Radiative Forcing and Energy Balance Models

Topics:
1. Radiative forcing
2. Climate sensitivity and feedbacks
3. Radiative equilibrium
4. Energy balance models

Reading: Liou 8; Thomas 12.1-12.7, 12.10, 12.11

Planetary radiative equilibrium:

Over some time the longwave radiation emitted by a planet must equal the solar radiation absorbed.
Therefore the top of atmosphere global average net flux is zero:

$$N = F_{net}^{TOA} = \frac{S}{4}(1 - \bar{r}) - F_{LW}^{TOA} = 0$$

where $\bar{r}$ is the global mean albedo, S is the solar constant $S = 1366\ \mathrm{W/m^2}$, and F_{LW}^{TOA} is the emitted longwave flux.

This can be used to define an effective blackbody temperature:

$$\sigma T_e^4 = \frac{S}{4}(1 - \bar{r}) \qquad T_e = \left(\frac{S(1 - \bar{r})}{4\sigma}\right)^{1/4}$$

For Earth the albedo is $\bar{r} \approx 0.30$, the absorbed solar flux is $\frac{S}{4}(1-\bar{r}) = 239\ \mathrm{W/m^2}$, and the effective temperature is $T_e = 255$ K.

Planet	S (W/m^2)	$\bar{r}$	T_e (K)	T_{sfc} (K)
Earth	1366	0.30	255	288
Venus	2640	0.78	225	730
Mars	590	0.17	216	218

Radiative Forcing

Our definition: Change in top of atmosphere net radiative flux (longwave and shortwave), ΔQ, due to an *external perturbation*.
Official definition (IPCC): Change in net radiative flux at the tropopause after stratosphere comes back into equilibrium.

Examples: change in solar constant, aerosol from pollution or a volcanic eruption, and increased greenhouse gases from industrial society, e.g. doubling CO_2 is $\Delta Q = 4\ \mathrm{W/m^2}$.

A change in water vapor due to climate change is *not* a radiative forcing (though it has an important radiative effect), because water vapor is an internal climate variable.

Positive radiative forcing: more absorbed shortwave or less outgoing longwave $\rightarrow$ climate warming. Negative radiative forcing $\rightarrow$ cooling.

Radiative forcing is a useful concept because it is often more accurate to compute radiative forcing than the global temperature response.

Climate Sensitivity and Feedbacks

Climate sensitivity is the relation between a change climate and a some radiative forcing.

Often climate sensitivity is simply defined to be the global temperature change from a radiative forcing, $\Delta T_s/\Delta Q$ ($\mathrm{K\ W^{-1}\ m^2}$).

The no feedback climate sensitivity is $G_0 = \Delta T_s/\Delta Q = 0.3\ \mathrm{K/(W/m^2)}$.

A *climate feedback* is a process that changes the climate sensitivity. Often a feedback changes the radiative balance.

Positive feedbacks enhance the climate change (for a fixed radiative forcing), while negative feedbacks diminish the climate change.
Examples:

1) Water vapor: warmer climate, more water vapor, less emitted LW
2) Snow-ice albedo: warmer climate, less snow/sea ice, lower albedo

