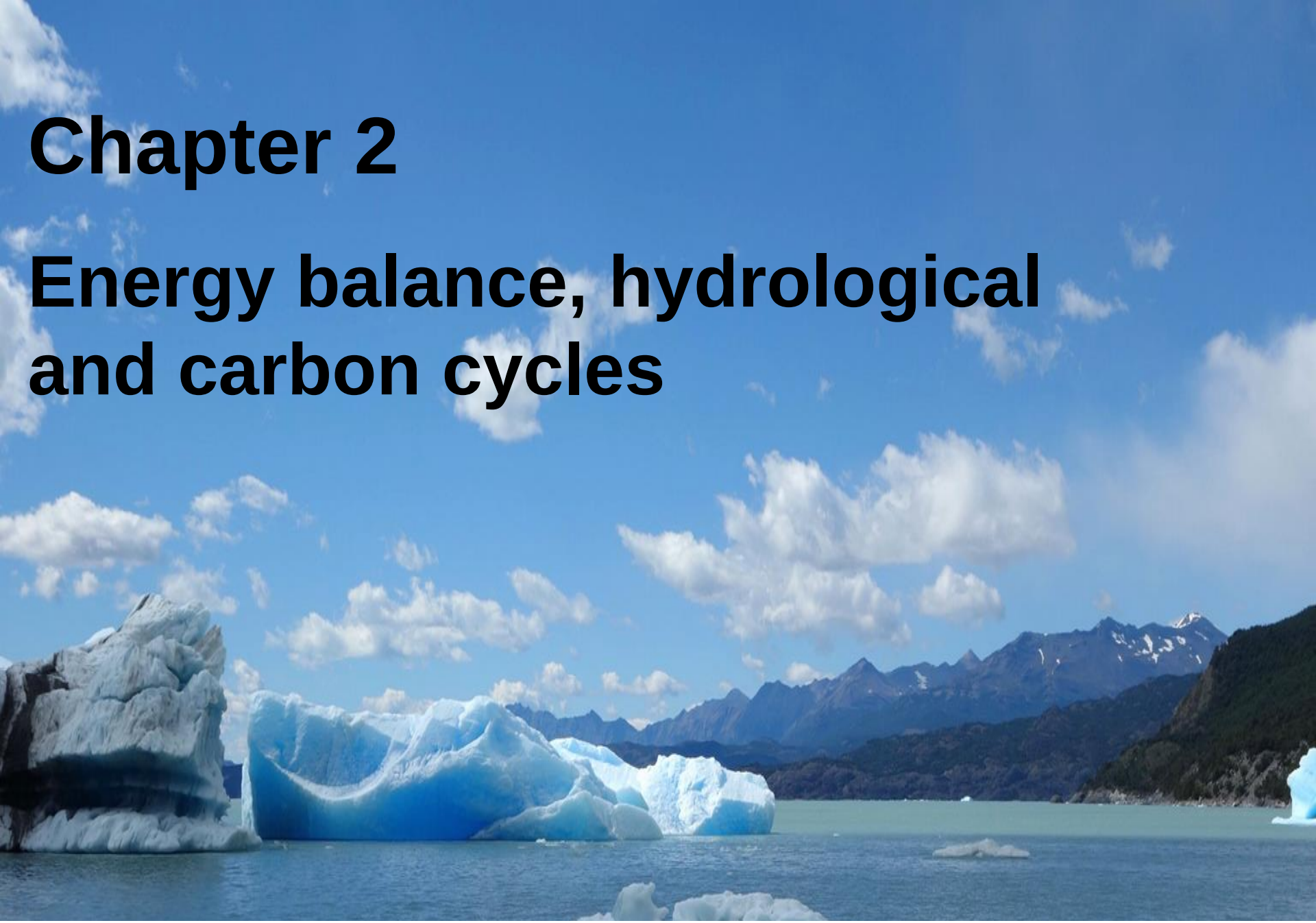
Chapter 2

Energy balance, hydrological and carbon cycles



Outline

Description of the global energy budget and of the exchanges of energy between the components of the climate system.

Spatial distribution for radiative fluxes and heat transport.

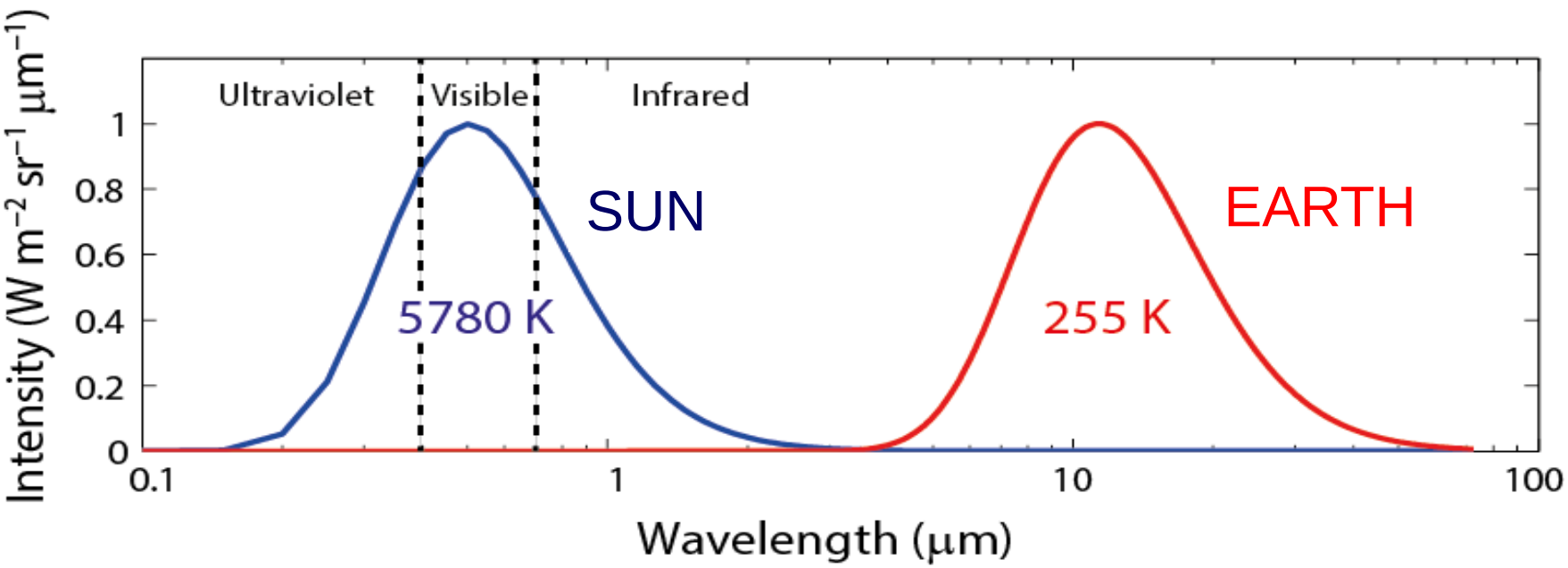
Description of the global water balance, local water balance and water transport.

Presentation of the carbon cycle, focusing on carbon dioxide and methane as they are major greenhouse gases.

The heat balance at the top of the atmosphere

At the top of the atmosphere, the energy received from the Sun (shortwave radiation) is balanced by the energy emitted by the Earth (longwave radiation).

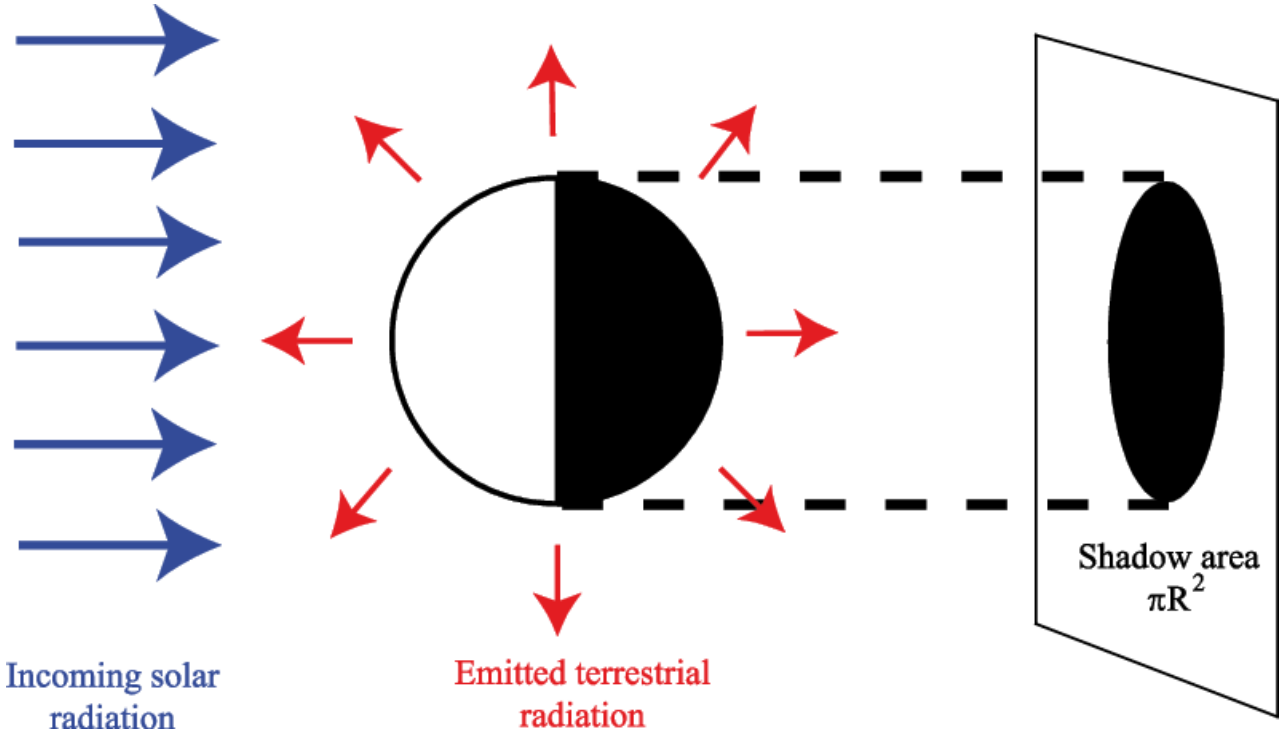
The total solar irradiance (TSI) is equal to 1360 W/m^2.



Normalized blackbody spectra for temperatures representative of the Sun (blue, temperature of 5780 K) and the Earth (red, temperature of 255 K).

The heat balance at the top of the atmosphere

On average, the total amount of incoming solar energy per unit of time outside the Earth's atmosphere is the TSI times the surface that intercepts the solar rays.



Schematic view of the energy absorbed and emitted by the Earth.

R, the Earth's radius, is equal to 6371 km.

The heat balance at the top of the atmosphere

The fraction of the incoming solar radiation that is reflected is called the albedo of the Earth or planetary albedo (α_p).

For present-day conditions it has a value of about 0.3.

The total amount of energy that is emitted by a 1 m^2 surface per unit of time by the Earth at the top of the atmosphere (A↑) can be computed following Stefan-Boltzmann's law:

$$A\uparrow = \sigma T_e^4$$

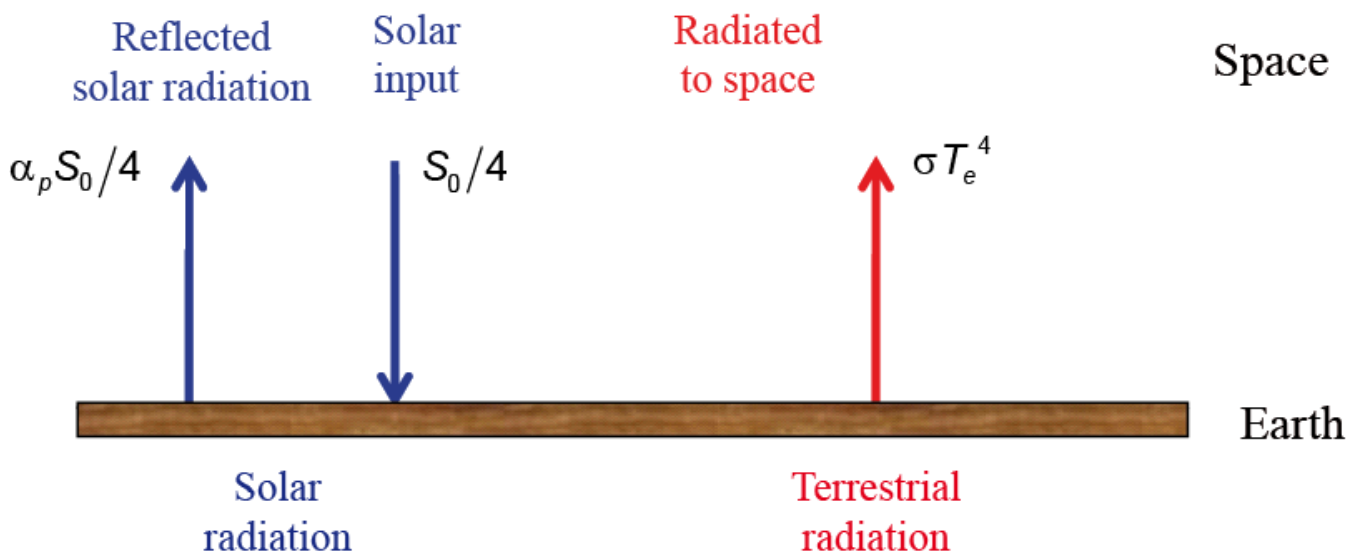
where T_e is the effective emission temperature of the Earth and σ is the Stefan Boltzmann constant (σ=5.67 10^{-8} W m^{-2} K^{-4}).

