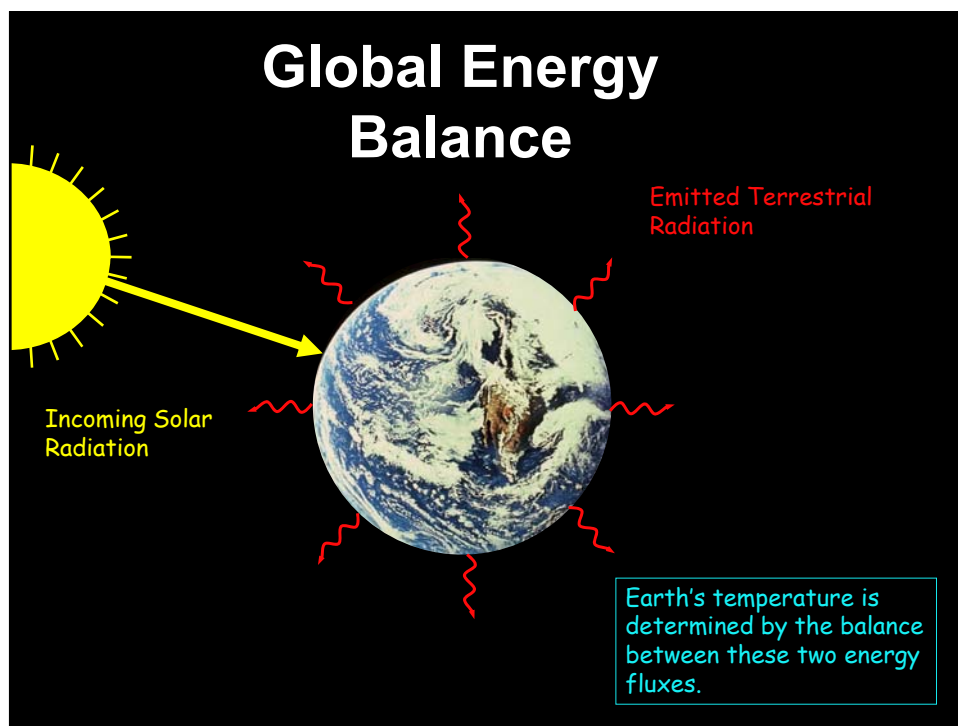# Estimating Climate Sensitivity From Past Climates

Outline

- Zero-dimensional model of climate system
- Climate sensitivity
- Climate feedbacks
- Forcings vs. feedbacks
- Paleocalibration vs. paleoclimate modeling
- Estimating past forcings
- Estimating past temperatures

Global Energy Balance

Earth's temperature is determined by the balance between these two energy fluxes.

Global Energy Balance

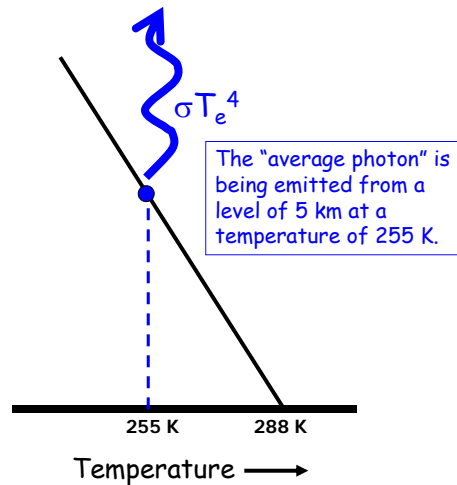| Incoming Solar Radiation	=	Outgoing Terrestrial Radiation

$$S_0 \cdot \pi r^2 \cdot (1-\alpha) \quad = \quad \sigma T^4 \cdot 4\pi r^2$$

solar irradiance — cross-sectional area of Earth — absorbed fraction — emitted infrared flux — surface area of Earth

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

Calculated temperature is 255 K, but actual global mean surface temperature is 288 K. Why is there is discrepancy?

