The manuscript *Internal Planetary Feedbacks, Mantle Dynamics, and Plate Tectonics* is a chapter in the book *Dynamics of Plate Tectonics and Mantle Convection* published by Elsevier.

Adrian Lenardic
ajns@rice.edu
Rice University

Johnny Seales
johnny.seales@gmail.com
Rice University.

Internal Planetary Feedbacks, Mantle Dynamics, and Plate Tectonics

Adrian Lenardic, Johnny Seales

Department of Earth Science, Rice University, Houston, TX 77251-1892. (ajns@rice.edu)

Abstract

Isolating planetary feedbacks, and feedback analysis, are prevalent aspects of climate and Earth surface process science. An under appreciation of internal planet feedbacks, and feedback analysis for plate tectonics research, motivate this chapter. We review feedbacks that influence the Earth's thermal evolution and expand them to include magmatic history and planetary water budgets. The predictions from feedback models are shown to be consistent with petrological constraints on the Earth's cooling. From there, we isolate feedbacks that connect structural elements within the mantle dynamics and plate tectonics system. The feedbacks allow for a reciprocal causality between plates, plumes, the asthenosphere, and mantle flow patterns, with each element being co-dependent on the others. The linked elements and feedbacks define plate tectonics are part of a self-sustaining flow system that can bootstrap itself into existence. Within that framework, plate tectonics involves the co-arising of critical system factors. No single factor is the cause of another. Rather, they emerge with the links between them and the generation of functional elements coincides, within relatively narrow time windows, with the co-emergence of factors that are critical for the maintenance of the elements themselves. What emerges is not a tectonic state but a process. That is, a set of feedbacks that can transform the tectonics of a planet and/or maintain plate tectonics. The feedback functions are not permanent but can operate over extended time frames such that plate tectonics can remain stable. The nature of the feedbacks, and their stability, can be studied at various levels of detail but questions of origin can become ill-defined. Observational tests of a feedback framework for plate tectonics and mantle dynamics are presented, along with research paths that apply feedback methodology to solid planet dynamics and comparative planetology.

Preprint submitted to TBD *March 18, 2022*

Keywords: Feedbacks, System Dynamics, Plate Tectonics

1. Introduction

The concept of feedback is prevalent within climate science, Earth surface science, and studies of the Earth's biosphere. Introductory textbooks on the subjects now routinely discuss feedback analysis. The same can not be said for solid-earth science. The relative under-appreciation of solid planet feedbacks motivates this chapter. Our pragmatic goal is to highlight several solid planet feedback processes. Our more meta-goal is to argue that feedback analysis is a useful tool for research into mantle dynamics and plate tectonics and can provide a means to cast new light on old questions.

Feedback occurs when a cause initiates a chain of events that ultimately leads to an effect on the initial cause itself. If the cause is enhanced, then the cycle is a positive feedback (also referred to as an amplifying feedback). If the initiating cause is damped, then its is a negative feedback (also referred to as a buffering and/or regulating feedback). Discussions of feedback appear throughout history but a modern appreciation started within engineering [Maxwell, 1868]. Feedback became a foundational concept for cybernetics and general systems theory [Weiner, 1948; von Bertalanffy, 1968]. Systems theory, systems science, and a systems approach expanded into a wide range of fields [Jantsch, 1980; Meadows 1982; 2009; Laszlo, 1996]. An historical overview of how systems theory expanded into, and intertwined with, the Earth sciences can be found in Steffan et al. [2020].

Systems can be defined in different ways. A general definition is that "A system is a set of interacting units or elements that form an integrated whole that performs some function [Skyttner, 1996]." Links and interactions lead to order, pattern, and structure that dynamically maintains itself and generates functions not inherent in any of the single elements, often expressed as the whole being greater than the sum of its parts and/or the idea that 'more is different' [Anderson, 1972]. A more precise definition comes from Ackoff [1981] who defines a system as a set of elements that satisfy the following conditions: 1) The behavior of each element has an effect on the behavior of the whole; 2) The behavior of the elements and their effects on the whole are interdependent; 3) However subgroups of the elements are formed, all have an effect on the behavior of the whole but none has an independent effect on it. All of the definitions implicate feedbacks as critical to distinguishing a system from an aggregate of elements. That feedbacks can generate collective system properties not inherent in system elements also connects feedback to studies of self-organization and emergent phenomena [Holland, 1998; Fromm, 2004].

Earth systems science, which rose to prominence in the 1980s, was founded on an appreciation of feedbacks in the natural world. An example is the silicate-weathering feedback, which provides a means for a planet to regulate atmospheric greenhouse gas concentrations [Walker et al., 1981]. That feedback is connected to the carbon cycle. Figure 1a shows a schematic of the cycle [Berner, 1983]. The schematic serves as a visual motivation for this chapter. The Earth's interior is conceptualized as a source and a sink for materials that enter into atmosphere, hydrosphere, and biosphere feedbacks. This example is not overly contrived and Earth systems literature often has the Earth's mantle conceptualized as an arrow that points to feedback loops outside of the mantle itself (Figure 1b). The potential of feedbacks within our planets interior has not been appreciated to the level that surface feedbacks have come to be appreciated.

The under appreciation of solid-planet feedbacks is perplexing given some historical serendipity. In 1974 a paper appeared that would set into motion an appreciation of feedbacks that could regulate the surface temperature of a planet [Lovelock and Margulis, 1974; Watson and Lovelock, 1983]. Two years earlier a paper appeared about a feedback that could regulate the internal temperature of a planet [Tozer, 1972]. The former is seen as foundational to an appreciation of feedbacks in the natural world and the rise of a systems approach to earth science [Steffan et al., 2020]. The later is less appreciated as a step toward the application of feedback thinking to natural systems. As such, it provides a good starting point for discussions of solid planet feedbacks. From there, we will spring board into discussions of additional feedbacks that can play roles in mantle dynamics and the operation of plate tectonics.

2. Thermal Cycles, Thermal-Hydro Cycles, and Internal Earth Cooling Feedbacks

The the cooling of the Earth is a problem with a rich history [Kelvin, 1863; England et al., 2007]. Plate tectonics set Earth cooling within a mobilist view. Plate tectonics is a surface expression of mantle convection and convective plate overturn is a principal means of interior cooling.

The first generation of thermal history models, based on convective cooling, came out in the late 1970s to early 1980s [Sharpe and Peltier, 1979; Schubert, 1979; 1980; Davies, 1980]. The models are based on a global energy balance that tracks the decay of internal mantle heat sources and surface heat loss. Heat loss is parameterized in terms of a relationship between convective heat flux (Nu) and a mantle Rayleigh number (Ra), a measure of convective vigor. That relationship is given by

$$Nu \sim Ra^{\beta} \tag{1}$$

where

$$Ra = \frac{\rho g \alpha \Delta T Z^3}{\kappa \eta} \tag{2}$$

and ρ is density, α is thermal expansivity, g is the acceleration due to gravity, ΔT is the driving temperature, Z is the thickness of the convecting layer, κ is the thermal diffusivity and η is the mantle viscosity. The scaling exponent, β, parameterizes the effects of physical factors on the efficiency of convective cooling. In particular, it depends on physical processes that resist convective motion and the overturn of tectonic plates. The first generation of thermal history models assumed that mantle viscosity is the dominant resistance to plate overturn. That leads to a β value near the high-end limit of 1/3 [Schubert et al., 1980]. Together with an exponential dependence of mantle viscosity on temperature, this allows for a feedback isolated by Tozer [1972]. Arguably, it was an appreciation of that feedback that set modern thermal history modeling into motion.

Figure 2a shows a causal loop diagram of the Tozer feedback. Arrows labelled as positive indicate that an increase/decrease of the factor at the base of the arrow leads to an increase/decrease of the factor at the head of the arrow. Arrows labelled as negative indicate that an increase/decrease of the factor at the base of the arrow leads to a decrease/increase of the factor at the head of the arrow. Multiplying positive and negative effects around an entire feedback loop leads to the overall sign of the feedback, which appears in parenthesis at the loop center. The Tozer feedback is a negative feedback: If heat loss becomes low/high relative to internal heat generation, then the mantle will heat/cool, viscosity will decrease/increase, and heat flux will increase/decrease (due to increased/decreased tectonic plate overturn associated with lower/higher viscous resistance). This regulates mantle temperature against large and/or long-lived fluctuations. For that reason it is also referred to as a thermostat feedback.

Although the Tozer feedback influenced thermal history modeling, wether it leads to mantle self-regulation has been called into question [Korenaga, 2016]. The ability of the feedback to regulate mantle evolution depends on two times scales (Figure 2a). One is associated with the decay rate of radiogenics in the mantle. That time scale, referred to as a secular time, is on the order of a billion years. The second is the time over which the feedback operates. That time scale relates to the reactance time of a system [Seely, 1964; Close et al., 2001]. Reactance time characterizes a systems response to perturbations.

Mantle reactance time depends on the relationship between heat loss and convective vigor. In thermal history models this relates to the assumed value of β in Equation 1. Figure 3a plots results from a reactance time analysis applied to a range of thermal history models [Seales, 2019]. Models with β near 1/3 have a relatively short reactance time that allows interior cooling to evolve along a self-regulated path [Davies, 1980]. Those models assume that viscosity is the dominant resistance to plate motions. If plate and/or plate margin strength also plays a significant role, then β has been argued to be closer to 0.15 [Conrad and Hager, 1999]. That still allows for self-regulation but fluctuations can be longer lived. If mantle viscosity plays no role in plate motions and plate strength remains constant, then β is zero and there is no feedback or self-regulation [Christensen, 1984; 1985]. If plate strength increases with convective vigor, then β can be negative [Korenaga, 2003; 2008]. A feedback exists but, unlike the Tozer feedback, it is a positive feedback (Figure 2b). Positive feedbacks do not regulate a system. Instead, they allow perturbations to be amplified and/or be very long lived; different initial conditions and/or fluctuations along a cooling path can influence a planets evolution over time scales longer than a secular time. The implications for planetary thermal histories are significant (Figure 3b).

If the Tozer feedback is operative, then the damping of thermal perturbations/fluctuations maintains the ratio of heat generation to heat loss, termed the Urey ratio (Ur), near unity [Schubert1980]. Stated another way, the mantle convection system has low thermal inertia such that any large deviations from thermal equilibrium are damped and interior cooling evolves along a series of quasi-equilibrium steps [Davies, 1980]. Such models cannot account for updated constraints on Earth's cooling history [Christensen, 1985; Korenaga, 2008]. In particular, data constraints place Ur between 0.2 and 0.5 [Jaupart et al., 2007], i.e., heat loss and heat generation are far from equilibrium. This has been used to argue that mantle convection is not self-regulated [Korenaga, 2016].

The argument that low Ur is not consistent with thermal self-regulation is robust but it does not rule out self-regulation altogether. The appreciation that mantle viscosity depends on temperature lead classic thermal history models to focus on thermal-regulation. What has gone under appreciated is that the critical assumption at their core is viscosity-regulation. That is, changes in viscosity dominate changes in the Earth's Rayleigh number and, over time scales shorter than secular decay times, viscosity, and by association the Rayleigh number, can be approximated as remaining constant (a quasi-equilibrium assumption). This is a critical assumption in using $Nu \sim Ra^{\beta}$ scaling relationships to begin with, as they are based on theory, experiments, and/or numerical simulations carried out under constant Ra values [Moore and Lenardic, 2015]. If viscosity depends only on temperature, then a lack of thermal-regulation rules out self-regulation. If that is not the case, then self-regulation remains viable. The dependence of mantle viscosity on water opens this possibility [Mackwell, 1985; Li et al., 2008].

