What makes a planet habitable?
Atmospheric science basics
For last time Lecture 2 key points: From Earth history to the Circumstellar Habitable Zone concept

Reminder: Class website is http://geosci.uchicago.edu/~kite/geos32060_2019/

• Small imbalances between the geologic release rate of $p_{CO_2}$ and the geologic consumption rate of $p_{CO_2}$ would have seriously threatened Earth’s habitability.

• This strongly suggests a negative feedback has regulated $p_{CO_2}$ over geologic time.
  – In the last ~10 years, the evidence for a negative feedback has gotten stronger - some people would go further and say a weathering feedback is required to explain the geologic data.

• A candidate mechanism for the weathering feedback is carbonate-silicate weathering.

• The Habitable Zone is defined as the range of distances from a star within which a weathering feedback might operate.
Definitions of a habitable planet

(Operational) definition of habitable planet used in this course:
Habitable planet \(\approx\) maintains surface liquid water over timescales relevant to biological macroevolution

On Earth, long-term climate stability involves the nonlinear temperature dependence of greenhouse gas drawdown by weathering.

(Walker et al., JGR, 1981; Kasting et al., Icarus, 1993)

Stabilizing feedback:

\[ F_{GHG,in} \quad \rightarrow \quad \text{increase} \quad \rightarrow \quad \text{increase GHG concentration} \quad \rightarrow \quad \text{increase temperature} \quad \rightarrow \quad \text{stabilize GHG concentration} \quad \rightarrow \quad \text{increase} \quad W_R \]

Habitable zone

Dependent on which atmospheric volatiles are available
Feedbacks in the carbonate-silicate cycle

Increased CO₂ in atmosphere (or solar luminosity)
- Warmer atmosphere
- More H₂O in atmosphere
- More acidic rainwater

Precipitation of CaCO₃ lowers CO₂

Decreased CO₂ in atmosphere (or lower temperature)
- Cooler atmosphere
- Less H₂O in atmosphere
- Less acidic rainwater

Less precipitation of CaCO₃ raises CO₂

Increased flux of Ca⁺⁺ to oceans

Decreased flux of Ca⁺⁺ to oceans
• **Volunteers needed** for presentation of section 3.1-3.6 of Catling & Kasting on Tue 16 April

• **Homework 1** will be on website Saturday (you will get a notification email), and due in class on Thu 18 April

• **Office hours:** Thursdays, 11a-noon (after class). Hinds 467, or by appointment [kite@uchicago.edu](mailto:kite@uchicago.edu). Please email at least 48 hours in advance with a range of suggested times.
Lecture 3: Atmospheric science basics: Key points

• Describe and qualitatively explain vertical temperature structure of rocky-planet atmospheres.

• Apply elementary models of radiation balance.

• Explain the greenhouse effect in terms of vertical temperature structure and opacity at visible and thermal wavelengths.

• (Explain the theoretical basis for expecting an atmosphere-surface temperature offset).
Horizontal structure

JunoCam on JUNO spacecraft (Jupiter polar orbiter)
Vertical structure

LORRI (Long Range Reconnaissance Imager) / New Horizons
Figure 1.1 The nomenclature for vertical regions of the Earth’s atmosphere, shown schematically.
For Earth, sunlight (most energy in 0.1-1 μm wavelengths) is mostly absorbed at the ground. Thermal emission (most energy at 4-20 μm for habitable surfaces) is mostly re-radiated to space in the upper troposphere.

Figure 1.2 Thermal structure of the atmospheres of various planets of the Solar System. The dashed line at 0.1 allows you to see the feature of a common tropopause near ~0.1 bar for the thick atmospheres, despite the differences in atmospheric composition. See Robinson and Catling (2014) for sources of data.
Hydrostatic balance and atmospheric pressure vs. height

\[ \Delta p = \frac{\text{weight}}{\text{area}} = -(\text{mass of column}) \times \frac{g}{1} = -(\rho \Delta z)g \]

\[ \frac{\partial p}{\partial z} = -g(z)\rho(z), \quad p(z) = \int_{z}^{\infty} g(z)\rho(z)\,dz \]

\[ \rho = \frac{m}{kT} \] ideal gas law

\[ \frac{\partial p}{\partial z} = -\left(\frac{m(z)g(z)}{kT(z)}\right)p \quad \Rightarrow \quad \frac{\partial p}{p} = -\left(\frac{\partial z}{H}\right), \text{ where } H = \frac{kT(z)}{m(z)g(z)} = \frac{\bar{R}(z)T(z)}{g(z)} \]

\[ p = p_s \exp\left( -\int_{0}^{z} \left( \frac{dz}{H} \right) \right) \]
Atmospheric vertical thermal structure

