

Why does Venus lack a magnetic field?

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ABSTRACT

Venus and the Earth have similar radii and estimated bulk compositions, and both possess an iron core that is at least partially liquid. However, despite these similarities, Venus lacks an appreciable dipolar magnetic field. Here I examine the hypothesis that this absence is due to Venus's also lacking plate tectonics for the past 0.5 b.y. (1 b.y.= 10^9 yr). The generation of a global magnetic field requires core convection, which in turn requires extraction of heat from the core into the overlying mantle. Plate tectonics cools the Earth's mantle; on the basis of elastic thickness estimates and convection models, it is argued here that the mantle temperature on Venus is currently increasing. This heating will reduce the heat flux out of the core to zero over ~ 1 b.y., halting core convection and magnetic field generation. If plate tectonics was operating on Venus prior to ca. 0.5 Ga, a magnetic field may also have existed. On Earth, the geodynamo may be a consequence of plate tectonics; this connection between near-surface processes and core magnetism may also be relevant to the generation of magnetic fields on Mars, Mercury and Ganymede.

Keywords: geodynamo, plate tectonics, core, convection.

INTRODUCTION

Venus has a similar composition to Earth (Taylor, 1992; Kaula, 1994) and an at least partially molten iron core (Konopliv and Yoder, 1996). Yet despite these similarities, Venus appears to lack an appreciable Earth-like dipolar magnetic field (Russell, 1980). This difference is a long-standing puzzle, which cannot be explained by the planet's slow rotation (Russell, 1980). Although a slightly higher mantle temperature and lower core pressure relative to those of the Earth may be the answer (Stevenson et al., 1983), such explanations rely on small differences in poorly known parameters. The explanation examined here is that the absence of a Venusian magnetic field may be caused by another fundamental characteristic of Venus, namely its lack of plate tectonics. Briefly, in the absence of plate tectonics, the mantle on Venus cannot cool rapidly enough to drive core convection and a geodynamo. Although qualitative links between core

behavior and near-surface conditions on Venus have been suggested before (Buffett et al., 1996; Nimmo and Stevenson, 2000; Jellinek et al., 2002), a quantitative explanation such as that presented here has been lacking.

The rest of the paper is organized as follows. First, I outline the physical arguments relating magnetic field generation and core heat flux to the behavior of the mantle. It is then argued that the mantle of Venus is heating up at the present day. A consequence of this heating is that the heat flux out of the core must be declining, and I show that this reduction in heat flux can account for the present-day lack of a geodynamo. Finally, the results are discussed with respect to other terrestrial bodies.

CORE HEAT FLUX AND CONVECTION

Planetary magnetic fields are produced by motion in a conductor, usually the planet's iron core. Such motion may be due to either thermal convection or compositional convection, driven by core solidification (Buffett et al., 1996). The maximum heat flux that can be extracted from the core without thermal convection is given by (Nimmo and Stevenson, 2000)

$$F_c = k \alpha g T / C_p, \quad (1)$$

where k and α are the thermal conductivity and expansivity, g is the acceleration due to gravity, T is the core temperature, and C_p is the specific heat capacity. The values adopted are shown in Table 1 and imply that F_c is in the range 11–30 $\text{mW}\cdot\text{m}^{-2}$. Thermal convection will cease if the heat being extracted from the core is less than F_c ; in the absence of core solidification, the geodynamo will halt. Compositional convection may continue (Buffett et al., 1996; Stevenson et al., 1983), but will certainly halt if the heat flux out of the core drops to zero or below (i.e., the core starts heating up). The rate at which the core loses heat is controlled by the temperature difference between core and mantle and, thus, on the rate at which the mantle is cooling (Stevenson et al., 1983; Buffett et al., 1996; Nimmo and Stevenson, 2000).

HEAT-FLUX ESTIMATES ON VENUS

On Earth, the mean rate of heat loss from the interior is $82 \text{ mW}\cdot\text{m}^{-2}$, of which ~65% is due to plate tectonics (Sclater et al., 1980). The rate of heat loss exceeds the rate at which radioactive heat is being generated within the interior by a factor of 1.7–2.7 (Galer and Goldstein, 1991), leading to mantle cooling. Venus currently lacks plate tectonics (Solomon et al., 1992), suggesting less efficient mantle cooling, although crater counts indicate that a global resurfacing event ended at 0.8–0.2 Ga (Schaber et al., 1992).

