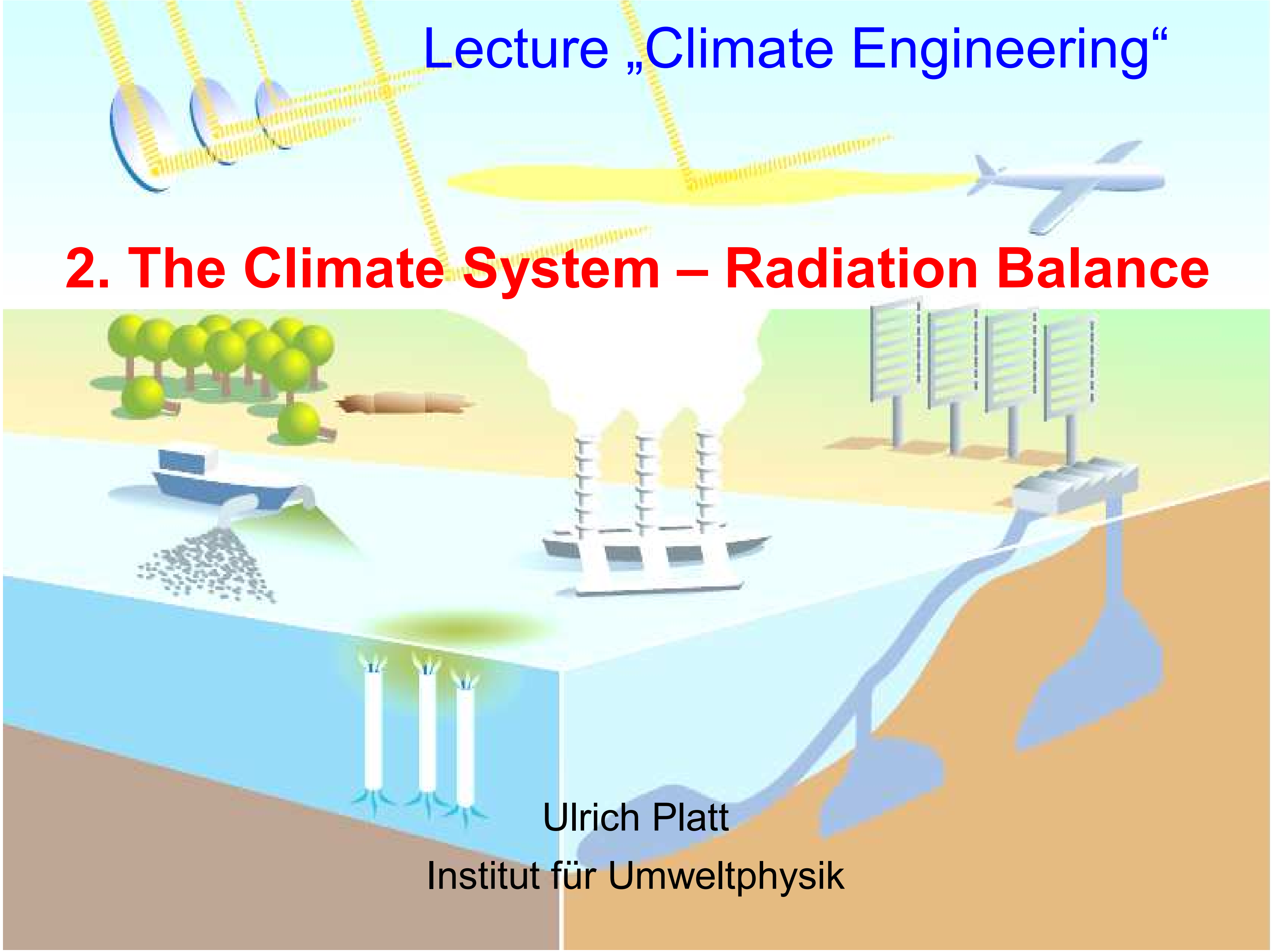


Lecture „Climate Engineering“

2. The Climate System – Radiation Balance

Ulrich Platt

Institut für Umweltphysik

Lecture Program of „Climate Engineering

Part 1: Introduction to the Climate System (4 sessions)

1. Introduction and scope of the lecture
2. The Climate System – Radiation Balance
3. Elements of the Climate System - Greenhouse Gases, Clouds, Aerosol
4. Dynamics of the Climate System - Sensitivity, Predictions

Part 2: Climate Engineering Methods - Solar Radiation Management, SRM

1. SRM – Reflectors in space
2. SRM – Aerosol in the Stratosphere
3. SRM – Cloud Whitening
4. SRM – Anything else

Part 3: Climate Engineering Methods – Carbon Dioxide Removal, CDR

1. Direct CO_2 removal from air
2. Alkalinity to the ocean (enhanced weathering)
3. Ocean fertilization
4. Removal of other greenhouse gases

Part 4: CE – Effectiveness, Side Effects (3 sessions)

1. Comparison of Techniques, characterisation of side effects
2. Other parameters than temperature
3. Summary

Outline for today

1. What is „Climate“?
2. Radiative balance
3. Greenhouse effect
4. Latitudinal energy budget – T
5. Vertical energy budget – T
6. Global energy budget

What is Climate?

Begriff und Aufgabe der Klimatologie. Unter Klima verstehen wir die Gesamtheit der meteorologischen Erscheinungen, welche den mittleren Zustand der Atmosphäre an irgend einer Stelle der Erdoberfläche charakterisieren. Was wir Witterung nennen, ist nur eine Phase, ein einzelner Akt aus der Aufeinanderfolge der Erscheinungen, deren voller, Jahr für Jahr mehr oder minder gleichartiger Ablauf das Klima eines Ortes bildet. Das Klima ist die Gesamtheit der „Witterungen“ eines längeren oder kürzeren Zeitabschnittes, wie sie durchschnittlich zu dieser Zeit des Jahres einzutreten pflegen; wir verstehen also unter einer Darstellung des Klimas die Schilderung des mittleren Zustandes der Atmosphäre.

Julius von Hann, Handbuch der Klimatologie (1883)

- Temperature
- Wind
- Humidity
- Precipitation
- …

Mark Twain: „Climate is what you expect, weather is what you get“

Earth in the Solar System

1368 W/m^2

The Laws of Radiation by Planck, Stephan-Boltzmann, Wien

1. **Planck's** Law:
   Planck Spectrum.
   Radiated power proportional to „Emissivity“ $\varepsilon(\lambda)$

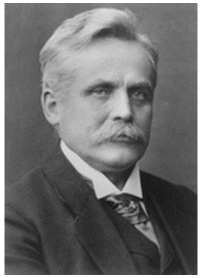
Max Planck

2. **Stefan-Boltzmann**'s Lawe:

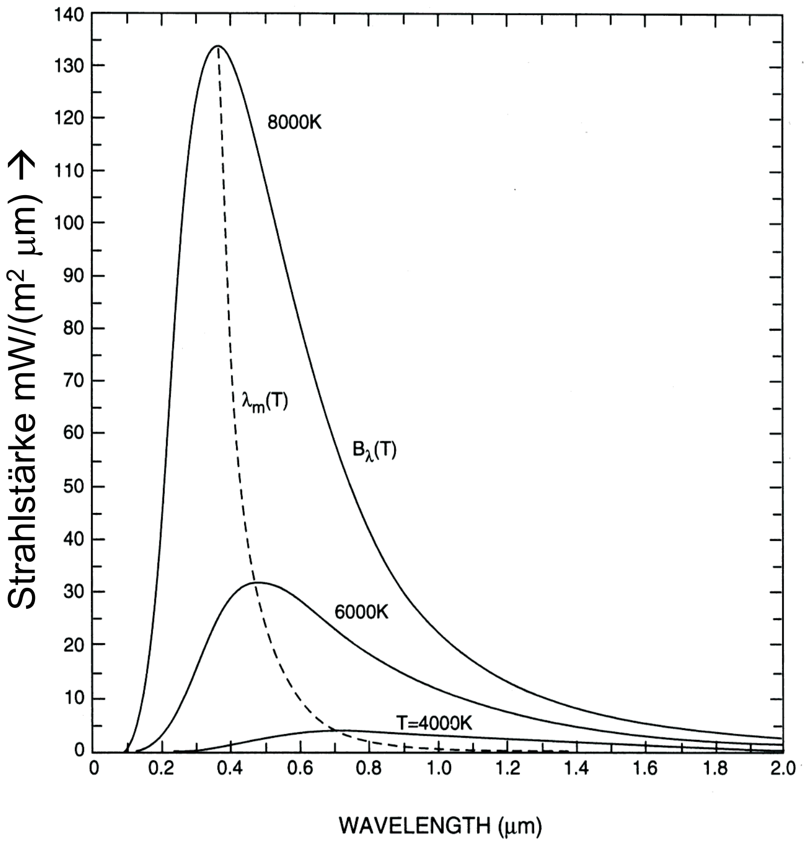
$$P = \varepsilon \sigma_{SB} T^4$$

Joseph Stefan & Ludwig Boltzmann

3. **Wien**'s Displacement law:
   The wavelength of the maximum is inversely proportional to the temperature of the radiating body.



Wilhelm Wien



1 m² bei 20°C:
T= 20+273.2=293.2K

Radiated power:
P=419 W/m²

Properties of Real Radiating Bodies

Real objects are no perfect absorbers/emitters.
Laws holding always:

Energy conservation:

$$\underset{\text{absorptivity}}{\alpha} + \underset{\text{reflectivity}}{\rho} + \underset{\text{transmissivity}}{\tau} = 1$$

Kirchhoff's Law: $\varepsilon_\lambda = \alpha_\lambda$

At any wavelength the emissivity exactly equals the absorptivity

Grey body: $\varepsilon_\lambda = \alpha_\lambda = \text{const.} < 1$

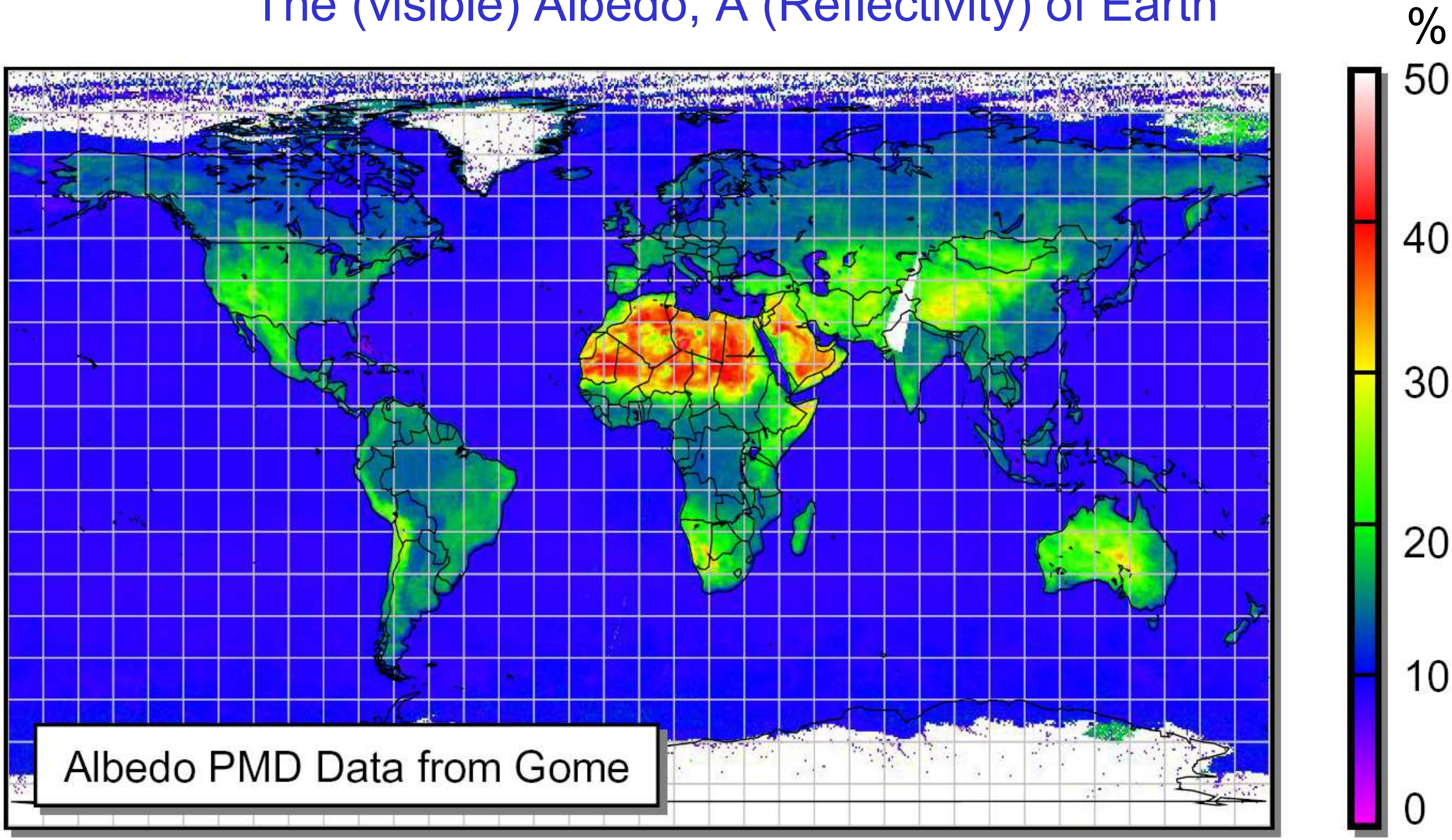
Emissivity < 1, independent of wavelength

Atoms and molecules in atmosphere have complex structure of $\varepsilon_\lambda = \alpha_\lambda$ (spectra with lines and bands)

Gustav Robert Kirchhoff
1824 - 1887

in Heidelberg:
1854 - 1874

The (visible) Albedo, A (Reflectivity) of Earth



Albedo (from albus = white) of the ground (%), Leue 2002
Global average of the cloud-free Earth about 13%
Clouds enhance Albedo to 30%

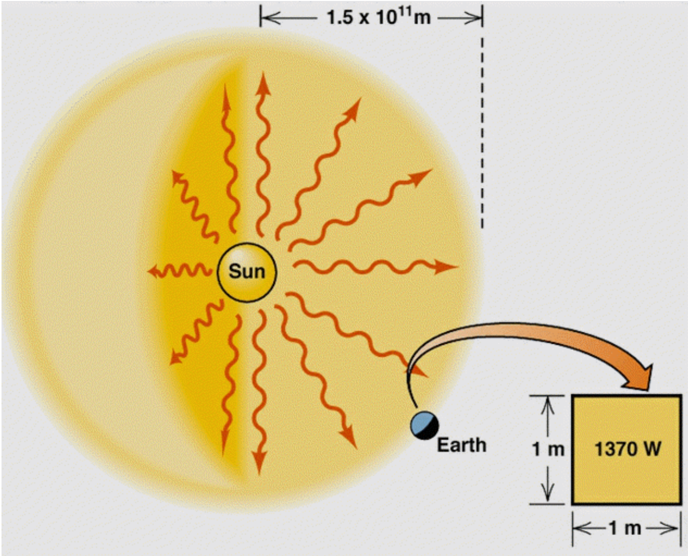
The Climate of Earth (Earth Temperature)

1st Approximation: No Atmosphere

(SW) Power received from the sun on earth:

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

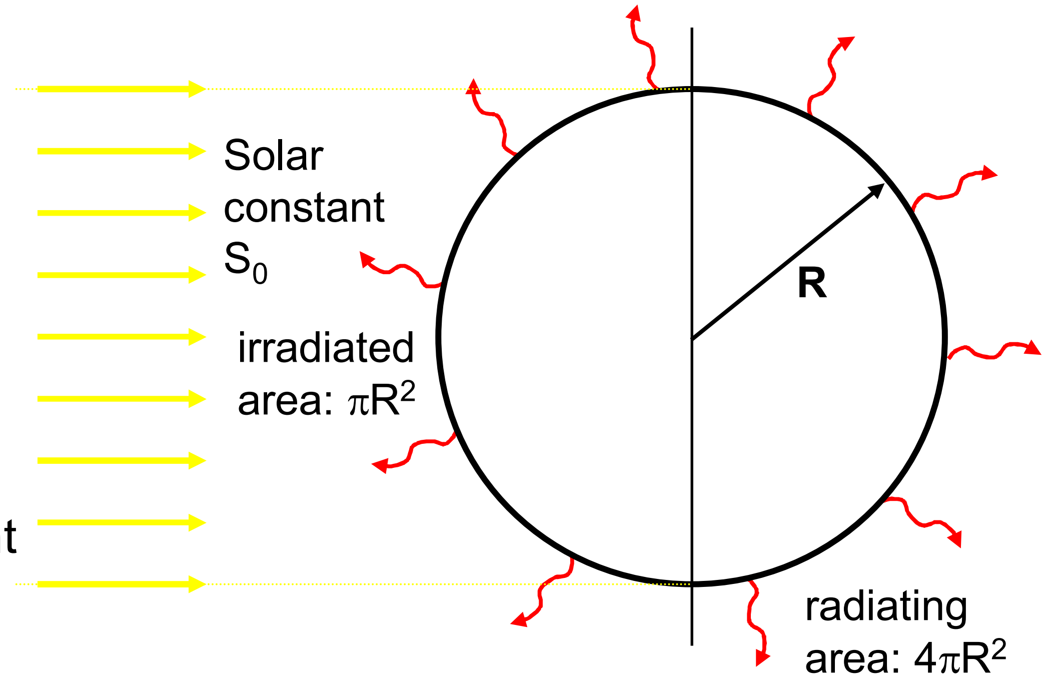
$B_{SC} = S_0$ = 1368 W/m^2 = Solar constant
A ≈ 0.3 = Albedo of earth
1-A = short-wave absorptivity of earth
R_e = 6740 Km = Earth radius



(IR) power radiated from earth:

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

ε ≈ 0.9 ... 1.0
= IR Emissisivity of earth
σ_{SB} = $5.67 \cdot 10^{-8}$ Wm^{-2}K^{-4}
= Stefan-Boltzmann constant



Surface Temperature of Earth. 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:

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

*Note however Earth is presently absorbing 0.58±0.15 W/m² more than it emits: Hansen J., Sato M., Kharecha P., and von Schuckmann K. (2011), Earth's energy imbalance and implications, Atmos. Chem. Phys., 11, 13421–13449, doi:10.5194/acp-11-13421-2011

Substituting for P_{out} and P_{in} we obtain the average of surface temperature of earth T_0 in 0th approximation:

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

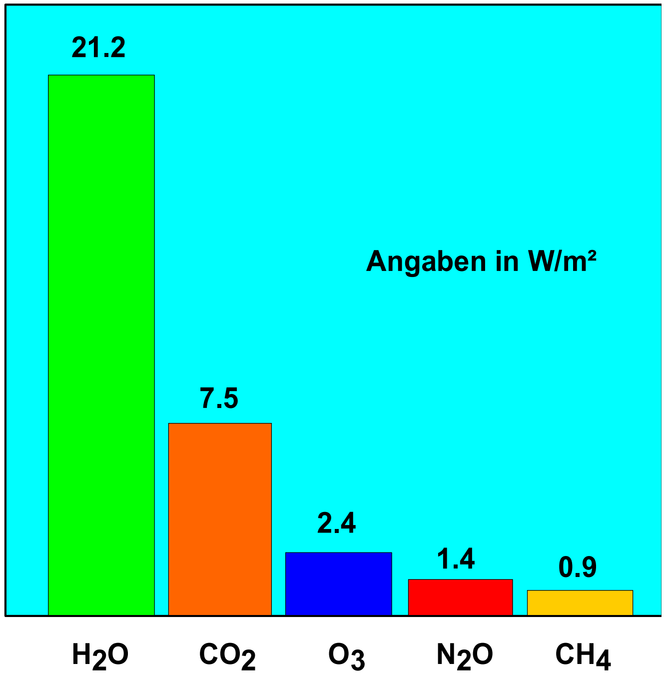
with the above numbers we obtain: $T_0 \approx 255$ K

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

Cause of the discrepancy:

The **natural greenhouse effect** of 33K.

Comment: the main constituents of the atmosphere, N_2 and O_2 provide a negligble greenhouse effect of ≈0.28 W/m² (global mean), see: Höpfner M., Milz M., Buehler S., Orphal J., and Stiller G. (2012), The natural greenhouse effect of atmospheric oxygen (O_2) and nitrogen (N_2), Geophys. Res. Lett. 39, L10706, doi:10.1029/2012GL051409.



