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Earth System Dynamics Discussions

Assessing life's effects on the interior dynamics of planet Earth using non-equilibrium thermodynamics

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Abstract

Vernadsky described life as *the* geologic force, while Lovelock noted the role of life in driving the Earth's atmospheric composition to a unique state of thermodynamic disequilibrium. Here, we use these notions in conjunction with thermodynamics to quantify biotic activity as a driving force for geologic processes. Specifically, we explore the hypothesis that biologically-mediated processes operating on the surface of the Earth, such as the biotic enhancement of weathering of continental crust, affect interior processes such as mantle convection and have therefore shaped the evolution of the whole Earth system beyond its surface and atmosphere. We set up three simple models of mantle convection, oceanic crust recycling and continental crust recycling. We describe these models in terms of non-equilibrium thermodynamics in which the generation and dissipation of gradients is central to driving their dynamics and that such dynamics can be affected by their boundary conditions. We use these models to quantify the maximum power that is involved in these processes. The assumption that these processes, given a set of boundary conditions, operate at maximum levels of generation and dissipation of free energy lead to reasonable predictions of core temperature, seafloor spreading rates, and continental crust thickness. With a set of sensitivity simulations we then show how these models interact through the boundary conditions at the mantle-crust and oceanic-continental crust interfaces. These simulations hence support our hypothesis that the depletion of continental crust at the land surface can affect rates of oceanic crust recycling and mantle convection deep within the Earth's interior. We situate this hypothesis within a broader assessment of surface-interior interactions by setting up a work budget of the Earth's interior to compare the maximum power estimates that drive interior processes to the power that is associated with biotic activity. We estimate that the maximum power involved in mantle convection is 12 TW, oceanic crust cycling is 28 TW, and continental uplift is less than 1 TW. By directly utilizing the low entropy nature of solar radiation, photosynthesis generates 215 TW of chemical free energy. This high power associated with life results from the fact that

photochemistry is not limited by the low energy that is available from the heating gradients that drive geophysical processes in the interior. We conclude that by utilizing only a small fraction of the generated free chemical energy for geochemical transformations at the surface, life has the potential to substantially affect interior processes, and so the whole Earth system. Consequently, when understanding Earth system processes we may need to adopt a dynamical model schema in which previously fixed boundary conditions become components of a co-evolutionary system.

1 Introduction

> To heat are also due the vast movements which take place on the Earth. It causes the agitation of the atmosphere... Even Earthquakes and eruptions are the result of heat.

Sadi Carnot, *Reflections on the motive power of heat*, Paris 1842.

From an early stage in the developments of thermodynamics, the Earth was characterised as a planetary heat engine. While it was not known that a proportion of this heat is not fossil heat left from the formation of the Earth, but rather is produced via radiogenic decay within the crust and mantle, the thermodynamic conception of the Earth as expressed by Carnot can be seen as surviving essentially intact to the current age (Backus, 1975). In order for a heat engine to perform work, a temperature gradient must exist. A substantial temperature gradient exists between the core and surface of the Earth as estimates for the temperature within the centre of the Earth range from 4500–5700 K (Anderson, 1989; Alfč et al., 2002) whereas the average surface temperature of the Earth is around 280 K. The work this heat engine produces is observed in processes of mantle convection, plate tectonics, continental uplift, earthquakes and vulcanism.

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Other heating gradients result from the uneven heating by absorption of solar radiation at the Earth's surface and drive processes such as the atmospheric circulation. Photosynthetic life plays a special role in that it does not follow the physical heat engine concept directly but uses photochemistry to directly utilize the low entropy nature of the incident solar radiation. The free energy generated by photosynthetic life can then be put to work in altering the Earth and act as the driving force for geology, as described in Vernadsky (1926). For example, the evolution of oxygenic photosynthesis resulted in high partial pressures of highly reactive molecular oxygen that would, without continual replenishment, oxidise to much lower levels (Goldblatt et al., 2006). These biologically-mediated effects can be understood as mechanisms that maintain the Earth system in particular thermodynamic states. Such observations can be seen as the starting point for the Gaia hypothesis which proposed that the Earth and its biota form a co-evolving, self-regulating system that is robust to perturbations (Lovelock, 1979). The initial Gaia hypothesis has developed with different studies considering it from different aspects such as evolution and natural selection (Lenton, 1998), theoretical ecology (von Bloh et al., 1997), dynamical systems (Lenton and van Oijen M., 2002) and thermodynamics (Kleidon, 2004). It is now accepted that the possession of an atmosphere far from thermodynamic equilibrium is a sign of widespread life on Earth. While this is not a sufficient indicator of life on other planets as abiotic processes are capable of producing atmospheric disequilibrium that could be detected from Earth, e.g. the abiotic flux of methane combined with an oxygen flux from the photodissociation of water (Schwartzman and Volk, 2004), the notion of biosignatures for extraterrestrial life is predicated on an appreciation of the planet-altering capabilities of life (Lovelock, 1965), (Grenfell et al., 2010).

In this paper we estimate the extent to which life can affect geologic processes through its effects on the weathering and erosion of continental crust. Our starting assumption is that life affects the intensity of erosion, which affects the rate of continental uplift and thereby the thickness of continental crust. This, in turn, affects oceanic crust recycling and the associated transport of heat from the upper mantle to the Earth's

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surface because altering the thickness of continental crust is analogous to altering the coverage of an insulating blanket over the Earth; continental crust is less dense and thicker than oceanic crust and so has greater resistance. Consequently, altering the thickness of continental crust will affect the heat transport from the upper mantle which should in turn affect the rate of mantle convection which is the dominant process for the transport of heat within the Earth's interior.

These results lead us to conclude that when considering the evolution of the Earth system it would behove us to consider it as a co-evolving system comprised of different interacting components and that, depending on the particulars of the study, it may be necessary to treat certain forces and fluxes as being dependent on other forces and fluxes. This may include the outgassing rate of buried and primordial carbon dioxide. Therefore, we propose that our simple modelling results indicates a new mechanism whereby surface life can alter the dynamics of the long-term carbonate-silicate cycle (Urey, 1952).

1.1 Structure of the paper

The major processes and the resulting interactions that affect the boundary conditions of the three processes of mantle convection, oceanic crust cycling, and continental crust dynamics are summarized in Sect. 2. In this section we describe these processes and interactions using non-equilibrium thermodynamics as it allows us to evaluate the rates at which processes perform work, move and transform material, and deplete driving forces on fundamental grounds. The basics of non-equilibrium thermodynamics are reviewed, as well as the proposed principle of Maximum Entropy Production (MEP) which states that sufficiently complex non-equilibrium thermodynamic systems are characterized by a state in which the rate of thermodynamic entropy production is maximized (Ozawa et al., 2003; Martyushev and Seleznev, 2006; Kleidon and Lorenz, 2005; Kleidon, 2009; Dyke and Kleidon, 2010). The MEP principle is controversial and not widely accepted as an organizational principle, we use it here to derive upper estimates of rates of work that can be extracted from heating gradients (which is ap-

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proximately proportional to entropy production). These quantifications will later allow us to compare and evaluate biotic activity to the power in geologic processes and in doing so quantitatively assess the effects of life on geological processes. In Sect. 3 a set of 3 models are formulated that describe the coupled nature of the mantle – oceanic crust – continental crust system. We will use these 3 models to quantify the maximum work that these processes can perform. We show that the proposed MEP principle provides reasonable estimates for certain characteristics of these processes, such as core temperature, seafloor spreading rates and continental crust thickness. In Sect. 4 we summarize our estimates of maximum power of interior processes in the form of a work budget for the rock cycle and compare it to the power associated with biotic activity. We then explore the sensitivity of the three models to the boundary conditions in order to test the hypothesis that surface life can affect the dynamics of mantle convection via altering rates of weathering and oceanic crust recycling. The paper concludes in Sect. 5.

2 Background on surface-interior interactions

Our perspective of surface-interior interactions is summarised in the diagram shown in Fig. 1. This diagram is a conceptual schematic that identifies and associates dissipative processes on the surface and within the interior of the Earth. Its primary purpose is to help convey how mechanisms that dissipate energy gradients deep inside the Earth can affect surface mechanisms which may, in turn, affect interior mechanisms. At the centre is radiogenic and fossil heat within the mantle and core. This heat flows through the Earth and radiates out into space. In doing so, energy gradients are established and dissipated. The boundary conditions for this system are flexible and can be altered by surface processes. Causation can be circular and what emerges is a co-evolving, non-equilibrium, thermodynamic system in which surface-interior interaction are critical to the overall state of the system.

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In the following sections we will first describe the primary mechanisms shown in Fig. 1 in more detail, then briefly review previous works on biotic effects on surface-interior effects.

2.1 Geological dynamics

The solid lines in Fig. 1 describe the driving forces for the primary geophysical processes of mantle convection and crust recycling.

2.1.1 Mantle convection

Radiogenic heating produced by the decay of isotopes of uranium, thorium and potasium, along with fossil heat left over from the formation of the Earth over 4.5 billion years ago, produces temperature gradients within the mantle (Schubert et al., 2001). These gradients are responsible for the changes in the density of mantle material which leads to long-term mantle convection. Although the mantle is composed of solid rock, over geological timescales density changes result in slow, creeping movement which leads to the bulk transport of material and the convection of heat from the hot interior towards the cool surface (Holmes, 1945). Large diapirs within the mantle may initiate at the mantle-outer core boundary and travel to the mantle-lithosphere boundary. Such core-mantle processes are important not only for the thermal evolution of the Earth but also the initiation and maintenance of the Earth's magnetic field (Stevenson et al., 1983) via the effects on the molten outer core.

