Climate Stability of Habitable Earth-like Planets

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Abstract

The carbon-silicate cycle regulates the atmospheric \(\text{CO}_2\) content of terrestrial planets on geological timescales through a balance between the rates of \(\text{CO}_2\) volcanic outgassing and planetary intake from rock weathering. It is thought to act as an efficient climatic thermostat on Earth and, by extension, on other habitable planets. If, however, the weathering rate increases with the atmospheric \(\text{CO}_2\) content, as expected on planets lacking land vascular plants, the carbon-silicate cycle feedback can become severely limited. Here we show that Earth-like planets receiving less sunlight than current Earth may no longer possess a stable warm climate but instead repeatedly cycle between unstable glaciated and deglaciated climatic states. This has implications for the search for life on exoplanets in the habitable zone of nearby stars.

1. Introduction

It is generally thought that the carbon-silicate cycle acts as a stabilizing feedback and a powerful thermostat for the Earth climate, guaranteeing surface liquid water conditions. Above freezing temperatures, rock weathering occurs faster at higher temperatures, which reduces the \(\text{CO}_2\) atmospheric partial pressure, \(p\text{CO}_2\), and cools down the climate. Conversely, on a frozen planet that temporarily lacks weathering, atmospheric \(\text{CO}_2\) builds up from continued volcanic outgassing, which warms up the climate until surface liquid water and weathering conditions are restored \([1, 2, 3]\). Such \(p\text{CO}_2\) build up is in fact the leading scenario for the deglaciation of Earth following a snowball event \([4, 5, 6]\). Generalizations of these concepts to Earth-like planets around other stars are central to the definition of their liquid water habitable zone \([2, 7, 8]\). In particular, planets subject to modest levels of
insolation are expected to achieve temperate conditions with liquid water at
the surface by building up massive enough $CO_2$ atmospheres [2, 3, 10].

Experimental data, theoretical arguments and paleoclimate modeling
suggest that the rate of $CO_2$ intake via rock weathering by a planet lacking
land vascular plants increases with the atmospheric $CO_2$ content, $pCO_2$ [3,
11, 12, 13, 14]. This feature of lifeless planets or planets with only primitive
forms of life is important because it will limit the buildup of $CO_2$ at high
values. As a result, the climate thermostat due to the carbon-silicate cycle
should become less efficient on weakly-insolated Earth-like planets located
in the outer regions of the habitable zone.

2. Climate-Weathering Models

To address this issue quantitatively, we model Earth-like climates with a
zero-dimensional, energy balance model that equates the net insolation and
thermal radiation fluxes,

$$\frac{S}{4} [1 - \alpha(T_{surf}, pCO_2)] = OLR(T_{surf}, pCO_2),$$

where $S$ is the insolation flux, $\alpha$ is the planetary Bond albedo and $OLR$
is the outgoing longwave radiation flux emitted by the planet. $OLR$ and
$\alpha$ are functions of the surface temperature, $T_{surf}$, and $pCO_2$, derived from
radiative-convective climate models [9] (see Appendix A for details).

While we assume that the climate reaches thermal equilibrium rapidly,
by virtue of Equation (1), the slower $CO_2$ compositional equilibrium is not
imposed a priori in our models. Rather, $pCO_2$ is evolved on the relevant
geological timescales according to

$$\frac{d}{dt} pCO_2 = V - W(T_{surf}, pCO_2),$$

where $V$ is the global $CO_2$ volcanic outgassing rate (estimated as $V_\oplus = 7$ bars/Gyr for Earth [15]) and $W$ is the rate of $CO_2$ intake by the solid
planet via rock weathering. The functional form of $W$ is adapted from
Earth studies for pre-vascular plant conditions [14]:

$$\frac{W}{W_\oplus} = \left(\frac{pCO_2}{p_\oplus}\right)^{\beta} e^{k_{act}(T_{surf} - 288)} \times \left[1 + k_{run}(T_{surf} - 288)\right]^{0.65},$$

where $p_\oplus = 330\mu$bar is the pre-industrial $pCO_2$ level, $W = W_\oplus \equiv V_\oplus$ for $T_{surf} = 288K$, $k_{act} = 0.09$ is related to an activation energy and $k_{run} =$
0.045 is a runoff efficiency factor. Varying $k_{\text{act}}$ in the range 0.06–0.135 and $k_{\text{run}}$ in the range 0.025–0.045 has only a minor quantitative impact on our results. Values of $\beta = 0.25–1$ have been considered in the literature for the dependence of weathering on $pCO_2$ in the absence of land vascular plants. We use $\beta = 0.5$ (default) and 0.25 in this work. Note that equations (1) and (2) are coupled through $pCO_2$ and $T_{\text{surf}}$.

