Key Points:
- Climate, mantle, and core dynamics are coupled
- Coupling between climate, mantle, and core can strongly influence evolution
- Various evolutionary scenarios are possible due to coupling

Correspondence to:
B. J. Foley,
bfoley@carnegiescience.edu

Whole planet coupling between climate, mantle, and core: Implications for rocky planet evolution

Bradford J. Foley[1] and Peter E. Driscoll[1]

[1]Department of Terrestrial Magnetism, Carnegie Institution for Science, Washington, District of Columbia, USA

Abstract Earth's climate, mantle, and core interact over geologic time scales. Climate influences whether plate tectonics can take place on a planet, with cool climates being favorable for plate tectonics because they enhance stresses in the lithosphere, suppress plate boundary annealing, and promote hydration and weakening of the lithosphere. Plate tectonics plays a vital role in the long-term carbon cycle, which helps to maintain a temperate climate. Plate tectonics provides long-term cooling of the core, which is vital for generating a magnetic field, and the magnetic field is capable of shielding atmospheric volatiles from the solar wind. Coupling between climate, mantle, and core can potentially explain the divergent evolution of Earth and Venus. As Venus lies too close to the sun for liquid water to exist, there is no long-term carbon cycle and thus an extremely hot climate. Therefore, plate tectonics cannot operate and a long-lived core dynamo cannot be sustained due to insufficient core cooling. On planets within the habitable zone where liquid water is possible, a wide range of evolutionary scenarios can take place depending on initial atmospheric composition, bulk volatile content, or the timing of when plate tectonics initiates, among other factors. Many of these evolutionary trajectories would render the planet uninhabitable. However, there is still significant uncertainty over the nature of the coupling between climate, mantle, and core. Future work is needed to constrain potential evolutionary scenarios and the likelihood of an Earth-like evolution.

1. Introduction

1.1. Overview

Recent discoveries have revealed that rocky exoplanets are relatively common [*Batalha*, 2014]. As a consequence, determining the factors necessary for a rocky planet to support life, especially life that may be remotely observable, has become an increasingly important topic. A major requirement, that has been extensively studied, is that solar luminosity must be neither too high nor too low for liquid water to be stable on a planet's surface; this requirement leads to the concept of the "habitable zone," the range of orbital distances where liquid water is possible [*Hart*, 1978, 1979; *Kasting et al.*, 1993b; *Franck et al.*, 2000; *Kopparapu et al.*, 2014]. Inward of the habitable zone's inner edge, the critical solar flux that triggers a runaway greenhouse effect is exceeded. The critical flux is typically estimated at ≈ 300 W m^{-2}, with variations of $\sim 10-100$ W m^{-2} possible due to atmospheric composition, planet size, or surface water inventory [*Ingersoll*, 1969; *Kasting*, 1988; *Nakajima et al.*, 1992; *Abe et al.*, 2011; *Goldblatt et al.*, 2013]. In a runaway greenhouse state, liquid water cannot condense out of the atmosphere, so any water present would exist as steam. Furthermore, a runaway greenhouse is thought to cause rapid water loss to space, and can thus leave a planet desiccated [*Kasting*, 1988; *Hamano et al.*, 2013; *Wordsworth and Pierrehumbert*, 2013]. Beyond the outer edge, insolation levels are so low that no amount of CO_2 can keep surface temperatures above freezing [*Kasting et al.*, 1993b].

However, lying within the habitable zone does not guarantee that surface conditions will be suitable for life. Variations in atmospheric CO_2 content can lead to cold climates where a planet is globally glaciated, or hot climates where surface temperatures are higher than any known life can tolerate (i.e., above ≈ 400 K [*Takai et al.*, 2008]). A hot CO_2 greenhouse can also cause rapid water loss to space [*Kasting*, 1988] (though *Wordsworth and Pierrehumbert* [2013] argues against this), or even a steam atmosphere if surface temperatures exceed water's critical temperature of 647 K. Moreover, the solar wind can strip the atmosphere of water and expose the surface to harmful radiation, unless a magnetic field is present to shield the planet [e.g., *Kasting and Catling*, 2003; *Griessmeier et al.*, 2009; *Brain et al.*, 2014].

Atmospheric CO_2 concentrations are regulated by the long-term carbon cycle on Earth, such that surface temperatures have remained temperate throughout geologic time [e.g., *Walker et al.*, 1981; *Berner*, 2004].
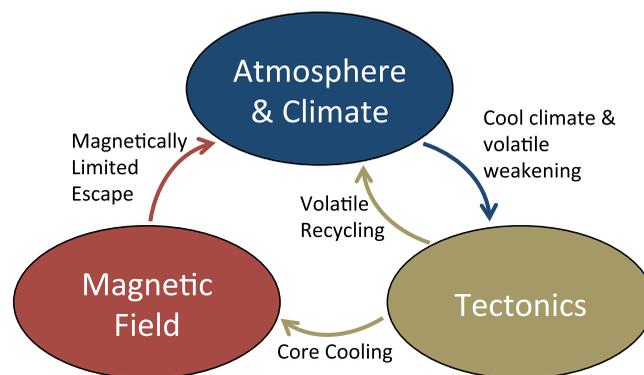
Figure 1. Flowchart representing the concept of whole planet coupling. Climate influences tectonics through the role of surface temperature in a planet's tectonic regime (i.e., stagnant lid versus plate tectonics), while the tectonic regime in turn affects climate through volatile cycling between the surface and interior. The tectonic regime also influences whether a magnetic field can be generated by dictating the core cooling rate. Finally, the strength of the magnetic field influences atmospheric escape, and therefore long-term climate evolution.

The long-term carbon cycle is facilitated by plate tectonics [e.g., *Kasting and Catling*, 2003]. Furthermore, the magnetic field is maintained by convection in Earth's liquid iron outer core (i.e., the geodynamo). As a result, interior processes, namely the operation of plate tectonics and the geodynamo, are vital for habitability. However, whether plate tectonics or a strong magnetic field is likely on rocky planets, especially those in the habitable zone where liquid water is possible, is unclear. The four rocky planets of our solar system have taken dramatically different evolutionary paths, with only Earth developing into a habitable planet possessing liquid water oceans, plate tectonics, and a strong, internally generated magnetic field. In particular, the contrast between Earth and Venus, which is approximately the same size as Earth and has a similar composition yet lacks plate tectonics, a magnetic field, and a temperate climate, is striking.

In this review, we synthesize recent work to highlight that plate tectonics, climate, and the geodynamo are coupled, and that this "whole planet coupling" between surface and interior places new constraints on whether plate tectonics, temperate climates, and magnetic fields can develop on a rocky planet. We hypothesize that whole planet coupling can potentially explain the Earth-Venus dichotomy, as it allows two otherwise similar planets to undergo drastically different evolutions, due solely to one lying inward of the habitable zone's inner edge and the other within the habitable zone. We also hypothesize that whole planet coupling can lead to a number of different evolutionary scenarios for habitable zone planets, many of which would be detrimental for life, based on initial atmospheric composition, planetary volatile content, and other factors. We primarily focus on habitable zone planets, as these are most interesting in terms of astrobiology, and because the full series of surface-interior interactions we describe involves the long-term carbon cycle, which requires liquid water. Each process, the generation of plate tectonics from mantle convection, climate regulation due to the long-term carbon cycle, and dynamo action in the core, is still incompletely understood and the couplings between these processes are even more uncertain. As a result, significant future work will be needed to place more quantitative constraints on the evolutionary scenarios discussed in this review.

1.2. Whole Planet Coupling

Several basic concepts illustrate the coupling between the surface and interior (Figure 1). (1) Climate influences whether plate tectonics can take place on a planet. (2) Plate tectonics plays a vital role in the long-term carbon cycle, which helps to maintain a temperate climate. (3) Plate tectonics affects the generation of the magnetic field via core cooling. (4) The magnetic field is capable of shielding the atmosphere from the solar wind. Cool climates are favorable for plate tectonics because they facilitate the formation of weak lithospheric shear zones, which are necessary for plate tectonics to operate. Low surface temperatures suppress rock annealing, increase the negative buoyancy of the lithosphere, and allow for deep cracking and subsequent water ingestion into the lithosphere, all of which promote the formation of weak shear zones. When liquid water is present on a planet's surface, silicate weathering, the primary sink of atmospheric CO_2 and thus a key component of the long-term carbon cycle, is active. However, silicate weathering also requires a sufficient supply of fresh, weatherable rock at the surface, which plate tectonics helps to provide via orogeny and uplift. As a result, the coupling between plate tectonics and climate can behave as a negative feedback mechanism in some cases, where a cool climate promotes the operation of plate tectonics, and plate tectonics enhances silicate weathering such that the carbon cycle can sustain cool climate conditions.

An additional coupling comes into play via the core dynamo and the magnetic field. The magnetic field is generated by either thermal or chemical convection in the liquid iron core. Thermal convection requires a superadiabatic heat flux out of the core, which is controlled in part by the style of mantle convection, while chemical convection is driven by light element release during inner core nucleation, which also relies on cooling of the core. Plate tectonics cools the mantle efficiently by continuously subducting cold slabs into the deep interior, thus maintaining a high heat flow out of the core. The magnetic field can in turn limit atmospheric escape, helping retain liquid surface water. The coupling between plate tectonics and the core dynamo, and the magnetic field and the climate, completes the concept of whole planet coupling (Figure 1). Moreover, the magnetic field and plate tectonics can also act as a negative feedback in cases where magnetic shielding is required to prevent rapid planetary water loss, and plate tectonics is needed to drive the dynamo.

1.3. Applications to the Evolution of Rocky Exoplanets

The concept of whole planet coupling discussed in this review is based on our knowledge of the Earth and the other rocky planets in the solar system, and is therefore limited to planets of a particular composition; i.e., planets made up mainly of silicate mantles and iron cores. It is unknown whether any of the processes discussed here are applicable to planets with more exotic compositions, such as planets primarily composed of carbides rather than silicates [e.g., *Madhusudhan et al.*, 2012]. Furthermore, we focus on H_2O and CO_2 as key volatiles; CO_2 is an important greenhouse gas for stabilizing planetary climate, and water is crucial for driving the carbon cycle, may play a role in the operation of plate tectonics, and is thought to be a necessary ingredient for life. Thus, planets must accrete a significant supply of both H_2O and CO_2 for whole planet coupling to be possible. However, volatile accretion is unlikely to be a problem as simulations indicate that planets commonly acquire large volatile inventories (i.e., Earth-like or larger) during late stage planet formation [*Morbidelli et al.*, 2000; *Raymond et al.*, 2004].

Another important consideration is the redox state of the mantle, which determines whether degassing via planetary magmatism releases H_2O and CO_2 to the atmosphere or reduced species such as H_2 and CH_4 [e.g., *Kasting et al.*, 1993a]. Carbon dioxide is the only major greenhouse gas known to be regulated by negative feedbacks such that it has a stabilizing influence on climate. Thus, having CO_2 as a primary greenhouse gas is important for the whole planet coupling discussed here to operate. An oxidized upper mantle favors CO_2 over CH_4 and other reduced gases. Earth's mantle has been oxidized at present-day levels since at least the early Archean [*Delano*, 2001], and possibly even since the Hadean [*Trail et al.*, 2011]. Oxidation of the mantle is thought to occur by disproportionation of FeO to Fe_2O_3-bearing perovskite and iron metal in the lower mantle during accretion and core formation. The iron metal is then lost to the core leaving behind oxidized perovskite that mixes with the rest of the mantle [*Frost et al.*, 2008; *Frost and McCammon*, 2008]. Disproportionation of FeO is expected to occur on rocky planets Earth sized or larger [*Wade and Wood*, 2005; *Wood et al.*, 2006], so CO_2 is likely to be an important greenhouse gas on exoplanets. Though other greenhouse gases can still be important, any planet where significant amounts of CO_2 are degassed by mantle volcanism will need silicate weathering to act as a CO_2 sink to avoid extremely hot climates.

1.4. Outline

The paper is structured as follows. We review the basic physics behind the generation of plate tectonics from mantle convection, highlighting the role of climate, in section 2. We then review the long-term carbon cycle and its influence on climate, and detail the specific ways plate tectonics is important for the operation of this cycle, in section 3. Next, we review magnetic field generation via a core dynamo and the physics of atmospheric shielding and volatile retention in section 4. In section 5, we integrate the discussion of the previous three sections to describe the range of evolutionary scenarios that can result from differences in orbital distance, initial climate state, volatile inventory, and other factors as a consequence of whole planet coupling. We conclude by listing some major open questions that must be addressed in order to further our understanding of how terrestrial planets evolve and the geophysical factors that influence habitability (section 6).

2. Generation of Plate Tectonics From Mantle Convection and the Role of Climate

2.1. General Physics of Plate Generation

Plate tectonics is the surface expression of convection in the Earth's mantle: the lithosphere, composed of the individual plates, is the cold upper thermal boundary layer of the convecting mantle, subducting slabs are

convective downwellings, and plumes are convective upwellings [*Davies*, 1999; *Bercovici et al.*, 2015]. However, mantle convection does not always lead to plate tectonics, as evidenced by Mercury, Mars, and Venus, each of which is thought to have a convective mantle but lack plate tectonics [e.g., *Breuer and Moore*, 2007]. On these planets, "stagnant lid convection," where the lithosphere acts as a rigid, immobile lid lying above the convecting mantle [e.g., *Ogawa et al.*, 1991; *Davaille and Jaupart*, 1993; *Solomatov*, 1995], is thought to operate [*Strom et al.*, 1975; *Phillips et al.*, 1981; *Solomon et al.*, 1992; *Solomatov and Moresi*, 1996, 1997; *Spohn et al.*, 2001; *O'Neill et al.*, 2007a], though Venus may experience occasional, short-lived episodes of subduction [*Turcotte*, 1993]. Stagnant lid convection is a result of the temperature dependence of mantle viscosity, which increases by many orders of magnitude as temperature decreases from the hot conditions that prevail in the planetary interior to the much cooler temperatures that prevail at the surface [e.g., *Karato and Wu*, 1993; *Hirth and Kohlstedt*, 2003]. When the viscosity at the surface is $\sim 10^3 - 10^4$ times that of the interior, the top thermal boundary layer is so viscous that it can no longer sink under its own weight and form subduction zones, and stagnant lid convection ensues [e.g., *Richter et al.*, 1983; *Christensen*, 1984; *Ogawa et al.*, 1991; *Davaille and Jaupart*, 1993; *Moresi and Solomatov*, 1995; *Solomatov*, 1995]. Laboratory experiments indicate that the viscosity ratio between the surface and mantle interior for typical terrestrial planets, including Earth, is $\sim 10^{10} - 10^{20}$, far larger than the threshold for stagnant lid convection. Stagnant lid convection is thus the "natural" state for rocky planets, and additional processes are necessary for plate tectonics to operate on Earth.

