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Convective thermal evolution of the upper mantles of Earth and Venus

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Abstract. On Earth the present-day rate of heat loss is about twice the heat generation rate; on Venus it is about half. Though this rough balance may be due to a feedback mechanism between mantle temperature and heat loss, it is difficult to see how such a mechanism can occur on timescales of 1 Ga or less when the upper mantle of the Earth is thought to be cooling at about 40 °C Ga^{-1}. On Venus a decrease in surface heat flux presumably occurred at the end of the catastrophic resurfacing event at $\sim$ 500 Ma. Parameterized convection models relate heat flux to Rayleigh number by the exponent β. Such models using a range of viscosities and values of β from 0.2 to 0.3 show that the effect of a sudden decrease in surface heat flux is to cause an independently convecting upper mantle to increase in temperature by $100-500$ °C over 1 Ga, whereas, if whole mantle convection occurs, the temperature change is less than 60 °C. An increase in mantle temperature of 200 °C or more will affect mantle viscosity, lithospheric thickness and melt generation rate, all of which may affect the feedback mechanism.

Introduction

On Earth 65 % or more of surface heat loss occurs by plate creation on ridges [Sclater et al. 1981]. The total rate of heat loss is about 4 $\times 10^{13}$ W, and is about twice the rate of radioactive heat production of 2 $\times 10^{13}$ W. The temperature of the interior must therefore be decreasing at an average rate of about 100 °C Ga^{-1} for the whole Earth. If the mantle convects as two layers, the cooling rate of the upper mantle is less than that of the lower mantle, and is probably about 40 °C Ga^{-1} [Richter 1984]. Since Earth's ocean floor, through which most of the heat is lost, is younger than 200 Ma, nothing is yet known about heat loss through oceanic plate creation at earlier times.

Whether the Earth's mantle convects in one or two layers is not clear. In view of these uncertainties, our models below explore both single and two-layered mantles which represent likely end-members of the real situation.

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Observations of Venus from the Magellan spacecraft have suggested that silicate planets with radii of about 6000 km do not all behave in the same manner. Venus appears to be in a different state to Earth, having no spreading ridges [Solomon et al. 1992]. Additionally, unlike Earth, Venus appears to have undergone a global resurfacing event which ended 300-500 Ma ago [Schaber et al. 1992].

The surface heat flux of Venus has not been measured directly, but can be estimated if the lithospheric thickness and mantle temperature are known. Estimates of the lithospheric thickness vary between about 100 and 400 km, giving rise to heat fluxes in the range 5-30 mW m^{-2}. Recent models of convection beneath plumes give similar values [Solomatov & Moresi 1996; Nimmo & McKenzie 1996]. In the models below, a value of 30 mW m^{-2} for the present-day heat loss on Venus is assumed, which is a likely upper limit.

The present day radiogenic heat generation rate on Venus is unknown. However, the K/U ratio measured by the Soviet landers is about the same as that on Earth, suggesting that potassium is present in similar quantities. Hence it is reasonable to assume that the Venus value of radiogenic heat generation is close to the terrestrial value of about 40 mW m^{-2}.

The rates of heat loss from both planets are therefore probably within a factor of two of their heat generation rates. This rough balance suggests that there may be some feedback mechanism which keeps the rates of heat loss and heat generation approximately equal. The most likely mechanism on Earth is that the changing temperature of the mantle alters the rate of plate creation or the average age of oceanic lithosphere, and hence the rate of heat loss. However, it is difficult to see how such processes could be affected by a change in mantle temperature of less than perhaps 100 °C, which according to the calculations above suggests that the feedback time might be of the order of the age of the Earth. The principal purpose of this paper is to show that the upper mantle temperature can in fact respond considerably more rapidly (within less than 1 Ga) to changes in surface heat flux such as may have occurred on Venus 300-500 Ma ago. This rapid adjustment only occurs if the mantle convects in two layers.

