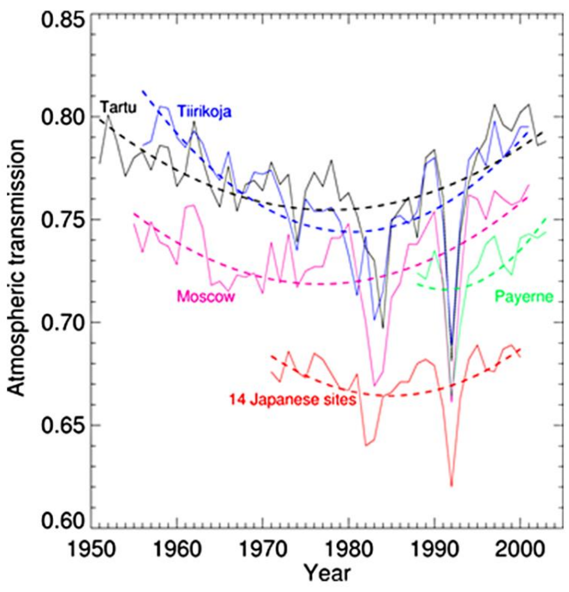
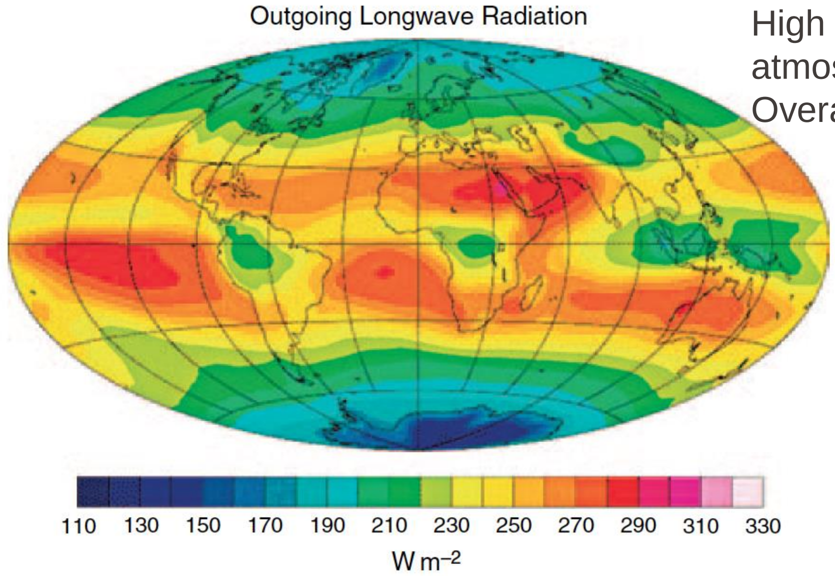
Recap of last lecture



Air pollution has led to a global dimming effect, masking the greenhouse effect



High clouds warme the atmosphere.
Overall clouds cool.

For each 1°C temperature increase, the atmosphere can hold 7 % more water vapor. Clausius-Clapeyron equation.

$$\frac{\Delta e_s}{e_s} = \frac{L_v}{T\,R_v} * \frac{\Delta T}{T} \approx 20\frac{\Delta T}{T}$$

General outline

No.	Date	Topics	Deadlines
1.	12.09.2024	Introduction	fill in Questionnaire in exercises (not graded)
2.	19.09.2024	Climate System, Radiation, Greenhouse effect	
3.	26.09.2024	Earth's energy balance, Radiative transfer,	
4.	03.10.2024	Aerosols & clouds, Radiative Forcing	Launch of poster assignment
5.	10.10.2024	**Feedback mechanisms, Climate Sensitivity**	
6.	17.10.2024	Paleoclimate	submission of Poster proposal (01.11.2024)
7.	31.10.2024	Climate variability	
8.	07.11.2024	Paris Agreement, Emission Gap, IPCC – present day climate change	
9.	14.11.2024	Extreme Events	
10.	21.11.2024	Climate scenarios (RCPs, SSPs), Tipping elements, 1.5 vs 2.0°C	submission of Poster draft
11.	28.11.2024	Carbon budget, carbon offsets, metrics	submission of assignment (graded)
12.	05.12.2024	Regional climate change	
13.	12.12.2024	Mitigation and adaptation, Climate Engineering	**Poster Conference** (graded)
14.	19.12.2024	Recapitulation of key points, questions and answers session	fill in Questionnaire in exercises (not graded)

Basics (1–6)

Present and future Climate change (6–11)

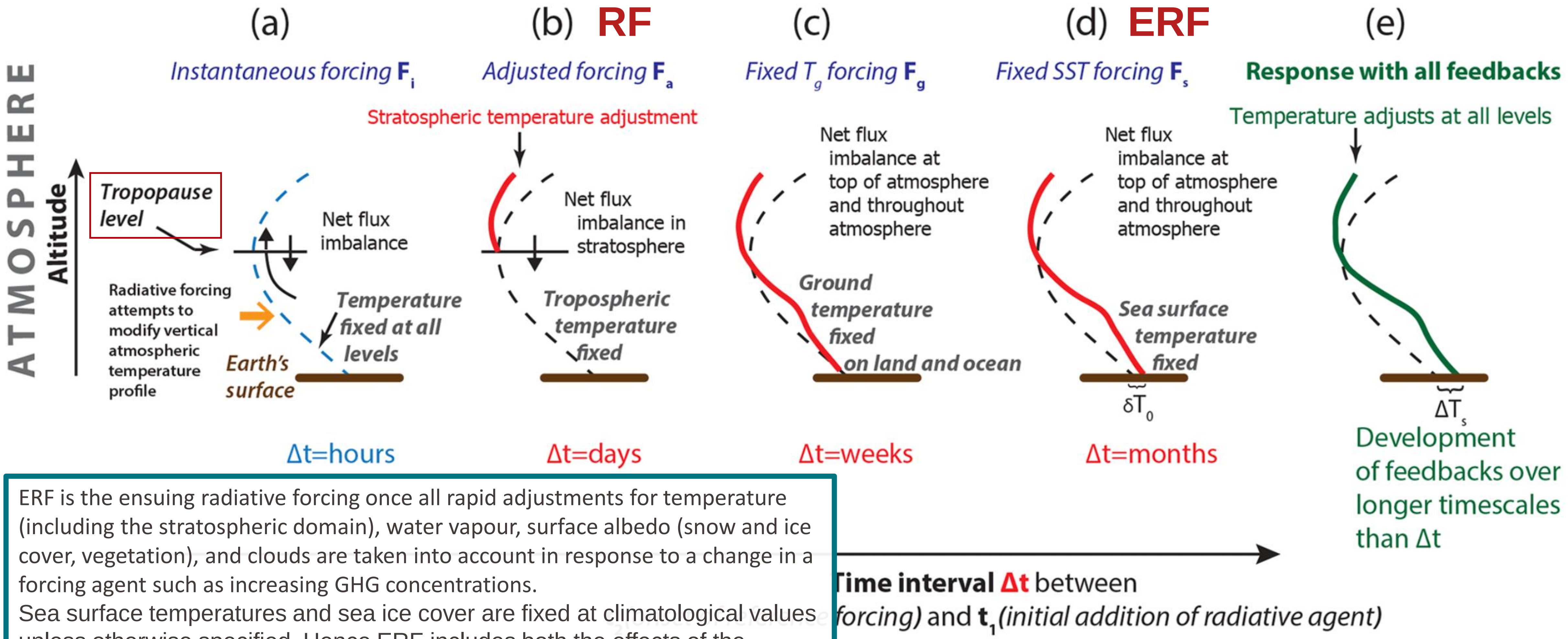
Actions (12–14)

Types of RF

RF = Radiative Forcing
ERF = Effective Radiative Forcing

used previously (and currently) → **RF**

current norm → **ERF**

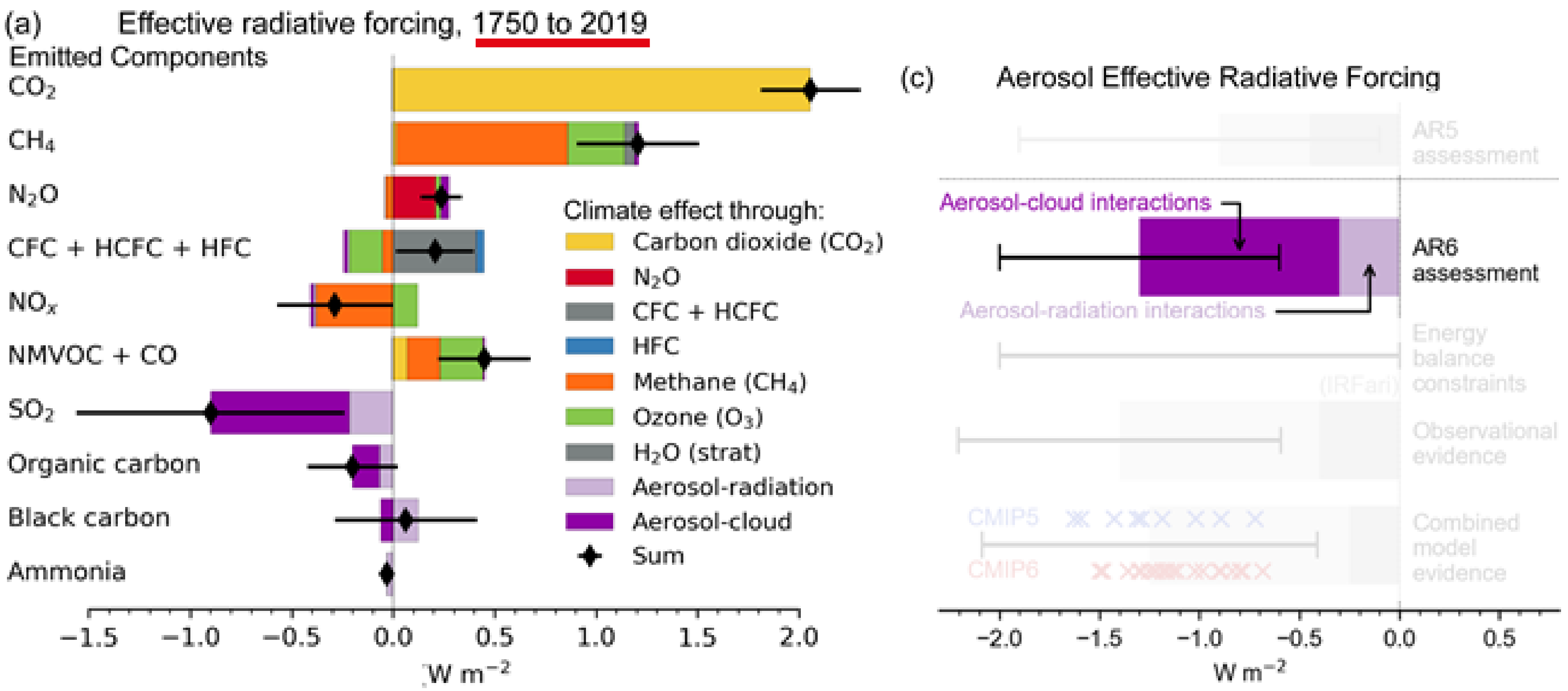


ERF is the ensuing radiative forcing once all rapid adjustments for temperature (including the stratospheric domain), water vapour, surface albedo (snow and ice cover, vegetation), and clouds are taken into account in response to a change in a forcing agent such as increasing GHG concentrations.
Sea surface temperatures and sea ice cover are fixed at climatological values unless otherwise specified. Hence ERF includes both the effects of the forcing agent itself and the rapid adjustments to that agent.

Heinze et al., ESD, 2019, https://doi.org/10.5194/esd-10-379-2019

ERF by component

GHG make up roughly 3.7 W m^{-2} radiative forcing
Compare: to TOA outgoing LW of -240 W m^{-2}



Negative leads to temperature decrease

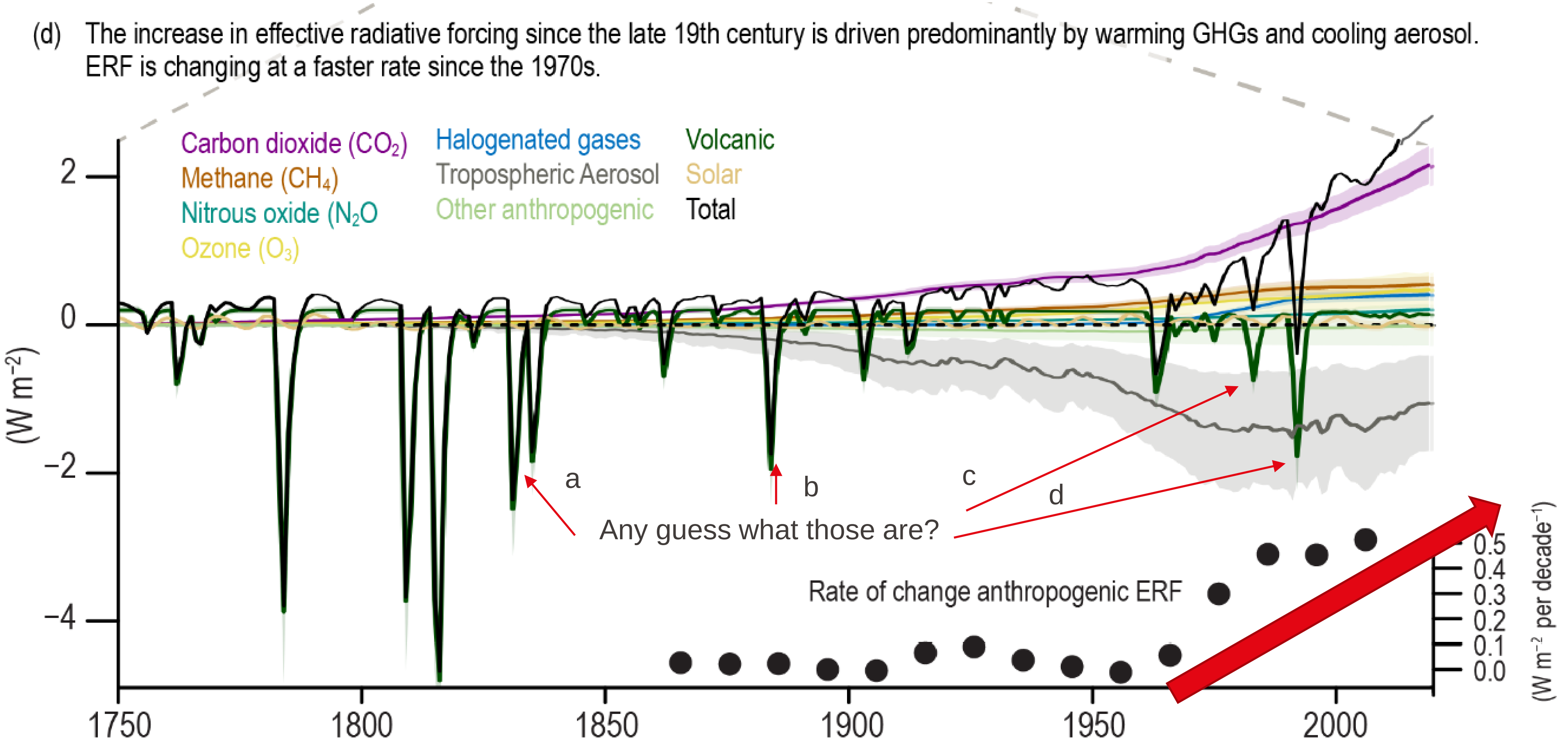
Positive forcing leads to temperature increase

- IPCC AR6, Fig. TS.15

Effects of most important GHG are well understood, the largest uncertainties are with aerosols and clouds.

