Climate sensitivity, radiative forcings and feedbacks : an introduction

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Climate sensitivity, radiative forcings and feedbacks : an introduction

Outlook

Part I

- A short history
- Why do we care about climate sensitivity?
- Forcing and feedback in a simple idealized model
- Radiative forcing - climate feedback analysis framework
- The various physical climate feedbacks
- How much individual feedbacks contribute to global warming

Part II

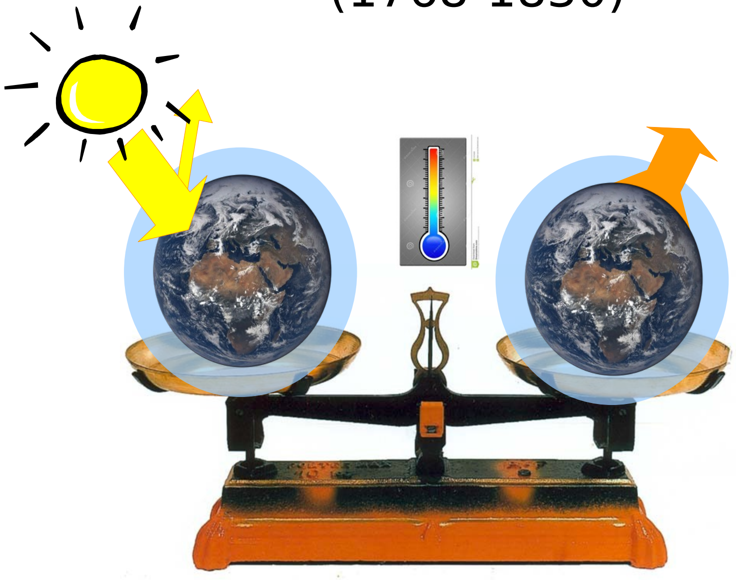
- The methods to estimate climate sensitivity and how to combine them
- Sensitivity to the geographical distribution of warming (pattern effect)

The first foundations of climate physics

J. Fourrier: *Mémoire sur les températures du globe terrestre et des espaces planétaires*, 1824

Joseph Fourrier
(1768-1830)

- He consider the Earth like any other planet
- The **energy balance equation** drives the temperature of all the planets
- The major heat transfers are
  1. **Solar radiation**
  2. **Infra-red radiation**
  3. Diffusion with the interior of Earth

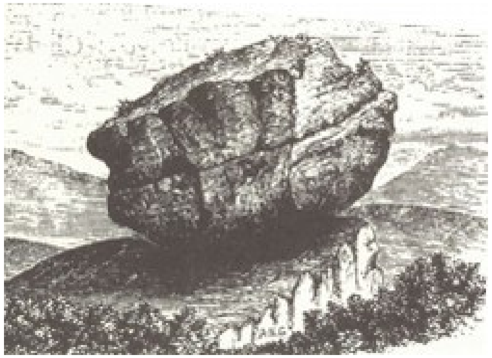


- He formulates the principle of the **greenhouse effect**
- He deduced that the **climate could change**, but refuted it and added an ad-hoc hypothesis.

Paleo climate changes

The discovery (1840-1860)

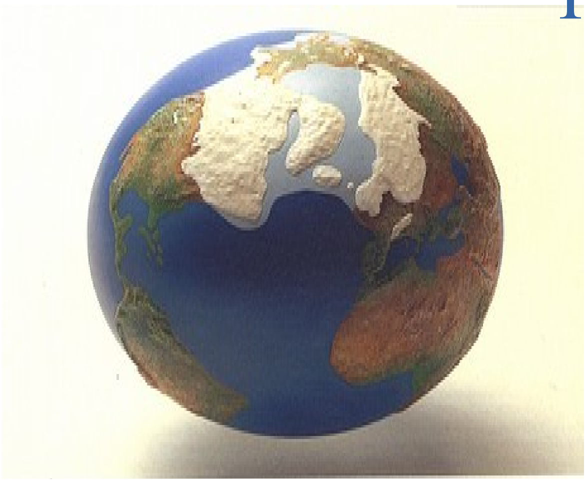
J. de Charpentier

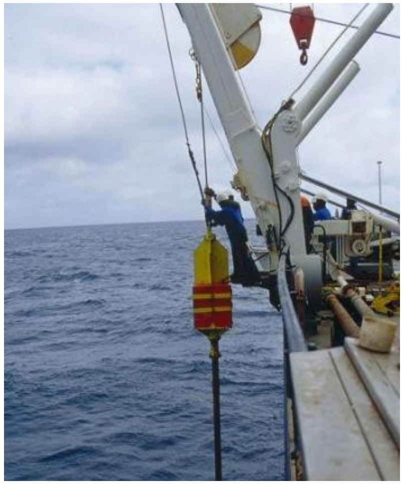


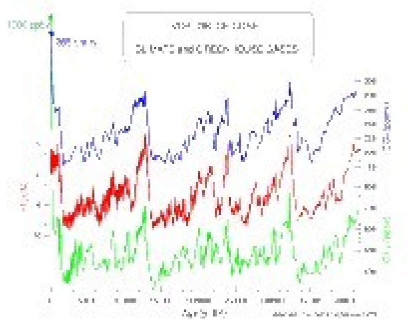
Blocs erratiques

L. Agassiz

The detailed description (1970-)







Paleo climate changes

The discovery (1840-1860)

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Blocs erratiques

L. Agassiz

Cause of these global temperature variations: sun or CO_2 ?

(1860-1900)

James Croll

Svante Arrhenius

Single layer greenhouse model

Surface temperature of an **isothermal planet** at equilibrium, with an **vertically uniform atmosphere**

$$\boxed{\sigma T_s^4 = \frac{(1-A)I_0}{1-\epsilon_a/2}}$$

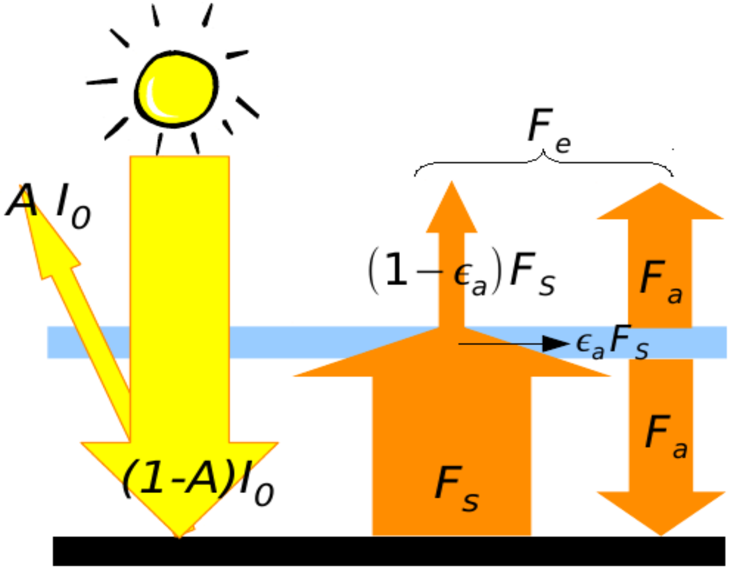
Svante Arrhenius

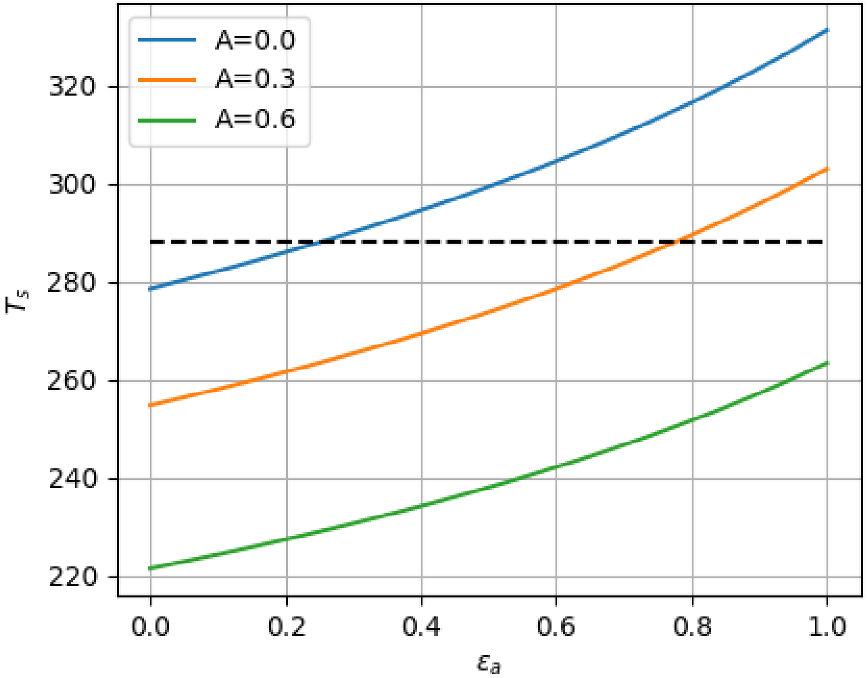
I_0: incoming solar radiation

A: planetary albedo

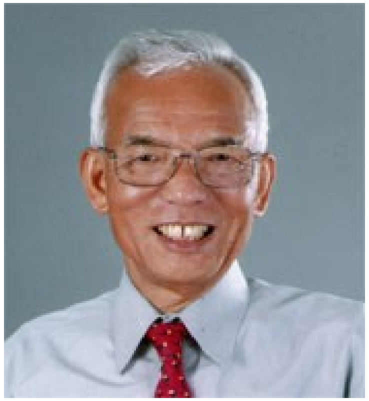
ε_a: planetary emissivity

$$T_s = f(\epsilon_a)$$



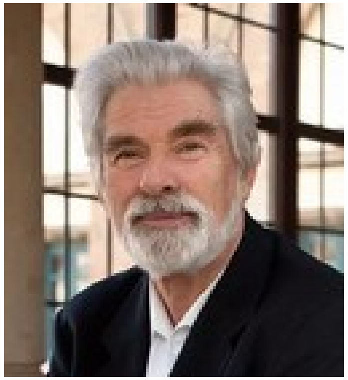


From energy balance models to general circulation models and dynamical systems



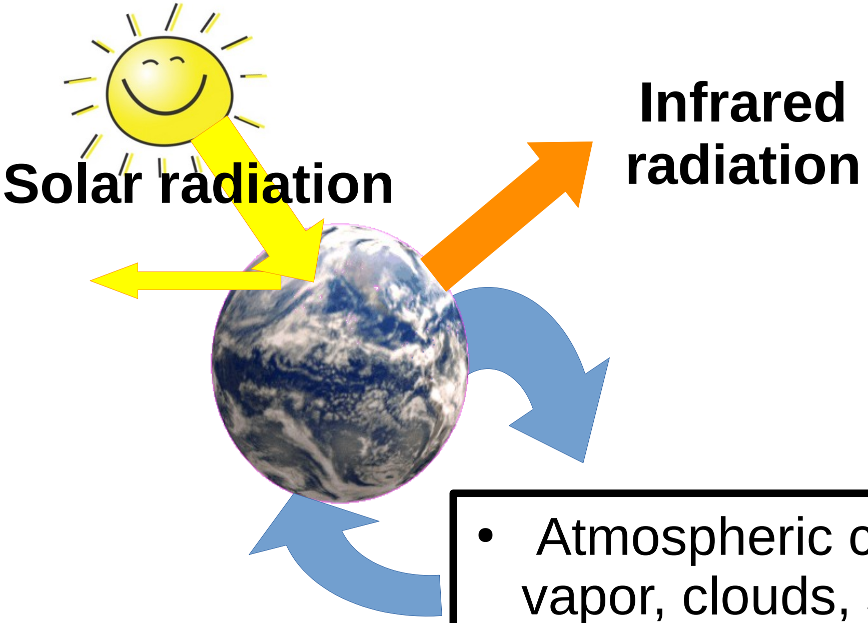
Syukuro Manabe

(1931-)



Klaus Hasselmann (1931-)

Nobel prize in Physics 2021



Solar radiation

Infrared radiation

	Time constant (years)
• Atmospheric circulation, water vapor, clouds, snow, surface ocean, sea ice, etc.	10
• Same + ocean circulation, CO2 continental biosphere, CH4, etc.	100
• Ice sheets, continental CO2	1000
• Geological" CO2 (continental erosion, volcanism)	$>10^6$

From energy balance models to general circulation models and dynamical systems

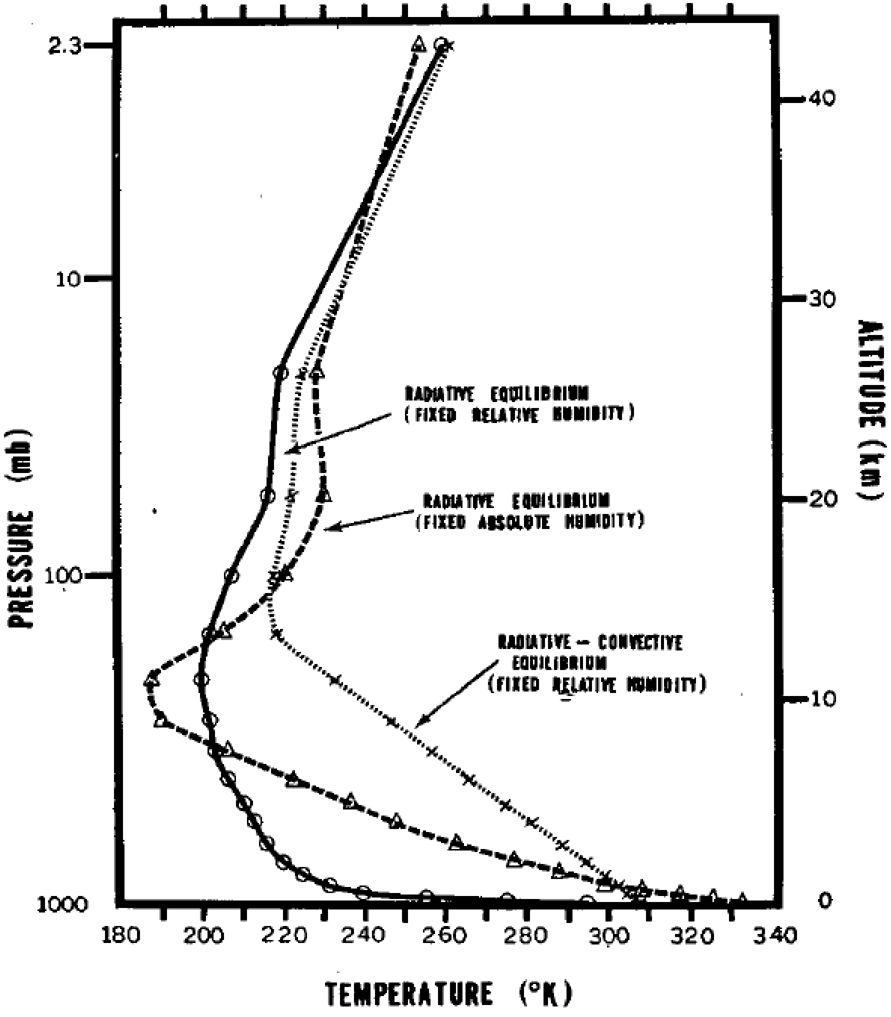


FIG. 5. Solid line, radiative equilibrium of the clear atmosphere with the given distribution of relative humidity; dashed line, radiative equilibrium of the clear atmosphere with the given distribution of absolute humidity; dotted line, radiative convective equilibrium of the atmosphere with the given distribution of relative humidity.

[Manabe & Wetherald, 1967]

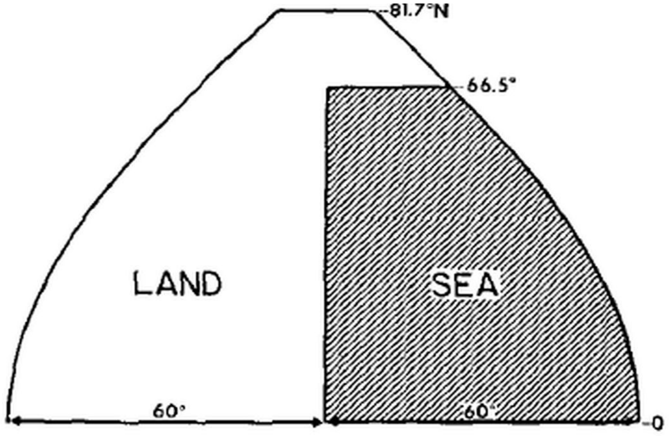
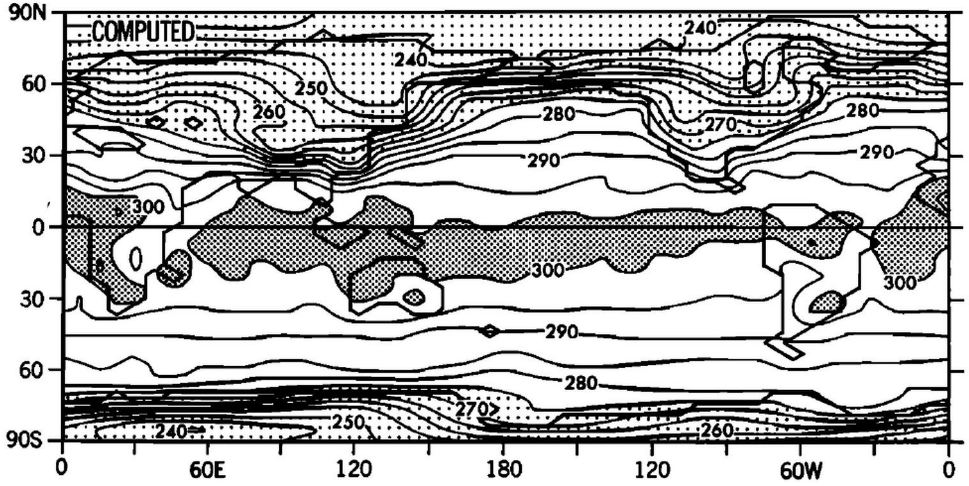


FIG. 1. Diagram illustrating the distribution of continent and "ocean." Cyclic continuity is assumed at the eastern and western ends of the domain.

