The Earth's Climate

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Physics of the Earth

Course ae4-876

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Contents

Introduction

- Atmospheric dynamics is very close to ocean dynamics, density of the atmosphere is 1000 times smaller than water
- Atmosphere dynamics is in essence also the Navier Stokes equations for a thin layer of gas on a rotating sphere
- Transport of energy is completely different for oceanic and atmospheric processes
- Time and length scales are completely different for the ocean and the atmosphere,
- You should be familiar with the difference between meteorology and climatology
- Human intervention is visible in the ocean and atmosphere, actually, both issues are coupled.

Hydrostatic equilibrium

$$dp = -\rho(z)g(z)dz$$

Ideal gas law

$$P = N * k * T \rightarrow P(z) = \frac{\rho(z)T(z)k}{m(z)}$$

P: pressure, ρ: density, g:gravity, T:temperature,
z: altitude, m: molecular mass, k: Bolzmann constant **Source: Ch4 PS book**

Consequence

$$dP = -\frac{m(z)g(z)}{T(z)k}P(z)dz$$

$$P(z) = P(0)e^{-\int_0^z \frac{dr}{H(r)}}$$

$$H(r) = \frac{kT(r)}{m(r)g(r)}$$

On Earth:

T : 288 Kelvin

H : 8.5 km

Source: Ch4 PS book

Thermal wind equations

Earlier in the tides lectures we found:

$$\frac{Du}{Dt} = \frac{-1}{\rho}\frac{\partial p}{\partial x} + 2\Omega\sin\phi v - 2\Omega\cos\phi w + F_x$$

$$\frac{Dv}{Dt} = \frac{-1}{\rho}\frac{\partial p}{\partial y} - 2\Omega\sin\phi u + F_y$$

$$\frac{Dw}{Dt} = \frac{-1}{\rho}\frac{\partial p}{\partial z} + 2\Omega\cos\phi u - g + F_z$$

For the discussion that follows we simplify the equations to:

$$\frac{1}{\rho}\frac{\partial p}{\partial x} = f.v_g \qquad \frac{1}{\rho}\frac{\partial p}{\partial y} = -f.u_g \qquad \frac{1}{\rho}\frac{\partial p}{\partial z} = -g$$

where u_g and v_g are geostrophic velocities:

Source: tides lectures

Thermal wind equations

We have:

$$-fv = -\partial\Phi/\partial x$$

and

$$fu = -\partial\Phi/\partial y$$

where Φ is called the geopotential which comes from the hydrostatic equation

$$d\Phi = -\frac{dP}{\rho} = v_s dP$$

where v_s is called the specific volume, for this there is a model that depends on pressure, salinity (only in oceanography) and temperature

Source: Gill book

The geopotential height H or Z_g

- The geopotential ϕ is describes the position dependent energy per unit mass.
- Position dependent energy is also called the potential energy in mechanics

$$\phi = \int_0^H g(\varphi,\lambda,z)\,dz$$

Formal definition

$$Z_g = \frac{\phi}{g_0}$$

Mean sea level gravity approximation

In oceanography and meteorology there is a so-called equation of state which describes the physical properties of water or air:

$$\delta = v_s(S,T,p) - v_s(S_0,T_0,p_0)$$

so that:

$$-\Phi' = \int_0^P \delta \, dp$$

And therefore we can generate equations like:

$$-f\{v_g(p_1)-v_g(p_2)\} = \frac{\partial\{\Phi'(p_1)-\Phi'(p_2)\}}{\partial x}$$

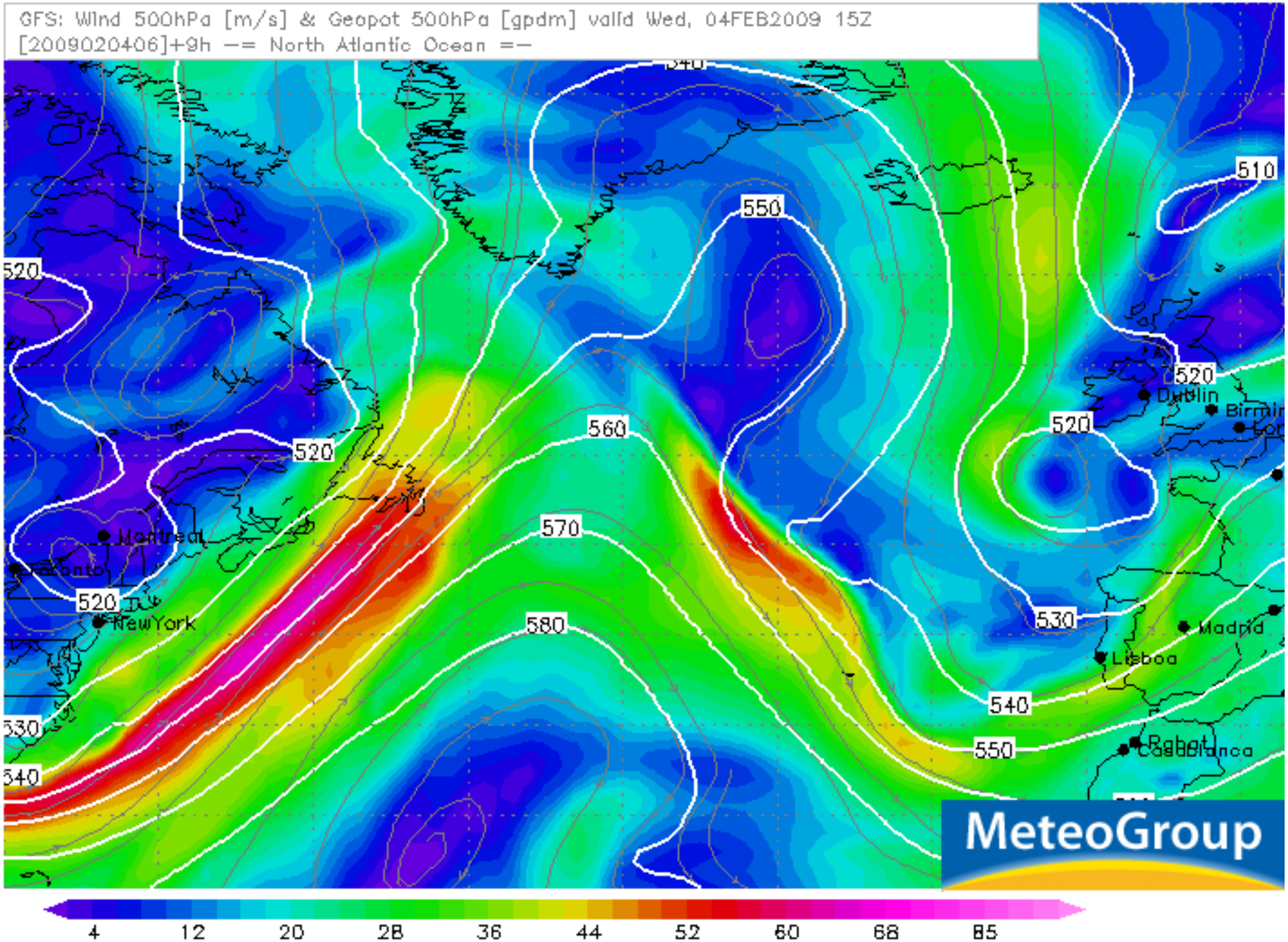
$$+f\{u_g(p_1)-u_g(p_2)\} = \frac{\partial\{\Phi'(p_1)-\Phi'(p_2)\}}{\partial y}$$

In meteorology:

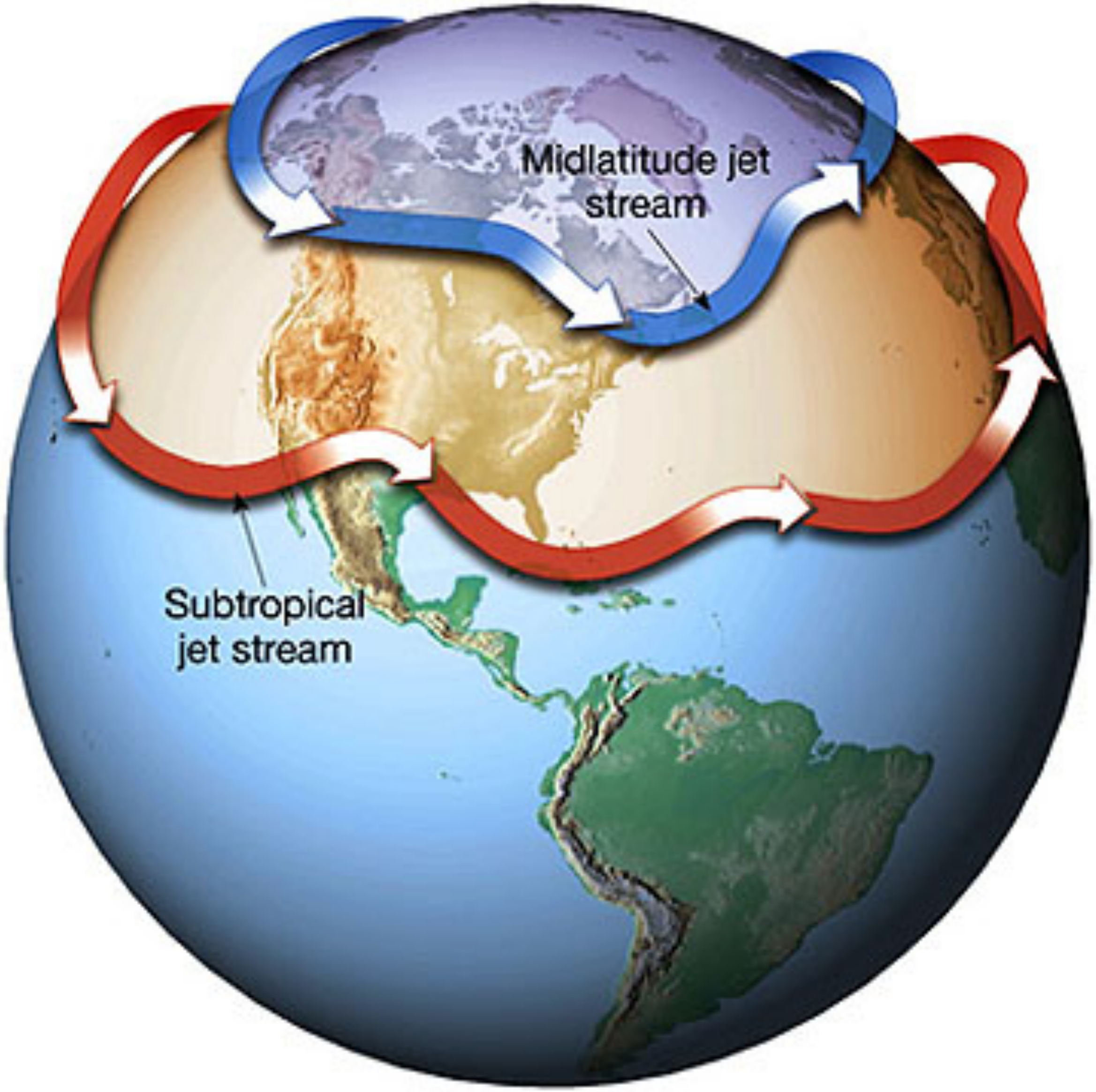
$$\Phi_1 - \Phi_2 = \int_{p1}^{p2} p^{-1} RT \, dp = R\overline{T} \ln(p_2/p_1)$$

Source: Gill book

500 hPa + wind 4-feb-2009 15Z



Jet Stream



Flight times (KLM):

AMS->SFO: 11h 15m

SFO->AMS: 10h 35m

Distance 8808 km

Speed1 = 783 km/h

Speed2 = 832 km/h

Delta = 49 km/h

Extreme jet stream speeds are much larger (>200 km/h)

Source: www.fas.org

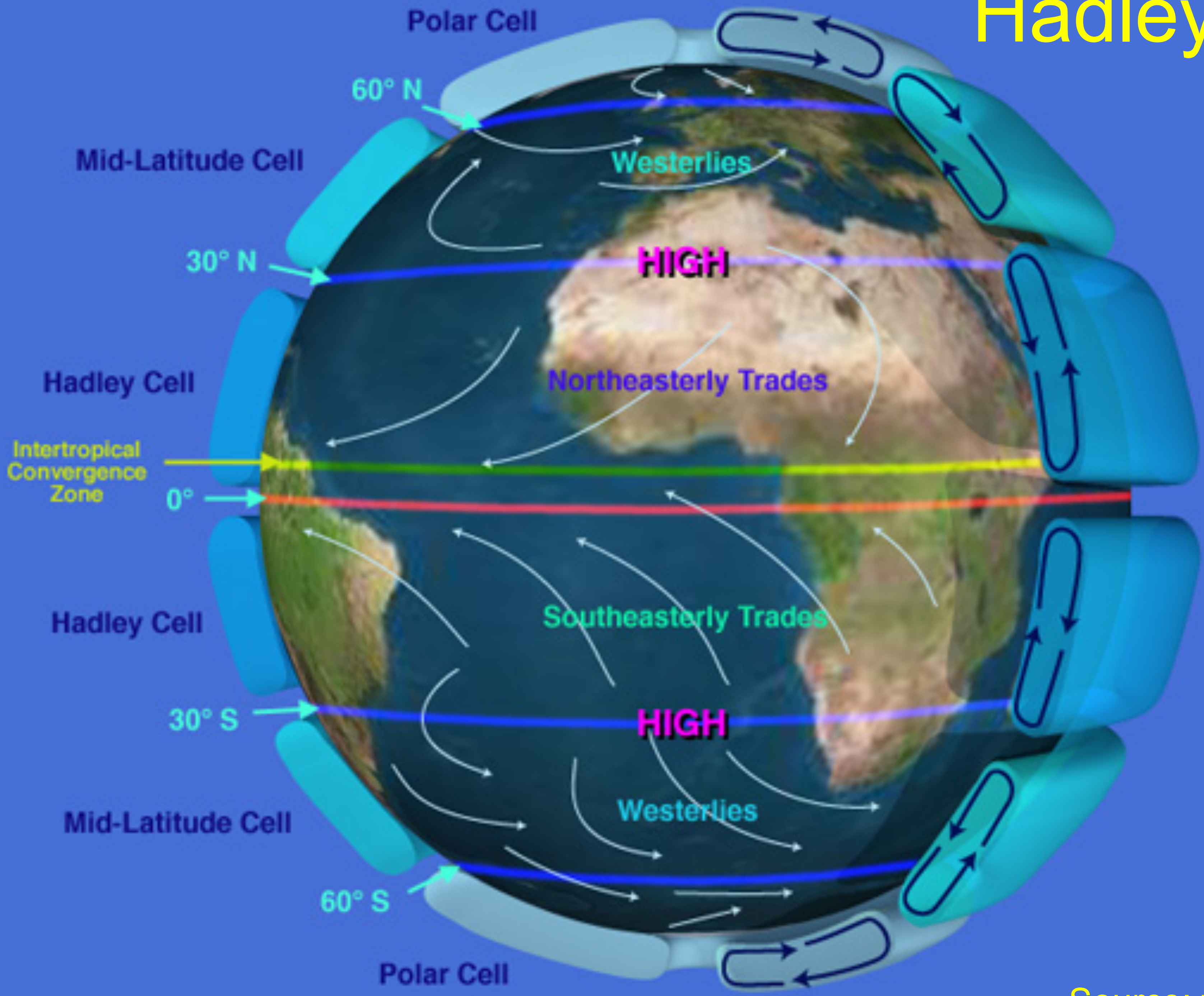
Summary thermal wind equation

- By a field experiment with XBTs or radiosondes you can always compute $\nabla\Phi$ which allows you to derive ***u*** and ***v*** profiles for the geostrophic wind speed

- In oceanography there is a level of no motion discussion, the deep ocean is motionless

- In meteorology you always have the ground level as a reference

Hadley cells



Source: wikipedia

Radiative equilibrium (holds for any object in space)