ERF longterm perspective

a – Kelud 1826, Indonesia
b – Krakatoa 1883, Indonesia
c – El Chichón 1982, Mexico
d – Pinatubo 1991, Philippines

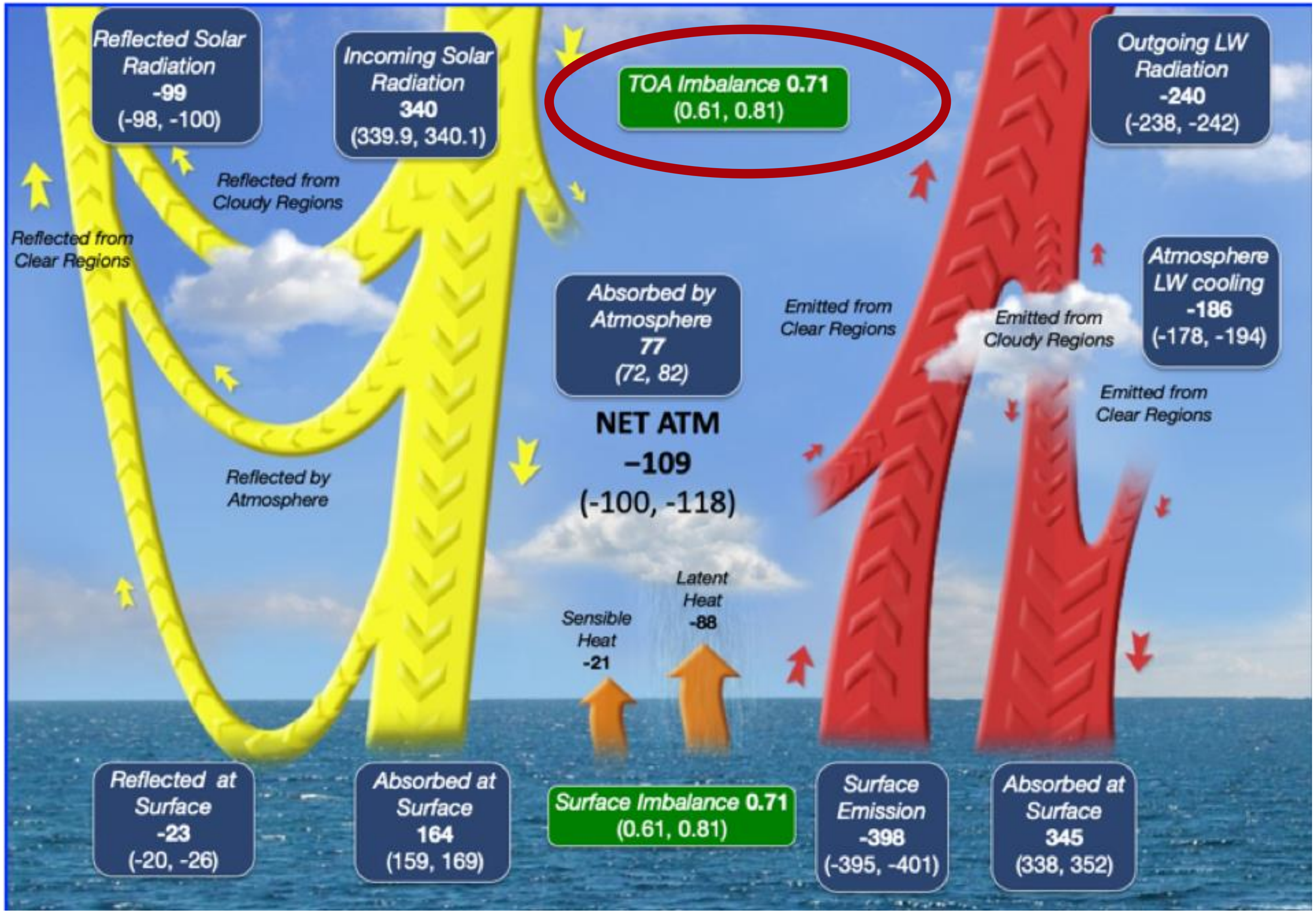


Rate of forcing is increasing as well!

- IPCC AR6, Fig. TS.9

Read here for more on volcanic eruptions

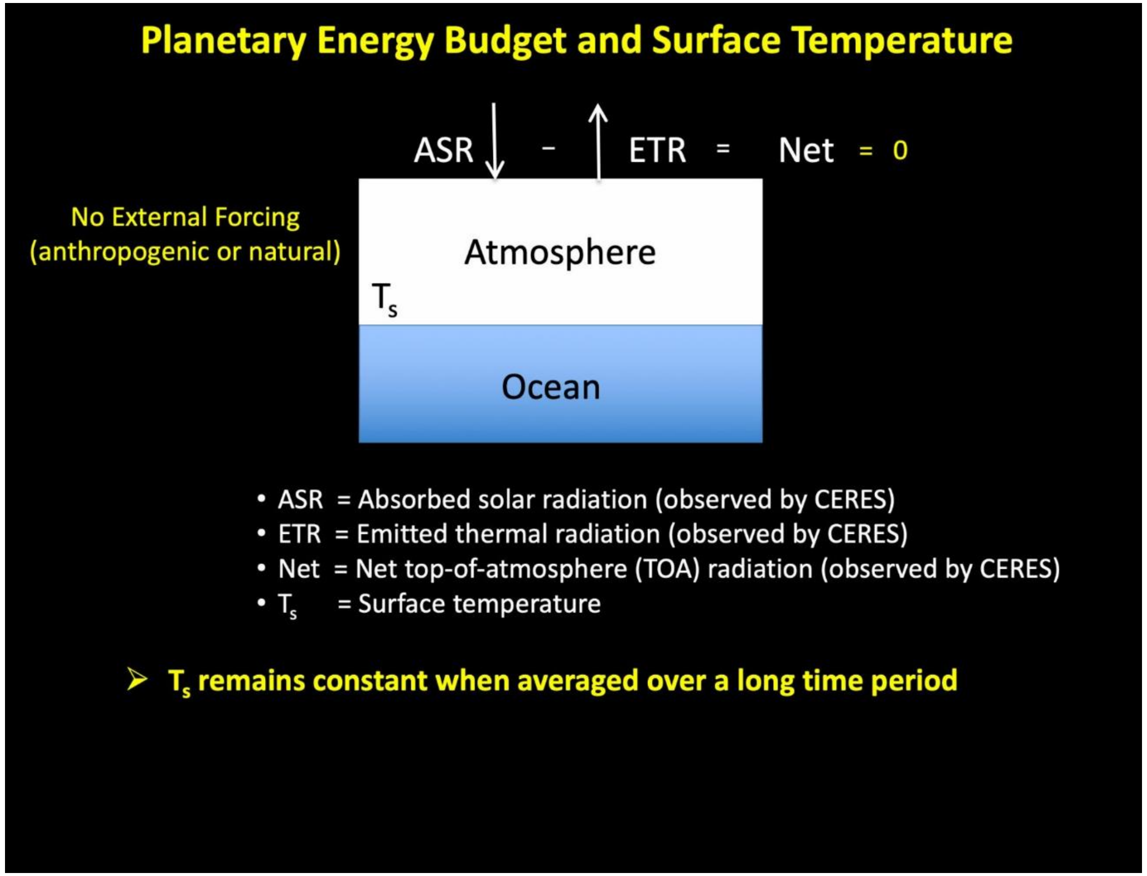
GHG ERF cannot be measured

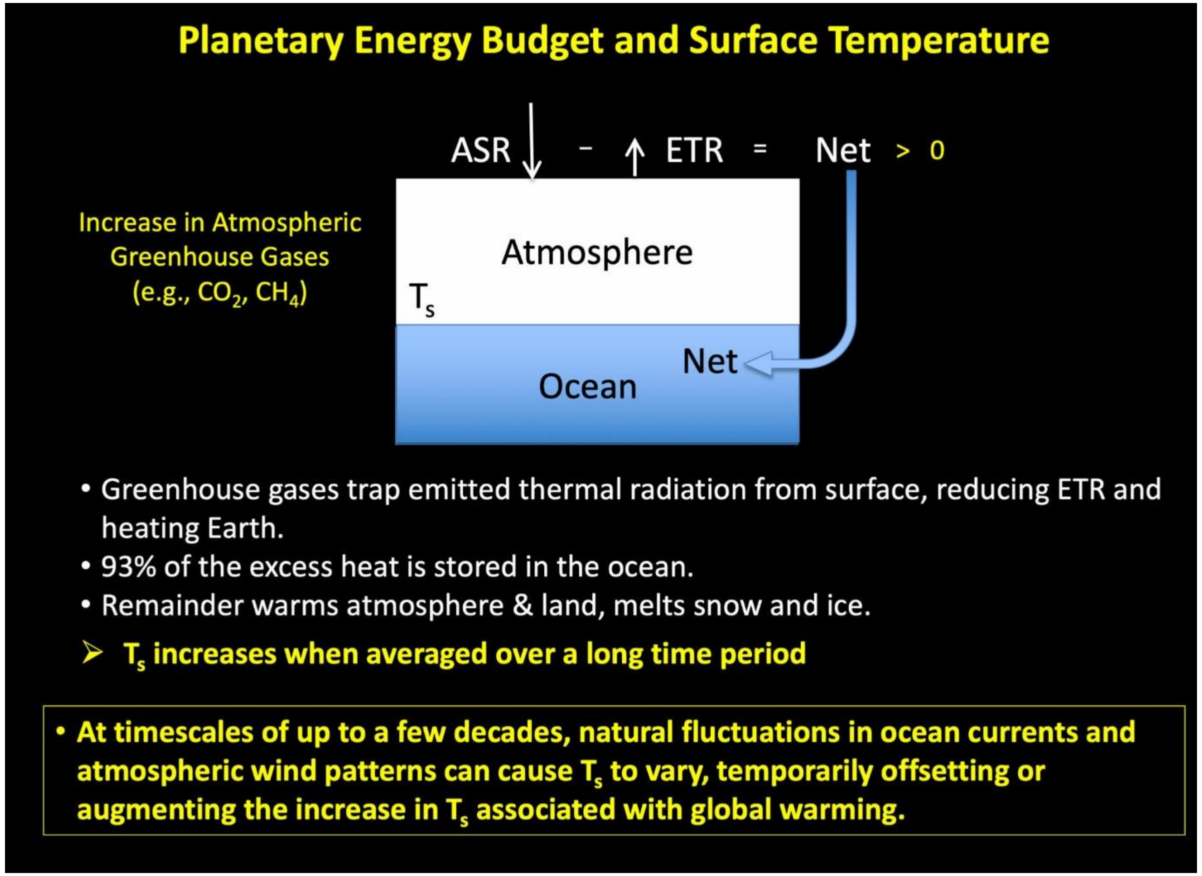


https://ceres.larc.nasa.gov/science/

Top of the atmosphere Earth's Energy Imbalance

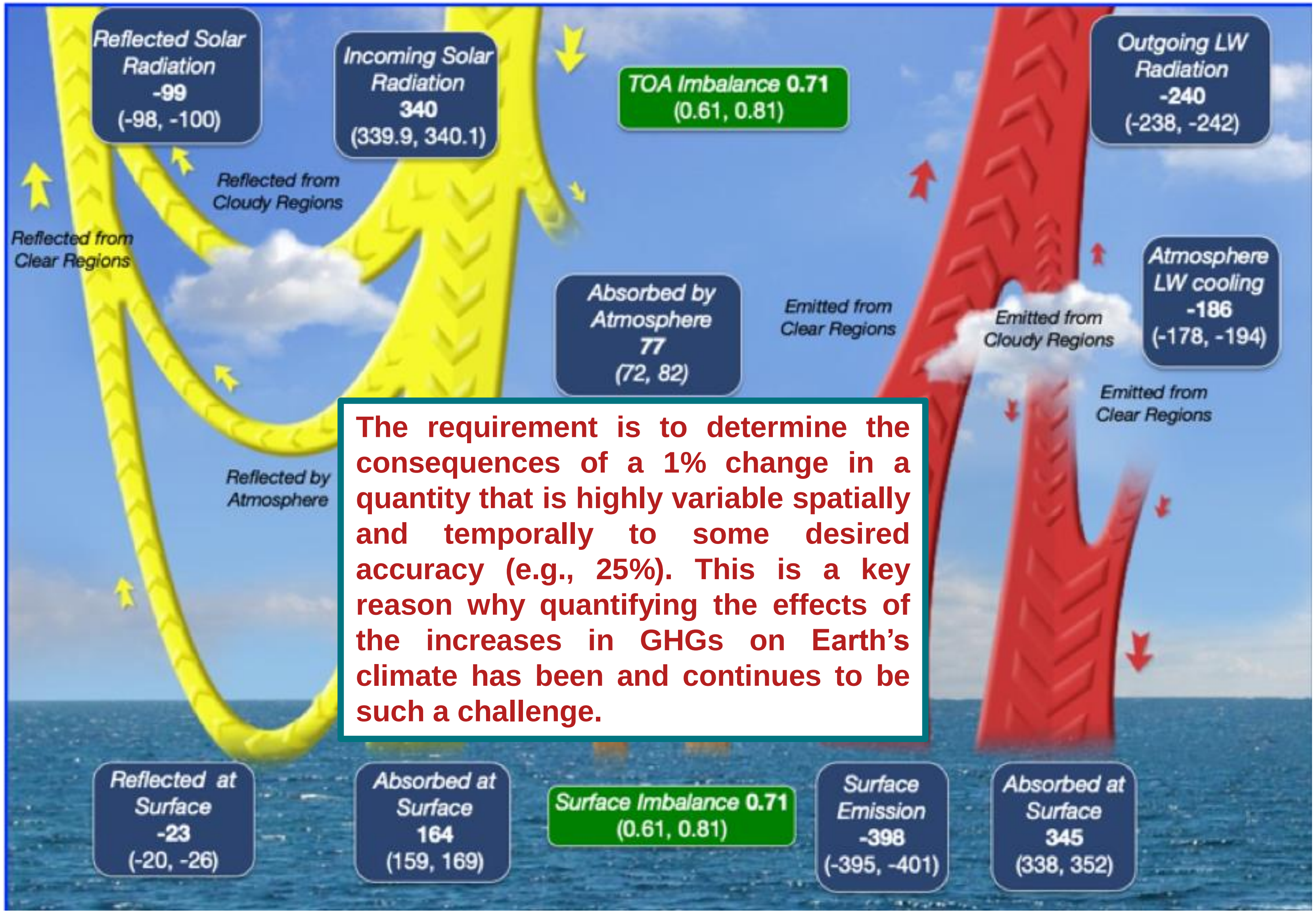
- The difference between incoming solar radiation and outgoing radiation determines the net radiative flux at TOA: If the imbalance is positive (i.e. less energy going out than coming in), energy in the form of heat is accumulated in the Earth system resulting in global warming.





https://ceres.larc.nasa.gov/science/

GHG ERF cannot be measured



The requirement is to determine the consequences of a 1% change in a quantity that is highly variable spatially and temporally to some desired accuracy (e.g., 25%). This is a key reason why quantifying the effects of the increases in GHGs on Earth's climate has been and continues to be such a challenge.

