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A climate induced transition in the tectonic style of a terrestrial planet

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ABSTRACT

We explore the possibility that an increase in the surface temperature of a terrestrial planet due to an enhanced concentration of atmospheric greenhouse gasses and/or increased solar luminosity could initiate a transition from an active-lid mode of mantle convection (e.g., plate tectonics) to an episodic or stagnant-lid mode (i.e., single plate planet). A scaling theory is developed to estimate the required temperature change as a function of the temperature dependence of mantle viscosity and the yield stress of the lithosphere. The theory relies on the assumptions that convective stresses scale with mantle viscosity and that a planet will adjust to surface temperature changes so as to maintain a surface heat flow that balances internal heat production. The theory is tested against a suite of numerical simulations of mantle convection. The comparisons are favorable. The combined theory and numerics suggest that if the yield stress for the earths' lithosphere is 30–35 MPa, then a surface temperature change of 60–120° could shut down an active-lid mode of convection assuming present day conditions. Lower values are predicted for higher yield stresses and for earlier times in the earth geologic evolution.

1. Introduction

Volcanic degassing links the interior dynamics of a terrestrial planet to its climatic evolution. A complementary link, in which climate change can affect the solid body dynamics of a planet, has also been proposed. Investigations into the nature of this link were motivated by exploration of the Earth's sister planet Venus.

The large amount of carbon dioxide in Venus' atmosphere has resulted in an extreme greenhouse climate and a surface temperature 450° hotter than that of Earth. Venus shows no evidence of recent plate tectonic activity. Eighty percent of its surface is covered by volcanic plains and interconnected linear deformation features indicative of plate boundaries are lacking. Wrinkle ridges are visible across significant portions of the plains and it has been suggested that the deformation event that formed these features resulted from atmospheric temperature changes (Solomon et al., 1999). Changes in atmospheric temperature can penetrate into the outer rock layer and cause thermal contraction/expansion induced stresses. By coupling a climate model to a model of thermal conduction in the Venusian crust, Solomon et al. (1999) showed that surface temperature changes could lead to stress levels capable of fracturing the upper portions of the solid planet.

The effects of a climatically driven temperature change can extend even deeper. If the change is long lived then its effect can penetrate beyond the lithosphere and into the convecting mantle. Mantle temperature affects melting conditions and thus the possibility exists of a two-way connection between volcanic and climatic history (Phillips et al., 2001). Phillips et al. (2001) explored the implications of this connection through a coupled model of climate and volcanic history. Although Venus shows no signs of geologically recent plate tectonic activity, it has been suggested that it may have experienced an early episode of plate tectonics (Moresi and Solomatov, 1998; Turcotte, 1993) (this has also been suggested for Mars (Sleep, 1994; Nimmo and Stevenson, 2000)). Phillips et al. (2001) explored models with lithospheric recycling, akin to the type associated with plate subduction on Earth, and models in which the entire lithosphere formed a single plate. The factors that could lead to a transition between these end-members were not specifically addressed.

Arguably, the largest potential effect climate change could have on the internal dynamics of a planet is to cause a change in its large-scale mode of surface tectonics. A surface temperature change, over time, can cause a corresponding change in the internal temperature of the convecting mantle. This will lower mantle viscosity which is exponentially dependent upon temperature (as a reference point, a 100° temperature increase results in an order of magnitude decrease in viscosity (Kohlstedt et al., 1995)). For mantle convection in an active-lid regime, akin to plate tectonics, mantle stress scales directly with viscosity (Moresi and Solomatov, 1998). For an active-lid regime to be maintained, convective mantle stress must be capable of reaching the yield strength of the lithosphere (Moresi and Solomatov, 1998; Trompert and Hansen, 1998).

The above suggests that increased surface temperatures could lower convective stress to the point that it approaches the lithospheric

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yield stress. Once this occurs, localized zones of lithospheric failure, i.e., plate boundaries, cannot be continuously maintained and plate tectonics could become episodic or cease altogether. The next section quantifies this idea via a scaling analysis. Scaling predictions are tested against the results from several suites of numerical simulations and some implications for terrestrial planets are discussed.

2. Scaling

We consider a thermally convecting mantle initially at statistically steady state, i.e., surface heat flow balances basal heat flow plus internal heat production. The surface temperature, $T_s(t)$, is spatially constant but can vary in time, t, due to changes in atmospheric conditions. Prior to a surface temperature change the mantle is in an active-lid mode of convection. We assume a viscoplastic rheology in which viscosity depends on temperature and the finite strength of the lithosphere is accounted for (Moresi and Solomatov, 1998).

For stresses below a specified yield, τ_y, the rheology law follows a temperature-dependent viscous branch given by

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

where A and θ are material parameters, T is temperature, and μ is mantle viscosity. For our scaling analysis, we will work with non-dimensional variables. A reference viscosity, μ_0, is defined from Eq. (1a) at the mantle's surface temperature prior to an imposed surface temperature change. Stress is non-dimensionalized by $d^2/\kappa\mu_0$, where d is the mantle depth and κ is the mantle thermal diffusivity. The non-dimensional yield stress is defined as:

$$\tau_y = \tau_{y0} + \tau_{yz} z, \tag{1b}$$

where τ_{y0} is a surface value, τ_{yz} is a depth-dependent term, and z is a non-dimensional depth coordinate. The τ_{yz} term accounts for increased normal stress with depth. It is non-dimensionalized by $(f_c Ra_0)/(\alpha\Delta T)$, where f_c is a friction coefficient, Ra_0 is the mantle Rayleigh number defined using the reference viscosity, and ΔT is the temperature drop across the mantle. At the yield stress, the flow law switches to a plastic branch with an effective viscosity given by

$$\mu_{\mathrm{plastic}} = \tau_y/I, \tag{1c}$$

where I is the second strain-rate invariant. This formulation allows localized zones of lithospheric failure, analogs to weak plate boundaries, to form in a self-consistent manner. The failure zones allow for lithospheric recycling and mantle stirring akin to plate tectonics on present day Earth. If convective stress levels fall below the yield stress, then weak margins cannot be generated and an active-lid mode of convection will cease (Moresi and Solomatov, 1998).