2.1.2 Oceanic crust recycling

New oceanic crust is being continually formed at mid oceanic ridges as part of the process of seafloor spreading (Dietz, 1961). Through cracks in the sea floor, hot oceanic crust material rises up from the underlying asthenosphere. Although oceanic crust moves away from ridges at speeds of up to 13 $cm^{-1}\,y^{-1}$ in the Pacific, the overall rate of change of the width of the oceanic plates is much less. The total mass in an oceanic

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plate is balanced by the production of new oceanic crust and the reentering into the mantle of older crust at subduction points. As oceanic crust cools, it increases in density until it founders on the softer underlying asthenosphere. Eventually the crust becomes sufficiently dense for it so re-enter the mantle via the process of subduction (Hess, 1965). The combination of ridge push as new material pushes away older crust from the mid oceanic ridge and slab pull as decending subducted oceanic crust pulls crust still on the surface leads to the long-term recycling of oceanic crust material (Condie, 2003).

2.1.3 Continental crust dynamics

While oceanic crust is never older than 100 million years, rocks within the continental crust may be over 3 billion years old. The continents are mainly comprised of the suite of granitic rocks which are less dense than oceanic crust or mantle rocks. Consequently they resist subduction by "floating" above the mantle on the asthenosphere much the same way as a boat floats on water. The distance between the top of a boat and the waterline will alter as the mass of the boat alters. Loading boxes onto a boat will displace more water and the boat sits lower in the water. Similarly, altering the mass of a column of continental crust will alter how high the crust sits within the mantle. The principle of isostacy describes the change in the position of continental crust within the asthenosphere as the mass of continental crust at particular places changes.

The exact mechanisms responsible for the formation of continental crust are disputed. For example it has been proposed that the continental crust was primarily formed by silicic magma during the Archean (Brown, 1977). More complex mechanisms that involve re-melting and high temperature metamorphism (in the presence of water) have also been proposed, e.g. Kay and Kay (1988). These theories require the recycling of lithosphere as it is the subduction processes which involves high pressure and hydration of maffic rocks that can lead to the partial melting and fractionating processes that lead to the production of granitic rocks.

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Once formed, continental crust is subject to physical attack in the form of wind, water and freezing/thawing erosion and chemical attack in the form of weathering whereby carbon dioxide in the atmosphere forms carboxylic acid which dissolves minerals. The Urey reaction (Urey, 1952) describes these reactions and is the chemical mechanism of the carbonate-silicate cycle. Eroded and weathered rock moves from higher to lower ground via the action of water; boulders, rocks and stones are washed down slope and dissolved minerals move through groundwater, streams and rives. The end of this journey comes when rocks and minerals enter the sea where they eventually settle out to form sediments on the sea floor. This results in the return of carbon to the mantle that was previously outgassed primarily in the form of carbon dioxide. The removal of material from the continents affects the overall mass of continental crust. It is estimated that some 14 000 million tons of continental crust is eroded away each year (Syvitski et al., 2005). This reduction in mass leads to the uplift of continental crust due to isostastic processes.

2.2 Biotic effects on geological processes

We can begin to assess Vernadsky's claim that life is the geological force by noting that by capturing just a fraction of the free energy of the incident 340 $W m^{-2}$ of solar energy, the biosphere can contribute significant amounts of free energy to geochemical cycling. However, limiting factors such as nutrient availability strongly determines the rate at which life can grow and reproduce. Eroded and ultimately dissolved rock provides nutrients required for biological organisms. Many life forms are limited in growth by the availability of phosphorous which is a component of DNA and can only enter the biosphere via the processes of erosion (Elser et al., 2007). Consequently rates of geochemical cycling can be seen as a limiting factor for the abundance of life on Earth with the cycling ratio – the number of times a molecule is consumed and excreted by different organisms before being lost from the biosphere – giving a good indication of how strongly limited life is with respect to that molecule (Volk, 1998). However, life is able to alter the flux of nutrients into the biosphere via a range of chemical attacks

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(Taylor et al., 2009). It has been proposed that life can enhance weathering and erosion rates by a number of direct and indirect mechanisms (Lovelock and Watson, 1982; Schwartman and Volk, 1989; Schwartzman, 1999; Lenton and Watson, 2004; Dietrich and Perron, 2006), while Arens and Kleidon (2008) investigated the effects of the biota on global weathering rates. Such effects may be primarily mediated via the climate which has been shown to affect tectonic evolution of mountains (Whipple, 2009). It has been argued that the effects of life on surface geological processes are so profound that a new discipline of evolutionary geomorphology (Corenblit and Steiger, 2009) has been proposed. One example of the potentially profound geological influence of life is that the rate of formation of continental crust may have significantly increased due to the effects of life and the disequilibrium produced in the Earth system by the evolution of photosynthesis (Rosing et al., 2006).

A fascinating question to consider is to what extent would the surface and the interior of the Earth be like in the complete absence of life. We make no direct comparisons between the actual biotic Earth and a proposed abiotic Earth in that respect. Rather, we first propose and quantitatively assess a number of ways in which life has affected and continues to affect the Earth. In the absence of a complete dynamical understanding of how the Earth system has evolved and with the current limitation of having available only one planet with widespread life to examine, we must resist temptations to generalise from particulars in the absence of an understanding of the principles and mechanisms involved. We return to this issue in Sect. 4. For now, we claim that it is neither sufficient to simply subtract out the effects of life in order to arrive at a hypothetical abiotic Earth, nor to assume that abiotic processes would perform the same rates of weathering and erosion with a fixed rate of outgassing from the interior.

2.3 Surface-interior interactions

The dashed lines in Fig. 1 describe the driving forces from the surface to the interior that may be significantly affected by surface life. These can be considered within a broader context of surface-interior interactions. The “top down tectonics” approach in

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Anderson (2001) is particularly relevant in that it regards the crust of the Earth as a component in a complex system in which the dynamics of the crust may be controlling interior processes rather than the crust being a passive component that only responds to interior processes. The previous section briefly reviewed a number of different ways in which life affects other components of the Earth system. Biotic activity has a significant affect on the rates of biogeochemical cycling via the increase in weathering and increase in the formation of sedimentary rocks and even on the distribution and overall mass of granitic rock and continental crust. Continental crust can be seen as an insulating lid that lays over the mantle convection system. Results returned by experiments with Benard convection cells found that the proportional coverage of insulating lid can strongly affects the dynamics of convection in the fluid below (Jellinek and Lenardic, 2009). Consequently, processes that alter the total mass and coverage of continental crust may also be affecting mantle convection and processes occurring at the core-mantle boundary.

3 Models and maximum estimates

To compare the strength of different interior processes to life, we examine the physical power involved in maintaining these processes from three simple models using thermodynamics and the proposed principle of Maximum Entropy Production. The models then allow us to derive the sensitivity of the different components to their respective boundary conditions. This allows us to test the hypothesis that surface processes can affect the strength of interior processes.

We first provide a brief review of thermodynamics and its relation to physical work and entropy production in non-equilibrium settings. We then provide an overview of the three models with some general definitions and conventions used in the mathematical formulations. This is followed by three subsections detailing the three models. In order to estimate maximum strengths of the processes, the formulations of the models require a consistent entropy balance in addition to the typical energy and/or mass

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balances. This entropy balance in steady state quantifies the rates of entropy production necessary for the application of MEP. For testing surface-interior interactions it is also important to formulate the boundary conditions at the mantle – oceanic crust and crust – continents interface in a flexible way. The use of MEP to estimate maximum rates of entropy production allows us to produce estimates for the maximum amount of work that a process can perform. Whatever the actual entropy produced is, it must (if the model has been constructed correctly) be lower than this maximum amount. This produces vital bounds for the parameterisation of the models.

3.1 Thermodynamics, entropy production and work

The laws of thermodynamics relate energy, heat and work. The ability to perform work is essential to move and transform mass within the Earth system. When we deal with systems that are continuously heated and cooled, such as the Earth's interior, we deal with non-equilibrium thermodynamic systems that are maintained away from a state of thermodynamic equilibrium. Processes within such systems can then continuously perform work by depleting gradients. In doing so, these processes produce entropy, following the natural direction given by the second law of thermodynamics. Here, we provide a very brief introduction to non-equilibrium thermodynamics that captures these statements in simple mathematical expressions. More detailed treatments of non-equilibrium thermodynamics and how it applies to Earth systems can be found in textbooks such as Kondepudi and Prigogine (1998) and review articles such as Kleidon (2009).

3.1.1 Maximum work

We start with considering the maximum work that can be extracted from the heat flow within a system that is between a hot and cold reservoir of temperatures T_h and T_c. If no change of internal energy occurs within the system ($dU = 0$), then the first law tells us that $dU = 0 = dQ - dW$, that is, the maximum amount of work dW is constrained by

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the addition of heat dQ. The second law tells us that the entropy of the system S can at best stay constant, $dS \geq 0$. Through the heat flux from the hot reservoir, entropy is imported into the system in the amount of $dS_{\mathrm{in}} = dQ/T_{\mathrm{h}}$, while entropy is exported by the flux to the cold reservoir, $dS_{\mathrm{out}} = (dQ - dW)/T_{\mathrm{c}}$. The entropy balance of our system is hence given by:

$$dS = \sigma dt + dS_{\mathrm{in}} - dS_{\mathrm{out}} \tag{1}$$

where σ is the entropy produced by irreversible processes within the system, and $dS_{\mathrm{in}} - dS_{\mathrm{out}}$ is the net exchange of entropy with the surroundings. In the best case, the entropy of the system does not increase ($dS = 0$) and no irreversible processes take place within the system, $\sigma = 0$. We get the maximum amount of work dW when $dS_{\mathrm{in}} = dS_{\mathrm{out}}$, which yields:

$$dW = T_{\mathrm{c}} \cdot dQ \cdot \left(\frac{1}{T_{\mathrm{c}}} - \frac{1}{T_{\mathrm{h}}} \right) \tag{2}$$

The corresponding work per unit time that can be extracted, or the extracted power P, is given by:

$$P = \frac{dW}{dt} = J \cdot \left(1 - \frac{T_{\mathrm{c}}}{T_{\mathrm{h}}} \right) = J \cdot \eta_{\max} \tag{3}$$

where $J = dQ/dt$ is the heat flux from the hot to cold reservoir and $\eta_{\max} = (1 - T_{\mathrm{c}}/T_{\mathrm{h}})$ is the well known Carnot efficiency of a heat engine. At steady state the power extracted is balanced by the dissipation D occurring in the system ($P = D$), resulting in entropy production σ. In this case, the entropy balance is reformulated from Eq. (1) to:

$$\frac{dS}{dt} = \sigma + \mathrm{NEE} \tag{4}$$

where NEE is the net entropy exchange associated with exchange fluxes of heat and/or mass with the surroundings. In steady state ($dS/dt = 0$), the entropy production as-

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sociated with the dissipation of the extracted power is balanced by the net entropy exchange:

$$\sigma = -\mathrm{NEE} = \frac{D}{T_\mathrm{c}} = J \cdot \left(\frac{1}{T_\mathrm{c}} - \frac{1}{T_\mathrm{h}}\right) \tag{5}$$

Hence, the extracted power from a gradient, the rate at which work is being performed, the dissipation and the resulting entropy production are tightly linked quantities.