Input :

$$F_{in} = (1 - A_b)\frac{L_{sun}}{4\pi r_{sun}^2}\pi R^2$$

L_{sun} : Luminosity of the Sun

r_{sun} : Earth Sun distance

R : radius Earth

A_b : Bond Albedo

Output :

$$F_{out} = 4\pi R^2 \varepsilon \sigma T^4$$

T : temperature

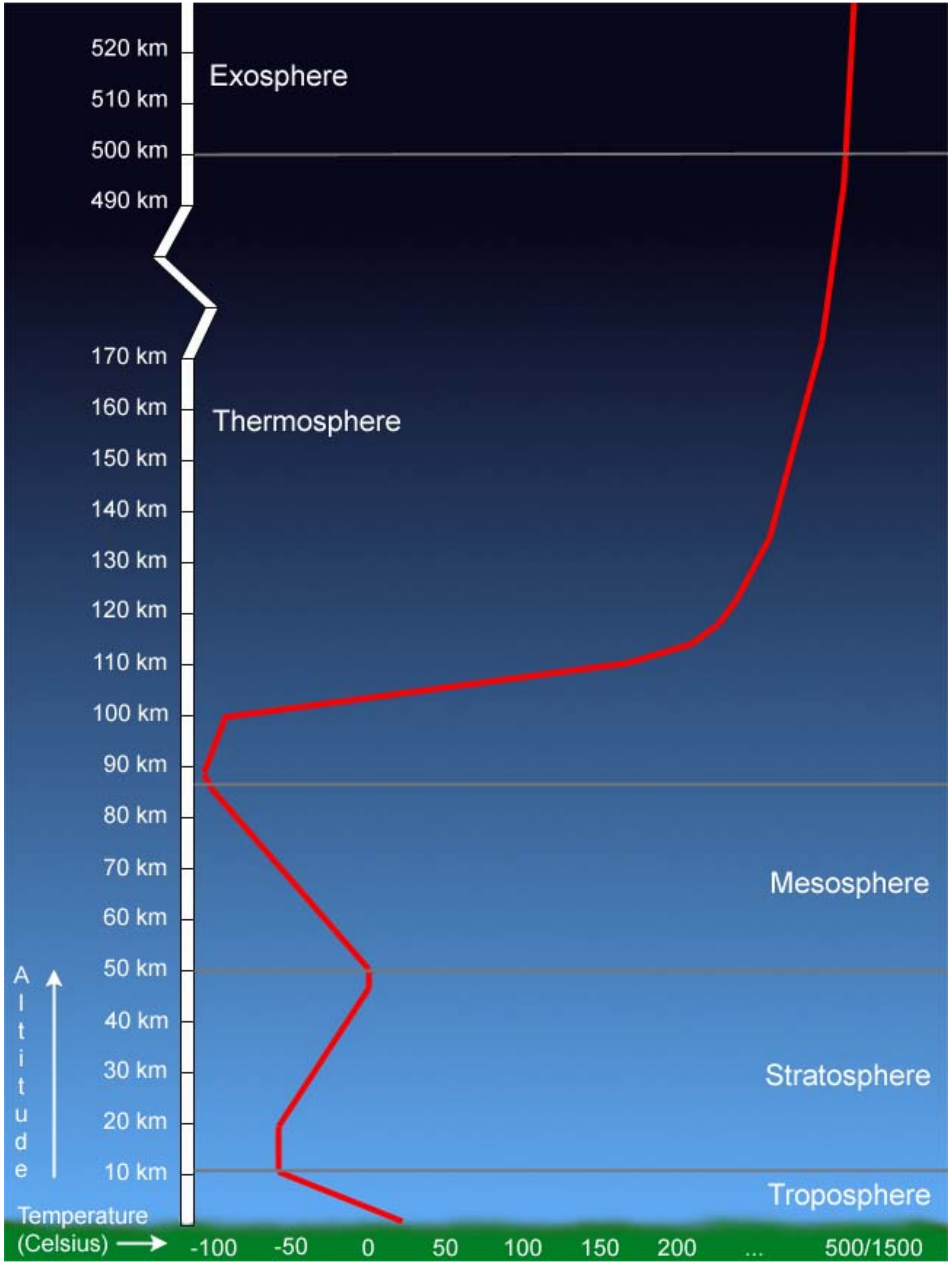
σ : Stefan - Boltzmann cst

ε : emissivity of the Earth

$$T_{eq} = \left(\frac{F_{sun}}{r_{AU}^2}\frac{(1 - A_b)}{4\varepsilon\sigma}\right)^{1/4}$$

⇒ 263K on Earth (it is too low)

Source: UCAR



spaceflight

Top of atmosphere

You

Thermal structure of the atmosphere

- The thermal structure of the atmosphere (dT/dz) is determined by sources and transport of heat energy.
- Major sources:
  - Sunlight is the dominant source, it is absorbed at the surface, the atmosphere where it is less opaque (optical depth), or dust in the atmosphere
  - The planet’s surface and dust are infrared radiators
- Minor sources:
  - Internal heat sources on the planet (Think of volcanoes and geysers, or the large gaseous planets)
  - Charged particle precipitation in the aural zones and thermospheric winds that heat the entire planet
  - Joule heating from electric currents in the ionosphere

Source: Chapter 4.2 PS book

Energy transport

- **Conduction:** This only happens in the upper thermosphere and the exosphere in the form of collisions between particles.
- **Convection:** The troposphere is governed by convection, dry adiabatic lapse rate, clouds, etc.
- **Radiation:** When energy is transported by absorption and re-emission of radiation. A good approximation is the radiative equilibrium of a planet, but more fundamentally you have to solve the radiative transfer problem.

Dry adiabatic lapse rate

$$\frac{dT}{dz} = -g(z)/c_p = -\frac{\gamma - 1}{\gamma}\frac{g(z)m(z)}{k}$$

$g(z)$: gravity

c_p : specific heat without changing pressure

c_v : specific heat without changing volume

$\gamma = c_p / c_v$

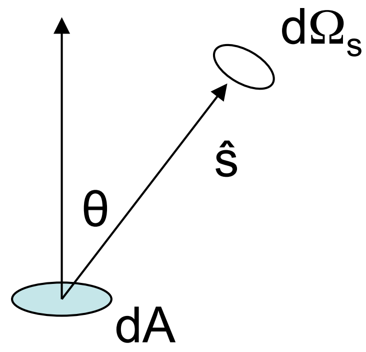
γ is usually 5/3, 7/5 or 4/3 depending on the gas

What is more efficient?

- For an atmosphere that is marginally unstable to convection we can define (dT/dz) by the dry adiabatic lapse rate equation
- An atmosphere is said to be super-adiabatic and convection causes dT/dz to be greater than the dry adiabatic lapse rate, this happens in the lower troposphere
- When the optical depth is not large and not too small energy is transported by radiation, this happens above the troposphere
- Upper part thermosphere: conduction.

Source PS book ch 3+4

Radiative transfer problem



$E = h\nu$: energy of a photon

$\overline{p} = \frac{E}{c}\hat{s}$: momentum of a photon

$dE = I_\nu \cos\theta \, dt \, dA \, d\Omega_s \, d\nu$

I_ν : specific intensity

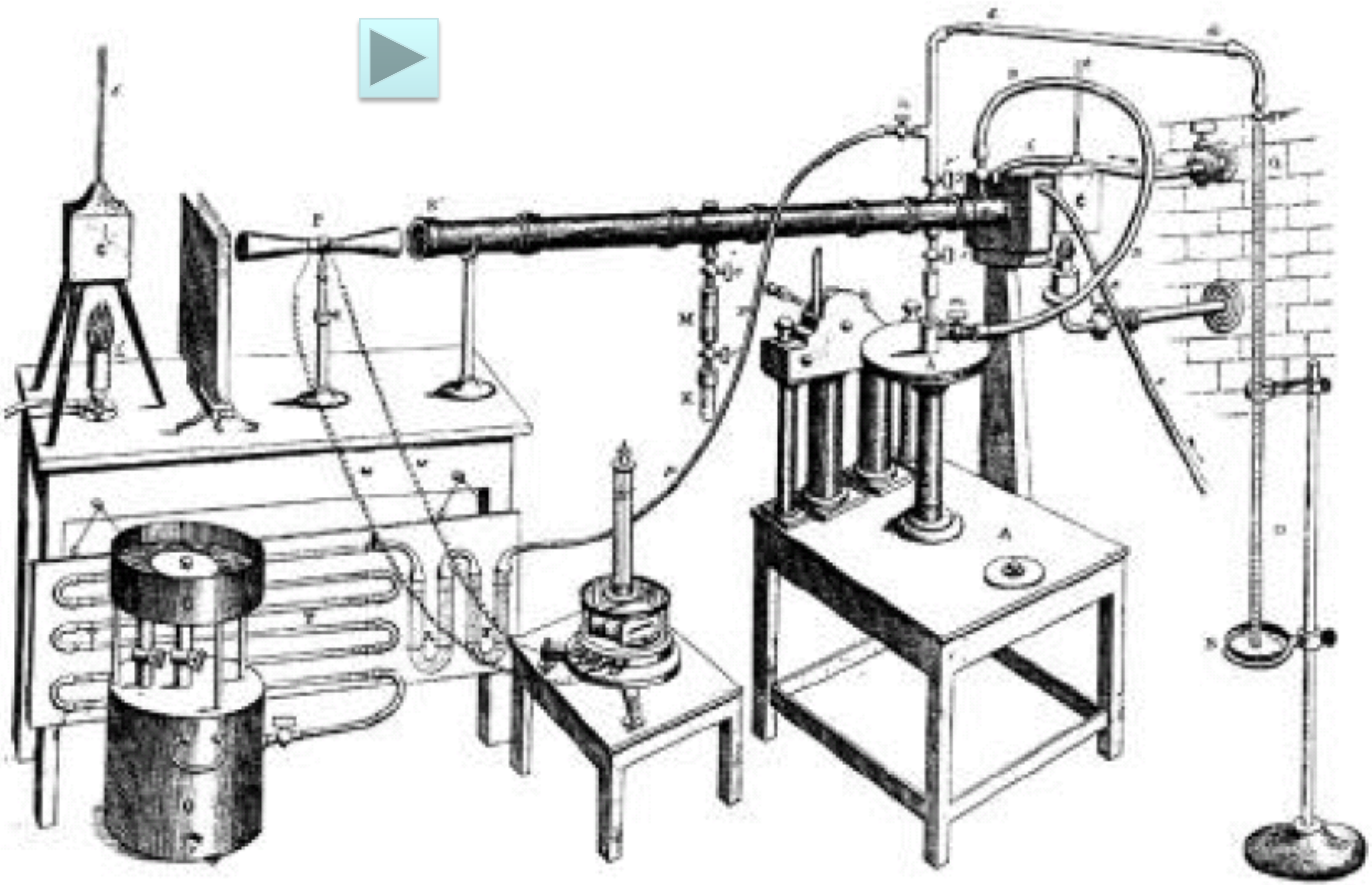
$I_\nu = B_\nu(T)$: specific intensity of a black body

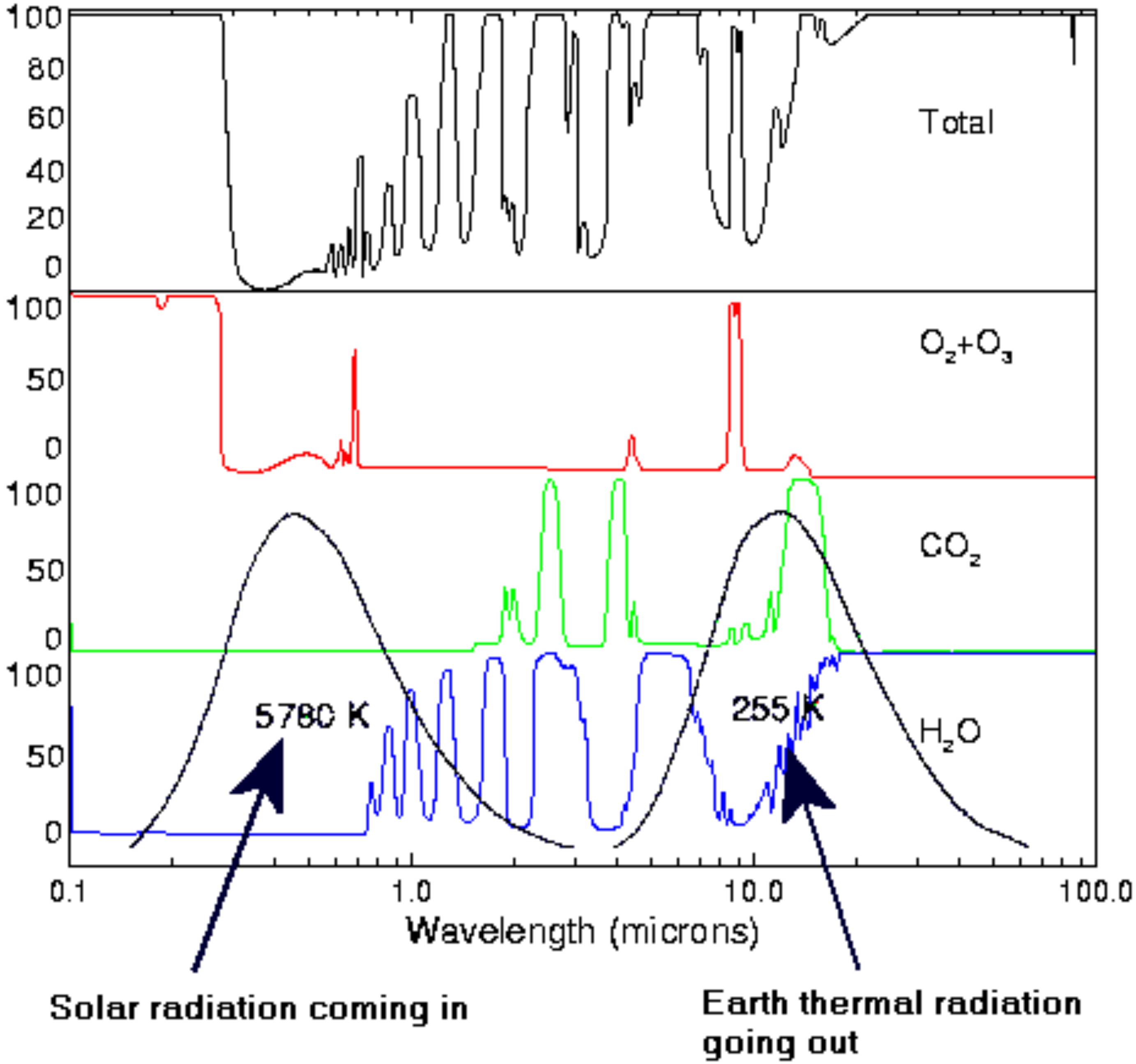
$dI_\nu = j_\nu \rho \, ds - I_\nu \alpha_\nu \rho \, ds$