3. Climate Solutions

3.1. Steady-State Solutions

Steady-state climate solutions, satisfying both radiative (Equation (1)) and weathering equilibrium ($d/dt \equiv 0$ in Equation (2)) are represented in Figure 1 by intersecting cooling and heating curves. Solid red lines in Figure 1 show the albedo-corrected insolation flux (LHS of Equation (1)) received by a planet located at 1, 1.25 and 1.6 AU from a Sun-like star as a function of surface temperature. Net insolation drops precipitously from 273K to 263K as the planetary surface freezes and the albedo reaches $\sim 0.65$. The various blue lines in Figure 1 represent the $OLR$ cooling flux (RHS of Equation (1)) according to various scenarios for the atmospheric $CO_2$ content.

A standard model with $pCO_2$ arbitrarily fixed at $p_{\oplus} = 330 \mu \text{bar}$ (i.e. not constrained by Equation (2)) is represented by the slanted solid blue line. As is well known, three steady-state climate solutions exist in such a model for an Earth-like planet at 1AU, as indicated by diamonds where the red and blue curves intersect. Earth’s current climate ($T_{\text{surf}} \approx 288 \text{ K}$) and a globally-frozen state ($T_{\text{surf}} \approx 229 \text{ K}$) are both stable, while the intermediate state ($T_{\text{surf}} \approx 270 \text{ K}$) is thermally unstable.

However, when $pCO_2$ is also required to satisfy weathering equilibrium (Equation (2)), steady-state climate+weathering solutions only exist above freezing temperatures since weathering stops operating on a frozen planet. A model including weathering without any $pCO_2$ dependence ($\beta = 0$ in Equation (3)) is represented by the vertical solid blue line in Figure 1. In such a model, weathering equilibrium enforces a unique surface temperature, set by requiring that weathering balances volcanic outgassing in

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$^1$Even though the planetary albedo $\alpha$ in Equation (1) depends on $pCO_2$, we find that this dependence is quantitatively negligible for $pCO_2 \ll 0.2$ bar. For simplicity, we plot heating (red) curves in this low $pCO_2$ limit in Figure 1, which is indeed satisfied by all the cooling (blue) curves shown. All our other results fully account for the $\alpha$–$pCO_2$ dependence.
Equation (2), and $pCO_2$ is only indirectly constrained by the constant $T_{surf}$ requirement.

On the other hand, when the weathering rate depends on both $pCO_2$ and $T_{surf}$, noticeable bends appear in the blue cooling curves shown in Figure 1. Indeed, in this class of models, the reduced efficiency of weathering at low temperatures must be balanced by large $pCO_2$ values to match the volcanic outgassing rate $V$. The weaker the weathering $pCO_2$ dependence, the stronger is the $pCO_2$ build-up at low surface temperatures (compare $\beta = 0.5$ and 0.25 models represented by the dashed and dash-dotted lines in Figure 1, respectively). A planet with a larger volcanic $CO_2$ outgassing rate [21] achieves a warmer stable climate ($T_{surf} \approx 292$ K for $V = 3V_⊕$ and adopting our default weathering parameters, which is the case shown as a dotted line in Figure 1).