Entropy production can also be calculated for fluxes of mass. To derive an expression for this, we consider a steady state in which there is no change in the free energy dF:

$$dF = 0 = T dS^{(\mathrm{m})} + \mu dN \tag{6}$$

where T is the temperature of the system (assumed to be constant), $dS^{(\mathrm{m})}$ is the change of entropy associated with mass redistribution dN between a gradient represented by a chemical potential μ. When mass is removed from a higher potential μ_h at a rate $J_\mathrm{m} = dN/dt$ and added to a location with a lower chemical potential μ_l, then this results in a net change in entropy $dS^{(\mathrm{m})}/dt$. Since we consider an isolated system in steady state, this change in entropy corresponds to the entropy production $\sigma^{(\mathrm{m})}$ associated with the mass transport J_m:

$$\sigma^{(\mathrm{m})} = \frac{dS^{(\mathrm{m})}}{dt} = J_\mathrm{m} \cdot \frac{\mu_\mathrm{h} - \mu_\mathrm{l}}{T} \tag{7}$$

Since we will deal only with gradients in potential energy in the following, we will refer to μ as the geopotential.

3.1.2 Maximum entropy production

For natural Earth systems, it is often the case that irreversible processes compete within the system to dissipate energy gradients. For example, heat is transported in the mantle by diffusion and mantle convection. The extraction of work to drive mantle convection competes with diffusion, which also depletes the temperature gradient

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that drives convection. Hence, the maximum power that can be extracted from the gradient is generally lower, resulting in a lower efficiency $\eta < \eta_{max}$. In this context, the proposed principle of Maximum Entropy Production (MEP) suggests that processes adapt to states at which the rate of entropy production is maximized. Several examples have demonstrated the feasibility of the MEP principle. For example, the prediction of poleward heat transport on Earth and other planetary atmospheres from simple considerations (Paltridge, 1975; Lorenz et al., 2001) and rates of mantle convection within the Earth (Lorenz, 2002) are two sets of examples where convective processes compete with diffusive and radiative processes.

To illustrate this example, let us write the heat flux as the sum of a conductive and convective heat flux: $J = J_c + J_v$. We assume a fixed heat flux J, and that the conductive flux can be expressed as a linear function of the temperature gradient, i.e. $J_c = k \cdot (T_h - T_c)$, where k is the material's conductivity and that the boundary conditions (i.e. the temperature gradient $\Delta T = T_h - T_c$) react to some extent to the value of flux J_v. With this we get for the entropy production by the convective flux σ_v:

$$\sigma_v = J_v \cdot \left(\frac{1}{T_c} - \frac{1}{T_h} \right) = \frac{J \Delta T - k \cdot \Delta T^2}{T_h T_c} \tag{8}$$

That is, σ_v is a quadratic function of ΔT, and since ΔT is some function of J_v, there is an optimum value of J_v that maximizes σ_v. The MEP principle applied to this example states that convection adopts this optimum flux $J_{v,opt}$ that maximizes σ_v. The associated maximum rate of work done and dissipation by the convective flux is then given in steady state, as above, by $P_{max} = D_{max} = T_c \cdot \sigma_{v,max}$.

3.2 Overview of the models

Our three models are set up to correspond to three thermodynamic subsystems that exchange heat at their boundaries. We neglect exchanges of mass for simplicity. The boundaries are illustrated in the conceptual diagram of the rock cycle shown in Fig. 3.

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In the mathematical formulation of the models, we use the naming convention for parameters and variables as shown Table 1. The indices used to identify variables in the different subsystems is given in Table 2. An overview of all variables used in the following is given in Table 3.

3.3 Model 1: mantle convection

Figure 5 represents the components of mantle convection. In this model we are concerned with capturing the dynamics of the flux of heat from the base of the mantle to the bottom of the lithosphere. For the purposes of this model we assume a uniform rate of heat production via the decay of radioactive elements within the mantle and latent heat produced by the freezing of the liquid outer core. Our results and analysis still apply if the mantle is instead subject to greater heat input from the core/mantle boundary and continental crust is modelled with higher concentrations of radiogenic elements. Also, while the conductivity of mantle material will vary as temperature varies, such changes in conductivity are sufficiently small to be ignored so that conductivity can be fixed for the range of temperatures under consideration. The production of entropy via mantle convection is conceptually the same as the simple system shown in Fig. 2. Reservoir 1 is the outer core, reservoir 2 is the lithospere. Heat is transported via conduction and convection within the mantle. While laboratory experiments can provide estimates for the rate of conduction through mantle rock, determining the rate of convection can be problematic. This is because the mantle over geological timescales behaves like a liquid with temperature dependent viscosity; the hotter it is, the more vigourous it will convect. As there are no direct measurements for the temperature of the mantle and as the temperature of the mantle will determine the rate of mantle convection we are posed with a system which defies analysis. As in previously cited studies on the application of the MEP principle to the planetary atmospheres, (Paltridge, 1975), (Lorenz et al., 2001), we are faced with a situation in which there are more unknowns than equations.

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(Lorenz, 2002) found that selecting the mantle convective heat flux so that entropy production was maximised, gave a temperature structure within the Earth that was consistent with other models and theory. The approach employed was essentially equivalent to two-box energy balance climate models that have been used to solve rates of latitudinal heat flux within the Earth's and other planet's climates. In this section we develop the Lorenz model with extensions into spherical geometry and analytical solutions for rates of heat convection and entropy production. We will show that plausible predictions for the temperature structure of the interior of the Earth are produced when it is assumed that the mantle convection system is in a steady state of maximum entropy production. In doing so, we will begin to demonstrate how interior processes can be affected by boundary conditions on the surface of the Earth.

3.3.1 Energy balance

We begin with the assumption of steady state, so heat production within the interior of the Earth is balanced by heat that is emitted at the surface of the Earth. Radiogenic heat is produced within the mantle and at higher concentrations in the continental crust. We assume that heat produced within the continental crust conducts away into space and so do not include this heat input into the upper part of the mantle. The remaining heat within the mantle is fossil heat left over from the formation of the Earth and latent heat delivered to the base of the mantle from outer core freezing. Over geological time the fossil and latent heat inputs will decrease. We assume steady state and that a proportion of the heat delivered from the mantle to the bottom of the lithosphere is equal to the fossil heat plus radiogenic heat produced within the mantle. Therefore the conductive heat flux through the mantle, $J_{m,c}$ and convective heat flux througth the mantle, $J_{m,v}$ equals the amount of fossil and radiogenically produced heat in the mantle, Φ_h.

$$\Phi_h = \nabla J_h = \nabla (J_{mc} + J_{mv}) \tag{9}$$

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We express the two heat fluxes in terms of the temperature gradient ∇T and the Nusselt number, N:

$$J_{\mathrm{m,c}} + J_{\mathrm{m,v}} = -k_m N \nabla T \tag{10}$$

where k_m is the conductivity, and $N = (k_m + c_m)/k_m$ is the Nusselt number and c_m an eddy conductivity characterizing the convective heat flux produced by mantle convection. Equations (9) and (10) and energy conservation $\Phi_{\mathrm{h}} = \nabla \cdot J_{\mathrm{h}}$ together in spherical coordinates yield the following heat conduction differential equation:

$$\Phi_{\mathrm{h}} = -\frac{2k_m N}{r}\frac{\partial T_{\mathrm{m}}}{\partial r} - k_m N \frac{\partial^2 T_{\mathrm{m}}}{\partial r^2} \tag{11}$$

The analytical solution of the diffusion equation in steady state ($\partial T_{\mathrm{m}}/\partial t = 0$) is given by:

$$T_{\mathrm{m}}(r) = T_{\mathrm{core}} - \frac{\Phi_{\mathrm{h}}}{6k_m N} r^2 \tag{12}$$

with the convective heat flux $J_{\mathrm{m,v}}$ given by:

$$J_{\mathrm{m,v}} = -k_m (N-1)\nabla T = \frac{\Phi_{\mathrm{h}} r (N-1)}{3N} \tag{13}$$

We now have an expression for temperature within the mantle as a function of the Nusselt number which in turn is a function of mantle convection. By altering the rate of mantle convection, we are able to produce different temperature structures within the Earth. In the following sections we will calculate rates of entropy production via mantle convection and then find that value of mantle convection that produces maximum rates of entropy production.