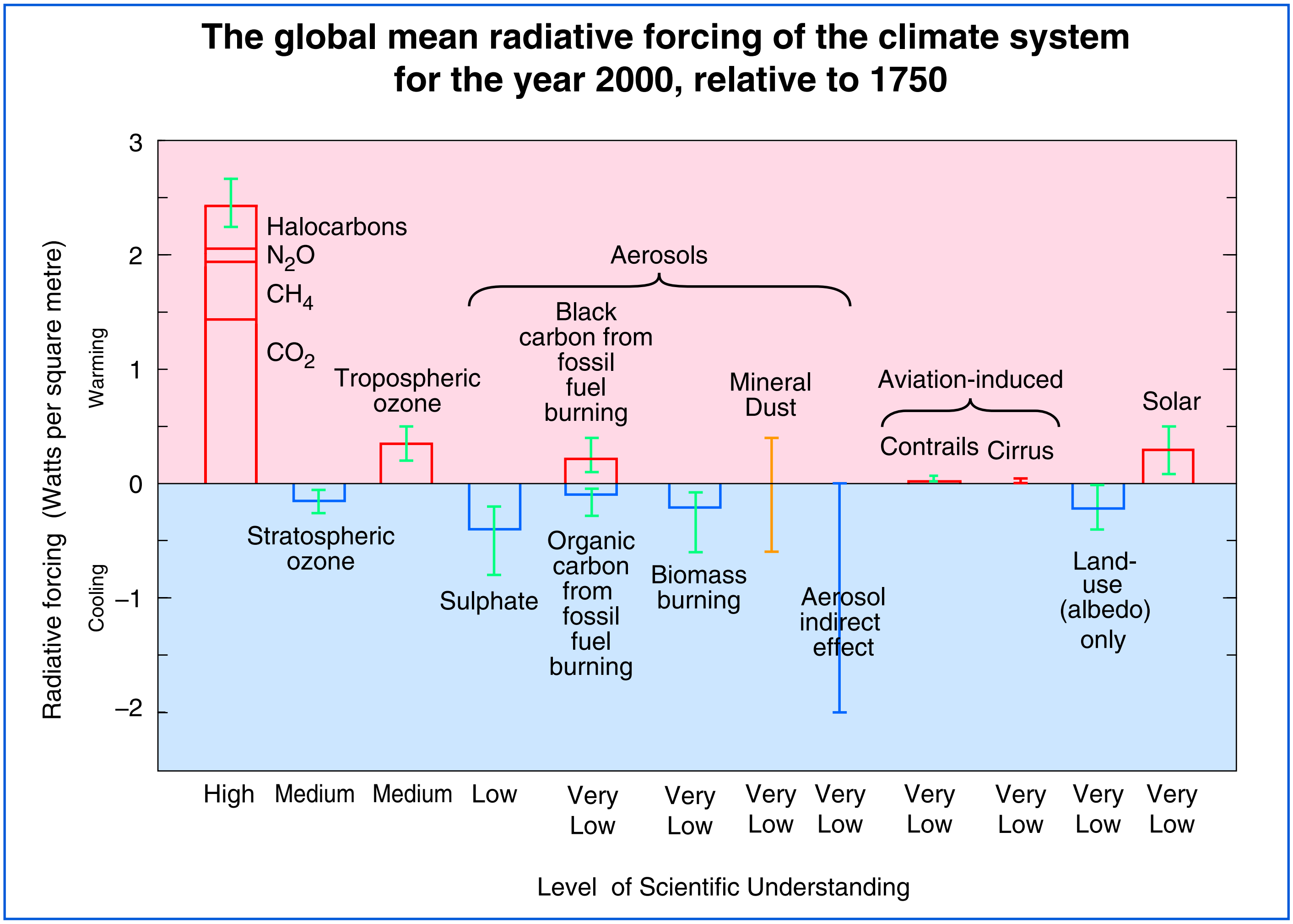
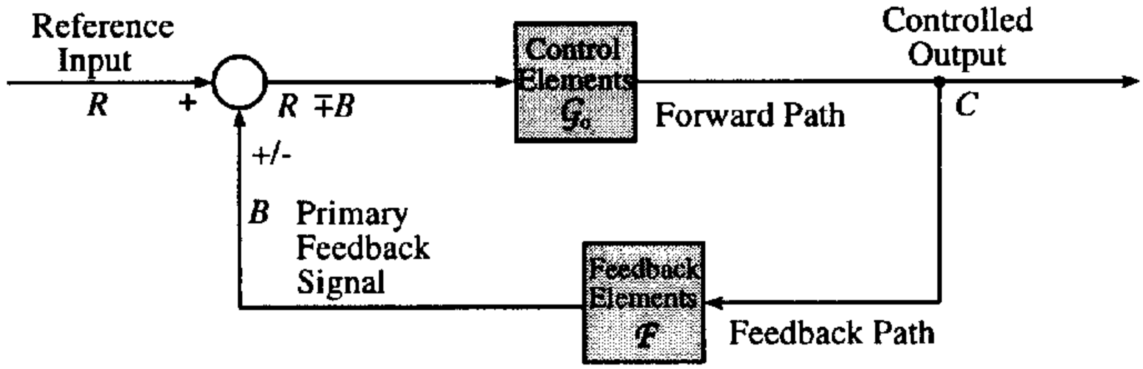


Figure 3: Many external factors force climate change.