[Manabe & Wetherald, 1975]



[Manabe & Stouffer, 1980]

“ Carbon Dioxide and Climate: A Scientific Assessment ”

(Charney et al. 1979)

- Now that the increase of the CO_2 concentration in the atmosphere has been observed, what are the implications?
- The US National Academy of Sciences asked a small work group of scientists to undertake a scientific assessment

Among the clonclusions: “We estimate the most probable warming for a doubling of CO_2 to be near 3°C with a probable error of 1.5°C.”

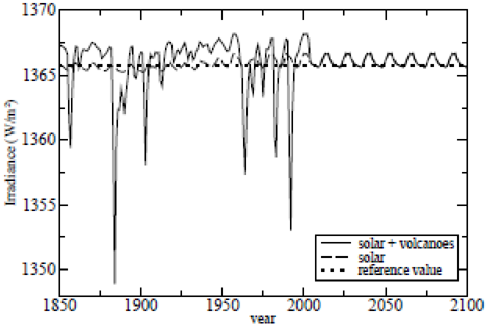
The ***Equilibrium climate sensitivity*** (ECS) is the equilibrium temperature change in response to a ***doubling*** of the atmospheric ***CO_2 concentration*** relative to pre-industrial levels.

Exemple of current climate model

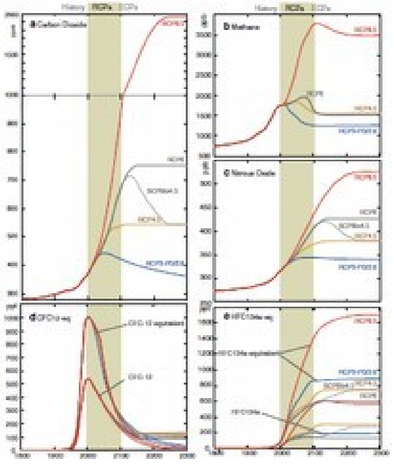
The IPSL « Earth System Model"

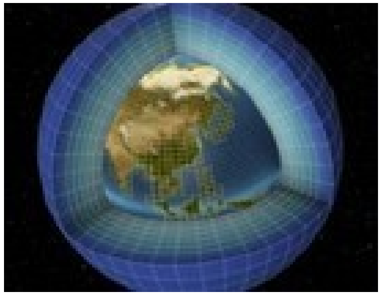
Natural and anthropogenic forcings

Sun and volcanos



Greenhouse or chemical reactive gases



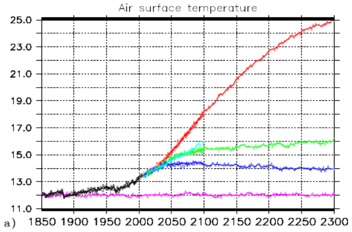


Climate model

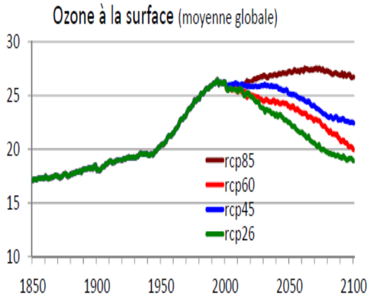
- 3D representation of the atmosphere, ocean, sea-ice, land-surface (coupling of different models)
- Representation and coupling with the biogeohemical cycles

Results

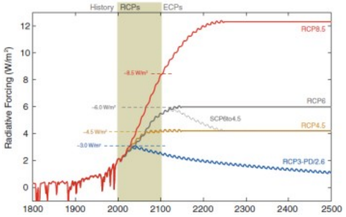
Climate changes



Atmosphere composition



Radiative forçings



The equilbrium climate sensitivity (ECS)

The definition of ECS, of which the interest is apparent, raises important difficulties:

- **Fundamental:** is there a climate in equilibrium?

- **Practical:**

• Computation time: as soon as coupled atmospheric-ocean dynamic models were used: reaching a new equilibrium requires very long simulations (several thousands of years), and therefore very expensive in

• How to make the link with observations?

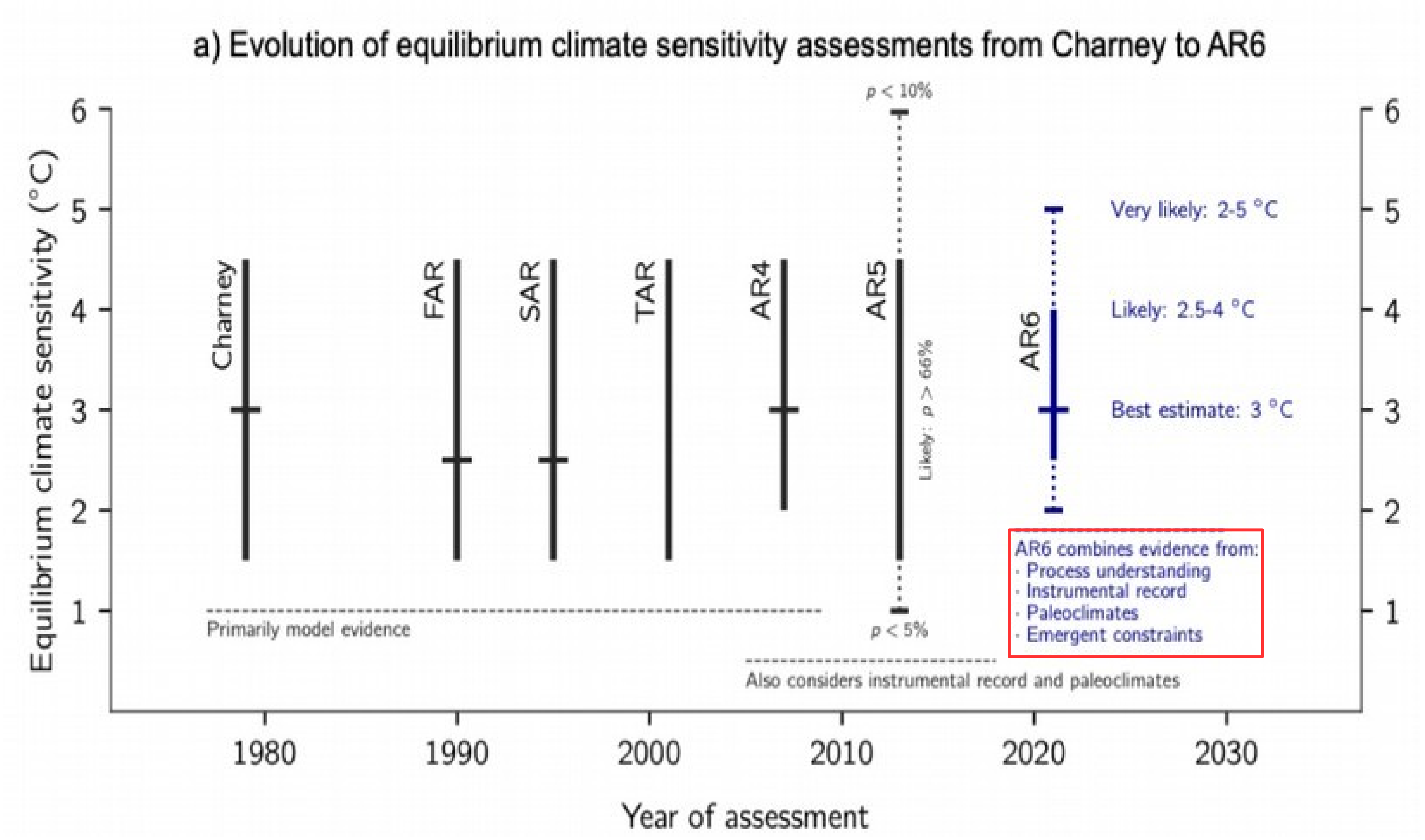
• How to assess the ECS?

• How does the response to CO_2 changes compare to the response to other perturbations (solar, volcanos, other GHGs, etc) ?

• How does this relate to future projections, for which the climate is out of equilibrium?

… and yet equilibrium climate sensitivity (ECS) is still essential in climate change studies

Estimate of the equilbrium climate sensitivity (ECS)



[IPCC AR6-WG1, TS]

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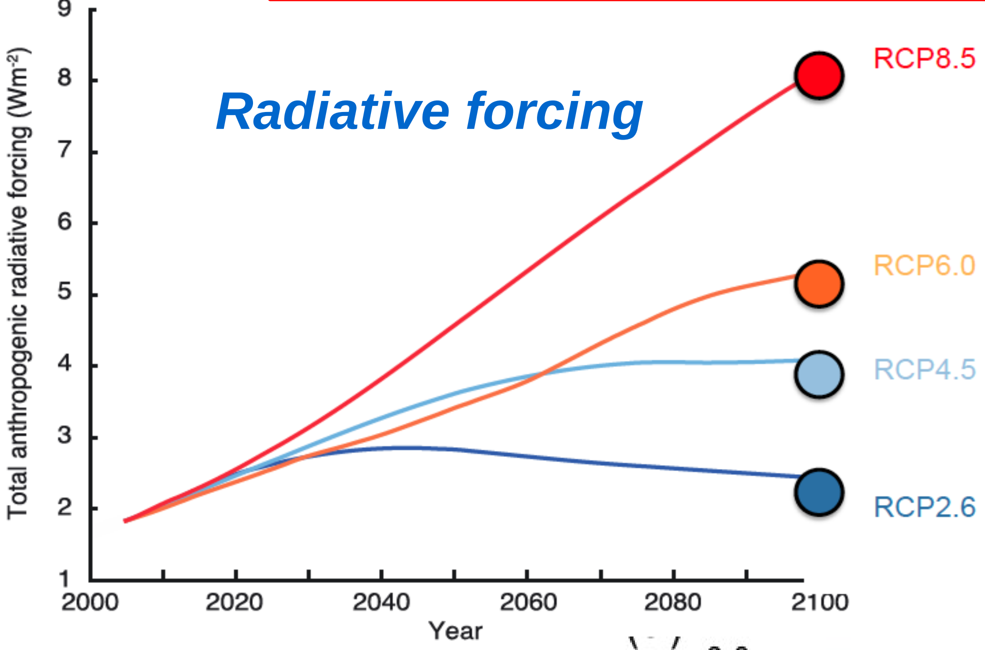
Why ECS is still used ?

Spread among model for future projections

Radiative forcing

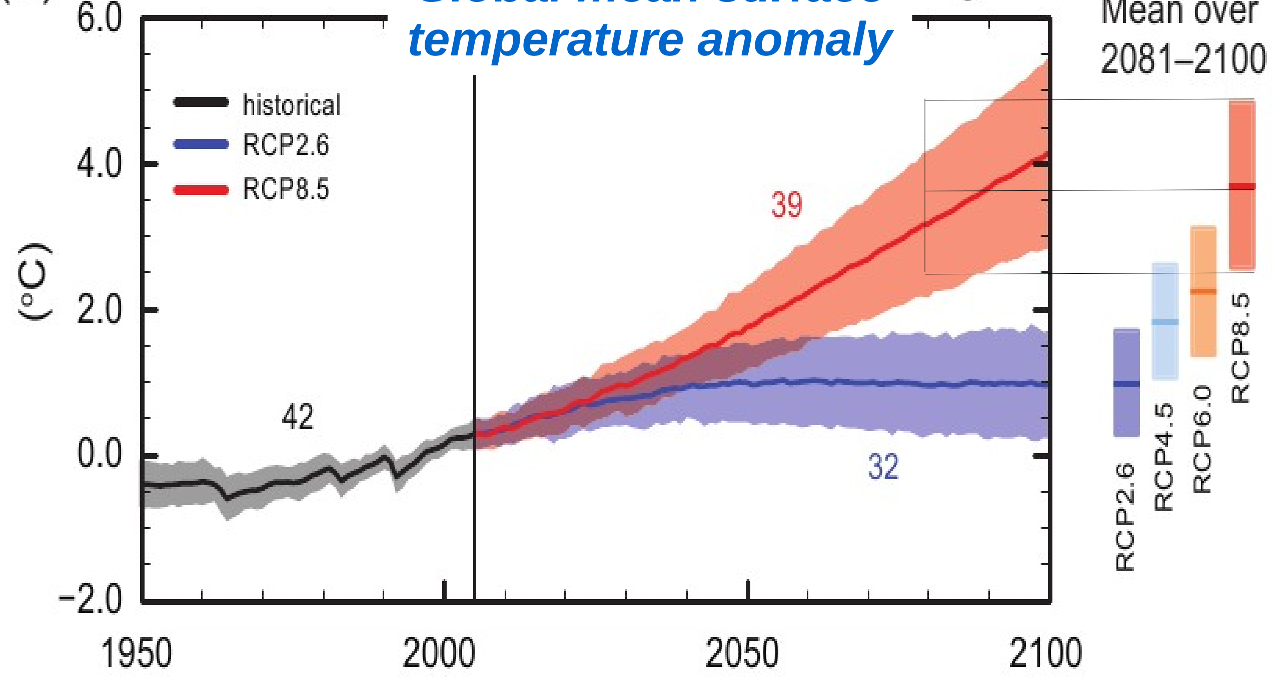


Strong emissions

Stabilisation

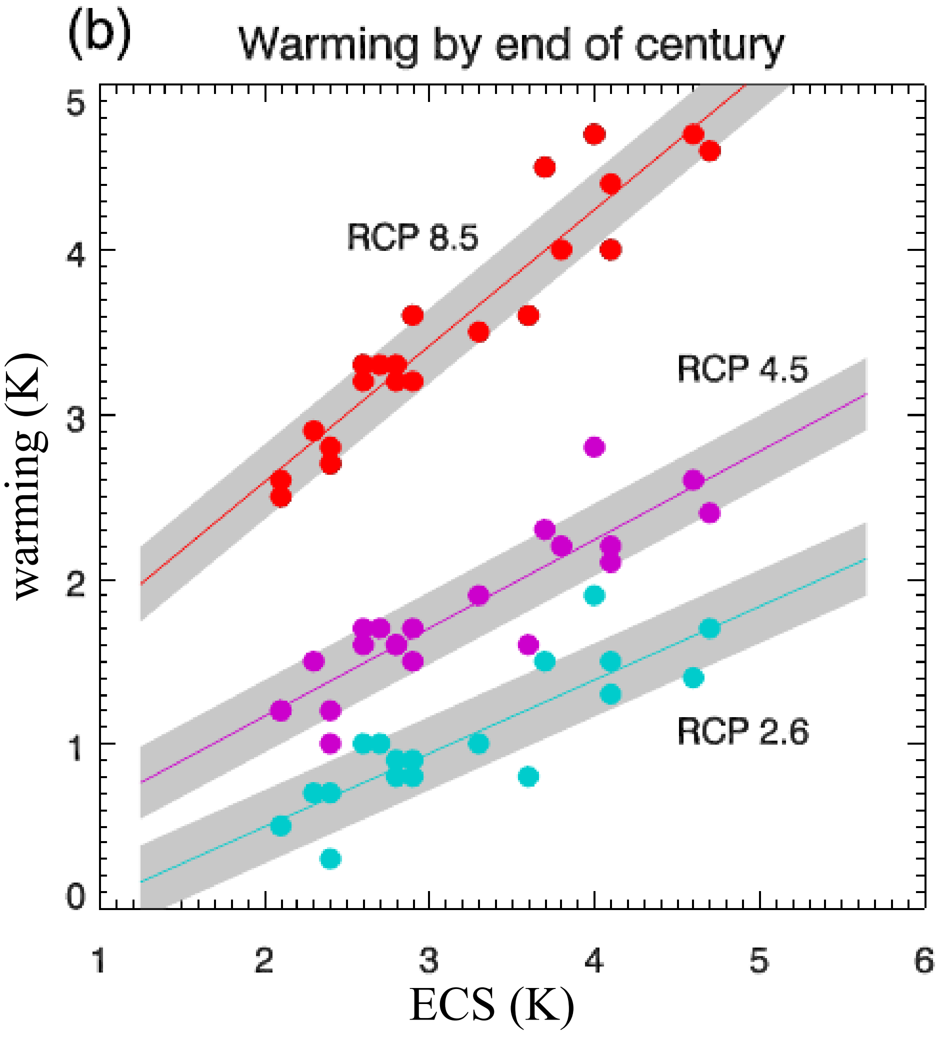
Emission control

Global mean surface temperature anomaly

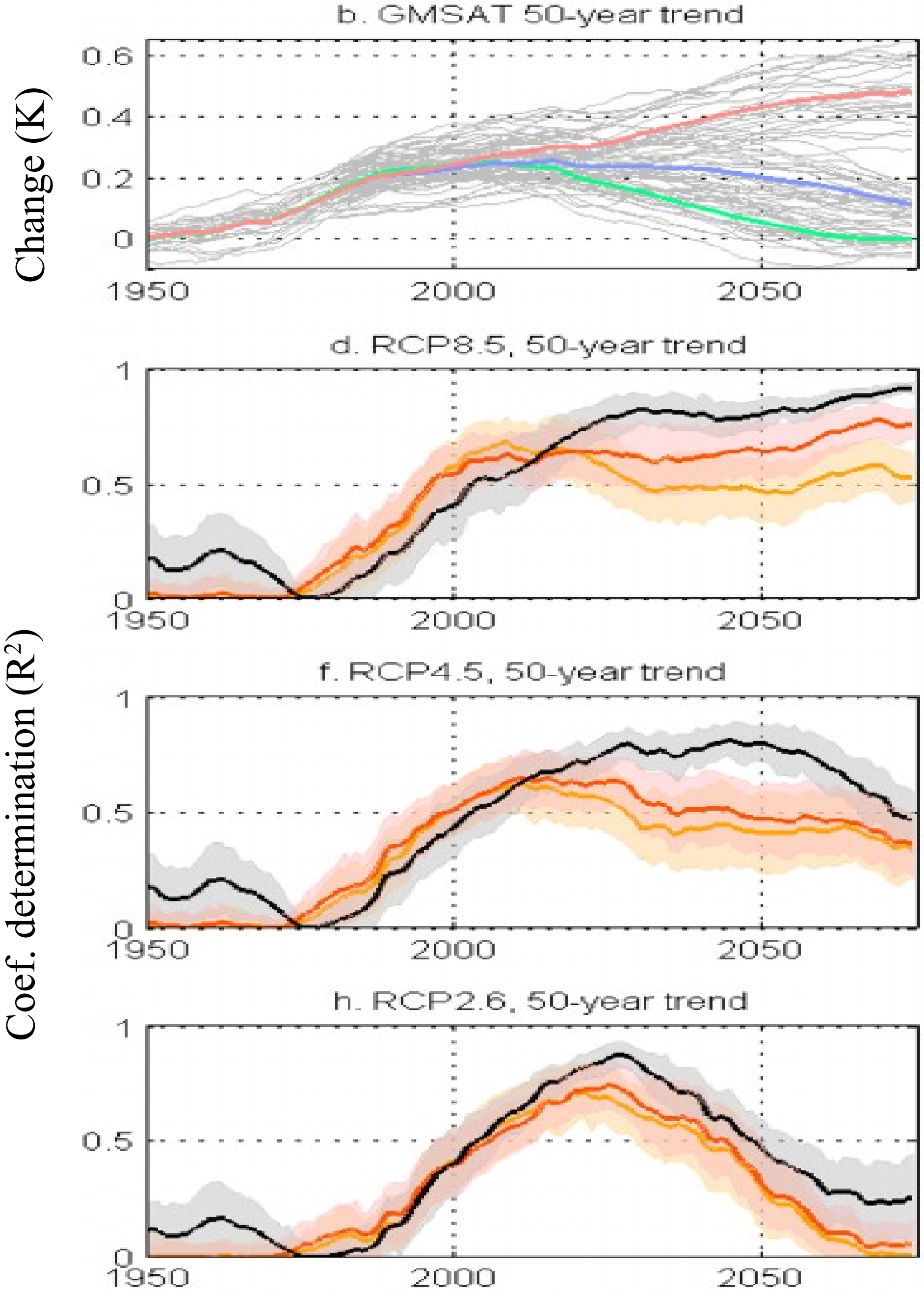


Why ECS is still used ?

