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Continental insulation, mantle cooling, and the surface area of oceans and continents

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Abstract

It is generally assumed that continents, acting as thermal insulation above the convecting mantle, inhibit the Earth's internal heat loss. We present theory, numerical simulations, and laboratory experiments to test the validity of this intuitive and commonly used assumption. A scaling theory is developed to predict heat flow from a convecting mantle partially covered by stable continental lithosphere. The theory predicts that parameter regimes exist for which increased continental insulation has no effect on mantle heat flow and can even enhance it. Partial insulation leads to increased internal mantle temperature and decreased viscosity. This, in turn, allows for the more rapid overturn of oceanic lithosphere and increased oceanic heat flux. Depending on the ratio of continental to oceanic surface area, global mantle heat flow can remain constant or even increase as a result. Theoretical scaling analyses are consistent with results from numerical simulations and laboratory experiments. Applying our results to the Earth we find, in contrast to conventional understanding, that continental insulation does not generally reduce global heat flow. Such insulation can have a negligible effect or even enhance mantle cooling, depending on the magnitude of the temperature dependence of mantle viscosity. The theory also suggests a potential constraint on continental surface area. Increased surface area enhances the subduction rate of oceanic lithosphere. If continents are produced in subduction settings this could enhance continental growth up to a critical point where increased insulation causes convective stress levels to drop to values approaching the lithospheric yield stress. This condition makes weak plate margins difficult to maintain which, in turn, lowers subduction rates and limits the further growth of continents. The theory is used to predict the critical point as a function of mantle heat flow. For the Earth's rate of mantle heat loss, the predicted continental surface area is in accord with the observed value.

Keywords: heat flow; continent–ocean area; mantle convection; continental growth

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1. Introduction

The chemical buoyancy of continental relative to oceanic lithosphere leads to a bimodal distribution of elevations on our planet. It also leads to two different modes of lithospheric heat transfer. Oceanic lithosphere is the active upper thermal boundary layer of mantle convection [1–3]. Buoyant continental lithosphere, on the other hand, does not participate in the convective overturn of the mantle. Heat transfer through stable continental lithosphere is principally by conduction [4–6] and continents thus act as thermal insulation above the convecting mantle [7–9]. Common sense experience with insulated systems suggests that this should lower the Earth's global rate of internal heat loss. This is the simplest, most intuitive of assumptions and has generally been adopted for thermal history studies [e,g,10–12]. In practice this assumption treats global mantle heat flow as a simple area weighted average of the heat flow through oceanic lithosphere and the heat flow into the base of continents. As subcontinental mantle heat flow is very low [13–16] relative to oceanic heat flow [17], this implies that increased continental area should decrease global mantle heat flow and retard mantle cooling. The goal of this paper is to explore the extent to which this assumption is generally true. In particular, we show that under certain conditions, relevant to the Earth, the presence of insulating continents has no effect on global heat flow and can actually enhance it.

Continental insulation not affecting, or even increasing, mantle heat flow initially seems counter-intuitive. However, a principal effect of continental insulation is to increase the internal temperature and, hence, to reduce the viscosity of the bulk mantle which leads, in turn, to higher convective velocities and enhanced heat flow through ocean basins. The enhanced oceanic heat flow can outweigh the lowered subcontinental mantle heat flow, due to continental insulation, provided the surface area covered by oceans is sufficiently large. Key physical conditions for global mantle heat flow to be unchanged or enhanced by insulating continents are: 1) that the mantle tends toward a thermally well mixed state; and 2) that the resistance to plate motion is predominantly governed by bulk mantle viscosity [18–22], which depends strongly on temperature [23]. The first condition ensures that the insulating effect of continents is also felt beneath oceanic plates. The second condition links the lowering of mantle viscosity as temperature increases to a higher plate velocity and higher oceanic heat flux. Both conditions imply that continental and oceanic heat flows are nonlinearly coupled to one another and that the continent–ocean system would need to be considered in full to address local heat flow through oceanic and/or continental lithosphere.

The remainder of this paper tests the plausibility of our hypothesis. First, we develop a scaling theory to quantify the effect of continents on heat flow. We then test the results of this analysis against both numerical simulations and laboratory experiments.

2. Theory

We consider a thermally convecting, bottom-heated and top-cooled mantle layer of depth D with spatially constant surface and base temperatures of T_s and T_b, respectively. We assume a temperature-dependent mantle viscosity (the exact form will be given below). The average internal mantle temperature, T_i, and associated viscosity, μ_i, depend on the relative surface area, A_c, average thickness, d, and thermal conductivity, K_c, of the continental lithosphere. We consider stable continental lithosphere for which the principal means of heat transfer is conductive. Continental deformation, magmatism, and hydrothermal circulation will not be addressed. These simplifications allow us to treat continents as conducting lids that float atop the convecting mantle. The heat flow scaling we develop will apply to this idealized system. The system isolates the global effects of partially insulating a thermally convecting layer with temperature-dependent viscosity free of added complexities. It also facilitates testing of our ideas with numerical simulations and laboratory experiments. For application to the Earth, the potential limitations imposed by our simplifications should be kept in mind.

Our theoretical approach is founded on thermal network analysis [24,25]. We model the solid Earth as a nonlinear thermal network (Fig. 1). The functional dependencies shown in Fig. 1 highlight the nonlinearity of the system in that parameters that affect

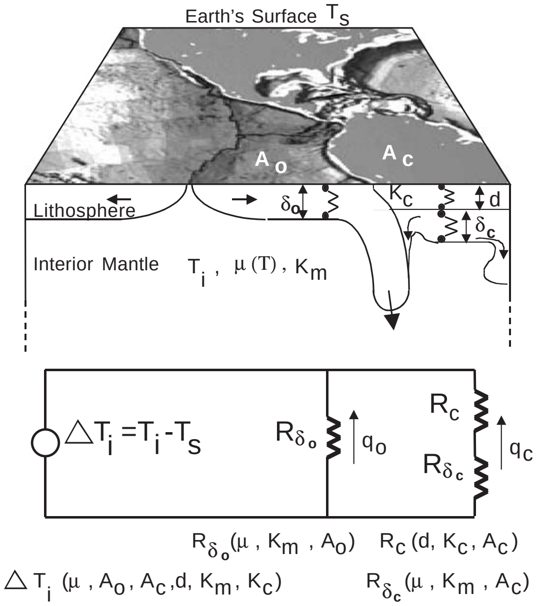


Fig. 1. Cartoon of the solid Earth heat transfer system (top) together with a schematic of our thermal network model (bottom). The simplest relevant network for the question we pose is composed of three components. The first represents conductive heat transfer within stable continental lithosphere of average thickness d_c, surface area A_c, and thermal conductivity K_c. The thermal resistance of this component, R_c, depends directly on d and inversely on K_c and A_c. A second component represents heat transfer from the convecting mantle into the base of continents across a convective sub-layer. The average thickness of this boundary layer, δ_c, and thus the effective resistance of this component, $R_{\delta c}$, depend inversely on convective mantle vigor and, by association, directly on internal mantle viscosity, μ_i. In series, components one and two form the continental path. This path is linked in parallel to an oceanic component. For oceanic regions, internal heat is transferred across an active thermal boundary layer of surface area A_0, i.e., oceanic plates. The average thickness of this boundary layer, δ_o and thus the effective resistance of this path, $R_{\delta c}$, depend inversely on convective mantle vigor and, by association, directly on, μ_i. The average internal temperature of the mantle, T_i, minus the temperature at the system surface, T_s, is the temperature drop, ΔT_i, driving heat transfer across the network.

the thermal resistances of the oceanic and continental lithosphere also affect the temperature drop driving heat loss across the lithosphere. This also highlights that introducing a continental thermal path to the network can affect the local resistance of the oceanic path and vice versa. We will develop a heat flow scaling for the full network in three steps. Steps one and two will consider the heat transfer properties of two end-members: 1) a mantle free of continental lithosphere and 2) a mantle completely covered by stable continental lithosphere. We will present expressions for surface heat flux and average internal temperature for each end-member. This will allow us to define effective thermal resistances for each. Step three will develop an expression for the lumped resistance of the composite system [e,g,24,25]. The changes that occur in the local resistances due to the parallel linking of the oceanic and continental paths will be addressed and these altered resistances will be used to define the effective resistance of the system.