Theory

A convenient way to investigate the time dependent convective history of planets is by using parameterized convection models [e.g. McKenzie & Richter 1981]. Unlike numerical solutions of the full convective equations in a spherical shell, such models are not based directly on the governing fluid dynamic equations. However, they are required to conserve heat, and in practice provide a fast method of exploring the behaviour of a large range of convective models.

All parameterized models are based on a relationship between the Rayleigh number, Ra, and the Nusselt number, Nu, where

$$Ra = \frac{g\alpha F d^4}{k\kappa\nu}, \quad (1)$$

g is gravity, F is heat flux across the layer, d is layer depth, k is conductivity, α is thermal expansivity, κ is diffusivity and ν is kinematic viscosity. Nu is the ratio of heat flux (or temperature drop) across the layer in the presence of convection to that in its absence [McKenzie & Richter 1981]. For constant viscosity fluids, using equation (1) to define Ra the relationship is of the form

$$Nu = Ra^{\beta}. \quad (2)$$

The question is the extent to which this relationship holds for temperature- dependent viscosity. Christensen [1984] shows that β is around 0.3 for a fixed viscosity ratio, but closer to 0.1 for a fixed surface viscosity. Recent studies with temperature-dependent viscosity find that β varies from 0.1 to 0.4 [Gurnis 1989; Giannandrea & Christensen 1993; Honda 1995; Honda & Iwase 1996]. In this study, values of β of 0.2, 0.25 and 0.3 were used.

Following the method of McKenzie & Richter [1981], the heat flux F due to convection across a thermal boundary layer is given by

$$F = G(\Delta T)^{1/(1-\beta)} \quad (3)$$

where

$$G = D^{1/(1-\beta)} k \left(\frac{g\alpha}{\kappa\nu}\right)^{\beta/(1-\beta)} d^{(4\beta-1)/(1-\beta)} \quad (4)$$

and ΔT is the temperature difference between the boundary and the isothermal core. D is a numerical constant with a value of $(1.324)^{4/3}$ for stress-free boundary conditions when β is 0.25 [McKenzie & Richter 1981].

Model

A simple model was used to study the time dependent behaviour, consisting of a sphere with a fluid core and either a single or a two-layer mantle. The initial model condition was that the heat transferred across each interface was equal to the total heat generated in the layers below the interface. From this equilibrium condition the model was then stepped forward by 50 Ma increments, with the heat transfer across each interface being governed by equation (3) and the resultant change in the average temperature of each layer calculated using the specific heat capacity of the material. The interior of each layer was assumed to be isothermal. The system was allowed to evolve for 4 Ga, at which point the surface heat flux was reduced from the value calculated using (3) to a constant value of 30 mW m^{-2}. The subsequent evolution of the system was followed for a further 1 Ga, after which the heat loss governed by equation (3) was turned on again and the model allowed to evolve for a further 2 Ga.

Table 1. Constants for radioactive element decay.

element	abundance by weight	energies J/kg s	decay constant years
K	2.255×10^{-4}	3.6×10^{-9}	5.31×10^{-10}
^{238}U	2.239×10^{-8}	9.4×10^{-5}	1.54×10^{-10}
^{235}U	1.612×10^{-10}	5.7×10^{-4}	9.72×10^{-10}
Th	8.344×10^{-8}	2.7×10^{-5}	4.99×10^{-11}

Abundances from Wasserburg et al. [1964].
Energies from Wetherill [1966].

Table 2. Constant parameters used in model.

quantity	symbol	value	units
gravity	g	9.8	ms^{-2}
surface temperature	T_s	450	°C
mantle			
thermal conductivity	k	3.11	$Wm^{-1}K^{-1}$
thermal expansivity	α	$4.\times10^{-5}$	K^{-1}
average density	ρ	4.8	Mg m^{-3}
heat capacity	C_p	1200	$Jkg^{-1}K^{-1}$
upper mantle thickness	z_u	700	km
lower mantle thickness	z_l	2485	km
core			
heat capacity	C_pc	530	$Jkg^{-1}K^{-1}$
density	ρ_c	10.5	Mg m^{-3}
thickness	z_c	3185	km

The heat generated by radioactive decay in the (lower) mantle was calculated from the abundances and half lives given in Table 1. For the three layer case, the heat generation rate in the upper mantle was assumed to be 0.1 of the lower mantle value. There was assumed to be no heat generation within the core. The viscosity of both layers was assumed to be constant, with the lower layer having a viscosity between 1 and 100 times that of the upper layer. Other constant model parameters are given in Table 2.

