Habitability from Tidally Induced Tectonics

Diana Valencia1,2, Vivian Yun Yan Tan3, and Zachary Zajac4

1 Department of Physical & Environmental Sciences, University of Toronto, Toronto, Canada
2 Department of Astronomy & Astrophysics, University of Toronto, Toronto, Canada
3 Department of Physics and Astronomy, York University, Toronto, Canada
4 Department of Medical Biophysics, University of Toronto, Toronto, Canada

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Abstract

The stability of Earth’s climate on geological timescales is enabled by the carbon–silicate cycle that acts as a negative feedback mechanism stabilizing surface temperatures via the intake and outgassing of atmospheric carbon. On Earth, this thermostat is enabled by plate tectonics that sequesters outgassed CO2 back into the mantle via weathering and subduction at convergent margins. Here we propose a separate tectonic mechanism—vertical recycling—that can serve as the vehicle for CO2 outgassing and sequestration over long timescales. The mechanism requires continuous tidal heating, which makes it particularly relevant to planets in the habitable zone of M stars. Dynamical models of this vertical recycling scenario and stability analysis show that temperate climates stable over timescales of billions of years are realized for a variety of initial conditions, even as the M star dims over time. The magnitude of equilibrium surface temperatures depends on the interplay of sea weathering and outgassing, which in turn depends on planetary carbon content, so that planets with lower carbon budgets are favored for temperate conditions. The habitability of planets such as found in the Trappist-1 system may be rooted in tidally driven tectonics.

Key words: planets and satellites: individual (Trappist-1) – planets and satellites: tectonics – planets and satellites: terrestrial planets

1. Introduction

In the search for habitable planets, we rely on our knowledge of the Earth to guide us. We understand that it is not enough for a planet to be located in the habitable zone for it to be habitable. It is equally important that its atmospheric response to insolation allows for liquid water on its surface, and this depends on the amount and type of greenhouse gases present.

The long-term stability on Earth has been attributed to the carbon–silicate cycle, which maintains atmospheric carbon dioxide levels at values that allow for surface liquid water over million-year timescales, while exchanging carbon between the different major reservoirs (atmosphere and ocean, continental and oceanic crusts, mantle). The main reason why this cycle brings climate stability is that weathering from the atmosphere depends on atmospheric temperature. When temperatures rise, weathering rates increase, drawing down CO2 from the atmosphere into the rocks, thus reducing the greenhouse effect and restoring temperature levels. Conversely, when the temperature is low, weathering is sluggish or non-existent (if the planet has gone into a snowball state), allowing for volcanism to increase levels of CO2 in the atmosphere.

Evidence in the geological record suggests that the Earth’s climate has been temperate over most of the past 3–4 billion years, despite the fact that the Sun has been brightening over time (Sagan & Mullen 1972). Owen et al. (1979) and Kasting et al. (1993) proposed that higher levels of CO2 in the past, possibly due to the carbon–silicate cycle, could offset the reduced insolation level. While this is the leading theory, studies suggesting that CO2 is unable to resolve the faint young Sun paradox include limits to the amount of atmospheric CO2 in the past derived from data on siderate palaeosols (Rye et al. 1995) as well as inferences from modeling vigorously convecting mantles and reactable ejecta in the early Earth that would draw down atmospheric CO2 to too low a value (Sleep & Zahnle 2001). While the need for other greenhouse gases such as NH3 (suggested by Sagan & Mullen 1972) might be the answer to these caveats, the carbon–silicate cycle on Earth, with the ability to regulate atmospheric CO2, has at least to some extent contributed to the long-term climate stability of our planet.

It is also true that in the case of the Earth, this cycle is enabled by the fact that plate tectonics connects the different reservoirs. Carbon is drawn from the atmosphere into the rocks via rock weathering on the continents and sea weathering on the ocean crust. Through rivers and streams, continental rocks get deposited into the oceanic crust, which gets subducted into the mantle at convergent margins, while continental crust is scraped and carried down by the subducting plate. Through volcanism, carbon is outgassed from the mantle into the ocean at mid-ocean ridges, and directly into the atmosphere at continental arcs and ocean islands. Thus, subduction, which is a central component of plate tectonics, closes the carbon–silicate cycle on Earth.

In addition, plate tectonics is also important to the carbon–silicate (C–Si) cycle because it assists the weathering process by constantly exposing fresh rock that is subsequently available for carbon sequestration, either by continually producing ocean crust at the mid-ocean ridges and ocean islands or by enabling erosion on the continents through persistent topographical changes derived from mountain building and orogeny processes (Turcotte & Schubert 2002). Given these reasons, plate tectonics has been tied to climate stability and hence to habitability on Earth (Walker et al. 1981; Gaillardet et al. 1999; Kump & Arthur 1999; West et al. 2005; West 2012; Maher & Chamberlain 2014).

However, it is debated whether or not plate tectonics can happen in exo-Earths. Some suggest it is possible (Valencia et al. 2007; Korenaga 2010; van Heck & Tackley 2011; Foley et al. 1995).
et al. 2012; Tackley et al. 2013), while others consider it to be unlikely (O’Neill et al. 2007; Stamenković et al. 2012; Noack & Breuer 2014). In this study, we explore a different type of tectonism, driven by tidal heating, that may serve in an analogous way to plate tectonics on Earth in assisting a carbon–silicate cycle. Thus, increasing the chances of finding planets that are habitable.

Inspired by the efforts to find planets in the habitable zone around M stars, where tidal heating can be important, we envision a tectonic scenario where volcanism is driven by tidal dissipation within the mantle, in a similar fashion to what has been proposed for Io (O’Reilly & Davies 1981).

This mechanism would be highly relevant for planets that have nonzero eccentricities orbiting in the habitable zones of M stars, such as three of the seven planets of the Trappist-1 system (Gillon et al. 2017). This recently discovered system of seven tightly packed planets near resonances includes three (d, e, f) in the habitable zone. Given the planets’ gravitational perturbations by their neighbors, we expect small nonzero values of eccentricity (Hansen & Murray 2015; Tamayo et al. 2017; Vinson & Hansen 2017), making these three planets highly suitable candidates to exhibit tidally driven tectonics, and perhaps a built-in thermostat for climate stability analogous to Earth.

In this article we investigate how this newly proposed mechanism may enable climate stability for planets that are tidally heated and are found in the habitable zone. It is organized as followed: in Section 2 we present the tidally driven tectonism we envision and how it can assist climate stability, as well as the governing equations; in Section 3 we discuss the results; in Section 4 we discuss our assumptions and implications, and we present a summary of our findings in Section 5.

2. Model

2.1. Scenario of a Tidally Driven C–Si Cycle

To come up with an alternative system, we need to break down the key elements of the carbon–silicate cycle on Earth that provide our planet with a viable thermostat for climate stability. (1) There needs to be a feedback mechanism that draws out an amount of CO₂ from the atmosphere that varies with CO₂ concentration (Walker et al. 1981), and deposits it in a different reservoir. In the case of the Earth this is both the rock and sea weathering processes that depend on CO₂ concentrations directly, and very importantly, indirectly via the atmospheric temperature, and store carbon in the rocks and ocean crust. (2) This mechanism has to supply fresh rock for weathering at a rate large enough not to produce a bottleneck in the system. On Earth this exposure of rock happens continuously thanks to persistent erosion and the production of mid-ocean ridges. (3) Lastly, the reservoir has to have a way of injecting CO₂ back into the atmosphere when atmospheric levels decrease. On Earth this happens because volcanism is a continuous source of CO₂ from the mantle, and the mantle in turn is continuously replenished with subducted carbonate rocks from the ocean and continental crust.

