Energy and Climate I: Earth atmosphere and the greenhouse effect

November 26, 2012
So far, have focused on **isolated energy systems**

**Many factors influence energy choices**

- Resource availability/scarcity
- Convenience (energy density, difficulty of conversion)
- Cost (generally controlled by availability/convenience)

Until recently, these factors have favored fossil fuels. But not indefinitely.

- Energy independence/politics
- External consequences of specific energy sources (nuclear waste, climate)

**Need to understand and incorporate consequences in use, cost.**

Latter factors motivate switching to renewables sooner, not later (opinion)
Energy plays a fundamental role in earth systems

- Energy flow connects and unifies earth systems, controls climate
- Human energy usage affects earth systems and climate

**Climate**: the statistical distribution of meteorological phenomena (temperature, precipitation, *etc.*) over long times.

Climate depends on Earth’s surface, atmosphere, hydrosphere (water), cryosphere (ice), and biosphere.

Human activity impacts all these “spheres”

In next 3 lectures, study energy-climate connections.

I. Earth atmosphere and the greenhouse effect (today)
II. The carbon cycle, feedbacks and climate history
III. Projections, consequences, and mitigation
Earth climate: simplified models — insolation and albedo

**Albedo:** $\alpha = \text{fraction of reflected incident solar energy}$

- Water, forest, etc.: low $\alpha$
- Desert, ice, sheep, etc.: high $\alpha$

*Earth in current state: $\alpha \approx 0.3$*

[~0.06 atmosphere, ~0.1 surface, ~0.15 clouds]
Earth climate–insolation and albedo

Start with some simple models to understand warming mechanism

**Simplest model:** average over surface, only O$_2$, N$_2$ in atmosphere
no atmospheric absorption of solar or IR radiation

\[
I = (1 - \alpha) I_0 \approx 0.84 I_0 \quad \text{w/o clouds}
\]
\[
I_0 = \langle I_{\text{in}} \rangle = 343 \text{ W/m}^2 = 1370 \text{ W/m}^2 \times \frac{\pi R^2}{4\pi R^2}
\]

In radiative equilibrium

\[
\sigma T_s^4 = 0.84 I_0 \approx 288 \text{ W/m}^2
\]
\[
\Rightarrow T_s \approx 267 \text{ K} = -6^\circ \text{C}
\]

(Actual temperature near surface \( \approx 287 \text{ K} \))

[Note: in reality this situation \( \rightarrow \) feedback: more ice \( \rightarrow \) \( \alpha \uparrow \rightarrow \) colder]
**Greenhouse effect**

Include IR absorption in atmosphere by greenhouse gases (GHG) \((\text{H}_2\text{O}, \text{CO}_2, \ldots)\)

**Simple model:**
- 1 layer, perfect IR absorption

$$
\begin{align*}
\sigma T_a^4 &= (1 - \alpha)I_0 \\
\sigma T_s^4 &= 2\sigma T_a^4 = 2(1 - \alpha)I_0 \\
T_a &\approx 267\text{K} \\
T_s &\approx \sqrt[4]{2} \times 267\text{K} \approx 318\text{K}
\end{align*}
$$

[don’t take “toy model” temperatures seriously; just qualitative effects]
Greenhouse effect: n level model

As greenhouse gases increase, IR absorbed/emitted multiple times

Model with 2 IR absorbing layers

\[
\sigma T_1^4 = I = (1 - \alpha) I_0 \\
\sigma T_2^4 = 2 \sigma T_1^4 \\
\sigma T_s^4 = \sigma (2T_2^4 - T_1^4) = 3 \sigma T_1^4 \\
T_s \approx \sqrt[4]{3} \times 267 K \approx 351 K
\]

Model with n IR absorbing layers: \( T_s = (\sqrt[4]{n+1}) \times 267 K \)

So temperature increases with more GHG—but unrealistically fast

Really, weaker dependence on GHG— not all IR absorbed
**Greenhouse effect: absorption bands**

GHG only absorb some frequencies

Model: layers absorb 1/2 IR

\[
\frac{1}{2} \sigma T_s^4 \rightarrow T_1 \\
\frac{1}{2} \sigma T_s^4 \rightarrow T_2 \\
I = 0.84 I_0 \rightarrow T_s
\]

• Better model, GHG dependence still incomplete

• Real atmosphere: continuous distribution, absorption bands, convection, ...
To go further . . .