2.1. Oceanic path end-member

We consider the end-member of a mantle completely free of continents. The majority of heat flow scalings developed for mantle convection over the last 30+ years have treated this limit. An attractive aspect of the thermal network approach is that it allows us to take advantage of results from this previous work.

It has long been argued that convection in a mantle with temperature dependent viscosity and an oceanic lithosphere that participates in convective overturn, should behave as an equivalent isoviscous system with a viscosity equal to that of the average internal viscosity of the temperature-dependent system [20,21]. If this is true, then the theoretically expected form of the heat flux scaling, for a bottom heated mantle, should be $q = a_o Ra_i^{1/3}$, where q is the nondimensional surface heat flux, a_o is a geometric scaling constant, and Ra_i is the Rayleigh number defined in terms of average internal viscosity [1,26]. This scaling has the potential to buffer global mantle heat flow against the insulating effect of stable continental lithosphere because it predicts that the effective resistance to advective heat transfer across oceanic lithosphere should decrease with increasing internal temperature and decreasing internal viscosity. The validity of this scaling has been confirmed via numerical simulations of mantle convection with a viscoplastic rheology [27]. A viscoplastic mantle rheology allows for an active lid mode of convection for the oceanic lithosphere, i.e., lithospheric recycling, and for an internal mantle viscosity that depends strongly on temperature [27]. The rheology law remains on a temperature-dependent viscous branch for stresses below a specified yield stress,

τ_{yield}. Along this branch, the viscosity function is given by

$$\mu = A\exp[-\theta T], \tag{1a}$$

where A and θ are material parameters, T is temperature, and μ is mantle viscosity. For stresses above a yield stress the flow law switches to a plastic branch with a nonlinear, effective viscosity given by

$$\mu_{plastic} = \tau_{yield}/I, \tag{1b}$$

where I is the second strain-rate invariant. This rheologic formulation allows zones of localized lithospheric failure, analogs to weak plate boundaries, to form in a self-consistent way. The failure zones allow for lithospheric subduction and mantle stirring akin to plate tectonics. If convective stress levels fall below the yield stress, then weak margins cannot be generated and a stagnant lid [28], i.e., single plate, mode of convection results.

Moresi and Solomatov [27] explored a range of numerical convection simulations employing the rheologic formulation above. They presented a scaling for mantle heat flux in the active lid regime based on simulation results. Bottom heating was assumed and continents were not incorporated into the simulations. The best fit scaling was found to be

$$q_o = 0.385Ra_{io}^{0.293} \tag{2}$$

where q_0 and Ra_{io} are, respectively, the nondimensional surface heat flux and the Rayleigh number defined in terms of average internal viscosity (we have added the subscript 'o' to make it clear that this scaling applies to the oceanic lithosphere only end-member case). This is reasonably close to the theoretically expected scaling if the temperature-dependent, active lid system behaves as an equivalent isoviscous system [20,26] (the slightly lower scaling exponent is not unexpected as the 1/3 exponent holds in the high Ra limit while simulations were run at intermediate to high Ra). The average internal temperature, T_{io}, is required to determine the internal viscosity and, thus, Ra_{io}. For high basal Rayleigh numbers, T_{io}, was shown to approach the mean of the surface and base temperatures which is theoretically expected [20,26]. For lower basal Rayleigh numbers, a relationship for T_{io}, as a function of known system parameters was presented [27]. This allows Ra_i to be determined and closes the heat flux scaling. We will use these results to determine the surface heat flux and average internal temperature for the oceanic lithosphere only end-member.

The viscoplastic formulation also allows us to explore conditions under which convective stress levels drop below the yield stress of oceanic lithosphere thereby "locking" plate margins. In an active lid regime, the upper boundary layer velocity was observed to be controlled by the bulk internal mantle viscosity, μ_{io} [27]. For such a case the appropriate stress scale is

$$\tau \sim (\mu_{io}\kappa)/\delta_o^2, \tag{3}$$

where κ is thermal diffusivity and δ_o is the thickness of the active upper mantle boundary layer. The dependence of convective stress on internal viscosity allows for the possibility that continental insulation could cause a transition from an active to a stagnant lid mode of convection. We will explore this possibility as it introduces an added limit on the degree to which global mantle heat flow can increase with added continental insulation.

2.2. Continental path end-member

We consider the end-member of a mantle completely covered by stable continental lithosphere of thickness d. Although we will not explicitly consider the conditions that lead to the stability of continental lithosphere, we do note that it is most likely achieved through a combination of chemical buoyancy and intrinsic strength [29–31]. Thus, d will parameterize continental lithosphere that is chemically buoyant and strong to the degree that it is not recycled into the convecting mantle, nor is it deformed by convection. The rheology assumed for the mantle is the same as that of the previous subsection.

Two key unknowns, for which we will develop theoretical expressions, are the average temperature at the base of the continental lithosphere, T_c, and the average internal temperature of the mantle, T_{ic}. As the continental lithosphere does not participate in mantle overturn its presence can reduce the effective Rayleigh number driving mantle convection, Ra_{eff}, by reducing the temperature difference across, and the overall depth of, the convecting layer. Conversely, the insulating effect of the continental lithosphere can increase Ra_{eff} by decreasing internal mantle viscosity, μ_{ic}. We thus define the Rayleigh number driving convection as $Ra_{\mathrm{eff}}=\rho_{0}g\alpha(T_{\mathrm{b}}-T_{\mathrm{c}})(D-d)^{3}/\mu_{\mathrm{ic}}\kappa$ where ρ_0 is reference mantle density, g is gravitational acceleration, α is the thermal expansion coefficient, D is mantle depth, and κ is thermal diffusivity. As Ra_{eff} is not known a priori, it will be useful to consider a Rayleigh number, Ra_{s}, based on the system temperature drop, ΔT, and the mantle viscosity defined using the surface temperature, μ_{s}. The two Rayleigh numbers are related by

$$Ra_{\mathrm{eff}} = \frac{(T_{\mathrm{b}} - T_{\mathrm{c}})}{\Delta T}(1 - d/D)^{3}\frac{\mu_{\mathrm{s}}}{\mu_{\mathrm{ic}}}Ra_{\mathrm{s}}. \tag{4}$$

By expressing Ra_{eff} in terms of μ_{ic} we are assuming the dominant resistance to sub-continental mantle convection is coming from the internal viscosity of the mantle. This implies that T_{c} is sufficiently high that the base of continents is within the rheological transition layer or that plastic failure occurs within the cold upper boundary layer of the mantle that forms below continents. Our approach does not rule out exploring the alternative possibility that a cold stagnant layer forms below chemically distinct continental lithosphere. For such a case, d will parameterize the thickness of chemically buoyant lithosphere plus the thickness of the stagnant mantle sublayer.