In order to generate plate-tectonic style mantle convection, rheological complexities capable of forming weak, localized shear zones in the high-viscosity lithosphere are necessary. These weak shear zones (essentially plate boundaries) remove the strong viscous resistance to surface motion that temperature-dependent viscosity causes and thus allow subduction and plate motions to occur. We discuss plate boundary formation in terms of viscous weakening in the lithosphere because the majority of the lithosphere, including the region of peak strength in the midlithosphere, deforms via ductile or semibrittle/semiductile behavior; the lithosphere only "breaks" in a brittle fashion at shallow depths of $<10 - 20$ km [*Bercovici et al.*, 2015]. Many mechanisms have been proposed for producing the lithospheric weakening and shear localization necessary for plate-tectonic style mantle convection [see reviews by *Tackley*, 2000a; *Bercovici*, 2003; *Bercovici and Karato*, 2003; *Regenauer-Lieb and Yuen*, 2003; *Bercovici et al.*, 2015]; while we will not detail each of these mechanisms, they do all share some general principles. Namely, shear-thinning non-Newtonian rheologies have been found to be successful at generating plate-like mantle convection.

In a shear-thinning fluid, the strain rate, $\dot{\varepsilon}$, is a nonlinear function of the stress, τ, (where $\dot{\varepsilon}$ and τ are the second invariants of the stress and strain rate tensors, respectively):

$$\dot{\varepsilon} \propto \tau^n, \tag{1}$$

where n is a constant. As the effective viscosity, μ_{eff}, is proportional to $\tau \dot{\varepsilon}^{-1}$,

$$\mu_{eff} \propto \tau^{(1-n)} \propto \dot{\varepsilon}^{\frac{(1-n)}{n}}. \tag{2}$$

With $n > 1$, viscosity decreases as stress or strain rate increases and high stress, high strain rate regions of the lithosphere, such as regions of compression or tension caused by drips breaking off the base of the lithosphere, are weakened (see Figure 2). Thus, lithospheric weakening occurs in exactly the regions necessary for facilitating subduction. Shear-thinning behavior can result from inherently non-Newtonian rheologies with $n > 1$, or more complicated rheologies that involve "memory," where a time-evolving state variable, itself a function of stress and/or strain rate, influences viscosity. In the case of memory, an effectively shear-thinning $\dot{\varepsilon} - \tau$ relationship can be calculated by taking the state variable in steady state (as we show for grainsize reduction in section 2.1.2). Moreover, rheologies where stress decreases at high strain rates, resulting in even stronger weakening, are possible but beyond the scope of this paper.

2.1.1. Viscoplasticity

One popular plate generation mechanism is the viscoplastic rheology, where the lithosphere "fails" and becomes mobilized when its intrinsic strength, the yield stress, is reached [*Fowler*, 1993; *Moresi and Solomatov*, 1998]. Typical implementations of the viscoplastic rheology assume that the mantle behaves like a Newtonian fluid (following (1) and (2) with $n = 1$) until stress becomes equal to the yield stress, τ_y (see Figures 2c and 2d). At this point, stress is independent of strain rate, fixed at τ_y, and the viscosity follows

$$\mu_{eff} \propto \frac{\tau_y}{\dot{\varepsilon}}. \tag{3}$$
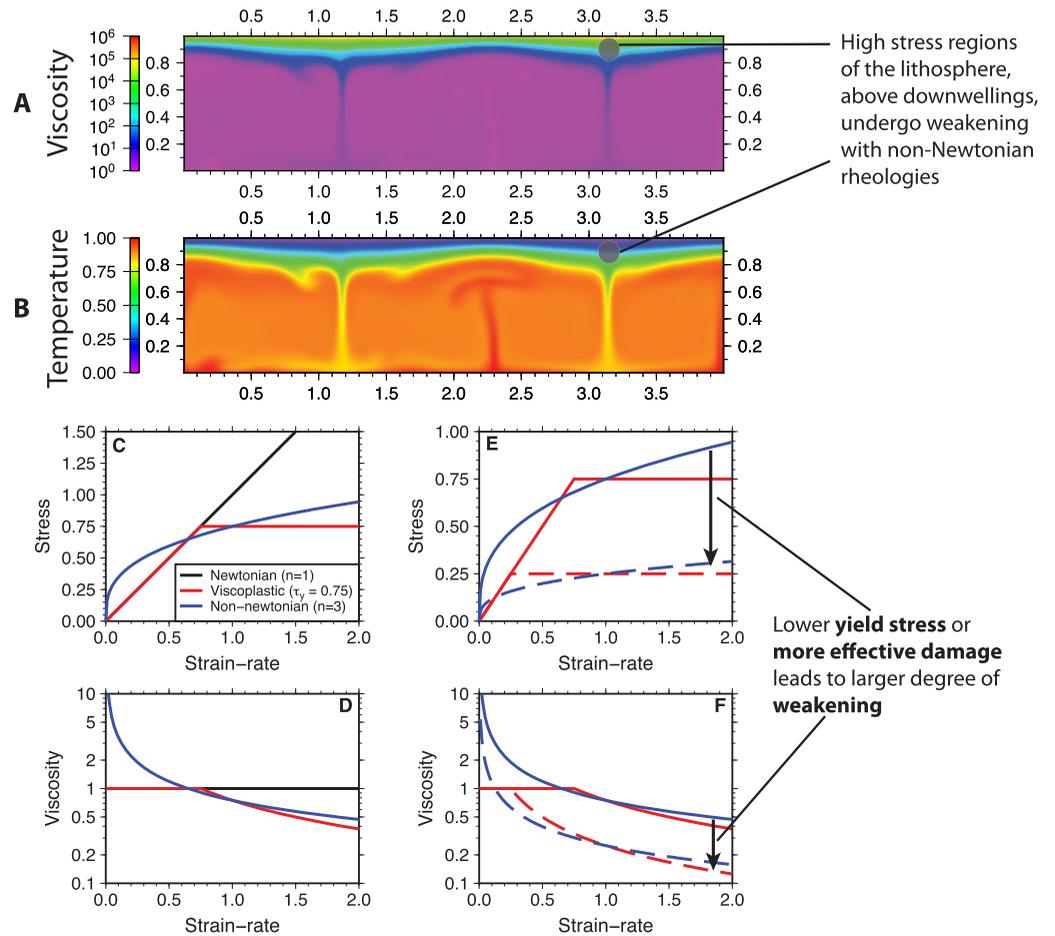
Figure 2. Illustration of plate generation using non-Newtonian (or effectively non-Newtonian) rheologies. Shown are a typical (a) viscosity field and (b) temperature field for stagnant lid convection and schematic (c,e) strain rate versus stress and (d, f) viscosity versus strain rate curves for viscoplastic (see section 2.1.1) and $n = 3$ power law (equations (1) and (2)) rheologies. Weakening of the high viscosity lithosphere present during stagnant lid convection is needed in order to allow subduction, and with non-Newtonian rheologies such weakening will occur at high stress regions (highlighted in grey in Figures 2a and 2b). Figures 2c and 2d illustrate how non-Newtonian rheologies cause viscous weakening at high stresses or strain rates, while Figures 2e and 2f schematically demonstrate how outside conditions that influence the effectiveness of a given plate generation mechanism change the (e) strain rate versus stress curve and the (f) viscosity versus strain rate curve.

When convective stresses in the lithosphere hit the yield stress, low-viscosity shear zones at convergent and divergent margins can form, and plate-like mantle convection can be obtained in two-dimensional [*Moresi and Solomatov*, 1998; *Richards et al.*, 2001; *Korenaga*, 2010b], three-dimensional Cartesian [*Trompert and Hansen*, 1998; *Tackley*, 2000b; *Stein et al.*, 2004], and three-dimensional spherical convection models [*van Heck and Tackley*, 2008; *Foley and Becker*, 2009]. Convective stresses lower than the yield stress do not form weak plate boundaries, and thus result in stagnant lid convection.

The convective stress, τ_m, has typically been found to scale as

$$\tau_m \propto \mu_m \dot{\varepsilon}_m, \tag{4}$$

where μ_m and $\dot{\varepsilon}_m$ are the effective viscosity and strain rate in the mantle, respectively [e.g., *O'Neill et al.*, 2007b]. The strain rate can be estimated as v_m/d, where v_m is convective velocity and d is the thickness of the mantle. Velocity scales as [e.g., *Solomatov*, 1995; *Turcotte and Schubert*, 2002]

$$v_m \propto Ra^{\frac{2}{3}}, \tag{5}$$

where the Rayleigh number (Ra) is defined as

$$Ra = \frac{\rho g \alpha \Delta T d^3}{\kappa \mu_m}, \tag{6}$$

and describes the convective vigor of the system (ρ is density, g gravity, α thermal expansivity, ΔT the non-adiabatic temperature drop across the mantle, and κ the thermal diffusivity).

However, despite a large Rayleigh number on the order of $10^7 - 10^8$, typical convective stresses are only $\sim 1 - 100$ MPa [e.g., *Tackley*, 2000b; *Solomatov*, 2004; *O'Neill and Lenardic*, 2007], approximately an order of magnitude lower than the strength of the lithosphere based on laboratory experiments [e.g., *Kohlstedt et al.*, 1995]. Thus, much work has focused on mechanisms for decreasing the lithospheric strength, usually involving water [e.g., *Tozer*, 1985; *Lenardic and Kaula*, 1994; *Regenauer-Lieb et al.*, 2001; *van der Lee et al.*, 2008], though some recent models have looked at mechanisms for increasing convective stresses [*Rolf and Tackley*, 2011; *Höink et al.*, 2012]. In section 2.2, we describe how the possibility of water weakening links the operation of plate tectonics to climate. Moreover, climate also influences convective stresses, by altering ΔT, providing another important link between climate and tectonics.

2.1.2. Damage and Grainsize Reduction

While viscoplasticity has shown some success at generating plate-like mantle convection, it fails to form localized strike-slip faults [*Tackley*, 2000b; *van Heck and Tackley*, 2008; *Foley and Becker*, 2009], a problem with all simple non-Newtonian rheologies [*Bercovici*, 1993]. Furthermore, viscoplasticity is an instantaneous rheology: when stresses fall below the yield stress weakened lithosphere instantly regains its strength. On Earth, however, dormant weak zones are known to persist over geologic time scales, and can serve as nucleation points for new subduction zones [*Toth and Gurnis*, 1998; *Gurnis et al.*, 2000]. An alternative mechanism, that allows for dormant weak zones (i.e., memory of past deformation) and is capable of generating strike-slip faults, is grainsize reduction [*Bercovici and Karato*, 2003; *Bercovici and Ricard*, 2012, 2013, 2014]. Grainsize reduction is commonly seen in field observations of exhumed lithospheric shear zones [*White et al.*, 1980; *Drury et al.*, 1991; *Jin et al.*, 1998; *Warren and Hirth*, 2006; *Skemer et al.*, 2010], and experiments show that viscosity is proportional to grainsize when deformation is dominated by diffusion creep or grain boundary sliding [e.g., *Hirth and Kohlstedt*, 2003].

One major problem for a grainsize reduction plate generation mechanism, though, is that grainsize reduction occurs by dynamic recrystallization, which takes place in the dislocation creep regime, while grainsize sensitive flow only occurs in the diffusion creep or grain boundary sliding regimes [e.g., *Etheridge and Wilkie*, 1979; *De Bresser et al.*, 1998; *Karato and Wu*, 1993]. Thus, a feedback mechanism, where deformation causes grainsize reduction, which weakens the material and leads to more deformation, is apparently not possible [e.g., *De Bresser et al.*, 2001]. Without such a feedback, only modest weakening can occur. However, a new theory argues that when a secondary phase (e.g., pyroxene) is dispersed throughout the primary phase (e.g., olivine), a grainsize feedback is possible, because grainsize reduction can continue in the diffusion creep regime due the combined effects of damage to the interface between phases and Zener pinning (where the secondary phase blocks grain growth of the primary phase) [*Bercovici and Ricard*, 2012].