DRY ADIABATIC LAPSE RATE

\[ \Gamma_a = -\left( \frac{dT}{dz} \right) = \frac{g}{c_p} \]

(1.27)

O(10^1) m/s^2 (for rocky planets, never or almost never
10^2 m/s^2, per Leslie Rogers’ work)

depends on atm. composition

This little equation provides the temperature change with altitude of a dry parcel of air moving up and down through an atmosphere in hydrostatic equilibrium.

Describes the troposphere under dry, convective conditions
Spectrum of blackbody radiation

Wien’s displacement law: \[ \lambda_{\text{max}} = \frac{b}{T} \]

\[ b = 2900 \text{ microns/K} \]

Consider the radiation field within a closed cavity that perfectly absorbs and emits radiation at all wavelengths. At equilibrium, production and loss of radiation balance and the intensity of the radiation field is uniform. Max Planck found the spectrum to be uniquely related to the cavity wall temperature, T, by the Planck function.
Figure 2.10  Blackbody curves of the Sun and Earth, where the solar flux density is that at 1 AU distance scaled by a factor of 1–$A_b$, where $A_b$ is Earth’s Bond albedo. The curves cross near ~4 μm. Solar and thermal spectra peak at ~0.5 and 10 μm, respectively.
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• **Apply elementary models of radiation balance.**

• Explain the greenhouse effect in terms of vertical temperature structure and opacity at visible and thermal wavelengths.

• (Explain the theoretical basis for expecting an atmosphere-surface temperature offset).
Definition of Bond/planetary albedo (used for climate calculations)

Kepler space telescope measures the dip in visible light from an exoplanet+star when the exoplanet passes behind the exoplanet’s host star.

(a) What is the phase angle for this measurement?
(b) Does this measurement constrain the Bond/planetary albedo or the geometric albedo?

Energy balance:
\[
(1 - A_b) \frac{S_p}{4} = \sigma T_{eq}^4
\]

Figure 2.7 Different types of reflection. (a) Pure specular reflection, where the angle of incidence equals the angle of reflection, e.g., a mirror. (b) A Lambertian surface reflects radiation evenly in all directions. This surface is to be diffusely reflecting.
Energy balance:

\[
(1 - A_b) \frac{S_p}{4} = \sigma T_{eq}^4
\]

Figure 2.5 (a) An elemental ring of surface defined by angle \( \phi \) on a planet of radius \( R \). The incident solar flux is \( S_\odot \). (b) The equivalent area of intercept for the solar flux is a projected disk of area \( \pi R^2 \) compared with a total sphere area of \( 4\pi R^2 \).

Table 9-2

<table>
<thead>
<tr>
<th></th>
<th>Mass of atmosphere kg/cm²</th>
<th>Distance from Sun 10⁶ km</th>
<th>Solar energy received 10⁶ watts/m²</th>
<th>Black-body temperature °C</th>
<th>Fraction sunlight reflected</th>
<th>Reflective cooling °C</th>
<th>Greenhouse warming °C</th>
<th>Actual surface temperature °C</th>
</tr>
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<tbody>
<tr>
<td>Mercury</td>
<td>0</td>
<td>58</td>
<td>9126</td>
<td>175</td>
<td>.068</td>
<td>-8</td>
<td></td>
<td>0</td>
</tr>
<tr>
<td>Venus</td>
<td>115°</td>
<td>108</td>
<td>2614</td>
<td>55</td>
<td>.90</td>
<td>-144</td>
<td>+553</td>
<td>464</td>
</tr>
<tr>
<td>Earth today</td>
<td>1.03**</td>
<td>150</td>
<td>1368</td>
<td>5</td>
<td>.30</td>
<td>-25</td>
<td>+35</td>
<td>15</td>
</tr>
<tr>
<td>Early Earth</td>
<td>150</td>
<td>958</td>
<td>-26</td>
<td>.30 (?)</td>
<td>-21</td>
<td>62(?)</td>
<td>15 (?)</td>
<td>-60</td>
</tr>
<tr>
<td>Mars</td>
<td>0.016*</td>
<td>228</td>
<td>589</td>
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<td>.25</td>
<td>-16</td>
<td>+3</td>
<td>-60</td>
</tr>
<tr>
<td>Moon</td>
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<td>150</td>
<td>1368</td>
<td>5</td>
<td>.11</td>
<td>-7</td>
<td></td>
<td>-160 – +130</td>
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</table>

~2/3 from H₂O; ~1/3 from CO₂

However, “H₂O is a slave to CO₂.”

