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Earth System Dynamics

Towards understanding how surface life can affect interior geological processes: a non-equilibrium thermodynamics approach

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Abstract. Life has significantly altered the Earth's atmosphere, oceans and crust. To what extent has it also affected interior geological processes? To address this question, three models of geological processes are formulated: mantle convection, continental crust uplift and erosion and oceanic crust recycling. These processes are characterised as non-equilibrium thermodynamic systems. Their states of disequilibrium are maintained by the power generated from the dissipation of energy from the interior of the Earth. Altering the thickness of continental crust via weathering and erosion affects the upper mantle temperature which leads to changes in rates of oceanic crust recycling and consequently rates of outgassing of carbon dioxide into the atmosphere. Estimates for the power generated by various elements in the Earth system are shown. This includes, inter alia, surface life generation of 264 TW of power, much greater than those of geological processes such as mantle convection at 12 TW. This high power results from life's ability to harvest energy directly from the sun. Life need only utilise a small fraction of the generated free chemical energy for geochemical transformations at the surface, such as affecting rates of weathering and erosion of continental rocks, in order to affect interior, geological processes. Consequently when assessing the effects of life on Earth, and potentially any planet with a significant biosphere, dynamical models may be required that better capture the coupled nature of biologically-mediated surface and interior processes.

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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 and the release of heat due to outer core freezing, 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; Alfé 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 volcanism.

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 but uses photochemistry to directly utilise the low entropy nature of the incident solar radiation. The emergence and evolution of widespread life has profoundly affected the Earth. For example, the evolution of oxygenic photosynthesis resulted in high partial pressures

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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, 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 (Kasting et al., 1997; 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).

Here, we consider a possible mechanism where the free energy generated by photosynthetic life can then be put to work in altering the Earth. Vernadsky described life as *the* geological force (Vernadsky, 1926). More recently, 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. Our starting assumption is that life affects the intensity of weathering and erosion, which affects the rate of continental uplift and thereby the thickness of continental crust. 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. We show that this, in turn, affects oceanic crust recycling and the rate of outgassing of CO_2 from mid-oceanic ridges.

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 (Walker et al., 1981).

Structure of the paper

We summarise 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 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. We review the basics of non-equilibrium thermodynamics, as well as the proposed principle of Maximum Entropy Production (MEP) which states that certain complex non-equilibrium thermodynamic systems can be successfully characterised as being in states in which the rate of thermodynamic entropy production is maximised (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 organisational principle, we use it here primarily to derive upper estimates of rates of work that can be extracted from heating gradients (which is approximately 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 we formulate a set of 3 models that describe the mantle, oceanic crust and continental crust systems. We 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, continental crust thickness and global erosion rates.

In Sect. 4 we explore the sensitivity of the three models to their boundary conditions and show that altering erosion rates of continental crust can affect upper mantle temperatures and oceanic crust recycling. We discuss this result in the context of the geological carbon cycle and propose a new feedback between biologically-mediated weathering/erosion and outgassing of CO_2. We situate the models and results within a broader assessment of power generation and dissipation within the Earth system while noting the comparatively large amounts of power produced by surface life.

We conclude the paper 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

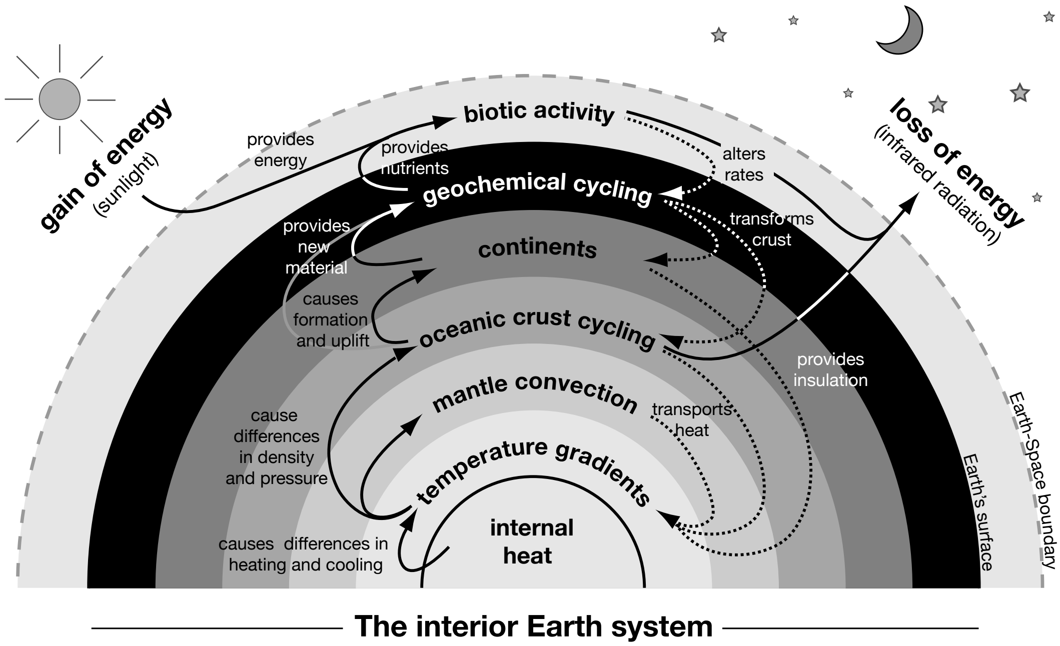

Fig. 1. A highly simplified schematic that represents the major pathways of surface-interior interactions. Interior processes can be seen as being driven by the flow of heat from the interior of the Earth. These processes are represented with solid lines. However, such processes are sensitive to alterations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

mechanisms. At the centre is heat the is left over from the formation of the Earth plus radiogenic and heat released as the outer core freezes. 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.

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 to geological interactions.

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 potassium, along with heat released by freezing of the outer core and 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 diapers 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 via the effects on the molten outer core (Stevenson et al., 1983).

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 material rises up from the underlying asthenosphere. Although oceanic crust moves away from ridges at speeds of up to 13 $\mathrm{cm}^{-1}\,\mathrm{yr}^{-1}$ in the Pacific, the overall rate of change of the width of the oceanic plates is much less. The total mass in an oceanic plate is balanced by the production of new oceanic crust and the reentering into the mantle of older crust at subduction points (Hess, 1965). The combination of ridge push as new material pushes away older crust from the mid oceanic ridge and slab pull as descending 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 mafic rocks that can lead to the partial melting and fractionating processes that lead to the production of granitic rocks.

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 1.4×10^{13} kg of continental crust is globally eroded away each year (Syvitski et al., 2005). This reduction in mass leads to the uplift of continental crust due to isostatic processes.

2.2 Biotic effects on geological processes

Vernadsky’s claim that life is *the* geological force is predicated on life’s ability to capture some of the 340 W m^{-2} that the Earth receives from the Sun. In doing so, 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 an element 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 element (Volk, 1998). However, life is able to alter the flux of nutrients into the biosphere via a range of chemical attacks (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). 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). This argument is predicated on an estimation that the energetic input by life into geological processes on Earth is three times larger than the contribution from Earth’s internal heat engine.

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 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 Bénard 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 potentially 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 examine how 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 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) and Dyke and Kleidon (2010).

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 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 $dS_{in} = dQ/T_h$, while entropy is exported by the flux to the cold reservoir, $dS_{out} = (dQ - dW)/T_c$. The entropy balance of our system is hence given by:

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

where σ is the entropy produced by irreversible processes within the system, and $dS_{in} - dS_{out}$ 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 $dS_{in} = dS_{out}$, which yields:

$$dW = T_c \cdot dQ \cdot \left(\frac{1}{T_c} - \frac{1}{T_h} \right). \tag{2}$$

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

$$P = \frac{dW}{dt} = J \cdot \left(1 - \frac{T_c}{T_h} \right) = J \cdot \eta_{max}, \tag{3}$$

where $J = dQ/dt$ is the heat flux from the hot to cold reservoir and $\eta_{max} = (1 - T_c/T_h)$ 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 associated with the dissipation of the extracted power is balanced by the net entropy exchange:

$$\sigma = -\mathrm{NEE} = \frac{D}{T_c} = J \cdot \left(\frac{1}{T_c} - \frac{1}{T_h} \right). \tag{5}$$

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.