The Greenhouse Effect—which is due to presence of water vapor, CO_2, CH_4, N_2O in the atmosphere. These gases absorb and re-emit infrared radiation, so photons emitted from the surface and lower atmosphere have a lower probability of escaping to space than do photons emitted from the upper atmosphere.

σT_e^4

The "average photon" is being emitted from a level of 5 km at a temperature of 255 K.

255 K   288 K

Temperature ⟶

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.
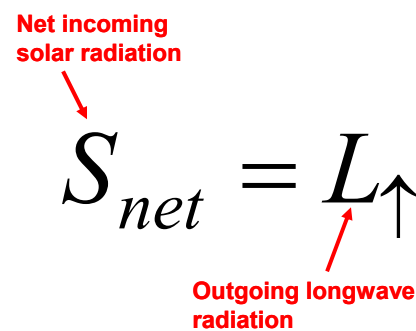
Zero-Dimensional Model of Earth's Climate

- Consider a very simple model of Earth's climate: a balance between incoming solar radiation and outgoing longwave radiation (i.e., thermal emission).

$$S_{net} = L_{\uparrow}$$

Zero-Dimensional Model of Earth's Climate

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Net incoming solar radiation

$$S_{net} = L_{\uparrow}$$

Outgoing longwave radiation

Zero-Dimensional Model of Earth's Climate

- We can write an expression for $F_{\uparrow}$ by assuming that Earth radiates as a blackbody and its temperature can be represented as a single value.

$$S_{net} = \sigma T^4$$

Zero-Dimensional Model of Earth's Climate

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Zero-Dimensional Model of Earth's Climate

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Temperature

Zero-Dimensional Model of Earth's Climate

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Net incoming solar radiation

Stefan-Boltzmann constant

$$S_{net} = \sigma T^4$$

Temperature

Differentiating with respect to T:

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

Rewriting in terms of dT/dS_{net}:

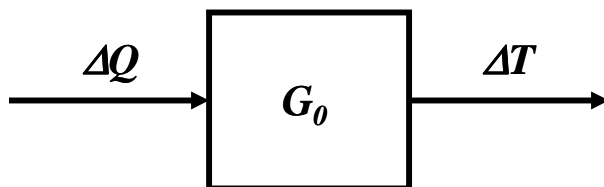$$\frac{dT}{dS_{net}} = \frac{1}{4\sigma T^3}$$

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

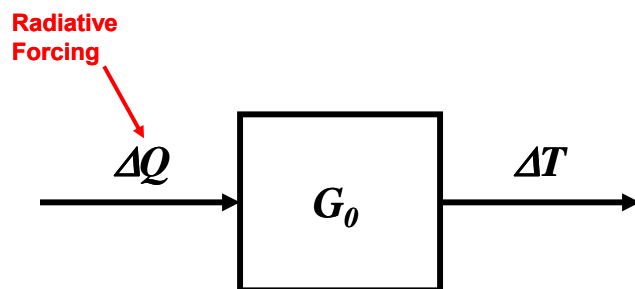$$\frac{\Delta T}{\Delta Q} = G_0 \equiv \frac{1}{4\sigma T^3}$$

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

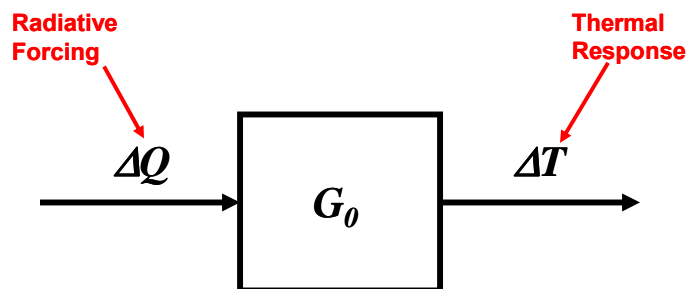$$G_0 = \frac{1}{(4)(5.67 \times 10^{-8})(255)^3} = 0.266 \; K \, W^{-1} \, m^2$$