Because direct measurements are lacking, the near-surface heat flux on Venus may be derived from the effective elastic thickness t_e of the rigid outer part of the planet (e.g., Solomon and Head, 1990). The temperature T_t governing the transition from rigid to ductile behavior may be estimated from experiments. If t_e , T_t , and the thermal conductivity are known, the near-surface heat flux may be calculated. The value of t_e recorded is the lowest since the episode of deformation, so for a thickening lithosphere such as that of Venus (Schubert et al., 1997), the heat flux derived is likely to be an overestimate.

Various authors have estimated t_e for Venus (e.g., Johnson and Sandwell, 1994; Phillips et al., 1997; Simons et al., 1997; Barnett et al., 2000) and obtained values in the range 5–48 km. In a global study of 34 localities, Barnett et al. (2002) concluded that there was no evidence that t_e anywhere on Venus is outside the range 29 ± 6 km and no correlation of t_e with surface age from crater counts. It thus appears that, to a first-order approximation, the elastic thickness on Venus (and hence its heat flux) may be characterized by a single value.

Figure 1 uses a theoretical model (Nimmo et al., 2002) to predict how t_e varies with the heat flux into the base of the crust for likely Venus conditions. The model assumes a temperature-dependent viscoelastic (Maxwell) rheology and includes radiogenic heat production within the crust. The predicted t_e increases with decreasing subcrustal heat flux and increases with increasing strain rate.

The observed range of t_e values are shown by the shaded region and imply a subcrustal heat flux F of $<10 \text{ mW}\cdot\text{m}^{-2}$. Using a stiffer rheology (dry olivine) or reducing the heating in the crust would increase F to $<20 \text{ mW}\cdot\text{m}^{-2}$. The temperature defining the base of the elastic layer, T_b , is 530–580 °C, which is similar to the upper bound found by Watts and Daly (1981) for the terrestrial oceanic mantle. This agreement is reasonable: Venusian crustal materials are nearly as rigid as terrestrial mantle (Mackwell et al., 1998) and probably contain less water. However, it should be noted that some authors (e.g., Phillips et al., 1997) have obtained heat fluxes of $\sim 40 \text{ mW}\cdot\text{m}^{-2}$ from the t_e observations.

An independent method of estimating near-surface heat flux comes from matching observations of inferred Venusian plume sites with convection models (Moore and Schubert, 1995; Nimmo and McKenzie, 1996; Reese et al., 1999). The inferred heat fluxes are generally $<20 \text{ mW}\cdot\text{m}^{-2}$, consistent with the estimates just described.

Surface measurements (Surkov et al., 1987) of U and Th abundances and K/U ratios on Venus suggest that the bulk concentrations of these elements on Venus are similar to those on Earth (Taylor, 1992; Kaula, 1994) and that the internal rate of heat generation may therefore be calculated. If a mean crustal thickness of 30 km is assumed (Simons et al., 1994), the surface measurements imply that up to one-third of the radiogenic elements may reside within the crust, rather than the mantle. This fraction is probably an upper bound because on Earth, the lower crust is less enriched in such elements than the upper crust. The heat production within the depleted Venus mantle is therefore equivalent to a heat flux H of $>30 \text{ mW}\cdot\text{m}^{-2}$, or at least 1.5 times the rate F at which heat is being lost. If the fraction of radiogenic elements in the crust is reduced, F increases (as already noted herein), but H increases more. The possibility that H exceeds F has been noted by other authors (Nimmo and McKenzie, 1997; Schubert et al., 1997) though not all agree (Phillips et al., 1997). A difference between F and H of $10 \text{ mW}\cdot\text{m}^{-2}$ over 1 b.y. results in a mantle-temperature increase of 30 K. This increase is likely to be at least 25% of the core-mantle temperature difference (see next section) and will thus reduce the core heat flux.

