Equilibrium Climate Sensitivity and Feedbacks

Consider a simple radiation energy balance of the Earth: the amount of energy the planet receives from the sun (intercepted by the planet's disk) equals the loss of energy by the thermal infrared (IR) radiation to space (emitted in all directions):

$$(1-\alpha)S\pi R^2 = \varepsilon\sigma T_s^4 4\pi R^2$$

or

$$\frac{1}{4}(1-\alpha)S = \varepsilon\sigma T_s^4 \qquad (1)$$

where α=0.3 is the Earth's shortwave albedo, S=1368 W/m^2 solar constant, ε bulk emissivity of the Earth, T_s mean surface temperature, and σ=5.67x10-8 Wm^{-2}/K^{-4} Stefan-Boltzmann constant. If the atmosphere does not absorb thermal radiation, then ε=1, and the surface temperature would be equal to the emission temperature of the Earth (assuming blackbody emission):

$$T_e = \left[\frac{(1-\alpha)S}{4\sigma}\right]^{1/4} = 255K \qquad (2)$$

The emission temperature is the temperature that a satellite would measure using broadband (integral over all wavelengths) measurements of outgoing radiation. Note that it only depends on the Earth albedo, and does not depend on any of the 'greenhouse effects', that is even if the amount of CO_2 in the Earth atmosphere was 10 times as large as today but Earth had the same albedo, it would not be detectable from broadband measurements from space as the emission temperature would be the same.

For the given albedo, εT_s^4=const, so for small climate perturbation (but still in equilibrium), the perturbation of emissivity and surface temperature are related as

$$\Delta T_s = -\frac{1}{4}T_s\frac{\Delta\varepsilon}{\varepsilon} = -\frac{1}{4}T_s\frac{\Delta\varepsilon\sigma T_s^4}{\varepsilon\sigma T_s^4} = \frac{1}{4}\frac{T_s}{T_e}F$$

where $F = -\Delta\varepsilon\sigma T_s^4$ is some forcing that is causing the climate change through the change in the planet's IR emissivity. Note that the forcing F is computed using undisturbed (present) surface temperature. From radiative transfer computations using the observed temperature and greenhouse gases (GHG) profiles, the forcing due to doubling the CO_2 is about 4 W/m^2 (the bulk Earth's emissivity decreases as atmosphere becomes more opaque to IR radiation). From observations, the mean surface temperature of the Earth is about 15^{o}C=288K, so for doubling CO_2 and the absence of all other changes, we would expect the surface temperature to warm by $\left(\Delta T_s\right)_{2xCO_2} \approx 1.2K$.

Besides the forcing F, the surface temperature T_s can be a function of some other climate-system parameters x_i, such as albedo, water vapor amount, cloud amount, etc, $T_s = f(F, x_i)$. If we assume that the parameters that affect the emissivity and hence T_s are independent of each other (which generally may not be true; for example, amount of water vapor can affect cloud amount), then, formally we can write for small climate perturbations using the differentiation rule

$$\Delta T_s = \frac{\partial T_s}{\partial F}F + \sum_i \frac{\partial T_s}{\partial x_i}\Delta x_i$$

Dividing by forcing F we get

$$\frac{\Delta T_s}{F} = \frac{\partial T_s}{\partial F} + \sum_i \frac{\partial T_s}{\partial x_i}\frac{\Delta x_i}{F} = \frac{\partial T_s}{\partial F} + \frac{\Delta T_s}{F}\sum_i \frac{\partial T_s}{\partial x_i}\frac{\Delta x_i}{\Delta T_s} \qquad (3)$$