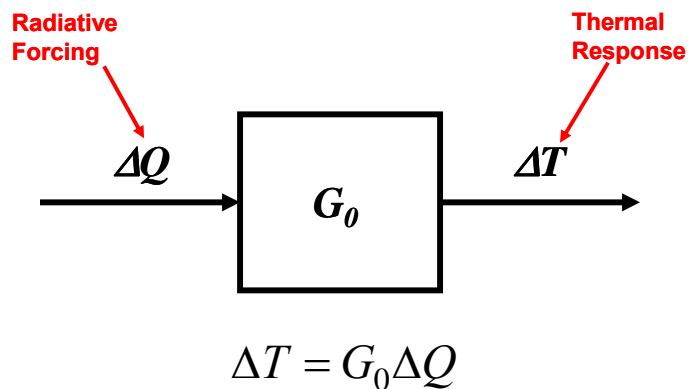## Schematic Diagram of Zero-Dimensional Climate Model

$\Delta Q \longrightarrow \boxed{G_0} \longrightarrow \Delta T$

Schematic Diagram of Zero-Dimensional Climate Model

Radiative Forcing

$\Delta Q \longrightarrow \boxed{G_0} \longrightarrow \Delta T$

Schematic Diagram of Zero-Dimensional Climate Model

Radiative Forcing

Thermal Response

ΔQ G_0 ΔT

Schematic Diagram of Zero-Dimensional Climate Model

Radiative Forcing

Thermal Response

ΔQ G_0 ΔT

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

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.

Radiative Feedbacks

- What would happen to each of these climate system properties if the global mean temperature were to increase?

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.
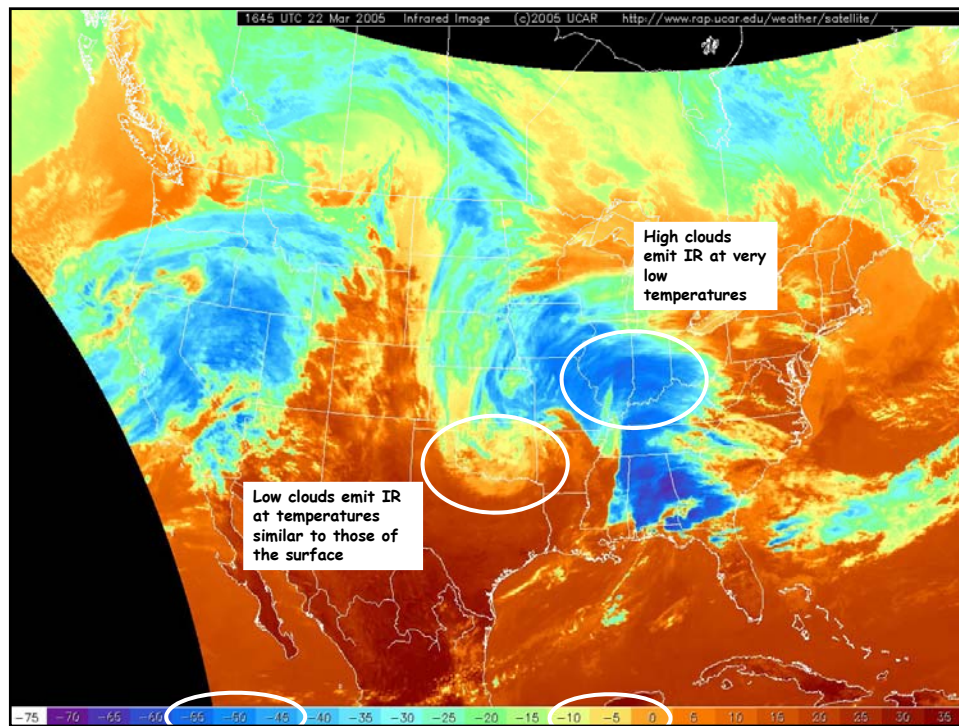- This is a positive feedback.

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.