Using a simplified version of the theory in *Bercovici and Ricard* [2012] [see also *Foley and Bercovici*, 2014], the grainsize reduction mechanism can be formulated as:

$$\frac{dA}{dt} = \frac{f_A}{\gamma} \Psi - h A^p, \tag{7}$$

where A is the fineness (or inverse grainsize), γ is surface tension, h is the grain-growth (or healing) rate, p is a constant, Ψ is the deformational work, defined as $\Psi = \varepsilon_{ij} \tau_{ij}$, where ε_{ij} is the strain rate tensor and τ_{ij} is the stress tensor, and f_A is the fraction of deformational work that partitions into surface energy, thus driving grainsize reduction. The grain-growth rate is strongly temperature dependent, and is given by

$$h = h_n \exp\left(\frac{-E_h}{R_g T}\right), \tag{8}$$

where h_n is a constant, E_h is the activation energy for grain growth, R_g is the universal gas constant, and T is temperature. The effective viscosity is

$$\mu_{eff} = \mu_n \exp\left(\frac{E_v}{R_g T}\right)\left(\frac{A}{A_0}\right)^{-m} = \mu(T)\left(\frac{A}{A_0}\right)^{-m}, \tag{9}$$

where μ_n and m are constants, E_v is the activation energy for diffusion creep, and A_0 is the reference fineness. Taking (7) in steady state, using the viscous constitutive law $\tau_{ij} = 2\mu_{eff}\dot{\varepsilon}_{ij}$, and the definition $\dot{\varepsilon}_{ij}\dot{\varepsilon}_{ij} = 2\dot{\varepsilon}^2$, the fineness can be written as a function of strain rate

$$A = \left(\frac{4f_A \mu(T) \dot{\varepsilon}^2 A_0^m}{\gamma h} \right)^{\frac{1}{p+m}}. \tag{10}$$

Combining (9) and (10), the viscosity is also a function of strain rate,

$$\mu_{eff} = \left(\mu(T)^p \left(\frac{4f_A \dot{\varepsilon}^2}{\gamma h A_0^p} \right)^{-m} \right)^{\frac{1}{p+m}}, \tag{11}$$

and using $\tau = 2\mu_{eff}\dot{\varepsilon}$, the effective $\tau - \dot{\varepsilon}$ relationship is

$$\dot{\varepsilon} = \frac{1}{2} \left(\mu(T)^{-p} \left(\frac{f_A}{\gamma h A_0^p} \right)^m \tau^{p+m} \right)^{\frac{1}{p-m}}. \tag{12}$$

With typical parameters of $m = 2$ and $p = 4$ [*Hirth and Kohlstedt*, 2003; *Bercovici and Ricard*, 2012], the grainsize reduction mechanism results in an effectively non-Newtonian rheology with $n = 3$ (Figures 2c and 2d). Furthermore, as detailed in section 2.2, climate can have a significant impact on plate generation with this mechanism, because grain growth is temperature dependent. At high surface temperatures, and thus higher temperatures in the midlithosphere, faster grain growth (i.e., higher h) impedes grainsize reduction.

2.2. Influence of Climate on Tectonic Regime

Both the viscoplastic and grainsize reduction mechanisms for generating plate-tectonic style mantle convection are linked to climate. Although we treat these mechanisms separately, they are not mutually exclusive. The viscoplastic rheology is most relevant to brittle deformation in the upper lithosphere, while grainsize reduction provides viscous weakening in the lower lithosphere. As a result both mechanisms can operate simultaneously, and both may be important for generating plate tectonics.

Climate can influence whether plate-like convection occurs with a viscoplastic rheology in three main ways: first, climate dictates whether water can exist in contact with ocean lithosphere at the surface, such that high pore pressure [e.g., *Hubbert and Rubey*, 1959; *Sibson*, 1977; *Rice*, 1992; *Sleep and Blanpied*, 1992] and weak, hydrous phases [*Escartín et al.*, 2001; *Hilairet et al.*, 2007] can lubricate faults and lower the lithosphere's yield strength; second, climate influences whether deep cracking of the lithosphere can occur, which is potentially important for hydrating the midlithosphere and lower lithosphere; and third, climate influences convective stresses. The first point, whether water can interact with rocks at the surface, provides only a weak influence of climate on tectonic regime, because hydrous phases can form over a wide range of surface temperatures. Geothermal heat flow can likely maintain a sub-ice ocean even at temperatures below the water freezing point [e.g., *Warren et al.*, 2002], and hydrous silicates are stable at temperatures up to $\approx 500-700°C$ [e.g., *Hacker et al.*, 2003], meaning these phases could even form under a steam atmosphere.

However, forming weak faults in the near-surface environment is not sufficient for plate tectonics; weakening of the midlithosphere and lower lithosphere, where strength is maximum, is also necessary. The deep lithosphere is initially dry due to dehydration during mid-ocean ridge melting [*Hirth and Kohlstedt*, 1996; *Evans et al.*, 2005], and is deeper than hydrothermal circulation can reach [*Gregory and Taylor*, 1981]. Thus, a mechanism capable of hydrating the midlithosphere is necessary for water weakening to be viable. One possible mechanism is the ingestion of water along deep thermal cracks [*Korenaga*, 2007]. Deep cracking is a result of thermal stresses that arise in the cooling lithosphere. The thermal stresses are proportional to the temperature difference between the cold lithosphere and the mantle interior, so climate can directly influence hydration of the midlithosphere. Higher surface temperatures could potentially lead to shallow cracks that leave the midlithosphere dry and strong. The surface temperature range where thermal cracking is effective has not been explored, but temperatures would likely have to be on the order of hundreds of degrees hotter than the present-day Earth to impede plate tectonics.
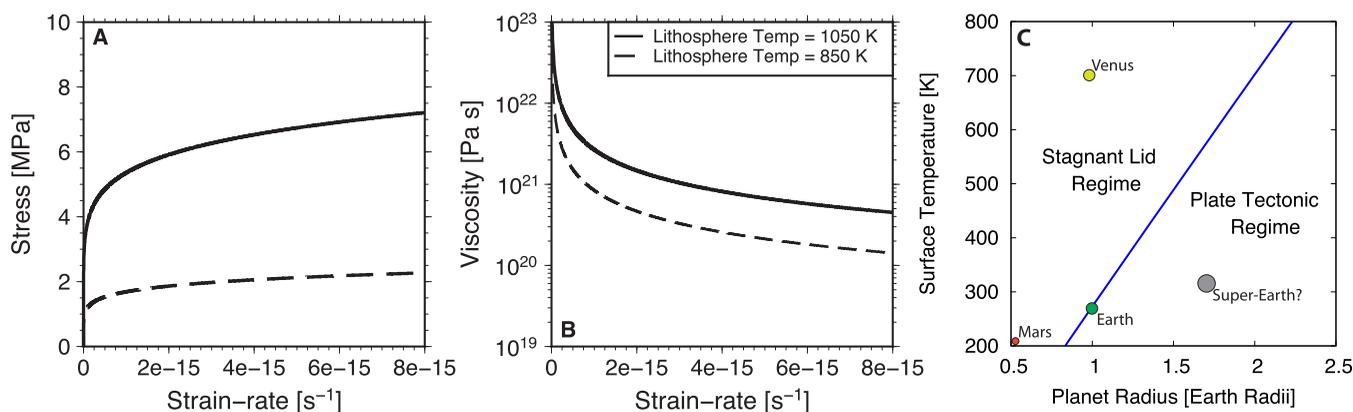
Figure 3. (a) Effective stress versus strain rate relationship and (b) viscosity versus strain rate relationship resulting from grain damage (equations (8)–(12) with $E_h = 500$ kJ mol^{-1}, $E_v = 300$ kJ mol^{-1}, $m = 2$, $p = 4$, and $f_A = 10^{-4}$). A lower temperature in the midlithosphere, which can result from a lower surface temperature, enhances viscous weakening, i.e., lower viscosities result from the same strain rate. As a result, the plate tectonic regime is found at low surface temperatures while the stagnant lid regime is found at high temperatures, as shown by (c) the regime diagram from *Foley et al.* [2012] (reprinted with permission). Note that the boundary between the plate-tectonic and stagnant lid regimes in Figure 3c is normalized to the Earth's present-day state. The exact position of the boundary between Earth and Venus could not be constrained by the simple models used in *Foley* et al. [2012].

An even stronger role for climate stems from the influence of surface temperature on convective stresses. High surface temperatures decrease stresses by lowering the negative thermal buoyancy of the lithosphere (e.g., ΔT decreases, resulting in lower τ_m from equations (4)–(6)). If the convective stress drops below the yield stress, then stagnant lid convection ensues. *Lenardic et al.* [2008] found that increasing Earth's present-day surface temperature by $\sim$100 K or more is sufficient to induce stagnant lid convection, even with a weak, hydrated lithosphere [see also *Weller et al.*, 2015].

When a grainsize reduction mechanism is considered, climate controls whether plate-like convection can occur through the influence of temperature on lithospheric grain-growth rates (Figure 3). Grain growth is faster at higher temperatures (see (8)), and higher grain-growth rates impede the formation of weak shear zones by acting to increase grainsize (see (10)). Thus, high surface temperatures, which lead to higher temperatures in the midlithosphere, promote stagnant lid convection [*Landuyt and Bercovici*, 2009; *Foley et al.*, 2012]. *Foley et al.* [2014] found that surface temperatures need to reach $\approx 500-600$ K for an Earth-like planet to enter a stagnant lid regime due to enhanced grain growth. Increasing surface temperature even further would eventually lead to a state where the lithosphere has a low enough viscosity to "subduct" even without any of the rheological weakening mechanisms discussed in this section (i.e., mantle convection would resemble constant viscosity convection). However, for this style of "subduction" to occur, the viscosity ratio between the surface and mantle interior must be less than $\approx 10^4$ [*Solomatov*, 1995], which requires a surface temperature of >1000 K with an Earth-like mantle temperature of ≈ 1600 K.

Summarizing the results from both mechanisms, cold surface temperatures are generally favorable for plate tectonics because they promote grainsize reduction and boost convective stresses. Moreover, cold surface temperatures are unlikely to prevent water weakening because geothermal heat flow can maintain liquid water beneath an ice covered ocean. However, high temperatures, in the range of 400–600 K, can cause stagnant lid convection by dropping convective stresses, increasing grain-growth rates, and possibly suppressing thermal cracking. The influence of climate on tectonic regime is also a leading a hypothesis for the lack of plate tectonics on Venus, as the Venusian surface temperature, 750 K, is easily hot enough to shut down plate tectonics based on the geodynamical models discussed in this section [*Lenardic et al.*, 2008; *Landuyt and Bercovici*, 2009].

2.3. Importance of Mantle Temperature and Other Factors

Naturally, climate is not the only important factor governing whether terrestrial planets will have plate tectonics. Mantle temperature, in particular, has been shown to have a major influence on a planet's tectonic regime. Mantle temperature modulates mantle convective stresses and lithospheric grain-growth rates, similar to the influence of surface temperature discussed in section 2.2, and mantle temperature controls the thickness of the oceanic crust formed at ridges, which affects the lithosphere's buoyancy with respect to the underlying mantle. All of these effects act to inhibit plate tectonics when interior temperatures are high, implying that planets with high internal heating rates or young planets, which are hotter because they have had less time

to lose their primordial heat [e.g., *Abbott et al.*, 1994; *Korenaga*, 2006; *Labrosse and Jaupart*, 2007; *Herzberg et al.*, 2010] (see also section 4.2), will be less likely to have plate tectonics. However, there is disagreement over just how important mantle temperature is in dictating a planet's mode of surface tectonics.

Increasing mantle temperature drops the mantle viscosity, μ_m. From equations (4)–(6), mantle stress scales as $\mu_m^{1/3}$. Thus, higher mantle temperatures lead to lower convective stresses, and, if stresses drop below the yield strength, stagnant lid convection in viscoplastic models [*O'Neill et al.*, 2007b; *Moore and Webb*, 2013; *Stein et al.*, 2013]; this effect may have even caused stagnant lid convection, possibly with episodic subduction events, on the early Earth [*O'Neill et al.*, 2007b; *Moore and Webb*, 2013]. The influence of mantle temperature on stress can also lead to hysteresis in viscoplastic mantle convection models [*Crowley and O'Connell*, 2012; *Weller and Lenardic*, 2012; *Weller et al.*, 2015]. Stagnant lid convection results in higher interior temperatures, and thus lower stresses, than mobile lid convection, due to low heat flow across the thick, immobile lithosphere. It is therefore harder to initiate plate tectonics starting from a stagnant lid initial condition than it is to sustain plate tectonics on a planet where it is already in operation. Such hysteresis would also mean that a planet that loses plate tectonics will be unlikely to restart it at a later time, even before the couplings with climate and the magnetic field are taken into account.

High mantle temperatures also play a role in plate generation via grainsize reduction, because they result in a warmer midlithosphere, and thus higher grain-growth rates. However, the influence of mantle temperature on lithospheric grain-growth rate is weaker than that of surface temperature, so elevated mantle temperatures alone are not capable of shutting down plate tectonics [*Foley et al.*, 2014], in contrast to viscoplasticity. Furthermore, the weak role of mantle temperature in plate generation means that the hysteresis loops seen in viscoplastic models may be less prominent with a grainsize reduction mechanism, though future study is needed to confirm this.

Mantle temperature also determines the thickness of oceanic crust generated at spreading centers, with hotter temperatures generally leading to a thicker crust [e.g., *White and McKenzie*, 1989]. As crust is less dense than the underlying mantle, cooling of the lithospheric mantle is necessary for the lithosphere as a whole to become convectively unstable [e.g., *Oxburgh and Parmentier*, 1977]. A very thick crust then requires a long cooling time, such that a thick lithospheric mantle root can form, in order to create a negatively buoyant lithospheric column. Forming a thick lithospheric mantle root is problematic, especially under high mantle temperature conditions, because the lithospheric mantle can delaminate before the lithosphere as a whole reaches convective instability, preventing subduction and plate tectonics [e.g., *Davies*, 1992]. The ability of crustal buoyancy to preclude plate tectonics on both the early Earth [*Davies*, 1992; *Vlaar*, 1985; *van Thienen et al.*, 2004b], and on exoplanets [*Kite et al.*, 2009], has been discussed by many authors. However, the transition from basalt to compositionally dense eclogite, which occurs at shallow depths of $\approx$ 50 km on Earth [e.g., *Hacker*, 1996], allows episodic subduction events to occur despite a thick, buoyant crust in models of the early Earth [*van Thienen et al.*, 2004a] and super-Earth exoplanets [*O'Rourke and Korenaga*, 2012]. Moreover, even stable, modern day plate tectonics can potentially still operate because the same melting process that creates the crust also dehydrates and stiffens the subcrustal mantle; this stiff lithospheric mantle then resists delamination and allows thick, negatively buoyant lithosphere to form through conductive cooling [*Korenaga*, 2006].

It is also important to point out that other factors, in particularly, composition and size, can potentially have a major impact on whether plate tectonics can take place on a rocky planet. However, both the influence of size and bulk composition on a planet's tectonic regime are not well understood. There is still significant disagreement over whether increasing planet size makes plate tectonics more or less likely, owing mainly to uncertainties in lithospheric rheology, the pressure dependence of mantle viscosity and thermal conductivity, and the expected radiogenic heating budget of exoplanets (see also section 5.2). Likewise, little is known about the key material properties for planetary mantles with a significantly different composition than Earth. Understanding how these factors influence a planet's propensity for plate tectonics and overall evolution is a vital area of future research.

