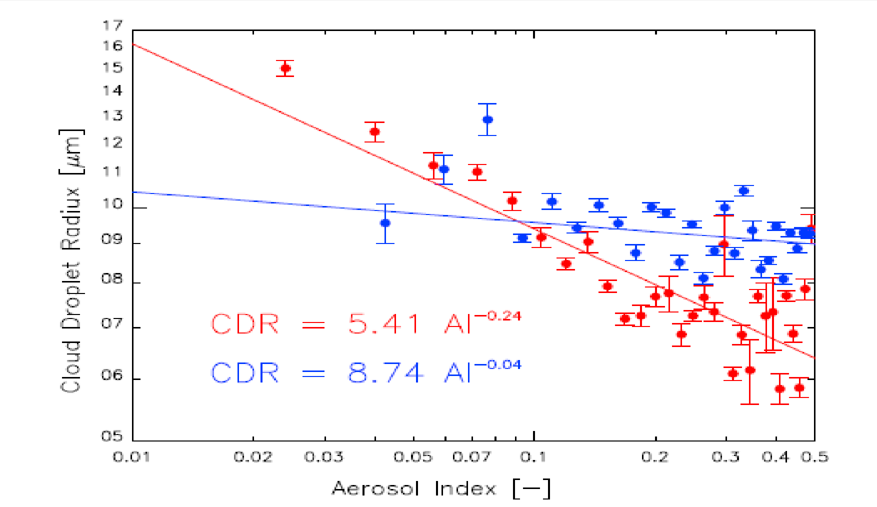


Radiative forcing positively counted in the upward direction

Indirect effect of aerosols

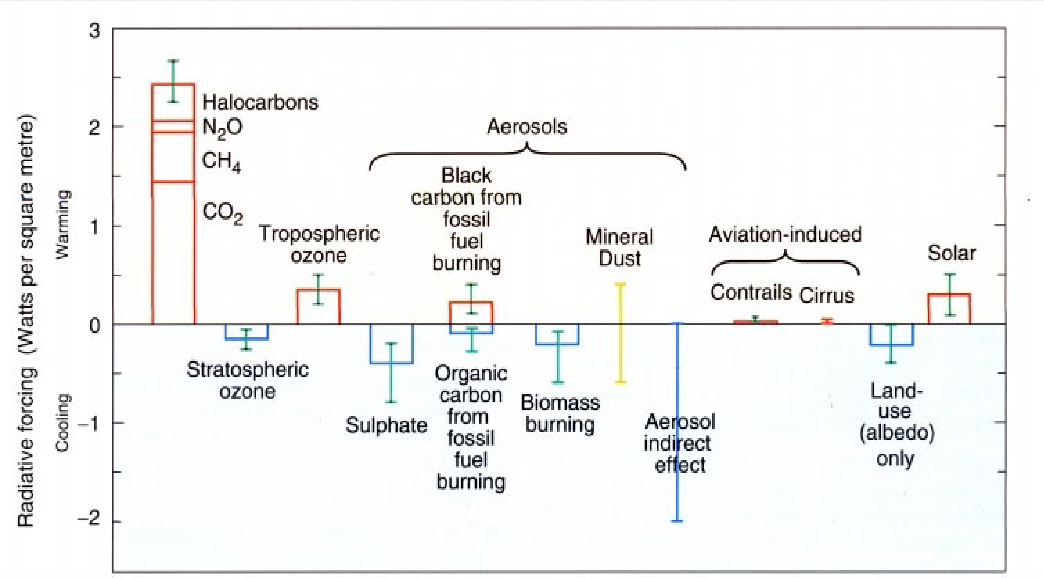
Costantino et Bréon, 2010

For a fixed amount of water, decrease of the size of droplets



Two cooling effects:

- Increase of the cloud albedo
- Increase of the life time of clouds



Climatology of the energy transfer in the atmosphere

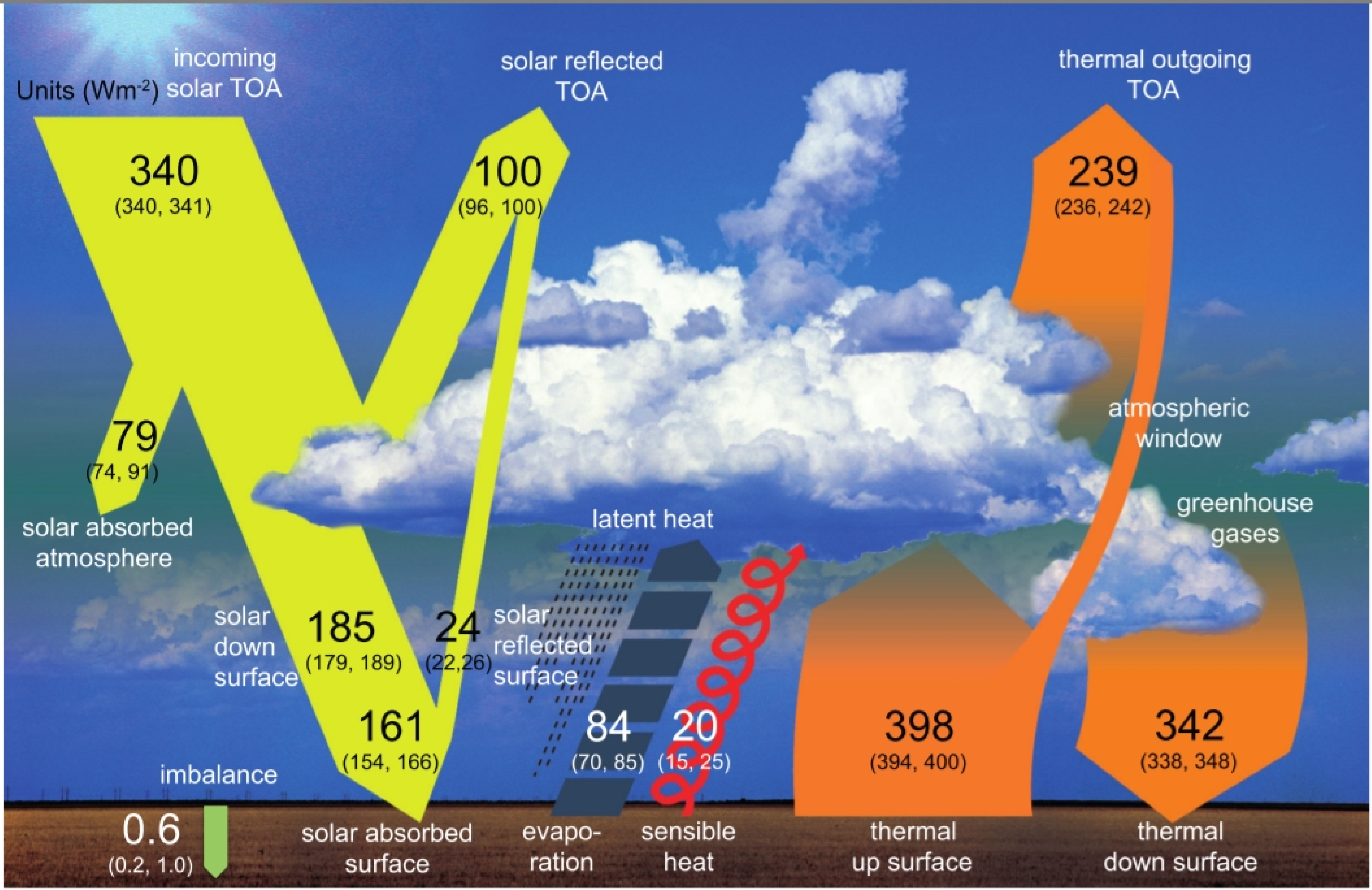


Figure 2.11: | Global mean energy budget under present-day climate conditions. Numbers state magnitudes of the individual energy fluxes in W m^{-2}, adjusted within their uncertainty ranges to close the energy budgets. Numbers in parentheses attached to the energy fluxes cover the range of values in line with observational constraints. (Adapted from Wild et al., 2013.)

