

## Today's Lecture (Lecture 11): Forcing and feedbacks

### Reference

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

# 4 – Forcing and feedbacks

## 1. Introduction

## 2. The climate system

## 3. Internal variability

## 4. Forcing and feedbacks

### 4.1 Radiative forcing

### 4.2 Feedbacks

### 4.3 Adjustments

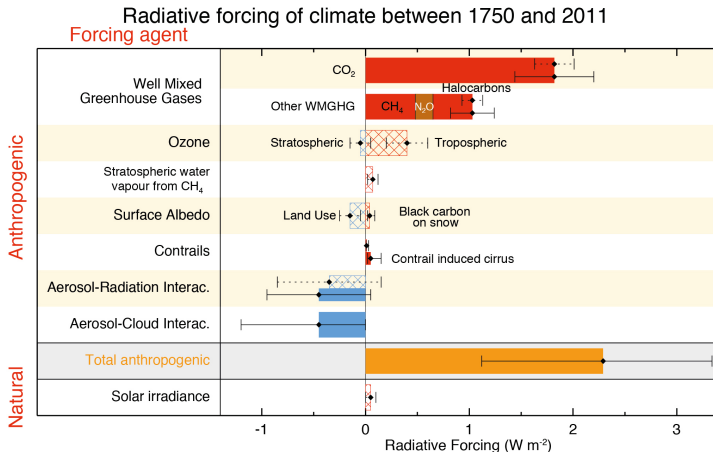
## 5. Anthropogenic climate change

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## 4.1 – Radiative forcing

- ▶ Change in net energy exchange between the climate system and the environment
- ▶ Examples:
  - ▶ Solar intensity cycles
  - ▶ Orbital parameter cycles
  - ▶ Volcanic eruptions
  - ▶ Anthropogenic
    - ▶ Greenhouse gases
    - ▶ Aerosols
    - ▶ Ozone depletion
    - ▶ Land-use change
- ▶ Utility: intercomparability of the different forcing agents; condenses the problem to a global number
- ▶ Drawbacks: ambiguity in definition; rapid adjustments; reference state; condenses the problem to a global number



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

## One-layer gray-body atmosphere model

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

$$\frac{S_0}{4}(1 - \alpha) = \epsilon_A \sigma T_A^4 + (1 - \epsilon_A) \epsilon_S \sigma T_S^4 \quad \text{TOA} \quad (4.1)$$

$$\epsilon_S \sigma T_S^4 = (1 - \epsilon_A) \epsilon_S \sigma T_S^4 + 2 \epsilon_A \sigma T_A^4 \quad \text{atmosphere} \quad (4.2)$$

$$\frac{S_0}{4}(1 - \alpha) + \epsilon_A \sigma T_A^4 = \epsilon_S \sigma T_S^4 \quad \text{surface} \quad (4.3)$$

Rearranging terms in (4.2), we recover the relationship between surface and atmospheric temperature:

$$\epsilon_S \sigma T_S^4 = 2 \sigma T_A^4 \quad (4.4)$$

Inserting (4.4) into (4.1) yields

$$\frac{S_0}{4}(1 - \alpha) = (2 - \epsilon_A) \sigma T_A^4 = \frac{2 - \epsilon_A}{2} \epsilon_S \sigma T_S^4 \quad (4.5)$$

(4.5) has the form of a radiative balance equation, using  $T_S$  as the response variable. **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 in (4.5) is

$$\epsilon = [(2 - \epsilon_A)/2]\epsilon_S \quad (4.6)$$

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 (4.5) with  $\epsilon_S \approx 1$ :  $T_S = 255$  K and  $T_S = 255\sqrt[4]{2}$  K, respectively.

### Increasing CO<sub>2</sub> concentration

If we model the atmospheric emissivity as a logarithmic function of CO<sub>2</sub> concentration [CO<sub>2</sub>] (which is itself an approximation),

$$\epsilon_A = k \log[\text{CO}_2] \quad (4.7)$$

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

$$\epsilon_A \approx k \log[\text{CO}_2] + k\Delta[\text{CO}_2]/[\text{CO}_2] \quad \text{for } \Delta[\text{CO}_2] \ll [\text{CO}_2] \quad (4.8)$$

Either way, increasing the CO<sub>2</sub> concentration increases  $\epsilon_A$  (so that the atmosphere becomes more opaque to thermal radiation). By (4.6), 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<sub>2</sub> concentration while holding  $T_S$  constant means that (4.5) is no longer satisfied. To reestablish energy balance, the other variables in the equation must change.

