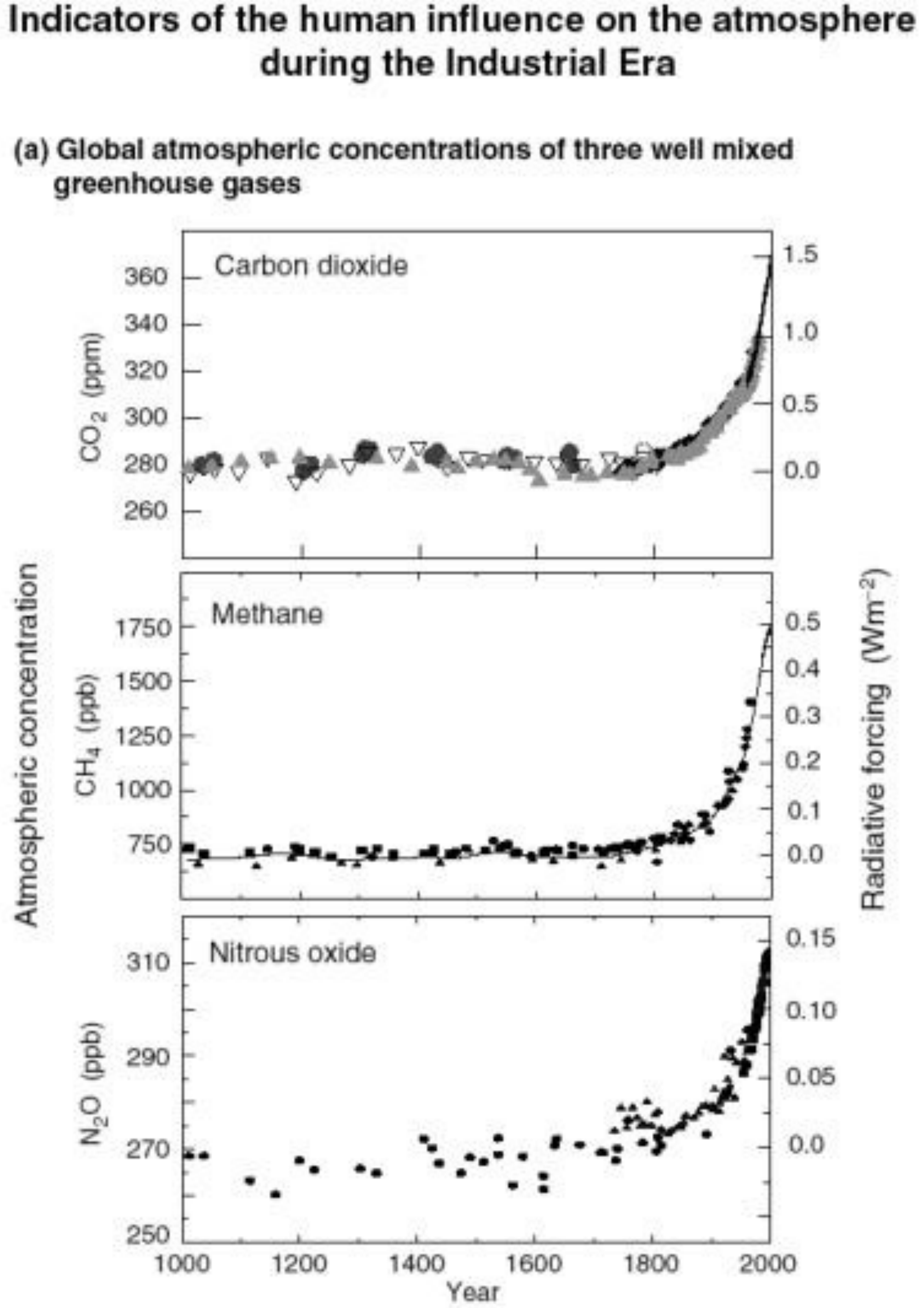
Climate Sensitivity

- Concept
- Estimates
- Applications

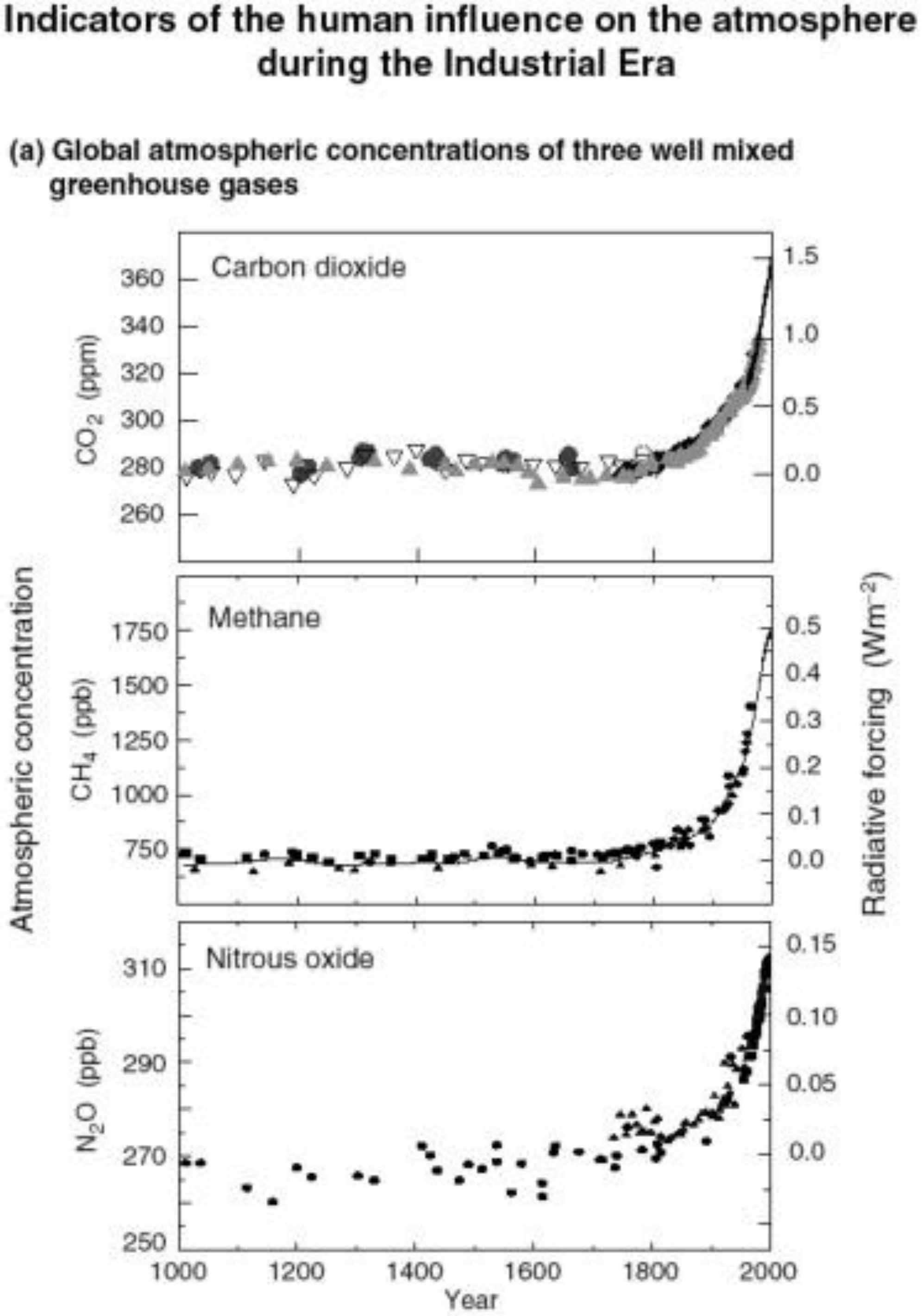
Greenhouse gas trends are large and can be associated directly with human actions.



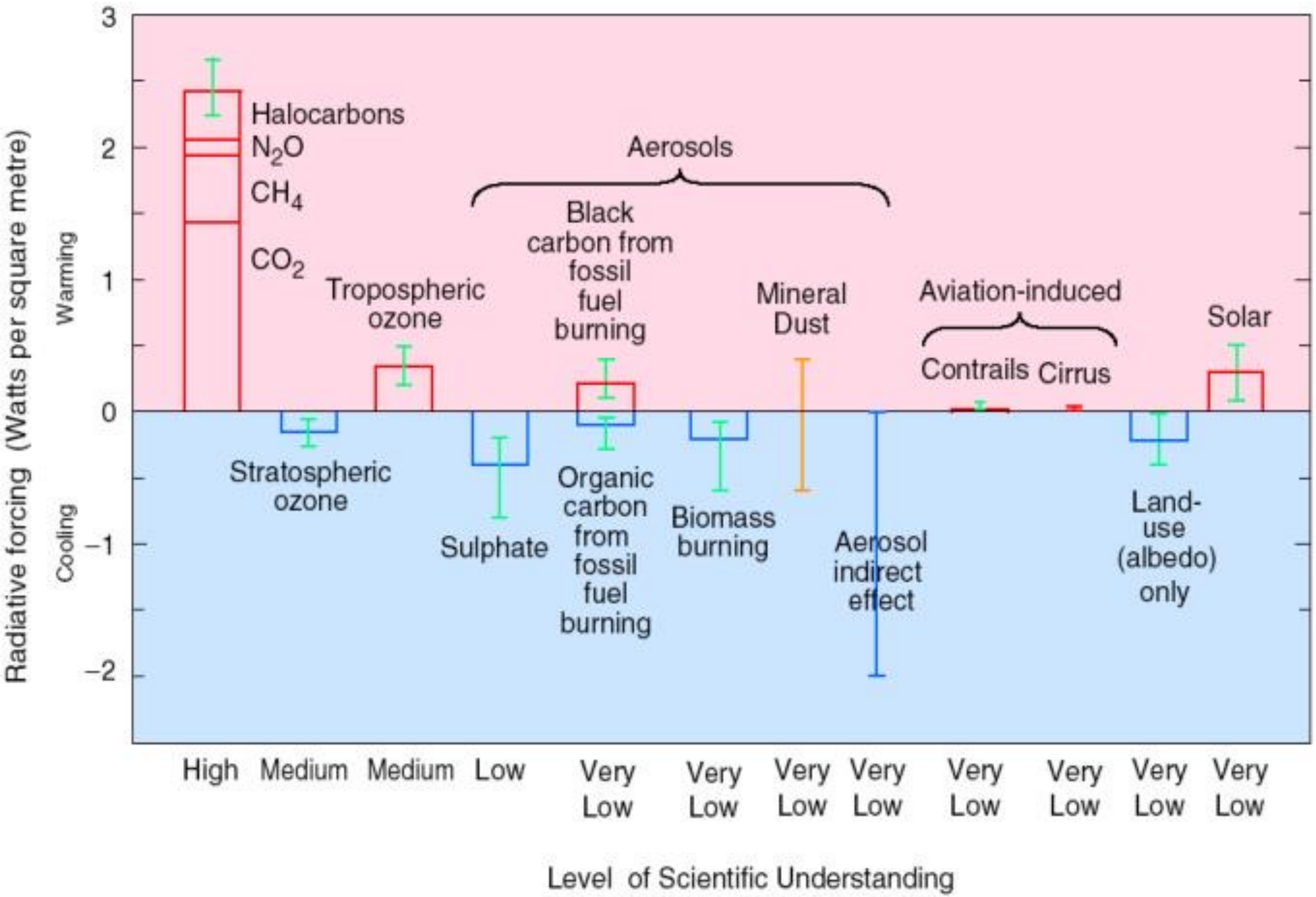
Greenhouse gas trends are large and can be associated directly with human actions.

Carbon dioxide trends Can be uniquely associated with fossil fuel burning through isotopes of carbon

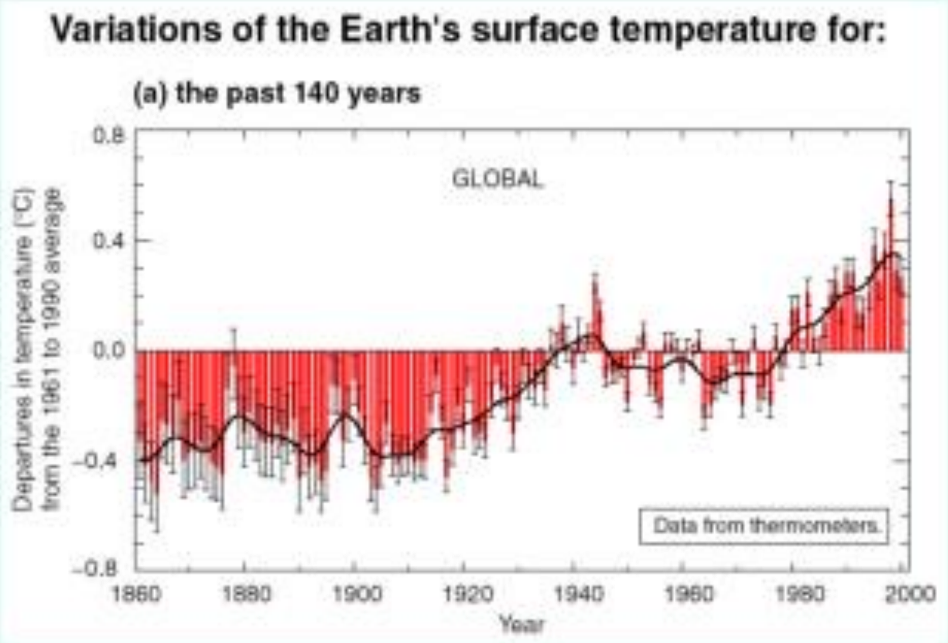
^{14}C



The global mean radiative forcing of the climate system for the year 2000, relative to 1750



To Project future climates by using the observed record of climate over the past century, we need to know three things to interpret the temperature time series:



$$\Delta Q = C\frac{\Delta T}{\Delta t} + \frac{1}{\lambda}\Delta T$$

Climate Forcing = ΔQ (Wm^{-2})

Heat capacity = C ($J\,^{o}K^{-1}\,m^{-2}$)

Climate sensitivity = λ (^{o}K per Wm^{-2})

Energy Equation:

