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# Rheological and petrological implications for a stagnant lid regime on Venus



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## ABSTRACT

Venus is physically similar to Earth but with no oceans and a hot dense atmosphere. Its near-random distribution of impact craters led to the inferences of episodic global resurfacing and a stagnant lid regime, and imply that it is not currently able to lose proportionately as much heat as Earth. This paper shows that a CO<sub>2</sub>-induced asthenosphere and decoupling of the mantle lid from the crust, caused by the elevated surface temperature, enables lid rejuvenation. Global hypsography implies a rate of  $4.0 \pm 0.5 \text{ km}^2 \text{ a}^{-1}$  and an implied heat loss rate of  $32.8 \pm 3.6 \text{ TW}$ ,  $\sim 90\%$  of a scaled Earth-like rate of heat loss of 36 TW. Estimates of the rate of lid rejuvenation by plume activity –  $0.07$  to  $0.09 \text{ km}^2 \text{ a}^{-1}$  – imply that ten times the number of observed plumes are required to equal this rate of heat loss. However, lid rejuvenation by convection allows Venus to maintain a stable tectonic regime, with subcrustal horizontal extension (half-spreading) rates of between 25 and 50 mm a<sup>-1</sup> determined from fits to topographic profiles across the principal rift systems. While the surface is largely detached from these processes, the association of rifting and other processes does imply that Venus is geologically active at the present day.

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## 1. Introduction

How Venus loses its internal heat has been uncertain since the realisation that its high surface temperature inhibits plate recycling that is so an effective heat loss mechanism on Earth (Anderson, 1981). The observation of a near-random distribution of impact craters (Phillips et al., 1992; Strom et al., 1994) led to the hypothesis of catastrophic, or episodic, global resurfacing (Turcotte, 1993), which proposes that for long periods (500 to 1000 Ma) the lithosphere cools and thickens whilst the interior heats up, the system eventually becoming unstable. In a geologically short period ( $\sim 50$  Ma) the whole lithosphere overturns and is replaced with new thin, buoyant lithosphere, whereupon subduction stops, and the cycle begins again.

Whether or not this happened, Venus is considered to now be in a stagnant lid regime (Solomatov and Moresi, 1996), in which the absence of an asthenosphere means that the lithospheric lid is coupled to a high viscosity mantle that convects slowly enough for conductive cooling to be the dominant heat transport mechanism. Some authors regard this regime as either complementary to episodic global resurfacing (Fowler and O'Brien, 1996; Sleep, 2000) or as a change in convective regime following the last

global resurfacing event (Reese et al., 1999). Whilst some lateral movements are expected in response to large-scale convection of the mantle (Grimm, 1994), such a lithosphere is not mobile in the terrestrial sense and is instead dominated by discrete small-scale plumes (Nimmo and McKenzie, 1998; Ogawa, 2000).

### 1.1. Global heat budget

Almost nothing is known about the internal properties of Venus; its moment of inertia is unknown and no seismic data have been obtained from which to constrain its internal structure. The lack of an intrinsic magnetic field and its  $k_2$  love number indicate that its core may be entirely liquid (Konopliv and Yoder, 1996) but there is little to constrain the core mass. Cosmochemical models (Basaltic Volcanism Study Project, 1981; Morgan and Anders, 1980) suggest core mass fractions between 23.6 and 32.0%—implying a mantle mass proportionately similar to or greater than Earth's. The Venera landers returned a number of K, U and Th measurements that imply bulk ratios, and hence internal radiogenic heating rates, comparable with Earth (Namiki and Solomon, 1998). While the Urey ratio may be different for Venus, the simplest assumption is to scale Earth's heat flux to Venus.

Earth's global heat flux, 44 TW (Pollack et al., 1993) scaled to Venus is 36 TW, or  $76 \text{ mW m}^{-2}$ . Turcotte et al. (1999) calculate that the hypothesis of episodic global resurfacing can remove only

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4.5 to 7.5 TW; their suggested mechanism for removing the remaining 28–31.5 TW is 'a very vigorous episode of tectonics and volcanism, following a global subduction event but prior to the subsequent stabilisation of a global lithosphere'. There are two problems with this reasoning: first, the global resurfacing event is the process that stabilises the lithosphere, by replacing unstable (> 250 km thick) lithosphere with thin, warm, buoyant new lithosphere; and second, the 'very vigorous' process is given no plausible physical explanation that can be tested either numerically or geologically. Thus more than three-quarters of the global heat budget is unaccounted for in this hypothesis.

A stagnant lid regime is able to remove 8–20  $\text{mW m}^{-2}$  (3.7–9.2 TW) depending on the model assumptions made (Reese et al., 1998; Solomatov and Moresi, 1996). Armann and Tackley (2012) explore the combination of magmatic heat pipes with stagnant lid regimes and conclude that while good fits may be obtained to observed topography, geoid and admittance ratios, the crustal production (magmatic resurfacing) rate is much too high, by about 2 orders of magnitude. Combining all three processes – episodic global resurfacing, magmatic heat pipes, and a stagnant lid regime – improves matters but still requires a magmatic flux at least an order of magnitude too high. However, the authors note that the inclusion of magmatic intrusion and a more realistic crustal rheology may be significant in reducing this discrepancy. This paper investigates the effect of realistic rheologies and the influence of mantle volatiles on a stagnant lid regime.