3.3.2 Entropy balance

We consider two mechanisms for entropy production within the mantle: conductive and convective heat flux. Calculating entropy produced via conductive heat flux should

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straightforward as rates of conduction will be an immediate result of the particular properties of the mantle (if we make the first order assumption that conduction does not vary with varying temperature). Convective heat flux and its associated entropy production is more challenging because rates of convection will vary with varying temperature and as neither the temperature nor rate of convection is known, the problem is poorly defined. Application of the MEP allow us to make predictions for rates of convection by assuming it is that rate which produces maximum entropy. Entropy production for the mantle system is:

$$\frac{dS_{\mathrm{m}}}{dt} = \mathrm{NEE}_{\mathrm{m}} + \sigma_{\mathrm{m}} \tag{14}$$

where S_{m} is the entropy of the mantle, σ_{m} is the total entropy production within the mantle, and $\mathrm{NEE}_{\mathrm{m}}$ is the net entropy exchange of the mantle to its surroundings. At steady state, $\sigma_{\mathrm{m}} = -\mathrm{NEE}_{\mathrm{m}}$. Entropy is exchanged with the surroundings by the heating rate h (entropy import) and by the export of entropy by the heat fluxes across the mantle-crust boundary. The entropy export is the heat flux out of the surface divided by the surface temperature. $J_{\mathrm{s}}A_{\mathrm{s}}/T_{\mathrm{s}}$. The calculation of the entropy import is not trivial because the temperature at which heat is added to the system is not constant. Consequently it is necessary to integrate over the whole interior, and the entropy flux into the system is: $\int_V h/T\,dV$. This leads to the formulation for entropy production in steady state as:

$$\sigma_{\mathrm{m}} = \frac{J_{\mathrm{s}}A_{\mathrm{s}}}{T_{\mathrm{s}}} - \int_V \frac{\Phi_{\mathrm{h}}}{T} dV \tag{15}$$

By definition of the Nuesselt number, the contribution of entropy production just by mantle convection is given by $(N-1)/N \cdot \sigma_{\mathrm{m}}$.

3.3.3 Maximum entropy production due to mantle convection

It is possible to formulate entropy production within the mantle as a function of mantle convection with Eq. (15). Figure 5 shows entropy production as a function of Nusselt number. When the Nusselt number ≈ 7.5 the greatest rates of entropy are produced. This equates to mantle conduction of $\approx 3\,\mathrm{WK}^{-1}$. and convection of $\approx 21\,\mathrm{WK}^{-1}$. When these values are used in Eq. (12) a temperature structure of the internal Earth can be constructed as shown in Fig. 5.

To calculate the maximum amount of work that can be extracted from mantle convection we multiply the entropy production by the upper mantle temperature:

$$P = \sigma T_{\mathrm{mo}} = 0.013\,\mathrm{TW\,K}^{-1} \cdot 980\,\mathrm{K} \approx 12\,\mathrm{TW} \tag{16}$$

Therefore, 12 TW is the maximum amount of work that can be performed by the mantle convection system.

3.4 Model 2: oceanic crust cycling

The processes of mantle convection and conduction delivers an amount of heat to the base of the lithosphere which finds its way to the surface and then radiates out into space. In model 2 we consider how the recycling of oceanic crust transports a proportion of this heat from mantle to surface. Continental crust is rigid and its thermal properties reasonably well known, so it is relatively straightforward to calculate rates of heat flux through the surface of continental crust as a function of upper mantle temperature. Oceanic crust transfers heat both via conduction and also via the bulk transport of heat as hot mantle material from the asthenosphere rises to the surface at mid oceanic ridges. The production of mid oceanic basalt (MORB) and its eventual subduction back into the mantle releases a significant proportion of heat from the interior. This process is conceptually similar to mantle convection in that an eddy convection process will transport a certain amount of heat given a certain temperature gradient. We will show in following sections that the rate of oceanic crust recycling has a significant affect on the temperature of the asthenosphere and so mantle convection.

Figure 6 is a schematic representation of oceanic crust cycling. To parametrize the heat flux through the oceanic crust we use the so called "half space cooling model" (Sclater et al., 1980; Stuwe, 2002). Hot MORB cools in contact with the cold ocean

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water. As new material is produced from mid oceanic ridges previously extruded material is pushed away from the ridge. Consequently, the distance from the ridge, the temperature and the time on the surface for oceanic crust are correlated.

3.4.1 Energy balance

We start with the heat balance between oceanic and continental crust. We assume that the heat flux through the surface of the Earth equals the heat flux from continental and oceanic crust.

$$J_{\mathrm{h}}(r_e) = J_{\mathrm{cc}} + J_{\mathrm{oc}} \tag{17}$$

The total heat flux through continental crust is a linear function of the temperature difference, volume and thermal properties of continental crust

$$J_{\mathrm{cc}} = f_{\mathrm{c}} k_{\mathrm{c}} \frac{T_{\mathrm{mc}} - T_{\mathrm{ca}}}{\Delta z_{\mathrm{c}}} \tag{18}$$

Heat transport through oceanic crust is modelled as heat diffusion

$$\frac{\partial T}{\partial t} = \kappa \frac{\partial^2 T}{\partial z^2} \tag{19}$$

where κ is the heat diffusivity of oceanic crust which is the ratio of its density and heat capacity. We ignore any horizontal diffusion of heat though the crust, so the time-dependent temperature profile, $T(z,t)$, is entirely determined by the vertical heat diffusion. We assume that the temperature of newly formed oceanic crust is that of the upper mantle, T_{mc}, so

$$T(z,0) = T_{\mathrm{mc}} \tag{20}$$

We are able to fix boundary conditions as the temperatures of the ocean and young oceanic crust (and so upper mantle) are known.

$$T(0,t) = T_{\mathrm{oa}} \qquad T(\infty,t) = T_{\mathrm{mc}} \tag{21}$$

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The solution of the heat conduction equation to these boundary conditions is:

$$T(z,t) = T_{oa} + (T_{mc} - T_{oa}) erf\left(\frac{z}{\sqrt{4\kappa t}}\right) \tag{22}$$

Having this temperature profile, it is straightforward to calculate the surface heat flow by taking the spatial derivative of T at $z = 0$

$$q_{oc} = k_o \left(\frac{dT}{dz}\right)_{z=0} = k_o \cdot (T_{mc} - T_{oa}) \sqrt{\frac{1}{\pi\kappa t}} \tag{23}$$

Where k_o is the heat conductivity of oceanic crust. This gives a relationship between the surface heat flux and the age of the oceanic crust t. We assume that oceanic crust moves with constant velocity from the ridge to the subduction zones. This allows us to replace t with x/v_o, where v_o is the crust velocity and x is the distance from the mid-oceanic ridge. To compute the overall heat flux, Q_{oc}, we integrate over x from 0 to the breadth of the plate L_p and multiply by the length of the ridge L_r.

$$Q_{oc} = L_r \int_0^{L_p} k_o \cdot (T_{mc} - T_{oa}) \sqrt{\frac{v_o}{\pi\kappa x}} dx \tag{24}$$

where L_r is the length of the ridge and L_p the length of the plate. This leads to a heat flux density of

$$J_{oc} = \frac{2}{A} L_r \sqrt{L_p} k_o \cdot (T_{mc} - T_{oa}) \sqrt{\frac{v_o}{\pi\kappa}} \tag{25}$$

We define $r = L_p/L_r$ as the ratio of plate length to breadth and set $r = 1$ (length of oceanic plate equals width). The following equation gives heat flux through oceanic crust:

$$J_{oc} = 2\sqrt[4]{\frac{f_o^3}{Ar}} k_o \cdot (T_{mc} - T_{oa}) \sqrt{\frac{v_o}{\pi\kappa}} \tag{26}$$

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This produces values for, $v_o = 0.01\,\mathrm{m}^{-1}\,\mathrm{y}^{-1}$ (velocity of oceanic crust from ridge to subduction zone) and the heat flux through oceanic crust as $0.1\,\mathrm{W\,m^{-2}}$, we find the temperature difference between the upper mantle and surface of oceanic crust is 1500 K. This also includes the assumption that the temperatures underneath the contintal crust and the oceanic crust are equal.

3.4.2 Entropy balance

Entropy production for the oceanic crust recycling system is:

$$\frac{dS_o}{dt} = \mathrm{NEE}_o + \sigma_o \tag{27}$$

When the surface temperature, the upper mantle temperature and the heat flux is know, the entropy production is simply:

$$\sigma_o = J_{oc}\left(\frac{1}{T_{oa}} - \frac{1}{T_{mc}}\right) \tag{28}$$

3.4.3 Maximum entropy production due to oceanic crust recycling

Rather than assume that the velocity of oceanic crust is fixed, we can instead make this parameter vary via its effects on the rate of entropy production that the oceanic crust recycling system produces. For the mantle convection model, we assumed that the rate of convection would be that which produced maximum rates of entropy. We can introduce a new parameter, γ, which is analogous to a diffusion rate, having dimensions of $\mathrm{m}^{-1}\,\mathrm{y}^{-1}\,\mathrm{K}^{-1}$, and is a temperature dependent rate of oceanic crust velocity

$$v_o = \gamma(T_{oa} - T_{mo}) \tag{29}$$

Figure 8 shows how entropy production varies as γ varies. Table 4 show values for the oceanic crust recycling model when γ is set to the value that produces maximum entropy production.

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To calculate the maximum amount of work that can be done by oceanic crust recycling we multiply the entropy production by the surface temperature:

$$P = \sigma T_s = 0.097\,\mathrm{TW\,K^{-1}} \cdot 293\,\mathrm{K} \approx 28\,\mathrm{TW} \tag{30}$$

Therefore, 28 TW is the maximum amount of work that can be performed by the oceanic crust recycling system.

3.5 Model 3: uplift and erosion

As continental material is eroded away into the sea, the mass of continental crust decreases and this reduction in mass leads to mantle pressure pushing the continental crust up. Erosion and uplift are related in that higher rates of erosion will lead to higher rates of uplift, with maximum rates of uplift being determined by the material properties of the asthenosphere.

