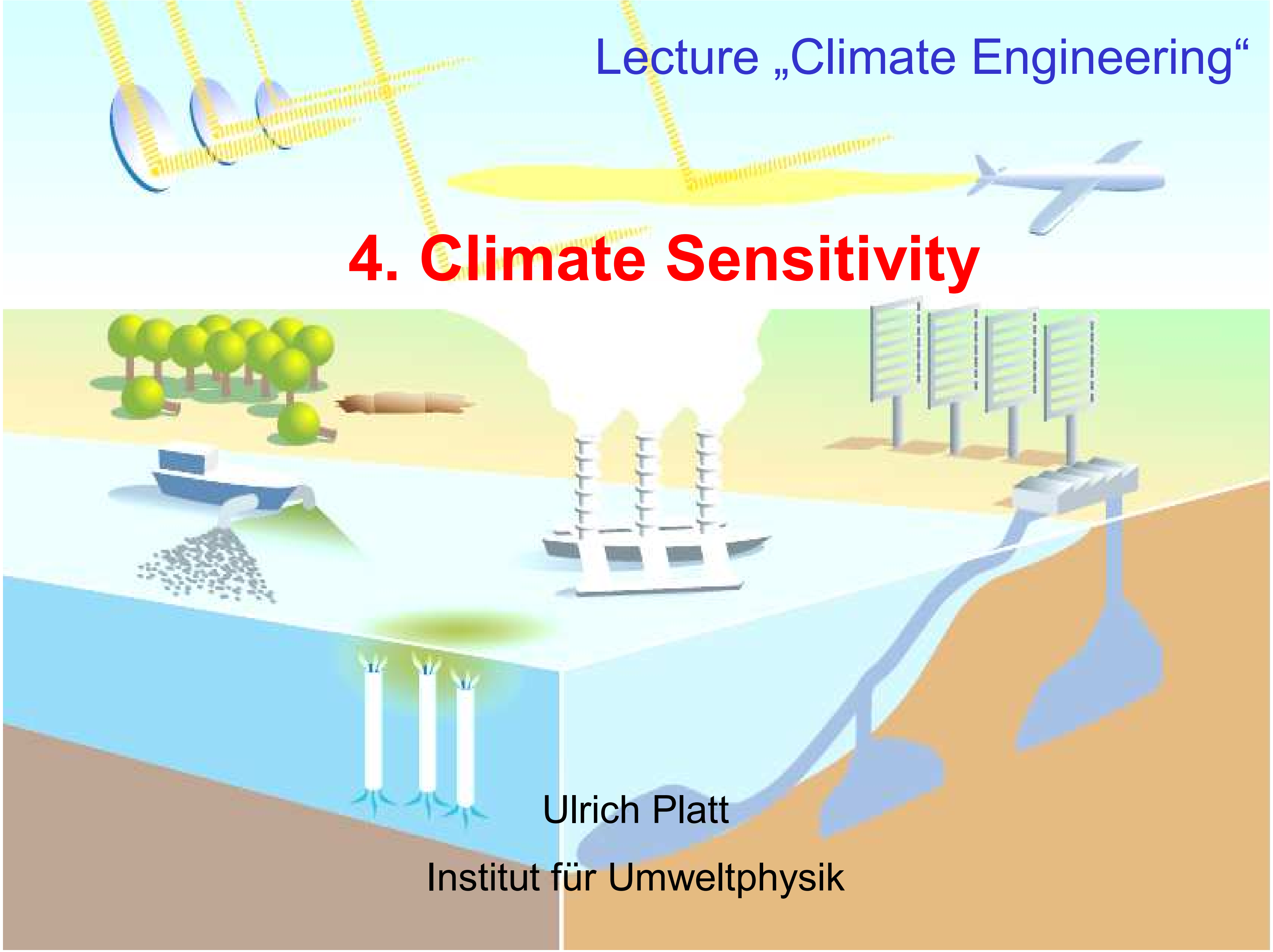


Lecture „Climate Engineering“

4. Climate Sensitivity

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. Climate System: Sensitivity, Predictions

Part 2: Climate Engineering Methods – SRM (4 sessions)

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

Part 3: Climate Engineering Methods – CDR (4 sessions)

1. Direct (Carbon dioxide) removal from air
2. Alkalinity to the ocean (enhanced weathering)
3. Ocean fertilization
4. 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

Literature

Textbooks

Marshall, J. and R. A. Plumb, 2008. Atmosphere, Ocean, and Climate Dynamics. An introductory Text. Elsevier Academic Press. Chap. 12.

Peixoto J.P., Oort A.H., 1993. Physics of Climate. American Institute of Physics, New York.

Roedel W. und Wagner T., 2017. Physik unserer Umwelt, Die Atmosphäre. Springer Verlag, Heidelberg, 5th Edition.

Online Resources

Goosse H. et al., 2010. Introduction to climate dynamics and climate modeling. Online textbook, http://www.climate.be/textbook

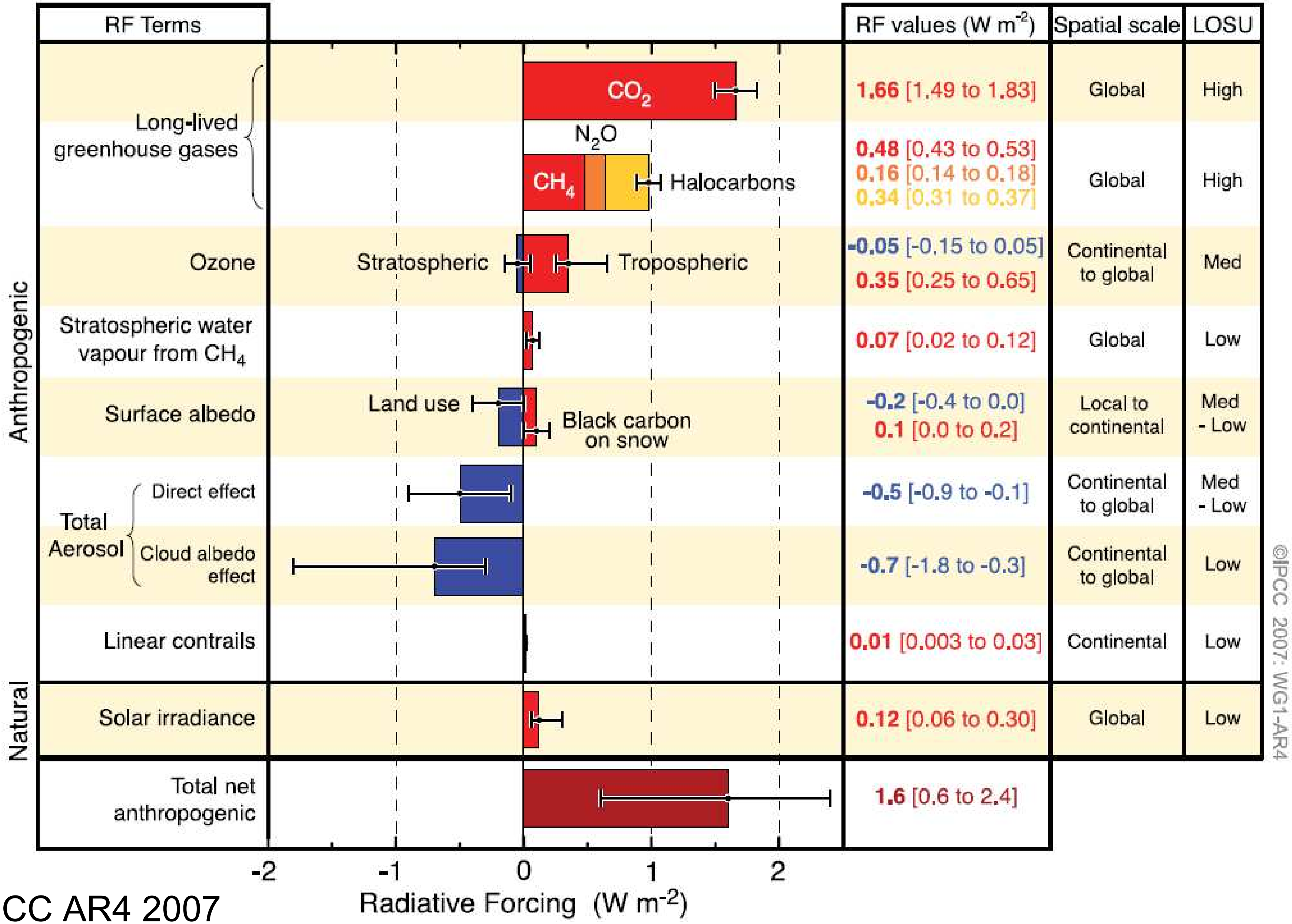
Stocker, T. 2008. Introduction to Climate Modelling. Lecture notes, University of Bern. See: https://climatehomes.unibe.ch/~stocker/papers/stocker08EKM.pdf

Contents of Today's Lecture

Climate Sensitivity, Feedbacks, and Predictions

- Climate sensitivity and feedbacks
- Heat capacity and response time
- Climate predictions

Global Annual Mean Radiative Forcing



IPCC AR4 2007

Climate Sensitivity

Anthropogenic changes (greenhouse gases, aerosols,…) lead to a radiative forcing F [W/m^2] of the climate system

Big question: How strongly (and how fast) will the climate system (global mean temperature T) react to the forcings?

Measure:

Equilibrium climate sensitivity λ: $\lambda = \frac{dT}{dF} \left[\frac{K}{Wm^{-2}} \right]$

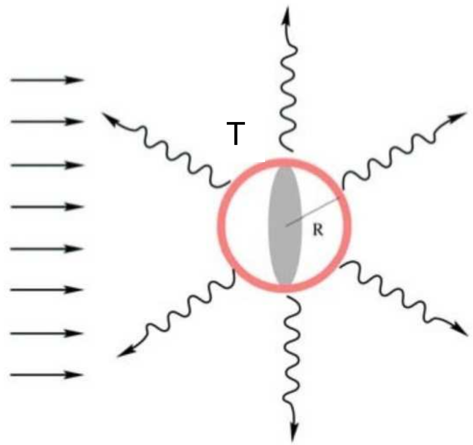
- Often the equilibrium temperature change for a doubling of atmospheric CO_2 (from 280 to 560 ppm) is referred to as climate sensitivity. We will call this quantity ΔT_{2xCO2}.
- ΔT_{2xCO2} can be calculated from λ, using the radiative forcing for CO_2 doubling of $F_{2xCO2} \approx 3.7$ W/m^2

$$\Delta T_{2xCO2} = \lambda F_{2xCO2}$$

$$\uparrow$$

$$F_{CO2} \approx 5.4 \cdot \ln\left(\frac{[CO_2]}{[CO_2]_{ref}}\right) \quad \left[\frac{W}{m^2}\right]$$

Climate Sensitivity in the „Black-Body“ (BB) Model



Radiative equilibrium: $\pi R^2 S_0 (1-A) = 4\pi R^2 \sigma_{SB} T^4$

$$\underbrace{\frac{S_0}{4}(1-A)}_{S} = \underbrace{\sigma_{SB} T^4}_{E} \quad \Rightarrow \quad T_{BB} = \left(\frac{S_0(1-A)}{4\sigma_{SB}}\right)^{\frac{1}{4}} = \left(\frac{S}{\sigma_{SB}}\right)^{\frac{1}{4}}$$

S: Net solar input per m²

E: IR emission per m²

With A = 0.3 one obtains T_{BB} = 255 K

New equilibrium with forcing:

$$S + F = E(T) = \sigma_{SB} T^4 \quad \Rightarrow \quad T = \left(\frac{S+F}{\sigma_{SB}}\right)^{\frac{1}{4}}$$

$$S \approx \frac{1368}{4}(1-0.3) = 342 \cdot 0.7 \approx 239.4 \frac{W}{m^2}$$

$$\Rightarrow \quad \lambda = \frac{dT}{dF} = \frac{1}{4} \cdot \left(\frac{S+F}{\sigma_{SB}}\right)^{\left(\frac{1}{4}-1\right)} \cdot \frac{1}{\sigma_{SB}} = \frac{1}{4\sigma_{SB}} \cdot \underbrace{\left(\frac{S+F}{\sigma_{SB}}\right)^{\frac{1}{4}}}_{T} \cdot \underbrace{\left(\frac{S+F}{\sigma_{SB}}\right)^{-1}}_{1/T^4} = \frac{1}{4\sigma_{SB} T^3}$$

For $F \rightarrow 0$:

(F<<S)

$$\lambda = \frac{1}{4\sigma_{SB}} \cdot \left(\frac{S}{\sigma_{SB}}\right)^{-\frac{3}{4}} = \frac{1}{4\sigma_{SB} T_{BB}^3} \approx 0.27 \frac{K}{Wm^{-2}}$$

Better way to calculate λ:

Response of climate system

$$S + F = E(T) \quad \Rightarrow \quad F = E(T) - S = R(T)$$

$$\Rightarrow \quad \lambda^{-1} = \frac{dF}{dT} \underbrace{\approx}_{F \ll S} \frac{dR}{dT} = \frac{d}{dT}\left(E(T) - S\right) = \frac{dE(T)}{dT} = 4\sigma_{SB} T^3$$

$$\Rightarrow \quad \lambda = \frac{1}{4\sigma_{SB} T^3} \text{ (see above)}$$

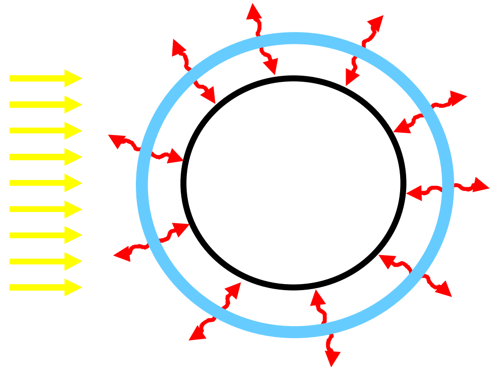
$$\text{Note:} \quad \lambda \propto \frac{1}{T^3}$$

However, Change in T is small!

Climate Sensitivity in Two-Layer Model

Radiative equilibrium for surface:

$$\underbrace{\frac{S_0}{4}(1-A)}_{\text{S: Solar input}} + \underbrace{\varepsilon_a \sigma_{SB} T_a^4}_{\text{B: Backradiation}} = \underbrace{\sigma_{SB} T_s^4}_{\text{E: Emission}}$$



New equilibrium with forcing (using $T = T_s = 2^{1/4}T_a = 288$ K and $\varepsilon_a = 0.77$):

$$S + B + F = E \quad \Rightarrow \quad F = R = E - S - B = \sigma_{SB} T^4 - \frac{S_0}{4}(1-A) + \frac{1}{2}\varepsilon_a \sigma_{SB} T^4$$

$$\Rightarrow \quad \lambda_{tot}^{-1} = \frac{dF}{dT} = \frac{dR}{dT} = \frac{dE}{dT} - \frac{dS}{dT} - \frac{dB}{dT} = \lambda_E^{-1} - \lambda_S^{-1} - \lambda_B^{-1}$$

with $\quad \lambda_E^{-1} = 4\sigma_{SB} T^3 = 5.41 \frac{W\,m^{-2}}{K}; \quad \lambda_S^{-1} = 0; \quad \lambda_B^{-1} = 2\varepsilon_a \sigma_{SB} T^3 = 2.08 \frac{W\,m^{-2}}{K}$

$$\Rightarrow \quad \lambda_{tot} = \frac{1}{\lambda_E^{-1} - \lambda_B^{-1}} = \frac{1}{5.41 - 2.08} = \frac{1}{3.33} = 0.30 \frac{K}{W\,m^{-2}} \equiv \lambda_0$$

The sensitivity for pure radiative forcing is $\lambda_0 = 0.30$ K/(W/m^2)

Climate Sensitivity and Feedbacks

With λ = 0.3 K/(W/m^2) and F_{2xCO2} ~ 3.7 W/m^2 we obtain:

$$\Delta T_{2xCO2} = \lambda F_{2xCO2} \approx 0.3 \cdot 3.7 \approx 1.1\ \text{K}$$

So what!?