$$\Delta Q = C\frac{\Delta T}{\Delta t} + \frac{1}{\lambda}\Delta T$$

Climate	**=**	**Heat**	**+**	**Heat**
Forcing		**Storage**		**Loss**

In Equilibrium, temperature is constant with time and so,

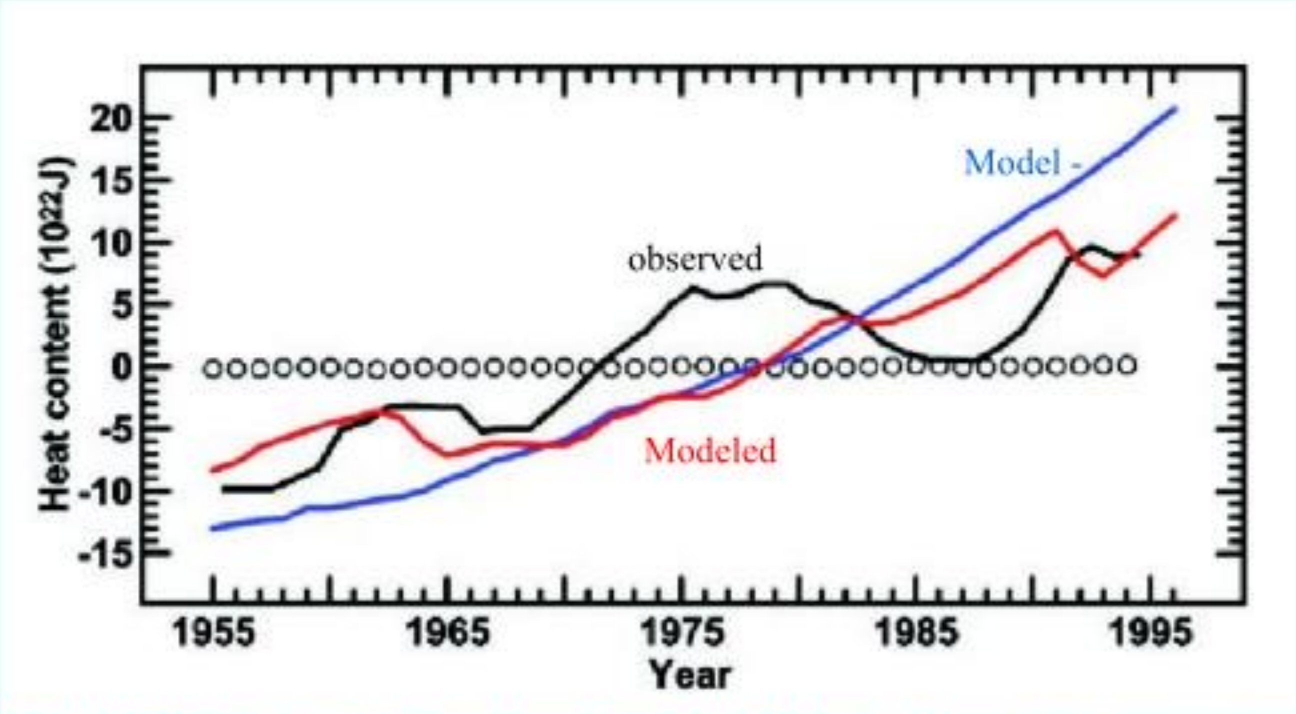
$$\Delta T = \lambda \cdot \Delta Q$$

λ is a measure of climate sensitivity;
K per Wm^{-2} of climate forcing

Heat Storage: Mostly the Oceans

1955-1996; Levitus et al. 2001: Science

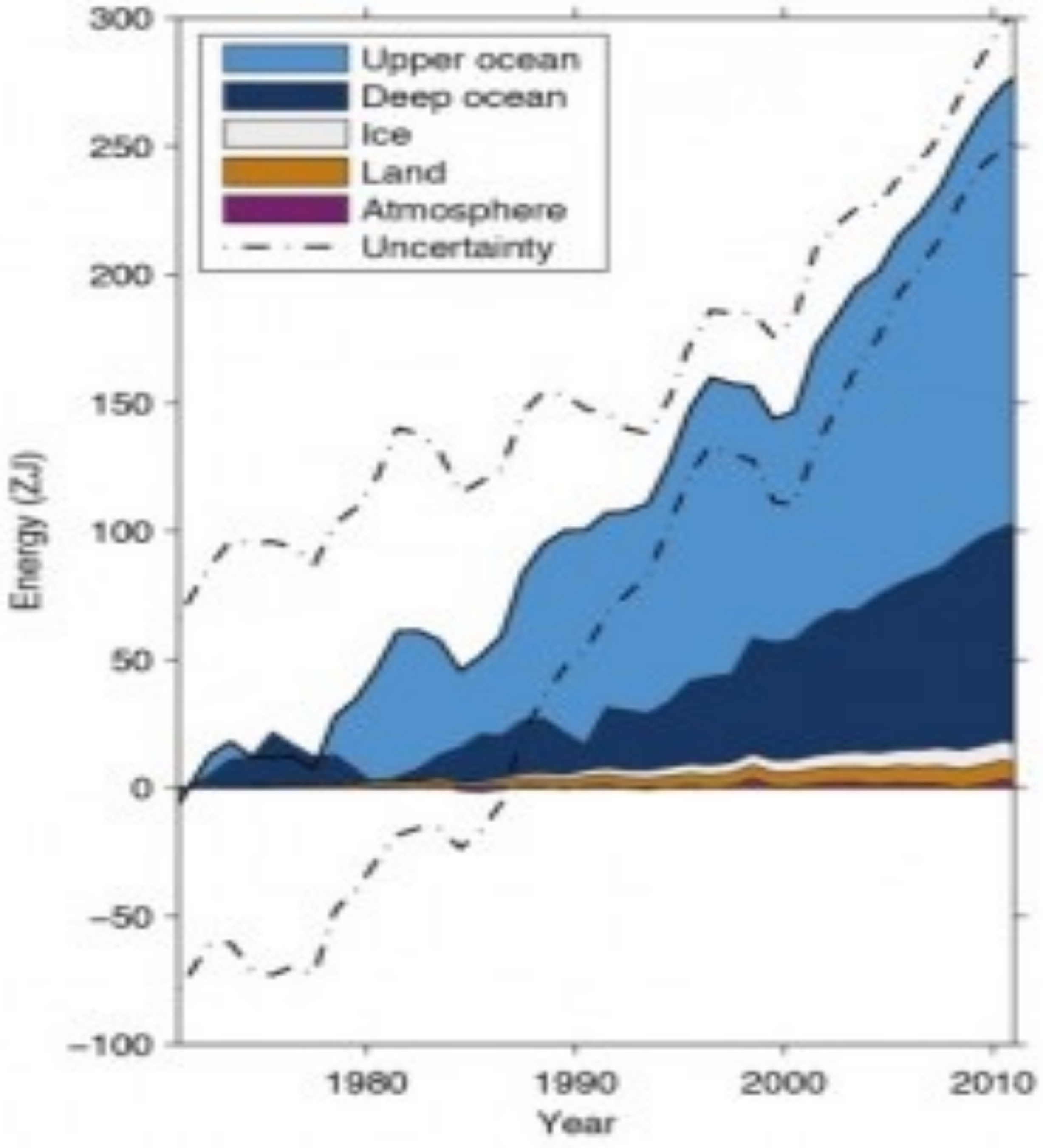
World Ocean = 18.2 x10^{22} Joules
Atmosphere = 0.7 x10^{22} Joules
Land Ice = 0.8 x10^{22} Joules



Model includes forcing from Greenhouse Gases, Sulfate Aerosols Solar irradiance changes, and volcanic aerosols.

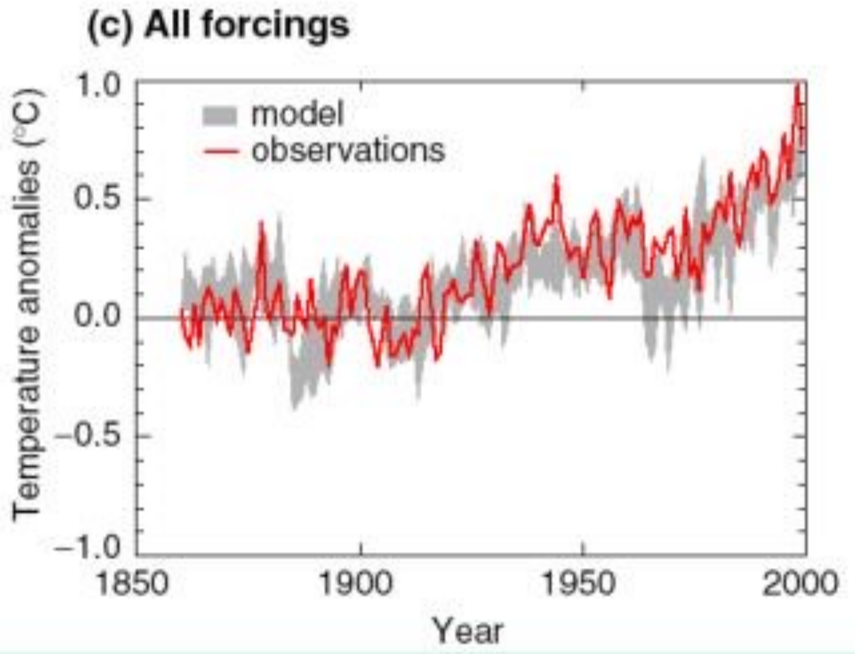
Model minus solar irradiance changes and volcanic aerosols.

Heat uptake:
The ocean matters



Top-Down Approach:

Determine sensitivity of climate from observed record over past 130 years. Use simple model to extrapolate into future.



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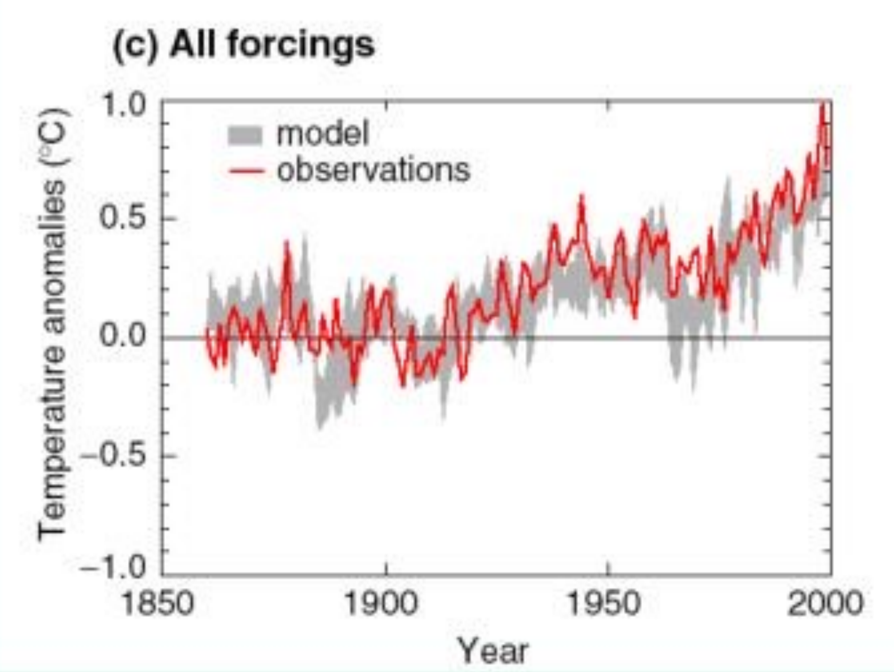


Problems: Need to know:

- Climate forcing - uncertain, especially solar and aerosol forcing.
- Heat storage - somewhat uncertain.
- Climate sensitivity - also uncertain.

No two of these are known with enough precision to usefully constrain uncertainty in the third, with the data available, although it is possible to fit the observations with fair precision using even a simple model.

Global Energy Balance

Incoming Solar Radiation = Outgoing Terrestrial Radiation

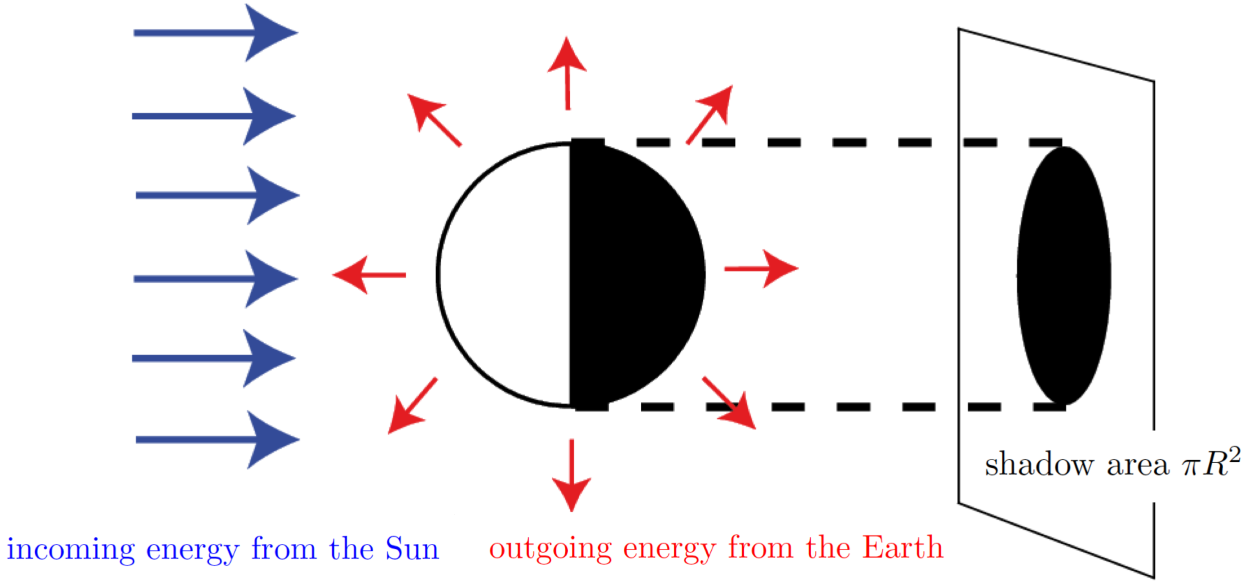
$$\underbrace{S_0}_{\text{solar irradiance}} \cdot \underbrace{\pi r^2}_{\text{cross-sectional area of Earth}} \cdot \underbrace{(1-\alpha)}_{\text{absorbed fraction}} = \underbrace{\sigma T^4}_{\text{emitted infrared flux}} \cdot \underbrace{4\pi r^2}_{\text{surface area of Earth}}$$

$$T = \left(\frac{S_0}{4\sigma} (1-\alpha) \right)^{\frac{1}{4}} = 255\,\mathrm{K}$$

Energy balance model: Concepts of climate

$$(1-\alpha)S\pi R^2 = 4\pi R^2 \epsilon\sigma T^4$$

$$T = \sqrt[4]{\frac{(1-\alpha)S}{4\epsilon\sigma}}$$



Heat capacity of the climate system

Fast rotation

Lohmann, 2020
doi:10.5194/esd-11-1195-2020

A Simple Question

- If we alter Earth's radiation balance by 1 W m^{-2} and allow the climate system to fully adjust, how much will the global average temperature change?
- This is a fundamental question in climate dynamics, and is relevant to both past and future climate change.

Effect of long wave radiation. S_{net}

Differentiating with respect to T:

$$\frac{dS_{net}}{dT} = 4\sigma T^3$$

Rewriting in terms of dT/dS_{net}:

$$\frac{dT}{dS_{net}} = \frac{1}{4\sigma T^3}$$

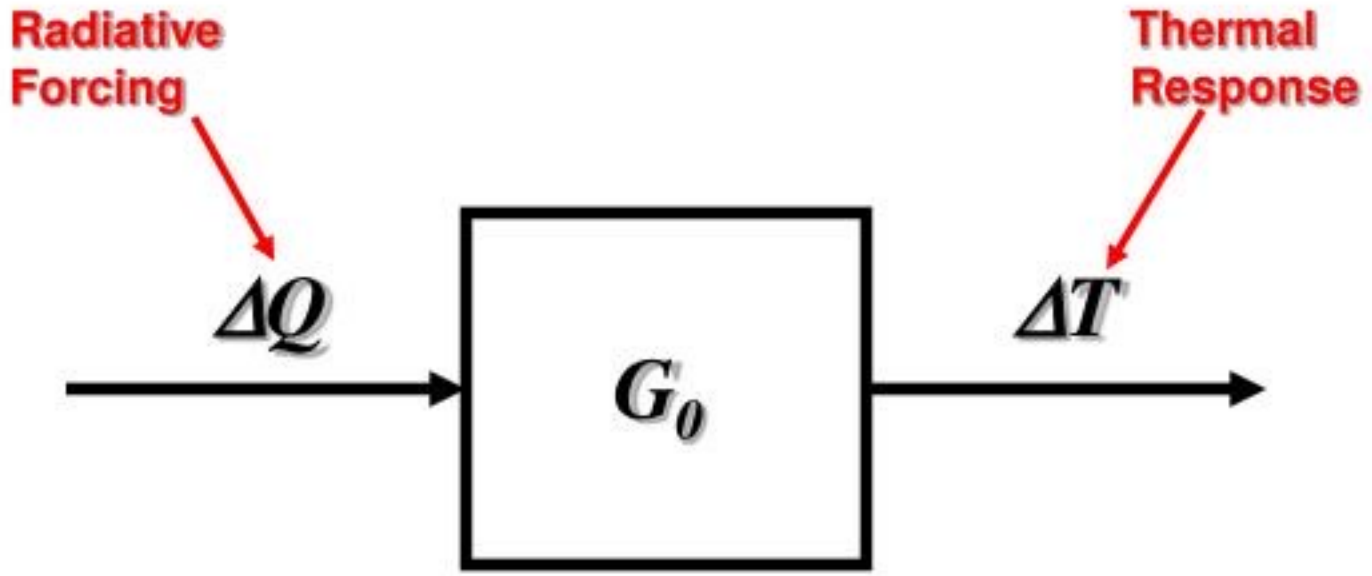
Expressing as finite differences and assuming that all perturbations to the global energy balance are equivalent:

$$\frac{\Delta T}{\Delta Q} = G_0 \equiv \frac{1}{4\sigma T^3}$$

In this simple model, G_0 is the gain of the climate system. For T = 255K,

$$G_0 = \frac{1}{(4)(5.67\times10^{-8})(255)^3} = 0.266\,K\,W^{-1}\,m^2$$

Schematic Diagram of Zero-Dimensional Climate Model



$$\Delta T = G_0 \Delta Q$$

Radiative Feedbacks

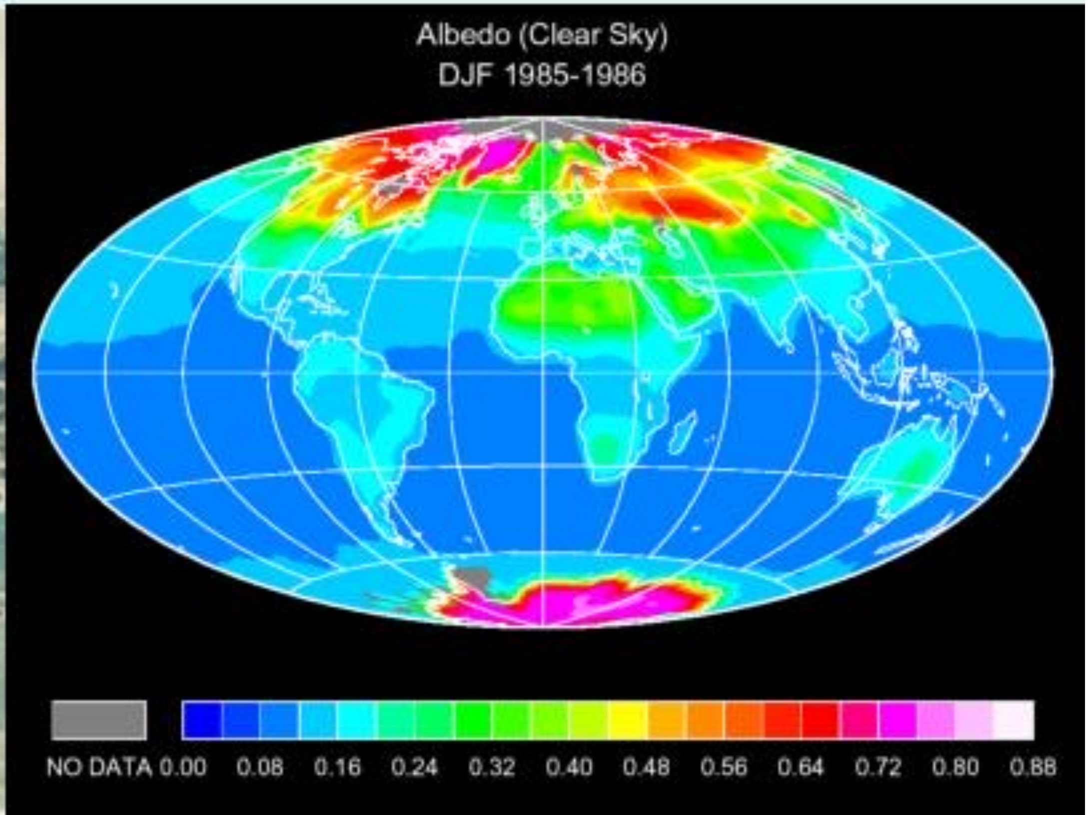
- Some properties of the climate system affect the global radiation balance.
- If these properties change as Earth warms or cools, they can lead to further changes in climate.
- Such changes are called radiative feedbacks.

Snow-Ice-Albedo Feedback

- In a warmer climate, snow cover and sea ice extent are reduced.
- Reduced snow cover and sea ice extent decrease the surface albedo of the earth, allowing more solar radiation to be absorbed.
- Increased absorption of solar radiation leads to a further increase in temperature.
- This is a positive feedback.