3. Climate Regulation and the Long-Term Carbon Cycle

A primary factor determining whether a planet's climate will be conducive to plate tectonics is the degree of greenhouse warming. Venus has $\approx$ 90 bar of CO_2 in its atmosphere resulting in a surface temperature of
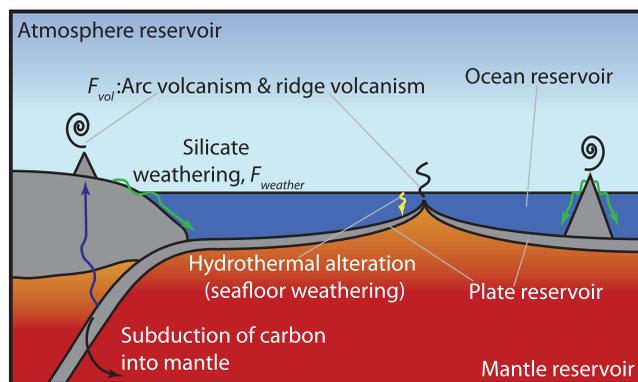
Figure 4. Schematic diagram of the global carbon cycle after *Foley* [2015].

≈ 750 K that is unfavorable for plate tectonics, while Earth only has ≈ 4× 10^{-4} bar of atmospheric CO_2, and thus has a temperate climate that allows for plate tectonics. However, on Earth, there is enough CO_2, anywhere from ≈ 60 to 200 bar [e.g., *Sleep and Zahnle*, 2001], locked in carbonate rocks at the surface and in the mantle to cause very hot surface temperatures of ≈ 400−500 K [*Kasting and Ackerman*, 1986]. Thus, a key factor for maintaining plate tectonics on a planet is the ability to prevent large quantities of CO_2 from building up in the atmosphere. On planets possessing liquid water, like Earth, this can be accomplished by silicate weathering at the surface and on the seafloor. Crucially, the weathering rate is sensitive to climate, with higher temperatures leading to larger weathering rates (and thus higher CO_2 drawdown rates), meaning that weathering acts as a negative feedback mechanism that works to maintain temperate climate conditions [*Walker et al.*, 1981; *Berner et al.*, 1983; *Tajika and Matsui*, 1990; *Brady*, 1991; *Tajika and Matsui*, 1992; *Berner and Caldeira*, 1997; *Berner*, 2004]. Planets that lack liquid water, like Venus, have no mechanism for regulating atmospheric CO_2. However, as we show in this section, simply possessing water does not guarantee that weathering can maintain climates within a range favorable for plate tectonics. A sufficient supply of fresh, weatherable rock at the surface is also needed. We further argue that plate tectonics enhances the supply of fresh rock to the surface, opening the possibility that plate tectonics and the long-term carbon cycle act as a self-sustaining feedback mechanism in some cases.

3.1. Modeling the Long-Term Carbon Cycle

Figure 4 gives a simple illustration of the global carbon cycle where four main reservoirs of carbon are considered (see *Berner* [2004] and *Ridgwell and Zeebe* [2005] for detailed reviews): the seafloor, the mantle, the atmosphere, and the ocean (the atmosphere and ocean reservoirs are often combined because equilibration between these two reservoirs is essentially instantaneous on geologic time scales [e.g., *Sleep and Zahnle*, 2001]). Carbon dioxide in the atmosphere is consumed by weathering reactions with silicate rocks, producing calcium, magnesium, and bicarbonate ions that flow from groundwater into rivers, eventually draining into the ocean (see *Gaillardet et al.* [1999] for a global compilation of CO_2 consumption via chemical weathering). Once in the ocean, these ions recombine to precipitate carbonates, leading to an overall net transfer of CO_2 from the atmosphere to carbonate rocks on the seafloor. Hydrothermal alteration of basalt can also act as a sink for CO_2 dissolved in the oceans, and given rapid equilibration between the atmosphere and ocean, a sink for atmospheric CO_2 as well [*Staudigel et al.*, 1989; *Alt and Teagle*, 1999; *Gillis and Coogan*, 2011]. Carbonates on the seafloor, both in the form of sediments and altered basalt, are subducted into the mantle at trenches. Here a fraction of the carbon devolatilizes and returns to the atmosphere through arc volcanoes, with the remaining carbon being recycled to the deep mantle. To close the cycle, mantle carbon is degassed through mid-ocean ridge and plume volcanism back to the atmosphere and ocean reservoirs.

The balance between weathering and volcanic outgassing dictates atmospheric CO_2 content. This balance is formulated as

$$F_{weather} + F_{sfw} = F_{arc} + F_{ridge}, \tag{13}$$

where $F_{weather}$ is the silicate weathering flux on land, F_{sfw} is the seafloor weathering flux, F_{arc} is the flux of CO_2 degassing from volcanic arcs, and F_{ridge} is the degassing flux at ridges. Balance between silicate weathering and degassing typically occurs rapidly, on a time scale of ∼1 Myr or less [e.g., *Sundquist*, 1991; *Berner and Caldeira*, 1997; *Driscoll and Bercovici*, 2013; *Foley*, 2015], so assuming that weathering always balances degassing is reasonable when studying long-term climate evolution. The arc degassing flux is typically written as

$$F_{arc} = \frac{f v_p L R_p}{A_p},$$ (14)

where f is the fraction of subducted carbon that degasses, v_p is the plate speed, L is the length of subduction zone trenches, R_p is the amount of carbon on the seafloor, and A_p is the area of the seafloor. The ridge degassing flux is given by

$$F_{ridge} = \frac{f_d v_p L d_{melt} R_{man}}{V_{man}},$$ (15)

where f_d is the fraction of upwelling mantle that degasses, L is the length of ridges (assumed equal to the length of trenches), d_{melt} is the depth where mid-ocean ridge melting begins, R_{man} is the amount of carbon in the mantle, and V_{man} is the volume of the mantle [e.g., *Tajika and Matsui*, 1990, 1992; *Sleep and Zahnle*, 2001; *Driscoll and Bercovici*, 2013; *Foley*, 2015].

Silicate weathering can exert a negative feedback on climate because mineral dissolution rates, and hence CO_2 drawdown rates, increase with increasing temperature and precipitation [e.g., *Walker et al.*, 1981; *Berner et al.*, 1983] (see *Kump et al.* [2000] and *Brantley and Olsen* [2014] for reviews on mineral dissolution kinetics and atmospheric CO_2). A similar link between climate and weathering rate can be seen in many field studies [e.g., *Velbel*, 1993; *White and Blum*, 1995; *Dessert et al.*, 2003; *Gislason et al.*, 2009; *White and Buss*, 2014; *Viers et al.*, 2014]. However, many other locations show no link between weathering and climate; instead, the weathering flux is linearly related to the physical erosion rate [*Stallard and Edmond*, 1983; *Edmond et al.*, 1995; *Gaillardet et al.*, 1999; *Oliva et al.*, 1999; *Millot et al.*, 2002; *Dupré et al.*, 2003; *Riebe et al.*, 2004; *West et al.*, 2005; *Viers et al.*, 2014]. These conflicting field observations are likely due to the distinction between "kinetically limited" (or "reaction limited") weathering and "supply limited" weathering. When weathering is kinetically limited, the kinetics of mineral dissolution controls the weathering rate. However, when weathering is supply limited, the supply of fresh rock brought to the weathering zone by erosion controls the weathering rate.

Supply limited weathering occurs when the reaction between silicate minerals and CO_2 runs to completion in the regolith, requiring physical erosion to expose fresh bedrock for weathering to continue. The supply-limited weathering flux (F_{w_s}) can be written as [*Riebe et al.*, 2004; *West et al.*, 2005; *Mills et al.*, 2011; *Foley*, 2015]

$$F_{w_s} = \frac{A_{land} E \chi \rho_{cc}}{\bar{m}},$$ (16)

where A_{land} is the surface area of all exposed land, E is the physical erosion rate, χ is the fraction of reactable elements in the crust, ρ_{cc} is the density of the crust, and $\bar{m}$ is the molar mass of reactable elements. The kinetically limited weathering flux, F_{w_k}, is sensitive to climate and can be written as [e.g., *Walker et al.*, 1981; *Berner et al.*, 1983; *Tajika and Matsui*, 1990, 1992; *Berner*, 1994; *Sleep and Zahnle*, 2001; *Berner*, 2004; *Driscoll and Bercovici*, 2013]

$$F_{w_k} = F_w^* \exp\left[\frac{E_a}{R_g}\left(\frac{1}{T^*} - \frac{1}{T}\right)\right]\left(\frac{P_{CO_2}}{P_{CO_2}^*}\right)^\beta \left(\frac{R}{R^*}\right)^\alpha \left(\frac{f_{land}}{f_{land}^*}\right),$$ (17)

where F_w^* is the present-day rate of atmospheric CO_2 drawdown by silicate weathering ($F_w^* \approx 6 \times 10^{12}$ mol yr^{-1}, or half the estimate of *Gaillardet et al.* [1999] because half of the CO_2 removed by weathering is rereleased to the atmosphere when carbonates form [*Berner*, 2004]), E_a is the activation energy for mineral dissolution, P_{CO_2} is the partial pressure of atmospheric CO_2, R is the runoff, f_{land} is the land fraction (subaerial land area divided by Earth's surface area), and β and α are constants. Stars represent present-day values.

The exponential term in (17) describes the temperature sensitivity of mineral dissolution rates, where typical values of E_a range from ≈ 40 to 50 kJ mol^{-1} [e.g., *Brady*, 1991]. The P_{CO_2} term describes the direct dependence of silicate weathering rates on atmospheric CO_2 concentration, which results from the role of pH in mineral dissolution. Specifically, more acidic pH values cause faster dissolution rates. However, when dissolution is caused by organic acids produced by biology (mainly plants on the modern Earth), soil pH is fixed, and the direct dependence of weathering rate on P_{CO_2} is weak (i.e., β in (17) is approximately 0.1 or less [*Volk*, 1987; *Berner*, 1994; *Sleep and Zahnle*, 2001]). Without biologically produced acids, carbonic acid

formed when atmospheric CO_2 dissolves in rainwater sets the soil pH, making the direct dependence of weathering rate on P_{CO_2} stronger ($\beta \approx 0.5$ [*Berner*, 1992]). The runoff term adds an additional climate feedback, as precipitation rates increase with temperature ($\alpha \approx 0.6-0.8$ [*Berner*, 1994; *Bluth and Kump*, 1994; *West et al.*, 2005]; see *Berner* [1994] for a detailed discussion of how silicate weathering scales with runoff). Finally, the weathering rate scales with land fraction, because decreasing the total area of land undergoing weathering lowers the total weathering flux.

The supply limited and kinetically limited weathering fluxes can be combined into a total weathering flux following *Gabet and Mudd* [2009], *Hilley et al.* [2010], and *West* [2012] (see *Foley* [2015] for a full derivation) as

$$F_{weather} = F_{w_s} \left[1 - \exp\left(-\frac{F_{w_k}}{F_{w_s}} \right) \right]. \tag{18}$$

When F_{w_k} is below the supply limit, F_{w_s}, the overall weathering rate is approximately equal to the kinetically limited weathering rate, i.e., $F_{weather} \approx F_{w_k}$. However, when F_{w_k} reaches or exceeds the supply limit, the overall weathering rate is fixed at F_{w_s}; i.e., $F_{weather}$ can no longer increase with increasing surface temperature or atmospheric CO_2 content once the supply limit to weathering has been reached. The present-day Earth consists of both regions undergoing supply limited weathering and those undergoing kinetically limited weathering. However, kinetically limited weathering appears to be the dominant component of the global silicate weathering flux [*West et al.*, 2005].

Finally, CO_2 drawdown also occurs via seafloor weathering on mid-ocean ridge flanks [*Alt and Teagle*, 1999]. Seafloor weathering is thought to be a smaller present-day CO_2 sink than weathering on land, and whether it exerts a strong feedback on climate is debated [*Caldeira*, 1995; *Brady and Gíslason*, 1997; *Berner*, 2004; *Coogan and Gillis*, 2013; *Coogan and Dosso*, 2015]. Many studies have assumed a weak dependence on P_{CO_2}, based on the experiments of *Brady and Gíslason* [1997], and formulate the seafloor weathering flux as [e.g., *Sleep and Zahnle*, 2001; *Mills et al.*, 2014; *Foley*, 2015]

$$F_{sfw} = F_{sfw}^* \left(\frac{v_p}{v_p^*} \right) \left(\frac{P_{CO_2}}{P_{CO_2}^*} \right)^{\xi}, \tag{19}$$

where the present-day seafloor weathering flux $F_{sfw}^* \approx 1.75\times10^{12}$ mol yr^{-1} [*Mills et al.*, 2014] and $\xi \approx 0.25$. However, both a stronger P_{CO_2} dependence and/or a direct temperature feedback are possible. Seafloor weathering can also become "supply limited," because only a finite amount of CO_2 can be stored in the ocean crust. *Sleep et al.* [2001, 2014] estimate that only the top ≈ 500 m of ocean crust can be completely reacted with CO_2, meaning no more than $\approx 4\times10^{21}$ mol (or ≈ 30 bar), $\approx 1/10–1/2$ Earth's total CO_2 budget, can be locked in the seafloor at any one time.

3.2. The Silicate Weathering Climate Feedback and Plate Tectonics

Kinetically limited weathering is able to prevent massive quantities of CO_2 from building up in the atmosphere, because the weathering rate increases with increasing atmospheric CO_2 content. Considering a balance between a given degassing flux, $F_{vol} = F_{arc} + F_{ridge}$, and weathering, the atmospheric CO_2 content as a function of degassing rate can be calculated (Figure 5). When the degassing rate is increased, only a small increase in atmospheric CO_2 concentration is needed to bring weathering back into balance with degassing, because the weathering rate is a strong function of P_{CO_2}. The stronger the dependence of weathering on atmospheric CO_2, the less sensitive P_{CO_2} is to variations in the degassing rate. It is this ability to balance the volcanic outgassing flux with minimal changes in P_{CO_2} that allows weathering to prevent extremely hot climates, and thus maintain conditions favorable for plate tectonics.