- TOA fluxes as well as most other fluxes within the atmosphere are one or several two orders of magnitude larger than the perturbation by GHG and aerosols.
- Hence, the RF by GHG and aerosols cannot be measured directly, but needs to be calculated.
- Calculation needs several ingredients:
  - Knowledge of concentration change (measurements)
  - Greenhouse gas properties: lifetime, effectiveness (next)
  - Models with realistic clouds, temperature and water vapor (upcoming)

https://ceres.larc.nasa.gov/science/

How sensitive is the climate system to the forcing?

Image source

Climate sensitivity

Generally, climate sensitivity refers to the amount of global mean surface warming (in °C or K) that will occur in response to a change of atmospheric CO_2 concentrations compared to preindustrial levels.

- «Magic» single number, which tells us how severe climate change will be?
- It's not that easy.
- But very important:
  - Hot or very hot? Makes a huge economic difference, health…

Climate sensitivity is a temperature.

- See also: Knutti et al., 2017; https://www.nature.com/articles/ngeo3017?proof=t

Climate Sensitivity Definitions

Equilibrium climate sensitivity (ECS) is the equilibrium annual global mean temperature response to a doubling of equivalent atmospheric CO_2 from pre-industrial levels and is thus a measure of the strength of the climate system's eventual response to greenhouse gas forcing. It takes into account changes in water vapour, lapse rate, clouds and surface albedo. The carbon cycle and other biogeochemical feedbacks, chemistry feedbacks, and slow feedback-like changes in vegetation types and ice sheets are deliberately not included in the concept of equilibrium climate sensitivity.

IPCC AR5: ECS of 1.5°C to 4.5°C;
CMIP6: ECS of 1.8°C to 5.6°C, **increased**

The **Earth system sensitivity (ESS)** includes very long-term Earth system feedbacks, such as changes in ice sheets or changes in the distribution of vegetative cover.

Transient climate response (TCR) is the annual global mean temperature change at the time of CO_2 doubling in a climate simulation with a 1% yr^{-1} compounded increase in CO_2 concentration. CO_2 doubling is reached after 70 years.
TCR is a measure of the strength and rapidity of the climate response to greenhouse gas forcing, and depends in part on the rate at which the ocean takes up heat. It differs from ECS because the distribution of heat between the atmosphere and oceans will not yet have reached equilibrium.

IPCC AR5: TCR of 1C to 2.5C;
CMIP 6: 1.7°C (1.3°C to 3.0°C), **increased**

TCR < ECS because of oceanic heat uptake, which is a slow process. In TCR experiments, the surface ocean remains cooler and hence IR emission remains lower.

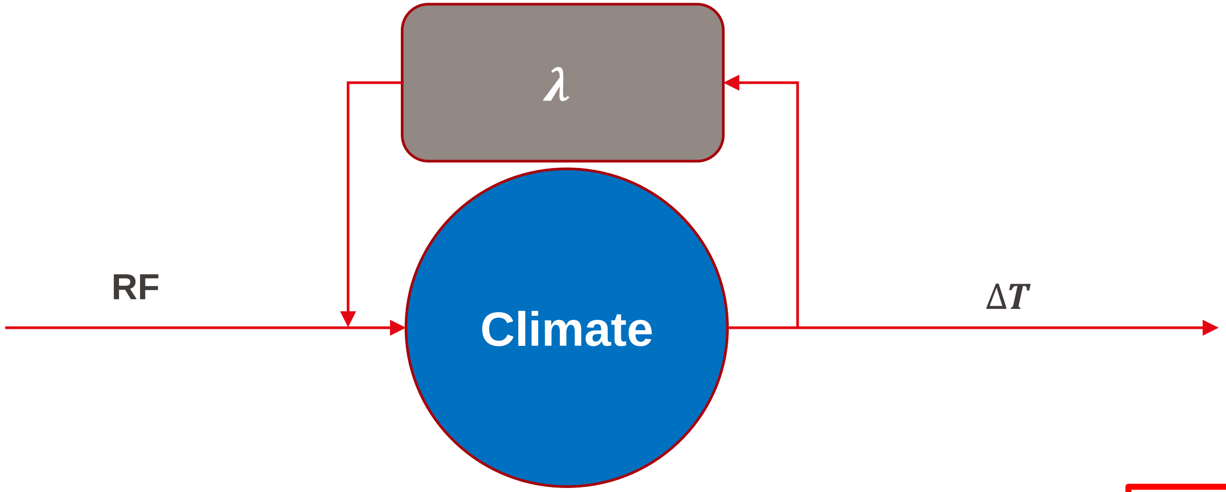
- https://www.carbonbrief.org/explainer-how-scientists-estimate-climate-sensitivity Heinze et al., ESD, 2019, https://doi.org/10.5194/esd-10-379-2019

Forcing, climate sensitivity and feedback

$$N = RF + \lambda \Delta T$$

N: net top of atmosphere energy balance,
RF: radiative forcing,
ΔT: global surface temperature response, climate sensitivity
λ : feedback factor

Positive feedback amplifies the response
Negative feedback dampens the response



Forcing, climate sensitivity and feedback are fundamental concepts to understand climate change.

To derive the climate sensitivity, we need to know the feedback factor.

Energy balance model: A simple climate model

Backbone of any climate model!

1. Net downward flux at TOA

$$F_{TOA} = (1-\alpha)Q - \sigma T_e^4$$

σ Boltzmann constant (5.67*10^{-8} Wm^{-2}K^{-4})
T_e equivalent temperature
Q area-weighted mean incoming solar flux (341.3 Wm^{-2})
α planetary albedo (0.3)

2. Planetary energy budget

$$C\frac{dT_s}{dt} = F_{TOA}$$

T_s average surface temperature
C atmosphere-ocean column effective heat capacity (Jm^{-2}K^{-1})

3. Adding radiative forcing

$$C\frac{d\Delta T_s}{dt} = RF + \Delta F_{TOA}$$

ΔT_s change of global mean surface temperature
RF = radiative forcing
T_s changes because of RF and resulting changes in radiative processes (internal to the climate system) ΔF_{TOA}

4. The **key assumption** in climate feedback analysis is that *changes in radiative flux are proportional to surface temperature changes*:

$$\Delta F_{TOA} = \lambda * \Delta T_s$$

λ is a constant of proportionality: climate feedback factor (W m^{-2} K^{-1})

$$C\frac{d\Delta T_s}{dt} = RF + \lambda * \Delta T_s$$

How much warming do we expect with a given radiative forcing?
Or, how much warming do we expect if we double the CO_2 concentration, which is about 3.7 W m^{-2}?

5. With sufficient time the system will equilibrate:

$$C\frac{d\Delta T_s}{dt} = 0 = RF + \lambda * \Delta T_s$$

$$\Delta T_s = -\frac{RF}{\lambda}$$

$$\lambda = \lambda_0 + \lambda_1 + \lambda_2 + \ldots = \sum_{i=0}^{n} \lambda_i$$

Many different climate feedbacks

λ>0, positive feedback
λ<0, negative feedback

What's the sign of λ?

Energy balance model: A simple climate model

Previous slide

$$C\frac{dT_s}{dt} = F_{TOA} = (1-\alpha)Q - \sigma T_e^4$$

Need to reconcile T_s and T_e

This is a simple climate model (only physics)!

6. Parameterize with T_s only.

$$C\frac{dT_s}{dt} = (1-\alpha)Q - \sigma(\beta T_s)^4$$

β measures the proportionality between the surface and emission temperature

$\beta = T_e/T_s$ = 255 K / 288 K = 0.885

7. Defnining λ

$$C\frac{d\Delta T_s}{dt} = RF + \lambda * \Delta T_s$$

8. Calculate λ

λ = **-3.3 W m^{-2} K^{-1}**

What does this mean?

For every 1 W m^{-2} radiative forcing, our planet must warm by -1/λ = 0.3 K to establish equilibrium.

Note:

This model only represents a **single feedback process**: the increase in longwave emission to space with surface warming.

This is called the **Planck feedback** because it is fundamentally due to the Planck blackbody radiation law (warmer temperatures = higher emission).

λ_0 = **-3.3 W m^{-2} K^{-1}**

$$N = RF + \lambda\Delta T$$

N: net top of atmosphere energy balance,
RF: radiative forcing,
ΔT: global surface temperature response,
λ : feedback factor

Types of feedback – fast physical

Atmospheric thermodynamic feedbacks *(most certain quantification)*
1 Planck response (black body radiation)
2 The combined water vapour and lapse rate feedback

Cloud feedbacks *(complex and large source of uncertainty)*
1 Rise of cloud top feedback
2 Tropical low -cloud feedback
3 Mid-latitude cloud reflectance feedback
4 Cloud water phase feedback

Fast surface feedbacks
1 Snow albedo feedback
2 Soil moisture evapotranspiration feedback and CO_2 stomata-water feedback

Fast ocean feedbacks
1 ocean mixed-layer and ocean thermocline feedbacks
2 Tropical circulation responses to a warming climate

Sea ice feedbacks
1 Sea ice albedo feedback

Color legend

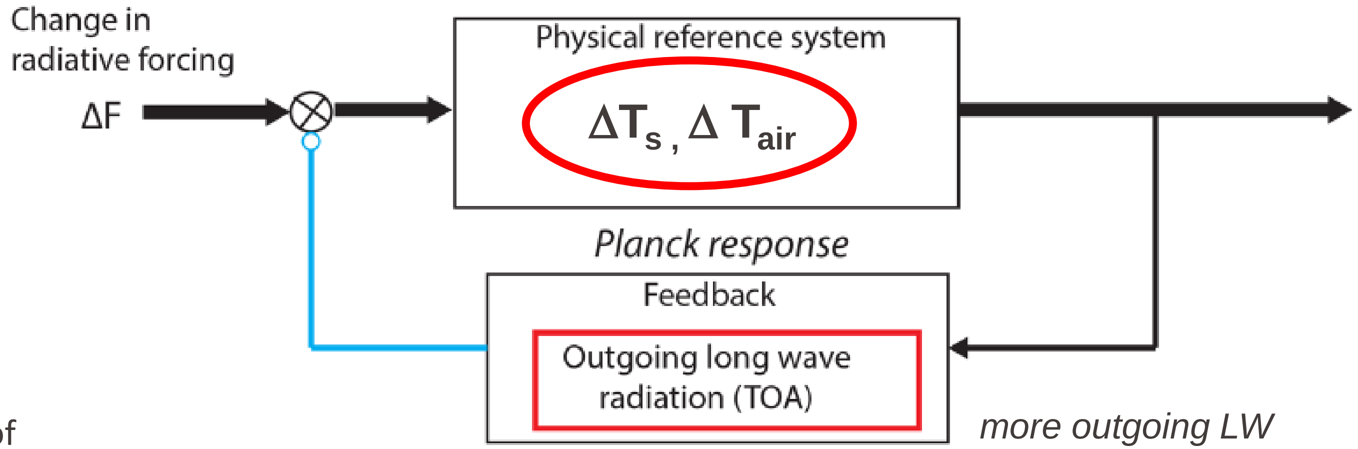
Thermal longwave (LW) heat redistribution including water vapour and moisture

Shortwave (SW) reflectivity / albedo

Both LW and SW effects

Heinze et al., ESD, 2019, https://doi.org/10.5194/esd-10-379-2019

Planck response



Arrows indicate positive coupling;
open circles indicate negative coupling.
Changes in state variables are indicated in ellipses.

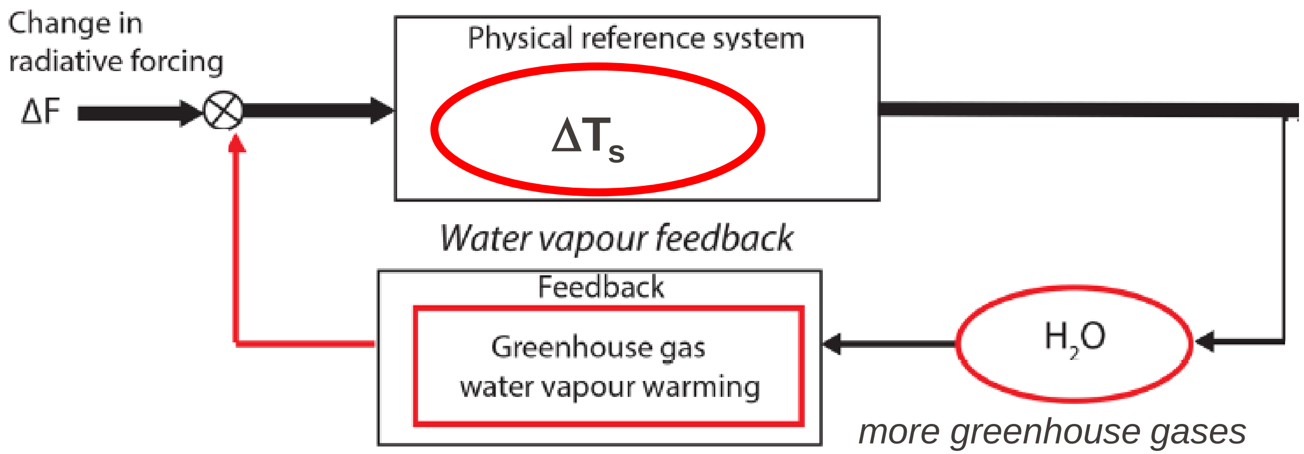
Red indicates increasing variable values, strengthening of processes, or positive feedback; blue indicates the opposite.

ΔT_s change in surface temperature
ΔT_{air} change in air temperature

- **Planck response**:
  - Single largest and negative feedback. Based on Stefan-Boltzmann law. Increase of LW radiation to space due to surface warming. Only if the Planck feedback is overcome by other positive feedbacks, a runaway greenhouse effect can be expected.

- Heinze et al., ESD, 2019, https://doi.org/10.5194/esd-10-379-2019

Water vapour



Arrows indicate positive coupling;
open circles indicate negative coupling.
Changes in state variables are indicated in ellipses.