We assume that long-lived surface temperature changes affect internal mantle temperature, T_i, and viscosity, μ_i. By long lived we mean that the spatial and temporal average of a climate induced temperature change must remain as a distinct atmospheric signal for a time scale comparable to the thermal response time of the mantle. An estimate for this scale is the thermal diffusion time across the lithosphere. For present day Earth, this is 10^8 yr and scales with the square of the lithospheric thickness. Our definition of a "long-lived change" is shorter for conditions that promote a thinner planetary lithosphere. The time scale estimate above represents the time required to communicate the change in surface boundary conditions to the mantle. It is not the time for the mantle to come to a new thermal equilibrium state. That later time scale will be longer and will depend on whether the new equilibrium state is one in which active-lid convection remains stable or if it is associated with a transition to an episodic or stagnant-lid state.

We consider the time scale for atmospheric thermal adjustment to solar variability or greenhouse forcing to be short relative to that of the mantle. Variations in atmospheric temperature can then be represented as a step function change in the mantle surface boundary condition from T_s^0 to T_s^1 at a time t_0. A superscript of zero or one indicates values prior to or after a surface temperature change.

We assume that a convecting mantle, subjected to a surface perturbation that lowers its heat loss, will experience an increase in internal temperature due to continued radiogenic heating and/or core heat flux. The mantle will follow a transient, disequilibrium evolutionary path that returns its average surface heat flux to equilibrium. We assume that, along the adjustment path, the internal temperature change will tend toward a final value that is proportional to the surface temperature change. We non-dimensionalize temperature by the non-adiabatic temperature drop across the convecting mantle before a surface temperature change and set T_s^0 to zero. In the limit of vigorous convection, the internal temperature in basally heated, active-lid convection tends toward the mean of the surface and base temperature (Moresi and Solomatov, 1998). Thus, the non-dimensional internal temperature increase due to surface temperature change of T_s^1 will tend toward a value of $T_s^1/2$.

We assume that the main effect of mixed heating is to shift the internal temperature to higher values. That is, the internal temperature prior to a climate change is increased but the effects of the climate change on internal temperature change will not be significantly different than for the bottom heated end-member. Specifically, we consider the following approximate relationship to hold for bottom and mixed heating cases

$$T_i^1 - T_i^0 \sim T_s^1/2. \tag{2}$$

We expect some error from this assumption in that a change in surface temperature will effect the ratio of the bottom to internally heated Rayleigh numbers. This ratio affects the internal temperature of a mixed heated system. The degree of error introduced by neglecting this effect can be assessed through comparisons of our scaling ideas to full numerical simulations.

During its adjustment to surface temperature changes, the mantle will be in thermal disequilibrium. No scaling theory based on statistically steady state behavior (all existing scaling theories for mantle convection) will apply until thermal equilibrium is re-established. Equilibrium scalings cannot constrain the path through which the system passes in response to external forcing, or the time it takes to reach the new equilibrium. They can tell us, however, whether the new equilibrium state is one in which an active lid of convection is or is not favored. The inability of our scaling analysis to constrain disequilibrium adjustment paths will be another potential source of error (as with internal temperature changes, whether the degree of error is tolerable can be assessed through comparison to numerical simulations).

In a plate tectonic mode, convective mantle stress, τ_m, scales as

$$\tau_m \sim \mu_i v/d \tag{3a}$$

where v is the characteristic plate velocity and d is a velocity length scale comparable to the depth of the convecting mantle. Numerical simulations, for bottom heated convection, have confirmed this scaling (Moresi and Solomatov, 1998). We assume that it carries over for basally and internally heated convection. The stress scaling can be expressed in terms of the internal mantle Rayleigh number, Ra_i. In an active-lid mode, velocity scales as $a_1 Ra_i^{2/3}$, where a_1 is a geometric constant that depends on the characteristic wavelength of mantle convection (Moresi and Solomatov, 1998). The Rayleigh number scales as the inverse of the internal viscosity. Thus, convective mantle stress scales as

$$\tau_m \sim a_1 Ra_i^{-1/3}. \tag{3b}$$

For an active-lid mode of mantle convection to exist prior to an atmospheric temperature change, the convective mantle stress, τ_m^0, must be able to reach the yield stress of the lithosphere, τ_y^L. For the situation in which the depth-dependent component of the yield stress in Eq. (1b) is relatively small, the lithosphere yield stress will be effectively determined by the surface value, τ_{y0}. For situations where the depth-dependent component is not relatively small, the lithosphere yield stress will be determined by the stress value at the brittle–ductile transition (i.e., the peak stress within the lithosphere). For fixed convective stress, a greater lithospheric yield stress will imply that a lower perturbation will be required to initiate a transition. For fixed yield stress, a greater level of convective stress will imply that a greater perturbation will be required to initiate a transition in the mode of mantle convection. The implication is that the ratio τ_y^L/τ_m^0 is expected to be a key scaling parameter. This ratio scales as

$$\tau_y^L/\tau_m^0 \sim \tau_y^L Ra_i^{1/3}/a_1 \quad (4)$$

where Ra_i is defined before the atmospheric temperature change.

We now need to relate mantle stress before and after a surface temperature change. We maintain our assumption that internal temperature changes will be related to surface changes. This leads to a lowering of convective stress directly through a reduction in viscosity. Lower mantle viscosity also raises the internal mantle Rayleigh number, Ra_i, which can effect plate velocity, v. Both of these effects are accounted for in Eq. (3b). What is not accounted for is the fact that a reduced temperature drop across the mantle, due to a surface temperature increase, lowers the mantle Rayleigh number which, in turn, lowers plate velocity. This effect will scale as $(1-T_s)^{2/3}$, where we maintain the convention of a non-dimensional mantle temperature drop of unity prior to a surface temperature change. Combining the assumptions above leads to an expression that relates mantle stress before and after a surface temperature change:

$$\tau_m^1/\tau_m^0 \sim \left(1-T_s^1\right)^{2/3} \exp\left[-\theta T_s^1/6\right]. \quad (5)$$

By equating Eqs. (4) and (5) we can relate the convective mantle stress after a surface temperature change to the yield stress of the lithosphere after the change which we assume to be equal to (not any smaller than) the yield stress before. The final step is to assume the existence of a critical surface temperature, T_{crit}^1, that causes a change in the mode of mantle convection from active-lid to episodic or stagnant lid. At this critical point the change in convective stress, Eq. (5), sends the mantle to a new stress state that is less than the lithospheric yield stress, Eq. (4). This leads to our main scaling relationship given by

$$\left(1-T_{crit}^1\right)^{2/3} \exp\left[-\theta T_{crit}^1/6\right] \sim \tau_y^L Ra_i^{1/3}/a_1. \quad (6)$$