Spread among model for future projections depend on the spread of their ECS



[Sherwood et al., 2020]



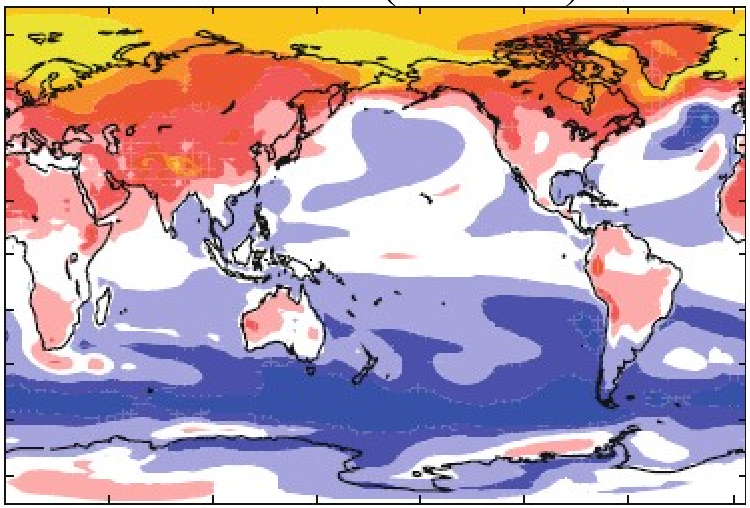
[Grose et al., 2018]

Why ECS is still used ?

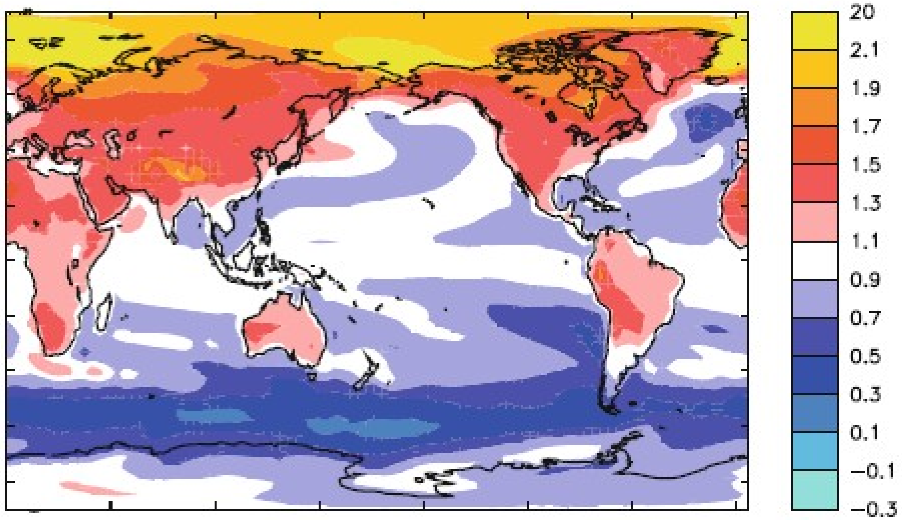
As a first approximation (pattern scaling): ΔX(space,time) = global ΔT(time) x pattern(space)

Local ΔT normalized by global ΔT (K/K)

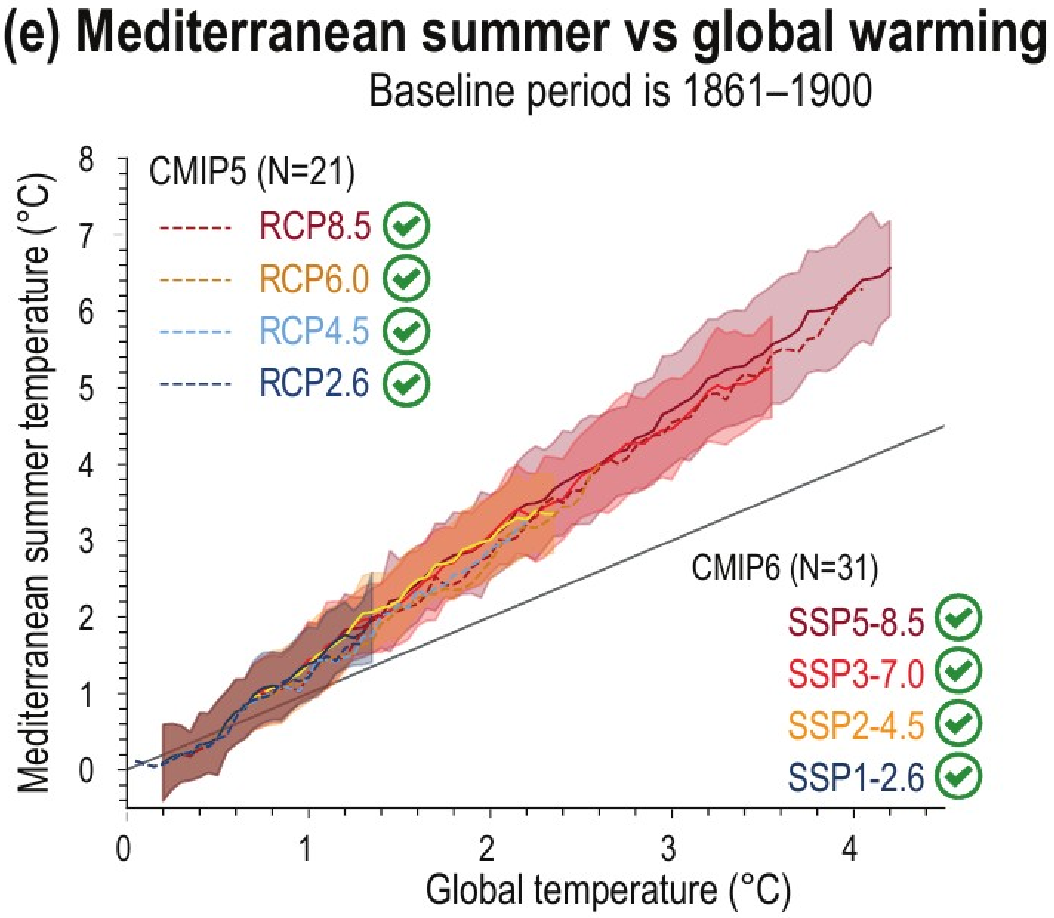
RCP 2.6 (ΔT = 2K)



RCP 8.5 (ΔT = 6K)



[IPSL]



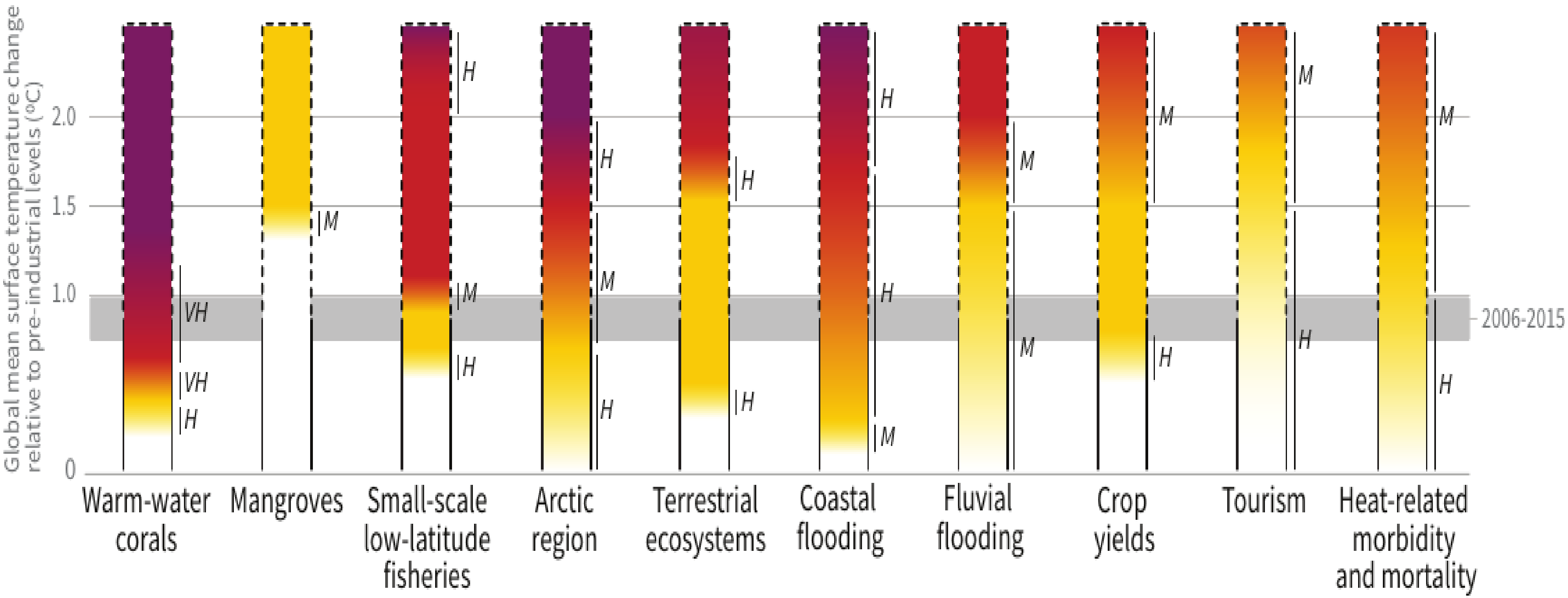
(e) Mediterranean summer vs global warming

Baseline period is 1861–1900

[IPCC AR6-WG1, TS]

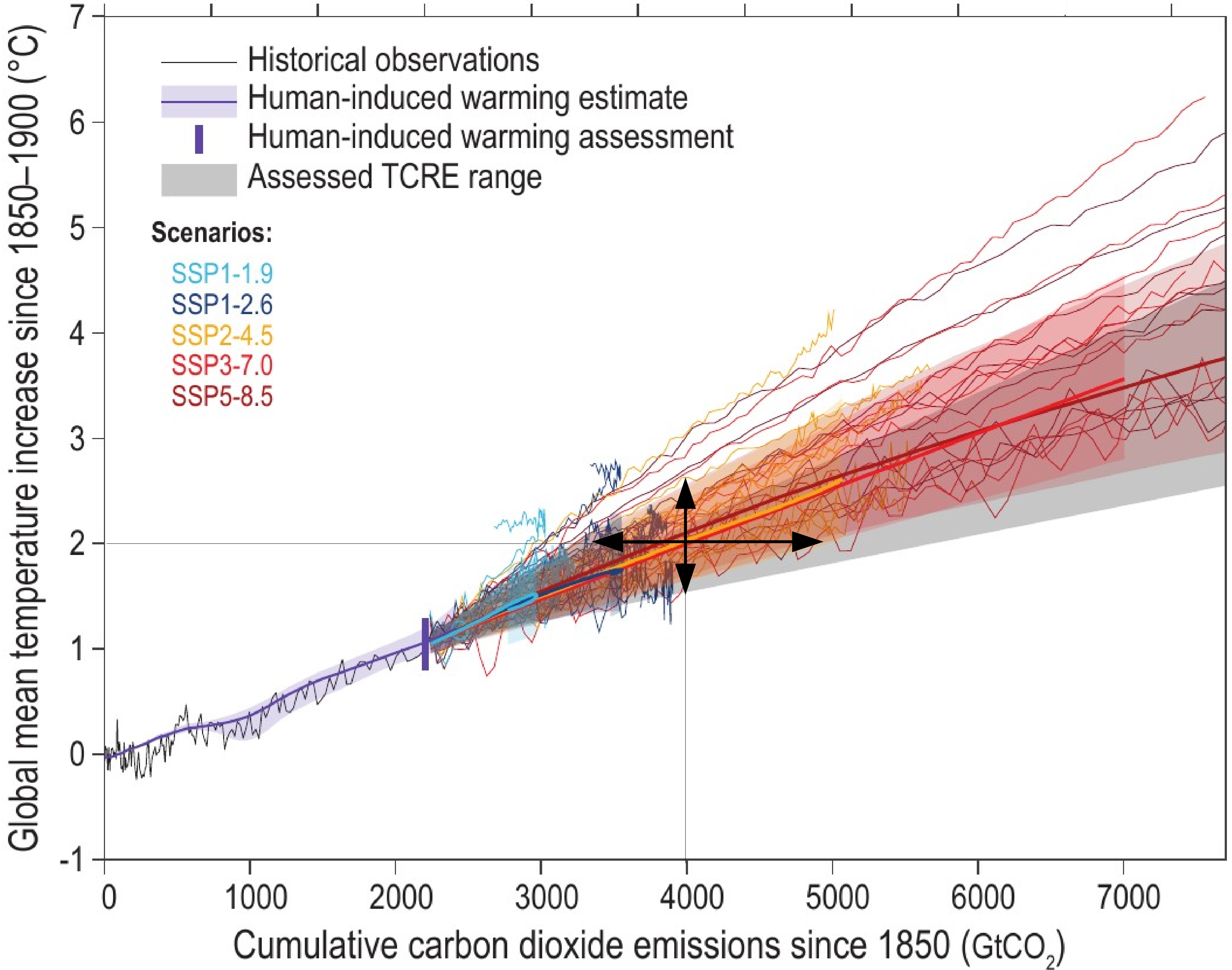
Why do we care about climate sensitivity?