Table 1. Naming convention used in the model formulations.

symbol	property	unit
ρ	density	$\mathrm{kg\,m^{-3}}$
A	area	$\mathrm{m^{-2}}$
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	$\mathrm{W\,K^{-1}}$

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:

$$\mathrm{d}F \;=\; 0 \;=\; T\;\mathrm{d}S^{(\mathrm{m})} \;+\; \mu\;\mathrm{d}N, \tag{6}$$

where T is the temperature of the system (assumed to be constant), d$S^{(\mathrm{m})}$ 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 $J_\mathrm{m} = \mathrm{d}N/\mathrm{d}t$ and added to a location with a lower chemical potential μ_l, then this results in a net change in entropy d$S^{(\mathrm{m})}/\mathrm{d}t$. 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:

$$\sigma^{(\mathrm{m})} \;=\; \frac{\mathrm{d}S^{(\mathrm{m})}}{\mathrm{d}t} \;=\; J_\mathrm{m}\;\cdot\;\frac{\mu_\mathrm{h}\;-\;\mu_\mathrm{l}}{T}. \tag{7}$$

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 that drives convection. Hence, the maximum power that can be extracted from the gradient is generally lower, resulting in a lower efficiency $\eta < \eta_\mathrm{max}$. 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 maximised. Several examples have demonstrated the utility of applying 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.

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

To illustrate this example, let us write the heat flux as the sum of a conductive and convective heat flux: $J = J_\mathrm{c} + J_\mathrm{v}$. 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. $J_\mathrm{c} = k\cdot(T_\mathrm{h} - T_\mathrm{c})$, where k is the material's conductivity and that the boundary conditions (i.e. the temperature gradient $\Delta T = T_\mathrm{h} - T_\mathrm{c}$) react to some extent to the value of flux J_v. With this we get the entropy production by the convective flux σ_v:

$$\sigma_\mathrm{v} \;=\; J_\mathrm{v}\;\cdot\;\left(\frac{1}{T_\mathrm{c}} - \frac{1}{T_\mathrm{h}}\right) \;=\; \frac{J\;\Delta T\;-\;k\;\cdot\;\Delta T^2}{T_\mathrm{h}\;T_\mathrm{c}}. \tag{8}$$

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 maximises σ_v. The MEP principle applied to this example states that convection adopts this optimum flux $J_\mathrm{v,opt}$ that maximises σ_v. The associated maximum rate of work done and dissipation by the convective flux is then given in steady state, as above, by $P_\mathrm{max} = D_\mathrm{max} = T_\mathrm{c}\cdot\sigma_\mathrm{v,max}$.

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.

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.

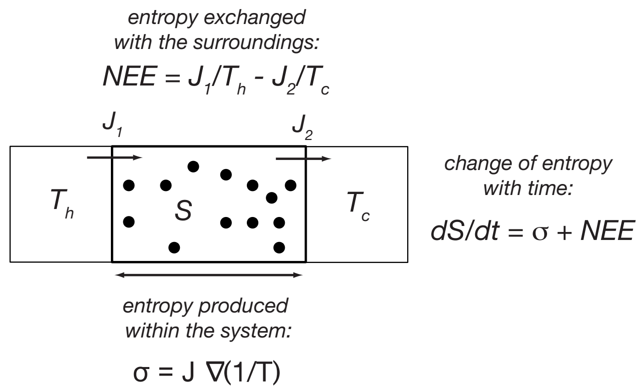

Fig. 2. The rate of change of entropy of a system over time dS/dt is a function of the entropy produced within the system σ and the entropy that is exchanged with its surroundings NEE. A heat flux J_1 from a hot reservoir at temperature T_h into the system imports entropy at a rate J_1/T_h, and the heat 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: $\sigma = -\mathrm{NEE}$.

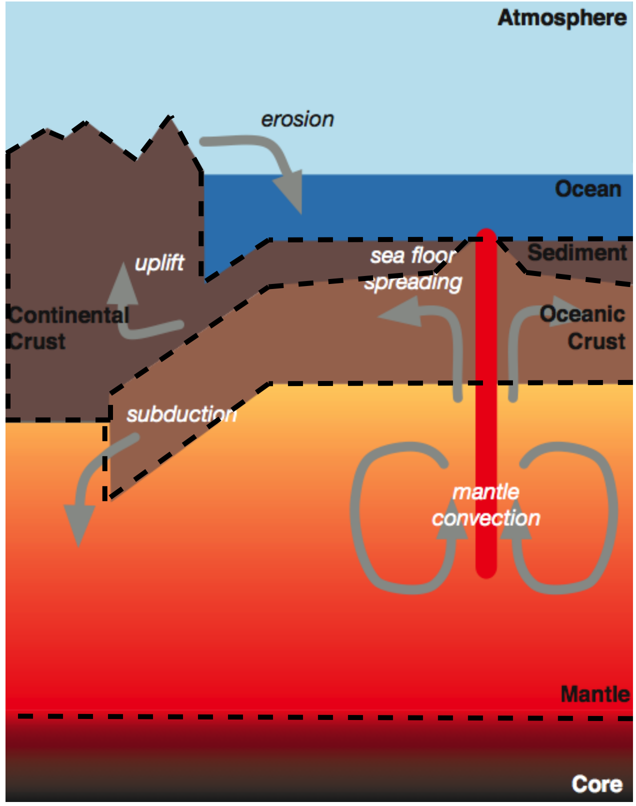

Fig. 3. The rock cycle's major components of: mantle convection, oceanic crust recycling and continental crust recycling are shown. The subsystem boundaries are delineated with dashed black lines.

3.3 Mantle convection

Figure 4 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 lithosphere. 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 viscosity of mantle material is temperature dependent; the hotter it is, the more vigorous 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.

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

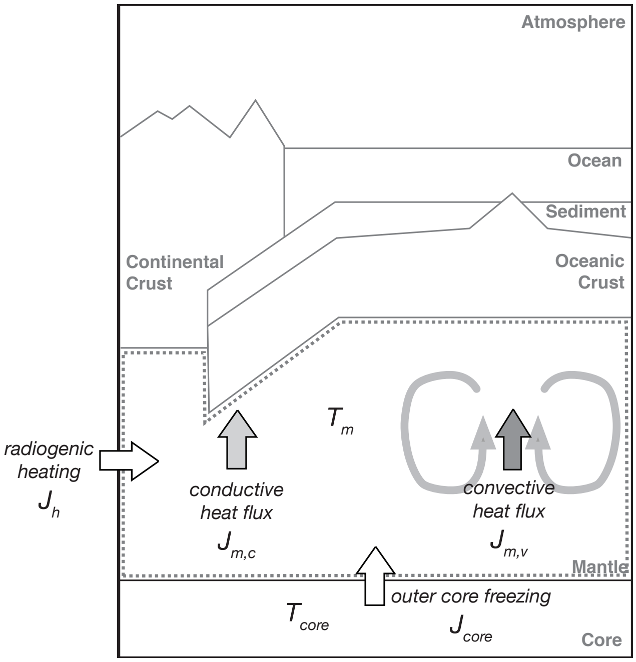

Fig. 4. Mantle convection: The mantle is heated via decay of radiogenic elements, J_h and release of heat from outer core freezing, J_{core}. This heat along with fossil heat flows to the surface via conduction, J_{mc}, and convection, J_{mv}. The temperature of the mantle and core are denoted with T_m and T_{core} respectively.

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 heat delivered to the base of the mantle from outer core freezing. We assume 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_{mc} and convective heat flux through the mantle, J_{mv} equals the amount of fossil, radiogenic and core freezing heat in the mantle, Φ_h. Consequently, the overall heat flux in the mantle, J_h, is found with:

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

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

$$J_{mc} + J_{mv} = -k_m N \nabla T, \tag{10}$$

where k_m is the conductivity, and $N = (k_m + c_m)/k_m$ is the Nusselt number which is a dimensionless ratio of conduction to convection. c_m is an eddy conductivity characterising the convective heat flux produced by mantle convection. Equations (9) and (10) and energy conservation $\Phi_h = \nabla \cdot J_h$ together in spherical coordinates yield the following heat conduction differential equation:

$$\Phi_h = -\frac{2 k_m N}{r} \frac{\partial T_m}{\partial r} - k_m N \frac{\partial^2 T_m}{\partial r^2}. \tag{11}$$

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, radioactive decay and core freezing	$W m^{-3}$
J_h	overall heat flux within the mantle	$W m^{-2}$
J_e	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	$W m^{-2}$
$J^{(m)}$	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	–

where T_m is the temperature of the mantle and $r = 6.371 \times 10^6$ m is the mean radius of the Earth. The analytical solution of the diffusion equation in steady state ($\partial T_m / \partial t = 0$) gives the temperature of the mantle at different depths, $T_m(r)$, as a function of the temperature of the core, T_{core} and the previous parameters:

$$T_m(r) = T_{core} - \frac{\Phi_h}{6 k_m N} r^2, \tag{12}$$

with the convective heat flux J_{mv} given by:

$$J_{mv} = -k_m (N - 1) \nabla T = \frac{\Phi_h r (N - 1)}{3 N}. \tag{13}$$

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.