The ratio of the realized surface temperature change over forcing that caused that change is called the *equilibrium* climate sensitivity

$$\lambda = \frac{\Delta T_s}{F} \qquad (4)$$

From (3) we get

$$\lambda = \frac{\lambda_0}{1 - \sum_i f_i} \qquad (5)$$

where $\lambda_0 = \frac{\partial T_s}{\partial F}$ is the base climate sensitivity to the forcing in the absence of feedbacks $f_i = \frac{\partial T_s}{\partial x_i}\frac{\Delta x_i}{\Delta T_s}$. For doubling-$CO_2$ forcing, λ_0=1.2/4=0.3 $KW^{-1}m^2$. The feedbacks can be positive or negative. If the net feedback of the climate system to the forcing $f = \sum_i f_i$ is positive, then the climate sensitivity is higher than the base, and vise versa. We expect the feedbacks to be relatively small for small perturbations of climate. However, unlike the base sensitivity, the feedbacks are mainly derived from the climate models. Let's rewrite (5) as the following

$$\frac{1}{\lambda} = \frac{1}{\lambda_0} - \sum_i p_i \qquad (6)$$

where p_i are feedback parameters. The values of these parameters can be estimated from the figure below taken from 2007 IPCC report. One can see that there are four major feedback parameters of the climate system: water vapor, cloud, albedo, and lapse-rate.

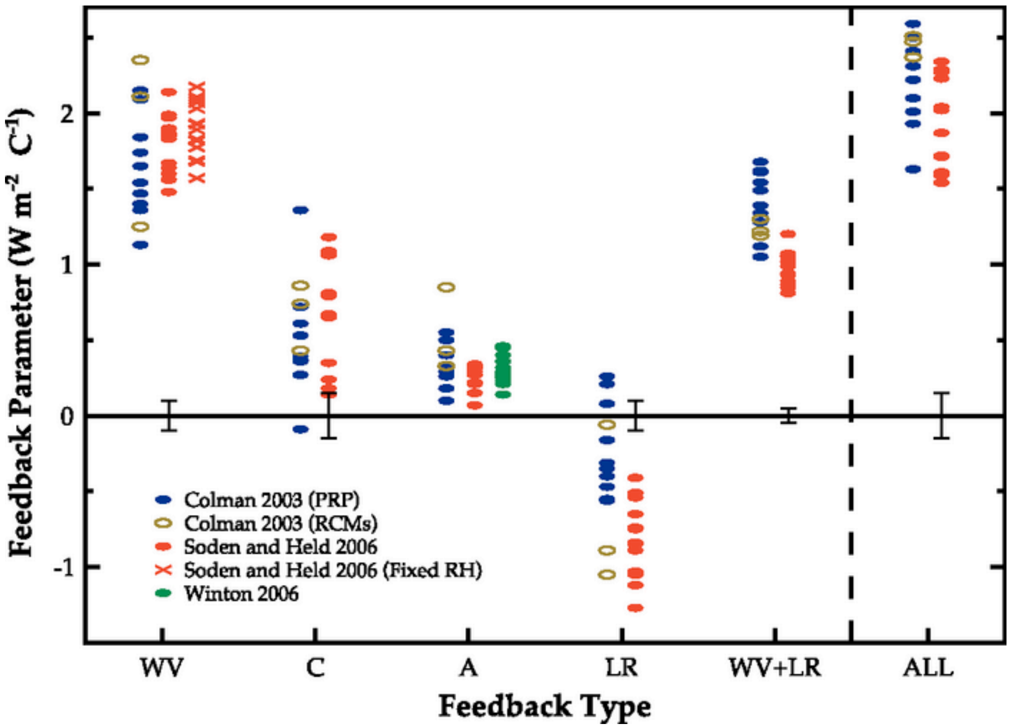


Figure 8.14. Comparison of GCM climate feedback parameters for water vapour (WV), cloud (C), surface albedo (A), lapse rate (LR) and the combined water vapour plus lapse rate (WV + LR) in units of W m^{-2} °C^{-1}. 'ALL' represents the sum of all feedbacks. Results are taken from Colman (2003a; blue, black), Soden and Held (2006; red) and Winton (2006a; green). Closed blue and open black symbols from Colman (2003a) represent calculations determined using the partial radiative perturbation (PRP) and the radiative-convective method (RCM) approaches respectively. Crosses represent the water vapour feedback computed for each model from Soden and Held (2006) assuming no change in relative humidity. Vertical bars depict the estimated uncertainty in the calculation of the feedbacks from Soden and Held (2006).