Interestingly, stable climate+weathering solutions can cease to exist at low insolation levels, such as the 1.25 and 1.6 AU cases shown in Figure 1, for a strong enough dependence of the weathering rate on $pCO_2$. For example, we find that blue curves no longer intersect with heating (insolation) lines beyond 1.077 AU if $\beta = 0.5$ and beyond 1.25 AU if $\beta = 0.25$. Conversely, a very weak weathering dependence on $pCO_2$ ($\beta < 0.1$), including the singular case $\beta = 0$ (vertical solid blue line in Figure 1), do permit stable climate+weathering solutions at almost arbitrarily low insolation levels [9]. Low $\beta$ values may be the relevant limit for planets where land vascular plants are widespread [3, 11, 12, 13, 14, 17]. On the other hand, for values of $\beta \sim 0.25$–0.5 appropriate for planets lacking land vascular plants [9, 11, 12, 13, 14, 17], the bending of cooling curves seen in Figure 1 also implies that at fixed volcanism rate, $V$, less insolated planets achieve climate+weathering equilibrium at gradually lower $T_{surf}$ values. For example, in the case $\beta = 0.25$, equilibrium is only marginally achieved above freezing temperatures at 1.25 AU, as shown in Figure 1. At low enough insolation levels (far enough away from the star), climate+weathering equilibrium is no longer possible above freezing temperatures, which implies that weathering equilibrium is unattainable.

On planets lacking weathering equilibrium, the climate must repeatedly cycle through a succession of radiative equilibria as illustrated in Figure 1: rapid transition from marginal to full glaciation ($A \to B$, at fixed $pCO_2$), slow build-up of $pCO_2$ caused by volcanic outgassing in the absence of weathering ($B \to C$), rapid transition to a deglaciated state ($C \to D$, at fixed $pCO_2$) and gradual $pCO_2$ decay under the action of weathering ($D \to A$), until the cycle repeats again with full glaciation. The general properties of the four critical points A-B-C-D of this climate cycle, which are independent of details of
the weathering model, are quantified in Figure 2 as a function of orbital distance from a Sun-like star. Blue curves correspond to the coldest cycle point with the lowest $pCO_2$ value, B, while the red curves correspond to the hottest point with the highest $pCO_2$ value, D.

Figure 2 shows that Earth-like planets at larger orbital distances glaciate and deglaciate at larger $pCO_2$ values. A deglaciation with $pCO_2 \simeq 0.14$ bar at 1 AU is consistent with values reported for snowball Earth deglaciation [6]. Planets beyond 1.3 AU support massive (> 0.3 bar) $CO_2$ atmospheres throughout their climate cycle. Planets in the frozen state have albedos $\simeq 0.65$, while the albedo of unfrozen planets rises from $\simeq 0.3$ to 0.45 in the range 1-1.8 AU, from an increasing atmospheric scattering contribution. The extremes of surface temperature along the cycle vary modestly with insolation level, with $T_{surf} \simeq 210-225$ K at the coldest point and 310-330 K at the hottest point.

### 3.2. Climate Cycles

Explicit time-dependent integrations of the system of Equations (1)-(3) reveal details of the climate cycle illustrated in Figure 1. We initiated these integrations at the hot, high $pCO_2$ (weathering-independent) point D and confirmed that Earth-like planets receiving sufficiently large insolation fluxes settle to a steady-state warm climate solution after relaxation to weathering equilibrium. By contrast, planets at large enough orbital distances (low enough insolation levels) experience large amplitude climate cycles, as anticipated from our discussion of equilibrium solutions in relation to Figure 1.

Figure 3 shows six illustrative examples (A-F) of such climate cycles, shown in terms of variable $T_{surf}$ and $pCO_2$ curves. Most of the cycle time is spent in the frozen state, during which $pCO_2$ build-up is slow compared to the fast weathering that occurs at above-freezing temperatures. In model 1, 4.6% of the 70 Myr cycle is spent in a warm state with surface liquid water. The corresponding numbers are 0.8% of 477 Myr in model 2, 7.5% of 76 Myr in model 3, 3.9% of 139 Myr in model 4, 28% of 312 Myr in model 5 and 17% of 517 Myr in model 6. Faster weathering at higher $T_{surf}$ also implies that most of the time in the unfrozen state is spent just above freezing temperatures, near the lowest $pCO_2$ levels covered during the cycle. For a fixed volcanic outgassing rate, $V$, the climate cycle duration increases with decreasing insolation because larger absolute $pCO_2$ values must be reached for climate transitions to occur (Figure 2). Decreasing insolation also reduces the fraction of cycle time spent with surface liquid water by the planet, although this can be compensated for by stronger volcanic outgassing. A weaker $pCO_2$ weathering dependence (lower $\beta$) lengthens the duration of
the unfrozen state since the decline in $pCO_2$ with time has less of an effect on the weathering rate.