IPCC: ΔT_{2xCO2} = 2.0 – 4.5 K (best estimate: 3.0 K)

What's wrong?

Problem: Amplifying feedbacks!

- Some climate system parameters such as Albedo A and IR-emissivity of the atmosphere ε_a are T-dependent

→ re-write balance equation:

$$F = R = E - S - B = \sigma_{SB} T^4 - \frac{S_0}{4}\left(1 - A(T)\right) - \frac{1}{2}\varepsilon_a(T)\sigma_{SB} T^4$$

Now:

$$\lambda_S^{-1} = \frac{dS}{dT} = -\frac{S_0}{4}\frac{dA}{dT}$$

and

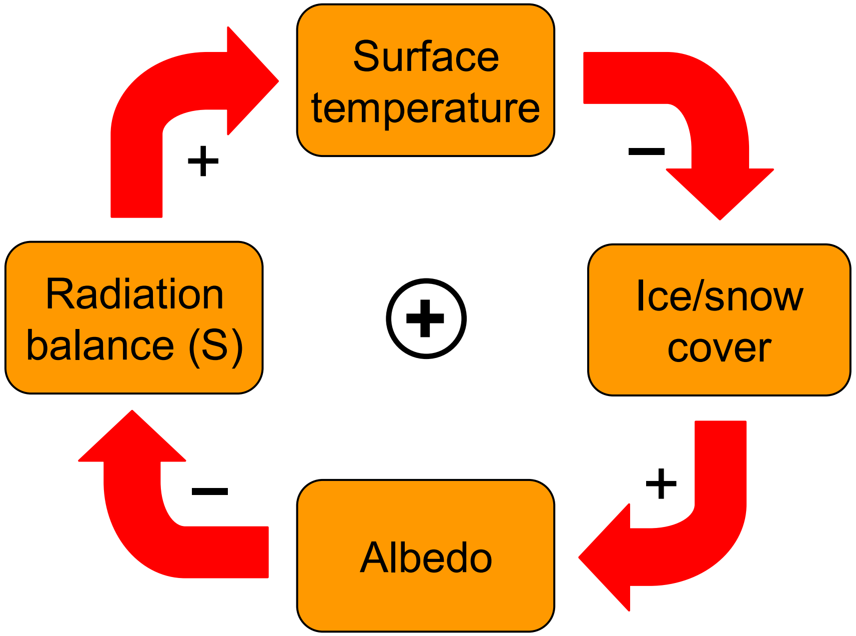
$$\lambda_B^{-1} = \frac{dB}{dT} = 2\varepsilon_a \sigma_{SB} T^3 + \underbrace{\frac{1}{2}\sigma_{SB} T^4 \frac{d\varepsilon_a}{dT}}_{\lambda_W^{-1}}$$

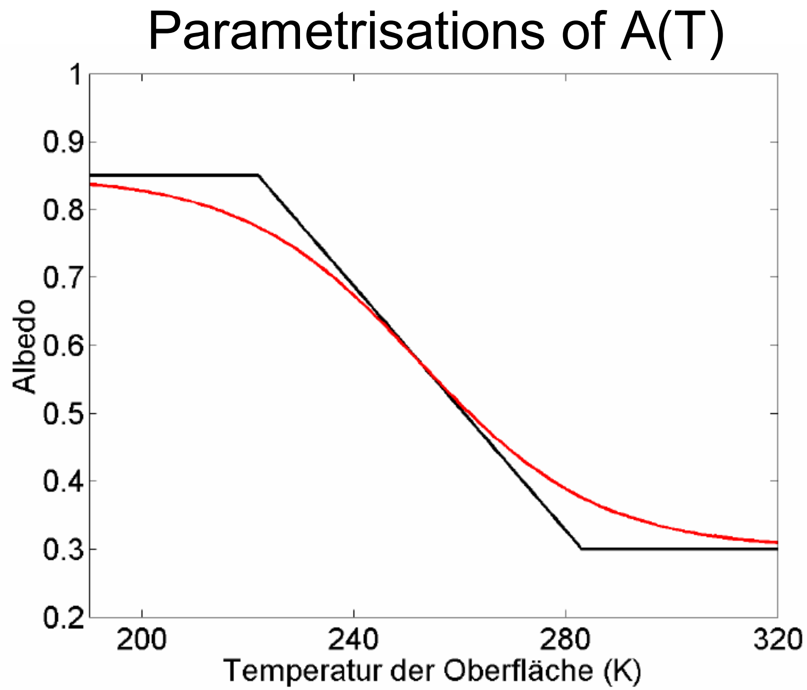
ice-albedo feedback ($\frac{dA}{dT}$)

cloud feedback ($\frac{dA}{dT}$ and $\frac{1}{2}\sigma_{SB}T^4\frac{d\varepsilon_a}{dT}$)

water vapour feedback ($\frac{d\varepsilon_a}{dT}$)

The Ice-Albedo Feedback





Parametrisation of A(T) proposed by Sellers, 1969 (J. Appl. Meteor. 8: 392-400):

$$A = 0.30 - 0.009 \cdot (T - 283K), \quad 222K \le T \le 283K$$

Thus:
$$\lambda_S^{-1} = -\frac{S_0}{4}\frac{dA}{dT} = 342\frac{W}{m^2} \cdot 0.009K^{-1} = 3.08\frac{W}{Km^2}$$

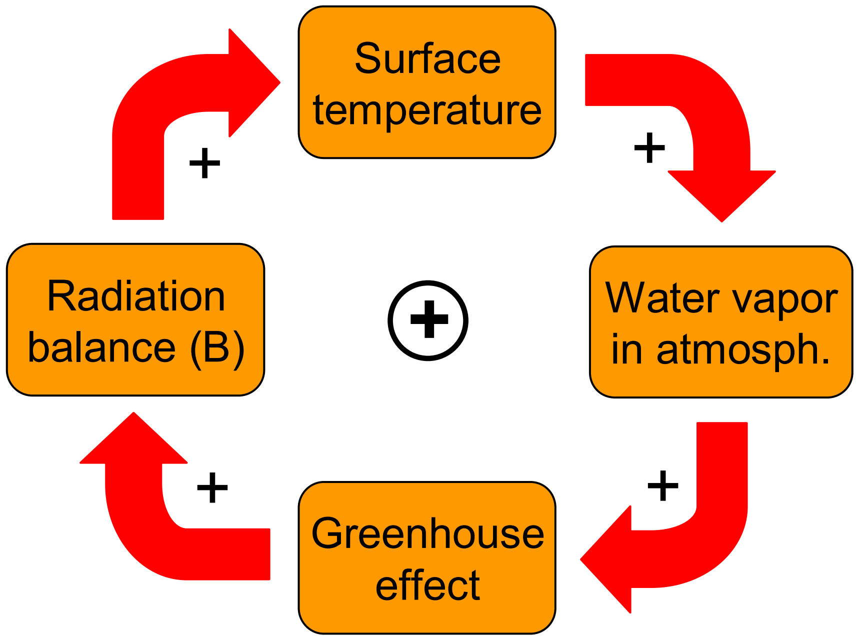
$$\Rightarrow \quad \lambda_{tot} = \frac{1}{\lambda_0^{-1} - \lambda_S^{-1}} = \frac{1}{3.33 - 3.08} = \frac{1}{0.25} = 4.0\frac{K}{W\,m^{-2}}$$

Purely radiative sensitivity

This is much too high!

after Stocker, 2009. Introduction to Climate Modelling. Lecture notes, University of Bern

Water Vapour Feedback



Water vapour pressure increases by 7 % per °C warming. ε_a should also increase, but it is difficult to estimate by how much (non-linear relationship and ε_a in two-layer model is not a realistic parameter).

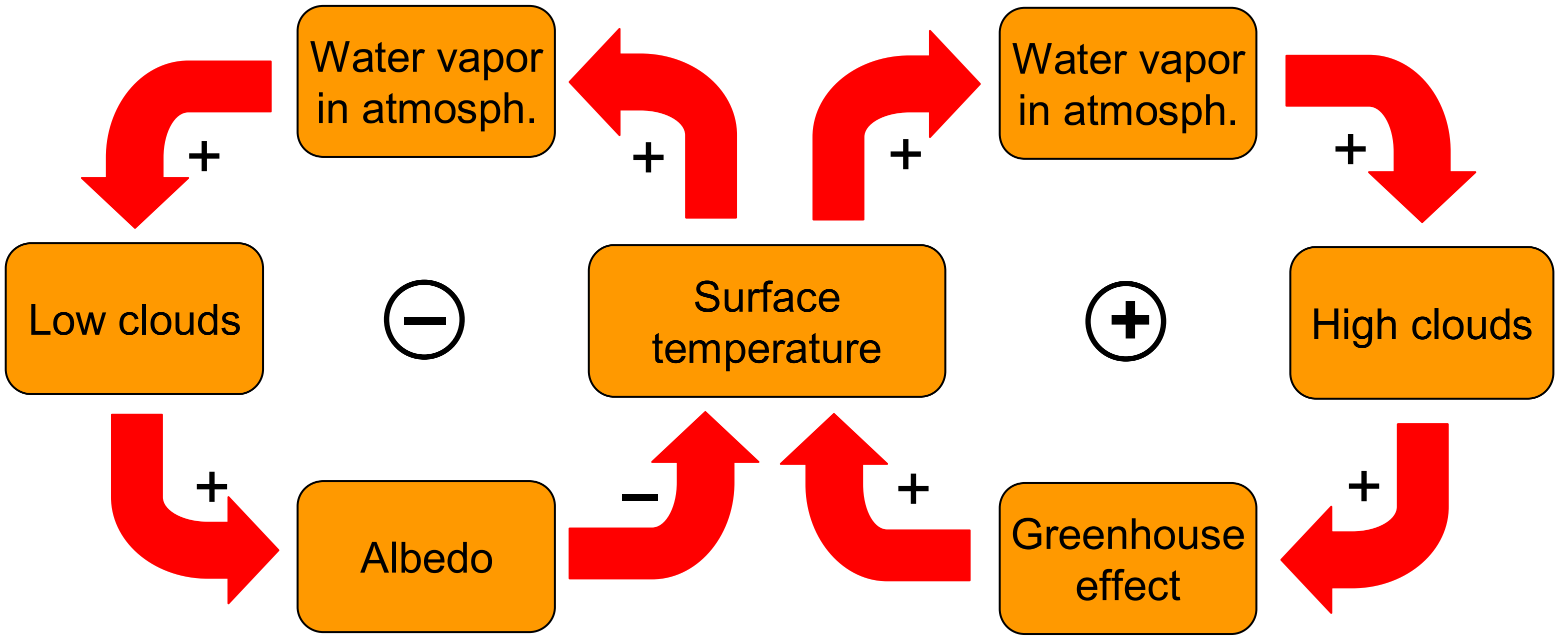
$d\varepsilon_a/dT = 0.009\ K^{-1}$ yields a result that is in agreement with results of complex climate models:

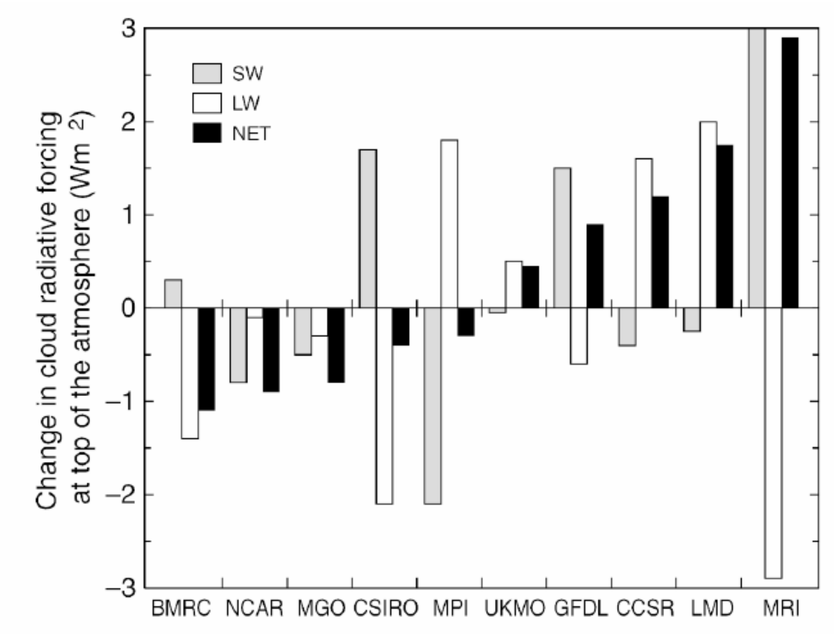
Thus: $$\lambda_W^{-1} = \frac{1}{2}\sigma_{SB}T^4\frac{d\varepsilon_a}{dT} = 1.75\frac{W}{K\,m^2}$$

$$\Rightarrow \quad \lambda_{tot} = \frac{1}{\lambda_0^{-1} - \lambda_W^{-1}} = \frac{1}{3.33 - 1.75} = \frac{1}{1.58} = 0.63\frac{K}{W\,m^{-2}}$$

Climate models have a significant sensitivity-enhancement due to the water vapour feedback. Verified by modeling effect of volcanoes.

