2. The Climate System – Radiation Balance
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 CO₂ 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?


- Temperature
- Wind
- Humidity
- Precipitation
- ...

Mark Twain: „Climate is what you expect, weather is what you get“
Earth in the Solar System

1368 W/m²
The Laws of Radiation by Planck, Stephan-Boltzmann, Wien

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

2. **Stefan-Boltzmann’s Law:**
   - $P = \varepsilon \sigma_{SB} T^4$

3. **Wien’s Displacement Law:**
   - The wavelength of the maximum is inversely proportional to the temperature of the radiating body.

1 m² bei 20°C:
$T = 20 + 273.2 = 293.2 K$

Radiated power:
$P = 419 \text{ W/m}^2$
Properties of Real Radiating Bodies

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

Energy conservation:
\[ \alpha + \rho + \tau = 1 \]
absorptivity  reflectivity  transmissivity

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)
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:

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

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

(IR) power radiated from earth:

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

- \( \varepsilon \approx 0.9 \ldots 1.0 \) = IR Emissivity of earth
- \( \sigma_{SB} = 5.67 \cdot 10^{-8} \text{ Wm}^{-2}\text{K}^{-4} \) = 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:

\[ P_{\text{out}} = P_{\text{in}} \]

*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 \( P_{\text{out}} \) and \( P_{\text{in}} \) we obtain the average of surface temperature of earth \( T_0 \) in 0\(^{th}\) approximation:

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

with the above numbers we obtain: \( T_0 \approx 255 \text{ K} \)

Measured average temperature of earth: \( T_e \approx 288 \text{ K} \)

Cause of the discrepancy:

The natural greenhouse effect of 33K.

Comment: the main constituents of the atmosphere, \( \text{N}_2 \) and \( \text{O}_2 \) provide a negligible greenhouse effect of \( \approx 0.28 \text{ W/m}^2 \) (global mean), see: Höpfner M., Milz M., Buehler S., Orphal J., and Stiller G. (2012), The natural greenhouse effect of atmospheric oxygen (\( \text{O}_2 \)) and nitrogen (\( \text{N}_2 \)), 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)

<table>
<thead>
<tr>
<th>Gas</th>
<th>Prominent Band</th>
<th>$\Delta T$ K</th>
<th>%</th>
</tr>
</thead>
<tbody>
<tr>
<td></td>
<td>$\mu$m</td>
<td></td>
<td></td>
</tr>
<tr>
<td>H$_2$O</td>
<td>6.3, &gt;16</td>
<td>20.6</td>
<td>62</td>
</tr>
<tr>
<td>CO$_2$</td>
<td>13 - 17</td>
<td>7.2</td>
<td>22</td>
</tr>
<tr>
<td>O$_3$ (in the troposphere)</td>
<td>9.6</td>
<td>2.4</td>
<td>7</td>
</tr>
<tr>
<td>N$_2$O</td>
<td>4.8, 7.8</td>
<td>1.4</td>
<td>4</td>
</tr>
<tr>
<td>CH$_4$</td>
<td>3.4, 7.3</td>
<td>0.8</td>
<td>2.5</td>
</tr>
</tbody>
</table>
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”

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

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

aus: W. Roedel, 2000
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.
Emission Spectra of Earth and Sun

Stratosphere: UV-absorption by O$_3$

Troposphere: IR-absorption by H$_2$O, CO$_2$, CH$_4$, etc.

Atmosphere reduces IR-radiation

Atmosphere absorbs and reemits IR radiation

atm. window: ~ground radiates to space
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 Atmospheric 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 Fouillet§; 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 and Motion,' 2nd ed. p. 405 (Lon., 1865).
§ § '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 $\varepsilon_G$,
VIS-(SW)Albedo $A = A_\rho$

**Atmosphere:**
Temperature $T_A$,
IR-Absorptivity = Emissivity $\varepsilon_A$,
Completely transparent for short wave radiation
The Effect of Absorbing Layers

\[ S'_1 = (1 - \alpha_A)S_1 \]

\[ P_1 = \varepsilon_G \sigma_{SB} T_G^4 \]

\[ P_2 = \varepsilon_A \sigma_{SB} T_2^4 \]

\[ T_2 < T_1 \]
Surface Temperature of Earth. 2\textsuperscript{nd} Approximation

Radiation equilibrium for both layers:

1) Ground: \[ \frac{S_0}{4} (1 - A) + \varepsilon_A \sigma_{SB} T_A^4 = \varepsilon_G \sigma_{SB} T_G^4 \]

where \( S_0 \) is the mean insolation, \( \varepsilon_A \) is the emissivity of the ground, \( \sigma_{SB} \) is the Stefan-Boltzmann constant, and \( T_A \) and \( T_G \) are the temperatures of the atmosphere and ground, respectively.