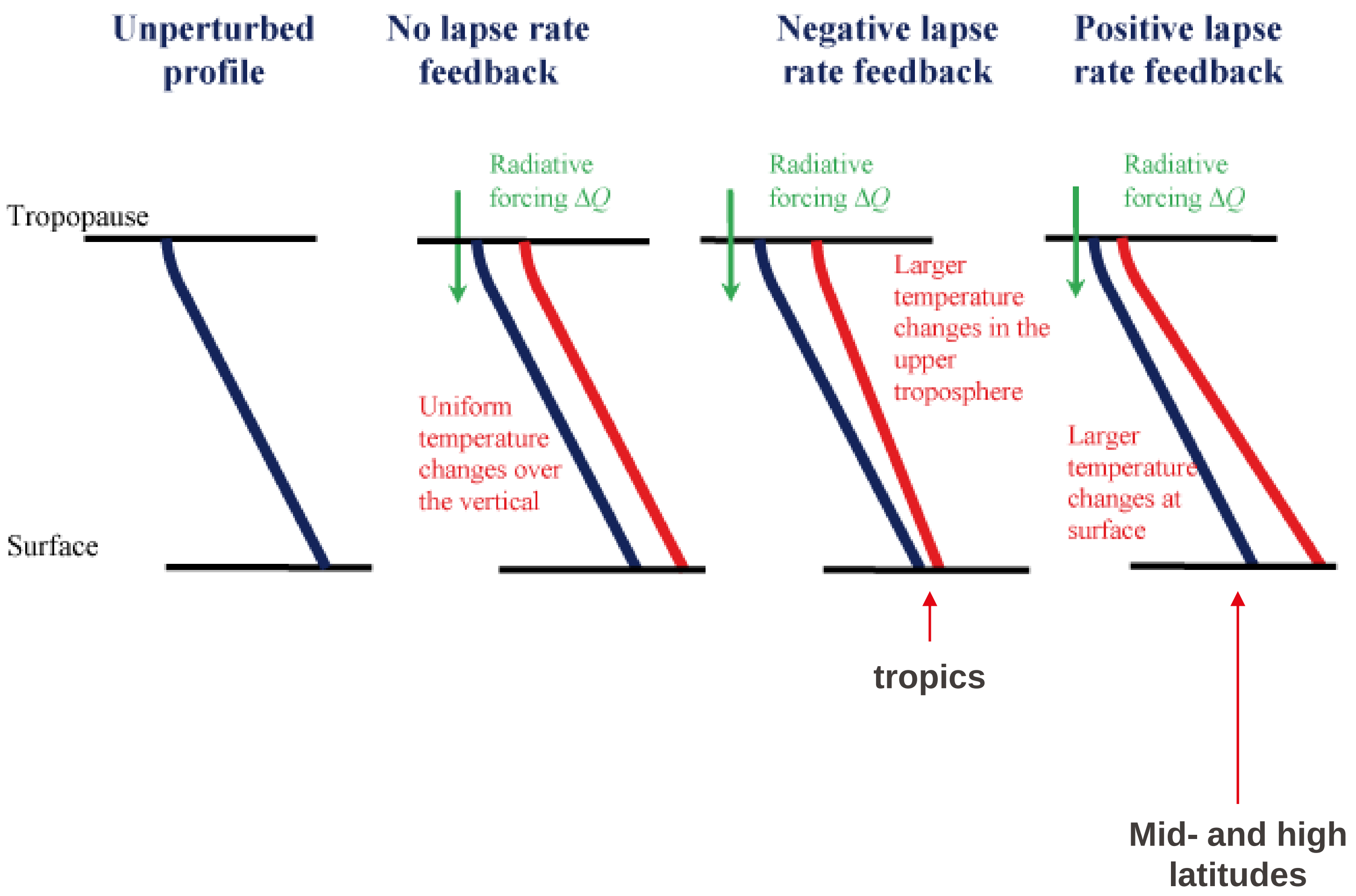
Red indicates increasing variable values, strengthening of processes, or positive feedback; blue indicates the opposite.

ΔT_s change in surface temperature
ΔT_{air} change in air temperature

- **Water vapour feedback**:
  - The warmer the atmosphere, the more water vapour it can hold. Water vapour is a greenhouse gas. The RF resulting from water vapour is roughly proportional to the logarithm of its concentration. Hence more warming occurs, where the unperturbed situation is rather dry (higher troposphere, cold regions).

- Heinze et al., ESD, 2019, https://doi.org/10.5194/esd-10-379-2019

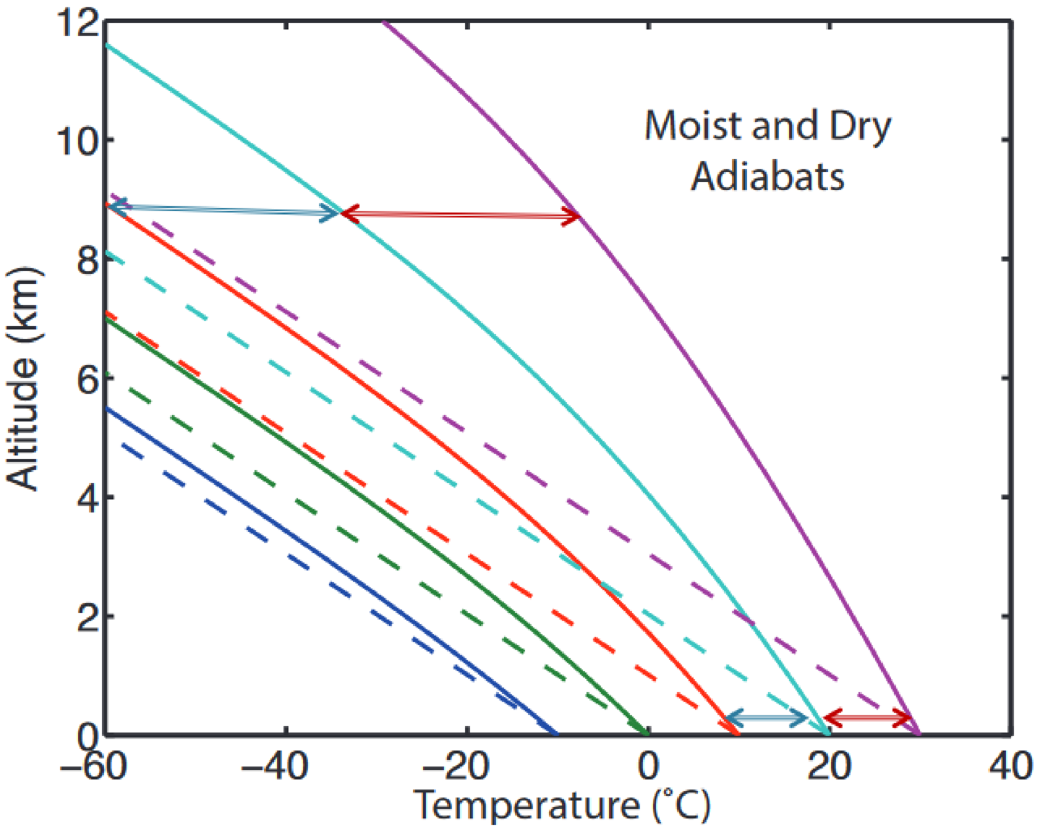
Lapse rate feedback



- Lapse rate: the moist adiabatic lapse rate is expected to decrease with warming tropospheric temperature. This means the outgoing LW radiation to space increases (loss of energy). → **First order effect.**
  - Above is particularly true **for tropical regions** where greenhouse gases lead to tropospheric warming (further aloft) due to increased convection from the surface.
- In the **mid- to high latitudes**, the surface tends to warm relatively more such that the lapse rate becomes steeper and results in a positive feedback (colder troposphere and less LW radiation loss to space).
- Overall, for the water vapour and lapse rate feedbacks the exact changes of temperature and humidity in high altitudes are not yet well known, but since the effects cancel each other out, their combined uncertainty is nevertheless relatively small.

- http://www.climate.be/textbook/chapter4_node7.html

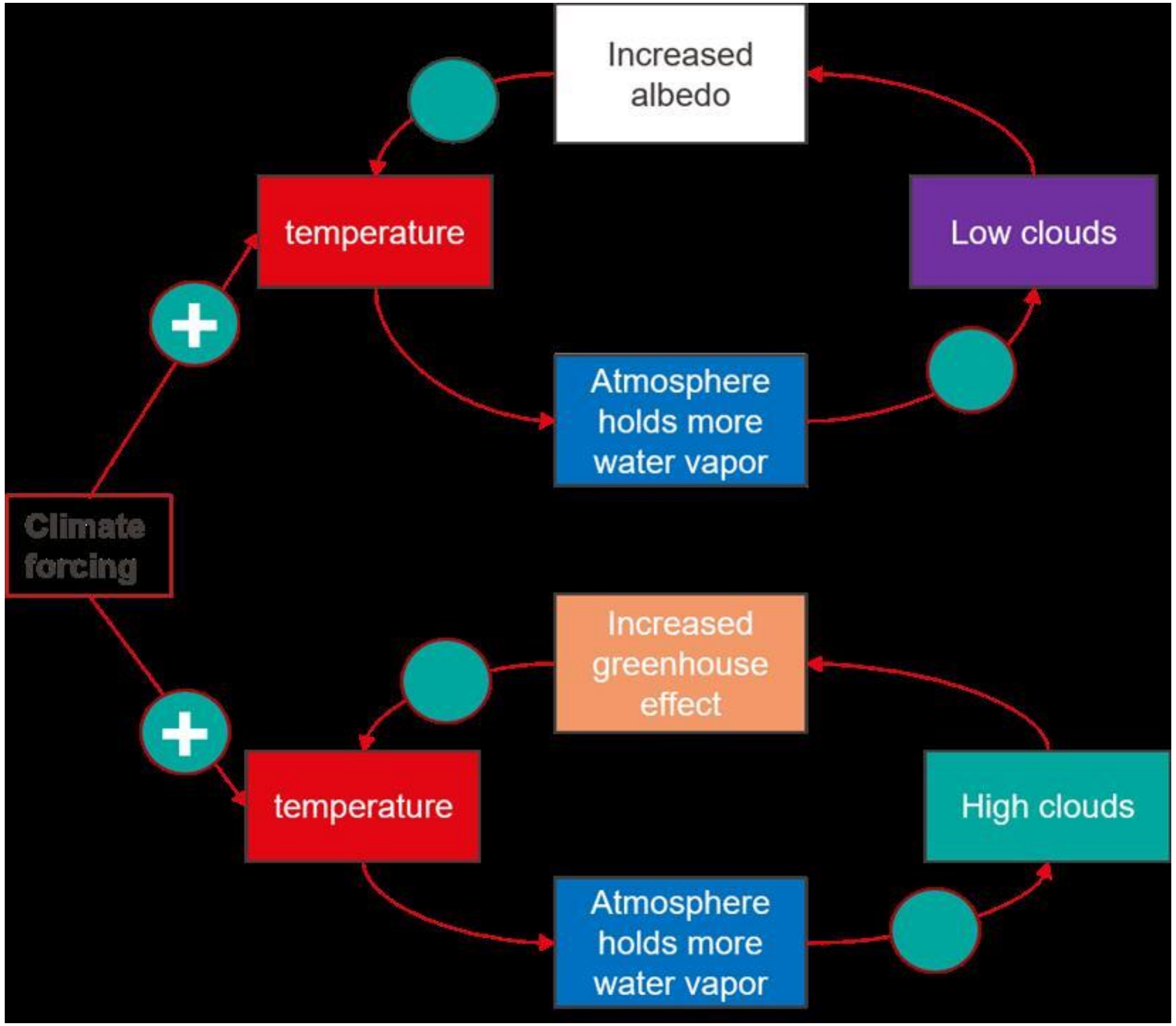
Moist adiabatic lapse rate in a warming climate

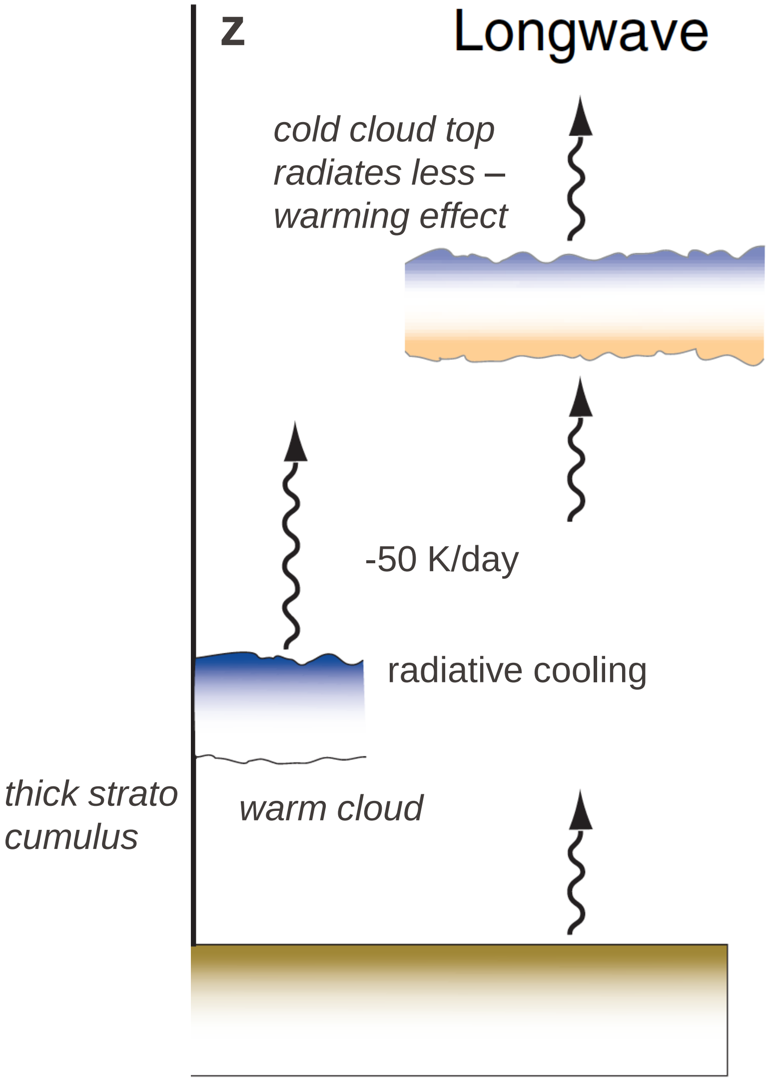


- The warmer the surface the smaller the moist adiabatic lapse rate.
- There is a warming amplification with altitude.