The relationship above predicts that the critical surface temperature change will decrease as activation temperature, lithospheric yield stress, and/or internal mantle Rayleigh number increase. The validity of these qualitative trends will be tested using numerical simulations. The simulations will also provide a quantitative test in that our analysis predicts that all parameter results, from simulation suites at fixed aspect ratio, should collapse to a single linear trend when expressed in terms of the specific parameter combinations of Eq. (6). The goodness of a linear fit will determine the level of error introduced by our simplifying assumptions. Provided the deviation from a linear trend is tolerable, numerical simulations employing variable convective wavelengths can be used to constrain the geometric constant within our scaling relationship which determines the slope of the linear trend. The intercept of the trend, at the point were $T_{crit}^1=0$, will constrain the lithospheric yield stress at which active-lid convection could not exist at the reference mantle state.

3. Numerical simulations

To test our scaling ideas, we compare scaling predictions to the results of two-dimensional numerical simulations. A finite element code (Moresi and Solomatov, 1995) is used to solve the coupled mass, momentum, and energy equations for infinite Prandtl number convection. The majority of simulations presented were preformed using mesh densities ranging from 64×64 to 128×128 finite elements over any 1×1 patch of the modeling domain with enhanced element concentrations within the upper and lower thermal boundary layers. For simulations with the highest activation temperature, several cases were preformed with 192×192 mesh densities. The resolution range was used to check that simulation determined critical surface temperature did not depend on mesh density. We were able to determine the critical temperature to two significant figures (that is, the lowest critical temperature we resolved was 0.01).

Thermal boundary conditions are a constant temperature base, $T=1$, and a spatially constant surface temperature that can vary in magnitude. Mechanical conditions are a free slip surface and base. Side boundaries are free slip and reflecting for the bulk of simulations. A relatively small suite of simulations will also consider wrap-around side boundaries. The convecting layer can be heated from below or through a combination of basal and internal heating. The viscosity function is given by Eqs. (1a)–(1c). The lithospheric yield stress was varied by changing the non-dimensional surface value, τ_{y0}, while holding the depth-dependent term fixed at a non-dimensional value

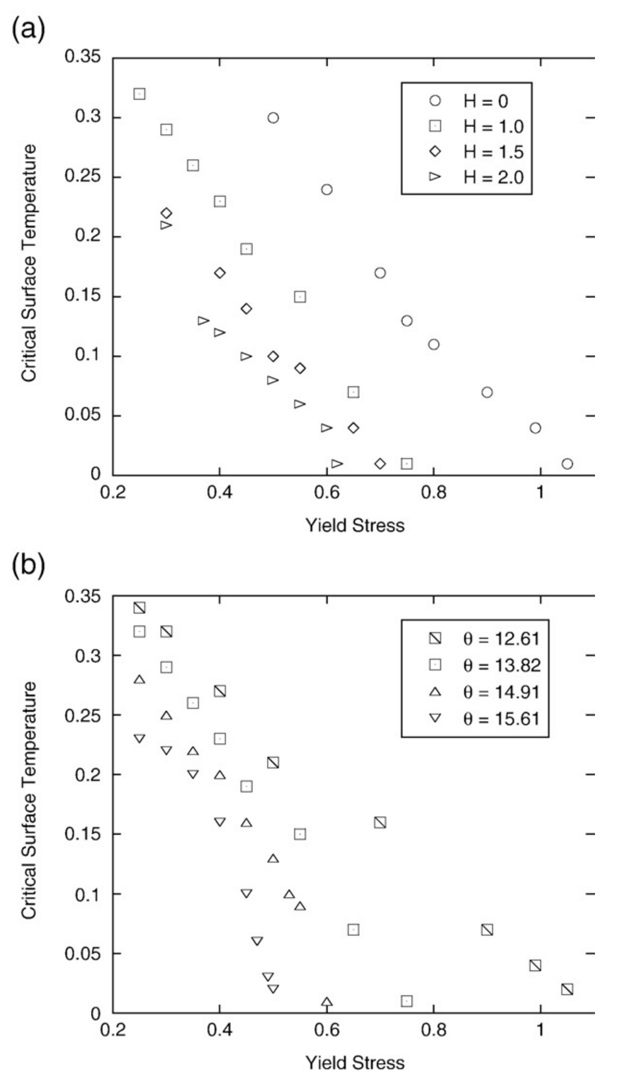

Fig. 1. (a) Critical surface temperature versus lithospheric yield stress from simulation suites with variable levels of internal heating. (b) Critical surface temperature versus lithospheric yield stress from simulation suites with variable activation temperatures.

of 0.1. The non-dimensional lithosphere thickness for all simulations was less than 1/5 of the full layer depth. In addition, τ_{y0} was always greater than 0.2. Thus the yield stress, Eq. (1b), is dominated by τ_{y0} and the lithosphere yield stress values noted in the remainder of this section are equal to τ_{y0}.

In addition to yield stress, the other simulation parameters are the activation temperature, θ, the surface Rayleigh number, Ra_s, the ratio of the internal to basal heating Rayleigh number, H, and the aspect ratio of mantle convection cells. The initial Ra_s is 100 for all suites. This value is defined using the viscosity at the initial surface temperature of zero. The internal Rayleigh number, Ra_i is not an input parameter. It can be determined from the simulations after they reach statistically steady state.

Numerical testing proceeded as follows. A reference state simulation, with a non-dimensional surface temperature of zero, was run to statistically steady state. The reference state was in an active-lid mode of convection. This reference state served as the initial condition from which the surface temperature could be increased and new simulations run to statistically steady state. The new states, after the temperature change, could remain in an active-lid mode of convection or transition to an episodic or stagnant-lid mode. For any initial reference case, roughly 7–10 simulations were run systematically to determine the critical surface temperature change needed to cause a transition. This was determined by monitoring average surface velocities for all simulations.

By varying activation temperature, surface yield stress, the ratio of internal to basal heating, and cell aspect ratio, we built up a range of different initial states. For each initial state, we ran a series simulations that varied the imposed surface temperature change. This allowed us to determine the value required to initiate a transition in the mode of convection. In this way a large number of simulation results were generated for comparison with scaling predictions.