$j_\nu = j_\nu(\text{scattering}) + j_\nu(\text{thermal excitation})$

α_ν : mass extinction = mass absorbtion + mass scattering

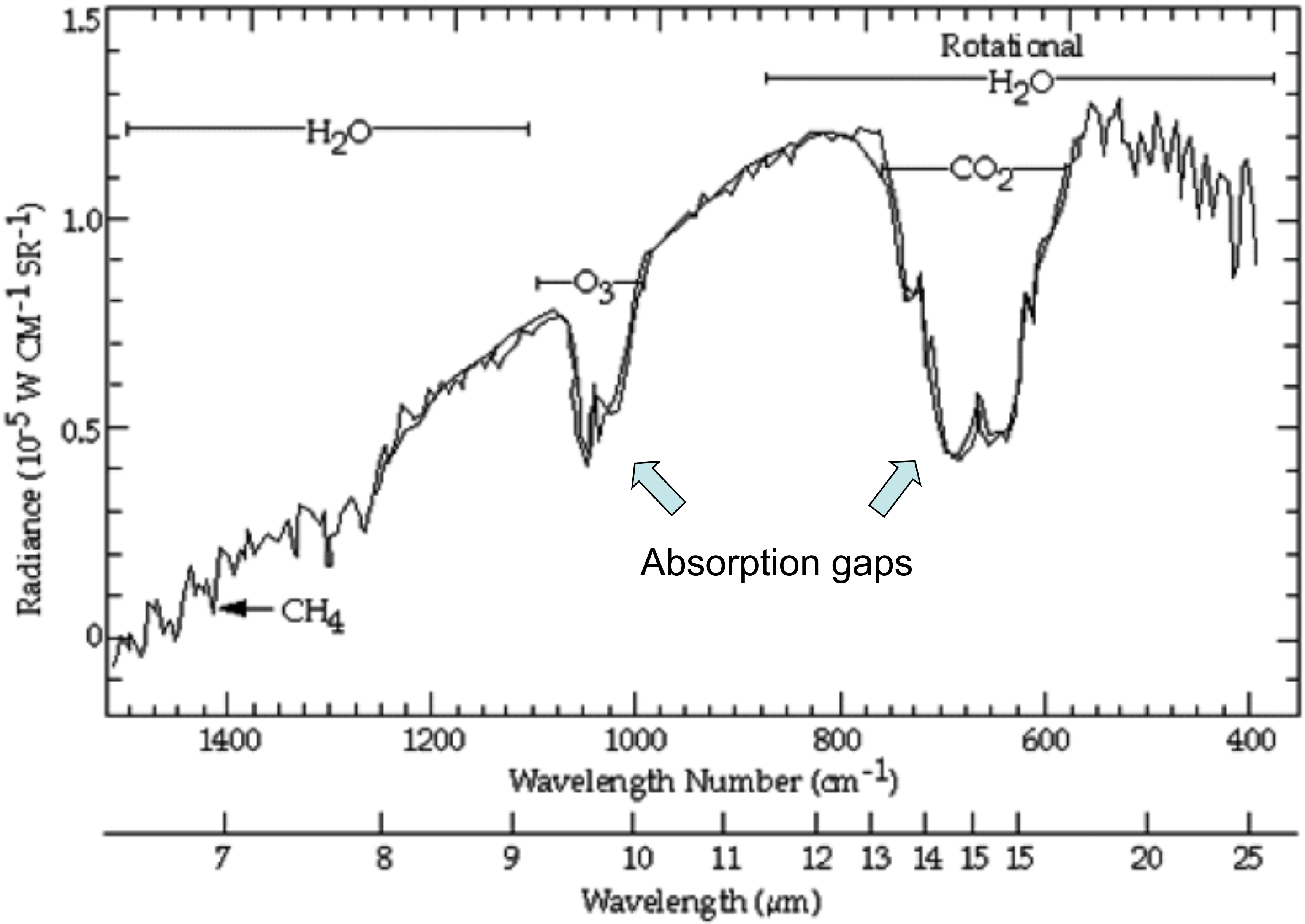
John Tyndall's experiment 1859



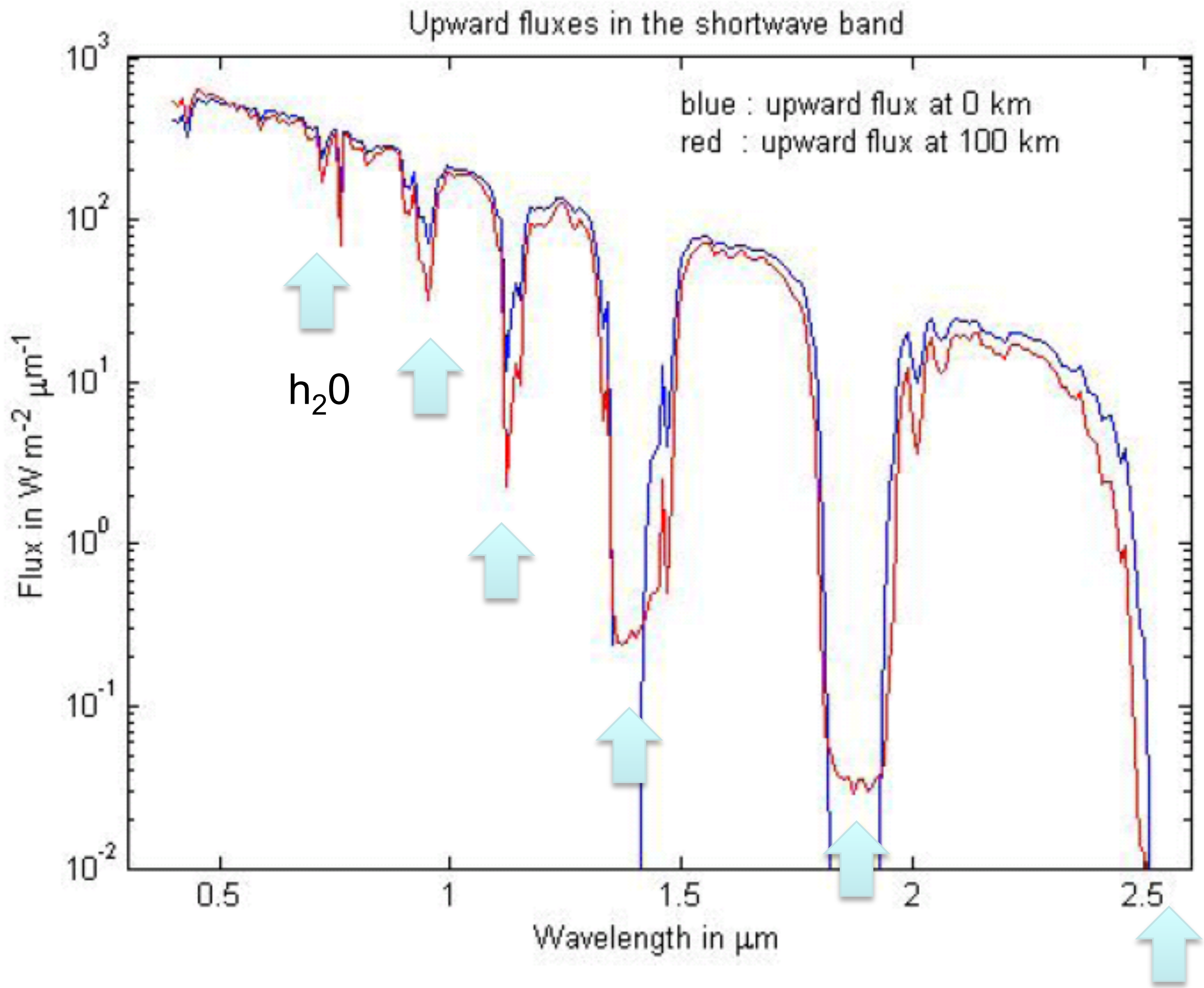


100
80
60
40
20
0
Total
100
50
0
O_2+O_3
100
50
0
CO_2
100
50
0
H_2O
5780 K
255 K
0.1
1.0
10.0
100.0
Wavelength (microns)
Solar radiation coming in
Earth thermal radiation going out

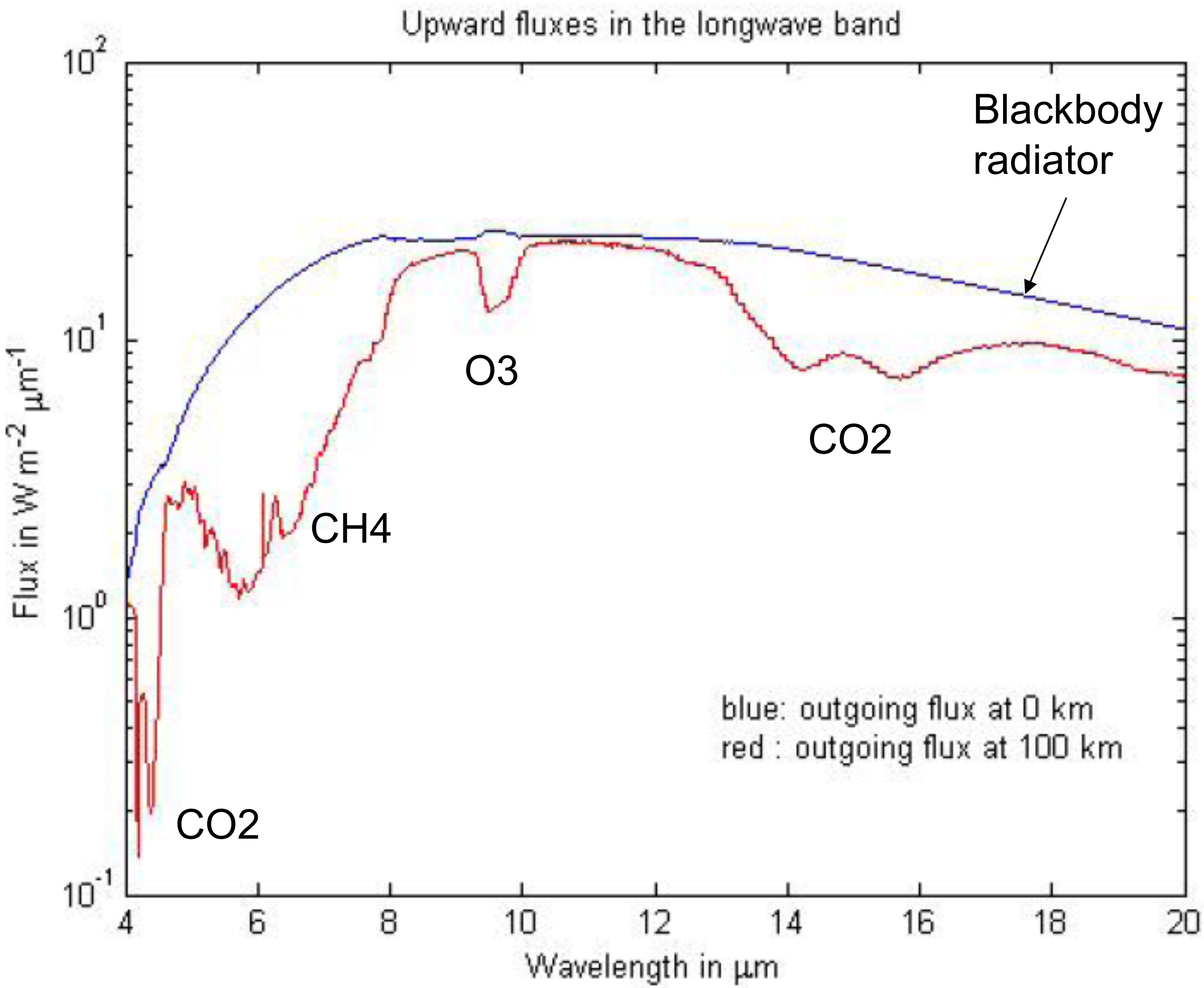
Infrared spectrum seen by Nimbus IV satellite



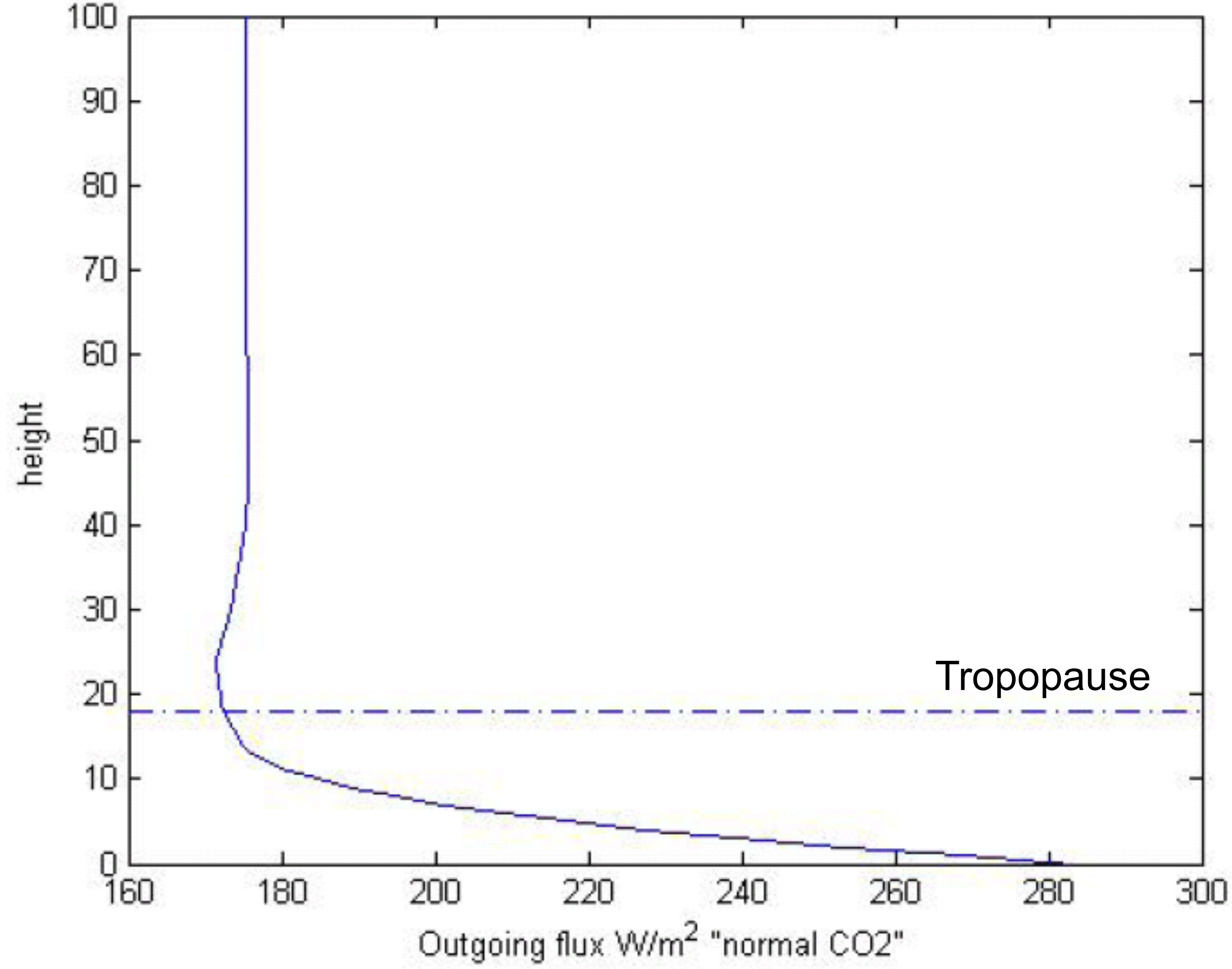
Outgoing flux of the Earth in the visible domain



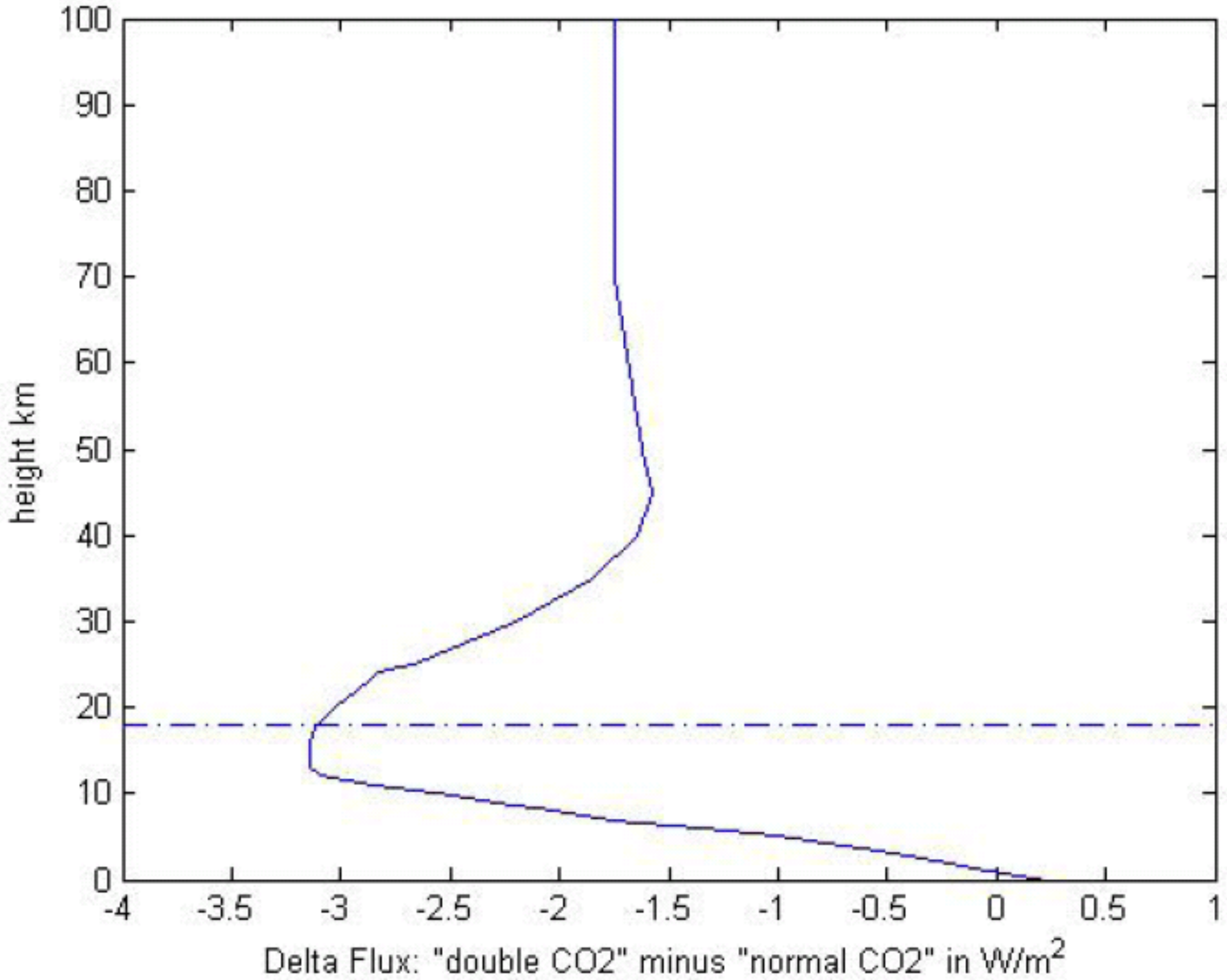
Infrared spectrum computed with COART code