An analogous system that accomplishes all four elements is inspired by the tidally driven tectonism suggested on Io (O’Reilly & Davies 1981), or pipe heating, and is depicted in Figure 1.

![Figure 1. Tidally driven tectonism and ρCO₂ long-term feedbacks. Continuous volcanism from a partially molten mantle degasses CO₂ in the ocean/atmosphere. The basaltic layer that forms from this volcanism can sequester C from the ocean/atmosphere reservoir via sea weathering. The basaltic crust grows at the top with resurfaced basalt, and founders at the bottom, moving C into the mantle. There is no need for continents, and continental weathering, for this mechanism to enable the C–Si cycle on tidally heated planets.](image)

Plants that are tidally heated can dissipate heat in their interior and exhibit partially molten mantles, which get rid of their heat by pushing melt through plumes to the surface. This melt continuously resurfaces in the form of basalt, and forms a layer that accumulates over time. This freshly advected rock can react with atmospheric or ocean CO₂ in a similar way to that on Earth and sequester C via a weathering reaction. With time, new basaltic crust gets deposited on top, so that the old basalt and carbonate rock get buried and move deeper within the planet. At some point, depending on the resurfacing rate, these rocks that were once at the surface will be buried and delaminated into the mantle, carrying down carbon with them and closing the cycle. Thus, recycling occurs in a vertical fashion instead of the horizontal character of plate tectonics on Earth.

We shall examine in more detail each of the components of this proposed system before presenting the climate model we built for these planets.

2.1.1. Tidal Heating

A planet with a continuous and substantial source of tidal heating may have a partially molten interior. Just like the volcanism on Io, this melt may reach the surface through plumes advecting heat out to the surface. It is thus important to calculate the amount of heating a planet can experience. Based on the theory of tides (Murray & Dermott 1999) the tidal heating available to a planet depends on the eccentricity of the system e, the semimajor axis a, the mass of the star Mₖ, the mean motion n, the radius of the planet Rₚ, the specific dissipation parameter Q, the ratio of elastic to gravitational forces \(\bar{\mu}(\sim(10^4 \text{ km}/Rₚ)^5)\) and the gravitational constant G:

\[
\frac{dE}{dt} = \frac{63\pi^2nRₚ^5GM_k^2}{4\bar{\mu}Qa}.
\]

The least well constrained parameter is Q although typical values of \(\sim 100\) are commonly used for icy/rocky planets. In reality the dissipation parameter is a function of the planet’s structure and should vary radially instead of being a single value.

\[\text{Footnote 5: We note that there is no need to invoke a global molten layer, but the presence of partial melt is enough.}\]
value. However, as a first step, we take the same approach as in most studies and use a constant value. Equation (1) can also be written as

\[
\frac{dE}{dt} \approx 10 \times \left( \frac{e}{0.01} \right)^2 \left( \frac{100}{Q} \right) \left( \frac{R_{p}}{a} \right)^7 \times \left( \frac{M_{e}}{M_{\odot}} \right)^{5/2} \left( \frac{0.1 \text{au}}{a} \right)^{15/2} \text{[TW]}.
\]

Therefore, an Earth-sized planet orbiting a Sun-like star at 0.1 au and with a low eccentricity of 0.01 would dissipate 10 TW or ~0.02 W m⁻², a tidal heat flux 100 times less than Io’s flux (Veerder et al. 1994; Spencer et al. 2000). For comparison, Earth’s flux is estimated at 0.09 W m⁻² (Davies & Davies 2010). To reproduce the measured heat flux for Io at its location with respect to Jupiter with Equation (1), the dissipation factor would have to be chosen to be \( \dot{Q}_{\text{b}} = 150 \).

Given that we are interested in planets in the habitable zone that can experience tidal heating, we need to consider planets around M stars or even brown dwarfs. Planets around G-type stars that are in the habitable zone are too far away to experience any tidal heating from the star, whereas M stars have their habitable zone close enough that tidal dissipation can matter. In fact, tidal dissipation in planets around M dwarfs has previously been studied in the context of orbital and thermal evolution (Driscoll & Barnes 2015). In fact, Barnes et al. (2009) claimed that there is a limit to how much tidal heating a habitable planet may experience based on the assumption that too much volcanism and high resurfacing rates would be inhospitable. In contrast, we propose that in these conditions a new mechanism for climate regulation may kick in, rendering the planet habitable.

For a star with \( M_{*} = 0.08 M_{\odot} \), such as Trappist-1, an Earth-sized planet orbiting at 0.02 au with an eccentricity of 0.01 would experience 3000 TW or 6 W m⁻² of tidal energy. For an assumed eccentricity of 0.01 planets d and e would have tidal fluxes of 0.96 and 0.30 W m⁻², respectively. These numbers quickly grow with eccentricity, so that planets around M stars can have large tidal heating fluxes. The important aspect for the model we propose is to determine whether there is enough energy to yield a partially molten mantle. One could use Io for a simple comparison to infer whether or not planets such as Trappist-1 d, e, and f could have a partially molten mantle by comparing the amount of tidal energy available to the tidal energy carried to the surface.

According to Equation (1) tidal production increases as \( K_{p}^5 \), while the heat flow carried from the mantle to the atmosphere increases as \( K_{p} \), so that to zeroth order larger planets should be more likely to have an interior melt region. Even at very low eccentricity values of 0.01, Trappist-1 d would experience similar tidal heat fluxes to Io. Thus, it is possible that the Trappist-1 planets in the habitable zone have persistent melt in their interior.

This melt would be advected in the form of volcanic conduits cooling at the surface, carrying with it volatiles and degassing them into the ocean/atmosphere (see Figure 1). This continuous basalt extraction would form a layer of a certain thickness, limited by a partially molten mantle at the bottom. To estimate both the thickness of the basalt layer and the resurfacing velocity that are used in the climate model we use the model of O’Reilly & Davies (1981).

The resurfacing velocity, taken to be the same as the subsidence velocity, is

\[
v = \frac{q_{f}}{\rho (L + C_{p} \Delta T)},
\]

where \( q_{f} \) is the tidal heat flux (tidal energy per unit surface area), \( \rho \) the density of the mantle, \( L \) the latent heat, \( C_{p} \) the heat capacity, and \( \Delta T = T_{m} - T_{s} \) the difference between the melting and surface temperatures. The thickness of the basalt layer is

\[
d_{bas} = \frac{\kappa}{v} \log \left( 1 + \frac{k v}{\kappa} \frac{\Delta T}{q_{f} \left( \frac{1}{a} - 1 \right)} \right).
\]
tectonics, decompression melting takes place near the surface as upper mantle material adiabatically ascends into mid-ocean ridges, and melt gets carried out as volcanism. However, if melt is produced at depth, it can stay within the mantle due to the pressure for the crossover density would be reached at shallower depths. This could pose a problem for our proposed mechanism if tidal dissipation creates melt deeper than the crossover density for the specific planet. The pressure for this crossover density depends on the composition of the solid (e.g., basalt versus peridotite, or iron content) and the water content of magma (Sakamaki et al. 2006). The higher the iron content of the solid or the drier the magma, the shallower the crossover density. Sakamaki et al. (2006) show that, for Earth, hydrated magma occurring from melting basalt rock in a mid-ocean ridge has a crossover density near 410 km (~24 GPa) if water is present at the 2% level, and deeper at the 8% level. By extrapolating their data and using the preliminary reference Earth model (Dziewonski & Anderson 1981), we obtain a depth for crossover density for 8% water in magma of 1400 km (~60 GPa) for Earth. Given as well that tidal strain is largest near the surface, it is reasonable to think that tidal heating may produce melt near the surface, and thus our proposed mechanism may work for a subset of planets.