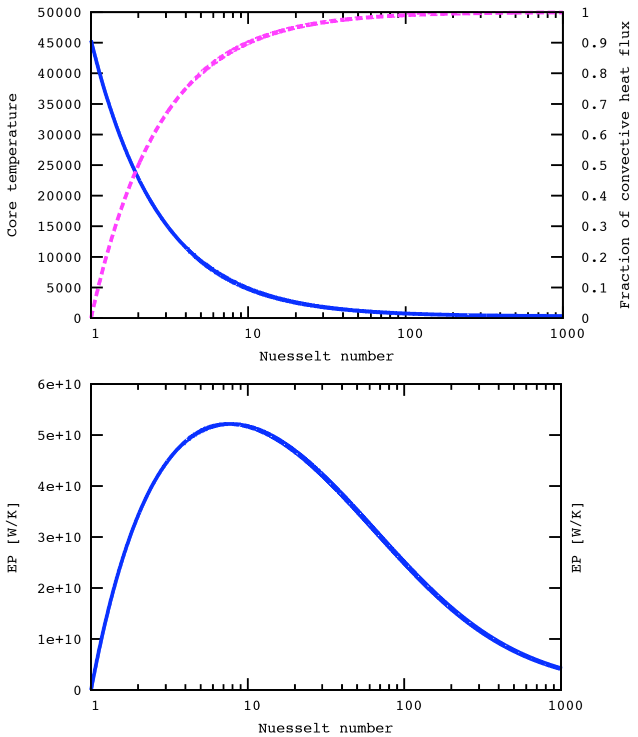

Fig. 5. Top plot: core temperature and fraction of convective heat flux with varying Nusselt number. The Nusselt number is a dimensionless value of the proportion of convection to conduction. Core temperature is plotted with a solid line (units on left vertical axis). Fraction of convective heat flux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

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 is 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:

$$\frac{\mathrm{d}S_\mathrm{m}}{\mathrm{d}t} = \mathrm{NEE_m} + \sigma_\mathrm{m}, \tag{14}$$

where S_m is the entropy of the mantle, σ_m is the total entropy production within the mantle, and $\mathrm{NEE_m}$ is the net entropy exchange of the mantle to its surroundings. At steady state, $\sigma_\mathrm{m} = -\mathrm{NEE_m}$. Entropy is exchanged with the surroundings by the heating rate Φ_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. $J_\mathrm{e} A_\mathrm{e}/T_\mathrm{s}$, where $A_\mathrm{e} = 5.1 \times 10^{11}$ m is the total surface of the Earth. 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: $\int_\mathrm{V} \Phi_\mathrm{h}/T\,\mathrm{d}V$, where $V = 4/3\pi r^3$ is the volume of the Earth. This leads to the formulation for entropy production in the mantle at steady state as:

$$\sigma_\mathrm{m} = \frac{J_\mathrm{e}\, A_\mathrm{e}}{T_\mathrm{s}} - \int_\mathrm{V} \frac{\Phi_\mathrm{h}}{T}\mathrm{d}V. \tag{15}$$

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

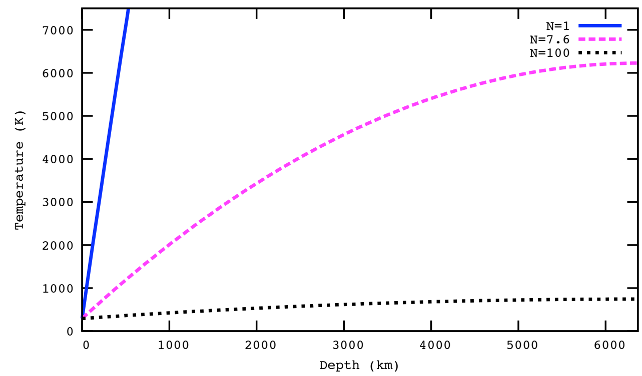

Fig. 6. MEP mantle convection temperature structure with three different Nusselt number values. Depth beneath the surface of the Earth is shown on the horizontal axis. Temperature in degrees Kelvin is shown on the vertical axis. With no mantle convection ($N = 1$ solid line) the core temperature is $>$40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

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 ≈3 W K^{-1} and convection of ≈21 W K^{-1}. When these values are used in Eq. (12) a temperature structure of the internal Earth can be constructed as shown in Fig. 6.

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

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

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

3.4 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 upper mantle 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 upper mantle and so mantle convection. We will also show that there are feedbacks both ways between these systems in that the upper mantle temperature can affect rates of oceanic crust recycling

Figure 7 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” (Stuwe, 2002). Hot MORB cools in contact with the cold ocean 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 crust, J_{cc}, and oceanic crust, J_{oc}.

$$J_{\mathrm{h}}(r_{\mathrm{e}}) = J_{\mathrm{cc}} + J_{\mathrm{oc}}. \qquad (17)$$

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

$$J_{\mathrm{cc}} = f_{\mathrm{c}}\, k_{\mathrm{c}}\, \frac{T_{\mathrm{mc}} - T_{\mathrm{ca}}}{\Delta z_{\mathrm{c}}}, \qquad (18)$$

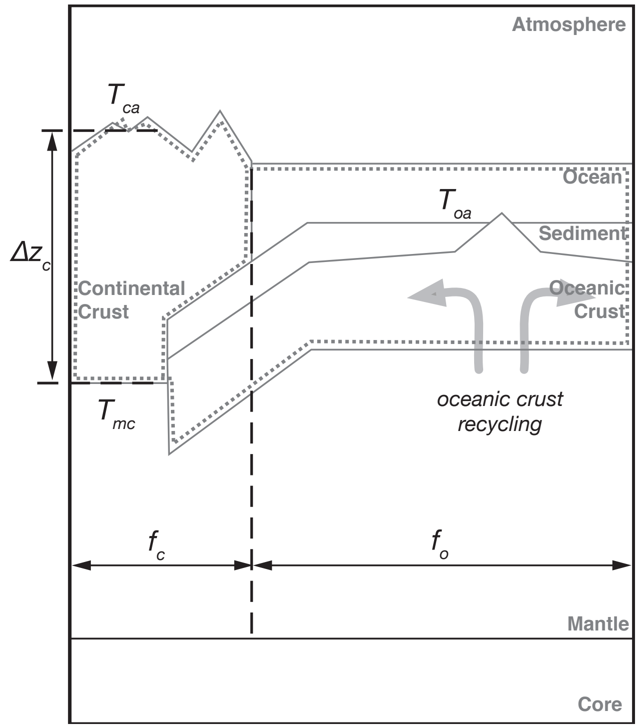

Fig. 7. Oceanic crust recycling: Heat flows from the mantle out into space via continental and oceanic crust. The proportion of continental to oceanic crust is denoted with f_{c} and f_{o} respectively where $f_{\mathrm{o}} = 1 - f_{\mathrm{c}}$. The total thickness of continental crust is denoted with Δz_{c}. The upper mantle temperature is denoted with T_{mc}, temperature of the oceanic crust – ocean boundary is denoted with T_{oa} and temperature of the continental crust – atmosphere boundary is denoted with T_{oa}

.

where f_{c} is the fractional cover of the continents, k_{c} the conductivity of continental crust, T_{mc} the temperature of the mantle-crust boundary, T_{ca} the temperature at the crust-atmosphere boundary and Δz_{c} the thickness of continental crust. Heat transport through oceanic crust is modelled as heat diffusion:

$$\frac{\partial T}{\partial t} = \kappa\, \frac{\partial^2 T}{\partial z^2}, \qquad (19)$$

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. Since we are interested in the steady state solution only, this lets us replace the time variable t with the distance from the mid-oceanic ridge x using the relation

$$t = \frac{x}{v_0}, \qquad (20)$$

where v_0 is the oceanic crust creep velocity. This way the differential Eq. (19) can be transformed into the following time-independent expression:

$$v_0 \frac{\partial T}{\partial x} = \kappa \frac{\partial^2 T}{\partial z^2}. \tag{21}$$