Fig. 1a and b shows, respectively, numerical simulation results that map the effects of variable degrees of internal to basal heating and variable activation temperatures. All the simulations are for aspect ratio one convection cells. The activation temperature is 13.82 for all simulations in Fig. 1a. The heating ratio is 1.0 for all simulations in Fig. 1b. The value of Ra_i, for the initial active-lid states, ranges from 4.74×10^5 to 1.76×10^7. The ratio of the surface heat flow due to internal versus basal heating ranges from 0 to 55%.

Increased internal heating increases the internal mantle temperature which lowers the internal viscosity and increases the internal mantle Rayleigh number. It is this increase in Ra_i that leads to a convective mantle stress nearer to the lithospheric yield stress, Eq. (4). Our analysis predicts that this leads to a lower critical surface temperature for tectonic transition, Eq. (6). Numerical simulation results are consistent with this prediction, Fig. 1a.

A higher activation temperature means that a temperature increase has a larger effect on the ratio of convective stress before and after a surface change, Eq. (5). This leads to the prediction that a higher activation temperature should lower the critical surface temperature change, Eq. (6). This prediction is also consistent with simulation results, Fig. 1b.

The majority of the simulation suites of Fig. 1 transition from an active-lid mode of convection to an episodic mode (relatively long time spans of stagnant lid with episodic pulses of whole lithosphere overturn). The exception was cases with the highest degree of internal heating. Those cases, for medium to high lithospheric yield stress values, transitioned from active to stagnant lid. Our scaling analysis is designed to isolate when active-lid convection becomes unstable. The conditions under which the transition is to episodic versus stagnant lid is an issue we have not addressed. Our analysis also does not map the thickness of the episodic regime in parameter space. Both of these are interesting issues that are left for future study.

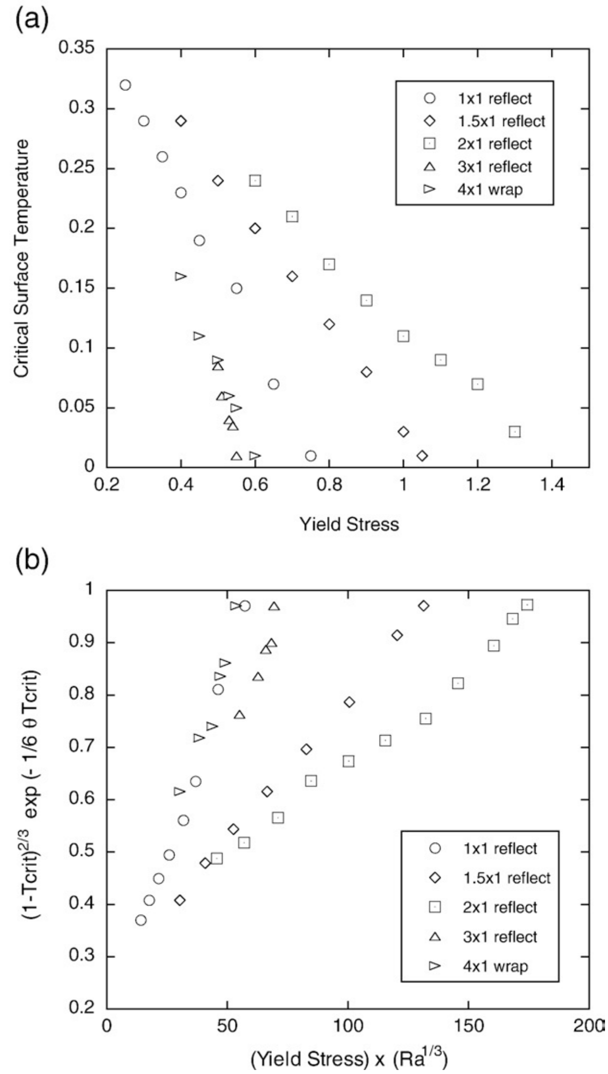

Fig. 3. (a) Critical surface temperature versus lithospheric yield stress from simulation suites with variable aspect ratios and side boundary conditions. (b) Results from all the simulation suites of (b) plotted in terms of the scaling analysis combination of parameters of Eq. (6).

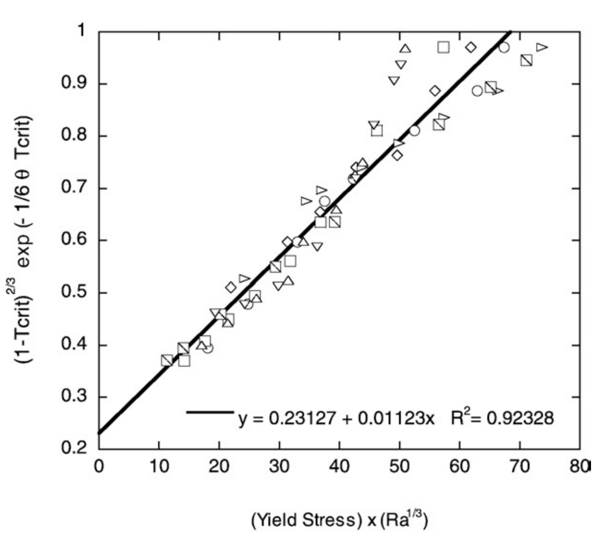

Fig. 2. Results from all the simulation suites of Fig. 1 plotted in terms of the scaling analysis combination of parameters of Eq. (6). A best linear fit through the data points is also shown with associated error.