2.1.2. Sea Weathering

In broad strokes, the carbon–silicate cycle is the process by which CO$_2$ is drawn from the atmosphere by removing C when it interacts with silicate rocks via water and carbonic acid, which releases bicarbonate, Ca$^{++}$ (and Mg$^{++}$) ions, either on the continents or directly with basalt on the newly formed ocean crust. If it is on the continents, these ions eventually find their way to the ocean floor, carried down by rivers and ground water. On the ocean floor they form carbonates that trap carbon (that was once in the atmosphere) into rocks, which eventually get subducted into the mantle at convergent margins. The net reaction is captured in the Urey reaction (Urey 1952):

$$\text{CO}_2(g) + \text{CaSiO}_3(s) \rightleftharpoons \text{SiO}_2(s) + \text{CaCO}_3(s),$$  

where the right arrow describes how carbonates are formed. Once at depth, these carbonate rocks can react back via metamorphism, releasing carbon dioxide that eventually makes it back into the atmosphere via volcanism (the reverse reaction, left arrow).

On Earth, this cycle is enabled by plate tectonics that continuously exposes fresh rock on the ocean floor, as well as occurring on the continents driven by erosion via orogeny and the build-up of topography. Rock weathering has been well studied (Walker et al. 1981; Berner & Caldeira 1997; Berner 2004), while sea weathering has gained more interest in recent years (Coogan & Gillis 2013; Coogan & Dosso 2015). The carbonate reactions of sea weathering may be controlled in a different manner to continental weathering, with secondary carbonates and alkalinity playing an important role (Coogan & Dosso 2015). The contribution of each weathering type to the total weathering process on Earth, or how it might have changed throughout Earth’s history, is still unknown (Mills et al. 2014), although recent estimates by Krissansen-Totton & Catling (2017) suggest that continental weathering has been dominant in the last 100 Myr.

For the planets proposed here, we envision that tidally driven tectonics can continuously expose fresh rock under an ocean, but that any continental shelves, if present, may not exhibit rock weathering owing to a sustained lack of erosion. In this case, the only mechanism for drawing down CO$_2$ in these planets would be sea weathering.

While some authors have considered sea weathering to be dependent on CO$_2$ concentrations alone (Sleep & Zahnle 2001; Foley 2015), laboratory experiments (Brady & Gíslason 1997) and isotopic constraints from oceanic carbonates (Coogan & Dosso 2015) suggest a temperature dependence in the form of an Arrhenius law. The temperature of the deep ocean, where Ca leaching and carbonate production are taking place, depends in turn on the atmospheric temperature, and hence on the atmospheric CO$_2$ concentrations, because atmospheric temperature determines how much cold surface water enters the thermohaline circulation system (Brady & Gíslason 1997). This dependence of sea weathering on atmospheric temperature and CO$_2$ is crucial in allowing for a climate-stabilizing feedback. We adopt the same equation as Mills et al. (2014) to describe sea weathering as a function of atmospheric CO$_2$ levels and atmospheric temperature $T$,

$$W_{\text{sea}} = \omega W_{\text{sea}}^E \left( \frac{p_{\text{CO}_2} \text{atm}}{p_E} \right) \exp \left( \frac{E}{R \left( \frac{1}{T_E} - \frac{1}{T} \right)} \right),$$  

where $R$ is the gas constant, $E$ is the activation energy, $W_{\text{sea}}^E$ is the baseline estimate of the sea weathering rate (in bar Myr$^{-1}$ or mol Myr$^{-1}$) for a reference state that we set at first to be the equilibrium atmospheric partial pressure of carbon dioxide $p_E$ and atmospheric equilibrium temperature $T_E$, and $\omega$ is a function that lumps together all other quantities that sea weathering might depend on. The feedback mechanism comes from the dependence on atmospheric temperature. Any deviations from the equilibrium temperature would drive much higher or lower levels of sea weathering, depending on whether the planet is hot with high levels of atmospheric CO$_2$ or cold, respectively.

In addition, sea weathering may include a dependence (through $\omega$) on the velocity at which fresh rock is exposed, which in the case of the Earth is the ridge spreading velocity, which has changed over time. For Earth this term is often written as $\omega = \left( \frac{\nu_{\text{sp}}}{\nu_{\text{sp}}^{\text{Earth}}} \right)^\beta$, where $\nu_{\text{sp}}$ is the present-day velocity. While many authors (Sleep & Zahnle 2001; Mills et al. 2014; Foley 2015) use $\beta = 1$, Krissansen-Tutton & Catling (2017) propose $\beta \sim 0.5$. In either case, this contribution is of order unity. Other factors may also be affecting the weathering rate (e.g., alkalinity, grain size, etc.), and thus for simplicity we take $\omega \sim 1$; or conversely our results for sea weathering can be taken as scaling with $\omega$.

Two terms in Equation (6) are poorly known: the direct dependence on atmospheric carbon dioxide pressure $\alpha$ and the activation energy $E$. Brady & Gíslason (1997) proposed $\alpha = 0.23$ and $E = 41$ kJ mol$^{-1}$, while Coogan & Dosso (2015) proposed $E = 92$ kJ mol$^{-1}$, and Krissansen-Tutton & Catling (2017) considered a range of $E = 40$–110 kJ mol$^{-1}$. We vary both $\alpha$ and $E$, as well as perform a stability analysis to
see the effect on the climate-stabilizing feedback we propose here.

Alkalinity—Coogan & Dosso (2015) have argued that carbonate mineral precipitation is largely controlled by alkalinity, and because rock dissolution involved in sea weathering increases alkalinity the effect is to efficiently drive carbon sequestration. We note that Equation (6) does not have an explicit dependence on alkalinity, but an implicit one, because it depends on global temperatures that influence deep-water ocean temperature, which in turn controls the dissolution rates of mafic and ultramafic rocks (Krissansen-Totton & Catling 2017).

### 2.2. Governing Equations

#### 2.2.1. Carbon Cycle

To build our model of the carbon cycle we divide our planet into three distinct reservoirs: the mantle, the basalt layer, and the ocean and atmosphere together (similar to Sleep & Zahnle 2001, Foley 2015). See Figure 1 for a cartoon representation of the tectonic process we propose in this study.

Magmatic volcanism originates in the molten mantle layer and carries melt within pipes that outgas CO₂ into the reservoirs of the ocean and atmosphere at a rate proportional to the resurfacing rate \( v \). The flux of CO₂ degassed into the atmosphere–ocean reservoir is

\[
D = pCO_2^{\text{man}} \left( \frac{f \Delta A_p}{V_{\text{man}}} \right),
\]

where \( pCO_2^{\text{man}} \) is the partial pressure of carbon dioxide in the mantle, \( V_{\text{man}} \) is the volume of the mantle, \( A_p \) is the area of the planet, and \( f \) is the fraction of CO₂ within the melt that gets degassed. On the other hand, the sink for the atmosphere–ocean reservoir is the flux of CO₂ that is weathered at the bottom of the ocean, namely the sea weathering rate described in Equation (6). By drawing out CO₂ from the ocean, sea weathering effectively draws out CO₂ from the atmosphere, given that the partitioning between the two reservoirs is set by how soluble CO₂ is. Following Foley (2015), we use Henry’s law for solubility to determine how much CO₂ is in the atmosphere (\( pCO_2^{\text{atm}} \)) versus the ocean (\( pCO_2^{\text{oc}} \)),

\[
pCO_2^{\text{atm}} = pCO_2^{\text{atm-oc}} - pCO_2^{\text{oc}} = \frac{K_H}{\mu} \left( pH_2O - \frac{pCO_2^{\text{oc}}}{\mu} \right),
\]

where \( K_H \) is the solubility constant, \( pH_2O \) is the content of water in the ocean, and \( \mu \) is the molar mass. The content of water in the ocean is calculated by assuming the mass of the Earth’s ocean, \( M_{\text{oc}} \).

