Physics of the atmosphere

LECTURE 1
RADIATIVE BUDGET

L3 etdiplôme de l'ENS Sciences de la Planète Terre
B. Legras, legras@lmd.ens.fr, http://www.lmd.ens.fr/legras
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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
I.1 Incoming solar radiation and terrestrial outgoing radiation

Incoming solar radiation
\[ S_0 = 1367 \text{ W m}^{-2} \]

Modulation of the solar constant (Froelich, 2009)

The Earth receives the solar flux and radiates back the energy to space.
I.2 Budget of the solar incoming radiation

- Earth's albedo: 30% reflected and scattered
- Incoming solar radiation: 100%
- 4% absorbed at the surface
- 20% absorbed by atmosphere and clouds
- 6% absorbed by atmosphere and clouds
- 51% absorbed at the surface
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 $a \ll \lambda$.
The diffused power varies as $\sim 1/\lambda^4$.

Lorenz-Mie diffusion for large parcels (droplets) such that $a \gg \lambda$.
The diffused power is independent of $\lambda$.
($a$: size of diffusing parcels)

Sky and cloud colour

Clean atmosphere:
- Midday: Sunlight is primarily in the blue region of the spectrum.
- Sunset: Sunlight is scattered, resulting in a reddish hue.

Dirty atmosphere:
- Midday: Sunlight is more dispersed, including red light.
- Sunset: The red light from the sun is scattered more efficiently, creating a deep red sky.

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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 emit and absorb 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.
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 $\nu$ 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.
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-example: laser
c) **Wien law**

\[ \lambda_{\text{max}} T = A = 2898 \, \mu \text{m} \, K \]

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

### II Kirchhoff law for the real systems

\( a_{\lambda} \) absorption, characteristic property of the material

For any system at thermal equilibrium, the emissivity is

\[ I_{\lambda}(T) = a_{\lambda} B_{\lambda}(T) \]

For a black body \( a_{\lambda} = 1 \) \( \text{et} \) \( I_{\lambda}(T) = B_{\lambda}(T) \)

For a real system \( a_{\lambda} < 1 \) \( \text{et} \) \( I_{\lambda}(T) < B_{\lambda}(T) \)

Any selective absorbant is a selective emettor

The emission at a given temperature is determined jointly by the properties of the system described by \( a_{\lambda} \) 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 lightest particles (electrons).

The lowest frequencies interact with the molecular structure (vibration, rotation).
I Introduction
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$_2$O molecule

$2.73 \mu m$  $2.65 \mu m$  $6.27 \mu m$

$v_1$  symmetric stretch  $v_3$  asymmetric stretch  $v_2$  bend

$x$  librations  $y$  

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

The absorption bands are composed of a multitude of individual rays.
Solar spectrum at ground level

Total absorption of the non reflected flux: ~ 30% (gas and clouds)
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
Comparison of a line by line radiative model with spectroscopic measurements of the descending thermal flux. Clear sky case of the ARM Great Plains station (Utah).
I Introduction

II Interactions of rayonnement with matter

III Gas absorption

IV Greenhouse effect

V Climatology of the radiative budget

VI Climate sensitivity
Incoming solar radiation
\( S_0 = 1367 \text{ W m}^{-2} \)

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

Albedo \( \alpha = 0.3 \)

Without greenhouse effect

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

Surface temperature \( T_e = 255 \text{ K} \)
Greenhouse effect

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

\( T_e = 255 \text{ K} \)

With an Atmosphere:
- \( T_s = 288 \text{ K} \)
- \( T_e = 255 \text{ K} \)

No Atmosphere Case:
- \( T_e = 255 \text{ K} \)
- \( T_s = 0 \text{ K} \)

Selective absorption of radiation by snow:
- Albedo = 0.9
- Short wave radiation

Earth's surface

Greenhouse gases

Cold

Warm

Daytime
Solar irradiance as measured at SIRTA

Descending infra-red flux as measured at SIRTA
SIRTA IPSL Ecole Polytechnique

**Dynamique - Turbulence :**
- Radar UHF
- T, H sol et air
- Sodar
- Lidar Doppler Vent
- Anémomètres Soniques

**Nuages - Aérosols :**
- Lidar Dépolarisation VIS-NIR
- Photomètre solaire
- Radiomètres solaire et IR
- Radar Doppler 95 GHz
- Radiomètre micro-onde
First model: Greenhouse effect for a one layer isothermal atmosphere

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

\[
\begin{align*}
S_0 (1-\alpha)/4 & \quad (1-\varepsilon)\sigma T_s^4 \\
\sigma T_s^4 & \quad \varepsilon \sigma T_A^4
\end{align*}
\]