2) Atmosphere:

\[ \varepsilon_A \varepsilon_G \sigma_{SB} T_G^4 = 2 \varepsilon_A \sigma_{SB} T_A^4 \]

\( \varepsilon_A = \alpha_A \) indicates that the atmosphere radiates both upwards and downwards.

2 Eq., 2 unknown variables: \( T_A, \varepsilon_A \) → Solution:

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

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

Note: Effektiere emission from higher, colder layer!

\( \Delta T = T_B - T_A = 49 \) K → \( z = \Delta T/\Gamma \approx 49/9.8 \) K km\(^{-1} \approx 5.0 \) km altitude
Gas Clouds also follow Kirchhoff’s Law

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 \((1-\alpha)\) + 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) $\tau$ [-]:

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

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

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

Discovered by:
Pierre Bouguer in 1729
Johann H. Lambert in 1760
August Beer in 1852
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$_2$O and O$_3$

Normal modes of vibration for linear molecules, e.g. CO$_2$ or N$_2$O

(Herzberg 1945)

Figure 8.11 Absorption spectra in (a) the 15-μm band of CO$_2$ and (b) the rotational band of H$_2$O 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:**
  \[ \sigma_{av} = S \sigma(v - v_0) = f(v - v_0) \int \sigma_v \, dv \]
  with \( S \) line strength and \( f \) shape factor (\( v_0 \) – line center)

- **natural** line broadening due to finite lifetime of excited state:
  Lorentz shape:
  \[ f_n(v - v_0) = \frac{\alpha_n}{\pi(v - v_0)^2 + \alpha_n^2} \quad \text{with} \quad \alpha_n = \frac{1}{2\pi \tau} \]

- **Doppler** broadening due to molecular motion in line of sight:
  Gaussian shape:
  \[ f_D(v - v_0) = \frac{1}{\alpha_D \sqrt{\pi}} \exp\left( -\frac{(v - v_0)^2}{\alpha_D^2} \right) \quad \text{with} \quad \alpha_D = \frac{v_0}{c} \sqrt{\frac{2kT}{m}} \]

- **Pressure** broadening due to collisions of molecules (\( \rightarrow \) lifetime reduction):

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

- **Note:** Pressure broadening: \( \Delta v \) independent of \( v \), Doppler: \( \Delta v \propto v \)

  Thus pressure broadening dominates at small \( v \) (long wavelengths, IR)
  Doppler broadening dominates at high \( v \) (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:
(\text{short-wave}) \text{ incoming solar radiation} = (\text{long wave}) \text{ IR outgoing radiation}

\[ S_0 \cdot (1 - A) / 4 = \sigma_{SB} \cdot T_s^4 \cdot (1 - B) \quad \text{[(1-B)=\varepsilon_G]} \]

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

\( T_s = 255 \text{ K} \) for \( B = 0 \) or \( 287.5 \text{ K} \) for \( B = 0.40 \), resp.

\( A \): Earth albedo in the visible (SW) Spectral range (= 0.298)
\( T_s \): Earth surface temperature
\( B \): IR albedo (= 0.40)
\( S_0 \): Solar constant (~ 1370 W/m²)
\( \sigma_{SB} \): Stefan Boltzmann constant (5.87 \times 10^{-8} \text{ W/(m}^4\text{K)}

Climate sensitivities (no feedback):

\[ \frac{\partial T_s}{\partial S_0} = T_s / 4 \cdot S_0 = 0.05 \text{ K} \cdot \text{m}^2 / \text{W} \text{ (cf. \sim 0.1 K} \cdot \text{m}^2 / \text{W from GCM's)} \]

\[ \approx 0.70 \text{ K} / 1\% \text{ change in } S_0 \]

\[ \frac{\partial T_s}{\partial A} = - T_s / \{4 \cdot (1-A)\} = 0.37 \text{ K} / 1\% \text{ change in } A \text{ (e.g. clouds, ice cover, aerosol, ...)} \]

\[ \frac{\partial T_s}{\partial B} = T_s / \{4 \cdot (1-B)\} = 0.51 \text{ K} / 1\% \text{ change in } B \text{ (e.g. greenhouse gases, cirrus, ...)} \]
Atmospheric Temperature Profile
Greenhouse Effect: Effective Emission height of the LW – Radiation

- Natural Greenhouse Effect
- Anthropogenic (enhanced) Greenhouse Effect
- No Atmosphere

Höhe [km]

Temp. [K]

250 300

0 5 10
Radiative equilibrium at TOA and surface

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.