Early thermal history models that considered the role of water predicted Ur values greater than one or comparable to classic models [Jackson and Pollack, 1987; McGovern and Schubert, 1989]. The former enforced a net loss of water from the Earth's interior. The latter assumed that Ur should be 0.8 and, as such, calibrated free parameters to keep mantle water content nearly constant. Crowley et al. [2011] showed that a larger range of behavior is possible if imbalances in mantle dewatering (D) and rewatering (R) are allowed for. Mantle dewatering occurs principally via melting at mid-ocean ridges. Mantle rewatering occurs at subduction zones, where descending slabs carry some of their bound water into the mantle. If mantle viscosity depends on temperature (T) and water content (χ), then the time rate of change of mantle viscosity can be written as

$$\frac{d\eta}{dt} = \frac{\partial \eta}{\partial T}\frac{dT}{dt} + \frac{\partial \eta}{\partial \chi}\frac{d\chi}{dt}. \tag{3}$$

Conservation of energy leads to

$$\frac{dT}{dt} = \frac{1}{\rho C_p V}(H - Q_s) \tag{4}$$

where C_p is specific heat, V is mantle volume, H is mantle heat production, and Q_s is surface heat flow. Conservation of mantle water content leads to

$$\frac{d\chi}{dt} = \frac{1}{\rho V}(R - D). \tag{5}$$

If viscosity remains statistically steady, relative to the time scale over which significant changes occur in internal heat generation, then the Urey ratio is given by

$$Ur \approx 1 - \frac{\eta_\chi}{\eta_T}\frac{C_p}{Q_s}(R - D), \tag{6}$$

where $\eta_\chi = \frac{\partial\eta}{\partial\chi}$ and $\eta_T = \frac{\partial\eta}{\partial T}$. If R exceeds D, then the Earth can be out of thermal equilibrium and low values of Ur are viable without requiring a weak, or negative, relationship between heat loss and Ra. A simplified causal loop diagram (Figure 4a) can help elucidate how imbalances in water cycling can drive the mantle out of thermal equilibrium. The analysis of Crowley et al. [2011] re-opened the possibility of planetary self-regulation but it did not investigate wether it was consistent with petrological constraints on Earth cooling. A recent study has addressed that question [Seales et al., 2021].

Seales et al. [2022] explored thermal and deep-water cycling models constrained to match thermal cooling paths consistent with petrological data [Herzberg et al., 2010, Condie et al., 2016, Ganne and Feng, 2017]. The models were also constrained by the present day Urey ratio and surface water content. Variable β values were allowed for. For each thermal path, one-hundred different combinations of Ur and β were sampled, within data bounds, and inverted for mantle water content. This involved converting a forward model of coupled thermal and water history [Sandu et al., 2011; Seales and Lenardic 2020b] into an inverse model [Seales et al., 2022]. Figure 4b shows the full causal loop diagram for the coupled thermal and water cycling model. With the data constraints, the evolution of mantle and surface water content was determined throughout Earth's history. This procedure produced over 10,000 evolution paths.

Figure 5a shows mantle cooling paths that met a goodness of fit criteria with petrological data, allowing for data uncertainties. Figure 5b shows the density of successful $Ur - \beta$ space. Successful models gathered towards the lower Ur bound with $\beta \geq 0.2$. Figure 5c shows mantle water evolution. Model and data uncertainties demand that outputs be calculated as probability distributions. The median of the distribution is depicted as a thick black line. The darker region is bounded by the upper and lower quartiles.

The lighter region extends to one and half times the interquartile range. Successful models experienced an early period of net mantle dewatering followed by net rewatering. The water-cycling switch is reflected in the evolution of mantle viscosity (Figure 5d).

The combined effects of thermal and water cycle feedbacks lead to mild variations in the mantle Rayleigh number over model evolution time (Figure 6a). Mild Rayleigh number variations, in the face of declining internal heat sources, is indicative of a self-regulated mantle evolution. Another measure of self-regulation is the ratio of the mantle geotherm to the mantle solidus, termed a homologous temperature (T_H). The greater the thermal distance between mantle temperature and the solidus, the greater the value of T_H and the greater the potential of mantle melt generation. When T_H drops below unity, melting is predicted to cease. Figure 6b shows T_H evolution for successful models. The decrease over the first few billion years coincides with net mantle dewatering, which increases the solidus [Katz et al., 2003]. The change from net mantle dewatering to rewatering alters the behavior of T_H. The flattening of the slope around 2 Ga indicates that the mantle geotherm becomes locked to the solidus and the two co-evolve, i.e., mantle melt potential is self-regulated.

One might assume that a drop in T_H over the first 2 billion years of evolution would be due to mantle cooling. However, data constraints show that, over this time, mantle cooling is mild, if at all (Figure 5a). That mild cooling stage is critical to a low present-day Urey ratio, which indicates that mantle heat flow is high relative to heat generation. This requires a period of low heat flow in the past to retain heat that is then available to supply elevated present day heat flow. Successful models allow for this via a switch from net mantle dewatering to rewatering, which also leads to an initial drop in T_H followed by a self-regulated phase. The change from net dewatering to rewatering is associated with a change in the dominant sign of the water-cycling feedback.

Figure 4c isolates the water cycle from the full thermal-hydro system (Figure 4b). It shows that the water cycle allows for two feedback loops. Under hotter mantle conditions, the negative loop dominates which allows melting to lower mantle water content. This works against thermal effects on mantle cooling. As a result, the full cycle maintains a near constant mantle temperature. As mantle water content drops, mantle cooling can accelerate. This leads to colder subduction which recycles more water into the mantle [Iwamori, 2007]. The water cycle becomes dominated by its positive feedback loop (right loop of Figure 4c). This increases mantle water content over time and lowers the mantle solidus which, together with coupled thermal effects, regulates melt potential (Figure 6). In short, coupled thermal and water cycling feedbacks allow for a self-regulated mantle evolution consistent with data constraints.

Before moving to the next feedbacks, it is worth taking an aside to point out how isolating feedbacks can expose ill-posed questions. If the feedbacks of Figure 2a or Figure 4 are operative, then the question "Does mantle viscosity regulate plate velocity or does plate tectonics regulate mantle viscosity?" is misleading. It is a 'flat-earth question'. That analogy stems from medieval times when people would point to the horizon and ask "does it go on forever or is there an edge?". Flat-earth questions assume mutually exclusive answers. Feedback shatters that assumption. Causal loop analysis provides a means to isolate layers of reciprocal causality versus linear causality (e.g., A causes B). Reciprocal causality, in turn, can highlight ill-posed research questions and/or cast new light on old questions. Readers may wish to pause and consider other questions that have been posed in a similar form, e.g., "Do plates drive mantle motions or do mantle motions drive plates?"

3. Mantle Dynamics and Mantle Viscosity Structure Feedbacks

The feedbacks of the previous section connected to the temperature- and hydration-dependence of mantle viscosity. Two additional viscosity-related factors allow for added mantle feedbacks: 1) Mantle viscosity allows for non-Newtonian behavior; 2) Topography, gravity, and geoid constraints show that mantle viscosity increases with depth [Richards and Hager, 1984].

Mineral physics constraints indicate that upper mantle viscosity is non-Newtonian (a power law viscosity that displays shear weakening) while the lower mantle is Newtonian [Burgmann and Dresen, 2008; Hirth and Kohlstedt, 2015]. A non-Newtonian upper mantle allows low viscosity regions to emerge in response to mantle flow. This leads to a feedback as changes in mantle velocity gradients alter local viscosity structure which, in turn, affects velocity and velocity gradients. The feedback allows localized zones of low viscosity to emerge within the mantle [Billen and Hirth, 2007; Andrews and Billen, 2009; Jadamec and Billen, 2010; Stadler et al., 2010; Alisic et al., 2012; Jadamec, 2016]. It also allows for a more global effect in which depth-variable mantle viscosity can be dynamically generated and maintained.

King [2016] showed that a non-Newtonian upper mantle could lead to a global viscosity structure characterized by a low viscosity upper mantle above a high viscosity lower mantle. Semple and Lenardic [2018; 2020a; 2020b] used a similar approach to explore how a non-Newtonian upper mantle could impact global mantle dynamics. The models showed that upper mantle flow channelization could emerge dynamically (Figure 7). It had previously been shown that upper mantle flow channelization, into an imposed low viscosity layer, would alter the global energy balance of mantle convection [Busse et al., 2006; Lenardic et al., 2006]. The non-Newtonian models showed that similar effects could emerge without a pre-existing low viscosity region. An added feedback was identified as the emerging channel promoted long wavelength mantle flow which increased mantle velocity and shear gradients. This, in turn, increased the viscosity variation between the upper and lower mantle. The enhanced viscosity variation favored longer wavelength flow. Figure 8a shows a loop diagram of the feedback cycles.

The two right side loops of Figure 8a are positive feedbacks. The system can not, however, runaway as geometry limits the maximum wavelength of mantle convection (the potential effects of continents can also provide wavelength limits [Zhong et al., 2007]). Another limiter can come from a negative feedback (the left loop of Figure 8a). Longer wavelengths will be associated with tectonic plates of greater lateral extent. Those plates will have greater ages when they enter subduction zones. The greater ages can lead to greater plate thickness and enhanced resistance to plate bending, which tends to lower plate velocity [Conrad and Hager, 1999]. That potential is in line with results from the previous section which showed that models with plate strength contributing to cooling efficiency, along with mantle viscosity, are consistent with data constraints (Figure 5).

Figure 8a provides an example of how isolating feedbacks can cast new light on old questions. Seismic tomography and geoid modeling are consistent with long wavelength mantle convection [Su and Dziewonski, 1992; Hager and Richards, 1989]. Long-wavelength convection is not the norm at the level of convective vigor inferred for the mantle. This presented a question: What allows for long-wavelength convection? A potential solution came from numerical simulations that generated long wavelength convection by imposing a viscosity increase from the upper to the lower mantle [Bunge et al., 1996, 1997; Hansen et al., 1993; Tackley, 1996; Zhang and Yuen, 1995; Zhong et al., 2000]. The physical mechanism behind this observation was elucidated via boundary layer theory [Busse et al., 2006; Lenardic et al., 2006]. In a convecting layer, with no internal viscosity variations, long-wavelength cells become unstable because lateral viscous dissipation dominates over vertical dissipation. Lateral dissipation increases with cell wavelength. Depth-variable viscosity changes the mantle global energy balance such that vertical dissipation can become dominant over a broader wavelength range. As mantle depth is fixed, the vertical term does not increase in the same way as the lateral term does with increasing flow wavelength. This allows long wavelength convection to remain stable over a broader wavelength band (a prediction confirmed via numerical stability analysis [Ahmed and Lenardic, 2010]). It also allows convective velocities to increase with wavelength [Hoink and Lenardic, 2008; 2010; Hoink et al., 2011].

Following the progression above, could lead to answering the question of 'what is the cause of long wavelength mantle convection?' with 'depth-variable viscosity'. Figure 12 shows what can be missed by applying that mode of thinking to a feedback process. Linear cause and effect thinking could limit one from considering the potential that depth-variable viscosity could, itself, be a property that is dependent on long wavelength flow. In the reciprocal causality view, depth-variable viscosity is connected to long wavelength flow but it is not necessarily the cause of it. The 'cause' could be an internal fluctuation that initiates an amplifying feedback which, in turn, leads to a system restructuring characterized by co-dependent depth-variable viscosity and long-wavelength flow. Readers may again wish to pause and consider other questions that have been framed in a linear cause and effect manner, e.g., 'What is the cause of plate tectonics?'

The feedback of this section can be connected to the previous section. A non-Newtonian upper mantle feedback provides an amplifier for a water cycling feedback (Figure 8b). Relatively small changes in upper mantle water content can enhance plate velocities and associated upper mantle shear. Increased shear will further lower viscosity on a relatively rapid time scale. This feeds back on the water cycle and lowers its reactance time. A lower reactance time, in turn, enhances the effectiveness of the water cycle feedback.

4. Boundary Layer Interactions and Plate-Plume Feedbacks

Mantle convection is driven from the decay of radioactive elements (internal heating) and heat flowing into the mantle from the core (basal heating). A range of studies have mapped similarities and differences between mixed mode heating and internal or basal heating end-members [Schubert and Anderson, 1985; Grasset and Parmentier, 1998; Sotin and Labrosse, 1999; Moore, 2008; Shahnas et al., 2008; Choblet and Parmentier, 2009; Wolstencroftet al., 2009; O'Farrell and Lowman, 2010; Choblet, 2012; Deschamps et al., 2010; 2012; O'Farrell et al., 2013; Stein et al., 2013; Weller et al., 2015; Korenaga, 2017; Vilella and Deschamps, 2018; Vilella et al., 2018]. A recently identified difference is that, in a mixed heating layer, surface velocity can decrease with increasing internal heat sources [Lenardic et al., 2020]. That observation connects to a mantle dynamics feedback and associated feedbacks between tectonic plates and mantle plumes.