The negative feedback between climate and weathering is also important for habitability, because it acts to stabilize climate in response to variations in solar luminosity. Stars increase in luminosity as they age; for example, the sun's luminosity was 70% its present-day level during the Archean [*Gough*, 1981]. The weathering feedback acts to partially counteract this change in luminosity by allowing higher CO_2 levels when luminosity is low and lower CO_2 levels when luminosity is high. However, this topic has been extensively studied [e.g., *Walker et al.*, 1981; *Franck et al.*, 1999; *Sleep and Zahnle*, 2001; *Abbot et al.*, 2012], and going into more detail is beyond our scope.
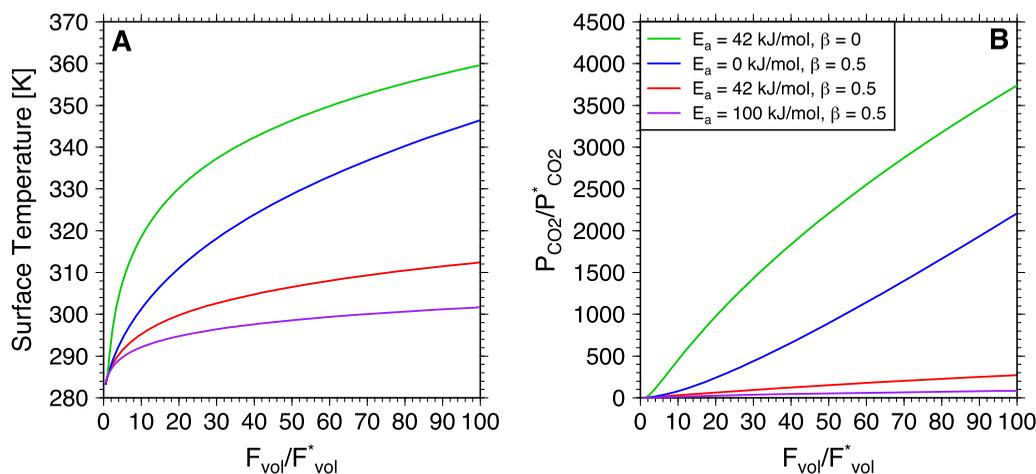
Figure 5. Response of (a) surface temperature and (b) partial pressure of atmospheric CO_2, normalized by the present-day value, to increases in the CO_2 degassing rate (where F_{vol}/F_{vol}^* is the volcanic outgassing rate normalized by the present-day value). Different values of E_a and β are used, as described in the legend, to show how stronger climate feedbacks cause climate to be less sensitive to degassing rate. Typical values for the Earth are $E_a = 42$ kJ mol^{-1} and $\beta = 0.5$ (red line).

On the other hand, when weathering becomes supply limited, it can no longer increase with atmospheric CO_2 level and is therefore unable to balance the degassing flux (globally supply limited weathering requires that $F_{vol} \geq F_{w_s}$, otherwise weathering would not be supply limited). As a result, atmospheric CO_2 accumulation from volcanic outgassing continues unabated until the mantle and plate reservoirs are depleted in carbon, and extremely hot climates, that are unfavorable for plate tectonics, prevail (Figure 6). Small land areas or low erosion rates lead to supply limited weathering, because both factors limit the supply of fresh rock to the surface, and hence lower F_{w_s} [see *Foley*, 2015, for details]. Alternatively, factors that increase the degassing rate can also drive a planet into the supply limited weathering regime, even with a large land area or high erosion rates. Planets with large total CO_2 inventories are more susceptible to supply limited weathering because degassing rates are higher. Another important factor is the fraction of subducted carbon that reaches the deep mantle, instead of devolatilizing and returning to the atmosphere at arcs (*f* from equation (14)). When more carbon can be subducted and stored in the mantle, F_{arc} is lower and kinetically limited weathering is easier to maintain (Figure 6d). The fraction of subducted carbon that degasses at arcs is not well constrained, and the physical and chemical processes controlling this number are poorly understood [e.g., *Kerrick and Connolly*, 2001; *Dasgupta and Hirschmann*, 2010; *Ague and Nicolescu*, 2014; *Kelemen and Manning*, 2015]. Moreover, *f* is likely a function of mantle temperature, as more slab CO_2 will devolatilize during subduction into a hotter mantle [e.g., *Dasgupta and Hirschmann*, 2010]. Thus, planets with hot mantles, as expected for young planets or those with high radiogenic heating budgets (see section 4.2), may be more susceptible to supply limited weathering. Better constraints on carbon subduction and devolatilization are clearly needed for understanding global climate feedbacks.

Figure 6 assumes that seafloor weathering follows (19), as in *Foley* [2015], which provides only a modest climate feedback. However, including a stronger climate feedback does not prevent extremely hot climates from forming when weathering on land is supply limited, because seafloor weathering will also become supply limited when all of the accessible basalt is completely altered [*Sleep et al.*, 2001, 2014]. With the ocean crust unable to take in any more carbon, a CO_2-rich atmosphere still forms [*Foley*, 2015]. Furthermore, the high surface temperatures predicted for the supply limited weathering regime could potentially lead to rapid water loss to space or even trigger a runaway greenhouse. Such scenarios are not explicitly modeled here, but would reinforce the fact that supply limited weathering causes a climate state that is unfavorable for plate tectonics; losing water would prevent volatile weakening of the lithosphere, and a runaway greenhouse climate results in temperatures even higher than those shown in Figure 6.

3.3. Extent of Coupling Between Climate and Surface Tectonics

Plate tectonics helps to sustain kinetically limited weathering, and therefore prevent extremely hot climates from forming, by acting to maintain high erosion rates and large subaerial land areas through orogeny and volcanism. Erosion rates are highest in rapidly uplifting, tectonically active areas [e.g., *Portenga and Bierman*, 2011],
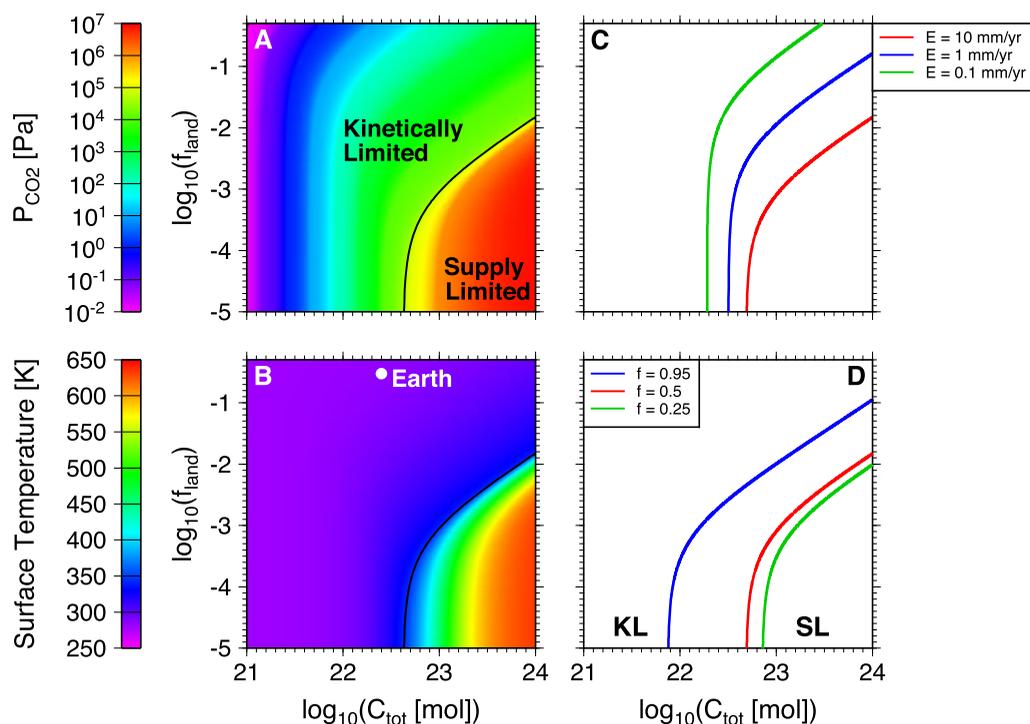
Figure 6. (a) Partial pressure of atmospheric CO_2 and (b) surface temperature as a function of total planetary CO_2 budget and land fraction (f_{land}) after *Foley* [2015]. Kinetically limited weathering occurs at large land fractions and low planetary CO_2 budgets, resulting in temperate surface temperatures, while supply limited weathering results in extremely hot climates when land fraction is low or total CO_2 budget is high. The estimated location of the Earth in land fraction-planetary CO_2 budget space is shown in Figure 6b. The boundary between the kinetically limited and supply limited weathering regimes shown (c) as a function of erosion rate, *E*, and (d) as a function of the fraction of subducted carbon that degasses at arcs, *f* (where the label KL refers to the kinetically limited regime and SL refers to the supply limited regime).

and orogenic processes are the primary cause of such uplift. Without tectonic uplift, erosion rates would decay to very low values because the surface can only erode as quickly as uplift creates topography. Orogeny results from plate tectonic processes, such as continent-continent collisions, island arc formation, and accretion of arcs to cratons, so plate tectonics is essential for keeping erosion rates on Earth high. Tectonic uplift and physical erosion have long been thought to be important for weathering and climate [e.g., *Raymo and Ruddiman*, 1992; *Maher and Chamberlain*, 2014], and the transition between kinetically limited and supply limited weathering provides an exciting new mechanism for explaining this influence [e.g., *Kump and Arthur*, 1997; *Froelich and Misra*, 2014].

Volcanism also creates topography and supplies fresh rock to the surface, so the fact that plate tectonics causes widespread subaerial volcanism (e.g., through arc, plume, and other hotspot melting), is important for sustaining high erosion rates as well. Perhaps the more important role for volcanism, however, is in creating exposed land, as large land areas also increase the supply of weatherable rock. In particular, continental crust, which makes up the large majority of Earth's exposed land, is thought to be predominantly created by subduction zone volcanism [e.g., *Rudnick*, 1995; *Cawood et al.*, 2013], a process that is unique to plate tectonics. Though hotspot or plume volcanism, the dominant mode of volcanism for a stagnant lid planet, can still create land and continental crust [e.g., *Stein and Goldstein*, 1996; *Smithies et al.*, 2005], plate tectonics clearly enhances subaerial land.

In fact, on a planet with a large area of exposed land, like Earth, the high erosion rates provided by plate tectonics may not be necessary for maintaining kinetically limited weathering. On Earth, typical erosion rates for flat-lying, tectonically inactive regions are on the order of 0.01 mm yr^{-1} [*Portenga and Bierman*, 2011; *Willenbring et al.*, 2013], which gives a supply limited weathering flux on the order of 10^{13} mol yr^{-1} when combined with Earth's large land area of $\approx 1.5 \times 10^{14}$ m^2. However, as half of the CO_2 removed by silicate weathering is rereleased to the atmosphere when carbonates form [e.g., *Berner et al.*, 1983], the net rate of CO_2 drawdown is $\sim 5 \times 10^{12}$ mol yr^{-1}. The present-day degassing flux is $\approx (6-10) \times 10^{12}$ mol yr^{-1}

[e.g., *Marty and Tolstikhin*, 1998; *Sleep and Zahnle*, 2001; *Burton et al.*, 2013], which would be significantly lower if Earth was in a stagnant lid regime. *Marty and Tolstikhin* [1998] estimate $\approx 3\times10^{12}$ mol yr^{-1} of CO_2 degassing from plumes, which is a reasonable approximation for Earth's degassing flux if it was in a stagnant lid regime. As the stagnant lid degassing flux is lower than the supply limited weathering flux, kinetically limited weathering is possible on a hypothetically stagnant lid Earth.

Therefore, plate tectonics and the long-term carbon cycle probably only act as a self-sustaining feedback on planets with small land areas or large planetary CO_2 inventories, where high erosion rates are needed to prevent supply limited weathering (i.e., planets that would lie near the boundary between kinetically limited and supply limited weathering in Figure 6). Unfortunately, a more quantitative estimate is not possible with our current level of understanding on how tectonics modulates erosion and weathering. Nevertheless, given that volatile acquisition during terrestrial planet formation can vary significantly [e.g., *Raymond et al.*, 2004], volatile-rich planets where plate tectonics and the carbon cycle are so tightly coupled may be common. Furthermore, even if a planet's exposed land area is large enough to sustain kinetically limited weathering at low erosion rates, plate tectonics may still be important. Plate tectonics may be responsible for creating most of the exposed land through continental crust formation, without which much higher erosion rates would be needed for kinetically limited weathering. Earth may even be an example of such a planet.

Another aspect of the coupling between plate tectonics and climate is that plate tectonics leads to long-lived CO_2 degassing by recycling carbon into the mantle at subduction zones. This process is important for habitability, because when CO_2 degassing rates are low, or cease entirely, snowball climates can form [*Kadoya and Tajika*, 2014]. However, snowball climates are unlikely to also shut down plate tectonics, so the role of plate tectonics in sustaining CO_2 degassing probably does not represent a self-sustaining feedback between tectonics and climate.

The different evolutionary scenarios that result from coupling between climate and surface tectonics are described in section 5. In particular, planets at different orbital distances can undergo different evolutions if one planet lies inward of the habitable zone's inner edge and thus lacks silicate weathering. Likewise different initial climate conditions can cause divergent evolutions if an initially hot climate prevents plate tectonics and kinetically limited weathering, or even weathering at all, from transpiring.

4. Generation of the Magnetic Field and Its Role in Atmospheric Escape

We have described the interactions between the surface environment (atmosphere plus ocean) and the mantle in sections 2 and 3. In this section, we demonstrate that additional interactions exists between the mantle and core, and the geomagnetic field (which is generated by the core dynamo) and atmosphere. The mantle controls the rate at which the core cools, thereby playing a crucial role in maintaining the energy flow necessary to drive convection and dynamo action in the core. The geomagnetic field provides a shield that holds the solar wind far above the surface (presently at about nine Earth-radii) so that most high energy particles are diverted and prevented from disrupting the near-surface environment. As a result, magnetic fields may limit the atmospheric escape rate under certain conditions. The magnetic field's influence on escape rate can then have an important control on long-term climate evolution, opening the possibility for an indirect influence of the core dynamo on mantle convection, in addition to the direct role mantle convection plays in driving the dynamo.