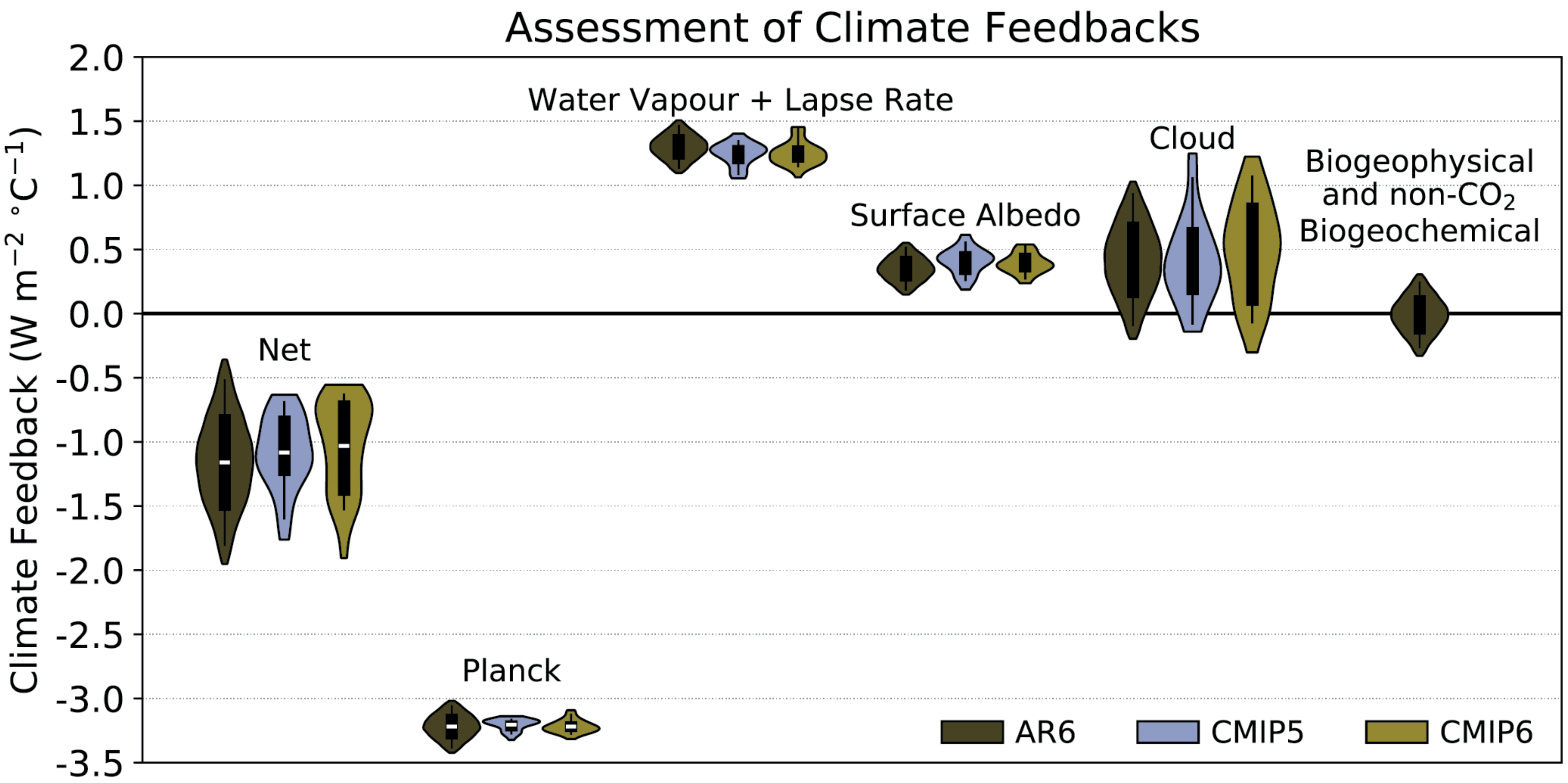
D. Hartmann UWashington

Cloud feedback: Where is the negative feedback?





Summary of physical feedback factors



- IPCC AR6, Fig. 7.10

Types of feedback – fast physical

Atmospheric thermodynamic feedbacks *(most certain quantification)*
1 Planck response (black body radiation)
2 The combined water vapour and lapse rate feedback

Cloud feedbacks *(complex and large source of uncertainty)*
1 Rise of cloud top feedback
2 Tropical low -cloud feedback
3 Mid-latitude cloud reflectance feedback
4 Cloud water phase feedback

Fast surface feedbacks
1 Snow albedo feedback
2 Soil moisture evapotranspiration feedback and CO_2 stomata-water feedback

Fast ocean feedbacks
1 ocean mixed-layer and ocean thermocline feedbacks
2 Tropical circulation responses to a warming climate

Sea ice feedbacks
1 Sea ice albedo feedback

Color legend

Thermal longwave (LW) heat redistribution including water vapour and moisture

Shortwave (SW) reflectivity / albedo

Both LW and SW effects

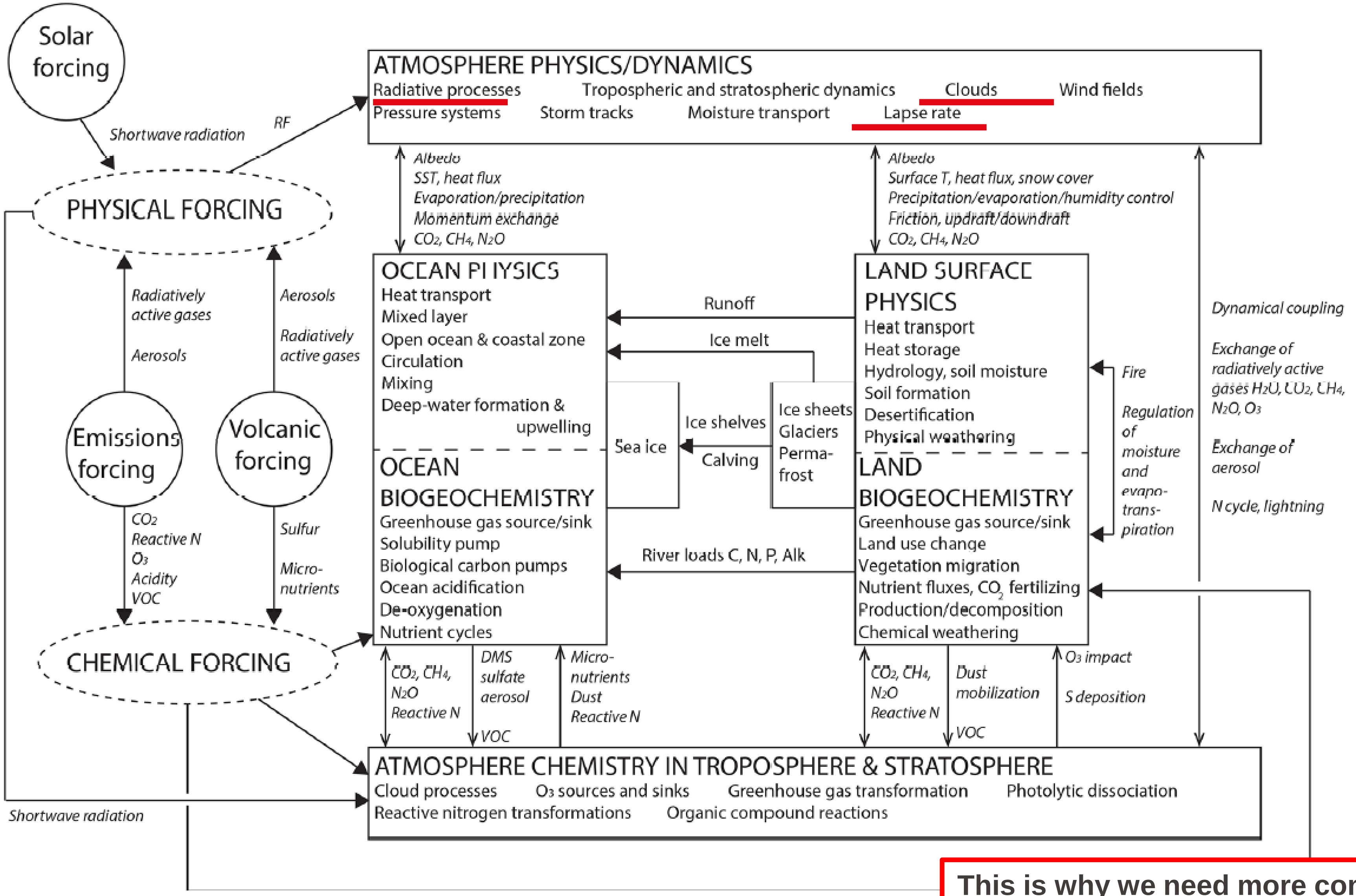
Heinze et al., ESD, 2019, https://doi.org/10.5194/esd-10-379-2019

EPFL

Upper ocean feedback

- Fundamental physical properties of the ocean are very different compared to the atmosphere: heat capacity, viscosity, and timescales of motion.
- The evolution of climate change depends critically on the penetration rate of the global warming signal into the ocean and the capacity of the ocean to uptake heat from the atmosphere.
- Ocean–climate feedback timescales range from the synoptic to seasonal, decadal, or even centennial.

There are many more feedbacks



Long time scale effects:
- Deep ocean
- Ice sheets

Biogeochemical effects
- Vegetation growth
- fires
- Marine emissions
- Inorganic ocean carbon cycle
- Aerosol effects

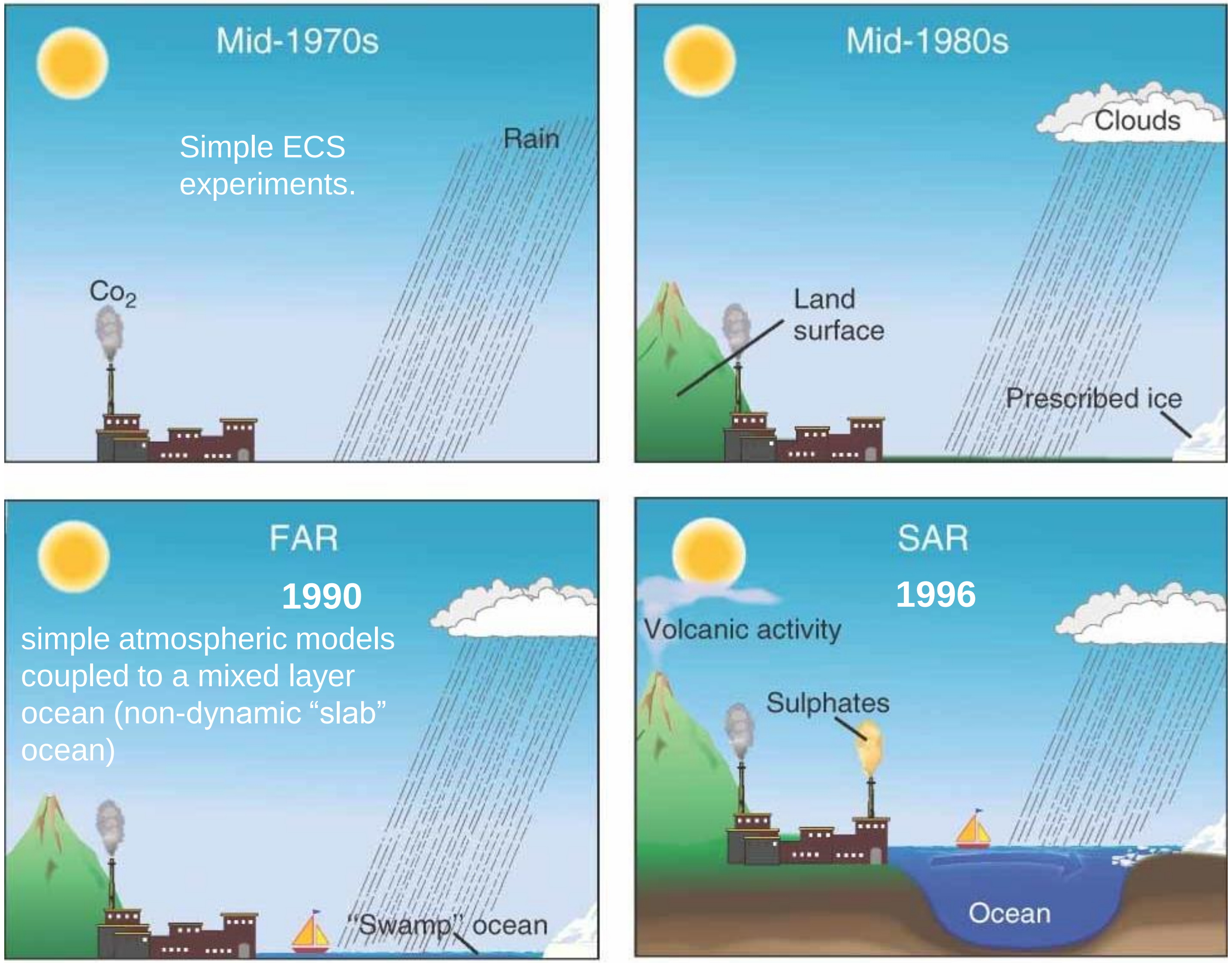
- Heinze et al., ESD, 2019, https://doi.org/10.5194/esd-10-379-2019

This is why we need more complex climate models including also biogeochemical processes.

Evolution of climate models

General circulation models (GCMs):

- «Dynamical core» simulates large-scale fluidic motion using primitive equations.
- «Model physics» simulate climate-relevant physical processes (e.g. radiative transfer).



First estimate of transcient climate response: increase CO_2 by 1% each year until doubled (ca. 70 years): TCR

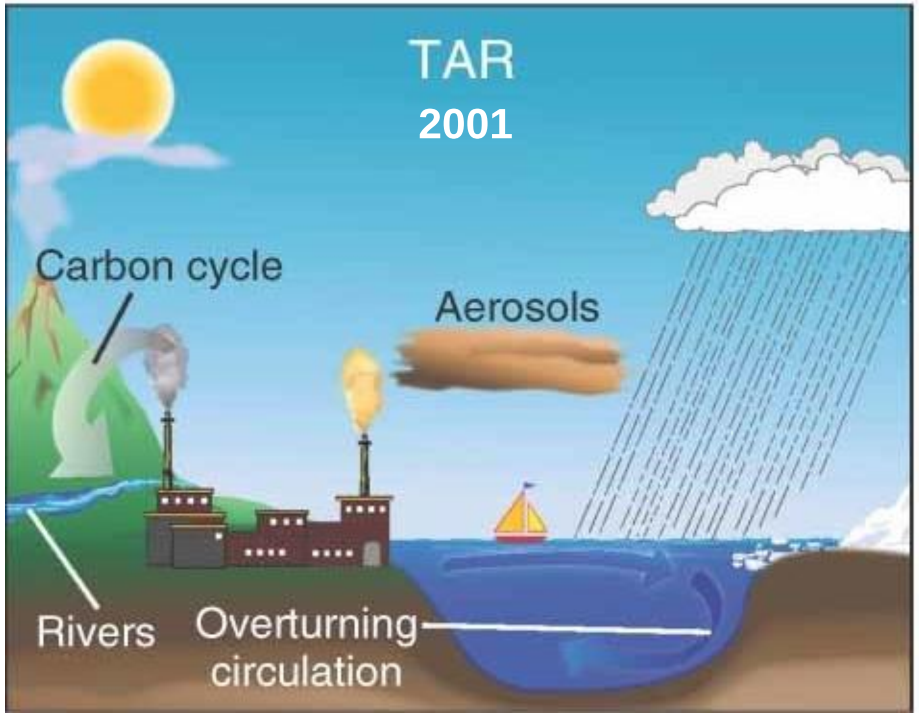
Established CMIP 2 for intercomparison of transcient responses. CMIP: coupled model intercomparison project

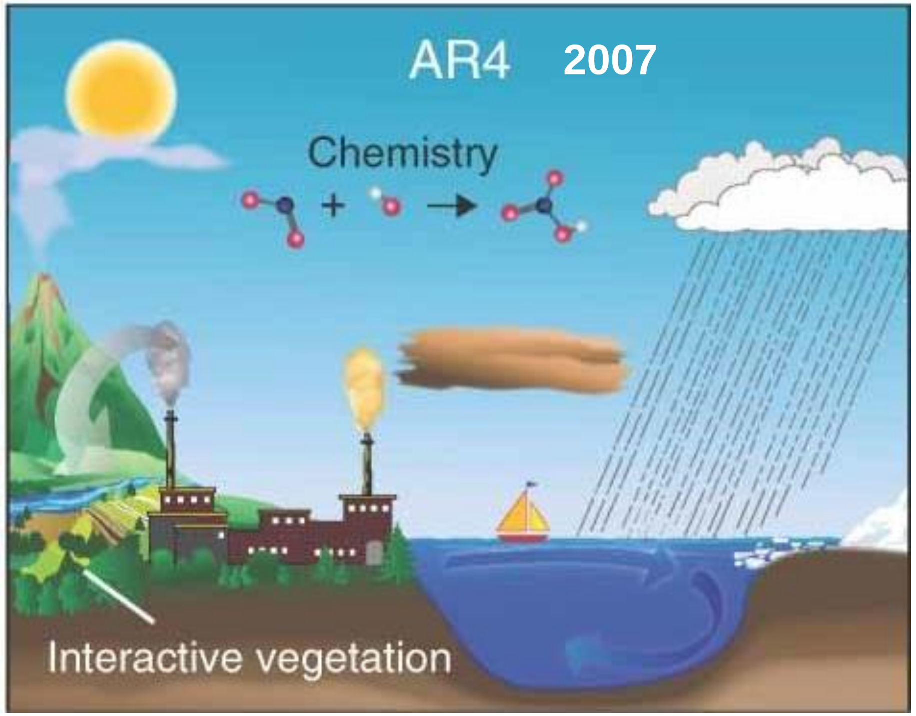
FAR: First assessment report
SAR: Second AR

Edwards, P.N. (2011), History of climate modeling. WIREs Clim Change, 2: 128-139. https://doi.org/10.1002/wcc.95

Evolution of climate models

Creation of time-dependent emission scenario runs.
Establish TCR experiments.