Heat source enrichment within continental lithosphere can be approximated by allowing the continental lid to have a relatively low thermal conductivity [32]. Heat flux into the continental base, q_{c} can then be equated to surface heat flux leading to $q_{\mathrm{c}}=K_{\mathrm{c}}T_{\mathrm{c}}/d$. A linear thermal gradient is assumed to hold across the active mantle boundary layer, of thickness δ_{c}, that forms below continents. Mantle heat flux, q_{m}, can be written as $q_{\mathrm{m}}=[K_{\mathrm{m}}(T_{\mathrm{i}}-T_{\mathrm{c}})]/\delta_{\mathrm{c}}$. Equating q_{m} and q_{c} leads to

$$\delta_{\mathrm{c}} = \frac{(T_{\mathrm{ic}} - T_{\mathrm{c}})K_{\mathrm{m}}d}{T_{\mathrm{c}}K_{\mathrm{c}}}. \tag{5}$$

We introduce a local boundary layer Rayleigh number, $Ra_{\delta\mathrm{c}}$, and assume it to remain near a constant value [26]. The critical Rayleigh number for convective onset is $\approx 10^{3}$ for a range of boundary conditions [33]. To allow for variation [34], we consider $Ra_{\delta\mathrm{c}}=a_{1}10^{3}$ where a_1 is a scaling constant. We thus have a second, independent, expression for δ_{c} given by

$$\delta_{\mathrm{c}}^{3} = \frac{a_{1}10^{3}D^{3}\Delta T}{(T_{\mathrm{ic}} - T_{\mathrm{c}})Ra_{\mathrm{eff}}}. \tag{6}$$

Continental lithosphere will affect internal mantle temperature through its thermal effects, which in our theory are expressed in the determination of T_{c}. There is also a mechanical effect that must be accounted for. That is, the mantle below stable continents is in contact with a rigid boundary. In contrast, the core-mantle boundary presents a free-slip mechanical boundary condition to mantle convection. Thus, an asymmetry is imposed on the mantle even in the limit of d going to zero and this will affect internal mantle temperature. The next paragraph considers this limit to isolate the effect of the mechanical asymmetry. From there we will reintroduce a finite thickness continental lithosphere to derive a final expression for internal mantle temperature.

Consider a mantle layer with a rigid surface and a free slip base. The temperature drops across the upper and lower boundary layers are denoted by ΔT_{u}, and ΔT_{l}. Their thickness by δ_{u} and δ_{l}. For high Ra, boundary layer thickness scales with the boundary layer Rayleigh number to the -l/3 power independent of mechanical conditions [26]. At low to intermediate Ra, numerical simulations and laboratory experiments have shown that for rigid boundary conditions the scaling exponent is less than 1/3 for both low and high Prandtl number fluids [35]. However, low Prandtl number experiments show that at high Ra the exponent does approach l/3 for rigid conditions [36]. Our own infinite Prandtl number simulations also show this trend (Fig. 2). Our focus is on high Ra convection and we thus assume the l/3 scaling which provides an absolute upper bound on heat flux in the infinite Prandtl number limit [37,38]. We note that this will lead to errors at lower Ra (Fig. 2). We further assume that the boundary layer Rayleigh numbers remain near constant values [26,34] and that the values are proportional to the critical Rayleigh numbers for convective onset (i.e., 1707 for rigid conditions and 657 for free conditions). The effects of boundary layer interactions on stability are not considered and this is another error source at low Ra. Making the noted assumptions and balancing heat flow across boundary layers leads to $\Delta T_{\mathrm{u}}/\Delta T_{\mathrm{l}}=b_{1}[1707/657]^{1/3}$ where b_1 is

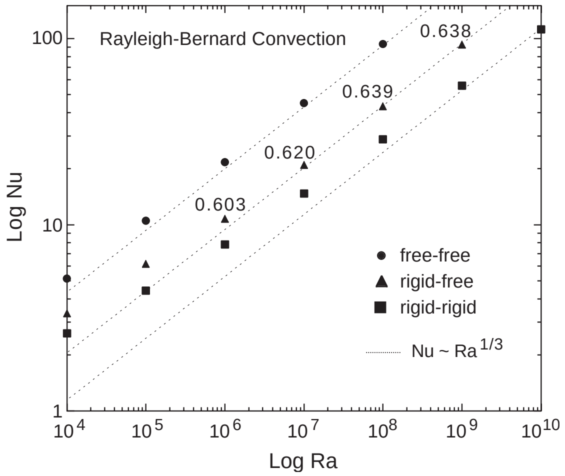


Fig. 2. Non-dimensional surface heat flux versus Rayleigh number for numerical simulations of Rayleigh–Benard convection in a 1×1 Cartesian domain with rigid/rigid, rigid/free, and free/free mechanical boundary conditions at the upper/lower surfaces of the system. Theoretical trends that predict a heat flux scaling with the Rayleigh number to the 1/3 power are also plotted [24] as are the average internal temperatures for the rigid/free cases.

a scaling constant of order unity. Numerical simulations constrained b_1 to 1.3 for the higher Ra cases explored (Fig. 2). Thus, a rigid surface and a free slip base cause the internal temperature of a convecting layer to be 0.64 times its base temperature plus 0.36 times its surface temperature.

We now reintroduce a finite thickness continental lid to the analysis above. The lid does not participate in convection so the surface temperature of the convecting layer is the lid base temperature, T_c, and the internal temperature of the convecting layer is then given by

$$T_{ic} = 0.64T_b + 0.36T_c. \tag{7}$$

Eqs. (4) (5) (6) and (7), together with rheologic assumptions, lead to an expression for T_c. The expression is simplified by nondimensionalizing with D as the length scale, K_F as the conductivity scale, and ΔT as the temperature scale (dimensionless surface and base temperatures are set to 0 and 1). The final expression is

$$\frac{(1-T_c)^5}{T_c^3\exp[-.36\theta T_c]} = \frac{a_1 5960.46K_c^3\exp[-.64\theta]}{(d-d^2)^3 Ra_s} \tag{8}$$

This allows us to solve for surface heat flux using the expression $q_c = K_c T_c / d$.

2.3. Linked continent–ocean system

We consider the linked system with both oceanic lithosphere that actively partakes in convective mantle overturn and continental lithosphere with a stable upper portion of thickness d that does not partake in convective overturn. We define A_c, and A_o, as the ratio of the planet's surface covered by continents and oceans, respectively (i.e., $A_c + A_o = 1$). The previous two subsections provide expressions for the average internal temperatures for the end-member limits of $A_c = 0$ and $A_o = 0$. We assume thorough thermal mixing such that the average internal temperature of the full thermal network, T_i, can be taken to be a volume weighted average of the internal temperatures for each end-member, T_{ic}, and T_{io}. Considering surface and volume ratios to be proportional leads to

$$T_i = A_c T_{ic} + A_o T_{io}. \tag{9}$$

End-member cases of a planet covered by, or devoid of, continents can be associated with thermal resistances of R_c, and R_o. Intermediate cases will have a total network resistance, R_t. Total system heat flux, q_t, can thus be expressed as $q_t = \Delta T_i / R_t$ where ΔT_i is the average temperature drop associated with heat transfer from the interior mantle to the surface. Similarly, for the end-members $q_c = \Delta T_c / R_c$ and $q_o = \Delta T_o / R_o$.

We consider oceanic and continental paths to be linked in parallel and the effective resistance of each path to decrease as its relative surface area increases. As well as surface area effects on local resistance, the effects of internal temperature must also be considered. The temperature of the full network is greater than that of the oceanic end-member and less than that of the continental end-member. Thus, the internal viscosity of the system will differ from that of either end-member. This, in turn, can alter the local resistance of the oceanic and continental path within the full system as compared to end-member cases (Fig. 1). We consider the effects of the oceanic path first. Higher temperature reduces internal mantle viscosity which facilitates more rapid overturn of oceanic lithosphere. Thus, the resistance of the oceanic path in the network will be less than its value for the oceanic lithosphere only end-member case. To account for this, we consider the internal viscosity for the oceanic end-member, $\mu_{io}=A\exp[-\theta T_{io}]$, and for the network, $\mu_i=A\exp[-\theta(A_cT_{ic}+A_oT_{io})]$. In a plate-like regime, surface heat flux for the oceanic end-member was found to scale inversely with the internal viscosity to the 0.293 power [27]. This scaling, together with the ratio of μ_i to μ_{io}, allows a weighting factor to be defined by $(\exp[\theta A_c(T_{io}-T_{ic})])^{0.293}$. This nondimensional factor, multiplied by R_o, accounts for the reduction in the thermal resistance of the oceanic path due to the higher internal mantle temperature and lower internal viscosity associated with continental insulation.