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

Climate Engineering Ideas:

Cool Earth by:

1) Reducing the „Solar Constant“S_0

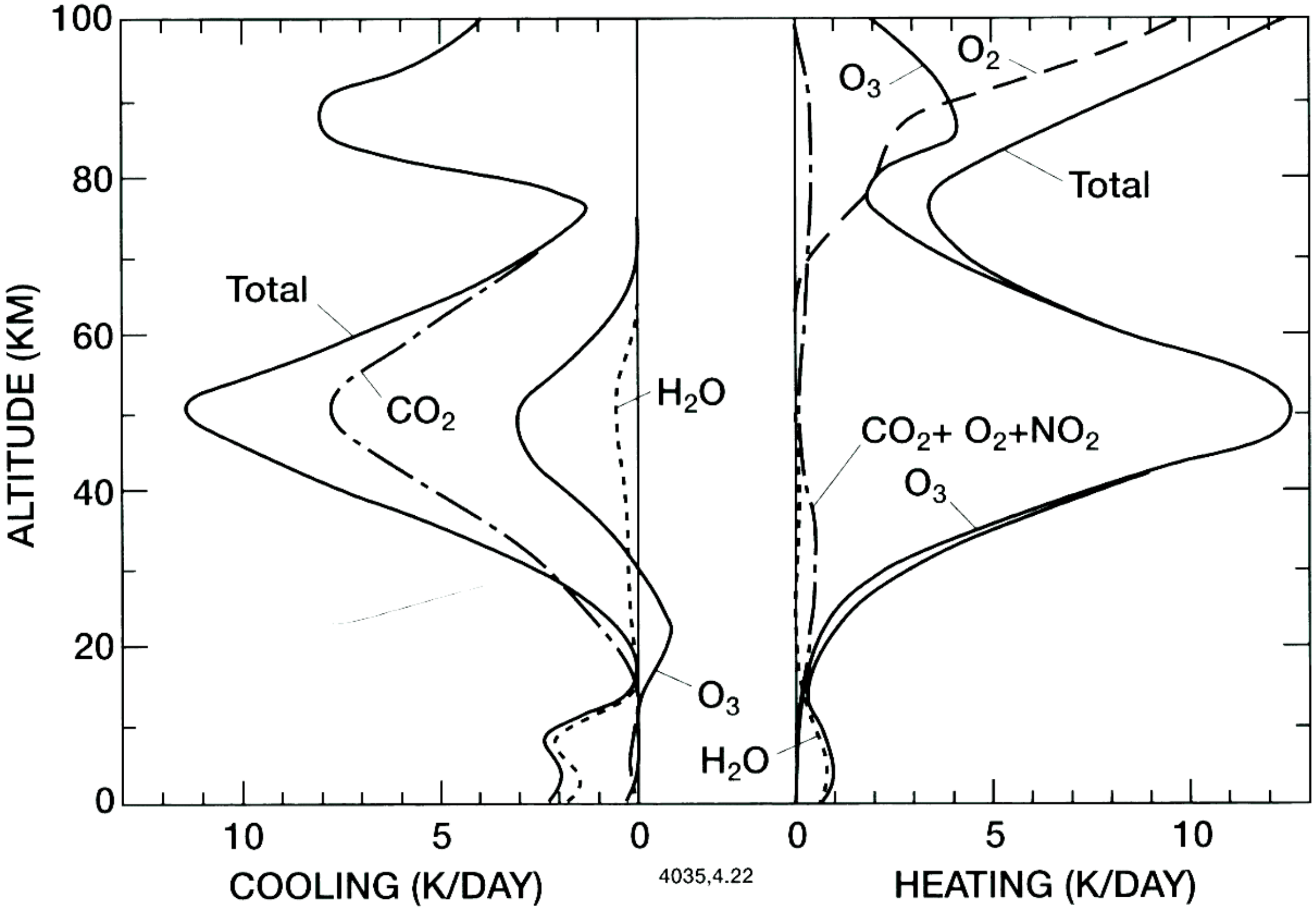
= „Solar Radiation Management

2) Increasing the Albedo A

Atmospheric Energy Balance

local heating rates:

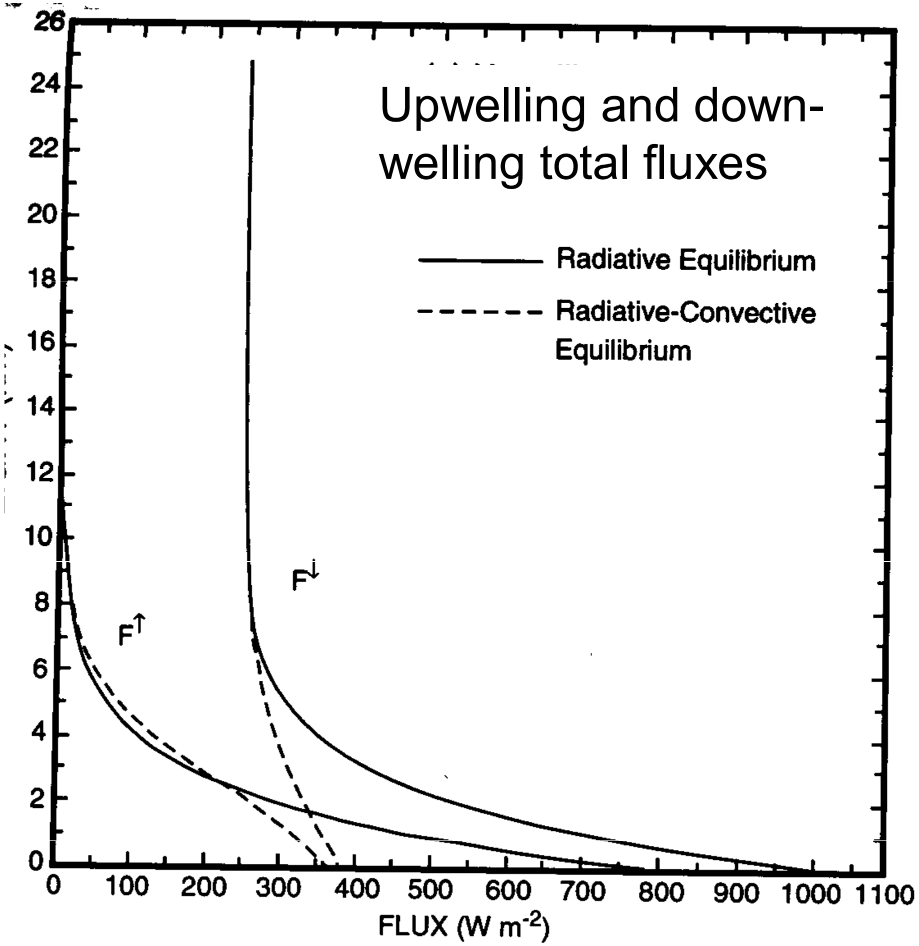
– stratosphere ~ radiative equilibrium
– troposphere ≠ radiative equilibrium - “convective adjustment”



Brasseur and Solomon, 2005 (IUP-Book 1968)

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

Radiative Equilibrium –
Radiative-Convective Equilibrium



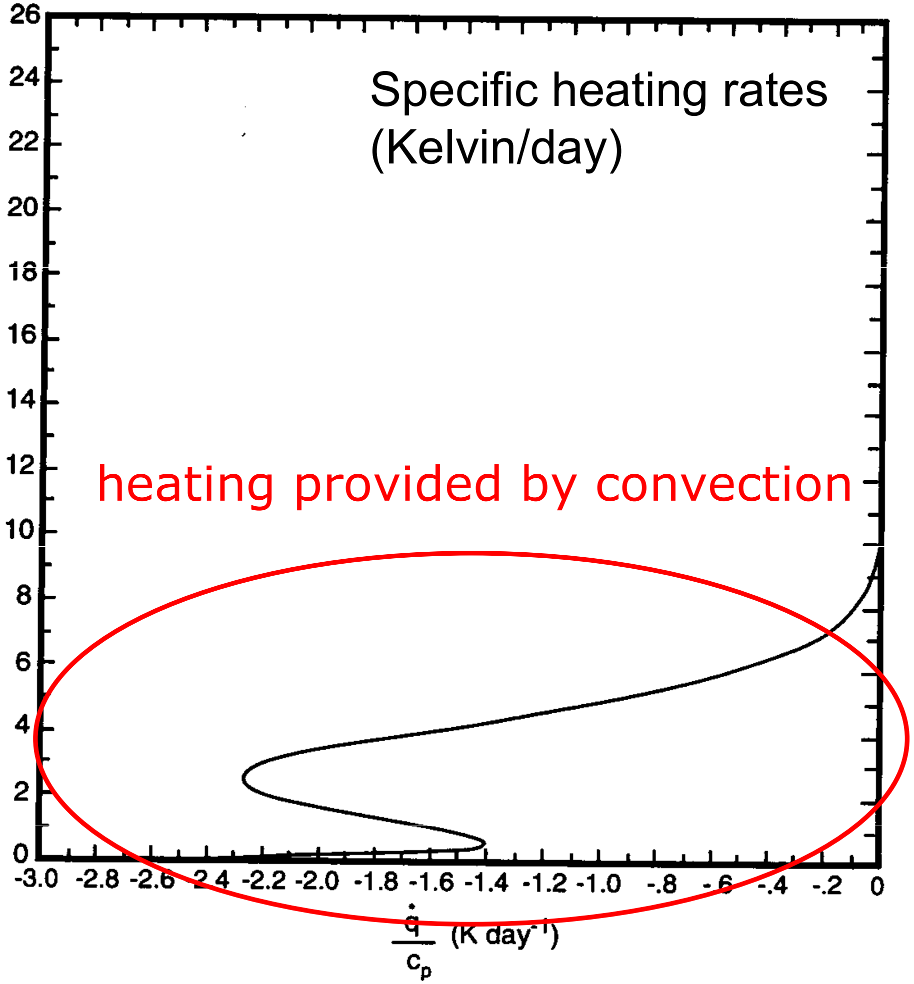
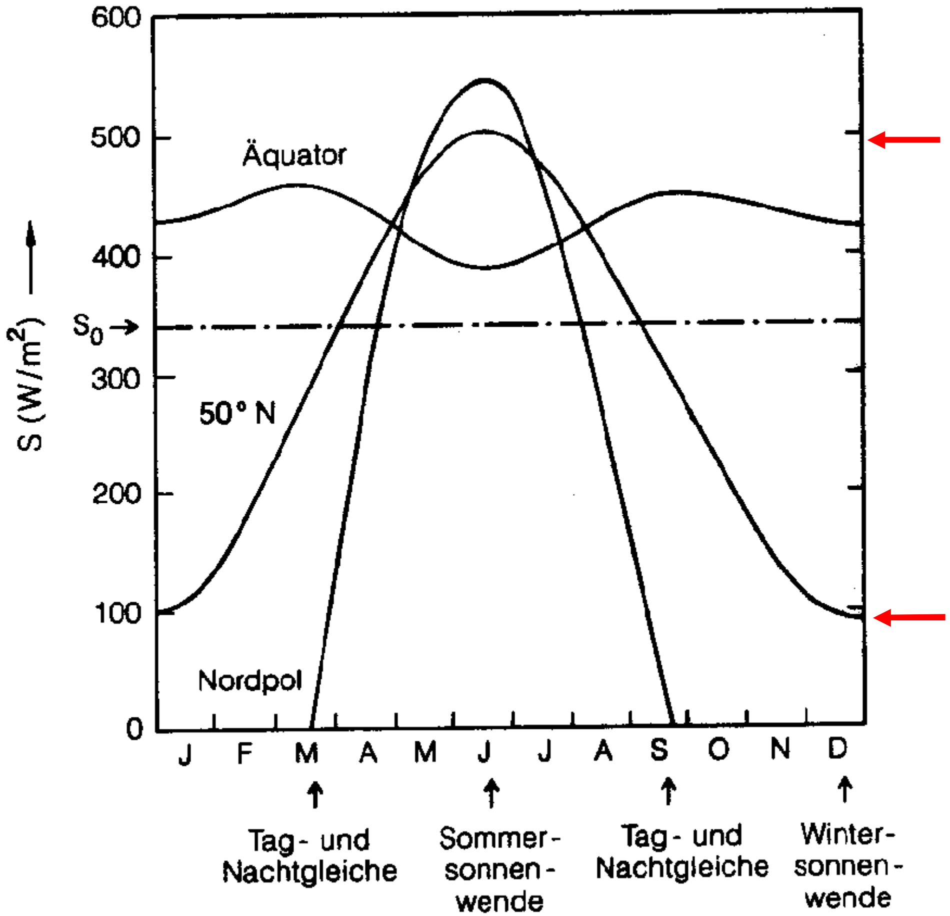


Figure 8.22 (a) Upwelling and downwelling fluxes as functions of height in the gray atmosphere in Fig. 8.20 for radiative equilibrium (solid lines) and radiative–convective equilibrium (dashed lines). (b) Specific heating rate under radiative–convective equilibrium.

Salby, 1996

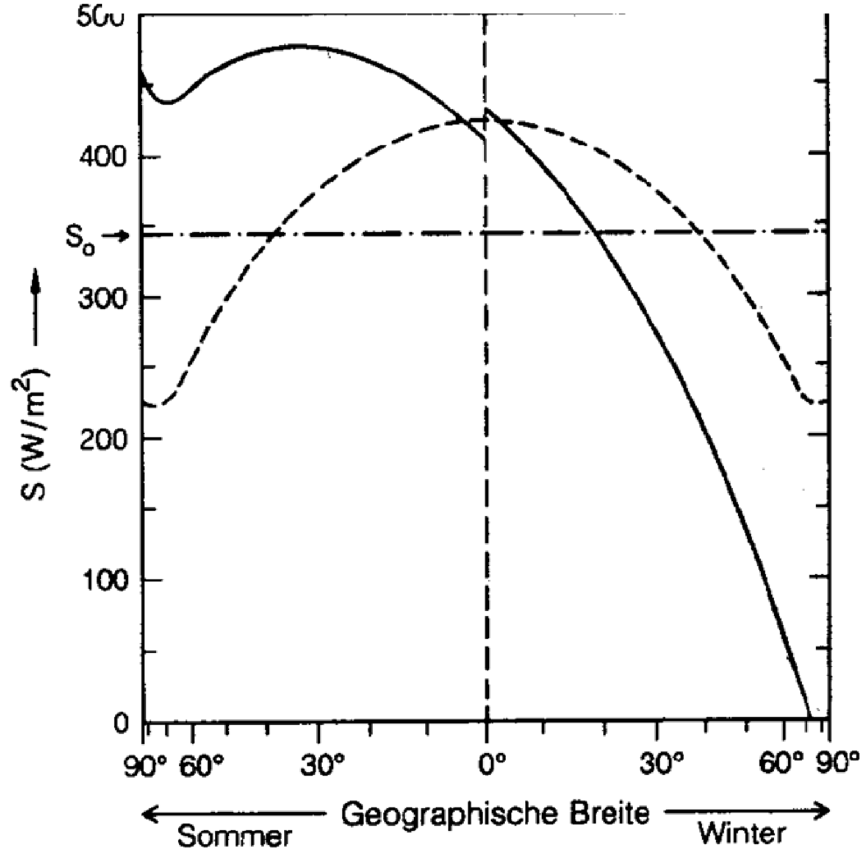
Latitudinal Insolation of Earth

24-hour average for
- Equator
- 50°N
- North pole



—— Mittlere Einstrahlung für Sommer- und Winterhalbjahr.

- - - - Jahresmittel der globalen Einstrahlung als Funktion der Breite



$$T_0 = \sqrt[4]{\frac{S_{Sommer}(50^\circ)}{S_{winter}(50^\circ)}} \approx \sqrt[4]{5} \approx 1.5$$

aus: W. Roedel, 2000

Latitudinal Energy Budget of Earth

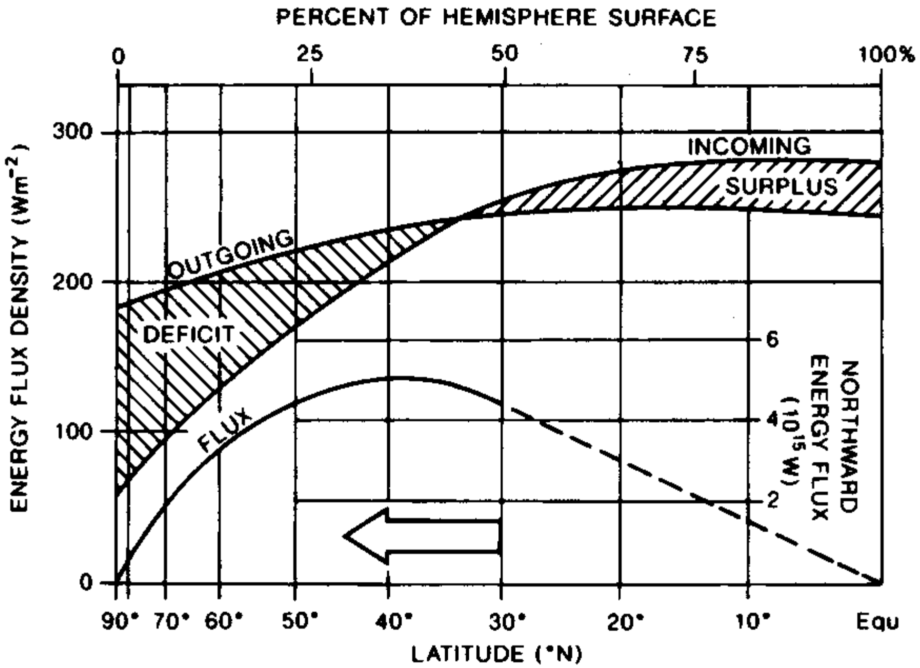
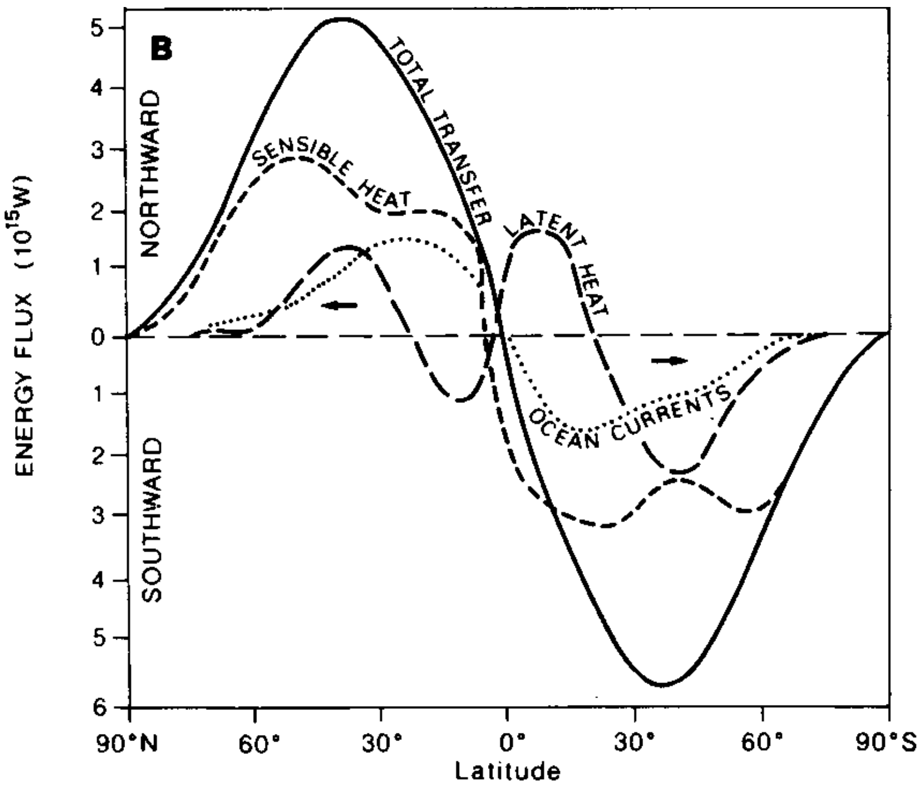


Figure 2.26 A meridional illustration of the balance between incoming solar radiation and outgoing radiation from the earth and atmosphere* in which the zones of permanent surplus and deficit are maintained in equilibrium by a poleward energy transfer.†

Sources: *Data from Houghton; after Newell 1964.

†After Gabites.