Atmosphere-ocean general circulation model

How do we now derive climate sensitivity?

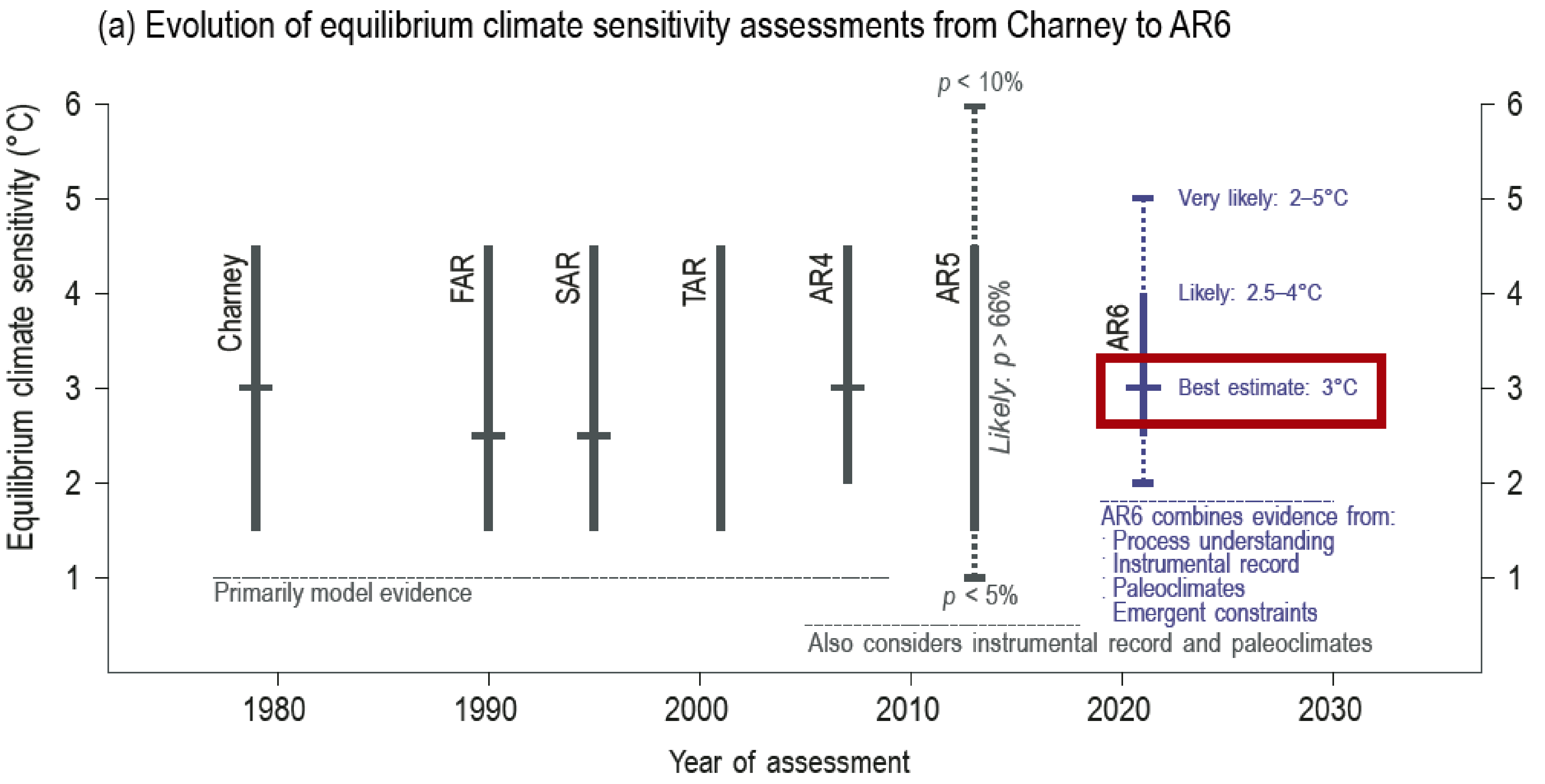
$$N = RF + \lambda\Delta T$$

N: net top of atmosphere energy balance,
RF: radiative forcing,
ΔT: global surface temperature response,
λ : feedback factor

- For a given forcing associated with a doubling of atmospheric CO_2 concentration (with a radiative forcing of about 3.7 W m^{-2}), at equilibrium N = 0, we can solve for ΔT, a quantity known as the “equilibrium climate sensitivity” (ECS).
- **Gregory method**:
  - The ECS calculated by the **Gregory method** is derived from a fully coupled Earth system model and does not require equilibrium to actually be achieved.
  - In the Gregory method, **CO_2 is instantaneously quadrupled** in a fully coupled Earth system model and **run for 150 years**.
  - As the surface temperature asymptotes toward equilibrium, the slope of the time-evolving curve of the net top-of-atmosphere radiance against the surface temperature is calculated to extrapolate the eventual temperature increase at equilibrium some time far in the future for a doubling of CO_2, assuming that there is a roughly linear response that is half of the warming from a quadrupling of CO_2.

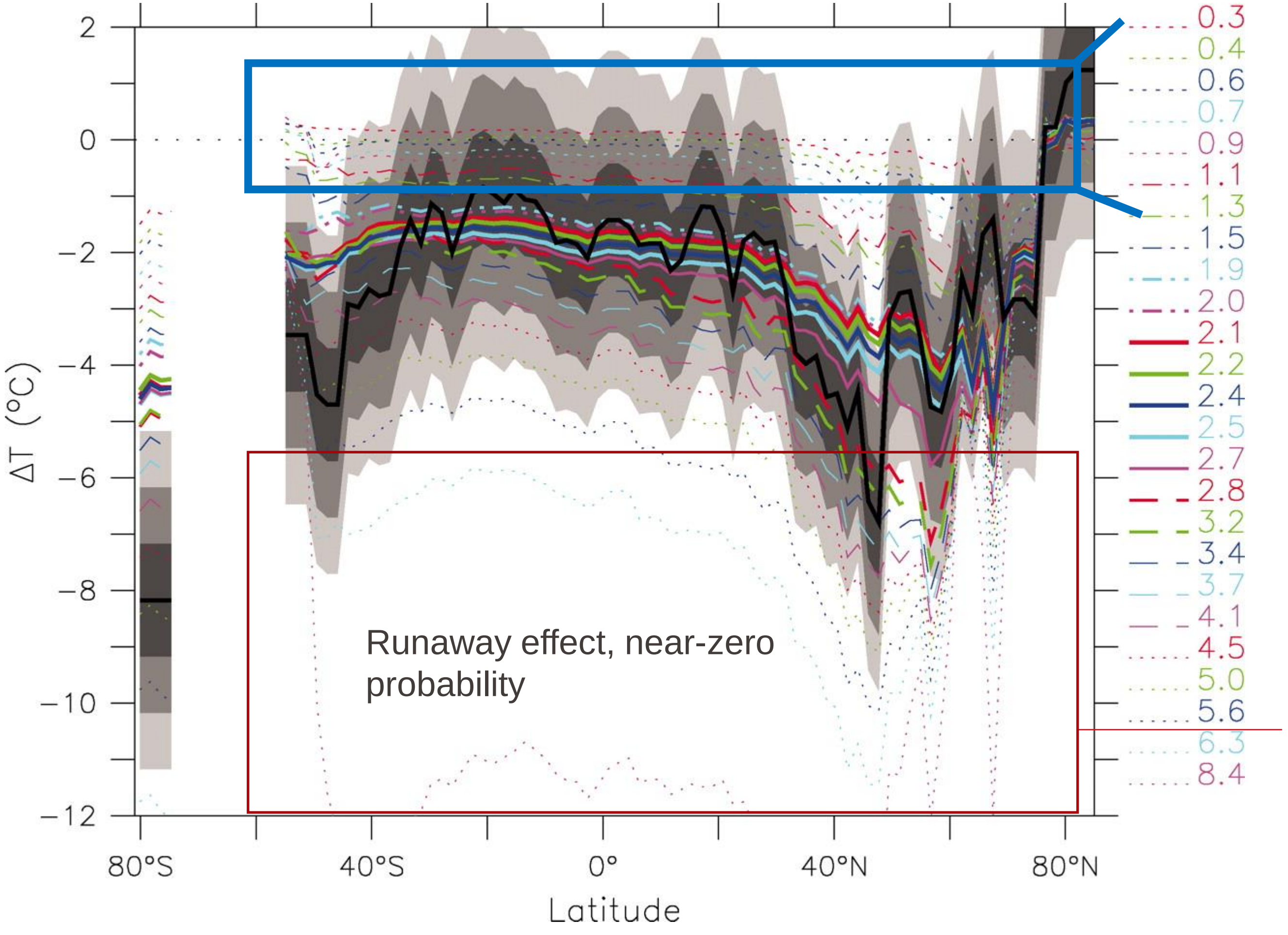
https://advances.sciencemag.org/content/6/26/eaba1981/tab-pdf

Evolution of climate sensitivity



- IPCC, AR6, Fig. TS16

Constraining with paleoclimate records



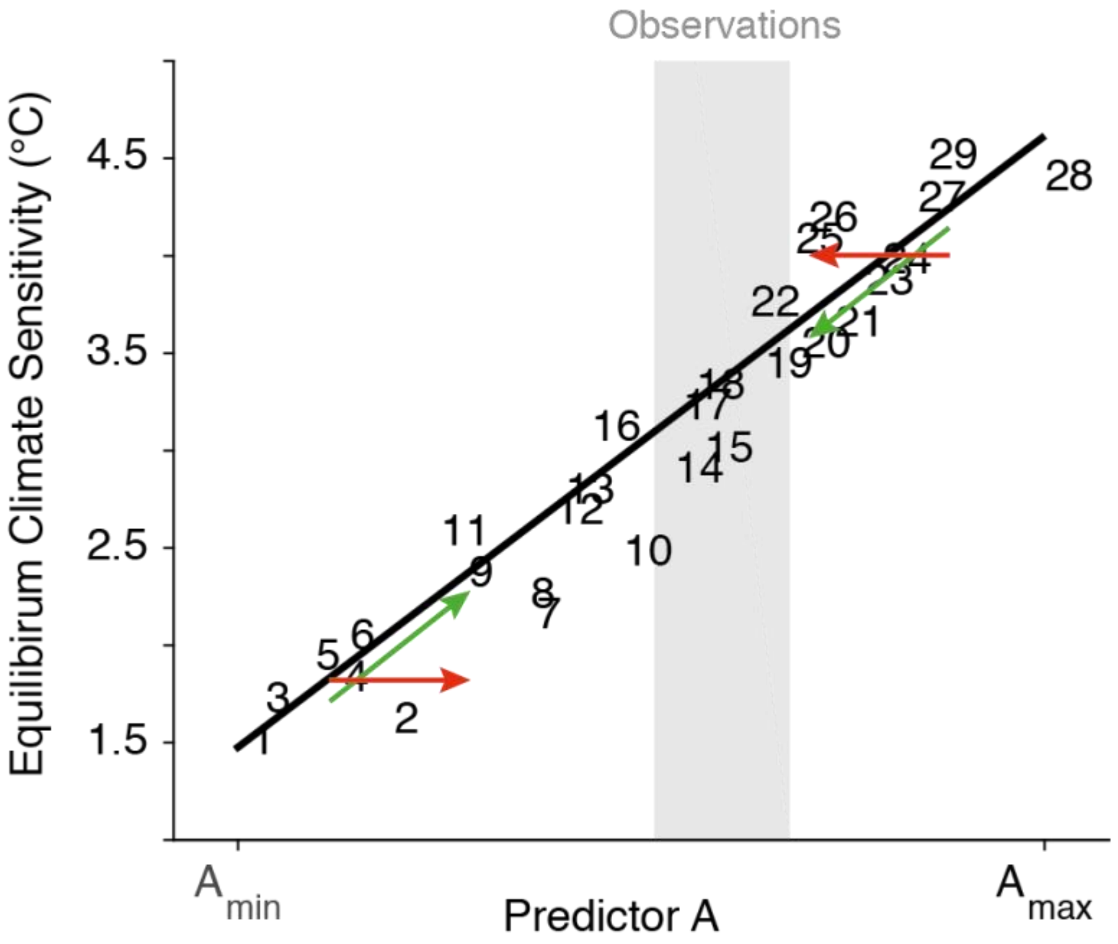
Zonally averaged surface temperature change between the Last Glacial Maximum and modern. Thick black line denotes the climate reconstructions, the gray shading the ±1, 2, and 3 K intervals around the observations. Modeled temperatures, averaged using only cells with reconstructions are shown as colored lines labeled with the corresponding ECS_{2xC} values.

- Models with ECS_{2xC} < 1.3 K underestimate the cooling at the LGM almost everywhere, particularly at mid-latitudes and over Antarctica,
- Models with ECS_{2xC} > 4.5 K overestimate the cooling almost everywhere, particularly at low latitudes.
- High-sensitivity models (ECS_{2xC} > 6.3 K) show a runaway effect resulting in a completely ice-covered planet. Once snow and ice cover reach a critical latitude, the positive ice-albedo feedback is larger than the negative feedback because of reduced longwave radiation (Planck feedback), triggering an irreversible transition.
- During the LGM, Earth was covered by more ice and snow than it is today, but continental ice sheets did not extend equatorward of ~40°N/S, and the tropics and subtropics were ice free except at high altitudes.