The discrepancy between mantle-heat production H and subcrustal heat flux F would be removed if the crust of Venus contained two-thirds of the total radiogenic element inventory, or if the total inventory were half that of Earth's. However, as noted earlier, neither of these alternatives is likely in view of the measured surface compositions. Unless some unrecognized mechanism of heat loss is occurring, the temperature of the mantle on Venus is increasing at $>30 \text{ K/b.y.}$ The temperature rise

probably only applies for the past 0.5 b.y.; before that time, plate tectonics (e.g., Turcotte, 1993) or some other mechanism of enhanced surface heat loss (e.g., Sleep, 2000) may have been operating, producing the resurfacing event deduced from crater counts (Schaber et al., 1992).

CONSEQUENCES OF MANTLE HEATING

The end of the resurfacing event is likely to have caused heat fluxes to drop to the present-day value deduced here. In Figure 2 the effect of a reduction in heat flux F at 0.6 Ga is illustrated, by using a parameterized thermal-evolution model (Nimmo and Stevenson, 2000). The heat flux out of the core is calculated by assuming that the lower thermal boundary layer has a mean thickness which keeps it at the critical Rayleigh number (here assumed to be 500; see Nimmo and Stevenson, 2000, for more details). This approach provides a good fit to experimental results (Manga et al., 2001). The core heat flux increases with the core-mantle temperature contrast, and the coefficient γ which determines the temperature dependence of the mantle viscosity. It decreases with increasing thermal boundary layer thickness.

The model starts with the surface heat flux equal to the heat generation rate. Prior to 0.6 Ga, plate tectonics is assumed to have been operating. This cools the mantle efficiently, driving a core heat flux of up to $13 \text{ mW}\cdot\text{m}^{-2}$. At 0.6 Ga the surface heat flux is reduced to $15 \text{ mW}\cdot\text{m}^{-2}$ to simulate the onset of stagnant lid convection. When the surface heat flow drops, the mantle temperature increases, and the heat flux out of the core decreases to zero over ~ 1 b.y. The shaded zone indicates the heat-flux range F_c below which core convection does not occur and shows that a geodynamo could not operate at the present day, but may have done so in the past.

The low value of the present-day core heat flux due to the reduction in F is simply a consequence of conservation of energy and is correspondingly robust. For instance, the time from 0.6 Ga to reach a core heat flux of zero varies by $<30\%$ for an-order-of-magnitude change in the reference viscosity or for a change in the viscosity exponent γ by a factor of 2. Mantle layering, if present, would increase this time scale (Nimmo and McKenzie, 1997), but not the magnitude of the effect. Sequestration of radiogenic elements into the crust would further reduce the present-day core heat flux. Calculating the present-day surface heat flux using a stagnant lid parameterization (Solomatov, 1995), rather than simply specifying it (as here), changes the present-day core heat flux by $<20\%$.

The model in Figure 2 starts with the core and mantle in thermal equilibrium, but the core of Venus may originally have been hotter due to the energy of accretion. An initially hot core increases the core heat flux and makes an early geodynamo more likely. However, because the mantle time constant is shorter than the age of the planet, initial conditions have little effect on the core after 4.6 b.y. A core which starts 200K hotter than in Figure 2 changes the core heat flux at the present day by less than $2 \text{ mW}\cdot\text{m}^{-2}$.

Other authors have also investigated the effect on mantle temperature of the cessation of plate tectonics and reached very similar conclusions, again illustrating the robustness of the result. Nimmo and McKenzie (1997) used a constant-viscosity mantle, resulting in a higher core heat flux but a similar time scale for the heat flux to decline. Schubert et al. (1997) found that the mantle temperature increased by 50 K over 0.5 b.y,

but did not consider the effect on the core. Neither of these papers addressed the likely consequences for the generation of a dynamo.

A major uncertainty not so far addressed is the effect of core solidification. Core solidification will occur under certain pressure and temperature conditions, and may drive a geodynamo even if the net heat flux out of the core is less than F_c (e.g., Buffett et al., 1996; Stevenson et al., 1983). The amount of sulfur present in the core is a major control on its solidus temperature, but for Venus this amount is unknown. The observational evidence for Venus suggests that the core is at least partially liquid (Konopliv and Yoder, 1996). A suite of models similar to Figure 2 were examined, incorporating core solidification using the parameters and method of Stevenson et al. (1983) and Schubert et al. (1988). Without a reduction in surface heat flux at 0.6 Ga, models producing partially liquid cores invariably resulted in a present-day dynamo. Some models with a completely liquid core lacked a geodynamo (c.f. model V1 of Stevenson et al., 1983). However, these models only occupied a restricted range of the parameter space explored, and in particular required high (>12%) core sulfur concentrations.

