The Earth as a Pair of Heat Engines

How Has the Earth Remained Habitable?

How it is that we are around to study the geological records of our few-billion-year evolution? How is it that the earth-surface temperatures have always remained within life’s rather narrow tolerances for at least 3.5 Ga despite a relatively huge 20% increase in the Sun’s luminosity over that time? Answering this question is a good starting point for studying the our planet’s evolution because it points up how the Earth System responds to changes in the Sun, organisms, the atmosphere, continents, and oceans interrelate in shaping our environment.

The Earth’s Heat Engines

A heat engine is a mechanism that transfers heat energy from a heat source to a heat sink by way of a working fluid and in the process extracts work. The working substance is cycled between the source and sink, the same material coming back for another load of energy after it has dumped the one before.

The Earth is two heat engines in one: the exogenic (“outside-generated”) heat engine of the atmosphere, hydrosphere, and crust; and the endogenic (“inside-generated”) heat engine of the core, mantle, and lithosphere. The exogenic heat engine’s energy source is the Sun; its heat sink is outer space; its working substances are the atmosphere and hydrosphere and, most important, the water they contain. The endogenic heat engine’s main energy source is the mantle’s radioactivity; its heat sink is the exogenic heat engine; and its working fluids are the outer core’s molten iron and the sublithospheric mantle’s fluid-like rock. The engines’ work is the rearrangement of earth materials.

The Global Heat Balance and Mean Global Surface Temperature

Why Do We Need to Know about Surface Temperature? If the Earth’s average surface temperature were 2-3°C lower, as it was a few thousand years ago, the Cornell campus (not to mention much of Europe and North America) would be buried beneath a continental ice sheet. Small changes in surface temperature make huge differences in the way water behaves and does geological work.

Life as we know it is confined between water’s freezing and boiling points. Temperature further restricts organisms by determining the rates of chemical reactions; for most organisms, unlike ourselves, take on the temperature of their surroundings. At earth-surface temperatures, a 10°C increase typically increases chemical reaction rates 1.5-5 times. You can see why environmental temperature is generally the single most important factor determining species’ geographic distributions: if temperatures fall outside the range to which the organism is adapted, the complex series of biochemical reactions that maintain the organism will not work; products from one reaction will not be delivered on schedule to join with products from other reactions as the reactants in the next reaction in the chain.

What Determines Surface Temperature? The Earth’s energy budget is very nearly balanced over the year: income essentially equals outgo; gains and losses show up as tiny changes in average temperature. Though energy goes through many transformations within the endogenic and exogenic heat engines, it enters and leaves the planet almost entirely as light. Thus, we can write down the Earth’s energy balance in terms of light absorption and reemission.

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1 As you remember from last time, earth scientists find it convenient to denote ages in years (ans) with the abbreviations ka (kiloans, as in kilobyte, 1 ka = 1000 years), Ma (megans, as in megabyte, 1 Ma = 1 million years), and Ga (gigans, as in gigabyte, 1 Ga = 1 billion years).

2 No doubt one of the most over-used words in the English language ever since the popularization of general systems theory in the 1950s and ‘60s, the term “system” is used here in the sense of a group of elements that stand in interaction to one another such that perturbing any one of them may bring about a response in any or all of the others.
As will be explained, the Earth’s overall surface temperature is determined by three quantities:

\[
\text{Surface Temperature} \propto [\text{Insolation} \cdot (1 - \text{Albedo}) \cdot (1 + \text{Atmosphere’s Optical Thickness})]^{1/4} \tag{2.10}
\]

1. **Insolation.** The solar power per unit area at the top of the atmosphere as measured perpendicular to the solar beam. Endogenic heat flow from the Earth’s interior, about 0.02% of the insolation, is negligible in comparison.

2. The Earth’s **albedo**, the fraction of solar power reflected back into space.

3. The Earth’s light absorption and reemission characteristics, a measured by the atmosphere’s **optical thickness**.

Here we will develop a simple, first-order statement of the energy balance which will enable us to assess the effects of the most major changes on global surface temperature— the increase in solar luminosity, the rise of oxygen-producing photosynthesizers, the impacts of massive bolides, and the spread of vegetation across continents.

**Thermal Radiation** — At high temperatures, bodies radiate light in a practically continuous spectrum of wavelengths. To conceptualize thermal radiation, 19th Century physicists invented the black body, a hypothetical entity that emits radiation in a perfectly continuous spectrum. Though obviously not black, the Sun, the Earth, the ocean, and the atmosphere behave very nearly as black-body radiators.