Water vapor feedback (WV) is strongly positive due to increased greenhouse effect of water vapor, which increases in warmer troposphere. The cloud feedback (C) is also positive as so is the albedo feedback (A). The latter makes sense as one would expect less ice and snow cover on warming planet, which would decrease albedo and hence contribute to further warming. The lapse rate (LR) feedback is negative and is the consequence of less steep temperature lapse rate in warmer world. This is because the temperature lapse-rate closely follows the moist-adiabatic lapse-rate, especially in tropical regions. The magnitude of the moist-adiabatic lapse-rate decreases as temperature increases, which would make the upper troposphere relatively warmer than at present. The upper troposphere is important for the IR emission to space, so that the warmer conditions mean higher emission, which would tend to cool the climate system, hence the negative feedback. As the water vapor is the major absorber and emitter of IR, the WV and LR feedback are not entirely independent and usually considered together.

From the figure we estimate that p_{WV}~1.75, p_C~0.75, p_A~0.25, and p_{LR}~-0.75 Wm^{-2}K^{-1}, so the total feedback is 2 Wm^{-2}K^{-1}. Substituting these values into (6), we obtain the climate sensitivity of the Earth climate system based on the consensus estimate from the climate models λ=0.75 KW^{-1}m^2, so for doubling CO2, the expected change of the mean surface temperature is 0.75 x 4 = 3 K, which is more than twice as large as 1.2 K due to direct effect of CO_2 doubling. The feedbacks are indeed important!

Cloud Feedback

Cloud feedback is generally regarded as the most uncertain. One of the reasons is that the climate models generally do not explicitly resolve clouds because of relatively coarse grid spacing, which are typically in the order of a hundred kilometers. Therefore, most of feedback estimates are from models that use cloud statistics derived from cloud parameterizations using grid-scale information.

Let's assume the balance of the top-of-atmosphere (TOA) outgoing longwave L and incoming shortwave S radiation fluxes: $L = S$. Due to some forcing F, a new TOA balance is reached. A perturbation to the TOA fluxes would balance the forcing as

$$\Delta L - \Delta S = F \quad (7)$$

The effect of clouds on TOA fluxes can be characterized by the so-called cloud radiative forcing, CRF, which sometimes (more correctly) also called cloud radiative effect. The CRF is defined separately for longwave and shortwave radiation. The longwave CRF is defined as