The heat balance at the top of the atmosphere

Heat balance of the Earth

Absorbed solar radiation = emitted terrestrial radiation

$$\pi R^2 \left(1-\alpha_p\right) S_0 = 4\pi R^2 \sigma T_e^4 \Rightarrow T_e = \left(\frac{1}{4\sigma}\left(1-\alpha_p\right) S_0\right)^{1/4}$$



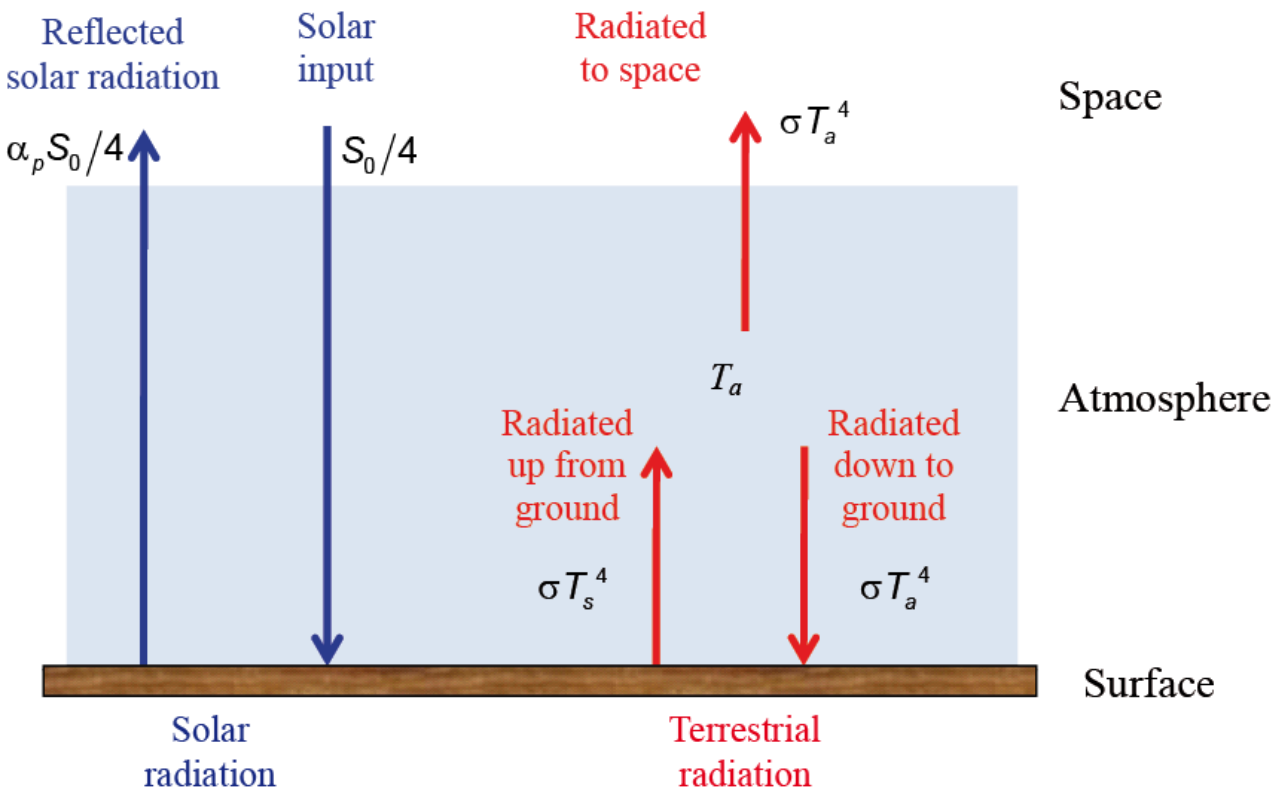
This corresponds to T_e=255 K (=-18°C).

The heat balance at the top of the atmosphere

Greenhouse effect

The atmosphere is nearly transparent to visible light.

The atmosphere is almost opaque across most of the infrared part of the electromagnetic spectrum because of some minor constituents (water vapour, carbon dioxide, methane and ozone).



Heat balance of the Earth with an atmosphere represented by a single layer totally transparent to solar radiation and opaque to infrared radiations.

The heat balance at the top of the atmosphere

Greenhouse effect

Representing the atmosphere by a single homogenous layer of temperature T_a, totally transparent to the solar radiation and totally opaque to the infrared radiations emitted by the Earth's surface, the heat balance at the top of the atmosphere is:

$$\frac{1}{4}\left(1-\alpha_p\right)S_0 = \sigma T_a^4 = \sigma T_e^4$$

The heat balance at the surface is:

$$\sigma T_s^{\,4} = \frac{1}{4}(1-\alpha_p)S_0 + \sigma T_a^{\,4}$$

This leads to:

$$T_s = 2^{1/4} T_e = 1.19 T_e$$

This corresponds to a surface temperature of 303K (30°C).

The heat balance at the top of the atmosphere

Greenhouse effect

A more precise estimate of the radiative balance of the Earth, requires to take into account

- the multiple absorption by the various atmospheric layers and reemission at a lower intensity as the temperature decreases with height.
- the strong absorption only in some specific ranges of frequencies which are characteristic of each component.

Furthermore, the contribution of non-radiative exchanges have to be included to close the surface energy balance.

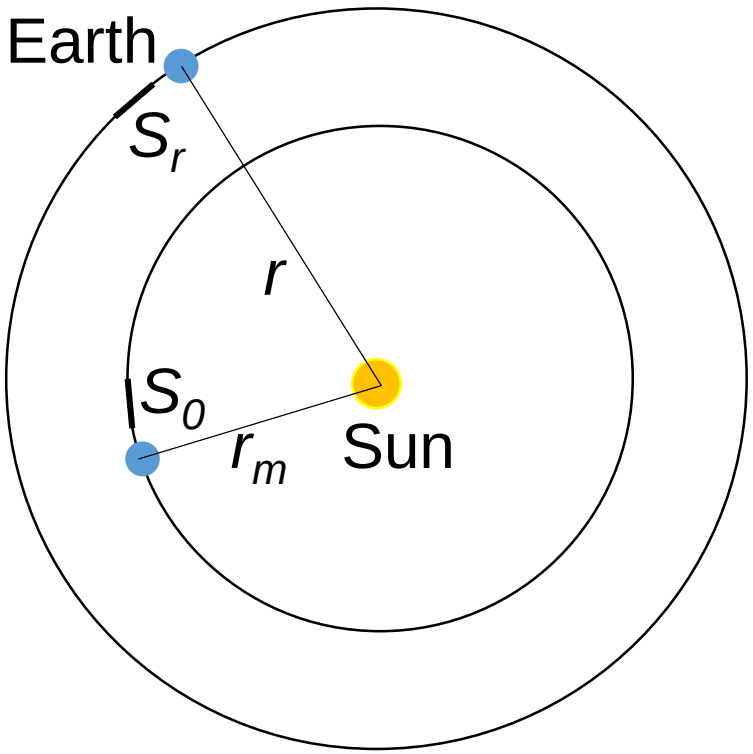
Present-day insolation at the top of the atmosphere

The irradiance at the top of the atmosphere is a function of the Earth-Sun distance.

Total energy emitted by the Sun at a distance r_m=
Total energy emitted by the Sun at a distance r

$$4\pi r_m^2 S_0 = 4\pi r^2 S_r$$

$$S_r = \frac{r_m^2}{r^2} S_0$$

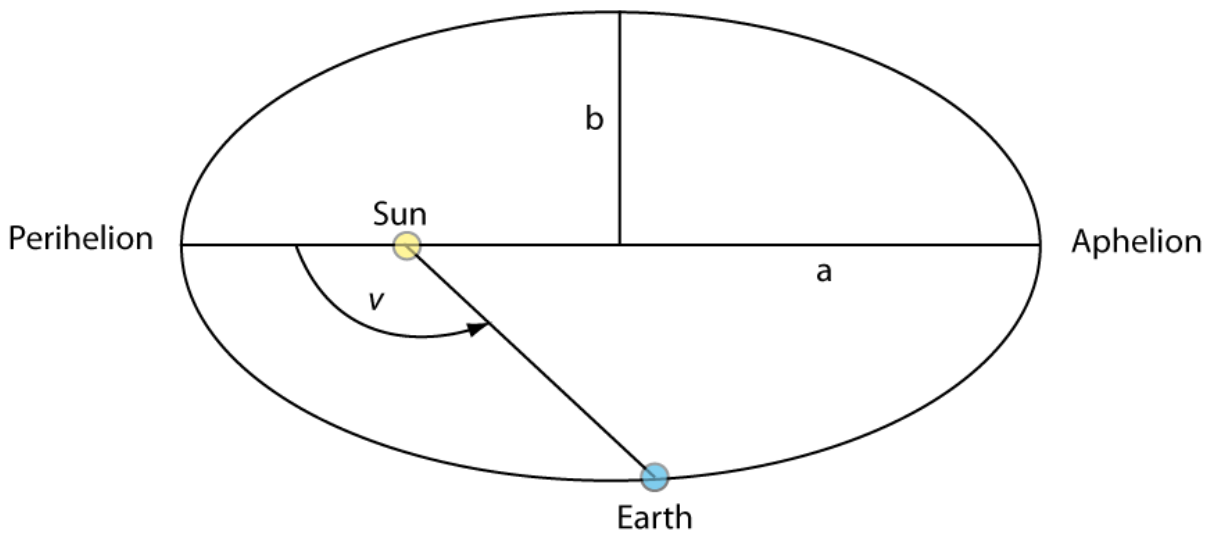


Present-day insolation at the top of the atmosphere

The Sun-Earth distance can be computed as a function of the position of the Earth on its elliptic orbit :

$$r = \frac{a\left(1 - ecc^2\right)}{1 + ecc\cos v}$$

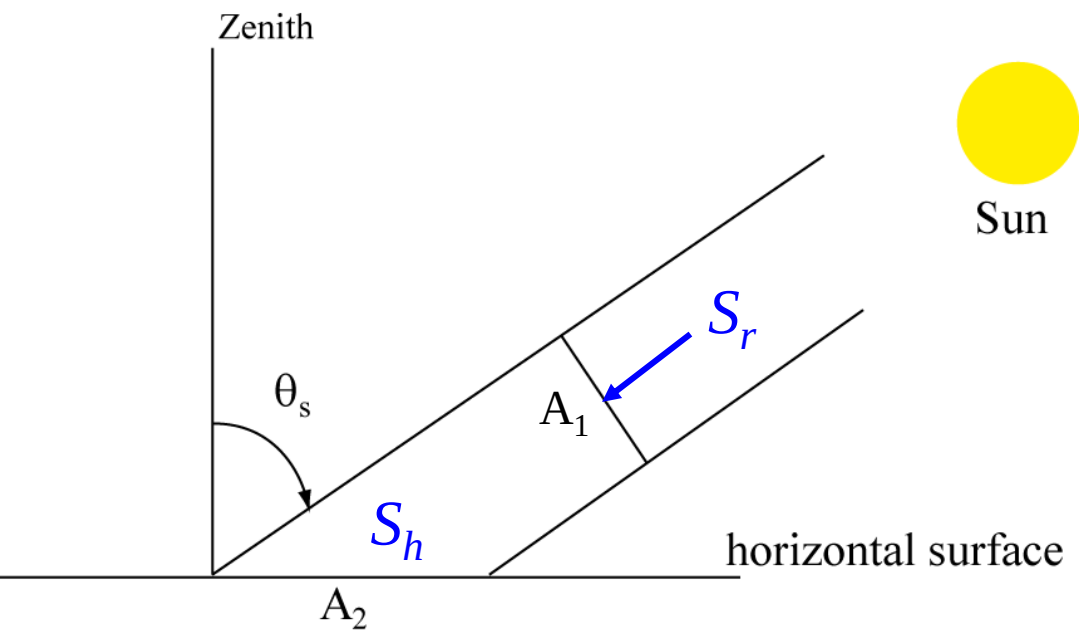
v is the true anomaly, *a*, half of the major axis, and *ecc* the eccentricity.



Schematic representation of the Earth's orbit around the Sun. The eccentricity has been strongly amplified for the clarity of the drawing.