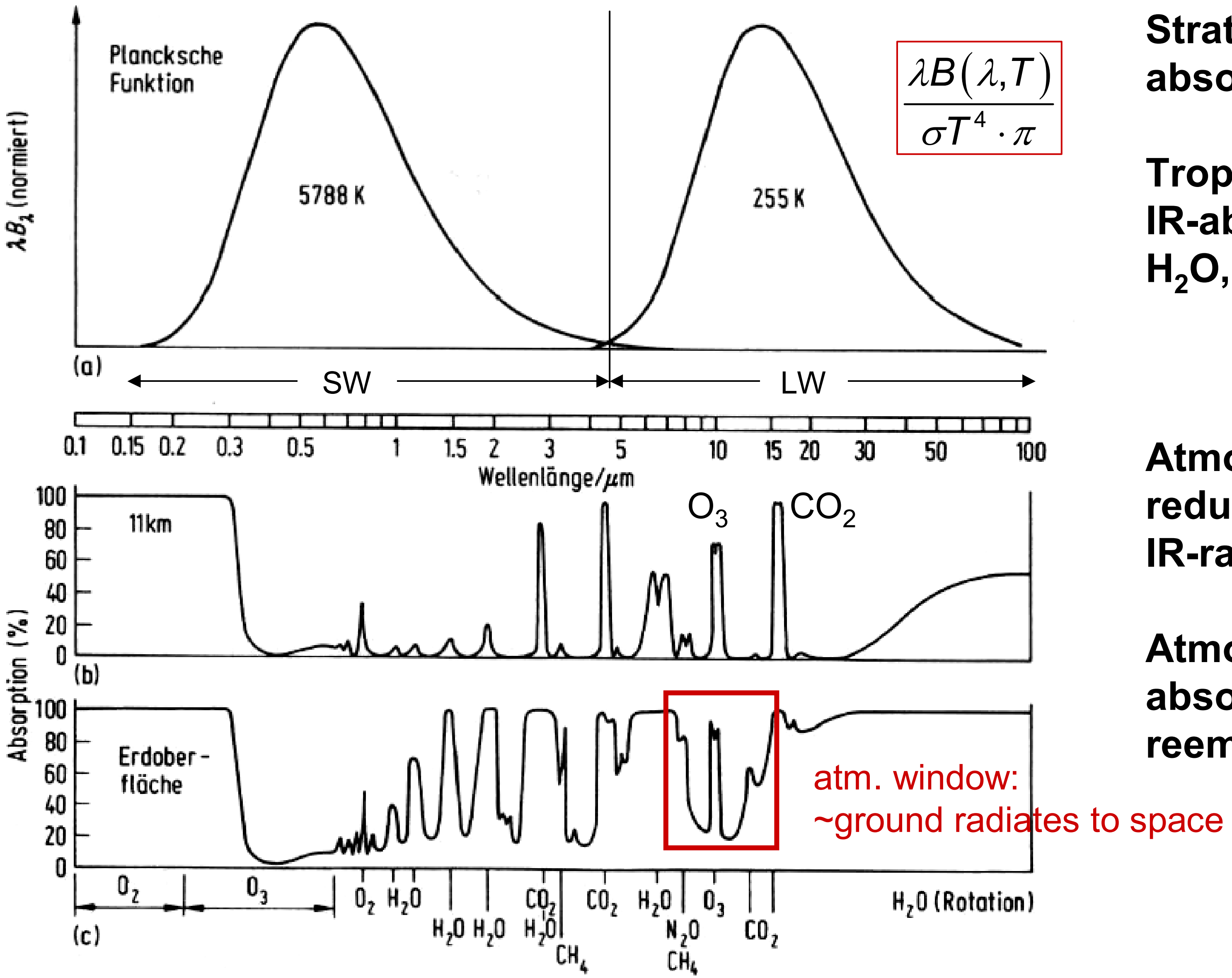


(B) The average annual latitudinal distribution of the components of the poleward energy transfer (in 10^{15}W) in the earth–atmosphere system.

Source: From Sellers 1965.

Barry and Chorley, 1998

Emission Spectra of Earth and Sun



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

The Greenhouse Effect – Svante Arrhenius

Svante Arrhenius
1859-1927

THE

LONDON, EDINBURGH, AND DUBLIN

PHILOSOPHICAL MAGAZINE

AND

JOURNAL OF SCIENCE.

[FIFTH SERIES.]

APRIL 1896.

XXXI. *On the Influence of Carbonic Acid in the Air upon the Temperature of the Ground. By* Prof. SVANTE ARRHENIUS *.

I. *Introduction : Observations of* Langley *on Atmospherical Absorption.*

A GREAT deal has been written on the influence of the absorption of the atmosphere upon the climate. Tyndall † in particular has pointed out the enormous importance of this question. To him it was chiefly the diurnal and annual variations of the temperature that were lessened by this circumstance. Another side of the question, that has long attracted the attention of physicists, is this : Is the mean temperature of the ground in any way influenced by the presence of heat-absorbing gases in the atmosphere? Fourier‡ maintained that the atmosphere acts like the glass of a hot-house, because it lets through the light rays of the sun but retains the dark rays from the ground. This idea was elaborated by Pouillet § ; and Langley was by some of his researches led to the view, that "the temperature of the earth under direct sunshine, even though our atmosphere were present as now, would probably fall to −200° C., if that atmosphere did not possess the quality of selective

* Extract from a paper presented to the Royal Swedish Academy of Sciences, 11th December, 1895. Communicated by the Author.

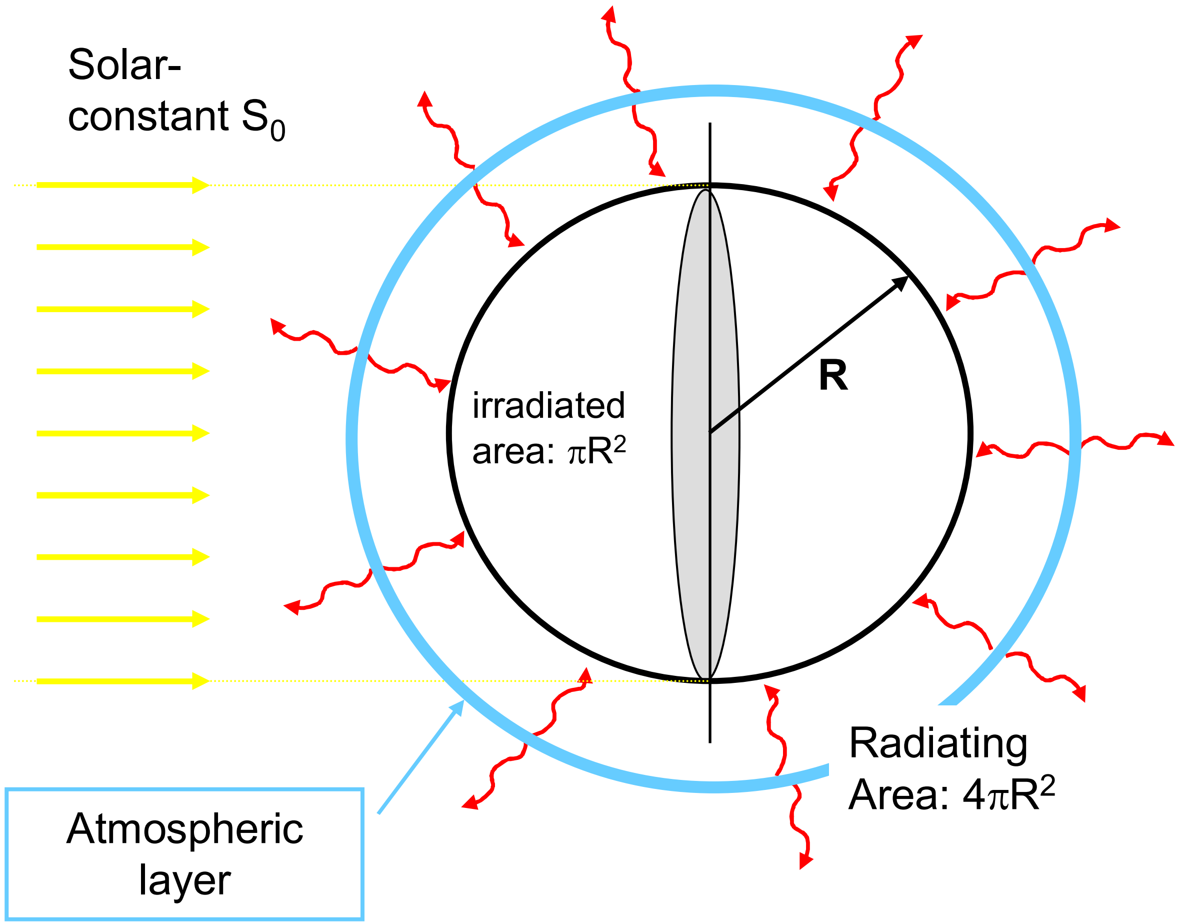
† 'Heat a Mode of Motion,' 2nd ed. p. 405 (Lond., 1865).

‡ *Mém. de l'Ac. R. d. Sci. de l'Inst. de France*, t. vii. 1827.

§ *Comptes rendus*, t. vii. p. 41 (1838).

Surface Temperature of Earth. 2nd Approximation

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

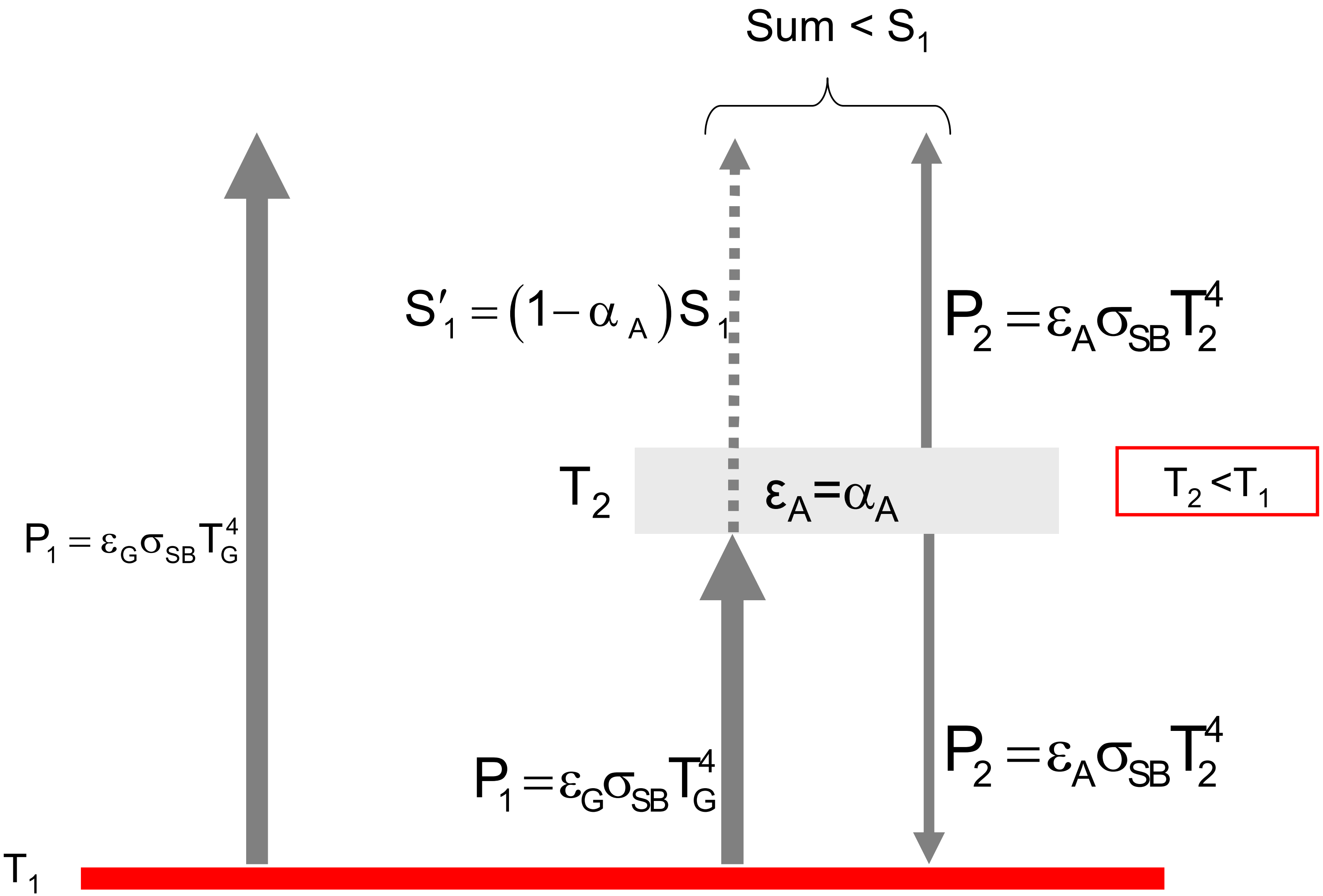
Atmosphere:

Temperature T_A,

IR-Absorptivity = Emissivity ε_A,

Completely transparent for short wave radiation

The Effect of Absorbing Layers



Surface Temperature of Earth. 2nd Approximation

Radiation equilibrium for both layers:

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

$\varepsilon_A = \alpha_A$

Atm. radiates upwards and downwards!

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

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

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

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

Note: Effektive emission from higher, colder layer!

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

Gas Clouds also follow Kirchhoff's Law

$$\alpha + \rho + \tau = 1$$

absorptivity reflectivity transmissivity

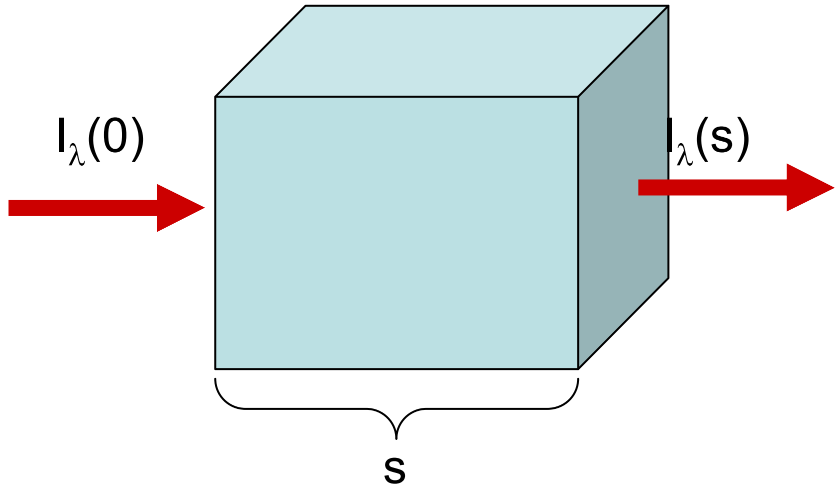
Kirchhoff's Law: $\varepsilon_\lambda = \alpha_\lambda$

At any wavelength the emissivity exactly equals the absorptivity

Absorptivity: Lambert-Beer's Law: $I_\lambda(s) = I_\lambda(0) \cdot e^{-k_{a,\lambda} \cdot s} = I_\lambda(0) \cdot e^{-\sigma_{a,\lambda} \cdot c \cdot s}$

$$\alpha = \frac{I_\lambda(s)}{I_\lambda(0)} = e^{-k_{a,\lambda} \cdot s}$$

Reflectivity: Thermal-Re-Emission + Back-Scattering



Transmissivity: Not absorbed (1-α) + Thermal-Re-Emission (backward direction) + Forward-Scattering

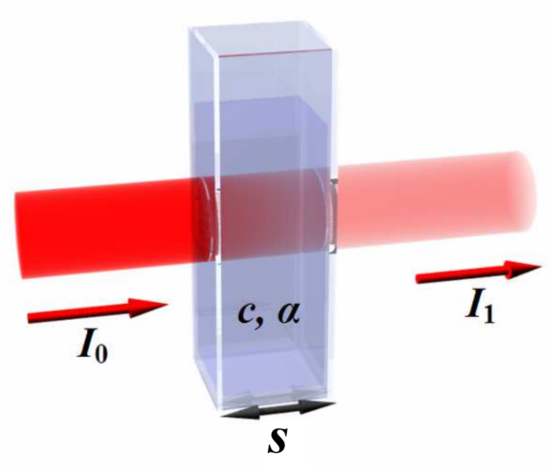
Absorption: Beer-Lambert Law

a) constant $k_{a,\lambda}$:

$$I_\lambda(s) = I_\lambda(0) \cdot e^{-k_{a,\lambda} \cdot s} = I_\lambda(0) \cdot e^{-\sigma_{a,\lambda} \cdot c \cdot s}$$

b) general:

$$I_\lambda(s) = I_\lambda(0) \cdot e^{-\int_0^s k_{a,\lambda} \cdot ds} = I_\lambda(0) \cdot e^{-\tau_\lambda(s)}$$



with the optical depth (density) τ [-]:

$$\tau_\lambda(s) = \int_0^s k_{a,\lambda} ds$$

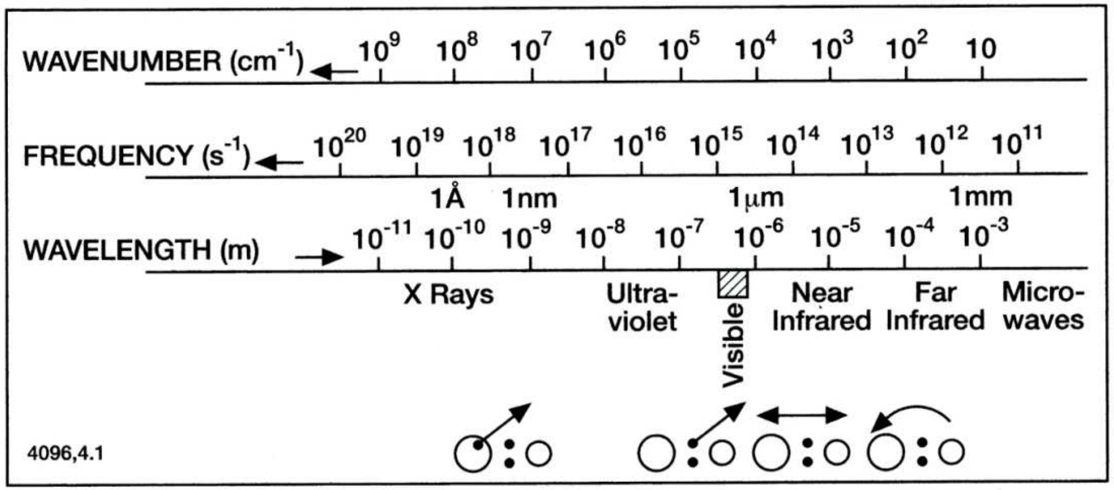
Discovered by:
Pierre Bouguer in 1729
Johann H. Lambert in 1760
August Beer in 1852

Transmissivity: $$T_\lambda(s) = \frac{I_\lambda(s)}{I_\lambda(0)} = e^{-\int_0^s k_{a,\lambda} ds} = e^{-\int_0^s \sigma_{a,\lambda} c ds} = e^{-\tau_\lambda(s)}$$

Optical depth: $$\tau_\lambda(s) = -\ln T_\lambda(s)$$

Line Absorption

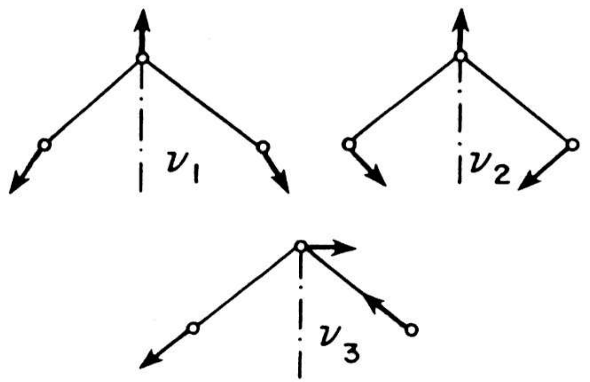
- absorption spectrum of gas consists of
  - continuum absorption (X-ray, short UV)
    - photoionization
    - photodissociation
  - complex arrays of lines corresponding to energy levels of:
    - discrete electronic - UV
    - vibrational
    - rotational



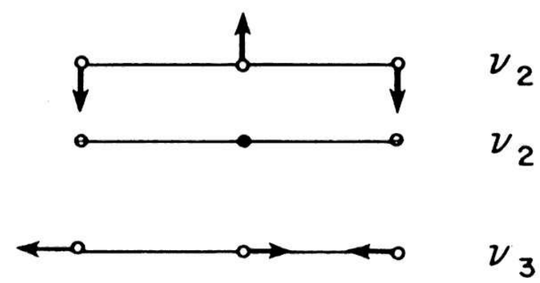
Brasseur and Solomon, 2005

Figure 4.1. Spectral regions and their effect on molecules: from left to right: ionization, dissociation, vibration, and rotation.