$$LWCF = L_c - L, \quad (8)$$

while shortwave CRF is defined as

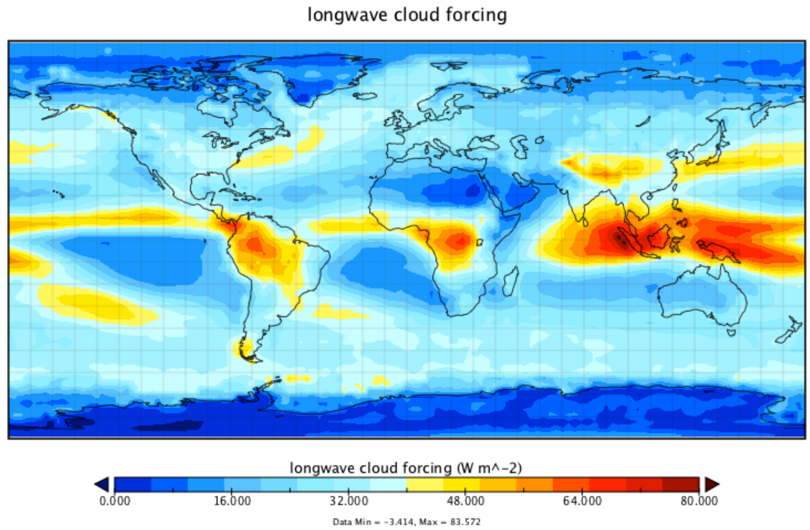
$$SWCF = S - S_c. \quad (9)$$

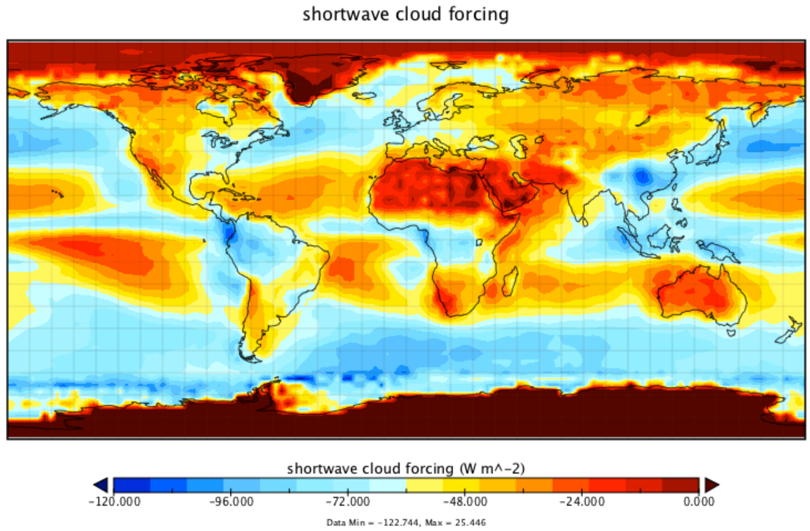
The fluxes with subscript 'c' denote the clear-sky fluxes, that is the fluxes at TOA with the clouds removed while all the other thermodynamic profiles kept unchanged.

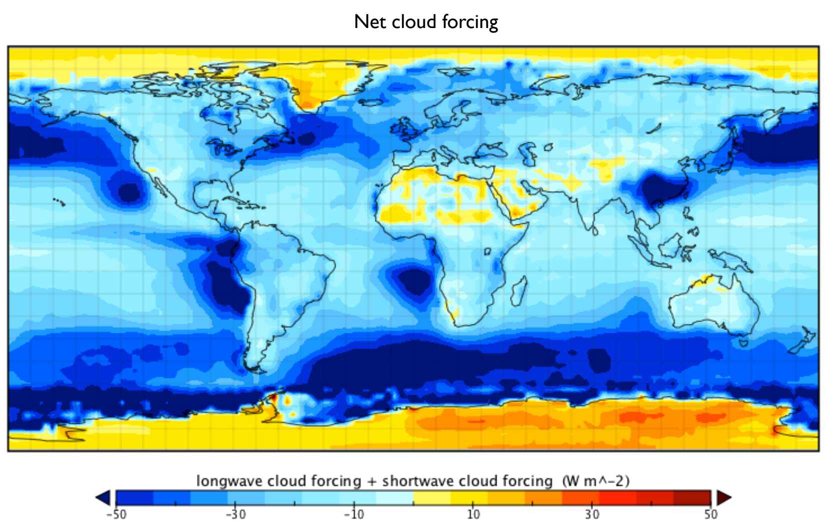
The LWCF is generally positive as clouds block the high IR from the surface and also by emitting low IR fluxes to space from their cold tops. Thus, in IR, clouds have warming effect on the surface temperature. The SWCF, on the other hand, is generally negative, as clouds have relatively high albedo compared to the surface, and reflect some of the incoming solar radiation. Hence, clouds tend to cool the surface in shortwave (solar) part of radiation. Overall, the net CRF is negative, that is clouds have cooling effect on climate. The annual mean LWCF, SWCF, and net CRF from the ERBE satellite measurements are shown in the figures below.