Figure 9a shows a representative case from a suite of numerical convection experiments driven by a mix of internal and basal heating [Lenardic et al., 2020]. The presence of two heat sources means that a bottom heating (Ra) and an internal heating Rayleigh number (Ra_i) characterize the system. Figure 9b plots surface velocities and rms system velocities, from the experimental suite, as functions of the two Rayleigh numbers. Decreasing velocities, with increased internal heating, is not an expectation based on classic ideas as to how velocity should scale with mantle heat sources [Schubert et al., 2001].

Classical convective scalings assume that thermal boundary layers behave in a self-determined manner [Howard, 1966]. Under that assumption, scaling trends can be derived via a local stability criteria that governs boundary layer thickness. A local Rayleigh number, Ra_δ, can be defined for the upper thermal boundary layer as

$$Ra_\delta = \frac{\rho g \alpha \left(T_i - T_s\right) \delta^3}{\kappa \eta} \tag{7}$$

where δ is the boundary layer thickness. For constant surface temperature (T_s), the internal temperature of the convecting layer (T_i) provides a measure for the temperature drop across the boundary layer. When Ra_δ exceeds a critical value, given by Ra_c, convective instabilities form and lower portions of the boundary layer detach. This maintains the boundary layer at a critical thickness and a statistically steady state is achieved with $Ra_\delta = Ra_c$. A prediction that follows is that $\delta \sim Ra^{-1/3}$. A second prediction is that $\delta \propto \Delta T^{-1/3}$. The connection to boundary layer velocity (u) comes from a balance of conductive and advective time scales. The time it takes heat to diffuse across the boundary layer is given by τ. This leads to $\delta = \sqrt{\kappa\tau}$. An advective time scale can be defined as aD/u, where a is the aspect ratio of a convection cell. If the boundary layer becomes unstable at a critical thickness, then the two time scales are equal and $\delta \propto u^{-1/2}$. This amounts to applying Howards criteria [1966] to cellular convection. Boundary layer theory, which explicitly solves for velocity as a function of aspect ratio, leads to an equivalent relationship between δ and u [Turcotte and Oxburgh, 1967]. In combination, the above predicts that $u \propto \Delta T^{2/3}$. Increasing internal temperature is predicted to increase boundary layer velocity. The thermal history community has leaned on this prediction for decades [Schubert et al., 2001].

A decrease in upper boundary layer velocity with increased internal heating (Figure 9b) suggests that the concept of a self-determined boundary layer may be incomplete. For bottom heated, high Reynolds number convection it has been shown that a self-determined boundary layer regime is not achieved even at what are considered to be very high Rayleigh numbers [Castaing et al., 1989]. This behavior connects to an inertial wind that shears the upper boundary layer and prevents it from achieving a critical thickness [Kadanoff, 2001]. Although inertial effects are absent in mantle convection, boundary layer interactions can occur [Weinstein et al., 1989; 1990; Lenardic and Kaula, 1994; Labrosse, 2002; Galsa and Lenkey, 2007; Vilella and Deschamps, 2018]. Moore [2008] has argued that such interactions will short-circuit self-determined boundary layer behavior.

Self-determined boundary layers require weak interactions between upper and lower boundary layers. At very high degrees of convective vigor this will be achieved as upwelling plumes, for example, will dissipate before they impact the upper boundary layer. Moore [2008] noted that discrepancies between classic scaling predictions and experimental results, even at values considered to be high Ra, indicate that mantle convection may not reach that limit. For example, the numerical experiments of Lenardic and Moresi [2008] indicated that classic scaling trends are approached asymptotically for $Ra > 10^9$. If boundary layers do interact, then the upper thermal boundary layer may not reach a critical thickness as per the theory of Howard [1966]. All other factors being equal, a decrease in boundary layer interaction could then lead to a decrease in boundary layer velocity. Increased internal heating, in a mixed heated layer, could progressively reduce boundary layer interactions as thermal upwellings become weaker and have less effect on the upper thermal boundary layer [Labrosse, 2003; Vilella and Deschamps, 2018].

The conceptual idea above can be formalized [Lenardic et al., 2020]. The theory of Moore [2008] predicts that the temperature drop across the upper thermal boundary layer (ΔT_{top}) scales as

$$\Delta T_{top} = 0.499 + 1.33 Q^{3/4} Ra^{-1/4} \tag{8}$$

and that the non-dimensional heat flow (Nu) across the upper boundary scales as

$$Nu_{top} - 1 = Nu_{bot} - 1 + Q = 0.5Q + 0.206(Ra - Ra_c)^{0.318} \tag{9}$$

where Q is the ratio of internal to bottom heating Rayleigh numbers. An upper boundary layer thickness (δ) can be derived from the ratio of ΔT_{top} and Nu_{top}. Recalling that $\delta \sim u^{-1/2}$, this predicts that the upper boundary velocity scales as

$$\frac{u}{U} = \left(\frac{0.5Q + 0.206(Ra - Ra_c)^{0.318}}{0.499 + 1.33 Q^{3/4} Ra^{-1/4}} \right)^2 \tag{10}$$

where U is a scaling constant. The theory predicts that regions exist, within Ra and Q space, over which increased internal heating leads to a decrease in upper boundary layer velocity. This occurs despite the fact that surface heat flux increases. The predictions were shown to be consistent with suites of numerical convection experiments [Lenardic et al., 2020; Weller and Lenardic, 2016; Weller et al., 2016].

Figure 10a places the ideas above into a feedback context. As an example sequence, consider an increase in upper boundary velocity. This lowers internal temperature, via enhanced cooling, which increases the temperature drop across the lower thermal boundary layer. That increases the velocity of plumes that form from the boundary layer. The plumes interact with the upper boundary layer and affect its velocity. That sequence applies to the inner, positive feedback loop at the right of Figure 10a. The outer loop, which connects to plume thickness, is a negative feedback. At very high degrees of convective vigor plumes become very thin and can dissipate before they can interact with the upper boundary layer. The other way a no boundary layer interaction limit can be hit is if the temperature drop across the lower boundary layer goes to zero, i.e., the pure internal heating limit. Before that limit is hit, the feedbacks of Figure 10a allow upper boundary layer velocity to decrease as the ratio of internal to basal heating increases.

Figure 10a applies to isoviscous convection. For application to the mantle, boundary layer feedbacks will connect to viscosity feedbacks. Figure

10b shows an example. The coupled feedbacks allow the effect of increased radiogenic heating on lowering mantle viscosity to outweigh its effect on increasing the internal to basal heating ratio. As such, plate velocities can increase with increased radiogenic heating, in a mixed heating mantle, for strongly temperature-dependent viscosity and weak plate margins [Lenardic et al., 2020]. None the less, boundary layer feedbacks remain operative and continue to affect plate velocity. The full trade-offs, between all the system parameters that come into play, remain to be worked out, i.e., it is an avenue for future research.

Boundary layer interactions relate to the potential influence of mantle plumes on tectonic plates. The motion of tectonic plates has long been connected to a low viscosity layer in the upper mantle, i.e., the asthenosphere. The idea that the viscosity of the asthenosphere could be influenced by mantle plumes was not originally considered as indicative of a mantle feedback but, we will argue, it is part of a larger feedback loop between tectonic plates and mantle plumes.

Figure 11a shows a conceptual model that attributes the Earth's asthenosphere to a plume generated thermal inversion [Deffeyes, 1972]. Morgan et al. [1995; 2013] expanded the model to include newer insights regarding plume dynamics [Olson and Singer, 1985; Richards et al., 1989; Davies, 1990; Loper, 1991]. Figure 11b is from Morgan et al. [1995; 2013] with an addition highlighting the type of plumes required for a plume fed asthenosphere.

To create a low viscosity asthenosphere, plumes must be hotter than the background mantle they rise through. Excess temperatures, inferred for present-day mantle plumes, are roughly 200 degrees [Jellinek and Manga, 2004]. This implies a viscosity variation of two orders of magnitude between plumes and the background mantle [Kohlstedt et al., 1995]. Under such conditions, plume morphology is characterized by a large plume head and a thin low viscosity tail that connects the plume to the basal thermal boundary layer it originates from. The tail is associated with upwelling velocities significantly greater than background flow [Loper and Stacy, 1983; Sleep, 2004; Thayalan, 2006]. Plumes of that type are termed cavity plumes [Olson and Singer, 1985].

Once a cavity plume head has risen and impacted the lithosphere, the tail can remain in place as a low viscosity conduit that maintains a high flow velocity and brings hot material from the core mantle boundary to the base of plates with little thermal loss [Richards et al., 1989]. If the tail moved slower, due to a viscosity value closer to that of the mantle, diffusion could lower the thermal anomaly plumes could maintain at the base of plates.

A plume-fed, or plume-influenced, asthenosphere is an added means by which the lower boundary layer of mantle convection (a plume source) can influence the motion of tectonic plates. On its own, that does not constitute a feedback processes. A feedback requires that plate motion, in turn, influences the lower boundary layer and plumes that originate from it. That link connects to the existence of plumes with a viscosity significantly lower than background mantle, i.e., to the existence of cavity plumes in the mantle.

Nataf [1981] first highlighted the difficulty of generating cavity plumes in a mantle with temperature-dependent viscosity. Temperature-dependent viscosity can lead to a stagnant lid mode of convection - an analog for a single plate planet. In that mode, the upper boundary layer absorbs the bulk of the total system viscosity contrast. As a result, mantle upwellings, in a single plate mode, are expected to have nearly the same viscosity as the mantle they rise through [Solomatov and Moresi, 2000; Labrosse, 2002]. Such thermals can not generate and maintain a low viscosity region below plates.

The lack of cavity plumes, in stagnant lid convection, points toward the need for plate subduction and associated mantle cooling [Lenardic and Kaula, 1994; Jellinek et al., 2002; Thayalan et al., 2006]. Figure 12a is from Robin et al. [2007] who used numerical and laboratory experiments to show how introducing overturn of the cold upper boundary layer, to a system that was initially in a stagnant lid mode, could lead to a morphological change in mantle upwellings. During the transition, thermals and cavity plumes coexisted in the experiments. As the system moved toward a statistically steady-state, thermals were globally replaced by cavity plumes due to the large temperature, and associated viscosity, variation across the lower thermal boundary layer. The numerical experiments included internal heating, as well as basal heating, which prevented thermals from reaching the upper boundary layer. Once cavity plumes formed they did reach the base of the upper thermal boundary layer.

The link above completes a feedback loop that connects tectonic plates to mantle plumes (Figure 12b). As well as isolating feedbacks, Figure 12b provides pointers to research questions that contain *a priori* assumptions about system dynamics that may not be correct and, as such, could lead to blind alleys. For example: 'Do mantle plumes initiate plate-tectonics?'; 'Do mantle plumes lead to an asthenosphere?'; 'Is the lack of plate tectonics on Venus due to the lack of an asthenosphere?'. The last question, as a specific example, would be changed by an appreciation of a feedback cycle in which the existence of an asthenosphere depends on plate tectonics and acts to maintain plate tectonics.

The discussion above also provides an example of how linear cause and effect thinking can influence the way observational data is interpreted. Observational data indicates that Venus lacks plate tectonics and lacks gravity-topography signatures indicative of an asthenosphere [Kaula and Phillips, 1981; Kiefer et al., 1986]. From the earliest days of mapping Venus' topography and gravity, and into the present day, the dominant interpretation has been that this indicates that a lack of an asthenosphere leads to a lack of plate tectonics on a terrestrial planet [Smrekar et al., 2007]. An equally valid interpretation is that a lack of plate tectonics leads to the lack of an asthenosphere. A third, data consistent, interpretation is that plate-tectonics and an asthenosphere are co-dependent components of a broader system. Our intent is not to argue that the last interpretation is correct but to show how an under-appreciation of feedbacks can limit the generation and exploration of multiple working hypotheses [Chamberlin, 1897].