Climate change impact scales with global temperature increase


Impacts and risks for selected natural, managed and human systems

[IPCC SR1.5]

Global warming depends on cumulative CO_2 emissions



[IPCC AR6-WG1, TS]

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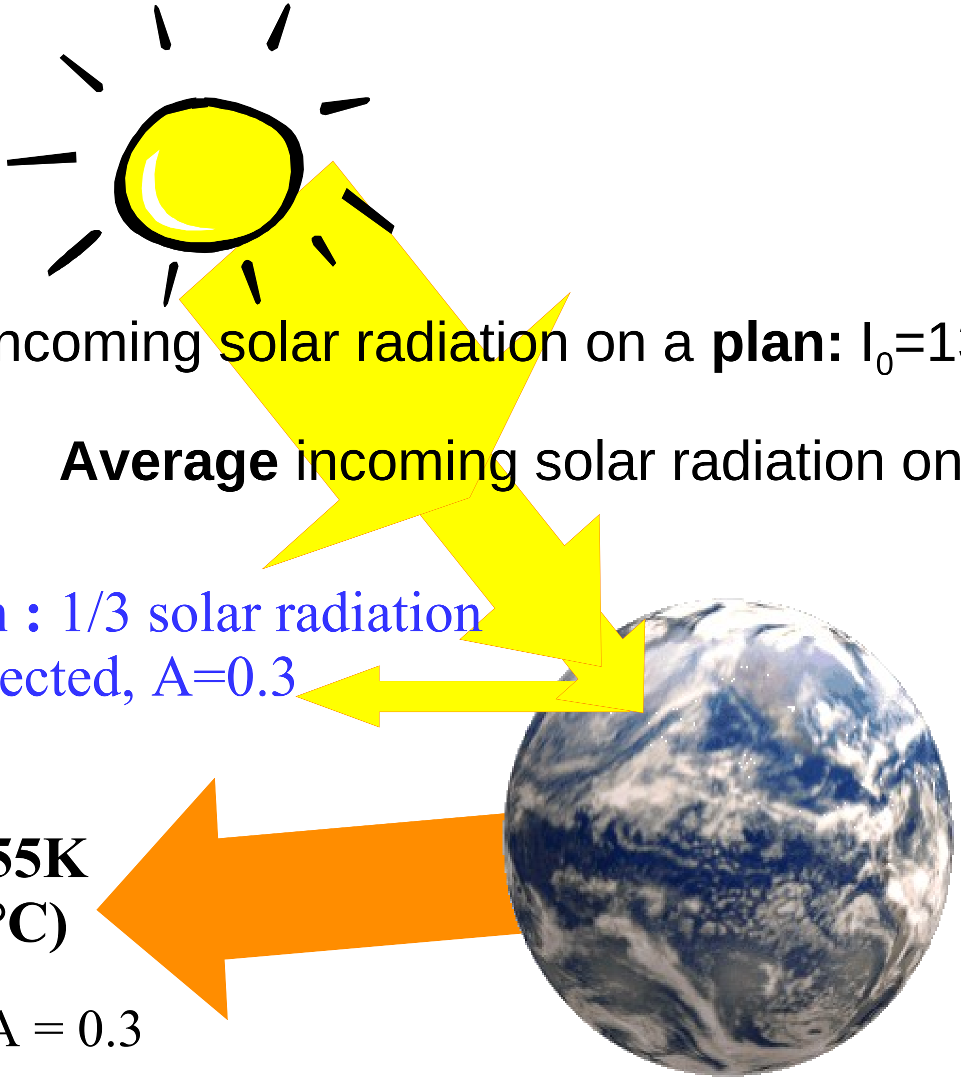
Equilibrium temperature of an **isothermal** planet



Incoming solar radiation on a **plan:** I_0=1364 W.m^{-2}

Average incoming solar radiation on a **sphere:** $I = I_0/4 = 341$ W.m^{-2}

Earth : 1/3 solar radiation is reflected, A=0.3

2/3 solar radiation is absorbed

T_s= 255K (-18°C)

$$\sigma \epsilon T_s^4 = (1-A) I$$

$\varepsilon = 1$, $A = 0.3$

- What happens when the incoming solar radiation I varies by δI ?
- What is the response δT_s of the surface temperature T_s

Response to a change in incoming solar radiation

Initial equilibrium state: $\sigma \epsilon T_s^4 = (1-A) I$

Equilibrium state after a change δI: $\sigma \epsilon (T_s + \delta T_s)^4 = (1-A)(I + \delta I)$

$$(T_s + \delta T_s)^4 \approx T_s^4 + 4 T_s^3 \delta T_s$$

$$\cancel{\sigma \epsilon T_s^4} + 4 \sigma \epsilon T_s^3 \delta T_s = (1\cancel{-A}) I + (1-A) \delta I$$

$$\boxed{-\lambda_P \delta T_s = (1-A) \delta I} \quad \text{with} \quad \boxed{\lambda_P = \frac{\partial F_{LW}}{\partial T_s} = -4 \sigma \epsilon T_s^3}$$

- The change of temperature δT_s leads to a change of the LW flux at the TOA of $\lambda_P \delta T_s$ that compensates the change in absorbed solar radiation $(1\text{-}A)\delta I$
- λ_P is called the Planck response parameter

$$T_s \approx 280\,K\,;\ \lambda_P \approx -5\,Wm^{-2}K^{-1};\ \delta T_s \approx 0.2(1-A)\delta I$$

$$T_s \approx 250\,K\,;\ \lambda_P \approx -3.5\,Wm^{-2}K^{-1};\ \delta T_s \approx 0.28(1-A)\delta I$$

Response to a change in incoming solar radiation

Previously we have assumed that the albedo did not changed.
What if the albedo depends on the surface temperature?

Equilibrium state after a change δI:

$$\sigma\epsilon(T_s+\delta T_s)^4=(1-[A+\frac{\partial A}{\partial T_s}\delta T_s])(I+\delta I)$$

$$\cancel{\sigma\epsilon T_s^4}+4\sigma\epsilon T_s^3\,\delta T_s=(1\cancel{-A})I+(1-A)\,\delta I-\frac{\partial A}{\partial T_s}\delta T_s I+\epsilon(\delta^2)$$

$$\boxed{-(\lambda_P+\lambda_A)\,\delta T_s\approx(1-A)\,\delta I}$$

with

$$\boxed{\lambda_A=\frac{\partial A}{\partial T_s}I=\frac{\partial F}{\partial A}\frac{\partial A}{\partial T_s}}$$

- The change of temperature δT_s leads to a change of the albedo and therefore the absorbed solar radiation and the surface temperature
- λ_A is called the “albedo feedback parameter”

Response to a change in incoming solar radiation

Without albedo feedback: $$\delta T_{s,P} = -\frac{(1-A)\,\delta I}{\lambda_P}$$

With albedo feedback: $$\delta T_s \approx -\frac{(1-A)\,\delta I}{\lambda_P + \lambda_A}$$

$$\delta T_s \approx -\frac{(1-A)\,\delta I}{\lambda_P(1+\lambda_A/\lambda_P)} = \delta T_{s,P}\frac{1}{1-g}$$

$$\delta T_s \approx G\,\delta T_{s,P}$$

with $\delta T_{s,P}$ the response with no feedback

$G = \frac{1}{1-g}$ the gain

$g = -\frac{\lambda_A}{\lambda_P}$ the feedback gain

- If $\lambda_A > 0$ then g >0, $\delta T_s > \delta T_{s,P}$, the feedback is positive, it amplifies the response without feedback

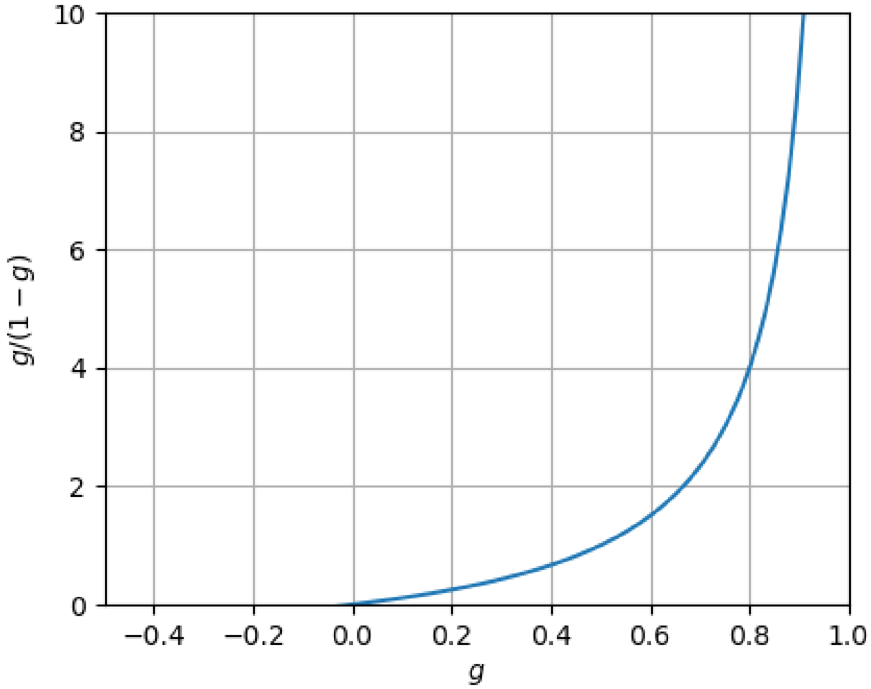
Response to a change in incoming solar radiation

Increase of temperature due to the albedo feedback:

$$\delta T_{s,A} = \delta T_s - \delta T_{s,P} = \delta T_{s,P}\left(\frac{1}{1-g} - 1\right) = \delta T_{s,P}\left(\frac{g}{1-g}\right) \qquad g = -\frac{\lambda_A}{\lambda_P}$$

Ex: If $\lambda_A = -0.2 * \lambda_P$ => then $0.25 * \delta T_{s,P}$

If $\lambda_A = -0.4 * \lambda_P$ => then $0.67 * \delta T_{s,P}$



Introducing the forcing and the response

Forcing (or perturbation): flux ΔQ (e.g. $(1\text{-}A)\,\delta I$ Heat budget : N

Surface temperature : Ts

Response of the whole system : $\Delta T_s = -\dfrac{\Delta Q}{\lambda}$ with $\lambda = \dfrac{dN}{dT_s}$

Planck response : $\Delta T_{s,P} = -\dfrac{\Delta Q}{\lambda_P}$ with $\lambda_P = \dfrac{\partial N}{\partial T_s}$

All feedbacks response : $\Delta T_{s,f} = -\dfrac{\Delta Q}{\lambda_f}$ with $\lambda_f = \dfrac{\partial N}{\partial f}$

Feedbacks :

- **increase the amplitude** of the response, relative to the Planck response, **if $\lambda_f > 0$**, i.e. if the energy balance increases with temperature because of the feedbacks
- **reduce the amplitude** of the response, compared to the Planck response, **if $\lambda_f < 0$,** i.e. if the energy balance decreases as the temperature increases due to feedbacks
- **make the system unstable if $g > 1$**, i.e. if the energy gained from feedbacks is greater than the energy lost through “Planck emission”

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Radiative forcing-feedback framework
(or forcing - response)

Radiative forcing **aims to compare the magnitude of different perturbations** that impact climate

The radiative forcing Δ Q is the **change in the net radiative flux** (in W.m^{-2}) at the top of atmosphere due to a change in an external forcing (a driver of climate change) **before surface temperature adjusts** to this perturbation

The “climate feedback parameter” λ is the **sensitivity of the net radiative flux** at the top of atmosphere **to a change in the global mean surface temperature T$_s$** (in W.m^{-2}.K^{-1})

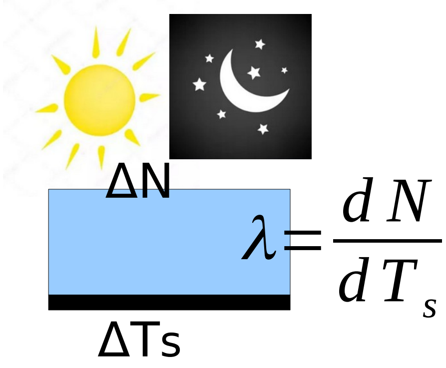
$$\boxed{\Delta N = \Delta Q + \lambda \Delta T_s}$$

Change in net flux at the TOA

radiative forcing

“climate feedback parameter”

Change in global mean **surface** temperature



$$\lambda = \frac{dN}{dT_s}$$

Radiative forcing-feedback framework (or forcing - response)

$$\boxed{\Delta N = \Delta Q + \lambda \Delta T_s}$$

Change in global mean **surface** temperature

Change in net flux at the TOA

radiative forcing

“climate feedback parameter”

When a new equilibrium is reached, $\Delta N=0$

$$\boxed{\Delta T_s^e = -\frac{\Delta Q}{\lambda}}$$

If λ is constant, ΔT is proportional to the radiative forcing

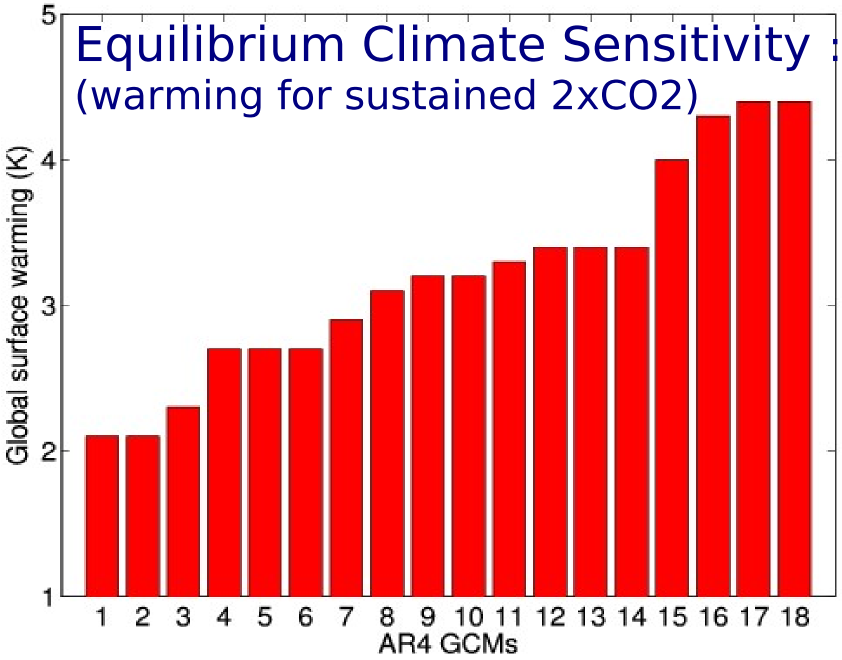
Equilibrium climate sensitivity (ECS) is the equilibrium change in global and annual mean surface air temperature after doubling the atmospheric concentration of CO2 relative to pre-industrial levels.

$$ECS = \frac{-\Delta Q(2xCO_2)}{\lambda}$$

Climate sensitivity and climate feedback parameters

Definition and ranges

Equilibrium climate sensitivity (ECS) is the equilibrium change in global and annual mean surface air temperature after doubling the atmospheric concentration of CO2 relative to pre-industrial levels.



At equilibrium: $\Delta Te = - \Delta Q/\lambda = S' \Delta Q$ (in K)

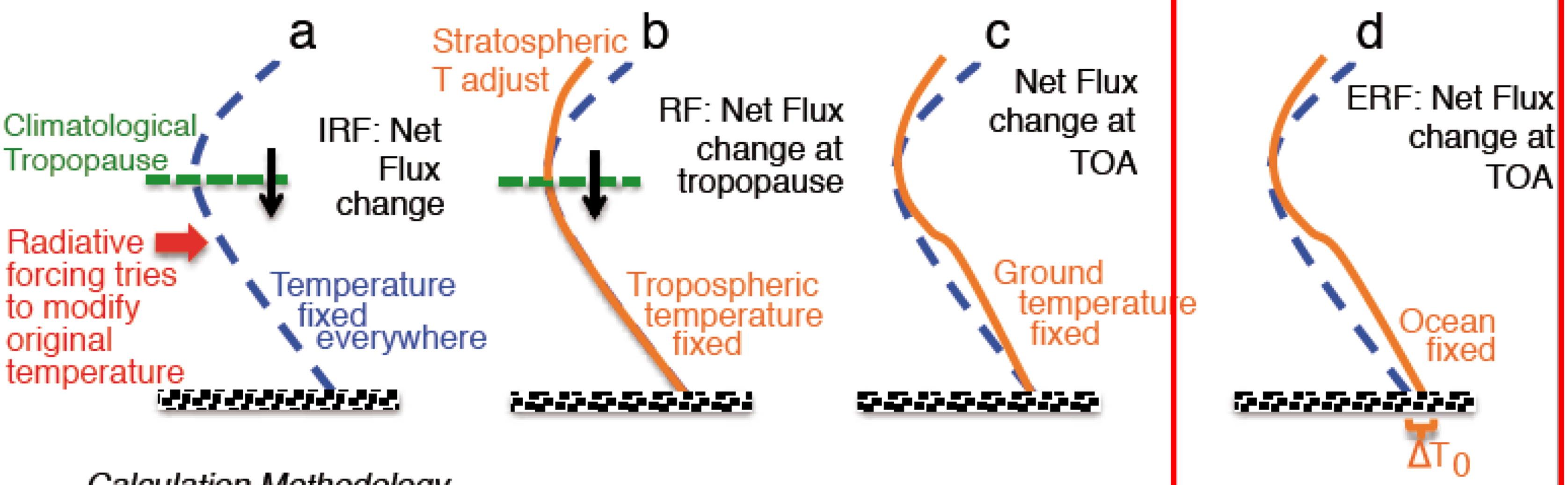
ΔQ : radiative forcing (in $W.m^{-2}$)

λ : climate feedback parameter (in $W.m^{-2}.K^{-1}$) ; range [-0.9 ; -1.8]

$S' = -1/\lambda$: climate sensitivity parameter (in $K.W^{-1}.m^{2}$); range [0.55 ; 1.1]

$ECS = - \Delta Q(2xCO_2)/\lambda$: climate sensitivity (in K) ; range [2 ; 4.5]

Radiative forcing: evolution of the definition to improve the proportionality between ***ΔQ*** and ***ΔT***



Calculation Methodology

Online or offline pair of radiative transfer calculations within one simulation

Difference between two offline radiative transfer calculations with prescribed surface and tropospheric conditions allowing stratospheric temperature to adjust