To summarize, a temperature-only dependence of the weathering rate ($\beta = 0$) uniquely ties the surface temperature to the volcanic outgassing rate $V$ via Equation (2). A more general dependence on $pCO_2$ ($\beta > 0$) leads to a richer set of climate solutions, including unstable climate cycles at low enough insolation levels, when weathering equilibrium ceases to exist. These results are not specific to the weathering functional form adopted in Equation (3), in the sense that other weathering laws with a positive dependence on $pCO_2$ and $T_{surf}$ would lead to qualitatively similar climate behaviors.

4. Conclusions

The key new feature of our analysis is the lack of stable climates on Earth-like planets lacking land vascular plants, at low enough insolation levels. This suggests that a subset of Earth-like planets located in the outer regions of habitable zones may be preferentially found in a frozen, rather than deglaciated, state. A globally frozen state might be observationally inferred from the very high albedo and the correspondingly low water content of the planet’s atmosphere. According to these results, some Earth-like planets in the outer habitable zone would also be caught in a transiently warm state with surface liquid water present only infrequently.

The link between unstable climate cycles and the emergence and evolution of life on weakly-insolated Earth-like planets is unclear but possibly important. A reduced amount of time with surface liquid water on planets experiencing climate cycles could in principle slow down the emergence and/or evolution of life. On the other hand, life itself could strongly impact the weathering process on weakly-insolated Earth-like planets, as it seems to have done on early Earth \[3, 11, 12, 13, 14, 17\]. In particular, the ability of land vascular plants to regulate the soil $pCO_2$ level that is relevant to the weathering process, well above atmospheric $pCO_2$ levels, is consistent with these plants effectively decoupling the weathering rate from the atmospheric $pCO_2$ level \[3, 11, 12, 13, 14, 17\], leading to $\beta \to 0$ in Equation (3). As a result (Figure 1, vertical line), the climate of weakly-insolated Earth-like planets could be stabilized against transient cycles once the presence of land vascular plants becomes widespread. This would constitute a strong example of life exerting a feedback on its environment.

It is worth noting that Earth’s geological record is qualitatively consistent with the evolutionary path one may envision for a habitable planet.
orbiting a star that is gradually brightening over time. Repeated snowball events should be restricted to early times, when insolation is weak and land vascular plants are absent. They should disappear at late times once insolation is strong enough and/or land vascular plants become widespread.

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**Appendix A. Energy Balance Climate Model.**

We model the climate of Earth-like planets with a zero-dimensional reduction of a one-dimensional energy balance model [9]. The model assumes Earth parameters unless otherwise specified (e.g., surface gravity, land/ocean fraction and nitrogen contribution to the total atmospheric mass). Greenhouse effect from atmospheric $H_2O$ and $CO_2$ are included, with an atmospheric vapor pressure set by surface evaporation (temperature).

The top-of-atmosphere albedo and the outgoing longwave radiation flux are modeled as polynomial fits to a large number of radiative-convective models [9]. The polynomial fits are functions of surface temperature, partial $CO_2$ pressure, solar zenith angle and surface albedo. Simple prescriptions for snow/ice coverage, surface albedo and water cloud coverage are adopted [9]. For simplicity, we fix the cosine of the zenith angle to $\mu = 0.4$ and the albedo of ice-free oceans to 0.07 in all the models presented here. We also smooth out the top-of-atmosphere albedo polynomials near the 280 K transition to improve the continuity of the albedo function with temperature.

Based on published results [2], we expect that our results would be quantitatively different, but remain qualitatively valid, for planets that differ modestly from Earth in terms of their surface gravity, land/ocean fraction and/or nitrogen atmospheric content. Note that it has been suggested that weathering does not strongly depend on land/ocean fraction on an Earth-like planet [16].