These radiative forcings arise from changes in the atmospheric composition, alteration of surface reflectance by land use, and variation in the output of the sun. Except for solar variation, some form of human activity is linked to each. The rectangular bars represent estimates of the contributions of these forcings – some of which yield warming, and some cooling. Forcing due to episodic volcanic events, which lead to a negative forcing lasting only for a few years, is not shown. The indirect effect of aerosols shown is their effect on the size and number of cloud droplets. A second indirect effect of aerosols on clouds, namely their effect on cloud lifetime, which would also lead to a negative forcing, is not shown. Effects of aviation on greenhouse gases are included in the individual bars. The vertical line about the rectangular bars indicates a range of estimates, guided by the spread in the published values of the forcings and physical understanding. Some of the forcings possess a much greater degree of certainty than others. A vertical line without a rectangular bar denotes a forcing for which no best estimate can be given owing to large uncertainties. The overall level of scientific understanding for each forcing varies considerably, as noted. Some of the radiative forcing agents are well mixed over the globe, such as CO_2, thereby perturbing the global heat balance. Others represent perturbations with stronger regional signatures because of their spatial distribution, such as aerosols. For this and other reasons, a simple sum of the positive and negative bars cannot be expected to yield the net effect on the climate system. The simulations of this assessment report (for example, Figure 5) indicate that the estimated net effect of these perturbations is to have warmed the global climate since 1750. [Based upon Chapter 6, Figure 6.6]

Control Theory

Feedbacks may be analyzed with linear control theory.



Basic bloack diagram including feedback elements. Negative feedback means that the summing point is a subtractor, i.e. $R - B$. Positive feedback means that the summing point is an adder, i.e. $R + B$. [Curry and Webster, 1999; Fig 13.1]

The input is the radiative forcing ΔQ, the gain of the system with no feedbacks is G_0, the output is the climate response (e.g. T_s), and the feedback $\mathcal{F}$ is the TOA radiative effect of the feedback process.

The *feedback factor* f is defined as $f = G_0 \mathcal{F}$ (unitless).

Temperature Response to Forcing

Imagine an instantaneous perturbation causing a radiative forcing ΔQ.

Initially the net flux is out of balance by $\Delta N = \Delta Q$.

The direct temperature response, i.e. with no feedbacks, restores the balance with a temperature change ΔT_0:

$$\Delta N = \Delta Q - \frac{\partial F_{LW}^{TOA}}{\partial T} \Delta T \ \rightarrow 0$$

$$\Delta T_0 = \frac{\Delta Q}{\partial F_{LW}^{TOA} / \partial T} = G_0 \Delta Q \qquad G_0 = \left(\frac{\partial F}{\partial T} \right)^{-1}$$

If there are climate feedbacks then the total derivative of N is needed:

$$\Delta N = \Delta Q + \frac{dN}{dT} \Delta T = 0$$

$$\Delta T = \frac{\Delta Q}{-dN/dT} = G \Delta Q \qquad G = -\left(\frac{dN}{dT} \right)^{-1}$$

The system gain G is the climate sensitivity.

Climate Response with Feedbacks

Climate feedbacks are analyzed by expanding the total derivative of net flux in terms of contributions from various internal variables X_j:

$$\frac{dN}{dT} = \frac{\partial N}{\partial T} + \sum_j \frac{\partial N}{\partial X_j}\frac{\partial X_j}{\partial T}$$

Therefore the climate sensitivity is

$$G = -\left(\frac{\partial N}{\partial T} + \sum_j \frac{\partial N}{\partial X_j}\frac{\partial X_j}{\partial T}\right)^{-1} \quad = \quad \left(\frac{\partial F}{\partial T} - \sum_j \frac{\partial N}{\partial X_j}\frac{\partial X_j}{\partial T}\right)^{-1}$$

The feedback factor for the j'th climate variable is

$$f_j = G_0 \frac{\partial N}{\partial X_j}\frac{\partial X_j}{\partial T}$$

Positive feedbacks have $f_j > 0$; negative feedbacks have $f_j < 0$.