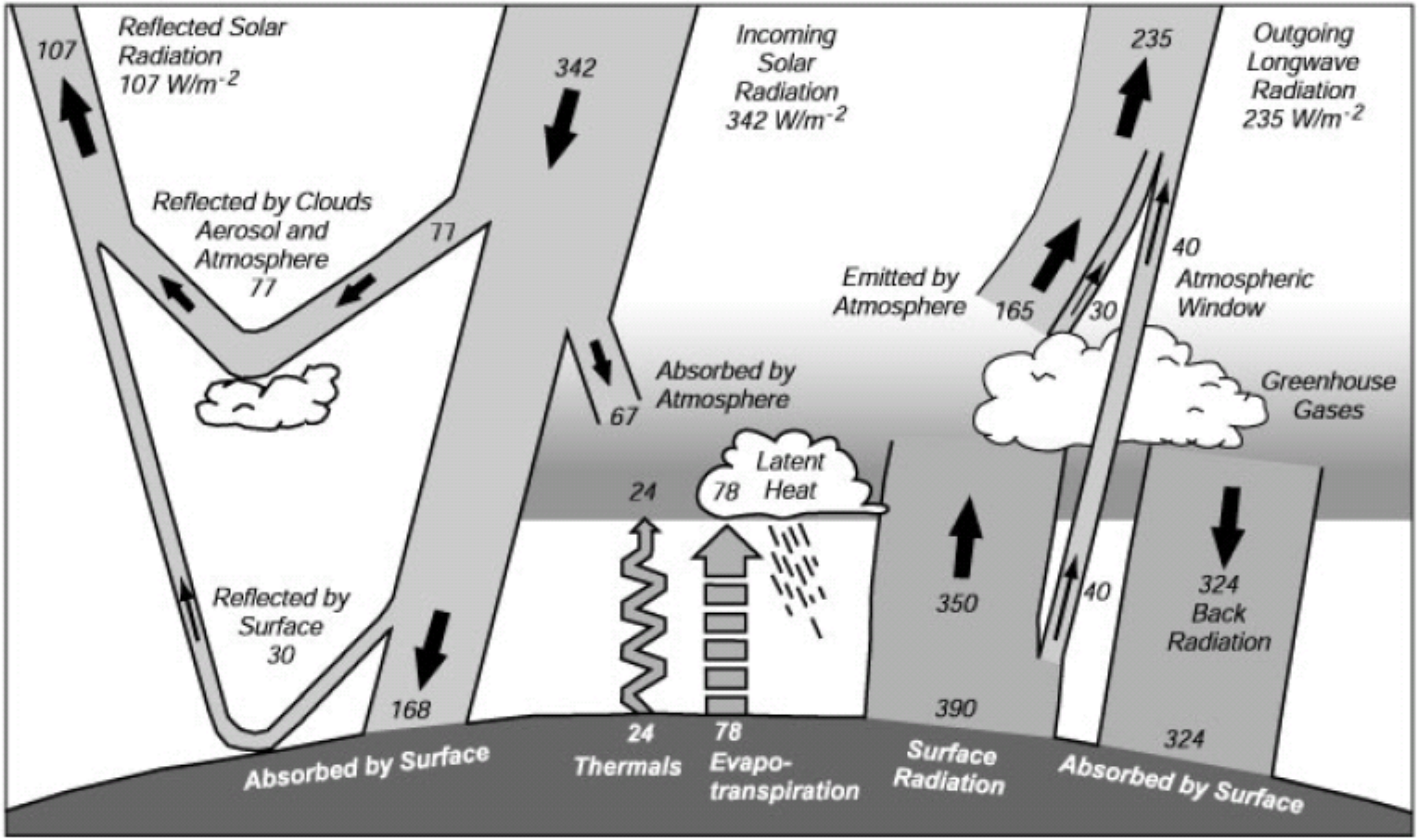
Outgoing long-wave forcing as a function of height



Effect on outgoing long-wave forcing wrt height as a result of doubling the present day carbon dioxide concentrations

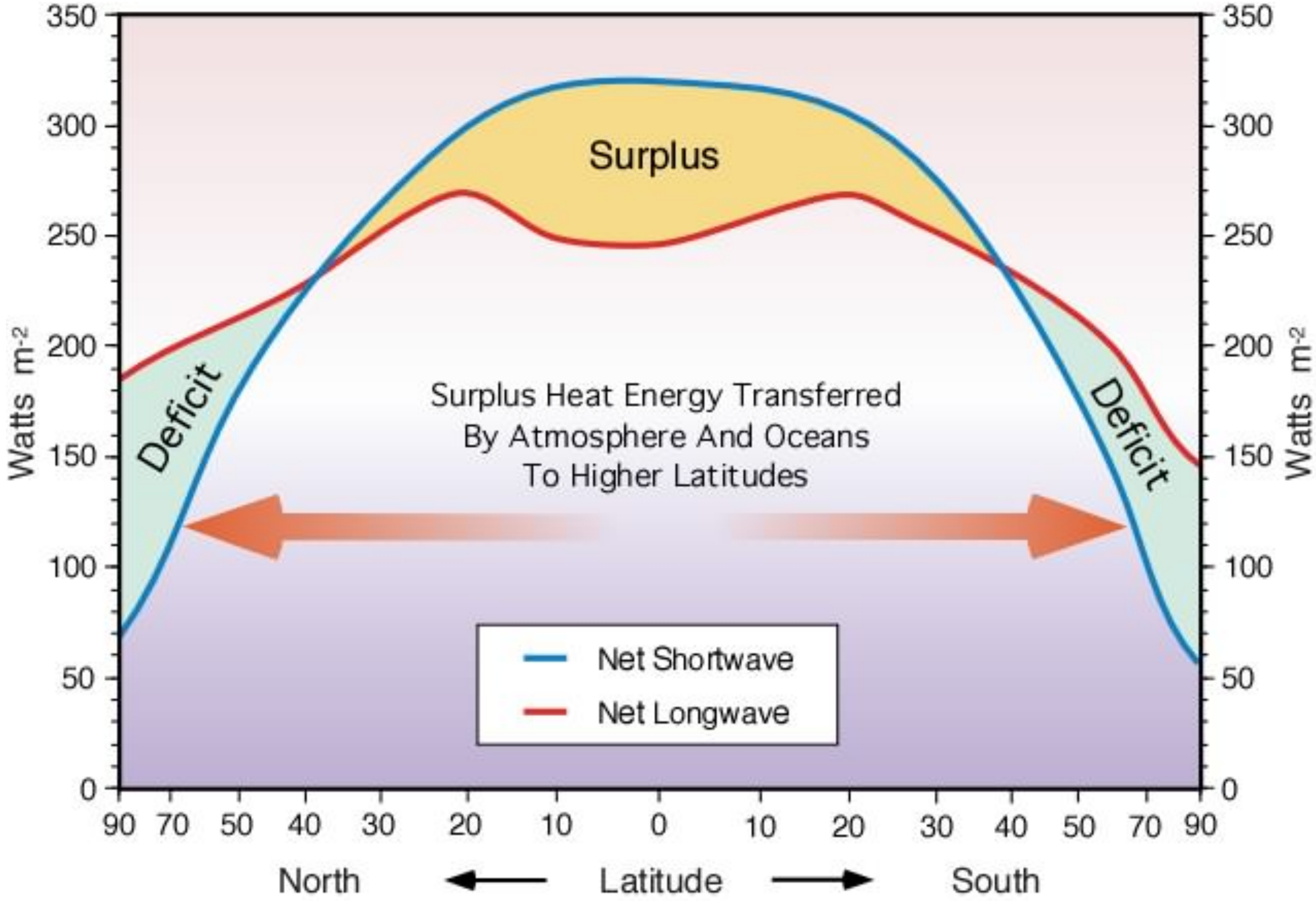


Earth's global radiation budget

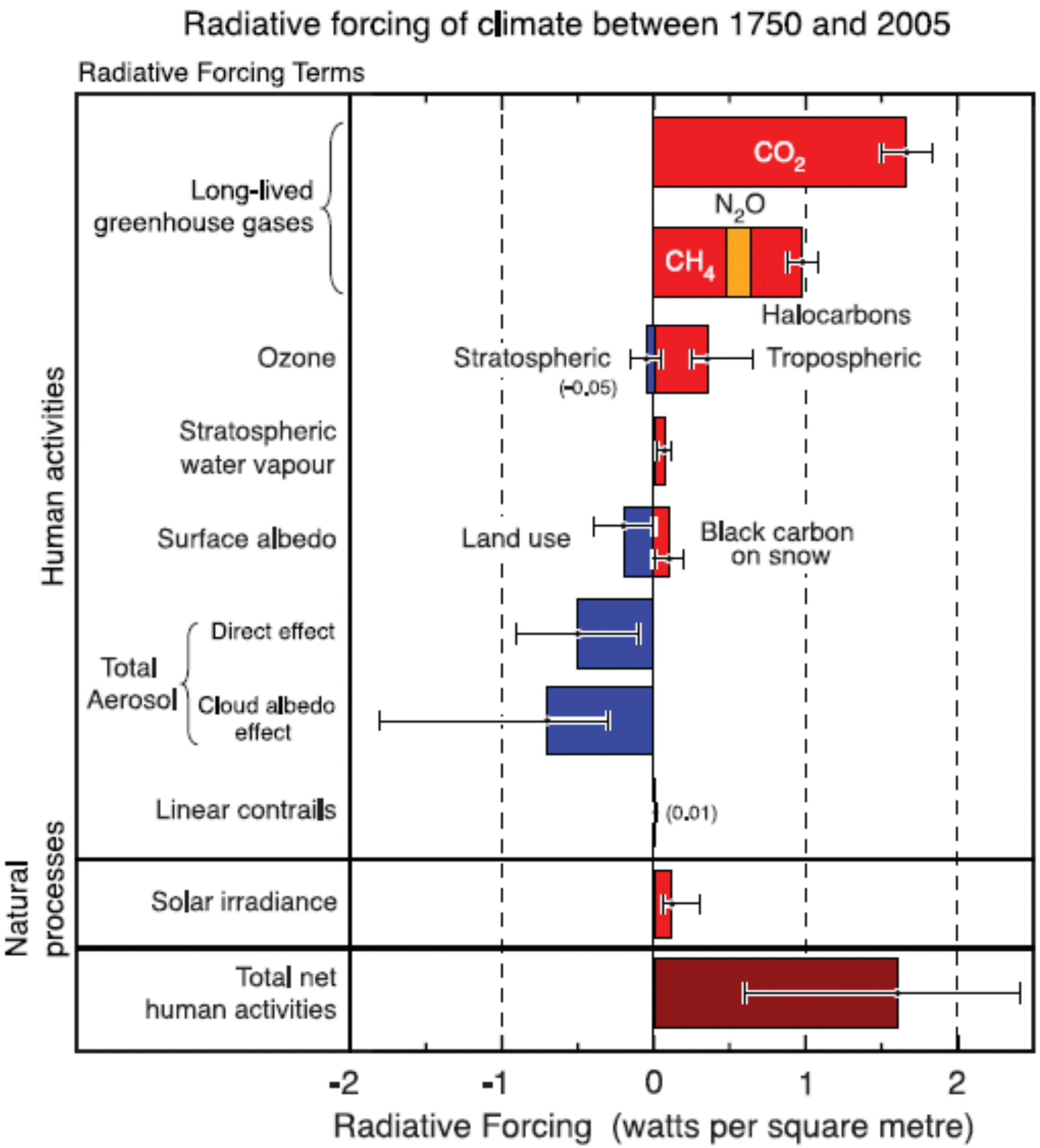


Source: Kiehl and Trenberth [1997]

Earth's radiation balance vs latitude



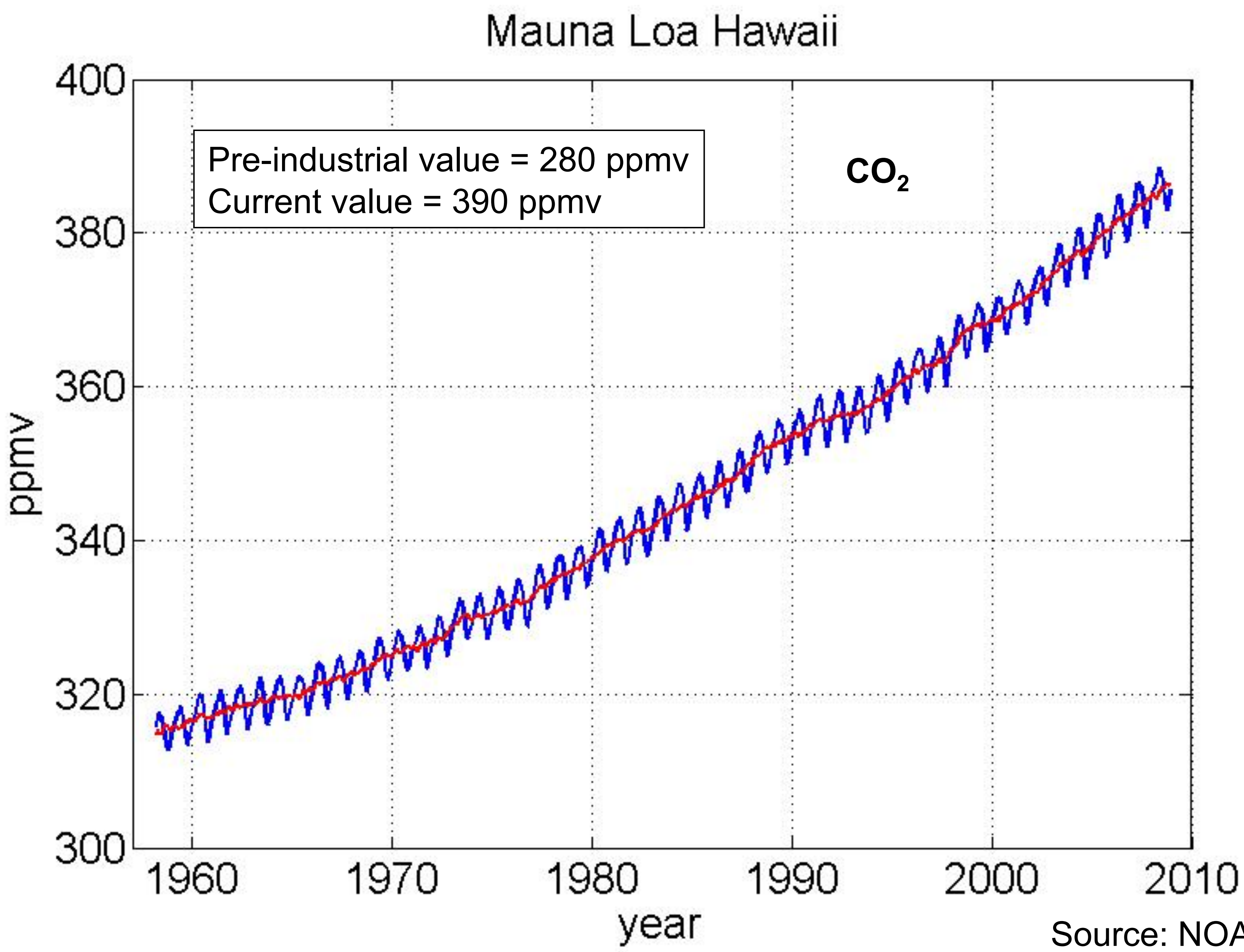
Source: http://www.physicalgeography.net/fundamentals/images/rad_balance_ERBE_1987.jpg