- \( S_0 \): solar constant
- \( \alpha \): albedo
- \( \sigma \): Stefan-Boltzmann constant
- \( T_s \): surface temperature
- \( T_A \): temperature of the atmosphere
- \( \varepsilon \): absorption - emissivity of the atmosphere

\[
\begin{align*}
\varepsilon \sigma T_A^4 & = S_0 (1-\alpha)/4 \\
\sigma T_s^4 & = \varepsilon \sigma T_A^4
\end{align*}
\]

Budget of the atmosphere:

\[
\varepsilon \sigma T_s^4 - 2 \varepsilon \sigma T_A^4 = 0
\]

Surface budget:

\[
\frac{S_0}{4} (1-\alpha) + \varepsilon \sigma T_A^4 - \sigma T_s^4 = 0
\]

\[
T_s = \left( \frac{S_0 (1-\alpha)}{2 \sigma (2-\varepsilon)} \right)^{1/4} = T_e \left( \frac{2}{2-\varepsilon} \right)^{1/4}
\]

For the Earth: \( T_e = 255^\circ K, T_s = 303^\circ K = +30^\circ C \),

The difference is due to the "greenhouse" 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 $\epsilon = \epsilon_{\text{CO}_2} + \epsilon_{\text{H}_2\text{O}}$ and $\phi = 241 \text{ W m}^{-2}$, the incoming solar flux. $\epsilon_{\text{CO}_2}$ is fixed and depends of the emissions while $\epsilon_{\text{H}_2\text{O}}$ depends on the temperature. We write $\delta \epsilon = \delta \epsilon_{\text{CO}_2} + \frac{d \epsilon_{\text{H}_2\text{O}}}{dT_S} \delta T_S$

Hence

$$\frac{2\phi}{(2-\epsilon)^2} \left( \delta \epsilon_{\text{CO}_2} + \frac{d \epsilon_{\text{H}_2\text{O}}}{dT_S} \delta T_S \right) = 4\sigma T_S^3 \delta T_S$$

$$\frac{2\phi}{(2-\epsilon)^2} \delta \epsilon_{\text{CO}_2} = \left( 4\sigma T_S^3 - \frac{2\phi}{(2-\epsilon)^2} \frac{d \epsilon_{\text{H}_2\text{O}}}{dT_S} \right) \delta T_S$$

Doing a more detailed and realistic calculation:

3.7 W/m$^2$ for a doubling of CO2

3.2 W/m$^2$/K \hspace{1cm} 1.5 W/m$^2$/K

with $\Gamma$ 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.
The outgoing long-wave radiation to space is reduced by the presence of the atmosphere

Greenhouse effect
\[ G = \sigma T_s^4 - OLR \]

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 \((\sigma T_s^4)\) if the atmosphere were transparent to infrared radiation.
We abandon here the one-layer simplification and consider here a still gray but stratified 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 \( \kappa \) 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* \( \tau \) between O and A.

\[
\tau = \int_O^A \kappa \rho \, dz \quad \text{hence} \quad I = I_0 e^{-\tau}
\]

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 transfer equation is thus

\[
dI = -I \kappa \rho \, dz + B \kappa \rho \, dz
\]

\[
\frac{dI}{d\tau} = -I + B
\]
We consider now the ascending and descending fluxes $F^\wedge$ and $F^\vee$:

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

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 $\pi B$. (the argument is based on the ray profile and the fact they are saturated at the center and are transmitting 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, $\frac{dF^\wedge}{dz} - \frac{dF^\vee}{dz} = -\rho C_p [\frac{dT}{dt}]_{\text{rad}}$

We define also the *optical depth* $\chi$, 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
\[ B(T) = \pi^{-1} \sigma T^4 \]  

Optical thickness
\[ d\tau = \kappa \rho \, dl \]

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

Equation for upward and downward fluxes (diffusion is neglected for the thermal IR) using \( \tau \) 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
\[
F^\wedge(\tau) = \sigma T_s^4 e^{-\tau} + \int_0^\tau \sigma T^4(\tau') e^{-\tau + \tau'} \, d\tau' \\
F^\vee(\tau) = \int_\tau^\infty \sigma T_s^4 e^{-\tau'} \, d\tau'
\]

Outgoing flux at the top of the atmosphere
\[ OLR = \sigma T_s^4 e^{-\tau_\infty} + \int_0^{\tau_\infty} \sigma T^4(\tau') e^{-\tau_\infty + \tau'} \, d\tau' \]

Pour la Terre \( \tau_\infty \approx 4 \), pour Vénus \( \tau_\infty \approx 80 \)
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

\[ OLR = \sigma T_S^4 e^{-\tau_S} + \int_0^{\tau_S} \sigma T_S^4 e^{-\tau_S + \tau'} d\tau' = \sigma T_S^4 \]