Cloud Feedbacks



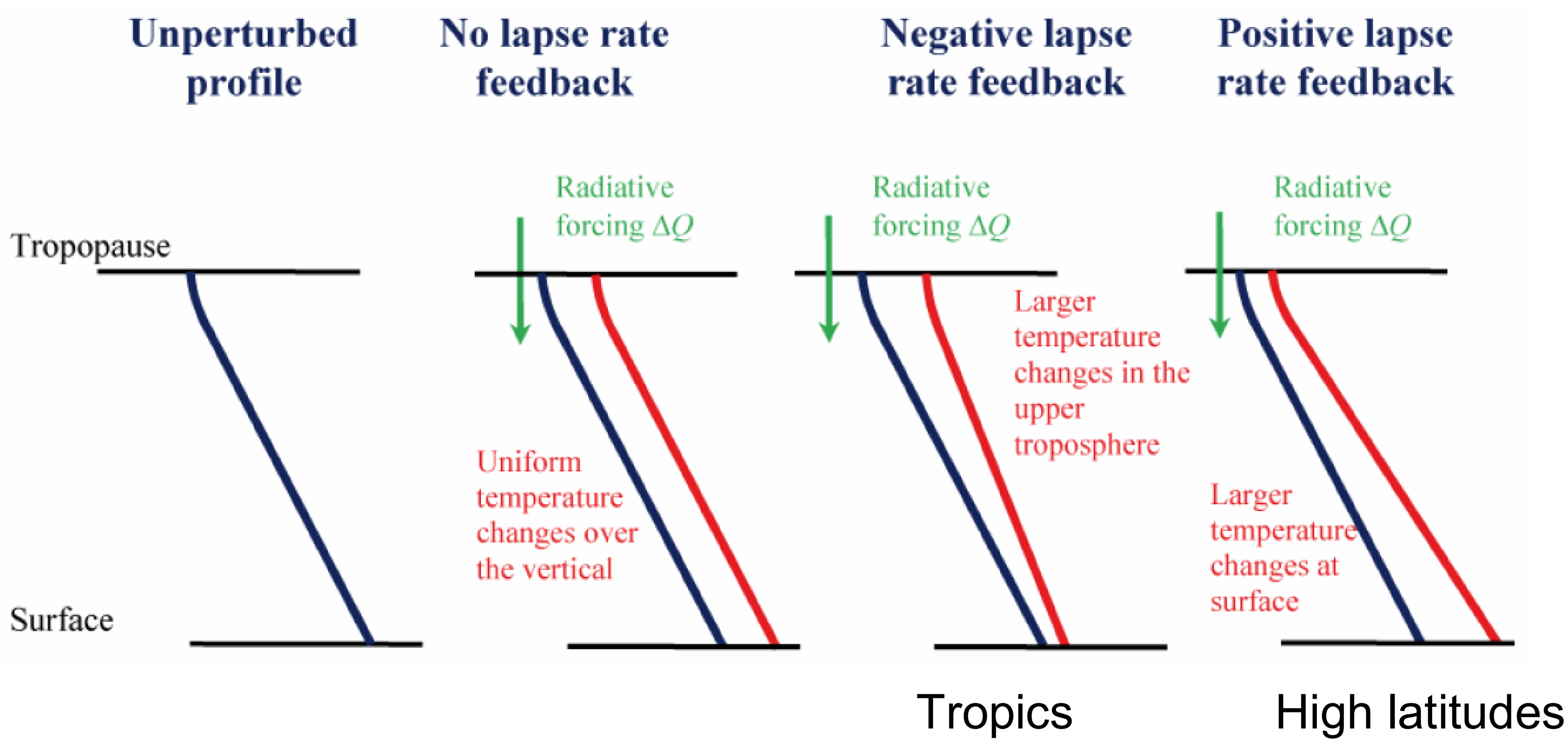


Climate models don't agree on the sign of the net cloud feedback. Figure: Cloud feedback (LW/SW/net) for $2xCO_2$ in 10 GCMs.

From IPCC 2007 and Stocker 2009.

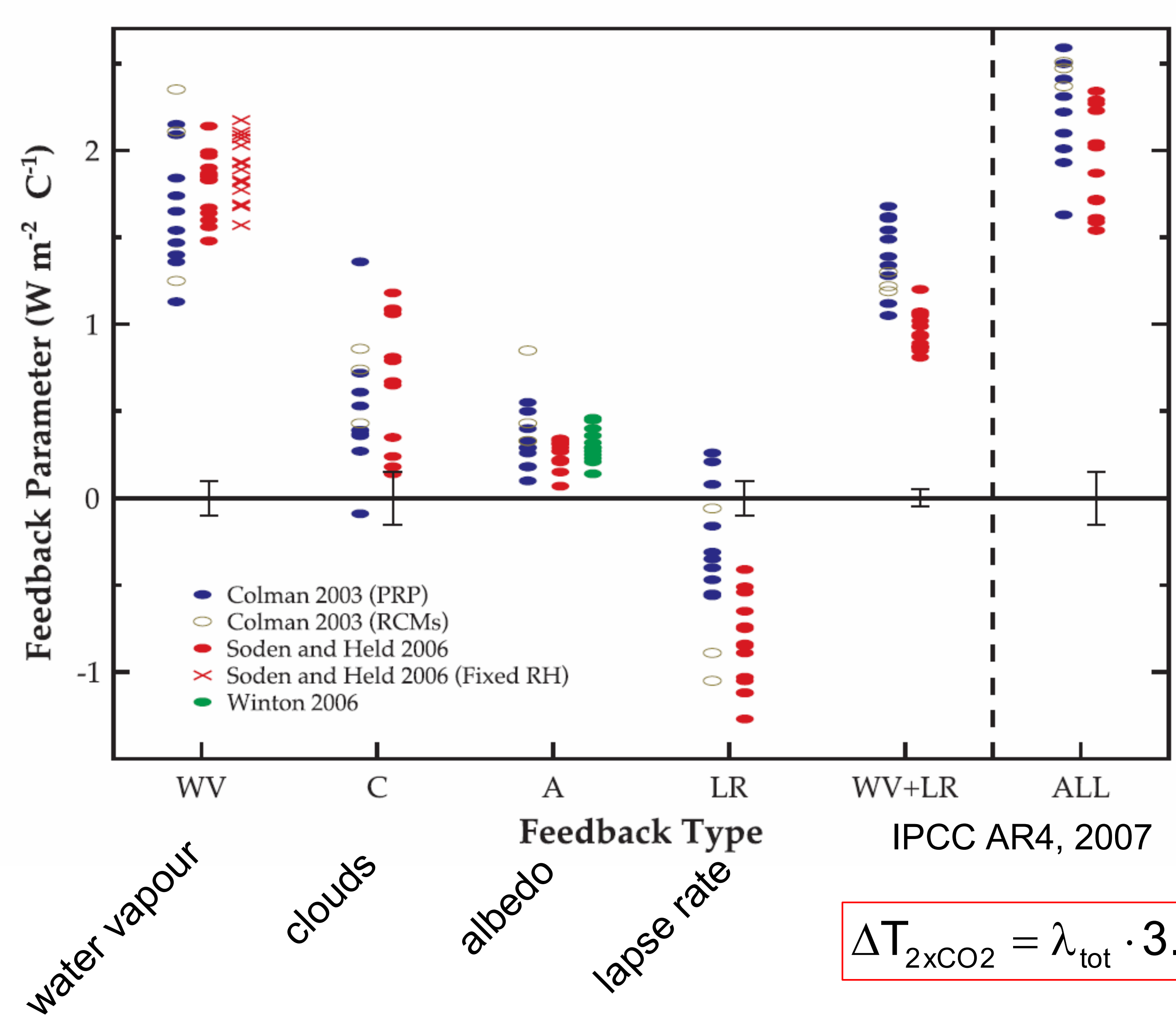
Lapse Rate Feedback

Changes of the lapse rate (T-gradient in the troposphere) also produce a feedback on radiative forcing. Overall estimated to be a negative feedback.



Goosse et al., 2010. Online textbook, http://www.climate.be/textbook

Feedbacks in Climate Models



IPCC AR4, 2007

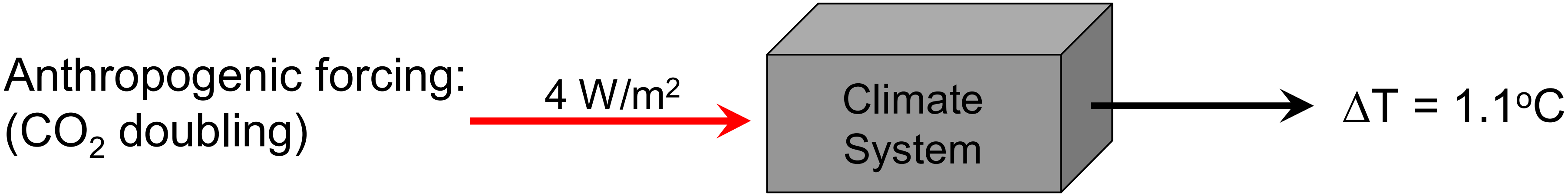
$$\lambda_{All}^{-1} \approx 2 \frac{W\,m^{-2}}{K}$$

$$\lambda_{tot} = \frac{1}{\lambda_0^{-1} - \lambda_{All}^{-1}} \approx \frac{1}{3.33 - 2} = \frac{1}{1.33} = 0.75 \frac{K}{W\,m^{-2}}$$

$$\Delta T_{2xCO2} = \lambda_{tot} \cdot 3.7\,Wm^{-2} = 2.8\,K$$

Effect of CO_2 Doubling without Feedbacks

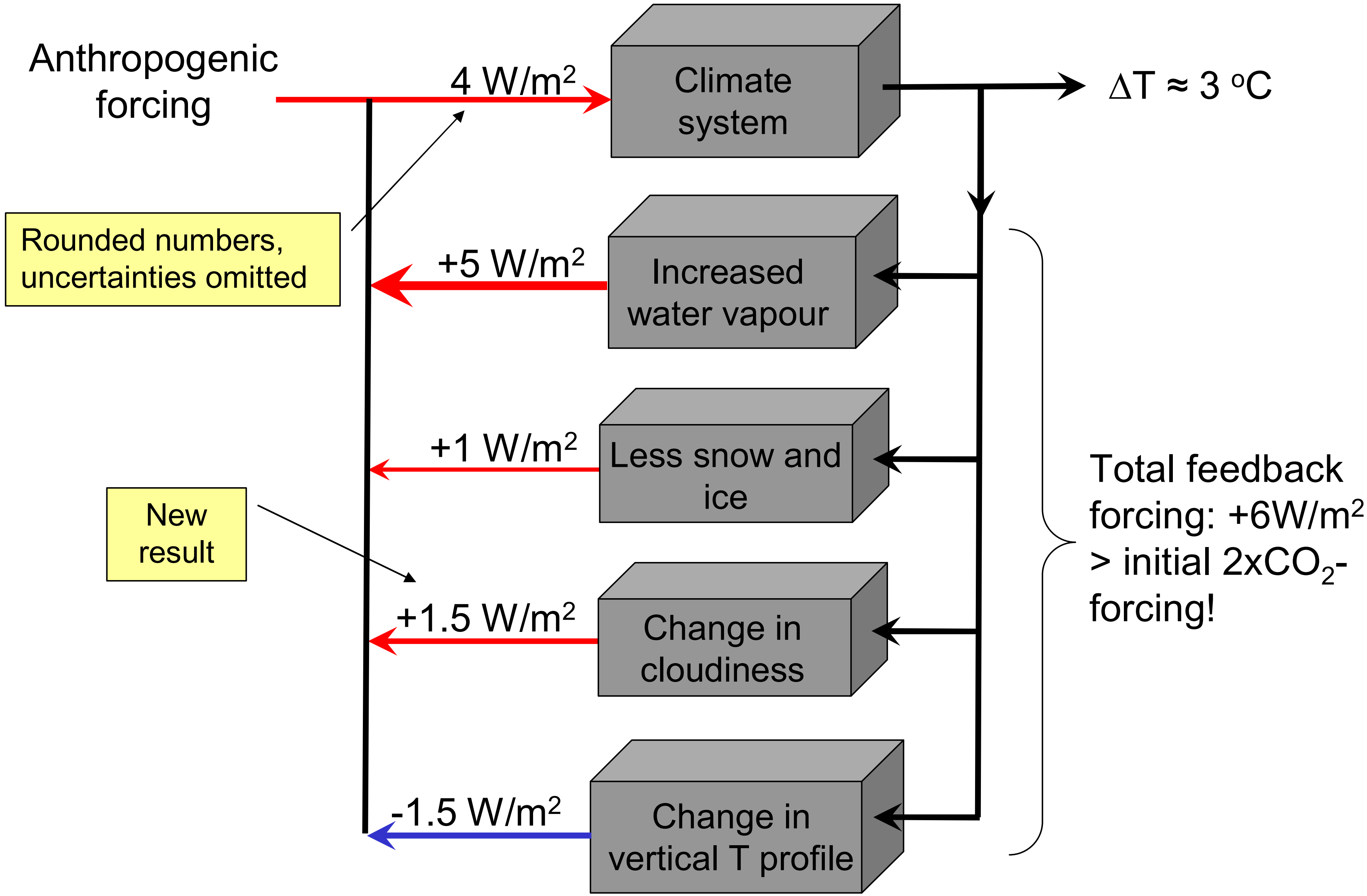
No feedback:



Without feedbacks, climate prediction would be rather simple.

Feedbacks complicate climate modeling and add uncertainty to predictions of global warming.

Effect of CO_2 Doubling with Feedbacks



Transient Temperature Response

Equilibrium warming for doubling of CO_2 is estimated at about 3 °C (2 to 4.5 °C, IPCC, AR4, 2007)

But how quickly does the climate system respond?

Assume a system at equilibrium at T_0: $R(T_0) = E(T_0) – S – B(T_0) = 0$

For $t \geq 0$, a constant forcing F is applied → temperature $T = T_0 + \Delta T(t)$

Net heat input Q to climate system is given by radiation imbalance:

$$Q = F - R(T_0 + \Delta T) \underset{\uparrow}{\approx} F - \Big[\underbrace{R(T_0)}_{0} + \underbrace{\frac{dR}{dT}}_{\lambda^{-1}} \Delta T \Big] = F - \frac{\Delta T}{\lambda}$$

linearisation 0 λ^{-1} (total sensitivity)

This heat input leads to warming, with C = effective heat capacity:

$$\frac{dE_{th}}{dt} = C\frac{d(\Delta T)}{dt} = Q = F - \frac{\Delta T}{\lambda} \quad \text{or} \quad \frac{d}{dt}\Delta T = \frac{F}{C} - \frac{1}{\lambda C}\Delta T$$

Solution: $\Delta T(t) = \lambda F\left(1 - e^{-\frac{t}{\lambda C}}\right) = \Delta T_\infty\left(1 - e^{-\frac{t}{\tau}}\right)$

$$\Delta T_\infty = \lambda F$$
$$\tau = \lambda C$$

Heat Capacity and Response Time

Effective heat capacity, sensitivity, and response time of the climate system are coupled through $\tau = \lambda C$

What is the heat capacity of the climate system?

C is the heat capacity per unit surface area [J m^{-2} K^{-1}].
For a uniform material with density ρ and height h it is given by

$C = h\rho c_P$ with c_P = heat capacity of the material [J kg^{-1} K^{-1}]

Atmosphere: h ≈ 8000 m, ρ ≈ 1.3 kg m^{-3}, c_P ≈ 1000 J kg^{-1} K^{-1}:

C = 1.0·10^{7} J m^{-2} K^{-1} → with λ = 0.30 K W^{-1} m^{2}: τ = 3.0·10^{6} s ≈ 35 d

→ with λ = 0.75 K W^{-1} m^{2}: τ = 7.5·10^{6} s ≈ 90 d

Upper ocean: h ≈ 100 m, ρ ≈ 1000 kg m^{-3}, c_P ≈ 4180 J kg^{-1} K^{-1}:

C = 4.2·10^{8} J m^{-2} K^{-1} → with λ = 0.30 K W^{-1} m^{2}: τ = 1.3·10^{8} s ≈ 4 a

→ with λ = 0.75 K W^{-1} m^{2}: τ = 3.1·10^{8} s ≈ 10 a

Simple model: If the effective heat capacity and the response time of the climate system were known, climate sensitivity could be determined.

But: Real climate system has multiple response times!