Figure 9 is a schematic representation of continental crust uplift and erosion. Model 3 characterises the process of continental crust uplift and erosion in terms of a competing processes that move material away and towards thermodynamic equilibrium. Mountains are manifestation of non-equilibrium geological processes in that they are not at isostatic equilibrium and so over time will sink back down into the asthenosphere. They also represent an energy gradient that erosion dissipates; material is moved from high above the surface of the Earth to the lower sea floor. Isostatic imbalance and erosion will both lead to a decrease in the height of any mountain. In the following sections, we will quantify theses processes in thermodynamic terms that will include the production of entropy via uplift and erosion.

3.5.1 Potential energy balance

Density differences are responsible for uplift as the density of continental crust is less than oceanic crust (which includes sediments and sedimentary rock). We ignore fluxes of heat as uplift and erosion are effectively irrelevant in determining the temperature of

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continental crust. The potential energy of continental crust material at the surface of the continents is expressed using the notion of a geopotential μ_{ca}:

$$\mu_{ca} = g \cdot z_{ca} \tag{31}$$

where of g is gravity and z_{ca} height above the mantle. The geopotential of continental crust material at the crust/mantle boundary, μ_{mc}, is given with:

$$\mu_{mc} = \frac{\rho_m - \rho_c}{\rho_c} \cdot g \cdot z_{mc} \tag{32}$$

where ρ_m and ρ_c are the densities of mantle and continental crust respectively and z_{mc} is the depth of the mantle/crust boundary.

3.5.2 Mass balance

We assume that the overall mass of continental crust is at steady state. We assume that all material that is eroded from continental crust ends up as sediments. We also assume that all sediments that are subducted end up as continental crust.

We express the total mass of material that forms the continental crust as the sum of the mass of continents, M_c, and the mass of sediments at the ocean floor, M_s. The two mass reservoirs are expressed per unit surface area in terms of a thickness or height of the reservoir, Δz_c and Δz_s, respectively, their respective densitites ρ_c and ρ_s and the fractional coverage of continents f_c and oceans f_o:

$$M_c = \rho_c \cdot \Delta z_c \cdot f_c \qquad M_s = \rho_s \cdot \Delta z_s \cdot f_o \tag{33}$$

The total thickness Δz_c of the continental crust consists of a contribution of crust above the zero line $\Delta z_{c,1}$ and the depth of the crustal root $\Delta z_{c,2}$ which is below the zero line.

The mass balance of continents per unit area is expressed as:

$$\frac{dM_c}{dt} = J_{sc}^{(m)} - J_{cs}^{(m)} \qquad \frac{dM_s}{dt} = J_{cs}^{(m)} - J_{sc}^{(m)} \tag{34}$$

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where $J_{sc}^{(m)}$ is the subduction flux from ocean sediment to crustal root and $J_{cs}^{(m)}$ is the erosional flux from elevated crust to oceanic sediment.

The subduction flux $J_{sc}^{(m)}$ is written as a function of mass per unit area of sediments $\rho_c \Delta z_s$ times the velocity v_o of the oceanic crust:

$$J_{sc}^{(m)} = \rho_c \cdot \Delta z_s \cdot v_o \cdot \sqrt{\frac{f_o}{rA}} \cdot 3 \tag{35}$$

where $r = 0.027$ is the ratio of oceanic crust length to breadth and the factor 3 reflects the fact that sediment is not distributed equally on the ocean floor but concentrated at the plate boundaries. We neglect the conversion of ρ_s to ρ_c during subduction and metamorphism.

The erosion flux $J_{cs}^{(m)}$ is written as a function of the topographic gradient between the continents and the sediments, $\Delta z_{c,1}$, where k_{cs} is an erosion parameter:

$$J_{cs}^{(m)} = k_{cs} \cdot \rho_c \cdot \Delta z_{c,1} \tag{36}$$

The movement of continental crust up and down within the asthenosphere, v_c, is determined by the more dense, displaced mantle producing a buoyancy force pushing the crust up, F_m, the resistive force of friction between the continental crust and asthenosphere, F_f, and the gravitational attraction pulling the continental crust back down, F_g.

$$\begin{aligned} z_c \cdot \rho_c \cdot \frac{dv_c}{dt} &= F_m + F_c + F_f \\ &= \rho_c \cdot \mu_{mc} - \rho_c \cdot \mu_{ca} - z_c \eta_c v_c \end{aligned} \tag{37}$$

In steady state, $dv_c/dt = 0$, the uplift velocity is:

$$v_c = \frac{\rho_c \cdot (\mu_{mc} - \mu_{ca})}{\eta_c \cdot z_c} \tag{38}$$

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3.5.3 Entropy balance

The entropy balance of the continental crust with respect to mass exchange consists of the entropy produced by irreversible processes within the system $\sigma_c^{(m)}$ and the net entropy exchange across the system boundary $\mathrm{NEE}_c^{(m)}$:

$$\frac{dS_c^{(m)}}{dt} = \mathrm{NEE}_c^{(m)} + \sigma_c^{(m)} \tag{39}$$

The entropy production $\sigma_c^{(m)}$ is associated with the depletion of the geopotential gradient associated with continental crust:

$$\sigma_c^{(m)} = J_m \cdot \frac{(\mu_{ca} - \mu_s)}{T} \tag{40}$$

3.5.4 Maximum entropy production due to continental erosion

We will show that there is a characteristic erosion rate at which entropy production by erosion is maximized. This maximization of erosional entropy production is equivalent with maximizing the uplift work or minimizing the frictional dissipation. In other words the uplift-engine would work with maximum efficiency. To calculate the entropy produced by uplift, we use Eq. (38) for the expression of v_c and replace $\Delta z_{c,1}$ and $\Delta z_{c,2}$ by the expressions of the chemical potentials μ_{mc} and μ_{ca}. Figure 10 shows numerical results for erosional entropy production as a function of the erosion parameters k_{cs}.

When the erosion parameter, k_{cs}, is set to $\approx 5\times10^{-16}\,\mathrm{s}^{-1}$, which produces maximum rates of entropy production for the erosion/uplift model. This produces an uplift velocity of 0.7 $\mathrm{mm}^{-1}\,\mathrm{y}^{-1}$ and a continental crust height above zero line of 4000 m.

We can directly calculate the amount of energy dissipated by erosional processes by $D = J_{er} \cdot g \cdot z$. Since dissipation equals work at steady state, we get an uplift work of 0.03 TW. It is important to note that this is a calculation of the power that is realized by the heat engine, while the values for work produced by the mantle convection

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and oceanic crust recycling systems are the theoretical maximum values of extracted power.

4 Discussion

In this section we first include the maximum power estimates from the three models into a work budget of the Earth's interior and the global rock cycle and compare them to biotic activity. We then show the sensitivity of the oceanic crust cycling and mantle convection models to continental crust thickness in order to substantiate our main hypothesis that biologically-mediated surface processes affect interior processes.

4.1 The work budget of interior processes and the rock cycle

We now summarize our results in the form of a work budget of the global rock cycle and interior processes. This work budget is summarized in Fig. 11. The maximum power associated with mantle convection (12 TW), oceanic crust cycling (28 TW), and continental uplift (< 1 TW) is taken from the previous three sections.

Also shown in the work budget are processes driven primarily by the climate system. For comparison we show the 900 TW of power involved in driving the global atmospheric circulation (Peixoto and Oort, 1992). This power drives the dehumidification of atmospheric vapor and therefore the hydrologic cycle. The strength of the hydrologic cycle is relevant here in that it (a) distills seawater, (b) lifts vapor into the atmosphere, and (c) transports water to land. The precipitation on land then contains chemical and potential free energy. The chemical free energy inherent in precipitation is used to chemically dissolve rocks and bring the dissolved ions to the oceans. The potential energy in precipitation at some height of the land surface generates stream power which can be used to mechanically transport sediments.

To estimate the available power to chemically weather rock by abiotic means, we consider the work necessary to desalinate the water when evaporated from the ocean

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and then the work that sea salt could perform upon mixing with seawater (Isaacs and Schmitt, 1980). The total flux of evaporation from the Eath's oceans is 361×10^{12} m^3/yr (Peixoto and Oort, 1992). Using a mean sea salt concentration of 35.153 g/kg (Schlesinger 1997), a mean molar weight of 70.629 g/mol, and an assumed vant Hoff factor of 2, this yields a power source of 27 TW generated by the evaporation of sea-water from the worlds oceans.

The desalinated water from the oceans that falls on land as precipitation is able to perform chemical work by dissolving rock-based minerals. Most of the rock minerals on land are not sodium chloride, so only a fraction of the work contained in the estimate of desalination can be used. Assuming that ocean water is in equilibrium with the composition of the crust, this work is estimated by combining the estimated value of continental runoff of 33×10^{12} m^3/yr^{-1} (Peixoto and Oort, 1992) (which in the mean equals the net moisture transport to land by the atmospheric circulation) with the above estimate of desalination, reduced by the fraction of salts other than sodium chloride in the total salt concentration of sea water of about 6%. This yields an estimate for the work by dissolution of minerals on land of 33/361·6%·27 TW=0.15 TW.

The power associated with biotic activity is derived from estimates of gross primary productivity (GPP) of 120 petagrams yr^{-1} on land and 50 petagrams yr^{-1} in the oceans (Solomon et al., 2007). If we assume all carbon is produced via photosynthesis and that all photosynthate is glucose, then 1.67×10^{15} and 6.91×10^{15} moles of glucose are produced each year on land and in the oceans respectively. One mole of glucose contains 2874 kJ. This gives an energy production for land and in the oceans of 4.79×10^{21} J yr^{-}1 and 2.11×10^{21} J yr^{-}1 respectively or 152 TW on land and 63 TW in the oceans giving a global biological power of 215 TW which is an order of magnitude greater than the power generated by mantle convection. From this energy budget, approximately 50% of land GPP and 20% of oceanic GPP will be used to drive autotrophic metabolisms, mainly via the process of respiration. The remaining energy is available to grow and to concentrate, move and transform geochemical material.