We assume that the temperature of newly formed oceanic crust is that of the upper mantle, T_{mc}, so:

$$T(z, 0) = T_{\mathrm{mc}}. \tag{22}$$

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

$$T(0, t) = T_{\mathrm{oa}}, \quad T(\infty, t) = T_{\mathrm{mc}}. \tag{23}$$

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

$$T(z, x) = T_{\mathrm{oa}} + (T_{\mathrm{mc}} - T_{\mathrm{oa}}) \operatorname{erf}\left(z \sqrt{\frac{v_0}{4 \kappa x}}\right), \tag{24}$$

where T_{oa} is the surface temperature of oceanic crust. Having this temperature profile, it is straightforward to calculate the surface heat flow through oceanic crust, q_{oc}, by taking the spatial derivative of T at $z = 0$

$$q_{\mathrm{oc}} = k_{\mathrm{o}} \left(\frac{\mathrm{d}T}{dz}\right)_{z=0} = k_{\mathrm{o}} \cdot (T_{\mathrm{mc}} - T_{\mathrm{oa}}) \sqrt{\frac{v_0}{\pi \kappa x}}, \tag{25}$$

where k_{o} is the heat conductivity of oceanic crust. This gives a relationship between the surface heat flux and the distance of the oceanic crust from the mid-oceanic ridge x. 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}:

$$Q_{\mathrm{oc}} = L_{\mathrm{r}} \int_0^{L_{\mathrm{p}}} k_{\mathrm{o}} \cdot (T_{\mathrm{mc}} - T_{\mathrm{oa}}) \sqrt{\frac{v_{\mathrm{o}}}{\pi \kappa x}}\, \mathrm{d}x, \tag{26}$$

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:

$$J_{\mathrm{oc}} = \frac{2}{A_{\mathrm{o}}} L_{\mathrm{r}} \sqrt{L_{\mathrm{p}}}\, k_{\mathrm{o}} \cdot (T_{\mathrm{mc}} - T_{\mathrm{oa}}) \sqrt{\frac{v_{\mathrm{o}}}{\pi \kappa}}, \tag{27}$$

where A_{o} is the surface area of oceanic crust. We define $r = L_{\mathrm{p}}/L_{\mathrm{r}}$ 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:

$$J_{\mathrm{oc}} = 2\sqrt[4]{\frac{f_{\mathrm{o}}^3}{A_{\mathrm{o}}\, r}}\, k_o \cdot (T_{\mathrm{mc}} - T_{\mathrm{oa}}) \sqrt{\frac{v_{\mathrm{o}}}{\pi \kappa}}, \tag{28}$$

where f_{o} is the fractional coverage of oceanic crust and is fixed at 0.7. This produces values for, $v_{\mathrm{o}} = 0.01\mathrm{m}^{-1}\,\mathrm{yr}^{-1}$ (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.

Table 4. Values for oceanic crust recycling when "diffusion" parameter $\gamma = 1.5 \times 10^{-11}$ which produces maximum rates of power generation (and dissipation and so entropy).

Variable	Description	Value
$T_{\mathrm{oa}} - T_{\mathrm{mo}}$	Temp gradient mantle-oceanic crust	600 K
J_{oc}	Heat flux through oceanic crust	38 TW
J_{cc}	Heat flux through continental crust	12 TW
v_{o}	Oceanic crust recycling velocity	$2.8\,\mathrm{cm}^{-1}\,\mathrm{yr}^{-1}$

3.4.2 Entropy balance

Entropy production for the oceanic crust recycling system is:

$$\frac{\mathrm{d}S_{\mathrm{o}}}{\mathrm{d}t} = \mathrm{NEE}_{\mathrm{o}} + \sigma_{\mathrm{o}}. \tag{29}$$

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

$$\sigma_{\mathrm{o}} = J_{\mathrm{oc}} \left(\frac{1}{T_{\mathrm{oa}}} - \frac{1}{T_{\mathrm{mc}}}\right). \tag{30}$$

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 $\mathrm{m}^{-1}\,\mathrm{yr}^{-1}\,\mathrm{K}^{-1}$, and is a temperature dependent rate of oceanic crust velocity:

$$v_{\mathrm{o}} = \gamma\, (T_{\mathrm{oa}} - T_{\mathrm{mo}}) \tag{31}$$

Figure 8 shows how entropy production varies as γ varies. Table 4 show values for the oceanic crust recycling model when $\gamma = 1.5 \times 10^{-11}$ which is the value that produces maximum entropy production.

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

$$P = \sigma\, T_{\mathrm{s}} = 0.089\ \mathrm{TW\,K^{-1}} \cdot 293\ \mathrm{K} \approx 26\ \mathrm{TW}. \tag{32}$$

Therefore, 26 TW is the maximum amount of work that can be performed by the oceanic crust recycling system.

3.5 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.

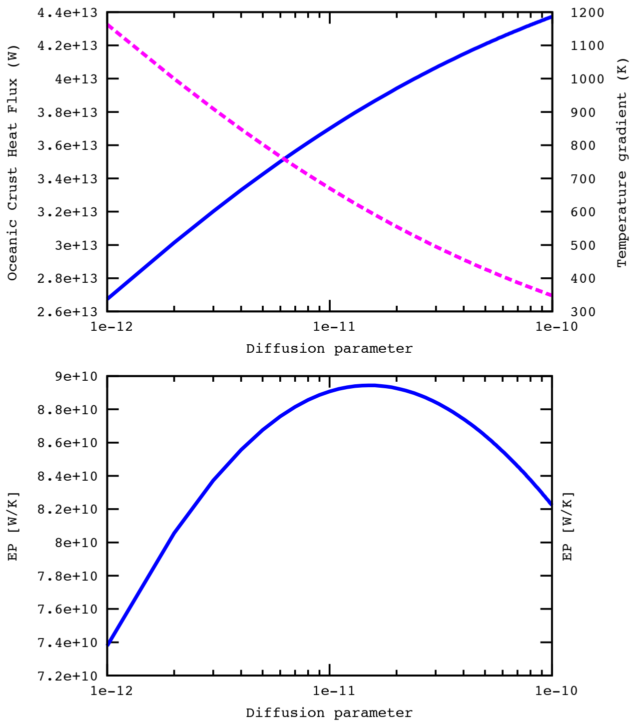

Fig. 8. Top plot: oceanic crust heat flux plotted with solid line (units on left horizontal axis) with varying values for the "Diffusion parameter", γ. Increasing γ increases the heat flux through oceanic crust. Temperature gradient, the difference in temperature between surface of oceanic crust and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

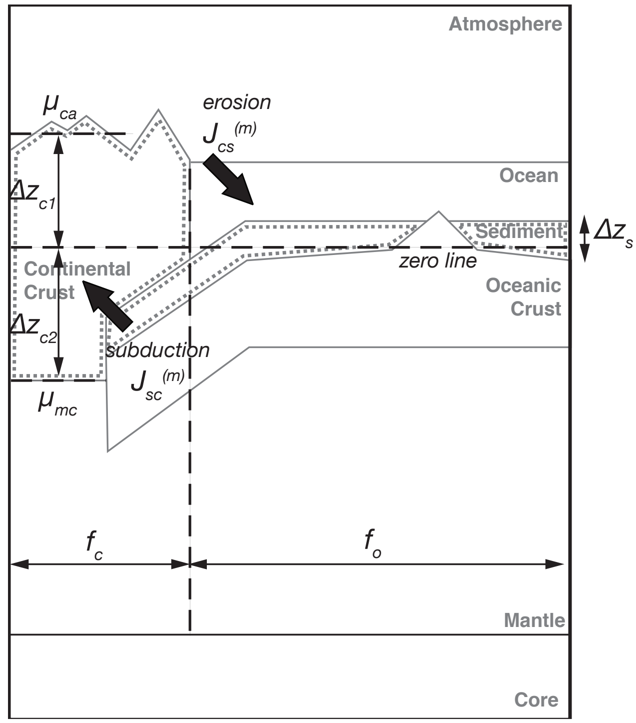

Fig. 9. Uplift and erosion: a simple model of the mass balance of continental crust driven by uplift and erosion. Weathering and erosion processes transfer continental crust material to the ocean where it is deposited as sediment. This flux of mass from continental crust to sediments is denoted with $J_{\mathrm{cs}}^{(\mathrm{m})}$. Continental crust material moves back to the continent though the process of subduction. This flux of mass from sediments to continental crust is denoted with $J_{\mathrm{sc}}^{(\mathrm{m})}$. The horizontal, black dashed line denotes the 'zero line' which marks the surface of a hypothetical Earth with no continental crust. $\Delta z_{\mathrm{c},1}$ denotes the thickness of continental crust above, $\Delta z_{\mathrm{c},2}$ denotes the thickness of continental crust below the zero line. $\Delta z_{\mathrm{c},1}+\Delta z_{\mathrm{c},2}=\Delta z$ the total thickness of continental crust. Δz_{s} denotes the thickness of sediments on the ocean floor. f_{c} and f_{o} denotes the proportional coverage of continental crust and oceanic crust respectively.