Radiative forcing of climate between 1750 and 2005
Radiative Forcing Terms
Long-lived greenhouse gases
CO_2
N_2O
CH_4
Halocarbons
Ozone
Stratospheric
(-0.05)
Tropospheric
Stratospheric water vapour
Surface albedo
Land use
Black carbon on snow
Total Aerosol
Direct effect
Cloud albedo effect
Linear contrails
(0.01)
Solar irradiance
Total net human activities
Human activities
Natural processes
-2
-1
0
1
2
Radiative Forcing (watts per square metre)

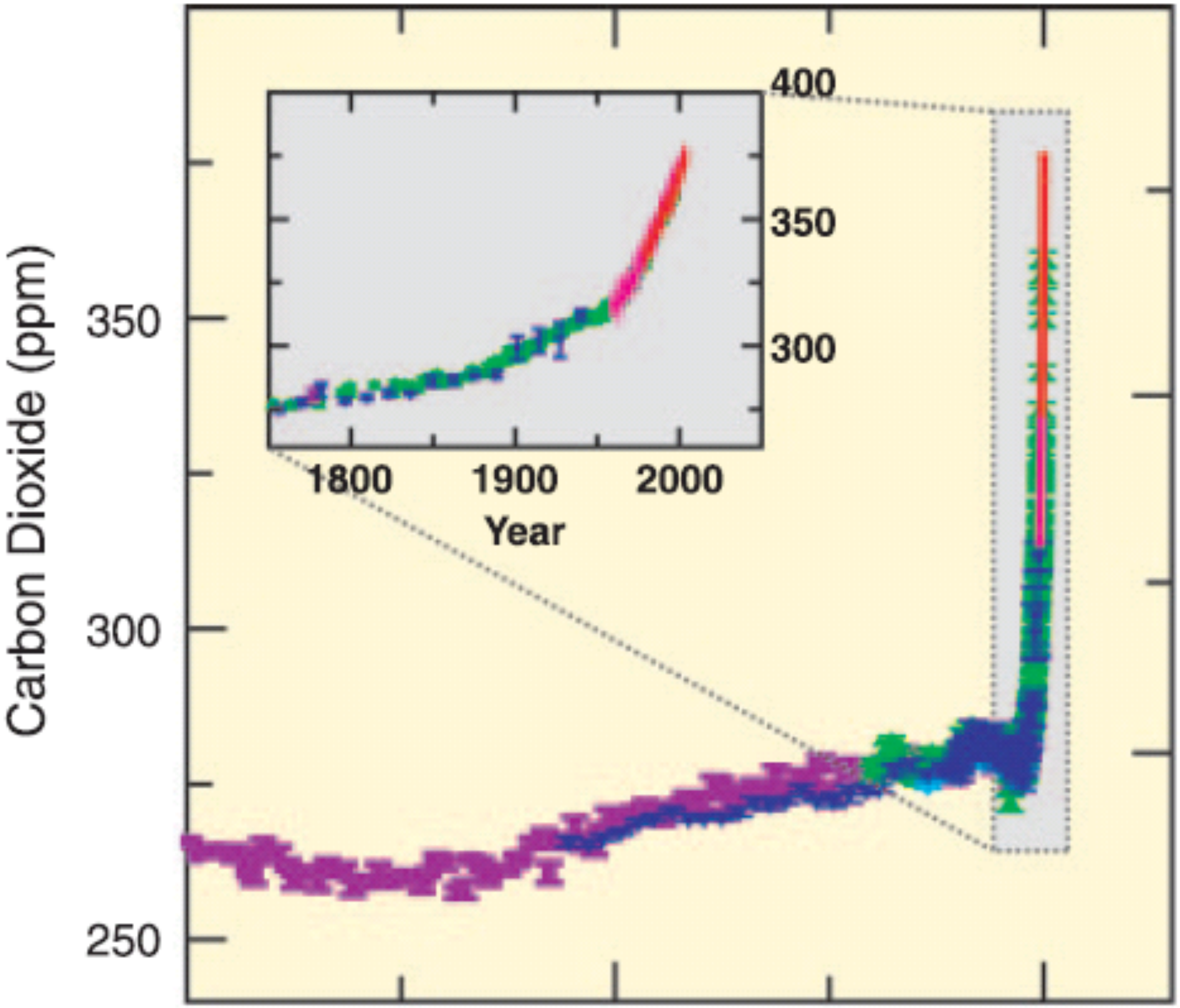
Source: IPCC

Climate change

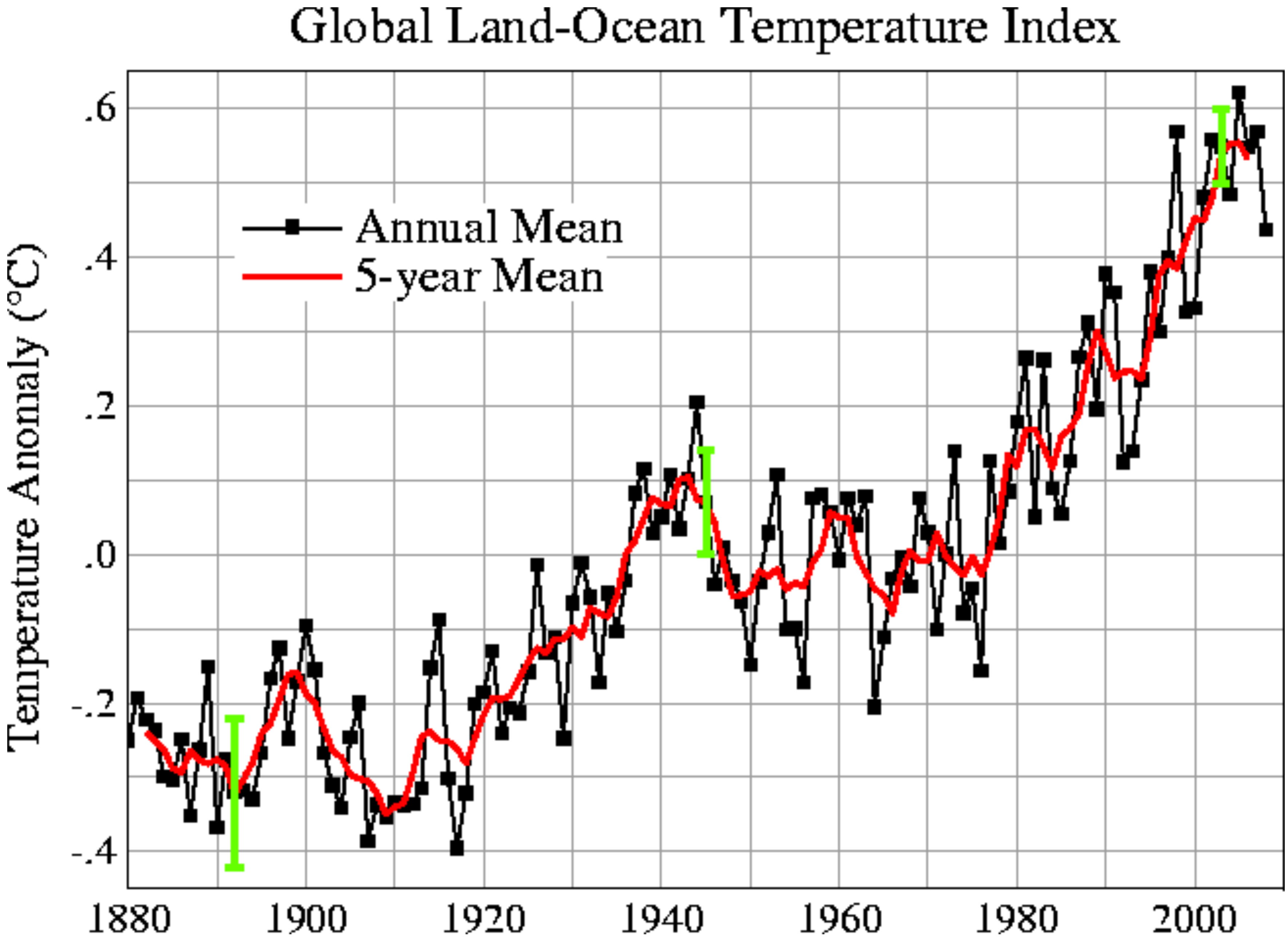
- The hockeystick curve(s)
  - CO_2 and temperature last 200 years
  - CO_2 and temperature last 500K years
  - Climate sensitivity (exercise)
- Changes in the cryosphere
  - GRACE, altimetry, InSAR
- Projected Sea level change


Mauna Loa Hawaii
Pre-industrial value = 280 ppmv
Current value = 390 ppmv
CO2
400
380
360
340
320
300
ppmv
1960
1970
1980
1990
2000
2010
year
Source: NOAA

Changes in GHGs from ice core and modern data

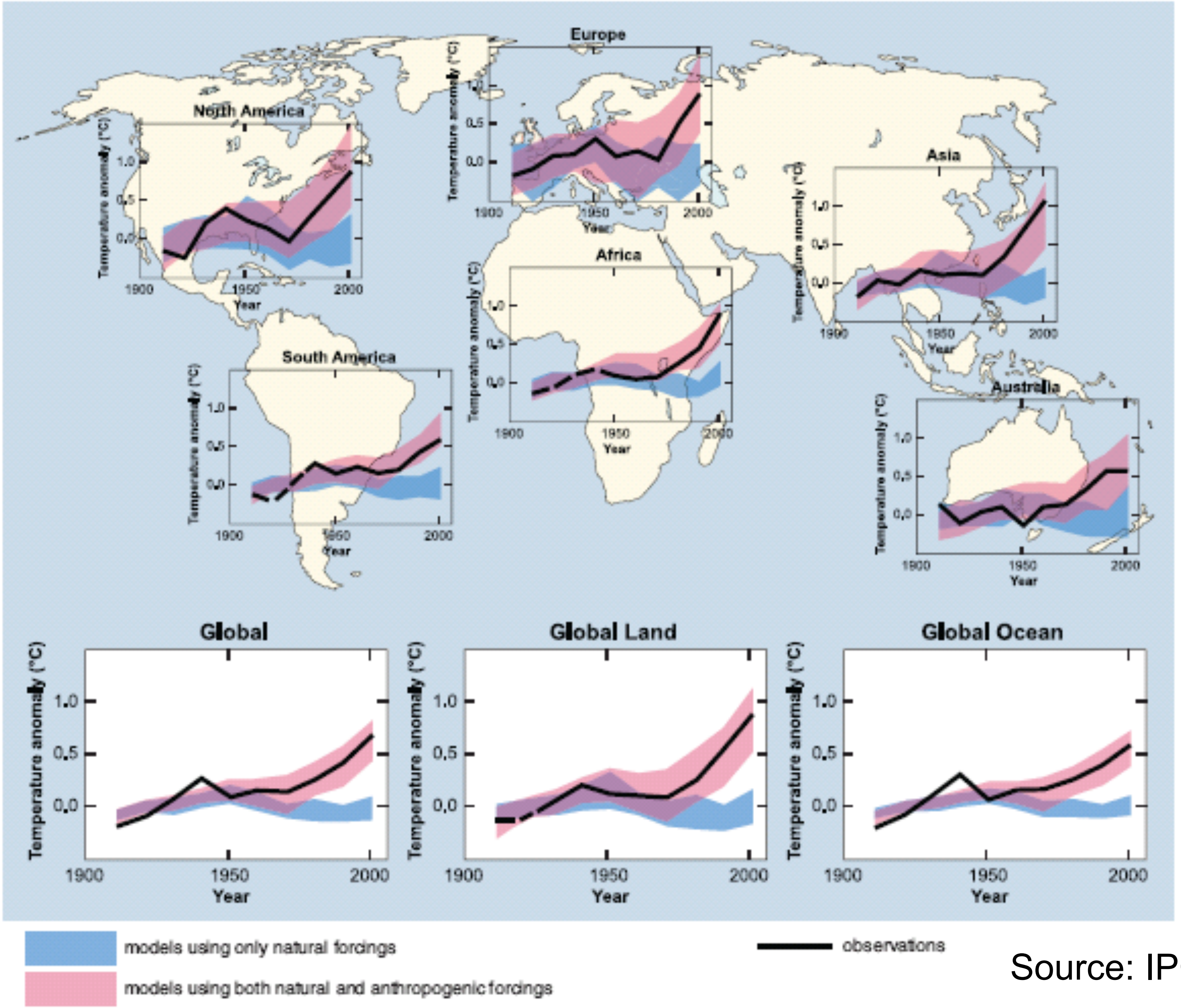


Source: IPCC


Global Land-Ocean Temperature Index
Annual Mean
5-year Mean
Temperature Anomaly (°C)
.6
.4
.2
.0
-.2
-.4
1880
1900
1920
1940
1960
1980
2000