The internal mantle temperature T_i for the full network will be lower than that for the continental end-member T_{ic}. This will increase mantle viscosity and increase the resistance of the continental path in the linked network relative to its value for the continental lithosphere only end-member case. As the expression for heat flux for the continental end-member involves a transcendental equation, Eq. (8), a simple expression cannot be written for this effect but it can be treated using an iterative approach. However, we already know that for the continental path temperature effects on thermal resistance will be weak relative to oceanic path. This is because the continental path is itself a composite of two resistance in series (Fig. 1). If the thermal resistance of stable continental lithosphere is much greater than that of the mantle sublayer below, then changes in internal mantle viscosity will have a negligible effect on the continental resistance. The thermal resistances of stable continental lithosphere relative to the mantle sublayer can be expressed as a local Biot number given by $(dK_m)/(\delta_cK_c)$. As noted in the continental end-member subsection, we consider $K_m/K_c>1$. For high Rayleigh numbers, d/δ_c will also be greater than one. In writing closed form expressions for the linked theory we assume that the high Rayleigh number limit holds so that the resistance of stable continental lithosphere dominates the total resistance of the continental path.

The assumptions above lead to an expression for the inverse resistance of the system given by

$$R_t^{-1} = R_c^{-1}A_c + R_o^{-1}A_o(\exp[\theta A_c(T_{ic}-T_{io})])^{0.293}. \tag{10}$$

The average system heat flux is then given by

$$q_t = A_cq_c\left[A_c + A_o\frac{T_{io}}{T_{ic}}\right] + A_oq_o\left[A_c\frac{T_{ic}}{T_{io}} + A_o\right](\exp[\theta A_c(T_{ic}-T_{io})])^{0.293} \tag{11}$$

where the surface temperature is set to 0. Notice that an increase in total heat flux, with added continental insulation, can occur if the increase in oceanic heat flux is great enough and if there is sufficient oceanic surface area. The equation shows that increased oceanic heat flux is due to two effects. The first is an increased temperature drop across the oceanic lithosphere. The second is reduced sublithospheric viscosity allowing for faster plate speeds and thinner oceanic lithosphere. This latter effect depends on our rheologic assumptions while the former effect does not.

Eq. (11) predicts global heat flux and local heat flux for the continental and oceanic paths within the network. The latter, together with T_i, is used to calculate δ_o, while μ_i is determined from the viscosity law. The stress scaling of Eq. (3) is used with δ_o and μ_i to determine when convective stress levels within the oceanic lithosphere drop below τ_{yield}. Beyond this point a stagnant lid mode of convection will operate and stagnant lid scalings can be used to predict heat flux [28]. The analysis for the full network is complete.

2.4. Additional considerations

In developing the theory we have made 3 significant assumptions that influence the heat transfer properties of the system. First, we have assumed thorough thermal mixing. The other extreme of no thermal mixing can also be considered. For such a case, the total system heat flux can be expressed as a simple area weighted average of the heat flux from each end-member, i.e., $q_{tnm}=A_cq_c+A_oq_o$ where the added subscript nm refers to no thermal mixing. It should be clear that for this case continental insulation will always decrease global mantle heat flow.

Second, we have assumed that, in an active lid mode of convection, plate boundaries are very weak [18,19]. Increased boundary zone strength can be introduced into our theory by shifting the exponent in

Eq. (2) from a value near l/3 to a lower value, which depends on the mechanical properties of the lithosphere [39]. This change would be incorporated into the linked system so that the exponent in Eq. (11) would also be lowered. Inspection of Eq. (11) shows that, for the case with strong plate boundaries, a stronger temperature dependence of mantle viscosity would be required for continental insulation to increase global mantle heat flow by the same level as for the case with weak plate boundaries.

Third, we have not explicitly considered internal heating. This reduces potential complexity and facilitates testing via laboratory experiments for the end-member case of a bottom heated mantle. For application to the Earth, however, the robustness of specific conclusions must be tested for the alternate end-member case of an internally heated mantle. We have done this via numerical simulations as discussed in the section to follow.

3. Numerical simulations

To test the theory, predictions are compared to 2-D numerical simulations. Simulations are similar to those of [27] except that a conducting surface layer is included to represent stable continental lithosphere [40]. For a simulation suite, the yield stress is set to 66% of the active to stagnant lid transition value assuming no continent present [27]. The thermal field from the no continent case is then used as the initial condition for a case that imposes a continent with a nondimensional lateral extent of 0.05. The calculation is run to a statistically steady-state and then used as the initial condition for the next case which increases the extent by 0.05. Thermal mixing within the mantle was achieved within one to two mantle overturn times (this does depend on the specific manner in which we "grew" continents and greater additions of continental material and/or thicker continental lithosphere than we employed could lead to longer mixing times). Numerical simulation suites were performed that imposed continents either above a zone of mantle upflow or a zone of downflow. For a compact description of parameter regimes, it will be useful to consider the effective thermal thickness of stable continental lithosphere, d_T, which is defined as $(dK_\mathrm{m})/(DK_\mathrm{c})$. Fixed parameters for any suite of simulations are d_T, the yield stress, the basal Rayleigh number, Ra_b, and the degree of temperature-dependence of mantle viscosity, θ (cf. Eq. (1a)).

Fig. 3a shows a transition in heat transport with increasing continental extent. Heat flux changes from relatively high values associated with plate-like convection to low values associated with a stagnant lid regime. Fig. 3a also shows theoretical predictions for heat flux and the critical continental extent that locks plate boundaries. For any suite, the reference stress is taken as the stress with no continent present. The continental extent at which the theory predicts convective stress to drop to 66% of this reference value is the point at which a stagnant lid regime is predicted. We consider this the point at which transition is complete. Although we can predict when a stagnant lid holds, the transition between the two regime limits is broad in the simulations. For Fig. 3a, the point at which the transition initiates, based on a pronounced decrease in heat flux, was at a stress value of approximately 75% of the reference stress. For subsequent testing we will use this criterion to determine transition initiation.

Fig. 3b compares results from several simulation suites to theoretical predictions. Simulation curves are shown up to the point at which any given suite has just entered the transitional regime. The theory predicts that for higher mantle Rayleigh numbers, global heat flow should increase more rapidly with continental extent and this is confirmed by the numerical simulations (Fig. 3b). Higher velocities associated with increasing Rayleigh numbers lead to a thinner mantle sublayer below continents. This, in turn, increases the effect stable continental lithosphere has on determining the thermal resistance of the continental path. In effect, the relative insulating power of continental lithosphere increases leading to greater internal temperature. The extreme thinning of the continental sublayer makes higher Rayleigh number simulations than shown demanding as very dense meshes are required to numerically resolve the mantle sublayer below continents.