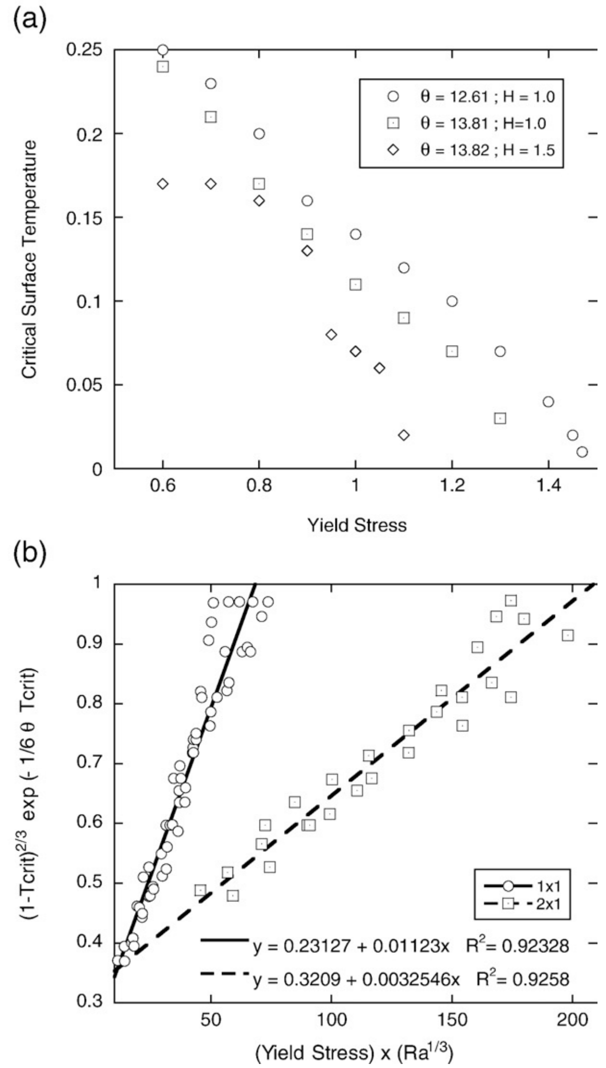

Fig. 4. (a) Critical surface temperature versus lithospheric yield stress from 2×1 simulation suites with variable activation temperatures and degrees of internal heating. (b) Results from all the simulation suites of Fig. 1 and 4b plotted in terms of the scaling analysis combination of parameters of Eq. (6).

Fig. 2 shows the results from all the simulations suites of Fig. 1 plotted in terms of the parameter combination of Eq. (6). A best fit linear trend, along with associated error, is also shown. The largest scatter occurs for simulations with relatively low critical temperatures. This indicates that the greatest scaling error occurs for situations in which the initial reference state is already near the transition in tectonic modes. We will return to this issue when we evaluate scaling errors.

Fig. 3a shows the results of four simulation suites that vary the aspect ratio of convection cells, maintaining reflecting side boundaries, and one simulation suite with wrap-around side boundaries. The activation temperature and heating ratio are, respectively, 13.82 and 1.0 for all simulations. Fig. 3b shows results from the same suites plotted in terms of the parameter combinations of Eq. (6). The slope of the trends, in the scaling parameter space, is seen to depend on cell aspect ratio. Aspect ratio two cells, with reflecting boundaries, are the most stable in the sense that they require the largest surface temperature change to induce a tectonic transition.

The simulation suite with wrap-around side boundaries (Fig. 3) starts with 2 cells within the 4×1 modeling domain. Unlike the reflecting boundary cases, the cells are free to drift laterally through the domain. This means that simulation analogs to plate margin zones are not fixed in space, as they are for the reflecting side boundary suites. The boundary conditions also allow for the potential of surface temperature changes altering convective wavelengths before a transition occurs. Based on the results of Fig. 3, this added system freedom leads to lower critical surface temperature being required to initiate tectonic transitions. This is an interesting observation that deserves future exploration. For now, the key is that this geometric effect leads to lower predicted critical temperatures. Thus by focussing on the results from 2×1 cells with reflecting boundaries we will remain conservative when it comes to applying scaling results to planets.

Fig. 4a shows that the qualitative trends associated with increased internal heating and or activation temperature are robust for 2×1 cells. For this simulation suite, Ra_i ranges from 4.4×10^6 to 4.0×10^7. The ratio of the surface heat flow due to internal versus basal heating ranges from 0 to 54%. Fig. 4b shows that predicted scaling trends are associated with similar error levels for variable convective cell wavelengths. Unlike the unit aspect ratio cases, the majority of 2×1 simulation suites transition directly from an active to a stagnant-lid mode of convection. Fig. 5 shows that crossing the critical temperature, for longer aspect ratio cases, leads to a change in convective wavelength. The shorter ensuing wavelengths lower convective stress levels (that wavelengths shorter than two are associated with lower critical temperatures, Fig. 3, suggests that convective stress is also lower at shorter wavelength). Thus, an episodic regime, which requires that the time variation of convective stress is such that it occasionally reaches yield stress values, becomes less likely.

The key assumption to our scaling analysis is that internal mantle temperature changes track surface temperature changes. Numerical simulations confirm the qualitative validity of this assumption (if this was not the case, numerical simulations would not show any tectonic transitions). Key assumptions were made in our analysis to determine a simple closed form relationship between surface temperature and internal temperature. The level or error introduced by our assumptions can be evaluated by comparing predictions from simulations

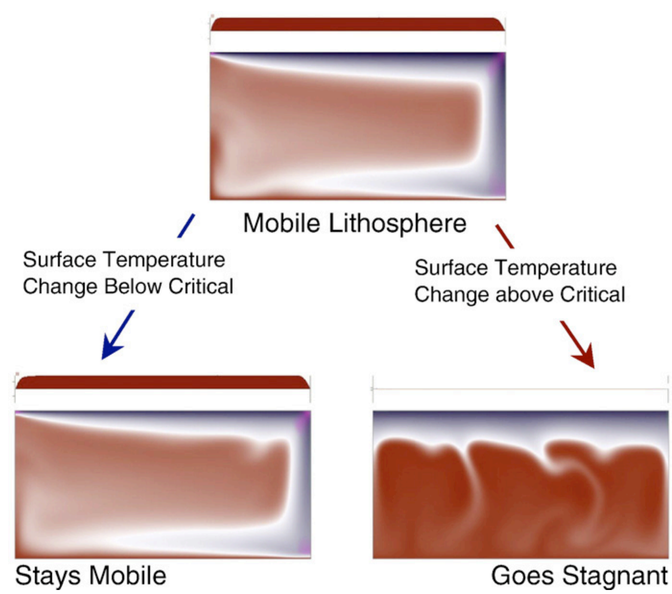

Fig. 5. Thermal fields from simulations with fixed activation temperature and yield stress values but variable surface temperatures. Horizontal surface velocity is shown above each image. The light zones in the upper corners of the top and bottom left image highlight regions of plastic failure (weak plate boundary analogs). The reference case (top) has a non-dimensional surface temperature of zero. From that reference case, the surface temperature is systematically increased to determine the conditions that allow an active-lid mode of convection to be maintained despite the temperature change (bottom left) and the conditions under which the temperature change induces a transition to a stagnant-lid mode of convection (bottom right).

that remain in an active-lid mode of convection but have variable surface temperatures. An example of this is shown in Fig. 6a. The simulations used for this figure all assumed a 2×1 domain with reflecting side boundaries, a heating ratio of 1.0, a non-dimensional yield stress of 0.8, and an activation temperature of 12.61. The error trends and amplitudes were robust for simulation suites with varied parameter values. The error levels can account for scaling misfits at the higher end of the predicted critical temperature (Fig. 4b). This error, which decreases at lower critical temperatures, cannot account for the largest scatter in simulation results.