Present-day insolation at the top of the atmosphere

The insolation on a unit horizontal surface at the top of the atmosphere (S_h) is proportional to the angle between the solar rays and the vertical.



energy crossing A_1
=energy reaching A_2

$$S_r A_1 = S_h A_2$$

$$= S_h A_1 / \cos\theta_s$$

$$S_h = S_r \cos\theta_s$$

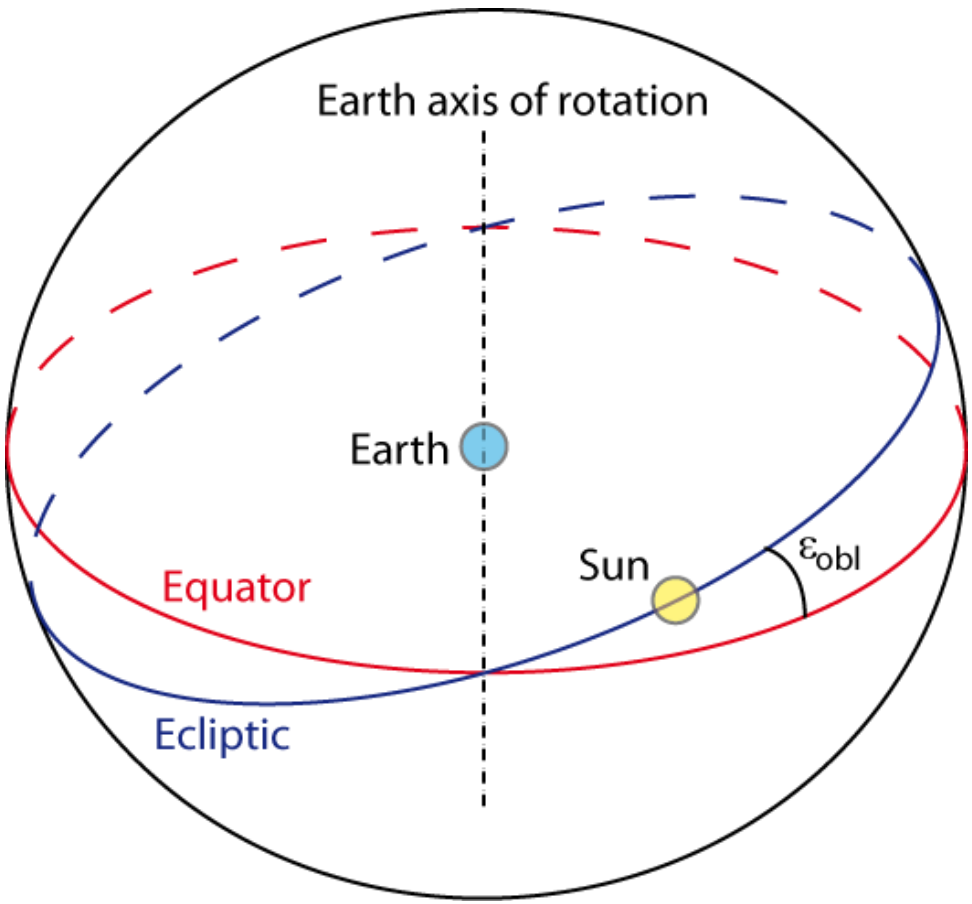
θ_s is the solar zenith distance

Present-day insolation at the top of the atmosphere

The solar zenith distance depends on the obliquity.

The obliquity, ε_{obl} , is the angle between the ecliptic plane and the celestial equatorial plane.

The obliquity is at the origin of the seasons.



Representation of the ecliptic and the obliquity ε_{obl} in a geocentric system.

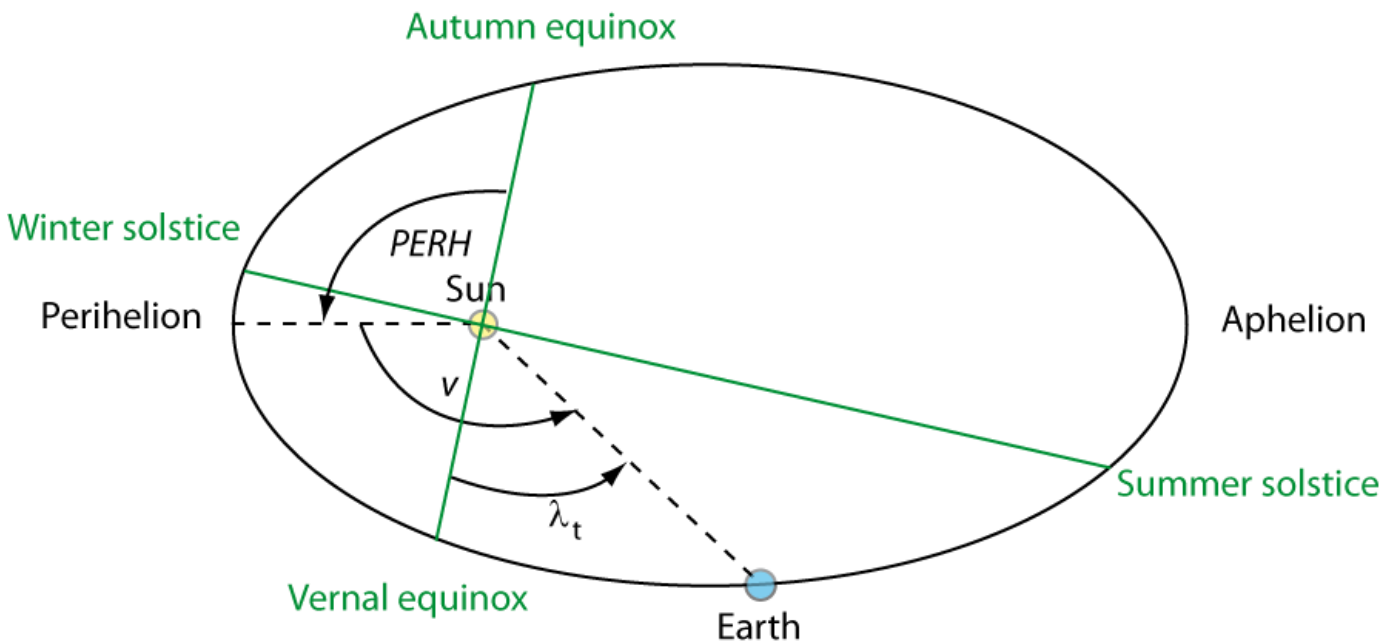
Presently

ε_{obl} =23°27'

Present-day insolation at the top of the atmosphere

The solar zenith distance depends on the position (true longitude λ_t) relative to the vernal equinox.

The vernal equinox corresponds to the intersection of the ecliptic plane with the celestial equator when the Sun "apparently" moves from the austral to the boreal hemisphere.



Representation of the true longitudes and the seasons in the ecliptic plane.

Present-day insolation at the top of the atmosphere

The solar zenith distance depends on the latitude (ϕ) and on the hour of the day (*HA*, the hour angle).

$$\cos\theta_s = \sin\phi \sin\delta + \cos\phi \cos\delta \cos HA$$

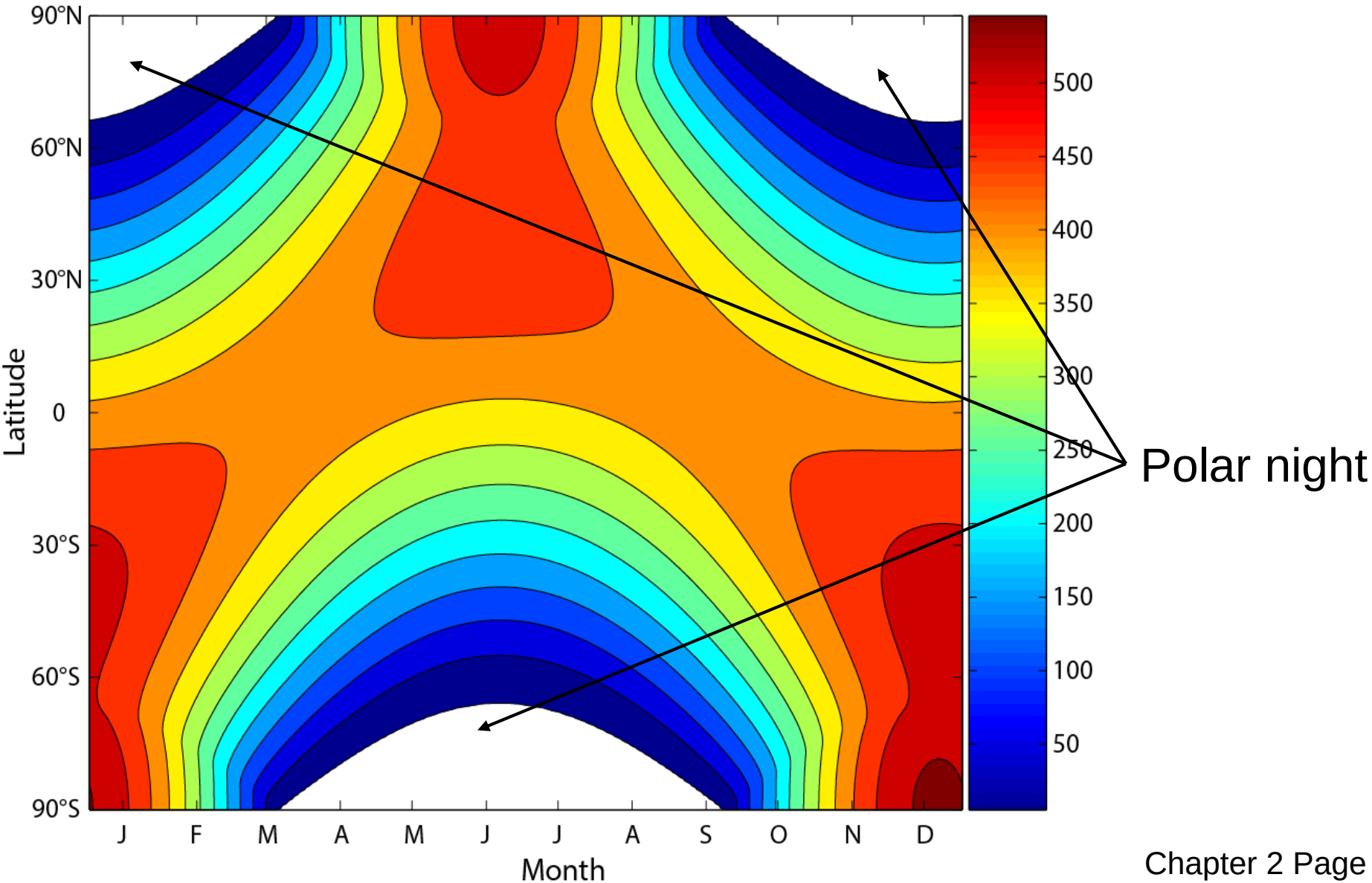
δ is the solar declination. It is related to the true longitude or alternatively to the day of the year.

$$\sin\delta = \sin\lambda_t \sin\varepsilon_{obl}$$

Those formulas can be used to compute the instantaneous insolation, the time of sunrise, of sunset as well as the daily mean insolation.

Present-day insolation at the top of the atmosphere

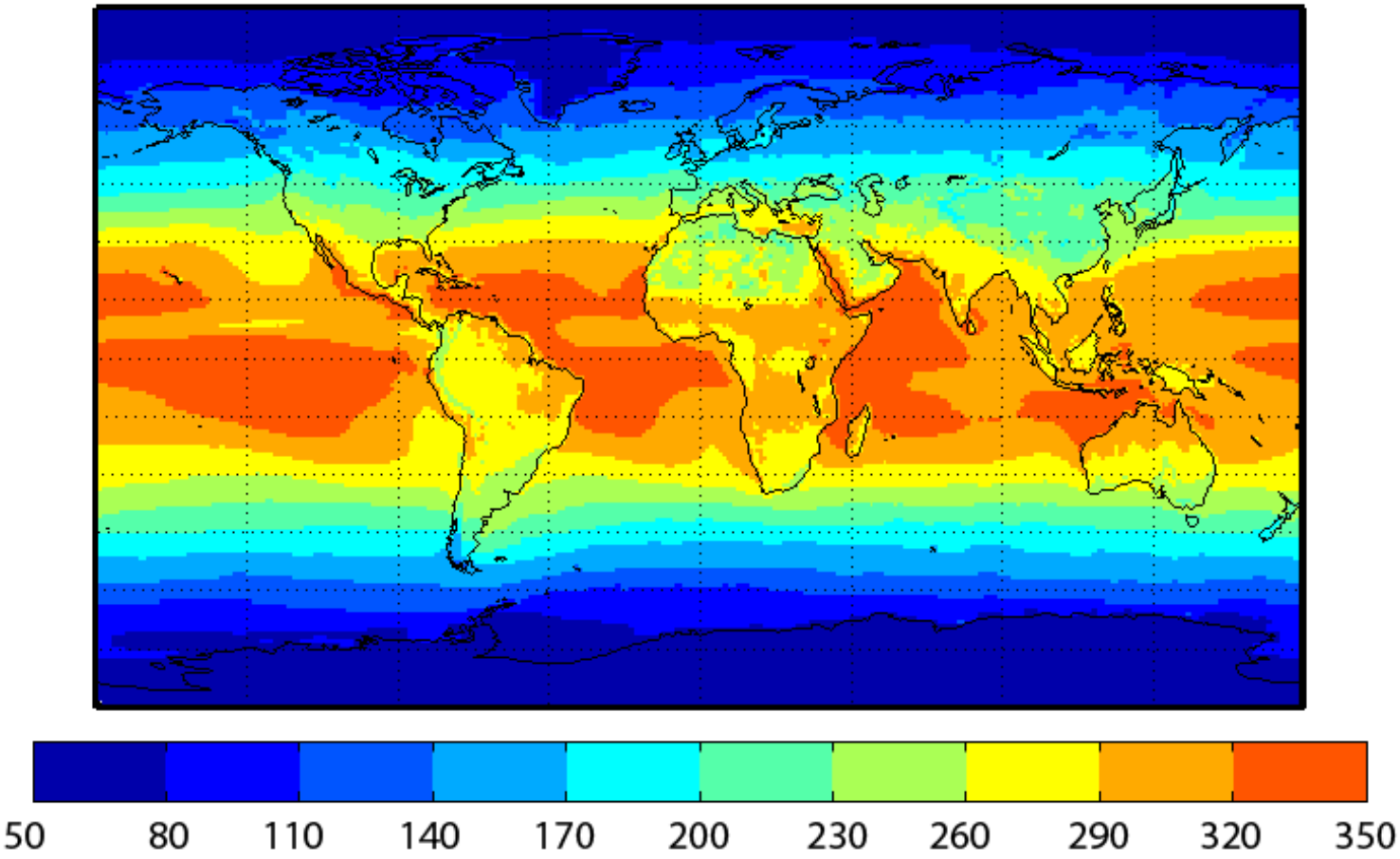
Daily mean insolation on an horizontal surface (W m^{-2}).