Molecular Bands



Normal modes of vibration for bent molecules, e.g. H_2O and O_3



Normal modes of vibration for linear molecules, e.g. CO_2 or N_2O

(Herzberg 1945)

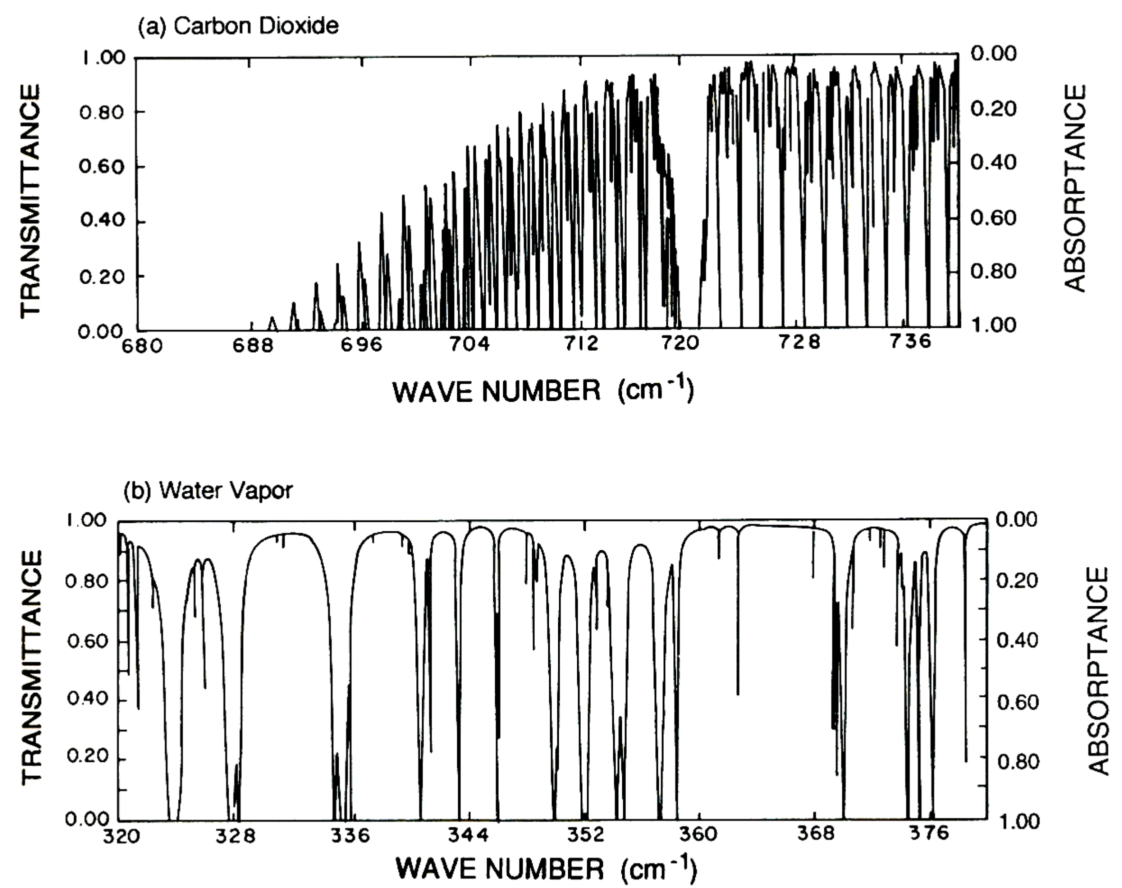


Figure 8.11 Absorption spectra in (a) the 15-μm band of CO_2 and (b) the rotational band of H_2O at 27 to 31 μm. Adapted from McClatchey and Selby (1972).

Salby, 1996

Line Broadening

Several effects lead to a broadening of spectral lines

- spectral width of line: $\sigma_{a\nu} = Sf(\nu - \nu_0) = f(\nu - \nu_0)\int \sigma_\nu d\nu$
  with S line strength and f shape factor (ν_0 – line center)

- natural line broadening due to finite lifetime of excited state:

  Lorentz shape: $$f_n(\nu - \nu_0) = \frac{\alpha_n}{\pi(\nu - \nu_0)^2 + \alpha_n^{\ 2}} \quad \text{with} \quad \alpha_n = \frac{1}{2\pi\tau}$$

- Doppler broadening due to molecular motion in line of sight:

  Gaussian shape: $$f_D(\nu - \nu_0) = \frac{1}{\alpha_D\sqrt{\pi}}\exp\left(-\frac{(\nu - \nu_0)^2}{\alpha_D^2}\right) \quad \text{with} \quad \alpha_D = \frac{\nu_0}{c}\sqrt{\frac{2kT}{m}}$$

- Pressure broadening due to collisions of molecules (→ lifetime reduction):

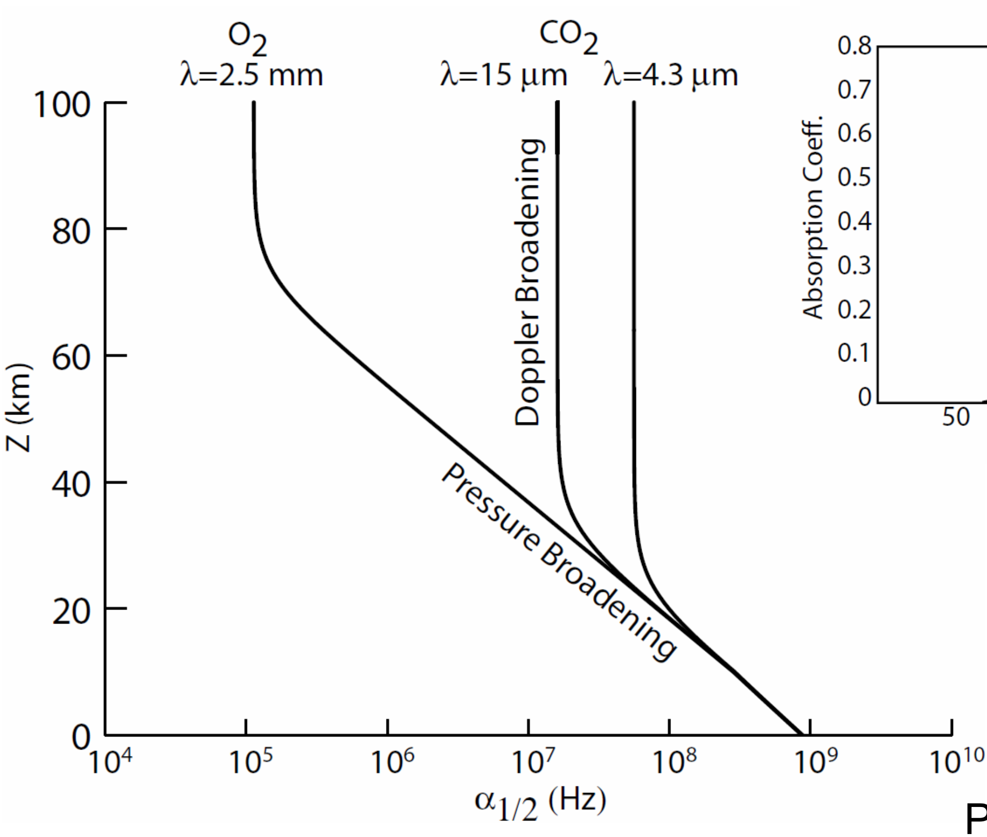
  Lorentz shape with $$\alpha_c = \alpha_0\left(\frac{p}{p_0}\right)\left(\frac{T_0}{T}\right)^{\frac{1}{2}}$$ α_0: half-width at standard T_0, p_0

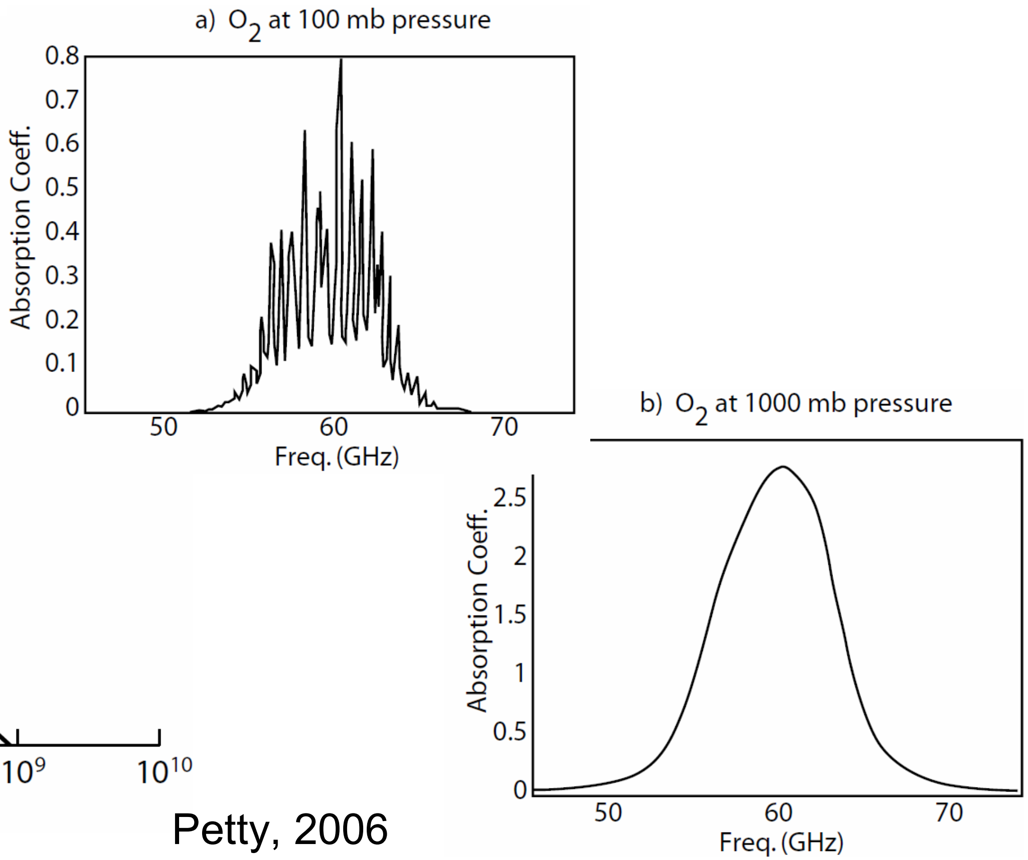
- Note: Pressure broadening: $\Delta\nu$ independent of ν, Doppler: $\Delta\nu \propto \nu$
  Thus pressure broadening dominates at small ν (long wavelengths, IR)
  Doppler broadening dominates at high ν (short wavelengths, UV)

Line Broadening – Band Models

- below 30 km: pressure broadening is dominant for IR
- above 30 km: natural and Doppler broadening important for VIS and UV
- complexity of lines make line-by-line calculation impractical for most applications

  → band models (e.g. assumption of random distribution of lines)





Petty, 2006

Taking into Account the Atmospheric IR-Albedo

At the top of the atmosphere we have:
(short-wave) incoming solar radiation= (long wave) IR outgoing radiation

$S_o \cdot (1 - A) / 4 = \sigma_{SB} \cdot T_s^4 \cdot (1 - B)$ $[(1\text{-}B)=\varepsilon_G]$

$$T_S = \sqrt[4]{S_0 \cdot \frac{(1-A)}{4(1-B)\sigma_{SB}}}$$

$$T_S = \left(S_0 \cdot \underbrace{\frac{(1-A)}{4(1-B)\sigma_{SB}}}_{=K} \right)^{0.25}$$

T_s = 255 K for B = 0 or = 287.5 K for B = 0.40, resp.

$$\frac{dT_S}{dS_0} = \frac{1}{4} S_0^{-0.75} \cdot K^{0.25}$$

$$= \frac{1}{4} S_0^{-1} \underbrace{S_0^{0.25} \cdot K^{0.25}}_{T_S}$$

$$= \frac{1}{4} \frac{T_S}{S_0}$$

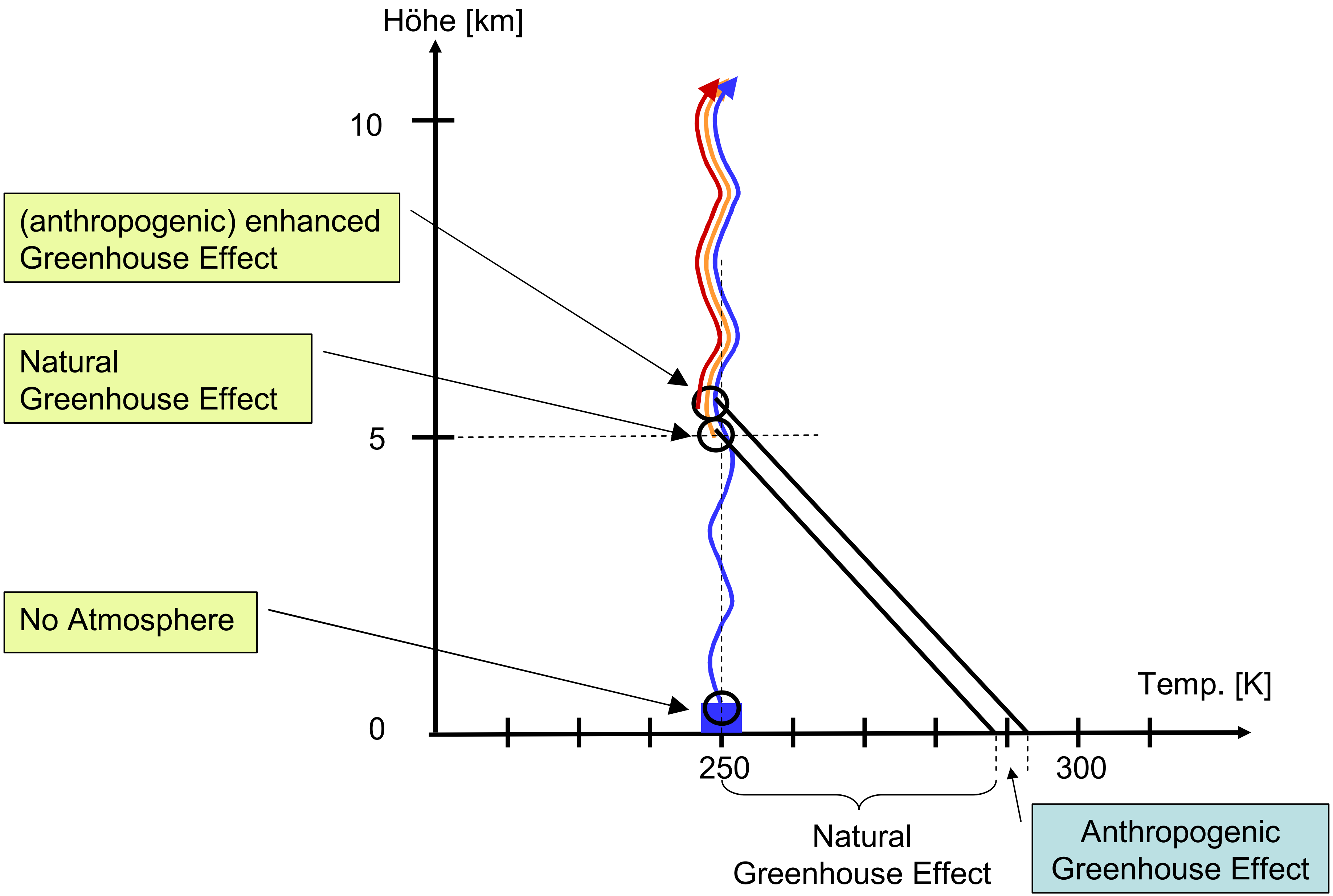
A: Earth albedo in the visible (SW) Spectral range (= 0.298)
T_s: Earth surface temperature
B: IR albedo (= 0.40)
S_o: Solar constant (~ 1370 W/m^2)
σ_{SB}: Stefan Boltzmann constant ($5.87 \cdot 10^{-8}$ W/(m$^4 \cdot$K)

Climate sensitivities (no feedback):

$\partial T_s / \partial S_0 = T_s / 4 \cdot S_o$ = 0.05 K·m^2/W (cf. ~ 0.1 K·m^2/W from GCM's)
$\approx$ 0.70 K / 1% change in S_0
$\partial T_s / \partial A = - T_s / \{4 \cdot (1-A)\}$ = 0.37 K / 1% change in A (e.g. clouds, ice cover, aerosol, …)
$\partial T_s / \partial B = T_s / \{4 \cdot (1-B)\}$ = 0.51 K / 1% change in B (e.g. greenhouse gases, cirrus, …)

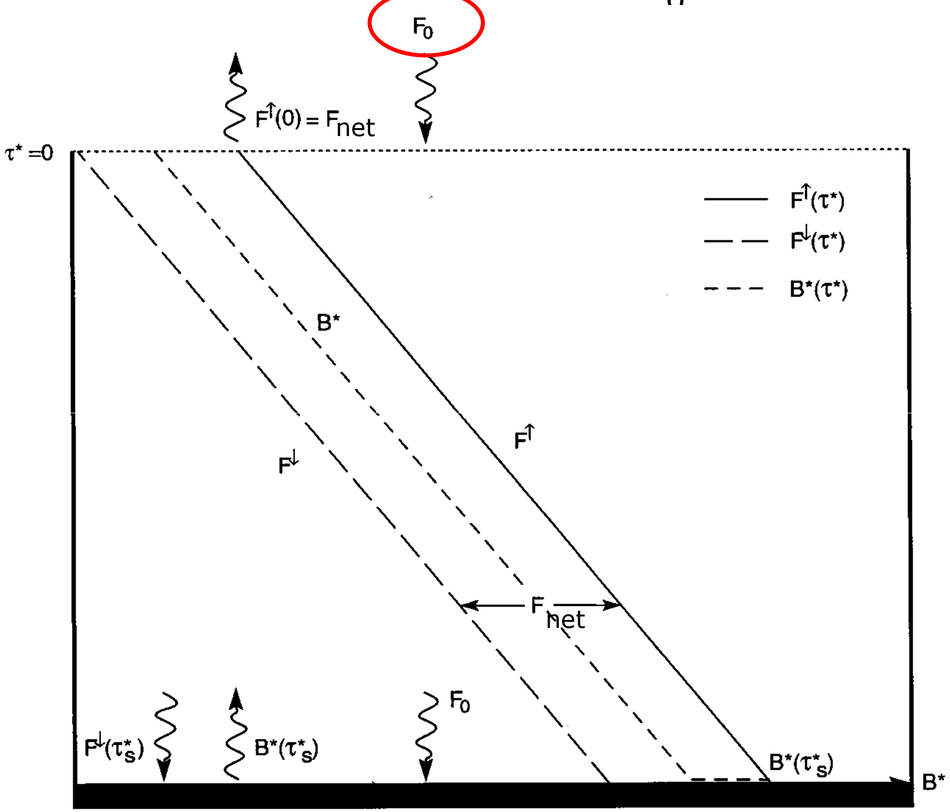
Atmospheric Temperature Profile

Greenhouse Effect:
Effective Emission height of the LW – Radiation



Radiative equilibrium at TOA and surface

solar SW: F_0



Upwelling and downwelling fluxes and emission in a grey atmosphere that is in radiative equilibrium with an incident SW flux F_0 and a black underlying surface. Atmosphere is assumed to be transparent to SW. Note: the emission profile is discontinuous at the surface.