4.1. Core Dynamo
Earth has maintained an internally generated planetary magnetic field through convective dynamo action in its core over much of its history [*Tarduno et al.*, 2015]. The presence of a long-lived magnetic field is another unique feature of our planet, as only Mercury and Ganymede also have active magnetic fields today among the terrestrial planets and moons of the solar system [*Schubert and Soderlund*, 2011]. Internally generated magnetic fields are created by convective dynamo action in a large rotating volume of electrically conductive liquid. In the terrestrial planets, this typically occurs in an iron core. The most commonly cited processes for driving dynamo action, and thus the focus of this section, are thermal or compositional convection. However, other driving mechanisms such as gravitational tides [e.g., *Cébron and Hollerbach*, 2014] or precipitation of Mg initially dissolved in the liquid iron [*O'Rourke and Stevenson*, 2016] are possible.

Thermal convection occurs when the core heat flow, Q_{cmb}, exceeds the heat conducted adiabatically through the core. Recent estimates place the core's conductive heat flow at 13–15 TW [*Driscoll and Bercovici*, 2014;

Pozzo et al., 2014]. A high core heat flow is also inferred from revisions to the dynamics of mantle plumes, which are thought to be generated at the core-mantle boundary (CMB) and be indicative of the CMB heat flow [*Bunge*, 2005; *Zhong*, 2006]. Even if the total core heat flow is less than that needed to keep the core adiabatically well mixed, compositional buoyancy may still be able to drive convection. Compositional convection, which can be driven a number of ways, usually involves a phase change in the liquid where a density gradient develops or by the dissolution of an incompatible element due to a change in the solubility. For example, as Earth's core cools the inner core crystallizes from below, releasing buoyant light element-rich fluid into the surrounding iron-rich fluid and driving compositional convection. In this example, core heat flow is still important for driving compositional convection, because core cooling is necessary for inner core growth.

The heat transferred across the core-mantle boundary, Q_{cmb}, is controlled by the temperature gradient in the viscous mantle above:

$$Q_{cmb} = A_{cmb} k_{LM} \frac{\Delta T_{LM}}{\delta_{LM}}, \quad (20)$$

where A_{cmb} is CMB area, and k_{LM}, ΔT_{LM}, and δ_{LM} are the thermal conductivity, temperature drop, and thickness of the lower mantle thermal boundary layer. The temperature drop, $\Delta T_{LM} = T_{cmb} - T_m$, measures the thermal disequilibrium between the mantle and core. If the mantle cools efficiently then ΔT_{LM} is large, resulting in a high Q_{cmb}. The thermal boundary layer thickness, δ_{LM}, can be derived by assuming the boundary layer Rayleigh number (i.e., $\rho g \alpha \Delta T_{LM} \delta_{LM}^3 / (k \mu_{LM})$, where μ_{LM} is the effective viscosity of the lower mantle boundary layer) is at the critical value for convection. Compositional differences between the lower mantle and the bulk mantle [e.g., *Sleep*, 1988] can influence the thickness of δ_{LM}; this influence can be incorporated by modifying μ_{LM} [*Driscoll and Bercovici*, 2014] or the critical Rayleigh number [*Stevenson et al.*, 1983]. Calculating δ_{LM} from the boundary layer Rayleigh number implies that a hotter mantle produces a thinner boundary layer, increasing Q_{cmb}. Therefore, the ratio $\Delta T_{LM}/\delta_{LM}$ in (20) implies that the core and mantle will evolve toward thermal equilibrium faster in a hotter mantle and that efficient mantle cooling leads to efficient core cooling.

However, cooling the core too fast can be detrimental to dynamo action in two ways: (1) rapid cooling can lead to complete solidification of the core, preventing fluid motion and dynamo action, and (2) rapid cooling can bring the core into thermal equilibrium with the mantle, which will eventually decrease the cooling rate below the threshold for driving convection. The latter is the eventual fate of every rocky body, but the longer a planet can avoid thermal equilibrium and maintain moderate cooling, the longer it will generate a magnetic field.

### 4.2. Coupling Between Core Dynamo and Tectonic Mode

As the core cooling rate is determined by the mantle, the surface tectonic mode, which determines the mantle cooling rate, dramatically influences the dynamo. A mobile lid can accommodate a larger surface heat flow by exposing the mantle's top thermal boundary layer to the surface, while a stagnant lid inhibits cooling by insulating the mantle with a thick conductive layer.

To model the cooling and convective power available to drive dynamo action over time, the energy balance of the core and mantle are solved simultaneously. Volume averaged temperature evolution can be derived using the secular cooling equation $Q_i = -c_i M_i \dot{T}_i$, where c is specific heat and i refers to either mantle ($i = m$) or core ($i = c$) in the mantle and core energy balances [e.g., *Driscoll and Bercovici*, 2014]. Solving for $\dot{T}_m$ and $\dot{T}_c$ in terms of sources and sinks gives the mantle and core thermal evolution equations,

$$\dot{T}_m = \frac{(Q_{cmb} + Q_{rad} - Q_{conv} - Q_{melt})}{M_m c_m}, \quad (21)$$

$$\dot{T}_c = -\frac{(Q_{cmb} - Q_{rad,c})}{M_c c_c - A_{ic} \rho_{ic} \epsilon_c \frac{dR_{ic}}{dT_{cmb}} (L_{Fe} + E_G)}, \quad (22)$$

where Q_{conv} is heat conducted through the lithospheric thermal boundary layer by mantle convection (in W), Q_{melt} is heat loss by mantle melt eruption, and Q_{rad} and $Q_{rad,c}$ are radiogenic heat production in the
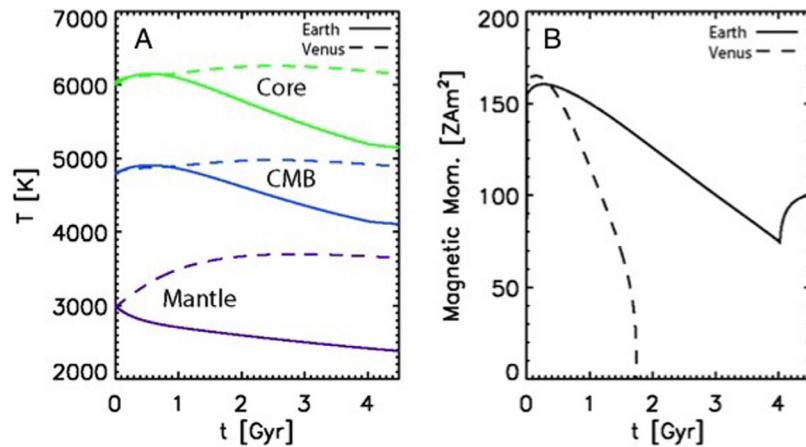
Figure 7. Comparison of the predicted (a) thermal and (b) magnetic histories of Earth (solid) and Venus (dashed). The only difference between the models is that the Earth model uses a mobile lid heat flow scaling while the Venus model uses a stagnant lid scaling (see (20)–(23); *Driscoll and Bercovici* [2014] contains additional details).

mantle and core, respectively. Crustal heat sources are excluded because they do not contribute to the mantle heat budget. The denominator of (22) is the sum of core specific heat and heat released by inner core growth, where A_{ic} is inner core surface area, ρ_{ic} is inner core density, ϵ_c is a constant that relates average core temperature to CMB temperature, dR_{ic}/dT_{cmb} is the rate of inner core growth as a function of CMB temperature, and L_{Fe} and E_G are the latent and gravitational energy released at the ICB per unit mass. Detailed expressions for heat flows and temperature profiles as functions of mantle and core properties are given in *Driscoll and Bercovici* [2014] and *Driscoll and Barnes* [2015].

Heat is lost from the mantle via conduction through the upper mantle thermal boundary layer (Q_{conv}) and by melt eruption (Q_{melt}). The rate at which this heat is lost is a complex function of mantle temperature, material properties, and style of convection. A mobile lid implies that the upper mantle thermal boundary layer reaches the planetary surface, so that the convective heat loss is controlled by conduction through this thin boundary layer. Assuming the boundary layer is at the critical Rayleigh number for convection, the convective heat flow is

$$Q_{conv} = a_m v_m^{-\beta} T_m^{\beta+1}, \qquad (23)$$

where $v_m = \mu_m/\rho$ is temperature-dependent mantle kinematic viscosity and $a_m = 8.4 \times 10^{10}$ W(m^{2}s^{-1}K^{-4})$^{1/3}$ and $\beta = 1/3$ are constants [see *Driscoll and Bercovici*, 2014, equation 43]. In steady state, a stagnant lid has a thicker conductive boundary layer, which can be modeled by decreasing a_m by a factor of $\sim$25 [*Solomatov*, 1995].

Mobile lid convection therefore favors dynamo action because it efficiently cools the mantle and thus boosts the core heat flow. Stagnant lid convection, on the other hand, can impede core cooling; stagnant lid convection on Venus is a leading hypothesis for why Venus lacks a magnetic field [*Nimmo*, 2002]. However, stagnant lid convection does not always prevent a core dynamo, as thermal history models commonly show a relatively short period of rapid cooling, regardless of the surface tectonic mode, during which core cooling rates are high and dynamo action is possible. This period of rapid cooling, or thermal adjustment period, results from initially hot interior temperatures that are a consequence of planetary accretion. The thermal adjustment period typically lasts for the first 1–2 Gyr as the internal temperatures and heat flows adjust to boundary conditions and heat sources. Similar early thermal dynamos have been proposed for the Moon [*Stegman et al.*, 2003] and Mars [*Nimmo and Stevenson*, 2000].

Figure 7 shows example thermal histories for Earth and Venus, where the Earth model uses a mobile lid heat flow scaling while the Venus model uses a stagnant lid scaling [see *Driscoll and Bercovici*, 2014]. The initial mid-mantle and CMB temperatures are assumed to be at the silicate liquidus, which is a reasonable starting temperature following the last large accretion event. Both models experience an initial thermal adjustment for 1–2 Gyr, but diverge soon after. After the initial adjustment, the Earth model cools

monotonically, maintaining a thermal dynamo and later a thermochemical dynamo after inner core nucleation around 4 Gyr, which can be seen as a jump in the predicted magnetic moment (Figure 7b). The core of the Venus model cools slightly during the thermal adjustment period, driving a transient thermal dynamo, but is slowly heated as the mantle and core heat up due to radiogenic heat trapped beneath the stagnant lid. In this case, the core is too hot to solidify, precluding a compositional dynamo. These models predict Venus maintained a magnetic field for about 1.5–4 Gyr [*Driscoll and Bercovici*, 2014]. During the thermal adjustment period, the core and mantle are still strongly coupled, but the cooling rate is less sensitive to the style of mantle convection. Thus, the primary role of plate tectonics is to extend the life of the dynamo beyond this thermal adjustment period.

In addition to normal convective cooling, volcanic heat loss can potentially extend core cooling beyond the thermal adjustment period on a stagnant lid planet, provided most of the melt can reach the surface of the planet and cool [*Morgan and Phillips*, 1983; *Nakagawa and Tackley*, 2012; *Armann and Tackley*, 2012; *Driscoll and Bercovici*, 2014]. Obviously, magmatic heat loss was not efficient enough for Venus, Mars, or the Moon to sustain dynamos, so we are left with the tentative conclusion that a stagnant lid reduces both the long-term convective and volcanic mantle heat loss.

The role of mantle dynamics in dictating magnetic field strength also indirectly links the core to climate. Climate influences the tectonic regime of a planet, and the tectonic regime dictates the core cooling rate through time and thus whether a long-lived magnetic field can be maintained. Therefore, a cool climate, being favorable for plate tectonics, will also be favorable for long-term magnetic field generation, and hot climates, to the extent that they lead to stagnant lid convection, will be unfavorable for the dynamo. An important question is then whether the core can exert any influence on climate through the role the magnetic field plays in limiting atmospheric escape. We address this topic next.

4.3. Magnetic Limited Escape

In describing the escape of a planetary atmosphere, it is helpful to think of the limiting escape mechanism rather than the specific physical escape process, as several escape processes typically occur concurrently with a single limiting bottleneck. The escape limit is typically characterized as being in either the diffusion or energy limited regime, with no connection to the presence of a magnetic field. However, it is commonly argued that magnetic field strength should play an important role in atmospheric escape [*Michel*, 1971; *Chassefiere*, 1997; *Lundin et al.*, 2007; *Dehant et al.*, 2007; *Lammer et al.*, 2012; *Owen and Adams*, 2014; *Brain et al.*, 2014]. Below we speculate about how a magnetic field could influence escape.

The hydrodynamic and diffusion limits to escape have been heavily studied [e.g., *Hunten*, 1973; *Watson et al.*, 1981; *Shizgal and Arkos*, 1996; *Lammer et al.*, 2008; *Luger and Barnes*, 2015]. Both of these escape limits depend on hydrogen number density. In the diffusion limit, escape is linearly proportional to hydrogen mixing ratio at the tropopause [*Hunten*, 1973], while hydrodynamic (or energy) limited escape has a weaker dependence on hydrogen mixing ratio but is expected to occur at much higher H concentrations [*Watson et al.*, 1981]. Hydrodynamic escape is fundamentally driven by incident energy absorbed at the base of the escaping region, but a bottleneck occurs as the escape rate grows and the absorbing species are lost, thereby limiting absorption and the energy available to drive escape. Regardless of their detailed dependences, diffusion and energy limited escape are expected to intersect at some H mixing ratio (therefore denoting a transition from one mechanism to the other) as depicted in Figure 8.