The largest scatter, when numerical simulation results are plotted in terms of Eq. (6) parameter combinations, occurs in the parameter range associated with the lowest critical temperatures. Errors in this parameter space range are due to the fact that our analysis is strictly valid only at thermal equilibrium points while the simulations are in disequilbrium during adjustment to surface changes. Fig. 6b shows a schematic diagram of how this serves as an error source for our analysis. Our analysis determines system characteristics at an initial equilibrium state, point A in the diagram. It also determines the equilibrium state associated with the highest surface temperature that allows plate tectonics to remain stable, point B, for the particular

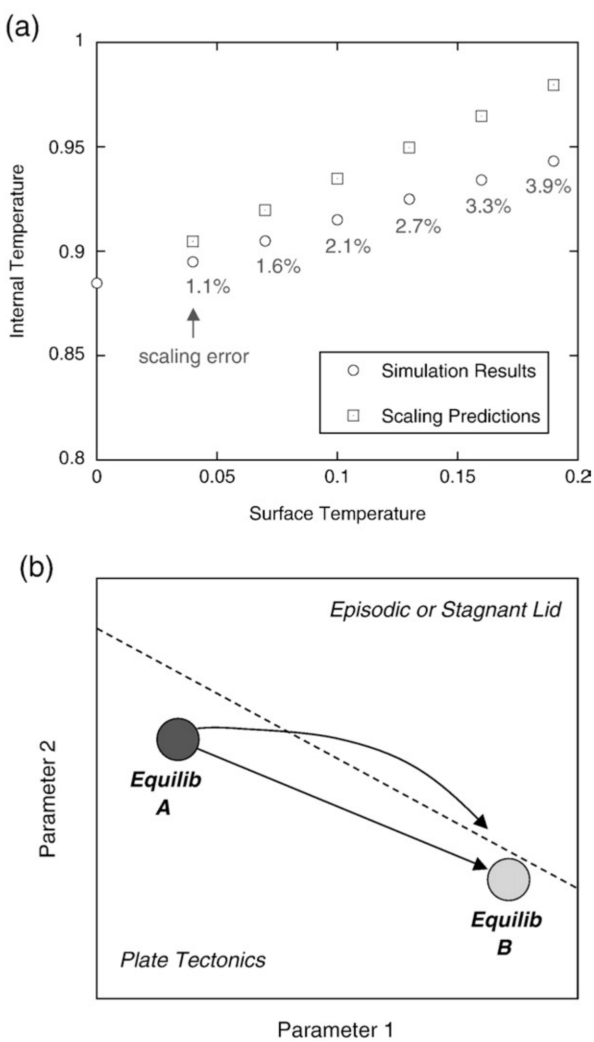

Fig. 6. (a) Numerical simulation results and scaling predictions for internal mantle temperature versus surface temperature for models that remain in an active-lid mode of convection. The error associated with the scaling prediction relative to simulation results is noted for each set of data points. (b) Schematic diagram of errors introduced to our equilibrium scaling analysis due to disequilibrium behavior during system adjustment to imposed surface temperature changes.

parameter values associated with the initial state. During disequilibrium adjustment, several factors that affect convective stress will be changing, e.g., mantle viscosity and surface plate velocity. If the key factors happened to all change at the same rate, then the adjustment path would qualitatively follow the straight arrow path of Fig. 6b. There is no reason to think this will be the case. It is more likely that different factors will adjust at different rates and that each adjustment rate will itself depend on the amplitude of the surface temperature change. This means that curved adjustment paths should be expected. The error this can introduce will become more pronounced if initial states are already near a transition point (all other factors remaining equal, if point A in Fig. 6b were farther from the transition line, then disequilibrium adjustment paths would be less likely to cause the system to transition). This is consistent with simulation results which show the largest scatter at small critical temperature values (Fig. 4b).

4. Discussion

The comparison of scaling predictions to numerical simulations showed that our analysis predicts correct qualitative trends. For application to planets, the quantitative errors inherent to our analysis are tolerable in the sense that the uncertainty they introduce in determining the critical surface temperature change is less than the uncertainties associated with our incomplete knowledge of key planetary parameter values. Specifically, uncertainties in the exact value of the activation temperature of planetary mantles, the internal mantle Rayleigh number, lithospheric yield stress, and the characteristic convective wavelength will introduce greater variations in the predicted critical surface temperature range than will the errors associated with our simplifying assumption. At this stage, we can take a conservative approach and assume parameter values that tend to maximize the predicted critical surface temperature change. Before doing this, however, several factors not included in our analysis deserve consideration.

Our analysis did not consider the effects of a zone of elastic behavior within the lithosphere. The depth to the base of the elastic component of the lithosphere will decrease with increasing temperature, all other factors remaining equal (Watts et al., 1980). This could decrease lithosphere strength which would act to counter a decrease in mantle convective stress. A surface temperature change will have a linear effect on lithospheric temperature profiles, and thus on the thickness of the elastic lithosphere, but will have an exponential effect on viscosity and thus on convectively generated stress. By neglecting elastic effects we are assuming that the temperature dependence of viscosity is so strong that the exponential effect on reducing stress, due to a surface temperature change, greatly exceeds the linear effect of increased surface temperature also reducing elastic lithospheric thickness.