Lapse Rate Feedback

- Moist convective processes control the vertical temperature distribution over much of the earth (i.e., tropics and much of summer hemisphere).
- The moist adiabatic lapse rate is smaller in a warmer climate, thus temperature changes in the upper troposphere are greater than those at the surface.
- Greater warming aloft increases the outgoing longwave radiation, thus cooling the atmosphere.
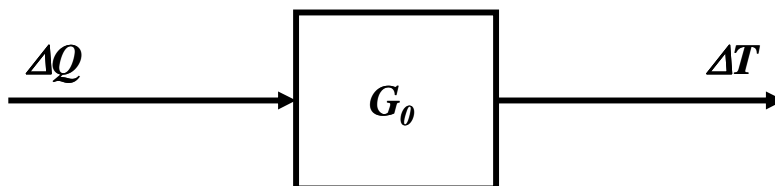- This is a negative feedback.

Cloud Feedback

- Low clouds and high clouds affect the earth's radiation balance differently.
- Both cloud types reflect solar radiation, but only high clouds decrease infrared emission.

High clouds emit IR at very low temperatures

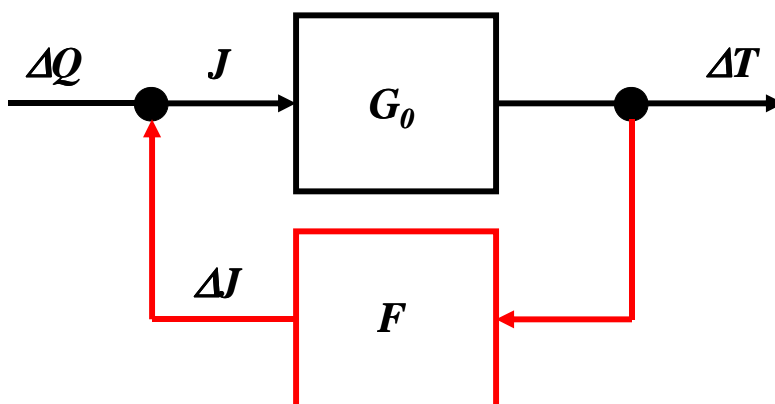Low clouds emit IR at temperatures similar to those of the surface

Cloud Feedback

- Low clouds and high clouds affect the earth's radiation balance differently.
- Both cloud types reflect solar radiation, but only high clouds decrease infrared emission.
- The net effect of low clouds is to cool the climate (reflect solar, but little effect on infrared).
- The net effect of high clouds is to warm the climate (reflect some solar, strongly decrease infrared emission).
- Sign of cloud feedback is uncertain because there is no simple relationship between cloud cover and global temperature and because of the interplay between the effects of high and low clouds.
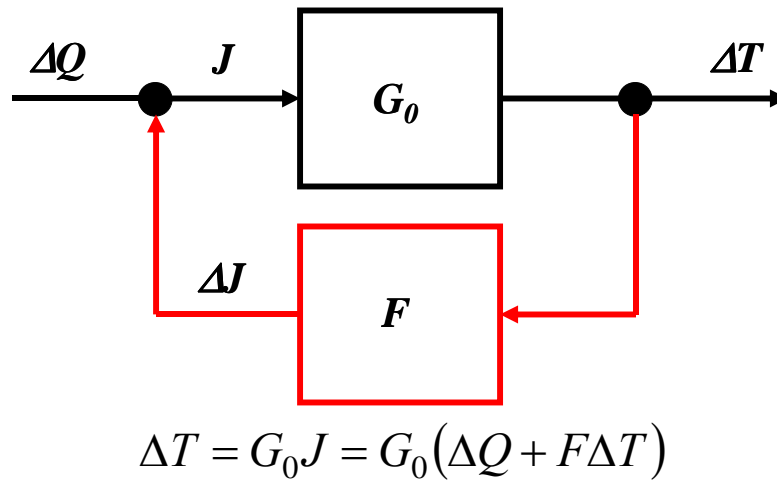
Zero-Dimensional Climate Model With Feedbacks