For example, for the water vapor feedback the variable is $X = u_v$, the water vapor amount. Increasing temperature increases the amount of water vapor $\frac{\partial u_v}{\partial T} > 0$, and an increase in water vapor increases the TOA net radiative flux $\frac{\partial N}{\partial u_v} > 0$ (since the outgoing longwave flux is reduced $\frac{\partial F_{LW}}{\partial u_v} < 0$). Therefore, the water vapor feedback factor is positive $f_v > 0$.

Adding Feedbacks

The feedback factors add for multiple feedbacks $f = \sum_j f_j$, but the individual temperature responses ΔT_j do not! The total system gain is that due to all the feedbacks, so the climate sensitivity is

$$G = \frac{G_0}{1 - \sum_j f_j}$$

If for a doubling of CO_2, $\Delta T_0 = 1.2$ K and a typical GCM gives $\Delta T = 4$ K, then this implies the total feedback factor is $f \approx 0.7$.

The effect of one feedback can be determined in a climate model with three runs:

1) A control run with current climate ($\Delta Q = 0$).
2) A run with radiative forcing ΔQ and all feedbacks on to get a temperature change, ΔT_{all}
3) A climate change run with feedback j turned off, to get ΔT_{-j}

The feedback factor is then

$$f_j = \left(\frac{1}{\Delta T_{-j}} - \frac{1}{\Delta T_{all}} \right) \Delta T_0$$

which results from $dN/dT = -\Delta Q/\Delta T$.

Radiative Equilibrium and Energy Balance models

Radiative Equilibrium: occurs when the total (shortwave + longwave) radiative heating is zero at all levels.

Energy balance models solve for planetary temperatures by balancing radiative energy (and perhaps other energy transports).

There is a hierarchy of energy balance models of differing complexity.
These models:

1) can explain temperature profiles in planetary atmospheres.
2) are used to analyze climate sensitivity to radiative forcing.

One-dimensional (latitude) energy balance model

Traditional, simple climate model (Budyko, 1969; Sellers, 1969; North 1975)

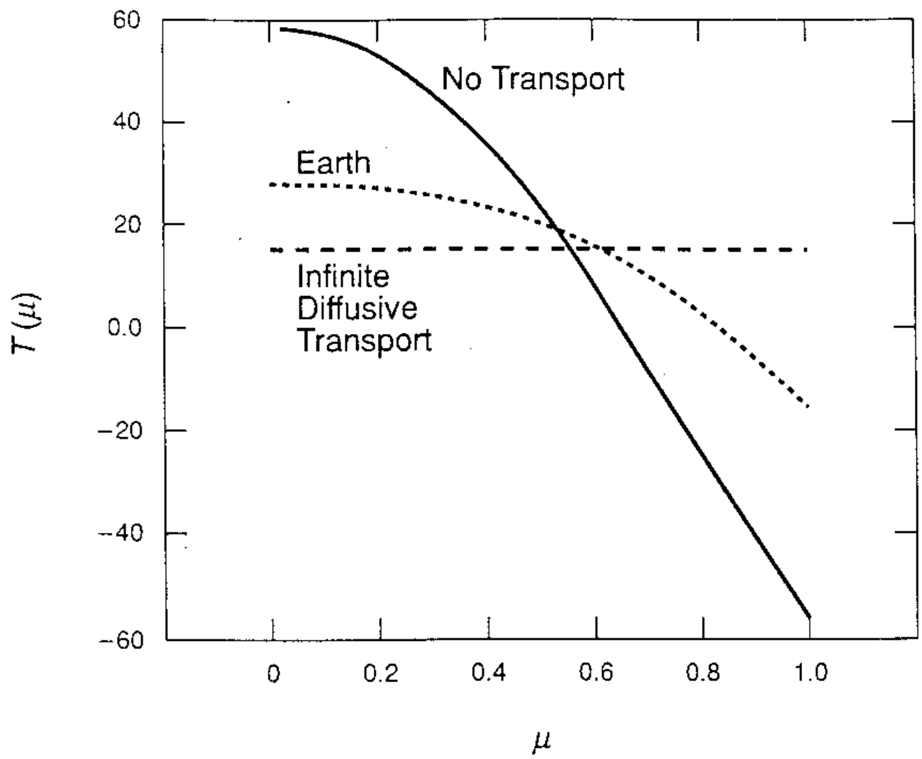
Atmosphere is only implicit:
Outgoing longwave flux is e.g. $F_{LW} = (1.55\ \mathrm{W/m^2/K})T_s - 212\ \mathrm{W/m^2}$

Annual mean TOA insolation fit well with $S(x) = S_0/4[1 - 0.482\mathcal{P}_2(x)]$,
where x is sine of latitude and $\mathcal{P}_2(x) = (3x^2 - 1)/2$.