While sea weathering is a sink for the atmosphere–ocean reservoir, it behaves like a source for the basaltic crust. Once C is sequestered into the basalt layer, it starts getting buried by subsequent melt deposited at the top. The C subsides until it reaches the bottom of the basalt layer, at which point it is delaminated into the molten mantle layer. To keep all the quantities in the same metrics, we use C and CO₂ interchangeably, knowing that C resides in rocks while CO₂ is the gaseous form (degassing from the mantle).

The rate at which CO₂ in the basalt is foundered or delaminated into the mantle is

\[
F_{\text{found}} = pCO_2^{\text{bas}} \frac{v}{d_{\text{bas}}},
\]

Therefore the source of carbon dioxide for the mantle is the carbon dioxide foundered from the basalt layer, and the sink is the flux being outgassed through volcanism. Thus, the equations describing this system are

\[
\frac{d}{dt} pCO_2^{\text{atm-oc}} = D(pCO_2^{\text{man}}) - W_{\text{sea}}(pCO_2^{\text{atm}}, T)
\]

\[
\frac{d}{dt} pCO_2^{\text{man}} = F_{\text{found}}(pCO_2^{\text{bas}}) - D(pCO_2^{\text{man}})
\]

\[
\frac{d}{dt} pCO_2^{\text{bas}} = W_{\text{sea}}(pCO_2^{\text{atm}}, T) - F_{\text{found}}(pCO_2^{\text{bas}}).
\]

(10)

Given that the total content of CO₂ for the planet is fixed, \( pCO_2^{\text{atm-oc}} + pCO_2^{\text{man}} + pCO_2^{\text{bas}} = pCO_2^{\text{tot}} \), one of the three equations is redundant.

The equilibrium carbon dioxide values for the mantle, atmosphere, and ocean are

\[
pE^{\text{man}} = W_{\text{sea}} p_{\text{man}} V_{\text{man}} / f \Delta A_p
\]

\[
pE^{\text{atm}} = pCO_2^{\text{tot}} - W_{\text{sea}} d_{\text{bas}} / v + V_{\text{man}} f \Delta A_p
\]

(12)

and \( pE^{\text{oc}} = pE^{\text{atm}} - pE^{\text{man}} \) in dimensional form.

For typical values refer to Table 1. To be able to solve the system of Equations (10) or (27), we need to specify the equilibrium conditions for the atmosphere, namely the equilibrium partial pressure of CO₂, \( pE^{\text{atm}} \), that sets the equilibrium atmospheric temperature. While Menou (2015) took \( pE^{\text{atm}} = 330 \) bar, the pre-industrialization partial pressure of carbon dioxide, as the equilibrium value for the Earth, Haqq-Misra et al. (2016) argued that a more appropriate value for an abiotic Earth would be \( pE^{\text{atm}} = 0.01 \) bar, by including all the carbon dioxide currently stored in the soil. We also take the equilibrium partial pressure to be \( pE^{\text{atm}} = 0.01 \) bar.

This value for the atmosphere, the solubility constant for the ocean, and the amount of water in Earth’s ocean, set the equilibrium CO₂ value for the atmosphere–ocean system via Equation (8) to be \( pE^{\text{atm}} = 0.038 \) bar.

We also note that in line with astrophysical studies we use the units of bars instead of moles, and the conversion factor we use is

\[
pCO_2 [\text{bar}] = 1.0197 \times 10^{-5} \times \mu_{CO_2} \frac{g}{4\pi R_p^2} \times C [\text{mol}],
\]

(13)

where \( g \) is the planet’s gravity and \( R_p \) is the planet’s radius.

#### 2.2.2. Climate Model

To obtain the atmospheric temperature of a planet provided the partial pressure of carbon dioxide, we use the energy balance model (EBM) of Haqq-Misra et al. (2016). They
provide a fit to the outgoing longwave radiation (OLR), and the Bond albedo at the top of the atmosphere for four different types of stars including G and M stars. These quantities depend on the partial pressure of atmospheric carbon dioxide, temperature, and zenith angle \( \mu \). We simplify the model to capture global parameters by fixing the zenith angle to a value \( \mu = 0.232 \) rad that would provide a global surface equilibrium temperature of \( T_E = 288 \) K for \( p_{\text{fl}} = 0.01 \) bar for an Earth-like scenario. The zeroth-order EBM we use equates the planet’s thermal radiation to the net insolination

\[
\text{OLR}(p_{\text{CO}_2}^{\text{atm}}, T) = \frac{S_\odot}{4}(1 - A(p_{\text{CO}_2}^{\text{atm}}, T)),
\]

where \( S_\odot \) is the solar constant at the planet’s location, and \( A \) is the albedo at the top of the atmosphere. This quantity also depends on the ground albedo, which we take from Williams & Kasting (1997) and modify for M stars. For the Earth, the ground albedo is taken to be a discontinuous function of \( T \), where for temperatures less than 273 K, the ground is considered to be frozen and exhibit a high albedo. At temperatures below 263 K the water in the atmosphere is assumed to condense out entirely, owing to the abrupt transition of the Earth entering a snowball state. However, a recent study by Checlair et al. (2017) on planets around M stars suggests that, unlike on Earth, ice coverage would proceed gradually, avoiding sudden snowball transitions, owing to the special spatial insolation pattern they receive. In our model, we use both the same albedo function for Earth and a modified version that precludes the snowball state following Checlair et al. (2017). To achieve this, we allowed for a gradual ice coverage as temperatures decrease below 273 K, retaining 30% of the land exposure at 150 K, and allowed for cloud albedo at all temperatures.

The timescale governing Equation (14) is much shorter than the long geophysical timescale involved in the carbon–silicate cycle of Equations (10) (Menou 2015). This means that in our model, surface temperature is calculated instantaneously from the amount of \( p_{\text{CO}_2}^{\text{atm}} \) at each timestep in the integration of Equations (10).

Figure 3 shows the results from the EBM model for the equilibrium case \( p_{\text{CO}_2}^{\text{atm}} = 0.01 \) bar. We show the case of the Earth (orange line) for comparison purposes. Earth’s climate exhibits two stable points, one near 225 K where the planet would be in a snowball state, and the temperate 288 K, as well as one unstable point, near 273 K. Furthermore, the snowball state at 225 K is thought to be transient, given that rock weathering is considered to be suppressed at these temperatures so that outgassing from volcanoes eventually deglaciates the planet. It is worth mentioning that if sea weathering can operate at these low temperatures in a large enough way to balance volcanism, this snowball state could be stable in the long term. Sea weathering would have to happen on the ocean floor below a thick crust of ice, which still allows for carbon dioxide to diffuse from the atmosphere to the ocean, and volcanism would most likely have to be sluggish.

Considering planets around M dwarfs, we also modified the EBM to restrict snowball states. This allows only one state with a stable temperature (purple line).