Neglecting elastic behavior is tied to our assumption of constant lithospheric yield stress before and after a surface temperature change. This may seem a liberal assumption in that it could, through the elastic effects noted above, predict a lower critical temperature than would occur in nature. There are, however, added effects which make this assumption a potentially conservative one. If the presence of water is key to plate margin weakness (Bird, 1978; Lenardic and Kaula, 1994; Bercovici, 1996), then there is the potential that increased surface temperatures, which could lead to partial dehydration, could increase lithospheric yield stress. Alternatively, if the presence of serpentinized oceanic lithosphere is key to maintaining a low effective friction coefficient in subduction settings and serpentinization depth is temperature limited (Evans, 1977; Schmidt and Poli, 1998), then increased temperature could lead to shallower serpentinization (Li and Lee, in press) which could also increase lithospheric yield stress. Both effects would allow a lower temperature change to induce a tectonic transition.

To show a representative application of our analysis, we will employ the scaling trend for 2×1 cells (Fig. 4b). This trend is associated with the highest predicted critical temperature and is, in this sense, conservative. To dimensionalize scaling results, we will use values representative of present day Earth. The yield stress of the lithosphere is associated with significant uncertainty that, as we will see, has a large effect of predicted critical temperatures. As such we will consider a range of potential yield stresses.

We set the mantle potential temperature to 1350°C and the average surface temperature, before a climate induced change, to zero. We consider a conservative activation temperature by using a relatively low non-dimensional value of 13.82. To dimensionalize viscosity we assume an internal viscosity, i.e., the value at the mantle potential temperature, of 10^{22} Pa s. This is conservative as lower values will predict lower critical temperatures. By setting a non-adiabatic mantle temperature drop, we can calculate the effective viscosity of the lithosphere which is required to dimensionalize yield stress. We choose a non-adiabatic temperature drop of 1800°. This is also conservative as larger values will predict lower critical temperatures. The temperature drop and internal viscosity are needed to determine the internal mantle Rayleigh number. Other required values are density (4000 kg/m^3), gravitational acceleration (9.8 ms^2), thermal expansion coefficient (3×10^5), mantle depth (3000 km), and thermal diffusivity (10^{-6} m^2/s). Fig. 7a plots the predicted critical surface temperature for a range of yield stress values.

The yield stress required to allow for plate tectonics, under present day Earth conditions, has been explored by several groups (Moresi and Solomatov, 1998; Trompert and Hansen, 1998; Tackley, 1998; O'Neill et al., submitted for publication). Moresi and Solomatov express the value in terms of an effective friction coefficient and find that it must be 0.03–0.13. This is well below dry rock friction but can be achieved with elevated pore fluid pressures. The effective friction coefficient value along major faults on the Earth has also been estimated from data constraints. For the San Andreas fault a recent study infers a value of 0.1 (d'Alessio et al., 2006). Constraints from subduction zones also lead to stress levels much lower than would be inferred based on dry rock friction (Bird, 1978). A recent study of this sort predicts effective friction coefficient values of 0.019–0.095 (Lamb, 2006). If we consider the average depth to the ductile portion of the lithosphere to be 20 km, then the range of field constrained effective friction coefficients above implies yield stress values from 11 to 59 MPa (assuming an average overburden density of 3000 kg/m^3). The average value of 35 MPa is in accord with an alternate estimate that considers the laboratory determined effective friction coefficient of the weakest materials that could potentially be present in fault zones (Moore and Rymer, 2007). The lowest friction coefficient material, talc, implies a stress at the brittle–ductile transition of ≈30 MPa.

If we consider the yield stress range of 30–35 MPa as a reasonable estimate, then Fig. 7a predicts that a long-lived surface temperature change of 60–120° could cause plate tectonics to become unstable. This temperature range is considerably less than that associated with present day Earth–Venus surface temperature differences and is also considerable less that the critical temperature beyond which free water could not exist at the Earths surface (Kasting and Ono, 2006). The implication is that climate driven surface temperature changes could cause plate tectonics to cease independent of major water loss and associated lithospheric dehydration. This is reinforced by the fact that we have been conservative in our parameter choices. Once plate tectonics ceases on a planet then the recycling of water and CO_2 will be altered and the potential exists of a further enhanced increase in surface temperature to the point of complete water loss. What is key, from the standpoint of our analysis, is that the trigger for this need not involve complete dehydration and associated lithosphere strengthening as is often assumed.

The illustrative example above considered parameter estimates for present day Earth. Estimates for yield stress on other planets are sparse but, given that Earth estimates are so low relative to dry rock friction values, it is possible that other terrestrial planets could be

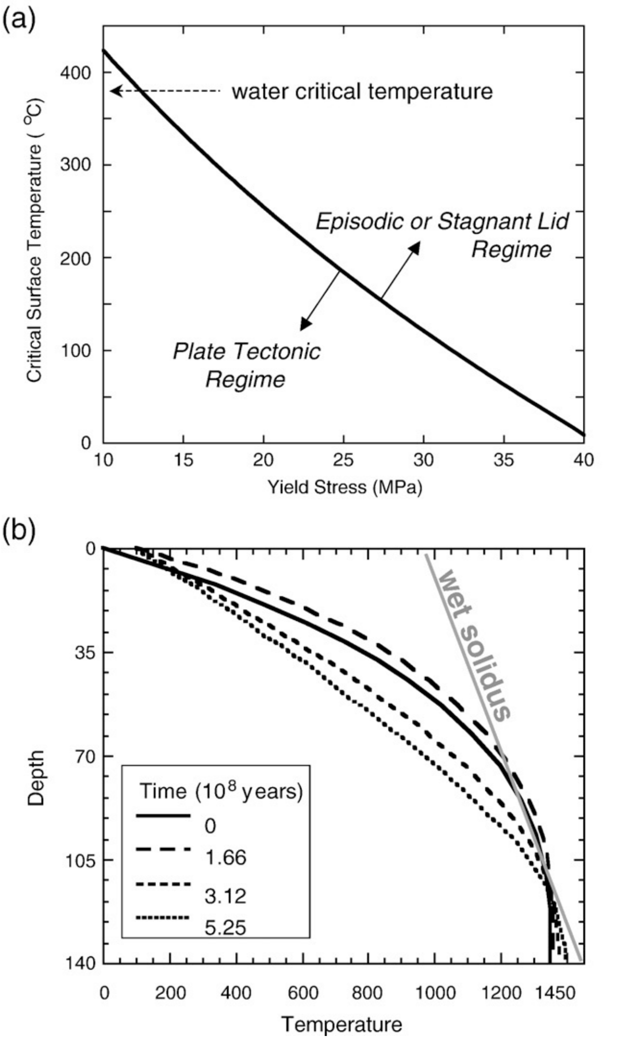

Fig. 7. (a) Dimensional critical surface temperature versus dimensional lithospheric yield stress. Parameter values used for dimensionalization are discussed in the body of the text. (b) Horizontally averaged vertical temperature profiles from a simulation that transitions from active to stagnant-lid convection in response to a surface temperature change. The wet solidus is for peridotite with a water content of 1000 ppm (Hirschmann, 2006).