Climate Engineering: Response Times

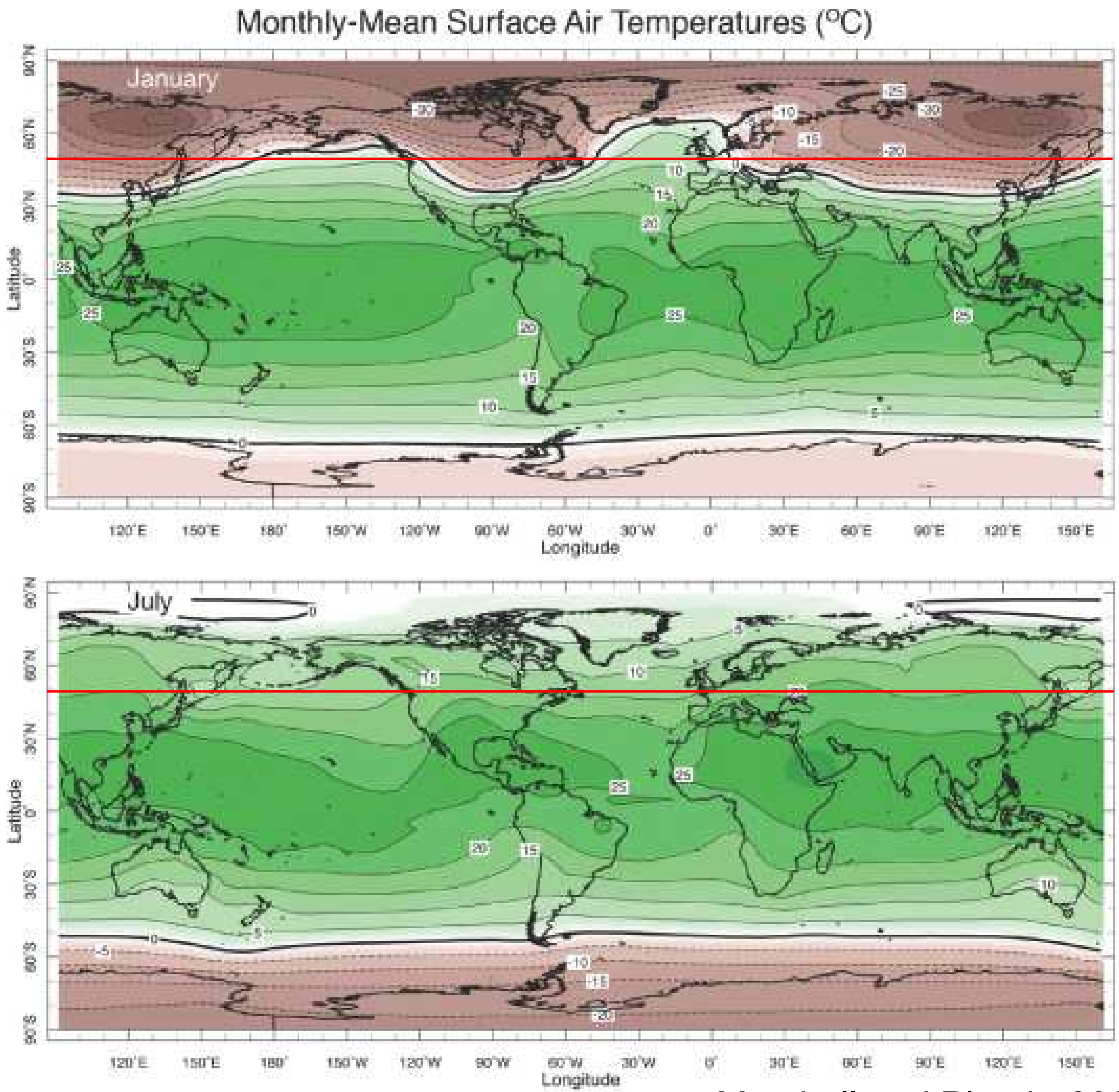
There are two response times:

1) Response of the CE-measure itself, for example:

- cloud whitening: days
- Stratospheric aerosol: ≈1-2 years
- CO_2-removal: depends on C-cycle, years to decades

2) Response of the climate system to CE-forcing
(or no more CE-forcing after CE-measures stopped).

Climatic Effects of Ocean Heat Capacity

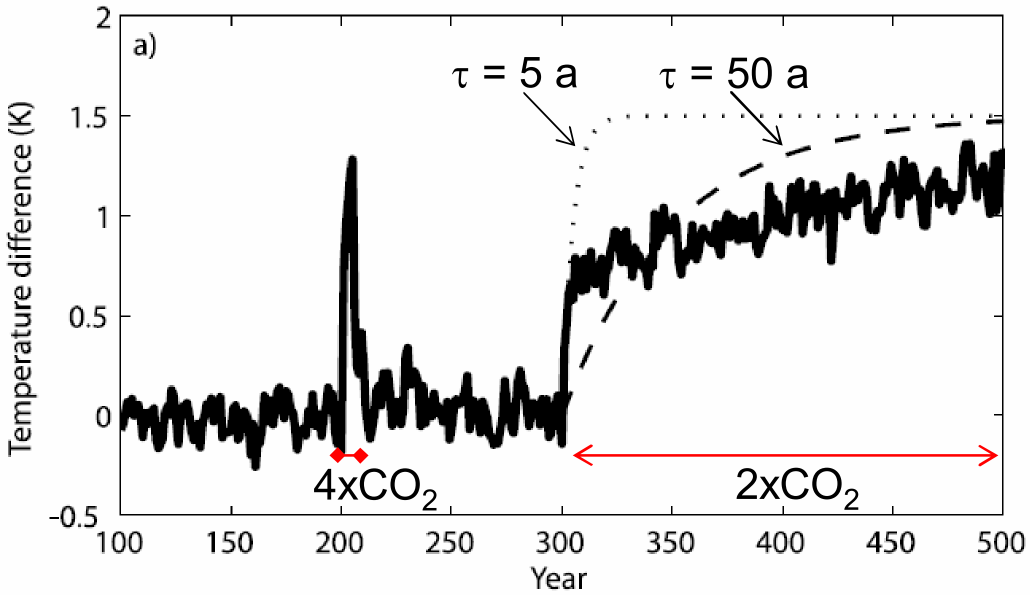


Mid-latitudes:
W-coast: maritime
E-coast: continental

Seasonal variation of local radiation balance:
Continents react much faster than oceans.

Marshall and Plumb, 2008

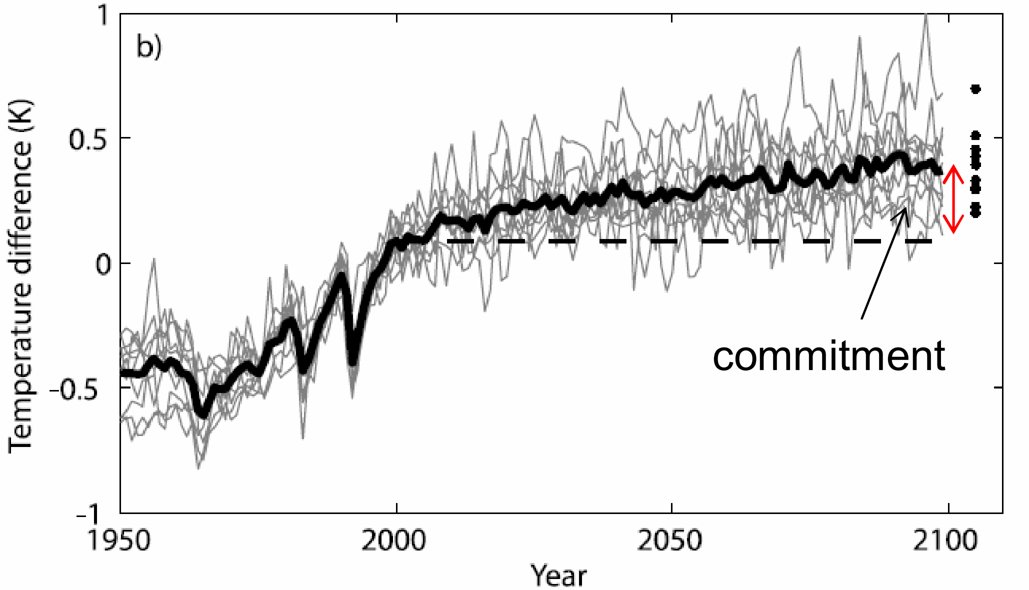
Response Times in Climate Models



Step-forcing experiment with one climate model.

$2xCO_2$ for t > 300 a.

→ No equilibrium at t = 500 a.



T-evolution 1950 -2100 in 10 climate models.

Forcing (CO_2) assumed to be constant after 2000.

Increase in 21st century: "Constant composition warming commitment"

Knutti et al., 2008, J. Geophys. Res., 113, D15103, doi:10.1029/2007JD009473

Sensitivity and Response in Climate Models

AOGCM	Equilibrium climate sensitivity (°C)	Transient climate response (°C)
1: BCC-CM1	n.a.	n.a.
2: BCCR-BCM2.0	n.a.	n.a.
3: CCSM3	2.7	1.5
4: CGCM3.1(T47)	3.4	*1.9*
5: CGCM3.1(T63)	*3.4*	n.a.
6: CNRM-CM3	n.a.	1.6
7: CSIRO-MK3.0	3.1	1.4
8: ECHAM5/MPI-OM	3.4	2.2
9: ECHO-G	3.2	1.7
10: FGOALS-g1.0	*2.3*	*1.2*
11: GFDL-CM2.0	2.9	1.6
12: GFDL-CM2.1	3.4	1.5
13: GISS-AOM	n.a.	n.a.
14: GISS-EH	2.7	1.6
15: GISS-ER	2.7	1.5
16: INM-CM3.0	2.1	1.6
17: IPSL-CM4	4.4	2.1
18: MIROC3.2(hires)	4.3	2.6
19: MIROC3.2(medres)	4.0	2.1
20: MRI-CGCM2.3.2	3.2	2.2
21: PCM	2.1	1.3
22: UKMO-HadCM3	3.3	2.0
23: UKMO-HadGEM1	4.4	1.9

Equilibrium climate sensitivity (ECS) and transient climate response (TCR) of the AOGCMs assessed in the 4th IPCC report.

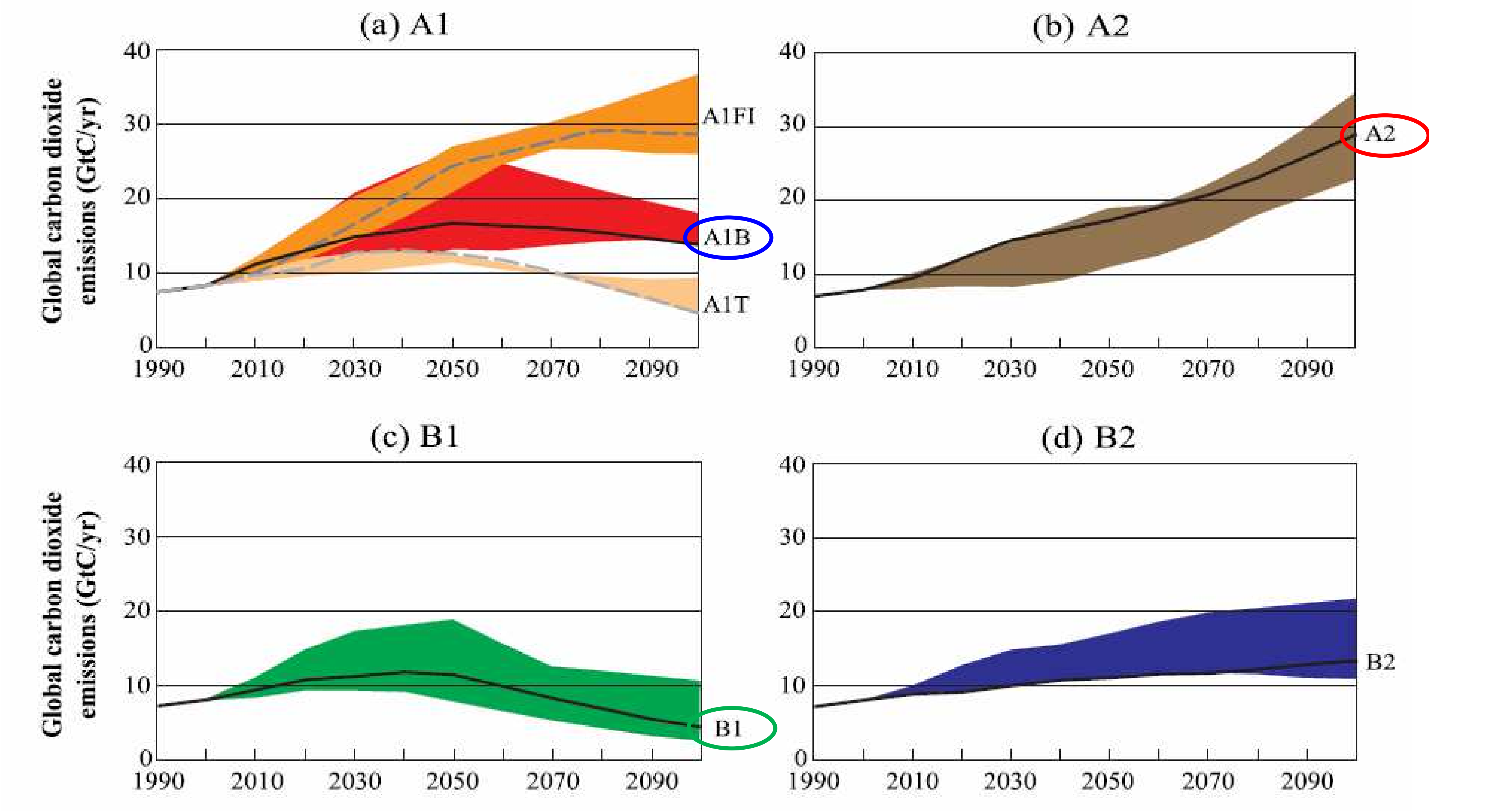
Equilibrium climate sensitivity is what we called ΔT_{2xCO2}.

Transient climate response is defined as the global annual mean surface air temperature change averaged over a 20-year period centred at the time of CO_2 doubling in a 1% yr^{-1} CO_2 increase scenario.