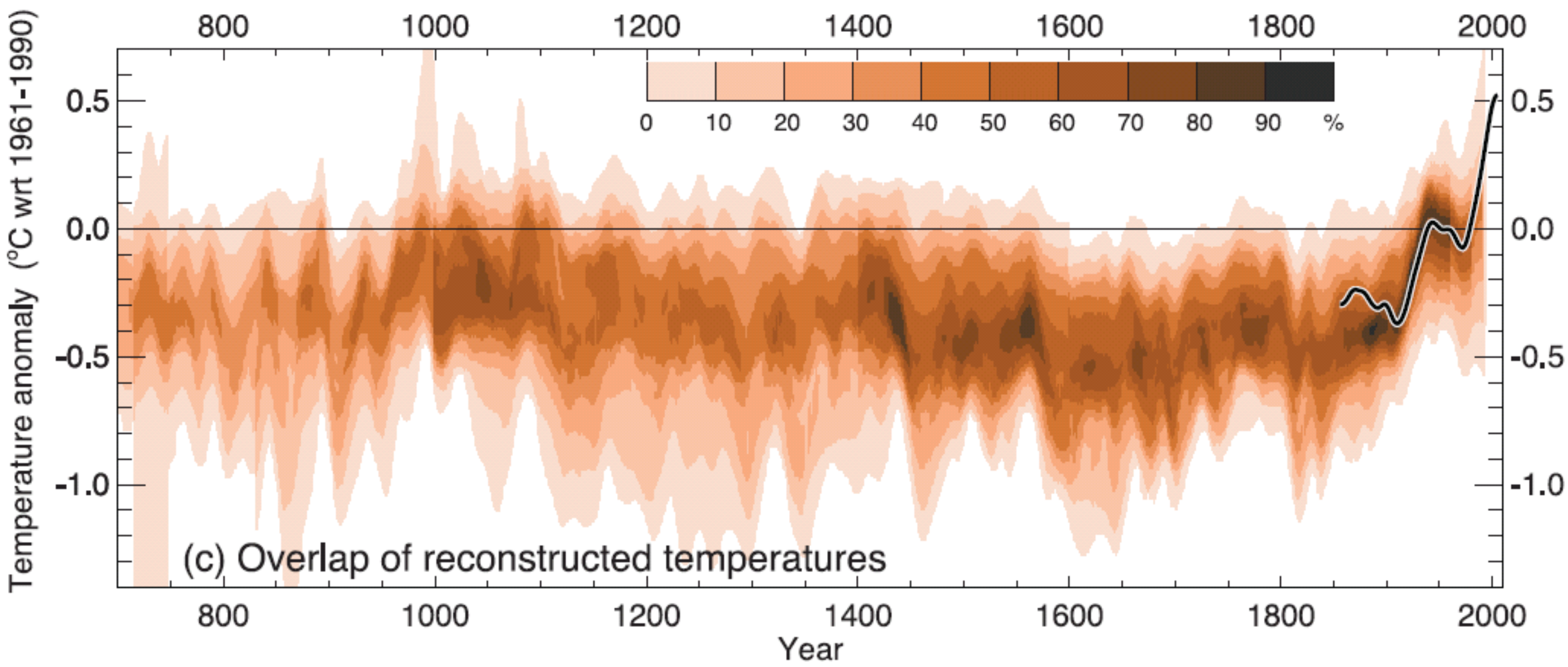
Source: GISS

Global and continental temperature change



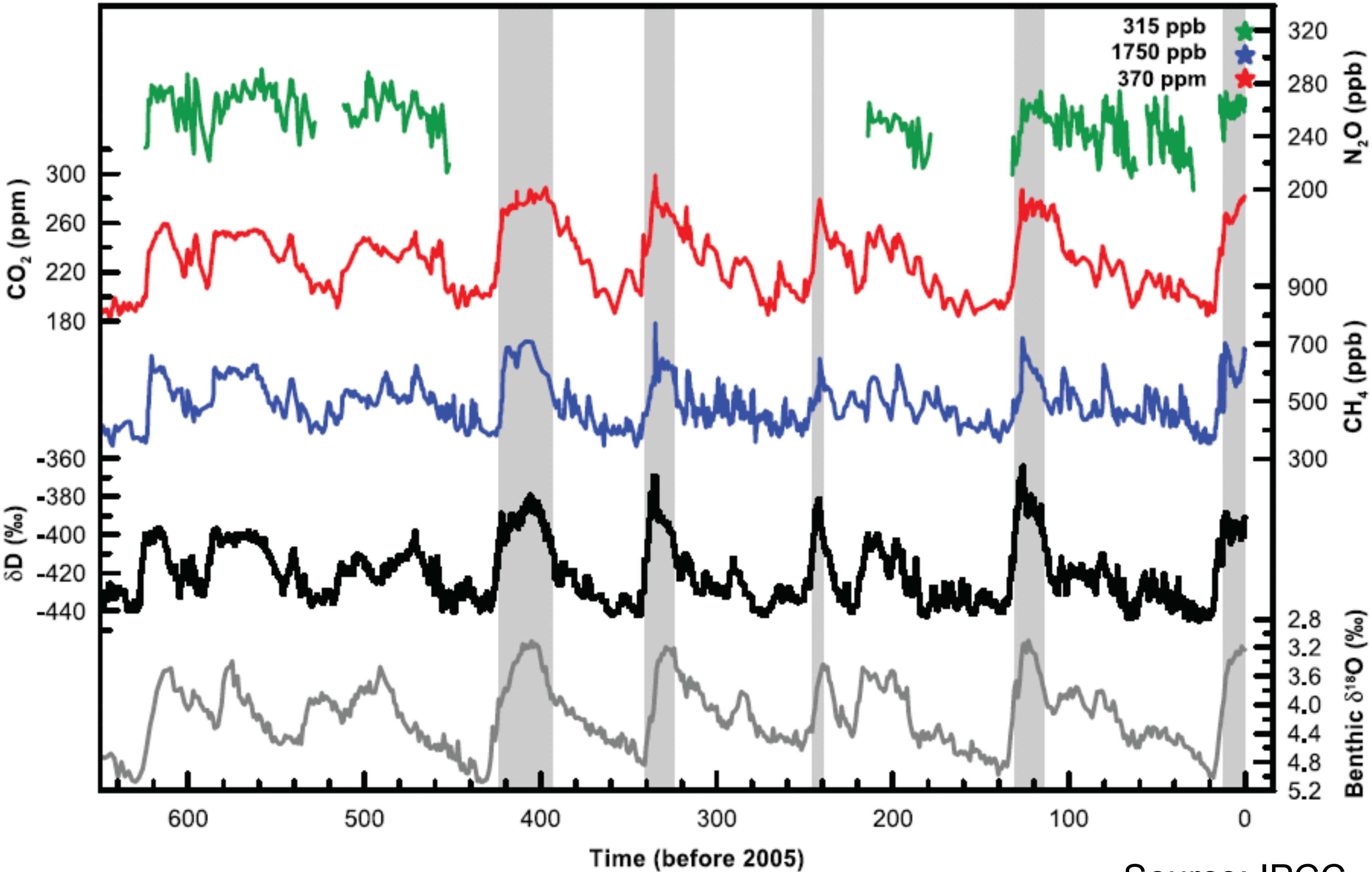
Source: IPCC

Tree ring proxy data



Source: IPCC

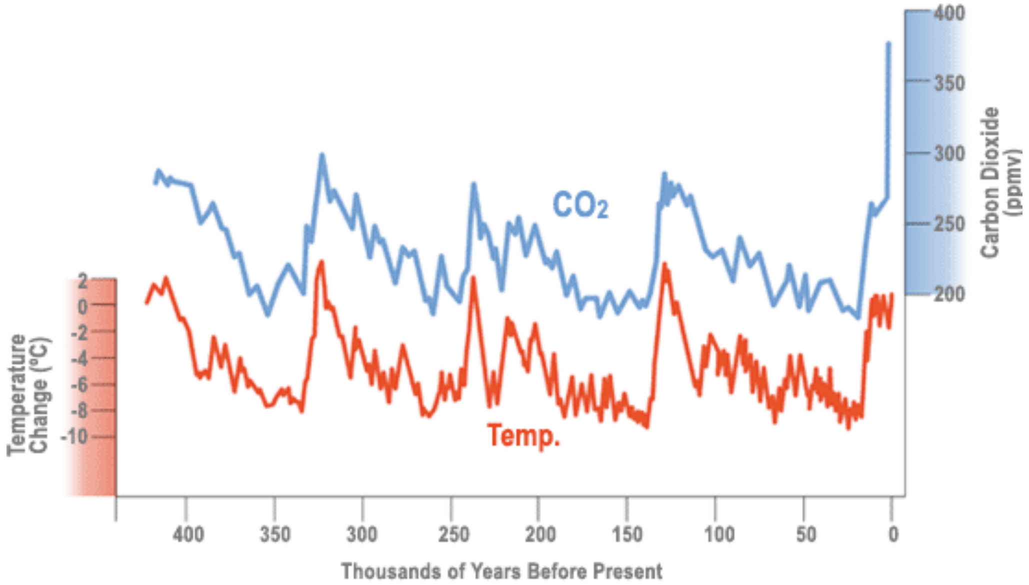
Paleo Climate

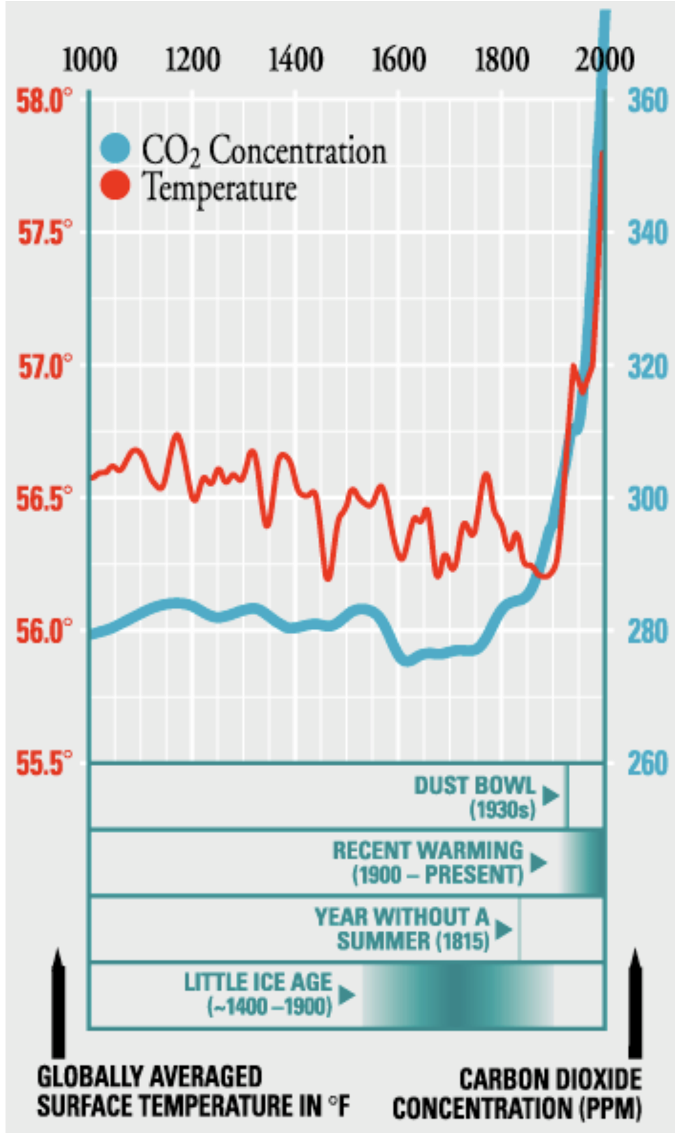


0.75+/-0.25 C/(W/m^2)

Source: IPCC

Exercise



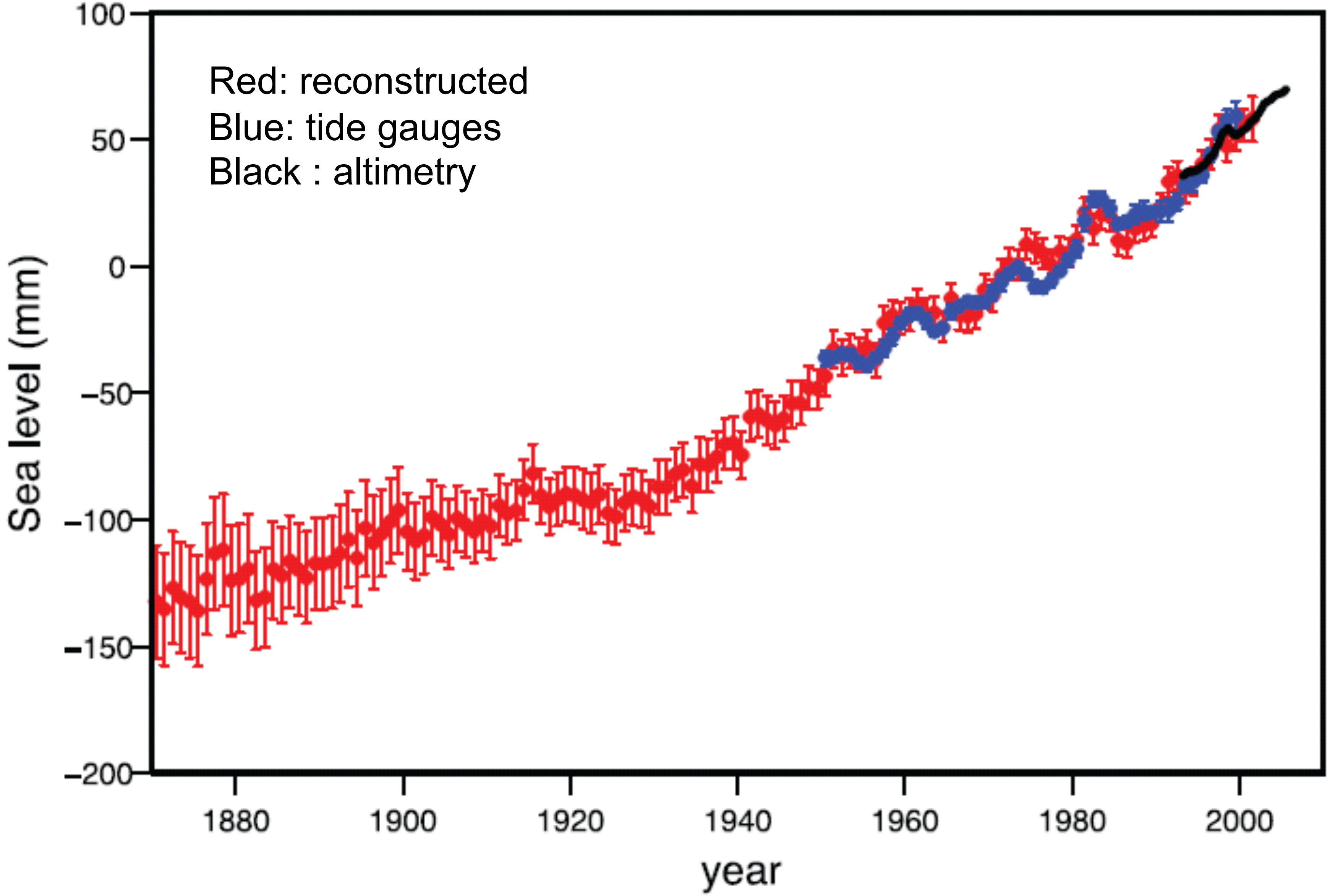


Given are GHG concentrations and temperature graphs, wanted is the climate sensitivity factor in K / (W/ m^2), also wanted is the projection of temperature for 2100

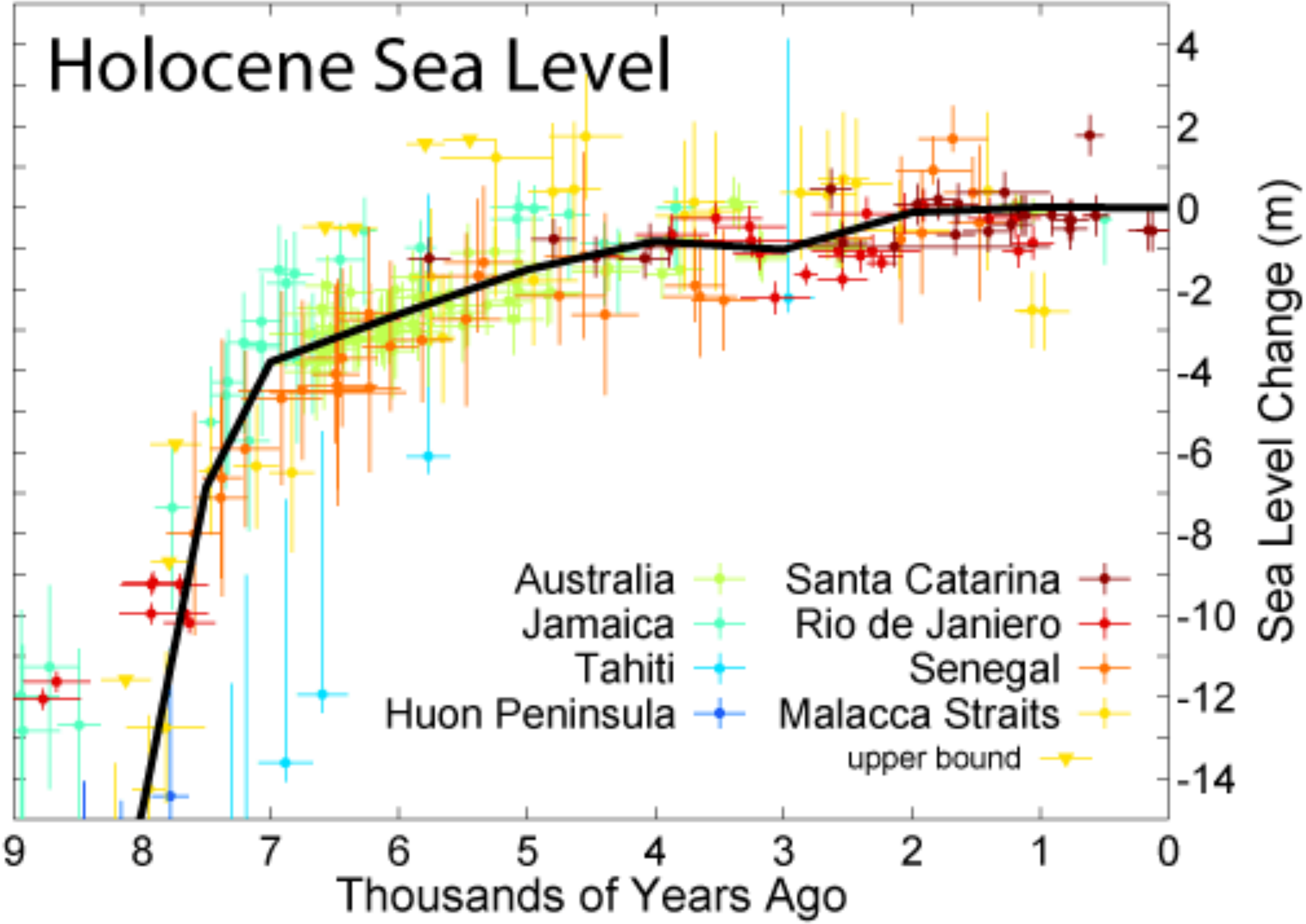
Consequence of global warming

- Sea level change
- De-glaciation
- Precipitation pattern changes
- Biodiversity changes
- Future warming