Need to understand something about atmospheres

- Where is material? (density distribution)
- What is temperature profile?
- How and where does atmosphere absorb/emit radiation
- What are the effects of convection + water vapor?

We will go over some basics,

For more details: Hartmann, “Global Physical Climatology”
Atmosphere: hydrostatic equilibrium

Assume $\rho, p$ in static equilibrium, no radiation/absorption

Pressure depends on weight of air above

$$z \uparrow: \quad dp = -g\rho dz$$

Combine with ideal gas law

$$pV = Nk_B T = V \frac{\rho}{m} k_B T \quad \Rightarrow \quad \rho = \frac{mp}{k_B T}$$

so $dp/p = -dz/H$ where $H = k_B T/mg =$ “scale height”

(pressure down by $1/e$ every $H$ in constant $T$ region;
  earth atm. $\approx 290$ K, $m \approx 29$ u: $H \sim 8.5$ km near surface, decreases w/T)

Solution:

$$p = p_0 e^{-\int_0^z \frac{dz}{H}}$$

[Note: for constant $T$, $\rho = \rho_0 e^{-mgz/k_B T} \sim$ Boltzmann!]
Atmosphere: lapse rate

Lapse rate: \( \Gamma = -\frac{dT}{dz} \) (rate of \( T \) change with altitude)

Consider motion of packet of air in hydrostatic equilibrium

First law: \( C_V dT + p \, dV = 0 \) (adiabatic)

[packet up, \( dV > 0 \) (expands), \( dT < 0 \) (cools)]

\[
p \, dV + V \, dp = Nk_B \, dT = (C_p - C_v) \, dT
\]

\[
\rightarrow C_p \, dT = V \, dp = -V \, g \rho \, dz
\]

so

\[
\frac{dT}{dz} = -\frac{V g \rho}{C_p} = -\frac{g}{c_p} = -\Gamma_d \quad \text{“adiabatic lapse rate”}
\]

Near surface: \( c_p \sim 1 \), \( \Gamma_d \approx 9.8^\circ \text{C/km} = \text{dry adiabatic lapse rate @ } 290 \text{ K} \)

\( \Gamma > \Gamma_d \Rightarrow \text{convective instability. } \ H_2O \text{ vapor } \rightarrow \Gamma_d \text{ decreases} \)

Troposphere: bottom \( \sim 10 \text{ km} \); global mean \( \Gamma \approx 6.5^\circ \text{C/km} \)

Heat transfer primarily convective, \( \Gamma \sim 3 - 10^\circ \text{C/km} \)
Atmospheric absorption

Molecules in atmosphere absorb in different parts of spectrum

**IR:** $\text{H}_2\text{O}, \text{CO}_2, \text{O}_3, \text{CH}_4, \text{N}_2\text{O}, \ldots$ in troposphere ($< \sim 15$ km)

have vibration-rotation spectra in thermal IR region

lines broadened (Doppler, pressure, QM) $\rightarrow$ bands

Small changes in minority constituents $\rightarrow$ affect T, climate

**UV:** $\text{O}_2, \text{O}_3$ in stratosphere ($\sim 50$ km)

$\text{O}_2$ dissociates with $< 246$ nm, (then $\text{O} + \text{O}_2 + \text{M} \rightarrow \text{O}_3 + \text{M}$)

$\text{O}_3$ has dissociation (Hartley, 200nm-300nm) in UV

Lambert-Beer absorption

\[
\frac{dI}{dz} = -\kappa I
\]

\[
\kappa = \rho k, \quad \rho = \text{density of absorber}, \quad k = \text{absorption x-section}
\]

(includes $z$-dependent $\rho$, most absorption $\tau \sim 1$)

Optical depth

\[
\tau = \int_{z}^{\infty} k\rho \, dz; \quad I = I_0 e^{-\tau}
\]
Putting it all together: 1D radiative-convective equilibrium

