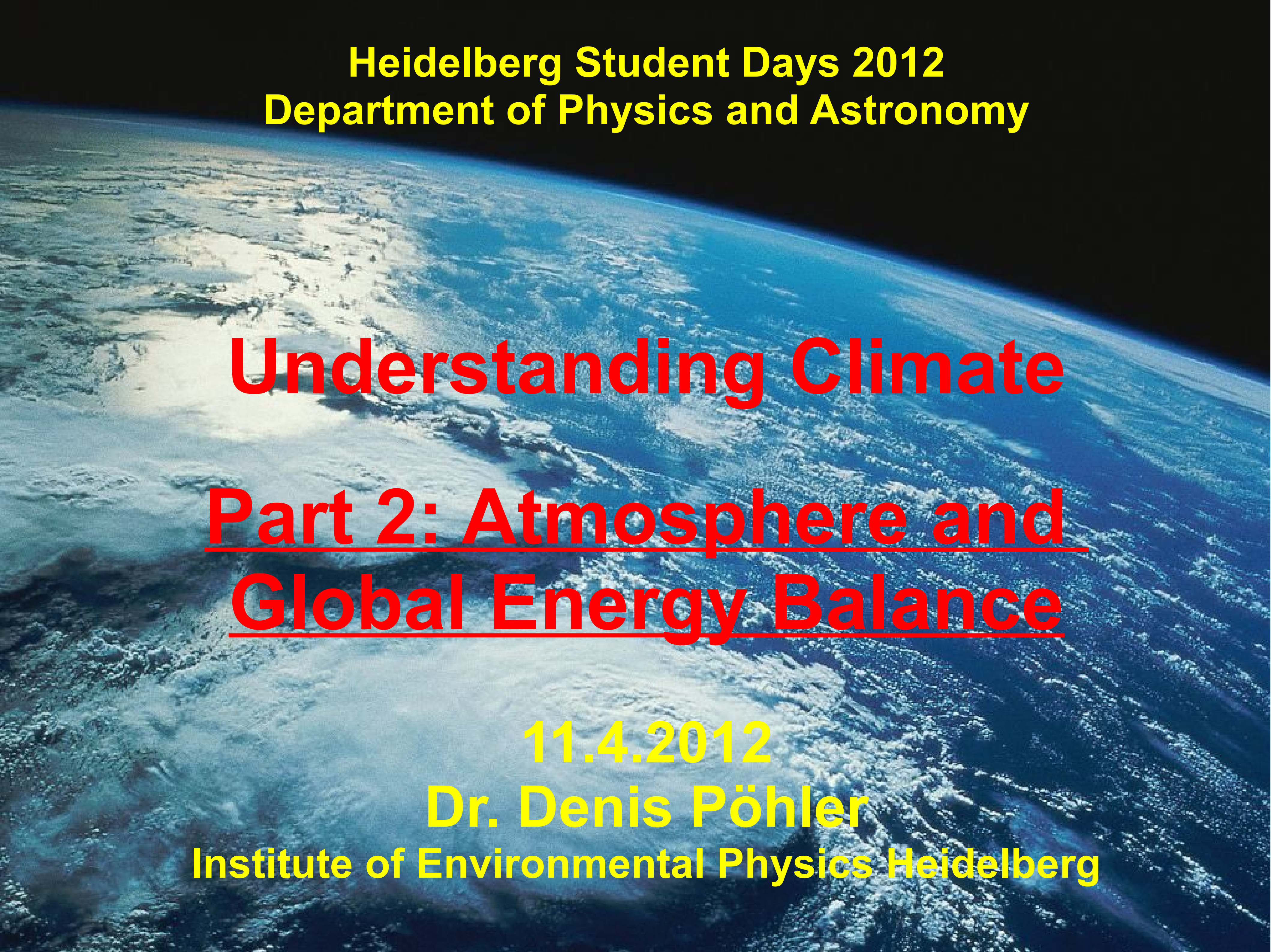


Heidelberg Student Days 2012
Department of Physics and Astronomy

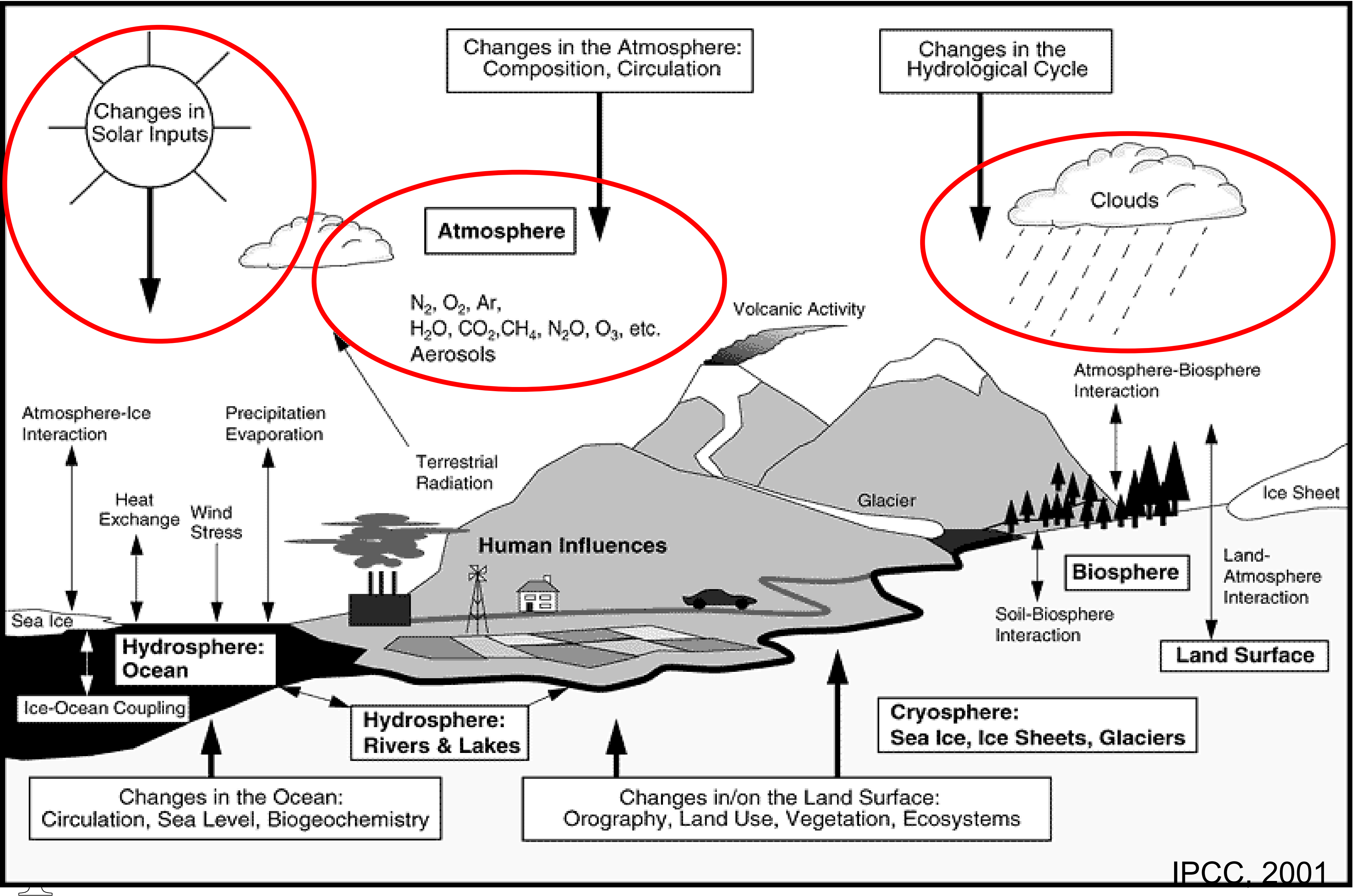
Understanding Climate

Part 2: Atmosphere and Global Energy Balance

11.4.2012
Dr. Denis Pöhler
Institute of Environmental Physics Heidelberg

Content

Overview Earth System Science



Changes in Solar Inputs

Changes in the Atmosphere: Composition, Circulation

Changes in the Hydrological Cycle

Atmosphere

N_2, O_2, Ar,
H_2O, CO_2,CH_4, N_2O, O_3, etc.
Aerosols

Clouds

Volcanic Activity

Atmosphere-Biosphere Interaction

Atmosphere-Ice Interaction

Precipitatien Evaporation

Terrestrial Radiation

Heat Exchange

Wind Stress

Glacier

Ice Sheet

Human Influences

Biosphere

Land-Atmosphere Interaction

Soil-Biosphere Interaction

Sea Ice

Hydrosphere: Ocean

Ice-Ocean Coupling

Hydrosphere: Rivers & Lakes

Land Surface

Cryosphere: Sea Ice, Ice Sheets, Glaciers

Changes in the Ocean: Circulation, Sea Level, Biogeochemistry

Changes in/on the Land Surface: Orography, Land Use, Vegetation, Ecosystems

IPCC, 2001

I. The Atmosphere: Basics

The atmosphere is a vital part of our environment:

- provides protective layer for life (stratospheric O_3)
- atmospheric compounds are essential for metabolism (e.g. O_2, CO_2)
- is part (compartment) of the Earth system
- connects different compartments & is driving force in climate system
- determines cycling of energy and matter
- is a complex dynamical system (chaotic motion)
- is a very thin & extended layer (7-8 km at surface pressure)

Composition of the Atmosphere

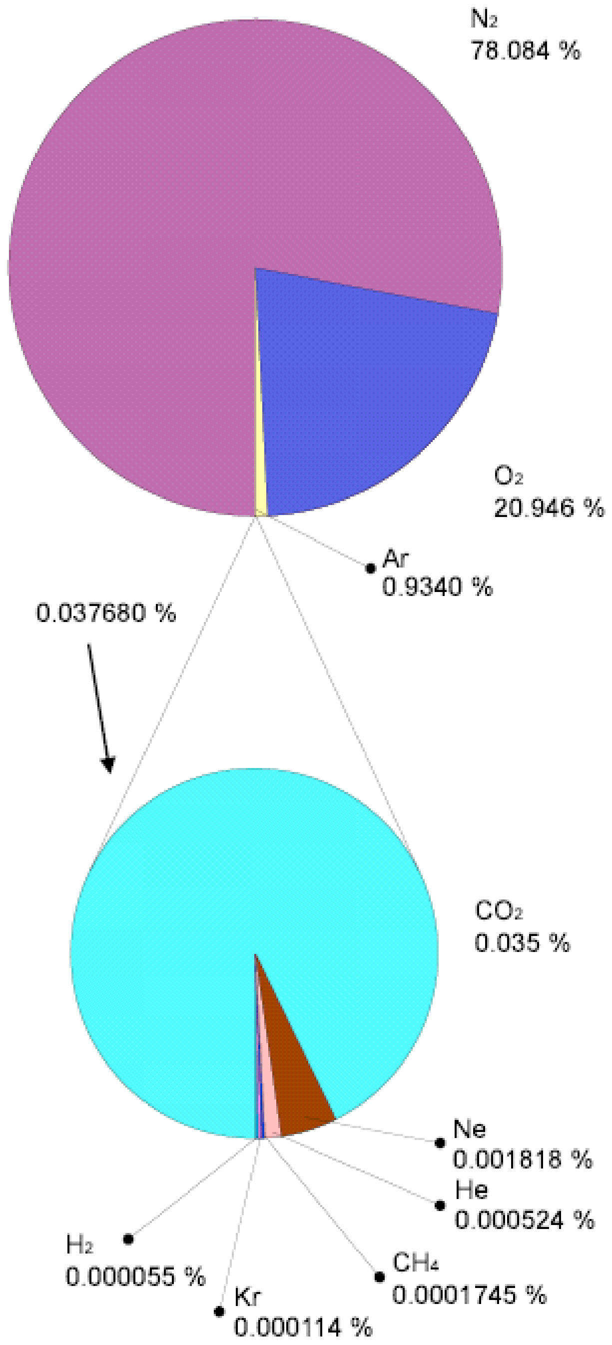
name	chemical formula	relative abundance [%]
nitrogen	N_2	77.9
oxygen	O_2	20.95
argon	Ar	0.93
carbon dioxide	CO_2	0.036
neon	Ne	0.0018
helium	He	0.0005
water vapour	H_2O	10^{-5} - 4
methane	CH_4	0.00017
krypton	Kr	0.00011
hydrogen	H_2	0.00005
ozone	O_3	$1 \cdot 10^{-6} - 1 \cdot 10^{-3}$

Nitrogen N_2: $M_{N2} = 28{,}015$ kg kmol^{-1}

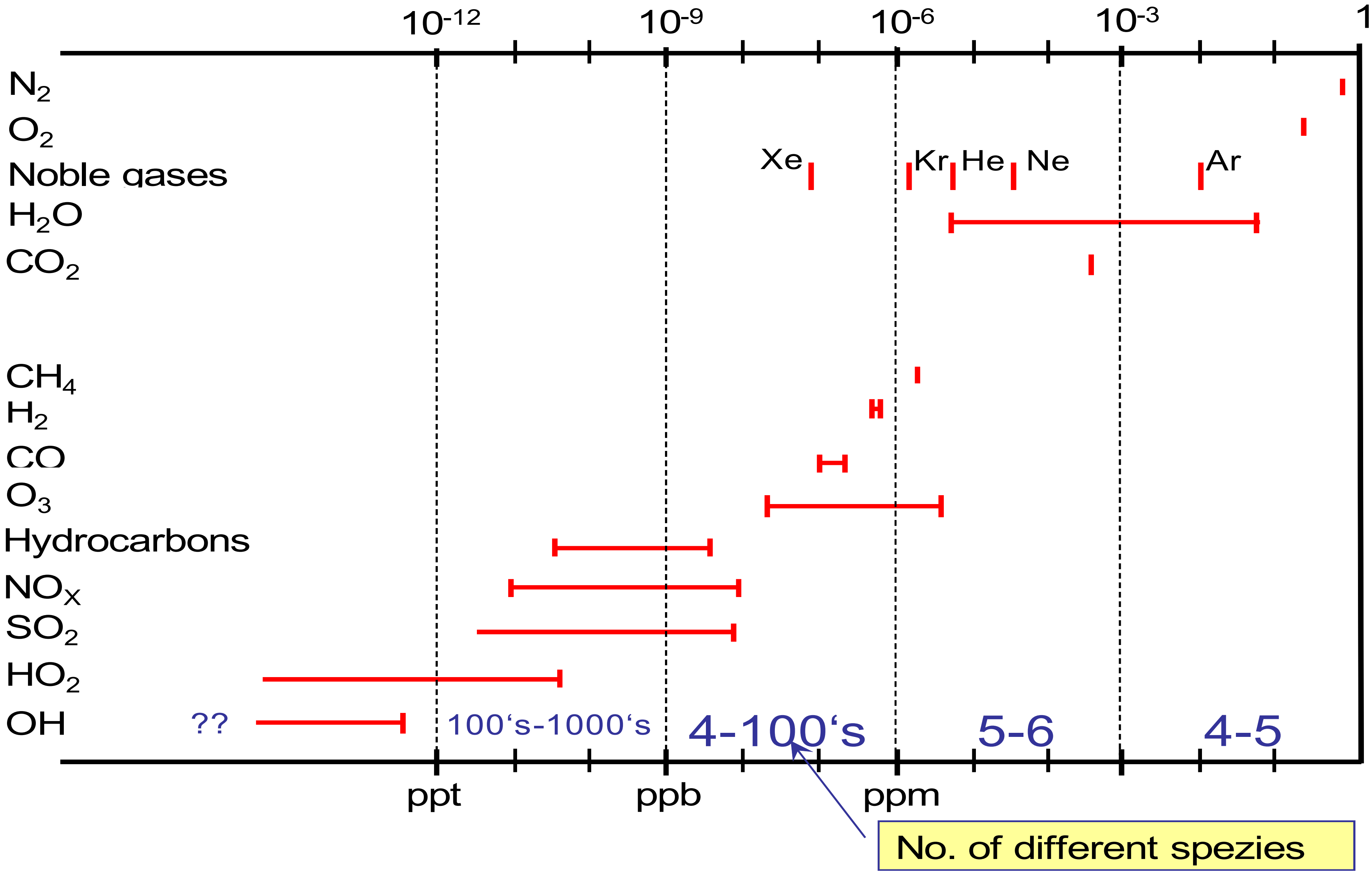
Oxygen O_2: $M_{O2} = 31{,}999$ kg kmol^{-1}

Argon Ar: $M_{Ar} = 39{,}942$ kg kmol^{-1}

Carbon dioxide CO_2: $M_{CO2} = 44{,}008$ kg kmol^{-1}

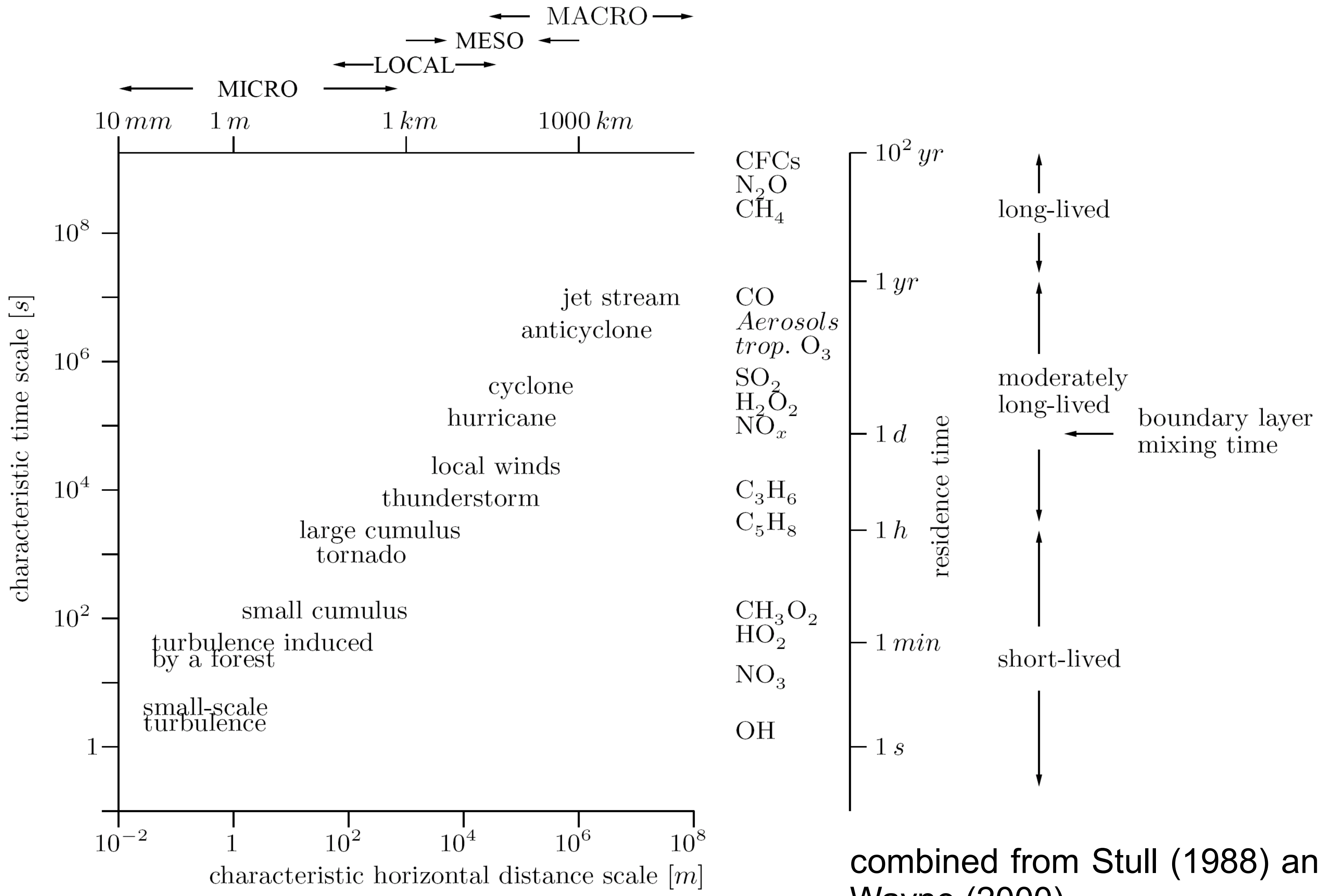


Mixing Ratios of Atmospheric Gases



→ The Atmosphere and its composition changed since formation of earth
See lecture on Thursday from Dr. Jens Fohlmeister and Mario Ruckelshausen

Scales of Turbulent Atmospheric Processes and Lifetime



combined from Stull (1988) and Wayne (2000).

Influence of Gases on the Atmosphere

Gas	Smog	Acid Rain	Turbidity of the atmosphere	Greenhouse Effect	Strat. Ozone Degradation	Influence on Self Cleaning of the Atmosphere
CO_2				+	+/-	
CH_4				+	+/-	+/-
CO	+					-
N_2O				+	+/-	
NO_X	+	+	+		+/-	+/-
SO_2	+	+	+	-		
VOC's	+		+			+
CFC's				+	+	
O_3	+	+		+		+

\+ Gas enhances the effect.

\- Gas reduces the effect.

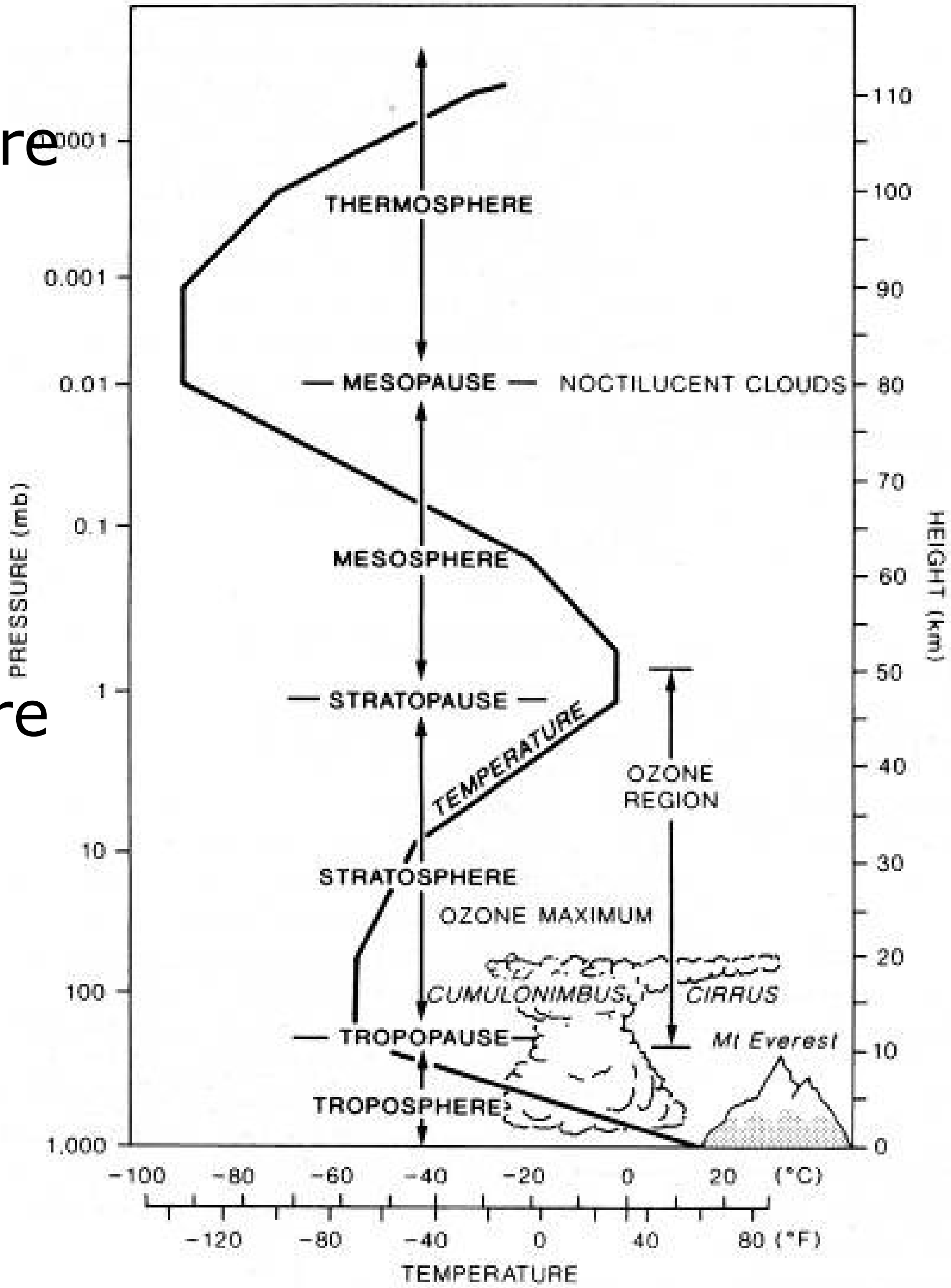
+/- Depending on conditions the influence can be positive or negative.
E.g. NO_X (= NO + NO_2) can both, enhance or reduce the stratospheric O_3 destruction.