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4.2 Sensitivity of the work budget to boundary conditions

In this section we perform a series of sensitivity analyses in order to assess the ability of surface processes to alter the boundary condition of interior processes and their rates of work. Figure 12 shows the temperature of the mantle-crust boundary as a function of continental crust thickness. Increasing the thickness of continental crust, increases the insulation on the surface of the Earth and increases the temperature of the mantle-crust boundary. Figure 13 shows how the rate of work performed by the oceanic crust recycling system and mantle convection system changes as the temperature of the mantle-crust boundary changes. As the mantle-crust temperature increases, the oceanic crust recycling system produces more work. As the cold reservoir of the oceanic crust recycling does not change (the temperature of the ocean), increasing the temperature of the hot reservoir (the upper mantle temperature) will lead to a greater possible Carnot efficiency for the heat engine and so greater rates of power. The same reasoning applies to the explanation for the decrease of work performed by the mantle convection system. We assume fixed heat production rates within the mantle (the hot reservoir). Increasing the temperature of the cold reservoir (the upper mantle) decreases the possible Carnot efficiency and amount of power the system can perform.

Therefore, altering the rate of continental crust recycling by, for example, increasing rates of biologically mediated wheathering, would alter the rate of oceanic crust recyling and so mantle convection. Life, via its attack on continental crust would be altering the boundary conditions for the planetary heat engine and so the dynamics of mantle convection deep within the interior.

4.3 Implications for long-term carbon cycle

The BLAG model of Berner et al. (1983), models interactions between tectonic processes such as sea floor spreading and partial pressures of CO_2 in the Earth's atmosphere. One conclusion is that faster rates of sea floor spreading would lead to faster

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outgassing of primordial CO_2 as this is largely released from mid oceanic ridges where new oceanic crust rises to the surface. If the causal relationships between the components of the Earth system are consistent with our highly simplified modelling, then surface life, in altering the thickness of continental crust could affect sea floor spreading rates and so the release of carbon dioxide into the atmosphere. In this way, life may lead to a change in the overall input of CO_2 into the atmosphere and so steady state fluxes of carbon to and from the surface of the Earth.

Schwartman and Volk (1989) and Schwartzman (1999) showed that life can greatly increase the rate of weathering of silicate rocks. The presence of life on Earth has lead to reduced CO_2 in the atmosphere and lower surface temperatures. Consequently, if the Earth was suddenly sterilised, then the decrease in weathering would lead to an increase of CO_2 and surface temperatures. This would in turn increase the rate of weathering and a new steady state would be reached in which weathering equaled outgassing. If that were the case, then the rates of weathering on the actual Earth would be the same on a hypothetical sterile Earth. This seems to falsify the hypothesis that surface life has affected mantle convection via altering continental crust thickness because there would, at steady state, be no difference between the weathering rates on a biotic and abiotic Earth.

However, this assumes that the total outgassing of CO_2 would be the same on a biotic and abiotic Earth. Our simple models show that an important difference between a biotic and abiotic Earth is the rate of oceanic crust recycling which arise from the thermodynamic consequences of life altering the boundary conditions for the planetary heat engine. The connections are thus: less life leads to less weathering and erosion of continental crust which leads to thicker continental crust which leads to a greater heat flux through oceanic crust which leads to faster rates of oceanic crust recycling which leads to greater release of primordial CO_2. Rather than CO_2 outgassing being fixed, it is ultimately determined by the temperature gradient within the Earth and the flux of heat through the surface. If life is able to alter the boundary conditions by altering where and at what rate heat flows through the crust of the Earth, then the rate of CO_2

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entering the atmosphere is not fixed but is in part determined by the actions of life.

4.4 Limitations

The models we developed here are clearly highly simplified and necessarily leave out many details. In the following discussion we aim to identify and defend the major assumptions and indicate a path for future work in this area.

The main assumption made in all the models is that in order to produce estimates for the maximum rates of entropy production and so work and power it is not necessary to capture the properties of the materials involved. This is a reasonable assumption as long as one does not specify the nature of the work that is performed. For example, if a certain amount of work in a system involves motion, then it is necessary to incorporate certain properties of the material in order to produce plausible rates of motion out of the estimates of maximum power. This explains why the mantle convection model did not feature any information about what the mantle is composed of. Such information is not necessary when one is able to specify the thermodynamic quantities of heat generation, temperature gradient and heat flux.

This assumption becomes hard to defend when considering continental crust recycling. We assumed that all eroded continental crust moves back to the continental crust with no change in its material properties. This would require all such material to be removed from subducted sediments at the surface at acretion prism boundaries. Consequently we assumed that there is no change to the material properties of crust material as it is weathered and eroded, deposited as sediment, subducted and then joined back to the continental crust. In reality this sequence of events could involve a number of metamorphic processes that leads to a change in thermodynamic fluxes. We also assumed that the mass balance of continental crust is zero and not altered by rates of erosion and continental crust formation. Erosion only altered the thickness of continental crust, not its proportional coverage.

In the oceanic crust recycling model we ignored any changes in the material properties of oceanic crust. Metamorphic processes resulting from the subduction of oceanic

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crust are important processes for the formation of granitic melts and so new continental crust. We assumed that the only difference between oceanic crust and the upper mantle is one of density. The rate of granitic melt formation may be significantly affected by the effects of life and so this is an important mechanism whereby life can affect geological processes. We did not include this process here.

We assume that all the systems we model are in steady state. This precludes any investigation into the evolution of these systems in which, for example, the heating rate due to radioactive decay and fossil heat would decrease over time. Although we have demonstrated the sensitivity of the mantle convection system to the thickness of continental crust, we did not explore the effects of altering the overall mass of continental crust and its proportional coverage on the surface of the Earth. Both have changed over time and if our main hypothesis is correct, then both will have had a significant effect on interior processes.

For the mantle convection model we assumed whole mantle convection and that the heat flux through the core is a component of that convective system. While this is a somewhat problematic assumption for the fluid outer core it is clearly erroneous for the solid inner core. We assumed uniform heat production due to radiogenic decay throughout the mantle. This is not accurate as the continental crust contains higher proportions of the lighter radioactive elements and so the continental crust represents not only a thermal "blanket" above the asthenosphere, but a "heated blanket". Secular cooling is also not a uniform process. An important heat flux into the mantle comes from the heat delivered through the outer core from latent heat release during inner core freezing.

While we would argue that atmospheric conditions represent an important boundary condition for the geological systems we model, we did not investigate the effects of one on the other by coupling atmospheric to interior processes. The outgassing of volatile components from the Earth's interior has had a profound influence on the evolution of the Earth's atmosphere and we would argue that this represents an important boundary condition for interior processes.

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These limitations certainly impact the accuracy of our estimates of maximum power as well as their sensitivity to changes in boundary conditions. While these estimates can be improved in future refinements in the model parameterizations, the order of magnitude of the power involved in geologic and biospheric processes should still be in the correct range. Also robust should be our notion of surface-interior interaction, since the maximum power of interior processes depend on gradients that are not only shaped by geologic processes, but also by the boundary conditions that are shaped by surface processes. Hence, our central hypothesis of biologically mediated surface-interior interactions should be unaffected by these limitations.

5 Conclusions

In this paper we explored the hypothesis that biologically-mediated processes on the surface of the Earth affect interior processes such as oceanic crust recycling and mantle convection. We formulated a series of simple models in terms of thermodynamic quantities, fluxes and forces. In doing so, we were able to quantify the upper bounds for the amounts of work that these systems perform by using the principle of Maximum Entropy Production. We showed how the work produced in one system can alter the boundary conditions for the other systems and so established a causal connection between surface to interior processes. Our justification for using thermodynamics in this respect, is that this represents the most principled way of assessing the influence of one system on another and provides quantitative estimates in terms of the power involved that is needed to move and transform matter. Our results are consistent with and can be seen as a quantitative extension of the observations and theories of Vernadsky that emphasised the capacity of life to affect the entire Earth system and of Lovelock in that such effects would be manifest in planetary systems that are not at thermodynamic equilibrium.

Our study is only a first step towards quantifying the effects of life on the functioning of the whole planet (that is, the coupled atmosphere-surface-interior system) and its

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evolution. Natural next steps would be to extend our set of models into a dynamic set of models that directly interact and explicitly capture the effects of life. This would require extensions to dynamic differential equations. Also, a more detailed treatment of the consequences of the power available to biotic activity on surface processes would make biotic effects more explicit. The coupling of the models to geochemical disequilibrium within the atmosphere and with respect to the redox gradient between the atmosphere and the crust should help us to investigate the driving forces of planetary disequilibrium, which, as Lovelock noted many years ago, should be a sign of widespread life on a planet. Hence, our thermodynamic approach to understand the Earth system processes and their interactions should be a central building block towards a better understanding of the fundamental signs of widespread life on a planet.

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Table 1. Naming convention used in the model formulations.

Symbol	Property	Unit
ρ	density	$\mathrm{kg\,m^{-3}}$
k	conductivity	$\mathrm{W\,m^{-1}\,K^{-1}}$
g	gravitational acceleration	$\mathrm{m\,s^{-2}}$
η	viscosity	$\mathrm{kg\,m^{-1}\,s^{-1}}$
f_c	fractional coverage of continents	–
f_o	fractional coverage of oceans	–
F	force	$\mathrm{kg\,m^{-1}\,s^{-2}}$
P	power	W
J	heat flux	$\mathrm{W\,m^{2}}$
$J^{(m)}$	mass flux	$\mathrm{kg\,m^{2}\,s^{-1}}$
D	dissipation	W
T	temperature	K
σ	entropy production	$\mathrm{W\,K^{-1}}$
S	entropy	$\mathrm{J\,K^{-1}}$
NEE	net entropy exchange	$\mathrm{W\,K^{-1}}$

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Table 2. Convention for the use of indices to identify subsystems as shown in Fig. 3.