Table 1

| Parameter                          | Symbol | Value  | Reference |
|-----------------------------------|--------|--------|-----------|
| Sea weathering dependence on CO₂ | \( \alpha \) | 0.23   | (1)       |
| Sea weathering at equilibrium     | \( W_{\text{sea}} \) | 0.12 bar Myr\(^{-1}\) | calculated |
| Equilibrium atmospheric temperature | \( T_E \) | 288 K | assumed |
| Equilibrium atmospheric CO₂ partial pressure | \( \rho_{\text{fl}} \) | 0.01 bar | (2) |
| Activation energy                 | \( E_{\text{sea}} \) | 41 kJ | nominal value, (1) |
| Outgassing CO₂ fraction           | \( f \) | 0.01 | nominal value, calculated |
| CO₂ molecular weight              | \( \mu_{\text{CO}_2} \) | 44.1 mol g\(^{-1}\) | calculated |
| H₂O molecular weight             | \( \mu_{\text{H}_2\text{O}} \) | 18.015 mol g\(^{-1}\) | calculated |
| Mass of ocean                     | \( M_\infty \) | 1.4 \times 10^{24} kg | calculated |
| Solubility constant               | \( K_\alpha \) | 235.48 bar | calculated |
| Radius of planet                  | \( R_p \) | 6371 km | (3) |
| Radius of core                    | \( R_c \) | 3400 km | (3) |
| Mass of planet                    | \( M_p \) | 5.972 \times 10^{24} kg | (3) |
| Zenith angle                      | \( \mu \) | 0.232 | calculated |
| Sun’s luminosity today            | \( S_{\odot} \) | 1361 W m\(^{-2}\) | calculated |
| Thickness of basalt crust         | \( d_{\text{bas}} \) | 10 km | nominal value, (4) |
| Resurfacing velocity              | \( \nu \) | 1 cm yr\(^{-1}\) | assumed |
| Total carbon content              | \( p_{\text{CO}_2}^{\text{atm}} \) | 215 bar | calculated |
| Characteristic timescale          | \( \tau \) | 3 Myr | calculated |

Note. References: (1) Brady & Gíslason (1997), (2) Haqq-Misra et al. (2016), (3) Stacey & Davis (2008), (4) Sleep & Zahnle (2001).
Figure 4. Stellar evolution tracks used. The brightness of the M dwarf (blue) decreases over time, while the Sun (orange) has been brightening since it reached the main sequence.

desiccated or wet after the intense EUV/XUV star fluxes. While earlier studies (Barnes et al. 2013) estimated planets around M dwarfs to be completely dry, recent works (Schaefer et al. 2016; Bolmont et al. 2017) that use a better prescription for the atmospheric loss (Tian 2015) suggest that some water is retained, making them prospective planets for habitability.

For the evolution of an M dwarf, we use a spline fit to the model by Baraffe et al. (2015) for a star of mass $M_{\text{star}} = 0.08 M_{\odot}$ such as Trappist-1. Figure 4 shows the evolution of Trappist-1 compared to that of our Sun.

3. Results

3.1. Sea Weathering

We look first at how an equilibrium may be established and the timescale associated with it. All the values for the parameters used in the model are shown in Table 1.

To integrate Equation (10) we can proceed in either of two ways: (1) determine the sea weathering rate needed to ensure the equilibrium of the system is at $p_E = 0.01$ bar and $T_E = 288$ K for the carbon content of the planet, and compare this weathering rate to values obtained for Earth, or (2) use Earth’s estimated weathering rate at present-day conditions of $p_{\text{CO}_2} = 0.01$ bar and $T = 288$ K as the reference state in Equation (6), include the term for velocity for other planets, derive the corresponding equilibrium states, and then ask whether they are suitable for surface liquid water.

In either case, the functional form of the governing equations remains the same, and from stability analysis (see the Appendix) we conclude that the equilibrium state is in fact stable.

We use both approaches but favor the first one to solve the equations for two reasons. We use a simple functional form for sea weathering that can be made more realistic with more understanding (e.g., alkalinity dependence, etc.), and there is considerable uncertainty behind calculating present-day sea weathering rates. Values used previously range over more than one order of magnitude: 0.225–0.675 Tmol C/yr (Krissansen-Totton & Catling 2017), 1.75 Tmol C/yr (Mills et al. 2014), and 3.4 Tmol C/yr (Alt & Teagle 1999; Sleep & Zahnle 2001). Thus, we are not confident enough in our understanding of the weathering rate on Earth to use it at face value for other planets.

To ensure the equilibrium conditions of $p_E = 0.01$ bar and $T_E = 288$ K are at met, the sea weathering at equilibrium must adjust itself to account for the planetary tectonic conditions and the total carbon dioxide content in the following way:

$$W_{\text{sea}}^E = (p_{\text{CO}_2}^{\text{tot}} - p_{E}^{\text{atm}}) \frac{f_{vA_p}}{V_{\text{man}} + f_{A_p} d_{\text{bas}}}.$$  (15)

If this sea weathering rate can be achieved, and there are no limiting processes hindering sequestration (i.e., reaction kinetics), then equilibrium conditions suitable for water on the planet’s surface are possible. Figure 5 shows how the sea weathering at equilibrium varies as a function of total carbon dioxide content for different resurfacing velocities and basaltic crust thicknesses.

We consider the total planetary carbon content to be the same as for Earth, estimated at $2.5 \times 10^{22}$ mol C (Sleep & Zahnle 2001) or 215 bar. Io’s estimated resurfacing rate is $0.1–1$ cm yr$^{-1}$ (McEwen et al. 2004; Phillips 2000) and crustal thickness is 14–30 km at a minimum (Jaeger et al. 2003; Carr et al. 1998). We take as our fiducial case a resurfacing velocity of $v = 1$ cm yr$^{-1}$ and a thickness of the basaltic crust layer $d_{\text{bas}} = 10$ km, in alignment with Io, which yields an equilibrium sea weathering rate of 120 bar Gyr$^{-1}$. This is above the range estimated for present-day Earth (1.9–29 bar Gyr$^{-1}$). For comparison, a less active planet, with a resurfacing rate of $v = 0.1$ cm yr$^{-1}$, would require an equilibrium sea weathering rate of 12 bar Gyr$^{-1}$, well within the estimates for present-day Earth.

It is clear that the most important factor is the carbon content of the planet and the resurfacing velocity. The shaded region shows present-day estimates for sea weathering rate on Earth.

If there is a limit to how much sea weathering can take place for a given total carbon content then lower resurfacing velocities are needed in order to have an equilibrium state, while the thickness of the basaltic layer is less important (see Figure 5). In turn, resurfacing velocities depend on the tidal heat flux, or tidal forcing. Thus, if there is a limit to sea weathering, there will be a limit to tidal heating, above which

---

This approach ensures that the reference state of the sea weathering in Equation (6) coincides with the equilibrium of the system.
the planet will not exhibit an equilibrium atmospheric state over long timescales that favors liquid water. It is beyond the scope of this paper to calculate the exact limits, because this would require building a model of thermal evolution for the planet.

Likewise, for a given resurfacing rate set by tidal heating, planets with less C are favored in maintaining habitable surface conditions via tidally induced tectonism.