## 1.2. Mantle volatiles

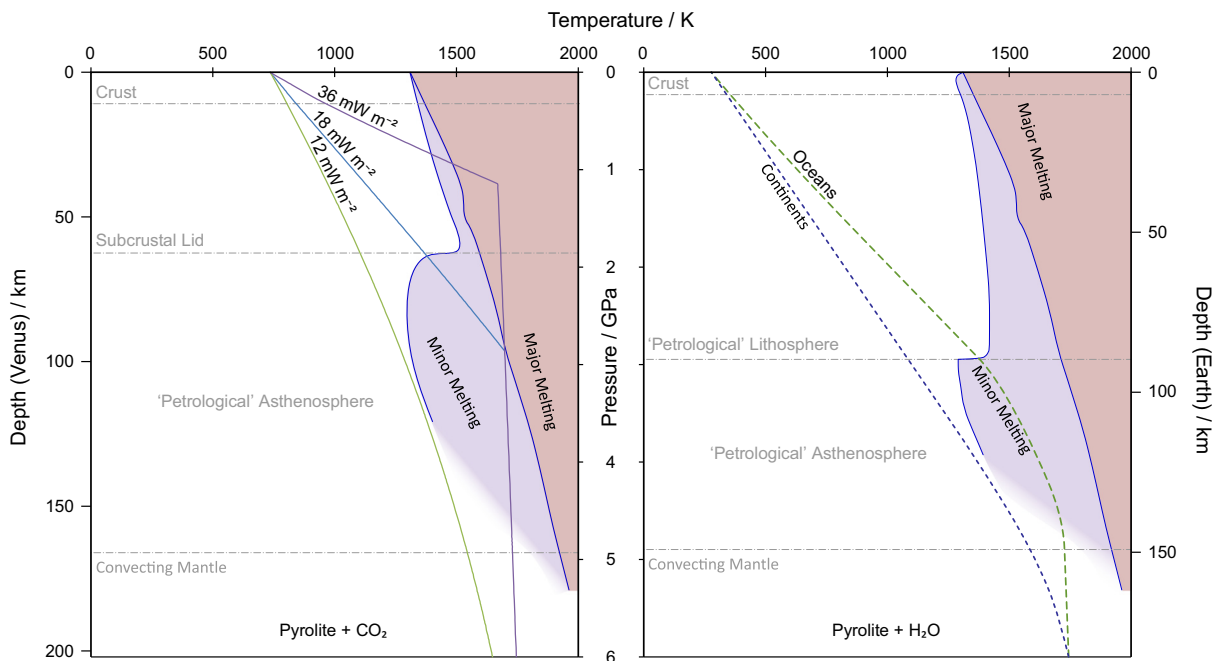
A compositionally Earth-like Venus would have volatiles, particularly  $\text{H}_2\text{O}$  and  $\text{CO}_2$ , in its mantle and consequently a weak low-viscosity asthenosphere. However, even from Pioneer Venus data it was clear that the 4 to 31  $\text{m km}^{-1}$  geoid–topography ratio is much higher than the terrestrial value of  $\sim 1\text{--}5 \text{ m km}^{-1}$  (Kucinskas et al., 1996; Smrekar and Phillips, 1991) implying the lack of an asthenosphere on Venus and a more-or-less constant viscosity mantle. Since a low-viscosity asthenosphere is considered to be essential for terrestrial plate tectonics (Richards et al.,

2001), which Venus lacks, the absence of an asthenosphere is perhaps not surprising.

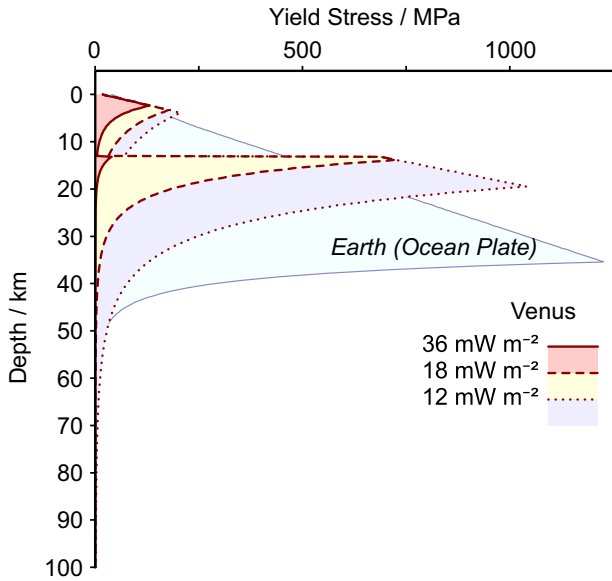
The extremely dry atmosphere, which contains only  $30 \pm 10 \text{ ppm H}_2\text{O}$  below the clouds (de Bergh et al., 2006), is consistent with the inference of a dry interior. Kaula (1999) argues that the deficiency of  $^{40}\text{Ar}$  in the atmosphere implies that the mantle must have been fully degassed very early in its history, before the accumulation of significant volumes of radiogenic  $^{40}\text{Ar}$  in the mantle. It is usually assumed that Venus lost its water through evaporation and hydrodynamic escape but given that even the Moon-forming impact did not fully devolatilise the terrestrial mantle and that extensive hydrodynamic erosion of a water-rich atmosphere is precluded by its noble gas inventory (Albarede, 2009), it seems more likely that Venus simply accreted less water than Earth.