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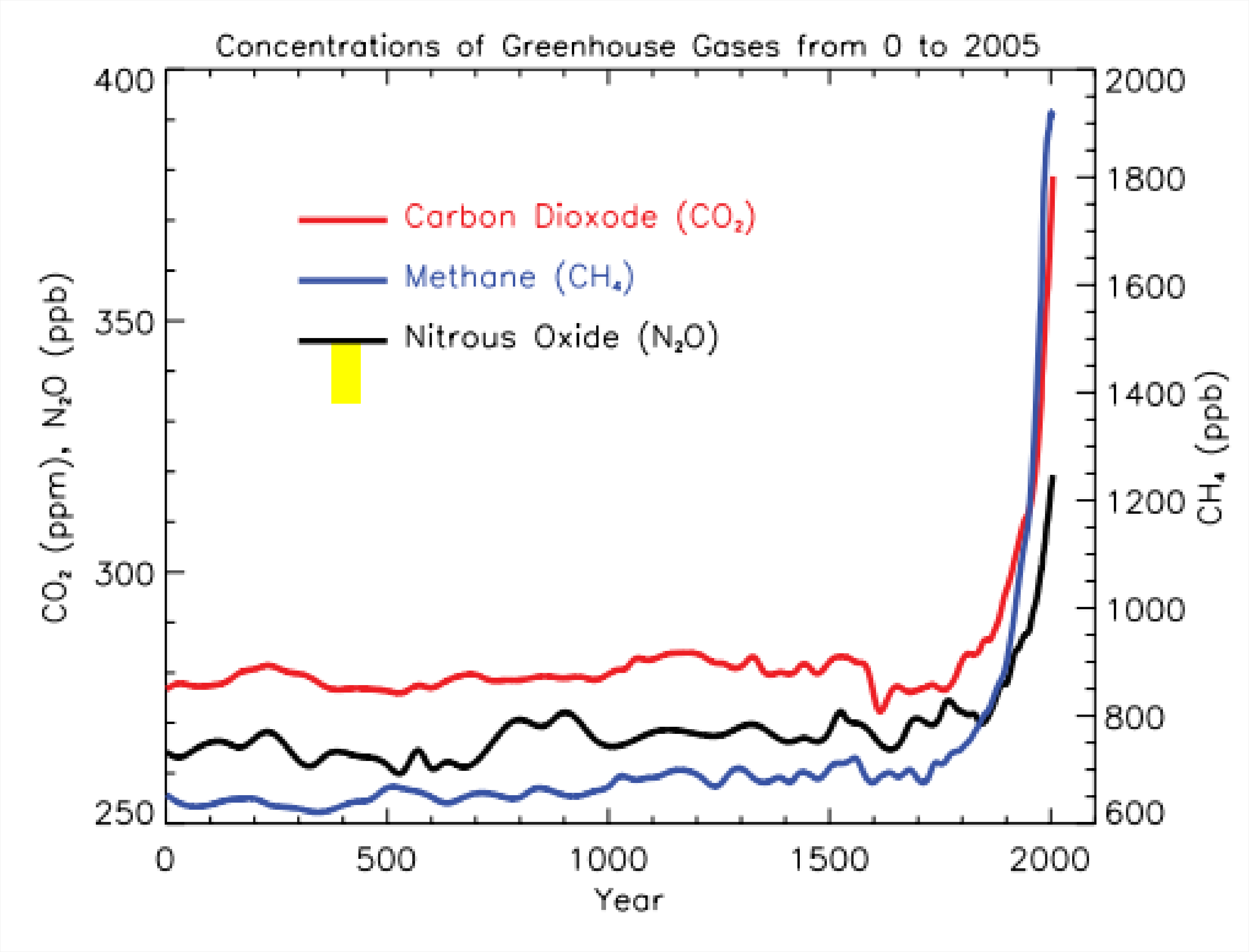
Increase of the concentration in greenhouse gases