$\Delta Q \longrightarrow \boxed{G_0} \longrightarrow \Delta T$

Zero-Dimensional Climate Model With Feedbacks

$\Delta Q \longrightarrow J \longrightarrow \boxed{G_0} \longrightarrow \Delta T$

$\Delta J \longleftarrow \boxed{F} \longleftarrow$

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

$$\Delta T = G_f \Delta Q$$

Very often, λ is used in place of G_f:

$$\Delta T = \lambda \Delta Q \rightarrow \lambda = \frac{\Delta T}{\Delta Q}$$

Less often, a different nomenclature is used:

$$\lambda = \frac{\Delta Q}{\Delta T}$$
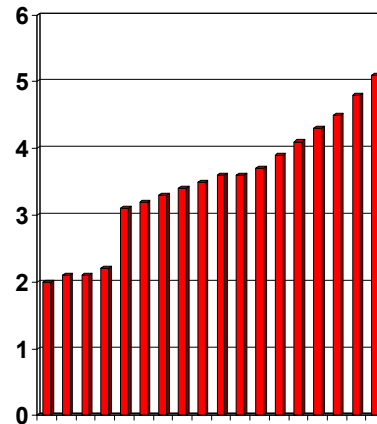
(This can be very confusing at times!)

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.
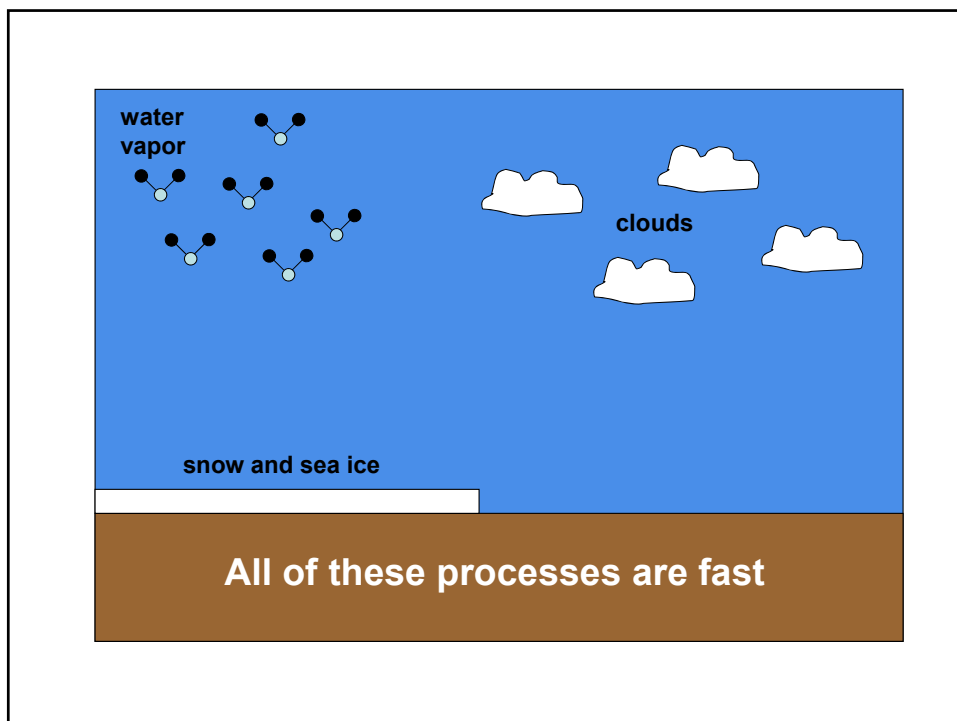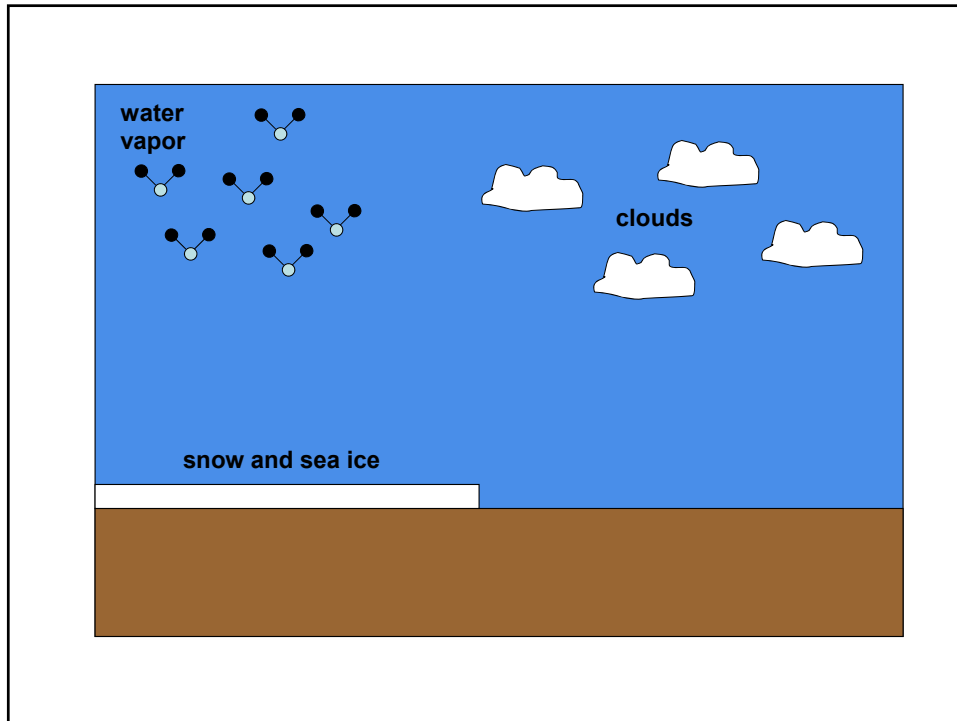