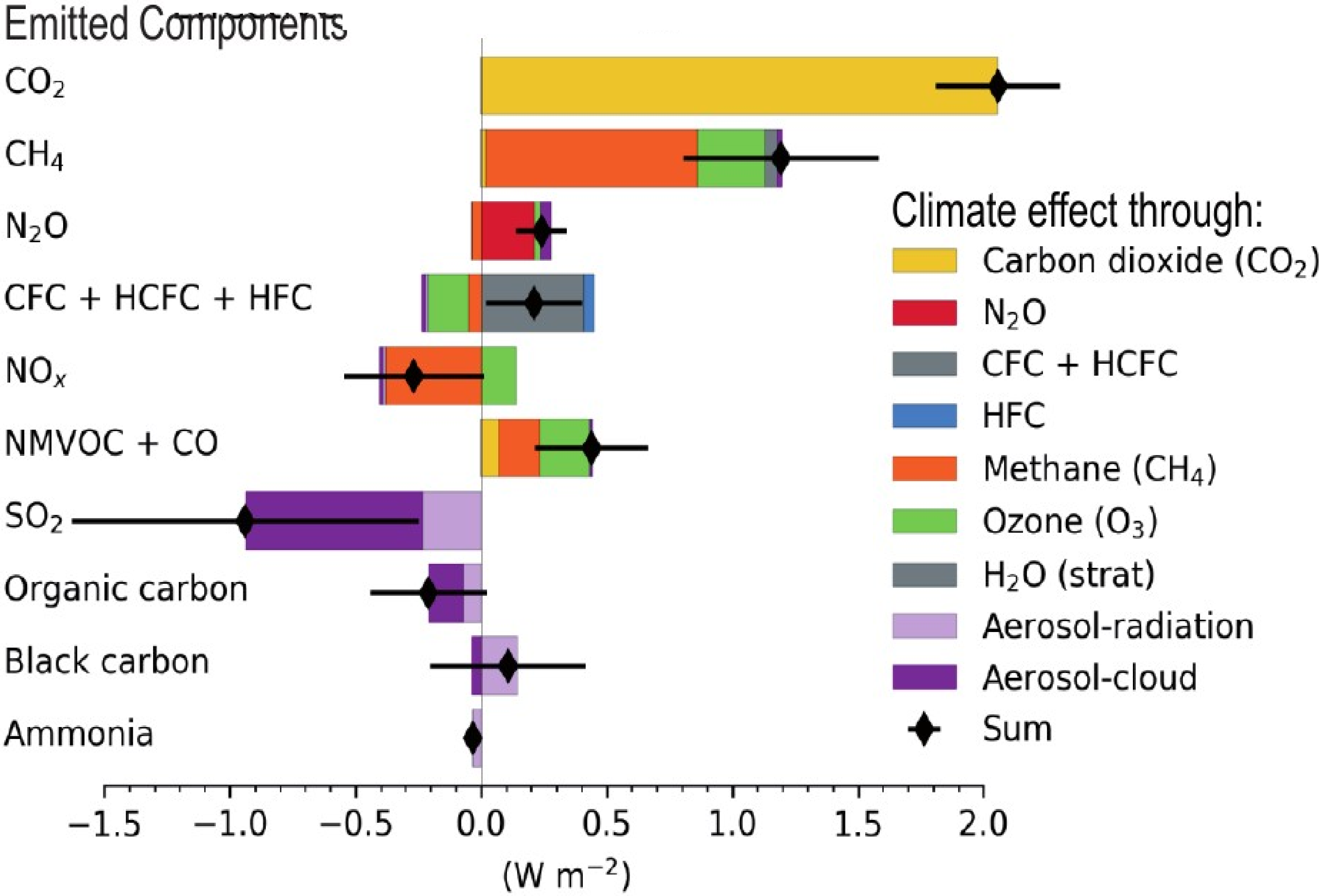
Difference between two full atmospheric model simulations with prescribed surface conditions everywhere or estimate based on regression of response in full coupled atmosphere-ocean simulation

Difference between two full atmospheric model simulations with prescribed ocean conditions (SSTs and sea ice)

Effective Radiative Forcing

Hansen 2005, updated IPCC 2013)

Radiative forcing from 1750 to 2019



[IPCC AR6-WG1, TS]

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Climate feedbacks

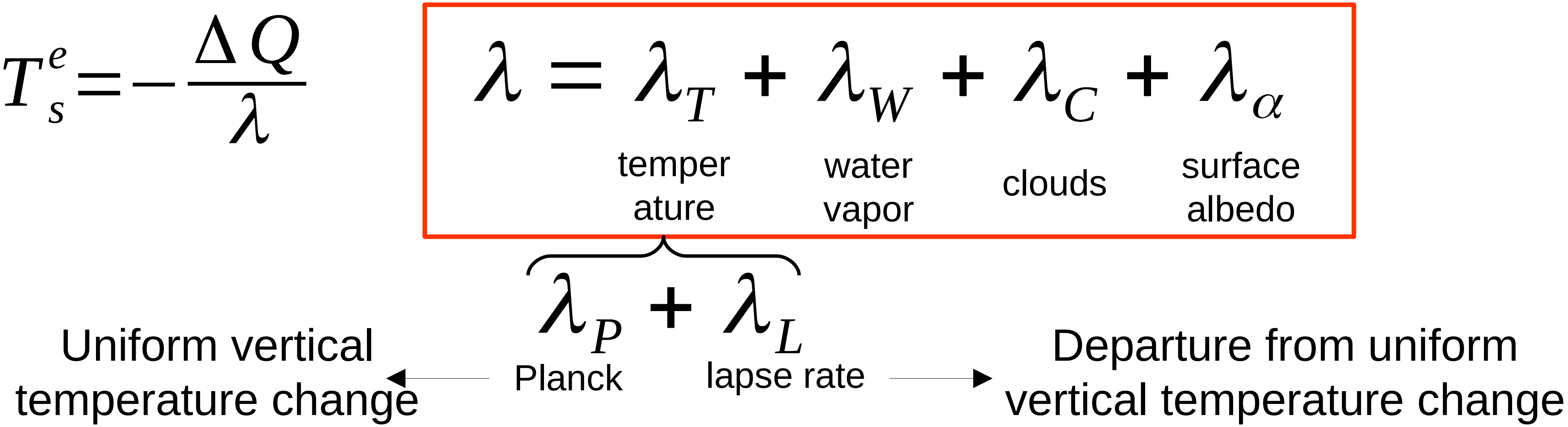
On Earth, Planck parameter $\lambda_p \approx -3.2$ W.m^{-2}K^{-1}

For a doubling of the CO_2 concentration, $\Delta Q \approx 3.7$ W.m^{-2}, the temperature increases by ≈ 1.2 K, if nothing change except the temperature

But feedbacks exist:

- Snow and sea ice reflect solar radiation; if they decrease, more solar energy will be absorbed ⇒ positive feedback
- Water vapour is the main greenhouse gas; if it increases, the greenhouse effect will be enhanced ⇒ positive feedback
- Clouds reflect solar radiation and contribute to the greenhouse effect; if they change, the energy budget will be modified ⇒ positive or negative feedback

$$\Delta T_s^e = -\frac{\Delta Q}{\lambda}$$



$$\boxed{\lambda = \underset{\text{temperature}}{\lambda_T} + \underset{\text{water vapor}}{\lambda_W} + \underset{\text{clouds}}{\lambda_C} + \underset{\text{surface albedo}}{\lambda_\alpha}}$$

$$\lambda_T = \underset{\text{Planck}}{\lambda_P} + \underset{\text{lapse rate}}{\lambda_L}$$

Uniform vertical temperature change ← Planck

lapse rate → Departure from uniform vertical temperature change

How to compute feedbacks ?

Diagnostic of feedback parameters through the Kernel approach

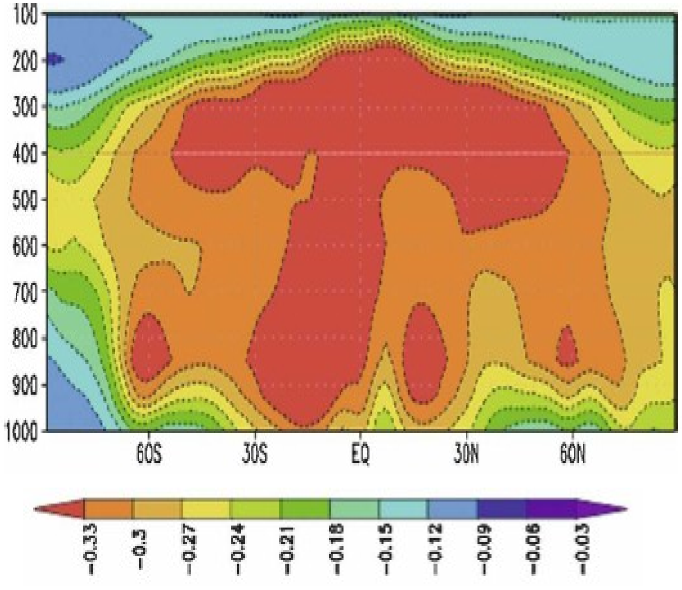
$$\lambda = \frac{dN}{dT_s} = \sum_x \frac{\partial N}{\partial x} \frac{\partial x}{\partial T_s}$$

radiative kernel computed by radiative codes

response to surface temperature change

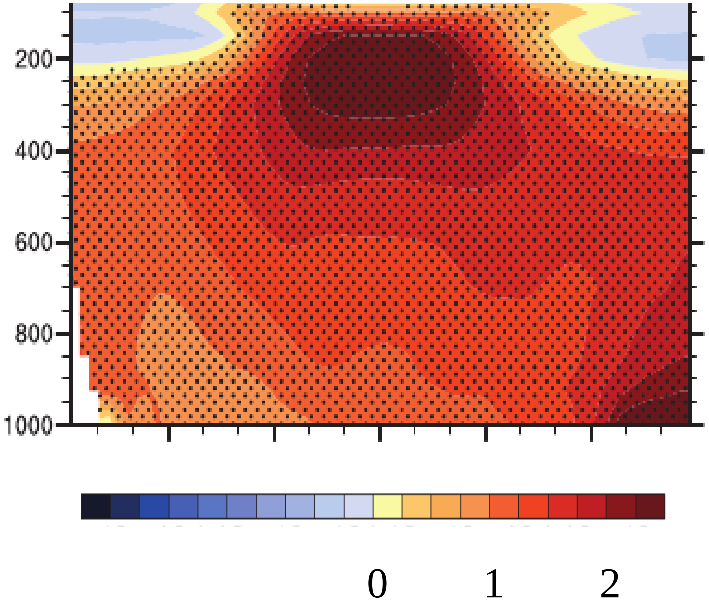
e.g. for *x* = *T* :

Temperature kernel $\frac{\partial N}{\partial T}$



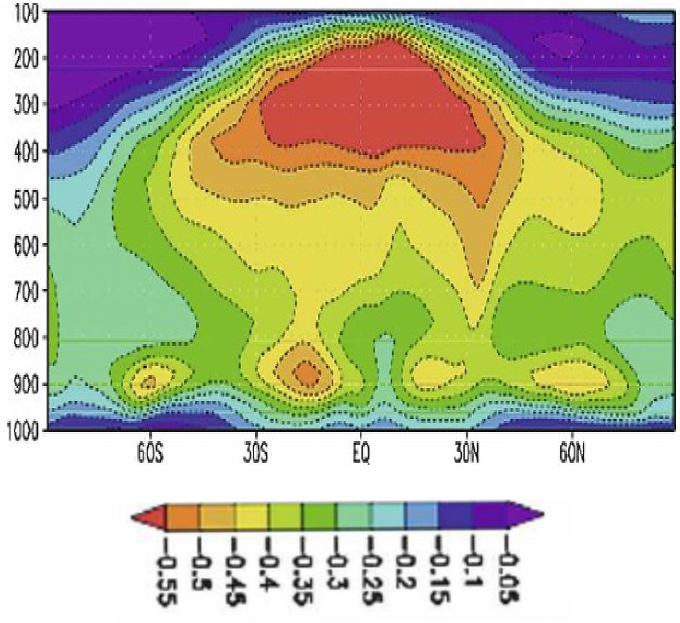
W/m²/K/(100hPa)

Temperature change $\frac{\partial T}{\partial T_s}$



K/K (approximate)

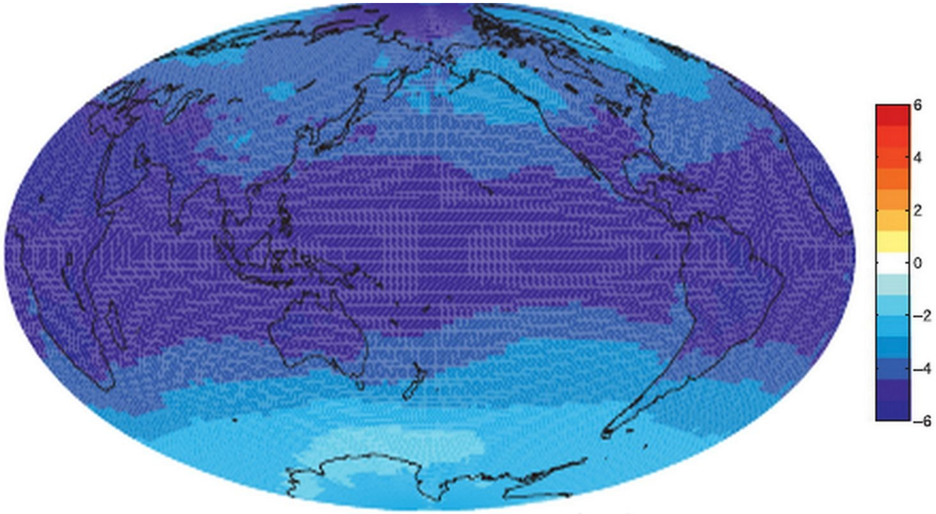
Temperature feedback parameter $\lambda_T = \frac{\partial N}{\partial T} \frac{\partial T}{\partial T_s}$



W/m²/K/(100hPa)

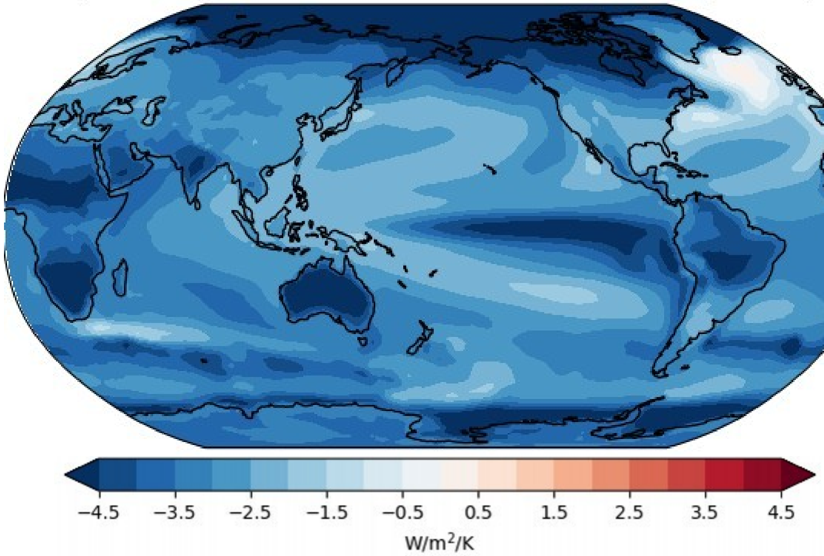
Soden et al., J. Climate, 2008

Temperature feedback



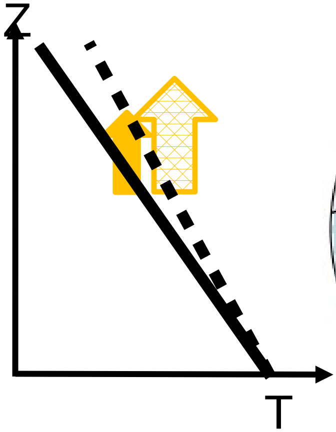
Planck feedback (uniform temp change)

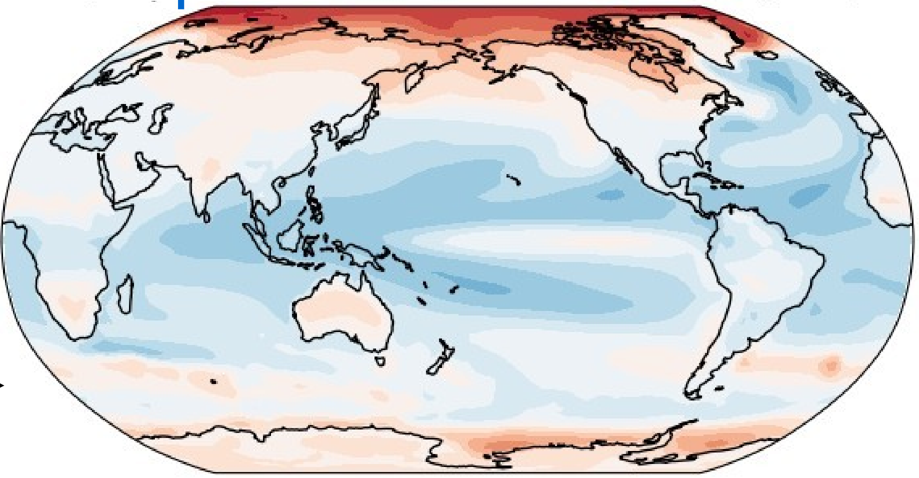
[-3.28]



Lapse-rate feedback

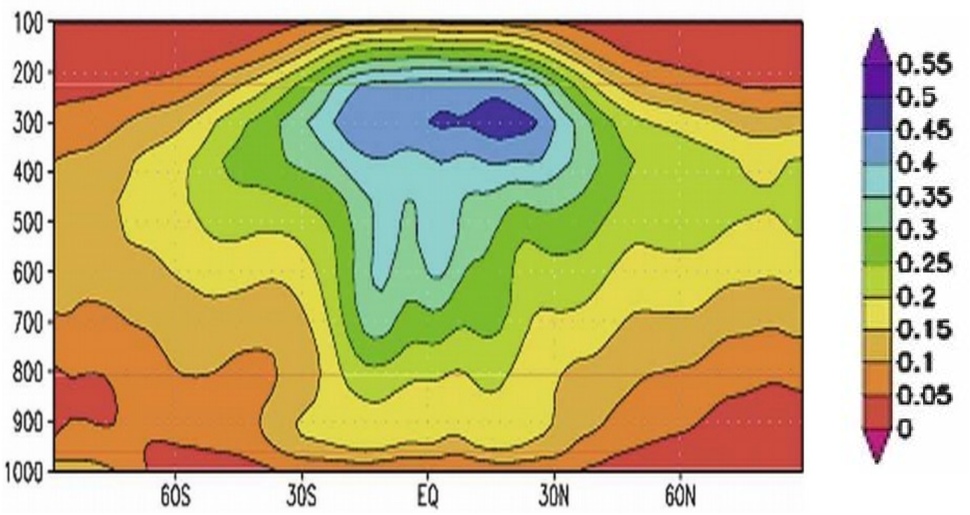
[-0.5]





[courtesy of M. Zelinka 2021]
(https://doi.org/10.5281/zenodo.5206851)