If a strong planetary magnetic field is present that balances the solar wind far from the top of the atmosphere, then the density of species exposed to the flow will be limited. In other words, the rate at which ionized planetary species are swept away by the stellar wind will decrease with increasing magnetopause distance and magnetic field strength. This effect adds a third limiting escape regime between the diffusion and hydrodynamic regimes. Figure 8 illustrates this scenario, where increasing hydrogen concentration leads to a transition from diffusion to magnetic limited, and then from magnetic to hydrodynamic limited escape. *Driscoll and Bercovici* [2013] propose that the limiting physical mechanism in the magnetic regime could be the removal of planetary ions from the magnetopause via Kelvin-Helmholtz instabilities, the occurrence of which are well documented [e.g., *Wolff et al.*, 1980; *Barabash et al.*, 2007; *Taylor and Lavraud*, 2008]. Mathematically, the magnetically limiting escape rate is predicted to be a function of hydrogen ion concentration, magnetopause surface area, and instability time scale, such that stronger magnetic fields expose a lower density of planetary species to the solar
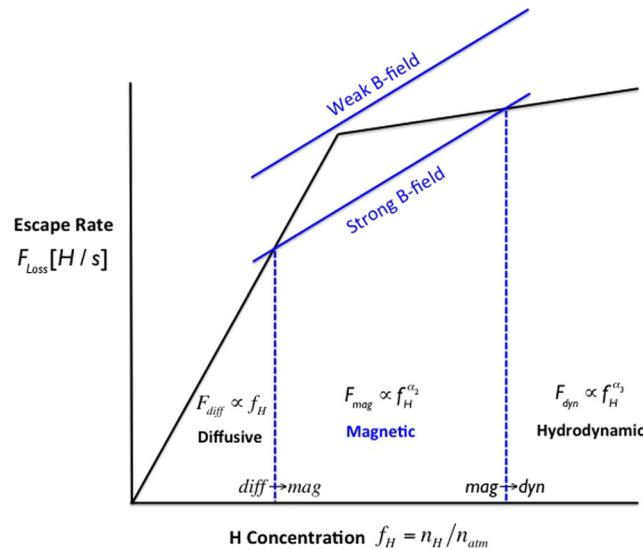
Figure 8. Schematic diagram of atmospheric limiting escape regimes. Hydrogen escape rate versus mixing ratio with the diffusive (black), magnetic (blue), and hydrodynamic (black) regimes. The transition from diffusive to hydrodynamic escape may be interrupted by magnetically limited escape if the planetary magnetic field is strong. See *Driscoll and Bercovici* [2013] for more details.

wind because their number density decreases faster than the magnetopause surface area [*Driscoll and Bercovici*, 2013]. Weaker (or nonexistent) planetary magnetic fields will allow the solar wind to interact with the "top" of the atmosphere (where escape occurs), potentially pushing the magnetic limit above the diffusion and hydrodynamic limits, rendering the magnetic limit irrelevant (Figure 8).

4.4. Extent of Coupling Between Mantle, Core, and Climate

The magnetic field is therefore most important for atmospheric evolution during the transition from hydrodynamic to diffusion limited escape. A strong magnetic field could reduce the escape rate during this transition, thereby helping to preserve a planet's water budget. Without magnetic shielding, significant water loss could occur, potentially preventing the silicate weathering feedback from acting to stabilize climate and, in turn, plate tectonics from operating.

Whether plate tectonics is itself necessary for magnetic shielding during the transition to diffusion limited escape depends on the timing of this transition (Figure 9). If the transition to the diffusion regime occurs quickly, i.e., f_H decreases rapidly, then it would occur during a planet's thermal adjustment period, when dynamo action is possible without plate tectonics (see section 4.2). On the contrary, if the magnetic limited escape regime is occupied longer than the mantle thermal adjustment period (e.g., f_H decreases slowly) then plate tectonics is probably important for preventing significant water loss through magnetic shielding. Unfortunately, the timing of the transition to diffusion limited escape is not well known, as f_H depends on the details of a planet's atmospheric structure. Future work in this area is needed to determine how long rocky planets occupy the magnetic limited escape regime, and the conditions under which plate tectonics is necessary for water retention.
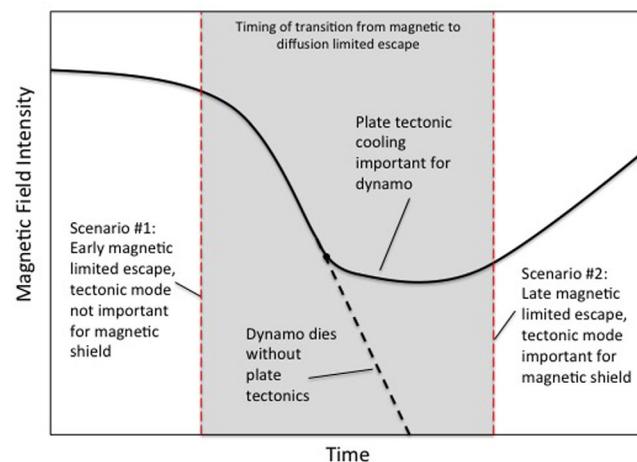
Figure 9. Schematic of magnetic field intensity over time. Initially, the dynamo is maintained by cooling during the thermal adjustment period (solid black) and tectonic mode is not important. This continues until core cooling begins to slow and the dynamo will either die without efficient mantle cooling (dashed black) or stay strong with efficient interior cooling associated with plate tectonics (solid black). If the timing of this event occurs after the transition from magnetic to diffusion limited escape (Scenario #1, vertical dashed red) then the shielding of the atmosphere is not dependent on the tectonic mode. If the dynamo divergence event occurs before the escape transition (Scenario #2, vertical dashed red) then the shielding of the atmosphere is coupled to the tectonic mode.

The timing of the transition to diffusion limited escape is also affected by stellar wind strength. In particular, magnetic shielding may be important over a wider range of stratospheric hydrogen mixing ratios, and thus over a longer period of time, for planets exposed to stellar winds much stronger than those at Earth today. Strong stellar winds are likely for planets orbiting active solar mass stars
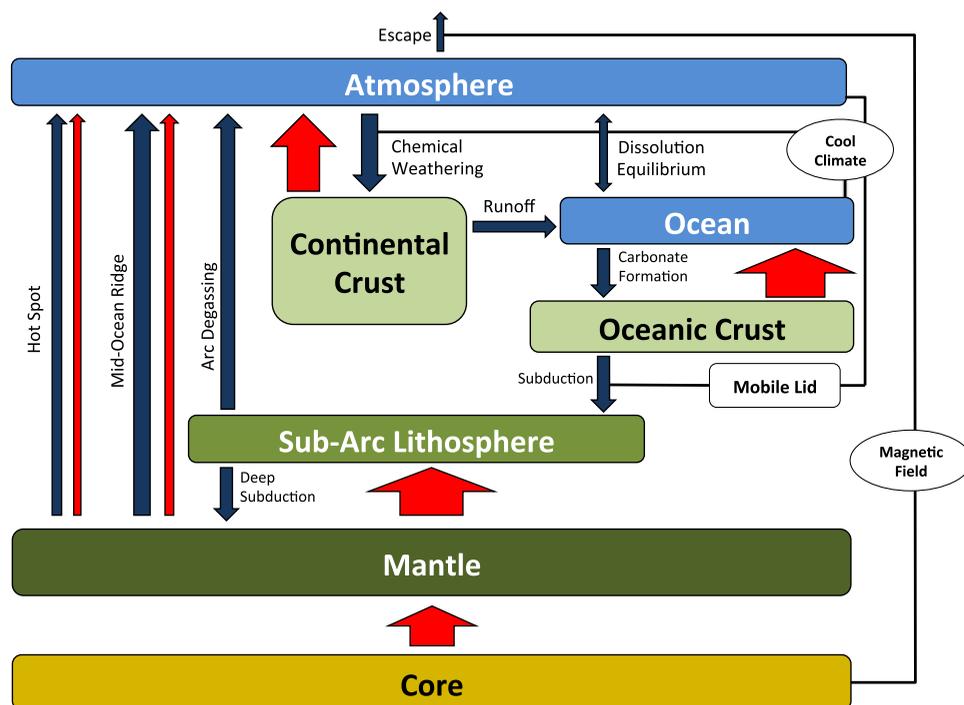
Figure 10. Flowchart of climate-tectonic-magnetic coupling for an Earth-like planet. Arrows between reservoirs indicate volatile (blue) and heat (red) fluxes, their width roughly in proportion to magnitude. Black lines indicate conceptual couplings, such as the influence of magnetic field strength on escape rate or chemical weathering on climate.

or orbiting close to moderately active small mass stars. In these cases, maintaining a magnetic field may be crucial for preserving liquid surface water over billion year time scales, and such a long-lived dynamo likely requires plate tectonics. In fact, the magnetic field, climate, and plate tectonics can act as a self-sustaining feedback in this case, where the magnetic field is required to prevent water loss, water is necessary for silicate weathering to keep the climate cool enough for plate tectonics, and plate tectonics is required to drive the dynamo.
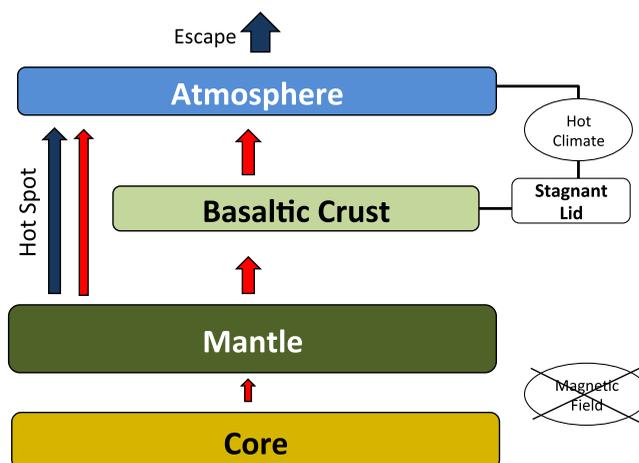
Our knowledge of how planetary magnetic fields influence atmospheric escape is still in its infancy with many unanswered questions. In fact, a strong magnetic field may even enhance escape in some instances, by producing a larger interaction cross section with the solar wind, which can concentrate incident energy flux by a factor of 10–100 [*Brain et al.*, 2014]. Clearly, a more thorough understanding of how the properties of the atmosphere, stellar wind, and magnetic field influence escape rates is needed to constrain the coupling between the core dynamo and climate.

5. Whole Planet Coupling and Planetary Evolution

Sections 2–4 together outline whole planet coupling between climate, plate tectonics, and the magnetic field. In this section, we describe how whole planet coupling can potentially explain the Earth-Venus dichotomy, and lead to a number different evolutionary scenarios for rocky exoplanets, many of which would be unfavorable for life. In particular, we focus on habitable zone planets, showing how events early in a planet's history, such as the initial atmospheric composition and the timing of the initiation of plate tectonics, play a major role in determining whether long-term habitable conditions can develop. A planet's volatile inventory is likely important as well. Given our incomplete understanding of plate tectonics, the carbon cycle, the geodynamo, and how they interact, our discussion is mostly qualitative, and only represents some initial hypotheses for the factors governing planetary evolution.

5.1. The Earth-Venus Dichotomy

Figures 10 and 11 illustrate how we propose that climate, mantle, and the core interact on Earth and Venus. Venus, being inward of the inner edge of the habitable zone, cannot have liquid water at its surface. As a

Figure 11. Flowchart of climate-tectonic-magnetic coupling on a Venus-like planet.

result, silicate weathering cannot draw CO_2 out of the atmosphere, so a hot, CO_2-rich climate forms via volcanic degassing and/or primordial degassing during accretion and magma ocean solidification. The hot climate prevents plate tectonics, and in turn the long-lived operation of a core dynamo due to a low core heat flux (Figure 11). For the Earth, being in the habitable zone allows liquid water, and in turn silicate weathering. Thanks to a temperate climate plate tectonics can operate, and the resulting high core heat flow powers the geodynamo (Figure 10). *Jellinek and Jackson* [2015] invoke a similar series of couplings to explain the Earth-Venus dichotomy, though they argue that the lack of plate tectonics on Venus is caused by high rates of radiogenic heating in the mantle, and that Venus' hot climate stems from the lack of plate tectonics. Their hypothesis contrasts with our interpretation, that Venus' orbital position is the key factor explaining its evolution.
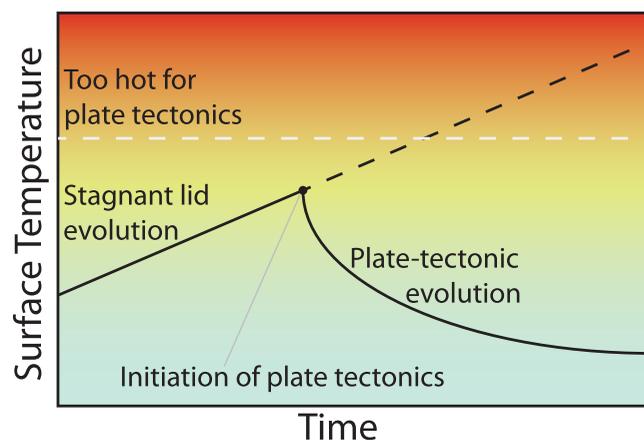
If surface-interior coupling is responsible for the divergent evolution of Earth and Venus, the runaway greenhouse climate, loss of plate tectonics (if it ever existed), and death of the magnetic field should all be correlated in time. Thus, determining the climate, magnetic, and tectonic history of Venus is a vital test. Unfortunately, our current knowledge of Venusian history is poor. The trace amount of water in the atmosphere requires a minimum of about 500 Myr for the H to escape to space [*Donahue*, 1999], so the runaway greenhouse must have occurred by at least $\approx 0.5-1$ Ga, but could have taken place much earlier. In fact, Venus may have entered a runaway greenhouse and lost its water during formation [*Hamano et al.*, 2013].