New evidence for pyroclastic volcanism (Ghail and Wilson, 2013) raises questions about whether the mantle is dry at all. However, water is not the only possible volatile:  $\text{CO}_2$  is abundant in the atmosphere and perhaps the interior too, while the variability in atmospheric  $\text{SO}_2$  (Esposito, 1984; Marcq et al., 2013) may require a volcanic source. Pauer et al. (2006) have shown that an Earthlike mantle with a 20 to 200 km thick high viscosity lid (lithosphere) above a 100 km thick low viscosity channel (asthenosphere), albeit less pronounced than on Earth, and a gradually increasing viscosity with depth below that, is at least equally consistent with the geoid–topography data. Armann and Tackley (2012) also find a mantle viscosity structure similar to Earth.

Like water,  $\text{CO}_2$  is known to depress the pyrolite solidus at depth (Falloon and Green, 1989) and induce minor melting to form an asthenosphere. The depression of the solidus by water in the mantle of Earth and  $\text{CO}_2$  in the mantle of Venus is similar (Fig. 1) until an abrupt increase in magnitude at 1.85 GPa on Venus caused by the breakdown of diopside to give dolomite. The equivalent increase in magnitude of solidus depression at 3.0 GPa on Earth is more abrupt but less severe and results from the breakdown of pargasite. The magnitude of the solidus depression in both cases reduces gradually above 4.5 GPa (deeper than



**Fig. 1.** Possible Venus geotherms (left) compared with Earth (right). The 'petrological' asthenosphere is produced by  $\sim 1\%$  partial melting in the presence of  $\text{CO}_2$  on Venus or  $\text{H}_2\text{O}$  on Earth (Falloon and Green, 1989). Major Melting curves approximate those for a 1650 K adiabat. All data are plotted against pressure; depths in Venus and Earth are approximate.



**Fig. 2.** The yield stress (strength) for model Venus crust and mantle, at strain rates between  $10^{-15} \text{ s}^{-1}$ . Brittle deformation is controlled by Byerlee's Law, assuming a friction angle of  $30^\circ$ , and ductile deformation is controlled by power law creep, assuming a dry Maryland diabase rheology (Mackwell et al., 1998) for the crust and a dry Anita dunitite rheology (Chopra and Paterson, 1984) for the mantle. The crust is  $12.5 \text{ km}$  thick.

$\sim 150 \text{ km}$ ). On both Earth and Venus, the geothermal gradient intersects the respective  $\text{H}_2\text{O}$ - or  $\text{CO}_2$ -saturated solidus at  $90$  to  $100 \text{ km}$  depth, resulting in minor melting (Falloon and Green, 1990) that lowers the viscosity of the mantle and forms an asthenosphere.

### 1.3. Crust and mantle rheology

The high surface temperature,  $735.3 \text{ K}$  (Seiff et al., 1985), has a significant effect on the rheology of the Venus crust and upper mantle (Arkani-Hamed, 1993). Whereas Earth's oceanic crust is strongly coupled to the upper mantle, such that interior forces are transmitted to the surface, the Venus crust is partly decoupled from its upper mantle for typical geological strain rates ( $\sim 10^{-15} \text{ s}^{-1}$ ) across a range of temperature gradients (Fig. 2), even with the likely strong dry diabase crustal rheology (Mackwell et al., 1998). For a detachment layer to form, the crustal thickness must be at least  $\sim 10 \text{ km}$  and no more than  $\sim 15 \text{ km}$  (Fig. 2); crust much thinner than this will not be detached from the mantle, while a thicker crust would result in a thick very low viscosity channel (Buck, 1992), which is unlikely. Although thinner than often assumed (Anderson and Smrekar, 2006; Lawrence, 2003; Nimmo and McKenzie, 1998; Romeo and Turcotte, 2008), this crustal thickness is within the bounds of uncertainty (Leftwich et al., 1999).