The theory also predicts that, as mantle viscosity becomes more temperature dependent, heat flow should progressively flatten and then increase with continental extent. This is also confirmed by the numerical simulations (Fig. 4a). Notice that the isoviscous version of the theory and isoviscous

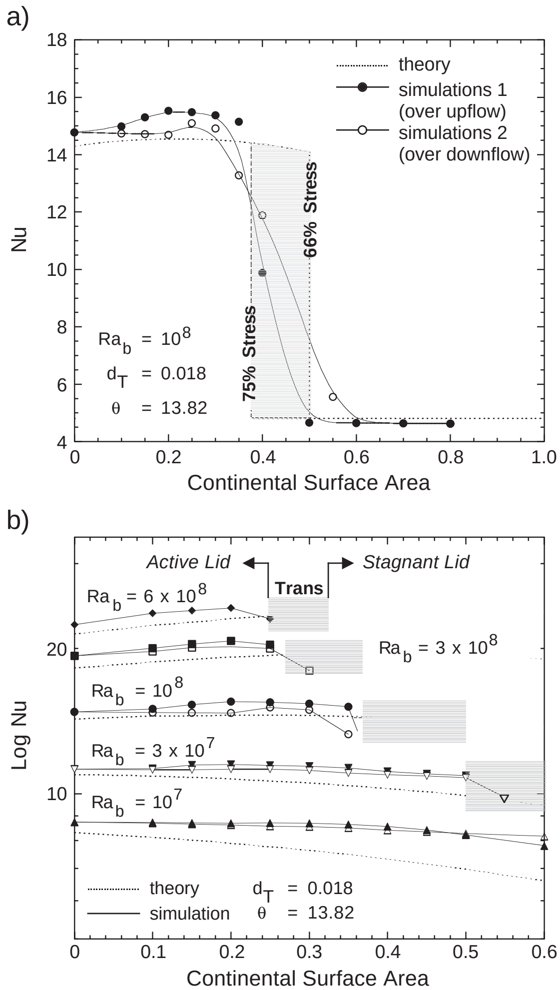


Fig. 3. Comparison of theory predictions to 2D numerical simulation results in a unit aspect ratio Cartesian domain. For the higher *Ra* cases, 256 × 256 finite element meshes were used to assure converged solutions. Results over the full range of behavior, from a partially insulated active lid to a transitional regime to a stagnant lid regime, are shown in (a). For (b), results are shown up to the point that simulation suites entered into the transitional regime. The scaling constant, a_1, was constrained to a value of 0.1 by direct comparison of theoretical predictions to simulation results. Two different simulation suites, 1 and 2, were performed for any given set of parameters to test model sensitivity to the end-member assumptions that continents preferentially grow and reside over zones of thermal mantle upwelling or over zones of thermal mantle downwelling.

simulations show a decrease in global heat flux with increasing continental extent. This follows the standard assumption that continental insulation should inhibit mantle heat loss. The degree of temperature dependence that leads to increasing heat flow with continental extent is mild compared to estimates for the Earth's value. For the Earth, θ is estimated to be between 20–45 for a range of potential creep mechanisms [27,28]. For such large degrees of temperature dependence, the theory predicts that continental insulation could increase global mantle heat flow by 25% (Fig. 4b). Simulations with more extreme temperature dependent viscosity than shown

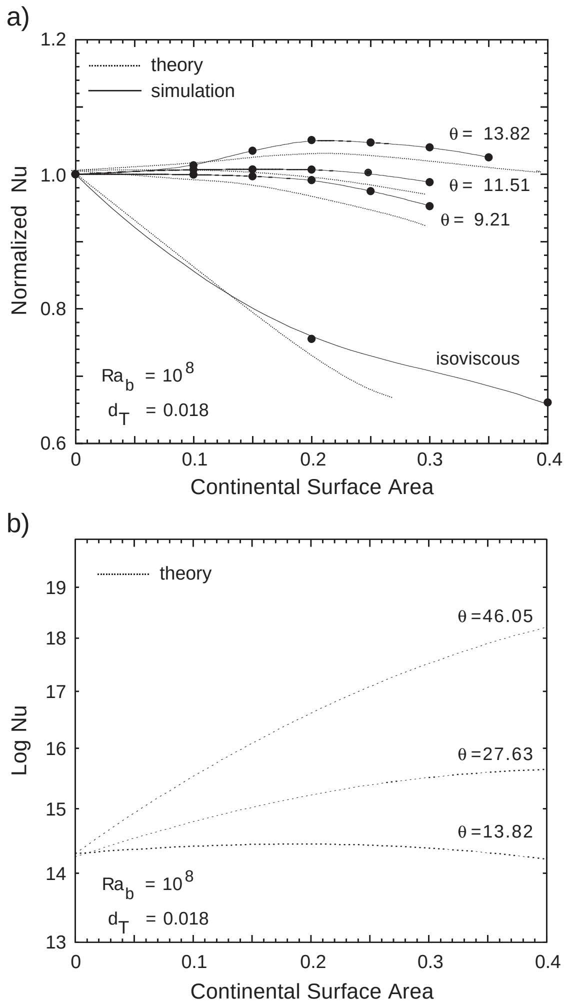


Fig. 4. (a) Normalized system heat flux versus percentage of continental lithosphere for four numerical simulation suites with different degrees of temperature-dependent viscosity compared to theory predictions. (b) Theory predictions for nondimensional surface heat flux versus percentage of continental lithosphere for mantle viscosities with stronger degrees of temperature-dependence.

are numerically demanding, particularly with a visco-plastic rheology, and we could not achieve converged, bottom heated solutions on finite elements meshes as dense as 256×256 elements over a unit aspect ratio domain for θ values approaching 20. Thus, the theoretical predictions at large degrees of temperature dependence remain untested at this stage.

The theory also predicts that increased thickness of stable continental lithosphere can increase mantle heat flow and this too is confirmed by the numerical simulations (Fig. 5a). Fig. 5b compares simulations that model continents as immersed conducting blocks, as done thus far, versus those that model continents as surface boundary condition zones of zero heat flux (i.e., perfect insulators). The latter approach can expedite numerical solutions. The two approaches lead to large relative differences for predicted heat flux in continental regions but relatively small differences for oceanic and global heat flow. The inset plot shows that, although increased insulation leads to only a small increase in global heat flow, it leads to a large increase in the velocity of the system. This is due to the fact that oceanic plates overturn at a faster rate with increased insulation. Imposing continents as zones of zero heat flux and having global heat flow remain unchanged also highlights how strong an effect partial insulation is having on oceanic heat flux; for the surface boundary condition simulations, adding 40% insulation increases oceanic heat flux by 66% (Fig. 5b).

The thermal mixing component of our theory leads to the prediction that the temperature below insulating continental lithosphere should not be appreciably greater than the mantle temperature, at equivalent depth, below oceanic lithosphere. This is confirmed by our own simulations and is in accord with previous simulations that explored the effects of continental insulation on lateral temperature variations in the mantle [41]. It should be re-stressed that our theory considers time averaged values once the system is in statistically steady state. Thus, the theory predicts that, over the time scale for which a near constant or mildly varying mantle Rayleigh number holds, the time averaged temperature below continents should not be elevated relative to the mean mantle temperature. The time scale over which the mantle Rayleigh number does not change appreciably, due to decaying heat sources, is often referred to as a secular time scale