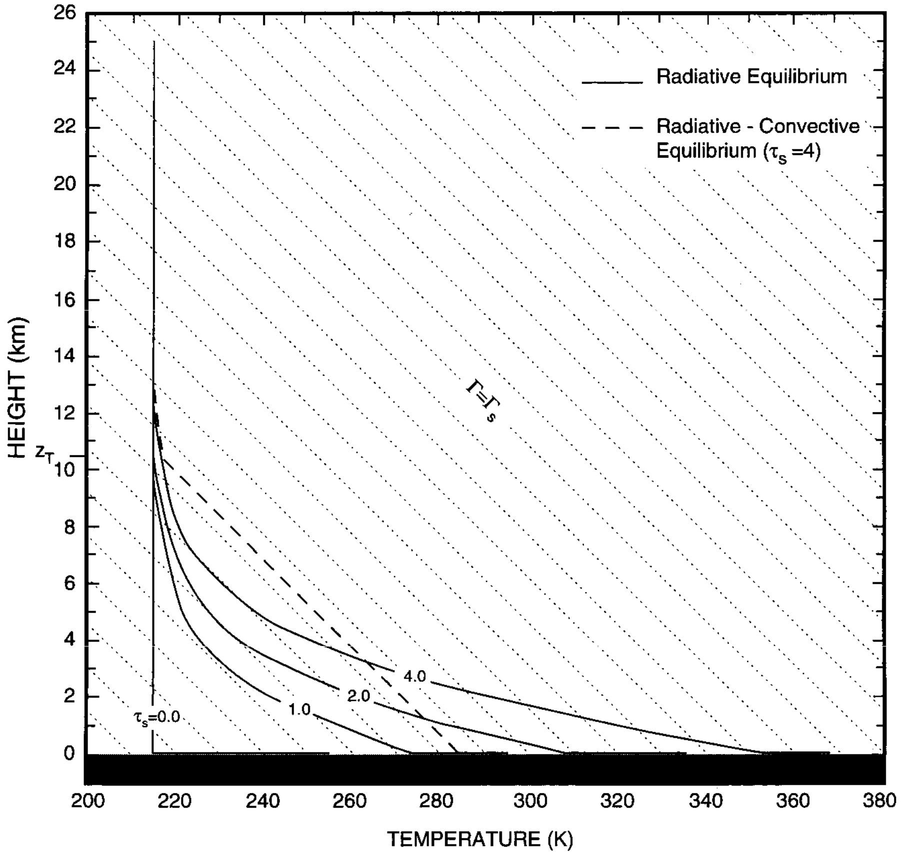
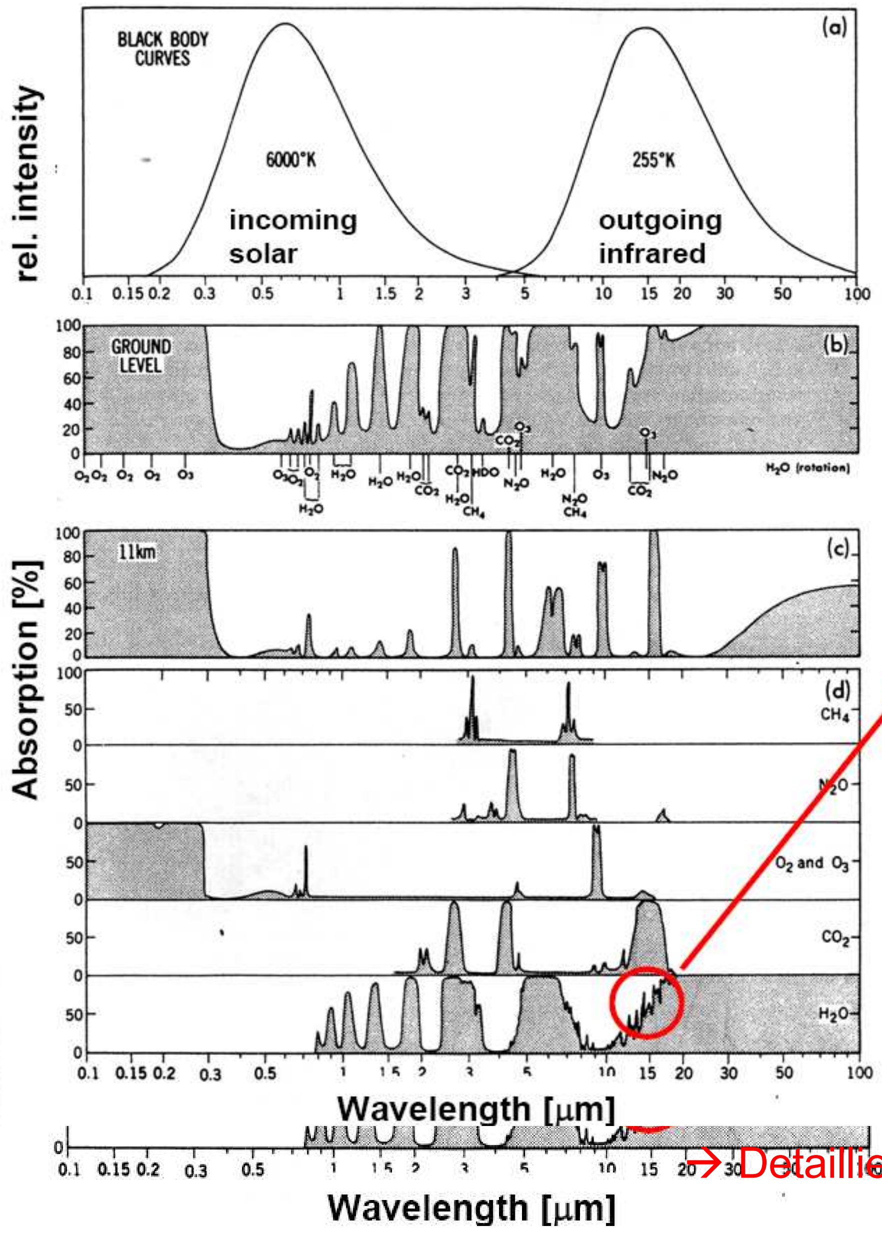


Figure 8.21 Radiative equilibrium temperature (solid lines) for the gray atmosphere in Fig. 8.20, with a profile of optical depth representative of water vapor (8.69), presented for several atmospheric optical depths τ_s. Saturated adiabatic lapse rate (dotted lines) and radiative–convective equilibrium temperature for $\tau_s = 4$ (dashed line) superposed.

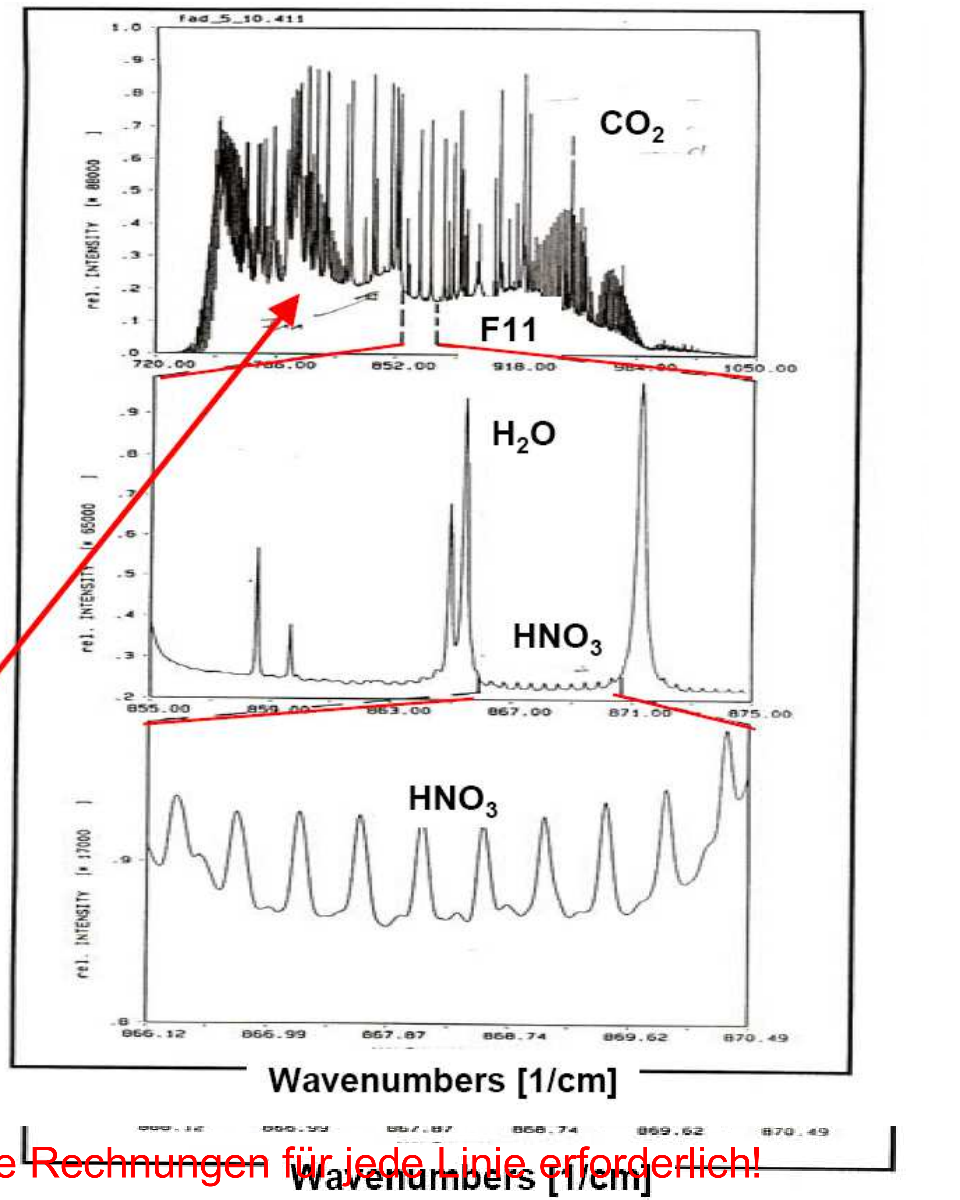
Salby, 1996

Surface Temperature of Earth. 3rd Approximation

Überblick, gesamter Wellenlängenbereich

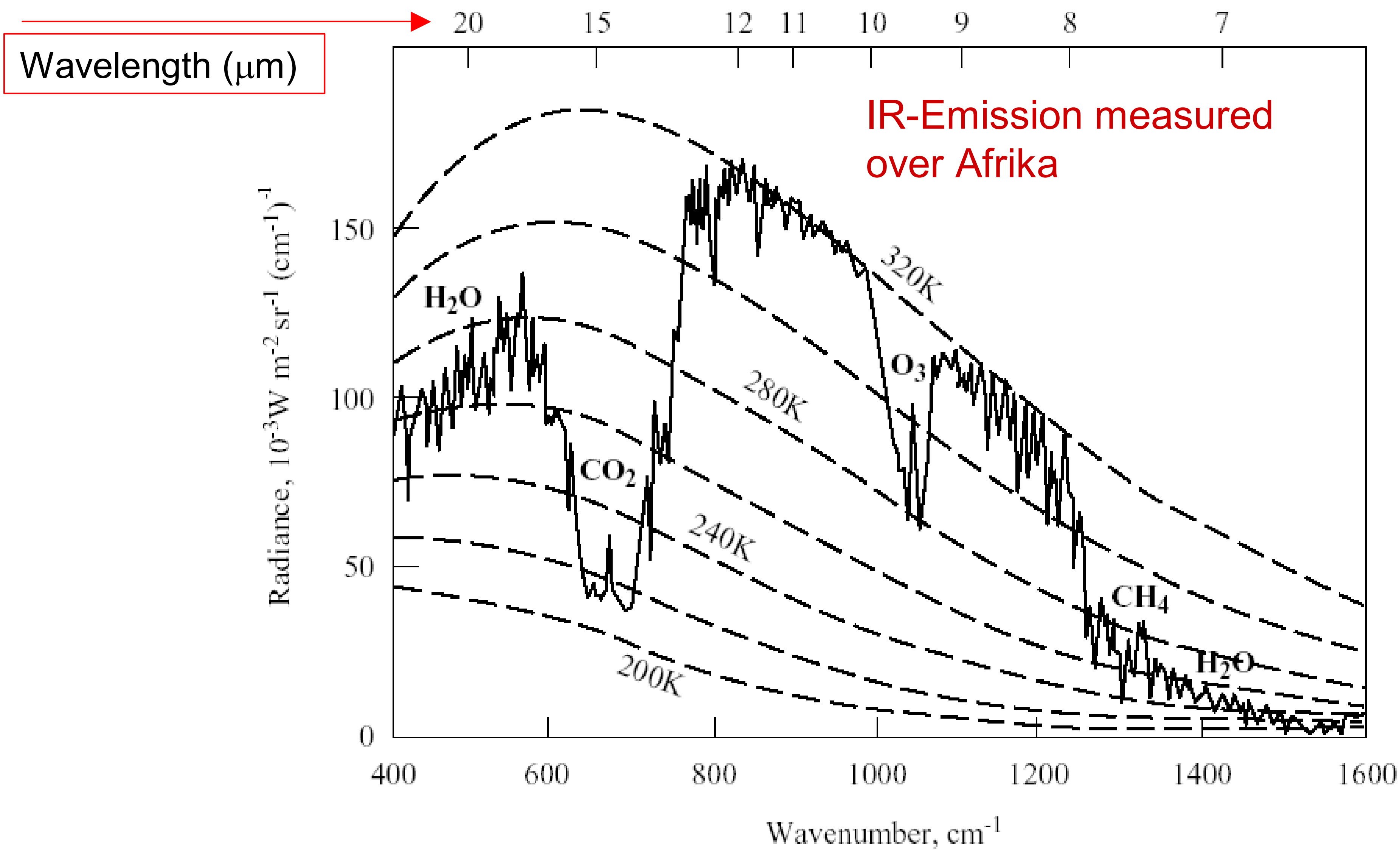


Zoom in den Bereich 886 – 870 cm^{-1}

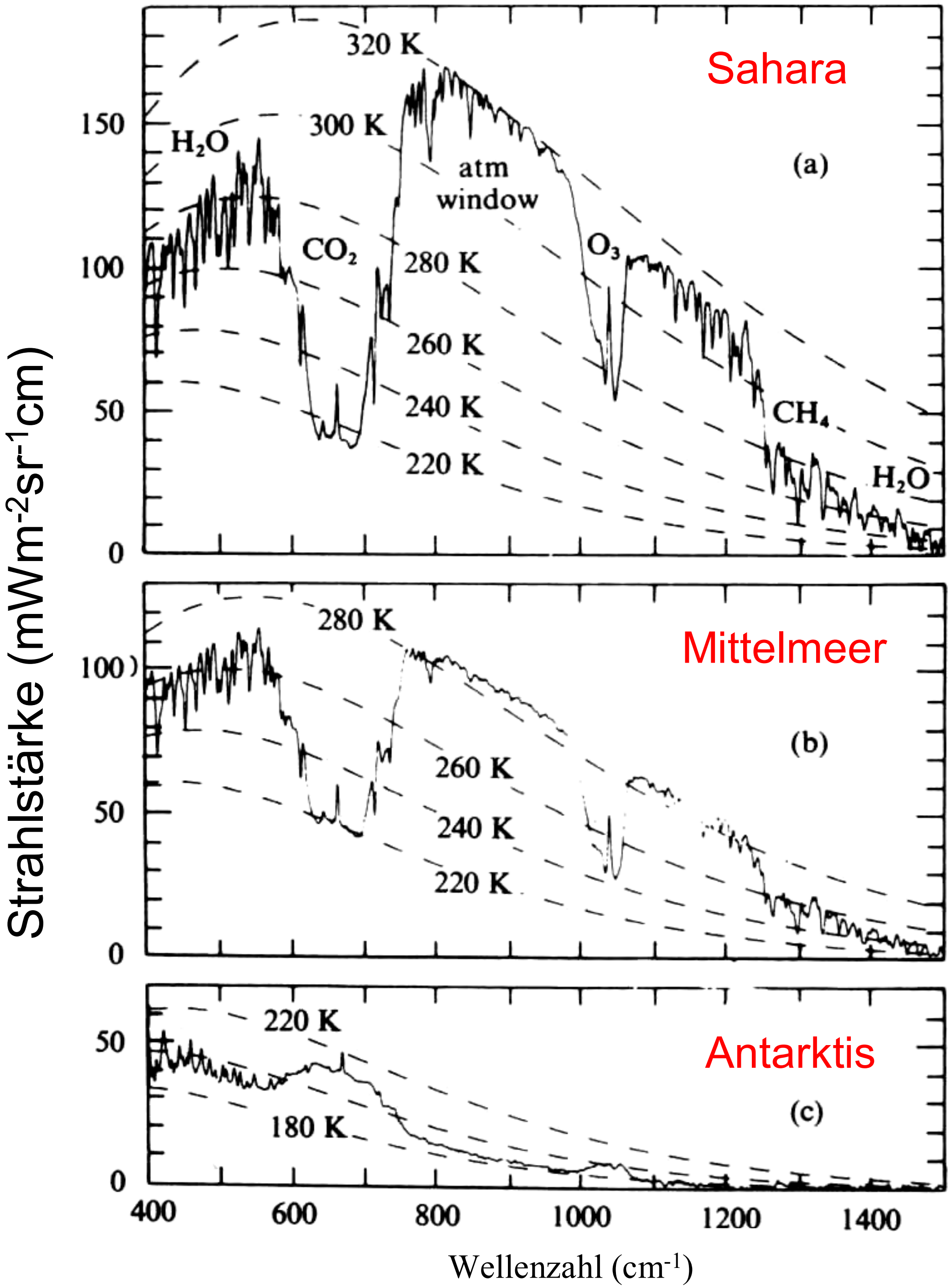


→ Detaillierte Rechnungen für jede Linie erforderlich!

Spectrum of the Terrestrial (LW) Emission



Spectra of Terrestrial (LW) Emission (NIMBUS 4)



Bergmann-Schäfer, 1997

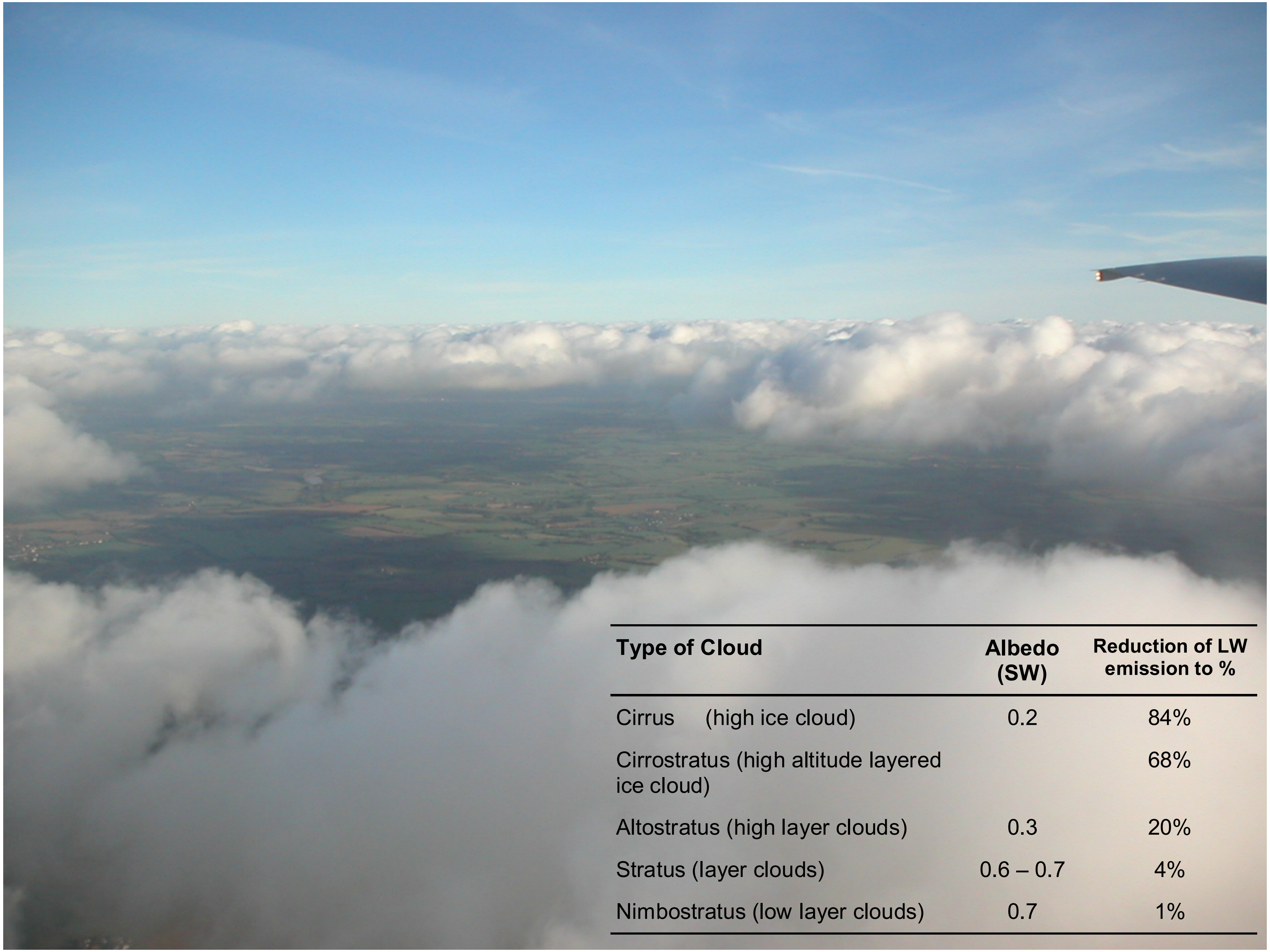
Climate Engineering Ideas:

Reduce Greenhouse Effect:

Reduce Greenhouse Gas (mostly CO_2) Concentrations

=Carbon dioxide Removal (CDR)

4^{th} 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%

What is a Cloud?