## 2. Lid rejuvenation

The hypsographic curve for Venus has long been known to correspond with thermal cooling of the lithosphere (Brass and Harrison, 1982), which need not imply plate tectonics (Morgan and Phillips, 1983; Rosenblatt et al., 1994). If the mantle part of the stagnant lid is detached from the crust, it may be rejuvenated by thinning and recycling. The simplest means for describing this process is as a cooling half-space in which the upper boundary condition is set by the temperature,  $T_c$ , at the crustal detachment layer. This temperature is controlled by conduction from below

and therefore changes as the lid cools but, assuming a constant mantle temperature  $T_m$ , it can be calculated from the thickness of the mantle lid  $l_m$ , which grows with the square root of time  $t$ :

$$l_m = 2.32\sqrt{\kappa t}$$

where  $\kappa$  is the thermal diffusivity,  $10^{-6} \text{ m}^2 \text{ s}^{-1}$ . For each time step,  $T_c$  is then simply:

$$T_c = \frac{l_c(T_m - T_s)}{(l_c + l_m)} + T_s$$

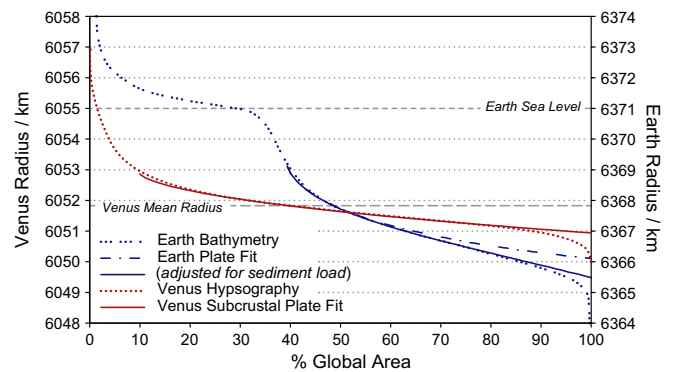
where  $T_s$  is the surface temperature and  $l_c$  is the thickness of the crust, i.e. the depth to the detachment layer, neglecting the slight increase in this depth as the crust conductively cools.

Using this value for the upper boundary condition the elevation  $h_s$  of the surface at each time step can then be calculated from the half-space cooling curve:

$$h_s = h_{crest} - \frac{2\alpha\rho_m(T_m - T_c)}{(\rho_m - \rho_a)}\sqrt{\frac{\kappa t}{\pi}}$$

where  $h_{crest}$  is the elevation of the crest formed when  $l_m$  is zero;  $\alpha$  is the thermal expansivity, which is close to  $3.5 \times 10^{-5} \text{ K}^{-1}$  for Venus;  $\rho_m$  is the mantle density,  $3300 \text{ kg m}^{-3}$ ; and  $\rho_a$  is the density of the atmosphere,  $64.8 \text{ kg m}^{-3}$  (Seiff et al., 1985). It is the atmospheric density that is significant here and not the crustal density because it is atmosphere that fills the void left by thermal contraction of the mantle lid, since although there is considerable variability in crust thickness, there is no obvious systematic change with time, so it is assumed that the crustal thickness is a constant and that it subsides along with the mantle lid. The implications of this assumption are discussed later.

The model is run at  $100 \text{ ka}$  intervals at a constant aerial rate of lid replacement (the rejuvenation rate,  $R_r$ ). For each time step, the heat flux and surface elevation (planetary radius) are calculated assuming an Earth-like mantle temperature of  $1650 \text{ K}$  (Doin and Fleitout, 1996; Johnson and Carlson, 1992; McKenzie et al., 2005; Stein and Stein, 1992) and assuming that the uppermost 10% of the hypsogram relates to features not generated by lid rejuvenation. The Venus hypsogram (Fig. 3) may be fitted well between 20% and 80% by adjusting the rejuvenation rate and the crest altitude; the best fit obtained by doing so is  $4.0 \text{ km}^2 \text{ a}^{-1}$ , at a crest elevation  $6052.9 \text{ km}$  radius. The misfit between this model and the hypsographic curve is  $< 0.1\%$ . The corresponding decompression melting magmatic volume production rate is  $29 \pm 4.2 \text{ km}^3 \text{ a}^{-1}$  (White et al., 1992) and implies a likely intrusion/extrusion ratio of  $\sim 5$  to 1 or more, assuming extrusion rate estimates for Venus (Bullock and Grinspoon, 2001; Romeo and Turcotte, 2010; Stofan et al., 2005).



**Fig. 3.** Global hypsography derived from Magellan altimeter data (Venus) fitted by a half-space cooling subcrustal lid model and ETOPO1 global relief model (Earth) fitted by a plate cooling model.

Conductive heat losses total  $7.3 \pm 0.3$  TW, in agreement with stagnant lid heat losses. However, lid rejuvenation – recycling of the subcrustal lid into the mantle – contributes by far the largest fraction to global heat loss, and may be calculated from the maximum lid thickness (corresponding to its lowest elevation on the hypsogram, i.e. in the last timestep), the mean temperature difference between this lid and the mantle, and the rejuvenation rate calculated above:

$$q_r = R_r l_m \rho_m C_p \frac{T_m - T_c}{2}$$

Assuming an average heat capacity,  $C_p$ , of  $1100 \text{ J kg}^{-1} \text{ K}^{-1}$ , the rejuvenation (replacement) of this lid accounts for  $32.8$  TW, including heat conducted through the crust. For comparison, the same assumptions and parameters imply that oceanic plate recycling is responsible for  $39.7$  TW of Earth's heat loss. Venus is able to lose so much heat because a much greater area of the planetary lid – 90% versus 60% – is involved in the process, and at a greater rate.