Radiative balance at the top of the atmosphere

Geographical distribution

Annual mean net solar flux at the top of the atmosphere (Wm^{-2})

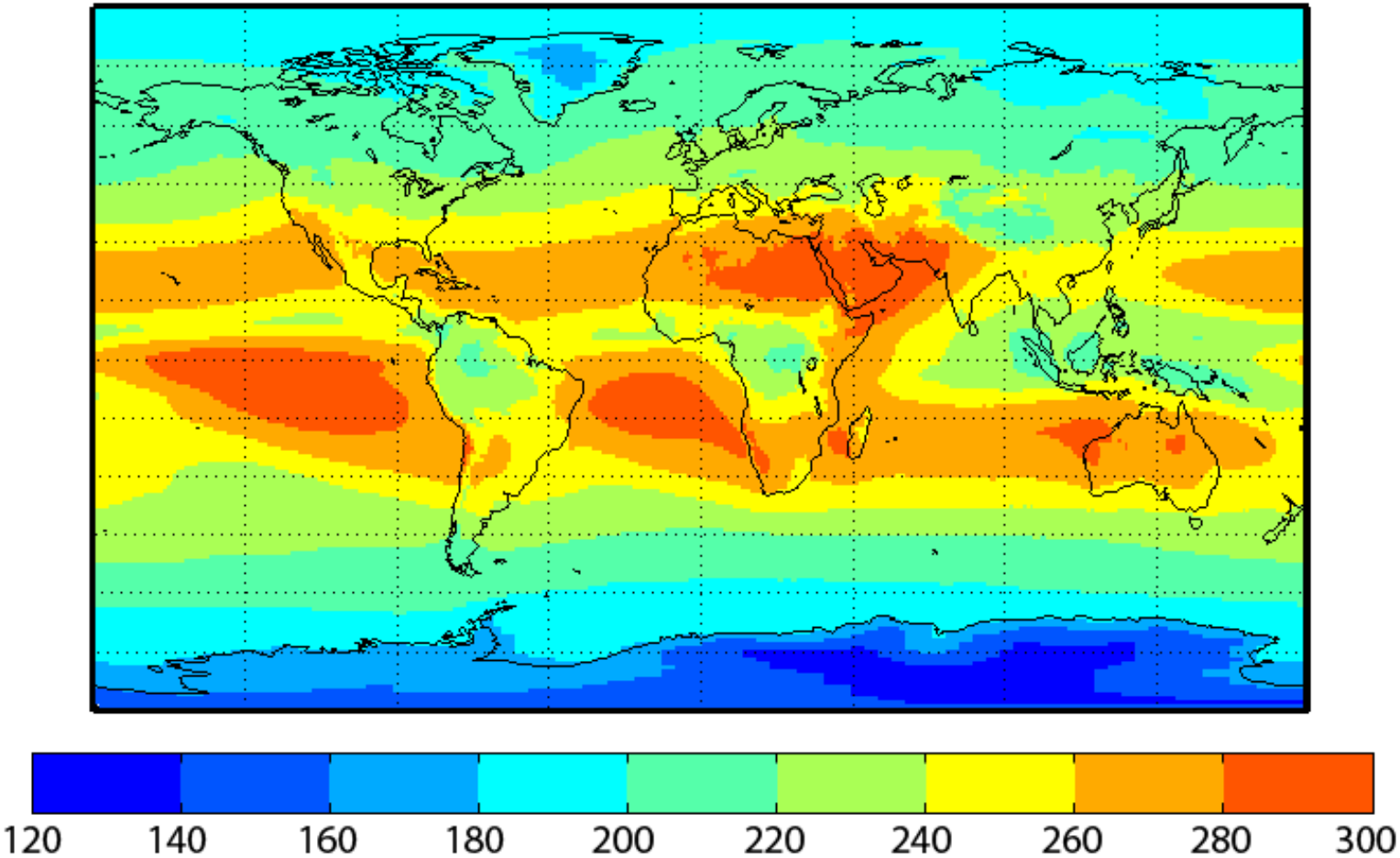


It is a function of the insolation and of the albedo.

Radiative balance at the top of the atmosphere

Geographical distribution

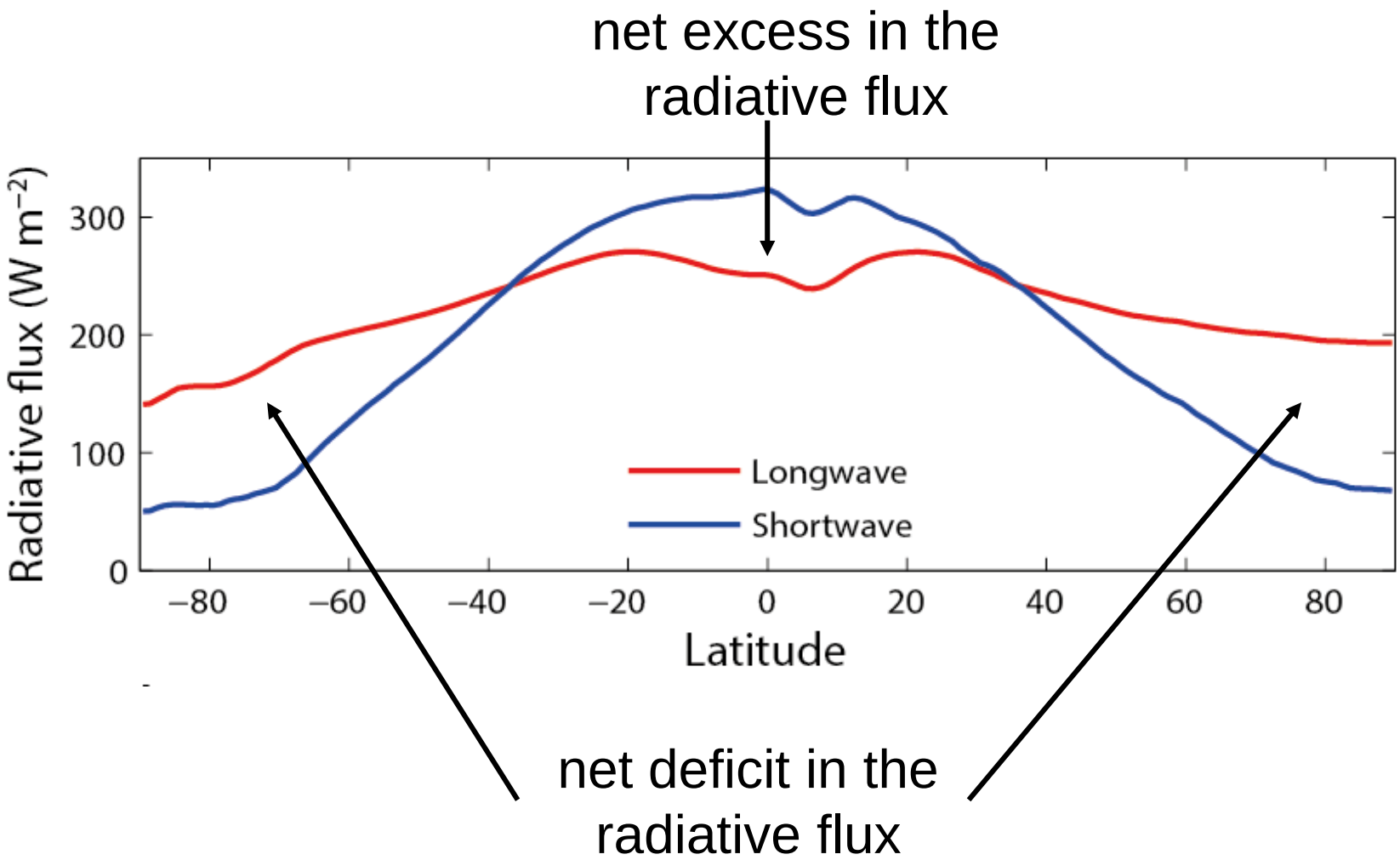
Net annual mean outgoing longwave flux
at the top of the atmosphere (Wm^{-2})



It is a function of the temperature and of the properties of the atmosphere.

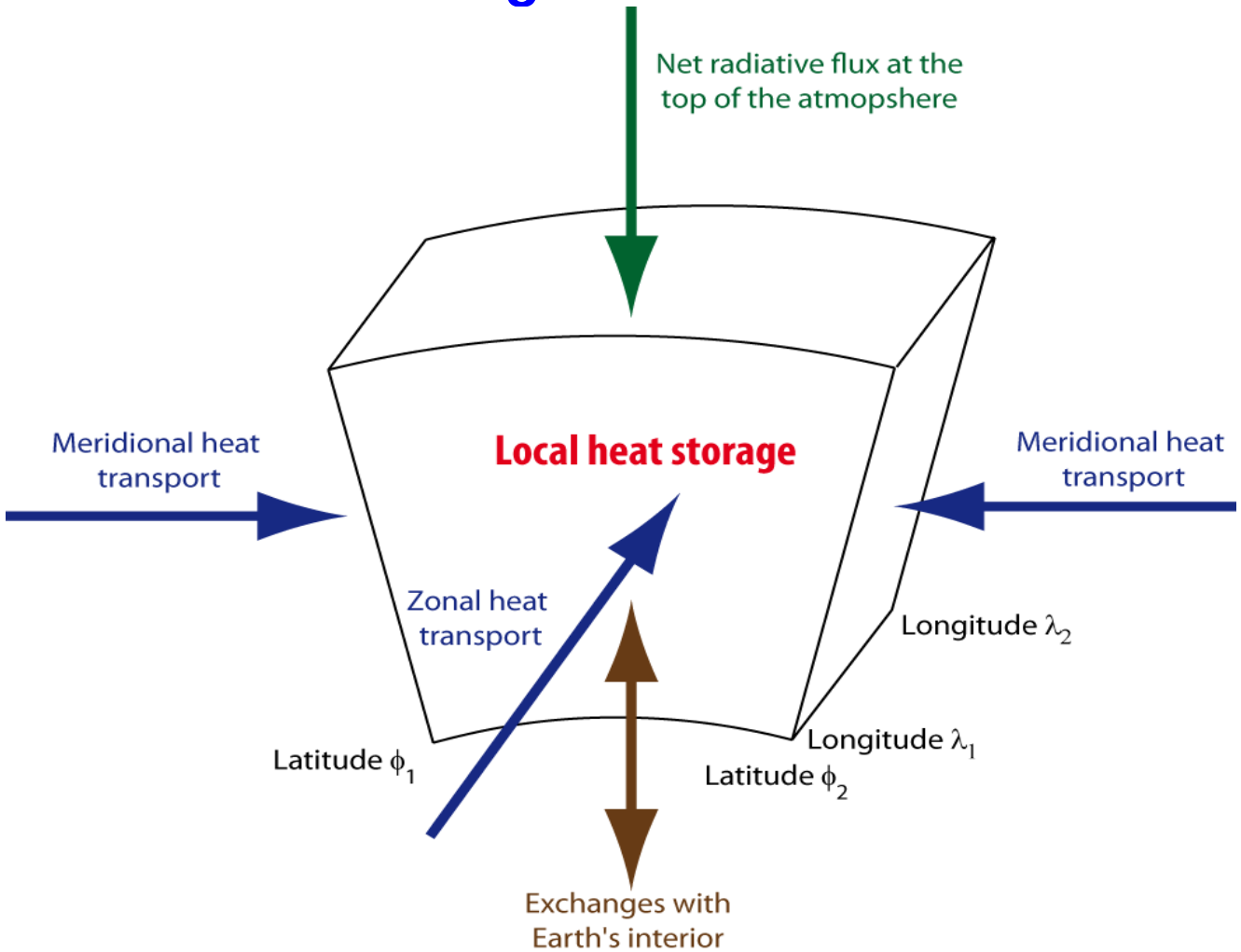
Radiative balance at the top of the atmosphere

Zonal mean of the absorbed solar radiation and the outgoing longwave radiation at the top of the atmosphere in annual mean (in W/m^2).



