Today's Lecture (Lecture 10): Forcing and feedbacks

Reference

- Hartmann, Ch. 9 and 12 (caution: note that Hartmann's sensitivity parameter λ_R is the reciprocal of the feedback parameter λ that we use)
- IPCC AR5, Ch. 7–8

How does the radiative budget change if we add GHG to the atmosphere?

One-layer atmosphere model

Recall our simple one-layer atmosphere model in radiative equilibrium (but this time allow atmospheric emissivity $\epsilon_A \neq 1$, surface emissivity $\epsilon_S \neq 1$) $\Longrightarrow$ as a result, $(1-\epsilon_A)$ of the surface emission will escape to space:

$$\frac{S_0}{4}(1-\alpha) = \epsilon_A \sigma T_A^4 + (1-\epsilon_A)\epsilon_S \sigma T_S^4 \qquad \text{TOA} \tag{7.1}$$

$$\epsilon_S \sigma T_S^4 = (1-\epsilon_A)\epsilon_S \sigma T_S^4 + 2\epsilon_A \sigma T_A^4 \qquad \text{atmosphere} \tag{7.2}$$

$$\frac{S_0}{4}(1-\alpha) + \epsilon_A \sigma T_A^4 = \epsilon_S \sigma T_S^4 \qquad \text{surface} \tag{7.3}$$

Rearranging terms in (7.3), we recover the relationship between surface and atmospheric temperature:

$$\epsilon_S \sigma T_S^4 = 2\sigma T_A^4 \tag{7.4}$$

Inserting (7.2) into (7.1) yields

$$\frac{S_0}{4}(1-\alpha) = (2-\epsilon_A)\sigma T_A^4 = \frac{2-\epsilon_A}{2}\epsilon_S \sigma T_S^4 \tag{7.5}$$

This result relates the radiative balance of the climate system to the T_S. (This is useful because T_S is of more consequence to ground-dwellers than T_A or T_e.)

How does the radiative budget change if we add GHG to the atmosphere?

Effective emissivity

The effective emissivity of the climate system is

$$\epsilon = \frac{2-\epsilon_A}{2}\epsilon_S \tag{7.6}$$

In the limiting cases $\epsilon_A = 0$ (no greenhouse effect) and $\epsilon_A = 1$ (fully opaque one-layer atmosphere), we recover the customary results from (7.5) with $\epsilon_S \approx 1$: $T_S = 255$ K and $T_S = 255\sqrt[4]{2}$ K, respectively.

Increasing CO_2 concentration

If we model the atmospheric emissivity as a logarithmic function of CO_2 concentration $[CO_2]$ (which is itself an approximation),

$$\epsilon_A = k\log[CO_2] \tag{7.7}$$

then we can linearize a change in emissivity around the unperturbed concentration as

$$\epsilon_A \approx k\log[CO_2] + k\frac{\Delta[CO_2]}{[CO_2]} \quad \text{for} \quad \Delta[CO_2] \ll [CO_2] \tag{7.8}$$

Either way, increasing the CO_2 concentration increases ϵ_A (so that the atmosphere becomes more opaque to thermal radiation). By (7.6), the result of the increased atmospheric emissivity is a decreased effective emissivity of the climate system: the atmosphere emits more strongly, but it also absorbs the surface emission more strongly, leading to a net decrease of TOA emission.

Increasing the CO_2 concentration while holding T_S constant means that (7.5) is no longer satisfied. To reestablish energy balance, the other variables in the equation must change.

TOA energy imbalance

In equilibrium, by rearranging (7.5) and using the effective emissivity (7.6),

$$\frac{S_0}{4}(1-\alpha) - \epsilon\sigma T_S^4 = 0 \tag{7.9}$$

Now consider perturbations due to the system: $\Delta\alpha, \Delta\epsilon, \Delta T_S$ occurring because of an anthropogenic activity Δx. (We could add ΔS_0 due to solar cycles or orbital cycles, but will restrict ourselves to anthropogenic forcing in the following discussion.) These perturbations will cause a TOA radiative imbalance ΔR:

$$\Delta R = -\frac{S_0}{4}\Delta\alpha - \Delta\epsilon\sigma T_S^4 - 4\epsilon\sigma T_S^3 \Delta T_S \tag{7.10}$$

Sign convention: $\Delta R > 0$ means the climate system is gaining energy.

