METR 5223: Atmospheric Radiation

GCM Sensitivity Analysis

1:

Lecture for Spring 2009
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2:

old:

GCM = General Circulation Model

new:

GCM = Global Climate Model

3:

Some jargon: *Climate sensitivity* usually means a derivative of equilibrium temperature:

$$\frac{dT}{dS_0}, \quad \frac{dT}{dc}, \quad \frac{\Delta T}{\Delta c} \quad \ldots$$

Or just ΔT with Δc implied to be a doubling of CO_2.

Note T is usually globally averaged surface air temperature.

4:

In the 1980s and 1990s, GCMS were often NOT used with variable CO_2 included in the radiative transfer scheme, and flux balance at the top of the atmosphere did NOT occur. The GCMs run with fixed SST and a perpetual season. A general *climate sensitivity* was deduced:

$$\lambda \equiv \frac{\Delta T}{\Delta R}$$

ΔT is the change in equlbrium temperature that would restore flux balance at the top of the atmosphere after an event caused a net inward radiative flux density ΔR at the top of the atmosphere. ΔT is the response that essentially removes the radiative forcing and restores radiative equilibrium. ΔR is called the *radiative forcing*.

5: Examples of agents of radiative forcing could be changes in the concentration of volcanic aerosols a or changes in concentration of CO_2 c. These agents are NOT added to the model.

Let Q be the net flux of absorbed radiation, for example $Q = S_0(1 - \alpha_p)/4$.

Let F be the flux of outward IR radiation.

6: In the global budget or global average,

$$0 = Q - F$$

With changes in a and/or c,

$$0 = \frac{dQ}{dT}dT - \frac{dF}{dT}dT + \frac{dQ}{da}da - \frac{dF}{dc}dc$$

Or

$$0 = \frac{dQ}{dT}dT - \frac{dF}{dT}dT + dR$$

where

$$dR = \frac{dQ}{da}da - \frac{dF}{dc}dc$$

7: Clever how GCMs do double CO_2 experiments. They first seek the *climate sensitivity*:

$$\frac{dT}{dR} = \frac{1}{\frac{dF}{dT} - \frac{dQ}{dT}} \equiv \lambda$$

Or seek

$$\lambda = \frac{1}{\frac{\Delta F}{\Delta T} - \frac{\Delta Q}{\Delta T}}$$

λ can be determined with ΔT imposed in fixed SST experiments. *Flux balance does not need to exist at the top of the atmosphere in the experiments.*

8: After λ is known, equilibrium prediction can be made with

$$\Delta T = \lambda \Delta R$$

where ΔR is known from external radiative transfer calculation.

For example, ΔR for CO_2 doubling is quoted by IPCC as 3.7 W m^{-2}.

9:

With feedbacks,

$$\lambda = \frac{1}{\frac{dF}{dT} - \frac{dQ}{dT}}$$

Without feedbacks,

$$\lambda_0 = \frac{1}{\frac{\partial F}{\partial T}}$$

$\frac{\partial F}{\partial T}$ can be approximated to be

$$\frac{\partial F}{\partial T_e} = 4\sigma T_e^3 = 3.76 \text{ W m}^{-2} \text{ K}^{-1}$$

Or, with T being the surface temperture T_s, a better estimate is

$$\frac{\partial F}{\partial T_s} = \frac{\partial F}{\partial T_e}\frac{dT_e}{dT_s} = 3.76 \text{ W m}^{-2} \text{ K}^{-1} \times \frac{T_e}{T_s} = 3.33 \text{ W m}^{-2} \text{ K}^{-1}$$

10:

$$\lambda_0 = \frac{1}{\frac{\partial F}{\partial T}} = .30 \text{ K W}^{-1} \text{ m}^2$$

ΔR for CO_2 doubling as quoted by IPCC is 3.7 W m^{-2}. So with *no feedback*, the change in global average temperature would be

$$\Delta T_{nf} = \lambda_0 \Delta R = .30 \frac{\text{K}}{\text{W m}^{-2}} \times 3.7 \text{ W m}^{-2} = 1.1 \text{ K}$$