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

IR-Emission measured over Afrika
Spectra of Terrestrial (LW) Emission (NIMBUS 4)

Bergmann-Schäfer, 1997
Climate Engineering Ideas:

Reduce Greenhouse Effect:

Reduce Greenhouse Gas (mostly $\text{CO}_2$) Concentrations

= Carbon dioxide Removal (CDR)
### 4th Approximation: The Role of Clouds

<table>
<thead>
<tr>
<th>Type of Cloud</th>
<th>Albedo (SW)</th>
<th>Reduction of LW emission to %</th>
</tr>
</thead>
<tbody>
<tr>
<td>Cirrus (high ice cloud)</td>
<td>0.2</td>
<td>84%</td>
</tr>
<tr>
<td>Cirrostratus (high altitude layered ice cloud)</td>
<td></td>
<td>68%</td>
</tr>
<tr>
<td>Altostratus (high layer clouds)</td>
<td>0.3</td>
<td>20%</td>
</tr>
<tr>
<td>Stratus (layer clouds)</td>
<td>0.6 – 0.7</td>
<td>4%</td>
</tr>
<tr>
<td>Nimbostratus (low layer clouds)</td>
<td>0.7</td>
<td>1%</td>
</tr>
</tbody>
</table>
What is a Cloud?

Kind of aerosol with particles of r around \(10 \mu m\)

→ 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

Cloud O.D. $\tau_C$ with:

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

(high) Clouds reduce IR emission → heat Earth

However:

(low) Clouds cool due to high albedo!

Net-effect: slight cooling

Wavelength ($\mu$m)

Net-long-wave cooling
Climate Engineering Ideas:

Enhance cloud Albedo:

Cloud seeding
Earth Radiation Budget Experiment – ERBE (Satellite)

Annul 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)

Keihl and Trenberth 1997
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 and 2400 nm
Gases: Ozone, NO\textsubscript{2}, BrO, OCIO, HCHO,
SO\textsubscript{2}, H\textsubscript{2}O, O\textsubscript{2}, O\textsubscript{4}, CO, CO\textsubscript{2}, CH\textsubscript{4}, N\textsubscript{2}O
Viewing Geometry: Nadir, Limb, direct sun
Satellite Measurements of the Global CH$_4$ Mixing Ratio
August - November 2003

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
General rule:
Low clouds tend to cool
High clouds tend to warm

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

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

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

IPCC 2007
Dependence of the SCIAMACHY H$_2$O Column on Temperature (ECMWF) (1996-2003)

Wagner et al. 2007

Change of the H$_2$O column [$10^{21}$ molec/cm$^2$] 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)

Higher Temperature $\rightarrow$ Fewer Clouds
$\rightarrow$ lower Albedo $\rightarrow$ positive feedback

Wagner et al. 2007
Dependence of the Cloud Top Height (from O$_2$) on Temperature 1996-2003

Higher Temperature $\rightarrow$ Higher Clouds $\rightarrow$ positive feedback on temperature

Change of cloud top height (km per Kelvin)

Wagner et al. 2007
Without feedbacks climate predictions would be rather easy

No feedback:

Antropogenic forcing: (CO$_2$ Doubling) 3.7 W/m$^2$ Climate System $\Delta T=1.1^\circ$C

Doubling of the CO$_2$ concentration:

→ temperature increase of about 1.1K

Because of Feedbacks Climate Predictions are Rather Difficult

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

Change in cloudiness

-1 W/m²

Change in vertical Temperature profile

-1.5 W/m²
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:

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

$\Delta F = \text{any climate forcing}$

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

We obtain (in linear approximation) for the resulting temperature change $\Delta T$:

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

For the present-day situation we obtain:

$$S_C = \frac{\Delta T_{2CO2}}{F_{2CO2}} \approx \frac{1.1}{3.7} \approx 0.3 \frac{\text{K}}{\text{W/m}^2} \text{ (For Glacial-Interglacial } S_C \approx 0.7 \text{K}(\text{W/m}^2)^{-1})$$
The Water Vapour Feedback

1. Direct CO$_2$ Greenhouse Effect
2. \( \Delta T > 0 \)
3. Water Vapor Feedback

- Latent Heat Release
- Increased Humidity
- Increased Evaporation
- Increased Sea Surface Temperature

Predicted Warming (°C) (model dependent)

<table>
<thead>
<tr>
<th>Process 1</th>
<th>Process 3</th>
<th>Process 3</th>
<th>Total</th>
</tr>
</thead>
<tbody>
<tr>
<td>0.17</td>
<td>0.33</td>
<td>1.7</td>
<td>2.2</td>
</tr>
</tbody>
</table>
Regional Consequences of Climate Change

A2: strong CO₂ increase
2100: ~850 ppm

B2: moderate CO₂ 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