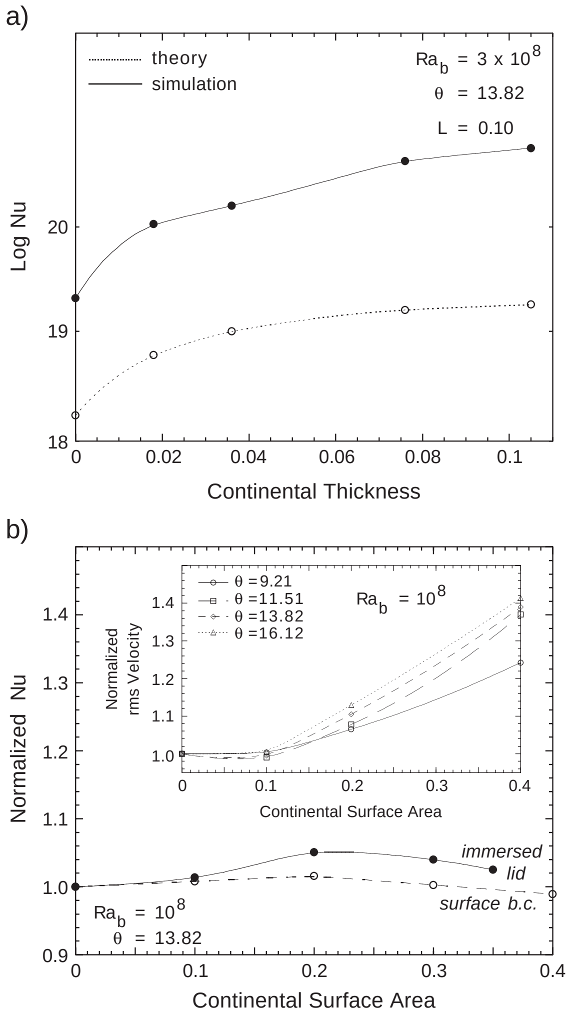
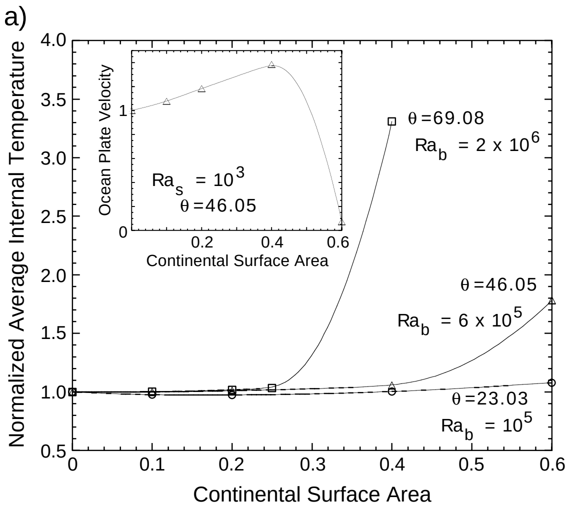


Fig. 5. (a) Comparison of theory predictions to 2D numerical simulation results for heat flux versus the thickness of stable continental lithosphere. (b) Comparison of normalized system heat flux versus percentage of continental lithosphere from numerical simulation suites that impose continents as immersed conducting blocks versus those that impose continents as surface boundary conditions zones of zero heat flux. The inset graph shows the effects of increasing continental extent on the normalized root mean square velocity of the mantle for simulation suites that impose continents as zone of zero heat flux.

in thermal history studies and it is estimated to be $\approx 10^9$ years [42,43]. Our results do not rule out the possibility that transient increases in mantle temperature can occur below continents on time scales more rapid than the secular time scale [7–9,44–46]. Such transient increases can most likely be associated with the Wilson cycle of supercontinent assembly and dispersal [9,46]. The Wilson cycle itself, provided it operates at a faster scale than the secular time scale, can potentially promote the thermal mixing required for our theory.

We have tested the robustness of two key conclusions via numerical simulations that consider an internally heated and top cooled layer with a variable extent of surface insulation. For these tests, continents were imposed as surface boundary condition zones of zero heat flux. The surface temperature outside of continental regions was fixed to a nondimensioanl value of zero. The basal thermal boundary condition was adiabatic as were side walls. Mechanical boundary conditions were free slip. The cooling efficiency of convection for an internally heated layer cannot be gauged by surface heat flux. Instead, the average internal temperature is an indication of convective efficiency. Fig. 6 shows the results of three internally heated simulation suites. The inset plots of Fig. 6a and b confirm that for internally heated convection, in a temperature-dependent material, partial insulation can significantly increase the average surface velocity in the noninsulated region. This allows for greater internal cooling via oceanic plate subduction which offsets the effects of partial insulation. As a result regimes can exist in which increased insulation does not decrease interior cooling (Fig. 6).

A second robust conclusion is that a critical extent of continental insulation exists that can cause convective stress to drop below the lithospheric yield stress and this critical extent is a decreasing function of convective mantle vigor. For internally heated convection, the basal temperature is not known a priori and thus neither is the basal Rayleigh number. Our simulation suites all have the same fixed yield stress and surface Rayleigh number. The suites increased the temperature dependence of mantle viscosity, i.e., increased θ. A higher θ leads to lower internal viscosity at equivalent temperature which can increase the internal and basal Rayleigh number. The basal Rayleigh numbers noted by the simulation suites of Fig. 6a are for reference cases free of continents. The higher θ cases could be resolved numerically as they lead to milder viscosity gradients for internal versus bottom heated convection (this is because the nondimensional temperature drop across the internally heated simulations was much less than that of the bottom heated simulations). The dramatic increase in internal temperature shown in Fig. 6a results when convective stresses drop below the yield stress and initiate a stagnant lid mode of convection (Fig. 6a,



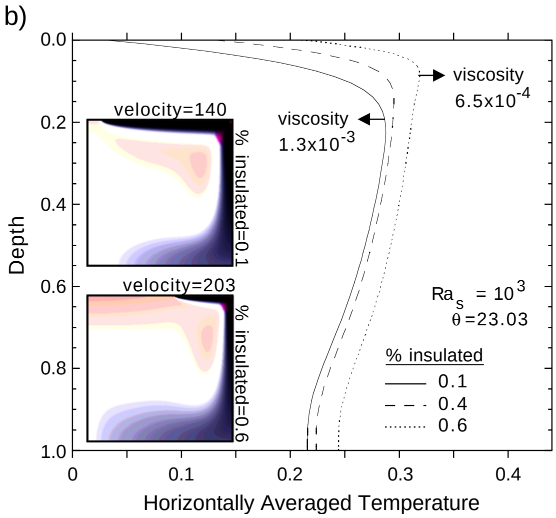


Fig. 6. Results from internally heated, partially insulated convection simulations. (a) Normalized average internal temperature versus percentage of continental lithosphere for simulation suites with different degrees of temperature dependent viscosity and different mean internal Rayleigh numbers. The inset displays normalized oceanic plate velocity versus percentage of continental lithosphere. (b) Average system geotherms from simulations with different degrees of continental insulation. The nondimensional viscosity below the upper thermal boundary layer is noted for two of the simulations (the surface viscosity is nondimensionalized to a value of unity). The inset shows thermal fields from two of the simulations along with average horizontal surface velocity.

inset). Consequently, the internal temperature rises until conduction across the stagnant lid can remove internally generated heat. The suites of Fig. 6a shows that, as for bottom heated convection, the critical continental extent that locks plate margins, in an internally heated mantle, decreases with increasing basal and bulk internal Rayleigh number.

4. Laboratory experiments

Although we have focused on continents and the mantle, the broader idea we have developed is that partial insulation of a convecting, temperature-dependent viscosity fluid can increase the heat flow out of fluid. To test the general validity of this idea, we conduct a series of laboratory experiments aimed at investigating the influence of a variable-width insulating lid (an analog "continent") on the heat transfer characteristics of thermal convection in a fluid with a temperature-dependent viscosity (Fig. 7a). We stress that the laboratory experiments provide a separate means of testing the general validity of our theoretical ideas. They are not designed for direct comparison to the numerical simulations.