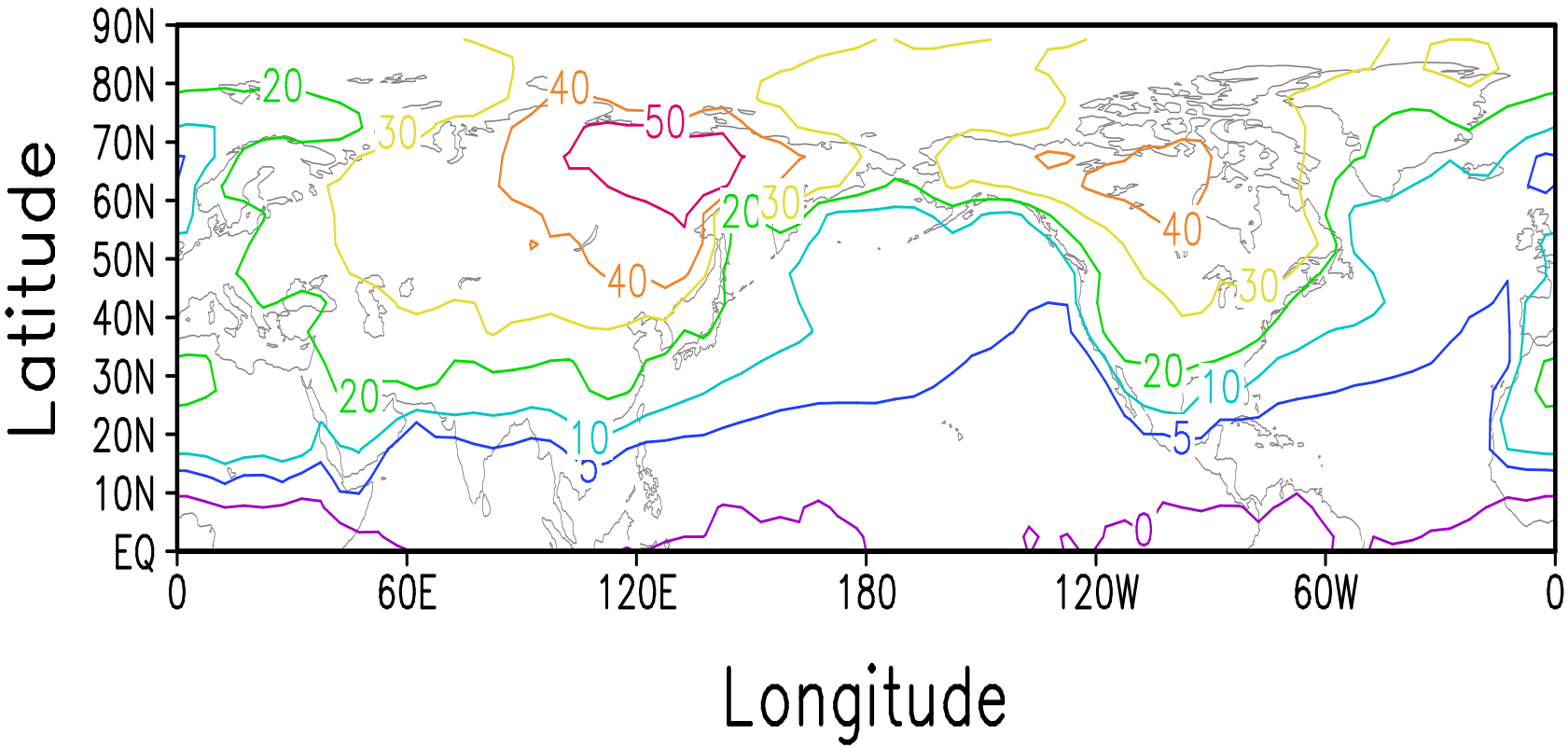
Heat storage and transport

The net radiative heat flux at the top of the atmosphere is mainly balanced by the horizontal heat transport and by changes in the heat storage.



Heat storage and transport

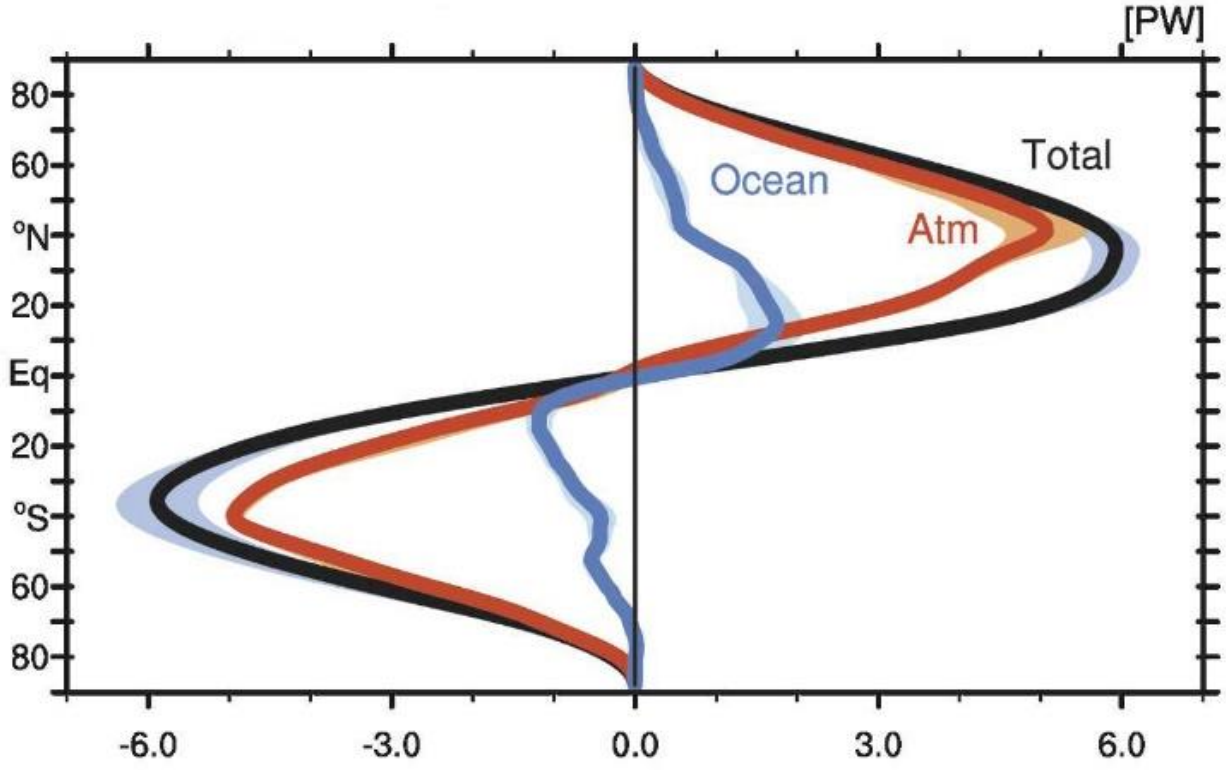
The heat storage strongly modulates the daily and seasonal cycles.



Amplitude of the seasonal cycle in surface temperature in the northern hemisphere measured as the difference between July and January monthly mean temperatures. Data from HadCRUT2 (Rayner et al., 2003).

Heat storage and transport

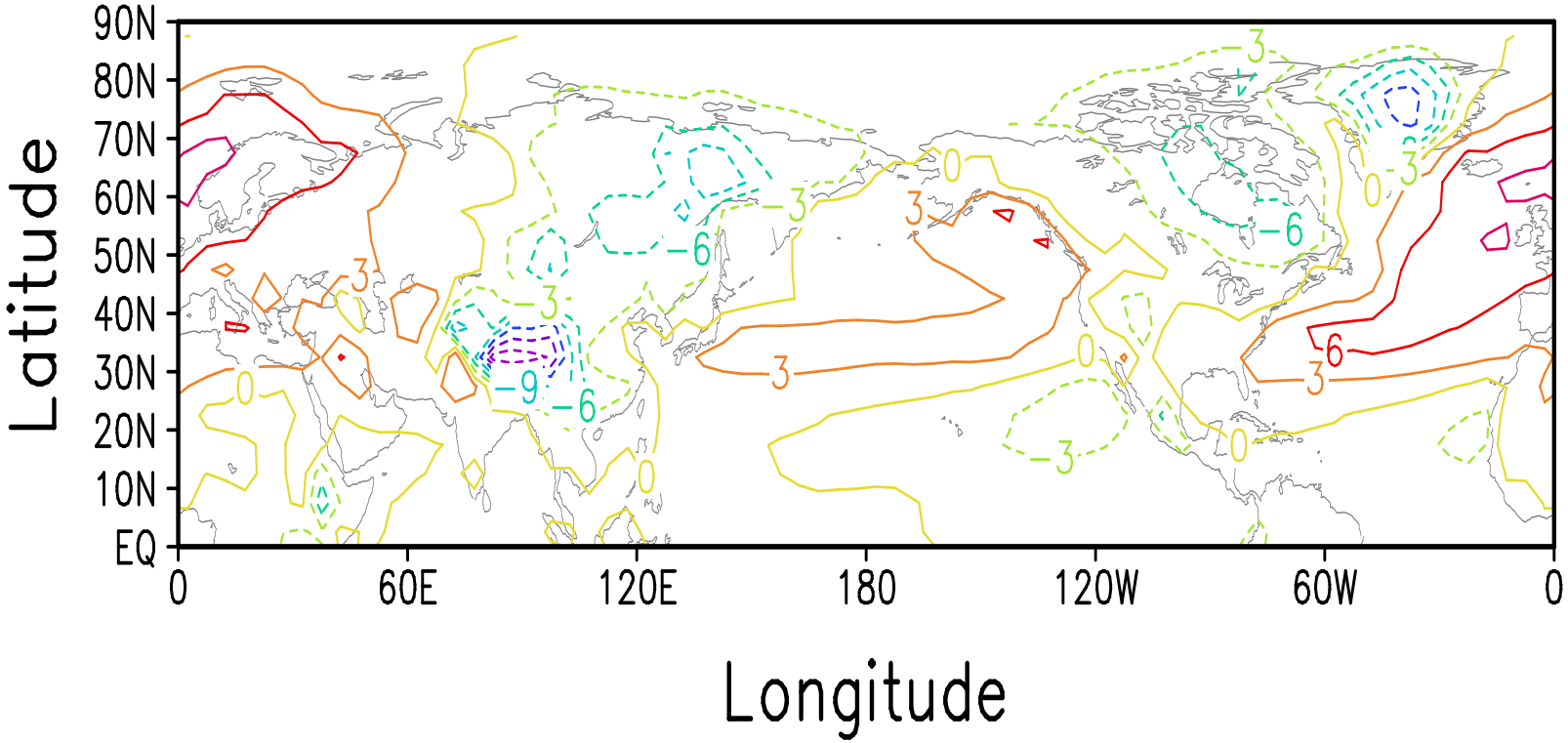
On annual mean, the net heat flux at the top of the atmosphere is balanced by the meridional heat transport.



The heat transport in PW (10^{15} W) needed to balance the net radiative imbalance at the top of the atmosphere (in black) and the repartition of this transport in oceanic (blue) and atmospheric (red) contributions. A positive value of the transport on the x axis corresponds to a northward transport. Figure from Fasullo and Trenberth (2008).

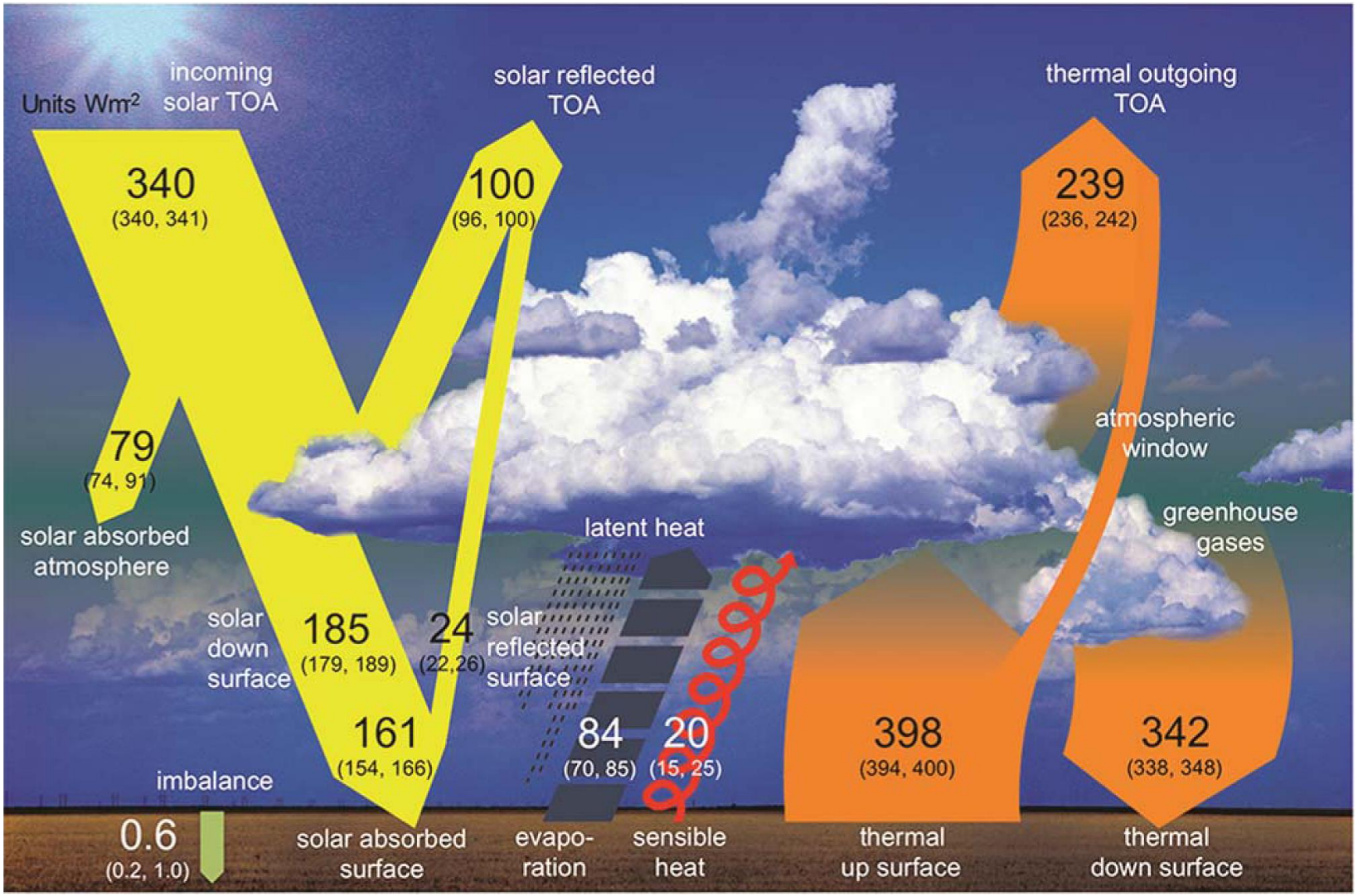
Heat storage and transport

The horizontal heat transport is also responsible for some temperature differences at the regional scale.