Rather than allowing convection to extend to the surface, a conductive lid overlying the top layer of fluid was included in the model to represent the lithosphere. The lid was incorporated using a method proposed by McKenzie & Bickle [1988, Appendix B], assuming a temperature at the base of the conducting lid of 900°C.

Model results

Fig 1a shows the evolution of the temperature of the layers against time for a three-layer model where β=0.25. The effect of the sudden drop in surface heat flux at 4 Ga is to cause an increase in the potential temperature of the upper mantle of about 180 °C in 1 Ga. Once the normal heat flux is switched back on again, this temperature anomaly decays within less than 1 Ga. The lower mantle temperature is not significantly affected.

The reason why the upper mantle temperature rises so rapidly is easily understood. When the surface heat flux is suddenly reduced, that from the lower mantle into the upper mantle is initially not affected. The difference between the heat flow into and out of the upper mantle causes the temperature of the upper mantle to rise. Because the thickness of the upper mantle is only 700-800 km, and the initial temperature difference between the upper and lower mantle is about 800 °C, the temperature of the upper mantle increases rapidly. When the surface heat flux is increased at 5 Ga, the upper mantle temperature again adjusts quickly because of its

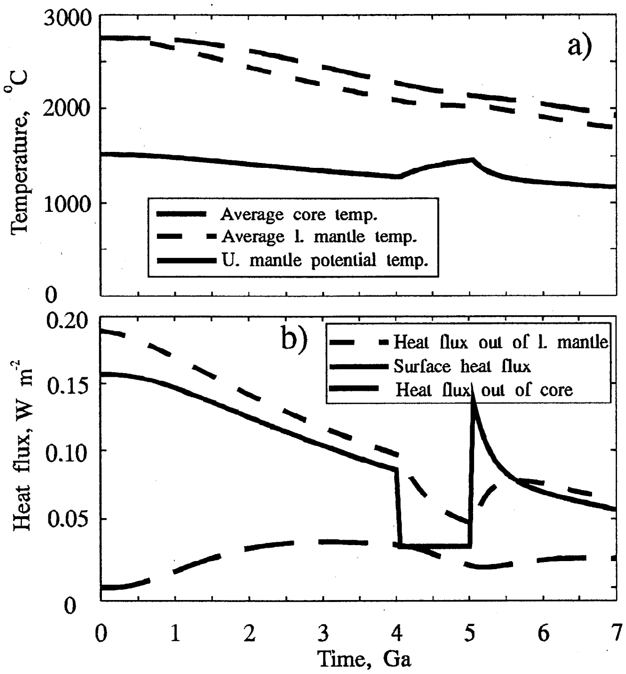

Figure 1. a) Variation in temperature with time for a three layer spherical planet, evolving from equilibrium with a conductive lid. The heat flux is fixed at 30 mW m^{-2} from 4 Ga to 5 Ga and β is 0.25 (see equation 2). The upper mantle viscosity ν_u is 5.5 $\times 10^{15}$ m^2 s^{-1} and the lower mantle viscosity ν_l is 1.65 $\times 10^{17}$ m^2 s^{-1}. The heat generation is calculated from Table 2. The upper mantle potential temperature T_p is calculated from $T_p = T_u e^{(\alpha g z_u / C_p)}$ where z_u, g, C_p and α are defined in Table 1 and T_u is the average temperature of the upper layer. The change in T_p from 4 to 5 Ga is 180 °C. b) Variation in heat flux for the same situation as Fig 1 a).