Ice-Albedo Feedback

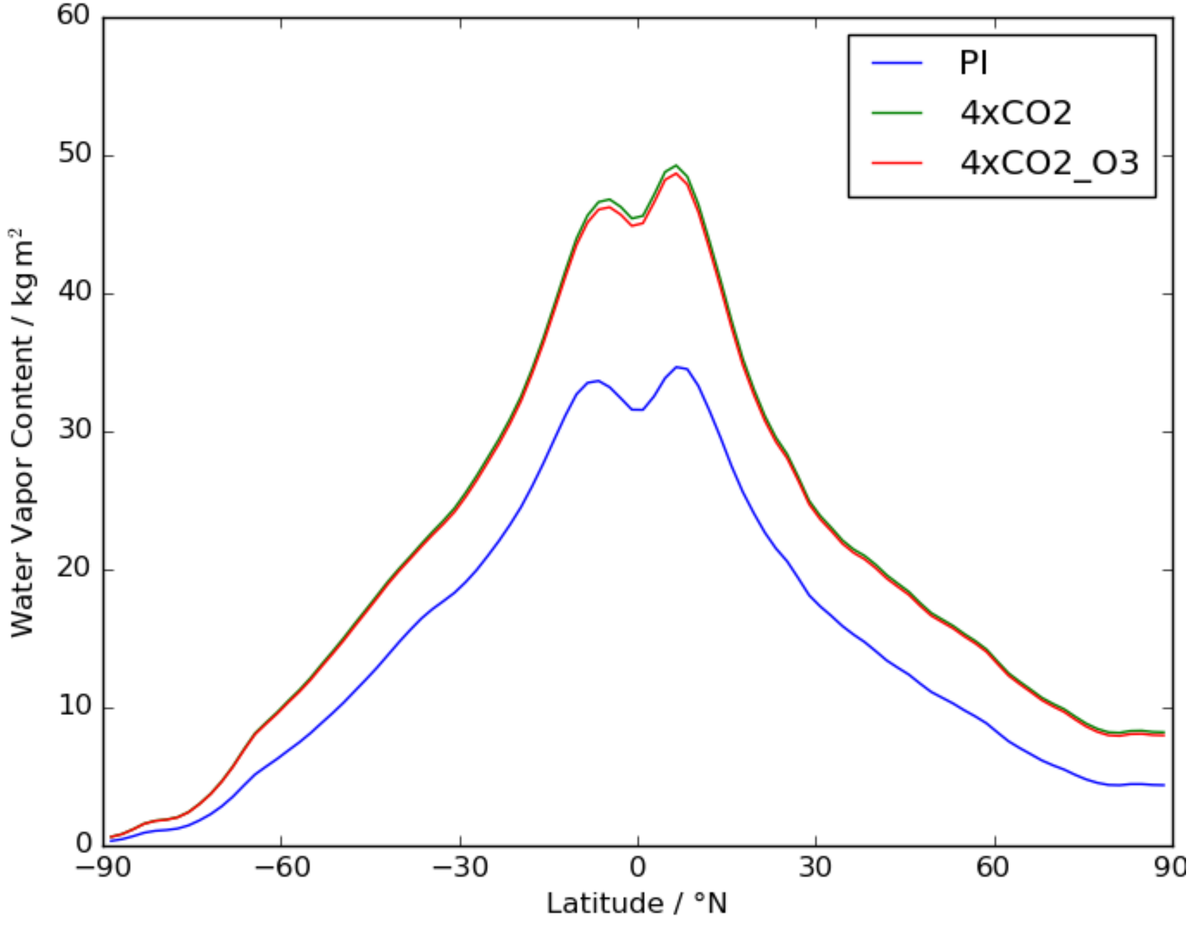
- Ice reflects more solar radiation than other surfaces
- As the Earth warms, ice melts in high latitudes and altitudes
- This lowers the albedo of Earth and leads to further warming.



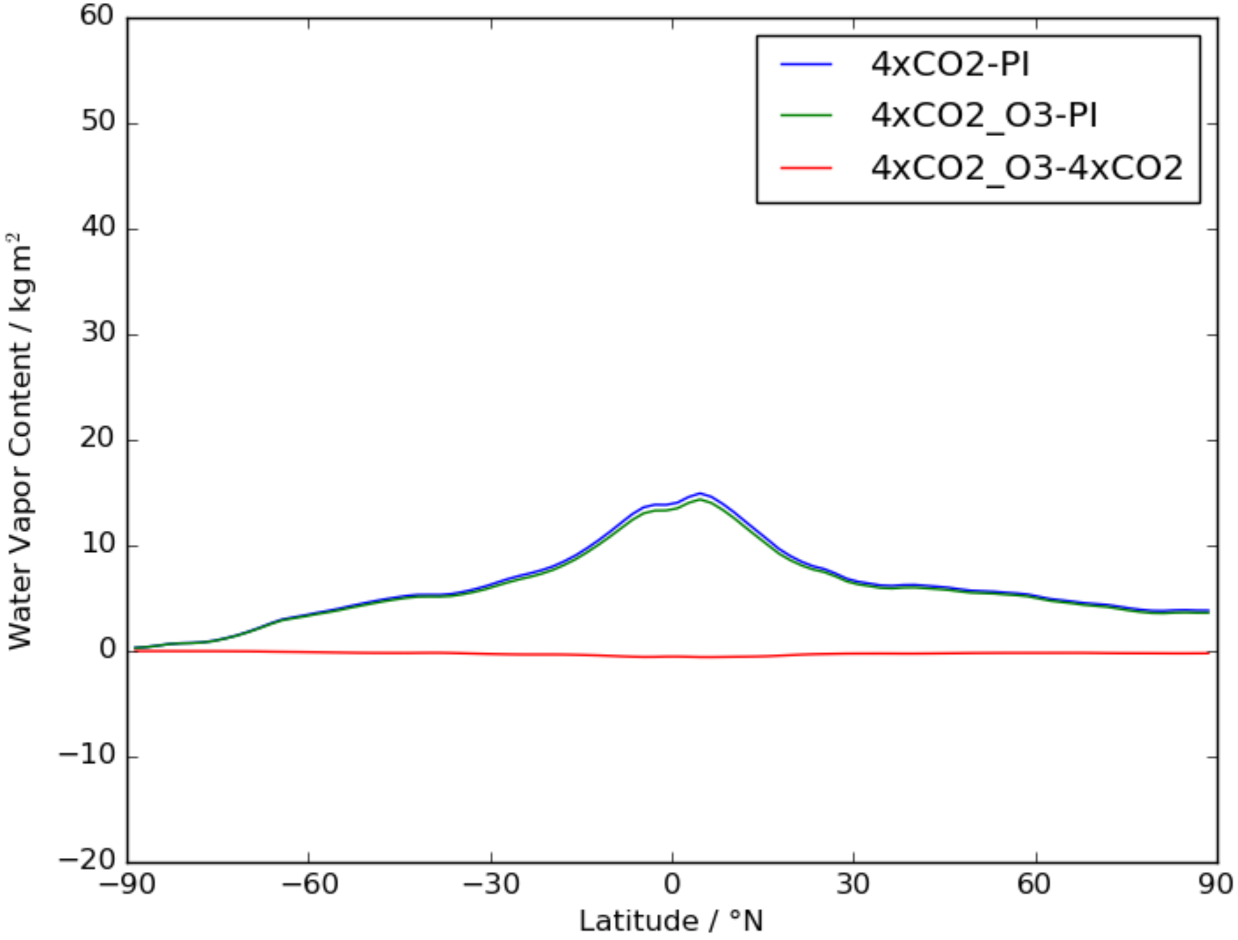
Water Vapor Feedback

- In a warmer climate, increases in saturation vapor pressure allow water vapor to increase.
- Increased water vapor increases the infrared opacity of the atmosphere.
- The reduction in outgoing longwave radiation leads to a further increase in temperature.
- This is a positive feedback.

Water vapor



PI
4xCO2
4xCO2_O3
Water Vapor Content / kg m^{2}
Latitude / °N



4xCO2-PI
4xCO2_O3-PI
4xCO2_O3-4xCO2
Water Vapor Content / kg m^{2}
Latitude / °N

Water Vapor Feedback

Effect on long-term response to doubled CO_2

$$\Delta T = \lambda \cdot \Delta Q$$

λ is a measure of climate sensitivity;
^{o}K per Wm^{-2} of climate forcing

λ_o = for fixed absolute humidity = 0.25 $^{o}K/(Wm^{-2})$

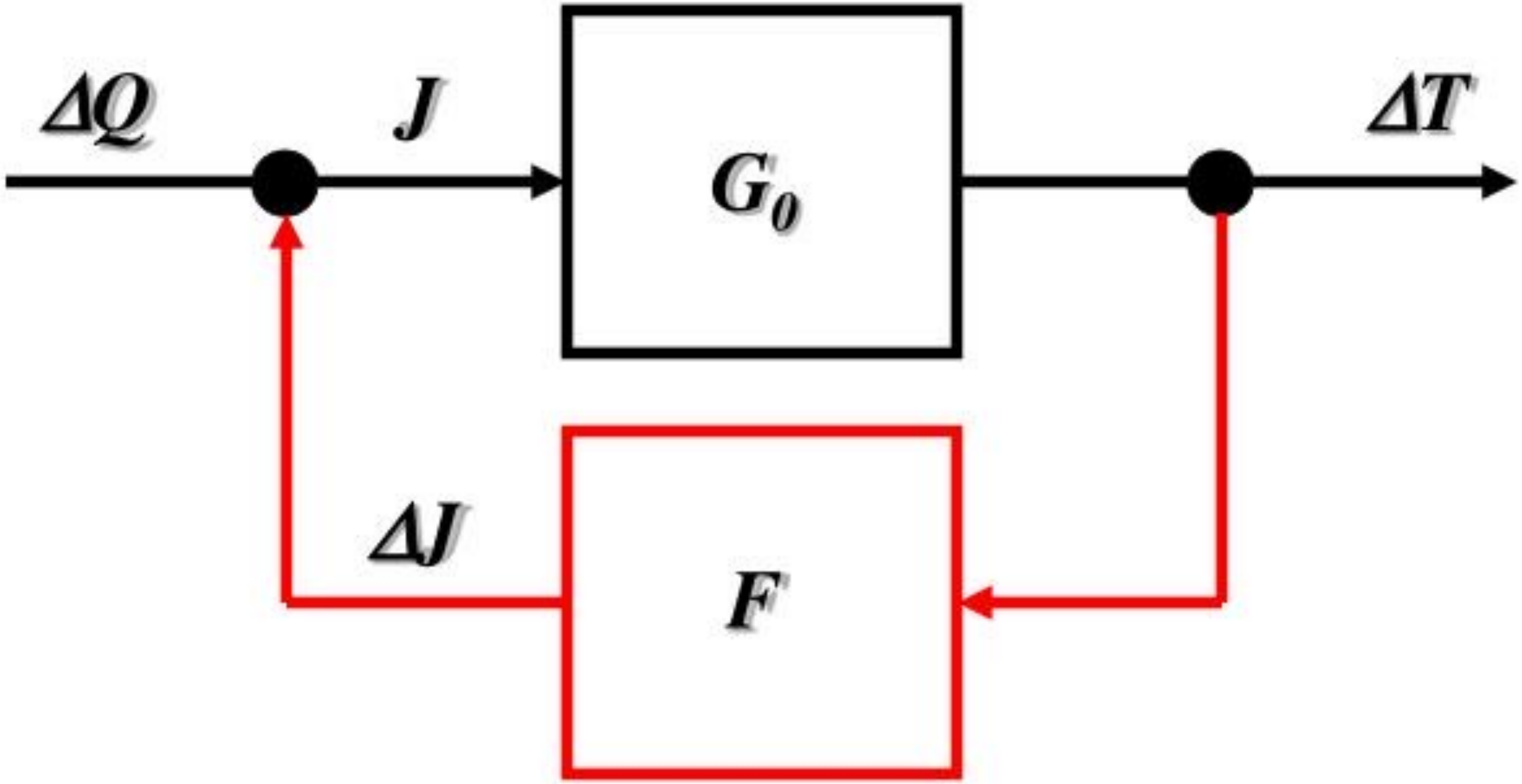
λ_{RH} = for fixed relative humidity = 0.50 $^{o}K/(Wm^{-2})$

$$\lambda_{RH}^{-1} = 2.0 \pm 0.5 Wm^{-2} K^{-1}$$ (NRC, 1979, still good?)

$$\Delta Q_{2\times CO_2} = 4 Wm^{-2} \quad gives$$

$$1.6C < \Delta T < 2.7C$$

Zero-Dimensional Climate Model With Feedbacks



$$\Delta T = G_0 J = G_0\left(\Delta Q + F\Delta T\right)$$

$$\Delta T = G_0\left(\Delta Q + F\Delta T\right)$$

Solving for ΔT:

$$\Delta T = \frac{G_0}{1-f}\Delta Q \qquad\qquad f \equiv G_0 F$$

This can also be written as

$$\Delta T = G_f \Delta Q \qquad\qquad G_f \equiv \frac{G_0}{1-f}$$

Larger positive F → larger G_f → larger ΔT

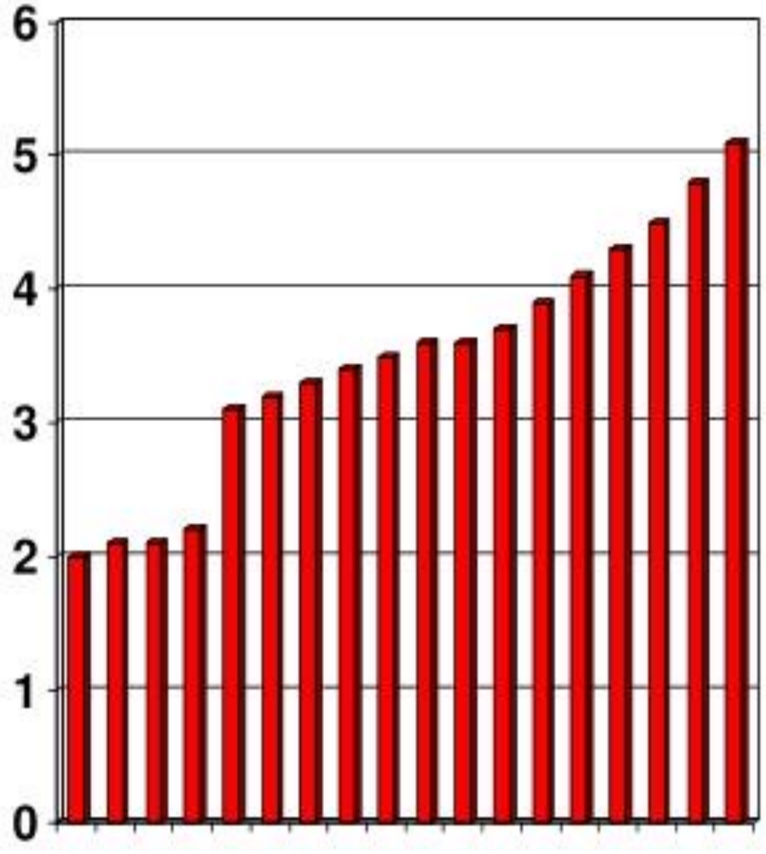
Climate sensitivity is sometimes expressed in terms of the equilibrium warming that would result from a doubling of atmospheric CO_2:

$$\Delta T_{2x} = G_f \Delta Q_{2x}$$

$$\Delta Q_{2x} \approx 4 W\, m^{-2}$$

Simulated Climate Sensitivity

- The equilibrium global warming to a doubling of CO_2 (ΔT_{2xCO2}) simulated by current climate models varies over a relatively wide range.
- IPCC: 66% chance that ΔT_{2xCO2} lies within 2.0-4.5 K; 95% chance that it is >1.5 K.



Forcings vs. Feedbacks

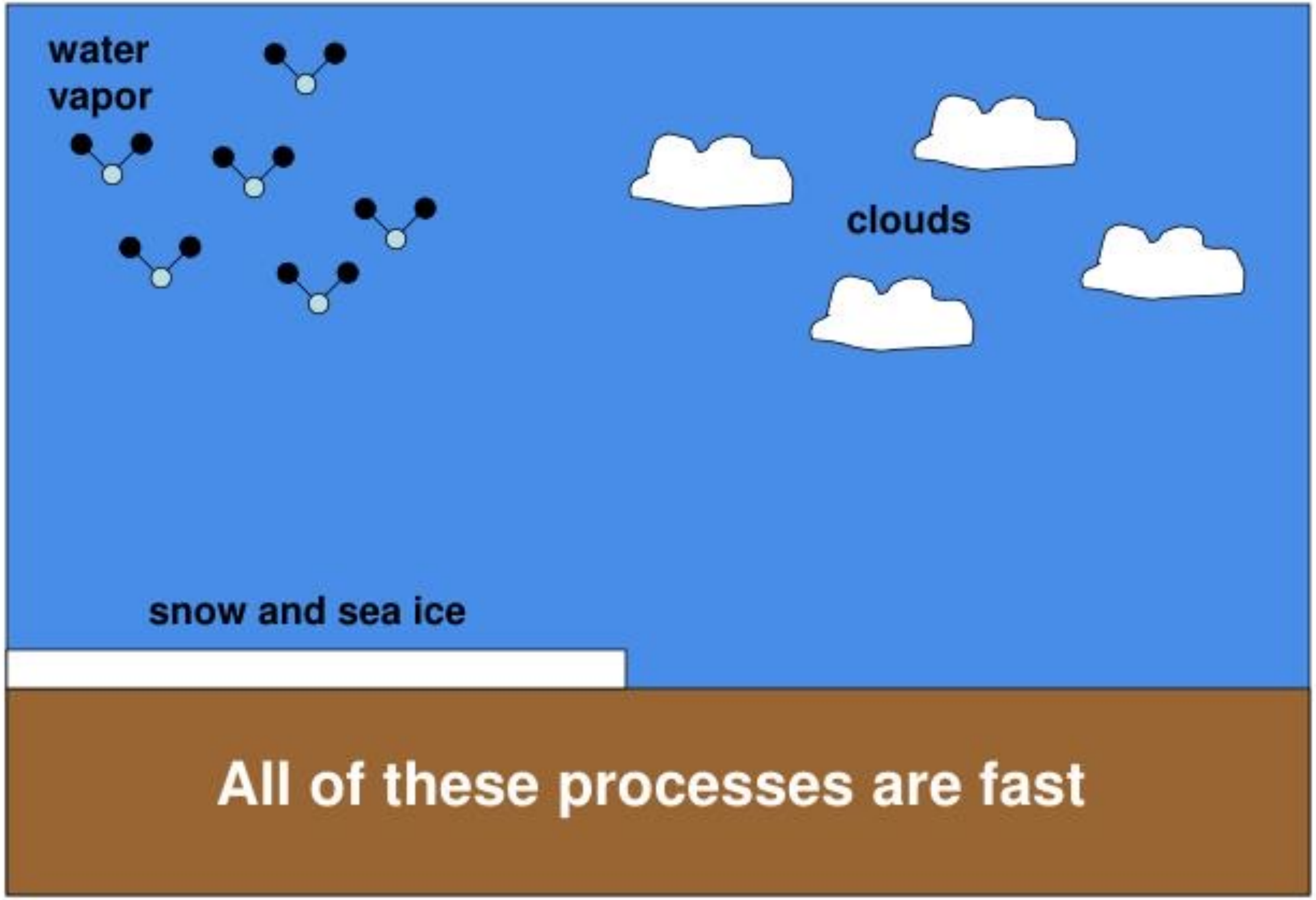
- When considering the real climate system, the distinction between forcings and feedbacks can sometimes be unclear.
- Example: CO_2 is regarded as an external forcing of future climate change, but natural, climate-dependent CO_2 variations have occurred in Earth's past.