5. Plate Tectonics-Mantle Dynamics Feedbacks and Bootstrap Hypotheses

A thermal inversion leading to a low viscosity upper mantle, as per Figure 11, does not require mantle plumes. It can also be generated by a sub-adiabatic thermal gradient [Stein and Hansen, 2008]. Several studies have argued that a sub-adiabatic thermal gradient exists in the Earth's mantle [Jeanloz and Morris, 1987; Lenardic and Kaula, 1994; Bunge et al., 2001; Matyska and Yuen, 2001; Sleep, 2003; Bunge, 2005; Sinha and Butler, 2007; Moore, 2008; Weller et al., 2016]. A consistent result is that increased internal heating favors a sub-adiabatic mantle, i.e., a hot upper mantle above a cooler lower mantle. This allows a thermal inversion to be maintained with increased internal heating even though that increase weakens the potential of a plume-fed inversion (Figure 10).

Increased internal heating favors a sub-adiabatic mantle via an asymmetry between upwelling and downwelling velocities. In a dominantly internally heated mantle, broad background upwelling balances concentrated downwellings associated with subducting slabs. Slabs can deposit cold fluid to the system base with little heating on descent. The slower moving, broad upwelling can experience heating as it rises. As a result, the mantle interior becomes sub-adiabatic [Jeanloz and Morris, 1987]. Mass balance suggests that longer aspect ratio cells can enhance sub-adiabatic gradients. The area of diffuse mantle upwelling increases with wavelength, which decreases its velocity. As a result, heating of the upwelling as it rises tends to increase. That prediction is consistent with numerical convection experiments [Höink and Lenardic, 2008; 2010; Lenardic et al., 2019]. Figure 13a shows thermal profiles from spherical geometry experiments. Figure 13b shows results from Cartesian experiments that allowed for a systematic exploration of wavelength effects by varying the lateral extent of the modeling domain.

The experiments of Figure 13 generated long wavelength flow by imposing a high viscosity lower mantle. Experiments with temperature-dependent viscosity showed that long wavelength flow could be maintained without an imposed viscosity variation [Lenardic et al., 2019]. Figure 14a (top) shows an example. The experiment has a temperature-dependent mantle viscosity and a rheology that allows for the formation of near surface weak zones that are analogs for plate margins [Moresi and Solomatov, 1998]. The weak margins allow the otherwise high viscosity upper boundary layer, a plate analog, to subduct and cool the interior mantle (an active lid mode of convection). Also shown is a case that did not allow for the formation of plate margins, i.e., a stagnant lid mode (Figure 14a, bottom). Internal viscosity variations are shown at the left of each image. The active lid case lead to an upper mantle with a viscosity that was significantly lower than the mid-mantle. That variation resulted from a sub-adiabatic mantle and was dynamically maintained by the active-lid mode of convection. In a stagnant-lid mode (single plate planet), the lack of cold sinking slabs lead to a nearly uniform internal temperature. As a result, internal viscosity variations were mild.

Experiments of the type in Figure 14a build off of Lowman et al. [2001] and King et al. [2002] who proposed the existence of a hot mantle layer below tectonic plates. Together with temperature-dependent viscosity, a hot upper mantle can generate a low viscosity layer and flow channelization below plates [Lenardic et al., 2019]. Channelization allows long-wavelength cells to remain stable, as per the theoretical expectations of Busse et al. [2006]. Also consistent with theory, flow channelization allows long wavelength cells to be more efficient in cooling the interior than would be the case if mantle flow did not channelize. The novel aspect, compared to experiments previously used to test theoretical predictions [Lenardic et al. [2006], is that no imposed increase in viscosity is required. Rather, depth-variable viscosity emerges dynamically and contributes to long wavelength flow as well as be influenced by long wavelength flow. An added effect relates to the maintenance of weak plate margins.

Figure 14b plots results from two experimental suites with different values of a convective stress level required to generate weak plate boundary zones. If plate margins can be created under any level of convective velocities, then wavelength effects on velocity will not feed into the generation of plate margins (the low margin strength suite of Figure 14b). If plate margins require a critical kinetic energy level, which relates to available work, then flow wavelengths associated with greater convective velocities will favor the generation of plate margins (the medium margin strength suite of Figure 14b). The specific tradeoffs between flow wavelength and margin generation will depend on the particular rheology used to model plate margin processes [Bercovici, 2015; Crowley and O'Connell, 2012]. However, the general results of Figure 14b only require that a critical condition on available convective work exists and that increased wavelength can lead to increased convective velocities.

Figure 15a shows a loop diagram for the feedbacks of this section. The feedbacks lead to the hypothesis that subduction generates a sub-adiabatic mantle and an associated increase of viscosity with depth. As a result, a low viscosity layer forms in the upper mantle which leads to flow channelization. Channelization feeds into generating long wavelength mantle flow and plate margins. Both stabilize plate tectonics. Plates, an asthenosphere, and a long wavelength component of mantle flow all depend on each other and are critical to the existence of each other. Hypotheses of that type are referred to as bootstrap hypotheses as they argue that no critical entity, for the particular aspect of nature a hypothesis is applied to, can exist independent of other entities, i.e., there are no 'fundamental entities' that the system can be reduced to [Chew, 1968; Cahill and Klinger, 2005; Kazansky, 2004; 2010].

Starting a self-sustaining system, whose operative entities are all co-dependent, can seem as difficult as pulling oneself up by ones own bootstraps. A well known example is that in order to boot up a computer, computer software must be loaded and initiated by computer software. When operational, the system is self-sustaining but the operational loops that allow for that need to be activated. To start a computer, a small amount of relatively simple code is needed to progressively load more complex code until the computer becomes self-operational. Once functioning, the boot up code is no longer needed and its presence does not show up in the system operation. The connection to plate tectonics will be taken up in the next section.

6. Discussion and Conclusion

The theory of plate tectonics, as formulated, is a kinematic one [McKenzie and Parker, 1967; Morgan, 1968; LePichon, 1968]. Extending it to a dynamic theory became a research avenue soon after the plate tectonics revolution and it remains so to this day [Cox, 1973; Coltice et al., 2017; Hawkesworth and Brown, 2018]. In efforts to extend plate tectonics, there are statements along the lines of "plate tectonics is due to water induced rock-weakening", "plate tectonics is due to a rheology that allows for weak plate margins", "plate tectonics is due to the presence of an asthenosphere", "plate tectonics is triggered by continental spreading", "mantle plume initiate plate tectonics", "impacts initiate plate tectonics". An even handed referencing, of hypotheses along those lines, would fill pages and our intent is to single out a commonality rather than particulars. The shared underpinning is the idea that there are some key factors that allow for plate tectonics - if a planet has those factors, then plate tectonics will follow. Tracking down the causal factors is one research avenue for addressing the question of what allows a planet to have plate tectonics. The existence of planetary feedbacks leads to an alternate way of framing the problem.

If the factors that allow for plate tectonics owe their existence to plate tectonics, then the solid Earth system cannot be broken down to causal chains. It needs to be approached from the standpoint of self-sustaining feedbacks and reciprocal causality. Figure 15b encapsulates that framework with a loop diagram that connects all the feedbacks previously discussed. A feedback framework does not imply that certain conditions are not required for the potential of plate tectonics: A planet must have sufficient internal energy to allow for surface deformation, the strength of rock can not be such that failure, and associated generation of plate margins, could not occur under any level of internal energy. A feedback framework does, however, lead to the conclusion that a planet can have all the "necessary conditions" and not have plate tectonics. The added requirement is that internal feedback loops become operative and stable.

The phrase "booting up plate tectonics" relates to the conjecture that plate tectonics is defined by internal feedbacks that, once operational, lead to the formation of non-separable system components [Lenardic et al., 2019]. Under this conjecture, plate tectonics involves the co-arising of critical system factors (one factor does not cause another - they emerge collectively with the links between them). Co-arising is not spontaneous but, from a practical standpoint, it can refer to situations in which functional elements that emerge coincide, within relatively narrow time windows, with the co-emergence of factors that are critical for the maintenance of the elements themselves [Chiatti, 2012]. From an observational standpoint, co-arising can lead to a rethinking of what is meant by 'the origin of plate tectonics'. The feedback loops of Figure 15b can be activated by a number of factors including internal fluctuations associated with chaotic convection in the mantle - fluctuations that, for all observational intent and purpose, would need to be viewed as stochastic [Weller et al., 2015; Wong and Solomatov, 2016; Lenardic et al., 2016]. Once the self-sustaining process is booted up, its operation can erase evidence of the boot up itself and what is left, observationally, is the workings of the self-sustaining process. The dynamics of the feedback loops can be studied at various levels of detail as can the factors that are required for their stability. Questions of "origin", on the other hand, require reconsideration.

A rethinking of origins for self-sustaining feedbacks, particularly in light of observational limits, is a common theme in the study of emergent phenomena [Holland, 1998; Fromm, 2004]. One can hear it stated that plate tectonics is an emergent phenomena or the closely related statement that plate tectonics is a self-organized system [Anderson, 2002; Morowitz, 2002]. Statements of that sort, and associated discussions, are often presented in a metaphorical form with no mention of feedback. If we ask 'what is it that emerges' and the answer is 'plate tectonics' then we run the risk of falling into regress. To avoid that pitfall, we need to be clear about what exactly emerges in phenomena we choose to label as emergent. We take the view that what emerges is structure/pattern. That is, a set of relationships and feedbacks resulting in functions with a level of continuity and the capability to transform and/or maintain phenomena. The feedback functions are not permanent but can operate over extend time frames such that a phenomena can remain stable over a related time frame. What emerges is not a thing, or a state, but a process [Bridgman, 1943]. If we apply this to plate tectonics, then a research avenue is to isolate operative feedback functions and determine their stability. Only then can we can ask how feedbacks could be activated and we must remain open to the possibility that the operative feedbacks could erase evidence of their initiation, i.e., multiple activation scenarios will be viable.

Figure 15b is a step towards the above. A couple of things are worth calling out. The first is system redundancies. Structural elements are maintained by more than a single feedback loop. Those redundancies enhance system stability. The value of redundancy can break a tendency toward looking for a single cause for a structural element and highlight the value in isolating new feedbacks, even if they feed into structures that can be maintained in other ways. The second thing of note is the co-existence of negative and positive feedbacks. In earth systems literature there tends to be a focus on isolating negative feedbacks as they can regulate system conditions in favorable ways. For example, a silicate-weathering feedback and/or a biology-albedo feedback can regulate planetary surface temperature in ways that are conducive to life [Walker et al., 1981; Watson and Lovelock, 1983]. Positive feedbacks, on the other hand, are associated with amplifications and instabilities that lead to system runaways. They can, however, be beneficial for the development of structure and patterns that define emergent processes. Positive feedbacks allow for time-scales needed for the emergence of co-depended elements that can lead to new system structure(s). Negative feedbacks, which often operate on slower time scales, can then stabilize the emergent structure(s) [Levin, 2000]. We note this to make it clear that isolating new positive feedbacks, in the plate tectonics system, does not necessarily lower the inferred stability of plate tectonics. It could bolster the hypothesis that plate tectonics is an emergent process.

To be clear, we can not, at this stage, say that plate tectonics is an emergent process. It may be that the fundamental entities for the operation of plate tectonics could be reduced to internal energy and lithosphere rheology and a linear cause and effect theory for the origin of plate tectonics may be viable. In that view plate tectonics may emerge in time but it is not an emergent phenomena if we hold to the view that emergent phenomena depend on feedback processes that maintain a particular structure. Not all hold to that view and, as noted above, one can find discussions of emergent plate tectonics that make no mention of feedbacks. However, if abandon the view that internal feedbacks are a defining characteristics of emergent phenomena, then anything that initiates at some point in time, or transforms in some way, would be termed emergent and one could well wonder if we are not wandering to tautology (i.e. what emerges is emergent).