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. The process of continental crust uplift and erosion are characterised in terms of competing processes that move material away and towards thermodynamic equilibrium. Mountains represent an energy gradient that erosion dissipates; material is moved from high above the surface of the Earth down to the sea floor. 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 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}:

$$\mu_{\mathrm{ca}} = g \cdot z_{\mathrm{c},1}, \tag{33}$$

where g is gravity and $z_{\mathrm{c},1}$ height above the zero line, where the zero line is a hypothetical surface on an Earth with no continental crust and can be thought of as a waterline on the side of a ship's hull in that the less dense continental crust "floats" on the asthenosphere. The geopotential of continental crust material at the crust/mantle boundary, μ_{mc}, is given with:

$$\mu_{\mathrm{mc}} = \frac{\rho_{\mathrm{m}} - \rho_{\mathrm{c}}}{\rho_{\mathrm{c}}} \cdot g \cdot z_{\mathrm{mc}}, \tag{34}$$

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 densities ρ_c and ρ_s and the fractional coverage of continents f_c and oceans f_o:

$$M_c = \rho_c \cdot \Delta z_c \cdot f_c \quad M_s = \rho_s \cdot \Delta z_s \cdot f_o. \tag{35}$$

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:

$$\frac{dM_c}{dt} = J_{sc}^{(m)} - J_{cs}^{(m)} \quad \frac{dM_s}{dt} = J_{cs}^{(m)} - J_{sc}^{(m)}, \tag{36}$$

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 $\rho_s \Delta z_s$ times the velocity v_o of the oceanic crust:

$$J_{sc}^{(m)} = \rho_s \cdot \Delta z_s \cdot v_o \cdot \sqrt{\frac{f_o}{r\,A_o}} \cdot 3, \tag{37}$$

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, $\Delta z_{c,1}$:

$$J_{cs}^{(m)} = k_{cs} \cdot \rho_c \cdot \Delta z_{c,1}, \tag{38}$$

where k_{cs} is an erosion parameter. The speed of continental crust moving 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.

$$\begin{aligned} z_c \cdot \rho_c \cdot \frac{dv_c}{dt} &= F_m + F_c + F_f \\ &= \rho_c \cdot \mu_{mc} - \rho_c \cdot \mu_{ca} - z_c\,\psi_c\,v_c, \end{aligned} \tag{39}$$

where ψ_c is a friction coefficient. Appendix I shows how the ψ value of $1.5 \times 10^{12}\,\mathrm{kg^{-1}\,m^3\,s^{-1}}$, was derived. In steady state, $dv_c/dt = 0$, the uplift velocity is:

$$v_c = \frac{\rho_c \cdot (\mu_{mc} - \mu_{ca})}{\psi_c \cdot z_c}. \tag{40}$$

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 $\mathrm{NEE}_c^{(m)}$:

$$\frac{dS_c^{(m)}}{dt} = \mathrm{NEE}_c^{(m)} + \sigma_c^{(m)}. \tag{41}$$

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

$$\sigma_c^{(m)} = J_m \cdot \frac{(\mu_{ca} - \mu_s)}{T}. \tag{42}$$

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 maximised. This maximisation of erosional entropy production is equivalent with maximising the uplift work or minimising the frictional dissipation. In other words the uplift-engine would work with maximum efficiency. To calculate the entropy produced by uplift, we use Eq. (40) 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 $\approx 1.8 \times 10^{-14}$ $\mathrm{s^{-1}}$ which produces maximum rates of entropy production for the erosion/uplift model. This produces an uplift velocity of $0.7\,\mathrm{mm^{-1}\,yr^{-1}}$ and a total continental crust thickness of 22 km. This would result in global amounts of erosion of $8.2 \times 10^{14}\,\mathrm{kg\,yr^{-1}}$ which is a little over one magnitude higher than the estimates of Syvitski et al. (2005).

We can directly calculate the amount of energy dissipated by erosional processes as it is proportional to the flux of material from continental crust of certain height. Since dissipation equals work at steady state, we get an uplift work of 0.03 TW.

4 Discussion

In this section we will show that the three models presented previously interact via their shared boundary conditions: altering the dynamics of one system results in a change of the boundary conditions and so dynamics of another system. In

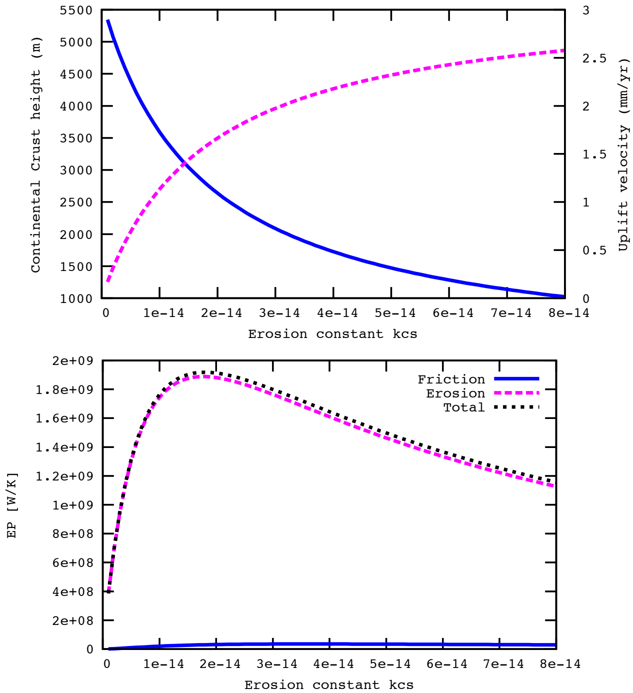

Fig. 10. Top plot: continental crust height plotted with solid line (units on left horizontal axis) and uplift velocity plotted with dashed line (units left horizontal axis) with varying values for erosion constant k_{cs}. Increasing k_{cs} decreases continental crust height and increases uplift velocity.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

particular, we will show that altering the rate of erosion and so thickness of continental crust alters the temperature of the upper mantle that in turn alters the rates of oceanic crust recycling. We then go onto to discuss the implications of these results for the geological carbon cycle. We propose an additional feedback mechanism to this cycle where altering the rate of continental crust erosion alters oceanic crust recycling and so leads to changes in the rates of outgassing of carbon dioxide. We situate the models and discussion within a broader assessment of free energy generation and dissipation within the Earth system. We show that surface life generates a magnitude more power than geological processes. Only a small fraction of this power would be required to alter the boundary conditions for interior geological processes. Finally, we discuss limitations of the models and propose future research to extend this study.

4.1 Assessing surface to interior interactions

We will show sensitivity analysis results that assess the ability of surface processes to alter the boundary condition of interior processes and their rates of work. Two sets of results are shown: *adaptive* and *non-adaptive* oceanic crust recycling. For non-adaptive simulations the γ "diffusivity" parameter was fixed. This parameter was adjusted in the oceanic crust recycling model in order that the rate of oceanic crust recycling was that which led to maximum rates of power in analogous fashion to the adjustment of the Nusselt number in the mantle convection model. With the γ parameter fixed, increasing the temperature of the upper mantle produced a monotonic, non-linear increase in the rate of oceanic crust recycling. For the adaptive simulations, we assumed that the oceanic crust recycling system would respond to changes in upper mantle temperatures and relax back to a state in which maximum rates of power were produced. Consequently, the γ term no longer remained fixed, but was adjusted in order to find that value of oceanic crust recycling that produced maximum rates of power generation for new values of upper mantle temperature.