## TOA energy imbalance

In equilibrium, by rearranging (4.5) and using the effective emissivity (4.6),

$$\frac{S_0}{4}(1 - \alpha) - \epsilon\sigma T_S^4 = 0 \quad (4.9)$$

Now consider perturbations to the system,  $\{\Delta\alpha, \Delta\epsilon, \Delta T_S\}$ , occurring because of an anthropogenic activity  $\Delta x$ . (We could add  $\Delta S_0$  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  $\Delta R$ :

$$\Delta R = -\frac{S_0}{4}\Delta\alpha - \Delta\epsilon\sigma T_S^4 - 4\epsilon\sigma T_S^3\Delta T_S \quad (4.10)$$

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

## Perturbations

Each of the perturbations can be decomposed into the change that is caused directly by the anthropogenic activity  $\Delta x$  and indirectly by the change in  $T_S$ :

$$\Delta\epsilon = \underbrace{\frac{\partial\epsilon}{\partial T_S} \Delta T_S}_{\substack{\text{water vapor} \\ \text{clouds} \\ \text{lapse rate}}} + \underbrace{\frac{\partial\epsilon}{\partial x} \Delta x}_{\text{anth. GHG}} \quad (4.11)$$

$$\Delta\alpha = \underbrace{\frac{\partial\alpha}{\partial T_S} \Delta T_S}_{\substack{\text{snow/sea ice melt} \\ \text{clouds}}} + \underbrace{\frac{\partial\alpha}{\partial x} \Delta x}_{\substack{\text{land-use change} \\ \text{anth. aerosol}}} \quad (4.12)$$

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

## Forcing and response

Inserting (4.11) and (4.12) into (4.10), we find

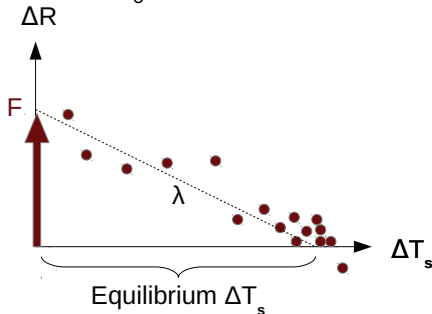
$$\begin{aligned} \underbrace{\Delta R}_{\text{imbalance}} &= \underbrace{\left( -\frac{S_0}{4} \frac{\partial \alpha}{\partial T_S} - \sigma T_S^4 \frac{\partial \epsilon}{\partial T_S} - 4\epsilon \sigma T_S^3 \right)}_{1/\text{sensitivity}} \underbrace{\Delta T_S}_{\text{response}} + \underbrace{\left( -\frac{S_0}{4} \frac{\partial \alpha}{\partial x} - \sigma T_S^4 \frac{\partial \epsilon}{\partial x} \right)}_{\text{forcing}} \Delta x \\ &= \lambda \Delta T_S + F \end{aligned} \quad (4.13)$$

(4.13) 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):

$$c \frac{\partial \Delta T_S}{\partial t} = \lambda \Delta T_S + F \quad (4.14)$$

(See also this week's homework.)





## 4.2 – Feedbacks

### How sensitive is the climate system to a forcing?

If  $\lambda \geq 0$ , we see from (4.14) that the system is unstable. If  $\lambda < 0$ ,

$$(\Delta T_S)_{\text{equilibrium}} = -\frac{F}{\lambda} \quad (4.15)$$

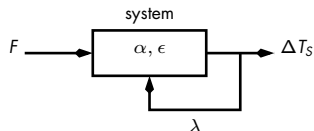
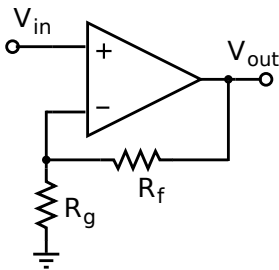
$|\lambda|$  large:  $(\Delta T_S)_{\text{equilibrium}}$  small, system insensitive

$|\lambda|$  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  $\Delta T_S$ .

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  $\lambda$  as the feedback parameter.