- **Results thus suggest that large climate sensitivities (ECS_{2xC} > 6 K) cannot be reconciled with paleoclimatic and geologic evidence and hence should be assigned near-zero probability.**

Schmittner et al., 2011: https://science.sciencemag.org/content/334/6061/1385

Constraining with emergent constraints (the concept)



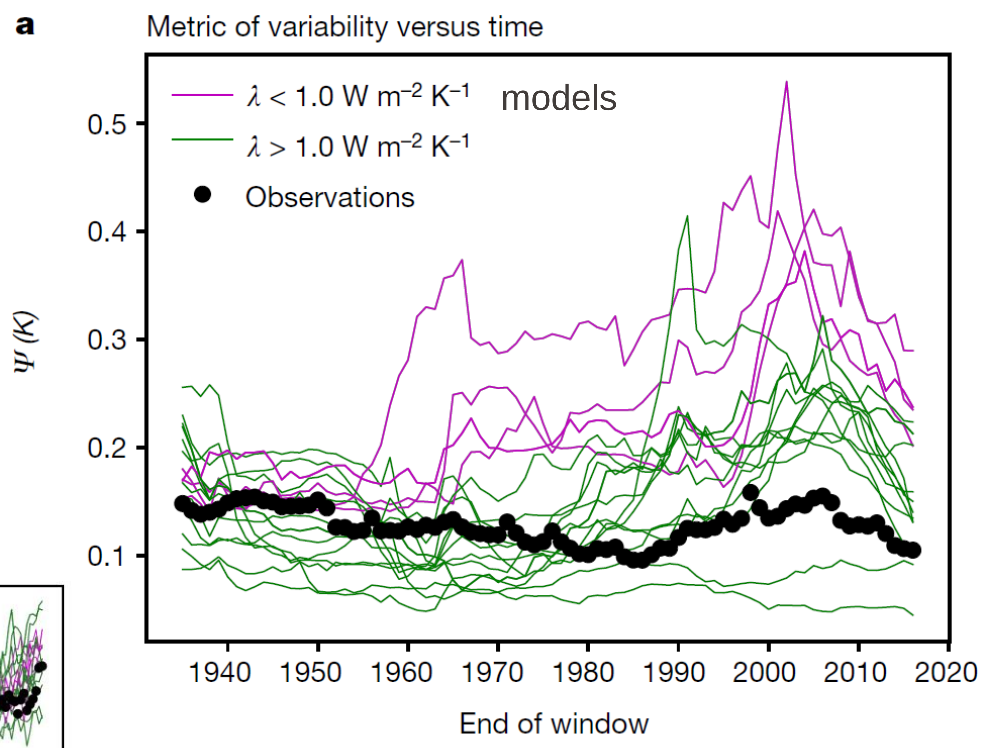
Hypothetical relationship between a predictor A and the equilibrium climate sensitivity (ECS) for 29 climate models.

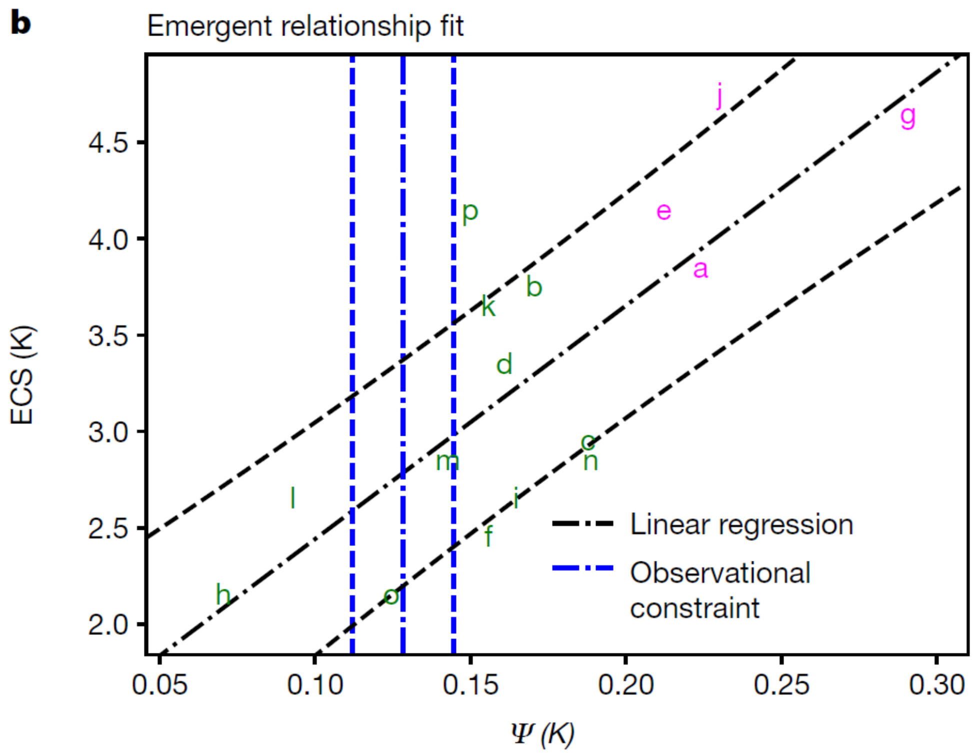
Predictor A may represent, for example, the variability of the surface temperature over time. On the y-axis, ECS may be replaced by any climate-change projection. The black line is the linear regression, and the grey vertical bar is the observed value of predictor A (with its uncertainty).
Arrows show the evolution of ECS after improving the representation of predictor A for two climate models having low (4) and high (24) ECS values.
If these climate models evolve following the red arrows, the relationship may have been found by chance. If they evolve following the green arrows, mechanisms underlying the relationship gain credibility.
Since predictor A can be observed, this relationship can be considered as an emergent constraint.

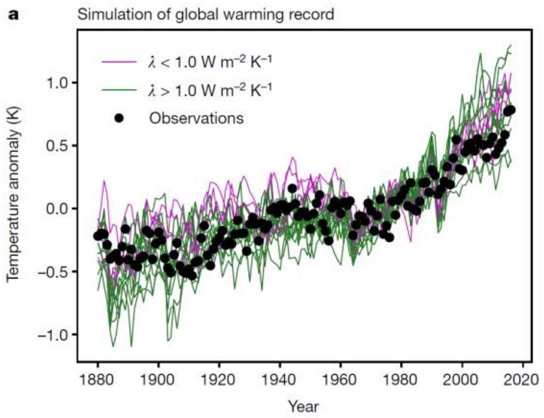
- https://climate-dynamics.org/reducing-uncertainties-in-climate-projections-with-emergent-constraints-part-1-concept/#more-1285

Constraining with emergent constraints

Temporal variability (Ψ) of the surface air temperature as observable metric to constrain ECS.







55 years as time windows to derive Ψ

Done for many time windows (not shown)

Reduces the probability of ECS being less than 1.5 degrees Celsius to less than 3 per cent, and the probability of ECS exceeding 4.5 degrees Celsius to less than 1 per cent.

- Cox et al., 2018, Nature, doi:10.1038/nature25450

Why larger range of ECS with the newest generation of climate models?

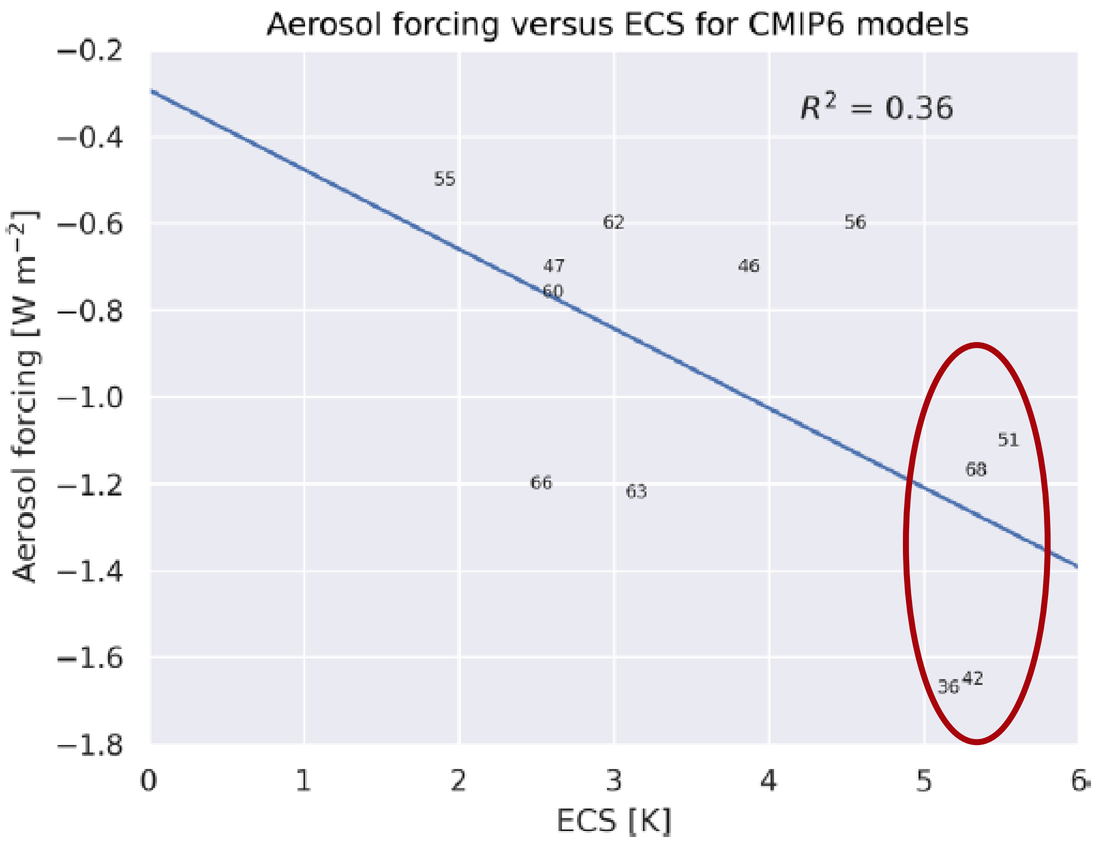


Fig. 4. Effective radiative forcing from aerosols versus ECS. Values supplied by the modeling groups (Table 3); black line is linear fit with R^2 of 0.36. The numbers denoting individual models are listed in Table 2.

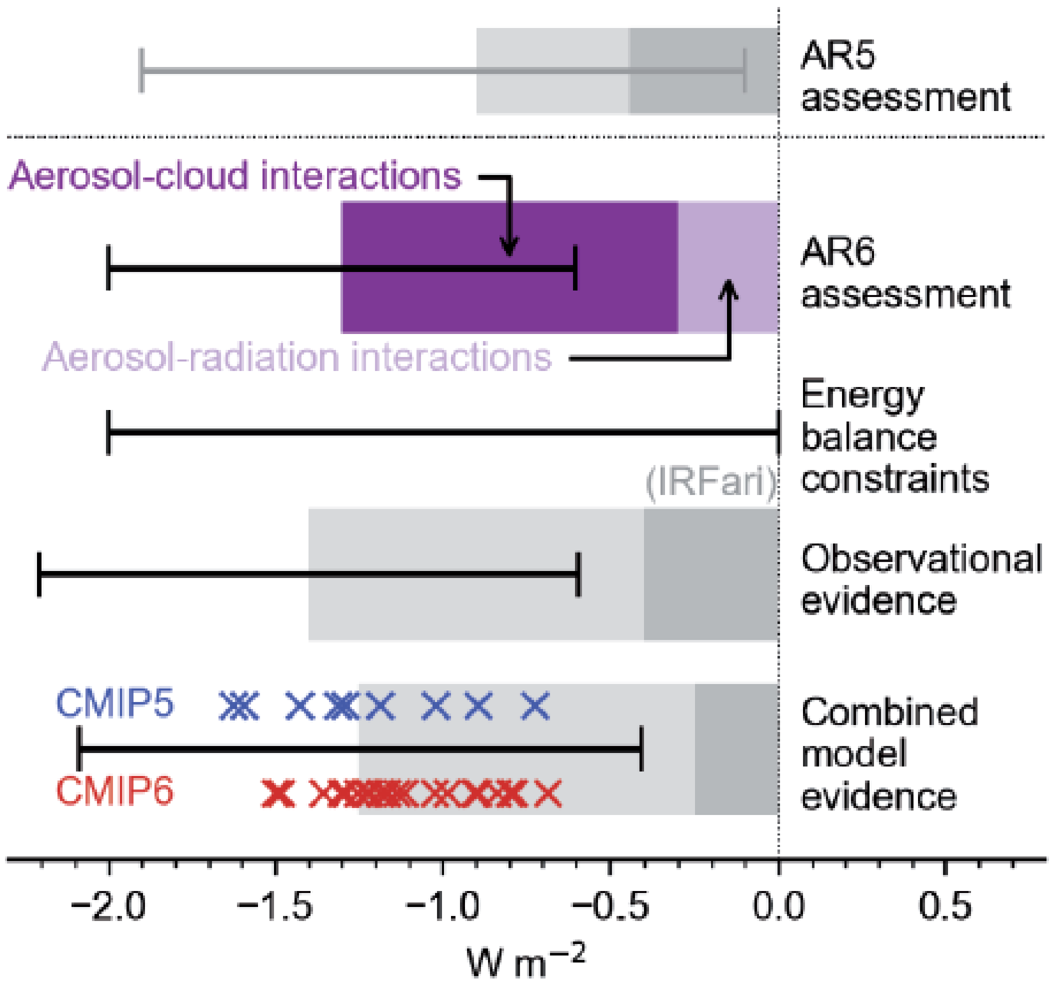
- Newer climate models include more complex treatment of aerosols.
- Aerosols interact with clouds which strongly influence the radiative forcing effect of aerosols.
- If the aerosol forcing is more negative, the climate sensitivity to CO_2 forcing needs to be higher to end up with the same rate of warming.
- However, CMIP6 models only show a week correlation ($R^2 = 0.36$).