Conversely, the cessation of plate tectonics provides a natural way of shutting off the geodynamo, which does not depend on the details of core solidification. For models with core sulfur contents between 0 and 15% in which a geodynamo was operating prior to 0.6 Ga, the reduction in surface heat flux reduced the magnetic dipole moment by more than a factor of 10 over 1 b.y., and in many cases stopped the geodynamo altogether. The reason, again, is simply energy conservation: the rate of freezing of the core must be reduced if the heat flux out of the core declines.

Whether a geodynamo could ever have operated on Venus is uncertain, given the uncertainties in the parameters controlling F_c and the simplifications in the model. The results presented here, and those of Stevenson et al. (1983), suggest that an early dynamo on Venus is likely, but by no means required, and that the major uncertainty is the amount of sulfur in the core. Thus, the link between a cessation in plate tectonics and the disappearance of the geodynamo is likely to be robust; the previous existence of a geodynamo is less so. It has been suggested that the highlands of Venus may contain magnetite below its Curie temperature (Starukhina and Kreslavsky, 2002). If so, the potential exists for a record of an ancient field; thus its existence may be testable, at least in principle.

COMPARISON WITH OTHER BODIES

It is argued above that Venus lacks a geodynamo because it lacks plate tectonics. On Earth, cold subducting plates probably penetrate to near the core-mantle boundary, thus increasing the core-mantle temperature contrast and the heat flux out of the core (Davies, 1993). Hofmeister (1999) concluded that the likely heat flux across the ~200 km thick layer at the base of the mantle (known as D'') was in the range 26-39 mW·m⁻², which would probably allow a geodynamo to operate even in the absence of core solidification. The origin of D'' and its relationship to subducting slabs is not clear; furthermore, its behaviour over time depends on the competition between thermal and chemical buoyancy (e.g., Montague and Kellogg, 2000), neither of which are well known. Nonetheless, it seems probable that plate tectonics helps to drive the terrestrial geodynamo. Similarly, on Mars it has been argued that an ancient episode of plate tectonics could have driven a geodynamo early in that planet's history (Nimmo and Stevenson, 2000).

The origin of the presumed present-day geodynamos of Mercury and Ganymede is obscure. Mercury has been tectonically inactive for ~4 b.y., so plate tectonics is obviously not required to generate a geodynamo (see Schubert et al., 1988). On Ganymede, an ancient episode of tidal heating (e.g., Nimmo et al., 2002) may have heated the mantle. Such an event would allow the core to stay partially liquid for longer, but by reducing the heat flux out of the core might temporarily halt the geodynamo. Clearly, more work is needed to address these issues.

CONCLUSIONS

The attraction of the hypothesis presented here is that one anomalous feature of Venus (its lack of plate tectonics) may help to explain another (the absence of a magnetic field). Whether plate tectonics occurs is probably governed by factors such as the presence of water and the rigidity of the lithosphere (e.g., Tackley, 2000). Yet it appears that these near-surface phenomena may also control whether or not a planet possesses a global magnetic field. Such a connection is not at first sight obvious, but may be of use in elucidating the thermal histories of other bodies possessing magnetic fields.