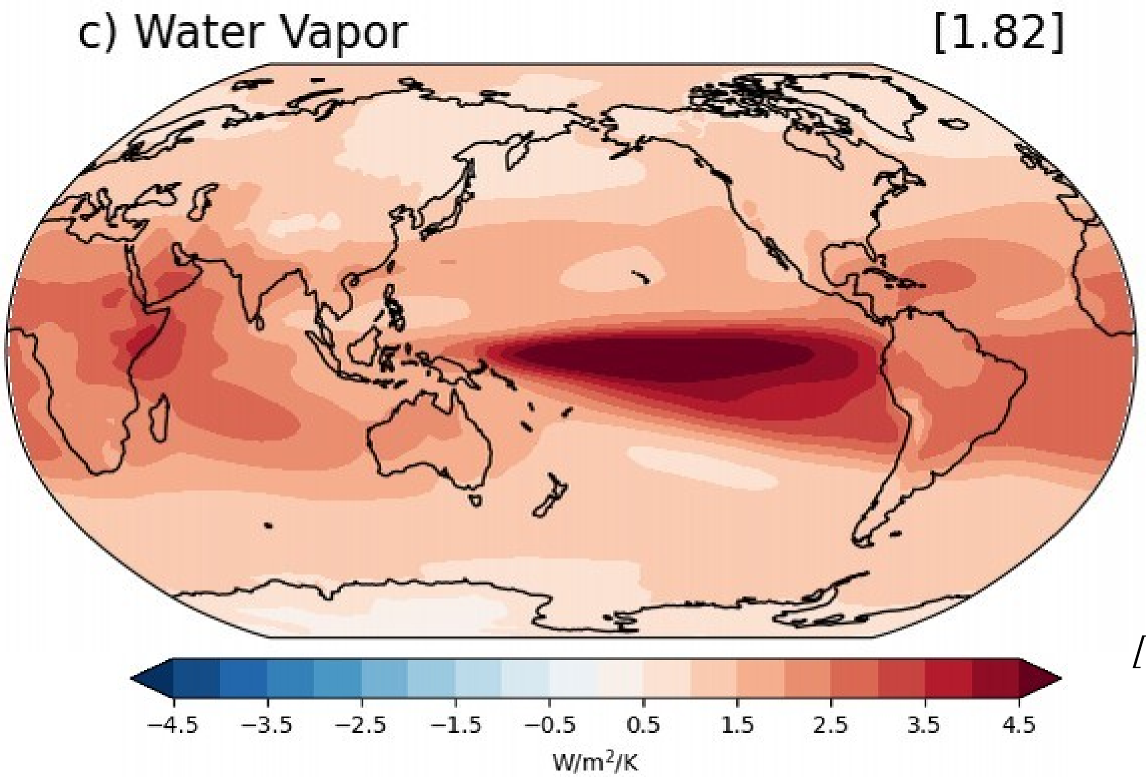
Water vapour feedback



$$\frac{\partial R}{\partial Q_a(P)}\frac{d\,Q_a(P)}{d\,T_s}$$

W/m^2/K/(100hPa)

Soden et al., J. Climate, 2008



[courtesy of M. Zelinka 2021]

Climate feedbacks

$$\Delta T_s^e = -\frac{\Delta Q}{\lambda}$$

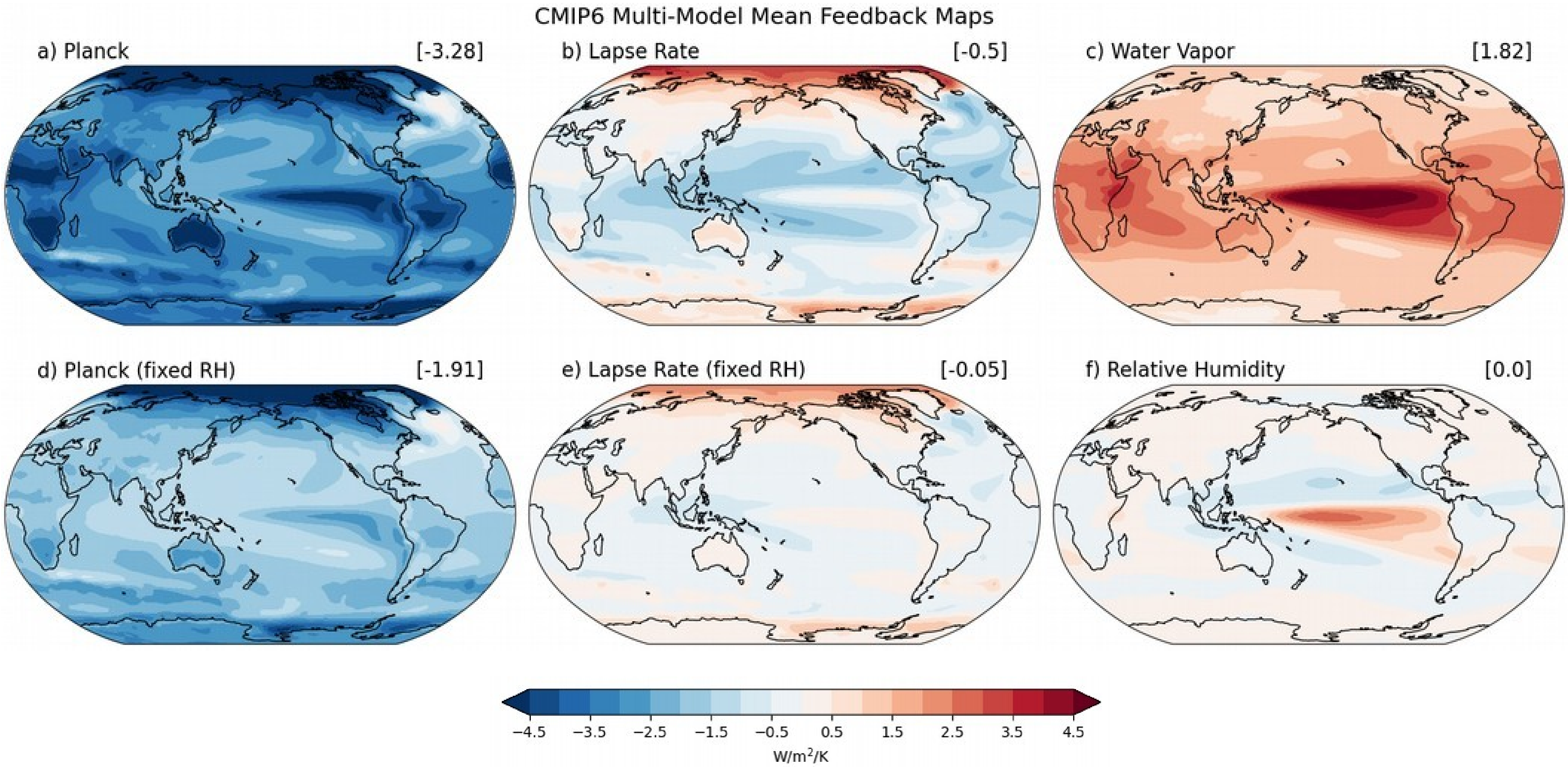
Classical decomposition (specific humidity)

$$\lambda = \underset{\text{Planck}}{\lambda_P} + \underset{\text{lapse rate}}{\lambda_L} + \underset{\text{water vapor}}{\lambda_W} + \underset{\text{clouds}}{\lambda_C} + \underset{\text{surface albedo}}{\lambda_\alpha}$$

Relative humidity decomposition (Held & Shell, 2012)

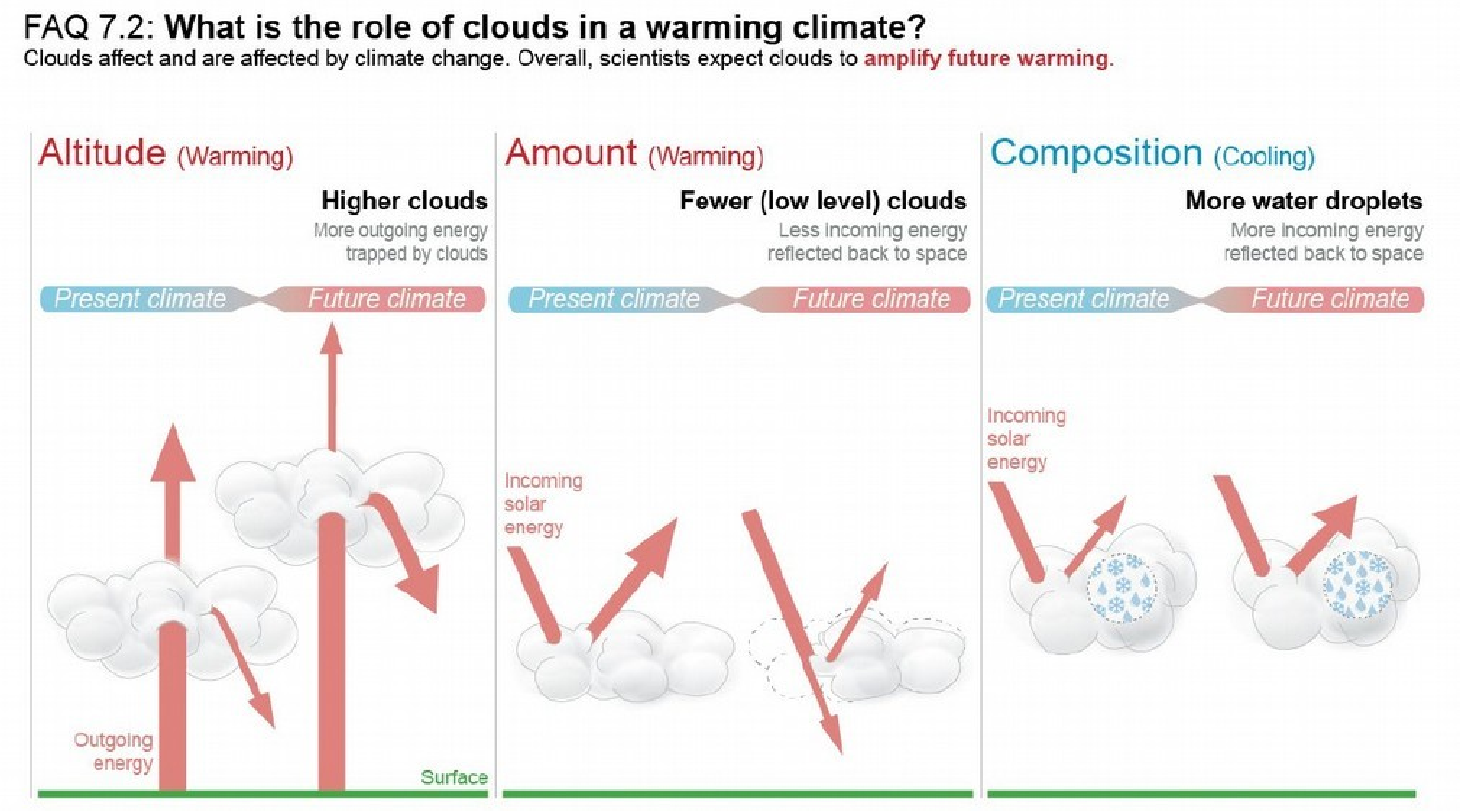
$$\lambda = \underbrace{\underset{\text{Planck}}{\lambda_P^*} + \underset{\text{lapse rate}}{\lambda_L^*}}_{\text{at constante relative himidity}} + \underset{\text{relative humidity}}{\lambda_R} + \underset{\text{clouds}}{\lambda_C} + \underset{\text{surface albedo}}{\lambda_\alpha}$$

Climate feedbacks with the absolute and relative humidity decompositions



[courtesy of M. Zelinka 2021]
(https://doi.org/10.5281/zenodo.5206851)

Cloud feedbacks



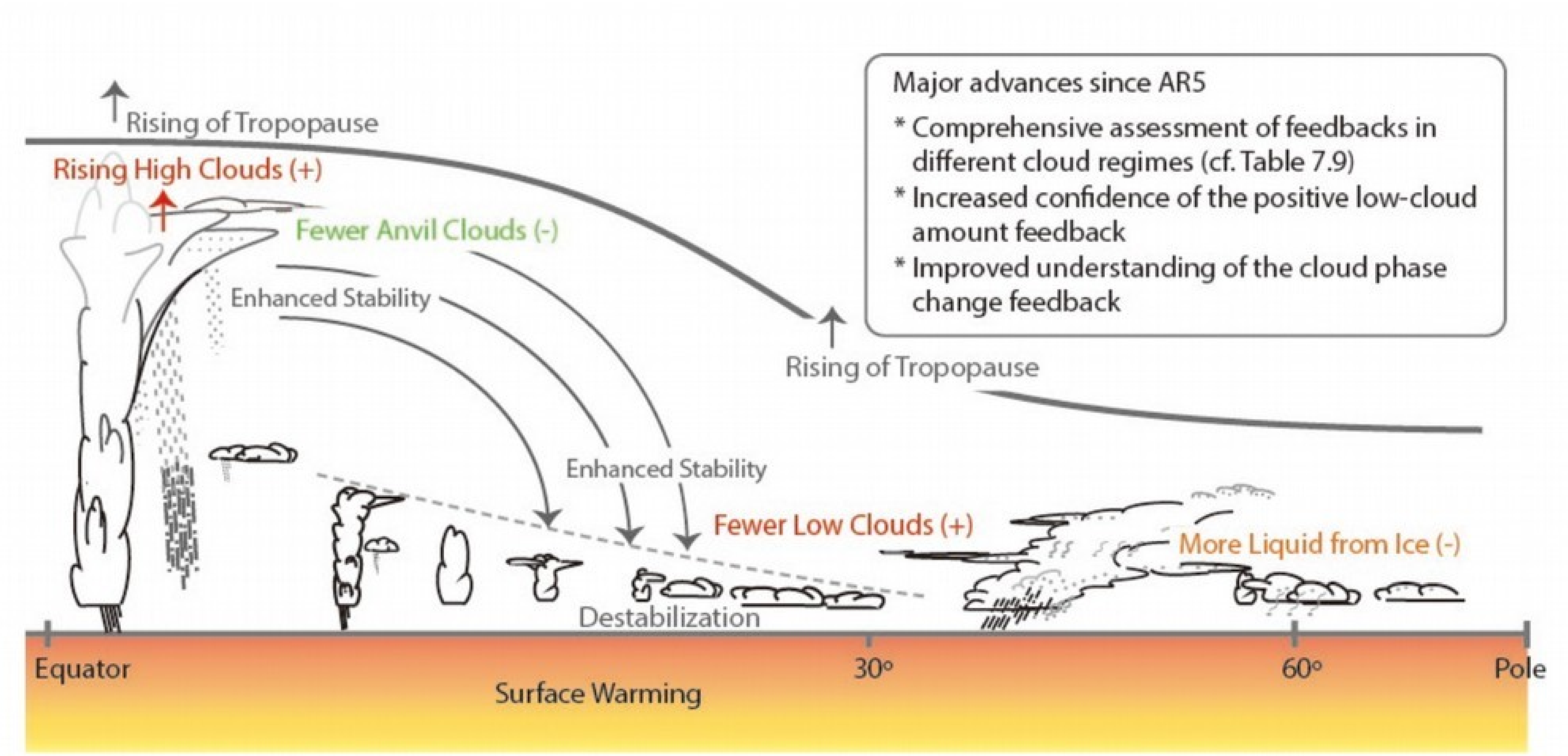
FAQ 7.2: **What is the role of clouds in a warming climate?**

Clouds affect and are affected by climate change. Overall, scientists expect clouds to **amplify future warming.**

Global warming is expected to alter the altitude (left) and the amount (centre) of clouds, which will amplify warming. On the other hand, cloud composition will change (right), offsetting some of the warming. Overall clouds are expected to amplify future warming.

[IPCC AR6 WG1, ch 7]

Cloud feedbacks



Major advances since AR5

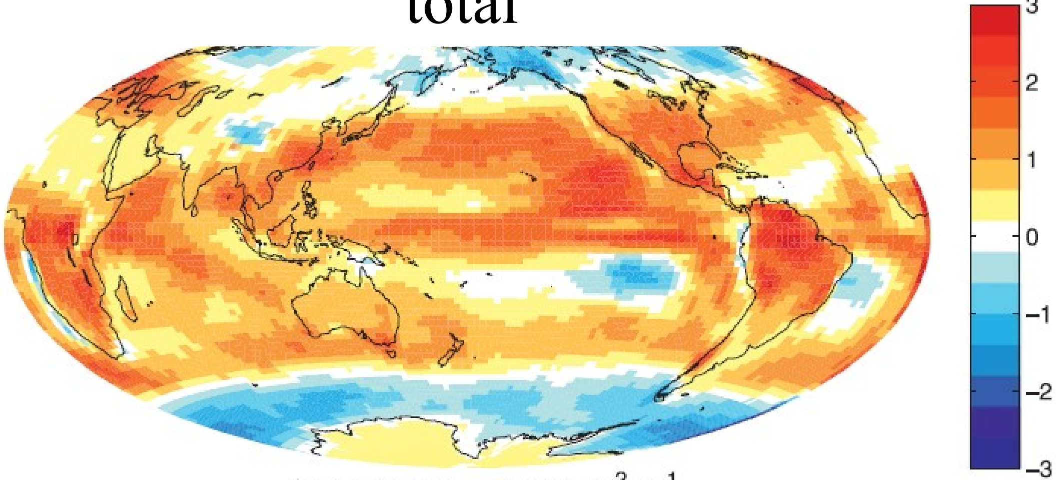
- * Comprehensive assessment of feedbacks in different cloud regimes (cf. Table 7.9)
- * Increased confidence of the positive low-cloud amount feedback
- * Improved understanding of the cloud phase change feedback

Schematic cross section of diverse cloud responses to surface warming. Thick solid and dashed curves indicate the tropopause and the subtropical inversion layer. Thin grey text and arrows represent robust responses. Text and arrows in red, orange and green show the major cloud responses assessed with high, medium and low confidence, respectively, and the sign of their feedbacks to the surface warming is indicated in the parenthesis.