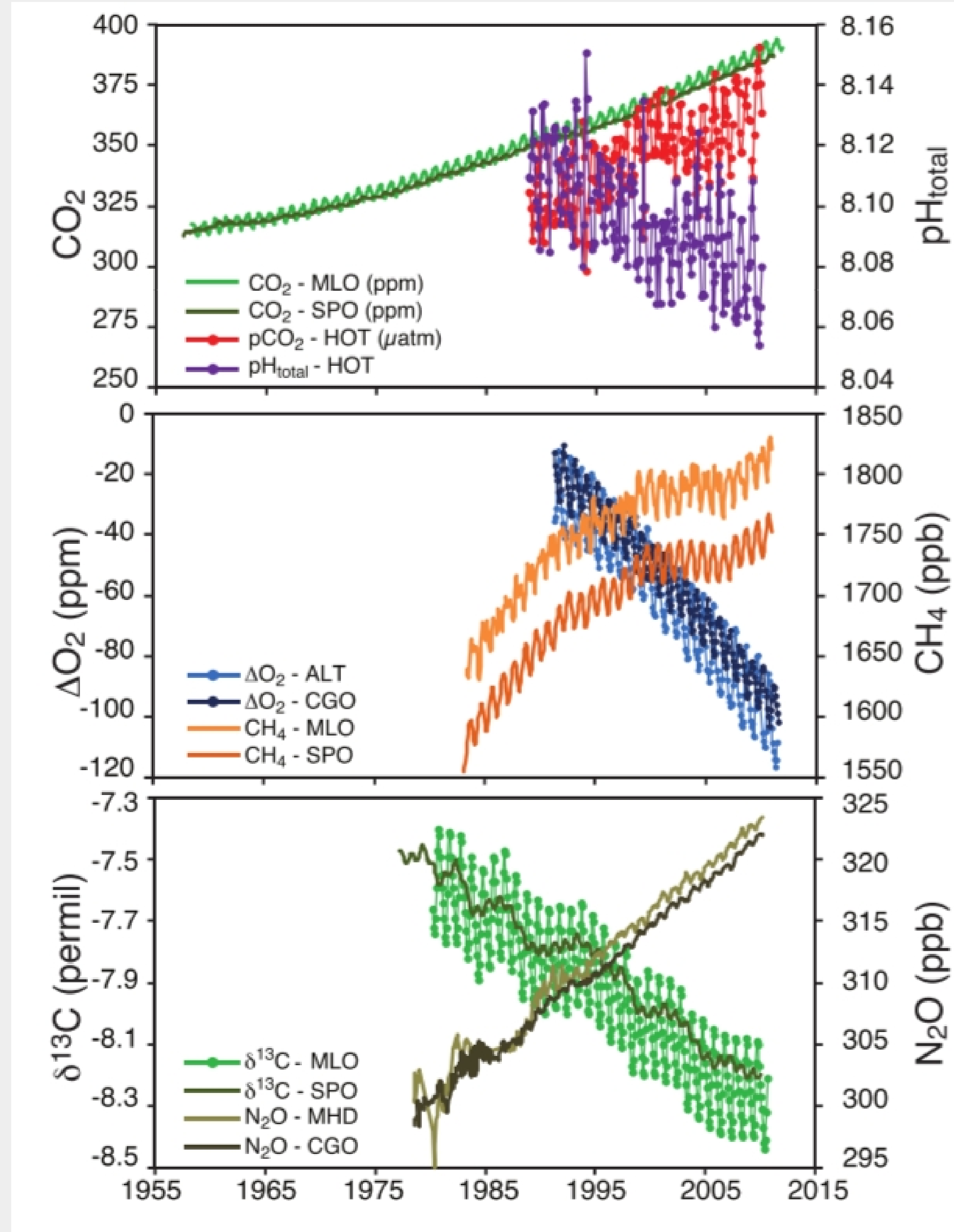


The anthropic origin of these variations is fully established

You might read sometimes on the web that CO2 has varied in the past and is coming from the ocean due to a warming where man takes no role.

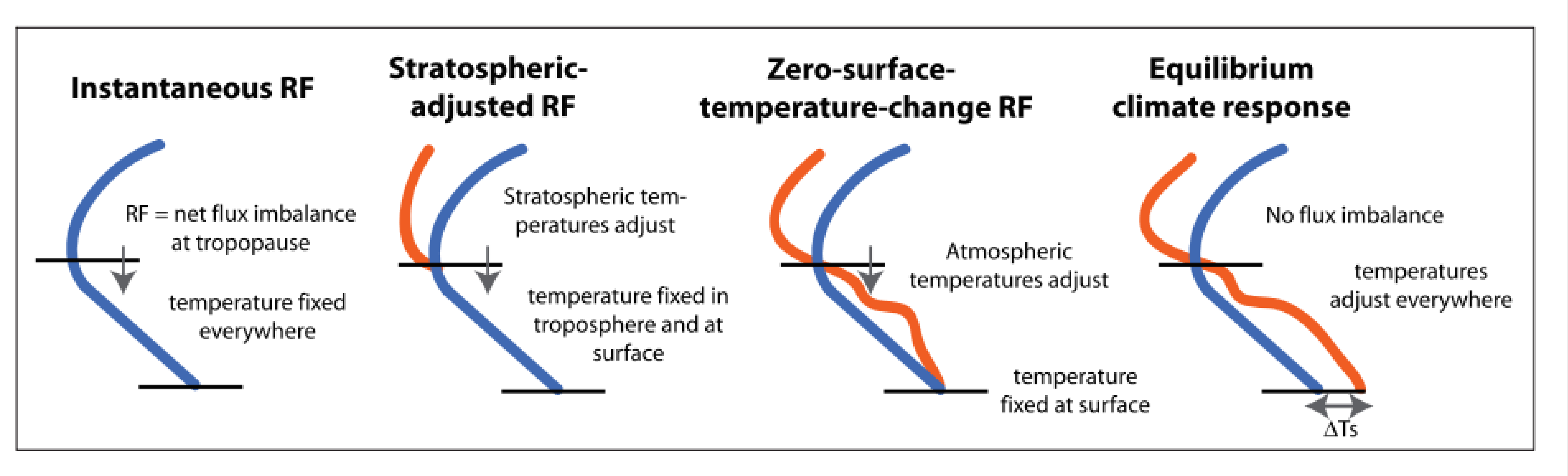
The last part is wrong: 1) CO2 and carbonate also increase in the ocean, reducing the pH, 2) Oxygen is reduced in the atmosphere due to combustion uptake, 3) The change in carbon isotopic ratio can only be explained by the combustion of depleted fossil fuel