The lid rejuvenation rate obtained is almost independent of the assumed mantle density. Reducing the detachment depth to 10 km has no effect on the rejuvenation rate but increases the total heat loss to  $33.6$  TW; conversely increasing the depth to 15 km reduces the rate to  $3.9 \text{ km}^2 \text{ a}^{-1}$  and the total heat loss to  $31.9$  TW. The area assumed not to be involved in lid rejuvenation has a significant effect on the rejuvenation rate but not on the total heat loss; if only 5% is not involved the rejuvenation rate falls to  $3.6 \text{ km}^2 \text{ a}^{-1}$  and the total heat loss to  $32.2$  TW. Conversely for 15% the rejuvenation rate increases to  $4.9 \text{ km}^2 \text{ a}^{-1}$  and the total heat loss to  $34.7$  TW. Mantle temperature has a more significant impact; at 1600 K the rejuvenation rate is  $3.5 \text{ km}^2 \text{ a}^{-1}$  and the total heat loss just  $29.2$  TW; at 1700 K the rate is  $4.5 \text{ km}^2 \text{ a}^{-1}$  and the total heat loss is  $36.4$  TW. None of these alternatives provide a better fit to the hypsogram but the misfits are all less than 0.15%. The model assumes a constant crustal thickness but if the crust is assumed to grow in thickness with the depth to detachment (i.e. its basal temperature is a constant 860 K), the rejuvenation rate drops to  $3.2 \text{ km}^2 \text{ a}^{-1}$ , the total heat loss falls to  $27.3$  TW and the maximum crustal thickness is  $20.3$  km. Such a scenario is considered unlikely because the crust is compositionally distinct from

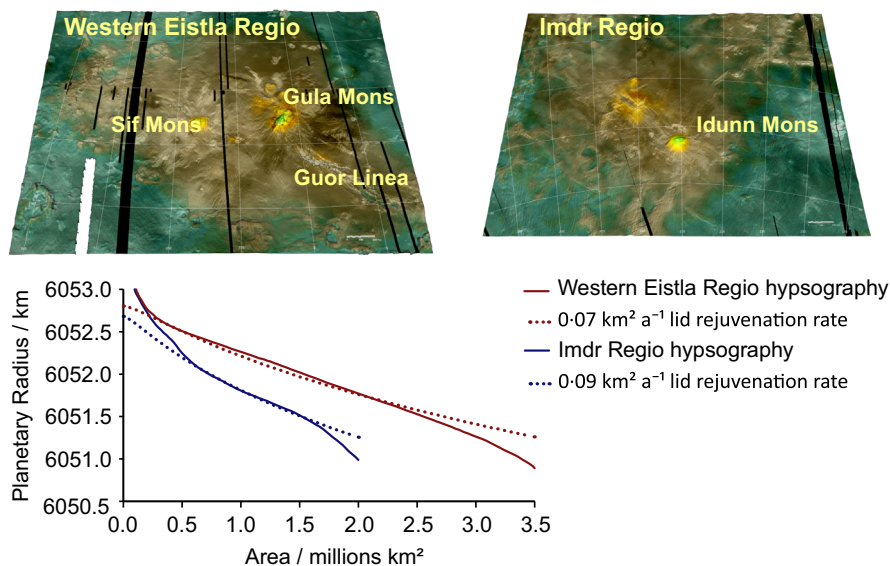
the mantle but it does illustrate the possible limits of crustal thickness and global heat loss.

The implication of these results is that Venus is able to maintain a stable tectonic regime (i.e. one without episodic global resurfacing) by rejuvenation of its subcrustal mantle lid, but how might this lid rejuvenation occur? The two most likely possibilities are replacement by plumes or entrainment (delamination) in mantle convection.

### 2.1. Lid rejuvenation by plumes

A plume or hot spot impinging on an old stagnant lid will warm, thin and uplift it, in effect rejuvenating it in the same way that oceanic hot spots do on Earth. Assuming that a plume initially impinges with a large head and is followed by a much narrower tail (or jet), a large topographic swell will form and then decay as the lid cools and thickens. Thus the height of the swell may be compared with the fit to global hypsography to estimate the length of time since the first impingement of the plume. Nine swells of approximately 1000 km diameter have been identified on Venus (Stofan et al., 1995) and interpreted as the surface expression of plumes. Of those located in lowland plains that might serve as examples of lid rejuvenation by plumes, two are inferred on the basis of their gravity signature to be young (Smrekar and Parmentier, 1996): western Eistla Regio and Imdr Regio. Western Eistla Regio is a  $2500 \times 1900$  km swell that rises to an elevation of  $6052.8$  km and includes the large complex volcanoes Sif and Gula Montes (Senske et al., 1992). Imdr Regio is a  $1900 \times 1400$  km swell rising to a mean elevation of  $6052.7$  km. It includes only one large volcano, Idunn Mons, which from Venus Express data is inferred to have been active within the last few million years (Smrekar et al., 2010b).