Kind of aerosol with particles of r around 10μm

→ Cloud particles scatter SW radiation $\lambda < 3.5\ \mu$m quite well

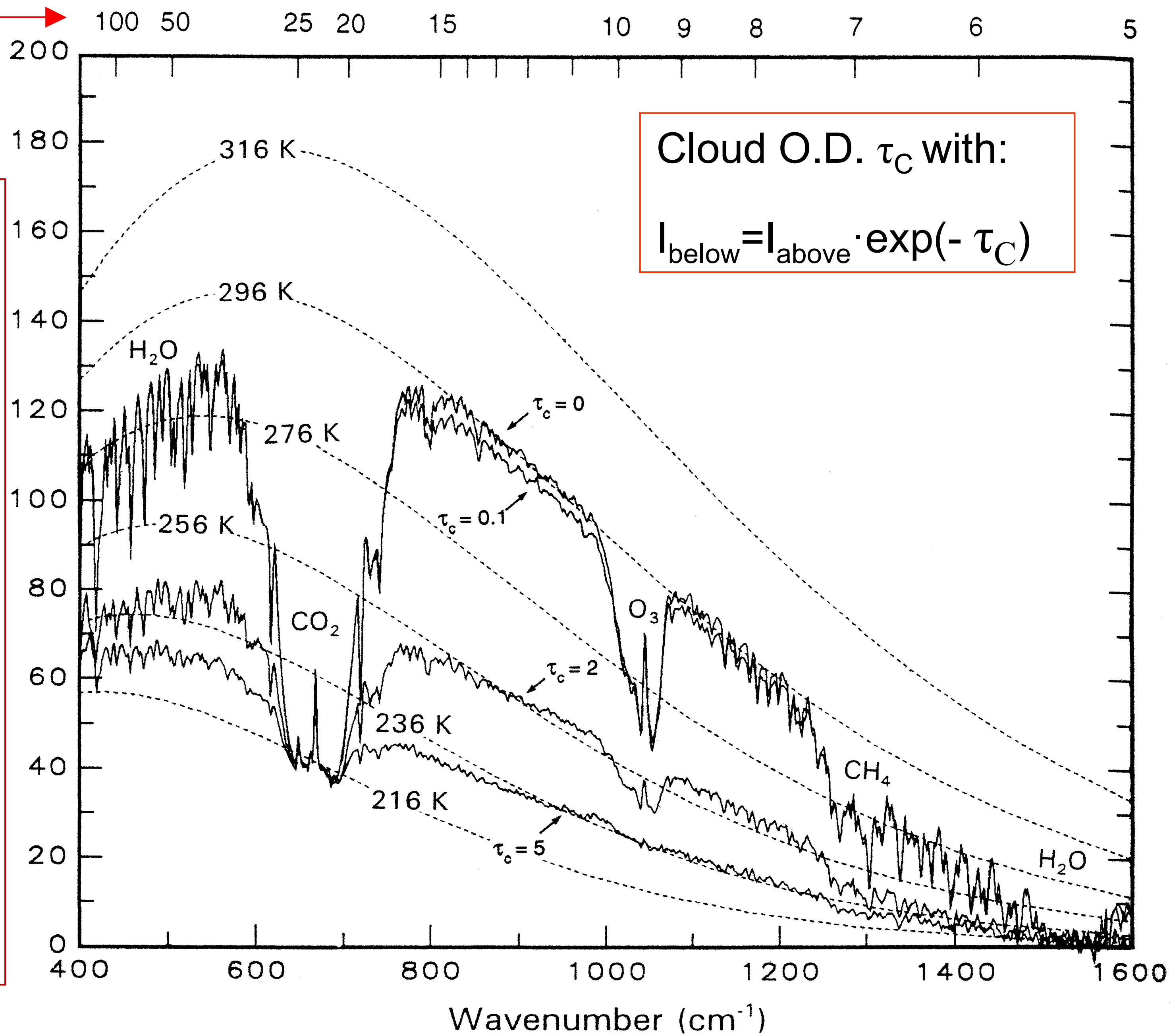
→ High SW albedo

→ Cloud particles scatter LW radiation also very well

→ Clouds act similar to greenhouse gases

Long-Wave Emission from Clouds

Wavelength (μm)



Cloud O.D. τ_C with:

$I_{below} = I_{above} \cdot \exp(-\tau_C)$

(high) Clouds reduce IR emission

→ heat Earth

However:

(low) Clouds cool due to high albedo!

Net-effect:

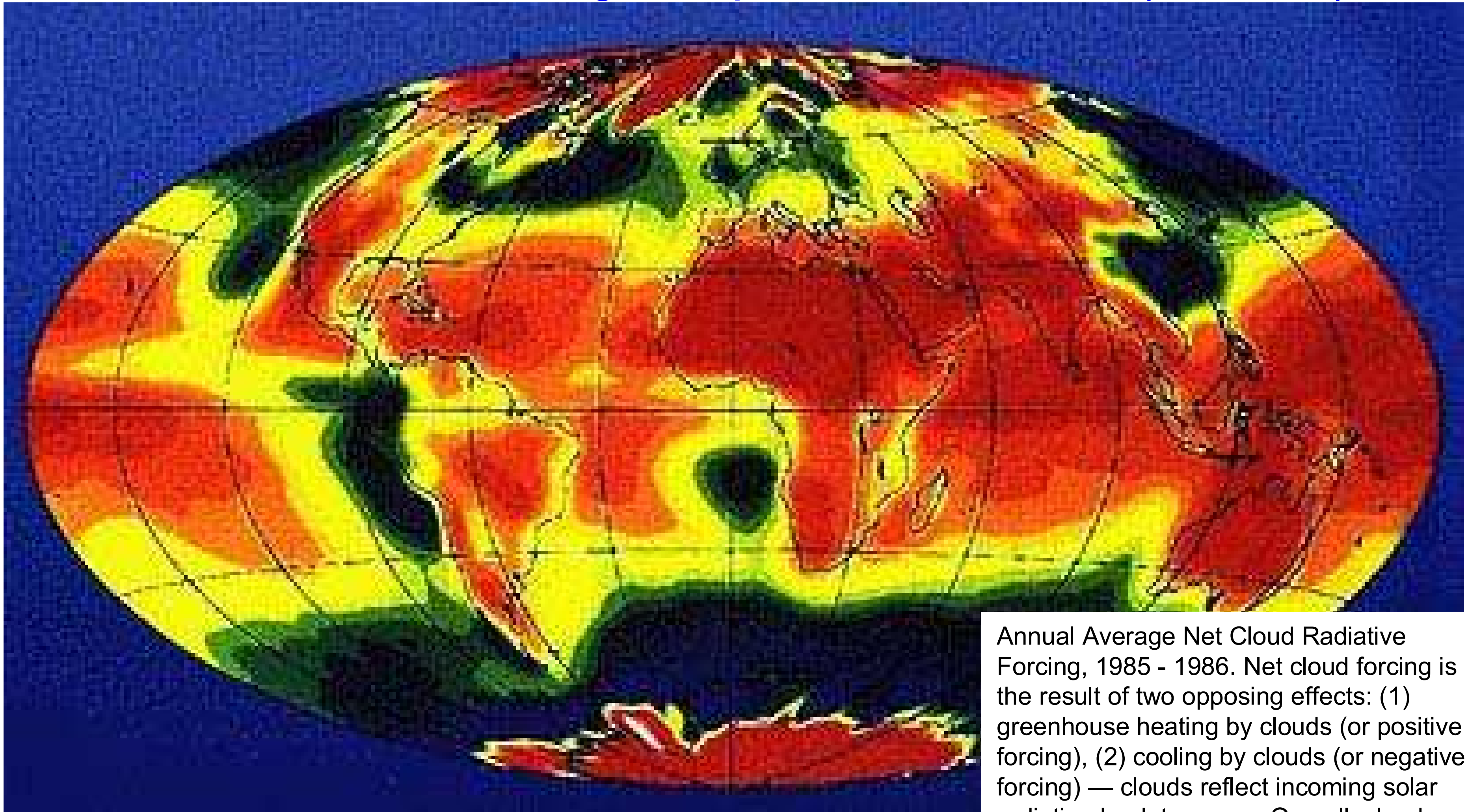
slight cooling

Climate Engineering Ideas:

Enhance cloud Albedo:

Cloud seeding

Earth Radiation Budget Experiment – ERBE (Satellite)

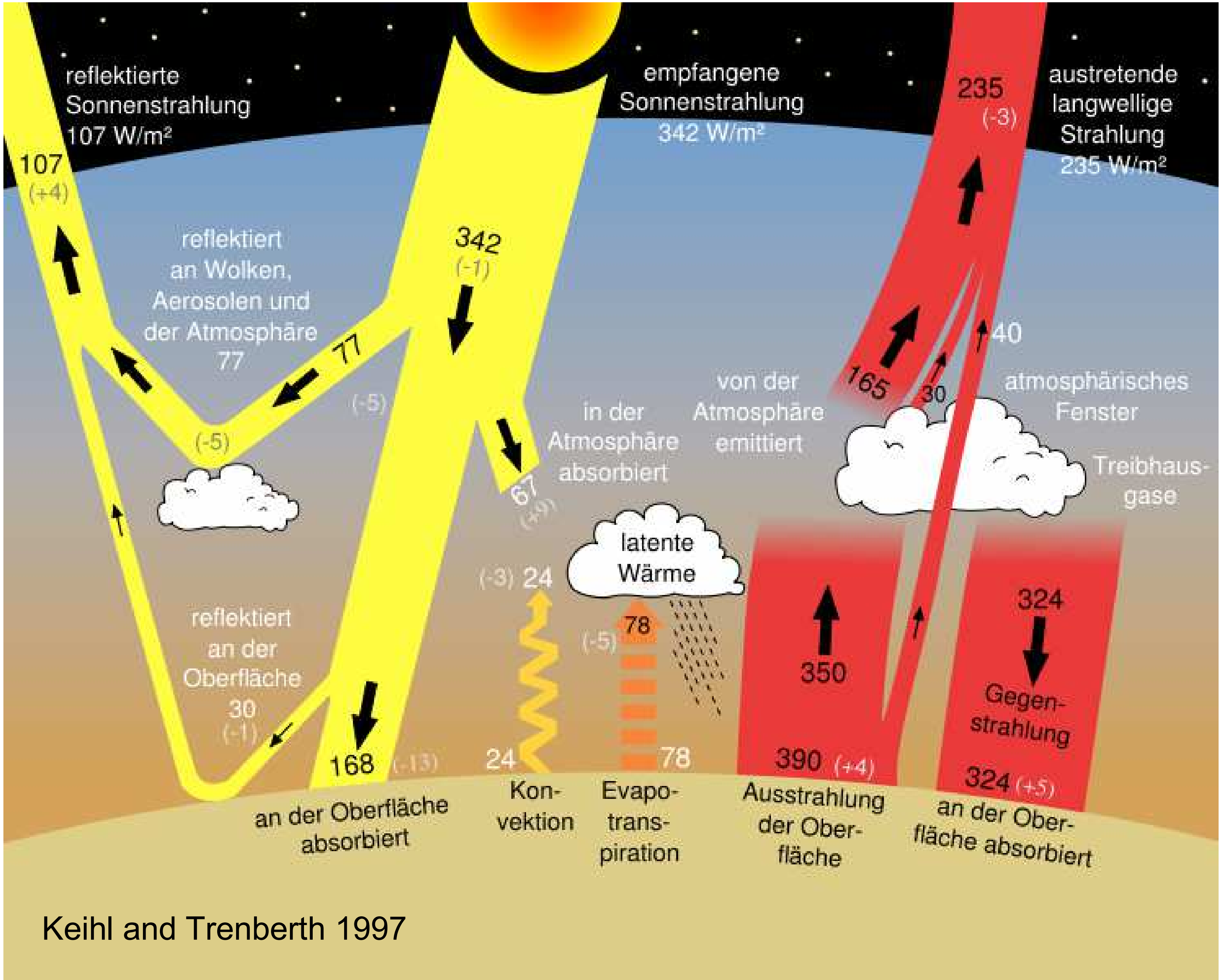


Annual Average Net Cloud Radiative Forcing, 1985 - 1986. Net cloud forcing is the result of two opposing effects: (1) greenhouse heating by clouds (or positive forcing), (2) cooling by clouds (or negative forcing) — clouds reflect incoming solar radiation back to space. Overall, clouds have the effect of lessening the amount of heating that would otherwise be experienced at Earth's surface-a cooling effect. (ERBE data on the Earth Radiation Budget Satellite and the NOAA-9 satellite. Data processed at NASA Langley Research Center; image produced at the University of Washington).

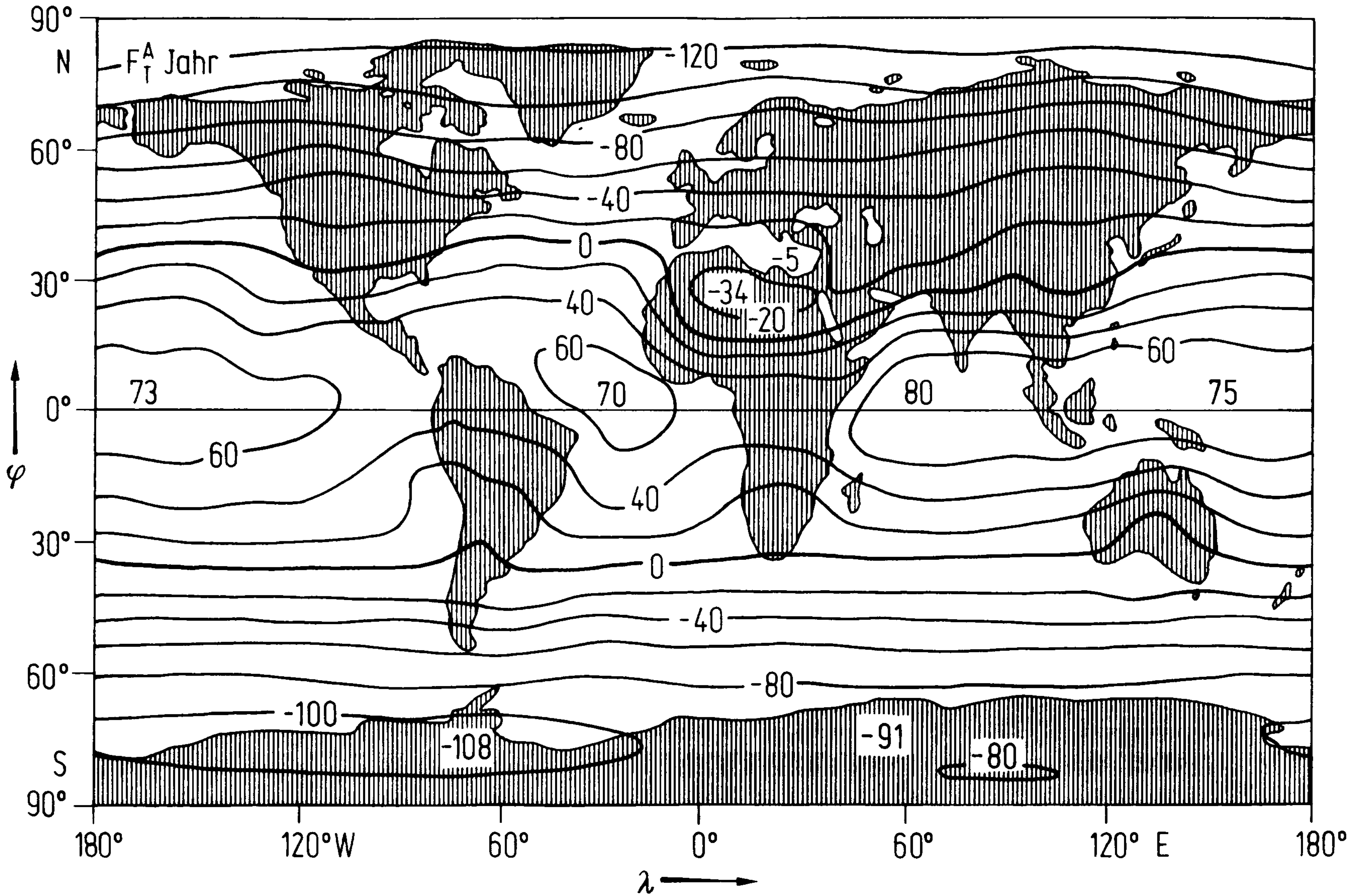
cooling - yellow to green to blue
heating - orange to red to pink
overall small net cooling effect by clouds

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

Energy Budget of Earth (Wm^{-2}, Global Mean)



Radiation Budget of Earth from Satellite



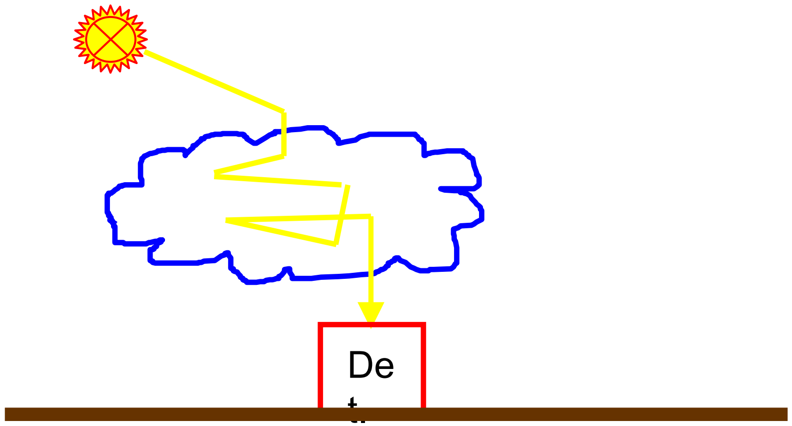
Yearly average of net radiation flux density in W/m², positive numbers: Gain of radiation energy (net radiation flux downwelling)

Summary

- Simple energy balance calculations reveal a lot about our climate:
  - presence of natural greenhouse effect
  - latitudinal and vertical structure of T and energy
  - radiative-convective equilibrium
- In more detail the line structure of atmospheric gases have to be taken into account
- Clouds are an important part of the climate system
- The global energy budget is in delicate balance, small changes have large effects

Contributions of the IUP to Climate Research

- SW – „extra“ energy absorption in the Atmosphere
- Light path lengths in clouds
- Greenhouse Gases: CH_4
- Cloud – feedback
- Carbon cycle
- Paleo climate



Satellite Data Evaluation

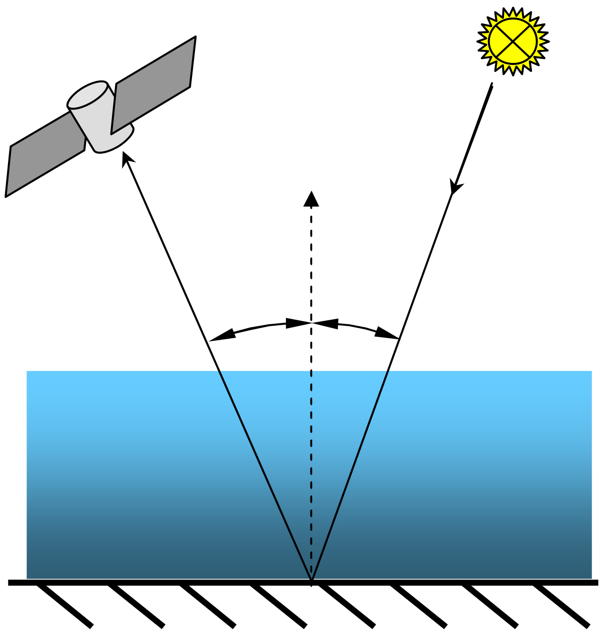
SCanning ImAging spectroMeter for Atmospheric CHartographY (SCIAMACHY)