Energy balance equilibrium ($\frac{\partial T}{\partial t} = 0$) with diffusive transport:

$$-D\frac{\partial}{\partial x}(1 - x^2)\frac{\partial F_{LW}}{\partial x} + F_{LW} = S(x)[1 - r(x)]$$

D is diffusion coefficient for energy transport and $r(x)$ is albedo.



Zonally average surface temperature (K) as a function of the sine of the latitude, μ, observed and for cases of no horizontal heat transport, and infinite horizontal heat transport. From North et al. (1981). [Fig 10.2; Kiehl's chapter in Trenberth, "Climate System Modeling"]

With no meridional transport, the poles are way too cold.

These models have been used to study the ice-albedo feedback by having the albedo $r(x)$ depend on temperature $T(x)$.

Single layer gray energy balance model

Assumptions: 1) Isothermal layer, 2) Gray atmosphere in LW,
3) Black surface in LW, 4) No atmosphere absorption in SW.

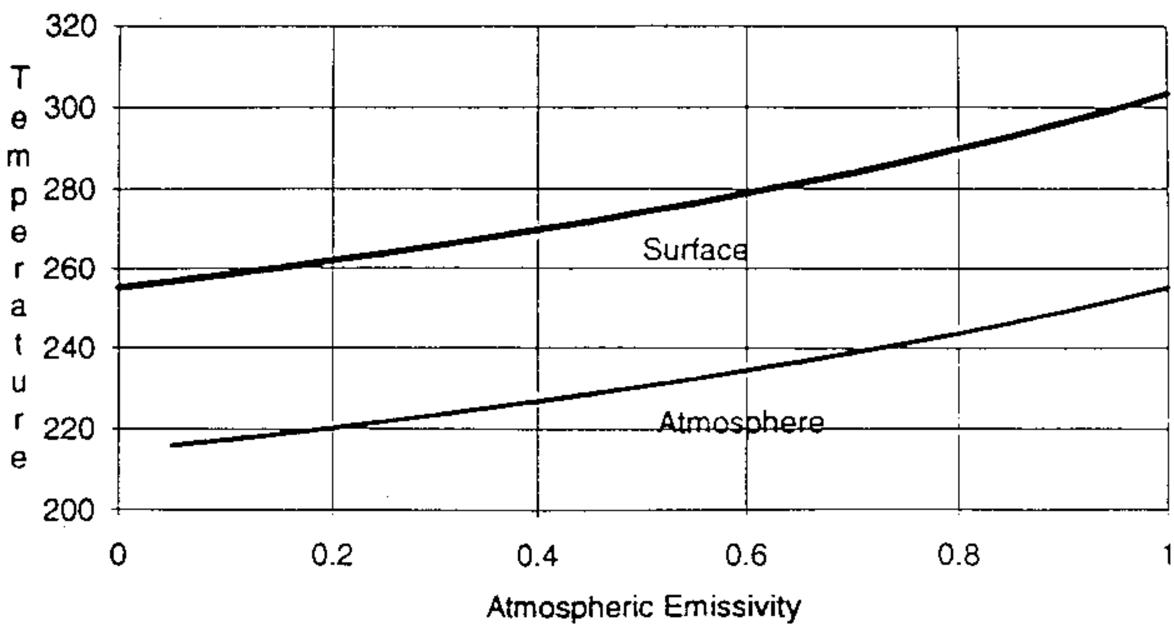
Absorbed solar flux is $F_{sun} = \frac{S}{4}(1-\bar{r}) = \sigma T_e^4$
S is solar constant, $\bar{r}$ is mean albedo, T_e is effective temperature.