Layers of the Atmosphere

Heterosphere

Homosphere

1% of Earth's Radius
99.98% of Atmosphere



Ionosphere

'hot' sphere

Neutrosphere

'intermediate' sphere

'layered' sphere

'weather' sphere

Barry+Chornley, fig 1.15

Vertical Pressure Distribution

The Atmospheric pressure p_0 at the earth's surface is given by:

$$p_0 = \frac{M_A \cdot g}{4\pi \ R_E^2}$$

where M_A denotes the total mass of the atmosphere, g the acceleration due to gravity of earth, and R_E the earth radius.

The atmospheric pressure decreases with height, in particular by changing the altitude by dz, the force K on an area F will change by the amount:

$$dF = - g \cdot \rho \cdot A \cdot dz$$

where ρ denotes the air density. The resulting pressure change dp is given by:

$$dp = dF/A = - g \cdot \rho \cdot dz$$

Expressing the air density ρ by molar-mass(M)/molar-volume(V) and substituting V = RT/p (R = gas constant, T = temperature) for one mole leads to:

$$\rho = \frac{M}{V} = \frac{Mp}{RT}$$

Substituting ρ in the expression for dp: $$dp = -\frac{Mg}{RT} pdz$$

After division by p and integration we obtain:

$$\ln p = \ln p_0 - \int_0^z \frac{Mg}{RT} dz$$

And for the pressure p(z) at the altitude z:

$$p(z) = p_0 \cdot \exp\left(- \int_0^z \frac{Mg}{RT} dz \right)$$

For an isothermal atmosphere, i.e. T (and g) being independent of z the above expression can be further simplified (**Altitude Pressure Relationship)** with z_s the Atmospheric Scale hight:

$$p(z) = p_0 \cdot \exp\left(- \frac{Mgz}{RT} \right) = p_0 \cdot \exp\left(- \frac{z}{z_s} \right)$$

$$z_s = - \frac{RT}{Mg} \propto \frac{T}{M}$$

For T = 273 K we obtain z_s = 7974 m, for more realistic T = 250 K we can state z_s = 7+/-1km best describes the atmospheric condition.

Atmospheric Scale Hight for some Gases

Gas	**Molecular Weight g/Mol**	**Scale Height z_s Km**	**p(20Km)/p(0) (hypothetical!)**
N_2	28	8.270	*0.089*
Air	29	7.984	
O_2	32	7.233	*0.063*
CO_2	44	5.260	*0.022*
$CFCl_3$	137.4	1.685	$7 \cdot 10^{-6}$

Adiabatic Processes and Potential Temperature

Scales are so large, that changes of state (e.g. lifting of an airmass) are adiabatic processes in good approximation:

$$T_1 = T_2 \times \frac{p_1^\gamma}{p_2^\gamma} = T_2 \times \left(\frac{p_1}{p_2}\right)^\gamma \Rightarrow T(z) = T_0 \times \left(\frac{p(z)}{p_0}\right)^\gamma$$

Air : Poisson's Equation of state :

$$\frac{T^\kappa}{p^{\kappa-1}} = \frac{T_0^\kappa}{p_0^{\kappa-1}} = \text{const.}$$

With the surface pressure $p_0 = 1013$ mBar.

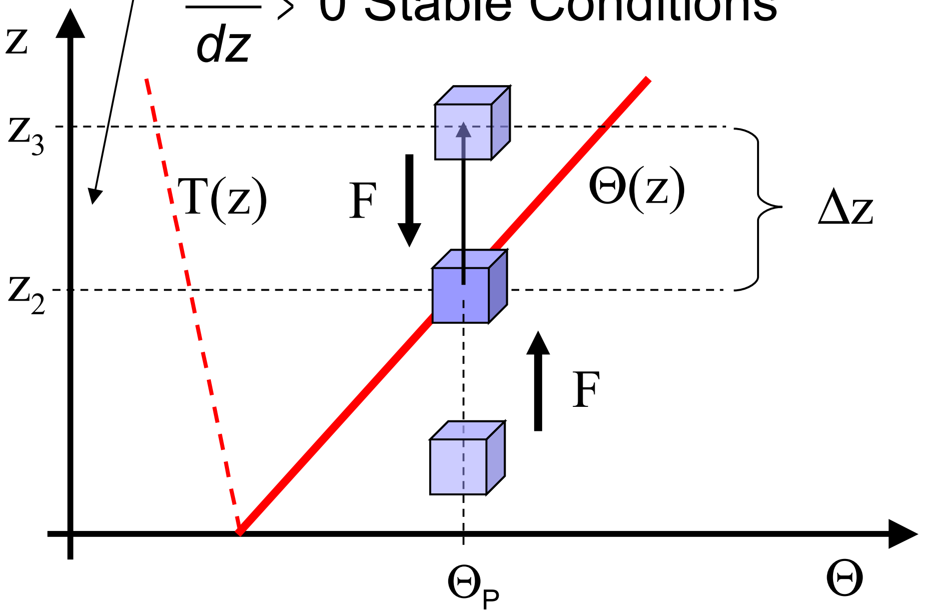
we obtain the potential Temperature θ :

$$\theta = T\left(\frac{p_0}{p}\right)^{\frac{\kappa-1}{\kappa}}$$

with $(\kappa-1)/\kappa \approx 0.286$ for air

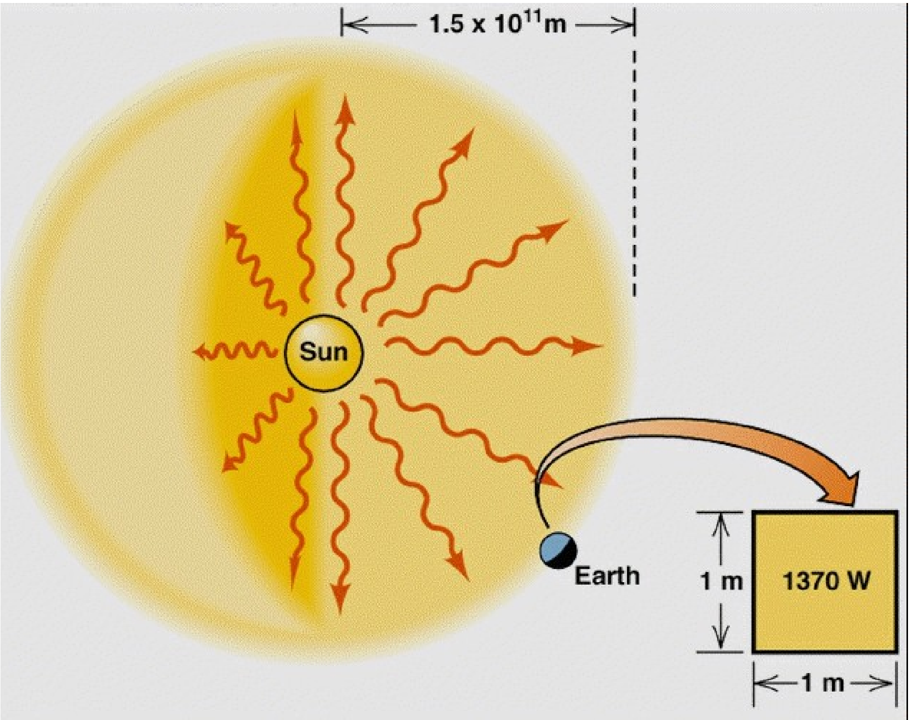
Displacement of a fluid parcel upwards (from z_2 to z_3):
Θ_P=const.
→ $\Theta_P < \Theta(z_3)$
→ Density of the fluid parcel larger than density of the surrounding fluid
→ restoring force F

$\frac{d\Theta}{dz} > 0$ Stable Conditions



II. Atmospheric Radiation - Basics

- **Sun is energy source**; radiative exchange of energy between sun-earth-space is key for energy budget
- Solar constant = Incident solar flux (through an area perpendicular to the direction of the sun): $S_C = 1368\ \mathrm{W\ m^{-2}}$
- Average incoming radiation: $S_0 = 342\ \mathrm{W\ m^{-2}}$ ($= S_C / 4$)
- Radiation is basis for most atmospheric processes including:
  - energy budget (heating, cooling)
  - photolysis → chemistry
  - large scale circulation
- Remote sensing: use of radiative measurements to probe the atmosphere



Basic Radiation Laws

Radiation Flux: $\Phi = \frac{\text{radiated energy}}{\text{time interval}} = \frac{dW}{dt}$

Irradiance: $B = \frac{\Phi}{A_e} \quad \frac{W}{m^2}$

A_e = Receiving area

Radiant intensity: $I = \frac{\Phi}{\Omega} \quad \frac{W}{sr}$

Ω = Solid angle

Radiance: $F = \frac{\Phi}{\Omega \cdot A_s} \quad \frac{W}{m^2 sr}$

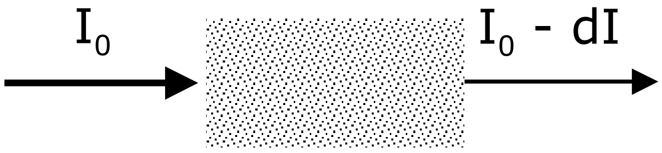
A_s = radiating area

(all areas assumed to be perpendicular to the direction of radiation flux)

Absorption

- Beer-Lambert's law:

$$dI_\lambda = -\sigma_{abs,\lambda} c I_\lambda ds$$



  - $\sigma_{abs,\lambda}$ - absorption cross section
  - fractional energy absorbed from a "pencil" of radiation
  - absorption coefficient

- Transmissivity

$$k_{abs,\lambda} = \sigma_{abs,\lambda} c$$

$$k_{abs,\lambda} = \sum_i \sigma_{abs,\lambda i} c_i$$

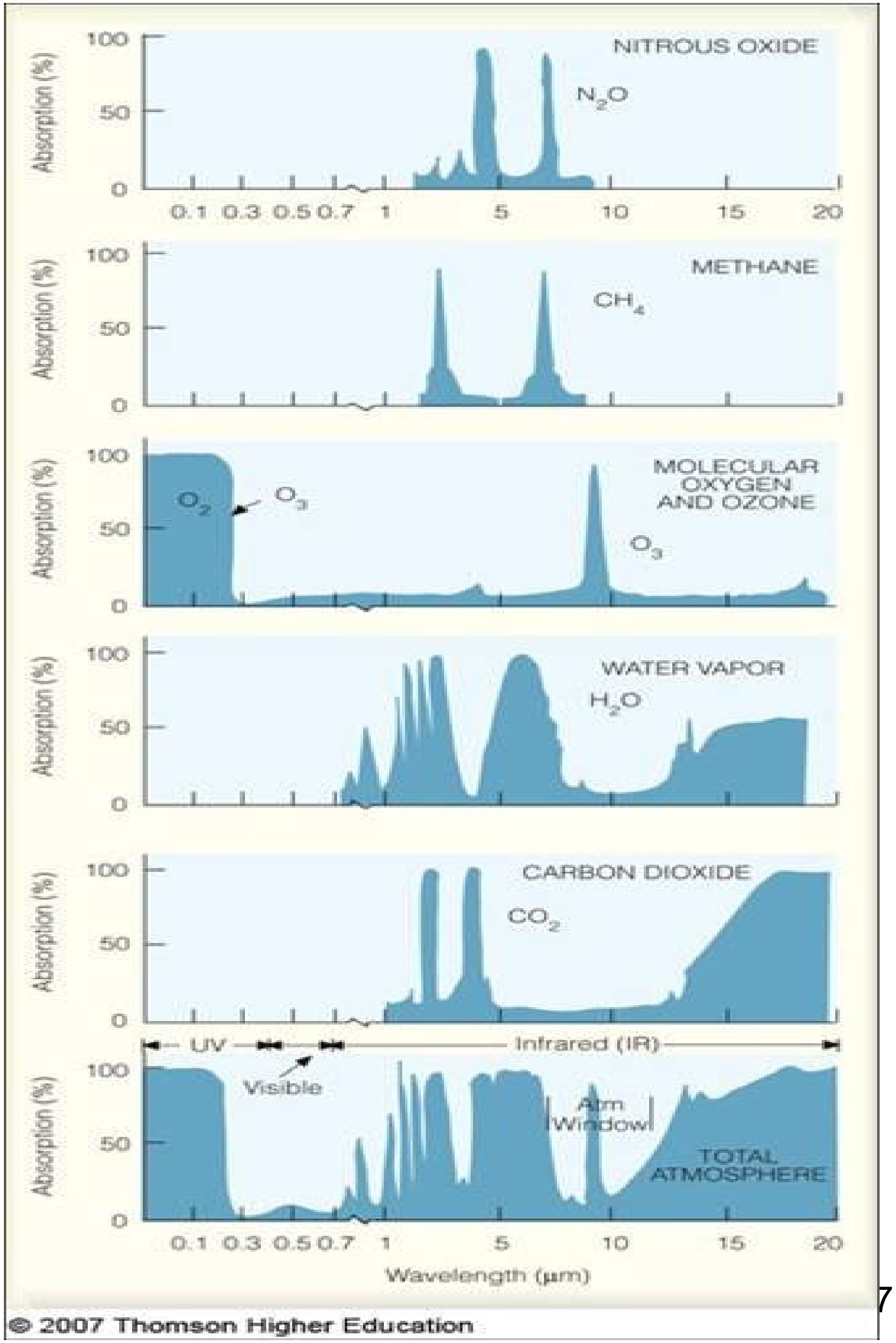
Discovered by:
Pierre Bouguer in 1729,
Johann Heinrich Lambert in 1760
and August Beer in 1852

$$T_\lambda(s) = \frac{I_\lambda(s)}{I_\lambda(0)} = e^{-\int_0^s \rho \sigma_{abs,\lambda} ds} = e^{-\tau_\lambda(s)}$$

$\tau_\lambda(s)$ – optical depth

Absorption of Gases in the Atmosphere

- absorption spectrum of gas consists of
  - continuum absorption (X-ray, short UV)
    - photoionization
    - photodissociation
  - complex arrays of lines corresponding to energy levels of:
    - discrete electronic (UV)
    - vibrational (IR)
    - rotational (IR and microwave)



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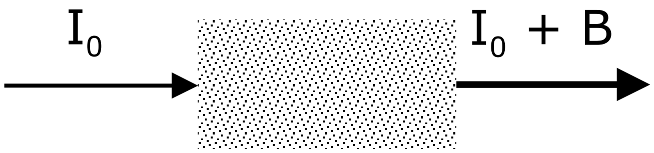
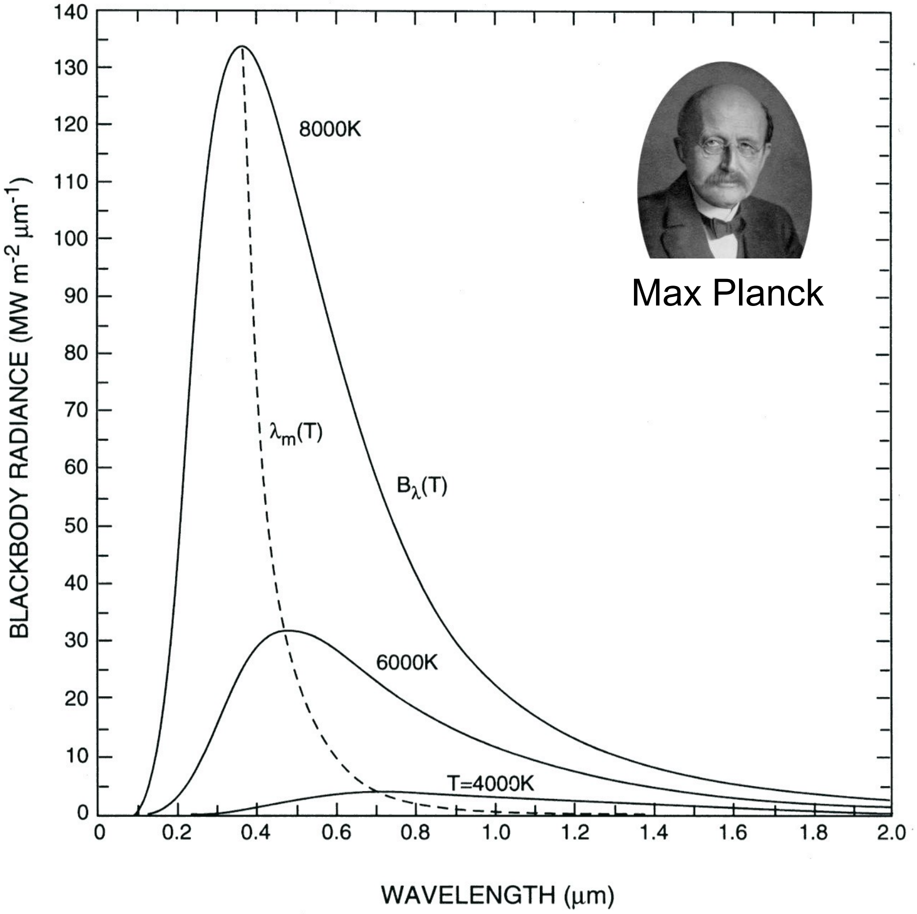
Emission

- Planck's law:
  - black body radiation
  - isotropic
  - radiance:

$$I(\lambda, T) = B_\lambda(T) = \frac{2hc^2}{\lambda^5\left(e^{\frac{hc}{\lambda kT}} - 1\right)}$$

  - unit: W/(m^2 sr)





$\lambda < \lambda_m$ → rapid increase in intensity with λ

$\lambda > \lambda_m$ → slow decrease "

Salby, 1996

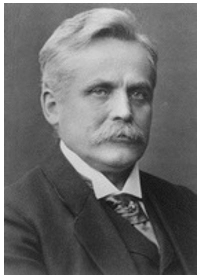
Wien's displacement Law

The maximum of the Planck Function $F_p(\lambda)$ is obtained from:

$$\frac{dI(\lambda, T)}{d\lambda} = 0$$

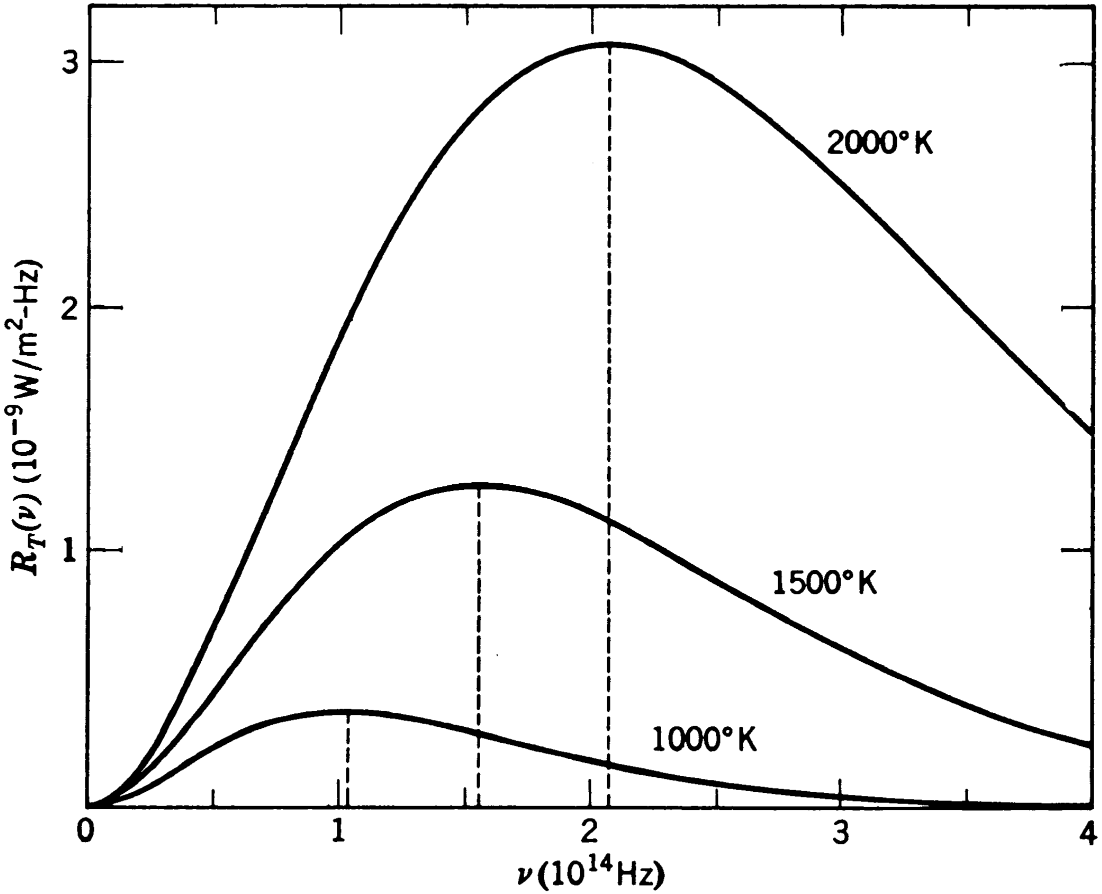
→ $\lambda_{max} = A/T$ with $A = 2898\ \mu m \cdot K$

or $\nu_{max} = c/A \cdot T \propto T$



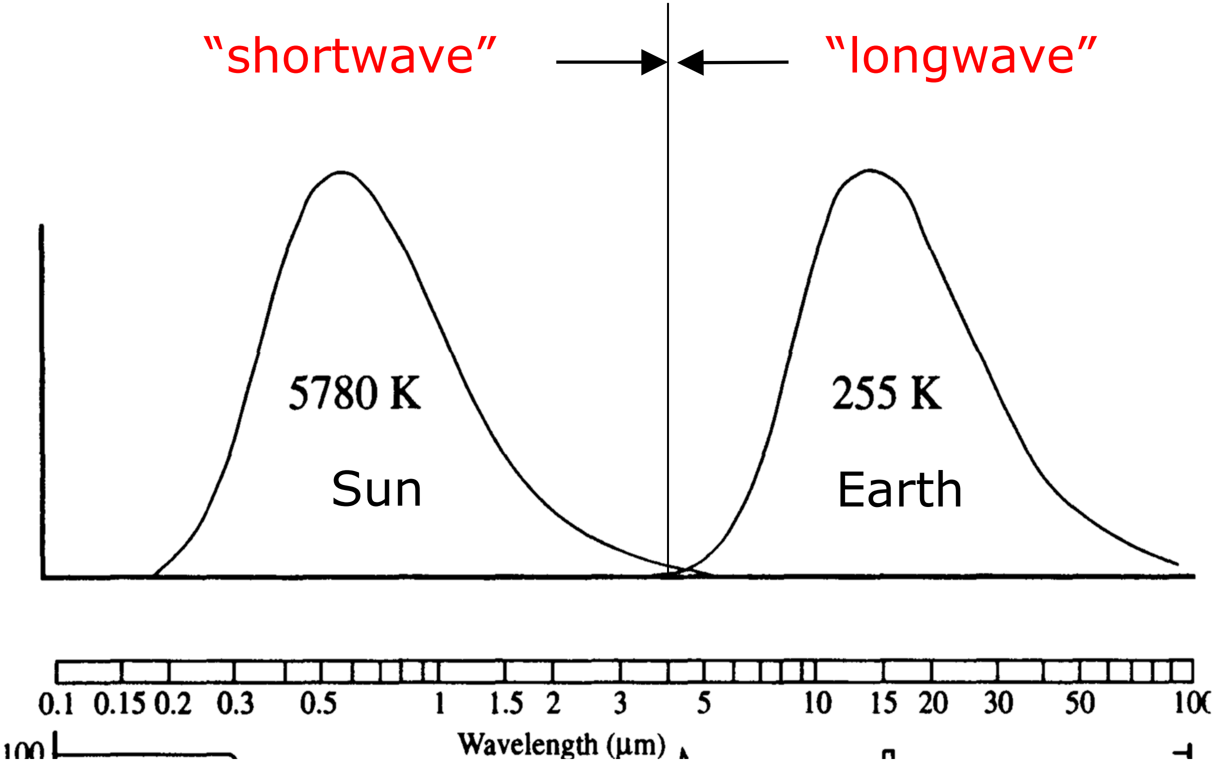
Wilhelm Wien

Planck function:
The radiance $F(\nu)d\nu$ of a black body as a function of frequency.
The frequency, where the maximum occurs (dashed vertical lines) increases linearly with temperature.