Forcings vs. Feedbacks

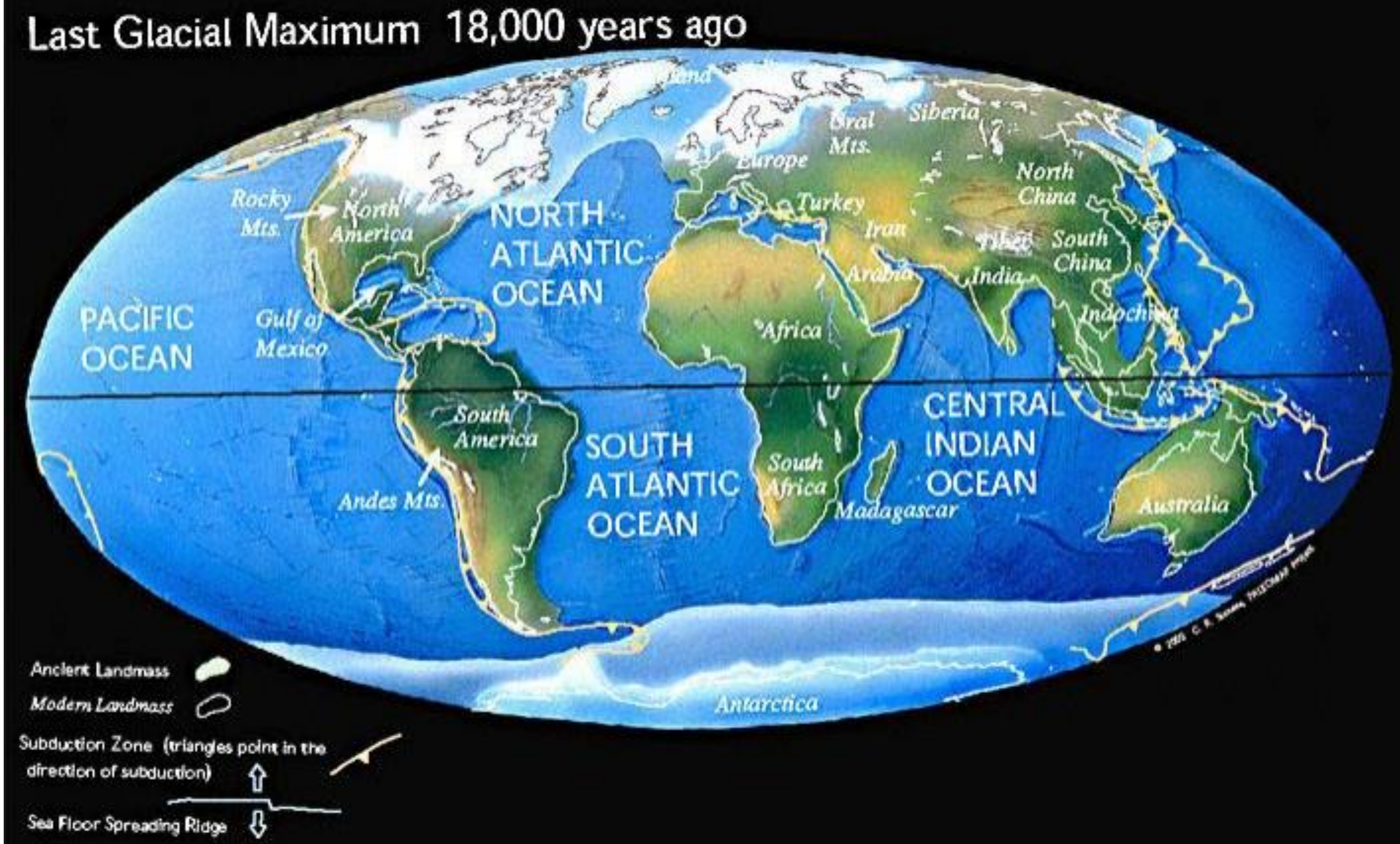
- Distinction depends on the definition of the climate system.
- In a model framework, forcings and feedbacks can be distinguished more readily.
- Forcing → process external to the system
- Feedback → process internal to the system

Fast vs. Slow Processes

- When using paleoclimate information to evaluate climate sensitivity for application to decadal-to-centennial scale climate change, it is useful to distnguish between “fast” and “slow” processes.
- Fast $\rightarrow$ time scales of years to decades
- Slow $\rightarrow$ time scales of centuries or longer


water
vapor
clouds
snow and sea ice
All of these processes are fast

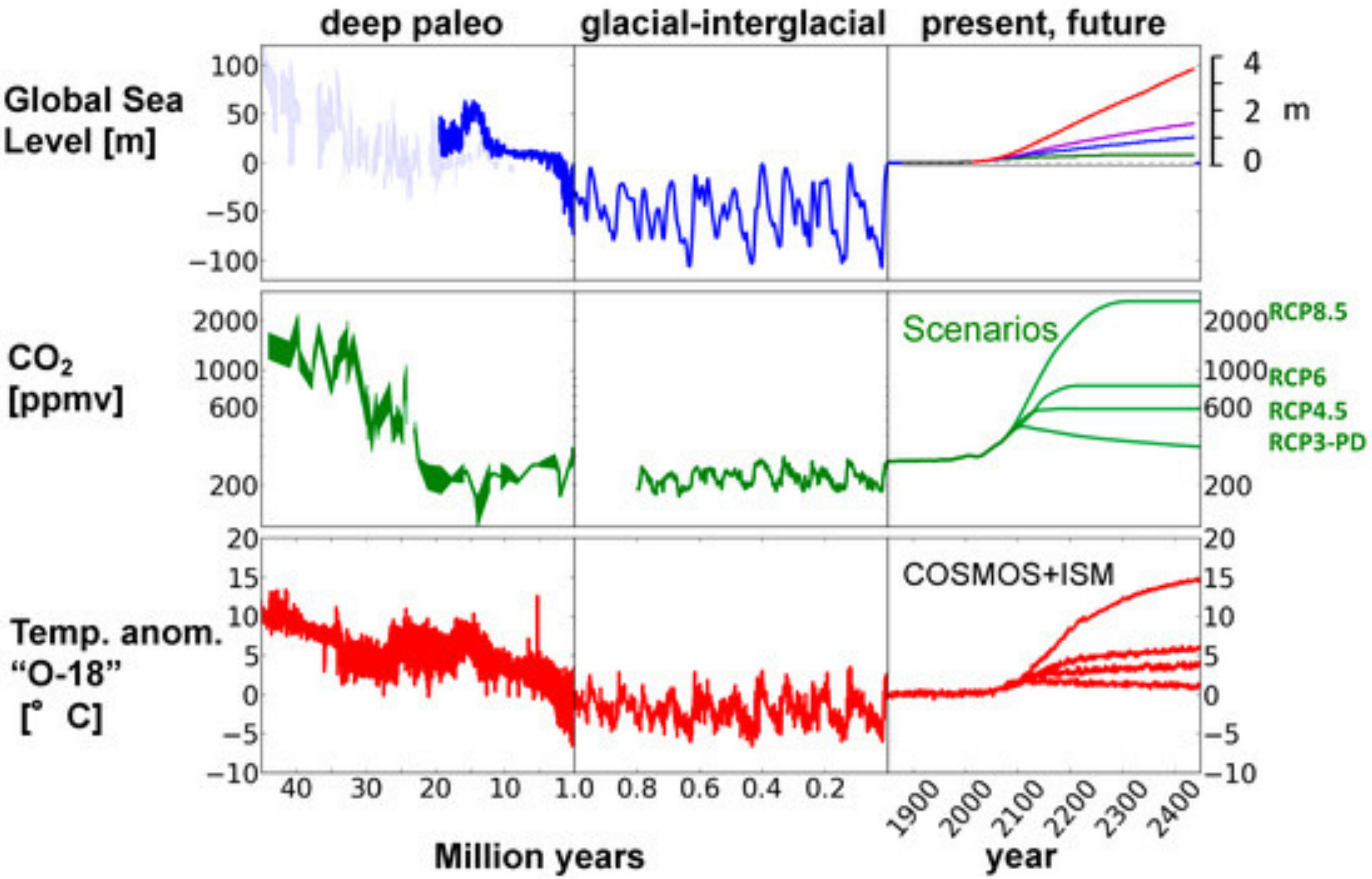
Learning from the Past



Radiative Feedbacks Involving Slow Processes

- Growth and decay of large continental ice sheets (albedo)
- Climate-dependent changes in vegetation (albedo)
- Biogeochemical changes in carbon cycle (atmospheric CO_2, CH_4)
- Tectonics (many indirect effects)

Natural variability and perturbed climate



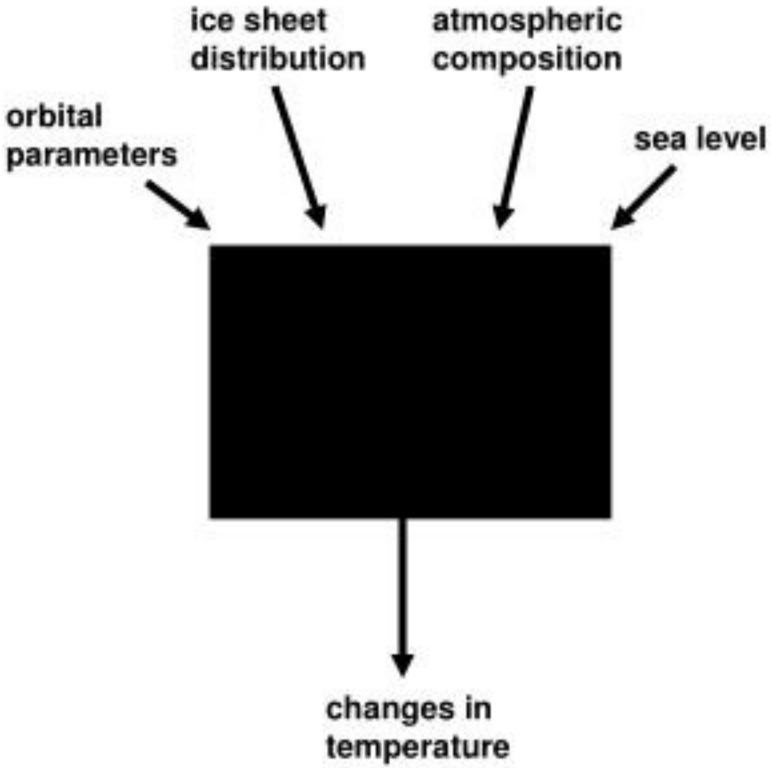
Lohmann et al., 2020b

Evaluating Climate Sensitivity Using "Paleocalibration"

- Determine ΔQ and ΔT from paleodata.
- Compute G_f (a.k.a. λ) from ΔQ and ΔT.
- Compare the "paleocalibrated" G_f value with model-derived or empirically derived estimates.

Evaluating Climate Sensitivity

- For evaluating climate sensitivity resulting from fast feedback processes (i.e., those most relevant to dec-cen climate change), external forcings and results of slow processes can be taken as inputs.



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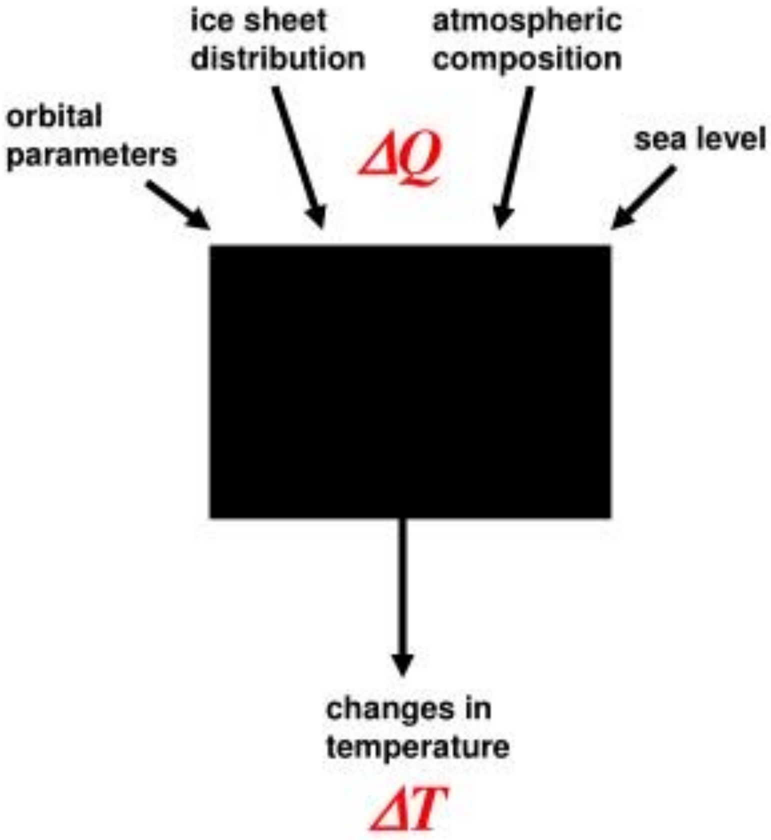

Evaluating Climate Sensitivity Using "Paleocalibration"

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Evaluating Climate Sensitivity Using Paleoclimate Modeling

- Determine required forcings (including those resulting from slow feedback processes).
- Apply these forcings to climate model.
- Compare resulting changes in temperature to those reconstructed from geological data.

Advantages and Disadvantages

“Paleocalibration”

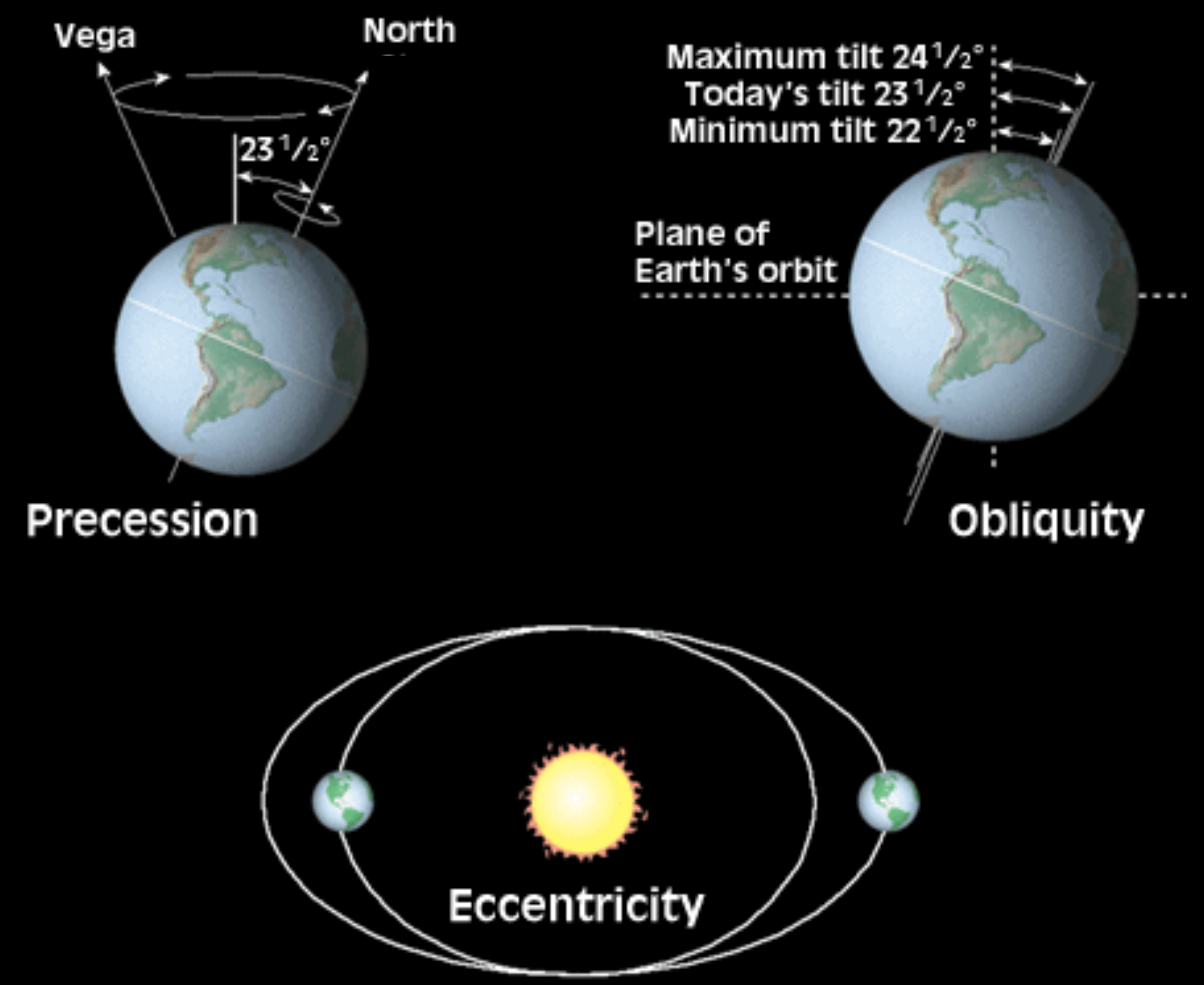
+ Results are independent of climate models.
+ Results can easily be revised when new estimates of forcing or response become available.
- Global mean temperature estimates are required.