The intensity and spectrum of radiation a black body emits is dependent on its temperature. The spectrum of wavelengths follows the Planck distribution (a skewed, bell-shaped curve). The modal wavelength is inversely proportional to the absolute temperature, which means that a red-hot black body glows bluer as it gets hotter.  

A body’s **excitance** — the power it emits per unit area — is extremely sensitive to absolute temperature, as the fourth-power dependence in Stefan’s Law shows:

\[
\text{Excitance} \propto (\text{Absolute Temperature})^4 \tag{2.1}
\]

The Sun’s light-emitting region behaves nearly as a black body with a temperature of about 5785 K and radiates largely in the visible region with a modal wavelength in blue-green (about 500·10^{-9} m). Being much cooler, the Earth is a less ideal black body radiator. Its radiation, predominantly infrared, comes primarily from the uppermost atmosphere. Lower atmospheric layers absorb these wavelengths so strongly that they are effectively opaque to the radiation lost to space. Owing to preferential absorption of certain wavelengths by so-called greenhouse gases such as carbon dioxide, the Earth radiates different parts of the spectrum as it were at somewhat different temperatures, roughly 260 ± 30 K (-13 ± 30°C) depending on the wavelength, as satellite observations show.

**Insolation** — The intensity of light that reaches us from a distant source such as the Sun decreases inversely with the square of the separation. To a very good approximation, the Sun is a spherical source that radiates energy equally in all directions. The total power it radiates is its **luminosity**. Because outer space is a near-perfect vacuum, the total flux of sunlight at a given distance from the source remains essentially constant as energy spreads out in all directions. That is, the total flux passing through a sphere centered on the Sun remains constant, so that:

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3 This generalization is known as Wien’s Displacement Law.
Light Intensity = \frac{\text{Sun's Luminosity}}{\text{Sphere's Area}} \quad (2.2a)

\begin{align*}
\text{Area Surfaces' Sphere} &= \text{Photosphere's Excitance} \cdot \frac{\text{Photosphere's Surface Area}}{\text{Sphere's Surface Area}} \quad (2.2b) \\
&= \text{Photosphere's Excitance} \cdot \left[ \frac{\text{Sun's Radius}}{\text{Distance from Sun's Center}} \right]^2 \quad (2.2c)
\end{align*}

The insolation is simply the light intensity in (2.2) at the Earth’s distance from the Sun. Relatively tiny, seasonal changes in insolation due to the slight ellipticity of the Earth’s orbit have a significant effect on climate, as will be discussed in Lecture 15.

**The Earth’s Effective Temperature** — For all practical purposes, the Earth is in thermal equilibrium: the solar power it absorbs equals the power it radiates back into space. The balance between income and outgo determines the Earth’s effective temperature (the Earth’s temperature as a black-body radiator, which, for practical purposes is the temperature of the outmost atmosphere, where nearly all of the Earth’s radiation takes place). The solar power absorbed is broken down in terms of the insolation (sunlight intensity at the atmosphere’s outer edge) and the planetary albedo (the fraction of power absorbed):

\begin{align*}
\text{Power Absorbed} &= \text{Insolation} \cdot (1 - \text{Albedo}) \cdot \text{Earth’s Cross-Sectional Area} \quad (2.3a) \\
&\propto \text{Insolation} \cdot (1 - \text{Albedo}) \cdot (\text{Earth’s Radius})^2 \quad (2.3b)
\end{align*}

Stefan’s Law (2.1) gives the power that the Earth radiates:

\begin{align*}
\text{Power Radiated} &= \text{Earth’s Surface Area} \cdot \text{Earth’s Excitance} \quad (2.4a) \\
&\propto (\text{Earth’s Radius})^2 \cdot (\text{Effective Temperature})^4 \quad (2.4b)
\end{align*}

Equating the power absorbed with the power radiated, one finds

\begin{equation}
\text{Effective Temperature} \propto [\text{Insolation} \cdot (1 - \text{Albedo})]^{1/4} \quad (2.5)
\end{equation}

Today, the insolation is about 1368 W/m², the planetary albedo is about 0.29, and the Earth’s effective temperature is about 256 K (-17° C).