IPCC AR4, 2007

Emission Scenarios

Basis of climate projections: Scenarios of greenhouse gas emission.
Standard: IPCC, 2000, Special Report on Emissions Scenarios (SRES)

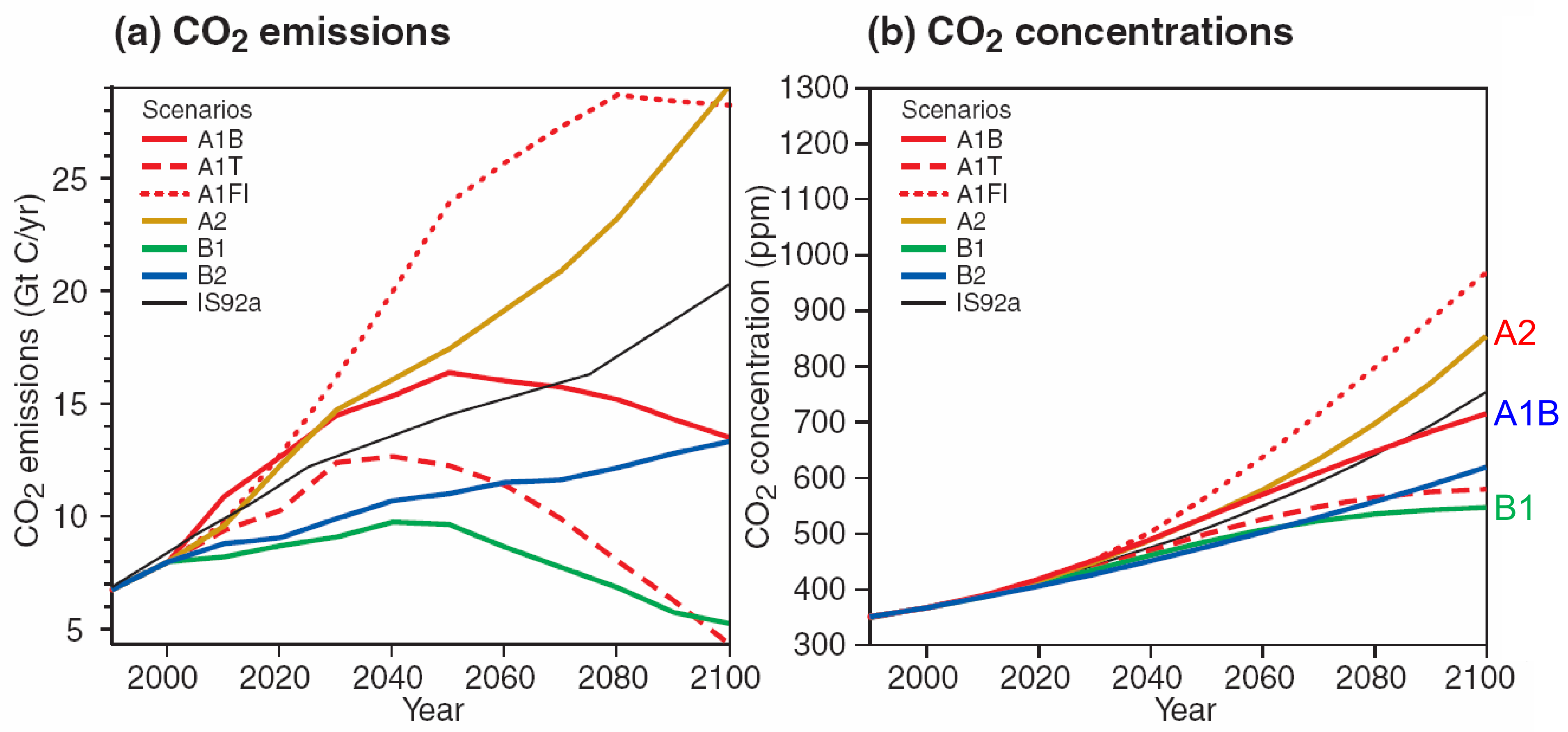


http://www.ipcc.ch/publications_and_data/publications_and_data_reports.shtml

Emissions and Concentrations

IPCC-Scenarios for CO_2 emissions and concentrations

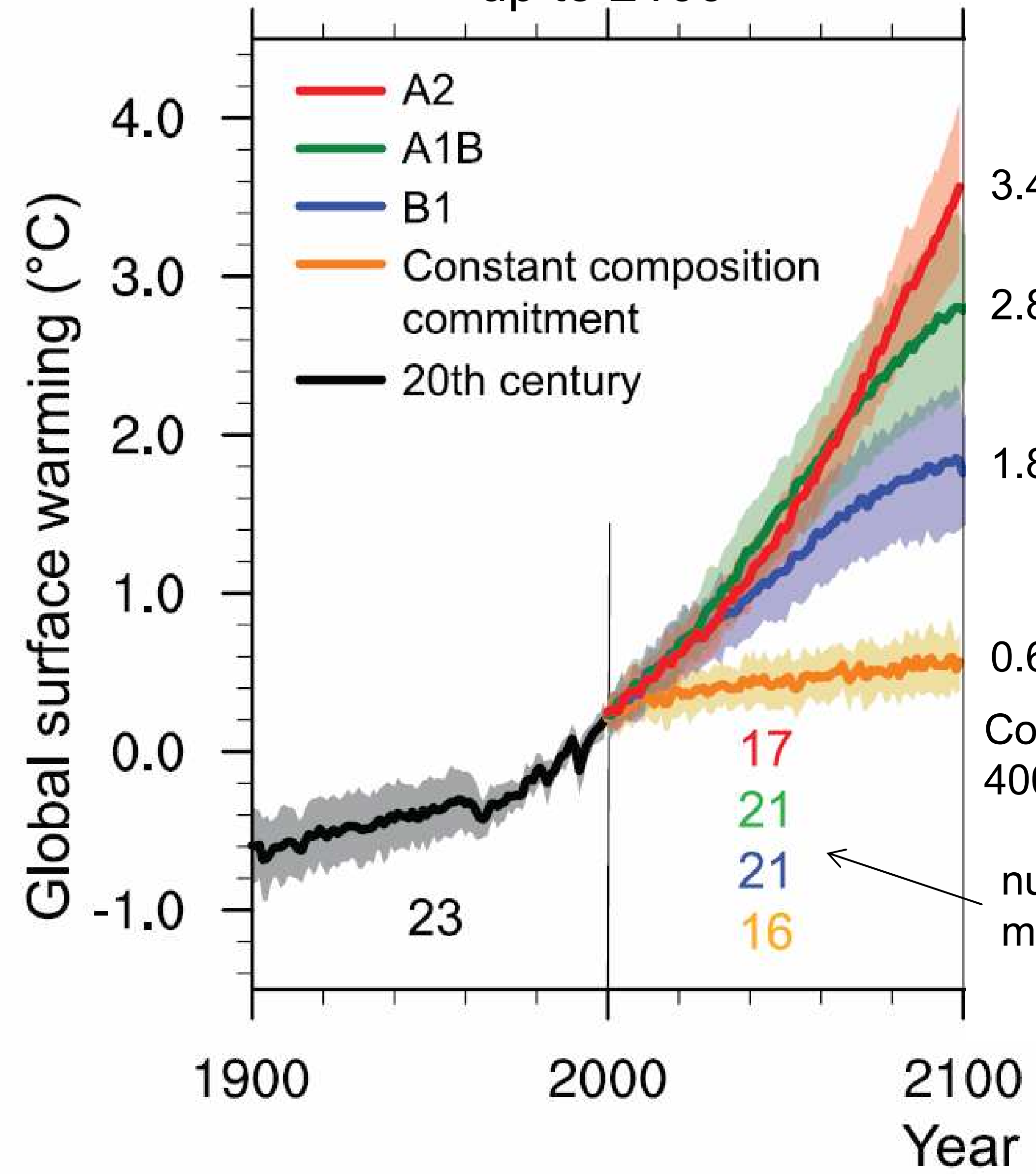
→ CO_2-concentration will increase significantly



IPCC TAR, 2001

Climate Predictions

up to 2100



3.4°C A2

2.8°C A1B

1.8°C B1

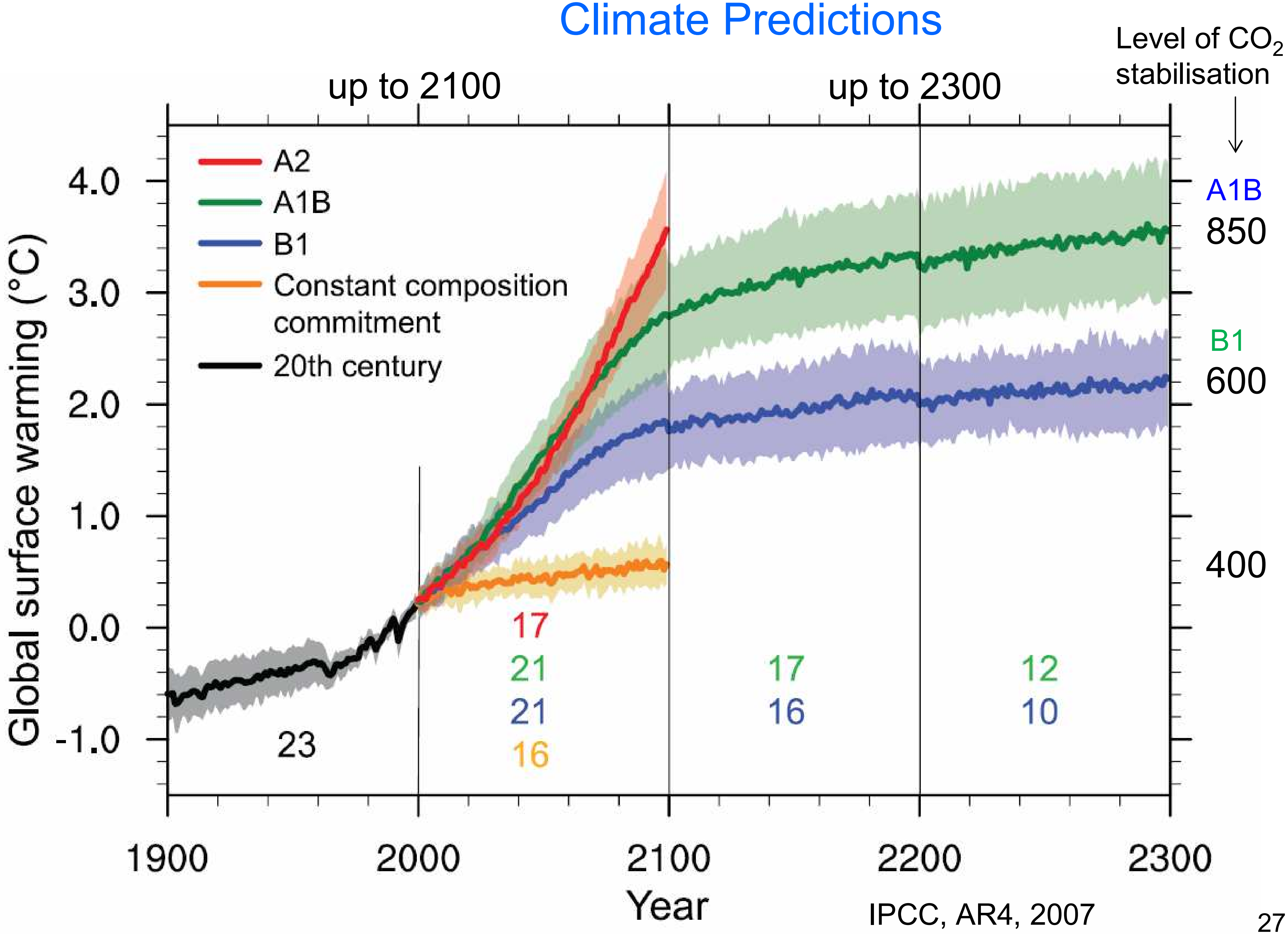
0.6°C constant CO_2

Commitment at 400 ppm CO_2

number of models

IPCC, AR4, 2007

Climate Predictions



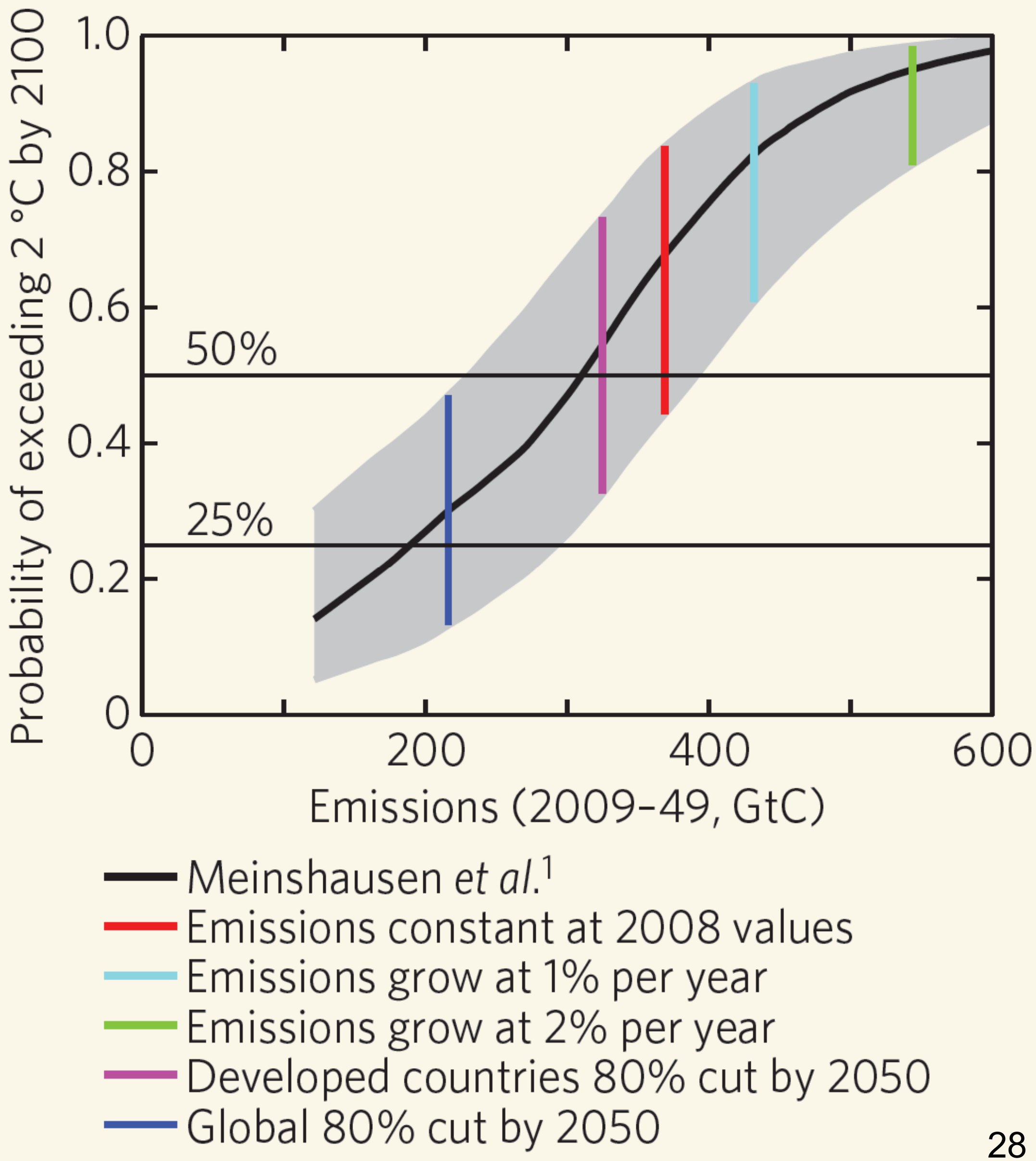
IPCC, AR4, 2007

The "2°C Lottery"

EU Policy:
To prevent the most severe impacts of climate change, … the world needs to limit global warming to **no more than 2ºC** above the pre-industrial temperature.
http://ec.europa.eu/clima/policies/brief/eu/

Black line: probability of peak global mean temperature exceeding 2°C above pre-industrial levels before the year 2100 vs. integrated emissions 2009 to 2049.