At this stage, we hold to the view that the feedbacks of Figure 15b, and the idea that plate tectonics depends on feedbacks, provide frameworks for future research. A general research path is to isolate added internal planet feedbacks and, as appropriate, connect them to surface feedbacks. This can enhance an appreciation that tools developed for general feedback analysis have utility for solid planet dynamics [Astrom and Murray, 2008]. A more specific path relates to how different the evolution of planets that allow for self-regulation is from planets that do not (Figure 3). This motivates a focus on collecting new data constraints, and re-examining existing ones, with the target of discriminating between the hypotheses that the Earth's evolution did or did not involve self regulation. Added research lines are to provide constraints on the operative times scales of feedbacks and stability analysis of individual loops in Figure 15b. A focus on feedback stability provides a useful flip on the question of what maintains plate tectonics by posing the question of what would be required to break the operation of plate tectonics.

The hypothesis of emergent plate tectonics provides for added research avenues that extend into planetary science. If plate tectonics is emergent, then it is possible that planets could experience episodes of subduction that do not activate and stabilize feedback loops needed to maintain plate tectonics. As the system is booting up, it is at its lowest state of resilience and nascent plate tectonics could become unstable. As feedbacks become operational, plate tectonics becomes more resilient. The implication is that boot up cycles may or may not run to completion based on the particulars of a planet's evolution path during the boot up sequence and when, in a planet's geologic life time, boot up sequences are activated. Stated another way, a planet can have the necessary factors for plate tectonics and still not have plate tectonics based on historical contingencies affecting the formation and stability of internal feedback loops. This connects the idea of emergent plate tectonics to the hypothesis of bistable tectonics [Sleep, 2000; Tackley, 2000; Crowley and O'Connell, 2012; Weller and Lenardic, 2012; Lenardic and Crowley, 2012; Lenardic et al., 2016].

Bistability allows different tectonic modes to exist under the same physical/chemical conditions and planetary age. Which mode is realized depends on historical contingency (not to be confused with randomness [Bohm, 1957]). The seeds of that idea go back to a model of subduction initiation - if multiple stable modes can exist, then subduction initiation could require a large amplitude perturbation to move from one attracting mode to another [McKenzie, 1977]. That initial shift away from one mode (e.g., a single plate planet) can lead to the dynamic enhancement of the attractor for another mode (e.g., plate tectonics) via the formation of internal feedback loops. Before feedbacks are stabilized, the tendency of the system will be to move back to a single plate mode, i.e., the plate tectonic attractor may be weak. The activation of feedback loops can deepen the plate tectonic attracting well making it progressively more stable. If the sequence runs to completion, then there is a transition. If not, then the system falls back to the single plate mode. There are two research lines that connect to observational support for the ideas of the last two paragraphs. One connects to Earth history. The other to forth coming planetary observations.

As laid out above, emergent plate tectonics allows for failed boot-up sequences in the Earth's past. The idea that the Earth has experienced episodes of 'failed plate tectonics' has recently been put forward [O'Neill et al., 2018]. A focus on collecting data to confirm or refute that idea could provide a fruitful research path for understanding our own planets evolution and for comparative planetology. Forth coming missions to Venus will provide for a research line aimed at confirming or refuting the idea that localized subduction occurred in Venus' past even though plate tectonics did not [Davaille et al., 2017]. As well as testing the idea of failed boot-up sequences, this could support the idea of planetary bistability. Forthcoming observations of terrestrial exoplanets also have the potential to test that hypothesis as the statistical distribution of planetary properties will be different if planets do or do not allow for bistability [Bean et al., 2017; Checlair, 2019; Lenardic and Seales, 2021]. Whatever the result, the range of potential solid planet feedbacks laid out in this chapter highlight the value of feedback analysis for a range of solid planet research problems. That serves as a conclusion and a motivation for future work that connects feedback to mantle dynamics and plate tectonics.

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Acknowledgements

We thank the editor for feedback.

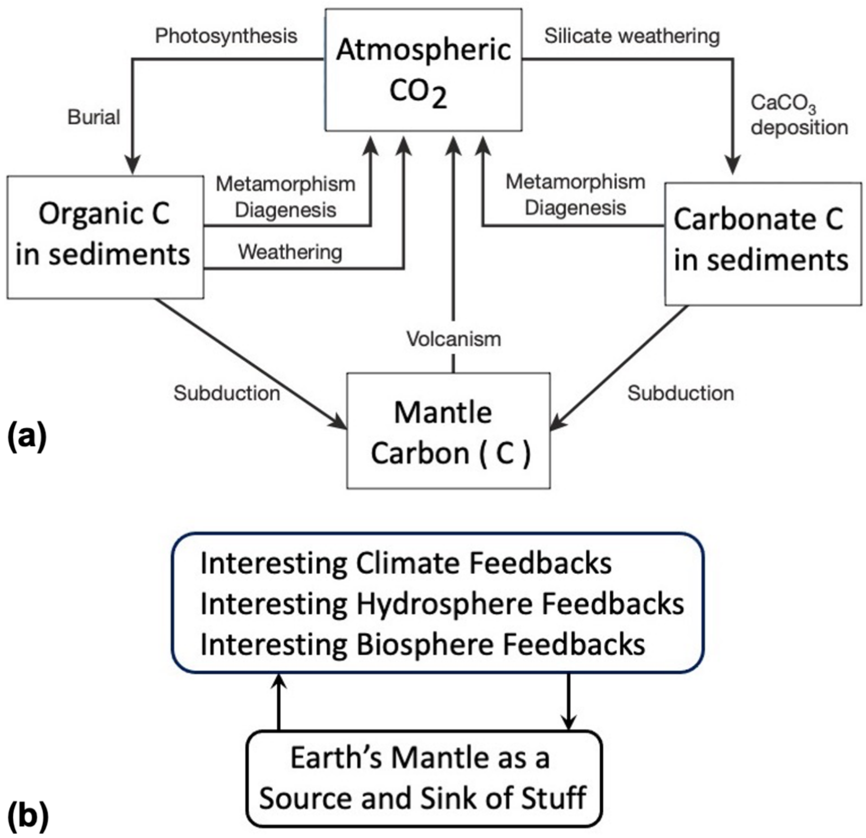
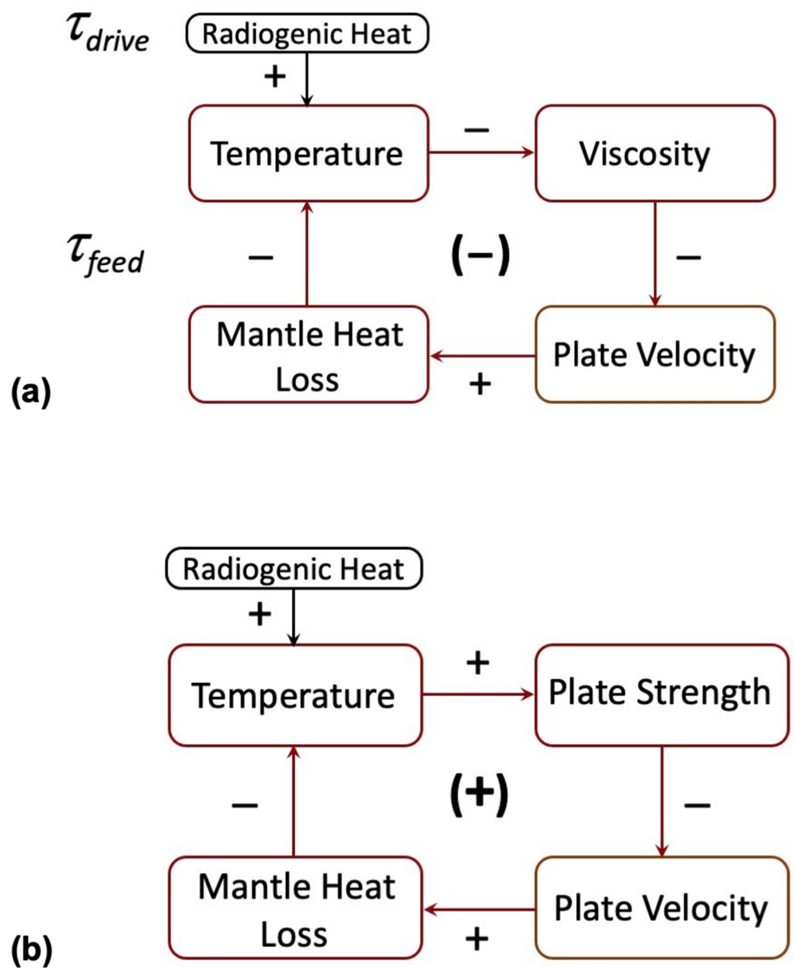

Photosynthesis
Atmospheric CO_2
Silicate weathering
Burial
$CaCO_3$ deposition
Organic C in sediments
Metamorphism Diagenesis
Weathering
Metamorphism Diagenesis
Carbonate C in sediments
Volcanism
Subduction
Subduction
(a)
Mantle Carbon (C)
Interesting Climate Feedbacks
Interesting Hydrosphere Feedbacks
Interesting Biosphere Feedbacks
Earth's Mantle as a Source and Sink of Stuff
(b)

Figure 1:


τ_{drive}
Radiogenic Heat
+
Temperature
–
Viscosity
τ_{feed}
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Mantle Heat Loss
Plate Velocity
+
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Radiogenic Heat
+
Temperature
+
Plate Strength
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(+)
–
Mantle Heat Loss
Plate Velocity
+
(b)

Figure 2:

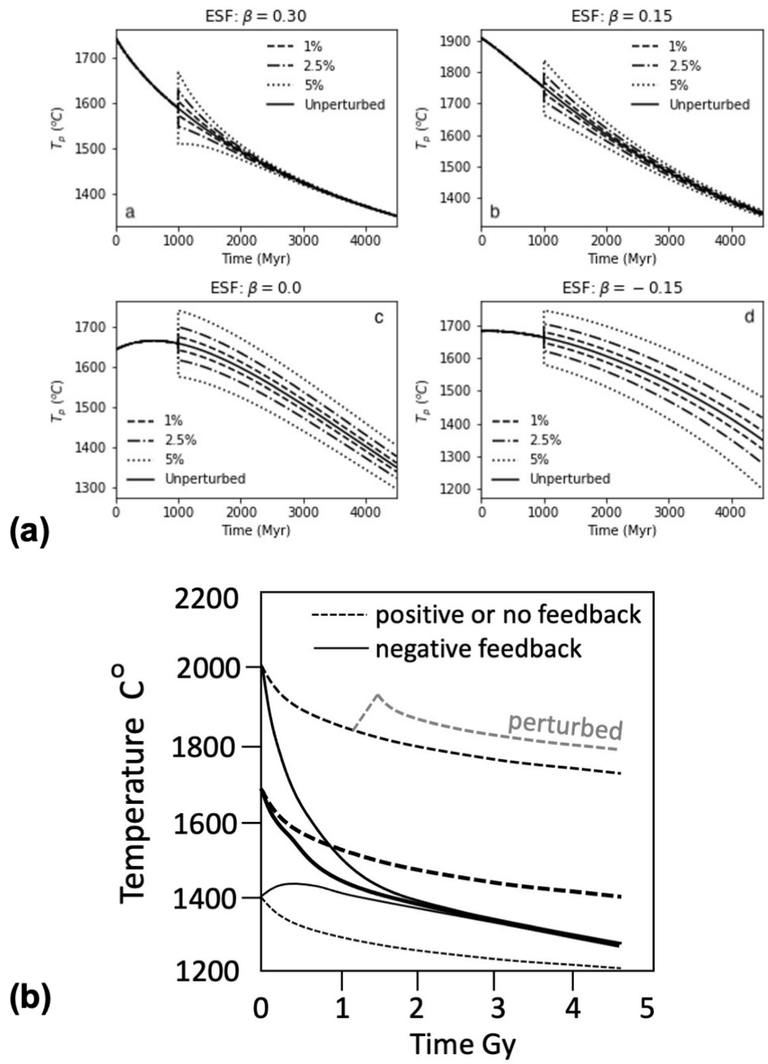

(a)
ESF: β = 0.30
ESF: β = 0.15
ESF: β = 0.0
ESF: β = − 0.15
1%
2.5%
5%
Unperturbed
a
b
c
d
T_p (°C)
Time (Myr)
(b)
positive or no feedback
negative feedback
perturbed
Temperature C°
Time Gy

Figure 3:

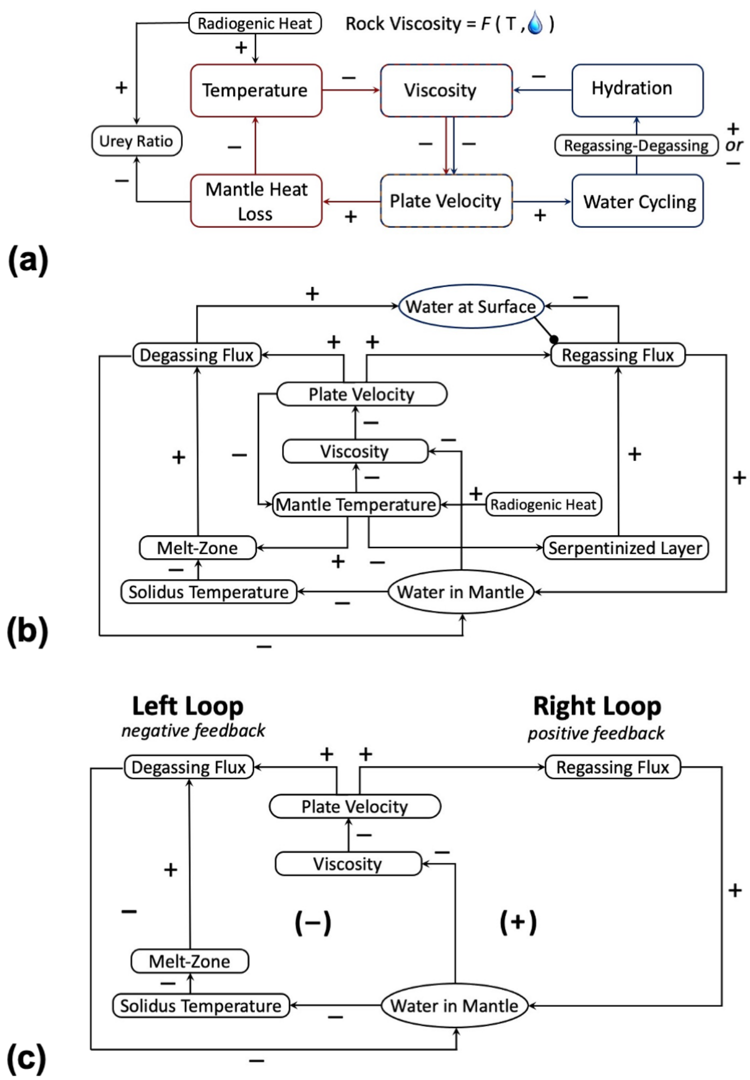

Radiogenic Heat
Rock Viscosity = F (T ,)
Temperature
Viscosity
Hydration
Urey Ratio
Regassing-Degassing
+ or −
Mantle Heat Loss
Plate Velocity
Water Cycling
(a)
Water at Surface
Degassing Flux
Regassing Flux
Plate Velocity
Viscosity
Mantle Temperature
Radiogenic Heat
Melt-Zone
Serpentinized Layer
Solidus Temperature
Water in Mantle
(b)
Left Loop
negative feedback
Right Loop
positive feedback
Degassing Flux
Regassing Flux
Plate Velocity
Viscosity
(−)
(+)
Melt-Zone
Solidus Temperature
Water in Mantle
(c)

Figure 4:

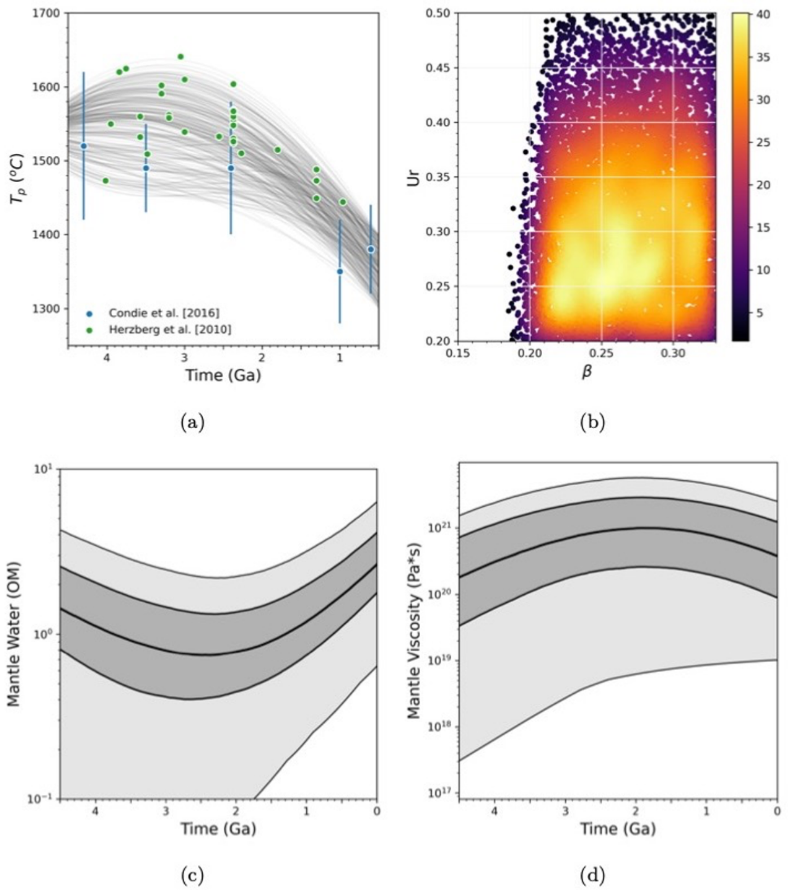

1700
1600
1500
1400
1300
T_p (°C)
Condie et al. [2016]
Herzberg et al. [2010]
Time (Ga)
(a)
Ur
β
0.50
0.45
0.40
0.35
0.30
0.25
0.20
0.15
40
35
30
25
20
15
10
5
(b)
Mantle Water (OM)
Time (Ga)
(c)
Mantle Viscosity (Pa*s)
Time (Ga)
(d)

Figure 5:

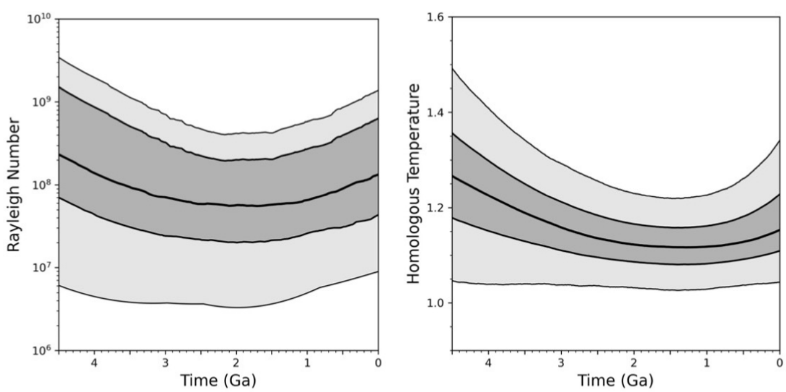


Figure 6:

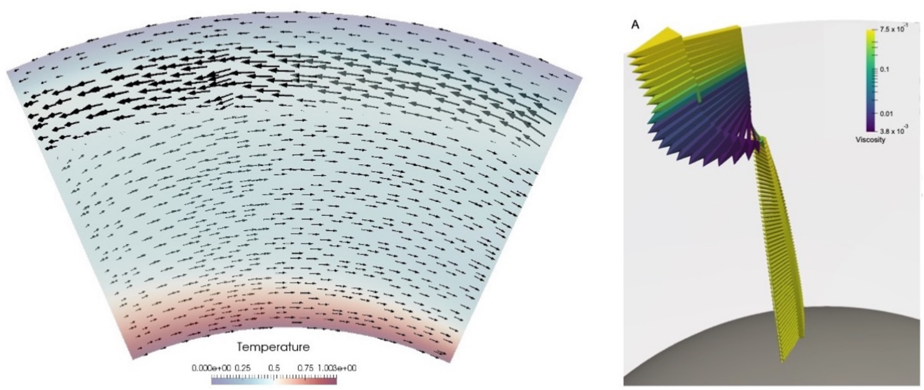

Temperature
0.000e+00 0.25 0.5 0.75 1.003e+00
A
7.5 x 10^-1
0.1
0.01
3.8 x 10^-3
Viscosity

Figure 7:

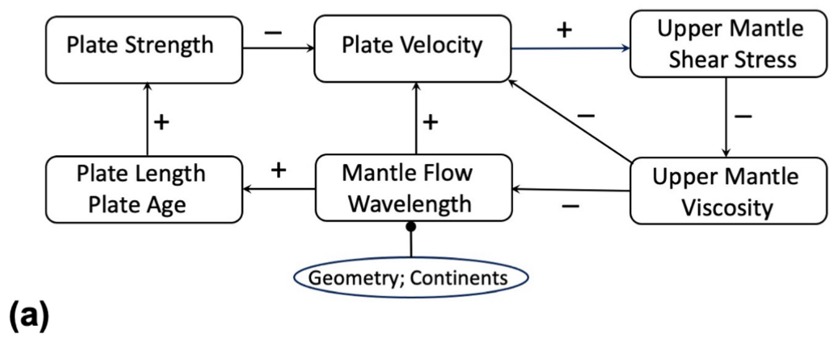


Plate Strength
Plate Velocity
Upper Mantle Shear Stress
Plate Length Plate Age
Mantle Flow Wavelength
Upper Mantle Viscosity
Geometry; Continents
(a)

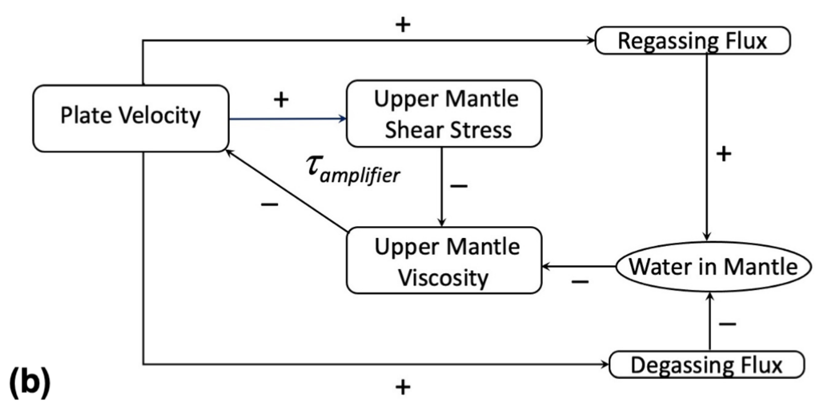
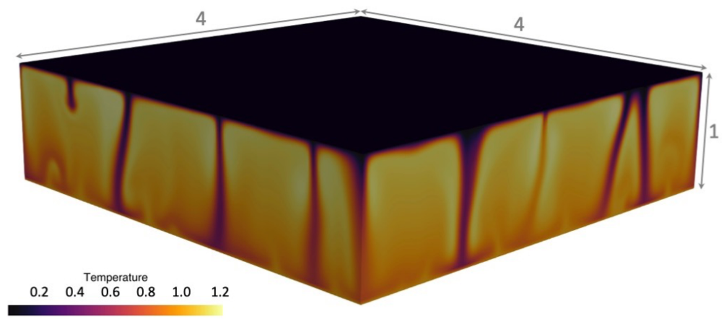
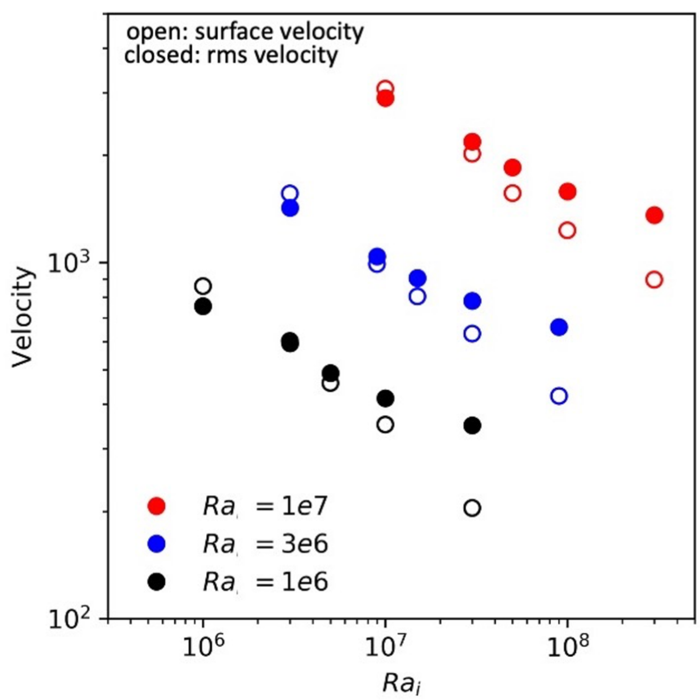