Perturbations

Each of the perturbations can be decomposed in the change that is caused directly by the anthropogenic activity Δx and indirectly by the change in T_S:

$$\Delta\epsilon = \underbrace{\frac{\partial \epsilon}{\partial T_S}\Delta T_S}_{\substack{\text{water vapor}\\ \text{clouds}\\ \text{lapse rate}}} + \underbrace{\frac{\partial \epsilon}{\partial x}\Delta x}_{\text{anth. GHG}} \tag{7.11}$$

$$\Delta\alpha = \underbrace{\frac{\partial \alpha}{\partial T_S}\Delta T_S}_{\substack{\text{snow/sea ice melt}\\ \text{clouds}}} + \underbrace{\frac{\partial \alpha}{\partial x}\Delta x}_{\substack{\text{land-use change}\\ \text{anth. aerosol}}} \tag{7.12}$$

Note: in both cases, we're converting the anthropogenic perturbation into a radiative flux change or radiative forcing

Forcing and response

Inserting (7.11) and (7.12) into (7.10), we find

$$\overbrace{\Delta R}^{\text{imbalance}} = \overbrace{\left(-\frac{S_0}{4}\frac{\partial \alpha}{\partial T_S} - \sigma T_S^4 \frac{\partial \epsilon}{\partial T_S} - 4\epsilon\sigma T_S^4\right)}^{1/\text{sensitivity}} \overbrace{\Delta T_S}^{\text{response}} + \overbrace{\left(-\frac{S_0}{4}\frac{\partial \alpha}{\partial x} - \sigma T_S^4 \frac{\partial \epsilon}{\partial x}\right)\Delta x}^{\text{forcing}}$$

$$= \lambda \Delta T_S + F \qquad (7.13)$$

(7.13) links the concepts relevant to climate change: forcing brings the climate system into an energy imbalance, to which the climate system responds through warming; the amount of warming required to bring the system back into balance ($\Delta R = 0$) is determined by the sensitivity $1/\lambda$. Sign convention: $F > 0$ causes warming.

To answer how the warming proceeds, we note that the energy imbalance is related to a warming rate through the effective heat capacity c of the climate system (dominated by the upper ocean on centennial timescales):

$$c\frac{\partial \Delta T_S}{\partial t} = \lambda \Delta T_S + F \qquad (7.14)$$

(See also this week's homework.)

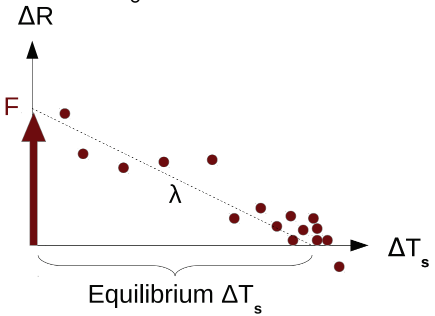


Figure: Karoline Block

How sensitive is the climate system to a forcing?

If $\lambda \geq 0$, we see from (7.14) that the system is unstable. If $\lambda < 0$,

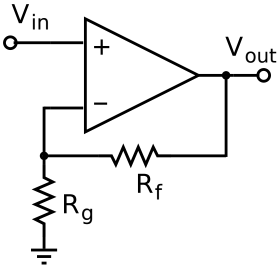
$$(\Delta T_S)_{\text{equilibrium}} = -\frac{F}{\lambda} \qquad (7.15)$$

$|\lambda|$ large: $(\Delta T_S)_{\text{equilibrium}}$ small, system insensitive

$|\lambda|$ small: $(\Delta T_S)_{\text{equilibrium}}$ large, system sensitive

The climate system from a systems analysis perspective

Think of F as an input that is processed by the climate system to result in a response in the form of ΔT_S.
By analogy with other systems where the response modifies the state of the system, the $\lambda \Delta T_S$ term is referred to as a feedback term, and λ as the feedback parameter.