[IPCC AR6 WG1, ch 7]

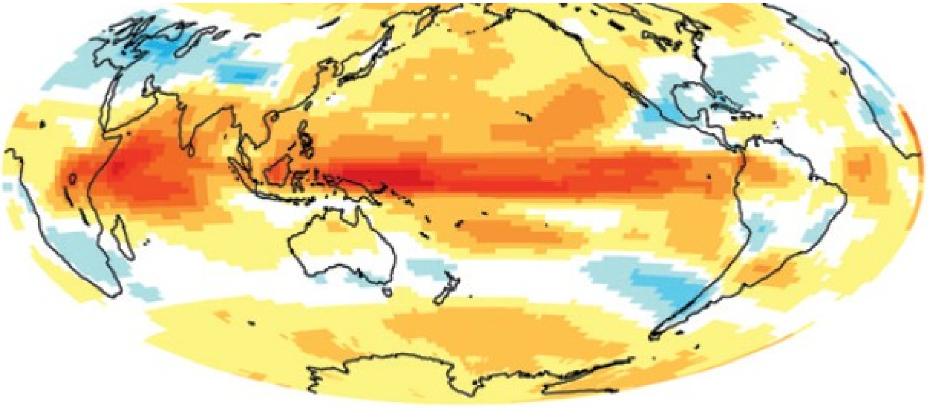
Cloud feedbacks

total



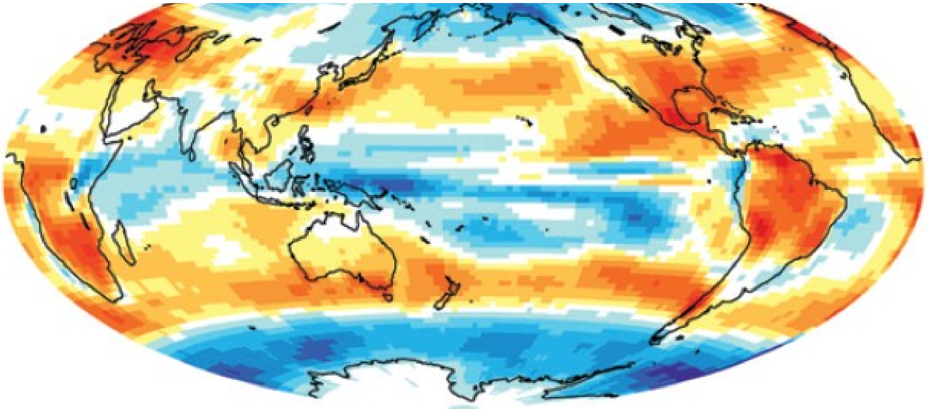
Global Mean = 0.61 W m^{-2} K^{-1}

LW (infrared)



Global Mean = 0.56 W m^{-2} K^{-1}

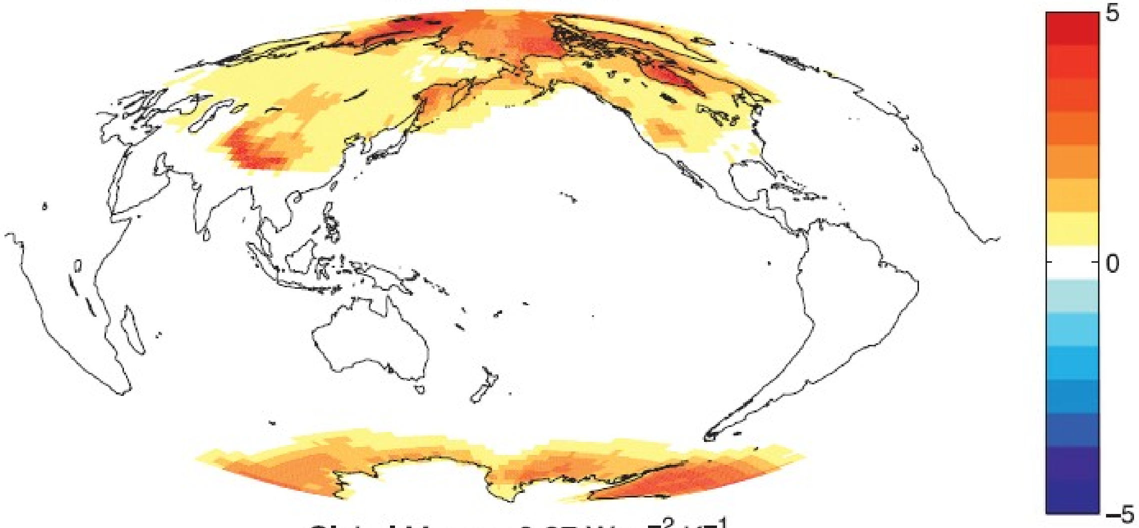
SW (solar)



Global Mean = 0.05 W m^{-2} K^{-1}

[Zelinka et al., 2012]

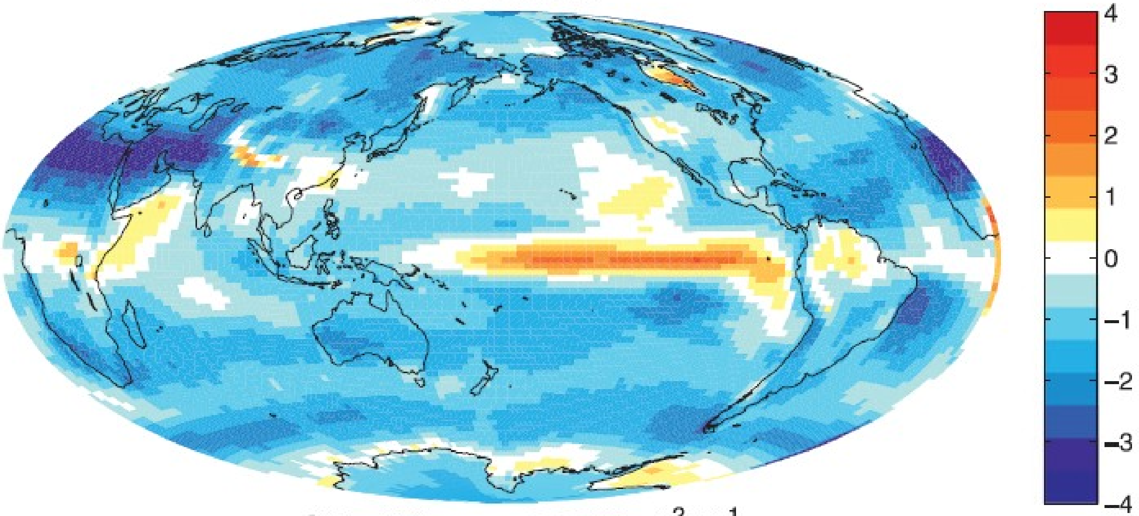
Surface albedo feedback



Global Mean = 0.27 W m^{-2} K^{-1}

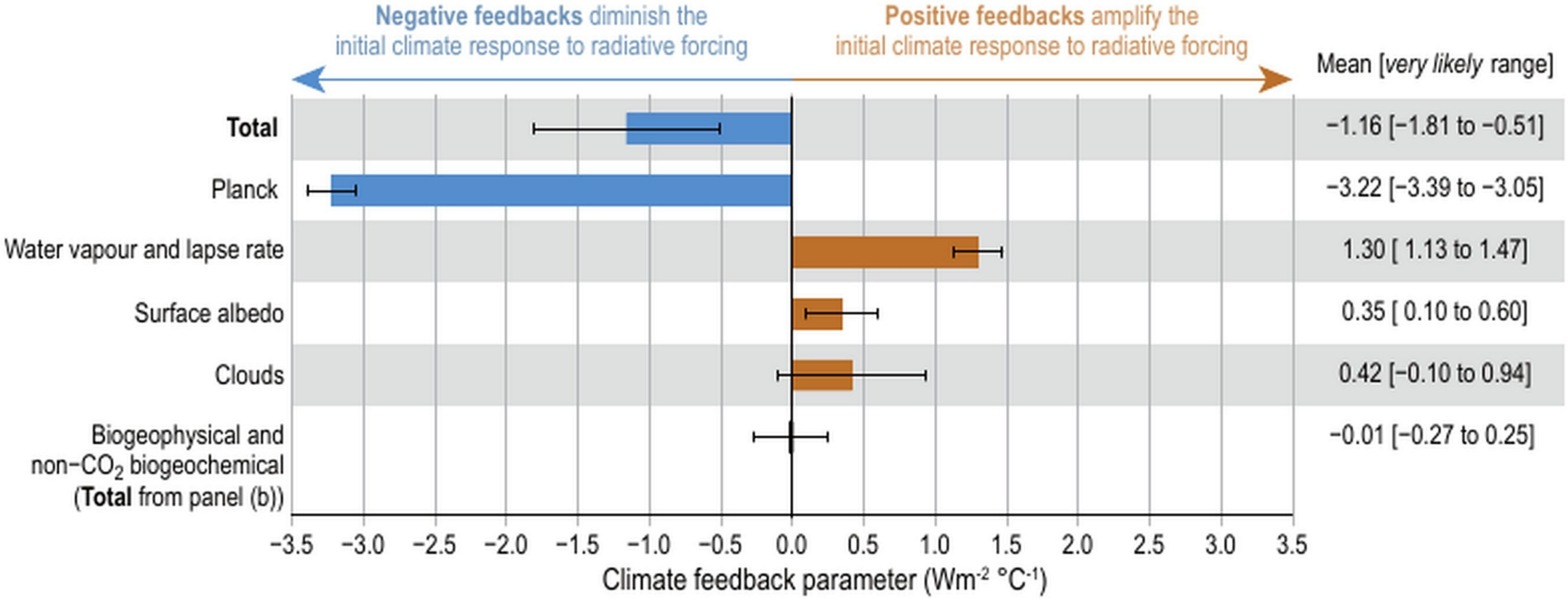
Climate total feedback

Net Feedback

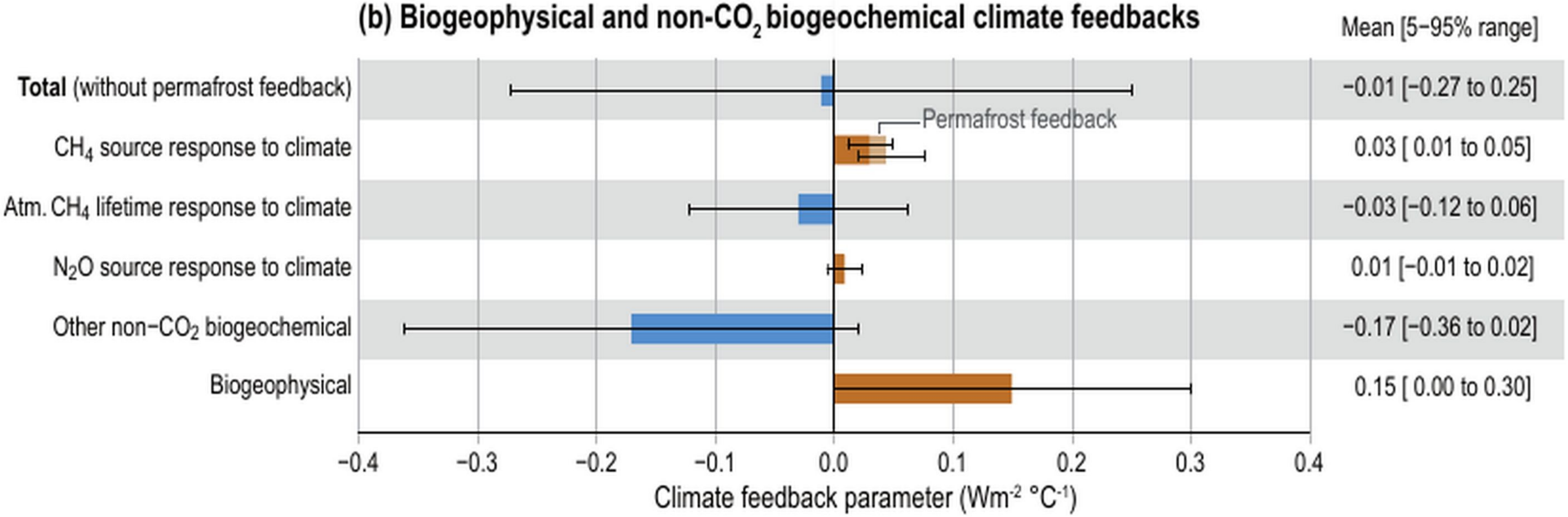


Global Mean = −0.98 W m^{-2} K^{-1}

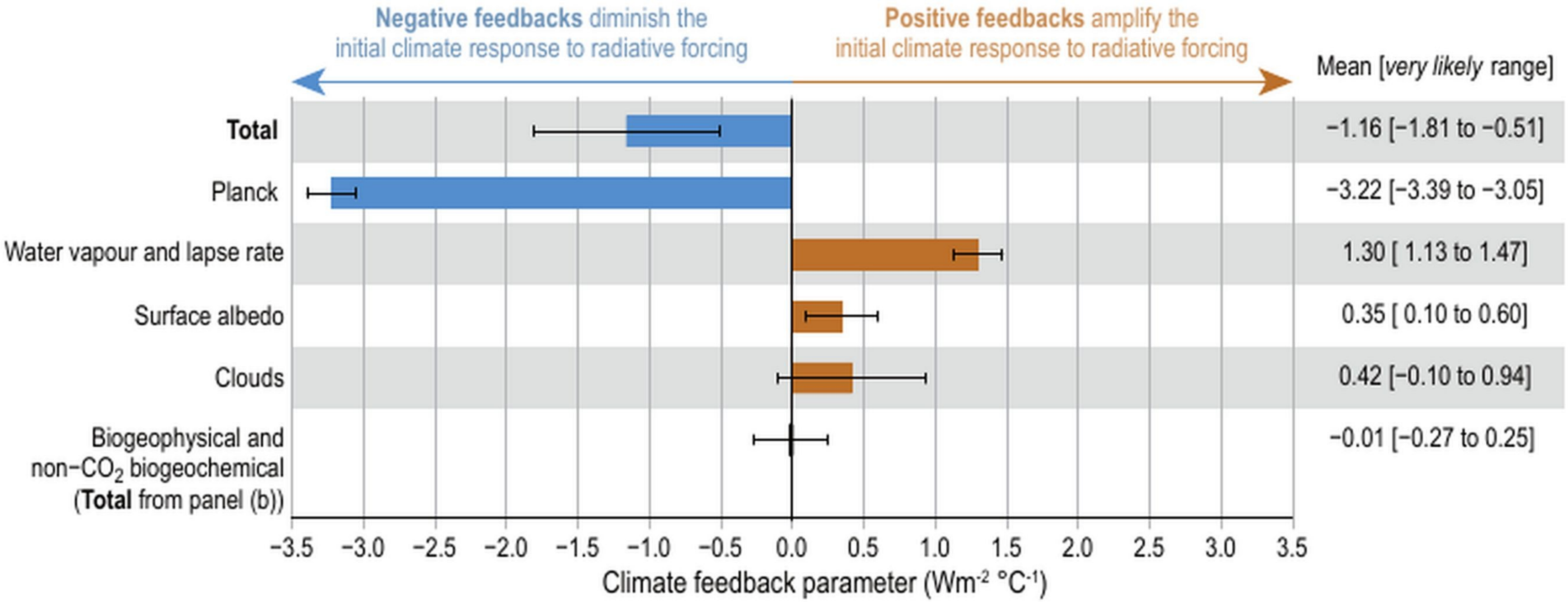
(a) Feedbacks in the climate system



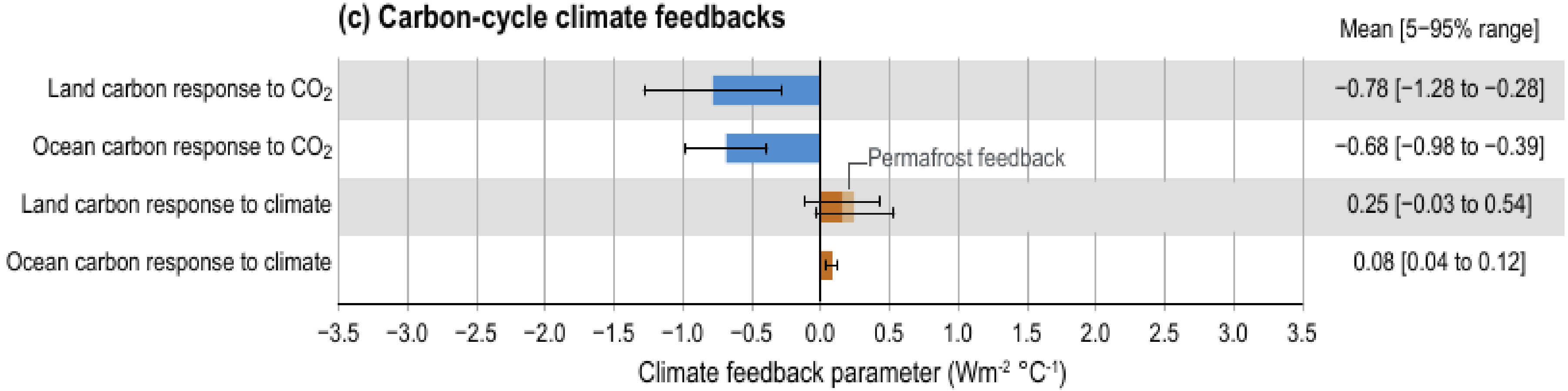
(b) Biogeophysical and non-CO_2 biogeochemical climate feedbacks



(a) Feedbacks in the climate system



(c) Carbon-cycle climate feedbacks



Outlook

Part I

- A short history
- Why do we care about climate sensitivity?
- Forcing and feedback in a simple idealized model
- Radiative forcing - climate feedback analysis framework
- The various physical climate feedbacks
- How much individual feedbacks contribute to global warming

How much individual feedbacks contribute to global warming

Equilibrium temperature response to a CO_2 doubling

Problem: The feedback parameters are additive, no the gains.