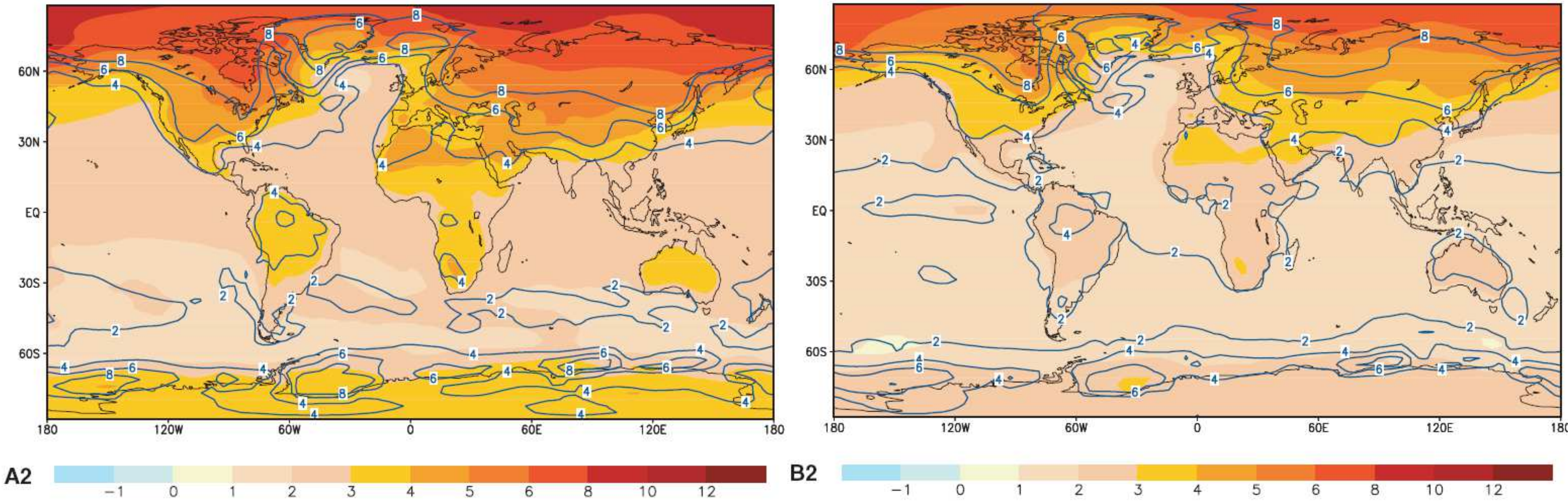
Schmidt & Archer, 2009, Nature 458: 1117-1118, adapted from Meinshausen et al., 2009 Nature 458: 1158-1163.



Predicted Regional Climate Change

A2: strong CO_2 increase
2100: ~850 ppm

B2: moderate CO_2 increase
2100: ~600 ppm

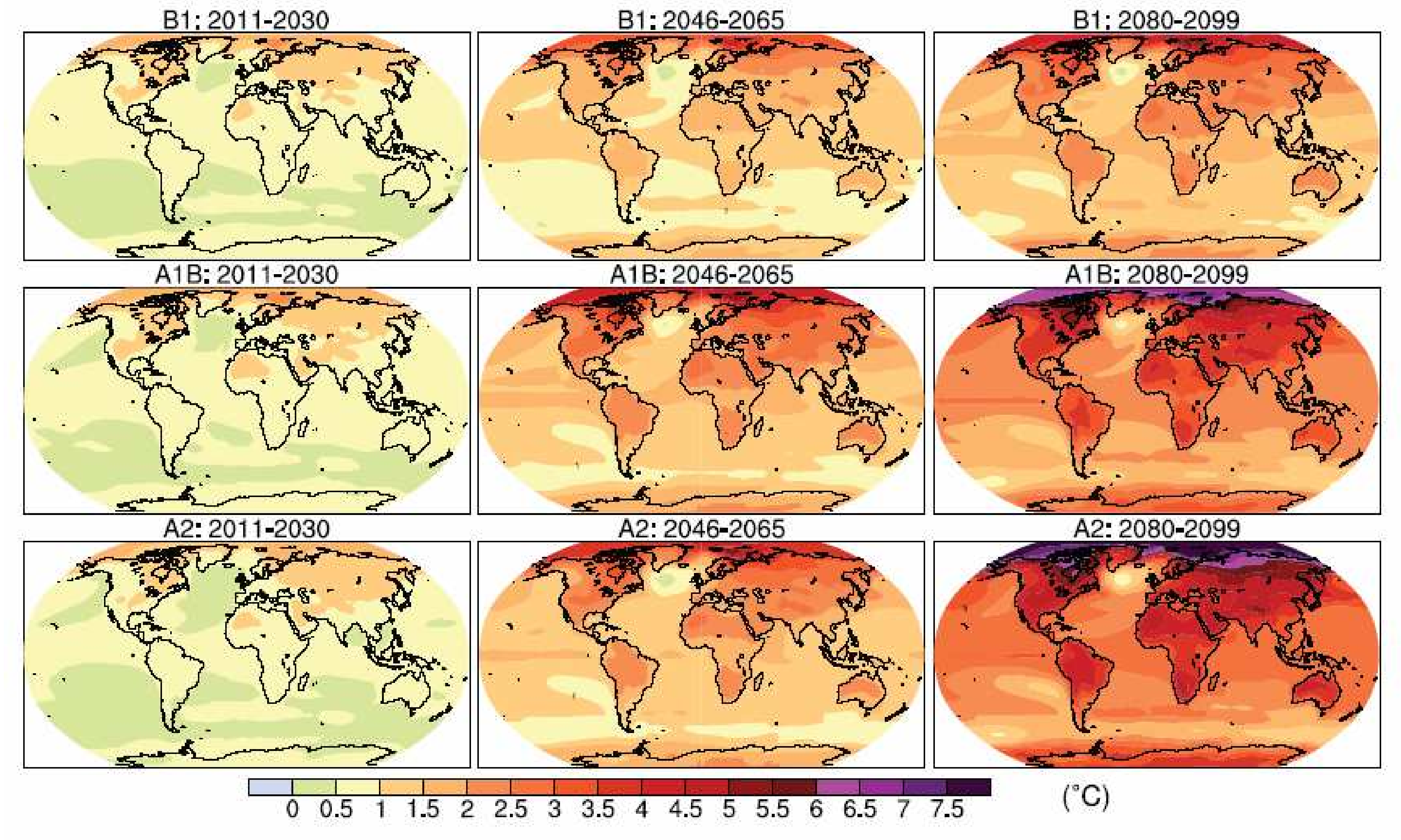


Annual multi-model mean change of temperature (colour shading) and its range (isolines, units: °C) from OAGCMs, for the period 2071 to 2100 relative to the period 1961 to 1990.

Patterns are similar, even though scenarios are very different.
Strong enhancement of warming in high latitudes!

IPCC, TAR, 2001

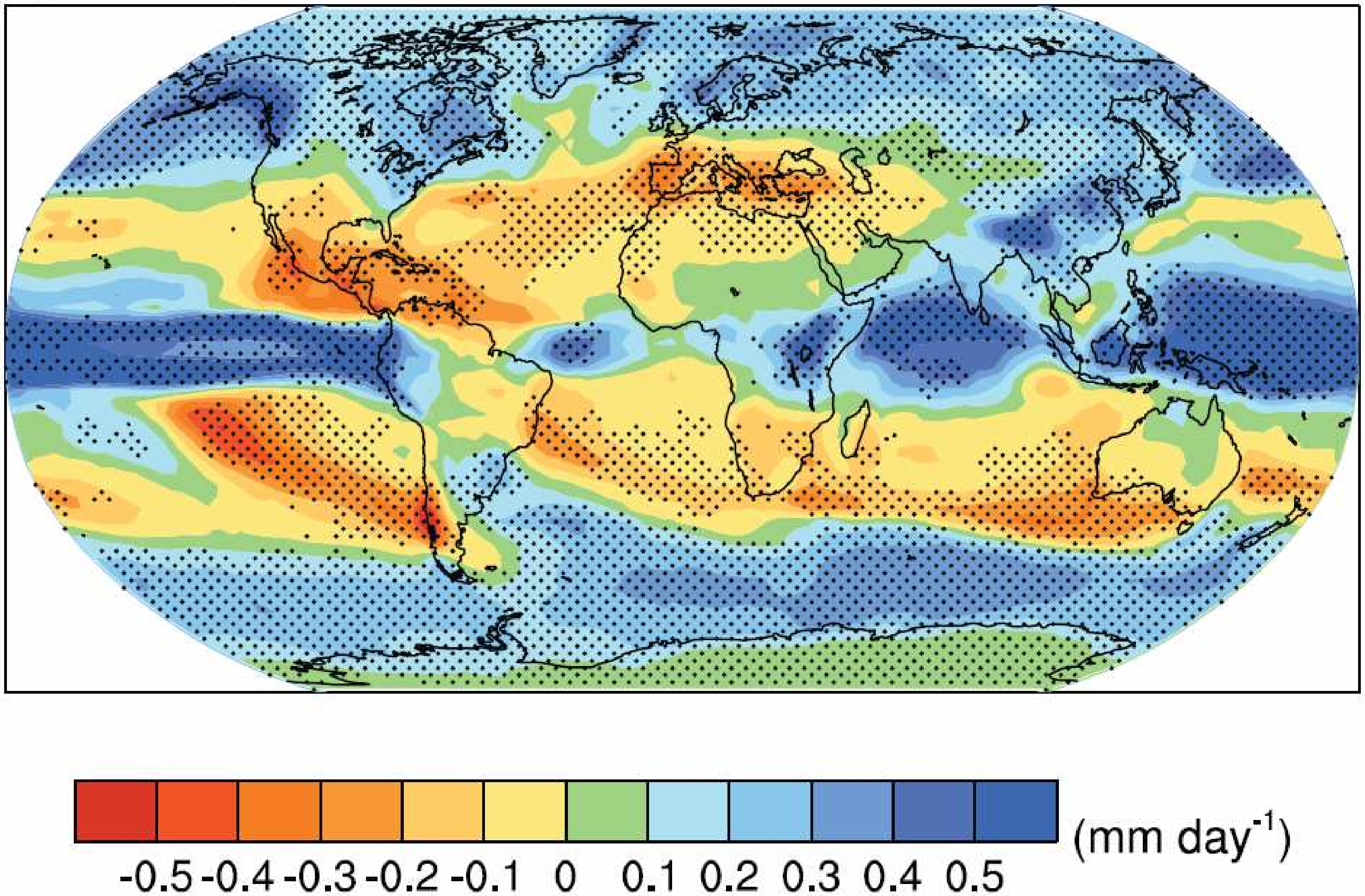
Predicted Warming Relative to 1980-1999



IPCC, AR4, 2007

Predicted Change in Precipitation

Changes for A1B scenario for the period 2080 - 99 relative to 1980 - 99.
Stippled areas: > 80 % of models agree on the sign of change.

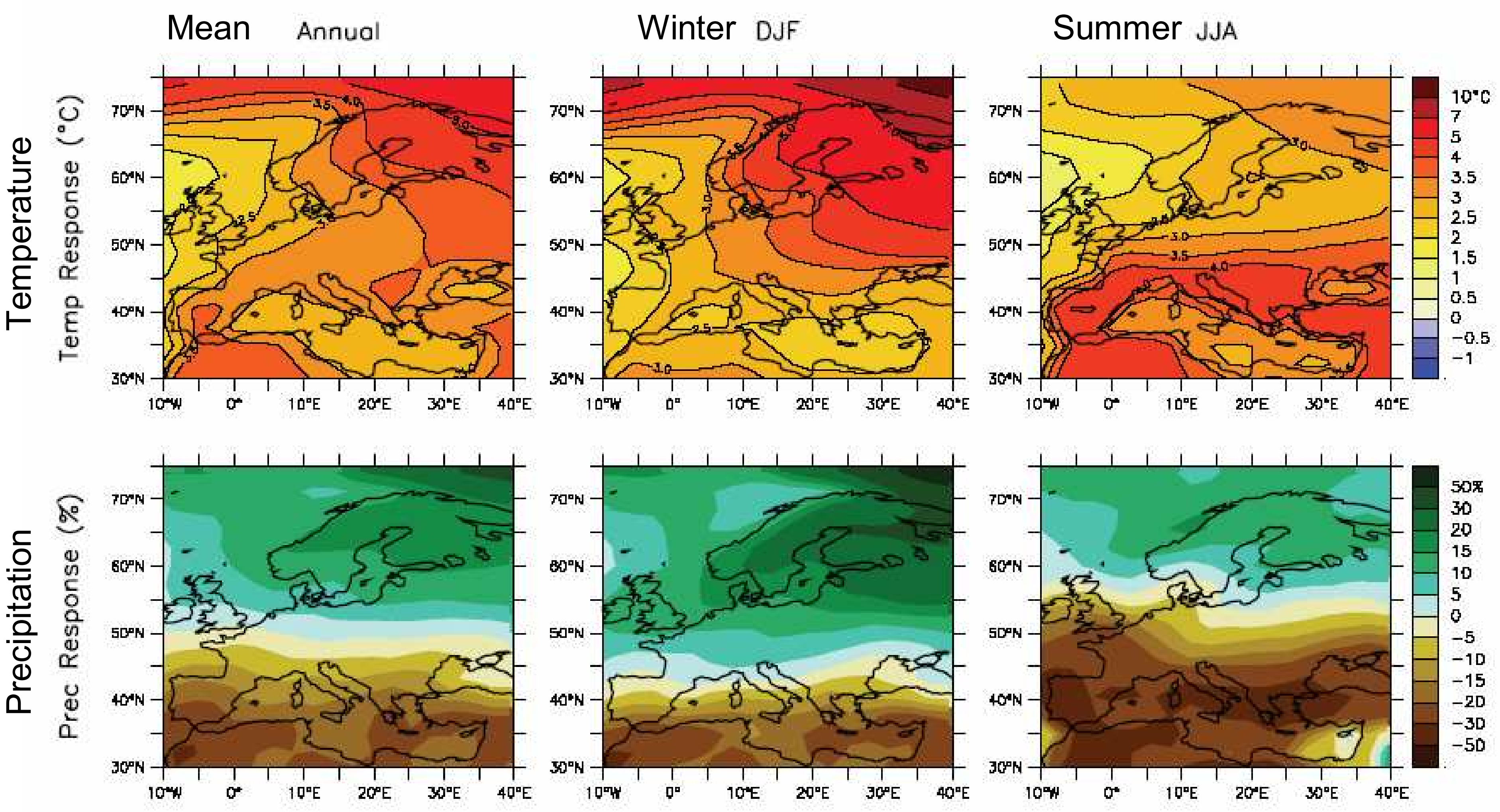


Drying in much of the subtropics, more rain at higher latitudes.