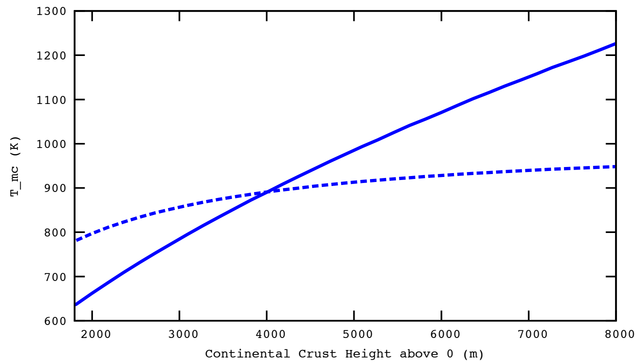

Fig. 11. Sensitivity of upper mantle temperature to continental crust thickness. Solutions assuming non-adaptive crust recycling are shown with a dashed line, solutions assuming adaptive crust recycling are shown with a solid line. The temperature at the mantle-crust boundary (T_{mc}) increases as the thickness of continental crust increases.

Figure 11 shows the temperature of the mantle-crust boundary as a function of continental crust thickness. Increasing the thickness of continental crust, increases the thickness of the insulation on the Earth's surface and increases the temperature of the mantle-crust boundary in both adaptive and non-adaptive simulations. The rate of change of temperature is higher with adaptive crust recycling. Figure 12 shows how the power generated by the oceanic crust recycling system changes with continental crust height. Higher continental crust increases the mantle-crust boundary temperature gradient and so more power can be generated by the oceanic crust recycling system. Figure 13 shows how the rate of oceanic crust recycling changes as the thickness of continental crust changes. Increasing the overall thickness of continental crust would lead to an increase in the rate of oceanic crust recycling in the non-adaptive assumption: a greater temperature gradient would drive the system faster. A greater change is observed with an adaptive system. Increasing the height of continental crust would lead to

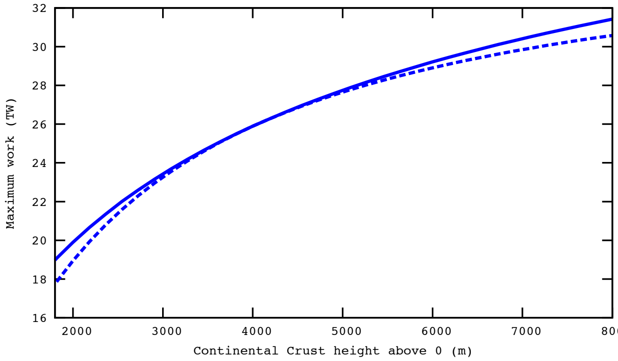

Fig. 12. Sensitivity of rates of maximum power (TW) produced in the oceanic crust system with different continental crust heights. Solutions assuming non-adaptive crust recycling are shown with a dashed line, solutions assuming adaptive crust recycling are shown with a solid line.

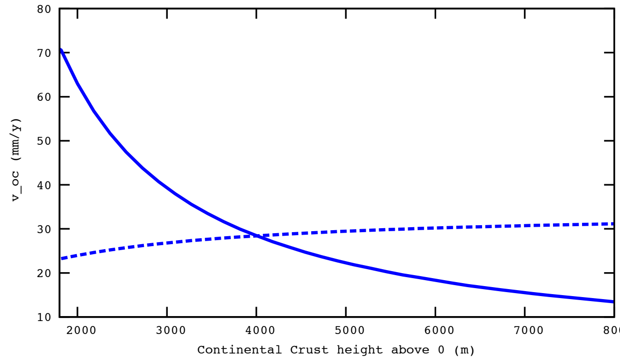

Fig. 13. Sensitivity of oceanic crust recycling rate to continental crust thickness. Solutions assuming non-adaptive crust recycling are shown with a dashed line, solutions assuming adaptive crust recycling are shown with a solid line.

a *decrease* in the rate of oceanic crust recycling. Decreasing the rate of oceanic crust recycling would increase the amount of heat that is conducted through continental crust. This heat is not able to drive the recycling system and so is "lost" to the oceanic crust recycling system. Therefore the decrease in recycling rates may seem counterintuitive. However, for a range of parameter values, the increase in the heat flux through continental crust is offset by the increase in the temperature gradient over the mantle-crust boundary. Consequently the greatest rates of power generation in the oceanic crust recycling system will be produced with a reduction in the rate of recycling as the thickness of continental crust increases. Equivalently, as continental crust thickness decreases, the temperature gradient will decrease and so the oceanic crust recycling system will respond by increasing the rate of recycling.

4.2 Implications for geological carbon cycle

In this section we consider how life may affect the long-term geological carbon cycle, via its effects on interior geological processes. We produce results for a hypothetical "dead Earth" in which the total rate of erosion of continental crust material to the oceans is reduced. We then consider how this change in erosion and so continental crust thickness would affect rates of oceanic crust recycling and outgassing of CO_2.

Life's effects on global weathering and erosion of continental crust is an important element in the geological carbon cycle as well as other biogeochemical cycles. Kelly et al. (1998) lists the ways plants can alter weathering rates which includes the generation of weathering agents, biocycling of cation and production of biogenic minerals. Moulton et al. (2000) analysed empirical data that suggests vascular plants produce a four fold increase of calcium and magnesium weathering. They conclude that the colonisation of land by plants some 410–360 million years ago would have had significant effects on atmospheric CO_2. Lichens are also important organisms with respect to the mechanical and chemical weathering of rocks (Chen et al., 2000). Schwartman and Volk (1989) and Schwartzman (1999) argue for strong biotic effects from microbes and plants such that weathering rates of silicate rocks on an abiotic Earth with the same average surface temperatures and partial pressure of CO_2 could be two to three magnitudes lower than current levels. We assume that extinction of life would result in a ten fold decrease in the rate of weathering of continental crust.

In order to determine what would be the effect on total erosion rates, we need to establish the ratio of eroded (via mechanical processes) to weathered (via chemical processes) flux. Following Walling and Webb (1983) we assume that the ratio of eroded/suspended and weathered/solution of flux from continents to oceans via rivers is 4:1. Erosion is more important for the transport of material from continents to oceans. Indeed, a ratio of 40:1 may be more appropriate. We simulate the total extinction of life on Earth (or at least all life involved in chemical weathering) by decreasing the erosion parameter k_{cs} from 1.6×10^{-14} to 1.31×10^{-14}. This leads to an increase in the thickness of continental crust from 22 km to 24 km. What would this increase in continental crust have on upper mantle temperatures, rates of oceanic crust recycling and outgassing of CO_2?

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 outgassing of primordial CO_2 as this is released from mid oceanic ridges where new oceanic crust rises to the surface. Gerlach (1989) estimates current outgassing of CO_2 are mid-oceanic ridges to be 10 to 37.8×10^{12} g yr^{-1} while a more recent analysis of empirical data (Chavrit et al., 2009) gives lower estimates of 1.4×10^{10} g yr^{-1}. Following Chavrit et al. (2009) we assume that the current rate of outgassing of CO_2 is 1.4×10^{10} g yr^{-1}. We scale this rate to our estimated rate of oceanic crust recycling of 2.84 cm/yr. For small changes of oceanic crust recycling, it would be reasonable to assume a linear change in outgassing. However the dynamics are complex and consequently a linear assumption becomes less tenable with large changes in oceanic crust recycling. For example, the degree of partial melt and solubility of CO_2 is unlikely to remain constant as the solubility depends on pressure, temperature and in part oxygen fugacity (Bottinga and Javoy, 1989, 1990).

For non-adaptive oceanic crust recycling (oceanic crust recycling increases as upper mantle temperature increases) removing the biotic component of weathering from total erosion of continental crust leads to an increase in the rate of oceanic crust recycling of $0.05\,\mathrm{cm\,yr^{-1}}$ and an increase in outgassing of CO_2 of $2\times10^8\,\mathrm{g\,yr^{-1}}$. For adaptive oceanic crust recycling (oceanic crust recycling decreases as upper mantle temperature increases) removing the biotic component of weathering from total erosion of continental crust leads to a decrease in the rate of oceanic crust recycling of $0.23\,\mathrm{cm\,yr^{-1}}$ and a decrease in outgassing of CO_2 of $0.12\times10^{10}\,\mathrm{g\,yr^{-1}}$. Within the context of the global carbon cycle, these numbers are small. However they represent a feedback on the outgassing of CO_2 and so how much carbon is delivered from the interior of the Earth to the surface. We discuss this feedback in the following section.