If instead, we take the value for sea weathering on Earth, \( W_{\text{sea}}^E \), at face value and valid only in the reference state, we can calculate the equilibrium conditions for atmospheric temperature and CO2 pressure for other planets. For this case we make explicit the dependence on the velocity by setting \( \omega = \frac{v}{v_0} \) and take \( v_0 = 3 \text{ cm yr}^{-1} \). We obtain the equilibrium conditions for a given set of resurfacing velocities, basaltic layer thicknesses, and total carbon content by solving for the value of \( p_E \) that satisfies the equation

\[
\frac{p_{\text{CO}_2}^{\text{tot}}}{P_{\text{man}}} = p_E^E \left( 1 + \frac{f d_{\text{man}} A_p}{V_{\text{man}}} \right) + P_E + P_{\text{CO}_2}^{oc} (P_E) \tag{16}
\]

given that

\[
p_{\text{man}}^E = \frac{V_{\text{man}}}{f v A_p E} W_{\text{sea}}^E \frac{v}{v_0} (p_E/p_0)^{\nu} \exp\left( E \left( \frac{1}{T_0} - \frac{1}{T_E} \right) \right) \tag{17}
\]

Figure 6 shows the range of equilibrium temperature values as a function of planetary carbon content for Earth’s average spreading rate of \( v_0 = 3 \text{ cm yr}^{-1} \) and three different estimates of Earth's sea weathering rate: \( W_{\text{sea}} = 0.55 \text{ Tmol/yr} = 1.9 \text{ bar Gyr}^{-1} \) (Krissansen-Totton & Catling 2017), \( W_{\text{sea}} = 1.75 \text{ Tmol/yr} = 15 \text{ bar Gyr}^{-1} \) (Mills et al. 2014), and \( W_{\text{sea}} = 3.4 \text{ Tmol/yr} = 29 \text{ bar Gyr}^{-1} \) (Sleep & Zahnle 2001). It can be seen that planets with modest amounts of total carbon can achieve habitable conditions more easily than those that have greater carbon content. For example, planets with carbon contents a few times that of the Earth can have liquid water (shaded region) when assuming an atmosphere of 1 bar, albeit in hotter conditions.

Figure 7. Evolution of CO2 toward equilibrium for tidally heated planets in the habitable zone. Temperature (top, green) changes as the partial pressure of atmospheric CO2 (blue) evolves. The CO2 content of the atmosphere-ocean reservoir (dashed blue) evolves commensurate with the atmospheric CO2 partial pressure given an assumed solubility independent of temperature. The C content in the basalt (gray) and mantle (red) reaches the equilibrium state at a few tens of millions of years once the sea weathering rate (pink, in bar Myr\(^{-1}\)) and volcanic degassing (brown, in bar Myr\(^{-1}\)) reach parity.

Whichever way sea weathering is calculated, it is clear that planets with low carbon contents and/or sluggish resurfacing velocities are favored for exhibiting a carbon–silicate cycle that keeps the atmospheric temperature habitable.

### 3.2. Equilibrium Timescale

Having established the equilibrium conditions, we proceed to solving the governing equations. Our approach is to set the sea weathering rate to a value that allows for an equilibrium similar to the Earth with \( p_E = 0.01 \text{ bar} \) and \( T_E = 288 \text{ K} \). However, the behavior is expected to be qualitatively similar had we fixed the sea weathering rate to the Earth’s value given the same functional form of the equations.

We first start with a fixed present-day Sun to find out how long it takes to reach the equilibrium state and how it changes with different parameters. We tested different initial conditions (similar to those used by Foley 2015) and show only the most extreme case of disequilibrium where there is no carbon in the mantle to begin with. While this case is not really realistic it does show that the system can recover equilibrium even from these extreme beginnings. We consider the cases (a) where these planets reach snowball states and (b) where they do not. We vary the initial atmospheric CO2 values to allow for cold or hot beginnings.

Figure 7 shows how the system reaches equilibrium starting from a cold state. With no C in the mantle, the system initially starts with no outgassing into the atmosphere. With low temperatures (\( T = 215.5 \text{ K} \) for \( p_0^{\text{atm}} = 0.001 \text{ bar} \), the weathering rate is very small and draws down little CO2 from the atmosphere (and ocean) into the basaltic layer, which starts as a rich C reservoir. Foundering from this basaltic crust slowly starts building the mantle’s C reservoir, as outgassing is outpaced. This little outgassing slowly builds up more CO2 in the atmosphere, so that it heats up and starts melting the ice on the surface. At some point there is enough CO2 in the atmosphere that the planet deglaciates completely; this changes the albedo suddenly to much lower values and the planet transitions into a hot state (the only permanent stable point in the EBM plus evolution equations is at high temperatures). The system overshoots from the equilibrium point because too much CO2 had accumulated in the deglaciation phase. This
overshoot is controlled by the rate of outgassing and the time it has taken the planet to deglaciate. In this very hot state, the sea weathering rate increases by an order of magnitude and quickly draws the excess CO₂ back down into the atmosphere, all the while decreasing C in the basaltic crust and increasing it in the mantle, relaxing into the equilibrium state. Analogous behavior has been discussed in the context of Earth and deglaciation mantles, relaxing into the equilibrium state. Analogous behavior has been discussed in the context of Earth and deglaciation.

If we restrict the ice coverage of the planet to eliminate snowball transitions, following Checlair et al. (2017), then the initial state is less cold (T = 243 K) and no overshooting happens. See Figure 8 top panel. The planet just heats up until it reaches the equilibrium state. Discontinuities in the temperature come from discontinuities in the ground albedo as a function of surface temperature via the amount of land, snow, and ice coverage, which has been modeled in a simple fashion.

The last case we considered was a hot beginning with \( p_{\text{atm}} = 0.05 \) bar (the hottest point allowed in the EBM model for M stars). The evolution from a snowball or no-snowball planet is the same. For this hot beginning, sea weathering starts at a high rate, drawing down CO₂ from the atmosphere, bringing down the surface temperature. In the meantime, the basaltic layer is foundering more C to the mantle than the mantle degasses, so that the mantle’s reservoir builds up. The system also slightly overshoots, but not to a point of glaciation, and then relaxes into the equilibrium state.

We find that the timescale to reach equilibrium is about 10–100 Myr and is independent of total CO₂ content, values of activation energy \( E_{\text{act}} \) or dependence of the sea weathering rate on the amount of atmospheric CO₂, \( \alpha \). The factors that change this timescale are the crustal thickness \( d_{\text{bas}} \), resurfacing velocity \( v \), and outgassing rate. An order-of-magnitude increase in crustal thickness from 10 to 100 km increases the timescale by about one order of magnitude from \( \sim 10 \) to \( \sim 100 \) Myr. An increase in

\[
\tau \approx \frac{d_{\text{bas}}}{v \sqrt{f}}.
\]

On the other hand, larger values of total carbon inventories, while not changing the timescale to reach equilibrium, may affect the timing of complete deglaciation (by a few Myr) while preserving the amount of overshoot in atmospheric CO₂ and thus the atmospheric temperature (to about 330 K). Different values of \( v \) have the same effect on changing the timing of the overshoot, but not the peaks and troughs in temperature. Thus, for reasonable values for the parameters in the model, the timescale to reach equilibrium is 10–100 Myr. Therefore, processes with a shorter or longer timescale are not expected to affect the planet’s ability to sustain equilibrium.

3.3. Long-term Evolution

For completeness, along the same lines, we also looked at how a tidally heated planet may evolve as the M star dims over time. Because the evolution of stars changes on a billion-year timescale, we find that planets reach equilibrium at 10–100 Myr, as expected. Thus, in Figure 9 we show only the evolution from a hot start because it is likely that planets that end up in the habitable zone around an M star started hot (unless there is some mechanism for migration that might have brought them from further out, à la Hansen & Murray 2012). To stay within the bounds of the EBM model of Haqq-Misra et al. (2016) our initial start is at \( p_{\text{atm}} = 0.033 \) bar. After 10–100 Myr the CO₂ contents of the reservoirs reach the equilibrium state for the solar luminosity at the time. Given that the star is brighter in the past, the CO₂ content is lower (\( p_{\text{atm}} = 0.007 \) bar at 2.9 Gyr ago), while the atmospheric temperature (\( T = 289 \) K) is slightly higher than the present-day value. Our choice for starting the evolution 3 Gyr ago is tied to the range in which the EBM model is valid. However, there is no reason to believe that this mechanism would not operate
further into the past at larger insolation values. The limit would be set by reaching the runaway greenhouse. In other words, planets in the habitable zone of M dwarfs may start hot, with a steam atmosphere that partially evaporates while the star is active. As luminosity decreases, the remaining water in the steam atmosphere that partially evaporates while the star is active.