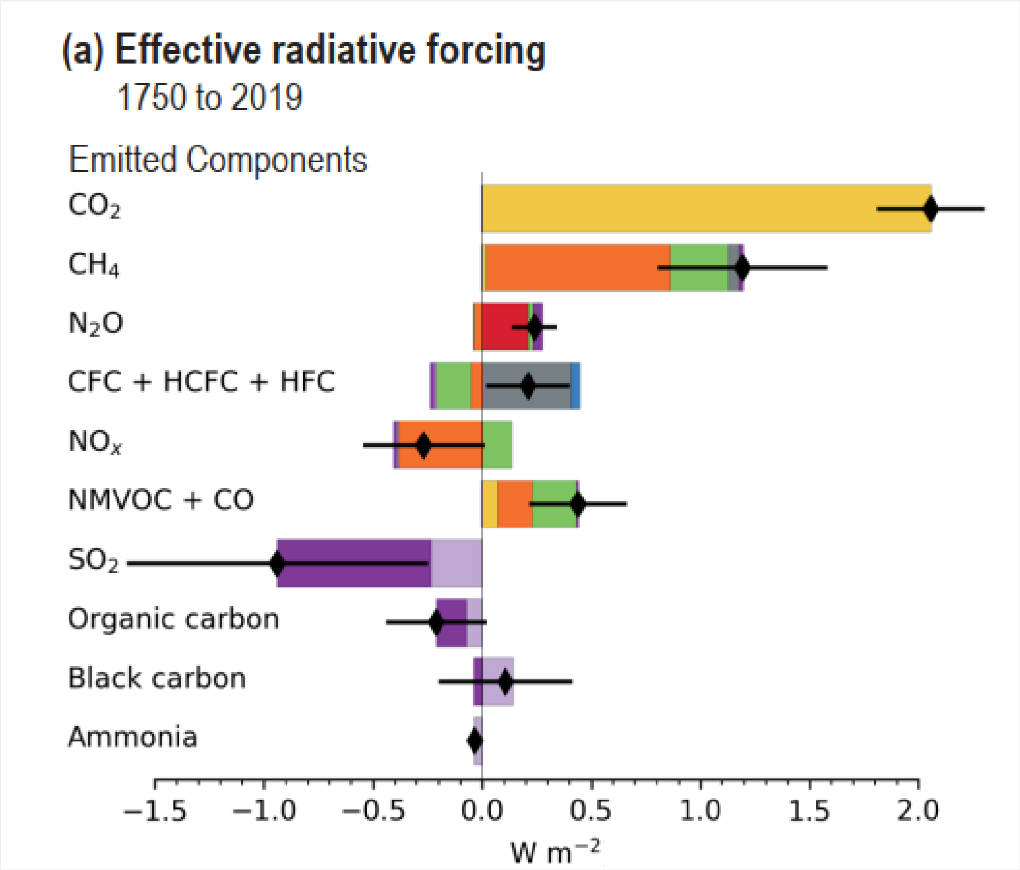
Figure TS.5 | Atmospheric concentration of CO_2, oxygen, $^{13}C/^{12}C$ stable isotope ratio in CO_2, as well as CH_4 and N_2O atmospheric concentrations and oceanic surface observations of CO_2 partial pressure (pCO_2) and pH, recorded at representative time series stations in the Northern and the Southern Hemispheres. MLO: Mauna Loa Observatory, Hawaii; SPO: South Pole; HOT: Hawaii Ocean Time-Series station; MHD: Mace Head, Ireland; CGO: Cape Grim, Tasmania; ALT: Alert, Northwest Territories, Canada. Further detail regarding the related Figure SPM.4 is given in the TS Supplementary Material. {Figures 3.18, 6.3; FAQ 3.3, Figure 1}