Difference between the annual mean surface temperature and the zonal mean temperature. This difference has been computed as the annual mean temperature measured at one particular point minus the mean temperature obtained at the same latitude but averaged over all possible longitudes.
Data from HadCRUT2 (Rayner et al., 2003).

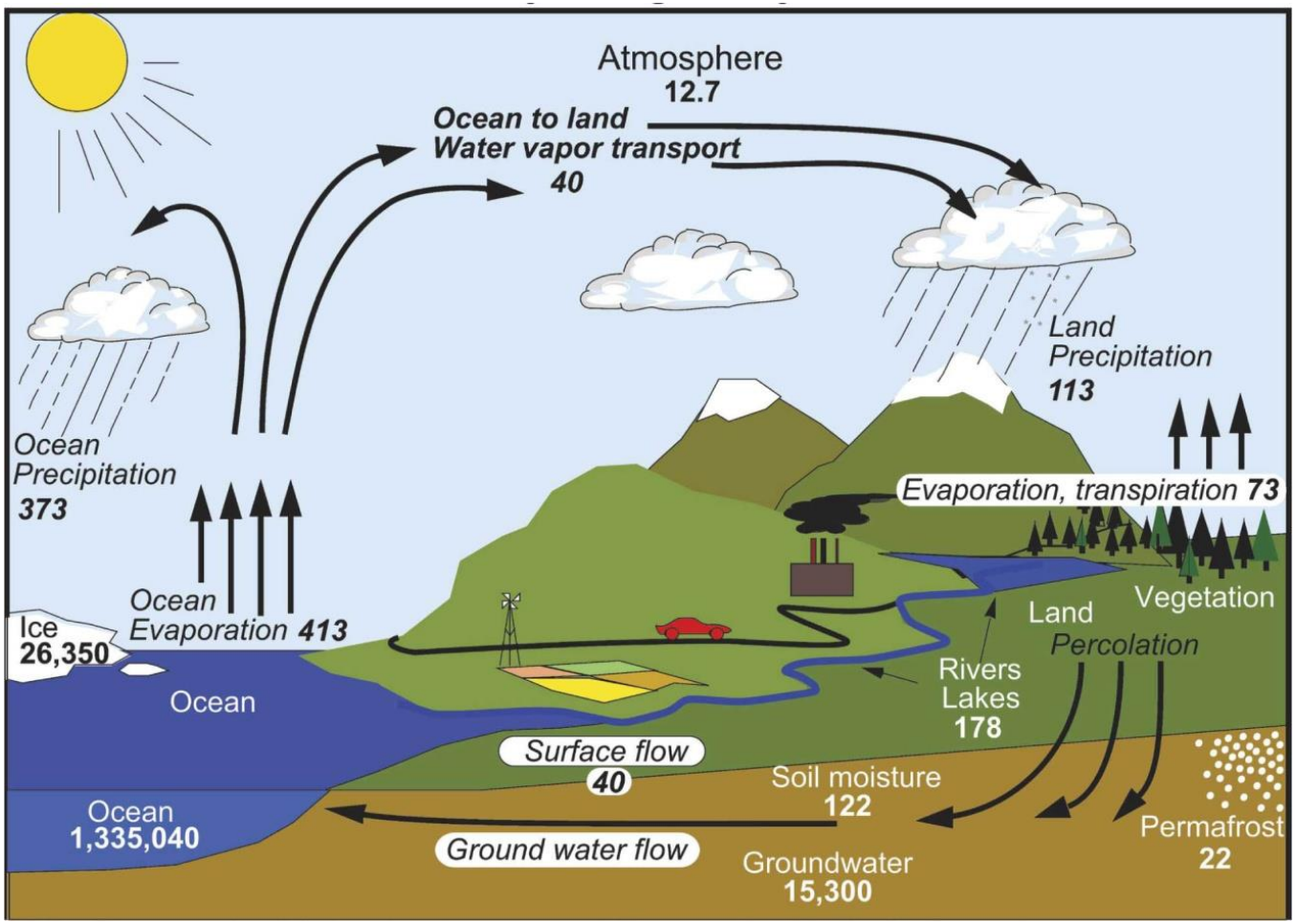
Heat balance at the surface



The numbers represent estimates of each individual energy flux whose uncertainty is given in the parentheses using smaller fonts. Figure from Hartmann et al. (2014) which is adapted from Wild et al. (2013).

Global water balance

Long-term mean global hydrological cycle



Estimates of the main water reservoirs in plain font (e.g. Soil moisture) are given in 10^3 km^3 and estimates of the flows between the reservoirs in italic (e.g. Surface flow) are given in 10^3 km^3/year. Figure from Trenberth et al. (2007)

Global water balance

Soil water balance

$$\frac{dS_m}{dt} = P - E - R_s - R_g$$

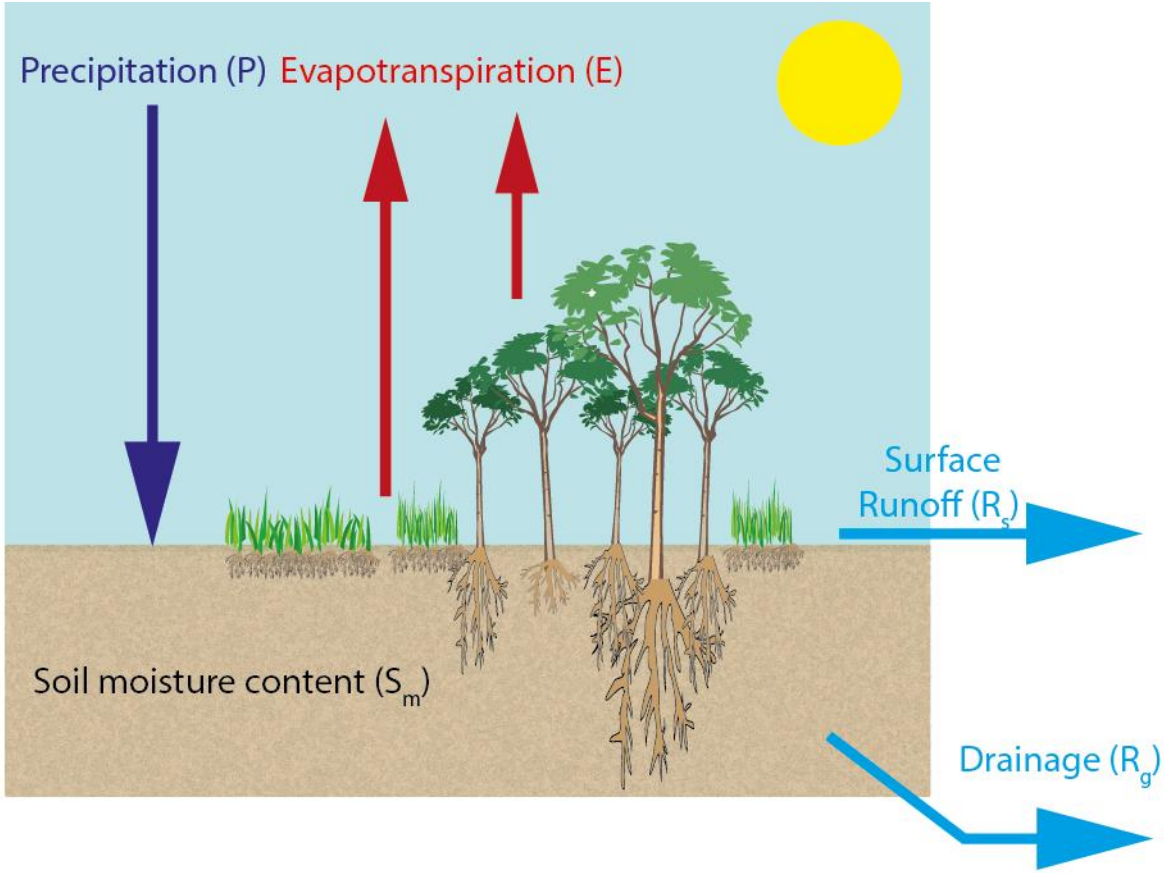
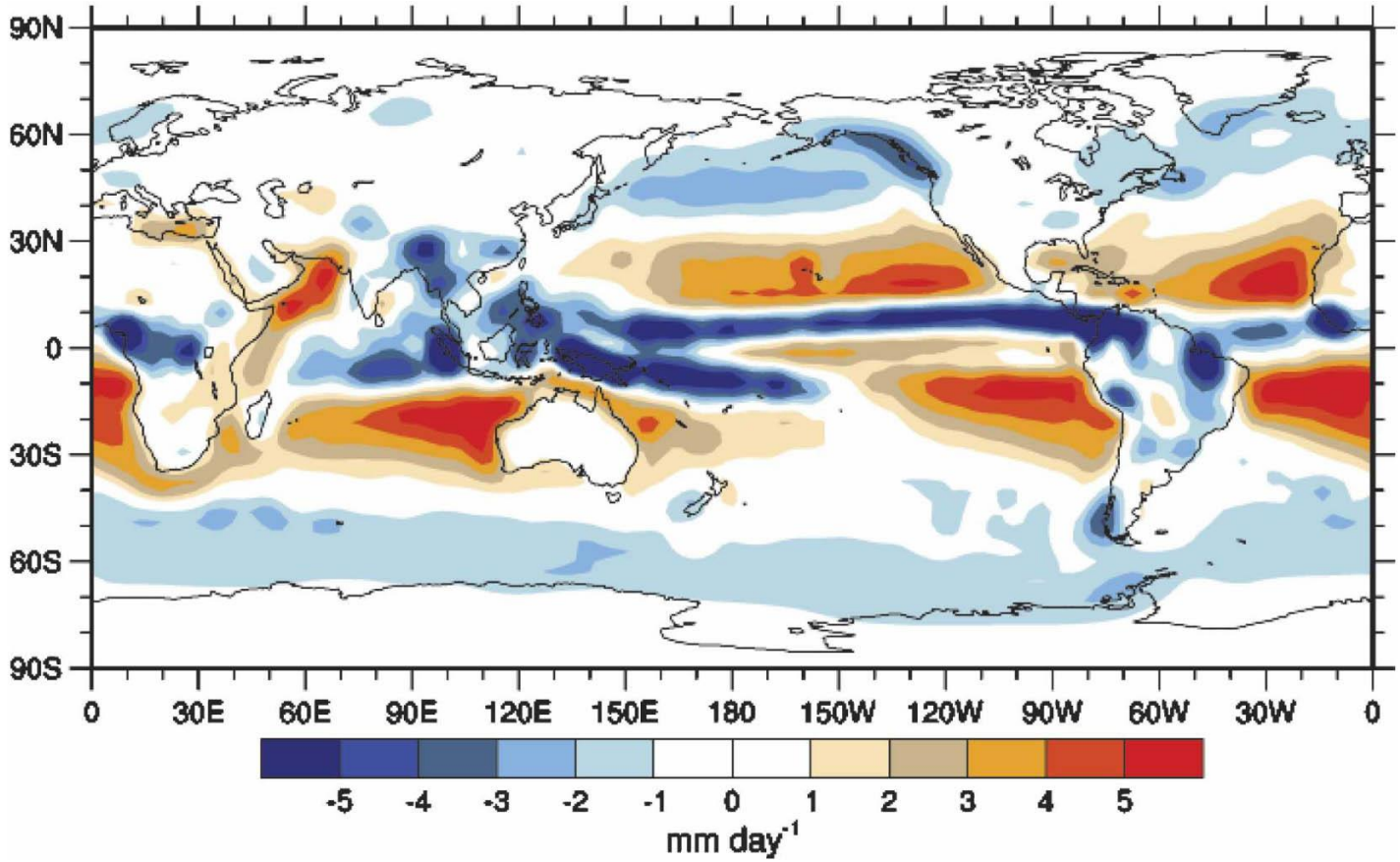


Figure Modified from Seneviratne et al. (2010).