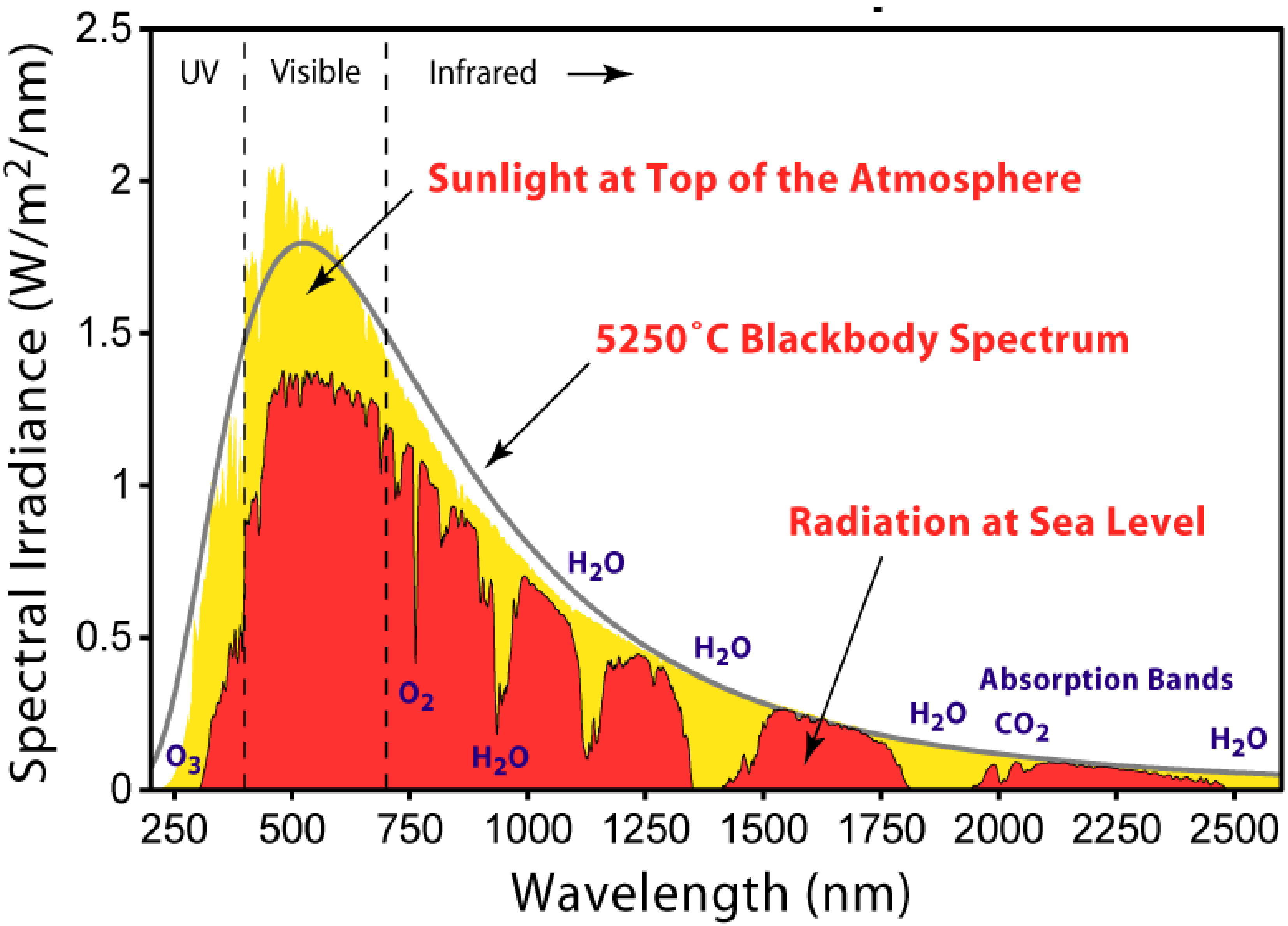
Emission Distribution

- Wien's displacement law:
  - allows to determine brightness temperature of a body from emitted radiation

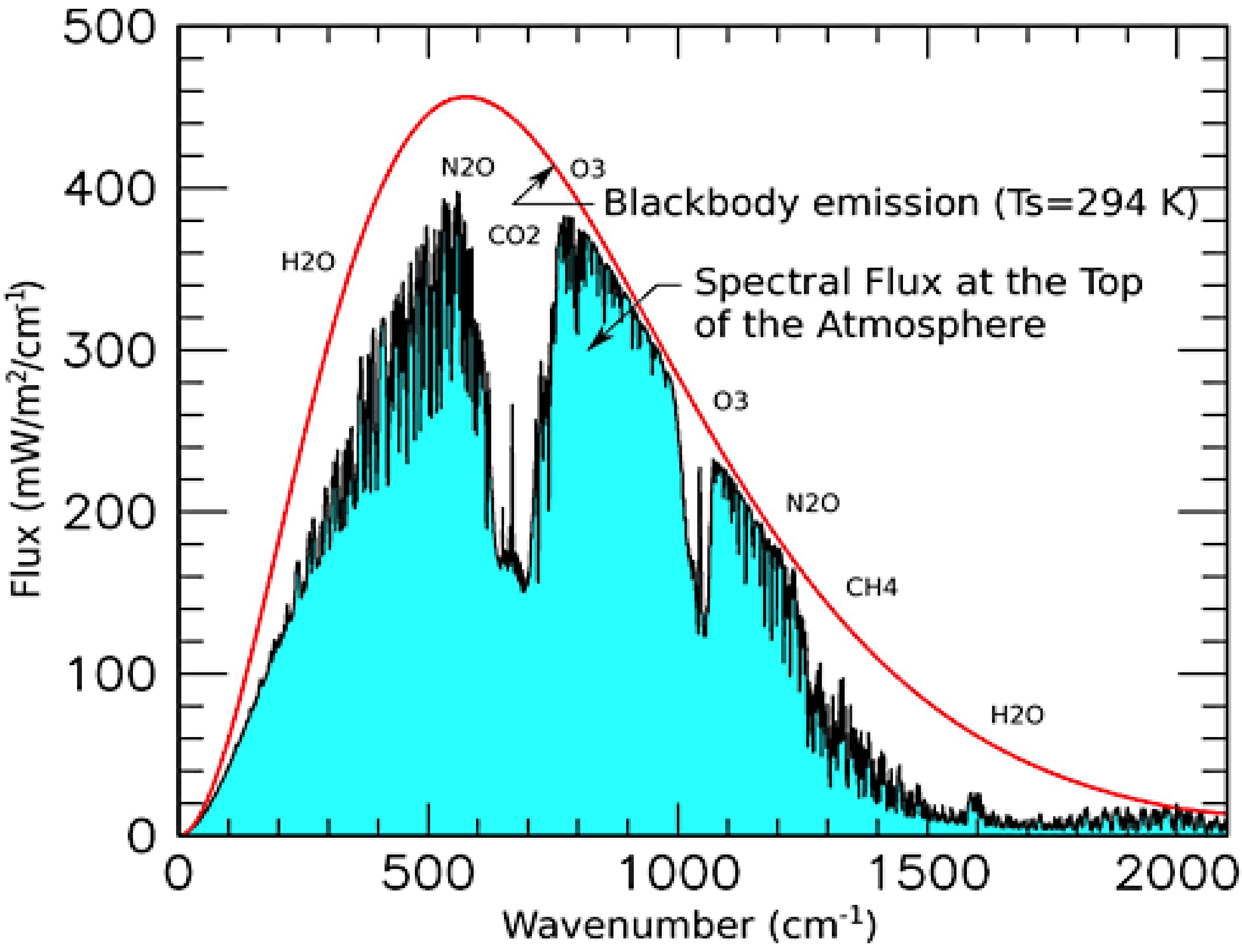


Salby, 1996

Solar Radiation Spectrum



Earth Radiation Spectrum



http://www.giss.nasa.gov/research/briefs/schmidt_05/

CO_2 Experiment

→ Start CO_2 demonstration experiment

Emission – Total Flux

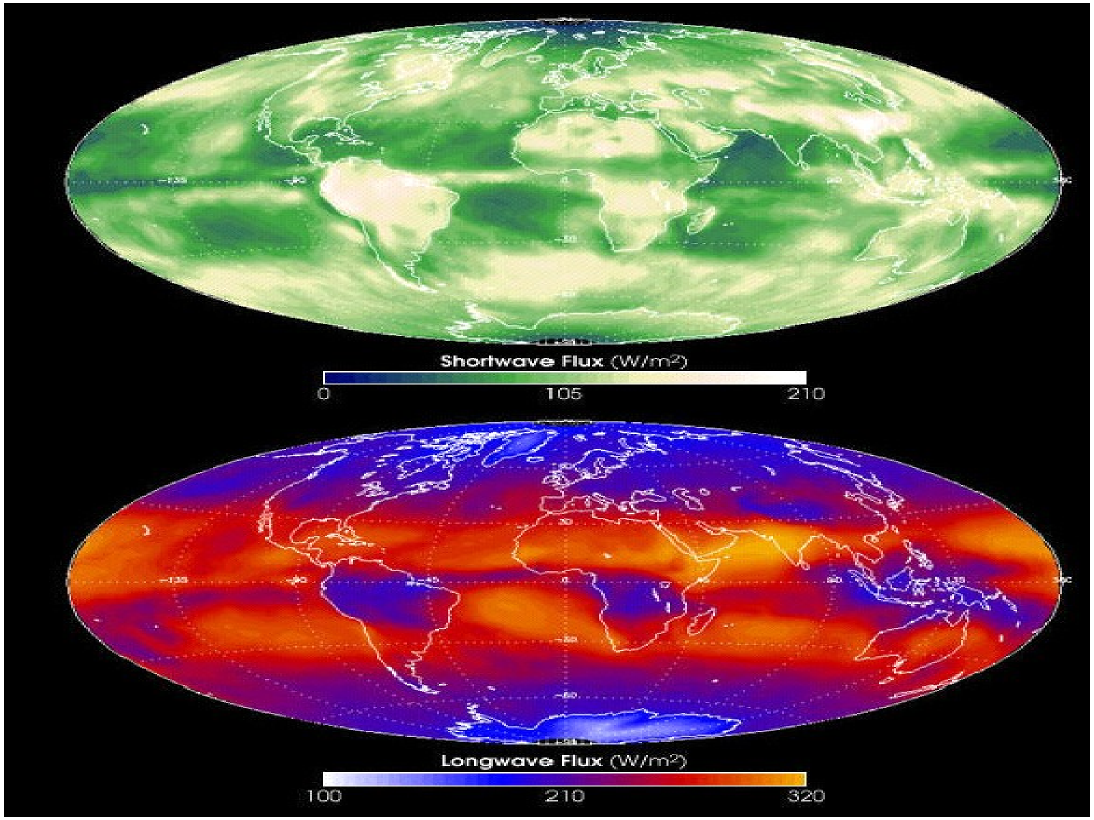
- Stefan-Boltzmann law:
  - total flux emitted by a black body:

$$F = \pi \int_0^\infty I_\lambda \, d\lambda = \sigma T^4$$

  - $\sigma = 5.67 \times 10^{-8}$ W m^{-2} K^{-4}: Stefan-Boltzmann constant
  - T-dependence:

emission from cold and warm objects differ sharply

Joseph Stefan & Ludwig Boltzmann



Grey Bodies

- real substances (=greybodys) are no perfect absorbers:
- absorptivity:

$$\alpha = \frac{\text{power absorbed from the actual radiating body}}{\text{power incident on the body}}$$

- reflectivity:

$$\rho = \frac{\text{power reflected by the actual body}}{\text{power incident on the body}}$$

- transmissivity:

$$\tau = \frac{\text{power transmitted through the actual body}}{\text{power incident on the body}}$$

- energy conservation: $\alpha + \rho + \tau = 1$
- Kirchhoff's law: $\varepsilon_\lambda = \alpha_\lambda$

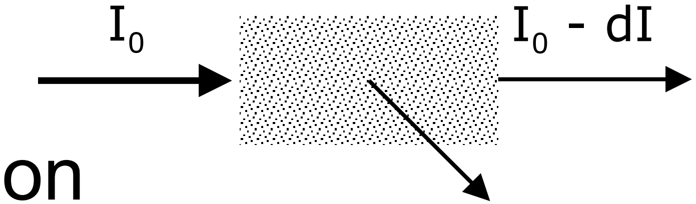
At any wavelength the emissivity of a body (or gas!) is exactly equal to its absorptivity

Gustav Robert
Kirchhoff
1824 - 1887

in Heidelberg:
1854 - 1874

Scattering

$$dI_\lambda = -\sigma_{scat,\lambda} c I_\lambda ds$$



- $\sigma_{scat,\lambda}$ - scattering cross section
- phase function p: fraction of radiation scattered from incident direction **Ω**' in direction **Ω**

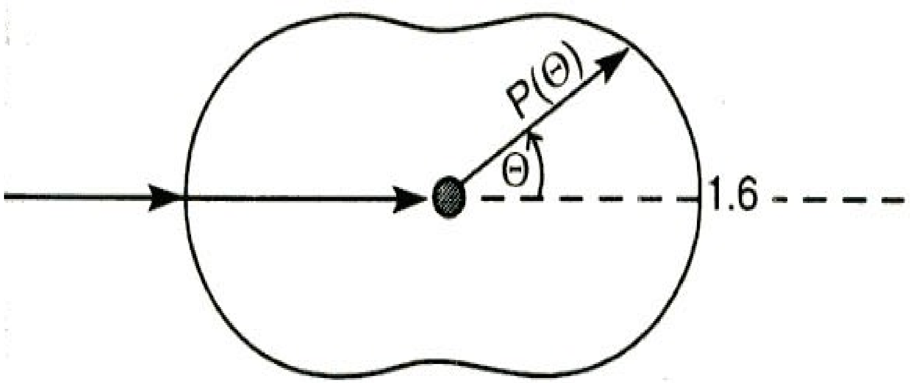
$$\frac{1}{4\pi}\int_0^\infty \int_{4\pi} p(\lambda' \to \lambda, \vec{\Omega}' \to \vec{\Omega})\, d\Omega\, d\lambda = 1$$

- most photons get scattered elastically by atoms and molecules – scattered photons have same wavelengths as incident photons:

$$\frac{1}{4\pi}\int_{4\pi} p(\lambda, \vec{\Omega}' \to \vec{\Omega})\, d\Omega = 1$$

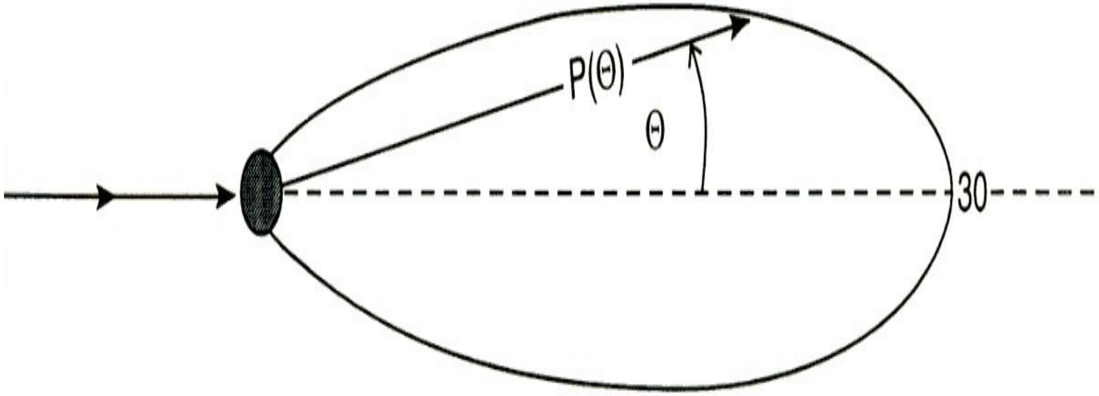
Rayleigh Scattering

- r << λ
- SW radiation
  (λ ≈ 100’s of nm) and
  gas molecules
  (r ≈ 0.1 nm)



Mie Scattering

- r ≥ λ
- SW radiation
  (λ≈100s nm) and aerosol
  particles/droplets (100nm < r < 50 μm)



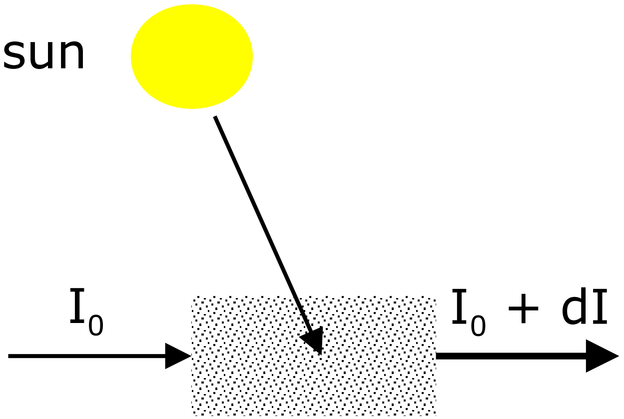
Salby, fig. 8.8

Extinction

- extinction := absorption + scattering
- extinction cross section: $\sigma_{ext,\lambda} = \sigma_{abs,\lambda} + \sigma_{scat,\lambda}$

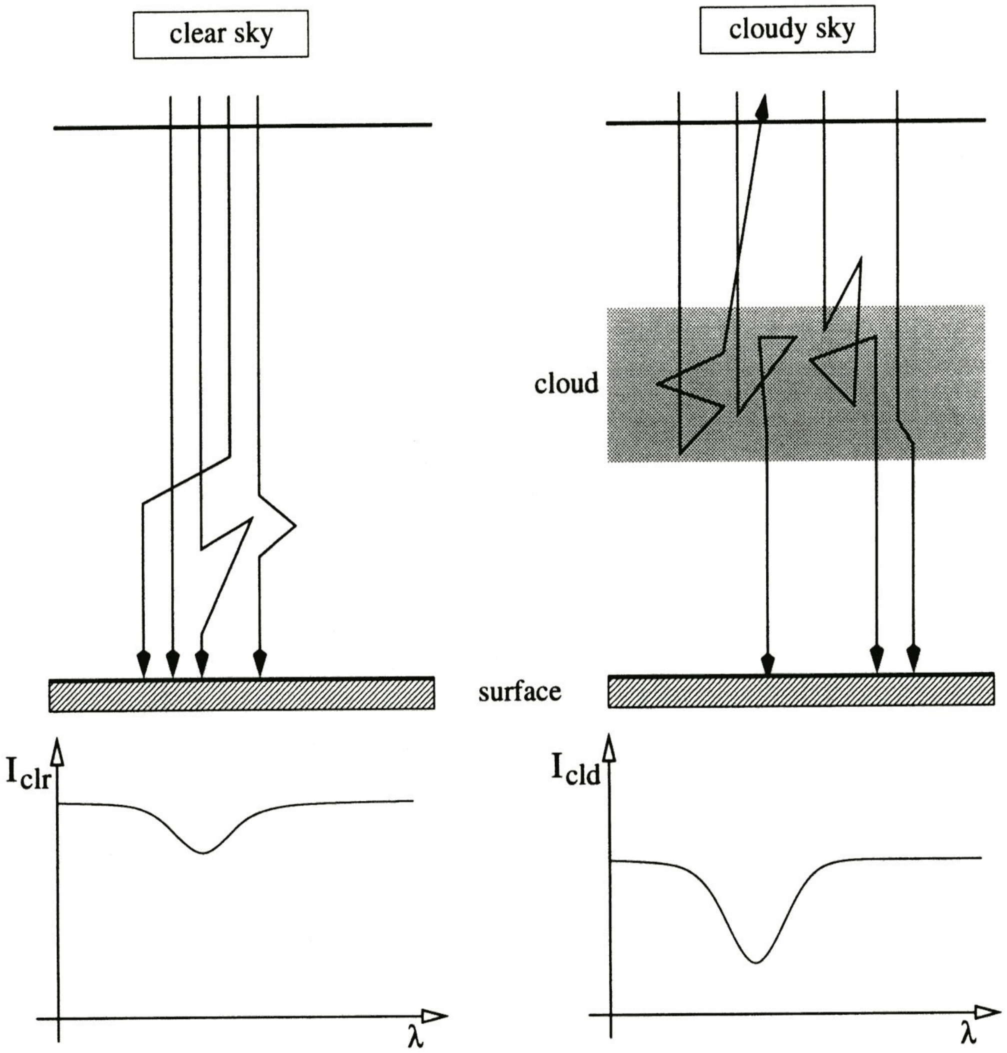
Scattering as Source of Radiation

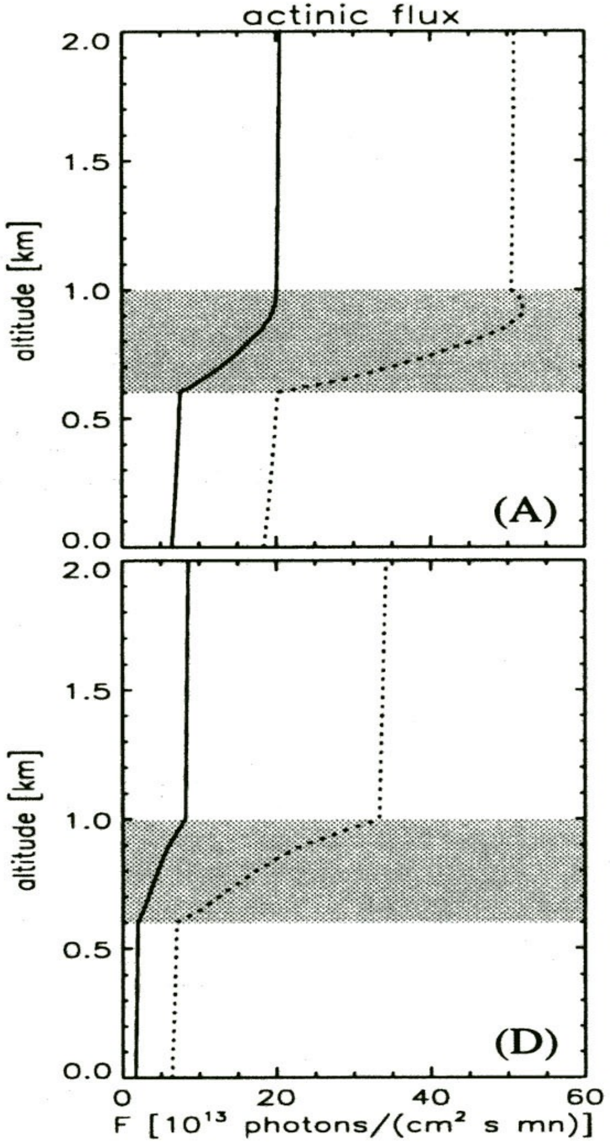
- also: increase of I due to direct scattering of solar radiation



$$\frac{dI_\lambda}{ds} = k_{scat,\lambda} \frac{1}{4\pi} I_{sun,\lambda}(\vec{\Omega}_0) e^{-\int_0^{s_0} k_{ext,\lambda}\, ds_0} p_\lambda(\vec{\Omega}_0 \to \vec{\Omega})$$

Impact of Clouds

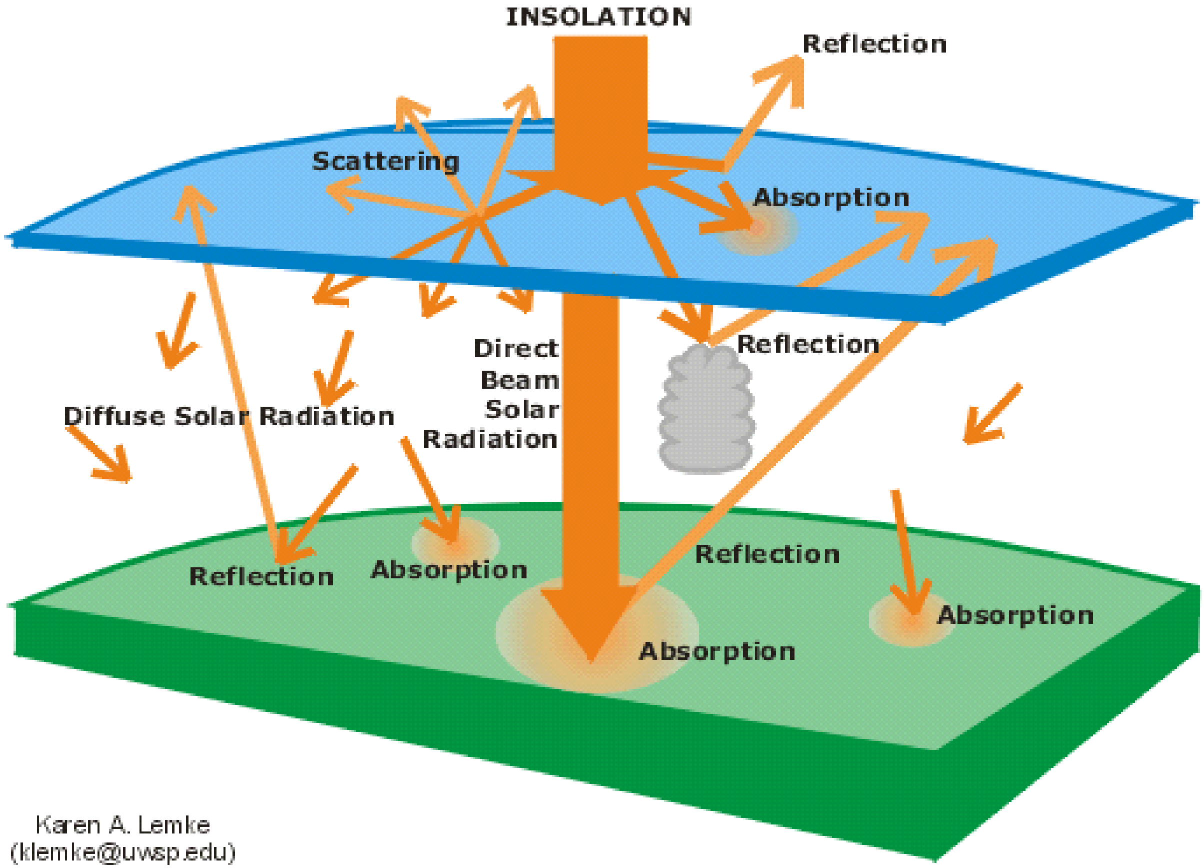




Landgraf, 1998

solid line: λ=320 nm
dotted line: λ=370 nm

Radiative Transfer Equation (RTE)