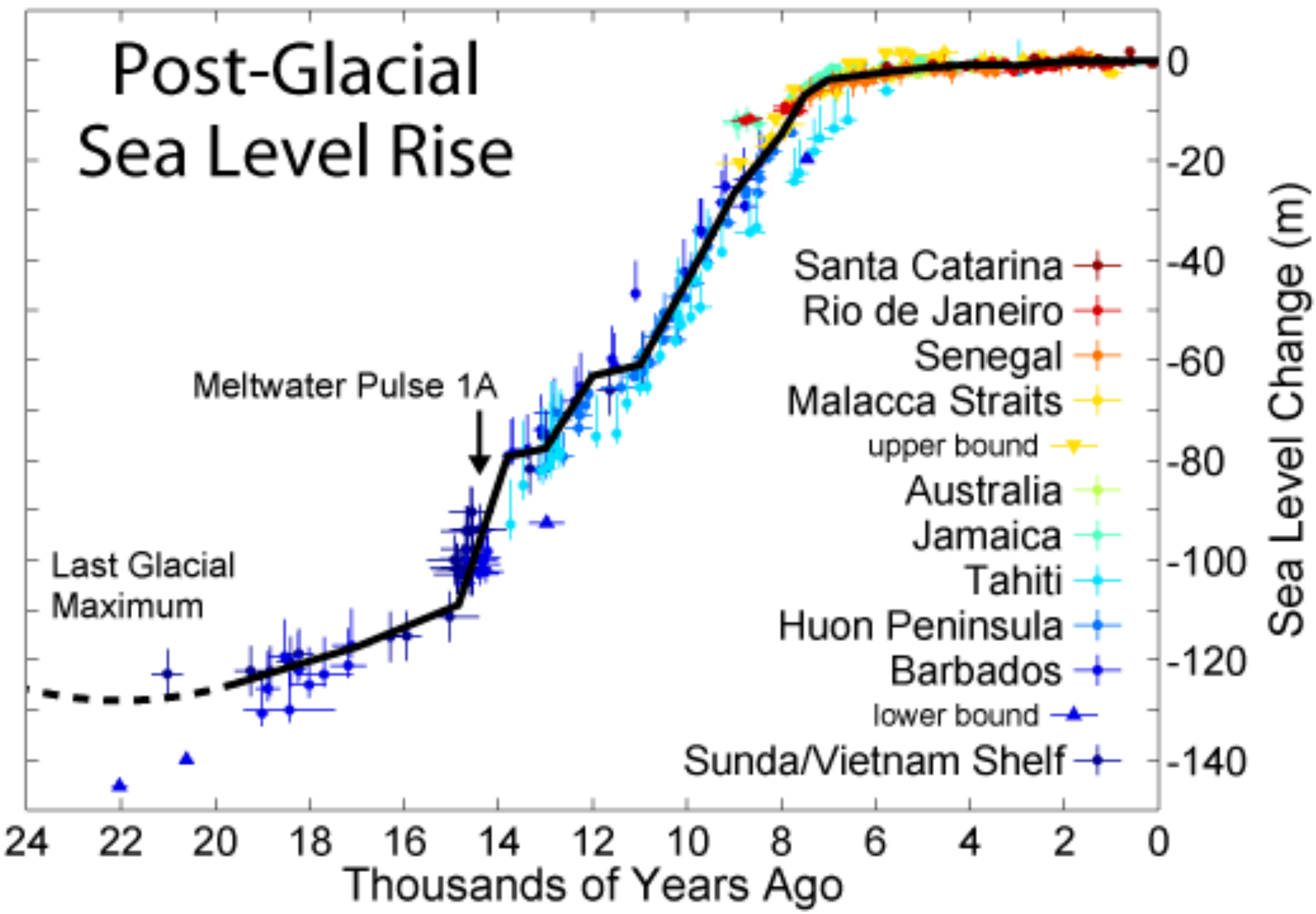
Sea level



Source: IPCC

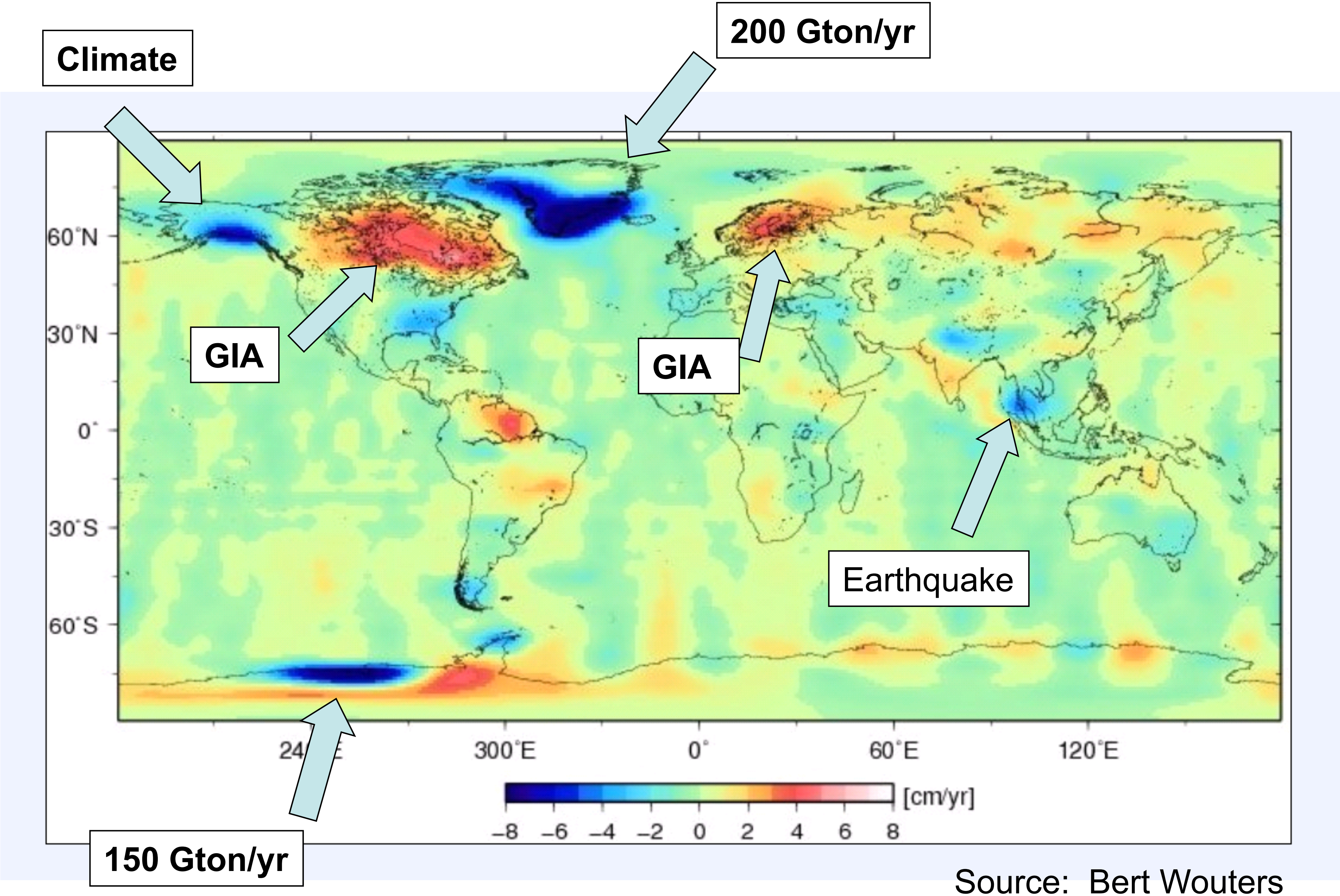


Source: wikipedia


Post-Glacial
Sea Level Rise
Meltwater Pulse 1A
Last Glacial
Maximum
Santa Catarina
Rio de Janeiro
Senegal
Malacca Straits
upper bound
Australia
Jamaica
Tahiti
Huon Peninsula
Barbados
lower bound
Sunda/Vietnam Shelf
0
-20
-40
-60
-80
-100
-120
-140
Sea Level Change (m)
24
22
20
18
16
14
12
10
8
6
4
2
0
Thousands of Years Ago

Source: wiki

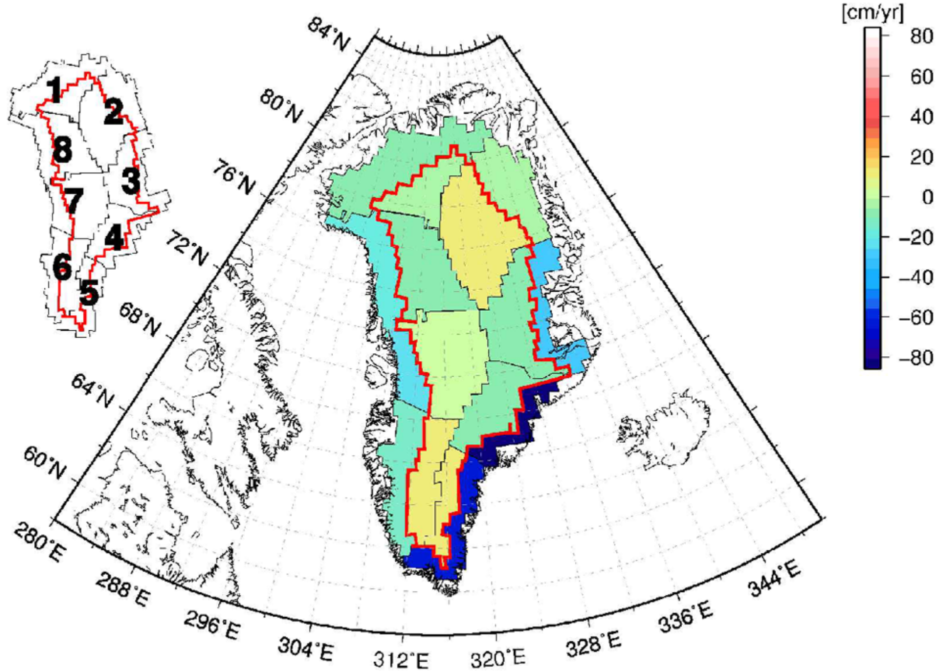
GRACE



Source: Bert Wouters

Local Trends (Gt/yr)

basin	< 2000 m	> 2000 m
1	-12 ± 4	-1 ± 4
2	-6 ± 4	19 ± 6
3	-25 ± 5	-10 ± 5
4	-49 ± 4	-7 ± 3
5	-51 ± 5	6 ± 6
6	-13 ± 5	11 ± 5
7	-14 ± 3	2 ± 5
8	-16 ± 4	-13 ± 5
Total	-186 ± 19	7 ± 18