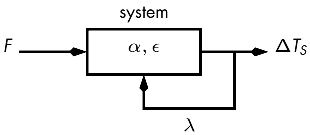


Figure: Wikipedia

Feedbacks

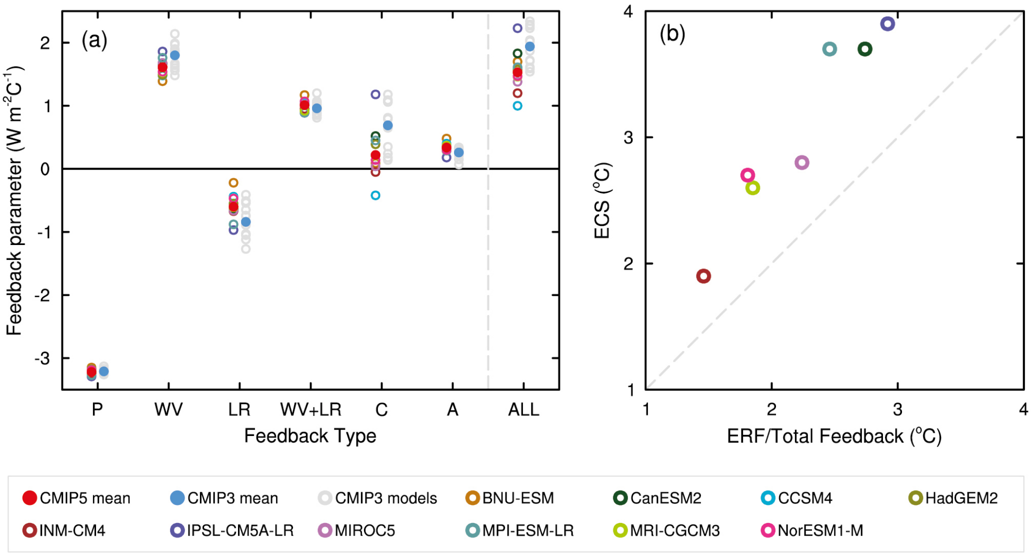
The constituent terms of λ in (7.13) tell us what feedbacks to expect in climate models. Depending on whether the contribution to λ is positive or negative, the feedback is called positive feedback or negative feedback. A positive feedback is destabilizing, a negative feedback stabilizing.

List of feedbacks

$-\frac{S_0}{4}\frac{\partial\alpha}{\partial T_S}$:	albedo change	$\xrightarrow{\text{positive}}$	surface albedo feedback: melting of sea ice and snow cover
		$\xrightarrow{\text{prob. positive}}$	cloud albedo feedback: change in SW CRE
$-\sigma T_S^4 \frac{\partial\epsilon}{\partial T_S}$:	emissivity change	$\xrightarrow{\text{positive}}$	water-vapor feedback: change in opacity of clear-sky atmosphere
		$\xrightarrow{\text{positive}}$	cloud emissivity feedback: change in LW CRE
		$\xrightarrow{\text{negative}}$	lapse-rate feedback: change in atmospheric temperature profile
$-4\epsilon\sigma T_S^3$:	surface black-body response	$\xrightarrow{\text{strongly negative}}$	"Planck" feedback: increased thermal emission by the surface

$\lambda_{\text{Planck}} \approx -3$ W m^{-2} K^{-1} is the strongest feedback and stabilizes the climate system. With $F_{2\times CO_2} \approx 4$ W m^{-2}, the temperature response in the absence of other feedbacks would be $\approx +1.5$ K. (The equilibrium temperature change due to a doubling of CO_2 is conventionally called equilibrium climate sensitivity (ECS).) The sum of the other feedbacks is likely positive, meaning the ECS is likely higher than 1.5 K. (IPCC AR5: likely in the 1.5 to 4.5 K range with high confidence.)