Paleoclimate Modeling

+ Global mean temperature estimates are not required. (More effective with good data coverage, though.)
+ Does not require the forcing-response relationship to be linear.
+ Provides additional insights beyond climate sensitivity.
- Requires extensive computation with a climate model.

Estimating Forcings: Orbital Parameters

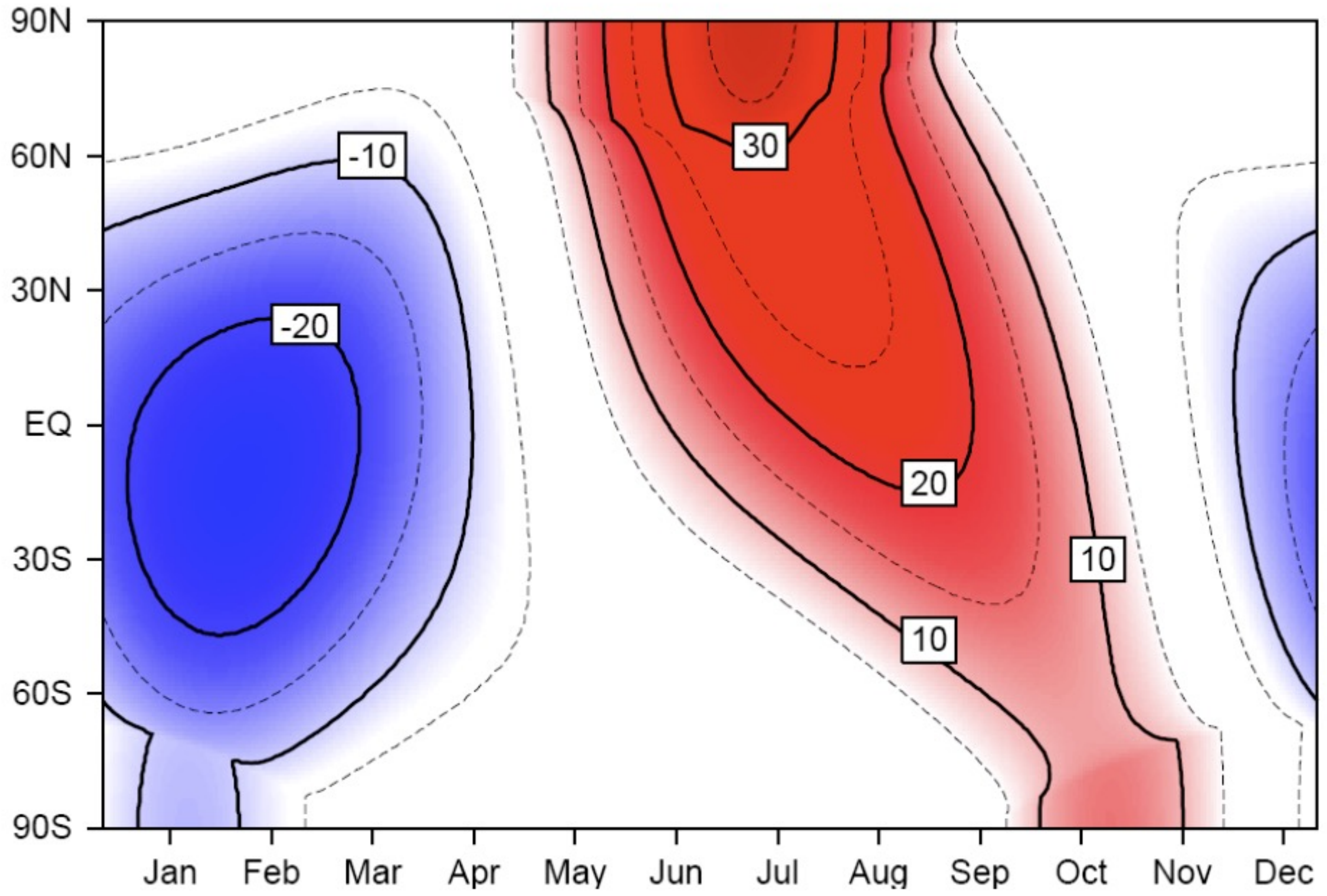
- Orbital parameters can be calculated accurately for millions of years based on orbital mechanics.
- Results of such calculations are widely available.



Vega
North
23 1/2°
Precession
Maximum tilt 24 1/2°
Today's tilt 23 1/2°
Minimum tilt 22 1/2°
Plane of
Earth's orbit
Obliquity
Eccentricity

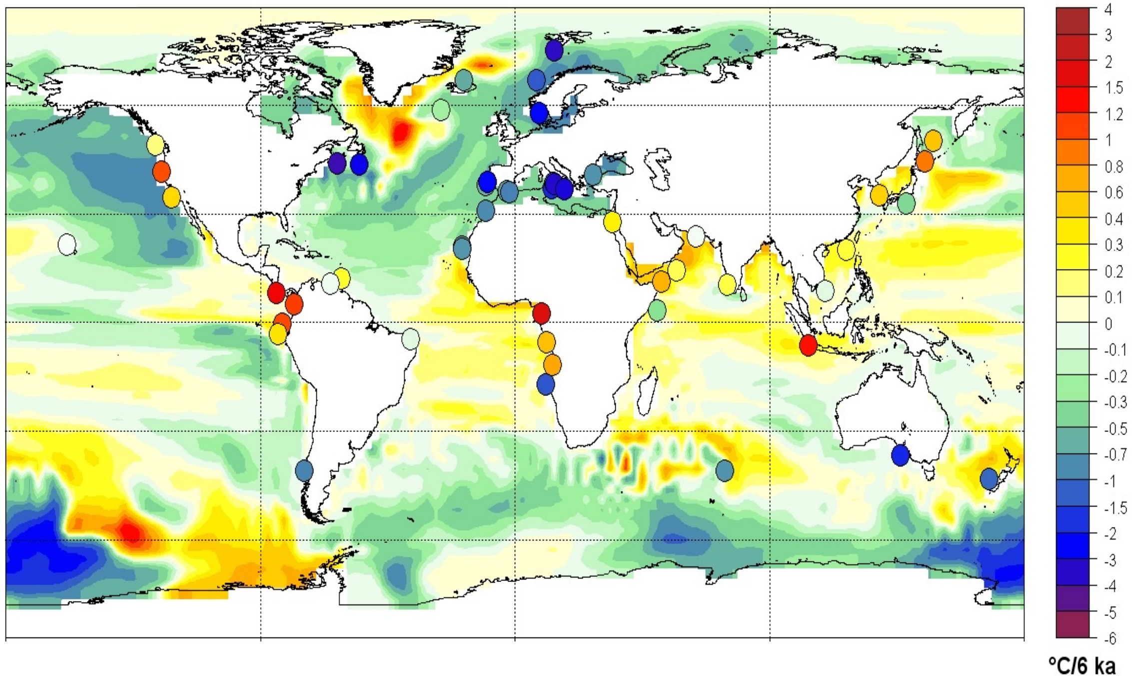
Example

Insolation (6k minus present)



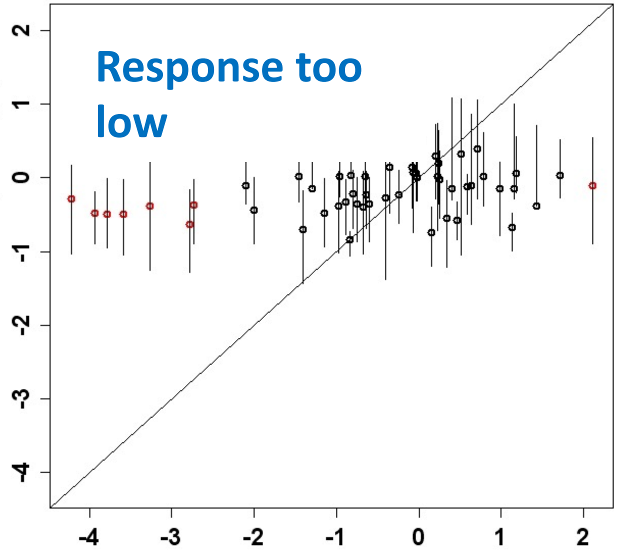
Marine temperature trends (last 6000 years)

Annual mean sea surface temperature trends



Alkenone-based temperature trends

ΔT



ΔT

Lohmann et al., 2013, CP

Estimating Forcings:
Ice Sheets and Sea Level

- Ice sheet extent can be estimated from terminal moraines.
- Evidence of ice sheet thickness may be available.
- Geophysical modeling (e.g., Peltier) of 3-D ice sheet distribution.

Estimating Forcings: Atmospheric Composition

- Fossil air can be recovered from ice cores.
- Chemical analysis of the air can yield concentrations of atmospheric constituents.

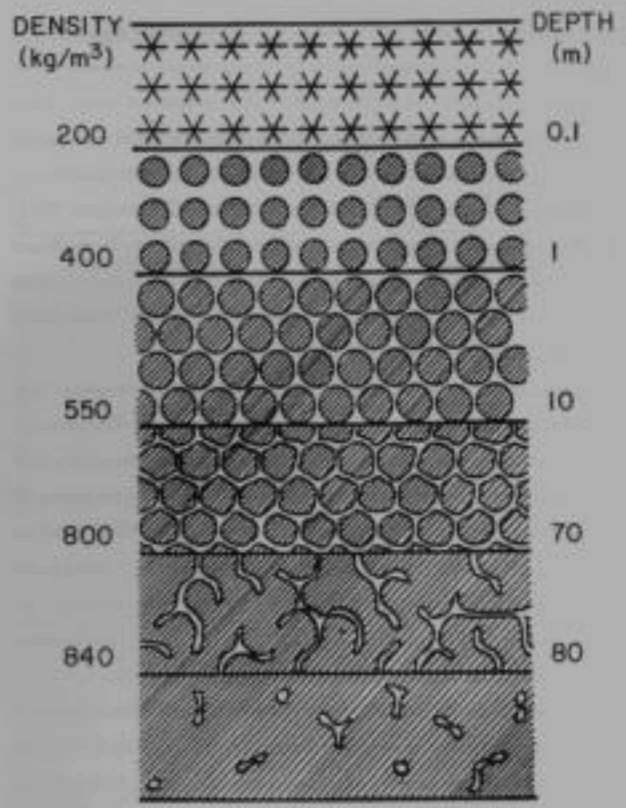


Figure 14. Diagrammatic depiction of the gradual lithification of the snow which falls on polar ice caps. The snow recrystallizes to firn which has passages through which air can circulate. When the firn is fully lithified, the air passages are sealed off becoming bubbles.

Estimating Temperature: Methods

- Mountain snowlines
- Isotopes in ice cores
- Distributions of marine microorganisms
- Alkenone molecules in marine flora
- Sr/Ca in corals
- Mg/Ca in planktonic foraminifera
- Pollen evidence of past vegetation
- Noble gases in aquifers

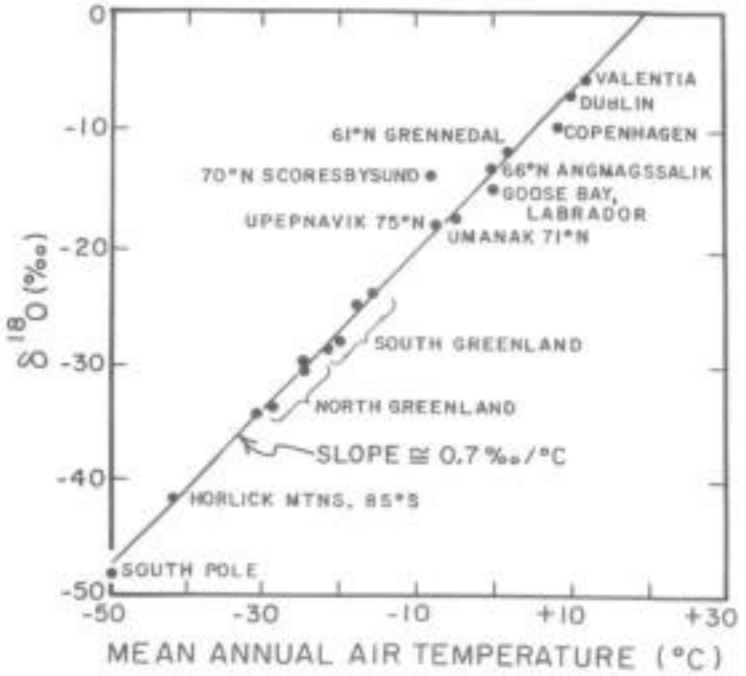
Mountain Snowlines

Mountain Snowlines

- Changes in the equilibrium lines of mountain glaciers, which can be inferred from moraines, can be interpreted in terms of temperature changes. (Ex: 200 m change x 0.6 K/100 m lapse rate = 1.2 K)
- Other factors, including moisture availability, also affect glacier growth and retreat.

Isotopes in Ice Cores

- Isotopes in precipitation have been empirically correlated with mean annual air temperature.
- Fractionation processes are responsible.



Observed $\delta^{18}O$ in average annual precipitation as a function of mean annual air temperature (Dansgaard 1964). Note that all the points in this graph are for high latitudes (>45°). (From Broecker 2002)

Distributions of Marine Microorganisms

- Determine where different species live in the modern ocean and their relationship to sea surface temperature.

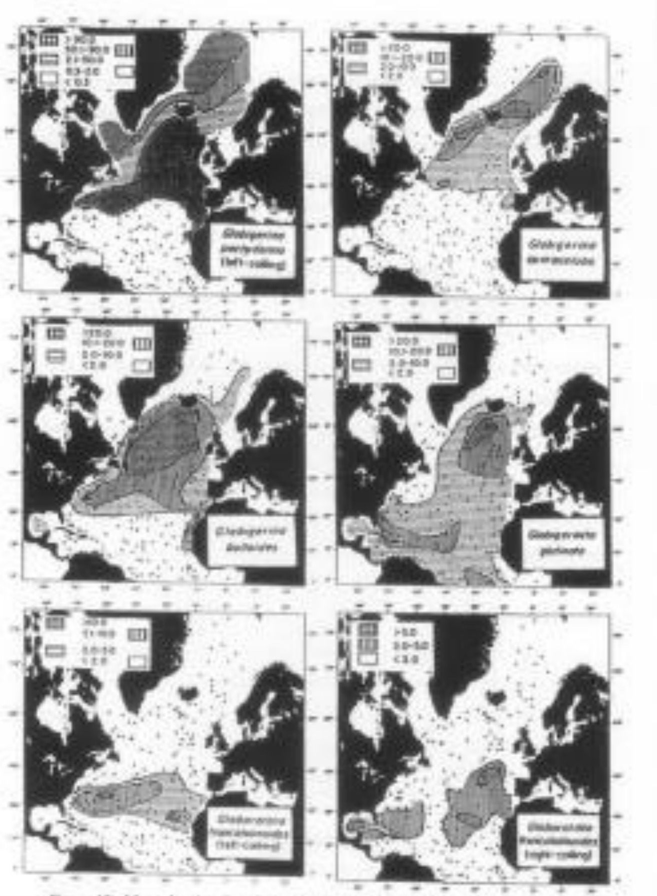
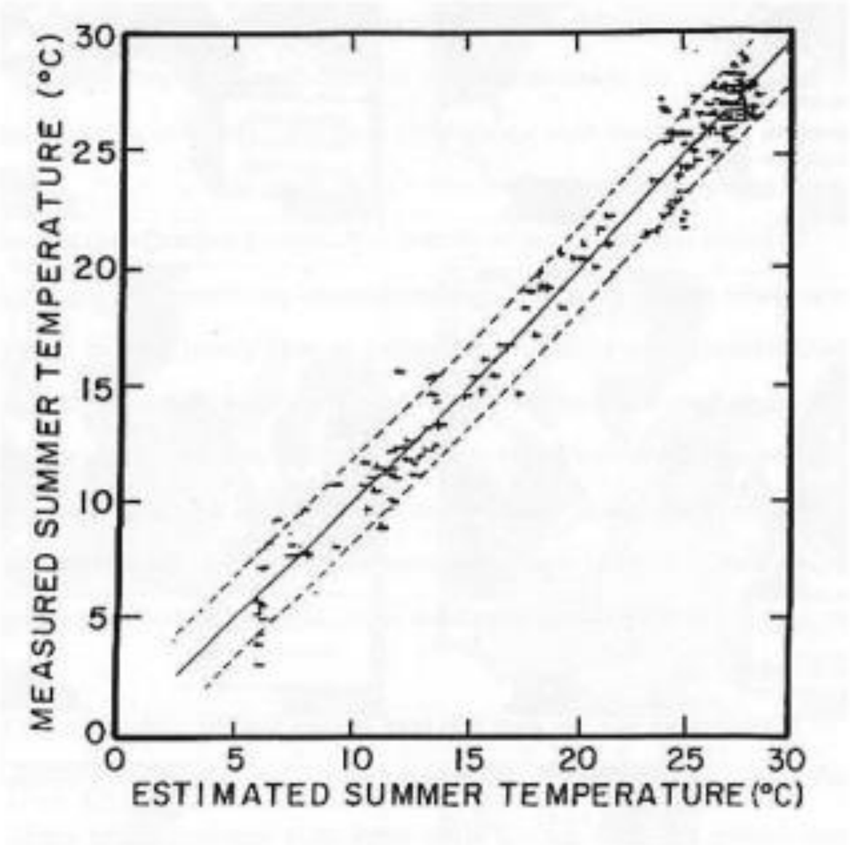


Figure 15: Maps showing the relative abundance patterns for the shells of six species of foraminifera in core top samples from the northern Atlantic (Kipp, 1976).