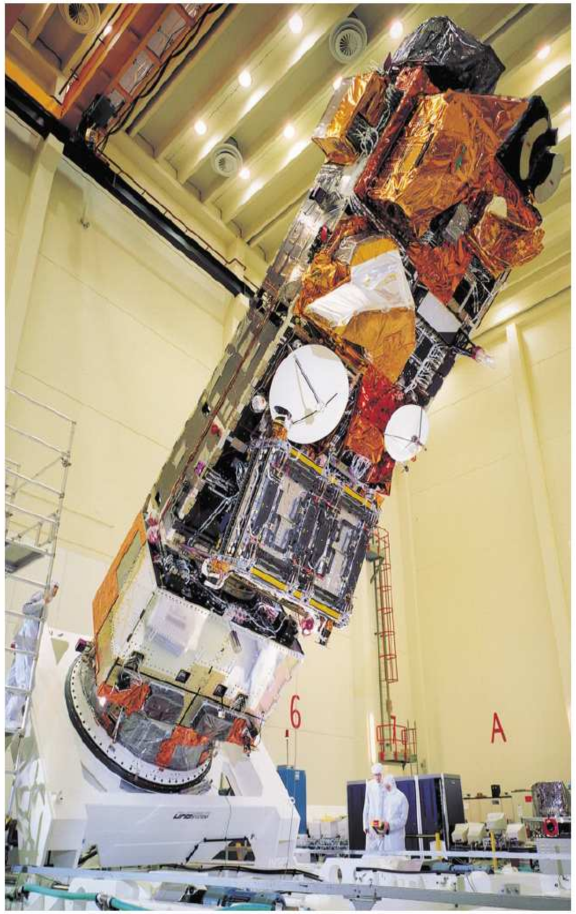
Launch: February 28, 2002 on ENVISAT

Spectral resolution 0.2 - 0.4 between 240 und 2400nm

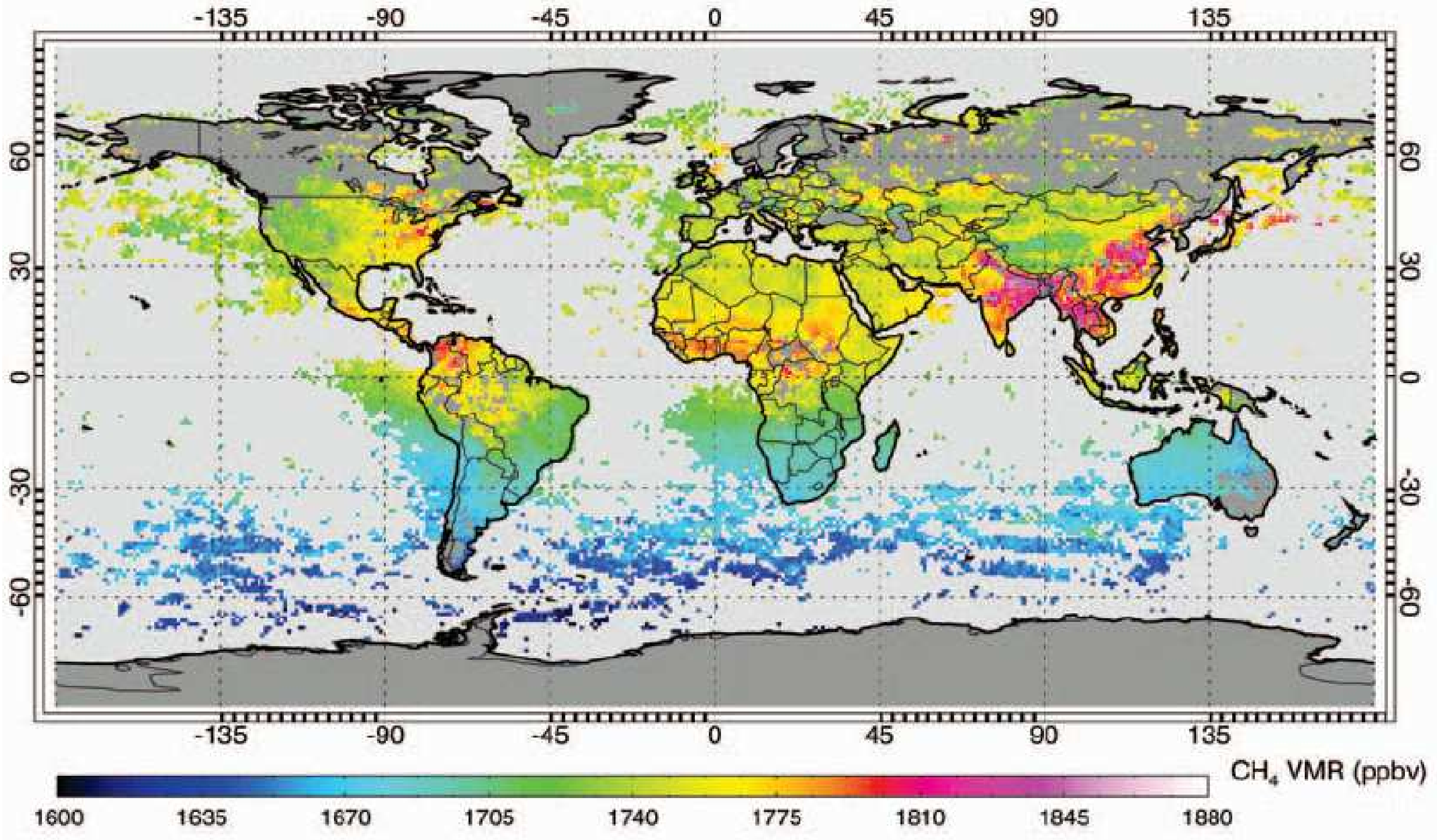
Gases: Ozone, NO_2 BrO OClO HCHO, SO_2 H_2O O_2 O_4 CO, CO_2 CH_4, N_2O

Viewing Geometry: **Nadir**, Limb, direct sun





Satellite Measurements of the Global CH_4 Mixing Ratio August - November 2003

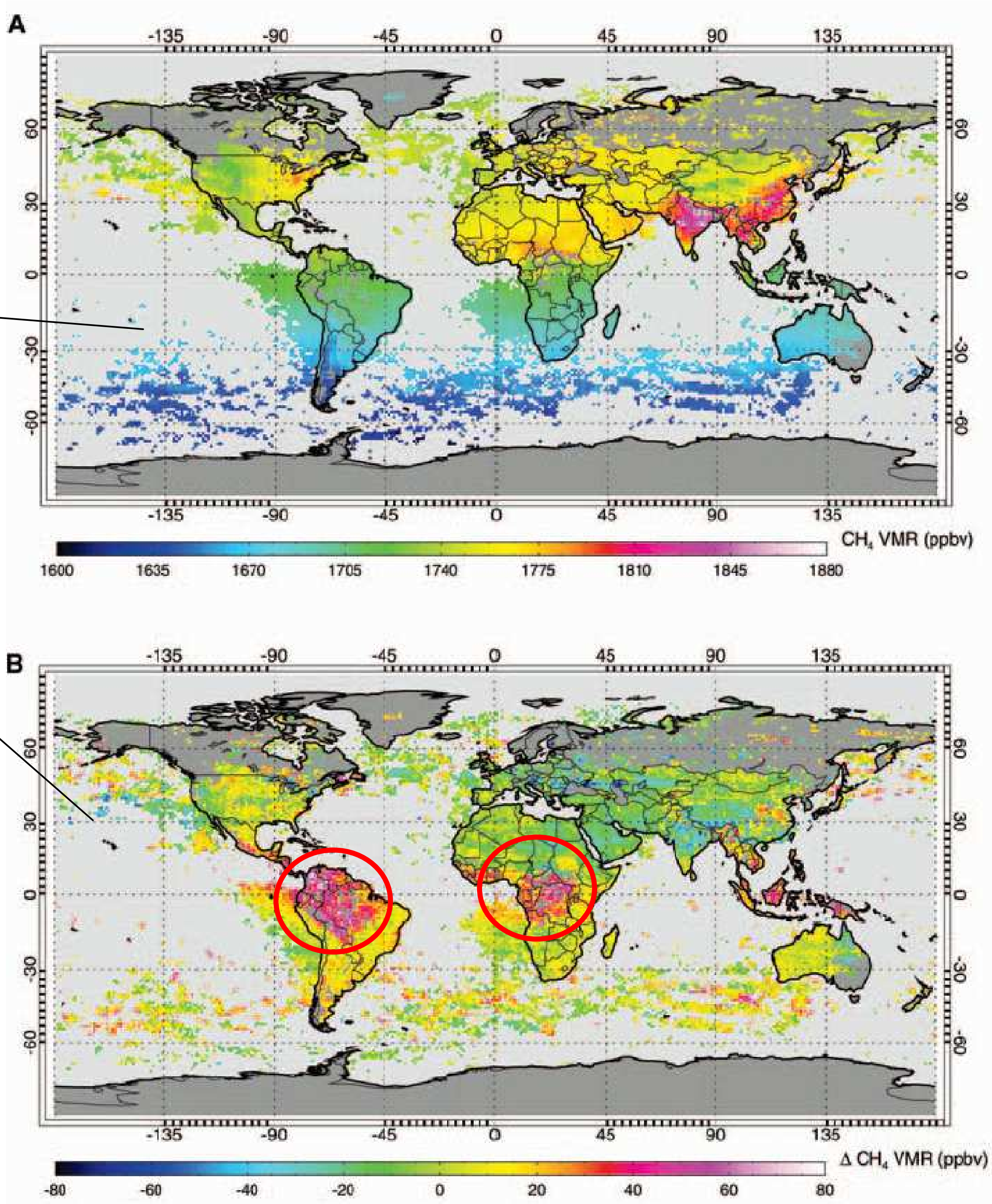


Frankenberg C., Meirink J.F., van Weele M., Platt U. and Wagner T., Assessing methane emissions from global space-borne observations, Science express, March 17, 2005

CH_4: Measurement - Model Comparison

TM3 Model Data, August - Nov. 2003

Difference: Model - Measured Data (ppb)



New CH4-Source: Emission from Rainforest

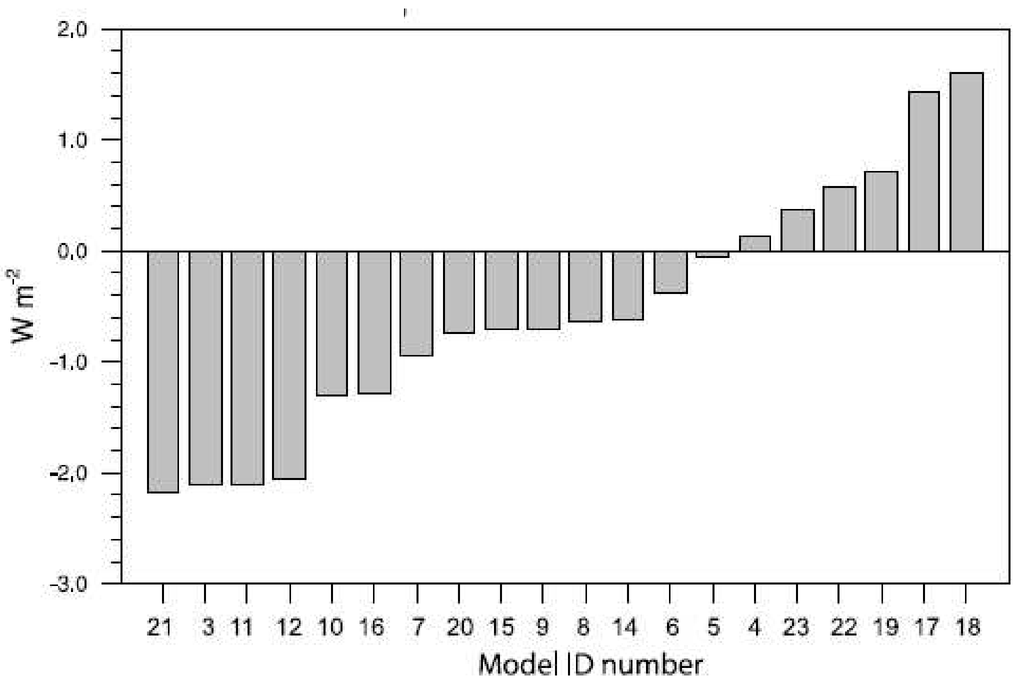
Frankenberg et al., Science 2005

Change in Cloud Forcing (1980-1999 vs. 2080-2099) Predicted by Different Models

General rule:

Low clouds tend to cool
High clouds tend to warm

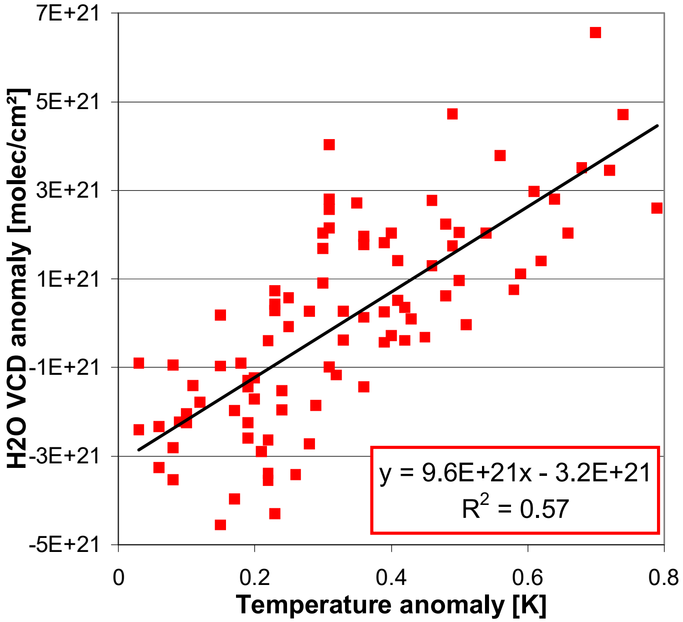
Main question:
Does cloud cover and/or distribution change when climate changes?



Changes in global mean cloud radiative forcing (Wm^{-2}) for the period 1980-1999 vs. 2080-2099

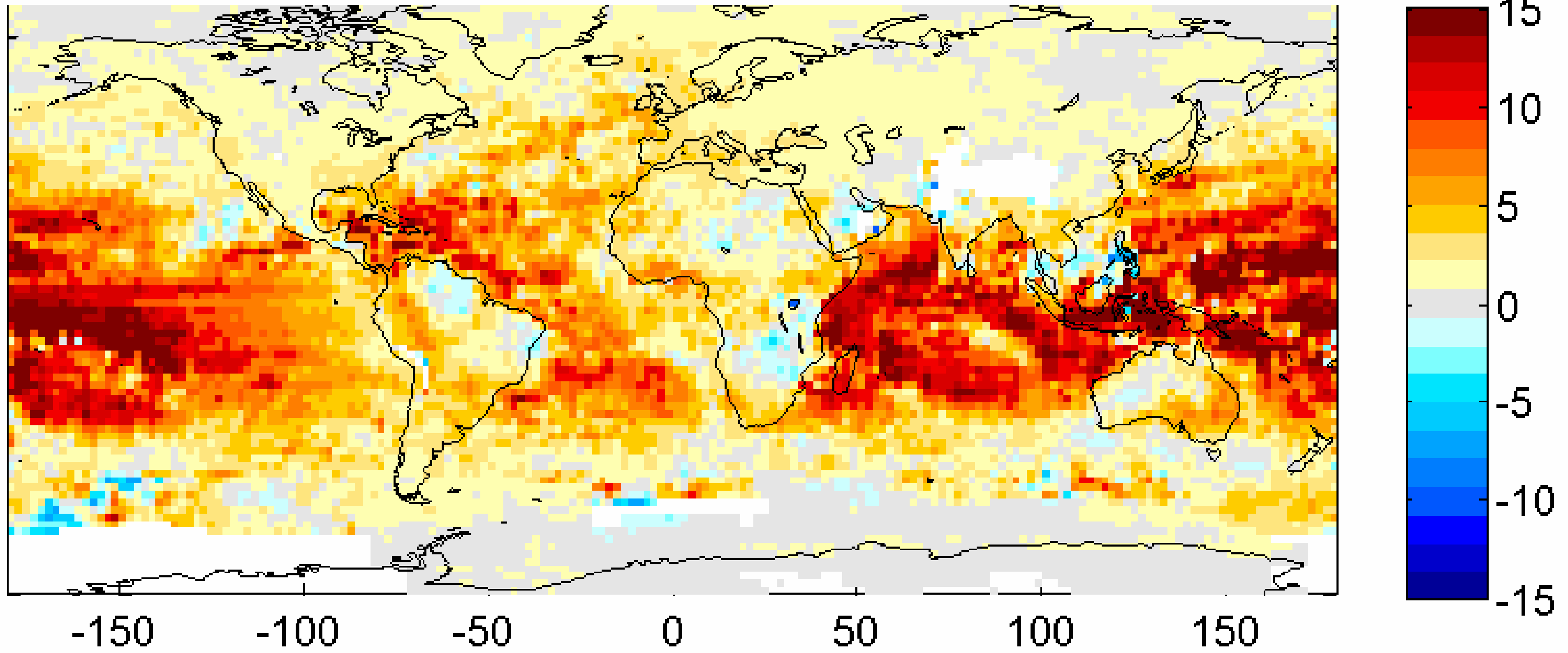
IPCC 2007

Dependence of the SCIAMACHY H2O Column on Temperature (ECMWF) (1996-2003)