Forcings vs. Feedbacks

- 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 → time scales of years to decades
- Slow → time scales of centuries or longer

water
vapor

clouds

snow and sea ice

water
vapor

clouds

snow and sea ice

All of these processes are fast
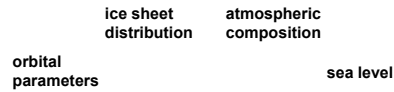
Radiative Feedbacks Involving Slow Processes

- 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)

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.
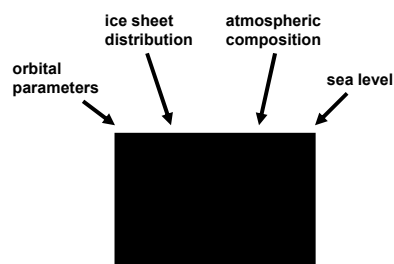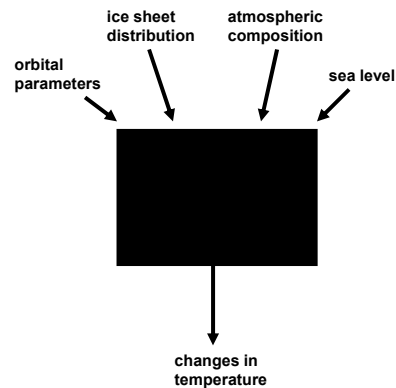
Evaluating Climate Sensitivity

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ice sheet distribution   **atmospheric composition**
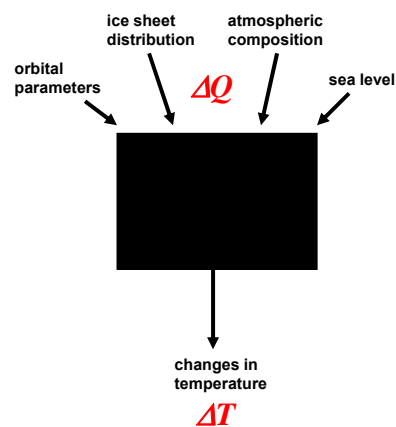
orbital parameters   **sea level**