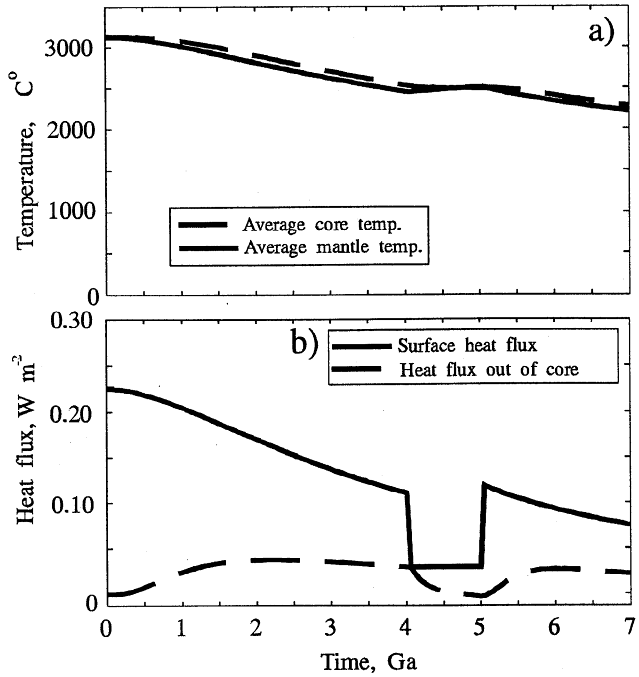

Figure 2. a) Variation in temperature with time for a two layer spherical planet, evolving from equilibrium with a conductive lid with $\nu_u = 5.5 \times 10^{15}$ m^2 s^{-1} ,$\nu_l =$ 1.65 $\times 10^{17}$ m^2 s^{-1}and $\beta = 0.25$. The change in mantle temperature from 4 to 5 Ga is 58 °C b) Variation in heat flux for the same situation as Fig 2 a).

short time constant. Fig 1b shows the heat flux across the three interfaces. The heat fluxes into and out of the upper mantle are almost equal at 4 Ga, their difference being due to the different surface areas of the two interfaces and the heat generation in the upper mantle. The equality of the heat fluxes is to be expected from equation 3, a consequence of which is that the upper mantle temperature will change so that the heat fluxes into and out of it are the same.

Fig 2a is the same as 1a but for a two-layer model, showing the mantle and core temperatures. The effect of reducing the surface heat flux for 1 Ga is to increase the mantle temperature by only 58 °C. This temperature change is smaller than in the three-layer system for two reasons. Firstly, the initial temperature difference between the core and mantle is zero, and thereafter it does not exceed 150 °C. It is this temperature difference which controls the net heat flux into the mantle after 4 Ga. Secondly, the greater heat capacity of the whole mantle compared with that of the upper mantle alone reduces the rate at which the layer heats up.

Fig 3 summarizes the results of experiments that all start from equilibrium, and whose surface heat flux is reduced at 4 Ga, using a value of β of 0.25. It shows the change in average internal temperature of the upper layer for a variety of upper mantle viscosities and viscosity contrasts, and demonstrates that the magnitude of the rise in upper mantle temperature is at least 100 °C over 1 Ga. An increase in β to 0.3 reduces all temperature changes, but the magnitude of the temperature rise is at least 100°C unless the viscosity is less than 10^{17} m^2 s^{-1}. The mantle temperature of two layer