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 troposphere but a consequence of this result is that the concentration of absorbing gases near the ground does not matter. In particular, we should not be misled by the domination of H\(_2\)O 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 \( \chi = \tau - \tau : T = T_s \left( \frac{\chi}{\tau} \right)^\gamma \)

\[
\text{OLR} = \sigma T_s^4 e^{-\tau_s} + \sigma T_s^4 \int_0^{\tau_s} \left( \frac{\chi}{\tau_s} \right)^4 e^{-\chi} d\chi
\]

\[
\text{OLR} = \sigma T_s^4 \left[ e^{-\tau_s} + \left( \Gamma(1+4\gamma,0) - \Gamma(1+4\gamma,\tau_s) \right) \tau_s^{-4\gamma} \right]
\]

For \( \tau_s \) large: \( T_s = \left( \frac{S_0}{4\sigma} (1-\alpha) \right)^{1/4} \Gamma(1+4\gamma,0)^{-1/4} \tau_s^\gamma \)

\[
\frac{\text{OLR}}{\sigma}
\]

Greenhouse effect

The greenhouse effect grows with the total optical thickness and depends on 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 \( \gamma \) par \( \gamma/(1+m) \) in the last formula.
Greenhouse effect in a stratified atmosphere

Net solar radiation $F_s$

Outgoing long-wave radiation $\text{OLR}=F_{ir}$

Doubling of CO2: +150m, -1K, -4W/m²

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

Return to equilibrium

Dufresnes, 2010
Emission spectrum of the Earth observed from space

Total atmospheric absorption

Unit: cm$^{-1}$
Cooling of the stratosphere when CO2 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 CO2).

At equilibrium

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

hence

\[ T_{STRATO} = \left( \frac{S_{O3} + \sigma T_E^4}{\varepsilon} \right)^{\frac{1}{4}} \]

If the concentration in CO2 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
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.
Greenhouse effect in an atmosphere in pure radiative equilibrium

If $\psi = F^\wedge + F^\vee$ et $\phi = F^\wedge - F^\vee$
then $\frac{d\psi}{d\chi} = \phi$ et $\frac{d\phi}{d\chi} = \psi - 2\pi B$.

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

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

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

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

Similarly:
$F^\wedge = \frac{\phi}{2}(\chi + 2)$ 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.

\[
\pi B_s = \frac{\phi}{2}(\chi_s + 2) = \sigma T_s^4
\]

$\chi_s = 0$ : 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 surface air temperature.
- the skin temperature is smaller than 255K, the Earth black-body temperature

\[
\chi_s = 0 
\]

\[
\phi
\]

$\rightarrow T_s \approx 336 K$
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Notice
- the ground temperature is larger than the surface air temperature.
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\[
\chi_s = 0 
\]

\[
\phi
\]
Radiative convective equilibrium calculated by including CO2, H2O and ozone

The stratosphere is (almost) in radiative equilibrium (basically: UV absorption by O3 and IR emission IR by CO2). In the whole troposphere, the radiative budget produces a net cooling. It is compensated by heating due to convective transport.

From Manabe & Strickler, JAS, 1966
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.
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 by 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

\[ \kappa = 10 \, m^2 \, kg^{-1} \]

leads to an absorption depth of 200m at 1000 hPa.

Widening of the rays by collision and Doppler effect

Lorentz and Voigt profiles

→ Absorption bands

Figure 4.7: The absorption coefficient vs. wavenumber for pure CO₂ 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.
Pour 300 ppmv de CO2, uniformément distribué, on a 3 kg/m²

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

Effect of the stratification and the shape of the bands

Note: for minor gases like CFC and CH₄, where saturation is not reached, the growth is linear → much larger impact per added molecule.
I Introduction
II Interactions of rayonnement with matter
III Gas absorption
IV Greenhouse effect
V Climatology of the radiative budget
VI Climate sensitivity
Radiative budget of the Earth

- Red: incoming radiation at the top of the atmosphere
- Blue: absorbed solar radiation
- Green: rIR radiation emitted to space

Excess in the low latitudes and deficit in the 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°.
Cloud effect:
During daytime, the dominating effect is to reflect the incoming radiation by the upper surface of clouds (→ cooling effect); during nighttime, only remains the blocking of IR emission by clouds (→ warming greenhouse effect).

Total radiative budget
(outgoing - incoming)

Visible channel
Radiative effect of the different type of clouds

LW radiative budget

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

Cloud radiative forcing at the top of the atmosphere

\[ CRF_{ir} = \epsilon \sigma (T_s^4 - Th^4) \]

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 albedo and low IR emission -> neutral or small effect
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 (strong 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

- **LW radiative forcing of clouds**
  
  \[ CRF_{\text{Ir}} = - \text{LW}_{\text{clear}} + \text{LW}_{\text{cloud}} \]

  \( \text{LW}_{\text{cloud}}: \) IR outgoing flux for a cloudy sky as measured by a satellite.
  \( \text{LW}_{\text{clear}}: \) IR outgoing flux for a clear sky as measured by a satellite or as calculated with radiative model.