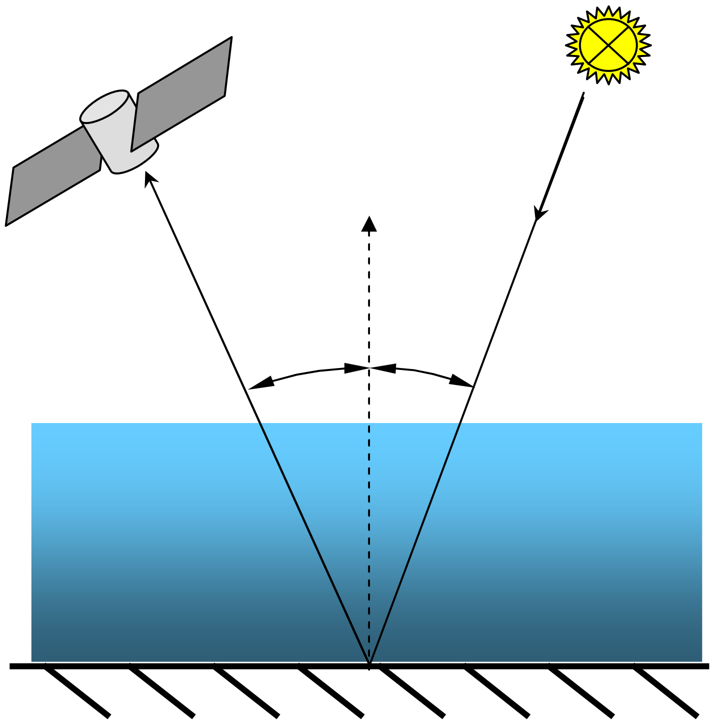
Wagner et al. 2007

Change of the H_2O column [10^{21} molec/cm²] per Kelvin temperature change

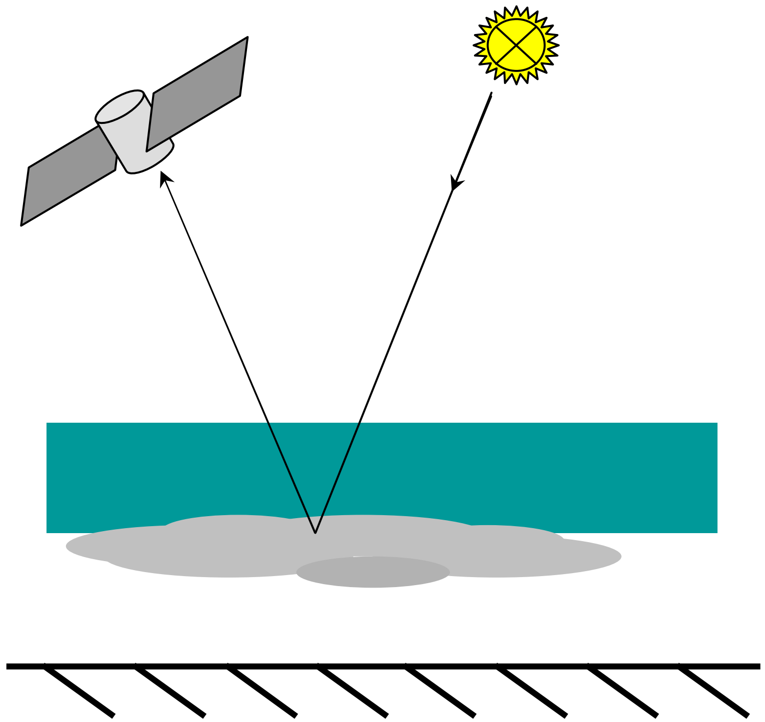


Cloud Fraction and Cloud (Top) – Height Observation from Satellite

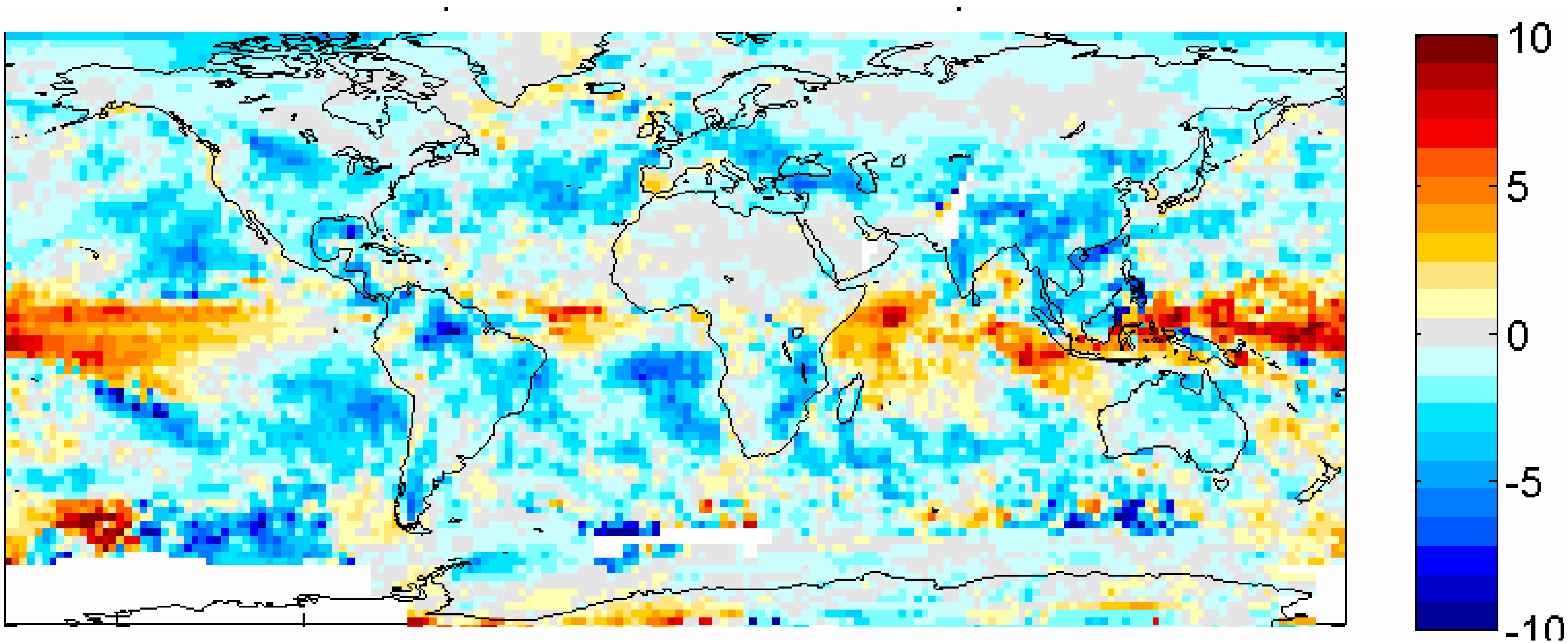
No Cloud: Large O_2-Column seen



Cloud: Small O_2-Column seen



Dependence of the cloud fraction on temperature derived from correlation analysis (1996-2003)

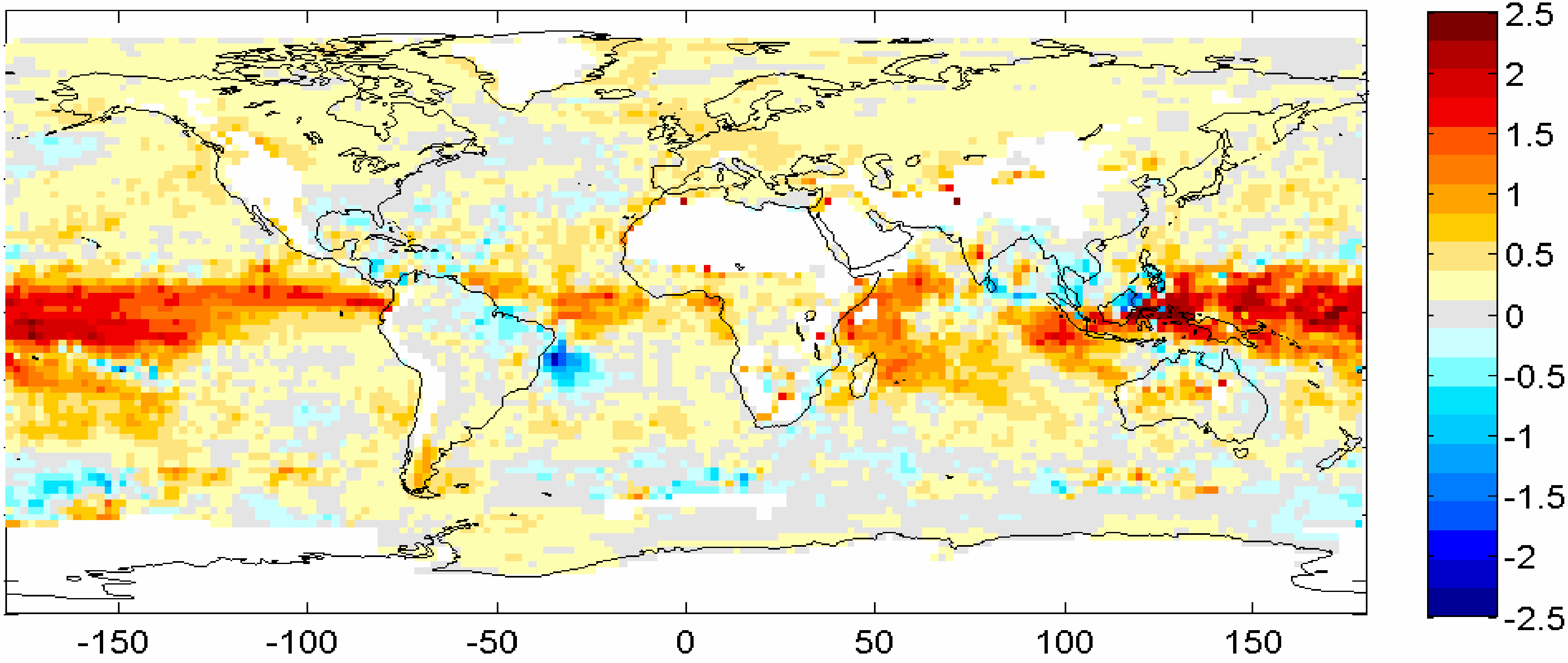


Change of the effective cloud fraction (% per K)

Higher Temperature → Fewer Clouds → lower Albedo → positive feedback

Wagner et al. 2007

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



Change of cloud top height (km per Kelvin)

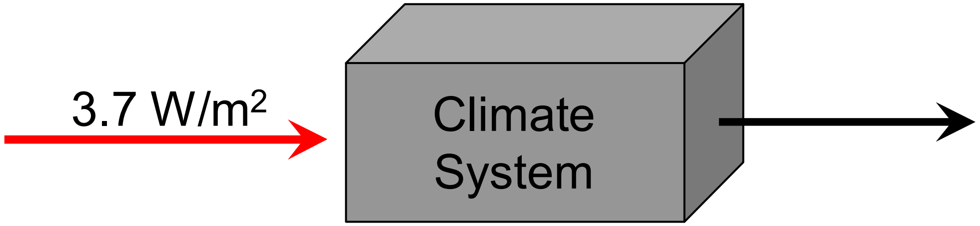
Higher Temperature → Higher Clouds
→ positive feedback on temperature

Wagner et al. 2007

Without feedbacks climate predictions would be rather easy

No feedback:

Antropogenic forcing:
(CO_2 Doubling)



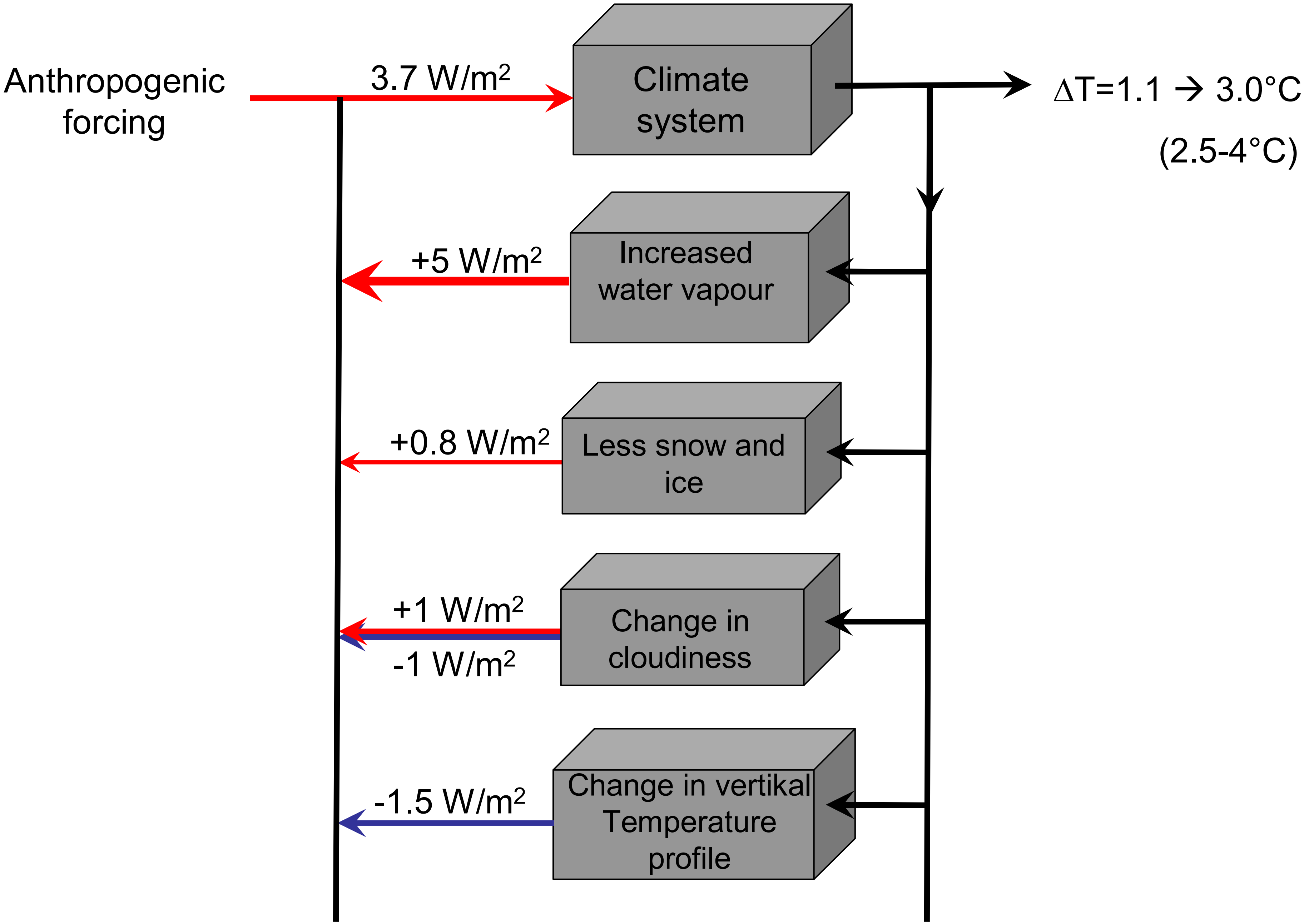
$\Delta T = 1.1°C$

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.

Because of Feedbacks Climate Predictions are Rather Difficult



Climate Sensitivity

equilibrium **climate sensitivity**:

Equilibrium change in global mean near-surface air temperature that would result from a sustained doubling of the atmospheric (equivalent) CO_2 concentration.

This value is estimated, by the IPCC Fourth Assessment Report (*AR4*) as likely to be in the range 2 to 4.5°C with a best estimate of about 3°C

With:

$F_{2CO2} \approx 3.7\ W/m^2$

ΔF = any climate forcing

$\Delta T_{2CO2} \approx 1.1\ K$

We obtain (in linear approximation) for the resulting temperature change ΔT:

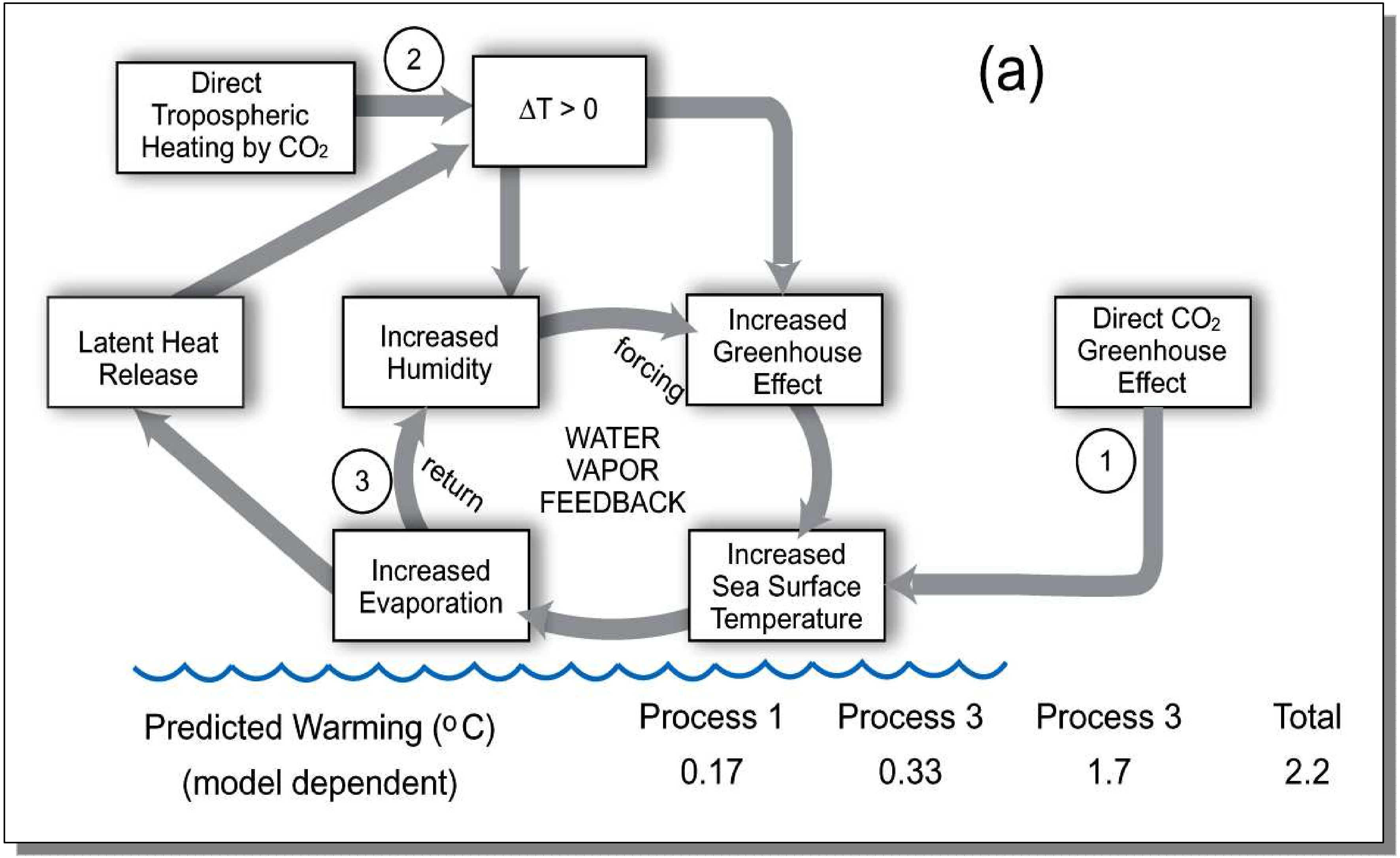
$$\Delta T = \frac{\Delta F}{F_{2CO2}} \cdot \Delta T_{2CO2} = \frac{\Delta T_{2CO2}}{F_{2CO2}} \cdot \Delta F = S_C \cdot \Delta F$$

For the present-day situation we obtain:

$$S_C = \frac{\Delta T_{2CO2}}{F_{2CO2}} \approx \frac{1.1}{3.7} \approx 0.3 \frac{K}{W/m^2}$$

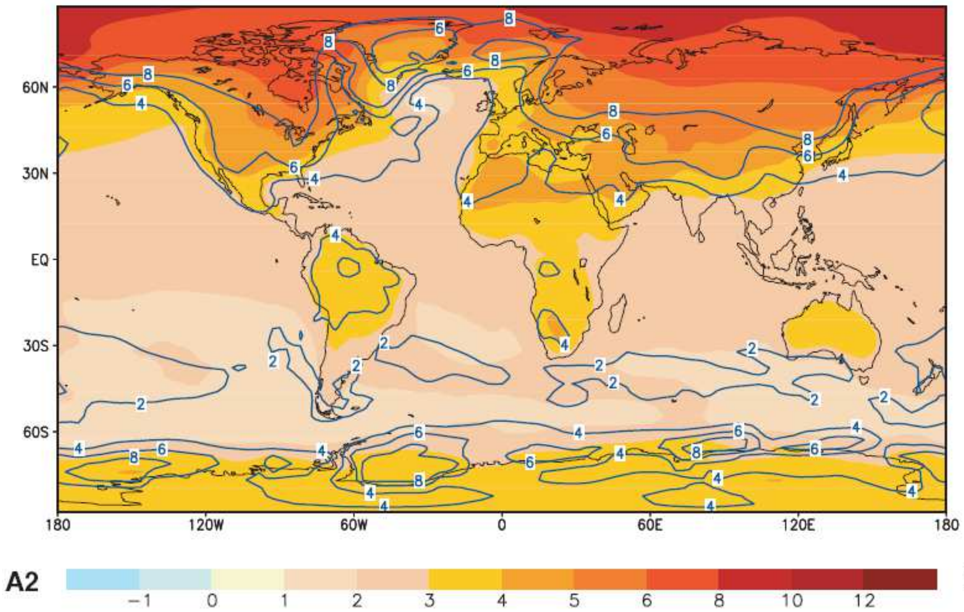
(For Glacial-Interglacial $S_C \approx 0.7 K (W/m^2)^{-1}$)

The Water Vapour Feedback

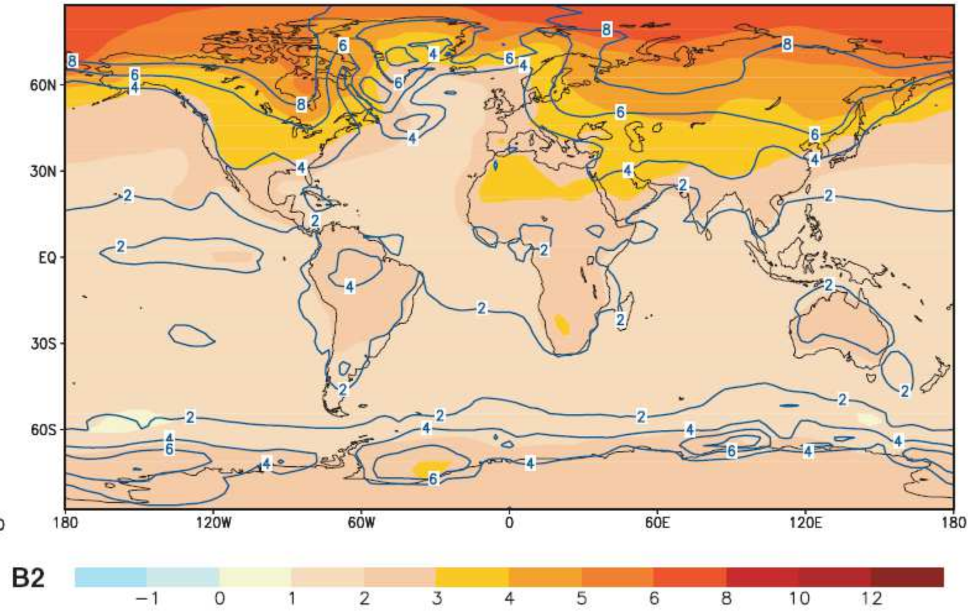


Regional Consequences of Climate Change

A2: strong CO_2 increase
2100: ~850 ppm



B2: moderate CO_2 increase
2100: ~600 ppm



The annual multi-model mean change of the temperature (colour shading) and its range (isolines) (Unit: °C) from OAGCMs.

Patterns are very similar, even though scenarios are very different.

IPCC