## Feedbacks

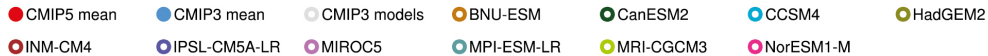
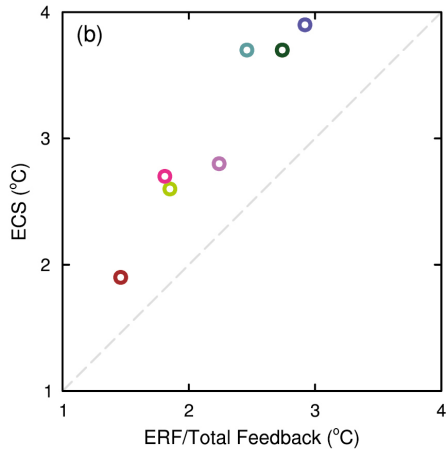
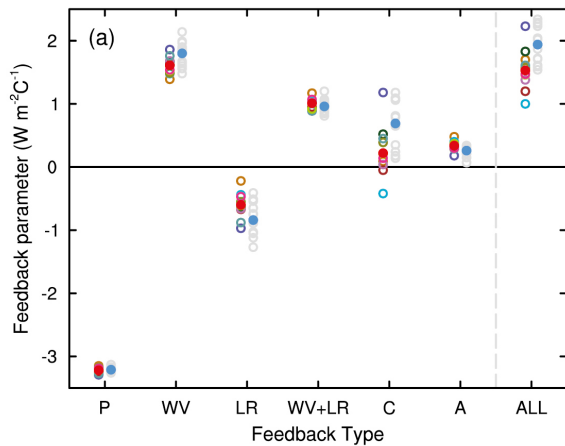
The constituent terms of  $\lambda$  in (4.13) tell us what feedbacks to expect in climate models. Depending on whether the contribution to  $\lambda$  is positive or negative, the feedback is called positive feedback or negative feedback. A positive feedback is destabilizing, a negative feedback stabilizing.

### List of feedbacks

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

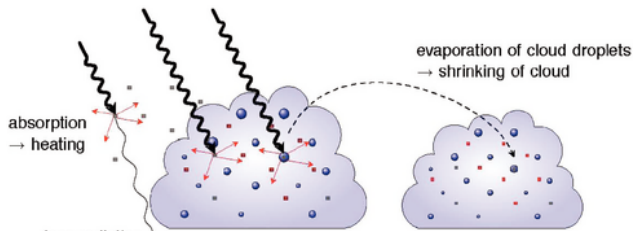
$\lambda_{\text{Planck}} \approx -3 \text{ W m}^{-2} \text{ K}^{-1}$  is the strongest feedback and stabilizes the climate system. With  $F_{2\times\text{CO}_2} \approx 4 \text{ W m}^{-2}$ , the temperature response in the absence of other feedbacks would be  $\approx +1.5 \text{ K}$ . (The equilibrium temperature change due to a doubling of  $\text{CO}_2$  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



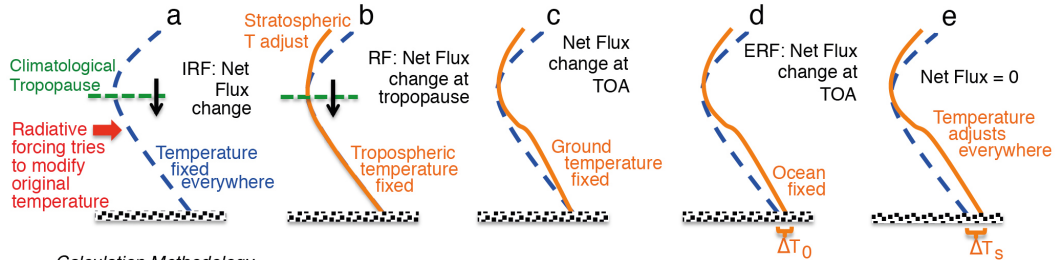
## 4.3 – Adjustments

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  $\Delta T_S$ .



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)**.

# 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

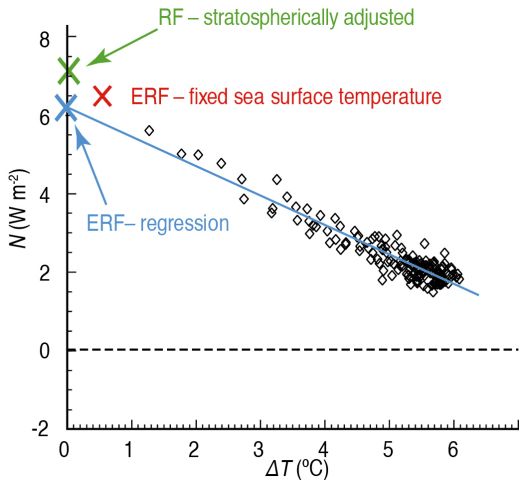
Difference between two full atmospheric model simulations with prescribed surface conditions everywhere or estimate based on regression of response in full coupled atmosphere-ocean simulation

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  $\Delta R$  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  $\Delta R$ -axis intercept (expensive: coupled model,  $> 100$  year run)<sup>a</sup>
  2. Run an atmosphere + land only model with an abrupt forcing and fixed SST; determine ERF from  $\Delta R$  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  $\text{CO}_2$  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.

<sup>a</sup>See Gregory et al. (2004), Andrews et al. (2012)