Our simple model shows that tidally locked planets may have a built-in mechanism to regulate the amount of CO2 in the atmosphere to allow for long-term stability of surface liquid water, similar to Earth. However, unlike Earth’s carbon–silicate cycle that relies on plate tectonics, these tidally heated planets recycle material vertically via continuous volcanism and foundering.

3.4. Sub-Earths and Super-Earths

A simple extension of this work is to consider planets that are less or more massive than Earth, while still being rocky. Low-mass exoplanets, including super-Earths are now known to be common in our Galaxy (Howard et al. 2010). We consider planets that have the same major element composition as Earth, including the same core-mass fraction, and assume that these planets experience constant volcanism. We use parameters for planetary radius, gravity, and core radius to account for the different mass according to the internal structure model of Valencia et al. (2006).

In terms of the planetary carbon inventory we consider two scenarios: (1) planets with the same total C content as Earth, and (2) planets with the same C content per unit mass. In reality, due to the volatile character of carbon, we do not fully understand how Earth acquired the amount it has, and how to extrapolate this accurately to other terrestrial planets. Thus our two scenarios may be thought of as possibilities, from which we can draw a few conclusions.

We use the same resurfacing velocity and crustal thickness for all planets, independent of their mass as a first-order approach. The timescale to reach equilibrium remains unchanged, although in the case of cold initial states with snowball transitions, the overshoot is delayed for larger planets. Thus, when considering only the effects of geometry and gravity, our proposed mechanism is robust.

Not surprisingly, when we allow a planet’s carbon inventory to scale with mass, it affects the system’s requirements. If we require an equilibrium around $T = 288 \text{ K}$, planets with lower carbon content require more reasonable sea weathering rates (Figure 10 top). If instead, we take Earth’s sea weathering at face value, smaller planets with lower carbon content experience equilibrium temperatures that are temperate whereas large planets would be too hot (Figure 10 bottom).

In general, in terms of a negative feedback mechanism that regulates surface temperature via CO2 sequestration, there needs to be a balance between outgassing that depends on planetary carbon content and weathering. Thus, lower planetary carbon contents are favored.

Another aspect of planetary mass is how it affects the ability of the planet to advect melt to the surface given the fact that there is a crossover density at some pressure, below which any melt produced is negatively buoyant and stays within the mantle (Ohtani et al. 1995). The depth at which this occurs depends on the gravity, and thus on planet mass. A simple gravity scaling based on Valencia et al. (2006) yields $g \approx g_E (M/M_E)^{1/2}$, which translates to a depth scaling of $h \approx h_E (M/M_E)^{-1/2}$. Thus, smaller planets are more likely to produce melt above the depth at which the crossover density occurs. Following Sakamaki et al. (2006) and this simple scaling (which ignores compression effects), we obtain for a $10 M_E$ planet (with the same core-mass fraction as Earth) depths for the crossover density of 110, 130, and 450 km for anhydrous magmas, and for hydrated magmas at the 2% and 8% levels, respectively. For a 0.1 $M_E$ planet the corresponding depths are 1100 km, 1300 km, and beyond the core–mantle boundary of the planet. Thus, details about the composition of the mantle and melt as well as where tidal dissipation takes place are needed to establish exactly which planets will produce melt that can be advected to the surface and sustain the mechanism we propose. These effects are beyond the scope of this paper and are left for future work. However, qualitatively, melt is more easily advected in smaller planets, with hydrated magmas, and thus makes small wet planets more suitable to exhibit habitability from tidally heated tectonism.

4. Discussion

The simple model we propose has three elements at its core that make for the built-in thermostat: (1) the sea weathering rate depends sensitively on the atmospheric temperature (via controlling the temperature of the deep ocean), (2) the atmospheric CO2 can be drawn out and sequestered into a reservoir when needed, and (3) this reservoir is connected back into releasing CO2 into the atmosphere via outgassing, which in our proposed model happens via foundering of the basaltic
layer into the mantle, which in turn continuously outgasses into the atmosphere by advecting melt onto the surface.

This model does not consider possible limits to the weathering rate. By construction we have invoked a mechanism that continuously exposes fresh basaltic crust available for weathering, in analogy to Earth. However, if volcanism is too infrequent on timescales that are longer than \( \sim 10-100 \text{ Myr} \), it could be a problem for the system to maintain equilibrium. Brady & Gíslason (1997) noted that seafloor weathering on Earth seems to occur mostly within the first 3 Myr and stops 10 Myr after crust production. Thus, for continuous seafloor weathering, sufficient volcanism should happen at least every million years or so.

Also, incipient volcanism may limit the amount of weatherable material in a similar way to the transport-limited scenario explored by West et al. (2005) and West (2012) for Earth. In the transport-limited regime, the replenishment of fresh Ca and Mg ions is the bottleneck to weathering. In the case of the Earth it can be due to low erosion rates (West et al. 2005; West 2012). In rocky exoplanets it can be due to limited land exposure (Foley 2015). In our case it can be due to limited volcanism. Including a transport-limited sea weathering rate would impose a limit on the amount of total planetary carbon and/or a resurfacing rate (which is dependent on tidal heating values) below which the planet has the ability to sequester C at a high enough rate to keep CO\(_2\) atmospheric values below a greenhouse atmosphere. Thus, understanding better what parameters control sea weathering would help us determine how ubiquitous this kind of C–Si cycle can be in exo-Earths.

In addition, we treated seafloor weathering in the simplest way possible. For example, our equation for seafloor weathering lumps alkalinities into the surface temperature dependence, and thus we omit a treatment for ocean chemistry (Krissansen-Totton & Catling 2017). These improvements are left for future work to bring focus to the main ingredients laid out in this study.

Our proposed scenario excluded the existence of rock weathering on continents. However, it may be that orogeny can happen as a secondary process to tidally induced volcanism as suggested on Io (McEwen et al. 2004 and references therein). If so, weathering may proceed both in the continents and on the seafloor.

We note that in our modeling we have taken the parameters for the thickness of the basaltic layer and the resurfacing velocity to be independent quantities, while in reality they are connected via the contribution of tidal heating to the total heat flux of the planet (factor \( \alpha \) in Equation (4)). We have taken this route because to properly assess this contribution one would have to model the thermal history of the planet, including an accurate description of the tidal dissipation parameter \( Q \), and the effect of melt on it, which is beyond the scope of this study.

Another refinement to our simple model may be to include the effects of phase transitions happening within the basaltic layer. On Earth, the basalt to eclogite transition may cause delamination within a thick crust and cool the surrounding mantle faster than otherwise, and limit the size of the basaltic crust. Including this effect on our model would require adding another layer between the basaltic crust and the mantle from which outgassing proceeds. Because of mass balance, adding another layer would not change the character of the equations, and thus the results presented.

Future work may be extended to include calculations of thermal history, sweeping of parameter space for planets in different orbital configurations, and different carbon contents, and may include the effects of transport-limited sea weathering via limited volcanism.

If our proposed mechanism to regulate atmospheric CO\(_2\) takes place on planets such as Trappist-1 d the CO\(_2\) content in the atmosphere would be around values that would yield liquid water on its surface, namely \( \sim 0.5 \text{ bar} \), assuming the planet is abiotic. Indeed, if there is a negative feedback mechanism that enables a thermostat taking place on other planets, enabled by either the mechanism proposed here or by plate tectonics, or even perhaps in stagnant lid (Foley & Smye 2017), the CO\(_2\) content of the atmosphere has to be commensurate with insolation values, something we can test for with enough atmospheric data. However, because planets that are substantially tidally heated are getting rid of heat via volcanic pipes, we can be guided by estimates of the tidal dissipation from orbital dynamics to pinpoint whether or not we expect pipe-heating to occur instead of plate tectonics or stagnant lid.