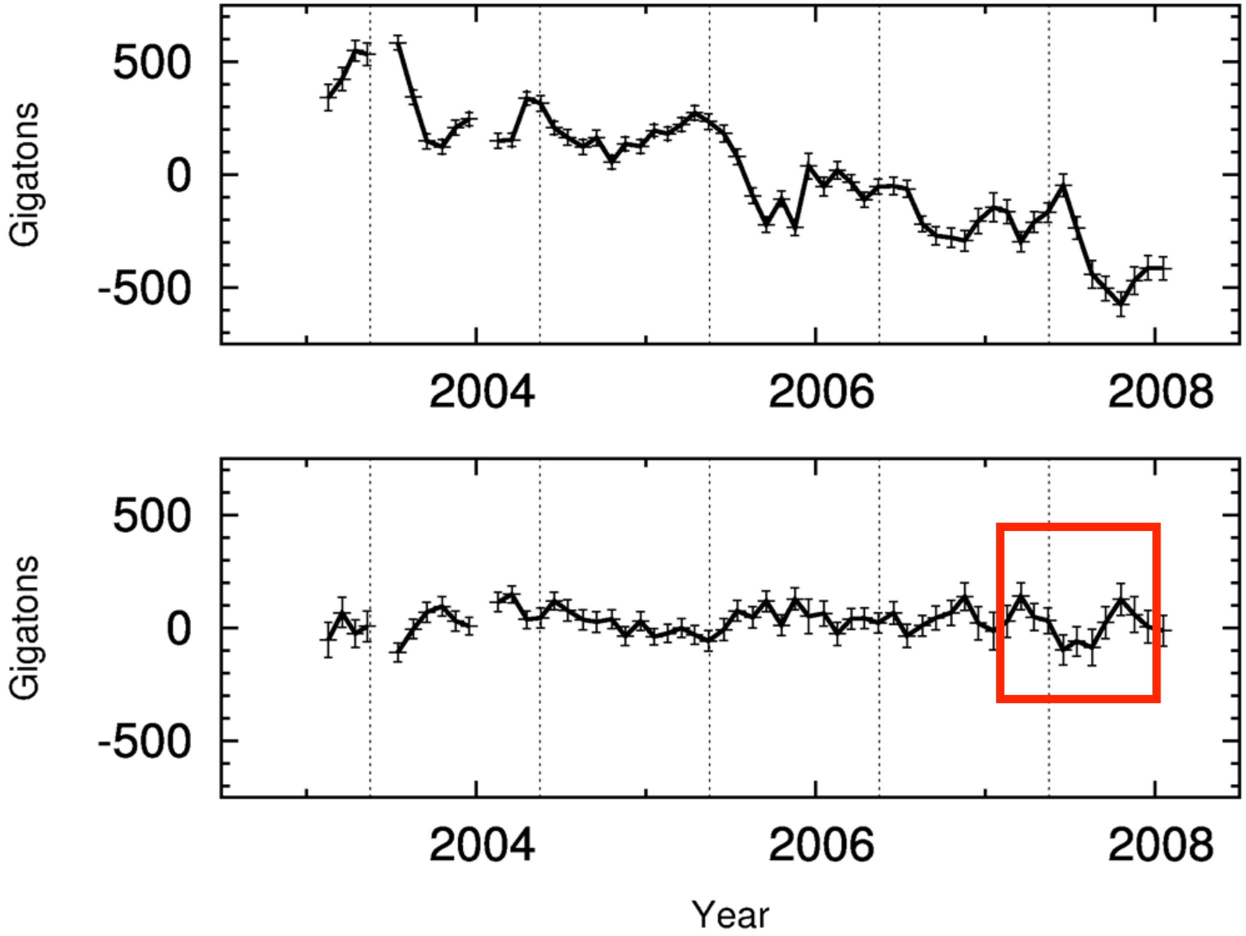


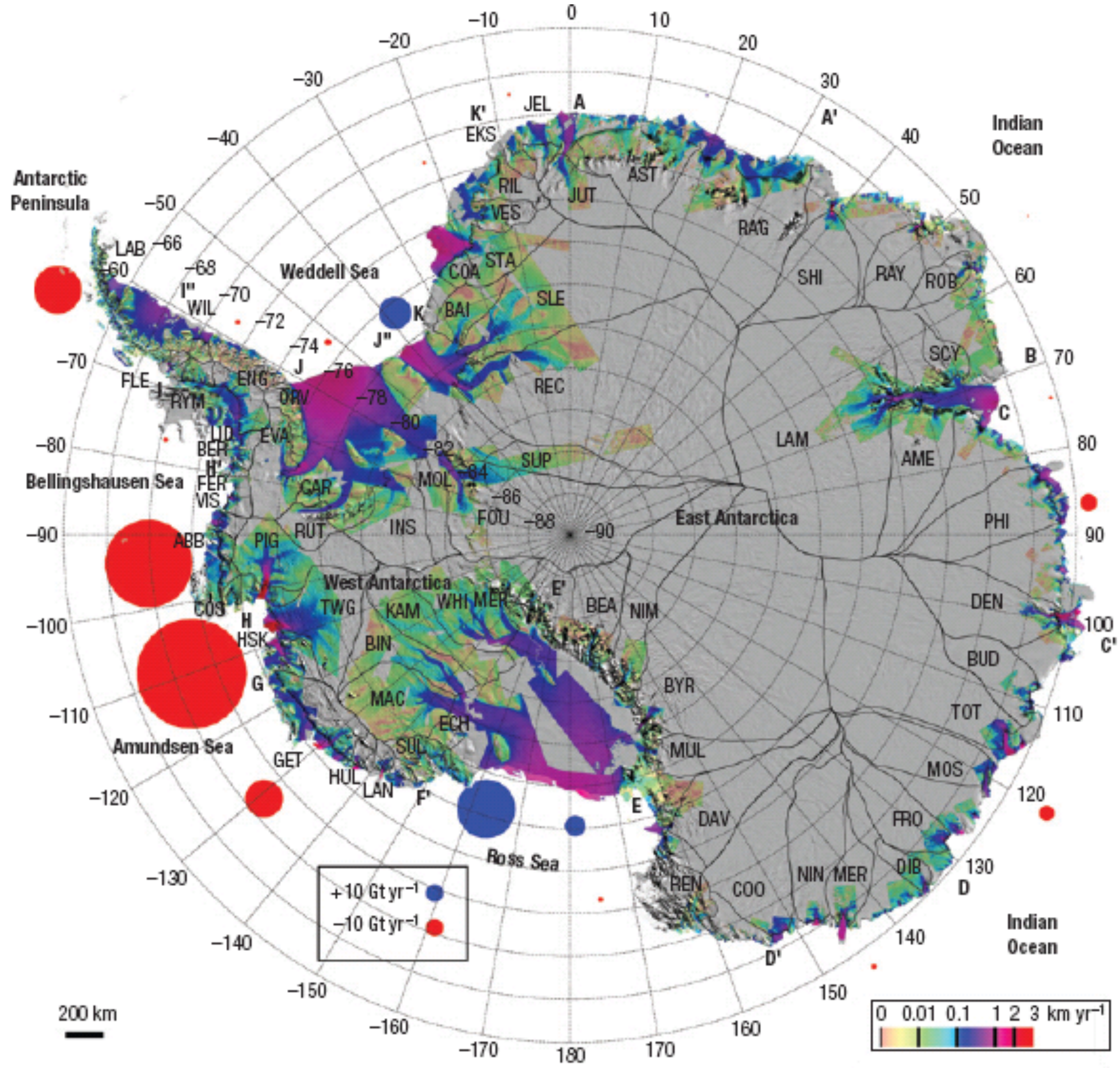
Overall Total:-179+/-25 Gt/yr

Regional mass loss

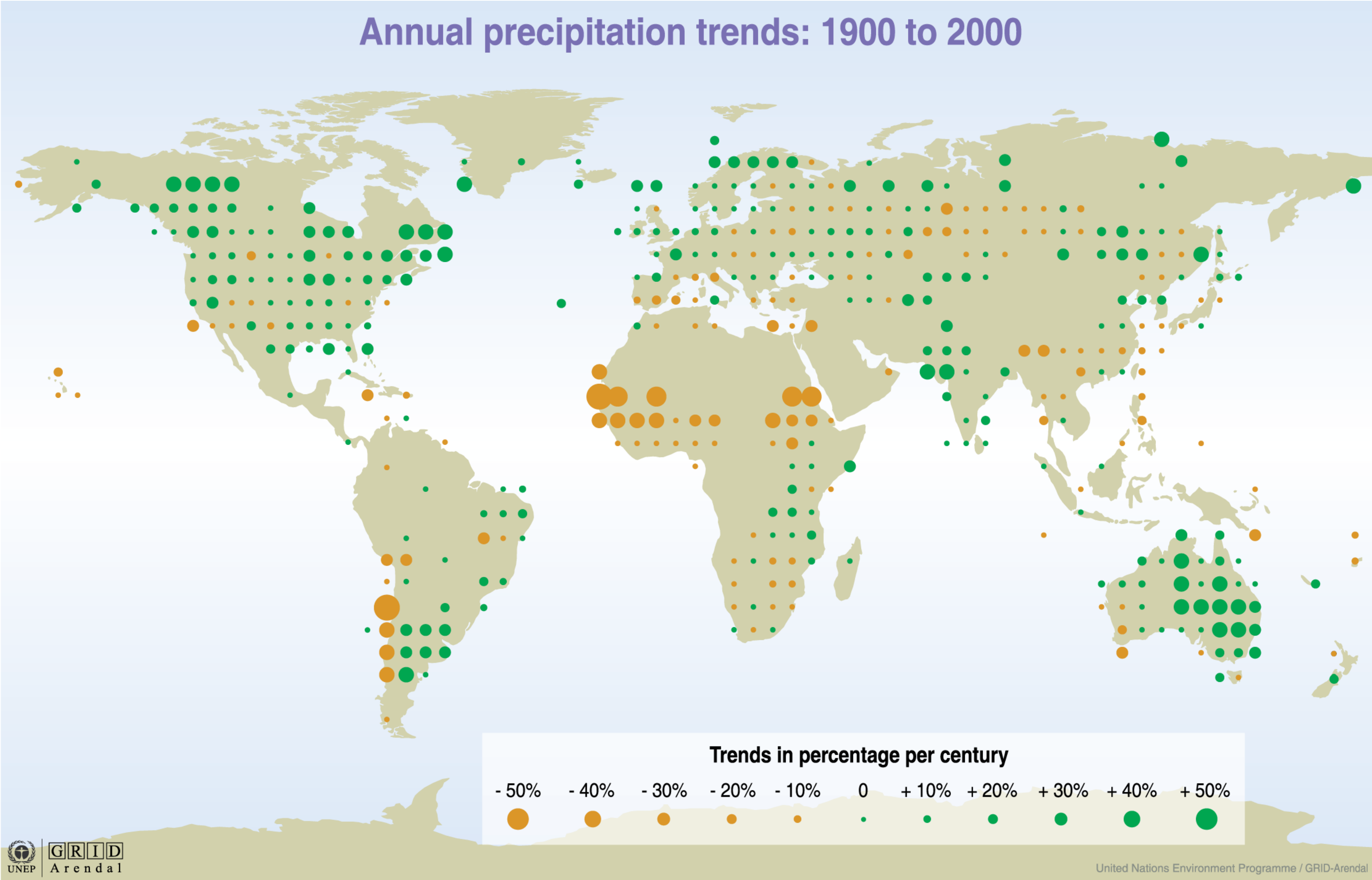
below 2000 m

above 2000 m

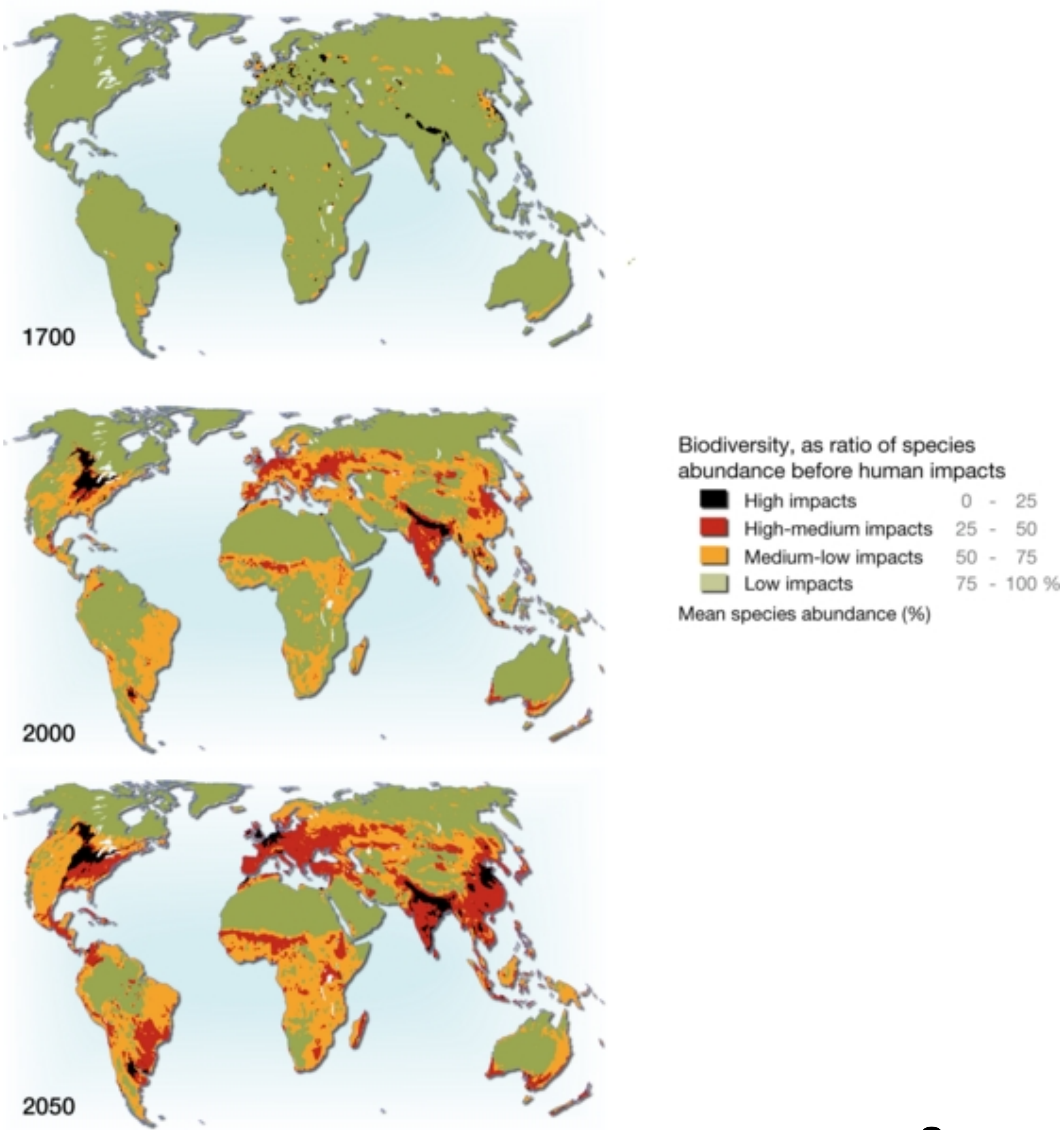



Antarctic Peninsula
Weddell Sea
Indian Ocean
Bellingshausen Sea
Amundsen Sea
Ross Sea
Indian Ocean
East Antarctica
West Antarctica
+10 Gt yr⁻¹
−10 Gt yr⁻¹
200 km
0 0.01 0.1 1 2 3 km yr⁻¹

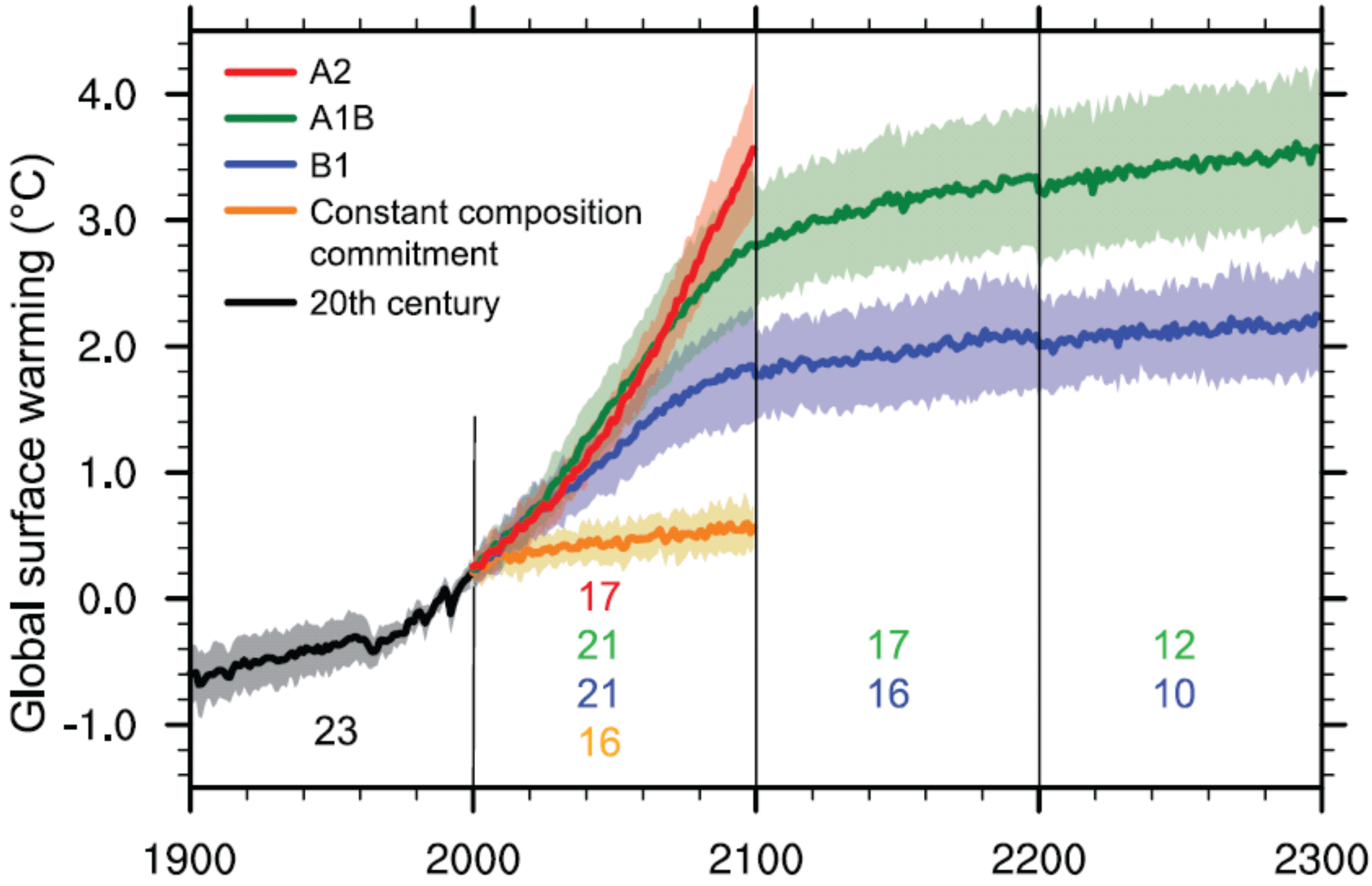
Rignot 2008



Source: United Nations Environmental Programme


1700
2000
2050
Biodiversity, as ratio of species abundance before human impacts
High impacts 0 - 25
High-medium impacts 25 - 50
Medium-low impacts 50 - 75
Low impacts 75 - 100 %
Mean species abundance (%)

Source: UNEP



A2
A1B
B1
Constant composition commitment
20th century
Global surface warming (°C)
4.0
3.0
2.0
1.0
0.0
-1.0
17
21
21
16
23
17
16
12
10
1900
2000
2100
2200
2300

Source: IPCC

Exercises

- Why is it necessary to distinguish between a longwave and shortware band when we discuss the Earth's radiation balance?
- How could you tell whether one Greenhouse gas is more effective than another?
- Explain why a satellite radiometer observes a peak inside all absorption gaps in the IR longwave band.
- Explain at least two independent techniques to reconstruct the mean atmospheric temperature up to 5000 years ago.
- How large is the contribution of Greenland's ice sheet ablation signal to the global mean sea level.
- Explain how INSar over glacier systems helps to estimate a mass balance of that ice sheet.