Index	Component
a	atmosphere
c	continental crust
s	sediments
o	oceanic crust
m	mantle

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Table 3. Characterization of system states and fluxes.

Symbol	State variable	Unit
T_{core}	core temperature	K
T_m	mantle temperature	K
T_{mc}	temperature at the mantle-crust boundary	K
T_c	temperature of continental crust	K
T_o	temperature of oceanic crust	K
T_{ca}	surface temperature of continental crust	K
T_{oa}	surface temperature of oceanic crust	K
Φ_h	heating rate due to secular cooling and radioactive decay	$W\,m^{-3}$
J_h	overall heat flux within the mantle	$W\,m^{-2}$
J_s	heat flux through surface of Earth	$W\,m^{-2}$
J_{mc}	conductive heat flux within the mantle	$W\,m^{-2}$
J_{mv}	convective heat flux within the mantle	$W\,m^{-2}$
J_{cc}	conductive heat flux through the continental crust	$W\,m^{-2}$
J_{oc}	conductive heat flux through the oceanic crust (and sediments)	$W\,m^{-2}$
J_{ov}	convective heat flux associated with oceanic crust recycling	$W\,m^{-2}$
$J^{(m)}$	mass flux	$kg\,s^{-1}$
Δz_c	thickness of continental crust	m
Δz_o	thickness of oceanic crust	m
Δz_s	thickness of sediments	m
M_c	mass of continental crust	kg
M_o	mass of oceanic crust	kg
M_s	mass in sediments	kg
v_c	vertical velocity of continents (uplift)	$m\,s^{-1}$
v_o	horizontal velocity of oceanic crust (seafloor spreading rate)	$m\,s^{-1}$
μ_{ca}	chemical potential at the surface of continental crust	$J\,kg^{-1}$
μ_{mc}	chemical potential at the mantle-continental crust interface	$J\,kg^{-1}$
μ_s	chemical potential of sediments	$J\,kg^{-1}$
N	Nusselt number	–

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Table 4. Values for oceanic crust recycling when "diffusion" parameter γ is selected to produce MEP.

Variable	Description	Value
$T_{oa} - T_{mo}$	Temp gradient mantle-oceanic crust	730 K
J_{oc}	Heat flux through oceanic crust	34 TW
J_{cc}	Heat flux through continental crust	16 TW
v_o	Oceanic crust recycling velocity	20 cm^{-1} y^{-1}

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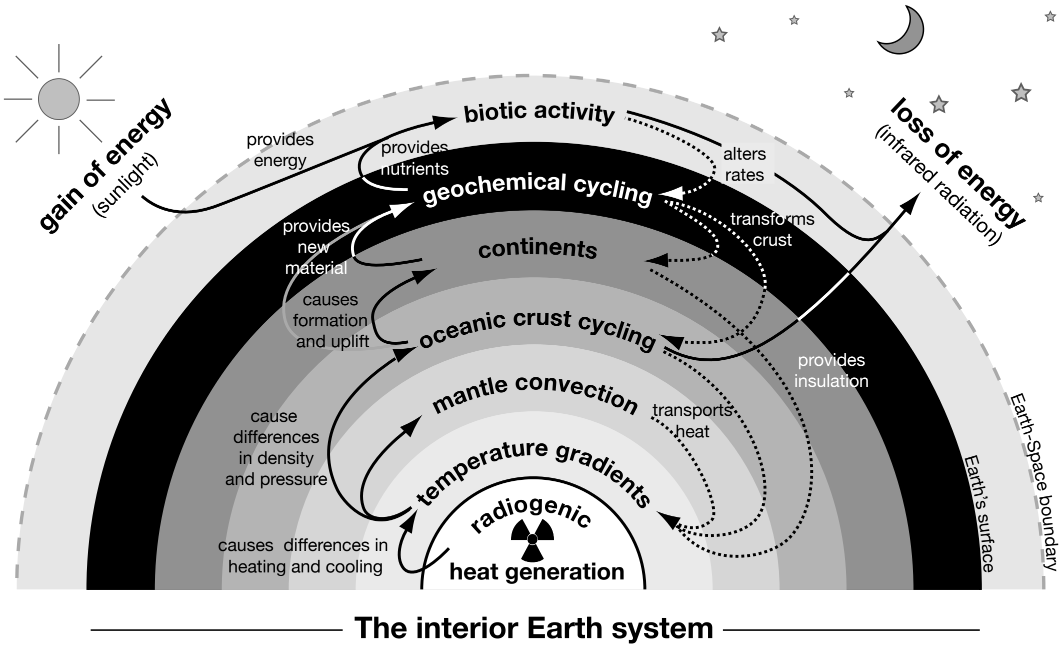

Fig. 1. A highly simplified schematic that represents the major pathways of surface-interior interactions. Interior processes can be seen as being driven by the production of radiogenic heat and fossil heat left from the formation of the Earth. These processes are represented with solid lines. However, such processes are sensitive to alterations to their boundary conditions that are represented by dashed lines.

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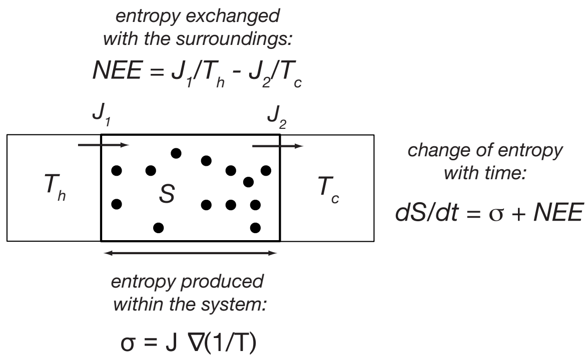

Fig. 2. The rate of change of entropy of a system over time dS/dt is a function of the entropy produced within the system σ and the entropy that is exchanged with its surroundings NEE. A heat flux J_1 from a hot reservoir at temperature T_h into the system imports entropy at a rate J_1/T_h, and the heat flux J_2 from the system to a cold reservoir at a temperature T_c exports entropy at a rate J_2/T_c. In steady state, $J_1 = J_2$ and the entropy produced within the system is balanced by the net entropy export: $\sigma = -\text{NEE}$.

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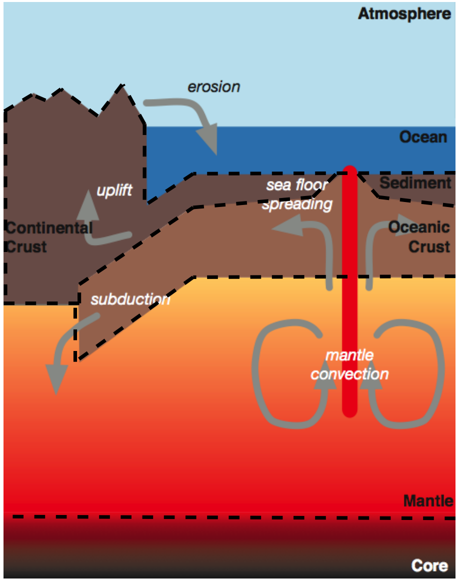

Fig. 3. The rock cycle's major components of: mantle convection, oceanic crust recycling and continental crust recycling are shown. The subsystem boundaries are delineated with dashed black lines.

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Model 1:
Mantle Convection

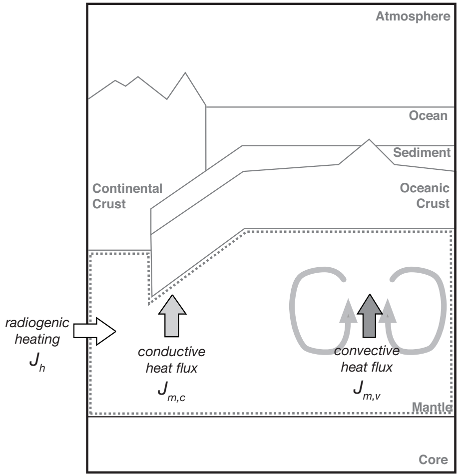

Fig. 4. Model 1: a simple model of mantle convection driven by a given flux of radiogenic heating and considering two modes of heat transport within the mantle by conduction/radiative exchange and convection.

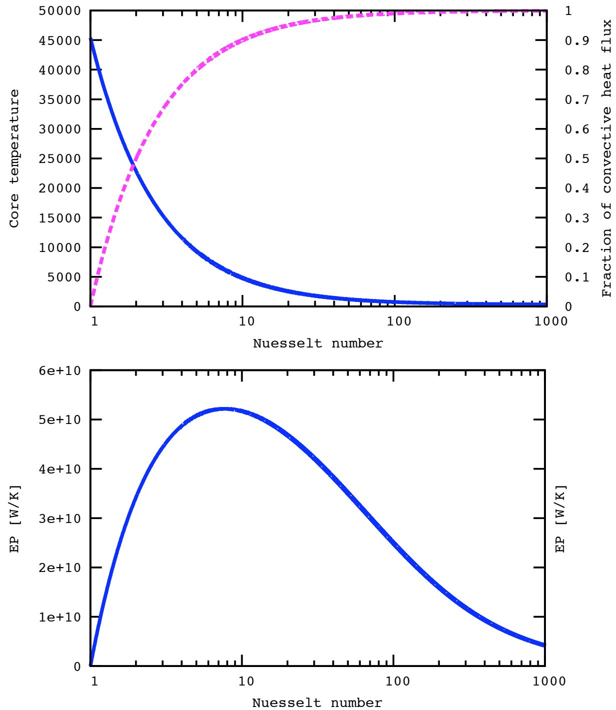

Fig. 5. Top plot: core temperature and fraction of convective heat flux with varying Nusselt number. The Nusselt number is a dimensionless value of the proportion of convection to conduction. Core temperature is plotted with a solid line (units on left vertical axis). Fraction of convective heat flux is plotted with a dashed line (units on right vertical axis). Bottom plot: entropy production via mantle convection with varying Nusselt number.