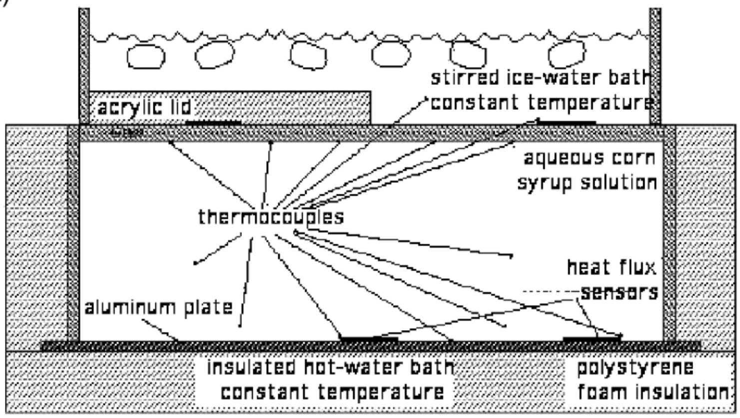
An aqueous corn syrup solution with a Newtonian viscosity that increases exponentially with decreasing temperature is heated from below and cooled from above. Viscosity varies by five orders of magnitude across the fluid layer. The surface bath is kept at a constant temperature of 273 ± 0.2 K while the hot bottom boundary temp was 337 K. The viscosity for the syrup solution, in Pa s, is given $\mu = \exp(a/T^2 + b/T + c)$, where T is the absolute temperature and $a = 8.85 \times 10^6$, $b = -4.18 \times 10^4$, and $c = 46.4$. An insulating lid is applied to the cold boundary. We characterize the influence of this boundary condition on the flow at thermal equilibrium with five external dimensionless parameters: (1) the Rayleigh number, 10^5, where the viscosity is based on the mean of the temperatures at the hot and cold boundaries; (2) the ratio of time scales for thermal to viscous diffusion 10^4; (3) the lid extent relative to the tank width, L; (4) the ratio of the thermal resistance of the lid to that of the glass, and $d_L K_g / d_g K_L = 9.3$; (5) the system aspect ratio of 4.2. Time-lapse video, shadowgraphs, still photographs, measurements of the local and global surface heat flux, floor and roof surface temperatures and interior fluid temperatures as a function of time are used characterize the flows quantitatively.

Measurements of the average heat flux across the layer (Fig. 7b) show that when $0 < L < 0.4$ (the "partial lid" case) heat transfer is increased relative to the no lid case, which is consistent with the theoretical results above. Because the flows and surface heat flow are unsteady, we plot the mean Nu (Nusselt number) and the errors show two standard deviations of the instantaneous Nu. Time-lapse video and analysis of time series of temperature data show that in the partial lid case horizontal temperature variations at the roof drive a tank-filling large-scale flow that enhances unsteady vertical motions from the hot and cold boundaries. In contrast, when $0.4 < L < 1$ (the "long lid" case) the flow is composed of two parts: beneath the lid, convection is governed by thermals ascending (or descending) from the hot (or cold) boundary; beneath the adjacent gap, thermals are present but the flow is dominated by approximately two-dimensional overturning motions that stir a region comparable in width to the gap.

Fig. 7b shows two theoretical curves for the nondimensional global heat flux or Nusselt number, Nu, as a function of L. For the "partial lid" curve (dashed line), the lid and gap heat transfer paths are assumed to be linked in parallel following our theory. In particular, the gap and lid sides are considered to be stirred by a tank-filling, large-scale flow so that the assumption of thorough thermal mixing, Eq. (9), holds. The solid curve is relevant to the long lid case and is obtained by assuming that both paths are thermally and mechanically independent of one another (poor thermal mixing). Both curves recover the end-member $L = 0$ and $L = 1$ case. The partial lid theory captures the initial increase in heat transfer with L while the long lid theory is consistent with data for $0.4 < L < 1$. The experimental results support the conclusion that a partial lid can enhance heat transfer in convecting systems.

5. Discussion

Scaling theory, numerical simulations, and laboratory experiments all support the hypothesis that continents, acting as thermal insulators above a convecting mantle, can enhance the global rate of a)



b)

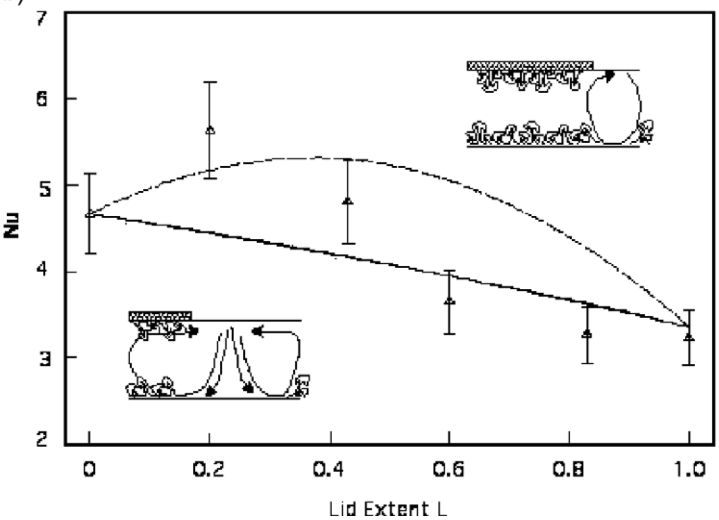


Fig. 7. (a) An illustration of our experimental apparatus. (b) Nusselt number versus lid extent for our experiments. Theoretical curves assume that the lid and gap heat transfer paths are linked in parallel (dashed line) or are independent of one another (solid line). Here, $Nu=0.45Ra^{0.28}$ for the no lid case. The temperature drop across the fluid layer and a viscosity based on the mean temperature in the fluid are used to define Ra. For the solid curve, an Ra_{eff} and corresponding Nu are obtained for each side and a global Nu reflects the weighted sum of the lid and gap sides. The dashed curve assumes that horizontal temperature variations at the roof drive a tank-tiling large scale flow that causes thorough thermal mixing of the fluid interior. Consequently, Ra_{eff} for the gap side is based on a viscosity at the mean interior temperature of the convecting system.

mantle heat loss. Even for situations in which the global increase is small, the increase in oceanic plate velocity and oceanic heat flux due to continental insulation is large. This is our most general result and it is a nonintuitive one: Regimes exist in which increased insulation above a thermally convecting layer does not keep heat in and can even help it get out, i.e., insulation can increase cooling rate. The key to understanding this result is treating the solid Earth as a thermal network with oceanic and continental paths linked nonlinearly in parallel. With rare exception [47], thermal history studies have not directly considered the thermal linking of oceanic and continental lithosphere. As an example of the potential significance of considering the linked system we note that the issue of continental growth altering sea-level has, to date, been addressed by only considering continental growth to lower the area of ocean basins [e,g,10,48,49]. Enhanced oceanic heat flux, due to continental insulation, is expected to reduce the average thickness of the oceanic lithosphere and, thus, reduce the average depth of ocean basins. This would have an added effect on sea level that has not been explored to date.

More specific conclusions are speculative at this stage but we can provide a preliminary assessment of one key implication of our theory: the existence of a limiting mechanism for the area of continental lithosphere. Increased continental surface area is predicted to enhance the velocity and subduction rate of oceanic lithosphere. If continents are produced in subduction settings this could enhance continental growth. However, continental insulation is also predicted to lower convective stress levels. If convective stress approaches the lithospheric yield stress, then weak plate margins will be difficult to maintain. This will lower the subduction rate of oceanic lithosphere and limit further continental growth. We can assess the viability of this idea by using our theory to predict the critical surface area and the self-consistently determined mantle heat flux for a range of potential parameter values. If the surface area and mantle heat flux are both in accord with present day observations then the existence of a limiting mechanism is viable.

To test this hypothesis we assume that some form of plate tectonics initiated before continents formed. Initiation occurs when convective stresses exceed the lithospheric yield stress. This critical convective stress depends on mantle flow velocities and, thus, can be related to Ra and the nondimensional vertical heat flux, Nu. At present $Nu \approx 40$ (for a mantle heat flux of 60 mW/m^2, a thermal conductivity of 3 W/mK, a lithospheric temperature drop of 1450 K, and whole mantle convection). If plate tectonics initiated early in Earth's geologic history, i.e., ≈ 4 bya, then the global heat flux was roughly 3–4 times higher than at present [42,43]. Assuming a heat flux value at plate tectonics initiation effectively fixes τ_{yield}. That is, a smaller initiation heat flux implies a later start time for plate tectonics and, by association, a higher lithospheric yield stress. Conversely, a higher initiation heat flux implies an earlier start and a lower lithospheric yield stress. For any parameter suite of models, we consider the yield stress to remain constant from initiation onward. As the Earth cools the Rayleigh number decreases as the internal viscosity of the mantle increases. For a thermal evolution model free of continents, convective stress, which scales with mantle viscosity, would remain above the yield stress. For the situation with continents, our theory predicts that this will not be the case if the surface area of continents exceeds a critical extent, which we can predict for any combination of continental thickness, viscosity law, initiation time of plate tectonics, and degree of convective vigor expressed through Ra.