Karen A. Lemke
(klemke@uwsp.edu)

RTE is expression for change of radiation in a medium (e.g. atmosphere):

- scattering
- absorption
- emission
- incident radiation (sun)

$$\frac{dI_\lambda}{ds} = -(k_{abs,\lambda} + k_{scat,\lambda})I_\lambda$$

$$+ k_{abs,\lambda} B_\lambda(T)$$

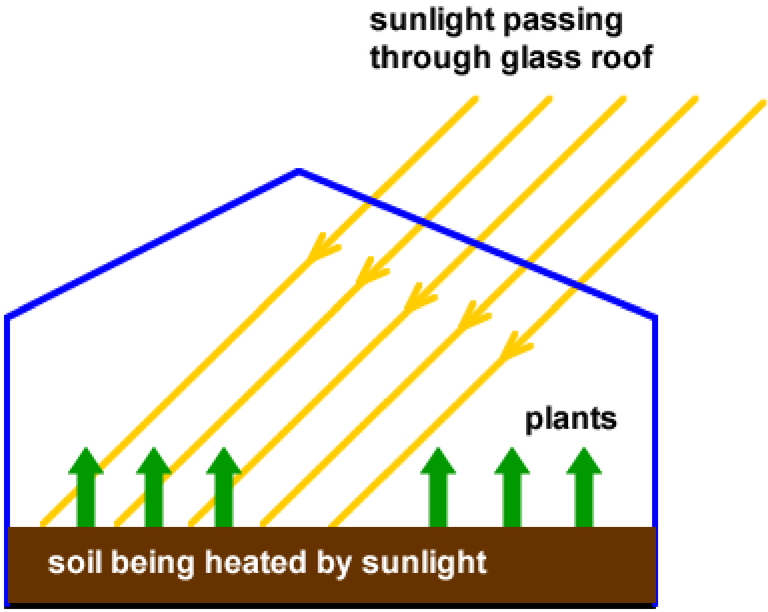
$$+ k_{scat,\lambda} \frac{1}{4\pi} I_{sol,\lambda} e^{-\int_0^{s_0} k_{ext,\lambda} ds_0} p_\lambda(\theta_0 \to \theta, \varphi_0 \to \varphi)$$

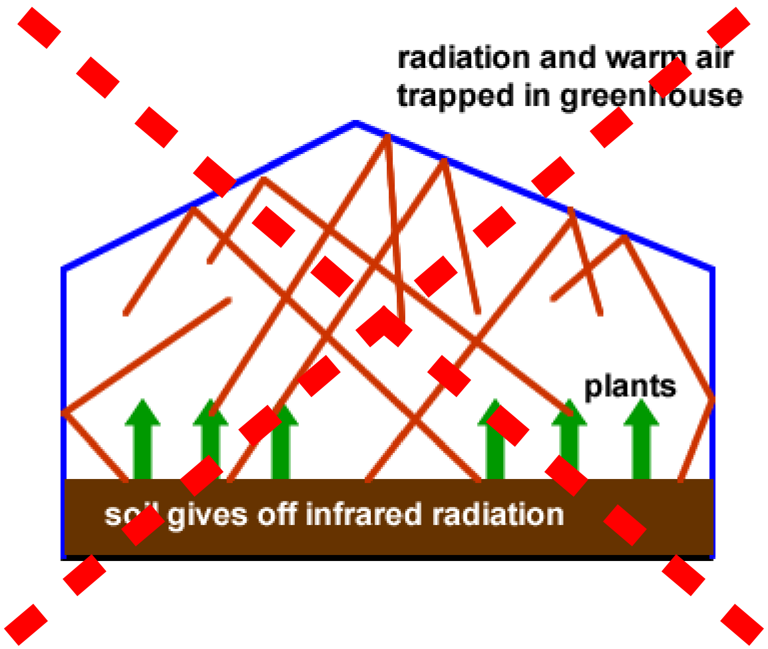
$$+ k_{scat,\lambda} \int_0^{\pi} \int_0^{2\pi} \frac{1}{4\pi} I'(\theta', \varphi') p_\lambda(\theta' \to \theta, \varphi' \to \varphi) \sin\theta' d\theta' d\varphi'$$

→ **can not be solved analytically**

→ **simplifications, and use of numerical models for idealised cases**

III. Warming in a Greenhouse





- Solar radiation is absorbed
- Ir radiation is absorbed by the glass

→ however this makes no significant effect as the surrounding has nearly the same temperature → no radiation gain

- suppression of convection result in heating!!! (glass is the barrier for the air flow)

→ This is not the atmospheric Greenhouse Effect!!!

IV. Atmospheric Radiation: Energy Budget

IV.a) 1st Approximation: No Atmosphere

(SW) Power received from the sun on earth:

$$P_{in} = \pi R^2 S_0 (1 - A)$$

$B_{SC} = S_0$ = 1368 W/m² = Solar constant

A ≈ 0.3 = Albedo of earth

1-A = short-wave absorptivity of earth

R_e = 6740 Km = Earth radius

(IR) power radiated from earth:

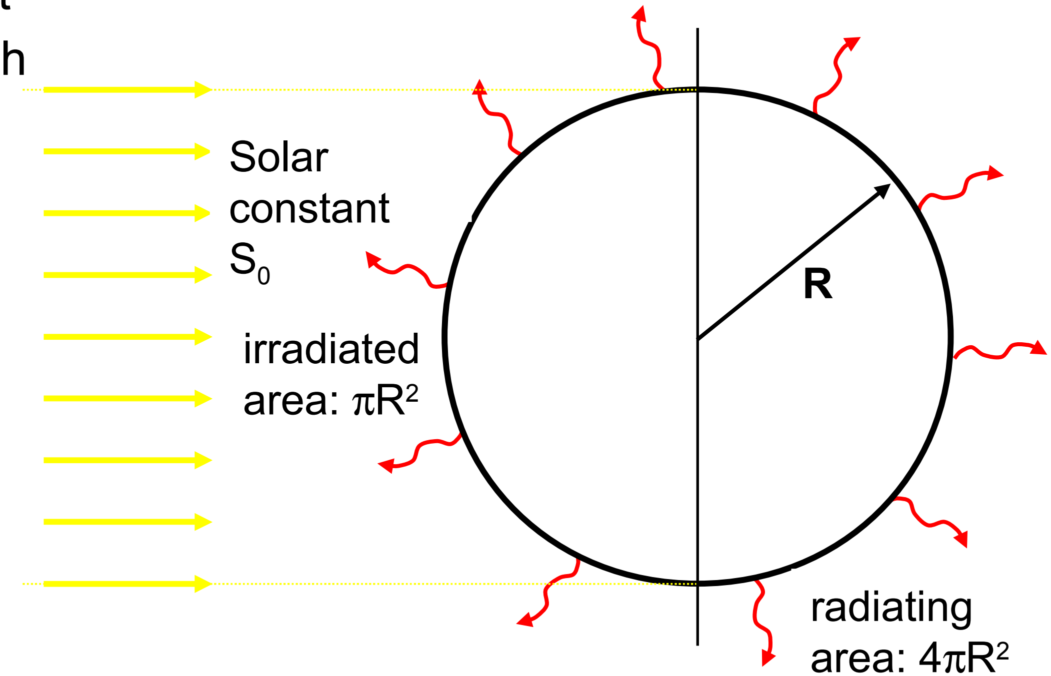
$$P_{out} = 4\pi R^2 \varepsilon \sigma_{SB} T_B^4$$

ε ≈ 0.9 ... 1.0

= IR Emissisivity of earth

σ_{SB} = $5.67 \cdot 10^{-8}$ Wm^{-2}K^{-4}

= Stefan-Boltzmann constant



Solar constant S_0

irradiated area: πR^2

radiating area: $4\pi R^2$

Surface Temperature 1st Approximation

Since earth is very close to thermal equilibrium and energy exchange can only take place via radiation we have in very good approximation:

$$P_{out} = P_{in} \quad *$$

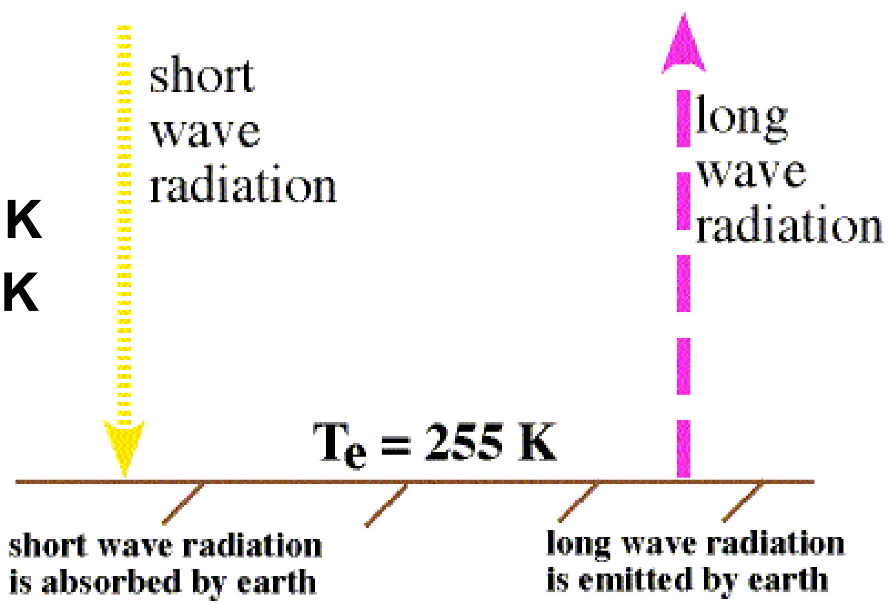
Substituting for P_{out} and P_{in} we obtain the average of surface temperature of earth T_0 in 1st approximation:

$$T_0 = \sqrt[4]{S_0 \times \frac{(1 - A)}{4 \varepsilon \sigma_{SB}}}$$

with the above numbers we obtain: **$T_0 \approx 255$ K**

Measured average temperature of earth: **$T_e \approx 288$ K**

Cause of the discrepancy:
The **natural greenhouse effect** of 33K.



No Atmosphere Case

*Note however: **"Earth is now absorbing 0.85±0.15 watts per square meter more energy from the Sun than it is emitting to space."** Hansen J., Nazarenko L., Ruedy R., Sato M., Willis J, Del Genio A., Koch D., Lacis A., Lo K., Menon S., Novakov T., Perlwitz J., Russell G., Schmidt G.A., Tausnev N. (2005), Earth's Energy Imbalance: Confirmation and Implications, SCIENCE 308, (3 JUNE), 1431-1435.

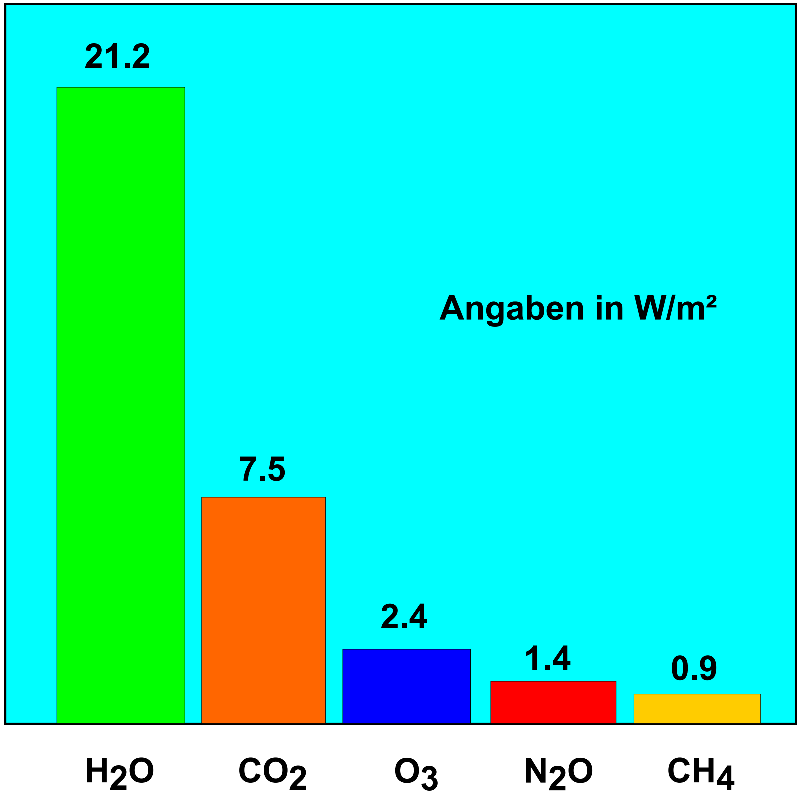
IV.b) 2nd Approximation: Basic Atmosphere

The 'Natural Greenhouse Effect'

In summary the “natural’ greenhouse effect amounts to about +33 K

Contribution of individual gases (after Kondratyev and Moskalenko, in J.T. Houghton (Ed.), IUP 957, 1984)

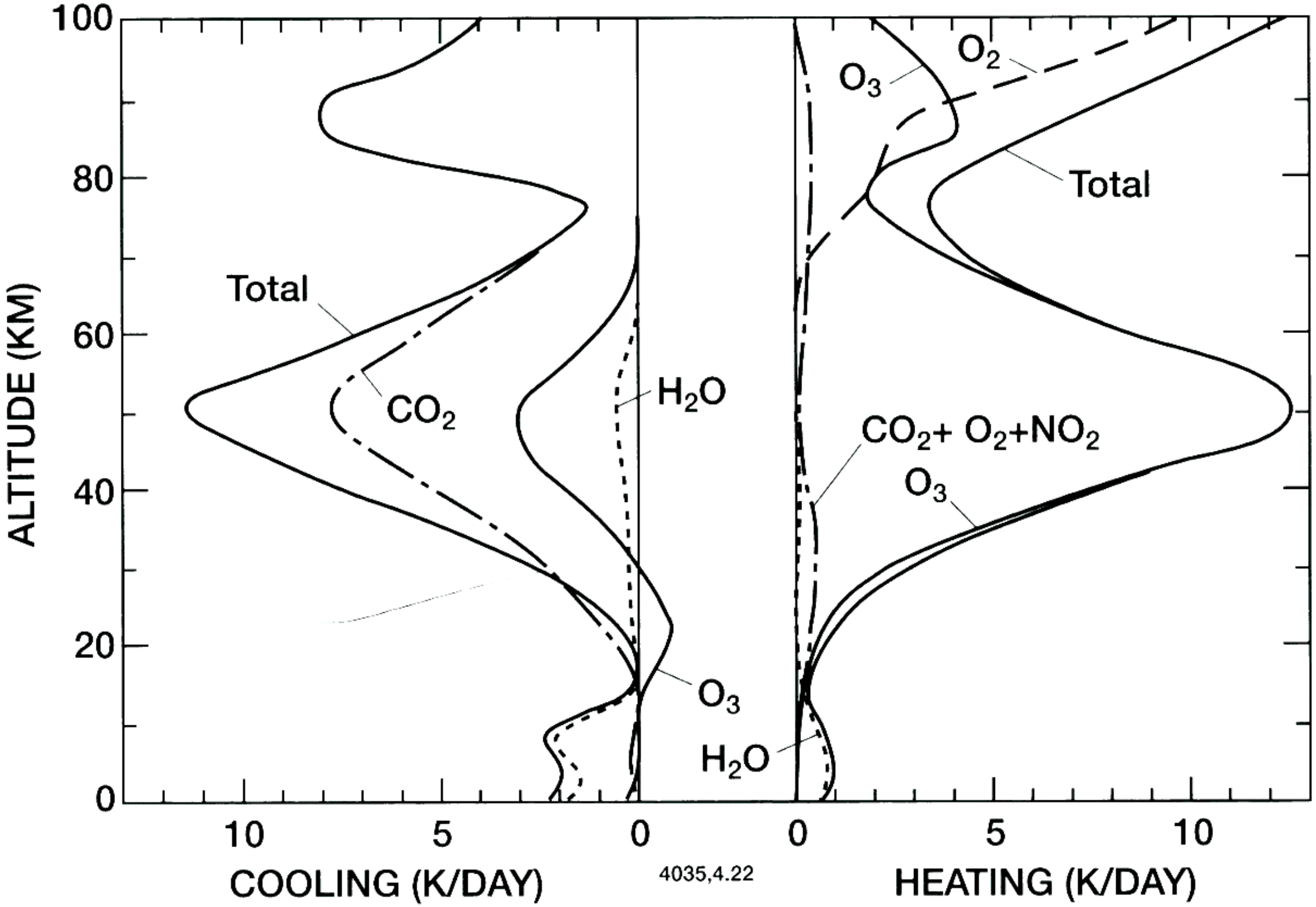
Gas	**Prominent Band μm**	**ΔT K**	**%**
H_2O	6.3, >16	20.6	62
CO_2	13 - 17	7.2	22
O_3 (in the troposphere)	9.6	2.4	7
N_2O	4.8, 7.8	1.4	4
CH_4	3.4, 7.3	0.8	2.5



Atmospheric Energy Balance

local heating rates:

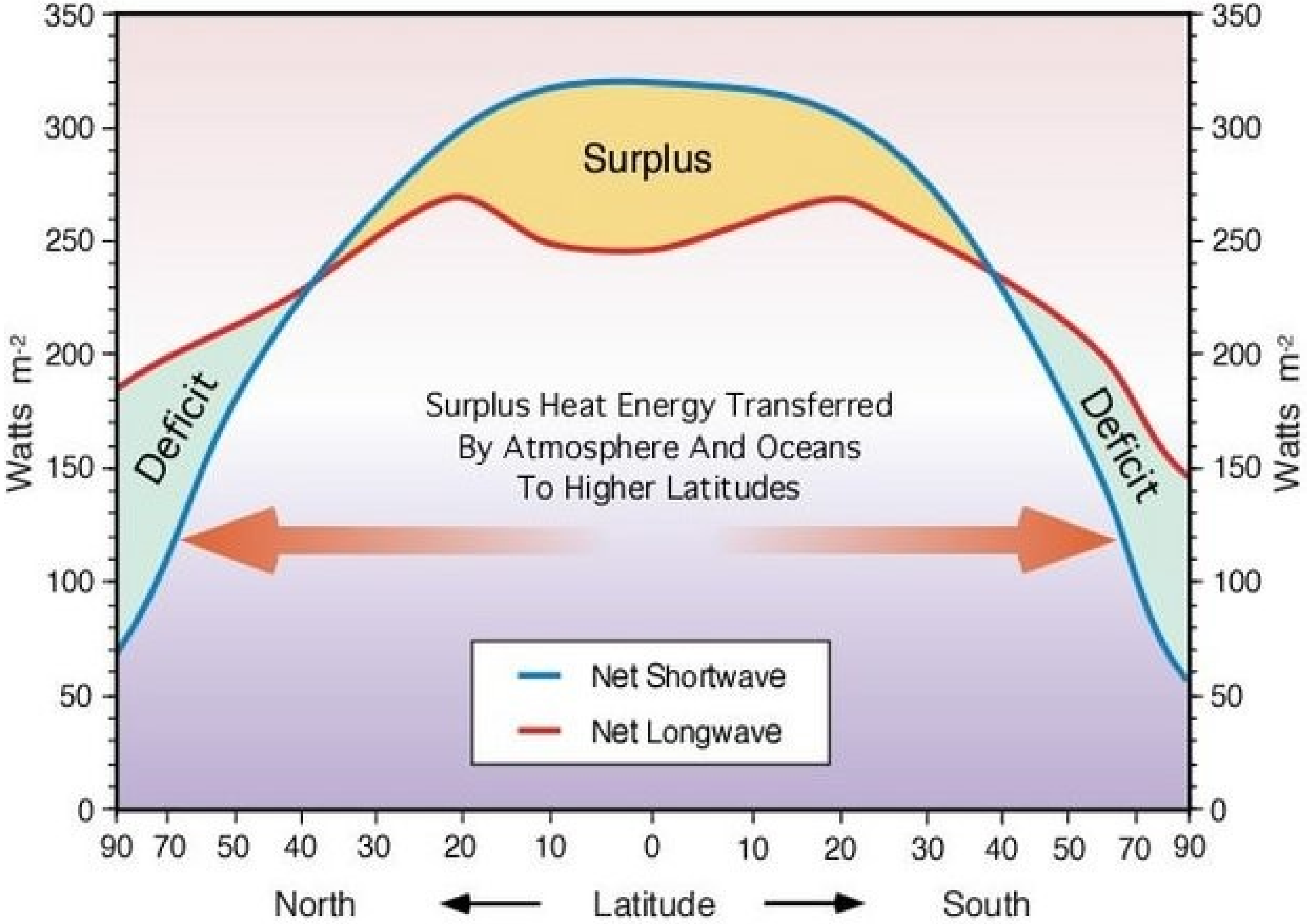
- stratosphere ~ radiative equilibrium
- troposphere ≠ radiative equilibrium - “convective adjustment”



Brasseur and Solomon, 2005 (IUP-Book 1968)

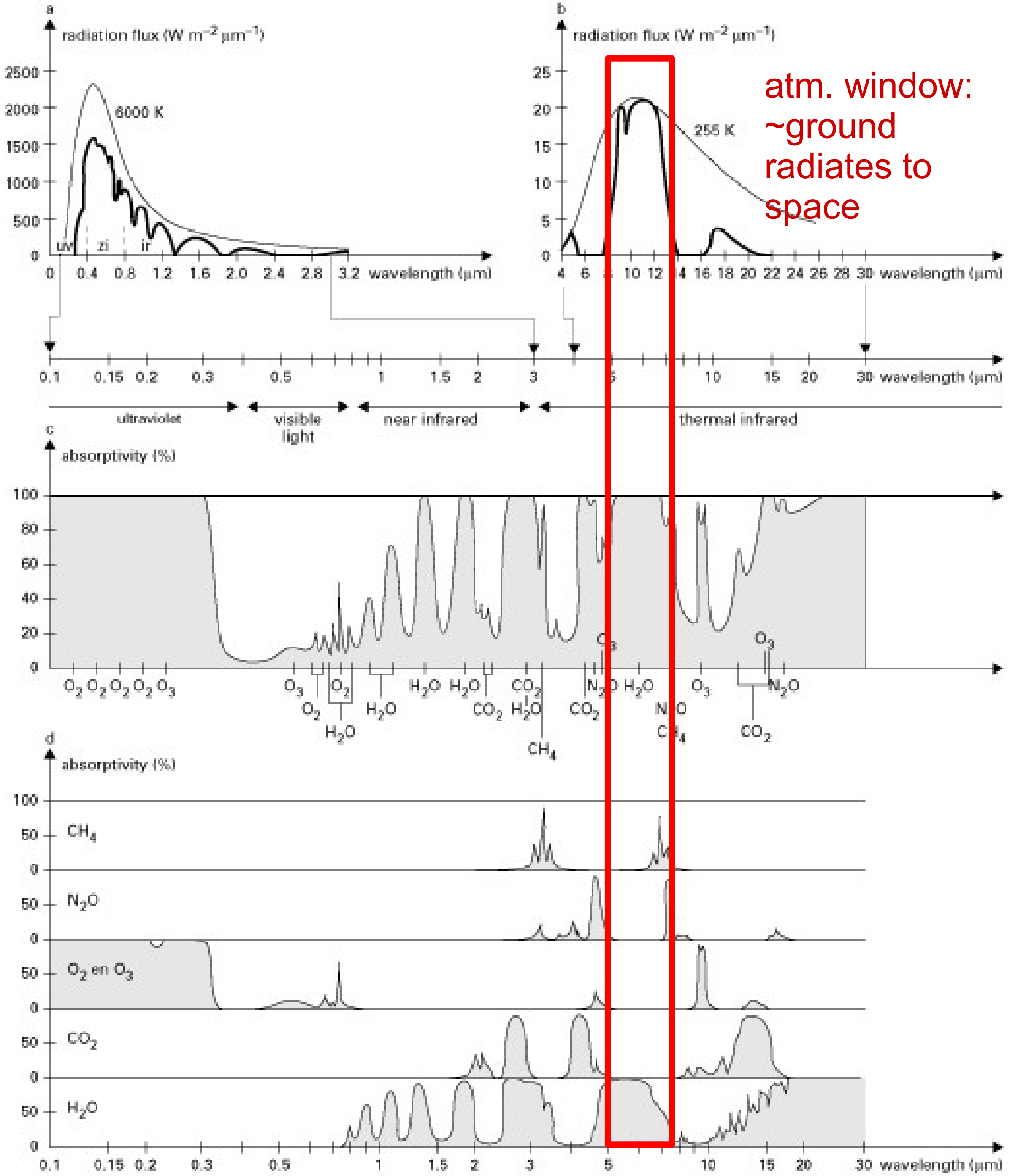
Higher atmosphere is (mainly) cooled by LW and heated by SW radiation