IPCC, AR4, 2007

Predicted Changes in Europe

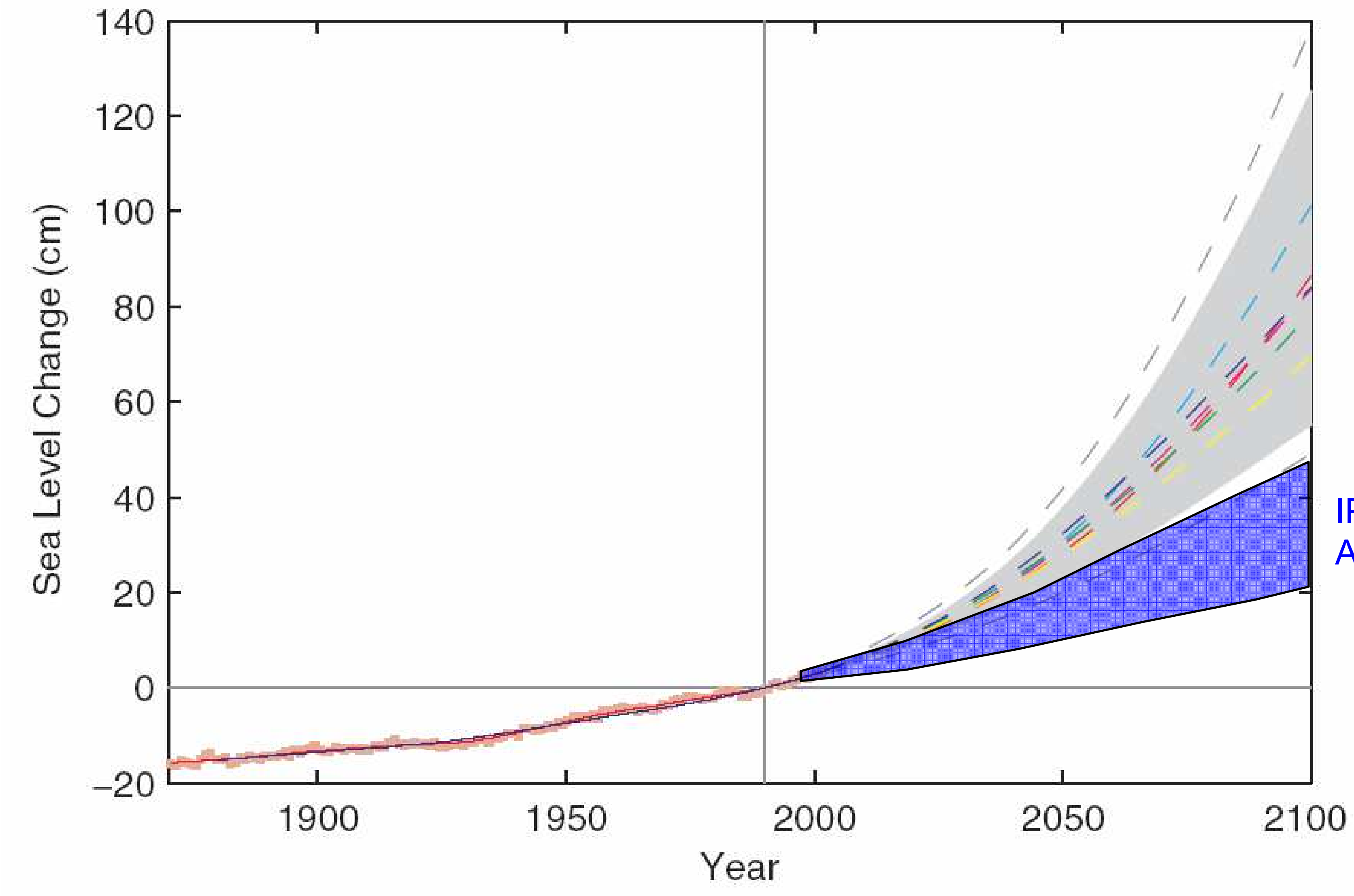
Changes for A1B scenario for the period 2080 - 99 relative to 1980 - 99.



IPCC, AR4, 2007

Predictions of Sea Level Rise

Sea level is predicted to continue to increase, but by how much is debated

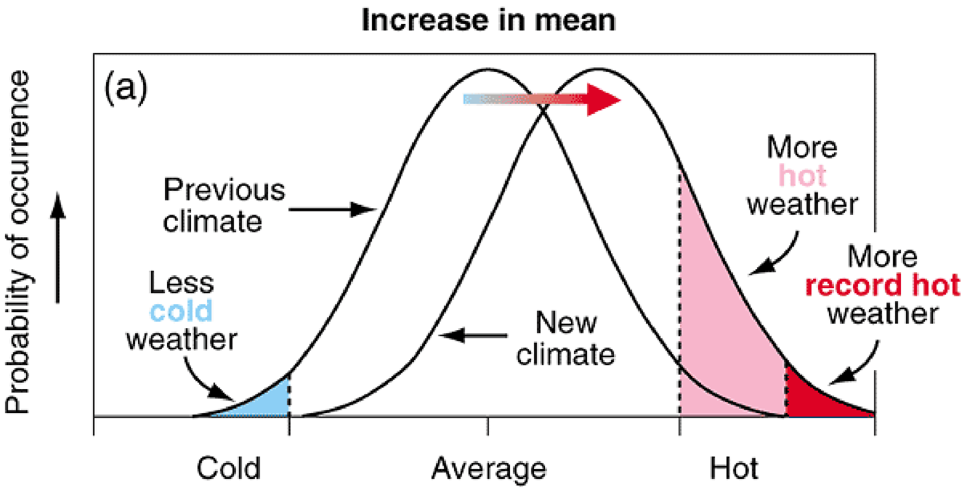


IPCC, AR4, 2007

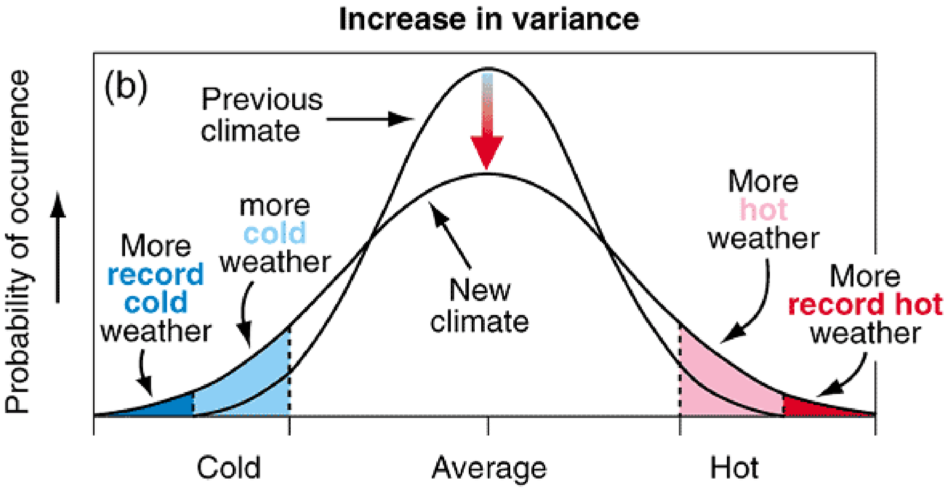
Rahmstorf, 2007, Science 315: 368-370

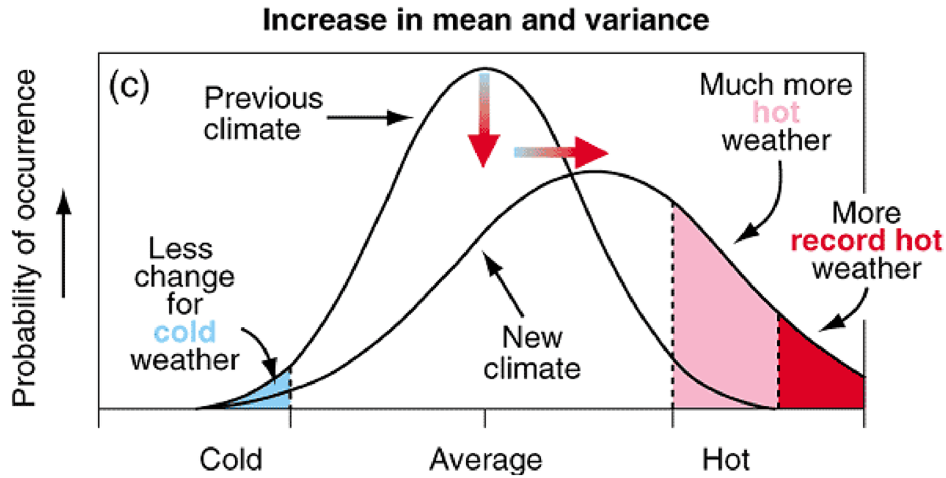
Mean, Variance, and Extremes

The mean of climate variables such as temperature and precipitation is expected to shift. The variability is expected to increase.



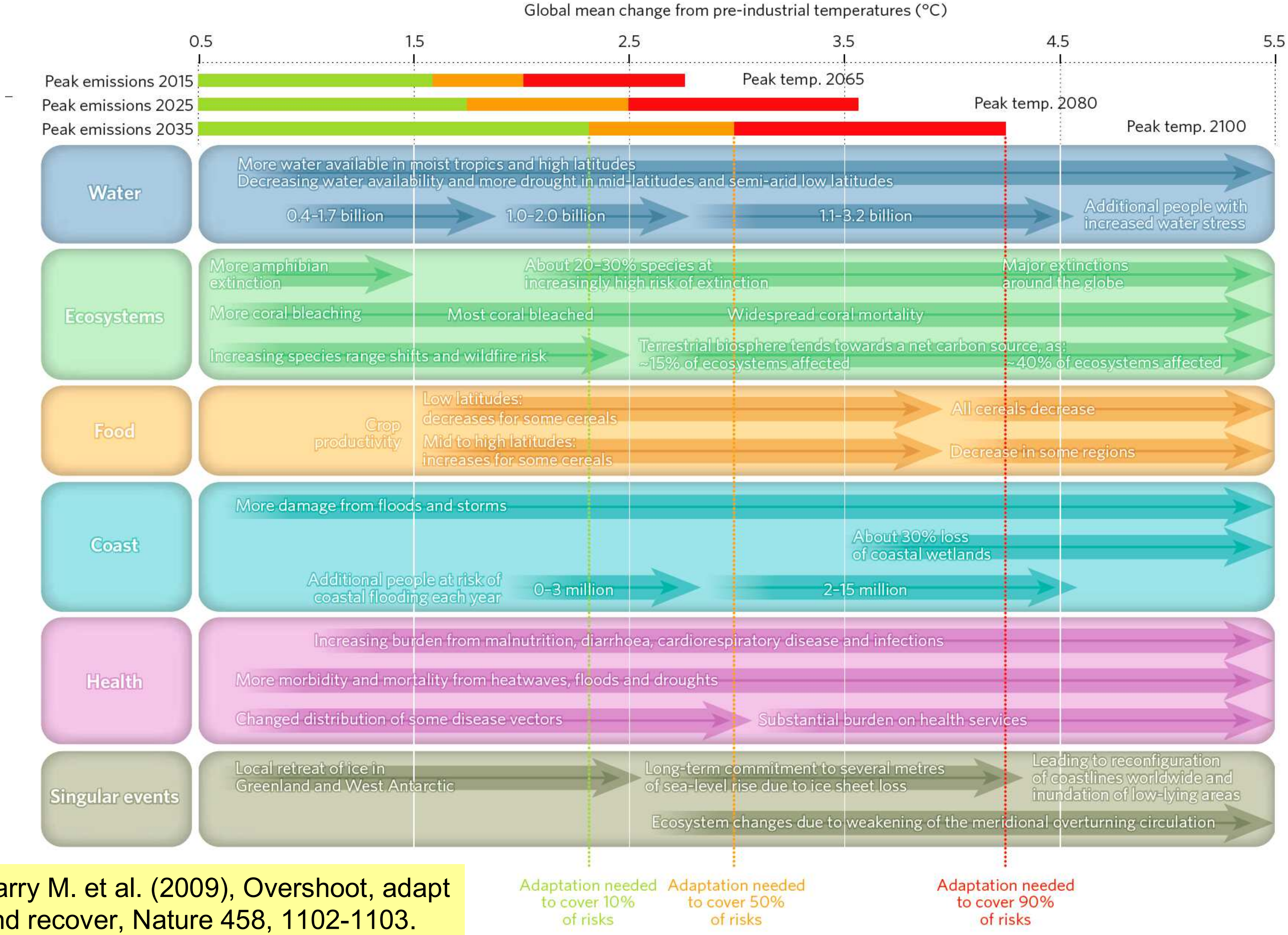
Both effects lead to a **disproportionate** increase of the frequency of extreme events (heat waves, floods, droughts,…)





IPCC, TAR, 2001

Consequences of Climate Change



Parry M. et al. (2009), Overshoot, adapt and recover, Nature 458, 1102-1103.

Natural vs. Anthropogenic Climate Change

Important questions related to climate change predictions:

- How significant is the anthropogenic influence on climate compared to natural drivers?
- Is the warming of the 20th century extraordinary? Can it be attributed to anthropogenic forcing?
- What is the natural variability of the climate system?
- What is the sensitivity of the climate system? How strong are the feedbacks?
- What are the impacts of climate change?

- Answers to these highly debated questions may come from information on the climate in the past (paleoclimate)

Summary

- Climate sensitivity: Measure for warming per forcing
- Sensitivity depends strongly on feedbacks
  - Without feedbacks: l = 0.30 K / (W m^{-2}), DT_{2xCO2} = 1.1 K
  - Main feedbacks: Water vapour, lapse rate, albedo, clouds
  - With all feedbacks (in models): l ≈ 0.75 K / (W m^{-2}), DT_{2xCO2} ≈ 3 K
- Transient climate response time: t = lC in simple model, but multiple time constants in the real climate system
- Climate predictions (IPCC summary of model results):
  - Warmer, esp. in high latitude regions
  - Dryer in subtropics, wetter in mid to high latitudes
  - Higher variability, more extreme events

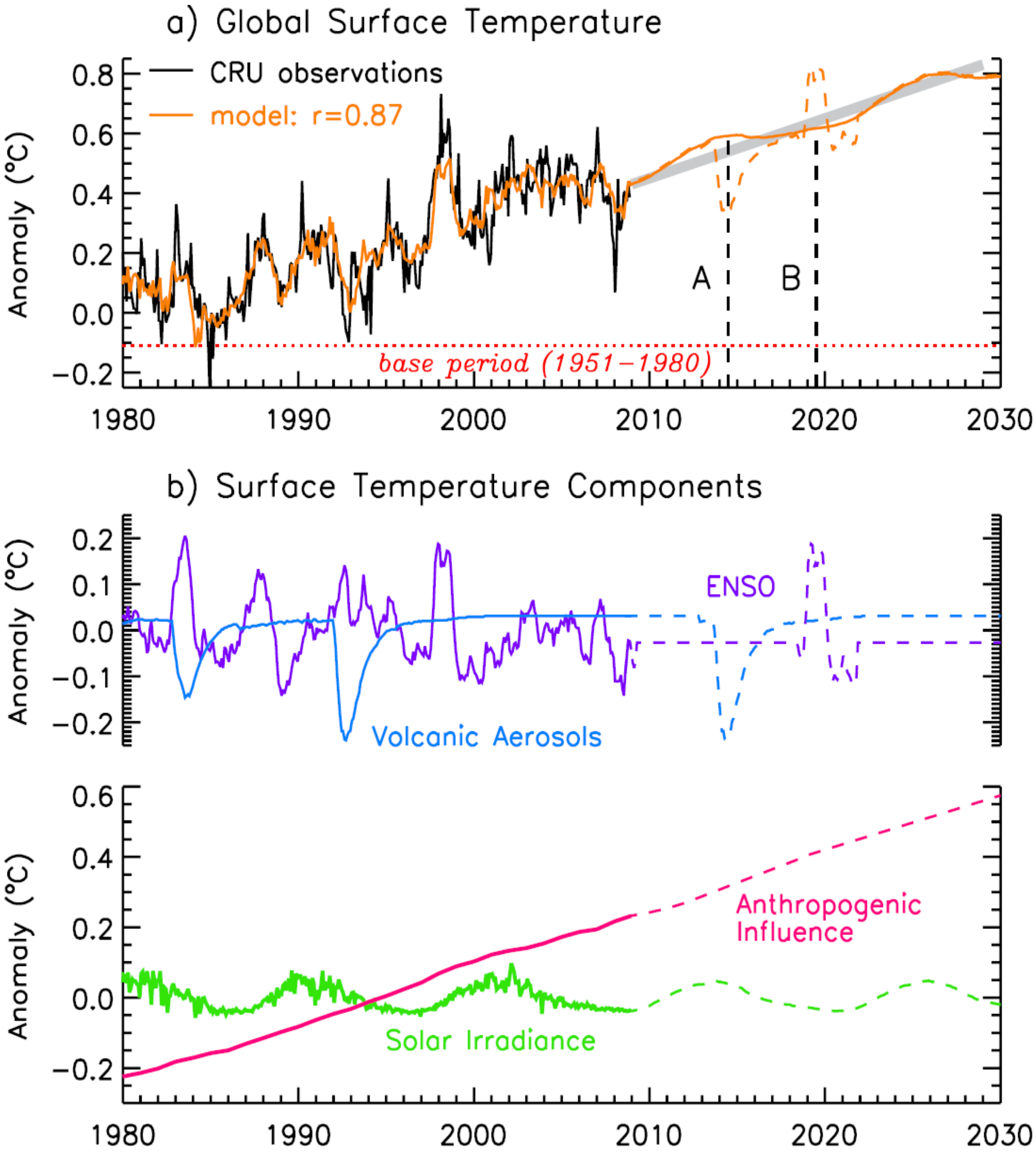
Do we see Global Warming?