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4 The distance from the Sun’s center is assumed greater than or equal to the Sun’s radius.
Atmospheric Light Absorption — For a simple description of how the atmosphere absorbs sunlight, we might imagine it to be made up of discrete layers like panes of glass. As light passes through a given pane, a certain fraction of its energy is absorbed. A pane’s optical thickness is measured in terms of the fraction of energy it transmits. A pane defined to have an optical thickness of unity transmits the fraction 1/e (or 36.79%) of the incident light, where e = 2.718 is the base of the natural logarithms. Suppose a light beam passes through a series of panes, each of unit optical thickness. The beam’s intensity falls to 1/e after passing through the first pane, to 1/e² after passing through the second, to 1/e³ after passing through the third, and so on. One sees that light intensity decreases exponentially with optical depth:

\[ \text{Light Intensity} = \text{Initial Intensity} \cdot e^{-\text{Optical Depth}} \]  

(2.6)

It is convenient to measure the overall intensity of the greenhouse effect (the relative amount of solar energy the atmosphere absorbs) as an optical depth. The modern atmosphere’s effective optical depth is about 0.60 (except around big cities and other smoggy places, where it is higher).

Atmospheric Light Absorption and Reemission — Imagine the atmosphere as a stack of glass plates with sunlight shining directly down through them. Each plate absorbs energy and, in addition, radiates as a black body. Light passes up and down through the stack. The top plate radiates energy back into space. This plate’s excitance corresponds to the Earth’s excitance, its temperature to the Earth’s effective temperature.

We assume that the atmosphere is in thermal equilibrium. After some mathematical manipulation, one finds that the excitance at a given optical depth in the atmosphere is

\[ \text{Atmosphere’s Excitance} = \text{Earth’s Excitance} \cdot (1 + \text{Optical Depth}) \]  

(2.7)

Applying Stefan’s Law (2.1) again, we find atmospheric temperature as a function of optical depth and the Earth’s effective temperature:

\[ \text{Atmosphere’s Temperature} = \text{Earth’s Effective Temperature} \cdot (1 + \text{Optical Depth})^{1/4} \]  

(2.8)

The fourth-root dependence indicates that that temperature increases very, very — four times “very” — insensitively with increasing optical depth from the atmosphere’s outer fringe. Notice that we do not assume that optical depth is directly proportional to the physical depth as measured from the top of the atmosphere, for indeed it is not. The atmosphere’s physical density increases with increasing physical depth from the top because the weight of overlying layers compresses the gas in layers below. Lower layers absorb relatively more energy because their molecules are more densely packed. One might say that progressively lower layers have a progressively deeper tint, that is, a progressively higher optical density.

Our simple model leads to a reasonable first-order description of atmospheric temperature. An important omission is atmospheric flow; for instance, clouds and storms. Flow, however, turns out to be important only in the 10 km or so nearest the ground. The upper parts of the atmosphere, with which we are primarily concerned here, are more or less stably stratified and, for our purposes, really do behave rather like a stack of glass plates.

Surface Temperature — Plugging in the atmosphere’s overall optical thickness as the optical depth in (2.8) gives

\[ \text{Surface Temperature} = \text{Effective Temperature} \cdot (1 + \text{Atmosphere’s Optical Thickness})^{1/4} \]  

(2.9)

Knowing the Earth’s effective temperature (256 K) and its mean surface temperature (288 K), we find the modern atmosphere’s apparent optical thickness to be about 0.60.
Defining an optical depth for the whole atmosphere across the whole spectrum of radiation is, to be sure, a gross generalization; but it is a very useful one. For our purposes, optical depth is a useful way to characterize atmospheric composition. An increase or decrease in atmospheric greenhouse gases, for instance, can be thought of as an increase or decrease in the atmosphere’s optical depth.

Finally, substituting (2.5) into (2.9) gives the basic relationship of surface temperature to the three factors that determine it:

\[
\text{Surface Temperature} \propto \left[ \text{Insolation} \cdot (1 - \text{Albedo}) \cdot (1 + \text{Atmosphere’s Optical Depth}) \right]^{1/4}
\]

(2.10)

Consider what this expression says about the Earth System’s operation, integration, and evolution: the character of life’s environment is determined by a complex of interrelated astronomical, atmospheric, oceanic, plate-tectonic, and biological factors.
The “Dim Sun” Paradox: How Come Life Wasn’t Frozen Out?