Latitudinal Energy Balance of Earth



http://www.eoearth.org/article/Energy_balance_of_Earth

Emission Spectra of Earth and Sun



atm. window:
~ground
radiates to
space

Stratosphere: UV-absorption by O_3

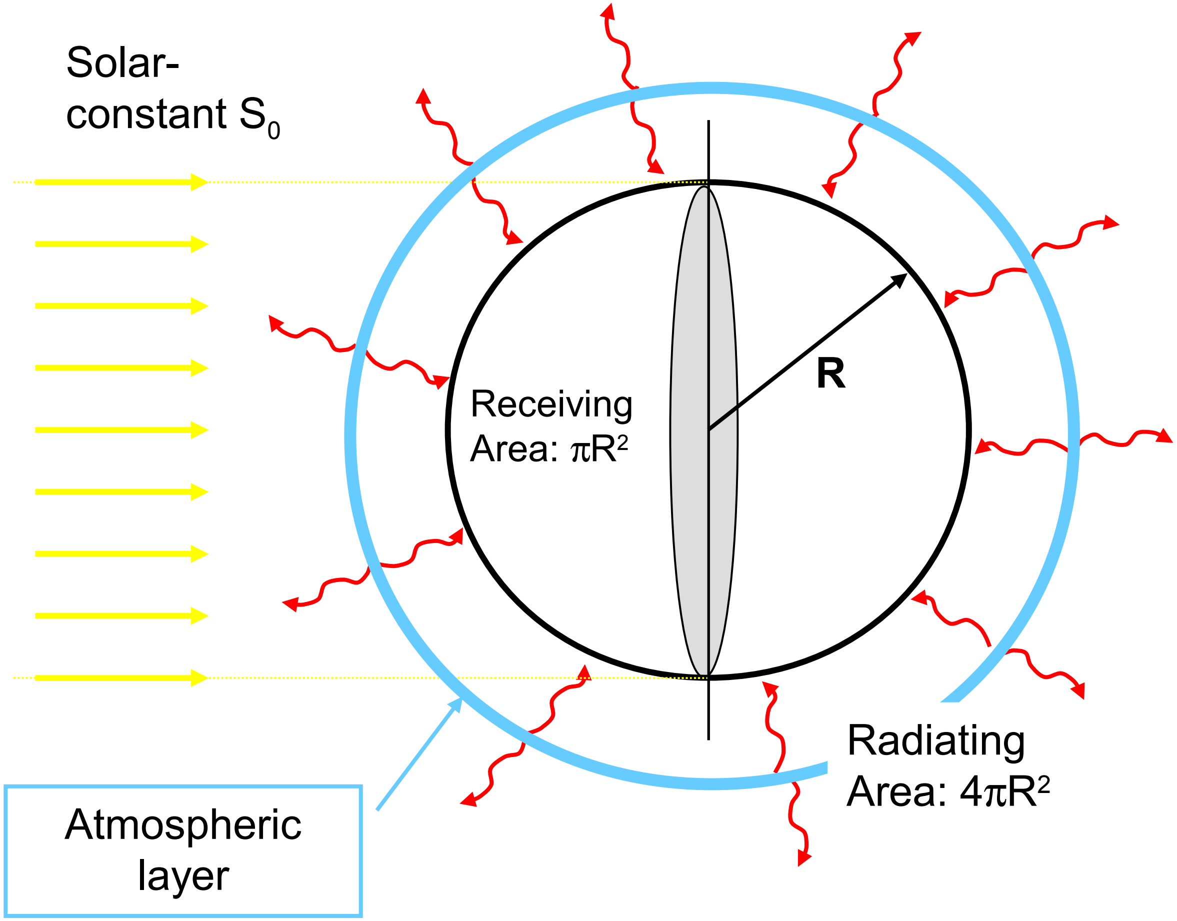
Troposhere:
IR-absorption by H_2O, CO_2, CH_4, etc.

Atmosphere reduces IR-radiation

Atmosphere absorbs and
reemits IR radiation

Calculation with Atmospheric Layer

Somewhat more realistic greenhouse-model: Atmosphere as thin, IR absorbing and emitting layer („glas roof").



Ground:

Temperature T_G,

IR-(LW) Absorptivity = Emissivity ε_G,

VIS-(SW)Albedo A = A_p

Atmosphere:

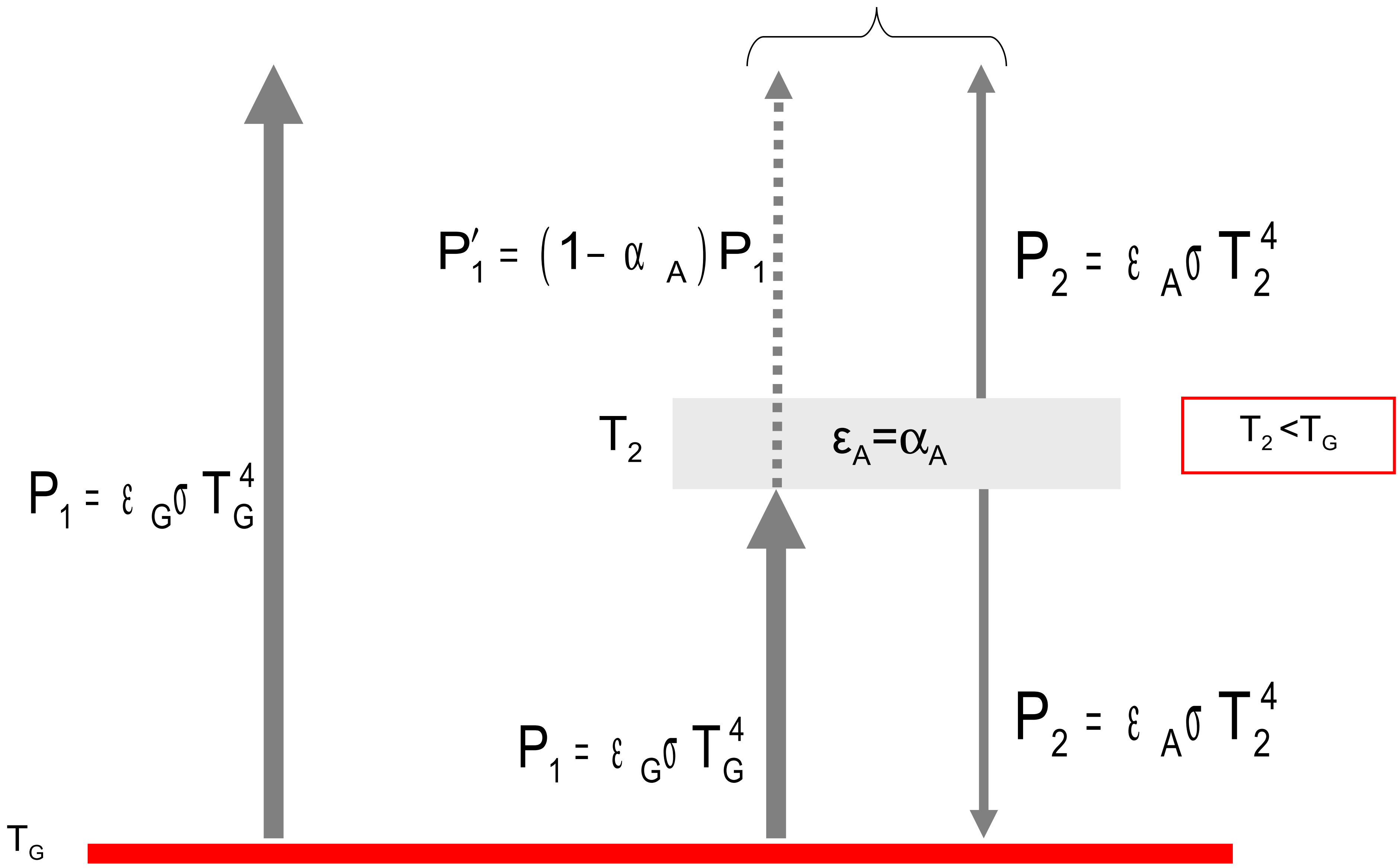
Temperature T_A,

IR-Absorptivity = Emissivity ε_A,

Completely transparent for short wave radiation

Calculation with Absorbing Layers

$\sigma = \sigma_{SB}$



Sum < P_1

$P_1' = \left(1 - \alpha_A\right) P_1$

$P_2 = \varepsilon_A \sigma T_2^4$

T_2

$\varepsilon_A = \alpha_A$

$T_2 < T_G$

$P_1 = \varepsilon_G \sigma T_G^4$

$P_1 = \varepsilon_G \sigma T_G^4$

$P_2 = \varepsilon_A \sigma T_2^4$

T_G

Surface Temperature 2nd Approximation

Radiation equilibrium for both layers:

1) Ground: $$\underbrace{\frac{S_0}{4}\left(1-A\right)}_{\text{mean insolation}} + \underbrace{\varepsilon_A \sigma T_A^4}_{\text{"counter-radiation"}} = \underbrace{\varepsilon_G \sigma T_G^4}_{\text{outgoing radiation from ground}}$$

$\varepsilon_A = \alpha_A$

Atm. radiates upwards and downwards!

2) Atmosphere: $$\underbrace{\varepsilon_A \varepsilon_G \sigma T_G^4}_{\text{IR-Absorption Atm.}} = \underbrace{2\varepsilon_A \sigma T_A^4}_{\text{Emission Atm.}}$$

2 Eq., 2 unknown variables: T_A, ε_A → Solution:

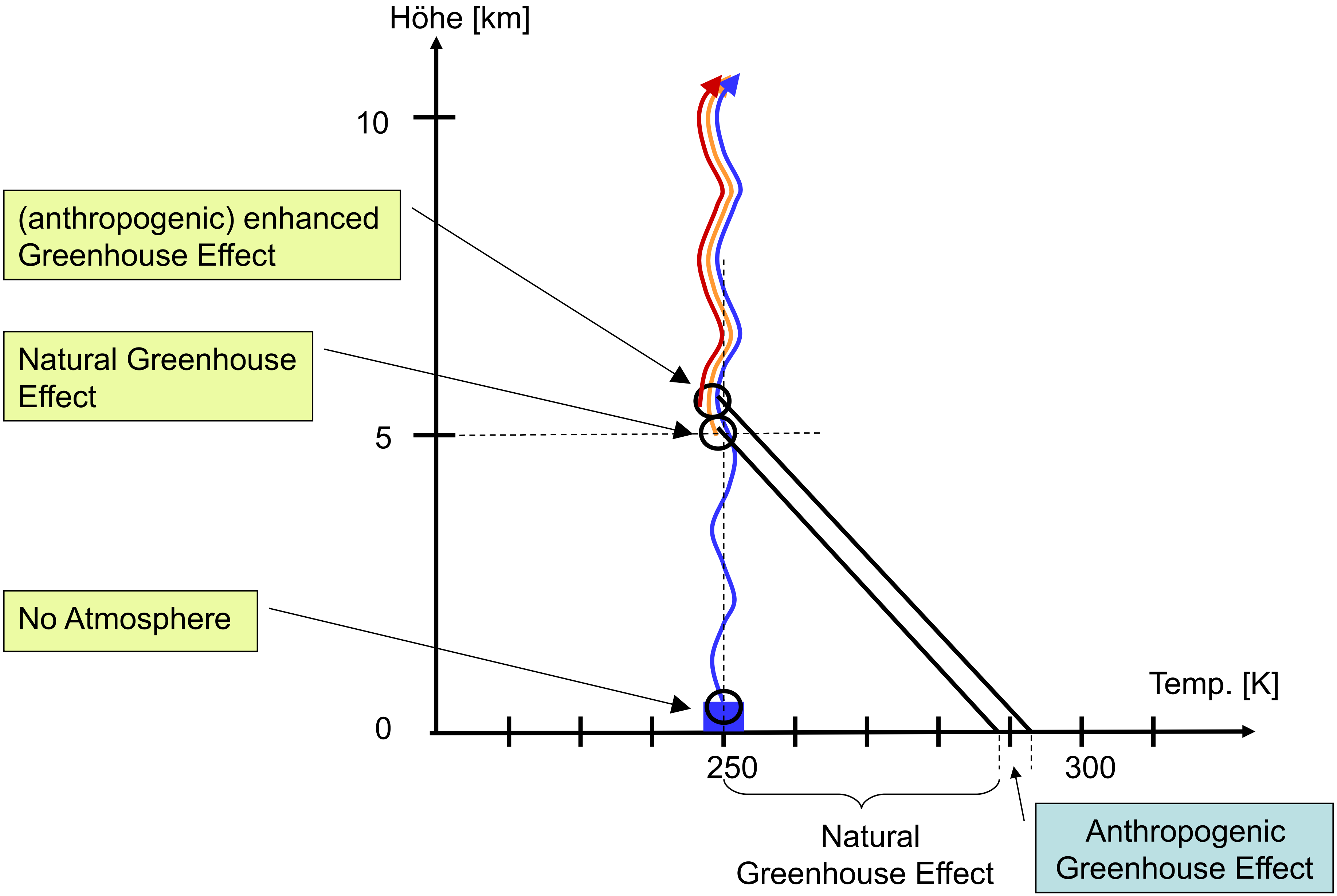
$$\varepsilon_A = 2 - \frac{S_0\left(1-A\right)}{2T_G^4\sigma\varepsilon_G}; \quad T_A = \left(\frac{S_0\left(1-A\right)}{4\sigma\left(2-\varepsilon_A\right)}\right)^{1/4}; \quad T_G = \left(\frac{S_0\left(1-A\right)}{2\varepsilon_G\sigma\left(2-\varepsilon_A\right)}\right)^{1/4}$$

→ $\varepsilon_A = 0.7$ and $T_A = 239$ K (-34 °C) for $T_B = 288$ K

Note: Effektive emission from higher, colder layer!

$\Delta T = T_B - T_A = 49$K → $z = \Delta T/\Gamma \approx 49\text{K}/9.8\text{Kkm}^{-1} \approx 5.0$km altitude

Effective Emission Height



(anthropogenic) enhanced Greenhouse Effect

Natural Greenhouse Effect

No Atmosphere

Natural Greenhouse Effect

Anthropogenic Greenhouse Effect

IV.c) 3rd Approximation: Atmosphere with Line Absorbers

The atmosphere is not ‘**grey’** but rather a **Line-Absorber**

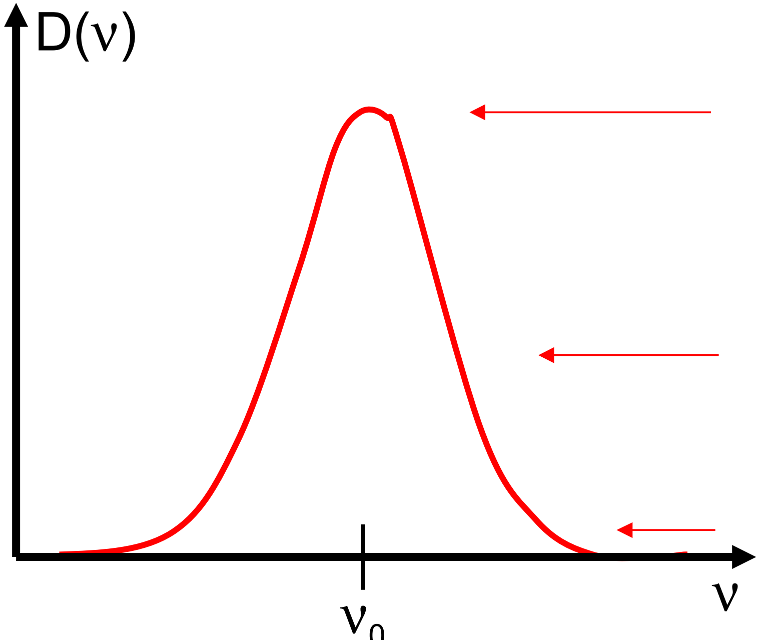
Frequently the optical density ($D = \sigma(\nu)\cdot c_{Gas}\cdot z_0$) at line centre ($\nu_0$) D_0 >>1.

Consequences:

1. The integral absorption S_L of a line:

$$S_L = \int_{-\infty}^{+\infty} e^{-D(\nu)} d\nu \ \mu \ \sqrt{D} \ \mu \ \sqrt{c_{Gas}}$$

2. At line centre (ν_0) the atmosphere only becomes optically dense above the tropopause, thus the line centre radiates at higher temperature than the wings of the line.