32

Distributions of Marine Microorganisms

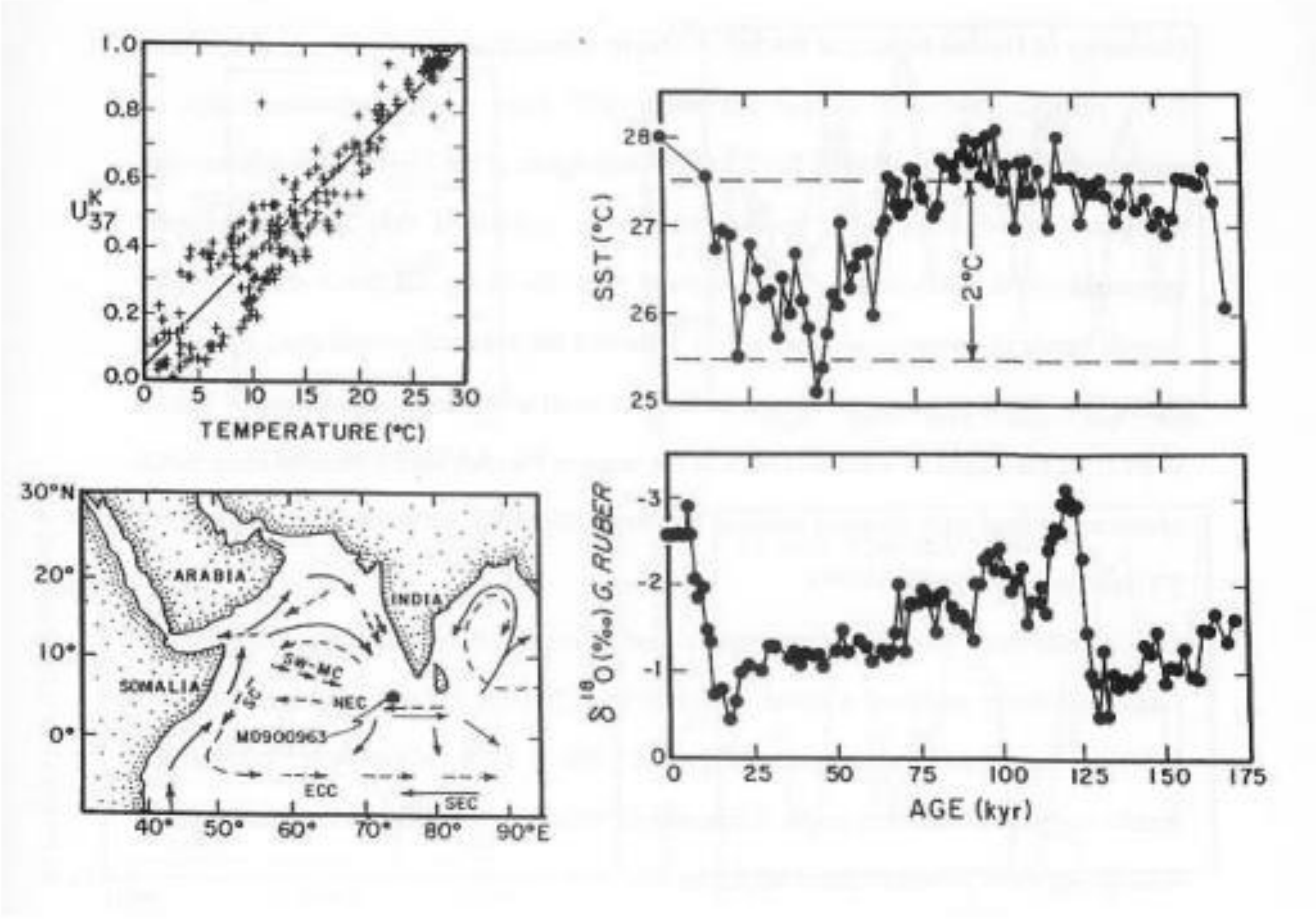
- Reconstruct past sea surface temperatures from shells recovered from deep sea sediment cores.



Alkenone Molecules in Marine Flora

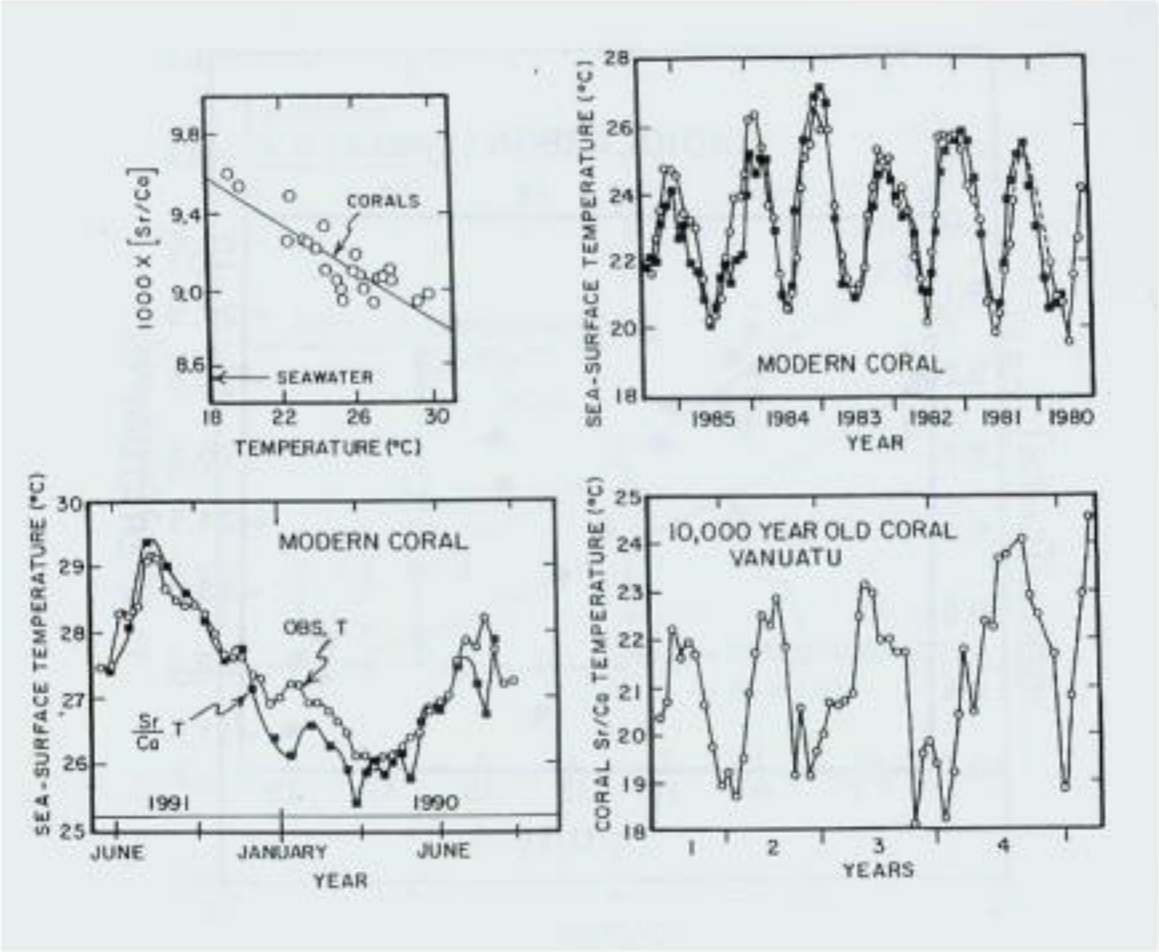
- A strong empirical relationship has been found between the ratio of two different molecules (each with 37 C atoms) and the temperature at which the macroscopic marine plants grew.
- These alkenone molecules are preserved in marine sediments.

Alkenone Molecules in Marine Flora



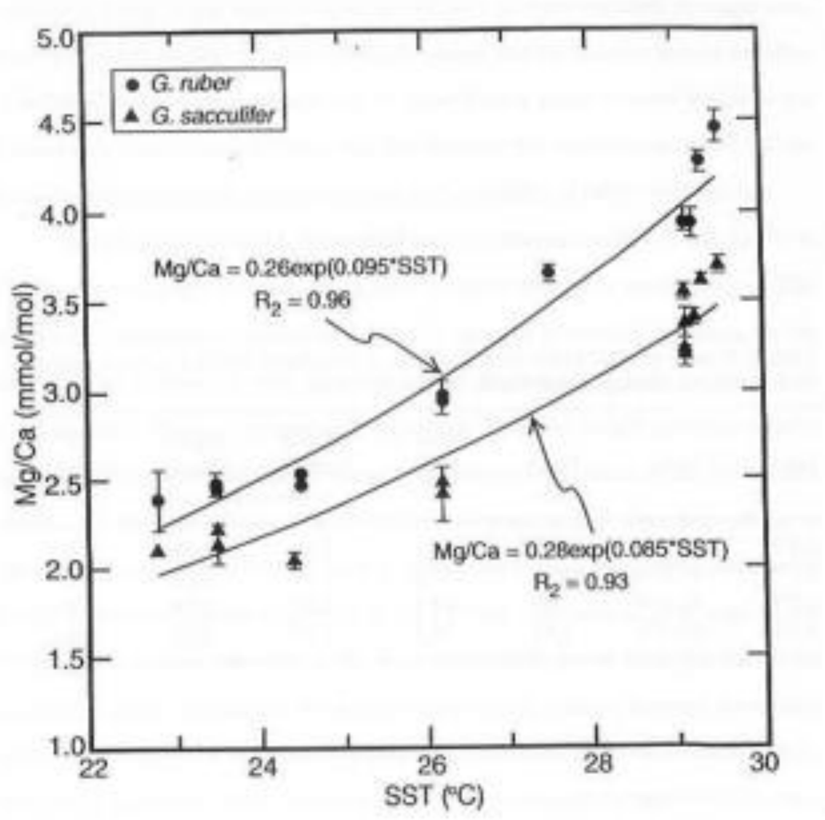
Sr/Ca in Corals

- Ratio of strontium to calcium in corals appears to be a function of temperature.



Mg/Ca in Planktonic Foraminifera

- The ratio of magnesium to calcium in planktonic foraminifera has been found to be a strong function of temperature.
- Mg/Ca and ^{18}O can be determined from the same shells.

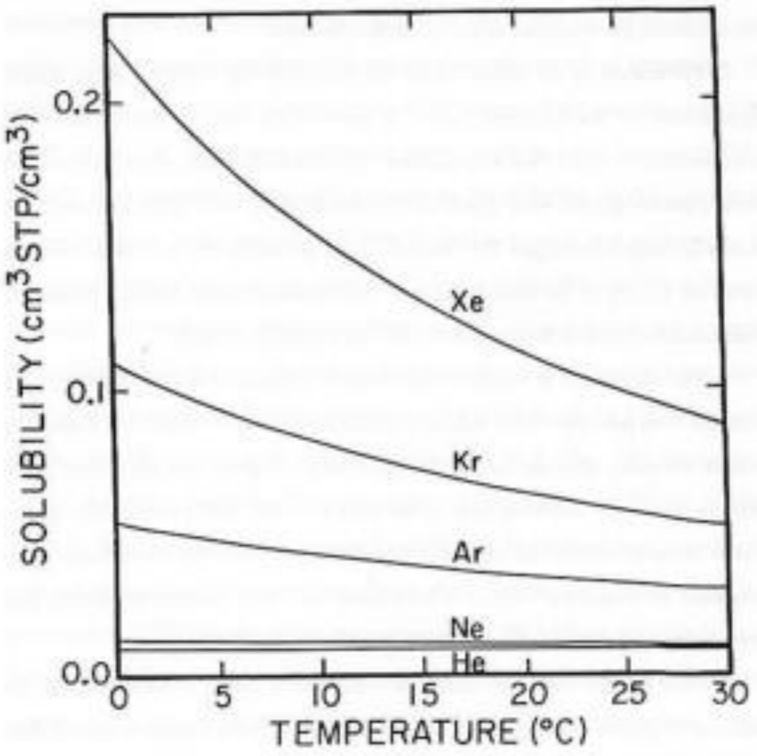


Pollen Evidence of Past Vegetation

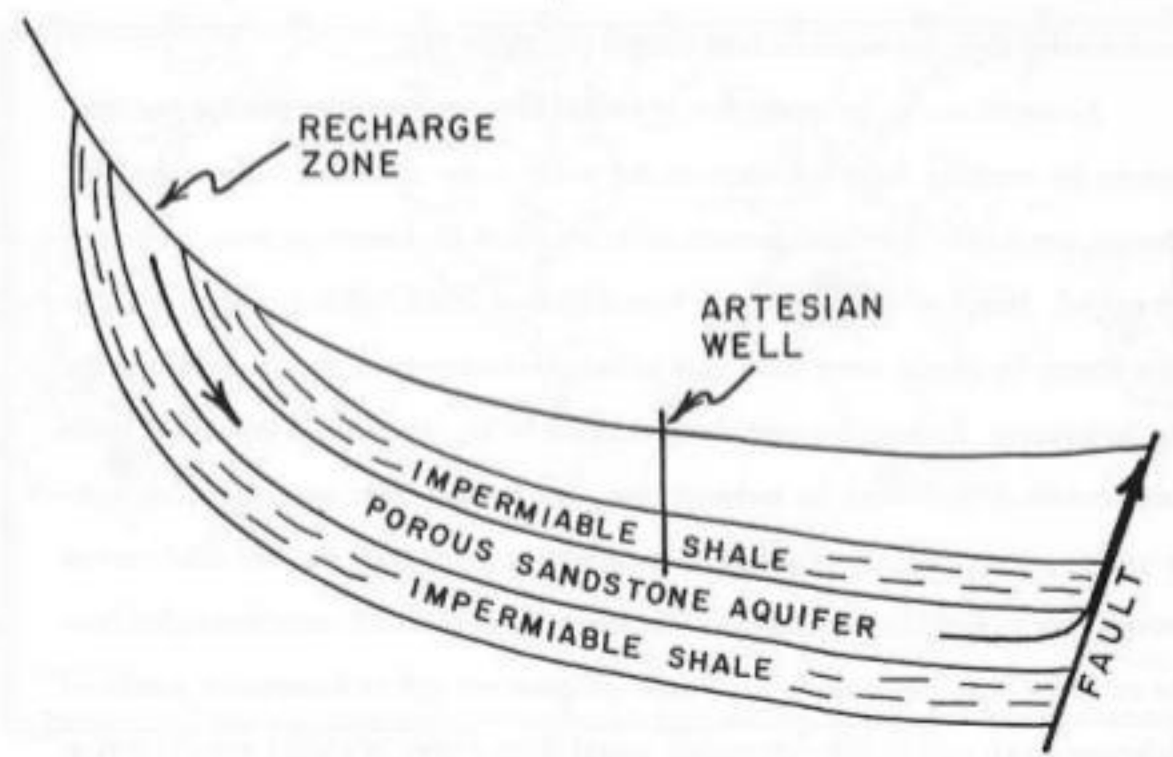
- Different plant species have different growth requirements that partly depend on climate.
- Pollen grains are distinctive and well-preserved in lake and wetland sediments.
- Changes in frequencies of pollen grains in a sediment core can be used to infer variations in climate.

Noble Gases in Aquifers

- Solubility of noble gases depends on temperature.
- Temperature dependence differs for each gas.
- Ratios can yield temperature information.