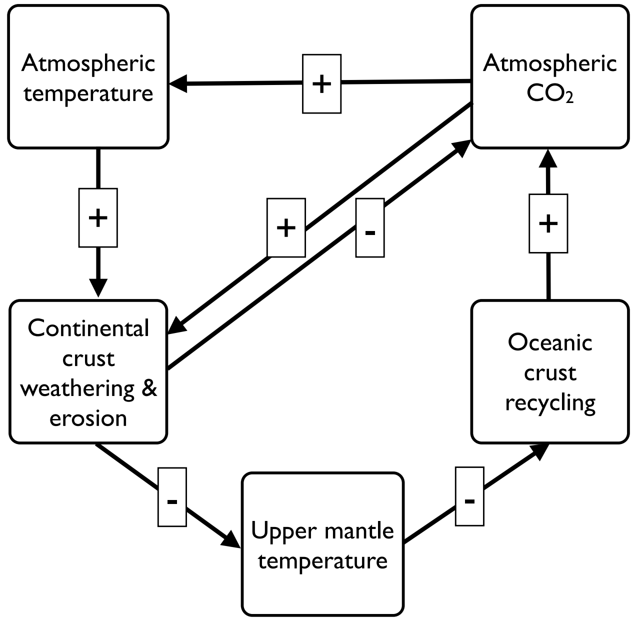

Fig. 14. Surface to interior interactions are shown schematically. Arrows denote the feedback relationships and their sign. The geological carbon cycle is represented by the Oceanic crust recycling, Atmospheric CO_2, Atmospheric temperature and Continental crust erosion boxes. We propose a new feedback between rates of continental crust erosion and oceanic crust recycling via upper mantle temperature. Increasing rates of continental crust erosion would lead to a decrease in upper mantle temperature. If we assume the oceanic crust recycling relaxes back to a state of maximum power generation, an increase in upper mantle temperature would lead to a decrease in rates of crust recycling and outgassing of CO_2, see Fig. 13. It is important to note that these different feedback mechanisms operate over a wide range of timescales.

Feedback and steady states

Figure 14 shows the feedbacks in the geological carbon cycle discussed in this paper. This cycle has two feedback loops. The first is the carbon – silicate feedback loop whereby increases of CO_2 in the atmosphere and/or surface temperatures would lead to increased intensity of weathering of continental crust and so reduction in CO_2 (Walker et al., 1981). It is this loop that is involved in the stabilisation of global temperatures over geological time as the increase in radiative forcing produced by the progressive increase in luminosity of the sun has been offset by the additional drawdown of atmospheric CO_2. This also puts a time limit or the possible age of the biosphere at approximately 1 billion years hence as increasing luminosity of the sun is offset by a reduction in CO_2 until there is no appreciable amounts of the greenhouse gas in the atmosphere and so further increases in luminosity would lead to progressively higher temperatures and the extinction of all life on Earth (Sagan and Mullen, 1972; Lovelock and Whitfield, 1982; Caldeira and Kasting, 1992; Lenton and von Bloh, 2001). The additional loop we propose is the effects that changes of continental crust have on upper mantle temperatures. Increasing continental crust would produce higher upper mantle temperature and so alter the rate of oceanic crust recycling. If we assume the oceanic crust recycling system were to relax back into a steady state that leads to maximum rates of power generation, then this would lead to a decrease in outgassing of CO_2.

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. What would differ, would be the partial pressures of CO_2 and global temperatures on an abiotic world. They would both be higher. However, this assumes that the total outgassing of CO_2 would be the same on a biotic and abiotic Earth. We have argued that processes happening on the surface of the Earth may affect interior processes and so the outgassing of CO_2. Rather than CO_2 outgassing being fixed or driven by interior geological processes, 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 entering the atmosphere is not fixed but is in part determined by the actions of life.

More abundant life, or rather more life directly involved in the weathering and erosion of continental crust, would lead to an increase in oceanic crust recycling and increase in outgassing of CO_2. In that respect this is a positive feedback loop as, all things equal, higher partial pressures of CO_2 leads to a general increase in growth and carbon accumulation in phytomass (Amthor, 1995).

4.3 Power generation in the Earth system

Another perspective of biotic effects on interior geological dynamics can be inferred from the supply rates of free energy that can fuel processes that lead to the physical and chemical transformation and transport of material on the Earth's surface. Estimates for the amount of power generated by life on the land and sea along with abiotic processes are shown in Fig. 15. The derivation of these numbers is discussed in the following sections.

4.3.1 Physical processes

Estimates for the maximum power associated with mantle convection (12 TW), oceanic crust cycling (26 TW), and continental uplift (<1 TW) are taken from the previous sections in this paper. 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 vapour and therefore the hydrologic cycle. The strength of the hydrologic cycle is relevant here in that it (a) distills seawater, (b) lifts vapour 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. Given a salinity of 3.5 %, the work required to desalinate a litre of seawater is approximately 3.8 kJ. For a net moisture transport of $37 \times 10^{12}\ \mathrm{m^3\ yr^{-1}}$, this corresponds to a power of approximately 4 TW. This power is potentially available to dissolve rock and bring the precipitated water to saturation with the continental rocks. However, since most of the salinity of the ocean is sodium chloride, which is only a relatively minor product of chemical weathering, the actual power for chemical weathering should be much less.

To estimate the power inherent in the potential energy in runoff and for a maximum estimate for physical weathering of continental rocks, we use estimates from a spatially explicit land surface model with realistic, present-day climatic forcing. With this model we estimate the power in

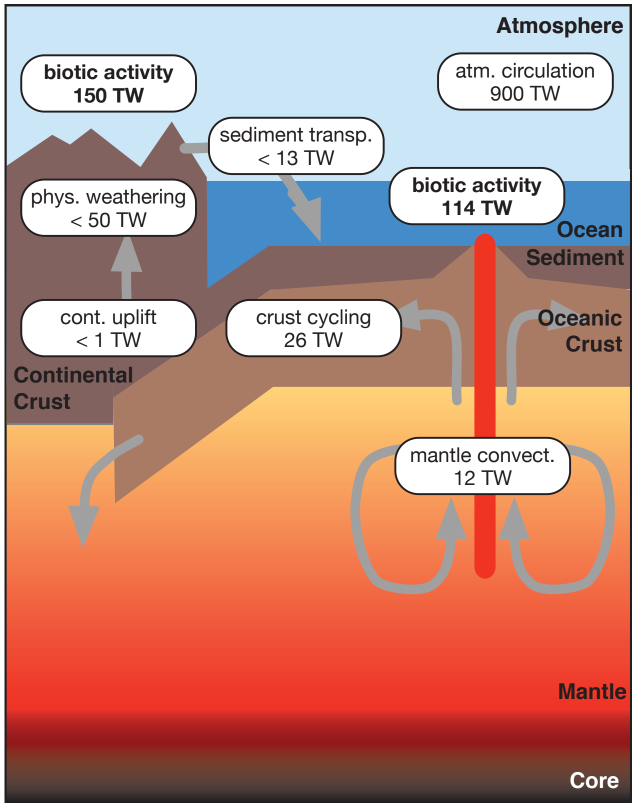

Fig. 15. Estimation of rates of work for components of the Earth system. Starting from estimates of net photosynthetic rates, biotic activity contributes a total of 264 TW with 150 TW and 114 TW being produced by life on land in the oceans respectively. These and the other estimates of power should be understood in terms of the absolute maximum possible.

continental runoff to be 13 TW, which sets the upper limit on the power available for sediment transport. The potential to physically weather bedrock by seasonal heating and cooling and freeze-thaw dynamics is less than 50 TW. This latter number is an upper estimate in that it assumes bedrock to be present at the surface.

4.3.2 Biological processes

The upper limit for the power associated with biotic activity can be derived from estimates of net photosynthetic rates. Following Kleidon (2009) we assumed net photosynthetic rates of $120\ \mathrm{GtC\ yr^{-1}}$ on land and $90\ \mathrm{GtC\ yr^{-1}}$ in the oceans. Photosynthesis consumes approximately 1710 kJ per mol of fixed carbon, which corresponds to approximately $1.8\ \mathrm{W\ m^{-2}}$ or 900 TW at the global scale. However, most of this energy is immediately respired by photorespiration. When we consider the carbon uptake in terms of the generation of chemical free energy in form of sugars (with a free energy content of $479\ \mathrm{kJ\ mol\ C^{-1}}$), these numbers translate into 150 TW of power in biotic activity on land and 114 TW in the oceans. These numbers represent the absolute maximum amount of power generated by life and are shown in Fig. 15.