The observed global Temperature evolution can be well explained by 4 main components (see e.g. Lean and Rind 2009):

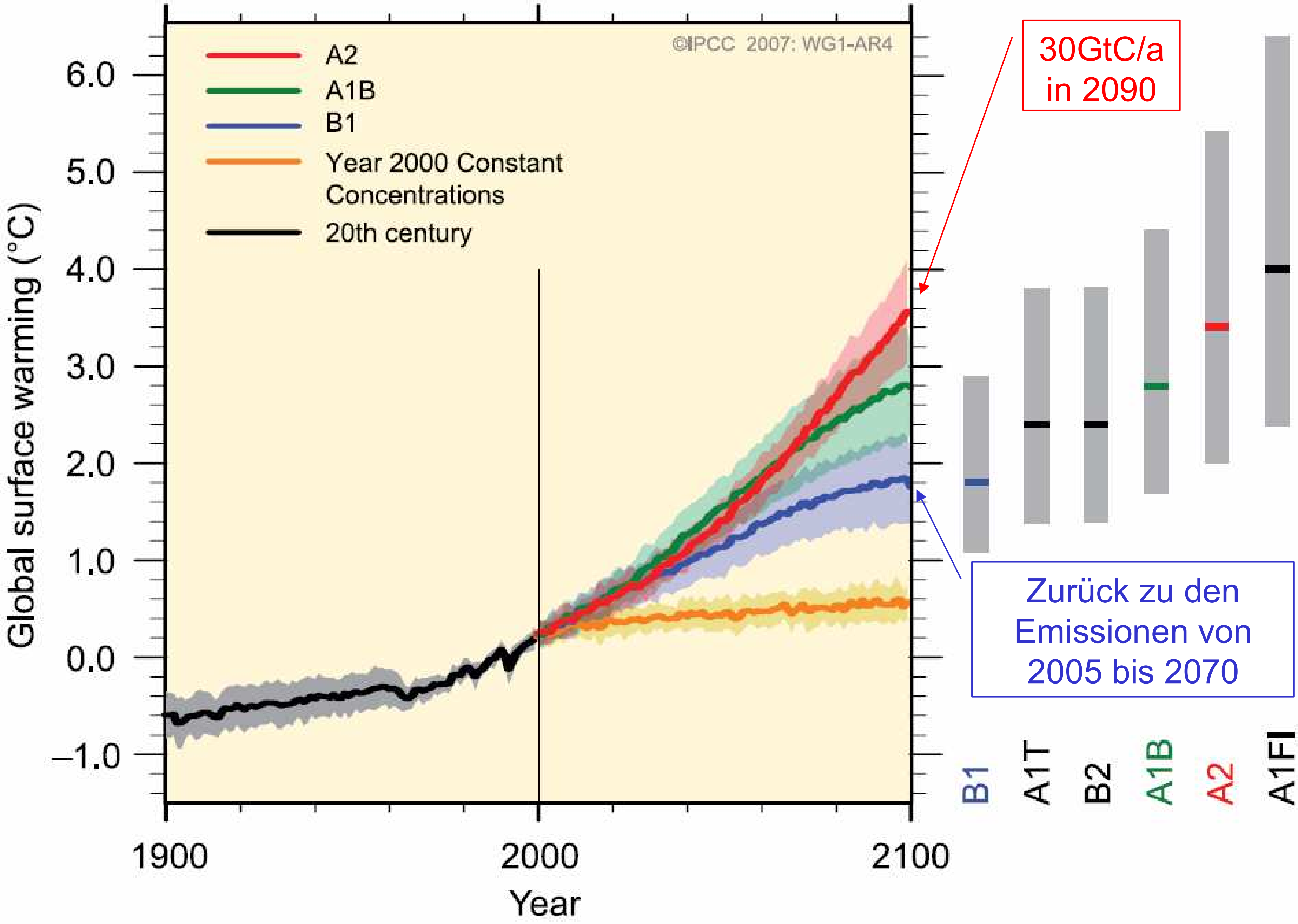
1) Anthropogenic Eff.
2) Volcanic aerosol
3) ENSO
4) Solr variability

However: How important is this question?

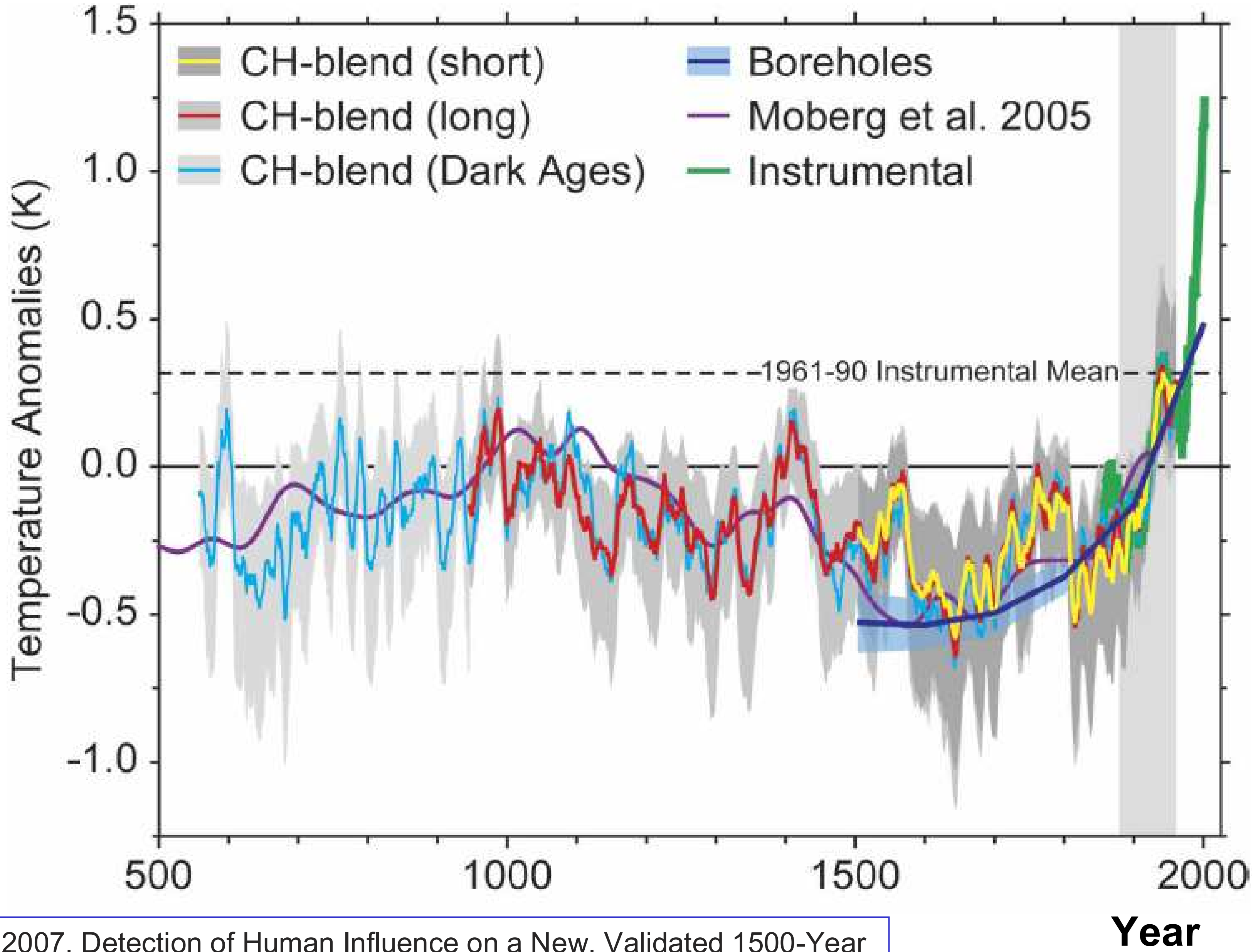
J.L. Lean and D.H. Rind, How will Earth's surface temperature change in future decades?, Geophys. Res. Lett.. 36, L15708, doi:10.1029/2009GL038932, 2009



Vorschau auf 2100 (IPCC 2007)



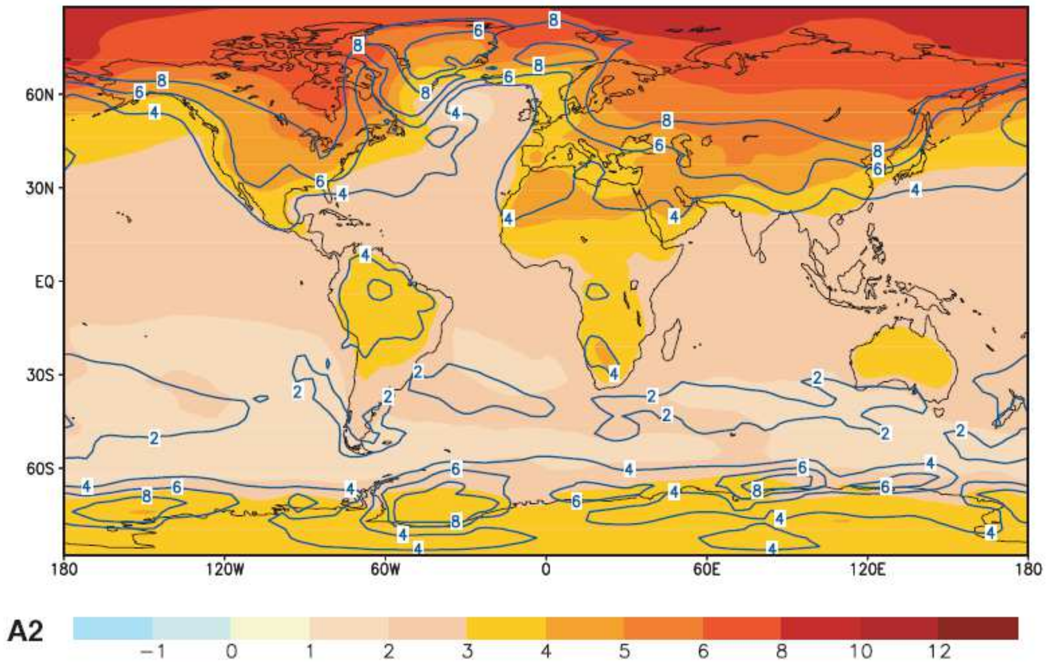
Rekonstruierte globale Temperatur über Land



Hegerl et al. 2007, Detection of Human Influence on a New, Validated 1500-Year Temperature Reconstruction, J. Climate 20, DOI: 10.1175/JCLI4011.1

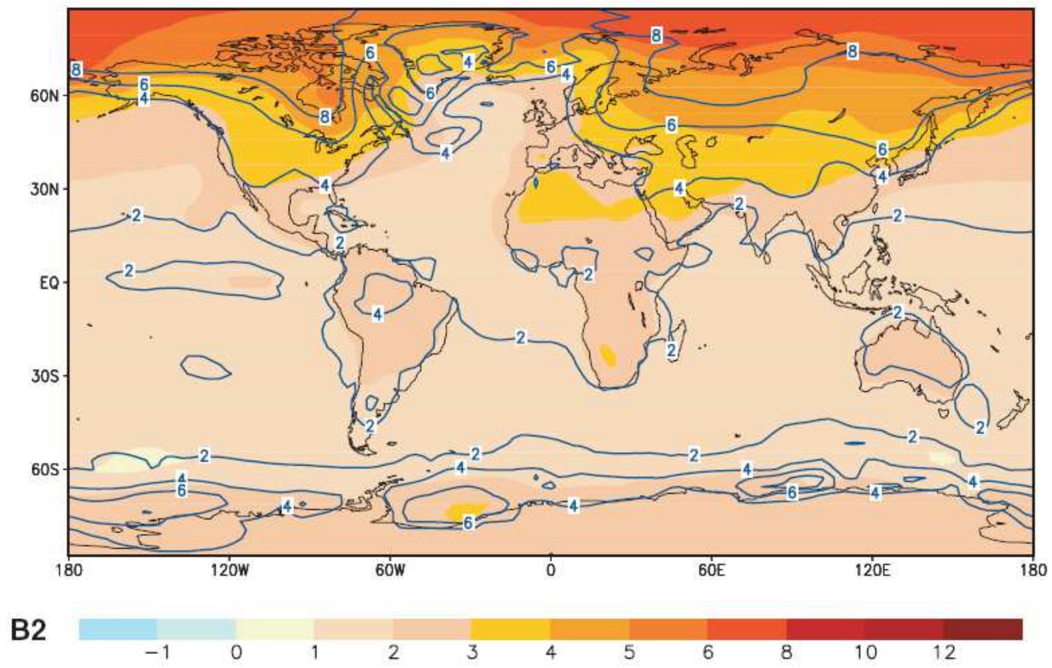
Regionale Konsequenzen des Klimawandels

A2: starker CO_2 Anstieg
2100: ~850 ppm



Jahresmittelwerte der Temperaturänderung bis 2100, Mittel über mehrere Modelle (Farbe) und Isolinien des Vertrauensbereiches (in °C) von OAGCMs.

B2: mäßiger CO_2 Anstieg
2100: ~600 ppm



→ Verschiedene Szenarien ergeben erstaunlich ähnliche Verteilungen

IPCC 2007