Optically dense in the stratosphere

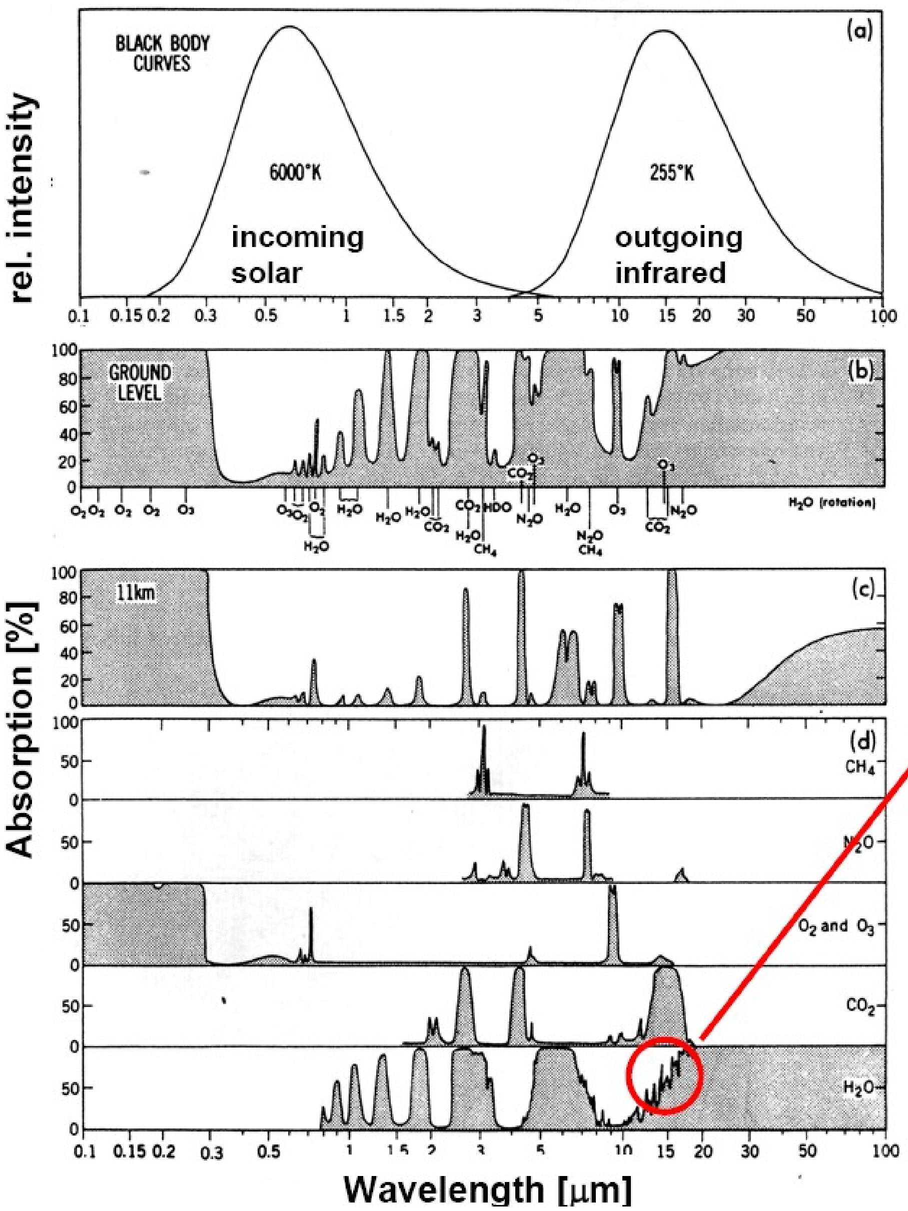
Optically dense in the troposphere

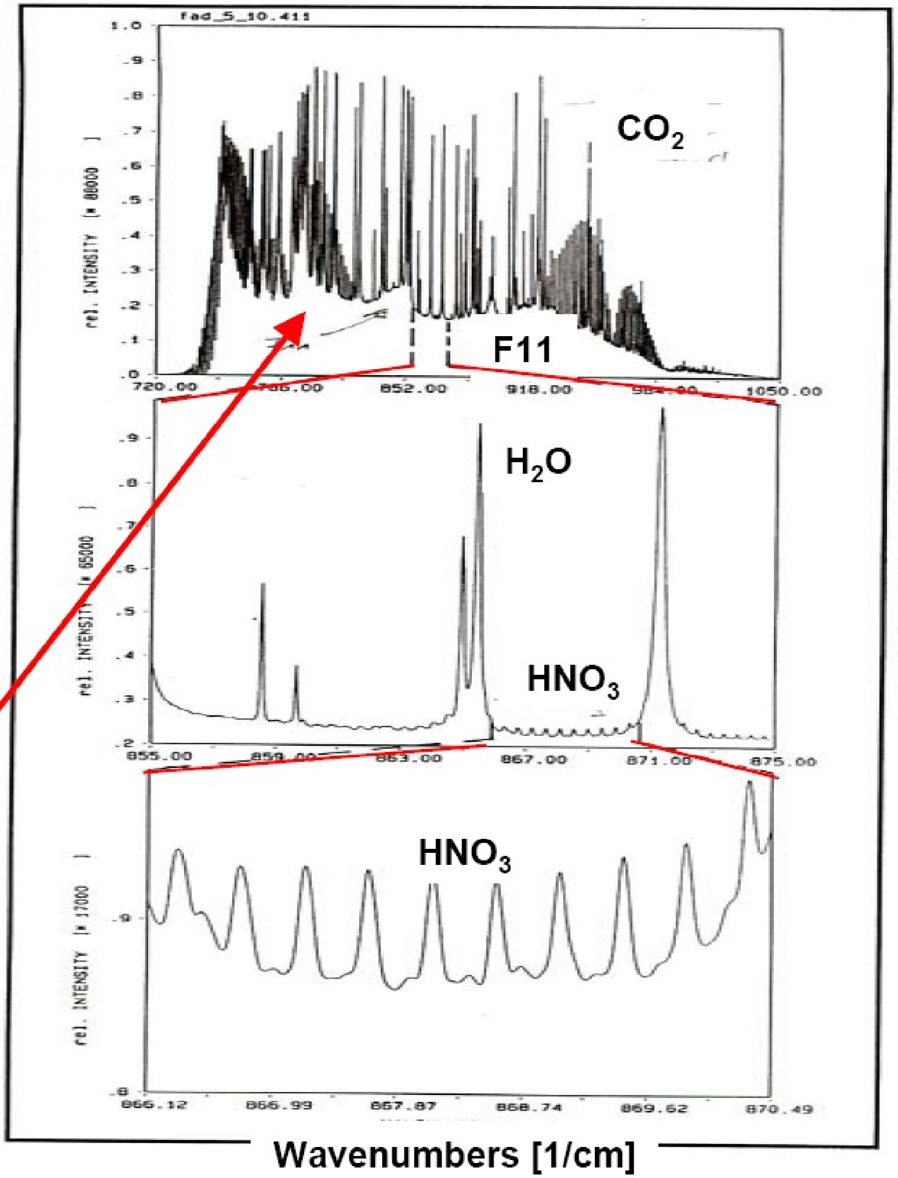
Optically dense near the ground

Detailed view on Absorption Lines

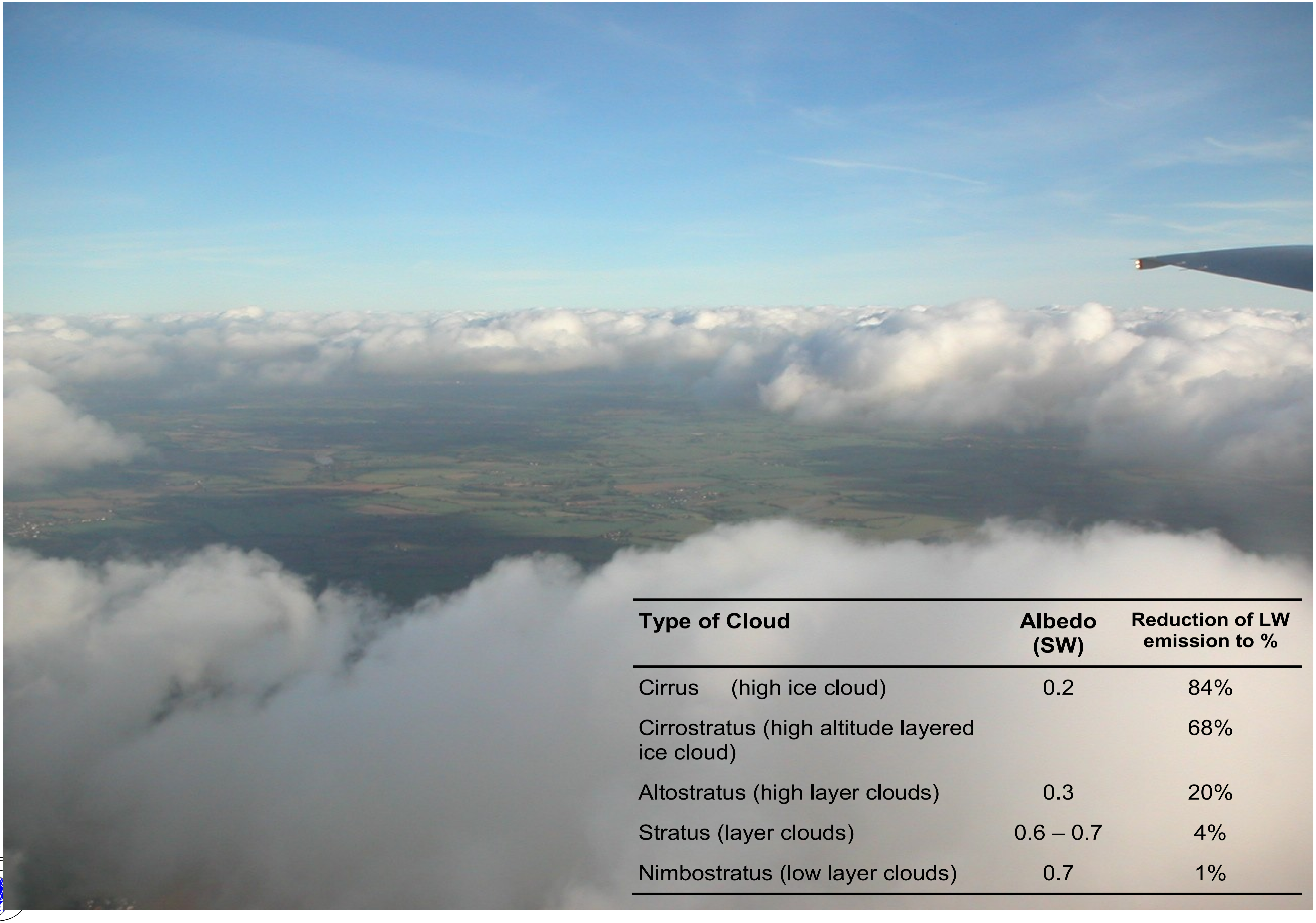
Overview over the atmospheric RT

Zoom into the 886.2 – 870.49 1/cm region



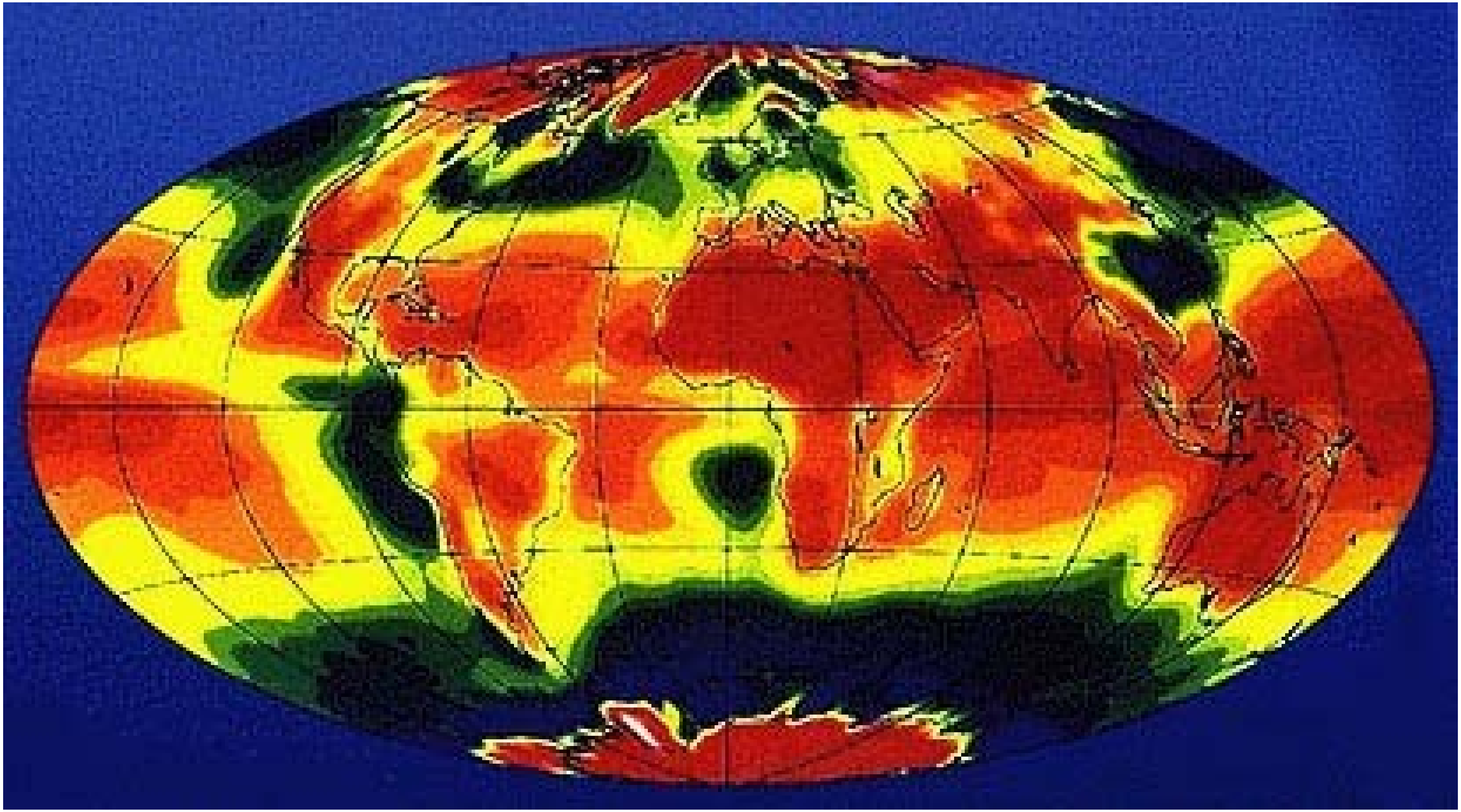


IV.d) 4th Approximation: The Role of Clouds



Type of Cloud	Albedo (SW)	Reduction of LW emission to %
Cirrus (high ice cloud)	0.2	84%
Cirrostratus (high altitude layered ice cloud)		68%
Altostratus (high layer clouds)	0.3	20%
Stratus (layer clouds)	0.6 – 0.7	4%
Nimbostratus (low layer clouds)	0.7	1%

Earth Radiation Budget Experiment – ERBE (Satellite)



cooling - yellow to green to blue
heating - orange to red to pink
overall small net cooling effect by clouds

(image produced at the University of Washington;
from NASA webpage, http://terra.nasa.gov/FactSheets/Clouds/)

Effect of Clouds

Results of ERBE:

Annual Average Net Cloud Radiative Forcing, 1985 - 1986.

Net cloud forcing is the result of two opposing effects:
(1) greenhouse heating by clouds (or positive forcing),
(2) cooling by clouds (or negative forcing) — clouds reflect incoming solar radiation back to space.

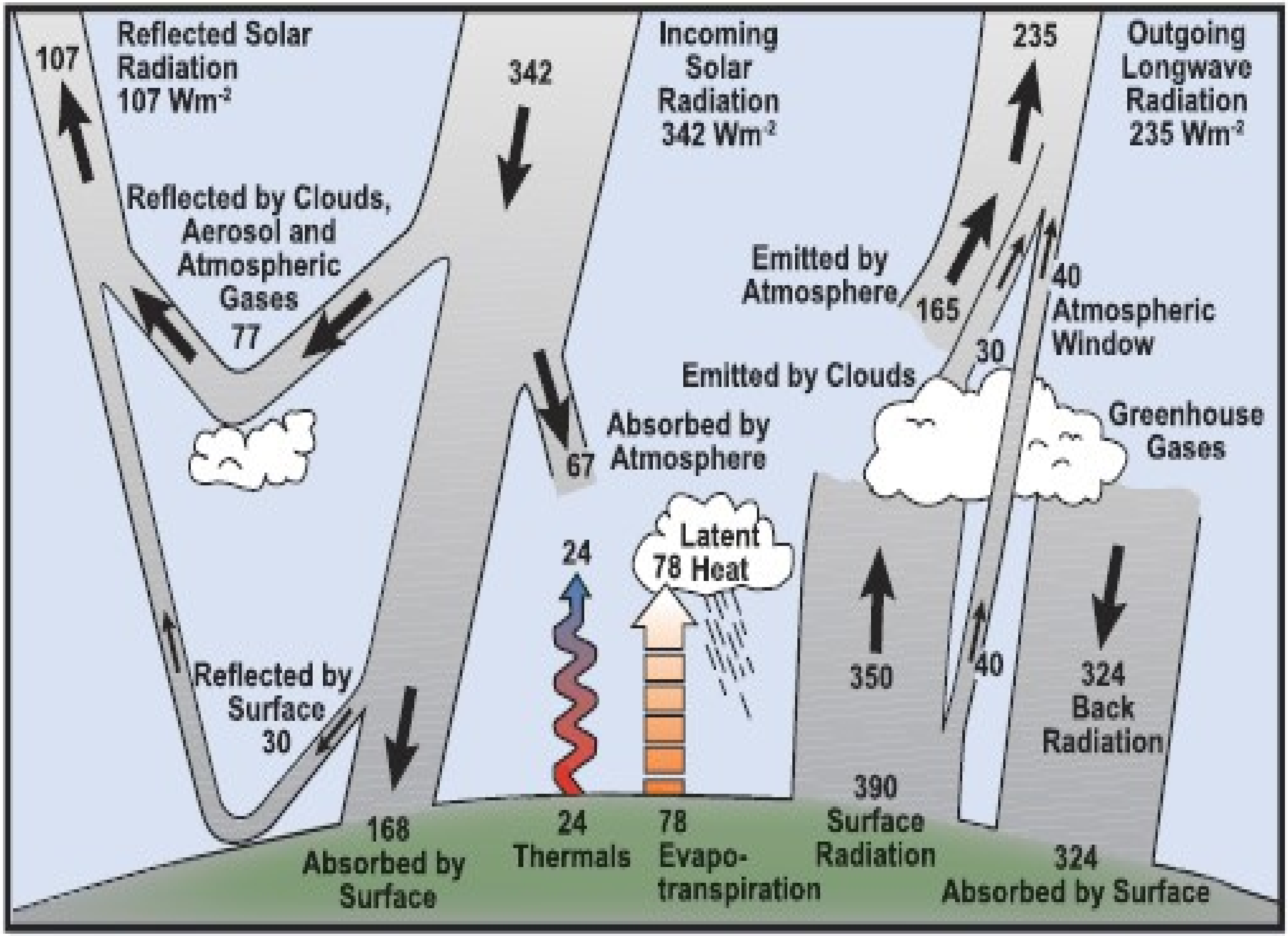
Overall, clouds have the effect of lessening the amount of heating that would otherwise be experienced at Earth's surface-a cooling effect.

(high) Clouds reduce IR emission → heat Earth

However:(low) Clouds cool due to high albedo!

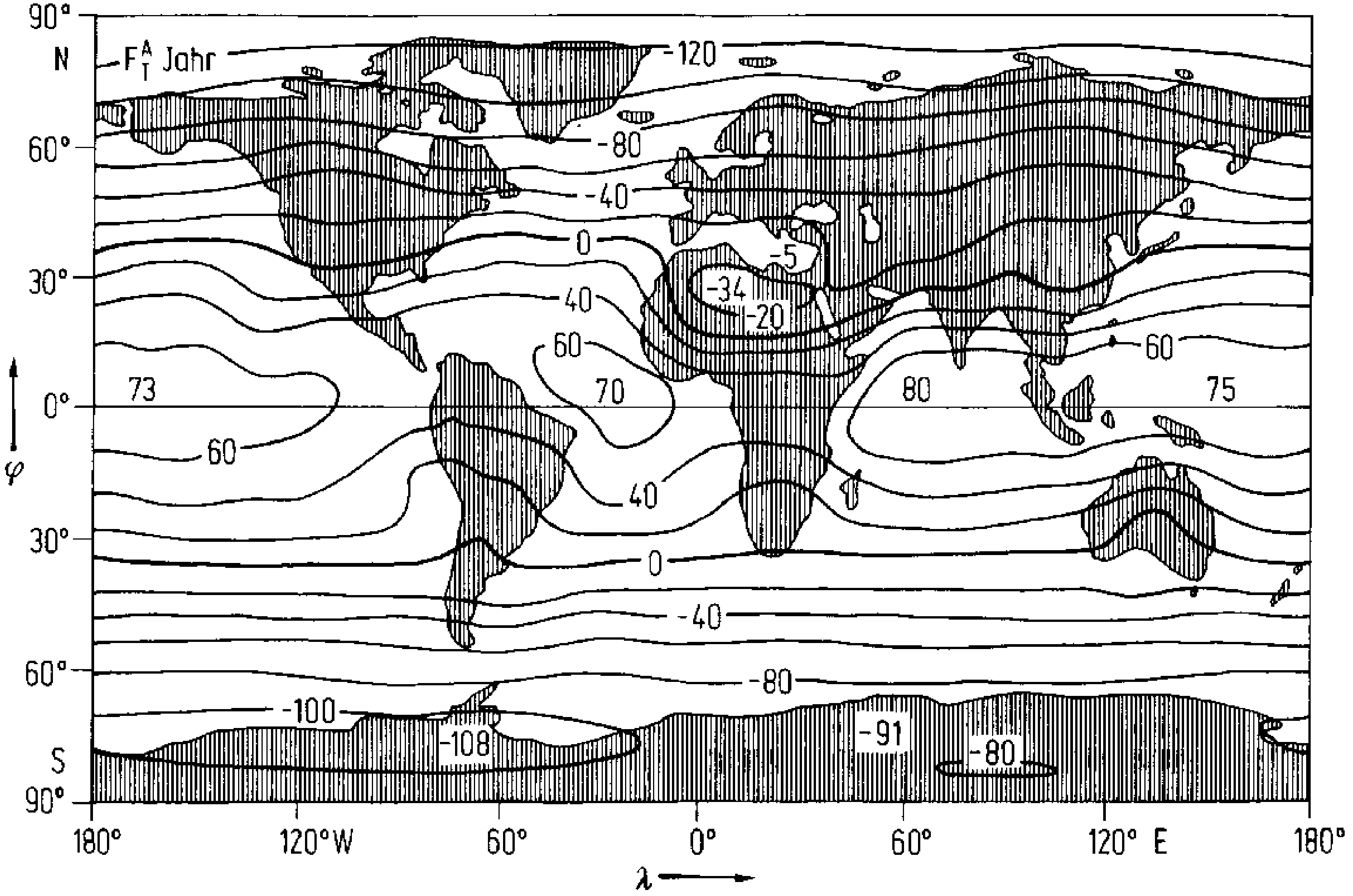
Net-effect: slight cooling

V. Global Mean Energy Budget of Earth



Keihl and Trenberth 1997

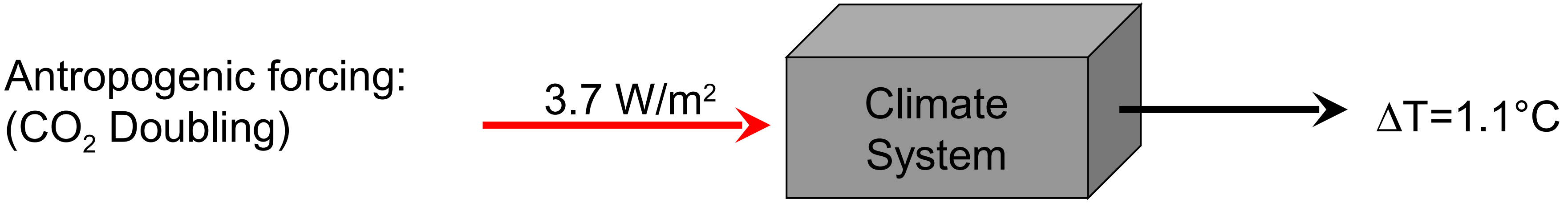
Radiation Budget from Satellite



Yearly average of net radiation flux density in W/m², positive numbers: Gain of radiation energy (net radiation flux downwelling)

VI. The Effects of Climate Feedback

- Without climate feedback climate predictions would be rather simple:



Antropogenic forcing:
(CO$_2$ Doubling)

3.7 W/m^2

Climate System

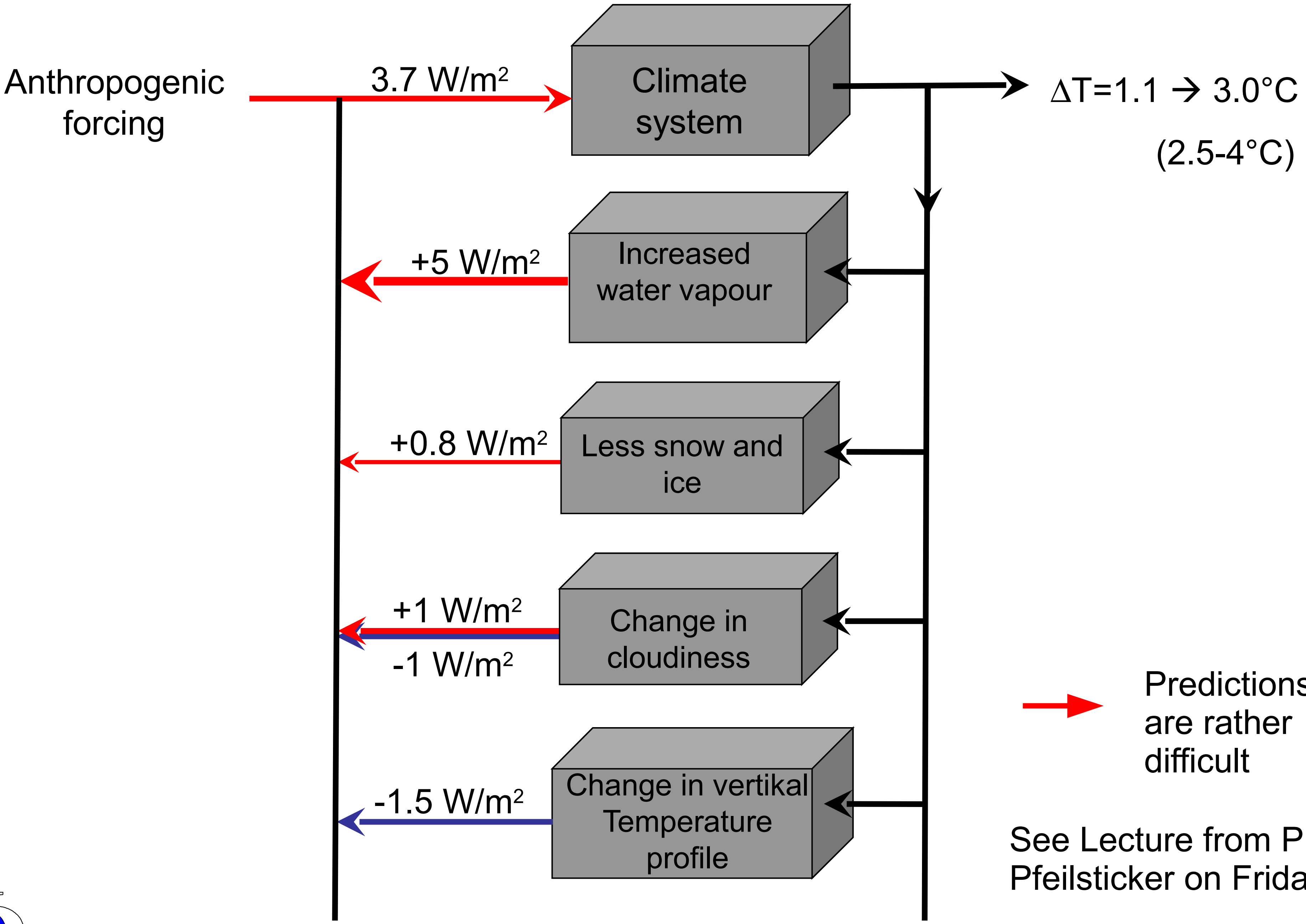
ΔT=1.1°C

Doubling of the CO$_2$ concentration:

→ temperature increase of about 1.1K

Schwartz, S. E. (2007), Heat capacity, time constant, and sensitivity of Earth’s climate system, J. Geophys. Res., 112, D24S05, doi:10.1029/2007JD008746.