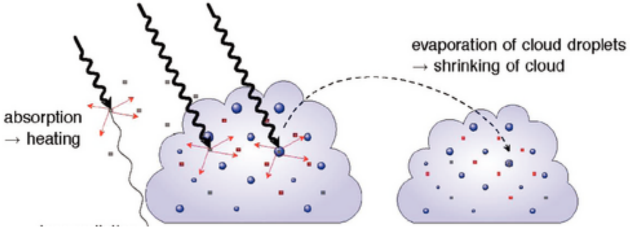
Feedbacks in models



Figures from IPCC AR5 unless noted

7.3 – Adjustments

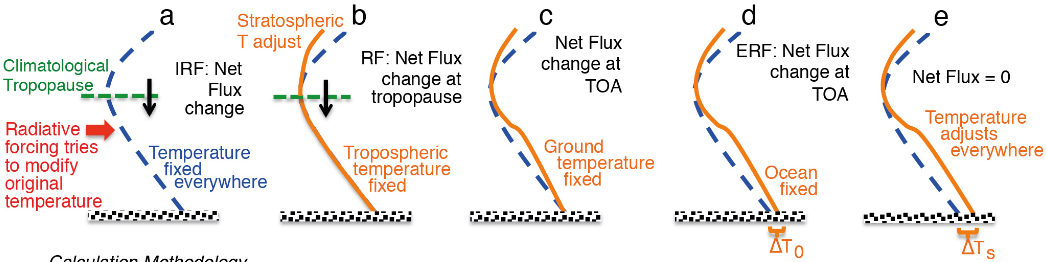
Feedbacks allow the system to respond to the perturbation.
But what about something like the effect of absorbing aerosols on clouds? This affects the radiative fluxes. It is not a feedback, because it does not depend on ΔT_S.



In the IPCC AR5, this and other fast responses of the atmosphere ("fast" meaning faster than the SST response) are called adjustments to the radiative forcing, and the combination of radiative forcing and adjustments is the effective radiative forcing (ERF).

Figure: IPCC AR4

Possible definitions of adjustments

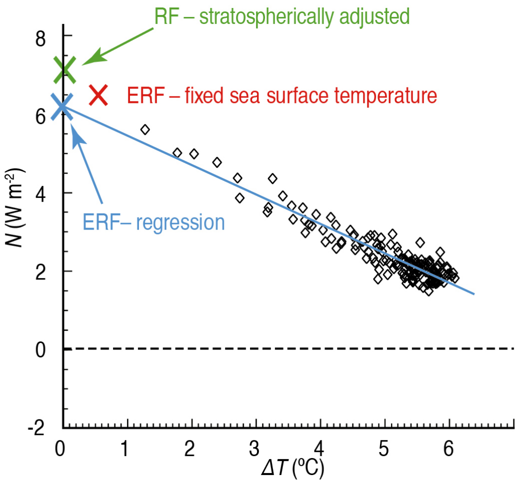


Calculation Methodology

Online or offline pair of radiative transfer calculations within one simulation	Difference between two offline radiative transfer calculations with prescribed surface and tropospheric conditions allowing stratospheric temperature to adjust	Difference between two full atmospheric model simulations with prescribed surface conditions everywhere or estimate based on regression of response in full coupled atmosphere-ocean simulation	Difference between two full atmospheric model simulations with prescribed ocean conditions (SSTs and sea ice)	Difference between two full coupled atmosphere-ocean model simulations

Since adjustments happen on many different time scales, it is a matter of convention which ones to include in the ERF. In IPCC TAR and AR4, the fast cooling of the stratosphere is included in RF, but the other adjustments are treated separately. In IPCC AR5, adjustments up to the land surface response (multi-year timescale) are part of ERF.

Effective radiative forcing



- Illustration of the difference between RF and ERF: adjustments are fast, so they affect ΔR before T_S increases.
- Note that practically, there are two approaches to determining the adjustments from climate models:
  1. Run a coupled (ocean, land, atmosphere) model with an abrupt forcing to equilibrium; determine ERF from ΔR-axis intercept (expensive: coupled model, $>$ 100 year run)[a]
  2. Run an atmosphere + land only model with an abrupt forcing and fixed SST; determine ERF from ΔR after atmosphere and land have come to equilibrium — note $\Delta T_S > 0$ (cheap: no ocean, $\mathcal{O}(10$ year$)$ run)
- Why is it important to know the ERF? For CO_2 emissions, the RF is fairly simple to determine (from a radiative transfer model); however, the initial TOA energy imbalance and therefore the heating rate of the climate system are determined by the ERF.
- The same is true for aerosol forcing — except that both the forcing and the adjustments are still very uncertain.

[a] See Gregory et al. (2004), Andrews et al. (2012)