associated with higher yield stress values. Our analysis would then suggest that even lower climate driven temperature changes could shut down an episode of plate tectonics on such planets. Our analysis also suggests that the surface temperature change needed to induce a tectonic transition depends on the internal mantle Rayleigh number. All other factors remaining equal, a higher Rayleigh number is associated with a lower critical temperature. This suggests that the stability of plate tectonics varies over the thermal history of a planet and smaller temperature changes could induce transitions earlier in a planets history.

Although the primary intent of this paper is to explore conditions that allow for the potential of a climate induced change in a planets tectonic mode, it is worth briefly considering the surface signatures of such a transition if it did occur. As convective stresses drop below yield stress, in our simulations, surface velocities experiences enhanced time dependence prior to a complete tectonic transition. This suggests that the transition might be preceded by tectonic pulses. Short of being on a planet long enough to monitor such tectonic bursts, the signatures they leave would be difficult to discern from planetary missions. A potentially more observable effect is associated with volcanism. For the present day Earth, the average oceanic geotherm almost nicks the wet solidus for peridotite with a water content of 1000 ppm (Hirschmann, 2006). Fig. 7b shows results from a simulation dimensionalized so that the initial average geotherm nicks the solidus. The simulation transitioned from an active to a stagnant-lid mode of convection at a dimensional time of 1.7×10^8 yr after a surface temperature change was imposed. Prior to the transition, internal mantle temperature increases. This can allow for a near global melt zone to form just prior to the shut down of plate tectonics. The melt zone can persist just past plate shut down but its extent should decrease as the thickness of the lithosphere increases after plate tectonics ceases. This suggests that enhanced volcanic activity can coincide with a climate induced shut down of plate tectonics and the level of volcanic activity should decay after the tectonic transition.

The illustrative example above depends on mantle water content and its thermal state prior to a climate induced tectonic transition. A dryer mantle would require higher potential temperatures and/or greater surface temperature changes to allow for a global melt event to accompany the end of an era of plate tectonics. These trade offs can be explored within the limits of our scaling theory as developed thus far but considerations of volcanism warrant extending our ideas. Enhanced volcanic activity could feedback into surface temperatures by releasing additional greenhouse gases into the atmosphere and could also lower the concentration of water in the mantle thereby increasing its viscosity. These two effects would work at odds in terms of altering convective stress in the mantle and thus the evolution of a planet, subsequent to plate tectonic shut down, could depend on which effect dominates. This motivates future extensions of our scaling ideas and numerical simulations that go beyond the scope of this paper.

5. Concluding remarks

The parameter conditions that allow for a climate induced transition in the global tectonic mode of a terrestrial planet have been explored by considering a simplified system. The system consists of a convecting mantle layer with a viscoplastic rheology. The layer can be subjected to surface temperature changes that mimic the effects of an enhanced greenhouse atmosphere and/or increased solar luminosity. Our main result is that, for parameter regimes pertinent to the Earth, an atmospheric temperature change of 60–120° could shut down plate tectonics.

Our conclusions are subject to our simplifying assumptions. However, the model system we consider retains essential aspects that govern the mode of mantle convection in terrestrial planets and is sufficiently simple that it may be understood with straightforward scaling theory augmented by a suite of numerical simulations. Together this forms a departure point for considering more complex models that involve potentially a rich variety of time-dependent processes that can influence surface temperature as well as the yield strength and response time of the lithosphere. Quantitatively, added complexities (e.g., depth-dependent viscosity, geometric effects associated with 3D spherical convection) will effect the exact value of the atmospheric temperature change required to initiate a tectonic transition. Qualitatively, our main conclusion is that such a value does exist: Climate change can shut down plate tectonics. This hinges on three assumptions: 1) convective mantle stress must reach lithospheric yield to allow for plate tectonics, 2) mantle stress scales with mantle viscosity which depends on temperature, and 3) increased surface temperature will lead to increased internal temperature in a convecting mantle. Provided that any unexplored effects do not cause one of these assumptions to be invalid, our general conclusion will hold for more complex models (of course, the required temperature change must remain within the range of possibility for mild to extreme greenhouse atmospheres in order for the theory to retain utility for our own solar system — for extensions to extra solar planets, specifically any found at the same proximity from their star as Earth is from ours, the required change must be compatible with the potential range of solar luminosities).

Another way of phrasing our general conclusion is that plate tectonics is potentially unstable to surface perturbations. This runs counter to a long standing idea that the mantle convection system of Earth, plate tectonics being the prime surface manifestation, is self buffered and thus stable (Tozer, 1972). The self buffering argument relies on a negative feedback effect that can be illustrated by thought experiment: If mantle heat loss decreases (for example, due to increased surface temperature), then the mantle would heat up which would reduce its viscosity allowing for enhanced convective vigor and, thus, increased surface heat loss which would act to offset the initial decrease. Like most negative feedbacks, this implies that the system is stable and can correct itself against perturbations to maintain a smooth running evolution until internal energy sources are eventually tapped. This idea was crucial to the development of thermal history models for the Earth and their extension to other planets (Davies, 1980; Schubert et al., 1980; Stevenson et al., 1983).

The self buffering effect discussed above relies on what is, effectively, an amplifier. That is, the exponential dependence of viscosity allows the effect a perturbation signal has on the interior mantle to be amplified such that the system can adjust relatively rapidly and damp potential perturbation growth. As with other buffered systems that rely on an amplifier, the potential exists for overcompensation and/or a compensation magnitude that adversely effects other aspects of the system (adversely in this sense means working counter to self regulation). We introduce a subtle but critical departure from the ideas of Tozer: Self-buffering feeds not only into heat loss but also into mantle stress, which has a threshold value beyond which self regulation cannot be maintained. It is this break from the conventional view that permits the hypothesis that mantle convection in a plate tectonic regime is potentially unstable to climate change.

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