Fig. 8a shows the predicted critical continental area, A_{cc} versus Nu for different assumptions as to plate tectonic initiation, i.e., different assumptions as to lithospheric yield stress. Parameter set curves all have a portion associated with increasing A_{cc} as Nu decreases and a second portion associated with decreasing A_{cc} as Nu decreases. For the former, the A_{cc} values represent the point at which initiation to a transitional regime between plate tectonics and a stagnant lid is predicted (Fig. 3). For the latter, the A_{cc}, values represent the point at which greater continental surface area is predicted to cause a decrease in global heat flow due to insufficient oceanic surface area. The area limit is plotted to fully delineate the conditions under which continental insulation can increase global mantle heat loss. It also delineates parameter regions where an added heat loss mode, not incorporated into our theory as it stands, can become significant. For regions above the portion of the curves where A_{cc} decreases with decreasing Nu, our theory predicts the potential of very large internal mantle temperatures due to low global heat flow and relatively large continental area. For many of the parameter sets the predicted temperatures imply large scale melting and heat loss due to melt transport would then need to be considered within the theory. Fortunately for application to the Earth the point at which this is predicted to occur in $Nu-A_{\mathrm{cc}}$ space is for continental surface areas greater than the observed present day value and/or for global heat fluxes lower than the observed present day value.

Fig. 8b shows the predicted critical continental area versus Nu assuming plate tectonics initiated at an Nu value ≈ 4 times the present day value [42,43]. Although there is a range of potential behavior, as the temperature-dependence of mantle viscosity becomes large, the spread in the parameter curves narrows, particularly for cases with larger d_{T}. For

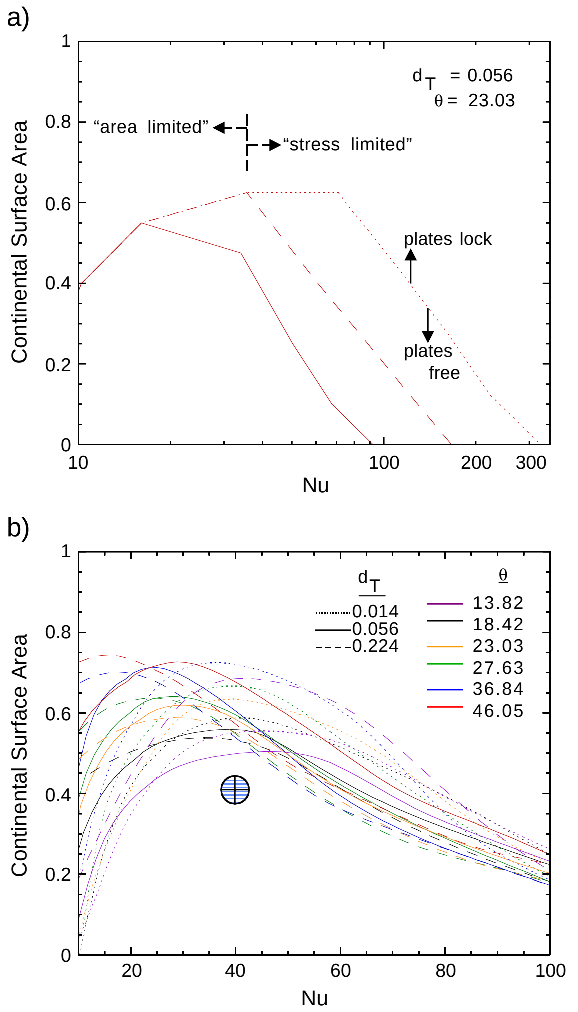


Fig. 8. (a) Predicted critical continental surface area versus predicted nondimensional heat flux, *Nu*, for different assumptions as to the initiation time of plate tectonics. For parameter regions above the portion of the curves labeled "stress limited", increased continental area would initiate a transition between plate tectonics and a stagnant lid mode of mantle convection. For regions above the portion of the curves labeled "area limited", increased continental area would cause a decrease in global heat flow. (b) Predicted critical continental surface area versus predicted *Nu* for variable d_T and θ. The crossed circle is an estimate of the present day of Earth values.

those cases the curves tend to collapse onto the lower end of the envelope of behavior shown in Fig. 8b. For that portion of the full envelope, our theory predicts that, for the Earth's current global heat flux, the critical continental surface area should be 35–50% for a yield stress range that allows plate tectonics to initiate at a global heat mantle flux between 2–4 times the present day estimate. The theory also predicts the partitioning of mantle heat flow between oceanic and continental regions. The predicted subcontinental mantle heat flux, for the noted parameter range, is 10–20 mW/m^2 which is in accord with values determined from heat flow measurements [13–16]. The predicted oceanic heat flux is 90–110 mW/m^2 which is also in accord with heat flow measurements [17]. Although not conclusive, this result does suggest that the idea of a limiting mechanism on the present day surface area of continents warrants further investigation.

Our theory also predicts that the critical continental surface area should decrease in the Earth's past, consistent with a hotter mantle having a lower viscosity. That is, the theory predicts that the critical area is a decreasing function of convective vigor. This allows us to construct theoretical continental growth curves based on the assumption that continental surface area has remained near a critical value over the Earth's history. To do this, we consider the decay of heat sources to cause the mantle Rayleigh number to drop by a factor of 10 every billion years [20,42,43]. Fig. 9 shows theory predictions as an envelope of behavior for cases with higher degrees of

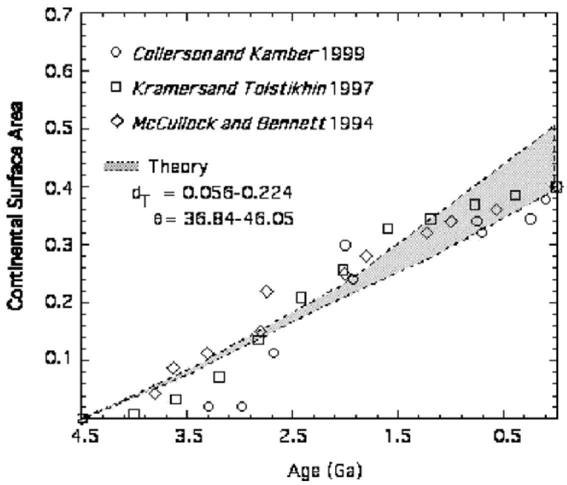


Fig. 9. Predicted critical continental surface area versus time shown as an envelope of behavior that applies to parameter cases with relatively large degrees of temperature-dependent mantle viscosity. Also plotted are surface area versus time trends from several geochemically based continental growth curves.

temperature-dependence. Also shown are several continental growth curves based on geochemical data [50–52]. As can be seen, our theory is not compatible with rapid initial growth of continents [53]. Nonetheless, the consistency between the theoretical envelope and several geochemically based curves suggests that the idea of a limiting mechanism on continental growth also warrants further investigation.

6. Concluding remarks

Theory, simulations, and laboratory experiments support the hypothesis that partial insulation of the convecting mantle, due to the presence of continents, does not lower the rate of mantle heat loss. Provided that the surface area of continents does not exceed a critical extent that causes plates to lock, continental insulation can actually help to cool the Earth by increasing the velocity and subduction rate of oceanic lithosphere. The strong coupling implied between oceanic and continental heat flow paths suggests that the local thermal structure of oceanic and/or continental lithosphere cannot be modeled independently of considering the global ocean-continent system.

Acknowledgments

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