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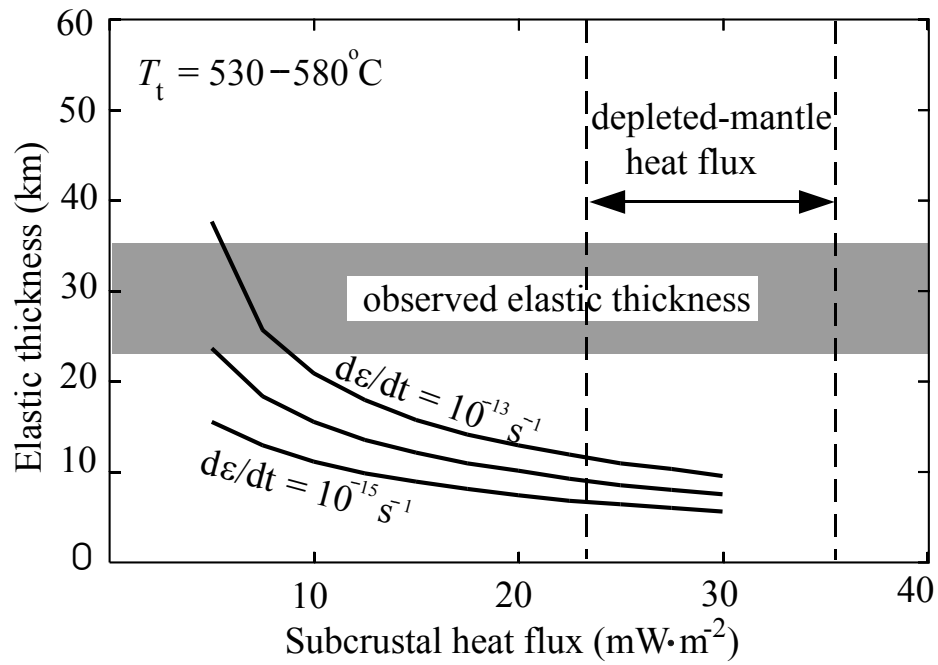
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FIGURE CAPTIONS

Figure 1. Variation in depth to base of elastic layer as a function of subcrustal heat flux and strain rate $d\varepsilon/dt$, by using method of Nimmo et al. (2002). Rheology used is dry diabase (Mackwell et al., 1998), Poisson's ratio is 0.25, Young's modulus is 100 GPa. To calculate geotherm, crust is assumed to be 30 km thick (Simons et al., 1994) and to have radiogenic concentrations 30 times undepleted mantle. Here and elsewhere, these

undepleted values are taken from Sun and McDonough (1989). Conductivity is $3 \text{ W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$, and surface temperature is $450 \text{ }^\circ\text{C}$. Shaded box denotes observed range of elastic thicknesses on Venus; vertical dashed lines show radiogenic mantle-heat flux H after crustal extraction, assuming $\pm 20\%$ error bars.

Figure 2. Model of thermal evolution of Venus, illustrating effect of reducing surface heat flux at 0.6 Ga. Model uses a parameterized convection scheme described in Nimmo and Stevenson (2000) and is started from equilibrium. It assumes a Newtonian mantle viscosity with a value of $10^{20} \text{ Pa}\cdot\text{s}$ at $1300 \text{ }^\circ\text{C}$ and a dependence on temperature of the form $\exp(-\gamma T)$, where $\gamma = 0.01 \text{ K}^{-1}$. Surface heat flux before 0.6 Ga is calculated by assuming that plate tectonics is operating (Nimmo and Stevenson, 2000); after 0.6 Ga, surface heat flux is set to a constant value of $15 \text{ mW}\cdot\text{m}^{-2}$. Core heat flux is calculated using a local critical Rayleigh number approach (see text). Mantle density, heat capacity, thermal diffusivity κ , thermal expansivity, and thickness are $4600 \text{ kg}\cdot\text{m}^{-3}$, $1200 \text{ J}\cdot\text{kg}^{-1}\cdot\text{K}^{-1}$, $10^{-6} \text{ m}^2\cdot\text{s}^{-1}$, $3 \times 10^{-5} \text{ K}^{-1}$ and 3000 km , respectively; core radius and density are 3185 km and $11500 \text{ kg}\cdot\text{m}^{-3}$, respectively; other parameters are given in Table 1. A: Evolution of core and mantle potential temperatures through time. A core potential temperature of $1300 \text{ }^\circ\text{C}$ equals $2300 \text{ }^\circ\text{C}$ at core-mantle boundary. B: Evolution of radiogenic heat production and heat fluxes across core and surface. Radiogenic concentrations are those given in Sun and McDonough (1989). Shaded box denotes estimates of F_c (see equation 1). Core heat fluxes above this range imply an active dynamo; heat fluxes less than this range should produce no dynamo.