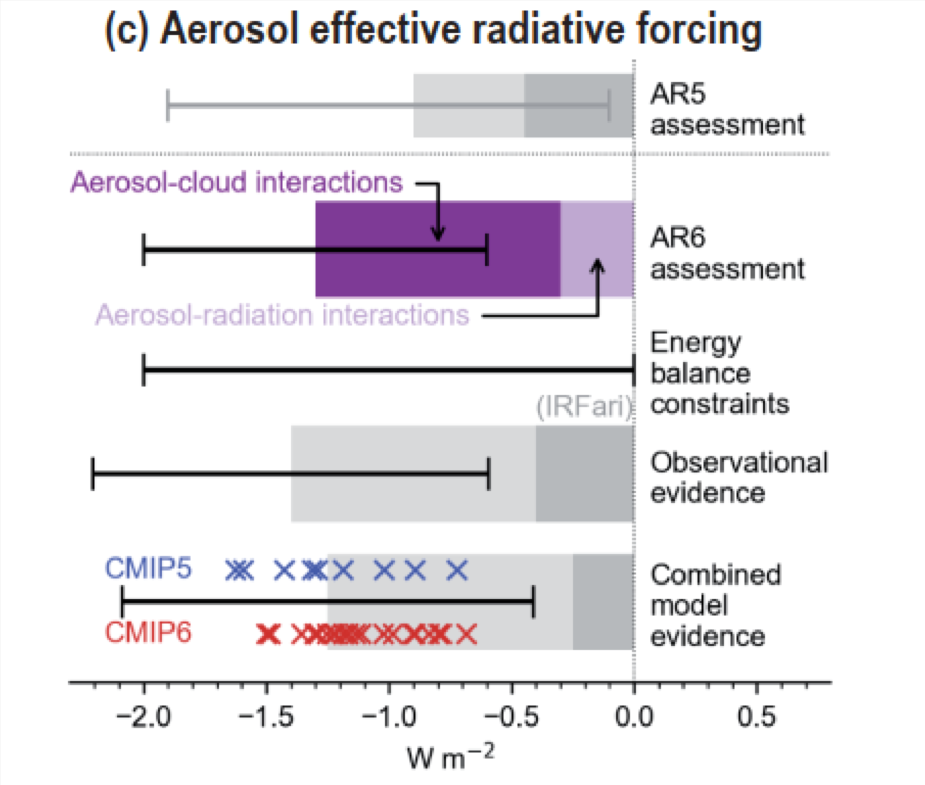


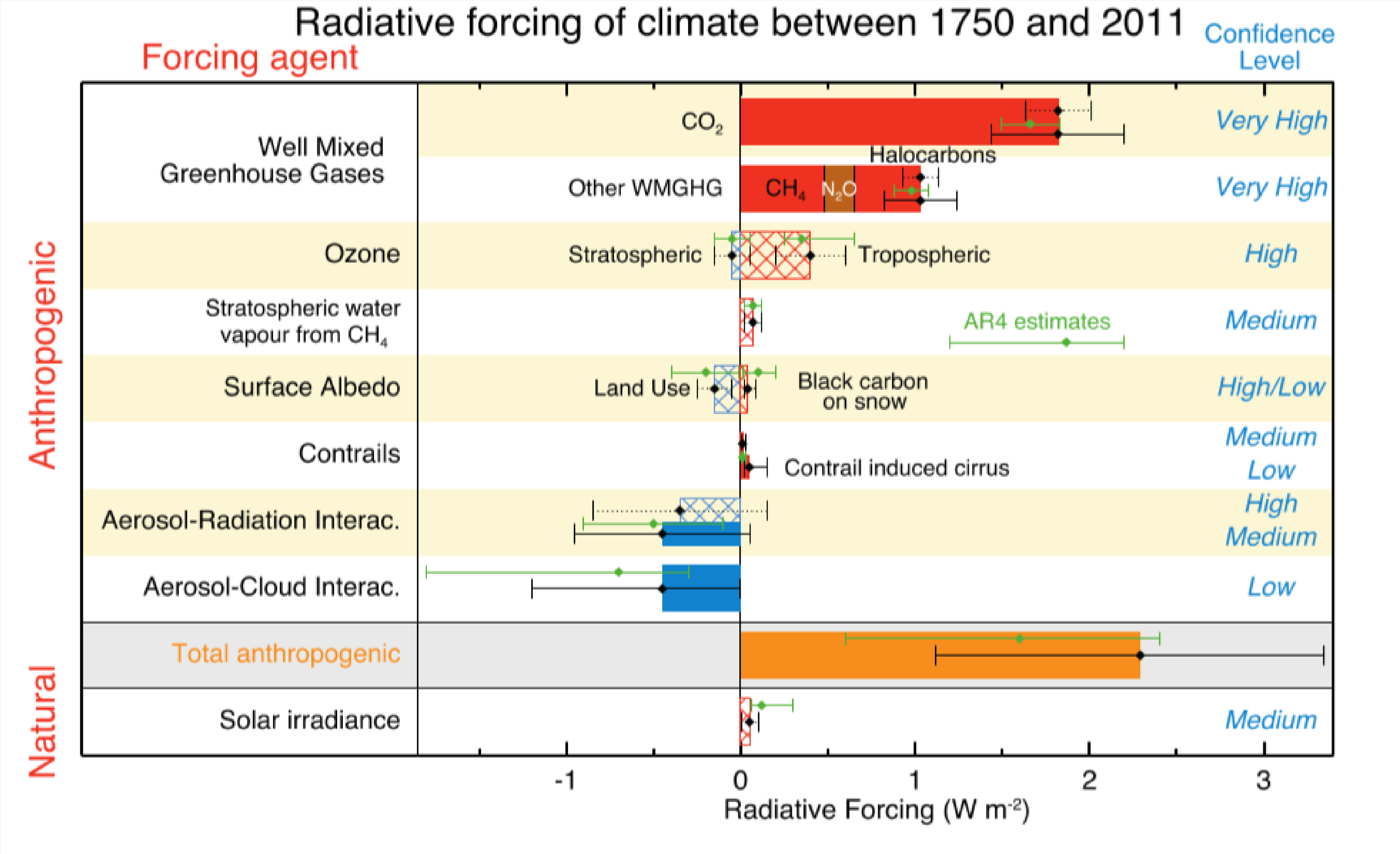
IPCC AR5 - TS

Calculation of the additional radiative forcing implied by a variation of teh concentration in greenhouse gases




(a) Effective radiative forcing
1750 to 2019
Emitted Components
CO2
CH4
N2O
CFC + HCFC + HFC
NOx
NMVOC + CO
SO2
Organic carbon
Black carbon
Ammonia
−1.5
−1.0
−0.5
0.0
0.5
1.0
1.5
2.0
W m−2


(c) Aerosol effective radiative forcing
AR5 assessment
Aerosol-cloud interactions
AR6 assessment
Aerosol-radiation interactions
Energy balance constraints
(IRFari)
Observational evidence
CMIP5
CMIP6
Combined model evidence
−2.0
−1.5
−1.0
−0.5
0.0
0.5
W m−2



Radiative forcing of climate between 1750 and 2011
Forcing agent
Confidence Level
Well Mixed Greenhouse Gases
CO$_2$
Very High
Halocarbons
Other WMGHG
CH$_4$
N$_2$O
Very High
Ozone
Stratospheric
Tropospheric
High
Stratospheric water vapour from CH$_4$
AR4 estimates
Medium
Surface Albedo
Land Use
Black carbon on snow
High/Low
Contrails
Contrail induced cirrus
Medium
Low
Aerosol-Radiation Interac.
High
Medium
Aerosol-Cloud Interac.
Low
Total anthropogenic
Solar irradiance
Medium
Anthropogenic
Natural
-1
0
1
2
3
Radiative Forcing (W m^{-2})