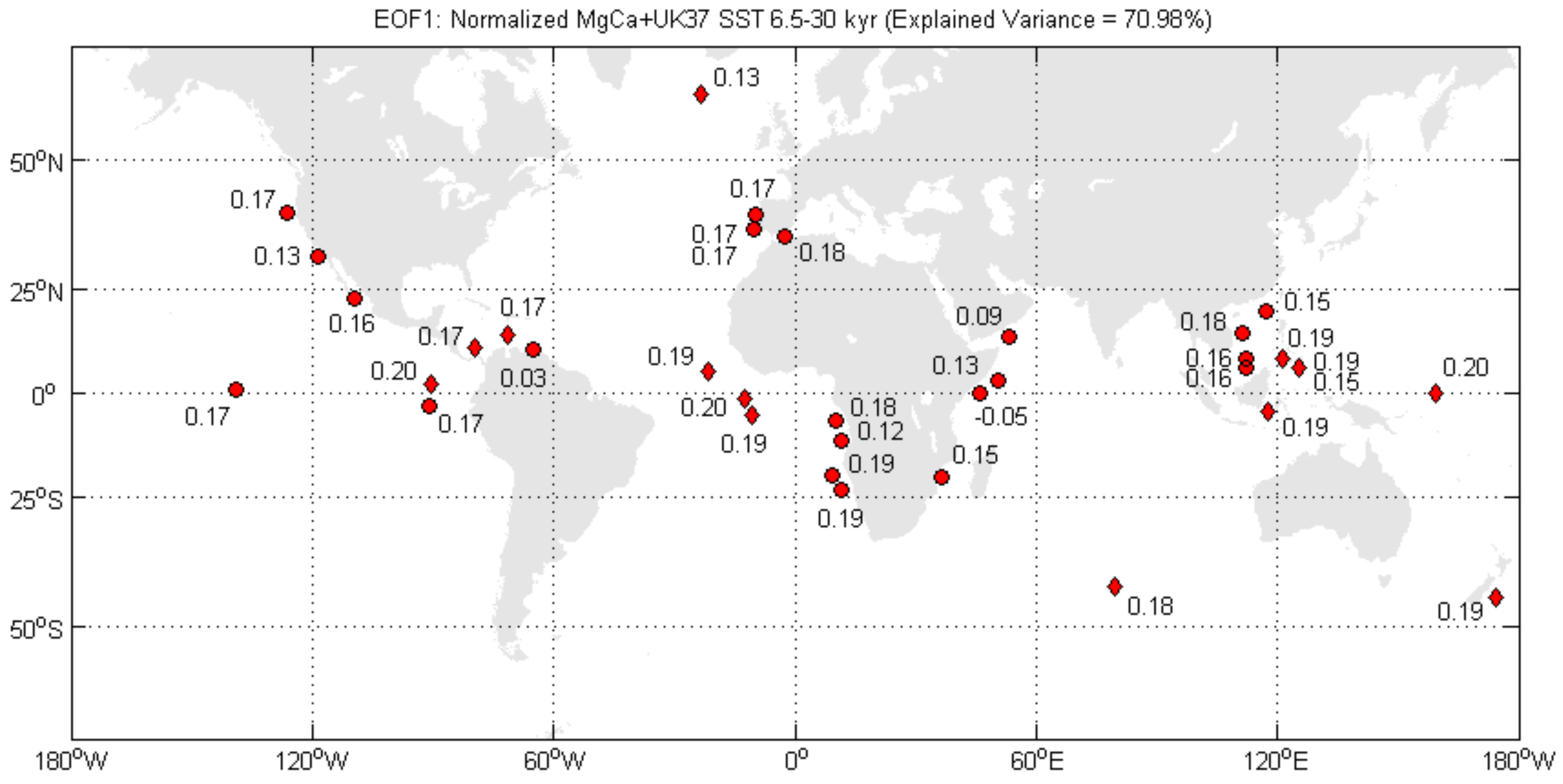
Noble Gases in Aquifers

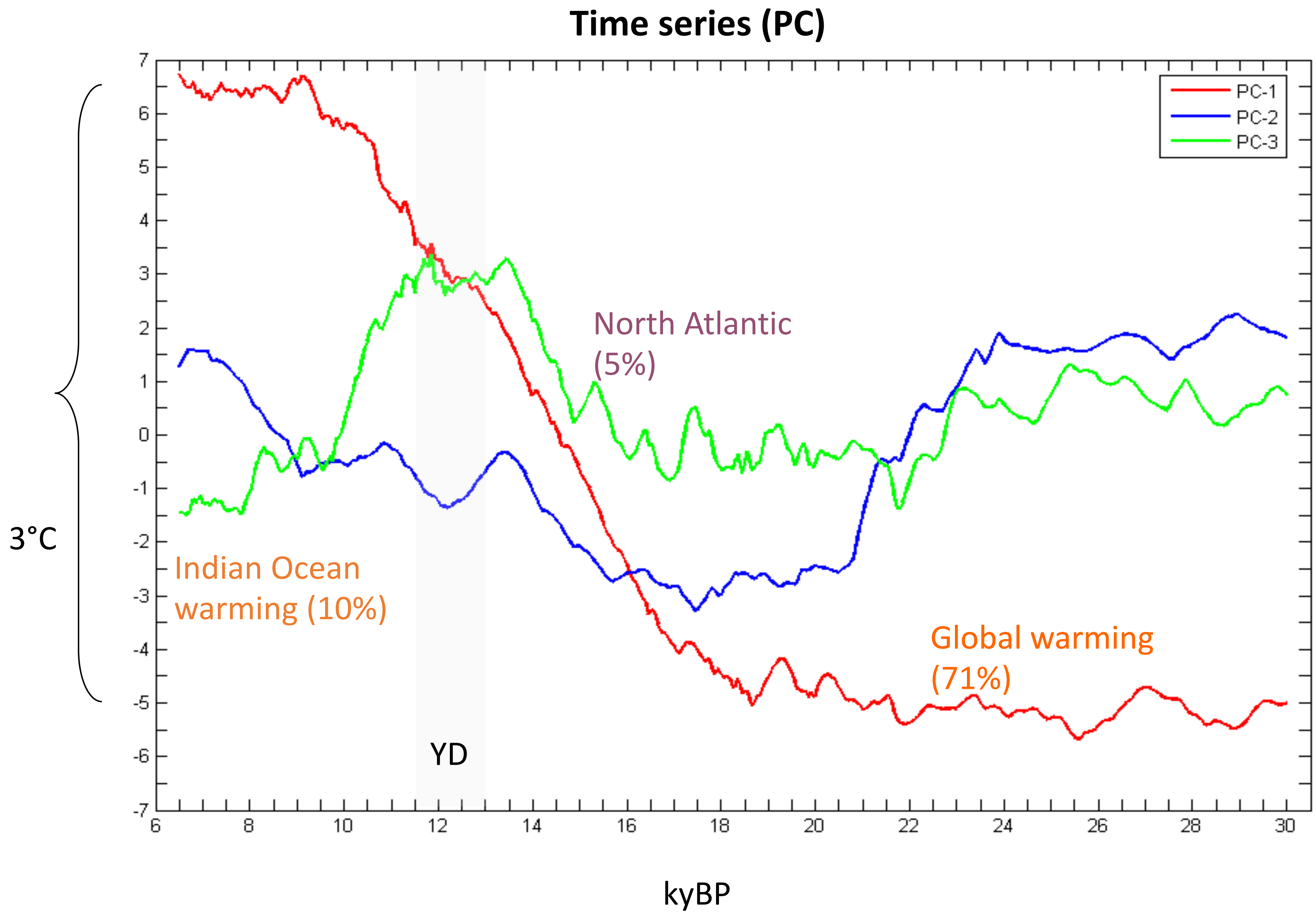


Deglacial temperatures and CO_2

Mg/Ca & Alkenones: SST proxy

EOF Analysis: temperature pattern & time series




Time series (PC)
PC-1
PC-2
PC-3
North Atlantic
(5%)
Indian Ocean
warming (10%)
Global warming
(71%)
YD
3°C
kyBP

Simple energy balance calculation

CS=3.5 K (includes water vapor, cloud feedbacks etc.):

3.5 K x $1/\ln(2)\ \ln(265/190)$ = 3.5 K x 0.48 = <u>1.68 K (CO2 effect)</u>

Global change of 3-5 K (or more)

Effects of orography, albedo, dust, other trace gases, etc. = 1-3

Consistent with modeling exercises

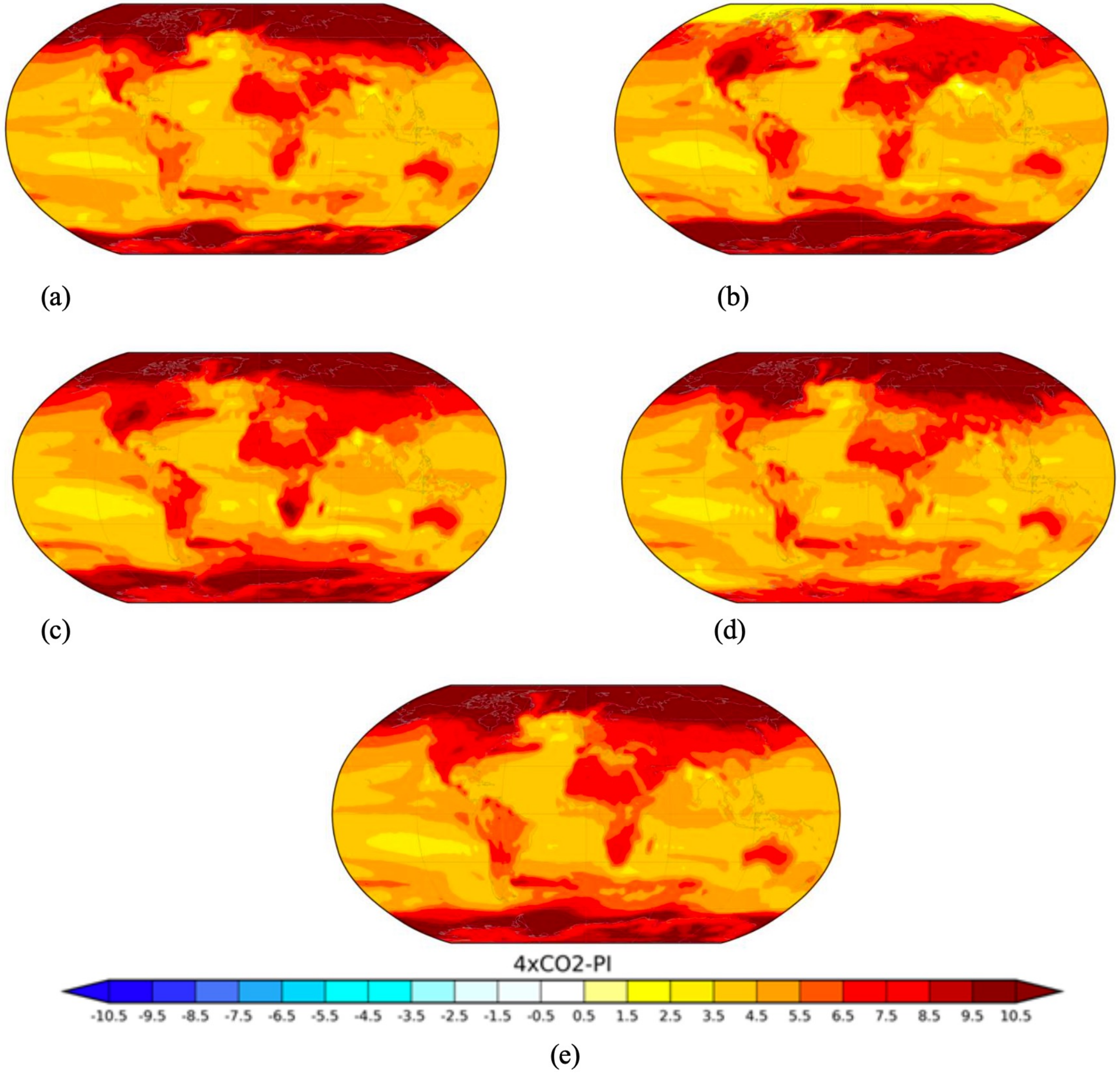


Figure 5: Surface temperature anomaly of 4xCO_2-PI for MAM (a), JJA (b), SON (c), DJF (d) and Annual (e).There are no insignificant anomalies here based on a t-test with 95% confidence interval. The unit is °C.

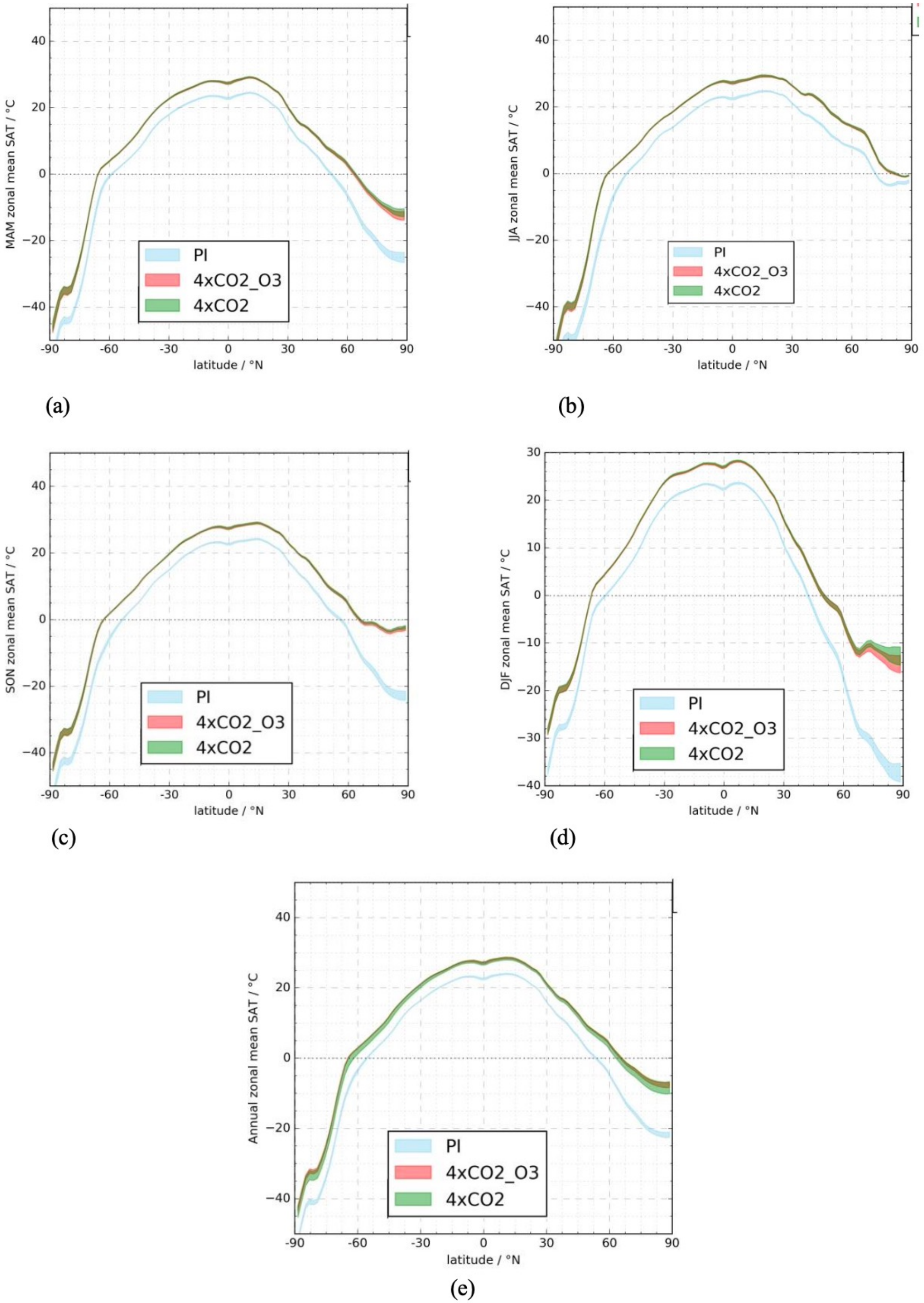
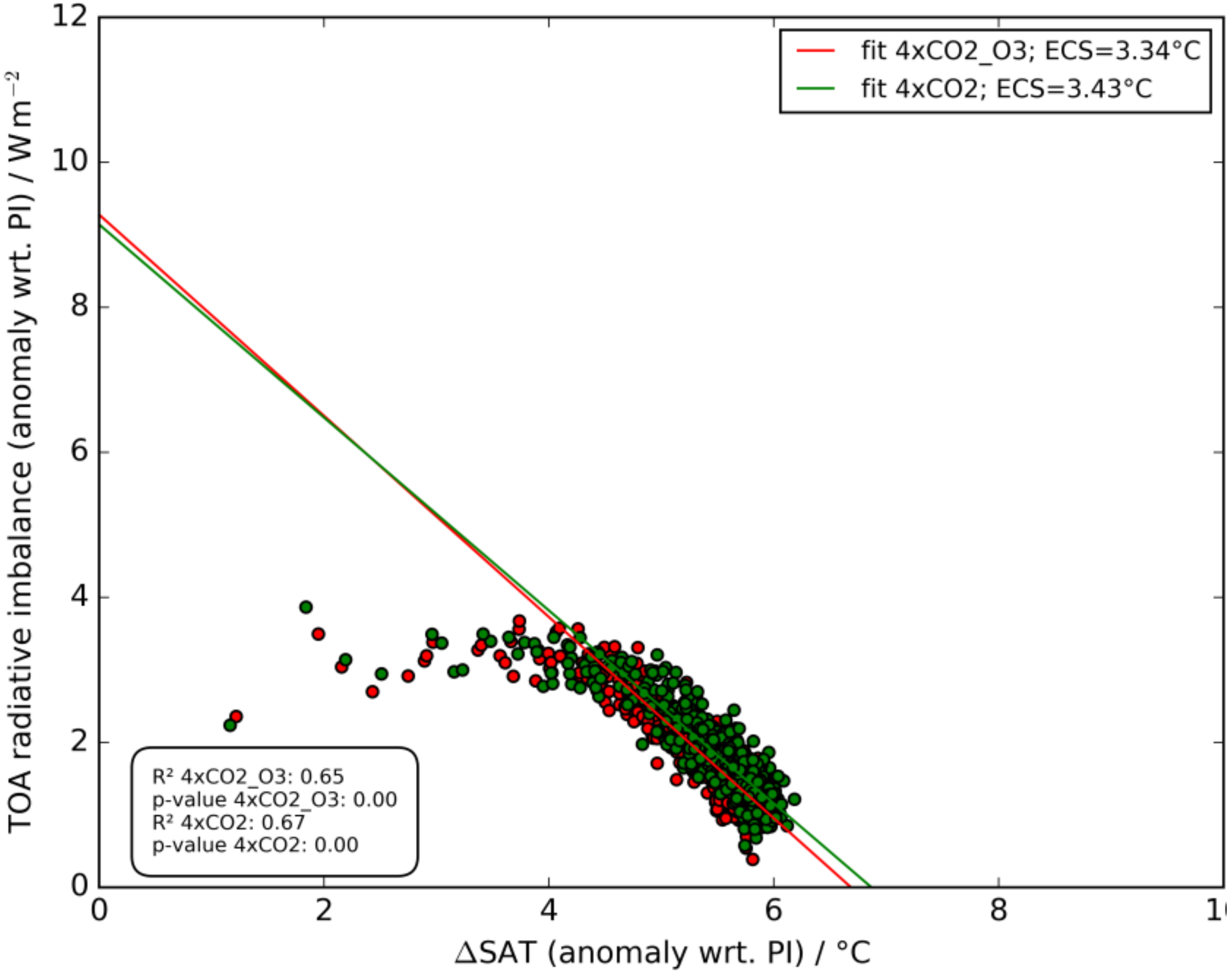


Figure 7: Meridional range of zonal mean surface air temperature (SAT) of PI, $4xCO_2$ and $4xCO_2\_O_3$ simulations for MAM (a), JJA (b), SON (c), DJF (d) and Annual mean (e).