Models with prognostic aerosol schemes and aerosol-cloud interactions.
It is difficult to pinpoint the exact feedback mechanisms in the models that lead to high ECS. But overall *”cloud feedbacks and cloud-aerosol interactions in models with prognostic aerosol schemes seem to be playing an important role”.*

Critical to research

- Meehl et al., Science Advances, 2021, https://advances.sciencemag.org/content/6/26/eaba1981/tab-pdf

Uncertainty from aerosols and clouds



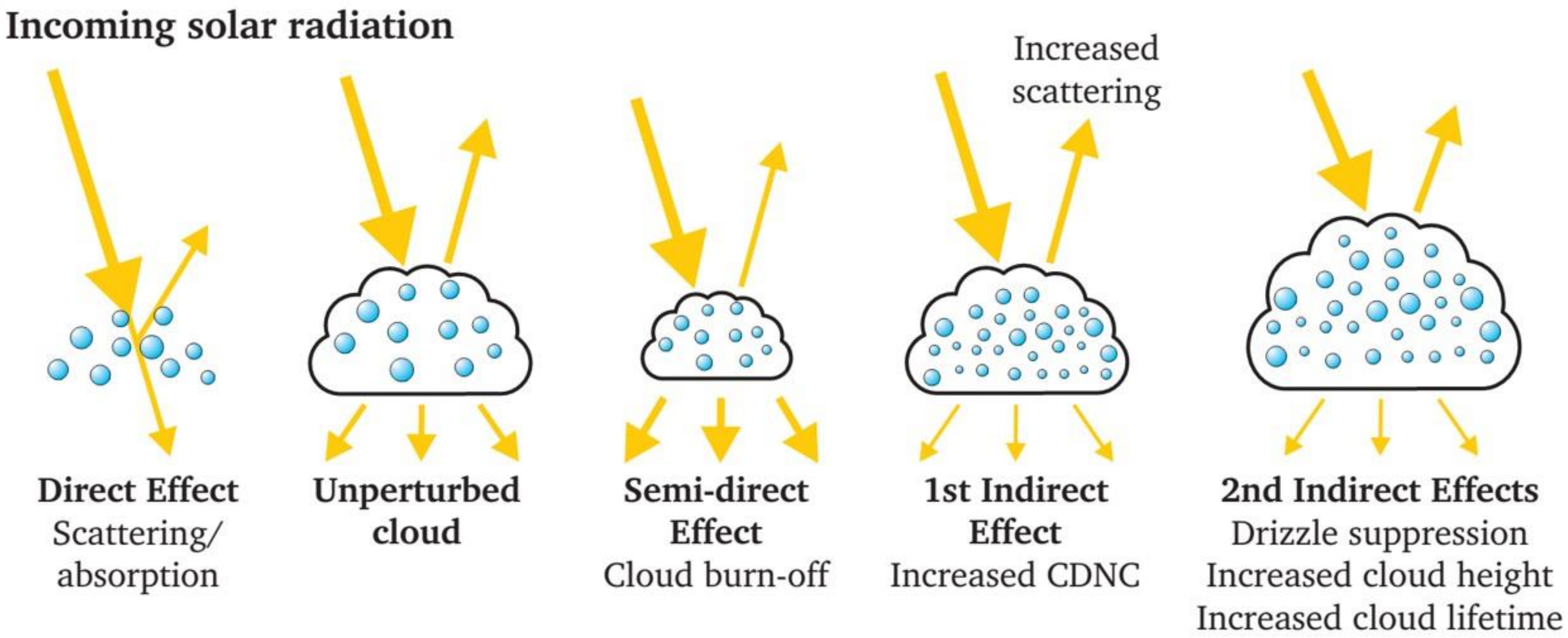
B. How can we reduce uncertainties from aerosol and cloud forcing?

➡ **defining the preindustrial**

Cloud feedbacks and cloud-aerosol interactions are the most likely contributors to the high values and increased range of ECS in CMIP6.

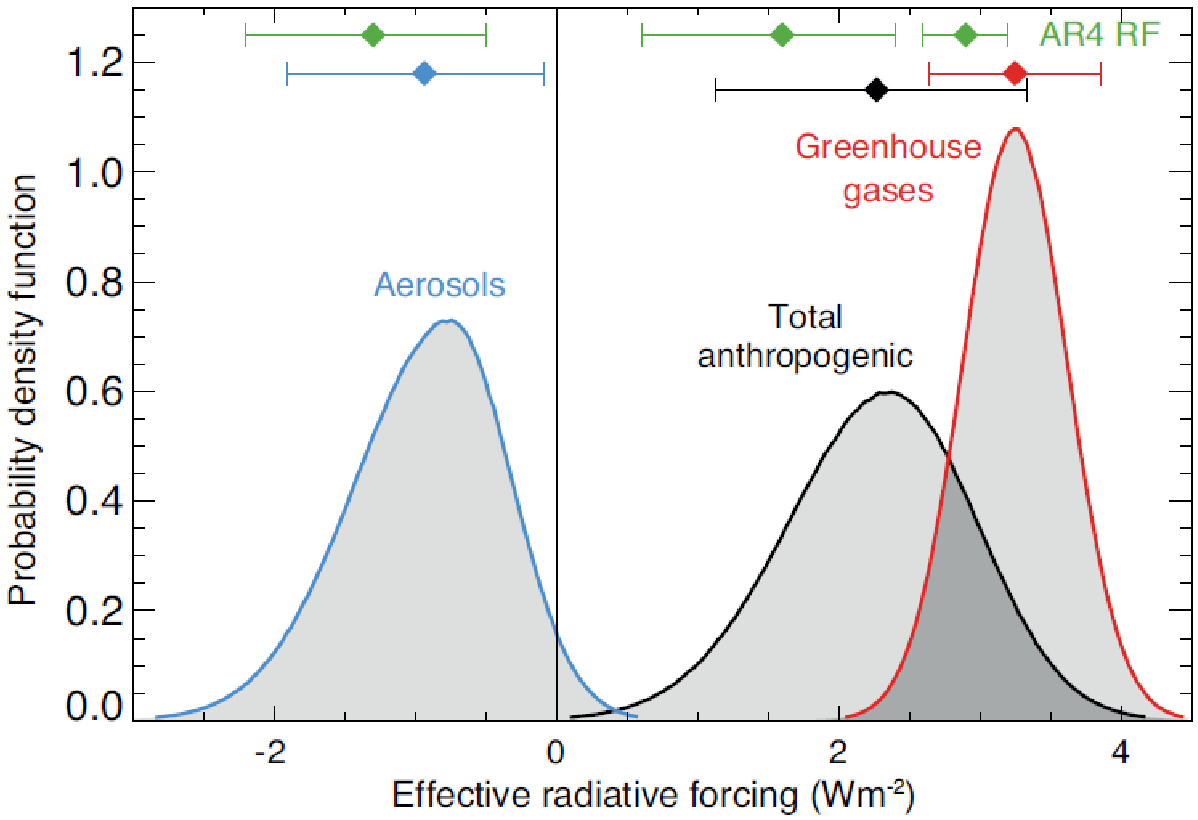
Meehl et al., Science Advances, 2021

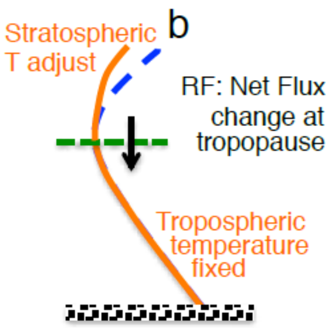
Aerosol-cloud interactions (ACI)

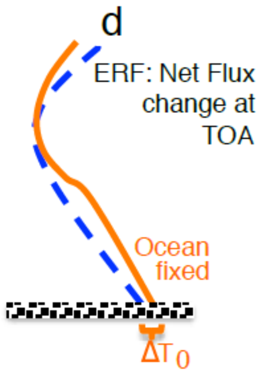


ERF from ARI and ACI

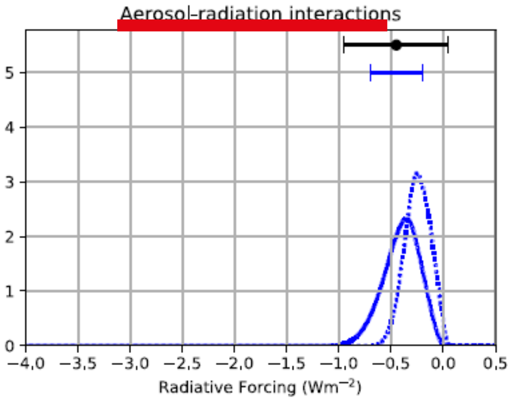
IPCC, 2013, chapter 8, Fig. 8.16



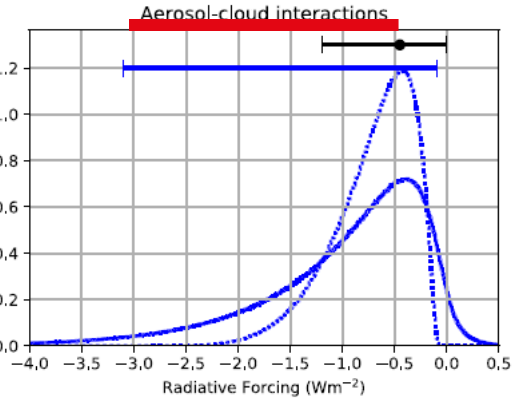


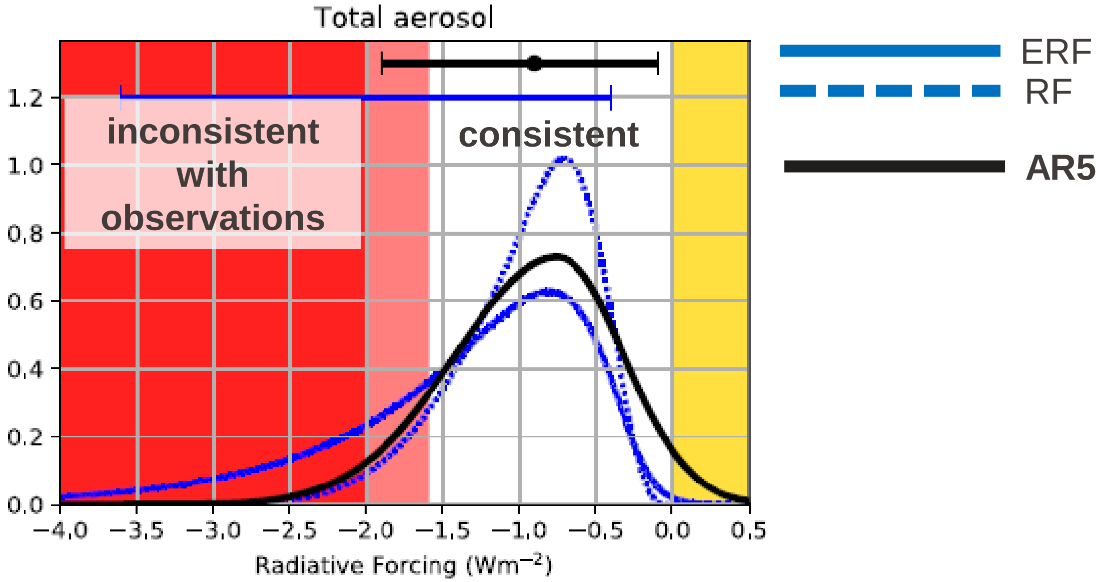


AR5 vs now:
better constrained
(less strong rapid adjustments)



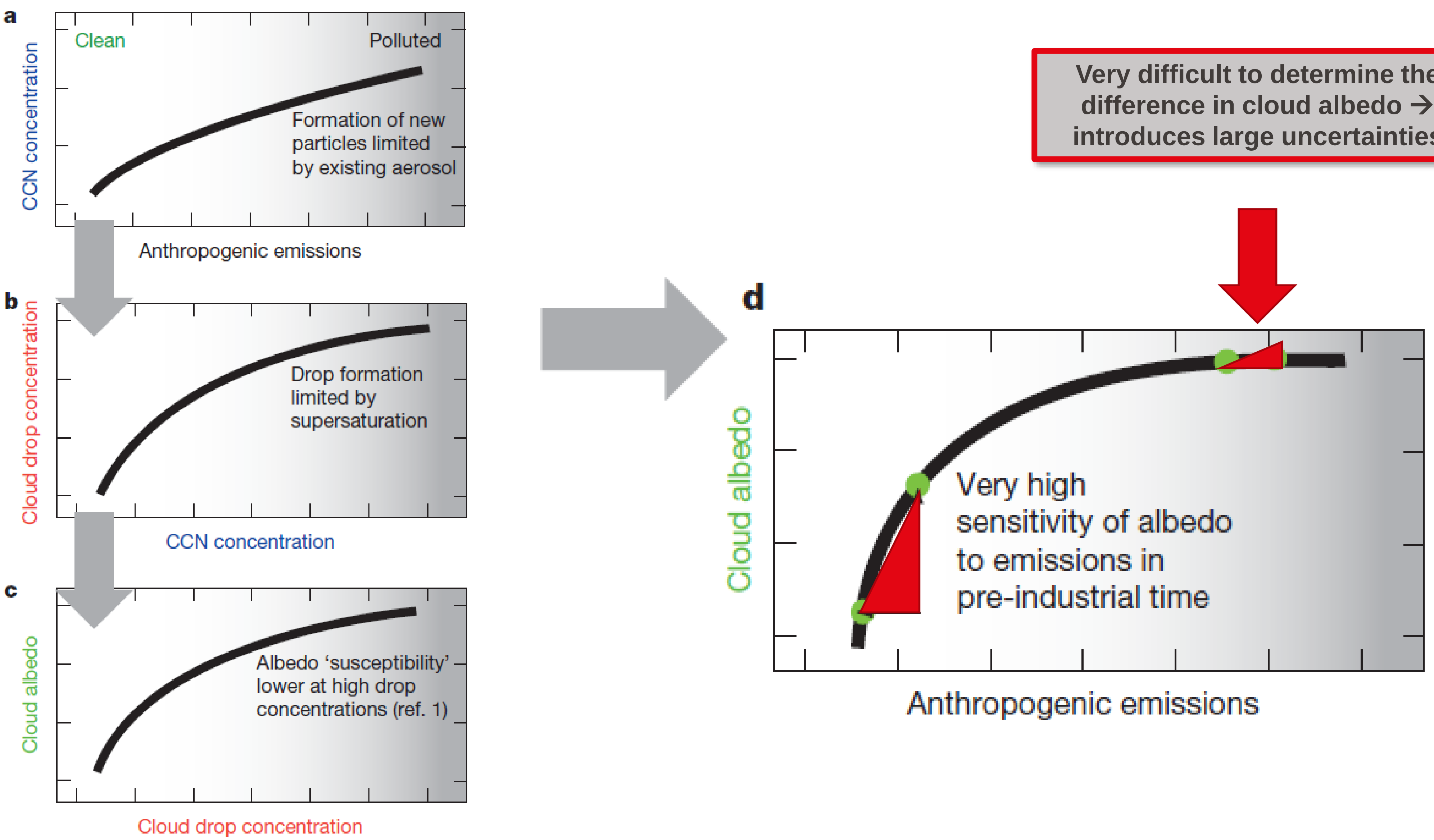
AR5 vs now:
less constrained
(wider assessment of rapid adjustments)





Bellouin et al., Rev. of Geophys., 2020

Determining the aerosol-cloud radiative effects



Very difficult to determine the difference in cloud albedo → introduces large uncertainties

CCN Cloud Condensation Nuclei

Carslaw et al., 2013, doi:10.1038/nature12674

EPFL

The challenge of climate science: small relative quantities!

- We deal with small magnitudes of the changes in radiative fluxes and global temperature relative to the magnitudes of the initial, unperturbed quantities to determine current and future climate change.
- The observed change in global means surface temperature of about 1.07 K represents a change of about 0.3% relative to the initial 287 K. Even the 2 K increase represents a change of less than 1%.
- The challenge to the climate change research community is to gain quantitative understanding of the changes in quantities influencing climate change and the expected response of the system to the accuracy necessary for informed decision making regarding prospective controls on future emissions of climate influencing substances.
- Such quantitative understanding is essential to answering “what if” questions regarding the consequences of future emissions of climate influencing substances.