(See `http://www.grida.no/climate/ipcc_tar/wg1/21`.htm9 for ΔR)

11:

Example from ECMWF (Cess,1990):

$$\frac{dF}{dT} = 2.46 \text{ W m}^{-2} \text{ K}^{-1} \qquad \frac{dQ}{dT} = 0.74 \text{ W m}^{-2}\text{K}^{-1}$$

$$\lambda = \frac{1}{\frac{dF}{dT} - \frac{dQ}{dT}} = \frac{1}{1.72 \text{ Wm}^{-2} \text{ K}^{-1}} = .58 \frac{\text{K}}{\text{Wm}^{-2}}$$

The ratio of the ΔT with feedback to the increase with no feedback is the *gain*:

$$\frac{\Delta T}{\Delta T_{nf}} = \frac{\lambda \Delta R}{\lambda_0 \Delta R} = \frac{\lambda}{\lambda_0} = \frac{.58}{.30} = 1.93$$

GCMs commonly have gain of order 2 or more, and so predict a ΔT of 2 K or more for a doubling of CO_2.

12:

Diagnosing the feedback. Write

$$\frac{dF}{dT} = \frac{\partial F}{\partial T} + \sum_i \frac{\partial F}{\partial I_i}\frac{dI_i}{dT}$$

where I_i is the amount of the i'th *thing* that effects F.

For example I_1 could be tons of water vapor, I_2 could be the average lapse rate.

We can easily find the sum of the feedback terms:

$$\sum_i \frac{\partial F}{\partial I_i}\frac{dI_i}{dT} = \frac{dF}{dT} - \frac{\partial F}{\partial T}$$

13:

In ECMWF,

$$\frac{dF}{dT} = 2.46\ \mathrm{Wm^{-2}\ K^{-1}} \qquad \frac{\partial F}{\partial T} = 4.14\ \mathrm{W\ m^{-2}\ K^{-1}}$$

(Note $\frac{\partial F}{\partial T}$ is different from the 3.33 $\mathrm{W\ m^{-2}K^{-1}}$ estimate earlier .) For ECMWF

$$\sum_i \frac{\partial F}{\partial I_i}\frac{dI_i}{dT} = -1.68\ \mathrm{W\ m^{-2}\ K^{-1}}$$

This negative number implies positive feedback:

$$\lambda = \frac{1}{\frac{\partial F}{\partial T} + \sum_i \frac{\partial F}{\partial I_i}\frac{dI_i}{dT} - \frac{dQ}{dT}}$$

$$= \frac{1}{(4.14 - 1.68 - .74)\ \mathrm{W\ m^{-2}\ K^{-1}}} = .58\frac{\mathrm{K}}{\mathrm{Wm^{-2}}}$$

14:

Note that the gain is

$$\frac{\lambda}{\lambda_0} = \frac{4.14}{4.14 - 1.68 - .74} = \frac{1}{1 - .406 - .129}$$

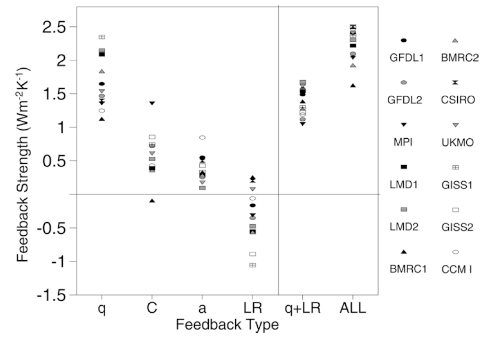
Both negative terms in the denominator make it smaller, and are positive feedback feedback factors, in this case .406 for water vapor feedback and .129 for albedo feedback.

We sometimes see gain written as:

$$g = \frac{1}{1 - \sum_i f_i}$$

where the f_i are feedback factors.

15:



Feedbacks are: q water vapor, C clouds, a albedo, LR lapse rate (Colman, 2003).