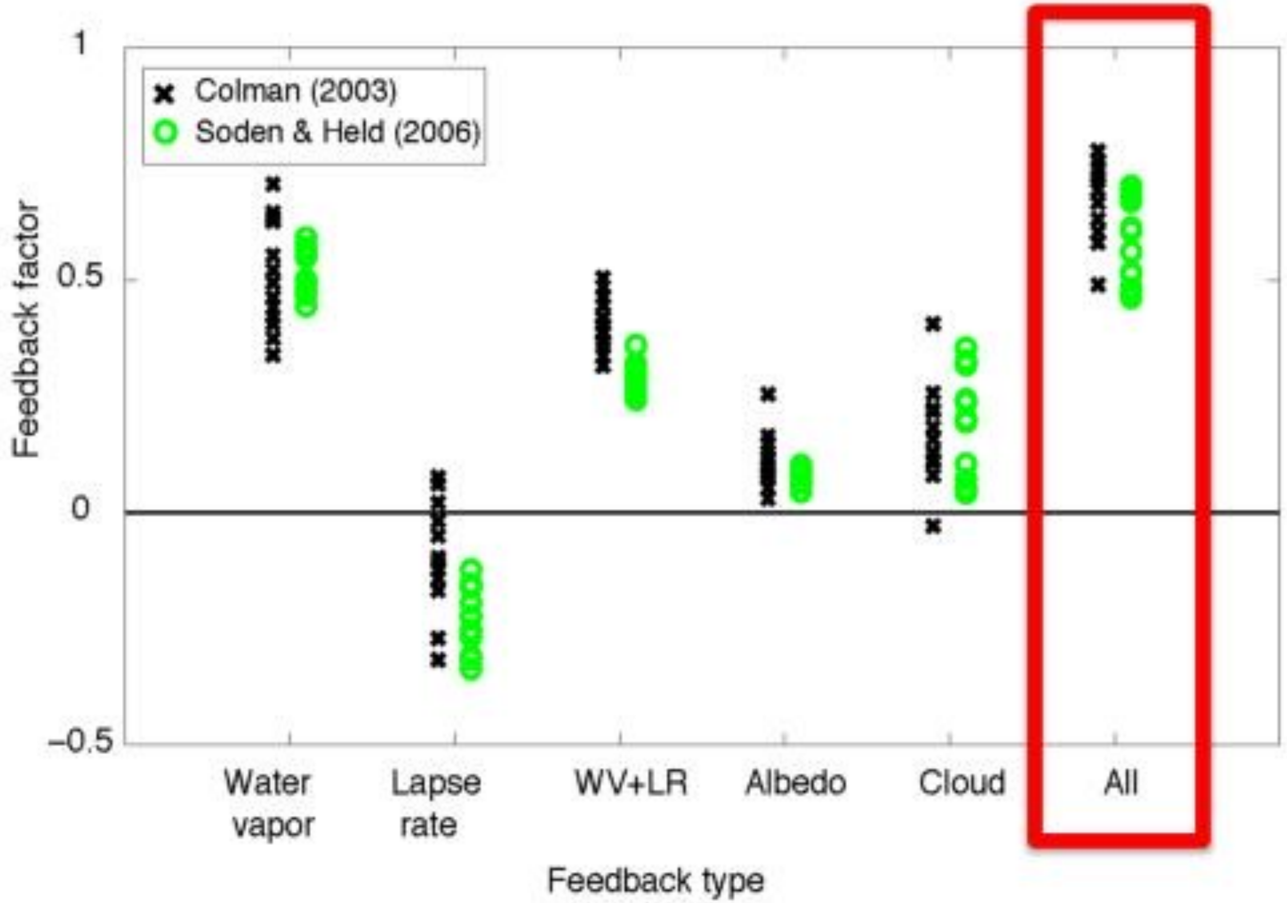


Gregory plot: regression of top of the atmosphere (TOA) radiative imbalance (sum of incoming shortwave and incoming and outgoing longwave radiation at top of atmosphere) versus change of global average surface air temperature(SAT)for(4xCO2-PI) (green)andfor(4xCO2_O3-PI) (red)basedonannualmeans.

Climate sensitivity.

Estimates from models.

$$\Delta T = \frac{\lambda_0 R_f}{1 - \sum f_i}$$



Individual feedbacks uncorrelated among models, so can be simply combined:

Soden & Held (2006):

$\bar{f} = 0.62; \sigma_f = 0.13$

Colman (2003):

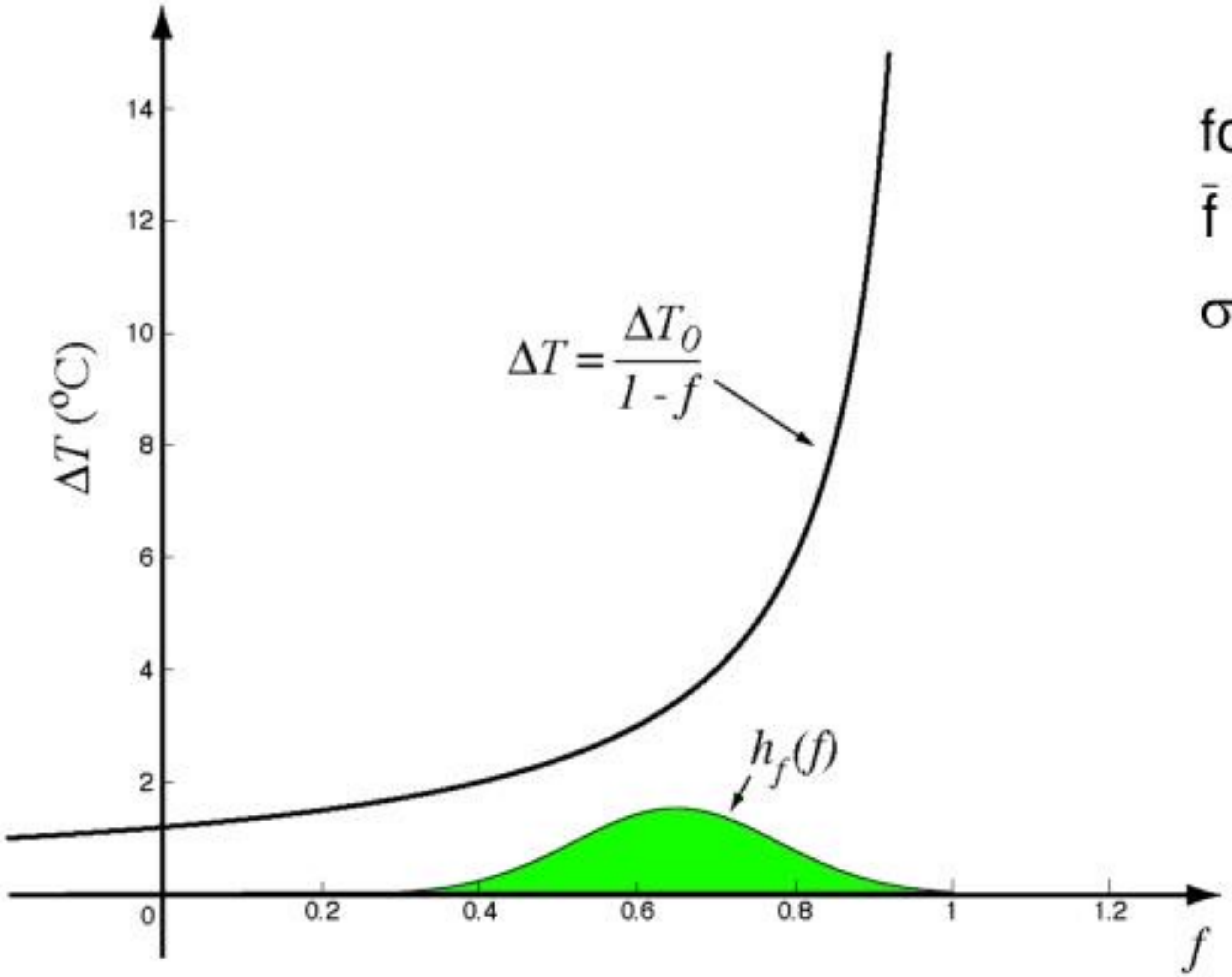
$\bar{f} = 0.70; \sigma_f = 0.14$

- How does this uncertainty in physics translate to uncertainty in climate sensitivity?

Climate sensitivity.

Estimates from models.

$$\boxed{\Delta T = \frac{\lambda_0 R_f}{1-\sum f_i}}$$



for:

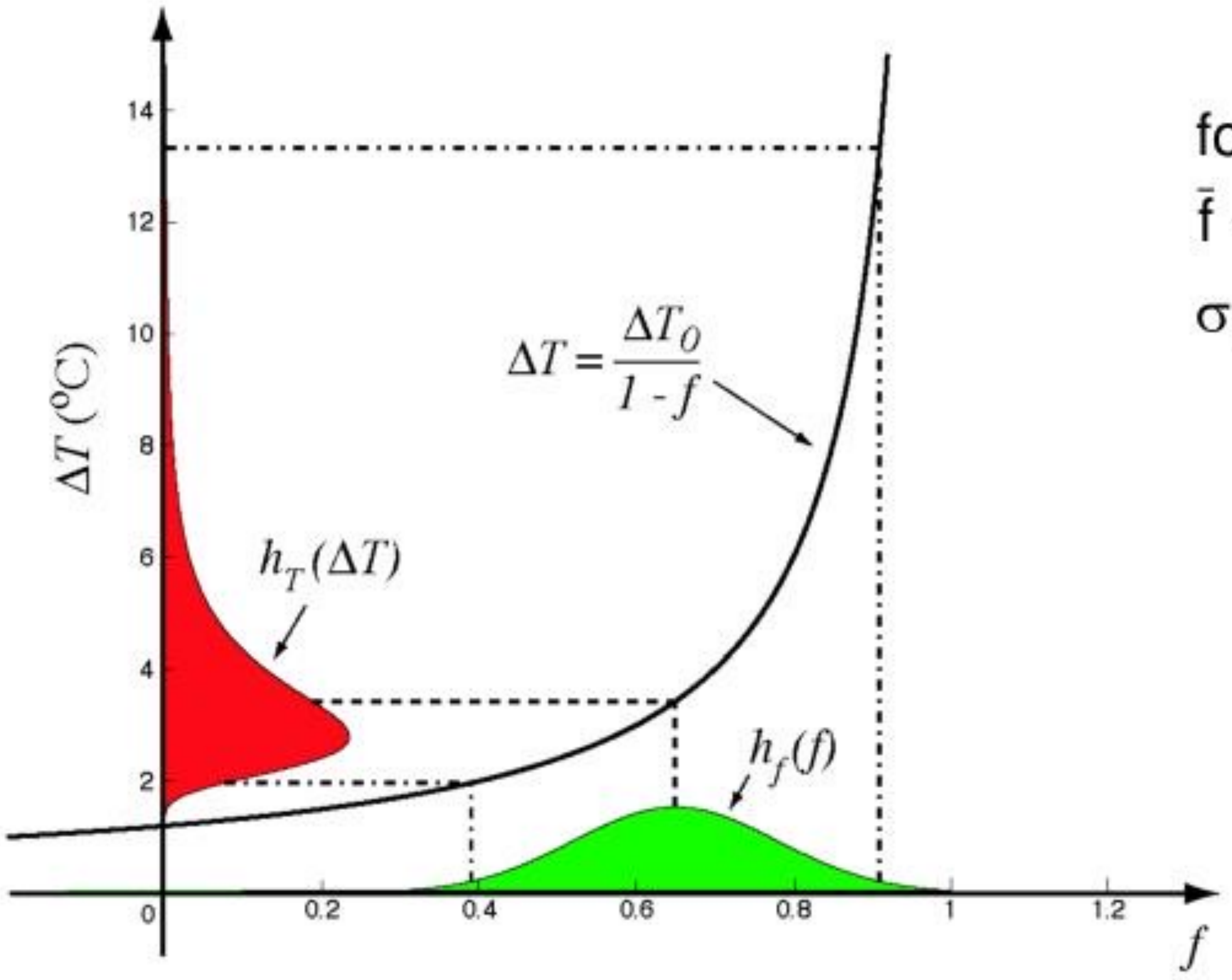
$\bar{f} = 0.65$

$\sigma_f = 0.14$

Climate sensitivity.

Estimates from models.

$$\Delta T = \frac{\lambda_0 R_f}{1 - \sum f_i}$$



for:

$\bar{f} = 0.65$

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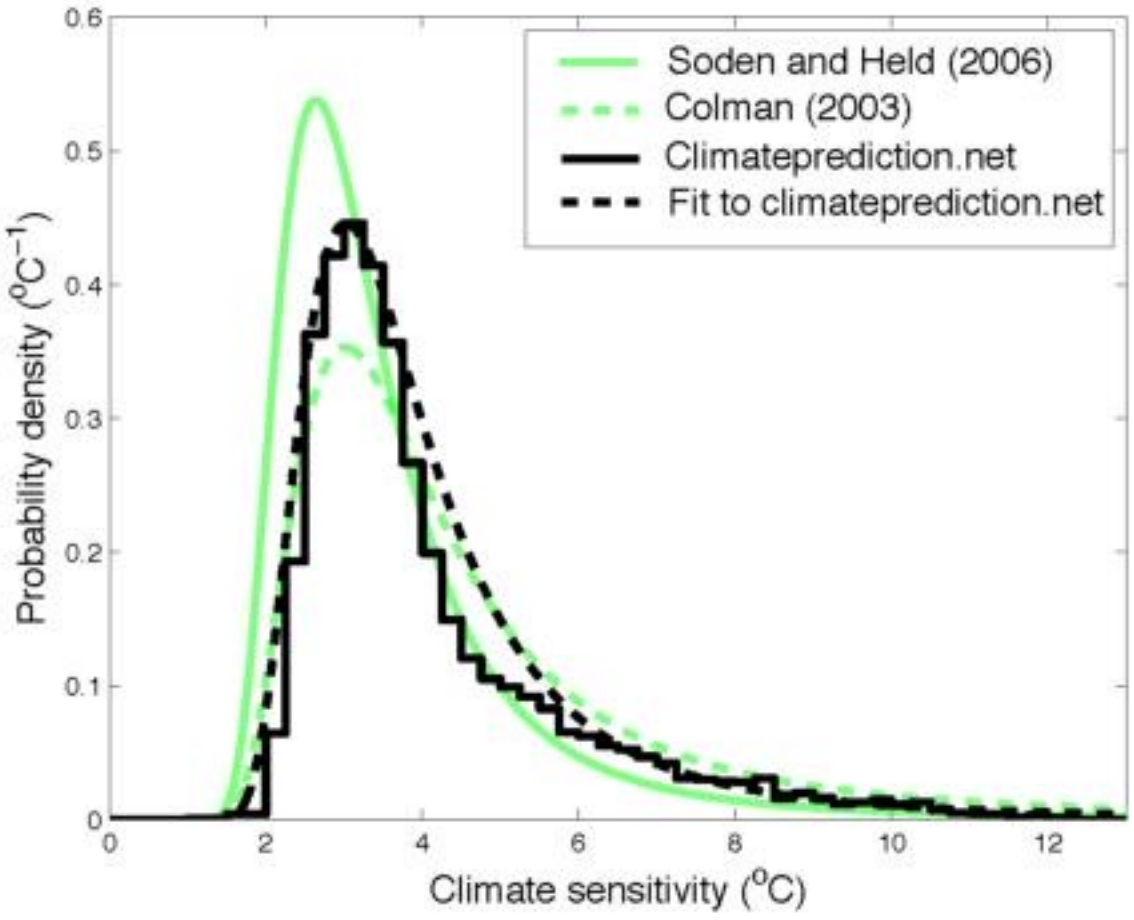
Climate sensitivity.

Estimates from models.

$$\Delta T = \frac{\lambda_0 R_f}{1 - \sum f_i}$$



- GCMs produce climate sensitivity consistent with the compounding effect of essentially-linear feedbacks.

An aside: nonlinearity of feedbacks

From basic analysis: $$\Delta R = \frac{dR}{dT}\Delta T + O(\Delta T^2)$$

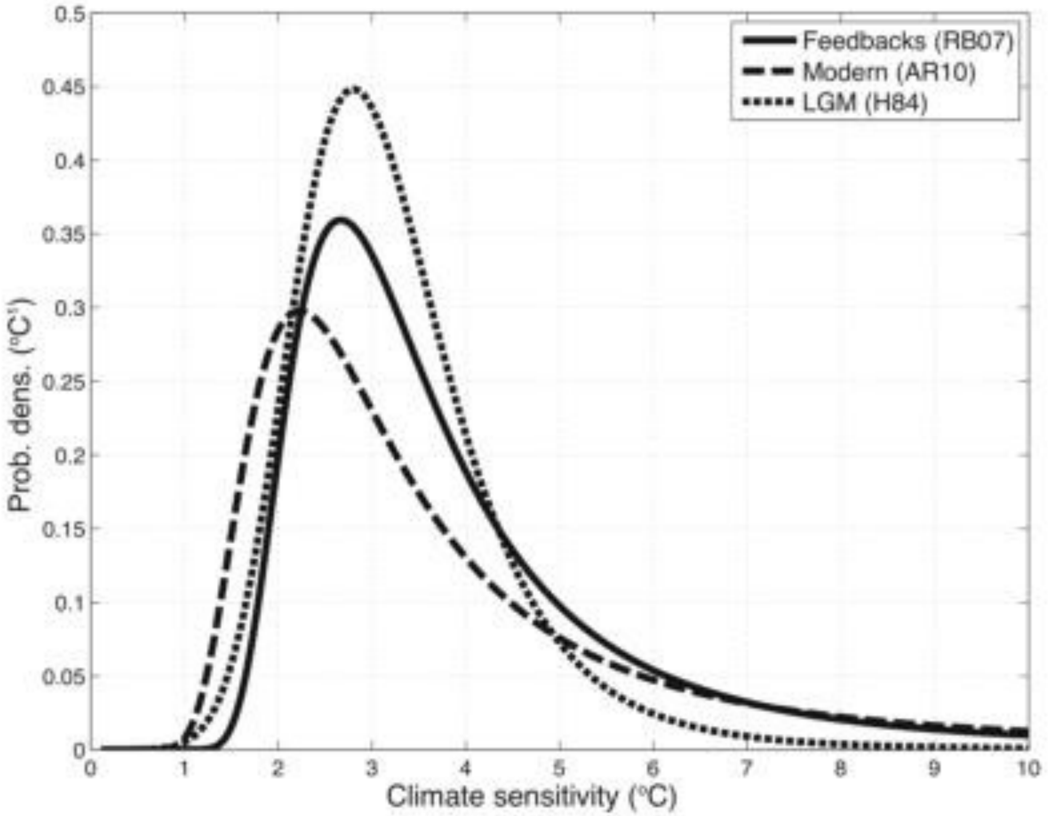
But can take quadratic terms… $$\Delta R = \frac{dR}{dT}\Delta T + \frac{1}{2}\frac{d^2R}{dT^2}\Delta T^2 + \cancel{O(\Delta T^3)}$$

giving… $$\boxed{G = \frac{1}{1 - f - \frac{\Delta T}{2}\frac{df}{dT}}}$$

Taking ~12 different studies: $$-0.04K^{-1} \le \frac{df}{dT} \le +0.04K^{-1}$$

Climate sensitivity.

Models and observations.



- All look pretty similar.
- How to do better?

Climate sensitivity.

How to do better?

1. Combine different estimates?

   Very hard to establish the degree of independence of individual estimates.

2. Use other observations?

   (e.g., NH vs. SH; pole-to-eq. ΔT; seasonality, trop. water vapor)
   Structural errors among models highly uncertain (see Knutti et al, 2010).

3. Transient climate response?

   Clim. Sens. is an equilibrium property, short observations only have limited resolving power.

Bottom-up approach

Understand and model key physical processes that affect climate sensitivity.
i.e. Feedback Processes

- Water vapor feedback
- Cloud feedback
- Ice-albedo feedback
- Many more