For thinking about the ancient Earth, it is convenient to rewrite the expression for mean surface temperature (2.10) in terms of the quantities’ present values:

\[
\text{Surface Temperature} = \frac{288 \text{ K}}{[\text{Insolation} \cdot \frac{1 - \text{Albedo}}{\text{Present Insolation}}] \cdot \frac{1 + \text{Optical Thickness}}{1 + 0.60}]^{1/4}
\]

(2.11)

The Paradox — For reasons to be explained in the next lecture, insolation has increased to about 1.2 times its value about 3.5 Ga ago, when the earliest known fossils occur. Supposing the albedo and the atmosphere’s optical depth were the same as today’s, what was the surface temperature at that time? Equation (2.11) gives this answer: \(288 \text{ K} \cdot (1/1.2)^{1/4} = 275 \text{ K} = 2^\circ \text{ C}\). Would the early Earth have been ice-covered? Yes. Does our retrospective prediction square with the the abundance of fossils and water-laid sediment over the last 3.5 Ga? No, it does not. An Earth with a 2\(^\circ\) C mean surface temperature would be in the process of freezing over and getting colder still. Its albedo would necessarily be higher than the present 0.29, and its temperature would necessarily be lower than 2\(^\circ\) C, assuming that the atmosphere’s optical depth was 0.60, as it is today. Hence the paradox.

Resolving the Paradox — Contrary to expectation, the Earth has always been a “greenhouse” and never an “ice house” (although, as we’ll see later on, it may have come close 600-900 Ma ago). Geological evidence of liquid water and life are abundant at all ages. Continental ice sheets are known only from a few periods; they seem to have formed whenever a continent has been at a pole. Glacial ages are another indication that average surface temperature probably has always hovered somewhere not far above water’s freezing point.

Suppose that the temperature 3.5 Ga ago were today’s 288 K. Supposing only the albedo were different, what must the albedo has been at the time? \([1 -1.2(1 - 0.29) = 0.15]\) What factors might have contributed to a lower albedo? [Less ice, fewer clouds, and the smaller continents deduced from geochemical evidence] If the Earth were cloud-free and entirely water-covered, its albedo could conceivably be as low as 0.1. Might albedo alone have compensated for reduced insolation? [Probably not] Supposing only the atmosphere’s optical depth were different, what must that depth have been? \([1.2 (1 + 0.6) - 1 = 0.92]\) What factors might have contributed to increased optical depth? [More carbon dioxide, water vapor, and perhaps other greenhouse gases such as methane [CH\(_4\)] and ammonia [NH\(_3\)]; limitation of necessarily carbon-dioxide-consuming photosynthetic organisms… but by what?]

The Remaining Problem — Despite the change in insolation and all the conceivable changes in albedo and the atmosphere’s optical depth, surface temperature over most of the Earth appears to have remained well within limits tolerable by life — a intriguing coincidence coming lectures will explore.
Study Questions

1. What is a heat engine? What is the exogenic heat engine? The endogenic heat engine? What are these engines’ respective heat sources, heat sinks, and main working fluids? Why is it helpful to conceptualize the Earth in these terms for attempting to address questions like the “dim sun” paradox?

2. What three factors determine the Earth’s overall surface temperature? Why can we write out the Earth’s energy balance in terms of the absorption and reemission of light?

3. How would the Earth’s overall surface temperature respond to each of the following well-precedented short-term changes, all else remaining the same? A volcanic eruption or bolide impact that throws large amounts of highly reflective dust into the atmosphere? A drop in sea level that exposed extensive, previously water-covered, continental shelves as barren land? Freezing over of the Arctic Ocean? Spread of an ice sheet across a continent? Massive release of methane through melting of the Arctic’s permafrost?

4. How would the Earth’s surface temperature respond to each of the following long-term changes, all else remaining the same? Increase in the Sun’s luminosity? Continental growth, with resulting replacement of ocean by land? Molecule-for-molecule replacement of the atmosphere’s CO₂ by O₂ (a much less effective greenhouse gas) through the activity of photosynthetic organisms. Spread of light-absorbing vegetation across the previously barren land?

5. Snow and ice are by far the most reflective substances exposed over any large area of the Earth. How would incomplete melting of Canada’s winter snow during a cool summer tend to affect the Earth’s overall surface temperature? How would the temperature change tend to affect the melting of snow and ice during the following summer? Do you see how a continental ice sheet could begin?