Surface balance: $F_{sun}^{\downarrow} + F_{atm}^{\downarrow} = F_{sfc}^{\uparrow}$
$F_{sfc}^{\uparrow} = \sigma T_s^4 \quad F_{atm}^{\downarrow} = \epsilon_a \sigma T_a^4 \qquad \epsilon_a$ is LW atmosphere emissivity.
TOA balance: $F_{sun} = F_{atm}^{\uparrow} + (1-\epsilon_a)F_{sfc}^{\uparrow}$

Two unknowns: surface and atmosphere temperature - T_s and T_s.
Eliminate $F_{sfc}^{\uparrow}$: $F_{atm} = \frac{\epsilon_a}{2-\epsilon_a} F_{sun}$
In terms of temperatures:

$$T_a = \frac{T_e}{(2-\epsilon_a)^{1/4}} \;\le T_e \qquad T_s = \frac{T_e}{(1-\epsilon_a/2)^{1/4}} \;= 2^{1/4} T_a \;\; \ge T_e$$

Limits: thin atmosphere $\epsilon_a \to 0 \;\Rightarrow\; T_a = T_e/2^{1/4} \quad T_s = T_e$
black atmosphere $\epsilon_a = 1 \;\Rightarrow\; T_a = T_e \quad T_s = 2^{1/4} T_e$



One-layer gray model temperatures as a function of atmospheric emissivitivy for the Earth's globally averaged absorbed solar energy (239 W/m^2).

Single black layer with spectral window

A perhaps more realistic way to deal with partial longwave transparency of the atmosphere, is to assume a fraction of the spectrum is clear.

Assume: 1) Fraction $f = \frac{1}{\sigma T^4}\int_{\nu_1}^{\nu_2} B_\nu(T)d\nu$ of LW spectrum is completely transparent, 2) Remainder of LW spectrum is black,
3) Black surface in LW, 4) No atmosphere absorption in SW.

Surface balance: $F_{sun} + (1-f)\sigma T_a^4 = \sigma T_s^4$
TOA balance: $F_{sun} = (1-f)\sigma T_a^4 + f\sigma T_s^4$

Add equations $[\sigma T_s^4 = 2F_{sun}/(1+f)]$ to get solution:

$$T_a = \left(\frac{1}{1+f}\right)^{1/4} T_e \ \leq T_e \qquad T_s = \left(\frac{2}{1+f}\right)^{1/4} T_e \ \geq T_e$$

Limits: no window $f = 0 \ \Rightarrow \ T_a = T_e \quad T_s = 2^{1/4} T_e$
all window $f = 1 \ \Rightarrow \ T_a = T_e/2^{1/4} \quad T_s = T_e$
Earth: $f \approx 0.3 \ T_a = 239$ K $\ T_s = 284$ K

Multiple black layers with spectral window

Need multiple atmosphere layers to explain atmosphere temperature profile.

Black layers $\rightarrow$ LW flux absorbed by a layer comes from adjacent layers.
Portion of surface flux emitted in window escapes directly to space.

TOA: $F_{sun} = \sigma T_e^4 \ = \ (1-f)\sigma T_1^4 + f\sigma T_s^4$
Layer 1: $(1-f)\sigma T_2^4 = 2(1-f)\sigma T_1^4$
Layer 2: $(1-f)\sigma T_1^4 + (1-f)\sigma T_3^4 = 2(1-f)\sigma T_2^4$
Layer n: $(1-f)\sigma T_{n-1}^4 + (1-f)\sigma T_{n+1}^4 = 2(1-f)\sigma T_n^4$
Surface: $F_{sun} + (1-f)\sigma T_N^4 = \sigma T_s^4$
Cancel $1-f$ factors: $T_{n+1}^4 = 2T_n^4 - T_{n-1}^4 \ \rightarrow \ T_N^4 = NT_1^4$
Two unknowns T_1 and T_s; solve to get

$$T_s = T_e\left(\frac{1+N}{1+fN}\right)^{1/4} \qquad T_1 = T_e\left(\frac{1}{1+fN}\right)^{1/4}$$

Thicker atmosphere (larger N) gives warmer surface and cooler TOA
$\rightarrow$ greater greenhouse effect.
Spectral window ($f > 0$) allows top of atmosphere to be colder than T_e.