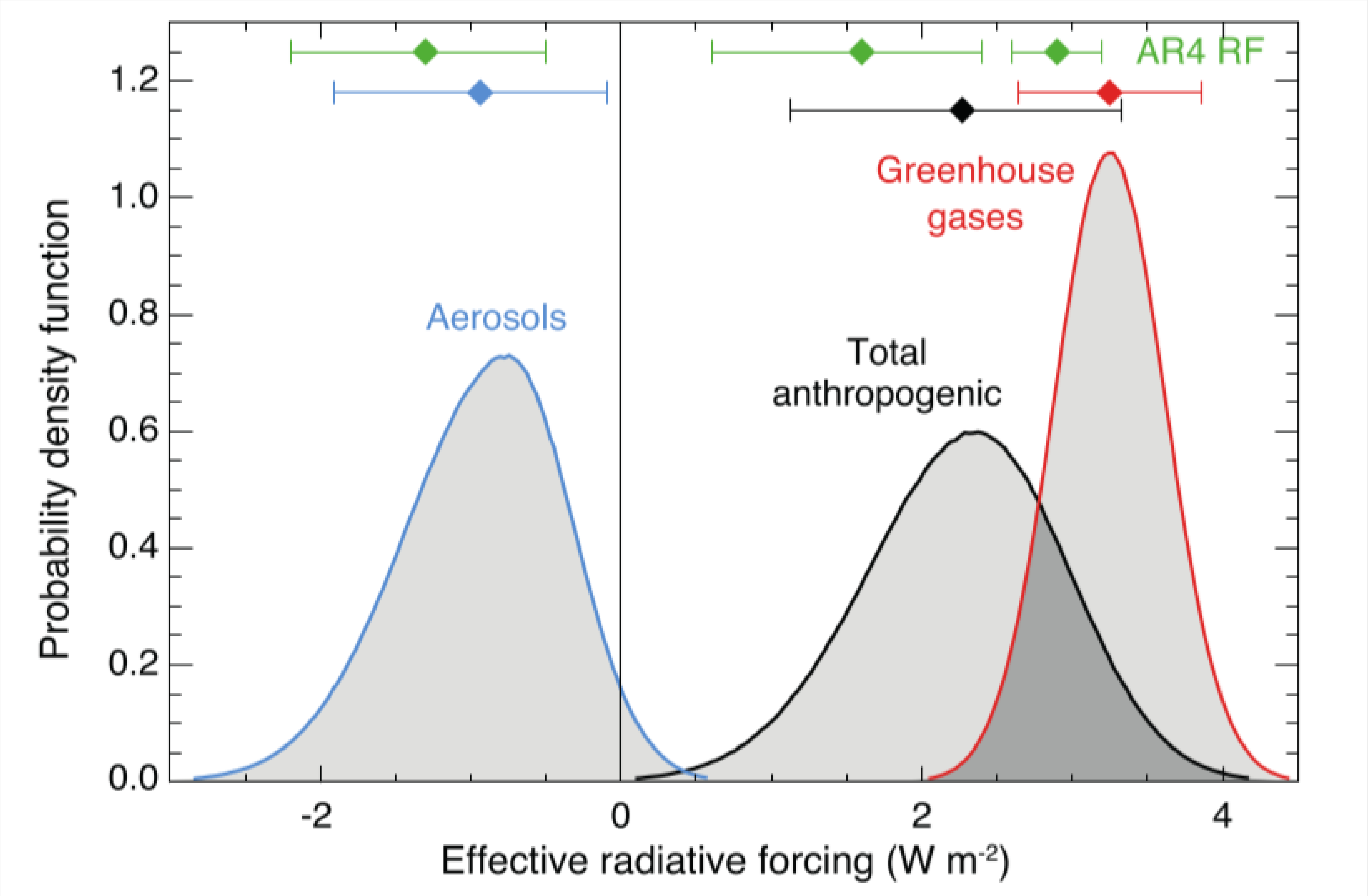


Figure TS.6 | Radiative forcing (RF) and Effective radiative forcing (ERF) of climate change during the Industrial Era. (Top) Forcing by concentration change between 1750 and 2011 with associated uncertainty range (solid bars are ERF, hatched bars are RF, green diamonds and associated uncertainties are for RF assessed in AR4). (Bottom) Probability density functions (PDFs) for the ERF, for the aerosol, greenhouse gas (GHG) and total. The green lines show the AR4 RF 90% confidence intervals and can be compared with the red, blue and black lines which show the AR5 ERF 90% confidence intervals (although RF and ERF differ, especially for aerosols). The ERF from surface albedo changes and combined contrails and contrail-induced cirrus is included in the total anthropogenic forcing, but not shown as a separate PDF. For some forcing mechanisms (ozone, land use, solar) the RF is assumed to be representative of the ERF but an additional uncertainty of 17% is added in quadrature to the RF uncertainty. {Figures 8.15, 8.16}