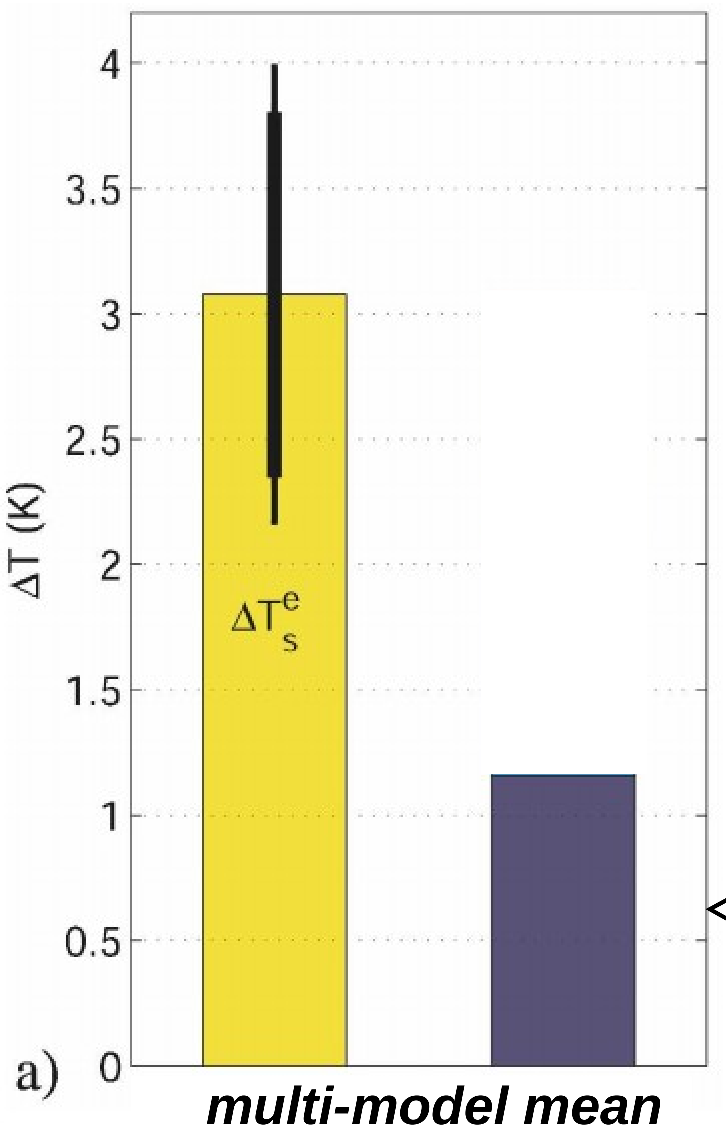
$$\Delta T = \Delta T_P + \sum_{x \neq P} \Delta T_x$$

$$\Delta T_x = \frac{g_x}{1-g} \Delta T_P \qquad g_x = -\frac{\lambda_x}{\lambda_P} \qquad g = \sum_x g_x$$

(Dufresne & Bony, 2008)

How much individual feedbacks contribute to global warming

Equilibrium temperature response to a CO_2 doubling



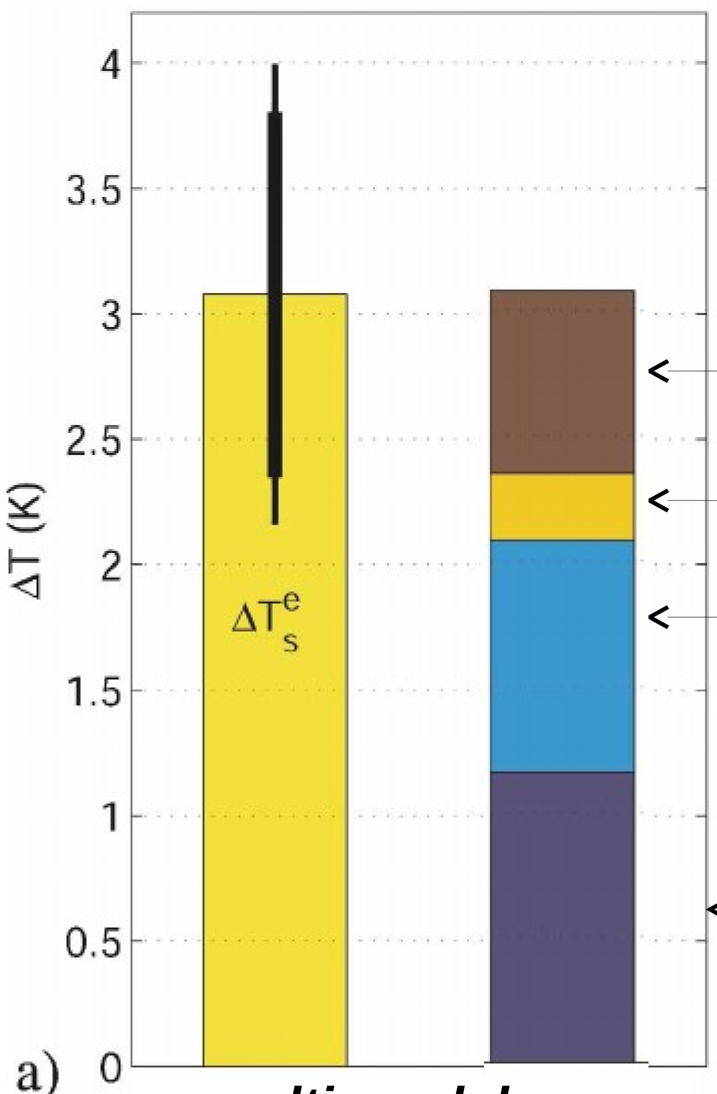
multi-model mean

Direct response to the forcing
Planck response

(Dufresne & Bony, 2008)

How much individual feedbacks contribute to global warming

Equilibrium temperature response to a CO_2 doubling



multi-model mean

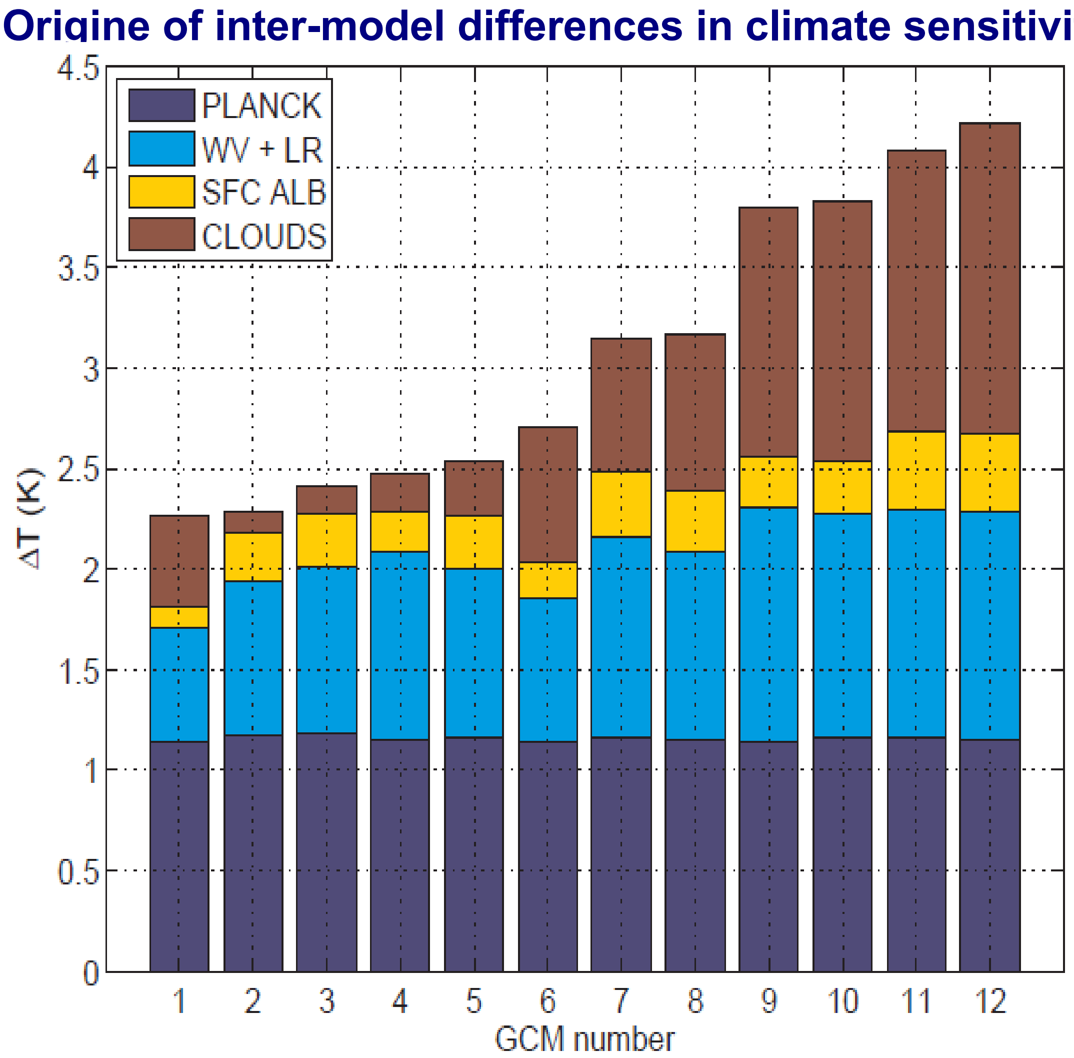
Climate feedbacks: Indirect response to the forcing
clouds
snow and ice (surface albedo)
water vapor

Direct response to the forcing
Planck response

(Dufresne & Bony, 2008)

How much individual feedbacks contribute to global warming

Equilibrium temperature response to a CO_2 doubling


Origine of inter-model differences in climate sensitivity :

ΔT

Cloud feedback

Surface albedo feedback

Water vapor +
lapse rate feedbacks

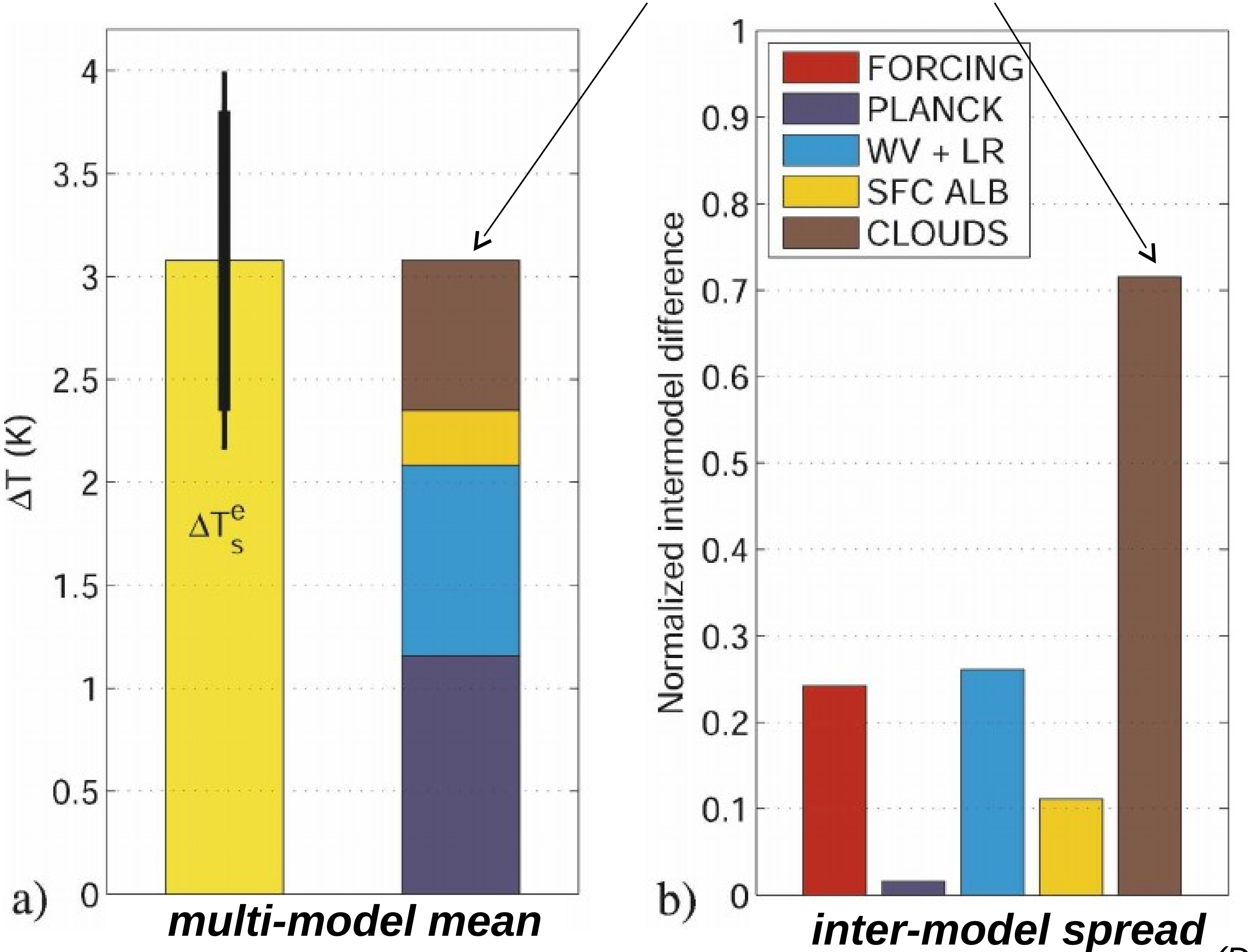
Planck response

(Dufresne & Bony, 2008)

How much individual feedbacks contribute to global warming

Equilibrium temperature response to a CO_2 doubling

Cloud feedback

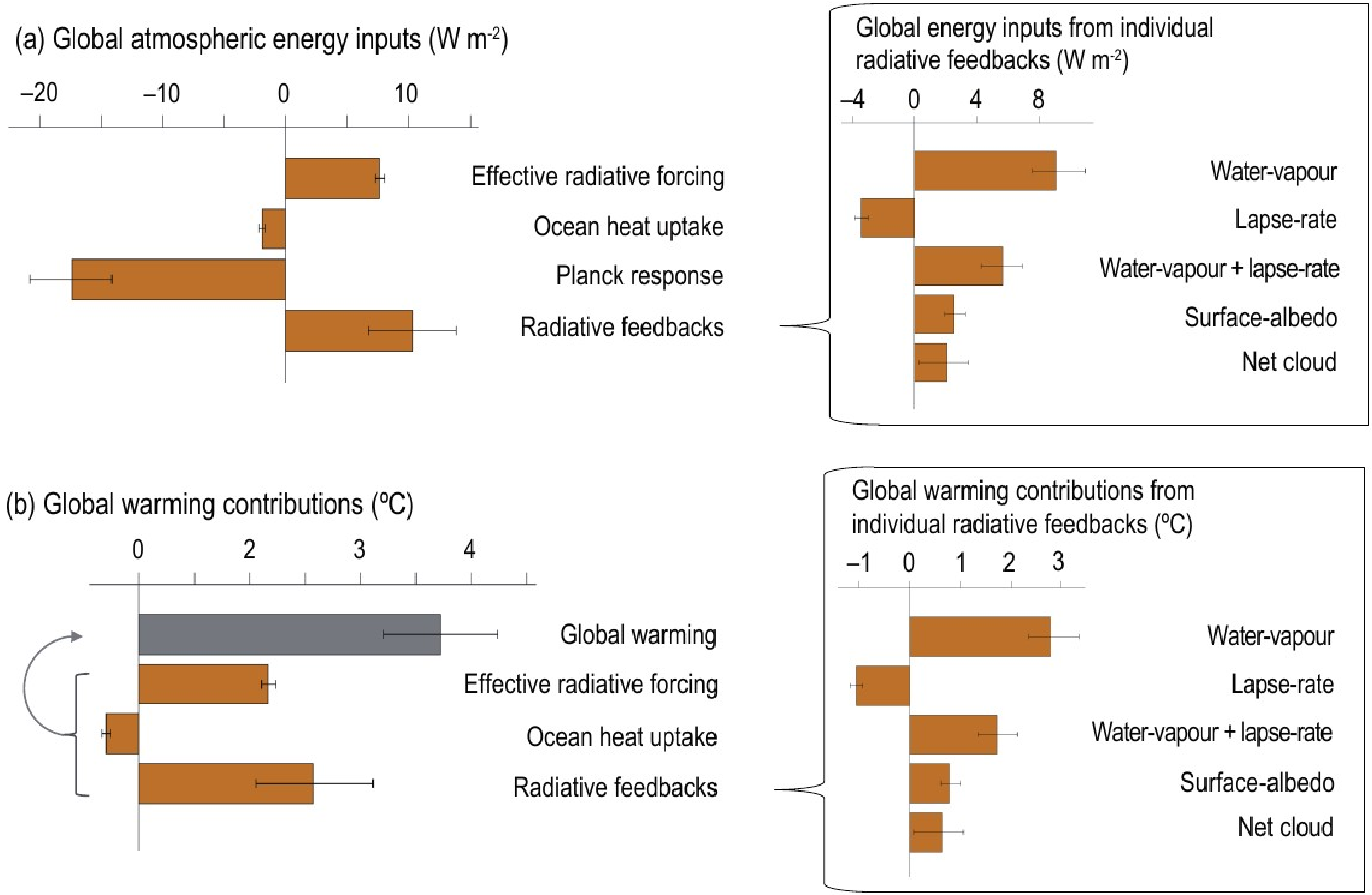


multi-model mean

inter-model spread

(Dufresne & Bony, 2008)

How much individual feedbacks contribute to global warming



[IPCC AR6 WG1, ch 7]

Thank you for your attention