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ice sheet distribution   **atmospheric composition**

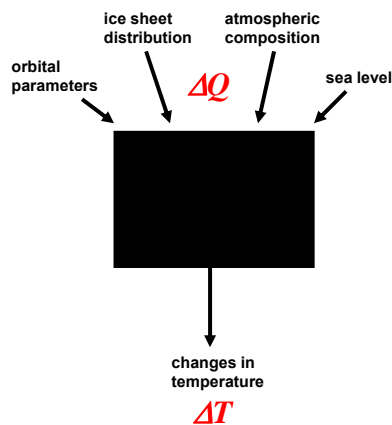**orbital parameters**   **sea level**

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orbital parameters

ice sheet distribution

atmospheric composition

sea level

changes in temperature

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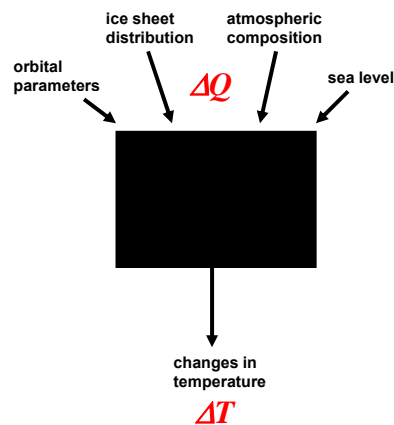
orbital parameters

ice sheet distribution

atmospheric composition

sea level

ΔQ

changes in temperature

ΔT

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 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.

ice sheet distribution   atmospheric composition

orbital parameters ΔQ sea level

changes in temperature

ΔT

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.

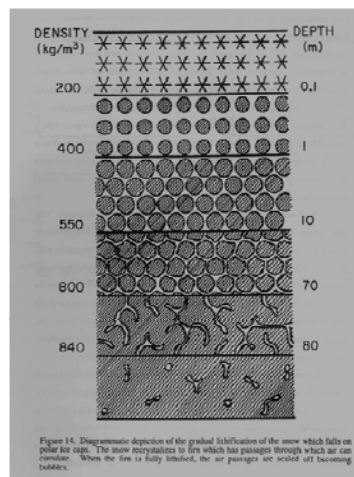- Results of such calculations are widely available.

Estimating Forcings:
Ice Sheets and Sea Level

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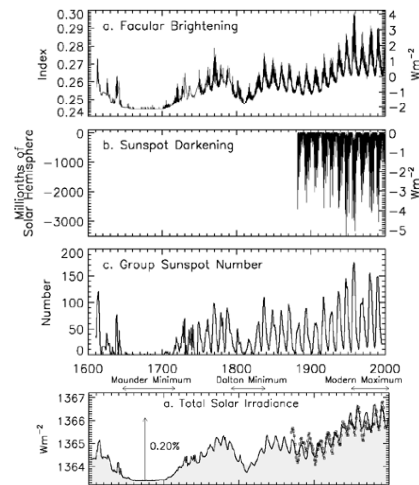- Geophysical modeling (e.g., Peltier) of 3-D ice sheet distribution.

Estimating Forcings:
Atmospheric Composition

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

Estimating Forcings: Solar Irradiance

- Proxies of solar geomagnetic activity are used as evidence of solar irradiance.
- Some questions about validity of relationships on long (i.e., > 11 years) time scales.