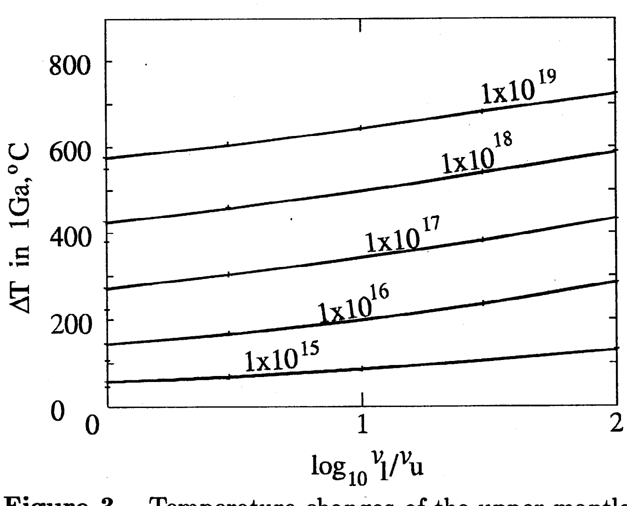

Figure 3. Temperature changes of the upper mantle, ΔT, from 4 Ga to 5 Ga for varying viscosities and a β of 0.25. A three-layer conductive-lid case starting from equilibrium is used, and the heat flux is fixed at 30 mW m^{-2} from 4 Ga to 5 Ga. Each line represents a different upper mantle viscosity (ν_u in m^2 s^{-1}) and the x-axis shows the log of the ratio of upper to lower mantle viscosities.

systems does not change by more than 60 °C over 1 Ga for the range of β and viscosities considered.

Discussion and Conclusions

There are a number of ways in which the simple model is not a good representation of what is likely to be happening in terrestrial planets. The most important is that the planet did not evolve from an initial equilibrium condition. The thermal evolution may also be affected by the temperature dependence of the viscosity, and the amount of heat generation within the core and upper mantle.

M^cKenzie & Richter [1981, equation 54] give an approximate analytical expression for the time for a three-layered system to adjust to a change in heat flux when β=0.25. Assuming an initial Rayleigh number for the whole Earth of around 8×10^{10} gives a time constant of 2.0 Ga, 2.8 Ga and 3.3 Ga for viscosity contrasts of 1, 30 and 100, respectively. For a two-layer mantle with a viscosity of 10^{18} m^2 $s^{-1}$and the same Rayleigh number the corresponding value is 0.9 Ga. Hence, for the systems modelled above, it is likely that the initial conditions will have little effect on the thermal behaviour of the system if the surface heat flux is reduced 4 Ga after the formation of the planet.

This study neglects heat generation within the core, which is a poorly constrained quantity. However, Breuer and Spohn [1993] find that the likely core contribution to radiogenic heat production is less than 13 % of the total, and can therefore be ignored to first order. Varying upper mantle heat production ratios between 0 and 1 of the lower mantle rate causes the temperature change over 1 Ga to vary between 145°C and 251°C for the case shown in Figure 1a.

The principal result from this study is that the temperature of the uppermost convecting layer will increase by $100 - 500$ °C over 1 Ga for Earth- or Venus-like conditions if a three-layer system suddenly has the heat loss out of the top layer reduced. The effect is much less pronounced for a model with whole mantle convection. The above results are only applicable if the behaviour of temperature-dependent viscosity fluids can be approximated by equation 2, with $0.2 < \beta < 0.3$; that such is the case is not yet clear.

An increase in upper mantle temperature of $\sim$ 200°C will have a major effect on the amount of melt produced, and probably also on the viscosity of the material immediately beneath the lithosphere. It is also likely to reduce the lithospheric and elastic thicknesses.

These results are some help in understanding how a planet can adjust its rate of heat loss to its heat generation rate. On Earth such an adjustment requires the mean age of the oceanic lithosphere to change, by changing either the velocity or the number of plates. It is, however, hard to understand how such feedback can occur when the rate of change of upper mantle temperature is as slow as 40 °C Ga^{-1}. This difficulty is reduced if the rate of change of temperature is more rapid, since plate velocities are likely to be strongly affected by changes in the upper mantle potential temperature of $100 - 200$ °C. The same is true of processes that cause new ridges and trenches to form. Both the rheology of mantle materials, and the rate of melt generation can be changed by an order of magnitude or more when the temperature increases by 200 °C. Detailed calculations on how the feedback process may occur are not straightforward, because they must be controlled by small scale processes, such as faulting, that cannot yet be properly modelled.

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