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.

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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
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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: $\alpha \approx 0.3$
[\sim0.06 \text{ atmosphere}, \sim0.1 \text{ surface}, \sim0.15 \text{ clouds}]
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Low albedo

High albedo
Simplified climate models

Earth atmosphere

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

\[
I = (1 - \alpha)I_0 \approx 0.84 \, I_0 \text{ w/o clouds}
\]

\[
I_0 = \langle I_{in} \rangle = 343 \, \text{W/m}^2 = 1370 \, \text{W/m}^2 \times \frac{\pi R^2}{4\pi R^2}
\]

In radiative equilibrium

\[
\sigma T_s^4 = 0.84I_0 \approx 288 \, \text{W/m}^2
\]

\[
\Rightarrow T_s \approx 267 \, \text{K} = -6 \, \text{°C}
\]

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

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

\[
\begin{align*}
0.84I_0 & \rightarrow T_a \\
T_s & \approx 267K \\
\sigma T_s^4 & \approx 2(1 - \alpha)I_0 \\
T_a & \approx 267K \\
T_s & \approx \sqrt[4]{2} 267K \approx 318K
\end{align*}
\]

[don’t take “toy model” temperatures seriously; just qualitative effects]
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Svante Arrhenius (1859-1927)
1896: predicted \(CO_2 \rightarrow\) warming
Greenhouse effect: n level model

As greenhouse gases increase, IR absorbed/emitted multiple times

Model with 2 IR absorbing layers

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

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

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

Equations: HW

- Better model, GHG dependence still incomplete
- Real atmosphere: continuous distribution, absorption bands, convection, ...
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\[ I = 0.84I_0 \]

\[ \frac{1}{2} \sigma T_s^4 \]

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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

\[ pV = N k_B T = V \frac{\rho}{m} k_B T \quad \Rightarrow \quad \rho = \frac{mp}{k_B T} \]

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. \( \approx 290 \text{ K}, \ m \approx 29 \text{ u}: \ H \approx 8.5 \text{ km near surface, decreases w/T} \)

Solution:

\[ p = p_0 e^{-\int_0^z dz/H} \]

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

Lapse rate: $\Gamma = -dT/dz$  (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)]

\[
p \, dV + V \, dp = N k_B \, dT = (C_p - C_v) \, dT
\]
\[
\rightarrow C_p \, dT = V \, dp = -V \, g \rho \, dz
\]

so

\[
\frac{dT}{dz} = \frac{V g \rho}{C_p} = -\frac{g}{c_p} = -\Gamma_d \quad \text{“adiabatic lapse rate”}
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Near surface: $c_p \sim 1, \quad \Gamma_d \approx 9.8^\circ \text{C/km} = \text{dry adiabatic lapse rate @ 290 K}$

$\Gamma > \Gamma_d \Rightarrow \text{convective instability.} \quad \text{H}_2\text{O vapor} \rightarrow \Gamma_d \text{ decreases}$

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

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

Molecules in atmosphere absorb in different parts of spectrum

**IR:** \(H_2O, CO_2, O_3, CH_4, N_2O, \ldots\) in troposphere \((<\sim 15 \text{ km})\) 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:** \(O_2, O_3\) in stratosphere \((\sim 50 \text{ km})\)
\(O_2\) dissociates w/ < 246 nm, (then \(O + O_2 + M \rightarrow O_3 + M\))
\(O_3\) has dissociation (Hartley, 200nm-300nm) in UV

Lambert-Beer absorption
\[
\frac{dI}{dz} = -\kappa I_0
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\(\kappa = \rho k\), \(\rho =\) density of absorber, \(k =\) absorption x-section
(includes \(z\)-dependent \(\rho\), most absorption \(\tau \sim 1\))

Optical depth
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\tau = \int_{z}^{\infty} k\rho \, dz; \quad I = I_0 e^{-\tau}
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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)
— $\text{H}_2\text{O}$ vapor $\Rightarrow$ critical lapse rate

Integrate using Schwarzschild’s equation

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

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

Simplifications: bands, LR

Cloudless atmosphere [Hartmann]
Critical LR 6.5°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
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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/m\(^2\)) 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/m\(^2\)
- 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
Simplified climate models
Earth atmosphere

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CO$_2$ 15 micron band

Radiative forcing grows logarithmically in CO$_2$ level

$$F \sim F_* + c \log_2 (M_{CO_2}/M_*)$$

[Temperature change $\Delta T$ also grows logarithmically if linear feedback.]

Why? Density increases $\rightarrow$ saturated band grows logarithmically

Given current understanding of atmosphere, logarithmic effect of increased CO$_2$ can be computed fairly accurately

Best estimate: $c \approx 3.7$ W/m$^2$
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Radiative forcing by CO₂

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

Define

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

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

Models say: if all else held fixed
\[ F_{2x} \approx 3.7 \text{ W/m}^2 \Rightarrow \Delta T_s \approx 1.2°C \]

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

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

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

\[ \sigma T^4 \approx 240.1 \text{ W/m}^2 \]
\[ \sigma (T + 0.98)^4 \approx 243.8 \text{ W/m}^2 \]
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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**

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- Greenhouse effect raises Earth temperature

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