The Venusian tectonic and magnetic histories are similarly uncertain. There is evidence for a massive resurfacing event at 0.5–1 Ga. Surface features related to this event are most consistent with volcanism and lava flows rather than subduction of old crust and creation of new crust at ridges [*Smrekar et al.*, 2013; *Ivanov and Head*, 2013]. The planet may still be volcanically "active" today [*Smrekar et al.*, 2010], but eruptions are likely sporadic. The style of tectonics before the resurfacing event is unknown, with stagnant lid convection, episodic overturns, or even Earth-like plate tectonics all possibilities. Unraveling the Venusian magnetic history is challenging because the preservation and in situ measurement of any remanent magnetization on Venus' hot surface is unlikely. However, another possible line of evidence that may imply the presence of a paleo-Venusian magnetic field would be the loss of H^+, He^+, and O^+ along polar magnetic field lines to the solar wind, a process known on Earth as the polar wind [*Moore and Horwitz*, 2007]. This ion escape mechanism relies on the presence of an internally generated magnetic field, and could potentially leave a chemical fingerprint in the Venusian atmosphere [*Brain et al.*, 2014]. However, our present-day knowledge provides no evidence for a strong, internally generated magnetic field on Venus. Future exploration of Venus is needed to place tighter constraints on its evolution.

5.2. Evolution of Rocky Exoplanets

Venus demonstrates one likely evolutionary scenario for planets lying inward of the habitable zone's inner edge. However, cooler climates, that still allow plate tectonics and a magnetic field, are also possible for such planets if they have a significantly smaller CO_2 inventory than Venus. In this case, a temperate climate can still exist, even without liquid water and silicate weathering to regulate atmospheric CO_2 levels, simply because there is not enough CO_2 to cause extreme greenhouse warming. In some cases, a planet that experiences a runaway greenhouse could even retain some water at the poles, and potentially remain habitable [*Kodama et al.*, 2015]. Planets lying beyond the habitable zone's outer edge will likely be cold, and thus can plausibly sustain plate tectonics and a magnetic field as well. However, whether complex life can develop on any of these planets is unclear.

Figure 12. Schematic diagram of the divergence point in planetary evolution involving the initiation of plate tectonics. Before plate tectonics, a planet's weathering rate may be supply limited, such that surface temperature climbs over time as a result of increasing luminosity. Once plate tectonics initiates, silicate weathering is enhanced by higher erosion rates and continent formation, and climate cools.
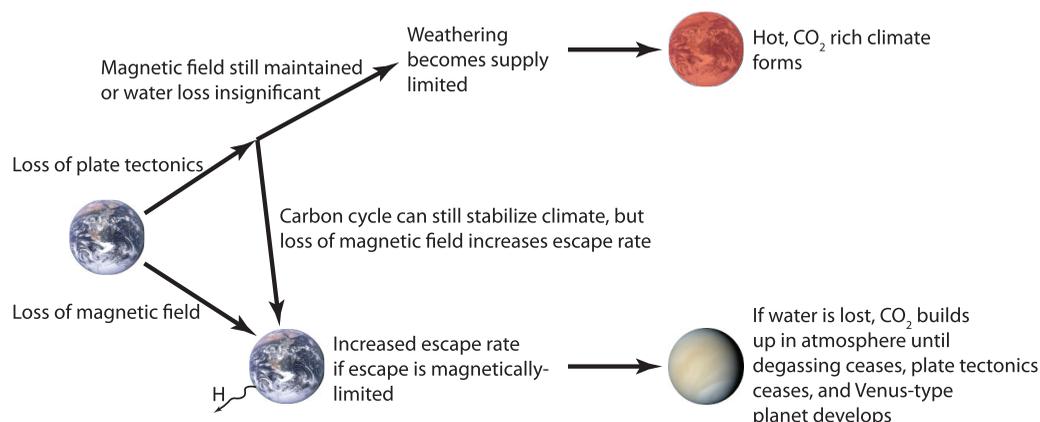
Likewise, an Earth-like evolution is one likely scenario for planets lying within the habitable zone, but a number of factors could lead to a different evolution that is unsuitable for life. Hot, CO_2-rich climates are expected after planet formation due to degassing during planetesimal accretion and magma ocean solidification [*Abe and Matsui*, 1985; *Zahnle et al.*, 2007; *Elkins-Tanton*, 2008]. If a planet's initial climate is so hot that liquid water is not stable (i.e., temperature and pressure conditions are beyond the critical point for water), then developing a temperate climate is probably not possible. Initial surface temperature and pressure conditions in excess of the critical point imply that silicate weathering would not occur, with or without plate tectonics (though some limited reaction between atmospheric CO_2 and the crust is possible). As a result, the climate would remain extremely hot, preventing both plate tectonics and a core dynamo. However, a climate hot enough to exceed the water critical point is an extreme case, requiring ≈ 500 bar of CO_2 for a planet with an atmosphere composed of CO_2 and H_2O and receiving the same insolation as the Hadean Earth [*Lebrun et al.*, 2013]. Earth's total planetary CO_2 budget is estimated at $\approx 100-200$ bar, so even with complete degassing during accretion or magma ocean solidification, liquid water was still stable [e.g., *Sleep and Zahnle*, 2001; *Zahnle et al.*, 2007]. Thus, a planet would need a significantly larger total CO_2 budget than Earth for initial atmospheric makeup to exceed the liquid water critical point.

Another possibility is a climate where liquid water is stable, but is still too hot for plate tectonics. In this case, low land fractions or erosion rates can potentially prevent plate tectonics, and a cool climate, from ever developing. High rates of volcanism would be needed to avoid this fate, by creating a sufficient supply of fresh rock at the surface for silicate weathering to cool the climate. Finally, even when climate conditions are amenable to plate tectonics, an uninhabitable state can be reached if plate tectonics does not initiate before increasing insolation warms the climate to the point where it is no longer possible (Figure 12). Before plate tectonics initiates on a planet, weathering may be supply limited and the atmosphere CO_2 rich. Thus, surface temperature will increase with increasing luminosity, and can potentially become hot enough to preclude plate tectonics from ever starting. If plate tectonics does not initiate before this divergence point is reached, a planet could become permanently stuck with a hot, uninhabitable climate, stagnant lid convection in the mantle, and no protective magnetic field.

Conversely, on a planet where plate tectonics does initiate before reaching this divergence point, continental growth and orogeny will increase both land area and erosion rates, enhancing the ability of silicate weathering to establish and maintain a temperate, habitable climate. If a large enough land area forms, weathering may even be capable of maintaining a temperate climate without plate tectonics to elevate erosion rates. Likewise, plate tectonics means that long-lived dynamo action, and therefore, volatile shielding from the solar wind is possible. The divergence point involving the initiation of plate tectonics could be particularly important if high mantle temperatures are a significant impediment to plate tectonics through low convective stresses or the creation of thick buoyant crust (see section 2.3). Mantle temperature may need to cool before plate tectonics can begin, even if surface temperatures are not hot enough to preclude plate motions. However, the increase in luminosity during a star's main sequence evolution is relatively gradual (the sun's luminosity was only $\approx 30\%$ lower at 4.5 Ga [*Gough*, 1981]), so there is a large time window, on the order of 1 Gyr, for sufficient mantle cooling to occur before climate becomes too hot for plate tectonics.

Another divergence point involves magnetic shielding and planetary water loss. The magnetic field can limit H escape, and thus help preserve surface water, when atmospheric escape transitions from being

Figure 13. Schematic diagram of possible scenarios where failure of the feedbacks between climate, plate tectonics, and the magnetic field leads to divergent evolution of terrestrial planets within the habitable zone.

hydrodynamically limited to diffusion limited (see sections 4.3 and 4.4.) If no magnetic field is available to shield the solar wind then massive amounts of H may be lost, leaving the planet desiccated and unable to regulate atmospheric CO_2 levels. As discussed in section 4.4, if the transition to diffusion limited escape is early in a planet's history, then magnetic shielding is possible regardless of tectonic regime, and many planets will likely be able to keep their volatiles and possibly develop habitable climates. However, if the transition occurs after the mantle's thermal adjustment period, then plate tectonics is likely necessary for magnetic shielding, and fewer planets will be able to retain surface water.

Additional divergence points are possible later in planetary evolution if climate, tectonics, and the magnetic field are tightly coupled. For example, if the carbon cycle and plate tectonics act as a self-sustaining feedback, where high erosion rates, supplied by plate tectonics, are required to maintain kinetically limited weathering and thus keep surface temperatures cool enough for plate tectonics (see section 3.3), then the loss of plate tectonics would lead directly to a hot climate state that likely precludes plate tectonics from reinitiating. Without plate tectonics to enhance erosion, the inhospitably hot climate that results from supply limited weathering would likely be permanent. Such tight coupling between plate tectonics and the carbon cycle is most likely on planets with small exposed land areas or high total CO_2 budgets. Another divergence point is possible for planets orbiting very active solar mass stars or very close to active small mass stars where stronger stellar winds can more efficiently strip atmospheric volatiles. For these planets, cessation of the core dynamo could cause significant water loss, in turn halting silicate weathering and, due to the ensuing hot climate, likely shutting down plate tectonics as well (see Figure 13 for a schematic illustration of the divergence points discussed in this paragraph).

Size is also a major factor in terrestrial planet evolution. Large planets are expected to have wider habitable zones due to the influence of higher gravity on atmospheric-scale height and the greenhouse effect [*Kopparapu et al.*, 2014], but also shallower ocean basins and thus less exposed land, unless feedbacks can regulate ocean volume such that continents are always exposed [*Cowan and Abbot*, 2014] (see also section 6). The influence of size on plate tectonics and magnetic field generation is also unclear. Previous studies have found that plate tectonics is more likely on larger planets [*Valencia et al.*, 2007; *Valencia and O'Connell*, 2009; *van Heck and Tackley*, 2011; *Foley et al.*, 2012], less likely on larger planets [*O'Neill and Lenardic*, 2007; *Kite et al.*, 2009; *Stein et al.*, 2013; *Noack and Breuer*, 2014; *Stamenković and Breuer*, 2014; *Miyagoshi et al.*, 2014; *Tachinami et al.*, 2014], or that size is relatively unimportant [*Korenaga*, 2010a]. Different studies reach very different conclusions because of the large uncertainties in the rheological mechanism necessary for generating plate tectonics, how key features such as internal heating rate scale with size, and how mantle properties are affected by pressure and temperature. The influence of size on magnetic field strength is likewise debatable. Several studies have found a weak dependence of field strength and lifetime on planet and core size, and possibly a peak in strength for Earth-sized planets [*Gaidos et al.*, 2010; *Tachinami et al.*, 2011; *Driscoll and Olson*, 2011; *Van Summeren et al.*, 2013]. Generally, larger dynamo regions are expected to produce stronger magnetic fields [e.g., *Christensen et al.*, 2009], but variations in more subtle properties, like mantle and core composition, likely play a fundamental role. The coupling between plate tectonics, climate,

and the magnetic field is expected to apply to planets of different size, but future work is needed to place tighter constraints on the influence of size on this coupling and on planetary evolution.

The issue of size highlights the many uncertainties remaining in our knowledge of planetary dynamics and evolution. Each aspect of planetary dynamics that is important for magnetic, tectonic, and climate evolution needs to be better understood before more rigorous predictions or interpretations can be made. In particular, the interactions between different components of the planetary system, specifically interactions among surface tectonics, mantle convection, and the long-term carbon cycle, between mantle convection and the core dynamo, and among the magnetic field, atmospheric escape, and climate, deserve significant attention. With a large number of rocky exoplanets already discovered, many of which are in their respective habitable zones [*Batalha*, 2014], and more certain to follow, improving our knowledge of planetary evolution is an important goal.

6. Future Directions

In addition to furthering our understanding of plate tectonics, magnetic field generation, climate evolution, and the interactions between these processes, there are a number of new questions and research topics that must be addressed to advance our knowledge of planetary evolution. We summarize a few of these important questions below.

1. *What are the material properties of Earth-like and non-Earth-like terrestrial planets at high-temperature, high-pressure conditions?* Without tighter constraints on basic properties like density, viscosity, and thermal conductivity models of exoplanet evolution will continue to be highly uncertain. Conducting both laboratory and numerical experiments at the conditions relevant for Earth and super-Earths is challenging, but is also vital for determining how terrestrial planets behave. Moreover, the composition of exoplanets could differ significantly from Earth, so constraints on the properties of non-Earth-like materials are necessary for modeling the evolution of these planets as well.

2. *What controls the volume of water at a planet's surface (i.e., the size of the oceans and the amount of subaerial land)?* Earth has maintained a relatively constant freeboard, or water level relative to the continents, throughout much of its history [*Wise*, 1974], despite the fact that there could be multiple oceans worth of water stored in the mantle [e.g., *Ohtani*, 2005]. Are there feedbacks in Earth's deep water cycle acting to keep a large continental surface area exposed, as proposed by some [*Kasting and Holm*, 1992; *Cowan and Abbot*, 2014]? Given the importance of exposed land to the long-term carbon cycle, this question is of fundamental importance to planetary evolution.

3. *What are typical volatile abundances, in particular water and carbon dioxide, for rocky planets, and how much of this volatile inventory resides in the atmosphere just after accretion and magma ocean solidification?* Both volatile inventory and initial atmospheric composition are important for planetary evolution. Thus, a better understanding of how and when volatiles are delivered during accretion, how much degassing occurs during accretion and solidification of a possible magma ocean, and how much atmospheric loss occurs during this early phase of planetary history are all necessary for constraining planetary evolution.

4. *Can strong gravitational tides render a planet uninhabitable?* Planets experiencing strong gravitational tides (caused by nearby stars, planets, or satellites) can generate significant internal heating via tidal dissipation, which can cause extreme surface volcanism and hinder dynamo action [*Driscoll and Barnes*, 2015]. Efficient cooling of the interior could allow the orbits of such planets to circularize faster, minimizing the length of time spent in a tidally heated regime, and could move the planet in (or out) of the habitable zone. Future work should explore the details of tidal dissipation in the mantle and core, and how tides can influence long-term evolution.

5. *How and when does plate tectonics initiate?* Factors other than climate, like mantle temperature, can have an important control over whether plate tectonics can operate. Thus, even with favorable climate conditions, other process such as mantle cooling may be necessary for Earth-like plate tectonics. Understanding the type of tectonics that might take place on a planet before plate tectonics, how much land and weatherable rock can be created through nonplate-tectonic volcanism, and the factors that then allow for Earth-like plate tectonics to develop will all be crucial for determining how likely planets are to follow an evolutionary trajectory similar to Earth's.

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