By applying the same methodology used for the global hypsography to the local hypsogram for each plume (Fig. 4) the rejuvenation rate for each plume can be estimated. A reasonable fit to the data is obtained for Imdr Regio with a lid rejuvenation rate of  $0.09 \text{ km}^2 \text{ a}^{-1}$ . The total heat flux associated with lid rejuvenation by the plume is  $0.45$  TW. The fit for western Eistla Regio is poorer; the data are fitted best with a deeper depth to detachment, at 15 km, and a lower rejuvenation rate of



**Fig. 4.** Perspective views across western Eistla Regio (left) and Imdr Regio (right). The plume swells (brown) rise above the surrounding plains (teal) by more than 1000 m. The horizontal scale bars are 200 km and the vertical exaggeration is  $25 \times$ . Black stripes are image data gaps, the white stripe an altimeter data gap. Imdr Regio is smaller and lower than western Eistla Regio but an apparently higher lid rejuvenation rate (bottom). These features may indicate an earlier stage of plume impingement or a smaller plume.



$0.07 \text{ km}^2 \text{ a}^{-1}$ , implying a rejuvenation heat flux of  $0.36 \text{ TW}$ . The higher rejuvenation rate of the Imdr Regio plume may indicate an earlier stage of development; the thicker crust at western Eistla Regio might result from a longer period of volcanic extrusion there, or from differences in crustal composition.

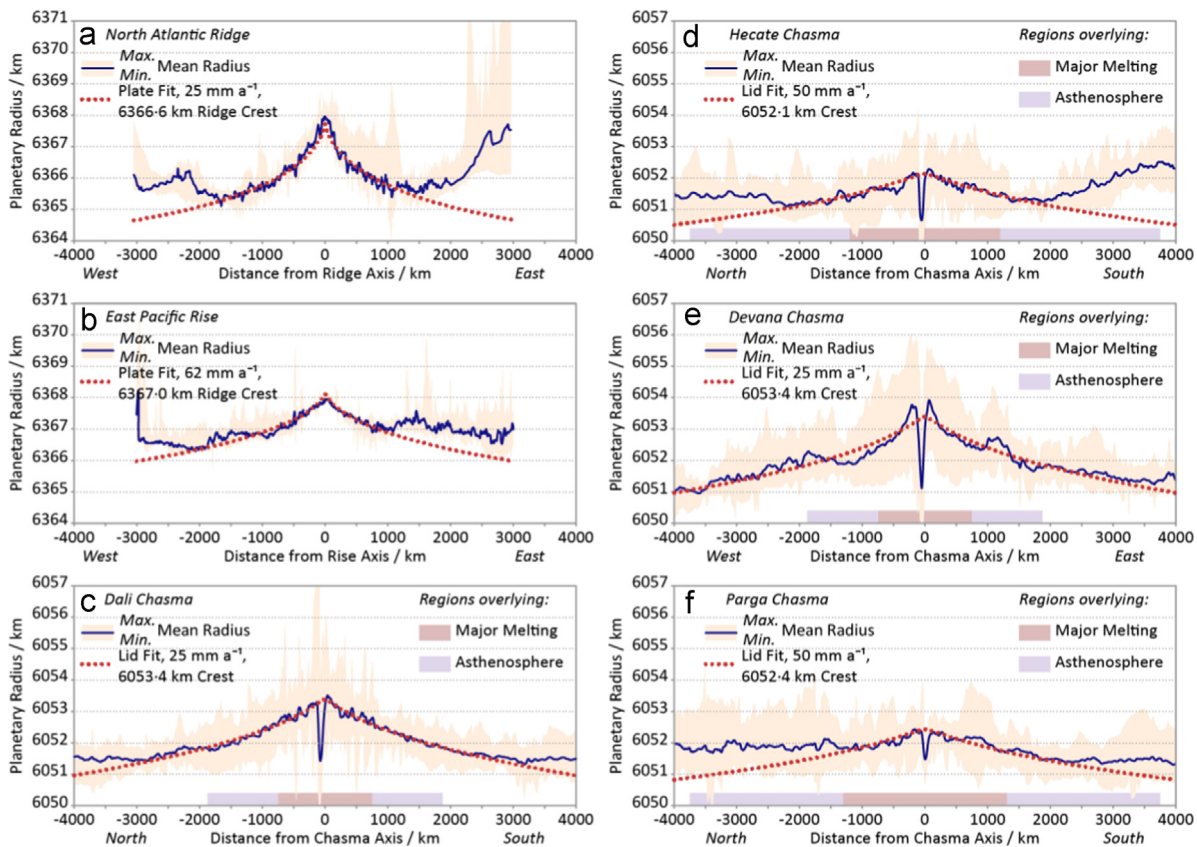
These rates are similar to those at the largest terrestrial plume, Hawaii, which has an estimated heat flux of  $0.36 \text{ TW}$  (Turcotte, 1995). For Venus to lose proportionately as much heat as Earth requires that  $75 \pm 2$  Hawaiian-sized plumes must be active at any time, more than ten times the estimated number of active swells on Venus (Smrekar and Parmentier, 1996).

