METR 5223: Atmospheric Radiation

# GCM Sensitivity Analysis

Lecture for Spring 2009

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old:

GCM = General Circulation Model

new:

GCM = Global Climate Model

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

$$\frac{dT}{dS_0}, \qquad \frac{dT}{dc}, \qquad \frac{\Delta T}{\Delta c} \qquad \cdots$$

Or just $\Delta T$ with $\Delta c$ implied to be a doubling of $CO_2$.

Note $T$ is usually globally averaged surface air temperature.

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:

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

$\Delta T$ is the change in equilbrium temperature that would restore flux balance at the top of the atmosphere after an event caused a net inward radiative flux density $\Delta R$ at the top of the atmosphere. $\Delta T$ is the response that essentially removes the radiative forcing and restores radiative equilibrium. $\Delta R$ is called the *radiative forcing*.

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.

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

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

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

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

$\lambda$ can be determined with $\Delta T$ imposed in fixed SST experiments. *Flux balance does not need to exist at the top of the atmosphere in the experiments.*

After $\lambda$ is known, equilibrium prediction can be made with

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

where $\Delta R$ is known from external radiative transfer calculation.

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

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

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

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

$\Delta R$ 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 $\Delta R$ )

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 $\Delta T$ 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 $\Delta T$ of 2 K or more for a doubling of $CO_2$.

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:

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

In ECMWF,

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

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

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

This negative number implies positive feedback:

$$\begin{aligned} \lambda &= \frac{1}{\frac{\partial F}{\partial T} + \sum_i \frac{\partial F}{\partial I_i}\frac{dI_i}{dT} - \frac{dQ}{dT}} \\ &= \frac{1}{(4.14 - 1.68 - .74) \text{ W m}^{-2} \text{ K}^{-1}} = .58 \frac{\text{K}}{\text{Wm}^{-2}} \end{aligned}$$

Note that the gain is

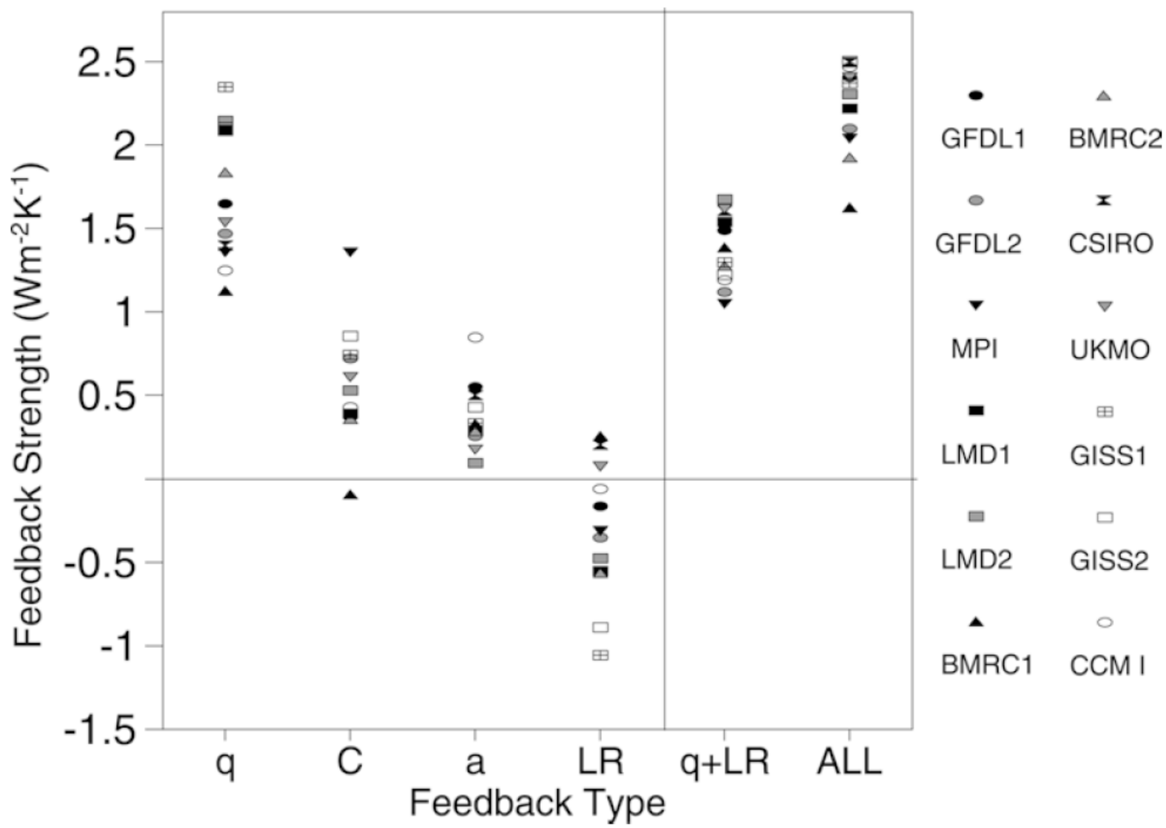
$$\frac{\lambda}{\lambda_0} = \frac{4.14}{4.14 - 1.68 - .74} = \frac{1}{1 - .406 - .129}$$

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.




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