The energy balance model employed here may not be fully reliable beyond 1.3-1.4 AU, where $CO_2$ clouds are expected to form and influence the climate [9, 10]. The most massive $CO_2$ atmospheres found in our models only marginally approach hard limits on $CO_2$ condensation [22].
Figure 1: Climate at global radiative equilibrium for an Earth-like planet. Red lines show the albedo-corrected insolation (heating) flux as a function of surface temperature, $T_{\text{surf}}$, at 1 AU (top), 1.25 AU (middle) and 1.6 AU (bottom) from a Sun-like star. Blue lines show the infrared cooling flux ($OLR$) according to various scenarios for the atmospheric $CO_2$ content (slanted solid line: fixed $pCO_2$ model; dashed: $\beta = 0.5$ weathering model; dotted: $\beta = 0.5$ weathering model with 3 times larger $CO_2$ outgassing rate; dashed-dotted: $\beta = 0.25$ weathering model; vertical solid: $\beta = 0$ weathering model). When no stable climate exists at large orbital distances (absent blue-red intersections), the climate must repeatedly cycle through points A-B-C-D shown in the 1.6 AU case, with a slow $pCO_2$ build-up (B-C), a transition to a hot climate (C-D), a weathering period with decreasing $pCO_2$ (D-A) and a transition to global glaciation (A-B).
Figure 2: Values of atmospheric partial CO$_2$ pressure (1), surface temperature (2) and planetary albedo (3) at extremes of the climate cycle illustrated in Figure 1, as a function of orbital distance from a Sun-like star. Blue curves correspond to point B (cold, high albedo, low pCO$_2$) and red curves to point D (hot, low albedo, high pCO$_2$) of the cycle shown in Figure 1.
Figure 3: Six examples of time-evolved climate cycles, with two full cycles shown in each case. For each model (1-6), the evolution of atmospheric partial CO$_2$ pressure (lower panel) and surface temperature (upper panel) are shown. Model 1: Default weathering model at 1.1 AU ($\beta = 0.5$ in Equation (3)). Model 2: Default weathering model at 1.8 AU. Model 3: 3 times larger CO$_2$ outgassing rate at 1.3 AU. Model 4: 3 times larger CO$_2$ outgassing rate at 1.6 AU. Model 5: Weaker pCO$_2$ weathering dependence ($\beta = 0.25$) at 1.3 AU. Model 6: Weaker pCO$_2$ weathering dependence ($\beta = 0.25$) at 1.6 AU. Increasing the CO$_2$ outgassing rate shortens the duration of the cold, CO$_2$ build-up phase. Weakening the weathering pCO$_2$ dependence ($\beta = 0.25$ rather than 0.5) lengthens the duration of the hot weathering phase, resulting in a much larger fraction of time spent with surface liquid water.
Appendix  B. Model Simplifications and Limitations.

Our models are idealized in a number of important ways. In addition to the simplified, zero-dimensional treatment of climate described above, which suggests the possibility of richer behaviors in higher complexity, three-dimensional climate models, our treatment of weathering processes is intentionally simple, in order to isolate the key factors that determine climate stability. We ignore seafloor weathering [16] and the mantle CO$_2$ cycle [23].

The absolute calibration of weathering rates in the absence of land vascular plants is unknown, but it is thought to be less than in their presence [11]. For concreteness, we have chosen to calibrate weathering fluxes in our models using current Earth [13, 14] (Equation (3)). We note that in a model admitting steady-state solutions, a factor three decrease in the weathering rate is equivalent to a factor three increase in the volcanic outgassing rate (see Equation (2)), which is one of the cases shown in Figure 1 (dotted blue line). Such a model retains the main qualitative feature highlighted in this work, which is the disappearance of stable climates solutions at low enough insolation levels (beyond 1.2AU for the dotted blue line shown in Figure 1). Different calibrations in weathering rates and/or volcanic outgassing rates will thus affect our results quantitatively, but our main conclusions should remain valid.

More generally, a planet is likely to change its weathering regime gradually over time, as different forms of life emerge and spread over its surface [3, 12]. Our models have intentionally focused on the distinction between the absence and presence of land vascular planets, which exemplifies the interplay between life, weathering processes and climate stability.

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