It may be argued that most of this energy is effectively lost via metabolic costs or stored as relatively inert biomass such as xylem in plants. However, these metabolic costs may have important geological consequences and while biomass stored in roots may be relatively chemically inert, it may have non-trivial effects on rates of mechanical weathering and erosion of rocks as roots grow into and expand crack in rocks. Marschner (1995) estimated that 5 % to 21 % of all photosynthetically fixed carbon is transferred to the rhizosphere through root exudates. Amino acids, organic acids, sugars, phenolics, and various other secondary metabolites are believed to comprise the majority of root exudates. The costs of producing these compounds is offset by the benefits accrued due to the regulation of the soil microbial community in their immediate vicinity, and changes in the chemical and physical properties of the soil (Walker et al., 2003). In this way plants are able to exert significant effects over their local environments and so will have important biological and ecological effects (Bais et al., 2006). There will also be significant geological impacts as root exudates will alter the chemical composition of the soil, for example increasing the concentration of CO_2 which would increase the chemical weathering of rocks and so increase rates of chemical weathering (Berner, 1997). Also of relevance is the production of acids by lichens in order to chelate minerals that can significantly alter the flux of elements from rocks into the biosphere (Schatz, 1963).

This first order estimate of biotic free energy generation represents the absolute maximum amount of work that life is able to do on the Earth's surface. By what processes this work occurs, what the actual amount of work done by life is, and what its effects are on particular geological processes are open questions which we have begun to address in this study.

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 many of 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, in part, 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 assumptions becomes harder 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 accretion 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.

The effects of life on erosion rates can vary and therefore we should not necessarily assume that a decrease in the amount of life would produce a monotonic decrease in erosion. For example, the overall effects of life may have been to *increase* the thickness and total amount of continental crust. Rosing et al. (2006) argued that the evolution of oxygenic photosynthesis in the oceans, led to an increase in hydration of the mafic oceanic crust and so production of partial melts that would have gone on to form continental crust. In the absence of the 'titration of the oceans' by photosynthetic organisms (Lovelock and Watson, 1982) rates of continental crust formation may have been much lower. In this sense, the emergence and evolution of life on the surface of the Earth may have had a profound impact on the boundary conditions for the thermodynamic processes operating within its interior. Over shorter timescales Viers et al. (2000) show that in flat tropical areas weathering rates are low despite the presence of a dense vegetal cover. Similarly, Vanacker et al. (2007) show that the erosion of steep slopes may be reduced by vegetation as the roots of plants can bind and maintain a protective top soil which acts as a protective layer over otherwise bare rock. Millot et al. (2002) conclude that weathering rates are closely coupled with erosion rates (their study finds a global power law between chemical and physical denudation rates). Consequently, topography and climate rather than biology may be significantly more important in determining the overall amount of material that is removed from continental crust. While this may be true, it can still be the case that life can have a detectable effect on continental crust dynamics.

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 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.

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”. However, this would amplify the effects of altering the thickness of continental crust on upper mantle temperatures and so increase the change in the rate of oceanic crust recycling and so outgassing of CO_2. 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.

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.

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 parameterisation, 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 interactions, 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 have taken the first step towards quantify the effects surface life can have on interior geological processes. 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. 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. 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. Specifically, we showed that changes in the rates of erosion of continental crust will lead to changes in upper mantle temperatures and so rates of oceanic crust recycling. Altering the rate of oceanic crust recycling would lead to changes in the outgassing of primordial CO_2.

We compiled estimates for power generation for different elements of the Earth system. Life in the oceans and on land generated 264 TW, much larger than any geological process. Consequently, it should come as no surprise that life has profoundly affected different aspects of the Earth system. We would argue that in that sense, surface life has already the potential to affect interior geological processes as life will be altering the boundary conditions for these interior processes. If we characterise the Earth as a heat engine, then altering the temperature of its surface, it’s cold reservoir, will lead to changes in the dynamics of the interior. These dynamics would include, tectonics, convection within the mantle and the freezing of the outer core. It is worth noting that processes in the Earth’s core may have been essential for the maintenance of widespread life on the surface of the Earth with respects to the generation of the Earth’s magnetic field via the ohmic dissipation produced in the fluid outer core (Stevenson et al., 1983; Anderson, 1989). In the absence of a magnetic field, the surface of the Earth would be subject to a range of harmful radiation from the Sun and the solar wind would eventually strip away the Earth’s atmosphere. It is interesting to speculate over the relationship between the generation of a magnetic field, mantle convection, mobile lid tectonics and a widespread biosphere. Earth possesses all four. Are the first three necessary for a widespread biosphere? Does a widespread biosphere affect the other three? The outgassing of volatile components such as CO_2 from the Earth’s interior along with the emergence and evolution of life has had a profound influence on the evolution of the Earth’s atmosphere and hydrosphere and we would argue that this represents an important boundary condition for interior processes. For example, the presence of large amounts of water in the Earth’s oceans and the resulting hydration of mafic rocks is an important element of metamorphic processes. It is possible that rates of subduction and the dynamics of tectonics would be significantly different in the absence of water. What would tectonic processes on an Earth with a runaway greenhouse and no oceans look like?

Our study is only a first step towards quantifying the effects of life on the functioning of the whole planet and its evolution. Natural next steps would be to develop 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 investigate the driving forces of planetary disequilibrium, which, as Lovelock noted many years ago, should be an indicator of widespread life on a planet. Hence, our thermodynamic approach to understand these Earth system processes and their interactions should be a central building block towards searching for the fundamental signs of life elsewhere in the universe.

Appendix A

Derivation of uplift friction coefficient

The total thickness of continental crust, $z_c = z_{c,1} + z_{c,2}$, where $z_{c,1}$ is the height of continental crust above the zero line and $z_{c,2}$ is the depth of continental crust below the zero line. The uplift force of the upper mantle acting on continental crust, F_m is found with:

$$F_m = (\rho_m - \rho_c)\, g\, z_{c,2} - \rho_c\, g z_{c,1}, \tag{A1}$$

where g is gravity and ρ_m and ρ_c are the densities of the mantle and continental crust respectively. The frictional force, F_f, that counteracts F_{up} is found with:

$$F_f = -\psi_c\, z_c\, v_c. \tag{A2}$$

The velocity of continental crust up and down within the asthenosphere, $v + c$, will vary over time in accordance with:

$$z_c\, \rho_c \frac{dv_c}{dt} = F_m + F_f \tag{A3}$$

$$= (\rho_m - \rho_c)\, g\, z_{c,2} - \rho_c\, g\, z_{c,1} - \psi_c\, z_c\, v_c.$$

We introduce $\delta = z_{c,1} - z_{c,1,eq}$, which is the displacement of the height of continental crust from its equilibrium value $z_{c,1,eq} = z_{c,2} \cdot \frac{\rho_m - \rho_c}{\rho_c}$. This means that for $\delta = 0$ no uplift would occur. We do the substitution in the differential equation and obtain a differential equation for δ:

$$z_c\, \rho_c\, \ddot{\delta} = -\rho_c\, g\, \delta - z_c\, \psi_c\, \dot{\delta}. \tag{A4}$$

At steady state, the velocity of continental crust will be:

$$v_c = -\frac{\rho_c\, g}{z_c\, \psi_c}\, \delta. \tag{A5}$$

$v_c = 0$ when $\delta = 0$. The only unknown here is the friction parameter ψ_c

A1 How to obtain ψ_c

We know that isostatic rebound has a timescale of about 10 000 yr so the relaxation time of the system described in Eq. (A4) should be in that order of magnitude. We solve Eq. (A4) and obtain, for the case that friction is dominating ($\frac{\psi_c}{2\,\rho_c} > \sqrt{\frac{g}{z}}$):

$$\delta(t) = c_1\, e^{\lambda_1 t} + c_2\, e^{\lambda_2 t} \tag{A6}$$

where

$$\lambda_1 = -\frac{\psi_c}{2\,\rho_c} + \sqrt{\frac{\psi_c^2}{4\,\rho_c^2} - \frac{g}{z_c}} \tag{A7}$$

$$\lambda_2 = -\frac{\psi_c}{2\,\rho_c} - \sqrt{\frac{\psi_c^2}{4\,\rho_c^2} - \frac{g}{z_c}}$$

We assume that gravitational forces are much smaller than frictional forces, which means that λ_1 is very close to 0, therefore dominates and results in a very slow decay. We identify the timescale $\tau = \frac{1}{\lambda_1}$ and do a Taylor expansion $\lambda_1 \approx \frac{4\rho_c g}{\psi_c z_c}$ and obtain the relation:

$$\tau = \frac{1}{\lambda} = \frac{\psi_c\, z_c}{4\,\rho_c\, g} \tag{A8}$$

This relation is used to determine ψ_c in a way that it produces the correct isostasy timescale.

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