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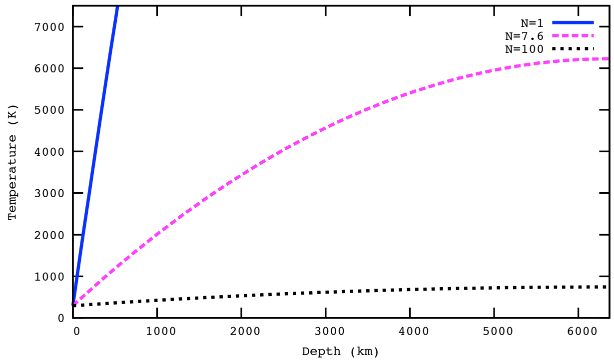

Fig. 6. MEP mantle convection temperature structure with three different Nusselt number values. Depth beneath the surface of the Earth is shown on the horizontal axis. Temperature in degrees Kelvin is shown on the vertical axis. With no mantle convection (N=1 solid line) the core temperature is >40 000 degrees Kelvin. With high rates of mantle convection ($N = 100$, dotted line) the core temperature is <1000 degrees Kelvin. When N is set to the MEP value of 7.6 (dashed centre line) the core temperature is ≈ 6000 degrees Kelvin.

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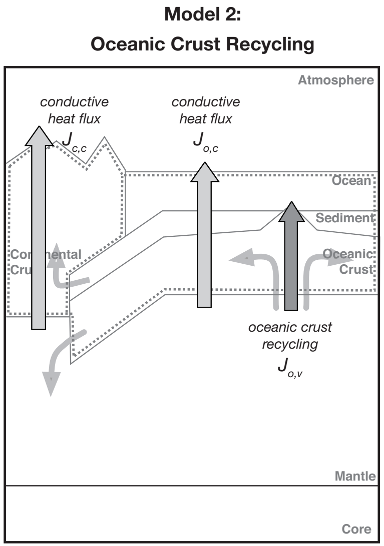

Fig. 7. Model 2: a simple model oceanic crust recycling. Heat flow from the mantle out into space via continental and oceanic crust. Heat is transferred through conductive heat flux only whereas heat flows through oceanic crust via conductive and convective processes that are generated by the bulk transport of oceanic crust material.

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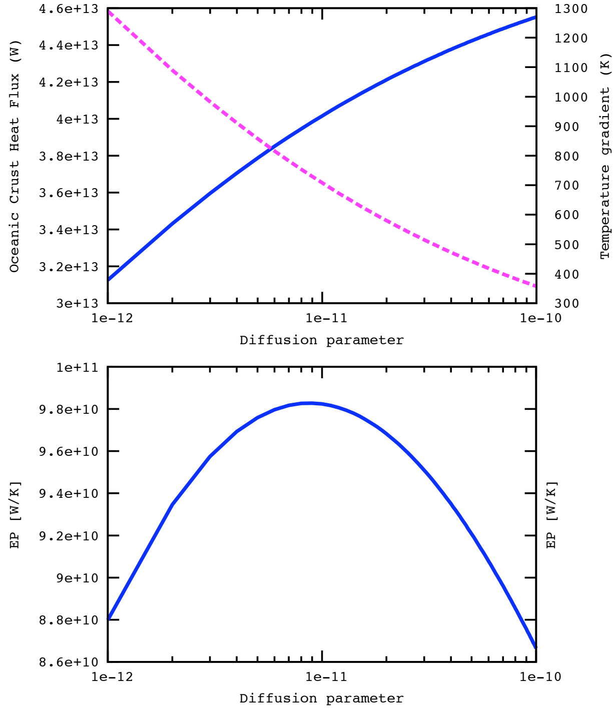

Fig. 8. Top plot: oceanic crust heat flux plotted with solid line (units on right horizontal axis) with varying values for the "Diffusion parameter", γ. Increasing γ increases the heat flux through oceanic crust. Temperature gradient, the difference in temperature between surface of oceanic crust and upper mantle, plotted with dashed line (units on left vertical axis) with varying values for γ . Increasing γ decreases the temperature gradient. Bottom plot: entropy production in oceanic crust recycling as a function of γ.

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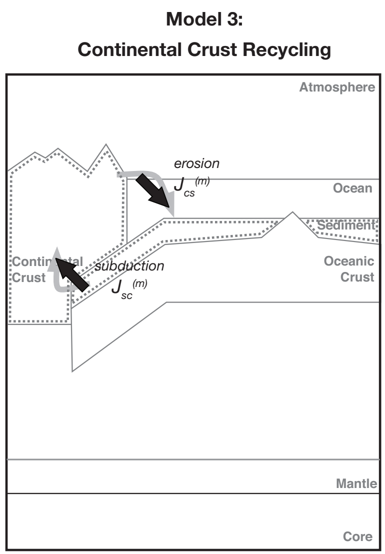

Fig. 9. Model 3: a simple model of the mass balance of continental crust driven by uplift and erosion. Weathering and erosion processes transfer continental crust material to the ocean where it is deposited as sediment. Continental crust material moves back to the continent though the process of subduction.

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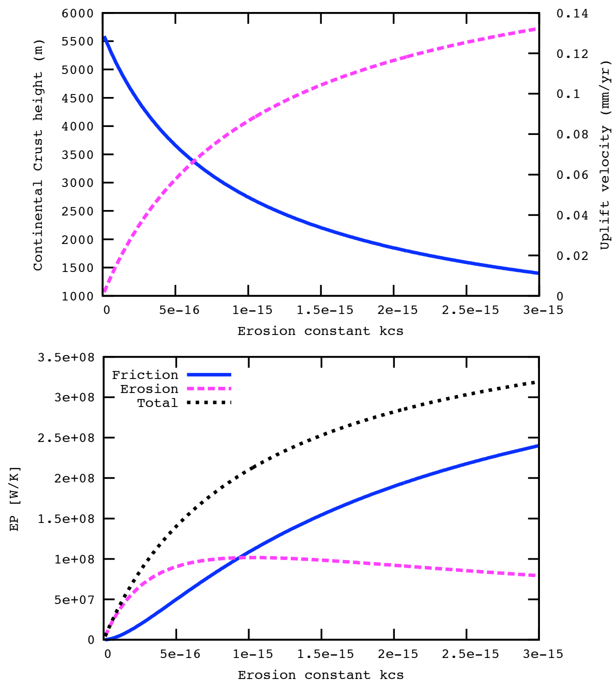

Fig. 10. Top plot: continental crust height plotted with solid line (units on left horizontal axis) and uplift velocity plotted with dashed line (units left horizontal axis) with varying values for erosion constant k_{cs}. Increasing k_{cs} decreases continental crust height and increases uplift velocity. Bottom plot: entropy production by erosion as a function of erosion rate, k_{cs}. Entropy produced by friction plotted with a solid line, erosion plotted by a dashed line and total entropy production with a dashed line (top line).

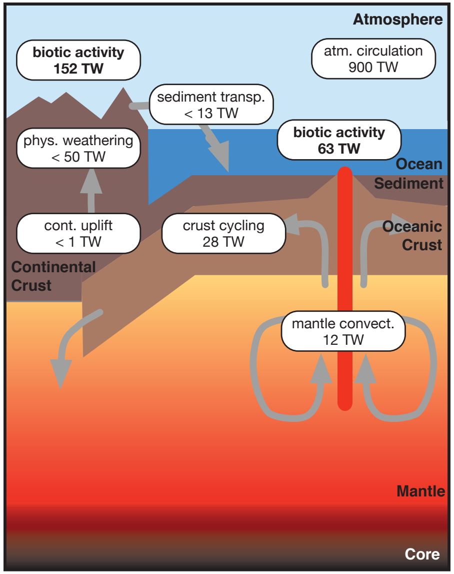

Fig. 11. Estimation of rates of work for components of the Earth system. Biotic activity contributes a total of 215 TW.

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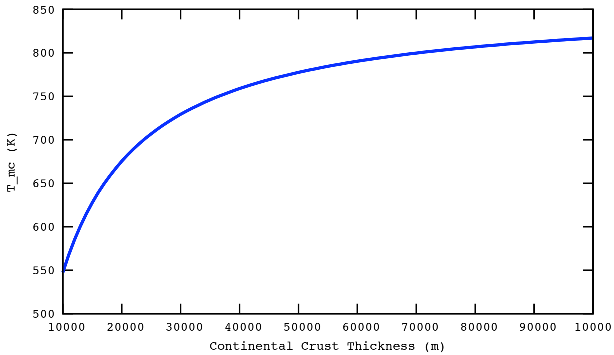

Fig. 12. Sensitivity of upper mantle temperature to continental crust thickness. The temperature at the mantle-crust boundary (T_{mc}) increases as the thickness of continental crust increases.

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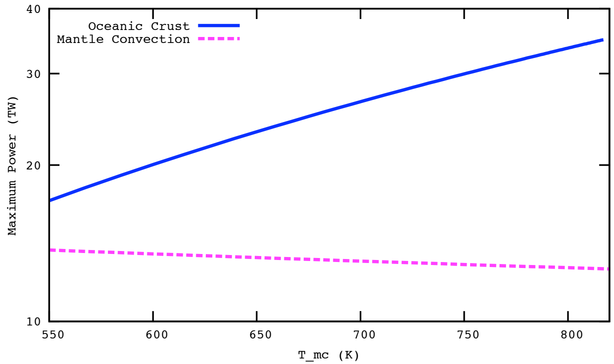

Fig. 13. Sensitivity of rates of maximum power (TW) produced in the oceanic crust (plotted with a solid line) and mantle convection (plotted with a dashed line) to varying upper mantle temperature (T_{mc}). Increasing the temperature of the upper mantle increases the maximum amount of power that can be generated in the oceanic crust system, while it decreases the maximum amount of power in the mantle convection system.

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