Overview of Climate Feedbacks



Anthropogenic forcing

3.7 W/m²

Climate system

ΔT=1.1 → 3.0°C

(2.5-4°C)

+5 W/m²

Increased water vapour

+0.8 W/m²

Less snow and ice

+1 W/m²

-1 W/m²

Change in cloudiness

-1.5 W/m²

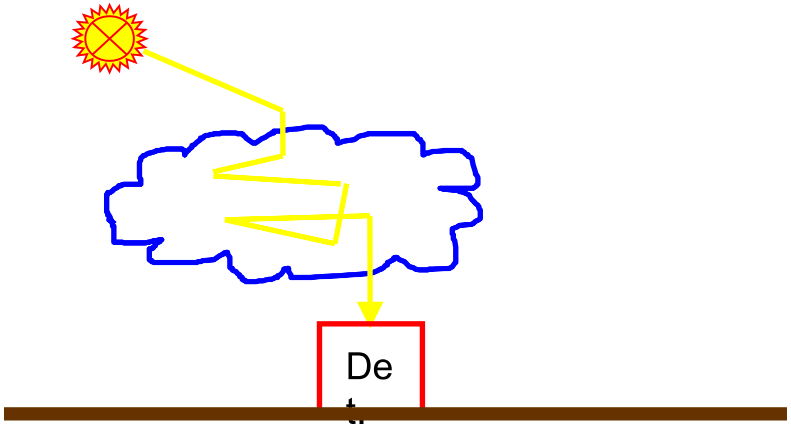
Change in vertikal Temperature profile

→ Predictions are rather difficult

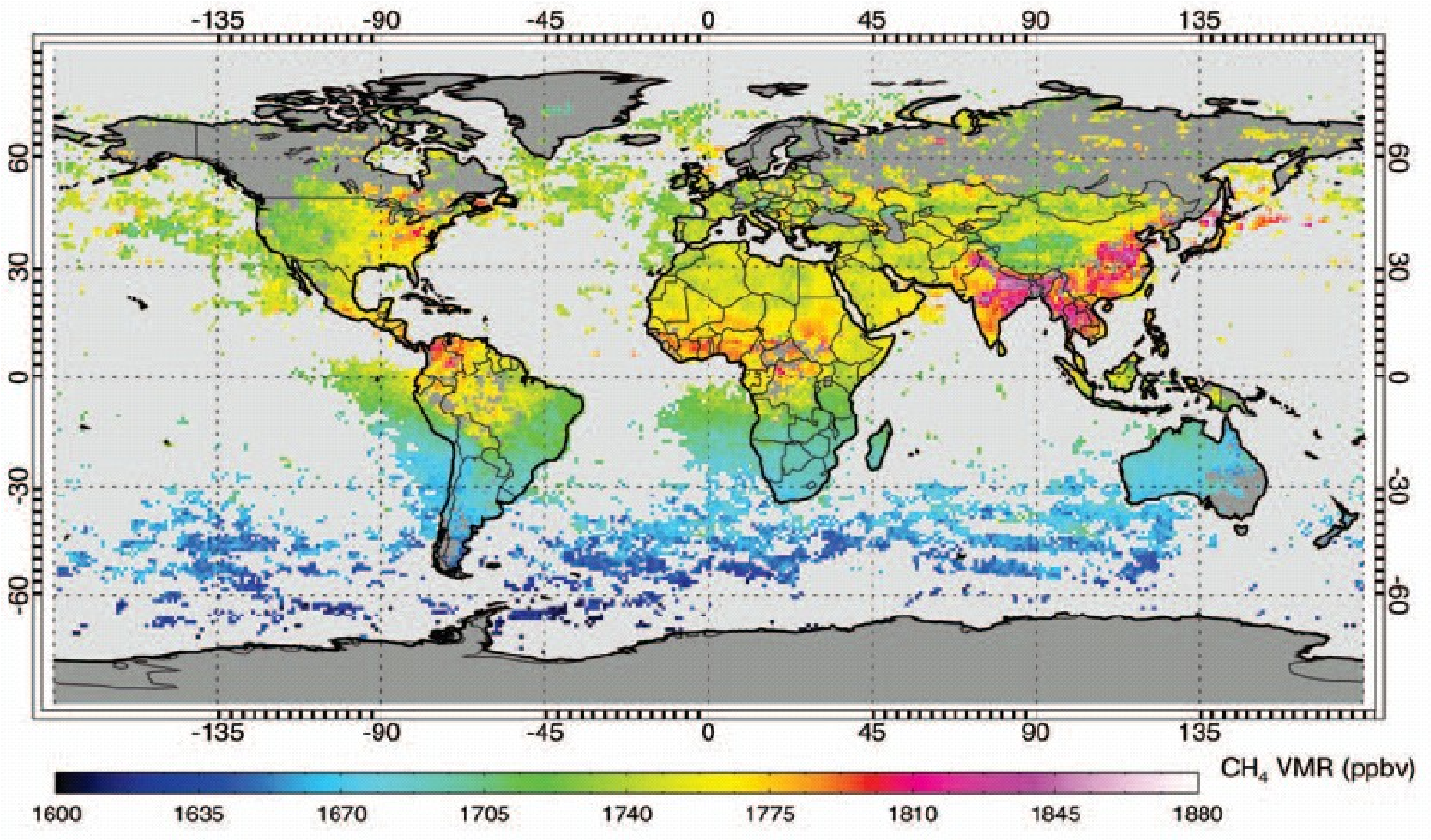
See Lecture from Prof. Pfeilsticker on Friday

VII. Contribution of the IUP to climate research (some examples)

- SW – „extra“ energy absorption in the Atmosphere
- Light path lengths in clouds
- Greenhouse Gases: CH_4 , O_3
- Cloud – feedback
- Carbon cycle
- Paleo climate
- Studying change of climate relevant trace gases



Satellite mesurements of Global CH_4 mixing ratio (Aug. – Nov. 2004)



Frankenberg C., Meirink J.F., van Weele M., Platt U. and Wagner T., Assessing methane emissions from global space-borne observations, Science express, March 17, 2005

Investigation change in Cloud Cover

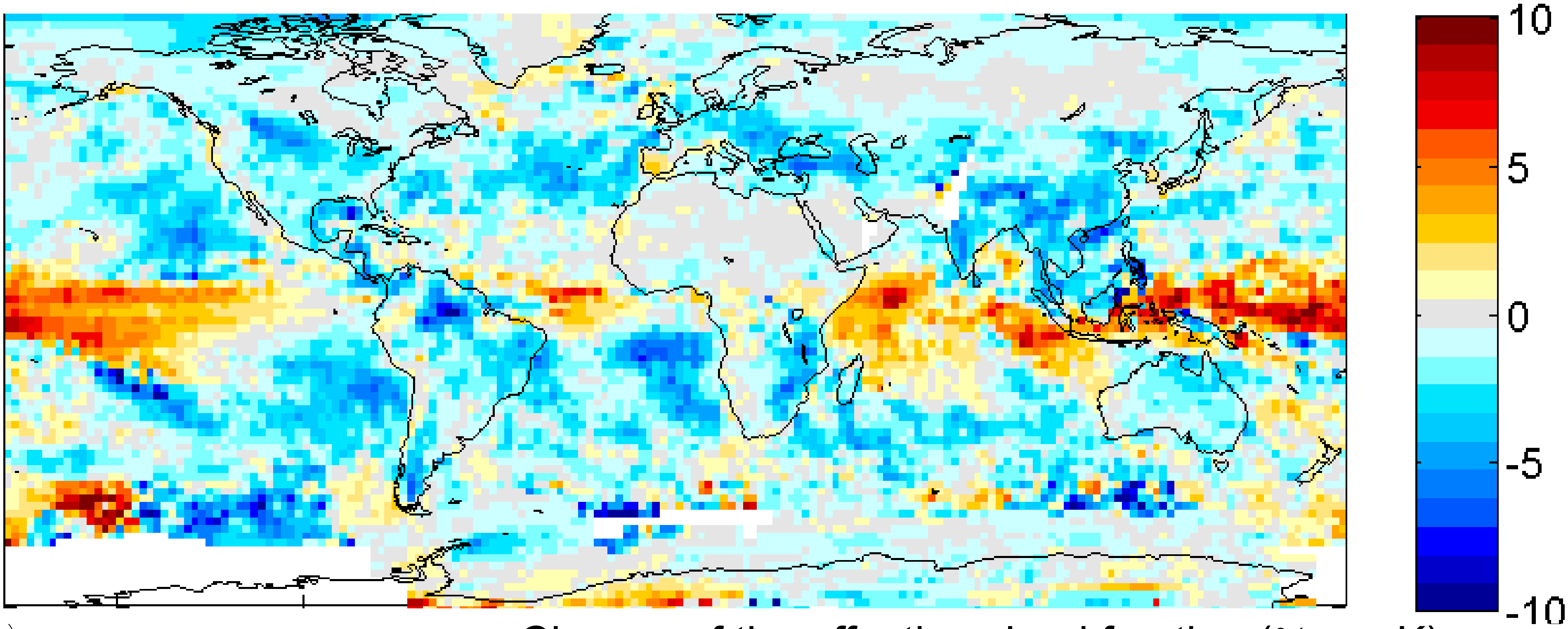
General rule:

Low clouds tend to cool
High clouds tend to warm

Main question:
Does cloud cover and/or distribution change when climate changes?

Cloud fraction vs. temperature (1996-2003)

Higher Temperature → Fewer Clouds → lower Albedo → positive feedback



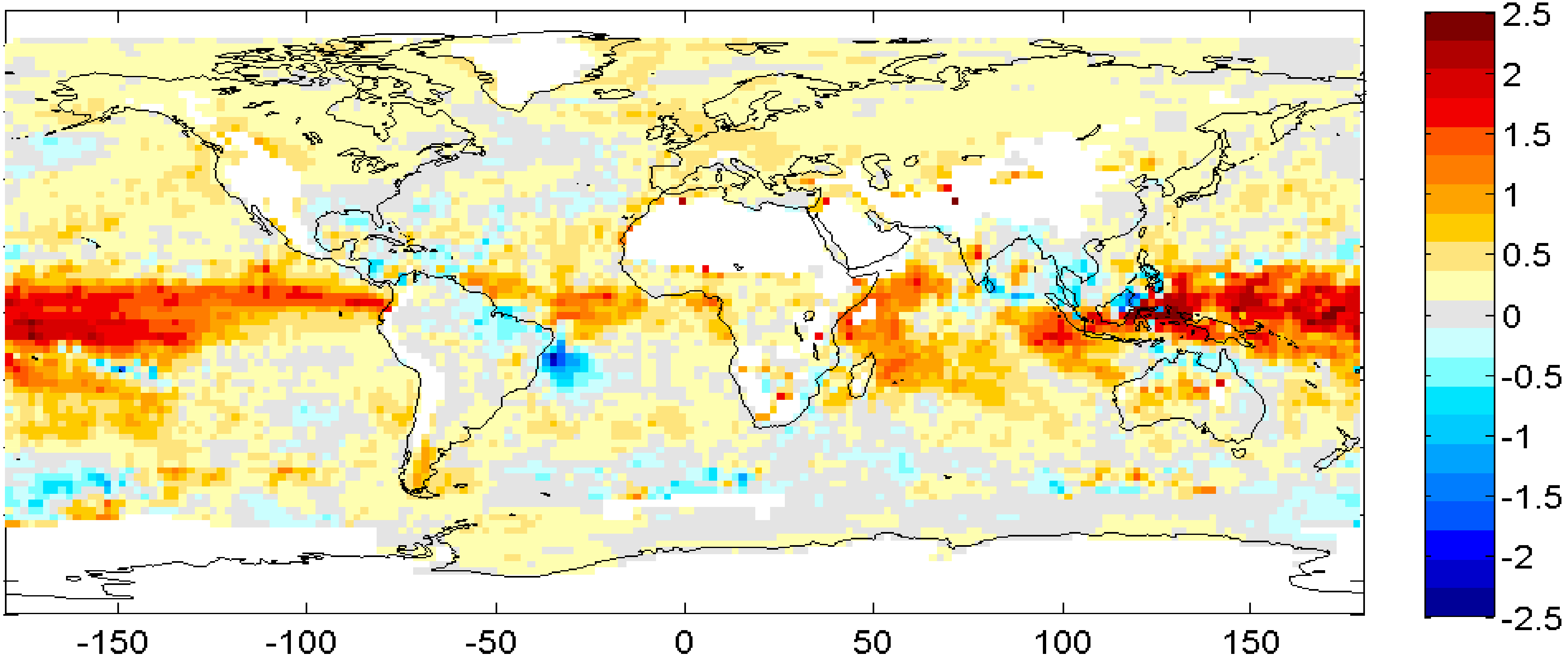
Change of the effective cloud fraction (% per K)

Wagner et al. 2007

Investigation of Cloud hight

Dependence of the Cloud Top Height (from O_2) on Temperature 1996-2003

Higher Temperature → Higher Clouds
→ positive feedback on temperature



Wagner et al. 2007

Change of cloud top height (km per Kelvin)

VIII. Summary

- The atmosphere consist of many gases with very different concentrations (% to less that sub ppt)
- According to their lifetime and atmospheric dynamics they are relevant on different scales
- The atmosphere is separated in layers
- Vertical air mass movement can be derived with the potential temperature
- Atmospheric radiation drives the energy budget of the earth
- Absorption, emission, scattering processes are essential to understand the processes in atmospheric radiation
- Wavelength dependent absorption of different trace gases have to be taken into account
- The emission spectrum depends on the temperature and is thus different for the sun and earth
- The earth can be described as a grey body
- Due to multi scattering events the photon propagation can only be studied in detail with Radiative Transfer Models

- Clouds enhance the light path and reduce especially intensity of short wave radiation
- Warming in a greenhouse arise mainly do to suppression of convection and not due to IR absorption
- Calculation of earth surface temperature without atmosphere would be 33°C lower than it is now
- Natural greenhouse effect lead to a earth with 33°C warmer average ground temperature
- The global energy budget give a net. gain at low latitudes and net. loss at high latitudes
- Atmosphere is invisible in many spectral ranges and has view spectral windows, variation in the absorption pattern may have large influences
- in a simple model the atmospheric hight would be related to 5km altitude
- A detailed investigation requires the analysis of absorption line structures and clouds
- Climate forecast are difficult due to feedbacks

IX. Websites on the Topic 1

Intergovernmental Panel on Climate Change (IPCC) **http://www.ipcc.ch/**
Presents the comprehensive reports (presently 4th report of 2007) of an international, independent group of scientists on climate and climate change.

European Ozone Research Coordinating Unit **http://www.ozone-sec.ch.cam.ac.uk/**
Results of European research on stratospheric chemistry and ozone loss.

NASA Life on Earth **http://www.nasa.gov/vision/earth/features/index.html**
Comprehensive site centred on remote sensing of planet earth.

German Weather Service (Deutscher Wetterdienst) **http://www.dwd.de/en/en.htm**
Information on weather and climate

National Oceanic and Atmospheric Administration (NOAA) **http://www.noaa.gov/**

The Federal Environmental Agency (Umweltbundesamt)
http://www.umweltbundesamt.de/index-e.htm
Has much information on the state of the environment in Germany.

IGBP - International Geosphere-Biosphere Programme
http://www.igbp.kva.se/cgi-bin/php/frameset.php
Mission: Deliver scientific knowledge to help human societies develop in harmony with Earth’s environment.

Institute for Environmental Physics - Institut für Umweltphysik
http://www.iup.uni-heidelberg.de Our own web-page

Websites on the Topic 2

Journal of Atmospheric Chemistry and Physics, **http://www.copernicus.org/EGS/acp**
This journal comes in two versions:
1) JACP – Discussion 2) JAPC The reviewed journal
both are completely in the internet and freely available

“The Master Chemical Mechanism”
http://www.chem.leeds.ac.uk/Atmospheric/MCM/mcmproj.html

JPL-Compilation: “Chemical Kinetics and Photochemical Data for Use in Stratospheric Modelling” **http://jpldataeval.jpl.nasa.gov/**

NIST-Compilation: **http://kinetics.nist.gov/index.php**

IUPAC-Compilation: **http://www.iupac-kinetic.ch.cam.ac.uk/**

NASA’s “Visible Earth” **http://visibleearth.nasa.gov**

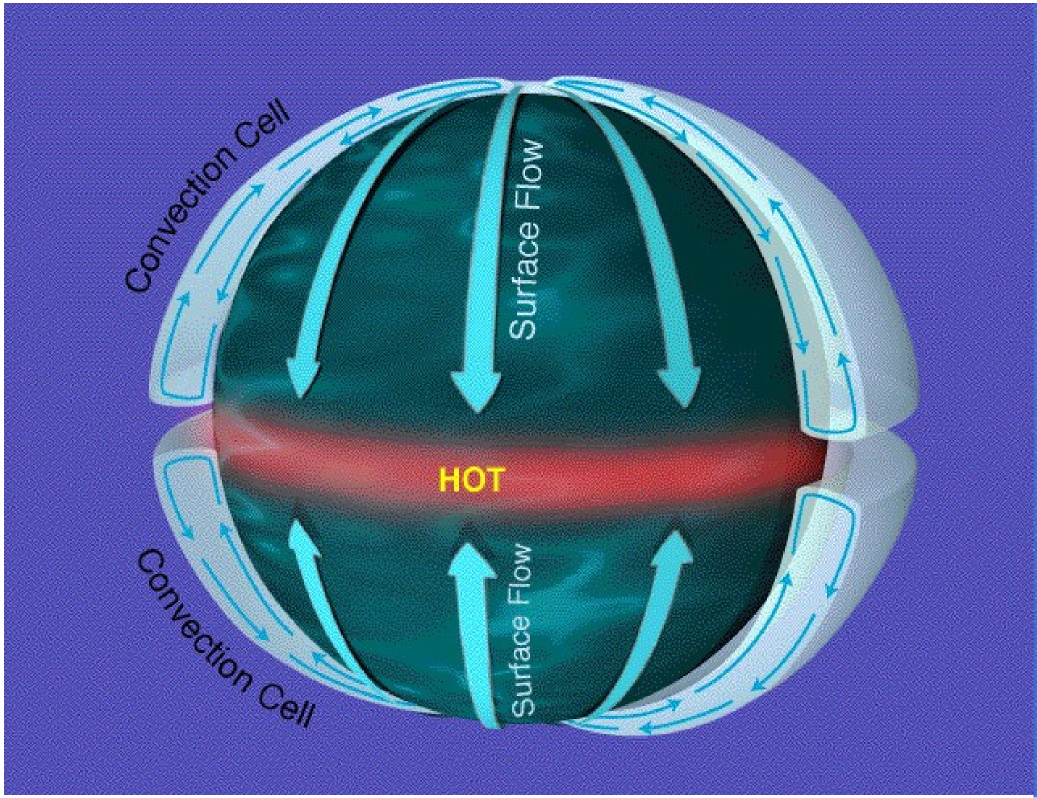
Aerosol Inorganic Modelling Home Page:
http://www.hpc1.uea.ac.uk/~e770/aim.html

Literature

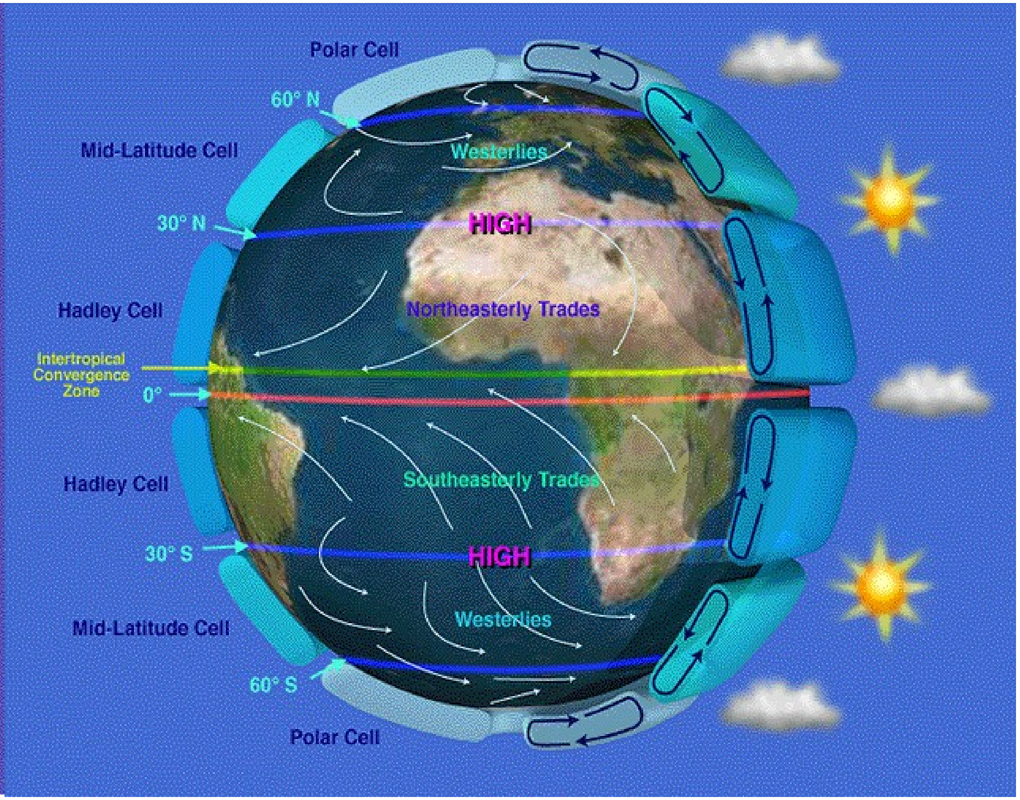
- (**IUP xxxx:** Book - Nr. in the library of the Institut für Umweltphysik, INF 229, 4th floor, R.410)
- **Physik unserer Umwelt, Die Atmosphäre**, Walter Roedel, Springer, Heidelberg, 3. Aufl., Eine ausgezeichnete Übersicht der physikalischen Prozesse in der Atmosphäre (**IUP 1511**).
- The Physics of Atmospheres, J. T. Houghton, Cambridge University Press, Cambridge 3rd Edition, 2002. Good, concise introduction.
- Atmosphere Chemistry and Physics, J.H. Seinfeld und S.N. Pandis, John Wiley & Sons, New York, 1998. Very comprehensive book on physics and chemistry of troposphere and stratosphere (**IUP 1724**).
- Theoretische Meteorologie, Eine Einführung, Dieter Etling, Springer Verlag Heidelberg, 2nd ed., 2002, ausführliche Einführung in Dynamik (UB LN-W 23-8691).
- Fundamentals of Atmospheric Physics, M. L. Salby, Academic Press, 1996: Failry comprehensive introduction to all aspects of atmospheric physics (**IUP 1647**).
- Aeronomy of the Middle Atmosphere, 2nd edition, G. P. Brasseur and S. Solomon, 2005 (**IUP 1211**).
- Fundamentals of Physics and Chemistry of the Atmosphere, G. Visconti, Springer-Verlag, 2001; Good and concise text, sometimes surprising mistakes.
- Fundamentals of Atmospheric Modeling, M. Z. Jacobson, Cambridge University Press, 2005; Covers physics and chemistry of troposphere and stratosphere with the goal to provide the relevant equations for numerical modeling. (**IUP 1925**)
- Physics of Climate, J. P. Peixoto, A. H. Oort, American Inst. of Physics, 1992; Dynamics, radiation, thermodynamics and a lot of observational data on climate (**IUP 1409**).

VII. Atmospheric Dynamics: Global Circulations Troposphere

Hypothetical circulation on a non-rotating Earth:
Just one convection cell between Equator and Pole



Real circulation on a rotating Earth:
Three convection cells between Equator and Pole



The Concept of Air Masses

The term „Air Mass“ denotes an extended volume of air with unique properties (e.g. temperature, humidity, PV, vertical stability).

Definition of an Air-Mass:

- horizontal extent > 500 km
- vertical extent > 1 km
- horizontal temperature gradient < 1K / 100km

Air Masses are generated if constant conditions prevail for sufficiently long times. Prerequisites are small pressure gradients and thus only slow motion of the air mass.

Owing to these conditions, homogeneous air masses predominantly form in the tropics and in polar regions. In mid-latitudes temperature- and pressure gradients are usually too strong to allow formation of distinct air masses.

The Motion of the Atmosphere

The subject of hydrodynamics is the investigation and description of the motion of fluids, (i.e. liquids and gases).

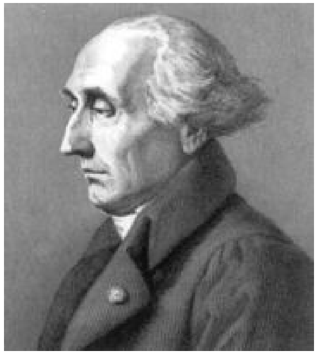
We distinguish between two fundamentally different types of representation:

Leonhard Euler,
1707-1783

1.) Description of flow as the temporal development of the local velocity field $\vec{v}(\vec{x},t)$ in a given point in space $\vec{x}$

→ Eulerian Representation

2.) Description of flow by the temporal development of the position $\vec{x}_m(t)$ of a particular element of mass of the fluid in space.