1D atmosphere model

— Lower atmosphere + minority constituents \( p = p_0 e^{-\int_0^z dz/H} \)
— Ozone 30 – 80 km (photochemical production)
— H\(_2\)O vapor \( \Rightarrow \) critical lapse rate

Integrate using Schwarzchild’s equation

\[
dI_\nu = (-I_\nu + B_\nu(T))k_\nu \rho \, dz
\]

\[
B_\nu(T) = \frac{2\pi h\nu^3}{c^2} \frac{1}{e^{\nu / k_B T} - 1}
\]

Simplifications: bands, LR

Cloudless atmosphere [Hartmann]
Critical LR 6.5\(^\circ\)C/km

Major component missing: clouds— less understood
More detailed modeling entails clouds, latitudinal variation, dynamics, including oceanic flow, etc. ⇒ General Circulation Models (GCM)

Actual global average radiation budget

Modeling is useful in understanding how changes affect balance.
Can now define a key concept in the science of climate change

**Radiative Forcing:**

“The change in net (down minus up) irradiance (solar + longwave; in W/m²) at the tropopause after allowing for stratospheric temperatures to readjust to radiative equilibrium, but with surface and tropospheric temperatures and state held fixed at the unperturbed values”

[IPCC 2001 + 2007 reports]

- Measured in W/m²
- RF $F$ related to $\Delta T$ by $\Delta T = \sigma F$ (assume linear response)
- Easy to calculate and compare $F$
- But not a full measure of climate change
CO₂ 15 micron band

Radiative forcing grows logarithmically in CO₂ level

\[ F \sim F_* + c \log_2 (M_{CO₂}/M_*) \]

[Temperature change ΔT also grows logarithmically if linear feedback.]

Why? Density increases → saturated band grows logarithmically

Given current understanding of atmosphere, logarithmic effect of increased CO₂ can be computed fairly accurately

Best estimate: \( c \approx 3.7 \text{ W/m}^2 \)
Radiative forcing by CO₂

- RF (in W/m²) quantitative measure of effect of increasing CO₂
- Expect RF of CO₂ depends logarithmically on CO₂

Define

\[ F_{2x} = \text{radiative forcing from doubling CO}_2 \ (e.g. \ 280 \text{ ppm} \rightarrow 560 \text{ ppm}) \]

Global models \( \Rightarrow F_{2x} \approx 3.7 \text{ W/m}^2 \)

Models say: if all else held fixed

\[ F_{2x} \approx 3.7 \text{ W/m}^2 \Rightarrow \Delta T_s \approx 1.2^\circ \text{C} \]

But feedbacks \( \rightarrow \Delta T_s \approx \sigma F_{2x} \)

Key question:
what is climate sensitivity \( \sigma \)

Simple computation:
If \( T = 255 \text{ K} \) (fits \( \alpha = 0.3 \)), what \( \Delta T \rightarrow 3.7 \text{ W/m}^2 \)?

\[ \sigma T^4 \approx 240.1 \text{ W/m}^2 \]
\[ \sigma (T + 0.98)^4 \approx 243.8 \text{ W/m}^2 \]
We have considered the theory, now let’s look at the data

Conclusive evidence: anthropogenic CO$_2$ increase

pre-industrial: 280 ppm, current: 391 ppm, increase: $\sim$ 2 ppm/year
SUMMARY

- Earth albedo $\alpha \approx 0.3 = \text{fraction of reflected solar energy}$
- Greenhouse effect raises Earth temperature
- Scale height $H = k_B T/mg = \text{vertical distance for pressure decrease by } 1/e$
- Lapse rate $\Gamma = -dT/dz = \text{rate of temperature decrease}.$
  Dry adiabatic lapse rate $\Gamma_d \approx 9.8^\circ \text{C/km}.$
  Water vapor decreases $\Gamma_d$
  $\Gamma > \Gamma_d \Rightarrow \text{convective instability}$
- Radiative forcing $= \text{increase in downward radiation at tropopause without including climate feedback}$
  Radiative forcing from doubling CO$_2$ $F_{2x} \approx 3.7 \text{ W/m}^2$