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
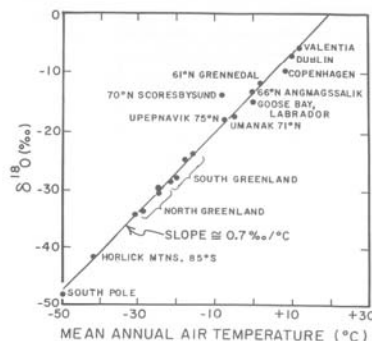- Noble gases in aquifers

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)
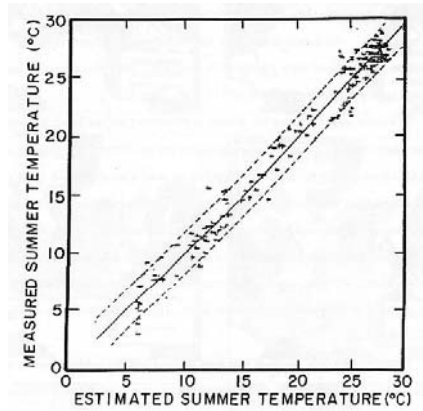
Distributions of Marine Microorganisms

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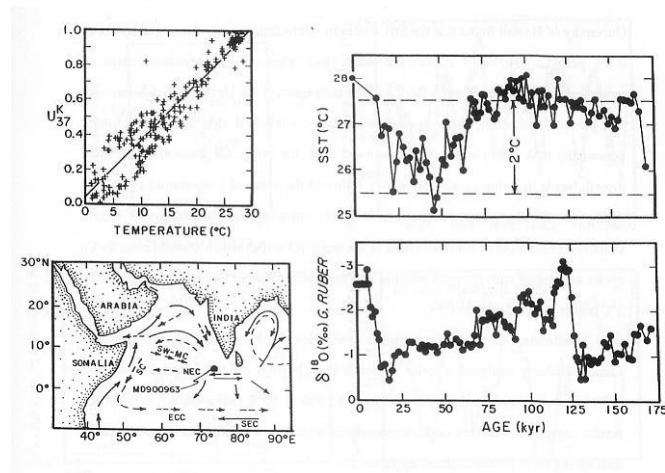
Distributions of Marine Microorganisms

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

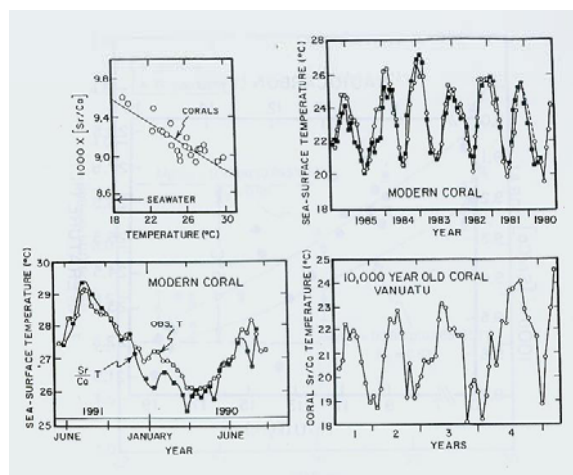# Alkenone Molecules in Marine Flora

- 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.
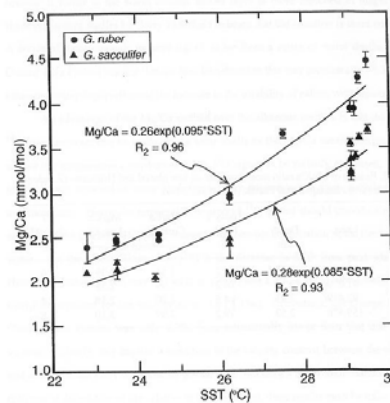
Alkenone Molecules in Marine Flora

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.

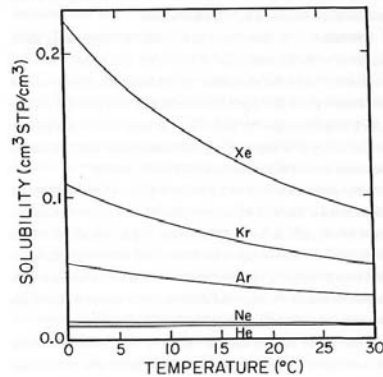- Mg/Ca and ^{18}O can be determined from the same shells.

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

clouds

snow cover and
sea ice decreases

Effect of increasing greenhouse gases