Joseph Louis Lagrange,
1736-1813

→ Lagrangian Representation

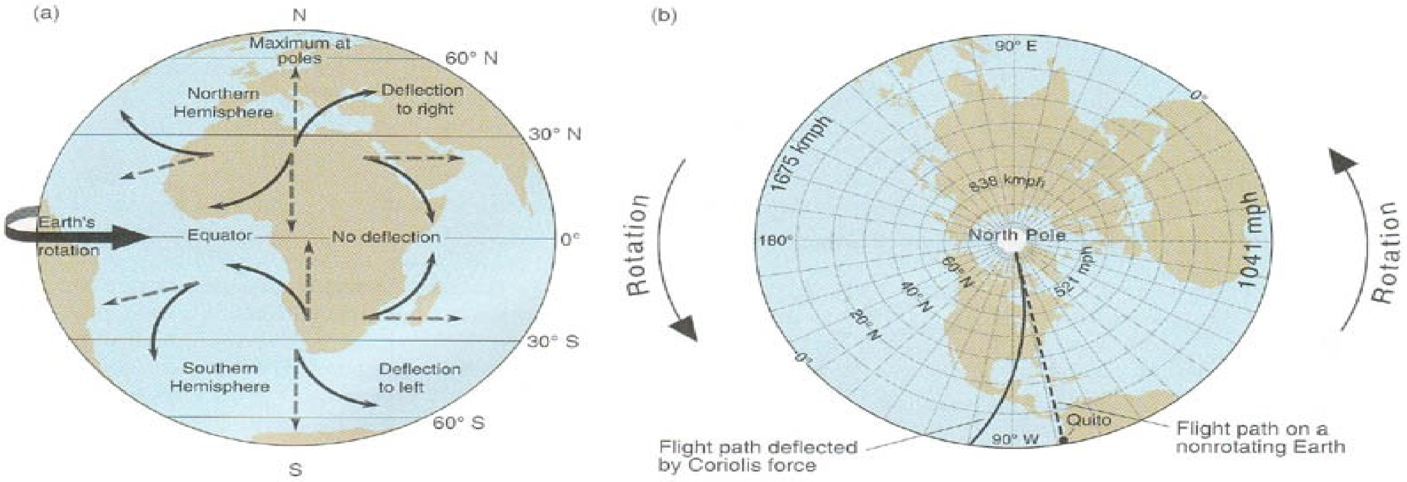
In the Lagrangian Representation the velocity is then obtained from the temporal derivative: $\vec{v}(\vec{x},t) = \frac{\partial}{\partial t}\vec{x}_m(t)$

Description of Atmospheric Motion

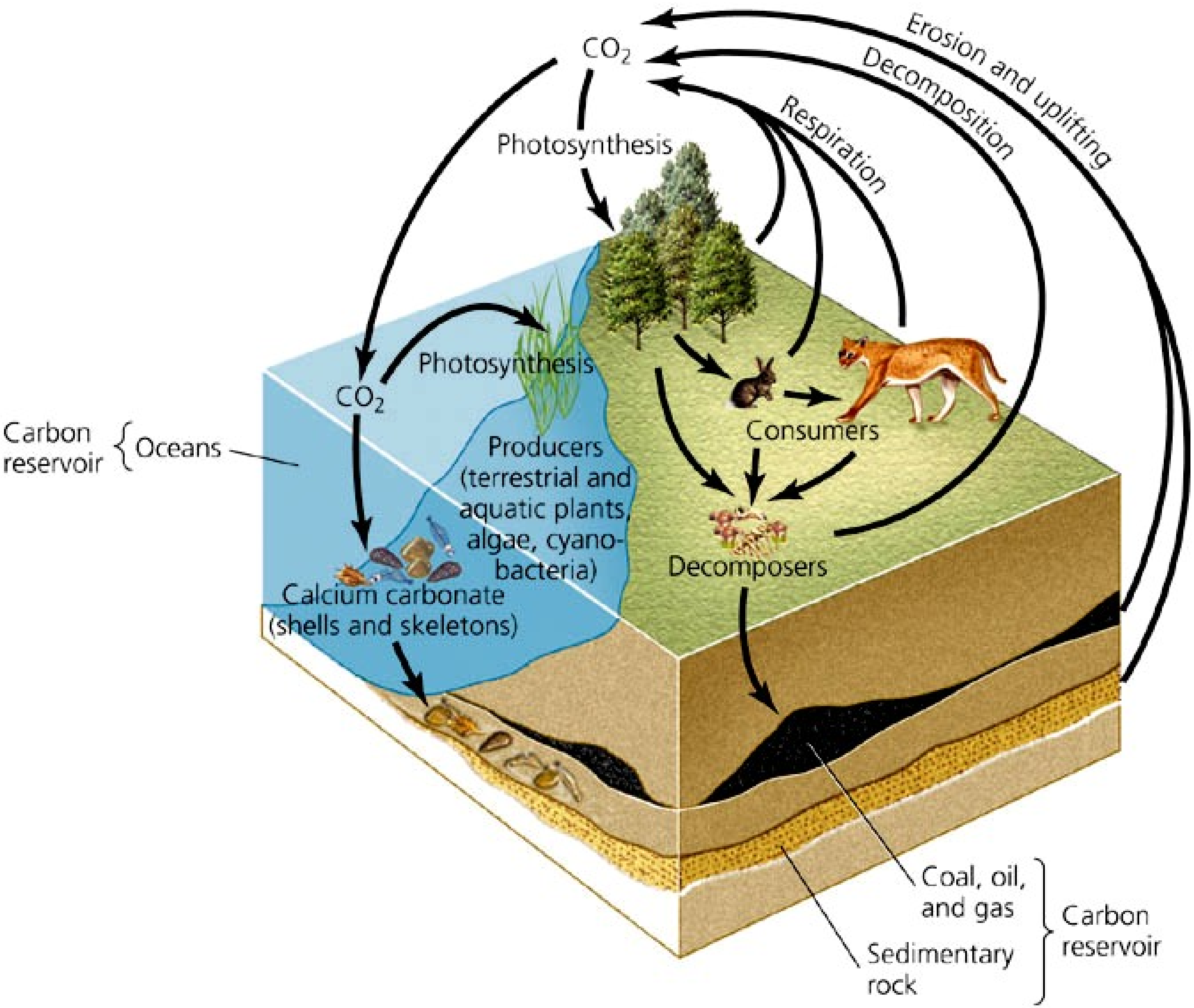
- various (volume)-forces
- Pressure gradient force
- Gravity force
- Inerial forces due to Earth's rotation
- Coriolis force
- Friction force

→ Summing all terms we obtain the Navier-Stokes Equation

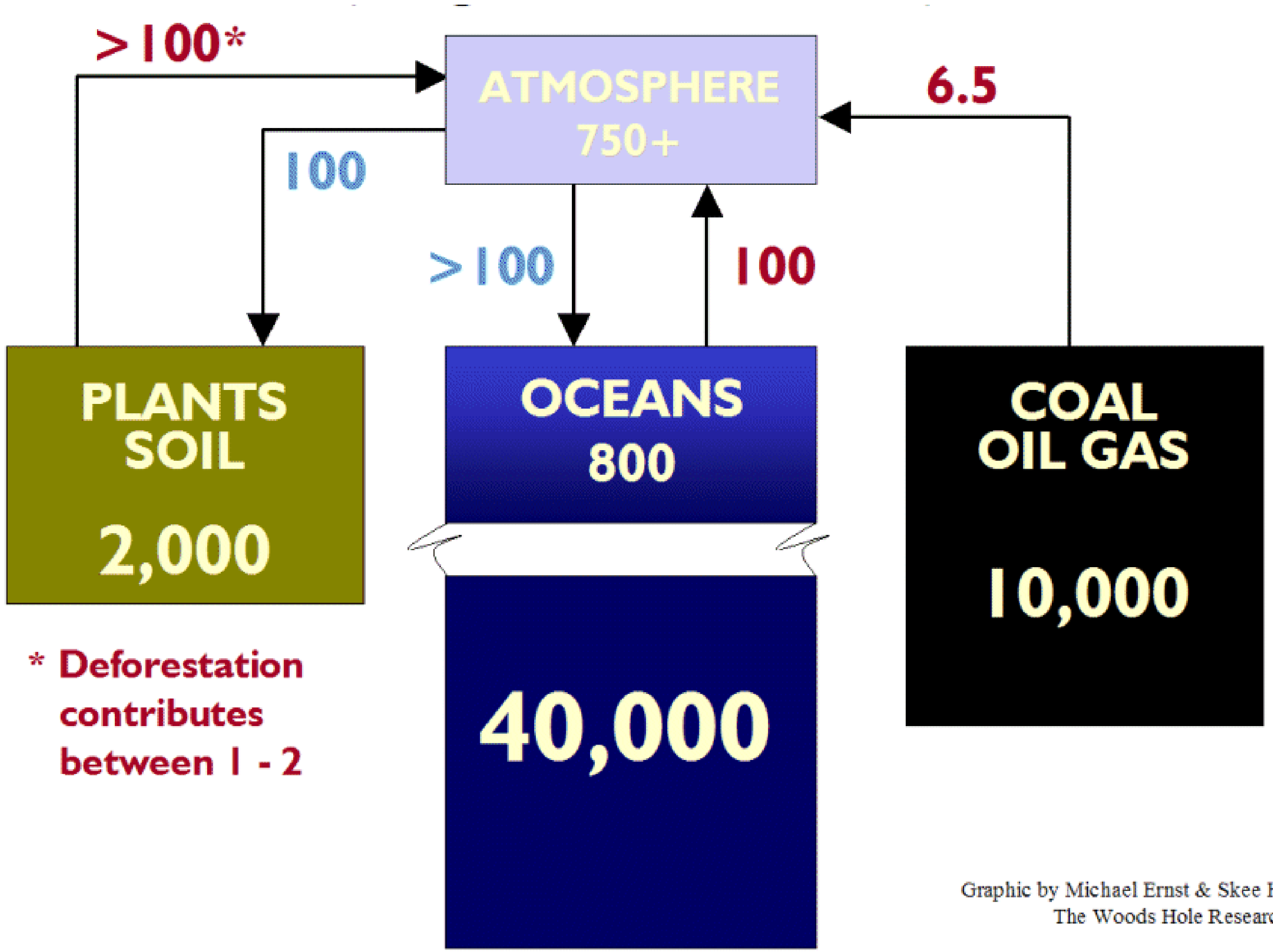
$$\rho \frac{\partial \vec{v}}{\partial t} + \rho \left(\vec{v} \times \vec{\nabla} \right) \vec{v} = -\vec{\nabla} p - \rho g + 2\rho \left(\vec{v} \times \vec{\Omega} \right) + \vec{\nabla} \left[\left(K + \nu \right) \rho \times \vec{\nabla} \right] \vec{v}$$



VII. Global Circulations
The Carbon Cycle



Carbon Cycle Quantities



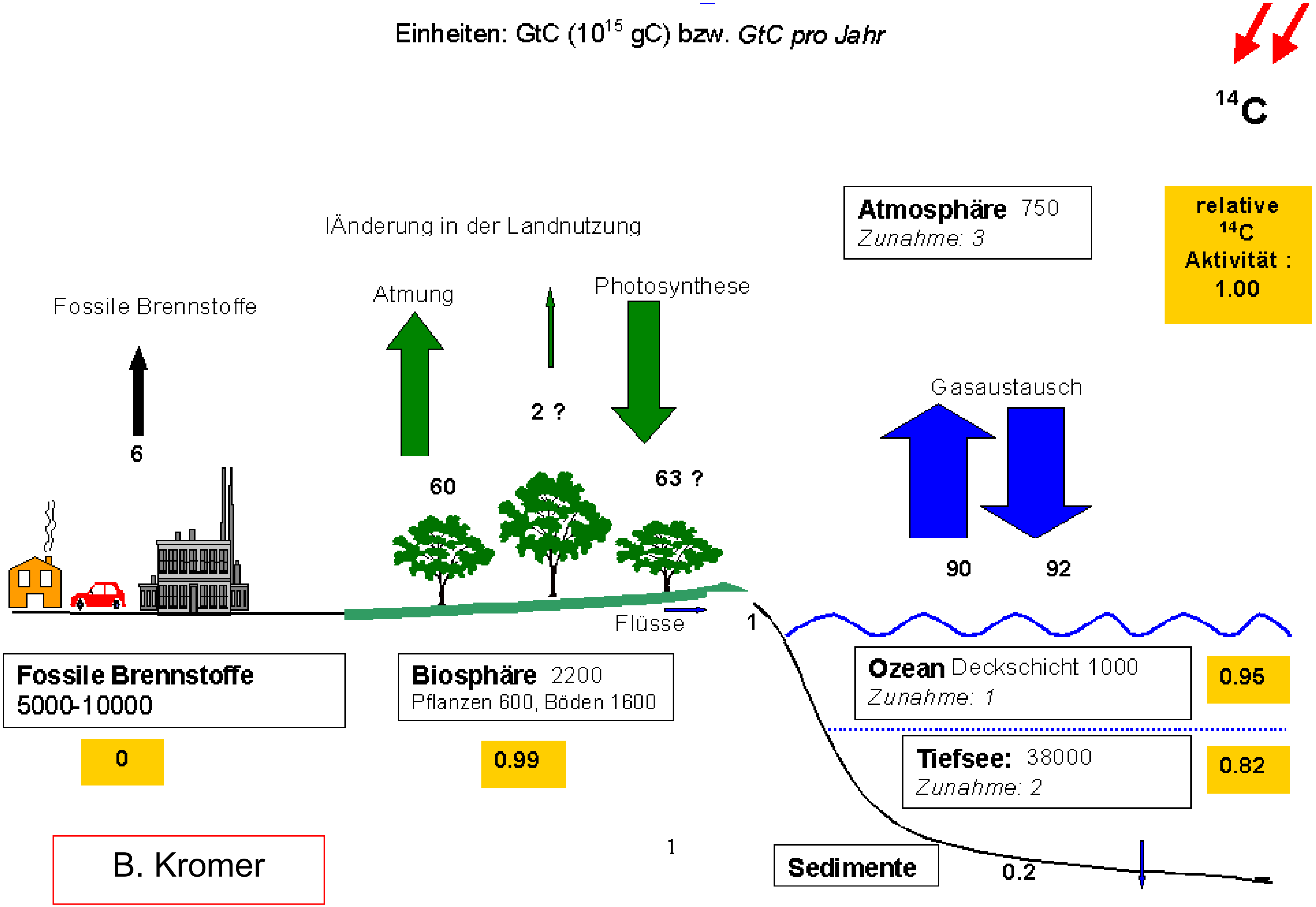
* Deforestation contributes between 1 - 2

Graphic by Michael Ernst & Skee Houghton
The Woods Hole Research Center

http://whrc.org/carbon/index.htm

The Global Carbon Cycle - Quantitative

Einheiten: GtC (10^{15} gC) bzw. *GtC pro Jahr*



^{14}C

Atmosphäre 750
Zunahme: 3

relative
^{14}C
Aktivität :
1.00

lÄnderung in der Landnutzung

Fossile Brennstoffe

6

Atmung

60

2 ?

Photosynthese

63 ?

Gasaustausch

90

92

Flüsse

1

Fossile Brennstoffe
5000-10000

0

Biosphäre 2200
Pflanzen 600, Böden 1600

0.99

Ozean Deckschicht 1000
Zunahme: 1

0.95

Tiefsee: 38000
Zunahme: 2

0.82

1

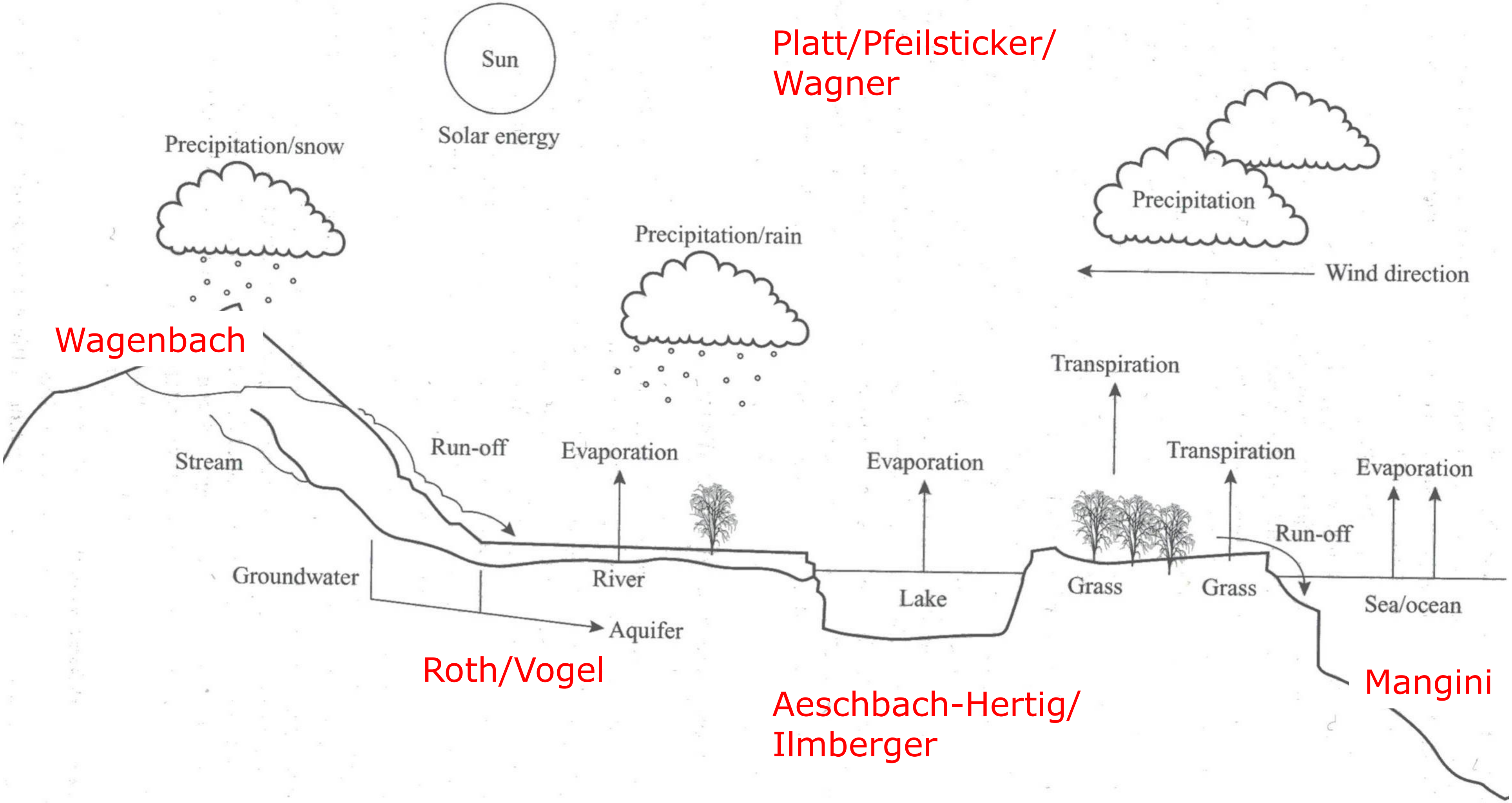
Sedimente

0.2

B. Kromer

The hydrological cycle

research at IUP



Platt/Pfeilsticker/
Wagner

Wagenbach

Roth/Vogel

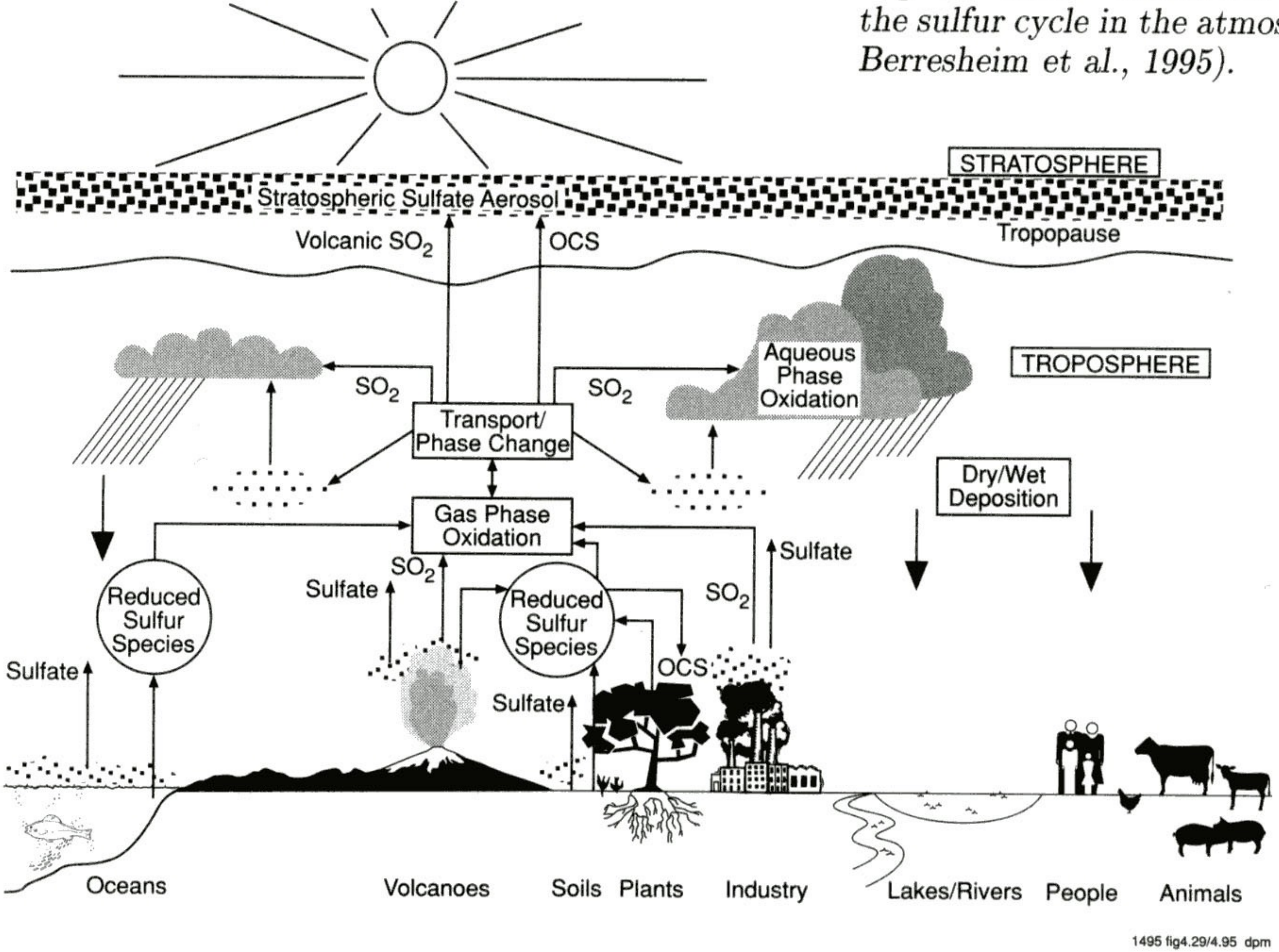
Aeschbach-Hertig/
Ilmberger

Mangini

Wright (2003)

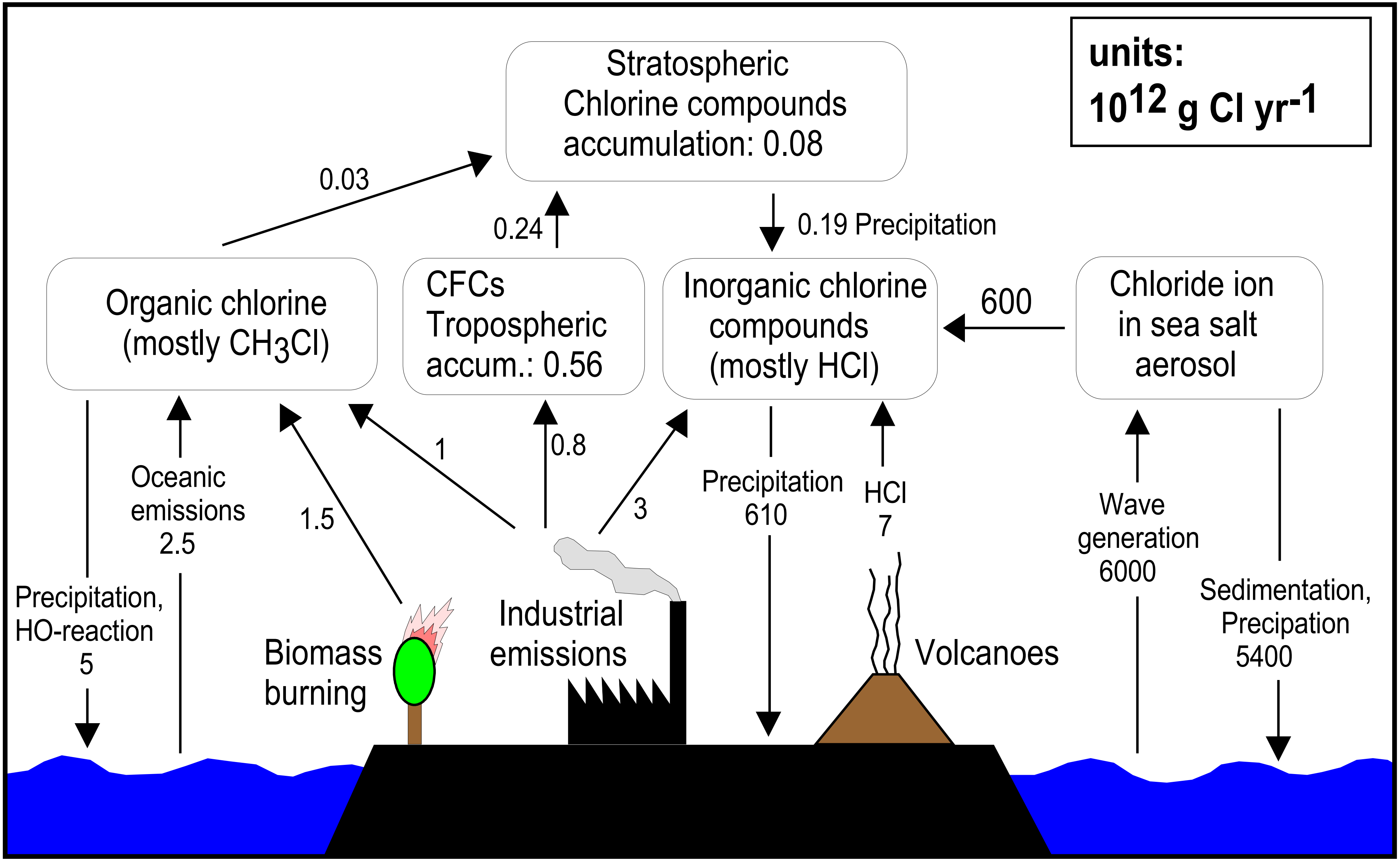
The global sulfur cycle

Figure 5.32. *Schematic diagram of the sulfur cycle in the atmosphere (from Berresheim et al., 1995).*



Brasseur et al., 1999

The Global Atmospheric Chlorine Cycle



units:
10^{12} g Cl yr^{-1}

Stratospheric Chlorine compounds accumulation: 0.08

Graedel and Crutzen [1993]