Global water balance

Water balance at the ocean surface

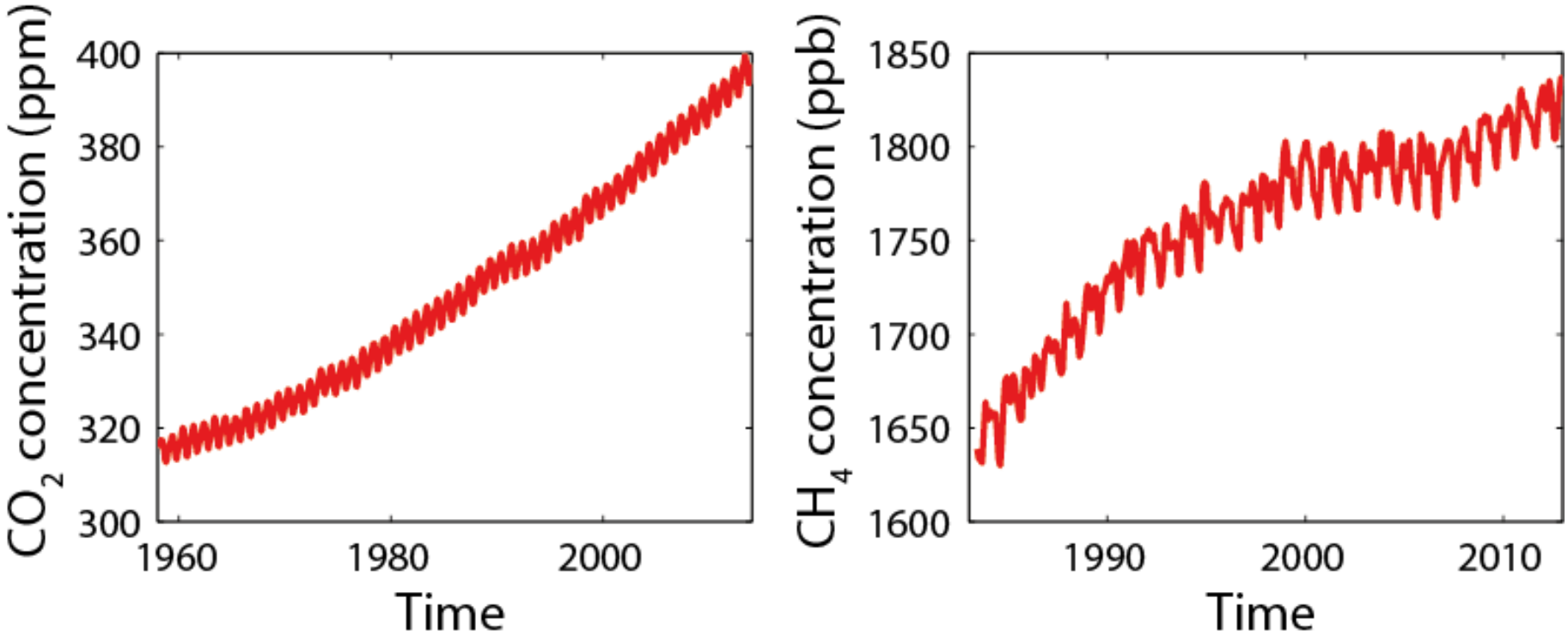


Long term annual mean evaporation minus precipitation (*E-P*) budget based on ERA-40 reanalyses. Figure from Trenberth et al. (2007).

The carbon cycle

CO_2 and CH_4 are two important greenhouse gases.

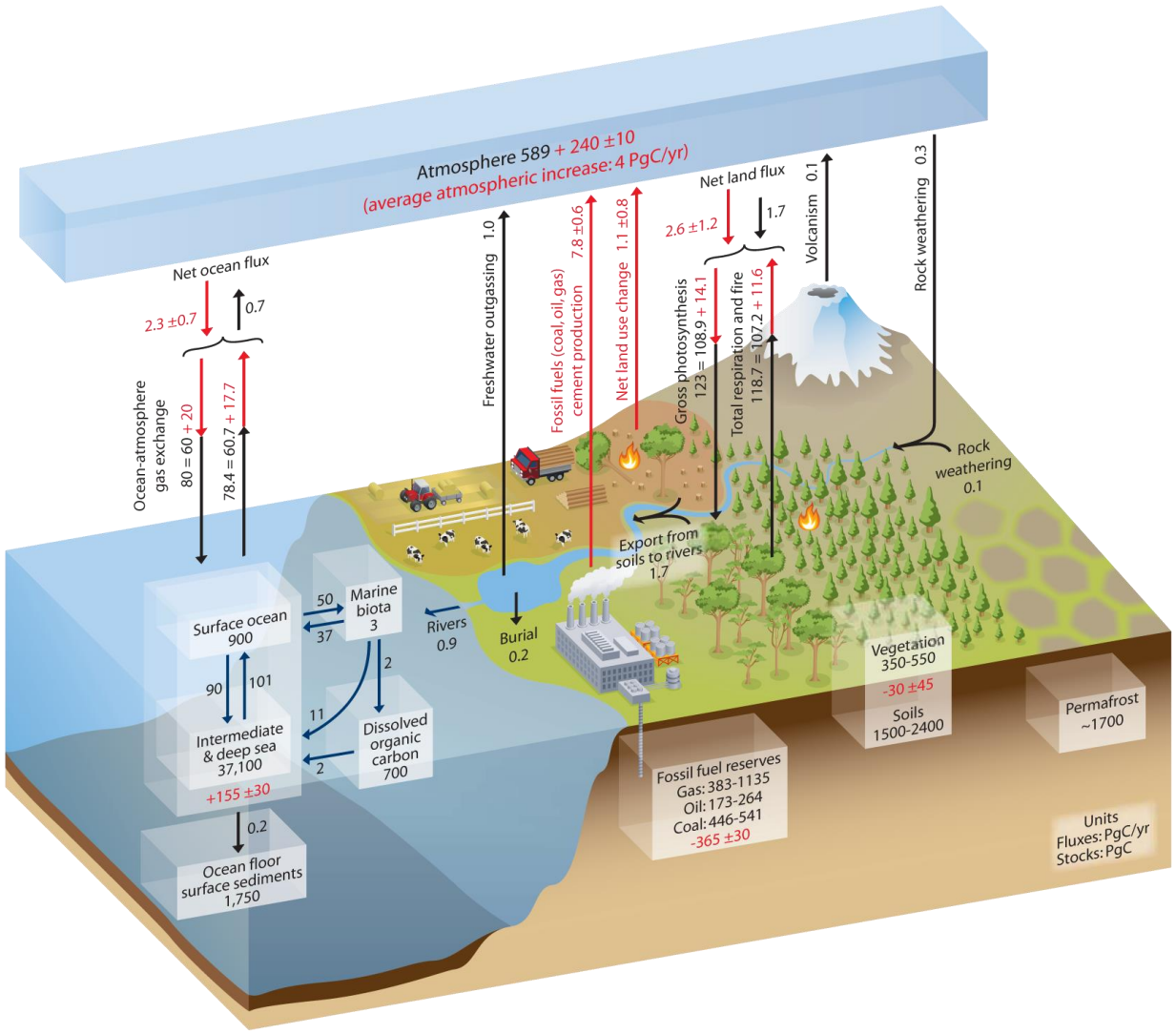
CO_2 and CH_4 concentration at Mona Laua observatory.



Sources: Dr. Pieter Tans, NOAA/ESRL (www.esrl.noaa.gov/gmd/ccgg/trends/) and Dr. Ralph Keeling, Scripps Institution of Oceanography (scrippsco2.ucsd.edu/), the NOAA Annual Greenhouse Gas Index (AGGI) (http://www.esrl.noaa.gov/gmd/aggi/), Dlugokencky et al. (2013).

The carbon cycle

Overview



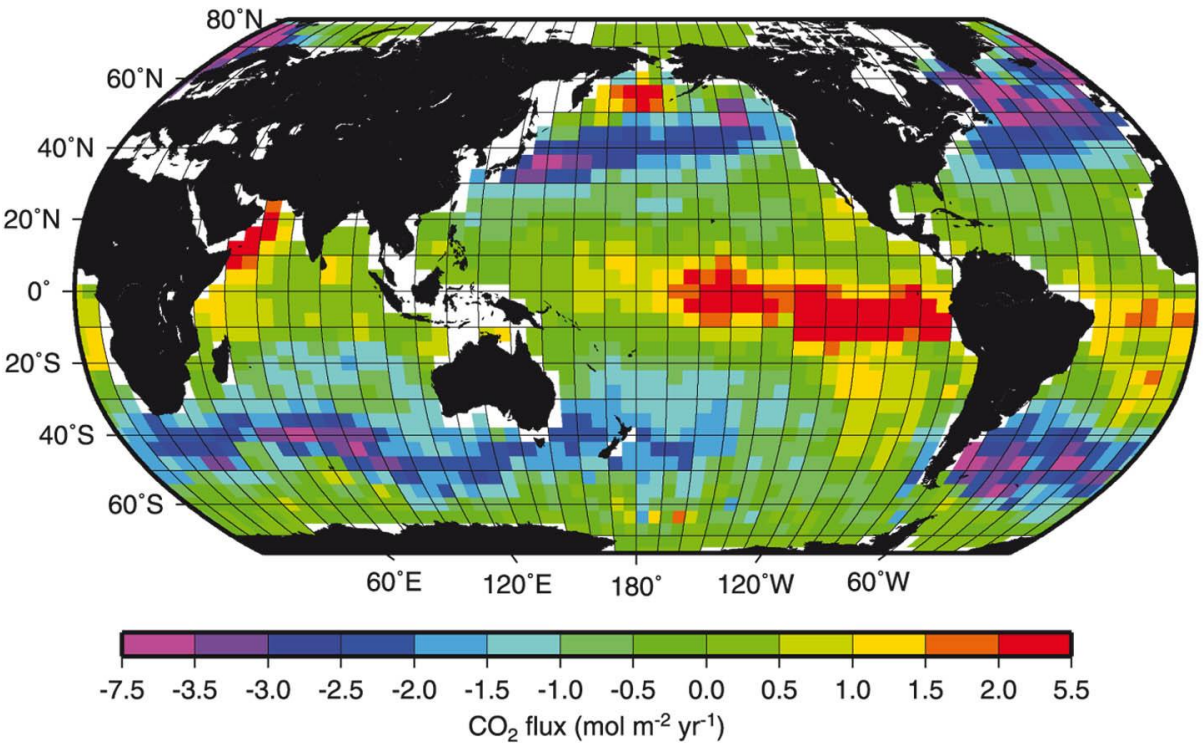
The annual fluxes are in PgC yr^{-1} , the carbon stocks in the reservoirs are given in PgC. Pre-industrial ‘natural’ fluxes are in black and ‘anthropogenic’ fluxes averaged over the 2000–2009 in red. Figure from Ciais et al. (2014).

The oceanic carbon cycle

The CO_2 flux from the ocean to the atmosphere is proportional to the difference of partial pressure (p^{CO_2}) between the two media:

$$\Phi^{CO_2} = k^{CO_2}\left(p_W^{CO_2} - p_A^{CO_2}\right)$$

Subscripts *A* and *W* refer to the air and the water, respectively. k^{CO_2} is a transfer coefficient.



Estimates of sea-to-air flux of CO_2 (Denman et al. (2007), based on the work of T. Takahashi).

The oceanic carbon cycle

The inorganic carbon cycle: balance between carbonic acid (H_2CO_3), bicarbonate (HCO_3^-) and carbonate ions (CO_3^{2-})

$$CO_{2(gas)} + H_2O \rightleftharpoons H_2CO_3$$

$$H_2CO_3 \rightleftharpoons H^+ + HCO_3^-$$

$$HCO_3^- \rightleftharpoons H^+ + CO_3^{2-}$$

The sum of the concentration of these three forms of carbon is referred to as the Dissolved Inorganic Carbon (DIC):

$$DIC = \left[H_2CO_3\right] + \left[HCO_3^-\right] + \left[CO_3^{2-}\right]$$

The oceanic carbon cycle

K_H, the solubility of CO_2 , relates the amount of the carbonic acid to the pCO_2 at equilibrium.

K_H is a strong function of the temperature.

$$K_H = \frac{[H_2CO_3]}{pCO_2}$$

At equilibrium, 90% of the dissolved inorganic carbon is in the form of bicarbonate, around 10% in carbonate form while carbonic acid represent only 0.5% of the DIC:

→ **Important for carbon storage in the ocean.**

The oceanic carbon cycle

The alkalinity *Alk* is defined as the excess of bases over acid in water:

$$Alk = \left[HCO_3^-\right] + 2\left[CO_3^{2-}\right] + \left[OH^-\right] - \left[H^+\right] + \left[B(OH)_4^-\right] + \text{minor bases}$$

where $\left[B(OH)_4^-\right]$ is the concentration of the borate ion.

The total alkalinity is dominated by the influence of bicarbonate and carbonate ions.

Conversely, changes in total alkalinity have a strong influence on the equilibrium of the reactions between the different carbon species.

The oceanic carbon cycle

Biological pumps: the soft tissue pump

During photosynthesis, phytoplankton uses solar radiation to form organic matter from CO_2 and water:

$$6CO_2 + 6H_2O \rightleftharpoons C_6H_{12}O_6 + 6O_2$$

The organic matter can be dissociated to form inorganic carbon by respiration and remineralisation of dead phytoplankton and detritus.

A fraction of the organic matter is exported downward out of the surface layer.

The oceanic carbon cycle

Biological pumps: the carbonate pump

Calcium carbonate, is produced by different species :

$$Ca^{2+} + CO_3^{2-} \rightleftharpoons CaCO_3$$

This production influences both the *DIC* and the *Alk* and thus has a large influence on the carbon cycle.

The dissolution of calcite and aragonite occurs mainly at depths, following the falling of particles and dead organism.

The oceanic carbon cycle

The solubility pump

The "solubility pump" is associated with the sinking at high latitude of cold surface water to great depth.

This cold water characterized by a relatively high solubility of CO_2 and thus high *DIC*.