Regassing Flux
Plate Velocity
Upper Mantle Shear Stress
$\tau_{amplifier}$
Upper Mantle Viscosity
Water in Mantle
Degassing Flux
(b)

Figure 8:


4
4
1
Temperature
0.2 0.4 0.6 0.8 1.0 1.2

(a)


open: surface velocity
closed: rms velocity
Velocity
10^3
10^2
10^6
10^7
10^8
Ra_i
$Ra_i = 1e7$
$Ra_i = 3e6$
$Ra_i = 1e6$

(b)

Figure 9:

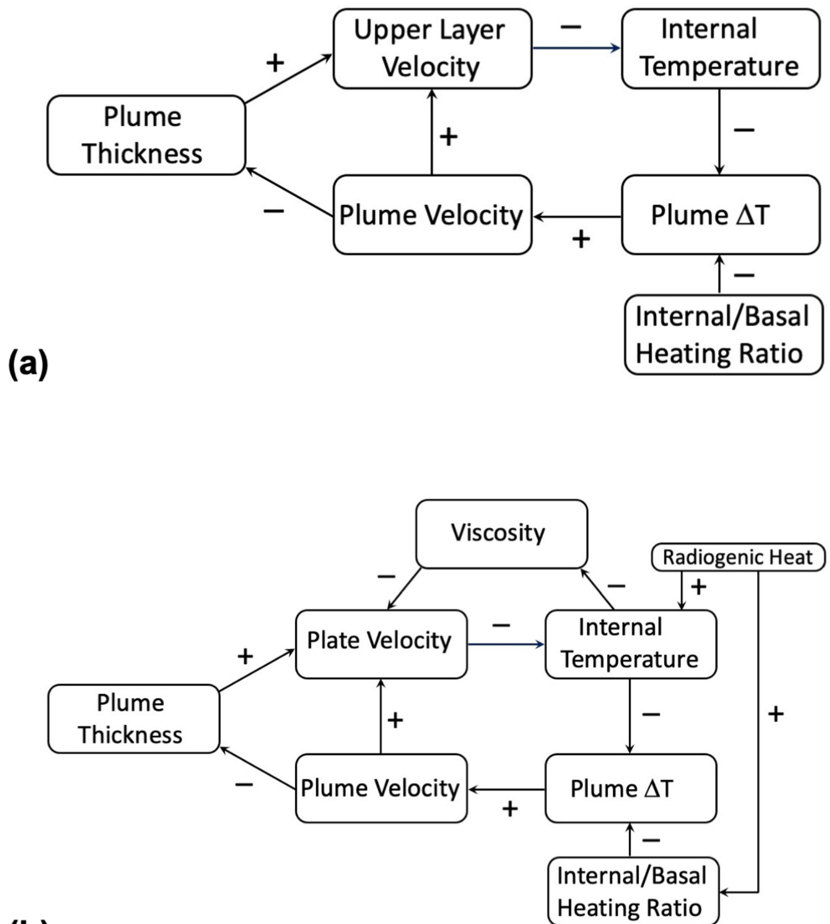

Upper Layer Velocity
Internal Temperature
Plume Thickness
Plume Velocity
Plume ΔT
Internal/Basal Heating Ratio
(a)
Viscosity
Radiogenic Heat
Plate Velocity
Internal Temperature
Plume Thickness
Plume Velocity
Plume ΔT
Internal/Basal Heating Ratio
(b)

Figure 10:

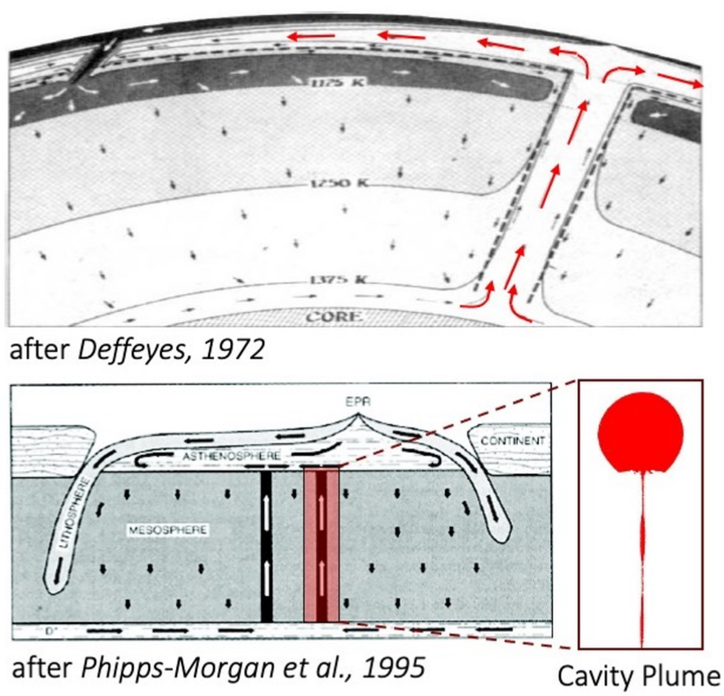

1175 K
1250 K
1375 K
CORE
after Deffeyes, 1972
EPR
CONTINENT
ASTHENOSPHERE
LITHOSPHERE
MESOSPHERE
D"
after Phipps-Morgan et al., 1995
Cavity Plume

Figure 11:

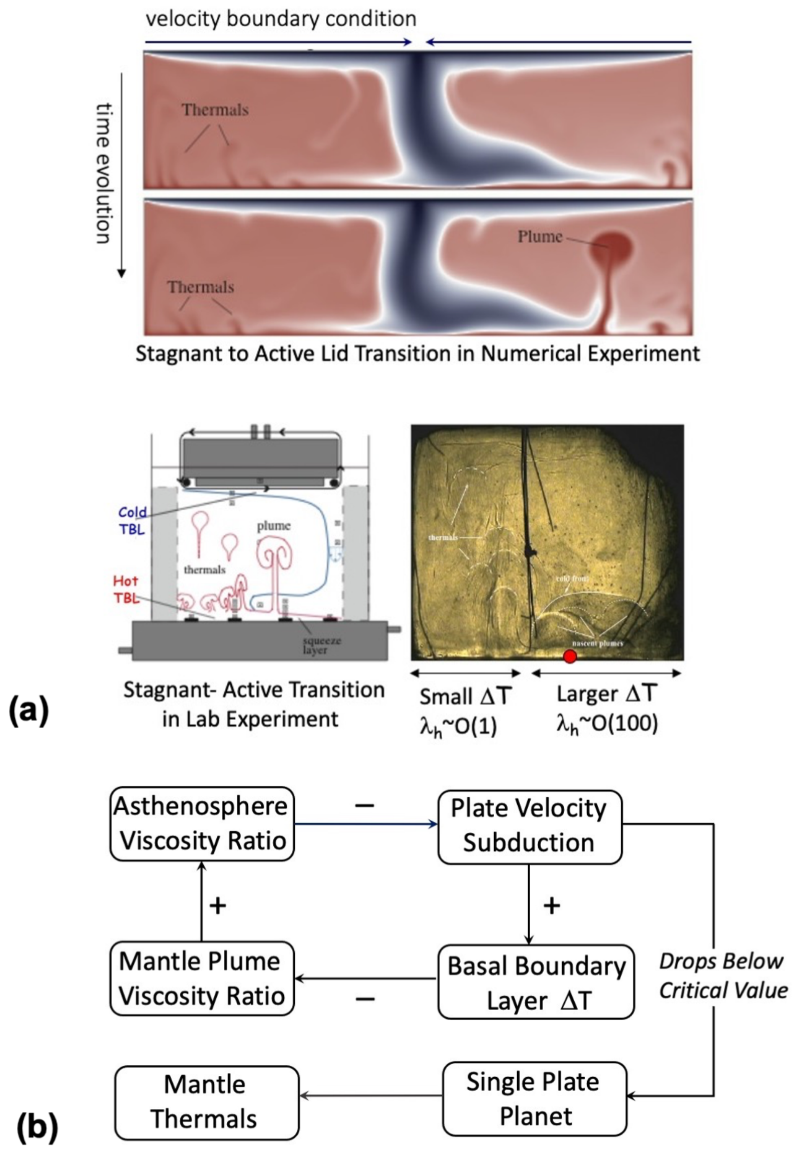
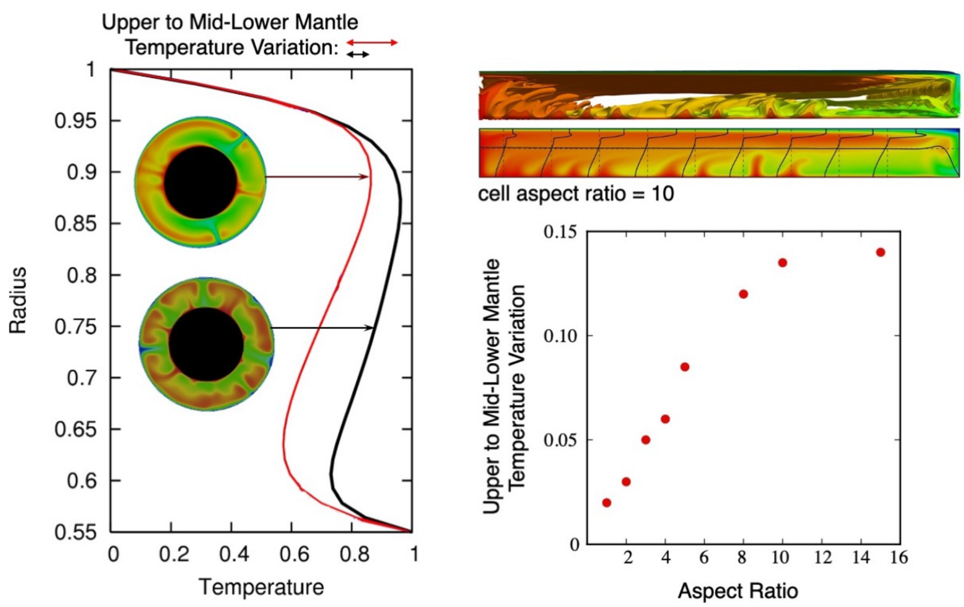

velocity boundary condition
time evolution
Thermals
Thermals
Plume
Stagnant to Active Lid Transition in Numerical Experiment
Cold TBL
Hot TBL
plume
thermals
squeeze layer
Stagnant- Active Transition in Lab Experiment
Small ΔT
λh~O(1)
Larger ΔT
λh~O(100)
(a)
Asthenosphere Viscosity Ratio
−
Plate Velocity Subduction
+
+
Mantle Plume Viscosity Ratio
−
Basal Boundary Layer ΔT
Drops Below Critical Value
Mantle Thermals
Single Plate Planet
(b)

Figure 12:



Upper to Mid-Lower Mantle
Temperature Variation:
Radius
Temperature
1
0.95
0.9
0.85
0.8
0.75
0.7
0.65
0.6
0.55
0
0.2
0.4
0.6
0.8
1
cell aspect ratio = 10
Upper to Mid-Lower Mantle
Temperature Variation
Aspect Ratio
0.15
0.10
0.05
0
2
4
6
8
10
12
14
16

Figure 13:

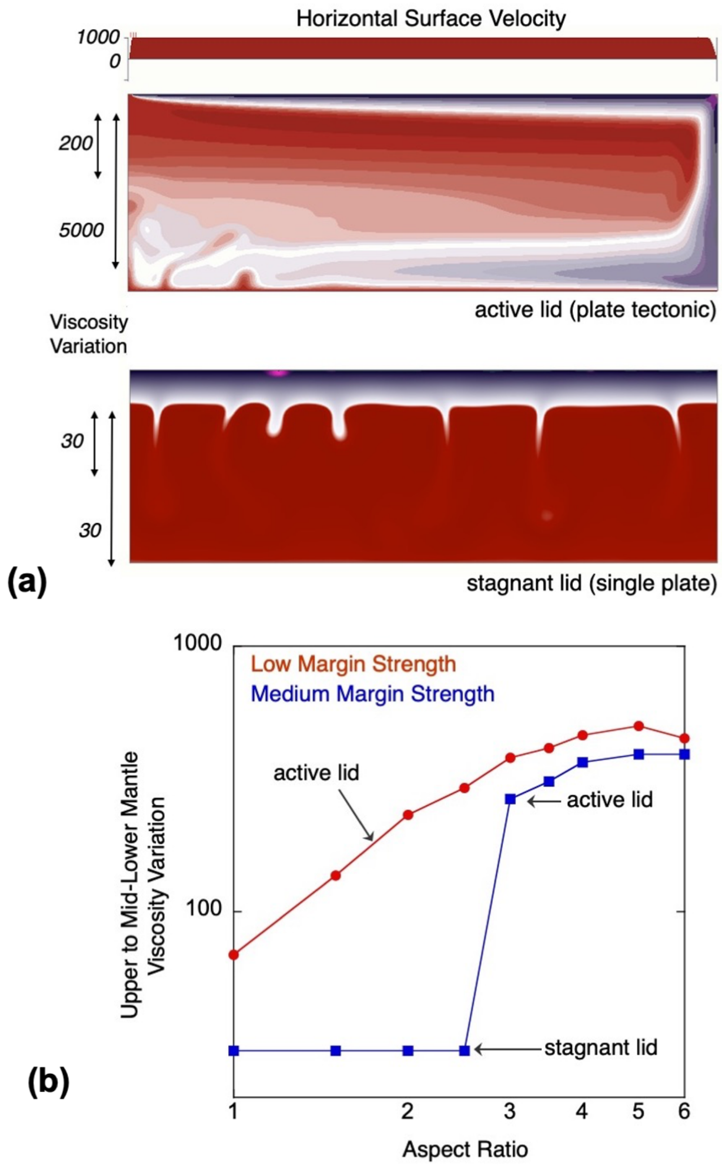
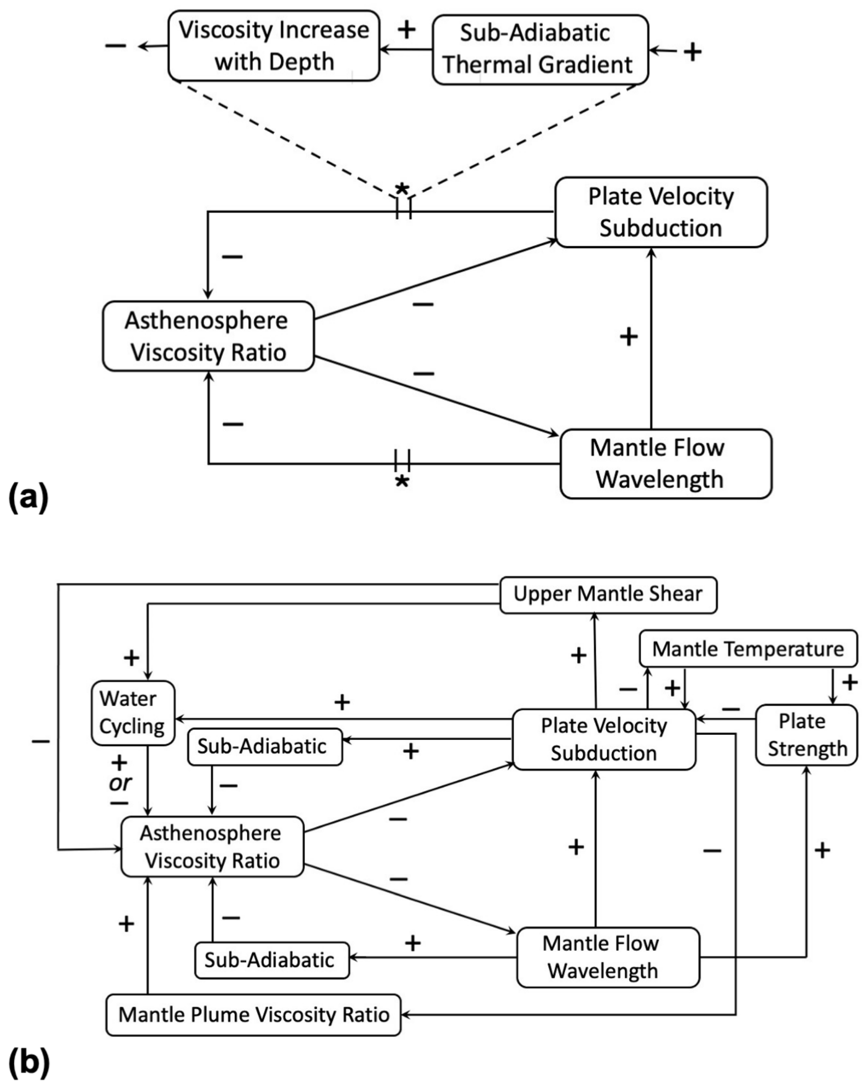

Horizontal Surface Velocity
1000
0
200
5000
active lid (plate tectonic)
Viscosity Variation
30
30
stagnant lid (single plate)
(a)
1000
Low Margin Strength
Medium Margin Strength
active lid
active lid
100
stagnant lid
Upper to Mid-Lower Mantle Viscosity Variation
(b)
1
2
3
4
5
6
Aspect Ratio

Figure 14:



Viscosity Increase with Depth
Sub-Adiabatic Thermal Gradient
Plate Velocity Subduction
Asthenosphere Viscosity Ratio
Mantle Flow Wavelength
(a)
Upper Mantle Shear
Mantle Temperature
Water Cycling
Sub-Adiabatic
Plate Velocity Subduction
Plate Strength
or
Asthenosphere Viscosity Ratio
Sub-Adiabatic
Mantle Flow Wavelength
Mantle Plume Viscosity Ratio
(b)

Figure 15:

Figure 1 (a) A diagram of the carbon cycle. (b) A schematic of how the Earth's interior is often conceptualized in Earth systems science papers and text books.

Figure 2 (a) A causal loop diagram of the Tozer feedback. Arrows labelled positive indicate a direct dependence between linked factors. Arrows labelled negative indicate an inverse dependence. The full feedback is negative and, as such, buffers/regulates mantle cooling. (b) A causal loop diagram of a positive feedback for mantle cooling. The feedback allows perturbations/fluctuations to be amplified and the effects of different initial conditions to be very long lived.

Figure 3 (a) Results from reactance time analysis for a number of thermal history models [Seales et al., 2019]. The analysis applies variable amplitude perturbations to thermal history paths and tracks perturbation decay. The slower the decay, the longer the reaction time of a model. Models with long reactions times have large structural uncertainties and the potential of being structurally unstable [Guckenheimer and Holmes, 1983]. That is, small fluctuations, that would result from physical factors not included in the models, would have long lived effects on model prediction (e.g., the effects of planetary impacts over the first billion years of Earth history could effect model predictions for the Earth's present day thermal conditions and, given the stochastic nature of impact histories, this leads to large model structural uncertainties). (b) A diagram of the qualitative difference between thermal history models that allow for self-regulation, via negative planetary feedbacks, versus models that do not.

Figure 4 (a) A simplified feedback loop diagram that highlights how coupled thermal and deep-water cycling can influence mantle cooling and the mantle Urey ratio. (b) A full feedback loop diagram of the coupled thermal and deep-water cycling system. The new symbol that appears between the surface water and regassing boxes is a limiter. If an the value of an element at the base of a limiter drops to zero then the element at the head of the limiter symbol, at the filled circle, can not be operative. (c) An loop diagram that isolates the water cycle component of the full thermal and deep-water cycling system. The existence of two feedback loops allows the ratio of mantle regassing to degassing to vary over system evolution, which feeds into the evolution of a mantle Urey ratio (Figure 4a).

Figure 5 (a) Mantle cooling trajectories consistent with petrological constraints. (b) Successful Ur-β parameter space colored by relative point density with higher values meaning the density of successful models is larger. (c) Evolution of mantle water content and (d) mantle viscosity from successful models, shown as distributions about their median values. The dark gray highlights values falling between the upper and lower quartiles and the lighter gray constraining the maximal and minimal limits.

Figure 6 Evolution of the mantle Rayleigh number (a) and a homologous temperature (b) from the models of Figure 5.

Figure 7 (left) A slice through a 3-D spherical mantle convection model with a non-Newtonian upper mantle. Velocity arrows are plotted over the model thermal field. (right) A depth profile of velocity from a full 3-D model. The profile is from the central region of a model plate and shows velocity arrows over a color plot of mantle viscosities. A low viscosity upper mantle results from mantle shear, driven by surface plate motion, together with a non-Newtonian rheology. The plots show that upper mantle flow channelizes into the low-viscosity region below plates. Convection remains of whole mantle type but upper mantle velocity is significantly greater than lower mantle velocity.

Figure 8 (a) A feedback loop diagram of a non-Newtonian upper mantle viscosity feedback process that links tectonic plate velocities, depth-variable mantle viscosity, and the wavelength of mantle flow. The right two loops are both positive feedbacks that can increase mantle flow wavelength towards a high-end limit that will be set by the geometric extent of the mantle and/or by the distribution of continents . The left loop is a negative feedback that can limit mantle flow wavelength via a feedback on plate velocity. (b) A loop diagram showing how a non-Newtonian upper mantle viscosity feedback can act as an amplifier for a mantle water-cycling feedback.

Figure 9 (a) Thermal field from a numerical mantle convection experiment driven by a combination of internal and basal heating. (b) Velocity results from a suite of numerical experiments of the type shown in Figure 9a. The experiments show that surface velocities and rms interior velocities can decrease with increased internal heating, i.e., with an increase in the internal heating Rayleigh number (Ra_i).

Figure 10 (a) A feedback loop diagram of boundary layer interactions in a convecting layer driven by internal and basal heating. (b) A feedback loop diagram of mantle plume and tectonic plate interactions.

Figure 11 Cartoons depicting conceptual models of how mantle plumes can potentially generate and maintain the Earth's asthenosphere.

Figure 12 (a) Numerical (upper) and lab tank (lower) experiments showing how plate subduction can alter the morphology of mantle plumes. (b) A feedback loop diagram of an internal mantle feedback that links tectonic plates, mantle plumes, and an asthenosphere. The plume viscosity ratio is mantle plume viscosity divided by background mantle viscosity. The asthenosphere viscosity ratio is asthenosphere viscosity divided by background mantle viscosity. Both viscosity ratios are assumed to always be less than or equal to unity.

Figure 13 (a) Temperature profiles from numerical experiments, along with temperature slices from the experiments shown as inset images. One experiment assumes an isoviscous mantle and leads to relative short wavelength convection and a mildly sub-adiabatic mantle. The other impose a factor of thirty viscosity increase from the upper to the lower mantle and leads to longer wavelength flow and a steeper sub-adiabatic mantle thermal gradient. (b) Thermal field from a numerical convection experiment (top) and a plot of temperature variations from the upper to the mid lower mantle as a function of cell wavelength from a suite of experiments (bottom).

Figure 14 (a) Thermal fields for numerical experiments with a six order of magnitude, temperature-dependent viscosity variation from the hottest to the coldest system temperature and a rheology that allows for the formation of weak plate margins. The top experiment had a yield condition that allowed for weak plate margins and associated plate like behavior, as reflected in the surface velocity plot. The bottom experiment had a higher yield condition and plate margins did not form. (b) Internal viscosity variations versus convective cell aspect ratio from two experimental suites.

Figure 15 (a) A feedback loop diagram that links plate subduction, a sub-adiabatic mantle, increasing mantle viscosity with depth, and mantle flow wavelength. (b) A causal loop diagram that links all of the feedbacks discussed in this chapter.