- **SW radiative forcing of clouds**
  
  \[ CRF_{\text{sw}} = \text{SW}_{\text{cloud}} - \text{SW}_{\text{clear}} = [\alpha_{\text{cloud}} - \alpha_{\text{clear}}] \text{E}_0 \]

  \( \text{SW}_{\text{cloud}}: \) net incoming solar flux in a cloudy sky.
  \( \text{SW}_{\text{clear}}: \) net incoming solar flux in a clear sky.
  \( \text{E}_0: \) Incoming solar flux at the top of the atmosphere.
  \( \alpha_{\text{cloud}}: \) cloudy sky albedo.
  \( \alpha_{\text{clear}}: \) clear sky albedo.
Cloud radiative effect
(calculated on each pixel: \( \text{average of cloudy cases} - \text{average of clear sky cases} \), Positive flux counted in the descending direction)

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

Long waves (infra-red)

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 reflection of incoming radiation by clouds generates a negative contribution to the budget. The high iced clouds are the more reflective. At mid latitudes 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.

CRF \( \sim -20 \text{ W/m2} \)
Radiative forcing of clouds in July 2000
Influence of El Nino on the radiative budget at the top of the atmosphere


Radiative forcing positively counted in the upward direction
Indirect effect of aerosols

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.)
I Introduction
II Interactions of rayonnement with matter
III Gas absorption
IV Greenhouse effect
V Climatology of the radiative budget
VI Climate sensitivity
Increase of the concentration in greenhouse gases

The anthropic origin of these variations is fully established
Figure TS.5 | Atmospheric concentration of CO₂, oxygen, $^{13}$C/$^{12}$C stable isotope ratio in CO₂, as well as CH₄ and N₂O atmospheric concentrations and oceanic surface observations of CO₂ partial pressure (pCO₂) and pH, recorded at representative time series stations in the Northern and 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]
Calculation of the additional radiative forcing implied by a variation of the concentration in greenhouse gases
### Radiative forcing of climate between 1750 and 2011

<table>
<thead>
<tr>
<th>Forcing agent</th>
<th>Confidence Level</th>
</tr>
</thead>
<tbody>
<tr>
<td><strong>Anthropogenic</strong></td>
<td></td>
</tr>
<tr>
<td>Well Mixed Greenhouse Gases</td>
<td>Very High</td>
</tr>
<tr>
<td>Other WMGHG</td>
<td>Very High</td>
</tr>
<tr>
<td>Ozone</td>
<td>High</td>
</tr>
<tr>
<td>Stratospheric water vapour from CH₄</td>
<td>Medium</td>
</tr>
<tr>
<td>Surface Albedo</td>
<td>High/Low</td>
</tr>
<tr>
<td>Contrails</td>
<td>Medium Low</td>
</tr>
<tr>
<td>Aerosol-Radiation Interac.</td>
<td>High</td>
</tr>
<tr>
<td>Aerosol-Cloud Interac.</td>
<td>Medium</td>
</tr>
<tr>
<td>Total anthropogenic</td>
<td></td>
</tr>
<tr>
<td>Solar irradiance</td>
<td>Medium</td>
</tr>
</tbody>
</table>

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)
Figure TS.7 | Radiative forcing (RF) of climate change during the Industrial Era shown by emitted components from 1750 to 2011. The horizontal bars indicate the overall uncertainty, while the vertical bars are for the individual components (vertical bar lengths proportional to the relative uncertainty, with a total length equal to the bar width for a ±50% uncertainty). Best estimates for the totals and individual components (from left to right) of the response are given in the right column. Values are RF except for the effective radiative forcing (ERF) due to aerosol-cloud interactions (ERF−i) and rapid adjustment associated with the RF due to aerosol-radiation interactions (RF−i Rapid Adjust.). Note that the total RF due to aerosol-radiation interaction (~0.35 Wm^-2) is slightly different from the sum of the RF of the individual components (~0.33 Wm^-2). The total RF due to aerosol-radiation interaction is the basis for Figure SPM.5. Secondary organic aerosol has not been included since the formation depends on a variety of factors not currently sufficiently quantified.

The ERF of contrails includes contrail induced cirrus. Combining ERF−i = ~0.45 (~−1.2 to 0.0) Wm^-2 and rapid adjustment of ~−0.1 (~−0.3 to +0.1) Wm^-2 results in an integrated component of adjustment due to aerosols of ~−0.55 (~−1.33 to −0.06) Wm^-2

CFCs = chlorofluorocarbons, HFCs = hydrochlorofluorocarbons, PFCs = perfluorocarbons, NMVOC = Non-Methane Volatile Organic Compounds, BC = black carbon. Further detail regarding the related Figure SPM.5 is given in the TS Supplementary Material. (Figure B.17)