Titan: complicated due to interplay of CH₄, H₂, and antigreenhouse haze

Langmuir & Broecker
The global mean surface temperature can be a lot less than the effective temperature. The arithmetic global mean surface temperature of Mars is estimated to be 202 K (Haberle, 2013), whereas the equilibrium temperature is 210 K (Table 2.2). This apparent paradox is resolved by realizing that one needs to compute the fourth power of the surface temperature $T_s^4$ at all locations on the planet and average this quantity globally and annually to obtain $\overline{T_s^4}$. Then, the global mean effective surface emission temperature $T_{se}$ and corresponding greenhouse effect are defined as follows:

$$T_{se} = \left[\overline{T_s^4}\right]^{1/4}, \quad \Delta T_g \equiv T_{se} - T_{eq} \quad (2.17)$$

For Mars, $T_{se}$ is 215 K so that the greenhouse effect is $\sim$5 K by eq. (2.17).
Top-of-atmosphere energy in = top-of-atmosphere energy out

is a very good approximation for the purposes of this course

• How quickly would Earth’s surface temperature double with a 1 W/m^2 (<1%) top-of-atmosphere energy imbalance?
• How quickly did Earth’s surface temperature return to normal following the Chicxulub impact (~15 km diameter asteroid, ~30 km/s)?
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Figure 3.5: Determination of a planet’s temperature by balancing absorbed solar energy against emitted longwave radiation. The horizontal line gives the absorbed solar energy per unit surface area, based on an albedo of .3 and a Solar constant of 1370 W/m². The OLR is given as a function of surface temperature. The upper curve assumes the atmosphere has no greenhouse effect ($p_{rad} = p_s$), while the lower OLR curve assumes $p_{rad}/p_s = .6$, a value appropriate to the present Earth.
Figure 3.6: Sketch illustrating how the greenhouse effect increases the surface temperature. In equilibrium, the outgoing radiation must remain equal to the absorbed solar radiation, so $T_{rad}$ stays constant. However, as more greenhouse gas is added to the atmosphere, $p_{rad}$ is reduced, so one must extrapolate temperature further along the adiabat to reach the surface.

Pierrehumbert, “Principles of planetary climate”
How to read a planet’s thermal emission spectrum

Earth, Niger valley, warm season

dashed lines: theoretical blackbody curves
solid line: spacecraft observations
chemical species: absorption bands

What could we conclude from this spectrum if we had obtained it for a habitable-zone 1-Earth-radius exoplanet?
ERBE = Earth Radiation Budget Experiment

Graph showing the upward flux (W/m²) at different latitudes.
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Relationship between surface temperature and atmospheric temperature.

Figure 2.17 A two-layer atmospheric model to understand the concept of atmospheric skin temperature. The lower atmosphere is represented by a layer that is optically thick in the infrared and emits at the effective temperature, $T_{\text{eff}}$. The upper atmosphere is represented by a layer that is optically thin in the infrared and has a temperature $T_{\text{skin}}$ and emissivity $\varepsilon$, which is equal to absorptivity by Kirchoff’s Law. By considering conservation of energy in the upper layer, we determine that $T_{\text{skin}} = 2^{3/8}T_{\text{eff}}$ (see text).
Relationship between surface temperature and atmospheric temperature

**Figure 2.18**
(a) Thought experiment with an infrared-transparent atmosphere and no convection. The surface reaches a temperature of the effective temperature, $T_{\text{eff}}$, while the optically thin atmosphere attains the skin temperature, $T_{\text{skin}}$. There is a temperature discontinuity at the surface. (b) Thought experiment with negligible greenhouse gases but convection. A convective profile sets in. The surface temperature $T_{s2}$ is $T_{\text{eff}}$. (c) Thought experiment with greenhouse gases and convection. The near-surface atmosphere is optically thick, so that emission at the effective temperature takes place at a higher level where the pressure is $p_{\text{emission}}$. The surface temperature $T_{s3}$ exceeds $T_{\text{eff}}$ because of a greenhouse effect.
Bonus slides