Accepting its interpreted origin above a mantle plume, it is useful to consider the rheological and petrological evolution of the region as the plume head first impinges on the crust and then spreads out and cools. By inference from the experimental data of Falloon and Green (1990), a plume saturated in  $\text{CO}_2$  ( $\sim 5 \text{ wt}\%$ ) and/or  $\text{H}_2\text{O}$  ( $2 \text{ wt}\%$ ) and rising from some depth, perhaps the core-mantle boundary, will start to melt at a depth of  $\sim 150 \text{ km}$ , forming a petrological asthenosphere with a melt fraction of  $1\sim 2\%$ . As noted earlier, even a volatile-poor plume will generate a low viscosity zone as it replaces the cold mantle lid. As it continues to rise and replace the mantle lid, a topographic swell will develop, raising the surface elevation to at least  $6052.6 \text{ km}$  radius by thinning of the lithospheric lid. Its thermal structure is approximated by the  $36 \text{ mW m}^{-2}$  ( $5 \text{ Ma}$ ) geotherm in Fig. 1, which shows that the plume undergoes major melting, producing picritic magmas from a depth of approximately  $100 \text{ km}$  and tholeiitic magmas nearer the surface. These volumetrically extensive melts may intrude at the base of the crust or extrude at the surface in

flood eruptions over a period of  $1$  to  $10 \text{ Ma}$  or more (Campbell, 2007). The rheological structure (Fig. 5) shows a thin weak crust overlying relatively low viscosity mantle. The plume head spreads out radially (Griffiths and Campbell, 1990) as it impinges on the weak crust, which readily stretches and thins in response.

Cooling of the plume head causes thermal subsidence of the swell and the cessation of flood eruptions, although the plume tail continues to supply discrete volcanic centres such as Sif and Gula Montes. Away from the tail itself the thermal structure of western Eistla Regio may now approximate the  $18 \text{ mW m}^{-2}$  ( $30 \text{ Ma}$ ) geotherm (Fig. 1), with an extensive asthenosphere still present. Minor melt in the asthenosphere may locally accumulate into diapirs capable of rising through the lithospheric lid and intruding the crust, perhaps forming coronae. The rheological structure now includes a stronger upper crust, a weak lower crustal detachment horizon and a thin, mainly brittle, upper mantle (Fig. 5). The detachment may be sufficiently weak to allow the continued lateral spread of the plume head but strong enough to transmit some degree of tractive stress to the crust, causing extension and rifting in the crust with little associated magmatism, as at Guor Linea.

As cooling continues, the melt fraction (away from the plume tail) will become more and more alkaline and carbonatitic (Hess and Head, 1990) such that the final volcanogenic features across the rise may be canali (Williams-Jones et al., 1998), although this is not suggest that all canali formed in this way. Assuming the extreme case of no background heat flux, continued cooling will eliminate the asthenosphere across all but the central core of the plume head after  $\sim 75 \text{ Ma}$  (the  $12 \text{ mW m}^{-2}$  geotherm in Fig. 1)



**Fig. 5.** Topographic profiles across (a and b) two terrestrial mid-ocean ridges and (c–f) four low-latitude great circle rift arcs (Jurdy and Stefanick, 1999). Each profile shows the mean elevation and range from at least seven across-axis profiles, fitted by a plate (Earth) or lid (Venus) half-space cooling curve, using the parameters given in each case. Additionally, the Venus profiles show the location of major and minor melting in the mantle; outside these regions there is no asthenosphere and the crust is strongly coupled to the mantle. However, note that the right-hand (south) side of (d) and left-hand (north) side of (f) overlap, in part explaining the poor fit in these regions.

although even a modest background heat flux will extend this timescale considerably. The rheological structure (Fig. 2) will then consist of a strong crust coupled to a stronger mantle lid with probably limited lateral displacement, although thermal subsidence will lead to the development of small contractional strains ( $\sim 2\%$ ), sufficient to induce wrinkle ridges. Continued thermal subsidence eventually leaves a basin up to  $\sim 1$  km deep (in the absence of any infilling by lava flows).

## 2.2. Convective lid rejuvenation

Geological evidence indicates a degree of tectonic organisation that may reflect mantle convection (Jurdy and Stefanick, 1999) and may also separate younger and older areas of the planet (Hauck et al., 1998; Price et al., 1996). Hot upwelling mantle underlying the major rift systems will generate both an asthenosphere and melt, and possibly induce ultra slow spreading (Stoddard and Jurdy, 2012). Differences in the temperature, rate and extent of the upwelling may be reflected in the axial topographic profile, the nature of the faulting, and the number of coronae associated with the rift system. However, if the crustal detachment layer is sufficiently weak to allow slip of the mantle lid in response to convection, advective heat loss by lid recycling can occur, in addition to magmatic advection and thermal conduction through the lid. The driving forces involved need not be prohibitive: the stress from subducting plates on Earth is sufficient to overcome resistive forces in the crust and mantle without any such weak detachment layer.