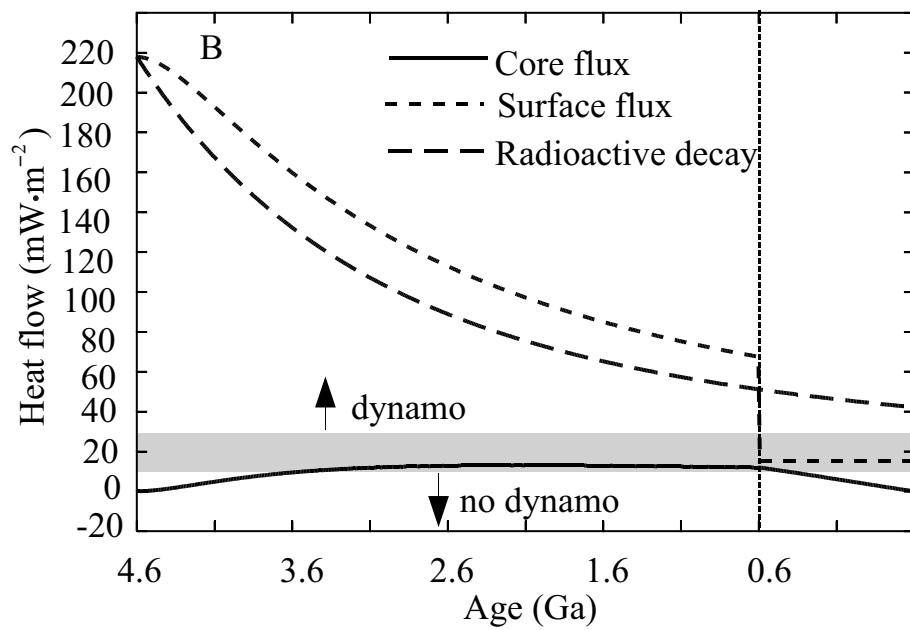
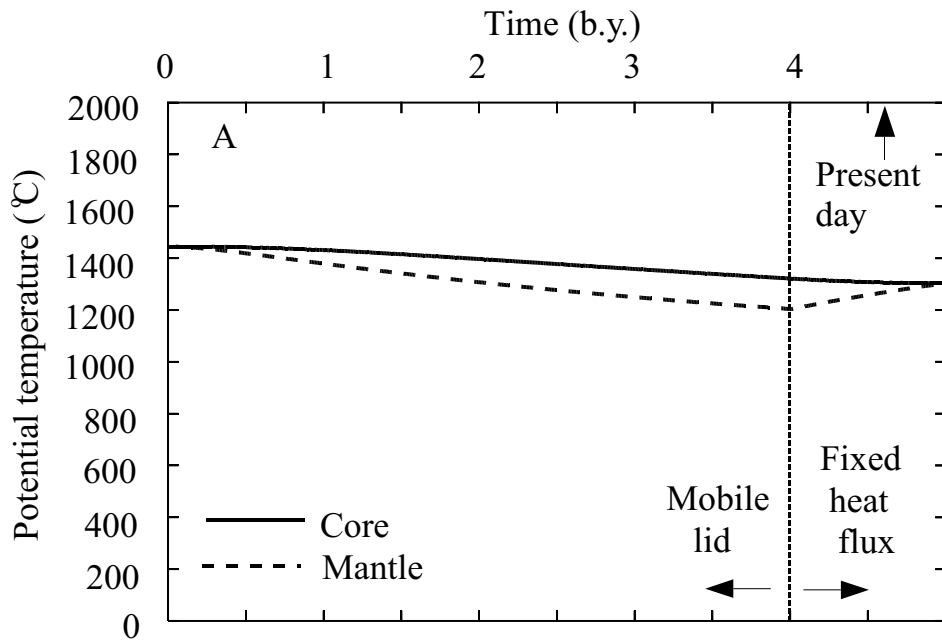


TABLE 1. VALUES OF CONSTANTS USED

Symbol	Value	Units	Reference
g	8.9	$\text{m}\cdot\text{s}^{-2}$	Kaula (1994)
k	36 +/- 8	$\text{W}\cdot\text{m}^{-1}\cdot\text{K}^{-1}$	Anderson (1998)
α	1.7 +/- 0.1	10^{-5}K^{-1}	Anderson (1998)
T	3000 +/- 500	K	Boehler (1996)
C_p	840 +/- 20	$\text{J}\cdot\text{kg}^{-1}\cdot\text{K}^{-1}$	Anderson (1998)