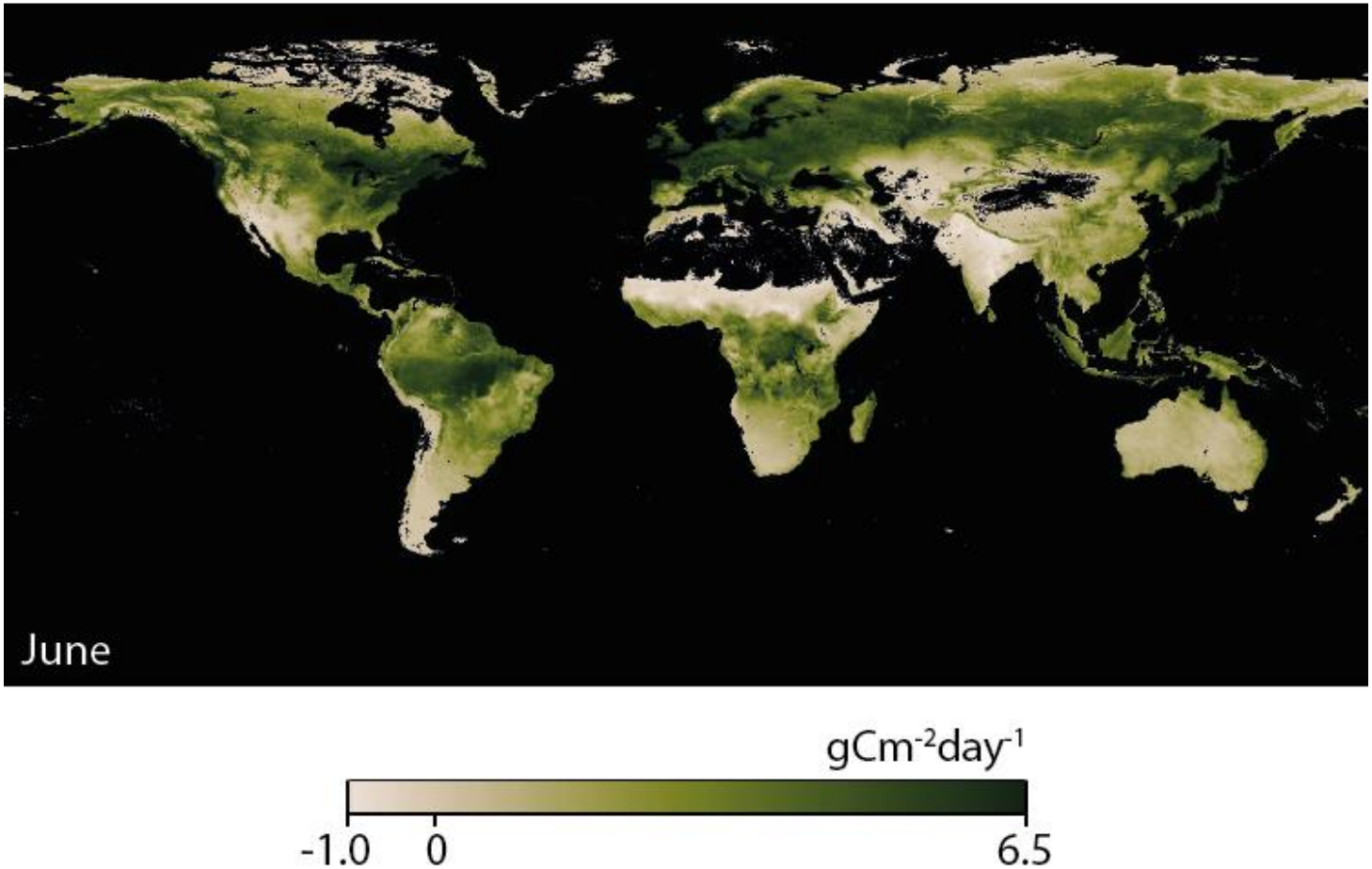
The oceanic carbon cycle

Because of those three pumps, DIC is about 15% higher at depth than at surface, inducing lower atmospheric CO_2 concentration compared to an homogenous ocean.

When deep water upwells to the surface, CO_2 will have the tendency to escape from the ocean because of the high *DIC*. But this is partly compensated by biological activity.

The terrestrial carbon cycle

The photosynthesis by land plants has a strong seasonal cycle.



Net productivity over land in December 2013 and June 2014 based on Terra/MODIS satellite data. Source: NASA Earth Observations.

Geological reservoirs

A fraction of the of $CaCO_3$ produced in the ocean is buried in the sediments to produce limestone, mainly in shallow seas.

During the subduction, limestone is transformed into calcium-silicate rocks (metamorphism) by the reaction:

$$CaCO_3 + SiO_2 \rightarrow CaSiO_3 + CO_2$$

The CO_2 that is realised by this reaction can return to the atmosphere, in particular through volcanic eruptions.

Geological reservoirs

If the calcium-silicate rocks are uplifted to the continental surface, they are affected by physical and chemical weathering.

$$CaSiO_3 + H_2CO_3 \rightarrow CaCO_3 + SiO_2 + H_2O$$

The products of this reaction are transported by rivers to the oceans where they could compensate for the net export of $CaCO_3$ by sedimentation.

The weathering tends thus to reduce atmospheric CO_2 by taking up carbonic acid to make $CaCO_3$ and increasing ocean alkalinity.

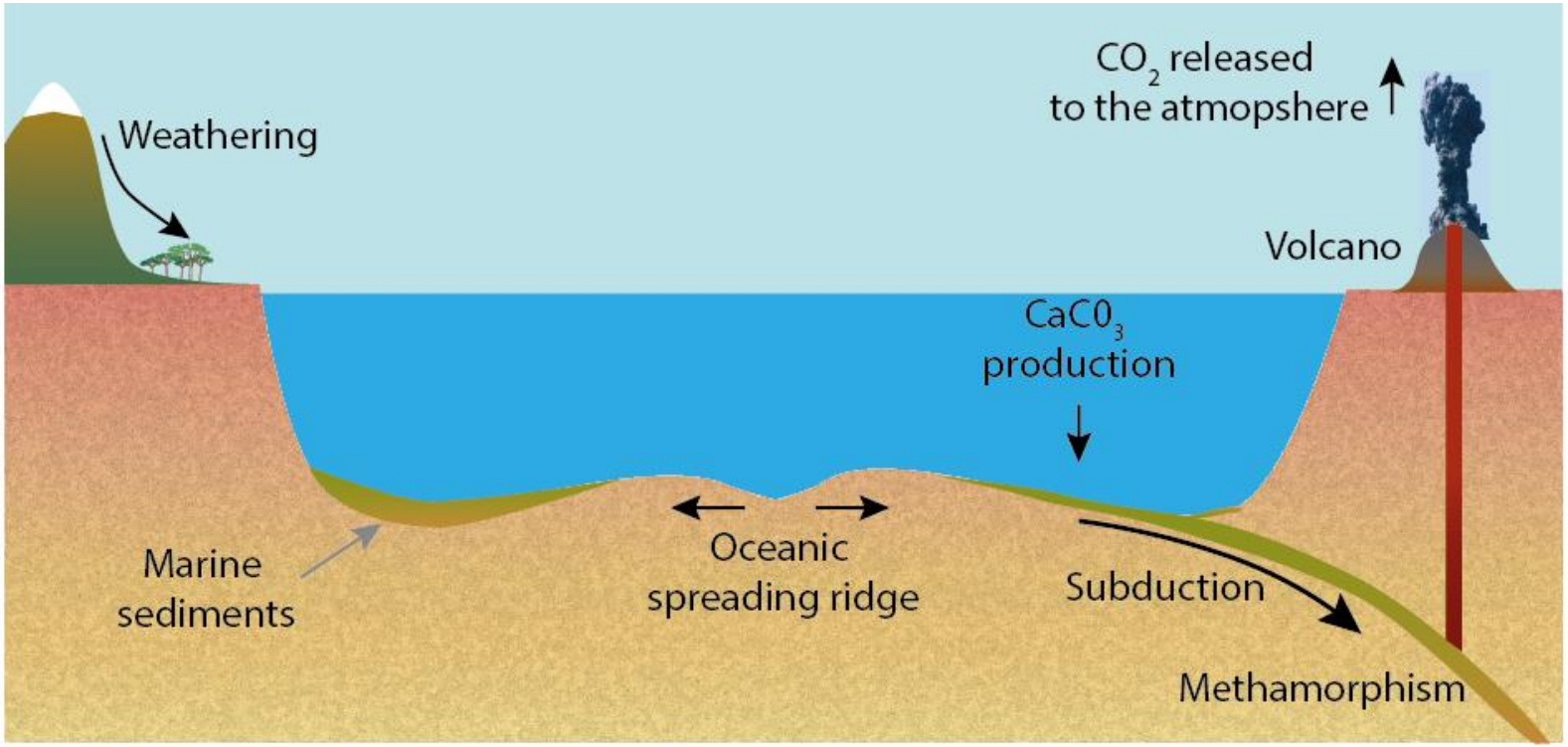
Geological reservoirs

Those equations describes the ‘long term inorganic carbon cycle” .

$$Ca^{2+} + CO_3^{2-} \rightarrow CaCO_3$$ Sedimentation

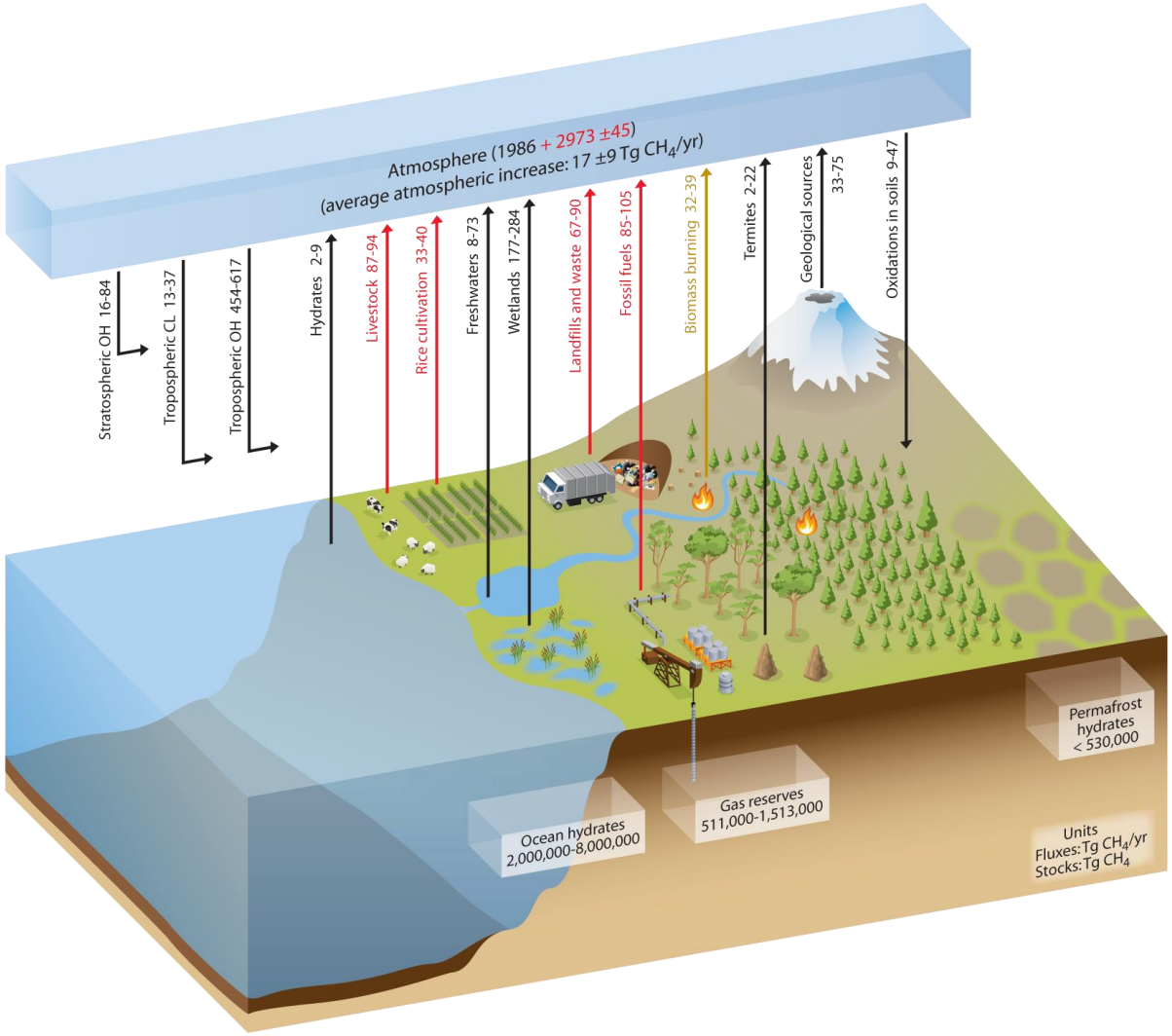
$$CaCO_3 + SiO_2 \rightarrow CaSiO_3 + CO_2$$ Metamorphism

$$CaSiO_3 + H_2CO_3 \rightarrow CaCO_3 + SiO_2 + H_2O$$ Weathering



Methane cycle

Overview



The annual fluxes are in Tg(CH_4) yr^{-1} for the period 2000–2009 and CH_4 reservoirs are in Tg(CH_4). Black arrows denote the natural fluxes red arrows the anthropogenic fluxes, and the light brown arrow denotes a combined natural + anthropogenic flux. Figure from Ciais et al. (2014).

Methane cycle

The observed atmospheric methane concentration results from the balance between sources and sinks due to methane oxidation.

$$CH_4 + 2O_2 \rightarrow CO_2 + 2H_2O$$

This reaction requires the presence of highly reactive constituents such as the hydroxyl radical (OH) .