Although the clouds have net cooling effect, it is not clear if the effect would increase or decrease as the result of warming due to CO_2 forcing. The climate sensitivity as the measure of the climate change can be affected by the CRF feedback. The change of CRF due to the forcing is given by

$$\Delta \mathrm{CRF} = = \Delta L_c - \Delta L + \Delta S - \Delta S_c = -(\Delta L - \Delta S) + \Delta L_c - \Delta S_c = -F + \Delta L_c - \Delta S_c$$


longwave cloud forcing
longwave cloud forcing (W m^-2)
0.000
16.000
32.000
48.000
64.000
80.000
Data Min = -3.414, Max = 83.572


shortwave cloud forcing
shortwave cloud forcing (W m^-2)
-120.000
-96.000
-72.000
-48.000
-24.000
0.000
Data Min = -122.744, Max = 25.446


Net cloud forcing
longwave cloud forcing + shortwave cloud forcing (W m^-2)
-50
-30
-10
10
30
50
Data Min = -82.455, Max = 59.21

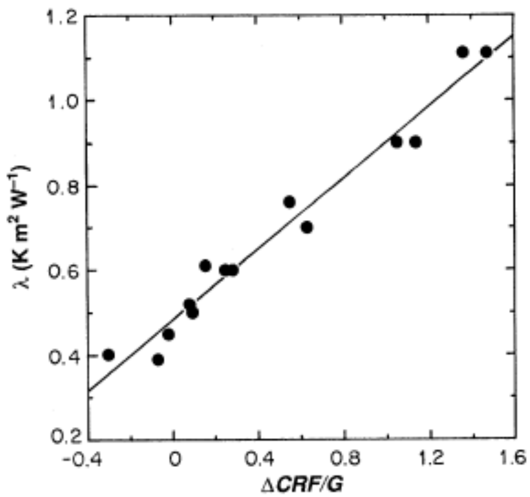


Fig. 1. The global sensitivity parameter λ plotted against the cloud feedback parameter Δ*CRF*/*G* for the 14 GCM simulations. The solid line represents a best-fit linear regression. Cess (1989)

Dividing by F we get

$$\frac{\Delta CRF}{F} = -1 + \frac{\Delta L_c - \Delta S_c}{F}$$

Let's define the clear-sky climate sensitivity λ_c as the ratio of temperature change to the forcing over clear-skies only. Obviously, to get the all-skies (with clouds) temperature change, the forcing should be different than F and equal to $\Delta L_c - \Delta S_c$, that is

$$\lambda_c = \frac{\Delta T_s}{\Delta L_c - \Delta S_c}$$

Then, we get that

$$\lambda = \lambda_c (1 + \frac{\Delta CRF}{F}) \qquad (10)$$

Thus, clouds can either increase or decrease sensitivity of climate to focing depending on the sign of the CRF change as the result of that forcing. The figure above, taken from the paper by Cess (1989), illustrates this dependence as simulated by 14 climate models. One can see that different models produce a wide range of CRF change. Moreover, the models can disagree not only on the magnitude of the cloud effect, but even on its sign! The best-fit line crosses zero CRF change at $\lambda = \lambda_c = 0.45$ K/W^{-1}m^2. That's the sensitivity of simulated climate with no cloud feedbacks. One can also see that the cloud feedback alone can increase the climate sensitivity by almost a factor of 2.5! Note that the results presented on that figure are for the state of the art of climate models in 1989. However, today, the range of predicted mean surface temperature change in response to doubling CO_2 is basically the same as in 1989, from 1.5 K to 4.5 K. The main reason for such a model diversity (which is not a good thing in this context) is believed to be still the uncertainty of the cloud feedback. Apparently, rather small progress has been made over the last 25 years to constrain the magnitude of the climate sensitivity despite the tremendous progress in computing technologies. The problem of climate sensitivity and particularly of cloud feedback is indeed difficult and still refuses to die!