5. Summary

In summary, we propose a new mechanism for rocky planets around M dwarfs to have a climate-controlling feedback mechanism that can keep liquid water stable for billions of years. An analogous carbon–silicate cycle can operate on planets by recycling carbon between the atmosphere, a basaltic crust, and the mantle via tidally induced volcanism, basaltic formation, and sea weathering, plus foundering. In contrast to plate tectonics, which enables the carbon–silicate cycle on Earth, these planets would achieve the same principle by recycling material vertically.

Basaltic crust exposed continuously through volcanism can be weathered to sequester C from the ocean and atmosphere, to draw down atmospheric levels of CO\(_2\) when values are too high, or outgas CO\(_2\) from the mantle to replenish atmospheric values when they are too low, as long as sea weathering depends on atmospheric temperature.

Therefore, the tidal properties of planets around M dwarfs, in the absence of plate tectonics, may enable stable climates suitable for habitability, making the search for these planets more attractive than it already is.

The equilibrium timescale of \( \sim 10-100 \text{ Myr} \) underlying this mechanism \( (\tau \approx \tau_{\text{bas}} / \sqrt{f}) \) is very different from the timescales for evolution of stellar brightness or flares, thus remaining impervious to these changes.

This mechanism may be tested by retrieving atmospheric CO\(_2\) values from planets with nonzero eccentricity in the habitable zone that are dissipating too much heat via pipe heating to exhibit plate tectonics. If this type of tectonism is happening and controlling climate, the values for CO\(_2\) should be consistent with insolation values. Trappist-1 planets in the habitable zone may be an example.

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Appendix

Linear Stability and Equilibrium Timescale

The system of ordinary differential equations that govern this system (Equation (10)) can be non-dimensionalized to yield

$$
\frac{d}{dt'} p_{CO_2}^{atm+oc} = \frac{W_{sea}}{p_{E}^{ao}} \tau (p_{CO_2}^{man} - (p_{CO_2}^{atm})^\circ) 
\times \exp\left(\frac{E}{R} \left(\frac{1}{T} - \frac{1}{T'}\right)\right)
$$

$$
\frac{d}{dt'} p_{CO_2}^{man} = \frac{fA_p}{V_{man} d_{bas}} \frac{p_{E}^{ao}}{W_{sea}^{E}} (1 - p_{CO_2}^{atm+oc})
+ \tau \left(\frac{fV_{p}}{V_{man} + V} + \frac{v}{d_{bas}}\right) (1 - p_{CO_2}^{man})
$$

with non-dimensional variables

$$
p_{CO_2}^{atm+oc} = \frac{p_{CO_2}^{atm+oc}}{p_{E}^{ao}}
$$

$$
p_{CO_2}^{atm} = \frac{p_{CO_2}^{atm}}{p_{E}}
$$

$$
p_{CO_2}^{man} = \frac{p_{CO_2}^{man}}{p_{man}^{ao}}
$$

$$
p_{CO_2}^{oc} = \frac{p_{CO_2}^{oc}}{p_{E}^{ao}}
$$

$$
t' = t/\tau.
$$

After inspection of the numerical results from solving the dimensional equations, as well as combining both equations in (A1) into a second-order differential equation for $p_{CO_2}^{man}$ and retaining the non-derivative terms, we conclude that the most appropriate timescale is

$$
\tau = \frac{1}{\sqrt{\frac{d_{bas} V_{man}}{fA_p}}},
$$

which in the limit of a thin basaltic crust layer $d_{bas}/R_p \ll 1$ becomes

$$
\tau = \frac{d_{bas}}{v} \frac{1}{\sqrt{f}}.
$$

The system of ordinary differential equations that govern this system (Equation (10)) can be non-dimensionalized to yield

$$
\frac{d}{dt'} p_{CO_2}^{atm+oc} = R_1 p_{CO_2}^{man} - (p_{CO_2}^{atm})^\circ
\times \exp\left(\frac{E}{R} \left(\frac{1}{T} - \frac{1}{T'}\right)\right)
$$

$$
\frac{d}{dt'} p_{CO_2}^{man} = \frac{1}{R_1} (1 - p_{CO_2}^{atm+oc})
+ R_2 (1 - p_{CO_2}^{man})
$$

where the dimensionless groups related to this problem are

$$
R_1 = \frac{W_{sea}^{E}}{p_{E}^{ao}} \frac{d_{bas} V_{man}}{fA_p}
$$

$$
R_2 = \left(\frac{fV_{p}}{V_{man} + V} + \frac{v}{d_{bas}}\right) \tau.
$$

In non-dimensional form it is easy to see the combination of parameters that govern the equation. For example, $\frac{v}{W_{sea}^{E}}$ comes as a block, as well as $\frac{d_{bas}}{d_{bas}}$ in the limit of a thin basaltic shell.

We are interested in determining whether the equilibrium point $p_{CO_2}^{atm} = p_{CO_2}^{atm+oc} = p_{CO_2}^{man} = 1$ and $p_{CO_2}^{oc} = 1 - \frac{p_{E}^{ao}}{p_{E}^{ao}}$ is stable.

Evaluating the Jacobian at this fixed point, we obtain

$$
J = \begin{bmatrix}
-A & R_1 \\
-1/R_1 & -R_2
\end{bmatrix}
$$

where

$$
A = R_1 \frac{p_{E}^{ao}}{p_{E}} \left(\alpha + \frac{E}{RT_E} \frac{dT}{p_{CO_2}^{atm}}\right) \left(1 - \frac{dp_{CO_2}^{oc}}{dp_{CO_2}^{atm+oc}}\right),
$$

which in the limit of thin basaltic crust layer becomes

$$
R_1 = \frac{W_{sea}^{E}}{p_{E}^{ao}} \frac{d_{bas}}{fA_p}
$$

$$
R_2 = \left(\frac{v}{d_{bas}}\right)^2 \frac{1+f}{f}.
$$

$$
K_H \mu_{CO_2} \mu_{H_2O} P_{H_2O} + 2 \mu_{CO_2}^2 P_{H_2O}^2 + 2 \mu_{CO_2} \mu_{H_2O} P_{H_2O} P_{E}^{ao} p_{CO_2}^{oc} + \mu_{H_2O}^2 P_{E}^{ao} P_{E}^{ao} p_{CO_2}^{oc2},
$$

a quantity that is always positive.

The derivative of temperature with respect to atmospheric carbon dioxide is always positive, so $A$ is always a positive quantity.

Obtaining the eigenvalues of the Jacobian matrix, we find

$$
\lambda_{+,-} = \frac{-1}{2}(A + R_2) \pm \sqrt{(A - R_2)^2 - 4}.
$$

There are two possibilities: the term under the square root is either negative or positive. If it is negative, given that $A > 0$ and $R_2 > 0$, the real part of the eigenvalue is negative and the equilibrium point is a stable solution even if there is decaying...
oscillatory behavior around it. If the term in the square root is positive, then we have to determine in which cases the eigenvalues are negative. Trivially, $\lambda$, is always negative. For $\lambda_n$, the condition is that $A + R > \sqrt{(R^2 - A^2)} - 4$ yields negative eigenvalues. As both quantities are non-negative, the condition can be reduced to $AR > -1$, which is always satisfied.

Therefore, we conclude that the equilibrium point is stable regardless of what assumptions are made about the amount of tidal heating reflected in values for $v$ and $d_{\text{bas}}$.

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