This situation can be approximated in the same manner as the global and plume hypsography but replacing the aerial rejuvenation rate with a linear rate and fitting to topographic profiles axial to the rift system. A series of such profiles taken across the four low-latitude extensional great circle arcs proposed by Jurdy and Stefanick (1999) is shown in Fig. 6; the North Atlantic Ridge and East Pacific Rise are shown for comparison. The greater topographic variability on Venus is obvious but not unexpected given that the cooling half-space mantle lid lies under more than 10 km of crust. The mean profiles are fitted with the cooling half-space model derived earlier from global hypsography by adjusting the crest elevation and rate of horizontal extension of the lid. Lid extension rates vary between 25 and 50  $\text{mm a}^{-1}$ , with a global average of 36  $\text{mm a}^{-1}$ , and are correlated with crest elevation: the higher the rate of extension, the lower the crest elevation, by up to 1000 m, but the average elevation is within 200 m of the crest elevation derived from the best fit to global hypsography. The

reason for this positive correlation may be that higher lid extension rates translate into broader zones of extension and thinning of the crust, offsetting any increase in magmatism and crustal thickening.

Applying the same logic as for the plumes, but noting that lid cooling is related to distance from the crest axis, regions of major and minor melting in the mantle can be mapped out geographically (Fig. 6). Differences in lid extension rates are immediately apparent; major melting extends for more than 1500 km from Parga Chasma but only 750 km from Devana Chasma. The extensive area of melt available under Parga and Hecate Chasmata may explain their greater concentration of coronae (Smrekar et al., 2010a), and within the Beta-Atla-Themis (“BAT”) region as a whole. Beyond the area of asthenosphere the crust and mantle are relatively strong and coupled. Wrinkle ridges overlie these areas (Fig. 6) and may correspond with convective downwelling where the thick, cold subcrustal lid is recycled.

## 3. Conclusions

The  $4.0 \pm 0.5 \text{ km}^2 \text{ a}^{-1}$  lid recycling rate inferred from global hypsometry implies a heat loss rate of  $32.8 \pm 3.6 \text{ TW}$ ,  $\sim 90\%$  of the scaled rate of global heat loss. The remaining  $\sim 10\%$  can plausibly be accounted for by crustal radioactivity and core heat loss, which leads to the principle conclusion that Venus is able to lose heat at proportionately the same rate as Earth and is therefore able to maintain a steady state without episodic global resurfacing.

This rate of heat loss depends only on the global lithospheric lid recycling rate; because of the weak crustal detachment horizon, this high recycling rate is consistent with relatively static surface processes. Plumes may certainly be responsible for a fraction of the lid rejuvenation but the number required ( $> 75$ ) is an order of magnitude greater than the number observed. It therefore appears that the globally extensive ( $\sim 55,000 \text{ km}$ ) network of rifts may be connected to lid rejuvenation by the creation and recycling of a subcrustal lid in a system analogous to plate creation and recycling on Earth.

While these processes are relatively disconnected from the surface by a rheological detachment horizon, the crust is displaced vertically by the cooling lid beneath it. Given the association of rifts and other features with these topographic rises, subcrustal lid rejuvenation very likely drives a range of more subtle geological process at the surface at the present day. New missions capable of change detection, such as interferometric SAR (Ghail et al., 2012;

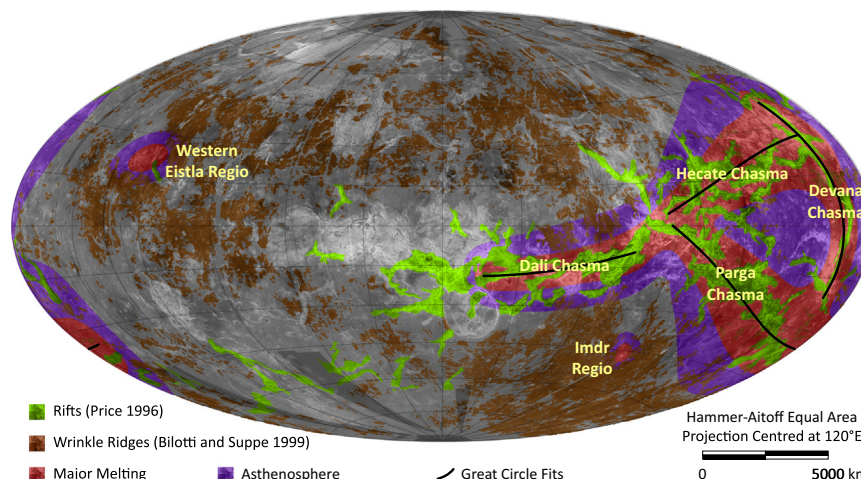


Fig. 6. Geographical extent of major melting and asthenosphere in relation to wrinkle ridges (Bilotti and Suppe, 1999) and rifts (Price et al., 1996) on Venus. Note that the wrinkle ridges avoid areas underlain by asthenosphere.

Meyer and Sandwell, 2012), offer the best opportunity to detect and measure these subtle effects and so finally unravel the behaviour of our enigmatic neighbour.

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